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STATUS OF THE BLACK SEA FLOOD HYPOTHESIS

William B.F. Ryan
Marine Geology and Geophysics, Lamont-Doherty Earth Observatory of Columbia University, 61 Route 9W, Palisades, NY 10964 USA

Abstract: Exploration of the Black Sea shelf reveals two major shelf-crossing unconformities. The older unconformity separates mostly-barren deposits of late glacial age from overlying Neoeuxinian sediment containing fresh to brackish fauna. This unconformity can be traced over the shelf edge to depths beyond –140 m. The substrate below is dry, firm, and contains unchallenged evidence of subaerial exposure at least to depths of –110 m. The Neoeuxinian cover is present on the outer shelf and is preserved, though incompletely, in depressions on the middle and inner shelf. It is even found as subsurface valley fill in the coastal limans and Sea of Azov. The Neoeuxinian on the shelf represents a transgression leading to a highstand at ~ –20 m below today's sea surface, which was reached by 10,000 BP (uncorrected). Sediments with marine fauna lie above the Neoeuxinian and are separated from it by a sand to gravel layer that represents a younger unconformity. In the limans, the hiatus between the Neoeuxinian and overlying Bugazian is called "peririf" and on the shelf a "washout." Dune fields between –65 and –80 m and wave-truncated terraces with beach-like berms at –90 to –100 m contain shell material dated between 9500 and 8500 BP, suggesting that the younger unconformity represents a post-Younger Dryas regression that took the surface of the Black Sea's lake below the level of the global ocean. Strontium isotopes document the first arrival of saltwater at 8400 BP. Objections to the rapid flooding hypothesis in which Mediterranean water initially poured into a low-lying enclosed lake are centered on the interpretation of the younger unconformity as evidence of either (1) subaerial erosion (and thus a major early Holocene regression) or (2) underwater erosion that does not require a regression. When examined, specific criticisms appear to be based on different interpretations of observations but do not as yet present a concrete refutation of a lowstand of the lake prior to the Mediterranean connection. The flooding hypothesis is today just as vulnerable as when it was first formulated. It serves to best account for the ubiquitous nature of the younger unconformity that not only appears in sediment cores but is also widely mapped by high-resolution reflection profiling. Greater attention needs to be paid in the future to a more comprehensive investigation to find the cause of the younger unconformity.

Keywords: regression, unconformity, shoreline, flood, dunes, pans, Neoeuxinian

1. BACKGROUND

The initial hypothesis of an abrupt saltwater flooding of the Black Sea's ice-age freshwater lake (Ryan *et al.* 1997a, b) was based on the confluence of seven observations obtained by a joint US-Russia-Turkey research collaboration begun in 1993. The first observation was a shelf-wide unconformity visible in high-resolution reflection profiles. The second was the presence of a uniform drape of sediment that begins simultaneously above the unconformity and that reveals practically the same thickness over nearby elevations and depressions while displaying no visible indication of coastal-directed onlap across the outer and middle shelf. The third was the presence of submerged shorelines with wave-cut terraces and cliffs, beach berms, offshore bars, and coastal dunes at depths between –70 and –120 m, elevations that lie below any known outlet sill to the global ocean (either the Bosporus or Dardanelles Straits). The fourth was the mapping of meandering river channels capped by the unconformity and extending seaward across the shelf to the vicinity of the –100 m isobath. The fifth was the recovery of strata immediately below the unconformity consisting of dense, low-water content mud containing desiccation cracks, plant roots, and sand lenses rich in freshwater molluscs (*Dreissena rostriformis*) with both valves still attached and coated with algal scum. The sixth was the appearance at the base of the mud drape of euryhaline molluscs and dinocysts that replace the fresh to brackish fauna and flora. The seventh was the measurement of stable isotopes that show light $\delta^{18}O$ (–6‰) in the sediment below the unconformity and a heavy value (as high as 1.1‰) above the unconformity.

The flood hypothesis raised considerable controversy and initiated much refutation (Görür *et al.* 2001; Aksu *et al.* 2002a, b; Hiscott and Aksu 2002; Hiscott *et al.* 2002; Yanko-Hombach *et al.* 2002; Yanko-Hombach and Tschepaliga 2003; Kaplin and Selivanov 2004; Chepalyga, this volume). As one critic (V. Yanko-Hombach) stated in her oral presentation at the annual meeting of the Geological Society of America in Seattle (2003), "It is impossible that such an event could have been missed by decades of Soviet research."

2. PRIOR AND SUBSEQUENT OBSERVATIONS

Did Soviet researchers, in fact, miss this event? As reprinted in Ryan *et al.* (1997b:Figure 2), Kuprin *et al.* (1974), Shcherbakov *et al.* (1978, 1983), and Kaplin and Shcherbakov (1986) had already documented the lowstand shoreline. At the time of the lowstand, the entire Sea of Azov was a terrestrial landscape with the mouth of the Don River 50 km south of the Kerch Strait. Coring and echo-sounding profiles had identified a littoral zone near the Black Sea shelf edge that extended along an offshore strip from Romania to the Caucasus. Dozens of cores penetrated the erosion surface, substantiating the substrate as either an alluvial, fluvial, aeolian, or Neogene outcrop. The Soviet researchers interpreted the presence of ancient river valleys traversing the shelf and confirmed the shoreline position with the recovery of sand, gravel, and freshwater molluscs typical of the coastal zone. Sand and gravel in the thalweg of an entrenched valley of the paleo-Don River contained fluvial gastropods (*Viviparus viviparous*). They were sampled from –62 m beneath the bottom of the Kerch Strait, which connects the Black Sea to the Sea of Azov (Popov 1973; Skiba *et al.* 1976). Semenenko and Sidenko (1979) charted this valley upstream across the floor of the Sea of Azov and offered interpretations of river confluences based drill cores calibrated with carbon-14 measurements.

Ostrovsky *et al.* (1977a) recognized the extensive down-cutting of coastal river valleys as evidence of a major water-level drop within the Black Sea's ice-age lake. Some of these entrenched river valleys continue across the shelf, reaching depths between –93 to –122 m (Ostrovsky *et al.* 1977b). Although the Soviet researchers had not published reflection profiles to document the exposed margin of the lake, their numerous piston and drill cores confirmed the ancient coast as once lying well beyond the –80 m isobath (Federov 1978; Balabanov and Izmailov 1988; Kaplin and Selivanov 2004). The lowstand shorelines prompted the US-Russia-Turkey team to examine the river valleys in more detail in 1993 with reflection profiling to search for coastal deltas at the lake edge. Our achievement is based on the prior Soviet investigations, and we are indebted to the Russian scientists who conducted them.

Objections to the flood hypothesis have raised little issue with the submerged shorelines. All subsequent reflection profiling has found the same shelf-wide erosion surface on the Romanian (Popescu *et al.* 2004; Lericolais 2001; Lericolais *et al.* 2003, this volume); Bulgarian (Genov *et al.* 2004; Coleman and Ballard, this volume), and Turkish margins (Okyar *et al.* 1994; Demirbağ *et al.* 1999; Okyar and Ediger 1999; Aksu *et al.* 2002b; Algan *et al.* 2002; Ergin *et al.* 2003; Algan *et al.*, this volume).

The objections focused on the timing, rate of submergence of the exposed margin, and the issue of a continuous connection with the Mediterranean. Until the hypothesis of Ryan *et al.* (1997a, b), the consensus of Black Sea researchers was that the lake's surface had risen in pace with global sea level via a relatively early connection through the Bosporus Strait (Shcherbakov 1982, 1983). Based on hydrologic considerations, Kvasov (1975), Kvasov and Blazhchishin (1978), and Chelpalyga (1984) had stipulated that outflow from the Black Sea through this strait had always been continuous, even at maximum lowstand conditions. For this to be the case, the lowstand shoreline had to be a measure of the level of the lake's outlet. However, as more became known about the shallowness of the Bosporus sill, Chepalyga (1995) used the suggestion of Pfannenstiel (1944) to place the lake's outlet in the Sakarya River valley, connecting what has been called the *Sakarya Bosporus* to the eastern arm of the Izmit Gulf, and from there across the Sea of Marmara to the Aegean Sea. The sill ultimately controlling the Black Sea lake level would therefore have been the Dardanelles bedrock at –85 m.

A deep outlet would permit inflow of Mediterranean water shortly after the connection of the Mediterranean with the Sea of Marmara around 12,000 BP. Such a deep connection would support the idea that the Neoeuxinian epoch between 18,000 and 9000 BP was a prolonged period of rising sea levels after a late glacial lowstand (Kaplin and Selivanov 2004). However, Major *et al.* (2002) and Myers *et al.* (2003) point out that although the salinity increase in the Sea of Marmara, as determined from the mollusc assemblage and stable isotopes (Çağatay *et al.* 2000; Sperling *et al.* 2003), started immediately after global sea level rose above the –85 m Dardanelles sill, salinity increase in the Black Sea was delayed for more than 3000 years (Deuser 1972, 1974; Wall and Dale 1974; Shcherbakov and Babak 1979). Although a vigorous outflow from the Black Sea could keep out Mediterranean saltwater, as argued by Lane-Serff *et al.* (1997), their hydraulic models prevented Mediterranean inflow only up to the moment when sea level rose 5 m above the sill. Global sea level at the time of the first marine signal in the Black Sea at 8400 BP was ~ –30 m, or more than 50 m above the deep inlet sill proposed by Chepalyga (1984, 1995). If the Black Sea outflow though a deep connection was truly so vigorous and persistent, it remains to be explained how this outflow could have permitted the early and sustained salinification of the Sea of Marmara at the downstream end of the water cascade. Thus, a number of researchers have recently rejected the hypothesis of a deep Black Sea outlet (Major *et al.* 2002; Myers *et al.* 2003; and Bahr *et al.* 2005).

A shallow outlet makes the Black Sea lowstand shorelines even more remarkable. Since the wave-cut terraces at –110 m off the Ukrainian coast (Ryan *et al.* 1997b), the littoral deposits at –122 m (Dimitrov 1982), and the –155 m beach off Sinop (Ballard *et al.* 2000) are beyond the limit of any realistic post-transgression subsidence, it seems necessary to consider the reality of interrupted outflow. Indeed, Soviet researchers had already raised the possibility that steady outflow was only a special characteristic of the Würm glaciation when the Black and Azov Seas would have had a strong positive moisture balance. Long before the catastrophic flooding hypothesis (Ryan *et al.* 1997a, b), Ivanov and Shmuratko (1983) had already proposed that, during interglacial warming, the level of the Black Sea's lake dropped below its outlet until the negative moisture balance was overcome by the inflow of Mediterranean salt water.

An enclosed lake is confirmed by isotopic evidence (Nikolaev 1995; Svitoch *et al.* 2000, and Major *et al.* 2002). Although intervals of enclosure may have been of relatively short duration, the lake level would have been dynamic during isolation and controlled only by the balance of evaporation versus inflow from rivers and precipitation. One expression of an enclosed lake is the lack of a stable surface. Witness the extreme regression in the Caspian Sea to –133 m (Chepalyga 1984; Svitoch 1999). Lake-level fluctuations might also account for the observed repetition of ‘cut and fill’ in the sediments of the river valleys that cross the shelf (Esin *et al.* 1986; Ryan *et al.* 2003; Popescu *et al.* 2004) as well as laterally-continuous wave-cut terraces at numerous levels from –44 to –121 m (Shimkus *et al.* 1980).

Another expression of an enclosed lake is increasing salinity as evaporation proceeds. The mollusc *Didacna moribunda*, found in the lowstand deposits, is thought to be an indicator of such increasing salinity (Chepalyga 1984) as is the appearance of the more brackish-tolerant *Dreissena polymorpha* and *Monodacna caspia*, which replaced the freshwater *Dreissena rostriformis* (Shcherbakov and Babak 1979). The concentration of solutes also leads to eventual precipitation. Authigenic calcite precipitation of calcareous mud of the “Seekreide” type appears following the deglacial meltwater delivery (Major *et al.* 2002; Ryan *et al.* 2003; and Bahr *et al.* 2005) and persists, except for an interruption during the Younger Dryas, until the eventual connection with the Mediterranean.

Krischev and Georgiev (1991) note “a drastic change of the sedimentation environment” coincident with the establishment of the connection with the Mediterranean though the Bosporus. They attribute this change to the “fast raising” of the water level during the transition from lacustrine to marine conditions. The change corresponds to a stratigraphic break (“washout”) in the cores that interrupts the lacustrine calcite precipitation and is followed by terrigenous mud with marine molluscs. They report the “washout” in more than 100 cores and propose that it is a regional occurrence across the entire western Black Sea. The “washout” continues from the shelf edge down to the basin floor, where it is certainly subaqueous in origin. Although they suggest that the cause could be slumping from earthquakes, they prefer hydrologic phenomena associated with the introduction of Mediterranean water. One mechanism is internal waves spawned on the interface between overlying freshwater and underlying saltier water. Internal waves on this density gradient would break against the slope while inflowing saltwater displaced the lake water upward and outward through the Bosporus (Lane-Serff *et al.* 1997). Calvert (1990) and Calvert and Fontugne (1987) discussed this freshwater flushing as a mechanism to lift nutrients to the surface to enhance productivity and eventually cause anoxia and sapropel deposition. There is a thin layer of precipitated aragonite at the sapropel base (Degens and Ross 1972; Jones and Gagnon 1994) that forms in calcite-saturated lake environments into which there has been an introduction of marine sulfate.

The second of the initial observations in formulating the flood hypothesis—a uniform mud drape above the unconformity—has also been found on other Black Sea margins (Algan *et al.* 2002; Lericolais *et al.*, this volume). In every case where this drape is resolvable in high-resolution reflection profiles and is not obscured by a long-duration sonic pulse, its thickness, when calculated from acoustic travel time to meters, corresponds in cores to the layer of terrigenous mud containing marine molluscs such as *Mytilus galloprovincialis*, *Mytilaster* (also known as *Mytilus*) *edulis*, *Cerastoderma edule*, and *Cardium edule* (Nevesskaya and Nevessky 1961; Nevesskaya 1965; Nevessky 1967; Kuprin *et al.* 1974; Shcherbakov *et al.* 1978; Shimkus *et al.* 1978; Shcherbakov 1979; Dimitrov 1982; Filipova *et al.* 1983; Shopov *et al.* 1992; Major *et al.* 2002; Lericolais *et al.*, this volume; Algan *et al.*, this volume). This lithologic and biostratigraphic interval on the shelf corresponds to the Bugazian, Old Black Sea, and New Black Sea stages of Soviet nomenclature and correlates with Units 1 and 2 in basin sediments as defined by Ross *et al.* (1970). Wall and Dale (1974) and Mayard (1974) investigated this interval in the basin cores. They report a replacement of stenohaline dinoflagellates and diatoms by euryhaline species at the contact between Unit 2 (sapropel without carbonate) and the subjacent Unit 3 (mud containing fine-grain calcite). The stratigraphic age equivalence of the basin Unit 2 (sapropel) and Unit 1 (sapropel with the calcareous nannofossil *Emiliana huxeli*) and the shelf sequence of the Bugazian, Old Black Sea, and New Black Sea stages is widely used and adopted by most researchers.

The third of the initial observations in formulating the flood hypothesis—shorelines, wave-cut terraces, beach berms and coastal dunes between –70 and –120 m—has already been partly discussed. These features have been richly described and documented in some detail (Ryan *et al.* 2003; Lericolais *et al.*, this volume). Most notable are the small depressions among linear sand ridges between the –64 and –80 m isobaths on the Danube shelf discovered during the 1998 BLASON expedition. The sand ridges are 4 to 5 m in relief with an average spacing of 750 m. They strike almost uniformly at an azimuth of 75±10°. The ridges are asymmetrical in cross-section with steeper sides facing to the southeast (Figure 1).

The ridges have a length to width ratio exceeding four. In addition to these features of positive relief, dozens of depressions with diameters from 100 to 1800 m and a negative relief of 3 to 9 m populate the southern half of a 100-km^2 corridor surveyed with multibeam sonar. The outer Ukrainian shelf also contains asymmetrical linear ridges with a similar orientation to those on the Romanian shelf (Ryan *et al.* 1997b), however, they have reduced heights of 1 to 2 m and reduced wavelengths of 150 to 250 m. They are distributed in a belt from 1 to 2 km in width lying between the present –70 and –80 m isobaths. The interiors of the ridges contain foreset-type clinoforms that dip steeply to the southwest and indicate a migration of the ridges in that direction.

Sampling into the interior of a ridge (BLKS9838) recovered sand rich in opaque heavy minerals and shell fragments. The minerals include quartz,

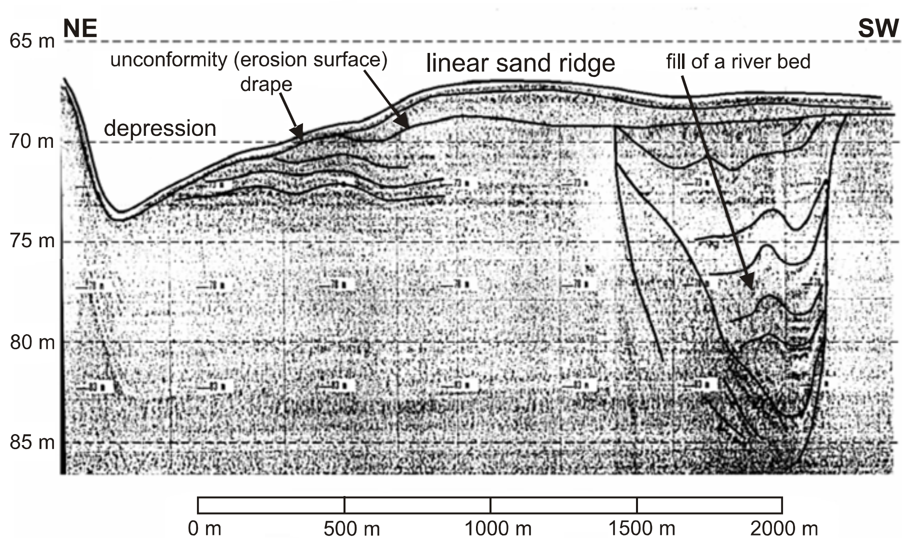

Figure 1. Reflection profile over a linear sand ridge showing its location above the unconformity and filled river channels below the unconformity. The profile was obtained on the BLASON expedition with the *R/V Suroît* using equipment provided by N. Panin.

garnet, and ilmenite. The shell fragments belong to the freshwater mussels of the genus *Dreissena*. Cores into the sediments on which the dunes have formed consist of shell-bearing sand and firm silty red and brownish clay with thin lenses containing specimens of *Monodacna* sp., dated at 9580 BP. Specimens of intact *Dreissena* within a sand matrix from material obtained at the top of a ridge have ages of 8360 BP for core BLKS9837 (–68 m of water depth) and 8275 BP for core BLKS9838 (–77 m of water depth).

The depressions (Figure 2) have a variety of configurations from nearly circular to kidney shaped. Depths of individual depressions are greatest at the base of their northeastern walls, and they shoal to the southwest. The shoaling slopes display terraces separated by low (< 1 m) scarps. Backscatter reflectivity is greatest on the northeast-facing wall of the depressions and on terrace scarps, whereas the steepest southwest-facing edges of the depressions are less reflective. Some of the smaller depressions appear to be strung together like pearls on a necklace, and a lesser number of the larger depressions connect together through shallow conduits. In the center of the surveyed corridor, some depressions align in troughs between the linear ridges.

The depressions are cavities cut into the shelf-wide erosion surface upon which the ridges have grown and migrated. The scarps that separate the terraces on the depression floors are located where bedding planes intersect the erosion surface. Beneath the erosion surface, one finds buried meandering channels with point-bars. Many of the small depressions appear to align along former braid plains (Popescu *et al.* 2004).

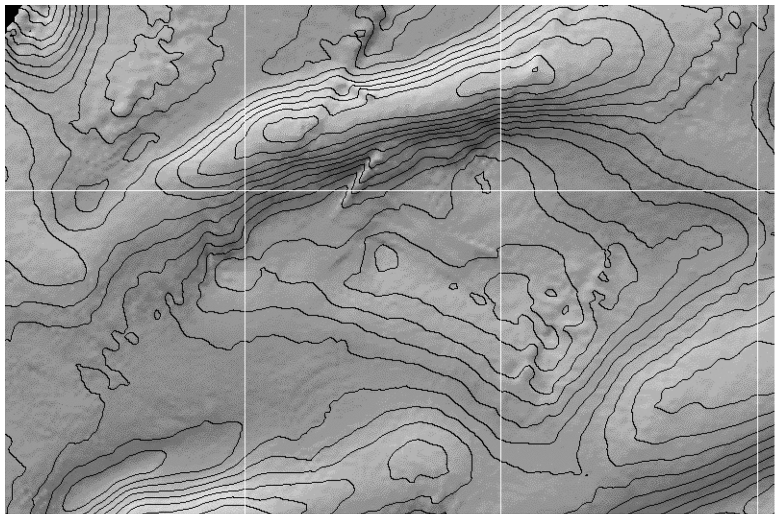

Figure 2. Shaded-relief map of linear sand ridges and depressions near 44° 04' N and 29° 59' E within the dune field on the Romanian shelf. The multibeam data were provided from IFREMER by G. Lericolais and processed by William F. Haxby. Distance between vertical lines is 1500 m, and the contour interval is 0.5 m. The floor of the central trough is at –71 m.

As an ensemble, the linear ridges and depressions are features of the terrestrial windswept landscape (Shaw and Thomas 1997). The cavities are eroded by wind deflation. The groundwater table limits the depth of the depression (Laity 1994). At the scale of those mapped on the Romanian shelf, the depressions are called pans. Pan initiation and growth depend on materials susceptible to deflation, such as poorly consolidated clay-rich material that curls, flakes, and blows away upon desiccation. Pans in dune settings may occur as a string of depressions aligned along a former river course and its braid plain. Pans often transform into ponds during wet phases, generally from groundwater recharge (Lancaster 1998). Pond and marsh sediments have been reported from between the dunes of sand seas during intervals of substantially increased moisture (Lancaster 1994). Present-day ponds are reported in the desert seas of northwestern China (Wang *et al.* 2002). In coastal domains, pans correspond to blowouts formed when onshore winds erode gaps in a single foredune or a series of beach ridges (Giles and McCann 1997). Leakage of sand from the beach to the interior occurs mainly through erosion of the dune front by storms or by landward movement of sand through blowouts or parabolic dune migration (Carter *et al.* 1990a, b), or combinations of the above.

The Romanian and Ukrainian fields of linear sand ridges and depressions are located foreshore of the bluff delineating the paleo-shoreline at –100 m of water depth. The height and spacing of the Black Sea linear ridges are representative of the coastal aeolian population (Lericolais *et al.* 2003, this volume).

The fourth of the initial observations—meandering river channels capped by the unconformity—has been subsequently observed as ubiquitous across the Romanian shelf (Popescu *et al.* 2004). Like those on the Ukrainian shelf (Ryan *et al.* 1997b:Figures 3 and 4), the channels extend right to the paleo-shoreline and pass under the belt of coastal sand ridges and depressions. These channels are invariably filled to the brim with thalweg and point bar deposits, which themselves have been beveled by a subsequent phase of erosion that includes wind deflation. Consequently, the regression that exposed the shelf surface into which the river channels were cut was followed by a transgression that led to the filling of the channels and then to another regression that deflated the channel fill and re-exposed the entire region to coastal dune and pan development.

The fifth observation of dense, dry mud below the erosional unconformity has since been reported on the Thrace margin by Algan *et al.* (this volume) in cores 4, 9, and 20 from the shelf edge. They note "a marked contact" between a 2-cm thick shell-enriched layer and a "stiff clay deposit with low water content at the base of these cores." The ^{14}C age of the shells (*Dreissena* sp.) is 8590±145 BP, comparable to the age of the shell material on top of the sand ridges on the Danube shelf. The authors write that "the lithological characteristic of this core indicates that the deposition starts with high-energy condition over the stiff eroded substrate at about –100 m, and continued with low-energy, suggesting a rapid deepening of a shallow environment." This firm clay with bulk densities in the range of 2.0±0.1g/cm^3 was also recovered near the shelf break on the Romanian margin in several cores (BLKS9801, BLKS9804, BLKS9806, BLKS9807, and BLKS9808) from the BLASON Expedition with ^{14}C dates in samples taken right below the erosion surface ranging from 10,250 to 24,160 BP (Major 2002). The dry substrate at a depth of –76 m also underlies the previously-described field of coastal dunes and pans as sampled in core BLKS9834. The physical properties of many of the BLASON cores were measured with a continuously-moving sensor track. In every core measured, the contact between overlying water-saturated mud and underlying stiff clay was abrupt and occurred in the span of 1 to 2 cm. The only measured cores without the dense substrate were those taken in water depths below –160 m.

The sixth observation—euryhaline molluscs and dinocysts at the base of the mud drape—has been widely reported elsewhere on the Russian (Nevesskaya 1965; Shcherbakov *et al.* 1978; Shimkus *et al.* 1978; Shcherbakov 1979; Shcherbakov and Babak 1979), Ukrainian (Kuprin *et al.* 1974; Semenenko and Sidenko 1979), Romanian (Lericolais *et al.*, this volume), Bulgarian (Khrischev and Shopov 1978; Dimitrov *et al.* 1979; Dimitrov 1982; Filipova *et al.* 1983; Shopov *et al.* 1986; Atanassova and Bozilova 1992; Shopov *et al.* 1992; Atanassova 1995), and Turkish margins (Görür *et al.* 2001; Aksu *et al.* 2002a; Algan *et al.*, this volume). Among the oldest reported ^{14}C dates on an individual valve of *Mytilus* sp. are 7770±70 BP in core MAR00-06 on the southwestern shelf (Aksu *et al.* 2002a), and 7415±115 BP from an 8-cm thick sandy shell layer overlying with a sharp contact the stiff clay described above near the entrance to the Bosporus Strait (Algan *et al.*, this volume). The oldest reported date for an individual shell of *Cardium* sp. is 7140±40 BP in core AK08-93 sampled just above the shell hash that separates the mud drape from the dry, firm clay on the outer Ukrainian shelf (Ryan *et al.* 1997b). Carbon 14 dates from bulk samples with mixed marine and lacustrine fauna often give older ages. Semenenko and Sidenko (1979) report ages of 7810±110 and 9100±130 BP for the base of the mud drape in the Sea of Azov for cores 23 and 8, respectively. The mollusc assemblage in the latter sample, however, is dominated by brackish species such as *Monodacna caspia*, *Adacna vitrea*, and *Dreissena polymorpha*. Yet, Semenenko and Kovalyukh (1973) present a ^{14}C age of 9280±200 BP for specimens of *Cardium edule* from a depth of –18 m in the Sea of Azov. For this specimen to indicate a marine connection as the authors propose, global sea level would have to have been at that height to deliver water to the Sea of Azov. Clearly, sea level in the global ocean was not that shallow and had not yet reached –40 m by 9000 BP (Siddall *et al.* 2003). As discussed by Major (2002), other specimens of *Cardium* and *Adacna* dated at 9850±90 BP have been found at shallow depths on the Romanian margin. Their position above the contemporaneous global sea level suggests that they are fauna from saline ponds or limans that were located landward of the shoreline of the Neoeuxinian lake. The oldest reported date on truly marine samples rich in *Mytilus* and *Chione gallina* in the Sea of Azov is 6200 BP. In the limans of Ukraine, the date of onset for the mud drape (i.e., the Bugazian stage of the stratigraphic interval called Q^1_{IV}) is placed at 8500 BP with the first euryhaline marine fauna (*Mytilus* and *Cardium* sp.) arriving after 7500 BP (Gozhik 1984:Table 2).

The seventh of the initial observations — a shift in the stable isotopes— has been advanced significantly in the PhD dissertation of Major (2002) and subsequent publications (Major *et al.* 2002; Ryan *et al.* 2003, Bahr *et al.* 2005). If one looks at an upper water layer proxy such as the $\delta^{18}O$ of fine-grain bulk carbonate as pioneered by Deuser (1972, 1974), the shift from light (–6‰) to heavy (+1.0‰) takes place abruptly at 8400±100 BP (Major *et al.* 2002). However, if one examines the benthic fauna, such as the mollusc species *Dreissena rostriformis*, the shift from –6.0‰ starts much earlier, around 14,000 BP in the lacustrine phase of the Black Sea. The shift in $\delta^{18}O$ starts at the same time as the onset of the first episode of calcite precipitation corresponding to the oldest of the carbonate peaks in cores from the slope and basin floor (Krischev and Georgiev 1991; Major *et al.* 2002; Bahr *et al.* 2005). Since the shift from light to heavy values cannot be a temperature effect of post-glacial warming (which should lighten the values), it must be either the input of a new source of water via precipitation and/or the effect of evaporative fractionation (Major *et al.* 2002). If the $\delta^{18}O$ shift measured in mollusc shells was exclusively a response to a new source of water, the shift would be expected to have occurred earlier, around 15,000 BP. This is the time of the first substantial delivery of meltwater from the northern ice sheets, which is recorded in the abrupt isotopic shift of $\delta^{87/86}$Sr (Major 2002; Ryan *et al.* 2003). Thus, the oxygen isotope measurements can be interpreted as a signal of evaporative fractionation during the two pulses of calcite precipitation. The return to the lighter glacial values during the trough between the first two carbonate peaks would happen when evaporation decreased during the Younger Dryas and the lake level rose to spill through its outlet to the Mediterranean. The δ^{18}O in the mollusc shells becomes suddenly heavier after 8400 BP, when the strontium isotopic composition of the Black Sea water also abruptly shifts to the global ocean value. Thus, the positive shift of δ^{18}O to near modern values beginning at 8400 BP is most certainly a compositional effect of arriving Mediterranean seawater, and neither a Holocene cooling event in post-glacial time nor further evaporative fractionation.

3. BLACK SEA CHRONOLOGY

A substantial number of well-described cores exist that have been ^{14}C-dated by Accelerator Mass Spectrometer (AMS). Consequently, it has been possible to assemble a useful Black Sea lithostratigraphic chronology reaching back to 25,000 BP. The chronological framework is built from observations of lithology (sediment composition, grain size, visual descriptions, and color changes), biostratigraphic observations (faunal and floral assemblages, including pollen), systematic isotopic variations, and absolute dating. For the ^{14}C dates, ages are in raw carbon-14 years before 1950, and, except when specifically stated, these dates are neither corrected for reservoir age nor calibrated to tree rings. By using raw values, comparisons can be made with ^{14}C dates published long before corrections and calibrations were applied. Furthermore, we do not have direct knowledge of reservoir ages prior to 1931– ~460 years as given by Jones and Gagnon (1994)–so any application of such corrections risks using assumptions that may not be valid.

The following are key lithostratigraphic descriptors from many authors (Kuprin *et al.* 1974; Krischev and Shopov 1978; Shopov *et al.* 1986; Krischev and Georgiev 1991; and my own experience with the *R/V Aquanaut* and BLASON cores) working from the glacial period though the Neoeuxinian stage: dark bluish-grey mud, dark grey mud with iron sulfides (hydrotriolite streaks), light brown and brown-red muds, light grey carbonate rich mud, dark grey mud again with iron sulfides, followed upward by light grey carbonate-rich mud with increasing grain-size. It is regularly observed that the late Pleistocene sediments are found along the shelf edge or are preserved in depressions within the shelf (facies N, L, NL, and P of Shopov *et al.* 1986). The sediments along the shelf edge are often of a littoral composition, accumulated in "bars near the shoreline of the early Neoeuxine Sea" (Shopov *et al.* 1986). The dark grey mud sometimes outcrops on the seabed near the shelf edge. On the shelf, the late Pleistocene sediments are terminated at their upper boundary by a "washout surface."

Only beyond depths of –150 m does one recover continuous sequences of late Pleistocene sediments, though rarely even on the middle and lower slope. One well-described core with apparently continuous deposition is A96, lifted from –630 m (Krischev and Georgiev 1991). It is the basis for correlation with BLKS9810 (Major *et al.* 2002) taken at –378 m depth and GeoB 7608-1 recovered from –1202 m (Bahr *et al.* 2005).

There is a succession of ten distinctive intervals (Figure 3) that can be recognized in each core:

(1) dark grey mud—G—followed by
(2) up to four distinct intervals of brown-red mud—R4-R1—interbedded with grey mud;
(3) dark grey mud with hydrotriolite streaks—H2;
(4) light grey carbonate rich mud—C2 of Major *et al.* (2002) or C3 of Bahr *et al.* (2005);
(5) dark grey mud with black hydrotriolite streaks—H1;
(6) light grey carbonate-rich mud—C1 of Major *et al.* (2002) or C2 of Bahr *et al.* (2005);
(7) grey-green silty mud—T of Major *et al.* (2002);
(8) light-speckled carbonate-bearing grey mud—C1 of Bahr *et al.* (2005);
(9) dark green carbonate-free sapropel—Unit 2 of Ross and Degens 1974); and
(10) a light green carbonate-rich sapropel banded with light grey coccolith-rich laminae—Unit 1 of Ross *et al.* (1970).

The intervals of dark grey mud with hydrotriolite streaks (H1 and H2) are coarser-grained with more than 10% of the material > 20 μm and 3% > 63 μm in size. These intervals also have high $\delta^{13}C$ isotopic compositions (> 1‰), whereas the brown-red mud intervals have the lowest values (< –2‰).

Bahr *et al.* (2005) have correlated the brown-red intervals to periods of warmer conditions recorded in the beginning of the post-glacial isotope record of the GRIP ice core from Greenland commencing at 17,900 calBP (calendar age). They have also used the interpretations of Major *et al.* (2002) and Ryan *et al.* (2003) to match the H1 carbonate trough with the Younger Dryas cold period in the same GRIP record. This correlation locates the earliest carbonate peak in the Bølling-Allerød warm period and the second peak in the warming trend of the earliest Holocene. The H2 dark grey clay with hydrotriolite streaks falls in the Oldest Dryas cold stage at 15,500 to 14,500 calBP in the GRIP calendar age chronology. The youngest carbonate peak (C3 of Bahr *et al.* 2005) is given a calendar age of 8200 to 7500 calBP and follows directly after the interval T with the highest detrital input as measured by the Ti/Ca proxy signal.

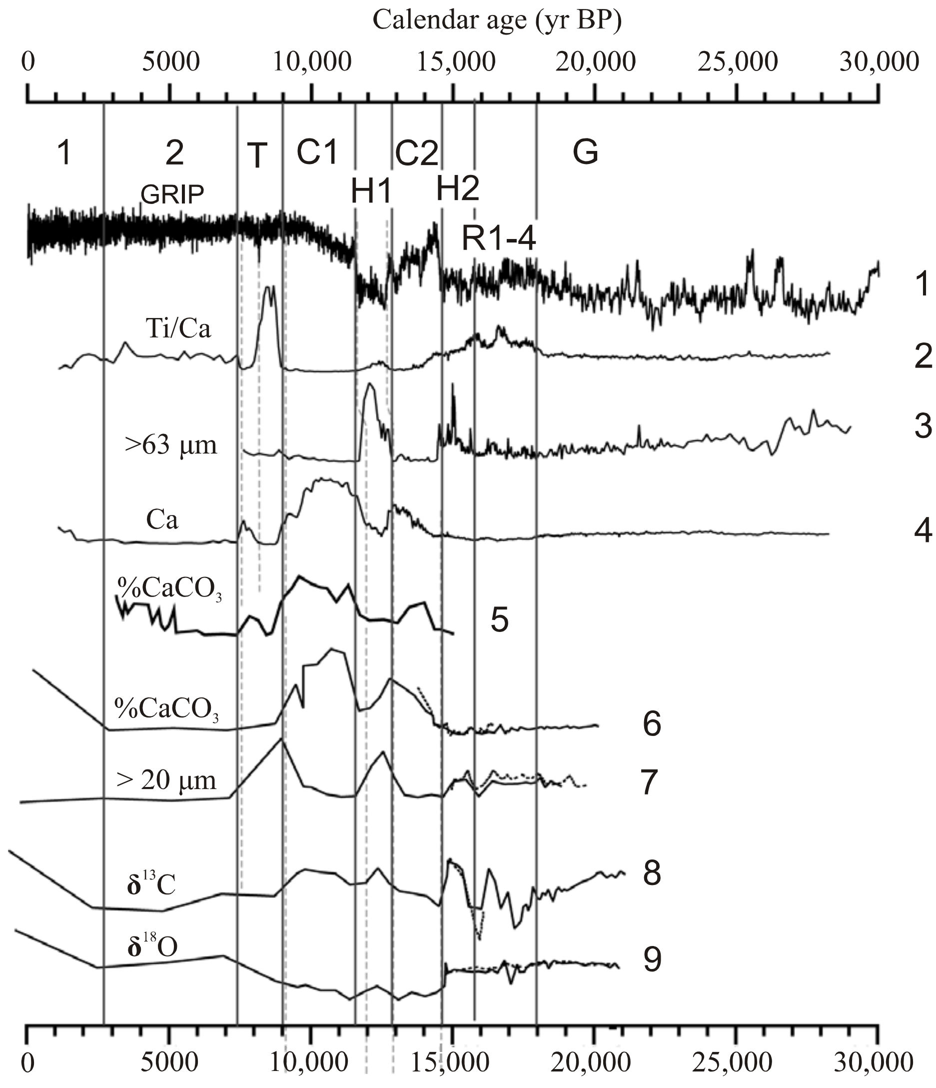

Figure 3. Correlation of Black Sea lithostratigraphic intervals with the GRIP chronology using the $\delta^{18}O$ signal in Greenland ice cores (Curve 1). Curves 2–4 are from core GeoB 7608-1 of Bahr *et al.* (2005). Curve 5 is from core A96 of Khrischev and Georgiev (1991). Curves 6–9 are from cores BLKS9809 and BLKS9810 of Major *et al.* (2002). The connection of the Black Sea with the Mediterranean occurs at the T/C1 boundary. H1 corresponds to the Younger Dryas cold stage.

If one correlates in detail the Ti/Ca and > 63 μm detrital signals in core GeoB 7608-1 (Bahr *et al.* 2005) and the > 20 μm detrital signals in cores BLKS9809 and BLKS9810 (Major *et al.* 2002) as well as the details of the Ca and $CaCO_3$ signals in these same cores to the GRIP $\delta^{18}O$ curve (a widely-accepted standard for calendar year chronology), it becomes possible to evaluate the differences between raw ^{14}C dates on the Black Sea materials and the matching calendar dates from the GRIP ice cores at many of the boundaries between the Black Sea lithologic units. Assuming that the age differences are the result of time-varying reservoir ages and dendrochronological calibration, one can then back-calculate the reservoir ages. The results are given in Table 1.

Table 1. Calculation of approximate reservoir age from correlation of Black Sea lithologic unit boundaries to the GRIP ice core record from Greenland. Note: * C1 and C2 are the peaks from Major *et al.* (2002) and correspond to peaks C2 and C3 of Bahr *et al.* (2005).

Unit Boundary	Raw ^{14}C Age	Calendar Age	Reservoir Age
1/2	3,090	2,700	460
2/T	7,160	7,540	400
T/C1	8,400	8,900	350
C1*/H1	10,100	11,700	0
H1/C2*	11,000	12,800	200
C2*/H2	12,700	14,700	200
H2/R1	13,400	15,500	300
R4/G	16,400	17,950	1600

The calculated reservoir ages show a decline from the onset of the brown-red muds that are deposited broadly across the continental slope; they are found even beneath the basin floor at DSDP Site 379. The origin of these muds has been attributed to overflow from the Caspian Sea through the Manych Depression to the Sea of Azov (Ryan *et al.* 2003; Bahr *et al.* 2005) and from megafloods caused by sudden thermofrost melting (Chepalyga, this volume). The Caspian overflow has been attributed to its Early Khvalynian transgression caused by pulses of meltwater from the Fennoscandia and/or Barents-Kara ice sheets (Grosswald 1980; Kroonenberg *et al.* 1997; Grosswald and Hughes 2002). During this transgression, the Caspian Sea accumulated its characteristic “chocolate clays” that resemble the color of the brown-red mud in the Black Sea cores. Core GeoB 7608-1 indicates four major flooding intervals, each lasting hundreds of years. As described by Bahr *et al.* (2005), each layer has mm-scale laminations that might correspond to annual pulses of meltwater discharge. The brown-red mud is rich in the clay minerals illite and kaolinite released from soils with the melting of permafrost. A high flux of these soils to streams ceased several thousand years later with the return of vegetation accompanying Bølling warming as shown in calculated sedimentation rates (Major *et al.* 2002). Thus, freshwater delivered in torrential floods from melting ice and permafrost would equilibrate effectively with the atmosphere and carry a negligible reservoir age. The lack of vegetation would also lead to minor inputs of organic carbon.

A dramatic dilution of the inherited 1600 year glacial reservoir age is observed in the calculated reservoir ages. Starting after 18,000 calendar years BP, the Black Sea's reservoir age diminishes to an eventual minimum in lithologic unit H1 that corresponds to the Younger Dryas, when climate cooled and the sediment mineralogy and the strontium isotopic composition indicate a trend back towards glacial conditions. Once the marine connection was established at 8900 calendar years BP, the Black Sea reservoir age increased toward the global ocean value of 400 to 500 years.

4. RESPONSE TO SPECIFIC CRITICISMS OF THE FLOOD HYPOTHESIS

In a series of publications, Aksu *et al.* (2002a, b) and Hiscott *et al.* (2002) offer data that they argue, "do not support the catastrophic refilling of the Black Sea by waters from the Mediterranean." Their proposition is two-fold. They use seismic reflection profiling and coring data from the southwestern shelf of the Black Sea and from the region immediately south of the Bosporus exit to the Sea of Marmara. From the Black Sea shelf they present a roughly orthogonal network of profiles with a 5-km line spacing that are strike- and dip-parallel. They recognized a lowstand system tract consisting of seaward-prograding shelf-edge wedges similar to those surveyed by Ryan *et al.* (1997a, b) on the outer Ukrainian margins and those mapped on the Romanian margin during the BLASON expedition. Their regional shelf-crossing erosion unconformity, α, truncates the top of these prograding clinoforms and extends seaward to depths beyond –150 m (Aksu *et al.* 2002b:Figure 11C) just like the erosion surface elsewhere in the Black Sea. In fact, they describe the unconformity as "a lowstand erosional surface."

On the middle and outer shelf north of the Bosporus outlet, at depths of –80 to –95 m, they profiled a series of asymmetrical bedforms atop this unconformity (Aksu *et al.* 2002b:Figures 13A, B, 20A, B, C, and 22). Although they interpret an orientation of these features parallel to the isobaths, an earlier report (Aksu *et al.* 1999) states that these features are normal to the isobaths. Figure 22 in Aksu *et al.* (2002b) is particularly informative because it displays a side-scan sonar image showing the crest of the bedform oriented north-northwest-south-southwest (shelf edge-oblique). Features that are oriented parallel to the isobaths should correlate from dip line to dip line, and such correlation is not documented.

Based primarily on geometry, Aksu *et al.* (2002b) interpret these asymmetrical bodies as consisting of "barrier islands with back-barrier washover fans." The deposits "landward of the barrier islands are believed to be lagoonal deposits blanketed by transgressive marine muds." They also write, "this interpretation is yet to be confirmed by coring." What they have shown by coring, however, is that the so-called washover fans and lagoonal sediments that supposedly make up these bedforms (their Unit 1C) were deposited in the marine stage of the Black Sea during the time since 7770 BP, as documented by several of their ^{14}C-dated gravity cores. The Black Sea had to be attached to the external Mediterranean at this time to support marine fauna with *Mytilus*. Global sea level at 7770 BP was above –20 m. There is no possible way that a sea surface at –20 m could be compatible with lagoons and washover fans at –80 to –95 m. To get around this conflict, they propose that the bedforms are composed of older deposits lying above the unconformity and reaching back to 11,320 BP, a dating which they obtain by extrapolation, not direct dating. However, one of their own cores, MAR95-04, has sediment directly above the unconformity dated at 5780 BP and resting on pre-erosion sediments dated at 33,550 BP. Algan *et al.* (this volume) present core 1 at a depth of –96 m with an 8-cm-thick sandy, shelly layer resting on the unconformity with a sharp contact. It contains *Mytilus* shells with a ^{14}C age of 7415±115 BP, which is close to the age of 7770 BP directly measured in MAR00-06 directly above the unconformity.

The argument for a gradual transgression of the southwestern Black Sea shelf is based on the extrapolated ages for the bedforms that cover the shelf-wide unconformity, α. Ryan *et al.* (1997b, 2003) do, in fact, document by numerous ^{14}C measurements a major transgression between 11,000 and 10,000 BP that reaches to depths as shallow as –30 m. However, these transgressive deposits with *Dreissena*-rich Neoeuxinian fauna were exposed by a subsequent regression that extended seaward to the –100 m shoreline of Ryan *et al.* (2003) and Lericolais *et al.* (this volume). It is upon this younger unconformity (α_1 of Aksu *et al.* 2002b) that the linear sand ridges were shaped on the Ukrainian and Romanian margins and that the asymmetrical bedforms on the southwestern shelf were subsequently deposited.

The first challenge to the abrupt flooding hypothesis is therefore based on the interpretation by Aksu *et al.* (2002b) of unsampled and thus undated "transgressive systems tract deposits" whose assumed isobath-parallel bedform orientation is unsubstantiated and may even be incorrect. An alternate interpretation that is consistent with both their reflection profiles and ^{14}C-dated cores as well as those of Algan *et al.* (this volume) is that the asymmetrical bedforms on the middle and outer southwestern Black Sea shelf in the vicinity of the Bosporus exit are post-flooding deposits that accumulated beneath the inflowing Mediterranean waters since the opening of the strait around 8400 BP.

The second challenge to the abrupt flooding hypothesis is based on the mapping and presumed dating of a delta deposit south of the Bosporus exit to the

Sea of Marmara. A reflection profile across this delta was first published in Ryan *et al.* (1997b:Figure 8, bottom panel, right side). Illustrated here in Figure 4, this profile shows two important features. One is a major erosion surface (unconformity) caused by the late Pleistocene lowstand of the Sea of Marmara, when global sea level fell below the –85 m Dardanelles outlet. The erosion cut into Paleozoic bedrock and formed a valley leading southward from the strait to the shelf edge. The valley was presumably shaped by Black Sea overflow that coursed though a river to the shore of the Sea of Marmara lake. The sediments above the unconformity accumulated after the Mediterranean re-connected with the Marmara Sea around 12,000 BP (Çağatay *et al.* 2000; Sperling *et al.* 2003). The second important feature is that the delta topset-foreset breaks lie at depths of –35 to –40 m. This level requires that sea level during the time of delta formation be even shallower (< –30 m) to allow for minimal wavebase. Global sea level reached this height only since 8000 BP, after the Mediterranean had already connected to the Black Sea.

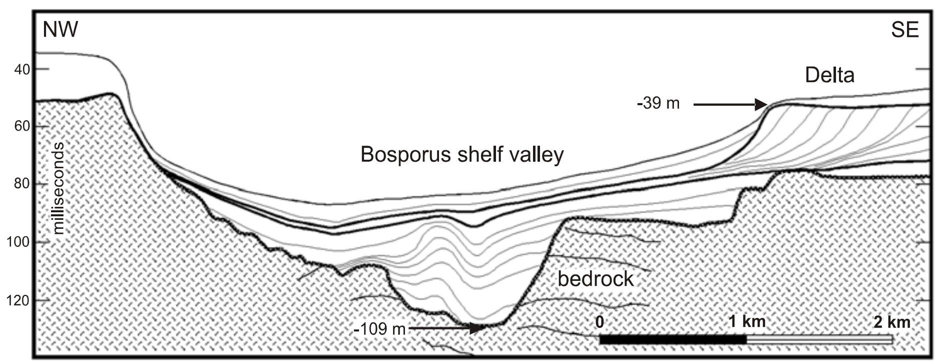

Figure 4. Reflection profile across the Bosporus shelf valley in the Sea of Marmara, south of the Bosporus Strait (from Ryan *et al.* 1997b). The prograding clinoforms of this subaqueous delta of the Kurbağalıdere River are shown on the southeastern side of the profile. The placement of the delta (outlined in bold) is in the upper half of the Holocene sediment succession that has accumulated above the late Pleistocene lowstand erosion surface.

However, Aksu *et al.* (2002a) and Hiscott *et al.* (2002) assert that the delta formed earlier, prior to the connection, and its sediments were sourced from the Bosporus Strait by a persistent Black Sea outflow. The crux of their argument refuting an abrupt Black Sea flooding is that during outflow, it is obvious that the Black Sea could not have been an enclosed lake susceptible to flooding from Mediterranean inflow. The argument for persistent outflow is constructed from two fronts. One is that the planview shape of the delta indicates a northern source. The second is that their Core 9 at a depth of –64 m sampled delta deposits at its base dated by ^{14}C between 10,000 and 9000 BP (Aksu *et al.* 2002a). If one looks critically at these propositions (shape and age), there are two flaws. First, the shape of the delta drawn in their illustration with a presumed northern source is unconstrained by the actual survey lines. The delta isopachs in Aksu *et al.* (2002a, b) are drawn well outside of the survey edges to imply a Bosporus source. Using a network of reflection profiles of the same vintage as in Figure 4, in combination with the published profiles of Aksu *et al.* (2002a, b), the delta is instead a lobate-shaped feature sourced from the Kurbağalıdere River to the east. Directions of foreset progradation can be determined from intersecting tracklines, and they clearly show a radial pattern with sediment delivered from the river mouth to the northwest, west, southwest, south, and southeast. A source of the delta from the Kurbağalıdere River was previously recognized by Oktay *et al.* (1992) and has been confirmed by similar progradation-direction measurements of Gökaşan *et al.* (2004).

The second flaw is the assertion that the basal sediments of core 9 belong to the delta. The core is located beyond the limits of the delta as shown even in Aksu *et al.* (2002a:Figure 4). Its basal layer, dated 10,000 to 9000 BP, is a fining-upward sequence rather than the coarsening upward sequence definitive for a prograding delta. Furthermore, if one looks at the stratigraphic position of the delta in Figure 4, the reflector that marks its base lies in the upper third of the post-12,000 BP sediment deposits. Such an elevation is more indicative of a middle Holocene origin rather than one in the earliest Holocene.

The criticism of Kaplin and Selivanov (2004) of a "fast but not catastrophic" water intrusion from the Mediterranean is based on evidence from the Kerch area that sea level can be reasonably estimated at –35 to –45 m for the period between 11,500 and 10,500 BP and was as shallow as –20 m at 9000 BP. Indeed, Neoeuxinian sediments of this time interval are preserved in depressions on the middle and inner shelf off Bulgaria and Ukraine (Dimitrov 1982; Ryan *et al*. 2003), in the limans of Ukraine (Shvebs 1988), and in the channels that cut across the Sea of Azov (Semenenko and Sidenko 1979). These sediments were deposited at relatively shallow levels when the Black Sea's lake was at its Younger Dryas highstand and, according to Major *et al*. (2002), spilled into the Sea of Marmara through a shallow outlet. During the Younger Dryas, the surface of the lake was well above the contemporaneous global ocean (Siddall *et al.* 2003). Kaplin and Selivanov (2004) consider a sea-level fall after 10,000 BP unlikely. According to them, the post-Neoeuxinian hiatus and its associated erosion surface is "problematic" despite the well-documented "peririf" ('break') above the Neoeuxinian in the coastal limans (Shvebs 1988), the hiatus observed in the core transect on the Ukrainian shelf (Ryan *et al.* 1997b, 1983), and the shelf-wide unconformity, α_1, of Hiscott *et al.* (2002). If this widespread break can be shown to have formed not in response to a regression, but instead to strong subaqueous currents all across an already-submerged shelf in association with a non-catastrophic intrusion of Mediterranean waters, then it may not be necessary to call upon a catastrophic flood phenomenon. As previously discussed, however, there is evidence that the dunes and pans at –65 to –80 m were formed by subaerial processes as recent as 8600 BP.

5. SUMMARY

Therefore, the criticisms of the abrupt flooding hypothesis by Aksu *et al.* (2002a, b), Hiscott *et al.* (2002), and Kaplin and Selivanov (2004) turn out to be differences in interpretation of facts and not the facts themselves. Does this mean that the hypothesis is not susceptible to nullification? No, the flood hypothesis is just as vulnerable as when first formulated. In the opinion of the author, this hypothesis serves only to explain the current observations. But there are still many potential obstacles in its path. Let me conclude by listing a few that concern me:

(1) Why, in an enclosed lake with its surface level at the whim of rapid climate change, did the lake level persist at –100 m sometime between 10,000 and 8400 BP to create such a pronounced shoreline with a characteristic beach profile and a belt of coastal dunes?

(2) Might climate have changed to wetter conditions just prior to the connection with the Mediterranean so that the terminal transgression was rapid, but entailed a freshwater flooding phenomenon instead of a saltwater event?

(3) Could phenomena associated with the rising interface between underlying saltwater and overlying freshwater following the connection with the Mediterranean have produced the erosional unconformity that separates the Neoeuxinian deposits from the overlying marine deposits? This horizon is the "washout" of Khrischev and Georgiev (1991). Though today, this watermass interface lies below the shelf edge, might it, right after the connection, have risen across the shelf all the way to the coast such that internal waves produced the entire unconformity? If the latter can be substantiated, a process other than wind must have shaped the belt of dunes and pans on the outer shelf, or these features are much older and date back to the pre-Neoeuxinian lowstand.

(4) Why during either the time of rapid ice sheet melting (R1-R4) or the Younger Dryas (H1) did vigorous Black Sea outflow not cut the Bosporus outlet so deep that passage was available for much earlier marine connection?

(5) Can we expect that during the beginning of other interglacial stages, climate in the Black Sea watershed went through a similar cycle, so that warming produced other evaporative drawdowns of past lakes, each terminated by an eventual marine flooding event?

7. ACKNOWLEDGMENTS

I thank Candace O. Major, Walter C. Pitman III, and Steven L. Goldstein for a close and fruitful collaboration at Lamont-Doherty over the years with data acquisition, analysis, discussions, and interpretations. Gilles Lericolais led the highly successful BLASON expedition on which I participated and provided core samples and bathymetry files. Namik Çağatay and Kadir Eriş at Istanbul Technical University have shared many ideas, offered warm hospitality, and provided much encouragement. Kazimeras Shimkus (deceased) was an inspiring leader of the initial expedition on *R/V Aquanaut* in 1993 and was generous with his rich knowledge of prior Soviet research and results. Nicolae Panin organized and hosted the International Workshop on Fluvial-Marine Interactions in Malnas, Romania in 1996 at which the flood hypothesis was first presented and debated. Irka Hajdas performed most of the AMS ^{14}C measurements at the Institut für Teilchenphysik in Zurich. The research was supported by the U.S. National Science Foundation (Grant OCE-97-11320).

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