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Marine Geology 190 (2002) 19–34

MARINE GEOLOGY
INTERNATIONAL JOURNAL OF MARINE GEOLOGY, GEOCHEMISTRY AND GEOPHYSICS

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Constraints on Black Sea outflow to the Sea of Marmara during the last glacial–interglacial transition

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Received 26 April 2001; accepted 19 February 2002

Abstract

New cores from the upper continental slope off Romania in the western Black Sea provide a continuous, high-resolution record of sedimentation rates, clay mineralogy, calcium carbonate content, and stable isotopes of oxygen and carbon over the last 20 000 yr in the western Black Sea. These records all indicate major changes occurring at 15 000, 12 800, 8400, and 7100 yr before present. These results are interpreted to reflect an evolving balance between water supplied by melting glacial ice and other river runoff and water removed by evaporation and outflow. The marked retreat of the Fennoscandian and Alpine ice between 15 000 and 14 000 yr is recorded by an increase in clays indicative of northern provenance in Black Sea sediments. A short return toward glacial values in all the measured series occurs during the Younger Dryas cold period. The timing of the first marine inflow to the Black Sea is dependent on the sill depths of the Bosporus and Dardanelles channels. The depth of the latter is known to be -80 ± 5 m, which is consistent with first evidence of marine inundation in the Sea of Marmara around 12 000 yr. The bedrock gorge of the Bosporus reaches depths in excess of -100 m (relative to present sea level), though it is now filled with sediments to depths as shallow as -32 m. Two scenarios are developed for the connection of the Black Sea with the Sea of Marmara. One is based on a deep Bosporus sill depth (effectively equivalent to the Dardanelles), and the other is based on a shallow Bosporus sill (less than -35 m). In the deep sill scenario the Black Sea's surface rises in tandem with the Sea of Marmara once the latter connected with the Aegean Sea, and Black Sea outflow remains continuous with inflowing marine water gradually displacing the freshwater in the deep basin. The increase in the $\delta^{18}O$ of mollusk shells at 12 800 yr and the simultaneous appearance of inorganic calcite with low $\delta^{18}O$ is compatible with such an early marine water influx causing periodic weak stratification of the water column. In the shallow sill scenario the Black Sea level is decoupled from world sea level and experiences rise and fall depending on the regional water budget until water from the rising Sea of Marmara breaches the shallow sill. In this case the oxygen isotope trend and the inorganic calcite precipitation is caused by increased evaporation in the basin, and the other changes in sediment properties reflect climate-driven river runoff variations within the Black Sea watershed. The presence of saline ponds on the Black Sea shelf circa 9600 yr support such evaporative draw-down, but a sensitive geochemical indicator of marine water, one that is not subject to temperature, salinity, or biological fractionation, is required to resolve whether the sill was deep or shallow.

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PII: S0025-3227(02)00340-7

Keywords: deglaciation; Younger Dryas; sapropel; stable isotopes; clay mineralogy

1. Introduction

The Black Sea lies at the interior end of a series of basins connected to the open ocean via straits and their sills (Ünlüata et al., 1990) (Fig. 1). High-salinity water from the Mediterranean travels from the Aegean Sea, through the Dardanelles Strait and the Sea of Marmara, to enter the Black Sea as an undercurrent in the narrow Bosporus Strait. The marine inflow has led to density stratification of the Black Sea water column. As a result, the Black Sea waters below ~200 m have been anoxic for the past 7160 yr[1] (Jones and Gagnon, 1994).

The Sea of Marmara and the Aegean Sea also show evidence of past but not current anoxia (Aksu et al., 1995). The onset of anoxia in these basins preceded that in the Black Sea and has been attributed to freshwater outflow from the Black Sea while it was isolated from the world ocean and existed as a freshwater lake (Olausson, 1961; Stanley and Blanpied, 1980; Aksu et al., 1999; Cagatay et al., 2000). Many researchers consider the Black Sea outflow to have been continuous through both glacial and interglacial time (Kvasov, 1968; Ross and Degens, 1974; Kvasov, 1975; Chepalyga, 1984). The outflow would have been augmented by displacement of the deep water stored behind the Bosporus Sill once the connection from the Mediterranean was established (Lane-Serff et al., 1997).

An alternative view suggests that outflow was intermittent and ceased when the Black Sea behaved like the inland Caspian Sea and dropped its surface below its outlet (Ozdogan, 1990). Shelf-beveling erosion surfaces (Ryan et al., 1997) and wave cut terraces at −80 to −155 m (Kuprin et al., 1974; Shimkus et al., 1980; Dimitrov, 1982; Ballard et al., 2000) require a significant Black Sea level fall that could have transformed the Bosporus Strait into a subaerial valley (Scholten, 1974). These terraces are assumed to have formed during the last glacial period, based mainly on the age of the sediments that overlie them.

Any interpretation of the connection history between the Black Sea and the Sea of Marmara is dependent upon assumptions about the Bosporus sill over which exchange took place. The tectonic activity along the north Anatolian fault, which runs just south of the Bosporus through the eastern Sea of Marmara, has led some researchers to suggest that vertical motion along the fault and not sea level has been the dominant control on connection between the basins (Demirbag et al., 1999). Others have suggested that there was a different gateway in the past, possibly through the Sakarya River valley, an idea supported by evidence for estuarine or marine sediments in this region (Pfannenstiel, 1944; Chepalyga, 1995; Meriç, 1995). We will not expand on these ideas here, because the sill we refer to is a generic feature whose depth is unknown.

Though a shallow sill similar to the present Bosporus configuration is most commonly proposed, seismic reflection profiles and boreholes clearly show a bedrock gorge beneath this strait partly filled with sediment (Yilmaz and Sakinc, 1990). Mollusk shells sampled in the fill of the gorge and 70 m below present sea level have been dated by electron spin resonance at 7400 ± 1300 yr (Göksu et al., 1990). Another borehole has 26 100-yr-old (radiocarbon dated) shells of the freshwater mollusk *Dreissena* (radiocarbon dated) above the bedrock at −105 m (Cagatay et al., 2000). Thus the possibility of a deep Black Sea/Marmara connection in the past cannot be excluded.

This paper addresses the question of a continuous or intermittent outflow from the Black Sea using a time series of paleoenvironmental indicators in sedimentary cores collected from the continental slope beyond the mouth of the Danube River. Our records indicate a strong climatic control on sedimentation, with particular variability

[1] All dates are reported as conventional radiocarbon years, using 1950 as the base year and the 5568 year Libby half-life. Dates are not corrected for reservoir age and are not calibrated to calendar years.

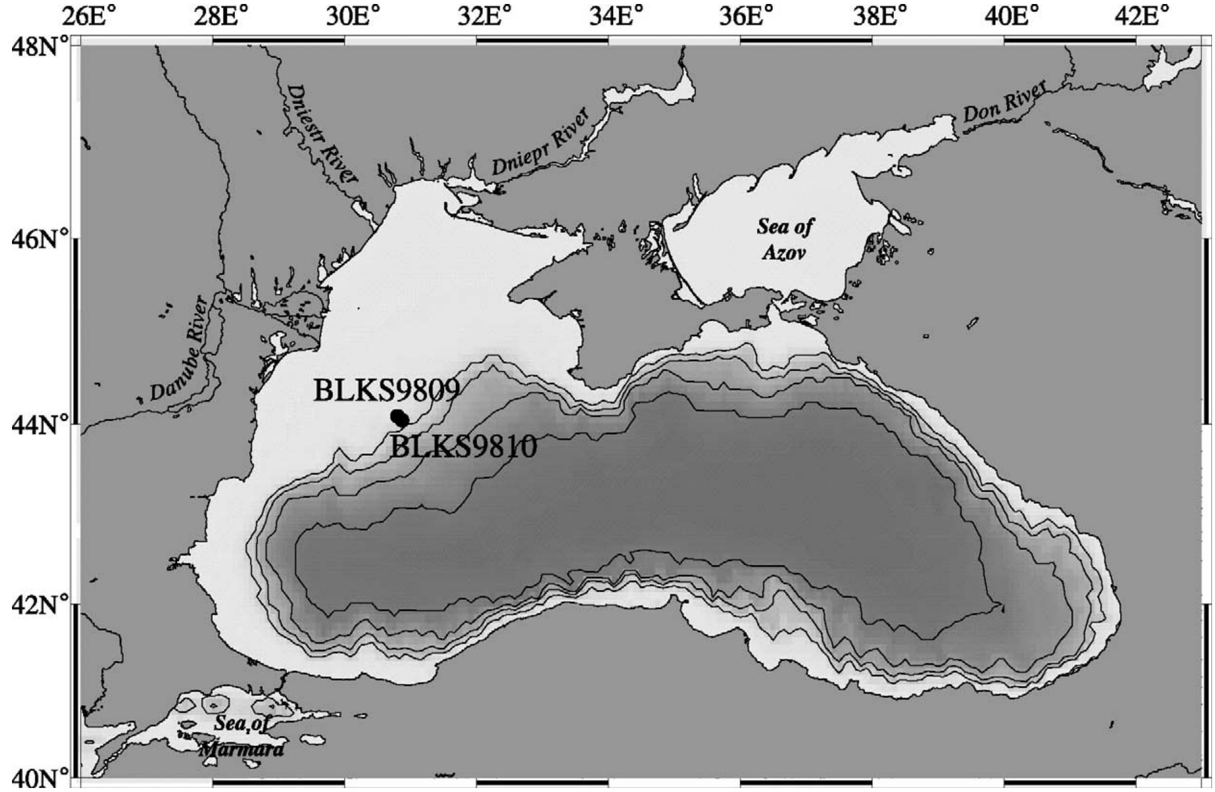


Fig. 1. Map of the Black Sea, showing major rivers and core locations. Contour interval is 500 m.

occurring over the period of the last deglaciation. Although there would be an earlier connection between the basins if the sill were deep, periods of outflow from the Black Sea prior to reconnection are controlled by the freshwater balance of the basin. The depth of the sill represents the upper limit of Black Sea level during its isolated phase.

1.1. Paleoclimatic context

The climate of the deglaciation brought about hydrological changes in the Black Sea drainage basin, altering runoff by ice melt contribution, reconfiguring drainages because of ice retreat, and enhancing groundwater storage with the elimination of permafrost. Freshwater input to the Black Sea would also be influenced by regional aridity, which is reflected in terrestrial records such as pollen assemblages and lake levels. Pollen data (Filipova et al., 1983; Peterson, 1983; Allen et al., 1999; Tzedakis, 1999) reveal that during the last glacial period the areas north and west of the Black Sea were populated with a steppe vegetation dominated by *Artemisia* and *Chenopodiaceae* and indicative of a cold, dry climate. At the same time, lake shorelines on the Anatolian Plateau were high (Roberts et al., 1979), suggesting low evaporation due to low temperatures at high altitudes. Because significant meltwater influx would be expected to bring the Black Sea to its outflow, regardless of the sill depth, it is important to investigate the timing of glacial retreat in the Black Sea watershed.

At glacial maximum, the southern margin of the Scandinavian ice sheet fed pro-glacial lakes in the Dnieper River drainage basin (Grosswald, 1980). Paleohydrology of the upper Dnieper reveals braided stream deposits indicative of ice-proximal high-energy discharge. These deposits are incised by underfit river-channel and overbank deposits dated 10 200 yr, typical of modern lower

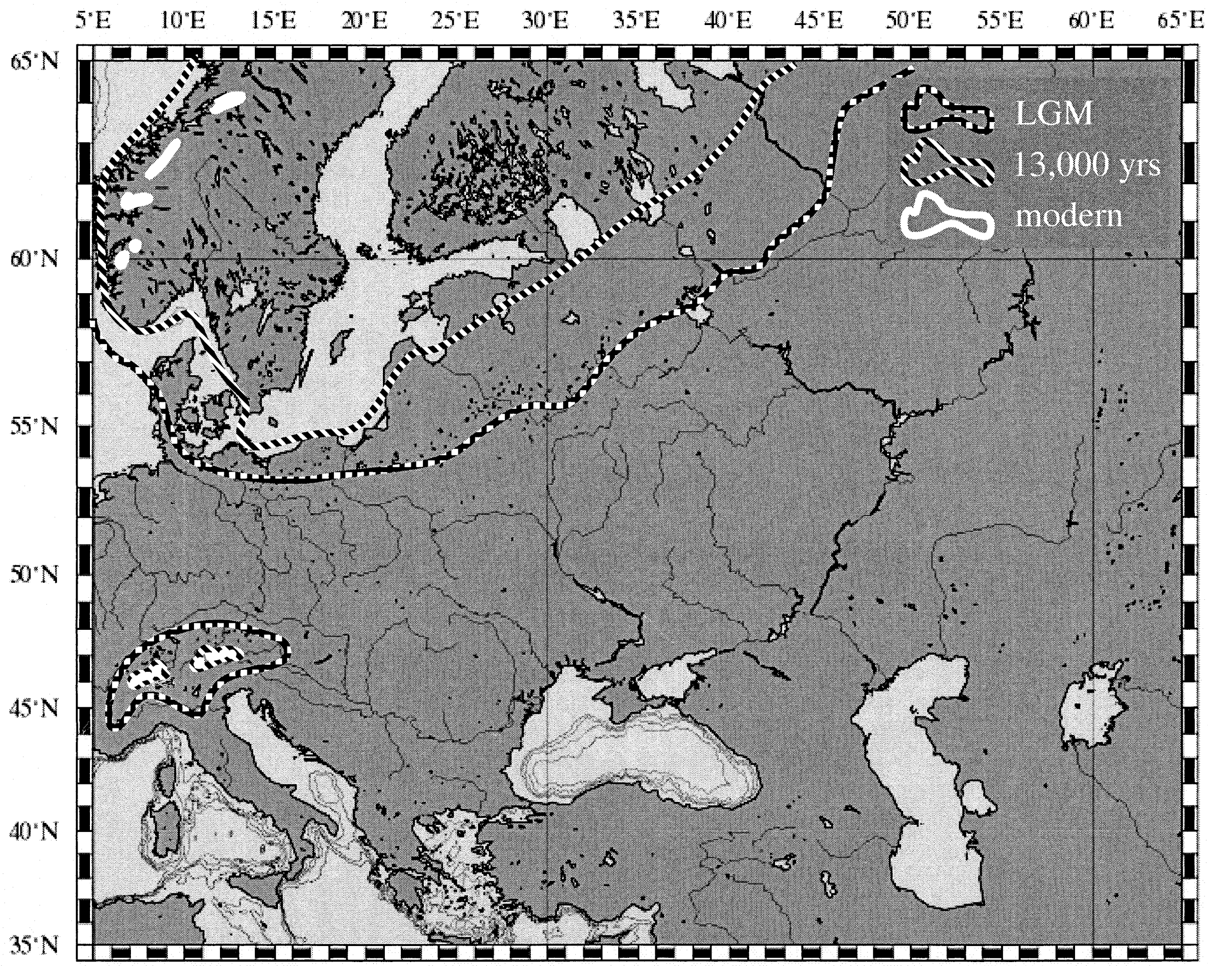


Fig. 2. Stages of ice coverage (adapted from Denton and Hughes, 1981), showing the retreat of ice out of the Black Sea watershed by 13 000 yr.

energy flow (Kalicki and Sanko, 1998). The Pomeranian moraine of the Scandinavian ice sheet, marking the last advance into areas south of the Baltic Sea into Poland and Germany, is dated between 18 000 and 15 000 yr (Denton and Hughes, 1981; Fig. 2). Beginning at about 15 000 yr ago, the rapid retreat of ice sheets and glaciers in northern Europe and the Alps would have contributed a burst of meltwater via the Dnieper and Danube rivers that was concluded in less than two millennia and perhaps only a few centuries (Denton et al., 1999). The Raunis interstadial in northeastern Europe, beginning at 14 000 yr, marks the end of meltwater delivered into the Dnieper watershed. The Alpine ice domes collapsed about the same time, retreating from the Swiss foreland to nearly their present positions between 14 600 and 14 200 yr (Denton et al., 1999). The Caspian Sea maintained a highstand through the last glacial maximum (Late Khvalynian transgression; Svitoch, 1999) that elevated its shoreline to the point of overflow through the Manysch depression (Chepalyga, 1984; Ferronsky et al., 1998) and allowed migration of Caspian fauna to the Black Sea. With the Caspian spilling into the Black Sea, it seems likely that the latter basin would have swollen to the level of its own outlet. However, the Caspian Sea level dropped

Table 1
Black Sea cores (BLaSON, *Le Suroit*, 1998)

Core name	Latitude	Longitude	Water depth (m)	Length (cm)
BLKS9809	44°05.23′N	30°47.98′E	240	840
BLKS9810	44°04.04′N	30°50.68′E	378	759

below the Manysch spill point (25 m above sea level) prior to 13 000 yr (Svitoch, 1999). The primary sources of freshwater to the Black Sea were therefore greatly diminished before world sea level reached even the deepest conceivable sill between the Aegean and Marmara seas, let alone the gateway to the Black Sea.

The onset of the Bølling–Allerød warm period at about 13 200 yr is marked by an increase in the percent of oak (*Quercus*) pollen, a sign of an expansion of woodlands and a reduction in the steppe vegetation (Traverse, 1974; Zonneveld, 1996). The cool Younger Dryas starting at 11 000 yr marks a return to the steppe environment. There is, however, little evidence to support a readvance of any of the ice caps south of the Baltic during the Younger Dryas cold period that could have resulted in significant enhancement of meltwater delivery to the Black Sea (Denton and Hughes, 1981). The Younger Dryas is followed by a brief return to warm, moister conditions (early Holocene), then cooler, drier conditions again (the so-called 8200 yr BP event[2]; Alley et al., 1993; von Grafenstein et al., 1999). Finally, there is an abrupt shift at ∼7100 yr from herb-dominated to tree-dominated assemblages coincident with the onset of anoxia (Traverse, 1974; Atanassova, 1995). Although there is an evolution in the Black Sea's surrounding vegetation beginning at 15 000 yr, the most pronounced shift accompanies the onset of sapropel deposition (Filipova et al., 1983). Dinoflagellates show a rapid replacement of freshwater (stenohaline) by marine (euryhaline) species with little overlap (Atanassova, 1995; Wall and Dale, 1974; J.-P. Suc, personal communication, 2001), and acritarchs, a planktonic stage of marine plants, show an abrupt 50-fold increase in abundance (Traverse, 1974) at this same stratigraphic level in a core from the eastern basin.

2. Methods

Two sediment cores were recovered by the Kullenburg method of piston coring in 240 and 378 m of water, respectively (Table 1). Navigation was acquired with differential GPS. The cores were positioned along high-resolution sub-bottom profiles used to correlate the strata from one site to the other. Radiocarbon age determinations were carried out at the ETH-Hoenggerberg AMS facility. Shells were leached prior to measurement in order to decrease the effects of diagenetic carbon contamination.

Samples from both cores were analyzed for clay mineral content. After treatment for removal of carbonate, amorphous metal oxides, and amorphous silica, each sample was split into three size fractions (>20, 2–20, and <2 μm), and each size fraction was weighed to determine grain size distribution. Oriented glass slide mounts were made of the <2-μm size fraction. The slides were each scanned three times on the Phillips X-Pert-MPRD X-ray diffractometer, the first a long, continuous scan from 1 to 50° (2θ) with a dwell time of 5 s and a 2θ step size of 0.02. Following the first scan, each sample was treated in a glycolated atmosphere for at least 24 h to expand the smectite interlayers. The last two scans were conducted on the glycolated samples over shorter 2θ ranges to determine the peak offset by expandable clays (smectite) and the ratio of kaolinite to chlorite. The clay mineral composition was calculated using the method of Biscaye (1965). The relative

[2] 8200 yr calendar (BP) ≅ 7500 yr relative to the chronology in this paper.

percentages should be viewed as semi-quantitative results.

Carbonate content was measured using a UIC-incorporated model 5011 CO_2 coulometer attached to a model 5130 acidification module. Carbon and oxygen stable isotopes of the carbonate fraction were measured on a Micromass Optima mass spectrometer with a Multiprep carbonate preparation device. Calibration to Vienna PeeDee Belemnite is via NBS-19 and an in house standard of similar isotopic composition, with a typical range in values for the standards of 0.05 for $\delta^{13}C$ and 0.08 for $\delta^{18}O$ (1σ). Owing to the lack of calcareous foraminifera or mollusks in most of the sediment samples, most isotopic analyses were obtained from bulk sediment samples. The carbonate subfractions (e.g., detrital, biogenic, and inorganic precipitate) were not isolated, and thus the measured isotopic composition is actually a function of three components. Samples were not treated to remove organic matter prior to acidification. Eight representative samples were viewed under a scanning electron microscope (SEM) to determine the type of carbonate material present. Eight of the dated shells were also analyzed for $\delta^{18}O$ and $\delta^{13}C$.

3. Results

The cores examined in this study cover the period from the freshwater stage of the last glacial period to the present. The chronology was established by nine radiocarbon ages measured on individual valves of freshwater mollusks (Table 2). The close similarity in appearance and physical properties of the two cores allowed us to construct a consistent age model for both cores using radiocarbon dates from each and lithologic boundaries as tie-points. We also incorporated the age of the base of the Black Sea sapropel (7160 yr; Jones and Gagnon, 1994) and the first appearance of brackish fauna on the mid-shelf at 8400 yr as additional age constraints. The latter is correlated with the base of a dark green mud just below the sapropel. Although core BLKS9809 has a gap in sedimentation between ∼12 000 and 7000 yr, the deeper water core, BLKS9810, contains an uninterrupted record from 18 000 yr to the present (Table 3). Measurements were focused on the intervals in the cores that show the greatest variability, namely, those representing the period between ∼15 000 and 7000 yr. The relatively stable early period prior to ∼15 000 yr is reflected in the low variability in most of the measured components. All proxies suggest a brief return between ∼11 000 and 10 000 yr toward values that characterized Unit 3 prior to 15 000 yr (homogeneous gray muds).

Sedimentation rates calculated from our age model show decreases at 15 000, 12 800 and 8400 yr (Fig. 3). The wider spacing of dates in the lower half of the cores is likely the cause of some of the disagreement in calculated rates prior to 12 800 yr, although broadly speaking our sedimentation rates are in the same range of magni-

Table 2
Radiocarbon ages of mollusks from cores BLKS9809 and BLKS9810

Core	Analysis #	Depth in core (cm)	Species	Age[a] (yr)	Error (±1σ) (yr)
BLKS9810	ETH-23298	94.5	*Turricaspia*	10 640	80
BLKS9810	ETH-23299	118.5	*Dreissena*	11 410	110
BLKS9810	ETH-23300	154.5	*Dreissena*	12 790	110
BLKS9810	ETH-23301	186.5	*Dreissena*	12 920	100
BLKS9810	ETH-23302	704	*Dreissena*	17 760	130
BLKS9809	ETH-22156	15	*Dreissena*	12 310	95
BLKS9809	ETH-22157	115	*Dreissena*	14 010	100
BLKS9809	ETH-21127	215	*Dreissena*	14 950	100
BLKS9809	ETH-21128	840	*Dreissena*	20 580	150

[a] Conventional radiocarbon years (uncorrected for reservoir).

Table 3
Ages of Black Sea units and subunits

Description	Thickness in cm (BLKS9809/BLKS9810)	Age range (conventional radiocarbon years)
Coccolith ooze (Unit 1)	(20[a]/16[a])	0–3330[b]
Black Sea sapropel (Unit 2)	(27[3a]/23.5)	3330[b]–7160[b]
Neoeuxine muds (Unit 3)		
Gray-green clay (T)	(–/7)	8400–7160[b]
Upper carbonate peak (C1)	(–/42)	10 000-8400
Carbonate trough	(–/38)	11 000–10 000
Lower carbonate peak (C2)	(5/42)	12 800–11 000
Homogeneous gray muds	(42/58)	13 400–12 800
Brown muds (B1/B2)	(195/112)	15 000–13 400
Homogeneous gray muds	(> 580/ > 420)	> 20 600–15 000

[a] Thickness from pilot core.
[b] From Jones and Gagnon (1994); conventional age of total carbonate carbon without 460-yr marine reservoir correction.

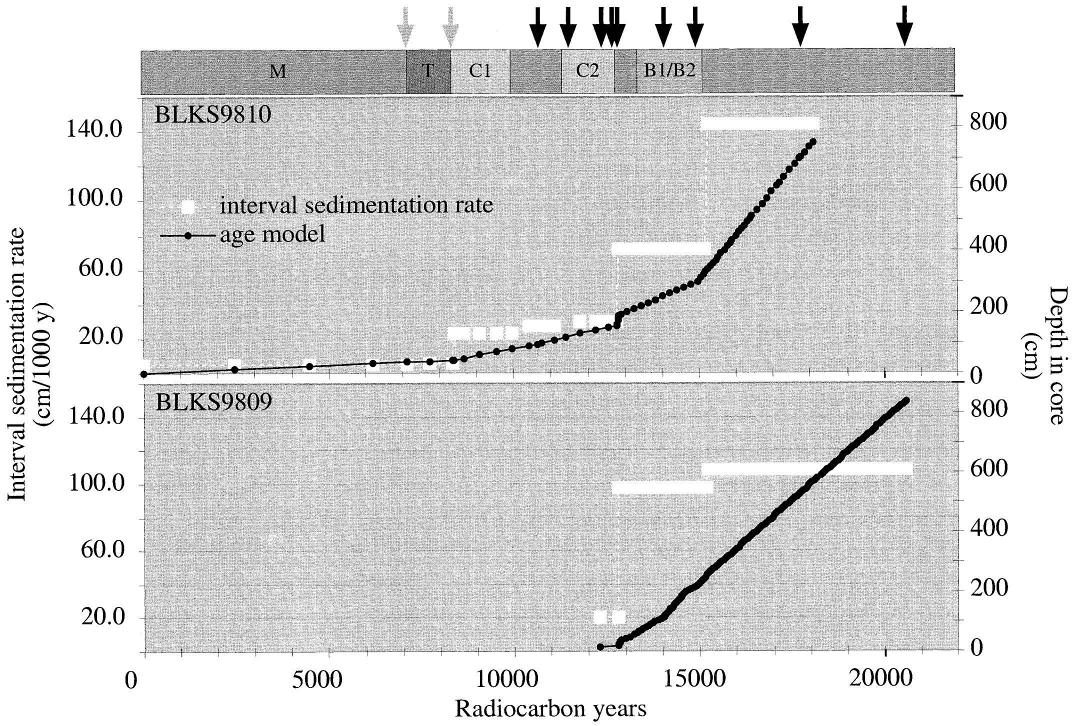


Fig. 3. Sedimentation rates of cores BLKS9809 and BLKS9810, showing marked changes at ~15 000 with the appearance of the brown muds (B1/B2), 12 800 yr with the onset of inorganic calcite formation (C2) and at 8400 yr with the transition (T) from inorganic calcite-rich sediment (C1) to the marine section (M). Interval sedimentation rate is the calculated average for each lithologic unit, whereas the age model is based only on the radiocarbon ages (indicated with black arrows) and independently determined lithologic boundary ages (gray arrow) (see text for discussion).

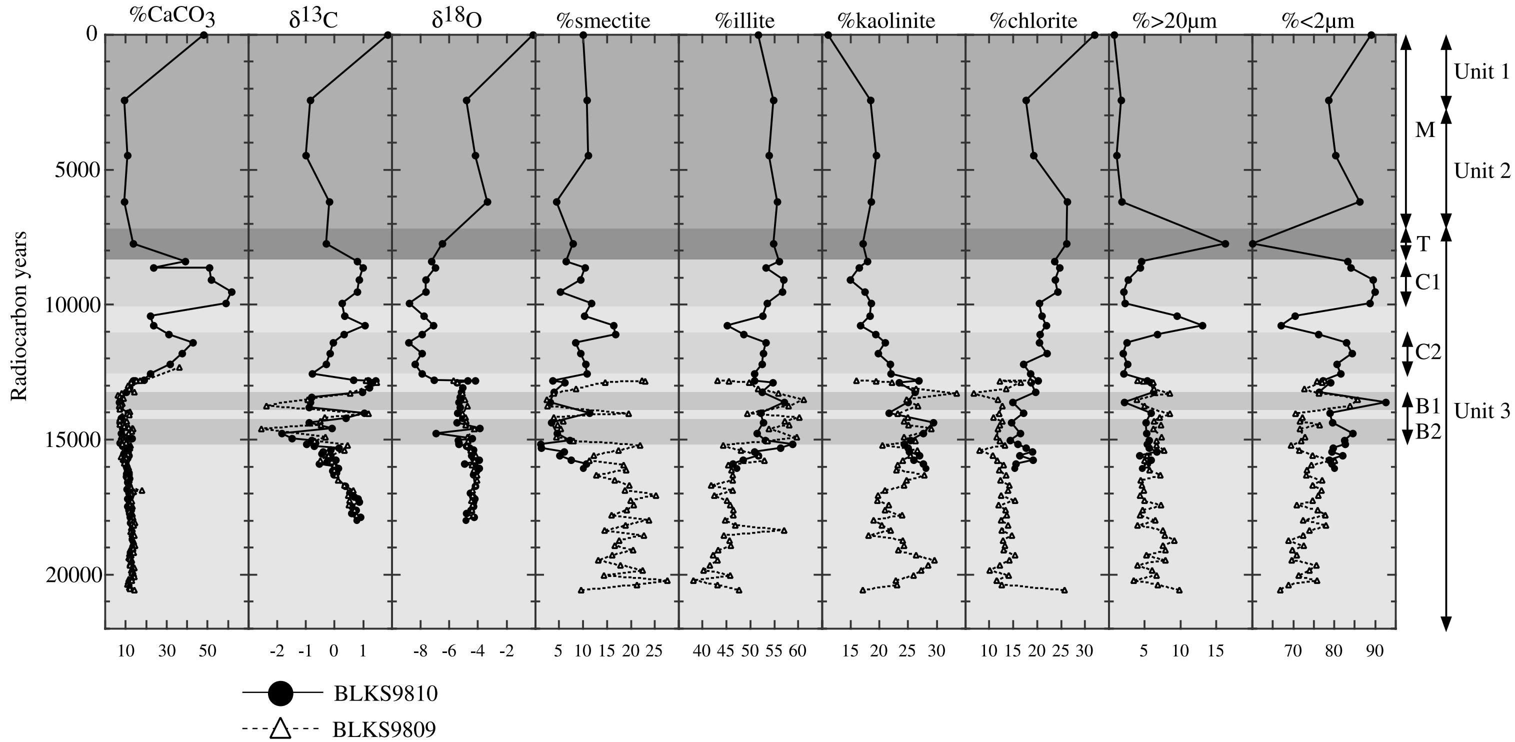


Fig. 4. Summary of core analyses plotted versus age in conventional radiocarbon years (uncorrected for reservoir). 'M' indicates the marine part of the section, including Units 1 and 2 of Ross and Degens (1974). 'T' indicates a dark green mud below the base of the sapropel (Unit 2). 'C1' and 'C2' are the upper and lower carbonate peaks, respectively. 'B1' and 'B2' are the brown clay beds. All scales are in percent, except $\delta^{18}O$ and $\delta^{13}C$, which have permil (relative to PDB) scales.

tude as those calculated by Ross and Degens (1974) for the deep basin cores.

3.1. Lithology

Both cores contain the lithologic units described by Ross and Degens (1974) in cores from the deep basin. These units, their thickness, and age ranges are included in Table 3. The uppermost lithology, Unit 1, is a coccolith-bearing, light olive-gray, organic-rich and finely laminated mud. Unit 2 is a dark olive-gray sapropel. Units 1 and 2 combined are thin (< 40 cm) in the slope setting compared to their thickness in the deep basin (~70 cm). Both represent deposition in the oxygen-depleted conditions of the semi-marine basin and are barren of macrofauna. The lowermost unit recovered, Unit 3, shows variation in color from dark green to light gray to reddish-brown and represents sedimentation in the pre-anoxic basin. The uppermost distinct layer in Unit 3 is a dark green mud (T in Fig. 4) with no apparent sedimentary structures and no macrofauna. Below that is an interval of light gray muds commonly streaked with black sulfides (this interval is further sub-divided on the basis of carbonate content; see below). Below this are homogeneous gray muds, which are interrupted by 2 intervals of brown muds (B1 and B2 in Fig. 4). Above the brown muds the boundaries between the sediment types are gradational, whereas the upper and lower bounds of the brown mud layers themselves are sharp. In the shallower core (BLKS9809) the uppermost part of Unit 3 down to the base of the light gray muds is missing, though the section in the deeper core (BKKS9810) is complete. All of the subunits of Unit 3 except the brown muds contain rare mollusks, usually individual valves of small specimens of the stenohaline species *Dreissena rostriformis* and more rarely *Turricaspia caspia*.

3.2. Carbonates and clays

Calcium carbonate content displays a narrow range of 9 to 13% in the Unit 3 homogeneous gray muds from the base of the cores up to 15 000 yr. Within the brown clays of Unit 3 the carbonate content is lower overall, dipping to 6%. Carbonate content rises gradually from 10 to 15% between 13 400 and 12 800 yr, then increases abruptly to a peak of 42% at 11 400 yr (C2 in Fig. 4). A second more pronounced peak in carbonate content (C1 in Fig. 4), reaching over 60% and lasting from roughly 10 000 to 8400 yr, follows an interval of low carbonate content between 11 000 and 10 000 yr. The carbonate peaks are accompanied by the appearance of euhedral, silt-sized calcite grains (identified both optically and by X-ray diffraction) as a major component of the sediment. Carbonate in the rest of the core is a mixture of reworked coccoliths, mollusk shell fragments, and detrital carbonate grains; none of these components is common, but the biogenic grains are more abundant than identifiable detrital grains, which were only seen once in the SEM scans.

Both cores show an increase in illite and a slight increase in kaolinite in the Unit 3 brown muds (~15 000 yr) and a decrease beginning above the brown muds (~13 400 yr). Smectite is a significant component prior to ~15 000 yr (Unit 3 gray muds) and in the interval between the high carbonate peaks (carbonate trough, ~11 000 to 10 000 yr). Smectite, like carbonate, is low within the brown muds. Unit 1 is low in kaolinite and high in chlorite, and Unit 2 has a clay mineral composition very similar to the upper part of Unit 3.

The grain-size data, which were collected after processing the sediment for removal of carbonate and amorphous silica and metal oxides and thus represent only the detrital component, show two major peaks in the coarse (> 20 μm) fraction within the carbonate trough and the dark green mud (T) at the top of Unit 3. The carbonate peaks and the base of Unit 2 have the highest percentage of clay-size (< 2 μm) detrital grains.

3.3. Stable isotopes

The $\delta^{18}O$ record of the bulk carbonate (Fig. 4) in Unit 3 shows little variation prior to 15 000 yr, ranging between −4.1 and −5‰ and averaging −4.5‰. The values drop to −5.3‰ between 15 000 and 12 800 yr (B1 and B2 and upper homogeneous gray muds), then rise briefly to earlier values of ~−4.5 before plunging to less than −7‰ just after 12 800 yr (C2). Values remain low, with the exception of a small excursion between ~11 000 and 10 000 yr, up to the top of Unit 3. The $\delta^{18}O$ increases across the dark green muds (T), reaches up to −3.4‰ within Unit 2, and is highest, −0.2‰, in Unit 1. The $\delta^{18}O$ of the mollusk shells (Fig. 5) shows an increasing trend starting at 12 800 yr, rising from a baseline value of −7‰ up to −3.1 at 10 640 yr.

The $\delta^{13}C$ values of the bulk carbonate fraction range from −1.8 to +1.4‰ within Unit 3, showing the maximum variation between 15 000 and 12 300 yr (Unit 3 brown muds and transition). There is an overall trend toward higher $\delta^{13}C$ values between 12 300 and 7880 yr, with a short positive excursion between 11 000 and 10 000 yr. The $\delta^{13}C$ of Unit 2 varies from −0.2 to −1.0‰, and reaches the highest value of +1.8‰ in Unit 1. The $\delta^{13}C$ of the mollusk shells displays two steps toward lower values, the first starting at 14 950 yr and the second at 12 800 yr.

4. Interpretation

The relatively constant values in the early part of most of the proxies described above support the idea that the Black Sea experienced a prolonged period of stability through the last glacial period. Sedimentation rates on both the upper continental slope and the deep basin were high. The clay mineral assemblage, which during this period is relatively high in smectite, suggest relatively greater contribution of sediment from the southern drainages than today, since smectite is the dominant clay of the Anatolian provenance. Permafrost in the northern drainage would inhibit erosion and thus limit the amount of clay brought in from that region until thaw. Modern core top sediments show illite and kaolinite dominating the northwestern quadrant of the Black Sea (Muller and Stoffers, 1974). The appearance of these northern provenance clays at ~15 100 yr (with the brown muds) suggests a linkage with the collapse of the Scandinavian and Alpine glacial ice and perhaps the melting of the permafrost. The

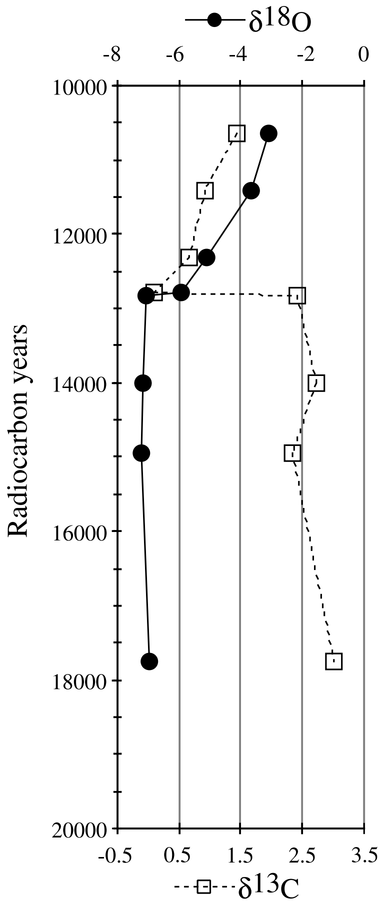


Fig. 5. $\delta^{18}O$ and $\delta^{13}C$ of mollusk shells (same as dated shells listed in Table 2). Both scales are permil (relative to PDB).

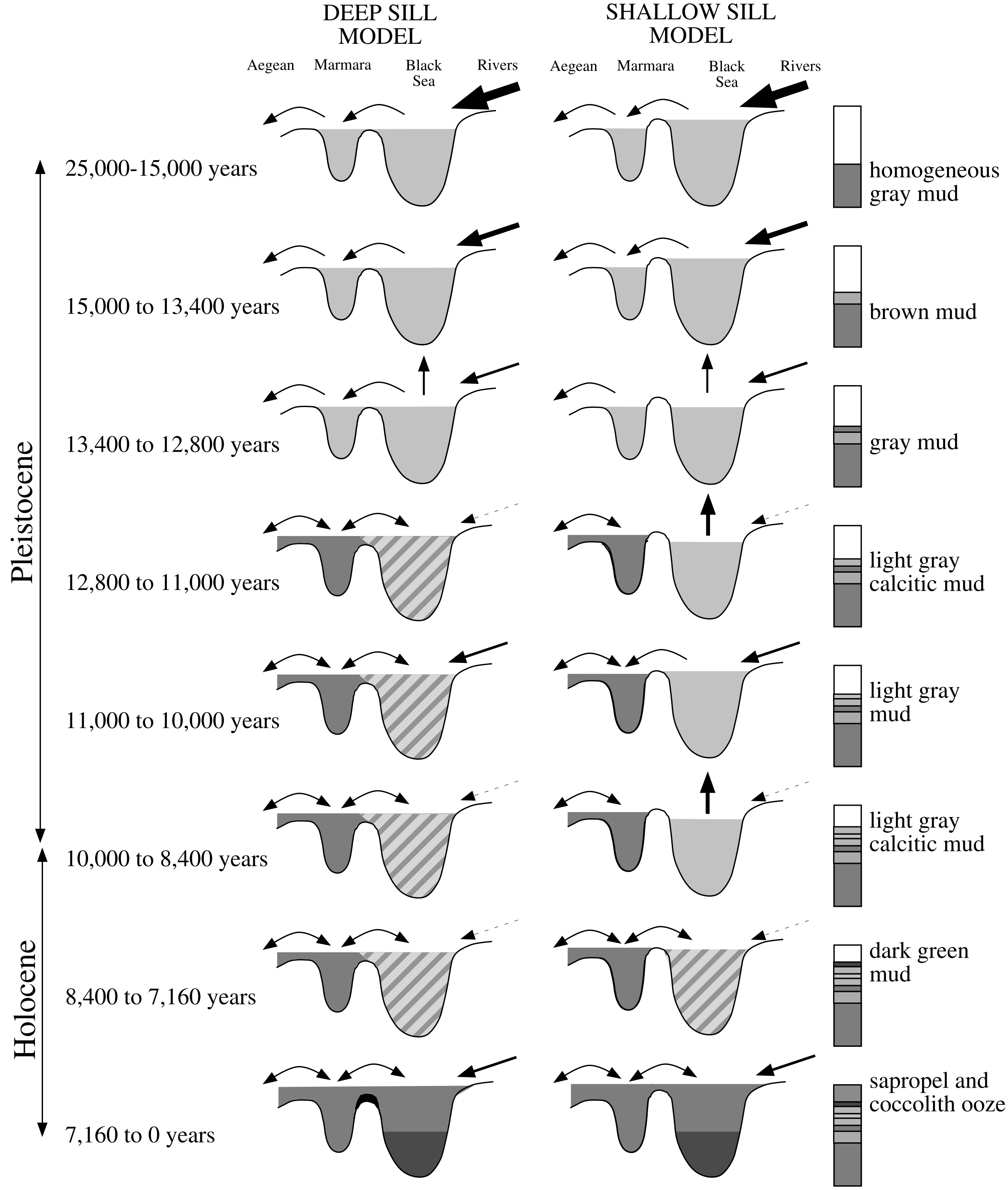
period of higher illite and kaolinite continues through the brown muds and might reflect high erosion rates which persist until vegetation is re-established in the drainage basin with the Bølling warming (Vanderberghe, 1995). Illite and kaolinite then decrease in the overlying gray mud.

There is an important signal on the Romanian slope commencing at 12 800 yr: (1) the sedimentation rate drops sharply by a factor of three to five, (2) the $\delta^{18}O$ of mollusk shells increases, (3) the $\delta^{18}O$ of the bulk carbonate fraction decreases, and (4) the carbonate content of the sediment increases from inorganic calcite precipitation. Soon afterwards marine fauna appear in the Sea of Marmara (Cagatay et al., 2000). It is possible to envision a scenario with a deep Black Sea outlet such that the Black Sea became connected in tandem with Marmara (Fig. 6). In support of such an hypothesis, the drop in sedimentation rate and decrease in grain size would signal a Black Sea coastal transgression concurrent with the rapid global sea-level rise at the time of meltwater pulse 1A (Fairbanks, 1989). Entry of Mediterranean water into the Black Sea might well explain the steady increase in $\delta^{18}O$ of the mollusk shells that begins at 12 800 yr. The light $\delta^{18}O$ of the inorganic calcite may indicate a weakly or seasonally stratified water column brought about by this introduction of denser high-salinity deep water. In this case the surface waters remain fresh, as marine water spills in and settles in the deeper part of the basin because of its greater density. Surface waters would warm in the milder climate of the Bølling–Allerød, which would also push the $\delta^{18}O$ toward lighter values.

On the other hand, mollusk faunal assemblages suggest that salinification of the Black Sea does not begin until 8400 yr (Shcherbakov and Babak, 1979). Could this observation be reconciled with a deep sill model, which would allow for early connection of the Black Sea with the global ocean but a delayed penetration of marine water? One possible mechanism for the delay in the Black Sea's marine signal would be a sufficiently vigorous freshwater outflow through the Bosporus to prevent influx of marine water for several thousand years after reconnection. The vigorous outflow hypothesis has been championed by Lane-Serff et al. (1997) and others, who envisage a strong freshwater flux through the Black Sea and out to the Sea of Marmara and the Aegean. The outflow of freshwater assumes that the hydrologic balance of the freshwater Black Sea was similar to that of the modern marine Black Sea (i.e., that precipitation plus runoff exceeds loss by evaporation). According to the hydraulic models presented in Lane-Serff et al. (1997), a freshwater outflow rate equal to today's would prevent marine inflow (establishment of two-way flow) until the water in the sill itself reaches a thickness of 5 m. World sea level at the time of the first marine signal (8400 yr) was ~ -30 m. An outflow regime similar to today's would then be consistent with a ~ -35-m sill depth. A sill depth of -80 m would require a factor of 10 greater freshwater inflow into the Black Sea to produce a corresponding strong outflow and delay marine input until 8400 yr. Such a vigorous outflow of freshwater from the Black Sea would have inhibited marine inflow through the Dardanelles Sill, but there is clear evidence of marine influence in the Sea of Marmara starting at around 12 000 yr (Cagatay et al., 2000) and persisting until today.

A second scenario that might explain a delayed marine signal despite early connection involves inhibited mixing between the incoming marine water and the freshwater of the Black Sea lake. In this scenario the much more dense marine water would descend to the deep basin upon entry. A density stratification, similar to that found in the modern Black Sea, would prevent mixing between the water masses. Assuming today's rate of marine inflow (312 m^3/yr, Ünlüata et al., 1990), the deep reservoir of the Black Sea lake could be replaced by marine water in roughly 1500 yr. A lower inflow rate, which would be expected because of the smaller cross section of the Bosporus Strait during lower sea level, would make this time longer. However, there are four major flaws in such a scenario. First, there is no evidence of sustained Black Sea stratification before the onset of sapropel deposition at 7100 yr (Jones and Gagnon, 1994). Second, diffusive exchange of salts could occur across the density interface even in the absence of convective mixing, and therefore a model with no mixing between the freshwater and marine layers is not realistic. Third, high bottom chlorinities prior to 8000 or 9000 yr are not supported by models of the pore water chemical gradients (Manheim and Chan, 1974) unless either initial bottom water chlorinity of the Black Sea lake was higher or the diffusion constant in the sediments is greater than the values used in the models (Shishkina, 1966). Fourth, marine water that enters the Black Sea today is rapidly entrained in surface and intermediate layers and does not slip unmixed into the deep basin (Özsoy et al., 1993).

A shallow Bosporus sill would delay the entrance of Mediterranean water until it was breached by a later, higher eustatic sea level. As discussed above, the first appearance of brackish mollusks at 8400 yr is consistent with a sill at



DEEP SILL MODEL
SHALLOW SILL MODEL
Aegean
Marmara
Black Sea
Rivers
Aegean
Marmara
Black Sea
Rivers
Pleistocene
Holocene
25,000-15,000 years
15,000 to 13,400 years
13,400 to 12,800 years
12,800 to 11,000 years
11,000 to 10,000 years
10,000 to 8,400 years
8,400 to 7,160 years
7,160 to 0 years
homogeneous gray mud
brown mud
gray mud
light gray calcitic mud
light gray mud
light gray calcitic mud
dark green mud
sapropel and coccolith ooze

−35 m, assuming effective mixing of entering marine water. Was the Black Sea outflowing over this shallow sill at the time of reconnection? A drop below the outlet sill would result from a shift toward a negative water balance in the Black Sea, i.e., a situation in which the input from rivers and precipitation is less than the export of water by evaporation. A warming climate with increased evaporation upon the conclusion of meltwater delivery at 14 000 yr could produce such a negative water balance. In this case the Black Sea would behave like the Caspian Sea, with highstands during cold intervals and low stands during warm intervals (Chepalyga, 1984; Svitoch, 1999). The increase in $\delta^{18}O$ of the shells at 12 800 yr would result from evaporative concentration of heavy oxygen. The peaks in inorganic calcite at ∼12 000 and ∼9500 yr would have been produced by evaporation leading to supersaturation of surface water. The persistence of the high carbonate up to the base of the dark green mud suggests that the evaporative stage lasted until the breaching of the sill by marine water. The decrease in inorganic carbonate and increase in grain size during the cold Younger Dryas, seen clearly in our cores, result from an increase in runoff. This change is possibly due to a reconfiguration in atmospheric circulation (Florineth and Schluchter, 2000), but is probably also influenced by inhibited evaporation in lower temperatures.

Deposits with Neoeuxine (freshwater) fauna have been reported in depths of −20 to −30 m on the inner shelf off Romania, Ukraine and Russia (Shcherbakov et al., 1978). Although these sediments are not radiocarbon dated, they have been used to infer a rise of the Black Sea lake which flooded most of the shelf prior to a marine connection (Kuprin et al., 1974; Shimkus et al., 1978). Other deposits from the Sea of Azov and the near shore part of the Black Sea shelf that have been radiocarbon dated contain Cardiids more than 9000 yr old (Shcherbakov and Babak, 1979). One of our BLaSON cores (BLVK9814) at a depth of −55 m contains such Cardiids dated at 9580 yr (Lericolais et al., 2002). These fauna are indicative of saline ponds, and their locations would necessarily have been above the Neoeuxine shoreline. Thus the Sea of Azov and the inner Black Sea shelf would have been transformed into a terrestrial landscape while the Black Sea lake shoreline retreated to the outer shelf. In summary, the presence of the freshwater fauna on the inner shelf suggest that the Black Sea lake may have risen to an outflow over a shallow sill at some point prior to marine connection, but then fallen to expose the shelf down to a depth of at least −55 m.

4.1. Implications for the Sea of Marmara

The deep and shallow sill scenarios predict very different histories of outflow from the Black Sea to Marmara after 12 800 yr (Table 4). In the early-entry and deep-sill model, continuous outflow would not produce density stratification until marine water entered the Sea of Marmara ∼12 000 yr ago. If Black Sea outflow to the marine-influenced Marmara is the cause of anoxia in that basin, then 1500 years are required to drop the oxygen levels in the Marmara deep water to levels that inhibit benthic colonization and effectively preserve organic carbon.

The late-entry and shallow-sill model, on the other hand, predicts that Black Sea outflow may have ceased altogether in the intervals from 13 400 to 11 000 yr and from 10 000 to 8400 yr. Outflow could have been significant during the Younger Dryas between 11 000 and 10 000 yr when the Sea of Marmara was already marine. Such outflow could have led to density stratification in the Sea of Marmara, but would have been too early

Fig. 6. Summary of shallow and deep sill models, indicating possible periods of exchange between the Black Sea, the Sea of Marmara, and the Aegean. Curved arrows indicate water exchange over the Dardanelles and Bosporus sills. Straight arrows indicate relative strength of river inflow (thickness is approximately proportional to flow). Lightest shading indicates freshwater, medium shading indicates marine water, and darkest shading indicates anoxic water. Hatched pattern indicates mixed fresh and marine water (i.e., water after two-way connection is established in the Dardanelles/Bosporus). Shaded bars indicate the lithologic unit deposited during each interval.

Table 4
Summary of regional climate indicators and predicted outflow for deep and shallow sill models

Time period[a]	Core lithology (this study)	Sedimentation rate	Authigenic carbonate	Clay provenance	Fauna	Flora	Caspian Sea level[b]	Sapropels	Mediterranean Sea level (end of interval)[c]	Konya Lake (Anatolia)[d]	River input	Evaporation	Outflow-shallow sill model	Outflow-deep sill model
7.16–0	Sapropel	low	none	mixed	marine[e]	woodland	low	Black Sea, Marmara	0 m	gone	moderate	moderate	moderate	moderate
8.4–7.16	transitional gray-green mud	low	none	mixed	brackish	steppe	low	Marmara, Aegean, E. Mediterranean	−16 m	gone	moderate	decreasing	yes	yes
10–8.4	upper carbonate peak (C1)	moderate	high	mixed	fresh	mixed woodland/steppe	low	Marmara, Aegean, E. Mediterranean	−30 m	gone	low	high	no	yes
11–10	light gray mud	moderate	low	south	fresh	steppe	low	none	−59 m	gone	moderate	low	likely	yes
12.8–11	lower carbonate peak (C2)	moderate	high	mixed	fresh	mixed woodland/steppe	falling	none	−68 m	gone	low	high	no	possible
13.4–12.8	gray muds	high	none	mixed	fresh	steppe	falling	none	−100 m	low	decreasing	increasing	no	yes
15–13.4	red muds (B1 and 2)	high	none	north	fresh	steppe	high	none	−110 m	low	decreasing	increasing	yes	yes
25–15	gray muds	high	none	south	fresh	steppe	highest	none	−114 m	high	high	low (cold)	yes	yes

[a] Conventional radiocarbon years (uncorrected for reservoir).
[b] From Svitoch (1999).
[c] From Fairbanks (1989), corrected for marine reservoir age of 400 yr.
[d] From Roberts (1983).
[e] Species of Mediterranean affinity.

to have influenced sapropel formation in the Aegean Sea and the eastern Mediterranean.

5. Conclusions

The Black Sea shows a marked change in sedimentation and oxygen isotopic composition at 12 800 yr. If this change reflects an early connection with the Mediterranean via the Sea of Marmara, the Neoeuxine lake would rise with the world sea level, and outflow via the Bosporus Strait would have remained uninterrupted. The Black Sea would never have stood above the level of the Dardanelles Sill (−85 m) during glacial times and would never have been catastrophically flooded. The delay in the introduction of marine fauna and flora in the Black Sea until 7100 yr requires a gradual salinification to a threshold suitable for these organisms.

If the aforementioned changes happen in a Black Sea isolated behind a shallow sill in glacial and early post-glacial time, the deep shorelines and perched saline ponds require episodes of evaporative draw-down and a transformation to a Caspian mode. The Black Sea cannot have outflowed during such periods. Although the stratification of the Sea of Marmara beginning as early as 10 600 yr favors the deep sill model, it could have been produced by outflow over a shallow sill during the Younger Dryas.

Acknowledgements

We thank Captain Michel Houmard and his officers, crew and technical support team on Le Suroit for the help and courtesy so generously provided during the BlaSON survey. IFREMER funded the expedition with the collaboration of GEOECOMAR in Bucharest. Support for U.S. collaborators was from grant OCE 97-11320 of the US National Science Foundation. We received valuable input from K. Shimkus, W. Pitman, R. Anderson, T. Takahashi, W. Broecker, P. de Menocal, J. Lynch, F. Guichard, I. Popescu and N. Panin. The manuscript also benefitted from reviews by R. Flood and F.Y. Oktay.

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