*Turkish J. Marine Sciences 5(3): 149-166 (1999)*

# Seismic Hazard Assessment in the Gemlik Bay Region following the 17 August Kocaeli Earthquake

# İzmit Körfezindeki 17 Ağustos Kocaeli Depremi ardından Gemlik Körfezi Deprem Çekincesinin Değerlendirmesi

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**Abstract**

Following the Kocaeli Earthquake (17 August 1999), the seismic reflection profiles gathered in the Gemlik Bay in 1984 were reinterpreted in order to identify the characteristics of the sedimentary depositional environments and their relations with the tectonic setting. A discussion followed about the implications for seismic hazard assessment in the Gemlik Bay region.

**Keywords:** Gemlik Bay, Marmara Sea, Earthquake, Seismic, Deltas

## Introduction

An earthquake of magnitude $M_w$=7.4 took place at Eastern Marmara Region (40.70°N, 29.91°E) on August 17th, 1999, at 3:01:39.07 am local time. The focal depth is rather shallow (about 16 km). It occurred on the northern strand of the North Anatolian Fault Zone and hereinafter will be termed as the Kocaeli Earthquake. Following the Kocaeli Earthquake, the tectonic setting of the Gemlik Bay region which is intersected by the middle strand of the NAFZ became more important (Figure 1).

The City of Gemlik (Kios) was founded by the Militians in 630 BC and for about 3000 years it was one of the centres of politics, economics and culture. The old city (acropolis, long city walls and quarries) was built further to the east than the present one. The acropolis was built on a hill at the highest point of the area, the city lying beneath it, near the beach. The city walls ran down to the beach.

Historical earthquake records of Turkey extend from 1900 to BC 2000. Even deficient or unreliable to some extend, the existing catalogues of historical earthquakes give some activities in 29, 128 (west of the İznik Lake), 362, 368, 1419 (Geyve), 1857 (west of the İznik Lake) in the surroundings of the İznik area (Ambresseys and Finkel, 1991; Guidoboni et al., 1994). There is not any important instrumental earthquake record ($20^{th}$ century) for the area.

On the other hand; according to *Stavron* book XIII page 628, the first earthquake in Kios (Gemlik) was on 17 AD during the Emperor Tiberius. Later David the Paflagonius refers that during 861 and 869 AD were strong earthquakes and the one in 869 was so strong that even the church of Agia Sophia in Istanbul was damaged. Most famous one was in 174 AD during the Emperor *Adrianus*, who visited Nikomedea, Nicea and Kios to offer great help, because the big disaster (Le Bas Philippe, Asie Mineure "L'Universe pittoresque" Paris 1869; Voyage archeologique en Grece et en Asie Mineure; Monuments d'antiquite figurew, Paris 1837; Tacide Annales, livre II).

As paleoseismology in Turkey is under development, source of most of prehistoric earthquakes were recognized by trenches excavated in the North Anatolian Fault Zone by a few researchers. Tsukuda et al. (1988) and Barka (1993) have studied the middle strand of the NAFZ in detail on which some trenching excavations have also been arranged to estimate the recurrence intervals of large earthquakes. Unfortunately, the results did not perform harmonious combinations. Ikeda (1988) and Ikeda et al. (1989, 1991) concluded that the last earthquake had occurred between 1500-1700 AD. Barka (1993) reported that the only large earthquake during the last 4000 years was taken place between the years 250 BC–700 AD.

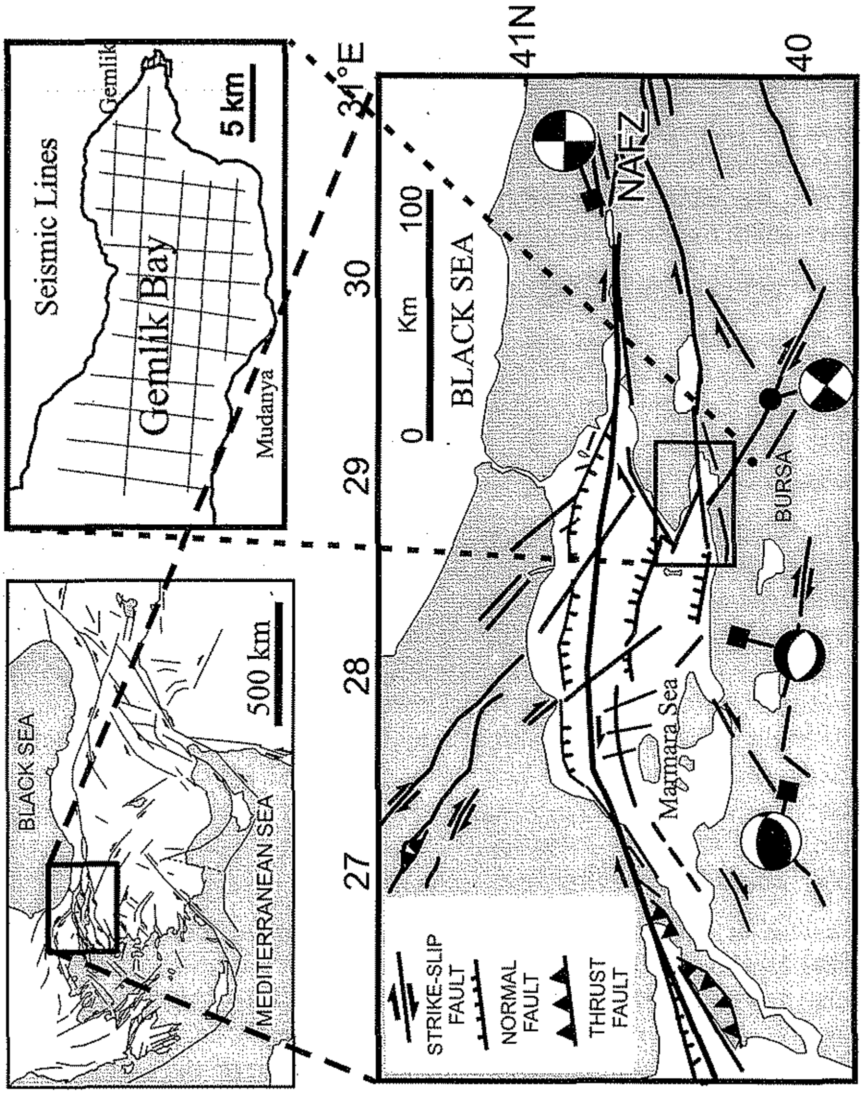


Figure 1. Study area. Modified from Yaltırak et al, in press. Fault plane solutions are from Taymaz et al. 1991).

In past, the geometry and kinematics of the Gemlik Bay were also studied by many authors (Kurtuluş, 1985; Özhan, 1986; Barka and Kuşçu, 1996) by interpreting the very same marine reflection seismic data set we utilised in this paper. Previous studies showed, in summary, that the Gemlik Bay area is a pull-apart structure. However, even though it is well known that the Gemlik Bay receives significant amount of detrital input from the rivers, the characteristics of the sedimentary depositional environments in the Gemlik Bay have not been known in detail.

## Paleogeographic evolution and deltas

Following the closing of the Intra-Pontide Ocean terminally (late Oligocene), a consequent suture zone formed largely in the area where the Marmara Sea developed (Okay and Tansel, 1992). The global sea level changes and the tectonic activities in forms of tilting, block rotation or subsidence, exercised control over all water movement in the basin.

The paleogeographic history of the Marmara Sea Basin began when the region was flooded for the first time by the Mediterranean waters in the late Serravalian. Later, from time to time, the saline Mediterranean and the brackish Paratethys / Black Sea occupied the basins of the Marmara Sea solely or conjointly. From late early-middle Pleistocene to present, at least 3 connection with the Mediterranean waters established via the valleys of the Strait of Çanakkale (Yaltırak et al., in press). The connection with the Paratethyan / Black Sea waters, on the other hand, was first established in the Pontian and except a short disconnection during the last glacial maximum, it appears to have continued up to the present (Görür et al., 1997; Emre et al., 1998).

Smith et al. (1995) observed a clustering of the maximum water depths of delta tops at about –100 m below the present sea level and inferred that relative sea-level rise in the southern margin of the Marmara Sea has been about 90 m since the last lowstand. In other words, during the late glacial maximum, under the control of the sill in the Strait of Çanakkale, the water depth at the fresh Marmara Lake was about 90 m lower than the modern sea level. Later, the global sea level started rising (18 $^{14}$C kyear BP) and the Aegean waters spilled over the Strait of Çanakkale into the Marmara Lake. The sea level rose to the sill depth of the Strait of Çanakkale at 11 kyear BP. The rising seawater in the Marmara Sea Basin reached to the sill depth in the Strait of Istanbul (~9.5 kyear BP). However, because of the vigorous outflow of the rising Black Sea waters, derived from melting of northern European ice sheets, through the Strait of

Istanbul, the seawater of the Marmara Sea was unable to penetrate the Black Sea until about 7 kyear BP (Aksu et al., 1999). Following the marine waters were finally able to penetrate into the Black Sea (~7 kyear BP), the single-layer flow turned into the modern two-layer flow.

Recently, using single-channel air-gun reflection profiles, Aksu et al. (1999) have studied the present-day Kocasu delta (Late Holocene delta, (Δ1) and older deltas (Δ2-Δ10) of the palaeo-Kocasu river). They defined the unconformities between the stacked deltas, assuming that these unconformities were time equivalent to the lowstand unconformities in the eastern Mediterranean and they dated them. A summary of the deltas and unconformities is given below as defined by Aksu et al. (1999);

(Δ1) : The 20-30-ms thick lobe of Δ1 occupies the deep bathymetric depression in the Gemlik Bay (Figure 2).

α0 : Beneath Δ1, the Holocene transgressive surface forms a strong unconformity reflection. This surface, which occurs within a few metres of the seabed, also represents the subaerial erosional surface (12 kyear BP), particularly in the modern nearshore zone.

(Δ2) : It is absent in the Gemlik Bay. This delta sequence deposited on the prodelta slopes across the southern shelf ~25 kyear BP when the Marmara Sea became isolated from the Aegean Sea because the global sea level dropped to below the sill depth of the Strait of Çanakkale.

(Δ3) : It is also absent in the Gemlik Bay, but deposited just outside the bay within the depressions of former deltas.

(Δ4) : At the western part of the Gemlik Bay, 50-ms thick delta lobe (Δ4) underlie Δ1 (Figure 2). The Δ4 is developed east of the present-day mouth of the Kocasu River.

(Δ5) : There are two separate lobes of Δ5; its eastern lobe is developed mainly in the Gemlik Bay and forms a basin fill along the central axis of Gemlik Bay (Figure 2). The central thickness of the eastern lobe of Δ5 reaches ~40 ms.

The deltaic successions of Δ2 and Δ6, between the unconformities of α1 and α0, developed between 65-12 kyear BP.

α1 : It is a shelf crossing unconformity (128 kyear BP), beneath the deltaic succession of Δ5 in the Gemlik Bay, approximately 25-35 m below the strong unconformity α0, at elevations ranging from -100 to -120 m.

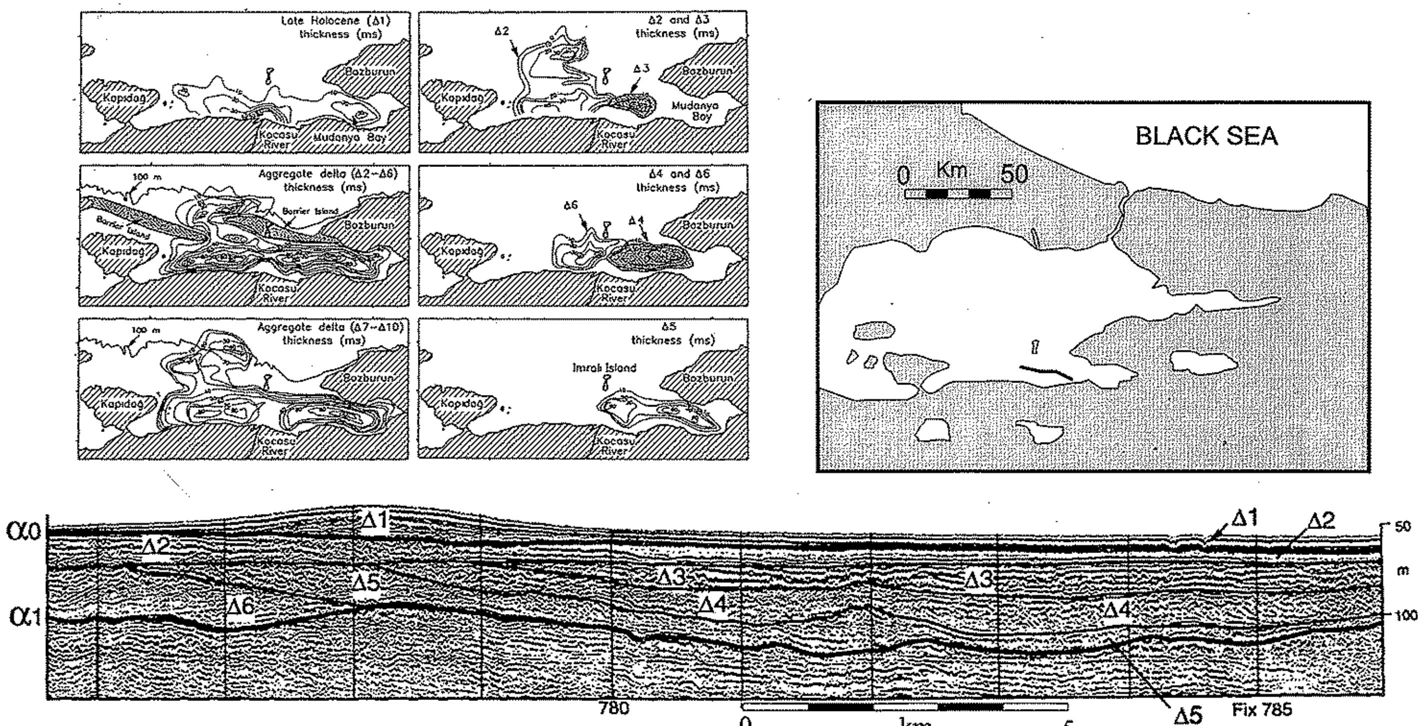

Figure 2. Deltas of Southern Marmara Sea. Inset shows the location of the study area and the location of the seismic section (Aksu et al., 1999).

(Δ7-10) : An aggregate delta thickness (Δ7-10) reach 100 ms in the Gemlik Bay (Figure 2). Prograded delta successions Δ7 to Δ10 developed from ~186 to 128 kyear BP.

α2 : It is another shelf crossing unconformity (~250 kyear BP) on the southern shelf, beneath the deltaic successions of Δ7-10 and at elevations ranging from -90 to –150 metres.

This stacked delta structure illustrates the effects of regional tectonic subsidence and a subsidence rate was calculated as 0.2 m/1000 year during Riss-Würm interglacial stage (Aksu et al, 1999).

In this paper, taking into account the recent studies, we will interpret the high resolution seismic reflection profiles in the Gemlik Bay, which were collected by the General Directorate of Mineral Research and Exploration (MTA) and brought out in some reports and papers, from a different viewpoint; the depositional environments in the Gemlik Bay, their relations with the tectonic setting and finally the seismic hazard assessment in the Gemlik Bay region.

## Material and Method

In 1984, approximately 560 line-km of single-channel high-resolution seismic reflection profiles were acquired during the cruise of the research vessel R/V MTA Sismik-1 of the Mineral, Research and Exploration in the Gemlik Bay (Figure 2). For positioning, a trisponder system was used with an accuracy of ±10 meters. During seismic works, depth data were also gathered using Atlas Deso-10 echo-sounder. The seismic reflection profiles were collected using a 500 joule sparker and a 20-element, 20-m-long hydrophone streamer. Analogue data were recorded on a EPC 4100 graphic recorder for 500 ms (two-way-time). The band-pass filter cut-off frequencies were 150-600 Hz. The vertical exaggeration of the seismic sections is about 15.6. These data were used in technical reports, academic studies and some papers (Kurtuluş, 1985; Özhan, 1986; Barka and Kuşçu, 1996).

Following the Kocaeli Earthquake, the authors have believed that the MTA's marine seismic data in the Gemlik Bay deserved much more detailed interpretation. Then, 20 high-resolution seismic profiles (560 km) were scanned (600 dpi) to bit-map images. In addition, a detailed isopach map was prepared using the precise depth data digitised from the echo-sounder charts and also considering classical navigation maps.

## Bathymetry

The maximum depth in the bay is ~110 m (Figure 3). The submarine valleys are the most interesting geomorphic and neotectonic features of the Gemlik Basin. Relative depths of the submarine valleys average 70-80 m. The most evident valley networks are placed off Kapaklı (28°58") to the north and wide-basin valley networks to the south and southeastern margins.

A NW-SE trending ecliptic trough is bounded by the –65 m isobath. It lies obliquely to the deep troughs in the Marmara Sea. While its north and south bounding edges are symmetrical, western and eastern sides are evidently at an angle.

## Seismic Data

The seismic profiles provide details on sedimentary deposits and erosional surfaces up to 150-200 m below the seabed. Throughout this paper, two-way travel times are converted to depths below sea level using a typical interval velocity of 1500 m/s for sea water and for the near-surface siliclastic sediments in open marine conditions.

The records were generally not in good quality and deteriorated in areas with steep sea-floor gradients. This is mainly because the source wavelet is not well defined during the survey. The acoustic basement, which should be deeper than 300 ms, can not be identified on the seismic sections due to the masking effect of the strong multiple reflections of the seabed. However, it can be suggested that the continental (possibly Upper Pliocene) and marine (Pleistocene) deposits lie on the basement that can not be easily traced on the seismic sections.

Even so, low-frequency and high-amplitude reflections within the upper Pliocene and Pleistocene deposits have been identified and delineated on the sections (Figure 4a, b). The layers are not always continuous and show some discontinuities, but not so important on a regional scale.

An E-W trending strike-slip fault cuts the alluvial fans at the eastern margin and just ahead of that it was overlain by younger fans (Figure 4a). Such kind of features are evidences of the tectonic activities in Plio-Pleistocene. The basinward slopes of the graben-bounding normal faults in the Gemlik Bay indicate dominant extensional forces (Figure 4a).

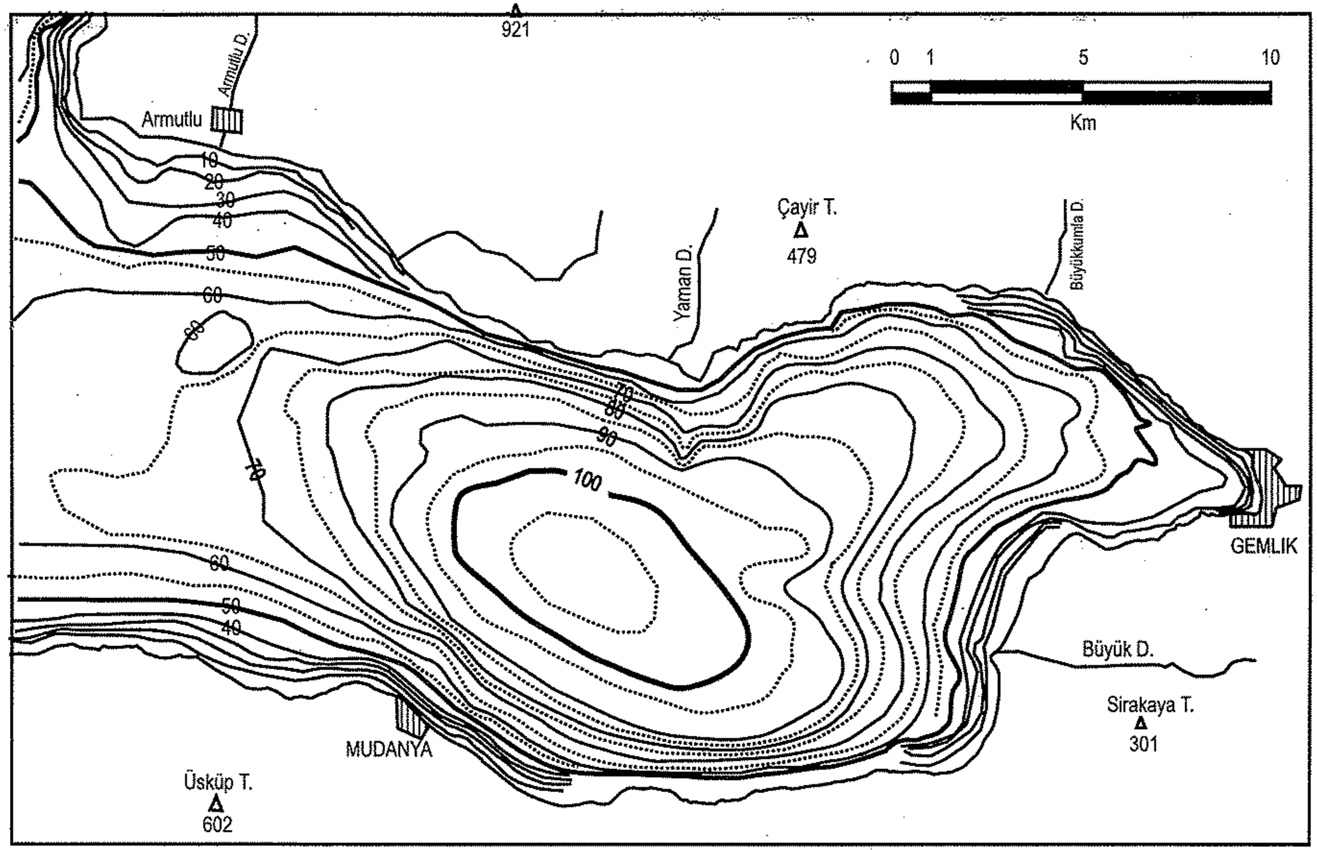


Figure 3. Bathymetry of the Gemlik Bay

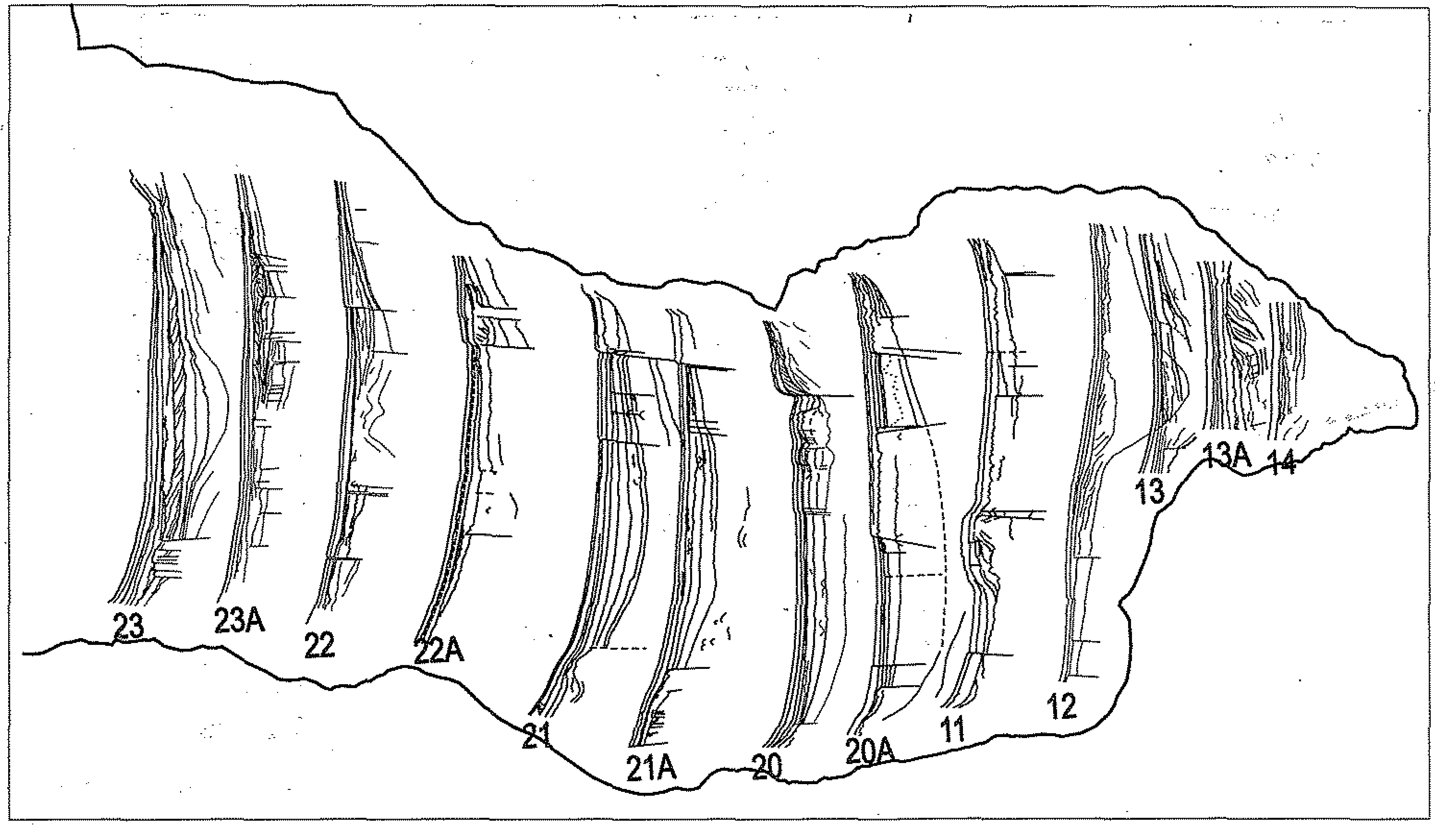


Figure 4a. Seismic fence diagram from east

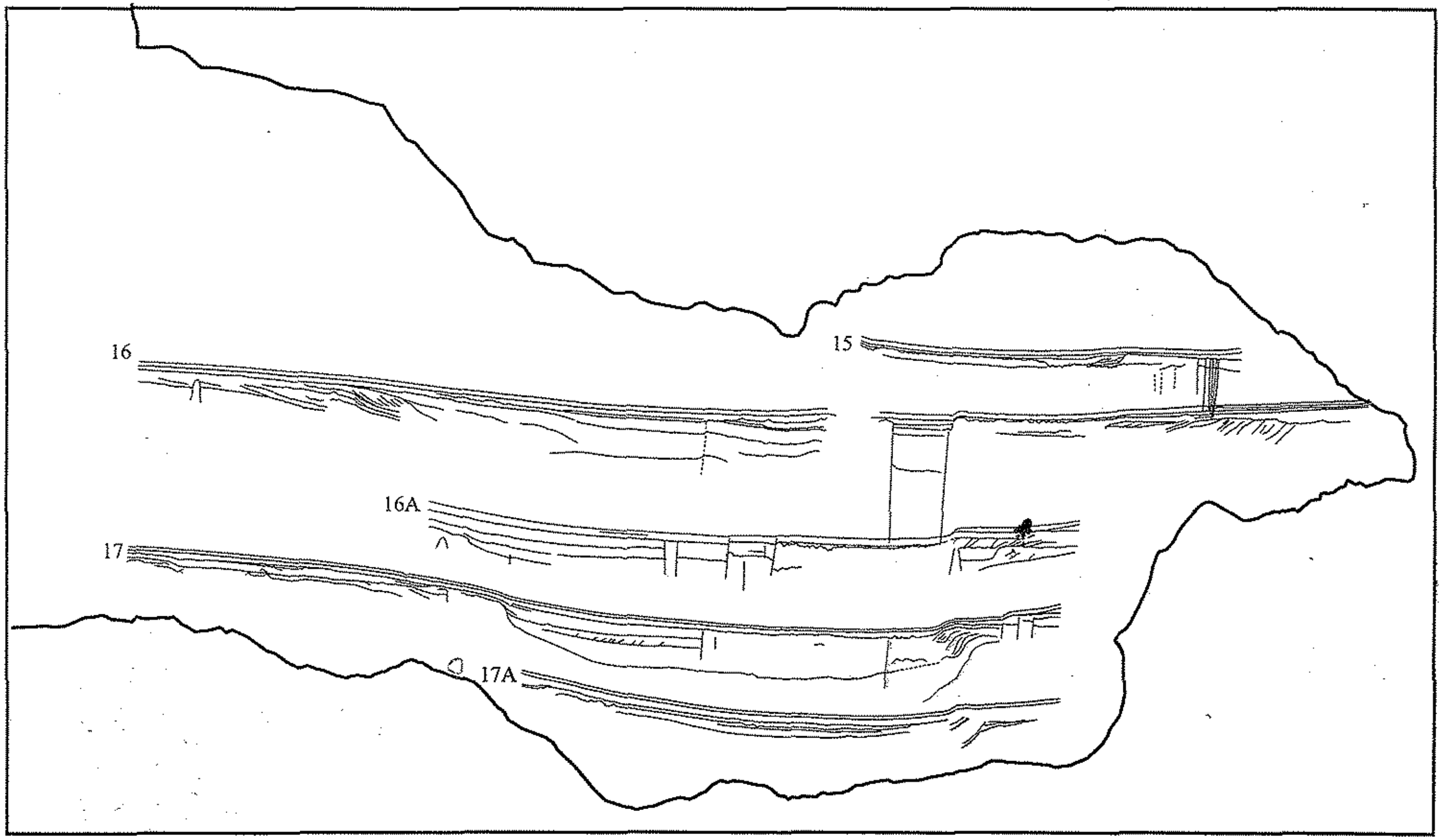


Figure 4b. Seismic fence diagram from south.

The seismic evidences of the lacustrine deposits, which deposited following the development of the subduction zone, indicate that there were palaeo-lakes in the area during Pliocene (Figure 5). Sea level lowstands in the Marmara Sea Basin possibly caused a palaeo-lake in the Gemlik Bay. Later, the Palaeo-Gemlik Lake formed a depositional area among the continental deposits. Deltaic units were placed as a result of drainage systems developed in this period. This depositional cones were affected by some faults which indicates the tectonic activities in Plio-Pleistocene.

The delineation of the deltaic sequences in the Gemlik Bay are given in Figure 5. The characteristic sigmoidal internal reflections in the deltaic successions of E, D and C indicate that they are basinward prograding fans and may be coincident with the deltaic fans (Δ4), (Δ5) and (Δ7-Δ10) described by Aksu et al. (1999) for the Kocasu delta, respectively. The distribution of the deltaic successions of E, D and C to the west show a smooth clockwise rotation with time while their equivalents at the eastern part of the Gemlik Bay occupy almost the same place (Figure 5).

On top of these deltaic successions is the seismic unit B with moderate-amplitudes. It is characterised by internally continuous and parallel reflectors with a thickness of about 30 m along the basin.

The uppermost seismic reflections on the records consist of a thin (20-30 ms) muddy blanket characterised by weak and internally parallel reflectors (unit A). This unit corresponds to Δ1 of Aksu et al. (1999) and deposited after the last important sea level lowstand (α0, 17-18 kyear BP) (Chappell and Shackleton, 1986; Table 2; -130 m / -150 m) until present. Since last Mediterranean transgression reached into the study area about 11 kyear BP, the unit A initiated deposition at this time.

## Tectonic Setting

The subsidence of the Marmara Basin was started from Oligocene and continued through the Neogene. However, the main tectonic movements took place between the early Pliocene and the last glacial period.

The middle strand runs from Geyve through Mekece and passing south of İznik Lake. From the İznik Lake towards the Gemlik Bay, the middle strand splays into two faults; NE-SE trending Gençali Fault and the E-W trending Gemlik Fault (Figure 6). The Gençali Fault forms the southern boundary of the Gemlik Bay (Tsukuda et al., 1988). The middle strand goes into the Marmara Sea, appearing near the Bandırma Bay and cutting

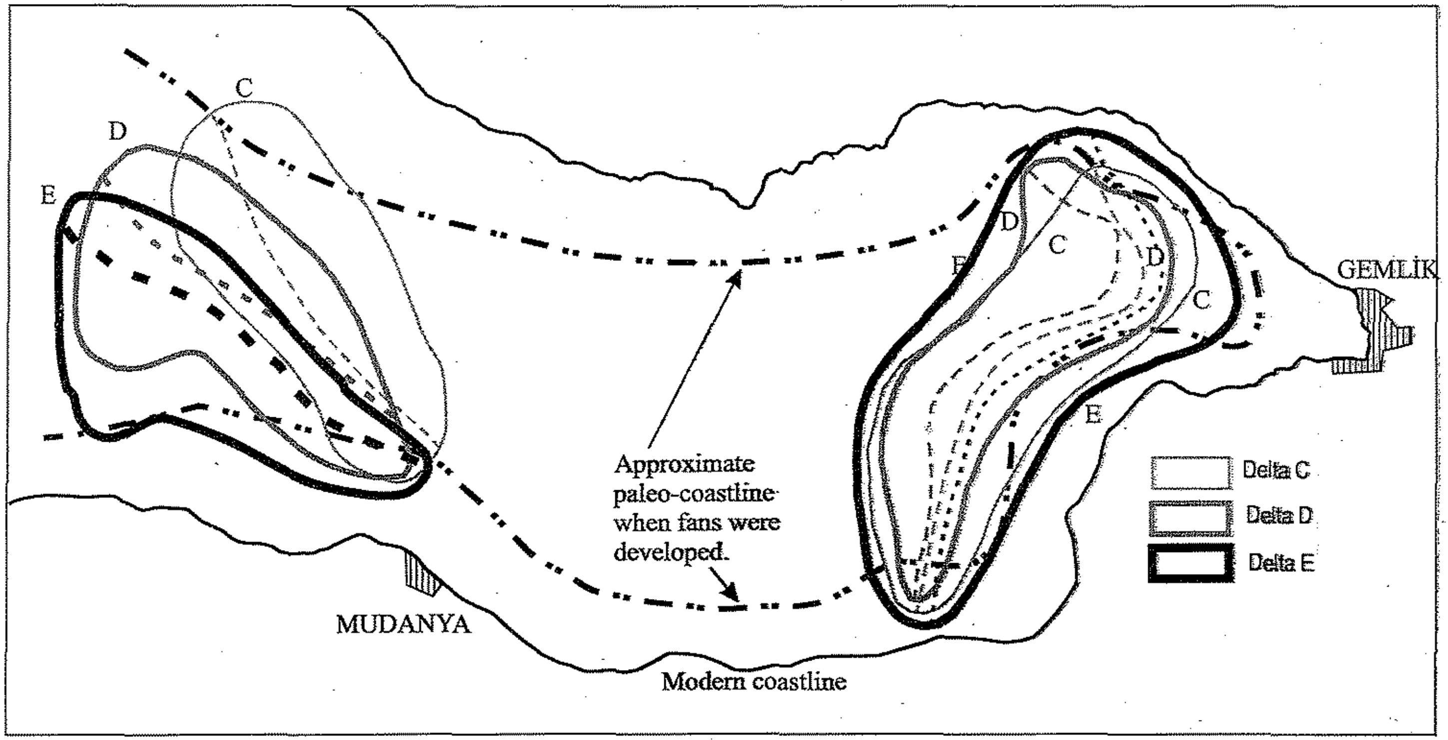


Figure 5. Distribution of the deltaic sequences. Figure also shows the approximate paleo-coastline during these deltas developed.

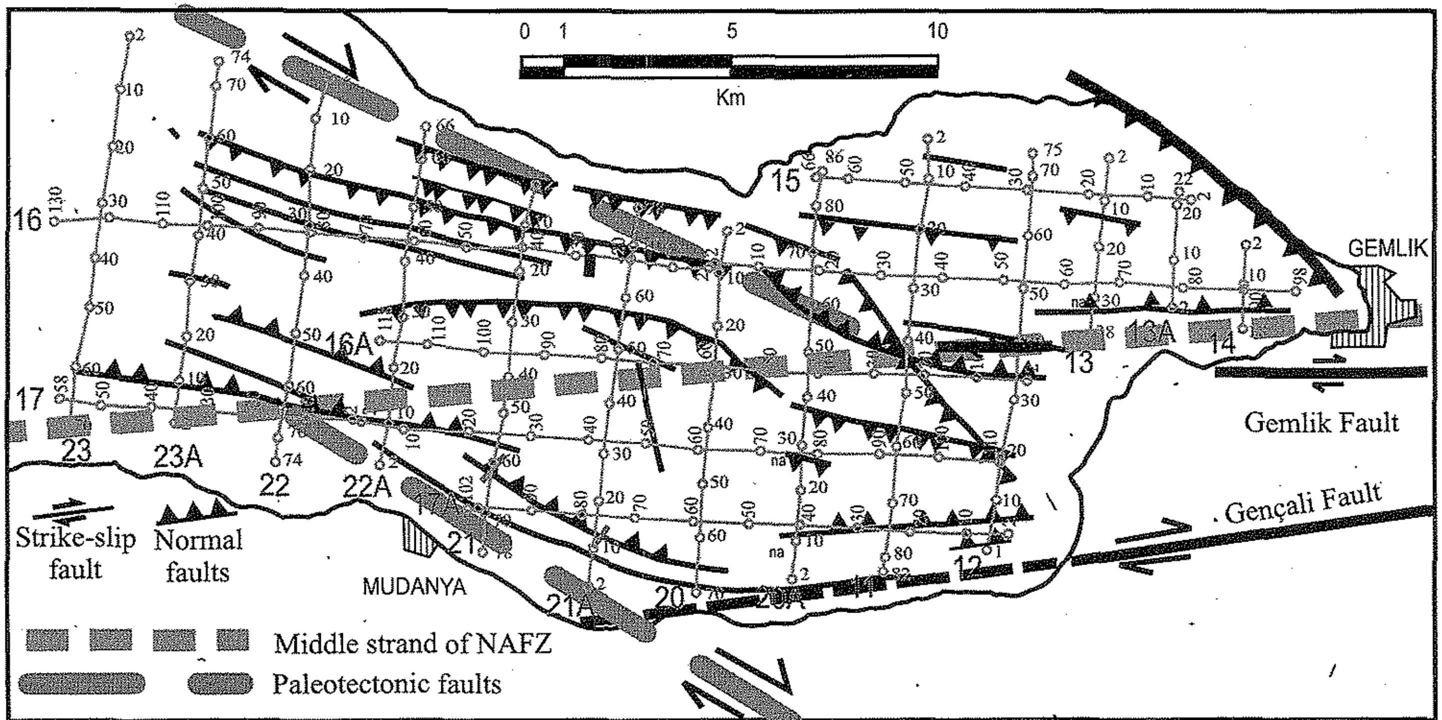


Figure 6. Structural elements in the Gemlik Bay. NW-SE trending faults have larger vertical offsets.

the Kapıdağ Peninsula, continues in the Biga Peninsula and then enters the Aegean Sea.

The Gemlik Bay forms a depression (graben) area in the Marmara Sea which itself is a depression basin as well. Most of the graben-bounding faults take place within the deformation area (Figure 6). The vertical components of the faults in the bay are dominant to the strike-slip ones. In other words, contrary to the well-known features of the NAFZ - the right lateral components of the NAFZ are dominant especially at the eastern parts of the region – the normal faults are dominant in the Gemlik Bay.

The spatial distribution of the graben-bounding normal faults indicates a sigmoidal distribution at the sea bottom.

The tectonic setting of the area, shearing forces of the strike-slip fault segments which were believed to be responsible for the evolution of the Gemlik Bay and the normal faults in the bay (Figure 6) should have played a deterministic role on the depositional distribution of the deltaic sediments with different slopes as given in Figure 5. In addition, the submarine valleys exhibit complicated scenes in these shear zones. The tectonic movements along the depression fields deeply affected the river network.

## Conclusion

From the end of Oligocene to the tectonic movements occurred at the end of Pliocene, the Gemlik Bay filled with the fluvio-lacustrine fine-grained clastics. Younger sediments overlie unconformably the upper Pliocene erosional truncation surface. This indicates that the tectonic activities also persisted in Quaternary. Seismic reflection terminations within the seismic sequences and the stacked fans lying on the Pliocene depositional environments indicate that global sea level fluctuations were important on sedimentation of the upper Pleistocene deltaic deposits (Figure 5).

The chaotic acoustic basement (upper Pliocene ?) was overlain by the deposits of calmer depositional environment (upper Pleistocene ?). Normal faults are dominant in the gently folded upper Pliocene deposits, showing compressional forces (Figure 6). On the same figure, the hypothetical paleotectonic faults (Yaltırak et al., 1999), which have been cut by the middle strand of the NAFZ and shifted about 15 km, have been superimposed.

The topsets of the deltaic sequences E, D and C which show clockwise rotational development at the western part of the Gemlik Bay (Figure 5) correspond with the western intersection point of the paleotectonic fault and the middle strand of the NAFZ (Figure 6). In addition, all of the topsets of the deltaic sequences at both sides in the Gemlik Bay (Figure 5) correspond with the eastern and western junctions of the graben-bounding faults (Figure 6).

Even though the Gemlik Bay is placed on the westward extension of the middle strand of the NAFZ (Figure 6), historical and paleosismological data indicate that the recurrence interval of large earthquakes around the Gemlik area is in the order of a thousand years (Ikeda et al., 1989, 1991; Ambresseys and Finkel, 1991; Barka, 1993 and Guidoboni et al., 1994). This confirms that the faults in the Gemlik Bay are less affected, if compared to the north strand, from the single North Anatolian Fault plane (Alpar, 1999) which occupies a certain position 6-7 km below the rhomboidal deep marine sub-basins in the modern Marmara Sea.

On the other hand, the micro-earthquake activities in the Gemlik Bay region indicate that the area is still active. The tectonic activities are mainly in case of normal faulting which are dominant in gently folded upper Pliocene deposits (Figure 6). However, considering the implications for seismic hazard assessment, the right lateral middle strand of the NAFZ, which lies below these graben-bounding normal faults, is believed being much more important for longer terms. The middle strand is also believed to continue on the westward side of the Gemlik Bay, possibly below the EW trending graben structure (4-5 km deep) modelled by Demirel (1999) using gravity data.

Assuming that the middle strand of the NAFZ shifted the NW-SE trending paleotectonic fault about 15 km westward (Figure 6) within 3-4 million years (fault zone initiation ?) period, the seismic hazard in the Gemlik Bay region may be estimated. If we consider the earthquake repetition time is roughly 1000 years, about 3000 large earthquakes should hit the Gemlik Bay region during last 3 million years. These assumptions give lateral displacements of about 3.7-5.0 m which should cause very large earthquakes. If the repetition time had been 2000 years, the lateral displacements would be as high as 7.5-10.0 m, which seems not so realistic. Hence, we may consider the earthquake repetition time to be about 1000 years for the Gemlik Bay region.

The stratigraphic features of the seismic units and deltaic successions in the Gemlik Bay, their spatial distribution, temporal displacements, relations with the regional tectonic setting and earthquake risk deserve further studying in the light of modern digital single and multi-channel high-resolution seismic prospecting and data processing.

## Özet

Toplandıkları 1984 yılından hemen sonra üzerinde çalışılan, ancak kesitlerin yeterli kalitede olmaması nedeniyle yayın haline dönüştürülmesi gelecek çalışmalara bırakılan Gemlik Körfezi sismik kesitleri, gerek bölgedeki deprem çekincesinin 17 Ağustos 1999 Kocaeli Depreminin ardından değiştiği düşüncesiyle, gerekse Gemlik Körfezi'nin Marmara Denizi tektonik yapısını çözecek kilit noktalardan birisi olduğu anlayışıyla, yeniden değerlendirilmiştir. Deprem tekrar sıklığı bin yıl mertebesinde olan Gemlik Körfezi'nde her ne kadar kısa dönemli periyotlarda fazla etkili olmayan normal faylar etkili ise de, uzun dönemde doğrultu atımlı fayların önemi vardır. Deprem çekincesinin daha detaylı belirlenebilmesi için, Gemlik Körfezi, sedimentolojik özellikleri, tektonik oluşumları ve bunların birbirleriyle olan ilişkileri açısından günümüzün modern yüksek çözünürlü sığ ve derin sismik sistemleri ile bir kez daha araştırılmalıdır.



## Acknowledgement

The authors wish to express their sincere thanks to the marvellous team (in 1980's) of the research vessels *MTA Sismik 1*; her captain, scientists, technicians and the crew for their unceasing and talented work at sea. They always free-lanced and knew thoroughly what they were doing.


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*Received 22.10.1999*
*Accepted 30.11.1999*

*Turkish J. Marine Sciences 5(3): 167-170 (1999)*

# The Bathymetry of the Izmit Bay

# İzmit Körfezi'nin Batimetrisi

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**Abstract**

The Bathymetrical Map of the Izmit Bay originally scaled 1/50.000 was prepared in 1997. The map provides information that shows the morphotectonic features of the sea-floor of the Izmit Bay.

**Keywords:** Bathymetry, isobath, Izmit Bay, Marmara Sea, morphotectonic features, sea floor, Northern Anatolian Fault

## Introduction

One of the resources used in order to determine the geomorphology of sea floor is a bathymetrical map. The features on this kind of maps shown by isobaths, such as getting closer, getting farther or curling etc. provide the information about the geomorphology of the area just the same as the topographical maps used in land studies.