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Lateglacial and Holocene sea level changes in semi-enclosed seas of North Eurasia: examples from the contrasting Black and White Seas

Pavel A. Kaplin, Andrei O. Selivanov*

Geography Department, Lomonosov Moscow State University, Moscow 119992, Russian Federation

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Abstract

A comparison of the Black and White Seas, which differ in their tectonic, glacial and climatic history but which share a strong dependence upon limited water exchange with the world ocean, represents an opportunity for the identification of major factors controlling sea level changes during the Lateglacial and Holocene and for the correlation of these changes. Existing data were critically analyzed and compared with the results of geological, geomorphological and palaeohydrological studies obtained by the present authors during the past two decades.

We conclude that glacioeustatic processes played a major role in relative sea level changes on most coasts of both areas. However, along several coastlines, other factors overwhelm glacioeustasy during some time intervals. In the Black Sea, water level rose from its minimum position, −100–120 m, at 18–17 ka BP, to −20–30 m at nearly 9 ka BP. In the White Sea, the decreasing trend in relative sea level is well illustrated on the Kola Peninsula and in Karelia, subject to glacioisostatic emergence. A drastic sea level fall from +15 to −25 m occurred with the drainage of glacial lakes in the eastern White Sea (12.5–9.5 ka BP).

The Black and White Sea histories changed drastically in the early Holocene or in the beginning of the middle Holocene (9.5–7.5 ka BP) due to the intrusion of water from the Mediterranean and the Barents seas, respectively. During this period, the White Sea developed under the strong influence of the formation of "ice shelves" and "dead ice" blocks, retreating glaciers, as well as of glacioisostatic and related processes. The Black Sea history, however, was determined by water exchange with the Mediterranean via the shallow Dardanelles and Bosporus straits (outflow from the Black Sea 10–9.5 ka BP and inflow from 9–7.5 ka BP according to various data), and, partially, by river discharge variations caused by climatic changes on the Russian Plain. The hypothesis of a catastrophic sea level rise from −120–150 to −15–20 m nearly 7550 calendar years BP is not supported by our data. Water intrusion from the Mediterranean was fast but not catastrophic.

In the Black Sea, periods of high sea levels after the intrusion of Mediterranean waters are dated from four sedimentary complexes, Vityazevian, Kalamitian, Dzhemetian and Nymphaean, from nearly 7.5, 7–6, 5.5–4.5 and 2.2–1.7 ka BP, respectively. A fluctuating pattern of sea level change was established in the White Sea after the drainage of proglacial lakes and intrusion of ocean waters at the end of the early Holocene (nearly 8.5–8.2 ka BP). Major periods of sea level rise in the White Sea are dated from the late Boreal–early Atlantic (8.5–7.5 ka BP), late Atlantic (6.5–5.2 ka BP), middle Subboreal (4.5–4 ka BP) and middle Subatlantic (1.8–1.5 ka BP). Fluctuations of relative sea level during the middle and late Holocene were possibly on the order of several meters (from +2–3 to −2–3 m in the Black Sea and from +3–5 to −2–3 m in the White

* Corresponding author. Fax: +7-95-9328836.
E-mail address: selivano@postman.ru (A.O. Selivanov).

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Sea). Lower estimates of regressive stages are principally derived from archaeological data on ancient settlements in tectonically submerging deltaic areas and cannot be regarded as reliable.

Palaeohydrological analysis does not indicate that intensive (15–25 m or greater) sea level fluctuations were present in the Black Sea or in the White Sea during the middle and late Holocene. Instead, such analysis provides independent evidence to support the argument that significant differences in water level between the Black Sea and the Mediterranean could not be maintained for an extended period of time.

Keywords: Transgression; Regression; Palaeohydrology; Archaeological settlements; Pleistocene; Holocene; Black Sea; Sea of Azov; White Sea

1. Introduction

Ancient sea coasts serve as unique reference sources for the reconstruction of past sea level positions, and therefore for the determination of the main factors controlling sea level changes over time (Kaplin and Selivanov, 1999). Among the most well-studied seas in Russia and the former USSR are the Black and White Seas (Fig. 1). The Black Sea and the White Sea are both semi-enclosed and are connected with the world ocean by shallow straits. Both seas underwent a drastic change in their volumes and salinity in the Early Holocene with the intrusion of water from the Mediterranean and the Barents Sea, respectively. However, the northwestern European coast of Russia, where the White Sea is situated, developed during this period under the strong influence of retreating glaciation and resulting glacioisostatic and other processes. In contrast, Black Sea history was determined by water exchange with the Mediterranean via the shallow and narrow Dardanelles (75 m) and Bosporus ($\sim$ 45 m) straits and by river discharge variations caused by climatic changes on the Russian Plain.

The Black Sea is an extensive body of water that has limited water exchange with the Marmara and Mediterranean seas. Under current conditions, it would take over 2600 years to totally exchange its water via the Dardanelles and the Bosporus Strait. Water in the Bosporus Strait is strongly stratified, with the outflow of freshened water (17–22‰) in the upper layer and the inflow of saline (34‰) below. This circulation governs the specific hydrological conditions and the unique biological environment of the sea, including that of the coastal zone. The medium salinity of surface waters, equal to 17–19‰, gives rise to complex mollusk fauna, which is a mixture of brackish and marine species. Endemic mollusk species serve as a principal basis for Quaternary stratigraphy of the region and, in many areas, constitute material suitable for radiocarbon dating.

On the other hand, coastal morphology generally precludes the formation of an adequate database for the Pleistocene and Holocene sea level history. Intensive neotectonic deformation and the relatively limited occurrence of typical morphosedimentary complexes of ancient shorelines prevent us from tracing along the coastline. An analysis of cores drilled at various coastal locations along the northern shores of the Black Sea (Fig. 2) forms the basis of a reconstruction of relative sea level changes.

The key issue in the sea level regime of the Black Sea in the Quaternary time frame is the shallowness of the Bosporus sill. Due to the low sill depth, the Black Sea repeatedly became an enclosed water basin (like the Caspian Sea) during the Pleistocene glacioeustatic regressions. However, a strong possibility exists that the altitude of the Bosporus bottom varied significantly during the Pleistocene because of tectonic movements, sedimentation, and erosion (Avenarius, 1979; Shcherbakov, 1983). This could add complexity to the pattern of sea level changes in the Black Sea. The Sea of Azov is actually an extensive gulf of the Black Sea with a mean depth of 7 m and a maximum depth of 15 m. The Sea of Azov largely depends on water exchange with the Black Sea via the shallow Kerch Strait and on water discharge from the Don River.

The coasts of the White Sea (Fig. 1) are characterized by broad variability in their morphology and recent history. This variability results primarily from the differences in the Late Pleistocene glacial history. The coasts of the Kola Peninsula and Karelia were covered by the thick Fennoscandian continental ice sheet and represent an area of intensive glacial erosion and isostatic emergence whereas the east coast devel-

Fig. 1. General scheme of Eastern Europe showing the Black and White Seas and inflowing rivers.

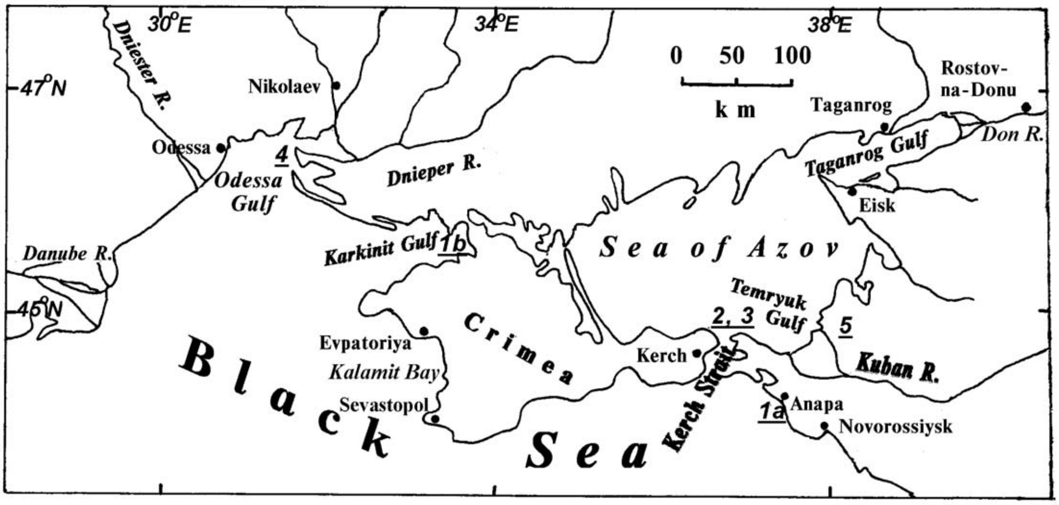

Fig. 2. General scheme of the Northern Black Sea. Areas of detailed sea level studies: (1a) the Caucasian coast near Anapa; (1b) Karkinit Gulf (Nevesskii, 1967); (2) the Kerch Strait and Caucasian coast (Fedorov, 1985); (3) Kerch-Taman area (Arslanov et al., 1982); (4) Odessa Gulf and adjacent areas (Voskoboinikov et al., 1982); (5) Kuban River delta.

oped in a marginal glacial area. The mean depth of the White Sea is nearly 67 m and the maximum depth exceeds 350 m. The salinity varies from 24‰ to 34.5‰ in different parts of the sea. The Gorlo (“Throat”) connecting the White Sea with the Barents Sea with a minimum depth of 75 m has much larger dimensions than the Bosporus.

Climate-driven fluctuations in the warming influence of the northeastern Atlantic current on the western sector of Russia’s European Arctic coast resulted in the successive migrations of Boreal malacofauna species. These species serve as palaeogeographic and stratigraphic markers in the area, and also provide reliable material for radiometric dating. A

Table 1
Correlation of Lateglacial and Holocene shorelines on the Kola Peninsula and in adjacent areas

Assemblages of malacofauna (Lavrova, 1969)	Regional shorelines in Fennoscandia (Tanner, 1930)	Elevation of shorelines in Kola Peninsula, m above MSL (Koshechkin, 1979)	Tentative correlation with climatic periods	Radiocarbon ages, uncal. years BP
Lateglacial	h	88–97, up to 120	Allerod	
Lateglacial	g (“main”)	80–100	Younger Dryas	
Portlandia 1	f	50–80	Preboreal	
Portlandia 2	e	25–40	Preboreal	10,030 ± 130
Littorina	d5–d1	16–40	Preboreal	9490 ± 100; 8980 ± 180
Folas (transgressive stage)	d	45–75, up to 90	Early Boreal	8890 ± 180; 8300 ± 140
Folas (regressive stage)	c4–c1	0–10 (?)	Late Boreal	7890 ± 150
Tapes 1	c	24–29	Early and Middle Atlantic	6790 ± 60
Tapes 2	a9–a7	20–25, up to 55	Middle and Late Atlantic	6400 ± 110; 5800 ± 105; 5000 ± 120
Trivia 1	a4	15–18, up to 27	Middle Subboreal	4170 ± 70; 4100 ± 70
Trivia 2	a3	10–15, up to 20	Late Subboreal	3590 ± 200; 3490 ± 200; 3090 ± 140
Ostrea 1	a2	5–8, up to 14	Early Subatlantic	1950 ± 150; 1800 ± 100; 1720 ± 80
Ostrea 2	a1	2,8–6	Late Subatlantic	730 ± 60

general stratigraphic scheme for the area in the Lateglacial and Postglacial periods was developed by Lavrova (1969) on the basis of the changes in mollusk assemblages. The successive basins were correlated with the principal shorelines (a–h) distinguished by Tanner (1930) on Fennoscandian coasts, presumably in the Finnmark region (Table 1).

This paper summarizes results achieved during the past two decades by many researchers, including the present authors. Many C-14 determinations have been obtained during the past decades and extensive studies of both coastal and bottom sediments and morphology were carried out. Some of the results are described in detail in other publications by the present authors (Kaplin and Selivanov, 1999; Kaplin et al., 1993, in press(a,b); Selivanov, 1996a,b; Svitoch et al., 1998).

2. Lateglacial and early Holocene sea levels

2.1. The Black Sea

In the traditional stratigraphic scheme of Arkhangelskii and Strakhov (1938), the period of the last continental glaciation in the Black Sea basin occurred in the Novoeuxinian (New Euxinian) regressive epoch, which was usually correlated with the Grimaldian epoch in the Mediterranean. The lowest sea level position during that period is estimated at −80 m (Balabanov and Izmailov, 1988), −80–90 m (Fedorov, 1978), or −100–110 m (Ostrovskii et al., 1977; Shimkus et al., 1980; Skiba et al., 1975). Shcherbakov (1983) found coastal pebbles and shells at −80–100 m on the shelf of South Crimea. Balabanov and Izmailov (1988) distinguished sediments of the maximum regressive period in bottom cores as Antian layers and dated them from 22–16 ka BP (Note: Unless otherwise noted, raw carbon-14 ages before reservoir corrections have been used throughout this text).

Novoeuxinian sediments fill deep erosional channels in the sediments of the previous regressive period (Fedorov, 1978; Shcherbakov, 1983). In the recent coastal zone the depth of these erosion features is 30–40 m (Palatnaya, 1982). Deeply entrenched erosional river valleys at depths of −93 to −122 m on the shelf edge (Ostrovskii et al., 1977; Skiba et al., 1975, etc.) are not reliably dated.

The general palaeogeographic situation in the area during the last glacial maximum was presented by Kaplin and Shcherbakov (1986). During this period, the whole Sea of Azov became a low-lying coastal plain, which was characterized by a cold and dry periglacial climate. The mouth of the Don River was situated 50 km south of the Kerch Strait, whereas the mouths of the Dnieper and Danube rivers lay 200 km south of the present ones. On the Caucasian coast, river mouths reached the heads of submarine canyons.

The Novoeuxinian epoch covered the whole period from 17–11 ka BP (Svitoch et al., 1998) and possibly to 9 ka BP (Fedorov, 1978). The whole range of radiocarbon dates for Novoeuxinian sediments is from 17,780 ± 200 to 9660 ± 70 years BP (Dimitrov, 1982; Shcherbakov, 1983; Shimkus et al., 1980). These sediments are usually found at −30–80 m (Chepalyga et al., 1989), whereas in some segments ancient coastlines of that age are situated at −20–25 m (Shcherbakov, 1983; Balabanov and Izmailov, 1988). It is worth noting that Novoeuxinian sediments at higher altitudes are usually younger than those at lower altitudes. Extreme estimates of −140-m sea level position during the Novoeuxinian epoch obtained from analysis of high-resolution seismic profiles and, partially, from dates on *Didacna rostriformis* bivalves (14,700 ± 65 to 10,400 ± 55 years BP) in ancient coastal and marsh sediments appeared recently (Ryan and Pitman, 1998; Ryan et al., 1997). However, these data need detailed confirmation by more extensive drilling.

Generally speaking, geological data support the idea that the Novoeuxinian epoch was a prolonged period of rising sea levels after the last glacial maximum lowstand (18–17 ka BP). An inflow of glacial meltwater from rivers and perhaps from the Khvalynian basin in the Caspian Sea (Svitoch et al., 1998) contributed much to the sea level rise in the Black Sea.

Based on mollusk fauna composition, water salinity was not over 5–6‰ (Nevesskaya, 1965). Moreover, the chemical composition of deep sediments allowed Degens and Ross (1972) to hypothesize that the salinity of Black Sea decreased to 3–7‰ 22–9 ka BP. Ryan et al. (1997) insist there was a “freshwater environment” at that time. It is unlikely that such a freshening could occur if there was a continuous connection with the Mediterranean as suggested by

Sholten (1974) and Shcherbakov (1983). Therefore, for a prolonged time period, the Novoeuxnian basin was an isolated "sea-lake".

The relative sea level rise in the late Novoeuxinian might have occurred at a high rate (Fig. 3). However, precise estimates of the sea level rise are not possible for many areas, e.g. for the Caucasian coast south of Novorosiisk because of the high amplitude of tectonic deformations, which are clearly seen from the inadequate altitudinal position of correlative surfaces along the coastline. To illustrate this, an ancient shoreline on the Caucasian shelf near Sochi at − 83 m is dated from 10.4 ka BP, whereas a shoreline dated from 9.5 ka BP is situated at − 30 m in the same area (Balabanov and Izmailov, 1988) but such a rate of sea level rise is not supported by data from more stable areas. Reasonable estimates can be obtained from the Kerch area and the northwestern Black Sea. For the period 11.5–10.5 ka BP, the sea level can be reasonably estimated at − 35–45 m and for the period nearly 9 ka BP, at − 20–30 m (Serebryanny, 1982; Varushchenko et al., 1987). These estimates are independently supported by our latest data on the Kerch Strait and adjacent areas (Kaplin et al., in press(a)). It is worth noting that the upper limit of the existence of sand, gravel, and shell coastal barrier features of the Novoeuxinian age is usually − 20–25 m (Shcherbakov, 1983), which is higher than the Bosporus sill.

Therefore, during the late stages of the Novoeuxinian epoch, nearly 10–9.5 ka BP, when the Black Sea was filled by low salinity water, a unidirectional outflow to the Sea of Marmara and the Mediterranean could have existed. This was initially proposed by Fedorov (1978) and later supported by independent palaeontological, geological and geomorphological data (Aksu et al., 2002; Hiskott et al., 2002; Mudie et al., 2002).

A further glacioeustatic global sea level rise inevitably resulted in the intrusion of saline water into the Black Sea from the Mediterranean. This palaeohydrologic event is dated from 10 ka BP (Ostrovskii et al., 1977), 9 ka BP (Degens and Ross, 1972), 9–8 ka BP (Kuprin and Sorokin, 1982) or 7.15 ka BP (7550 calendar a) (Ryan et al., 1997). The present pattern of two-layered exchange between the two seas was perhaps established 1–2 ka later than the intrusion of saline water into the Black Sea (Degens and Ross, 1972).

Evidence exists that a significant water–level drop in the Black Sea occurred during the early Holocene (10–8 ka BP). However, the principal evidence for this event is a problematic erosional surface in Novoeuxinian sediments and a change in mollusk fauna, e.g. in coastal areas of the Kerch Strait region (Fedorov, 1978) or in bottom cores in the northern part of the western depression (Khristev and Georgiev, 1991). Ryan et al. (1997) and Ryan and Pitman (1998) assume that extremely dry climate conditions were dominant in the Black Sea area and its drainage area nearly 8 ka BP. However, other palaeoclimate reconstructions give no evidence for this dryness in pollen records from the Black and Marmara seas aquatories (Mudie et al., 2002) and from the drainage area in the Russian Plain (Klige et al., 1993). In our opinion, a water-level drop did not occur at all or was not significant and the facts cited by the former researchers may be successfully explained in terms of local climatic and hydrological changes due to the intrusion of Mediterranean waters. No submerged coastal feature representing this regression is currently known.

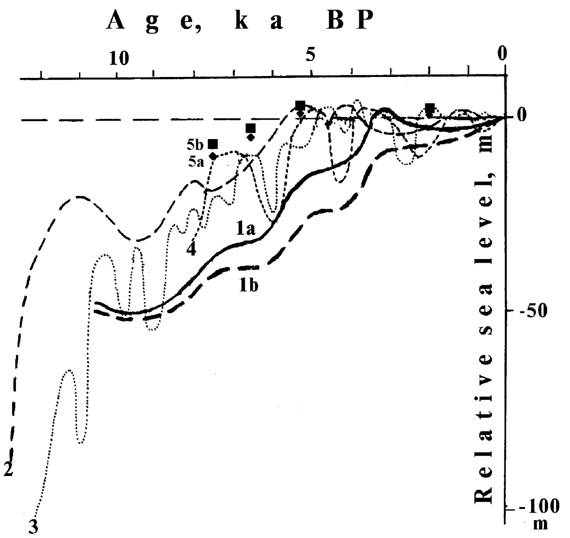

Fig. 3. Fluctuations in the Black Sea water level during the last 12 ka according to various authors. (1a) the Caucasian coast near Anapa; (1b) Karkinit Gulf (Nevesskii, 1967); (2) the Kerch Strait and Caucasian coast (Fedorov, 1985); (3) Kerch-Taman area (Arslanov et al., 1982); (4) Odessa Gulf and adjacent areas (Voskoboinikov et al., 1982); (5) Kuban river delta (this paper): (5a) without corrections for tectonic deformations; (5b) with corrections for tectonic deformations included. See Fig. 2 for locations.

2.2. The White Sea

During the Lateglacial, almost the entire basin of the present White Sea was filled by lobes of the Fennoscandian continental ice sheet. In the lower reaches of the Severnaya Dvina, Onega, Mezen and some other rivers, extensive lakes existed during the Lateglacial due to the barrier effect of the glacial lobes. These lakes at elevations of 50–140 m above the present sea level outflowed southwards from the ice sheets. Their flow could have reached Central Russia, presumably through the existing river valleys (Fig. 1). Additionally, they could have fed the Dnieper, Don and other rivers discharging into the Black Sea and the Sea of Azov (Gerasimov and Velichko, 1982).

Anomalously elevated shorelines of fluvioglacial lakes (127–138 m, up to 260 m in some locations) date from the Allerod and are typical for the White Sea coast of the Kola Peninsula, particularly for the Kandalaksha Gulf (Koshechkin, 1979; Fig. 4A). Later, the relative sea level decreased with a series of fluctuations (Fig. 5). The largest fall in relative sea level occurred between the Younger Dryas and Late Preboreal (10.5–9 ka BP) and was as fast as 45–50 m per thousand years (Fig. 5).

On the coasts of Onega Peninsula in the eastern White Sea, ancient coastal terraces of fluvioglacial or glaciomarine origin are traditionally distinguished at elevations of up to 60–80 m (Lavrova, 1969). However, our field observations do not prove the existence of terraces at elevations exceeding +20 m (Selivanov, 1996b; Kaplin et al., in press(b)). Moreover, the highest depositional terrace (13–20 m) is clearly of lacustrine origin.

In coastal bluffs in the east and southeast corner of the White Sea (Fig. 4A), tills of the last glaciation are usually covered at 10–18 m above the present sea level by laminated grey clays, aleurites with Arctic freshwater and mixed diatom and mollusk assemblages and periglacial sporo-pollen spectra (Selivanov, 1996b; Fig. 4B). In coastal segments with a less intensive sediment supply, an erosion of glacial tills during that time is marked by erosional escarpments up to 16–20 m, and in rare cases, 25–30 m above the present sea level. However, the maximum sea level fall in the area occurred in the early Holocene, 10–9 ka BP (Fig. 6), as in the Kola Peninsula. Southwest of the Onega Peninsula, in the Solovki Archipelago, a series of erosional terraces in pre-Quaternary rocks lies at elevations of up to 21–23 m (Nikishin, 1981).

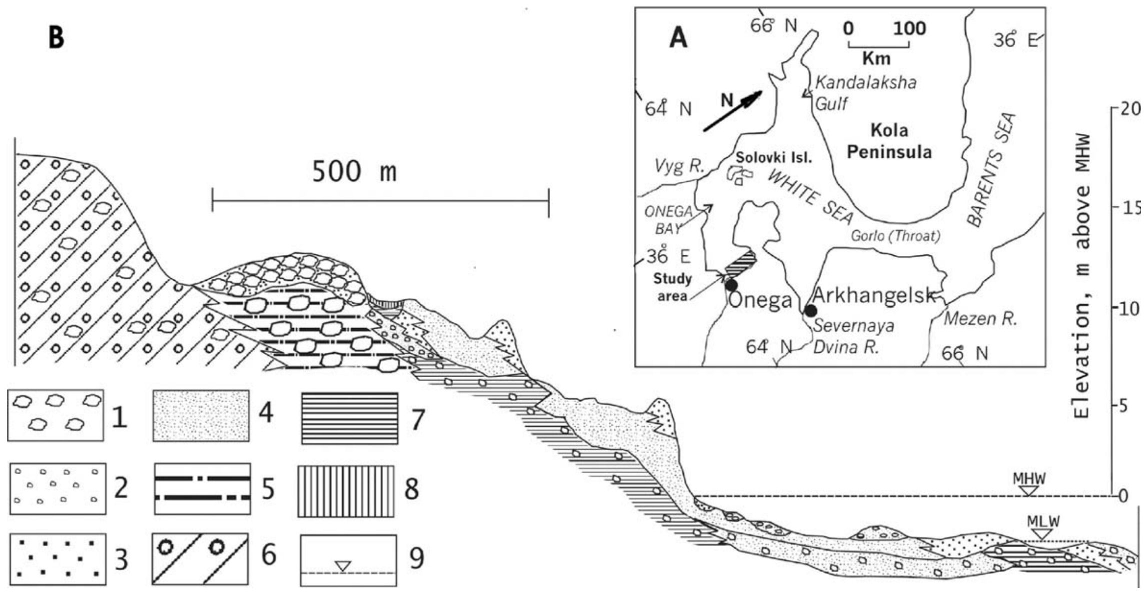

Fig. 4. (A) General scheme of the White Sea. (B) Schematic geological profile of the coastal zone in the eastern part of the Onega Bay, the White Sea: (1) boulders; (2) pebble and gravel; (3) laminated coarse sand; (4) laminated medium and fine sand; (5) sandy loam; (6) loam; (7) clay or gyttja; (8) peat; (9) present sea level position during the mean high water (MHW) and mean low water (MLW) periods.

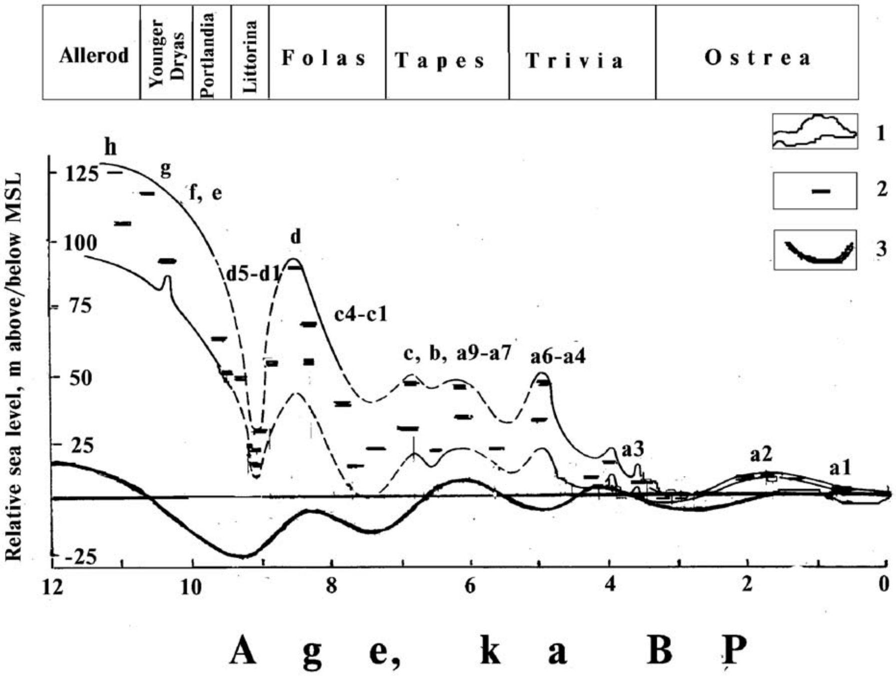

Fig. 5. Changes in the relative sea level on the Kola Peninsula during the Latest Pleistocene and Holocene (Koshechkin, 1979): (1) highest and lowest position of shorelines; (2) radiocarbon dates of marine sediments; (3) radiocarbon dates of continental sediments. Correlation with the "stadial shorelines" of Tanner (1930) is shown by letters (a–h).

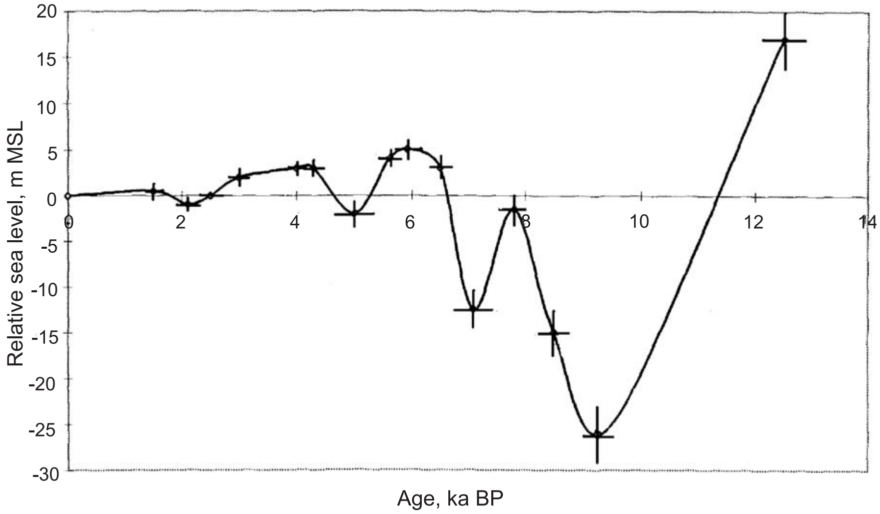

Fig. 6. Possible changes in the relative sea level in the eastern part of the White Sea during the Holocene. Uncertainty bars for age and height ranges are shown.

Therefore, the maximum water level of the periglacial lake was possibly +15–20 m (Fig. 6).

Direct radiocarbon measurements from lacustrine sediments have not yet been done. However, on the east coast of Onega Bay, conformably overlying peats are dated from 7980 ± 270 and 7570 ± 250 years, thus confirming the possible Early Holocene age for the lacustrine clays and aleurites.

According to data from sea floor cores, in the Boreal period (9–8 ka BP) the central basin of the White Sea and Onega Bay was still covered by extensive glacial masses (Nevesskii et al., 1977; Pavlidis et al., 1998). Intrusion of typical marine fauna into the central basin of the sea is dated from 8.5–8.2 ka BP (Koshechkin, 1979) and the steady bidirectional water exchange of the White Sea and the Barents Sea as a part of the Arctic Ocean was established at 7.7–7 ka BP (Nevesskii et al., 1977). However, a marine mollusk fauna was found on the Kola Peninsula in the sediments of the "main postglacial shoreline" ("g" in Table 1) dated from the Younger Dryas. Bottom cores in the Gorlo reveal marine fauna and diatoms beginning from the same period (Sobolev et al., 1995). This discrepancy may be preliminarily explained by the "dead ice" character of glacial masses in the central part of the sea basin and in Onega Bay that could survive in a periglacial marine basin for several millennia (Pavlidis et al., 1998).

3. Middle and late Holocene sea levels

3.1. The Black Sea

3.1.1. General stratigraphic schemes

The traditional stratigraphy of Holocene bottom sediments in the Black Sea is based on mollusk assemblages and includes Bugazian, Vityazevian, Kalamitian, and Dzhemetian layers (Nevesskaya, 1965). A rough age estimate for the first two layers is 8–6 ka BP and for the last two layers, 6–3 ka BP, respectively (Shcherbakov, 1983). Balabanov and Izmailov (1988) dated all the layers from earlier ages but the database is very small (three to five datings for each layer) and the data dispersion is too high to reliably estimate age and sea level position.

An alternative stratigraphic scheme based on coastal sedimentary sections includes Drevnechernomorian (Ancient Black Sea), Novochernomorian (New Black Sea), and Nymphean layers (Fedorov, 1978). According to the latest radiocarbon estimations, they can be dated approximately at 9.5–5.7, 5.7–3.0 and the last 3 ka, respectively.

3.1.2. Geological and geomorphological data

A series of submerged and emerged Holocene coastal terraces, mainly of erosional character, on the Caucasian coast of the sea was tentatively correlated to the stratigraphic scheme of Nevesskaya and Fedorov (Ostrovskii et al., 1977; Balabanov and Izmailov, 1988, etc.). The Bugazian, Vityazevian, and Kalamitian terraces, which were correlated with the Drevnechernomorian layers, were found at depths of 16, 10, and 3–4 m below sea level and dated from 9.5–7.9, 7.9–7, and 7–5.9 ka BP, respectively. Ancient estuaries formed at 8–7 ka BP were inundated in the northwestern Black Sea coast at altitudes of up to −8 m (Voskoboinikov et al., 1982). In the Kerch Strait area, submerged ancient coastal barriers at −30 m are dated from 7 ka BP and those at −17 m are dated from 6.5 ka BP (Shcherbakov, 1983). However, all these areas were obviously subjected to neotectonic deformation and can hardly serve as references.

Some researchers believe that at the end of this period, 6.5–5.8 ka BP, the sea level fell to −25–27 m and incised valleys formed on the Caucasian shelf (Ostrovskii et al., 1977; Arslanov et al., 1982; Fig. 3). However, these valleys are not dated and their age cannot be determined.

Depositional coastal features of the Novochernomorian stage are most extensive in the Black Sea basin. They are represented by gravel and sand terraces and associated barriers that separate lagoons in the northwestern corner of the sea and in Bulgaria, as well as by spits in West Crimea and the Kuban River delta in the Sea of Azov. They usually elevate up to 4–5 m above sea level both in tectonically stable and unstable coastal areas. Considering wave run-up, the highest sea level during the Novochernomorian stage was most likely +2–3 m. Based on cores taken in subsiding areas, the Novochernomorian transgression can be correlated with Dzhemetian layers (5.7–3 ka BP) (Nevesskaya, 1965).

It is only in the Kolchis Lowland in Georgia and in the Karkinit Gulf in the northwest of the Black Sea that the Novochernomorian coastal sediments are covered by the younger Nymphean sediments and no signs of sea level indicators higher than the present one are known. Both these regions are well known for tectonic subsidence.

In different areas, the Novochernomorian transgression dates from different times. On the Bulgarian coast it is dated from 5.5–4.5 ka BP (Svitoch et al., 1998). In the northwestern corner of the sea, it is dated from 4.5 ka BP (Molodykh et al., 1984), on the northern coast, from 5.5–4 ka BP (Fedorov, 1985), and in the Kerch region and on the Caucasian coast, from 4.2–3.8 ka BP (Arslanov et al., 1982) or nearly 3 ka BP (Nevesskii, 1967). Based on bottom drill cores, the transgression was possibly of a two-peaked character. This may be proved both by several coastal morphological indicators (Fedorov, 1985) and by the existence of two transgressive–regressive series in the Novochernomorian sediments in the Kolchis Lowland (Tvalchrelidze, 1989). In the last area, transgressive phases are dated from 5.7–5.2 and 4.2–3.9 ka BP, respectively.

It is evident that sea level curves for the different coastal stretches of the Black Sea vary greatly. Possible reasons for these differences are briefly discussed below. Valuable additional information on sea level changes may be obtained from coastal areas of steady deposition. The Kuban River delta (Fig. 2) is one of the most representative areas of this type. The Kuban River deltaic plain is an extremely gently sloping surface up to 150 km in width. Elevations do not surpass +3–5 m. Extensive areas are covered by swamps, lagoons, active and inactive deltaic channels. Holocene depositional complexes overlie Late Pleistocene loess sediments and are 11–14 m, or at places up to 20-m thick. Holocene sediments were studied by an extensive series of drilling cores supported by the analysis of mollusk fauna, lithological studies and radiocarbon dating. Holocene sediments are represented by the intricate intercalation of alluvial, deltaic, lagoon and coastal sediments (Fig. 7). They form several coastal complexes, both submerged and emerged. The most indicative features of these complexes are coastal depositional barriers composed primarily of shells and detritus (up to 70–80%) with fine quartz sand. Radiocarbon dating was carried out

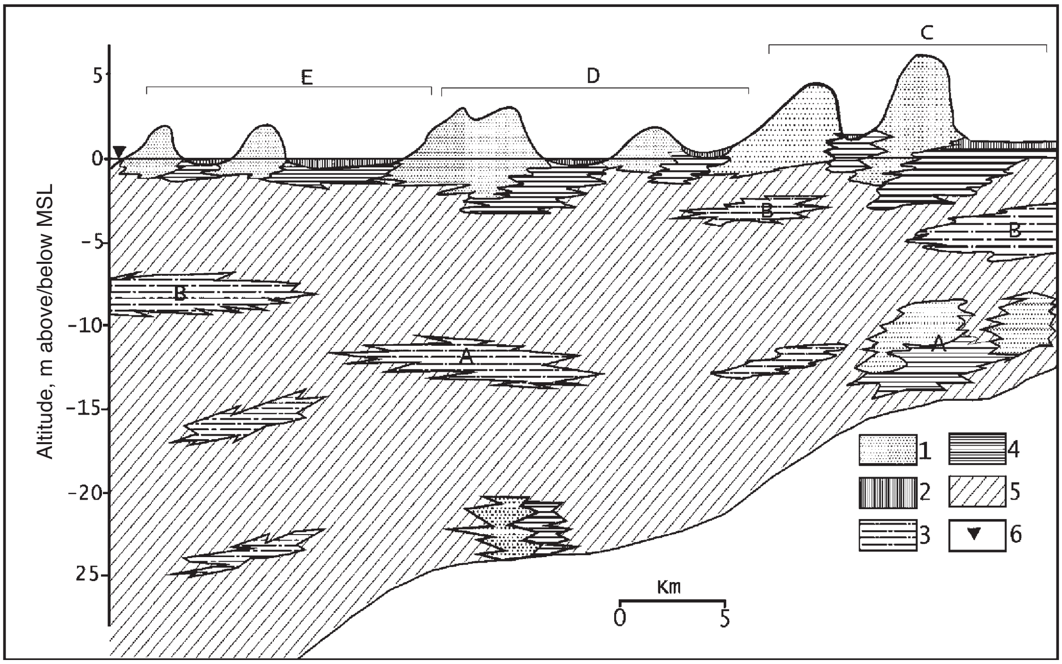

Fig. 7. Schematic geological profile of the Holocene sediments in the Kuban River delta: (1) sandy shells of coastal barriers; (2) lagoon silts and gyttja; (3) lagoon peats; (4) deltaic silts and sandy silts with peat layers; (5) deltaic shell sands and loams; (6) present mean sea level. Ancient coastal complexes (A–E) are described in the text. See 5 in Fig. 2 for location.

on inner layers of thick *Cerastoderma glaucum* (*Cardium edule*) shells from these barriers (Izmailov et al., 1989 and later).

Based on morphological and altitudinal position, as well as radiocarbon dates, we distinguish five primary coastal complexes (Fig. 7):

(A) a submerged coastal complex at − 8.6–12.5 m dated from 7380 years BP (Vityazevian stage);
(B) a submerged coastal complex at − 3–8 m. It is preliminarily correlated with the Kalamitian stage of the Black Sea (approximately 7–6 ka BP). However, direct radiocarbon assays have not yet been obtained;
(C) an emerged coastal complex with lagoon surfaces at + 0.5–2.5 m with ages from 5.7 to 4.5 ka BP and coastal barriers up to 4.5–5.5 m (Dzhemetian stage). The mollusk fauna of this period (*C. glaucum*, *Chione gallina*, *Mytilus galloprovincialis*, *Corbula mediterranea maeotica*, *Hydrobia ventrosa*, etc.) indicates the highest salinity occurred during the Holocene. This fact confirms the existence of intensive water exchange with the Black Sea, perhaps as a result of the high Holocene sea level;
(D) an emerged coastal complex with lagoon surfaces at 0 to + 1.5 m aged from 2.2–1.7 ka BP and coastal barriers up to + 2.5 m (Nymphean stage). This complex is similar in morphology and altitudes to the previous one but is clearly differentiated from it by its position nearly 8–10 km to the west;
(E) the present coastal barriers with elevations of up to + 2.5 m.

In general, landward migration of the shoreline occurred until the Middle Atlantic period (Kalamitian stage, 7–6 ka BP) and changed to its seaward migration since that time. This phenomenon is typical for many coasts of the world and is possibly conditioned by the deceleration of sea level rise during that period (Selivanov, 1996a,b). Information on the Kuban River delta may be used to estimate sea level position during different periods of its Holocene history (Fig. 3). Ancient lagoon surfaces are the most reliable indicators of sea level position. It should be noted that complexes (4), and possibly (3), may have been partially modified in recent times by storm surges exceeding 3 m, according to direct observations. Additionally, changes in the sedimentary budget of particular deltaic areas due to the migration of channels or to the position of barrier forms cannot be excluded. It is reasonable also to allow for tectonic subsidence of the deltaic area. According to direct observations, rates of subsidence in the 20th century in the central part of the Kuban delta are nearly 4.5 mm/year, and 3 mm/year in peripheral parts (Selivanov, 1995). With the moderate correction of 3 mm/year included, the mean sea levels during the middle and late Holocene transgressive stages may be tentatively estimated at the following marks (Fig. 3): − 6 m during the Vityazevian stage (7380 years BP); − 1 to − 2 m during the Kalamitian stage (7–6 ka BP); + 3 m during the Dzhemetian stage (5.7–4.5 ka BP); + 1.5 to + 2 m during the Nymphean stage (2.2–1.7 ka BP).

No direct indicators of low sea levels during the Holocene regressive stages are available for the Kuban River delta. However, beds of lagoon silts and gyttja since the Kalamitian stage are situated not lower than − 2–2,5 m. This level can perhaps be regarded as the “base” for minimum sea level position during the past 6 ka.

3.1.3. Archaeological data

The late Holocene history of sea level in the Black Sea is reliably known owing to the abundance of Greek and Roman archaeological sites in the northwest of the sea, in Crimea, and in the Kerch region. It is traditionally cited in the northern Black Sea as Nymphaean layers (Fedorov, 1978). On sea coasts of the area, depositional features and, rarely, erosional surfaces of the Novochernomorian transgression in the middle 1st millennium BC were occupied by the ancient towns of Tira, Olbia, Khersonesus, Panticapaeum, Phanagoria, Dioskuria, etc. The lowermost builtup level of the 5th–3th centuries BC lies at 3–4 m below sea level and in Dioskuria (near present-day Sukhumi), even at − 10 m (Agbunov, 1992). The ancient Istria in the Danube River delta was also inundated (Stefan, 1987).

Therefore, the sea level during that time, known as the Phanagorian regression, might have fallen by several meters. However, estimates of its lowermost position vary from − 5–7 m (Fedorov, 1985) to − 8–10 m (Ostrovskii et al., 1977), − 10 m (Shilik, 1997) or even to −13 m (Arslanov et al., 1982). We attribute most of the archaeological sites of that time to the specific river mouths definitely known as areas of recent submergence (Fig. 8). The more reliable estimate of −3 m may be deduced from the studies of ancient Khersonesus, near present-day Sevastopol in Crimea (Blagovolin and Shcheglov, 1968).

The most recent Nymphaean coastal features in the Black Sea were first distinguished by Pavel Fedorov near the ancient town of Nymphei on the eastern side of the Kerch Strait. Correlative coastal depositional features (terraces, barriers, rarely spits) elevated up to +2.5–3 m are known from the northwest corner of the sea and from the coast of Bulgaria. In some places, transgressive series overlie Greek (not Roman) cultural remnants and, therefore, may be dated from the early 1st millennium AD. Roman archaeological sites are always situated above the present sea level. From archaeological data, it may be deduced that the rise of sea level after the Phanagorian regression began from the 1st–3rd centuries AD and was not significant. The highest level of the transgression was possibly in the range of +1.5–2 m (Fedorov, 1985) and, by our data (Kaplin et al., in press (a)), +1 m. The few radiocarbon dates range from 1.7 to 1.2 ka BP (Arslanov et al., 1982; Fedorov, 1985).

Later, the Black Sea water level gradually decreased until the middle of the 19th century. A number of researchers believe it fell in Medieval times (Korsunian regression, 1.4–1.5 ka BP) to −2–3 m (Shilik, 1997) or −3–5 m (Ostrovskii et al., 1977). However, reliable geological and archaeological data on this regression are few in number.

Our own independent data on the Kerch Strait and Kuban River delta do not prove any significant sea level fall during that period. It was probably not lower than −0.5 m (Kaplin et al., in press (a)).

3.2. The White Sea

Middle and late Holocene coastal complexes on the Kola coast of the White Sea are represented by staircases of depositional and erosional–depositional terraces with marine mollusk fauna at elevations of up to 90 m (Table 1). The best developed terraces date from the Folas period (8.9–8.3 ka BP), Tapes 2 (6.4–5 ka BP) and Trivia 1 (4.1 ka BP) and are traced along the altitudes of 45–75, 20–25 and 15–20 m, respectively. Shorelines of the second stage of the Tapes transgression are typical for the most abundant and best-studied mollusk assemblage and the pumice layer. The shoreline of the last significant transgressive period Ostrea 1, at +5–8 m, can perhaps be dated from the first two or three centuries AD.

In embayments, glacial tills form a significant part of outcrops in the higher terraces whereas the lower terraces are usually characterized by exclusively depositional character. On headlands, terraces are usu-

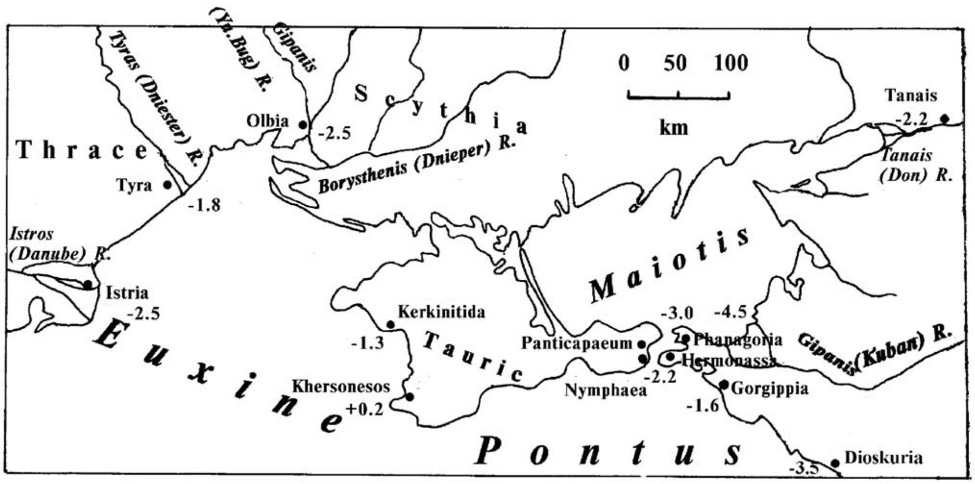

Fig. 8. The Black Sea in the late 1st millennium BC (approximately 2200–2000 a BP). The figure covers the same area and has the same scale as in Fig. 2 but with historical names. Present rates of vertical tectonic deformations by instrumental observations are shown in mm/year.

ally poor in sedimentary cover and characterized by higher elevations.

Near the Vyg river mouth, in the south of the sea, a detailed stratigraphy of Holocene sea level and climate changes was established by archaeological investigation and pollen analysis of coastal terraces up to 45–50 m above the present sea level (Devyatova, 1976). It was assumed that sea level fluctuations during the Middle and Late Holocene were as high as 15–20 m in amplitude. However, this conclusion was not supported by detailed lithological analysis and by tracing of the terraces along the coast.

On the Onega Peninsula, a 3-m terrace is widely developed in embayments. Its sections usually consist of two transgressive–regressive sequences of sand and mud facies of the upper shelf, inter-tidal flat, beach and lagoon (Selivanov, 1996b). Peats indicating successive falls in the relative sea level were dated from 8705–7825 and 4030 ± 70 years BP (Koshechkin et al., 1977).

On the open east coast of Onega Bay stretching along the belt of end moraines dated from 13–12 ka BP, erosional features are obvious in the seaward slope of the moraine ridge, which is situated several hundred meters inland from the present beach (Selivanov, 1996b; Fig. 4B). Sediments of the ancient lagoon at altitudes of + 6–9 m are reliably dated from the middle Atlantic period: 6455 ± 80, 5940 ± 250, 5600 ± 250, 5240 ± 200 years BP (Boyarskaya et al., 1986). During the following sea-level fall, a series of coastal ridges and a coastal dune formed at the present altitude of + 7–10 m. The subsequent transgressive–regressive cycle in relative sea level is marked in this coastal segment by the layer of beach gravel and sand at 3–4 m above the present sea level and a series of coastal sand ridges and a coastal dune at + 5–6-m altitude. This coastal depositional terrace may be tentatively attributed to the Subboreal period by the information on the Late Neolithic archaeological site (approximately 4.5–4 ka BP in the regional chronology) found in its upper sediments. This age is confirmed by two radiocarbon dates of the upper transgressive layer from the Holocene terrace in the adjacent embayment 4800 ± 180, 4100 ± 150 years BP. Pollen analysis reveals a general coincidence of the periods of rising sea levels and climate amelioration (Boyarskaya et al., 1986).

Summarizing the data above, in the White Sea the relative sea level during the transgressive stages may be estimated as follows: – 2–5 m during the late Boreal–early Atlantic period (8.5–7.5 ka BP), + 3–5 m during the Late Atlantic period (6.5–5.2 ka BP) and + 2–3 m during the Subboreal transgressive period (4.5–4 ka BP) (Fig. 6). During the Subatlantic period, a sea level significantly higher than the present one is doubtful. The highest sea level of + 0.5 m was perhaps reached at 1.8–1.5 ka BP. The possible influence of relatively slow coastal emergence on the altitudinal position of these shorelines cannot be excluded. Similar to Boyarskaya et al. (1986), Selivanov (1996b) did not find any traces of the relative sea level lower than – 1–2 m during the late Preboreal–early Boreal (9.5–8.5 ka BP) and late Subboreal–early Subatlantic (3–2 ka BP) regressive periods. Therefore, drastic sea level fluctuations by over 10 m during the past 8 ka, presumed to be the case for the Kola Peninsula and the Vyg River mouth, are not supported by these data. These drastic fluctuations could have resulted from the local block deformation of coastal areas (Selivanov, 1996a).

4. Discussion

Both the Black Sea and the White Sea have limited water exchange with the world's oceans. In the White Sea, the situation is additionally complicated by the existence of continental ice sheets and blocks of "dead ice" in the early Holocene (Pavlidis et al., 1998). The existence of large "dead ice" blocks in the White Sea is supported by independent changes of its water level at the western and eastern coasts of the sea during the lateglacial and early Holocene periods.

The possibility of long-term differences in water level between the Black Sea and the Mediterranean is under serious question. The hypotheses of Fedorov (1978, 1985), Avenarius (1979), and Shcherbakov (1983) on the preservation of significant differences in water level between the Black Sea and Mediterranean for several centuries and even millennia have already been mentioned. To illustrate, Shcherbakov (1983) supposes that during regressive periods the transverse flow area of the Bosporus decreased to such an extent that a 30–40-m difference was established between the water levels in the Black Sea and the Sea of Marmara, while the outflow from the Black Sea was retained during the whole regressive stages. Conversely, Fedorov (1978, 1985) believes that a unidirectional outflow from the Black Sea to the Sea of Marmara existed for several millennia, particularly in the Late Pleistocene.

The mean width of the Bosporus is over 2 km now, whereas the maximum width is 3.8 km and the minimum one is 700 m, its length being 30 km and the present sill depth 45 m (Gorshkov, 1980). It may have been that the depth of the stream decreased to 5 m, but due to the existence of rapids, it is hardly probable because the bottom of the Bosporus down to at least −85 m is composed of loose Quaternary sediments, which would have been intensively eroded under the water-level lowering. Making a comparison with the strait that connects the Kara Bogaz Gol Bay with the main water body of the Caspian Sea (Selivanov, 1998), it seems likely that the rate of bottom erosion is as high as several mm per year, i.e. comparable to the possible rate of the water-level fall.

The mean annual water flow from one basin to another can be estimated using the well-known Chezy equation:

$$Q = cw\sqrt{(RI)},$$

where c is the dimensional coefficient equal to 20–50 for such a stream (Zheleznyakov, 1981), w is the transverse flow area, R is its hydraulic radius, and I is the mean water surface slope. Even if the extreme assumption of a decrease in depth to 5 m was true, for a difference in water levels equal to 20 m, the mean annual flow along the strait would be as high as several hundred km^3/year and 100 km^3/year for a water-level difference of 5 m. the present water budget of the Black Sea is estimated by instrumental data at +165 km^3/year (Klige, 1985) or even +300 km^3/year (Aksu et al., 2002). As cited earlier in this paper, based on palaeoclimate data, water inflow into the sea could have been higher, evaporation from its surface could have been lower, and precipitation could have been similar to present-day conditions. In this case, the realistic positive water budget of the Black Sea estimated at 300–400 km^3/year would have resulted in the equalizing of water levels between the neighboring seas in several years or decades.

Based on the salt budget of the Black Sea, the existence of a unidirectional outflow from the Black Sea to the Sea of Marmara for more than several centuries is also improbable. The mean salinity of the Black Sea varied from the present value or a slightly higher value during the transgressive stages (Nevesskaya, 1965) to 5‰ or less during periods of low water level. In the event of the annual water flow along the Bosporus, such a drastic change in water salinity would have occurred over several centuries.

Therefore, the unidirectional water exchange between the Black Sea and the Mediterranean would inevitably have ceased on a 100–1000-year time scale. In any case, during a geologically short time period of several centuries, the unidirectional outflow from the Black Sea to the Sea of Marmara and, further, to the Black Sea could have occurred and was possibly the case nearly 10–9.5 ka BP.

However, the abovementioned estimates cannot be used as a confirmation for the idea of catastrophic filling of the Black Sea at a rate of several dozens of cm per day (Ryan and Pitman, 1998; Ryan et al., 1997). According to our data, at the period of water intrusion into the Black Sea (9–7.2 ka BP), its water level rose to nearly −20–30, possibly −35 m, which differed from the mean global sea level by not more than 10–15 m. The greatest rate of sea level rise could only have been on the order of several cm/year.

Moreover, a hypothesis exists that significant sea level fluctuations occurred during the middle and late Holocene. The inadequate facies interpretations of sediments in bottom cores can be the principal reason for extreme estimates of sea level fluctuations by Voskoboinikov et al. (1982) and some other researchers (Fig. 3). However, migration of the mollusk fauna could occur even during such a geologically short time period. Moreover, water-level falls in the Black Sea to −25 m at 6.5–5.8 ka BP and to −8–12 m at 3–2 ka BP as proposed by several researchers (Fig. 3) are improbable during the periods of water exchange with the Mediterranean and the world's oceans. The same is true for the middle and late Holocene history of the White Sea (Selivanov, 1996a).

The assumption that sea level fell to −10–15 m during the Phanagorian transgression and to −2–5 m in the Medieval period, as proposed by Balabanov and Izmailov (1988), possibly resulted from the unjustified comparison of terraces of questionable coastal genesis for the transgressive phases with bottom cores for the regressive stages. Pirazzoli (1991) reasonably suggested that several sea level curves with fluctuations of over a dozen meters (Serebryanny, 1982; Fedorov, 1985; Chepalyga et al., 1989) could have resulted from the unjustified comparison of ancient sea level indicators from areas differing in tectonic regime. Deep sea level falls in the middle and late Holocene of the Black sea as indicated by data on submerged ancient towns (Fedorov, 1985; Shilik, 1997) obviously resulted from tectonic submergence of several coastal stretches (see Section 3.1.3; Fig. 8).

Bearing in mind the higher water exchange of the White Sea in comparison with the Black Sea, we cannot assume the existence of large (in the order of over 5–7 m) water-level fluctuations in the White Sea during the middle and late Holocene (Selivanov, 1996a).

5. Conclusions

A comparison of the Black and White Seas, which differ in their tectonic, glacial and climatic history, but which share a strong dependence upon limited water exchange with the world ocean, represents an opportunity for the identification and correlation of major factors controlling sea level changes during the Lateglacial and Holocene. Existing data were critically analyzed and compared with the results of geological, geomorphological and palaeohydrological studies obtained by the present authors during the past two decades.

We conclude that glacioeustatic processes played a major role in relative sea level changes on most coasts of both areas. However, along several coastlines other factors overwhelm glacioeustasy during some time intervals. In the Black Sea, water level rose from its minimum position of −100–120 m at 18–17 ka BP to −20–30 m at nearly 9 ka BP. From 18 to 10 ka BP the Black Sea was possibly an isolated "sea-lake". At nearly 10–9.5 ka BP, a unidirectional outflow from the Black Sea to the Sea of Marmara and further to the Mediterranean occurred. After the intrusion of Mediterranean waters into the Black Sea in the early Holocene (9–7.5 ka BP according to various data), the water-level changes depended upon global sea levels. However, water exchange with the ocean remained limited. The hypothesis of a catastrophic sea level rise from between −120 and −150 m to between −15 and −20 m in a couple of years nearly 7550 calendar years BP (Ryan and Pitman, 1998; Ryan et al., 1997) is not supported by our data. Water intrusion from the Mediterranean was fast but not catastrophic.

The lateglacial and postglacial history of sea level in the White Sea depended to a large extent upon the glacial history of the region. Various interpretations of this history, from continental glaciers spreading onto the shelf to "glacier shelves" and extensive masses of "dead ice" in sea basins, as well as the low reliability and the lack of radiocarbon dates, result in a broad variety of possible sea level scenarios for the region. The situation becomes more complicated due to difficulties in tracing shorelines along the coast due to both intensive glacioisostatic rebound in the western sector of the region and to the block character of tectonic movements in the whole area.

In the White Sea, the decreasing trend in relative sea level is well illustrated on the Kola Peninsula and in Karelia subjected to glacioisostatic emergence. A drastic sea level fall from +15 to −25 m occurred with the drainage of glacial lakes in the eastern White Sea (12.5–9.5 ka BP).

In general, the decreasing trend in relative sea level in the Kola Peninsula and Karelia was superimposed by fluctuations on the order of 8–10 m in the Lateglacial and early Holocene (12–8 ka BP). In contrast, in the eastern White Sea, fluctuating patterns of sea level changes were established after the drainage of proglacial lakes in the Younger Dryas or Early Holocene. The trend of increasing sea level in the Black Sea since the Lateglacial is well documented.

Smaller-scale sea level changes of nearly 1.5–2.5 ka in duration are documented both for northern and southern European Russia in the middle and late Holocene (the last 8 ka BP). In the Black Sea, the periods of high sea levels are dated from four sedimentary complexes (Vityazevian, Kalamitian, Dzhemetian and Nymphean) from nearly 7.5, 7–6, 5.5–4.5 and 2.2–1.7 ka BP, respectively. In the White Sea, the major periods of sea level rise are dated from the late Boreal–early Atlantic (8.5–7.5 ka BP), late Atlantic (6.5–5.2 ka BP), middle Subboreal (4.5–4 ka BP) and middle Subatlantic (1.8–1.5 ka BP).

In tectonically stable areas, fluctuations of relative sea level during the middle and late Holocene were possibly on the order of several meters (from +2–3 to −2–3 m in the Black Sea and from +3–5 to −2–3 m in the White Sea). The higher estimates for the Kola Peninsula and some other coastal stretches reflect the inadequacy of methodology, namely the comparison of areas with differing tectonic regimes.

The hypothesis that significant differences in water level occurred between the Black Sea and the Mediterranean, the White Sea and the Barents Sea in the Middle and Late Holocene is not supported. The hypothesis of large-scale, 15–25-m, water-level fluctuations in these seas during the Middle and Late Holocene is rejected both by geological and palaeohydrological data.

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