Chapter 16

Geological and Geomorphological Factors and Marine Conditions of the Azov-Black Sea Basin and Coastal Characteristics as They Determine Prospecting for Seabed Prehistoric Sites on the Continental Shelf

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Introduction

The Black Sea lies at the junction of three major cultural areas: Europe, Central Asia, and the Near East. The history of primary occupation and cultural exploitation of the Black Sea basin goes back to 1.89 million years ago (Dmanisi, Georgia), as is documented by numerous open-air archaeological sites, the frequency of which indicates a high concentration of human activity from the Lower Paleolithic to the Early Iron Age (Özdoğan 2007). Comprehensive study of these sites contributes to some of the most interesting debates in European prehistory, among which are the spread of anatomically modern humans, the transition to an agricultural economy, the repercussions of early urbanization across Eurasia, and others, which play a crucial role in enduring discussions about the impact of complex Near Eastern societies on European societies. Fluctuations in sea level and the commensurate shrinking and expansion of littoral areas had considerable impact on the settlement pattern of prehistoric societies of the Black Sea region, and submerged archaeological landscapes are highly possible (Stanko 2007).

The details and taphonomic conditions of the Black Sea are unusual. It is the world’s largest anoxic (oxygen-free) marine basin. Its strongly stratified water column possesses (1) a thin, well-oxygenated surface layer (20–30 m) with low salinity and warm temperatures, (2) a low-oxygen (suboxic) transition layer (30–150 m), and (3) a thick bottom layer of colder, denser, and more saline water lacking oxygen but high in sulfides. Few organisms feed on organic material in its oxygen-starved depths; this highly favorable underwater environment preserves archaeological material, such as shipwrecks, creating the world’s largest underwater museum (Ballard 2008).

The Black Sea is the easternmost of the seas of the Atlantic Ocean basin. If we take the ratio of the sea volume to the summary area of the cross-sections of all its straits (which is 0.04 km² for the Bosphorus and 0.02 km² for the Kerch Strait) as a measure of isolation of a sea basin, then the Black Sea can be considered the most isolated sea of the Global Ocean (Zubov 1956). Its maximum length (along 42° 29’ N lat) and width are 1148 km and 611 km, respectively. Its surface area (excluding estuaries, such as the Dnieper-Bug liman — liman is a local term for ancient estuaries in the Black Sea and Sea of Azov) and its volume are about 416,790 km² and 535,430 km³, respectively, and the maximum depth is 2212 m (Ivanov & Belokopytov 2013). The Sea of Azov is connected to the Black Sea via the Kerch Strait and has an area of 39,000 km² and a volume of 290 km³. The maximum length, width, and depth of the Sea of Azov are 360 km, 180 km, and 14 m, respectively. On average, the level of the Black Sea is 7–11 cm lower than that of the Sea of Azov and 30 cm higher than that of the Sea of Marmara.

Climatic conditions of the Black Sea basin are determined by its geographical location and general atmospheric circulation. The northern part of the basin is located in the temperate climate zone while the southern part is in the subtropical climate belt. In January, mean air temperature above the central basin is 8°C; it decreases to 0–3°C with an absolute minimum of −30°C in the northwest. Mean air temperature in July is 22°C to 24°C, sometimes reaching 35°C. In the Sea of Azov, cold winters (down to −33°C) can be followed by dry and sultry summers (up to 40°C). Annual precipitation over the Sea of Azov and the western and northwestern Black Sea is 300 mm to 500 mm; in the southern Black Sea, it is 700 mm to 800 mm; and to the east, it can be 1800 mm to 2500 mm. Over most of the Black Sea, northerly and northeasterly winds prevail, whereas southerly and southwesterly ones are typical in the south-east. Average monthly wind speed is maximal in January-February (7–8 m/sec) and minimal in June-July (4–6 m/sec). Constant maximum wind speed is observed in the western Black Sea, and the minimum wind speed is in the south-east.

During the Late Quaternary, the Black Sea was repeatedly isolated from the ocean due to varying climatic conditions (Yanko-Hombach 2007; Yanko-Hombach et al. 2007; 2011a,b; 2012a,b; 2013a,b). Geographical location and periodic connection of the Black Sea either with the Mediterranean or Caspian seas (Fig. 16.1) predetermined specific hydrogeological regimes in the basin, making it an excellent paleoenvironmental amplifier and a sensitive recorder of climatic events.
This chapter is focused on the description of the Late Pleistocene–Holocene environmental factors defining the Azov-Black Sea basin, and the identification of potential sample areas for seabed prehistoric site prospecting and landscape exploration on the Black Sea's northwestern continental shelf.

**Earth Sciences Data**

**Main sources of data**

Within the International Bathymetric Chart of the Mediterranean (IBCM) (www.ngdc.noaa.gov/mgg/ibcm/ibcm.html) there is a set of geological maps for the Black Sea (Sheet 5) with a scale of 1:1,000,000, published by the Charts Division of the Head Department of Navigation and Oceanography in Russia under the authority of the Intergovernmental Oceanographic Commission (IOC) of UNESCO. The set of maps with explanatory notes (Emelyanov et al. 2005) includes the basemap for a Geological/Geophysical series with Bouguer gravity anomalies [IBCM-G], seismicity [IBCM-S], unconsolidated bottom-surface sediments [IBCM-SED] (www.ngdc.noaa.gov/mgg/ibcm/ibcmsg.html), thickness of Plio-Quaternary sediments [IBCM-PQ] (www.ngdc.noaa.gov/mgg/ibcm/ibcmsg.html), and magnetic anomalies [IBCM-M].

The IBCM-SED map of the Black Sea is satisfactory for both the deep-sea area and for the western, northwestern, and northern areas of the shelf. The Anatolian shelf is characterized by numerous gaps, as no detailed data were available to the compilers. For the Anatolian slope, the results of thorough lithological investigations were conducted by IORAS (Shirshov Institute of Oceanology Russian Academy of Sciences), its Siberian Branch (SBIORAS), the Institute of Geology, NASU (National Academy of Science of Ukraine) (Shnyukov 1981; 1982; 1983; 1984, 1985; 1987), and M.V. Lomonosov Moscow State University (Kuprin & Sorokin 2007).

The Black Sea is surrounded by six countries: Ukraine, Russia, Georgia, Turkey, Romania and Bulgaria. Each country keeps the most recent earth science database for its own economic zone. Much of these data are not a public record.

The best source of most recent earth science data for the Ukrainian economic zone (Shchiptsov 1998) of the Black Sea and Sea of Azov is in a map series (each 594 X 841 mm) with a scale of 1:500,000. An overview of map locations (marked by “S”) is shown in Fig. 16.2.
Fig. 16.2 Overview geological map (1:1,000,000) of pre-Quaternary sediments with identification of mineral resources of the Ukrainian economic zone of the Black Sea, Shnyukov (2011). The map includes seven-sheet set (each 594 × 841 mm) 1:500,000 scale. For the overview of their location see ‘S’ on the map. The map, geological profile A–A3 and tectonic chart Б1–Б8, utilize color schemes based on standards that are related to the timescale. FGDC (2006).

The map series includes: pre-Quaternary and Quaternary geology of the coast, shelf, continental slope, and abyssal plain, supplemented by information on the distribution of mineral resources, including lists of mineral fields and their projected areas; lithological and geomorphological maps of recent, Quaternary, and pre-Quaternary sediments; and charted neotectonic movements. The maps are compiled by the state-owned enterprise Krymgeologiya [Industrial Geological Association Crimean Geology] (crimeageology.ru) and Prichernomor SRGE [Black Sea Area State Regional Geologic Enterprise] (www.pgrgp.com.ua). The explanatory notes and legend were complied by Kakaranza et al. (2007).

The maps are based on geological survey results at different scales (from 1:1,000,000 to 1:50,000). The survey’s primary goal was to identify promising areas for mineral exploration in the Azov-Black Sea basin. In total, 41 reports were written between 1972 and 2011.
(e.g. Podoplelov et al. 1973-1975; Sibirchenko et al. 1983; Ivanov 1987; Avrametz et al. 2007). For the full list of reports see Pasynkov (2013). The survey area was divided into quadrats, within each of which a substantial number of gravity cores and/or vibrocores up to 5 m in length were recovered on grids of $2 \times 2$ km and $0.5 \times 0.5$ km. In addition, each quadrat also contained about 25 mapping boreholes (Fig. 16.3).

The survey was supplemented by geophysical data and studied by a multidisciplinary team (Shnyukov 1981; 1982; 1983; 1984a,b; 1985; 1987; Shnyukov et al. 2011 among many others), providing a vast amount of lithological, paleontological, micropaleontological and palynological data from the sediment column, and radiocarbon dated in many cases (for dates see Appendix 1 and 2 in Balabanov 2007; Yanko-Hombach 2007a,b; Yanko-Hombach et al. 2013a,b).

Today, Prichernomor SRGE and Krymgeologiya are primary archives of geological and geophysical data. Other institutions possessing great amounts of earth science data include the Institute of Geological Sciences and Department of Marine Geology and Mineral Resources of the National Academy of Sciences of Ukraine; the Departments of Physical and Marine Geology, Engineering Geology and Hydrogeology, and Physical Geography, all at Odessa I.I. Mechnikov National University, Ukraine (onu.edu.ua/en/) (Konikov 2007; Shmuratko 2007; Shuisky 2007; Yanko-Hombach 2007a); the

Fig. 16.3 Chart of medium- (1:200,000) to large- (1:50,000) scale geological survey of the Ukrainian part of the northwestern shelf: (A) shelf is divided into quadrats; (B) an example of the quadrat L-36-XIV where cores are shown by black dots. Citations provided in text.
Evidence from around the world shows that wave energy and breaking surf are the principal forces that can destroy prehistoric sites, both during inundation and afterwards. Given the limited dimensions of the Black Sea basin, the wind fetch is limited, and the waves cannot be as large as in the open Mediterranean, and those affecting the Atlantic coast are larger still. The regions with the highest wave energy and the greatest threat of destruction of submerged archaeological sites (if any) include the southwestern part of the Black Sea and the Thracian shores of Turkey, especially the west side of Istanbul. The eastern part of the Black Sea is the least energetic in terms of wave power. In general, wave energy decreases along the coast from west to east. More information about annual wave energies (kWh/m) and breaking surf for different regions of the Black Sea can be found in Aydogan et al. (2013).

Geodynamic settings of the Black Sea

The Black Sea lies within the Anatolian sector of the Alpine-Himalayan orogenic system, located between the Eurasian plate to the north and the African-Arabian plates to the south (Fig. 16.4). Global plate models (DeMets et al. 1990; 1994) and recent space geodetic measurements (Smith et al. 1994; Reilinger et al. 1997) indicate that in the surrounding region, the northward-moving African and Arabian plates are colliding with the Eurasian plate. From this collision, the Anatolian block is moving westward with a rotation pole located approximately to the north of the Sinai Peninsula (Tari et al. 2000). The northward movement of the Arabian plate and westward escape of the Anatolian block along the North and East Anatolian faults have been accompanied by several episodes of extension and shortening since the Permian (Yilmaz 1997; Robertson et al. 2004) until now, as can be seen in seismic reflection data (McKenzie 1972; McClusky et al. 2000).

The crustal structure of the Black Sea reveals western and eastern deep basins (Fig. 16.4). The former is older than the latter; it was rifted with the dissection of a Late Jurassic to Early Cretaceous carbonate platform that had been established on the southern margin (Moesian Platform) of the northern supercontinent, Laurasia. The strong western Black Sea lithosphere apparently rifted for about 30 Myr, while rifting and spreading in the eastern Black Sea, characterized by a much weaker lithosphere, were completed within 8 Myr (Spadini et al. 1996).

The western Black Sea basin is underlain by oceanic to suboceanic crust and overlain by sediment cover with a thickness of about 19 km. The eastern Black Sea basin is underlain in the center by oceanic to suboceanic crust, surrounded by thinned continental crust approximately 10 km in thickness and overlain by sediment cover about 12 km thick (Nikishin et al. 2011). The basins are separated from each other by the Andrusov Ridge, formed from continental crust and overlain by sedimentary cover 5 km to 6 km thick (Tugolesov et al. 1985; Finetti et al. 1988; Belousov & Volovsky 1989; Robinson 1997; Nikishin et al. 2003). The compressional tectonic regime is still active in the eastern Black Sea region as can be seen from geological and geophysical evidence, including offshore seismic reflection profiles (Finetti et al. 1988), offshore morphology (Meisner et al. 1995), onshore geology and morphology (Okay & Sahinturk 1997 and references therein), and recent seismic activity (Neprochnov & Ross 1978; Barka & Reilinger 1997). The north–south motions of several mm/year are seen mostly in the eastern Black Sea, while in the Anatolian region, they are approximately 10 mm/year to 20 mm/year (Tari et al. 2000).

The isostatic anomaly and the isostatic residual anomaly along two transects in the western and eastern Black Sea basins, based on high-quality regional gravity data, show a clear discrepancy indicating a non-local isostatic compensation of the Black Sea lithosphere. The western basin appears to be in an overall upward state of flexure (undercompensated basin center and overcompensated basin flanks), and the eastern basin is
in a downward state of flexure (overcompensated basin center and undercompensated basin flanks). This suggests that the level of necking involved in the deformation is deep in the west but shallow in the east (Spadini et al. 1996).

**Vertical earth movements and rates of vertical coastal displacement**

Geodetic and tide-gauge observations are probably the best sources of measured vertical movements. They provide information describing secular movements of the Earth’s crust from north of the Scandinavian Peninsula to the Black Sea basin. The rate of recent crustal movements in this area is on average 2 mm/year to 4 mm/year. The maximum rate of uplift (~10 mm/year) occurs in the central regions of Fennoscandia (the Baltic Shield), in the region of Krivoy Rog (the Ukrainian Shield), and in the near-Carpathian region (Artyushkov & Mescherikov 2013).

The map by Blagovolin and Pobedonostsev (1973) shows the crustal movements of the northern Black Sea region. Vertical earth movements are clearly visible in parts of the Black Sea bottom. However, studies of local tectonics in the Black Sea region encounter serious problems due to: (1) a limited number of reliably dated marine terrace deposits and insufficient use of oxygen isotope (d18O) oceanic records in regional stratigraphic schemes; and (2) complicated tectonic pulses that, in addition to climatically-induced sea-level changes, contributed
Fig. 16.5 Chart of neotectonic movements of the northwestern shelf. Compiled by Pasynkov (2013), based on data of Morgunov (1981). (1) Area of intensive subsidence; (2) Area of weak subsidence; (3) Area of relative uplift in areas (1) and (2). Dotted red line = faults.

significantly to repeated erosion of older marine terraces by younger transgressions resulting in numerous gaps in the geological record. Shelf morphology is intricately linked to major climatic fluctuations over the past million years (Shmuratko 2001; 2003). The number of steps in the staircase of marine terraces corresponds to the number of Ice Age cycles, each with an average duration of about 100,000 years. The lower marine terraces are useful for identifying shorelines over the course of the postglacial transgression (Shmuratko 2007).

The areas of intensive and weak subsidence, as well as relative uplifting of the continental crust in the northwestern Black Sea area calculated from thicknesses of the Pliocene–Quaternary sediments and instrumental observations, is shown in Fig. 16.5.

The rates of subsidence on the northwestern shelf vary between 1 and 5 mm/year to the north-west of Odessa; 0.8 and 0.9 mm/year in areas of Stanislav and Ochakov; and 0.5 mm/year in the area of Belgorod-Dniestrovsky. The shelf area adjacent to the western coast of Crimea also experienced an intense subsidence. Especially vulnerable are the tectonically-fractured sedimentary coasts of the Taman Peninsula and the estuaries in southern Ukraine. The Taman Peninsula shows subsidence rates of 0.4 m/kyr to 1.6 m/kyr (Brückner et al. 2010).

The greater part of the Caucasian shelf edge is marked by a scarp of presumably tectonic origin (Glebov & She笠ting 2007, fig. 9). Depth at the shelf edge varies between 85 m and 115 m BSL due to different rates of tectonic movement.

Some areas of the continental slope experienced catastrophic subsidence due to seismotectonics from the movement of the Arabian plate, which generates strong earthquakes along the Caucasus and Anatolian coast. Supporting catastrophic subsidence is the wide range of elevations of the surfaces of Maikopian (Oligocene to Early Miocene) sediments on the upper continental slope. They vary between 1000 m and 1300 m BSL on the slope and 4000 m and 6000 m BSL in the deep basin.

The presence of lacustrine sediments around Karaburun on the southern shelf indicates an uplift of about 75 m along the northwestern coast of the Istanbul Peninsula during the Late Quaternary (Oktay et al. 2002).

Among other institutes, primary datasets on Black Sea geotectonics are held by the S.I. Subbotin Institute of Geophysics, National Academy of Sciences, Kiev, Ukraine (Starostenko et al. 2004), the Geological Faculty of Moscow State University (Nikishin et al. 2001; 2003; 2011), the Department of Marine Geology and Mineral Resources of the National Ukrainian Academy of Sciences (Shnyukov 1987), the Crimean Department of Ukrgeofizika (www.ukrgeofizika.kiev.ua/en/company.html), Kiev, Ukraine, Chernomorneftegaz [the State Joint Stock Company] (gas.crimea.ru), Simpheropol, Crimea, and the Institute of Marine Geology and Geocology, GeoEcoMar, Romania (www.geocomar.ro).

Solid geology of the Black Sea

Projects involving scientists from countries around the Black Sea as well as Canada, France, Germany, the USA and others, have been developed to study the solid geology of the Black Sea. It is difficult to indicate the main institutes holding extensive data on this subject.
So far, there is no single concept of foundation time, origin, and sediment infill for the Black Sea, although discussions have been ongoing since the end of the nineteenth century.

The geological history of the Black Sea basin can be divided into three major stages: (1) a pre-Middle Cretaceous time, prior to the commencement of the Black Sea opening; (2) a Middle–Late Cretaceous and/or Paleocene–Eocene phase of complex basin opening; and (3) a post-rift development in Oligocene–Quaternary times, accompanied by orogenic activity in the surrounding areas, such as the Greater Caucasus, Balkans, and Pontides. Any reconstruction of the regional kinematics of the Black Sea area must also take into account the mechanism responsible for the profound subsidence of the western Black Sea basin (16 km) since the Early Tertiary (Robb et al. 1998).

Information on the age and lithology of stratigraphic units in the Black Sea region comes from drilling and onshore geologic mapping. Sediments as old as the Late Miocene have been sampled in the center of the Black Sea at three Deep Sea Drilling Project (DSDP) sites, and sediments as old as the Late Jurassic have been recovered from industrial wells at the margins of the Black Sea (Zonenshain & Le Pichon 1986; Banks et al. 1997).

In the eastern Black Sea (Fig. 16.7), seismic stratigraphic horizons have been tied to well control at the edges of the basin using 2D and 3D industry seismic
data sets in order to estimate the ages and lithologies of sedimentary units (Shillington et al. 2008).

The oldest geological formations are represented by Late Jurassic through Late Cretaceous sedimentary rocks recovered by drilling at the margins of the Black Sea, as well as time-correlative onshore units. They comprise a variety of lithologies, notably including shallow-water carbonate rocks with significant volcanic material dated to the following intervals: Early Paleocene to Middle Eocene (65–45 Ma); Middle Eocene to the top of the Eocene (45–33.9 Ma); top of Eocene to Early Miocene (33.9–20.5 Ma); Early Miocene to Middle Miocene (base of Sarmatian) (20.5–13 Ma); Middle Miocene (base of Sarmatian) to Late Miocene (top of Sarmatian) (13–11 Ma); and Late Miocene (top of Sarmatian) to Pliocene (11–1.8 Ma). The latter are represented by sands and conglomerates of Pliocene age recovered by drilling onshore in Georgia and mapping in northeastern Turkey.
These units are typically non-marine and unlikely to be representative of lithologies in the basin center where chalk, siderite, clay and limestone were recovered by DSDP drilling. This interval also contains a thin unit comprising algal mats and peletal limestones, indicative of very shallow water depths. Although interpretations regarding the age and causes of these deposits are disputed, it appears that they correspond to a drop in sea level of over 2000 m, possibly related to the Messinian Salinity Crisis that affected the entire Mediterranean region (for references, see Shillington et al. 2008).

Pleistocene deposits have been recovered in many gravity cores and drilling; the former are also well studied in coastal outcrops. Lithologically, they are variable: clays, marls and occasional turbidites (for references, see Yanko-Hombach 2007a). High-resolution seismic and sonar images show primarily flat-lying, undisturbed sediments in the basin center, although the shallowest sediments often show disruption by gas (Ergün et al. 2002; Shnyukov & Yanko-Hombach 2009).

Bathymetry of the Black Sea

The most comprehensive source is the International Bathymetric Chart of the Mediterranean (IBCM), including the Black Sea (Sheet 5), which is also the basemap for a Geological/Geophysical series of maps (Fig. 16.8). Each of these chart series now has a published explanatory brochure available as a pdf file. All published maps, some 70 sheets at 1:1 million, and 7 sheets at 1:5 million, have now been scanned and are available digitally at different resolutions (Hall & Morelli 2007). The digitized marine contours are available on CD-ROM as part of the British Oceanographic Data Centre's

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**Fig. 16.8** Map of the Black Sea bottom relief, based on the International Bathymetric Chart of the Mediterranean Sea, reproduced with permission from IBCM (IBCM Project, Intergovernmental Oceanographic Commission, International Hydrographic Organization, and Head Department of Navigation and Oceanography, Russian Federation – Sheet 5).
Modern coastline

The Black Sea coastline is slightly indented. Together with nearshore bars and spits, it is 4725 km in length, based on measurements from the 1:100,000 scale topographic map. The total length of the Sea of Azov shoreline is 1860 km. The best source for determination of the coastline length and area of the Black Sea using Geographic Information System (GIS) methods and Landsat 7 satellite images can be found in Stanchev et al. (2011).

The northwestern coastline is characterized by lowland shores indented by bays, gulfs, inlets, and estuaries, most of which do not penetrate far inland. The northeastern coast is mainly high mountains except for the lowland area between the Panagia and Anap's'kyi capes on the coast of the Taman Peninsula. In the east, the vast Kolkhida lowland meets the sea. The southern coast is high and precipitous almost everywhere along the northern Tavr mountain system. It is interrupted by flat lowlands in the vicinity of the estuaries of the large rivers, such as the Kızılırmak, Yeşilırmak, Yenice, and Sakarya. The coast west of the Bosporus Strait is rather low and descends into lowlands. The Strandzhys'ke and Medenrudnits'ke high coasts begin at the Rezovs'ka Estuary and reach Burgas Bay. Then, there is a low coast
as far as the Varnensky Gulf. The capes in this area drop steeply toward the sea. From Cape Kaliakra to the great plain of the Danube Delta, the coast gradually descends. The coast noticeably rises to the east of Sevastopol Bay. The shores of the Kerch Peninsula are precipitous for almost the entire length, except for localities with barrier spits enclosing limans and lagoons. The shores of the Sea of Azov in the west, north, and east are mainly low, whereas in the south, they are steep. As in the Black Sea, the existence of limans and lagoons is a characteristic feature, especially in the northwestern part. The formation of large sand and shell spits, as well as barrier spits of different kinds is typical. They separate a number of shallow bays and limans from the sea. The largest of them, the Arabats'kaya Strelka, separates the shallow Syvash Gulf.

Among the largest bays and gulfs are: Odessa Bay and Karkinitsky Gulf on the northwestern coast; Novorossiysk Bay on the Caucasian coast; Sinop Bay, Bay of Samsun, Vaughn Bay, and the Gulf of Iğneada on the Turkish coast; and Burgas and Varna bays on the Bulgarian coast. Except for the Crimean Peninsula, which projects well out into the sea, there are no other peninsulas and large capes. The largest island is Zmeinyi (Snake) Island (also known as Fidonisi) with a total area of 0.17 km² located not far from the mouth of the Danube River. The main geographical locations are shown in Fig. 16.10.

The main source of information on the Black Sea wetlands is the Directory of Azov-Black Sea Coastal Wetlands (Marushevsky 2003). The largest Black Sea wetlands are found in the coastal lowlands of Romania, Russia, and Ukraine, where the massive catchments of the rivers Danube, Dniester, Dnieper, Don, and Kuban support river deltas. In contrast, the Black Sea wetlands of Bulgaria, Turkey, and Georgia tend to be much smaller and have much more limited catchments, reflecting the mountainous hinterlands of these countries. There are about 94 wetlands with a total area of 2,486,372 ha. Thirty-five Black Sea coastal wetlands, totalling 1,953,576 ha, are of international importance and are designated as Ramsar sites. The best source on the Black Sea wetlands (Fig. 16.11) can be found at blacksearegion.wetlands.org/Portals/9/BlackSea%20map.jpg.

The vast majority of limans and lagoons are present along the northwestern coastal zone (Fig. 16.12). The Black Sea limans are unique geographical and geological features. Study of their depositional structure can aid in establishing the most recent stages of the basin's geological history. The limans are unique because:
1) the thickness of Neoeuxinian and Holocene deposits greatly exceeds their thickness on the shelf, continental slope, and the deep sea; 2) the lithological structure of deposits is diverse, and based on lithological and faunistic composition the deposits are clearly stratified; 3) the major faunal complexes range from freshwater to polyhaline; and 4) limans belong to the first belt of avalanche sedimentation and represent a zone of geochemical barrier (Konikov & Pedan 2006). There is an extensive bibliography on aspects of Black Sea liman geology (Shnyukov 1984b; Zaitsev et al. 2006).

There are two major deltas in the northwestern Black Sea: the Danube and Dnieper deltas (Fig. 16.13). The former is larger and divided into Romanian and Ukrainian parts, 4400 km² and 1240 km², respectively. A large data set on the Romanian part of the Danube Delta is held by GeoEcoMar, Bucharest, Romania (www.geocomar.ro).

For the Ukrainian part of the delta, a large database is archived by the Ukrainian Scientific Center of Ecology of the Sea, Odessa, Ukraine (UkrSCES) (www.sea.gov.ua) (Berlinkov et al. 2006).

The Danube Delta is located between 44°24’ and 45°40’ N, and 28°14’ and 29°46’ E. It is marshy, contains a dense network of branches and lakes, and has three main arms: the St. George (I and II), the Sulina, and the Chilia or Kilia (Fig. 16.13a).

The delta is divided into two parts: ancient riverine and young marine. It includes a series of lakes, e.g. the Yalpug, Kugurluy, Katlabukh, Kitay, as well as a large area between the St. George I and Chilia branches, the Dranov Peninsula, and the Razelm-Sinoe lake-lagoon complex. All these areas are closely connected hydrologically, and in this sense they represent a single territory with similar climate, lithology, soil, and vegetation characteristics.
Sedimentological features of the delta are determined by its geographic position. Bounded on the north and south by the Budzhak and Dobrudza plateaux, it has a limited amount of sediment coming from those areas. Runoff from the plateau slopes into the lakes comes mainly during snowmelt or heavy rains via small rivers. Total annual runoff from these small rivers does not exceed 2000 million m$^3$, which is less than 0.1% of the average annual flow in the upper reaches of the Danube Delta.

The evolution of the Danube Delta during the Holocene and corresponding coastline changes are shown in Fig. 16.13c. The Dniester Delta is much smaller than the Danube Delta. It formed at the outlet of the river into the Dnieper-Bugsky liman (Fig. 16.13b). Sediment discharged by the river accumulates in the inner part of the delta forming spits, bars, and alluvial islands. The width of the floodplain terrace in the lower part between western Cairo and Kherson ranges from 2.5–3 km to 5 km. It widens below Kherson, at the outlet into the Dnieper Estuary, where the river splits into numerous arms forming a modern delta. Filling with sediment to varying degrees, the delta extends seaward. Deltaic sediments are represented by an alternation of sub-aerial and subaqueous facies. In the mouth of delta arms the crescent-shaped sand bars were formed. They collapse during floods with development of new bars seaward.

Sedimentological features of the delta are determined by its geographic position. Bounded on the north and south by the Budzhak and Dobrudza plateaux, it has a limited amount of sediment coming from those areas. Runoff from the plateau slopes into the lakes comes mainly during snowmelt or heavy rains via small rivers. Total annual runoff from these small rivers does not exceed 2000 million m$^3$, which is less than 0.1% of the average annual flow in the upper reaches of the Danube Delta. The evolution of the Danube Delta during the Holocene and corresponding coastline changes are shown in Fig. 16.13c.

The Dniester Delta is much smaller than the Danube Delta. It formed at the outlet of the river into the Dnieper-Bugsky liman (Fig. 16.13b). Sediment discharged by the river accumulates in the inner part of the delta forming spits, bars, and alluvial islands. The width of the floodplain terrace in the lower part between western Cairo and Kherson ranges from 2.5–3 km to 5 km. It widens below Kherson, at the outlet into the Dnieper Estuary, where the river splits into numerous arms forming a modern delta. Filling with sediment to varying degrees, the delta extends seaward. Deltaic sediments are represented by an alternation of sub-aerial and subaqueous facies. In the mouth of delta arms the crescent-shaped sand bars were formed. They collapse during floods with development of new bars seaward.

The morphometry and bottom topography of the Black Sea are important oceanographic features because they determine major characteristics of thermohaline structure and water circulation. Configuration of the shoreline, width of the shelf and continental slope, as well as the shape of the bottom profile, the presence of valleys, canyons, ridges, and depressions, influence the distribution of water masses, the direction and speed of currents, and the position and intensity of the topogenic eddies and coastal upwelling.

The Black Sea exhibits the standard oceanic provinces of continental shelf, slope, and abyssal plain (Fig. 16.14). The continental shelf is part of the submerged coastal land and covers 25% of the sea area; isobath 200 m is commonly taken as the shelf boundary of the Global Ocean. The northwestern shelf extends outward 220 km, occupies 16% of the sea area (68,390 km$^2$) and 0.7% of the water volume (3555 km$^3$) between capes Chersonesus and Kaliakra (nos. 4 and 33 in Fig. 16.10). In the flattened and gently sloping part of the shelf adjacent to the shore, depths are 30 m to 40 m, and the bottom slope is 1° to 2°. Its steepness increases toward the shelf break to 10° to 12°. Against the flat plain of the shelf, several large, shallow paleoriver valleys are visible in Fig. 16.14, separated by low underwater hills (Ivanov & Belokopytov 2013). The bottom relief is largely smooth due to sediment discharge and distribution provided by
major lowland European rivers, such as the Danube, Dnieper, Dniester, and the Southern Bug, that together discharge 56.8 million tonnes of sediments annually (Panin & Jipa 1997). There are no known expressions of active tectonic movements that would influence the ancient shoreline positions and deposition of sediments in any appreciable way.

The other, less extensive shelf areas of the Black Sea include: the coastal zone of Bulgaria and western Turkey from Cape Kaliakra to the city of Ereğli (shelf width up to 50 km); the Kerch-Taman shelf (shelf width up to 50 km); the central Anatolian coast from Cape Kerempe to the city of Giresun (shelf width up to 35 km); the southern Crimean coast between capes Chersonesus and Ai-Todor (shelf width up to 30 km); and the Gudauta Bank in the vicinity of Ochamchira town (shelf width up to 20 km) (Ivanov & Belokopytov 2013).

Narrow shelves with widths of several kilometers are located along the Caucasian and Anatolian coasts, as well as along the southern Crimean coast from Yalta to Cape
Meganom. Their bottom slope is considerably steeper compared to broader shelves, ranging from 5–6° to 30°. The shelf break lies at depths from 100 m to 200 m, the slope is 1° to 2°. The depth of shelf break is close to 100 m while in areas with a broader shelf it can deepen to over 200 m. Most gas seepages occur along the shelf breaks in areas of fault scarps on the sea floor, shallow sub-surface faults, small grabens, and conical depressions or domes of gently sloping anticlinal rises (Fig. 16.14) (Shnyukov & Yanko-Hombach 2009).

The predominantly flat bottom of the Sea of Azov descends gradually to the depression at its center. At the bottom, there are a few positive relief forms, the largest of them being the Pischana Bank.

The continental slope descends down to 1600 m to 1900 m of water depth with a considerable gradient from...
11° to 13°, sometimes reaching 38° in the regions along the southern Crimean and Turkish coasts. The surface of the continental slope is complicated with blocks of the Earth’s crust that often give it a graduated profile, and most of it reveals underwater canyons of different origin. They can begin in the coastal zone at a depth of 10 m to 15 m and reach a lower depth of 1600 m. The canyons are the most important route for the transfer of sedimentary material from the coast to the abyssal depression of the Black Sea (Fig. 16.14). In the deepest part of the canyons, at depths of 1600 m to 1900 m, the transferred sedimentary material forms big cones. Individual cones can coalesce to form the continental sub-slope. Thus, the morphogenesis of the slope is directly linked to selective erosion and denudation of rocks with different physical and mechanical properties. Erosive and denudation activities in the canyons caused the emergence of huge underwater amphitheaters forming deepwater fans and plumes of terrigenous sediments on the footslope.

The abyssal plain is bounded by isobath ~2000 m and occupies about 35% of the total sea area. It is a flat accumulative plain with a slight slope to the south. The bottom of the abyssal basin is characterized by hilly relief; slope angles vary from 0° to 1°. According to echo-sounding surveys, significantly large features of submarine relief are absent (Fig. 16.14). Deposits covering the abyssal plain form eleven material-genetic types. Six types are shallow, and five are deep water. Between all types of deposits, there is a continuous transfer conditioned by gradual change in their grain-size and material composition. The mean rate of accumulation at the bottom of the central abyssal depression is 30 mm/kyr to 40 mm/kyr.

There are about seventy mud volcanoes located mainly in the northern part of the western Black Sea, Sorokin trough, Tuapsinskaya trough, Shatskiy arch, and Kerch downfold (the area south of the Kerch Peninsula) (Fig. 16.14; Shnyukov et al. 2010, 2013).

Today, the Bosphorus Strait (Fig. 16.14) is the only passage for exchange of water and organisms between the Black Sea and Sea of Marmara. This zigzagging strait is about 35 km in length, 0.7 km to 3.5 km in width, and 35.8 m deep on average, with a few elongate potholes (about 110 m in depth each) on the bottom. It possesses two sills, one in the north at a water depth of 59 m and one in the south at a water depth of 34 m, each located about 3 km from the corresponding entrance to the strait. The two directions of waterflow within the strait overlap each other: the northward underflow (inflow) from the Sea of Marmara has an average salinity of 38 psu and a velocity of 5 cm/sec to 15 cm/sec, and the southward overflow (outflow) from the Black Sea has an average salinity of 18 psu and a velocity of 10 cm/sec to 30 cm/sec. Due to the sills, the interface between the two flow directions rises from −50 m at the northern end to −20 m at the southern end. The underflow is initiated by the difference in water density between the Black Sea and the Sea of Marmara; the pressure gradient pushes against the Black Sea and acts to power the underflow. The outflow is initiated by two main factors: (1) the 30 cm elevation of the Black Sea surface above that of the Sea of Marmara, which, in turn, is 5 cm to 27 cm above the level of the northern Aegean Sea, and (2) the positive balance of the Black Sea, where precipitation (575 km³/year) exceeds evaporation (350 km³/year), producing a discharge of about 600 km³ of brackish water annually (Yanko-Hombach et al. 2007).

The Kerch Strait connects the Black Sea with the Sea of Azov (Fig. 16.14) and is 45 km long, 4.5 km wide, and up to 6 m deep. The shallowness of the strait results in reduced water exchange between the two basins, which is five to ten times smaller than that of the Bosphorus.

Coastal geomorpho-dynamics, erosion, accumulation

Coastal zones of the Black Sea and Sea of Azov can be considered in two parts: surface and underwater. The former is referred to as the coastal zone, the latter as the submarine sea slope. Both parts are closely connected in their origin and develop simultaneously under the influence of common types and sources of energy. The main processes of coastal zone development are morphodynamic (topographic features) and lithodynamic (deposits).

Coastal geomorpho-dynamics are defined largely by the wave and current regimes that give the coastal zone
a variety of borders and widths. The width of the coastal zone depends upon the gradient of the coast and continental slope. The form of coastal cross-section can be highly diversified: first, the shape of the profile depends upon the physical and mechanical properties of the rock and deposits, as well as their resistance to abrasion. In general, the coastline is significantly wider in plains areas and narrower in mountainous ones.

The coastal zone in the Black Sea and Sea of Azov basins is prone to constant change along its entire length. In some places, rough seas destroy the coast, while in other areas, they smooth the coast, creating new land areas and changing continental terrace topography. The mechanical energy of sea waves and wave currents encourages a variety of morphodynamic and lithodynamic processes leading to various topographic features, as well as the amount, composition, and mobility of the deposits and diversity of coastal zone components. In the Azov-Black Sea coastal zone, abrasion generally prevails over accumulation. Capes and coastal promontories that project into the sea undergo significant destruction; rough seas and landswells constantly seek to level the coastline. General coastline erosion and degradation is slower in estuary areas with intense solid suspended sediment inflow; in the estuaries of big rivers (Danube, Dnipro, Psou, Bzyb, Inguri, Rioni, Chorokh), drifts and sand spits can grow up to 10 m during the river flood season. The formation of coastal landforms depends on the strength of the rocks. Igneous and some particularly strong metamorphic rocks are only slightly susceptible to the damaging effects of waves. Coastal slopes composed of such rocks are practically unaffected by waves, and thicknesses of generated sediment are negligible. In contrast, weakly cemented sedimentary rocks clearly show traces of wave and chemical weathering by sea water. Here, on the coastal sea edge, niches, caves, cracks, and ‘stone boilers’ are formed. Upon further exposure to waves, the overhanging niches collapse leading to cliff formation or abrasion ledges. Cliffs exposing non-cohesive sediments of loam, clay and sand are destroyed very quickly, and retreat at a rate of several meters per year. Different morpho-dynamic types of coastline are shown in Fig. 16.15.

The best source and digital archive for morphostructural zoning of the shelf is presently at the Department of Marine Geology and Mineral Resources of the National Academy of Sciences of Ukraine, as well as the Department of Physical Geography and Department of Engineering Geology and Hydrogeology at Odessa I.I. Mechnikov National University. Detailed characteristics of leading geological processes in the Black Sea coastal zone of Ukraine can be found in Pasynkov (2013).

**Landscape regions of the northwestern Black Sea shelf**

Based on distribution of statistical parameters describing water depth, Holocene sediment thickness, and percentage of silt and clay within the sediments, the following recent landscape areas have been delineated (Larchenkov & Kadurin 2007a,b; 2011): erosional coastal offshore slope; Danube pro-delta and paleovalley; depressions of river paleovalleys; terraces of the inner shelf; central shelf plain; and the outer shelf plain (Fig. 16.16).

Similar methods were used to reconstruct paleo-lands-}
Fig. 16.15  (A) abrasional coast of Cape Tarkhankut; (B) abrasional-avalanched coast of the western Crimea; (C) landslide coast of the Kerch Peninsula; (D) abrasional-creek coast of Cape Fiolent, Crimea; (E) abrasional-landslide southern coast of Crimea with shore protection structures; (F) the ‘Eltigen’ neostratotype exposed on abrasional-avalanched coast of tectonically elevated terrace, Kerch Peninsula, Crimea. From Yanko-Hombach et al. (2012b). This neostratotype contains the most complete and well-preserved marine sequence of the Karangatian transgression.
the Mediterranean and Caspian seas, are largely based on the study of outcrops exposed on tectonically elevated terraces of the Kerch and Taman peninsulas, Crimea, and Caucasus (Yanko 1990; Yanko-Hombach & Motnenko 2011; 2012).

The Russian subdivision of the Quaternary system includes the Eopleistocene (1.8–0.78 Ma), the Neopleistocene (0.78–0.01 Ma), and the Holocene (0.01–0.0 Ma) (Zhamoida 2004). The boundary between the Eopleistocene and Neopleistocene coincides with the Brunhes-Matuyama reversal (i.e. 780 kyr). This boundary is readily traced in the Black Sea at the bottom of the Chaudian horizon (Fig. 16.6).

The classical Quaternary stratification in the Black Sea region includes (from oldest to youngest) the Eopleistocene (Gurian), Neopleistocene (Chaudian, Drevneuxinian [Old Euxinian], Uzunlarian, Karangatian, Novoeuxinian], and the Holocene (Drevnechernomorian and Novochernomorian) beds (Fig. 16.6). The Chaudian, Old Euxinian, and Karangatian beds were first proposed by Andrusov (1918). They are exposed in stratotypes (except for the Old Euxinian which does not have a stratotype) in tectonically elevated terraces. These classic names were incorporated into the stratigraphic scale of Arkhangel'sky and Strakhov (1938), who also described the Uzunlarian horizon and identified its stratotype. Previously designated Euxinian sediments (Andrusov 1918) bearing brackish, Caspian-type fauna were divided by Arkhangel'sky and Strakhov (1938) into Drevneevskian [Old Euxinian] and Novoevksinian [Neoeuxinian] beds. The former presently lies above sea level on tectonically elevated terraces, occasionally on the sea bottom, and it contains Didacna pontocaspia. The latter is distributed below sea level only and contains the mollusks Dreissena and Monodacna. The initial stratigraphic framework was later improved based on mollusks (Nevesskaya 1965), foraminifera and ostracoda (Yanko 1990; Yanko & Gramova 1990).

The absolute age of the sediments in the stratotypes varies depending on the method used, e.g., 230 U/Th,
thermoluminescence, etc. For example, based on a few dozen thermoluminescence dates and 12 magnetic-polarity datum planes, Zubakov (1988) has assigned a numerical age to the horizons as follows: Chaudian = 1.1 Ma–600 ka; Uzunlarian = 580–300 ka; Karangatian = 300–50 ka. Tchepalyga (1984), Yanko (1990), Yanko-Hombach and Motnenko (2011; 2012), Yanko-Hombach et al. (2013b) suggest slightly different numerical ages for the Eo- and Neopleistocene geological sequences in the Black Sea.

Ecostratigraphy of the Pleistocene and Holocene sediments is based on taxonomy and spatial distribution of bivalve molluscs (Nevesskaya 1965), gastropods (Il'ina 1966), benthic foraminifera and ostracoda (Yanko 1990; Yanko & Gramova 1990; Yanko-Hombach 2007a,b) as well as palynological data supplemented by 14C datings (e.g. Appendix 1 and 2 in Yanko-Hombach et al. 2007 and Balabanov 2007, respectively).

The paleosalinity in this chapter is described as follows: freshwater (<0.5 psu), semi-fresh (0.5–5 psu), brackish (>5–12 psu), semi-marine (>12–18 psu), and marine (>18–26 psu). The use of ‘psu’ (Practical Salinity Units) instead of former ‰ is explained in Yanko-Hombach et al. (2013a).

The Gurian (or Gurian Chauda) beds correspond to Marine Isotope Stage (MIS) 23–22 (Fig. 16.6). Its stratotype has a thickness exceeding 1000 m located in Tsvermaghala Mountain, Georgia. Except for the Bulgarian outer shelf (Malovitsky 1979), no Gurian deposits were discovered in other places of the Black Sea bottom. The Gurian basin was semi-freshwater with sea level somewhere around 100 m to 150 m BSL.

The Chaudian beds (780–500 ka) correspond to MIS 21–15 (Fig. 16.6). Its stratotype has a thickness exceeding 1000 m located in Tsermaghala Mountain, Georgia. Except for the Bulgarian outer shelf (Malovitsky 1979), no Gurian deposits were discovered in other places of the Black Sea bottom. The Gurian basin was semi-freshwater with sea level somewhere around 100 m to 150 m BSL.

On the Bulgarian outer shelf, the thickness is 13.7 m (Kuprin et al. 1984), while on the Caucasian shelf, it reaches 17 m (Yanko 1989). The level of the Chaudian basin varies according to different authors: either below the present one (Fedorov 1978; Yanina 2012); similar to the present one (Inozemtsev 2013); or above the present one at 15 m to 18 m above sea level (ASL) (Fedorov 1997) or 20 m ASL (Chepalyga 1997; Svitoch et al. 1998).

Chaudian beds are overlain by Drevnejuxinian [Old Euxinian] beds corresponding to MIS 14–8 (Fig. 16.6). The latter are widely distributed in tectonically elevated terraces on the Kerch and Taman peninsulas. However, their most complete geological section with a thickness of 12.8 m was recovered on the northwestern Black Sea shelf. The roof of the Drevnejuxinian beds (here and elsewhere calculated as a sum of water depth and sediment thickness) lies between isobaths 32.4 m and 55.7 m BSL on the northwestern shelf and between 18 m and 23 m BSL on the northeastern shelf, where most likely it is tectonically elevated. Sea level was likely around 30 m BSL (Yanko 1989), while Svitoch et al. (1998) consider it was slightly above the present one. The hydrological regime of the basin was as follows. In the Early Drevnejuxinian, the basin was connected to the Caspian Sea, which discharged its water into the Drevnejuxinian basin. The salinity did not exceed 7 psu. In the Late Drevnejuxinian, it increased to ~18 psu due to connection with the Mediterranean Sea (Fig. 16.6) (Yanko 1989; 1990).

The Drevnejuxinian beds with an erosional unconformity corresponding to the Paleouzunlarian regression are overlain by Uzunlarian beds (Fig. 16.6) corresponding to MIS 7. Uzunlarian beds are rarely found on the shelf but usually in coastal outcrops. The hydrological regime of the Uzunlarian basins was as follows: at the beginning, the basin was connected to the Caspian Sea, and its salinity did not exceed 8 psu to 10 psu; then it increased, probably
to 18 psu, and dropped again to 12 psu to 13 psu by the end of the Uzunlarian. According to Fedorov (1997), the level of the Late Uzunlarian basin was about 10 m ASL. Svitoch et al. (1998) consider it was close to recent sea level, while Yanina (2012) believes it exceeded slightly the present one.

The Late Pleistocene is represented by the Karangatian (MIS 5), Tarkhankutian (MIS 4), and Neoeuxinian (MIS 3–2) beds (Fig. 16.6). Detailed descriptions of the Karangatian neorstratotype ‘Eltigen’ (shown in Fig. 16.15c) are provided in Motnenko (1990) and Yanko et al. (1990). New 14C dates for the Eltigen can be found in Nicholas et al. (2008). The Karangatian beds, which accumulated during the Riss-Würm (Mikulino Interstadial interglacial — a term proposed by A. I. Moskvitin in 1947), are widely distributed in both coastal areas and the sea bottom. Their stratigraphic position is clearly identified by the highest numbers of Mediterranean molluscan and foraminiferan species that do not live in the Black Sea today and are the warmest dwelling species among all others in Pleistocene and Holocene. The basin was connected to the Mediterranean Sea, and its salinity reached 30 psu. Sea level might have been around 7 m BSL (Fedorov 1978; 1997; Svitoch et al. 1998; Yanina 2012) or between –5 m and –10 m (Chepalyga 1997).

Tarkhankutian beds (Fig. 16.6) were first reported by Nevesskaya and Nevessky (1961) from Karkinitsky Bay, northwestern shelf, at water depths of 30 m to 35 m as sediments containing a mixture of Caspian and Mediterranean mollusks. Later, they were discovered in other places on the Black Sea coast, e.g. the Colchis Plain (Georgia) where they are overlain by sub-aerial peats dated ca. 31 ka BP at a sampling depth of 60 m (Dzhanelidze & Mikadze 1975). Popov and Zubakov (1975) and Popov (1983) recognized similar sediments as Surozhian. Svitoch et al. (1998) considered Tarkhankutian and Surozhian sediments as coeval, with an age of 40 ka to 25 ka BP. The Tarkhankutian transgression at 31,330 ± 719 ka BP (Chepalyga 2002a,b) brought Mediterranean water and organisms into the Black Sea and increased salinity to about 8 psu to 11 psu (Nevesskaya 1965; Yanko 1989; 1990; Yanko-Hombach 2007a). Submerged accumulative coastal bars of synchronous age are located at water depths of 22 m to 30 m on the Ukrainian (Chepalyga et al. 1989; Chepalyga 2002a,b) and Romanian (Caraiam et al. 1986) shelf, indicating that Tarkhankutian sea level was about 25 m BSL (Chepalyga 1997; Yanina 2012), 30 m BSL (Yanko-Hombach 2007a), or 45 m BSL (Fedorov 1997). Temporally, the Tarkhankutian sediments correspond to Unit 3 (Çağatay 2003) in the Sea of Marmara. This unit contains some marine mollusks and benthic foraminifera indicating a weak Mediterranean marine incursion during the early part of MIS 3. Interestingly, no similar sediments have yet been found in the Bosporus. Instead, they have been recovered in Izmit Bay and the Sakarya valley (Meriç et al. 1995; Yanko-Hombach et al. 2004).

The Tarkhankutian is overlain by Neoeuxinian beds (Yanko-Hombach 2007a). The latter correspond to MIS 2 and were formed during and soon after the Last Glacial Maximum (LGM). They can be divided into Lower and Upper Neoeuxinian beds. The Lower Neoeuxinian beds (Fig. 16.17) were formed between 27 ka BP and 17 ka BP and can be found below isobath ca. 100 m (Yanko-Hombach 2007a; Mudie et al. 2014; Yanko-Hombach et al. 2013a,b). Lower Neoeuxinian sediments are distributed everywhere in the Black Sea below isobath 100 m (Kvasov 1975; Fedorov 1977; 1978; 1988; Shcherbakov et al. 1978; Abashin et al. 1982; Shcherbakov 1983; Shnyukov 1985; Svitoch et al. 1998; Kuprin 2002; Kuprin & Sorokin 2007; Yanko-Hombach 2007a; Yanko-Hombach et al. 2013b, fig. 3). Lithologically, they are represented by alternations of gray silt and gray striped clays enriched with hydrotriolite, sand (minor), and shells of D. rostriformis distincta. The level of the Early Neoeuxinian lake was ca. 100 m BSL (Kuprin & Sorokin 2007; Yanko-Hombach 2007a; Yanko-Hombach et al. 2011a,b; 2013a). Their recovered thickness ranges between 5 cm and 80 cm (Fig. 16.17).

The Lower Neoeuxinian beds are often overlapped by sub-aerial loams and further on by aquatic sediments with ostracoda Candona, Candoniella, and foraminifera A. novoaeuxinica. This change indicates the transformation of the bottom from an erosional to subaquatic accumulative phase at the beginning of the Late Neoeuxinian transgression (Gozhik 1984b; Shnyukov 1985).
The Upper Neoeuxinian beds overlie the Lower Neoeuxinian beds, covering the Black Sea floor almost everywhere below isobath 39 m BSL on the northwestern shelf (Fig. 16.17) (Larchenkov & Kadurin 2011; Mudie et al. 2014; Yanko-Hombach et al. 2013a,b), 30 m BSL on the Bulgarian (Filipova-Marinova 2007), Crimean (Shnyukov 1985), and Caucasian shelves (Balabanov et al. 1981; Yanko & Gramova 1990), and 18 m BSL on the Turkish shelf. In some places (e.g. the western part of the Golitsin Uplift located at the mouth of Karkinitsky Bay), they are exposed on the sea floor (Tkachenko et al. 1970; Ishchenko 1974; Tkachenko 1974; Yanko 1974; 1975; 1989).

Their thickness can reach 25 m (Put’ 1981). Lithologically, Upper Neoeuxinian beds on the shelf are rather monotonous and represented by light gray sandy coquina and/or bluish gray stiff clays that fill pre-Neoeuxinian depressions and paleoriver valleys (e.g. Arkhangel’sky & Strakhov 1938; Neveskaya 1965; Semenenko & Kovalykh 1973; Ostrovsky et al. 1977; Malovitsky 1979; Balabanov et al. 1981; Yanko 1982; Gozhik 1984a,b,d; Shnyukov 1985; Fedorov 1988; Gozhik et al. 1987; Yanko 1989; 1990; Yanko & Gramova 1990; Glebov et al. 1996). The stiff clay has a massive structure, high density (about 2.7 g/cm³), and low water content. The interstitial water salinity is 7 psu (Konikov 2007). Mollusks are dominated by D. polymorpha and D. rostriformis on the inner and outer shelf, respectively. A paleosalinity for the Late Neoeuxinian lake was about 5 psu in the shallow areas; it could have reached 7 psu, which is typical of interstitial salinity (Manheim & Chan 1974), and even 11 psu to 12 psu (Neveskaya 1965; Mudie et al. 2001; 2011; Marret et al. 2009; Yanko-Hombach et al. 2013a,b) in deeper parts of the basin. Despite a relatively high salinity, Mediterranean species are absent, and Caspian immigrants are abundant.

Holocene sediments cover the Black Sea shelf almost everywhere (Fig. 16.18). Their thickness and lithological characteristics vary significantly in different parts of the basin due to morphodynamics of the shelf (e.g. width, bottom relief), as well as tectonic activities, amount of sediment discharged into the Black Sea by rivers, as well as subsidence, compaction, and erosion of sediments during regressive stages — all greater on the narrow shelves compared to the wider northwestern shelf (Malovitsky 1979; Kuprin et al. 1980; Shcherbakov 1983; Shnyukov 1984b; Kuprin et al. 1985; Krystev et al. 1990; Yanko & Gramova 1990; Devdariani et al. 1992; Sorokin et al. 1998; Aksu et al. 2002; Shimkus 2005; Glebov & Shel’ting 2007; Konikov 2007; Kuprin & Sorokin 2007; Yanko-Hombach 2007a,b).

The spatial distribution of Holocene deposit thicknesses is non-uniform, with a distribution skewed toward smaller values; they can be subdivided into three general groupings: bathymetric rises, paleovalleys, and mid- to outer shelf deposits (Fig. 16.18). The normal thickness for each region corresponds to a mean value with a standard deviation, with deviations in thickness as positive or negative anomalies (Table 16.1).

The rate of sedimentation on the shelf varies significantly from place to place. On the northern shelf, it is 50 cm/kyr to 200 cm/kyr in the Danube pro-delta and Dneprovsky gutter; on the western Crimean shelf, it is 30 cm/kyr to 50 cm/kyr; and in the central part of the shelf at the 50 m isobath it is 10 cm/kyr to 20 cm/kyr. On the outer northwestern shelf at water depths of 70 m to 100 m, the rate of sedimentation decreases to 5–15 cm/kyr, remaining higher in its eastern part. In the upper part of the continental slope at water depths of 100 m to 400 m, the rate of sedimentation does not exceed 10 cm/kyr, falling to 5 cm/kyr and lower in the areas of intensive sediment erosion (Fig. 16.19).
Sedimentologically, the Romanian shelf (between the Danube Delta and Cape Kaliakra) is a continuation of the Ukrainian shelf. Since solid sediments of the Danube River are carried out to the south by currents, this part of the shelf is sediment starved. Therefore, the rate of sedimentation on most of the Romanian shelf (excluding submerged parts of the Danube valley and narrow coastal zone) is 5 cm/kyr to 10 cm/kyr. At depths of 60 m to 70 m along the east–west boundary between Romania and Bulgaria, the 5-m-thick Holocene sediments indicate a sedimentation rate of 60 cm/kyr to 70 cm/kyr.

On the Bulgarian shelf, the highest sedimentation rates occur between capes Kaliakra and Emine. Between 30 m and 60 m depth, the rates of sedimentation vary between 200 cm/kyr to 350 cm/kyr. They decrease to 10–15 cm/kyr and less toward the shelf break and coastal slope. A high rate of sedimentation is typical for the southern shelf between Cape Emine and the Turkish–Bulgarian border. At Burgas Bay, it varies between 10 cm/kyr to 20 cm/kyr and 150 cm/kyr to 200 cm/kyr, reaching its maximum in buried river valleys. At depths of 40 m to 60 m, the sedimentation rate is about 100 cm/kyr. Toward the outer shelf, the rates decrease to 10–25 cm/kyr. In the upper continental slope, similar to the Romanian shelf, the distribution of sedimentation rates has a mosaic character due to a large number of underwater river valleys where erosion takes place. If no erosion of sediments occurred, sedimentation would be 15 cm/kyr to 20 cm/kyr.

The Turkish western shelf (west of the Bosporus) is covered by low thickness Holocene sediments. Judging from the limited number of studied cores and seismoacoustic data, the rate of sedimentation does not exceed 30 cm/kyr to 40 cm/kyr. At the shelf break and upper continental slope, it decreases to 10–20 cm/kyr.

On the Crimean shelf (between the Khersones Peninsula and Kerc Strait), the sedimentation rates vary significantly. The lowest values are typical of underwater coastal slopes, where a bench thinly covered by sandy-pebbly sediments is present almost everywhere. At depths between 15 m and 30 m, sedimentation reaches 250–400 cm/kyr. It decreases toward the middle part of the shelf to 40–60 cm/kyr and to 10–15 cm/kyr toward the shelf break at depths of 80 m to 100 m. The lowest sedimentation rates are in the eastern Crimean shelf.

The Caucasian shelf between Anapa and Batumi is characterized by reduced thickness of Quaternary sediments in general, and Holocene sediments in particular. In most of the shallow coastal zone, at depths of 5 m to 10 m, the bedrock is exposed. In the area of Novorossiysk-Gelendzhik, Adler-Pitsunda, and Poti-Kobuleti, sedimentation rate increases to 100–400 cm/kyr. It decreases to 50–100 cm/kyr and 10–40 cm/kyr in the central part of the shelf and at the shelf break, respectively. On the continental slope, the rates of sedimentation change from

### Table 16.1 Average and Anomalous Thicknesses of Upper Pleistocene–Holocene Sediments in Different Geomorphological Areas of the Shelf. From Larchenkov & Kadurin (2011).

<table>
<thead>
<tr>
<th>Region</th>
<th>Sediment thickness (m)</th>
<th>Negative anomaly</th>
<th>Positive anomaly</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bathymetric rise</td>
<td>0.4–3.1</td>
<td>&lt;0.4</td>
<td>&gt;3.1</td>
</tr>
<tr>
<td>Paleovalley</td>
<td>3.2–7.9</td>
<td>&lt;3.2</td>
<td>&gt;7.9</td>
</tr>
<tr>
<td>Middle to outer shelf</td>
<td>0.3–1.8</td>
<td>&lt;0.3</td>
<td>&gt;1.8</td>
</tr>
</tbody>
</table>

**Fig. 16.19** Chart of distribution of sedimentation rate of Neoeuxinian and Holocene sediments on the northwestern shelf calculated on the basis of the $^{14}$C data, cm/kyr: (1) <10; (2) 10–20; (3) 20–30; (4) 30–40; (5) 40–50; (6) >50; (7) black dots = sampling stations where sedimentation rate was calculated. Shnyukov (1985).
0 cm/kyr to 20–30 cm/kyr due to its strong dissection and erosion by turbidity currents.

The best data sources on the spatial distribution of sedimentation rate and thicknesses of Late Pleistocene/Holocene sediments, including maps, can be found at the Department of Marine Geology and Mineral Resources of the National Academy of Sciences of Ukraine, the Department of Lithology and Marine Geology at Moscow State University, Prichernomor SRGE, and the Department of Engineering Geology and Hydrogeology, Odessa I.I. Mechnikov National University, Ukraine.

Post-LGM Climate, Sea Level, and Paleoshorelines

During the LGM between 27 ka BP and 17 ka BP, ice sheets spread over much of North America and northern Asia. Over Hudson Bay in Canada, the ice was 4 km thick. In Europe, ice sheets extended over much of the UK and as far south as Germany and Poland (Yanko-Hombach et al. 2012b). The spread of ice profoundly impacted the Earth’s climate, causing drought, desertification, and a dramatic drop in sea levels in basins connected to the ocean. About 21 ka BP, the world’s oceans were 120 m lower than today. So too was the Mediterranean Sea, which was connected to the Atlantic Ocean through the Gibraltar Strait. The north Adriatic Sea and the north Aegean Sea were completely exposed. On the edges of glacial areas, the entire East European Platform was covered with tundra-steppe vegetation (Yanko-Hombach et al. 2012b).

Early Neoeuxinian palynological diagrams are dominated by Artemisia, Chenopodiaceae, Adonis, and Thalictrum and are similar to those of the dry pine forests of Romania (Komarov et al. 1979; Pop 1957), the pine/birch forests and xerophytic steppe in southern Ukraine and Moldova (Artyushchenko et al. 1972; Kyrvel et al. 1976), and the steppe and forest-steppe on the Balkan Peninsula (Bottema 1974). All indicate a cold and dry climate (Nikonov & Pakhomov 1993). By implication, such a climate should lead to a dramatic decrease in river discharge into the Early Neoeuxinian lake, causing in turn a dramatic drawdown of the lake level.

The Early Neoeuxinian lake was isolated from both the Mediterranean and Caspian seas. It was semi-fresh or brackish, aerobic (Degens & Ross 1974), and heavily populated by organisms, in particular those with calcareous shells (CaCO₃ in sediments: ~50%), e.g. mollusks, ostracoda, and on a much smaller scale, by foraminifera. During the Early Neoeuxinian, a large portion of the present shelf above present isobath 100 m was exposed and eroded. The northwestern shelf was downcut some 40 m to the basement by the Pre-Danube, Pre-Dnieper, and Pre-Dniester rivers, and covered by sub-aerial loams (e.g. Shcherbakov et al. 1978; Shcherbakov 1983; Inozemtsev et al. 1984; Fedorov 1988). River mouths were relocated 80 km to 100 km seaward (Gozhik 1984b; Shnyukov 1985), where they possessed poorly developed deltas and opened directly into the canyons on the continental slope (Fig. 16.20).

The river valleys and canyons were filled with thick (22–40 m) alluvial sediments (Kuprin & Sorokin 2007) of the stratigraphic unit Ant age (22,800–16,900 BP) containing 27 freshwater and 14 brackish water shallow ostracods dominated by Cyprideis littoralis and Ilyocypris bradyi (Gozhik 1984c). Direct palynology evidence from high-resolution marine cores (Mudie et al. 2002; 2007) suggests a marshy and mosquito-infested shoreline subject to periodic river flooding. This might have provided good hunting and fishing but poor conditions for settled farming because of brackish water and soils prone to salinization and waterlogging; these problems still limit coastal farming today.

Late Neoeuxinian palynological diagrams are dominated by Quercus, Carpinus, Ulmus, Salix, and Betula, with reduced concentration of Pinus and grass (Komarov et al. 1979; Kravadze & Dzeiranshvili 1989). They are similar to the Late Glacial diagrams of the Balkan Peninsula (Bozilova 1973; 1975) and the Prichernomor horizon (Veklich & Sirenko 1976) formed before 10.5 ka BP (Ivanova 1966). The climate warmed during Late Neoeuxinian times, as indicated by the replacement of pine by broad-leaved forests. In the warming climate of about 16 ka BP, a massive water discharge, most likely from the Caspian Sea via the Manych Spillway.
GEOLOGICAL AND GEOMORPHOLOGICAL FACTORS AND MARINE CONDITIONS OF THE AZOV-BLACK SEA BASIN

Fig. 16.20 Schematic geomorphological map of the northwestern shelf during the Early Neoeuxinian: (1) paleovalleys and large alluvial plains; (2) gently sloping hills formed by Late Pleistocene marine and continental landforms; (3) gently sloping piedmont uplands; (4) steep banks and seabed folded bedrock; (5) large relict accumulative landforms formed by sand; (6) paleoriver bed; (7) underwater debris cone; (8) coastline of 18 ka BP; (9) shelf edge: (a) outer shelf, (b) continental slope. Modified after Shnyukov (2002). Reconstructions were based on results of the 1:200,000 marine geological survey outlined in Fig. 16.3, as well as numerous onshore and offshore drillings, and geophysical profiles which enabled the discovery that deltas and valleys of the Pliocene Don, Dnieper, and other rivers migrated across the northwestern coast and adjacent shelf. Two buried paleovalleys of the Paleo-Kanchak and Pre-Dnieper were discovered as fragments of the migration of the ancient valley of the Dnieper River.

(Chepalyga 2007) increased the level of the Late Neoeuxinian lake to ca. –20 m. The latter must have overflowed, pouring its excess semi-fresh-to-brackish water into the Sea of Marmara and from there into the Mediterranean (Fig. 16.21).

The Late Neoeuxinian lake was aerobic and heavily populated by organisms with carbonate shells. Late Neoeuxinian sediments seem to be partially synchronous with the upper part of Unit 2 (Çağatay 2003) of the Sea of Marmara sediment column. The level of the Late Neoeuxinian lake was much higher (about –20 m) than that of the Sea of Marmara (about –85 m), and by implication, the Late Neoeuxinian lake discharged its waters into the Sea of Marmara. The Bosporus continued to be a semi-fresh lake and might have served as a channel for southward water discharge from the Neoeuxinian lake. However, this discharge could have occurred through the Izmit Gulf–Sakarya valley, as indicated by the presence of fresh/brackish facies with an age of 14.6 ka BP in borehole KS2 (Kerey et al. 2004; Yanko-Hombach et al. 2004; 2007).

Upper Neoeuxinian beds, often with erosional unconformity, are overlapped by Bugazian beds containing the first Mediterranean immigrants. Bugazian sediments are widely distributed below the 17 m isobath. Their thickness increases from 0.03 m to 0.20 m on the slopes of submerged river valleys to 2.5 m on their bottoms. Bugazian palynological diagrams are characterized by a sharp decrease in grassy elements (e.g. wormwood, goosefoot) and conifers (Pinus, Picea, Juniperus). Instead, broadleaf Quercus, Corylus, Ulmus, Betula, and even beech become dominant, indicating moderate climate conditions typical of the Boreal Ecozone (Komarov et al. 1979). A similar palynological diagram of the deep sediments of the Black Sea and peats of the Ril Massif in Bulgaria have $^{14}$C ages of 10,737±315 BP (Shimkus et al. 1977) and 10,035±65 BP (Bozilova 1973).

A summary of the palynological data from lakes over a wide area west and south of the Black Sea shows...
that oak-pistacio (Quercus–Pistacia) forests were present over most of the region by 10 ka BP, although local desert-steppe vegetation persisted until about 7 ka BP in the south-east, from Lake Van (Eastern Turkey) to the Caspian Sea (Mudie et al. 2002). These forests indicate the early establishment of mesic climatic conditions characterized by >600 mm/year of precipitation in excess of evapotranspiration, as is presently found in most of central and western Europe (Hiscott et al. 2007a).

Since the Bugazian, a series of low amplitude transgressive and regressive phases is clearly manifested on the inner shelf of the Black Sea (Fig. 16.6). Due to the low amplitude of the regressive phases, they cannot be traced in cores recovered from a depth of more than 50 m, thus giving the impression of a gradual increase in sea level and salinity. In general, the dynamic nature of the recent Black Sea level changes is a result of the joint influence of isostatic, eustatic, and tectonic processes, and to a lesser degree, anthropogenic factors.

The paleolandscapes of the northwestern shelf since the LGM are shown in Fig. 16.22.

There are three main scenarios for the post-LGM development of the Black Sea: gradual, oscillating, and catastrophic (also called rapid or prominent); they are discussed elsewhere (e.g. Yanko-Hombach 2007b; Yanko-Hombach et al. 2011a,b; see also Chapter 17, pages 485–487).

According to the gradual scenario, the Late Neoeuxinian lake was gradually transformed into a marine basin and re-colonized by Mediterranean organisms in the course of the postglacial transgression (Arkhangelskii & Strakhov 1938; Neveskaya 1965; Il’ina 1966; Aksu et al. 2002; Hiscott et al. 2007a,b). The latter started at ca. 9.5 ka BP when a two-way interchange was established via the Bosporus (Stanley & Blanpied 1980).

According to the oscillating scenario (e.g. Ostrovsky et al. 1977; Fedorov 1978; 1982; Balabanov et al. 1981; Tchepalyga 1984; Shnyukov 1985; Balabanov & Izmailov 1988; Yanko 1990; Yanko & Gramova 1990; Shilik 1997; Svitoch et al. 1998; Chepalyga 2002a,b; Balabanov 2007; Konikov 2007; Yanko-Hombach 2007a,b; Mudie et al. 2014; Yanko-Hombach et al. 2011a,b; 2013a), the transformation of the Late Neoeuxinian lake into a marine basin was gradual but fluctuating.

According to the catastrophic (Ryan et al. 1997; Ryan 2007) / rapid (Lericolais et al. 2007a,b; 2011) / prominent (Nicholas et al. 2011) scenario, the transformation of the Neoeuxinian lake into a marine basin was not gradual but catastrophically rapid. At a rate in excess of 50 km$^3$ per day, Mediterranean salt water funneled through the narrow Bosporus and hit the Black Sea at 200 times the force of Niagara Falls, thereby sharply increasing salinity, refilling the Neoeuxinian lake with a rate of rise of level of 15 cm/day over two years, and replacing freshwater biota with marine organisms (Ryan et al. 1997). Later, instead of a single inundation, two lowstands (~120 m at 13.4–11 ka BP; and ~95 m at 10–8.4 ka BP) and two catastrophic floods (sea-level rise from ~120 m to ~30 m at 11–10 ka BP; and from ~95 m to ~30 m at 8.4 ka BP) were proposed (Ryan et al. 2003). The second of these two major transgressions was labeled the Great Flood as described in the Bible. The initial Flood Hypothesis was based on evidence from seven short (about 1.25 m each), low-resolution sediment cores and 350 km of seismic profiles collected within a fairly restricted area of the Black Sea’s northwestern shelf at water depths between 49 m and 140 m during a single mission in 1993 (for their location see Fig. 16.16).

Our latest multidisciplinary study of geological material recovered in areas of the northwestern, northeastern, and southwestern Black Sea shelf (Yanko-Hombach et al. 2013b; Mudie et al. 2014) confirm our previous data.

1 The level of the Late Neoeuxinian lake prior to the Early Holocene Mediterranean transgression stood around 40 m BSL, but not 100 m BSL or more, as suggested by advocates of the catastrophic/rapid/prominent flooding scenario.

2 Microfossil data examined from multiple shelf sites show that at all times, the Neoeuxinian lake was brackish with a salinity of about 7 psu prior to the Initial Marine Inflow (IMI) and Mediterranean transgression.

3 By 8.9 ka BP, the outer Black Sea shelf was already submerged by the Mediterranean transgression. An increase in salinity took place over 3600 years, with the rate of marine incursion estimated on the order of 0.05 cm/year to 1.7 cm/year.
Fig. 16.22 Paleogeographic scheme of the Pontic Lowland and northwestern Black Sea shelf: (A) LGM (27–25 ka BP); (B) in the Late Neoeuxinian (15.5–15 ka BP); (C) at the Younger Dryas (11–10 ka BP); (D) just before the Holocene transgression (10–9.4 ka BP); (E) at the beginning of the Holocene transgression (9.4 ka BP); (F) during the Kalamitian (4 ka BP). Legend in Fig. 16.22(a) is also for Figs. 16.22(b–e). Yanko-Hombach et al. (2011a).
The combined data from sedimentological characteristics and microfossil salinity evidence establish that the Holocene marine transgression was of a gradual, progressive but oscillating nature. Lately, Badertscher et al. (2011) attempted to reconstruct the time and number of water intrusions into the Black Sea from the Mediterranean and Caspian seas using speleothems from Sofular Cave. According to their data, there were twelve and seven intrusions of Mediterranean and Caspian water respectively into the Black Sea. If their data are correct, we must then see an alternation of molluscan and foraminiferal assemblages in the coastal outcrops. Instead, we see that connections between the Mediterranean and Black seas occurred six times, while connections between the Caspian and Black seas occurred four times since the Brunhes-Matuyama reversal (i.e. the last 780 kyr), and in most cases, these connections did not occur synchronously with those of Badertscher et al. (2011) (Fig. 16.6). Namely, the Early Chaudian, Early Old Euxinian, Early Uzunarian, and Late Neoeuxinian basins were connected to the Caspian Sea. The Karadjenizian, Late Old Euxinian, Middle–Late Uzunarian, Karangatian, Tarkhankutian, and Chernomorian (Old and New Black Sea) basins were connected to the Mediterranean Sea. In all stages, these connections had an oscillating character. As such, the speleothem-based conclusions for reconstructing water intrusions from the Mediterranean and Caspian into the Black Sea must be used with great caution (Yanko-Hombach & Motnenko 2011; 2012).

Evidence for Submerged Terrestrial Landforms and Ecology

Submerged terrestrial features are represented in the Black Sea by a variety of landforms. From the Danube Delta to Odessa Bay and further to Karkinitsky Bay, they are erosive-accumulative and accumulative sandy bars, spits, and barriers (Fig. 16.23a), with the exception of Snake Island, which was formed tectonically (Fig. 16.23b). Underwater ridges, caves, and tunnels are widely distributed within the South Crimea littoral zone.
Fig. 16.23 (A) Dzharylgachsky Gulf (for location see no. 38 in Fig. 16.10): (1) emerged barrier spit; (2) barrier beach; (3) lagoon; (4) barrier island; (5) dunes; (6) barrier shallows; (7) emerging spit; (B) Snake Island; (C) Cape Aya with underwater caves and freshwater springs; (D) Adalary Rocks mentioned in Homer’s *The Odyssey*; (E) Protrusions formed by Sarmatian limestones in the form of islands called ‘Stones-Ships’, off the Kerch-Taman Peninsula; (F) Sonar profile across the southern underwater extension of Parpachsky Peak formed by folded Maikopian deposits with uplifted ‘Stones-Ships’; (G) Seismic profile across a barchan dune inside a trough. Lericolas et al. (2007b). Reproduced with permission.

(Fig. 16.23c). Erratic stacks, separated by erosion from the slopes of the Crimean Mountains, form underwater rock massifs sometimes rising above sea level in the form of islands, such as the Adalary Rocks mentioned in Homer’s *The Odyssey* (Fig. 16.23d). A strong illustration of tectonic activity can be seen on the southeastern coast of Crimea, expressed as intense mudflows and landslide-crumbling landforms onshore, and canyons and landslides offshore. Protrusions in the form of islands can be seen off the coast of the Kerch-Taman Peninsula near the crest of the submarine continuation of Parpachsky Full (Figs. 16.23e,f).

A number of underwater shorelines were digitally reconstructed based on relief trend analysis, statistical analysis of recent sea bottom relief, and GIS modeling using Kriging methods of interpolation within grid areas of $100 \times 100$ (Larchenkov & Kadurin 2006). One shoreline corresponding to the LGM is especially well...
pronounced. Ballard et al. (2000), Algan (2003), Algan et al. (2007) and Lericolais et al. (2007a,b; 2010; 2011) described a submerged coastline with wave-cut terraces and coastal paleodunes (Fig. 16.23g) at various depths ranging from 90 m on the Romanian shelf to 155 m on the Turkish shelf near Sinop, in support of rapid transgression at the beginning of the Holocene. However, paleodune studies of Badyukova (2010) in the Caspian–Black Sea corridor established the following principles: (1) there are no opportunities for dunes to persist on the
sea bottom in any transgressive scenario for the Black Sea corridor; (2) transgressive parasequences involve the accumulation of lagoonal and marine deposits during rising relative sea level and landward migration of a coastline over coastal plain deposits. Our core data are in full agreement with Badyukova’s conclusion, and we find no lithological signs of drowned windblown dunes described by the above-mentioned authors (Yanko-Hombach et al. 2013b).

Potential for Archaeological Site Survival

The Pontic steppes are extremely rich in archaeological finds, which have been systematically studied by several generations of Soviet, Russian, and Ukrainian archaeologists (Dolukhanov 1979; Kremenetski 1991; Shilik 1997; Smyntyna 1999; Dergachev & Dolukhanov 2007; Dolukhanov & Shilik 2007; Stanko 2007). During the Upper Paleolithic, between 32,000 and 10,000 years ago, a diverse array of hunter-gatherer societies occupied the Pontic–Caspian steppes (Soffer & Praslov 1993; Sinitsyn 2003; Smyntyna 2004; Stanko 2007). The southern fringe of the Eurasian forest zone was, at that time, a periglacial steppe known as the famous ‘mammoth steppe’, where Ice-Age hunters built hide-covered huts over frameworks of mammoth tusks at places such as Mezhirich and Mezin on the middle Dnieper drainage and at Kostenki II on the middle Don. Reindeer hunters lived in the Carpathian foothills at Molodova V (Grigor’eva 1980). In what is now the steppe zone north of the Black and Caspian seas, bison and horse hunters lived at sites such as Amvrosievka and Anetovka II (Stanko 2007). The hunting patterns and tool kits of these hunters in the south were quite different from those of the colder mammoth steppe to the north. The area of the southern steppes was also much larger than it is now (Anthony 2007).

Over recent decades, geoarchaeological studies in the Black Sea region carried out within the framework of multidisciplinary international networking projects, such as IGCP 521 “Black Sea–Mediterranean corridor during the last 30 ky: Sea level change and human adaptation” (2005–2010), INQUA 0501 “Caspian-Black Sea-Mediterranean Corridor during last 30 ka: Sea level change and human adaptive strategies” (2005–2011), and IGCP 610 “From the Caspian to Mediterranean: Environmental Change and Human Response during the Quaternary” (Yanko-Hombach et al. 2012a,b). These geoarchaeological projects collected substantial material and focused on the possible influence of environmental change on human adaptive strategies. These investigations leave no doubt that, since early prehistoric times, subsistence and social dynamics of human groups in that area were directly affected by changes in climate, vegetation, and sea-level fluctuations, with commensurate shrinking and expansion of littoral areas, and had considerable impact on the settlement patterns of the prehistoric societies (Stanko 2007). Most researchers agree that during the LGM, the level of the Black Sea was at least 100 m BSL, and the area of the shelf above isobath 100 m was dry land. Geological and palynological data show it was an exposed steppe region (Yanko-Hombach et al. 2011a,b; 2013a) that could be exploited extensively by hunter-gatherers migrating in from west to east, and from northwest to south (Stanko 2007). This hypothesis is supported by (1) the similarity in artifacts from several Upper Paleolithic archaeological sites located today across the northwestern Black Sea in Romania, Ukraine, and Crimea (Fig. 16.24) — namely Liubymovka in the Lower Dnieper area, Ukraine, and Movileni and Lespezi in Romania, Mitoc-Malul Galben and Dobrudja in Romania, and Siuren 1 in Crimea, Molodova in the Middle Dniester region and Vishenke 2 in Crimea (Otte et al. 1997; Demidenko 2000-2001; Chabai 2007) and (2) the finding of several flint tools within boreholes taken in various places on the northwestern shelf (Stanko 2007).

However, no submerged prehistoric sites have been discovered so far, except one ceramic plate from –90 m off Varna, and photographs of boulders at –90 m depth off Sinop that, according to Ballard (2001), have a Neolithic age of over 8000 years. However, underwater artifacts and shipwrecks recovered to date from this region are of historical date (Ward & Ballard 2004; Ward & Horlings 2008).

There are some submerged archaeological sites of much younger age at water depths shallower than 10 m (Fig. 16.9), such as the Greco-Roman ruins near the Sea of
Possible migration routes of Upper Paleolithic populations ca. 25 ka BP based on the similarity in artifacts from several Upper Paleolithic archaeological sites located today across the northwestern Black Sea in Romania, Ukraine, and Crimea, and described in Otte et al. (1997), Demidenko (2000-2001) and Chabai (2007).

Potential Areas for Future Work

At the LGM (27–17 ka BP), when the level of the Early Neoeuxinian lake was at least 100 m BSL (Figs. 16.20; 16.22a), on the northwestern coast, we would expect Upper Paleolithic sites to be located within the deep valleys of small rivers. These valleys were flooded in the course of the Late Neoeuxinian transgression (17–10 ka BP); however, they are well expressed geomorphologically and can be easily traced on the present Black Sea shelf. This topographic information can be used to search for submerged Upper Paleolithic sites on the shelf, thereby helping to locate evidence for the transition among ancient human groups from hunting large herd animals to small non-gregarious species. The beginning of the Mediterranean transgression occurred around 9.5 ka BP. Both the transgression and faunal migration occurred over the course of six transgressive-regressive stages. Mesolithic sites continued to be located along river valleys; they bear some evidence of the transition from hunting to gathering of edible plants. No signs of catastrophic flooding of the Black Sea in the Early Holocene have been found (Yanko-Hombach et al. 2011a).

One should bear in mind that searching for prehistoric archaeological objects, especially of Paleolithic age, will be difficult because the artifacts are small flints. Some approaches should be implemented: (1) to define the main principles of distribution of already known terrestrial archaeological sites; (2) to show their distribution on the coast with a focus on geomorphological features of the study area; (3) to define conformities in their terrestrial distribution.

Geoarchaeological modeling of the northwestern Black Sea region over the last 25 kyr with respect to paleoenvironment and settlement pattern mapping as an instrument for submerged prehistoric site prospecting was presented by Yanko-Hombach et al. at SPLASHCOS meetings in Berlin (2011) and by Yanko at Szczecin (2013); it was further developed by Kadurin and Kiosak (2013). GIS-aided mathematical modeling has indicated the most favorable areas on the northwestern shelf to search for submerged archaeological sites of Upper Paleolithic age based on the distribution of terrestrial sites and their locations along submerged valleys of small rivers as shown on geomorphological maps (such as Fig. 16.20, 16.22a for the Upper Paleolithic and Figs. 16.22b-f for younger ages).

Conclusion

The northwestern shelf of the Black Sea seems to be the most suitable area to search for submerged archaeological sites. It is the widest part of the shelf and the best studied geologically and geomorphologically. The finding of several flint tools retrieved from boreholes in various...
places on the northwestern shelf by Stanko (2007) is encouraging, and

“...archaeological surveys targeted at final Paleolithic and Mesolithic sites on the northwest Black Sea shelf and along the submerged river valleys might be deemed promising...[they] might solve a number of major problems related to the character and chronology of the submergence, migrations, and the interrelationship between prehistoric groups of the Balkans, Central Europe, and Crimea (Stanko 2007:374).”

Recent archaeological studies in Turkey, Georgia, and Bulgaria, together with investigations of karstic caves on the Crimean Peninsula in Ukraine allow us to assume that steep, as well as abrupt and rocky Black Sea coasts, also deserve attention with respect to searches for submerged prehistoric sites, especially of Lower and Middle Paleolithic times, and of the Mesolithic period.

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In this paper, we have transliterated Cyrillic letters into the Latin alphabet according to the BGN/PCGN Romanization system for Russian used by Oxford University Press. Exceptions are the names of authors, which we have left in their own preferred transliterations, as well as geographical names as presented most commonly in the majority of English papers.

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