Mediterranean Marine Science

Vol 2, No 2 (2001)

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doi: 10.12681/mms.269

Mediterranean Marine Science
Vol. 2/2, 2001, 95-118

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Abstract

The Enez-Evros Delta, NE Aegean Sea, is located in one the most important wetlands in the world with its sandy offshore islands, abandoned channel mouths, sand-dunes, shoals, marshlands, saline lagoons and saltpans. It comprises very well developed sedimentary units and a prodelta lying on an older submarine delta. The present day elevations of the middle-late Pleistocene marine terraces indicate a regional tectonic uplift in the area. Due to lack of geophysical and bore hole data and partly due to its strategic position, the structural and stratigraphic features of the submarine extension of the delta are not known in detail. In this paper, Plio-Quaternary history of this delta and its submarine part on the Turkish shelf was explored by using high-resolution shallow reflection seismic profiles. The delta is formed by the alluvial deposits of the Enez-Evros River and shaped by their interaction with the sea. It takes place in front of a large and protected ancient bay which was filled rapidly over millennia. The sediments in the plateau off the river are principally pro-deltaic with muddy areas near the river mouths changing to muddy sand further out. The sea-level changes in Plio-Quaternary were characterised by three different seismic stratigraphic units on the folded Miocene limestone basement. In the late Pleistocene, the shelf area over an Upper Miocene basement was flooded during the Riss-Würm interglacial period, exposed in the Würm glacial stage, and flooded once again during the Holocene transgression.

Keywords: North Aegean Sea, Seismic stratigraphy, Coastal changes, Relative sea level.

Introduction

Six rivers discharge into the North Aegean Sea; Pinios, Axios, Aliakmon, Strymon, Nestos and Enez (Evros) (Fig. 1a). The Enez-Evros River (ER) is one of the longest (480 km) rivers in the Balkan Peninsula. In ancient times, it was first called Rhombus and later came to be known as Ebros. It is also known as Maritsa, Marica or Meric. The ER rises in the Rila Mountains and flows southeast across a fertile valley before forming a short (15 km) section of the Bulgaria-Greece border (Fig. 1b). The river then subsequently turns south at Edirne, Turkey, and flows southwest through the flat Thracian plain. Since 1923 its lower course has formed the 215-km boundary between Greece and Turkey, except for the small (12.5 km) Karagathc triangle. It is navigable for small boats as far as Edirne. Its

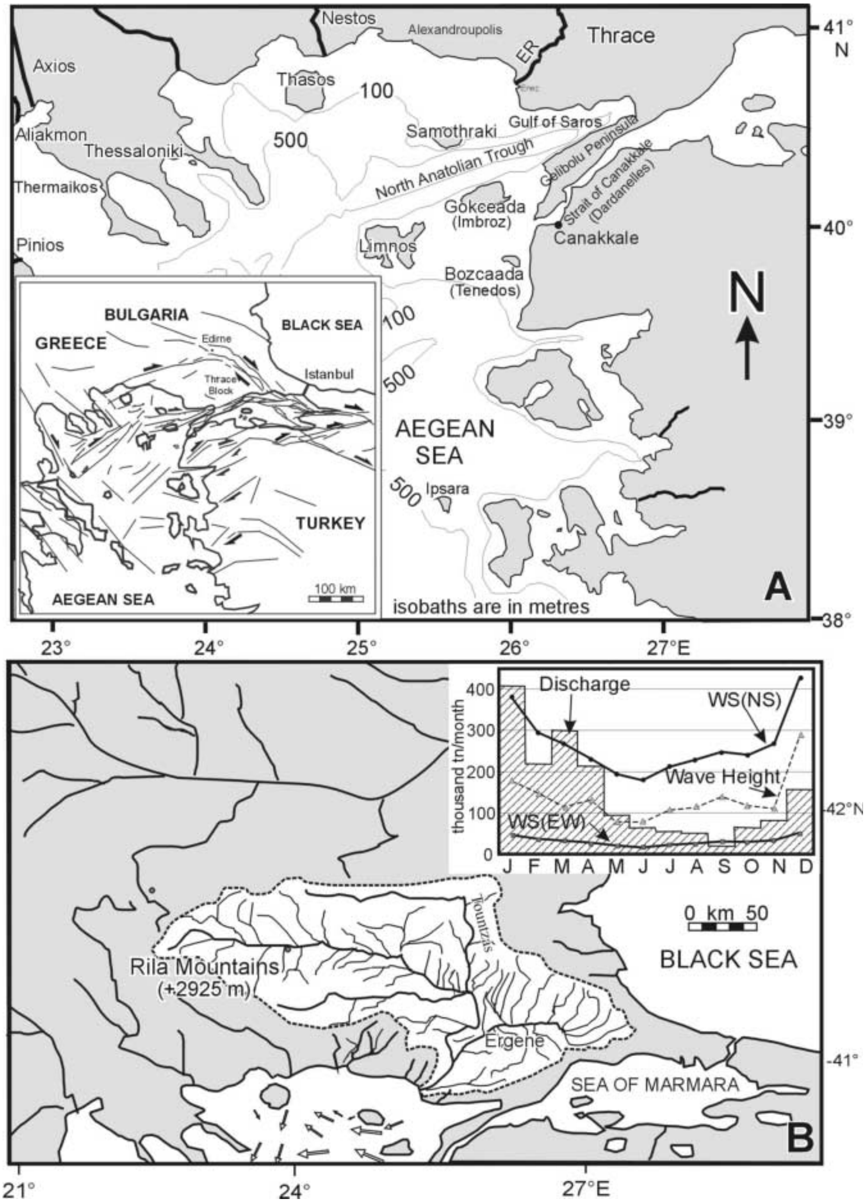

Fig. 1: (A) North Aegean rivers and present-day bathymetry. Inset shows the general tectonic setting of the Marmara and Aegean seas (modified from YALTIRAK *et al.,* 2000). (B) Drainage area of the Enez-Evros River. The currents are under the control of the cyclonic surface circulation in the Aegean. The Black Sea water flows westward mostly along the northern coast of Limnos island. In summer, they enter the Gulf of Saros which is characterised by longshore currents and anticyclonic eddies. Also shown is a graph on a monthly basis summarising 20-year averages (1971-1990) of sediment yield (thousand tons per month) of the eastern tributary (compiled from EIE, 1993 and after ALPAR *et al.,* 1998), the relative influence of NS and WE components of wind stress and also wave height (after ZODIATIS, 1994; GEZGIN, 2001).

tributaries are Arthas (Ardas) River in Greece and Bulgaria, Erithropotamos in Greece, Tountzas in Bulgaria (Tundja) and Turkey (Tunca) and Ergene (Erginis) in Turkey. The catchment area of the ER is surrounded by the Strandzha massif to the north, Rhodopi Mountains to the west and Thracian plain to the east (Fig. 1b). It is 52,500 km^2 of which 66% lies in Bulgaria, 27.5% in Turkey and 6.5% in Greece.

The ER discharges into a broad gulf on the Thracian (Thracean) Sea, east of Alexandrou poli, where its delta deposits have formed an extensive delta with a width of about 11 km and an area of 150 km^2. This delta comprises a distinguished and very important hydro-biotope in Europe, and in this study it is called the Enez-Evros Delta (ED) as a common geological element of two neighbouring countries. Round its twin mouths, noted by Strabo, are several swamps and lakes, the largest of which is Gala Lake, the ancient Stentoris of Herodotus, on the Turkish side.

Its lower alluvial plain is located in a low potential seismic hazard zone on the southern shelf of the Thracian plain (PERISSORATIS & PAPADOPOULOS, 1999). It is of triangular shape with its base located on the shoreline of Aegean Sea and its apex at the point where the upper delta begins and the ER is divided into two main branches. The multifarious morphology of the lower alluvial plain and its coast has great ecological importance. It holds a diversity of habitats. It is a breeding - and overwintering - place for thousands of aquatic birds. It includes quite a wide variety of biotopes. However, river pollution and the hydrologic condition of the biotope has caused the reduction of the quantity and the number of fish species. Fresh water from the ER has been increasingly taken off for irrigation. Its waters show a high nitrate load due to agricultural fertiliser runoff in the catchment area. Since the 1970's, the ED has, unfortunately, undergone severe degradation, due to dikes, canals, drainage works and irrigation ponds. Channel construction and embankment of the river have caused major hydrological and land use changes in the delta. The greater part of the delta is covered with saline vegetation. Parts of the wetland habitats have been lost.

Sediment transportation and oceanographic interactions

The seaward part of the ER (Lower Delta) has changed its shape very rapidly as a result of gradually increasing fluvial deposition. At present, the western tributary of the ER, known as "Efthigrammisi", carries the bulk of the river's water, while the eastern one is the old riverbed, which marks the political border between Greece and Turkey. The latter is mostly responsible for the aggradation and seaward progradation of the present delta, which is a typical Mediterranean coastal type. As a result, an almost regular shoreline with lagoons, separated from the sea by slightly inclined bars, became irregular (GOCMEN, 1977). All geomorphologic units representing different stages of development became mixed in the region. They include shallow coastal waters, brackish lagoons (still non-filled) and smaller freshwater lakes and channels, sandy beaches and islets, temporarily flooded mudflats, saltmarshes, and to a lesser extent, freshwater marshes. On the delta, the terrain is characterised by vast open spaces, either dry or flooded with shallow water, full of small plants that prefer brackish water. The monotony is broken here and there by the Tamarisks, which form clumps on the banks of the swamps and trenches, and by dense beds of reeds. The delta has developed in a very short time and is growing with two conspicuous lobes on the old submarine delta of the ER, forming a marked feature on the shelf (ARDEL, 1959).

The ER has an annual discharge of 6.8 km^3 yr^{-1} with an increasing trend in late winter and early spring (AKSU *et al.,* 1995). Considering sediment yield, the ER is the most important river discharging into the Aegean Sea. A total of 1.8 x 10^6 tn yr^{-1} of sediment load is discharged by the eastern tributary (Fig. 1b). In the inner shelf prodelta deposits have been developed in "prismatic" geometry receiving almost 80-90% of the discharged sediments (LYKOUSIS & CHRONIS, 1989). The outer shelf consists mainly of relict muddy sands and sands. The thickness of the late Quaternary sediments (128,000 years-today) in the slopes and basins ranges from 20 to 50 m, implying mean sedimentation rates from 10-30 cm ka^{-1}. For the Holocene sediments, the equivalent values are 0.5 - 1.5 m and 5 - 15 cm ka^{-1}, respectively (PIPER & PERISSORATIS, 1991). In fact, the subsidence of coastal margins of Mediterranean deltas during the Holocene (late 10,000 years) is considerably greater (>100 cm ka^{-1}) than in adjacent coastal plains (STANLEY, 1997). Similarly, the tectonic subsidence rates for the North Aegean basins have been calculated for the eastern part (95 - 180 cm ka^{-1}) and western part (30 - 150 cm ka^{-1}) (LYKOUSIS, 1991).

Since the deltaic development is controlled by the interaction of the transported alluvial deposits with the sea, a knowledge of the dominant climatic and oceanographic conditions is important. The climate is humid, mesothermic Mediterranean with a great deal of rain and snow. Average rainfall is 95.2 mm in December and 12.9 mm in August, with an annual average of 580 mm (second half of the 20th century). The winters are often cold (4-8 °C) with strong northeast winds, while the summers are hot and dry (CELENK, 1973; AKGUN, 1990). The prevailing winds are northerly, dry and cold in the greater part of the year (Fig. 1b).

Tidal range is low with a mean spring range of 20.3 cm at Bozcaada (ALPAR & YUCE, 1996). There has been little attempt at in modeling of the cyclonic water mass circulation in the North Aegean Sea, due to sparse hydrographic data. The warm (16 – 25 °C) and saline (39.0 - 39.5 psu) Mediterranean water enters the Aegean Sea from offshore of Rhodes and moves northward along the western coasts of Turkey. Then it is displaced westward south of Canakkale by the cooler and less saline (22 - 23 psu) Black Sea surface water outflow (YUCE, 1992). The water mass flowing from the Black Sea to the Aegean Sea is 874 km^3 yr^{-1} (STASHCHUK & HUTTER, 2001), with a minimum (200 km^3 yr^{-1}) in March and a maximum (1,000 km^3 yr^{-1}) in September (BESIKTEPE *et al.,* 1993). Therefore the influence of the Black Sea waters is greatest in summer and fall. The main pathway of the Black Sea water follows the periphery of the cyclonic gyre existing in the North Aegean, deflecting branches in the Samothraki and Thermaikos plateaus.

In the summer, strong, cold and dry, northerly winds (Etesians), are particularly influential in the hydrodynamics of the Aegean Sea giving rise to intense convective movements of water. This leads to upwellings along the western Turkish coast and western coasts of the Greek islands in the Eastern Aegean Sea. In the winter, an overall cyclonic surface circulation exists with a northerly current along the Asian coast and a southerly current along the Greek coast. Between them, surface water flows westward over the Samothraki and Limnos plateaus (Fig. 1b). Resultant dense waters sink down to deep troughs and basins, contributing to the renewal of the deep water (GEORGOPOULOS *et al.,* 1992; THEOCHARIS & GEORGOPOULOS, 1993; ZODIAKIS, 1994; GEZGIN, 2001).

Secular sea level and related vertical tectonic changes

The secular sea-level changes in the Black Sea are due to the joint action of eustatic movements and tectonic subsidence. The continual rising trends of the mean annual sea-levels vary between 1.3 to 8.2 mm yr^{-1} along the Black Sea coast (BONDAR, 1989). Meanwhile, the mean subsidence speeds of the Black Sea coasts are about 5.2, 1.1 and 6.5

mm yr-1 at Odessa, Ialta (Sevastopol) and Poti, respectively (LISITZIN, 1974; BONDAR, 1989).

Similar rising trends of the mean annual are valid for the Aegean coasts. The quantitative analyses indicate mean sea level rises in Thessaloniki (2.1 mm yr^{-1}), Izmir (0.2 mm yr^{-1}) and Leros (3.7 mm yr^{-1}) (MOSETTI & PURGA, 1991). However, the percentage of the eustatic and vertical tectonic movements is not known. From paleogeographical evidence, KAYAN (1988), EMERY & AUBNEY (1991) have found a tectonic subsidence of 4.6 mm yr^{-1} along western Anatolian coasts for the late Holocene that might be because of the compression of marine sediments. FLEMMING *et al.,* (1973) reported an about 2 m relative sea level rise in southwestern Turkey during the last 4,000 years but they did not state if it was because of eustatic or tectonic movements. On the other hand, some decreasing mean sea-level slopes were reported in Alexandroupolis (-0.2 mm yr^{-1}) and Kavalla (-3.6 mm yr^{-1}) (MOSETTI & PURGA, 1991). These findings may imply that the northern coastal area of the North Aegean Sea is exposed to a regional tectonic uplift. This is in agreement with palaeogeographic results obtained previously (SAKINC & YALTIRAK, 1997), even though lack of dating prevented them from predicting the uplift rate.

Geological basement and late Quaternary deposits

The structural formation of the lower valley started to evolve in the Oligocene (Fig. 2). The Neogene sediments are composed of various similar-aged lithofacies with lateral and vertical transitions (DERMITZAKIS & PAPANIKOLAOU, 1981; YALTIRAK,

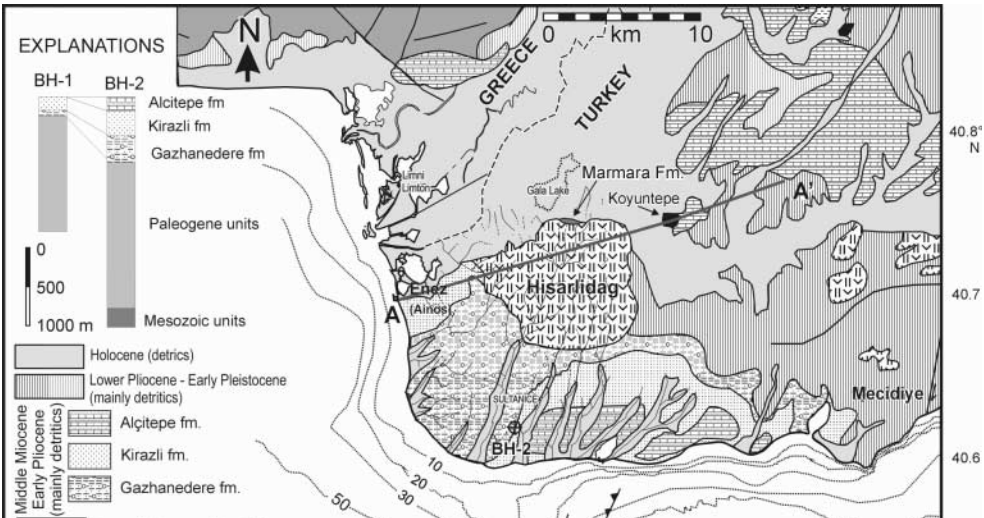

Fig. 2: Map showing geology of study area (modified from YALTIRAK *et al.,* 1998; YALTIRAK & ALPAR, 2002). The faults in the Gulf of Saros were compiled from shallow seismic investigations of CAGATAY *et al.* (1998); YALTIRAK *et al.* (1998). Heavy lines are faults with ticks on downthrown side. Borehole data were compiled from YALTIRAK & ALPAR 2002.

1996; YALTIRAK *et al.,* 1998; SAKINC *et al.,* 1999; YALTIRAK & ALPAR, 2002). On the basis of these previous studies and also depending on our field studies, a generalised stratigraphic section representing the study area has been produced in this study (Fig. 3). The Quaternary deposits discordantly overlie a basement consisting of Hisarlidag volcanic rocks (late Oligocene, 35±0.9 My; SUMENGEN *et al.,* 1987), siliclastic rocks (Oligocene-Eocene), carbonated sandstones (Upper Pontsian) and Ostrea-bearing carbonate-clastics (Lower Pliocene). The Hisarlidag volcanic rocks have been spread out over a wide area with a diameter of about 25 km (DEMIREL *et al.,* 1998). Pillow lavas at some localities characterise quiescent submarine volcanic activity. The Hisarlidag Formations were overlaid unconformably by the Gazhanedere Formation (Middle Miocene) representing fluvial and lagoonal environments (Fig. 3). On top of the Gazhanedere Formation, with lateral and vertical transitions, rests the Kirazli Formation, made of massive and semi-consolidated sandstone representing beach and back-beach lithofacies of the late Miocene (SAKINC *et al.,* 1999). Both formations are conformably

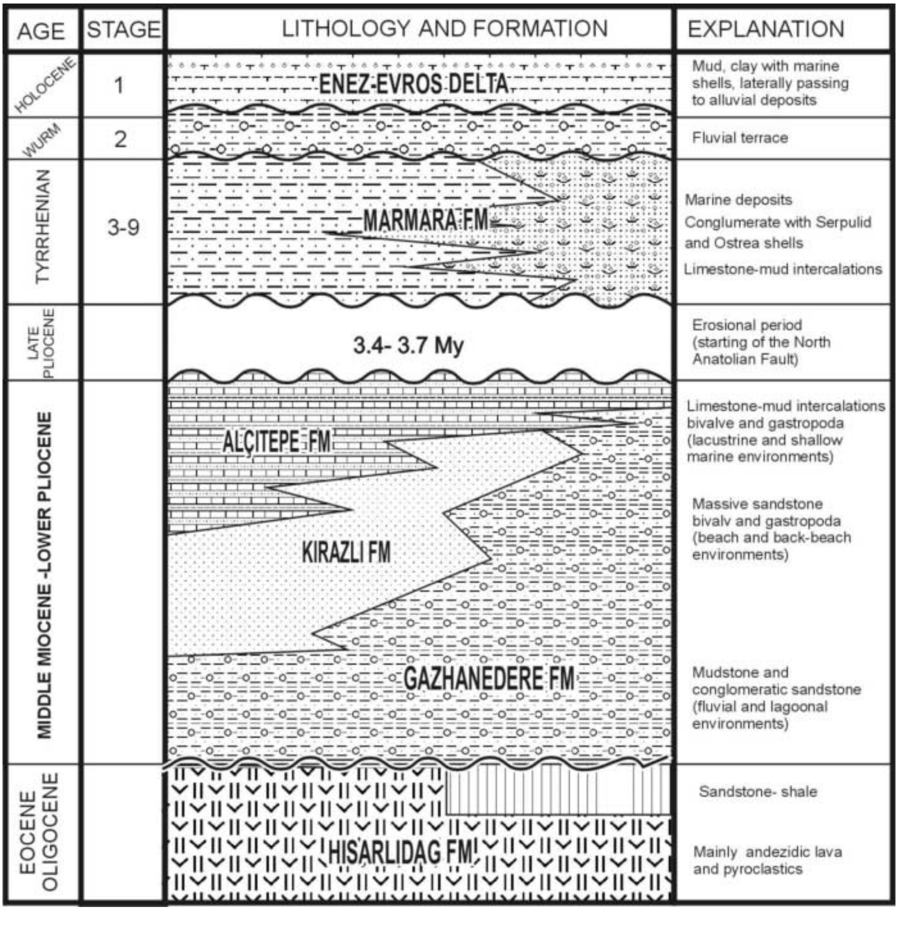

Fig. 3: Generalised stratigraphic section of the study area. All formations observed on land and isotopic stages of their equivalent underwater units can be compared. The Miocene sequence near Enez (Fig. 2) comprises 23 m-thick light grey bioclastic limestone, containing Paratethys fauna.

overlaid by the Alcitepe Formation with lateral and vertical transitions, which is made of carbonated sandstones with cross-bedding and gravels, representing shallow marine and lacustrine depositional environments. The basement of the Alcitepe Formation contains Paratethys fauna. Ostrea-bearing fossils (Upper Messinian) at the upper parts of the Alcitepe Formation indicate normal marine depositional conditions (SAKINC *et al.,* 1999). The town of Enez (Ainos) is located on top of a remnant hill composed of the Upper Miocene formations rising above the alluvial fill. These layers also form the area south of the Hisarlidag massive (Fig. 2). Farther NE, the Ergene Group rocks (Pliocene) were made from fluvial-lacustrine siliclastics. Along the eastern part of the ER, the angular unconformity deposited on top of the lower units (KASAR, 1995). Koyuntepe Basaltic Rocks (Miocene) can be observed only in a small area in the village Koyuntepe (Fig. 2) and is typical of alkali basalt outcrops around the flat Thracian plain (SUMENGEN *et al.,* 1987).

A non-depositional and denudation period continued until the Pleistocene (Fig. 3). During the Mediterranean transgression in the middle-late Pleistocene (stages 9-3, YALTIRAK *et al.,* 2002), the valleys and broad coastal plains were occupied by warm sea conditions. In this period, the Marmara Formation (TYRRHENIAN, YALTIRAK, 1996; SAKINC *et al.,* 1999; YALTIRAK *et al.,* 2002) was deposited along the coastal area (Fig. 4). Its thickness may vary depending on the morphologic characteristics of the shoreline. The terraces consist mainly of conglomerates and sandstones, rich in Ostrea edulis shells. In general, they overlie angular uncomformity the Miocene rocks and are underlaid disconformably by the Holocene shallow-marine sands, containing Cardium sp. and Murex sp. (SAKINC & YALTIRAK, 1997). With the continuation of tectonic activity, the Marmara Formation attained a regressive character in a short period and tectonically uplifted 5-35 meters above the present sea level. The Marmara Formation in the study area can be observed as a terrace unit at the north foot of the Hisarlidag volcanics, 24 m above present sea level (Fig. 2). Its average thickness is 12 m. It consists wholly of detritic material (clastics) and generally terminates with beachrock facies (Fig. 3). Upper parts of this formation have a transgressive characteristic developed in normal marine conditions, indicating a tectonic uplift (YALTIRAK *et al.,* 1988; 2000; 2002).

Depending on the sea level fall during the Postyrrhenian regression (Stage 2, YALTIRAK *et al.,* 2002), the palaeo-river incised its valley. A river channel was developed on the shelf, reaching to the North Anatolian Trough. There are some fluvial terraces caused by the palaeo-river. Traces of entrenched meanders at the sides of these terraces are related to the displacements of the stream flow. They are associated with the coastal marine terraces (4-5 m) lying 200-500 m away from the shoreline to the south of the town of Enez and parallel to it (GOCMEN, 1977). Their positioning may imply that they were deposited at the early deglaciation phase following the last glacial period.

Topmost deposits are marine, lagoonal, coastal plain and alluvial sediments accumulated during and after the Holocene transgression. The boring of the General Directorate of State Hydraulic Works (DSI) indicated that the Holocene deposits were as thick as 28 m along the coast (30 m inland). The fossils in these deposits *(Ostrea edulus, Hydrobia sp., Vermetus sp., Dentalium sp., Mytilus edulus, Gastrana fragilis, Bittium reticulatum, Corbula gibba, Mactra subtrancata, Cerithium vulgatum)* indicate a transient depositional environment from brackish to saline water (GOCMEN, 1977; SAKINC & YALTIRAK, 1997). Furthermore, 26-km inland near the town of Ipsala, gravel deposits placed on top of the basaltic layers (-110 m) represent a typical braided channel (GOCMEN, 1977).

Late Quaternary evolution of the delta

The region was occasionally affected by the Black Sea (a part of the Neogene Paratethys until Quaternary) during the Middle Miocene - Lower Pliocene forming a passage area between Mediterranean and Paratethys (SAKINC *et al.,* 1999). After Upper Miocene, the basement rocks folded along NE trending axes (YALTIRAK *et al.,* 1998) due to the compressional forces caused by the dextral movement along the Thrace-Eskisehir right-lateral strike-slip fault and its branches (Fig. 1a) (SAKINC *et al.,* 1999; YALTIRAK *et al.,* 2000; YALTIRAK, 2002; YALTIRAK & ALPAR, 2002). In later periods (Plio-Pleistocene), the Aegean Sea, which was associated with the Black Sea through the Ergene-Euxine Basin, left the area gradually leaving behind lakes and marshes (STEININGER & ROGL, 1984). The lower valley was a connecting channel between the continental basin to the northeast and the marine area to the southwest during the high sea level stands in Pleistocene. Meantime, the folded strata defined the catchment area of the ER (Fig. 1b) and its tributaries located on an erosion plain (Lower Pliocene, GOCMEN, 1977).

Since the mean sea level in Tyrrhenian (stage 5, YALTIRAK *et al.,* 2002) is not very different from the present one, the terraces (Marmara fm) located at the northern foot of the Hisarlidag mountain (Fig. 2) indicate a regional uplift of 24 m (SAKINC & YALTIRAK, 1997). The existence of these terraces implies that the Tyrrhenian (stage 5) Aegean Sea invaded a large bay (Fig. 4). During the last glacial age, the sea retreated back down to the present shelf edge (120-130 m below present sea level, CHAPELL & SHACKLETON, 1986). The fossils indicate different Holocene sea levels during the post-glacial rise. The braided channels on top of

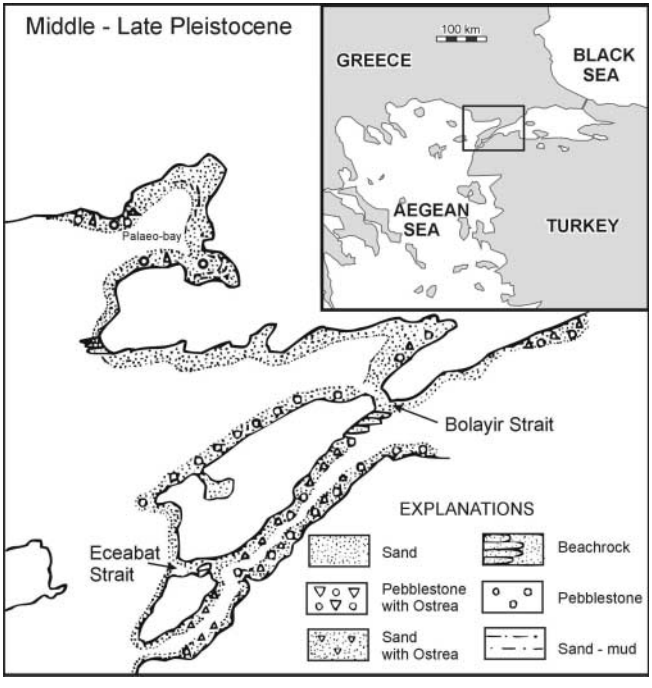

Fig. 4: Middle-late Pleistocene palaeogeography of the northeast Aegean Sea (after SAKINC & YALTIRAK, 1997).

the basaltic layers (-110 m) near Ipsala imply that the warm Aegean Sea waters invaded the delta and flood plains along the ER at least 26 km inland from the present coast and even approached the Ergene junction, forming a large, shallow bay. The sea level was stable at these levels for a long time during the Holocene (stage 1) transgression (EROL, 1991).

Parallel to the tectonic uplifting during the Holocene, the erosion and the transportation of the ER increased and rapidly filled the large bay at its mouth. Meanwhile, the ER located along the valleys parallel to the fold axes has frequently changed its route, causing a stack of deltas. Beyond extensive land use in the form of agriculture, dikes, irrigation channels and drainage work on both sides of the ER, the Landsat 5 Thematic Mapper (TM) images also indicate these changes (GAZIOGLU *et al.,* 1998). The ER and the mouth of its western tributary changed position several times. The delta first developed around the Drakus Lake. The secondary evolution of the delta front sandy deposits coincides with the location of the present delta. Later on, a new delta was developed southward at Psevdorhion, and deltaic construction retreated northward back to its present location. During these evolutions of the western tributary, the eastern one was the main riverbed filling a large protected embayment at its mouth. The delta fronts only became evident after the bay was filled with the alluviums. In spite of this delay, this delta lobe is better developed, partly because the material transported by the eastern tributary was deposited on the old delta lobe and other deltaic successions of the western tributary (GOCMEN, 1977; GAZIOGLU *et al.,* 1998). The last geological evolution could also be supported with the most recent archaeological data and historical geographic maps.

The well-watered coastal plains of the Northern Aegean may have been inhabited or little used until 9,000 BP when the first Neolithic settlements appear. The area may have furnished subsistence for a plains population. Many historical artefacts representing the Thracian (8-6th centuries BC), Persian, Hellenistic Period, Roman, Byzantine and Ottoman cultures have been excavated in the area (BASARAN, 1996). Neither historical (immigrations, invasions) nor natural adverse factors (river overflows, floods, torrents) interrupted the continuous settlement in the region. The first site of Ainos (Enez) is 3 km far from the present coast line and on top of a hill, rising to a height of 25 m. Some structural relics belonging to the late Paleolithic (7th century BC) have been found around the acropolis (citadel in ancient Greek towns) on top of this hill. This structure was placed on a hard basement (Miocene carbonated sandstone) 7.5 m deep from the present surface (ERZEN, 1974, 1987). The relics of the Roman, Byzantine and Ottoman periods are distributed in layers down on the plain area. On the basis of the dated archaeological information and drilling works carried out around the acropolis, it may be suggested in this study that 3 ± 0.5 m isopach above present mean sea level represents the coast line dominant at the end of the late Paleolithic (Fig. 5a). For comparison a geographic map is given with the drainage network and bottom sediment classification (Fig 5b). Some available geographic maps show both tributaries of the ER. The oldest is the chart drawn by a Turkish Admiral (PIRI REIS, 1526) (Fig. 5c). This chart should be included in the "Book on Navigation" due to the geopolitical importance of the region. A natural port is located at the junction of the mainland and sea routes, from the Balkans and southern Europe to the Anatolia and Aegean Sea. It is also well-known that river transportation was possible along the ER (EYICE, 1956). On Kiepert's map (Fig. 5d), the port is located on the shore of a large lagoon, separated from the sea by two spits opposite each other (KIEPERT, 1890). The eastern branch forms an estuary discharging into this lagoon. Farther NW, the western branch forms a sharp point seaward.

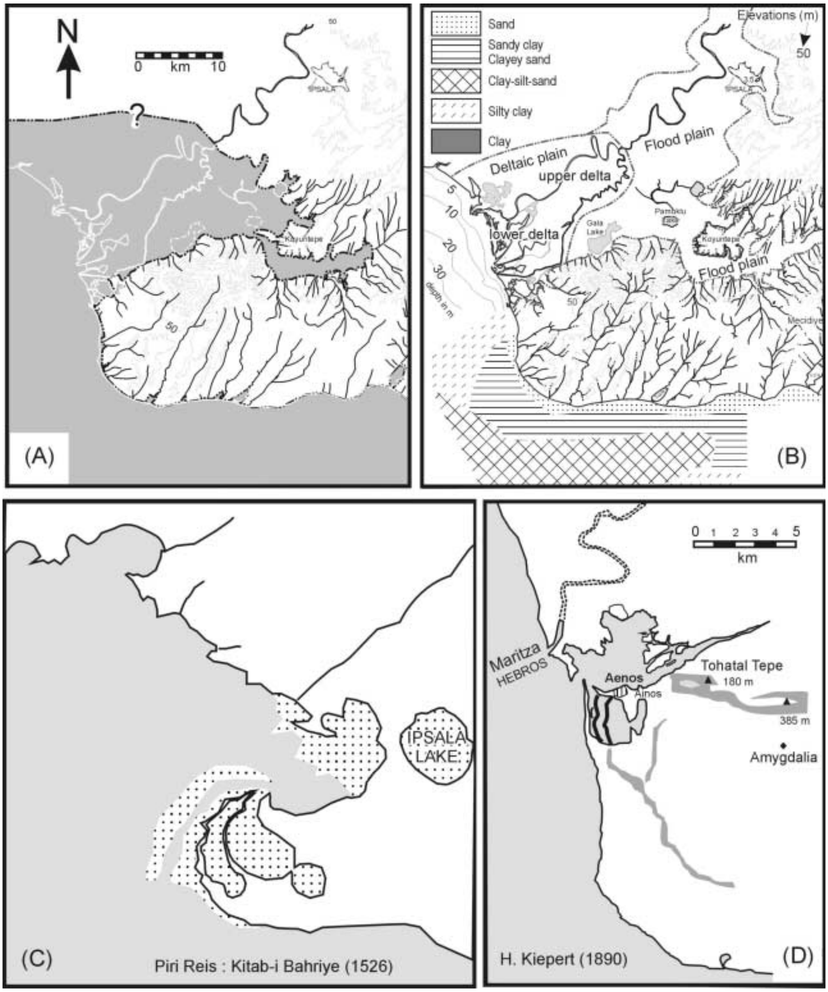

Fig. 5: (A) Palaeoshores presenting the end of the late Paleolithic period (about 7th century BC). A base height (about +3 m above msl) was estimated in this study from the dated archaeological data and drilling works carried out around the acropolis. The coastal line was about 10 km inland from the present one and the hill on which the acropolis is located may occasionally have been turned into an island or a headland depending on the river floods. (B) Present-day geography for comparison. The drainage network (modified from data in GOCMEN, 1977) was superimposed. Bottom sediment classification is compiled from SANER, 1985; CAGATAY *et al.,* 1997, 1998; SARI, 1997; YALTIRAK *et al.,* 1998; ALPAR *et al.,* 1999; SARI and CAGATAY, 2001; and this study. Also shown are two historical maps showing the coastal development of the area. (C) A map from the Naval Book of Piri Reis, a Turkish Admiral (1470-1554). The date of the book is given in verse in the traditional way. From the final couplet one makes the date out to be 1526 A.D. (D) The map of H. KIEPERT (1890).

On the basis of previous studies, the present deltas at the mouth of the ER and on the respective late glacial shelf area, attained their final configuration with the Holocene transgression. All available data and the coastal bars indicate that the progradation processes were especially dominant in the present eastern tributary and frequently changed its direction. A platform structure 35-40 m below the mean sea level is even evident on navigation charts, indicating that marine and fluvial processes are in effect at present on a main delta. Even though the ED forms an interesting unit with its old abandoned mouths, sand shoals developed when the sea level lowered below present levels at the river mouth, the structural and stratigraphic features of its submarine prolongation are not known in detail due to the lack of geophysical and bore hole data. This is partly because the ED is positioned in a politically strategic area. Therefore, the main scope of this study, in the hope that the results may be the key to the tectonic and geomorphologic evolution of the study area, is to determine the delta's evolutionary phases in the Plio-Quaternary.

Data

In October 1997, a high-resolution seismic cruise was carried out offshore the ED, on board R/V. Arar of Istanbul University. The database consists of 224 km-line of seismic profiles (Fig. 6). Unfortunately, the seismic lines stuck within the continental waters (6 miles) due to the policies of the Ministry of Foreign Affairs. A few sediment gravity cores up to 3.1 m-long were also obtained on the submarine ED (ALPAR *et al.,* 1998).

A 1.25 kJ multi-electrode sparkarray (6 kV and 30 mF) and a 11-element, 10-m-long surface-towed hydrophone streamer (single channel) were used. The seismic source was less than one meter in length with 30 discharging electrodes (6 kV and 30 mF), spaced about 5 cm apart. The sparkarray and streamer were kept floating about a half metre below sea surface. The sampling interval was 1/4 ms, the recording window length was 250 ms (two-way-travel time) and the shot interval was 2 sec (about 4.1 m). Positioning was carried out by using an integrated GPS with an accuracy of ± 50 m for moving objects. The preliminary results (ALPAR et al., 1998) of these non-processed sections provided some details on sedimentary deposits more than 150 m below the seabed, which is in good agreement with the practical applications given by DES VALLIERES *et al.* (1978) for seismically hard environments. However, some parts of the sections were masked by reverberations. Therefore, within the scope of this paper, and since the raw data is digital, some data processing methods were applied. The electric noises which are emitted from the sparker and observed in the beginning of the traces before the first breaks, were muted. Trim-statics helped the elimination of the arrival time shifts caused by rough sea states. Band-pass filtering was made by zero-phase filter with a band width of approximately 200 to 500 Hz. In addition, multiple reflection attenuation, which uses the simple M. Backus principle, was applied to suppress the seabottom multiples, as the most serious problem in the sections. Even though the signal distortions due to amplification were somewhat high, they did not significantly prevent us from decreasing the multiple reflection level remarkably on the final sections. These processes have contributed a great deal to the more precise definition of the depositional sequences.

Results

On seismic sections, the infill in the study area is characterised by two depositional sequences separated by an angular unconformity. In descending order, these are the sedimentary units (late Quaternary sequences) and the Miocene basement. The reflection characteristics of the topmost sediments imply that the eastern part of the study area and the northern shelf of the Gulf of Saros generally consist mainly of coarse detritic materials, which indicate high near bottom physical energy of currents and waves. On the contrary, well developed sedimentary sequences are dominant westward, owing to the high sediment inputs. Therefore mud is dominant westward on the shelf area closer to

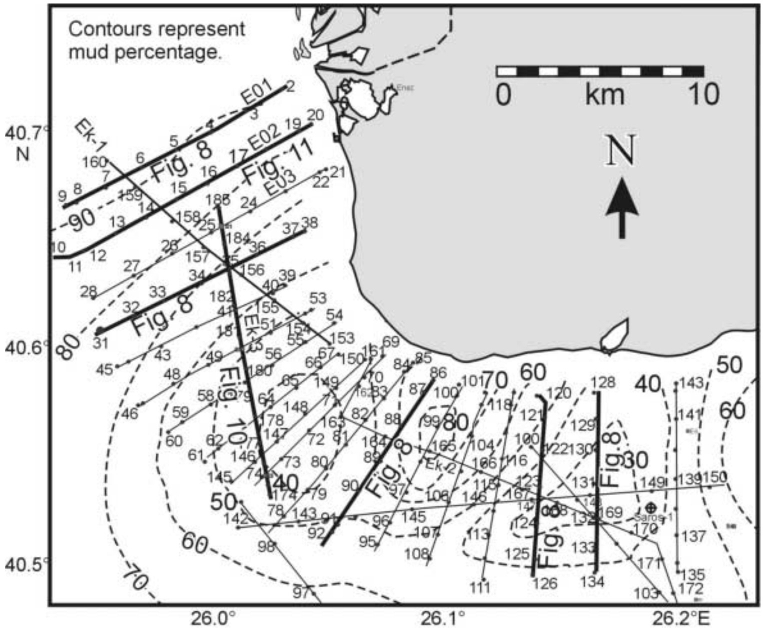

Fig. 6: Location of sparker profiles. Thick lines indicate seismic lines given in Figs. 8, 10 and 11. Contours at sea represent mud ratio of the bottom deposits (after SARI, 1997; CAGATAY *et al.,* 1998; SARI & CAGATAY, 2001).

the ER and on the slopes as well. Although grain size decreases from shore to offshore, many high impedance reflections (fine grained material) and some discontinuous reflections (mixed or chaotic material) can be observed. The main morphologic and geologic features can be classified as below.

Bathymetry

The North Aegean Sea is characterised by a relatively extensive shelf and the North Anatolian Trough (Fig. 1a), down to 1,600m deep. The shelf gradually dips southward, a typical deltaic platform developed by delta progradation during late glacial sea-level stand (progressive type margin). The Island of Samothraki is located on this shelf. The broad shelf area continues up to –100 m isobath where a deep trough continues towards WSW between the islands of Bozcaada (Tenedos) and Samothraki to join the North Anatolian Trough. The regions along the outer shelf and the shelf-break (100-150 m depth) adjacent to the rapidly subsiding North Anatolian Trough are erosion areas.

A bathymetry chart of the study area (Fig. 7) was prepared from the seismic data. Even though the isobaths on the shelf show a parallel extension to the shore in its general outlines, they represent a morphology caused by delta lobes. The -40 and -50 m isobaths, which open up like a fan in the NW part of the map, indicate the submarine delta of the ER. The –50 m isobath extends at a distance of 35-45 km from the mouth of the ER. The average bottom slope gradients on the shelf area are 0.26° for the ED and 0.46° for the eastern region of the study area. The thickness of the submarine delta decreases offshore and it terminates where the water depth is between 40-42 m as it is seen on the seismic sections (Fig. 8). The shelf-break denotes the topset to foreset transitions of the ED prograded during the end of last glacial period, immediately prior to the Holocene transgression.

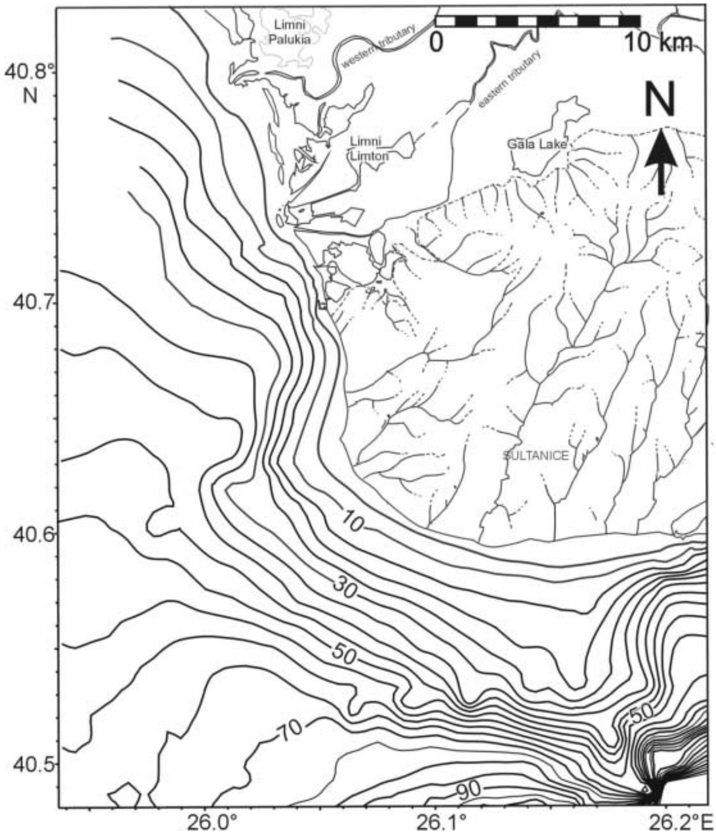

Fig. 7: Map showing bathymetry of study area. Depths are in metres as converted mainly from seismic data using 1500 m s-1 sound velocity.

Sedimentary units
(late Quaternary sequences)

Widespread and continuous reflections with satisfactory amplitudes could be traced between the high-amplitude sea bottom and an acoustic basement. They are well developed westward on the erosional surface, owing to the high sediment input by the ER. On land, the late Quaternary sedimentary units were defined in detail along the southern coastal areas of Thrace. Later, the offshore sedimentary deposits corresponding to these formations in the Gulf of Saros, NE Aegean Sea and along the western coasts of Turkey were defined by many researchers (AKSU & PIPER, 1983; AKSU *et al.,* 1987, 1990; 1995; YALTIRAK, 1996; SAKINC & YALTIRAK, 1997; SKENE *et al.,* 1998; CAGATAY *et al.,* 1998; SAKINC *et al.,* 1999; YALTIRAK *et al.,* 1998, 2000; YALTIRAK & ALPAR, 2002). The sequences outlined in our reflection data are subdivided into three seismic stratigraphic units; A, B and C from top to bottom. Each unit is separated from the other by moderate reflections. These units could be compared with those previously studied units, and could possibly be dated. On the basis of previous studies, these units are interpreted as the late Quaternary sequences on the erosional surface of the Upper Miocene basement. In order to be able to correlate the units defined in shallow seismic reflection data, the Pleistocene units will be referred to as Marmara Formation in this study.

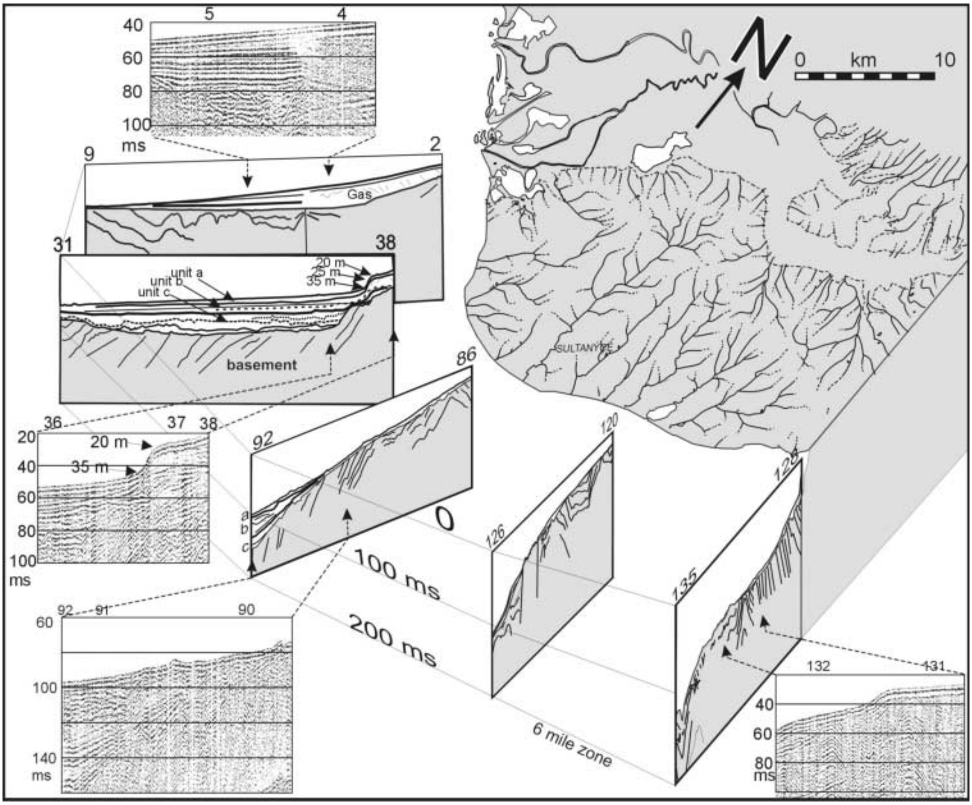

Fig. 8: Fence diagram of interpreted line drawings. See Fig. 6 for detailed location. The basement, which is mainly made up of detritics of Alcitepe, Kirazli and Gazhanedere formations (Early Miocene- Early Pliocene), is shaded. View is looking northwest; northeast is to the right. During the worldwide Holocene delta initiation when sea-level rise began to decelerate from 8,000 to 6,500 yrs BP, the sea level was 25 m below the present stand. This means that the palaeo-coastline, which can also be observed on the sections by a sudden dip change around the same depth, was about 7 km seaward from the present coast.

Unit A

This depositional sequence consists of a thin (< 9 m) sediment blanket characterised by weak and internally parallel reflectors. It is thickest western part of the study area especially along the lines E02 and Ek-3. The prodelta platform becomes narrow and Unit A thinner to the south (Fig. 9a). Two depositional sub lobes can be recognised within Unit A. A few sediment gravity cores (1-3 m) on the shelf area indicated rich gastropode and bivalve macrofauna (ALPAR *et al.,* 1999). Core data and its reflection characteristics (Fig. 10) show that the river Unit A is principally deltaic with green to grey muddy areas near the river mouths changing to muddy sand further out offshore (Fig. 5b). Farther east, the surface sediments on the northern shelf consist mainly of sand, whereas those on the slope and the deep trough are mainly silt and clay (SARI, 1997; CAGATAY *et al.,* 1997). Although surface sediments in the Gulf of Saros indicate a pristine nature in terms of metal pollution, organic carbon distribution shows the effect of anthropogenic and natural inputs from the ER (SARI & CAGATAY, 2001).

This topmost and widely distributed unit is interpreted as a Holocene posttransgression unit deposited during the most recent high stand of sea-level (late 6,000 yrs). Since, the exposed lands were inundated for the last time and become the site of deposition for Unit A during the Holocene, maximum sedimentation rate for the Holocene sediments (late 10,000 yrs) in front of the ER could be calculated as high as 90 cm ka^{-1}, 6 times higher than that for northern Aegean continental margin given by PIPER & PERISSORATIS (1991).

Unit B

This depositional sequence underlies Unit A and onlaps Unit C. It is thickest (about 22 m) at the mouth of the ER and then at the shelf edge side of the line Ek-3 (Fig. 9b). It gradually thins towards the east of the area and the basement comes close to the sea bottom. The distribution and thickness of Unit B suggests that the river discharge occurred near the present distributary mouth and the shoreline was probably at a position comparable to that of the present day. However, the river mouth may have shifted during the depositional period of Unit B, giving rise to two partly overlapped delta progradational lobes in this depositional sequence (Fig. 9b). This unit also gradually thins towards shelf-break and possibly disappears out of the study area. At the shelf-break, it reappears again with sigmoid-oblique configuration. This unit is defined by sigmoid to oblique and basinward prograding clinoforms dipping with low angles. Its inner weak-amplitude reflections terminate updip by toplap near a horizontal reflector and downlap against the older, near-horizontal reflectors. The topset to foreset transition is outside the study area to the southwest, about 100 to 120 m below the present sea level.

Unit B is interpreted to have been deposited as a deltaic sequence progradation during a sea-level lowstand between the Würm glacial period (stage 2) and the following early phase of deglaciation (25,000-13,000 yrs BP) when the sea level dropped as much as 120 m below the present stand (VAN ANDEL &

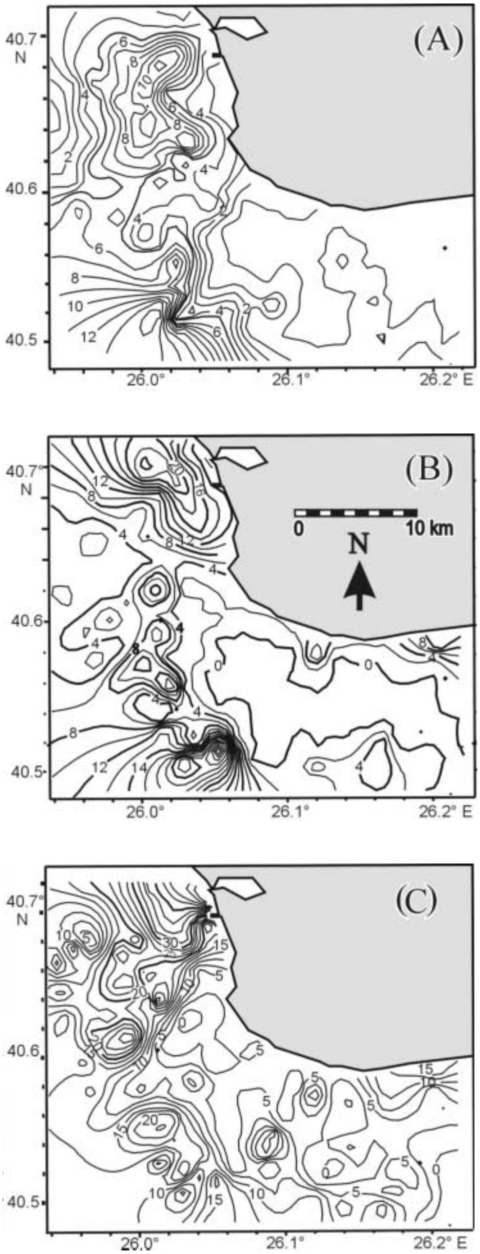

Fig. 9: Isopach maps showing sediment thickness of the units A, B and C. Thickness is in ms two-way travel time. Contours are at (A) 1-ms intervals for unit A, (B) 2-ms intervals for unit B, and (C) 2.5-ms intervals for unit C.

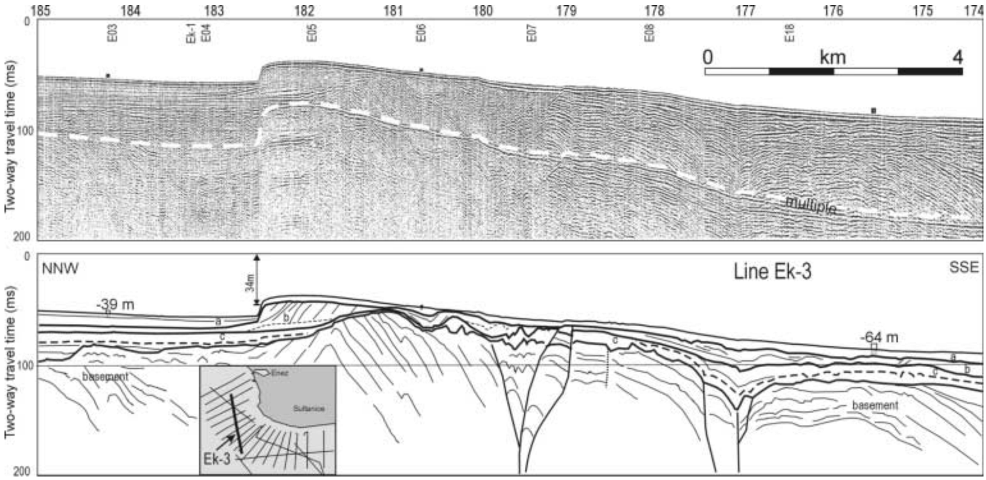

Fig. 10: Seismic data (Line-EK3) crossing the analysed core samples. A mud layer of at least 3 m was recovered around fix number 184. The entire sediment core was fine grained with a complete absence of shells and shell debris. Between fix numbers 175 and 176, homogeneous mud and silt are dominant along a 1.75 m-long core. A thin layer 0.8-1.0 m below the sea floor contain very rich gastropode and bivalve macrofauna, and overlies a silty clay matrix with silt and sporadic sand bands that contains some marine macrofauna (ALPAR *et al.,* 1999). The vertical scale in two-way travel time. 100 ms is about 80-90 m by considering an average velocity of the water column and uppermost sediments. Inset shows the location of the line. See also Fig. 6 for detailed location.

LIANOS, 1984; AKSU *et al.,* 1987) and thus caused the exposure of the shelf areas once again. Similar progradational deltaic sequences occur along the coasts of western Anatolia as well as in the NW Aegean (Thermaikos Gulf, LYKOUSIS, 1991) where they are found below the present day deltas at 100-110 m water depths. Their upper parts are dated by the 14C method to be 14,000-10,000 years BP (AKSU & PIPER, 1983; AKSU *et al.,* 1987).

Unit C

This depositional sequence is represented by parallel to quasi-parallel reflections and its small-scale depositional sub lobes. These sub lobes are bounded by strong reflectors and will be named as C1, C2 and C3 (Fig. 11). They downlap against the erosional surface of the Upper Miocene basement. Their total thickness changes between 0-50 m. It is thickest at the present distributary mouth and also to the southwest part of Line E05 (Fig. 9c). The depositional sub lobes represent marine sediments deposited in the late Pleistocene when the shelf areas were flooded and truncated by the Mediterranean waters. This transgression probably took place during the Riss-Würm interglacial period (stages 3-5) (128,000-25,000 yrs BP). Occasional weak reflection packages encountered at the base of some sub-units may represent coarser materials such as gravels which are followed towards the top by finer material. Soft sediment deformations related to faulting can also be identified within this unit (Fig. 11). In the area close to shore (right side) the screening effect may be caused by the gas saturated silty sediments (shallow gas), because of the high absorption and backscattering effects in these sediments. The downward inclination or sagging of the deep reflectors in that area may be caused by the velocity decrease in gas saturated sediments.

As a consequence, the total thickness of the late Quaternary sediments (135,000 yr-today)

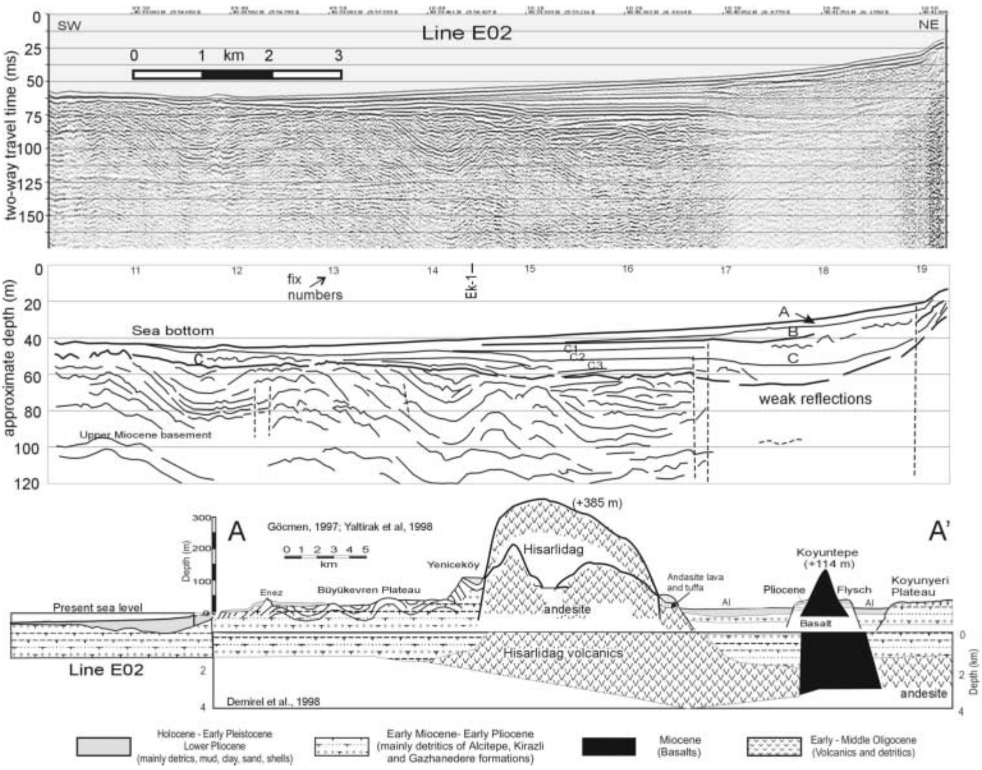

Fig. 11: Seismic section (Line-E02) and its interpreted line drawing. Also shown at bottom is a generalised (NE-SW) geological section along the profile A-A' onland (Fig. 2), modified from GOCMEN (1977) and YALTIRAK *et al.,* (1998). Deeper data are from the gravity modeling studies (after DEMIREL *et al.,* 1998).

ranges from 20 to 74 m considering an average velocity of 2,000 ms^{-1} (Fig. 12). Therefore, the mean sedimentation rate varies between 15 and 55 cm ka^{-1}. The late Quaternary sediments are thickest off the present distributary mouth of the ER, western and southwestern regions of the study area. Eastward, they are rather thin (5-10 m) on the northern shelf of the Gulf of Saros (south of Sultanice), which may be caused by lack of terrigenous inputs from the surrounding land.

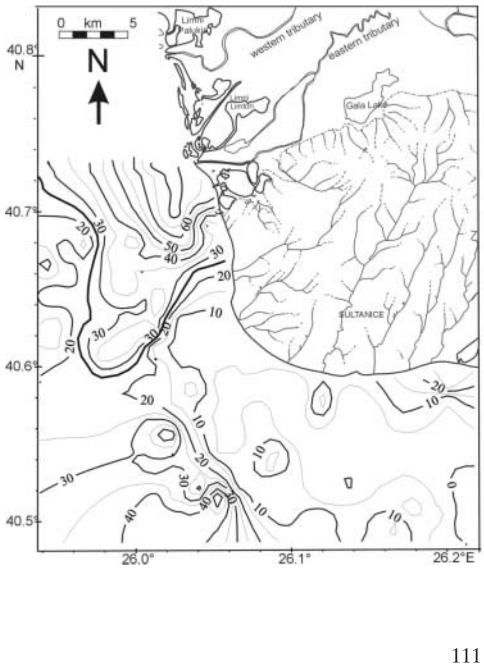

Fig. 12: Isopach map of the late Quaternary deposits (135,000 yr-today) showing sediment thickness. Contours are at 5-ms intervals, considering an average velocity of 2,000 ms^{-1}.

Unit D (basement)

Beneath all these depositional sequences, the deepest sedimentary unit constitutes the acoustic basement of the seismic sections. This sequence (Unit D) has high amplitude reflections and generally stayed within the seismic penetration limits. Based on the well-known discontinuities observed on land, as well as the cuttings of the BH-1 (Fig. 2), the top of this sedimentary unit highly possibly corresponds to the upper levels of the Alcitepe Formation (Upper Messinian) (Fig. 3). It is made up of folded layers which were probably formed during the Pliocene-early Pleistocene in the tranpression zones due to the westward propagation of the Anatolian Block (DEWEY & SENGOR, 1979; SENGOR *et al.,* 1985; YALTIRAK *et al.,* 1998). These folds, which occasionally outcrop on the sedimentary units and sea bottom, were defined as Upper Miocene in the seismic studies conducted in the Gulf of Saros (SANER, 1985; CAGATAY *et al.,* 1998; YALTIRAK *et al.,* 1998) and western part of the Gelibolu Peninsula (YALTIRAK *et al.,* 2000).

Within the study area, the average depth to the acoustic basement from the present sea level varies between 60 and 90 m (Fig. 13). The maximum is calculated as 121 m to the southeast end of Line Ek-2 (Fig. 6). From shore to offshore, the basement tends to become deeper at first and then become shallower (Fig. 13). However, the imaginary arc connecting the points where the depth to the basement is maximum, comes close to the coast at the western part of the study area and approaches the present distributary mouth of the ER 100 m offshore. Inadequate seismic penetration prevented us from classifying weak inner reflections of the offshore Miocene sequence in detail.

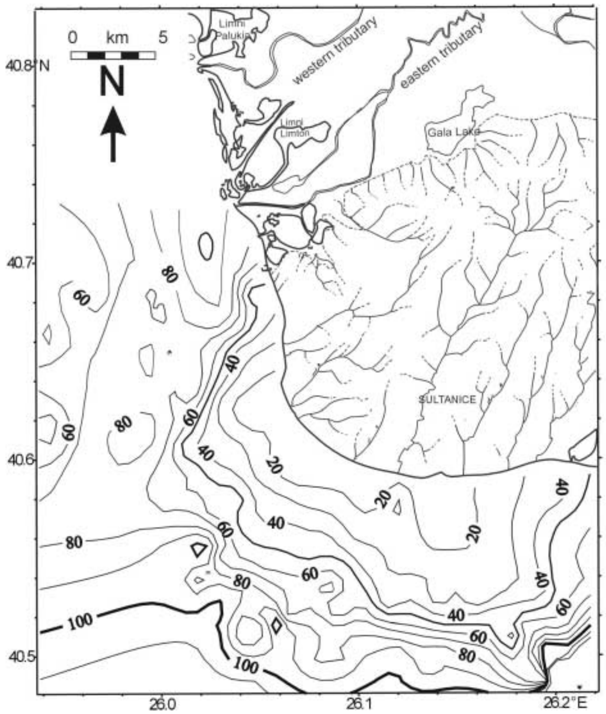

Fig. 13: Map showing the depth of the upper Miocene acoustical basement. Depths are given in m relative to present sea level. Contours are at 10-m intervals. Since there was no available velocity data, the average velocities were assumed to be 1600, 1800 and 2200 ms^{-1} for the depositional units A, B and C, respectively.

Conclusions

The ED is formed by the alluvial deposits (calcareous origin) of the ER incised into the lower valley which connects the Ergene Basin with the Aegean Sea. Its structural formation started to evolve around the end of the Oligocene. The initial volcanic activity phase started simultaneously with the folding of the Oligocene strata at the marginal parts of the valley (Figs. 2, 3, 11). Eroded deposits cut through the Oligocene formations fill the surrounding depressions. Intensive volcanic eruptions have taken place within the Miocene. In the late Miocene, the study area was relatively shallow and filled mostly with shallow water/coastal sediments. Tectonic movements caused open folds within the Miocene strata, with crest lines mainly tending N50˚E. Miocene seas turned into shallow swamps. Structural and lithological characteristics of the Miocene formations have seriously affected the establishment and development of the drainage pattern. A typical parallel pattern exists towards the Aegean Sea in the NE-SW direction, which conforms with the general direction of the folding axes of the Miocene strata, on the plateau surface (Fig. 2). In the late Miocene-early Pliocene, co-seismic uplifting resulted in dissection of the lower valley area. Meanwhile, transported sediments have been deposited in the basin located between the Hisarlidag mass and Samothraki island.

In the Pliocene, a basinwide transgression took place throughout the Aegean Sea and there was a break in sedimentation on the shelf areas (Fig. 3). Pliocene strata were deposited along the primitive valley of the ER. The ER with its tributaries were formed on the original inclination of the Lower Pliocene erosion surface. Faulting within the Middle Pliocene deformed the erosion surface and caused tectonic uplifting. These deformations triggered the second phase of volcanic activity, dispensing andesitic and basaltic lavas in the Upper Pliocene (Fig. 2). After the second phase of volcanic activity triggered by tectonic movements, the Hisarlidag massive completed its development in about the Middle Pliocene. The lower plateau surrounding the valley corresponds to the Upper Pliocene erosion surface, which was dissected with various streams later on.

The most prominent feature of the area is a non-depositional and denudation period which continued until the Pleistocene (Fig. 3). This hiatus meant that the area started to uplift after the early Pliocene. At present, even the Pleistocene units are located tens of meters above their depositional environments. Some recent research proposed that the ER was flowing into the Sea of Marmara through the Ergene River till about 1.5 Ma and filling the 2.5 km thick syn-transform sediments in the deep West Marmara sub-basin (OKAY & OKAY, 2001). The drainage diversion from this paleo-Maritsa to the present ER was inferred to have been related to the uplift of the northern margin of the Sea of Marmara region.

In the late Pleistocene during the Riss-Würm (stages 3-5) interglacial stage, 130,000-25,000 yrs BP), sea level fluctuations have resulted in dissections and formation of some depositional surfaces along the valley of the ER. The shelf area over an Upper Miocene basement was flooded and truncated by the Aegean waters. Meanwhile, the marine Unit C was deposited with a maximum of 55 m offshore of the present distributary mouth of the ER (Fig. 9c).

In late Palaeolithic (40,000-17,000 yrs BP) and Mesolithic (17,000-8,000 yrs BP) times the site overlooked a wide coastal plain. The northern Aegean formed broad coastal plains traversed by many rivers at 18,000 yrs BP (VAN ANDEL & SHACKLETON, 1982). Many Aegean islands, such as Samothraki and Gokceada (Imbros), were connected with the continent, and most of them were joined (ALPAR & DOGAN, 1999).

Between the last glacial stage and the following early phase of deglaciation (25,000-

13,000 yrs BP), the sea-level dropped as much as -120 m. The lowering of the sea level (Posttyrrhen Regression; stage 2) have caused dissections in between –110 and –115 m at the coastal part. With the shelf being exposed, the ER was able to erode its channel and transport most of its sediment load rapidly into the deep basins. Hence, the prograding deltaic sequences of Unit B were deposited, with a maximum of about 22 m west of the present distributary mouth of the ER (Figs. 9b and 11).

The post-glacial rise of sea level started at 16,000 yrs BP and the shelf area was inundated for the last time. On the basis of fossil contents of the cores recovered from the seismic survey area (ALPAR *et al.,* 1999), and correlating them with the known units distributed along the northern margin of the Gulf of Saros and dated by previous studies in the Strait of Canakkale (YALTIRAK *et al.,* 2000; 2002), a relative chronology can be established.

Rich gastropode and bivalve macrofauna in Unit A represent a transgressive unit deposited as beach facies on the shelf area 50 to 120 m below present sea level and during a time span of 16,000-10,000 yrs BP. On the basis of the 17,000-year glacio-eustatic sea level record (FAIRBANKS, 1989) and also the data from the northern margin of the Gulf of Saros (CAGATAY *et al.,* 1998), this sediment layer could be dated at about 12,000 yrs BP. In other words, by 11,000 to 12,000 yrs BP the transgression processes covered most of the previously exposed shelves. The shoreline nearly approached the present one at around 9,000 yrs BP (late Paleolithic), causing loss of coastal plains. An extensive Neolithic settlement may have been located 8,000-9,000 yrs BP along the river banks or bay shore. Sea-level rise began to decelerate from 8,000 to 6,500 yrs BP (STANLEY & WARNE, 1994). The existence of braided channels at –34 to –40 m depths support this event. Meanwhile, fine-grained transgressive deltaic sequence (Unit A) was deposited (Fig. 8). About 6,000 yrs BP, river flow decreased, but still filling the previously dissected valleys.

When the sea level reached -5 m. ca. 3,000 years ago and remained there until late Roman times (VAN ANDEL *et al.,* 1980), the late Holocene alluvial sedimentation was accelerated. This alluvial progradation has caused some classical archaeological sites along the western Turkish Aegean coasts, such as Ephesus and Troy to be located inland now. The sedimentation of the ER increased greatly during that period, filling the valleys. The comparison of the historical maps (Figs. 5c,d) and the chart reconstructed on the basis of archaeological data (Fig. 5a) indicates that the shallow and protected estuary at the mouth of the eastern ER filled rapidly and sand shoals were developed as a result of late Holocene alluvial accumulation.

All this depositional sequence is the Holocene posttransgressional unit, characterised by weak to moderate reflections within the continuous seismic records. Its sediments are principally pro-deltaic with muddy areas near the mouth of the ER and changing to muddy sand further out. This sequence becomes gradually thicker towards the present distributary mouth of the ER, depending on the redistribution by marine coastal processes and sediment input of the ER (Figs. 8, 9a, 10 and 11). The average thickness of Unit A the distributary mouth of the ER (Fig. 9a) changes between 3 and 6 m offshore, with a maximum of 9 m. This gives an average sedimentation rate of 40-90 cm ka^{-1} for the Holocene delta sediments, which may be as high as 140 cm ka^{-1} in front of the ER.

In conclusion, the ED was shaped as a result of late Holocene alluvial progradation interacting with the sea, which slowed below the present sea level stand. The tectonic uplift in the region may also give rise to this rapid fill. Even tectonic subsidence is common in the area with mean rates of 110-650 cm ka^{-1} for the Black Sea, 30-180 cm ka^{-1} for the northwest Aegean Sea and 50-460 cm ka^{-1} along the southwestern and western Anatolian coasts for the late Holocene, a regional tectonic uplift is a matter for the study area.

Elevated marine terraces, wide-spread Hisarlidag volcanics (Fig. 11), related palaeogeographic results (SAKINC & YALTIRAK, 1997), the decreasing trend of mean sea-levels in Alexandroupolis and Kavalla and also the tectonic settings and kinematics of the bending Thrace-Eskisehir right-lateral strike-slip fault to the north, and the North Anatolian Trough controlled by the North Anatolian right-lateral strike-slip fault to the south (YALTIRAK *et al.,* 1998) may support this idea. Beyond the effects of sea level and tectonic uplift, clearance of hinterland forests and land degradation might have accelerated these alluvial accumulation processes.

The protected embayment (or estuary) shown on the reconstructed chart (Fig. 5a) is now one of the most important wetlands in the world. Its delta and flood plains (Fig. 5b) are characterised by sandy offshore islands (Asanis, Karaviou, Xiradi), sand-dunes, halophytic marshes interspersed with saline lagoons and saltpans (Drana, Laki, Monolimni or Paloukia, Gala Lake, Pamuklu Lake, Sigirci Lake, Dalya Lake-Tasalti and Bucurmana), scattered freshwater areas fringed with reed-swamp, rivers, strips of gallery woodland and Tamarix scrub bordering the rivers, grassland or seasonal marshy areas. The enormous fill stuffing the present lower valley is not only formed by the fluvial processes, but also by the marine deposits that represent high stands. In other words, the evolution of the ED was due to fundamental changes in the relative influence of sediment input of the ER and redistribution by marine coastal processes. Even the tidal range is narrow, the little hypsometrical difference between the delta and the sea surface as well as the quiet flow of the river may occasionally cause flooding of the land areas. When the flow of the river is slow, especially in the summer, sea water enters the river and canals reaching a significant level up stream. Upwelling-favorable northerly winds generally generate a setdown. This circulation forcing mechanism allows for offshore expansion of the plume in an anticyclonic flow over the shelf. The isopach maps showing sediment thickness of the units (Fig. 9) may mean that this condition was also valid in the past. Meanwhile, occasional southerly winds (late winter and early spring) enhance the coastal currents and restrict offshore removal of low-salinity waters. These processes may have destructive effects on the coastal area. However, this seasonal period corresponds to the duration that sediment discharge rate of ER is the highest. That means that downwelling-favorable winds and erosive effects of coastal processes may not easily be able to wipe out the deltaic development, unless there is a substantial increment in human impacts.

Acknowledgements

This paper was partly supported by the National Geology and Geophysics Programme (YDABCAG-600/G, Coordinator : N. Görür) of Turkish Scientific and Technical Research Council and the University of Istanbul, Research Fund (O-593/160399). Warm thanks are also due to the logistic support of Ertugrul Dogan. The author appreciates the great care exerted in the field measurements by the crew of the R/V ARAR of the Institute of Marine Sciences and Management, Istanbul University. The manuscript benefited from the thoughtful comments of Daniel Jean Stanley, Valeri Gainanov, Cenk Yaltirak and Namik Cagatay. The author thanks J. Papailia and other anonymous reviewers of Mediterranean Marine Science for their helpful suggestions.

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