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Chapter 15

Geology of the Eastern Pontides

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ABSTRACT

The 500-km-long Eastern Pontide belt shows several common stratigraphic features resulting from a common Mesozoic–Tertiary tectonic history. There is a heterogeneous pre-Jurassic basement comprised of Devonian? high-grade metamorphic rocks, Lower Carboniferous granodiorites and dacites, Upper Carboniferous–Lower Permian shallow-marine to terrigeneous sedimentary rocks and an allochthonous Permo–Triassic metabasite-phyllite-marble unit. The Mesozoic sedimentary sequence starts with a widespread Liassic marine transgression coming from the south. The Lower and Middle Jurassic rocks of the Eastern Pontides make up a 2000-m-thick sequence of tuff, pyroclastic rock, lava, and interbedded clastic sedimentary rock; the volcanism is probably related to rifting leading to the opening of the Neotethyan Ocean in the south. The Upper Jurassic–Lower Cretaceous is characterized by carbonates, showing a transition from platform carbonate deposition in the north to pelagic carbonates and calciturbidites in the south; this indicates the development of a south-facing passive continental margin. During the Cenomanian, there was uplift and erosion throughout the Eastern Pontides. Rocks of this stage are not present, and in many localities the Senonian deposits lie unconformably over Jurassic carbonates and even over the Carboniferous granitic basement. This compressive event is associated with the northward emplacement of an ophiolitic melange over the passive continental margin of the Eastern Pontides. The obduction of the ophiolitic melange is probably caused by the partial subduction of the Eastern Pontides continental margin in a south-dipping intra-oceanic subduction zone. This was followed by the flip of the subduction polarity during the Cenomanian–Turonian, which led to the development of a Senonian volcanic arc in the outer Eastern Pontides above the northward-subducting Tethyan Ocean floor. The volcanic arc is represented by >2-km-thick succession of volcanic and volcaniclastic rocks and interbedded limestones and marls. There are also intrusive granodiorite plutons with isotopic ages of 95 to 65 m.y. The volcanism shows a general silica enrichment, with time, ranging from basalts and andesites to dacites. The

Senonian sequence in the inner Eastern Pontides is made up of a tuffaceous flyschoid series representing the fore-arc succession. The Eastern Black Sea Basin probably opened during the Maastrichtian through the rifting of the volcanic arc axis.

During the late Paleocene–early Eocene, there was north-vergent thrust imbrication of the inner Eastern Pontides with the development of a major foreland flysch basin in front of the northward moving thrust sheets. Folding and uplift occurred in the outer Eastern Pontides during this period. This compressive deformational event, the strongest Mesozoic–Tertiary orogenic phase in the Eastern Pontides, was probably caused by the collision between the Pontide arc and the Tauride microplate in the south.

Widespread calc-alkaline volcanism and shallow-marine sedimentation occurred throughout the Eastern Pontides during the middle Eocene. The middle Eocene rocks are essentially undeformed and lie unconformably over a folded and thrust-faulted basement. This major middle Eocene extensional event is probably related to an accelarated phase of opening of the Eastern Black Sea Basin. From the end of the middle Eocene onward, the Eastern Pontides stayed largely above sea level, with minor volcanism and terrigeneous sedimentation.

INTRODUCTION

The Eastern Pontides form a mountain chain 500 km long and 100 km wide along the southeastern coast of the Black Sea. Geologically, the Eastern Pontides are well known as one of the best preserved examples of a paleo-island arc (e.g., Akin, 1978; Şengör and Yılmaz, 1981; Akıncı, 1984), which was formed during the Senonian above the northward-subducting Tethyan Ocean floor. The object of this chapter is to document the evolution of the Eastern Pontides from its position along the southern continental margin of Laurasia, its transformation into an active continental margin, and its eventual deformation during the Early Tertiary continental collision.

Geographically, the Eastern Pontides is a loosely defined term used for the region skirting the Eastern Black Sea coast of Turkey. Its western boundary is taken arbitrarily either as the Yeşılırmak or Kızılırmak river near Samsun. Tectonically, it forms the eastern part of the Sakarya Zone of the Pontides (Figure 1) (Okay, 1989). It is bounded in the south by the Ankara-Erzincan Neotethyan suture and in the north by the East Black Sea Basin; in the east, it extends without a break into the Lesser Caucasus, where it can be correlated with the Achara-Trialet (Banks et al., this volume) and Artvin-Karabakh zones (Figure 1) (Khain, 1975). Its western geological boundary with the Central Pontides is stratigraphic and corresponds to a facies change in the Cretaceous sequence.

The Sakarya Zone, which includes the Eastern Pontides, is characterized by a general absence of in-situ Paleozoic sedimentary rocks, by the presence of Paleotethyan Permo–Triassic accretion-subduction complexes (the Karakaya Complex), and by a ubiquitous Liassic transgression (Okay, 1989; Okay et al., 1996a). In contrast, the Taurides in the south show a well-developed Paleozoic sedimentary succession and do not comprise Paleotethyan accretion-subduction complexes. In northeast Turkey, these two paleogeographic realms are separated by the Ankara-Erzincan suture zone, which is marked by large bodies of peridotite and ophiolitic melange (Figure 2).

The Eastern Pontides is commonly divided into an inner/southern and an outer/northern part (Akin, 1978; Gedikoğlu et al., 1979; Özsayar et al., 1981). The outer Eastern Pontides are dominated by Senonian and middle Eocene volcanic and volcaniclastic rocks, which hide much of the pre-Senonian geology (Figure 2). On the other hand, pre-Senonian rocks are widely exposed in the inner Eastern Pontides, which occupied a fore-arc position during the Senonian and underwent much more intensive deformation than did the outer Eastern Pontides during the Early Tertiary continental collison. The transitional boundary between these two parts follows approximately the Niksar-Gümüşhane-Ardanuç line (Figure 2).

THE PRE-JURASSIC BASEMENT OF THE EASTERN PONTIDES

The isolated exposures of the pre-Jurassic basement in the Eastern Pontides can be grouped into four types: a high-grade metamorphic complex of pre-Carboniferous age, an early Carboniferous granodiorite-dacite complex, an Upper Carboniferous–Lower Permian shallow-water to terrigeneous sedimentary sequence, and a Permo–Triassic metabasite-phyllite-marble unit.

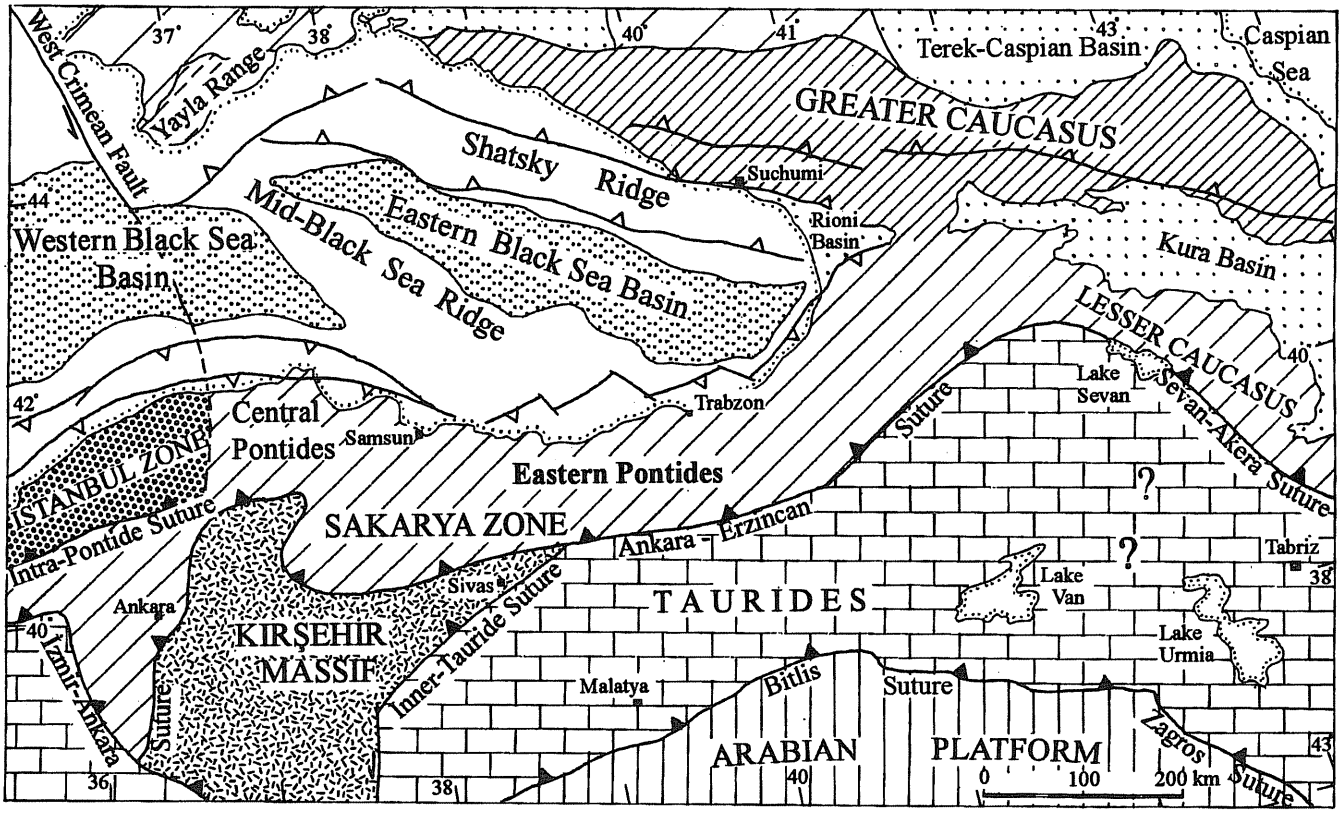

Figure 1. Tectonic map of the Eastern Black Sea region. Lines with black triangles indicate Neotethyan sutures with the original subduction polarity. Lines with open triangles are major post-Eocene thrusts.

High-Grade Metamorphic Complex—Pulur Massif

A heterogeneous crystalline basement of cordierite-sillimanite-garnet gneiss, microgneiss, migmatite, metaquartzite, banded amphibolite, diopside-plagioclase rock, and metadiorite is exposed in a major north-vergent Eocene thrust slice in the inner Eastern Pontides (Figures 2, 3). This metamorphic complex, called the Pulur Massif, outcrops in a 5–10 km wide and 60 km long northeast-trending belt and is overlain unconformably by the Liassic volcanosedimentary rocks (Ketin, 1951; Tanyolu, 1988; Keskin et al., 1989; Okay, 1996). The high-grade metamorphic rocks of the Pulur Massif are associated with medium-grained, banded metagranitic rocks that may represent synmetamorphic granites. In the Pulur Massif, there are also rare metadunite and metaharzburgite bands, a few meters thick, interlayered with amphibolites (Okay, 1996). The metamorphic rocks of the Pulur Massif are extensively mylonitized during the Alpide events and are intruded by Eocene dioritic and andesitic stocks, dykes, and sills. The gneisses and amphibolites, which form small outcrops under the Jurassic volcanosedimentary rocks south of Artvin (Figure 2) (Maden Tetkik ve Arama Enstitüsü, 1977; Tarhan, 1982), are probably part of the same high-grade metamorphic complex.

Lower Carboniferous Granodiorite-Dacite Complex—Gümüşhane and Köse Granodiorites

Granitic rocks occupy large areas south of Gümüşhane forming the Gümüşhane and Köse composite plutons (Figures 2, 3). They are generally hornblende-biotite granodiorites (Zankl, 1962a; Yılmaz, 1976) and intrude a metasedimentary basement of greenschist facies schists and phyllites of unknown age (Yılmaz, 1972, 1977). The Köse pluton has yielded a well-defined earliest Carboniferous (360 ±2 Ma) Rb/Sr isochron age (Bergougnan, 1987).

Probable extrusive equivalents of the Gümüşhane and Kelkit granodiorites occur south of the Olur region, where Yılmaz (1985) and Bozkus (1992a) describe a thick sequence of dacite, rhyodacite, and rhyolite lavas and tuffs unconformably overlain by the Jurassic volcaniclastic rocks (Figure 2).

Upper Carboniferous–Lower Permian? Sedimentary Sequence

A coherent and thick Upper Carboniferous sequence is exposed in two thrust sheets in the inner Eastern Pontides (Figure 3). The sequence shows a simple monoclinal structure and is unconformably overlain by Jurassic sandy limestones. The base of the sequence is not exposed, but probably lies over the high-grade metamorphic rocks of the Pulur Massif, which outcrops in the overlying thrust sheet. This Upper Paleozoic sequence, known since Ketin (1951), has been studied by several geologists with conflicting views regarding its stratigraphy and age (Ağar, 1977; Akdeniz, 1988; Keskin et al., 1989; Okay, 1993; Robinson et al., 1995). A map and cross section of the

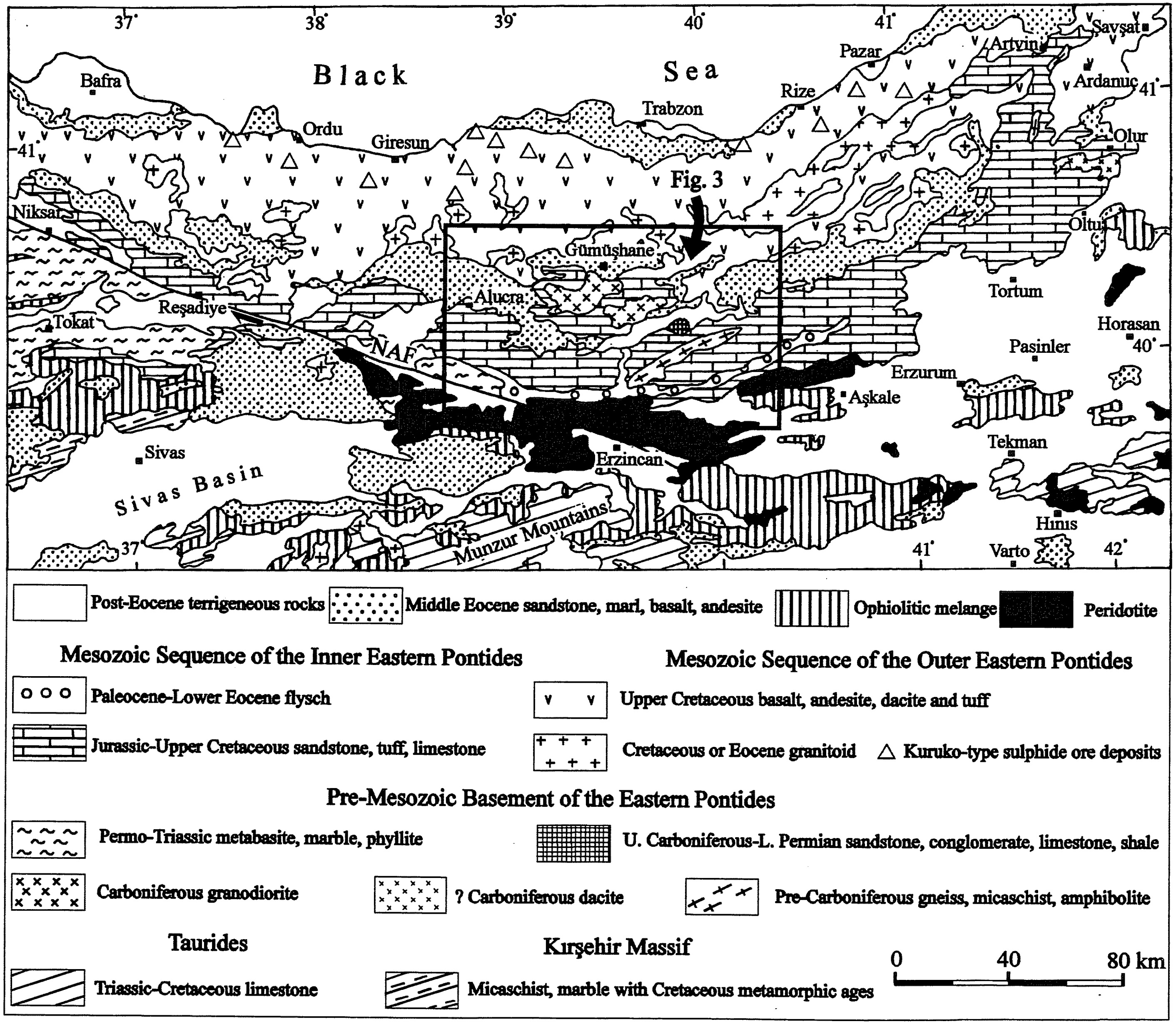

Figure 2. Simplified geological map of the Eastern Pontides. For clarity, no tectonic boundaries, except the North Anatolian fault (NAF), are shown. The locations of the sulfide deposits are from Pejatovic (1979) and Akıncı (1984). The region shown in Figure 3 is outlined.

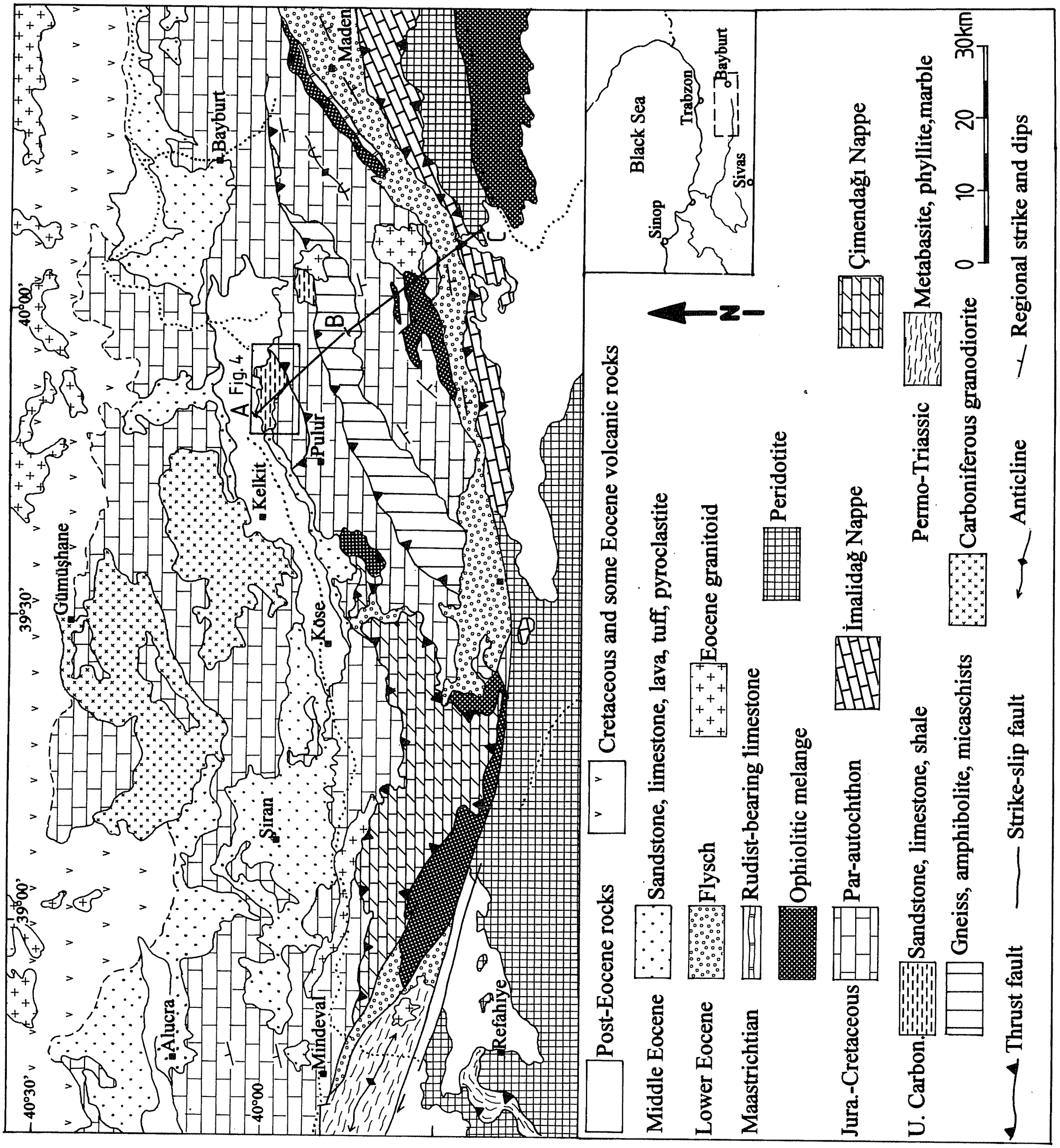

Figure 3. Geological map of the inner Eastern Pontides. For location, see Figure 2. Lines of sections refer to Figure 8.

region, based on recent mapping (Okay, 1993; Okay et al., in press), are shown in Figure 4. The Paleozoic sequence starts near Çatalçeşme with a ~1100-m-thick, heterogeneous series of sandstone, pebbly sandstone, quartzite, dark limestone, siltstone, and shale with rare thin coal seams; these lithologies are intimately intercalated on a few meters to a few tens of meters scale. The sandstones are medium to thickly bedded arkosic arenites with well-rounded pebbles of acidic magmatic rocks, possibly derived from the equivalents of the Gümüşhane granodiorite and Olur dacites. They are intercalated with medium to thickly bedded, dark-gray to black limestones locally rich in brachiopods, coral, gastropod, algae, and fusulinids. Thinly bedded, bioturbated siltstones with plant fragments and black-gray shale, with a few centimeter thick discontinuous coal seams, occur locally between the limestone beds. The fusulinid fauna in the limestones conclusively indicate a Late Carboniferous (Upper Kasimovian–Lower Gzelian) age. The diagnostic fusulinid species include *Eostafella* sp., *Ozawainella* cf. *angulata* (Colani) sp., *O. nikitovkensis* (Brazhnikova), *Pseudoendothyra* cf. *timanica* (Rauser), *Schubertella obscura* Lee & Chen, *S. pseudomagna* Putrya & Leont, *S. parvifusiformis* Lin, *Quasifusilina* ex gr. *longissima* (Moeller), *Q.* cf. *praecursor* Rauser, *Triticites gissaricus* Bensch, *T.* cf. *sinuosus* Rosovskaya, *T.* aff. *simplex* Schellwien, *T. petschoricus* Rauser, *T.* ex gr. *karlensis* Rosovskaya *Rugofusulina prisca ovoidea*, and *R. cf. praevia* Shlykova (Okay and Leven, 1996).

This heterogeneous Upper Carboniferous series is conformably overlain by ~1000-m-thick monotonous red terrigenous sandstones (Figure 4). The sandstones are thickly bedded to massive arkosic arenites and pebbly arenites. The pebbles in the sandstones are generally 2–5 cm large and consist of quartz and acidic magmatic rocks. No fossils have been found in the sandstones; however, considering that even the uppermost parts of the heterogeneous series are of Kasimovian–early Gzelian age, these conformably overlying terrigenous sandstones should be latest Carboniferous (Gzelian) and possibly earliest Permian in age.

Permo–Triassic Metabasite-Marble-Phyllite Unit—Ağvanis and Tokat Massifs

A thick sequence of closely intercalated metabasite, marble and phyllite forms a 35-km-long by ~8-km-wide metamorphic block, called Ağvanis Massif, immediately north of the Ankara-Erzincan suture north of Refahiye (Figures 2, 3) (Nebert, 1961; Okay, 1984). The metamorphic rocks form a large anticlinorium and are bounded in the north and south by strike-slip faults of the North Anatolian fault zone.

The Ağvanis Massif is dominated by metabasic rocks (60%–70% of the sequence), largely metatuffs and metalavas, locally with preserved pillow structures. The metabasic rocks are intimately intercalated with marble and phyllite horizons, each horizon usually <50 m thick. There are also very rare serpentinite lenses, which are a few tens of meters long (Okay, 1984). The total structural thickness is ~4.5 km. The whole sequence has undergone a greenschist facies metamorphism with the development of actinolite/barroisite-albite-epidote-chlorite-sphene paragenesis in the metabasic rocks.

The metamorphic rocks of the Ağvanis Massif are unconformably overlain by a Paleocene olistostrome formation. It is likely that the metamorphic rocks also form the basement to the Liassic sedimentary and volcanic rocks exposed north of the massif, although a post-Miocene strike-slip fault now constitutes the surface contact between these two formations (Figure 3).

No fossils are found in the Ağvanis Massif. Similar metabasite-marble-phyllite sequences occur in the Tokat Massif in the west (Blumenthal, 1950; Alp, 1972; Özcan et al., 1980; Tutkun and Inan, 1982; Aktimur et al., 1992) and are widespread in the western part of the Sakarya Zone forming part of the Karakaya Complex (the Nilüfer Unit of Okay et al., 1991, 1996a). In the Tokat Massif, metaclastic rocks locally comprise Permian and Triassic limestone blocks (Blumenthal, 1950; Öztürk, 1979; Özcan et al., 1980) indicating Permo–Triassic depositional and latest Triassic metamorphic ages.

The close intercalation of basic volcanic rock and tuff with limestone and shale, and absence of sheeted dyke complex or gabbro, indicates that the deposition of the Ağvanis Massif occurred in an arc-related basin, where there is close intermingling of volcanic and sedimentary rocks. This interpretation is in contrast to the ophiolite interpretation of the Ağvanis Massif by Şengör et al. (1980).

PRE-JURASSIC EVOLUTION OF THE EASTERN PONTIDES

The high-grade metamorphic basement of the Eastern Pontides is exposed in the Pulur Massif. This basement was intruded by granodiorites during the earliest Carboniferous, represented by the Gümüşhane and Köse plutons. Acidic volcanism as observed in the Olur region (Figure 5) was also associated with the granitic intrusions. The high-grade metamorphism could plausibly be associated with the thermal event that produced the calc-alkaline magmatism, and hence could be of Devonian or earliest Carboniferous age. The high-temperature–low-pressure metamorphic rocks and the calc-alkaline magmatic rocks probably formed in a Late Devonian–Early Carboniferous magmatic arc, although the polarity of subduction and the setting of the magmatic arc are obscure. In the Late Carboniferous, shallow-marine to fluviatile sedimentary rocks transgressed over the metamorphic-magmatic basement. Plant fossils in the Upper Carboniferous sequence show Euroamerican affinities (Şengör, 1990). Similar Upper Carboniferous–Lower Permian facies are widespread in the Greater Caucasus and represent molasse deposition at the end of the Hercynian orogeny (Khain, 1975; Adamia et al., 1982), and can be compared with the Rotliegende of Europe. Furthermore, early Hercynian sillimanite-cordierite-bearing gneisses, migmatites, and intrusive plutons (similar to those from the basement of the Eastern Pontides) are reported from the Greater

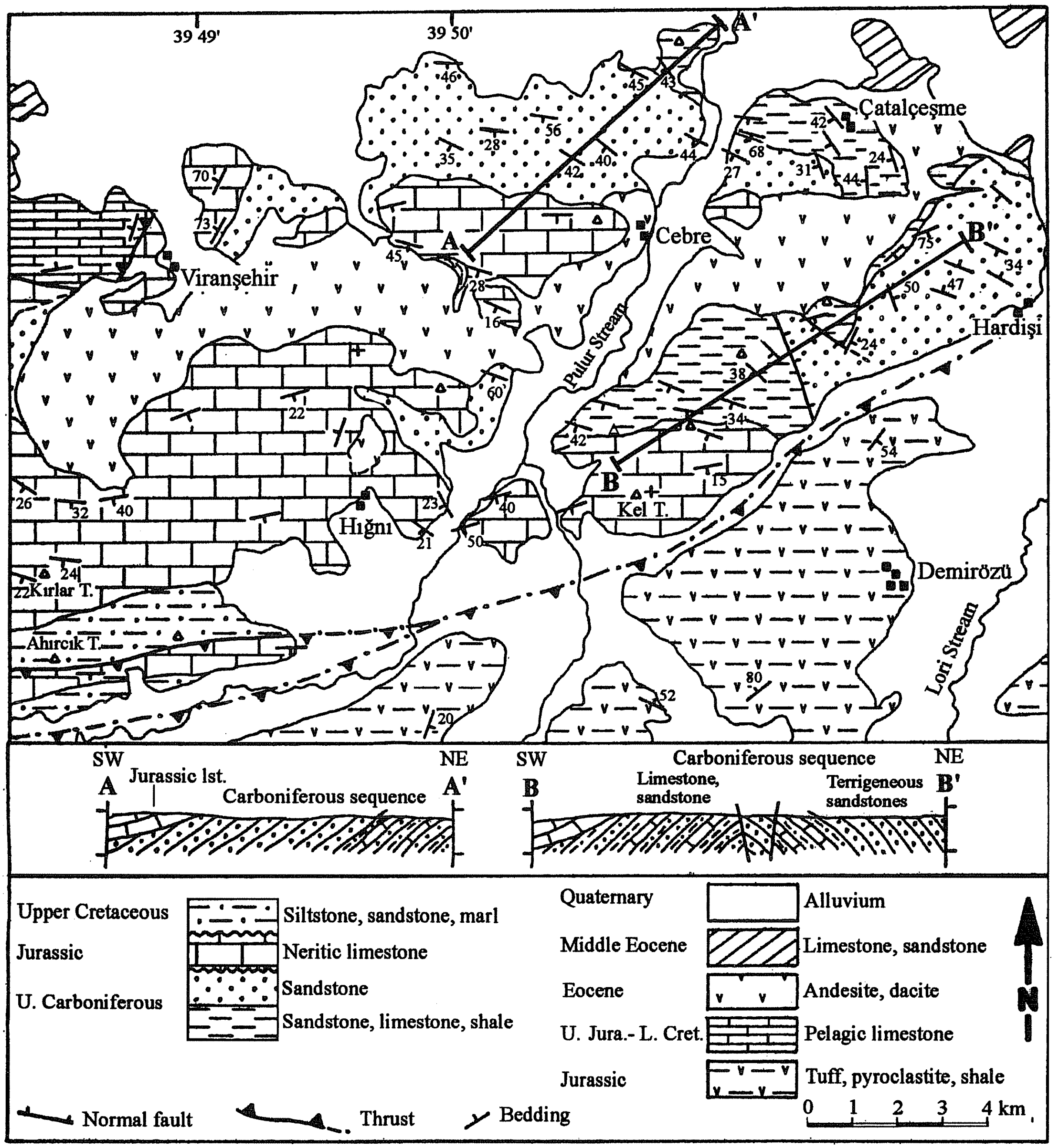

Figure 4. Geological map and cross sections of the Demirözü region showing the stratigraphic and tectonic setting of the Upper Carboniferous sequence. For location see Figure 3.

Caucasus (fore-range and main range zones) and from the Dzirula, Khrami, and Loki salients of the Transcaucasian Median Massif (Abesadze et al., 1982; Adamia et al., 1982, 1983). In contrast, no Hercynian metamorphism or magmatism is known in the Taurides, indicating that during the Late Paleozoic, the Eastern Pontides were possibly part of the Caucasian realm located along the southern margin of Laurasia (Adamia et al., 1982; Robinson et al., 1995). A contrasting view is given by Şengör (1990), who places the Eastern Pontides and the Dzirula, Khrami, and Loki salients at the northern margin of Gondwana. This is largely based on a continuous Middle Devonian to Middle Triassic clastic series (the Dizi Series, Adamia et al., 1982), with no record of a late Paleozoic deformation, located in the Greater Caucasus north of the Transcaucasian Median Massif. However, the Dizi Series could be completely allochthonous and derived from the south during the Late Triassic.

The Permian and Triassic events are poorly recorded in the Eastern Pontides. The various reports of the presence of Permo–Triassic ophiolites in the

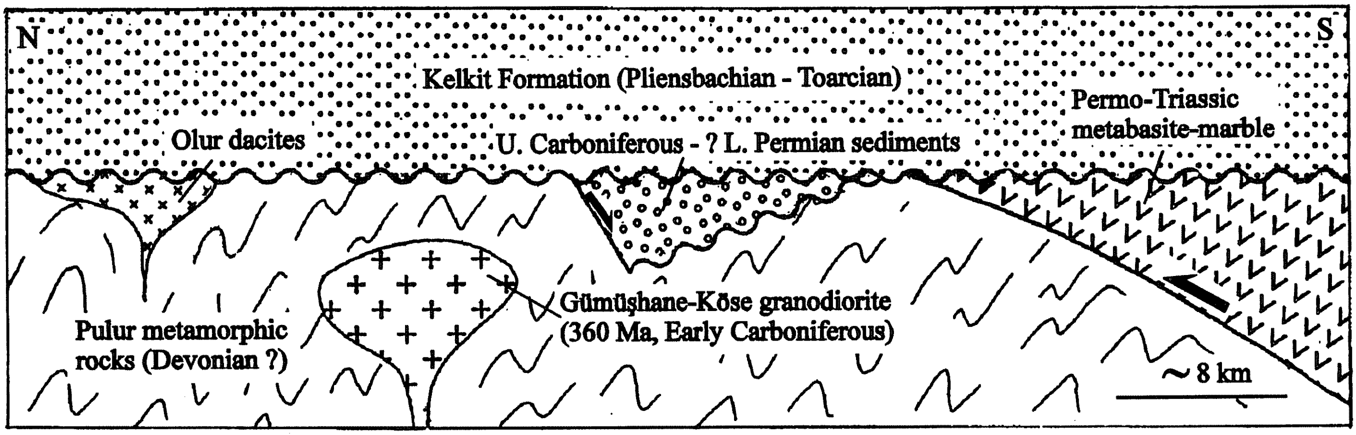

Figure 5. Speculative sketch illustrating possible relations among the pre-Jurassic basement units of the Eastern Pontides.

Eastern Pontides (Seymen, 1975; Bektas et al., 1984; Koçyiğit, 1990) are not verified. In the Biga Peninsula in the western part of the Sakarya Zone, a Permo-Triassic metabasite-phyllite-marble unit, similar to the Ağvanis Massif, tectonically overlies a high-grade metamorphic basement with mid-Carboniferous (308 ±16 Ma) zircon Pb ages (Okay et al., 1996a). This relationship is interpreted as the tectonic emplacement of a Permo–Triassic ensimatic fore-arc sequence over a continental margin during the latest Triassic final closure of the Paleotethys (Okay et al., 1996a). A similar tectonic relationship and interpretation can be envisaged for the Eastern Pontides. The location of the Paleotethyan fore-arc sequences along the inner margin of the Eastern Pontides and their apparent absence in the north suggest that the Paleotethys was located to the south of the Eastern Pontides. However, using the available data from the Eastern Pontides, little else can be said about the initiation, subduction polarity, and destruction of the Paleotethys.

JURASSIC–EARLY CRETACEOUS—EVOLUTION OF A SOUTH-FACING PASSIVE CONTINENTAL MARGIN

Lower–Middle Jurassic Rift Sequence—The Kelkit Formation

Triassic sedimentary rocks are virtually absent in the Eastern Pontides, and the Mesozoic deposits begin with a major Liassic transgression coming from the south (Akin, 1978). The Jurassic in the Eastern Pontides is characteristically represented by a predominantly volcaniclastic formation (the Kelkit Formation), which rests unconformably over a heterogeneous basement. It consists of basaltic and andesitic lithic tuffs, volcanogenic sandstone, shale, basaltic and andesitic lavas, and conglomerate. There are also thin, discontinuous coal and ammonitico rosso horizons within the sequence (Figure 6). Although the sequence is dominated by volcanogenic sandstones, there are also true pyroclastic rocks and basaltic and andesitic lava flows (Bergougnan, 1976, 1987). The clinopyroxene composition from the pyroclastic rocks and lavas suggests a tholeiitic parent magma (Bergougnan, 1987). The Kelkit Formation has a very wide distribution in the Eastern Pontides and extends from the Destek-Reşadiye area (Seymen, 1975; Öztürk, 1979; Aktimur et al., 1992) 500 km eastward to the Olur and Yusufeli regions (Maden Tetkik ve Arama Enstitüsü, 1977; Yılmaz, 1985; Bozkus, 1992a). Its northward extension is largely concealed by the Cretaceous and Eocene deposits north of Gümüşhane; however, small inliers in this region (Tasli, 1984; Bektas et al., 1987; Korkmaz, 1993) show that a similar facies, albeit richer in volcanic rocks, characterizes the Lower–Middle Jurassic throughout the Eastern Pontides (Figure 7A).

The Kelkit Formation is ~1500 to 2000 m thick (Ketin, 1951; Nebert, 1961; Pelin, 1977; Bergougnan, 1987; Gürsoy, 1989; Okay, 1993) and shows vertical and lateral facies changes. Three main facies are recognized: (1) a turbiditic volcaniclastic sandstone-shale unit showing graded bedded and flow and slump structures (Okay, 1984), (2) a widespread lithic tuff-volcaniclastic sandstone-conglomerate unit including several coal horizons representing a paralic environment (Pelin, 1977), and (3) the ammonitico rosso facies showing condensed deposition on top of seamounts. The Kelkit Formation becomes generally finer grained and deeper marine toward the south (Yılmaz, 1985; Okay, 1993). For example, south of Pulur, the basal Kelkit Formation overlies the high-grade metamorphic rocks of the Pulur Massif containing a 20-m-thick horizon of pebbly arenites with thin coal seams and grades up into a 1500-m-thick sequence of interbedded fine-grained dark tuff and shale; this includes Jurassic limestone blocks several hundred meters large (Okay, 1993). The deposits south of Pulur do not include the volcaniclastic sandstones and coarse pyroclastics common in the Kelkit Formation farther north.

The age of the Kelkit Formation is largely based on ammonites from the ammonitico rosso horizons, which show the presence of all the Liassic stages from Early Pliensbachian to Toarcian (Otkun, 1942;

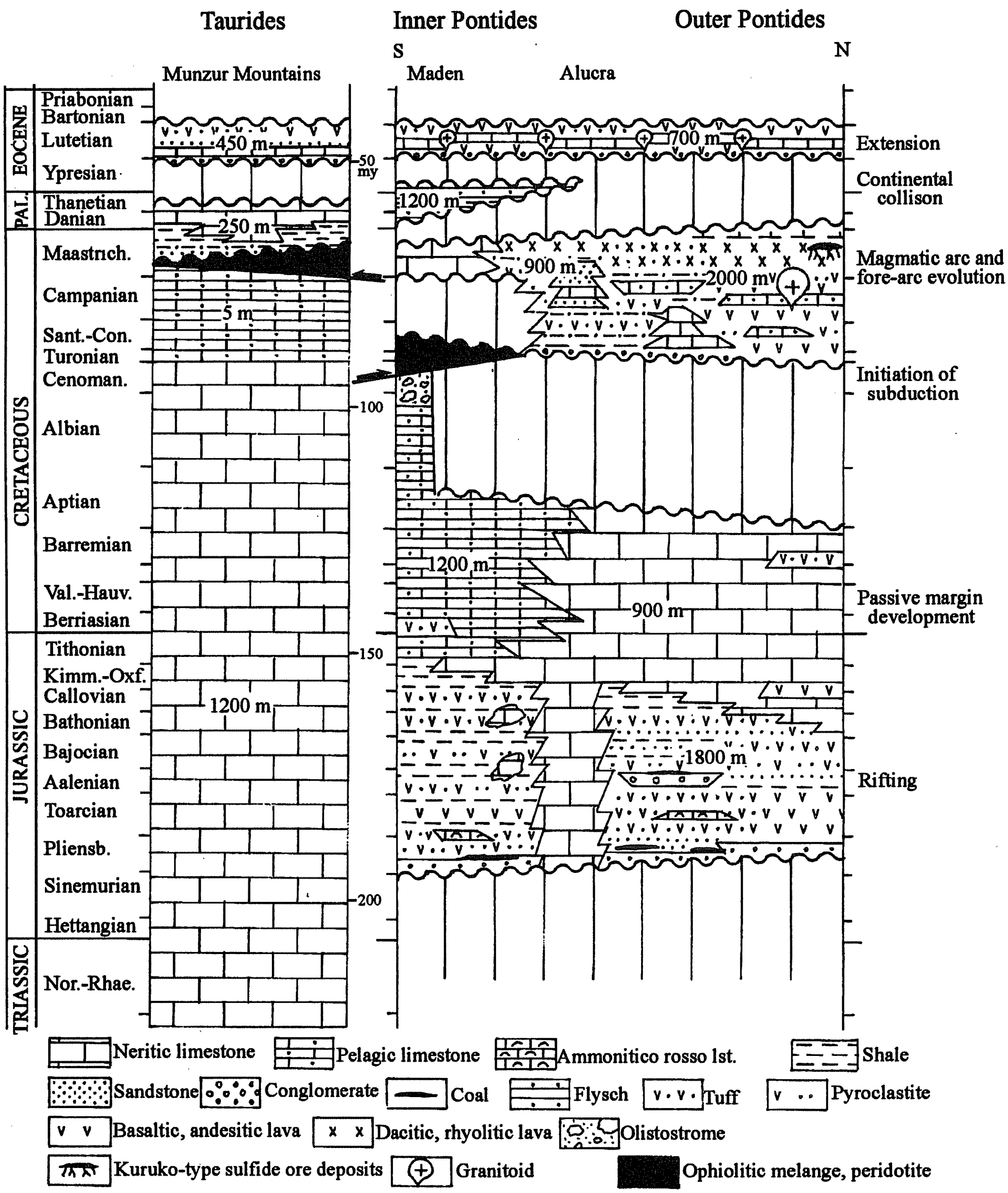

Figure 6. Stratigraphic columns for the Eastern Pontides and the Munzur Mountains of the Taurides. The stratigraphy of the Taurides in the Munzur Mountains is based on Özgül and Tursucu (1984), for the numerous stratigraphic data on the Eastern Pontides, see the text.

Stchepinsky, 1945; Ketin, 1951; Alp, 1972; Yılmaz, 1972; Bassoullet et al., 1975; Özer, 1984; Bergougnan, 1987). In addition, palynology of separate coal horizons gives Liassic (Agrali et al., 1966; Pelin, 1977) and Dogger ages (Agrali et al., 1965). The presence of Dogger age is further suggested by rare macrofossils (Wedding, 1963) and by the recent determination of dinoflagellate and palynomorph assemblages (Robinson et al., 1995). Thus, the age span of the Kelkit Formation is from early Pliensbachian up to at least the end of the Bathonian.

A different Jurassic facies occurs in the Pulur region in the relative authochthon at the base of the thrust sheets. Here, Liassic shallow-water sandy limestones

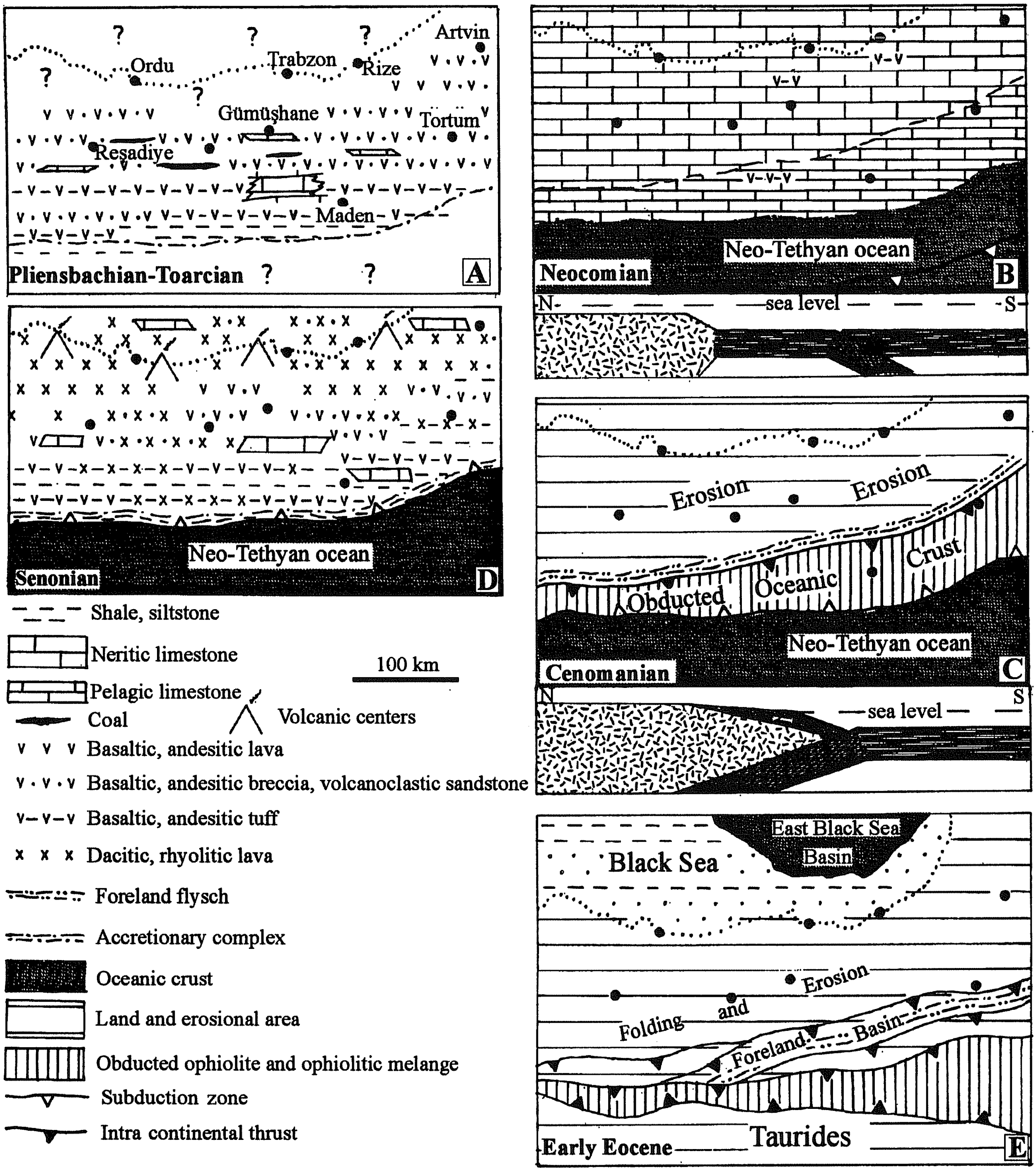

Figure 7. Mesozoic paleogeographic and tectonic maps of the Eastern Pontides. The scale is only for the continental crust, assuming a conservative 50% crustal shortening in the inner Eastern Pontides. (A) Pliensbachian–Toarcian, ≤2000-m-thick volcaniclastic rocks with basic lavas characterize much of the Eastern Pontides. Associated with the basic volcaniclastic rocks are thin coal seams and ammonitico rosso-type limestones. In contrast, neritic limestone deposition occurs in a major horst south of Gümüşhane. The paralic to shallow-marine environment in the north changes to deeper marine conditions in the south. Possible rifting leading to the development of the Neotethyan Ocean in the south. (B) Neocomian, extensive carbonate deposition over the whole of the Eastern Pontides. The carbonates range from neritic to pelagic toward the south and suggest the development of a south-facing passive continental margin. Minor tuff deposition from a possible intra-oceanic subduction zone to the south. (C) Cenomanian, obduction of oceanic crust and accretionary complex is caused by the partial subduction of the Eastern Pontides continental margin in an intra-oceanic subduction zone. This is followed by the flip of the subduction polarity. Uplift and erosion characterize the outer Eastern Pontides. (D) Senonian, a volcanic arc forms above the northward-subducting Tethyan Ocean floor. The Eastern Black Sea Basin develops during the Maastrichtian through the splitting of the arc axis. (E) Early Eocene, continental collision with the Taurides results in the thrust imbrication of the inner Eastern Pontides continental margin. A major foreland flysch basin develops in front of the north-vergent thrust sheets. Strong uplift and folding in the Eastern Pontides.

(Akdeniz, 1988; Okay, 1993; Robinson et al., 1995) lie unconformably over the Carboniferous sedimentary sequence (Figure 4). They grade up through an oolitic limestone horizon to thickly bedded, cherty micrites. The whole sequence is ~750 m thick and ranges in age from Liassic to Berriasian–Valanginian. It is unconformably overlain by Senonian deep-water sediments (Akdeniz, 1988; Okay et al., in press). Lower and Middle Jurassic rock in the overlying thrust sheets is developed in the volcaniclastic facies, suggesting that the Pulur Jurassic rocks are deposited on top of a horst and thus avoided the volcaniclastic influx (Figure 6).

The similarity in the Mesozoic stratigraphy between the Eastern Pontides and the Caucasus, as shown by the general absence of Triassic deposits and similar volcanic and coal-bearing Jurassic facies (Khain, 1975; Adamia et al., 1982), indicates, contrary to some suggestions (Bektas et al., 1984), that the Eastern Pontides were located along the southern margin of Laurasia before the opening of the Eastern Black Sea Basin in the Maastrichtian.

Görür et al. (1983) interpreted the Kelkit Formation of the Eastern Pontides as a rift facies related to the opening of the Neotethyan Ocean in the south. The lateral facies and thickness changes in the Kelkit Formation support this interpretation. The southward deepening inferred from the Kelkit Formation is probably related to the opening of the Neotethyan Ankara-Erzincan Ocean in the south (Görür et al., 1983). However, a discrepancy arises in this interpretation due to the dissimilarity of the pre-Jurassic stratigraphy between the Eastern Pontides and the Taurides to the south. In the Munzur and Keban units of the Taurides there is a well-developed Triassic and Permian carbonate succession not seen anywhere in the Pontides (Figure 6) (Özgül and Tursucu, 1984). Furthermore, during the Liassic, there were faunal differences between the two sides of the suture (Bassoullet et al., 1975; Enay, 1976); Liassic ammonites from the Pontides are similar to those from the southern Laurasian margin, whereas Liassic fauna from the Munzur Mountains of the Taurides resemble those from the southern margin of the Tethys. It is quite possible that there was major post-Liassic strike-slip movement along the southern margin of Laurasia, so that the Taurides were not contiguous with the Eastern Pontides during the pre-Liassic. One possibility is that the Sanandaj-Sirjan Zone in Southwest Iran, which shows evidence for Paleozoic and Triassic orogenic activity and has an uppermost Triassic to Middle Jurassic unconformable cover of thick coal-bearing clastic rocks, similar to the Kelkit Formation of the Eastern Pontides (Stöcklin, 1968; Şengör, 1990), could represent the continental fragment that rifted off from the Eastern Pontides during the Jurassic.

Upper Jurassic–Lower Cretaceous Carbonate Platform

The Upper Jurassic–Lower Cretaceous carbonates, which conformably overlie the Kelkit Formation, outcrop throughout the Sakarya Zone. In the Eastern Pontides, the carbonate deposition begins slightly earlier in the north (Oxfordian–Tithonian), where it is represented by a ~900-m-thick neritic limestone and dolomite (Pelin, 1977, 1981; Gürsoy, 1989; Yılmaz, 1992; Robinson et al., 1995), than in the south (Tithonian–Berriasian), where there is a >1000-m-thick pelagic radiolarian biomicrite and calciturbidite sequence (Ketin, 1951; Bursuk, 1975, 1981; Tutkun and Inan, 1982; Okay et al., in press). Minor and local basic volcanism during the Late Jurassic–Early Cretaceous is evidenced by thin tuff beds in the limestone sequence north of Gümüşhane (Zankl, 1962b) and south of Pulur (Okay et al., in press). In the Late Jurassic, there is a clear facies differentiation between the outer and inner Eastern Pontides. The boundary between these two realms follows approximately the Niksar-Demirözü-Bayburt-Tortum line (Figure 2). The picture is that of a carbonate platform passing southeast into a carbonate ramp representing the south-facing continental margin of Laurasia (Figure 7B) (Görür, 1988). This phase in the Eastern Pontides represents the development of a south-facing passive carbonate continental margin.

The Upper Jurassic–Lower Cretaceous carbonates are bounded above by a major Senonian unconformity associated in places (for example, in parts of the Alucra region) (Pelin, 1977) with the erosion of the whole carbonate sequence. The carbonates are karstified with locally developed paleovalleys removing the complete carbonate sequences. The uppermost age of the preserved carbonate sequence is late Barremian in the north (Pelin, 1977; Bergougnan, 1987) and Aptian in the south (Bursuk, 1981; Okay et al., in press). Albian ages from the pelagic carbonates are only reported from the Maden region near the suture zone (Elmas, 1994; Robinson et al., 1995).

CENOMANIAN–TURONIAN—REGIONAL UPLIFT RELATED TO THE EMPLACEMENT OF OPHIOLITIC MELANGE

The Cenomanian represents a major episode of uplift and erosion throughout the Eastern Pontides (Ketin, 1977; Pelin et al., 1982), which is associated with the northward emplacement of an ophiolitic melange over the inner Eastern Pontides (Figure 7C). Cenomanian deposits are absent throughout the Eastern Pontides, and in most localities Campanian limestones lie unconformably over Jurassic sediments (Pelin, 1977; Yılmaz, 1985; Bergougnan, 1987; Gürsoy, 1989; Robinson et al., 1995); in the Gümüşhane region, Senonian rocks rest directly on the Carboniferous granites (Yılmaz, 1972). In this respect, the Eastern Pontides differ from the central Pontides, where the Aptian–Albian is characterized by shallow to deep marine clastic rocks interpreted as synrift deposits associated with the opening of the Western Black Sea Basin (Görür, 1988; Robinson et al., 1995).

The ophiolitic melange in the inner Eastern Pontides forms a 200-km-long thrust sheet from north of Erzincan to north of Tortum (Figure 3) and rests tectonically over the Lower Cretaceous pelagic carbonates (Bilgin, 1984; Bergougnan, 1987; Inan, 1988; Okay et al., in press). Due to the Early Tertiary erosion, the ophiolitic melange is preserved as small klippen over the Neocomian pelagic carbonates, which contrasts with the large peridotite and ophiolitic melange masses in the suture zone and in the Taurides (Figure 2). Locally, an olistostromal foreland flysch sequence occurs between the melange and the carbonates. The youngest age recorded in the carbonates underneath the melange is Late Albian in the Maden region (Elmas, 1994). The ophiolitic melange is composed of, in order of abundance, spilitized basalt, radiolarian chert, pelagic and neritic limestone, sandstone, shale, siltstone, and serpentinite blocks. It includes lithologies from both the oceanic crust and the passive continental margin of the Eastern Pontides. The various blocks in the melange are juxtaposed along fault contacts without any discernible matrix. Pelagic limestone blocks in the melange, derived from the Eastern Pontides continental margin, range in age up to Aptian (Okay et al., in press), and the melange is unconformably overlain by Maastrichtian rudist-bearing neritic limestones in the Maden region (Figure 3) (Ketin, 1951; Fenerci, 1994). Farther south in the Askale region (Bozkus, 1992b) and northwest of Erzurum (Inan, 1988), Maastrichtian pelagic limestones and calciturbidites lie unconformably over the melange. This constrains the age of the melange formation and its emplacement onto the continental margin to the Cenomanian–Campanian (Figure 6). Considering that the Late Cretaceous transgression in the Eastern Pontides starts in the Turonian–Coniacian, the emplacement age of the melange is further constrained to Cenomanian–Turonian (Figure 6). The ophiolitic melange thrust sheet apparently did not reach farther north than the line connecting Siran-Kelkit-Bayburt (Figure 3), although the compression was felt throughout the Eastern Pontides as a period of major uplift and erosion (Figure 7C). In the west, in the region of the Tokat Massif, Aktimur et al. (1992) also describe the emplacement of an ophiolitic melange during this time. In the Lesser Caucasus, ophiolite and ophiolitic melange were emplaced northward over the Sevan-Akera Zone also during the Cenomanian–Coniacian (Knipper, 1980).

Thus, oceanic crust and/or oceanic accretionary complex were obducted northward over the 1000-km-long Eastern Pontides–Lesser Caucasus continental margin during the Cenomanian–Turonian. This obduction may be related to the partial subduction of the Eastern Pontides passive continental margin in a south-dipping juvenile intra-oceanic subduction zone (Figure 7C). This was probably followed by the flip of the subduction polarity from a south-to-north–dipping subduction, which explains the close temporal relation between the obduction of the ophiolitic melange and the start of the northward-dipping subduction as deduced from the Turonian subduction-related volcanic activity in the outer Eastern Pontides. An alternative mechanism for the emplacement of the melange is the back-thrusting of the accretionary complex (Silver and Reed, 1988) above a northward-dipping subduction zone, which has been suggested for the Eastern Pontides by Elmas (1995). However, this mechanism is unlikely in the present case, as there is little evidence for subduction activity in the Eastern Pontides prior to the emplacement of the ophiolitic melange.

SENONIAN—BUILDUP OF THE PONTIDE MAGMATIC ARC

In the outer Eastern Pontides, which is dominated by Senonian and Eocene volcanic and sedimentary rocks, there is no clear break between the Lower and Upper Cretaceous. In the inner Eastern Pontides, the Senonian transgression youngs southward from Late Turonian–Coniacian in Alucra (Schiftah, 1967; Pelin, 1977; Robinson et al., 1995), to Coniacian–Santonian in the Kelkit (Gürsoy, 1989) and finally to Early Maastrichtian in the Maden region (Ketin, 1951; Fenerci, 1994), suggesting that the transgression came from the north.

A >2-km-thick Cretaceous volcano-sedimentary sequence, representing the upper surficial deposits of a magmatic arc (Dewey et al., 1973; Boccaletti et al., 1974; Akin, 1978), outcrops throughout the outer Eastern Pontides (Figure 6). Although the outer Eastern Pontides region is heavily vegetated, it has been intensely studied due to the widespread polymetallic sulfide mineralization in the volcanic rocks. The structure of the Mesozoic series of the outer Eastern Pontides is characterized by block-faulting and gentle seaward dips. To date, no major folds or thrusts have been mapped. The steep-dipping faults defining complex horsts and grabens are generally conjugate and follow NE and NW directions (Schultze-Westrum, 1962; Zankl, 1962a; Kronberg, 1970; Buser and Cvetic, 1973; Akin, 1978; Gedikoğlu, 1978).

The Cretaceous volcanic cycle in the outer Eastern Pontides starts with basaltic and andesitic lavas and passes up to dacitic and rhyolitic lavas, breccias, and tuffs with minor limestone intercalations (Schultze-Westrum, 1962; Zankl, 1962a; Eğin et al., 1979; Özsayar et al., 1982; Ercan and Gedik, 1983; Akıncı, 1984; Gedik et al., 1992; Korkmaz, 1993). The cycle ends with intercalations of mudstone, tuff, radiolarian chert, marl, and limestone from which a late Maastrichtian to Danian pelagic fauna was described (Hirst and Eğin, 1979; Korkmaz, 1993). Although there is some evidence for local basic volcanism during the Early Cretaceous (Zankl, 1962b; Okay, 1993), in most regions the volcanic and volcaniclastic rocks unconformably overlie neritic limestones of Kimmeridgian to Neocomian age and are thus younger than Cenomanian (Pelin, 1977; Terlemez and Yılmaz, 1980; Tasli, 1984; Korkmaz, 1993). The basal parts of the volcanic sequence are precisely dated by Taner and Zaninetti (1978), who describe a pelagic Middle Turonian fauna in the limestones intercalated with the basic volcanic rocks stratigraphically 1500 m below the dacites. Farther up in the sequence, all the stages of the Senonian are recognized in the intercalated limestones (Özsayar, 1971), such that the volcanism was continuous up to the end of the Maastrichtian and in places passes into the Early Paleocene (Korkmaz and Gedik, 1988; Korkmaz, 1993).

Thus, contrary to many views, which initiate the subduction under the Eastern Pontides during the Jurassic and earlier (Adamia et al., 1977; Kazmin et al., 1986), we follow Şengör and Yılmaz (1981) and Görür (1988) in placing the beginning of subduction into the Cenomanian–Turonian.

The thickness of the Cretaceous volcanic cycle is >2000 m; the volcanism was wholly submarine and, judging from the intercalated *Globotruncana*-bearing pelagic and rudist-bearing neritic limestone lenses, occurred over an uneven submarine topography. Possible volcanic centers were separated by deep marine basins (Robinson et al., 1995). Numerous massive and stockwork-type polymetallic (Fe, Cu, Pb, and Zn) sulfide ore deposits occur within the dacitic-rhyolitic lava, breccia, and tuff in the Maastrichtian part of the volcanic sequence (Figure 2). These deposits are intensely studied (Kraeff, 1963; Koprivica, 1976; Altun, 1977; Eğin et al., 1979; Hirst and Eğin, 1979; Pejatovic, 1979; Çağatay and Boyle, 1980; Akıncı, 1984). They are remarkably similar to the Japanese Miocene Kuruko sulfide ores (Akin, 1978), formed during arc volcanism around the volcanic centers (Mitchell and Garson, 1976). The sulfide ores are usually overlain by exhalative-sedimentary manganese deposits thought to be produced by volcanic hot springs discharging directly onto the sea floor (Hirst and Eğin, 1979). The distribution of the Kuruko-type sulfide ores in the Eastern Pontides (Vujanovic, 1974; Pejatovic, 1979) gives an indication of the volcanic centers during the Senonian, and shows that the Maastrichtian volcanic axis was located close to the present-day Black Sea margin (Figure 2). The sulfide mineralization was partly controlled by the NE-SW– and NW-SE–trending conjugate faults (Schultze-Westrum, 1962; Koprivica, 1976; Hirst and Eğin, 1979).

Cretaceous volcanic rocks are generally subalkaline and give typical island-arc geochemical signatures (Peccerillo and Taylor, 1975; Akin, 1978; Gedikoğlu, 1978; Eğin et al., 1979; Eğin and Hirst, 1979; Manetti et al., 1983; Akıncı, 1984; Köprübasi, 1993). Only toward the end of the volcanic cycle was there a minor alkaline, shoshonitic basic volcanism producing leucite and nepheline-bearing basanites and tephrites (Gümüs, 1978; Korkmaz et al., 1993).

Associated with the Cretaceous volcanism was the intrusion of a large number of granitoids (Figure 2). Their isotopic ages range from ~95 to 65 m.y. (Taner, 1977; Gedikoğlu, 1978; Moore et al., 1980). Although they show a wide modal scatter, they are dominantly hornblende-biotite, granodiorite, and quartz-diorite; geochemically, the granitoids show a typical calc-alkaline trend (Çoğulu, 1975; Taner, 1977; Gedikoğlu, 1978; Moore et al., 1980). The granitoids generally intrude the lower basic volcanic rocks and are locally overlain by dacitic lavas.

During the Senonian, the inner Eastern Pontides were in a fore-arc position. In some places, such as north of Kelkit, there was a ridge between the arc and fore-arc, characterized by the deposition of shallow-marine sandstones and rudist-bearing limestones (Schiftah, 1967; Gürsoy, 1989). However, in most regions, the transition was gradual, with a southward decrease in the amount of lava flows at the expense of volcaniclastic rocks. The Senonian, in the inner Eastern Pontides is characteristically represented by a 500- to 900-m-thick tuffaceous flyschoid sequence with pelagic limestone intercalations (Seymen, 1975; Pelin, 1977; Yılmaz, 1985; Gürsoy, 1989; Okay et al., in press).

The Senonian volcanic arc of the Eastern Pontides was an extensional arc, as shown by the submarine nature of the volcanism. The marginal basins behind extensional arcs usually develop by the splitting of the volcanic arc axis (Karig, 1971). The distribution of the Kuruko-type ore deposits in the Eastern Pontides shows that during the Maastrichtian, the volcanic arc axis was located close to the present-day Black Sea coast. Thus, the Eastern Black Sea Basin probably started to open during the Maastrichtian by the splitting of this arc axis. An earlier opening of the Eastern Black Sea Basin would have resulted in a major volcanic and volcaniclastic apron to the north of the present-day volcanic axis, which is not observed in the very narrow Black Sea shelf to the north of the Eastern Pontides.

PALEOCENE–EARLY EOCENE THRUST IMBRICATION—CONTINENTAL COLLISION

During the Paleocene–early Eocene, there was major shortening along the inner Eastern Pontides, and the continental margin was telescoped into a series of stacked north-vergent thrust slices. The thrusting did not reach farther north than the Bayburt-Kelkit-Mindeval line (Figure 3). In the Alucra region, which is located between the inner and outer Pontides, there was upright folding and a major break in sedimentation during the Paleocene–early Eocene. Lutetian limestones, marls, and sandstones in this region lie with the angular unconformity on folded Jurassic and Cretaceous sediments (Nebert, 1961; Pelin, 1977) and even overlie the metamorphic basement of the Ağvanis Massif (Okay, 1984). In the outer Eastern Pontides, sedimentary or volcanic rocks of Late Paleocene–early Eocene age are also not recognized, and Lutetian volcano-sedimentary rocks lie unconformably over the Senonian rocks (Gedikoğlu et al., 1982; Korkmaz and Gedik, 1988; Korkmaz, 1993), suggesting that the entire Eastern Pontides was above sea level during the Paleocene–Early Eocene.

South of the town of Kelkit, three north-vergent thrust slices are recognized (Figure 3) (Okay et al., in press). The basement rocks were involved in thrusting, and the high-grade Pulur metamorphic rocks were emplaced over the Jurassic Kelkit Formation (Figure 8). The topmost thrust slice in the structural sequence is rootless and is named the Çimendaği nappe in the west and the İmalidağ nappe in the east (Figure 3) (Bergougnan, 1975, 1976). The Çimendaği nappe consists mainly of Oxfordian to Berriasian shallow-water carbonates overlain unconformably by Santonian to Campanian red pelagic limestones, while the İmalidağ nappe is made up of Tithonian flysch overlain by Neocomian radiolarian biomicrites (Bergougnan, 1976,

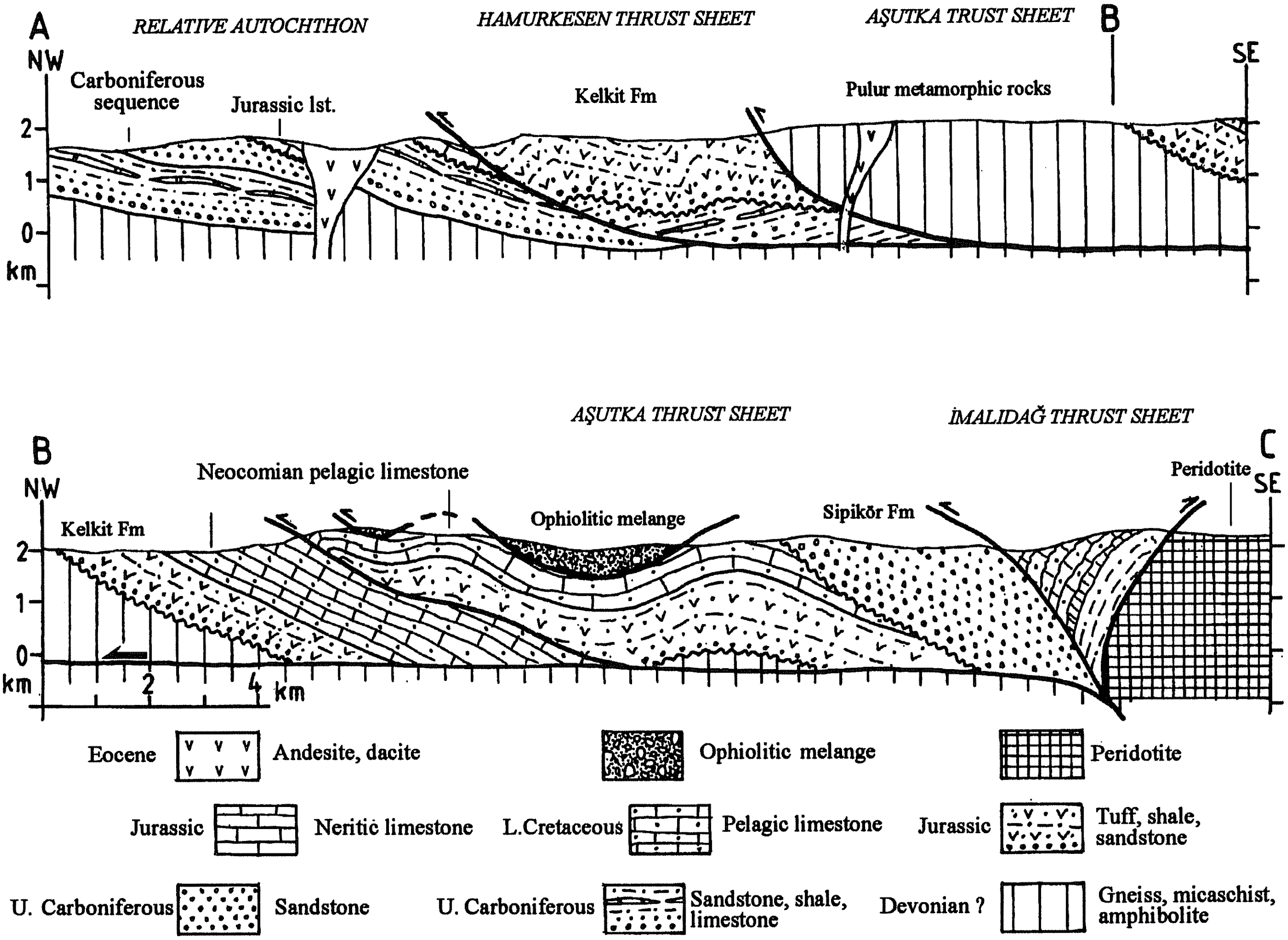

Figure 8. Geologic cross sections across the inner Eastern Pontides. For the location of the cross section, see Figure 3. Note the overlap between the top and bottom cross sections.

1987; Okay, 1984; Yılmaz, 1985). A major clastic foreland basin developed in front of these two northward-moving nappes (Figures 7, 8). Judging from its isolated outcrops, the foreland basin must have originally extended at least up to the Köse region (Gürsoy, 1989); however, it is now largely preserved as an 180-km-long and 3–5-km-wide east-to-northeast–trending belt below the İmalidağ and Çimendaği nappes (Figure 3). The basin fill consists of a few hundred-meters-thick conglomerates overlain by a >1000-m-thick upper Paleocene–lower Eocene turbidite sequence with large limestone olistoliths derived from the overlying nappes (Ketin, 1951; Bergougnan, 1976; Norman, 1976). Norman (1976) made a systematic study of the paleocurrent directions in the turbidites (the Spikör Formation) and showed that the currents in the basin were largely axial and were coming from the southwest. These data imply that the collision probably occurred along a northeast-southwest direction, which is oblique to the continental margin.

This major Paleocene/early Eocene compressive event, the strongest post-Jurassic deformational event in the Eastern Pontides, is probably related to the collision of the Eastern Pontides island arc with the Anatolides–Taurides along the Ankara–Erzincan suture in the south (Figure 7E). There is a clear northward decrease in deformation from a thrust belt in the south to a region of upright folding in the Alucra-Gümüşhane area and finally to a subtle unconformity in the outer Eastern Pontides in the north. Regional uplift caused deep erosion, which in places removed sections down to the pre-Mesozoic basement. All these are typical features of a continental collision: however, most workers (Tokel, 1977; Akin, 1978; Robinson et al., 1995) relate the Lutetian calc-alkaline magmatism to the northward subduction of the Tethys Ocean and consider the continental collision to be of Oligocene age. Others (Şengör and Yılmaz, 1981) initiate the collision in the Paleocene–early Eocene, but still relate the Lutetian magmatism to ongoing subduction. However, Lutetian rocks are largely undeformed and lie with a major angular discordance over all the older units, indicating that there was no major regional compression in the Eastern Pontides after the middle Eocene, which is difficult to believe if the continental collision had occurred in the Oligocene. The Senonian arc volcanism is confined to a relatively narrow belt, a characteristic feature of the present-day arc volcanism (Hamilton, 1995). In contrast, Lutetian magmatism, although much more restricted in time, is spread over the 100 km breadth of the Eastern Pontides and extends south to the Ankara-Erzincan Tethyan suture. Such a diffuse magmatism cannot be related to the activity of a single subduction system over a restricted time. A third line of evidence comes from the Kösedağ syenite pluton of middle-late Eocene age (42–37 Ma Rb/Sr whole-rock age) (Kalkanci, 1974) located northeast of Sıvas (Figure 2). It is an alkaline pluton with within-plate and postorogenic geochemical features (Boztuğ et al., 1994), and intrudes the suture zone. Thus, the weight of the geological evidence is for a Paleocene–early Eocene collision between the Pontides and the Anatolide-Tauride Platform (Elmas, 1995).

MIDDLE EOCENE—A NEW SEDIMENTARY AND VOLCANIC CYCLE

In the Eastern Pontides, Lutetian deposits transgressed over an existing folded and thrust-faulted surface (Figure 6). The Lutetian deposits are generally not folded and are not involved in thrust faults. They mark a new cycle of marine sedimentation and volcanism, with ages confined largely to the Lutetian. Lutetian rocks in the Eastern Pontides were deposited in shallow-water and are characterized by the presence of abundant nummulites in the limestones and sandstones intercalated with the volcanic rocks. The volcanism was more basic in composition than the Senonian volcanism, and occurred throughout the Eastern Pontides. The base of the Eocene transgression is generally dated as Lutetian (Ketin, 1951; Nebert, 1964; Bergougnan, 1987; Elmas, 1995), although in some regions it might have been as early as late Ypresian (Bursuk, 1975). The Lutetian in the inner Eastern Pontides is represented by ~700-m-thick shallow-marine succession consisting of basal conglomerates overlain by nummulite-bearing sandy limestones and sandstones, which pass up to a volcaniclastic series of sandstone, siltstone, basaltic and andesitic tuff, marl and rare basaltic flows (Ketin, 1951; Nebert, 1961; Ağar, 1977; Pelin, 1977; Özer, 1984; Gürsoy, 1989; Bozkus, 1992a). In the outer Eastern Pontides, a similar sequence, ~1000 m thick but richer in volcanic rocks, constitutes the Lutetian sequence (Tokel, 1977; Korkmaz and Gedik, 1988; Korkmaz, 1993). Lutetian volcanic rocks represent a short-lived volcanic cycle known as the Upper Basic Series (Schultze-Westrum, 1962; Zankl, 1962a; Tokel, 1977; Akıncı, 1984) and show a similar but less pronounced silica enrichment with time as the Senonian volcanic cycle. Geochemically, Lutetian volcanic rocks in the outer Eastern Pontides are calc-alkaline and fall in the field of island-arc basalts and andesites (Tokel, 1977; Eğin et al., 1979; Eğin and Hirst, 1979; Akıncı, 1984; Terzioğlu, 1984). Several hornblende-biotite granodiorites, which intrude Eocene sediments and volcanics, constitute the plutonic members of this magmatic cycle (Schultze-Westrum, 1962; Zankl, 1962a). The few isotopic ages from these granitoids range from 45 to 30 m.y. (Çoğulu, 1975; Moore et al., 1980). However, unlike the Senonian magmatism, the Lutetian magmatic rocks are not restricted to the outer Eastern Pontides but occur throughout the Eastern Pontides and adjacent Taurides. They intrude the late Paleocene–early Eocene thrust contacts in the inner Eastern Pontides north of Erzincan (Figure 3). In the Ankara-Erzincan suture zone northeast of Sıvas, the middle–late Eocene (42–37 m.y.) Kösedağ pluton (Kalkanci, 1974) intrudes the Lutetian volcano-sedimentary formations (Figure 2). Lutetian deposits south of the Ankara-Erzincan suture also comprise andesitic lavas, tuffs, and pyroclastic rocks intercalated with shallow-water sedimentary rocks (Özgül, 1981; Yılmaz, 1985). This and the major Paleocene–early Eocene unconformity show that the Senonian and Lutetian volcanic sequences represent two distinct cycles (Manetti et al., 1983, 1988).

The cause of the widespread Lutetian volcanism is not clear. A Tibetan-type postcollisional magmatism triggered by crustal thickening is unlikely, as the Lutetian magmatism was wholly submarine, suggesting regional extension only a few million years after the collision. This regional extension and the associated magmatism could have been related to the accelerated opening of the oceanic Eastern Black Sea Basin. Okay et al. (1994) suggested that the Eastern Black Sea Basin has opened by the counterclockwise rotation of a large continental block north of the Eastern Pontides. Although the rotation and thus the opening of the Eastern Black Sea Basin probably started in the latest Cretaceous, an increase in the rotation rate during the Lutetian might have resulted in the extension and magmatism throughout the Eastern Pontides. However, Eocene magmatism is also very widespread in the Lesser Caucasus and Iran (Kazmin et al., 1986), which suggests a more global cause for this enigmatic magmatic event.

OLIGOCENE AND YOUNGER DEPOSITS

None of the marine sequences in the Eastern Pontides reach to the late Eocene, showing that the Eastern Pontides were uplifted en bloc by the end of the mid-Eocene and have stayed largely above sea level since the middle Eocene, except for an Early Miocene marine incursion along the Ankara-Erzincan suture zone (Gedik, 1985; Inan, 1988; Yılmaz et al., 1988; Gökten, 1993). To the north of the Eastern Pontides, minor marine deposits of late Miocene (Sarmatian) and early Pliocene (Pontian) ages form small outcrops along the present-day Black Sea coast and represent small bays of the Paratethys (Özsayar, 1971, 1977).

The Oligocene to early Miocene history of northeastern Turkey is characterized by continuing compression caused by the ongoing collision of Gondwana and Laurasia after the complete elimination of the intervening oceanic basins. Most of this compression was accommodated by general south-vergent thrusting and folding along the Ankara-Erzincan suture to the south of the inner Eastern Pontides (Aktimur et al., 1990; Bozkus, 1992b). Small foreland basins, characterized by terrigenous sedimentation, developed during the Oligocene and Miocene in front of these thrust slices along the Ankara-Erzincan suture (Inan, 1988). There was also largely basaltic alkaline volcanism, which lasted up to the Recent (Terzioğlu, 1985). In the middle Miocene, the continuing compression resulted in the initiation of the North Anatolian fault, which started the westward escape of the Anatolian block, a tectonic regime that continues today (Şengör, 1979).

THE OTHER SIDE OF THE TETHYS OCEAN— THE TAURIDE MARGIN

Further insight on the evolution of the Tethyan Ocean south of the Eastern Pontides can be gained from the geology of the northern margin of the Taurides, south of the Tethyan suture. Pre-Tertiary lithologies south of the suture occur intermittently from south of Sıvas to south of Erzincan (Figure 2). They are best exposed in the Munzur Mountains southwest of Erzincan, where the Mesozoic is represented by Upper Triassic–Cenomanian platform carbonates overlain by Turonian–upper Campanian pelagic biomicrites (Figure 6) (Özgül 1981; Özgül and Tursucu, 1984). In the early Maastrichtian, ophiolitic melange with large peridotite slices was emplaced over the carbonates, and the carbonate platform was internally sliced along major south-vergent thrusts. In a 350-km-long region between south of Sıvas and Horasan, the ophiolitic melange and the peridotite slices are unconformably overlain by Maastrichtian to upper Paleocene shallow-water limestones (Özgül, 1981; Gedik, 1985; Yılmaz et al., 1988; Inan et al., 1993). In the Munzur region, this was followed by a period of uplift and erosion, and the Lutetian sandstone, limestone, tuff, basalt, and andesites unconformably overlie all the older units (Figure 6). In the Munzur region, the upper Eocene rocks are not present as in the Pontides, and the Oligocene is made up of terrigenous clastics and evaporites; the marine sequence starts again with Lower Miocene shallow-marine deposits (Gökten, 1993).

Geology of the northern margin of the Tauride block does not well constrain the age of the collision with the Eastern Pontides. Here, the major compressive event is of late Campanian–early Maastrichtian age and is related to the obduction of ophiolite and ophiolitic melange. This deformational event is precollisional and is similar to that observed in Oman, where the emplacement of the Semail ophiolite led to the imbrication of the Arabian continental margin (Robertson and Searle, 1990; Michard et al., 1991). Based on data from the Taurides, the collision with the Eastern Pontides could have been either during the early Eocene or in the late Eocene–Oligocene, during which there was folding and regional uplift.

CONCLUSIONS

The major stratigraphic and tectonic features of the Eastern Pontides can be summarized as follows.

There is a heterogeneous pre-Jurassic basement consolidated during the early Hercynian and Cimmeride orogenic events. Triassic sedimentary rocks are absent, possibly due to metamorphism during the Cimmeride orogeny. The Mesozoic sequence starts with a widespread Liassic transgression and continues essentially uninterrupted until the mid-Cretaceous, when there is a major break in sedimentation, with uplift and erosion of the entire Eastern Pontides. This compressive event caused the emplacement of an ophiolitic melange over the southern continental margin of the Eastern Pontides during the Cenomanian–Turonian. The compression and obduction of the melange is probably related to the partial subduction of the Eastern Pontides passive continental margin in a south-dipping intra-oceanic subduction zone. This was followed by a flip in the subduction polarity and, consequently, a volcanic arc developed during the Turonian to Maastrichtian-Danian in the outer Eastern Pontides above the northward-subducting Tethyan Ocean floor. The volcanic arc was extensional and wholly submarine. The inner parts of the Eastern Pontides were in a fore-arc position during the Senonian. The Eastern Black Sea Basin probably started to open during the Maastrichtian through the splitting of the volcanic arc axis.

Major thrust imbrication of the southern continental margin of the Eastern Pontides occurred during the late Paleocene–early Eocene. The thrusting involved the pre-Jurassic basement; thick foreland flysch basins have developed in front of the northward-moving nappes. In the outer Eastern Pontides, the late Paleocene–early Eocene is characterized by folding, uplift, and erosion. This orogenic event, the strongest in the Eastern Pontides during the Mesozoic and Tertiary, marks the continental collision between the Eastern Pontide arc and the Taurides.

Essentially undeformed basaltic and andesitic volcanic rocks and shallow-marine sedimentary rocks of Middle Eocene age occur throughout the Eastern Pontides; these are transgressive over a folded and thrust-faulted basement. They mark a regional extension, probably related to an accelerated phase of the opening of the Eastern Black Sea Basin. From the middle Eocene onward, the Eastern Pontides stayed above sea level, with minor volcanism and terrigenous sedimentation.

The Eastern Pontides stratigraphic sequence is rich in volcanic and volcaniclastic rocks. Three major Mesozoic–Tertiary magmatic cycles are recognized. The first cycle is Early to Mid-Jurassic, is probably of tholeiitic character, and is probably related to rifting. The second cycle is Turonian–Maastrichtian, is dominantly subalkaline, and is subduction related. The third cycle is Middle Eocene, is calc-alkaline, and is probably related to a regional extension.

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