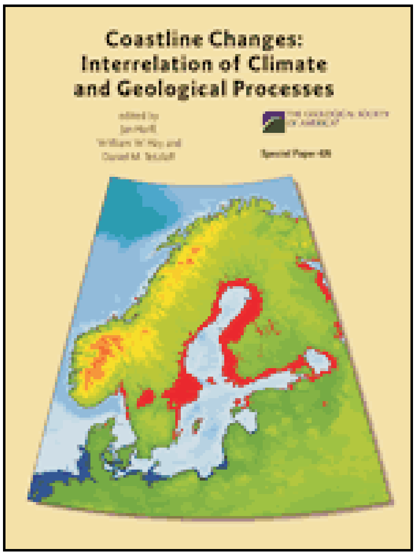
Special Papers

Coastline Changes: Interrelation of Climate and Geological Processes



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This volume contributes to the current discussion of the role of natural and anthropogenic driving forces for coastal processes and their socioeconomic consequences. Special attention is paid to computerized tools that allow us—based on reconstruction of paleodevelopments—to predict the interference of processes on different time scales. On the one hand, the book provides an overview of the current model developments in describing vertical crustal movement, climate change forcing sea-level variations, the genesis of the basin fill along continental margins, and the interference of these processes in coastal development. On the other hand, it describes coastal development in key areas for different climate zones and geological settings. *Coastline Changes* is addressed to students and professionals in the geosciences, archaeology, social sciences, economy, and computer sciences. It will foster interdisciplinary discussion for the purpose of developing integrated concepts for sustainable development of the coastal zones.

Coastline Changes: Interrelation of Climate and Geological Processes

edited by Jan Harff, William W. Hay, and Daniel M. Tetzlaff

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A Black Sea lowstand at 8500 yr B.P. indicated by a relict coastal dune system at a depth of 90 m below sea level

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ABSTRACT

Oceanographic surveys in the Black Sea during 1998, 2002, and 2004 in the framework of a French-Romanian joint project, and recently in the framework of the European project ASSEMBLAGE, complement previous seabed mapping and subsurface sampling studies undertaken in the Black Sea by various international expeditions. Until the Ryan and Pitman flood theory and prior to this project, it was proposed that the Black Sea was predominantly a fresh-water lake interrupted by possible marine invasions coincident with high sea level during the Quaternary.

From the recent surveys carried out on the western part of the Black Sea it is evident that the Black Sea's lake level rose on the shelf to at least the isobath –40 to –30 m as ascertained by the landward limit of extent of the *Dreissena* layer characteristic of brackish to fresh-water conditions. This rise in the lake level could coincide with the answer of the Black Sea catchment's basin to the meltwater drained from the thawing of the ice cap ensuing Melt Water Pulse 1A (Bard et al., 1996). It is possible that at that time the lake level filled by fresh water reached the level of its outlet and spilled into the Mediterranean Sea. Later, in the mid-Holocene at 7.5 k.y. B.P., the onset of salt-water conditions is clearly evident in the Black Sea. From these observations Ryan et al. (1997) came to the conclusion that the Black Sea could have been filled by salt water cascading from the Mediterranean. Even though this hypothesis has been challenged (Aksu et al., 2002b, 1999b), the recent confirmation of the excellent preservation of drowned beaches, sand dunes, and soils during Ifremer (Institut français de recherche pour l'exploitation de la mer) surveys seems to support the Ryan and Pitman hypothesis (Ryan and Pitman, 1999).

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The multibeam echo-sounding and the seismic reflection profiles acquired on the Romanian margin during our surveys revealed wave-cut terraces at an average water depth of 100 m. More evidence of seawater penetration is marked at the Bosphorus outlet by the presence of recent canyon heads mapped during the last cruise in 2002. The cores recovered on the Romanian continental shelf penetrated an erosion surface, indicating subaerial exposure well below the level of the modern Bosphorus outlet. The ^{14}C ages documented a simultaneous colonization of the terrestrial surface by marine mollusks at 7.1 k.y. B.P. The most recent palynology analysis and studies of the dynocyst population (Popescu, 2004) document a real onset of fresh-water arrival during the Younger Dryas and abrupt replacement of Black Sea dynocyst by Mediterranean population, coincident with the onset of the marine mollusks.

Keywords: rapid transgression, Younger Dryas, seismic stratigraphy, multibeam geomorphology, Bosphorus outlet, Black Sea continental shelf.

INTRODUCTION

In 1997, Ryan et al. published results of a joint Russian-American expedition carried out in 1993 on the continental shelf south of the Kerch Strait and west of the Crimea (Major, 1994; Ryan et al., 1997). They had evidence in support of a catastrophic flood of the Black Sea 7500 yr ago. Their interpretation was deduced from high-resolution seismic reflection profiles, and ^{14}C accelerator mass spectrometry (AMS) dating of faunas sampled from cores targeted on these profiles. This joint Russian-American survey revealed a buried erosional surface of shelly gravel extending across the broad continental margin of the northern Black Sea to beyond the shelf break (Evsylekov and Shimkus, 1995; Major, 1994). The cores recovered evidence of subaerial mud cracks at −99 m, algae remains at −110 m, and the in situ roots of shrubs in desiccated mud at −123 m. Each site lay well below the −70 m level of the Bosphorus bedrock sill (Algan et al., 2001; Gökasan et al., 1997). From these results Ryan et al. (1997) proposed that a drowning event in the Black Sea 7500 yr ago may have been the consequence of Mediterranean water penetration into a lowstand lake. This rapid Black Sea transgression is characterized by the deposition of a uniform marine mud drape on the terrestrial surface equally as thick in depressions as on dune crests with no sign of landward-directed onlap of the sedimentary layers in the drape (Ryan et al., 2003). The ^{14}C ages documented a simultaneous subaqueous colonization of the terrestrial surface by marine mollusks at 7100 yr B.P.[1] This age was assigned to the Holocene flooding event. However, flooding precludes the possibility of outflow to the Sea of Marmara during the prior lowstand lake stage. Recently, Aksu et al. (2002b) presented arguments for persistent Holocene outflow from the Black Sea to the eastern Mediterranean and (2002c) for noncatastrophic variations in Black Sea water level during the last 10,000 yr B.P. This contradicts the flood hypothesis.

In the meantime, international and European projects in the Black Sea facilitated two Ifremer oceanographic surveys onboard RV *Le Suroît* in 1998 and 2002 and onboard *Le Marion Dufresne* in 2004, completing previous international studies of seabed mapping and of subsurface sampling. The main objectives of these cruises, which were supported by the European ASSEMBLAGE project, are the assessment of the Black Sea sedimentary system since the Last Glacial Maximum (LGM) and the quantification of the impacts of climate change and the sensitivity of the Black Sea system to external force. These surveys, which were carried out on the northwestern continental shelf of the Black Sea, complement systematic high-resolution seismic prospecting done in the 1990s by the Romanian GeoEcoMar institute (National Institute of Marine Geology and Geo-ecology). The preliminary subsurface analysis of the Romanian continental shelf (Popescu, 2002; Popescu et al., 2004) correlated with the results of these last surveys point out that Black Sea lake levels rose on the shelf to at least the –40 to −30 m isobath as indicated by the landward limit of extent of the brackish fauna encountered in the *Dreissena* layer. Since the Black Sea was an important catchment basin for the meltwater drained from the Fennoscandian ice cap ensuing Melt Water Pulse 1A (MWP 1A) in the Bølling-Allerød period (Bard et al., 1990), it is possible that at that time the water level of the lake filled with melt freshwater, rose to the level of its outlet, and spilled into the Mediterranean. However, the onset of salt-water conditions in the Black Sea during the mid-Holocene (7500 yr B.P.) has been clearly indicated. Although contradictory hypotheses have been discussed (Aksu et al., 1999b, 2002a, 2002b, 2002c), lowstand-incised anastomosed channels mapped on the Romanian shelf and recent discoveries of well-preserved drowned beaches, sand dunes, and soils provide new support for the Ryan and Pitman flood hypothesis.

BACKGROUND

General Setting

The northwestern part of the Black Sea receives water and sediment discharge from the largest European rivers—the Danube, the Dniepr, the Dniestr, and the Southern Bug. 817,000 km^2 of the drainage basin for the Danube alone represents the expanse of these river basins. The shelf is particularly wide in this part of the basin (~140 km with a maximum of 170 km off the mouth of the Dniepr River) and narrows to both the east and west (Fig. 1). The total average sediment discharge from the Danube subsequent to damming is estimated to be around 30–35 million t/yr, out of which only 4–6 million t/yr consists of sandy material (Panin, 1997). During the glacial lowstands and especially at the beginning of interglacials, sediment discharge from these rivers was probably much higher.

The Black Sea is a marginal basin, connected to the external Mediterranean Sea over a sill located in the Bosphorus Strait (with a current 32 m sill depth). It has long been recognized that the Black Sea was isolated from the Marmara Sea and the Mediterranean during glacial intervals when levels of the latter seas fell below the sill depth of the Bosphorus (Degens and Ross, 1974). It has been postulated that the Black Sea water level oscillated independently of the global eustasy, as the Black Sea was an isolated basin for any water level below the depth of the strait. Consequently, lowstand periods in the Black Sea do not necessarily correspond to lowstands in the world ocean, but are related to regional wet-dry cycles (Major, 2002; Tchepalyga, 1984). The role that the Bosphorus Strait has played in controlling the salinity and stratification of both seas and for the production intervals of anoxia has been widely discussed (Arkhangelskiy and Strakhov, 1938; Lane-Serff et al., 1997; Muramoto et al., 1991; Rohling, 1994; Scholten, 1974). The general view is that during periods of low global sea level, connection between the Black Sea and the ocean was lost. The Black Sea, freshened from river discharge, expanded to become a well-ventilated vast lake and established

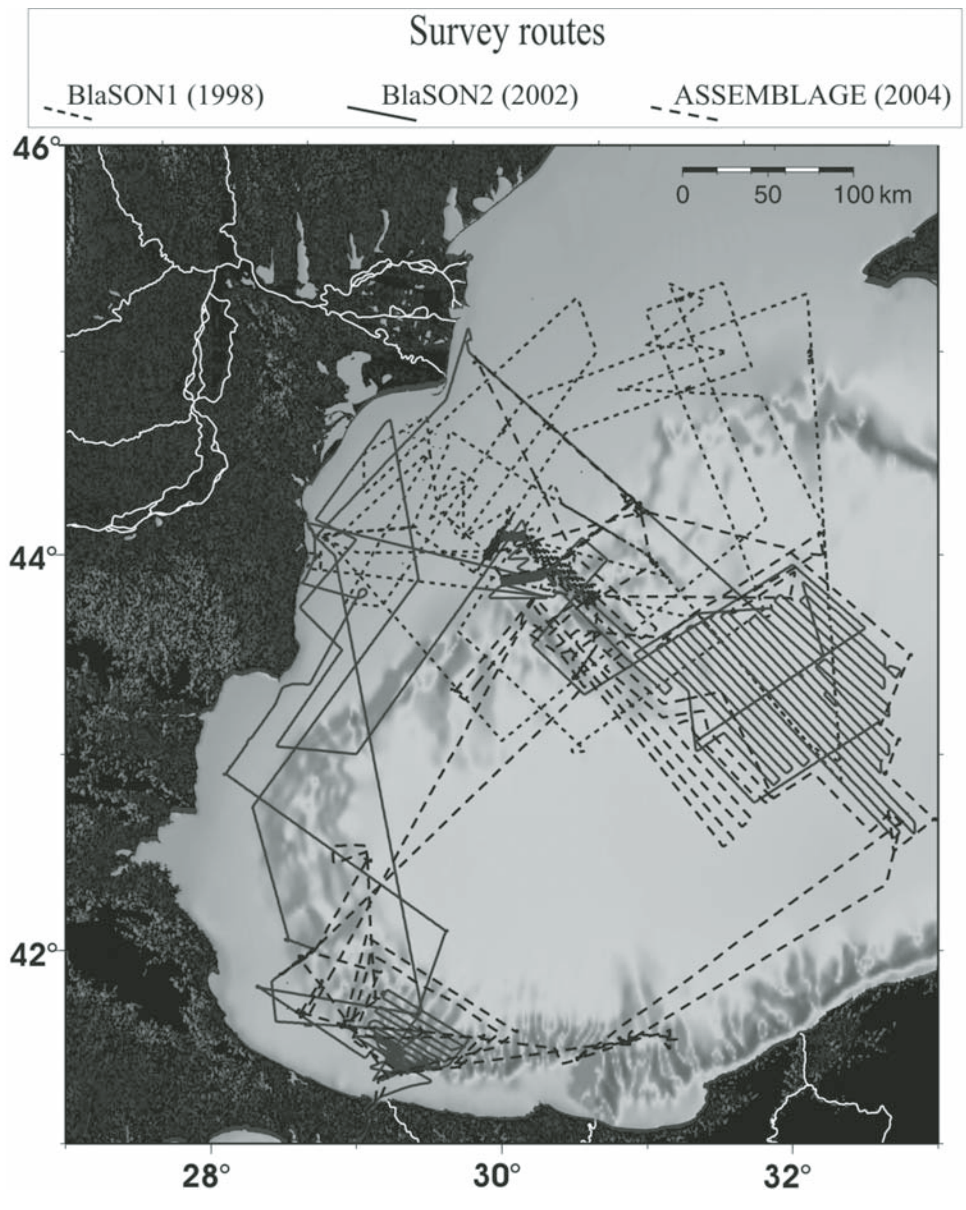


Figure 1. Bathymetry of the semi-enclosed Black Sea basin and Ifremer survey route locations.

a shoreline at the level of its outlet (Hodder, 1990; Tchepalyga, 1984) in order to export excess water to the Mediterranean. Another widely accepted hypothesis regarding the connection of the Black Sea with the external oceans postulates that this inland sea had always maintained a continuous outflow through the Bosphorus and Dardanelles Straits, even during the highly arid glacial intervals (Kvasov and Blazhchishin, 1978; Tchepalyga, 1984). According to this hypothesis, precipitation and river input into the Black Sea exceeded any loss from local evaporation. Indeed, meltwater from former ice caps in Fennoscandia, northern Asia (Grosswald, 1980), and the central Alps transformed the Black Sea into a giant fresh-water lake a number of times in the past (Federov, 1971; Ross et al., 1970), most recently during the Neoeuxinian stage of the Late Pleistocene (Arkhangelskiy and Strakhov, 1938; Federov, 1971; Nevesskaja, 1965; Nevesskaja and Nevesskiy, 1961; Ross et al., 1970).

In 1997, Ryan et al. underlined the occurrence of a widespread unconformity interpreted as an erosional surface subaerially exposed during the last glacial. Using new AMS ^{14}C dates, and abrupt changes in the organic carbon content, water content, and δ^{18}O from core material, these authors presented evidence that the Black Sea was a fresh-water lake 7500 yr ago (Ryan et al., 1997). In 2003 Ryan et al. proposed using strontium isotopes to show that the salinization was initiated earlier at 8.4 k.y. B.P. and that the 7.14 k.y. B.P. only reflected a threshold in salinity.

Variously aged sapropels sampled from the eastern Mediterranean and the Black Sea are used as arguments against the hypothesis favoring a catastrophic flooding. Sapropel S_1 in the Aegean is generally interpreted to have been deposited between ~8000 and 5500 yr B.P. (Aksu et al., 1999a, 1999b; Fontugne et al., 1994), although deposition may have lasted until 5300 yr B.P. (Rohling and de Rijk, 1999). To support the noncatastrophic flood hypothesis, Aksu et al. (1999a) propose that fresh water from the Black Sea rich in nutrients reduced the surface salinity of the eastern Mediterranean. This phenomenon, which decreases the deep circulation of this sea, favors high surface productivity with restricted circulation, providing good conditions for sapropel formation. However, Rohling (1994) suggests that sapropel formation in the Black Sea started ~550 yr later than in the eastern Mediterranean, when the denser Mediterranean waters displaced the nutrient-rich waters in the Black Sea toward the surface (Calvert, 1990; Calvert and Fontugne, 1987). This lag is probably too large to be accounted for by the catastrophic flooding hypothesis.

Paleorivers on the Romanian Continental Shelf

From previous Romanian surveys carried out by the GeoEcoMar institute, several recent paleoriver channels have been identified that incise the continental shelf down to a 90 m water depth (Popescu et al., 2004) (Fig. 2). These paleochannels are completely filled by sediments and are no longer visible in the bathymetry. These erosive features reach 400–1500 m in width and 20–30 m in depth; they present conventional asymmetry on some cross sections with point-bar– and cut-side– like structures (Fig. 3). These paleochannels are also sealed by the mud drape described by Major et al. (2002), Popescu et al. (2004), and Ryan et al. (2003), which is parallel to the sea bottom and which corresponds to the marine sedimentation (Lericolais et al., 2003). For Popescu et al. (2004) the stratigraphic position of these incisions lying directly under the discontinuity at the base of the Holocene strongly suggests that they formed during the last lowstand. The cartography of these buried channels shows that they are concentrated around two main directions. This distribution leads to their interpretation as anastomosed fluvial systems corresponding to two distinct drainage systems (Fig. 2). These would correspond to former paleo–Danube River flooding on the shelf to the outer shelf where they apparently split into several arms similar to a fluvial deltaic structure comparable in size to the modern Danube delta and lie close to the Danube Canyon (Popescu et al., 2004), which is also known as the Viteaz Canyon.

Paleocoastline on the Romanian Continental Shelf

Terraces have been recognized on many Black Sea margins, including the narrow Caucasus shelves (Ostrovskiy et al., 1977a; Shimkus et al., 1980) and the Northern Turkish shelf (Algan et al., 2002; Ballard et al., 2000). Among these terraces, shells belonging to past littoral environments were dated between 19–9 k.y. B.P. (Dimitrov, 1982; Ostrovskiy et al., 1977a; Shcherbakov et al., 1978; Tchepalyga, 1984). On the Romanian continental shelf, Popescu et al. (2004) have noticed the absence of incised river channels below 90 m water depth where a wave-cut terrace–like morphology was mapped ~100 km far from the Danube delta (Fig. 2). Wave-cut terraces are erosional surfaces created by erosion from wave action indicating the vicinity of the shoreline. Since rivers do not always generate continuous incised valleys along the entire shelf (Lericolais et al., 2001; Talling, 2000; Wescott, 1993), their absence below the isobath –90 m does not necessarily indicate the location of the paleocoastline. A good indicator of the paleocoastline is the wave-cut terrace wrapping around the head of Viteaz Canyon and present between isobath –98 and isobath –112 m (Popescu et al., 2004). Northward of Viteaz Canyon the terrace deepens again to –97 m while the height increases to 10–15 m and splits into two distinct steps. The last lowstand paleocoastline should thus have been situated between this submerged terrace and the deepest buried fluvial channels (Fig. 2).

OBSERVATIONS

Acquisition and Processing of Data

During the two campaigns on board the RV *Le Suroît*, precise positioning was given by a Differential Global Positioning System (DGPS). In some areas (i.e., dune field mosaics), realtime navigation processing within a few meters' accuracy made it possible to follow parallel track lines spaced 200 m apart at a speed of 4 knots. The vessel was equipped with EM1000 and EM300 swath bathymetry systems, and high-resolution seismic

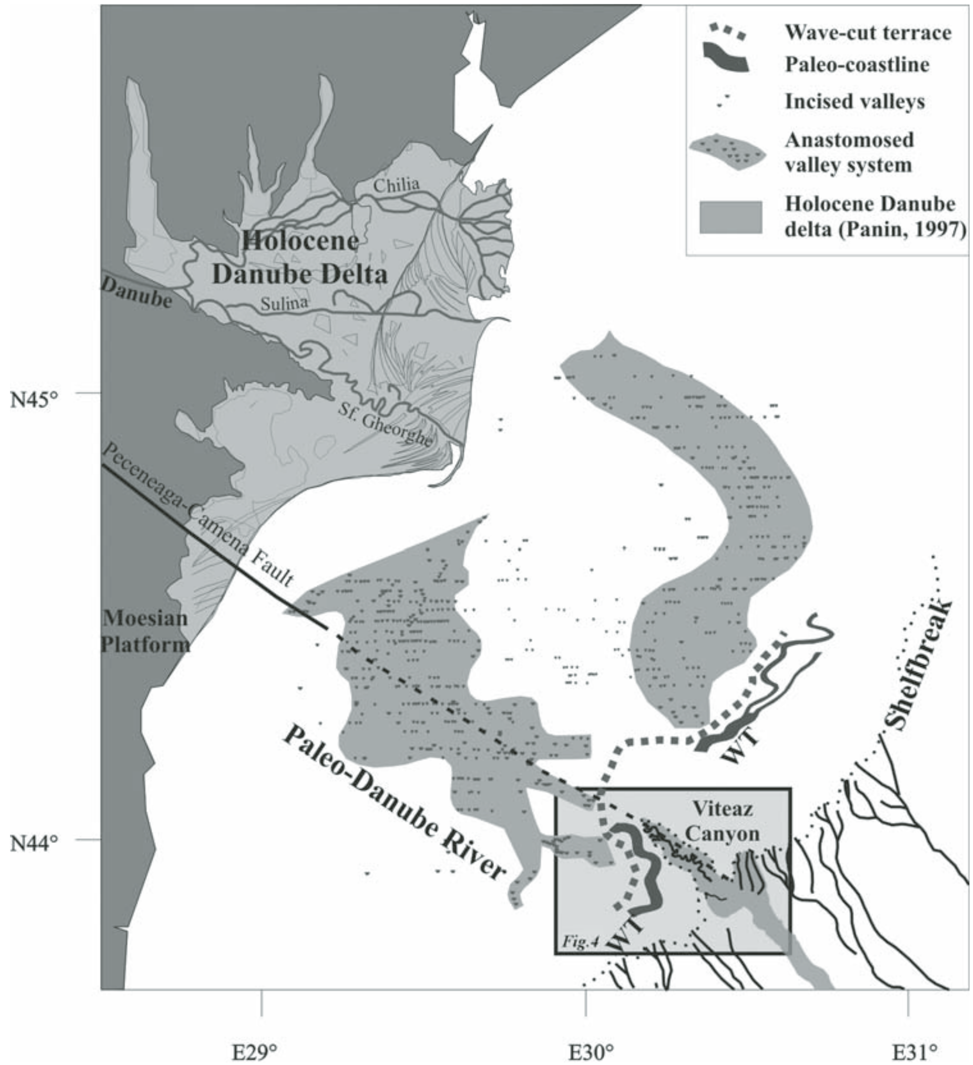


Figure 2. Paleogeographic map of the northwestern Black Sea margin during the last sea-level lowstand. Individual incised paleochannels identified on shallow seismic profiles were shot by the GeoEcoMar institute and interpreted by Popescu et al. (2004). Areas characterized by dense occurrence of buried channels cluster in two main paleo–drainage systems. The presumed position of the Peceneaga-Camena fault is drawn after Winguth et al. (2000) and Dinu et al. (2002, 2005). WT—Location of the submerged wave-cut terrace.

lines were shot simultaneously using a Chirp sonar system. All data acquisition was synchronized and digitally recorded.

The Simrad multibeam echo-sounders provided the mapped bathymetry by processing the returned echo of each sonic transmission through a selection of preformed beams trained across a strip of seafloor effectively four to seven times as wide as the water depth. Each beam gives a depth resolution in the order of 10 cm in a footprint 5 m across and 2 m long along the swath at the selected 4 knots survey speed. An image of backscatter reflectance energy was generated along with the digital elevations. Navigation, bathymetry, and image data were processed and synthesized in order to plot positions automatically, to produce bathymetric maps, and to produce an image mosaic.

The high-resolution seismic reflection source was a Chirp sonar single channel, with a frequency sweeping from 1.8 to 5.3 kHz. The digital data acquisition was done in real time on the Delph PC-based system.

More than 90 cores were recovered during these cruises. They were done using a Kullenberg piston core and in some cases by using a vibrocorer.

Surficial Features

Bathymetry data were provided by multibeam echo-sounder (Fig. 4). In 1998, the Viteaz Canyon (Popescu et al., 2004) and zone A (Fig. 4) were surveyed. On the zone A mosaic, a sea-

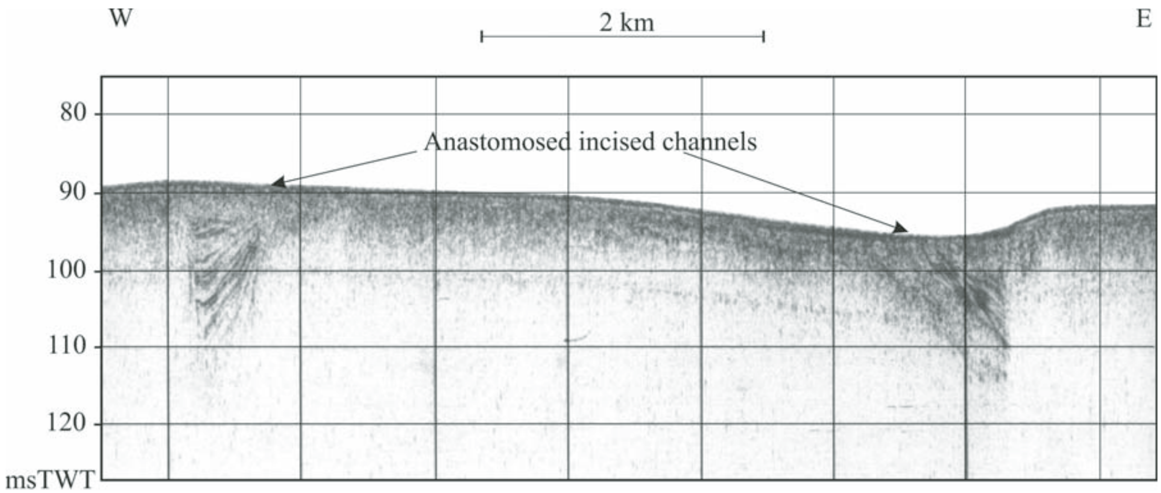


Figure 3. Paleovalleys with small cut side and point bar incised in the brackish layer (from Popescu et al., 2004). msTWT—Two Way Time (ms).

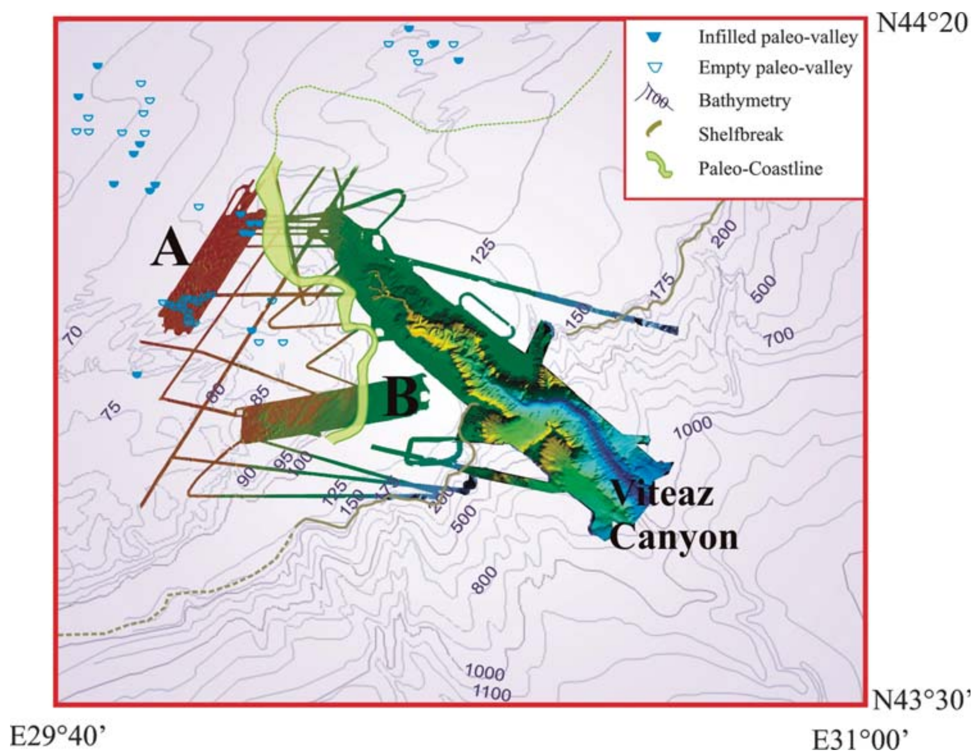


Figure 4. The multibeam bathymetry showing the 1998 (zone A) and 2002 (zone B) mosaic realized around the Viteaz Canyon head.

bed populated by sand ridges and small depressions overlying and sculpted into the eroded remains of a former terrestrial floodplain was revealed. Results were presented by Lericolais et al. (2006). These ridges located at the crest and landward of a shoreface recognized at depths of 85 to 100 m have stronger correspondence to aspect ratios of modern linear beach ridges than to those of underwater sand waves. The depressions among this first mosaic are similar in size and shape to blowouts formed through wind deflation. The ridges and depressions sit on a surface exposed by a lowstand of the Black Sea's glacial and postglacial lake. Seismic sections present forced regression sequences eroded by a wave-cut terrace at a depth of 100 m. Submergence without destruction and infilling suggests a rapid rise in the lake's surface.

In 2002, zone B (Fig. 5) was recognized. In the northeastern part of this mosaic, linear ridges 4–5 m in relief and with an average spacing of 250 m are prominent. They strike almost uniformly obliquely to a berm-like step along a north-south axis of the mosaic. In addition, depressions with a diameter from 100 to 500 m and a negative relief of 5 to 10 m are present in the southwestern half of the corridor. Parabolic features have been revealed inside these depressions (D on Fig. 5). The wave-cut terrace described as unique on the outer shelf in previous works (Major et al., 2002; Popescu et al., 2004; Ryan et al., 2003) is clearly evident on the presented mosaic. The upper surface of the berm varies around 90 m b.s.l. This is consistent with a major lowstand level situated somewhere around 100 m below the sea surface.

Subsurface Features

The ridges and depressions can be viewed in cross section using high-resolution seismic reflection profiles. The Chirp sonar provides seismic penetration to tens of meters and defines layering to the sub-meter scale. The profiles indicate that the ridges are superimposed on a reverberant "bottomset" constituted of prograding reflectors whose morphology is interpreted as forced reflection system tract (Posamentier et al., 1992) (Fig. 6). These last reflectors deepen seaward and are truncated by an erosional surface described as the wave-cut terrace on the multibeam mosaic. From the seismic profiles, it is clear that the entire area is covered by a drape of less than one meter thick, confirming that the dune system is no longer active. Everywhere across the mid and outer shelf the ridges, mounds, and depressions are draped by this thin layer of sediment with a remarkably uniform thickness of no more than a meter (Fig. 6).

Seismic profiles across the parabolic features show that these sand bodies developed inside a trough. Their seismic facies is similar to sand dune facies and can be interpreted as barkhane sand dunes (Fig. 7).

Sediment Cores

Sediments, obtained by coring, provide evidence in support of the reflection profiles. During the BlaSON (Black Sea Over Neoeuxinian) and ASSEMBLAGE (ASSEssMent of the BLAck Sea sedimentary system since the last Glacial Extreme) campaigns, an important set of gravity cores has been recovered. Here we present only results obtained on the dune fields bringing important information on the age of the onset of these sand bodies. Sampling into the interior of a ridge on the zone A dune field mosaic done by Lericolais et al. (2006) (Fig. 4) (i.e., core BLKS9837 taken in a water depth of 68 m, core length 190 cm, position N44°00.54'–E29°58.87') recovered dark sand rich in opaque heavy minerals and shell fragments (Fig. 8). Sampled minerals include quartz, garnet, and ilmenite. Shell fragments belong to the fresh-water mussels of the Dreissena species. Cores into the bedded sediments, on which the dunes have formed, sampled silty red and brownish clay with thin lenses containing fresh to slightly brackish water mollusks (Dreissena and Monodacna sp., respectively). These mollusk specimens return AMS radiocarbon dates spanning from 8585 ± 50 yr B.P. to 10,160 ± 90 yr B.P. (without reservoir and dendrochronologic calibration).

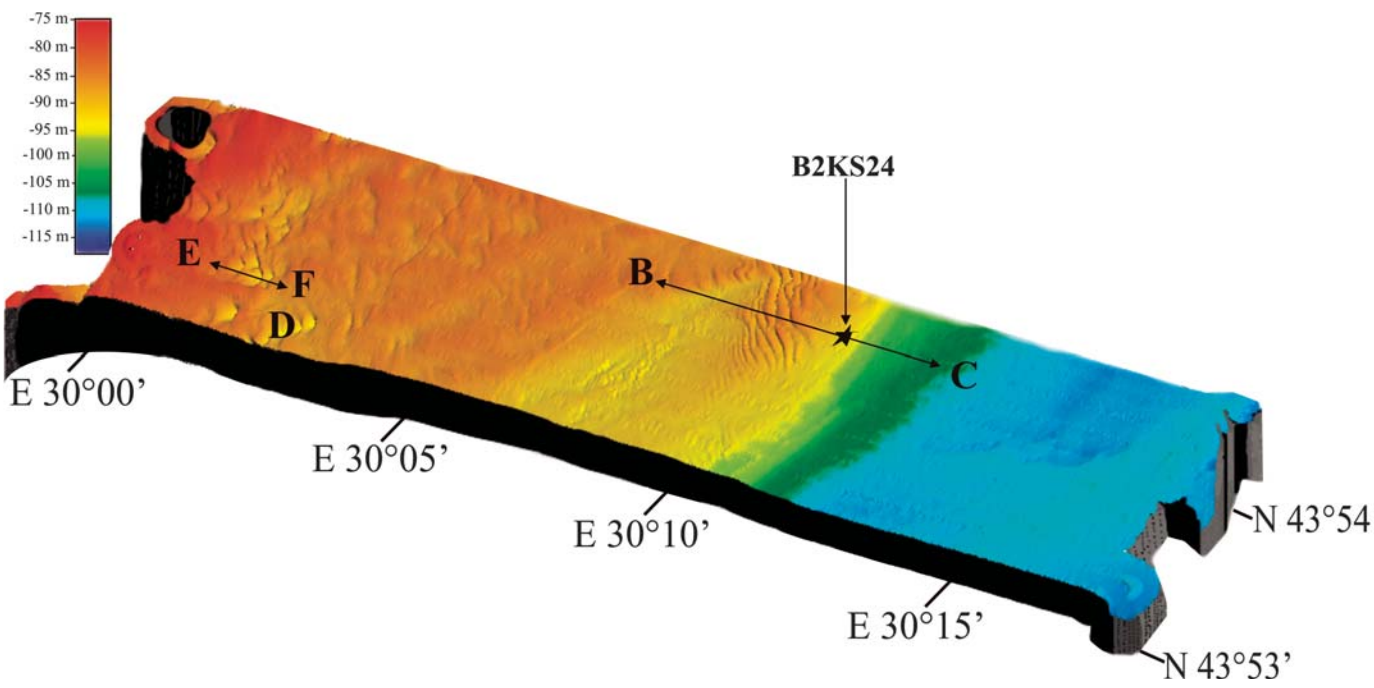


Figure 5. The multibeam bathymetry covering a sand dune field (zone B, Fig. 4). B-C—seismic profile across the sand dunes (Fig. 6); B2KS24—name and location of the core (Fig. 9); D—depression location; E-F—seismic profile across a trough with a parabolic dune (Fig. 7).

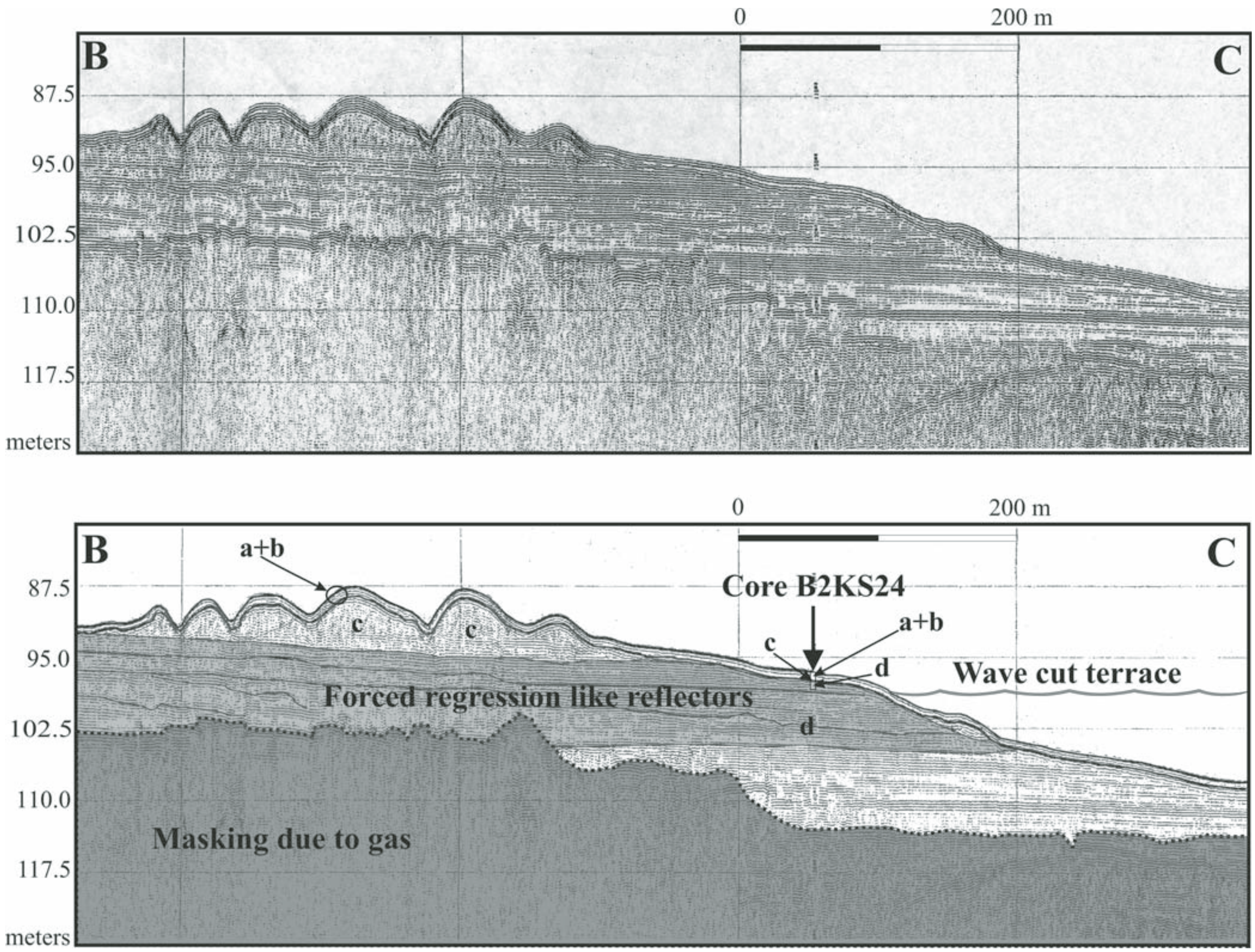


Figure 6. Seismic profile across a sand dune (B-C on Fig. 5). Core B2KS24 is presented in Figure 9; a+b corresponds to the marine layer; c—hash layer with *Dreissena*; and d—limnic sediment layer.

Sampling the top of the wave-cut terrace by piston coring (B2KS24 taken in a water depth of 96 m, core length 401 cm, position N43°53.79'–E30°11.00') provides evidence in support of the reflection profile (Fig. 9). Coring of the bedded sediments underlying the dunes and upon which they formed extracted a complete section similar to the description made by Major et al. (2002) of a core obtained during the first French-Romanian survey in 1998. The sequences can be described as follows: (1) the bottom of the core presents silty laminated clay (unit d) containing fresh to slightly brackish water mollusks (*Dreissena distincta*). These specimens return AMS radiocarbon dates of 11,040 ± 50 C^{14} B.P. (without reservoir and dendrochronologic calibration). (2) Above, marked by a sharp basal contact is a 15-cm-thick light gray layer rich in *Dreissena* sp. detritus dated between 8760 ± 40 C^{14} B.P. and 8620 ± 50 C^{14} B.P. (described as the hash layer by Major et al., 2002). Above the layer rich in *Dreissena* sp. detritus is a marine sequence characterized by a mollusk assemblage that is exclusively salt-water species, such as *Mytilus edulis* (also known as *Mytilaster*) and *Cerastoderma edule*. Those last shells sampled near the base of the uppermost drape, which is dated 2820 ± 30 C^{14} B.P., date between 6520 ± 40 C^{14} B.P. and 5525 ± 35 C^{14} B.P. Recent palynological and dynocyst analyses (Popescu, 2004) on samples from within these cores indicate an abrupt fresh-water transition during the Younger Dryas followed by an abrupt replacement of the endemic Black Sea dynocysts by a Mediterranean population at 7150 C^{14} B.P.

INTERPRETATION

Interpretation of the Linear Ridges

Are They Sand Waves?

In underwater environments the asymmetrical linear ridges would be classified as sand waves (Allen, 1982) or large dunes (Ashley, 1990). One observes sand waves in many places on

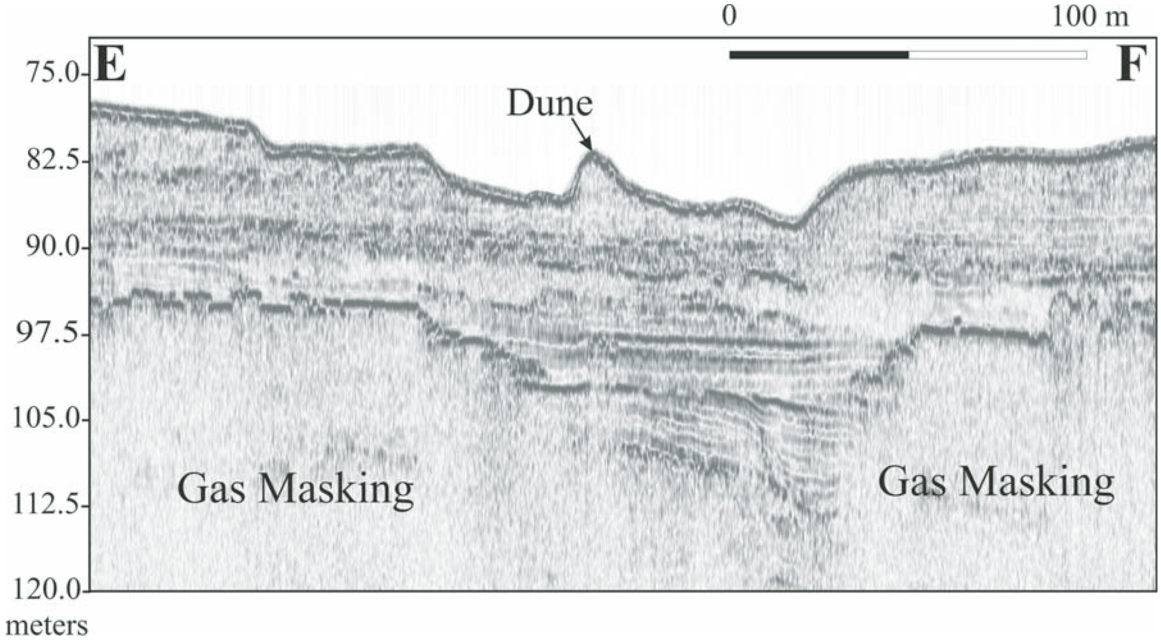


Figure 7. Seismic profile across a barkhane dune inside a trough; E-F on Figure 5.

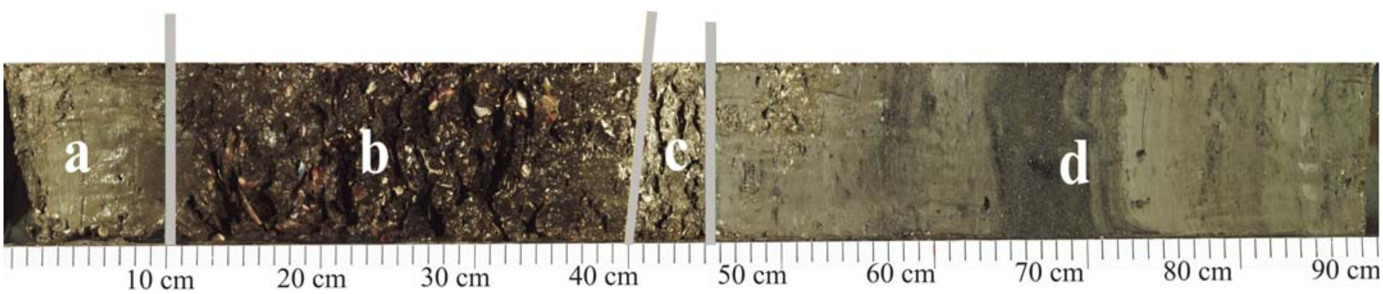


Figure 8. Core BLKS9837 (N 44°0.54'–E 29°58.87') recovered on the dune field at a depth of 68 m (Lericolais et al., 2006): (a) Layer with *Modulus* dated from 3000 yr B.P. to present; (b) layer with *Mytilus edulis* dates between 6900 and 3000 yr B.P.; (c) layer with *Dreissena*—the youngest is dated 7500 yr B.P.; and (d) limnic sediment layer dated 10,160 yr B.P.

the continental shelf. Examples include the Bay of Fundy (Dalrymple, 1984), the English Channel off the Normandy coast of France (Berné et al., 1988), the North Sea (McCave, 1971), and the East Coast of the United States (Swift and Freeland, 1978). Sand waves are products of sedimentary environments characterized by high-energy bottom currents (Allen, 1982). The strongest currents are generated by the ebb and flood of tides through inlets (Berné et al., 1998; Kenyon and Stride, 1968; McCave, 1971; Terwindt and Brouwer, 1986). Such currents tend to amplify within estuaries (Kostaschuk and Villard, 1996). Sand waves also form in wave-dominated environments (Berné et al., 1988) or by oceanic currents (Flemming, 1988). In seas without significant tide, such as the Mediterranean, sand bodies are generally found as wedges built of shoreface deposits. But dunes are also observed around major delta systems such as the Rhône (Amorosi et al., 1999) and the Nile (Stanley, 1996).

Sand waves in estuary settings have varying asymmetries depending upon whether they are flood or ebb dominated. Their internal inclined beds show reversals in dip direction as flood-dominated sand ridges become ebb-dominated and vice versa. Active sand waves are features of the shallow inner shelf (Swift and Field, 1981) and its coastal bays. Sand waves on the outer shelf in depths of 110 to 140 m on the southern Celtic margin (Berné et al., 1998) are relict bedforms deposited when the sea surface lay more than 100 m below its present level. Sand waves are flow-transverse bedforms whereas linear ridges are flow parallel. The Black Sea's linear ridges are also relict as shown by the drape that covers them.

Are They Longshore Bars?

Longshore bars are described in lake environments, such as Lakes Michigan, Huron, and Erie (Davidson-Arnott et al., 1996). These bars form depths less than 4 m. They are typically 10 km long, 25–50 m wide, and 0.2–0.75 m high (Wattrus and Rausch,

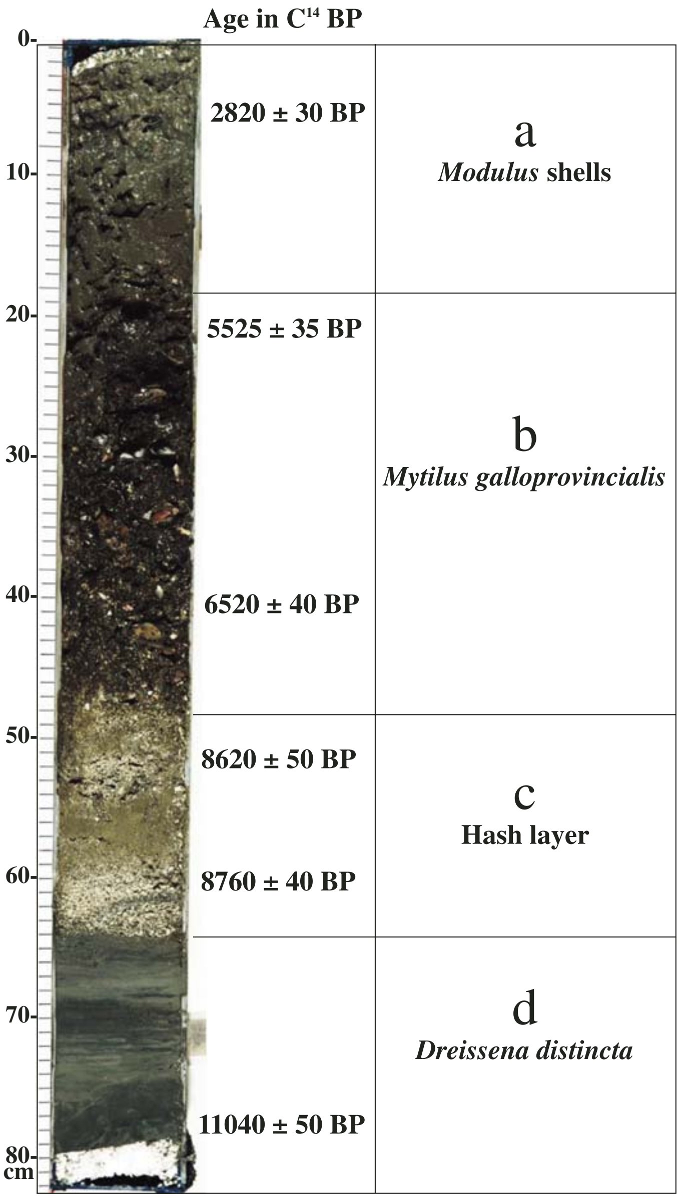


Figure 9. Core B2KS24 (N43°53.79'–E30°11.00') recovered on dune field B (Fig. 4) and on top of the wave-cut terrace (Fig. 6).

2001) and develop as multiple sets of ridges, which are aligned parallel to the shore. The linear ridges surveyed in this study are aligned somewhat obliquely to the regional bathymetric contour and to the paleoshoreline outlined by wave-cut terraces (Figs. 4 and 5). The striking characteristic of shoreface-attached ridges is that they orient at an oblique angle to the shoreline with their seaward tips directed toward the prevailing current (Snedden and Bergman, 1999). If the Black Sea's linear ridges correspond to relict-drowned, shoreface-attached sand ridges, then they should have formed when the Black Sea level lay 5 m above the crests. The level of the Black Sea would have been approximately –60 to –75 m at the time of their formation.

Are They Delta Mouth Bars?

The Black Sea's linear ridges could be interpreted as successive mouth bars of a deltaic body constructed during a lowstand of sea level. Such mouth bars are present today within the Saint George II distributary channel of the Danube delta (Panin, 1997). Wave-dominated deltas present a series of shore-parallel sand ridges forming as mouth bars built up and out to form a new beach. This results from a net supply of bedload material from the river. In any case, the utilization of such ridges as water-level indicators presupposes a recognizable interface between the wave-built foreshore and the overlying aeolian forest within a given ridge (Otvos, 2000). For the Black Sea's ridges to be mouth bars, they had to have been constructed during a regression that brought the shoreline to where they are now found, confirming a lowstand at that position.

Are They Coastal Dunes?

Coastal dunes sensu lato result from the accumulation of sand transported along the shore by the combined action of winds and waves (Carter et al., 1990a) in littoral environments (Carter et al., 1990b). Foredunes and parallel dunes, beach ridges and cheniers are long, linear, ridge-forming formations representing the locations of ancient shorelines (Meldahl, 1995). They occur on low-relief coastal plains, standing with several meters of relief above their surroundings. In terrestrial sand seas, aeolian dunes are classified on the basis of shape and the number of facies (McKee, 1979). The abundant dune type is linear, with the long axes oriented parallel to the prevailing or dominant wind and resultant sand drift direction (Wilson, 1972). Linear dunes, accumulating rather than migrating during active phases, tend to develop in wind regimes with a significant degree of directional variability and where sand supply is minimal (Thomas, 1997).

Coastal dune formation is enhanced by the input of littoral sediments during periods of low sea level. Foredune ridges have an aeolian (foredune) cap on the accreted berm that indicates the occurrence of winds sufficiently strong to transport beach sand (Roy et al., 1994). Beach ridges are low-relief, wave-formed berms that rarely rise more than 3 m above mean sea level, but foredune ridges and swale have greater amplitude (3–5 m) and crest elevations of 7–10 m (Thom and Hall, 1991). For the linear ridges on the Black Sea's shelf to be coastal dunes, they had to have been constructed during a regression that exposed the surface upon which they grew, and they had to survive intact any subsequent shoreface erosion as they became submerged.

Interpretation of the Hollows

In subsea environments, small enclosed and unfilled hollows are rare. The exception is for substrate which is easily eroded, where bedforms may climb at a negative angle and cannibalize the substrate. If these bedforms are sinuous or crescentic, they will produce swale by scouring to the lee face of the depression (Berné et al., 1998). Another exception is the small circular pockmark left by venting gas. Bathymetric surveys in other oceans and seas have not revealed hollows displaying the size and shape interconnectedness as those mapped on the Romanian shelf. However, enclosed depressions like those in the surveyed corridor are characteristic of windy and arid environments (Shaw and Thomas, 1997). The cavities are eroded into the substrate by deflation processes, with the magnitude of the excavation ultimately limited by the groundwater table (Laity, 1994). The depressions at the scale of those mapped on the Romanian shelf are similar to pans. Pan initiation and growth depend on materials susceptible to deflation such as poorly consolidated clay-rich material that curls, flakes, and blows away upon desiccation. Pans in dune settings may occur as a string of depressions aligned along a former river course and its braid plain. Pans often transform into ponds during wet phases, generally from groundwater recharge. Pond and marsh sediments have been reported from between the dunes of sand seas during intervals of substantially increased moisture (Lancaster et al., 1994). Present-day ponds are reported in the desert seas of northwestern China (Wang et al., 2002). In coastal domains, pans correspond to blowouts formed when strong onshore winds erode gaps in a single foredune or series of beach ridges (Giles et al., 1997). Leakage of sand occurs mainly through erosion of the dune front by storm waves or by landward movement of sand through blowouts or parabolic dune migration (Carter et al., 1990a) or by a combination of these factors.

Subsea or Subaerial Origin?

To evaluate an origin for the Black Sea's linear ridges their dimensions and spacing are compared to subaerial dunes and subsea sand waves. In Figure 10, the height and spacing of the linear ridges on the Romanian and Ukrainian margins fall within and along the trend of the population defined for sedimentary coasts, where combined wind and wave action shape coastal ridges on the barrier surface with dimensions and spacing situated at the border between aeolian and subaqueous features. Various subsea sand ridges on the European Atlantic margin have an average height around 4.7 m (Berné et al., 1998) and are comparable to the 4.25 m calculated from a bibliographic synthesis (Allen, 1982).

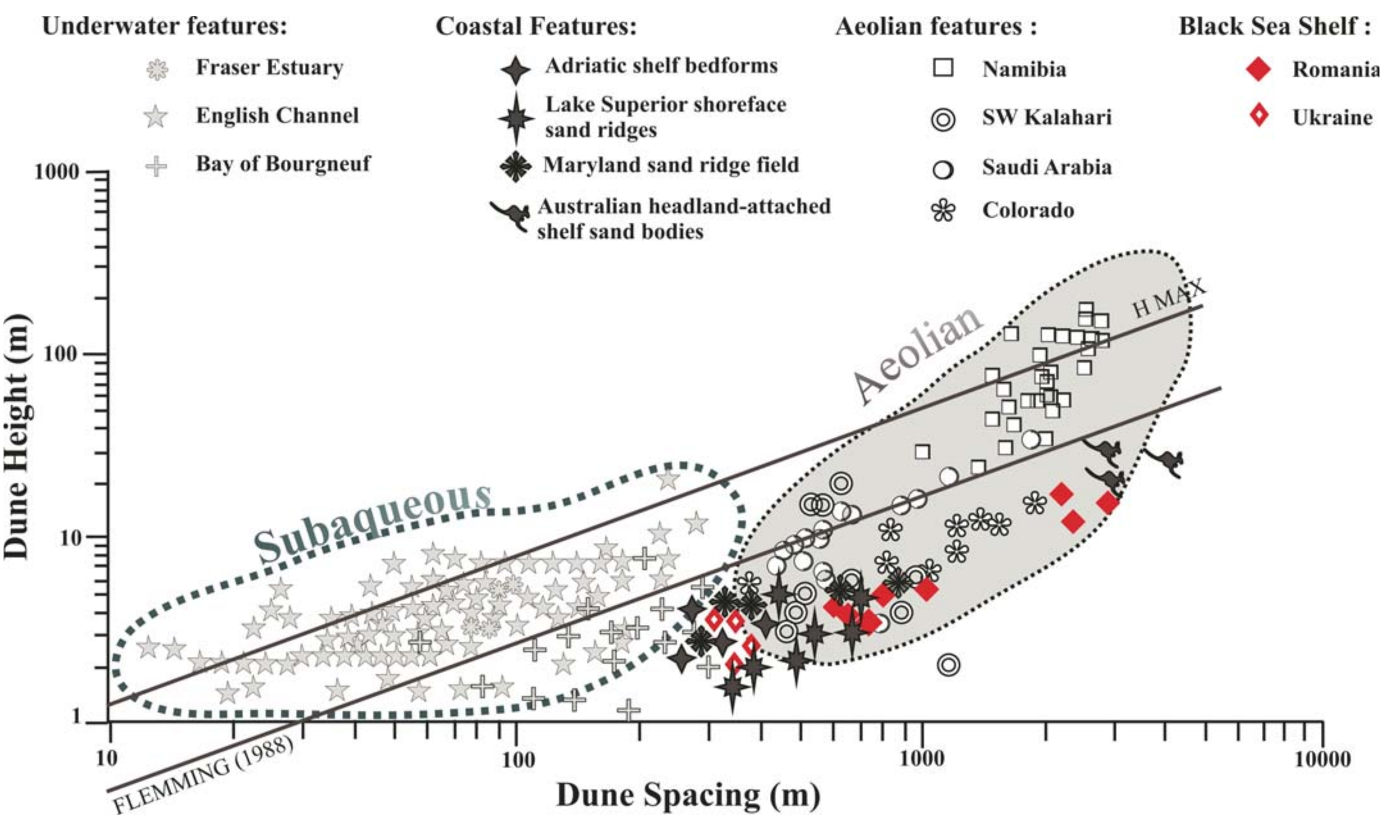


Figure 10. Dune height (m) versus dune spacing (m). Logarithmic diagram of aeolian and subaqueous dunes. The surveyed dunes of the Romanian and Ukrainian shelf are presented among other dunes from both origins (aeolian and subaqueous). The lower and the upper lines have been established by Flemming (1988) for subaqueous dunes; H—$0.0677_{*}L^{0.8098}$ and H_{max}—$0.16_{*}L^{0.84}$. References used for:

- Underwater features: Kostaschuk and Villard (1996) for Fraser River, Berné et al. (1998) for English Channel, Berné (1991) for Bay of Bourgneuf;
- Coastal features: Correggiari et al. (1996) for Adriatic shelf bedforms, Wattrus and Rausch (2001) for Lake Superior shoreface sand ridges, Swift and Field (1981) for Maryland sand ridge field, and Roy et al. (1994) for Australian headland-attached shelf sand bodies;
- Aeolian features: Lancaster (1998) for Namibian dunes, Thomas (1997) for dunes from the Kalahari desert, Fryberger et al. (1983) for the Saudian sand sea, and Fryberger et al. (1979) for Colorado eolian "sand-sheet" deposit; and
- Black Sea shelf: this paper for the Romanian shelf and Ryan et al. (1997) for the Ukrainian shelf.

The height/spacing ratio of subsea ridges (Berné et al., 1998; Flemming, 1988) is parallel to but offset from the trend of the aeolian features. The height and spacing of the linear ridges on the Romanian and Ukrainian margins are more representative of the coastal population than the subsea one. Similar features to the studied dune formation have been reported at the southern part of Cape Flattery (Australia), where currently active parabolic dunes are transgressing over an older strandplain surface (Pye, 1993). The Romanian-surveyed dune fields are located directly foreshore the wave-cut bluff delineating the paleoshoreline situated at100 m deep. Waves, wave-dominated currents, and winds are mechanisms for moving and depositing sand on shorefaces and beaches of the open coast. In the case of a rapid sea-level rise, coastal processes only have time to build low, discontinuous foredunes on the barrier surface that provide little impediment to storm washover and wind deflation resulting in the formation of blowouts or pans.

Age of the Ridges and Depressions

A grip on the age of the Black Sea's linear ridges is obtained by analysis of the materials beneath and above them. The uniform drape of the euryhaline mollusk–bearing mud across the erosion surface provides an upper minimum limit, and the brackish mollusks in the shell hash with a marine strontium isotopic signature (Ryan et al., 1997) refine the end of ridge and depression activity at 8.5 k.y. B.P. The materials within the dunes consisting of shell-bearing sand with *Dreissena* fragments give dates of 8.5 k.y. B.P., but dune formation could have commenced earlier because core penetration into the sand was very limited. The prograding reflectors upon which the dunes sit and into which the depressions have eroded as well as how the wave-cut terrace is marked provide a lower limit. The red-brown clay in core BLKS9814 at –55m and lying 1.2 m below the erosion surface contains a lens rich in *Monodacna* dated at 9580 yr B.P. (Ryan et al., 2003). Since the linear ridges in the surveyed corridor lie landward of this core and their substrate may not have been as deeply eroded as that of the outermost shelf, it is probably safe to infer that the ridges have been constructed after this later date and that they migrated over floodplain deposits as young as 9580 yr B.P.

Evidence of Subaerial Exposure

An erosion surface is present everywhere across the Romanian, Ukrainian, and Turkish shelves (Demirbag et al., 1999). The erosion truncates the sediment fill of river channels, some of which display the point-bar bedding geometry of meandering streams (Fig. 3). The sediments corresponding to the eroded substrate are found to possess high bulk density (1.8–2.2 g/cm^3) and low water content (<30%), compatible with de-watering in a terrestrial setting (Ryan et al., 1997). Root casts and desiccation cracks have also been observed in soils below the erosion surface (Major, 1994). The strata cut by erosion are stiff clays with occasional silt/sand lenses containing mollusks that have been dated on the Ukrainian and Bulgarian shelf to glacial periods in the late to middle Pleistocene. The erosion has even cut into Neogene strata (Kuprin et al., 1974; Shcherbakov and Babak, 1979; Shopov et al., 1986).

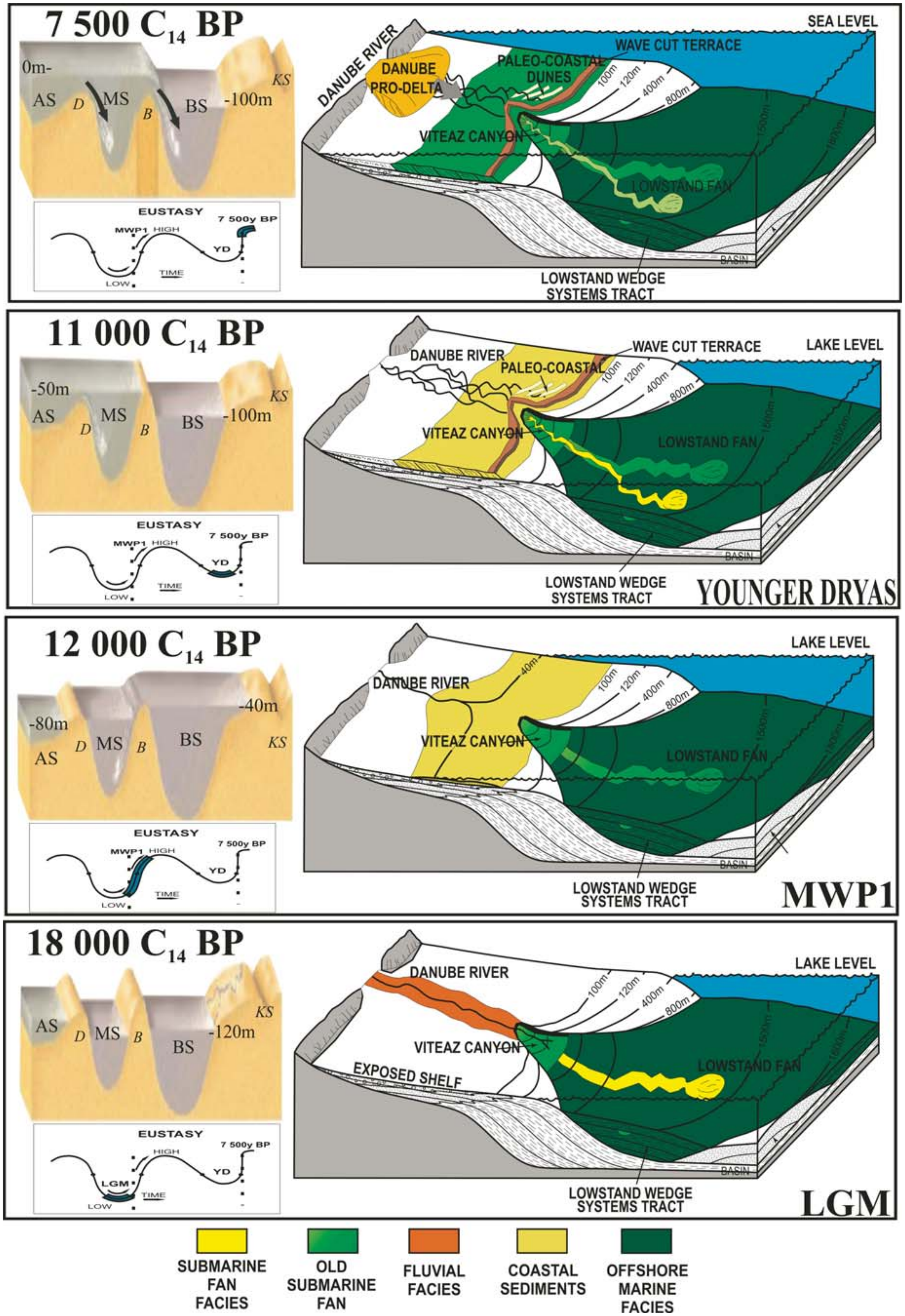
The dry sediment and desiccation features on the erosion surface are compatible with the superimposed ridges and depressions forming in coastal, river mouth, and shoreface environments.

DISCUSSION

The transition of the Black Sea system from a fresh-water lake to a marine environment was perhaps one of the most dramatic Late Quaternary environmental events in the world. During the Last Glacial Maximum, 21,000 yr ago, the Black Sea was probably a giant fresh-water lake as proposed by Arkhangelskiy and Strakhov (1938) or at least a brackish enclosed basin; its water level stood more than 120 m below today's level. As early as 2001, analysis of high-resolution seismic reflection profiles, Chirp and side-scan data, together with piston cores from the first BlaSON survey realized on the Danube fan (Popescu et al., 2001), provided new insight into the recent sedimentation processes in the deep northwestern Black Sea. It was determined that the last channel-levee system on the Danube fan developed during the Neoeuxinian lowstand (stage 2) in a semi–fresh-water basin with a water level ~120 m lower than today. Sediments supplied by the Danube were transported to the deep basin through the Viteaz Canyon (Fig. 4). A definite relationship exists between water level and Danube fan sedimentation: When water level is close to the shelf break during lowstands, fluvial sediments are transported to the deep-sea fan, while fan construction is essentially interrupted during water-level highstands (Fig. 11). Functioning of the deep-sea fan is a good indicator of lowstand periods (Popescu et al., 2001; Winguth et al., 2000; Wong et al., 1997).

Since the Black Sea was in close vicinity to the Scandinavian-Russian ice cap, the supply of the melting water from the glaciers into the Black Sea through the major drainage system constituted by the large European rivers (Danube, Dniepr, Dniestr, and Bug) was assimilated into the brownish layers described in cores (Bahr et al., 2005; Major et al., 2002). The water brought to the Black Sea after Melt Water Pulse 1A (MWP 1A) at ~12,500 C^{14} B.P. (14,500 yr cal. B.P.) (Bard et al., 1990) was supposed to have a large enough impact that the water level rose to between

Figure 11. Schematic scenario inspired by Posamentier and Vail (1988) of the water-level fluctuations in the Black Sea since the Last Glacial Maximum deduced from geomorphological results, supported by the Danube deep-sea fan functioning (Popescu et al., 2001), the results from palynology and dinoflagellates (Popescu, 2004), and the paleocoastline position. LGM—Last Glacial Maximum; MWP1—Melt Water Pulse 1; YD—Younger Dryas; AS—Aegean Sea; MS—Marmara Sea; BS—Black Sea; KS—Kerch Strait; D—Dadanelles; B—Bosporus.
1 yr b.p. means years before present (1950) with neither correction for reservoir age nor calibration to calendar years. In Ryan et al (1997, 1999) ages were expressed in calendar years with 7500 cal yr b.p. equivalent to 7100 yr b.p.


7 500 C14 BP
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LOW
TIME
SEA LEVEL
DANUBE RIVER
WAVE CUT TERRACE
DANUBE PRO-DELTA
PALEO-COASTAL DUNES
100m
120m
400m
800m
VITEAZ CANYON
LOWSTAND FAN
BASIN
LOWSTAND WEDGE SYSTEMS TRACT
11 000 C14 BP
-50m
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-100m
KS
LAKE LEVEL
DANUBE RIVER
PALEO-COASTAL
WAVE CUT TERRACE
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LOWSTAND FAN
LOWSTAND WEDGE SYSTEMS TRACT
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LAKE LEVEL
DANUBE RIVER
40m
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LAKE LEVEL
DANUBE RIVER
VITEAZ CANYON
EXPOSED SHELF
LOWSTAND FAN
LOWSTAND WEDGE SYSTEMS TRACT
LGM
SUBMARINE FAN FACIES
OLD SUBMARINE FAN
FLUVIAL FACIES
COASTAL SEDIMENTS
OFFSHORE MARINE FACIES

–40 m to –20 m, where the *Dreissena* layers were deposited. The –40 m upper limit is interpreted from our records, which are not exhaustive, and the –20 m limit is provided by Valentina Yanko-Hombach (Yanko, 1990). This last value for the transgression upper limit would have brought the level of the Black Sea even higher to the Bosphorus sill, and possible inflow of marine species like Mediterranean dinoflagellate population can be envisaged (Popescu, 2004). The rise in the Black Sea water level, which stayed between fresh and brackish conditions, stopped the deep-sea fan sedimentation (Fig. 11).

Palynological studies done on BlaSON cores (Popescu, 2004) show that during the Younger Dryas a cool and drier climate prevailed. Northeastern rivers converged to the North Sea and to the Ancylus Lake (Baltic Sea) (Jensen et al., 1999), giving reduced river input to the Black Sea and resulting in a receding shoreline. This is consistent with some evaporative drawdown of the Black Sea and correlates to the evidence of an authigenic aragonite layer present in all the cores studied (Jermannaud, 2004). This drawdown is also confirmed by the determination of the forced regression–like reflectors recognized on the dune field mosaics and dated from this period. This lowered sea level in the Black Sea persisted afterward as implied by (1) the continuously dry climatic conditions in the region starting around 13 k.y. B.P. and lasting till 8 k.y. B.P. (large percentages of herbs and steppe elements were described in the cores [Popescu, 2004]), and (2) the dune formation between 9.7 k.y. B.P. and 8.5 k.y. B.P. on the desiccated northwestern Black Sea shelf at –100 m. The Younger Dryas climatic event had lowered the Black Sea water level and the presence of the coastal sand dunes and wave-cut terrace confirmed this lowstand (Fig. 11). Numerous Russian authors (e.g., Ostrovskiy et al., 1977b; Shcherbakov et al., 1978, and 1979; Shimkus et al., 1980) have indicated a sea-level lowstand at about –90 m, based on the location of offshore sand ridges described at the shelf edge south of Crimea. The anastomosed buried fluvial channels described by Popescu et al. (2004), which suddenly disappear below 90 m depth, and a unique wave-cut terrace on the outer shelf, with an upper surface varying between –95 and –100 m, are therefore consistent with a major lowstand level situated somewhere around 100 m deep. Around Viteaz Canyon the paleocoastline was forming a wide gulf in which two rivers were flowing (Fig. 2). Previous studies have proposed a depth of 105 m for this lowstand according to a regional erosional truncation recognized on the southern coast of the Black Sea (Demirbag et al., 1999; Gorur et al., 2001) and also based on evidence from a terrace on the northern shelf edge (Major et al., 2002).

Preservation of these sand dunes and buried small incised valleys can be linked with a rapid transgression during which the ravinement processes related to the water-level rise have no time to sufficiently erode the sea bottom (Benan and Kocurek, 2000; Lericolais et al., 2004). Around 7.5 k.y. B.P., the surface waters of the Black Sea suddenly attained present-day conditions owing to an abrupt flooding of the Black Sea by Mediterranean waters, as shown by dinoflagellate cyst records (Popescu, 2004) and supposed by Ryan et al. (2003, 1997). This can also be related to the beginning of the sapropel deposit, which was widespread and synchronous across slope and basin floor. Popescu (2004) demonstrated that at 7160 yr B.P. a sudden (<760 yr according to the resolution of their data) inflow of a large volume of marine Mediterranean waters caused an abrupt increase in salinity, which attained the present-day euxinic values. This inflow of marine waters is confirmed by the abrupt replacement of fresh to brackish species by marine species. Furthermore, the model developed by Siddall et al. (2004) shows that ~60,000 m^3/s of water must have flowed into the Black Sea basin after the sill broke and it would have taken 33 yr to equalize water levels in the Black Sea and the Sea of Marmara. Such a sudden flood would have preserved lowstand marks on the Black Sea's northwestern shelf.

CONCLUSION

The results presented here confirm that the Holocene climate modifications in the intercontinental setting of Eastern Europe had significant implications for the behavior of the Black Sea water-level fluctuations. Rare preservation of an intact regressional surface, a big drainage basin fed by meltwater from the Fennoscandian ice cap, and the two-way exchange of water through the Bosphorus and Dardanelles Straits are the major consequences resulting from reconnection between the Marmara Sea and the Black Sea. The sand dune fields and associated wave-cut terrace are interpreted as coastal zone relicts that persisted between 9580 and 8585 yr B.P. A wave-cut terrace presently at –100 m and dunes and pans between –80 and –65 m would have lain well below the level of the external ocean (Fairbanks, 1989; Lambeck and Bard, 2000). As the Black Sea moved from an enclosed to a semi-enclosed basin, and as it was the receiving basin of the meltwaters, it is clear that water fluctuation was more directly linked to climate change while the global ocean had a stronger hysteresis. The water issued from the melting of the Fennoscandian ice cap ensuing from the Bølling-Allerød warming was drained to the Black Sea by major European rivers (Dniepr, Dniestr, and Bug) and the Danube drained part of the meltwater from the Alps. The Black Sea, whose brackish water level was under –120 m at that time, rose up to –40 or even –20 m. The onset of the Younger Dryas cool and drier climate favored the regression processed in the Black Sea and the level went down to –100 m as witnessed by the coastal dunes and the anastomosed river system. These conditions prevailed till the water level of the Marmara Sea reached the sill of the Bosphorus, initiating the entrance of marine waters into the Black Sea. The resulting transgression was so rapid that the coastal features were preserved.

A Black Sea lake preserved below global sea level would have needed a Bosphorus barrier shallower than the external ocean. Burial of the dunes and pans by a drape of mud is not sufficient to imply a sudden infilling of the depression once the Bosphorus barrier was breached. But taken with evidence of an abrupt transition from shell hash (the very condensed layer with brackish fauna) to mud, and the impressive preservation of dunes and pans with no preferential infilling of the depression, it seems likely that the Black Sea experienced a rapid terminal transgression. The Caspian Sea, which seems to have encountered a similar phenomenon, with the exception of the last reconnection, has similar sand dunes as high as 20 m that lie parallel to the seashore and that contain sandy particles and fragments of shells in the coastal regions of Mazandaran and Gilan (Iran). The relict Black Sea dunes are occasionally cut across by wind blows witnessing a lowstand. The Black Sea water-level fluctuations appear to be directly linked to climate variability (Kvasov, 1975; Svitoch et al., 2000). The Black Sea, like other enclosed basins, when not connected to the Mediterranean, reacted by reaching highstand after the meltwater pulses, which were directly connected to the drainage of the Fennoscandian ice cap and lowstand through evaporation in dry periods as was the Younger Dryas in this region (Popescu, 2004). Such behavior is additionally reported for Turkish lakes during the same period, which followed a water-level fluctuation directly linked to climate variability and encountered a drawdown in the Younger Dryas (Fontugne et al., 1999; Kashima, 2003), as did the giant Black Sea lake. In addition, the fresh water provided by the melting of the ice was stopped during the Younger Dryas and the former fresh-water–providing rivers were diverted to the north—the North Sea and the Baltic Sea were free of ice.

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