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LATE QUATERNARY SEDIMENTATION IN THE STRAIT OF BOSPORUS: HIGH-RESOLUTION SEISMIC PROFILING

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Abstract

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Uniboom profiling supported by borehole data in the southern part of the Bosporus revealed that close to its banks Holocene sediments are underlain by the faulted slopes of a valley cut in many places into the Palaeozoic bedrock. Away from the banks the sediments thicken and their boundary with the underlying Pleistocene deposits is defined by an erosional surface at about the 80 ms TWT. This surface marks the initiation of marine currents in the valley as its deeper parts began to be submerged in the late Pleistocene to early Holocene (marine spillway phase). The upward-coarsening Holocene sequence records the subsequent evolution of the present two-way marine strait. The sequence is divided into three seismostratigraphic units. The lowest one constitutes an onlapping fill deposited under salt-wedge estuarine conditions when sediment input and runoff discharge were high and sealevel was still rising but the flow of Mediterranean waters into the Black Sea was not continuous. It fills low areas and is well developed under a sill close to the southern entrance of the Strait. As sealevel rose closer to its present level and the rate of terrigenous input decreased in the middle Holocene, this unit began to be eroded and reworked by strong bottom currents. The second unit records the transitional stage between the estuarine phase and the present two-layered current system. Its seismic facies and sedimentary features indicate deposition under high-energy conditions and shifting loci of accumulation. The third (surficial) unit is a thin depositional lag rich in shell fragments blanketing the other two units. It records deposition under fully marine conditions and flow of powerful bottom currents during the last few millenia.

Introduction

The Strait of Bosporus is a 31 km long meandering waterway carved out in a Palaeozoic and Cretaceous terrain between the Black Sea and Marmara Sea (Figs.1 and 2). It has a variable width (0.7–3.5 km) and a very irregular bottom profile with several major deeps and sharp bends along its course (Gunnerson and Ozturgut, 1974; Scholten, 1974). Its maximum depth (110 m) occurs in one of these depressions about 10 km to the north of its southern entrance. Its minimum depth (26 m) is registered on the crest of a major sill about 3 km to the north of the same entrance (IMS METU, 1985a, figs.1 and 4). The continuation of the course of the Strait of Bosporus in the form of submerged valleys across the shelf at its junctions with the adjacent seas*. These features, the meandering course and the localized deeps are usually considered as relics from the times of fluvial erosion in an emergent Quaternary valley before transformation into a two-way transit marine strait during the Holocene transgression (Caspers, 1957; DAMOC, 1971; Scholten, 1974; Erinc, 1978).

However, as Scholten (1974, p.117) men-

*Admiralty Chart No. 2604 and Turkish Navy Chart No. 2921.

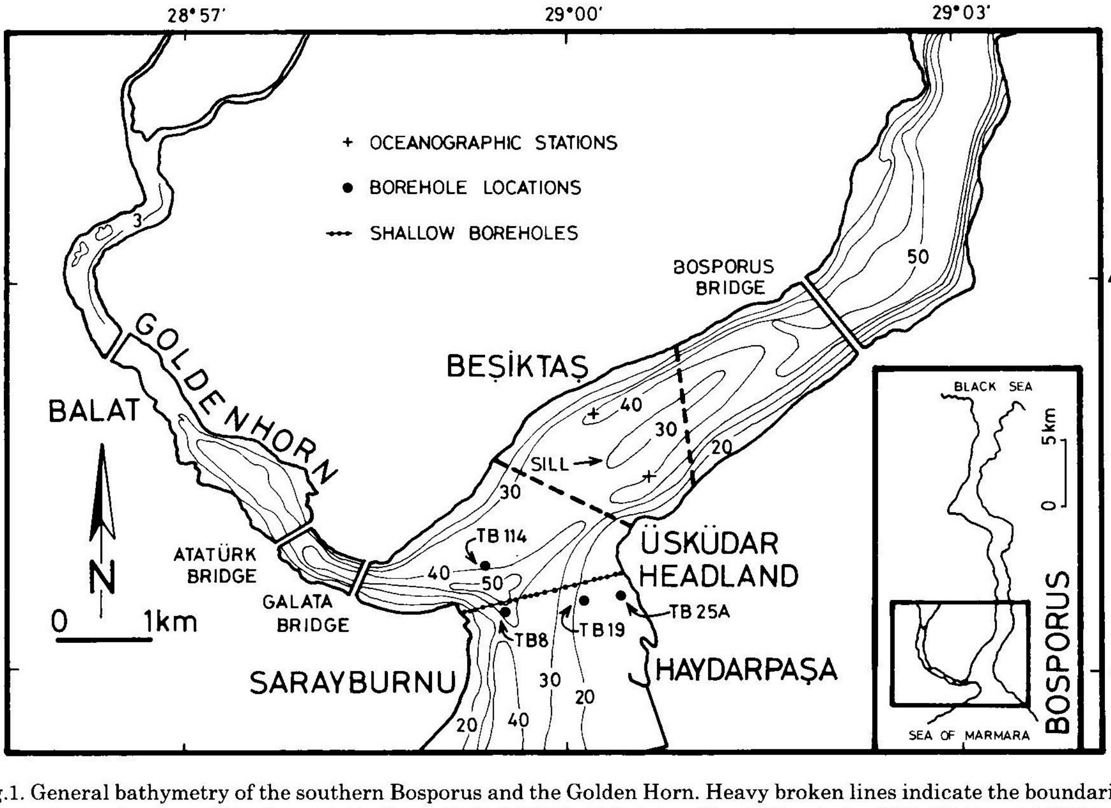

Fig.1. General bathymetry of the southern Bosporus and the Golden Horn. Heavy broken lines indicate the boundaries of the surveyed area between Uskudar and Besiktas (IMS, METU, 1985a) (see Fig.4). Boreholes TB8, TB19 and TB25A are from Eroskay and Kale (1986). TB114 is from Meric et al. (1988). Shallow borehole data from Timur (Unpubl.). Current measurements at the oceanographic stations are reported in IMS, METU (1985b). Contours in metres.

tioned, the geological evolution of the Bosporus valley goes beyond fluviatile erosion and flooding of a "stream valley" in the late Quaternary. The available marine geophysical and geological data show that the thickness of the sub-bottom sediments vary considerably from place to place and may exceed over 100 m at some localities (Scholten, 1974, fig.3; Timur, 1977). Deep-penetration (airgun) seismic profiling between Sarayburnu and Haydarpasa (Fig.1) has shown that the sedimentary sequence on the Carboniferous rocks can be up to about 110 m thick (Ulug et al., 1987). Drillholes from this area (Fig.1) (Eroskay and Kale, 1986; Meric et al., 1988) also show that the Holocene sequence is over 40 m thick in some places. However, the detailed stratigraphy and sedimentary evolution of the succession beneath the floor of the Strait of Bosporus remains to be studied.

In this paper we discuss the late Pleistocene and Holocene morphological and sedimentary evolution of the southern part of the waterway in the light of bathymetric, sidescan sonar and high-resolution (Uniboom) seismic reflection data from an area close to the entrance to the Golden Horn (Fig.1) (IMS, METU, 1985a; Okyar, 1987; Okyar and Alavi, 1987) and another area farther south between Haydarpasa and Sarayburnu and off Kurbagali River on the Anatolian bank (Fig.2) (IMS, METU, 1986). The seismic profiles and sonographs have been interpreted in relation to the results of the borehole studies and the characteristics of the surface sediments in the area (DAMOC, 1971, and Unpubl.). This has enabled us to present a model of the sedimentary evolution of the Strait of Bosporus since the time of inception of the present system of water exchange between the Black Sea and the

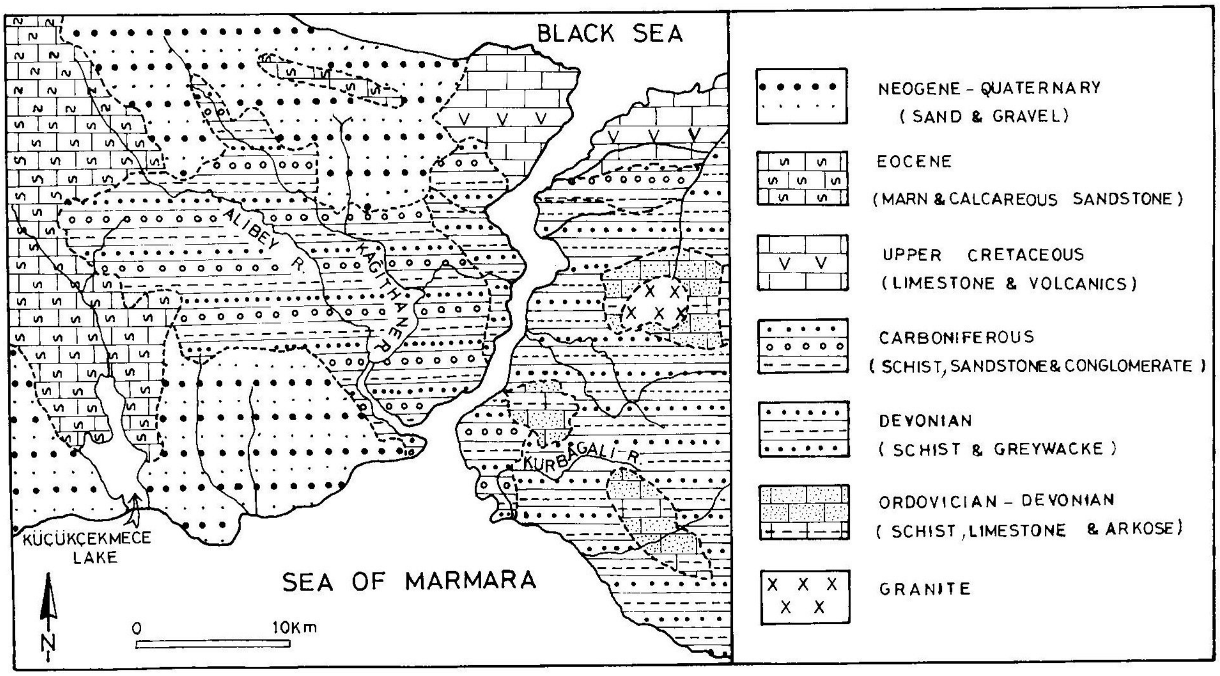

Fig.2. General geology and drainage system of the Bosporus region (from Sayar, 1976).

Mediterranean Sea at about 10 ka B.P. (Ross and Degens, 1974; Scholten, 1974; Degens and Stoffers, 1980).

Methods

In the area between Uskudar and Besiktas 35 Uniboom and 13 lines of sidescan records were collected (IMS, METU, 1985a) with an average distance of 25 m between fix points (Fig.4). Position finding was accomplished with a Trisponder system and the locations of the transponders were pin-pointed on the 1/1000 plans of the banks of the Strait. Bathymetric data were obtained with a portable echosounder (100 kHz) and a PDR (210 kHz) and subsequently corrected for the variations in the sound velocity. The margin of error was calculated as varying between −15 to +100 cm depending on depth and variations in the physical properties of the water column measured at the time of the survey (March 1985) (IMS, METU, 1985b; Okyar, 1987). Tidal fluctuations of the sea surface are less than this margin of error (Gunnerson and Ozturgut, 1974).

In the area between Uskudar and Besiktas the Uniboom (300 J) and sidescan sonar were towed along the course of the Strait because of the heavy traffic and strong surface currents (up to 3 knots). However, in the southern area traverses were made across the Strait. We have followed the basic concepts and methods of seismostratigraphic interpretations, summarized by Brown and Fisher (1980). These have been applied to high-resolution seismic profiles from shelf areas in the western Mediterranean Sea by a number of investigators (e.g., Stefa-

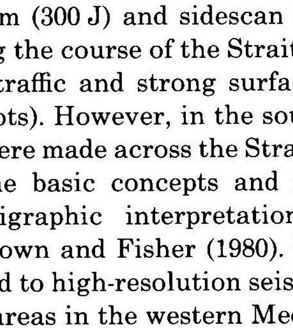

Fig.3. The trends of the major faults in the region and their relationship with the course of the Bosporus (based on a satellite photointerpretation by Eroskay and Kale, 1986).

non, 1985; Got et al., 1987; Canals et al., 1988). Three main seismostratigraphic units (depositional sequences) within a maximum TWT sweep of 100 ms were distinguished on the basis of their lithoseismic characteristics, geometry and mutual contact relationships. The nature of the acoustic basement varies depending on depth, local geology and distance from the shore. Each unit was mapped and its vertical and lateral facies variations were studied. The depth of water at each fix point was added to the calculated thickness of each unit and isopach maps were constructed assuming an average sound velocity of 1700 m/s for the unconsolidated to semi-consolidated sediments penetratable by the high-frequency acoustic signal (Ross et al., 1978).

Morphology and structure of the Bosporus

Figure 3 shows the intimate relationship between the trends of the regional tectonic lineations and the course of the Bosporus. This relationship suggests that the meanders of the Strait are primarily controlled by the trends of the major faults and their mutual intersections. Indeed, a number of investigators have already pointed out that the drainage pattern in this region is controlled by the structural lineations (Ketin, 1967; Scholten, 1974; Ilhan, 1976; Meric et al., 1988). On the basis of deep-penetration reflection profiles Ulug et al. (1987) inferred that the structure of the Palaeozoic basement in the southern sector of the Bospo-

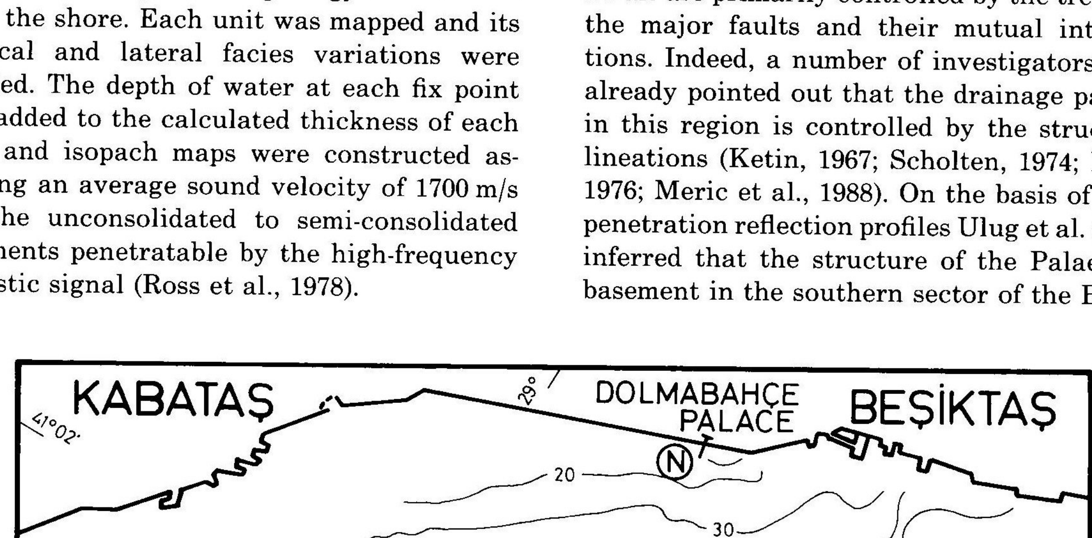

Fig.4. Bathymetry of the surveyed area between Uskudar (Anatolian side) and Besiktas (European side). "N" and "S" indicate each end of the cross section shown in Fig.5. Heavy lines are the cruise tracks of the illustrated Uniboom records in Figs.8–12. Contours in metres.

rus resembles that of a rift valley (or a graben) with a substantial sedimentary fill along its axis. The presence of such a thick (about 100 m) sequence on the Palaeozoic bedrock strengthens the view that the origin and evolution of the Bosporus involves more than the drowning of a Pleistocene fluvial valley. In addition if it is accepted that the Turkish Straits region had reached its geomorphological maturity (peneplanation stage) in the late Miocene to early Pliocene (Sayar, 1976; Erol, 1987), when down-cutting by rivers was intensified due to the start of a new phase of tectonic uplift and/or eustatic regression, the sedimentary sequence beneath the floor of the Bosporus and Dardanelles may be older than Quaternary. This question remains to be answered and is beyond the scope of this paper. However the sedimentary evolution of the straits cannot be divorced from the structural origin and the active tectonics of the region (Sengor et al., 1985; Crampin and Evans, 1986). In the remainder of this section we will attempt to demonstrate some relationships between the morphology of the southern sector of the Bosporus and its structural background.

In the area between Uskudar and Besiktas (Figs.1 and 4) bathymetric data (IMS, METU, 1985a) have confirmed the presence of two channels (or grooves) close to the banks of the strait separated by an elongated sill along the thalweg. The crestal part of the sill is defined by the 30 m isobath. This is the shallowest and the largest sill in the Bosporus (there is another smaller and deeper sill close to the northern exit). The channels gently deepen towards the north and join together as the sill gradually tapers out in the same direction (Fig.1). The coastal slopes of the channels are much steeper (4°–13°) than those on the flanks of the sill (1.3°–2°) (Fig.5). The Uskudar headland has a rocky coastline and sonographs have revealed rock exposures at a depth of about 25 m (Fig.6) close to the shore. These

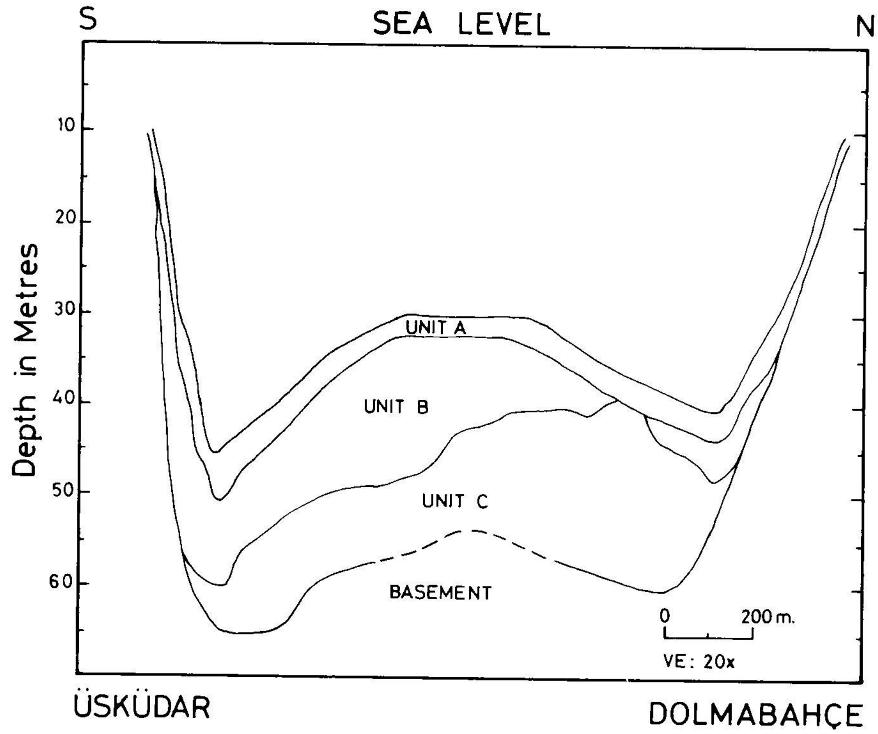

Fig.5. A geological section across the Bosporus showing thickness variations of the three seismic units on the acoustic basement. See Fig.4 for location. See text for details (based on isopach maps for each unit by Okyar, 1987).

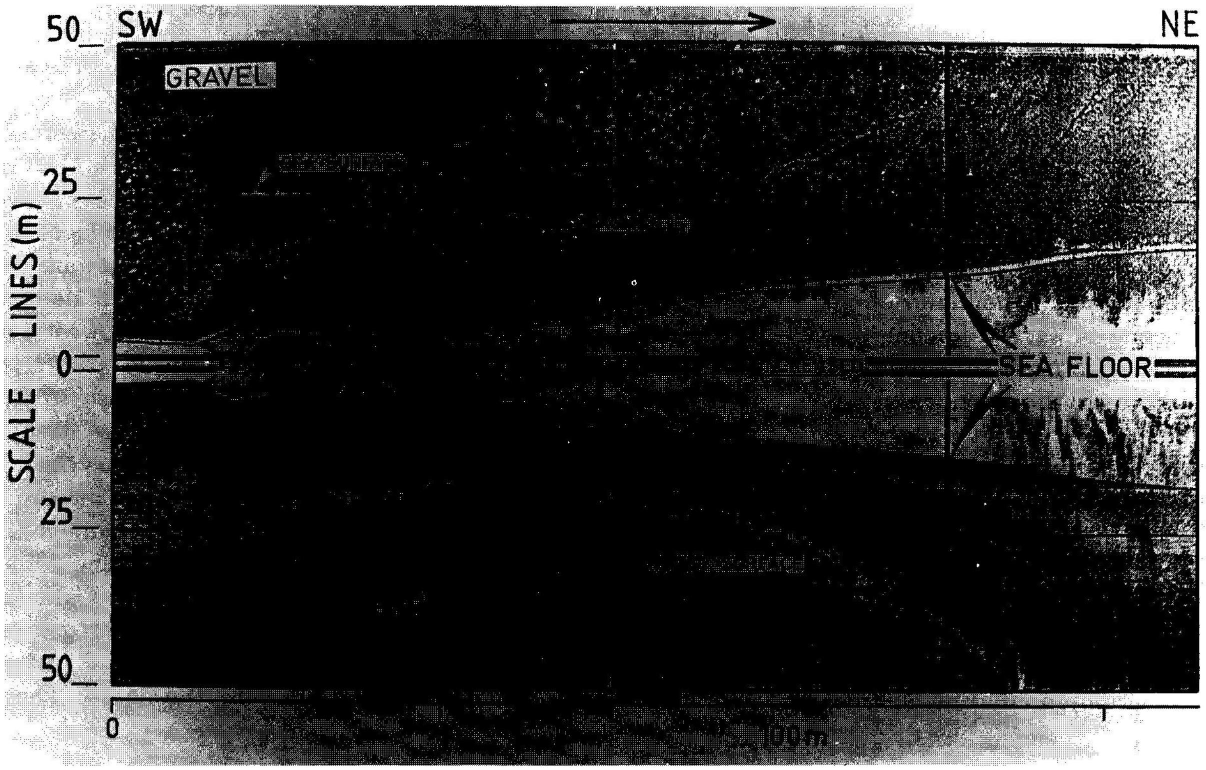

Fig.6. Sonograph from an area close to the cost of Uskudar showing a rocky bottom at a depth of about 25 m. See Fig.4 for location.

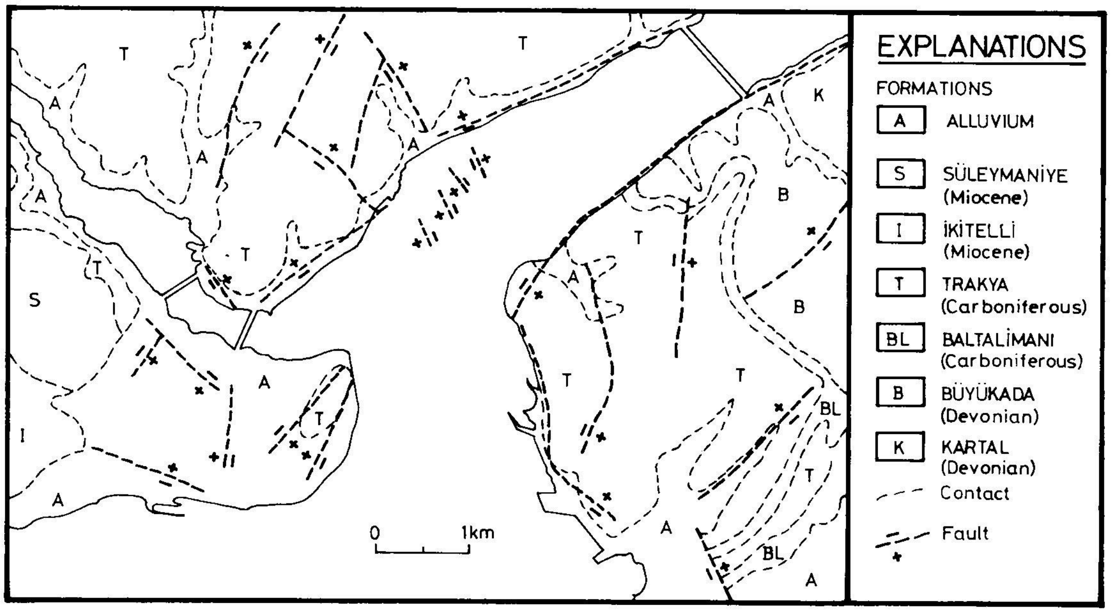

Fig.7. Details of land geology around the southern part of the Bosporus (from a 1:50,000 map (IRTC, 1986)). Trends and senses of relative motions along some offshore faults in the Palaeozoic basement are indicated.

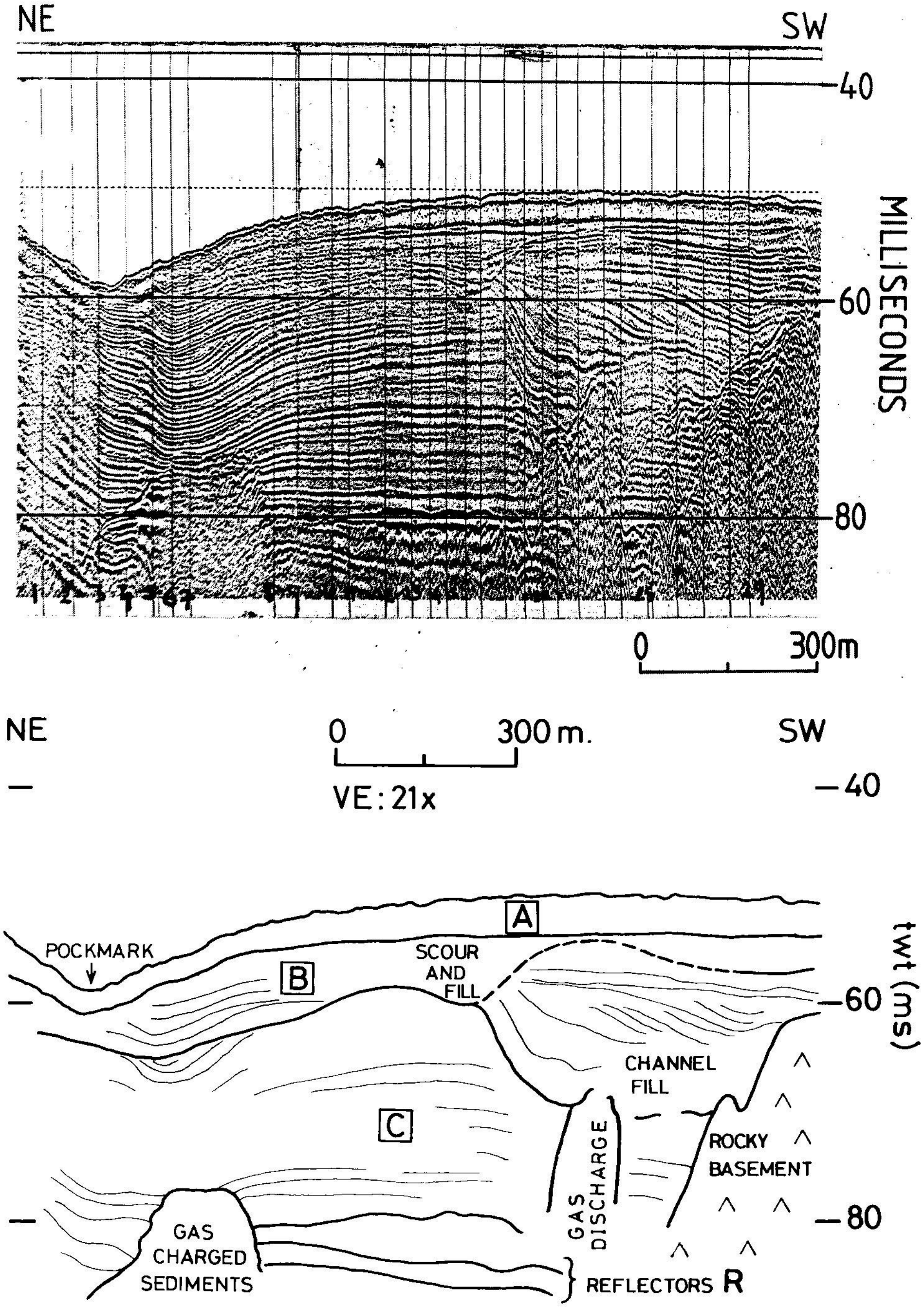

Fig.8. Uniboom profile and interpretation from the European side of the channel (see Fig.4 for location). Units A–C are shown.

findings and the steep (approximately 9°) gradient of the Palaeozoic bedrock along this bank are attributed to faulting (Okyar, 1987). A detailed geological map (Fig.7) of the area shows that both coasts of the Strait are fault controlled in this area. Eroskay and Kale (1986) have also reported basement faulting along this part of the Anatolian bank.

The southern end of the sill may have also been affected by recent differential vertical

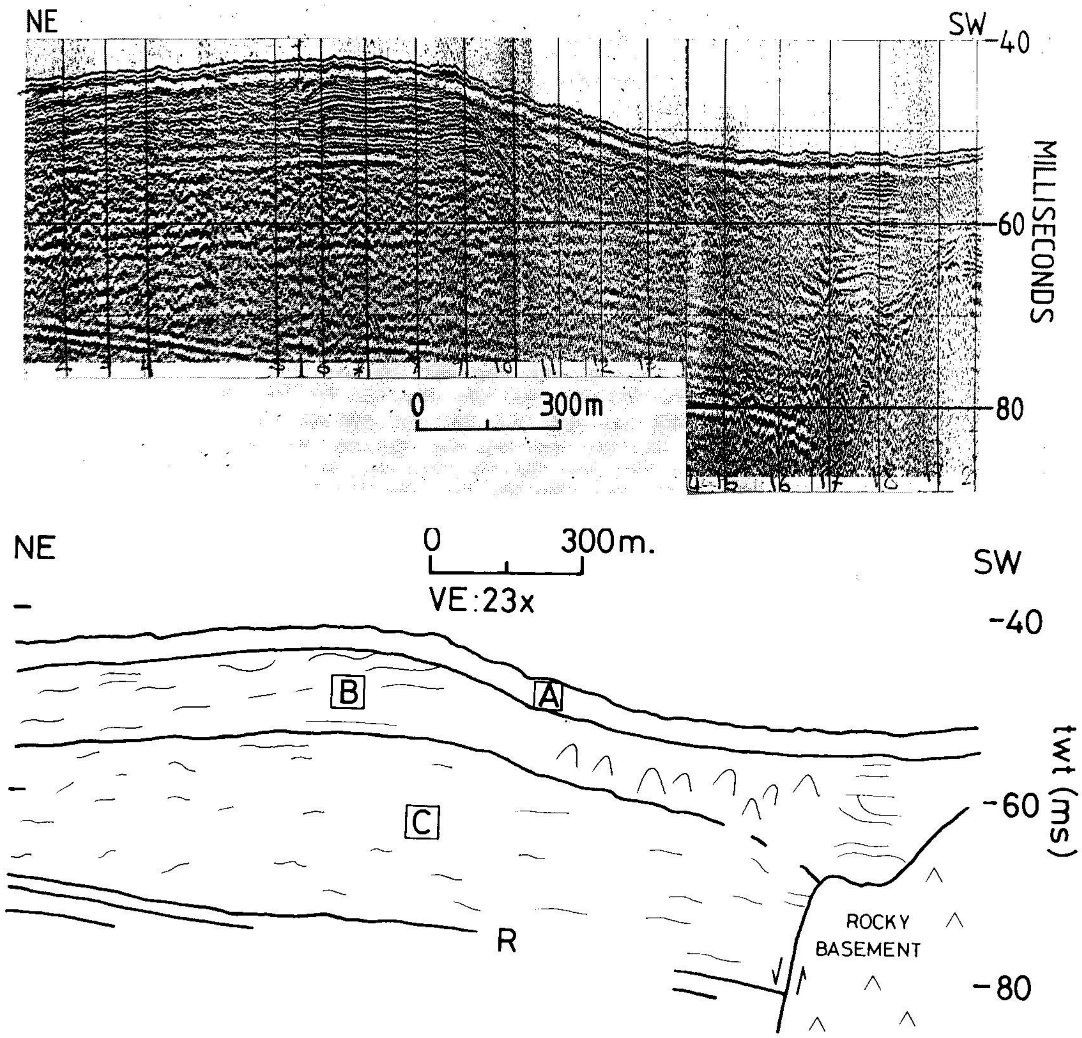

Fig.9. Uniboom profile and interpretation from the European side flank of the sill. Note the abrupt terminations of reflectors (*R*) against the basement high to the southwest. Units A–C are shown. (See Fig.4 for location.)

movements. The rugged surface of the Palaeozoic rocks shoals along the NW–SE trend of a fault running across the Strait under the southern slope of the sill (Figs.3, 8 and 9). The southern termination of this sill is abrupt in comparison with its gradual northward deepening and tapering (Figs.1 and 4) and it occurs at a sharp bend in the course of the Strait (from a N–S trend to a NE–SW one) at a line joining the Uskudar headland and Kabatas. The bend is accompanied by a significant constriction in the width of the strait towards the Black Sea. At this bend the deepest parts of the coastal channels become relatively flat and then begin to slope towards the south. These morphological changes occur at the meeting point of three sets of faults (or fault zones) (Fig.3).

The termination of the sill to the south can be partly due to intensified erosion of its crestal parts as a result of greater turbulence close to the bend (Ozsoy et al., 1986), where jet flow and internal hydraulic "jump" conditions prevail (Unluata et al., in press). These conditions essentially arise out of the sudden constriction of the width of the Strait and a change in its depth. The wavy character of the seabed (Figs.10 and 11) at some localities and the presence of small fields of sand waves and ripples on sonographs indicate active erosion of the shallowest parts of the sill (Okyar, 1987). The thickness of the surficial sediments (unit A) is also considerably reduced at these places.

The coastal slopes and bottoms of the channels show many short topographic features with a wavelength of up to several tens of metres. Some of these are about 10 m shallower or deeper than their immediate surroundings and seem to be controlled by the rugged surface of the bedrock, which is thinly buried (< 20 m) over the shallower parts of the landward slopes of the channels (Fig.5). These features become much less frequent on the slopes and crestal parts of the sill where the sedimentary sequence thickens under this feature. Since faulting appears to have most severely affected the banks of the Bosporus parallel to its course (Figs.3 and 7), the channels are probably more intensely affected by neotectonic movements along these mobile zones. The shallower parts of these channels have also been affected by turbulence preventing the development of thicker sedimentary sequences. The increasing thickness of the sediments away from the coastal zones may have been controlled by the pattern of deformation to a graben-like valley undergoing faster subsidence in its central parts (Ulug et al., 1987); however, it is more likely to be the case that hydrodynamic, physiographic and climatic conditions were more favourable for sediment accumulation in the early to middle Holocene, when the deeper parts of the valley were already submerged and bottom currents were still sluggish (see below). Syn-depositional deformation by neotectonic movements may have produced some of these small-scale undulations through differential compaction, liquefaction and related phenomena. Some of the deeps are certainly pockmarks. These are small depressions produced as a result of gas or porewater escaping from soft sandy sediments. They have been reported from several shelf areas around the word (e.g., Harrington, 1985). An example is shown in Fig.8. The escape of gas is indicated by the local downwarp of the reflectors and their overlying depression on the seafloor. The wipe-out effect or blurring of the reflectors beneath the downwarped reflector is most probably due to a rising gas column and can easily be confused with a basement ridge. Very similar features have been reported from the northern parts of the Adriatic Sea close to the Po Delta (Stefanon, 1985).

Thus the larger morphological features and the tortuous course of the Strait of Bosporus appear to be essentially controlled by the trends of the structural lineations in the region. The ancestral valley may have actually been initially formed as a segmented and linked series of grabens between parallel sets of intersecting faults. This implies that its major deeps (or potholes (DAMOC, 1971; Gunnerson and Ozturgut, 1974)) and embayments may not all be relict fluvial features from the past periods of aerial erosion before the post-glacial rise in sealevel. The (palaeo)relief of the basement and its interaction with the hydrodynamics and processes of sedimentation in the strait have certainly played some role in determining the morphological evolution of the valley in the late Quaternary when it was transformed into a marine strait.

Recent sediments

During various engineering projects at least three series of boreholes have been drilled on the floor of the Bosporus (Fig.1). The first series (1973) includes nineteen shallow (less than 10 m) holes at an average distance of 120 m from each other between Uskudar and Sarayburnu. The logging data show that the sediments consist of shelly sand richer in coarser fractions in the uppermost few metres at most locations, particularly in shallow waters. Only one of these holes (at a depth of 38 m and a distance of 200 m off Sarayburnu) penetrated through a clayey and sandy silt facies beneath a 4 m thick layer of coarse shelly sand. The same facies was reported from another borehole at a sub-bottom depth of nearly 3 m in the same area (Eroskay and Kale

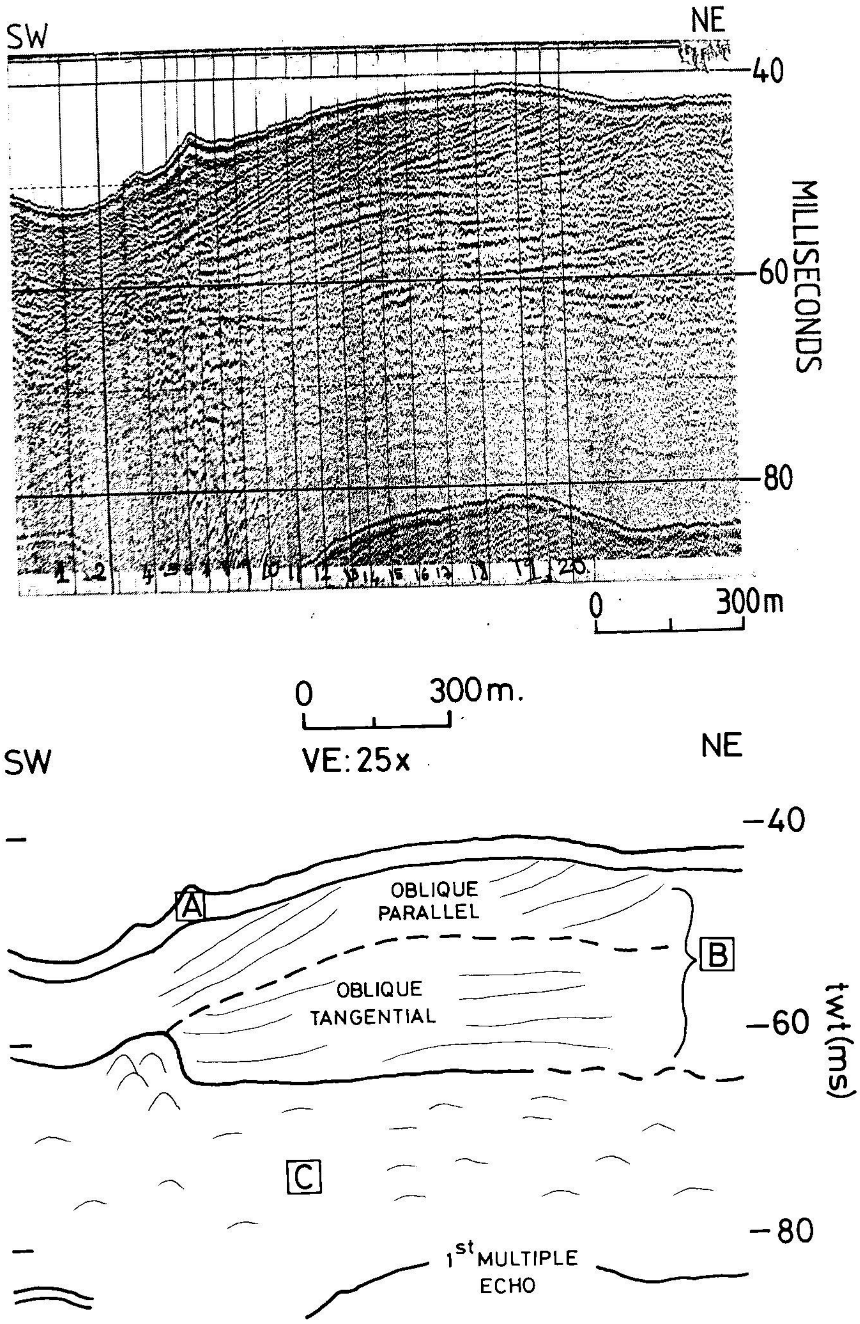

Fig.10. Uniboom profile and interpretation from the crest and southern slope of the sill. Notice the reduced thickness of unit A on the shallower parts of the sill. (See Fig.4 for location.)

(1986); hole TB8, Fig.1). The surficial shelly sand layer thickens to about 15 m at site TB19 approximately 900 m to the east of the first hole. While hole TB8 reached the Carboniferous clastics of the Trakya Formation at a sub-bottom depth of 37.5 m, TB19 did not, despite penetration down to a sub-bottom depth of 73 m. At this site (depth of 12 m) the silty unit is covered by a 23 m thick layer of medium sand with the surficial shelly sand at the top.

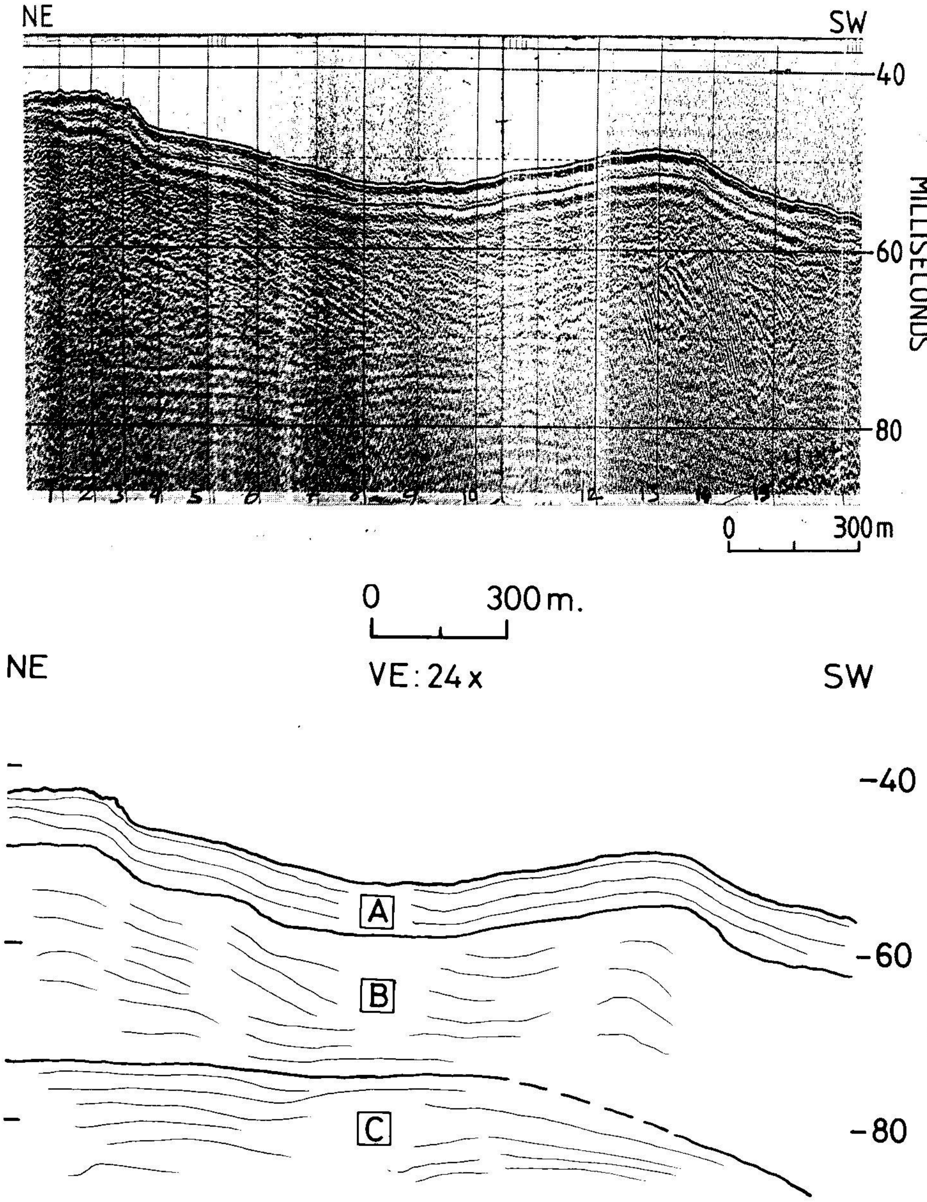

Fig.11. Uniboom profile and interpretation from the Anatolian side flank of the sill. Notice the wavy nature of the seabed at the northeastern end. (See Fig.4 for location.)

Farther to the east, less than 500 m from the Anatolian bank at a depth of nearly 9 m (site TB25A), the shelly unit and the upper part of the underlying medium sand layer are missing. Here the borehole reached the Palaeozoic rocks at a sub-bottom depth of almost 19 m. This was inferred to be due to basement faulting between the two sites with the latter site being thrown upwards (Eroskay and Kale, 1986).

Micropalaeontological studies on a large series of samples from a third group of boreholes nearly 1 km to the north of the three deep boreholes mentioned above (Meric et al., 1988) show that the subsurface sediments down to 40 m are entirely of Holocene age. The benthic foraminiferal assemblages that are principally composed of various morphotypes of *Ammonia beccarii* Linné become increasingly poorer in specific diversity with depth. The sediments are less than 20 m thick and lie on the Trakya Formation about 600 m away from the coast of Uskudar, but in the deepest part of the Strait along the traverse (50 m deep) the basement was not reached at a maximum sub-bottom depth of nearly 40 m (site TB114, Fig.1) (Meric et al., 1988). At this site the upper 8 m of the hole penetrated through a coarse sandy unit rich in calcareous shell fragments accompanied by assemblages of benthic foraminifera including a variety of typical shelf-dwelling species (textularids, miliolids, cibicidids, and discorbids) (E. Meric, pers. commun., 1988). However, the sediments below this unit are poor in macrobenthos shell fragments and consist of sand and silt often with a yellowish colour. These are correlated with the above-mentioned subsurface facies rich in silt and fine sand beneath the surficial shelly sand close to the coast of Sarayburnu. Its foraminiferal fauna has a low diversity and is dominated by *Ammonia beccarii*. The assemblages can be compared with those reported from the shallower parts of the Black Sea shelf of Romania (Gheorghian, 1974). They have been deposited under brackish-water conditions close to the seabed (salinity less than 30‰), and comparable assemblages dominated by this species characterize the prodeltaic muddy areas near major rivers in the Mediterranean Sea (Alavi, 1980). The lowest part of borehole TB114 penetrated through some gravelly and shelly material at a sub-bottom depth of about 40 m. However, because of poor penetration and lack of data from the other sites, the age of these deposits could not be determined (Meric et al., 1988). They are probably late Pleistocene and formed at the time of the maximum drop in sealevel during the last glacial stage (see below).

On the basis of these data, the uppermost 30 to 35 m of sediments from the deeper parts of the southern Bosporus can be divided into two principal facies; a gravelly and shelly sand layer 3 to 15 m thick overlying a silty sand facies poor in shell fragments. The upper layer tends to become richer in coarser shell fragments in its uppermost part and is better developed towards the Anatolian coast in areas deeper than 30 m. Over the coastal slopes it can be mixed with coarse-grained terrigenous material and anthropogenic waste at some localities. The underlying finer grained sediments often contain an impoverished benthic foraminiferal fauna and seem to be better developed close to the entrance to the Golden Horn. The upper facies includes diverse assemblages of benthic foraminifera and is comparable with the carbonate-rich shelf sediments from the less than 50 m deep eastern parts of the Sea of Marmara (Ergin and Alavi, in prep.). It records deposition under normal marine conditions and may not be older than the "Climatic Optimum", when a tongue of the saline Mediterranean waters had penetrated well into the Strait of Bosporus as the undercurrent began to be fully established and the rate of runoff discharge had decreased together with the surface outflow of low-salinity water from the Black Sea which decreased from its higher levels in the early Holocene (Ross and Degens, 1974; Erinc, 1978; Degens and Stoffers, 1980; Stanley and Blanpied, 1980).

The system of currents

The circulation system of the Bosporus has been the subject of study for a long time and the results are reviewed by Unluata and Oguz (1983), Tolmazin (1985) and Ozsoy et al. (1986). In recent years the Turkish Straits system has received more attention by the Turkish Government authorities. A long-term monitoring program of the hydrochemical properties of the water masses between the Black Sea and the Aegean Sea is underway, the results of which will substantially increase our knowledge of the system. In this section some of the most salient aspects of the system of water exchange through the Bosporus will be discussed on the basis of the early results of the monitoring program (Ozsoy et al., 1986, 1988; Latif et al., 1988).

The system of currents is primarily governed by the outflow of brackish (17–19‰) surface waters from the Black Sea above saline (38–38.5‰) bottom waters that flow from the Sea of Marmara in the opposite direction

(Gunnerson and Ozturgut, 1974). It is believed that the system reached its present state between 3 to 1 ka B.P. when the volume of the surface outflow decreased from its early to middle Holocene high values in the present value (Ross and Degens, 1974; Degens and Stoffers, 1980; Stanley and Blanpied, 1980). The rate of outflow from the Black Sea is two to three times higher than that of the subsurface flow of Mediterranean waters in the opposite direction (Ozsoy et al., 1986). The system is essentially density driven with the hydraulic head being about 35 cm higher at the northern end of the Strait of Bosporus (Sokorin, 1983). The level of no motion between the two currents and their salinity interface (halocline) slopes from a depth of about 15 m in the south to as much as 45 m at the northern exit of the strait. Thus, the thickness of the bottom layer is reduced by net upward entrainment as the undercurrent flows towards the Black Sea. However, there is no doubt that the undercurrent continuously flows into the Black Sea (Tolmazin, 1985; Latif et al., 1988). When the surface outflow is at its seasonal peak in early summer the flow of the undercurrent may, however, be temporarily halted under the influence of strong northerly winds which pile up water at the northern entrance of the strait (Latif et al., 1988). The situation becomes comparable with that of a salt-wedge estuary (De Fillipi et al., 1986). These events only last up to a few days and are followed by a phase of strong northward flow of saline waters into the Black Sea.

Recent studies show that the intensity of turbulence and surface current velocities increases in the southern parts of the Strait of Bosporus essentially due to sharp changes in the width of the Strait and the depth of water (Ozsoy et al., 1986). Current measurements by a moored system of current meters at two stations in the northern survey area (Fig.1) (IMS, METU, 1985b) for a period of 1 month have shown that the average velocity at 5 m above the seafloor is nearly 50 cm/s in the European side channel and 80 cm/s in the Anatolian side channel. The main direction of flow was found to be towards the north as expected and surface currents were found to show greater lateral variability, being on average up to three times faster on the Anatolian side of the sill (approximately 250 cm/s). The presence of a previously reported (DAMOC, 1971) small clockwise surface gyre off Besiktas is confirmed and it is suggested that this feature may be the reason for slower surface currents on the European side of the sill. Similar gyres develop in other embayments along the course of the Bosporus farther to the north, where the axis of the undercurrent may also shift from one side to another temporarily and spatially (Caspers, 1957; Gunnerson and Ozturgut, 1974). A detailed seasonal hydrographic and bathymetric study at the northern entrance of the Strait of Bosporus has shown that the dense and saline subsurface waters flow over the deepest parts of the Strait through a canal which extends towards the north on the shelf of the Black Sea (Latif et al., 1988). This study clearly demonstrates the steering effect of the bottom topography on the course of the bottom currents.

Seismostratigraphy

In the lower part of the sequence penetrated on a maximum TWT sweep of 100 ms a band of two or three strong reflectors (R) appears at about 75 ms in the European side channel and at close to 80 ms beneath the sill, deepening to about the 90 ms on the floor of the Anatolian side channel (Figs.8, 9, 11 and 12). The sequence above these reflectors is estimated to be between 30 and 35 m thick on the floor of the channels, becoming thicker beneath the sill (Fig.5) and towards the north, where the R-reflectors are not always encountered (e.g., Fig.10). The reflectors often abut against the rocky (Palaeozoic) basement on the coastal slopes of the Bosporus and in some places on the southern slope of the sill (Fig.9). The surface of the Palaeozoic basement is usually associated with prominent hyperbolic echoes characteristic of uneven rocky surfaces covered by soft and semiconsolidated sediment

(McQuillin and Ardus, 1977). This surface appears to be faulted in some places, in alignment with the trends of some faults on the banks of the Strait (Okyar, 1987) (Fig.7). The relationship between the relief of this surface and its overlying sediments clearly indicates that the former has exerted some influence on the thickness variations of the latter and the topography of the seafloor in the coastal channels.

The R-reflectors are underlain by a distinct hummocky seismic facies beneath the deepest

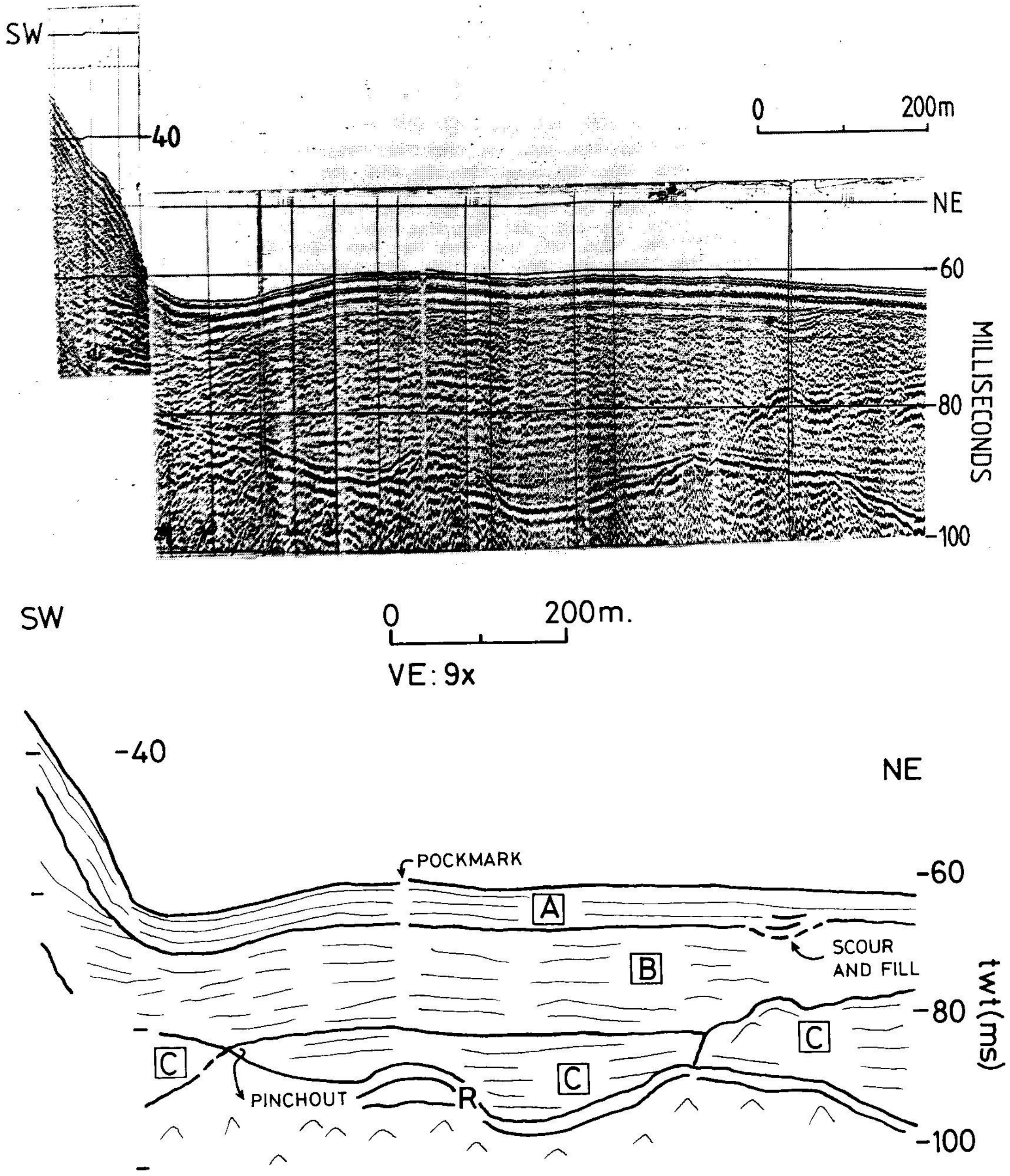

Fig.12. Uniboom profile and interpretation from the floor of the Anatolian side channel. Note the undulatory nature of the erosional surface at the bottom of unit C and the hummocky to chaotic character of the reflectors beneath it. Compare with Fig.8. (See Fig.4 for location.)

parts of the Anatolian side channel (Fig.12). This facies is interpreted as consisting of poorly stratified sediments containing a high proportion of coarse-grained material deposited under unstable and relatively high-energy conditions probably as channel fill (or chaotic fill) (Brown and Fisher, 1980; Got et al., 1987). The upper surface of this facies is undulating but smooth and without the prominent hyperbolic echoes associated with the uneven surface of the rocky basement. It represents an erosional level registering a major change in the depositional regime of the Strait. The sediments beneath it may have been deposited under very shallow water conditions, when sealevel was still too low for the initiation of marine currents in the Bosporus. The same surface was found farther south between Haydarpasa and Saraybrunu at a subsurface depth of nearly 60 m underlain by a similar hummocky seismic facies in deep (> 50 m) waters (IMS, METU, 1986). The lateral discontinuity of the reflectors and their wavey characteristics as well as the presence of frequent overlapping hyperbolic echoes suggest deposition in nearshore or beach environments. They probably represent the shelly and gravelly deposits encountered close to the bottom (subsurface depth of 40 m) of borehole TB114 (Fig.1) (Meric et al., 1988). If so, they could be the proposed late Pleistocene (Karangat phase) shallow marine sediments deposited in the deepest parts of the Bosporus valley under marine strait conditions before the maximum drop in sealevel and the subaerial exposure of the valley at the peak of the last glacial stage (Neoeuxinian phase) (Scholten, 1974, p.123).

Although we have no evidence or data to confirm this correlation, considering its subbottom depth and the results of micropalaeontological studies mentioned above the erosional surface cannot be older than the early Holocene to late Pleistocene. It occurs at a depth of about 65 m below the mean sealevel in the deepest parts of the Uskudar channel, shoaling to about 60 m beneath the floor of the European side channel (Fig.5). This is probably due to post-depositional tectonic movements, variable rates of sedimentation and palaeotopographic controls. If a total subsidence of 10 m is assumed to have occurred during the last 10 ka and the depth of water was not more than about 5 m at the time of erosion, a depth of 50 to 45 m below the present sealevel would give an age of nearly 10 ka B.P. on the global curve of the post-glacial eustatic rise in sealevel (Van Andel and Lianos, 1984). This is close to the time of the earliest incursions of the Mediterranean waters into the Black Sea (Ross, 1977; Deuser, 1974). Therefore the erosional surface was probably formed by marine currents as the eustatic sealevel was still rising rapidly in the early Holocene and the Bosporus was at its latest marine spillway phase. Shortly after this time the Mediterranean Sea was flooding the lower parts of coastal valleys in the Aegean region, among them the Strait of Bosporus and the Golden Horn (Erinc, 1978; Kraft et al., 1980; Aksu et al., 1987; Meric et al., 1988). In addition, the Black Sea outflow is believed to have been at its peak almost at the same time due to the increased rate of precipitation in the catchment basins, connecting with the Caspian Sea, and meltwater discharge from the dwindling Fenno-Scandinavian ice sheet (Ryan, 1972; Stanley and Blanpied, 1980; Berger et al., 1985). Hence, the succession above the erosional surface is considered to have recorded the post-glacial sedimentary evolution of the marine strait period. It is divided into three major seismic units. These are briefly described, from top to bottom.

Unit A

This unit usually appears as a band of two or three strong reflectors just beneath the seabed (Figs.8–12). It occurs all over the area except at a locality off the Uskudar headland where rock exposures occur on the seafloor (Fig.6). The reflectors in it tend to be weaker and discontinuous over much of the European side channel. In the deepest parts of this channel the unit is acoustically more transparent and thinner than in the other channel (Fig.8). This is an indication of a more homogeneous grain-size composition of the surficial sediments or of their greater content of sand and finer grained material.

The sidescan sonar survey showed that the surface sediments are mainly sandy with gravel patches in some places. The results of grain-size analyses on the sediments (DAMOC, 1971) support this interpretation. They show that most of the sediments consist of gravelly sand with the proportion of shell fragments increasing towards the Anatolian side of the sill. This could be an indication of greater near-bottom current activities in the latter area (IMS, METU, 1985b). It may also be a result of lesser dilution of bottom waters by upward mixing on this side of the sill. Surface currents are also reported to be weaker off Besiktas. The presence of a permanent gyre on this side may lead to the entrapment of finer grained suspensate materials and the enrichment of deeper water sediments in these materials.

The thickness of unit A often increases in restricted topographic lows, while it is reduced to its lowest values over the shallowest parts of the sill. It is much thinner than its underlying other two units (Fig.5), having a maximum thickness of about 10 m in a small depression near the Uskudar headland at a depth of 45 m (Okyar, 1987). The unit is better developed on the floor of this channel and includes two strong continuous subsurface reflectors (Fig.12), suggesting better stratification and greater contents of sand. In areas shallower than about 15 m it lies directly on the Palaeozoic basement and in deeper waters its contact with the second unit is often erosional, showing scour and fill features in some places (Figs.10–12).

Unit B

This unit shows a distinct pattern of variation in its lithoseismic characteristics across the Strait. Its reflectors are generally weaker showing greater continuity and regular spacing in the European side channel (Figs.8 and 9). On the southern slope of the sill they are oblique and still fairly continuous (Figs.10 and 11), but towards the Anatolian side channel they become discontinuous and locally chaotic to hummocky with frequent hyperbolic echoes (Fig.12). This is probably due to reworking under intense current activity and inclusion of a larger amount of sand-size material (Got et al., 1987). The obliquity of the reflectors also increases upwards in some places on the southern slope of the sill (Fig.10) suggesting progradation at the expense of episodic erosion of the crestal parts of this feature. The thickness of the unit tends to decrease on the flanks of the sill presumably due to slope instability. Channel filling and scour and fill structures were noticed in the unit on the floor of the European side channel (Fig.8). These features and the erosinal contact of the unit with the underlying unit indicate high energy conditions of deposition. The thickness of the unit in this channel is less than in the Anatolian side channel. In the former it thins out towards the northwest (off Besiktas) (Okyar, 1987). It is completely missing from areas shallower than about 30 m on the coastal slope of this channel, pinching out at a combined water and sediment depth of about 35 m a few hundred metres off the European bank. However, on the Anatolian coastal slope the unit is found to occur in waters as shallow as about 15 m. Thus, similar to its seismic facies characteristics, the asymmetrical thickness distribution of this unit across the Strait also suggests that depositional conditions differed from each other on either side of the sill at the time of deposition of the unit. As was noted above, this is also the case now.

Unit C

A comparison of the seismic facies characteristics of unit B with those of unit C suggests that the latter was deposited under lower energy conditions. Unit C shows the criteria of an onlapping marine fill. Note the even, parallel, continuous nature of the reflectors in the European side channel (Figs.8) where the unit is better developed and stratified due to greater sediment supply and/or more protected conditions. Over the sill the continuity of the reflectors is reduced probably because of the greater variability of the hydrodynamic conditions (Figs.9–11). This unit also becomes thicker on the Anatolian flank of the sill (Fig.5) and its lower contact with the underlying material cannot always be clearly discerned. However, in the deepest parts of the European side channel it is separated from underlying materials by R-reflectors. In this channel unit C displays the features of a complex fill consisting of overlapping lenticular sediment bodies (Fig.12). Again, depositional regimes differed from each other on each side of the sill.

In contrast with the thickness distribution pattern of unit B this unit appears to be thicker over the northern parts of the sill (Fig.5). This is an indication of temporal shifts in loci of rapid deposition and erosion by the laterally migrating axis of currents at each stage as well as palaeotopographic controls. Such variabilities are common in those parts of estuaries close to bends in their courses (Dyer, 1979). The unit is restricted to areas presently beneath a total depth of 35 m from the mean sealevel and its contact with the overlying units is erosional and discordant.

Interpretations and discussion

Unit C is considered to have been deposited in the early to middle Holocene when the eustatic sealevel rose from 30 to 35 m below the present level to slightly above the present level (Kraft et al., 1980; Van Andel and Lianos, 1984). As the rate of precipitation was higher and stream erosion was prevalent in the region (Erinc, 1978; Brakenridge, 1980; Kraft et al., 1980; Rossignol-Strick, 1987) the rate of terrigenous sediment supply must have also been greater. This is clearly evident from prograding deltaic sigmoid sets off the mouth of Kurbagali River (Fig.2) showing that seaward accretion continued as sealevel was still rising rapidly (IMS, METU, 1986; see Beard et al., 1982). The foreset beds of these deltaic sequences gradually thin out towards the deepest parts of the Bosporus and tangentially downlap over the same erosional surface represented by the R-reflectors in the area between Uskudar and Besiktas. At that time (10–7 ka B.P.) the undercurrent was not yet fully established as a coherent and continuously flowing current, because sealevel had not yet risen well above the threshold of 35 m in the Strait (Ross and Degens, 1974) and the Black Sea outflow was stronger than at present (Stanley and Blanpied, 1980). Under these conditions a salt-wedge estuary may have developed in the southern parts of the Bosporus and the Golden Horn.

This situation must have followed the initial phase of strong current activity on the floor of the Strait as the Mediterranean waters were infiltrating into the Black Sea (the marine spillway phase of Scholten, 1974). However, it is more likely that the erosion of this surface took place by means of early Holocene spillover currents from the Black Sea before the establishment of the salt-wedge estuarine conditions. Flushing of large amounts of freshwater from the Black Sea is believed to have contributed towards a sharpening of the water-column stratification and development of widespread deep-water anoxia in the eastern Mediterranean basins between 9 and 7 ka B.P. (e.g., Anastasakis and Stanley, 1986). Even if the Black Sea outflow was not as strong as has been proposed (e.g., Cramp and Collins, 1988) increased regional precipitation and runoff in the early Holocene may have contributed towards the development of a salt-wedge system in the southern part of the Bosporus valley as Mediterranean waters were penetrating into it from the south. The high level of runoff from the drainage system of the Golden Horn, which by this time was flowing towards the south (Scholten, 1974), may have played an important role in these developments. As sealevel was still too low for the continuous flow of the undercurrent, most of the sediments were trapped in the deeper parts of the Strait closer to the Sea of Marmara. This situation can explain the development of the sill and the increased thickness of unit C beneath it, as well as a thick (70–65 m) sedimentary fill in the deepest parts of the Golden Horn (DAMOC, 1971; Sayar, 1976). The latter is reported to lie over Palaeozoic rocks and its benthic microfaunal assemblages (ostracods and foraminifera) indicate deposition under brackish conditions in its lower parts (Meric et al. (1988) and pers. commun. (1988)). Thus, in this part of the Bosporus unit C may represent a part of the subsurface silty facies containing foraminiferal assemblages rich in *Ammonia beccarii*.

Together with the gradual reduction in the rate of the Black Sea outflow and increased aridity in the region since about 6 ka B.P. (Bonatti, 1966; Erinc, 1978; Farrand, 1979; Stanley and Blanpied, 1980) riverine sediment supply became restricted and sealevel reached close to its present level. An increasing part of riverloads flowing into the Golden Horn was becoming trapped upstream. Progradation of the deltaic platform off Kurbagali River came to a halt and its shallower parts began to be reworked and covered with sediments rich in biogenic carbonates. Meanwhile, by this time the flow of Mediterranean waters into the Black Sea as a coherent undercurrent had begun (Ross and Degens, 1974; Degens and Stoffers, 1980). Thus the floor of the Strait began to be subjected to more erosion and reworking, particularly in areas where large sediment accumulation occurred during the previous stage and in the restricted parts of the Strait close to its southern sill. Under vigorous hydrodynamic conditions active sediment accumulation must have been restricted only to those areas deeper than the interface between the two currents (e.g., in the Anatolian side channel where unit B is thicker) or to those areas close to those rivers still debouching their loads directly into the Strait. The deeper parts of the Golden Horn were by this time occupied by sea and most of the input into them could not reach the Bosporus. Unit B records deposition under these conditions. It represents the transitional phase of the transformation of the Bosporus from an early Holocene salt-wedge estuary into its present two-way marine strait phase. This unit may be correlated with the shelly sand facies in the upper parts of the boreholes containing diverse assemblages of benthic foraminifera which indicate deposition under conditions of salinity which were higher than in the previous stage. It may also be tentatively correlated with the middle Holocene sapropel in the Black Sea (unit 2 of Ross and Degens, 1974).

By the time of the beginning of the deposition of the coccolith-rich surficial sediments in the Black Sea between 3 and 1 ky ago (Ross and Degens, 1974; Degens and Stoffers, 1980; Calvert et al., 1987) the rate of the outflow from the Black Sea had decreased to close to its present value and sealevel was probably standing a few metres higher than its present position in the Marmara Sea and Black Sea (Scholten, 1974; Kraft et al., 1980; Hay, 1988). The deeper parts of the Golden Horn were transformed into an inlet as a result of which most of the input in to it was trapped upstream to the north of Balat (Fig.1). Rapid alluviation and expansion of the deltaic plains in some river valleys during this time period has been well documented along the Aegean coast of Turkey (Eisma, 1978; Aksu et al., 1987). A phase of more intense erosion and reworking of the older sediments on the floor of the Strait had thus begun. This led to the development of a surficial veneer of gravelly and shelly sand several metres thick in some places. These deposits record sedimentation under fully marine conditions over the deepest parts of the Strait occupied by dense saline Mediterranean waters, providing a safe passage for the migration of some Mediterranean faunal elements into the Black Sea (Furnestin, 1979; Tolmazin, 1985). Borehole data show that the sediments become richer in calcareous shell fragments in their upper few metres, which may be correlated with seismic unit A as an erosional lag. Recovery of similar facies containing abundant remains of benthos in the Quaternary sediments of the Turkish Straits can be regarded as evidence of earlier phases of marine strait conditions.

Conclusions

The geological origin and evolution of the Bosporus goes well beyond the post-glacial drowning of a late Pleistocene emergent river valley and may extend back to pre-Quaternary times, since the sediments beneath its floor are up to 150 m thick in some places. Its course shows a close relationship with the trends of regional tectonic lineations and seems to follow a segmented series of grabens between two main sets of faults in the Palaeozoic and Mesozoic basement. The late Quaternary morphological and sedimentary evolution of the southern part of the Bosporus has been affected by neotectonic activity and the (palaeo)relief of the basement and the interaction of this activity with the sedimentary conditions. The rate of sedimentation has varied from place to place depending on hydrodynamic conditions, rate of supply and shifting loci of accumulation.

Towards the end of the last glacial period, when sealevel was standing at about 100 m below its present level in the Black Sea and Marmara Sea (Erinc, 1978; Kraft et al., 1980; Kaplin and Shcherbakov, 1986), the emergent valley gradually became inundated either from the north or the south and the Pleistocene deposits in its deepest parts were eroded during a marine spillway phase at about 10 ka B.P. As sealevel rose in the early Holocene to about 35 m below its present position, the southern part of the valley was transformed into a salt-wedge estuary resulting in the entrapment of the bulk of the riverine silt and sand material washed into it via the Golden Horn and other rivers. With the inception of the present system of water currents in the middle Holocene, reworking and erosion of previous sediments began and a new phase of sedimentation started under higher energy conditions. By the time of the full development of the undercurrent sealevel probably stood a few metres above its present level in the Strait of Bosporus and the Black Sea (Kraft et al., 1980; Hay, 1988). The present conditions have been prevailing since then. These have led to the blanketing of older sediments with a veneer of coarse sand rich in shell fragments. Strong bottom-current activities have removed most of the finer materials into the adjacent deep basins at each end of the waterway (Evans et al., 1989) or into the protected deeper parts of the Strait itself. These events are registered by an upward-coarsening sequence of Recent sediments on top of a late Pleistocene–early Holocene erosional surface at about 60 m below the mean sealevel. They have left an overprint of fine morphological features and subsurface sedimentary facies changes concealing the basement structure and older sediments of the deeper parts of the valley.

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