

Paratethyan–Mediterranean connectivity in the Sea of Marmara region (NW Turkey) during the Messinian

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Abstract

The Sea of Marmara region is thought to have been a gateway between Paratethys and the Mediterranean since the Middle Miocene, and is therefore an important control on water mass exchange between the two realms. The Miocene successions in the northeastern Aegean and northwestern Marmara regions indicate that the first Mediterranean marine transgression to affect these areas occurred during the late Serravallian.

In the northeastern Aegean region, frequent marine incursions occurred during the Tortonian and Messinian stages. The Messinian stage in this area is represented by a package of brackish- to fresh-water carbonates with some marine sandstone–siltstone interbeds (Alçıtepe Formation), which conformably overlies the Tortonian Kirazlı Formation. The Messinian sequence is overlain with an erosional contact by a shallow marine siliciclastic sequence (Göztepe Formation) of Zanclean age. With its brackish- to fresh-water carbonates and broadly constrained age, the Messinian sequence is interpreted as being coeval with the Upper Evaporite–Lago Mare sequence observed in western Mediterranean basins.

In the western Marmara region, the Pontian (Messinian) Alçıtepe Formation consists of bioclastic and oolitic limestones with basal clastic rocks. It conformably overlies the fluvio-lacustrine siliciclastic sediments of the Middle to Upper Miocene Kirazlı Formation and is overlain by fluvio-lacustrine sediments of the Kimmerian (5.5–3.2 Ma) Truva and Tefikiye formations with an erosional contact.

The bioclastic limestones of the Alçıtepe Formation in the western Marmara region contain a molluscan and ostracod fauna that are endemic to Paratethys. These fauna indicate deposition in a shallow, brackish- to fresh-water environment. Faunal and paleomagnetic analyses of a section of the Alçıtepe Formation at Yenimahalle (Çanakkale) confirm that the formation is of Pontian age and represents chron C3r (6.04–5.24 Ma). The ostracod analysis indicates that during deposition of the Alçıtepe Formation, salinity increased from brackish in the lower part to more saline conditions in the upper part. Ostracod valves have low ⁸⁷Sr / ⁸⁶Sr values relative to coeval Late Miocene ocean water. This indicates that exchange between the Sea of Marmara region and the global ocean was restricted throughout this period. Fossil and Sr-isotope evidence suggests, however, that there was a Paratethyan–Marmara connection during the deposition of the lower part of the Alçıtepe Formation, with Paratethyan influence reaching the north Aegean. Connection via Marmara between Paratethys and the Mediterranean was not re-established until the late

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Aktchagylian (Late Pliocene). The re-connection was caused by both increased activity on the North Anatolian Fault and global sea level rise.

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Keywords: Paratethys; Aegean Sea; Marmara region; NW Turkey; Mediterranean; Messinian salinity crisis; Water exchange

1. Introduction

The Sea of Marmara and its surrounding region has been recognized as a gateway which episodically linked Paratethys and the Mediterranean since the Middle Miocene. It is therefore an important area for constraining the timing and nature of water-mass exchanges between the two realms (Fig. 1). The Messinian Salinity Crisis (MSC) in the Mediterranean was a major oceanographic event which resulted in the deposition of thick evaporite sequences (e.g., Hsü et al., 1973; Ryan et al., 1973; Hsü et al., 1978; Cita et al., 1978). Although the cause of the event has now been resolved mainly in favor of the tectonic closure of the Atlantic connections, timing (onset and termination in different parts of the Mediterranean basin) and depositional conditions (drastic salinity changes) are still matters of controversy. The impact of the MSC on the Paratethys is also poorly

constrained (Cita et al., 1978; Kojumdjieva, 1979; Hsü and Giovanoli, 1979/1980).

The sediments deposited during the MSC in the Mediterranean marginal basins include the Lower and Upper Evaporites and the overlying brackish- to fresh-water “Lago-Mare” sediments (Hsü et al., 1973, 1978; Cita et al., 1978; Rouchy, 1982; Rouchy and Saint Martin, 1992; Krijgsman et al., 1999, 2001). In some basins (e.g., Nijar Basin in Spain), however, the Lago-Mare sediments are equivalent to and interbedded with the Upper Evaporites (Fortuin and Krijgsman, 2003). The Mediterranean Messinian sequence is covered by the marine sediments deposited during the abrupt flooding of the Mediterranean in the earliest Zanclean. The cause of fresh water input during the deposition of the Lago Mare sediments is controversial; a rapid freshwater influx from Paratethys (Hsü et al., 1973; Cita et al., 1978), or orbitally forced climate change and

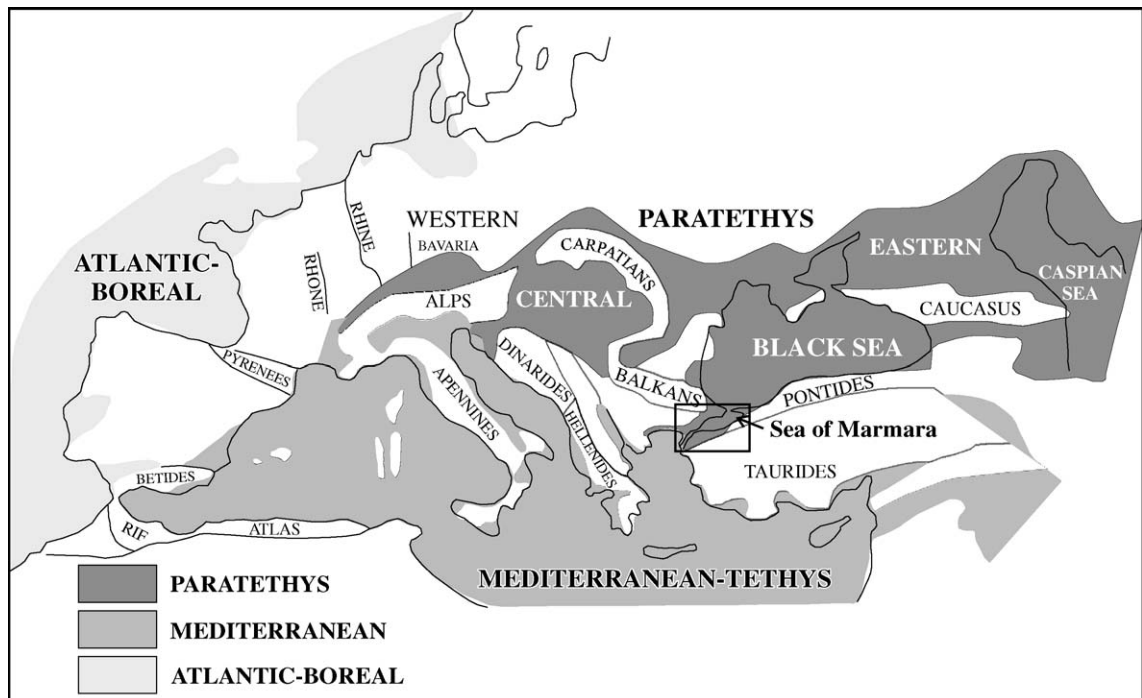


Fig. 1. Map showing the location of Sea of Marmara region, NW Turkey in relation to Paratethys and Mediterranean during the Neogene (modified from Rögl and Steininger, 1983).

increased river runoff (Krijgsman et al., 2001; Rouchy et al., 2001) are among the proposed hypotheses.

The Sea of Marmara region is in a strategic location to test these hypotheses. It is also part of a tectonically active region, which has been influenced by the dextral North Anatolian Fault (NAF) and N–S extensional regime of the Aegean since the Neogene (Fig. 2; McKenzie, 1972; Dewey and Şengör, 1979; Le Pichon and Angelier, 1981; Görür et al., 1997; Çağatay et al., 1999). The NAF developed as a broad shear zone in the Sea of Marmara region during the late Serravallian (Şengör et al., 1985), and continued to play a decisive role over the paleogeographic evolution of the region and Paratethyan–Mediterranean connectivity during the Neogene. During the Pliocene, intense tectonic dextral strike-slip activity along the NAF caused the development of deep basins, push-up structures and block rotations in the Sea of Marmara region (Şaroğlu, 1988; Görür et al., 1997; Sakıncı et al., 1999; Okay et al., 1999; Le Pichon et al., 2001; Şengör et al., 2004). The E–W trending grabens in western Anatolia and the Aegean also formed mainly since the Pliocene (Görür et al., 1997; Çağatay et al., 1999; Yılmaz et al., 2000).

The Late Miocene sediments in the Sea of Marmara region are in general characterized by siliciclastic shoreline sediments and bioclastic carbonates containing brackish-water mollusks and ostracods of Paratethyan affinities (Gillet et al., 1978; Sümengen et al., 1987; Siyako et al., 1989; Görür et al., 1997; Sakıncı et al.,

1999; Görür et al., 2000; Sakıncı and Yaltrak, 2005). The bioclastic carbonates with an endemic Paratethyan fauna (Alçıtepe Formation) are widely distributed around the Sea of Marmara and along the northern coast of the Gulf of Saros (Figs. 3 and 4). These rocks extend further south covering wide regions in the Aegean regions (Papp et al., 1978; Papp and Steininger, 1979). In the Gulf of Saros area, which was a part of the Aegean Sea since the early Neogene, Pontian brackish-water carbonates, sandstones and mudstones are interbedded with normal marine *Ostrea*-bearing units (Figs. 5 and 6; Ternek, 1949). Except for the *Ostrea*-bearing interbeds in the Gulf of Saros region, on the basis of their faunal content and broadly constrained age, the fresh- to brackish-water carbonates and their clastic interbeds in the Sea of Marmara regions are interpreted as being coeval with the sediments attributed to the Upper Evaporite–Lago Mare sequence of the Mediterranean.

The main objectives of this paper are to investigate the stratigraphic records of connectivity between Paratethys and the Mediterranean and the source of fresh- to brackish-water in the Sea of Marmara region during the Messinian. To fulfill these objectives we carried out stratigraphic analysis of outcrop sections in the Gelibolu and Biga peninsulas in the western Marmara region (Fig. 4). We studied in detail a section of the Alçıtepe Formation of Pontian age at Yenimahalle (7 km south of İntepe and 18 km south of Çanakkale) involving biolitho-stratigraphic, magnetostratigraphic

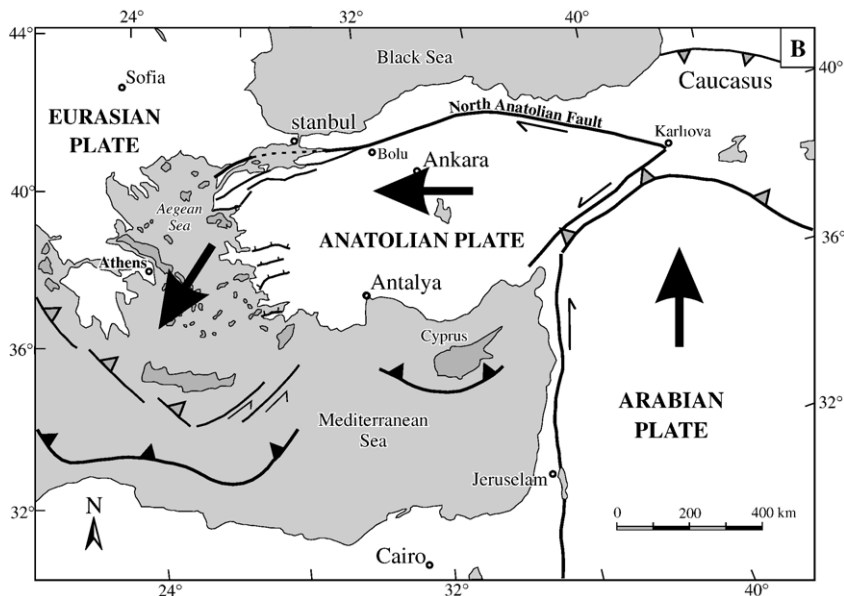


Fig. 2. Map showing main tectonic elements of eastern Mediterranean regions (modified from McKenzie, 1972; Şengör et al., 1985; Okay et al., 1999).

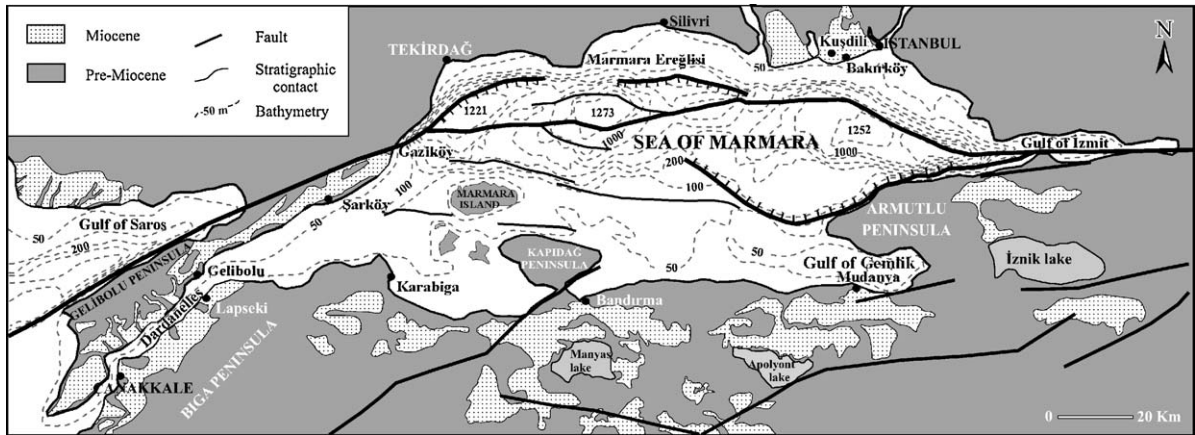


Fig. 3. Map showing outcrops of Miocene rocks in the Sea of Marmara regions.

and strontium isotope ($^{87}\text{Sr}/^{86}\text{Sr}$) analyses. We also studied the Late Miocene stratigraphy of the northern coast of the Gulf of Saros as part of the Aegean and compared it with that of the western Sea of Marmara region (Fig. 4). Such a comparison is important for correlating events in the Sea of Marmara region and the Mediterranean.

2. Laboratory methods

Ostracods from the Alçıtepe Formation exposed in the Yenimahalle Section on the Çanakkale–İzmir road (Fig. 4) were picked from washed and sieved whole rock samples and repeatedly ultrasonicated in 18.2 MΩ Milli-Q water to remove clay particles. The upper part of the section (Truva and Tevfikiye formations) did not yield sufficient unaltered shells for the isotope analysis. The $>63\ \mu\text{m}$ material was studied under binocular microscope for species identification and counting. Ostracods were studied under scanning electron microscope (SEM) to check for any visual evidence of diagenetic alteration.

Ultrasonicated ostracod valves were leached with 1 N ammonium acetate (Gorokhov et al., 1995) to remove readily exchangeable Sr, and then dissolved in 2.5 M HCl (Flecker and Ellam, 1999). Sr was separated using conventional cation exchange procedures. Total procedure blanks were below 1 ng, which is negligible compared to the $>1\ \mu\text{g}$ Sr sample sizes used. Samples were loaded on Re filaments with a Ta activator similar to that described by Birck (1986), but prepared from high purity Ta metal. Sr isotope ratios were measured on a VG Sector 54-30 mass spectrometer in dynamic multi-collection mode. Instrumental mass fractionation was corrected using an exponential law and $^{86}\text{Sr}/^{88}\text{Sr}=0.1194$. NIST SRM987 values over the period the measurements

were carried out were: $^{87}\text{Sr}/^{86}\text{Sr}=0.710252\pm 20\ (2\ \sigma\ \text{SD})$, $n=27$.

To establish a magnetostratigraphy for the Yenimahalle Section, a total of 36 cores from 20 different levels were sampled with a water-cooled drill and oriented with a compass. The paleomagnetic sampling covered a wide variety of lithologies and comprised a stratigraphic interval of 35 m. In the laboratory, the cores were cut into standard size ($11.4\ \text{cm}^3$) specimens. One specimen per sampling level was thermally demagnetised according to standard techniques (e.g. Langereis et al., 1989) with small temperature increments of 30–50 °C up to a maximum temperature of 400 °C, in a magnetically shielded, laboratory-built furnace. The natural remanent magnetisation (NRM) was measured after each increment on a 2G Enterprises DC SQUID cryogenic magnetometer. Least-square analysis (Kirschvink, 1980) was applied to determine the component directions of the NRM, chosen by inspection of vector end-point demagnetisation diagrams (Zijderveld, 1967).

3. Results and discussion

3.1. Stratigraphic framework

The Neogene succession in the Sea of Marmara and northern Gulf of Saros regions starts with fluvio-lacustrine sediments of the Gazhanedere Formation above a pre-Early Miocene unconformity (Fig. 5). In the Biga Peninsula in the western Marmara region, the first stratigraphic unit above the unconformity consists of a fanglomerate. In the Gelibolu Peninsula, the northern margin of the Gulf of Saros and Gaziköy–Şarköy areas, the Gazhanedere Formation is composed of vari-colored (green, grey, red and purple) sandstone and mudstone

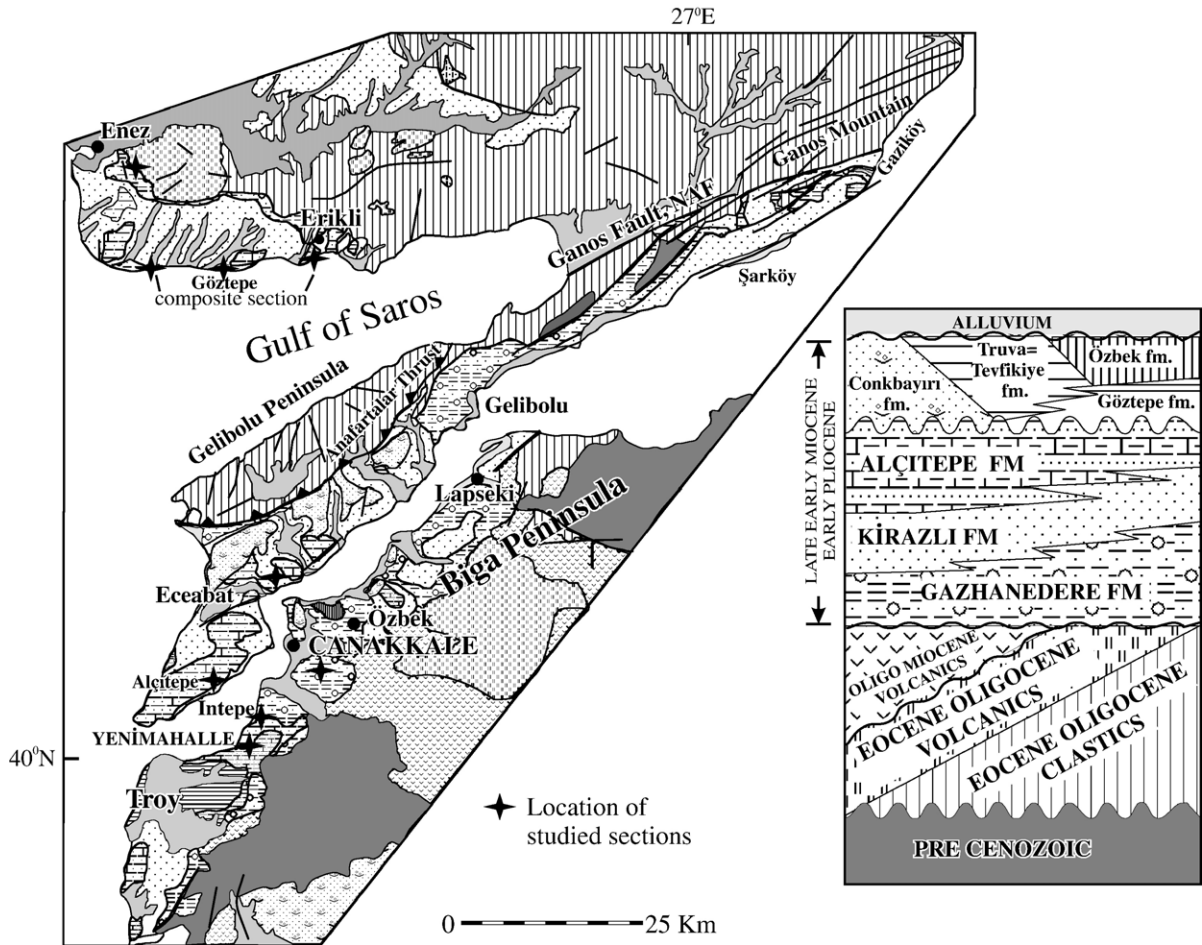


Fig. 4. Geological map and simplified stratigraphic column section showing the locations of the studied sections in the Çanakkale (Dardanelles) and Gulf of Saros area (modified from Sümengen et al., 1987; Siyako et al., 1989; Sakiñç et al., 1999). Pattern for different formations on the map is the same as that on the stratigraphic section.

with some rare conglomerate interbeds, coal seams and marl. The marls and mudstone beds contain the fresh water bivalve, *Unio* sp. The Gazhanedere Formation is about 400-m thick in the Gaziköy–Şarköy area, with micro-mammal fauna indicating a late Orleanian age (equivalent to early Serravallian in Mediterranean chronostratigraphy; Ünay and de Bruijn, 1984) and an Astaracian age (early Burdigalian to late Serravallian) in the Gelibolu Peninsula (Sümengen et al., 1987).

The Kirazlı Formation conformably overlies the Gazhanedere Formation and consists predominantly of cross-bedded, yellow to buff sandstones with rare mudstone and conglomeratic intercalations and *Ostrea*-banks in the northern Gulf of Saros. In the Gelibolu Peninsula it is about 200 m thick, and its base is dated as Vallesian age (equivalent of Tortonian in Mediterranean chronostratigraphy; Ünay and de Bruijn, 1984; Sümengen et al., 1987; Kaya, 1989). Around Çanakkale and

Lapseki in the western Sea of Marmara region, the formation consists of cross-bedded sandstones at the base, *Unio*-bearing green mudstones and sandstones in the middle, and *Macra*-bearing sandstones towards the top (Sakiñç et al., 1999; Sakiñç and Yaltrak, 2005). On the northern Saros coast, the Kirazlı Formation consists of fine to medium, cross-bedded sandstones interbedded with 5 cm to 6 m thick *Ostrea*-banks in the upper part and some gypsum lenses in lower part. The *Ostrea* banks contain a euryhaline molluscan fauna of mainly *Ostrea cuculata* and *O. gingensis*, and are underlain by *Cardium edule*-bearing sandstone.

The Alçıtepe Formation is widely distributed in the Sea of Marmara and the Gulf of Saros regions (Fig. 4; Sayar, 1987; Sümengen et al., 1987; Siyako et al., 1989; Görür et al., 1997; Çağatay et al., 1999; Sakiñç et al., 1999; Görür et al., 2000). Its type locality is in Alçıtepe, west of the Gelibolu Peninsula (Önem, 1974). It is assigned different

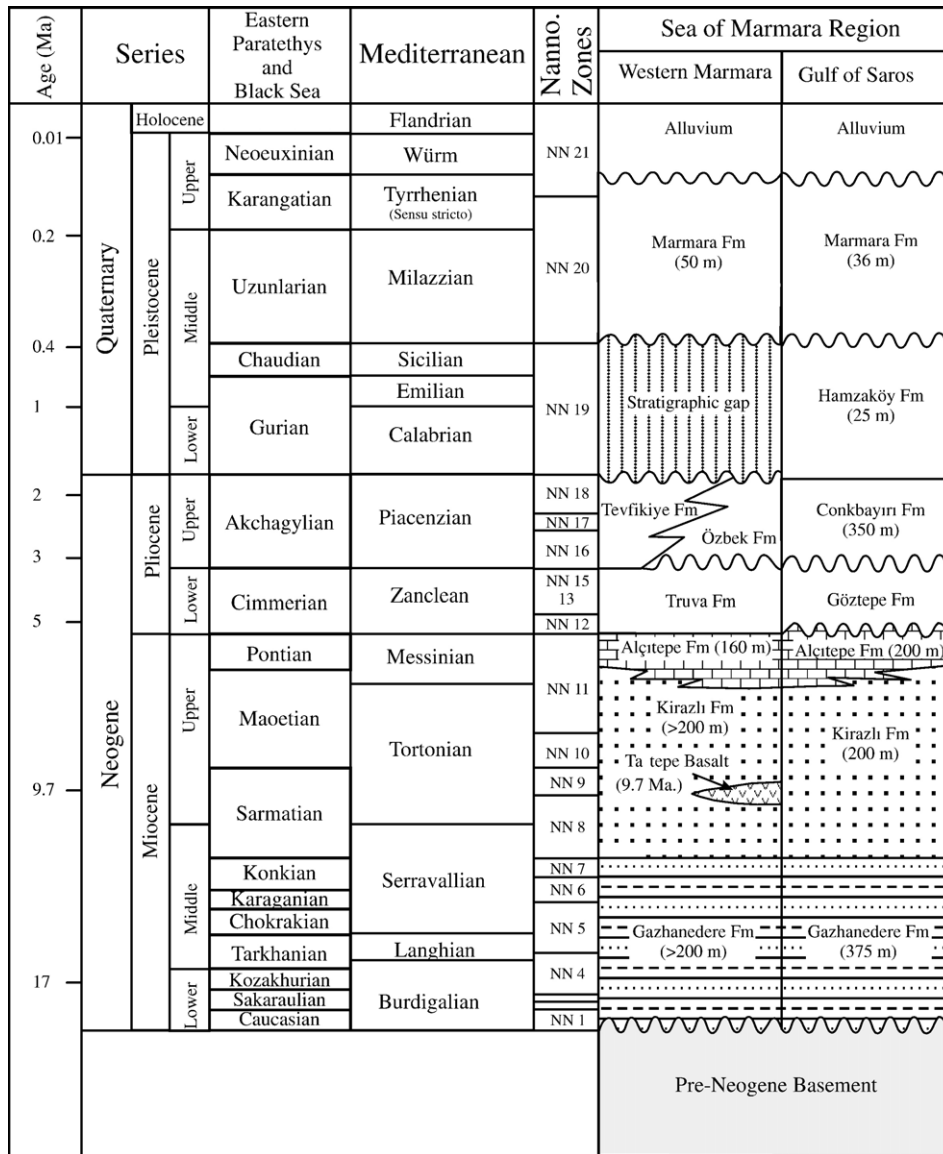


Fig. 5. Stratigraphy of Neogene sections in NW Turkey.

names in other parts of the region, such as the Bakırköy Formation in west of İstanbul and Bayraktepe Formation, south of Çanakale (Şentürk et al., 1987; Sümengen et al., 1987). The formation lies conformably over the Kirazlı Formation into which it appears to be laterally and vertically transitional in the northern coast of the Gulf of Saros area. Its ostracod fauna in the lower part of the formation in its type locality yielded a Pannonian age and the mammal fauna from its uppermost part produced a Turolian age for the formation (Şentürk et al., 1987; Kaya, 1989). The age of the formation in the northern coast of the Gulf of Saros is between late Sarmatian to Pontian (Fig. 5; Sümengen et al., 1987).

The Alçıtepe Formation on the northern coast of the Gulf of Saros and Gelibolu and Biga peninsulas shows different facies characteristics. Along the northern coast of the Gulf of Saros, the formation consists of *Mastra*-bearing bioclastic and cross-bedded limestones, interbedded with *Cardium*-bearing sandstones and *Ostrea* banks (Fig. 6). This change in the fauna of the sandstone interbeds from *Cardium* to *Ostrea* indicates fluctuations in the salinity from brackish water to normal marine conditions. The Alçıtepe Formation in the Gulf of Saros is overlain with an erosional unconformity by the Göztepe Formation of Zanclean age (zone NN12), consisting of shallow marine siltstone and sandstone

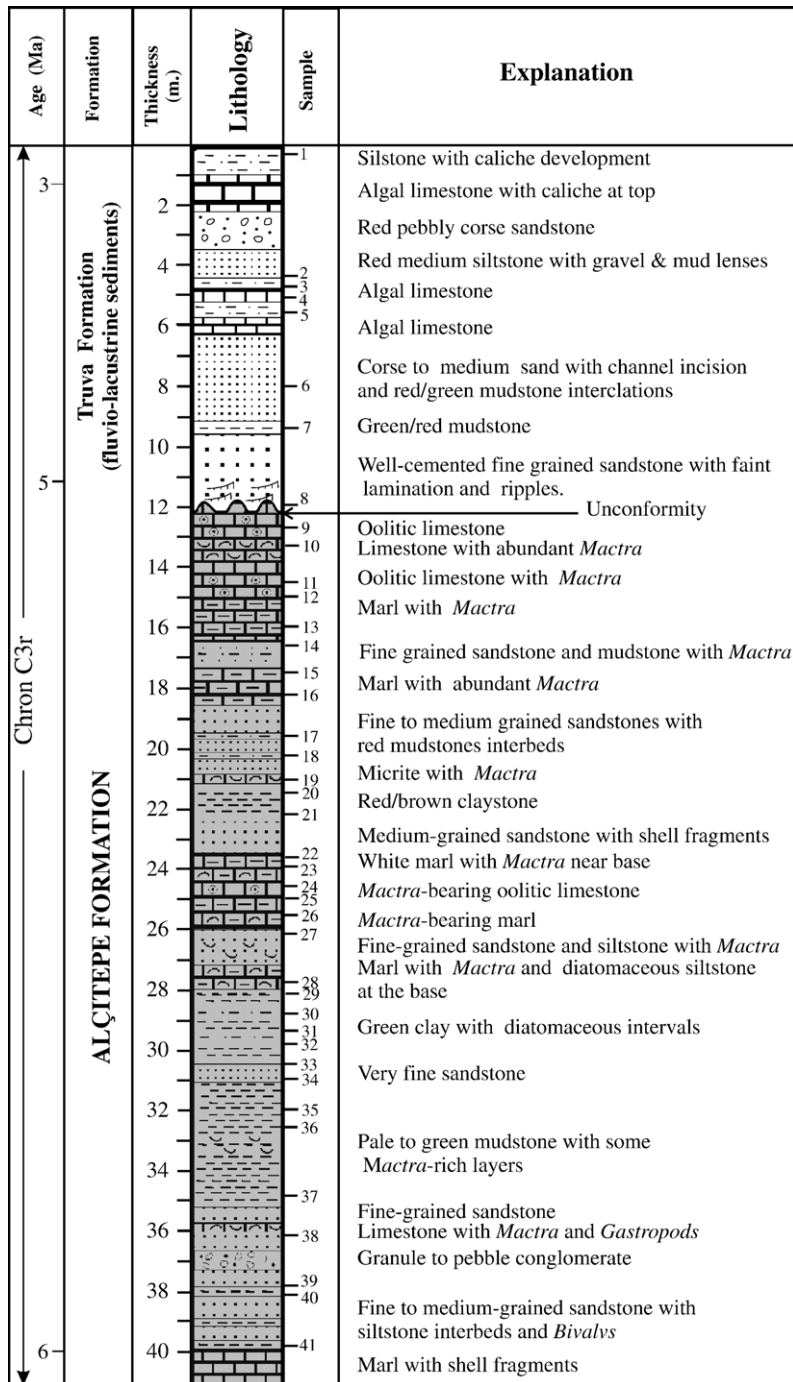


Fig. 7. Simplified stratigraphic section at Yenimahalle, south of Çanakkale.

bivalve fauna (*Paradacna abichi* sinz, *Dreissena* sp.). In the upper part, this limestone unit contains *Cardium* sp., *C. edulis*, and *Paradacna*, and is overlain by an *Ostrea*-bank unit.

The Alçitepe Formation in the Gelibolu and Biga peninsulas consists of mudstone and marl in the lower part

and bioclastic and oolitic limestones with marl, mudstone, and sandstone intercalations in the upper part (Fig. 4). It conformably overlies the *Unio*-bearing mudstones and *Maetra*-bearing sandstones of the Kirazlı Formation, and is overlain unconformably by the alluvial fan deposits of the Conkbayırı Formation in the Gelibolu Peninsula, the

Table 1
Results of ostracod analysis in Yenimahalle Section, south of Çanakkale

Sample number	Depth (m) from section top	Species (number of ostracod valves)
I-8	12.20	<i>Xestoleberis</i> sp. (29) <i>Cyprideis torosa</i> Jones, 1850 (1) <i>Heterocypris salina salina</i> Brady, 1868 (1)
I-9	12.50	<i>Cyprideis torosa</i> Jones, 1850 (18) <i>Xestoleberis</i> sp. (15) <i>Loxoconcha</i> sp. (7)
I-12	14.90	<i>Xestoleberis</i> sp. (21) <i>Limnocythere</i> sp. (75) <i>Cyprideis torosa</i> Jones, 1850 (4) <i>Loxoconcha</i> sp. (8)
I-16	18.25	<i>Cyprideis torosa</i> Jones, 1850 (12) <i>Limnocythere</i> sp. (27) <i>Xestoleberis</i> sp. 1 (19) <i>Loxoconcha</i> sp. (5) <i>Xestoleberis</i> sp. 2 (6) <i>Loxoconcha cartaensis</i> Tunoğlu, 1984 (7)
I-19	20.75	<i>Xestoleberis</i> sp. 1 (35) <i>Loxoconcha</i> sp. (23) <i>Limnocythere</i> sp. (7) <i>Cyprideis torosa</i> Jones, 1850 (6) <i>Cyprideis pannonica</i> (Mehes, 1908) (6)
I-20	21.25	<i>Cyprideis torosa</i> Jones, 1850 (54)
I-23	23.50	<i>Cyprideis pannonica</i> (Mehes, 1908) (35) <i>Heterocypris salina salina</i> Brady, 1868 (15)
I-24	24.65	<i>Loxoconcha</i> sp. (12) <i>Xestoleberis</i> sp. (11) <i>Cyprideis torosa</i> Jones, 1850 (7)
I-25	25.10	Sterile
I-26	25.40	<i>Loxoconcha</i> sp. (10) <i>Xestoleberis</i> sp. (13) <i>Cyprideis torosa</i> Jones, 1850 (6)
I-27	25.90	<i>Xestoleberis</i> sp. 1 (13) <i>Xestoleberis</i> sp. 2 (12) <i>Limnocythere</i> sp. (8) <i>Loxoconcha</i> sp. (2) <i>Cyprideis</i> sp. (3)
I-28	27.75	Sterile
I-29	27.95	<i>Loxoconcha</i> sp. 1 (5) <i>Loxoconcha</i> sp. 2 (3) <i>Xestoleberis</i> sp. 1 (3) <i>Xestoleberis</i> sp. 2 (1)
I-30	28.75	Sterile
I-32	29.35	Sterile
I-37	34.80	<i>Heterocypris salina salina</i> Brady, 1868 (75) <i>Cyprideis torosa</i> Jones, 1850 (22)
I-38	36.10	<i>Heterocypris salina salina</i> Brady, 1868 (32) <i>Cyprideis torosa</i> Jones, 1850 (55) <i>Candona</i> sp. (18)
I-39	37.50	<i>Cyprideis torosa</i> (337) abundance zone <i>Heterocypris salina salina</i> Brady, 1868 (112) <i>Cyprideis hexatuberculata</i> Ünal, 1996 (57) <i>Ilyocypris bradyi</i> Sars, 1890 (25)
I-40	37.85	<i>Cyprideis torosa</i> Jones, 1850 (27) <i>Heterocypris salina salina</i> Brady, 1868 (119) <i>Cyprideis tetratuberculata</i> (9)
I-41	39.65	<i>Candona neglecta</i> SARS, 1888 (26) <i>Candona candida</i> (O.F. Müller, 1776) (19) <i>Heterocypris salina salina</i> Brady, 1868 (33) <i>Cyprideis torosa</i> Jones, 1850 (29) <i>Ilyocypris</i> sp. (5) <i>Darwinula</i> sp. ? (1)

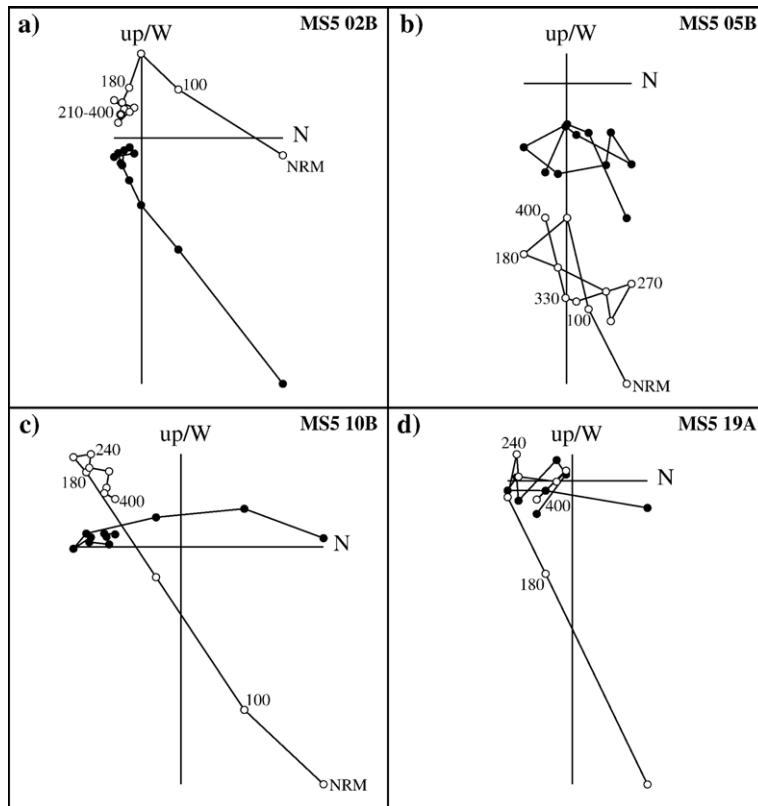


Fig. 8. Thermal magnetization of selected samples from Yenimahalle Section. Closed (open) circles denote the projection on the horizontal (vertical) plane. Values are temperatures in degrees Celsius.

fluvio-lacustrine sediments of the Kimmerian (5.5–3.2 Ma) Truva and Tevfikiye formations in the Biga Peninsula, and by the marine sediments of the late Akchagilian Özbek Formation east of Çanakkale (Görür et al., 2000). The Alçitepe Formation in the Gelibolu and Biga peninsulas has rich brackish-water bivalve (*Maetra* sp., gastropods) and ostracod (*Cyprideis pannonica*, *Cyprideis torosa*, *Candona neglecta*, *C. candida*, *Loxoconcha* sp., *Heterocypris salina salina*) faunas endemic to Paratethys.

3.2. Litho- and bio-stratigraphy of the Yenimahalle Section

The 41-m thick sampled part of the Yenimahalle Section along the Çanakkale–İzmir road comprises the upper part of the Alçitepe Formation and lower part of the Truva Formation (Fig. 7). Here, the Alçitepe Formation is about 30-m thick and consists of beige and white *Maetra*-bearing marl and limestone interbedded with green, grey and less commonly red mudstones and sandstones. The sandstones and mudrocks are more abundant in the lower part of the section, which also contains a few finely-

laminated diatomite beds. Oolitic limestone units are more common in top 2.5 m of the formation. The thickness of individual marl and limestone units vary from 0.2 to 2.5 m. The Alçitepe Formation is overlain with an erosional contact by fluvio-lacustrine sands, shales and algal (red algae) and oolitic limestones of the Truva Formation, which contains several caliche horizons.

The Alçitepe Formation contains a rich bivalve and ostracod fauna. However, the bivalves are almost completely dissolved with only their moulds remaining. It is therefore impossible to identify *Maetra* species. The ostracod fauna is well preserved and includes an abundance of brackish-water (*Heterocypris salina salina*, *Cyprideis torosa*) and fresh-water (*Candona neglecta*, *Candona candida*, *Ilyocypris* sp., *Darwinula* sp., *Ilyocypris bardyi*) species (Table 1). There is an abundance zone at 37.7 m and barren zones at 29–29.5, 25 and 26.5–27.5 m intervals (Fig. 9). Starting from about 28-m upwards, species characteristic of high salinities similar to that of seawater (*Xestoleberis* sp., *Loxoconcha* sp., *Loxoconcha cartaensis*) appear for the first time and become abundant in most of the upper part of the Alçitepe Formation. In the upper part, brackish salinity

Table 2

Sr isotope data from the Yenimahalle Section, south coast of the Çanakkale Strait (Dardanelles), Sea of Marmara region

Sample number	Depth (m) from section top	$^{87}\text{Sr}/^{86}\text{Sr}$	2σ ($\times 10^6$)
I-16	18.3	0.70827	19.83156
I-20	21.3	0.708379	21.25136
I-26	25.2	0.708281	19.83187
I-37	34.7	0.708729	18.42695
I-37	34.7	0.708723	15.59191
I-37	34.7	0.708656	21.25968
I-37	34.7	0.708715	15.59173
I-39	37.6	0.708836	17.01206

species (*Cyprideis torosa*, *Heterocypris salina salina*, *Cyprideis pannonica*) are abundant in the 21–24 m interval and fresh water species (*Candona neglecta*, *Candona candida*, *Ilyocypris* sp.) in 19–12.5 m interval (Fig. 9, Table 1). The overall similarity of the ostracod fauna, and especially the common presence of *Loxiconcha cartaensis*, in the Yenimahalle Section and the Pontian succession of the Eastern Paratethys in the Sinop Peninsula on the Black Sea coast of Turkey, strongly suggests a Pontian age for the Alçitepe Formation of the Yenimahalle Section (Tunoğlu, 1984; Tunoğlu and Gökçen, 1985, 1991).

3.3. Magnetostratigraphy of the Yenimahalle Section

NRM intensities of the Yenimahalle Section show significant differences depending on lithology and range from 0.2 to 13 mA/m. In general, the best results are obtained from red and grey silts and clays of the Alçitepe

and Truva formations, while beige-white marls give less reliable signals. Thermal demagnetisation (Zijderveld diagrams) reveals that in all samples a normal polarity component is removed at temperatures between 100 and 240 °C (Fig. 8). This relatively low-temperature component has a present-day field direction and can be regarded as a secondary overprint, probably caused by viscous behaviour and sub-recent weathering. Demagnetisation at higher temperatures shows that a second component is gradually removed at temperatures higher than 240 °C, which can be interpreted as the primary component or the characteristic remanent magnetisation (ChRM). The thermal demagnetisation diagrams are of mixed quality throughout the section. In approximately 50% of the sampled levels, the ChRM shows a progressive decay toward the origin (Fig. 8a, c). This component is always reversed polarity and directions can be accurately determined. Other samples, especially the ones with a relatively low NRM intensity, do not show a decay toward the origin but tend to move successively toward the reversed quadrant of the Zijderveld diagram (Fig. 8b, d). In these cases it can be assumed that the primary (reversed) component is partly overprinted by a secondary (normal) component with overlapping blocking temperature spectra. Based on the thermal demagnetisation results, there is no evidence for normal polarities in the temperature range 240–400 °C.

Thermal demagnetisation data indicates that the sampled interval of the Alçitepe and Truva formations in the Yenimahalle Section is entirely of reversed polarity. Consequently, no age constraints can be given to the Yenimahalle Section based on the paleomagnetic results.

Table 3

$^{87}\text{Sr}/^{86}\text{Sr}$ ratio of Yenimahalle Section and various waters and fluids

Material analysed	$^{87}\text{Sr}/^{86}\text{Sr}$	Reference
Yenimahalle section lower part (32–37 m interval from top)	0.708656–0.708836	This study
Yenimahalle section middle part (18–28 m interval from top)	0.708270–0.708379	This study
Kocasu (south Marmara)	0.7084	Major et al. (2004)
Gönen (south Marmara)	0.7078	Major et al. (2004)
Karabiga (south Marmara)	0.7068	Major et al. (2004)
Present-day ocean water	0.709172	Palmer and Edmond (1989)
Black Sea at present	0.793	Cox and Faure (1974)
Average ocean water during deposition of Lower Evaporites	0.708999	(Howarth and McArthur, 1997)
Average ocean water during deposition of Upper Evaporites	0.709012	(Howarth and McArthur, 1997)
Messinian sea water	0.708983–0.709028	Howarth and McArthur (1997)
Messinian brine	0.70865	McCulloch and De Deckker (1989)
Rhone river water	0.708700	(Brass, 1976; Albarède and Michard, 1987)
Nile river water	0.706000	(Brass, 1976; Albarède and Michard, 1987)
Global river water	0.712000	Palmer and Edmond (1989)
Danube	0.7089	Palmer and Edmond (1989)
Dniepr	0.7084	Palmer and Edmond (1989)
Caspian Sea	0.7085	Palmer and Edmond (1989)
Oceanic ridge hydrothermal fluids	0.7029–0.7047	Palmer and Edmond (1989)

However, considering the biostratigraphic age for the Alçitepe Formation in Yenimahalle is Pontian, only one magnetostratigraphic correlation seems valid. The lower and upper Pontian boundaries are determined in the Black Sea region in the Taman peninsula of Russia at the top of chron C3An.1n (~6.1 Ma) and the top of C3n.4n (~5.0 Ma), respectively (Trubikhin, 1990; Popov et al., 1996). Magnetostratigraphic results from the Focsani basin of Romania indicate that the Pontian stage in the Carpathian foredeep extends from the base of chron C3r to the base of C3n.3n (5.9 to 4.9 Ma) (Vasiliev et al., 2004). Consequently, the long reversed polarity interval of the Yenimahalle Section most likely corresponds to chron C3r, with an age range of 6.04–5.24 Ma. This range encompasses the entire Messinian salinity crisis (5.96–5.33 Ma).

3.4. Strontium isotopes in the Yenimahalle Section

All the strontium isotope ratios of ostracod valves collected from the Yenimahalle Section are lower than coeval Messinian global ocean values (McArthur et al., 2001; Tables 2 and 3; Fig. 9). The samples from the middle part of the section (i.e., upper part of the Alçitepe Formation) contain ostracods that require salinities similar to that of seawater. These have Sr isotope ratios lower than those at the bottom of the section where brackish water ostracod species predominate and saline forms are absent.

If the increase in salinity inferred from the appearance of ostracods usually considered to be of marine affinity was controlled by an influx of water with marine salinities from the global ocean, the Sr-isotope ratios of these

ostracods should be closer to Messinian global ocean values (Table 2) than the brackish and fresh water ostracod samples. The reverse is true (Fig. 9). If, however, the influx of salty water came from the Mediterranean, then Yenimahalle ostracod Sr isotope ratios should be deflected towards Mediterranean Sr isotope ratios. Mediterranean $^{87}\text{Sr}/^{86}\text{Sr}$ during the MSC varies significantly from ratios within error of Messinian ocean water (Table 2) to lower ratios (ca. 0.7086, Müller and Mueller, 1991) from Upper Evaporites and Lago Mare samples (Flecker et al., 2002). Even these lowest Sr isotope ratios are significantly higher than the values measured in the middle part of the Yenimahalle Section. An influx from the isolated Mediterranean alone therefore cannot account for the Sr isotope values measured at Yenimahalle. Further, the correlation between low Sr isotope ratios and higher salinities indicates that the variation in salinity reflected in the ostracod population was not controlled simply by an influx of seawater. Independent behaviour of salinity and ocean water pulses are documented from Late Miocene Mediterranean sequences (Flecker et al., 2002; Flecker and Ellam, 2006-this volume). Typically, this occurs in systems with only limited connection to the ocean and where the evaporation flux exceeds precipitation.

The Sr isotope ratio of Paratethys during the Miocene is poorly constrained. However, we can assume that it reflected the flux of river water feeding the basin. The $^{87}\text{Sr}/^{86}\text{Sr}$ values for the lower part of the Alçitepe Formation at Yenimahalle are lower than that of Sr isotope ratio of the River Danube (Table 2), which today supplies about half of the total freshwater runoff to the Black Sea (Shimkus and Trimonis, 1974). Other large

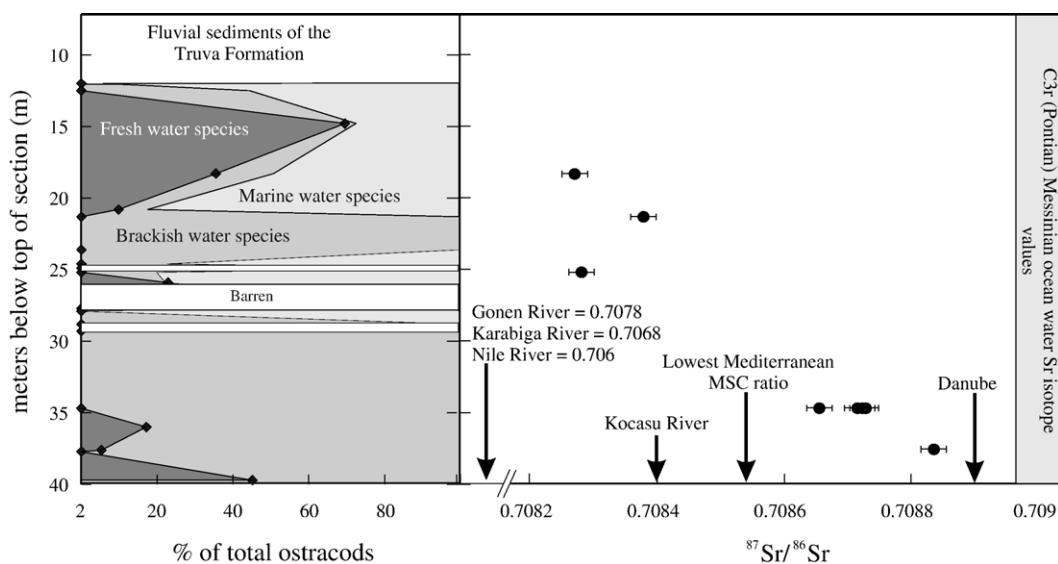


Fig. 9. Sr isotope ratio of ostracod shells and abundance of ostracod species in Yenimahalle Section. See Table 3 for sources of Sr isotope data.

ivers emptying into the Black Sea (Dniepr, Don) and the Caspian Sea have lower values than the River Danube (Table 2; Fig. 9), but without knowing run-off fluxes of these rivers for the Messinian stage from global climate simulations it is not possible to assess the impact these rivers may have had on lowering the Sr isotope ratio of Paratethyan water. Nevertheless, overall range of the Sr-isotope ratio of these waters probably represents Paratethyan brackish waters and overlaps well with the Sr-isotope values in the lower part of the Yenimahalle Section. These Sr-isotope data therefore suggests that during the deposition of the lower part of the section, the Yenimahalle area was probably connected with the Eastern Paratethys. This conclusion is corroborated by the faunal content of the section, which is endemic to the Paratethys.

The $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the fresh water input into the present day Sea of Marmara by the southern rivers ranges between 0.7068 and 0.7084 (Major et al., 2004). This range overlaps with the low Sr isotope ratios in upper part of the Alçitepe Formation (Tables 2 and 3; Fig. 9). One interpretation of the Yenimahalle Sr isotope record is therefore that at this time, the western Sea of Marmara region was dominated by local river input.

3.5. Palaeogeographic evolution

In the late Serravallian, shallow marine conditions were established in the northeast Aegean, as shown by the deposition of siliciclastic facies containing *Ostrea* and foraminifera in the upper part of the Kirazlı Formation in the northern coastal area of the Gulf of Saros and Gaziköy–Şarköy areas. These facies briefly extended eastwards into the Marmara region along the northern part

of the Sea of Marmara, following the broad shear zone of the incipient NAF (Fig. 10; Şengör et al., 1985; Görür et al., 1997). The Messinian Alçitepe Formation in the Gulf of Saros area shows evidence of marine incursions by the presence of some *Ostrea* banks and *Cardium*-bearing sands in a mainly *Maestra*-bearing limestone sequence deposited under brackish-water conditions.

In contrast, the Kirazlı and Alçitepe Formations south of Çanakkale and in the Gelibolu Peninsula do not show any clear faunal evidence of the Mediterranean incursion from the late Serravallian to Late Miocene. Instead, here, we see a late Serravallian to late Messinian siliciclastic to carbonate sequence that is mainly characterised by a brackish-water, endemic Paratethyan fauna. Sr isotope data from the Yenimahalle Section suggest that the Sea of Marmara area was dominated by local fluvial run-off. The independence of Sr isotope data with respect to salinity fluctuations inferred from ostracod assemblage changes suggests that salinity was controlled by changes to the hydrologic balance in a setting where evaporation exceeded precipitation. Connection with the Paratethys is possible on basis of the Sr-isotope data during the deposition of the lower part of the sampled Yenimahalle Section. The extensive *Maestra*-bearing limestone facies in the northern Gulf of Saros area suggests that during the late Messinian, there was a Paratethyan outflow that probably extended westwards and southwards into the Aegean (Fig. 11; see also Papp and Steininger, 1979; Syrides, 1998; Popov et al., 2004; Sakıncı and Yaltrak, 2005). The connection of Sea of Marmara region with both the Paratethys and Mediterranean was severed during the early Pliocene, and the fluvio-lacustrine deposits of the Truva and

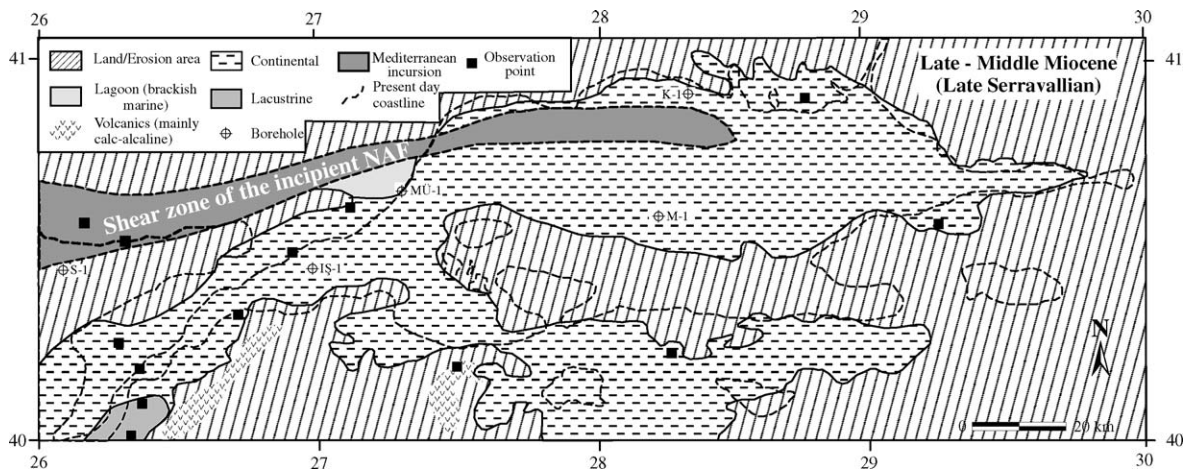


Fig. 10. Middle Miocene palaeogeographic map of Sea of Marmara region (modified after Görür et al., 1997). Observation point denotes a location where a stratigraphic section is measured, or a significant stratigraphic information pertaining to palaeogeography is obtained. Note that petroleum exploration boreholes provide additional subsurface information on general stratigraphy.

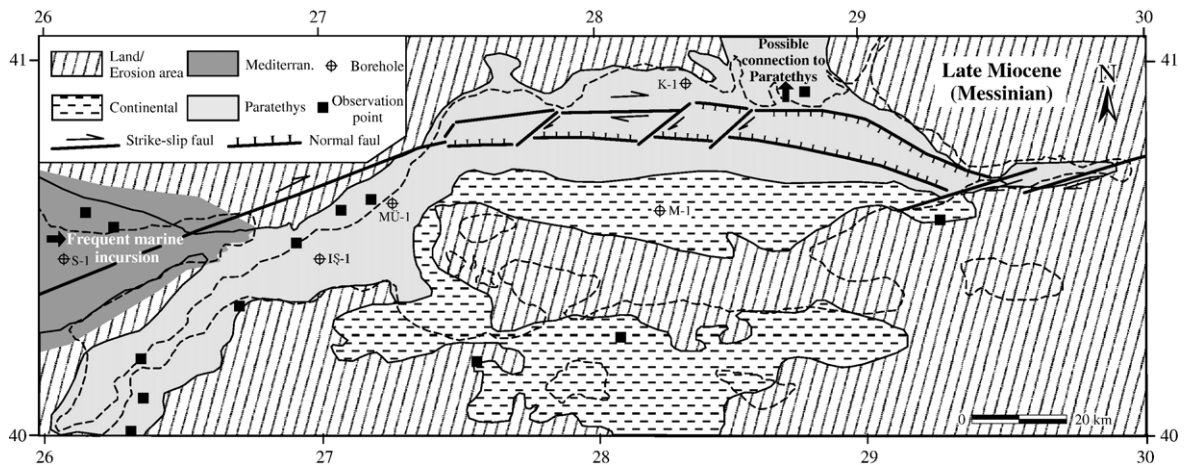


Fig. 11. Messinian palaeogeographic map of Sea of Marmara region (modified after Görür et al., 1997).

Tevfikiye formations were deposited in the region with an erosional unconformity (Görür et al., 1997). Despite the widespread transgression (the Zanclean flooding) in the Mediterranean, marine influence in the Sea of Marmara was obstructed by horsts and transpressional push-up structures generated by the NAF in the western Sea of Marmara region (Görür et al., 2000). It was not until the late Akchagilian (later Late Pliocene–early Pleistocene) that the Sea of Marmara basin was flooded by Mediterranean water (Toker and Şengüler, 1995) for the first time since the late Serravallian, depositing the Özbek Formation (Fig. 12; Görür et al., 1997). In Eastern Paratethys, the presence of NN11 and NN12 nannoplankton zones is strong evidence for two Mediterranean marine water influxes during the Late Miocene and Early Pliocene period (Semenenko and Olejnik, 1995; Marunteanu and Papaianopol, 1998).

The faunal and Sr isotope evidence presented here from the Sea of the Marmara region suggests that these Mediterranean water influxes accessed Eastern Paratethys via a gateway other than the Sea of Marmara region.

4. Conclusions

During the late Serravallian, shallow marine conditions were established in the northern Aegean and briefly affected the northwestern Sea of Marmara regions. The northern Aegean region was frequently flooded by marine waters during the rest of the Neogene, but in contrast, there is no clear faunal evidence of Mediterranean incursion into the western Marmara region (Gelibolu and Biga peninsulas) from the late Serravallian to Late Pliocene. Instead, this region is characterized by a Late

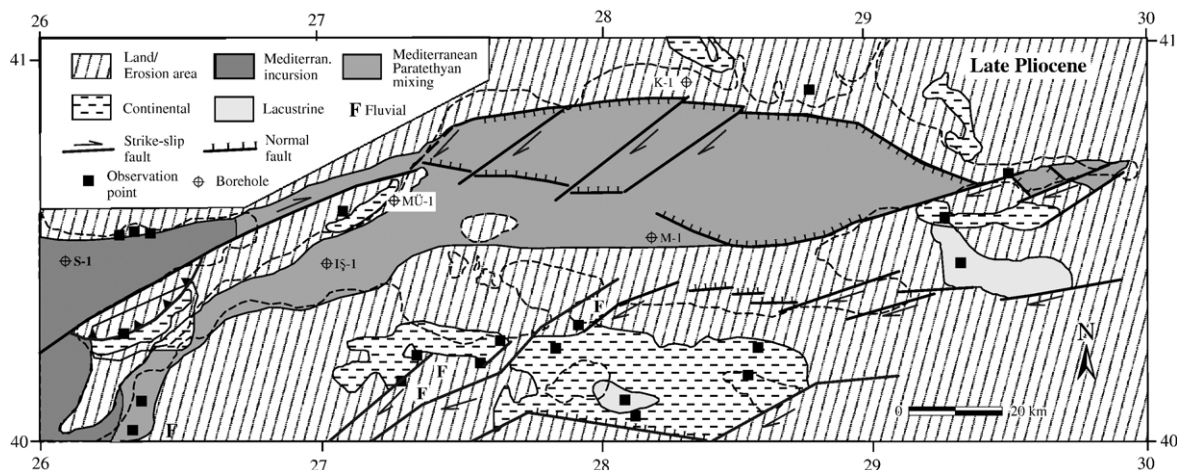


Fig. 12. Latest Pliocene palaeogeographic map of Sea of Marmara region (modified after Görür et al., 1997).

Miocene siliciclastic to carbonate sequence (Kirazlı and Alçitepe formations) containing predominantly a brackish-water fauna typical of Paratethys.

Magnetostratigraphic and Pontian biostratigraphic data from one key section of the Alçitepe Formation at Yenimahalle, south of Çanakkale, indicate that it was deposited during chron C3r (6.04–5.24 Ma). This corresponds to the time of the MSC in the Mediterranean.

The Sr-isotope data from the Yenimahalle Section suggest that (a) during the deposition of the Pontian Alçitepe Formation, exchange with the Mediterranean was reduced or terminated and that this area was dominated by local river water discharge and (b) the salinity of the Alçitepe Formation in the Sea of Marmara basin was controlled by the dominance of the evaporation flux, and not by fluctuation in exchange either with the global ocean or Mediterranean. Connection between Paratethys and part of the Sea of Marmara region cannot be excluded on the basis of the Sr isotope data. The presence of some ostracod species of marine affinity in the upper part of the Alçitepe Formation at Yenimahalle are considered as endemic, probably adopted to high salinities under continental conditions.

Faunal evidence suggests that during deposition the middle-upper part of the Alçitepe Formation, Paratethys extended into the Aegean and caused the deposition of the fresh-to brackish-water facies coeval with the Upper Evaporite–Lago Mare sequence in the Mediterranean. Connection between Marmara and Paratethys with the rest of the Mediterranean was not re-established until the Late Pliocene. This connection was the result of increased activity on the North Anatolian Fault and global sea level rise.

Acknowledgements

This work was funded by TUBITAK under project YDABÇAG 199Y077. Vinny Gallagher and Anne Kelly assisted with the Sr analysis which was carried out in SUERC, East Kilbride. RF acknowledges additional support from a Royal Society Wolfson Dorothy Hodgkin Fellowship. We thank the editors of this volume and two reviewers, Fabienne Orszag-Sperber and Fritz Steininger, for useful suggestions that improved the paper.

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