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# The Seaway Connection between the Sea of Marmara and the Mediterranean: Tectonic Development of the Dardanelles

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**Abstract**

A basin consisting of alluvial-fan, fluvial clastics, and lacustrine sediments dating from the late middle Miocene was developed in the southern part of the Thrace Basin, Gelibolu Peninsula, and the northwestern part of the Biga Peninsula. This basin was controlled by a southeastern fault that lies parallel to the northwestern rim of the Biga Peninsula. Shallow-marine sediments that occur widely in the upper parts of the sedimentary sequence in the basin indicate an initial connection between the Sea of Marmara and the Mediterranean during the middle to late Pannonian, extending from southern Thrace to west of the Biga Peninsula.

The Saros-Ganos segment, forming a branch of the North Anatolian fault, developed as a positive flower structure in the late Miocene to early Pliocene. This structural high obstructed the seaway connection around Saros Bay. Clastics, originating from this positive feature, filled the Dardanelles Strait area of today.

A late middle Miocene fault that controlled the geometry of the basin from the eastern shoulder of the Biga Peninsula was reactivated as a right-lateral strike-slip fault during the late early Pliocene (Ezine-Sarköy fault). Synthetic faults associated with this major tectonic structure caused the opening of the Dardanelles. This tectonism resulted in the reconnection of the Mediterranean with the Sea of Marmara. The ages of foraminifer, ostracod, and nannoplankton assemblages, determined from wells in the Gulf of Izmit, indicate intermittent instead of continuous connection. The oldest Mediterranean connection was in late Pliocene–early Pleistocene time, the next connection occurred in the early to middle Pleistocene, and the final connection occurred during the late Pleistocene to Holocene.

## Introduction

THE SEAWAY CONNECTION between the Sea of Marmara and the Mediterranean has been investigated extensively. However, determination of when, where, and how this connection occurred has remained elusive. Most previous investigators (Penck, 1917; Yalçinlar, 1949; Ardel and Inandik, 1957; Erol, 1968; Sentürk and Karaköse, 1987) viewed the Dardanelles as a fluvial valley, and Önem (1974) proposed that it was a graben. According to Erol (1992), the Dardanelles is an epigenic fluvial valley developed under the control of Pliocene–early Pleistocene faulting, with the sea invading the valley during the late Pleistocene–Holocene. Stanley and Blanpied (1980) argued that the connection between the Sea of Marmara and the Mediterranean via the Dardanelles occurred during the last 12,000 to 9,500 years. The data from the surface sediments of the Gulf of Izmit show that the latest sea connection goes back at least 35,000 years (Çetin et al., 1995).

We analyzed Gulf of Izmit samples from nine boreholes that were drilled between the Hersek and Kaba promontories (Fig. 1A). Data obtained from the fossil assemblages indicate that three transgressions occurred in the region—late Pliocene–early Pleistocene, early to middle Pleistocene, and late Pleistocene–Holocene—during which Mediterranean water reached the Gulf of Izmit (Meriç, 1995).

The North Anatolian fault (NAF) extends between Karliova in the east and the

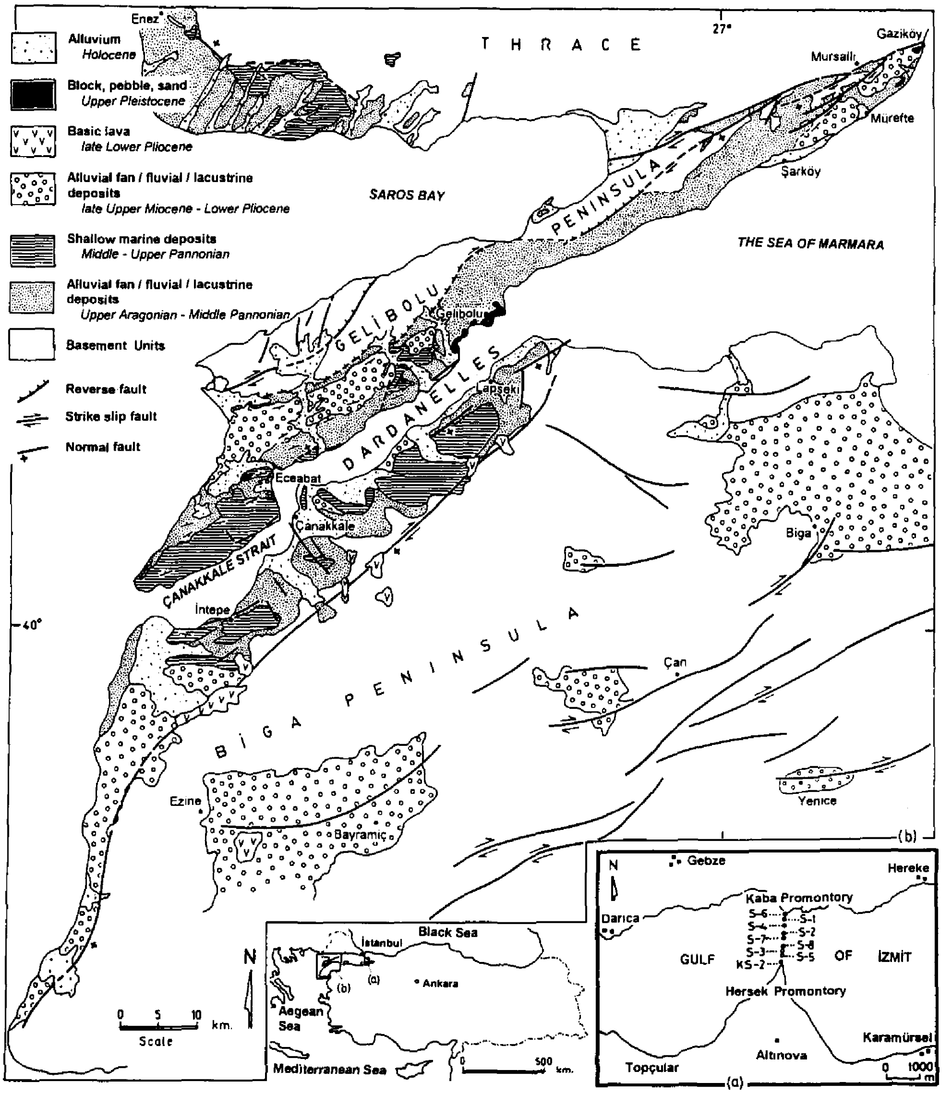


FIG. 1. A. Localities of wells in Izmit Bay. B. Simplified geological map of southern Thrace and the Gelibolu and Biga peninsulas (modified from Sentürk and Karaköse, 1987; Sümengen et al., 1987; Siyako et al., 1989; Erkal, 1991; Yaltirak, 1995a, 1995b).

Mudurnu Valley in the west. West of the Mudurnu Valley, it divides into three branches and displays a horsetail structure in northwestern Anatolia (Dewey and Sengör, 1979; Lyberis, 1984; Sengör et al., 1985; Barka, 1992). In the northern part of the Sea of Marmara, many pull-apart basins developed along the fault zone (Sengör et al., 1985; Barka and Kadinsky-Cade, 1988; Barka, 1992). The Neogene–Quaternary basins (Irrlitz, 1972; Barka and Hancock, 1984; Sengör et al., 1985; Barka and Gülen, 1988, 1989; Koçyigit, 1989) situated along the NAF zone were developed from the early middle Miocene and late Miocene to the Pliocene. The early middle Miocene deposits appear to be unrelated to the NAF (Barka, 1992). In the basins that were developed around releasing bends of the fault zone, sedimentation started during the late Miocene to early Pliocene. The ages of these basins range from Tortonian (Irrlitz, 1972) to lower Pleistocene.

It is obvious that the study area is a tectonically active area controlled by the NAF. For this reason, the investigation of neotectonic structural entities could help provide answers to problems such as which way(s) and how the seaway connections between the Sea of Marmara and the Mediterranean developed in the three aforementioned intervals, and if other connections existed apart from these. This paper describes the stratigraphic and structural entities of the region from the middle to late Miocene to the recent period, then discusses a possible connection scenario.

Units older than middle Miocene will be treated as basement units. Units developed in the basins that were affected directly by the NAF will be discussed under the heading of middle–upper Miocene/Quaternary units.

## Basement Units

The oldest unit in the Gelibolu Peninsula is the Upper Cretaceous–Paleocene ophiolitic mélange (Okay et al., 1990). It is unconformably overlain by the Ypresian–lower Lutetian (Siyako et al., 1989) regressive sequence (Önal, 1986; Sümengen et al., 1987). An upper Lutetian–upper Eocene transgressive sequence overlies these units. The late Eocene–late Oligocene epoch is represented by strata displaying a regressive character (Sümengen et al., 1987). The entire assemblage forms the basement of the middle–upper Miocene/Quaternary sequence in the Gelibolu Peninsula.

In the Biga Peninsula, three tectonic zones of pre-Tertiary age extend from northeast to southwest—a Permo-Carbonifereous metasedimentary sequence and an overlying ophiolite (the Ezine Zone), the Upper Cretaceous–Paleocene ophiolitic mélange (Çetmi Ophiolitic Mélange), and a gneiss-amphibolite-marble complex (Kazdag Group) overthrust by a Triassic blocky metavolcanic-sedimentary sequence (Karakaya Formation) (Sakarya Zone) (Okay et al., 1990). The Jurassic–Upper Cretaceous sequence overlies the Karakaya units. The middle Eocene neritic carbonates and the upper Eocene turbidites with andesitic tuff-lava intercalations form the lowermost parts of the Tertiary sequence. These are overlain by the lower–middle Miocene calcalkaline volcanic and lacustrine deposits. The entire sequence forms the basement of the middle–upper Miocene/Quaternary sequence in the Biga Peninsula.

## The Middle–Upper Miocene/Quaternary Sequence in the Biga Peninsula

The middle–upper Miocene/Pliocene sequence (Dardanelles Group) is represented by disordered blocky conglomerate-sandstone-siltstone at the base (Sariyar Member [Sentürk and Karaköse, 1987]; Gazhanedere Formation [Saltik, 1974]). They dip toward basement units on the east (Fig. 1B). Coarse-grained, poorly sorted material becomes finer grained and well sorted toward the west. Lateral and vertical transitions of the units are abrupt, and pebbles within channel fills are imbricated. The unit contains cut-and-fill structures, mudflow, mass-flow, sheet-flood, and stream-flood deposits. Thus, this unit—whose age is early Pannonian on the basis of stratigraphic position—indicates deposition in an alluvial fan and, less commonly, in a fluvial environment.

Toward the top, the sequence grades laterally and vertically into crossbedded, laminated sandstones with shale intercalations (Kirazli Formation) (Saltik, 1974) (Fig. 2). The age of the unit deposited in a nearshore environment is Tortonian–Sarmatian on the basis of vertebrates and other fossils (Erguvanli, 1957; Kopp et al., 1969; Sümengen et al., 1987) and middle to late Miocene on the basis of palynology (Siyako et al., 1989). The age of the fluviatile parts (Anafartalar Member/Çanakkale Formation) of the unit is late

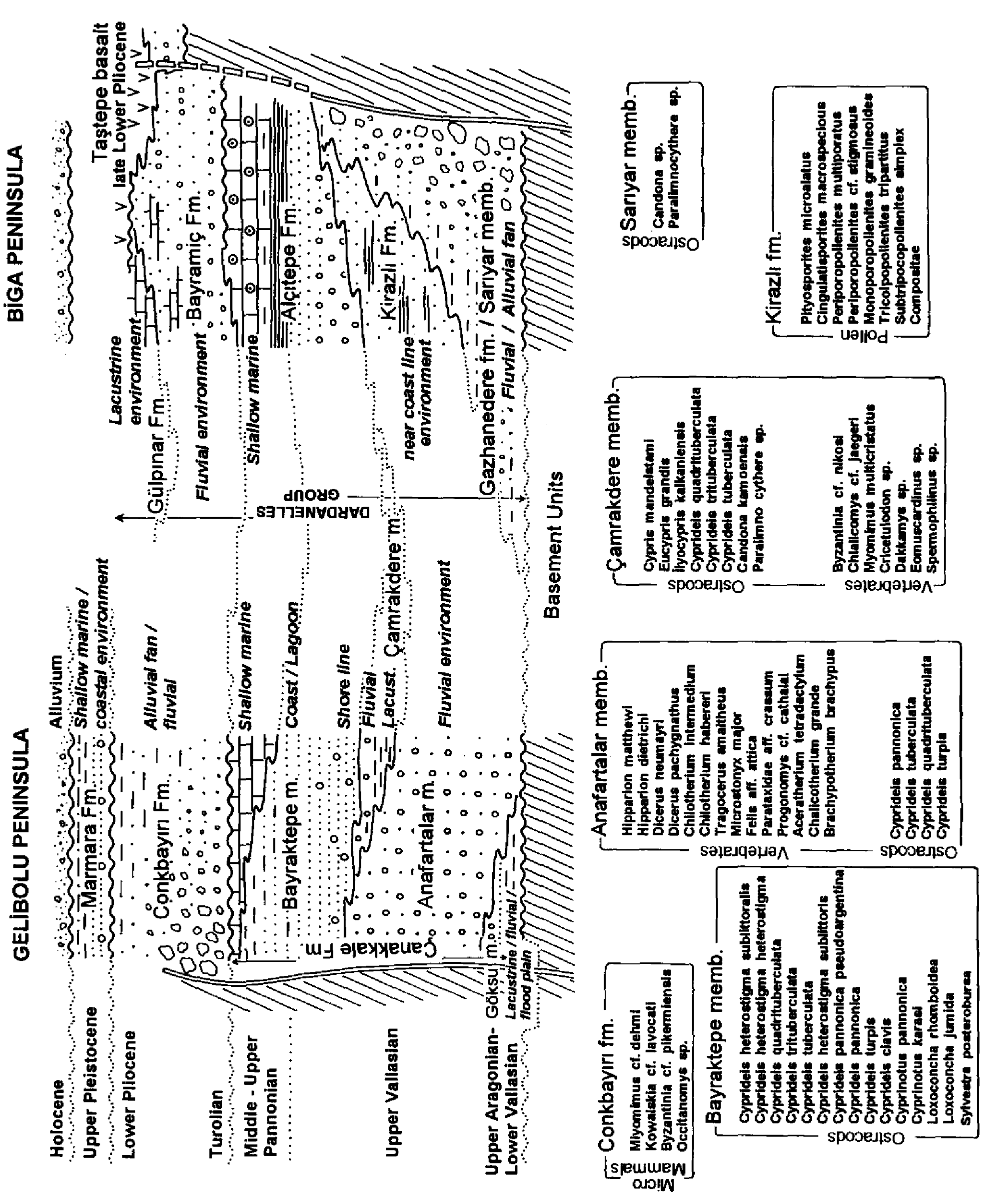


FIG. 2. Columnar section showing late middle Miocene to Holocene units in the Gelibolu and Biga peninsulas.

Aragonian–middle Pannonian (Sentürk and Karaköse, 1987).

All of the units gradually pass upward into a sequence consisting of coarse clastics at the base and finer clastics and oolithic carbonates toward the top (Alçitepe Formation) (Önem, 1974) (Fig. 2). Shallow-marine carbonates forming the upper parts of the sequence become widespread toward the inner parts of the Biga Peninsula inland from the coast of the Dardanelles. On the basis of ostracods, the unit is middle Pannonian in age (Sentürk and Karaköse, 1987) but extends to the late Pannonian (A. Nazik, 1996, pers. commun.).

Fluvial deposits (Bayramiç Formation) (Siyako et al., 1989) consisting of conglomerate, sandstone, and shale rest unconformably on shallow-marine carbonates; the upper parts of this unit include some lacustrine carbonates (Gülpinar Formation) (Fig. 2). The unit interfingers with basaltic rocks (Tastepe Basalt) extruded along faults; radiometric dating of these basalts yields an age of 3.8 Ma (Y. Yilmaz, 1994, pers. commun.). The shallow-marine equivalents of the Bayramiç Formation, determined in oil wells in Edremit Bay and outcrops situated in the southern parts of the Çanakkale city center, near the Dardanelles, are conformable with the underlying upper Miocene units (Siyako et al., 1989). The age of the unit is latest Miocene–Pliocene.

## The Middle–Upper Miocene/Quaternary Sequence in Gelibolu Peninsula

Middle–upper Miocene fluvial, lacustrine, lagoonal, coastal, and offshore deposits (Çanakkale Formation) (Sentürk and Karaköse, 1987) unconformably overlie basement rocks that are composed of Upper Cretaceous–Oligocene sedimentary rocks and an ophiolitic mélange on the Gelibolu Peninsula and in southern Thrace and volcanic rocks around Enez (Fig. 1B). The lower parts of the middle–upper Miocene/Quaternary sequence are represented by coal-bearing claystone, clayey limestone, and sandstone on the Gelibolu Peninsula (Göksu Member) (Önem, 1974). These levels include freshwater fauna and flora in the laminated fine-grained material. There is no marine biota in the unit, which was deposited in shallow water and displays a regressive character. The lower parts of the sequence are represented by lacustrine deposits, changing into fluviolacustrine deposits near the top (Sentürk and Karaköse, 1987).

Eastward, the pre- and lower Pannonian sequence (Göksu Member) gradually passes into fluvial deposits composed of claystone-mudstone-sandstone-conglomerate and coal-bearing floodplain deposits (Anafartalar Member) (Sentürk and Karaköse, 1987). These units are represented by periodically upward-fining clastics and rare meandering-river deposits. Erosional surfaces mark the base of the cyclic sequence. Marine fauna are absent, but numerous terrestrial vertebrates and rare freshwater fauna are present (Fig. 2). Rose diagrams of the paleocurrent directions from the Anafartalar Member demonstrate that the flow directions of the meandering streams were toward the west. These currents also indicate a northward flow west of the Gelibolu Peninsula (Sentürk and Karaköse, 1987).

The sequence gradually passes upward into strata composed of claystone-sandstone-mudstone with conglomerates (Çamrakdere Member) (Sentürk and Karaköse, 1987) (Fig. 2), which are characterized by thin, well-defined bedding and laminations. Rarely, crossbedding and symmetrical ripples and grading also occur. Vertebrate fossils (Fig. 2) suggest that deposition took place in a marsh environment; exclusively freshwater biota are present. Because of these properties, the lower layers of the sequence indicate deposition in a lacustrine environment, whereas the upper levels reflect lacustrine, fluvial, and floodplain environments. According to the vertebrate fossils, fluvial deposits (Anafartalar Member) are upper Aragonian to lower Vallasian in age. On the other hand, the lacustrine units (Çamrakdere Member) that are vertically transitional with them have a late Vallasian lower age limit. On the basis of ostracods, the ages of both units extend to the early to middle Pannonian (Sentürk and Karaköse, 1987) (Fig. 2).

Fluvial (Anafartalar Member) and lacustrine (Çamrakdere Member) deposits gradually pass into a sequence consisting of clastics at the base, which becomes finer and grades into carbonates toward the top (Bayraktepe Member) (Sentürk and Karaköse, 1987) (Fig. 2). The coarse clastics forming the lower levels of the unit (middle to late Pannonian in age on the basis of ostracods) indicate a high-energy coastal environment with an unstable shoreline. These levels consist of a mixture of carbonate-cemented sandstone, shells of brackish-freshwater forms, sand derived from land, and vertebrate bones. The situation indicates material transportation by two flow directions. The middle levels, consisting of finer clastics and including fresh-brackish-haline water fauna, indicate deposition below wave base in a low-energy shallow-water environment (generally lagoon) separated from the open sea. These clastics are laterally transitional with carbonate mounds. The upper levels, consisting of calcarenites and limestones, indicate that deposition occurred in a low-energy, shallow-marine, nearshore environment influenced by intermittent transgressions and regressions (Sentürk and Karaköse, 1987). Except for carbonates forming the upper levels of the late middle to upper Miocene sequence, the beds lack a marine fauna.

A unit (Conkbayiri Formation) (Kellog, 1973) composed of intercalated mudstone, sandstone, silt-claystone, and conglomerate rests unconformably over fluvial deposits (Anafartalar Member) and shallow-marine carbonates (Bayraktepe Member) (Sentürk and Karaköse, 1987, Sümengen et al., 1987) (Fig. 2). Channel-fill deposits in the unit were derived from Eocene–Oligocene units cropping out west of the Gelibolu Peninsula. Mudflow, mass-flow, sheet-flood, and stream-flood deposits are abundant (Sentürk and Karaköse, 1987). The thickness of the unit and the grain size (block–fine sand–silt–clay) of the clastics decreases from west to east. The regressive sequence (Yaltirak, 1995a) forms a syncline whose limb is vertical near the western boundary and dips gently in the east (Sentürk and Karaköse, 1987). The unit, including vertebrate and rare freshwater ostracods, is regarded as of early Pliocene–early Pleistocene age on the basis of stratigraphic position (Yaltirak, 1995a) and Turolian age on the basis of micromammals (Sentürk and Karaköse, 1987). Considering the stratigraphic and sedimentologic characteristics, the unit appears to have been deposited in an alluvial-fan environment on the slopes of topographically high areas of the western parts of the Gelibolu Peninsula (Saner, 1985). In the area between Mürefte and the Gaziköy fault (Fig. 1B), Pliocene coarse clastics (Hosköy Formation) (Bargu, 1989) constitute the equivalent of the Conkbayiri Formation.

A Bacunian–Tyrrhenian sequence is situated in the upper parts of the middle to late Miocene–Quaternary sequence in the Gelibolu Peninsula (Fig. 2). The sequence (Marmara Formation) (Yaltirak, 1995b)—composed of fluvial deposits at the base and upper parts and shallow-water, warm-marine, fauna-bearing carbonates and clastics in the middle parts—unconformably overlies the Conkbayiri Formation. North of the Gelibolu Peninsula, along the shore between Gaziköy and Mürefte, the lower to middle parts of the sequence are represented by transgressive deposits, whereas the upper parts are regressive (Gaziköy Formation) (Bargu, 1989). The unit is overlain unconformably by a Holocene succession.

## Paleontologic Data

In the Gulf of Izmit, samples comprising a unit 118.45 m in total thickness were obtained by drilling nine boreholes (Figs. 1A,

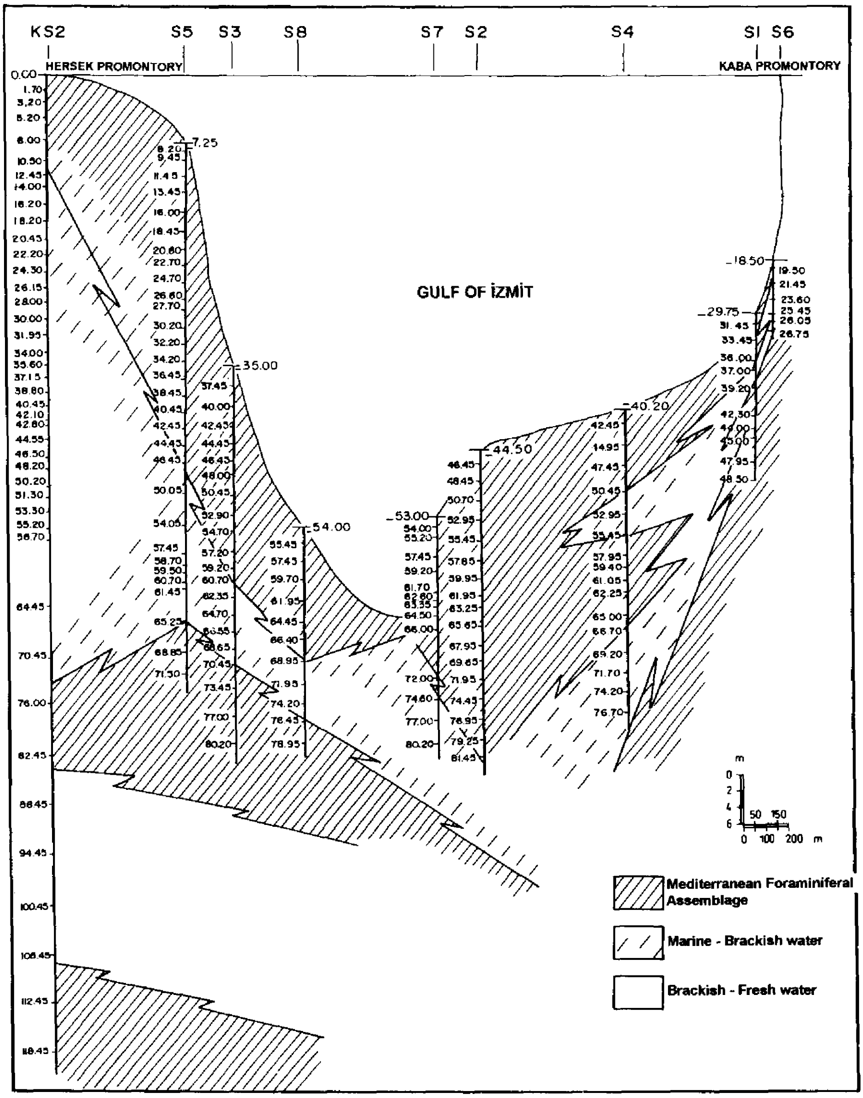


FIG. 3. The distribution of Mediterranean foraminifers in Izmit Bay for late Pliocene and Quaternary time.

3) between the Hersek and Kaba promontories. As a result of analysis, the times of arrival of Mediterranean waters to the Gulf of Izmit were determined as late Pliocene, Pleistocene, and Holocene.

*Nannoplankton assemblage*

At the base of the 118.45 m section, between 112.45 and 112.00 m, a *Discoaster brouweri* zone was identified (Toker and Sengüler, 1995). Thus, part of the sequence was deposited during the latest Pliocene. On the other hand, the existence of genera and species belonging to the Pleistocene *Pseudoemiliana lacunosa* zone, between 106.45–106.00 m and 70.45–70.00 m in depth, indicates that the Gulf of Izmit was influenced by Mediterranean waters during this period.

*Foraminiferal assemblage*

The deepest well between the Hersek and Kaba promontories is the KS-2 borehole (Fig. 3). Between 118.45–118.00, 112.45–112.00, and 76.00–75.55 meters of this drill-

| | | FORAMINIFERA ASSEMBLAGE | Wells No | (m) | NUMERIC VALUES (Year) |
|---|---|---|---|---|---|
| QUATERNARY | UPPER PLEISTOCENE - HOLOCENE | Spiroplectinella sagittula, Textularia aglutinans, T. sagittula, Adelosina cliarensis, A. mediterranensis, Spiroloculina excavata, S. depressa, Siphonaperta aspera, S. dilatata, Cyclofonna colomi, C. juleana, C. rugosa, Lachianella bicormis, L. undulata, Massilina secans, Quinqueloculina jugosa, Q. laevigata, Q. seminula, Miliolinella labiosa, M. subrotunda, Pseudotriloculina laevigata, Pyrgo elongata, Triloculina marioni, T. tricarinata, Sigmoilinita costata, Parrina bradyi, Amphicoryna scalaris, Favulina hexagona, Brizalina alata, B. spathulata, Cassidulina carinata, Bulimina aculeata, B. elongata, B. marginata, Ruessella spinulosa, Valvulineria bradyana, Eponides repandus, Stomatorbina concentrica, Neoconorbina orbicularis, Rosalina bradyi, Discorbinella berthelotiana, Cibicidina walli, Lobatula lobatula, Planorbulina mediterranensis, Sphaerogypsina globula, Asterigerinata mamilla, Nonionella turgida, Ammonia compacta, A. parkinsoniana, Elphidium aculeatum, E. advenum, E. complanatum, E. crispum, E. macellum. | KS-2<br>S-5<br>S-3<br>S-8<br>S-7<br>S-2<br>S-4<br>S-1<br>S-6 | 10.50-2.75<br>30.20-9.00<br>48.00-37.00<br>57.45-55.00<br>55.20-53.50<br>55.45-48.00<br>47.45-42.00<br>33.45-31.00<br>21.45-19.00 | Holocene<br>10.000<br>Upper Pleistocene<br>130.000 |
| | LOWER - MIDDLE PLEISTOCENE | Spiroloculina excavata, Quinqueloculina laevigata, Q. seminula, Miliolinella subrotunda, Brizalina alata, B. spathulata, Cassidulina carinata, Rosalina bradyi, Hyalinea baltica, Cibicides floridanus, Lobatula lobatula, Planorbulina mediterranensis, Asterigerinata mamilla, Nonionella atlantica, N. turgida, Ammonia compacta, A. parkinsoniana, Elphidium advenum, E. crispum, E. macellum. | KS-2<br>S-3<br>S-2 | 42.10-41.60<br>70.45-70.00<br>79.25-78.80 | Middle Pleistocene<br>500.000<br>Lower Pleistocene<br>750.000 |
| TERTIARY | U. PLIOCENE LOWER PLEISTOCENE | Spiroloculina excavata, Quinqueloculina seminula, Brizalina alata, B. spathulata, Cassidulina carinata, Bulimina elongata, B. marginata, Valvulineria bradyana, Lobatula lobatula, Asterigerinata mamilla, Nonionella atlantica, N. turgida, Chilostomella mediterranensis, Cribroelphidium poeyanum, Elphidium complanatum, E. macellum. | KS-2 | 76.00-75.55<br>112.45-112.00<br>118.45-118.00 | Eopleistocene<br>1.800.000 |

FIG. 4. Foraminifera assemblage (from Meriç, 1995) for the late Pliocene–Quaternary period, Izmit Bay. Numerical values are from Yanko (1990).

hole, foraminifers of Mediterranean origin (Meriç and Sakinç, 1990; Cimerman and Langer, 1991; Sgarella and Montcharmont-Zei, 1993) were found together with late Pliocene nannoplankton in the younger sedimentary rocks (Fig. 4). This indicates that Mediterranean waters reached the Gulf of Izmit during this period. Mollusc shells in the KS-2 well between 55.20 and 54.75 m yielded an age of 817.000 ± 105.000 yr by the electron spin resonance (ESR) method. Therefore, the age of these parts of the sequence is Eopleistocene (Koreneva and Kartashova, 1978; Yanko, 1990).

The existence of early to middle Pleistocene sediments in the upper parts (KS-2, S-3, S-2, and S-4 wells) of the investigated sequence (Fig. 3) was determined by the ESR method (Çetin et al., (1995). Çetin et al. (1995) obtained ages of: 693,000 ± 126,000 yr (S-4; 71.70–71.25 m); 664,000 ± 94,000 yr (KS-2; 42.10–41.60 m); 320,000 ± 37,000 yr (S-3; 70.45–70.00 m); 306,000 ± 39,000 yr (S-2; 79.25–78.80 m); 254,000 ± 34,000 yr (S-7; 77.00–76.55 m); 199,000 ± 22,000 yr (S-8; 74.20–73.75 m); 198,000 ± 23,000 and 195,000 ± 20,000 yr (S-4; 55.45–55.00 and 52.45–52.00 m); and 186,000 ± 20,000 yr (S-8; 64.45–64.00 m). In the KS-2 (42.10–41.60 m), S-3 (70.45–70.00 m), and S-2 (79.25–78.80 m) wells, characteristic Mediterranean foraminifers (Meriç and Sakinç, 1990; Cimerman and Langer, 1991; Sgarella and Montcharmont-Zei, 1993) are found (Fig. 4). These data show that the Gulf of Izmit again received Mediterranean waters in early to middle Pleistocene time.

Finally, the region was influenced by Mediterranean waters from 35,200 ± 8,100 yr. The obtained age data are: 35,200 ± 8,100 yr (50.05 m), 33,700 ± 7,200 yr (40.45–40.00 m), and 31,700 ± 5,400 yr (38.45–38.00 m) in the S-5 drillhole; 24,800 ± 3,700 yr (44.45–44.00 m) in the S-3 drillhole; 15,400 ± 3,700 yr (26.60–26.15 m) in the S-5 drillhole; 11,600 ± 2,800 yr (5.20–4.75 m) and

| | | OSTRACODA ASSEMBLAGE | Wells No | (m) | (Year) NUMERIC VALUES |
|---|---|---|---|---|---|
| QUATERNARY | UPPER PLEISTOCENE - HOLOCENE | Cyprideis sorbyana,<br>Cythereis jonesi,<br>Carinocythereis antiquata,<br>C. carinata, C. quadridentata,<br>Costa batei, C. edwardsii,<br>Urocythereis britannica | KS-2<br>S-5<br>S-3<br>S-8<br>S-7<br>S-2<br>S-4<br>S-1<br>S-6 | 10.50-2.75<br>30.20-9.00<br>48.00-37.00<br>57.45-55.00<br>55.20-53.50<br>55.45-48.00<br>47.95-42.00<br>33.45-31.00<br>19.50-19.00 | Holocene<br>10.000<br>Upper Pleistocene<br>130.000 |
| | LOWER - MIDDLE PLEISTOCENE | Carinocythereis quadridentata,<br>Costa batei, C. edwardsii,<br>Urocythereis britannica,<br>Loxoconcha rhomboidea,<br>Pseudocytherura calcarata | S-3<br>S-2 | 70.45-70.00<br>79.25-78.80 | Middle Pleistocene<br>500.000<br>Lower Pleistocene<br>750.000 |
| TERTIARY | U. PLIOCENE LOWER PLEISTOCENE | Loxoconcha rhomboidea,<br>Loxoconcha gibberosa,<br>Costa edwardsii | KS-2 | 55.20-54-75<br>94.45-94.00<br>118.45-118.00 | Eopleistocene<br>1.800.000 |

FIG. 5. Ostracod assemblage (from Meriç et al., 1995) for the late Pliocene–Quaternary period, Izmit Bay. Numerical values are from Yanko (1990).

6,600 ± 700 yr (3.20–2.75 m) in the KS-2 drillhole; 5,000 ± 900 yrs (8.20–8.00 m) in the S-5 drillhole; and 500 ± 200 yr (37.45–37.00 m) in the S-3 drillhole (Çetin et al., 1995).

The foraminifera observed in the upper levels of the wells were evaluated collectively. Results show the abundance of foraminifers of Mediterranean origin in the Gulf of Izmit during 12,000 ± 1,900 and 500 ± 200 yr B.P. (Fig. 4). As a result, the seaway connection between Mediterranean and the Sea of Marmara was established in the late Pliocene and Quaternary.

*Ostracod assemblage*

Samples from the wells between the Hersek and Kaba promontories contain a rich ostracod fauna. Especially the ostracod assemblage (Gülen et al., 1995) from the KS-2 well (118.45–118.00 m, 94.45–94.00 m, and 55.20–54.75 m) (817,00 ± 105,000 yr) (Çetin et al., 1995) confirmed the arrival of ostracods of Mediterranean, Aegean, and Adriatic origin to the Sea of Marmara during late Pliocene (Toker and Sengüler, 1995) to early Pleistocene time (Fig. 5).

In the upper parts of the young sediments, Mediterranean, Adriatic, and Aegean ostracods (Fig. 5) were obtained (Gülen et al., 1995), which yielded ages of 320,000 ± 37,000 yr (S-3, 70.45–70.00 m) and 306,000 ± 29,000 yr (S-2, 79.25–78.80 m) (Çetin et al., 1995). These results indicate that the ostracods lived in the Sea of Marmara during the early to middle Pleistocene. Those reported by Gülen et al. (1995) from the upper Pleistocene–Holocene sediments also are of Mediterranean, Aegean, and Adriatic origin (Puri et al., 1969; Barbeito-Gonzalez, 1971; Bonaduce et al., 1975; Ruggieri, 1976; Yassini, 1979) (Fig. 5).

An eastern Mediterranean form, *Urocythereis britannica* (Athersuch), is present in Pliocene sediments of the Aegean region (Sissingh, 1972), in Cyprus (Athersuch, 1977), and in some areas in the Aegean Sea such as Edremit Bay and the area west of Bozcaada (C. Kubanç, 1994, pers. commun.). Thus, the species reached the Mediterranean and Aegean seas in the Pliocene and the Sea of Marmara in the early to middle Pleistocene through a seaway connection from the Atlantic.

## Discussion and Evolution

At the lower parts of the middle to upper Miocene/Quaternary sequence (Dardanelles Group and Çanakkale Formation) in the

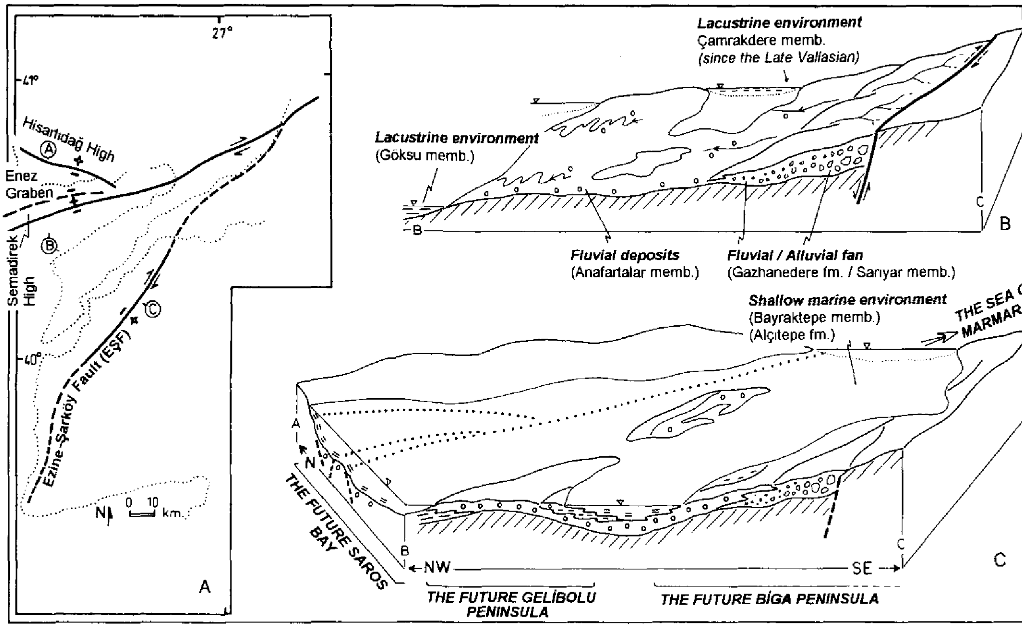


FIG. 6. Structural map (A) and block diagrams of the study area for late Aragonian–early Pannonian (B) and middle to late Pannonian (C) time.

region, the observed east-to-west facies changes (Figs. 2 and 6) indicate the existence of a topographic high to the east of a line from Lapseki to Ezine on the Biga Peninsula during late Aragonian–middle Pannonian time. An alluvial fan (Sariyar Member) (Sentürk and Karaköse, 1987) formed toward the west from the topographic high produced by a fault (Ezine-Sarköy fault). Reflecting this topography, streams flowed from east to west (Anafartalar Member) (Sentürk and Karaköse, 1987), generating downstream lacustrine environments (Göksu Member) (Sentürk and Karaköse, 1987) (Fig. 6B). On the other hand, the lacustrine basins (Çamrakdere Member) (Sentürk and Karaköse, 1987), which developed since the late Vallasian, formed in the central parts of the area produced by river deposits. The sites of these basins coincide with the modern-day Dardanelles (Fig. 6B).

In the eastern Marmara region, the North Anatolian fault (NAF) splays into two branches. The Ezine-Sarköy fault, which caused the facies changes east to west from the western Marmara region during late Aragonian to middle Pannonian time, is a strike-slip fault connecting the northern branch beneath the northern part of the Sea of Marmara with the southern branch in the Biga Peninsula of the NAF. The vertical slip component of this fault exceeded the strike-slip component east of the Dardanelles during this period, as a result of the westward movement of the southern block of the northern branch, the eastward movement of the northern block of the southern branch, and a releasing bend in the northwestern part of the Biga Peninsula (Fig. 6A).

The middle parts of the Dardanelles Group are represented by a transgressive sequence (Alçitepe Formation [Önem, 1974]; Bayraktepe Member [Sentürk and Karaköse, 1987]) that begins with clastics and passes upward to carbonates. The middle to upper Pannonian shallow-marine carbonates forming the upper parts of the sequence also overlie basement rocks in the Biga Peninsula (Fig. 6C). The unit forms a unique part of the Dardanelles Group in which marine fauna existed. Shallow-marine carbonates crop out almost everywhere in the region, except for tectonically uplifted areas such as north of the Ganos fault, west of the Gelibolu Peninsula, and just east of the Dardanelles (Fig. 1B). The following deductions can be made on the basis of the overall characteristics of the unit: (1) following the middle Pannonian, areas of lacustrine environment became a shallow sea; (2) this shallow sea covered an extensive area; (3) in the middle (?) to late Pannonian (9–10 Ma), the Sea of Marmara and the Mediterranean were connected by a broad seaway around the modern-day Dardanelles, Gelibolu Peninsula, and Saros Bay; and (4) lagoonal deposits, however, are present in the western parts of the Gelibolu Peninsula (Yaltirak, 1995b). In addition, the fluvial deposits progressively filled the lacustrine basins (Çamrakdere Member) in the northwestern part of the Biga Peninsula and the area of the Gelibolu Peninsula during the Pannonian. Thus, land areas developed in the center of the wide seaway. During the middle (?) to late Pannonian, the connection between the Sea of Marmara and the Mediterranean occurred essentially via the modern-day Dardanelles and Saros Bay (Fig. 6C). The western part of the Gelibolu Peninsula consisted of small islands within the broad seaway in the middle (?) to late Pannonian.

Accompanying the development of the late middle Miocene to late Miocene units (Dardanelles Group and Çanakkale Formation), the Marmara trough also began to subside as a result of the development of several pull-apart structures along the northern branch of the NAF in the region (Sengör et al., 1985; Barka and Kadinsky-Cade, 1988; Barka, 1992). According to Sentürk and Karaköse (1987), the development of the pull-apart basins started in the pre–late Miocene, whereas Barka (1992) suggested a Pliocene age. The ages of non-marine and lacustrine deposits resting on basement units around the Dardanelles, Saros Bay, and Sea of Marmara indicate that the western prolongation of the NAF, which controlled the development of these basins, existed since the late middle Miocene. According to Sengör et al, (1985), the westward movement of the Anatolian plate caused by the NAF began in the middle to late Miocene (late Serravalian–Tortonian). In addition, the early to middle Miocene basins in the vicinity of the NAF are unrelated to the fault, but the late Miocene–Pliocene basins were controlled by releasing bends of the fault zone (Barka, 1992).

Diverse structural entities may be distinguished around Saros Bay. From northwest to southeast these are the Hisarlidag high, Enez graben, Semadirek high, Saros graben, and Gelibolu block (Saner, 1985).

During the late Miocene, a basin (Enez graben) developed between the Semadirek and Hisarlidag highs (Fig. 6A). Seismic sections indicate that the Dardanelles Group/Çanakkale Formation is the initial unit covering the upper Oligocene basement units in the Enez and Saros grabens, as is the case in land areas (Saner, 1985). The lower parts of Mio-Pliocene deposits (the equivalents of the Dardanelles Group/Çanakkale Formation), identified in the seismic sections, are cut by synsedimentary faults and transgressively overlie the Semadirek and Hisarlidag highs (Saner, 1985). Thus, subsidence in the Enez graben began since the development of the Dardanelles Group/Çanakkale Formation (in the late middle Miocene) and ended in the late Miocene. The horizontal position and the stability of the Quaternary deposits in the region imply the absence of recent active faulting (Saner, 1985).

The fault bounding the Semadirek high to the south constitutes the dextral northern branch of the NAF in the region during the late middle Miocene–late Miocene. The fault bordering the Enez graben on the north is compatible with the normal faults defined by Ramsay (1967) and developed in the strike-slip fault zones.

During the development of the fault bounding the Semadirek high on the south, the existence of a graben (Saros graben) to the south is suspected. In other words, there

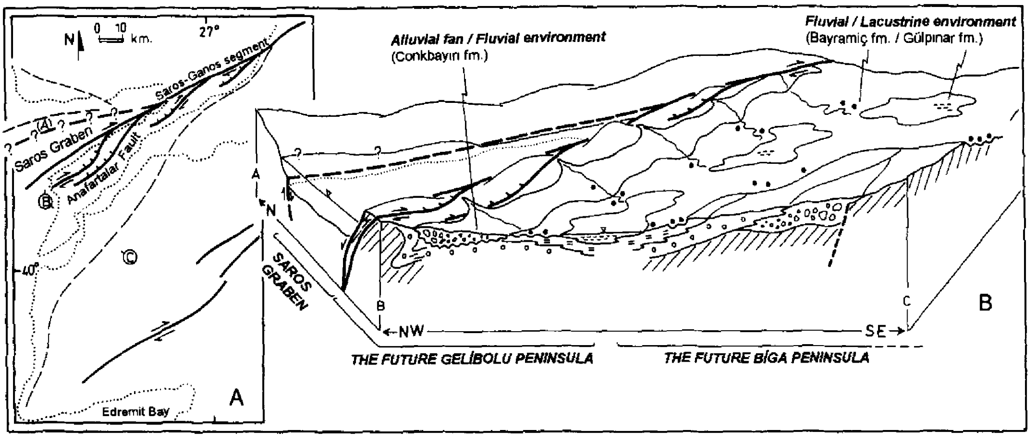


FIG. 7. Structural map (A) and block diagram (B) of the study area for the late upper Miocene–early Pliocene.

are no data supporting the presence of a fault bordering the Saros graben to the south. In the basin bounded by Ezine-Sarköy fault to the east (Fig. 6B), there is a transition from the alluvial fan units (Sariyar Member) to lacustrine deposits (Göksu Member). Considering the presence of the Saros graben and the fault bounding it to the south: (1) the topographic high created by this fault did not disturb the stability of lacustrine basins; and (2) these basins are bounded in the west by slightly dipping slopes of the eastward-tilted rising block. However, paleocurrent analyses of stream deposits of this period indicate the presence of streams flowing from the east into lakes locted in the western part of the Gelibolu Peninsula. No stream flowed from west to east. The Saros graben has been shaped since the development of a fault (Saros-Ganos segment) bounding it on the south after deposition of the Çanakkale Formation.

The development of the fault bounding the Saros graben to the south since the late Miocene(?)–early Pliocene resulted in a decrease or possible termination in the activity of the fault bounding the Semadirek high to the south (Fig. 7). Therefore, the Enez graben was filled at the end of the Miocene. The compression in the region produced by the fault bounding the Saros graben to the south is compensated by the Anafartalar fault.

The Anafartalar fault (Yaltirak, 1995a), which formed as a positive flower structure (Christie-Blick and Biddle, 1985), elevated the western parts of the Gelibolu Peninsula. The seaway connection has been obstructed for this reason around Saros Bay since the latest Miocene. The uplifted area between the Anafartalar fault and the fault bounding the Saros graben to the south resulted in the development of alluvial-fan deposits (Conkbayiri Formation) (Kellog, 1973) in the eastern areas (Saner, 1985; Yaltirak, 1995a). The Conkbayiri Formation is located east of the uplift (Figs. 1B, 7B).

The compression and uplift in the western parts of the Gelibolu Peninsula have not resulted in faulting everywhere in the region (Fig. 7), although Sentürk and Karaköse (1987) did report the existence of a few reverse faults. In the middle parts of the Gelibolu Peninsula, the units forming the basement of the middle to upper Miocene deposits are overturned toward the southeast (Saner, 1985). Erkal (1991) reported the presence of strike-slip faults branching from the NAF north of Sarköy. The northern blocks of these faults are overthrust toward the southeast (Sümengen et al., 1987). Thus, the uplift and occasional break-up of the western parts of the Gelibolu Peninsula are restricted to the initial stage of development of the late upper Miocene (?)–lower Pliocene Conkbayiri Formation.

The fault bounding the Saros graben to the south, as well as the Anafartalar fault, also are responsible for the termination of the Ezine-Sarköy fault, which was active from the late middle Miocene–early late Miocene until the early Pliocene (Fig. 7). Late upper Miocene–lower Pliocene stream/lacustrine deposits (Bayramiç and Gürpinar formations) (Siyako et al., 1989) in the Biga Peninsula are the temporal equivalents of the Conkbayiri Formation, controlled by the Anafartalar fault (Fig. 2). The lithological characteristics of the lower and middle parts of these deposits indicate that the Ezine-Sarköy fault terminated its activity during the period of their development. The shallow-marine equivalents of the Bayramiç and Gülpinar formations—found in the well in Edremit Bay and outcrops near the Dardanelles south of Çanakkale—are conformable with the underlying upper Miocene units (Siyako et al., 1989). Therefore, a shallow-marine environment has persisted since the late Miocene around Edremit Bay. The shallow seaway (also encompassing the Dardanelles) that existed between the Sea of Marmara and the Mediterranean in the late Miocene during the latest Miocene–early Pliocene was obstructed by the erosion of the area raised by the Anafartalar fault and the filling of the trough situated to the east (Fig. 7B). However, in areas farther east (the western part of Biga Peninsula), where the trough filling material could not have reached, deposition continued in a fluviolacustrine environment since the late Miocene (Fig. 7B). Despite the shallowing, the continuity of the deposition was provided by the westward escape of the southern block of the fault bounding the Saros graben on the south. This escape subsequently caused reactivation of the fault (Ezine-Sarköy fault) northwest of the Biga Peninsula during late early Pliocene time.

In the Biga-Bayramiç-Çanakkale area, the nature of volcanic activity was calc-alkaline in the Miocene (Ercan, 1981), alkaline basaltic at the end of the Miocene (Borsi et al., 1972), and alkaline in the late Pliocene (Ercan, 1981). The lavas (Tastepe basalt) (Siyako et al., 1989) extruded along faults in the Biga Peninsula are intercalated with the upper parts of the Bayramiç Formation and rest on it. Radiometric dating of these basalts, which extend from north of Ezine to south of Lapseki (Fig. 1), yields ages of 3.8 Ma (Y. Yilmaz, 1994, pers. commun.).

The Ezine-Sarköy fault, which extruded the basic lavas, forms the boundary between the basement units in the east and middle to upper Miocene/Pliocene units in the west. Thus, this fault, which previously affected the development of the late middle Miocene to early upper Miocene alluvial-fan/stream deposits (Sariyar Member and Gazhanedere Formation) in the Biga Peninsula, was reactivated in the late early Pliocene (Fig. 8). Subsidence on the western part of the fault caused development of a trough. The Ezine-Sarköy fault is the prolongation of a dextral strike-slip fault (Ambraseys, 1970; Allen, 1975; Sümengen et al., 1987) that resulted in the development of a compressional trough (Sengör et al., 1985) offshore from Sarköy and Kumbag (Figs. 9, 10). The location of this fault is compatible with the strike-slip faults of pull-apart basins (Sengör et al., 1985; Barka and Kadinsky-Cade, 1988; Barka, 1992) beneath the Sea of Marmara (Fig. 9).

The opening of the Dardanelles probably was controlled by the structural entities related to these strike-slip fault zones. These structural features are simple synthetic faults that have the same sense of slip as the main fault (Wilcox et al., 1973) (Fig. 8). These structures apparently caused the seaway connection between the Sea of Marmara and the Mediterranean, because drillhole data from the Gulf of Izmit indicate that such a connection has existed since the late Pliocene.

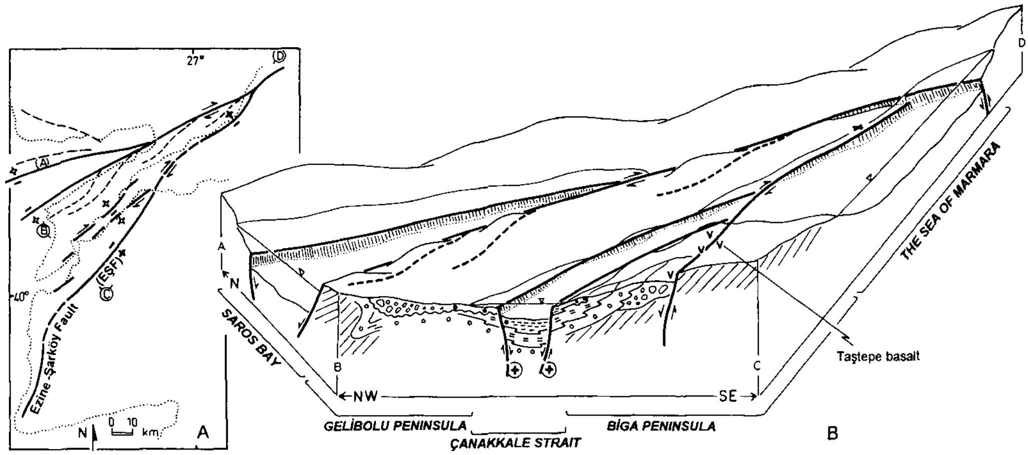


FIG. 8. Structural map (A) and block diagram (B) of the study area for the late lower Pliocene–late Pliocene.

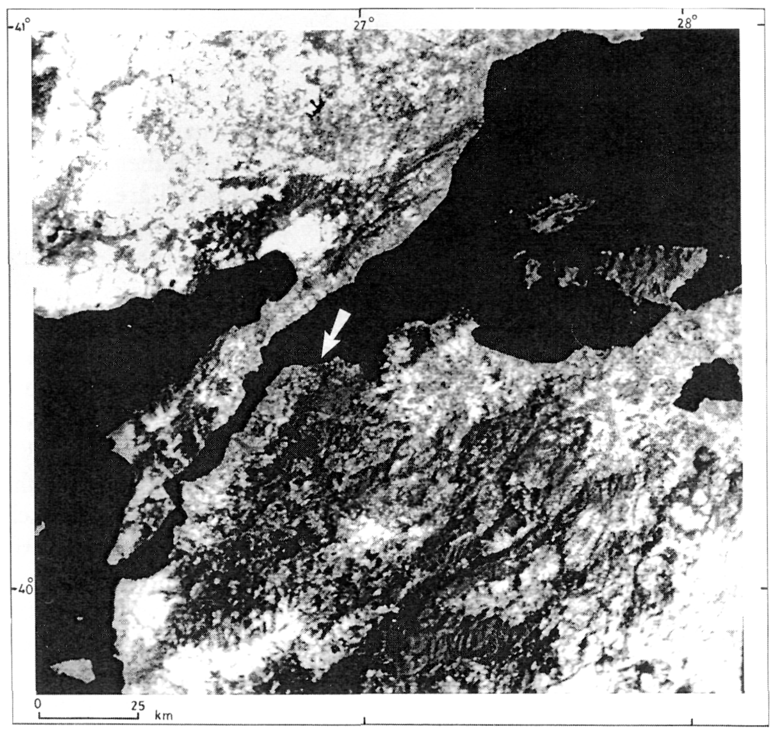


FIG. 9. Enhanced Landsat-5 Thematic Mapper image of the Dardanelles area. White arrow points to trace of the Ezine-Sarköy fault.

Although these structures can be observed on land, they are obscured in the Dardanelles. Moreover, westward escape of the southern block of the fault bounding the Saros graben to the south accelerated the extension and the subsidence in the region.

It has been proposed that the Dardanelles was a fluvial valley (Penck, 1917; Yalçinlar,

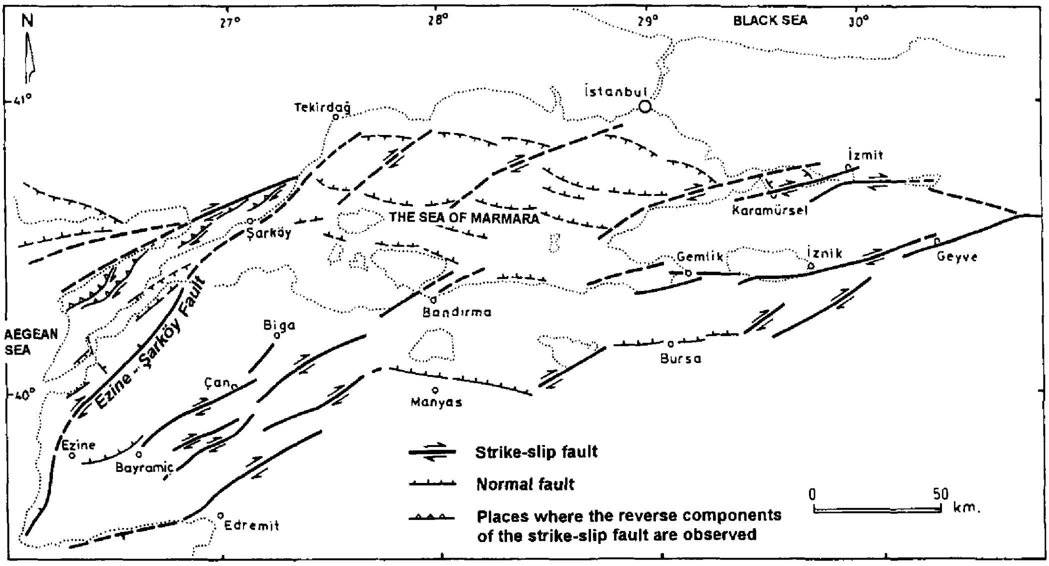


FIG. 10. Neotectonic faults of the Marmara region (adapted from Saner, 1985; Barka and Kadinsky-Cade, 1988; Siyako et al., 1989; Smith et al., 1995).

1949; Ardel and Inandik, 1957; Erol, 1968; Sentürk and Karaköse, 1987). However, on the Biga Peninsula, streams such as Karamenderes-Kepez-Sariçay-Yapildak-Umurbey and Çinarli, which flow into the Dardanelles, have V-shaped valley profiles drowned by alluvium. The valley bottom of the Karamenderes River is characterized by a buried meandering river. Furthermore, the hydraulic slopes of the streams, which flow from the Gelibolu Peninsula to the Dardanelles, increase suddenly upon approaching the strait (Sentürk and Karaköse, 1987). In addition, Quaternary marine and non-marine terraces are situated on both sides of the Dardanelles (Ardel and Inandik, 1957; Ardel, 1960; Erol, 1968, 1973; Erol and Inal, 1980). The marine terraces of 4–5 m, 12–15 m, 30–35 m, and 115 m are pebbly or sandy, and occasionally include biogenetic sand and pelecypods. Moreover, normal faults lying parallel to the coast exist on both sides of the Dardanelles, and those portions of the faults that are adjacent to the strait evidently have subsided. All of these characteristics indicate that both sides have been actively faulted in recent time. Önem (1974) proposed that the Dardanelles is a graben.

The extensional areas and troughs in the region probably were formed by strike-slip faults, owing to their pronounced vertical components. The faults bounding the Saros graben are characterized by both horizontal and vertical components (McKenzie, 1978). In addition, the distribution of $M > 2.5$ earthquakes recorded by the Kandilli Observatory between 1976 and 1990 in the Marmara region indicates that the pull-apart basins along the northern branch of the NAF are undergoing continuous activity (Üçer et al., 1985). However, the connecting strike-slip faults and the branches of the NAF in the Biga Peninsula are characterized by little or no activity (Barka, 1992). In this case, the strike-slip components of the faults, which caused the opening of the Dardanelles, now dominate.

The fault system that gave rise to the opening of the Dardanelles follows the boundary between two tectonic units welded in the early Tertiary. One of these is the Upper Cretaceous–Paleocene ophiolitic mélange (Çetmi Ophiolitic Mélange/Geli-

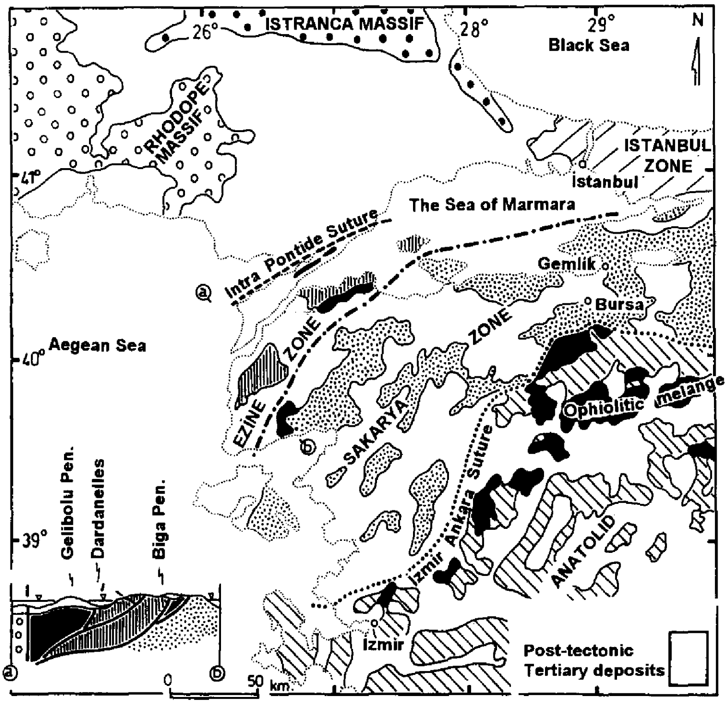


FIG. 11. Geotectonic map of the region surrouding the Sea of Marmara (from Okay et al., 1990).

bolu Zone) (Okay et al., 1990), which forms the basement in the Gelibolu Peninsula. The other unit consists of Permo-Carbonifereous metasedimentary rocks and ophiolite nappe thrust onto them (Ezine Zone) (Okay et al., 1990) in the basement of the northwestern parts of the Biga Peninsula (Fig. 11).

On the eastern coast of the Gelibolu Peninsula, the Bacunian-Tyrrhenian sequence (Marmara Formation) (Yaltirak, 1995b) unconformably rests on upper Miocene and lower Pliocene sedimentary rocks. The sequence, which is transgressive in the lower and middle parts and regressive in the upper (Bargu, 1989), includes shallow-marine strata. The sequence on land provides evidence for the connection between the Sea of Marmara and the Mediterranean via the Dardanelles during the late Pleistocene. In addition, the Tyrrhenian marine units (Erinç, 1956) of the graben in the Gulf of Izmit were developed by the influence of mesoscale strike-slip and other faults (Sengör et al., 1982). The equivalents of the Tyrrhenian deposits (Altinova Formation) (Sakinç and Bargu, 1989) in the Izmit-Karamürsel region are present south of the Gaziköy fault. The upper Pleistocene deposits on the southern block of the Gaziköy fault were uplifted by as much as 80 m (Bargu, 1989).

Uplift and subsidence of fault blocks in the graben and sea-level changes during the Pleistocene glacial period intermittently obstructed the shallow-marine connection around the Dardanelles. According to Sümengen et al. (1987), sea level began to rise at the end of the glacial period, since Holocene time, and had risen by up to 40 m by 10,000 years B.P.

## Conclusions

Stratigraphic and paleontologic data obtained from adjoining land areas indicate an initial seaway connection—extending from southern Thrace to west of the Biga

Peninsula—between the Sea of Marmara and the Mediterranean Sea during middle to late Pannonian time. This seaway connection is not directly related to tectonism. The Saros-Ganos segment, forming a branch of the North Anatolian fault, developed as a positive flower structure in the late Miocene to early Pliocene; this structural high obstructed the seaway connection around Saros Bay. Furthermore, clastics originating from this positive feature filled the Dardanelles Strait area of today.

In addition, foraminifer, ostracod, and nannoplankton assemblages of Mediterranean origin have been found in late Pliocene–Holocene deposits in the Gulf of Izmit. These data suggest that the Gulf of Izmit and the Sea of Marmara were connected to the Mediterranean, but that this connection was related directly to tectonism and occurred since the late Pliocene via the Dardanelles. The data and the foraminiferal assemblage of sediments in Izmit Bay indicate intermittent connections instead of a continuous one. The oldest Mediterranean connection occurred in the late Pliocene–early Pleistocene. The next connection was in the early middle Pleistocene, and the final one occurred during the late Pleistocene–Holocene.

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