A. Yin & T.M. Harrison (eds.), The Tectonic Evolution of Asia, 420-441
1996 Cambridge University Press.

19 Paleo- and Neo-Tethyan events in northwestern Turkey: Geologic and geochronologic constraints

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Abstract

Western Turkey is made up of several continental fragments with independent Paleozoic and Mesozoic geologic histories that had become assembled by the early Tertiary, following the closure of the intervening Tethyan oceanic basins. The imprint of a Paleozoic Tethyan ocean, the Paleo-Tethys, is preserved in the northern tectonic zones of Turkey and particularly in the Sakarya zone. The Sakarya zone comprises a Paleozoic continental basement of metamorphic rocks and granitoids that have been isotopically dated using a single-zircon–Pb evaporation technique. The granodiorite shows an earliest Devonian age (~400 Ma), and the gneisses of the basement show mid-Carboniferous ages (~310 Ma), interpreted as the time of Hercynian high-grade metamorphism. These ages indicate the Laurasian affinity of the continental basement of the Sakarya zone. During the late Triassic, that basement was overthrust from the south by a Permo–Triassic volcanic arc and related accretionary complexes of the Paleo-Tethys, represented by the Karakaya Complex. The exotic Middle Carboniferous and Lower Permian radiolarian chert blocks in the Triassic accretionary complexes indicate a minimum Carboniferous age for the Paleo-Tethys. The more abundant Upper Permian shallow-water limestone blocks in the Karakaya Complex, on the other hand, probably were derived from a small continental sliver rifted from the margin of Gondwana during the late Permian–early Triassic. This back-arc rifting, caused by the southward subduction of the Paleo-Tethys, developed in the late Triassic to produce a Neo-Tethyan ocean.

During the Mesozoic, three oceanic basins, separated by continental blocks, existed between the two megacontinents Laurasia and Gondwana in the western Turkey transect. In the Cretaceous, these three oceanic basins began closing by northward subduction. The $^{40}Ar/^{39}Ar$ isotope ages from the amphibolites below the peridotite thrust sheets indicate that the intra-oceanic slicing leading to the obduction had started by the Aptian (~118 Ma) in the northern ocean and by the Albian (~101 Ma) in the central ocean. The northward subduction of the northern ocean led to the Cretaceous opening of the Black Sea as a back-arc basin, with the rifting of a small continental sliver from the Laurasia margin. That small continental fragment collided during the early Eocene with the continental block in the south, thus obliterating the northern ocean and resulting in relatively minor deformation and localized regional metamorphism. On the other hand, the closure of the central ocean gave rise to large-scale ophiolite obduction, blueschist metamorphism, and internal slicing of the central Anatolide–Tauride–Pelagonian block that lasted from the late Cretaceous to the Miocene. The irregular northern margin of the central Anatolide–Tauride–Pelagonian block had a major influence on the timing and direction of ophiolite obduction and the consequent blueschist metamorphism.

Introduction

Turkey, forming an east–west bridge between Europe and Asia, also straddles the geologic boundary between Gondwana and Laurasia along a north–south transect. It was not a single entity until the early Tertiary, when several continental fragments with independent Paleozoic and Mesozoic geologic histories were assembled during a complex sequence of events leading to the collision of Gondwana and Laurasia. The convergence between Gondwana and Laurasia involved the collision and amalgamation of these continental fragments, as well as the opening of new back-arc basins, such as the Black Sea, which in turn created new continental slivers. Although Gondwana and Laurasia presently are in direct contact in eastern Turkey along various sutures, two remnant oceanic basins, the Black Sea and the eastern Mediterranean, are interposed between the two megacontinents in the western Turkey transect (Figure 19.1). As such, the orogenic history of Turkey is more complex than the relatively simple continent–continent-collision orogeny such as that of the Himalaya or the Qinling orogen in China (e.g., Mattauer et al., 1985; Le Fort, 1989).

Another feature that distinguishes the Alpide orogen in Turkey is the strongly curved nature of the fold belts, as opposed to the long straight segments observed in the Himalaya and Qinling orogens. An example is the Vardar–Izmir–Ankara suture zone, which makes two large loops when traced from

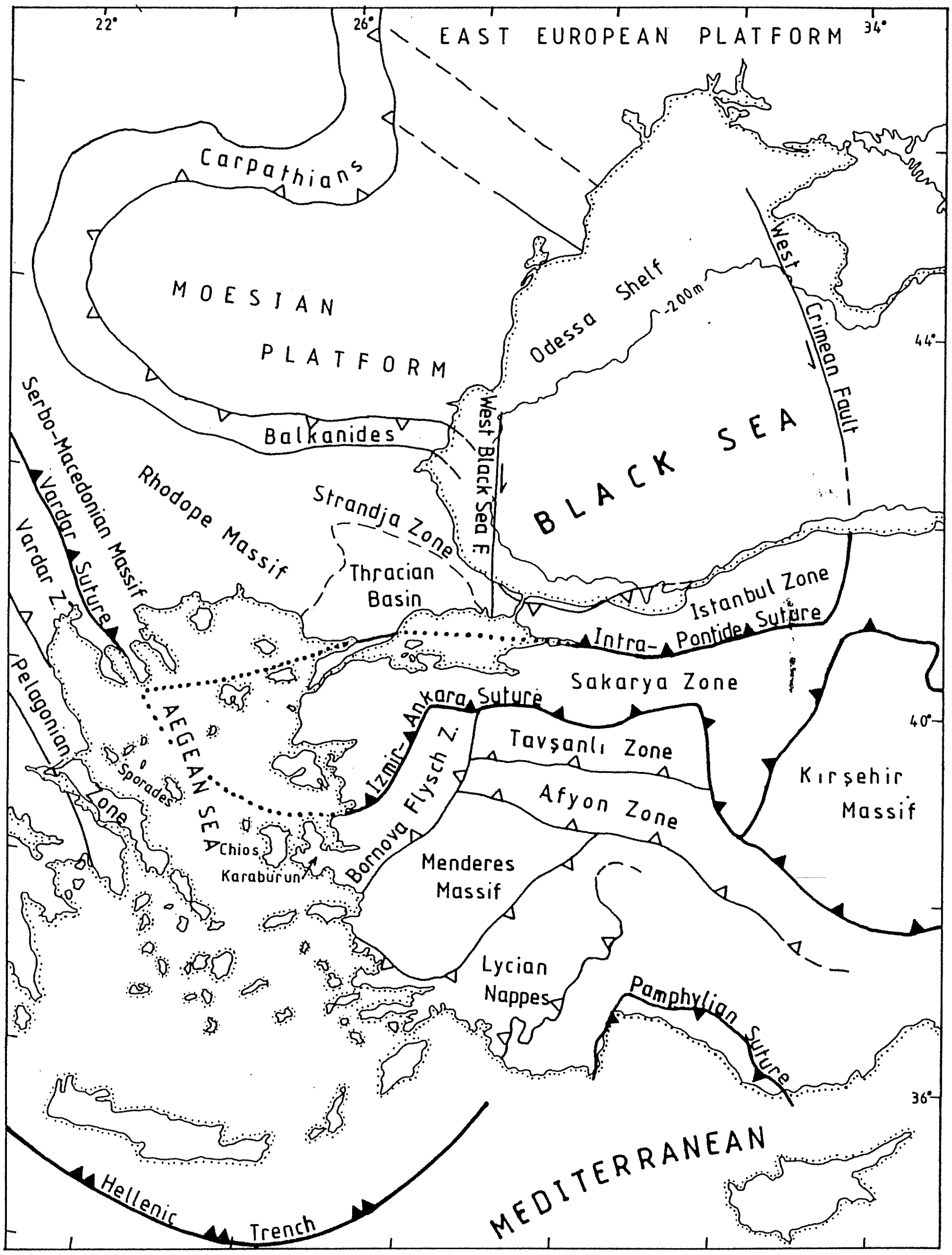


Figure 19.1. Tectonic map of western Turkey and surrounding regions, showing major continental blocks and tectonic zones. Thick barbed lines indicate the Mesozoic sutures with the former subduction polarities. The line with the double triangles shows the Hellenic subduction zone south of Crete. Thinner lines with open triangles denote major intracontinental thrusts. Data from Şengör and Yılmaz (1981), Okay (1989), and Okay and Siyako (1993).

Greece to central Turkey (Figure 19.1). The origin of these loops, whether they reflect the original irregular margins of the former plates or whether they were formed later during the continental collision, is not clear.

A controversial point in the geologic evolution of the eastern Mediterranean is the significance of the Paleo-Tethys, a Tethyan ocean of Paleozoic age. Permian paleogeographic reconstructions of the world (e.g., Smith, Hurley, and Briden, 1981; Scotese and Golonka; 1992) show a westward-narrowing Tethys ocean separating Gondwana and Laurasia in the region of the eastern Mediterranean. However, as initially pointed out by Smith (1973), almost all the known ophiolites and continental-margin sequences in this region are of Mesozoic age and are regarded as remnants of a Mesozoic Tethys, the Neo-Tethys. Although field evidence for a Paleozoic Tethyan ocean has subsequently been presented (e.g., Şengör, Yılmaz, and Ketin, 1980; Şengör, Yılmaz, and Sungurlu, 1984a; Tüysüz, 1990; Stampfli, Marcoux, and Baud, 1991; Ustaömer and Robertson, 1994), these data, as well as the locations of the Paleo-Tethyan suture and the suggested discontinuity between the Paleo-Tethys and Neo-Tethys, have been disputed at various times (e.g., Bergougnan and Fourquin, 1982). During the past 10 years, a large amount of information on the geology of northwestern Turkey has become available as a result of a coordinated systematic mapping program supported by the Turkish Petroleum Company. This information, which exists in reports or has been published in local journals, has major implications for the Paleo-Tethyan and Neo-Tethyan evolution. We shall synthesize this information and present new isotope age data that will shed new light on Paleo-Tethyan and Neo-Tethyan events in the eastern Mediterranean.

Tectonic framework of northwestern Turkey

Northwestern Turkey between the Black Sea in the north and the Aegean Sea in the west lies within the Alpide orogenic belt. Two major Alpide sutures divide northwestern Turkey into three continental blocks (Figure 19.1) (Şengör and Yılmaz, 1981). The Istanbul and Strandja zones in the north can be regarded as parts of the late Mesozoic Laurasian active continental margin. They are separated by the Intra-Pontide suture from the Sakarya zone, which had an important Carboniferous (Hercynian) and Triassic (Cimmeride) orogenic history, but during the late Mesozoic was an appendage of Laurasia (Figure 19.2). The various tectonic zones to the south of the Izmir–Ankara suture, differentiated by distinctive stratigraphic, metamorphic, and deformational features (Figure 19.3), constitute the Anatolide–Tauride block, which extends in the west to the Pelagonian zone in Greece. The Anatolide–Tauride block shows a stratigraphy similar to that of the Gondwana margin, but was separated from it by a minor oceanic basin during the late Mesozoic (Figure 19.2) (Şengör and Yılmaz, 1981). The present-day eastern Mediterranean is a possible relict of that ocean. The final amalgamation of these continental blocks in western Turkey was completed by the middle Eocene. During the Miocene and Pliocene terrigenous sedimentary and volcanic rocks covered large areas in northwestern Turkey (Figure 19.4). That was also the time of initiation of the dextral strike-slip North Anatolian fault (Şengör, 1979). The North Anatolian fault, which is confined to a narrow west-trending zone in central Turkey, changes its strike to southwest and expands to a 40-km-wide zone in the Biga Peninsula in northwestern Turkey (Figure 19.4). These two factors, the extensive Neogene cover and the effects of the post-Miocene dextral strike-slip faulting, have made it difficult to unravel the pre-Eocene geologic history of northwestern Turkey.

Istanbul Zone: Hercynian fragment rifted from Laurasia

The Istanbul zone, located along the western Black Sea coast (Figure 19.1), consists of a Precambrian crystalline basement overlain by a continuous, well-developed transgressive sedimentary sequence extending from the Ordovician to the Carboniferous that represents the southward-facing Paleozoic passive continental margin of Laurasia (Figure 19.3) (Tokay, 1952; Haas, 1968). The Paleozoic sequence was folded and possibly thrust-faulted during the late Carboniferous–Permian Hercynian orogeny and was then unconformably overlain by Triassic and younger sedimentary rocks, particularly well developed east of Istanbul (Figure 19.3) (Özdemir, 1981). The Istanbul zone is separated in the south from the Sakarya zone, by the Intra-Pontide suture, and in the west, a pre-Eocene dextral strike-slip fault, the West Black Sea fault, constitutes its boundary with the Strandja zone (Figure 19.1).

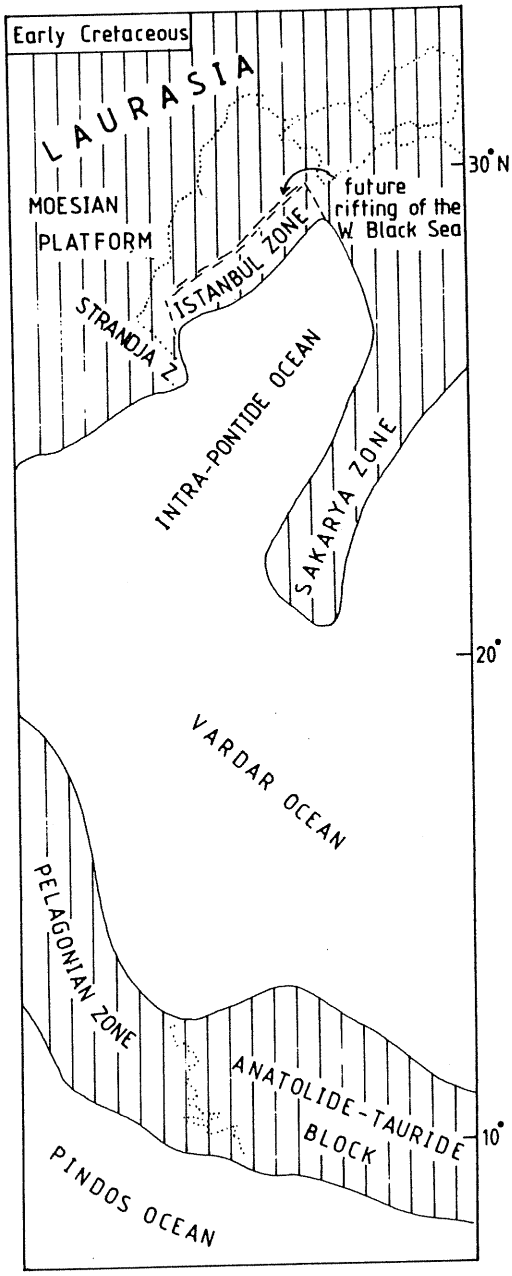


Figure 19.2. Sketch map of the Aegean region during the early Cretaceous, showing the preferred locations and orientations of the continental blocks and oceanic areas discussed in the text. The map is a compromise between the single-ocean model of Dercourt et al. (1986) and the many-ocean model of Robertson et al. (1991) and is similar to that of Şengör and Yılmaz (1981), except in the position of the Sakarya zone. The orientations of the blocks are adapted from Smith et al. (1981).

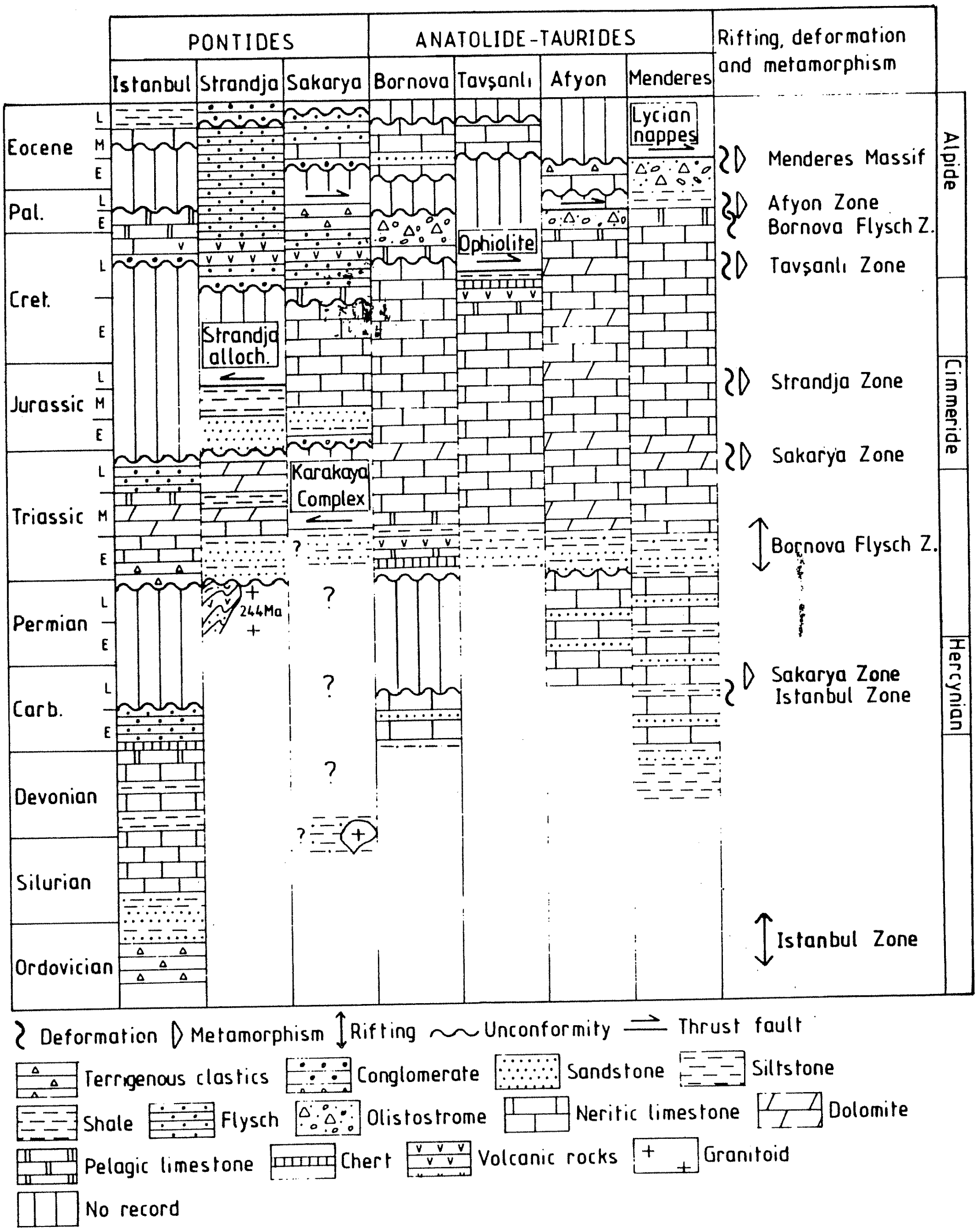


Figure 19.3. Synthetic stratigraphic sections of the tectonic zones of western Turkey. The Istanbul, Strandja, and Sakarya zones traditionally are collectively termed the Pontides (Ketin, 1966).

The Paleozoic stratigraphy of the Istanbul zone is exotic to the rest of northwestern Turkey, where in situ pre-Permian sedimentary sequences are not known, and it bears close resemblances to the Laurasian margin sequences of the Moesian platform. The paleo-latitudes for the Paleozoic (Evans et al., 1991) and Triassic rocks (Sarıbudak, Sanver, and Ponat, 1989) of the Istanbul zone are also compatible with a location along the southern margin of Laurasia. To explain these features, Okay, Şengör, and Görür (1994) suggested that the Istanbul zone rifted from the southern margin of Laurasia during the late Cretaceous, following the opening of the western Black Sea as a back-arc basin above the northward-subducting Intra-Pontide ocean. It moved south, opening the oceanic western Black Sea basin in the north and closing the Intra-Pontide ocean in the south, and eventually collided in the early Eocene with the Sakarya zone (cf. Figure 19.2) (Okay et al., 1994). The evidence for this model is the close stratigraphic similarity between the Istanbul zone and the Moesian platform, the morphological fit of the Istanbul zone to the Odessa shelf between Moesia and Crimea, and the presence in the western Black Sea of the strike-slip faults that have guided the Istanbul zone to the south (see Figure 19.1). Thus, prior to the late Cretaceous, the Istanbul zone was located farther north than the Strandja zone, and it reached its present position only in the early Eocene.

Strandja massif: a late Cimmeride zone with a Hercynian basement

The Strandja massif forms the eastern extension of the poorly understood Rhodope massif in Greece and Bulgaria (Figure 19.1). It is bordered on the south by Middle Eocene and younger clastic sediments (>8 km thick) of the Thrace basin (Turgut, Türkaslan, and Perinçek, 1991). In the east, its contact with the Istanbul zone is covered by a 17-km-wide strip of undeformed Middle Eocene limestone and marl, under which the West Black Sea fault is presumed to lie.

The Turkish part of the Strandja zone consists of a basement of felsic gneiss and migmatite, intruded by Hercynian granitic plutons, one of which shows a 244 ± 11 Ma Rb-Sr isochron age, based on three rock samples (Aydın, 1974). The basement is unconformably overlain by Lower to Middle Triassic continental to shallow-marine clastics and carbonates that show stratigraphic affinities with Germanic Trias facies (Kasar and Okay, 1992). The sequence in the Bulgarian part is better known and extends up to the middle Jurassic (Figure 19.3) (Chatalov, 1988). Recent studies in the Turkish part of the Strandja zone (A. Okay, O. Tüysüz, and S. Akyüz, unpublished data, 1994) have shown the presence of granitic basement nappes several kilometers thick that tectonically overlie deformed and metamorphosed cover sequences across thick mylonite zones. Higher allochthonous structural levels are exposed in the Bulgarian part, where the autochtonous sequence is tectonically overlain by Triassic graywacke-shale series with volcanic-rock and carbonate intercalations (Chatalov, 1988).

Both the autochthon and the allochthon of the Strandja zone were deformed and metamorphosed to greenschist facies during the mid-Jurassic, as reflected by the 155–149-Ma biotite K-Ar ages in the basement granites (Aydın, 1982). The metamorphic rocks of the Strandja zone are unconformably overlain by Cenomanian conglomerates and shallow-marine limestones that progress up to a thick Upper Cretaceous andesitic volcanic and volcaniclastic series. Granodiorite plutons and stocks were intruded into the metamorphic rocks as part of the calc-alkaline magmatism; the largest of the plutons (the Demirköy pluton) shows K-Ar hornblende and biotite ages of 84–78 Ma (Moore, McKee, and Akıncı, 1980). The late Cretaceous calc-alkaline magmatism in the northern parts of the Strandja zone represents, in continuity with the Srednogorie zone in Bulgaria, the establishment of an Andean-type magmatic arc above the northward-subducting Intra-Pontide ocean (e.g., Boccaletti, Gocev, and Manetti, 1974; Şengör and Yılmaz, 1981).

Microstructural features at the base of the granite nappe in the Turkish part of the Strandja zone indicate top-to-the-north shearing during thrusting (A. Okay, O. Tüysüz, and S. Akyüz, unpublished data, 1994.) The presence of Triassic shallow-water carbonate sequences to the north of the Strandja zone in Bulgaria (e.g., Ganev, 1974; Chatalov, 1991) also points to a southward derivation of the Strandja cover allochthons, presumably from a Triassic active continental margin. The latest Triassic to middle Jurassic orogeny in the Strandja zone was related to the collision of that active margin with the passive continental margin of Laurasia, represented by the autochthonous series in the Strandja zone.

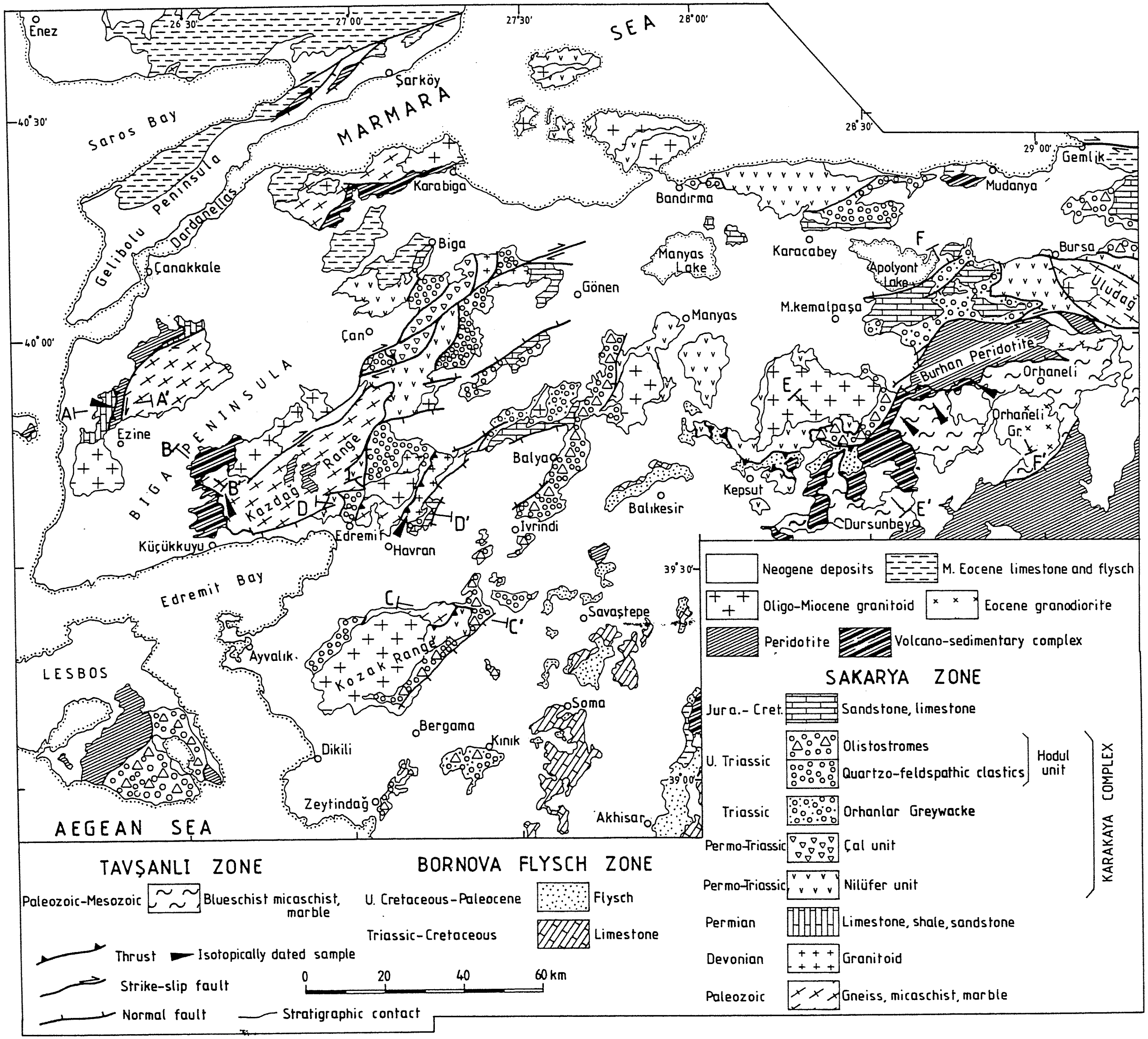
Intra-Pontide suture and related oceanic rocks

The southern boundary of the Strandja and Istanbul zones is constituted by the Intra-Pontide suture (Şengör and Yılmaz, 1981). The actual trace of the suture is generally covered by the Middle Eocene and younger sedimentary rocks of the Thrace basin and by the Marmara Sea, and its location is constrained by the small phyllite inlier of the Rhodope/Strandja massif north of Saros Bay and the elongated fault-bounded uplifts of the volcano-sedimentary complex north of Şarköy in Thrace (Figure 19.4). The Intra-Pontide suture trends from north of the Gelibolu Peninsula westward and joins the İzmir–Ankara–Vardar suture in the Aegean Sea (Figure 19.1)

Large peridotite and volcano-sedimentary-complex bodies were thrust up to 50 km south and east from the Intra-Pontide suture during the late Cretaceous–Paleocene. The volcano-sedimentary complex in the Biga Peninsula and Thrace consists of mafic volcanic and pyroclastic flows, radiolarian chert, Triassic, Upper Cretaceous, and Middle Paleocene pelagic limestone, siliceous shale, sandstone, and minor serpentinite. Exotic blocks of eclogite and blueschist are also present in the volcano-sedimentary complex west of the Kazdağ range (Okay, Siyako, and Bürkan, 1991) and west of Şarköy in Thrace (Kopp, Pavoni, and Schindler, 1969; Şentürk and Okay, 1984), respectively. West of the Kazdağ range, the volcano-sedimentary complex rests over felsic gneisses; with its normal-fault contact marked by a 2-km-thick mylonite zone with serpentinite lenses, thus forming the carapace of an Eocene extensional core complex (Figures 19.4 and 19.5B) (Okay, Satir, and Metzger, in press). The volcano-sedimentary complex, whose internal structure ranges from imbricate tectonic slices to tectonic mélange, represents a sediment-starved accretionary complex made up of the upper layers of an oceanic-crustal section. The insignificance of graywackes in this accretionary complex, probably related to the presence of extensive carbonate platforms surrounding the continental side of the trench, distinguishes it from the turbidite-dominated accretionary complexes of Franciscan or Makran type.

Large peridotite bodies north of Ezine in the Biga Peninsula and on the island of Lesbos (Figure 19.4) were also obducted from the Intra-Pontide ocean. North of Ezine, the peridotite rests

Figure 19.4. Geologic map of northwestern Turkey south of the Marmara Sea, using data from Okay et al. (1991) and Okay and Siyako (1993). The tips of the small black triangles point to the sample localities isotopically dated in this study and those dated by Okay and Kelley (1994) and Harris et al. (1994).



Enez
26°30'
27°00'
27°30'
28°00'
28°30'
29°00'
40°30'
40°00'
39°30'
39°00'
Saros Bay
Gelibolu Peninsula
Dardanelles
Çanakkale
MARMARA SEA
Şarköy
Karabiga
Biga
Bandırma
Mudanya
Gemlik
Bursa
Uludağ
Manyas Lake
Karacabey
Apolyont Lake
M.kemalpaşa
Gönen
Manyas
Çan
BIGA PENINSULA
Ezine
Kazdağ Range
Edremit
Havran
Balya
Ivrindi
Balıkesir
Kepsut
Dursunbey
Burhan Peridotite
Orhaneli
Orhaneli Gr.
Küçükkuyu
Edremit Bay
Ayvalık
Kozak Range
Savaştepe
Soma
Bergama
Kınık
Dikili
Zeytindağ
Akhisar
LESBOS
AEGEAN SEA
A
A'
B
B'
C
C'
D
D'
E
E'
F
F'
Neogene deposits
M. Eocene limestone and flysch
Oligo-Miocene granitoid
Eocene granodiorite
Peridotite
Volcano-sedimentary complex
SAKARYA ZONE
Jura.- Cret.
Sandstone, limestone
U. Triassic
Olistostromes
Quartzo-feldspathic clastics
Hodul unit
Triassic
Orhanlar Greywacke
Permo-Triassic
Çal unit
Permo-Triassic
Nilüfer unit
KARAKAYA COMPLEX
Permian
Limestone, shale, sandstone
Devonian
Granitoid
Paleozoic
Gneiss, micaschist, marble
TAVŞANLI ZONE
Paleozoic-Mesozoic
Blueschist micaschist, marble
BORNOVA FLYSCH ZONE
U. Cretaceous-Paleocene
Flysch
Triassic-Cretaceous
Limestone
Thrust
Isotopically dated sample
Strike-slip fault
Normal fault
Stratigraphic contact
0
20
40
60 km

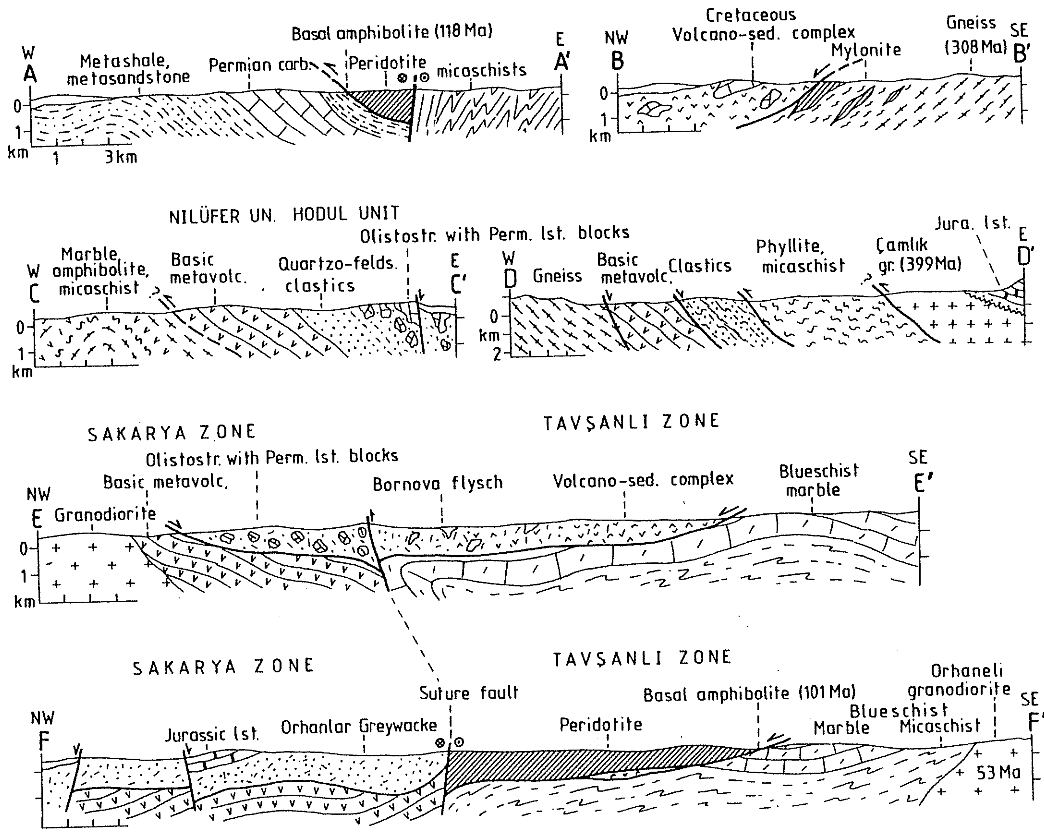


Figure 19.5. Simplified cross sections illustrating the large-scale structures in various parts of northwestern Turkey. For locations of cross sections, see Figure 19.4.

tectonically, through an intervening 350-m-thick shale-siltstone-calciturbidite layer, on a 1600-m-thick coherent sequence of shallow-water limestone, shale, siltstone, and sandstone (Figure 19.6). Although the whole sequence has undergone low-greenschist-facies metamorphism, Permian fusulinids are present in the upper parts of the carbonate sequence. A Permo–Triassic ophiolite obduction has been suggested for this region, based on the tectonic position of the peridotite above the Permian carbonates (Okay et al., 1991). To test that hypothesis, two amphibolite samples (4328 and 4329) from the base of the peridotite (Figure 19.6) were isotopically dated by the $^{40}Ar/^{39}Ar$ method. The amphibolites occur as discontinuous coarse-grained lenses of hornblende and plagioclase, up to 50 m thick, directly at the base of the peridotite and have been interpreted as composing a syn-thrusting metamorphic sole. Field and petrologic studies have shown no inverted metamorphic gradient. The analytical procedures used for isotope dating were the same as those described by Maluski, Costa, and Echtler (1991). The $^{40}Ar/^{39}Ar$ analysis of hornblende and plagioclase from the two samples gave complex argon-release spectra (Figure 19.7, Table 19.1). Hornblende ages were 117.0 ± 1.5 Ma and 118.3 ± 3.1 Ma, and a plagioclase age from one sample was 109.6 ± 2.4 Ma. The 117-Ma and 118-Ma hornblende ages from the two samples have been interpreted to date the time of cooling below the hornblende-closure temperature. This indicates, contrary to the earlier suggestion, an early Cretaceous (Aptian) emplacement age for the peridotite.

Sakarya zone: Hercynian fragment with Paleo-Tethyan active-margin units

Most of the evidence pertaining to the Paleo-Tethyan evolution in Turkey lies in the complexly deformed and partly metamorphosed pre-Jurassic basement of the Sakarya zone. As the Sakarya zone was the upper plate during the Cretaceous northward subduction of the Neo-Tethyan Vardar ocean, Jurassic and younger rocks of the Sakarya zone show relatively little Alpide deformation and no Alpide metamorphism, except in the vicinity of the Intra-Pontide suture in the north (Göncüoğlu and Erendil, 1990; Yılmaz et al., 1994).

Şengör and Yılmaz (1981) saw the Sakarya zone as an isolated continent during the late Mesozoic, surrounded by the Intra-Pontide ocean in the north and the Vardar ocean in the south. However, the Red Sea–type termination of the Intra-Pontide ocean toward the east, as inferred from its suture pattern, and the presence of Triassic sequences in Crimea similar to those observed in the basement of the Sakarya zone (e.g., Kotanski, 1978) indicate that the Sakarya zone was an appendage of Laurasia during the Jurassic and Cretaceous (Figure 19.2).

The basement of the Sakarya zone can be subdivided into two tectonic assemblages that were juxtaposed during the late Triassic: a lower assemblage of Paleozoic granitic and metamorphic rocks, and an upper assemblage of accretion-subduction units of the Paleo-Tethys, termed the Karakaya Complex.

Table 19.1. *Ar-isotope data for amphibolite samples from the base of the peridotite in the Biga Peninsula*

Temp. (°C)	$^{40*}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$ (*100)	$^{37}Ar/^{39}Ar$ (*100)	$^{39}Ar/^{40}Ar$ (*100)	% ^{39}Ar (cumul.)	Age (Ma)
4328C hornblende, J = 0.011819						
500	3.11	16.57	564.464	1.935	0.45	65.1 ± 22.9
600	1.78	2.24	518.534	12.018	3.19	37.6 ± 3.4
700	5.41	0.66	138.907	13.666	11.35	112.8 ± 1.7
800	0.00					
850	5.42	0.39	0.000	15.352	24.00	112.0 ± 1.4
900	6.03	0.20	122.820	15.202	38.25	124.2 ± 1.3
950	5.77	0.05	449.041	17.074	47.85	119.0 ± 1.2
1000	5.68	0.44	2156.114	14.523	55.06	117.2 ± 1.9
1050	0.00	3.62	5346.028	9.823	61.24	0.0
1100	0.00	5.22	6773.713	8.561	68.49	0.0
1150	5.96	1.82	7968.903	9.022	75.59	122.9 ± 5.2
1450	3.24	3.19	8222.577	8.070	100.00	67.9 ± 6.9
4328C plagioclase, J = 0.010199						
500	7.05	29.75	84.550	1.061	0.10	125.3 ± 34.3
550	9.73	5.83	159.681	3.737	0.59	170.6 ± 7.3
600	7.90	1.17	267.279	8.868	1.80	139.7 ± 2.3
650	5.40	0.76	196.155	13.168	4.47	96.6 ± 1.6
700	5.58	0.33	112.754	15.371	8.19	99.9 ± 1.2
750	5.75	0.57	81.893	13.548	16.85	102.8 ± 1.4
800	6.02	0.32	56.838	14.485	28.81	107.4 ± 1.2
850	6.26	0.43	61.286	13.406	45.18	111.6 ± 1.4
900	6.09	0.57	233.865	12.951	72.68	108.7 ± 1.5
950	6.23	0.41	732.359	13.550	84.93	111.1 ± 1.4
1000	0.00	3.71	7839.077	9.473	91.67	0.0
1050	0.00	7.87	22067.220	7.950	94.60	0.0
1100	2.56	2.81	7838.881	9.405	96.46	46.6 ± 5.5
1150	2.27	2.30	2988.866	11.194	97.72	41.3 ± 3.7
1450	14.15	6.00	32941.060	3.230	100.00	243.1 ± 13.2
4329C hornblende, J = 0.011819						
500	6.13	16.49	808.364	1.837	0.36	126.2 ± 22.3
600	5.12	2.48	823.245	8.093	2.03	106.1 ± 4.1
700	4.94	1.58	340.650	10.481	6.31	102.3 ± 2.8
800	6.22	0.20	328.583	14.809	13.40	128.0 ± 1.4
850	5.85	0.14	128.403	16.115	21.71	120.6 ± 1.3
900	5.75	0.01	199.874	17.406	30.20	118.6 ± 1.1
950	5.55	0.03	454.168	17.876	35.78	114.6 ± 1.1
1050	4.75	1.37	2708.113	11.538	39.58	98.4 ± 3.1
1100	5.65	1.78	4154.842	9.307	48.42	116.7 ± 4.1
1150	5.83	1.52	4552.486	9.837	68.43	120.3 ± 3.9
1200	5.60	2.29	4791.471	8.221	74.13	115.6 ± 4.9
1450	7.65	1.39	4585.693	8.647	100.00	156.1 ± 3.8

Note: $^{40*}Ar$ = radiogenic argon; ^{40}Ar = total argon.

Paleozoic continental rocks of the Sakarya zone

Granitic and metamorphic rocks make up the Paleozoic continental basement of the Sakarya zone. They are well exposed in the tectonic windows of the Uludağ, Kazdağ, and Kozak ranges, where they are tectonically overlain by the Karakaya Complex (Figures 19.4 and 19.5C,D). As no isotope age data exist for the pre-Jurassic granitoids in northwestern Turkey, we have dated a granodiorite body north of Havran, the Çamlık granodiorite (Figure 19.4), using the single-zircon step-wise-evaporation technique of Kober (1986; 1987). The Çamlık granodiorite is a medium-grained, leucocratic body that consists dominantly of quartz, plagioclase, chloritized biotite, and minor K-feldspar. In the east it is unconformably overlain by the late Triassic–Jurassic sandstones and limestones, and in the west it overlies, probably tectonically, a meta-sedimentary sequence belonging to the

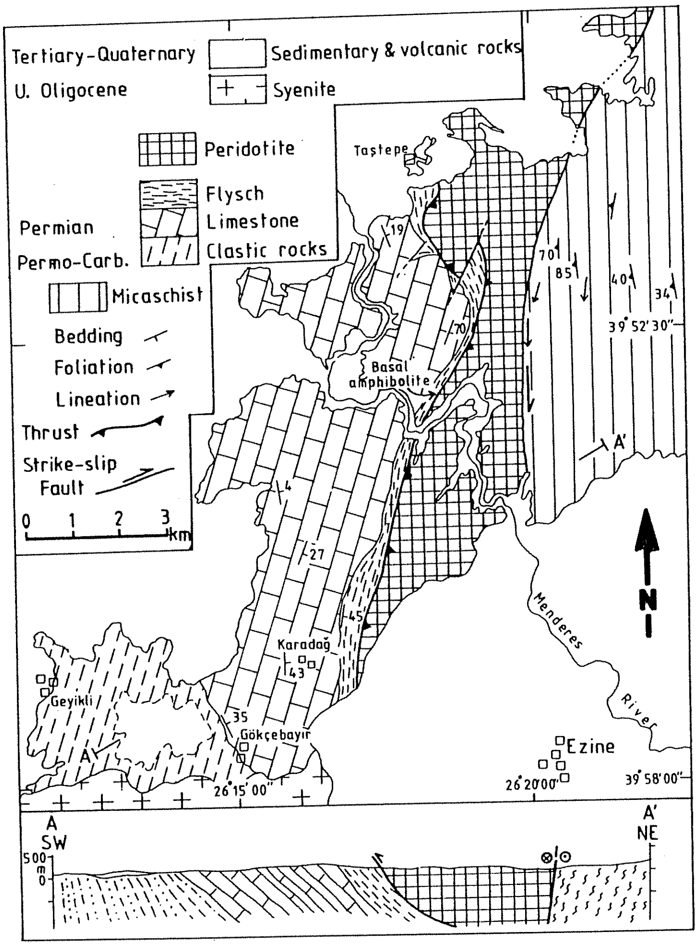


Figure 19.6. Geologic map of the Ezine region in the Biga Peninsula, showing the peridotite slab over the Permian neritic limestones, with the intervening syn-thrusting flysch. The locations of the dated amphibolite samples from the base of the peridotite are indicated. For orientation, see Figure 19.4.

continental basement of the Sakarya zone (Figure 19.5D). Along that western contact, the Çamlık granodiorite shows a distinct tectonic foliation that gradually dies away toward the east.

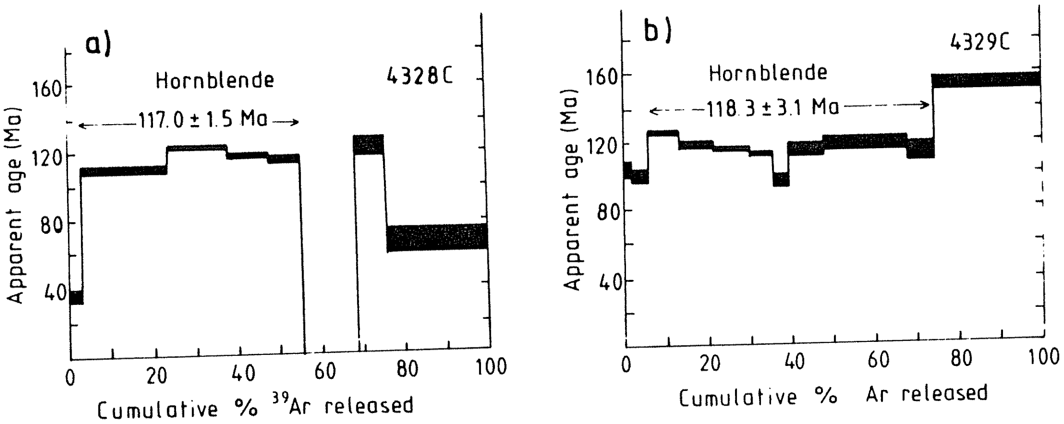
The single-zircon step-wise-Pb-evaporation dating method has the advantages that only small zircon samples are needed, no chemical procedures are involved and thus there is no possibility of blank contamination. This approach has given results that agree with those from conventional dating techniques (e.g., Chocherie, Guerrot, and Rossi, 1992). The analytical technique is described in the Appendix to this chapter.

Sample 93/1 from the Çamlık granodiorite contains distinct populations of brown and pale zircons, although a few zircons show intermediate colors. All the zircons are turbid and euhedral, although many are broken. A few zircons, especially the brown variety, show black spots and overgrowths. Five pale and brown unbroken euhedral zircons without inclusions or overgrowths chosen from a population of over 50 zircons were used for the dating. The results are shown in Table 19.2 and are summarized in the histogram of Figure 19.8A. The five individual $^{207}Pb/^{206}Pb$ step ages range from 378 ± 13 Ma to 415 ± 10 Ma, with a mean of 399 ± 13 Ma (Figure 19.8A). We interpret this earliest Devonian age as the time of crystallization of the Çamlik granodiorite, making it the oldest known granitoid from the Sakarya zone. No Paleozoic granitic rocks are known from the Anatolide–Tauride block, although Lower Paleozoic granitoids have been described from the Strandja zone in Bulgaria (Chatalov, 1988), suggesting a Laurasian affinity for the Paleozoic continental basement of the Sakarya zone.

The Paleozoic metamorphic rocks consist mainly of medium- to coarse-grained felsic gneiss, quartzofeldspathic mica schist intercalated with banded amphibolite and marble. The metamorphism was at high-grade-amphibolite facies to granulite facies, with local anatexis. The structural thickness of the metamorphic rocks is more than 10 km (Bingöl, 1969; Okay et al., 1991).

Two gneiss samples from the northwestern part of the Kazdağ range have been dated (Figure 19.4). The first sample (K13) is a cordierite gneiss in which three euhedral zircon populations with grain sizes larger than 0.125 mm can be differentiated; K13/1 is the transparent and colorless variety (locally with turbid cores), and K13/2 and K13/3 are turbid zircons that are colorless and yellowish white, respectively. From these populations of over 50 grains, three unbroken representative zircons without inclusions were selected for dating. The second sample (R25), a quartzofeldspathic gneiss, contains a single population of yellowish white, turbid, euhedral zircons. The three individual $^{207}Pb/^{206}Pb$ step ages from sample K13 and the one age from sample R25 range from 292 ± 8 Ma to 323 ± 14 Ma, with a mean of 308 ± 16 Ma (Table 19.1, Figure 19.8B). We interpret these mid-Carboniferous ages as the time of high-grade metamorphism and migmatization in the Kazdağ range (Metzger, Bracke, and Satır, in press). This provides the first evidence for the long-suspected Hercynian metamorphism in the continental basement of the Sakarya zone. Similar Carboniferous K-Ar and Rb-Sr ages were reported for the migmatites of the Serbo–Macedonian massif (Borsi, Ferrara, and Mercier, 1964), again emphasizing the connection between the continental basement of the Sakarya zone and the Laurasian margin. Bingöl (1971) reported a 233 ± 24-Ma Rb-Sr isochron age from the gneisses of the Kazdağ range, probably reflecting the time of a second, lower-grade regional metamorphism associated with the emplacement of the Karakaya Complex. The Oligocene (27–23 Ma) K-Ar ages and some Rb-Sr mineral ages (Bingöl, 1971) reflect the heating associated with the strong Oligo–Miocene magmatism in northwestern Turkey and the post-Pliocene up-doming of the Kazdağ range.

The Paleozoic continental metamorphic rocks of the Sakarya zone had a complex thermo-tectonic history, with mid-Carboniferous (Hercynian), late Triassic (Cimmeride), and Oligo–Miocene (Alpide) thermal events. It is noteworthy that the earliest Devonian Çamlık granodiorite does not show Hercynian metamorphism, although it lies only 20 km east of the Hercynian migmatitic gneisses of the Kazdağ range, suggesting major Hercynian and/or Cimmeride crustal shortening (cf. Figure 19.5D).



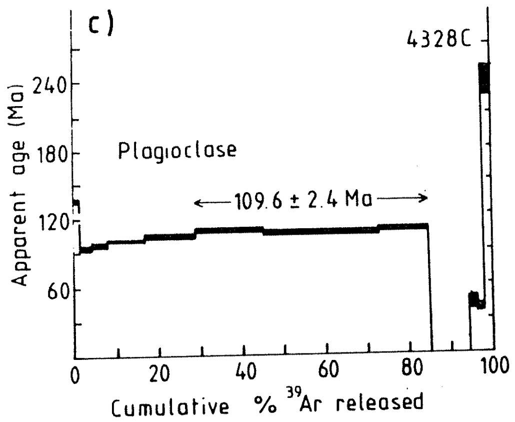


Figure 19.7. $^{40}Ar/^{39}Ar$ **ages for hornblende and plagioclase concentrates from two amphibolite samples (4328C and 4329C) from the base of the peridotite in the Ezine region, northwestern Turkey. Plateau ages (plateau increments outlined by arrows) are listed on each spectrum. See also Table 19.1.**

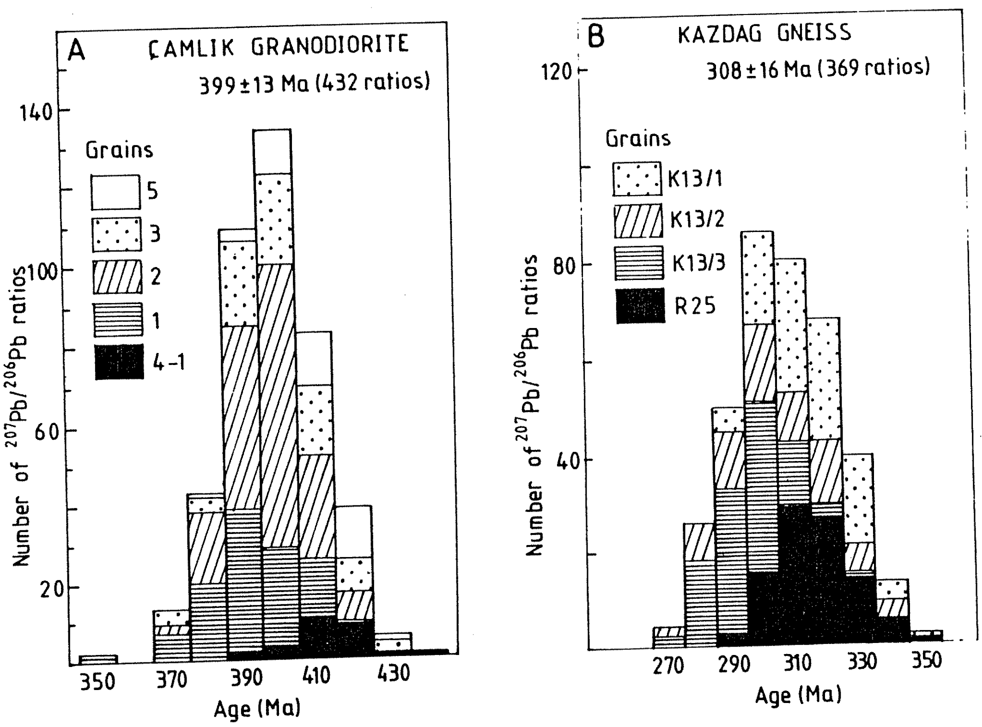


Figure 19.8. Histograms showing the distributions of radiogenic Pb-isotope ratios derived from evaporation of five individual zircon grains from the Çamlik granodiorite (A) and four zircon crystals from the migmatic gneisses of the Kazdağ range (B) in the Sakarya zone.

Table 19.2. *Pb-isotope data for zircons from Kazdağ gneisses and Çamlık granodiorite*

Grain number	Evaporation steps (°C)	Number of ratios	$^{206}Pb/^{204}Pb$ (mean)	$^{207}Pb/^{206}Pb$ [$\pm\ \sigma$ (mean)]	Age $\pm\ \sigma$
Kazdağ migmatitic gneiss (samples K13 and R25)					
K13/1	1305	33	9554	0.05268 ± 26	315 ± 11
K13/1	1356	33	5208	0.05268 ± 27	316 ± 11
K13/1	1380	33	3709	0.05266 ± 31	315 ± 13
K13/2	1305	18	1776	0.05259 ± 58	307 ± 25
K13/2	1331	16	3314	0.05221 ± 19	295 ± 8
K13/2	1380	38	4289	0.05255 ± 33	310 ± 14
K13/3	1331	37	6594	0.05219 ± 27	294 ± 12
K13/3	1356	36	8630	0.05235 ± 27	301 ± 12
K13/3	1380	33	7671	0.05213 ± 18	292 ± 8
R25	1343	36	20957	0.05254 ± 19	309 ± 8
R25	1377	32	26612	0.05275 ± 21	319 ± 9
R25	1408	24	18802	0.05286 ± 34	323 ± 14
	sum	369		mean	308 ± 16
Çamlık granodiorite (sample 93/1)					
1	1308	17	3000	0.05416 ± 31	378 ± 13
1	1343	47	21500	0.05443 ± 18	389 ± 7
1	1377	47	22000	0.05471 ± 22	401 ± 9
2	1308	30	3700	0.05448 ± 28	391 ± 12
2	1343	52	18000	0.05466 ± 17	399 ± 7
2	1377	55	40000	0.05454 ± 16	394 ± 6
2	1408	33	56000	0.05482 ± 19	405 ± 8
3	1377	18	20000	0.05469 ± 51	400 ± 21
3	1408	64	16000	0.05471 ± 25	401 ± 10
4–1	1440	26	20000	0.05506 ± 25	415 ± 10
5	1343	36	7000	0.05494 ± 27	410 ± 11
5	1377	7	19000	0.05488 ± 34	408 ± 14
sum/mean		432			399 ± 13

Paleo-Tethyan active-margin units: Karakaya Complex

The Karakaya Complex consists of Permo–Triassic intra-oceanic fore-arc deposits and late Paleozoic–Triassic accretionary complexes with abundant exotic limestone blocks (Figure 19.9) (Okay et al., 1991). Similar assemblages extend for more than 1,000 km eastward in the Sakarya zone and northward to the southern Crimea (e.g., Kotanski, 1978; Tekeli, 1981; Okay, 1984; Tüysüz, 1990; Ustaömer and Robertson, 1994).

Intra-oceanic fore-arc rocks: the Nilüfer unit

The Nilüfer unit, more than 3 km thick, is a semicoherent sequence of mafic tuffs, pyroclastic rocks, and pillow lavas interstratified with carbonate and shale bands up to 50 m thick (Figure 19.9). In some regions, such as south of Manyas and on the island of Marmara, the volcaniclastic rocks are stratigraphically overlain by carbonates several hundred meters thick. The Nilüfer unit rests, with a tectonic contact, over the Carboniferous gneisses of the Kazdağ and Uludağ ranges (Figure 19.4); these contacts, which initially may have been thrust faults, are presently large normal faults. The Nilüfer unit is overlain, generally with tectonic contacts, by the Upper Triassic clastic sequences of the Karakaya Complex (Figure 19.9).

The Nilüfer unit has undergone high-pressure greenschist-facies metamorphism, with the development of albite, chlorite, epidote, actinolite, and sphene in the fine-grained mafic tuffs. Sodic amphibole occurs in iron-rich tuffs, and the massive, coarse-grained pyroclastic flows retain most of their igneous texture and their igneous clinopyroxene. The greenschist-facies meta-basites of the Nilüfer unit include very rare tectonic lenses of ultramafic rock, generally a few tens of meters thick, and a 40-m lens of glaucophane eclogite east of Bandırma. The deformation in the Nilüfer unit is characterized by the development of cleavage in the phyllites and fine-grained meta-tuffs and by mesoscopic upright isoclinal folds. The more massive marble bands have been boudinaged, giving a "broken-formation" character to the Nilüfer unit. Middle Triassic conodonts have been described in the carbonates interstratified with the meta-basites from the uppermost part of the Nilüfer unit in the Kozak range (Kaya and Mostler, 1992), indicating a middle Triassic and earlier age for the sequence. That and the unconformably overlying Liassic clastic rocks indicate that the regional metamorphism of the Nilüfer unit occurred in the late Triassic.

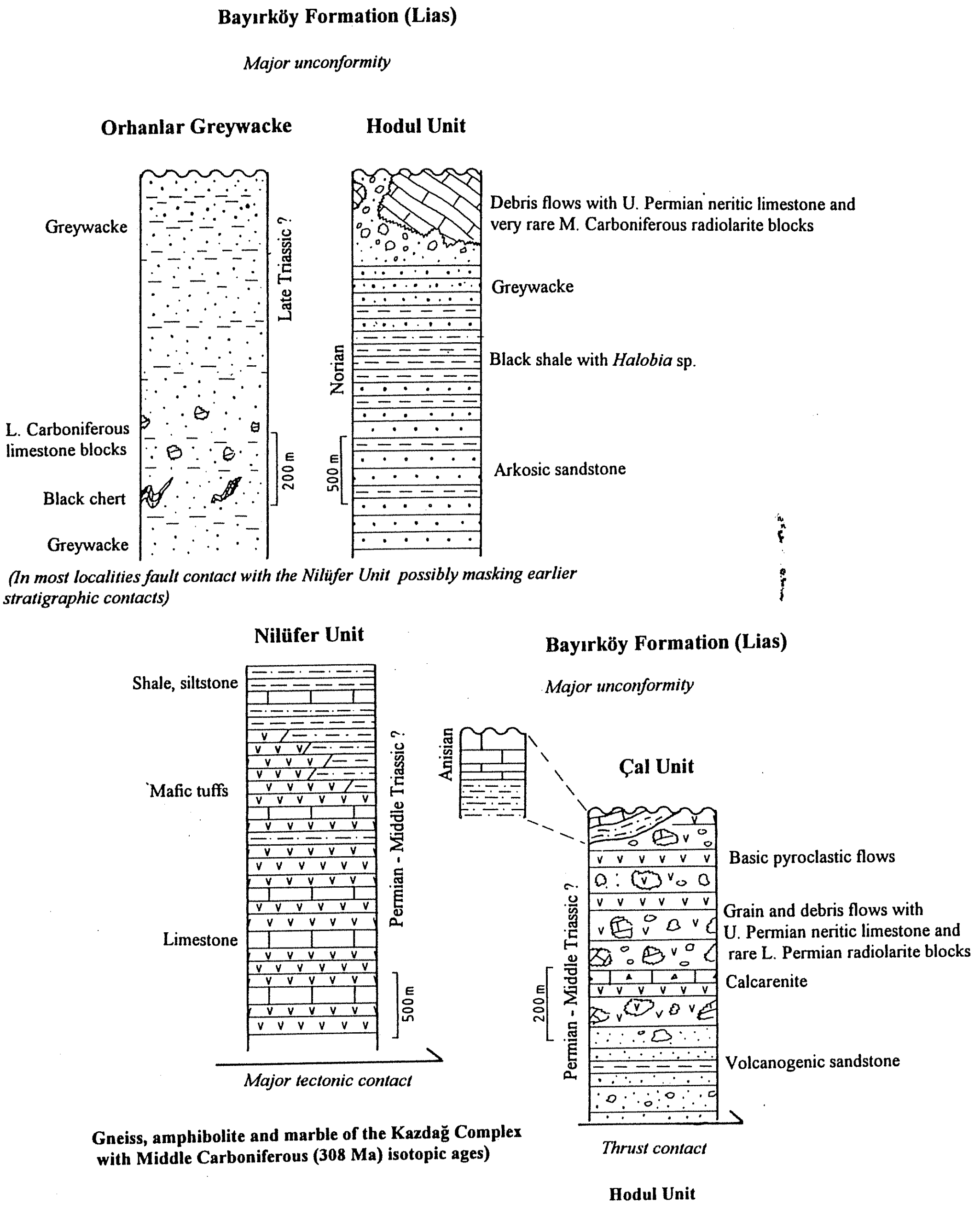


Figure 19.9. Generalized stratigraphic sections for the units of the Karakaya Complex. The thicknesses are highly approximate.

The volcaniclastic rocks make up more than 80% of the Nilüfer unit. Such thick and laterally extensive volcaniclastic sequences characterize sedimentary basins flanking active island-arcs (e.g., Dickinson and Seely, 1979). The generally fine-grain sizes of the volcaniclastic rocks, the fine laminations observed in some interstratified carbonates, and the apparent absence of intrusive magmatic rocks, such as dikes and sills, in the sequence suggest a depositional environment away from the active volcanic arc and in an intra-arc or fore-arc setting. The presence of rare eclogite and serpentinite lenses, probably tectonically introduced from an underlying accretionary complex, favors a fore-arc setting for the Nilüfer unit. The carbonate intervals represent pelagic sedimentary deposition during the volcanic quiescence. Similar volcaniclastic sequences interstratified with carbonates have been described for Permian (Houghton and Landis, 1989; Miller, 1989) and Recent (e.g., Hathway, 1994) volcanic arcs. One apparent anomaly in this picture is the alkaline nature of some of the basalts of the Nilüfer unit, as inferred from the common presence of Ti-augite and kaersutite in the flows and pyroclastic rocks, and from geochemical analysis (Pickett et al., 1993). Detailed geochemistry studies of similar basaltic sequences from the Ankara region farther east have also indicated that the alkali basalts are the predominant lava type (Floyd, 1993). Such alkali basalts are generally regarded as being typical of intraplate oceanic islands and seamounts. However, the Nilüfer unit outcrops for more than 1,000 km in the Sakarya zone with little change in its lithology and stratigraphy (Okay, 1984; Okay et al., 1991). A seamount origin for such a laterally extensive unit is implausible. On the other hand, it is increasingly recognized that alkaline magmas can occur in a variety of settings, including in intra-oceanic and continental-margin arcs, especially during the subduction of young, hot oceanic lithosphere. For example, the majority of basalts from the Cascade arc in Washington are geochemically indistinguishable from ocean-island basalts (Leeman et al., 1990).

Çal unit

The Çal unit consists dominantly of mafic volcanic flows and pyroclastic rocks, sheetlike debris-flow conglomerates, and volcanogenic sandstone and shale (Figure 19.9). It has steeply dipping fault contacts, mostly of Miocene age, with the other Karakaya Complex units, except that southeast of Çan, it lies, with a thrust contact, over the late Triassic clastic rocks (Figure 19.4). The Çal unit is unconformably overlain by late Liassic basal conglomerates (Okay et al., 1991).

The debris-flow conglomerates, which account for the bulk of the Çal unit, comprise poorly sorted Upper Permian shallow-water limestone clasts in a mafic volcanic or volcanogenic sandstone matrix. The Upper Permian limestones with fusulinids range from a few millimeters to a maximum of a few hundred meters in size. Well-bedded calciturbidites with transported Upper Permian limestone clasts, pelagic limestone, radiolarian chert, and Middle Triassic shallow-water limestone also occur in minor amounts in the Çal unit (Okay et al., 1991). Like most of the Karakaya Complex units, the Çal unit has a highly disrupted internal structure that ranges from broken formation to mélange. In most cases it is not clear whether the more competent lithologies are exotic blocks or represent an original part of the now-disrupted stratigraphic sequence. A "block" of radiolarian chert (a few meters in diameter) in siliceous shales from southeast of Çan (Figure 19.4) yielded early Permian (Sakmarian–Artinskien) radiolarian fauna of *Latentibifistula* cf. *triacanthophora, Holdisphaera* sp., *Praedeflandrella* sp., and *Copicyntra* sp. (Okay and Mostler, 1994), providing the first evidence of pelagic Permian facies in Turkey.

The close intermingling of mafic volcanic rocks and Upper Permian limestone clasts suggests that the limestone deposition was penecontemporaneous with the mafic volcanism, indicating an Upper Permian age for part of the sequence. The Middle Triassic limestones may represent an interval of carbonate deposition following cessation of the volcanism. The Liassic clastic rocks that unconformably overlie the Çal unit constrain its age to the Permo–Triassic.

The dominance of mafic pyroclastic rocks and debris-flow deposits in the Çal unit suggests deposition on the flanks and aprons of a mature oceanic seamount or an island-arc that was partly capped by Upper Permian limestone (Pickett et al., 1993). Similar volcanic and volcaniclastic sequences associated with debris-flow conglomerates with Permian limestone clasts have been described from the eastern Klamath terrane in California and are interpreted as having been deposited on the flanks of an early Permian island-arc (e.g., Watkins, 1993). The flanks and aprons of the ancient and present-day seamounts are characterized by the preponderance of debris-flow deposits, slumps, and pyroclastic rocks (e.g., Staudigel and Schmincke, 1984; Moore et al., 1989). These Permo–Triassic oceanic deposits were incorporated during the late Triassic into an accretionary complex that is now represented by the Çal unit.

Karakaya clastics: Hodul unit and Orhanlar graywacke

The oceanic volcanic-arc rocks of the Nilüfer unit are associated with highly disrupted, turbiditic clastic sequences several kilometers thick, mostly representing Triassic accretionary complexes. The contacts between the two are almost always steeply dipping faults, except that in the eastern part of the Kozak range the meta-volcanic rocks of the Nilüfer unit are stratigraphically overlain by late Triassic quartzofeldspathic sandstones and shales (Figures 19.4 and 19.5C). The coherence of the clastic sequences has largely been destroyed, and they range from broken formations to mélange; locally they show dynamo-thermal metamorphism and cleavage development. Two major types of clastic sequences can be distinguished in the Karakaya Complex. One is a quartzofeldspathic sandstone-shale sequence (Hodul unit), with a continental granitic source, and the other is a graywacke-shale sequence (Orhanlar graywacke) (Figure 19.9) (Okay et al., 1991).

The quartzofeldspathic clastic sequence ranges from proximal to distal turbidites and in regions near the İzmir–Ankara suture passes up to extensive late Triassic (Norian) debris flows with exotic Upper Permian limestone blocks in a graywacke matrix (Figure 19.9). This olistostrome belt can be traced from the mainland to the island of Lesbos in the Aegean Sea, where a disrupted graywacke-shale sequence with Lower Carboniferous and Permian limestone blocks has been described (Figure 19.4) (Hecht, 1972). The neritic massive to thickly bedded Upper Permian limestone blocks, which can reach up to several kilometers in scale, make up over 95% of the olistoliths and are characterized by rich fusulinid faunas (Okay et al., 1991). Smaller blocks of Carboniferous limestone and fine-grained aphyric mafic volcanic and pelagic sedimentary rocks also occur in the olistostromes. A 2-m-scale block of intercalated red pelagic limestone and radiolarian chert in the graywackes northeast of Balya has yielded middle Carboniferous (Bashkirian) conodonts: *Idiognathoides* cf. *optimus, Ozarkodina* sp., and *Hindeodus* sp. (Okay and Mostler, 1994).

The clastic rocks of the Orhanlar graywacke are homogeneous graywackes, with very poorly sorted angular quartz, plagioclase, opaques, lydite, radiolarian chert, mafic volcanic rock, and phyllite fragments in an argillaceous matrix. They contain rare, exotic blocks, up to a few meters thick, of Lower Carboniferous dark limestone rich in corals, brachiopods, and foraminifera (Okay et al., 1991). Southeast of Mustafakemalpaşa, the

Orhanlar graywacke passes laterally into olistostromes with several-hundred-meter-thick Upper Permian limestone blocks and resembles the upper parts of the Hodul unit (Figure 19.4).

The olistostrome facies dominated by Upper Permian limestone blocks form a northeast-trending belt 175 km long and 10 km wide that extends from the island of Lesbos to Manyas, immediately northwest of the Izmir–Ankara suture (Figure 19.4). The olistostromes were deposited during the late Triassic, as indicated by the Norian bivalves, brachiopods, corals, and cephalopods in the siltstone sequences under the olistrostromes (Okay et al., 1991). Triassic clastic rocks farther west in the Biga Peninsula contain only minor debris flows, with small and rare Upper Permian limestone fragments. This suggests that the blocks were derived from a thrust slice of Upper Paleozoic limestone that was approaching from the southeast, from the direction of the Anatolide–Tauride block. This is compatible with the late Permian Tethyan paleogeography, which was characterized by neritic carbonate deposition on the Gondwana margin in the south and terrigenous clastic sedimentation and non-deposition on the Laurasian margin in the north (Argyriadis, 1975). However, no vestiges of that thrust slice are present to the southeast of the olistostrome belt. The Permian limestone and clastic rocks north of Ezine, forming the only coherent large Permian sequence in northwestern Anatolia, could be preserved remnants of that once very extensive thrust sheet. The absence of Lower and Middle Triassic blocks in the Upper Triassic olistostromes suggests that the Permian limestone thrust sheet was an area of non-deposition during the Triassic.

Jurassic and younger cover rocks of the Sakarya zone

Following the latest Triassic Karakaya orogeny, molasse-type continental to shallow-marine Liassic clastic rocks were deposited over the entire western Sakarya zone (e.g., Altıner et al., 1991). They lie unconformably over various Karakaya Complex units, as well as over granitic rocks (Figure 19.4). The Liassic clastic rocks are disconformably succeeded by neritic Middle Jurassic to uppermost Jurassic carbonates, which in turn are unconformably overlain by Albian–Cenomanian pelagic limestones (Figure 19.3). In northwestern Turkey, in the region shown in Figure 19.4, the rest of the Cretaceous and Paleocene sedimentary rocks are missing, and Middle Eocene neritic limestones unconformably overlie units as old as the Karakaya Complex (Figure 19.4) (Siyako, Bürhan, and Okay, 1989). The late Cretaceous–Eocene was the period of southward ophiolite emplacement from the Intra-Pontide suture. However, farther east, in the region east of Bursa, the Middle Cretaceous pelagic limestones are succeeded by an Upper Cretaceous tuffaceous flysch sequence, more than 1,000 m thick, with serpentinite, blueschist, and Jurassic limestone olistoliths, that progresses up to continental-to-fluviatile Paleocene clastic rocks (Saner, 1980). The first compressive movements were in the Paleocene, when Jurassic limestones were thrust over the Paleocene terrigenous sandstones (Yılmaz, 1981). The progression from flysch to molasse sedimentation and the subsequent thrusting reflect the Paleocene continent–continent collision between the Sakarya zone and the Anatolide–Tauride block.

Izmir–Ankara suture and related oceanic rocks

The Izmir–Ankara suture, generally regarded as the main Neo-Tethyan suture in western Turkey, forms the southern boundary of the Sakarya zone. It consists of two segments, with different features (Figures 19.1 and 19.4). The northeast-trending western segment between İzmir and Balıkesir juxtaposes the Sakarya zone against a latest Cretaceous–Paleocene flysch unit containing very large Mesozoic limestone blocks, the Bornova flysch zone (Okay and Siyako, 1993). The contact between these two zones is largely covered by Neogene deposits, and only in the region north of Kepsut can the Upper Cretaceous–Paleocene flysch be seen tectonically to overlie the Karakaya Complex along a Paleocene back-thrust (Figures 19.4 and 19.5E). The east-trending eastern segment, marked by a post-Miocene strike-slip fault, constitutes the boundary between a major Cretaceous blueschist-ophiolite belt (the Tavşanlı zone) and the Sakarya zone. Large bodies of ultramafic rock and volcano-sedimentary complex were emplaced during the Cretaceous from this segment southward over the Anatolide–Tauride block. Some of these peridotite thrust slices are now located in the Lycian nappes 350 km south of the suture (Figure 19.1). The volcano-sedimentary complex, like its counterpart in the Intra-Pontide suture zone, consists of mafic volcanic and pyroclastic flows, radiolarian chert, pelagic shale, serpentinite, and rare Triassic and Cretaceous pelagic limestones. These lithologies form imbricate thrust slices and represent sediment-starved oceanic accretionary prisms. The mafic volcanic rocks in the volcano-sedimentary complex locally show incipient blueschist-facies metamorphism, with lawsonite, aragonite, and sodic pyroxene in the veins and amygdales of the mafic volcanic rocks, which otherwise retain their magmatic texture and mineralogy (Okay, 1986). Large peridotite bodies lie tectonically over the volcano-sedimentary complex along sub-horizontal faults. Isolated garnet-amphibolite slices at the base of the Burhan peridotite south of Bursa (Figures 19.4 and 19.5F), similar to the metamorphic soles described from the bases of several ophiolites (e.g., Spray et al., 1984), indicate Albian (101.1 ± 3.8 Ma) $^{40}Ar/^{39}Ar$ laser-spot isochron ages (Harris, Kelley, and Okay, 1994). This gives the age for the initial slicing of the oceanic crust prior to its emplacement over the passive continental margin of the Anatolide–Tauride block.

Bornova flysch zone: a foundered Mesozoic carbonate platform

The Bornova flysch zone is a 50–90-km-wide zone of chaotically deformed Upper Maastrichtian–Paleocene graywacke and shale, with Mesozoic neritic limestone and rare small blocks of mafic volcanic rock, chert, pelagic shale, and peridotite (Okay and Siyako, 1993). The Mesozoic limestone blocks can reach 10 km or more in scale. The proportions of mafic volcanic and pelagic sedimentary rocks in the flysch increase eastward, and the flysch passes laterally to the volcano-sedimentary complex in the region of Kepsut (Figures 19.4 and 19.5E). In the east, the Bornova flysch zone is in contact with the Menderes massif along post-Eocene normal faults.

The Bornova flysch zone was formed by the rapid foundering and destruction of a Mesozoic carbonate platform during the Maastrichtian–early Paleocene. Large sections of relatively intact carbonate platform are exposed on the island of Chios and on the adjacent Karaburun Peninsula (Figure 19.1), where the stratigraphy is also most nearly complete (Besenecker et al., 1968; Brinkmann et al., 1972; Erdoğan et al., 1990). In the Karaburun Peninsula, Carboniferous neritic limestones at least 300 m thick are unconformably overlain by a very heterogeneous, Lower–Middle Triassic sequence (>2,000 m thick) of pelagic cherty limestone, radiolarian chert, mafic lava, tuff, sandstone, mudstone, and debris flows with Carboniferous limestone blocks (Erdoğan et al., 1990). This Scythian–Anisian sequence, which formed during the rifting episode of the Neo-Tethyan Vardar ocean, is overlain by Ladinian–Albian platform carbonates more than 4 km thick (Figure 19.3). The ages of the platform carbonates range up to Turonian in the limestone blocks farther north (Poisson and Sahincı, 1988). Upper Cretaceous red pelagic limestones, representing the initial foundering of the carbonate platform, unconformably overlie the platform carbonates, cutting down to as low as the Norian limestones. The ages of the pelagic limestones change progressively from Santonian–Late Campanian in the Savaştepe region in the north to Campanian–Middle Maastrichtian in the Karaburun Peninsula in the south, reflecting the progressive southwestward foundering of the carbonate platform (Okay and Siyako, 1993). The pelagic limestones originally must have been overlain by flysch sediments; however, because of later strong deformation, the carbonates now occur as blocks in the sheared graywacke and shale. The undeformed Lower to Middle Eocene neritic limestones, which are transgressive over the flysch and the blocks north of Akhisar (Figure 19.4), constrain the timing of the breakup of the carbonate platform to the late Paleocene–early Eocene. It is noteworthy that although the Bornova flysch zone must have been located at a Mesozoic continental margin, Mesozoic continental-slope and rise deposits are rare in this zone.

Tavşanlı zone: a subducted passive continental margin

The Tavşanlı zone is a 50–60-km-wide east-trending belt of regionally metamorphosed blueschists tectonically overlain by volcano-sedimentary complex and large peridotite slabs (Okay, 1986). Undeformed Lower to Middle Eocene plutonic rocks (48-Ma and 53-Ma $^{40}Ar/^{39}Ar$ isochron ages), the Orhaneli and Topuk granodiorites, intrude the blueschists and the overlying peridotite (Figures 19.4 and 19.5E). (Harris et al., 1994). The blueschists consist of basal metaclastics more than 1,000 m thick, with jadeite, chloritoid, lawsonite, and glaucophane overlain by several kilometers of marbles that progress up to a meta-basite, meta-chert, and meta-shale, sequence, again with well-preserved blueschist-facies minerals. On the basis of broad stratigraphic comparisons with Tauride–Anatolide units, the time of deposition of the blueschist metaclastics probably was the Paleozoic, and that of the marbles was Mesozoic (Figure 19.3). The time of the blueschist metamorphism, based on phengite $^{40}Ar/^{39}Ar$ dating was Turonian–Coniacian (88.4 ± 0.5 Ma) (Okay and Kelley, 1994).

The blueschists of the Tavşanlı zone represent the subducted north-facing passive continental margin of the Anatolide–Tauride block and show patterns of tectonic evolution similar to those seen in Oman (e.g., Goffe et al., 1988; Searle et al., 1994). They were partly exposed or were at high crustal levels by the latest Cretaceous, prior to the continent–continent collision, as evidenced by blueschist detritus in latest Cretaceous clastic sequences in the Sakarya and Afyon zones to the south and north of the Tavşanlı zone, respectively. The continental collision between the Sakarya zone and the Anatolide–Tauride block occurred in the Paleocene, as deduced from the transition from flysch to molasse sedimentation in the Sakarya zone during the Paleocene, and the Paleocene southward thrusting of the Tavşanlı zone over the Afyon zone. The continuing compression in the post-continental-collision stage in western Turkey is characterized by southward-younging thrusts, from the Paleocene in the Afyon zone to the early Miocene in the Lycian nappes, as is the case in other major continent–continent collision belts, such as the Himalaya.

Afyon zone and Menderes massif: Alpide metamorphosed shelf sequences

The Afyon zone has stratigraphy similar to that of the blueschists in the Tavşanlı zone, and because of the low-grade greenschist-facies regional metamorphism, the sequence can be dated paleontologically (Figure 19.3) (Özcan et al., 1988; Göncüoğlu et al., 1992). Permo–Carboniferous clastic rocks, limestones, and minor tuffs are the lowest formations exposed in the Afyon zone; they progress up to Lower Triassic shallow-water clastics and dolomites, which are succeeded by Middle Triassic–Campanian platform carbonates, overlain by the Lower Maastrichtian pelagic micrites, radiolarian cherts, and siliceous shales (Figure 19.3). Those are followed by an olistostrome unit of late Maastrichtian–early Paleocene age more than 3 km thick, with ophiolite, blueschist, radiolarite, and limestone blocks tectonically overlain by peridotite and gabbro thrust sheets (Göncüoğlu et al., 1992). That deformed and metamorphosed sequence is unconformably overlain by Upper Paleocene–Lower Eocene sandstone and marine limestones that constrain the timing of the southward thrusting of the Tavşanlı zone over the Afyon zone to the late Maastrichtian–early Paleocene.

The Menderes massif is a dome-shaped metamorphic complex located to the south of the Afyon zone and east of the Bornova flysch zone (Figure 19.1). It shows essentially the same stratigraphy as the Afyon zone (Figure 19.3), but is distinguished in that the transition to pelagic carbonate sedimentation occurred in the Paleocene, and the olistostrome facies are of late Paleocene–early Eocene age (Dürr, 1975; Çağlayan et al., 1980; Şengör, Satır, and Akkök, 1984b). Because of a major Oligocene up-doming, it also exposes a Precambrian gneissic basement in its core (Satır and Friedrichsen, 1986). The middle Eocene Barrovian-type amphibolite to greenschist-facies regional metamorphism shows a gradual decrease from the core upward to the tectonically overlying Lycian nappes that were emplaced from the north during the middle Eocene.

Paleo-Tethyan evolution

The strong effects of the later Alpide deformations, the probable but unconfirmed lateral movement of the Sakarya zone during the Paleozoic, and the paucity of radiometric data make it impossible at present to formulate a convincing model for the Paleo-Tethyan evolution in northwestern Turkey and adjoining areas. However, the data that have been presented place constraints on our estimates of the location and age span of the Paleo-Tethys and its relationship to the Neo-Tethys (Figure 19.10).

Location of the Paleo-Tethys

The Paleo-Tethys was initially believed to have been to the north of the Istanbul zone (Şengör et al., 1980, 1984a). However, recent evidence demonstrating the continuity of the Istanbul zone and Laurasia until the late Cretaceous opening of the Black Sea (Okay et al., 1994) shows that that could not have been the case. On the other hand, Robertson and Dixon (1984) and Okay (1989) have placed the Paleo-Tethys between the Istanbul and Sakarya zones and have left the Sakarya zone as part of the Gondwana margin, in continuity with the Anatolide–Tauride block. However, the absence of late Triassic deformation, Hercynian metamorphic basement, and Paleozoic granites in the Anatolide–Tauride block indicates that it could not have been contiguous with the Sakarya zone. We would locate the Paleo-Tethys to the south of the Sakarya zone, leaving the Paleozoic basement of the Sakarya zone in possible continuity with the Moesia/Istanbul and Strandja zones (Figure 19.10A), all representing a post-Hercynian continental area along the southern margin of Laurasia.

Age and subduction history of the Paleo-Tethys and its relationship to the Neo-Tethys

Middle Carboniferous and Lower Permian radiolarian cherts from the Karakaya Complex in northwestern Turkey indicate the presence of a pelagic basin and possibly some oceanic crust by the late Paleozoic, which is compatible with the paleogeographic reconstructions showing a large oceanic area in the region of northwestern Turkey (Smith et al., 1981; Scotese and Golonka, 1992). The ages of the volcanic arc and the associated clastics suggest that subduction of the Paleo-Tethys probably had begun by the Permian. The northwestward emplacement of the late

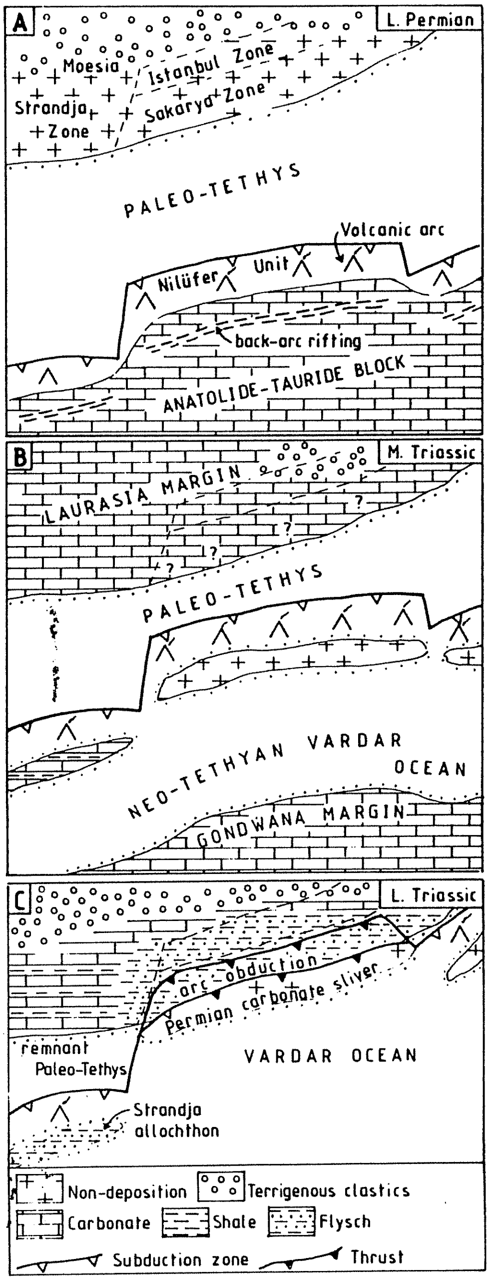


Figure 19.10. Schematic paleogeographic diagrams illustrating a possible Paleo-Tethyan tectonic evolution for northwestern Turkey. (A) Late Permian: The Istanbul, Sakarya, and Strandja zones form part of the post-Hercynian continental margin of Laurasia, which is characterized by non-deposition or terrigenous clastic sedimentation. Southward subduction of the Paleo-Tethys resulted in back-arc rifting in the Anatolide–Tauride block, where carbonate sedimentation prevailed during the late Permian. The fore-arc and accretionary complex of the subduction zone are represented by the units of the Karakaya Complex. (B) Middle Triassic: The back-arc rift has developed in the Neo-Tethyan Vardar ocean and has left a small continental sliver capped by Permian carbonates in the north adjacent to the volcanic arc. (C) Late Triassic: The volcanic arc and the continental sliver with the Permian carbonates are obducted on the Laurasian continental margin. A remnant Paleo-Tethyan oceanic crust is present in the west, whose elimination will give rise to the mid-Jurassic deformation and metamorphism in the Strandja zone.

Paleozoic carbonate thrust sheet and the inferred early Triassic back-arc-related rifting in the Anatolide–Tauride block indicate a southeastward-dipping subduction (Figure 19.10A). The early Triassic rifting must have left a small sliver of continental-to-transitional crust, capped by Permian carbonates, to the northwest of the Vardar ocean (Figure 19.10B). The lack of late Triassic deformation in the Bornova flysch zone, where the massive Norian reefal limestones pass up without interruption into Rhaetian and Liassic shallow-water carbonates (Erdoğan, 1990; Erdoğan et al., 1990), indicates that by the late Triassic, that rift must have developed into the Neo-Tethyan Vardar ocean.

The Karakaya orogeny, reflecting the closure of the Paleo-Tethys, was caused by the collision of the oceanic volcanic arc and associated accretionary complexes with the Laurasian passive continental margin (Figure 19.10C). The final collision and obduction of the upper-crustal part of the arc over the Sakarya zone occurred in the latest Triassic. That was followed by the northwestward thrusting of the Paleozoic carbonate sheet over the obducted arc. In northwestern Turkey, all of the deformation and metamorphism (and thus the Paleo-Tethyan evolution) were over by the Liassic, when widespread molasse-type clastics were deposited over the Karakaya Complex. However, the closure of the Paleo-Tethys was slightly later in the east, in the central Pontides, where the sandstones that unconformably overlie the Karakaya Complex are of Dogger age (Tüysüz, 1990), as well as in the west, in the Strandja zone, where the deformation and metamorphism occurred in the middle Jurassic (Chatalov, 1988).

A direct continuity between the Paleo-Tethys and Neo-Tethys, as claimed by Robertson and Dixon (1984) and Robertson et al. (1991), is unlikely because of the absence of late Paleozoic pelagic sedimentary rocks in the continental-slope sequences of the Anatolide–Tauride block. The sedimentary rocks in the Lycian nappes, to the south of the Menderes massif, record the establishment of a Triassic–Jurassic continental slope to the north of the Anatolide–Tauride block.

Vardar ocean: switching obduction events

Data from the Karaburun Peninsula and the Lycian nappes indicate that the rifting that led to the development of the Neo-Tethyan Vardar ocean began in the early Triassic. Görür et al. (1983), on the basis of sedimentology of the Liassic clastic rocks in the Sakarya zone suggested an early Jurrassic opening of the Vardar ocean, with the implication that the Sakarya zone and the Anatolide–Tauride block were contiguous up to the Liassic. However, that is not compatible with the very different Triassic stratigraphic patterns shown by the two units (Figure 19.3) and the absence of latest Triassic deformation in the northern margin of the Anatolide–Tauride block. The rifting event reflected in the Liassic clastic rocks of the Sakarya zone probably was related to the opening of the Intra-Pontide ocean between the Sakarya and Strandja–Istanbul zones. The Intra-Pontide suture cannot be traced east of the Istanbul zone (Figure 19.1), suggesting that the Intra-Pontide ocean was an embayment of the Vardar ocean (Figure 19.2). This is compatible with the paleomagnetic data from the Jurassic limestones of the Sakarya zone, which indicate that the Sakarya zone was not far removed from the Laurasian margin during the late Jurassic (Evans and Hall, 1990). The pre-Eocene 90° clockwise rotation recorded in some Upper Jurassic limestones from the Sakarya zone (Evans and Hall, 1990) may suggest that the full-scale development of the Intra-Pontide ocean, resulting in the counterclockwise rotation of the Sakarya zone, occurred in the middle Jurassic or earlier (Figure 19.2).

The trend of the Vardar–Izmir–Ankara suture and the orientations of the tectonic zones to the south suggest that the Anatolide–Tauride–Pelagonian block had an irregular, north-facing passive continental margin (Figure 19.11A), which had a major influence on its geologic history. During the earliest Cretaceous, ophiolite and oceanic sedimentary rocks were emplaced over the Vardar and Pelagonian zones (e.g., Burchfiel, 1980; Jacobshagen, 1986; Robertson et al., 1991). Although the direction of the ophiolite emplacement is controversial, comparisons with the late Cretaceous events in the Anatolide–Tauride block suggest a southwestward emplacement from the Vardar ocean. That earliest Cretaceous ophiolite obduction, probably caused by partial subduction of the Pelagonian margin in an oceanic subduction zone, can be traced as far southeast as the Sporades Islands (Jacobshagen and Wallbrecher, 1984), but no evidence for an early Cretaceous orogenic event is present on the Karaburun Peninsula, where the carbonate sedimentation continues, without a break, from the Ladinian to the Albian (Erdoğan et al., 1990). This indicates that a transform fault located along the outer margin of the Bornova flysch zone, termed the Soma transform, has relayed the movement associated with the ophiolite obduction to a subduction zone north of the Anatolide–Tauride block (Figure 19.11A). An origin for the Bornova flysch zone in a Californian-type transform-fault margin would also provide an explanation for the scarcity of Mesozoic continental-rise and slope deposits in this zone.

The earliest Cretaceous ophiolite obduction in Greece was followed by exhumation and erosion of the Pelagonian zone and deposition of Cenomanian–Eocene sedimentary rocks that locally rest unconformably on the obducted ophiolites. As there is no evidence for an early Cretaceous collision in the Pelagonian zone (cf. Burchfiel, 1980), the oceanic subduction zone must have been relocated farther north during the late Cretaceous, when the orogenic activity jumped eastward to the Anatolide–Tauride block (Figure 19.11B).

The ages of the metamorphic soles and the sedimentary rocks underlying the ophiolites in the Tavşanlı zone point to a late Cretaceous (Senonian) ophiolite obduction over the Anatolide–Tauride block (Okay and Kelley, 1994; Harris et al., 1994). The northern margin of the platform, which was deeply subducted, underwent blueschist-facies metamorphism. One anomaly in this picture is the Bornova flysch zone which shows no metamorphism and in its western part shows no evidence of ophiolite obduction. That and its position oblique to the other Anatolide–Tauride tectonic zones (Figure 19.1) suggest that the late

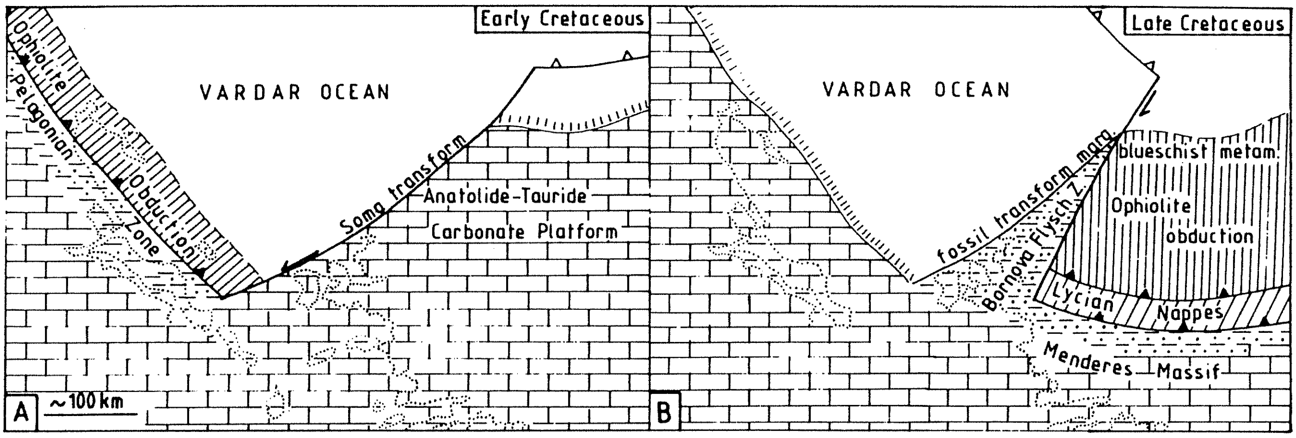


Figure 19.11. Cretaceous evolution of the northern margin of the Anatolide–Tauride–Pelagonian block. (A) Early Cretaceous: Ophiolite obduction occurs from the Vardar ocean during the earliest Cretaceous over the Vardar and Pelagonian zones in Greece. The obduction is relayed to an intra-oceanic subduction zone north of the Anatolide–Tauride block via the Soma transform fault, and neritic carbonate deposition continues uninterrupted on the Anatolide–Tauride block. (B) Late Cretaceous: In the early Senonian, a very large ophiolite slab is obducted over the Anatolide–Tauride carbonate platform. That leads to blueschist-facies metamorphism of its northern margin and southward transportation of the continental-slope and rise sequences (Lycian nappes). A triangular area in the western part of the Anatolide–Tauride carbonate platform (the Bornova flysch zone) is free from ophiolite obduction, but is engulfed in flysch sediments derived from the obducted ophiolite and continental-margin sequences. Carbonate sedimentation continues in the Pelagonian zone and farther south. The present Aegean coast is dotted for reference; however, the major late Mesozoic–Tertiary intracontinental shortening in Greece and western Turkey means that the Anatolide–Tauride–Pelagonian block was much wider in the Cretaceous than is indicated by its present-day size.

Cretaceous ophiolite obduction bypassed the Bornova flysch zone, leaving it as a major carbonate ramp bounded by a fossil transform margin in the northwest and by a major dextral shear fault in the southeast (Figure 19.11B). The progressive southwestward foundering of the Bornova carbonate platform, from the Santonian in the north to the Maastrichtian in the south, reflects the southward passage of the ophiolite and Lycian nappes over the Tavşanlı and Afyon zones, which also supplied detritus for the Paleocene flysch that engulfed the disrupted Mesozoic carbonate sequences.

The continental collision between the Sakarya zone and the Anatolide–Tauride block occurred during the Paleocene, and that was followed by progressive southward internal slicing of the shelf sequences of the Anatolide–Tauride block. The timing of the thrusts ranged from late Paleocene in the north to early Miocene in the south, when the Lycian nappes and the Menderes massif were thrust south over the most southerly shelf sequences (Figure 19.1).

The presence of Middle Paleocene pelagic limestones in the Intra-Pontide suture zone in Thrace (Okay and Tansel, 1994) and Middle Eocene shallow-water carbonates that unconformably overlie the accretionary complex in this region indicates an early Eocene closure of the Intra-Pontide ocean. Its closure was related to the late Cretaceous to early Eocene southward drift and eventual collision of the Istanbul zone with the Sakarya zone, concomitant with the opening of the oceanic western Black Sea basin (Figure 19.2). The collision of the Istanbul zone resulted in a southward shift of the western part of the Sakarya zone, probably compensated by increased oceanic subduction in the eastern Mediterranean area, and caused a major southward deflection of the Izmir–Ankara suture west of Ankara (Figure 19.1), where the north–south-trending suture is marked by a sinistral strike-slip fault (O. Tüysüz, personal communication). With that collision, the last vestiges of the oceanic crust were eliminated in western Turkey; however, a remnant Vardar ocean probably existed in the area of the present-day Aegean Sea until the middle Eocene, when its closure led to high-pressure low-temperature metamorphism in the Cyclades blueschist belt.

Conclusions

Evidence for a Paleozoic Tethyan ocean can be found in the Sakarya zone of northern Turkey in the Middle Carboniferous and Lower Permian radiolarian chert blocks in accretionary clastic complexes and in the volcanic-arc-related Permo–Triassic mafic pyroclastic rock, limestone, and debris-flow sequences. That Paleo-Tethyan episode ended by the latest Triassic with the emplacement of the volcanic arc over the passive margin of Laurasia. The Neo-Tethyan episode began with the early Triassic rifting of the Gondwana margin, which led to development of the Vardar ocean.

Different rock associations and sequences of events characterize the Paleo-Tethyan and Neo-Tethyan evolutions in northwestern Turkey. Intra-oceanic volcanic arcs, highly disrupted voluminous clastic sequences with exotic blocks, a scarcity of ophiolites, and the presence of steeply dipping faults distinguish the late Paleozoic–Triassic Paleo-Tethyan evolution of northwestern Turkey. In these aspects it shows close similarities to the rock associations and structures described for the Klamath terranes in the western Cordillera of North America. In contrast, the

Permian-to-present Neo-Tethyan evolution of Turkey has been typically Alpine in character and has been characterized by obduction of large, flat-lying ophiolite slices over passive-continental-margin sequences and by nappe tectonics involving passive-continental-margin assemblages and a protracted period of deformation. Active-margin sequences are rare and are represented by sediment-starved accretionary complexes made up of mafic volcanic rock, radiolarian chert, and siliceous shale.

The large oroclines that are distinctive features of the Alpide structures in western Turkey and the Aegean have pre- and post-collisional origins. The large offset of the Vardar–Izmir–Ankara suture between Izmir and Balikesir was an original irregularity, possibly a transform fault, in the northern margin of the Anatolide–Tauride–Pelagonian block that had a major influence on its Mesozoic orogenic history. It functioned as a hinge, such that ophiolite obduction and metamorphic events could occur at alternating times on the two sides. On the other hand, the large deflection of the Izmir–Ankara suture west of Ankara was produced during the Eocene continental collision between the Istanbul and Sakarya zones.

Strike-slip faults at high angles to the orogen, such as the West Black Sea fault or the Izmir–Balıkesir fault, played major roles in reorganization of the orogenic architecture. For example, the Istanbul zone, which was initially located outside the Alpide orogen and to the north of the Strandja zone, was translated during the Eocene, with the opening of the western Black Sea basin to the south of the Strandja zone inside the Alpide orogen. As a consequence, during the post-Eocene circum–Black Sea thrusting, it formed the hinterland, although its former lateral equivalent, the Moesian platform, was in a foreland position (Figure 19.1). Translations along orogen-parallel strike-slip faults, which undoubtedly must have had a major influence on the orogenic development, are at present impossible to quantify.

Acknowledgments

We thank the Turkish Petroleum Company and especially the late Ozan Sungurlu, former head of its exploration division, for supporting our field work in northwestern Turkey during the past 10 years, and we thank TÜBITAK for funding over the past 2 years (TÜBITAK–Glotek grant). The first author thanks the Alexander von Humboldt Foundation for a grant during preparation of the manuscript. For helpful discussions over many years, we thank the late Ozan Sungurlu, Celal Şengör, Leopold Krystyn, Okan Tekeli, Cemal Göncuoglu, Naci Görür, Kerem Ali Bürkan, Sancar Kasar, and Demir Altıner. The manuscript benefited from constructive reviews by Lothar Ratschbacher and An Yin.

References

Altiner, D., Kocyigit, A., Farinacci, A., Nicosia, U., and Conti, M. A. 1991. Jurassic, Lower Cretaceous stratigraphy and paleogeographic evolution of the southern part of north-western Anatolia. *Geologica Romana* 28:13–80.

Argyriadis, I. 1975. Mesogee permienne, chaine hercynienne et cassure tethysienne. *Bull. Soc. Géol. France* 17:56–66.

Aydın, Y. 1974. Etude petrographique et geochimique de la partie centrale du Massif d'Istranca (Turquie). Ph.D. thesis, Université de Nancy.

Aydın, Y. 1982. Geology of the Yildiz mountains (Strandja) (in Turkish). Habilitation thesis, Istanbul Technical University.

Bergougnan, H., and Fourquin, C. 1982. Remnants of a pre-Late Jurassic ocean in northern Turkey: fragments of Permian–Triassic Neo-Tethys? Discussion and reply by Şengör, A. M. C., Yılmaz, Y., and Ketin, İ. *Bull. Geol. Soc. Am.* 93:929–36.

Besenecker, H., Dürr, S., Herget, G., Jacobshagen, V., Kauffmann, G., Lüdtke, G., Roth, W., and Tietze, K. W. 1968. Geologie von Chios (Agais). *Geologica et Palaeontologica* 2:121–50.

Bingöl, E. 1969. Geology of the central and southeastern sections of the Kazdag Massif (in Turkish). *Bull. Mineral Res. Exp. Inst. Turkey* 72:102–23.

Bingöl, E. 1971. Classification of age determination methods and an application of Rb-Sr and K-Ar methods in Kazdag (in Turkish). *Türkiye Jeoloji Kurumu Bülteni* 14:1–16.

Boccaletti, M., Gocev, P., and Manetti, P. 1974. Mesozoic isopic zones in the Black Sea region. *Bull. Soc. Geol. Italiana* 93:547–65.

Borsi, S., Ferrara, G., and Mercier, J. 1964. Determination de l'age des series metamorphiques de Massif Serbomacedonien au Nord-Est de Thessalonique (Grece) par les methodes Rb/Sr et K/Ar. *Ann. Soc. Geol. du Nord* 84:223–5.

Brinkmann, R., Flügel, E., Jacobshagen, V., Lechner, H., Rendel, B., and Trick, P. 1972. Trias, Jura und Unterkreide der Halbinsel Karaburun (West-Anatolien). *Geologica et Palaeontologica* 6:139–90.

Burchfiel, B. C. 1980. Eastern European Alpine system and the Carpathian orocline as an example of collision tectonics. *Tectonophysics* 63:31–61.

Çağlayan, A., Öztürk, E. M., Öztürk, A., Sav, H., and Akat, U. 1980. New data on the southern part of the Menderes massif and a structural interpretation (in Turkish). *Jeoloji Mühendisligi* 10:9–17.

Chatalov, G. A. 1988. Recent developments in the geology of the Strandzha Zone in Bulgaria. *Bull. Tech. Univ. Istanbul* 41:433–65.

Chatalov, G. A. 1991. Triassic in Bulgaria – a review. *Bull. Tech. Univ. Istanbul* 44:103–35.

Cocherie, A., Guerrot, C., and Rossi, P. 1991. Single-zircon dating by step-wise Pb evaporation: comparison with other geochronological techniques applied to the Hercynian granites of Corsica, France. *Chem. Geol.* 101:131–41.

Dercourt, J., et al. 1986. Geological evolution of the Tethys belt from the Atlantic to the Pamirs since the Lias. *Tectonophysics* 123:241–315.

Dickinson, W. R., and Seely, D. R. 1979. Structure and stratigraphy of the fore-arc regions. *Am. Assoc. Pet. Geol. Bull.* 63:2–31.

Dürr, S. 1975. Über alter und geotektonische Stellung des Menderes-Kristallins, SW-Anatolien und seine Aequivalente in der mittleren Aegaeis. Thesis, University of Marburg.

Erdoğan, B. 1990. Stratigraphic features and tectonic evolution of the Izmir–Ankara Zone in the region between Izmir and Seferihisar (in Turkish). *Türkiye Petrol Jeologları Dernegi Bülteni* 2:1–20.

Erdoğan, B., Altiner, D., Güngor, T., and Sacit, O. 1990.

Stratigraphy of the Karaburun Peninsula (in Turkish). *Bull. Mineral Res. Exp. Inst. Turkey* 111:1–24.
Evans, I., and Hall, S. A. 1990. Paleomagnetic constraints on the tectonic evolution of the Sakarya continent, northwestern Anatolia. *Tectonophysics* 182:357–72.
Evans, I., Hall, S. A., Saribudak, M. A., and Aykol, A. 1991. Preliminary palaeomagnetic results from Palaeozoic rocks of the İstanbul–Zonguldak region, N.W. Turkey. *Bull. Tech. Univ. Istanbul* 44:165–90.
Floyd, P. A. 1993. Geochemical discrimination and petrogenesis of alkalic basalt sequences in part of the Ankara melange, central Turkey. *J. Geol. Soc. London* 150:541–50.
Ganev, M. 1974. Stand der Kenntnisse über die Stratigraphie der Trias Bulgariens. In *Die Stratigraphie der alpin-mediterranean Trias,* ed. H. Zapfe, pp. 93–6. Wein: Springer.
Goffe, B., Michard, A., Kienast, J. R., and Le Mer, O. 1988. A case of obduction-related high-pressure, low-temperature metamorphism in upper crustal nappes, Arabian continental margin, Oman: *P–T* paths and kinematic interpretation. *Tectonophysics* 151:363–86.
Göncüoğlu, M. C., and Erendil, M. 1990. Pre–Late Cretaceous tectonic units of the Armutlu Peninsula. In *Proceedings of the 8th Petroleum Congress of Turkey,* p. 161–8.
Göncüoğlu, M. C., Özcan, A., Turhan, N., and Isik, A. 1992. Stratigraphy of the Kütahya region. In *Field Guide Book for the Symposium on the Geology of the Black Sea Region*, pp. 3–11. Ankara:
Görür, N., Şengör, A. M. C., Akkök, R., and Yılmaz, Y. 1983. Sedimentological data regarding the opening of the northern branch of the Neo-Tethys in the Pontides (in Turkish). *Türkiye Jeoloji Kurumu Bülteni* 26:11–20.
Haas, W. 1968, Das Alt-paläozoikum von Bithynien (Nordwest Türkei). *Neues Jb. Geol. Paläont. Abh.* 131:178–242.
Harris, N. B. W., Kelley, S. P., and Okay, A. I. 1994. Post-collision magmatism and tectonics in northwest Turkey. *Contrib. Mineral. Petrol.* 117:241–52.
Hathway, B. 1994. Sedimentation and volcanism in an Oligocene–Miocene intra-oceanic arc and fore-arc, southwestern Viti Levu, Fiji. *J. Geol. Soc. London* 151:499–514.
Hecht, J. 1972. Zur Geologie von Südost-Lesbos (Griechenland). *Zeitschrift der deutschen geologischen Gesellschaft* 123:423–32.
Houghton, B. F., and Landis, C. A. 1989. Sedimentation and volcanism in a Permian arc-related basin, southern New Zealand. *Bull. Volcanology* 51:433–50.
Jacobshagen, V. 1986. *Geologie von Griechenland.* Berlin: Gebrüder Borntraeger.
Jacobshagen, V., and Wallbrecher, E. 1984. Pre-Neogene nappe structure and metamorphism of the North Sporades and the southern Pelion peninsula. In *The Geological Evolution of the Eastern Mediterranean,* ed. J. E. Dixon and A. H. F. Robertson, pp. 591–602. Geological Society London special publication no. 17.
Kasar, S., and Okay, A. I. 1992. Geology of the region between Silivri Kıyıköy and Bosphorus (in Turkish). Unpublished report of the Turkish Petroleum Company, no. 3119.
Kaya, O., and Mostler, H. 1992. A Middle Triassic age for low-grade greenschist facies metamorphic sequence in Bergama (Izmir), western Turkey: the first paleontological age assignment and structural–stratigraphic implications. *Newsletter for Stratigraphy* 26:1–17.
Ketin, İ. 1966. Tectonic units of Anatolia (Asia Minor). *Bull. Mineral Res. Exp. Inst. Turkey* 66:23–34.
Kober, B. 1986. Whole grain evaporation for $^{207}Pb/^{206}Pb$-age-investigations on single zircons using double-filament thermal ion source. *Contrib. Mineral. Petrol.* 93:482–90.
Kober, B. 1987. Single zircon evaporation combined with Pb+ emitter bedding for $^{207}Pb/^{206}Pb$-age investigations using thermal ion mass spectrometry, and implications to zirconology. *Contrib. Mineral. Petrol.* 96:63–71.
Kopp, K., Pavoni, N., and Schindler, C. 1969. Geologie Thrakiens. V: Das Ergene Becken. *Beihefte der geologischen Jahrbuch* 76:1–136.
Kotanski, Z. 1978. The Caucasus, Crimea and their foreland (Scythian platform), Black Sea and the Caspian Sea. In *Geological Atlas of Alpine Europa and Adjoining Alpine Areas,* ed. M. Lemoine, pp. 545–76. Amsterdam: Elsevier.
Leeman, W. P., Smith, D. R., Hildreth, W., Palacz, Z., and Rogers, N. 1990. Compositional diversity of late Cenozoic basalts in a transect across the southern Washington Cascades: implications for subduction zone magmatism. *J. Geophys. Res.* 95:19,561–82.
LeFort, P. 1989. The Himalayan orogenic segment. In *Tectonic Evolution of the Tethyan Region*, ed. A. M. C. Şengör, pp. 289–386. Dordrecht: Kluwer.
Maluski, H., Costa, A., and Echtler, H. 1991. Late Variscan tectonic evolution by thinning of earlier thickened crust. An ^{39}Ar-^{40}Ar study of the Montagne Noire, southern Massif Central, France. *Lithos* 26:287–304.
Mattauer, M., Matte, P., Malavieille, J., Tapponier, P., Maluski, H., Qin, X. Z., Lun, L. Y., and Qin, T. Y. 1985. Tectonics of the Qinling Belt: build-up and evolution of eastern Asia. *Nature* 317:496–500.
Metzger, R., Bracke, G., and Satir, M. in press. Single-zircon dating of gneisses from the Kazdağ Massif (NW Turkey) by step-wise Pb-evaporation technique.
Miller, M, M, 1989. Intra-arc sedimentation and tectonism: late Paleozoic evolution of the eastern Klamath terrane, California. *Bull. Geol. Soc. Am.* 101:170–87.
Moore, J. G., Clague, D. A., Holcomb, R. T., Lipman, P. W., Normark, W. R., and Torresan, M. E. 1989. Prodigious submarine landslides on the Hawaiian Ridge. *J. Geophys. Res.* 94:17465–84.
Moore, W. J., McKee, E. H., and Akıncı, Ö. 1980. Chemistry and chronology of plutonic rocks in the Pontid mountains, northern Turkey. In *European Copper Deposits,* pp. 209–16. Belgrade:
Okay, A. I. 1984. The geology of the Agvanis metamorphic rocks and neighbouring formations. *Bull. Mineral Res. Exp. Inst. Turkey* 99/100:16–36.
Okay, A. I. 1986. High-pressure/low-temperature metamorphic rocks of Turkey. *Geol. Soc. Am. Mem.* 164:333–47.
Okay, A. I. 1989. Tectonic units and sutures in the Pontides, northern Turkey. In *Tectonic Evolution of the Tethyan Region*, ed. A. M. C. Şengör, pp. 109–13. Dordrecht: Kluwer.
Okay, A. I., and Kelley, S. 1994. Jadeite and chloritoid schists from northwest Turkey: tectonic setting, petrology and geochronology. *J. Metamorphic Geol.* 12:455–66.
Okay, A. I., and Mostler, H. 1994. Carboniferous and Permian radiolarite blocks in the Karakaya Complex in northwest Turkey. *Turkish J. Earth Sci.* 3:23–8.
Okay, A. I., Satır, M., and Metzger, R. in press. An Eocene core extensional complex in western Turkey.
Okay, A. I., Şengör, A. M. C., and Görür, N. 1994. Kinematic history of the opening of the Black Sea and its effect on the surrounding regions. *Geology* 22:267–70.
Okay, A. I., and Siyako, M. 1993. The new position of the Izmir–Ankara Neo-Tethyan suture between Izmir and

Balıkesir (in Turkish). In *Tectonics and Hydrocarbon Potential of Anatolia and Surrounding Regions, Proceedings of the Ozan Sungurlu Symposium, Ankara,* ed. S. Turgut, pp. 333–55.

Okay, A. I., Siyako, M., and Bürkan, K. A. 1991. Geology and tectonic evolution of the Biga Peninsula, northwestern Turkey. *Bull. Tech. Univ. İstanbul* 44:191–256.

Okay, A. I., and Tansel, I. 1994. New data on the upper age of the Intra-Pontide ocean from north of Şarköy in Thrace. *Bull. Mineral Res. Exp. Inst. Turkey* 114:21–24.

Özcan, A., Göncüoğlu, M. C., Turan, N., Uysal, S., Sentürk, K., and Isik, A. 1988. Late Paleozoic evolution of the Kütahya–Bolkardag belt. *Middle East Technical University J. Pure Appl. Sci.* 21:211–20.

Özdemir, U. 1981. Microfauna and biostratigraphy of the Tepeköy Triassic in the Kocaeli Peninsula. *Bull. Mineral Res. Exp. Inst. Turkey* 77:57–98.

Pickett, E., Robertson, A. H. F., Sharp, I. R., and Dixon, J. E. 1993. Seamount build-up and accretion processes in the late Paleozoic–early Mesozoic Tethys of northwest Turkey. *Terra Abstracts* 5:266.

Poisson, A., and Sahıncı, A. 1988. La serie mesozoique de Kemalpaşa et le flysch Paleocene d'Izmir au Nord-Quest du Menderes (Anatolie occidentale Turquie). Un jalon du microcontinent taurique. *C. R. Acad. Sci. Paris* 307: 1075–80.

Robertson, A. H. F., Clift, P. D., Degnan, P. J., and Jones, G. 1991. Palaeogeographic and palaeotectonic evolution of the Eastern Mediterranean Neotethys. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 87:289–343.

Robertson, A. H. F., and Dixon, J. E. 1984. Introduction: aspects of the geological evolution of the Eastern Mediterranean. In *The Geological Evolution of the Eastern Mediterranean,* ed. J. E. Dixon and A. H. F. Robertson. *Geol. Soc. London Spec. Pub.* 17:1–74.

Saner, S. 1980. Paleogeographic interpretation of the Jurassic and younger sediments of the Mudurnu–Göynük basin (in Turkish). *Türkiye Jeoloji Kurumu Bülteni* 23:39–52.

Sarıbudak, M., Sanver, M., and Ponat, E. 1989. Location of western Pontides, NW Turkey, during Triassic time: preliminary palaeomagnetic results. *Geophys. J.* 96:43–50.

Satır, M., and Friedrichsen, H. 1986. The origin and evolution of the Menderes Massif, W-Turkey: a rubidium/strontium and oxygen isotope study. *Geol. Rundschau* 75:703–15.

Scotese, C. R., and Golonka, J. 1992. *Paleogeographic Atlas.* PALEOMAP project, Department of Geology, University of Texas at Arlington.

Searle, M., Waters, D. J., Martin, H. N., and Rex, D. C. 1994. Structure and metamorphism of blueschist-eclogite facies rocks from the northeastern Oman Mountains. *J. Geol. Soc. London* 151:555–76.

Şengör, A. M. C. 1979. The North Anatolian transform fault: its age, offset and tectonic significance. *J. Geol. Soc. London* 136:269–82.

Şengör, A. M. C., Satır, M., and Akkök, R. 1984b. Timing of Tectonic events in the Menderes massif, western Turkey: implications for tectonic evolution and evidence for Pan-African basement in Turkey. *Tectonics* 3:693–707.

Şengör, A. M. C., and Yılmaz, Y. 1981. Tethyan evolution of Turkey: a plate tectonic approach. *Tectonophysics* 75: 181–241.

Şengör, A. M. C., Yılmaz, Y., and Ketin, İ. 1980. Remnants of a pre-late Jurassic ocean in northern Turkey: fragments of Permian–Triassic Paleo-Tethys? *Bull. Geol. Soc. Am.* 91:599–609.

Şengör, A. M. C., Yılmaz, Y., and Sungurlu, O. 1984a. Tectonics of the Mediterranean Cimmerides: nature and evolution of the western termination of Palaeo-Tethys. In *The Geological Evolution of the Eastern Mediterranean,* ed. J. E. Dixon and A. H. F. Robertson. *Geol. Soc. London Spec. Pub.* 17:77–112.

Şentürk, K., and Okay, A. I. 1984. Blueschists discovered east of Saros bay in Thrace. *Bull. Mineral Res. Exp. Inst. Turkey* 97/97:152–5.

Siyako, M., Bürkan, K. A., and Okay, A. I. 1989. Tertiary geology and hydrocarbon potential of the Biga and Gelibolu peninsulas (in Turkish). *Türkiye Petrol Jeologlari Derneği Bülteni* 1:183–99.

Smith, A. G. 1973. The so-called Tethyan ophiolites. In *Implications of Continental Drift to the Earth Sciences,* ed. D. H. Tarling and S. K. Runcorn, pp. 997–86. New York: Academic Press.

Smith, A. G., Hurley, A. M., and Briden, J. C. 1981. *Phanerozoic Paleocontinental World Maps.* Cambridge University Press.

Spray, J. G., Bebien, J., Rex, D. C., and Roddick, J. C. 1984. Age constraints on the igneous and metamorphic evolution of the Hellenic-Dinaric ophiolites. In *The Geological Evolution of the Eastern Mediterranean,* ed. J. E. Dixon and A. H. F. Robertson. *Geol. Soc. London Spec. Pub.* 17:619–27.

Stampfli, G., Marcoux, J., and Baud, A. 1991. Tethyan margins in space and time. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* 87:373–409.

Staudigel, H., and Schmincke, H. U. 1984. The Pliocene seamount series of La Palma/Canary Islands. *J. Geophys. Res.* 89:11,195–215.

Tekeli, O. 1981. Subduction complex of pre-Jurassic age, northern Anatolia, Turkey. *Geology* 9:68–72.

Tokay, M. 1952. Geologie de la region de Bartin (Zonguldak, Turquie de Nord). *Bull. Mineral Res. Exp. Inst. Turkey* 46/47:46–63.

Turgut, S., Türkaslan, M., and Perinçek, D. 1991. Evolution of the Thrace sedimentary basin and its hydrocarbon prospectivity. In *Generation, Accumulation, and Production of Europe's Hydrocarbons,* ed. A. M. Spencer, pp. 415–37. Oxford University Press.

Tüysüz, O. 1990. Tectonic evolution of a part of the Tethyside orogenic collage: the Kargi Massif, northern Turkey. *Tectonics* 9:141–60.

Ustaömer, T., and Robertson, A. H. F. 1994. Late Paleozoic marginal basin and subduction-accretion: the Paleotethyan Küre Complex, Central Pontides, northern Turkey. *J. Geol. Soc. London* 151:291–305.

Watkins, R. 1993. Permian carbonate platform development in an island-arc setting, eastern Klamath terrane, California. *J. Geol.* 101:659–66.

Yılmaz, Y. 1981. Tectonic evolution of the southern margin of the Sakarya continent (in Turkish). *Istanbul Yerbilimleri* 1:33–52.

Yılmaz, Y., Genç, S. C., Yigitbaş, E., Bozcu, M., and Yılmaz, K. 1994. Geological evolution of the late Mesozoic continental margin, NW Anatolia (in Turkish). In *Proceedings of the 10th Petroleum Congress of Turkey,* pp. 37–55. Istanbul.

Appendix: Analytical method for single-zircon stepwise-Pb-evaporation dating

The selected representative zircons typically were about 0.05 mm in width and 0.12 mm in length. Isotope measurements