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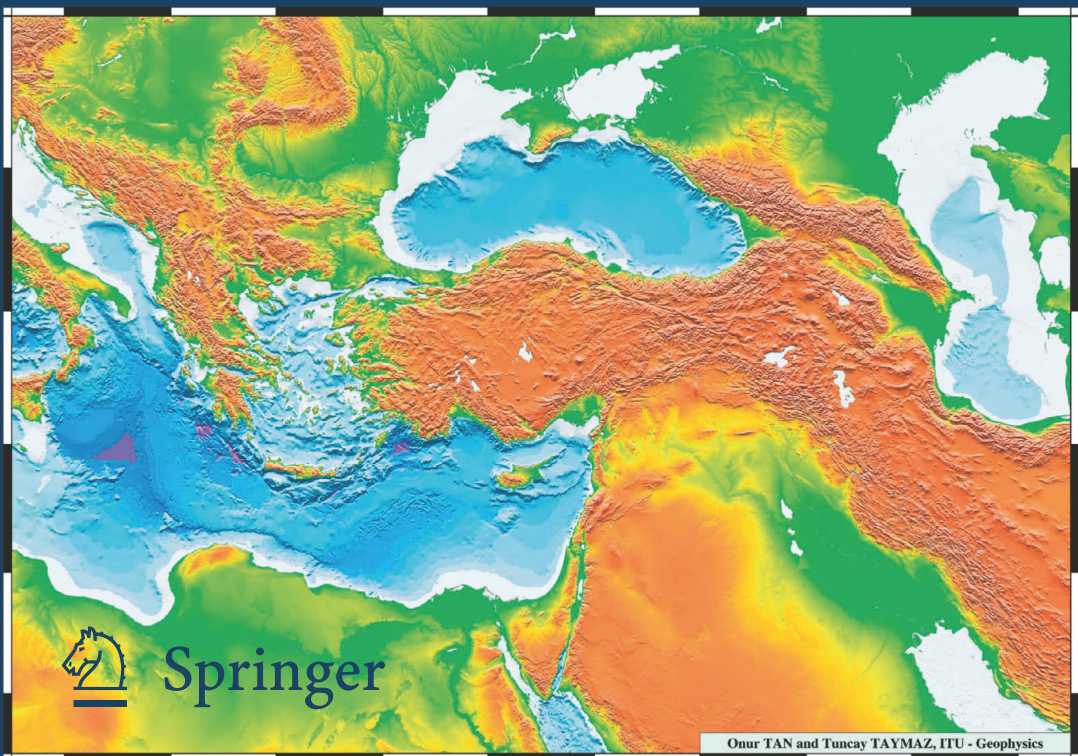
Editors

The Black Sea Flood Question

Changes in Coastline, Climate and Human Settlement



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The Black Sea Flood Question:
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The Black Sea Flood Question: Changes in Coastline, Climate, and Human Settlement

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Front cover: Physiographic map of the Caspian-Azov-Black-Marmara-Aegean Sea corridor produced using Generic Mapping Tools (GMT) Software based on topographic data from USGS–NOAA/GTOPO30 and NASA-SRTM (Shuttle Radar Topography Mission), and bathymetric data from GEBCO/97–BODC (General Bathymetric Chart of the Oceans). Map appears courtesy of O. Tan and T. Taymaz, Istanbul Technical University.

Back cover: A view northward along the central part of the Turkish Black Sea coast showing the discontinuous, zigzagging coastline formed by conjugated pairs of oblique faults. The village sits on an elevated ancient sea terrace. Photo appears courtesy of Y. Yılmaz, Kadir Has Üniversitesi, Istanbul.

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PREFACE

A lengthy book does not need a lengthy preface, so these opening words will convey only some essential matters, including the circumstances that led to the present publication, some of the background to the research it contains, and thanks to those who helped in the effort.

The Black Sea is one of the largest marginal seas: as deep as 2250 m and over 420,000 km^2 in area. Its coastline visits seven nations and links Europe with Southwestern Asia, while its water is the product of Eurasian rivers and rainfall mixing with immigrant Mediterranean saline flowing in through the Bosphorus. Due to its semi-isolation from the world ocean, the Black Sea tends to amplify environmental changes, and thus its detailed and sensitive paleoclimatic record has become a focus of oceanographic research. It is also the world's largest anoxic basin, enabling sophisticated studies of marine oxygen depletion and the exploration of ancient shipwrecks preserved in near pristine condition.

The earliest marine explorations in the Black Sea, dating to 1890–1891, were undertaken by the Black Sea Fleet's *R/V Chernomorets* to study the basin's hydrology and bottom sediments at water depths from 150 to 730 m (Andrusov 1890; Murray 1900). It quickly became clear that the bottom of the Black Sea is lifeless below the 150-m isobath, and that the sediments often contain shells of *Monodacna* and *Dreissena*, molluscan genera that no longer live in the Black Sea but are widely distributed in the slightly brackish limans, the local term for estuaries or submerged lower parts of river valleys.

Marine explorations continued under J.M. Shokalsky (hydrology) and A.D. Arkhangel'sky (sedimentology) on the Black Sea Fleet's *R/V "Pervoe Maya"* between 1925 and 1927. Using a one-meter corer, Arkhangel'sky discovered that the character of the bottom sediments had completely changed during the most recent millennia (Arkhangel'sky 1927). This transformation was explored in detail over the course of the next marine campaign, undertaken on *R/V "Pervoe Maya"* and *R/V "Hydrograph"* between 1928 and 1933. A new, improved corer of 6 m length was introduced, and based on the results of numerous expeditions, the geological structure and history of the Black Sea was described, and the first stratigraphic scheme for the Quaternary was developed (Arkhangel'sky and Strakhov 1938).

Previously designated Euxinian sediments (Andrusov 1918) bearing brackish, Caspian-type fauna were divided by Arkhangel'sky and Strakhov (1938) into Neoeuxinian (Novoevksinskie) and Old Euxinian (Drevneevksinskie) beds, the former distributed below sea level and containing the molluscs *Dreissena* and *Monodacna* and the latter presently lying above sea level on tectonically elevated terraces and containing *Didacna pontocaspia*. In today's terminology, Neoeuxinian deposits were laid down during the Middle to Upper

Würm, from Oxygen Isotopic Stage 2 to the beginning of Stage 1 (*ca.* 23 to 9.1 ky BP). The Old Euxinian is much older, Mindel and Mindel-Riss in date, *ca.* 400–260 ky BP (Yanko 1990). This initial stratigraphic framework was later improved by Nevesskaya (1965) based on molluscs and by Yanko (1990) based on foraminifera.

In 1970, a large-scale, systematic investigation of the floor of the Black Sea and Sea of Azov began under the authority of the USSR government to intensify the search for mineral, petroleum, and gas resources. Participating in the exploration were the USSR Academy of Sciences, the USSR Ministry of Education (Odessa and Moscow Universities), as well as various geological industries. At the about same time, western initiatives were undertaken, first in 1967 aboard the *R/V Pillsbury* of the Rosenstiel School of Marine and Atmospheric Science, Miami, and then in 1969 aboard the *R/V Atlantis II* of the Woods Hole Oceanographic Institution. The Atlantis II expedition was an extensive seven-week, two-leg cruise with international participation, results of which yielded additional information on the geological history of the basin that was widely distributed in the west (Degens and Ross 1974).

By 1997, a marine geological survey (1:500,000 to 1:10,000) of the Black Sea shelf had been largely completed by Eastern European scientists. Thousands of cores and tens of thousands of kilometers of high-resolution seismic profiles, taken across the shelf from the northern exit of the Bosphorus Strait in the west to the city of Batumi in the east, were studied as part of a multi-disciplinary effort. A methodology for Black Sea shelf investigation was developed, and using it, the paleoclimatic, tectonic, and sedimentary history of the basin was investigated. A high-resolution Quaternary biostratigraphy was established based on molluscs and foraminifera, supported by hundreds of radiocarbon assays (Appendices 1 and 2, this volume), and sea-level dynamics were reconstructed (for the references, see Balabanov, this volume; and Yanko-Hombach, this volume).

In all this work, no evidence was observed for a rapid sea-level rise in the Black Sea during the Holocene. Thus, when marine geologists William Ryan and Walter Pitman of the Lamont-Doherty Earth Observatory, Columbia University, announced their discovery of an abrupt flooding of the Black Sea in the early post-glacial and linked it to the biblical legend about Noah's Flood, the scientific community was surprised. It has now been a decade since they published their flood hypothesis. In it, they and their research collaborators proposed that, during the interval 14.7–10 ky BP (^{14}C uncorrected), the Black Sea was a freshwater Neoeuxinian lake with a level about 140 m below that of today. A rapid rise of the lake during the early Holocene transgression, which they initially dated to around 7.15 ky BP, submerged more than 100,000 km^2 of exposed continental shelf, rapidly and permanently flooding human settlements along the coast. In their view, this catastrophe accelerated the dispersion of early Neolithic foragers and farmers into the interior of Europe and might have formed the historical basis for the biblical story of Noah's Flood (Ryan *et al.* 1997).

The "Noah's Flood Hypothesis" triggered tremendous interest by the public, the scientific community, and the media. Major newspapers carried stories, based mostly on Ryan and Pitman's popular book (1998) and conference presentations, and some religious people exulted in the prospect that scientific evidence for a major Biblical story had at last been found. Many archaeologists with knowledge of the Pontic region's prehistory expressed skepticism that enough was known to link such a flood with the expansion of agriculture and Indo-European language spread, and many Near Eastern historians wondered how it could ever be proven that a Neolithic event taking place so far to the north was related to the growth of religious myth at the dawn of written history in Mesopotamia over 3000 years later. Many marine geologists were skeptical that such a flood had ever occurred, and a vigorous discussion over the matter ensued. The overall effect was salubrious to Black Sea studies in that the flood question very quickly aroused new interest in the region and encouraged fresh research ventures.

Unfortunately, the abundance of Black Sea data obtained by ex-USSR and former Eastern Bloc scientists was largely ignored in the global discussion. Neither the language barrier posed by the literature nor the general lack of west-east scientific dialogue could be overcome. Most of the information available in the west about the Black Sea came from western research initiatives. In addition, the flood problem was multi-disciplinary and needed a coordinated examination of both the geological and archaeological records by earth scientists and anthropologists. Such a strategy had not been implemented by the end of the 20th century, though some attempts had been made to do so on a small scale.

Ancient Mesopotamians would surely have agreed that astronomical alignments and favorable omens accounted for the coincidental planning of the conferences in the fall of 2003. V. Yanko-Hombach (with N. Panin) organized a meeting on late Pleistocene-Holocene climate and coastline migration of the Black Sea in Bucharest with NATO support in early October, and she assembled a topical session at the Geological Society of America's Annual Meeting in Seattle in early November. A. Gilbert independently organized a conference for mid-October at Columbia University under the auspices of the University Seminars to examine the archaeological and geological implications of the flood. Only with the increasing overlap in participants as planning progressed did the parallel efforts become mutually apparent. The three meetings were eventually coordinated and, with the wide geographic spacing of venues, they accommodated researchers from Eastern Europe as well as other western scientists:

(1) the NATO Advanced Research Workshop "Climate Change and Coastline Migration" (October 1-5, 2003, Bucharest, Romania; http://www.avalon-institute.org/NATO_ARW.html);

(2) the International Conference "The Black Sea Flood: Archaeological and Geological Evidence" (Columbia University Seminar on the Ancient Near

East, October 18-20, 2003, New York, USA; http://www.columbia.edu/cu/seminars/special-event/black-sea-conference); and

(3) the GSA Topical Session "'Noah's Flood' and the Late Quaternary Geological and Archaeological History of the Black Sea and Adjacent Basins," (Geological Society of America Annual Meeting, November, 4, 2003, Seattle, USA; http://gsa.confex.com/gsa/2003AM/finalprogram/session_9644.htm).

Over 50 papers were presented; the original programs are provided in Appendix 3.

The present volume was initially contracted to cover only the Bucharest meeting with NATO subvention. The advantages of including the diverse papers from the other meetings convinced the backers and the publisher Springer to allow the book to accept wider participation. Though the intention was a speedy report of conference proceedings to be submitted in camera-ready form, the substantial issues at stake and the magnitude of the interpretive differences prompted participants to rewrite their papers, incorporating extensive data presentations and discussions. At the same time, other Eastern European scientists with significant research investment in Black Sea studies volunteered contributions though they had not been conferees at any of the meetings. Some of these papers had to be fully translated from the Russian. The inclusion of so much new information eventually outstripped the publication limits of the NATO Scientific Series, and so with NATO's approval, a new contract was signed with Springer's Geosciences division leading to the present expanded volume. In the end, the book was transformed into a much more extensive review of the problems, and it contained a long overdue introduction into the western literature of a substantial amount of data previously locked away behind the Cyrillic in which it had originally been published. The goal of bringing east and west together to share perspectives as well as findings succeeded beyond expectation.

This collaboration will continue to grow in the research programs of the individual scientists, but also in a new five-year IGCP Project 521 entitled "Black Sea-Mediterranean Corridor during the last 30 ky: Sea-level Change and Human Adaptation" (www.avalon-institute.org/IGCP), which will be funded by UNESCO and IUGS. Further testing of the catastrophic flood hypothesis is among the main tasks of the project, but it will be more far-reaching in examining issues of regional climate, dynamics of human settlement, economic resources, and future environmental stability.

This book brings together 35 papers on geological, hydrological, climatological, archaeological, and linguistic aspects of the Black Sea flood hypotheses. Data and discussions reflect efforts at discerning and understanding paleoenvironment, climate dynamics, sea-level changes and coastline migration, regional hydrological variations, active tectonics, and geomorphology as parameters influencing human adaptation to the Circum-Pontic Region since the Last Glacial Maximum. Only empirical evidence recovered through accepted scientific methods was considered, and speculative implications linking Black

Sea events to biblical narrative, as has increasingly happened in popular books and media reports, was avoided.

No final answer to the Black Sea flood question appears here. Each paper in this book marshals its own evidence and offers its own interpretations, and there is no summary at the end with an overall resolution. The goal has been to provide access to information on a broad scale that crosses previously impenetrable language barriers, so that new work in the region can proceed with the benefit of greater perspective. The three fundamental scenarios for late glacial to Holocene rise in the level of the Black Sea—catastrophic (Ryan), gradual (Hiscott *et al.*), and oscillating (Chepalyga and Yanko-Hombach)—are presented early in the book, with the succeeding papers organized by geographic sector: northern (Ukraine), western (Moldova, Romania, and Bulgaria), southern (Turkey), and eastern (Georgia and Russia), as well as three papers on the Mediterranean. We hope that the contributions of this volume will serve as a foundation for designing more inclusive collaborative investigations and building greater consensus about what the past century's discoveries in the Circum-Pontic Region mean.

Each paper in the book underwent a lengthy review process (three reviewers as a rule per paper) and both language and graphics editing. Only one paper (W. Ryan) did not go through the reviews because it was submitted late. The complex editorial work (done mostly by Gilbert) took longer than expected, which accounts for why the publication was delayed over two years.

Acknowledgment must first be given for the financial assistance that made the conferences and book possible. First, the Avalon Institute for Applied Science provided much encouragement as well as release time to V. Yanko-Hombach for her conference organization and book editing. We thank NATO for supporting the Advanced Research Workshop in Bucharest and for kindly permitting recontracting of the present volume with Springer Publishers. Much of the cost of the Columbia University conference and many of the editorial expenses for graphics and translation were generously underwritten by the Office of University Seminars with the enthusiastic encouragement of its director, Prof. Robert Belknap, and the invaluable administrative help of Amanda Roberts, Alison Garforth, and Meredith Davis. Additional funding needed to cover travel and logistical costs incurred by the New York participants was provided by The Institute for Aegean Prehistory, The Trust for Mutual Understanding, The Joukowsky Family Foundation, Turkish Airlines, and one anonymous donor. The Geological Society of America also provided grant sponsorship to support several foreign presenters at the Seattle meeting. For all this financial backing, we express our sincere appreciation.

At various stages in the book preparation, we called upon the help of others. Russian translations were provided by Marianna Taymanova, Valentina Yanko-Hombach, Irena Motnenko, and Pavel Dolukhanov. All transliterations of cited sources in Cyrillic followed Library of Congress style for both

consistency and compatibility with the Online Computer Library Center's World Catalogue, to maximize ease of location for the references in question. Most of this standardization was done by Kira Haimovsky of Walsh Library, Fordham University. The only exception has been with proper names, of people and some geographic places, where 'y' has been used instead of 'ii' or 'i' according to preference or a pre-existing Romanization in common use. Citations of already published sources containing a transliterated Russian name retain that transliteration style, whether or not it follows the conventions of this book. All article and book titles cited in eastern languages have been provided with English translations for the reader's convenience. The reader will also find among the papers two spellings of the Istanbul strait (Bosphorus and Bosporus), used according to author preferences. Radiocarbon dates appear often as ky BP, or thousand years before present, and are to be understood in this form as uncalibrated. When calibrated, they are rendered as ky calBP or ky calBC.

Grateful acknowledgment is offered for the thoughtful efforts of many external reviewers: Ali Aksu, Canada; Norman Alavi, Australia; David Anthony, U.S.A.; Igor Balabanov, Russia; Ofer Bar-Yosef, U.S.A.; Zvi Ben Avraham, Israel; John Bintliff, The Netherlands; Elizaveta Bozilova, Bulgaria; Rodolfo Coccioni, Italy; Carlos Cordova, U.S.A.; Italo DiGeronimo, Italy; Pavel Dolukhanov, U.K.; Walter Dragoni, Italy; Rhodes Fairbridge, U.S.A.; Kevin Fleming, Germany; Victor Friedman, U.S.A.; Liviu Giosan, U.S.A.; Alexander Glebov, Russia; Erkan Gökaşan, Turkey; Ze'ev Herzog, Israel; Michael Kaminski, U.K.; Akio Kitoh, Japan; Johannes Koder, Austria; Hayrettin Koral, Turkey; Michal Kowalewski, U.S.A.; Peter Kuniholm, U.S.A.; Frank Lamy, Germany; Suzanne Leroy, U.K.; Mike MacCracken, U.S.A.; Yossi Mart, Israel; Bryant McAvaney, Australia; Peta Mudie, Canada; James Murray, U.S.A.; Mehmet Özdoğan, Turkey; Nicolae Panin, Romania; Jürgen Richter, Germany; Karen Rubinson, U.S.A.; Curtis Runnels, U.S.A.; William Ryan, U.S.A.; Gavin Schmidt, U.S.A.; David Scott, Canada; Andrew Sherratt, U.K.; Jim Teller, Canada; Tjeerd van Andel, U.K.; Claudio VitaFinzi, U.K.; Alasdair Whittle, U.K.; How Kin Wong, Germany; Herbert Wright, Jr., U.S.A.; Aslıhan Yener, U.S.A.; and Erdinç Yiğitbaş, Turkey.

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INTRODUCTION

The Earth sciences and catastrophic flooding have been linked for a very long time. The *Oxford English Dictionary* reports the etymology of the word 'geology' as it serves to label a branch of science. The first use of the word was in the title of the 1690 book *Geologia: or, a Discourse Concerning the Earth Before the Deluge*, written by Erasmus Warren. Warren's *Geologia* concerned the literal truth of the Book of Genesis. However, like biblical literalists before and since, Warren had to resolve a paradox: the mixing in Genesis of two very different accounts of the Noachian Debacle. In one account, The Flood derives from "foundations of the great deep" (Genesis 7:11). In the other, The Flood derives from "the windows of heaven," such that it rained continuously for 40 days and 40 nights (Genesis 7:12). Warren privileged the first account, postulating that water burst from great caverns. It is interesting that modern biblical scholars (Cohn 1996) hold that the phrase "fountains of the great deep" can refer to underground waters, but it also can refer to water from the world ocean. The latter source would be relevant to the controversy that is the subject of the present book.

Although the first geology was arguably "flood geology," it soon became apparent that geologists would not be worthy of the appellation "scientists" (coined in 1840 by the Cambridge mineralogist, William Whewell) if their activity were to consist solely of bearing witness to authoritative pronouncements, including those presumed to come from a deity. It is unfortunate that today "flood geology" commonly refers to a branch of "creation science," sharing with that enterprise an erroneous use of the word "science." Science is no more and no less than an unrestricted inquiry into nature. To have its answers ordained in advance is a restriction on free inquiry, and the result is sham reasoning (Haack 1996), not science.

By the early nineteenth century, geology had evolved to a science concerned with observations of nature on a path to whatever could be discovered about causal patterns in regard to those observations. The inquiries of geologists had to be free to lead anywhere the observations and their implications required, unconstrained by prior notions of what was true, or even of what might be proper in the pursuit of that truth. Unfortunately, there also emerged confusion over the last point, and vestiges of that confusion linger even to the present day. This confusion involves the notion of "uniformitarianism" (another word coined by the famous polymath and logician, William Whewell). As a prohibition against the valid inference of cataclysmic processes, uniformitarianism is invalid as a concept in science, i.e., it blocks the path of inquiry (Baker 1998). Indeed, there is nothing wrong, scientifically speaking, with invoking cataclysmic flooding as a natural explanation, if the facts, rather than some preconceived belief, lead the inquiry in that direction.

This volume was stimulated by just such a series of facts and the inquiry that followed. The fascinating thing, however, was that the inquiry led back to a geological conclusion, with similarities to the one that had been argued erroneously 300 years earlier. Based on remarkable marine science data from the Black Sea, W.B.F. Ryan and colleagues (Ryan *et al.* 1997) proposed that the Black Sea basin had been catastrophically flooded during the early Holocene, now thought to have been about 8400 years ago (Ryan, this volume). Ryan and Pitman (1998) subsequently elaborated that, prior to this flood, the Black Sea basin held an isolated freshwater lake, which was separated from the world ocean (then at a much reduced sea level) by the mountains of Turkey. Moreover, a large population of people inhabited the shores of this lake.

Rising world sea level eventually resulted in the breaching of the mountain divides that separated the freshwater lake of the Black Sea from the world ocean. As the water burst through the modern Bosporus Strait, the water rose 15 cm per day in the Black Sea, filling its basin in about 2 years. The human population that experienced this cataclysm was forced to disperse, carrying with it a memory of the great flooding, and conveying that story to the many other cultures that were encountered. Given that one of those cultures provided the Mesopotamian influence on the author(s) of Genesis, it was appropriate to label the model for this event, the "Noah's Flood Hypothesis."

The papers in this volume are all concerned, at least peripherally, with the "Noah's Flood Hypothesis" of W.B.F. Ryan and colleagues. The current status of this hypothesis, modified from the original by Ryan *et al.* (2003), is defended by Ryan (this volume), who outlines seven observations that are key to his model of abrupt early Holocene saltwater flooding of the late ice-age lake that occupied the Black Sea basin. Hiscott *et al.* (this volume) present an alternative model, the "Outflow Hypothesis," involving a gradual transition in salinity of the late Quaternary Black Sea. These authors do not accept the early Holocene evaporative drawdown of the freshwater lake in the Black Sea basin that preceded the 8.4 ky BP cataclysmic saltwater inundation of the "Noah's Flood Hypothesis." However, they do accept one of the modifications made in the original Ryan *et al.* model, specifically that late-glacial, meltwater-induced inflow to the Black Sea basin induced it to spill freshwater through the Bosporus to the Sea of Marmara. This late Pleistocene freshwater flooding was on an immense scale, such that Chepalyga (this volume) claims that "The Flood" was not the 8.4 ky BP saltwater inundation of the Black Sea basin (derived from the world ocean via the Bosporus, Sea of Marmara, etc.). Instead, there was an earlier, much larger catastrophe in which a cascade of spillings occurred from the Aral to the Caspian basins, and ultimately to the Black Sea via the Manych Spillway. Additional water was supplied by "superfloods" in the river valleys of European Russia, the Don, Dnieper, and Volga.

In the 1970s Mikhail G. Grosswald recognized that the Late Quaternary ice-sheet margins of northern Eurasia, like those of northern North America, held huge proglacial lakes. Great spillways developed for the diversion of drainage. In North America immense flows were successively diverted into the Mississippi, Mackenzie, and St. Lawrence Rivers (Teller *et al.* 2002). A final outburst of the megalake Agassiz-Ojibway released about 160,000 km^3 of freshwater into the Labrador Sea via the Hudson Strait about 8400 years ago (Clarke *et al.* 2003). Grosswald (1980) envisioned Eurasian meltwater diverted to the south-flowing Dnieper and Volga Rivers, leading to the Caspian and Black Sea basins. However, he more controversially hypothesized impoundment of the great north-flowing Siberian rivers, the Irtysh, Ob, and Yenisei, by ice sheets that covered the modern Barents and Kara Seas. More recent work confirms these impoundments and ice sheets, though there remains considerable controversy over their extent, timing, and genesis (Mangerud *et al.* 2004). These flows would have contributed to the system described by Chepalyga (this volume), but the discharges would have been much larger than he proposes. Indeed, many of the late-glacial cataclysmic flood systems had flows immensely larger than those proposed by Ryan and Pitman in the "Noah's Flood Hypothesis" (Baker 2002).

With much of North America and Eurasia experiencing huge diversions of drainage by glacial meltwater flooding during the period of major ice-sheet decay, it is not surprising that many human cultures developed narrative traditions involving "world-wide flooding." Certainly, "the world" for a local human society of 12,000 years ago involved a much smaller geographical extent than that word would convey to the global human society of today. There is no mystery that the most impressive events in the lives of many late ice-age cultures would have been "world-wide flooding."

Was the Black Sea inundation the source of a flood myth, specifically one that inspired western Asiatic peoples to the beliefs that inspired the account of Noah in Genesis? The anthropological implications of the "Noah's Flood Hypothesis" were greeted with considerable skepticism by many archaeologists. If the papers in this volume can be considered a test of the model, their conclusions range from equivocal (Chabai, this volume; Filipova-Marinova, this volume) to negative (Dolukhanov and Shilik, this volume; Anthony, this volume; Dergachev and Dolukhanov, this volume; and Bailey, this volume). Moreover, evidence for human dispersion after "The Flood" cannot be gleaned from language patterns (Nichols, this volume).

The physical aspects of the Ryan *et al*. model of early Holocene saltwater flooding of the Black Sea basin receive some support from Coleman and Ballard (this volume), Algan *et al*. (this volume) and Lericolais *et al*. (this volume). These studies document spectacular evidence for submerged paleo-shorelines, including drowned beaches, sand dunes, and wave-cut terraces. Some radiocarbon dates (Ryan, this volume) support the proposed early Holocene age for these presumed shorelines of the freshwater lake that existed prior to the cataclysmic inflow of marine water. Other studies find no evidence in preserved fauna or sediments that there was a cataclysmic flood (Yanko-Hombach, this volume; Kuprin and Sorokin, this volume; Shuisky, this volume; Shmuratko, this volume; Panin and Popescu, this volume; Balabanov, this volume; Glebov and Shel'ting, this volume).

Too much can be made in science of the current philosophical fad of testing (falsifying) hypotheses. As long recognized in geological investigations, hypotheses about past phenomena cannot function as propositions to be experimentally manipulated in a controlled laboratory setting. Because geologists study a past that is inaccessible to experimentation, they follow "working hypotheses," testing for their consistency and coherence with the whole body of collected evidence. Applying methods described by T.C. Chamberlin, G.K. Gilbert, and W.M. Davis (see Baker 1996), geologists have long used their working hypotheses to advance a path of inquiry toward the truth of the past, while avoiding the blockage of that inquiry by privileging any particular take on that past. It is certainly within this tradition that the various studies in this volume have operated. For both its advocates and detractors, the "Noah's Flood Hypothesis" of Ryan *et al*. has been a stimulus to further inquiry, made more productive by having a target to consider for the investigation. That the target involved considerable inspiration to the popular imagination just made the inquiry more intense and compelling. For what more could one ask in a scientific controversy?

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1

OXIC, SUBOXIC, AND ANOXIC CONDITIONS IN THE BLACK SEA

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Abstract: The Black Sea is the classic marine anoxic basin. It possesses an oxygenated surface layer that overlies a sulfide containing (anoxic) deep layer. This condition has evolved because the water column displays a strong density stratification arising because water with high salinity enters from the Bosporus Strait and mixes with water from an overlying cold intermediate layer (CIL). The CIL forms in the winter on the northwestern shelf and in the western gyre, and its rate of formation varies in response to changing climate. This mixture of Bosporus outflow and CIL produces the Bosporus Plume, which ventilates the deep layers of the Black Sea. New data about biogeochemical distributions of oxygen, sulfide, nitrate, and ammonium were obtained during *R/V Knorr* research cruises in 2001 and 2003. Oxygen is consumed by respiration of sinking organic matter, and sulfate reduction in the deep water results in the accumulation of hydrogen sulfide. Distributions in the upper layers reflect a classic example of the connection between climate forcing, physical regime, chemical fluxes, and biological response.

Keywords: suboxic zone, ventilation, temperature, salinity, oxygen, sulfide

1. INTRODUCTION

The Black Sea is located between latitudes 40° 55' and 46° 32' N and longitudes 27° 27' to 41° 42' E in the east-west oriented depression between two alpine fold belts, the Pontic Mountains to the south and the Caucasus Mountains to the northeast. The topography of the northwestern coast (except for Crimea) is relatively low. The Black Sea is the world's largest semi-enclosed marginal sea, and its physical and chemical structure is determined by its hydrological balance (Neuman 1942; Caspers 1957; Sorokin 1983). Values for area, volume,

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and depth are summarized in Table 1. The continental shelf is widest in the northwest, but the rest of the Black Sea is surprisingly deep for a marginal sea.

Table 1. Physical characteristics of the Black Sea.

Total Area	423,000 km^2
Area of Northwestern Shelf	101,600 km^2
Total Volume	534,000 km^3
Deep Water Volume (>50 m)	520,000 km^3
Depth of Permanent Halocline	50 to 200 m
Maximum Depth	2243 m

The only pathway for water exchange between the Black Sea and the Mediterranean is the narrow (0.76–3.60 km) and shallow (<93 m) Bosporus Strait. The sill depths of the Bosporus are 32–34 m at the southern end and 60 m at the northern end (Gunnerson and Özturgut 1974; Latif *et al.* 1991). The seawater flowing northward out of the Bosporus Strait is the only source of salty water to the Pontic basin, and as a consequence, deep-water salinity increases to 22.33‰. Freshwater inflow from several European rivers (especially the Danube, Dniester, Dnieper, Don, and Kuban) keeps the salinity low in the surface layer (S = 18.0 to 18.5‰ in the central region), and for this reason, the water column is strongly stratified with respect to salinity, and thus density. The main water fluxes are summarized in Table 2. These values show that evaporation exceeds precipitation and that the surface outflow is about twice as large as the deep inflow through the Bosporus. The currents in both directions are very strong.

Table 2. Present-day water fluxes of the Black Sea.

River Input		+350 $km^3\ y^{-1}$
Danube	250 $km^3\ y^{-1}$	
Dniester	8	
Dnieper	51	
Don	28	
Kuban	12	
Precipitation		+300 $km^3\ y^{-1}$
Bosporus Inflow to Black Sea		+313 $km^3\ y^{-1}$
Average Salinity	34.9 ‰	
Temperature	14.5° C–15.0° C in summer 12.5° C–13.5° C in winter	
Evaporation		–353 $km^3\ y^{-1}$
Bosporus Outflow to Marmara Sea		–610 $km^3\ y^{-1}$
Slope of water surface along the Bosporus from north to south	35 cm	
Current	~2 $m\ s^{-1}$ (surface) ~0.5 $m\ s^{-1}$ (at depth, but reaching ~1.5 $m\ s^{-1}$ over the sills)	

A consequence of the vertical stratification is that the surface layer (about 0 to 50m) is well oxygenated, while the deep layer (100m to 2000m) is anoxic and contains high sulfide concentrations. At the boundary between the oxic surface and anoxic deep layers, there is a suboxic zone (at approximately 50 to 100 m in depth), where the concentrations of both O_2 and H_2S are extremely low and do not exhibit any perceptible vertical or horizontal gradients (Murray *et al.* 1989; Codispoti *et al.* 1991; Jørgensen *et al.* 1991).

The suboxic zone in the Black Sea (Murray *et al.* 1989, 1995) is an important biogeochemical transition zone between the oxic surface layer and sulfidic deep waters. This layer, where O_2 and H_2S do not overlap, was first observed during the 1988 *R/V Knorr* Black Sea Expedition (Murray and Izdar 1989; Murray 1991). Its boundaries were chosen from the vertical distribution of oxygen and sulfide observed in the central gyre. After its discovery, these distributions were confirmed by others, and the processes controlling its origin and variability have been extensively discussed. When the suboxic zone was first observed, Murray *et al.* (1989) suggested that it might be a new feature resulting from reduced fresh water input from rivers. Subsequent research has shown that it is most likely a permanent feature of the Black Sea, at least since the early 1960s (Buesseler *et al.* 1994; Murray *et al.* 1995). The average thickness of this zone varies several-fold on a time scale of decades (Konovalov and Murray 2001), and this variability appears to be driven by variability in climate (Oguz and Dippner nd). The balance between oxygen injected due to ventilation of the thermocline with surface water and oxygen consumed by oxidation of organic matter governs the depth of the upper boundary of the suboxic zone (Konovalov and Murray 2001). The injection of oxygen into the upper part of the sulfide zone by water from the Bosporus is also an important control for the depth of the lower boundary, which marks the first appearance of sulfide (Konovalov and Murray 2001). Redox processes involving nitrate-manganese-sulfur are important for cycling of those elements in the lower part of the suboxic zone (Oguz *et al.* 2001).

The Black Sea is important to geochemists for several reasons.

(1) It is the classic anoxic marine ocean basin and is considered a prototype for the earth's ancient ocean. The ocean was considered to be initially totally anoxic. As atmospheric oxygen increased, the ocean contained an oxic surface layer and anoxic deep water from about 2.5 to 0.7 bya (Holland 1984; Berner and Canfield 1989).

(2) It has a well developed suboxic zone at the interface between the oxic and sulfidic layers where many important redox reactions involving Fe, Mn, N, and other intermediate redox elements occur.

(3) Similar redox reactions take place in sediments throughout the world's oceans, but they are easier to study in the Black Sea because they are spread out over a depth scale of 10s of meters (rather than cm or mm as in sediments). The various reactions have been shown to occur on similar density

(or depth) horizons from year to year, making them easy to study on repeated cruises.

(4) The Black Sea is an ideal site to study the effect of climate on ocean distributions. It is small enough in scale that variability in climate can vary physical forcing, and thus chemical fluxes and biological processes.

2. NEW DATA SETS FOR THE BLACK SEA

New hydrographic (T, S, and density) and oxygen/sulfide data were collected during two *R/V Knorr* research cruises to the Black Sea in 2001 and 2003. The cruises were divided into multiple legs, which allowed participation by 48 scientists from the US, Turkey, Ukraine, Russia, and Romania. One goal of these cruises was to analyze spatial and temporal variability in the suboxic zone in the southwestern part of the Black Sea in order to determine the effect on biogeochemical properties of the Black Sea caused by the intrusion of high salinity waters from the Bosporus (Konovalov *et al.* 2003). At the same time, new data were also collected at the northeastern coast of the Black Sea, near Gelendzhik, Russia, by researchers from the Southern Branch of the P.P. Shirshov Institute of Oceanology (SBSIO) (Yakushev *et al.* nd). The station locations were well situated to study the continental margin areas in the southwestern, northwestern, and northeastern regions.

Hydrographic data were obtained by standard CTD (conductivity, temperature, depth) procedures using SeaBird sensors. Oxygen and sulfide were determined by both wet chemical (volumetric) and electrochemical (voltametric) techniques (Luther *et al.* 2002; Konovalov *et al.* 2003). Nutrients were analyzed using standard autoanalyzer techniques. The vertical distribution of properties was sampled with rosette-CTD and pump profiling techniques (Codispoti *et al.* 1991; Konovalov *et al.* 2003). Charts of station locations, tables of participants, the analyses, and all data are available on the Knorr2001 and Knorr2003 web sites.[1]

3. HOW DOES THE BLACK SEA WORK?

Like the open oceans, the Black Sea possesses wind driven circulation with gyres, eddies, deep water thermohaline circulation, and shallower ventilation into the thermocline. Neuman (1942) described the surface circulation of the Black Sea as consisting of two large cyclonic (counterclockwise) central gyres that define the eastern and western basins. These gyres are bounded by the wind-driven Rim Current (Oguz *et al.* 1998), which flows along the abruptly varying continental slope all the way around the basin. The Rim Current exhibits large meanders and filaments that protrude into the regions of the central gyres. The geostrophically calculated currents typically have speeds of 25 cm s^{-1} along the axis of the Rim Current. Inshore or coastal of the Rim Current, there are several anticyclonic (clockwise) eddies (Oguz 2002) (Figure 1). Some of these eddies are permanently controlled by topography (e.g., the Sakarya Eddy located over the Sakaraya submarine canyon), while others are more temporally and spatially variable (e.g., the Sevastopol Eddy).

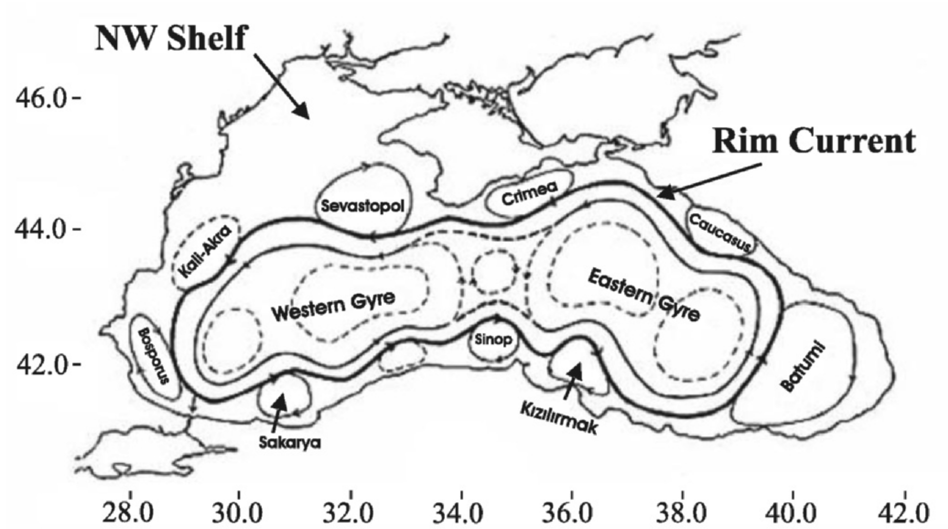

Figure 1. Chart of the Black Sea showing the wind-driven counterclockwise (cyclonic) Rim Current and several of the main anticyclonic gyres. The northwestern shelf and Rim Current are indicated.

The Bosporus Strait is the Black Sea's only connection with the Marmara and Mediterranean Seas, making it the only source of salt water to the Pontic basin. This salty water is also relatively warm (~15° C). The rivers are the main source of fresh water (300 km^3 y^{-1}) and mostly drain onto the northwestern shelf. The surface water can become relatively cold in winter, especially on this northwestern shelf. On average, the lower layer inflow from the Bosporus to the Black Sea is about 300 km^3 y^{-1} and the upper layer outflow is about 600 km^3 y^{-1}, which yields 300 km^3 y^{-1} for the vertically integrated transport driven by the Bosporus inflow (Özsoy *et al.* 1998) (Table 2).

These inputs result in strong vertical stratification with a fresh, lower density layer at the surface and a salty, higher density layer in the deep water. The keys for understanding the distributions are to remember that the only source of salt (and warm) water is through the Bosporus, and the only source of cold (and fresh) water is from the surface.

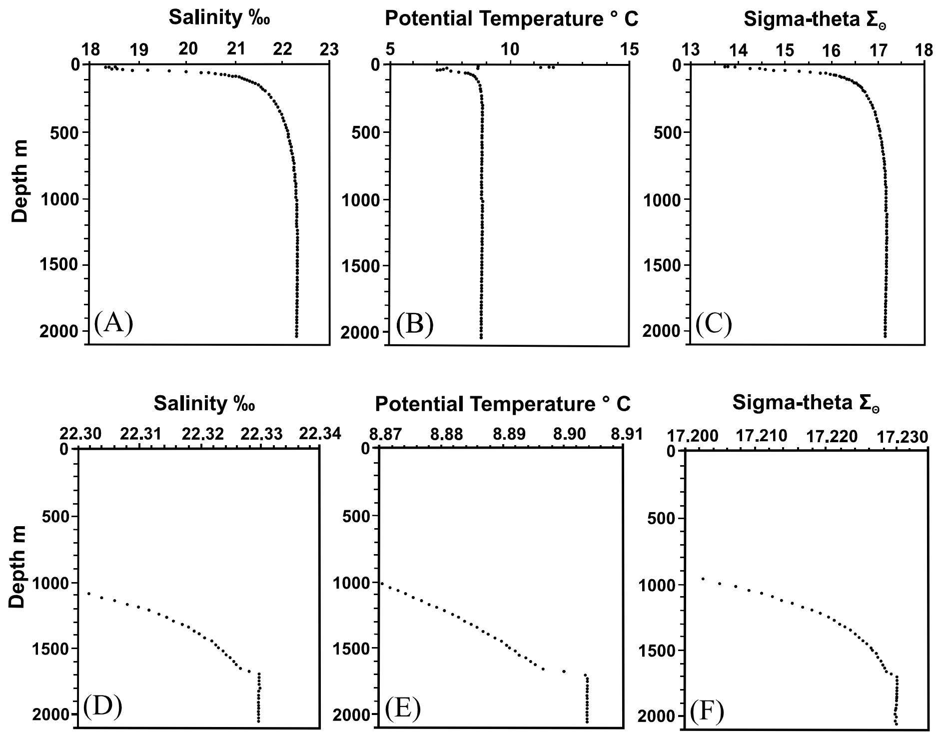

Figure 2. Salinity, potential temperature, and density (sigma-theta) from *R/V Knorr* 1988 Stn# BS3-2 HC-20 the Black Sea. A, B, C = Full scale water column 0 – 2200 m; D, E, F = Expanded scale to illustrate bottom boundary layer (from Murray *et al.* 1991).

The Black Sea has an estuarine-type circulation, which means the water flows in at depth and out at the surface. Full scale (0–2200 m) salinity, potential temperature, and density (sigma-theta) are shown in Figures 2A, B, and C. These CTD data were obtained during a research cruise on the *R/V Knorr* in the center of the western gyre in May of 1988 (Murray *et al.* 1991). Salinity increases continuously from low values of about S = 18‰ at the surface to deep water values of over S = 22.33‰. Density (σ_Θ) is controlled primarily by the salinity, and it increases similarly. Temperature is seasonally variable at the surface and decreases with depth to a feature called the cold intermediate layer (CIL) with a temperature minimum at about 50 m (Figure 2B). The water in this layer forms in the winter on the northwestern shelf and in the center of the eastern and western gyres. Its extent of replenishment varies from year to year depending on the intensity of the winter (Oguz and Dippner nd). Below the CIL, the temperature increases continuously all the way to the bottom. The properties of salinity, temperature, and density are extremely uniform in the deep water, from about 1700 m to the bottom, and form a homogeneous bottom boundary layer (Figure 2D, E, and F) (Murray *et al.* 1991). The top of the layer can be identified by a sharp density step. This layer appears to be formed due to bottom heating of the Black Sea by the upward flux of geothermal heat flow (which destabilizes density) superimposed on the downward increasing salinity (which stabilizes density). The situation when the gradients of temperature and salinity have the same sign, and thus opposite effects on density, results in a transport process called double diffusion (Imboden and Wuest 1995).

A temperature-salinity diagram can be used to illustrate the relationships between the distributions of temperature and salinity. The data from Figure 2A are shown as a T-S plot in Figure 3.

The high temperature and low salinity data on the left are from water near the surface. Temperature decreases to a minimum of about 7° C in the cold intermediate layer, and then both salinity and temperature increase continuously into the deep water.

When T-S data from the Black Sea are plotted with data from the Bosporus (which has maximum values of about T = 15° C and S = 36‰), it is apparent that, to a first approximation, the deep water of the Black Sea forms from linear, two-end member mixing of the Bosporus inflow with the cold intermediate layer (Figure 4). The magnitude of the Bosporus inflow averages 350 $km^3\ y^{-1}$, but current measurements suggest large variability in response to changing local winds. This implies that local synoptic meteorological conditions exert strong controls on the magnitude of transport. Numerical model results have estimated short-term (Oguz *et al.* 1990) and longer-term, decadal time-scale (Stanev and Peneva 2002) variability in transport through the Bosporus. The net transport varies from 200 to 350 $km^3\ y^{-1}$ over decadal time scales.

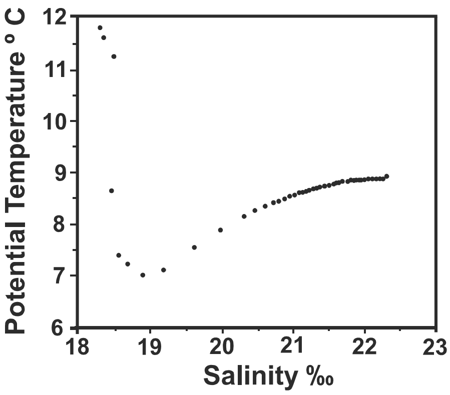

Figure 3. Potential temperature-salinity diagram for the center of the western basin from *R/V Knorr* 1988 Stn# BS3-2 HC-20 (from Murray *et al.* 1991).

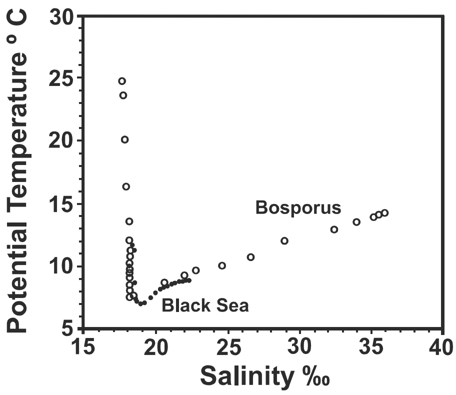

Figure 4. T-S data from the Black Sea and the Bosporus plotted together (from Murray *et al.* 1991).

The cold intermediate layer has two sources that are highly variable in intensity depending on climate. The first is the shallow northwestern shelf, where the water becomes very cold (<5.5° C) in winter (Figure 5) (Tolmazin 1985a). The second site is in the central gyre regions. An example of rejuvenation of the CIL in the western gyre was observed during a series of *R/V Knorr* cruises from March to May, 2003. The distribution of temperature versus depth for this period of time is shown in Figure 6. During the cruise in March, the water displayed a uniformly cold temperature (T = 6.1° C) from the surface to the depth (density σ_Θ = 14.5) of the CIL (Gregg and Yakushev 2005). The relative intensity of the northwestern shelf and central gyre sources is probably variable on a year-to-year basis depending on climatic conditions.

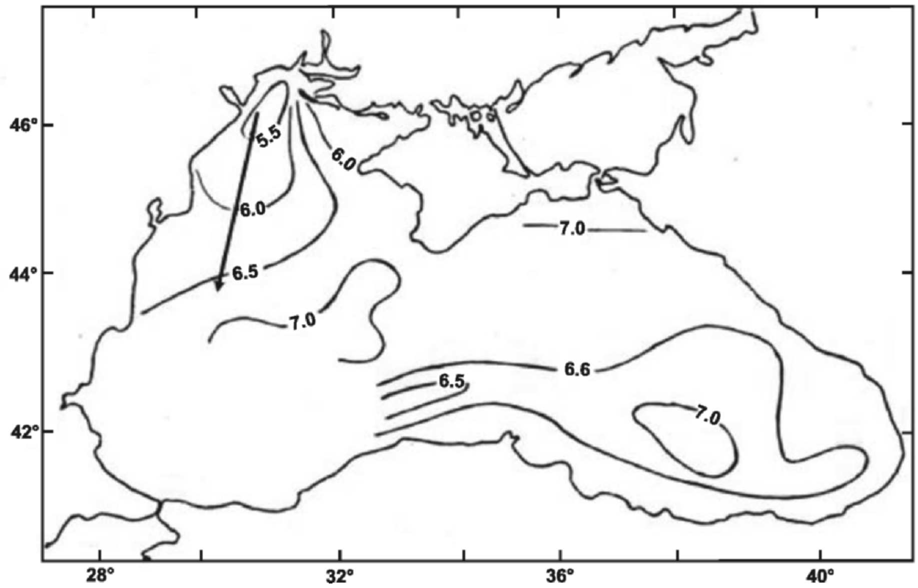

Figure 5. Isotherms in the cold intermediate layer for spring (from Tolmazin 1985a). The lowest temperatures are in the source area on the northwestern shelf.

Most of the mixing between the Bosporus outflow and the overlying cold intermediate layer occurs on the continental shelf just north of the Bosporus (Tolmazin 1985b). This can be seen in the salinity and temperature sections that extend along the axis of the Bosporus from the Marmara Sea to the continental shelf of the Black Sea (Figure 7) (Gregg and Özsoy 1999). Salinity and temperature are plotted versus thalweg distance, which is the distance along the axis of the Bosporus, measured relative to the southern entrance. The southern and northern sills are located at ~3 km and ~34 km, respectively. The bottom layer with high salinity water from the Marmara Sea comes in from the south and thins as it enters the Black Sea. Salinity gradients are sharp at its upper boundary

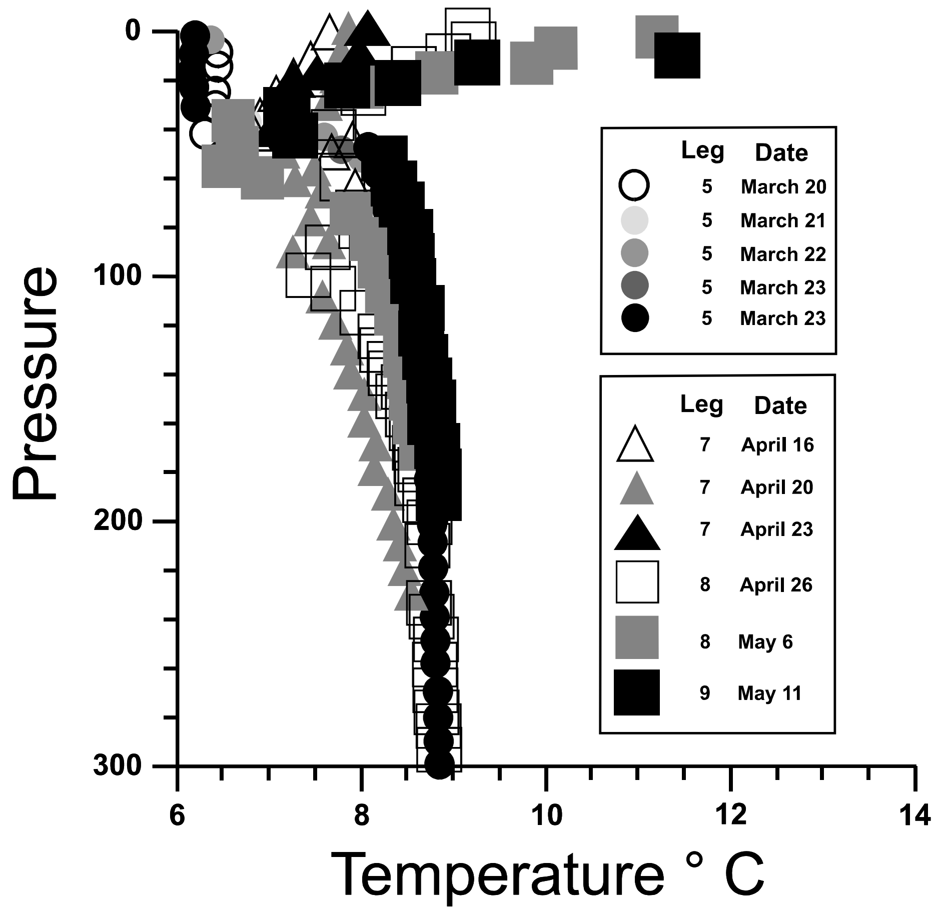

Figure 6. Temperature versus depth at the same location in the center of the western gyre of the Black Sea in March-May 2003. The legend gives the *R/V Knorr* Leg and date of the CTD casts. The chief scientists were Mike Gregg (UW) for Leg 5, James Murray for Leg 7, and George Luther for Legs 8 and 9.

indicating mixing with overlying water. The overlying water is characterized by the temperature minimum characteristic of the CIL. This mixing results in the linear, two end-member mixing characteristics for T and S discussed above (Figure 7). Most mixing occurs before the Bosporus outflow reaches the shelf break. The resulting Bosporus Plume ventilates the interior of the Black Sea at the depth represented by its density when it reaches the shelf break (Özsoy *et al.* 1993; Stanev *et al.* 2004). The most common mixing conditions result in ventilation of the upper 500 m but there must be occasional or rare ventilation events that reach the bottom. We know this because the only source for relatively warm and salty water is the Bosporus Plume, while S and T increase continuously all the way to the bottom. From the salinity balance of the deep Black Sea (50 to 2200 m), the ventilating water is composed of an average CIL to

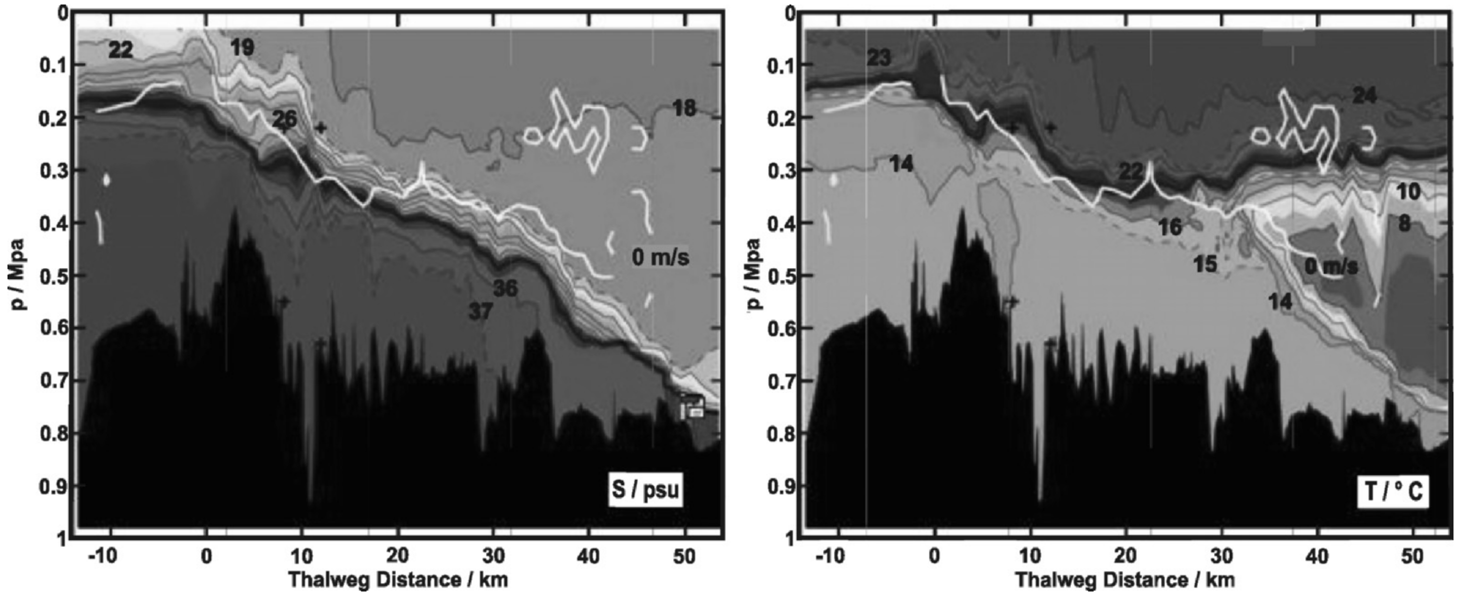

Figure 7. Section of salinity (left) and temperature (right) along the axis of the Bosporus, showing bottom topography (from Gregg and Özsoy 1999). The depth scale is from 0 to 100m. The southern entrance of the Bosporus is at a thalweg distance of 0, and outflow into the Black Sea occurs at 34 km. The southern and northern sills are seen at 4 km and 34 km, respectively. High salinity water from the Marmara Sea (S > 37‰) comes in from the south at the bottom. The temperature minimum of the CIL can be seen at about 40 m north of the Bosporus on the shelf of the Black Sea.

Bosporus entrainment ratio of ~4:1 (Murray *et al.* 1991). Thus, the average composition of the Bosporus Plume consists of a mixture of 4 parts CIL with 1 part high salinity Bosporus inflow from the Mediterranean.

In detail, this ratio must vary with depth; it is higher in the upper few 100s of meters and lower in the deeper water. Buesseler *et al.* (1991) used Cs isotope data to estimate an entrainment ratio of 10 for depths shallower than 200 m. Lee *et al.* (2002) used chlorofluorocarbon (CFC) data to model the decrease in ventilation and increase in residence time at depths shallower than 500 m. The residence time of water in different layers calculated from CFC data is shown in Table 3. Zone 1 includes the suboxic zone. The lower boundary of Zone 10 is at about 500 m. The entrainment ratio of CIL to Bosporus inflow decreases from 9.95 in Zone 1 to 3.78 in Zone 10. The CFC residence time increases from 4.8 years to 625 years over the same interval.

Table 3. Entrainment ratios (CIL/Bosporus outflow) and residence times of water in different density intervals determined through the use of CFC data (from Lee *et al.* 2002).

Zone	Density Interval	Entrainment Ratio	Residence Time (years)
1	15.450 – 16.178	9.95	4.8
2	16.178 – 16.451	5.73	12.5
3	16.451 – 16.610	4.93	21.1
4	16.610 – 16.717	4.55	30.8
5	16.717 – 16.799	4.32	44.4
6	16.799 – 16.858	4.1	59.7
7	16.858 – 16.910	4.03	85.1
8	16.910 – 16.950	3.91	129
9	16.950 – 16.986	3.84	235
10	16.986 – 17.016	3.78	625

Residence time for the deep water has been a subject of debate. If the only source of water to the deep Black Sea (volume >50 m = 5.20 x 10^5 km^3) was Bosporus inflow (313 km^3 y^{-1}), the residence time would be 1661 years. But, as the Bosporus inflow entrains cold intermediate water in a ratio of 4:1, the correct inflow to the deep water is 313 x 5 = 1565 km^3 y^{-1}, with a corresponding average residence time for water greater than 50 m depth of 332 years. A salinity balance gives the same answer (Murray *et al.* 1991). In detail, the residence time varies with depth, so this value is not incompatible with the values calculated from CFC given in Table 3.

^{14}C has been used to calculate an age of ~2000 years for the deep water of the Black Sea (Östlund and Dyrssen 1986). The difficulty with this approach is that the age was calculated relative to an input value that was assumed to be $\Delta^{14}C$ = –50‰. This value is expected for surface seawater in equilibrium with the atmosphere in pre-nuclear times (before ~ 1950). Atmospheric equilibration of new surface water with ^{14}C takes on the order of 10 years (Broecker and Peng 1974). If the residence time is really close to 330 years as calculated above, and the ventilating water is composed in part of Bosporus inflow with $\Delta^{14}C = -50‰$, the value of the ^{14}C for the CIL water entrained would have to be $\Delta^{14}C = -200‰$. In other words, the CIL would have had a $\Delta^{14}C$ value much older than expected for atmospheric equilibrium. This could occur if "old" deep water was upwelled, cooled at the surface during the winter to make CIL, then sent back down to the deep Black Sea without coming to equilibrium with the atmosphere.

4. BIOGEOCHEMICAL DISTRIBUTIONS

Because of the strong vertical stratification, the deep water is not replenished fast enough to replace the oxygen consumed by respiration of organic matter. Thus, the Black Sea has an oxygen-containing surface layer and a sulfide-containing deep layer. An example of the oxygen and sulfide distributions versus depth and density in the center of the western gyre is shown in Figure 8 (*R/V Knorr* 2003, Leg 7, Stn 12). Oxygen is at atmospheric saturation in the upper 40 m, then decreases sharply to near zero by 60 m. The same profiles are shown versus density in the figure on the right. Because of the sur-

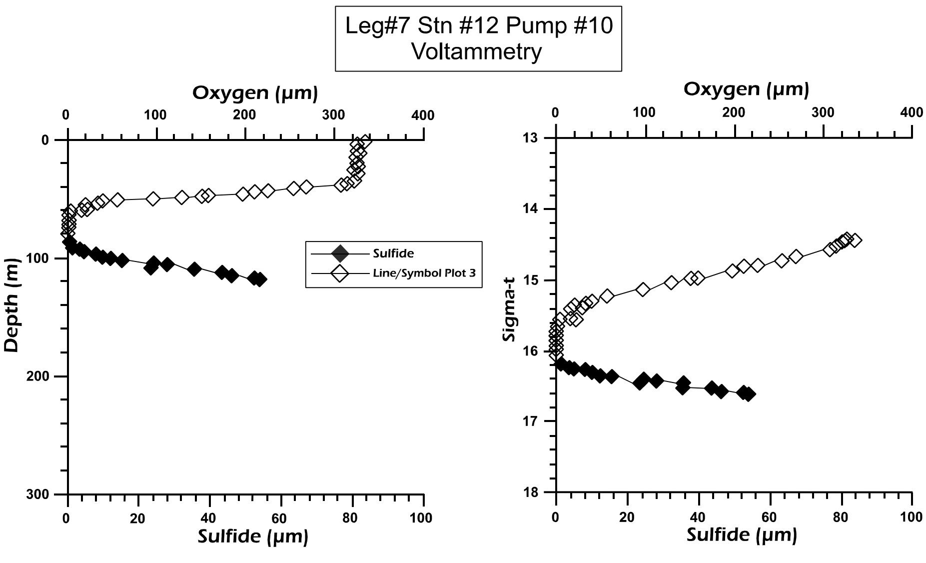

Figure 8. Oxygen and sulfide data from the center of the western gyre (Leg 7, Stn 12) during *R/V Knorr* 2003, courtesy of G. Luther (U. Delaware) and S. Konovalov (MHI, Sevastopol, Ukraine). Depth on the left and density on the right.

face circulation patterns discussed earlier, density surfaces are deeper around the margins, and they dome upward in the central regions. The shorthand notation is to express density as $\sigma = (\rho - 1) \times 1000$ where ρ is the density. For water with a density of $\rho = 1.016$ kg m^{-3}, the value of $\sigma = 16.0$. Thus, density is usually used as a vertical coordinate in the Black Sea rather than depth in m. All characteristic features tend to be deeper near the margins and shallower in the central gyres, but they almost always fall on the same density levels. For this reason, plotting against depth in the Black Sea produces a scatter of data, but when plotted against density, the same data show much less variability (e.g., Codispoti *et al.* 1991).

The first appearance of sulfide occurs at about 90 m or $\sigma = 16.15$ (Figure 8), and then sulfide increases continuously to maximum values of about 400 µM by 2200 m (Luther *et al.* 1991). One of the intriguing questions for the Black Sea is: what is the sink for the upward flux of sulfide? Sulfide decreases to zero before the first appearance of oxygen, thus sulfide is apparently not oxidized directly by the downward flux of oxygen. Several hypotheses have been proposed to explain the removal of sulfide. Millero (1991) and Luther *et al.* (1991) suggested that Mn cycling plays an important role and that the upward flux of sulfide is oxidized by a downward flux of oxidized Mn (III, IV). Konovalov and Murray (2001) estimated that significant upward flux of sulfide is oxidized by O_2 injected horizontally by the Bosporus Plume. Oxygen containing intrusions from the Bosporus Plume (deeper than $\sigma_t = 14.5$) are easily seen in profiles from the stations of the 2001 and 2003 KNORR cruises in the southwestern part of the sea, close to the Bosporus (Konovalov *et al.* 2003).

The suboxic zone is defined as the region between the depth at which oxygen decreases to near zero ($O_2 < 10$ µM) and that at which sulfide first appears ($H_2S > 10$ nM) (Murray *et al.* 1989, 1995). This layer varies in thickness from year to year (Figure 9) mostly due to climate driven variability in the depth of complete oxygen depletion (Konovalov and Murray 2001; Oguz and Dippner nd). The distribution of O_2 is determined by a balance between the oxygen consumed by respiration of sinking particulate organic matter (enhanced by increasing eutrophication during the 1970s and 1980s) and the input of O_2 by ventilation of the CIL and deeper layers (Konovalov and Murray 2001). Ventilation of the CIL sets the upper oxygen concentration and fundamentally determines the steepness of the vertical gradient and, thus, the downward flux of oxygen. The first appearance of sulfide has been much less variable with time. There was a period during the late 1980s and early 1990s, when the oxycline moved deeper (Figure 9) because a series of severe winters produced favorable climate conditions (cold) for ventilating the CIL. The thickness of the suboxic zone decreased as oxygen penetrated deeper. During this period, the temperature minimum (T_{min}) of the CIL was low, and it moved to deeper density layers. Ventilation of the CIL was enhanced, and oxygen concentrations on $\sigma_t = 15.4$ were higher. The periods of 1987 and 2001 had lower oxygen concentrations on the $\sigma_t = 15.4$ density surface and followed warm periods with less CIL formation, resulting in higher temperatures. As a consequence, the suboxic layer became thicker. Data from the 2001 KNORR cruise (Figure 9) demonstrate that the

Black Sea remains highly eutrophic and will undergo further perturbations during the anticipated future warming climate conditions.

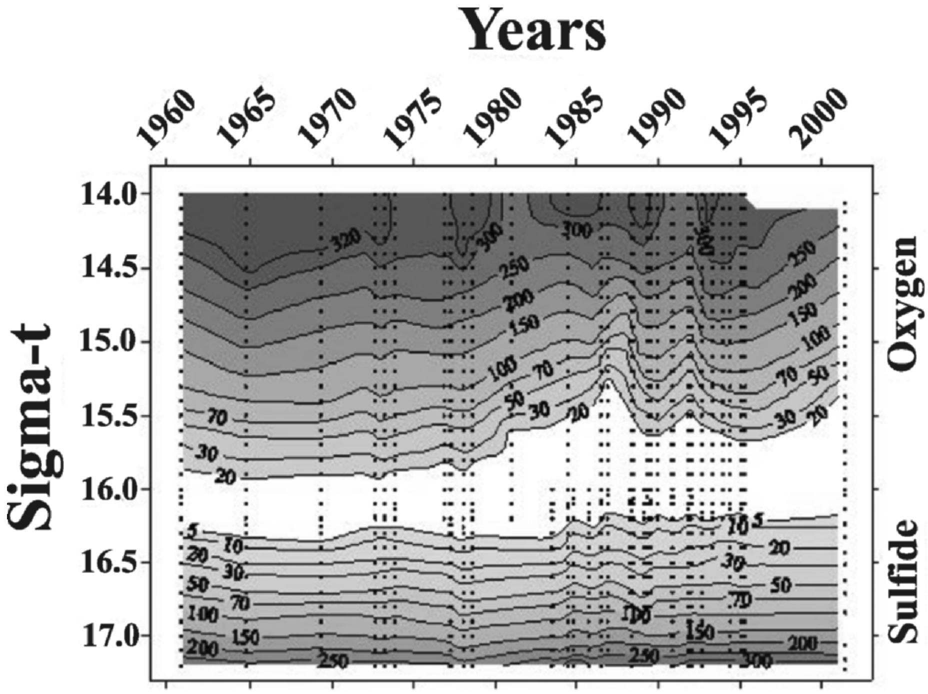

Figure 9. Temporal variability in the distribution of oxygen and sulfide versus sigma-t from 1961 to the present (from Konovalov and Murray 2001). In this example, the suboxic zone is the region between $O_2 < 10\mu M$ and $H_2S < 5$ nM.

The distributions of nitrate and ammonium in the center of the western gyre (*R/V Knorr* 2003, Leg 7, Stn 12) are shown versus depth and density in Figure 10. Nitrate is depleted at the surface due to biological uptake. It starts to increase at 40 m and reaches a maximum at 65 m ($\sigma_t = 15.5$) (approximately where O_2 decreases to 0). After oxygen has decreased to zero, nitrification of ammonium released from organic matter no longer occurs. Nitrate then decreases to zero at 75 m ($\sigma_t = 15.95$). Ammonium starts to increase at the same depth (density) and increases progressively into deep water. The disappearance of NO_3 and NH_4 at the same depth is consistent with a downward flux of NO_3 and an upward flux of NH_4 that are consumed over a narrow depth interval by the anammox reaction ($NO_2^- + NH_4^+ = N_2 + 2H_2O$). Note that the anammox reaction reduces NO_2^-, not NO_3^-, so there must be some denitrification also occurring that reduces NO_3^- to NO_2^- in order for anammox to occur. Kuipers *et al.* (2003) used 16S RNA gene sequences, RNA probes, ^{15}N label experiments and ladderane membrane lipids to show that anammox bacteria are indeed present at this level in the Black Sea. This reaction where ammonium is oxidized anaerobically to N_2 is important in the nitrogen cycle of the Black Sea (Fuchsman and Murray nd). Such a reaction has long been inferred from chemical distributions (Richards 1965; Brewer and Murray 1973) and is now confirmed.

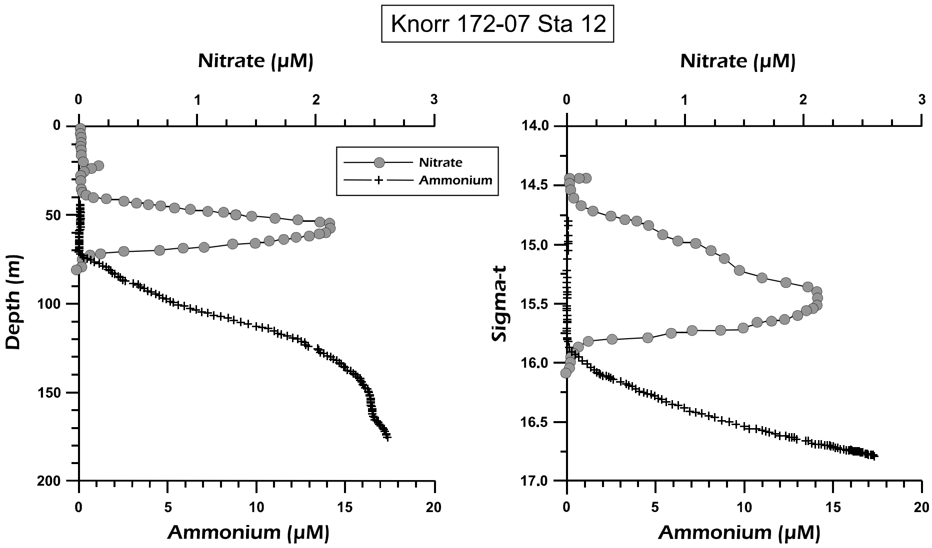

Figure 10. Nitrate and ammonium from the center of the western gyre (Leg 7, Stn 12) during *R/V Knorr* 2003. Depth on the left and density on the right.

The suboxic zone has interesting distributions for many other elements with redox chemistry intermediate between oxygen and sulfide. Examples include manganese and iron (Spencer and Brewer 1971; Spencer *et al.* 1972; Lewis and Landing 1991; Tebo 1991), arsenic and antimony (Cutter 1991), rare earth elements (Schijf *et al.* 1991), and iodine (Luther and Campbell 1991; Truesdale *et al.* 2001).

Methane (CH_4), which is produced after sulfate has been totally reduced in the classic redox sequence, is also high in the deep Black Sea. Initial studies of rates by Reeburgh *et al.* (1991) could detect only anaerobic methane consumption; the source of methane was uncertain. Recent studies have discovered numerous cold seeps of CH_4 on the northwestern shelf and slope that result in large plumes of CH_4 rich bubbles (C. Schubert, EAWAG, Switzerland, personal communication 2003). The source must be CH_4 production deep in the sediments of the northwestern shelf, slope and rise region, and especially the Danube Fan, after sulfate has been reduced to zero.

Bacteria mediate most of these redox reactions, yielding *in situ* consumption of CO_2 (Brewer and Murray 1973; Yılmaz *et al.* nd). Chemo-synthetic bacteria that grow on carbon dioxide and water obtain energy from reduced compounds like H_2S, NH_4, Mn(II), Fe(II), and CH_4. In addition, the discovery of high concentrations of bacteriochlorophyll (BChl) e in the suboxic zone by Repeta *et al.* (1989) suggests that anoxygenic photosynthesis occurs (Jørgensen *et al.* 1991). This is surprising because the light availability at these depths (<4 mEinst m^{-2} s^{-1}) is equal to 0.0005% of the surface irradiance (Overmann *et al.* 1992). The BChl e is associated with brown phototrophic *Chlorobium* bacteria, which probably compete with other bacteria under conditions of severe light limitation. The role these bacteria play in elemental cycling is still unclear.

The density values that were characteristic of many water column features during the 1988 Knorr Expedition are shown in Table 4. These values have served as a benchmark for subsequent cruises to evaluate the stability of the redox features. There has been some variability from place to place and time to time, but the general picture has remained the same.

Table 4. Characteristic density values of features associated with the biogeochemistry of the Black Sea. The uncertainty of each value is about 0.05 density units.

Feature	Density (σ_θ)
PO_4 shallow maximum	15.5
O_2 < 10 μM	15.7
NO_3 maximum	15.4
Mn_d < 200 nM	15.85
Mn_p maximum	15.85
PO_4 minimum	15.85
NO_2 maximum	15.85
NO_3 < 0.2 μM	15.95
NH_4 > 0.2 μM	15.95
Fe_d < 10 nM	16
H_2S > 1 μM	16.15
PO_4 deep maximum	16.20

5. ORGANIC MATTER PRESERVATION

Conditions that control the preservation, accumulation, and burial of organic matter have been the focus of considerable debate. Central to this controversy is whether there is enhanced preservation of organic matter under anoxic versus oxic conditions. One theory is that anoxic decomposition of organic matter is intrinsically slower than oxic decomposition. Lee (1992) conducted lab experiments that showed there was little difference in the rates of decomposition of specific types of organic compounds (e.g., amino acids, carbohydrates) under oxic versus anoxic conditions. Several geochemical studies (e.g., Pederson and Calvert 1990; Calvert *et al.* 1992; Ganeshram *et al.* 1999) have argued that organic matter preservation is enhanced not by anaerobic conditions but by the magnitude of the source function and grain size effects. Smaller grain size sediments contain more organic matter because they have a larger surface area to volume ratio, and organic matter coats the surfaces. Hartnett *et al.* (1998) suggested that the organic matter preservation efficiency is inversely correlated with the length of time organic matter is exposed to O_2 before burial under anoxic conditions. In the Black Sea, oxygen exposure time is minimal as sinking particles reach anoxic conditions just below the euphotic zone. Finally, a comparison of organic carbon preservation in sediments off Washington State and off Mexico led to the conclusion that oxygen rather than carbon input controls the extent of preservation (Hartnett and Devol 2003).

So this debate continues, but we do know from observations that wooden shipwrecks are better preserved in the anoxic layer than in the oxic layer of the Black Sea (D. Coleman and R. Ballard, personal communication 2003). The anoxic conditions in the Black Sea do enhance the preservation of archaeological remains.

ACKNOWLEDGMENTS

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ENDNOTES

1. Knorr2001 and Knorr2003 websites are at oceanweb.ocean.washington.edu/cruises/Knorr2001 and oceanweb.ocean.washington.edu/cruises/Knorr2003.

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2

MOLLUSCAN PALEOECOLOGY IN THE RECONSTRUCTION OF COASTAL CHANGES

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Abstract: Coastline migration significantly affects the evolution of marine and continental environments. Tracing sea-level changes that have occurred in the past can be achieved using evidence from marine benthic associations of fossil molluscs. This paper outlines a strategy for documenting recent molluscan benthic assemblages in order to detect abrupt or gradual variations in their depositional succession that could reflect changes in the land-sea relationship. Characteristics of molluscan fossil assemblages are reviewed and an example of such an analysis is offered from research conducted in Iskenderun Bay in the northeastern Mediterranean Sea.

Keywords: coastline migration, molluscan paleoecology, marine benthic associations, taphonomic signature, Iskenderun Bay

1. INTRODUCTION

Sea-level changes influence the equilibrium profile of rivers and their sediment load, and therefore they have the potential to affect the evolution of both marine and continental environments. Coastline migration can be traced by using evidence from historical heritage and archaeological remains as well as landscape geomorphology and marine geophysical survey, but the effects it imposes upon marine benthic associations can also be seen along the continental shelf and down into the basins.

Paleoecology integrates data from fossil biota and surrounding sediments, and, with the aid of distributional and ecological models derived from the study of modern environments, it identifies the most important ecological variables that previously characterized the paleoenvironment. Of the many potential tools for coastal and shelf studies, benthic molluscs are particularly

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useful for several reasons.

First, molluscs are long-lived. Longevity estimates for some common coastal bivalves have been extended to 50 years or more, while certain offshore species are known to live for over 100 years (Ropes *et al.* 1984; Allison *et al.* 1994; Faldborg *et al.* 1994; Schöne *et al.* 2001; Keller *et al.* 2002; Sejr *et al.* 2002). A well-structured, autochthonous molluscan shell assemblage (i.e., local populations composed of juveniles and adults) needs years or even decades to manifest itself, and thus, a more or less stable paleoenvironment, at least at the scale of the molluscan life span, must have existed (Staff and Powell 1988). Moreover, such molluscan assemblages acquire their character within biotopes that are in dynamic equilibrium, demonstrating both short- and medium-term variability. For this reason, the assemblages cannot reflect an ideal, static situation–unless an exceptionally high sedimentation rate occurs–but instead are the result of a longer time scale that filters out shorter-term variation and local oscillations.

Second, molluscan taxonomy has a comparatively sound basis. The ecological requirements of the most common species are well known, allowing relatively easy data exchange between workers from different regions.

Third, molluscs make up the greatest part of preserved shelf assemblages (Staff and Powell 1988; Basso *et al.* 1988; Flessa *et al.* 1993), and they are protagonists in several models of benthic zonation, whether based on communities (Petersen 1915) or on biocoenoses (Möbius 1877; for a discussion of the different approaches, see Basso and Corselli 2002). In particular, the bionomy of the modern Mediterranean Sea (Pérès and Picard 1964; Pérès 1982) has been successfully tested as a model and conceptual framework in dozens of papers devoted to paleoecological reconstructions of Quaternary marine benthic environments (e.g., DiGeronimo 1985).

Another advantage of molluscs is the extensive amount of information available about their taphonomic behavior. Taphonomy has been defined as a conceptual subsystem of paleontology (Fernández López 1988) that examines the "laws of burial" in order to understand how organic remains make the transition from the biosphere into the lithosphere (Efremov 1940). The taphonomic approach is fundamental in identifying the main post-mortem processes that modify the original source community and transform it into a fossil assemblage. Several papers on the theory and application of molluscan taphonomy have provided a solid framework for interpretation (e.g., Allison and Briggs 1991; Kidwell 2001; Powell *et al.* 2002; Kowalewski *et al.* 2003).

This paper describes a practical strategy for identifying successions of molluscan benthic assemblages within recent marine deposits in order to detect gradients or abrupt changes in their depositional environments that might reflect modifications in land–sea interaction within the coastal zone.

2. RATIONALE

Molluscan fossil assemblages may appear to be incomprehensible jumbles of skeletal remains, but the paleontological literature is crowded with attempts to determine how reliably these naturally accumulated ancient death assemblages (or thanatocoenoses) reflect their formerly living source associations. Modern thanatocoenoses recovered on the present sea floor, represent, in turn, the precursors for future fossil assemblages that will be identifiable, like their ancient counterparts, as layers in outcrops or cores. The maximum age of a shell in an active sedimentary environment is a measure of time-averaging. Based on radiocarbon data reported in the literature, dates compiled from many active deposits reveal a median duration of time-averaging in nearshore (<10 m) deposits of 1250 years, with 22% of the values less than 500 years and the mode equal to zero (Flessa and Kowalewski 1994).

Relative species abundance in fossil assemblages was previously thought to vary significantly as a result of modifications brought about by destructive taphonomic processes or time-averaging of multiple generations (Walker and Bambach 1971; Kowalewski 1996). Recently, however, statistical meta-analysis of molluscan data sets from different habitats has shown that time-averaged death assemblages retain a strong signal of the original rank order of species (Kidwell 2001). In particular, species rank-order correlations and live-dead agreement are higher for data sets recovered from meshes >1 mm and for muddy, coastal habitats. Live specimens in the size range of 1 mm often represent ecologically transient juveniles that can predominate within the associations. Similarly, dead shells in the size range of 1 mm are prone to transport out-of-habitat or chemical dissolution soon after burial. Therefore, sieving with 1 mm mesh retains the most volatile component of both biocoenosis and thanatocoenosis (Kidwell 2001).

Biostratinomy deals with taphonomic processes that affect organisms from death until final burial (Lawrence 1968). Biostratinomic processes that are relevant to molluscs and paleoenvironmental reconstructions are mainly: necrolysis, chemical dissolution, bivalve disarticulation, fragmentation, encrustation, bioerosion, and transport. Each process is independent of the others and produces observable features on molluscan shells.

Some features are the result of a combination of processes, but caution is required in interpreting such features. For example, necrolysis and transport in combination can lead to the disarticulation of bivalve shells. On the other hand, deducing the age of a shell from its degree of abrasion can be misleading, since old shells can escape abrasion by early burial (Powell and Davies 1990). Radiocarbon and amino acid dating have been used to test the true age of beach-collected shells that appeared old or young on the basis of relative abrasion, dissolution, wholeness, edge-rounding, bioerosion, encrustation, sheen, and color preservation. The only characteristics that were significantly correlated with age were discoloration and sheen loss (Powell and Davies 1990; Flessa *et al.* 1993).

Some processes are primarily environment-dependent (e.g., dissolution of carbonates occurs in undersaturated waters), whereas others are mainly time-dependent (e.g., abrasion and edge-rounding are observable features of mechanical strain due to transport). In the latter example, however, it is practically impossible to discriminate between the effects of short distance transport of long duration or long distance transport of short duration. Further, given equal distances, mass transport might not affect shells at all, while rolling and saltation of individual shells would be expected to cause considerable taphonomic effects.

The taphonomic signature is the overall picture of attributes and features observable on a dead or fossil shell. Most taphonomic studies proceed under the assumption that taphonomic features co-occur predictably, thereby defining "taphofacies." In this way, different depositional environments are hypothesized to generate different taphonomic signatures, which can then be used to characterize major environments of deposition (EODs) (Brett and Baird 1986; Powell *et al.* 2002). Experiments with shells deployed over two years in different EODs along onshore-offshore transects off Lee Stocking Island (Bahamas) and in the Gulf of Mexico apparently failed to confirm the hypothesis, however. In contrast, taphonomic processes appear to be unexpectedly complex and largely related to specific edaphic factors. For example, taphonomic alteration is greater on hard ground and on brine-seep sites than on terrigenous mud, regardless of water depth. Similarly, dissolution is less effective where burial is deeper, regardless of the specimen's position along the transects (Powell *et al.* 2002).

Attempts to quantify the taphonomic attributes of a thanatocoenosis or fossil assemblage include the use of taphonomic grades, which represent an arbitrary ranking of the quality of specimen preservation on an ordinal scale (Kowalewski *et al.* 1994). Taphonomic grades can describe the degree of preservation in different areas of the same shell, such as the umbo, the margin, and the inner sides of a bivalve (e.g., Davies *et al.* 1989), across different specimens of one species (e.g., Holland 1988), or within a fossil assemblage in general (Brandt 1989). There is no agreement on how many taphonomic grades should be recognized, though the literature reports two to seven ranks for different taphonomic features (e.g., Holland 1988; Brandt 1989; Powell and Davies 1990; Kowalewski *et al.* 1994; Kowalewski *et al.* 1995; Powell *et al.* 2002). Taphonomic grades can be defined separately for each taphonomic feature (Kowalewski *et al.* 1995), they can be combined by averaging several individual grades (e.g., Feige and Fürsich 1991), or they can be assigned directly (e.g., Flessa *et al.* 1993).

It is well known that shell morphology influences its mechanical behavior during transport (e.g., Muller 1957; Seilacher 1976; Kidwell *et al.* 1986). Though it is obvious that a thin, fragile bivalve is more susceptible to fragmentation than a thick-shelled gastropod, species have a less obvious, more variable susceptibility to discoloration, abrasion, and dissolution (Powell *et al.* 2002). In sum, every species is susceptible to different taphonomic processes in different ways.

Varied environments host species assemblages that are ecologically compatible with existing conditions. When dealing with recent and modern biota, knowledge of the ecological requirements of mollusc species can be used to infer their ecological compatibility with the rest of the association and the embedding sediments. Thus, knowing about ecological compatibility could help to discriminate between a transport of long duration over a small distance and a single, long-distance transport. It also explains why thanatocoenoses are usually dominated by poorly significant species with a wide environmental distribution (Picard 1965).

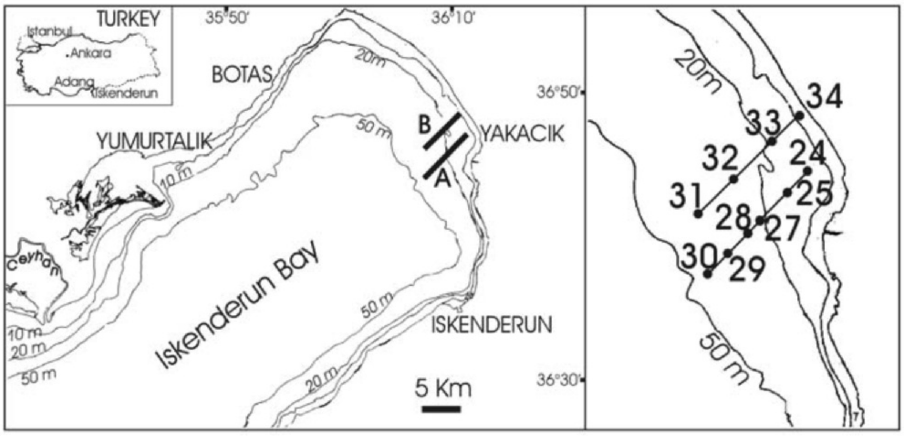

Figure 1. Map of Iskenderun Bay and location of the sampling sites; on right, details of the transects, showing the position of sampling sites.

3. SITE DESCRIPTION

Iskenderun Bay is a relatively long, narrow, and shallow basin extending into the coast of Turkey in the northeastern corner of the Mediterranean Sea (Figure 1). Along the southwestern coast of the bay, the Ceyhan River has formed a delta, which was abandoned and replaced by one immediately to the south. The Ceyhan River was a major source of terrigenous sediments deposited within the bay, since it drains a basin of 20,466 km^2 with an average annual discharge of 303 $m^3 s^{-1}$ and an annual sediment yield of 5462 x 10^3 t (Aksu *et al.* 1992). The maximum depth along the main axis of the bay is about 70 m, and it gradually increases to the southwest, toward the open sea, where the bottom descends to about 90 m.

The geomorphology and sedimentology of Iskenderun Bay are described in previous papers that resulted from a comprehensive, multidisciplinary investigation of its benthic environment (Yanko *et al.* 1994; Koral *et al.* 1997; Spezzaferri *et al.* 2000; Basso and Spezzaferri 2000). This paper will focus on two sampling transects (A and B) situated at the eastern end of Iskenderun Bay, not far from the village of Yakacık.

4. METHODS

Bottom sediment was sampled using a 10 liter Van Veen Grab. Sampling sites were spaced about 0.5 nautical mile (nm) apart, with the exception of sites 31-32-33 and 25-27, which were 1 nm away from one another. Only undisturbed samples in which the sediment surface was still recognizable and preserved have been considered for this study. Separate samples were taken from each grab for grain-size analyses, which were performed by sieving (Koral 1995). About 1 liter was taken for molluscan study, sieved on board using 1 mm mesh and immediately treated in a 4% formalin solution with sea water buffered with 20 grams of $Na_2B_4O_7$ (Na-borate hydrate) per liter in order to prevent dissolution of calcium carbonate. Total organic matter was measured as percentage of weight lost from the surface sediments upon ignition. Water transparency was evaluated with a Secchi disk. Depth, salinity, and temperature were measured with a CTD probe.

A ship-operated gravity core was recovered at site 29 in transect A. Details of the core study are given in Spezzaferri *et al.* (2000).

Entire shells and fragments were picked from the sediment. Since species differ in their morphological distinctiveness, the inclusion of easily recognized fragments may distort the taxonomic composition of the studied thanatocoenoses, possibly depressing estimates of diversity and evenness. In order to avoid this "*Chlamys* effect" (Kowalewski *et al.* 2003), bivalve fragments were retained only when they included the umbo and the hinge. Gastropod fragments were retained only when they consisted of at least 2/3 of the abapical portion including the aperture, or the apical portion with the apex (DiGeronimo and Robba 1976).

4.1 Specimen Counting

All species were identified and specimens counted, with articulated bivalves counted as one specimen. Abundance analysis of assemblages containing both bivalves and gastropods must be pursued with the understanding that the probability of sampling a disarticulated bivalve shell is twice the probability of sampling a gastropod shell, given equal original proportions, since gastropods do not undergo disarticulation (Gilinsky and Bennington 1994). For bivalves, the left/right valve ratio is useful for assessing possible transport (Schäfer and Craig 1972). As shell damage will occur much faster after disarticulation, left and right disarticulated valves were counted separately, then a total number of specimens was obtained by taking the number of the most abundant valve and adding half the number of the other. Each gastropod fragment was counted as one specimen (DiGeronimo and Robba 1976). The dominance (or percent abundance) of each species in the thanatocoenosis was then calculated.

4.2 Preservation Code

For statistical purposes, a preservation code corresponding to a combined taphonomic grade was assigned to each shell in a sample. The taphonomic features considered in the code assignment were: loss of periostracum, breakage, discoloration, dissolution, and abrasion (including edge-rounding).

The preservation codes range from 5 to 1 in order of decreasing preservation (Table 1). Only one preservation code (combined taphonomic grade) was assigned to each species in each sample, in order to allow comparison of several samples. Code 5 was reserved for species with at least one specimen found live in the sample.

Table 1. The four taphonomic features observed in molluscan shell assemblages and the combined preservation codes (1 to 4)

Feature \ Code	4	3	2	1
Loss of periostracum and/or ligament	Not complete	Complete	Complete	Complete
Breakage	Absent	< 25 %	25%< >75%	> 75%
Discoloration and loss of sheen	Original color and sheen	Fading color to white	White, complete loss of sheen	White to brown/black
Dissolution and/or abrasion	Absent	On < 25% of shell surface	On 25%< >75% of shell surface	On > 75% of shell surface

A natural, autochthonous molluscan population, in equilibrium with its environment, could show all degrees of shell preservation among dead molluscs, from fragments to poorly preserved to recently deceased. In contrast, species occurring only as poorly preserved shells might have reached that condition as a result of transport or because they were members of a pre-existing molluscan population that had undergone its own taphonomic deterioration. Giving such a species a mean preservation value by averaging values observed for each specimen would be misleading, since the same code could result from very different situations (e.g., a shell assemblage in which all specimens were uniformly recorded as preservation code 3 or a shell assemblage showing all degrees of preservation that ultimately averaged out to 3).

Therefore, in the case where conspecific specimens reveal different preservation codes, only the highest (best preserved) code was considered, since shells in better condition represent evidence of autochthony and a relatively recent death for the molluscs in question.

4.3 Status Code

Another code was assigned to each species reflecting its population structure in the death assemblage. If total population demographics were represented (shells of juveniles and adults mixed together), status code 3 was assigned. If adult shells only were found, status code 2 was applied, and if only juveniles were recovered, the designation was status code 1. The structure of a natural, autochthonous molluscan population in equilibrium with its environment will show a majority of juveniles mixed with some adult shells. A shell assemblage with juveniles and no adults of the same species can result from selective transport, but it could also be the consequence of immigration by larvae that, after metamorphosis, could not mature outside their appropriate environment. Plausible explanations for such a situation include (1) the inability of the biotope of immigration to satisfy the increasing trophic requirements of the immigrant molluscs, or (2) after settlement by the immigrant species, some variable within the new environment oscillates beyond its limits of tolerance. The status code, associated with the preservation code, better defines the autochthony of the shell assemblage and its equilibrium within the biotope.

4.4 Ecological Categorization

The third attribute assigned to each species in the thanatocoenosis concerns its ecology, including components based on (in order of increasing fidelity): substrate preference, occurrence in a particular marine benthic biocoenosis, and preferential or exclusive association with a biocoenosis (Pérès and Picard 1964; Picard 1965; Pérès 1982). Species were grouped into seven categories according to their substrate preference: (1) mixticolous species, including those characteristic of the coastal detritic biocoenosis (DC); (2) gravel-related species, including those characteristic of the biocoenosis of coarse sands and fine gravels under bottom currents (SGCF); (3) sand-related species, including those characteristic of the biocoenosis of muddy sands in sheltered areas (SVMC) as well as those of fine well-sorted sands (SFBC); (4) mud-related species, including those characteristic of the coastal muddy bottom biocoenosis VTC); (5) species occurring with photophilic algae and seagrass meadows (AP biocoenosis and HP complex); (6) species related to biogenic hard bottoms (coralligenous biocoenosis); and (7) species characteristic of a heterogeneous assemblage that testifies to sedimentary or trophic instability (PE of Picard (1965). Where the benthic environment undergoes an abrupt change, the PE group of species typically appears opportunistically before the establishment of the new biocoenosis (Di Geronimo and Robba 1989).

Species without precise ecological meaning ("lre" and "ssp" species of Picard 1965) and those for which no relevant ecological information was available were excluded from further consideration in order to simplify the structure of the data base and its interpretation (Basso and Corselli 2002). Interpretation of the results considered three groups of species separately: (1) those represented by well preserved or living populations (preservation code 4 and 5, status code 3); (2) those represented by well preserved shells of juveniles (preservation code 4 and 5, status code 1); and (3) those specimens corresponding to preservation codes 1 to 3.

Radiocarbon dating was performed at the Department of Physics, University of Rome "La Sapienza." Results were calibrated (Stuiver and Reimer 1993) and rounded in function of the error.

5. RESULTS

At the investigated stations, total organic matter ranged from 5 to 14%, salinity ranged from 38 to 39 ‰.

A short description of each grab sample follows.

5.1 Transect A

Grab 24 (14.5 m) contained olive-brown, gravelly sand (Figure 2) with coarse remains of molluscs and very abraded echinids, sparse bryozoan colonies among which were *Myriapora truncata*, *Sertella beaniana*, and *Margaretta cereoides*, fragments of serpulids and wood, and rare pebbles. The sediment was 49% biogenic (carbonate). Secchi depth: 3.5 m.

Grab 25 (15 m) contained gravelly sand with coarse remains of molluscs and echinids. Also in the residue were fragments of crustaceans, bryozoan colonies (*M. truncata*, *S. beaniana*, *M. cereoides*, and *Cellepora punicosa*), serpulids, and rare pebbles. The sediment was 55% biogenic. Secchi depth: 3.5 m.

Grab 27 (24.5 m) contained brown, muddy sand with rare biogenic remains, among which were small (1–2 cm) buried rhodoliths of *Peyssonnelia* sp. and *Lithothamnion minervae*, sparse molluscs, rare serpulids and bryozoans (*C. punicosa*). Secchi depth: 7.3 m. The scarce recovery prevented statistical elaboration of the molluscan assemblage, which is, however, related to the biocoenosis of the coastal detritic (DC). The shallow depth explains the influence from the AP-HP biocoenosis (photophilic algae and *Posidonietum*), attested by the identification of *Glans trapezia* and *Columbella rustica*. The sediment was 23% biogenic.

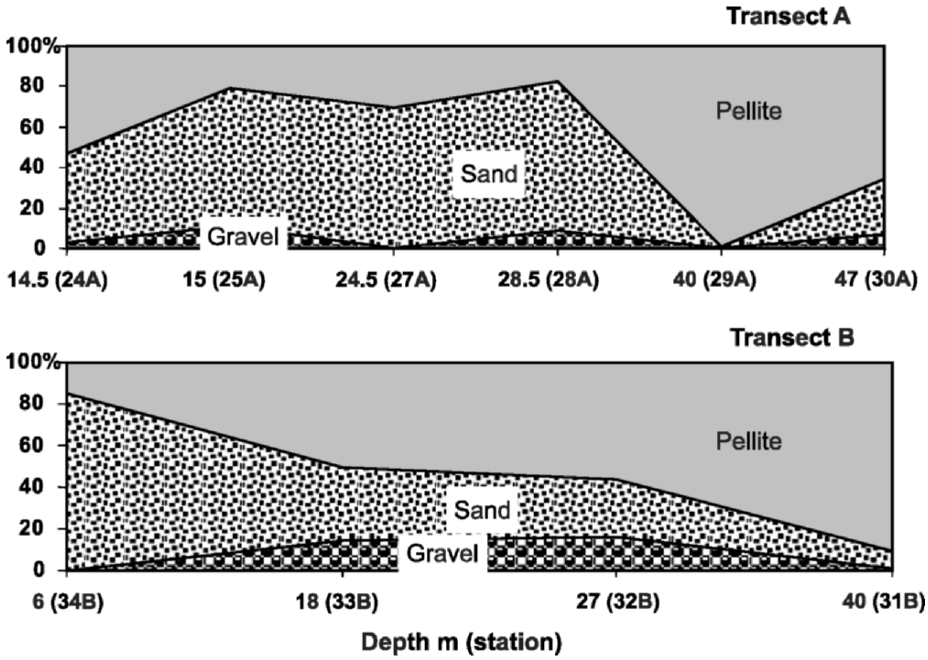

Figure 2. Grain size distribution. Pellite includes silt and clay (grains <64 µm).

Grab 28 (28.5 m) contained brown, gravelly sand and mud with shell fragments of molluscs and echinids. About 20% of the residue after sieving comprised pluricentimetric, buried rhodoliths. Also present were fragments of crustaceans, bryozoans (*S. beaniana* and *C. punicosa*), and fragments of serpulids as well as the scleractinian *Cladocora caespitosa*. The sediment was 6% biogenic. Secchi depth: 7.2 m.

Core 29 (39 m) contained greenish-grey pellite in the first 35 cm, with a concentration of bivalves (*Nucula sulcata*, *Nuculana pella*, and *Corbula gibba*) about 5 cm from the top. From 5 to 30 cm from the top, a sandy mud with remains of *C. caespitosa* overlay a 60 cm high coral colony (*C. caespitosa*) in life position. Radiocarbon dating gave an age of 5280±60 calBP for the base of the coral colony, and 3790±90 calBP for the top (Spezzaferri *et al.* 2000). The surface sediment was 25% biogenic.

Grab 30 (47 m) contained sandy mud with bioclasts. After sieving, more than 50% of the residue comprised molluscan (among which *Arca noae*) and echinid skeletal remains, about 15% calcareous red algae, and a diversified association of bryozoan colonies (*S. beaniana*, *M. truncata*, *M. cereoides*, and *C. punicosa*). Also present were some fragments of *C. caespitosa*, carbonized wood, serpulids, and rare pebbles. A decimetric valve of a poorly preserved *Spondylus* gave a radiocarbon age of 300±50 calBP (Spezzaferri *et al.* 2000). The sediment was 65% biogenic. Secchi depth: 7.1 m.

5.2 Transect B

Grab 31 (40 m) contained olive-gray sandy pellites with sparse biogenic remains. The scarcity of recovered materials did not allow statistical treatment of data, however, most of the best preserved shells belong to mud-related species like *Nucula sulcata* and *Abra alba*, while the other poorly preserved specimens belong to sand-loving species like *Tellina pulchella*. The sediment was 49% biogenic.

Grab 32 (27 m) contained brown gravelly-sandy pellite. The coarse, biogenic fraction was made up of bryozoan colonies (*S. beaniana*, *M. cereoides*, and *C. punicosa*). Crustacean remains, fragments of scleractians (*C. caespitosa*), and serpulids were also identified in the residue. The sediment was 33% biogenic.

Grab 33 (18 m) contained light brown sandy-gravelly pellite. In the residue, molluscs dominated with remains of echinids, about 20% calcareous red algae, 20% large pebbles. Fragments of crustaceans, bryozoans (*S. beaniana*), and scleractinian (*C. caespitosa*) also occurred in the residue. The sediment was 23% biogenic. Secchi depth: 4 m.

Grab 34 (6 m) contained brown, muddy sand with sparse molluscan and echinid remains but abundant plant fibers. The sediment was 17% biogenic.

The molluscan species identified in these 10 grab samples from the two transects, their abundance, dominance, preservation, and status are listed in Table 2. Since no living mollusc was recovered, the preservation code of 5 is absent from our data base. Only the stations with more than 1000 counted specimens have been studied further on the basis of their substrate preference. The shell assemblage at each station was divided into three groups: shells with preservation code <4 (group C<4); those well preserved, occurring as populations (group S3C4); and those well preserved, occurring as juveniles only (group S1C4) (Figure 3).

We observe a parallel trend in the two transects: the coralligenous-related species occur in both transects within the S3C4 and C<4 species groups. No coralligenous-related species occur in the S1C4 groups in either transect. The opposite trend is observed for the mud-loving species. They are absent or minor components in the C<4 groups, but increasing in the S3C4 and S1C4 groups. At station 25, the change from a well-preserved coralligenous-related population to mud-loving juveniles is particularly dramatic.

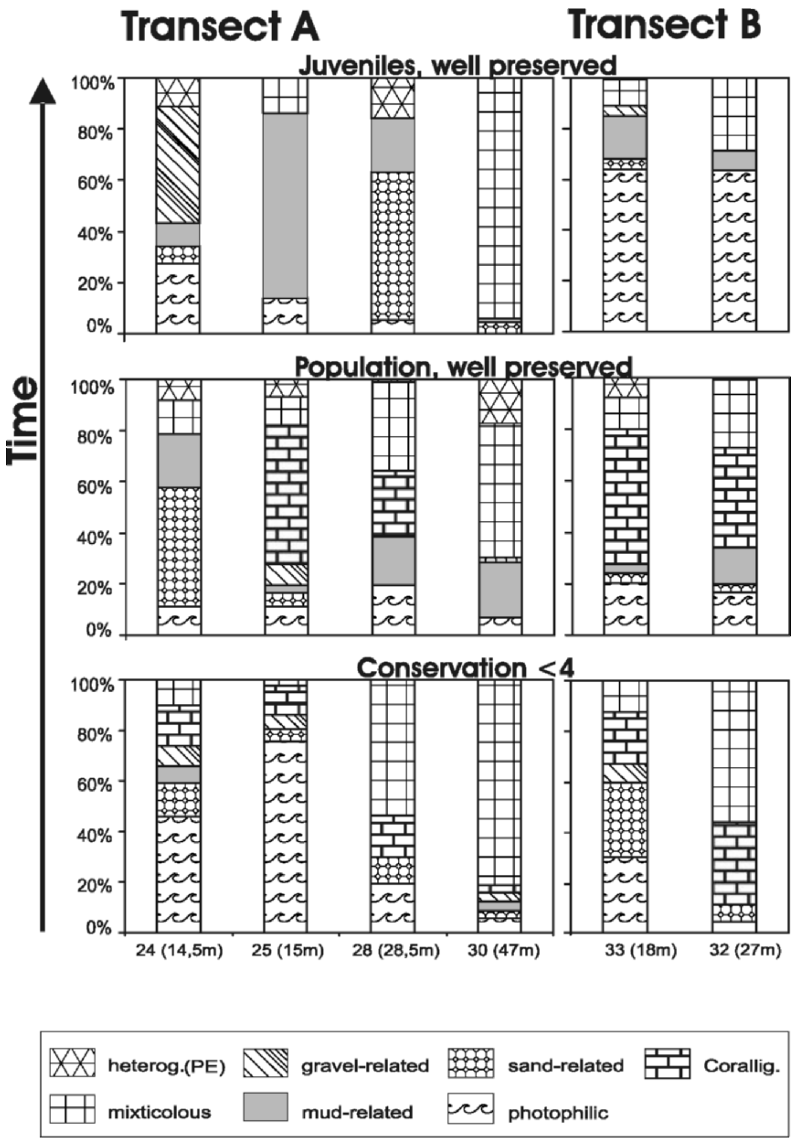

Figure 3. Distribution of molluscan associations as a function of Status and Preservation.

Table 2. Molluscan death assemblages at the investigated stations indicating abundance (A), dominance (Dm), status (St), and preservation (C). Lessepsian immigrant species in **bold**.

TRANSECT A

Station:	24				25				27				28				30			
Species:	A	Dm	St	C	A	Dm	St	C	A	Dm	St	C	A	Dm	St	C	A	Dm	St	C
Abra alba	27	1.91	3	4	4	0.08	3	4					38	2.96	3	4	203	7.67	3	4
Abra prismatica	2	0.14	2	3																
Abra tenuis	5	0.35	3	4													1	0.04	1	4
Acanthocardia echinata																	3	0.11	1	4
Acanthocardia paucicostata	8	0.57	3	4													4	0.15	1	4
Acmaea virginea	2	0.14	1	2	2	0.04	1	4												
Alvania lineata	6	0.43	1	3	50	0.98	3	4					66	5.15	2	4	10	0.38	2	3
Alvinia subareolata	8	0.57	3	4									1	0.08	1	3				
Anachys savignyi					4	0.08	2	4												
Anadara diluvii																	3	0.11	1	3
Anodontia fragilis	104	7.38	3	4	6	0.12	3	3					5	0.39	2	4	3	0.11	2	3
Anomia ephippium					1	0.02	1	4												
Aporrhais pespelecani																	1	0.04	1	4
Arca noae	2	0.14	2	3	4	0.08	3	3					13	1.01	3	3	14	0.53	3	4
Arca tetragona					6	0.12	1	4									1	0.04	1	4
Barbatia scabra	1	0.07	1	3					2	14.29	1	4					2	0.08	1	3
Bittium reticulatum reticulatum	529	37.52	3	4	2787	54.75	3	4	5	35.71	3	4	143	11.15	3	4	276	10.43	3	4
Bulla striata	25	1.77	1	4	144	2.83	1	4	4	28.57	1	4	37	2.89	1	4	20	0.76	3	4
Callista chione													1	0.08	1	3				
Cerithidium submamillatum	13	0.92	1	3	57	1.18	1	2												
Cerithiopsis tubercularis	2	0.14	1	3	31	0.61	1	2					4	0.31	1	3	1	0.04	1	2
Cerithiopsis tubercularis horrida	15	1.06	3	3	407	8.00	3	4					142	11.08	3	4	24	0.91	3	4
Cerithium scabridum	3	0.21	1	3	311	6.11	1	4					13	1.01	1	4				
Cerithium vulgatum	4	0.28	1	4	56	1.10	1	4					4	0.31	1	4				
Chama gryphoides									1	7.14	2	4								
Chlamys glabra													3	0.23	1	4	13	0.49	1	4
Chlamys multistriata	1	0.07	2	2									15	1.17	3	4	29	1.10	3	4
Chlamys pesfelis					1	0.02	1	3												
Chlamys varia													2	0.16	1	3				

Station:	24				25				27				28				30			
Species:	A	Dm	St	C	A	DM	St	C	A	DM	St	C	A	Dm	St	C	A	Dm	St	C
Circulus striatus	1	0.07	1	4																
Chrysallida excavata					4	0.08	1	3												
Chrysallida maiae					26	0.51	3	4									3	0.11	2	4
Clanculus corallinus					4	0.08	1	2					1	0.08	1	2				
Columbella rustica									1	7.14	1	3					2	0.08	3	3
Conus ventricosus	24	1.70	3	2	24	0.47	3	2					38	2.96	3	2	19	0.72	3	4
Corbula gibba	21	1.49	3	4	51	1.00	3	4					6	0.47	3	4	232	8.77	3	4
Ctena decussata	12	0.85	3	4	4	0.08	3	4					4	0.31	2	3				
Dentalium inaequicostatum	1	0.07	1	2	1	0.02	1	2					1	0.08	2	4	5	0.19	3	4
Dentalium vulgare	11	0.78	1	2	16	0.31	3	3												
Diplodonta rotundata	7	0.50	3	3	7	0.14	1	4									2	0.08	1	3
Divaricella divaricata	26	1.84	3	4	20	0.39	3	4					1	0.08	2	4				
Donax venustus	2	0.14	1	3																
Dosinia lupinus	2	0.14	1	4	1	0.02	1	3												
Emarginula elongata	2	0.14	1	2	2	0.04	1	1					3	0.23	1	2				
Epitonium clathratulum					2	0.04	1	2					1	0.08	2	4				
Eulimella scillae					2	0.04	1	4												
Evalea subulata																	1	0.04	2	4
Fenella virgata	3	0.21	1	4	3	0.06	1	4					5	0.39	1	3	3	0.11	1	4
Fusiturris undatiruga																	2	0.08	1	3
Gibbula albida	2	0.14	1	3	2	0.04	1	3												
Gibbula ardens	19	1.35	1	3	19	0.37	1	3					14	1.09	1	3	1	0.04	1	3
Glans trapezia	28	1.99	3	4	83	1.63	3	4	1	7.14	2	4	110	8.58	3	4	98	3.70	3	4
Glycymeris glycymeris	18	1.28	1	4	60	1.28	3	4												
Gouldia minima	34	2.41	3	4	81	1.59	3	4					196	15.29	3	4	760	28.72	3	4
Haminoea navicula	3	0.21	1	4									3	0.23	1	4	3	0.11	1	3
Hiatella rugosa																	3	0.11	3	4
Hinia reticulata	5	0.35	1	4	5	0.10	1	4					3	0.23	1	4	21	0.79	1	4
Jujubinus corallinus	1	0.07	1	1	1	0.02	1	1					4	0.31	1	3	4	0.15	3	3
Lepton nitidum	26	1.84	1	4	6	0.12	1	4					1	0.08	2	4	3	0.11	1	3
Lepton squamosum	4	0.28	3	4	4	0.08	3	4												
Limatula sulcata	2	0.14	2	3	1	0.02	1	3												
Lissopecten hialinus																	5	0.19	1	3
Loripes lacteus	3	0.21	1	3																

Station:	24				25				27				28				30			
Species:	**A**	**Dm**	**St**	**C**	**A**	**Dm**	**St**	**C**	**A**	**Dm**	**St**	**C**	**A**	**Dm**	**St**	**C**	**A**	**Dm**	**St**	**C**
Mamilloretusa mamillata					18	0.35	2	1												
Mangelia attenuata					2	0.04	1	2												
Mangelia wareni	3	0.21	1	2	3	0.06	1	2					7	0.55	1	3	13	0.49	1	3
Melanella polita					6	0.12	1	4												
Merelina tessellata	2	0.14	1	3	74	1.45	3	1					1	0.08	1	2				
Mitra cornicula	15	1.06	1	4									16	1.25	3	3	13	0.49	1	4
Musculus subpictus													1	0.08	1	4	7	0.26	1	3
Mysella bidentata	2	0.14	1	3	1	0.02	2	4									4	0.15	1	4
Nassarius mutabilis	1	0.07	1	4																
Naticarius punctatus													3	0.23	1	4				
Neverita josephinia	1	0.07	2	4																
Nucula sulcata	26	1.84	3	4	18	0.35	3	4					71	5.54	3	4	152	5.74	3	4
Nuculana pella	5	0.35	1	4									3	0.23	1	4	48	1.81	3	4
Odostomia plicata	1	0.07	1	1	22	0.43	2	4					3	0.23	2	3	1	0.04	1	3
Ostrea edulis													4	0.31	1	4				
Parvicardium exiguum	8	0.57	1	3	3	0.06	1	3					6	0.47	1	3	96	3.63	1	3
Pitar rudis rudis																	236	8.92	1	4
Plagiocardium papillosum	1	0.07	1	3	2	0.04	3	4									85	3.21	3	4
Raphitoma echinata	4	0.28	1	3	4	0.08	1	3					4	0.31	1	3	6	0.23	3	4
Raphitoma leufroyi													1	0.08	2	4				
Retusa mariateresae					3	0.06	1	3												
Ringicula auriculata																	1	0.04	2	4
Rissoina bruguierei	10	0.71	1	3	10	0.20	1	1					1	0.08	1	4				
Smaragdia viridis	12	0.85	1	4	12	0.24	1	4												
Spondylus gaederopus																	1	0.04	2	1
Striarca lactea					1	0.02	1	4					4	0.31	2	3	4	0.15	1	4
Strombus persicus	10	0.71	1	4	40	0.79	3	4												
Tellimya ferruginosa	1	0.07	2	3																
Tellina nitida	4	0.28	3	4													1	0.04	2	3
Tellina pulchella	78	5.53	3	4	17	0.33	3	4					11	0.86	1	4	11	0.42	1	4
Tellina serrata																	1	0.04	2	4
Timoclea ovata																	4	0.15	1	4
Tornus subcarinatus					16	0.31	2	3												
Tricolia pullus	5	0.35	1	1	5	0.10	1	1												

Station:	24				25				27				28				30			
Species:	**A**	**DM**	**St**	**C**	**A**	**Dm**	**St**	**C**	**A**	**Dm**	**St**	**C**	**A**	**Dm**	**St**	**C**	**A**	**Dm**	**St**	**C**
Triphora perversa	1	0.07	1	3	1	0.02	1	3					3	0.23	1	3				
Trophonopsis muricata					2	0.04	1	3					18	1.40	3	4	3	0.11	1	3
Turboella incospicua	1	0.07	2	4																
Turboella lineolata					9	0.18	1	4												
Turboella marginata	22	1.56	1	3	22	0.43	1	3					7	0.55	1	4	6	0.23	1	3
Turboella radiata	101	7.16	3	4	101	1.98	3	4					8	0.62	3	4	68	2.57	3	4
Turbona cimex	8	0.57	1	1	160	3.14	3	3					111	8.66	3	4	22	0.83	3	3
Turbonilla lactea	13	0.92	1	3	13	0.26	1	3					2	0.16	1	3	3	0.11	1	3
Turbonilla rufa					4	0.08	1	3												
Turritella comunis																	3	0.11	1	3
Turritella turbona					2	0.04	1	2					30	2.34	1	3	20	0.76	1	3
Typhinellus sowerbyi													1	0.08	1	4	7	0.26	1	4
Venericardia antiquata	46	3.26	1	3	204	4.01	2	3									1	0.04	1	3
Venerupsis geographica	12	0.85	1	3	10	0.20	1	4												
Venus casina	2	0.14	1	4	2	0.04	1	2	14				5	0.39	3	4	5	0.19	3	3
TOTAL species 111																				
TOTAL specimens 10291	1409				5087				28				1199				2596			

TRANSECT B

Station:	31				32				33				34			
Species:	**A**	**Dm**	**St**	**C**	**A**	**Dm**	**St**	**C**	**A**	**Dm**	**St**	**C**	**A**	**Dm**	**St**	**C**
Abra alba	2	15.38	1	4	6	0.45	2	4								
Abra prismatica	1	7.69	1	3												
Abra tenuis	2	15.38	1	4												
Acanthocardia echinata									1	0.05	2	3				
Acmaea virginea					14	1.05	1	4	1	0.05	1	2				
Alvania lineata					101	7.55	3	4	38	1.92	3	2				
Alvinia subareolata									1	0.05	1	2				
Anachys savignyi					1	0.07	2	4	1	0.05	2	4				
Anodontia fragilis					3	0.22	2	3	1	0.05	2	3				
Arca noae					3	0.22	3	4	5	0.25	1	3				
Barbatia scabra					6	0.45	3	4								
Bittium reticulatum reticulatum					122	9.12	3	4	1457	73.58	3	4				

Station:	31				32				33				34			
Species:	A	Dm	St	C	A	Dm	St	C	A	Dm	St	C	A	Dm	St	C
Bulla striata					51	3.81	1	4	64	3.23	1	4				
Callista chione									1	0.05	1	3				
Cerithidium submamillatum					14	1.05	3	3	5	0.25	1	2				
Cerithiopsis tubercularis					6	0.45	1	3	5	0.25	1	2				
Cerithiopsis tubercularis horrida					145	10.84	3	4	199	10.05	3	4				
Cerithium scabridum					27	1.93	1	3	39	1.97	1	4				
Cerithium vulgatum					5	0.37	1	4	23	1.16	1	4				
Chlamys glabra					1	0.07	2	4								
Chlamys multistriata					2	0.15	2	4								
Chrysallida excavata					1	0.07	1	2								
Chrysallida maiae									5	0.25	3	4				
Circulus striatus									7	0.35	1	2				
Columbella rustica					1	0.07	1	3								
Conus ventricosus					19	1.42	1	4	55	2.78	1	3				
Corbula gibba									29	1.46	3	4				
Ctena decussata					10	0.75	3	4	15	0.76	3	4				
Dentalium vulgare									1	0.05	2	3				
Divaricella divaricata					1	0.07	2	4	4	0.20	2	4				
Donax venustus													1	20	1	4
Dosinia lupinus									6	0.30	1	4				
Emarginula elongata					4	0.30	3	3								
Fenella virgata									1	0.05	1	3				
Gibbula ardens					31	2.21	1	2	35	1.77	1	4				
Glans trapezia					58	4.33	3	4	45	2.27	3	4				
Glycymeris glycymeris									7	0.35	3	2				
Gouldia minima					98	7.32	3	4	47	2.37	3	4				
Haminoea navicula					7	0.52	1	4	2	0.10	1	3				
Hinia reticulata					9	0.67	1	4	16	0.81	1	4				
Jujubinus corallinus					19	1.42	1	3	5	0.25	1	1				
Lepton nitidum									4	0.20	1	4				
Lima lima					1	0.07	1	4								
Mamilloretusa mamillata									1	0.05	2	2				
Mangelia wareni					13	0.97	1	3	10	0.51	1	3				
Melanella polita									3	0.15	1	3				

Station:	31				32				33				34			
Species:	A	Dm	St	C	A	Dm	St	C	A	Dm	St	C	A	Dm	St	C
Merelina tessellata					5	0.37	3	3	23	1.16	3	2				
Mitra cornicula					14	1.05	1	3	26	1.31	1	3				
Naticarius punctatus					1	0.07	1	4								
Nucula sulcata	5	38.46	1	4	52	3.89	3	4	14	0.71	3	4				
Nuculana pella	1	7.69	1	4	3	0.22	3	4	1	0.05	1	4				
Odostomia plicata					3	0.22	2	2	19	0.96	3	4				
Parvicardium exiguum									8	0.40	1	4				
Pitar rudis rudis					4	0.30	1	4								
Plagiocardium papillosum					14	1.05	1	4	6	0.30	1	4				
Raphitoma echinata					11	0.85	1	3	7	0.35	1	2				
Retusa mariateresae									2	0.10	2	3				
Rissoina bruguierei					41	2.91	1	4	88	4.44	1	4				
Spondylus gaederopus					1	0.07	2	1								
Smaragdia viridis									15	0.76	1	4				
Striarca lactea									4	0.20	1	2				
Strombus persicus					1	0.07	1	4	29	1.46	1	4				
Tellina crassa									2	0.10	2	4				
Tellina pulchella	2	15.38	1	3	3	0.22	3	4	1	0.05	2	3	3	60	3	4
Timoclea ovata					1	0.07	1	3								
Tricolia pullus					2	0.15	1	1	30	1.52	1	3				
Triphora perversa					5	0.37	1	1	12	0.61	1	2				
Trophonopsis muricata					21	1.49	1	4	22	1.11	1	2				
Turboella marginata					5	0.37	1	4	84	4.24	1	4				
Turboella radiata					150	11.21	3	4	324	16.36	3	4				
Turbona cimex					165	12.33	3	3	57	2.88	3	3				
Turbonilla lactea					5	0.37	1	3	33	1.67	1	2				
Turbonilla rufa					4	0.30	1	3	3	0.15	1	3				
Turritella turbona					40	2.85	1	3	11	0.56	1	2				
Typhinellus sowerbyi					2	0.15	1	2								
Venericardia antiquata					3	0.22	1	4	32	1.62	3	4				
Venus casina									6	0.30	1	4	1	20	1	3
TOTAL species 77																
TOTAL specimens 4351	13				1335				2998				5			

6. DISCUSSION

Several constraints affect the temporal reconstruction of coastal changes in Iskenderun Bay. The first is the absolute dating of *Spondylus* in station 30 (300±50 calBP). Molluscan shells buried under 1–3 cm of mud at other recovery sites in Iskenderun Bay gave consistent radiocarbon ages (calcareous red algae, *Spondylus*, *Glycymeris* from 250 to 900 calBP). This line of evidence allows us to assign approximately this age interval to the C<4 groups of Figure 3.

The second is the occurrence of Lessepsian immigrants, which entered the Mediterranean Sea after the opening of the Suez channel in 1869 (Table 2). Their appearance in the study area should not pre-date this event, even if some accidental transport cannot be excluded. The four Lessepsian immigrants recognized in the area are mostly well preserved populations and juveniles (79% and 14% respectively), allowing us to estimate that the groups S3C4 and S1C4 represent approximately the last two centuries. Among the immigrants, *Cerithium scabridum* and *Fenella virgata* occur only as juveniles, mainly in the AP-HP assemblages, which provides evidence of their incomplete adaptation to their new environment. In fact, the whole S1C4 group is interpreted as being composed of species that recently attempted to settle at the studied stations, but failed to reach the adult size. Whether they represent some potential, stable components of a future benthic association will depend upon several factors, but such a question requires further investigation.

Two stations located at approximately the same depth (27–28 m) in the two transects (Figure 3) can be compared. Station 28 was characterized by dominant mixticolous species, with AP-HP-, sand- and coralligenous-related species (C<4) totalling together about 40%. Increasingly muddy sedimentation resulted in the appearance of mud-related species in the S3C4 group, replacing the sand-related ones. Well preserved juveniles (S1C4) were dominated by sand-related species, accompanied by mud-related and PE-related species. Mixticolous species disappeared. The sediment of station 28 shows a peak of terrigenous sand apparently not consistent with the "normal" onshore-offshore gradient (Figure 2, compare with Transect B), and this peak is accompanied by species related to sediment instability (PE). These elements suggest that the sand has increased at Station 28 very recently, probably during the last century.

At station 32, the largest biogenic remains composing the gravel fraction allowed the settlement of the coralligenous-related species, becoming dominant in the S3C4 group. The mud increase is responsible for the appearance of the mud-related species, which partially substitute for the mixticolous molluscs in the S3C4 group. The coralligenous species disappear in the S1C4 group, where AP-HP-related species become dominant. Apparently, this is ecologically inconsistent with the mixed sediment of station 32. However, since any taphonomic evidence of transport is lacking, and since only two species (*Rissoina bruguierei* and *Venericardia antiquata*) make up the photophilic fraction of the S1C4 group, they should be considered accidental immigrants from neighboring biotopes into the station 32 environment.

Previous investigations on the Holocene environmental evolution of Iskenderun Bay identified three pulses of muddy sedimentation into the bay: an older pulse (about 3700 calBP) possibly related to the large-scale forest clearing called the Beyşehir Occupation Phase (defined by Bottema and Woldring 1990); another pulse corresponding to a major delta progradation of the Ceyhan River at about 2140 calBP; and a younger pulse following the diversion of the Ceyhan River mouth toward Yumurtalik, from the Middle Ages to 1935 (Spezzaferri *et al.* 2000). This younger pulse is recorded in the molluscan thanatocoenoses of the studied stations.

7. CONCLUSIONS

The use of combined codes in the paleoenvironmental study of fossil molluscan assemblages allows a large amount of data to be treated relatively simply. The results of taphonomic analysis, interpreted in the light of the ecological compatibility of the species, allows one to discriminate gradients and changes on a time-scale of centuries.

More detailed discussion of paleoenvironmental evolution at each station was intentionally avoided as being too far from the primary aim of this paper. In the particular case history discussed here, a few absolute radiocarbon dates combined with evidence of recovered Lessepsian immigrants allowed the establishment of a temporal framework for our study. Comparison with data from the literature then confirmed the validity of our results. When applied to other sites with different geological/environmental settings, analysis might have been conducted also on the basis of trophic guilds, tier categories, etc., or it might have required additional radiocarbon data, or comparison of groups on the basis of different combinations of preservation and status codes, and ecological categories.

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3

CLIMATE MODELING RESULTS FOR THE CIRCUM-PONTIC REGION FROM THE LATE PLEISTOCENE TO THE MID-HOLOCENE

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Abstract: Three-dimensional mathematical climate models (referred to as general circulation models or GCMs) have been used (1) to assess past and future climates under prescribed boundary conditions and (2) to determine climatic equilibrium responses to component changes. With geographic focus on the Circum-Pontic region, this paper examines the results of several GCMs for periods of extreme climate during the late Pleistocene and mid-Holocene. Climatic characteristics were prepared to simulate conditions during the Last Glacial Maximum (21 ky calBP/18 ky BP) and the Holocene warm event (6 ky calBP/5.3 ky BP). The results yield possible temperature and precipitation distributions for the times in question compared to those of today.

Keywords: paleoclimate modeling, PMIP (Paleoclimate Modeling Intercomparison Project), Circum-Pontic region, general circulation models, Last Glacial Maximum, Holocene warm event

1. INTRODUCTION

Climate is caused by the interaction of solar radiation with the atmosphere and terrestrial geography. Long-term orbital variations produce quasi-periodic variations in the seasonal and latitudinal distribution of solar radiation (insolation), while atmospheric composition influences terrestrial heating by controlling the absorption and transmission of incoming insolation and outgoing terrestrial radiation.

The configuration of the continents did not change during the Quaternary, apart from the waxing and waning of continental ice sheets and associated

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changes in sea level and coastal zone geography. Boundary conditions (insolation, greenhouse gas concentrations, land surface changes, and aerosol concentrations) combined to affect atmospheric and oceanic circulation, as well as the extent and height of the ice sheets. Global ice volume lags behind insolation changes by some thousands of years, so that within the relatively short time scale of the Late Pleistocene and Holocene, insolation and ice sheets acted as independent controls on the atmosphere and ocean.

To evaluate the effects of changing CO_2 and insolation, as well as other factors, on climate, one must rely upon the use of mathematical models. The most complete climate system models are constructed by discretizing, then numerically solving equations representing the basic laws governing the behavior of the atmosphere, ocean, and land surface. Such three-dimensional models have come to be known as general circulation models (or GCMs). Because of the large computing requirements of current climate numerical experimentation, horizontal resolution has been limited to about 2–4° of latitude and longitude, and vertical resolution has been limited to about 10–20 layers in both atmosphere and ocean. This level of resolution is very coarse, however. It does not define regional and local peculiarities of the surface very well, and it creates problems of parameterization for important motions and effects whose scales are smaller than the resolvable size.

Many experiments relevant to paleoclimate problems have been carried out, mostly to determine the equilibrium response to changes in solar energy flux, CO_2, and properties of the surface, but because of the large heat capacity of the deep ocean, it is very expensive to run a coupled atmosphere-ocean model to equilibrium. A simpler and less costly approach is to use a climate model that includes only the atmosphere, the land, and an ocean consisting of a single inert layer with a thickness of about 50 m. Simpler yet is to prescribe the spatial distribution of the sea surface temperature (SST) as boundary conditions for a model of the atmosphere. For many practical applications, it is sufficient to represent the seasonal variation of a meteorological regime.

These methods allow one to study the sensitivity of the equilibrium response, including the role of various feedback cycles that amplify or decrease the basic response to forcing. Among the feedback cycles working solely within the framework of the atmosphere are:

(1) *Water vapor increase as temperature rises.* As water vapor is a greenhouse gas, it enhances warming.

(2) *Enhanced poleward heat flux* decreases the latitudinal temperature gradient, which then decreases the intensity of atmospheric waves and cyclones, thus decreasing the poleward heat flux, etc.

(3) *Cycles connecting cryosphere and atmosphere.* Decrease in the extent of snow and ice cover associated with warming also enhances warming by decreasing the reflectivity of the surface, and vice versa.

(4) *Climate and global carbonate system.* Temperature change effects changes in the carbonate system, which is reflected in atmospheric CO_2 concentration, which in turn enhances the changes in surface temperature, etc.

Such experiments are relatively inexpensive compared with a full coupled run to equilibrium or a transient experiment where modeled climate response to prescribed changes of insolation, greenhouse gases, or land surface albedo is simulated over thousands of years. The assumption that the ocean does not change is a potential source of uncertainty because a fixed SST simulation does not permit the full expression of many kinds of feedback, but on the other hand, choosing real SSTs as a boundary condition limits modeling errors.

Global modeling studies have recently been compared across a number of special projects, such as the Coupled Model Intercomparison Project (CMIP), the Atmospheric Model Intercomparison Project (AMIP), and the Paleoclimate Modeling Intercomparison Project (PMIP). In 1995, the JSC/CLIVAR (Joint Scientific Committee/Climate Variability and Prediction) Working Group on Coupled Models, part of the World Climate Research Program, established CMIP in order to provide climate scientists with a database of coupled GCM simulations under standardized boundary conditions (see Meehl *et al.* 2000). CMIP may be regarded as an analog of AMIP (see Gates *et al.* 1999). In AMIP simulations, sea ice and sea surface temperatures are prescribed to match recent observations, and the atmospheric response to these boundary conditions is studied; in CMIP, the complete physical climate system, including the oceans and sea ice, adjusts to prescribed atmospheric concentrations of CO_2. PMIP was intended to model past climate events, with a focus on data-model and model-model comparison (Joussaume and Taylor 1995).[1]

Running several models under the same conditions helps explain how differences in model parameters and resolution influence responses, and it better assesses a model's ability to simulate climate changes. The starting point for model-data comparison is an assumption that climate models form a statistical ensemble. At each site, a model's calculated average values and the inter-model dispersion can be characterized by the standard deviation of value scores.

This paper gives an overview of the modeling of extreme climates during the last ~20,000 years, focusing on the Circum-Pontic Region (henceforth CPR). The CPR extends from the Manych-Kerch Gateway through the Black Sea, the Marmara Gateway, and the Aegean Sea to the Levantine basin. Climate model results, based on PMIP data, allow an evaluation of the range of possible climate variations affected by natural external forcing. Important results of paleoclimate modeling have been discussed by Hewitt and Mitchell (1998), Joussaume (1999), Joussaume *et al.* (1999), Otto-Bliesner (1999), Valdes and Glover (1999), and Braconnot *et al.* (2000). These authors have noted both the planetary-scale effects of orbital forcing as well as the regional-scale effects due to feedback from the ocean and vegetation.

2. THE LAST QUATERNARY COLD EVENT

The Quaternary has been characterized by both cold and warm phases. One candidate for a forcing agent that could produce such pronounced global climate variations is the Milankovitch mechanism (Berger and Loutre 1992). According to this theory, the Earth's orbital parameters change due to the influence of the Moon, Sun, and planets. Over 100,000- and 400,000-year periods, eccentricity slowly varies, inducing small changes in the annual mean total insolation received by the Earth. Obliquity oscillates from 22° to 25° over a 41,000-year period, and the position of the equinoxes precesses relative to the perihelion with 19,000- and 23,000-year periodicities. Obliquity and precession do not lead to global changes in annual mean energy but strongly modulate the seasonal pattern of insolation. Orbitally-induced changes in insolation can be amplified by internal feedbacks generating behavior resembling the so-called 'red' noise, however, climatic response to variation in insolation can be distinguished from noise at times when orbital agents work synchronously.

Table 1. PMIP models that provided data for this paper. Resolution designations depend on kind of GCM model: either spectral modes or number of cells determined by lat. x long. grid points.

Model Designation	**Resolution & # of Vertical Levels (L)**
BMRC Bureau of Meteorology Research Center (Australia)	R21L9
CCC2 Canadian Center for Climate Modeling and Analysis (Canada)	T32L10
CCM3 NCAR Climate Community model (USA)	T42L18
CCSR1 Center for Climate System Research (Japan)	T21L20
CSIRO Commonwealth Scientific and Industrial Research Organization (Australia)	R21L9
ECHAM3 Max-Plank Institut für Meteorologie (Germany)	T42L19
GEN2 National Center for Atmospheric Research (USA)	T31L18
GFDL Geophysical Fluid Dynamics Laboratory (USA)	R30L20
GISS Goddard Institute for Space Studies (USA)	72*46L9
LMCELMD5 Laboratory of Dynamic Meteorology (France)	64*50L11
MRI2 Meteorological Research Institute (Japan)	72*46L15
MSU Moscow State University (Russia)	18*24L3
UGAMP Universities Global Atmospheric Modeling Programme (UK)	T42L19
UIUC11 University of Illinois (USA)	72*46L11
UKMO UK Meteorological Office Unified model (UK)	96*73L19
YONU Yonsei University (Korea)	72*46L8

Such an effect occurred at the transition from the cold late Pleistocene to the warm Holocene. Another example is the transition from marine isotope stage 5e to 5d (126 and 115 ky BP), when radiative forcing, amplified by a change in CO_2, was sufficient to generate a fundamental climatic shift from interglacial to glacial mode (Schlesinger and Verbitsky 1996). Of course, such transitions were not gradual; they were complicated by short-term contingent (possibly random) events (e.g., the Allerød-Younger Dryas cycle). Many authors link the origin of these cycles to the behavior of the Atlantic thermohaline circulation (Sarnthein *et al.* 2000).

The last cold event of the Late Pleistocene (a cold extreme during the Last Glacial Maximum, or LGM) occurred approximately 21 ky calBP, i.e., 18 ky BP in radiocarbon years (Bard *et al.* 1990). (This paper employs calBP dates, which allow use of the absolute solar insolation calendar.) It involved substantial changes in surface boundary conditions: ice sheet extent and height, SSTs, albedo, sea level, and concentration of greenhouse gases and aerosols, but only minor changes in solar radiation (for example, during the northern summer, the solar energy deficit received by the Earth was about 2–4 Wm^{-2}).

Within the PMIP, a specific set of boundary conditions was prepared to simulate the final Pleistocene cold event. Designations and brief descriptions of the PMIP GCMs used in this paper are listed in Table 1.

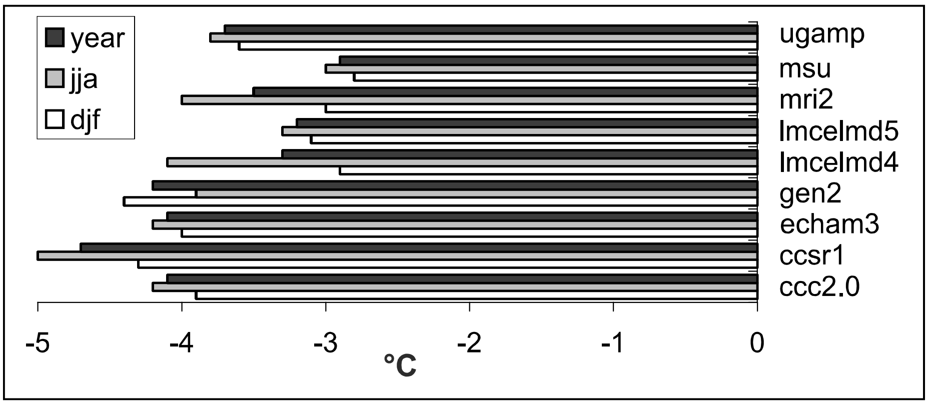

Figure 1. Annual and seasonal mean global cooling (° C) simulated for the late Pleistocene cold event (21 ky calBP vs. modern control) from different PMIP GCMs forced by CLIMAP SSTs. Model designations as in Table 1; jja = June, July, August; djf = December, January, February.

At about 21 ky calBP, the latitudinal distribution of insolation and its relative seasonal strength were similar to today. The ice sheet extent and height has been provided by Peltier (1994). CO_2 concentration was estimated at 200 ppm as inferred from Antarctic ice cores (Raynaud *et al.* 1993). Over the oceans, two sets of experiments were defined: (1) prescribing SST changes from estimates given by CLIMAP (CLIMAP 1981), and (2) computing SST using coupled atmosphere-mixed layer ocean models. Clearly, these experiments show only a limited range of climatic variations that are affected by selected changes. Other kinds of intrinsic variability (e.g., due to ocean circulation changes) or other kinds of natural forcing (e.g., solar irradiance and volcanic forcing) are not encompassed by these experiments.

Note that the boundary conditions used by the PMIP contain some uncertainties. For example, the location of ice sheets reconstructed by Grosswald (1999) has been extended over all parts of the Arctic shelf to cover large portions of the continents, but from the point of view of Vasil'chuk and Vasil'chuk (1995), the "glacial maximum" is expressed by an increase in mountain glaciation and spreading of permafrost conditions. Another example touches upon an underestimation of CLIMAP SST anomalies in the tropics (Anderson and Webb 1994; Guilderson *et al.* 1994; Hostetler and Mix 1999) and another upon the problem of location and seasonal behavior of sea-ice cover (Weinelt *et al.* 1996; Sarnthein *et al.* 2000).

An annual mean global cooling of about 4° C is obtained by all PMIP models forced by the CLIMAP SST estimates (Figure 1). This cooling is not uniform; it is stronger in the northern hemisphere, over and near the ice sheets, and stronger over land than over oceans due to snow/ice albedo feedback.

Global model results have been evaluated against terrestrial proxy data (Joussaume 1999; Kohfeld and Harrison 2000; Kislov *et al.* 2002). These results clearly demonstrate that temperatures modeled in the PMIP experiment reproduce the main peculiarities of reconstructed temperature fields, but over the tropics, the simulations with prescribed CLIMAP SSTs produce too weak a cooling effect over land. All models produce drying in the extratropical zone, although the extent and location of the regions of increased aridity vary between models. Comparing results of two sets of experiments with prescribed and computed SSTs indicates that the prescribed SSTs should yield a better model-data fit over land. This is an interesting fact, because there is no *a priori* reason to expect it to be so given the uncertainties in SST reconstructions.

Using PMIP data, we now calculate ensemble mean and standard deviation values for the CPR.

Winter mean cooling is obtained in the CPR by all PMIP models forced by the CLIMAP SST estimates (Figure 2A). Maximum cooling $|\Delta T| \geq 10°$ C occurs in the northern part of the CPR. Over other parts of the CPR, the anomaly $|\Delta T| \approx 5–7°$ C is obtained. The spatial distribution of the inter-model standard deviation (σ) is much more uniform and typically $|\Delta T|/\sigma \gg 1$ (Figure 3A). This finding means that the temperature patterns simulated by the different models have many principal details in common.

Simulated winter precipitation is generally low over the CPR (Figure 4A). In the central and western CPR, the models indicate drying, which can be

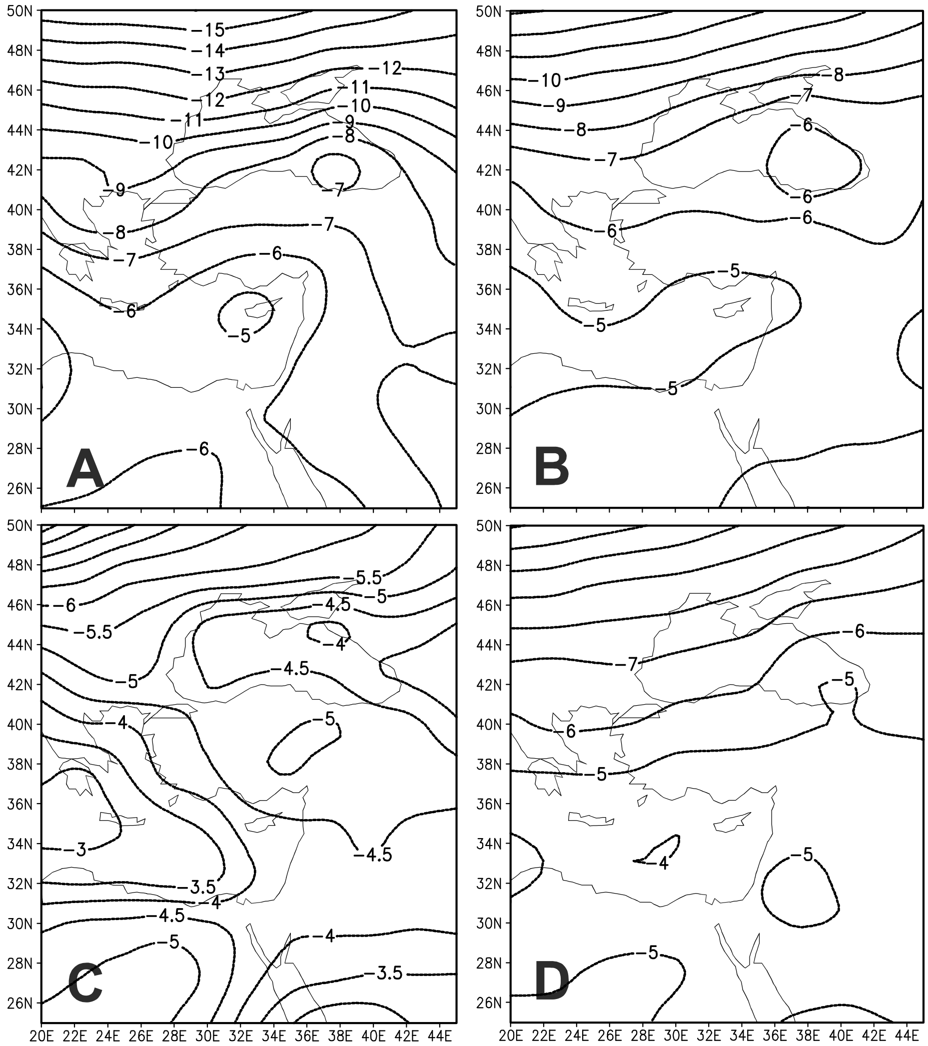

Figure 2. Winter (A), spring (B), summer (C) and autumn (D) mean cooling (° C) simulated for the LGM (21 ky calBP vs. modern control) from the ensemble of PMIP GCM simulations forced by CLIMAP SSTs.

characterized by $\delta = (P_{21} - P_0)/P_0 \approx -20\%$ (where P_{21} and P_0 denote the precipitation rates at 21 and 0 ky BP, respectively). The simulations show conditions wetter than present in Northern Africa ($\delta \approx 20\%$). These temperature and precipitation anomalies are the same in main aspects of their patterning for the months of October to April (Figure 2B, D and Figure 4B, D).

Maximum summer mean cooling $|\Delta T| \geq 7°$ C takes place in the northern part of the CPR (Figure 2C). Over other parts of the CPR, the anomaly

$|\Delta T| \approx 5°$ C is obtained. The models produce drying (Figure 4C) in the western part of the region ($\delta = -30\%$), with the most extensive changes toward the center ($\delta = -60\%$), and they show positive anomalies in the southeast (δ = 20–40%).

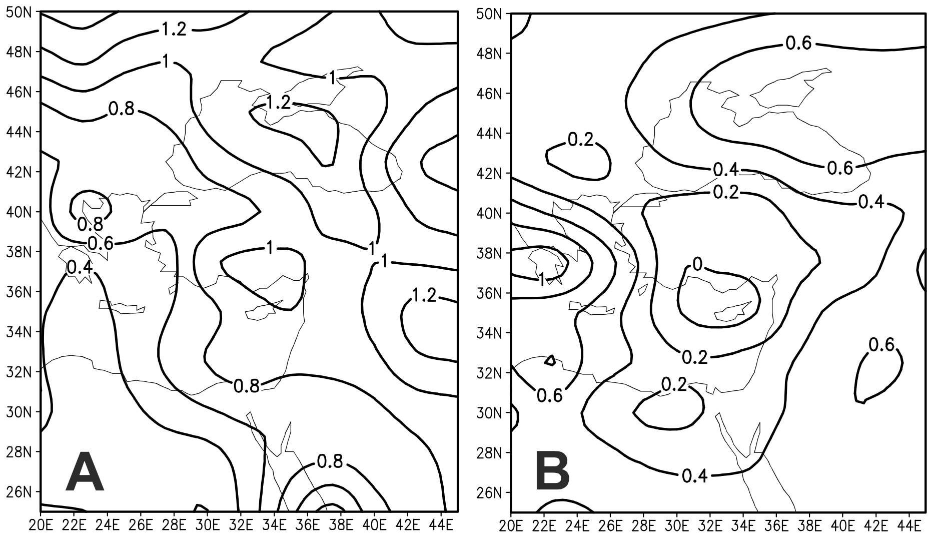

Figure 3. Winter (A) and summer (B) inter-model standard deviation (° C) for the LGM from the ensemble of PMIP GCM simulations forced by CLIMAP SSTs.

Regional climate differences have been widely interpreted in terms of prescribed changes through external causes and internal factors induced by the model. The positive greenhouse gas forcing pattern, in accordance with the assumption that such greenhouse gases were well-mixed, is geographically smooth. The negative ice-sheet forcing patterns are coupled to the locations of the North American and European glaciers. Temperature and precipitation response patterns are more complex, however. They are determined by the influence of feedback processes, which are related mainly to changes in the atmospheric circulation pattern.

During cold seasons of present climate, a low-pressure cell develops over the relatively warm North Atlantic Ocean (Icelandic Low). Surface westerly flow around its southern margin, amplified by the influence of the subtropical Azores High, brings frontal precipitation to Southern Europe and the Mediterranean. This precipitation is amplified by the strong temperature contrast between the warm Mediterranean Sea and the cold European land mass.

At 21 ky calBP, simulations of the winter circulation pattern show that deep cyclones developed over the North Atlantic (south of the present Icelandic Low). The presence of a substantial ice sheet over Scandinavia and the consequent air cooling led to the formation of a high-pressure cell over Northern

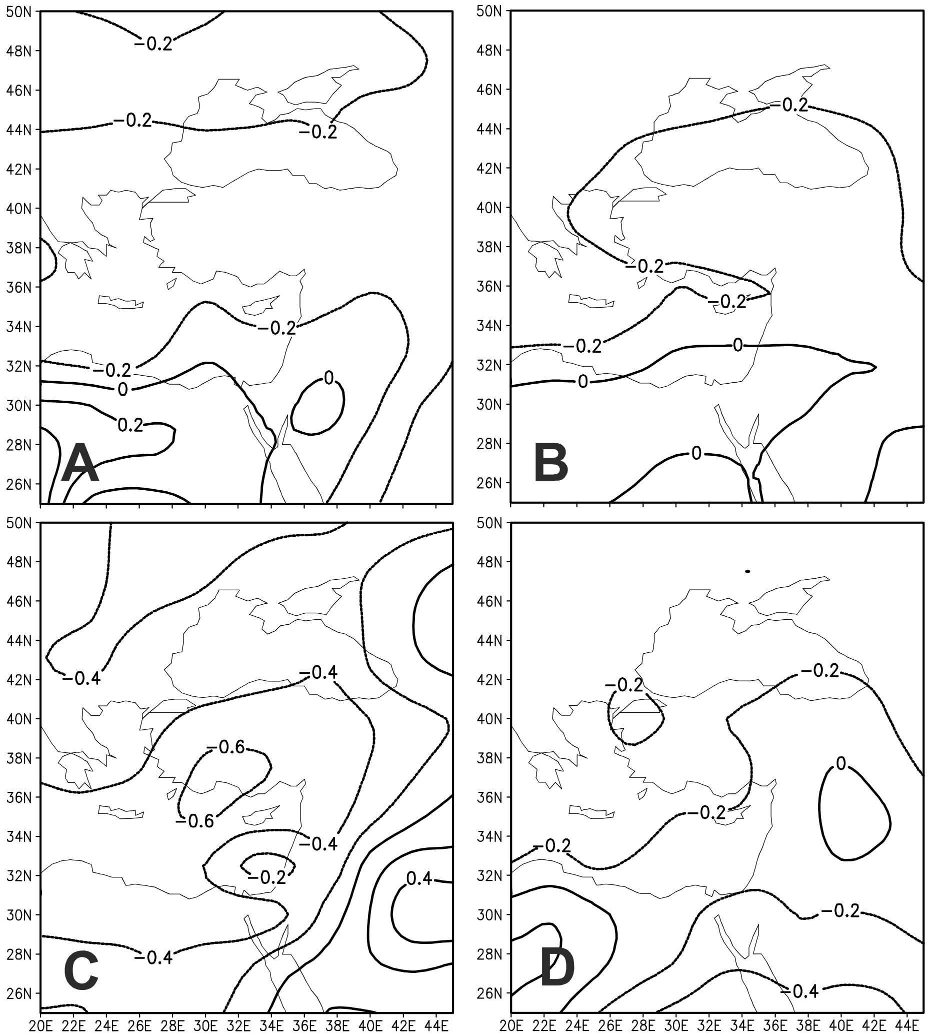

Figure 4. Winter (A), spring (B), summer (C) and autumn (D) mean precipitation changes ($\delta \equiv (P_{21} - P_0)/P_0$) simulated for the LGM (21 ky calBP vs. modern control) from the ensemble of PMIP GCM simulations forced by CLIMAP SSTs.

and Central Europe. Northerly or northeasterly flow on the eastern flank of the anticyclone would have brought cold, dry air masses to the south of the East European Plain from the continental interior and from the Arctic. Therefore, the most extensive temperature changes were located in the northern part of the CPR (10–14° C) as a result of this northeasterly flow around the glacial anticyclone. It made the CPR winters extremely cold and dry. We stress that, in the modern climate, the extremes of winter cooling observed in the southern part of the East European Plain are similarly associated with northeastern air mass invasions.

The influence of the glacial anticyclone did not penetrate very far to the south, and therefore, there are no strong anomalies over the central and southern parts of the CPR. Anomalies develop due to cooler and drier (compared to modern) air advection from the Atlantic Ocean into the Eastern Mediterranean.

The glacial anticyclone over Northern and Central Europe shifts the Mediterranean branch of the polar front to the south, increasing precipitation over North Africa (Figure 4A).

At 21 ky calBP, models show summer conditions in the CPR cooler and drier than today. Simulated summer temperature anomalies are not as large as those of the winter. This is determined by the fact that during the warm season, in response to the reduced equator-to-pole temperature gradient, the westerlies become weaker and are displaced northward. The southern part of Europe is occupied by the Azores High, whose ridge extends over the Mediterranean region. Under these conditions, local radiative and air-land and air-sea heat exchanges play an important role in temperature field formation together with the influence of cooler air advection.

3. THE MID-HOLOCENE WARM EVENT

The transition from glacial (the late Pleistocene cold event) to interglacial (the Holocene) was not gradual but was interrupted by several distinct cold events lasting several hundred years. The Holocene is marked by a relatively stable climate experiencing weak changes on a global average. By the Early Holocene (~9 ky BP), CO_2 was up to its preindustrial level, the Scandinavian ice sheet was almost gone, and the Laurentide ice sheet had shrunk considerably (Dyke and Prest 1987; Svensson 1991). Sea surface temperature was not significantly different from today. The Northern Hemisphere experienced a strong seasonal contrast in insolation, which was near its maximum at this time. In the mid-Holocene, at 6 ky calBP (5.3 ky BP), insolation anomalies were +5% in summer and –5% in winter, less than at the beginning of the Holocene when climatic response was influenced by remnant ice sheets of the last glaciation.

Recent studies within the PMIP have focused on the 6 ky calBP climate. As a first approximation, SSTs were prescribed to be same as today, the CO_2 concentration was similar to its preindustrial value of 280 ppm (Raynaud *et al.* 1993), and vegetation and land-surface characteristics were held constant. PMIP simulations of the 6 ky calBP climate were thus designed as pure sensitivity experiments to changes of insolation forcing.

All PMIP models simulate an increased seasonal cycle of temperature over the continents in the Northern Hemisphere, reaching about ±1° C on global

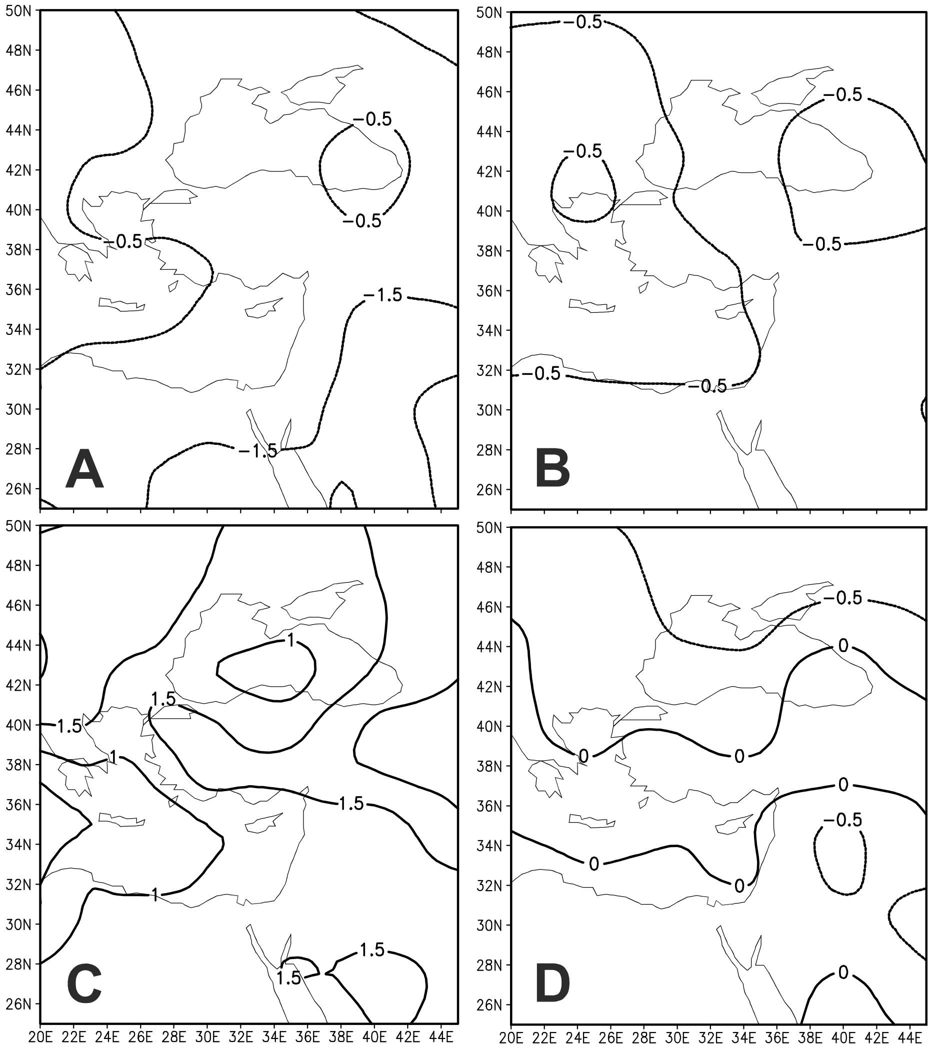

Figure 5. Winter (A), spring (B), summer (C), and autumn (D) mean temperature changes (° C) simulated for the mid-Holocene warm event (6 ky calBP vs. modern control) from the ensemble of PMIP GCM simulations forced by insolation changes.

average for summer/winter seasons. In summer, all models demonstrate a temperature increase over land in the temperate zone. All models simulate an increase of African and Asian monsoon rains in response to additional summer warming of the land (Joussaume *et al.* 1999).

Global model results have been evaluated (Joussaume 1999; Kohfeld and Harrison 2000) against terrestrial proxy data (lake levels, pollen- and plant macrofossil-based reconstructions of 6 ky calBP vegetation). Comparing these paleoenvironmental estimates of temperature and precipitation changes at 6 ky calBP with modeling data indicates their close agreement but shows that modeled precipitation is underestimated in several arid and semi-arid regions. Model-data discrepancies may result in part from the omission of vegetation and ocean temperature feedback on climate (Joussaume 1999; Otto-Bliesner 1999; Braconnot *et al.* 2000).

CPR winter mean cooling (Figure 5A) is obtained by all models forced by the orbitally-induced negative anomalies of insolation, however, the simulated changes are very small. The patterns of spatial distribution of spring and autumn mean temperature (Figure 5B,D) are essentially similar to today. The models show winter conditions drier than today ($\delta \approx -10\%$) in the northern and eastern part of the region, whereas conditions wetter than today ($\delta \approx 40\%$) form in the north of the Arabian Peninsula (Figure 6). This is very small from the point of view of absolute values, but taking into account the weather peculiarities of the area, it indicates an increased frequency of rainstorms.

At 6 ky calBP, the models show insolation-induced summer conditions in the CPR (Figure 5C) that were warmer than today ($\Delta T \approx 1.5°$ C, and again $|\Delta T|/\sigma \gg 1$). Maximum summer mean warming $\Delta T \geq 2°$ C occurred near the Caucasus Mountains. PMIP simulation of the 6 ky calBP summer shows conditions drier than today over practically the whole CPR except in the southeast (Figure 6C). Extensive precipitation increases over the Arabian Peninsula ($\delta \approx 180\%$) were due to a strengthening South Asian monsoon (Joussaume *et al.* 1999), which allowed rainfall to penetrate into the Eastern Mediterranean.

4. CONCLUSION

Simulated mid-Holocene climate differs radically from that simulated for the late Pleistocene, and it differs in more subtle ways from present climate. Climate maps presented in this paper demonstrate that the signs of temperature change over the CPR and the Northern Hemisphere are similar. Precipitation response is a more complex issue. Wetter (or drier) anomalies correlate with both increasing and decreasing temperature. These results characterize the typical range of climate variations affected by natural external forcing.

Recent decades have witnessed many environmental changes, among which the most appreciable have been an increase in near surface global temperature together with a decrease in air temperature within the lower stratosphere, a reduction in the area of glaciers, and a rise in sea level (IPCC 2001). The main candidate for forcing agent in current and near future climate variation is the mounting level of atmospheric CO_2 (IPCC 2001). Variations in ozone,

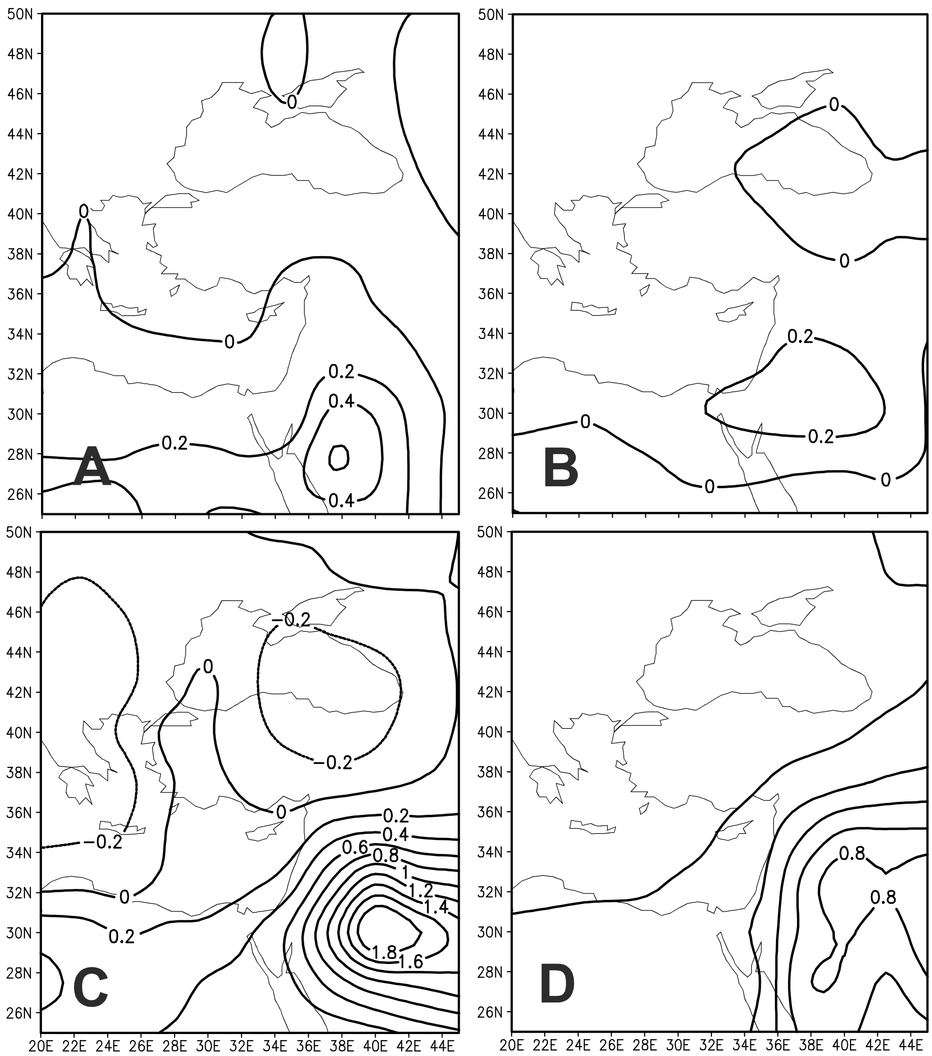

Figure 6. Winter (A), spring (B), summer (C) and autumn (D) mean precipitation changes ($\delta \equiv (P_{21} - P_0)/P_0$) simulated for the mid-Holocene warm event (6 ky calBP vs. modern control) from the ensemble of PMIP GCM simulations forced by insolation changes.

solar constant, and optical depth of the stratosphere due to large volcanic eruptions, together with the anthropogenic flux of SO_2, are considered among the additional driving processes of modern climate variability.

The method of assessing equilibrium response considered above is not realistic for evaluating the effects of increasing CO_2, changing insolation, etc., on climate because the gases will increase gradually, while the ocean and sea ice will not respond immediately. Recently, several scientific teams have modeled time-dependent future climate changes based on the most frequently used A2 and B2 scenarios of future economic and environmental changes (IPCC 2001). Experiments with the CMIP models (Covey *et al.* 2003) demonstrate that there is statistical confidence for warming in practically every site of the globe, but it is not uniform. Warming tends to be stronger in the high latitudes, and it is stronger over land than over oceans due to snow/ice albedo feedback. Statistical confidence for precipitation changes is found in only some places. One such place is the Mediterranean basin, where precipitation is projected to decrease.

Räisänen *et al.* (2004) have explored the transient dynamic of climate based on a series of regional climate change simulations focusing on Europe. Regional numerical experiments were driven by different global models. Comparison between two slices of time, 1961–1990 and 2071–2100, indicates that in all the scenario runs, warming in the CPR peaks during the 2071–2100 summers when average temperature reaches ~6° C higher than the 1961–1990 summers. In winter it is ~4° C higher. All simulations agree on a general decrease in precipitation in the CPR, which declines ~40–50% in summer. In winter, the simulated changes of precipitation are not so large. These changes are very important because they are projected to exceed those experienced during the mid-Holocene warm event.

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ENDNOTES

1. The first phase of the PMIP terminated in 2002. For detailed information on its second phase, see: http://www-lsce.cea.fr/pmip2.

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4

STATUS OF THE BLACK SEA FLOOD HYPOTHESIS

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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.

V. Yanko-Hombach et al. (eds.), The Black Sea Flood Question, 63-88.

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 fresh-water 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 (Fedorov 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 Chepalyga (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 through 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.

Khrischev 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 Maynard (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 huxleyi*) 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,

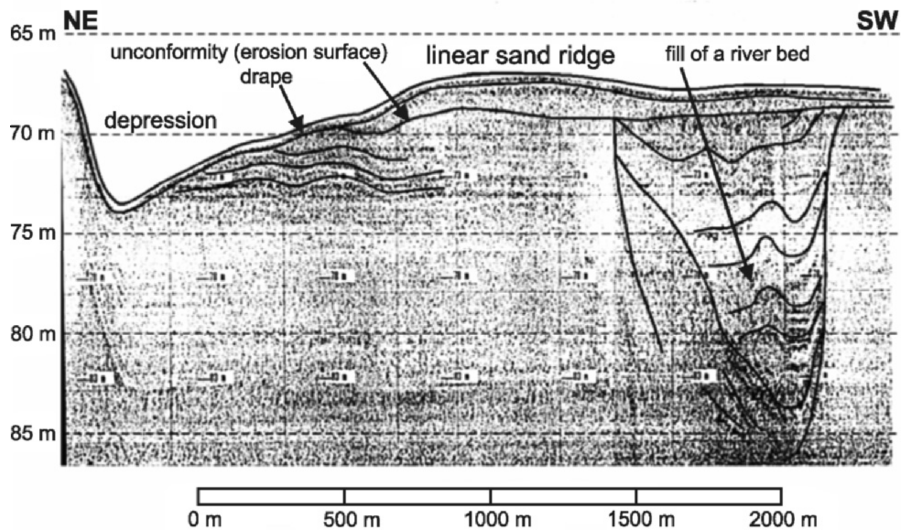

Figure 1. Reflection profile over a linear sand ridge showing its location above the unconformity and filled river channels below the unconformity. The profile was obtained on the BLASON expedition with the *R/V Suroît* using equipment provided by N. Panin.

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).

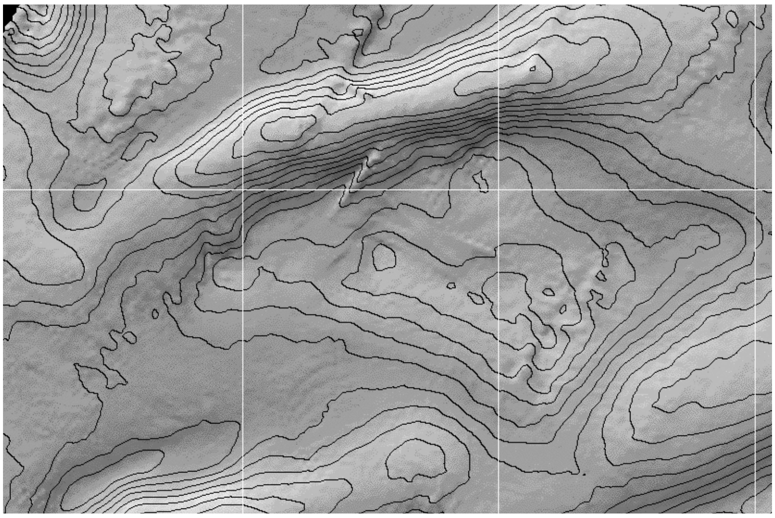

Figure 2. Shaded-relief map of linear sand ridges and depressions near 44° 04' N and 29° 59' E within the dune field on the Romanian shelf. The multibeam data were provided from IFREMER by G. Lericolais and processed by William F. Haxby. Distance between vertical lines is 1500 m, and the contour interval 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 (Khrischev 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; Khrischev and Shopov 1978; Shopov *et al.*1986; Khrischev 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 (hydrotroilite 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 (Khrischev 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 hydrotroilite 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 hydrotroilite 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 hydrotroilite 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 hydrotroilite 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.

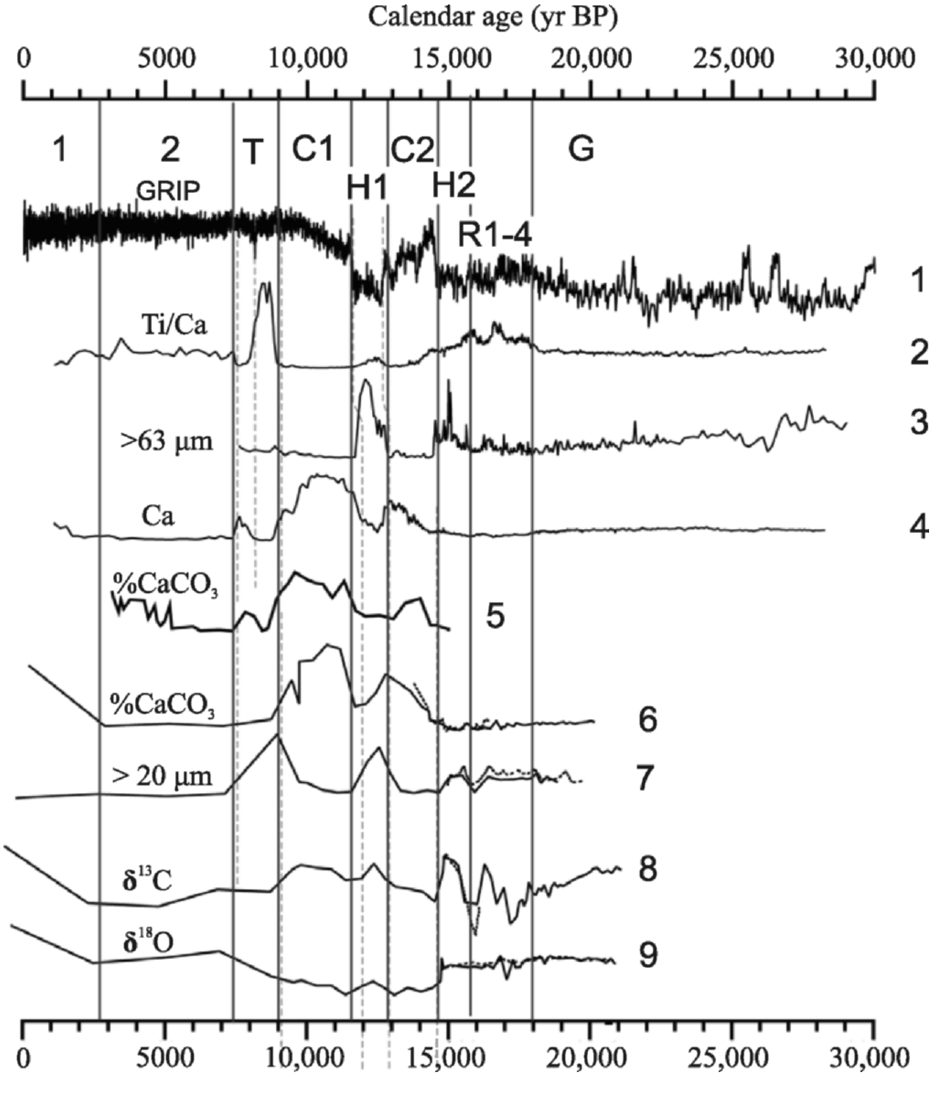

Figure 3. Correlation of Black Sea lithostratigraphic intervals with the GRIP chronology using the $\delta^{18}O$ signal in Greenland ice cores (Curve 1). Curves 2–4 are from core GeoB 7608-1 of Bahr *et al.* (2005). Curve 5 is from core A96 of Khrischev and Georgiev (1991). Curves 6–9 are from cores 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.

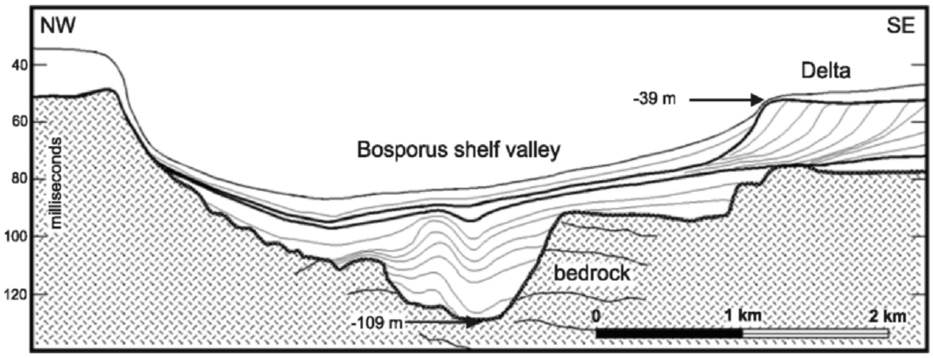

Figure 4. Reflection profile across the Bosporus shelf valley in the Sea of Marmara, south of the Bosporus Strait (from Ryan *et al.* 1997b). The prograding clinoforms of this subaqueous delta of the Kurbağalıdere River are shown on the southeastern side of the profile. The placement of the delta (outlined 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?

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5

THE MARMARA SEA GATEWAY SINCE ~16 KY BP: NON-CATASTROPHIC CAUSES OF PALEOCEANOGRAPHIC EVENTS IN THE BLACK SEA AT 8.4 AND 7.15 KY BP

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Abstract: The Late Quaternary history of connection of the Black Sea to the Eastern Mediterranean has been intensely debated. Ryan, Pitman and coworkers advocate two pulses of outflow from the Black Sea to the world ocean at ~16–14.7 ky BP and ~11–10 ky BP. From ~14.7–11 ky BP and from ~10–8.4 ky BP, they suggest that the level of the Black Sea fell to ~ –100 m. At 8.4 ky BP, they further claim that a catastrophic flood occurred in a geological instant, refilling the Black Sea with saline waters from the Mediterranean. In contrast, we continue to gather evidence from seismic profiles and dated cores in the Marmara Sea which demonstrate conclusively that the proposed flood did not occur. Instead, the Black Sea has been at or above the Bosphorus sill depth and flowing into the world ocean unabated since ~10.5 ky BP. This conclusion is based on continuous Holocene water-column stratification (leading to sapropel deposition in the Marmara Sea and the Aegean Sea), proxy indicators of sea-surface salinity, and migration of endemic species across the Bosphorus in both directions whenever appropriate hydrographic conditions existed in the strait. The two pulses of outflow documented by Ryan, Pitman and coworkers find support in our data, and we have modified

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our earlier interpretations so that these pulses now coincide with the development of mid-shelf deltas: Δ2 (16–14.7 ky BP) and Δ1 (10.5–9 ky BP) at the southern end of the Bosphorus Strait. However, continued Black Sea outflow after 9 ky BP prevented the northward advection of Mediterranean water and the entry of open-marine species into the Black Sea for more than 1000 years. Sufficient Mediterranean water to change the Sr-isotopic composition of slope and shelf water masses was not available until ~8.4 ky BP (along with the first arrival of many varieties of marine fauna and flora), whereas euryhaline molluscs did not successfully populate the Black Sea shelves until ~7.15 ky BP. Instead of relying on catastrophic events, we recognize a slow, progressive reconnection of the Black Sea to the world ocean, accompanied by significant time lags.

Keywords: Marmara Sea Gateway, Bosphorus Strait, Black Sea Flood Hypothesis, Outflow Hypothesis, climate change

1. INTRODUCTION

The “Marmara Sea Gateway” connects the Black Sea and Eastern Mediterranean (Figure 1A, B). The gateway consists of (1) a linked set of narrow straits with shallow bedrock sills (the Bosphorus Strait with a sill depth of ~ –40 m; and the Dardanelles Strait with a sill depth of ~ –70 m) and (2) the inland Marmara Sea. The Marmara Sea fills a rugged, tectonically active depression comprising three abyssal basins reaching depths of >1200 m and separated by cross-basin ridges (Aksu *et al.* 2000).

The Marmara Sea Gateway provides an unparalleled natural laboratory in which to study the evolution of Quaternary climate in central and northern Europe. This is because the narrow straits regulate all communication between the Black Sea and the world ocean through the small Marmara Sea, which acts as a sediment trap, registering, like a sensitive tape recorder, paleoceanographic events and paleoenvironmental changes in the catchment area of the rivers that drain into the Black Sea. The volume of the Marmara Sea is only 0.65% of the volume of the Black Sea, so there is enormous amplification in the Marmara sediments of the effects of Quaternary water exchange between the Aegean and Black Seas. The geometry of this gateway is analogous to a vacuum line in a chemistry laboratory, with its series of control valves and condensation traps.

Today, the Black Sea is swollen by the discharge of major European rivers (the Danube, Don, Dnieper, Dniester, Southern Bug), resulting in the net export of ~300 km^3/yr of water through the gateway. This volume represents fifty times the cumulative annual discharge of the small rivers entering the Marmara Sea. Satellite altimetry shows that the surface of the Black Sea is ~30 cm above the level of the Marmara Sea, which, in turn, varies between 5 and 27

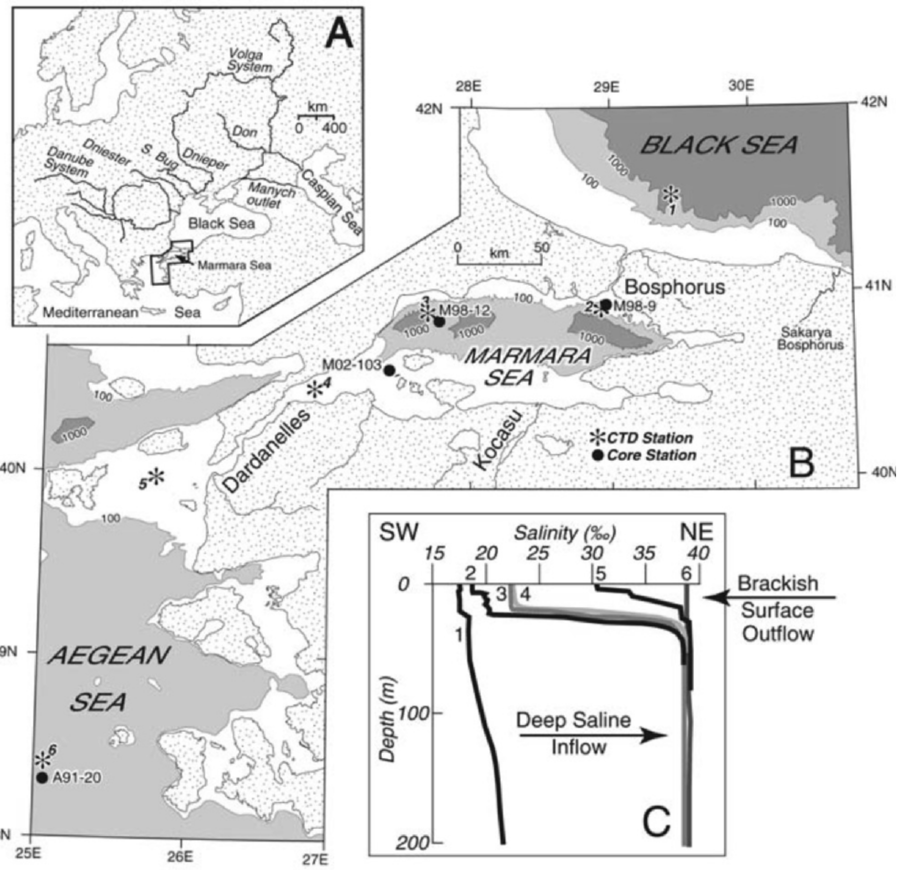

Figure 1. **A:** Location of the Marmara Sea and major European rivers entering the Black Sea. During deglaciation, the Volga system was connected to the Black Sea through the Manych outlet and a link to the Don River. **B:** Simplified bathymetry in meters (Aksu *et al.* 1999); location of cores M98-9, M98-12, M02-103 and A91-20; and location of conductivity-temperature-depth (CTD) stations 1–6, corresponding to data plotted in part C. The Sakarya Bosphorus no longer connects to the eastern end of the Marmara Sea because of Quaternary offsets along the North Anatolian Transform Fault. **C:** Salinity *versus* water depth in the upper 200 m (stations 1, 3, 6) or to the seabed (stations 2, 4, 5), showing sharp salinity-controlled pycnocline at ~20–25 m except in the Black Sea (subtle pycnocline at ~130 m) and the southern Aegean Sea (no low-salinity layer). The low-salinity surface lid originates from Black Sea outflow and promotes permanent stratification which in turn promotes sub-pycnocline oxygen depletion. Salinity (CTD) data from archives of Institute of Marine Sciences and Technology (IMST), Dokuz Eylül University, Izmir.

cm above the level of the northern Aegean Sea (Polat and Tuğrul 1996). These elevation differences drive the outflow across the gateway.

The present water exchange across the Bosphorus Strait is a two-layer flow. A cooler, lower salinity (17–20‰) surface layer exits the Black Sea, while warmer, higher salinity (38–39‰) Mediterranean water flows northward through

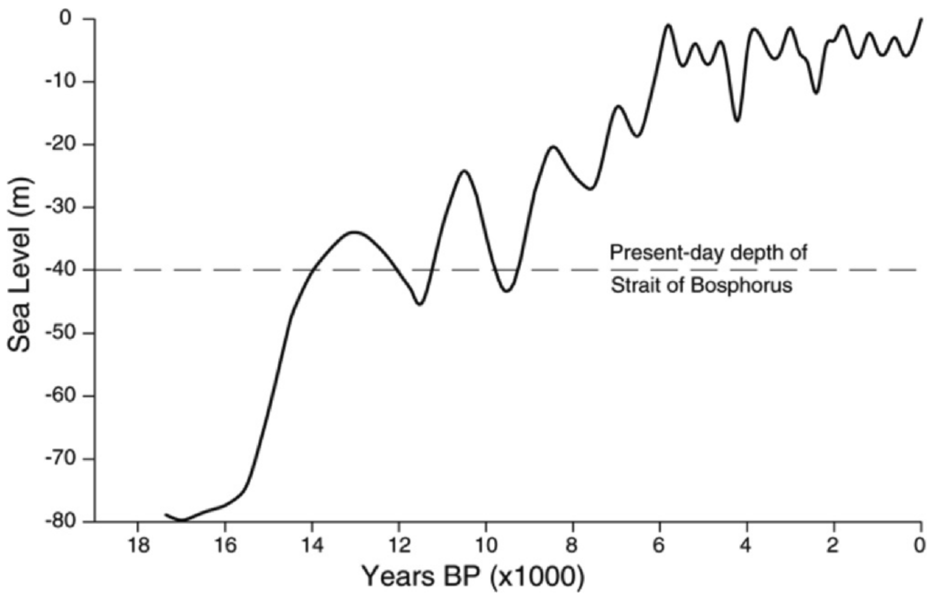

Figure 2. Record of Black Sea water levels compiled from radiocarbon dates on shells collected from former shorelines around the Black Sea coast and inner shelf (after Chepalyga 2002).

the strait at depth (Polat and Tuğrul 1996). The upper tens of meters of the water column in the gateway area (but not in the open Aegean Sea) are strikingly fresher than the deep water because of the Black Sea outflow (Figure 1C). The low-density surface layer prevents ventilation of the deeper water column, promotes organic-matter preservation, and fosters benthic communities adapted to suboxic to dysoxic conditions with 0.1–1.3 ml/l dissolved oxygen.

Throughout the Quaternary, the Black Sea experienced a complex series of transgressions and regressions (Figure 2) (Chepalyga 2002). Transgressions resulted from a combination of melting ice sheets and permafrost in Central Asia (east of the Ural Mountains), wet periods with increased rainfall, cooler time intervals with reduced evaporation across eastern Europe and west-central Asia, and periodic discharges of water from the Caspian Sea through the Manych Depression (e.g., the 16–14 ky BP late Khvalynian flood peak of Chepalyga, this volume). Whenever the sill in the Bosphorus Strait (or Sakarya Bosphorus as in Figure 1B) was subaerially exposed because of low global sea level (glacial stages 6, 4–2; see Yaltırak *et al.* 2002), the Black Sea oscillated independently, whereas at other times like today (interglacial stages), the level of the Black Sea was effectively pinned to the level of the global ocean because of free exchange across the Marmara Sea Gateway. Throughout its Quaternary history, the salinity of the Black Sea has varied from marine to semi-fresh. Salinity estimates used here follow the terminology of Chepalyga (1984): marine (30–40‰ salinity), semi-marine (12–30‰), brackish (5–12‰), semi-fresh (0.5–5‰), and fresh (<0.5‰).

2. RIVAL HYPOTHESES

The means by which the Black and Mediterranean Seas were reconnected after the last glaciation is intensely debated. Ryan *et al.* (1997) and Ryan and Pitman (1998), modified by Ryan *et al.* (2003), proposed their *Flood Hypothesis*, which entailed a catastrophic refilling of the Black Sea basin by marine water at ~8.4 ky BP, an event they estimate to have taken place in less than two years. They link this controversial deluge with the biblical account of Noah's Flood and explain a low Black Sea (Neoeuxinian Lake) persisting well into the Holocene by advocating a dry central European climate. Before this flood, Ryan and coworkers argue for the scenario diagrammed in Figure 3A:

(1) a ~16–14.7 ky BP meltwater-induced outflow from the Black Sea through the Bosphorus channel (= late Khvalynian flood peak of Chepalyga, this volume),

(2) an evaporative drawdown of the Black Sea to an elevation of ~ –105 m from 14.7–12 ky BP,

(3) a ~11.5–11 ky BP Black Sea transgression to –25 to –30 m that caused a second outflow into the Marmara Sea from ~11–10 ky BP, and

(4) a 10–8.4 ky BP evaporative drawdown of the Black Sea until it reached ~ –95 m.

Because the Mediterranean and Marmara water levels had by ~8.4 ky BP already risen to ~ –25 m, well above the present sill depth of the Bosphorus Strait, Ryan *et al.* (1997) hypothesized the existence of a sediment dam to hold back the global ocean. They concluded that the Marmara and Mediterranean catastrophically flooded into the depressed Black Sea basin when this hypothetical sediment dam in the Bosphorus channel was scoured away.

Initially, Ryan *et al.* (1997) relied on radiocarbon dates of the first euryhaline molluscs to populate the drowned shelves of the northern Black Sea as a marker for the time of flooding; these dates clustered at ~7.15 ky BP. Euryhaline molluscs require semi-marine to marine salinity of ~20–40‰ (Knox 1986), conditions only marginally attained in the Black Sea today (Figure 1C). Subsequently, Major (2002) and Ryan *et al.* (2003) shifted the date for flooding to ~8.4 ky BP based mainly on Sr-isotopic data, and they reinterpreted the ~7.15 ky BP shell ages as having resulted from an approximately 1300-year salinization time lag, after which the Black Sea shelf waters attained a salinity level appropriate for colonization by euryhaline fauna well after the actual reconnection took place.

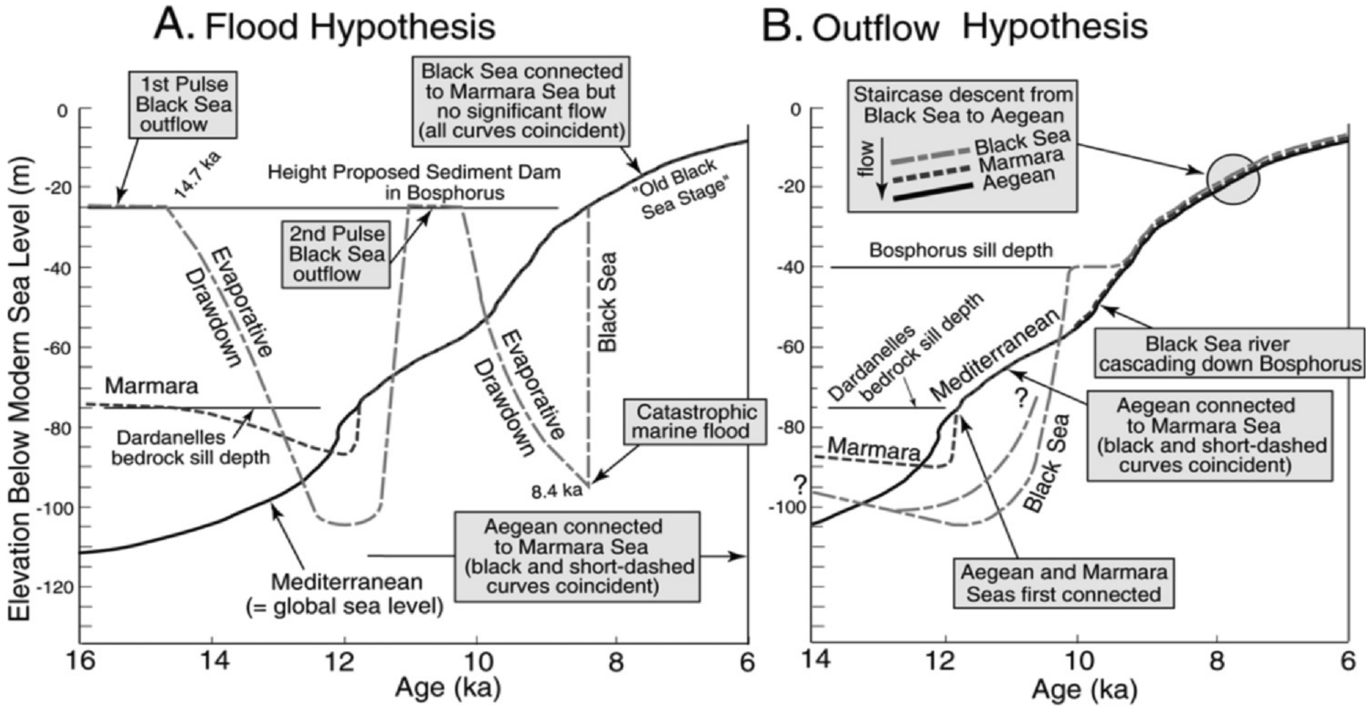

Figure 3. Schematic water-level histories of the Black, Marmara, and Mediterranean (Aegean) Seas according to A. the *Flood Hypothesis* of Ryan *et al.* (2003) and B. the original *Outflow Hypothesis* of Aksu *et al.* (2002a) and Hiscott *et al.* (2002). In both plots, the Mediterranean curve (solid dark line) is the Barbados (global) curve of Fairbanks (1989). When Mediterranean and Marmara curves (dotted dark line) are superimposed (e.g., from ~12–10 ky BP in both plots), the Marmara Sea was an embayment of the Mediterranean.

The flood hypothesis was initially challenged by several authors. Shifting the flood date from ~7.15 to ~8.4 ky BP satisfied the objections of Görür *et al.* (2001), but an extensive Russian literature on regional water-level variations still concludes that the Black Sea reached a level of ~ –30 m by ~10–9 ky BP and has never been lower since (Figure 2). In the Marmara and Aegean Seas, sensitive proxy indicators in sediment cores have been interpreted by Aksu *et al.* (1995, 2002a, b, c), Çağatay *et al.* (2000), Kaminski *et al.* (2002), and Abrajano *et al.* (2002) as evidence that the Black Sea has been continuously exporting low-salinity water through the Marmara Sea Gateway since ~10 ky BP, or perhaps earlier. This is the basis for our *Outflow Hypothesis* (Figure 3B). Physical sedimentological evidence from bedform asymmetry and deltaic progradation at the southern end of the Bosphorus Strait supports this interpretation (Aksu *et al.* 1999; Aksu *et al.* 2002a; Hiscott *et al.* 2002). These last authors named the youngest delta (~10–9 ky BP) at the strait exit "delta 1," or Δ1.

We began to study the gateway area in 1995 and have acquired approximately 8500 line-km of airgun, sparker, and Huntec boomer profiles (vertical resolution ~10–20 cm), about 126 gravity and piston cores, and 78 radiocarbon dates from cores. A case has been built for continuous Black Sea outflow since ~10 ky BP that is rooted mainly in our large dataset from the Marmara and Aegean Seas, and less so from the Black Sea itself. Ryan *et al.* (2003) criticize our efforts to deduce the behavior of the Black Sea largely from adjacent waterways, but the unique configuration of the Marmara Sea Gateway perhaps makes it a better place to monitor the reconnection history than within the Black Sea itself. As an analogy, if one wants to know the number of spectators in a stadium, a count at the exit as people enter or leave is more reliable than an attempt to estimate the number of spectators from within the stadium. Our evidence from the connecting link between the Aegean and Black Seas leaves little doubt that the last phase of Black Sea outflow began as early as ~11–10 ky BP and has continued to the present.

Ryan *et al.* (1997) and Ryan and Pitman (1998) instead proposed that the Black Sea was in a protracted phase of evaporative drawndown from ~14.5–7.15 ky BP, ending with a catastrophic flood when the rising Mediterranean and Marmara Seas breached the sediment dam in the Bosphorus Strait. In their present version of events, Major (2002) and Ryan *et al.* (2003) acknowledge outflow at ~11–10 ky BP but insist that it then stopped, first because evaporation lowered the level of the Black Sea and terminated the earlier connection during the interval 10–8.4 ky BP, and subsequently because the Black Sea remained sufficiently evaporative that it continued to receive net inflow from the Aegean Sea through the Bosphorus Strait. They suggest that this latter situation continued until ~3 ky BP, when the Black Sea's tributary rivers eventually attained near-modern discharges and the climate shifted to modern humidity and rainfall

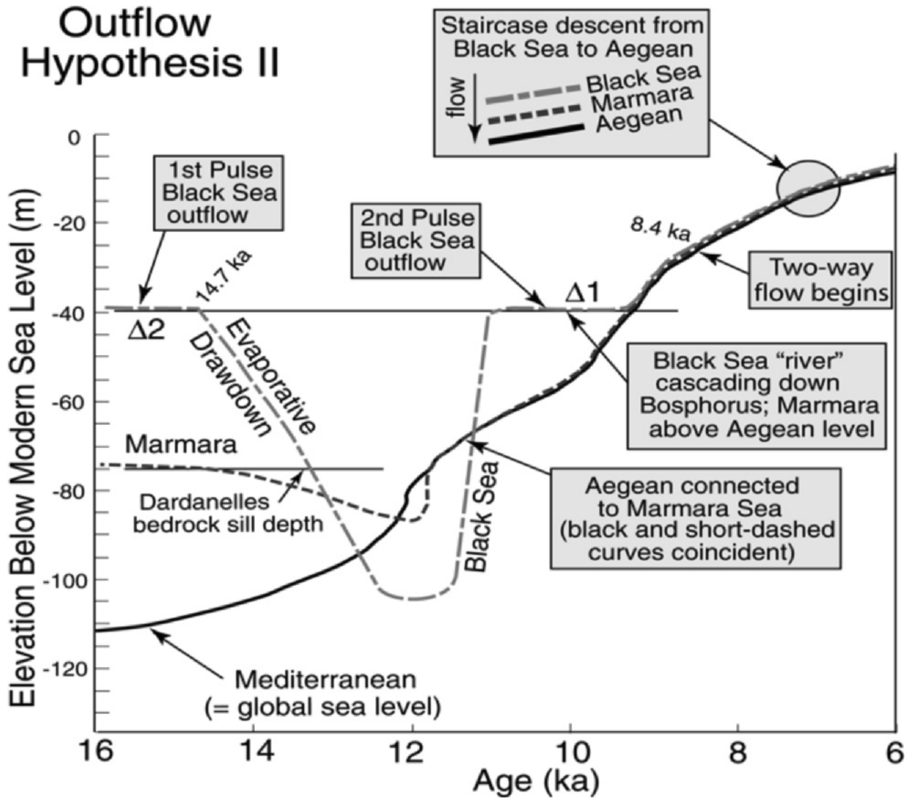

Figure 4. Modified *Outflow Hypothesis*, designated as version II. An ~16–14.7 ky BP episode of Black Sea outflow is incorporated, and the second pulse of Black Sea outflow is initiated somewhat earlier than in Figure 3B.

levels. Only since ~3 ky BP (not shown in the plot of Figure 3A), in what is known as the New Black Sea stage, do Ryan *et al.* (2003) believe that water exchange between the Black Sea and the Mediterranean has been like that of the present, with significant Black Sea outflow across the gateway.

In the Marmara Sea, the *Flood Hypothesis* predicts thorough mixing of the water column between ~10 and 8.4 ky BP, with a final turbulent stirring and flushing in of Mediterranean fauna and flora over a short period at 8.4 ky BP. According to the original *Outflow Hypothesis*, the Black Sea reached the –40 m bedrock sill depth in the Bosphorus Strait first, initiating a cascade downslope into the rising Marmara Sea from ~10–9 ky BP and building delta Δ1 (Figure 5). This hypothesis does not involve a catastrophic flood but instead predicts stratification and low oxygen conditions in the Marmara Sea since ~10 ky BP, similar to today (Figure 1C).

Is there any common ground between the *Flood Hypothesis* and our

Outflow Hypothesis? Yes, but only for the period before 10 ky BP. For example, we see considerable merit in the proposal of Ryan *et al.* (2003) that there were two outflow events at ~16–14.7 and ~11–10 ky BP, and we have incorporated these into a modified *Outflow Hypothesis* (version II, Figure 4). This modification does not violate any of the data presented by Aksu *et al.* (2002b) and Hiscott *et al.* (2002) and provides an explanation for the double unconformity present locally on the southwestern Black Sea shelf (α and α1 of Aksu *et al.* 2002b). Also, the ~16–14.7 ky BP outflow event provides an attractive explanation for the older delta 2 (Δ2) described by Hiscott *et al.* (2002) at the southern exit of the Bosphorus (Figure 5). With new cores, we have now established that the widespread mud drape on top of Δ2 and below the β3 unconformity was forming at 10,950 yr BP based upon radiocarbon dating from Core M02-111 (Table 1, Figures 6 and 7B). A comprehensive reanalysis of the development of Δ2 and the younger Δ1, and the constraints they place on Black Sea outflow, is presented in a later section.

In this paper, we marshal previously published and new evidence from the Marmara and Aegean Seas to evaluate the history of connection between the Black Sea and the open ocean. The critical debating points between Ryan and coworkers and our research team can be reduced to two fundamental issues:

(1) Is there any evidence in the Marmara Sea Gateway that the strong Black Sea outflow that began at ~11–10 ky BP was reduced to nothing (~10–8.4 ky BP) and then very little (8.4–3 ky BP) as Ryan *et al.* (2003) have proposed?

(2) Is there credible evidence that early Holocene climate throughout the region was sufficiently arid to promote net evaporation from the Black Sea surface from ~10–3 ky BP?

We first review the basis for the agreed early strong outflow (before ~10 ky BP). Then, we evaluate whether environmental and paleoceanographic proxy data from younger deposits are compatible with a continuation of this outflow to the present day, or instead suggest a predominant Mediterranean influence across the Marmara Sea Gateway from ~8.4–3 ky BP with significant net Black Sea outflow resuming only after that time.

2.1 The ~16–9 ky BP History of Connection and Black Sea Outflow

Until ~12 ky BP, global sea level was lower than the spill depth of the Marmara Sea at the Dardanelles Strait (e.g., Fairbanks 1989), effectively limiting water level in the Marmara Sea to a maximum height of ~ –70 m. Ryan *et al.* (2003) describe highstand deposits from the Black Sea dating to ~16–14.7 ky BP that indicate overspill into the Marmara Sea at that time. This influx of water into the small Marmara basin (with <1% of the volume of the Black Sea) would have

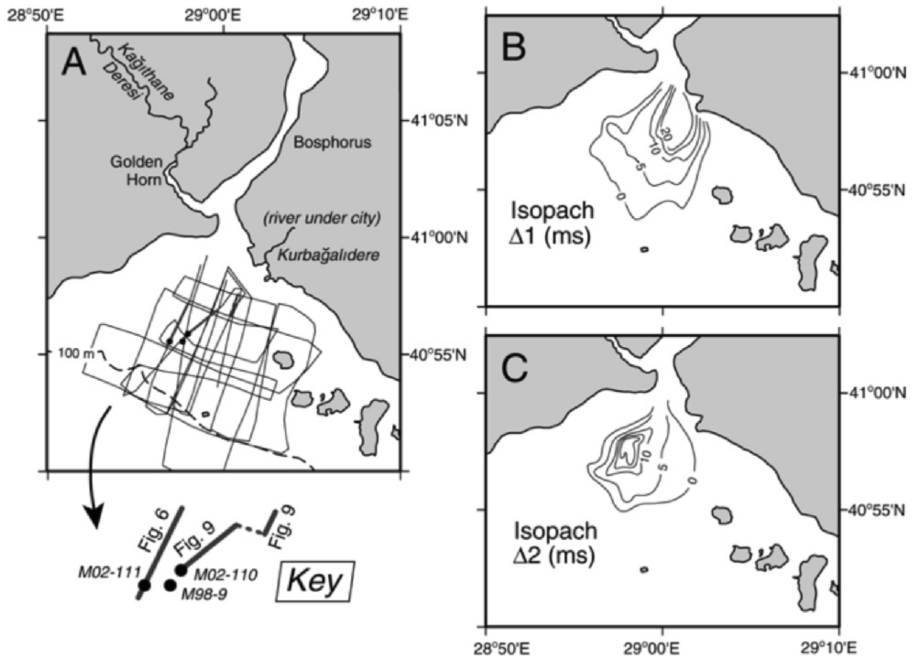

Figure 5. Maps of the area south of the Bosphorus Strait.
A. Boomer survey tracks (~280 line-km), core locations, and key to the locations of seismic profiles shown in other figures (see enlarged labels for cores and seismic profiles below the map). The Kağıthane Deresi and Kurbağalı Dere are small rivers described in the text and Table 1.
B. and **C.** Isopach maps of strait-mouth deltas, the younger Δ1 and the older Δ2, with sediment thicknesses in milliseconds of two-way travel time (10 ms ≈ 7.5 m).

maintained its water level at the Dardanelles spill depth until ~14.7 ky BP, creating a staircase descent from the Black Sea at ~ –40 m (Figure 4) or ~ –25 m (Figure 3A), to the Marmara Sea at ~ –70 m, to the gradually rising Aegean Sea.

The isolated mid-shelf delta (Δ2 of Hiscott *et al.* 2002) at the southern end of the Bosphorus Strait is here reinterpreted to have formed during the ~16–14.7 ky BP outflow event (Figures 5C and 6). The youngest topset-to-foreset transition of this delta is at a modern elevation of –69 m. Ideally, the topset elevation should be several meters below the contemporaneous sea level. Given potential uncertainty in the estimate of the depth of the Dardanelles sill, and possible uplift of Δ2 because of proximity to the North Anatolian Transform Fault (C. Yaltırak, personal communication 2004), the agreement between the topset-to-foreset elevation (–69 m) and the spill depth (~ –70 m) is remarkably good. When the first phase of Black Sea outflow ceased at ~14.7 ky BP (Ryan *et al.* 2003), Δ2 was abandoned and a marine drape filled depressions adjacent to the delta lobe (Figure 6 between β4 and β3). At this time, the Marmara Sea

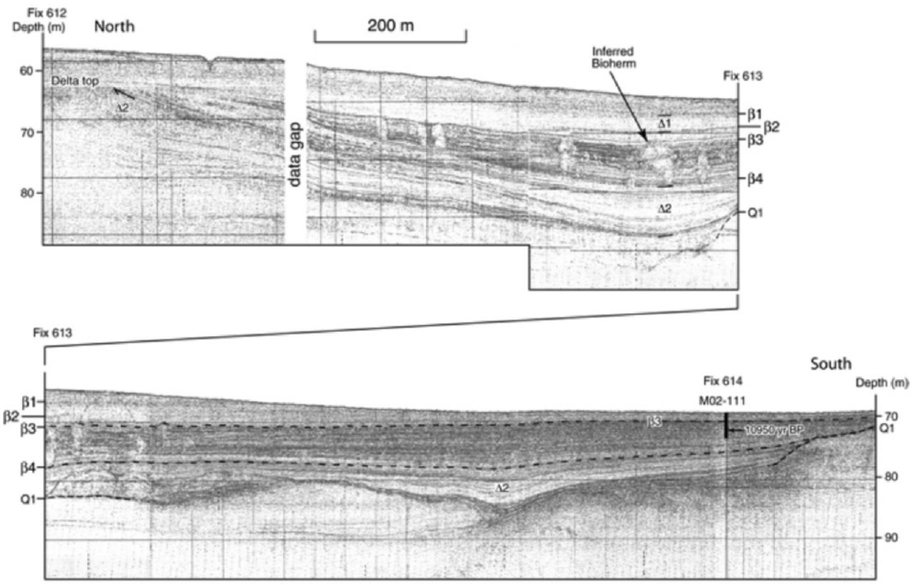

Figure 6. Huntec deep-tow boomer profile (location in Figure 5A) across Δ2, showing reflections β1–β4 and Q1 defined by Hiscott *et al.* (2002). Strata equivalent to Δ2 are confined between reflections β4 and Q1, whereas the younger Δ1 is represented in this profile only by its prodelta deposits between reflections β1 and β2. The mounds along the β3 surface are interpreted as algal-serpulid bioherms. Core M02-111 penetrates well below β3; facies, radiocarbon dates, and picks of reflectors are shown in Figure 7B.

was a landlocked lake, and evaporation from ~14.7–12 ky BP caused its surface to drop to ~ –100 m (Aksu *et al.* 1999) forming the β3 unconformity in the vicinity of Δ2. Our own observations on the shelf of the southwestern Black Sea confirm a lowstand of ~ –120 m during the same time period (Aksu *et al.* 2002b), so that there would have been no connection or water exchange between the Black Sea and the landlocked Marmara Sea. Proxy paleoclimate data are consistent with dry conditions at this time to account for the net evaporation (Mudie *et al.* 2002b).

During the ~16–14.7 ky BP outflow event, sands and gravels at the shallow western end of the Marmara Sea were reworked into west-directed bedforms (Aksu *et al.* 1999). We recently cored through the homogeneous mud drape that overlies these bedforms and recovered, from below the mud, a unit of gravelly sand of which 10–20% consisted of <2 cm pebbles (Figures 8 and 7A). Marine shells at the base of the mud provide a minimum age of 11,340 yr BP for this reworked gravel (Table 1), but the gravel itself was non-fossiliferous and potentially several thousand years older. It could have been formed contemporaneously with Δ2 during the ~16–14.7 ky BP outflow event when the Marmara Sea stood at –65 to –70 m. No erosional break was found within the

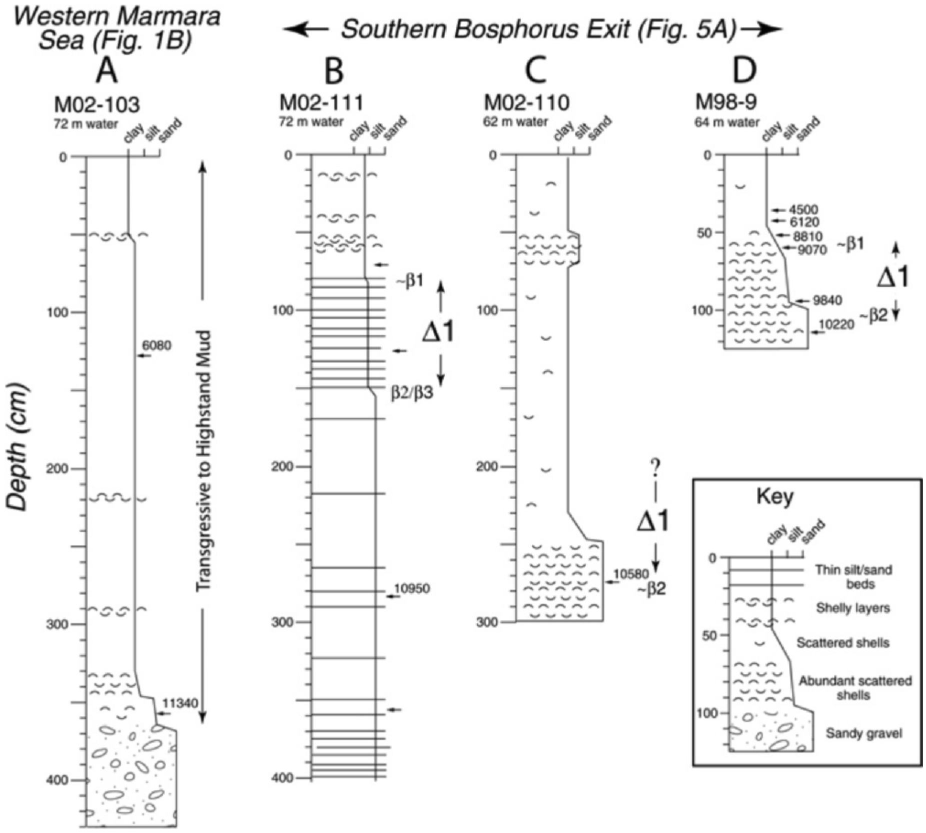

Figure 7. Graphic core logs for three new cores not described in earlier publications, and the previously described Core M98-9 (see also Figure 10). Radiocarbon dates are presented more fully in Table 1. Interpreted reflector depths are determined from seismic profiles at each core site and from changes in lithology.

mud drape, contrary to what would be expected if there had been a violent northward flow of Mediterranean water in the early Holocene as proposed by Ryan *et al.* (1997), Ryan and Pitman (1998) and Ryan *et al.* (2003).

At ~12 ky BP, the world ocean had risen to the depth of the Dardanelles sill, and Mediterranean waters rapidly refilled the small Marmara basin to its sill depth (–65 to –70 m). Because the Black Sea was still isolated from the world ocean at this time (Figure 3A) (Ryan *et al.* 2003), the Marmara Sea developed into an arm of the Aegean Sea and became fully saline. Its shelves were colonized by algal-serpulid bioherms, which are found on the crests of submerged barrier islands and bedforms in the western Marmara Sea, and on the β3 unconformity south of Bosphorus Strait (Figure 6) (Hiscott *et al.* 2002). The hiatus in Black Sea outflow during the interval ~14.7–10 ky BP provided a

Table 1. Radiocarbon ages reported as uncalibrated conventional ^{14}C dates in years BP (half-life of 5568 years; errors are 68.3% confidence limits). Data are for cores considered in this paper only; see Aksu *et al.* (2002b) for additional dates.

Core	Depth (cm)	Latitude	Longitude	Water Depth (m)	Dated Material	Radiocarbon years BP	Lab Number
M98-9	35	40°55.36'N	28°56.80'E	–64	*Anomia* spp.	4500±60	TO-7789
	42				*Nuclea nucleus*	6120±70	TO-8455
	52				*Varicorbula gibba*	8810±100	TO-8456
	60				*Turritella* spp.	9070±70	TO-7790
	94				*Turritella* spp.	9840±80	TO-7791
	113				*Mytilus* spp.	10,220±70	TO-7792
M98-12	50	40°50.54'N	27°47.68'E	–549	Bivalve fragment	4200±100	TO-8457
	130				*Nuculacea* spp.	10,660±130	TO-8458
M02-103	128	40°34.85'N	27°27.81'E	–72	*Turritella* spp.	6080±80	TO-11148
	358				*Parvicardium exiguum*	11,340±80	TO-11011
M02-110	275	40°55.61'N	28°57.10'E	–62	*Anadara* spp.	10,580±100	TO-11149
M02-111	284	40°55.31'N	28°36.13'E	–72	*Anadara* spp.	10,950±100	TO-11150
A91-20	120	38°26.00'N	24°58.00'E	–630	foraminifera	9830±70	TO-3742

unique environment for bioherm development that has not recurred since.

A second pulse of brackish-water outflow from the Black Sea began perhaps as early as ~11 ky BP (Ryan *et al.* 2003) but certainly by 10 ky BP (Aksu *et al.* 2002a). This outflow suppressed the growth of algal-serpulid bioherms and reactivated the deposition of a mud drape across Δ2 and the β3 unconformity. Soon, the outflow intensified and a second mid-shelf delta (Δ1 of Hiscott *et al.* 2002) began to develop on the shelf (Figures 5B and 9), precisely at the elevation of global sea level from 10–9 ky BP (Fairbanks 1989). At 10 ky BP, the Marmara Sea was still some 20 m below the depth of the Bosphorus sill, so the burgeoning Black Sea fed a brackish-water river that nourished Δ1. This delta is unusual because it grew upward into the rising sea level, so that the topset-to-foreset transition is shallowest in the youngest deposits (Figure 9). The radiocarbon age of the reflector that defines the top of Δ1 is 8810–9070 BP (Table 1, Figure 7D), whereas the base of the delta is younger than β3 and therefore younger than 10,950 BP (Figures 7B and 6). This base is dated at ~10,580 BP in core M02-110 (Table 1, Figures 7C and 9).

Core M02-110 sampled the bottomsets of Δ1 (Figure 9) and provides a much more reliable indication of the age of the delta than the extrapolated age previously published by us for core M98-9 (see criticism in Ryan *et al.* 2003). This second episode of Black Sea outflow, which created Δ1, was triggered by the swelling discharges of the Danube, Dniester, Southern Bug, Dnieper, and Don Rivers, augmented at times by Volga River discharge through the Manych

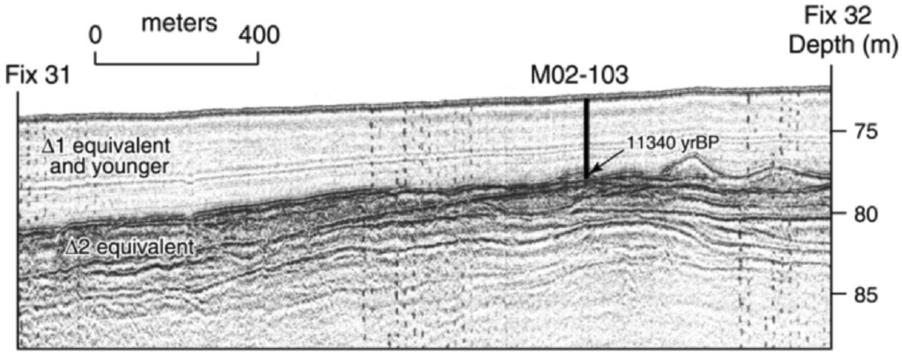

Figure 8. Huntec deep-tow boomer profile showing typical characteristics of the essentially homogeneous mud drape that overlies lowstand deposits around the Marmara Sea. The position of Core M02-103 on Figure 1B provides location. Core facies and ages are given in Figure 7B and Table 1. The highly reflective deposits below the cored interval are interpreted as fluviatile to shallow marine sands and gravels. Local crossbedding in this sediment indicates westward paleoflow (Aksu *et al.* 1999).

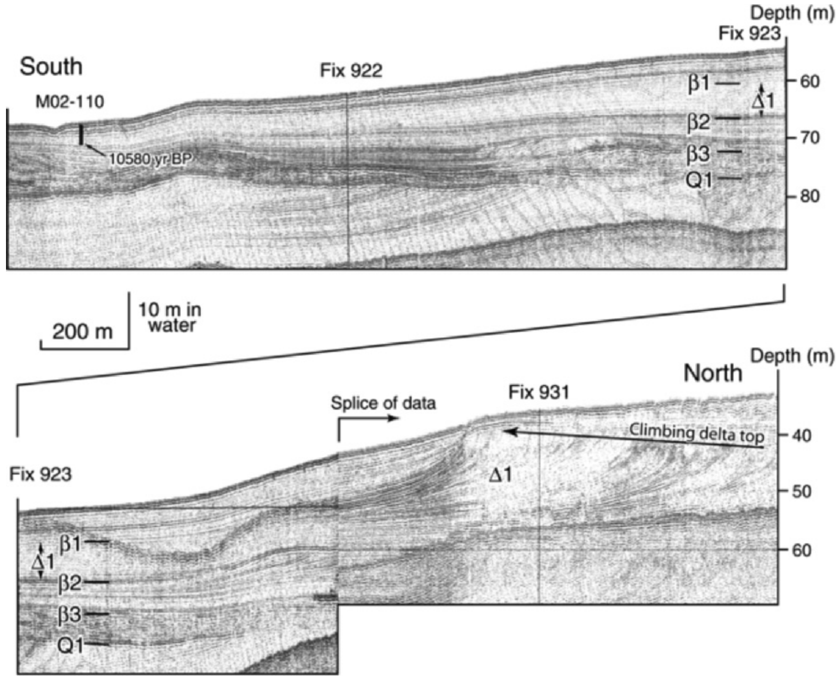

Figure 9. Huntec deep-tow boomer profile (location in Figure 5A) showing Δ1 delta front and prodelta, and the position of Core M02-110. The topset–foreset transition (offlap break) climbs consistently from right to left in the lower panel (north to south), indicating progradation during a relative sea-level rise. Foresets dip ~2°. Core M02-110 penetrates approximately to β2 at the base of the Δ1 succession; facies, radiocarbon dates, and picks of reflectors appear in Figure 7C.

outlet (Mamedov 1997; Chepalyga, this volume). The combined drainage area of these six rivers equals that of the Mississippi-Missouri system in America.

Ryan (2003) challenged our assertion that Δ1 and Δ2 were fed by Black Sea outflow through the Bosphorus Strait, suggesting instead that they are the deltas of either the Kağıthane Deresi or the Kurbağalı Dere, two small rivers that presently drain into the Marmara Sea near Istanbul (Figure 5A). This suggestion is untenable because the modern sediment discharge rates of these rivers are so small that it would take ~25,000–70,000 years to construct Δ1 (Table 2), a time span clearly at odds with the duration of delta progradation (~1000 years from ~10–9 ky BP). Hiscott *et al.* (2002) calculated Δ1sediment mass at 6.2 x 10^8 t.

Table 2. Comparison of water and sediment discharges of selected rivers with the modern Bosphorus (EIE 1999).

River or Strait	Drainage Area (km^2)	Mean Discharge (m^3s^{-1})	Suspended Sediment Discharge ($t\ yr^{-1}$)
Kocasu	21,611	151	1986 x 10^3
Kurbağalı Dere	41	1.6	9 x 10^3
Kağıthane Deresi	183	3.3	24 x 10^3
Modern Bosphorus	N/A	~10,000	surface outflow only

2.2 Evidence for Sustained Black Sea Outflow 10–3 ky BP

We have published several papers outlining why Black Sea outflow into the Aegean Sea through the Marmara Sea must have started ~10 ky BP and persisted to the present day. Black Sea outflow from ~10–6.4 ky BP is absolutely required to account for the intense water-column stratification and dysoxyia that accompanied deposition of contemporaneous sapropels S1 in the Aegean Sea (Aksu *et al.*1995) and M1 in the Marmara Sea (Ça ğatay *et al.*2000; Aksu *et al.* 2002a, c). This assertion has not been seriously challenged or rebutted by Ryan and coworkers, even though it is a fundamental inconsistency with their flood hypothesis. We remind the reader of the analogy of counting spectators as they leave a stadium in order to ascertain attendance. The large volumes of brackish water that entered the Marmara Sea from ~10–6.4 ky BP are an unambiguous indicator of persistent Black Sea outflow, because there is absolutely no other conceivable source for this low-salinity influx. Today, the volume of the semi-marine Black Sea outflow is ~50 times the combined volume of all rivers entering the Marmara Sea.

Benthic foramininfera allow estimation of bottom water oxygenation, using the benthic foraminiferal oxygen index (BFOI) of Kaiho (1994). Low values indicate dysoxic conditions at the seabed below a stratified water column (Kaminski *et al.* 2002). In the central Marmara Sea and northern Aegean Sea (Cores M98-12 and A91-20), low values confirm profound stratification through the ~10–6.5 ky BP interval (Figure 10). Dinoflagellate cysts *Brigantedinium simplex* and *Spiniferites cruciformis* (Figure 10) are sensitive indicators of low-salinity marine and fresh/brackish water conditions, respectively, and can be used to trace water masses. Mildly brackish water conditions prevailed in the Marmara Sea before its reconnection with the Aegean at ~12 ky BP (Core M98-12) and accompanied the development of sapropel in the Aegean Sea from ~10–6.5 ky BP (Core A91-20) (Figure 10). The only reasonable source of significant amounts of brackish water in the northern Aegean Sea is outflow from the Black Sea because small rivers in the region have insufficient catchments and discharges. This ouflow is confirmed in Core M98-12 by a broad peak in *Peridinium ponticum* (endemic to the Black Sea) from ~11–6 ky BP (Mudie *et al.* 2002a). Thus, Black Sea overflow began prior to 8.4 ky BP and continued into the Holocene.

Palynology indicates increased terrigenous supply of pollen from rivers beginning at ~11–10 ky BP and declining in the central Marmara and Aegean Seas by ~6 ky BP (Cores M98-12 and A91-20) (Mudie *et al.* 2002b). The persistent moderate pollen abundances at the southern exit from the Bosphorus Strait (Core M98-9) since ~9.5 ky BP are ascribed to pollen input into the Black Sea from major European rivers, and throughput of this pollen via outflow to the Marmara Sea (Figure 10).

Paleo sea-surface salinity (SSS) was calculated (as explained in Aksu *et al.* 1995) using (1) a Mediterranean-based transfer function to determine sea-surface temperatures, (2) the paleotemperature equation of Shackleton (1974) to determine $\delta^{18}O$ of the ancient surface waters, and (3) empirical data to relate these latter values to salinity. In the Aegean and central Marmara Seas, SSS dropped dramatically during deposition of approximately time-equivalent sapropels S1 and M1 (Cores A91-20 and M98-12). SSS at the southern exit from the Bosphorus Strait was depressed from ~10–9 ky BP, consistent with high abundances of the fresh/brackish-water dinocyst *S. cruciformis*.

Elevated TOC (total organic carbon) in Aegean Sea sediments coincides with lowered SSS, increased terrigenous supply of pollen, and increased stratification of the water column (low BFOI). In the Marmara Sea, TOC has been persistently high since ~11–10 ky BP, with a moderate decrease since ~5–3 ky BP away from the Bosphorus exit (e.g., Core M98-12). Pollen abundance mimics these trends, confirming the terrestrial origin of the organic matter (Mudie *et al.* 2002b; Abrajano *et al.* 2002). The core data indicate development of a brackish-water surface layer in the Marmara Sea by ~11–10 ky BP, with the strongest water-column stratification from ~10–6.5 ky BP, when widespread sapropel developed in the gateway area. The ~11 ky BP onset of stratification is consistent with the timing of the second outflow event of Ryan *et al.* (2003), but its continuation well beyond ~10 ky BP is not (Figures 3A and 4). Instead, this continuation requires unabated Black Sea outflow as proposed by Aksu *et al.*

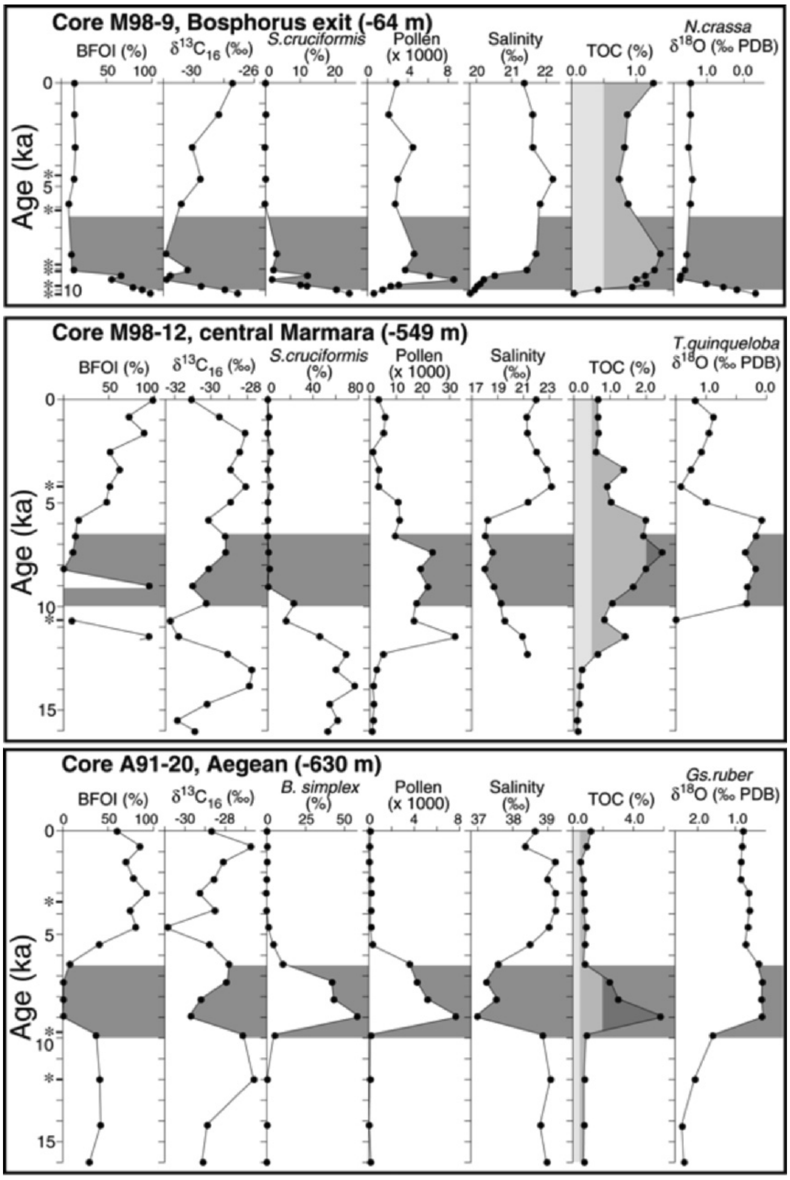

Figure 10. Downcore plots of key proxy variables; core locations in Figures 1B and 5A. Water depth at each site is indicated (e.g., -630 m). Samples were transposed into a time domain using radiocarbon dates tabulated in Aksu *et al.* (2002c), and the oxygen-isotope and ash record in Core A91-20. Control points are marked by asterisks (*) along the age scale. Where filled dots (sample positions) cluster, the accumulation rate was highest (e.g., deltaic strata at the base of Core M98-9). Where filled dots are missing for some variables, the species required for determinations were absent (e.g., *T. quinqueloba* before ~10 ky BP for Core M98-12). The gray band from ~10–6.5 ky BP coincides with sapropel deposition and significant changes in several proxy variables.

(2002a) and Hiscott *et al.* (2002).

Today, the semi-marine-water surface layer that is responsible for bottom-water dysoxyia originates entirely from Black Sea outflow. Local riverine input into the Marmara Sea is ~2% of the Black Sea outflow, and it is entirely inadequate to explain the profound water-column stratification that has prevailed in the Marmara Sea since ~10 ky BP. There was no interruption in the degree of stratification at ~8.4 ky BP, as would surely have accompanied a catastrophic flood. Similarly, the failure of open-marine foraminifera to colonize the Marmara Sea at ~8.4 ky BP is inconsistent with a major flood (Aksu *et al.* 2002c; Kaminski *et al.* 2002).

2.3 Evidence Against a Dry Early Holocene Climate

In their revised catastrophic flood model, Ryan *et al.* (2003) proposed that from ~10–8.4 ky BP, water level in the isolated Black Sea dropped by ~70 m from ~ –25 m to ~ –95 m (Figure 3A). They presumed that the drawdown of the sea surface occurred in a manner "akin to the Caspian Sea," where evaporation exceeded all inputs during warm periods of the Quaternary (Chepalyga 1984), although Mamedov (1997) reported that sea level in the Caspian was reduced by only about 15 m during the period of ~10–7.8 ky BP. Ryan and Pitman (1998) initially cited unpublished palynological data from cores collected along the Bulgarian coast as evidence for the persistence of cold, dry conditions similar to those of the Younger Dryas until 7.5 ky BP. They later presented a summary of Atanassova's pollen stratigraphy (1995) to bolster their argument for an early Holocene interval of cold, dry climate in the western Black Sea, characterized by herb-grass vegetation (Figure 11 of Ryan *et al.* 2003).

In contrast, Mudie *et al.* (2002b, their Table 2) summarized palynological data from lakes in a wide area west and south of the Black Sea and showed that oak-pistacio (*Quercus-Pistacia*) forests were present over most of the region by 10 ky BP, although local desert-steppe vegetation persisted until ~7 ky BP in the southeast, from Lake Van to the Caspian Sea. These forests indicate the early establishment of mesic climatic conditions characterized by >600 mm/year of precipitation (P) in excess of evapotranspiration (E), as is presently found in most of central and western Europe. Pollen diagrams for lakes in Bulgaria (Bozilova and Beug 1994) and for a long core from the Bay of Sozopol (Filipova-Marinova and Bozilova 2002) also show the presence of mesic deciduous oak forest before 8.4 ky BP, which accords with the establishment of a mixed coniferous-deciduous forest in Romania by about 9.5 ky BP (10,750 calBP; Björkman *et al.* 2003). Velichko *et al.* (1997) also report that broadleafed oak, lime, and elm trees were established in parts of the central Russian plains by 9.6 ky BP, following an interval between 9.9 and 9.5 ky BP during which winters were about 2° C colder.

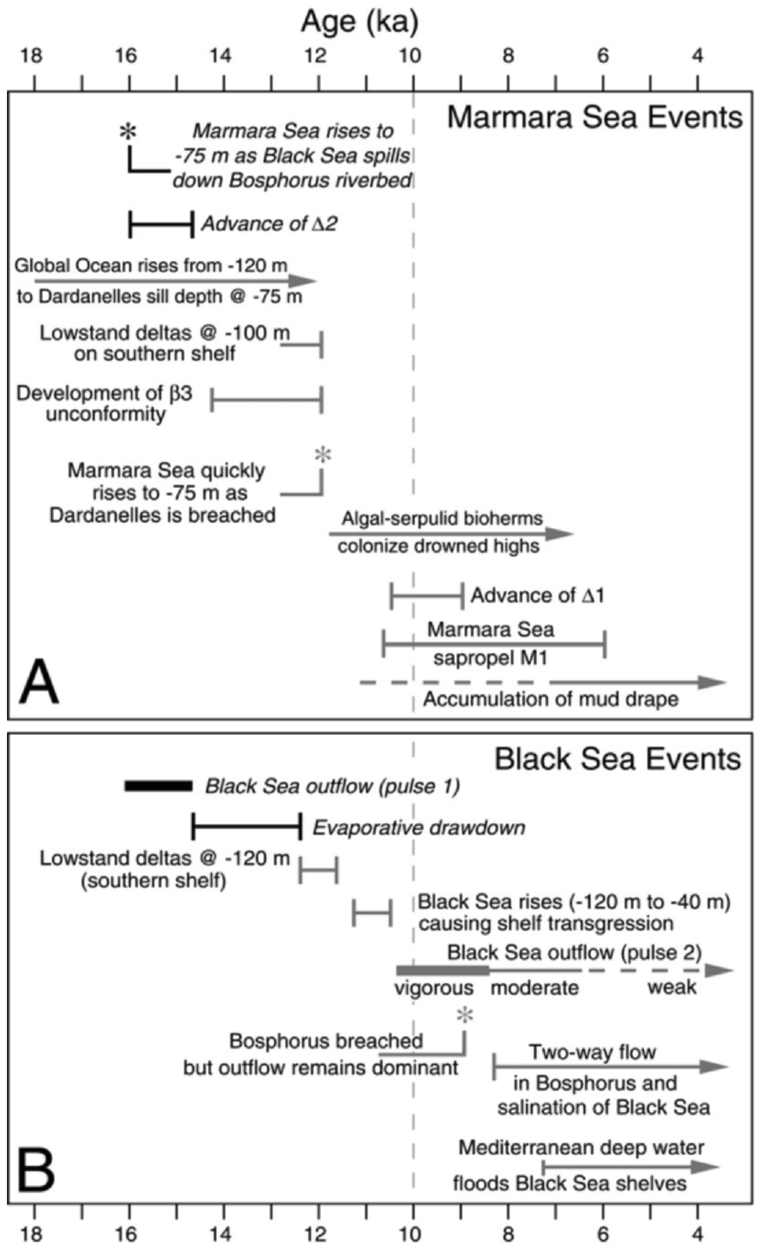

Figure 11. Summary of the paleoceanographic history of the Marmara Sea (A) and Black Sea (B). Timing and duration of some of the events colored gray are slightly modified from Hiscott *et al.* (2002, their figure 15) in order to conform with the *Outflow Hypothesis II* (Figure 4). Events in black with italics labels are additions to our earlier work. Asterisks (*) indicate events which occurred very rapidly, in tens of years to ~100 years. The Dardanelles sill depth is specified as –75 m rather than the modern value of –70 m to account for ongoing uplift of the area around the strait (Yaltırak *et al.* 2002).

Mudie *et al.* (2002b) studied marine pollen data from a deep-water core in the south-central Black Sea and from three cores in the Marmara Sea. They found an interval of ameliorating climatic conditions from 10–9 ky BP when a broadleafed oak forest tolerant of colder (–2° C) winters but requiring >600 mm per year excess moisture replaced a pre-existing, drought tolerant steppe-forest and herb-grass-shrub vegetation. They also documented increased warmth and humidity supporting a mesic, temperate, deciduous, oak-pistacio forest from 9 ky BP to the present, which was interrupted from ~8–3.2 ky BP by a climatic optimum with warmer and wetter conditions than today. Their reconstruction of the Black Sea area is consistent with precipitation and evaporation levels that Dolukhanov (1997) outlines for eastern Europe since ~11 ky BP, and it also matches the quantitative temperature and precipitation reconstructions of Velichko *et al.* (1997) for the central Russian plains from 50°–60° N.

Since the early Holocene, the Pontic Basin has apparently supported a range of vegetation types broadly similar to those found today (see Mudie *et al.* 2002b). The modern vegetation zones include the western subregion of Mediterranean woodland with a Csb climate (mesothermal, summer dry), the southern subregion of mesic Euxinian-type forest with a Cfa climate (humid subtropical conditions with year-round precipitation), and aridic steppe or savannah grassland subregions in the north and east with a semi-arid BSk climate (see Finch *et al.* 1957, for details). The eastern side of the Caspian Sea is a mid-latitude desert with a BSw climate. The Cs and Ca climate zones have an excess of precipitation over potential evaporation; the dry climate BS zones have a deficiency of rainfall (Finch *et al.* 1957).

Ryan *et al.* (2003) advocated using the water-level excursions of the Caspian Sea as a guide for interpreting the Late Quaternary changes in climate and sea level in the Black Sea basin. At the simplest level, the extreme aridity east of the Caspian Sea makes it a questionable model for their proposed 10–8.4 ky BP drawdown in the Black Sea. Nevertheless, the underlying principles that control the behavior of the Caspian Sea (Chepalyga 1984) can easily be applied to the Black Sea basin, even if the boundary conditions differ. The Black Sea basin presently has a large excess of river inflow, R, over net evaporation, P – E. (P and E are annual precipitation and evaporation over the surface of the water body, respectively. If P – E = a positive quantity, the difference is called net precipitation; if negative, we use net evaporation). This large fresh-water supply from the Danube, Dniester, Dnieper, Southern Bug, and Don Rivers drives a net outflow of ~300 km^3/year through the Bosphorus Strait (Özsoy *et al.* 1995; Polat and Tuğrul 1996). To evaluate the climatic implications of the proposal of Ryan *et al.* (2003) for a rapid early Holocene sea-level fall in the Black Sea, let us assume a closure of the Bosphorus at 10 ky BP and try to account for a subsequent drop in water level of ~70 m in the brief cold, dry interval from 10–9 ky BP. After 9 ky BP, our data indicate a climate much like today, and we see no reason to expect a negative water balance in the Black Sea at that time. If the Black Sea was indeed low at ~9 ky BP, then we would predict no further net evaporation; instead, water level might have been static or slowly rising.

We used the Black Sea bathymetric map to calculate that a water loss of ~25,600 km^3 is required to depress the sea surface from –25 m to –95 m. The average annual loss rate over a 1000-year interval would therefore have been ~25 km^3 if the Ryan *et al.* (2003) proposal of evaporative drawdown is valid. In a closed basin, a water-volume loss must be explained by an annual excess of output (evaporation, E) over inputs (precipitation, P, and river inflow, R):

$$\Delta V = R + P - E$$

If we set ΔV to -25 km^3/year, and assume that river input at ~10 ky BP was two-thirds of the modern ~350 km^3/year (to account for drier conditions), then a P – E value of ~ –258 km^3/year will satisfy the mass-balance equation for ΔV.

P – E for the modern Black Sea has been estimated by many researchers. The average of nine estimates published between 1970 and 1992 is –117 km^3/year (Jaoshvili 2002, his Table 5.1 and references therein). A parameter that can be compared among the modern Black Sea, the ~10 ky BP Black Sea, and other Eurasian marginal seas and lakes, is the *net potential evaporation* (NPE), which is obtained by normalizing P – E to the surface area of the water body: NPE = (P – E)/area. For the modern Black Sea (area = 441,000 km^2), NPE = –0.27 m/year. For the ~10–8.4 ky BP Black Sea described above (average area during fall from –25 m to –90 m elevation ~368,000 km^2), NPE = –0.70 m/year. For comparison, Table 3 lists NPE values for selected marginal seas and enclosed water bodies of central Eurasia. The Red Sea and Persian Gulf are at the most evaporative end of the spectrum, with NPE of ~ –2 m/year.

For the Black Sea to have dropped to a lowstand of ~ –95 m in 1000 years with ~66% of its modern fluvial input, the evaporative conditions over the sea surface had to be intermediate between those of the modern Caspian and Aegean seas (Table 3), leading to a net potential evaporation more than double the modern value. The Caspian region has a much drier, continental climate than the Black Sea, and the Aegean region is much warmer year-round, explaining the significantly higher rates of evaporation in both these areas relative to the Black Sea basin. Other researchers (e.g., Chepalyga 1984) assume changes in P – E of perhaps 20–30% during transitions from forest to steppe conditions, much less than the factor of two required to explain a 10–9 ky BP rapid evaporative lowering of the Black Sea. As a result, we conclude that the evaporative drawdown proposed by Ryan *et al.* (2003) cannot be explained by reasonable boundary conditions. The only alternative scenario that might permit such a rapid drawdown without recourse to unrealistic aridity would be a reduction in river flow to far less than 66% of modern discharges. However, neither the pollen data nor lake-status studies for the early Holocene (Harrison *et al.* 1996) support such a profound reduction in river flow throughout central and eastern Europe at that time.

Table 3. Estimates of net potential evaporation (NPE) for marginal and inland seas

Name	Age	P – E (km^3/yr)	Area (km^2)	NPE (m/year)	Source
Black Sea	modern	–117	441,000	–0.27	9 estimates (Jaoshvili 2002)
Aegean Sea	modern	–104	174,000	–0.60	Josey (2003)
Black Sea	~10 ky BP	–258	368,000	–0.70	This paper, to evaluate proposed drawdown of Ryan *et al.* (2003)
Caspian Sea	1900-1985	–296	371,000	–0.80	Golubev (1998), Kosarev and Makarova (1988)
Aral Sea	1926-1960	–56	65,780	–0.85	Micklin (1988)
Persian Gulf	modern	–482	241,000	~ –2.00	Johns *et al.* (2003)
Red Sea	modern	–924	440,000	–2.10	Sofianos *et al.* (2002)

As a final cautionary note regarding paleoclimate models, it is now evident that the interpretation of early Holocene herb-grassland vegetation requires careful re-evaluation in light of quantitative studies of the Younger Dryas *Artemisia*:chenopod pollen signal, formerly interpreted as indicating cold, dry conditions (Prentice *et al.* 1992). The ratio of these herbs was formerly used as a drought index, but it is now clear that the same pollen ratios can result from an increase (not decrease) in surface winter runoff. This finding has especially important implications when trying to track the switch from steppe to steppe-forest vegetation based on pollen signals.

3. NON-CATASTROPHIC EXPLANATION OF THE 8.4 KY BP AND 7.15 KY BP PALEOCEANOGRAPHIC SHIFTS

There is encouraging convergence between the Flood Hypothesis (Ryan *et al.* 1997; Ryan and Pitman, 1998; Ryan *et al.* 2003) and our Outflow Hypothesis II (Figure 4). The suggestion of a significant time lag between the first marine connection at the Bosphorus and colonization of the Black Sea shelves by euryhaline fauna (Aksu *et al.* 1999 and 2002a, based on arguments in Lane-Serff *et al.* 1997) was incorporated by Ryan *et al.* (2003) into their modified flood hypothesis (Figure 3A), so it is no longer an issue for debate. We have incorporated the two periods of Black Sea spillover identified by Ryan *et al.* (2003) into our modified outflow hypothesis (Figure 4). Where we continue to disagree is on the post-10 ky BP history of the Marmara Sea Gateway. Ryan *et al.* (2003) and Major (2002) base their requirement for a catastrophic inundation of the Black Sea at 8.4 ky BP entirely on a dramatic shift in the Sr-isotopic composition of mollusc shells to open-marine values. We see no need for such an interpretation because the isotopic shift can also be the result of a time lag in the connection process.

Lane-Serff *et al.* (1997) described the way in which two-way flow becomes established in narrow and shallow straits, using the Bosphorus channel as their example. From ~10–9 ky BP, the Black Sea was significantly higher than the Marmara Sea and the Bosphorus acted like a river (Figure 4), so there was no possibility of open-ocean water reaching the Black Sea to change its Sr-isotopic signature. Once the Marmara Sea reached the Bosphorus sill depth at ~9 ky BP, the depth of water in the strait would have been too shallow to allow sustained two-way flow. The presence of small quantities of brackish-water dinoflagellate cysts in the southern Black Sea by ~9 ky BP (Mudie *et al.* 2004) may indicate that some Mediterranean water periodically succeeded in reaching the Black Sea by that time, transporting dinoflagellate cysts northward. However, this volume was so small that the chemical composition of Black Sea shelf waters did not change (Major 2002).

The first clear evidence of marine dinoflagellate cysts (*Spiniferites mirabilis*) in the Black Sea occurs at ~8.5 ky BP (Mudie *et al.* 2004). Kaminski *et al.* (2002) independently concluded from benthic foraminiferal data that effective and sustained two-way flow was not established until ~8.5 ky BP. This dating coincides almost exactly with the dramatic shift in the Sr-isotopic composition of Black Sea molluscs that Major (2002) and Ryan *et al.* (2003) ascribe to a catastrophic inundation of the Black Sea. Instead, we propose that the 8.4 ky BP event simply marks the onset of sustained two-way flow, permitting sufficient Mediterranean water into the Black Sea basin to induce a rapid change in the Sr-isotopic composition of its waters. Today, the Black Sea has a sharp chemocline that separates more saline (~22‰) and anoxic basinal waters from less saline (~18‰) and oxygenated surface waters. A comparable chemocline likely characterized the Black Sea as it began to receive saline inflow from the Marmara Sea. Disruption of this early stratification, perhaps by a major storm or set of storms, might have been enough to trigger an homogenization of Sr-isotopic values between the deep and surface water masses at ~8.4 ky

BP. There is, we believe, no need to advocate a catastrophic flood to affect such a change.

4. CONCLUSIONS

The Outflow Hypothesis II (Figure 4) represents a rethinking and consolidation of our interpretation of the evolution of the Marmara Sea Gateway since ~16 ky BP, incorporating two pulses of Black Sea outflow suggested by Ryan *et al.*(2003) and Major (2002) for the period 16–10 ky BP. For younger times, we reject the notion that the level of the Black Sea again fell to ~ –90 m, followed by a catastrophic incursion of marine water at ~8.4 ky BP. Instead, our previously published multiproxy data from cores in the Marmara Sea, and interpretation of seismic facies in high-resolution boomer profiles, point conclusively to unabated Black Sea outflow from ~10.5 ky BP to the present. This outflow formed a cascading river from ~10.5–9 ky BP until the level of the world ocean (and Marmara Sea) reached the elevation of the sill in the Bosphorus Strait. From 9–8.4 ky BP, the outflow was sufficiently strong to prevent any measurable amount of saline water from entering the Black Sea, and so during this time, there was only one-way (outward) flow through the strait. Beginning at ~8.5 ky BP, two-way flow was established leading quickly to (1) a sharp shift in the Sr-isotopic composition of the Black Sea (Major, 2002), and (2) arrival of the first Mediterranean immigrants (dinoflagellates and benthic foraminifera). Considerably later at ~7.15 ky BP, the salinity of shelf waters in the Black Sea became high enough to permit euryhaline molluscs to thrive.

Our perception of the history of the gateway is summarized in Figure 11, modified from an earlier version published by Hiscott *et al.* (2002, their Figure 15). Except for minor shifts in the timing of events needed to conform more closely to the Outflow Hypothesis II, the only substantive change is the incorporation of a ~16–14.7 ky BP pulse of outflow which nicely accounts for the development of the older Δ2 south of the Bosphorus Strait. Previously, we had interpreted Δ2 to be much older (marine isotopic stage 3), but new radiocarbon dates show that the β3 unconformity represents a very short hiatus. When β3 was flooded by the rising Mediterranean via the Dardanelles Strait at ~12 ky BP, marine conditions were established and algal-serpulid bioherms developed on the unconformity.

We have intimately linked the development of the younger Δ1 with the onset of water-column stratification and sapropel deposition in the Marmara and Aegean Seas (Aksu *et al.* 2002a; Hiscott *et al.* 2002). When Δ2 was actively prograding during pulse 1 (Figures 4 and 11), global sea level was still depressed and the Marmara Sea was a brackish to semi-fresh lake. As a result, the water column in the Marmara Sea remained unstratified, and no sapropel developed there. However, we predict that the contemporary Aegean Sea should have become stratified as a consequence of the low-salinity discharge through the gateway from ~16–14.7 ky BP. We intend to look for organic-rich deposits of that age in the Aegean to support the proposed early history of the gateway.

The evidence for persistent Holocene Black Sea outflow is, in our view, unambiguous. The main arguments are summarized below.

(1) Calculated sea-surface salinity in the Marmara Sea has been low since ~11 ky BP, and requires a large fresh- to brackish-water input like what is provided by the Black Sea today. Local rivers cannot account for the calculated dilution of surface waters.

(2) Poor oxygenation of bottom waters in the Marmara Sea, based on the characteristics of the benthic foraminiferal communities, points to a strong and persistent Holocene pycnocline. Such intense water-column stratification can result only from the presence of a low-salinity surface layer. As before, Black Sea outflow is implicated.

(3) A climbing mid-shelf delta is present on the Marmara Sea shelf directly south of the Bosphorus exit. Local rivers cannot possibly account for its development because of their small sediment discharges. Deltaic growth is securely dated in cores from ~10.5–9 ky BP, precisely when rising global sea level was ~5 m shallower than the elevation of the topset–foreset transition of the delta.

(4) The brackish-water outflow through the Bosphorus Strait was sufficiently strong (and initially sufficiently thin) to prevent the establishment of two-way flow in the strait until ~8.5 ky BP (cf. Lane-Serff *et al.* 1997), so that salinization of the Black Sea was delayed and Mediterranean fauna and flora were unable to migrate northward.

(5) Even after the onset of two-way flow, the quantity of saline water which was advected northward across the Bosphorus Strait did not have an immediate impact on salinity of the shelf water masses. Hence, euryhaline molluscs did not successfully colonize the Black Sea shelves until ~7.15 ky BP, well after the initial connection.

(6) The widespread and homogeneous mud drape which blankets lowstand deposits in the Marmara Sea began to accumulate at ~11.5 ky BP and is uninterrupted by facies or faunal changes that would surely have accompanied a northward-flowing marine flood. Because the volume of the Marmara Sea is tiny compared with the volume of water needed to raise the level of the Black Sea from ~ –90 m to ~ –25 m (Figure 3A), a flood of the magnitude proposed by Ryan *et al.* (2003) should have left an indelible record in the Holocene succession on the shelves of the Marmara Sea. No such record exists.

More work is needed to understand fully the evolution of the Marmara Sea Gateway and Black Sea. We believe that the Outflow Hypothesis II (Figure 4) is a major step toward resolving the incompatible models of Aksu *et al.* (2002a) and Ryan *et al.* (2003). It is encouraging that new research appears to have reduced the number of disputed events. The time interval 10–7 ky BP now holds the final challenge for advocates and detractors of a catastrophic flood. Concerted study of cores spanning this time interval will eventually unlock the final mysteries of the Holocene paleoceanography of this exciting area.

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6

THE LATE GLACIAL GREAT FLOOD IN THE PONTO-CASPIAN BASIN

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Abstract: Evidence from geology, lithology, paleontology, and geomorphology reflecting the Great Eurasian Floods in the Ponto-Caspian basin is discussed. These flood events (17 to 10 ky BP) left traces on coastal plains (marine transgressions), in river valleys (superfloods) and on watersheds (thermokarst lakes) and slopes. The linkage of marine and lacustrine water bodies formed the Cascade of Eurasian Basins (the Vorukashah Sea) extending from the Aral to the Marmara Sea. It included various current and former spillways (Uzboi, Manych-Kerch, Bosphorus, and Dardanelles), covered as much as 1.5 million km^2, contained a combined water volume of about 700,000 km^3, and maintained a salinity of between 5 and 10‰. At the peak of the flood, sea level in the Caspian basin reached 190 to 200 m above the level of the previous basin. The flood's history may be divided into 10 oscillations (each lasting 500–600 years), which may be grouped into three superflood waves that have been identified in river valleys, each lasting as long as 2000 years. Such dramatic changes in sea level must have imposed substantial stresses upon coeval human populations, and the inundations probably remained in cultural memory as the Great Flood. These events might have stimulated the beginning of shipping, as well as horse domestication.

Key words: Khvalynian transgression, superfloods, spillways, Noah's Flood, civilization

1. INTRODUCTION

This paper was inspired by numerous publications that deal with the discovery of evidence for a major flood in the Black Sea. Among the many researchers who have contributed to the subject should be mentioned Bill Ryan and Walter Pitman (1998; Ryan *et al.* 1997, 2003) as well as Petko Dimitrov (Dimitrov and Dimitrov 2003). Subsequent discussion and critical remarks by Valentina Yanko-Hombach and Tschepaliga (2003), Görür *et al.* (2001), and Ali

V. Yanko-Hombach et al. (eds.), The Black Sea Flood Question, 119-148.

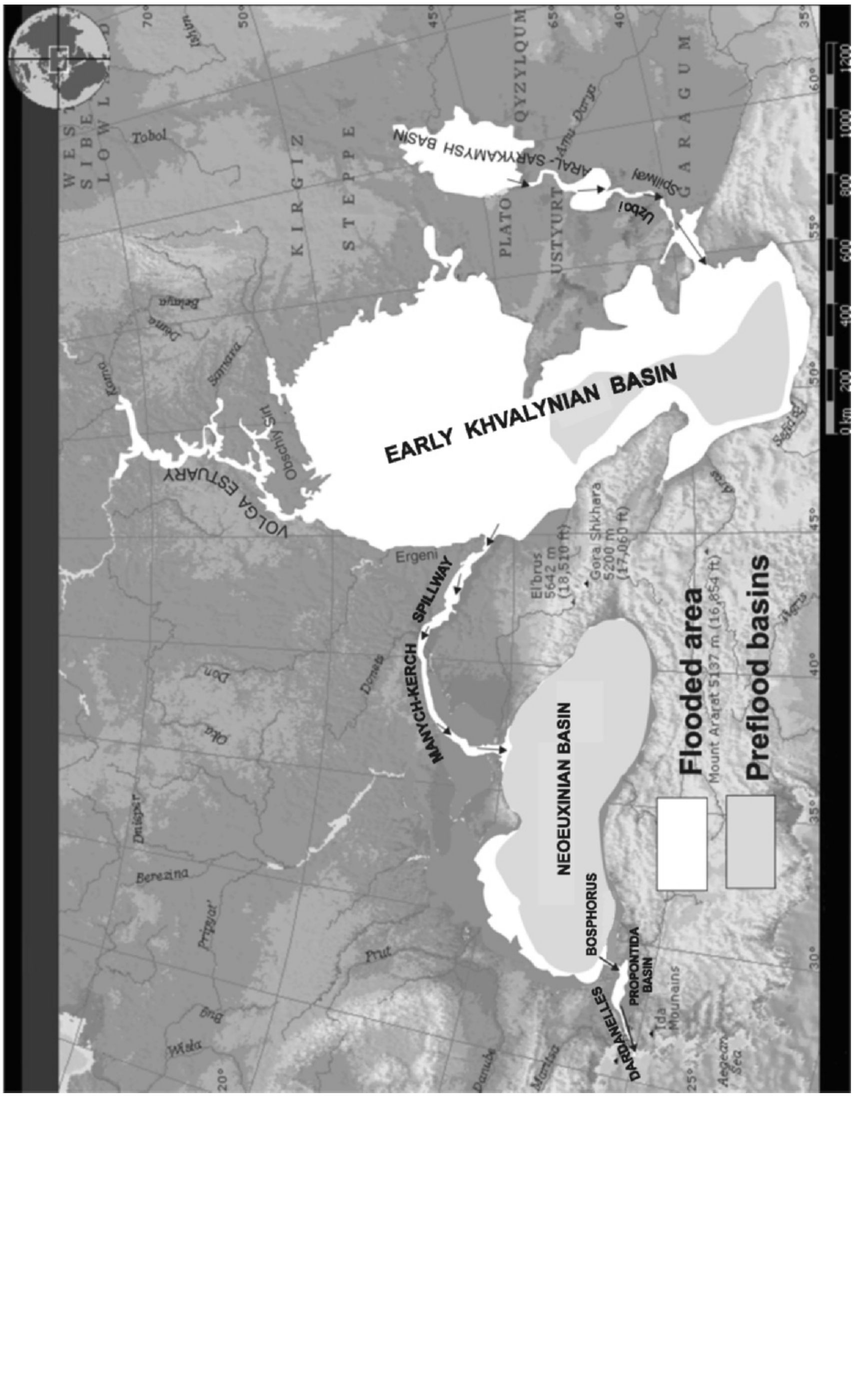


Figure 1. Ponto-Caspian Great Flood basins.

E. Aksu *et al.* (2002) in particular, aroused still greater interest in this problem.

The author's work involves the search for events which, in their dimensions and antiquity, would not be unlike the description of Noah's Deluge, retained in human memory within the Bible. The evidence presented in this paper for massive late-glacial flooding in the Black Sea-Caspian Sea region and its drainage basin within Eastern Europe derives from long-term field investigations and laboratory research, and details have been presented at a number of scientific conferences in 2003 dedicated all or in part to the subject of the Black Sea flood: the XVI INQUA Congress in Reno, Nevada (23–30 July 2003); the Advanced Research Workshop (ARW) "Climate Change and Coastline Migration," in Bucharest, Romania (1–5 October 2003); the international conference "The Black Sea Flood: Archaeological and Geological Evidence" organized by the Columbia University Seminar on the Ancient Near East in New York City (18–20 October 2003); and the Geological Society of America Annual Meeting, Session 189: "'Noah's Flood' and the Late Quaternary and Archaeological History of the Black Sea and Adjacent Basins" in Seattle, Washington (2–5 November 2003) (Chepalyga 2002, 2003, 2004a, b; Yanko-Hombach and Tschepaliga 2003; Chepalyga and Yanko-Hombach 2003).

The first stage of the overall research strategy entailed a search for extreme hydro-climatic events, such as marine transgressions, during the last 18–20,000 years in the Pontic and Caspian basins. Attention was focused on finding possible sources of water for such events–for example, overfloods in river valleys and relict permafrost thawing on watersheds and slopes. The second stage involved chronocorrelation of the events using geomorphological and stratigraphic evidence together with available radiocarbon dates. This effort was followed by paleohydrological reconstructions of the basins, including their level, areas, water mass volumes, and the nature of the water exchange between basins. Particular emphasis was placed on calculations of flood dynamics: rates of water-level rise, coastal lowland flooding and coastline shifts, and related hydrographic changes that could have resulted either in population migrations from flooded territories or barriers preventing ancient cultures from interacting with each other.

Finally, on the basis of archaeological data, the possible influence of these events on late Pleistocene human societies of the Black Sea littoral areas was examined. The aim of this investigation was to develop a comprehensive concept of the flood and, possibly, to link it with the events engraved in human memory. With the use of "Great Flood" (or "Flood"), this writer refers specifically to the late glacial inundation within the Ponto-Caspian basin at ~17 to 10 ky BP (with its maximum at 17–14 ky BP).

If the Biblical Flood was a real historical event, then besides tales and myths, it had to leave certain traces in bottom sediments of the sea, in fossils, landforms, old coastlines, and other aspects of the geologic record. The author's

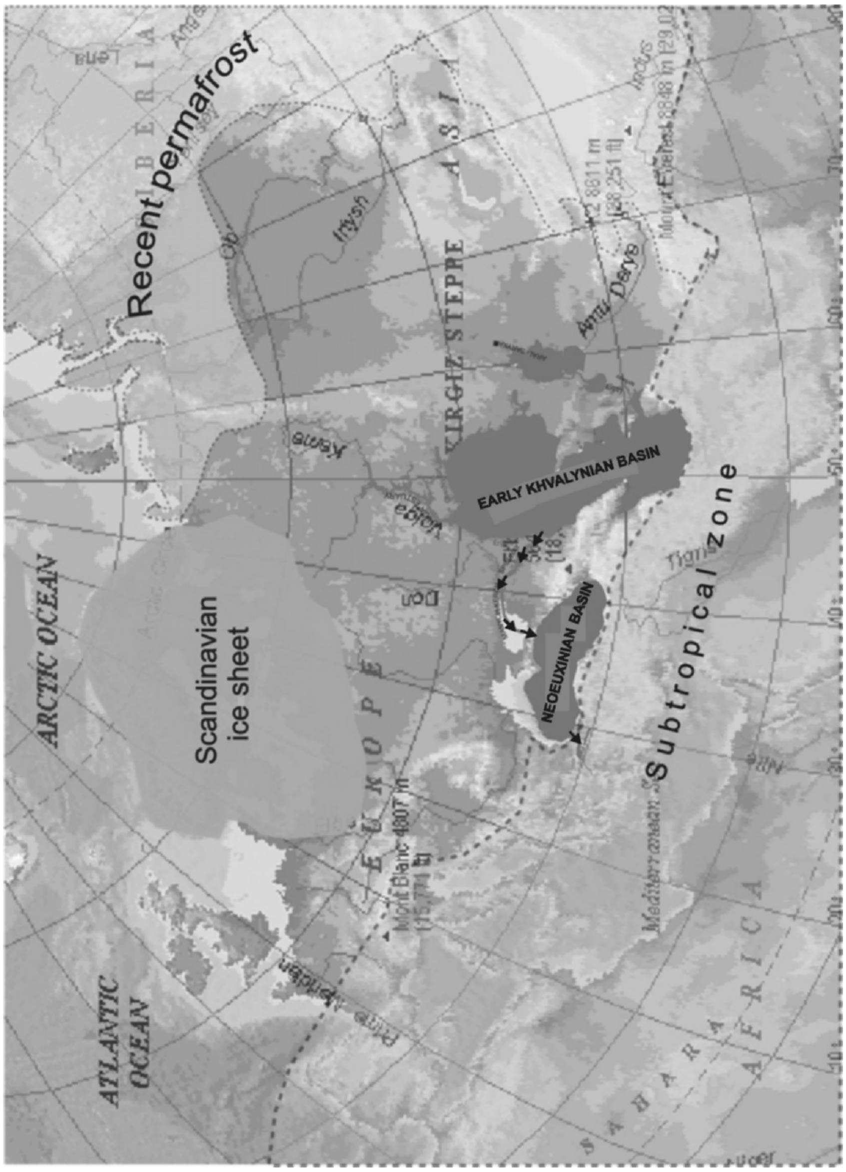


Figure 2. Great Flood events area.

investigations revealed traces of massive flooding in the Ponto-Caspian region (Figure 1) and its neighboring drainage basin at the time of the melting of the last (Valdai) ice sheet, about 17 to 10 ky BP. This flooding left its mark on various landscapes, including coastal plains, river valleys, interfluvial surfaces, and even on slopes. The following sections will discuss the geological, geomorphic, and lithological traces of the Flood in greater detail.

2. FLOOD GEOLOGY

Bottom and littoral sediments of the different basins, and the fossils they contain, constitute the geological evidence of the Flood. Detailed analysis of their lithology, mineralogy, and geochemistry, as well as isotopic composition of the sediments and fossils, make possible the reconstruction of sedimentary environments, composition of the flood water, and the sequence of events.

2.1 Flood Sediments

In the epicenter of the Flood, which lay within the Caspian basin, bottom sediments attributable to this event are dated to the Khvalynian interval (the maximum phase of the Flood was in the Early Khvalynian). Khvalynian sediments differ from the immediately under- and overlying layers in many respects (Badyukova 2000; Chistyakova 2001; Leonov *et al.* 2002). Typically, Khvalynian layers contain "chocolate clays," so-called because of their characteristically reddish-brown color. Locally, these clays are interlayered with thinly laminated (1–2 cm) greenish-gray and dark gray clays. The chocolate clays also interlayer with–and pass laterally into–silts, sandy loams, and occasionally sands possessing a distinctly high proportion of clayey matter together with marine molluscs of Caspian type.

The chocolate clays and related Khvalynian sediments do not usually exceed a few meters (3–5 m) in thickness, but sometimes they reach 20–25 m or more. They are mostly confined to the Caspian Lowland, from the modern Caspian coast to the foot of bordering elevations (Ergeni, Obshchiy Syrt, Privolzhskaya, Stavropol), but they are also found in the Volga and Ural estuaries. Surface area of exposed Khvalynian sediments amounts to 0.5 million km^2, while their total area of distribution is 1 million km^2 (Figure 2).

Dominant within the clastic sediments are poorly rounded quartz grains, with less common micas, carbonate clasts, feldspars, coal, and occasional epidote, hornblende, zoisite, tourmaline, and zircon. Authigenic minerals, such as iron hydroxide (up to 65%), gypsum (single crystals and aggregates), and glauconite are less frequent (Chistyakova 2001). Clay minerals are represented by

Table 1. Caspian Sea stratigraphy for the Late Pleistocene and Early Holocene.

Caspian stratigraphy			Khvalynian terraces of the Caspian	Caspian sea-level oscillations		Great Flood events	
				Highstands (transgressions)	Lowstands (regressions)	Oscillations	Waves
HOLOCENE					Mangyshlak –50 m	Deep drying	
LATE PLEISTOCENE	KHVALYNIAN	Late	Terrace –16 m	Sartas –12 m		X –16 m	Wave III of Flood Complete isolation; 4 oscillations Flooded area = 150–200,000 km²
					Regression –30 m	Drying	
			Terrace –5 to –6 m	Kuma –5 to –6 m		IX –5 to –6 m	
					Bekgash –28 m	Drying	
			Terrace 0 to –2 m	Mahachkala 0 to –2 m		VIII 0 to –2 m	
					Regression –28 m	Drying	
			Sediments –6 m	Transgression –6 m		VII –5 to –6 m	
					Enotaevka –100 to –120 m	Deep drying	
		Middle	Terrace +6 m	Unnamed +6 m		VI +6 m	Wave II of Flood Partly isolated; 3 oscillations Flooded area = 400–450,000 km²
					Regression –20 to –30 m	Drying	
			Terrace +16 m	Turkmen +16 m		V +16 m	
					Regression –20 to –30 m	Drying	
			Terrace +20 to +22 m	Bujnaksk +20 to +22 m		IV +22 m	
					Elton –50 m	Deep drying	
		Early	Terrace +35 m	Talginka +35 m		III +35 m	Wave I of Flood Outflow to Black Sea via Manych; 3 oscillations Flooded area = 850,000 km²
					Regression 0 to –20 m	Drying	
			Terrace +48 to +50 m	Maximal +50 m		II +50 m	
					Regression 0 to –20 m	Drying	
			Yashkul sands +40 m	Yashkul +40 m		I +40 m	
	ATELIAN				Atel –100 m	Pre-Flood desiccation	

smectite, kaolinite, montmorillonite, chlorite, and hydromica. In sections along the Lower Volga (Middle Akhtuba), chlorite is typical of the lowermost Khvalynian sediments but almost disappears upwards (Chistyakova 2001); this may indicate changes in the sources of clay material.

The characteristic reddish-brown color of the chocolate clays cannot be attributed to free iron oxides; it is most probably related to iron-containing clay minerals. The low content of carbonates in the clays or their complete absence suggests a cold climate, as solubility of carbonates increases at low temperatures, permitting them to remain in solution. On the other hand, abundance of chemogenic dispersed carbonates with no secondary changes recorded in the terrigenous pellitomorphic clays suggests sedimentation under arid climatic conditions. The beginning and peak of the transgression reveal evidence of aridity and increased evaporation. Judging from the sediment geochemistry and composition of authigenic minerals, therefore, the Khvalynian transgression is more likely to have developed in an arid rather than a humid climate. This information conflicts with the existing climatic hypothesis of the Khvalynian transgression, which attributes it to a wetter climate. A model that avoids the contradiction will be suggested below.

Recently, the Khvalynian sediments (and the chocolate clays, in particular) have been considered "cryo-suspensites" resulting from rapid melting of permafrost and activation of solifluction processes during warmer phases of the Valdai deglaciation (Chistyakova and Lavrushin 2004).

2.2 Stratigraphy

In the Caspian basin sequence, Khvalynian layers occur above the Late Khazarian (dated to the last interglacial) and below the New Caspian (Holocene) deposits (Table 1). They are separated from the Lower Khazarian series by continental Atelian layers synchronous with marine sediments of the Atelian regressive basin. The level of the latter was 110–120 m below today's Caspian sea level, that is –140 to –150 m (Lokhin and Maev 1990; Maev 1994; Maev and Chepalyga 2002). In the Caspian Lowland, Khvalynian sediments occur mostly close to the surface. Younger (and higher in the sequence) are Holocene floodplain lacustrine and marine (New Caspian) sediments.

Khvalynian sediments are divided into three parts: Lower, Middle, and Upper. The Lower Khvalynian rests on a base of Atelian loams and is overlain by Elton continental sediments. In the Caspian Lowland, they are exposed at the surface in the range of +48 to +25 m. Middle Khvalynian sediments are confined between the Eltonian and Enotaevka regressions, and they outcrop between isohypses 25 and 0 m. Finally, the Upper Khvalynian crowns the marine sequence and outcrops below the zero isohypse, though above the New Caspian transgression limit (–22 to –25 m).

In the Manych Depression, reddish-brown clays and silts of the Abeskun layers (Popov 1983) may be considered an analog of the chocolate clays. They are exposed on the present-day surface and contain fossil molluscs of the Khvalynian basin, including *Didacna*, *Monodacna*, *Adacna*, *Hypanis*, *Dreissena*, and *Micromelania*. The sediments make up constructional landforms (ridges) in the Manych Spillway and may be correlated only with the Early Khvalynian sediments and the main flood event at 17–14 ky BP. The absence of younger deposits with Caspian fauna suggests the spillage flow from the Caspian Sea had stopped by then.

Flood deposits in the Black Sea occur within the Neoeuxinian series. On the continental slope and within the deep-sea basin, they form light reddish-brown and pale yellow muds (red clays) about 0.5–1.0 m thick (Ryan *et al.* 2003). In color, they are not unlike the chocolate clays of the Caspian basin, and their age of 15 ky BP is also close to that of the latter. They have been identified (by the author together with W. Ryan) in core 1 (section 3) of the DSDP hole 380 (drilled north of the Bosphorus Strait) within the 384–450 cm depth interval.

2.3 Flood Fossils

The main indicators of the Flood are specific brackish-water mollusc species close to modern ones from the North Caspian. Among them are endemic Caspian species belonging to the Limnocardiidae family, such as the genus *Didacna* Eichwald (Nevesskaya 1965). This genus is not presently found anywhere outside the Caspian Sea, while it occurred widely in the Azov-Black Sea basin during the Pleistocene up to Karangatian time. The genus is represented by *Didacna praetrigonoides* (dominant), *D. parallela*, *D. delenda*, *D. subcatillus*, *D. ebersini*, *D. pallasi*, as well as relatively deep-water (>25 m) *D. protracta*. Other endemic limnocardiides characteristic of the region are *Monodacna caspia*, *M. laeviscula*, *Adacna vitrea*, and *Hypanis plicata*. Of the Early Khvalynian elements, molluscs of the *Pontodreissena* subgenus are the most common outside the Caspian Sea (*Pontodreissena rostriformis* and *Dreissena polymorpha* in semi-freshwater basins). Gastropods are usually represented by the endemic Caspian genera *Caspia* and *Micromelania*.

Shells of the Early Khvalynian complex are distinctive in their small size (2–3 times smaller than those of today) and thin walls. The complex is usually considered to represent an adaptation to cold climate and low salinity. It is known, however, that a cold climate fosters the development of larger individuals. As for salinity, it is unlikely to have been much lower than it is in the North Caspian at present (10‰ and greater) as is indicated by the rich species composition. It seems more likely that small size and thin walls result from a considerable turbidity of water and lack of oxygen at the bottom of the basin. The higher turbidity, in turn, could be a consequence of intensified solifluction on slopes under conditions of permafrost decay.

Neoeuxinian sediments contain mollusc fauna of the Caspian type. Dominant are *Pontodreissena rostriformis*, rarer are *D. polymorpha*, the limnocardiides *Monodacna caspia*, *M. colorata*, *Adacna*, *Hypanis*, and the gastropods *Caspia* and *Micromelania*. The genus *Didacna* is entirely absent from the Black Sea, though it has been traced along the Manych Depression as far as the Western Manych River mouth (Manych-Balabinka village). This may indicate a lower salinity in the Neoeuxinian basin (5 to 6‰). Fauna of Caspian type, close to those described above in composition, were found in the Sea of Marmara (Propontida basin) and within Bosphorus bottom sediments in hole 14 at water depths of 800 to 100 m, and dated to 26–10 ky BP (Algan *et al.* 2001). The Caspian mollusc *Pontodreissena rostriformis* is dominant.

The Khvalynian and Neoeuxinian marine sediments also contain microfossils, such as foraminifera, ostracoda, and diatom algae, with endemic Caspian elements.

2.4 Flood Geomorphology

The floodwater left distinct traces in the morphology of landforms, such as marine terraces, specific coastlines, flattened seafloor surface, as well as sculptured and constructional landforms within the former spillways: the Manych-Kerch Spillway, and the Bosphorus and Dardanelles Straits.

2.4.1 Spillways

The Manych-Kerch Spillway is a large trough, deeply eroded into solid rock, that connected the Caspian and Black Seas (Figure 3). It was inherited from an older strait between the two seas, which existed (with interruptions) since the Late Pliocene Akchagylian basin (Popov 1983). It follows a tectonic depression that skirts the southern periphery of the Karpinsky Swell (an elevated Mesozoic structure confined between the Donbass and Mangyshlak). The total length of the spillway amounted to 950–1000 km (depending on the location of sea level), with maximum and minimum widths of 50–55 and 10 km, respectively (Chepalyga *et al.* 2004). Its depth attained 30–50 m. The spillway bottom gradient was 0.0001 (10 cm/km), and the drop in water level from the Caspian Sea (+50 m) to the Black Sea (–80 to –100 m) reached 150 m at the beginning of the excess water flow; by the end of this flow, the drop was 100 m.

The spillway began from the Khvalynian coastline at the head of Chograi Bay between the Ergeni and Stavropol uplands. The bay is 60 km long and 30 km wide at its entrance, and its depth is between 40 and 50 m. The narrowest section of the spillway (about 10 km) was near the Zunda-Tolga village, where the water flowed over a sill at 20 m above sea level (Figure 4).

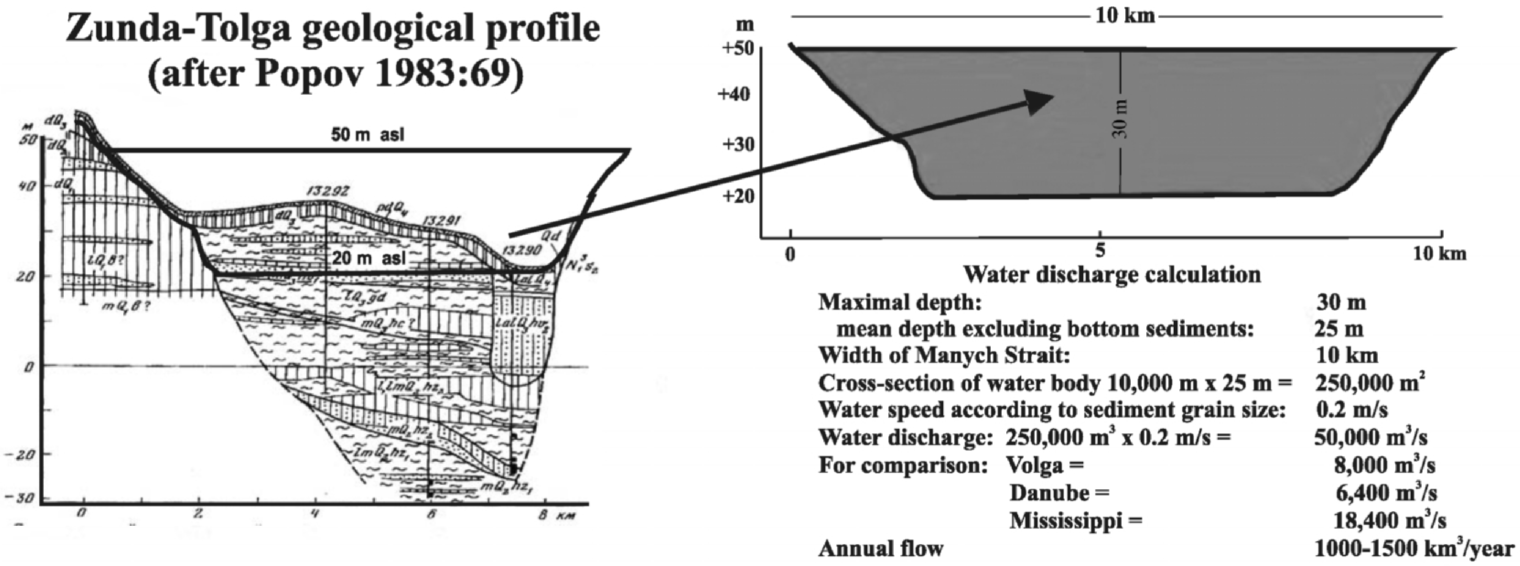


Figure 4. The Zunda-Tolga profile.

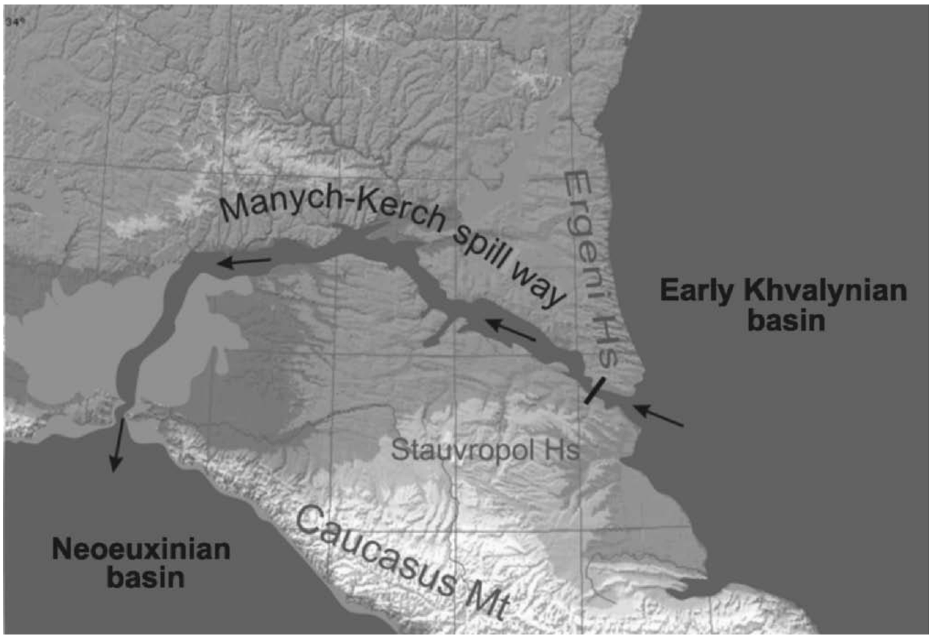


Figure 3. The Manych-Kerch Spillway; bar is located near the Zunda-Tolga section of Figure 4.

When the Khvalynian transgression was at its maximum (50 m asl), the spillway depth was up to 30 m (its average depth was 20–25 m). The spillway bed is covered with silt and clay 5 to 10 m thick. The cited data enable us to estimate the flow velocity at ~0.2 m/s and maximum discharge through the Manych Spillway at 40 to 50 thousand m^3/s. The total runoff would have amounted to more than 1000 km^3/year. This output is six times greater than the Volga River runoff and three times that of the Mississippi River.

In the above calculations, sill depth was assumed to be constant at 20 m asl, but with a stream depth of 30 m, the flow velocity would have been much greater. This scenario does not agree with the fine composition of the sediments, however. Such a contradiction may be explained by assuming a higher initial sill level, about 40 m asl or even higher, in which case, the flow discharge would be reduced by several orders of magnitude to near the modern Volga discharge (8 to 10 thousand m^3/s).

A short distance downstream, at the Kalaus River mouth, a plug was formed by merged fans of the Kalaus and Zapadny Manych Rivers (Badyukova 2001). This plug blocked the spillway and formed a divide between the Caspian and Black Sea basins at 25 m asl. Farther west lay the widest part of the spillway, a 180-km-long section up to 50–55 km wide, which is now occupied by the Manych-Gudilo Lake. This section of the channel was braided, as is indicated by alluvial landforms. Among the latter, there are several (up to 5–7) parallel ridges 10 to 15 km long and 20–30 m high. Ridge widths range from a few hundred meters to 1–2 km. The ridges are composed of clays and silts, less frequently of sands with Khvalynian marine fauna (analogous to that recovered from the chocolate clays). At present, the interridge hollows are occupied by water bodies. Near the Salsky Swell, the spillway narrowed to 15–20 km due to tectonic uplift, and the stream flowed within a single channel, without braiding.

At the Manych River mouth, the spillway occupied the entire Don River valley, its surface level reduced to 0 m and its bed resting at –15 to –20 m. Within the limits of the modern Azov Sea, the spillway occupied an over-deepened valley, with its bed at –40 to –50 m and its width varying from 10 to 20 km. In the Kerch Strait, the stream narrowed to 6–8 km but again widened on the Black Sea shelf, forming a delta within the depth interval of –80/–100 to –40/–50 m.

2.4.2 Coastlines

The coastline of the Early Khvalynian basin was fundamentally different from that of today because the high sea level had risen to the slopes of the surrounding uplands (Ergeni, Obshchiy Syrt, Privolzhskaya). Instead of depositional coasts with shallow, flat-bottomed bays of intricate shape typical of the Caspian Lowland–the so-called Caspian type (Leont'ev *et al.* 1977)–and the large deltas of the Volga and Ural Rivers, abrasion-embayed coasts were created with deep embayments of the liman type, where the sea ingressed into the valleys of dissected uplands. An example is the bay in the Yashkul valley that penetrates the Ergeni Upland for 50 km; it is filled with chocolate clays bearing Khvalynian fauna. The most spectacular coastline feature dated to the time of the Flood is the Volga estuary (Figure 1), which is filled with chocolate clays upstream beyond the Zhiguli Ridge. The estuary at the time was about 800 km long, 50–80 km wide (narrowing locally to a few km) and up to 20–30 m deep. The backwater reached Cheboksary in the Kama valley. Other estuaries (of the Ural, Terek, and Emba Rivers) were smaller in size but still differed markedly from their modern river mouths, which are mostly of deltaic type.

2.4.3 Marine Terraces

Marine terraces mark the position of sea level and coastline at every individual oscillation during the regression of the Khvalynian Sea. Because the flood basin level was unusually high, its sediments mantle much older terraces. In tectonically stable regions (Daghestan), they form as many as nine marine terraces at elevations of 48, 35, 22, 16, 6, –5, 0, –6, and –16 m (Leont'ev *et al.* 1977; Rychagov 1997). These terraces mark episodes of temporary sea-level stability during the general reduction of the basin. The stable episodes were separated by regressive phases during which sea level fell by tens of meters.

The largest regressive phases (within the Khvalynian stage) were the Eltonian (to –50 m) and the Enotaevka (to –100 m). The presence of the terrace steps suggests a series of the Khvalynian basin fluctuations in its later (regressive) phases. The levels are marked by submarine fans at the Mangyshlak sill (Lokhin and Maev 1990; Maev 1994).

2.5 Results: Geography of the Flood

Studies of the spatial distribution of the flood events definitely show that "the Great Flood" was local, not global, in occurrence, but that it was nevertheless widespread and impacted four types of landscape: (1) coastal plains (as marine transgressions of the Ponto-Caspian basin), (2) river valleys (as superfloods and associated macromeanders), (3) interfluves (with thermokarst lakes of alas type), and (4) slopes (as solifluction flows).

In Northern Eurasia, such events took place over a substantial range within the temperate zone, from the AtlanOcean to the Yenisei River (Figure 1). To the north, they reached the Scandinavian ice sheet and the present-day limit of permafrost. To the south, they were bounded by the subtropics and the Alpine-Himalayan mountain belt.

3. CHRONOLOGY OF THE FLOOD

It has not been easy to date this Great Flood and determine its duration and characteristics. Recently, however, new evidence, including radiocarbon and other dates, has permitted the age of the Flood to be determined and its dynamics to be reconstructed (Svitoch *et al.* 2000; Leonov *et al.* 2002). Estimates of the historical age of the Flood of legend, as defined by various authors, varies from 4.5 to more than 10 thousand years ago. In Mesopotamia, it is dated to 4500–6000 years BP (Rohl 2002), but this flood was not on the scale of "Noah's Flood" according to the written sources. It seems more like a local, though extensive, inundation. As for Noah's deluge, recent research has proposed placing it between the 12th to 9th millennia BC (Balandin 2003), which is older than 13 to 12 ky BP. According to this thinking, the Flood occurred well within late glacial time, and not at its very end. The duration of the Flood also varies with the sources, from a fortnight to a few months. Theological sources from the Bible have offered a precise date for the Flood: 9545 years BC (Leonov *et al.* 2002), which is 11,949 years ago.

Dates close to these have been obtained for flood deposits in the Caspian (Khvalynian) basin, from Neoeuxinian sediments of the Black Sea, and fluvial

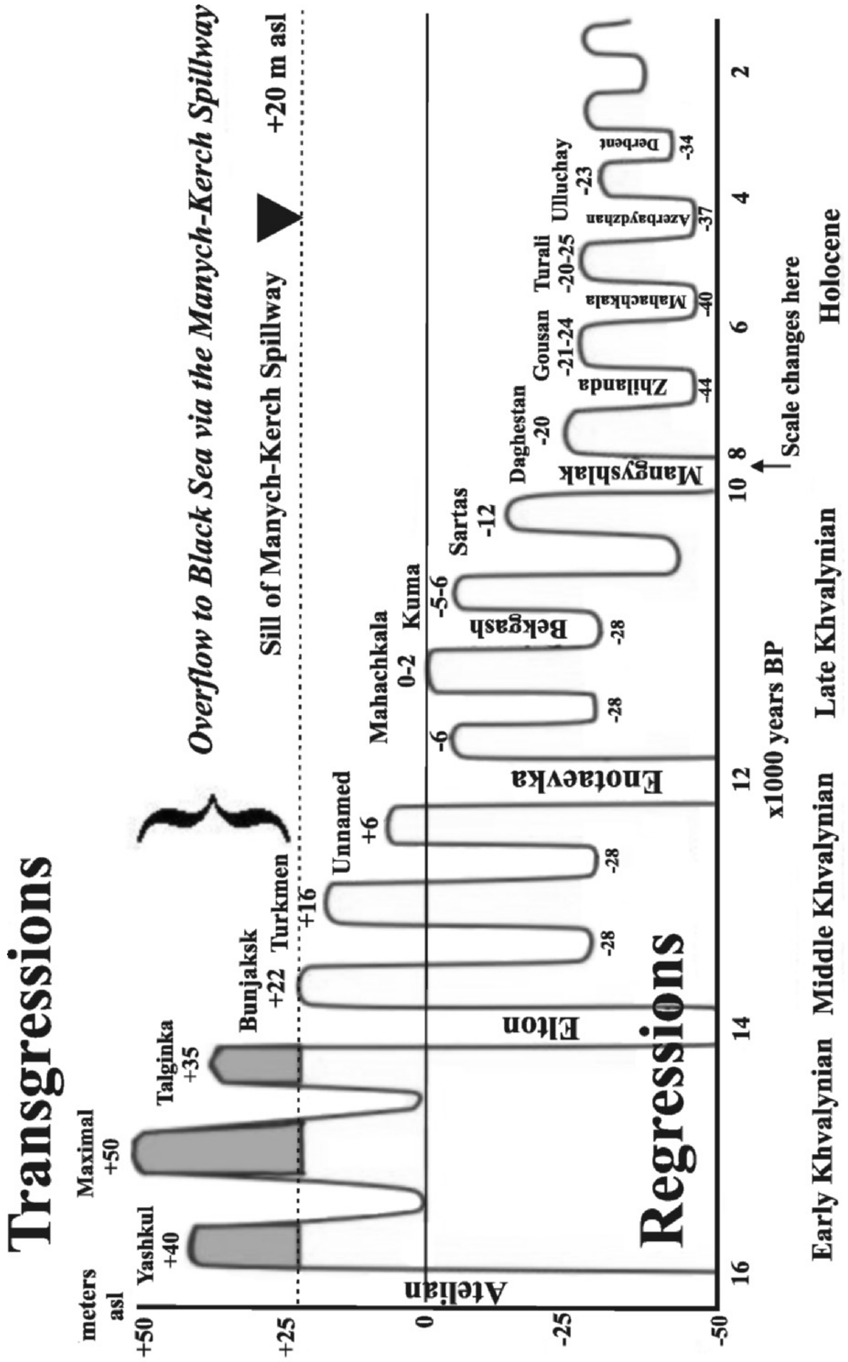


Figure 5. Chronology of the Caspian transgressions 16–0 ky BP.

deposits filling macromeanders in river valleys. The Khvalynian transgression of the Caspian is the most extensively studied of these flood residues (Varushchenko *et al.* 1987; Rychagov 1997; Maev and Chepalyga 2002; Leonov *et al.* 2002). More than 50 ^{14}C dates have been obtained; some (except for extreme ones that must be erroneous) are shown in Table 2.

Most dates cluster in the range of 17–9 ky BP, with a flood duration estimated at 5–6 thousand years. Early Khvalynian sediments date to 17–14 ky BP, Late Khvalynian are 11–9 ky BP, and Middle Khvalynian layers fit between them, at 13–11 ky BP. Of the three subdivisions, only the Early Khvalynian interval may be considered properly a period of "Great Flood." At this time, the Caspian level rose by 180–190 m (Maev 1994).

Table 2. Radiocarbon dates from Khvalynian mollusc shells.

Lab Number	Uncalibrated ^{14}C	Lab Number	Uncalibrated ^{14}C
MGU-1039	10,770±330	LU-1353	12,690±100
MGU-1037	11,280±700	MGU-19	12,600±240
LU-425V	11,740±180	LU-1359	12,010±340
MGU-98	11,600±140	LU-1357	12,210±150
MGU-1034	11,290±380	MGU-99	12,050±190
LU-841	11,490±380	LU-490A	12,520±140
LU-1358	11,390±200	MGU-19	12,600±240
LU-426B	11,600±1000	LU-1353	12,690±100
MGU-792	11,760±200	MGU-25	13,100±300
LU-864a	11,830±200	LG-93	14,080±100
MGU-793	11,820±250	MGU-18*	15,600±300
MGU-IOAN-38	12,150±200	MGU-18	15,500±350
GIN-66	12,500±140	MGU-97	16,000±330

* Two different shell fractions of MGU-18 were analyzed

The system of Khvalynian terraces in the Caspian basin served as a chronological scale to determine the time of the Flood and its position among other "Flood-like" events (it should be considered as part of the entire sequence of events). The terraces mark highstands of Caspian water against a background of its intermittent lowering. There are as many as nine Caspian terraces, with one relatively lower level (the Yashkulian transgression at 35 to 40 m) preceding the highest stand at 50 m. During Khvalynian time (estimated at 5–6 thousand years), about 10 cycles of sea-level fluctuations occurred, with a periodicity of 500–600 years (Table 1). They can be grouped into three series of ~2000 years each: Early Khvalynian levels (40, 50 and 35 m); Middle Khvalynian levels (22, 16, 6 m); and Late Khvalynian levels (–6, 0, –5, –12 m). They are separated by two regressive phases, the Eltonian and Enotaevka.

Water-level fluctuations in the Khvalynian basin led to coastline migrations over hundreds and thousands of kilometers, leading to large-scale flooding and then drying of the sea bottom. The major oscillations may be considered as waves of the Flood (Figure 5). The first wave (the Early Khvalynian) began 17–16.5 ky BP and lasted about 2000 years. It was complicated by three superimposed smaller oscillations, with sea level rising to 40 m, 50 m, and 35 m. The sill in the Manych Spillway at about 20 m asl was surpassed by all three transgressions, which overflowed into the Black Sea. It is this first wave, and its rising phases in particular, that should be considered the Great Flood in the Ponto-Caspian region.

The second wave (the Middle Khvalynian) did not exceed 22, 16, and 6 m, even during its oscillation peaks. Thus, Caspian water did not flow into the Black Sea, and the Manych Depression, in all probability, did not function as a spillway at this time.

The third wave of the Flood (the Late Khvalynian) did not surpass modern ocean level (0 m). All its oscillations (–6, 0, –5, and –12 m) remained below it, though above Caspian level during the Holocene.

Caspian water-level fluctuations demonstrate two distinct trends. The first trend, exhibited during the period of 16 to 9 ky BP, was marked by a progressive lowering of sea level over successive oscillation peaks from +50 m to –12 m; this represents a drop of 62 m over 6 ky. This period marks the occurrence of the Flood. The second trend is one of relative stability during the Holocene, with a 6 m difference in peak height (–20 and –26 m) over 10 ky.

Therefore, the Flood phase itself (i.e., the active sea-level rise) occurred sometime between 16 and 15 ky BP. The Caspian rose by 180–190 m over 100–150 years. The last value is inferred from a demimillenial cycle of 500–600 years, based on the assumption that sea-level rise, highstand, and subsequent sea-level drop each lasted for approximately the same duration. The rise in sea level could have taken even less time if it resulted from a sudden warming during the glacial Heinrich event N 1 (15–14.3 ky BP). This event was of global character, and the warming was accompanied by surging of arctic glaciers (Grosval'd 1999), a high rate of glacier decay, and eustatic rise in ocean level.

4. FLOOD HYDROLOGY: MARINE BASINS

The most sizeable events, comparable with the ancient floods recorded in legend, occurred in the inner seas and lakes of the Ponto-Caspian basin.

4.1 The Kvalynian Sea

The Khvalynian Sea appears to have been the epicenter of the Flood and the most sensitive indicator of the related events (sea-level rise, coastline shift, and coastal lowland flooding). This basin concentrated the bulk of the Flood water, altered the water composition and marine environment, while excess water escaped into the Black Sea.

In the process of flooding, the Khvalynian Sea expanded over an area of about one million km^2, up to 1.1 million km^2 if the Aral-Sarykamysh basin is included. The total area was three times that of the present Caspian Sea, and the accumulated water volume (which reached 48–50 m asl at peak flood stage) was twice that of today's Caspian (130,000 km^3). The type of basin changed as well: the isolated and closed lake of the Atelian basin was transformed into a gigantic through-flow lake-sea, with one-way discharge into the adjacent basin. Despite being repeatedly washed with freshwater, the basin's chemical composition and water mineralization did not vary much (within 10 to 12‰), as indicated by the lack of appreciable changes in the molluscan fauna and the composition of other biotic assemblages. It seems probable that the through-flow phases were short-lived.

Yet, judging from the low $\delta^{18}O$ (10‰), the Khvalynian water was colder than that of the Caspian Sea–4° C in the north and 14° C in the south (Nikolaev 1995; Shkatova and Arslanov 2004). Khvalynian water might also have been turbid enough to affect sediment composition and produce smaller mollusc shells. High turbidity could have been due to the heavy solifluction and increased solid runoff from the drainage basin (Leonov *et al.* 2002).

4.2 The Neoeuxinian Sea

At the time of the Flood, the Pontic depression was occupied by the Neoeuxinian lake-sea, which was not higher than 80 to 100 m below sea level in its early stages. Floodwater discharge from the Caspian basin brought the level rapidly up to –50 or –40 m, increasing its area from 350,000 to 380,000 km^2. The flooded shelf area did not exceed 20–30 thousand km^2. Water volume at that time was as much as 545,000 km^3, which represents somewhat less than the Black Sea of today, but the origin of the water was quite different from that of today. It came uniquely from the Caspian Sea and river basins.

Paleontological data on molluscs and foraminifera suggest that they were Khvalynian fauna, though without *Didacna*, which would indicate a lower salinity (6 to 8‰). The slightly mineralized water of through-flow basins is usually termed semi-fresh water (Chepalyga 1984, 2002a; Kessel and Chepalyga 2002). Oxygen isotope composition (–10 to –11‰) suggests a low temperature (Nikolaev 1995; Shkatova and Arslanov 2004), and the lithological characteristics of the sediments (reddish-brown and pale yellow clays) indicate a high degree of oxygen saturation, which probably resulted from mixing by turbidity currents.

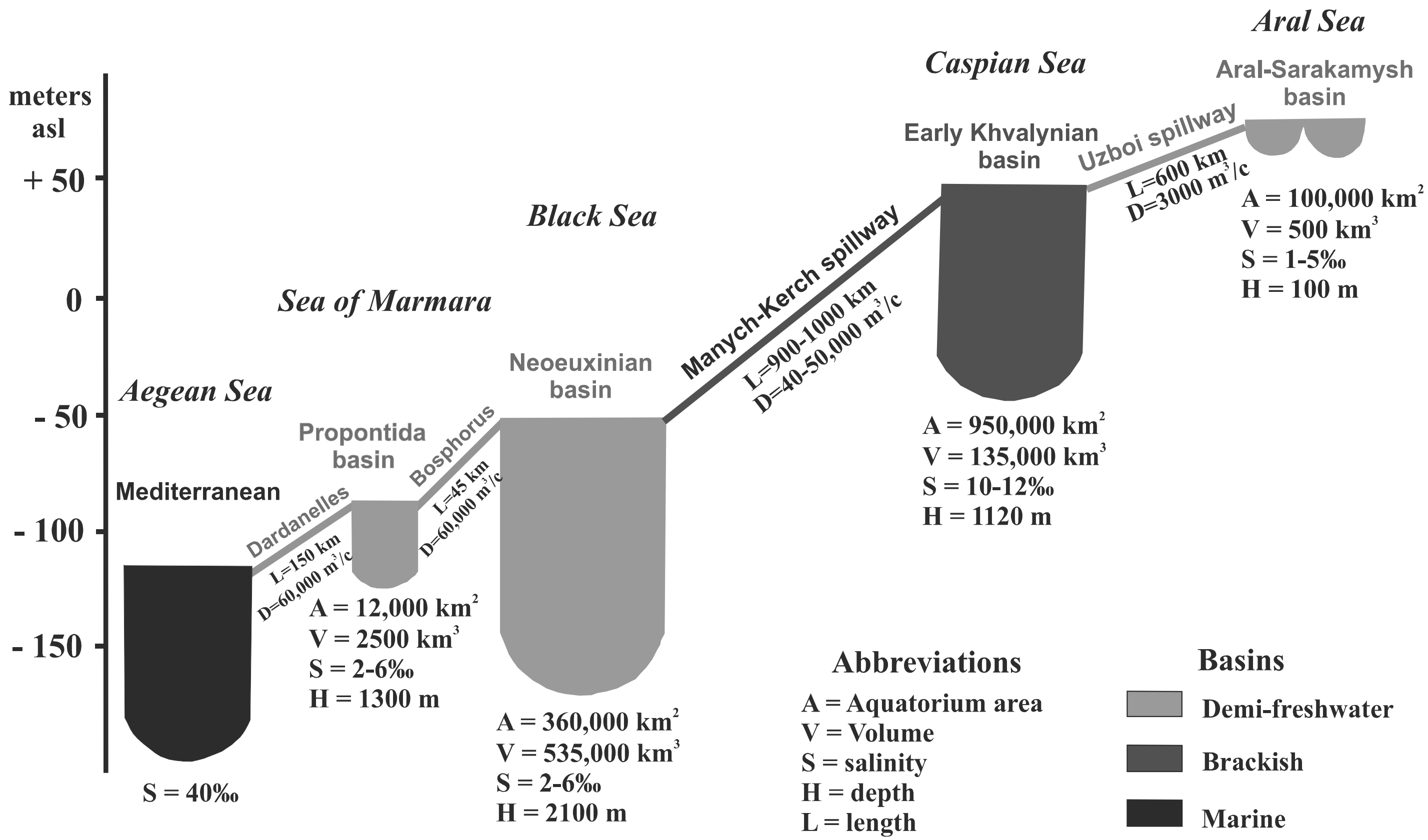


Figure 6. Cascade of Ponto-Caspian Great Flood basins (after Chepalyga 2002).

4.3 Propontida Basin

At the time of the flooding, a semi-fresh water basin of Caspian type existed in the Marmara Sea, as it received excess water from the Neoeuxinian basin together with mollusc fauna of *Dreissena rostriformis* and *D. polymorpha*. Its salinity also did not exceed 6 to 8‰, and the water flowing into it continued outward through the Dardanelles into the Mediterranean Sea.

4.4 Cascade of Eurasian Basins (the Vorukashah Sea)

The Great Flood created a system of interrelated basins in inner Eurasia that have been studied using various tracers, including lithology (reddish-brown interlayers of chocolate clay type), paleontology (Caspian endemic molluscs, foraminifera, and ostracoda) and isotopes (of oxygen and other elements). These markers have linked together the entire drainage system, providing evidence for a Cascade of Eurasian Basins, beginning in the Aral-Sarykamysh basin, then draining through the Uzboi Spillway into the Khvalynian (Caspian basin) Sea, from there through the Manych-Kerch Spillway into the Neoeuxinian (Pontic basin) Sea, and finally through the Bosphorus, the ancient Sea of Marmara, and the Dardanelles into the Mediterranean Sea (Figures 6 and 1).

Parameters of this superbasin were as follows:

Area	about 1.5 million km^2
Drainage basin area	more than 3 million km^2
Water volume	up to 700 thousand km^3
Water discharge	more than 60 thousand m^3/s
Salt resource	5000 km^3, or 10 billion tons
E-W extension	3000 km (from the Mediterranean to Central Asia)
N-S extension	2500 km (from 57° to 35° N. lat.)

The Eurasian cascade system of seas and lakes is unparalleled in water area. The largest intracontinental lake system of today–the Great Lakes of North America–ranks below it in all parameters; its area (245,000 km^2) is six times smaller, its water volume (22,700 km^3) is 30 times smaller, its discharge (14 thousand m^3/s) is over four times less, and its drainage basin area (1 million km^2) is about three times smaller.

The Eurasian Cascade would have been an impressive phenomenon to late Paleolithic humans and could have been reflected in old epic poems and mythology. In particular, a similar basin was described in the "Avesta" (the Zoroastrian Holy Scriptures) under the name of Vorukashah Sea.

5. DISCUSSION: WAS THE GREAT FLOOD A CATASTROPHE?

The rate of water-level rise may be inferred from the duration of the whole cycle, which is estimated at 500–600 years. Assuming all phases were equal in length, that the rising, highstand, and lowering were each about 150–200 years long, the sea level would have risen by 180–190 m at a rate of at least 1 m/year. Such a rate would have been 1000 times faster than the modern ocean rise (about 1 mm per year). The recent rise of the Caspian Sea amounted to 2.5 m since 1978, the rate thus approaching 10 cm/year, which is still 10 times slower than that calculated for the late glacial Caspian. Yet, the recent Caspian transgression had considerable adverse impact on human activities.

Therefore, the Khvalynian transgression (representing the main event of the Great Flood) must have been all the more catastrophic, especially when considering the rate of coastline shifts over the plains of the North Caspian region. The coastline migrated from the Atelian coast (near the Mangyshlak sill) northwards over 1000 km, which would produce a 5 to 10 km yearly advance. Such a rate would have been appreciable for coastal dwelling populations. The northward shift in the mouth of the Volga River would have proceeded even faster, as it shifted upstream by more than 2000 km within 150–200 years, i.e., more than 10 km/year or about 30 m/day.

Such a rate of change must have been beyond inconvenient, becoming very likely dangerous for local human populations. In addition, the coastline not only shifted but underwent substantial qualitative changes. First, the deltas of the large rivers–such as the Volga, Ural, Terek, Kura, and others–completely disappeared as the upstream shift of the river mouths transformed them into deep estuaries. The deltaic ecosystems were thereby wiped out, eliminating a highly productive and hospitable area of settlement. Their drowning may have seriously disrupted foraging patterns in the late glacial human economy. The marine transgression reduced considerably the usable land resources, and this loss was aggravated by superfloods in the river valleys and the inundation of interfluves by growing thermokarst lakes.

5.1 Sources of Water for the Flood

The provision of water for the Flood events must have involved some additional sources. Filling the Caspian basin to a level of +50 m would have taken as much as 70,000 km^3 of water, an amount equal to 200 years of present-day river discharge to the Caspian. Besides, some water flowed through the Manych Spillway–perhaps 250 to 1000 km^3/year–and some was lost through surface evaporation (possibly >100 km^3/year). The water for all these processes could have been supplied from various sources (excluding the Scandinavian ice sheet), namely: (1) superfloods in the river valleys, (2) permafrost melting, (3) higher runoff coefficient under conditions of permafrost, (4) increased catchment area (including the Central Asia area, which is closed today), and (5) lower surface evaporation due to winter ice cover.

The existence of a **superflood** was initially inferred from studies of Flood age macromeanders in river valleys (Sidorchuk *et al.* 2003), dimensions of which considerably exceeded modern ones. In addition, their widths tended to increase from north to south. They are similar to modern meanders in the tundra zone, but two or three times greater than those in the forest-tundra, three to five times larger than those in the taiga, five to eight times larger that those of the mixed forest zones, 10 times greater that those in the broadleaf zone, and 13 times larger than those in the forest-steppe and steppe (Sidorchuk *et al.* 2003) (Figure 7). No superflood effects have been noted within recent permafrost areas. Ancient annual runoff values calculated from the dimensions of the macromeanders are accordingly also well above those of modern runoff. They indicate a figure twice the modern value for the Volga River, three times the modern value for the Kama River, and almost four times more than the Don River (Table 4).

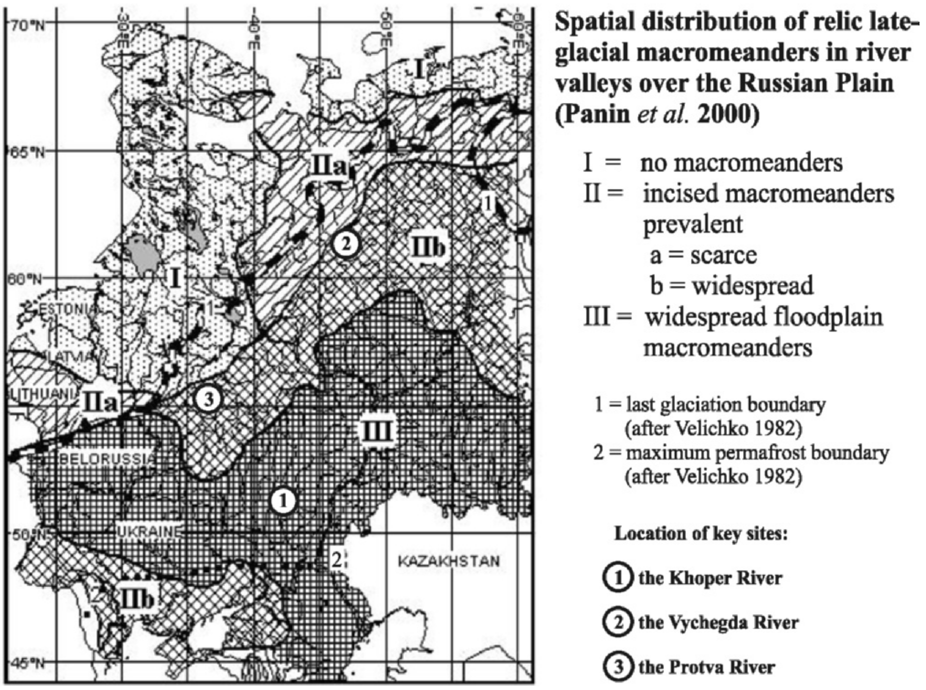


Figure 7. Spatial reconstruction of the superfloods.

Table 4. Annual water flow volume of the Late Glacial rivers in the Russian Plain and in West Siberia (Sidorchuk *et al.* 2003)

River Basin	Basin Area (10^3 km^2)	Annual Water Flow Volume (km^3)	
		Late Glacial	Modern
Severnaya Dvina	380	115	107
Mezen’	78	45	20
Pechora	322	220	126
Upper Volga	220	93	85
Oka	245	147	41
Kama	507	260	88
Middle/Lower Volga	249	85	40
Don	422	110	28
Pur	95	50	28
Taz	100	38	34

It was not only the amount of precipitation that accounted for the greater discharge and anomalous high runoff factor of the superfloods, but also its redistribution throughout the year. The cold late glacial climate favored large snow mass accumulation during the long winter. Rapid spring melting resulted in very high floods. This effect was enhanced by permafrost (due to poor permeability of frozen ground), and as a result, the runoff coefficient increased by a factor of 2 or 3 (that is, to 0.9–1.0). Depth of runoff was in excess of 800 mm in the upper and middle Volga drainage basin and that of the flood flow, more than 500 mm. The age of the macromeanders and the superfloods has been reliably determined by dozens of radiocarbon dates, which suggest a correspondence with the time of the Ponto-Caspian transgressions (16 to 10 ky BP). The largest macromeanders are dated to 16–14 ky BP, smallest ones to 12–10 ky BP, and those of medium size to 14–12 ky BP (Sidorchuk *et al.* 2003). So, the main source of water for the Flood events–the Khvalynian and Neoeuxinian transgressions–is the superfloods of the river valleys.

The next question, however, is from where the great water masses of the superfloods came. It was earlier thought that meltwater from the decaying Scandinavian ice sheet contributed considerably to the flood (Kvasov 1979). Recent reconstructions of the ice sheet (Velichko 2002; Leonov *et al.* 2002) show, by contrast, that at the time of the Flood (16–15 ky BP), the ice margin had receded from the Volga drainage basin, and meltwater could not reach the Caspian Sea. Besides, the most distinct superfloods were located in the valleys of small rivers (in the Don and Seim drainage basins), which had no connection with the Scandinavian glaciers. It seemed logical to look for additional water sources in higher rainfall, yet most of the existing reconstructions clearly testified to considerable aridity in the periglacial zone (Velichko 1973; Gerasimov and Velichko 1982), with annual precipitation several times less than that of the present (100–150 mm/yr). Recently, new data (Sidorchuk *et al.*

2001a, b; Sidorchuk *et al.* 2003) indicate a much higher precipitation in Eastern Europe (up to 600–800 mm per year). A considerable amount of water could be coming into rivers also due to a threefold increase in the runoff coefficient under conditions of permafrost (up to 0.9–1.0). Finally, duration of the superfloods could be much shorter, with their discharge increased accordingly. Total runoff could have been as high as 800 mm per year, and with a catchment area of 2 million km^2, the total water volume delivered yearly to the Caspian basin would have amounted to 1000–1500 km^3. That would be enough for the Caspian Sea level to rise to +25 m, and even to +50 m, though insufficient to create 1000 km^3 of excess water to outflow through the Manych-Kerch Spillway.

Further research led us to consider interfluves, including their gentle slopes and terrace plains. All those surfaces bear distinct relict cryogenic microsculpture: polygonal networks (pseudomorphs on ice wedges) and small, flat-bottomed depressions, 50–60 m in diameter and 1.5–2 m deep. Judging from the interpretation of aerial photographs (Velichko *et al.* 1996) and calculations (Porozhnyakova nd), such depressions occupied as much as 50% of interfluvial terrain. Excavations into their bottom exposed thin (20–30 cm) layers of sand showing the wave-like and horizontal bedding typical of shallow thermokarst lakes in the modern cryolithozone (alasses of Yakutia). In Eastern Europe, such depressions are referred to as paleo-alasses and micro-alasses (Porozhnyakova 1997, nd). The depressions were initiated by the thawing of permafrost and could have supplied water for the superfloods. The question remains, however, how much water would have been released by permafrost thawing and whether this amount would have been sufficient to supply the Flood processes. The calculations of O.M. Porozhnyakova (nd) indicate that total ice content (in formerly frozen ground) amounted to 27% and did not exceed 40%, including macro-ice (in ice wedges) estimated at 20% and segregation ice estimated at 7%. If a permafrost layer 1 m thick thawed, it could yield as much as 250–400 mm of runoff. On the whole, permafrost thawing over the entire Caspian catchment (~2,000,000 km^2) could supply about 1000 km^3 of runoff.

6. ARCHAEOLOGY OF THE FLOOD

The Flood could have had considerable impact on humans from the rise of sea level and the flooding of vast areas, including fertile lands, river deltas, and floodplains. These transformations could have stimulated a mass exodus from the flooded areas and perhaps the appearance of new ethnic communities. Within river valleys, settlements tended to move upslope. Thus, in the Seim River valley, the pre-Flood Late Paleolithic site Avdeevo (dated to 20–18 ky BP) was at the very edge of the water, while younger sites were located much higher, possibly because of the superfloods (Leonova 1998). On the other hand, basins of the Vorukashah system formed water barriers several thousands of kilometers long and a few hundreds of kilometers wide, which could have hampered contacts and exchanges between peoples. Even the relatively narrow Manych-Kerch spillway prevented cultural exchange. This can be seen in the sequence of Paleolithic cultures at the Kamennaya Balka site on the northern coast of the spillway (Leonova 1998). Of the three occupation layers, the lower and upper ones, dated to 20–17 and 13–12 ky BP, respectively, contain implements typical of the Caucasian (Imeretian culture) and Near Eastern (Shanidar) types, with a prevalence of microliths, a discovery that suggests close connections with regions to the south. The middle layer, which is synchronous with the Flood peak (17–14 ky BP), yielded mostly autochthonous tools of local type, without microliths. Such a difference may be attributed to the fact that the site was isolated from the Caucasus by the Manych-Kerch Spillway (Figure 8).

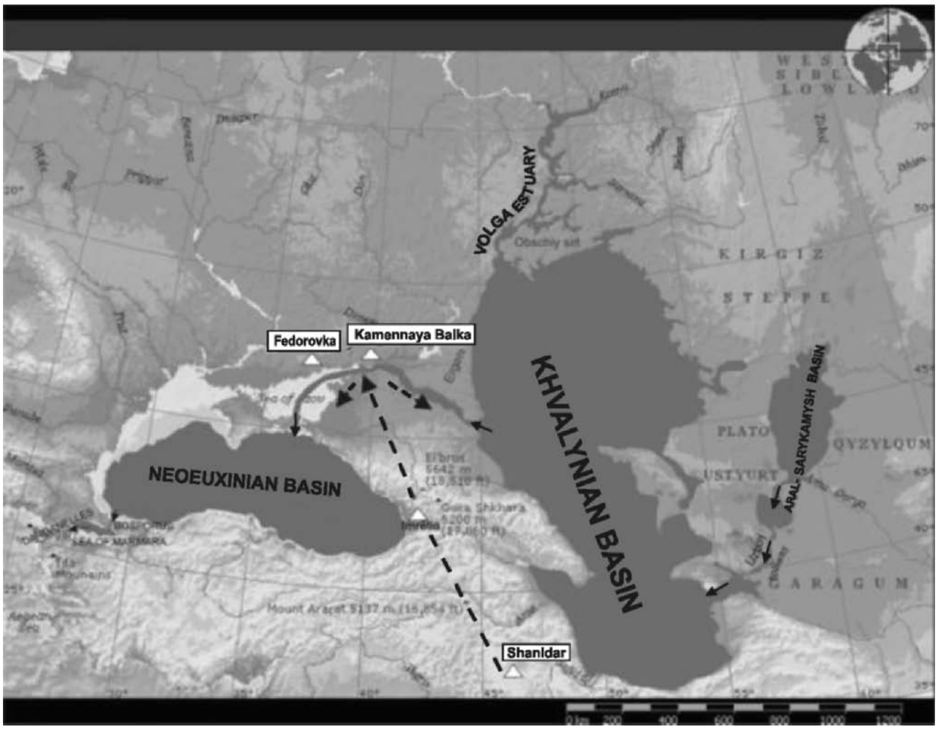
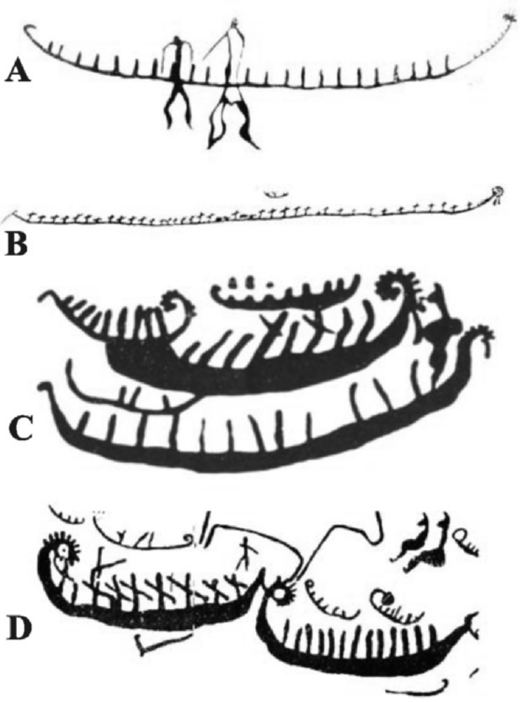


Figure 8. Second stage of Kamennaya Balka cultural evolution: late Paleolithic migrations from Caucasus and the Near East blocked by the Manych-Kerch Spillway.

The dramatic reduction in fertile land area together with highly dynamic environments on the river banks and the Vorukashah Sea coasts could have

Figure 9. Origin of sailing: the world's oldest rock art depicting Mesolithic boats in Gobustan.



given impetus to the development of a producing economy and the appearance of ancient civilizations. In fact, the oldest boat images dated to 8–9 ky BP were found in Gobustan, on the Caspian coast, south of the Kura River Delta (Dzhafarzade 1973). These rock depictions show flat-bottomed boats (Figure 9A, B) and keel-built vessels (Figure 9C, D) suitable for marine navigation, some with as many as 37 oarsmen (Figure 9B). The earliest ships appeared in the Caspian region immediately after the Flood, which may be interpreted as a consequence of this event. Other evidence for the appearance of a productive economy as a result of the Flood is that horses were domesticated there earlier than anywhere else in the world (Matyushin 1976).

The Ponto-Caspian Flood is not the Flood described in the Bible. The relation may be only indirect. The collective memory of mankind may have retained these events for thousands of years, until they were later written in ancient Aryan scriptures, such as the Rigveda and Avesta, and the concept of the Flood was adopted by the ancient inhabitants of Mesopotamia, from whom it eventually came to the Bible.

7. CONCLUSION

(1) A period of a Great Flood has been recognized in the Ponto-Caspian basin of the Late Pleistocene (17 to 10 ky BP). The flooding events affected wide areas in Eurasia and left their imprint on coastal plains (in the form of marine transgressions), on river valleys (in the form of superfloods), and in interfluvial areas (in the form of thermokarst lakes).

(2) The most conspicuous floods centered on the Khvalynian basin of the Caspian Sea. Several waves of flooding occurred, during which sea level

fluctuated over a range of more than 100 m, and the 1500–2000-year duration of each wave is tentatively correlated with Shnitnikov's cycles. Smaller oscillations (10 known to date) were 500–600 years long, and these reveal a sea-level fluctuation range of 50–100 m.

(3) The Great Flood itself was associated with a rapid rise of the Caspian level (almost 200 m over the course of ~100 years). It was followed by a number of cyclical fluctuations with progressively lower peaks. At the Flood's maximum (16–15 ky BP), a Cascade of Eurasian Basins developed that was more than 3000 km long and 1.5 million km^2 in area. The submerged area totaled ~1 million km^2.

(4) The Flood was fed from a few sources, namely: Scandinavian ice sheet melting (only in the initial stage); superfloods in the river valleys; permafrost melting; higher runoff coefficient under permafrost conditions; increased catchment area to include Central Asia; and lower evaporation from the water surface (due to winter ice cover).

(5) Drastic changes in sea level (up to 2 m per year) and associated coastline migration (as great as 10–20 km per year) resulted in expansive flooding, so that fertile lands were lost to submergence. This, in turn, must have brought about major stresses and migrations of people, leading to population density increase and perhaps stimulus toward the development of a more advanced economy.

(6) The Flood appears to have been of greater importance for ancient humans than the Last Glacial Maximum (LGM). It did not destroy cultures, but on the contrary, the cyclical and progressive environmental changes may have promoted the appearance of the first signs of a productive economy, such as shipping or domestication of horses.

(7) The events of the Eurasian flooding were perhaps kept in memory by ancient Proto-Aryans and were written in their ancient scriptures, as well as those of the ancient inhabitants of Mesopotamia, from whom the Flood narrative came to the Bible.

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7

CONTROVERSY OVER NOAH'S FLOOD IN THE BLACK SEA: GEOLOGICAL AND FORAMINIFERAL EVIDENCE FROM THE SHELF

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Abstract: This paper reviews the geological and foraminiferal evidence collected during the course of extensive paleoceanographic and geological studies in the Black Sea, conducted largely by Eastern European scientists since 1970. Though this research has a long history, its most recent focus has been examining the "Noah's Flood" Hypothesis proposed by William Ryan and Walter Pitman, which proposes an abrupt and catastrophic flooding of the Pontic basin in the early Holocene. Specifically, the hypothesis states that the Black Sea was a freshwater lake with a surface about 140 m below present sea level between 14.7 and 10 ky BP, while at 7.2 ky BP (initial hypothesis) or 8.4 ky BP (modified hypothesis), the lake was rapidly inundated by Mediterranean water flowing through the Bosphorus, which forced the dispersion of early Neolithic people into the interior of Europe. The hypothesis further suggests that the event formed the historical basis for the biblical legend of Noah's Flood. This paper considers the period between 28 and 7 ky BP, and three crucial points are discussed: (1) the level and salinity of the Neoeuxinian lake; (2) the re-colonization of the Black Sea by Mediterranean immigrants–and by implication sea level and salinity changes due to connection/isolation between adjacent basins; and (3) an alternative to the Bosphorus as an inter-basin conduit. It will be shown that, prior to the moderately warm Würm Paudorf (Middle Weichselian) Pleniglacial (prior to *ca.* 27 ky BP), a brackish Tarkhankutian basin was connected with the Sea of Marmara. At the Last Glacial Maximum (LGM), this connection was interrupted, and the level of the Tarkhankutian basin dropped to about –100 m, transforming this basin into a closed Early Neoeuxinian lake. In the warming climate of *ca.* 17 ky BP, a massive water discharge originating most likely from the Caspian Sea and arriving via the Manych Spillway increased the level of the Late Neoeuxinian lake to about –20 m. Excess semi-fresh to brackish water must have spilled into the Sea of Marmara and from there into the Mediterranean. During the short climatic cooling episode of the Younger Dryas, the level of the lake dropped from –20 to –50 m and then rose again to about –20 m. After *ca.* 10

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ky BP, the level of the Black Sea never again dropped below the –40 m isobath, nor exhibited a maximum amplitude of fluctuation greater than approximately 20 m. At *ca.* 9.5 ky BP, the water level reached –20 m again and Mediterranean water with its inhabiting organisms entered the Late Neoeuxinian lake. This re-colonization of the Black Sea occurred in an oscillating manner. It was slow at the beginning, becoming most prominent by 7.0 ky BP. The connection between adjacent basins was probably not through the Bosphorus Strait but via an alternative route, e.g., that following Izmit Bay–Sapanca Lake–Sakarya River. On average, sea level rose gradually, but in an oscillating manner, to its present level, and perhaps slightly higher, averaging 3 cm per 100 years but certainly not 15 cm per day (almost 55 m per year) as postulated by the "Noah's Flood" hypothesis. A rate of sea-level increase of 3 cm per 100 years would not be noticed by local inhabitants and would not have accelerated their dispersion into the interior of Europe. This brings us to the conclusion that "Noah's Flood" in the Black Sea is a contemporary legend.

Key words: Black Sea, Late Pleistocene, Holocene, sea level, flood, salinity, benthic foraminifera, re-colonization

1. INTRODUCTION

Whether it is referred to as Noah's Flood or the Great Flood, this disastrous event is so deeply rooted in the collective memory of humankind that it is reported in the Epic of Gilgamesh (Keller 1981) and other similarly ancient texts–e.g., the Epic of Atrahasis (Lambert and Millard 1969) and the Epic of Ziusudra (Best 1999)–and it is reflected in several world religions. It comes as no surprise that scientists have a strong interest in the historical reality, if any, behind the story of Noah's Flood.

Although the story likely originated in Mesopotamia, and the Epic of Gilgamesh was recorded at the northwestern end of the Persian Gulf (King 1918; Magnusson 1977; Keller 1981), Ryan *et al.* (1997) locate the Great Flood in the region of the Black Sea, quite far from Mesopotamia, on the other side of a large mountain chain. Based upon 350 km of high-resolution seismic profiles, a few short sediment cores obtained at water depths of –49 to –140 m within a fairly restricted area of the Black Sea's northwestern shelf, and ^{14}C dates on *Dreissena* shells (all radiocarbon ages in this paper are uncorrected), they concluded that the Black Sea was a freshwater Neoeuxinian lake with a level 140 m below present between 14.7 and 10.0 ky BP. According to their Flood Hypothesis, in the course of the post-glacial transgression, at 7.2 ky BP (dates based on *Mytilus galloprovincialis*), saltwater broke through a barrier within the narrow Bosphorus Strait and funneled through this channel at a speed of 50 mph, hitting the Black Sea at 200 times the force of Niagara Falls, thereby rapidly refilling the lake and increasing its salinity. A single, structureless and uniform layer of jelly-like sapropel was formed, draping the undulating surface of the Neo-euxinian unconformity. At a rate of 15 cm per day, the sea level rose 100 m within two years, catastrophically submerging more than 100,000 km^2 of exposed shelf and flooding coastal farms. This catastrophe accelerated the dispersion of early Neolithic foragers and farmers into the interior of Europe, forming the historical basis for the biblical story of Noah's Flood (Ryan and Pitman 1998).

This Flood Hypothesis spurred tremendous interest by the public, the scientific community, and the media–e.g., BBC (1996); *New Scientist* (Mestel 1997; Hecht 2003); *New York Times* (Wilford 1999, 2001); *Scientific American* (Morrison and Morrison 1999); *Washington Post* (Gugliotta 1999, 2000); *Der Spiegel* (2000); *National Geographic* (Ballard 2001); *GSA Today* (Aksu *et al.* 2002a); *Frankfurter Allgemeine Zeitung* (2003)–encouraging a new wave of research in the Black Sea-Mediterranean Corridor.

In support of the Flood Hypothesis, Ballard *et al.* (2000), Lericolais (2001, 2003, 2004), Lericolais *et al.* (this volume), and Algan *et al.* (2003, this volume) described a submerged coastline with wave-cut terraces, coastal dunes, and drowned beaches, enriched with *Dreissena* at various depths, and ranging from –90 m at the Romanian shelf to –155 m at the Turkish shelf near Sinop. This coastline was overlapped by a uniform drape of mud containing *Mytilus galloprovincialis*. The age of the *Dreissena* samples ranges between 24.2 and 7.9 BP (Major 2002), in particular, 11.8–7.9 ky BP (Algan *et al.* this volume), 10.2–8.6 ky BP (Lericolais *et al.*, this volume), and 15.5–7.4 ky BP (Ballard *et al.* 2000). The age of *M. galloprovincialis* varies between 7.8 and 4.0 ky BP, in particular, 7.8–6.6 ky BP (Lericolais *et al.*, this volume), 7.5–4.0 ky BP (Ballard *et al.* 2000), and 7.4–5.9 ky BP (Algan *et al.*, this volume).

Görür *et al.* (2001) posed a contradiction to the Flood Hypothesis. After studying the coastal plain and offshore sedimentary successions on the southern Black Sea coast around the mouth of the Sakarya River, about 130 km east of the Bosphorus, they suggested that the water level of the lake rose gradually from some time prior to 8.0 ky BP to 7.2 ky BP, when it attained a surface level of –18 m, and the most recent influx of Mediterranean water began. Further evidence for a higher level within the Black Sea in the early Holocene has been proposed by Aksu *et al.* (1999, 2002a, b), who suggested that the Black Sea was higher than the Sea of Marmara and has been flowing out into the world ocean unabated since 10.5 ky BP. In so doing, it prevented the establishment of a two-way flow in the Bosphorus Strait and delayed the salination of the Black Sea and the immigration of Mediterranean organisms northward until 8.5 ky BP (the Outflow Hypothesis of Aksu *et al.* 2002a, b). This interpretation is based on physical sedimentological evidence from bedform asymmetry and, directly south of the Bosphorus exit, a climbing mid-shelf delta in the Sea of Marmara that was formed by the Bosphorus outflow (Aksu *et al.* 1999; Aksu *et al.* 2002a; Hiscott *et al.* 2002). Similarly, Çağatay *et al.* (2000) proposed that the formation of a sapropel layer in the deep Marmara Sea during the period 10.6–6.4 ky BP reflects water column stratification and seafloor anoxia, which they attribute to prolonged freshwater influx from a Black Sea whose surface at that time must have been at or above the Bosphorus sill depth of –35 m.

To address these arguments, the Flood Hypothesis was eventually modified. The initially formulated lowstand of –140 m at 14.7–10.0 ky BP and the abrupt drowning of the Black Sea shelf at 7.2 ky BP were replaced by two lowstands and two floods (Ryan *et al.* 2003). The first lowstand at –120 m occurred between 13.4 and 11 ky BP and was followed by the first flooding event, which increased the level of the Neoeuxinian lake from –120 m to –30 m at 11.0–10.0 ky BP (i.e., the Younger Dryas), when the surface of the lake rose well above that of the contemporaneous global ocean (Siddall *et al.* 2003). The level of the Neoeuxinian lake during the Younger Dryas was explained by decoupling the lake from the world ocean by a shallow (less than 30 m) sill in the Bosphorus under moist climatic conditions. The second lowstand (–95 m) of the Neoeuxinian lake occurred at 10–8.4 ky BP due to evaporation of the isolated lake under arid conditions. At 8.4 ky BP, Mediterranean water topped the Bosphorus sill and flooded the Neoeuxinian lake. Simultaneously with the world sea-level rise, the second flood raised the surface level from –95 to –30 m and replaced the relict Caspian biota with Mediterranean organisms. The mid-shelf climbing delta at the southern Bosphorus exit was attributed to a small stream, the Kurbağalıdere River (Ryan *et al.* 2004, this volume).

Hiscott *et al.* (this volume) criticized this modified Flood Hypothesis, arguing that modern sediment discharge from the Kurbağalıdere River is so small that it would take about 100,000 years to construct the delta, a time span clearly at odds with the duration of delta progradation (about 1000 years from *ca.* 10–9 ky BP). They insist that the Black Sea has been at or above the Bosphorus sill depth and flowing into the world ocean uninterrupted since 10.5 ky BP, making flooding of the basin impossible.

As an alternative to both hypotheses, Kerey *et al.* (2004), and Yanko-Hombach *et al.* (2004) argue that the Bosphorus Strait is too young to have played a cataclysmic role in water exchange before *ca.* 5.5 ky BP, and an alternative route between the basins must be considered.

The western scientists have based their hypotheses largely on material obtained outside the Black Sea (e.g., Aksu *et al.* 2002a, b) or on limited data from the outer shelf (e.g., Ryan *et al.* 1997). The abundant scientific data recovered directly from the Black Sea by USSR and Former Eastern Bloc scientists (among them, Andrusov 1918; Arkhangel'sky and Strakhov 1938; Nevesskaya and Nevessky 1961; Nevesskaya 1963, 1965; Il'ina 1966; Nevessky 1967; Semenenko and Kovalyukh 1973; Tsereteli 1975; Shilik 1977; Fedorov

1978; Shcherbakov *et al.* 1978; Dimitrov *et al.* 1979; Komarov *et al.* 1979; Malovitsky *et al.* 1979; Kuprin *et al.* 1980; Balabanov *et al.* 1981; Shnyukov 1981, 1982; Popov 1983; Shcherbakov 1983; Shnyukov 1983, 1984a, b; Kuprin *et al.* 1985; Shnyukov 1985, 1987; Gozhik *et al.* 1987; Yanko and Troitskaya 1987; Balabanov and Izmailov 1988; Panin 1989; Yanko 1989, 1990a; Yanko and Gramova 1990; Khrischev and Georgiev 1991; Gorshkov *et al.* 1993; Glebov *et al.* 1996; Mel'nik 1997; Shilik 1997; Stanko 1997; Kuprin 2002) have unfortunately been largely ignored in the global scientific debate, apparently due to language barriers and the lack of west-east scientific dialogue.

By 1997, a large (1:500,000 to 1:10,000) marine geological survey of the Black Sea shelf had been nearly completed (e.g., Kuprin *et al.* 1980; Balabanov *et al.* 1981; Panin 1983; Shnyukov 1981, 1982, 1983, 1984a, b, 1985; Esin *et al.* 1985; Shnyukov 1987; Dmitrienko *et al.* 1988; Yanko-Hombach 2003). As part of these projects, thousands of cores and tens of thousands of kilometers of high-resolution seismic profiles across the Black Sea shelf from the northern exit of the Bosphorus Strait on the west to the city of Batumi on the east (Figure 1) had been collected and studied in a multi-disciplinary effort.

A methodology for the Black Sea shelf investigation had been developed (Shnyukov 1982), and the paleoclimatic (e.g., Komarov *et al.* 1979), tectonic (e.g., Shnyukov 1985), and sedimentary (e.g., Fedorov 1978; Kuprin *et al.* 1980) history of the basin had been investigated. A high-resolution Quaternary biostratigraphy based upon molluscs (Nevesskaya 1965; Fedorov 1978) and foraminifera (Yanko 1989, 1990; Yanko and Gramova 1990), all supported by hundreds of radiocarbon assays (Appendices 1 and 2, this volume), had been established, and sea-level dynamics had been reconstructed (e.g., Tsereteli 1975; Shilik 1977; Balabanov *et al.* 1981; Shilik 1997; Balabanov, this volume; Chepalyga, this volume).

This paper focuses on the reconstruction of sea level and salinity in the Black Sea since the Last Glacial Maximum (LGM) using benthic foraminifera as the main tool. These organisms, ubiquitous in marine environments, are well known as reliable paleoenvironmental indicators. Their tremendous taxonomic diversity allows for a wide range of biological reactions to varied environmental factors, including many species-specific responses to ecological conditions (Fursenko 1978), which adds to their potential as index species for monitoring sea-level and salinity changes. They have very short reproductive cycles–six months to one year (Boltovskoy 1964)–and rapid growth (Walton 1964), making even their community structure particularly responsive to environmental change. Their tests are readily preserved and can record evidence of environmental variability through time, thus providing historical baseline data even in the absence of background studies. They are small and abundant compared to other larger, hard-shelled taxa (such as molluscs), which makes them particularly easy to recover in statistically significant numbers (Yanko *et al.* 1999a).

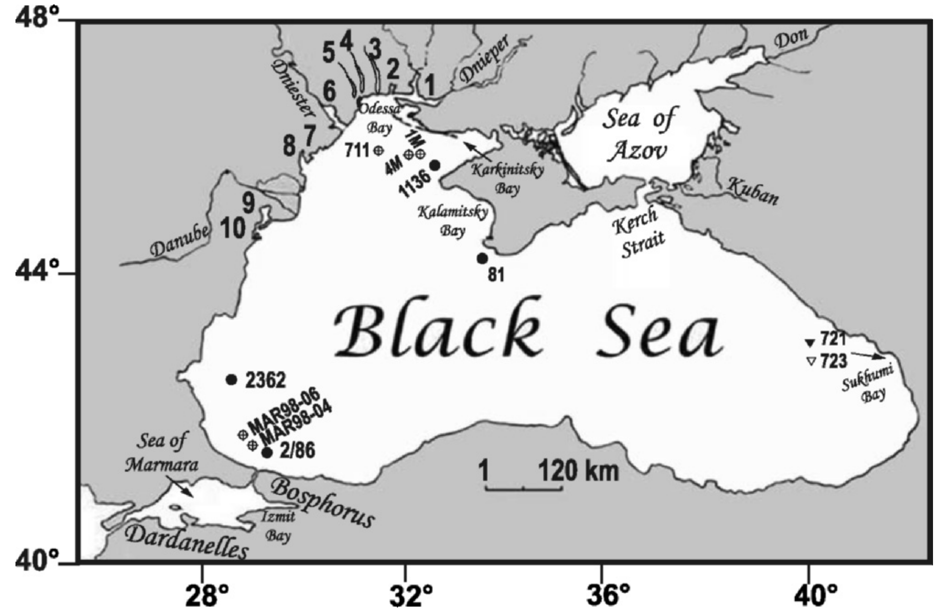

Figure 1. Map showing location of sample cores in the Black Sea. **Key:** Solid circles = gravity cores; solid triangles = boreholes; cross within circles and hollow triangles = cores and boreholes, respectively, dated by ^{14}C from the literature: Core 4M and 1M (Balandin and Mel'nik 1987); Core 723 (Kvavadze and Dzeiranshvili 1989); Core 711 (Gozhik *et al.* 1987); Cores MAR98-04 and MAR98-06 (Aksu *et al.* 2002b); 1 to 10 = limans/lagoons: 1– Dnieper-Bugsky, 2 – Berezansky, 3 – Tiligulsky, 4 – Kuyalnitsky, 5 – Khadzhibeysky, 6 – Dniestrovsky, 7 – Alibey, 8 – Sasyk, 9 – Razelm, 10 – Golovitsa, Sinoe, Nuntash.

In examining the flood hypotheses, this paper emphasizes the interval of 28–7 ky BP, and three crucial points are discussed: (1) level and salinity of the Neoeuxinian lake; (2) re-colonization of the Black Sea by Mediterranean immigrants, and, by implication, sea-level and salinity changes due to connection/isolation between adjacent basins; and (3) an "alternative" to the Bosphorus connection between adjacent basins.

It will be shown that the increase in the level of the Black Sea was not catastrophic, nor was it gradual. It occurred in an oscillating manner from its lowest point at about –100 m during the LGM (27–18 ky BP) to –20 m at *ca.* 10 ky BP. After this date, it never again dropped below the –40 m isobath, nor did it exhibit a maximum amplitude of fluctuation greater than about 20 m. During the last 10,000 years, sea level rose in a steady, oscillating manner to its present level, averaging only 3 cm/100 years, not enough to force human groups into the interior of Europe between 8.4 and 7.2 ky BP. The re-colonization of the Black Sea by Mediterranean immigrants began at 9.5 ky BP, and by 7.2 ky BP, the shelf was completely re-colonized. This evidence contradicts the "Noah's Flood" Hypothesis, suggesting that it is a contemporary legend.

2. MATERIALS AND METHODS

2.1 Data Collection

The data were collected since 1971 over the course of a large-scale geological survey of the Black Sea shelf (e.g., Malovitsky *et al.* 1979; Yanko 1979; Dmitrienko *et al.*1988; Yanko 1989, 1990a; Yanko and Gramova 1990). Materials from the Eastern Mediterranean (Basso *et al.*1994; Yanko *et al.*1998; Basso and Spezzaferri 2000; Koral *et al.* 2001), the Sea of Marmara (Yanko *et al.* 1998), and the Caspian Sea (Yanko 1989, 1990a) were used as supporting evidence for the origin of the Black Sea foraminifera.

In the Black Sea, approximately 30,000 samples from 1325 grabs, 4000 gravity/piston cores, and 56 boreholes were investigated. The cores were obtained in limans, lagoons, river deltas, the Kerch Strait, the Sea of Azov, and across the shelf. In the Caspian Sea, the Sea of Marmara, and the Eastern Mediterranean, 302, 73, and 512 sediment samples, respectively, were studied. The location maps of studied materials are provided in Yanko (1989, 1990a), Basso *et al.* (1994), and Yanko *et al.* (1998).

Locations of six exemplary cores discussed in this paper appear in Figure 1. They were chosen because they represent the larger population of cores particularly well. The working half of the sediment column was examined at 2 cm intervals within the gravity cores (max. length 5 m), and in the uppermost 2 cm of each 10 cm interval within boreholes (max. length 28 m).

2.2 Grain Size Analysis

Grain size analysis of the sediments was performed at Odessa State University in Ukraine, Tel Aviv University in Israel, and Istanbul University in Turkey. Sediments were divided into clay (<0.0039 mm), silt (0.0039–0.0625 mm), sand (0.0625–2.0 mm), and gravel (>2 mm) fractions, and the methods used included wet sieving for sediments with a grain size greater than 63 µm, and pipette analyses for clay and silt fractions as described in Folk (1974).

2.3 Radiocarbon Dating

Conventional radiocarbon dating of peat, wood, and mollusc shells was performed at various USSR (Appendices 1 and 2, this volume) and foreign (Appendix 2, this volume) laboratories. Throughout this text, ^{14}C data are expressed as uncorrected years BP in order to remain comparable with world-wide Pleistocene/Holocene chronology and sea-level curves (e.g., Fairbanks 1989).

2.4 Foraminifera

Live (Rose Bengal stained) and fossil foraminifera were investigated separately as described in Yanko *et al.*(1998) and Yanko and Troitskaya (1987), respectively. Samples were soaked and washed in distilled water, and passed through 63 μm mesh sieve. Live foraminifera were counted in wet samples equivalent to 50 g (Black Sea) and 5 g (Eastern Mediterranean) of dry sediment mass. Fossil foraminifera were studied in samples that were dried at room temperature to avoid destruction of agglutinated species. Dried samples were split with a microsplitter to avoid sample bias; about 300 fossil foraminifera were picked by hand (flotation in CCl_4 was sometimes used) and counted for population statistics. The total number of foraminifera was calculated in dry samples of 50 g (Black Sea) and 5 g (Sea of Marmara, Eastern Mediterranean).

All species were morphologically examined, taxonomically identified, and SEM pictures were taken. In our taxonomic work, we followed the suprageneric classification in *Osnovi Paleontologii* (Rauzer-Chernousova and Fursenko 1959), in combination with the generic classification of Loeblich and Tappan (1987). All identified taxa were systemized as belonging to Protozoa (Class Sarcodina, Subclass Foraminifera). Direct comparison with the original collections of d'Orbigny, Schlumberger, and Le Calvez in the Museum of Natural History, Paris, was used for most of the species.

The collection of Black Sea, Caspian Sea, and Sea of Azov foraminifera (155 species) is stored in the Paleontological Museum of Odessa National University, Ukraine. The original collection of foraminifera (~ 500 species) from the Eastern Mediterranean and Sea of Marmara is stored at the Avalon Institute of Applied Science. A partial (Israeli shelf) duplicate of the collection is kept at University College London, UK, and at the Museum of Natural History, Paris.

The foraminifera were divided into dominant (<50% of a given population) and accessory species. Species that occur at 50% of all studied locations are considered to be widely distributed, 49–10% are considered frequent, 9–1% rare, and <1% trace. According to their ecological preferences (Table 1), foraminifera are divided into oligohaline (1–5‰), strictoeuryhaline (11–26‰), polyhaline (18–26‰), euryhaline (1–26‰), shallow (0–30 m), relatively deep (31–70 m), and deep (71–220 m) species (Yanko and Troitskaya 1987; Yanko 1989, 1990a).

2.5 Ecostratigraphic Techniques

Ecostratigraphy is the biostratigraphic application of ecological and paleoecological principles to develop an understanding of the global external forcing agents that drive ecological change. The ecostratigraphy of the Black

Sea addresses biotic responses to isolation from and connection to the neighboring Sea of Marmara and Caspian Sea, and to related sea-level changes and salinity oscillations.

Our ecostratigraphic technique is based largely on alternation of foraminiferal assemblages and their ecological characteristics in geological sections, supported by ^{14}C and palynological assays. An increase in the number of Mediterranean immigrants, especially strictoeuryhaline and polyhaline species, in sediment sequences indicates an increase of Mediterranean influence and salinity, and *vice versa*. The complete replacement of Mediterranean immigrants by oligohaline Caspian species shows separation between the Black Sea and Mediterranean, followed by desalination of the Black Sea. This conclusion is based on a generally accepted observation, fully supported by our ecological study (Yanko 1989, 1990a, b), that foraminifera are not well adapted to freshwater environments (Sen Gupta 1999). The classification of Tchepalyga [also spelled as Chepalyga] (1984) is used to describe paleobasin salinity: fresh <0.5‰, semi-fresh 0.5–5‰, brackish 5–12‰, semi-marine 12–30‰, and marine 30–40‰.

3. RESULTS AND INTERPRETATION: ECOSTRATIGRAPHY AND PALEOENVIRONMENTAL RECONSTRUCTIONS

3.1 Live and Fossil Foraminifera

Planktonic foraminifera do not live in the Ponto-Caspian basins–the Black Sea, the Sea of Azov, and the Caspian Sea (Yanko 1989, 1990a; Yanko and Vorob'eva 1990, 1991)–while they abound in the neighboring Sea of Marmara (Alavi 1988; Kaminski *et al.* 2002) and the Mediterranean Sea (Cimerman and Langer 1991).

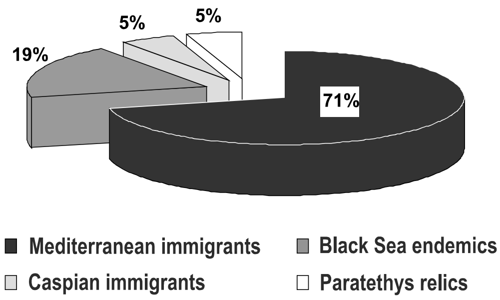

Figure 2. Origin of the Black Sea foraminifera.

Benthic foraminifera live on the shelf to a maximum depth of 220 m in the Black Sea and 70 m in the Caspian Sea. In the Black Sea, they are represented by 101 species: 19 Black Sea endemics, 5 Paratethys relics, 5 Caspian, and 72 Mediterranean immigrants (Figure 2). In the shallow (maximum depth 13 m) Sea of Azov, they are represented by 24 Black Sea immigrants. In the Caspian Sea, only 27 (mostly endemic) species were found (Yanko 1989, 1990a).

Taxonomic diversity in Black Sea benthic foraminifera is low compared to that of the Eastern and Western Mediterranean, where 250 and 400 species have been identified, respectively (Cimerman and Langer 1991; Sgarrella and Moncharmont Zei 1993; AVICENNE Annual Report 1995, 1996; Basso and Spezzaferri 2000). The Black Sea average salinity (17‰) is only half that of the Eastern (39‰) and Western (34‰) Mediterranean, and the maximum salinity of the Sea of Azov and Caspian Sea is almost the same, about 13‰. However, the Caspian Sea has a continental type of salinity with a dominance of Ca^{2+} and SO_4^{2-} ions (Bruevich 1952).

In the Black Sea, the number of species and their abundance decreases progressively with decreasing salinity (Figure 3, Table 1), and no live foraminifera exist in salinities below 1‰.

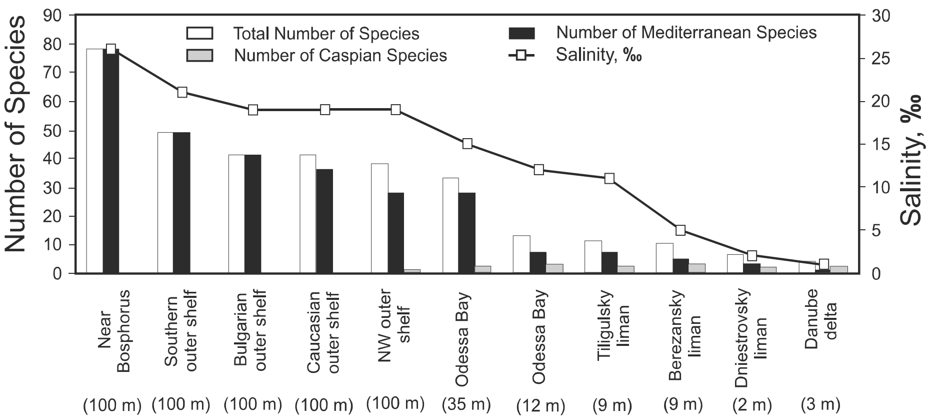

Figure 3. Decrease in the number of foraminiferal species with decreasing salinity in various areas of the Black Sea (average depths of sampling areas shown in parentheses).

Black Sea foraminifera are dominated by 10 species of *Ammonia* (Yanko 1989, 1990a, b). The Black Sea endemic *A. novoeuxinica* together with Caspian endemics *Mayerella brotskajae* and *Elphidium caspicum caspicum* inhabit river deltas. Together with other Caspian endemics *A. caspica* and *Porosononion martkobi tschaudicus*, they indicate a semi-fresh regime for the Black Sea. The euryhaline Mediterranean species *A. tepida* lives everywhere on the shelf, while the polyhaline *A. compacta* and *A. ammoniformis* dominate on the outer shelf where salinity is >18‰. Together with the other Mediterranean species, *A. beccarii*, *A. parkinsoniana*, and *A. agoiensis*, they indicate a semi-marine and marine regime within the Black Sea.

There is strong variation in the taxonomy and diversity of foraminiferal assemblages across the Black Sea shelf. The western assemblages differ from the northwestern ones in having low numbers of *Haynesina anglica*, miliolids, and a total absence of *E. caspicum caspicum*. The eastern assemblages are distinguished by the presence of *Canalifera nigarensis*. The outer shelf assemblages differ from those of the inner shelf by having a greater number of Mediterranean immigrants of Lagenida, Buliminida, *Acervulina*, *Gavelinopsis*, *Planorbulina*, *Pateoris*, and *Pyrgo*, which are especially abundant near the Bosphorus (Yanko and Troitskaya 1987; Yanko 1989, 1990a, b; Yanko and Vorob'eva 1991). In general, diversity among foraminifera increases from the northwest toward the east and west, reaching maximum values near the Bosphorus. The main foraminiferal assemblages in the Black and Caspian Seas are given in Tables 1 and 2, at the end of this paper.

3.3 Late Pleistocene Ecostratigraphy

3.3.1 Tarkhankutian Beds (40–27 ky BP)

In the cores, the Late Pleistocene is represented by Tarkhankutian and Neoeuxinian beds. The Tarkhankutian beds were recovered in Core 2362 (Bed 1, dated 27,295 BP) on the Bulgarian shelf at a water depth of –103 m (28°24'2" N by 42°°12'6" E). They are represented by dark-grey terrigenous silt and clay (Figure 4A) with $CaCO_3$ at 18–32%, C_{org} at 0.3–0.7% (Malovitsky *et al.*1979), and a monospecific (*Dreissena rostriformis distincta*) mollusc assemblage. While no Mediterranean species are found among the molluscs (Govberg *et al.* 1979), they are present among foraminifera, represented by 12 species and 74 specimens (Figure 4B). Mediterranean holeuryhaline *A. tepida* dominates (Figure 4C), while strictoeuryhaline *Nonion matagordanus* and *Elphidium ponticum* play an accessory role (Figure 4D). At present, the closest foraminiferal assemblage, Od-2, with elements of NW-1 inhabits Odessa Bay (Table 1), indicating that the paleosalinity and paleodepth of the basin during the accumulation of Bed 1 was around 11‰ and >35 m, respectively.

Nevesskaya and Nevessky (1961) first reported Tarkhankutian sediments with a mixture of Caspian and Mediterranean molluscs from Karkinitsky Bay, on the northwestern shelf, at water depths of 30–35 m. Later, they were discovered in many places in the Pontic region, e.g., the Colchis Plain (Georgia) where they are overlain by subaerial peats dated *ca.* 31 ky BP at sampling depth –60 m (Dzhanelidze and Mikadze 1975). Popov and Zubakov (1975) and Popov (1983) recognized similar sediments as Surozhian. Svitoch *et al.* (1998) considered Tarkhankutian and Surozhian sediments as coeval, with an age of 40–25 ky BP. The Tarkhankutian transgression at 31,330±719 ky BP (Chepalyga 2002a, b) brought Mediterranean waters and organisms (Figure 4) into the Black Sea and increased salinity to about 8–11‰ (Nevesskaya 1965; Yanko 1989, 1990a). The submerged accumulative coastal bars of synchronous age are lo-

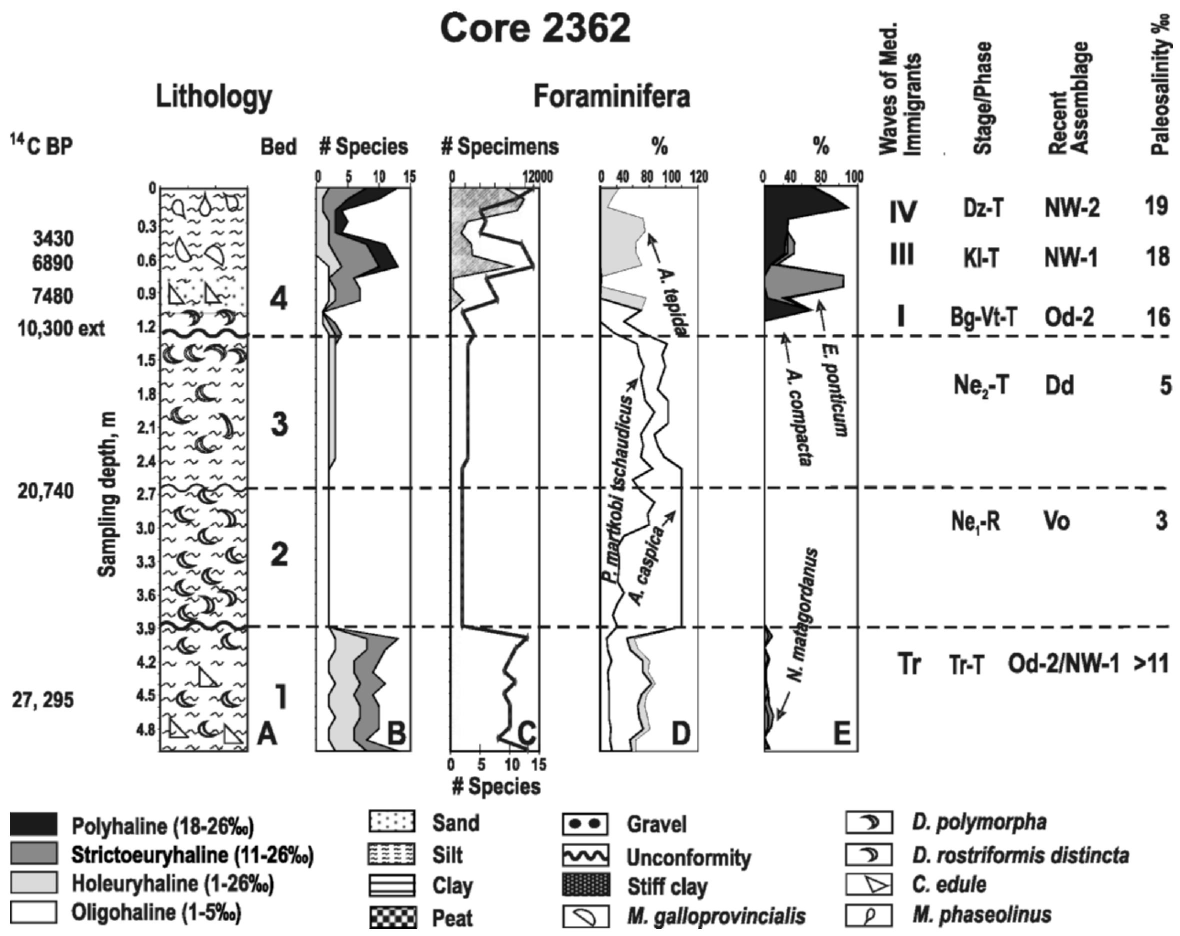

Figure 4. Diagrams for Core 2362 (Bulgarian shelf, 42°12'6" E by 28°24'2" N, water depth –103 m) showing down-core changes in lithology and foraminifera. Dotted lines indicate boundaries between the stratigraphic units. For explanations of recent assemblages, see Tables 1 and 2. Identification of the phases will be found in the caption to Figure 8.

cated at water depths of –22 to –30 m on the northwestern (Chepalyga *et al.* 1989; Chepalyga 2002a, b) and Romanian (Caraivan *et al.* 1986) shelf, indicating that Tarkhankutian sea level was about –30 m compared to the present. Temporally, the Tarkhankutian sediments correspond to Unit 3 (Çağatay 2003) in the Sea of Marmara. This unit contains some marine molluscs and benthic foraminifera indicating a weak Mediterranean marine incursion during the early part of MIS-3.

Interestingly, no similar sediments have yet been found in the Bosphorus. Instead, they have been recovered in Izmit Bay and the Sakarya Valley (Meriç *et al.* 1995; Yanko-Hombach *et al.* 2004).

The palynological diagrams are dominated by arboreal *Pinus* (subgenus *Diploxyton*) but contain broad-leaved *Quercus*, *Acer*, *Carpinus*, and grassy *Artemisia*, as well as Asteraceae, e.g., Bed 1, Core 2362 (Komarov *et al.* 1979). In the view of this author, they are similar to the palynological diagrams of the Schtilfrid soil in Austria (Frenzel 1964), which are typical of the moderately warm Würm Paudorf (Middle Weichselian) Pleniglacial, dated 27,990–28,120 BP (Fink 1962). This period of increased Eastern Mediterranean pluviality together with northwestern European permafrost degradation and climatic warming is associated with increased fluvial discharges (Huijzer and Vandenberghe 1998) and must have been accompanied by a noticeable increase in river discharge flowing into the Black Sea (Aksu *et al.* 2002b). By implication, the level of the Black Sea must have been high (Çağatay *et al.* 2000), and the Black Sea should have been connected to the Marmara via a south-flowing river (Aksu *et al.* 2002a, b). The presence of Mediterranean species in the Tarkhankutian sediments indicates northward flow from the Sea of Marmara as well.

3.3.2 Lower Neoeuxinian beds (27–17 ky BP)

In the cores recovered below isobath –100 m, the Tarkhankutian beds are separated from the overlying Lower Neoeuxinian beds (27–17 ky BP; Figure 5D) by an erosional unconformity (Figure 4A). According to Ross and Degens (1974), the Black Sea was in the process of evolving from a marine basin to a more freshwater environment by about 23 ky BP. Our data show that this process started earlier at *ca.* 27 ky BP, during the accumulation of the Lower Neoeuxinian beds. The latter are represented by alternations of grey silt and grey striped clays enriched with hydrotriolite, sand (minor), and shells of *D. rostriformis distincta* (e.g., Bed 2, Core 2362, Figure 4A; Bed 1, Core 2/86, Figure 6; Bed 1, Core 81, Figure 7A); $CaCO_3$ is about 50% and C_{org} >1% (Malovitsky *et al.* 1979). The foraminiferal assemblage consists of *A. caspica* and *P. martkobi tschaudicus* (Figure 4B, D; 6, 7D). The number of specimens is low (12–29, Figure 4C, 7C). A similar assemblage, Dd (Table 1), inhabits

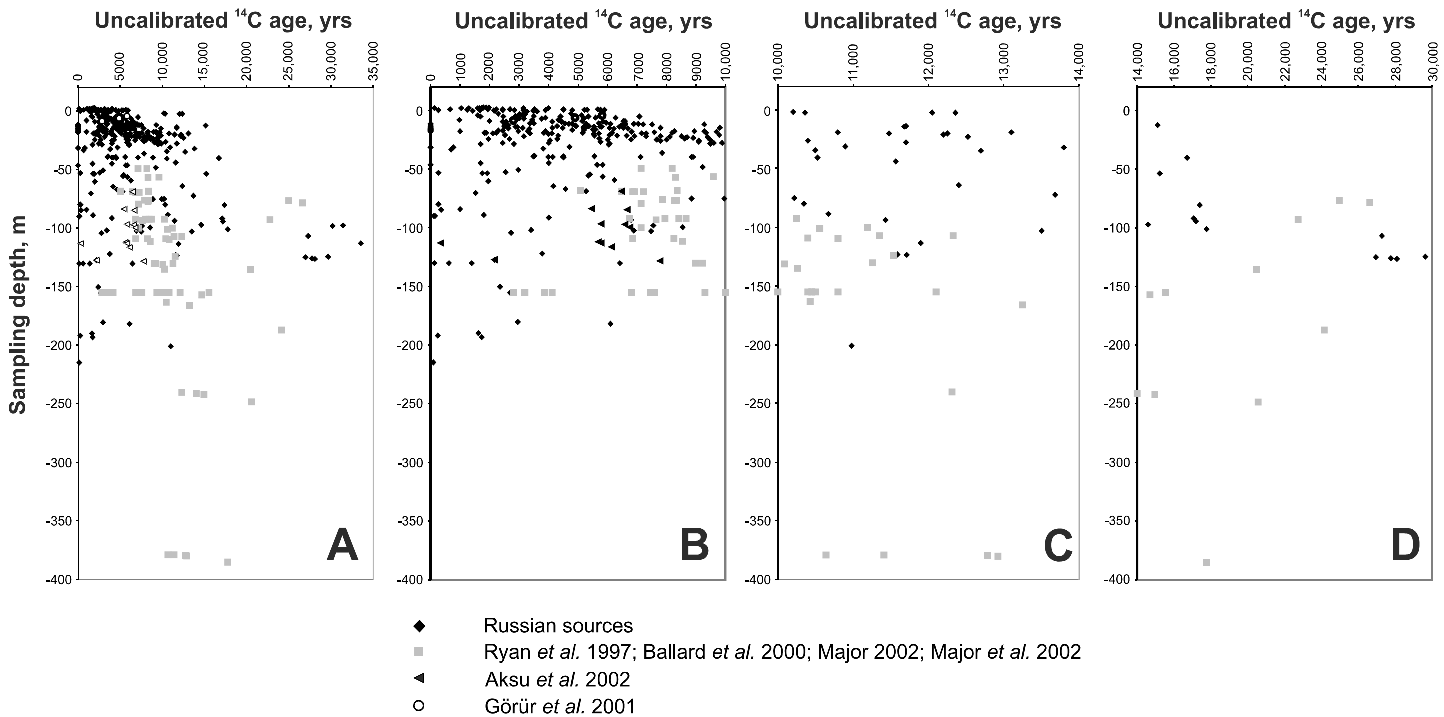

Figure 5. Scatter diagrams compiled from Appendices 1 and 2 showing uncalibrated ^{14}C records for the time intervals (A) 0–35 ky BP, (B) 0–10 ky BP, (C) 10–14 ky BP, and (D) 14–30 ky BP. Only age determinations obtained from mollusc shells are considered. For the time interval 35–10 ky BP, ^{14}C analysis was performed on *Dreissena* and occasionally *Monodacna* and *Viviparus*. For the time interval 10–0 ky BP, other mollusc shells, e.g., *Cardium edule*, *C. exiguum*, *Mytilus galloprovincialis*, and some others (see Appendices 1 and 2) were used in addition to *Dreissena*. The total data set consists of 424 radiocarbon records obtained by conventional (Russian sources) and AMS (western sources) methods.

Core 2/86

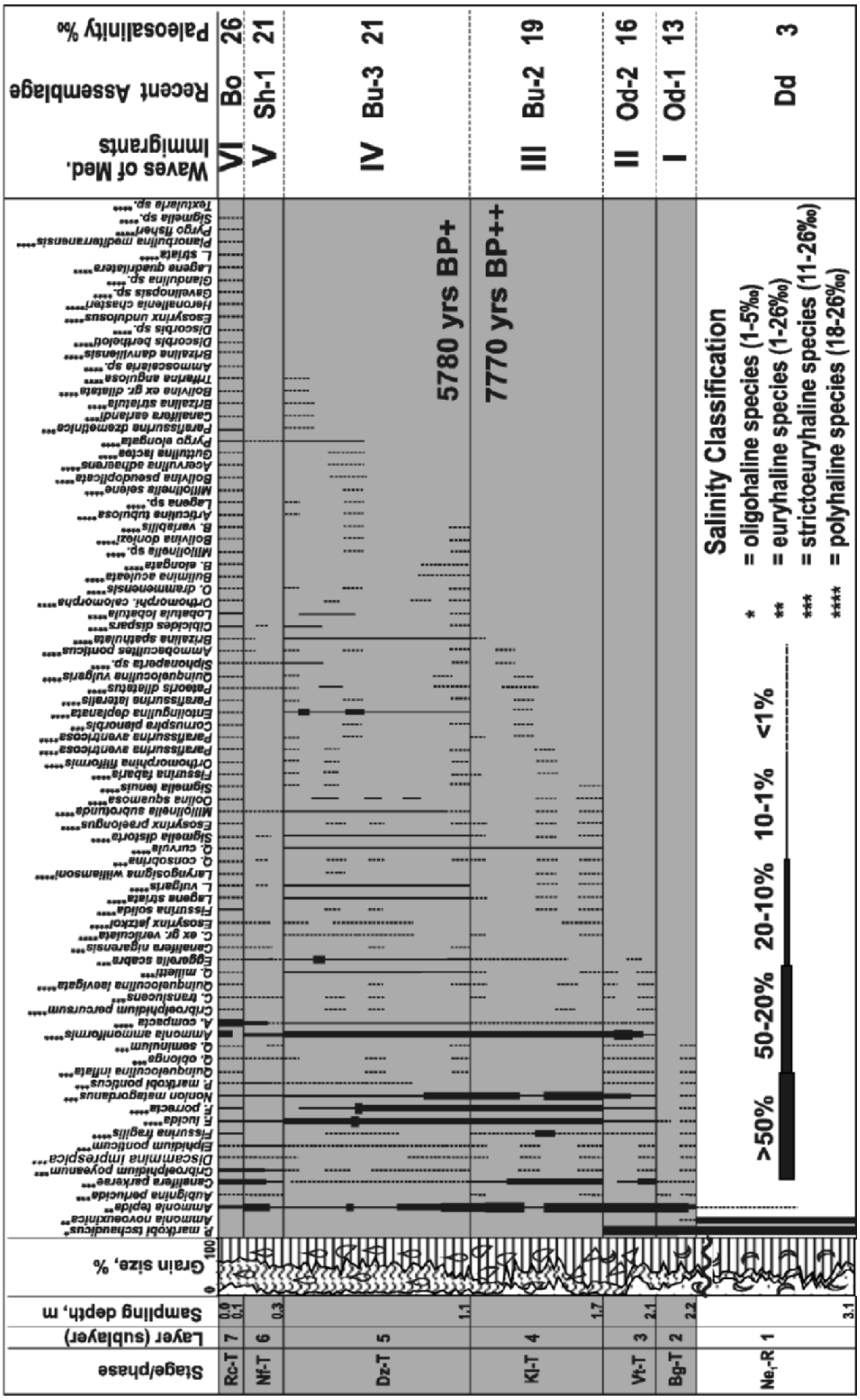
Stage/phase
Layer (sublayer)
Sampling depth, m
Grain size, %
Waves of Med. Immigrants
Recent Assemblage
Paleosalinity %
VI Bo 26
V Sh-1 21
IV Bu-3 21
III Bu-2 19
II Od-2 16
I Od-1 13
Dd 3
Rc-T
Nf-T
Dz-T
Kl-T
Vt-T
Bg-T
5780 yrs BP+
7770 yrs BP++
>50% 50-20% 20-10% 10-1% <1%
Salinity Classification
* = oligohaline species (1-5‰)
** = euryhaline species (1-26‰)
*** = strictoeuryhaline species (11-26‰)
**** = polyhaline species (18-26‰)

Figure 6. (Previous page) Diagram for Core 2/86 (Northern Bosphorus exit, 41°31'30" N by 28°59'32" E, water depth –103 m). The sediment column is subdivided into seven beds showing downcore changes in grain size and percentage of foraminiferal species. ^{14}C on *Mytilus* from cores MAR98-04 (+) and MAR98-06 (++) (Aksu *et al.* 2002b). For an explanation of the recent assemblages, see Table 1. Identification of the phases appears in the caption to Figure 8. For further explanations, see Figure 4.

present-day river deltas, indicating a shallow, semi-fresh paleoenvironment during the accumulation of the Lower Neoeuxinian beds. There are no Mediterranean species among the foraminifera, ostracoda (Yanko and Gramova 1990), and molluscs (Nevesskaya 1965). Similarly, there are no Mediterranean species in synchronous sediments of the Sea of Marmara (Unit 2, Çağatay 2003), indicating that they were deposited in a fresh/brackish environment (salinity <6‰), corresponding to the late part of MIS-3 and MIS-2. During the deposition of Unit 2, the water level was at –85 m, and the shelf areas were subaerially exposed or occupied by small, isolated lakes (Çağatay 2003). The Bosphorus was a freshwater lake (26–5.3 ky BP) with a sandy bottom containing freshwater molluscs of Black Sea Neoeuxinian affinity–*D. rostriformis*, *D. polymorpha*, *Monodacna pontica* (Algan *et al.* 2001). The level of the Aegean Sea, –115 m (Aksu *et al.* 1987), and the Sea of Marmara, –100 m (Smith *et al.* 1995) was about the same as the level of the Black Sea, assuming a lack of connection between the basins during early Neoeuxinian times (Svitoch *et al.* 1998).

The Early Neoeuxinian basin was semi-fresh, aerobic (Degens and Ross 1974), and heavily populated by benthic organisms, in particular by those with calcareous shells ($CaCO_3$ in sediments: ≥50%), e.g., molluscs, ostracoda, and on a much smaller scale, by foraminifera.

The Early Neoeuxinian palynological diagrams are dominated by *Artemisia*, Chenopodiaceae, *Adonis*, and *Thalictrum* (e.g., Bed 2, Core 2362). They are similar to those of the dry pine forest of Romania (Komarov *et al.* 1979; Pop 1957), the pine/birch forest and xerophyte steppe in southern Ukraine and Moldova (Artyushchenko *et al.* 1972; Kyrvel *et al.* 1976), and the steppe and forest-steppe on the Balkan Peninsula (Bottema 1974), all indicating a cold and dry climate (Nikonov and Pakhomov 1993). By implication, the river discharge into the Early Neoeuxinian lake must have decreased, causing a dramatic drawdown of the water level below the –100 m isobath (Kvasov 1975; Skiba *et al.* 1975; Fedorov 1977, 1978; Shcherbakov *et al.* 1978, Abashin *et al.* 1982; Shcherbakov 1983; Shnyukov *et al.* 1985; Fedorov 1988; Svitoch *et al.* 1998). A large portion of the present shelf was exposed, eroded, downcut some 40 m into the basement by the Pre-Danube, Pre-Dnieper and Pre-Dniester Rivers, and covered by subaerial loams (e.g., Shcherbakov *et al.* 1978; Shcherbakov 1983; Inozemtsev *et al.* 1984; Fedorov 1988). The river mouths were relocated 80–100 km seaward (Gozhik 1984b; Shnyukov *et al*. 1985), where they possessed poorly developed deltas and opened directly into the canyons on the continental

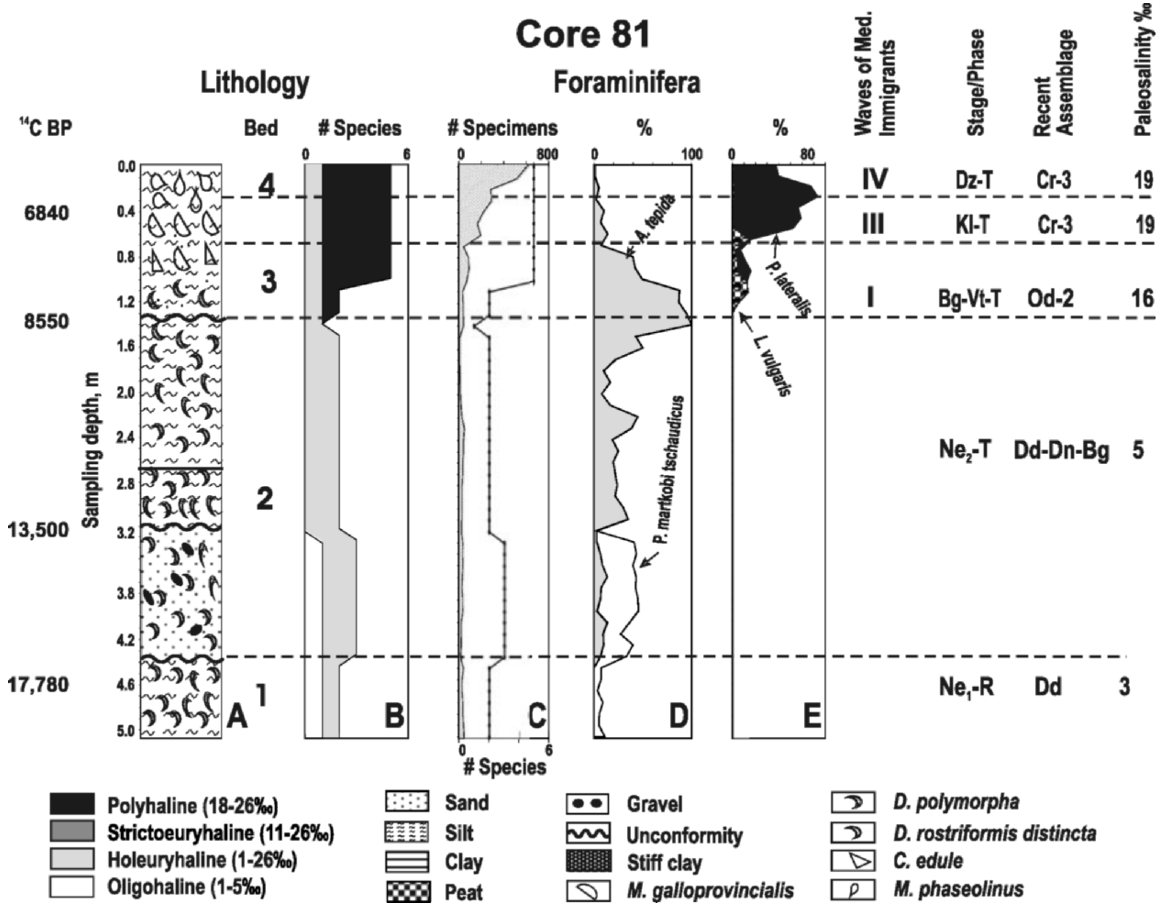

Figure 7. Diagrams for Core 81 (Crimean shelf, 44°24' N by 33°28'40" E, water depth –100 m). The sediment column is subdivided into four beds that show the downcore changes in lithology and number of foraminiferal species, specimens, and percentage of dominant species. Dotted lines indicate boundaries between stratigraphic units. For an explanation of the recent assemblages, see Table 1. Identification of the phases appears in the caption to Figure 8. For further explanations, see Figure 4.

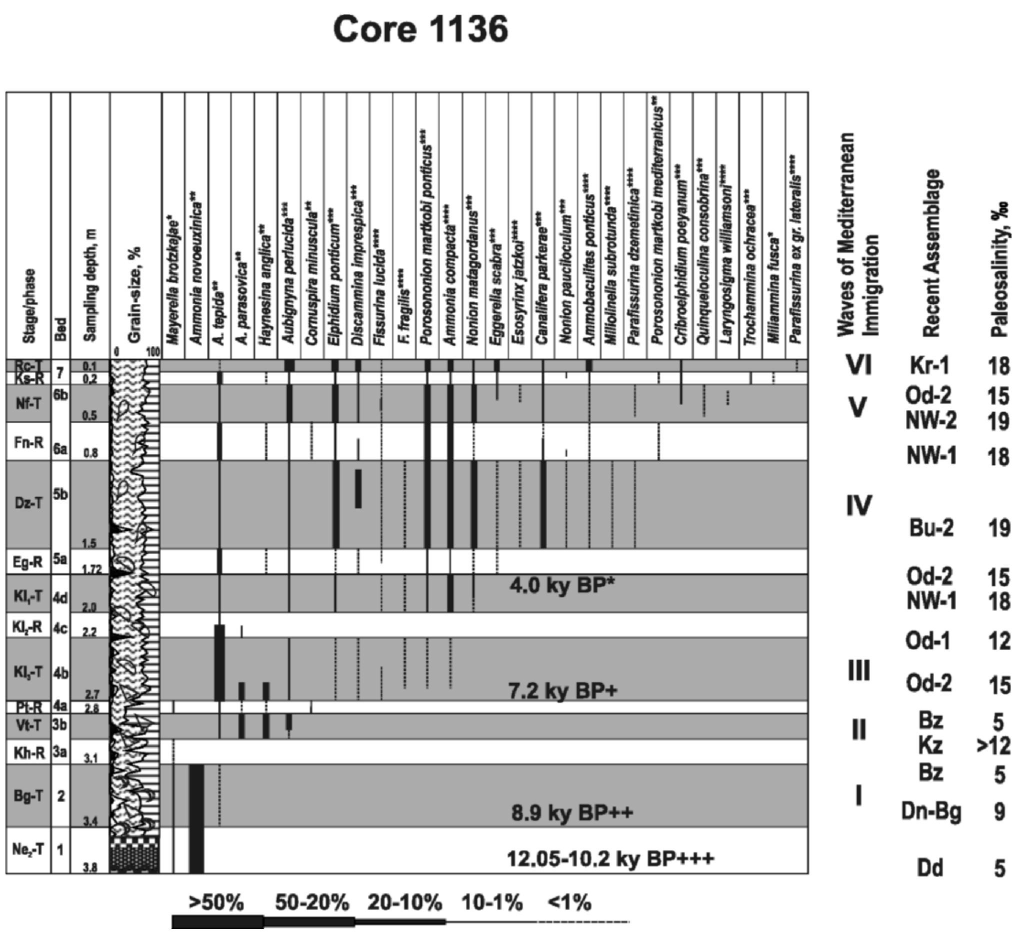

Figure 8. Diagrams for Core 1136 (Northwestern shelf, Karkinitsky Bay, 45°35'02" N by 32°25'07" E, water depth –31 m) showing downcore changes in grain size and percentage of species. ^{14}C on *Mytilus* (+), *Cardium* (++) and *Viviparus*/*Dreissena* (+++) from cores 4M, 1M (Balandin and Mel'nik 1987) and 711 (Gozhik *et al.* 1987) located in close vicinity to Core 1136 (for location, see Figure 1). For an explanation of the recent assemblages, see Table 1. For further explanations, see Figure 4. Transgressive (T, in grey) and regressive (R) phases: Tr = Tarkhankutian, Ne1 = early (lower) Neoeuxinian; Ne2 = late (upper) Neoeuxinian, Bg = Bugazian, Kh = Kolkhidian, Vt = Vityazevian, Pt = Pontian, Kl = Kalamitian, Eg = Eggrisian, Dz = Dzhemetinian, Fn = Phanagorian, Nf = Nymphaean, Ks = Korsunian, Rc = Recent.

slope (Konikov, this volume: his Figure 12). The river valleys and canyons were filled with thick (22–40 m) alluvial sediments (Kuprin and Sorokin, this volume) of the Ant age (22,800–16,900 BP) containing 27 freshwater and 14 brackish water shallow ostracods dominated by *Cyprideis littoralis* and *Ilyocypris bradyi* (Gozhik 1984c).

3.4.2 Late Neoeuxinian (17–10 ky BP)

Above isobath –100 m, the Lower Neoeuxinian beds are often overlapped by subaerial loams and further on by aquatic sediments with ostracoda *Candona*, *Candoniella*, and foraminifera *A. novoeuxinica*. This change indicates transformation of the bottom from an erosional to a subaquatic accumulative phase at the beginning of the Late Neoeuxinian transgression (Gozhik 1984b; Shnyukov 1985).

The Upper Neoeuxinian beds (e.g., Bed 3, Core 2362, Figure 4A; Bed 1, Core 1136, Figure 8A; Bed 1, Core 711, Figure 9A) cover the Black Sea floor below isobath –20 m almost everywhere: –18 m on the Turkish shelf (Görür *et al.* 2001), –30 m on the Bulgarian shelf (Filipova-Marinova, this volume), –20 m on the northwestern shelf (Gozhik *et al.* 1987; Konikov, this volume), –30 m on the Crimean shelf (Shnyukov 1985), –30 m on the Caucasian shelf (Balabanov *et al.* 1981; Yanko and Gramova 1990), and –11 m in the Kerch Strait (Put' 1981). In some places (e.g., the western part of the Golitsin Uplift located at the mouth of Karkinitsky Bay, see Figure 1 for location), they are exposed on the seafloor (Tkachenko *et al.* 1970; Ishchenko 1974;Tkachenko 1974; Yanko 1974, 1975, 1989). Their thickness varies up to 25 m (Put' 1981).

Lithologically, the Upper Neoeuxinian beds on the shelf are rather monotonous. They are represented by light grey sandy coquina and/or bluish-grey stiff clays that fill pre-Neoeuxinian depressions and paleoriver valleys (e.g., Arkhangel'sky and Strakhov 1938; Nevesskaya 1965; Semenenko and Kovalyukh 1973; Ostrovsky *et al.* 1977; Malovitsky *et al.* 1979; Balabanov *et al.* 1981; Skryabina 1981; Yanko 1982; Gozhik 1984a, b, d; Voskoboinikov *et al.* 1985; Shnyukov *et al.* 1985; Fedorov 1988; Gozhik *et al.* 1987; Yanko 1989, 1990a; Yanko and Gramova 1990; Glebov *et al.* 1996). The stiff clay has a massive structure, high density (about 2.7 g/cm^3), and low water content. The interstitial water salinity is 7‰ (Konikov, this volume: his Figures 3 and 4).

Molluscs are dominated by *D. polymorpha* and *D. rostriformis* on the inner and outer shelf, respectively. Other Caspian molluscs, such as *M. caspia*, are also abundant (Figure 9F). The foraminiferal assemblage is rather uniform, being dominated by oligohaline Caspian *M. brotzkajae* and *E. caspicum*, and holeuryhaline Black Sea endemic *A. novoeuxinica*. The number of specimens was found to be >100 (Figure 4B–D, 6, 7, 8B–D, 9B–D, 10B–D). Today, a similar foraminiferal assemblage, Dd, with elements of the Dn-Bg assemblage,

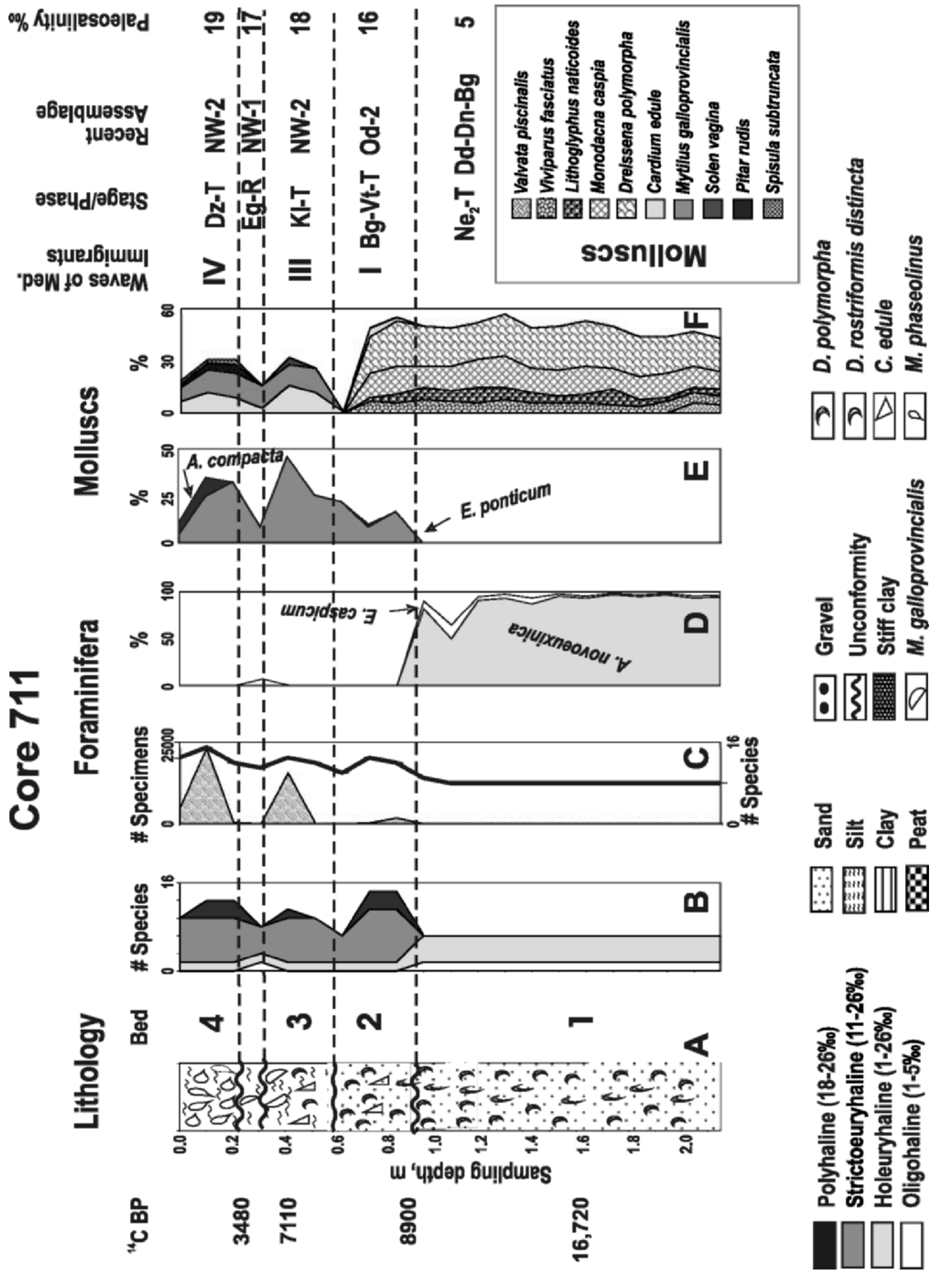
Core 711
Lithology
Foraminifera
Molluscs
¹⁴C BP
3480
7110
8900
16,720
Sampling depth, m
Bed
4
3
2
1
A
Species
B
Specimens
C
Species
D
A. novoeuxinica
E. caspicum
E
A. compacta
E. ponticum
F
%
Waves of Med. Immigrants
Stage/Phase
Recent Assemblage
Paleosalinity ‰
IV Dz-T NW-2 19
Eg-R NW-1 17
III Kl-T NW-2 18
I Bg-Vt-T Od-2 16
Ne2-T Dd-Dn-Bg 5
Molluscs
Valvata piscinalis
Viviparus fasciatus
Lithoglyphus naticoides
Monodacna caspia
Dreissena polymorpha
Cardium edule
Mytilus galloprovincialis
Solen vagina
Pitar rudis
Spisula subtruncata
Polyhaline (18-26‰)
Strictoeuryhaline (11-26‰)
Holeuryhaline (1-26‰)
Oligohaline (1-5‰)
Sand
Silt
Clay
Peat
Gravel
Unconformity
Stiff clay
M. galloprovincialis
D. polymorpha
D. rostriformis distincta
C. edule
M. phaseolinus

Figure 9. (Facing page) Diagrams showing downcore changes in lithology, number of species, specimens, and percentage of dominant foraminiferal and molluscan species for Core 711. For an explanation of the recent assemblages, see Table 1. Identification of phases appears in the caption to Figure 8. For further explanations, see Figure 4.

inhabits river deltas and semi-fresh limans (Tables 1 and 2), indicating a paleosalinity for the Late Neoeuxinian lake of about 5‰ in the shallow area; it could have reached 7‰, which is typical of interstitial salinity (Manheim and Chan 1974), and even 11‰ (Nevesskaya 1965) in deeper parts of the basin. Despite a relatively high salinity, no Mediterranean species are present. Instead, Caspian immigrants are abundant. The Late Neoeuxinian lake was aerobic and heavily populated by organisms with carbonate shells. The Upper Neoeuxinian sediments seem to be partially synchronous with the upper part of Unit 2 (Çağatay 2003) of the Sea of Marmara sediment column. The level of the Late Neoeuxinian lake was much higher (~ –20 m) than that of the Sea of Marmara (~ –85 m), and by implication, the Late Neoeuxinian lake discharged its waters into the Sea of Marmara. The Bosphorus continued to be a semi-fresh lake and might have served as a channel for southward water discharge from the Neoeuxinian lake. However, this discharge could have occurred through the Izmit Gulf-Sakarya Valley, as indicated by the presence of fresh/brackish facies with an age of 14.6 ky BP in borehole KS2 (Kerey *et al.* 2004).

Late Neoeuxinian palynological diagrams are dominated by *Quercus*, *Carpinus*, *Ulmus*, *Salix*, and *Betula*, with decreased concentration of *Pinus* and grass (Komarov *et al.* 1979; Kvavadze and Dzeiranshvili 1989). They are similar to the late glacial diagrams of the Balkan Peninsula (Bozilova 1973, 1975) and the Prichernomorian soil horizon (Veklich and Sirenko 1976) formed before 10.5 ky BP (Ivanova 1966). The climate warmed during Late Neoeuxinian times, which is indicated by the replacement of pine by broad-leaved forests.

In many places, the Upper Neoeuxinian beds are overlapped by peats (Figures 8 and 10A) of *ca.*10 ky BP: 10,600–9900 BP (Inozemtsev *et al.* 1984); 10,550 BP (Kind 1976); 10,130 BP (Balabanov *et al.* 1981); 9580 BP (Yanko and Troitskaya 1987), and/or very coarse sediments (Figure 10A). The maximum sampling depth (water depth plus depth in the core) of the peats is about 50 m. They were formed at the end of the Younger Dryas (*ca.* 10.2 ky BP) when the level of the lake dropped to about –50 m (Figure 5C and Balabanov, this volume: his Figure 3).

3.4.3 Holocene (*ca.* 10 ky BP–present)

The Late Neoeuxinian beds, often with erosional unconformity, are overlapped by Bugazian beds containing the first Mediterranean immigrants. Bugazian sediments are widely distributed below the –17 m isobath. Their

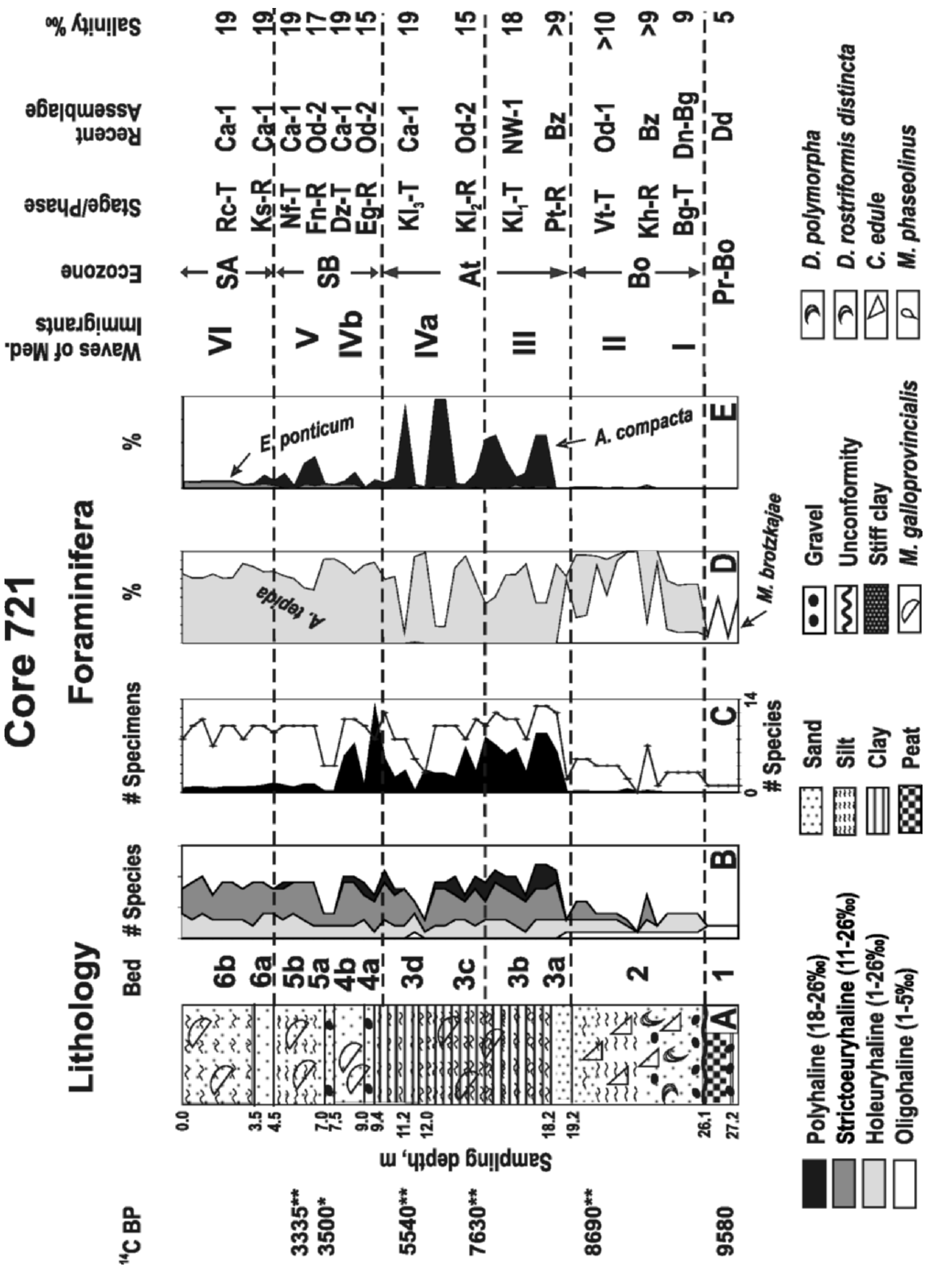
Core 721
Lithology
Foraminifera
14C BP
Sampling depth, m
Bed
Species
Specimens
%
%
Waves of Med. Immigrants
Ecozone
Stage/Phase
Recent Assemblage
Salinity ‰
3335**
3500*
5540**
7630**
8690**
9580
0.0
3.5
4.5
7.0
7.5
9.0
9.4
11.2
12.0
18.2
19.2
26.1
27.2
6b
6a
5b
5a
4b
4a
3d
3c
3b
3a
2
1
A
B
C
D
E
E. ponticum
A. tepida
A. compacta
M. brotzkajae
VI
V
IVb
IVa
III
II
I
SA
SB
At
Bo
Pr-Bo
Rc-T
Ks-R
Nf-T
Fn-R
Dz-T
Eg-R
Kl3-T
Kl2-R
Kl1-T
Pt-R
Vt-T
Kh-R
Bg-T
Ca-1
Ca-1
Ca-1
Od-2
Ca-1
Od-2
Ca-1
Od-2
NW-1
Bz
Od-1
Bz
Dn-Bg
Dd
19
19
19
17
19
15
19
15
18
>9
>10
>9
9
5
0
14
Species
Polyhaline (18-26‰)
Strictoeuryhaline (11-26‰)
Holeuryhaline (1-26‰)
Oligohaline (1-5‰)
Sand
Silt
Clay
Peat
Gravel
Unconformity
Stiff clay
M. galloprovincialis
D. polymorpha
D. rostriformis distincta
C. edule
M. phaseolinus

Figure 10. (Facing page) Diagrams showing downcore changes in lithology, number of species, specimens, and percentage of dominant foraminiferal species for Core 721. * = ^{14}C from archaeological source; ** = ^{14}C from Core 723 (Figure 1). For recent assemblages, see Table 1. For phases, see caption to Figure 8. For further explanations, see Figure 4.

thickness increases from 0.03–0.2 m on the slopes of submerged river valleys to 2.5 m on their bottom. They are represented by light grey or greenish-grey fine sand (Bed 2, Core 711, Figure 9A; Bed 2, Core 721, Figure 10A) or bluish-grey clayey silt (Bed 2, Core 1136, Figure 8) or silty clays (Bed 2, Core 2/86, Figure 6) with $CaCO_3$ at 31–36% and C_{org} at 0.5–1.1%. The sediments have rudimentary lamination expressed as alternating light grey and dark grey microlayers of 1–2 mm thickness (Konikov, this volume: his Figure 3).

The Bugazian palynological diagrams are characterized by a sharp decrease in grassy elements (e.g., wormwood, goosefoot) and conifers (*Pinus*, *Picea*, *Juniperus*). Instead, broadleaf *Quercus*, *Corylus*, *Ulmus*, *Betula*, and even beech become dominant, indicating moderate climate conditions typical of the Boreal Ecozone (Komarov *et al.* 1979). A similar palynological diagram of the deep sediments of the Black Sea and peats of the Ril mountain massif in Bulgaria have ^{14}C ages of 10,737±315 BP (Shimkus *et al.* 1977) and 10,035±65 BP (Bozilova 1973), respectively, which is close to the extrapolated ^{14}C age (10,300 BP) of Bugazian beds in Core 2362 (Figure 4A). A summary of the palynological data from lakes in a wide area west and south of the Black Sea shows that oak-pistacio (*Quercus*-*Pistacia*) forests were present over most of the region by 10 ky BP, although local desert-steppe vegetation persisted until about 7 ky BP in the southeast, from Lake Van (Eastern Turkey) to the Caspian Sea (Mudie *et al.* 2002). These forests indicate the early establishment of mesic climatic conditions characterized by >600 mm/year of precipitation in excess of evapotranspiration, as is presently found in most of central and western Europe (Hiscott *et al.*, this volume).

On the inner shelf, Bugazian mollusc assemblages are dominated by *D. polymorpha*, but rare *C. edule* are present as well. The foraminiferal assemblage includes *M. brotzkajae*, *A. novoeuxinica*, and *A. tepida* (up to 30 specimens per sample in total) and resembles the recent Dn-Bg assemblage distributed at depths >9 m and salinities of about 14.0‰. The interstitial water salinity is 15‰ (Konikov, this volume: his Figures 3 and 4). On the outer shelf, below isobath –40 m, Bugazian mollusc assemblages are dominated by *D. rostriformis distincta* (Govberg 1979), but rare *C. edule* are also present. The number of foraminiferal species and specimens increases to 13 and 7000, respectively (Figure 6, Bed 2). The euryhaline Mediterranean *A. tepida* dominates. The similar foraminiferal assemblage Od-1 lives today in Odessa Bay under a salinity of 13‰, indicating a brackish/semi-marine regime within the Black Sea during the accumulation of the Bugazian beds.

The Bugazian beds overlap the Upper Neoeuxinian beds with an erosional unconformity (Figure 10A) corresponding to a drop in sea level from –20 to –53 m below present (Balabanov, this volume: his Figure 3). The boundary between the two is clearly marked by the appearance of the first Mediterranean immigrants among the foraminifera (Figure 4E), molluscs (Figure 9F) and ostracoda (Yanko and Gramova 1990) in the Bugazian beds. This indicates the beginning of a Mediterranean transgression that transformed the semi-fresh Late Neoeuxinian lake into the semi-marine Black Sea. This transgression started *ca.* 9.8 ky BP (Grigor'ev *et al.* 1984) or 9.5 ky BP (Yanko and Troitskaya 1987), when sea level and salinity were about –42 m (Balabanov, this volume: his Figure 3) and 7‰, respectively. The increase in sea level and salinity was neither rapid nor catastrophic. Rather, it was gradual and occurred in a fluctuating manner (Balabanov *et al.* 1981; Yanko 1989, 1990a; Chepalyga 2002a, b; Balabanov, this volume: his Figure 3). A series of low amplitude transgressive and regressive (**in bold**) phases is clearly manifested on the inner shelf of the Black Sea (Figures 8 and 10). Nevesskaya (1965), Fedorov (1978), and Tsereteli (1975) named them Bugazian (9.4–8.1 ky BP), **Kolkhidian** (8.1–7.9 ky BP), Vityazevian (7.9–7.4 ky BP), **Pontian** (7.3–7.0 ky BP), Kalamitian (7.0–6.4 ky BP), **Eggrisian** (6.4–6.0 ky BP), Dzhemetinian (6.0–2.8 ky BP), **Phanagorian** (2.8–2.4 ky BP), Nymphaean (2.4–1.6 ky BP), **Korsunian** (1.6–1.2 ky BP), and Recent (1.2 ky BP–present). Due to the low amplitude of the regressive phases, they cannot be traced in cores recovered from a depth of more than 50 m (Figures 6 and 7), thus giving an impression of a gradual increase in sea level and salinity.

The first wave of Mediterranean immigration was poor. It slowed and even stopped during the Kolkhidian regression (Figures 8 and 10A, B, D). The re-colonization became more or less prominent during Vityazevian times, reaching its maximum in the course of the Kalamitian transgression (the third wave of Mediterranean immigration). On the inner shelf, the Kalamitian beds are represented by silt and sand with predominantly obliquely and horizontally stratified structure and C_{org} <2.6% (Bed 4b, Core 1136, Figure 10A). On the outer shelf, they form a single structureless and uniform bed of jelly-like sapropel that drapes all of the undulations of the unconformity surface as was correctly noted by Ryan *et al.* (1997). Among the molluscs, *D. polymorpha* disappears and *C. edule* becomes rare. Instead, *M. galloprovincialis* takes on a dominant role, being widely distributed across the shelf. Foraminifera, molluscs, and ostracoda indicate an increase in salinity to approximately 19‰, in full agreement with data on interstitial water salinity of the sediments (Konikov, this volume: his Figures 3 and 4). The Kalamitian pollen diagrams are dominated by mixed broadleaf trees–*Quercus*, *Corylus*, *Ulmus*, beech, lime, nutwood, and alder–with a small component of coniferous and herbaceous trees, indicating a

Holocene climatic optimum corresponding to the Atlantic Climatic Ecozone of about 7.4 ky BP (Shimkus *et al.* 1977; Komarov *et al.* 1979).

4. DISCUSSION

4.1 Level of the Neoeuxinian Lake

During the extensive studies conducted by Soviet and Eastern Bloc researchers, no evidence for a catastrophic flooding of the Neoeuxinian lake by Mediterranean waters was ever discovered. This writer stated at the 2003 annual meeting of the Geological Society of America (and cited by Ryan, this volume), "It is next to impossible that that such an event could have been missed by decades of Soviet research" (Yanko 2003). The "event" in question was the rapid flooding of the Neoeuxinian lake; it was not, as presented by Ryan, the lowstand shoreline or down-cutting of coastal river valleys that indicate a major water-level drop in the Black Sea's ice-age lake. These latter features were well known to Soviet geologists since at least the time of Arkhangel'sky and Strakhov (1938) and cannot under any circumstances be used by themselves as support for catastrophic flooding of the Neoeuxinian lake as presented by Ryan (this volume). The details are as follows.

Shallow foraminiferal assemblages in sediments of 27–17 ky BP distributed below the –100 m isobath demonstrate that the level of the Early Neoeuxinian lake stood at approximately that elevation, –100 m (Figure 5D). They also show that the water level reached approximately –20 m at about 10 ky BP (Figure 5B-C). If the level of the Neoeuxinian lake had been –120 m below present between 13.4 and 11 ky BP, or –95 m between 10 and 8.4 ky BP, no sediments with the foraminiferal and molluscan assemblages of these periods would have been recovered above these depths. However, such sediments cover most of the shelf up to the –20 m isobath (Figure 5B; Balabanov, this volume: his Figure 3; Konikov, this volume: his Figure 2; Kuprin and Sorokin, this volume: their Figure 5). These results preclude the first flooding event of the modified Flood Hypothesis (Ryan *et al.* 2003) that allegedly raised the level of the lake from –120 to –30 m during the Younger Dryas. Instead, the level of the lake dropped from –20 to approximately –50 m (Figure 5C; Balabanov, this volume: his Figure 3), as indicated by peats in numerous cores.

Subsequently, the lake level rose again to approximately –20 m by 9.5 ky BP owing to the warmer, post-Younger Dryas climate, coinciding with the first wave of Mediterranean immigrants into the Black Sea (Figure 10). After about 10 ky BP, the level of the Black Sea never again dropped below the –40 m isobath (Figure 5C; Balabanov, this volume: his Figure 3), nor exhibited a maximum amplitude of fluctuation greater than about 20 m, leaving no room for the second lowstand (–95 m) and flooding event described by Ryan *et al.* (2003) in the modified Flood Hypothesis.

The age of the submerged coastline at about –100 m was obtained on shells of *D. polymorpha* and/or *D. rostriformis* (Ryan *et al.* 2003; Lericolais *et al.*, this volume; Algan *et al.*, this volume). These species, however, are not exclusively Neoeuxinian indicators as suggested by Ryan *et al.* (1997, 2003), Major (2002), Algan *et al.* (2002), and Lericolais *et al.* (2004). In fact, their stratigraphic distribution is much wider, as they are present in all semi-fresh to brackish facies of the Pontic region from the Neogene (Meotis, *D. polymorpha*) and Pliocene (Apsheron, *D. rostriformis distincta*) (Nevesskaya 1965; Il'ina *et al.* 1976) to about 7 ky BP. They can also be found together with Mediterranean species (Figure 11). The AMS radiocarbon age of *Dreissena* varies, e.g., between 24.2 and 7.9 ky BP (Major 2002). Thus, *Dreissena* is not a reliable age-marker for the submerged coastline.

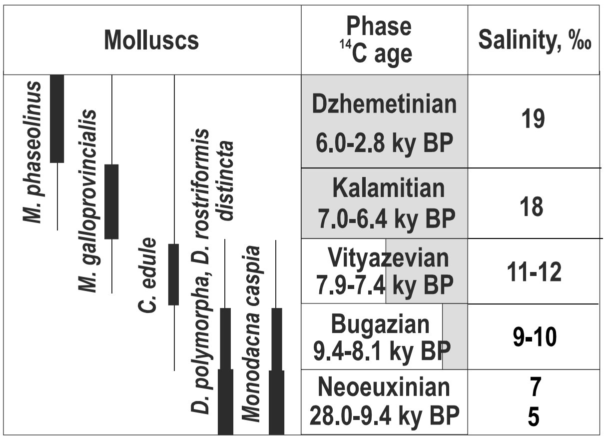

Figure 11. Stratigraphic distribution of key molluscs in the Late Pleistocene - Holocene sediments of the Black Sea (modified after Nevesskaya 1965; Grigor'ev *et al.* 1984). **Key**: white = semi-fresh, grey = semi-marine, white and grey = brackish water body. Thickness of the black vertical lines is proportional to abundance of the species.

Dreissena is also not a reliable paleobathymetric indicator because the distribution of fossil *Dreissena* is much wider than that of its living specimens. The difference between the two distributions becomes wider with increasing distance from shore, and it reaches its maximum on the continental slope due to reworking (Arkhangel'sky and Strakhov 1938). The alternation of transgressive and regressive phases intensifies reworking, leading to the mixing of paleontological material of different ages and ecological affinities (Shnyukov *et al.* 1985; Yanko and Gramova 1990). For example, the gravel-pebble beach sediments described by Ballard *et al.* (2000) as relics of the Neoeuxinian paleoshoreline at a water depth of –155 m contain a mixture of Caspian (*Dreissena*, *Turricaspia*) and Mediterranean (e.g., *Modiolus phaseolinus*) molluscs (see their Table 1). *M. phaseolinus* immigrated into the Black Sea at about 3 ky BP, however. Today, it lives at a water depths of 40–120 m under salinities ≤18‰ (Nevesskaya 1965). Consequently, these two molluscs do not indicate the same environment. Either one or both of them must be reworked, e.g., transported down to the bottom together with coarse material by underwater currents (Kuprin and Sorokin, this volume). A similar phenomenon was observed on the Crimean (Arkhangel'sky and Strakhov 1938) and Caucasian continental slopes in the canyons of the Kodori and Bzyb' Rivers (Kuprin *et al.* 1985; Solov'eva and Sorokin 1993).

In defending his dating of the lowstand and its accompanying arid landscape to the late Neoeuxinian (14–11 ky BP) based on radiocarbon dating of associated shells, Ryan (this volume) explains that both Kuprin *et al.* (1974) and Shcherbakov *et al.* (1978) had already drawn attention to the submerged shoreline. Based on more than 250 sediment cores containing shallow, brackish water foraminifera, ostracoda, and molluscs of Caspian affinity (e.g., Govberg *et al.* 1979; Yanko 1990a; Yanko and Gramova 1990; Yanko-Hombach 2003, 2004) without the appearance of a single Mediterranean species, however, Shcherbakov *et al.* (1978) attributed this lowstand to the LGM (20–18 ky BP). Indeed, Russian geologists did not miss or ignore the occurrence of a major regression within the Pontic basin. They simply assumed that it was earlier in date and that the subsequent transgression was neither abrupt nor catastrophic.

Foraminifera serve as a more powerful tool for paleoenvironmental reconstructions than molluscs, as demonstrated herein for the Tarkhankutian beds. However, their use still requires caution. For example, it is methodologically wrong to use *Ammonia* as indicating a shallow and low salinity paleoenvironment as was done by Kaminski *et al.* (2002). There are at least 10 species of *Ammonia* in the Black Sea, each of which has its own ecological preferences, which vary from oligohaline to polyhaline conditions (Yanko 1990b).

Following the second scenario of the Flood Hypothesis, one would expect that the older (13.4–11 ky BP) submerged coastline lies at a water depth of about –120 m, while the younger (10–8.4 ky BP) would be located at a water depth of about –95 m. However, the water depth and age of the submerged shorelines vary:

Turkish shelf (near Sinop): –155 m, 7.5–6.8 ky BP (Ballard *et al.* 2000);
Turkish shelf (near Sakarya Valley): –90 m, 11.8 ky BP (Algan *et al.*, this volume);
Romanian shelf: –95 m, 10.2–8.6 ky BP (Lericolais *et al.*, this volume);
Romanian shelf: –100 m, Last Glacial Maximum, MIS-2 (Winguth *et al.* 2000);
Romanian shelf: –120 m, 13.4–11 ky BP (Ryan 2003);
Northwestern shelf (between Karkinitsky and Kalamitsky Bays: –140 m, 14.7–10.0 ky BP (Ryan *et al.* 1997);
Crimean shelf: –100 m, 20–18 ky BP (Shcherbakov *et al.* 1978);

Bulgarian shelf: –100 m, Chaudinian* (Krystev *et al.* 1990; Yanko-Hombach *et al.* 2004) [*age of Chaudinian beds is *ca.* 800 ky BP];

Caucasian, Bulgarian, and Kerch shelf: –147 to –70 (Skiba *et al.* 1975; Esin *et al.* 1980; Goncharov and Evsyukov 1985; Glebov 1987; Glebov *et al.* 1996; Glebov and Shel'ting, this volume).

Abashin *et al.* (1982) show that the depth of the outer edge of the Early Neoeuxinian terrace (with coastal bars constituted by shallow lacustrine facies) changes from east to west from –110 m in the vicinity of the Kerch Strait to –90 m in the central part of the southern shelf of Crimea, and from –160 to –200 m in the western part.

Most probably, variations in the water depths of submerged coastlines relate to neotectonic processes that have caused differential uplift/subsidence of different blocks of the seafloor (Tkachenko *et al.* 1970; Tkachenko 1974; Abashin *et al.* 1982; Tkachenko, personal communication, January 5, 2005), vertical movements similar to those that were noticed recently in Izmit Bay (Sea of Marmara) during the 1999 earthquake (Öztürk *et al.* 2000). The displacement of the shelf break and development of submerged staircase terraces and off-shore cliffs and bars that cluster at certain bathymetric levels (Glebov and Shel'ting, this volume) support this conclusion. Variations in age are likely related to reworking of the mollusc shells used for radiocarbon dating.

At the beginning of the Late Neoeuxinian transgression, the level of the lake must have risen rather rapidly from –100 m at 17 ky BP to –40 m at 12 ky BP (12 mm per year); for comparison, the recent rate of sea-level rise is 2 mm per year (Tushingham and Peltier 1991). Under climate warming, the melting of the Scandinavian Ice Sheet and massive river discharge increased the level of the Caspian Sea to +50 m. Such a large amount of water could not be retained in the Caspian depression and was discharged into the Neoeuxinian lake through the Manych-Kerch Outlet (Mamedov 1997; Chepalyga 2003, this volume).

Such a relatively rapid increase in the Late Neoeuxinian lake level prevented the stabilization of the shoreline and the development of accumulative coastal bars (Mel'nik 1997). Therefore, the coastal dunes of Ryan *et al.* (1997) and Lericolais *et al.* (this volume) must be older than *ca.* 17 ky BP.

4.2 Salinity of the Neoeuxinian Lake

Foraminifera are not adapted to freshwater. As a rule, the boundary between brackish and freshwater environments is marked by their disappearance, with the exception of the organic-walled Allogromiida (Sen Gupta 1999). For this reason, the presence of relatively diverse calcareous foraminifera implies that the Neoeuxinian basin could not be fresh.

However, the advocates of the Flood Hypothesis (Ryan *et al.* 2003;

Lericolais *et al*, this volume) insist that the Neoeuxinian lake was fresh due to drainage from the melting of the ice cap after Melt Water Pulse 1A. This assumption is based on the belief that *D. polymorpha* and *D. rostriformis* may serve as freshwater indicators (Ryan *et al.* 1997; Ryan and Pitman 1998; Ballard *et al.* 2000; Major *et al.* 2002; Filipova-Marinova and Christova 2004; Filipova-Marinova *et al.* 2004; Lericolais *et al.* 2004, this volume; Algan *et al*, this volume; Ryan, this volume).

Nevesskaya (1963, 1965), Il'ina (1966), and Yanko-Hombach (2004) argue that the Neoeuxinian basin was semi-fresh to brackish. This difference of opinion results from an erroneous interpretation of the salinity requirements for *D. polymorpha* and *D. rostriformis*, which, in the Caspian Sea today, tolerate salinities $\leq$ 13‰, similar to other molluscs (*M. caspia*), ostracoda (*Leptocythere bacuana*, *Loxoconcha lepida*) and foraminifera (*E. caspicum*, *M. brotzkajae*) that coexist with *Dreissena* species in sedimentological sequences (Davitashvili and Merklin *et al.* 1966; Shornikov 1972; Yanko and Gramova 1990).

Paleontological and geochemical data are in full agreement. The Neoeuxinian salinity of the interstitial water is 4.0‰ (Bruevich 1952; Glagol'eva 1961), about 5.0‰ (Markov 1965), 6.0‰ (Manheim and Chan 1974), 6.8‰ (Shishkina 1962), and even 10-12‰ (Konikov, this volume: his Figures 3 and 4). Whatever reasonable set of assumptions one uses, it is impossible to equate salinity patterns in the interstitial waters with a totally fresh Neoeuxinian lake (Manheim and Chan 1974). Thus, the Neoeuxinan lake was brackish with a continental type of salinity where Ca^{2+} and SO_4^{2-} dominate over K^+ and Cl^-, respectively (Nevesskaya 1970). This is probably a consequence of the Caspian flood described by Chepalyga (this volume).

No oxygen depletion has been found in the Neoeuxinian sediments, thus signifying that the lake was aerobic (Arkhangel'sky and Strakhov 1938; Nevesskaya 1965; Degens and Ross 1974; Fedorov 1978). Our data are in full agreement with this conclusion.

4.3 Sea Level and Salinity in the Holocene

If there had not been northward inflow from the Sea of Marmara prior to 7.2 or 8.4 ky BP (as required under the first and second flood scenarios), then no Mediterranean immigrants would be present in the sediments of about 10 ky BP. However, the first Holocene wave of Mediterranean immigrants appears in the sediment sequences at about 10 ky BP (e.g., Figure 4). If re-colonization of the Neoeuxinian lake by Mediterranean organisms had been as rapid as is proposed by the advocates of the Flood Hypothesis (Ryan *et al.* 1997, 2003), we would observe a dramatic increase of foraminiferal species and specimens in Bugazian sediments. However, their diversity and abundance are low. Moreover, upcore variation leads to almost complete disappearance during the Kolkhidian regression, which is counterindicative of the alternative gradual re-colonization of the Black Sea by Mediterranean organisms suggested by the opponents of the Flood Hypothesis (Kaminski *et al.* 2002). Indeed, there are six major waves of re-colonization reflecting the fluctuating character of the Holocene Mediterranean transgression. A strong shift of the Sr isotope signature toward marine values at 8.4 ky BP (Major 2002) most likely indicates progressive salination of the Black Sea by increasing Mediterranean inflow during the Vityazevian transgression, but not at the time of the Black Sea–Sea of Marmara reconnection. The prominent foraminiferal re-colonization occurred at about 7.0 ky BP and coincides with a massive immigration of *M. galloprovincialis* when salinity increased to its present value. It was, however, still not high enough to support an immigration of planktonic foraminifera.

Thus, the immigration was neither rapid nor catastrophic, and so, neither the sea-level nor the salinity increase was catastrophic. Instead, the change occurred in a fluctuating manner that can be seen only above the –50 m isobath due to the low amplitude of the regressive phases. A high-resolution micropaleontological study with detailed knowledge of the ecological preferences of each foraminiferal species is needed to reconstruct the low amplitude fluctuations of sea level.

4.4 Alternative to the Bosphorus Connection Between Adjacent Basins

If the Upper Neoeuxinian sediments are found at water depths of about –20 m prior to 10 ky BP, then one must assume that the level of the lake was higher than the level of the world ocean at that time: –60 m (Fairbanks 1989). If so, the Neoeuxinian lake should have discharged an excess of brackish water into the Sea of Marmara, as was correctly pointed out by many authors (e.g., Fedorov 1978; Aksu *et al.* 2002a, b). If this persistent outflow occurred through the Bosphorus Strait (Aksu *et al.* 2002a, b), it should have prevented migration of Mediterranean organisms northward. Their presence in the Black Sea sediments, however, suggests that an alternative connection between the Black Sea and Sea of Marmara could have existed at about 9.5 ky BP (Yanko-Hombach *et al.* 2004).

If Mediterranean immigrants with an age of about 10 ky BP are present in the Black Sea, one would expect to find them in the Bosphorus, however, no Mediterranean immigrants younger than 5.3 ky BP have been found to date. The Bosphorus was a freshwater lake between 26 and 5.3 ky BP when the first euryhaline Mediterranean molluscs entered its sediments (Algan *et al.* 2001). Deposition of coarse *Mytilus*-bank and *Ostrea*-bank units suggests that the establishment of the present dual-flow regime in the Bosphorus occurred at 4.4

ky BP. A clear stratification of the water in the Bosphorus is apparent from about 4 ky BP (Yanko *et al.*1999b). Thus, the Bosphorus seems not to have been the main link between adjacent basins prior 5.3 ky BP

It can be argued that the record of the Bosphorus Strait boreholes is incomplete and/or marine sediments in the strait were eroded. However, the paleotopographic relief in the strait, at least below the mid-Holocene unconformity disclosed by geophysical profiles, contradicts this interpretation (Kerey *et al.*2004). At the same time, marine sediments with an age of about 30 ky BP (Tarkhankutian?) and about 8 ky BP (Vityazevian?) are found in Izmit Bay and the Sakarya Valley, suggesting an alternative route between the Black Sea and Sea of Marmara (Meriç 1995; Kerey *et al*2004; Yanko-Hombach *et al.*2004).

5. CONCLUSIONS

During the moderately warm Würm Paudorf (Middle Weichselian) Pleniglacial (prior to *ca.* 27 ky BP), there was a brackish Tarkhankutian basin connected with the Sea of Marmara. At the LGM, this connection was interrupted, and the level of the Tarkhankutian basin dropped to about –100 m, transforming it into the Early Neoeuxinian lake. The lake did not have a connection with the Caspian Sea.

In the warming climate of about 17 ky BP, a massive water discharge, most likely from the Caspian Sea via the Manych Spillway, increased the level of the Late Neoeuxinian lake to about –20 m. The latter must have overflowed, pouring its excess semi-fresh to brackish water into the Sea of Marmara and from there into the Mediterranean. During the short climatic cooling episode occurring at the Younger Dryas, the level of the lake dropped from –20 to about –50 m and then rose again to about –20 m. After *ca.* 10 ky BP, the level of the Black Sea never again dropped below the –40 m isobath, nor exhibited a maximum amplitude of fluctuation greater than 20 m.

At *ca.* 10 ky BP, the lake level reached –20 m again, allowing Mediterranean water and organisms to enter the Late Neoeuxinian basin. This re-colonization of the Black Sea occurred in an oscillating manner. It was slow at the beginning, becoming most prominent at about 7.0 ky BP. The connection between adjacent basins was probably not through the Bosphorus Strait, but via an alternative route, e.g., Izmit Bay–Sapanca Lake–Sakarya River.

On average, sea level rose gradually, but in an oscillating manner, to its present level, and perhaps slightly higher, averaging 3 cm per 100 years but certainly not 15 cm per day (almost 55 m per year) as postulated by the Noah's Flood Hypothesis. An increase in sea level of 3 cm per 100 years would not be noticed by the region's inhabitants and would not have accelerated their dispersion into the interior of Europe, bringing us to conclude that “Noah’s Flood” in the Black Sea is a contemporary legend.

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Table 1. Live foraminiferal assemblages of the Black Sea and Sea of Azov. Locations appear in Figure 1.

Connection with the Black Sea	River Discharge	Area	Depth, m	Salinity, ‰	No. of Stations	No. of Species	Dominant Species, max %	Accessory Species	Assemblage Index
Permanent	Strong	Danube delta	>5	1–5 (avg 1)	28	4	*A. novoeuxinica*, 76 *A. tepida*, 16	*Ma. brotzkajae*, *Ma. kolchidica*	Dd
		Dniestrovsky liman	>9	0.03–9 (avg 2)	10	6	*A. novoeuxinica*, 70 *A. tepida*, 17 *H. anglica*, 12	*M. fusca*, *H. anglica*	Dn
		Dniepro-Bugsky liman	>9	0.1–14.3 (avg 2.5)	10	6	*A. novoeuxinica*, 70 *A. tepida*, 17 *H. anglica*, 12	*M. fusca*, *H. anglica*	Dn-Bg
	Weak	Golovitsa liman	>9	5	20	5	*A. novoeuxinica*, 76 *A. tepida*, 19 *H. anglica*, 14	*M. fusca*, *H. anglica*	Gv
		Berezansky liman	>9	1–14 (avg 5)	10	10	*A. novoeuxinica*, 76 *A. tepida*, 18 *H. anglica*, 4	*M. fusca*, *H. anglica*	Bz
	Absent	Sinoe liman	>2	5–18 (avg 5)	13	6	*A. tepida*, 86 *H. anglica*, 10	*J. polystoma dacica*	Si
Restricted	Temporary (season-dependent)	Tiligulsky liman	>19	1–76 (avg 11)	10	11	*A. tepida*, 47 *H. anglica*, 41	*A. novoeuxinica*	Tl
		Khadzhibeysky liman	>14	2–63.5 (avg 12)	6	11	*A. tepida*, 47 *H. anglica*, 41	*A. novoeuxinica*	Kz
		Nuntash lagoon	?	21	9?	6	*A. tepida*, 63 *N. matagordanus*, 20 *P. martkobi ponticus*, 16	*Q. seminulum*	Nu
Absent	Absent	Alibey lagoon	>2.5	1–140 (avg 27.4)	10	14	*H. anglica*, 50 *A. tepida*, 40	*E. caspicum azovicum*, *A. parasovica*, *Au. perlucida*	Al
		Techirghiol lagoon	>1	75	5	7	*A. tepida*, 30 *T. aguajoi*, 30 *J. polystoma dacica*, 30	*M. fusca*	Tg

Free	Strong	Odessa Bay	>10	3–14 (avg 12)	100	13	*A. novoeuxinica*, 55 *H. anglica*, 20 *A. tepida*,14	*E. caspicum azovicum*	Od-1
	Significant	Odessa Bay	11–25	11–16 (avg 15)	200	33	*A. tepida*, 48 *H. anglica*, 19 *P. martkobi ponticus*, 12	*Au. perlucida*, *E. ponticum*, *A. parasovica*	Od-2
	Absent	Karkinitsky Bay	>35	18.2–19 (avg 18.6)	50	40	*A. tepida*, 31 *A. caucasica*, 10 *Q. seminulum*, 11	*Q. laevigata*, *Q. bicornis*	Kn
		Kalamitsky Bay	>35	18.3	20	39	*A. tepida*, 20 *E. scabra*,15 *A. caucasica*, 10	*M. secans*, *Ab. ponticus*	Kl
Open shelf	Absent	NW and Crimean (western) outer shelf	36–70	18.3	200	38	*A. compacta*, 29 *E. ponticum*, 20 *C. parkerae*, 9 *Cr. poeyanum*, 12	*N. matagordanus*	NW-1
		NW and Crimean (western) outer shelf	71–150	19	160	33	*A. compacta*, 27 *P. martkobi ponticus*, 12 *N. matagordanus*, 9 *Pa. dzemetinica*, 6	*L. vulgaris*, *Es. jatzkoi*, *E. deplanata*, *Lr. williamsoni*	NW-2
		Crimean (eastern) inner shelf	>35	18	10	37	*A. tepida*, 43 *E. scabra*, 13 *P. martkobi ponticus*, 10	*C. nigarensis*	Kr-1
		Crimean (eastern) outer shelf	36-70	19–20	10	40	*A. compacta*, 29 *C. parkerae*, 16 *Cr. poeyanum*, 13	*N. matagordanus*	Kr-2
		Crimean (eastern) outer shelf	71–200	20–21	10	32	*A. compacta*, 34 *N. matagordanus*, 12 *F. lucida*, 10	*L. vulgaris*, *Lr. williamsoni*, *Es. yatzkoi*	Kr-3
Restricted	Absent	Kerch Strait (central part)	>10	14–15	10	17	*A. parasovica*, 34 *A. tepida*, 22	*H. anglica*, *Au. perlucida*	Ke-1
		Kerch Strait (southern part)	>10	17.1–17.5	10	17	*A. tepida*, 30 *A. parasovica*, 25	*H. anglica*, *P. martkobi ponticus*	Ke-2
	Moderate	Sea of Azov (southern part)	>12	12–13	20	17	*A. parasovica*, 40 *A. tepida*, 23	*E. caspicum azovicum*, *H. anglica*	Az-1
		Sea of Azov (central part)	>12	11–12	20	17	*A. parasovica*, 45 *A. tepida*, 16 *A. novoeuxinica*, 10	*E. caspicum azovicum*, *H. anglica*	Az-2

Connection with the Black Sea	River Discharge	Area	Depth, m	Salinity, ‰	No. of Stations	No. of Species	Dominant Species, max %	Accessory Species	Assemblage Index
Restricted	Strong	Taganrogsky Bay (southern part)	>7	7–9	10	11	*A. novoeuxinica*, 79 *A. tepida*, 10	*E. caspicum azovicum*, *H. anglica*	Tg-1
		Taganrogsky Bay (northeastern part)	>2	4.3	10	9	*A. novoeuxinica*, 82 *A. tepida*, 9	*Ma. brotzkajae*	Tg-2
Open shelf	Minor	Caucasian inner shelf	>35	18–19	80	42	*A. compacta*, 30 *P. martkobi ponticus*, 22	*A. caucasica*, *E. scabra*	Ca-1
	Absent	Caucasian outer shelf	36–70	19–20	65	41	*A. compacta*, 46 *P. martkobi ponticus*, 18 *N. matagordanus*, 12	*A. ammoniformis*, *E. scabra*	Ca-2
		Caucasian outer shelf	71–220	20–21	75	34	*A. compacta*, 57 *F. solida*, 13	*A. ammoniformis*, *N. matagordanus*, *E. scabra*	Ca-3
Open shelf	Moderate	Bulgarian inner shelf	8–35	17–19	14	38	*A. tepida*, 40 *A. compacta*, 14 *A. ammoniformis*, 13	*A. caucasica*, *Cr. poeyanum*, *Au. perlucida*	Bu-1
	Absent	Bulgarian outer shelf	36–70	19–19.6	30	41	*A. ammoniformis*, 35 *A. compacta*, 21 *C. parkerae*, 12	*A. tepida*, *P. martkobi ponticus*, *Au. perlucida*	Bu-2
		Bulgarian outer shelf	71–220	21–22	60	43	*A. compacta*, 30 *A. ammoniformis*, 15 Lagenida, 18	*N. matagordanus*, *Pa. dzemetinica*, *F. lucida*	Bu-3
		Southern shelf	71–220	21–23	20	49	*A. ammoniformis*, 32 *A. compacta*, 28 Lagenida, 25	*Py. elongata*, *N. matagordanus*, *Cr. poeyanum*	Sh-1
		Northern Bosphorus exit	100–120	26.2	10	79	*A. ammoniformis*, 32 *A. compacta*, 28 Lagenida, 13	*Bolivina*, *Brizalina*, *Pyrgo*	Bo

Table 2. Live foraminiferal assemblages of the Caspian Sea.

Part of Caspian Sea	Connection with Caspian Sea	River Discharge	Area	Depth, m	Salinity, ‰	No. of Stations	No. of Species	Dominant Species, max %	Accessory Species	Assemblage Index
Northern	Free	Very strong	Volga River delta	>3	0.1–7.5 (avg 2.3)	10	3	*A. caspica*, 96	*Ma. brotzkajae*, *M. fusca*	Vo
		Strong	Northeastern inner shelf	>17	7–9	11	9	*A. caspica*, 73 *Am. verae*, 17	*Ma. brotzkajae*	NC-1
		Weak	Northwestern inner shelf	>22	9–12	11	9	*A. caspica*, 66 *Am. verae*, 19	*E. caspicum caspicum*, *M. fusca*	NC-2
Central	Free	Strong	Western inner shelf	>35	11–12.5	11	11	*Am. verae*, 31 *A. caspica*, 23	*C. minuscula*	CC-1
		Weak	Western outer shelf	36–70	12.4–12.9	11	3	*A. caspica*, 88	*M. fusca*, *C. minuscula*	CC-2
		Absent	Eastern inner shelf	>35	12.7–13	11	14	*A. caspica*, 50	*E. caspicum caspicum*	CC-3
			Eastern outer shelf	36–70	12.7–13	11	3	*A. caspica*, 89	*M. fusca*, *C. minuscula*	CC-4
			Krasnovodsky Bay	>5	14–15	11	17	*Am. verae*, 55	*S. perexilis*	Kr
	Restricted in 1968	Absent	Kara-Bogaz–Gol Bay	>2	13–14	11	13	*A. caspica*, 54 *B. macrostoma*, 28	*T. aguajoi*	KBG-1
	Absent in 1981		Kara-Bogaz–Gol Bay	>2	60–65	11	4	*T. aguajoi*, 80	*B. macrostoma*	KBG-2
	Free in 1968		Kara-Bogaz–Gol Strait	>2	12.2–13.3	11	12	*A. caspica*, 43	*Am. verae*	KBG-s
Southern	Free	Very strong	Kura delta	>10	>3	6	3	*A. caspica*, 97	*Ma. brotzkajae*	Kd
		Absent	Western inner shelf	>35	12.2	11	18	*E. caspicum caspicum*, 20	*E. shohinae*	SC-1
			Western outer shelf	36–70	12.8	11	3	*A. caspica*, 91	*M. fusca*	SC-2
		Weak	Turkmensky Bay	>35	12.6–13.2	11	12	*A. caspica*, 70	*E. caspicum caspicum*	Tu
		Absent	Eastern inner shelf	>35	13	11	18	*A. caspica*, 66 *E. caspicum caspicum*, 22	*E. shohinae*	SC-3
			Eastern outer shelf	36–70	13.1–13.8	11	3	*A. caspica*, 58	*M. fusca*	SC-4

Addendum: Faunal reference list of benthic foraminiferal species included in the text and Tables 1 and 2 (in alphabetical order).

***Ammobaculites ponticus* Mikhalevich** (1968:15, pl. I, fig. 4). Yanko (1979: pl. 24A, fig. 6; 1982: pl. V, fig. 3; 1989:8–10, pl. 1, figs 7, 8). Yanko and Troitskaya (1987:14, pl. 1, fig. 1).

***Ammonia ammoniformis* (d'Orbigny)** = *Rotalia* (turbinuline) *ammoniformis* d'Orbigny (1826:174, pl. 12, fig. 149). Yanko and Troitskaya (1987:42, pl. X, figs 1–10). Yanko (1989:169–177, pl. XXVII, figs 1–12; pl. XXVIII, figs 1–9; pl. XXIX, figs 1–5).

***Ammonia agoiensis* Yanko** (1990b:24, pl. I, fig. 1; pl. II, fig. 1; 1989:167–169, pl. XXIX, fig. 6).

***Ammonia caspica* Shchedrina** = *Ammonia beccarii caspica* Mayer (1968:28, fig. 48) = *Ammonia neobeccarii caspica* Shchedrina (Shchedrina and Mayer 1975:255, figs 1–8). Yanko (1989:177–179, pl. XXX, figs 1–9).

***Ammonia caucasica* Yanko** (1990b:25, pl. I, figs 2, 3; pl. II, fig. 5; 1989:179, pl. XXXI, figs 1–6; pl. XXXII, figs 1–12).

***Ammonia compacta* (Hofker)** = *Streblus compactus* Hofker (1969:99, figs 242, 243) = *Ammonia neobeccarii pontica* Yanko (1979:82, pl. 24Б, fig. 2). Yanko and Troitskaya (1987:44, pl. XI, figs 1–10). Yanko (1989:182–185, pl. XXXIII, figs 1–9).

***Ammonia novoeuxinica* Yanko** (1979: pl. 24Б, fig. 1; 1990b: pl. 2, fig. 7; 1989:185–187, pl. XXXIV, figs 1–11). Yanko and Troitskaya (1987:46, pl. XII, figs 1–3).

***Ammonia parasovica* Shchedrina and Mayer** (1975:255, pl. 2, figs 4–6). Yanko and Troitskaya (1987:47, pl. XII, figs 4–6). Yanko (1989:187, 188, pl. XXXV, figs 1–10; pl. XXXVI, figs 1–9).

***Ammonia tepida* (Cushman)** = *Rotalia beccarii* (Linnaeus) var. *tepida* Cushman (1928:79, pl. 1). Yanko and Troitskaya (1987:48, pl. 12, figs 7–12). Yanko (1989:192–195, pl. XXXVIII, figs 1–9; pl. XXXIX, figs 1–9).

***Ammoscalaria verae* (Mayer)** = *Ammotium* (?) *verae* Mayer (1968:21, fig. 40). Yanko (1989:12, 13, pl. II, fig. 3).

***Aubignyna perlucida* (Herron-Allen and Earland)** = *Rotalia perlucida* Herron-Allen and Earland (1913:139, pl. 13, figs 7–9). Yanko (1979: pl. 24Б, fig. 5; 1982: pl. 4, fig. 1; 1989:226–229, pl. XLVIII, fig. 3; pl. XLIX, figs 1–4). Yanko and Troitskaya (1987:36, pl. VII, figs 6–9; pl. VIII, fig. 1).

***Birsteiniolla macrostoma* Mayer** (1974:25, fig. 19). Yanko (1989:17–19, pl. III, fig. 1).

***Canalifera nigarensis* (Cushman)** = *Elphidium nigarense* Cushman (1939:63, pl. 17, fig. 19). Yanko and Troitskaya (1987:50, pl. XIII, figs 1–5). Yanko (1989:197–200, pl. XL, figs 6–8; pl. XLI, figs 1–3).

***Canalifera parkerae* (Yanko)** = *Cribroelphidium parkeri* Yanko (1974:24, pl. 1, fig. 1) = *Nonion* sp. B (Parker 1958:191, pl. 1, figs 40, 41). Yanko and Troitskaya (1987:51, pl. XIV, figs 1–6). Yanko (1989: 201–204, pl. XLII, figs 1–9; pl. XLIII, fig. 1).

***Cornuspira minuscula* (Mayer)** = *Cyclogyra minuscula* Mayer (1972:33, fig. 5). Yanko (1982: pl. 5, fig. 5; 1989:25, 26, pl. IV, fig. 1).

***Cribroelphidium poeyanum* (d'Orbigny)** = *Polystomella poeyana* d'Orbigny (1839:55, pl. 6, figs 25, 26). Yanko (1979:84, pl. 24Г, fig. 2; 1982: pl. 1, fig. 4; 1989:262, pl. LXI, figs 1–6; pl. LXII, fig. 1). Yanko and Troitskaya (1987:58, pl. XXI, figs 4–6). Yanko *et al.* (1998: pl. 1, fig. 15).

***Eggerella scabra* (Williamson), 1858** = *Bulimina scabra* Williamson (1858:604–605, pl. 5, figs 136, 137). Yanko (1989:22–24, pl. III, figs 5–7).

***Elphidium caspicum azovicum* Yanko** (Yanko 1989:243–246, pl. LIII, figs 3–7; pl. LIV, figs 1–4).

***Elphidium caspicum caspicum* Yanko** (1989:242–243, pl. LIII, figs 1, 2) = *Elphidium littorale caspicum* Mayer (1968:31, fig. 50) = *Elphidium caspicum* (Yanko and Troitskaya 1987:55, pl. XV, fig. 4).

***Elphidium ponticum* (Dolgopol'skaya and Pauli)** = *Elphidium advenum* var. *pontica* Dolgopol'skaya and Pauli (1931:36, pl. III, fig. 14) = *Elphidium ponticum* (Mikhalevich 1968:19, pl. 6, fig. 2). Yanko (1979: pl. 24Г, fig. 1; 1982: pl. 1, fig. 1; 1989:254–257, pl. LVII, figs 1–4; pl. LVIII, figs 1–4). Yanko and Troitskaya (1987:56, pl. XVI, fig. 3; pl. XVII, fig. 1–3).

***Elphidium shohinae* Mayer** (1968:32, fig. 51; 1974:34, fig. 26). Yanko (1989:257–258, pl. LIX, figs 4, 5).

***Entolingulina deplanata* Yanko** (1979: pl. 24A, fig. 4; 1982:130, pl. III, fig. 5; 1989:113–114, pl. XV, figs 5–7). Yanko and Troitskaya (1987:27, pl. IV, figs 7, 8).

***Esosyrinx jatzkoi* Yanko** (1974:28, fig. 3; 1979: pl. 24A, fig. 7; 1989:107–108, pl. XIV, figs 4–6). Yanko and Troitskaya (1987:26, pl. IV, figs 3, 4).

***Fissurina lucida* (Williamson)** = *Entosolenia marginata* (Montagu) var. *lucida* Williamson (1858:17, pl. 2, fig. 17). Voorthuysen (1973:46, pl. 5, fig. 9). Yanko and Troitskaya (1987:30, pl. V, figs 1–12). Yanko (1989: 122–125, pl. XVI, figs 6–17).

***Fissurina solida* Seguenza** (1862:56, pl. 1, fig. 42) = *Fissurina* ex gr. *solida* (Yanko and Troitskaya 1987:33, pl. VI, figs 6–9). Yanko (1989:128-129, pl. XVII, figs 3–10).

***Haynesina anglica* (Murray)** = *Protelphidium anglicum* Murray (1965:149, pl. 25, figs 1–5). Yanko and Troitskaya (1987:54, pl. XX, figs 1–3). Yanko (1989:232–235, pl. L, figs 1–7; pl. LI, figs 1–6).

Jadammina polystoma dacica Tufescu (1973:28, pl. I, fig. 2a-b).

***Lagena vulgaris* Williamson** (1858:3, pl. 1, fig. 5). Yanko (1979: pl. 24A, fig. 2; 1982: pl. 3, fig. 6; 1989:101–103, pl. XIII, figs 12–14). Yanko and Troitskaya (1987:25, pl. III, figs 13, 14).

***Laryngosigma williamsoni* (Terquem)** = *Polymorphina lactea* var. *oblonga* d'Orbigny (Williamson 1958:71, pl. 6, fig. 149) = *Polymorphina williamsoni* Terquem (1878:37). Mikhalevich (1968:18, pl. V, fig. 3). Yanko (1982: pl. III, fig. 7; 1989:111–113, pl. XV, figs 3, 4). Yanko and Troitskaya (1987:27, pl. IV, figs 5, 6).

***Massilina secans* (d'Orbigny)** = *Quinqueloculina secans* d'Orbigny (1826:303, pl. 43, fig. 96). Yanko (1982: pl. 3, fig. 2; 1989:65–68, pl. X, fig. 5). Yanko and Troitskaya (1987:22, pl. II, fig. 9).

***Mayerella brotzkajae* (Mayer)** = *Elphidiella* (?) *brotzkajae* Mayer (1968:33, fig. 52; 1974:35, fig. 28). Yanko and Troitskaya (1987:60, pl. XXII, figs 1–3; pl. XXIII, figs 1–4). Yanko (1989:272–274, pl. LXIV, figs 2–4; pl. LXV, figs 1–4; pl. LXVI, fig. 2).

***Mayerella kolchidica* Yanko** (1989:274–275, pl. LXVI, fig. 3) = *Mayerella* ex gr. *brotzkajae* (Yanko and Troitskaya 1987:61, pl. XXIV, fig. 2).

***Miliammina fusca* (Brady)** = *Quinqueloculina fusca* Brady (1870:286, pl. XI, figs 2, 3). Mayer (1968:23, fig. 41; 1974:23, fig. 18). Yanko (1989:15–17, pl. II, fig. 5).

***Nonion matagordanus* Kornfeld** = *Nonion depressulus* (Walker and Jacob) var. *matagordana* Kornfeld (Cushman 1939:21, pl. 5, figs 23–25). Yanko (1979: pl. 24Б, fig. 4; 1982: pl. IV, fig. 2; 1989:154–157, pl. XXIV, figs 1–5). Yanko and Troitskaya (1987:40, pl. 9, figs 7–9).

***Parafissurina dzemetinica* Yanko** (1979: pl. 24A, fig. 3, 1982:130, pl. 5, fig. 7; 1989:132–134, pl. XIX, figs 6–13). Yanko and Troitskaya (1987:34, pl. VII, figs 6–12).

***Parafissurina lateralis* Cushman** = ***Parafissurina* ex gr. *lateralis* Cushman** (Yanko and Troitskaya 1987:35, pl. VI, figs 13–15). Yanko (1989:134–137, pl. XIX, figs 14–16).

***Porosononion martkobi ponticus* Yanko** (1989:210–214, pl. XLIV, figs 1–4) = *Nonion martkobi* Bogdanovich (1947:30, pl. IV, fig. 4a-c) = *Nonion stelligerum* Dolgopol'skaya and Pauli (1931:31, pl. 3, fig. 12a, b) = *Porosononion martkobi* (Yanko and Troitskaya 1987:52, pl. 18, figs 1–4) = *Protelphidium martkobi* (Yanko 1979: fig.24Г, fig. 3; 1982: pl. 2, fig. 2) = *Cribroelphidium martkobi* (Mikhalevich 1968:20, pl. 7, fig. 1).

***Porosononion martkobi tschaudicus* Yanko** (1989:215–218, pl. XLV, figs 1, 2).

***Pyrgo elongata* (d'Orbigny)** = *Biloculina elongata* d'Orbigny (1826:298, fig. 4). Yanko and Troitskaya (1987:21, pl. 2, figs 5, 6). Yanko (1989:70–72, pl. XI, figs 1, 2).

***Quinqueloculina bicornis* (Walker and Jacob)** = *Serpula bicornis* Walker and Jacob (1978:633, pl. 14) = *Quinqueloculina* ex gr. *bicornis* (Yanko and Troitskaya 1987:17, pl. 1, fig. 9). Yanko (1982: pl. III, fig. 4; 1989:3–34, pl. IV, figs 5–10).

***Quinqueloculina laevigata* (d'Orbigny)** = *Triloculina laevigata* d'Orbigny (1826:134, pl. IV, fig. 1) = *Quinqueloculina laevigata* (Cushman 1922:65, pl. 13, fig. 2). Mikhalevich (1968:17, pl. III, fig. 2). Yanko and Troitskaya (1987:18, pl. 1, fig. 10). Yanko (1989:42–45, pl. VI, figs 6–9; pl. VII, figs 1, 2).

***Quinqueloculina seminulum* (Linnaeus)** = *Serpula seminulum* Linné (1767:1264, fig. 1) = *Quinqueloculina seminulum* (Williamson 1858:86, pl. 7, figs 183–185) = *Quinqueloculina pseudoseminula* (Mikhalevich 1968:17, pl. 4, fig. 1). Yanko (1979: pl. 24A, fig. 1; 1982: pl. 3, fig. 1). Yanko and Troitskaya (1987:20, pl. 2, figs 3, 4). Yanko (1989:55–57, pl. VIII, figs 5–9).

***Spiroplectinata perexilis* Mayer** (Mayer 1968:24, fig. 44). Yanko (1989:21–22, pl. III, fig. 4).

***Trichochyalus aguajoi* (Bermúdez)** = *Discorbis aguajoi* Bermúdez (1935:204, pl. 15, figs 10–14) = *Discorbis instans* Mayer (1968:26, fig. 46). Tufescu (1974: pl. IV, fig. 19). Yanko (1989:163–165, pl. XXVI, figs 5, 6).

8

ON THE POST-GLACIAL CHANGES IN THE LEVEL OF THE BLACK SEA

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Abstract: This paper summarizes the lithological and stratigraphic investigations that have been conducted on Late Quaternary sediments recovered within more than 300 cores removed from the Black Sea shelf. Using mollusc assemblage studies and radiocarbon dates, the Neoeuxinian, transitional, Old and New Black Sea horizons were identified, and their areas of extension defined. The studies confirm that during the Neoeuxinian period (from 17–15 until 8.5–8.0 ky BP), sea level rose to an elevation of –25 to –30 m. During the transitional period (8.0–6.8 ky BP), which marked the change from freshwater floral and faunal assemblages to those typical of a marine environment, sea level reached –25 to –20 m, and later, it attained present-day level. During the entire post-glacial period, the rise in Black Sea level exhibited a gradual character.

Keywords: Black Sea shelf, Late Quaternary sediments, Post-glacial transgression

1. INTRODUCTION

In recent years, the problems surrounding the character and rate of post-glacial rise in the Black Sea and the influx of Mediterranean Sea water have been intensely discussed in scholarly publications. Based on the study of several cores and seismic-acoustic profiles in limited areas of the northwestern Black Sea shelf and Kerch Strait, Ryan *et al.* (1997) proposed a catastrophic flooding of the shelf by Mediterranean saline water that led to an abrupt sea-level rise at *ca.* 7 ky BP akin to that of the Biblical Flood. Several scholars (Ballard *et al.* 2000; Major *et al.* 2002) have supported this hypothesis, while others (Aksu *et al.* 2002a, b, c; Demirbağ *et al.* 1999; Görür *et al.* 2001; Hiscott and Aksu 2002; Hiscott *et al.* 2002; Kaminski *et al.* 2000; Yanko-Hombach *et al.* 2002) have strongly

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disagreed, arguing for a gradual rise in Black Sea level during the course of the post-glacial transgression (Arkhangel'sky and Strakhov 1938; Nevessky 1967; Shcherbakov *et al.* 1976, 1978; Fedorov 1978). Publications of the present writers were quoted by several scholars who argued in favor of the catastrophic flooding, and for this reason, we consider it necessary to outline our position on the matter using all available data.

2. MATERIALS AND METHODS

More than 700 cores containing Late Quaternary bottom sediments were obtained by the Laboratory of Marine Geology of Moscow State University during the course of its field studies. Over 300 of these contain shelf sediments extracted from the southwestern, northern, northeastern, and eastern areas of the Black Sea (Figure 1). Materials from numerous survey cores taken in the same areas were also used (Kuprin and Sorokin 1982; Kuprin *et al.* 1984a, b; Shnyukov 1985). Detailed biostratigraphic and lithological analyses were performed on samples from these cores. Stratigraphic subdivision of the deposits, originally advanced by Arkhangel'sky and Strakhov (1938), includes Neoeuxinian, Old Black Sea (with the transition horizon at its bottom), and New Black Sea layers, of which the boundaries are defined by bivalve mollusc assemblages. ^{14}C dates were obtained for the deposits in several cores.

Figure 1. Location of the studied cores: **A**. Turkish, Bulgarian and Romanian sectors; **B**. Northwestern, Crimean, and Kerch-Taman sectors. Hatched line = the shelf edge.

2.1 Main Features of the Study Area

The main relief features of the investigated areas of the Black Sea shelf are determined by bottom morphology and position within the shelf structure. From Cape Kaliakra to the Bosphorus, shelf width varies from 30 to 70 km. The northwestern shelf, located within stable platform-type structures, exhibits the maximum width (reaching 250 km) as well as a gentle gradient. The outer shelf edge, eroded by the upper reaches of numerous underwater valleys and canyons on the continental slope, generally lies at a depth of 130–180 m. Judging from seismic-acoustic evidence, part of this edge consists of marine terraces formed during periods of low sea level.

The shelf consists of a low plain with an evenly undulating surface that was formed by accumulation and denudation over the course of numerous shoreline migrations. The smooth shelf just beyond the folded mountains of Crimea is generally narrow (less than 5–10 km) and steep. Its edge lies at 100–150 m, where a structural unconformity with the continental slope can be observed. In the eastern direction, the shelf widens to 50 km, and its edge is consistently found at a depth of 100 m.

2. RESULTS

The composition of the shelf sediments in several key areas is outlined below.

2.1 The Southern Crimean Shelf

Due to their extensive thickness, the Neoeuxinian, transitional, and Black Sea sediments were fully recovered only in the central and outer zones of the shelf, at depths exceeding 60 m. Compact clay, presumably of Pliocene age, was found at the bottom of several cores in the eastern area (Figure 2). The clay is overlaid by Neoeuxinian sediments with the fragmented remains of *Dreissena polymorpha* (Pall.), *D. rostriformis* Andrus., and *Monodacna caspia* (Eichw.). In shallow areas, the lower deposits consist of inequigranular sand with an admixture of coquina and, rarely, pebbles (as in the core 219). Above this follows a thin (< 18 cm) transitional horizon of predominantly fine-grained mud (with an admixture of coquina, as in core 211) and a mixed assemblage of molluscs: *Dreissena polymorpha*, *D. rostriformis*, *Monodacna caspia*, *Cerastoderma glaucum* (Poir.), *Mytilaster lineatus* (Gm.), and *Mytilus galloprovincialis* (Lam.).

The top of the sequence consists of thick silty-clayey mud with Mediterranean-type mollusc fauna: *Mytilus galloprovincialis* in the Old Black Sea and *Modiolus phaseolinus* (Phil.) in the New Black Sea layers. In cores 215 and 216, the Old Black Sea mud is rich in organic matter, acquiring the character of shelf sapropel.

Farther west, in the Yalta and Laspi area, the Neoeuxinian pebbles and

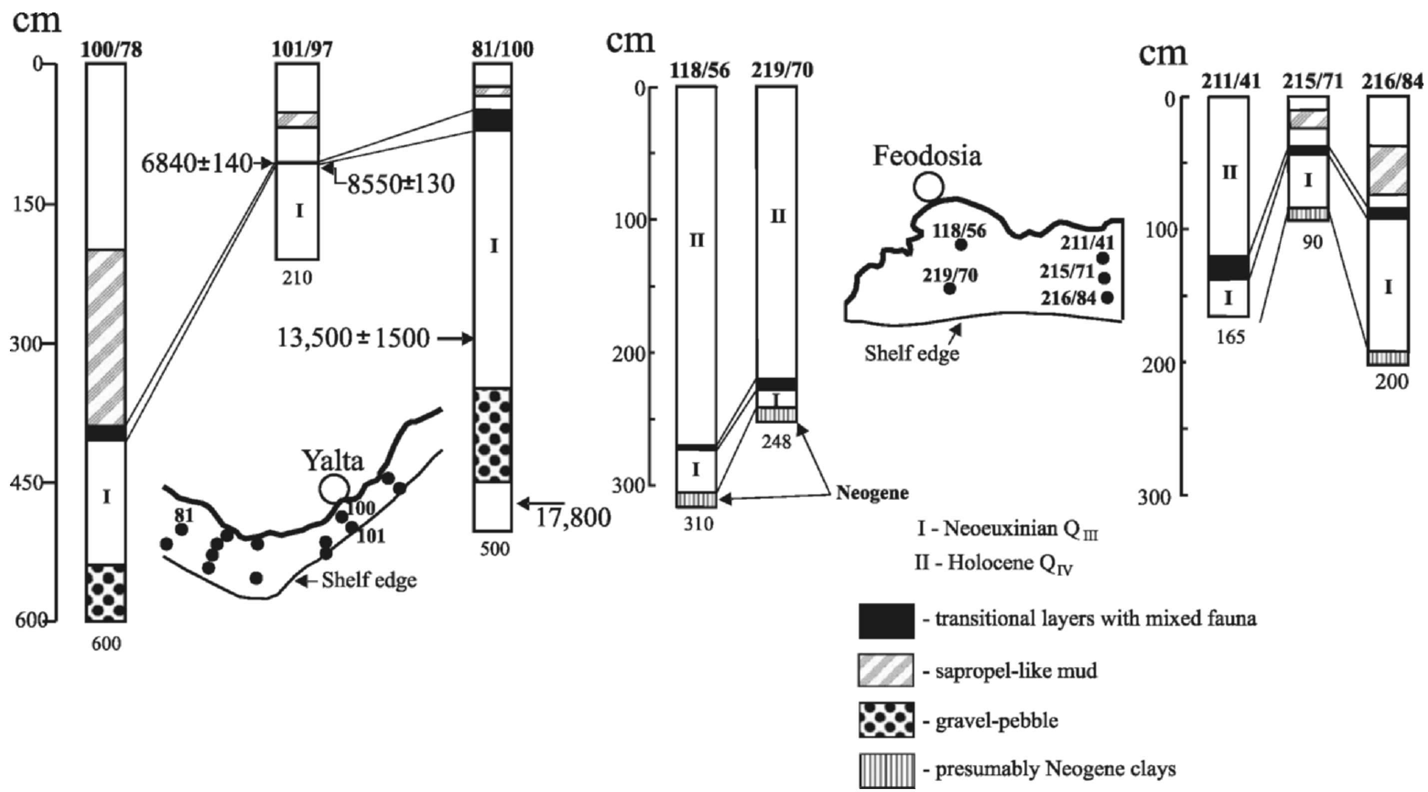

Figure 2. Sequences of Late Quaternary sediments on the Crimean (left) and Kerch (right) shelf.

gravel were found at the depth of 80–100 m; their present-day analogues are found in bottom sediments at depths less than 5 m (Figure 2). The ^{14}C dates show that, in the area of core 81, these deposits formed between 17 and 15 ky BP. In areas 90–100 m deep, the upper Neoeuxinian sediment sections consist of finer silty-clayey mud with the corresponding molluscan fauna. Landwards, the sand fraction increases. Radiocarbon dates for *Dreissena* shells from the upper Neoeuxinian layer yielded an age of 8550±130 BP (Shcherbakov *et al.* 1978).

The Neoeuxinian sediments are universally overlain by a layer, up to 22 cm thick, that includes a mixed marine and freshwater fauna. In core 101, the interval between 105 and 110 cm was radiocarbon-dated to 6840±140 BP.

The overlying Old and New Black Sea sediments with fragmentary marine shells, contain an organic-rich interval which may be attributed to the sapropel facies. These sediments are found both directly overlying the transitional layer with a mixed fauna, and at various distances from it.

3.2 The Northwestern Shelf

The wide northwestern shelf is covered by a thin drape of Late Quaternary sediments. Significantly, marine sediments overlie both continental deposits and Neogene sediments of various origins.

Over wide areas, the Neoeuxinian deposits consist of coarse-grained coquina, biogenic-detritus sand with mollusc shells, and, rarely, silt. Their thickness on the shelf's outer edge is usually 10–12 cm (Figure 3). The spread of these deposits is delimited by the depth contours of 25 and 30 m. They wedge out landward. A zone of submerged late Neoeuxinian river valleys with distinct traces of a step is acknowledged at –29 to –19 m (Morgunov *et al.* 1973). There, Holocene deposits often lie above the Pleistocene continental sediments. In some areas, at a lesser depth, lagoon and prodelta deposits have been found with remains of freshwater fauna: *Monodacna caspia* (predominant), *Dreissena* sp., *Viviparus*, and *Limnea*.

Thin transitional layers with both freshwater and marine molluscs lie above. They are also found in a wider area and at depths of –22 to –25 m. Sediments of a younger age contain characteristically marine mollusc assemblages, identifying a gradual salinization. The Old Black Sea mud and coquina include predominantly *Mytilus galloprovincialis*, *Abra ovata* (Phil.), *Cardium* sp., *Mytilaster lineatus*, and others. The New Black Sea Mud in the area of the outer edge includes *Modiolus phaseolinus*, *Spisula subtrancata* (Costa), and a variegated pelecypod and gastropod assemblage in a shallow area. As has been previously mentioned, sediments with a marine fauna in areas less than 25 m deep directly overlie either continental, lagoonal, or deltaic deposits. This is the area of offshore inflection, or slope change seaward of the beach; it is of a

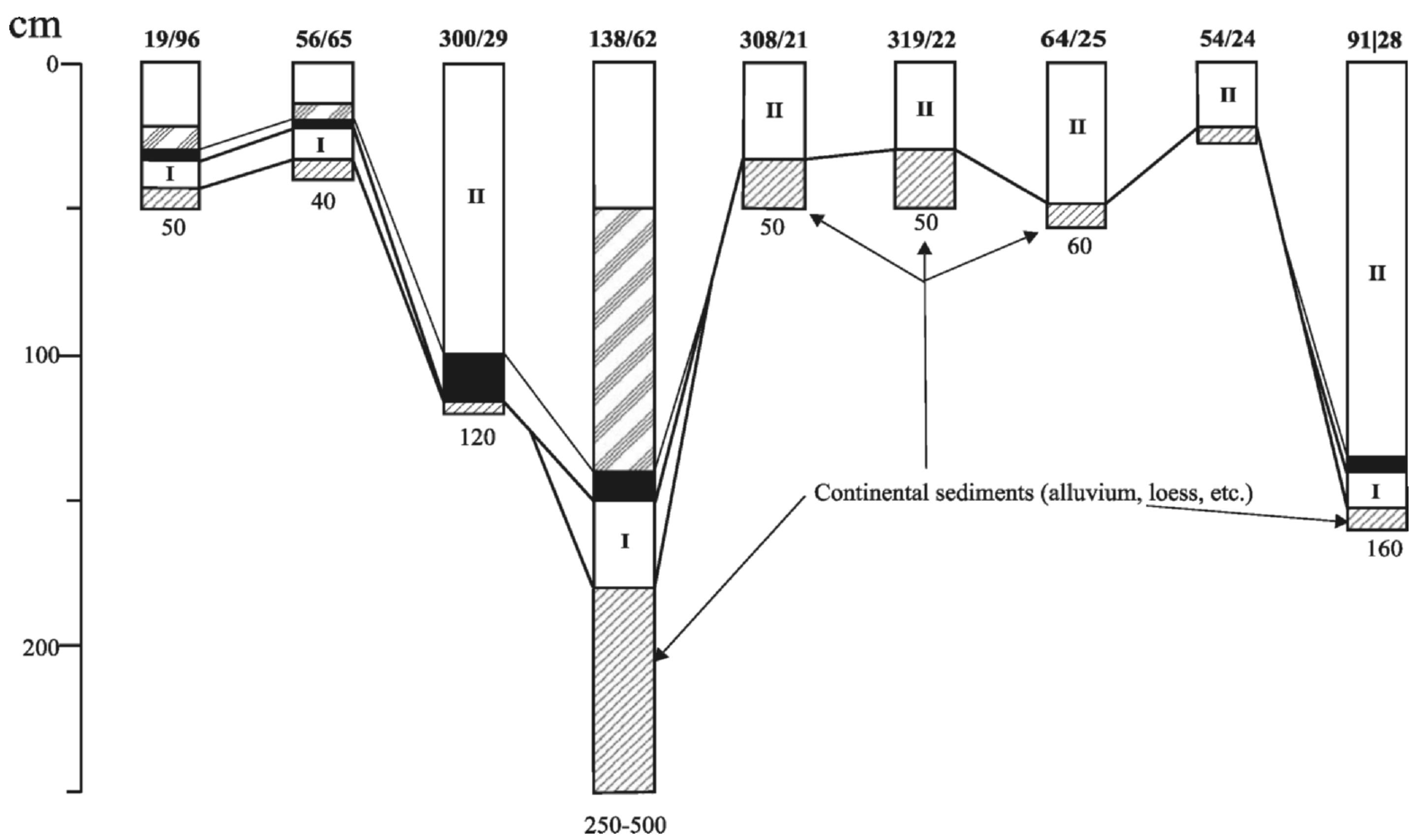

Figure 3. Sequences of Late Quaternary sediments on the Northwestern shelf. For core locations, see Figure 5; for key, see Figure 2.

variable width, where the drift from major rivers–the Danube, Dniester and Dnieper–was deposited.

3.3 The Western Shelf

In the central and southern areas of the Bulgarian shelf, recovered cores contain Late Pleistocene and Holocene deposits at depths between 90 and 120 m (Figure 4). The sequences are fairly similar to those found in other areas of the Black Sea. Survey coring off the Bulgarian coast has revealed the considerable thickness of the sediments of this age (Kalinin *et al.* 1984; Kuprin *et al.* 1984a, b).

The composition of the Neoeuxinian sediments between Capes Kaliakra and Emine (Figure 4, cores 52, 55, 56, 2) changes from a fine-grained (core 2) to a sandy fraction with substantial shells and shell detritus (core 52), and a silty or silty-clayey mud with mixed freshwater and marine fauna.

The sediments with Mediterranean-type fauna consist predominantly of clayey mud with mollusc shells. As in other areas, the Old Black Sea sapropel-like deposits vary in thickness. They have been found both directly overlying the transitional layer and separated some distance from it.

The composition of the sediments in the areas facing Bourgas Bay is similar (Figure 4, cores 60, 62, 63). It should be noted that a large amount of gravel and small-sized pebbles was found in the New Black Sea detrital sand recovered in core 62 below 191 cm.

Farther south (cores 83-86), the Chaudinian clay was recovered beneath the Neoeuxinian horizon (Kuprin *et al.* 1980; Kuprin 1988). The Neoeuxinian sediments proper consist of thin layers of biogenic-detritus sand with gravel and small-sized pebble, numerous *Dreissena* and *Monodacna* shells, and redeposited Chaudinian molluscs.

The transitional horizon is usually thin, its thickness increasing landwards. The Old and New Black Sea clayey mud is identifiable based on its typical molluscan assemblage. The sapropel-like clayey horizon directly overlies the transitional layer and makes up a considerable part of the Holocene sequence.

Radiocarbon dates have been obtained from several Bulgarian shelf cores (Nikolaev *et al.* 1980; Kuptsov *et al.* 1979) (Table 1). The quoted dates suggest considerable erosion of the upper part of the Neoeuxinian deposits between 11 and 9 ky BP. The most recent Neoeuxinian deposits are about 8.5 ky BP. The age of the transitional horizon with the fauna of mixed Caspian-Mediterranean type is 7.5 to 7 ky BP. Radiocarbon dates from the Old Black Sea horizon lie within the time span of 6.8 to 3.5 ky BP. The age of the New Black Sea horizon is less than 3.5 ky.

The above-outlined sediment sequences are basically similar to those of

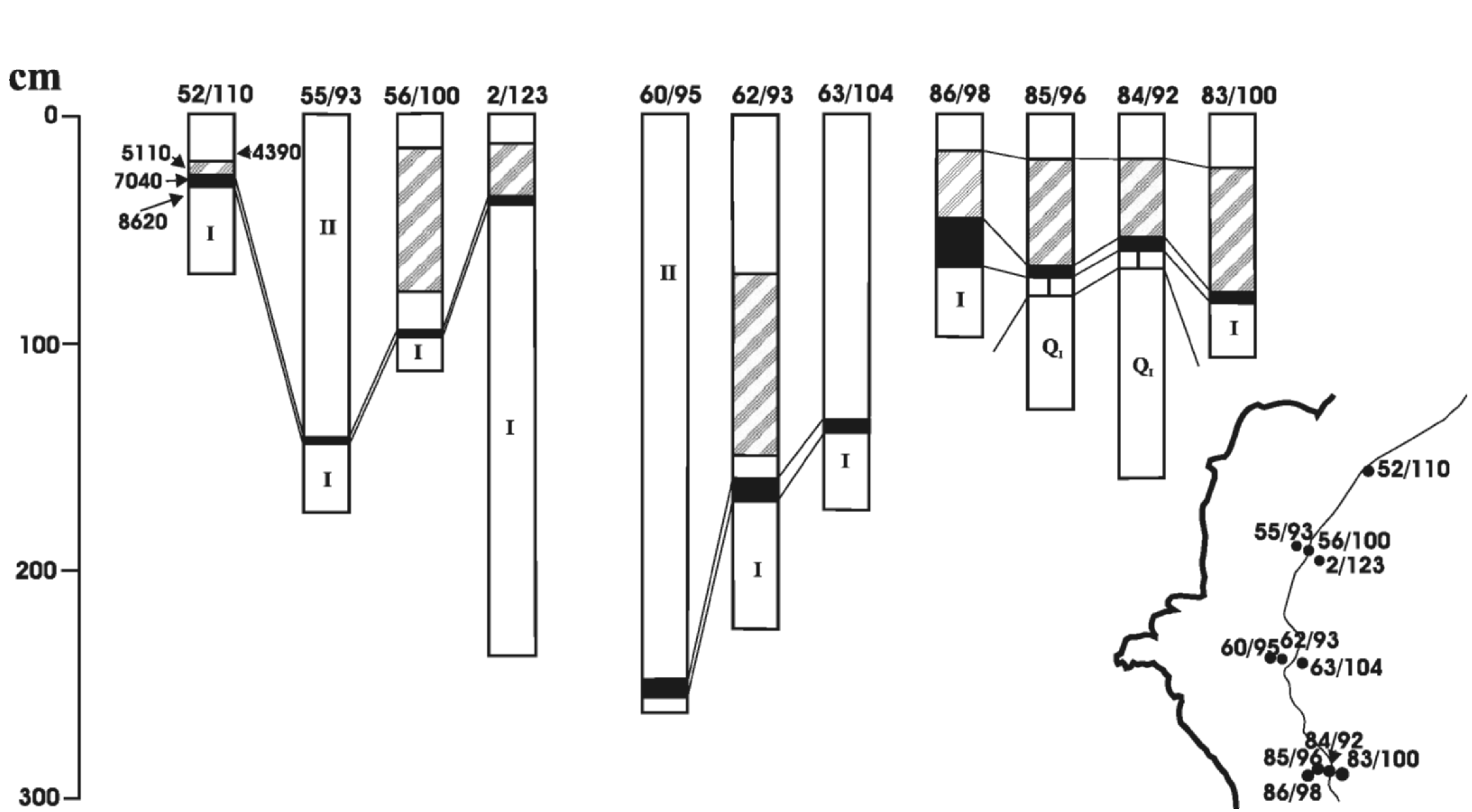

Figure 4. Sequences of Late Quaternary sediments on the Bulgarian shelf. For core locations, see Figure 5; for key, see Figure 2.

other studied areas of the Black Sea, namely, the extensive Kerch-Taman shelf, the Caucasus shelf stretching from Tuapse to Batumi, the outer shelf off the coast of northwestern Turkey, and the Romanian shelf.

Table 1. ^{14}C dates for shelf sediments.

Core no.	Sea depth (m)	Sample depth (cm)	^{14}C dates	Geologic Age
81	100	465–470	17,780±200	Neoeuxinian
81		290–300	13,500±1500	
62	93	140–150	17,180±300	
62		90–100	11,430±330	
2345	122	120–125	11,590±240	
101	97	110–115	8550±130	Neoeuxinian, top level
52	110	32–35	8620±70	
101	97	105–110	6840±140	Transitional
52	110	27–30	7040±200	
2362*	102	70–90	7480±540	
2345	122	85–95	6880±260	Old Black Sea, bottom level
2362*	102	60–70	6890±630	Old Black Sea
62	93	55–65*	6800	
52	110	20–27	5110±70	
52		13–20	4390±65	New Black Sea, top level
62	93	10–20	3450	New Black Sea

*according to Kuptsov *et al.* (1979).

4. DISCUSSION

The chronology of water level changes in the Black Sea since the last Ice Age has been established using the lithological and paleontological evidence summarized above, as well as radiometric dates.

Based on the finds of littoral gravelly-pebbly deposits at the outer edge of the southern Crimean shelf, the level of the Neoeuxinian Sea during the interval 17 to 15 ky BP (i.e., during the maximum stage of the last Ice Age regression) may be estimated as about –100 m. These deposits are both underlain and overlain by more fine-grained sediments accumulated in the environment of a deeper sea that preceded and followed the lowest layer of a freshwater basin.

The regular change in grain-size composition upward through the sequences of Neoeuxinian sediments on the outer shelf, from the coarse-grain fractions to the fine-grained silty-clayey muds, suggests the gradual landward migration of the shoreline. As shown above for the easternmost area studied (Figure 2), Neoeuxinian deposits lie directly on the supposedly Neogene clay within the interval between 56 and 84 m. In that eastern area and on the Romanian shelf, cores taken at the depth of 40–65 m show the thin (less than 5 cm) biogenic-detritus Neoeuxinian sand directly overlying either the Neogene or continental Pleistocene sediments. Consequently, our materials fail to provide evidence of a repeated and considerable fall of the Black Sea to the level of –120 to –150 m, as suggested by Ryan *et al.* (1997) and Ballard *et al.* (2000).

In this respect, one should note that while the basin remained at the level of *ca.* –100 m, solid discharge from the rivers was deposited at the edge of the northwestern shelf, in the upper portion of the continental slope, where the prodeltas and alluvial fans were formed. Therefore, the alluvial deposits found by Ryan *et al.* (1997) at the depth of –140 m cannot be viewed as proof of a shoreline occurring at that depth. Neither can the pebbly sediments found by Ballard *et al*. (2000) in the southern area of the Turkish Black Sea shelf at –155 m prove the pre-existence of a beach. If one scrutinizes the graphs in this paper, one can easily see on Figure 4 that the pebbles include shells of *Modiolus phaseolinus*, which populated the Black Sea at about 3000 BP. It seems probable that these sediments resulted from the clastic torrents in the upper reaches of the canyons, commonly observable on the Caucasian continental slope in the upper stretches of the Kodori, Bzyb', and other submerged rivers (Kuprin *et al.* 1985; Solov'eva and Sorokin 1993). It is equally possible that the pebbles were redeposited from Early Quaternary deposits outcropping in the Sinop area, or from older sediments exposed in the littoral area and recovered by cores (Okyar *et al.* 1994).

As our evidence shows, the level of the freshwater Neoeuxinian basin rose to an elevation of –25 to –30 m, corresponding to the maximum extension of Neoeuxinian sediments on the stable northwestern shelf within the confines of the Russian platform (Figure 5 and 6). This observation is confirmed by data from cores on the southern Crimean shelf near the city of Yalta at a depth of *ca.* –30 m (Shnyukov 1985). Judging from the radiocarbon dates cited above, the Neoeuxinian sea-lake existed until 8.5–8 ky BP. The rate of uninterrupted sea-level rise may be estimated as 1 cm per year.

The Neoeuxinian deposits with semi-fresh water fauna are universally overlain by sediments with fauna of transitional type. Its extension from the coast side is limited by the 20–25 m contours. Based on the available radiocarbon measurements, the period of accumulation for these sediments lies between 8550±130 and 6840±140 BP.

At the start of this period, *ca.* 8 ky BP, Mediterranean water penetrated into the Black Sea basin, leading to its salinization and fundamental change in its flora and fauna. Sea level rose from –25/–30 to –25/–20 m, since transitional deposits in the cores were never found above –20 m (Figures 5 and 6). The rate of sea-level rise dropped to 0.5–1.0 cm per year.

At about 6.8 ky BP, marine deposits of Old and New Black Sea age

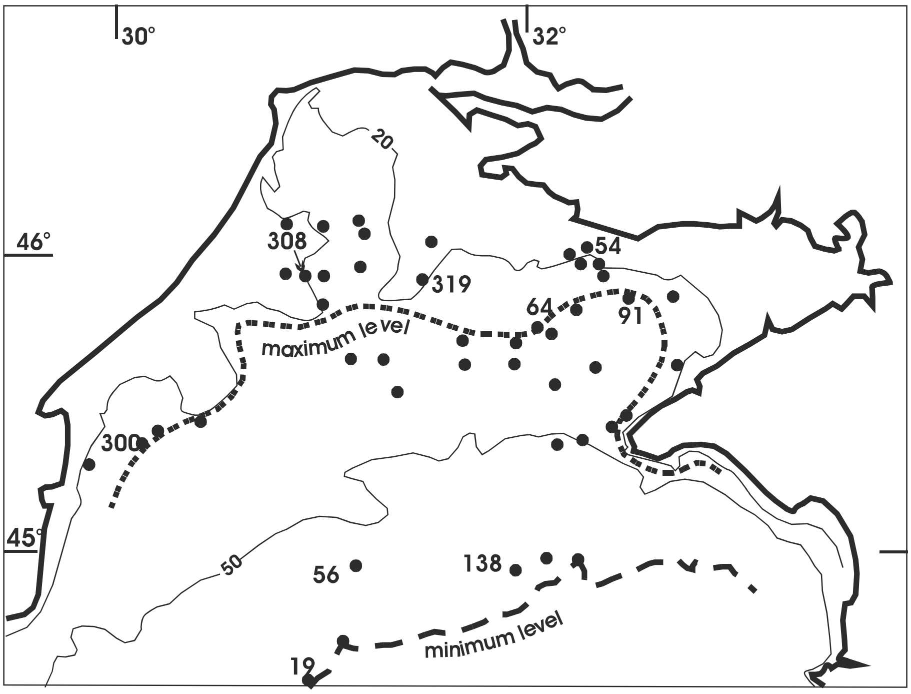

Figure 5. Location of the studied sequences and the position of the Neoeuxinian shorelines on the northwestern shelf. Hatched lines = the minimum (–100 m) and maximum (–30 m) levels.

proper started accumulating in an environment of continued Black Sea level rise, now including a two-way water exchange with the Mediterranean Sea and an average rate reduced to 0.3 cm per year.

The consecutive migration of Neoeuxinian sea-lake sediments, as well as those of the transitional and marine Old and New Black Sea phases, from the shelf's outer edge landwards indicates a gradual and progressive sea-level rise with small-scale oscillations during the post-glacial period. The hydrodynamic features reflecting these processes are observable in the morphology of the sea bottom. Elements of sea-bottom relief on the northwestern shelf include sub-latitudinal belts succeeding each other from the shelf's edge landward. At the shelf's junction with the continental slope, in water 100 to 250 m deep, canyon mouths and submerged river deltas create a belt of complex relief.

Farther north lies a band of a smoothed relief at depths ranging from *ca.* 100 to 60 m; at 70–60 m, its upper edge truncates valleys of pre-Neoeuxinian rivers. The bottom deposits consist of marine-littoral biogenic-detritus sand with alluvium in the paleoriver valleys (Kuprin 2002). A belt of intensively dissected relief with many submerged river valleys is found on the northern side at depths between 65 and 45 m. Linear steps beneath a drape of ancient alluvium have been identified at 26 to 19 m (Morgunov *et al.* 1973).

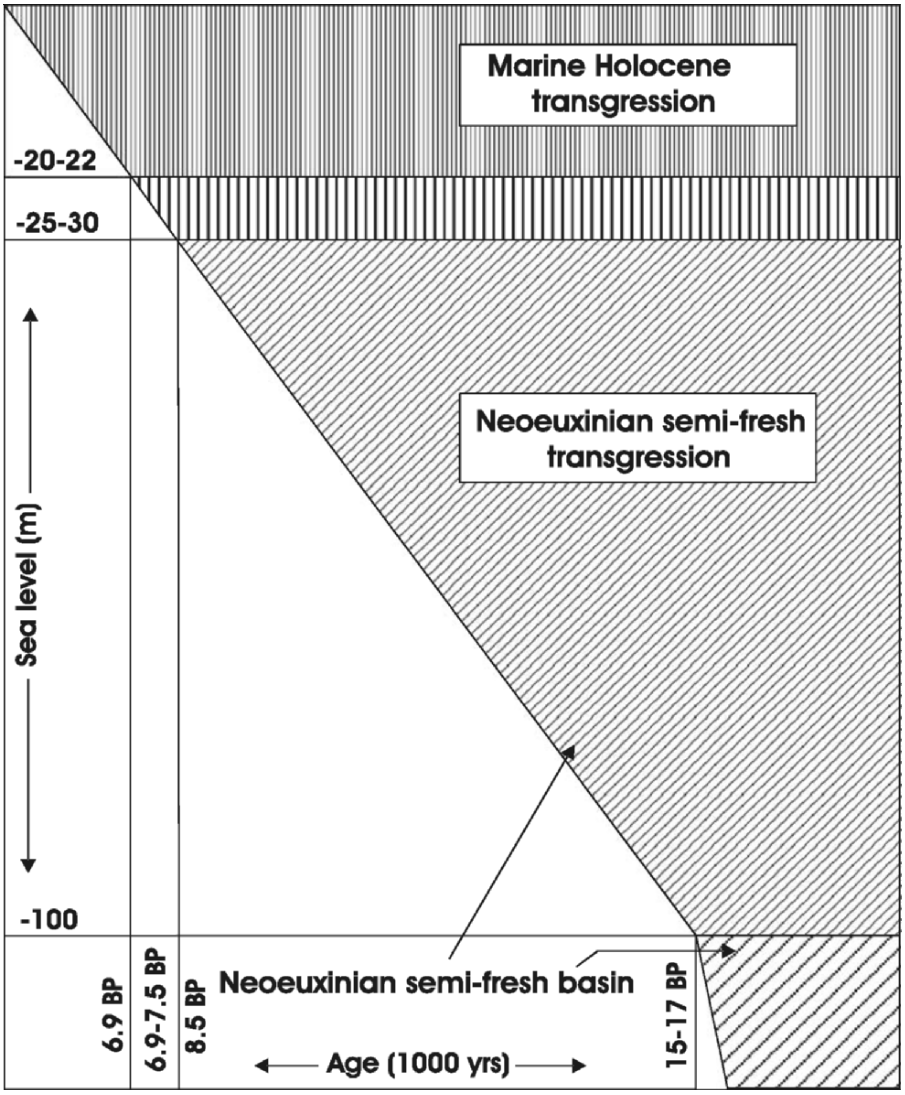

Figure 6. The Late Quaternary evolution of the Black Sea.

The relationship of the Old Black Sea Sapropel within the depression with that on the shelf needs some comment. Numerous radiocarbon dates obtained by Soviet and foreign laboratories (Ross and Degens 1974) suggest that the basal age of the deep-sea sapropel was about 7 ky BP. Earlier dates of 9 ky BP were obtained by Soviet geologists for the Bosphorus area (Vinogradov *et al.* 1962). This time limit marks the change from a freshwater diatom assemblage to a marine one. Lithologically, it represents an abrupt transition from Neoeuxinian clayey muds to sapropel. Previously, we had argued (Kuprin *et al.* 1980) that the bottom deposits of the sapropel within the depression, poor in organic matter and with thin aragonite interbeddings, corresponded to the transitional horizon.

Following from our evidence, sapropel accumulation on the shelf started after salinization of the basin, as the sapropel lies above the transitional horizon. As had been noted above, in some cases, the sapropel is separated from the transitional horizon by normal sediments, a finding that suggests its asynchronous formation on the shelf.

5. CONCLUSIONS

The available evidence strongly suggests a gradual rise in the level of the Black Sea with oscillations during the post-glacial period. The post-glacial transgression began subsequent to the maximum regression at 17–15 ky BP and synchronously with the melting and recession of the ice sheets. It proceeded in two stages. During the first stage during the later Neoeuxinian period, sea level supposedly rose from –100 to –25/–30 m. In the final Neoeuxinian stage, freshwater rose above the Bosphorus sill and spilled over into the Marmara and Mediterranean Seas.

Between 8.5 and 7 ky BP, the rising Mediterranean Sea led to a two-way exchange flow between the two basins, causing an intensive salinization of the Black Sea. The following period (that lasted 1.0–1.5 ky) saw the formation of Mediterranean-type marine fauna and a sea-level rise to an elevation of *ca.* –20 m. The marine transgression proper began around 6.8 ky BP and ultimately resulted in the present-day sea level.

In view of this evidence, it would be inadmissible to argue for a momentary catastrophic influx of saline Mediterranean water into the Black Sea depression. This assertion is possible only if one ignores the Late Quaternary transitional horizon that contains a mixed Caspian-Mediterranean fauna and accumulated over a period of 1.0–1.5 ky. One would also have to ignore all the above-cited radiocarbon dates on mollusc shells from Late Quaternary horizons that have been conducted by highly reliable laboratories of the USSR and the USA. This would also mean the rejection of reliable geological evidence for the aquatic origin of the Late Glacial Neoeuxinian sediments at depths of 25–30 m over the entire Black Sea shelf.

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9

THE POST-GLACIAL TRANSGRESSION OF THE BLACK SEA

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Abstract: This paper discusses four problems, resolution of which will help to determine a realistic scenario for the post-glacial transgression in the Black Sea: (1) the post-glacial history of possible changes in the Black Sea's own water balance, under the hypothetical condition of its isolation from the global ocean; (2) the geologic history of areas through which the possible linkage of the Black Sea with the Mediterranean and Caspian basins might have occurred; (3) the regional peculiarities of recent vertical tectonic movements in coastal and shelf areas surrounding the Black Sea depression; and (4) the history of global ice volume changes and glacio-isostatic sea-level fluctuations. It is clear that both existing evidence and theoretical concepts better substantiate the traditional model of post-glacial history in the Black Sea than the recently proposed hypothesis of an abrupt flooding of the Black Sea depression, which allegedly generated the biblical Flood.

Keywords: Black Sea, water balance, Bosphorus, recent tectonic movements, ice volume, eustatic sea-level fluctuations

1. INTRODUCTION

Establishing a comprehensive and realistic scenario that accurately describes the post-glacial transgression in the Black Sea is feasible only if consensus can be reached in the resolution of four interrelated problems. These problems involve the following matters and their dating:

(1) the Water Balance problem: the post-glacial history of possible changes in the water balance of the Black Sea considered by itself (i.e., local to the Black Sea drainage basin and under the hypothetical condition that it was isolated from the global ocean; hereafter referred to as "local Black Sea water balance"),

V. Yanko-Hombach et al. (eds.), The Black Sea Flood Question, 221-250.

(2) the Straits History problem: the post-glacial geologic history of the areas through which the Black Sea might have been linked with the Mediterranean and Caspian basins,

(3) the Regional Tectonic problem: the regional peculiarities of recent tectonic movements, including those related to geoidal deformations and isostasy, and

(4) the Ice Volume problem: the history of changes in global ice volume as well as glacio-isostatic variation in the level of the world ocean.

Ryan *et al.* (1997, 2003) and Major *et al.* (2002) have recently put forward a hypothesis suggesting an extremely rapid rise in sea level during one stage of the Black Sea's post-glacial transgression, and that this inundation could have been the basis for the biblical 'Flood'. The following paragraphs discuss existing evidence and whether it argues for or against this unorthodox model.

In theory, any of the problems noted above could have provided a mechanism to generate such a 'Flood', e.g., an unusually rapid rise in the level of the global ocean. Cases of abrupt climatic transitions are currently being discussed by Alley and Clark (1999), Behl and Kennett (1996), and others, as it has become increasingly accepted that rapid transformations are more the rule than the exception in the evolution of climate. Yet, if such a 'Flood' had been triggered by a global cause, signs should have been manifested worldwide, and not just within the Black Sea.

Other mechanisms might have produced a more geographically restricted effect. For example, one might theoretically admit that an abrupt sea-level rise in the course of the post-glacial transgression could have been caused by a rapid tectonic submergence in some areas of the Black Sea shelf, or by an abnormal–or very high amplitude–change in the Black Sea water balance in combination with an influx of Mediterranean water. It might also have been a consequence of geologic processes occurring in some areas, distinct from the Bosphorus and Dardanelles, that might have created channels through which the Mediterranean Sea could have flooded the Pontic basin.

Uncertainties embedded within the research methods employed might lie at the root of the problem here. Several writers (Shcherbakov *et al.* 1978, and others) have acknowledged geologic events occurring with abnormally high frequencies at the Pleistocene-Holocene boundary, both on a worldwide scale and within the Pontic basin. Yet, in no case did the interpretation of available evidence prompt any of these researchers to propose a catastrophic model. One has to admit that the solution suggested by the traditional interpretation cannot be viewed as either final or totally satisfactory, as many issues remain unclear. However, even if the 'Flood' hypothesis is finally rejected, analysis of the evidence from a new perspective will be highly informative, not only for paleoclimatology and paleogeography, but also for practical purposes: the improvement of long-range climate change forecasts.

2. WATER BALANCE

Water balance in the Black Sea has been discussed repeatedly for a long time (Makarov 1950 [1885]; Solyankin 1963; Ostrovsky 1982; Altman 1991; and many others). The general consensus is that the present water balance is positive, yet its quantitative assessment by various writers varies substantially (Table 1). One recent publication (Altman 1991) considers the total annual river discharge from the Black Sea catchment to be equal to a water layer of 80 cm covering the entire sea surface of 423,000 km^2. The rivers of the northwest, including the Danube (49 cm/year), the Dnieper (12 cm/year), and the Southern Bug (3 cm/year) contribute 80% of this total, while Caucasian (10 cm/year) and Turkish (6 cm/year) rivers contribute much less. Rainfall constitutes the second most important fresh water input. According to Altman's (1991) estimate, it adds the equivalent of a 56 cm deep water layer every year.

Table 1. Local Black Sea water balance estimates (after Altman 1991).

Sources	A	B	C	D	E	F
Altman 1984	80	56	95	+41	–39	2.05
Rozengut & Sitnikov 1973	96	50	90	+56	–40	2.4
Sitnikov 1972	100	35	90	+45	–55	1.82
Rojdestvenski 1971	70	60	71	+59	–11	6.36
Tixeront 1970	95	43	93	+45	–50	1.9
Solyankin 1963	82	28	78	+32	–50	1.64
Berenbeim 1960	80	28	66	+42	–38	2.11
Leonov 1960	73	54	86	+41	–32	2.28
Neumann & Roseman 1954	101	57	94	+64	–37	2.73
Bruevich 1953	83	53	83	+53	–30	2.77
Sverdrup 1942	78	57	86	+49	–29	2.69
Möller 1928	78	55	84	+49	–29	2.69
AVERAGE*	86	47	86	+47	–39	2.2

Values reflect depth of equivalent water layers in cm. Key: A = river discharge; B = rainfall; C = evaporation; D = fresh water excess [(A + B) – C]; E = evaporation-rainfall difference (B – C); F = relative magnitude of river discharge over evaporation-rainfall difference (A / |B – C|).
* Average values ignore Rojdestvenski's 1971 estimate as strongly deviating from other data.

Hence, the total fresh water input averages 136 cm/year. Estimated annual evaporation from the Black Sea surface is equivalent to a 94 cm layer, and thus, the total fresh water balance is positive. If the Black Sea were totally isolated from the world ocean, its sea level would rise at an average rate of 42 cm/year (based on an input of 136 cm and evaporative loss of 94 cm).

Though considerable discrepancies exist in the assessment of water balance by various writers (Table 1), it should be noted that, under present conditions and viewed according to an interglacial climatic model, local water balance for the Black Sea (i.e., considered independent of outside marine influences) is calculated to be positive. If one assumes that the riverine contribution of 80 cm is totally evaporated from the sea surface (as columns A and C in Table 1 are often nearly equivalent), the positive balance in the Black Sea would be exclusively due to rainfall. With no outflow via the Bosphorus, water level in the Black Sea would then rise by more than 0.5 m annually. As follows from column F, river input is usually greater than twice the difference between rainfall and evaporation, and thus, if river input were to disappear completely, evaporation would plunge the water balance into negative range (in this case, the annual deficit would be equivalent to a water layer 39 cm deep). Therefore, the positive balance sustained under present conditions could be transformed into a negative one if (1) river input is halved, (2) rainfall totally disappears, or (3) evaporation increases by 50–60%.

These data are insufficient to provide a realistic water balance estimate during the Last Glacial Maximum (LGM). Yet, some indications suggest that the Black Sea's own local water balance remained positive not only during interglacials, but during other phases of the glacial-interglacial climatic cycle as well.

(1) If one accepts that, during the LGM, local water balance within the Black Sea was negative, and the basin was in a regressive stage, the behavior of the Caspian Sea, which had a high water level at the time, is totally unclear. This paradox was discussed by Avenarius (1979), who reiterated a reasonable solution previously offered by Fedorov (1978). In this writer's view, both the Caspian and Black Seas had independent positive water balances (particularly the latter since it is located closer to the Mediterranean Sea and the Atlantic Ocean, and consequently receives more rainfall). The main difference between the Caspian and Black Seas lies in the morphology of the straits through which they were connected to the ocean. In contrast to the Bosphorus, the Manych Depression was not in active use, while outflow via the Bosphorus prevented the Black Sea from experiencing a Caspian-type transgression.

(2) If one accepts that, during the LGM, water balance within the Black Sea was negative, and the basin was isolated and regressing, it is unclear what might have caused its desalination (Avenarius 1979). According to the reliable stratigraphic scheme of Shcherbakov *et al.* (1978), the transgression maximum, and the Tarkhankutian layers in the Late Pleistocene Black Sea sequence, correspond to OIS 3. In such a case, the desalination of the Black Sea should have lasted only a few millennia. Yet, the Tarkhankutian layers include Mediterranean mollusc shells (*Cardium edule* and *Abra ovata*), which suggest high salinity and the existence of a two-layer water flow.

(3) If one accepts that, during the LGM, water balance within the Black Sea was negative, and the basin was isolated and regressing, the mechanism accounting for the erosion observable within the Pontic limans remains unexplained. The erosional downcutting reached a level of –40 to –45 m (Gozhik *et al.* 1984), after which the cut was filled with transgressive sediments of Holo-

cene age. A similar phenomenon is seen in the Kerch Strait, where downcutting reached –70 m (Shnyukov and Alenkin 1981). It is also worth mentioning that thick deltaic deposits were laid down widely over marginal areas of the north-western Pontic shelf during regressive stages throughout the Pleistocene, including the LGM (Zelinsky *et al.* 1987). All this evidence suggests that, during the LGM, rivers were sufficiently powerful to produce large-scale stream erosion. If this was the case, if evaporation did not exceed present values, and if rainfall remained at its present level, a negative water balance could not be achieved.

(4) If one accepts that, during the LGM, local water balance in the Black Sea was negative, it becomes difficult to explain the finding of marine and even oceanic diatoms reported in several brackish water Black Sea layers of Upper Pleistocene age (Shcherbakov *et al.* 1978). If the basin had a one way outflow through the Bosphorus, sporadic penetration of Mediterranean water at the initial stages of post-glacial transgression could have been possible because the difference between water levels in the Black and Mediterranean Seas is never big. In this case, even a small decrease in the positive local water balance of the Black Sea would be enough for the underflow to become stronger than the overflow. Such fluctuations in water balance could relate to the Shnitnikov cycles with a periodicity of 1850–2000 years. It is important to note that the short-lived penetration of Mediterranean water into the Black Sea basin occurred at ~16–15 ky BP (Svitoch *et al.* 1998:122). Significantly, a relatively abrupt increase in sulphur isotope (S^{34}) has been reported within the Neoeuxinian sulphates, where it rose from 3–4‰ (typical of continental freshwater run-off) to 14–15‰. In S.D. Nikolaev's view (Svitoch *et al.* 1998), this could have been caused only by the penetration of Mediterranean water into the Black Sea basin, which initially occurred at ~16–15 ky BP, after which a brief desalination took place. Judging from the sulphur isotopes, the major penetration of Mediterranean water occurred at ~13.5–13 ky BP.

(5) In the Pontic climate of the LGM, the thickness of the evaporation layer covering the Black Sea could hardly have exceeded that of the Interglacial period. Under the low temperature conditions of the surface water during the LGM , evaporation should have been correspondingly less. As follows from the reconstructed estimate of the deviation in mean annual precipitation values during the 'Late Würm cryochrone' at 18,000 ky BP (Zubakov 1986), the rainfall would have led to a higher water level. According to the quoted estimate, the greater part of the Black Sea would have experienced a positive deviation in mean annual precipitation. This means that during the LGM, the difference between rainfall and evaporation would have exceeded that of the present, and water level would have risen higher. Bearing in mind that water input could not have been totally absent at that time, one must acknowledge that a model allocating a negative local water balance to the isolated Black Sea is unrealistic.

Hence, both theoretically and based on the available evidence, one has to conclude that in no case over the last 20 ky, could the Black Sea's own local water balance have been negative.

As one may judge from the cited references, this conclusion was arrived at long ago. If the newly advanced theories related to the Late Pleistocene-Holocene history of the Black Sea contradict this assertion, they should, first and foremost, find and present arguments against a model in which the local water balance of the Black Sea, considered in isolation, is positive. Plainly according to this model, however, the Black Sea basin could not have been isolated and must have had at least a unidirectional linkage to the ocean; it was not land-locked during the LGM.

This implies certain corollaries that may be used to test the model of the positive local water balance for the Black Sea.

First, one has to admit that at no time could the Black Sea level lie above that of the global ocean, neither at the LGM, nor at the interglacial climatic optima, and still less so during the period of rapid ice-sheet melting.

Second, with a positive local Black Sea water balance, and taking into account the glacio-eustatic fall in the level of the Black Sea following that of the ocean, the uninterrupted outflux of surplus Black Sea water via the Bosphorus was unable to erode its channel bed to the level predicted by water current dynamics under conditions of maximum oceanic regression. The corresponding erosional downcutting should have been on the order of ~100 m (accepting that the Black Sea regression reached –80 to –90 m). One should remember that P.V. Fedorov had estimated an erosional downcutting of that magnitude long before the geology of the Bosphorus became known in any detail.

Third, the rate of the Black Sea glacio-eustatic transgression should match approximately that of the ocean. Only during the limited intervals of abnormally rapid ice-sheet melting could the Black Sea transgression exceed the normal values of ocean-level rise. This implies that 'flood' effects in the Black Sea basin could emerge only during times of abnormally rapid ice-sheet melting. If similar effects are linked to other episodes of deglaciation, then this model necessitates additional arguments, specifically against the model of a positive local Black Sea water balance.

3. STRAITS HISTORY

Studies focused on the geologic history of the straits linking the Black Sea with the global ocean have been ongoing for more than 320 years (the initial investigation of the Bosphorus that demonstrated its two-layer current were performed by L. Marsigli, the Italian scholar, in 1681). Yet, new evidence bearing serious consequences for traditional concepts of the paleogeography of the straits began to appear only recently. One cannot dismiss the unorthodox scenario of the Black Sea 'flood', as it has been a stimulus for generating new evidence.

Among the recent investigations, one should especially mention those that shed new light on the geological structure of the Bosphorus and the Marmara Sea. Kerey *et al.* (2004) provide new evidence on the sedimentary deposits filling the erosional downcutting within the Bosphorus channel. Aksu *et al.* (2002a) investigated the Bosphorus's geologic structure at its Black Sea and Marmara Sea gateways. Aksu *et al.* (2002b) focused on the paleoenviroment of the Marmara Sea's Holocene sediments.

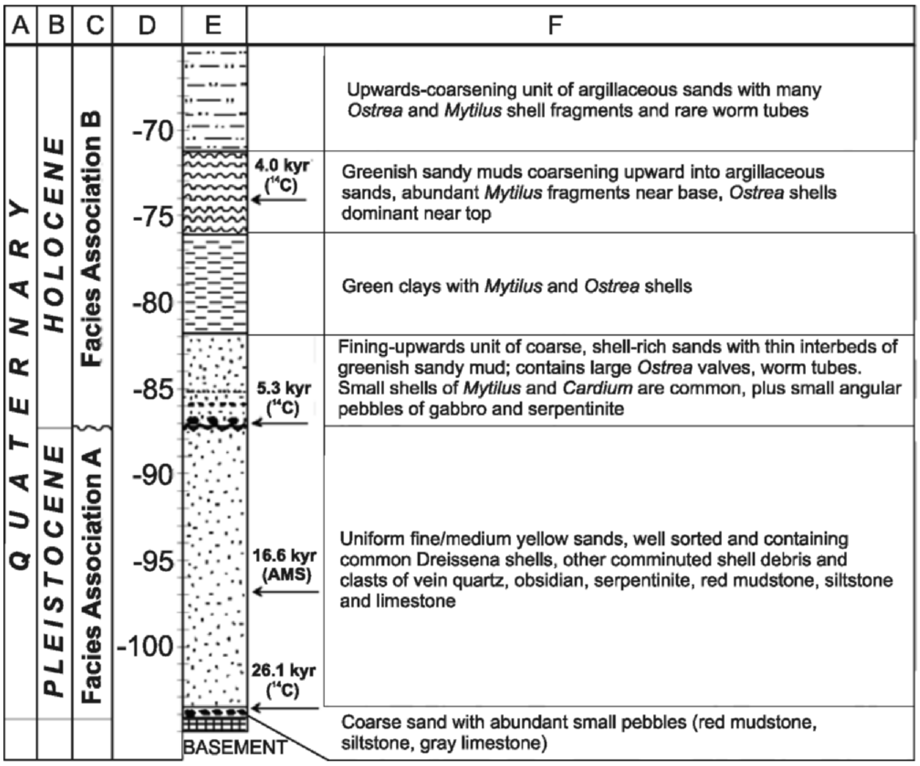

Figure 1. Litho-stratigraphy of core BPMP-14 from the Bosphorus central area (after Kerey *et al.* 2004). **Key:** A = system; B = series; C = facies; D = depth in m (water depth 65 m); E = litho-stratigraphy; F = lithological description.

Kerey *et al.* (2004) describe the litho-stratigraphy of a sequence in the Bosphorus's erosion valley, situated at the latitude of +41°08'20" (Figure 1). The core was located in the strait's central northwestern area, which separates its northern and southern parts by an elbow-like bend. The present writer cannot fully concur with the statement that

> the northern and north-central sectors of the Bosphorus formed an estuarine/lagoon arm of the Black Sea during much of the Pleistocene. Presumably at that time these sectors were separated from the southern, fault-bounded portion of the Strait (and the Marmara Sea) by a topographic barrier of some sort. (Kerey *et al.* 2004:17).

In the present writer's view, this sequence clearly demonstrates two stages in the recent geologic history of the Bosphorus. The former encompasses the Late Pleistocene and early Holocene (the lower part of the sequence, up to a depth of –87 m). During this stage, the core sediments do not reflect 'an estuarine/lagoon arm of the Black Sea', but a river that brought surplus Black Sea water into the Mediterranean Sea, and in so doing, caused a deeply eroded channel to a depth of –104 m.

The second stage recorded in the cored sequence corresponds to the Holocene Climatic Optimum. In the present writer's opinion, the underlying Mediterranean inflow through the Bosphorus at that time possessed maximum water mass and velocity, and consequently, it eroded the previously accumulated sediments (Facies Association B). Judging from the radiocarbon dates of the mollusc shells from the lower part of the Facies Association A, this occurred at about 3.5 ky BP.

Two main arguments support this view. Several writers (Mörner 1971; Badyukov 1982; and others) have shown that due to the geoid's deformations and glacio-hydrostatic effects, the post-glacial sea-level rise proceeded differently in various parts of the global ocean. In several areas, sea level never exceeded the present one, while in the others, it attained the highest elevation during the various stages of the Holocene Climatic Optimum, and gradually fell subsequently. In the Black Sea basin, maximum level was reached about 5–6 ky ago according the existing ^{14}C dates (Fedorov 1978; Badyukov 1982).

Based on the preceding, one might suggest that, in the Bosphorus, the older erosion was produced by the southbound flow (the outflow of fresher Black Sea water into the ocean), while the later erosion occurred during the Holocene Climatic Optimum 5–6 ky ago, when the two-layer flow was in place and the lower current was moving from south to north.

This model does not contradict recent geological and geophysical evidence (Aksu *et al.* 2002a, c), which identifies the existence of two systems of prograding shelf-edge wedges at the northern and southern Bosphorus gateways. According to this writer's interpretation, the southern system of prograding shelf-edge wedges in the Marmara Sea was formed during the Ice Age, resulting from the spillage of brackish Black Sea water, while the northern one, resulted from the erosion and accumulation of the lower Bosphorus flow, which became active and acquired maximum energy during the Pleistocene interglacials.

Existing evidence on the hydrology of the Marmara Sea (Aksu *et al.* 2002b; Akçer *et al.* 2004) does not contradict this model, although some questions remain unresolved. The problems related to Izmit Bay and the 'Ancient Channel' need further studies. Nonetheless, in view of the newly available evidence, the model of the gradual emergence of the two-layer flow seems to be much more substantiated than that of the 'Flood'. The two systems of prograding shelf-edge wedges at the northern and southern Bosphorus gateway, as well as the fact that bedrock erosion exceeded 100 m in depth and increased in the southern direction (Kerey *et al.*2004:5, Figure 4) seriously undermine the Black Sea 'Flood' scenario.

Hence, one can conclude that evidence relating to both the Water Balance and Straits History problems does not justify rejection of the traditional scenarios of post-glacial Black Sea history in favor of a catastrophic model.

4. REGIONAL TECTONICS

Studies of recent tectonics in the northwestern Pontic area (discussed predominantly in this section) encounter serious problems due to a limited number of reliably dated marine terrace deposits and insufficient use of oxygen-isotope ($\delta^{18}O$) oceanic records in regional stratigraphic schemes. Also, complicated tectonic pulses have led to repeated erosion of older marine terraces by younger transgressions resulting in numerous gaps in the geologic record.

A remarkable feature of the northwestern Pontic shelf is the prevalence of downward tectonic movements resulting in the development of a 'reversed staircase' of Pleistocene marine terraces, which may provide a key to the solution of several outstanding problems.

Shelf morphology is intricately linked to major climatic fluctuations over the past million years (Shmuratko 2001, 2003). The number of steps in the staircase of marine terraces corresponds to the number of Ice Age cycles, each with an average duration of about 100,000 years. The lower marine terraces are useful for identifying shorelines over the course of the post-glacial transgression.

Confronted with the nearly total absence of reliably dated early marine terraces, the writer attempted to assess the local tectonic features using a model based on the $\delta^{18}O$ records of deep-sea sediments. Two promising strategies in the study of global ocean-level fluctuations may be mentioned. The first, based on the dating of the sediments of early shorelines, is aimed at plotting the regional curves of relative sea level (RSL). The main objective of the second strategy consists of using $\delta^{18}O$ deep-ocean core records to plot the ice-volume equivalent sea level (ESL), which is understood to represent sea-level fluctuations caused exclusively by changes in ice volume, excluding the effects of lithology, glaciology, isostasy, magnetic pole displacements, and regional tectonics.

The empirical method of plotting the RSL curve has a long history of use (Broecker *et al.*1968; Bloom *et al.*1974; Dodge *et al.* 1983; Shackleton *et al.*

1984; Aharon and Chappell 1986; Fairbanks 1989; Vacher and Hearty 1989; Hearty 1998; Chappell 2002). It is based on the elevation and age of marine terraces either in tectonically stable areas or in those where the rate and direction of tectonic movements have been deemed constant over considerable intervals of time. The latter method of using terraces in places where tectonics are constant is regarded as the simplest option.

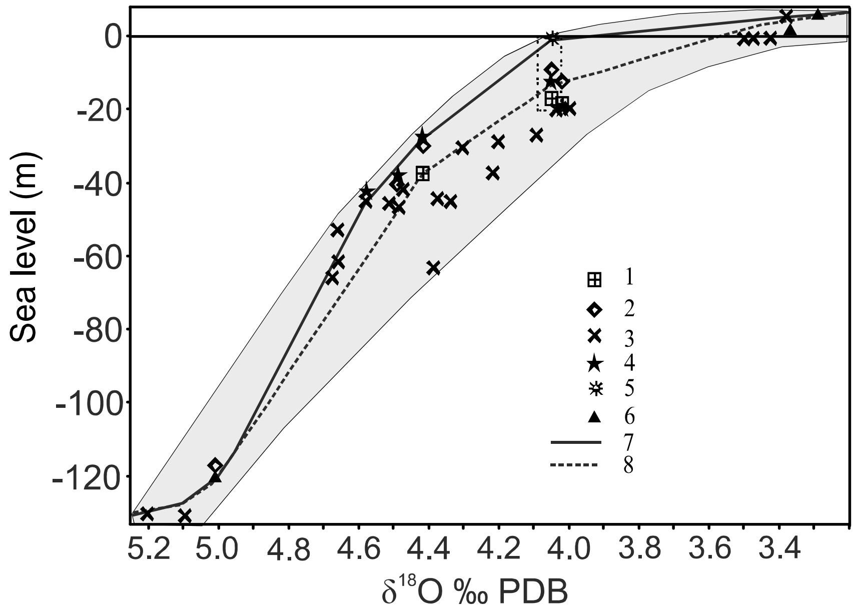

Figure 2. Relations between oceanic ESL and the $\delta^{18}O$ signal (Shmuratko 2001). **Key:** 1–5 = heights of dated terraces in tectonically stable areas: 1 – after Harmon *et al.* (1983); 2 – after Shackleton *et al.* (1984); 3 – after Shackleton (1986); 4 – after Aharon and Chappell (1986); 5 – after Vacher and Hearty (1989) and Hearty (1998)]; 6 – sea levels corresponding to the Late Pleistocene Climatic Optimum ($\delta^{18}O$ stage 5.5), the maximum regression of the global ocean ($\delta^{18}O$ stage 2), and the Holocene Climatic Optimum ($\delta^{18}O$ stage 2); 7 – line-segment interpolation according to the 'Bahamas' ESL-$\delta^{18}O$ signal model; 8 – line-segment interpolation according to to the 'Barbados' ESL-$\delta^{18}O$ signal model.

Note: The 'Bahamas' and 'Barbados' models differ mainly in the uncertainty related to the oceanic ESL position at $\delta^{18}O$ stage 5.1. Based on Mesolella *et al.* (1969), Bloom *et al.* (1974), and Dodge *et al.* (1983), it is usually accepted that at that time, sea level was 13–19 m below present. Yet Vacher and Hearty (1989) have reported the 5.1 stage sea level of –7 m at Nicholas Island, California. Based upon studies on Eleuthera Island, Bahamas, Hearty (1998) concludes that the stage 5.1 sea level was either close to the present one (0 to +1 m) or 4–5 m below it.

This method of plotting RSL provided evidence for the two-stage pattern of post-glacial transgression (Fairbanks 1989), with the maximum rates of deglaciation at 12 and 9.5 ky BP. As was later shown (Bard *et al.* 1990), the period of increased ice-sheet melting should be pushed back to 14 and 11 ky BP corres-

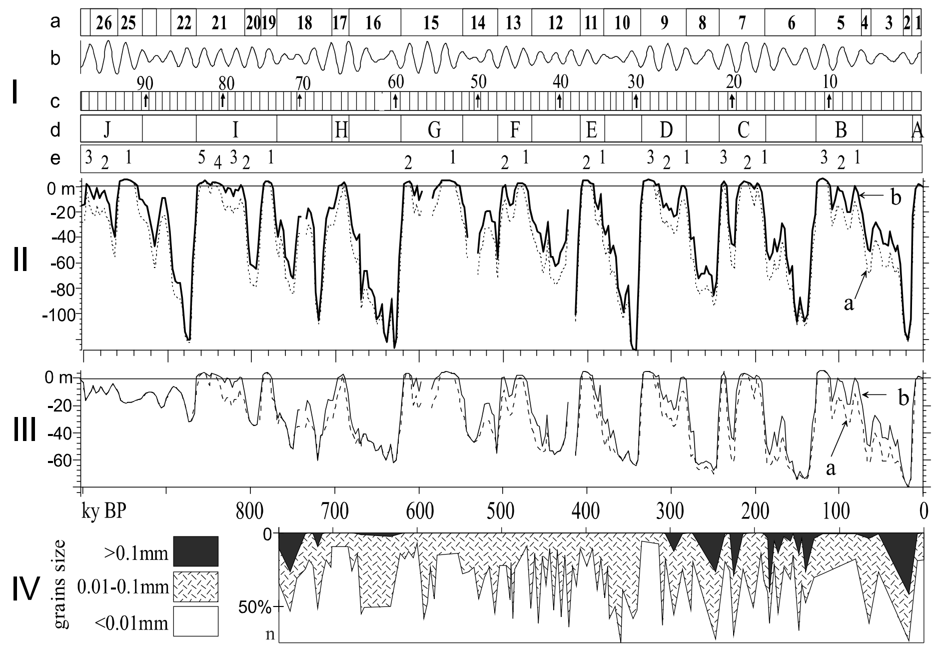

Figure 3. Oceanic and Black Sea ESL during the Pleistocene (Shmuratko 2001). **Key:** I = Astronomic 'clock' and Earth's climatic system: a – $\delta^{18}O$ stages after Emiliani (1955), Shackleton and Opdyke (1976); b – precession factor (the Sun-Earth distance at solstice); c – precession cycles; d – glacio-eustatic transgressions; e – transgression phases. II = oceanic ESL (a – 'Barbados model'; b – 'Bahamas model'). III = Black Sea ESL calculated from oceanic curves with allowance for effects of Bosphorus and local Pleistocene water balance (Shmuratko 2001). IV = grain-size distribution of deep-sea ooze in core 379A, central Black Sea depression, in Trimonis and Shimkus (1980), with the present writer's geochronology.

Note: According to the transitional functions accepted here, $\delta^{18}O$ and ESL records vary considerably for interglacial periods. Sea level reaches its highest position and subsequently changes more slowly than would be expected from the linear interpretation of $\delta^{18}O$ records. If one accepts the nonlinear character of the transitional function, the duration of the sea-level highstand during interglacials increases considerably. Arguments for this have been advanced by Aharon *et al.* (1980), Johnson and Andrews (1986), Aharon and Chappell (1986), and Vacher and Hearty (1989), who argue that the sea-level highstand during $\delta^{18}O$ stage 5.5 lasted longer than the related peak of the isotope curve, 15 ky and 3–6 ky correspondingly.

pondingly. Yokoyama *et al.* (2000) have demonstrated that, following the LGM, the first abrupt sea-level rise occurred at 19 ky BP. Hence, the empirical evidence suggests three relatively rapid glacio-eustatic sea-level rises in the world ocean during the interval between 19 and 9 ky BP, each followed either by a slowing down of, or a temporary halt in, the post-glacial transgression.

The theoretical approach (with the ESL estimation based on the $\delta^{18}O$ records) implies several assumptions which need robust substantiation; for this reason, that method is rarely used. In this writer's view, the best strategy should include both the empirical and theoretical methods.

Mix and Ruddiman (1984) argue for a nonlinear relationship between the $\delta^{18}O$ signal and ESL. Nonetheless, in the major publications on the global Pleistocene-Holocene sea-level changes, both the $\delta^{18}O$ and ESL curves are deemed identical.

The present writer (Shmuratko 2001), accepting the nonlinear character of the transitional function $F_{\delta H}$ between $\delta^{18}O$ and ESL, has transformed the $\delta^{18}O$ curve into the ESL curve for global ocean and the Black Sea basin for the last million years, and plotted the graphs for the transitional function $F_{\delta H}$ (Figure 2), based on the marine terrace evidence (Harmon *et al.* 1983; Shackleton *et al.* 1984; Shackleton 1986; Aharon and Chappell 1986; Vacher and Hearty 1998; Hearty 1998). The standard $\delta^{18}O$ curve (based on cores V19-30 and ODP677; Shackleton and Pisias 1985; Shackleton *et al.* 1990) and the 'Bahamas' and 'Barbados' transitional functions were used to obtain corresponding ESR values for the global ocean and the Black Sea basin (Figure 3).

Lacunae in the geologic record of the northwestern Pontic shelf have prevented the development of a reliable quantitative model for vertical tectonic movements (VTM) in that area during the Pleistocene. Existing evidence permits only ambiguous stratigraphic and paleogeographic interpretations. The optimal solution includes the development of several models providing for both regional VTMs and ESL changes in the Black Sea basin (Table 2; Shmuratko 2003).

Within the northwestern Pontic shelf, fragments of Pleistocene marine terraces were identified at eight elevations: about –7, –14, –20 to –21, –25, –31, –32, –37, and –49 m (in the Danube Delta area, correspondingly at –3, –6, –11, –15, –20, –30, –35, and –43 m) (see references in Shmuratko 2003). As one can observe, in practically all areas of the shelf, the depth of the fossil terraces increases with age (thus forming a 'reversed staircase'). In the first approximation, the shelf may be likened to a giant amphitheatre, facing the Black Sea basin and consisting of several extensive step blocks.

Several climatic-stratigraphic schemes that have been suggested for the Pleistocene deposits of the Black and Azov Sea areas (Veklich 1989; Zubakov 1986, 1989; and others) are not easily correlated with each other in view of the limited number of reliably dated marine terraces. For this reason, one must develop several models for VTM. Table 2 includes the parameters for seven quantitative models obtained with the use of the 'Bahamas' version of ESL changes. The mean VTM rate for the entire Pleistocene in Models 1–5 was calculated based on the estimated age of a single terrace (either '–45 m', or '–7 m'). The estimated mean rates of tectonic submergence in models 1–5 are: 4.6, 6.3, 7.0, 9.1, and 10.8 cm/millennium, respectively.

Models 6 and 7 were developed based on the stratigraphic (and, consequently, geochronological) estimates for all terraces. This permitted acquisition of mean VTM rates for various time intervals and assessment of possible changes in regional tectonic movements during the course of the last million years (Figures 4 and 5). These models carry several implications.

Table 2. VTM Models: Hypothetical geochronology of marine terraces in the northwestern Pontic shelf area (based on the present-day elevation of marine terraces, in m).

GLACIO-EUSTATIC TRANSGRESSIONS				VTM MODELS						
E	F	T	H_{bg}	1	2	3	4	5	6	7
B	1	81	+0.4				-7			-7
	2	102	0.0							-14
	3	122	+6.0					-7	-7	–20/–21
C	1	197	+2.4							
	2	213	+3.8						-14	-25
	3	237	+3.8							
D	1	285	+0.8							
	2	306	–13.2							
	3	327	+5.5						–20/–21	–31
E	1	384	–5.9							
	2	402	+5.2						–25	–32
F	1	477	+2.7						–31	
	2	498	+0.9							
G	1	567	+4.9						–32	
	2	612	+3.9							
H		690	+3.0			–45			–45	
I	1	780	+4.2		–45					–45
	2	813	+1.0							
	3	825	–0.7							
	4	844	+3.4							
	5	855	+4.6							
J	1	948	+5.5							
	2	975	–2.9	–45						
	3	996	+2.2							

Key: E = glacio-eustatic transgression stages; F = glacio-eustatic transgression phases; T = age (ky BP) of highest ocean levels based on $\delta^{18}O$ curves for cores V19-30 and ODP677 (Shackleton and Pisias 1985; Shackleton *et al.* 1990); H_{bg} = Black Sea level (in m) according to the 'Bahamas' model (Shmuratko 2001).

(1) The VTM rate in the studied area could not remain constant throughout the Pleistocene, as the latter suggestion contradicts the observable elevations of the fragmented terraces.

(2) The most probable rate of tectonic submergence of the shelf during the Pleistocene ranges between 3–4 and 60–70 cm/millennium.

(3) Theoretically, the most probable number of the Pleistocene terraces is 10–12 (Figure 4B).

(4) In the present-day relief of the northwestern Pontic shelf (Figure 4C), one can reliably identify 10–12 steps, which coincide with the theoretically assessed number of Pleistocene terraces. This implies that each Black Sea transgression stage was accompanied by active tectonic submergence in the shelf area. Allegedly, this genetically relates to the ongoing extension of the Black Sea basin at a recent stage of its tectonic evolution (Shmuratko 2001).

(5) As follows from model 6 (regarded as the most reliable), four stages of accelerated submergence are identifiable over the last 800 ky (1–4, Figure 5B), with a periodicity of 200 ky.

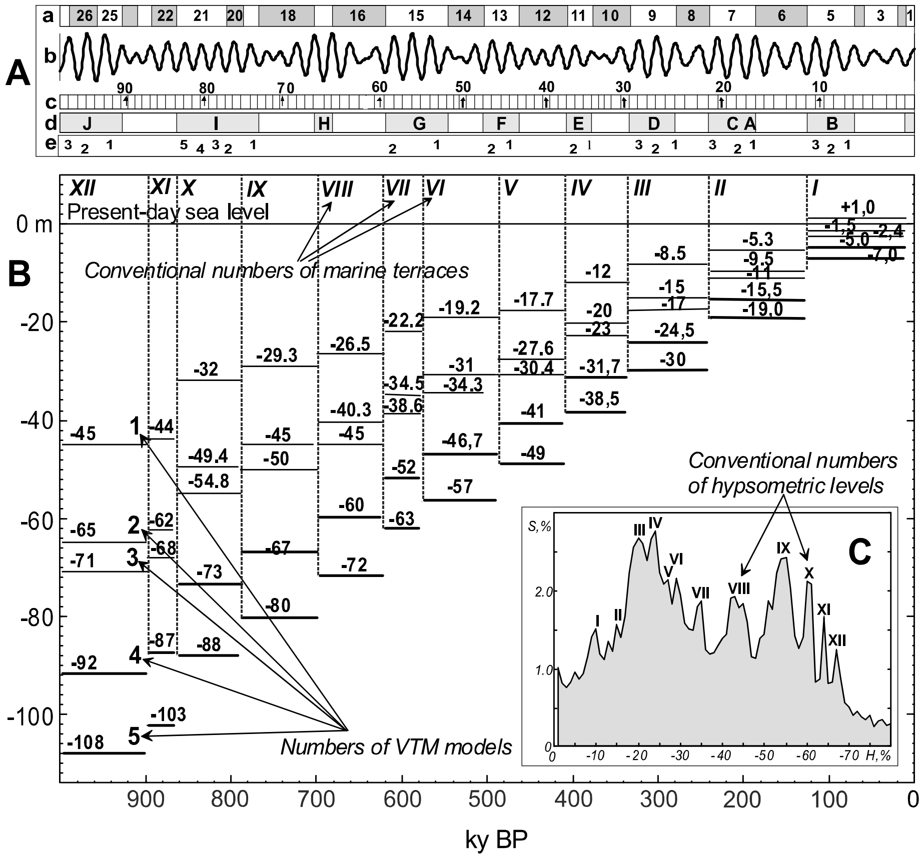

Figure 4. Theoretically expected depths of marine terraces on the northwestern Pontic shelf in accordance with models 1–5. **Key:** A = Astronomic 'clock'; B = VTM models (figures above horizontal lines represent terrace depth (in m) in relation to present sea level); C = hypsometric curve of the present relief of the northwestern Pontic shelf.

(6) An unequivocal choice in favor of one of the suggested models will be possible only when a sufficient number of reliable dates for shelf sediments become available.

The transitional function used to plot the ESR curves can be viewed as neither robust nor the only one possible. First, while calculating these functions, we were unable to use the entire evidence from radiocarbon dates. Second, the VTM pattern in the areas of coral terraces should be taken into consideration. Potter and Lambeck (2003) have demonstrated the impact of isostasy, seriously distorting the ocean level during OIS 5a. It therefore becomes obvious that our 'Bahama-based' model unjustifiably enhances ESR values and prolongs the periods of highstand at each Pleistocene interglacial.

Recently, Waelbroeck *et al.* (2002) have statistically estimated the transitional functions based on much greater factual evidence. Using the dates for coral

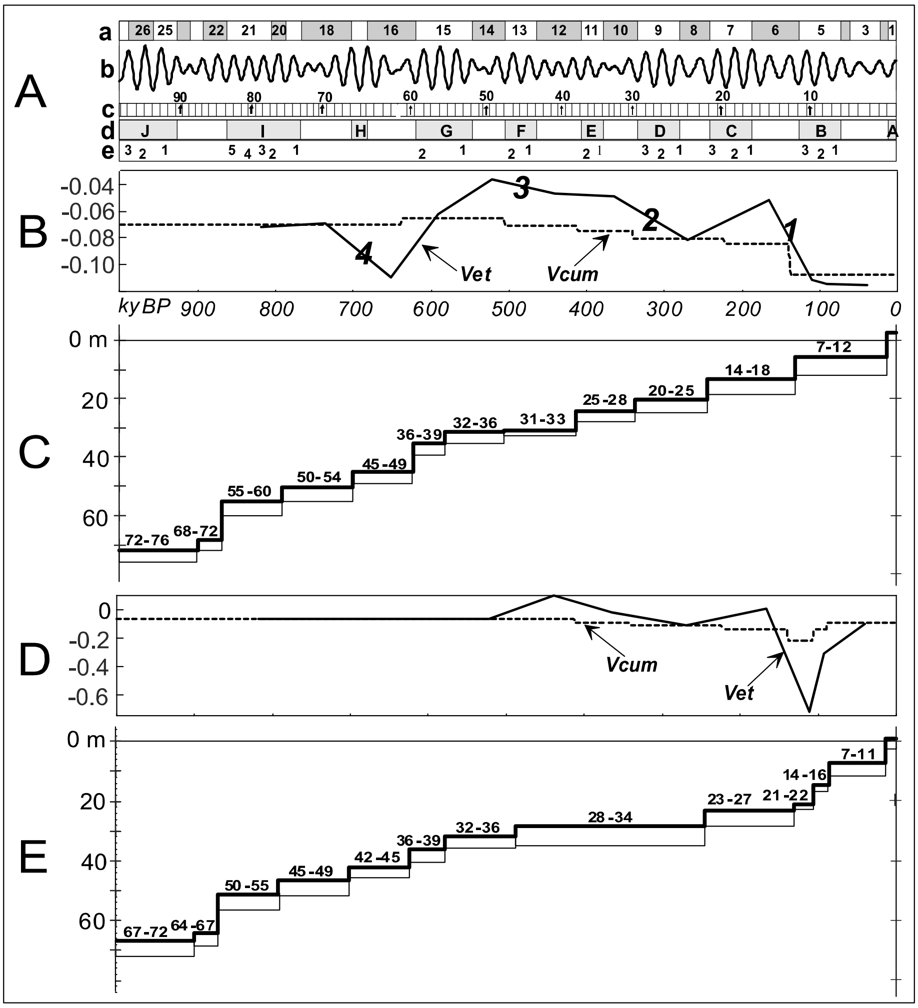

Figure 5. Theoretically expected depths of marine terraces on the northwestern Pontic shelf in accordance with models 6 (C) and 7 (E). **Key:** A = Astronomic 'clock'; B, D – mean 'stage' (V_{et}), and 'cumulative' (V_{cum}) VTM rates (mm/year).

$V_{et} = (H_i-H_j) / (T_i-T_j)$, where H_i and H_j = height a.s.l. of 'i' and 'j' terraces (in m); T_i T_j = age of 'i' and 'j' terraces (in ky). $V_{cum} = (H - H_0)_i / T_i$,, where H = height a.s.l. of 'i' terrace, H_0 = sea level at the time of formation of 'i' terrace (see Table 2, column H_{bg}), T_i = age of 'i' terrace (in ky).

C, E = theoretically expected depths of marine terraces on the northwestern Pontic shelf in accordance with models 6 (C) and 7 (E).

terraces, and $\delta^{18}O$ records for cores NA 87-22 and NA 87-25 (North Atlantic) and V19-30 (Pacific Ocean), and the assessed transitional functions, the writers have plotted the ESL curve for the last 800 ky (Figure 6).

The development of a model for recent tectonic movements in the north-western Pontic shelf using the transitional functions as suggested by Waelbroeck

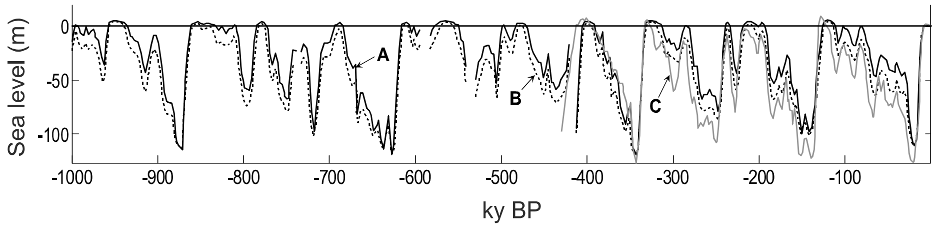

Figure 6. Glacio-eustatic fluctuations in global ocean levels based on a nonlinear relation between oxygen isotope signal and ice volume. **Key:** A = 'Bahamas' model; B = 'Barbados' model (Shmuratko 2001); C = Waelbroeck *et al.* (2002).

et al. (2002) will somewhat modify our own estimates, and this needs further investigation. Yet, it is highly improbable that these modifications will necessitate a revision of the main conclusions related to the recent tectonics of the area being discussed. The most likely changes will involve the assessment of amplitude for the identified tectonic 'wave' of 200 ky periodicity, which is likely to become still more articulate. One may expect minor changes in the number of terrace steps estimated in the models, yet taking into account the uncertainties related to the lack of reliable dates for marine terraces, all these changes will be of minor importance.

Much more important is the fact that the suggested approach permits an estimation of the possible range of VTM within the shelf. According to our estimate, at certain stages of the Pleistocene, this rate might have reached 60–70 cm/millennium (Figure 5D). Hence, the provisional conclusion that, during the last 17 ky, the shoreline in the northwestern Pontic area corresponding to the lowermost level of the Neoeuxinian basin could not have been tectonically submerged by more than 15–17 m. This means that the alleged position of sea level at –150 to –155 m at the LGM (Ryan *et al.* 1997; Ballard *et al.* 2000; Ryan *et al.* 2003) cannot be interpreted in terms of tectonic submergence (within the model of the positive local water balance in the Black Sea basin).

If one accepts the position of the Neoeuxinian basin at a level of –80 to –90 m, its causal mechanism should not be sought in recent tectonics. In other words, the tectonic nature of the 'Flood' model is unacceptable, at least for the northwestern Pontic shelf. This model contradicts the history of recent tectonic movements and also (to a certain extent) the existing evidence on the depth of erosion in the Kerch Strait and North Pontic limans. The lowest erosion level of the Kerch Strait, even in the vicinity of the shelf edge, lies at –70 m. This corresponds closely to the traditional view that the sea level of the Neoeuxinian basin was at 80–90 m below that of the present one. The same applies to the North Pontic limans. The erosion level within the river valleys, reaching –40 to –45 m at a distance of 180–220 km landwards from the Neoeuxinian shoreline which stood at –80 to –90 m, agrees well with the equilibrium profile of rivers such as the Dniester, Dnieper, and Southern Bug. Hence, from the point of view of recent tectonics, the traditional scenario of Black Sea post-glacial transgression fits the evidence better than the 'Flood' model.

5. ICE VOLUME

In this section, the following matters will be discussed:

(1) plausible variants of the transition functions between $\delta^{18}O$ signals from deep-ocean sediments to glacio-eustatic sea-levels, and

(2) plausible models of post-glacial glacio-eustatic ocean and Black Sea level fluctuations, based on the detailed $\delta^{18}O$ records of core MD952042.

These records are unique and detailed in many aspects, but they still fail to provide unambiguous solutions to all questions of interest.

The main objective consists in estimating the order and rate of possible changes in sea level controlled by climate changes of short period and high frequency on a millennium scale. Admittedly, the post-glacial sea-level curves (both of the global ocean and the Black Sea) are never smooth. Apart from two or three accelerated transgressions, they include short-period 'wiggles' or rapid sea-level rises and falls with the characteristic periodicity of 1.5–2.5 ky.

Another obvious objective includes a model of post-glacial Black Sea level change based on the detailed $\delta^{18}O$ oceanic record. Although the existing uncertainties prevent the plotting of a curve of the post-glacial Black Sea transgression based directly on the $\delta^{18}O$ oceanic record, this approach remains promising and should be implemented in future, when more empirical evidence becomes available.

In the framework of the theoretical approach, the key issue lies in the definition of function $F_{\delta H}$, linking changes in the $\delta^{18}O$ signal from oceanic records to glacio-isostatic sea-level fluctuations. The problem is to estimate the contribution of temperature to the isotopic signal. As Shackleton and Opdyke (1973), Nikolaev *et al.*(1989), and Chappell (2002) have shown, the $\delta^{18}O$ value, apart from regional and random noise, is mainly controlled by the $\delta^{18}O$ content in sea water (the global signal) and its temperature, which depends on the fractionation of the oxygen isotopes. At first approximation, the change in isotopic signal ($\Delta\delta^{18}O$) during a time interval may be shown as a sum of two components:

$$\Delta\delta^{18}O = \Delta\delta^{18}O_H + \Delta\delta^{18}O_T = k_H \cdot \Delta H + k_T \cdot \Delta T, \quad (1)$$

where ΔH = sea-level change (in m), ΔT = sea water temperature change (in °C), k_H = glacio-eustatic coefficient: the change of isotopic signal corresponding to the glacio-eustatic sea-level change by 1 m (in ‰/m), and k_T = temperature coefficient: the change of isotopic signal corresponding to sea water temperature change by 1°C (in ‰/°C).

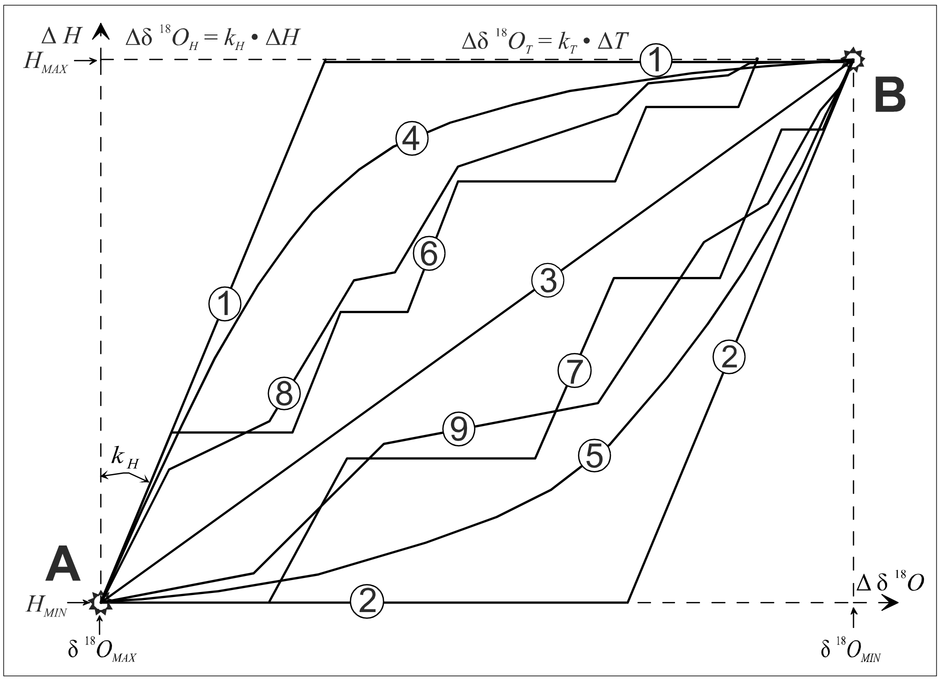

Figure 7. Theoretically possible transition functions $F_{\delta H}$ between $\delta^{18}O$ signal and ESL (1–9). **Key:** A,B = points corresponding to the maximum ice volume in continental ice sheets (A) and at iceless periods (B). A to B transition: deglaciation; B to A transition: glaciation.

The accepted values of k_H and k_T for the period of the last deglaciation are, respectively, –0.008‰/m (Shackleton *et al.* 2000) and –0.23‰/°C (Shackleton and Opdyke 1973). The value k_H depends on the isotopic composition of melting ice and, strictly speaking, is changeable over time. Yet these changes may be ignored, as they are, first, insignificant, and, second, their time pattern is well known. As follows from equation (1), the transition function linking the $\delta^{18}O$ signal to ESL depends on the temperature change of the sea water and the relative significance of glacio-eustatic and temperature factors at various ice age stages. It is convenient, in this case, to plot the $F_{\delta H}$ function in oblique coordinates.

Figure 7 shows theoretically admissible $F_{\delta H}$ functions, part of which cannot be realized. Thus, as follows from model 1, the deglaciation should proceed in two stages. In the initial stage, the isotopic signal change is controlled exclusively by the melting ice, with temperature remaining constant. In the later stage, when the ice is melted and sea level has attained its maximum, the isotopic signal change is controlled exclusively by water temperature. Apparently, this model is inappropriate. Equally unrealistic is model 2. More realistic are the models providing for simultaneous changes of both sea-water temperature and glacio-eustatic sea level. Yet, the relative input of these factors does not remain constant during the ice age, as model 3 stipulates. Therefore models 1, 2, and 3 may be excluded from further analysis. As for models 4–9, they should be tested against existing evidence.

In the present paper, both the theoretical ESR model and that of the post-glacial Black Sea transgression are based on the $\delta^{18}O$ values for the last 20 ky, obtained from core MD952042 in the Northern Atlantic (Shackleton *et al.* 2000; Shackleton *et al.* 2004). The main difficulty lies in the abnormally high input of temperature in the $\delta^{18}O$ values of that core. Yet, its main advantage is high resolution, permitting the identification of minor climatic fluctuations with a periodicity of 1.5–2 ky (Shnitnikov's cycles) that are superimposed upon the major cycles. There is evidence (Ostrovsky *et al.* 1977; Ivanov and Shmuratko 1982; Voskoboinikov *et al.* 1982; Zubakov 1992) that these minor fluctuations played an important role in the post-glacial Black Sea transgression, particularly during the last 14 ky. In this respect, due to its high resolution, the record of core MD952042 may be deemed a standard for the geochronology of climatic-eustatic ocean level fluctuations (and those of the Black Sea, with some reservation), on a scale comparable to those of Shnitnikov's cycles. As mentioned above, the climate and sea-level changes on that scale may be used as an instrument for testing the 'Flood' hypothesis, as abnormally high rates of climate and sea-level change may have occurred during these minor cycles.

Taking the above into account, it would seem logical to use the record of core MD952042 and the transition function calculated for the Pleistocene to estimate post-glacial sea-level changes. Following from Figure 8, however, not a single transition function applicable for 100,000-year cycles can be used with glacio-eustatic fluctuations of 1000-year periodicity. Apparently, models 1 and 2 (for which transition functions were calculated based on the limited evidence) lengthen the interglacial periods. In accordance with these models, global ice volume reached its present values already by 14–13 ky BP, which contradicts the evidence. Moreover, these models smooth out Shnitnikov's cycles. Model 3 is also inapplicable to the benthic records of core MD952042, although it generates a glacio-eustatic curve that agrees generally with the coral dates. This means that the Holocene glacio-eustatic level never reached +20 m, as the model predicts.

One can hardly consider the results shown in Figure 8 as unexpected. In general, the observed value of the $\delta^{18}O$ signal is controlled not only by the isotope content of the sea water and the temperature, but it is also distorted by several local factors. Strictly speaking, each core has a transitional function F of its own. One may assert that transitional functions of various cores differ to the same degree as their $\delta^{18}O$ records. Notwithstanding their common pattern, $\delta^{18}O$ curves obtained for various areas are not identical, yet, we are endeavoring to plot a single global ESR curve based upon them. This uncertainty may be overcome only when more high resolution $\delta^{18}O$ records are available.

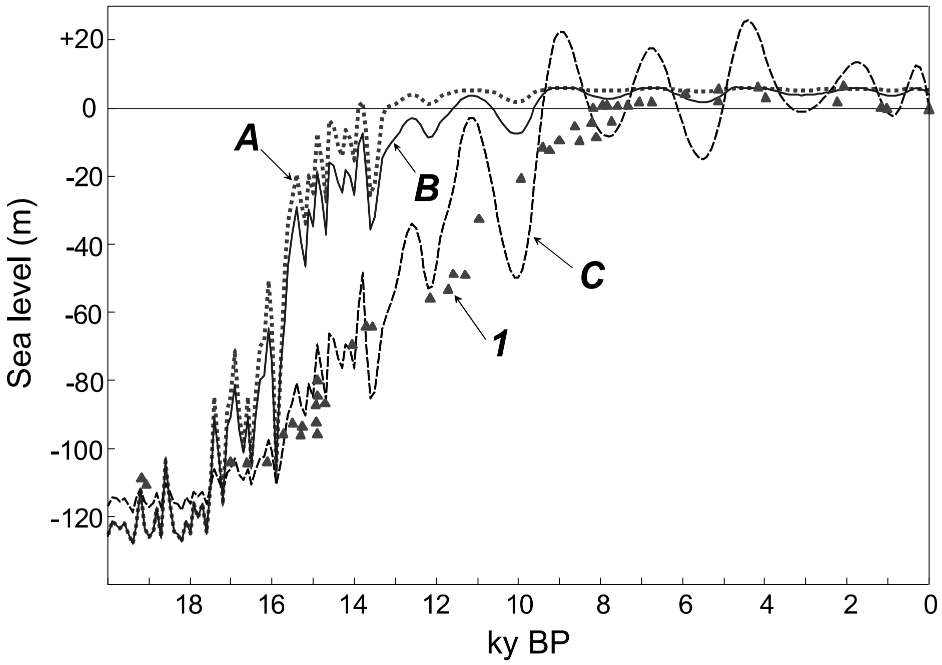

Figure 8. Models of post-glacial glacio-eustatic oceanic transgression based on the benthic $\delta^{18}O$ records of core MD952042 with the use of transitional functions: A = 'Bahamas' model; B = 'Barbados' model (Shmuratko 2001); C = following Waelbroeck *et al.* (2002); 1 = sea level according to coral reef dates (Fairbanks 1989; Bard *et al.* 1996; McManus *et al.* 2004:Figure 2).

Even presently, one may, nonetheless, reach some understanding of certain properties of transitional functions that link $\delta^{18}O$ records to ESL. Thus, Waelbroeck *et al.* (2002) note the differences in transitional functions for transgressive and regressive phases, as well as variations depending on the geographical locations of the respective cores.

In general, the main property of transitional functions lies in their unequal, step-like pattern. This follows from recent investigations proving the leap-like character of climatic transitions from one state to another (Dansgaard *et al.* 1993; Behl and Kennett 1996; Alley and Clark 1999; Cannariato *et al.* 1999; and others). This observation is applicable to climatic events of various scales, as well as the Earth's Cenozoic ice age as a whole (Lear *et al.* 2000; Bohaty and Zachos 2003). The diagram of ESL-$\delta^{18}O$ (Figure 9, model D) provides an additional indication of the step-like character of transitional function $F_{\delta H}$.

Figure 10, curve 1, shows the ESL fluctuations based on the transitional function D. As expected, this curve usually precedes one based on the coral dates, lying between the curves of models B and C.

In this respect, it maybe interesting to solve an opposite problem, i.e., to seek a transitional function for the benthic record of core MD952042 based on the

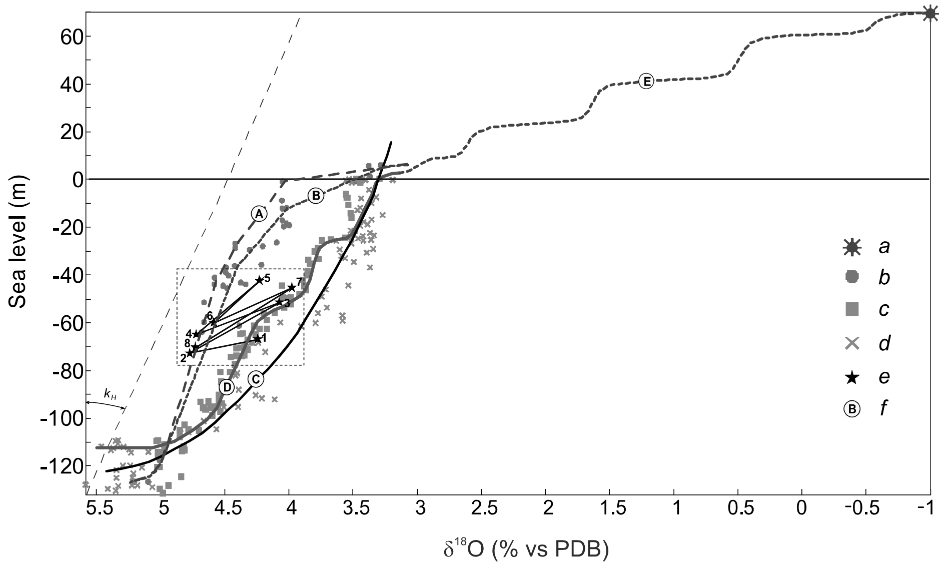

Figure 9. Transitional functions $F_{\delta H}$ for the deglaciation period. **Key:** a = iceless epoch point (Nikolaev *et al.* 1989; Lear *et al.* 2000); b = ESL and δ^{18}O signal relationship (Harmon *et al.* 1983; Shackleton *et al.* 1984; Shackleton 1986; Aharon and Chappell 1986; Vacher and Hearty 1989; Hearty 1998, whose data were used by the writer (Shmuratko 2001); c-d = ESL and δ^{18}O relationship during the last deglaciation (after Waelbroeck *et al.* 2002:Figure 2): c = core NA 87-22; d = core V19-30; e = relationship between ESL (Huon Peninsula, Papua New Guinea, coral terraces) and δ^{18}O signals from core MD952042 (Chappell 2002). Point numbers correspond to the consecutive transition stages from one level to another (Chappell 2002:Figure 5); f = transitional functions $F_{\delta H}$: A = 'Bahamas' model; B = 'Barbados' model (Shmuratko 2001); C = 'deglaciation' model (Waelbroeck *et al.* 2002); D = 'stepped' model; E = 'stepped' model to total deglaciation (up to the iceless state).

admission that evidence of coral dates from tectonically stable areas adequately reflects ESL changes. A correct solution to this problem is hampered by various factors, such as the minor fluctuations (1000-year periodicity) visible in the δ^{18}O curve of core MD952042, fragments of which are hardly identifiable with coral dates. Nonetheless, it is still possible to obtain a satisfactory match between the sea-level curve and the dated terraces (Figure 10, curve 2). Yet, in this case, the transitional function will have a linear-fragmentary pattern, as shown in Figure 11 (model F). It will be inserted into the 'statistical' transitional function as suggested by Waelbroeck *et al.* (2002). Transitional function F enables one to exclude the unsubstantiated high sea level predicted by the model, based on the transitional function C.

One might suggest the following interpretation of function F. If one accepts that glacio-eustatic coefficient k_H is unchanged, then the isotopic signal from core MD952042 would be controlled not only by the isotopic composition of sea water but also by temperature (the angle of the linear fragments of function F exceeds k_H, Figure 11). At the same time, this angle diminishes with the transi-

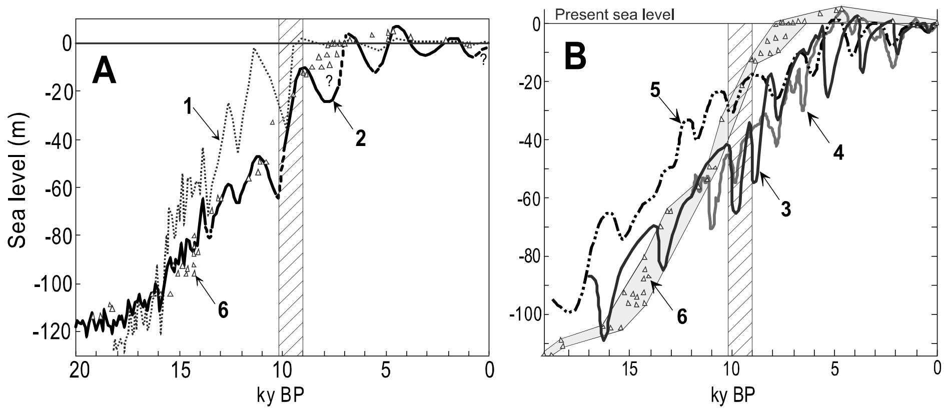

Figure 10. Post-glacial Black Sea transgression. **Key:** A = theoretical models in accordance with transitional functions D (1) and F (2); B = empirical curves: 3 = after Ivanov and Shmuratko (1982), 4 = after Ivannikov *et al.* (2000), 5 = after Konikov (this volume), 6 = oceanic level according to coral date evidence after Fairbanks (1989), Bard *et al.* (1996), McManus *et al.* (2004:Figure 2). Hatched rectangle: most probable periods with maximum rates of sea-level rise.

tion from one linear fragment of function F to another, moving upwards. This means that with the melting of the ice, the input of temperature gradually diminishes. As the glaciers melt, the rise in temperature of the bottom water slows down. Ruptures of linear fragments of function F can be viewed as the phases of a rapid decrease in bottom water temperature without a noticeable glacio-eustatic sea-level rise. Most likely, the transition from one phase (a rapid melting of ice and a rapid glacio-eustatic sea-level rise) to another (either stoppage or drop in sea level and decrease in bottom water temperature) proceeded at a rapid rate, like all similar climate changes on a millennium scale. Supposedly, this character of transitional function reflects the specific oceanography of the North Atlantic and will remain in place until the Greenland ice sheet is melted. Following that, it will transform into transitional function E (Figure 9).

Figure 10 relates to the possible rates of glacio-eustatic rise in the level of the global ocean and the Black Sea on a millennium fluctuation scale. The hatched rectangle shows the period of maximum transgression rate. Even with an obvious overestimation, this rate does not exceed 6–7 cm/year (according to curve 2, sea level rose from –70 to –10 m over the course of 1000 years). It is hard to imagine that even such a high rate (in the geological sense) may have led to anything like a 'flood'.

Figure 10 shows several curves of sea-level fluctuations in the Black Sea against the background of coral dates. If one accepts that the latter reflect the real trajectory of the oceanic glacio-eustatic transgression and then compares them with the Black Sea curves, one can identify two periods the boundary between which coincides with the emergence of the lower current in the Bosphorus. Until 10–9 ky BP, sea-level curves for the Black Sea lie either above or even with the

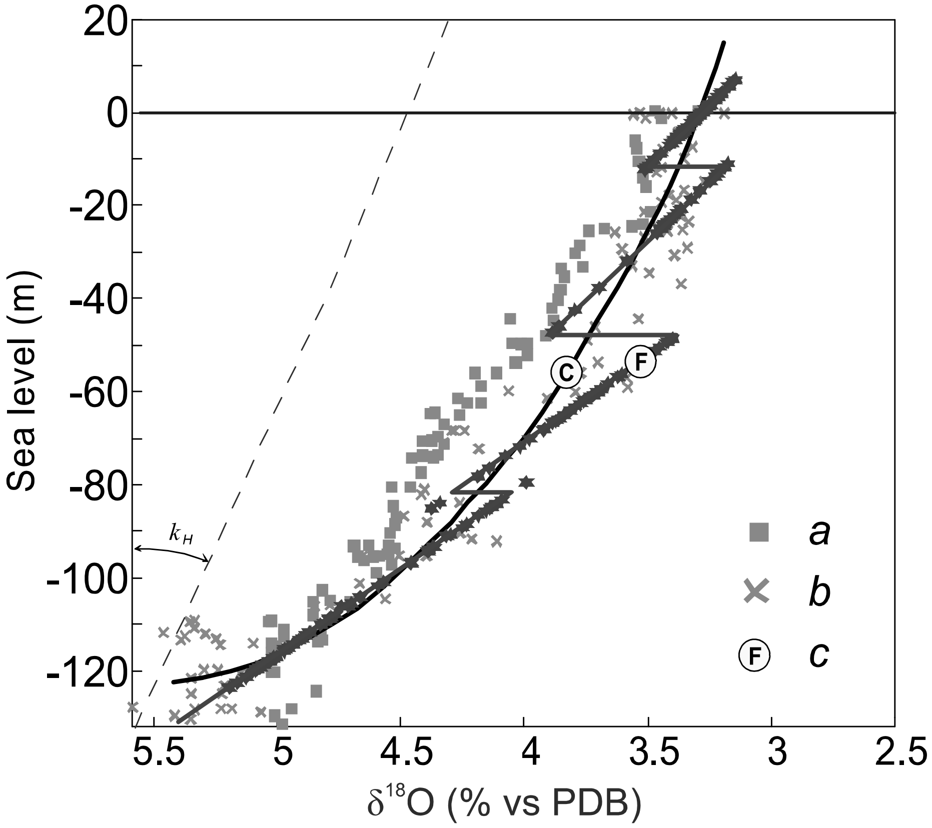

Figure 11. 'Linear-fragmentary' transitional function $F_{\delta H}$. **Key:** a-b = ESL-δ^{18}O signal relationship during the last deglaciation, following Waelbroeck *et al.* 2002:Figure 2; c = transitional function $F_{\delta H}$: C= after Waelbroeck *et al.* 2002 (deglaciation model for the North Atlantic); F = 'linear-fragmentary' transitional function.

oceanic trajectory. Subsequently, the ocean level precedes the Black Sea curves. In relation to curve 5 (Figure 10), one might suggest that the higher level of the Black Sea is due to ice-sheet melting and increase in the positive local Black Sea water balance. In that case, the situation that arose during the second period may be explained only by a tectonic submergence of the northwestern Black Sea shelf (as curve 5 ignored any possible tectonic movements). Supposedly, these active tectonic movements might have had a hydro-isostatic nature: the tectonic submergence of the last 5–6 ky having been accelerated by a rapid marine transgression.

If one compares recently plotted curves 4 and 5 (Figure 10), it becomes obvious that the problem of a comprehensive curve adequately reflecting the post-glacial glacio-eustatic transgression of the Black Sea remains open both on the empirical and theoretical levels. Much remains to be accomplished. Nonetheless, one can firmly acknowledge that there is presently no reason to relinquish traditional concepts in favor of the hypothesis of the Black Sea ‘flood’.

6. CONCLUSIONS

(1) The local Black Sea water balance (under the condition of its hypothetical isolation from global ocean during the last 20 ky) could only be positive. This implies that during the Last Glacial Maximum, the Black Sea necessarily had a unidirectional link with the ocean, i.e., it was a flowing, not a land-locked, basin. This further implies that, during the last 20 ky, the Black Sea level could not fall below that of the global ocean. The only exception could be a phase shift over a periodicity of 1000 years, yet this needs additional investigation.

(2) Recent geological and geophysical investigations of the Bosphorus and neighboring areas (Kerey *et al.* 2004; Aksu *et al.* 2002a, 2002c; and others) confirm the model of the Black Sea as a flowing lake not only for the last 20 ky, but also for several glacial-interglacial cycles on a 100,000 year time scale. Particularly important are the data identifying the occurrence of two systems of prograding shelf-edge wedges at the northern and southern Bosphorus gateways, as well as the evidence of erosion more than 100 m deep into the bedrock of the Bosphorus as well as the two-layer composition of infilling deposits.

(3) As studies of the recent tectonic movements and geomorphology of the northwestern Pontic shelf suggest, the rate of vertical tectonic movements during the Pleistocene changed with a periodicity of ~200,000 (or possibly 100,000) years in accordance with the Milankovitch orbital cycles. During the last million years, the northwestern Pontic shelf was formed from differentiated tectonic movements as a giant stepped amphitheater facing the Black Sea basin, with a reverse terraced staircase, similar to the ‘direct’ staircase of the Caucasus coast. Ten to twelve steps are reliably identifiable in the present relief of the northwestern Pontic shelf, corresponding to the number of theoretically expected shorelines. Supposedly, each Black Sea transgression stage coincided with an active tectonic submergence, which might be related to the expansion of the Black Sea basin. The most probable estimate of the mean rate of tectonic submergence during the Pleistocene is 6–7 cm/millennium; this value might have reached 60–70 cm/millennium at certain stages. Hence the alleged position of the Black Sea level at –150 to –155 m during the LGM (Ryan *et al.* 1997; Ballard *et al.* 2000; Ryan *et al.* 2003) cannot be explained by a tectonic submergence (within the framework of a positive local Black Sea water balance). Reliable dating of ancient shorelines using state-of-the-art techniques is essential for understanding the nature of recent tectonics in the northwestern Pontic shelf area.

(4) Both empirical data and modeling with various transitional functions of 'oxygen isotope signal vs. glacio-eustatic sea level' based on core MD952042 suggest that, over the last 20 ky, the rate of post-glacial glacio-eustatic Black Sea transgression never exceeded 6–7 cm/year. This value is insufficient to generate any sea-level rise that could contribute to a Flood legend.

(5) Interfacing the evidence of coral terrace dating (Fairbanks 1989; Bard *et al.* 1996) with the fragmented shorelines of the northwestern Pontic shelf suggests that the time break at 10–9 ky, when the main lower flow within the Bosphorus was initiated, became reflected in the pattern of recent tectonic movements on the shelf. Prior to that time (and different from its aftermath), the level of the Black Sea was higher than that of global ocean. This preliminary conclusion needs further substantiation.

(6) The main conclusion is that existing evidence encourages further detailed elaboration of the traditional concepts of post-glacial Black Sea history rather than acceptance of the new 'Flood' scenario. It should be emphasized yet again, that even if the idea of a 'Flood' is finally rejected, it had a positive impact, stimulating review of the Late Pleistocene-Holocene history of the Black Sea and prompting further development of fresh studies focused on it.

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10

CLIMATE DYNAMICS, SEA-LEVEL CHANGE, AND SHORELINE MIGRATION IN THE UKRAINIAN SECTOR OF THE CIRCUM-PONTIC REGION

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Abstract: The Black Sea is part of the Mediterranean circulation. It possesses a deep floor and a relatively narrow shelf, except in the northwest where the shelf is quite wide, revealing clear indicators of shoreline changes over the past 20,000 years. Morphological and geological features on this shelf (relief, granulometric and mineralogical composition of sediment, presence of wood and peat, shell horizons) as well as palynological, ^{14}C, and ^{18}O records have helped to reconstruct sea-level changes. The effects of climate, primary relief, sea-level equilibrium, and mobilization of coastal sediment account for the various phases of post-glacial transgression and the Holocene evolution of the Black Sea coast. Similar patterns of coastal evolution took place along the shores of other seas in the Circum-Pontic region.

Keywords: Black Sea, coastal zone, transgression, bottom relief, lithology

1. GENERAL GEOGRAPHICAL FEATURES OF THE BLACK SEA

The Black Sea is an internal sea and part of the Atlantic Ocean system; it is connected to the Atlantic through the Bosphorus, Dardanelles, and Gibraltar Straits and through the Marmara, Aegean, and Mediterranean Seas (Figure 1). Its area is 423,000 km^2, its water volume is 538,000 km^3, and its depth ranges from an average of –1272 m to a maximum of –2211 m (Terziev 1991). Total shoreline length, including nearshore bars and spits, is 4431 km. Surface water salinity varies from 14 to 19‰, while that of deep water remains between 22.3

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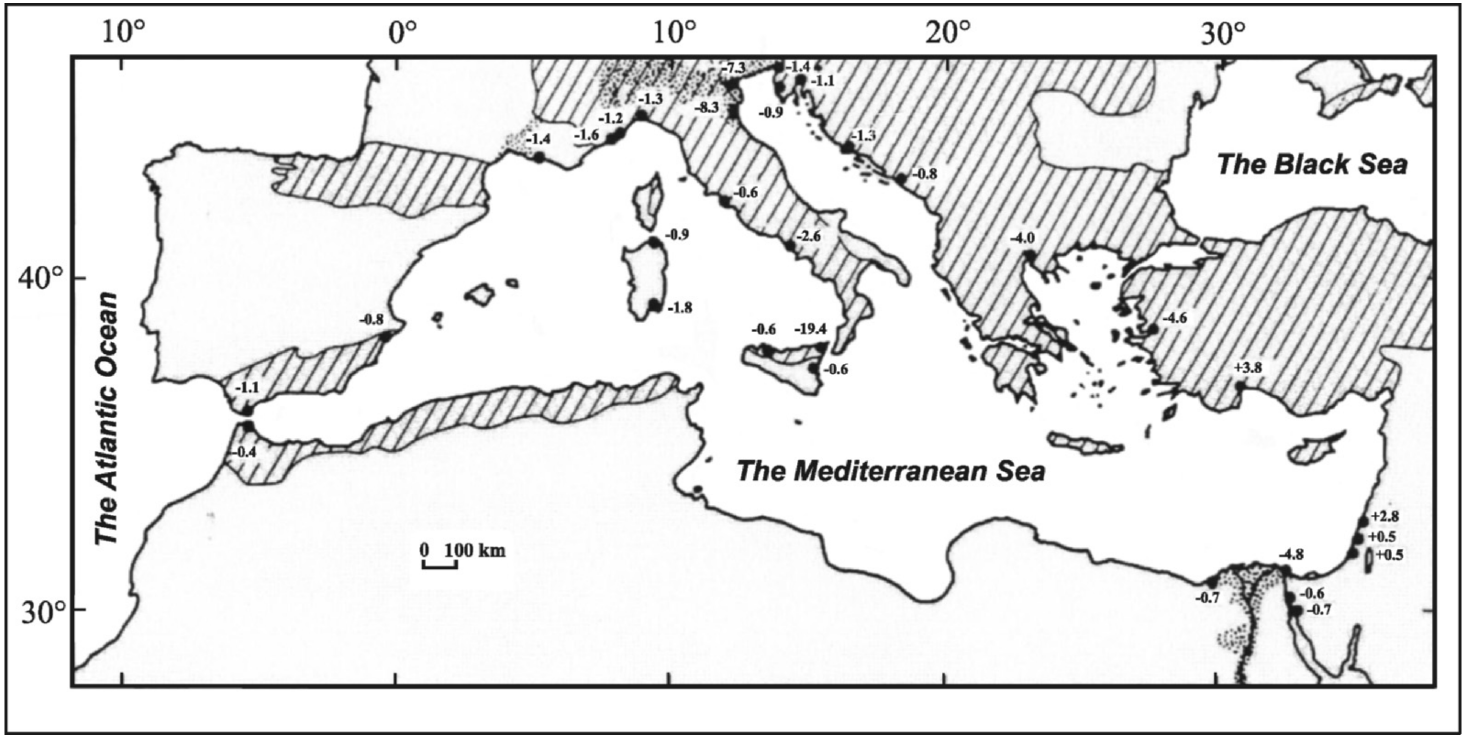

Figure 1. The Black Sea and Mediterranean basin system. Numbers located near large dots indicate the relative rate of sea-level change during recent decades. Shaded areas represent territory occupied by Alpine age geostructures.

and 22.6‰; the overall average is 21.8‰. Surface water temperature rises in the summer to 24–26° C near the shore, with a maximum reaching 29° C in the wider expanses of shallow water. In the basin's center, far from shore, water temperature is 22° C. In winter, temperature generally falls to 8–9° C in the surface layer, although negative temperatures occur and the sea may freeze over in the shallow northwestern part in occasional years. At depths below 150 m, temperature changes are insignificant, varying within the range of 8.6 to 9.1° C.

The Black Sea depression formed about $5 \cdot 10^6$ years ago as a subsiding intermontane zeugosyncline located between the orogenic systems of Crimea and Caucasus on one side, and the Pontic Mountains of Anatolia on the other. Beginning in the Oligocene, and following the termination of the geosynclinal stage, orogenic movements were activated within the Mediterranean intermediate belt, triggering the subsidence of the Black Sea basin, which continues today. The Azov-Black Sea basin achieved its present contours in the Cimmerian, when the base for the present-day hydrological network was established. During the Pontic and early Cimmerian stages, the Black Sea did not yet exist. Only in the middle Cimmerian, following the Zancian-Cimmerian transgression, did the catastrophic submergence of the sea floor begin, accompanied by the deposition of siderite (Semenenko 1987, 1999). During subsequent development, the size, shape, and depth of the sea were continuously modified under the impact of hydro-climatic and tectonic processes. Connections with the Caspian and Mediterranean Seas were repeatedly established and disrupted.

Prior to and subsequent to the emergence of the Euxinian basin, the following stages are recognized: (1) Middle Miocene marine basin, (2) Sarmatian saline basin, (3) Meotic basin, (4) Pont brackish lake, (5) Akchagyl basin, developed from the Cimmerian and Kuyalnitskii basins, and (6) Apsheron basin, freshwater, developed from the Akchagyl basin.

The tectonically unstable Aegean-Bosphorus zone was formed essentially during the Cimmerian stage. Several geostructures developed along its trend, including the North Aegean trough, the Dardanelles, the Marmara Sea, and the Bosphorus Strait. Practically all negative relief forms are of tectonic origin, reflecting disruptions in the crystalline basement and bedrock (Malovitsky and Kazakov 1975). These depressions were infilled by sub-horizontally layered deposits of Pliocene-Quaternary age. The Mediterranean-Black Sea transition zone has a block structure, with separate blocks experiencing variably directed movements. As a result, the fault lines are usually deformed, either widening or shrinking, as noted by Garkalenko, Vartanov, Malovitsky, A.P. Milashin, and others (Shnyukov 1984b). The Bosphorus Strait is located in one of these fault zones, its size, width, and depth having been repeatedly modified, with concomitant changes to the water discharge flowing through it.

Large-scale transgressions of the Black Sea occurred during the Chaudinian, early Old-Euxinian, late Old-Euxinian, Paleo-Uzunlarian, Uzunlarian, Karangatian, and Surozhian periods (Figure 2). During these transgressions,

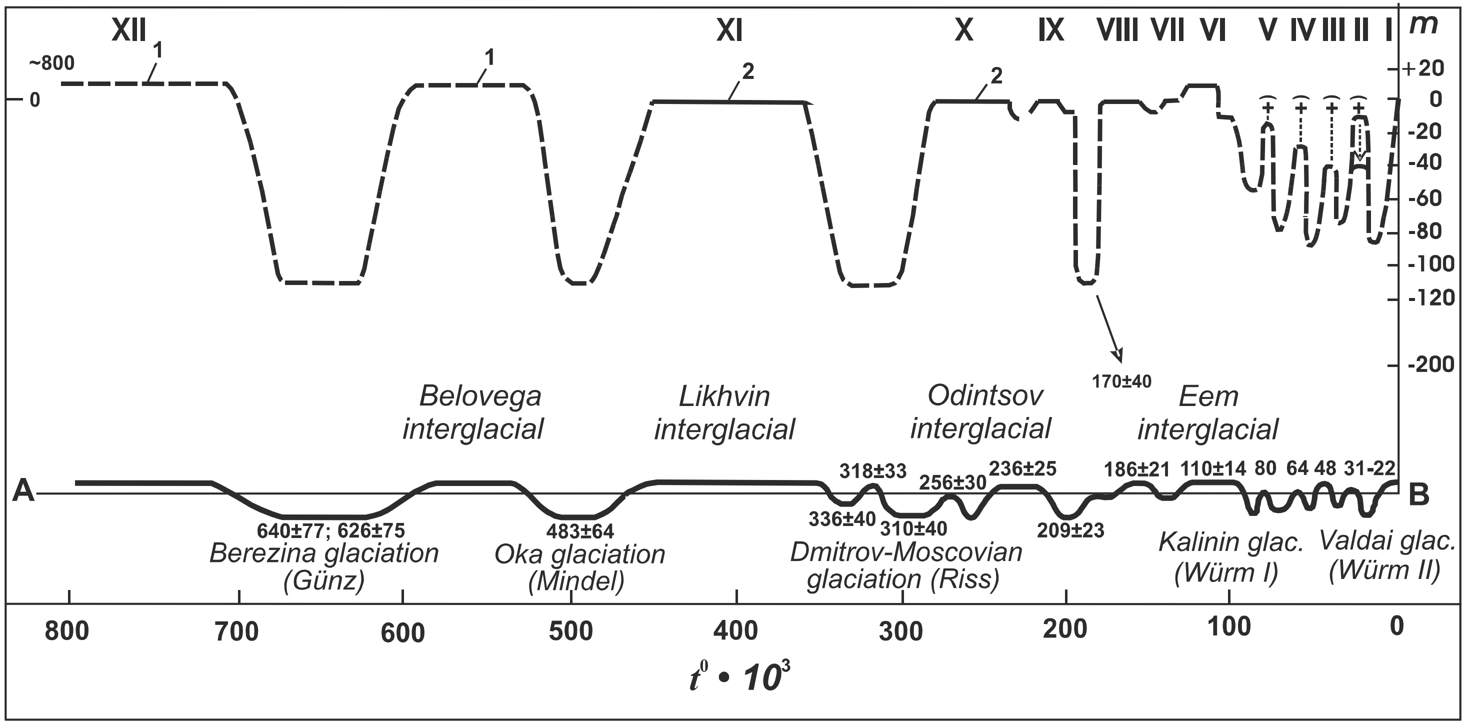

Figure 2. Comparison of world ocean levels during glacial and interglacial stages of the Pleistocene, according to P.A. Kaplin (1976, 1982). **Key:** upper curve — proposed variations in the level of the global ocean, showing elevation of dated terraces (XII-I). 1 — sea-level variations and transgression peaks, 2 — terraces; lower curve A-B — glacial and interglacial periods with ages in thousands of years based on various dating methods.

sea level did not exceed 11–12 m above present (Fedorov 1982). The deepest regressions (up to –100 m) that separated these transgressions were: the Post-Chaudinian, the Post-Old-Euxinian, and the Post-Karangatian (Figure 3). Hence, Pleistocene-Holocene fluctuations in the level of the Black Sea were a common occurrence, having had an influence on shoreline formation during each phase of recent geologic history. Such an observation suggests that serious examination be accorded to the accelerated sea-level rise of the past 100–150 years, in view of the potentially unpredictable future variations.

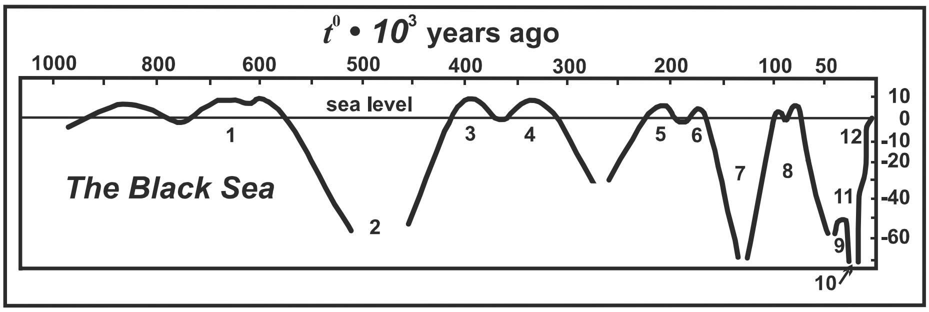

Figure 3. Schematic curve of variations in the level of the Black Sea during the Pleistocene (according to Fedorov 1988); numbers designate stages. **Key:** 1 — Chaudinian transgression; 2 — Post-Chaudinian regression; 3 — Early Old-Euxinian transgression; 4 — Paleo-Uzunlarian transgression; 5 — Late Old-Euxinian transgression; 6 — Uzunlarian transgression; 7 — Post-Uzunlarian regression; 8 — Karangatian transgression; 9 — Surozhian transgressive phase; 10 — Post-Karangatian regression; 11 — Neoeuxinian transgression; 12 — Black Sea Holocene transgression.

During the late Pleistocene and Holocene, three main factors controlled the formation of Black Sea shorelines and the continental shelf: (1) fluctuations in the level of the world ocean, (2) crustal movement in the present-day Bosphorus area, and (3) tectonic movements of the Black Sea depression. These factors acted against a background of changes in atmospheric moisture and water balance of the world ocean. An analysis of sea-level changes in the world ocean during the geological past is a prerequisite for understanding fluctuations in the level of the Black Sea.

2. CLIMATE CHANGES OVER THE PAST 20,000 YEARS

Complex changes in climatic parameters have been observed from the beginning of the Holocene transgression that were directly dependent upon climate-forming factors. At present, most authors link fluctuations in the level of the world ocean with the degradation of the ice sheets under the impact of climate changes (Kalinin 1976; Kalinin *et al.* 1976; Kaplin 1976; Klige 1982, 2000). Since ice-sheet degradation and the ensuing transgressions were controlled by increasing mean temperature and evaporation, as well as decreasing compensatory atmospheric precipitation, climate changes are viewed as a prime mover for eustatic sea-level fluctuations. This reflects a common pattern in natural processes, including those on geological time scales (Drozdov and Lugina 1998).

A dry and cold climate generally prevailed at the beginning of the Holocene transgression, particularly in the Circum-Pontic area. Until recently, the Verbitsky model (1982) has been used to estimate climate changes based on oxygen isotope and pollen evidence from the present-day areas of mainland deglaciation. In developing mathematical models, several assumptions were made that (1) the marginal areas of Pleistocene glaciation were located mainly on landmasses, (2) the influx of atmospheric precipitation to the ice-sheet surface was compensated exclusively by melting in the marginal zone, and (3) climate changes resulted in marked variations in the extent of the ice sheet.

It was presumed that a stationary ice sheet would have spread along a basement of any shape, and that icebergs formed directly off the margin of the ice sheet. Thus, one could assume that the size of an ice sheet corresponds to that of a landmass until warming reduces iceberg formation to zero and the ice sheet no longer progresses seaward but recedes. The distribution of atmospheric precipitation and air temperature over the surface of the ice sheet and the geothermal flux have to be considered as well. It is necessary to estimate ice-sheet volume, and for this purpose, a system of equations describing the thermodynamic regime of the ice sheets has been developed.

Consequently, the regime of glacial development under conditions of increasing temperature can be estimated. Climate warming modifies the area of atmospheric precipitation. As the oxygen isotope evidence from ice cores taken from various points on the glaciers has shown, at 15–20 ky BP, temperature was 5–6° C lower, and precipitation was reduced by a factor of 1.9 compared to the present. A drop in temperature of 1° C, therefore, produces a decline in precipitation by up to 14% on average. Extrapolating this estimate to temperature increase, one obtains a 14% increase in precipitation with a temperature rise of 1° C.

Modeling has confirmed that a temperature rise decreases ice viscosity and enlarges the melting zone, and that the size of the meltwater flux reduces the volume of the ice sheet proportionately. An increase in atmospheric precipitation as snow on the surface of the glacier increases its volume. When an increase in temperature is accompanied by an increase in precipitation in the form of rain, it is possible to develop a scenario in which the loss of water may be compensated by surface precipitation on mainland glaciers. Hence, global sea level may change slightly, oscillating with consecutive rises and falls.

As they are part of the world ocean, the Black and Azov Seas should be subject to the same sea-level changes that global water balance imposes on oceanic sea level. However, because both seas are presently connected to the world ocean via the narrow and shallow Bosphorus Strait, the link between these basins and the ocean is quite limited. Consequently, forecasting future sea-level variation within the Pontic basin will require consideration of not only world ocean influences, but other factors as well.

3. CHANGES IN OCEAN LEVEL

Analysis of the earth's geological history using isotopic methods has made it possible to generate reliable curves documenting ocean-level changes for the late Pleistocene-Holocene, i.e., the past 35–30 k years. These recent ocean-level changes occurred against a background of long-term fluctuations over the past 400 k years. Figure 4 graphs several curves (from Klige 1980, 1985) based on evidence from (1) coastal deposits, (3) Mediterranean terraces, and (4) coral reefs adjusted for evidence from sea floor deposits. It is clear that, in the Pleistocene, rising sea level corresponds to a warming climate, and falling sea level corresponds to cooling. The Fairbridge curve (2) based on evidence from tectonically stable coastal areas is deemed the most reliable.

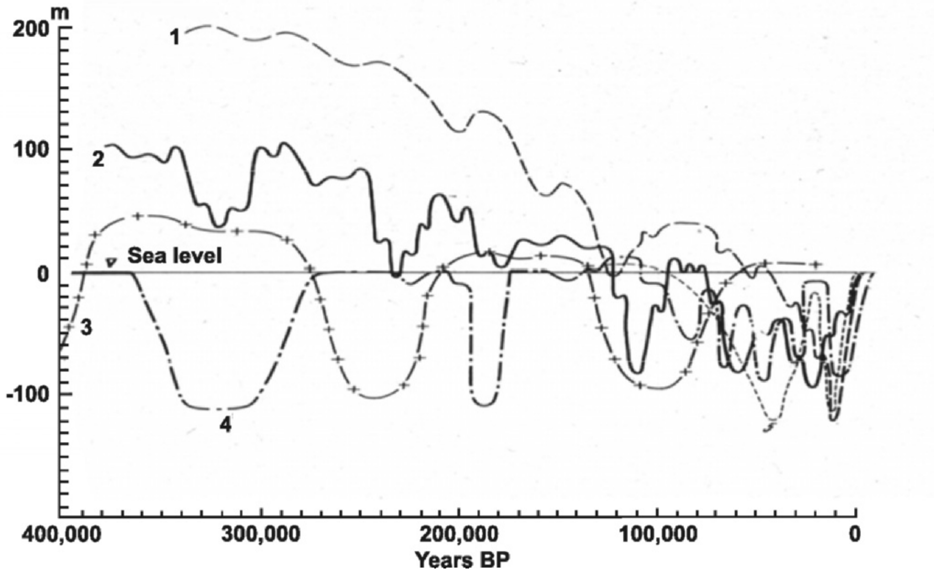

Figure 4. Changes in ocean level during the past 400,000 years according to: 1 – H. Nakagawa; 2 – R.W. Fairbridge; 3 – E. Ceiner; 4 – P.A. Kaplin; 5 – F. Shepard and J. Curray (from Klige 1980, 1985).

In Figure 5, the Fairbridge curve for the late Pleistocene (curve 1) was adjusted by Klige (1982) to eliminate the effects of tectonics. Interfacing Klige's data with Emiliani's temperature curve (curve 3, t° C) makes it apparent how oceanic transgressions and regressions are related to thermal conditions on the surface of the earth. Climatically dependent fluctuations of the world ocean level confirm Milankovitch's theory and show the changes in solar radiation expressed as equivalent changes of latitude φ° (curve 2). As may be observed on the graph, the radiation and thermal curves 2 and 3 correlate satisfactorily with those of sea-level changes, thus proving that such changes are climatically controlled. Besides, fluctuations in the radiation balance directly affect temperature variations in the lower atmosphere and thereby enhance the periodicity of eustatic sea-level changes in the Black Sea, since it is part of the system (Komarov 1979; Shmuratko 1991; Zubakov 1992). Apparently, the same pattern is applicable to the Holocene and may be used for developing a scenario for the next 50–100 years. The right side of the graph demonstrates that the Holocene is characterized by more pronounced sea-level rises (Figures 5 and 6).

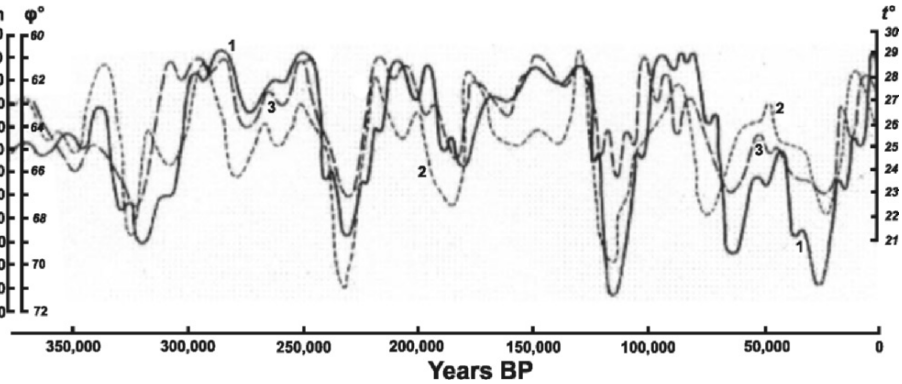

Figure 5. Fluctuations in the level of the world ocean, excluding tectonic impact: 1 — according to R.W. Fairbridge (1961); 2 — curve of solar radiation changes expressed in equivalent change in latitude by M. Milankovitch (Shmuratko 1991); 3 — temperature equilibrium of the surface layer of the world ocean in tropical latitudes, according to C. Emiliani (Klige 1980).

At present, various authors in different countries and using different methods have proposed more than 100 sea-level fluctuation curves, some of which are shown here (Figure 7). In spite of regional differences, most of the curves follow a similar trend, which becomes apparent subsequent to the decay of the last continental ice sheet (Kaplin 1976; Klige 1980, 1982; Kaplin 1982; Kaplin *et al.* 1991). Glacial melting was accompanied by a rapid rise in sea level, especially during the interval between 8–6 ky BP.

A major regression began 30,000 years ago during the Late Pleistocene, and though it was complicated by fluctuations, water level dropped to about 110 m below present (Figures 2 and 3). This was due to disturbances in global water

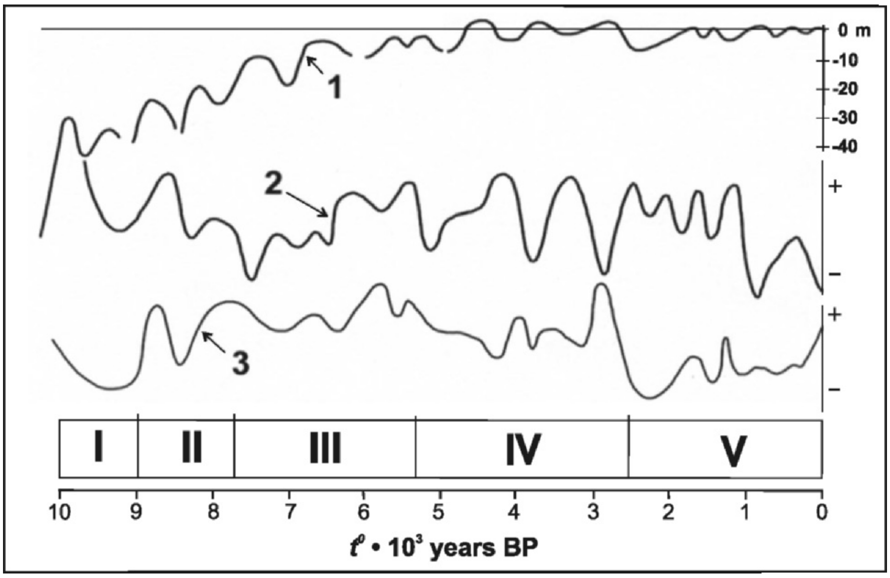

Figure 6. Comparison of the dated Black Sea level (1) with Holocene climatic changes, expressed as curves of moisture (2) and heat supply (3). Climatic stages: I — Preboreal; II — Boreal; III — Atlantic; IV — Subboreal; V — Subatlantic; time scale in uncalibrated years BP (according to V.I. Shmuratko).

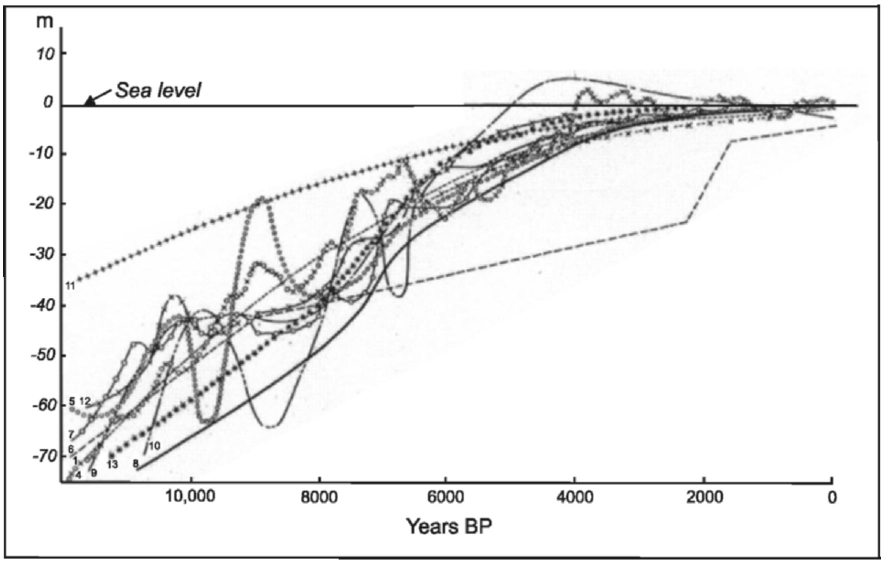

Figure 7. Curves of sea-level changes in the world ocean using different methods in different areas according to: 1 — F. Shepard and J. Curray; 2 — S. Jelgersma; 3 — V. Auer; 4 — R.W. Fairbridge; 5 — T. Kenney; 6 — K. Emery; 7 — N. Mörner; 8 — H. Godwin; 9 — H. Flohn; 10 — F. Shepard and H. Suess; 11 — M. Bloch; 12 — P. Farrand; 13 — H. Faure and P. Floward (from Klige 1985).

exchange caused mainly by changing thermal conditions on the earth's surface, and it led to transformations in the earth's water balance (Kalinin 1976; Klige 1980, 1982,1985; Siddall *et al.* 2003). During this time, the connection between the Black Sea and the Mediterranean was blocked, and the former may have been transformed into a large freshwater lake (Balandin *et al.* 1978; Fedorov 1982; Kuprin and Sorokin 1982; Ostrovsky 1982; Zubakov 1992; Sperling *et al.* 2003; Kerey *et al.* 2004).

During the entire interglacial period, the semi-enclosed water body in the Pontic basin was connected to the world ocean via the Mediterranean Sea. During glacial periods, the shallow Bosphorus Strait (depth 40 m) dried up. Available data indicate, however, that depending on the position of sea level, a small current in the form of a one-way outflow from the Black Sea into the Bosphorus was sustained (Ostrovsky 1982). On the other hand, a connection between the Black Sea and the Caspian basin was established as a result of the pluvial hydrologic regime within the catchment basin of the East European Plain that was controlled by variation in precipitation and temperature. In consequence, the levels of the Black and Caspian Seas changed, mostly in antiphase with the world ocean. Over the past $1.5 \cdot 10^6$ years, the Caspian and Black Seas were connected via the Manych Strait seven times, the last of which occurred during the Old-Euxinian stage. Between the Caspian transgressions, the western part of the Manych Strait was transgressed by the Black Sea; this took place twice during the Uzunlarian and three times during the Karangatian stages (Balandin *et al.* 1978; Fedorov 1982; Semenenko 1987; Zubakov 1992; Semenenko 1999).

The eustatic nature of post-glacial fluctuations is generally acknowledged (Blagovolin *et al.* 1976; Nikiforov 1977; Klige 1985; Semenenko 1987, 1999; Klige 2000). Most writers consider that warming of the lower atmosphere and the ensuing rapid melting of the ice sheets beginning around 16,000 years ago (Figures 4 and 5) led to a simultaneous eustatic rise in sea level. Most authors also argue that sea level has been rising progressively over the past 16,000 years and that the average rate of increase reached or even exceeded 10 mm/year (Shcherbakov *et al.* 1976; Shcherbakov 1982a, b, 1984). Yet, even higher rates were possible during short-lived fluctuations in the next phase of sea-level history. The irregular pattern of this transgression resulted first from temperature and moisture variations in the atmosphere and ocean, and subsequently from the degradation of glaciers, water saturation of the atmosphere, and variations in global water balance (Klige 1980; Shmuratko 1991; Zubakov 1992; Klige 2000). This led to rhythmic water-balance changes and fluctuations in the world ocean level, including that of the Black Sea, during the past 18,000–16,000 years. Within the 20th century, sea level has been rising eustatically at a rate of 20–26 cm/100 years, and several coastal hydrological stations have fixed rates of relative sea-level rise at more than 10 mm/year.

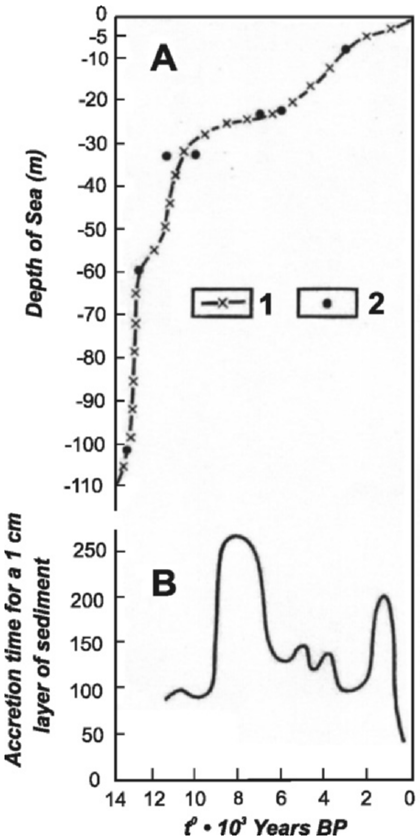

Figure 8. Graph of the Holocene transgression (A) on the Ukrainian shelf of the Black Sea based on radiocarbon dating: 1 — isobath of lower part of dated Holocene sediments; 2 — sampling sites on the shelf yielding absolute dates. Graph of Holocene deposit rates (B) for the deep water part of the Black Sea (based on Balandin *et al.* 1978).

How the post-glacial transgression was initiated in the Black Sea remains an important issue. Based on radiometric dates as well as their structure, bottom sediments at 100–110 m depths on the outer margin of the north-western Black Sea shelf (Figure 8) have been estimated to date to 18,000–17,000 BP (Komarov 1979; Shcherbakov 1982a, b, 1984). According to Shimkus (1975), this occurred earlier in the Mediterranean Sea (about 20,000–19,000 years ago), which should be taken into consideration when discussing the interaction between the Mediterranean and Black Sea basins. It should be emphasized that the Black Sea was not totally isolated from the Mediterranean (Ostrovsky 1982) during the Pre-Neoeuxinian stage, at the beginning of warming when the Black Sea waters remained fresh. Shimkus (1975) also believes in a high probability that the two seas were connected. A.B. Ostrovsky assumed that during a deep regression, highly freshened water could exist in the Black Sea only when there was outflow into the Mediterranean Sea. This could not be prevented by the presence of the –40 m Bosphorus sill because, at that time, it was eroded to about –120 m (Scholten 1974), determined on the basis of geophysical soundings along the Bosphorus thalweg and columnar drilling with microfaunal analysis. This connection between the two seas should have affected late Pleistocene fauna and flora in the Bosphorus area.

By the time of Anosov's thermophase (16,600–15,200 years ago), almost the entire shelf in the Ukrainian sector of the Black Sea and most of the Azov Sea bottom became dry land, exposed to subaerial conditions and subject to processes of denudation and water erosion (Shcherbakov *et al.* 1976).

During the Azylian cryophase (15,200–12,300 years ago), a cooling interval within the context of general warming occurred, and average temperature dropped by 1° C as recognized in pollen spectra by an increase of cold-resistant herbs (Chenopodiaceae and *Artemisia* making up 62%) and a decline of arboreal species (*Pinus* by 29%) (Kuprin and Sorokin 1982; Zubakov 1992). Experimental evidence from Verbitsky (1982) and others suggests a corresponding lowering of atmospheric precipitation (by 15%), i.e., an increased dryness of climate.

The climate of the Azov-Black Sea basin became warmer during the short-lived Borshchevo thermophase (12,300–11,000 years ago), at the beginning of which sea level began to rise rapidly, invading depressions in the initial relief. Sediment accumulation rates on the sea bed reached 90–95 cm/1000 years. This was followed by the next cryophase at the beginning of the late Neoeuxinian, or Younger Dryas, stage (11,000–10,500 years ago), and this stage was soon followed by a mild and wet one at the end of the late Neoeuxinian (10,500–9300 years ago). In the palynological spectra, herb pollen and pine markedly declined (up to 15%), whereas mixed-broadleaf forest species rose up to 65%, and that of alder up to 8%. The Black Sea level rose to –60/–50 m, as shown by Balandin *et al.* (1978) (Figure 8) and observed by Shcherbakov *et al.* (1976) based on the radiometric age of bottom sediments sampled in the cores along a profile stretching from the outer margin of the shelf to the nearshore.

From 9300 BP, following a short yet strong deceleration in the rise of the Black Sea at the elevation of –30/–35 m, there occurred a short-term cooling with an increased dryness of climate. The Bugazian phase (Nevessky 1967) followed, which lasted until approximately 6200 years ago. Black Sea water became increasingly saline, with hydrogen sulfide contamination of the deep layers. The pollen spectra show an increase in forest species: pine, spruce, fir, with maximum values of alder, linden, beech, elm, hornbeam, oak, hazelnut (up to 85%), and reduced values of herbs (7–10%). The conditions of the post-glacial climate optimum were attained during the Bugazian and Vityazevian phases of the Holocene (7000–4500 years ago), when climate became warmer and increasingly wetter. By its end, sea level had reached its present elevation and even exceeded it, as manifested by the Nymphaean (New Black Sea) terrace.

The modern climate of the Pontic region emerged over the last 3500 years during the late Holocene Dzhemetinian stage. In the northern and western part of the Black and Azov Seas, the climate was close to ultra-continental, but it became markedly different in the east, where conditions became warmer and wetter. Pollen spectra show a reduction in grains of arboreal and shrub pollen and the development of xerophytic vegetation. Over the entire North Pontic area, the number of arboreal species declined by about 60–65%, with a corresponding rise in herbs (by 20–25%), with *Artemisia*, Chenopodiaceae, and Cerealia in the northwestern Azov Sea area. During the Dzhemetinian stage, changes in vegetative cover reflected minor variations in temperature and moisture. Regular temperature changes did not exceed 0.5–1.5° C, although occasionally they could have been greater. Correspondingly, moisture changes within the last 3500 years lie within 20–25%. In general terms, the forested area was reduced and replaced by steppe and forest-steppe associations.

Changes in climate and the ensuing changes in Black Sea water balance discussed above resulted in a general sea-level rise, as had been noted in relation to the influence of the world ocean on the formation of the Black Sea basin (Figures 2, 5, and 8). Similar curves have been obtained by several authors using different methods (Nevessky 1967; Blagovolin *et al.*1976; Shcherbakov *et al.* 1976; Fedorov 1982; Serebryanny 1982; Mel'nik 1999) (Figure 9B).

Similar phenomena were noted by Shmuratko (1991) in an examination of many deep ocean oxygen isotope (^{18}O) records; he estimated the beginning of the post-glacial transgression at 19,000 years ago, as well as the period of accelerated sea-level rise at 13,000–10,000 years ago (Figure 10). This was also noted by Shimkus (1975) with regard to the Mediterranean Sea. Significantly, as indicated above, the period of 12,300–11,000 years ago (the Borshchevo thermophase), was marked by substantial climate warming in the entire Black Sea catchment basin and considerable deepening of the Bosphorus Strait.

These and other writers report very high rates of Black Sea level rise at certain stages, reaching 10 mm per annum. Even higher rates, exceeding 25 mm/year, may be expected if one takes into account the natural rhythm of sea-level rise and the acceleration of sea-level rise during the initial parts of various stages (Figure 6). This approach may be instrumental when attempts are made to develop scenarios of sea-level fluctuations for future decades.

5. TECTONICS AND GEOSTRUCTURAL PARAMETERS OF THE COASTAL ZONE AND BOTTOM OF THE BLACK SEA

Until now, the tectonic scheme for the Azov-Black Sea basin suggested by Tkachenko *et al.* (1970) has been neither criticized nor rejected. Indeed, its main elements have been confirmed by other investigators (Shnyukov 1984a, 1984b; Shnyukov *et al.* 1999). The evidence obtained has clarified the structure of the upper parts of the earth's crust and the initial pattern of the Azov-Black Sea bottom (Figure 11). It should be stressed that recent tectonics greatly influenced the structure of both modern and ancient coastal landforms in the Azov Sea littoral zone (Klyuva and Furtes 1970).

Based on research results, a schematic map of neotectonic structures was compiled (Figure 11). As one can see, the majority of shorelines follow fault lines and tectonically weakened zones. At a number of sites, particularly in western Crimea, one will note fault coasts of linear direction that arose during the Alpine Orogenic Phase. In general terms, the submergence of the Black Sea bottom follows the fault line that resulted in the formation of an extremely steep continental slope (up to 33°). The slope is particularly steep along the coast of Asia Minor, Crimea, and Caucasus (Yaltırak *et al.* 2002). As a consequence,

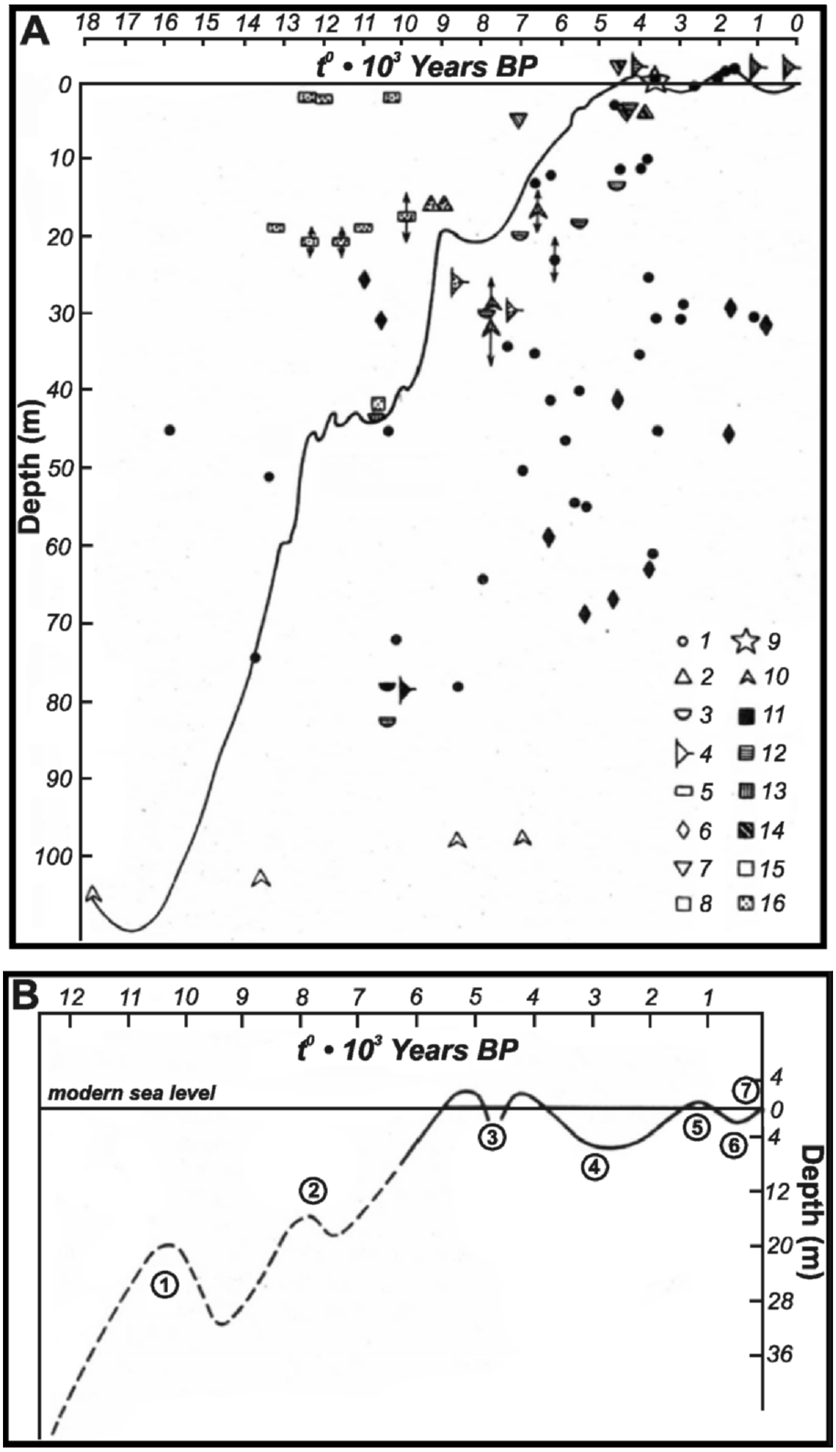
A
t° • 10³ Years BP
Depth (m)
B
t° • 10³ Years BP
modern sea level
Depth (m)

Figure 9. (Facing page) Graphs of sea-level changes in the Black Sea during Late Würm and Holocene times. **A.** Compared to a curve of glacial-eustatic rise in the world ocean, data points represent absolute dates; shapes indicate different laboratories (1–10) and shading fill indicates Black Sea locations from where they were taken: 11 — Pitsunda; 12 — Sochi-Adler; 13 — Kerch Strait; 14 — Gagra Bay; 15 — between Sarich and Meganom capes; 16 — northwestern shelf of the Black Sea. Data from L.R.Serebryanny (Blagovolin *et al.* 1976). **B.** Phases of the postglacial transgression according to stratigraphic and geomorphological investigations: 1 — Neoeuxinian; 2 — Old Black Sea; 3 — New Black Sea; 4 — Phanagorian; 5 — Nymphaean; 6 — Korsunian; 7 — modern (according to Fedorov 1982, 1988).

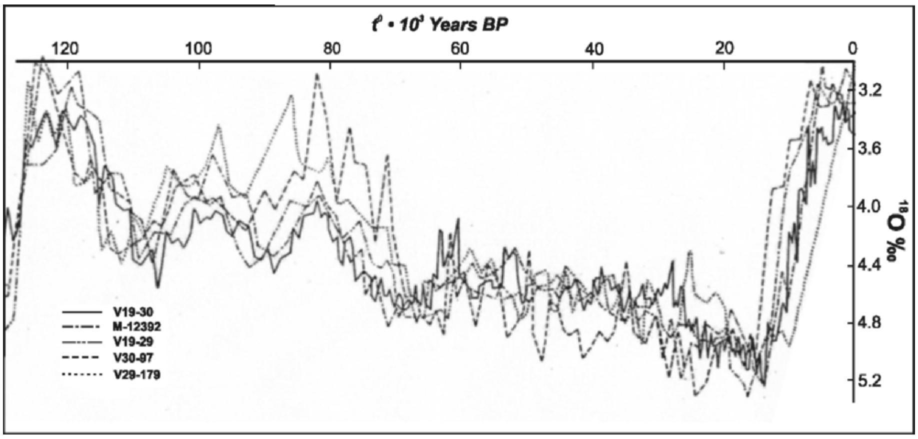

Figure 10. Graphs of changes in oxygen-isotope content in shells from benthic foraminifera of different ages in bottom deposits of the world ocean: V19-30, M-12392, V19-29, V30-97, V29-179 represent typical core samples taken in different areas (Shmuratko 1991).

recent sediments have been rapidly removed from these continental slopes (Fairbridge 1946). They have been localized within underwater canyons and carried down the continental slope to the deep sea bottom. Frequent underwater landslides result in slippages of largely intact sedimentary blocks.

The initial tectonic dislocation of the coastal zone in the contact region of the Scythian Plate, Russian Platform, and Black Sea depression led to the fragmentation of southwestern Crimea and the formation of arcs within Sebastopol Bay. The block disjunction either sinking to the Black Sea bottom or remaining as part of the Crimean anticlinoria resulted in the emergence of the steep mountain slope with landslide shores, high cliffs, and steep submarine slope. Block structure of the coastal area is equally typical of the Bosphorus Strait.

Therefore, relief fragmentation during the pre-transgression period, and the Pleistocene in general, greatly affected the formation of the shelf and coastal landforms in the Azov-Black Sea basin. The East European Platform was the main tectonic element of the northern Black Sea zone. In the process of its interaction with the Hercynian geosynclines of Dobruja, the Scythian Plate, and the Alpine geosynclines of Crimea and Caucasus, a complex system of faults,

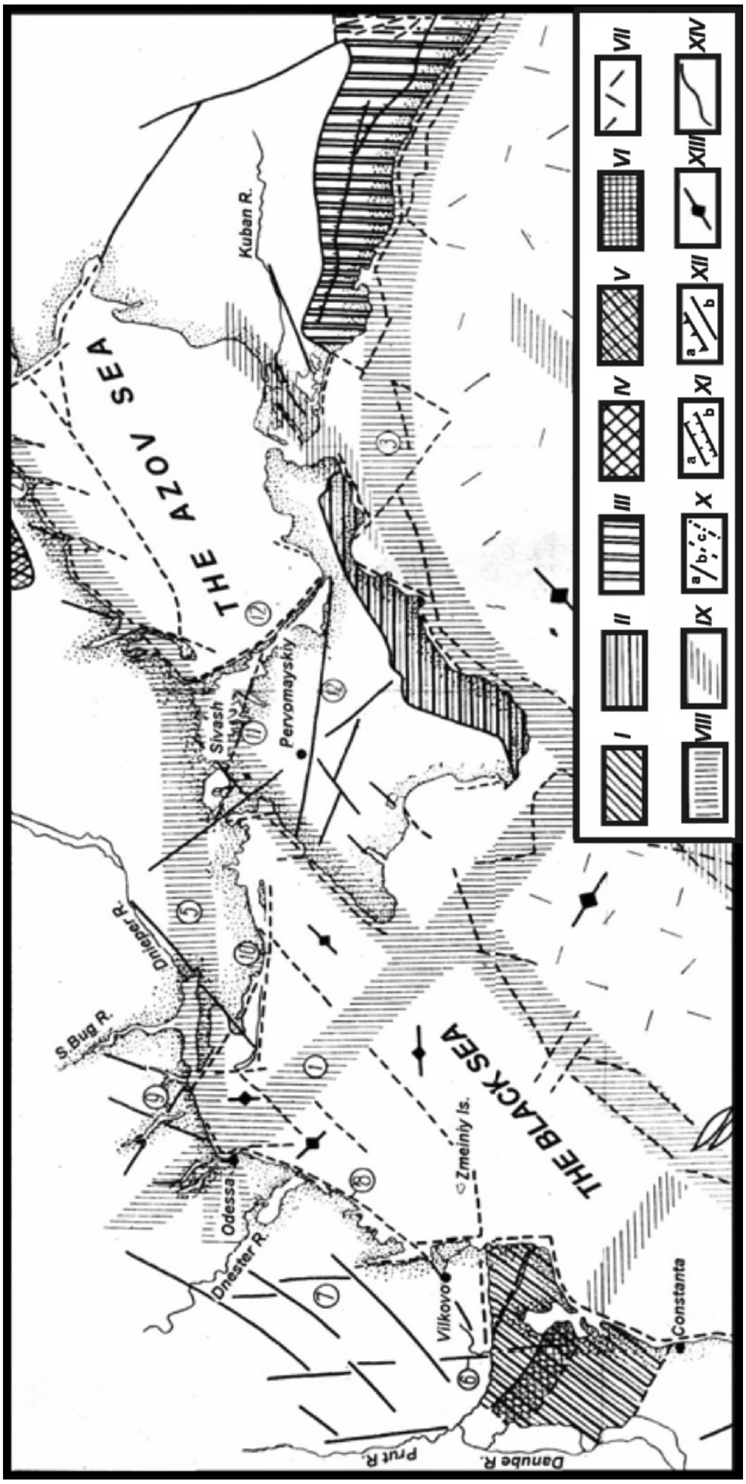
THE AZOV SEA
THE BLACK SEA
Kuban R.
Sivash
Pervomayskiy
Dnieper R.
S.Bug R.
Odessa
Dnester R.
Zmeiny Is.
Vilkovo
Constanta
Prut R.
Danube R.
I
II
III
IV
V
VI
VII
VIII
IX
X
XI
XII
XIII
XIV

Figure 11. (Facing page) Schematic of the neotectonic structure of the Azov and Black Sea basins (Tkachenko *et al.* 1970). Basic tectonic elements: I — Dobruja; II — meganticlinorium of the Crimean Mountains; III — meganticlinorium of the Big Caucasus; IV — meganticlinorium of the Old Planina; V — Ukrainian Crystalline massif; VI — platform mantle, closely folded by Alpine folding; VII — the Black Sea basin; VIII — new tectonically unstable zones concurrent with deep faults of Pre-Cambrian age; IX — the same zones, but conjectural; X — faults of different age and depth of initial stage: *a*) legible/reflected in relief of land, *b*) legible/reflected in shorelines and relief of sea-bottom, *c*) conjectural neotectonic faults; XI — upthrown faults and pivotal faults: *a*) large overthrusts; XII — borders of tectonic kettle depressions (*a*) and faults of unstated types (*b*); XIII — strike direction of structural plan of consolidated base; XIV — shoreline. Largest new tectonically unstable zones: 1 — the main Black Sea; 2 — Burgas-Zhdanovskaya; 3 — Central Crimean and Caucasian; 4 — Coastal Anadolu; 5 — Odessa-Sivash. The basic faults: 6 — Kishinev-Constanţa; 7 — Frunze-Arciz; 8 — Vilkovo-Odessa; 9 — Kinburnskii; 10 — Tendrovskii; 11 — Southern Sivash; 12 — Pervomaiskii; 13 — Arabatskii.

numerous shallow anticlines, synclines, and disjoint structures arose. All negative structures attracted the existing streams and stimulated active erosion. They became especially active during the regression phases leading to the exposure of the modern shelf and increased humidity.

North of the coastal territory, there appeared a system of consequent and submeridional compression faults. They laid the foundation for river valleys with incision of up to 50–60 m. The tectonically weakened Odessa-Sivash and Burgas-Zhdanov zones developed along the Black and Azov Sea coastlines.

Within the North Crimea Plain and the adjoining sea floor, uplifted sections of the folded basement form the broad Tarkhankut-Novoselovo uplift zone. Sublatitudinally-directed folds—the Peschanaya, Mezhvodnenskaya, and Bakalskaya brachyanticlines—are located on its northern flank, while smaller sublatitudinally-directed synclinal structures are located in between. The Kerch Peninsula has a similar structure. In western Crimea, some positive ridges (Tarkhankut, Dzhangul, Yarylgach, Oyrat, and others) as well as negative geostructures (Yarylgach, Karadzha, Ribachya, etc.) are faulted capwise. A complex Alma depression lies to the south, and the Indol foredeep lies to the east of the Novoselovo anticlinal dome. Hence, due to the complexity of the structure and geomorphology of the coastal area, its various parts reacted differently during the transgression (Shuisky and Bertman 1975; Nikiforov 1977).

Initial differentiation led to the alternation of positive and negative structures on various scales. During the Neoeuxinian regression, the Black Sea shelf was exposed to a depth of –110 m. Surface structures of the broad shelf area have undergone morphological changes, mainly erosion and denudation, partially marine and aeolian in nature. The rivers have produced a complex hydrological network with terraces along the fault lines (Figure 12), and thick terrigenous sedimentary sequences accumulated on the Neoeuxinian surface.

During the post-glacial transgression (13,000–11,000 years ago), marine waters invaded the mouths of river valleys to form estuaries on the northern coasts of the Black and Azov Seas. A similar situation arose where the trans-

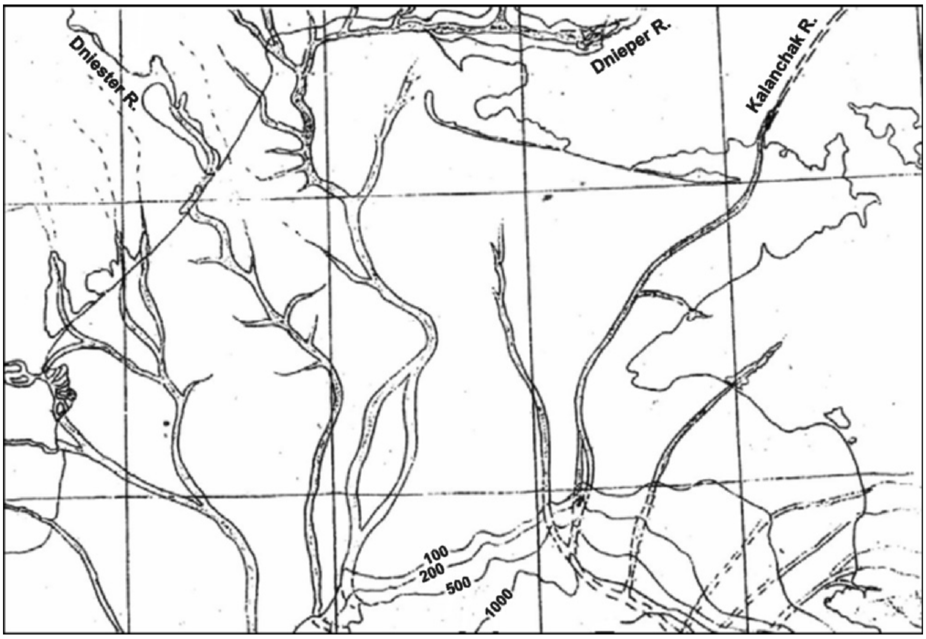

Figure 12. A schematic map of the Post-Pleistocene river network on the surface of the wide northwestern Black Sea shelf (from Shnyukov *et al.* 1999).

gressive waters flooded structural depressions. Hence, a sinuous shoreline was formed. Subsequently, the estuaries were isolated from the open seas by baymouth barriers, spits, and barrier islands, forming limans, lagoons, and rias. The relatively flat coastal profiles with high cliffs and steep underwater slopes emerged in coastal areas intersected by faults of high amplitude (up to 30–80 m).

6. RECENT CONCEPTS REGARDING POST-GLACIAL SEA-LEVEL FLUCTUATIONS

The aforementioned data pertaining to the development of the Black Sea basin, its relief, and long-term fluctuations of sea level have generally been supported in recent publications (e.g., Aksu *et al.* 2002a, b; Siddall *et al.* 2003; Kerey *et al.* 2004). At the same time, new evidence has been obtained based on investigations of the structure and texture of the bottom deposits, the content and properties of pore water in the bottom sediments, and the geological structure of the bottom and upper crust in the vicinity of the Bosphorus Strait. In several cases, man-made structures of varied date have been recovered from the Black Sea bottom at different depths. Based on all this, Ryan *et al.* (1997) have proposed their 'Flood' hypothesis.

The supporters of this hypothesis dwell on the assumption that, as the melting glaciers of northern Europe constantly supplied the Black Sea with fresh water during its 'lake stage', sea level could have dropped to –140 m, and that the Flood catastrophe, occurring during the post-glacial period at 6700 BC, was followed by a period of rapid warming at 5800 BC (Haarmann 2004). This resulted from an increase in air temperature by 5–10° C that caused an intensive melting of glaciers and an influx of fresh water into the Black Sea (Preisinger and Aslanian 2004). During the proposed catastrophic inflow of saline water into the Black Sea depression through the Bosphorus Strait, the powerful entry stream up to 2 km wide moving at a velocity of 10 m/sec eroded the Bosphorus submarine canyon, the depth of which presently exceeds 80 m (Kerey *et al.* 2004; Siddall *et al.* 2004). As several of these conclusions partly coincide with the results of previous investigations, the Flood hypothesis may be partly correct, though its many points meet serious objections.

First, one should define the meaning of 'catastrophic flooding'. As the most reliable genetic criterion, one should consider the 'negative' reaction of the coast to particular rates of sea-level rise. At various points along the Black Sea coast and at various stages of Holocene sea-level history, these values varied from 18 to 27 mm/year (Shcherbakov *et al.* 1976; Shuisky *et al.* 1995; Shuisky 1999). The rate was affected by numerous factors: water balance, tectonic movements of offshore and coastal areas of the Black Sea, sedimentation of clastic deposits, temperature variations within the active dynamic water layer, compaction of clastic deposits, and others. Under natural conditions, Black Sea water intake cannot exceed that specified by the above cited critical rates. In view of that, the nearly instantaneous sea-level rise suggested by the Flood hypothesis is deemed purely speculative and not consistent with the realities of the relative sea-level fluctuations.

The proposal for catastrophic level change in the Black Sea and flooding of its shelf is based on the assumption that a huge water stream entered the Pontic basin from the Marmara Sea. Even if one accepts that Niagara-like cascades might naturally exist, velocities of up to 10 m/sec are unattainable for the cross-section of the Bosphorus (Ryan *et al.* 1997; Siddall *et al.* 2004). Besides, most researchers share the view that Holocene sea-level changes in the Black Sea followed the pattern discussed in sections 3. and 4. of this paper.

For example, Siddall *et al.* (2003) plotted the sea-level curve for the world ocean in comparison to that of the Red Sea (Figure 13). These curves are similar to those discussed above (Figures 4 and 7). Other authors also accept that all these curves are similar and reliable. None of them show a sea-level rise in the world ocean on the order of hundreds of meters. The fluctuations in the Black Sea over the past 20,000 years, including the second half of the 20th century, proceeded within the context of these changes in the level of the world ocean.

As to the Black Sea, the evidence of its water-level fluctuations is often

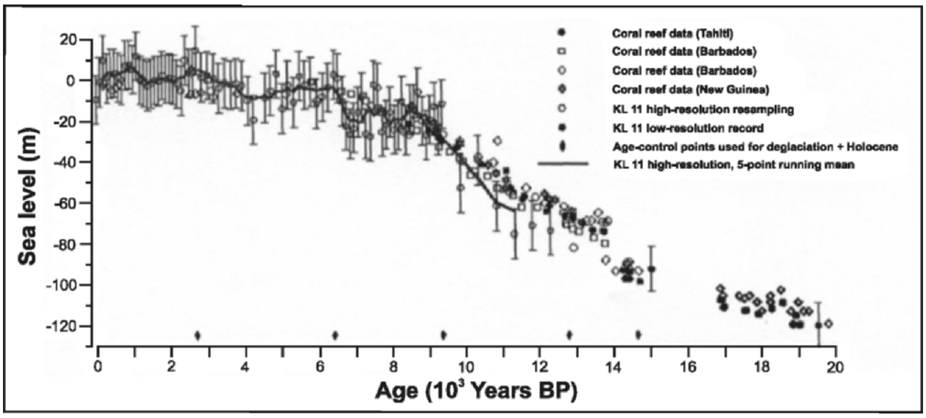

Figure 13. Sea-level reconstruction since the start of the Younger Dryas for the Red Sea, based on the $\delta^{18}O$ record from core KL11 (18º 44.5´ N, 39º 20.6´ E), including error bars of ±12 m. Chronology is based on calibrated AMS radiocarbon dates. The record is zeroed to modern sea level by removing the mean KL11 record for the past 7 kyr. The point from the low-resolution record of KL11 reported in the LGM is based on intercalibrated benthic $\delta^{18}O$ data. Black symbols are sea-level values obtained from coral reef studies (according to Siddall *et al.* 2003).

equivocal. For several decades until the 1990s, intensive studies were conducted on its geological history, paleogeography, and Pleistocene-Holocene sea-level changes, as indicated in previous sections of this paper. They yielded reliable and convincing evidence outlined in sections 3. and 4. Figures 6, 8, 9, and 10 graph the curves of Black Sea level changes following Würm II, plotted by using various techniques and different theoretical approaches. Recently, curves were obtained by other writers (cf. Aksu *et al.* 2002a) that appear fundamentally the same as those generated by previous researchers some 30 years ago (Figures 8 and 9). These convergences affirm the validity of previous research, and they totally contradict the Flood hypothesis.

First, one must consider the geographical position of the Black Sea. The melting Fennoscandian ice sheet and Alpine mountain glaciers could fill in the Black Sea with only a limited quantity of water. During Würm I and II, the Scandinavian ice sheets and their meltwater had no access to the Black Sea. The discharge of glacial meltwater from the Alps was directed mainly to the north and south; it had only minimal access to the east. Hence, during the last 20,000 years, fluctuations in the level of the Black Sea constituted a regional phenomenon.

Other authors also cite convincing evidence that contradicts the Flood hypothesis, such as that emerging from detailed studies of shelf deposits (Aksu *et al.* 2002a, b). Preisinger and Aslanian (2004) argue that, at about 11,500 years ago, temperature increased by 5–10° C over a span of 100–200 years. The physical properties of the atmosphere make this suggestion improbable, however. Neither can this be confirmed by evidence from pollen analysis of sea floor deposits, which does not support the contention and, in contrast, indicates that the succeeding cryophase began *ca.* 11,000 years ago, at the beginning of the late Neoeuxinian (Shcherbakov *et al.* 1976; Zubakov 1992). Siddall *et al.* (2004) argue that later, at about 7300–7000 years ago, an enormous stream of saline water invaded the Black Sea with a velocity of up to 10 m/sec, a rate that is impossible due to the Black Sea's physical geography at the time (Blagovolin *et al.* 1976; Nevessky 1967; Shcherbakov *et al.* 1967). It is difficult to accept the conclusion that, under the impact of this intrusion of saline water, sand bars were formed on the Black Sea bottom, and the Bosphorus underwater canyon was eroded. With a current velocity of 5–10 m/sec, sand bars would have been washed away; the canyon has a tectonic-subsidence origin.

Based on biostratigraphical and lithological studies of bottom deposits in the Bosphorus Strait from both the Black and Marmara Sea Gates and radiocarbon dating, several writers (Kerey *et al.* 2004) arrived at the unusual conclusion that the Bosphorus bottom deposits are only as old as the Early-Middle Holocene. As Ryan *et al.* (1997) and Siddall *et al.* (2004) argue, water from the Marmara Sea was pouring into the Black Sea as a powerful stream at the time, but under such conditions, the character of the sediments would be different and inconsistent with the existing sedimentary environments within the Bosphorus Strait.

As the analysis of numerous recent scholarly publications shows, there exist considerable differences in the conclusions and interpretations offered for the last 20,000 years of Black Sea history, with wide discrepancies on several points including that of sea-level fluctuations (Figure 13). It is quite natural if one considers the wide spectrum of methods used. Yet, the general picture presented by Aksu *et al.* (2002a) is quite similar to that of a great number of other researchers (Figure 14).

7. CONCLUSIONS

Fluctuations in the level of the Black and Azov Seas during the Neoeuxinian and Black Sea stages were complex. The general trend consisted of a sea-level rise from –100/–110 m to the present level following the pattern of the world ocean. A glacial eustatic character is deemed the most probable for the Holocene fluctuations.

During the Holocene transgression, the fluctuations of the Black and Azov Seas proceeded unevenly, with accelerations and decelerations. During periods of acceleration, the rate of sea-level rise could reach 50–65 mm/year. Possibly, these rates, and even higher ones, may occur in the near future.

Coastal landforms of today resulted from the effect on various tectonic structures surviving from the Pleistocene of surficial processes and wave-induced

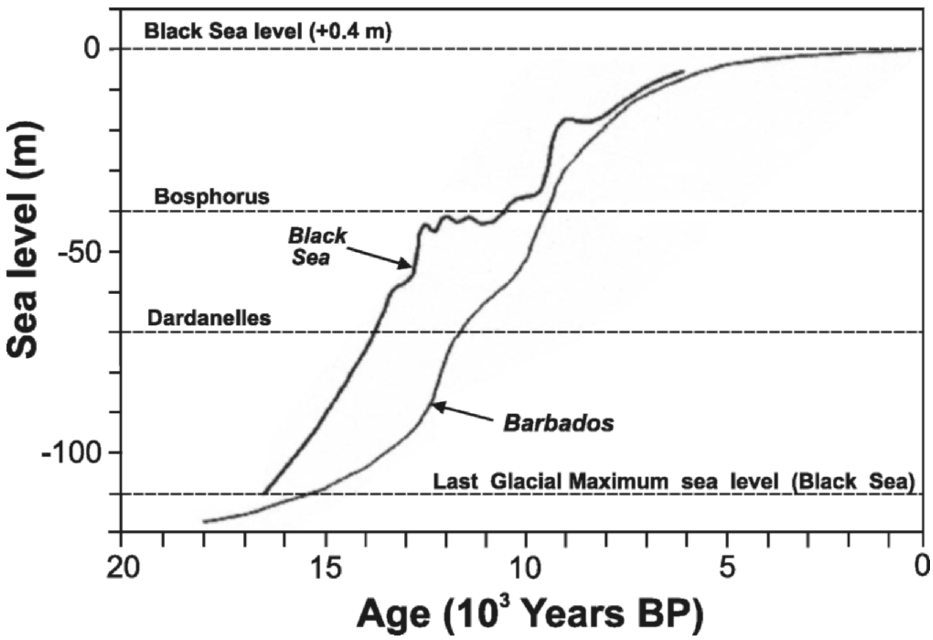

Figure 14. Post-glacial sea-level curve for the Black Sea compared to that of the Atlantic Ocean as represented by Barbados Island (after Aksu *et al.* 2002a).

reworking over the course of the Holocene transgression. They are characterized by a variety of erosional and accretionary landforms, geological structures, and variations in wave impact.

The evolution of modern shorelines led to their structural differentiation. Shorelines migrated differently during various stages of the Holocene transgression, and hence, one might suggest that an accelerated sea-level rise in the future could affect various parts of the coast differently.

Available evidence does not support the Black Sea 'Flood' scenario, but the findings of all research on this problem are important contributions to the database of the NATO Project "Climate Change and Coastal Migration."

ACKNOWLEDGMENTS

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11

THE MIDDLE PALEOLITHIC AND EARLY UPPER PALEOLITHIC IN THE NORTHERN BLACK SEA REGION

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Abstract: Typological and technological characteristics of the lithic industries produced during the Middle and early Upper Paleolithic in Eastern Europe are summarized. Chronologically, the assemblages span the interval between the Last Interglacial and the Arcy/Denekamp Interstadial. Their regional distribution is described as well as the paleoclimatic implications of environmental indicators.

Keywords: Eastern European Middle and Upper Paleolithic, Micoquian, Levallois-Mousterian, Blade Mousterian, Spitsynskaya, Streletskaya, Gorodtsovskaya, Aurignacian, Early Gravettian

1. INTRODUCTION

As in Western Europe, there is considerable diversity within the Middle and Upper Paleolithic industries defined for Eastern Europe. The Middle Paleolithic of Eastern Europe may be viewed as including at least three quite distinct techno-complexes that can be distinguished by their technological and typological characteristics: the Micoquian, Levallois-Mousterian, and Blade Mousterian (Chabai 2003).

The early Upper Paleolithic of Eastern Europe possesses five named techno-complexes: the Spitsynskaya, Streletskaya, Gorodtsovskaya, Aurignacian, and Early Gravettian (Rogachev and Anikovich 1984), as well as at least two lithic assemblages yet unnamed (Sinitsyn 2000).

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2. TECHNOLOGICAL AND TYPOLOGICAL CHARACTERISTICS OF THE MIDDLE PALEOLITHIC INDUSTRIES

Micoquian assemblages have been found in all archaeologically investigated regions of Eastern Europe (Kolesnik 1994; Chabai 1996; Sytnyk 2000; Golovanova and Hoffecker 2000). Symmetric and asymmetric plano-convex bifacial tools have been recognized as characteristic of the Micoquian (Bosinski 1967; Kulakovskaya *et al.* 1993), and their appearance together with specific evidence of their manufacturing method has become the diagnostic criterion that distinguishes the Micoquian from other Central and Eastern European Middle and early Upper Paleolithic industries.

In general, the Micoquian of Eastern Europe is technologically homogeneous but proportionately variable across its range of tools (Chabai 2003:91). The raw material exploitation technologies are based on both non-Levallois flake core reduction and the plano-convex method of bifacial tool production. All the recognized proportional variations within tool kits are found in Crimea, where the greatest number of Micoquian occupations is known, and from which the most material has been published. The tool kit varieties of Micoquian assemblages in other regions probably reflect the same economic pursuits and differential availability of raw material as in Crimea. In fact, the Crimean facies of the Micoquian exhibits more typological variability than all other Eastern European Micoquian assemblages combined.

The Levallois-Mousterian is found in two areas: the Prut-Dniester basins and Crimea (Chabai 2000a; Sytnyk 2000). Prut-Dniester industries with a pronounced Levallois component are called the "Molodova Mousterian Culture"; in Crimea, they are the "Western Crimean Mousterian." In both areas, common technological and typological characteristics of Levallois-Mousterian assemblages appear: (1) a combination of Levallois tortoise with uni- and bidirectional blade technologies, (2) a dominance of simple scrapers with a relative rarity of convergent scrapers, denticulates, and notches, and (3) the use of flat, non-invasive scalar retouch. No evidence exists for any bifacial tool technology in the Levallois-Mousterian of Eastern Europe. The main difference between the Levallois-Mousterian and Micoquian in Eastern Europe lies in the fundamentally different technologies used for both blank and tool production.

Blade Mousterian assemblages are found only in the Don River basin and its tributary, the Seversky Donets basin (Kolesnik 1994; Nehoroshev and Vishnyatsky 2000). Here, the core reduction strategy is based solely on unidirectional and bidirectional volumetric cores. Tool kits include points, simple and double scrapers, denticulates, and notches, while convergent scrapers are rare. The most distinctive types are the widely used truncated-faceted and bitruncated-faceted pieces. Thus, Blade Mousterian assemblages have both a typologically and technologically distinctive repertoire compared to the Micoquian. Unlike Micoquian and Levallois-Mousterian assemblages, however, those of the Blade Mousterian are very homogeneous and do not show any significant technological, typological, chronological, or geographic variations.

3. TECHNOLOGICAL AND TYPOLOGICAL CHARACTERISTICS OF THE EARLY UPPER PALEOLITHIC INDUSTRIES

The Streletskaya reveals the broadest distribution of any early Upper Paleolithic industry. They are found in the Middle and Lower Don, Crimea, and the Central and Northern Urals (Rogachev and Anikovich 1984; Guslitzer and Pavlov 1993; Marks 1998; Matioukhine 1998). Common technological and typological features include the production of thin, biconvex bifacial tools, the presence of bifacial leaf-shaped and triangular points (many of the latter with concave bases), and fan-shaped, laterally retouched end scrapers on flakes, sometimes with a thinned base. Though defined as Upper Paleolithic, the Streletskaya reveals no blade technology and contains very few burins.

There is only one reliable Spitsynskaya assemblage: Kostenki 17, layer II, in the Middle Don Valley (Boriskovsky *et al.* 1982). The Spitsynskaya employed only a blade technology based on parallel, “prismatic,” single and double platform cores with volumetric flaking surfaces. Burins dominated, representing about 48% of the tool kit. Bone and ivory artifacts are represented by awls made from hare and polar fox *humeri*, three fragmentary bone points, and one fragment of worked mammoth tusk. Personal adornments included pendants made from polar fox teeth, stone, belemnites, shells, and fossil coral.

The Gorodtsovskaya assemblages are recovered from the Middle Don region (Sinitsyn 1996, 2000). Technologically, both flakes and blades were produced from unsystematic and parallel cores, the latter usually exhibiting a volumetric flaking surface. Bifacial tools are rare and unstandardized. The most common parallel-edged end scrapers are thick and abrupt, while the fan-shaped end scrapers often have ventral thinning. Double end scrapers with retouched lateral edges are common and resemble limaces, which are also present in Gorodtsovskaya assemblages. Other tools include scaled pieces of various forms and both dihedral and angle burins. The “archaic Mousterian element” includes simple scrapers, transverse scrapers, canted and double-canted scrapers, convergent scrapers, points, limaces, and small bifacial tools. The bone and ivory artifacts reveal a striking variety of shapes: “shovels” with nail-like heads made from mammoth long bones, points of different types (including needle-like ones), awls, retouchers, polishers, and pendants. The shafts of the "shovels," bone points, and tubular bone fragments were decorated with complex bands of geometric incisions. In sum, the Gorodtsovskaya stone and bone industries have a distinct character, both technologically and typologically.

Stratified *in situ* Aurignacian assemblages in Eastern Europe occur in the Prut River Valley, in Crimea, and in the Middle Don Valley (Rogachev 1957; Hahn 1977; Otte *et al.* 1997; Demidenko *et al.* 1998). Those from Crimea and the Middle Don belong to the Krems-Dufour variant of the Aurignacian and possess a pronounced bladelet and microblade component, which accounts for about 50% of all blanks, while true blades do not represent even 20%. These bladelets and microblades are associated with carinated cores, end scrapers, and "regular" bladelet/microblade cores. Dufour and pseudo-Dufour bladelets and microblades make up about half of all tool kits. A few Krems points occur, while the rest of the tools are typically Aurignacian: blades with "Aurignacian-like" retouch, carinated end scrapers and burins, thick nosed/shouldered scrapers, bone points and awl fragments, and marine shell pendants.

The typological definition of Mitoc Malul Galben from Prut is more complicated. According to M. Otte, it was a specialized workshop for bladelet production, utilizing both bladelet and carinated core reduction (Otte *et al.* 1997:282): "It must be noted that no Dufour bladelet has ever been found at Malul Galben; but this workshop and other finds from the recent excavations confirm that they were certainly an important part of the lithic production." Some carinated burins and Mladeè points were recovered from Aurignacian occupations as well. If the Mitoc Malul Galben workshop was really oriented toward *lamelle Dufour* production, the Krems-Dufour type was probably the only Aurignacian variant in Eastern Europe (Demidenko 2000–2001:161).

The Early Gravettian occurs in two regions only: the Prut-Dniester basins and the Middle Don (Rogachev *et al.* 1982; Borziak and Kulakovskaya 1998). Assemblages from the Prut-Dniester area are chronologically and typologically true Early Gravettian, based on volumetric blade core exploitation. Those of the Middle Don reflect only its chronological position, however. The technological differences are striking, as the "Gravettoid" Middle Don assemblage of Kostenki 8, layer II, employed microblades.

4. CHRONOLOGICAL AND ENVIRONMENTAL CONTEXT OF MIDDLE AND EARLY UPPER PALEOLITHIC VARIABILITY

Chronological evidence in Eastern Europe indicates that (1) no known Middle Paleolithic assemblage can be dated before the Last Interglacial (Chabai

2003), (2) the latest manifestation of the Middle Paleolithic occurred during the Arcy Interstadial, *ca.* 32–28 ky BP (Chabai 2000b), (3) the earliest known Upper Paleolithic dates to about 38–36 ky BP (Sinitsyn *et al.* 1997), and (4) from about 38 to 28 ky BP, Middle and early Upper Paleolithic industries co-existed (Chabai 1996, 2000b).

Crimea currently possesses the most complete, the most detailed, and the best dated Middle Paleolithic chronology of Eastern Europe, based upon numerous AMS, U-series, and ESR dates, as well as biostratigraphic sequences (geological, pollen, snail, and microfauna). Sites such as Kabazi II, Starosele, Zaskalnaya V, and Buran-Kaya III have all produced significant chronological and environmental data (Hedges *et al.* 1996; Rink *et al.* 1998; McKinney 1998; Chabai *et al.* 1998; Pettitt 1998; Gerasimenko 1999; Markova 1999; Mikhailesku 1999; and Chabai 2000b).

This evidence, combined with information from other regions, permits a subdivision of the Eastern European Middle Paleolithic into three temporal units (Tables 1 and 2): (1) a First Period (*ca.* 125 to 60 ky BP) dating from the Last Interglacial through the Moershoofd Interstadial, (2) a Second Period (*ca.* 60 to 38 ky BP) including the Hengelo Interstadial and the previous Stadial, and (3) a Third Period (*ca.* 38 to 28 ky BP) including the Arcy/Denekamp Interstadial and the preceding stadial.

Taking into account the probable geographic isolation of Crimea during the Last Interglacial (Lazukov *et al.* 1981), the First Period can be divided into two stages (Figures 1A and B): (1) the Last Interglacial and (2) the Early Glacial. According to Chepalyga (1984:230), the Black Sea (the Karangatian transgression) was 8 to 12 m higher than today during the Last Interglacial, and the Caspian basin may have been connected to the Black Sea through the Manych Depression. If so, the Northern Caucasus was separated from the rest of Eastern Europe by this Manych Strait, and Crimea may have been an island. Much of Eastern Europe was covered by broad leaf forests together with open grass meadows (Bolikhovskaya and Pashkevich 1982; Grichuk 1984; Pashkevich 1987; Gerasimenko 1999, 2003; and Boguckyj *et al.* 2001).

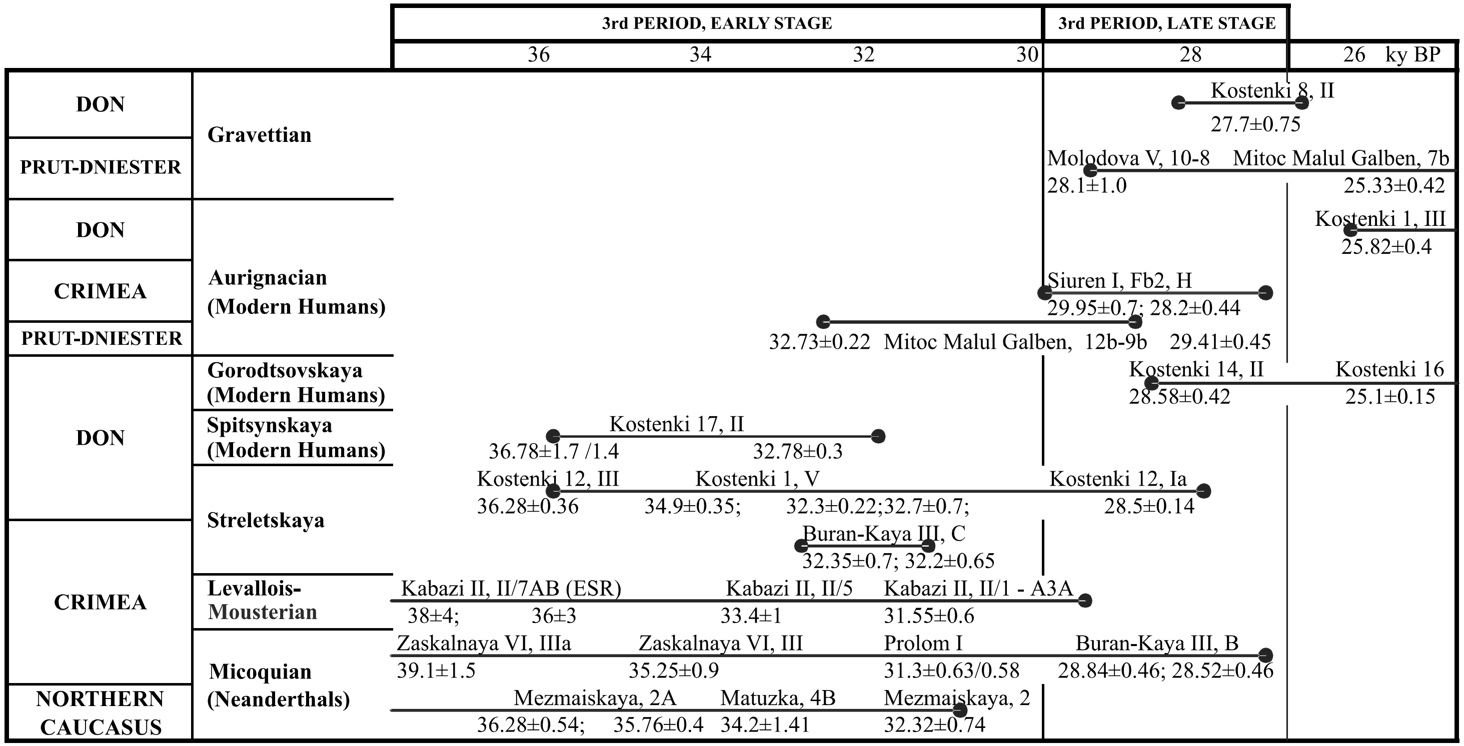
The well-dated Last Interglacial Micoquian is known only from Kabazi II. Micoquian deposits from the Northern Caucasus, Donbass, and Lower Volga likely date to the Last Interglacial as well. In the Prut-Dniester region, the earliest well-documented Middle Paleolithic manifestation is found in layer III of Yezupil, which contains a Levallois-Mousterian industry (Figure 1A).

During the Early Glacial and Early Pleniglacial, the post-Karangatian (pre-Surozhian) Regression dropped the Black Sea 100 to 110 m below its current level. The Caspian basin shrank, the Azov Sea disappeared, and the Northern Caucasus was joined to an enlarged northern Black Sea Plain that included Crimea (Alekseev *et al.* 1986). The climate of Eastern Europe became significantly colder and drier.

Table 1. The chronological position of the Middle Paleolithic industries of Eastern Europe: from the Eemian Interglacial to the Hengelo Interstadial.

ca. BP	OIS-18	CHRONO-STRATIGRAPHY		MICOQUIAN					LEVALLOIS-MOUSTERIAN		BLADE MOUSTERIAN
				Prut-Dniester	Crimea	Northern Caucasus	Donbass	Lower Volga	Prut-Dniester	Crimea	Seversky Donets-Don
40	3	Hengelo	2nd PERIOD	Ripiceni-Izvor, V	Starosele, 1, 2					Kabazi II, II/8-IIA/1	
50		Stadial		Ripiceni-Izvor, IV	Zaskalnaya V, III Zaskalnaya V, IV Kabazi II, IIA/2-3 Chokurcha I, IV	Monasheskaya, 2-4 Barakaevskaya Mezmaiskaya, 2B-1, 2 Mezmaiskaya, 2B-3; 3				Kabazi II, IIA/2	
60		Moershoofd	1st PERIOD		Kabazi V, II/3-II/4A Kabazi II, IIA/3-IIA/4B				Ripiceni Izvor, III Molodova V, 11a		
70	4	Stadial			Kabazi V, III/1-III/3 Kabazi II, III/1A-III/1						Shlyakh, 8C Belokuzminovka, 3 Zvanovka
80 90 100 110	5a 5b 5c 5d	Odderade ? Brörup ? Amersfoort ?		Yezupil, II Kolodiev, 12.5-12.9	Starosele, 4 Kabazi II, III/2 Zaskalnaya V, V Kabazi II, III/2A-III/7	Ilskaya I	Nosovo I Antonovka II Antonovka I	Sukhaya Mechetka	Ripiceni Izvor, II Molodova I, 4 Molodova V, 11 Molodova V, 12 Ripiceni Izvor, I Proniatin Igrovitsa I, II Bugliv V, II		Kurdumovka Belokuzminovka, 2
120	5e	Eem			Kabazi II, V/3-VI/17		Belokuzminovka, 1	Cheluskinets (?)	Yezupil, III		

Table 2. The chronological position of Middle Paleolithic and early Upper Paleolithic industries of Eastern Europe from 38 to 26 ky BP.



The Early Glacial and especially the Early Pleniglacial were associated with a sharp decline in broad leafed trees as well as with a general reduction of forested areas. Arcto-boreal vegetation and fauna became dominant in almost all regions of Eastern Europe. No such harsh climatic conditions existed in Crimea at this time, as boreal species of fauna and flora were uncommon (Gerasimenko 1999, 2003; Markova 1999; Mikhailesku 1999).

During the Early Glacial and Early Pleniglacial, the Micoquian was the only Middle Paleolithic industry known in Crimea, while the Prut-Dniester region was occupied mainly by groups using Levallois-Mousterian industries, except in two cases where Micoquian was found. Also, two different industries have been found in the Donbass region: Micoquian and Blade Mousterian (Figure 1B).

The Middle Pleniglacial (from Moershoofd to Arcy, corresponding to the Surozhian transgression) witnessed a significant rise in the Black Sea, reaching perhaps to that of today or 10 to 15 m lower (Chepalyga 1984:234). During most of the Early and Middle Pleniglacial, the Azov Sea probably did not exist. The Don River extended into the area, and the northeastern flowing Crimean rivers became its tributaries (Alekseev *et al.* 1986:172–176).

Middle Pleniglacial environments differed considerably by region, and in general, they were not as harsh as those of the Early Pleniglacial. No boreal flora or fauna have been found in Crimea, where the landscapes have been characterized as forest-steppe/steppe, with pine as the main arboreal genus (Gerasimenko 2003). Pine was also important in the Prut-Dniester forest-steppe and steppe landscapes together with spruce (Bolikhovskaya and Pashkevich 1982; Pashkevich 1987; Pãunescu 1993). The reconstruction of forest-steppe and steppe environments for Crimea and the Prut-Dniester region has been supported by faunal remains, which are represented by *Saiga tatarica*, *Equus hydruntinus*, *Equus caballus* (sp.), *Mammuthus primigenius*, and rarely *Cervus elaphus* and *Rangifer tarandus* (Alekseeva 1987; López Bayón 1998; Burke *et al.* 1999; Patou-Mathis 1999). According to Alekseeva (1987:160), the appearance of reindeer in the Dniester Valley might be explained by the autumn-winter migration out of the northern warm season habitats.

In Crimea, the Levallois-Mousterian appears as the Western Crimean Mousterian, and the Micoquian continues. Both industries have been found at a number of localities. In the Prut-Dniester region, the Levallois-Mousterian disappears, but a clear Micoquian has been found in layers IV to V at Ripiceni-Izvor, the temporal position of which was established by biostratigraphic studies and radiocarbon dates. In the Northern Caucasus, three sites of this period contain Micoquian assemblages, which are thus well represented (Figure 2). The absence of Middle Paleolithic materials from the Donbass, Lower Volga, and the Middle Don regions, particularly during the Stadial preceding Hengelo, may well reflect the extremely cold and dry conditions in those areas.

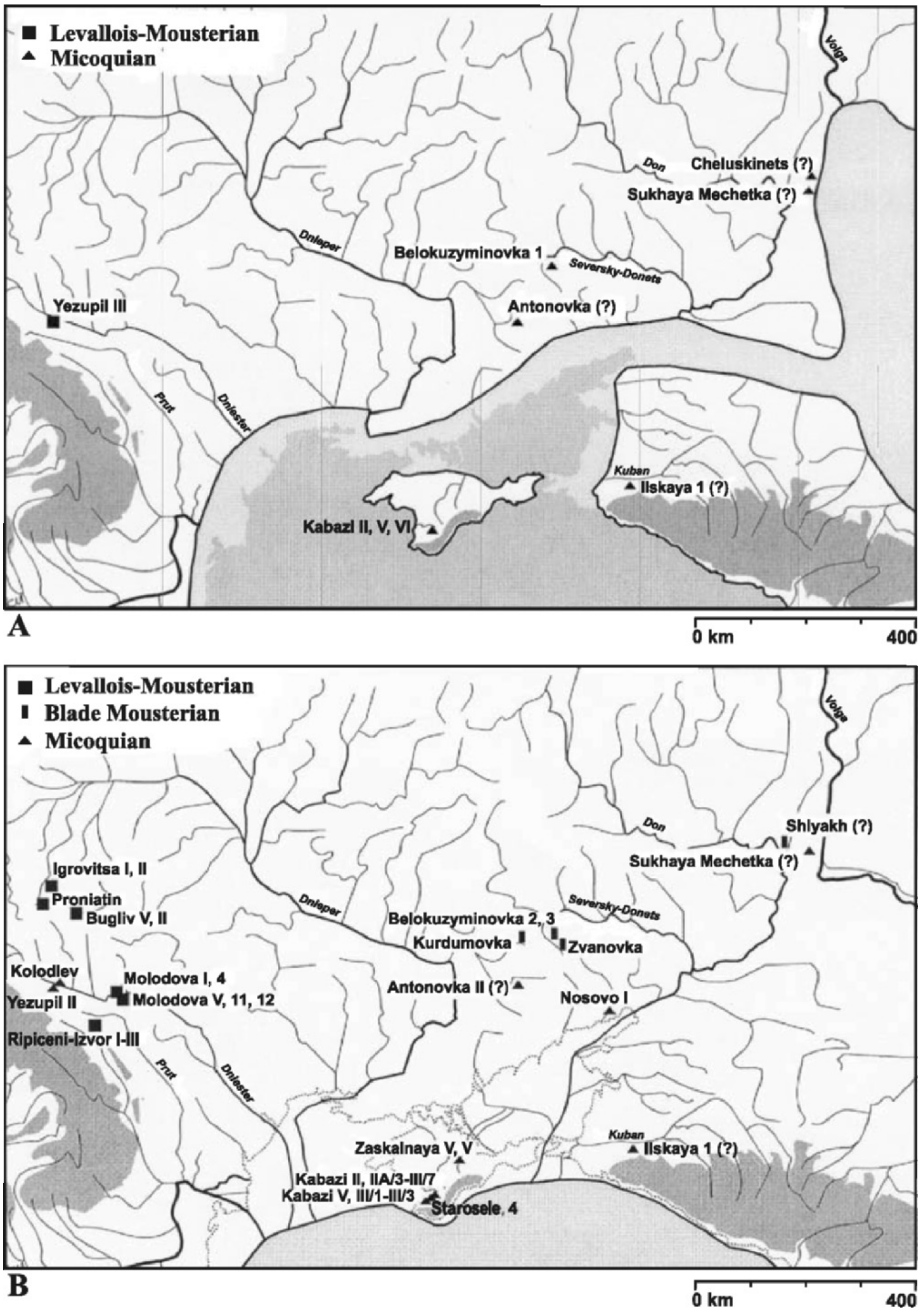


Figure 1. The First Period of Eastern European Middle Paleolithic; A: the Last Interglacial, B: the Early Glacial through the Moershoofd Interstadial.

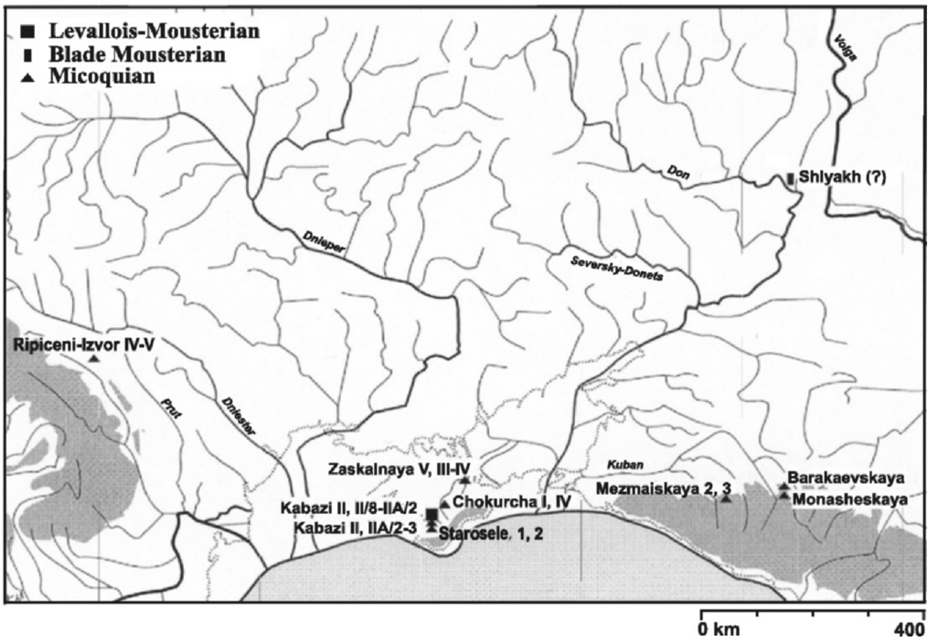


Figure 2. The Second Period of the Eastern European Middle Paleolithic: post-Moershoofd Stadial through Hengelo Interstadial.

In the Middle Don, pollen evidence indicates a humid forest vegetation of northern taiga type at 37–36 ky BP (Malyasova and Spiridonova 1982:237–238). Arboreal pollen dominates, principally spruce, birch, and pine. By the time of the local Bryansk Interstadial (Arcy) at 30–28 ky BP, this taiga forest was somewhat drier and of southern type. According to Markova *et al.* (2002:394):

> Taiga communities did not have a continuous distribution during Bryansk time but were restricted to "island" configurations… These "islands" continued into the highlands of the Russian Plain (the Valdai, the Middle Russian, the Donetsk highlands, etc.).

Also, the dominant faunal remains (*Equus latipes*) correlate well with the taiga environment. These horses were adapted to the "relatively soft surface of forest-steppe and forest biotopes" (Vereshchagin and Kuzmina 1982:227). During the Upper Pleistocene, the range of *Equus latipes* extended into the southern borders of the Urals and Siberia (Vereshchagin and Kuzmina 1982:229). Hare (*Lepus tanaiticus*) played the second most important role in the Middle Don taiga faunal assemblages. Its distribution at the time included the Russian Plain, Urals, and Eastern Siberia extending up to the shore of the Arctic Ocean (Vereshchagin and Kuz'mina 1982:231).

Thus, during the Middle Pleniglacial, the Middle Don was ecologically quite distinct from the coeval forest-steppe environments of Crimea and the Prut-Dneister. With the onset of the Late Pleniglacial, however, the Middle Don taiga forests were replaced by open forest-steppe and steppes (Malyasova and Spiridonova 1982:245).

The Third Period ranges from *ca.* 38 ky BP until somewhat after 28–27 ky BP. Human groups of the time employed a number of Middle and early Upper Paleolithic industries, including Micoquian, Levallois-Mousterian, Gorodtsovskaya, Spitsynskaya, Streletskaya, Aurignacian, and Gravettian. In Crimea, the period begins with the appearance of a clear Upper Paleolithic Streletskaya assemblage, coeval with Levallois-Mousterian and Micoquian occupations. Micoquian occupations are also known from the Northern Caucasus. In the Prut-Dniester region, the beginning of the Third Period is marked by the appearance of Aurignacian at Mitoc Malul Galben, while the Middle Don region was populated by Upper Paleolithic peoples producing Streletskaya and Spitsynskaya industries (Figure 3A).

The late phase of the Third Period in the Prut-Dniester region saw Aurignacian and Gravettian industries at Mitoc Malul Galben and Molodova V. Aurignacian assemblages were widespread at the time in Crimea, where they co-existed with Micoquian and Levallois-Mousterian, and in the Middle Don region, where they were accompanied by other Upper Paleolithic industries, such as the Gorodtsovskaya, Streletskaya, and Gravettian (Figure 3B).

5. DISCUSSION

Between the Last Interglacial and the Moershoofd Interstadial, Crimea was populated by Micoquians, while in the Azov-Donbass and Prut-Dniester areas, Micoquians were accompanied by users of Blade Mousterian and Levallois-Mousterian tools, respectively (Table 1 and Figure 1). The geographical situation of Crimea during the Last Interglacial is not clear. If it was an island, it will be difficult to explain the appearance there of the same industries that appear in neighboring territories of Eastern Europe. After the Moershoofd Interstadial, the Levallois-Mousterian sequence in the Prut-Dniester region ends, and comparable assemblages are found in Crimea (Table 1 and Figure 2). Is this perhaps evidence of migration? If so, this population movement might have been caused by harsh climatic conditions in the Prut-Dniester region. Evidence for environmental severity is seen in the post-Moershoofd Molodova V sequence and also throughout Poland during OIS 4 and 3 (Madeyska 1996). According to Kozlowski (2000:77), there were two "hiatuses" in the human occupation of southern Poland that correspond to the Early Pleniglacial and First Stadial of the Middle Pleniglacial. Unlike southern Poland where Micoquian

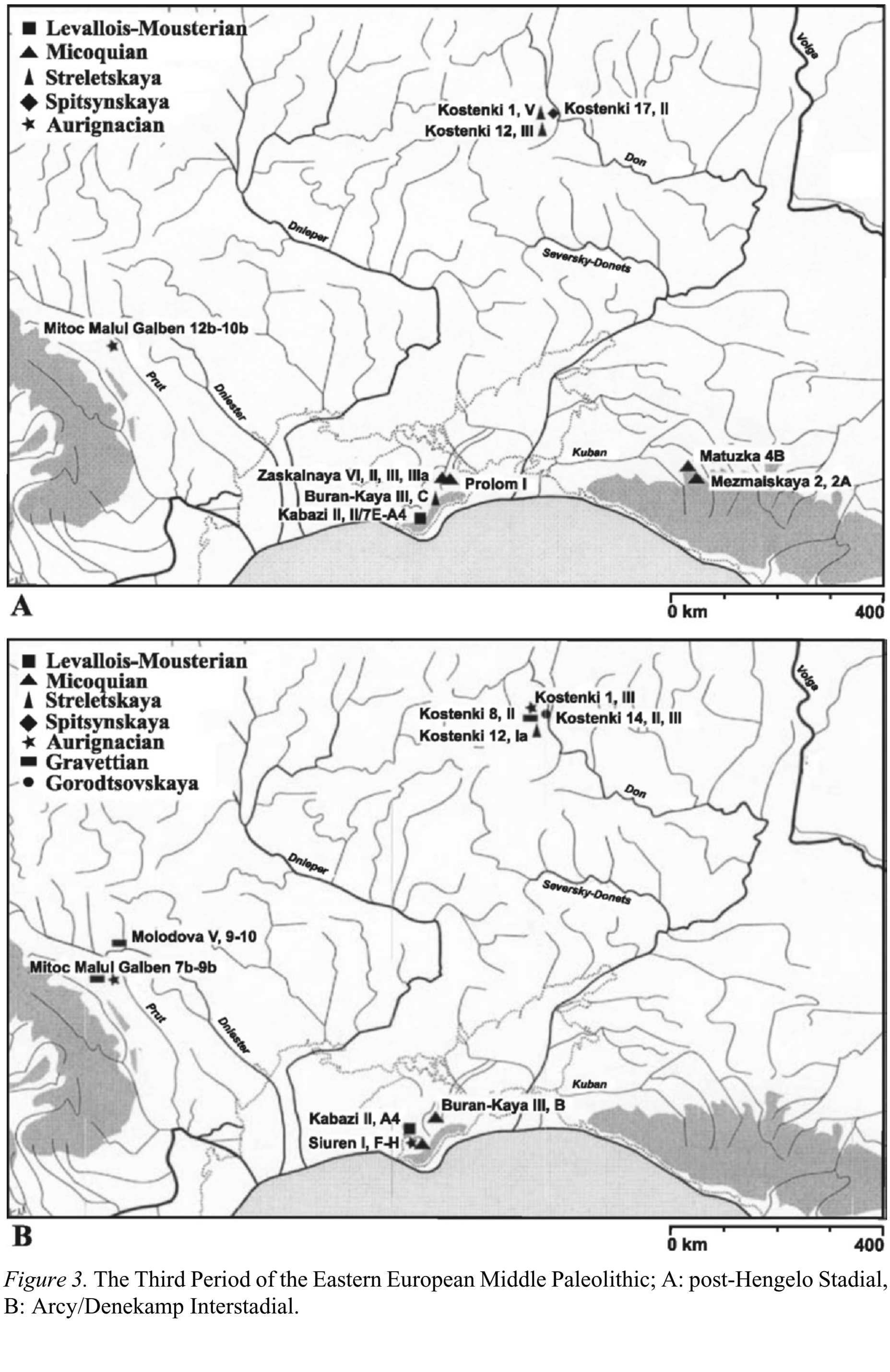


Figure 3. The Third Period of the Eastern European Middle Paleolithic; A: post-Hengelo Stadial, B: Arcy/Denekamp Interstadial.

assemblages disappeared before the maximum of the Early Pleniglacial (OIS 4), Micoquian replaced Levallois-Mousterian at the beginning of OIS 3 in the Prut-Dniester region. Also, after the Moershoofd began, the first well-documented appearance of Micoquian in the Northern Caucasus is found, and both Micoquian and Levallois-Mousterian occur in Crimea. Thus, the climatic deterioration during the Early Pleniglacial (OIS 4) may have caused population movements toward milder southern climates. It would appear that the onset of OIS 3 and the Middle Pleniglacial was a time of changes in Eastern Europe as well as other areas in Europe (Gamble and Roebroeks 1999).

The appearance of early Upper Paleolithic assemblages in the western (Prut-Dniester) and central (Middle Don) parts of Eastern Europe during the Stadial preceding Arcy/Denekamp was a major event (Table 2 and Figure 3A). At the time, peoples producing Middle Paleolithic industries still inhabited Crimea and the Northern Caucasus. Seemingly, the Upper Paleolithic emergence had little initial affect on the Micoquian residents, and this may relate to Early Upper Paleolithic choice of settlement locations. The Middle Don early Upper Paleolithic Spitsynskaya from Kostenki 17, layer II, was associated with humid forests of northern taiga type, and there is no reason to believe that Streletskaya occupations of about the same time and same area existed in different environments: they both appear to have been limited to the taiga forest. The Micoquian distribution, on the other hand, lay mainly across the forest-steppe environments of Crimea and the Northern Caucasus.

The connection between the Middle Don and Crimea was determined geographically by the Black Sea regression, which caused the Azov Sea to disappear and transformed the eastern Crimean rivers into tributaries of the Don.

After about 30 ky BP, the Gorodtsovskaya, another group of “taiga people” who were undoubtedly anthropologically modern, entered the Middle Don region (Table 2 and Figure 3B). At least two Gorodtsovskaya assemblages from Kostenki 14, layers II and III, are associated with forested landscapes, but these were neither as humid nor as cold as Kostenki 17, layer II, though still of taiga type. The Gorodtsovskaya reflects a relatively “primitive” stone technology but shows an incredibly developed and variable bone industry, adornments, burial customs, and dwelling structures for that period.

The first appearance of Aurignacian in Eastern Europe is documented in the Prut Valley at the beginning of the Arcy/Denekamp Interstadial and the end of the previous Stadial. It then appears in Crimea during the Arcy/Denekamp and finally in the Middle Don at the end or after the Arcy/Denekamp. Thus, the Aurignacian co-existed not with the local early Upper Paleolithic but with the Crimean Micoquian. The first documented Aurignacian “visit” to Crimea and the Middle Don was after Streletskaya, Spitsynskaya, and, at least, after the first manifestation of the Gorodtsovskaya.

The appearance of Aurignacian in the Middle Don was connected to environmental changes that took place after deposition of the Upper Humic Bed. Pollen samples from loess deposited above the Upper Humic Bed show open forest-steppe/steppe-desert landscapes in the Middle Don area suggesting that the Eastern European Aurignacian is associated with the open landscapes exploited by Middle Paleolithic populations. The Aurignacian exploitation of the "Don Passage" between the Kostenki-Borshchevo area and Crimea is documented by Black Sea shells found in Kostenki 1, layer III (Hahn 1977: 144).

In summary, archaeological data for the Upper Pleistocene indicate close links between populations in Crimea and the Eastern European Plain. Such evidence suggests that the Crimean peninsula might not have been so geographically isolated then as it is currently.

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12

ENVIRONMENT, SEA-LEVEL CHANGES, AND HUMAN MIGRATIONS IN THE NORTHERN PONTIC AREA DURING LATE PLEISTOCENE AND HOLOCENE TIMES

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Abstract: Initial spread of anatomically modern humans occurred during the Megainterstadial of the Last Ice Age, when the Neoeuxinian basin was in a deep regressive stage. The gradual transition to pottery-making and agriculture in the northern Pontic area occurred during the Holocene Climatic Optimum, 7000–5200 calBC, and was not related to a catastrophic flooding. Transgression/regression cycles in the Black Sea during the Holocene occurred with a periodicity of 1250 years.

Keywords: Black Sea paleoenvironment, sea-level changes, human migrations, early agriculture

1. INTRODUCTION

The northern Pontic area is extremely rich in archaeological resources, which have been systematically studied by several generations of Soviet, Russian, and Ukrainian archaeologists. These investigations, now being conducted with increasing participation by Western scholars, leave one in no doubt that, since early prehistoric times, subsistence and social dynamics of human groups in that area were directly affected by changes in climate, vegetation, and sea-level fluctuations. Over the last decades, archaeological studies in the northern Pontic area have included multidisciplinary projects targeted at reconstructing in detail the ancient environments of investigated sites (Dolukhanov 1979; Kremenetski 1991; Pashkevich 1997; Shilik 1997; Smyntyna 1999;

V. Yanko-Hombach et al. (eds.), The Black Sea Flood Question, 297-318.

Dolukhanov 2001; Dergachev and Dolukhanov, this volume). Recently, interest in the latter problem has been whipped up by the 'Flood' hypothesis (Ryan *et al.* 1997; Ryan and Pitman 1998). In its most recent revision (Ryan *et al.*2003), this proposal argues for the catastrophic flooding of the exposed Black Sea shelf at *ca.* 7400 calBC, which led to an accelerated spread of farming along the major rivers of southeastern Europe.

The aim of the present article is to analyze the existing evidence on Late Quaternary changes in the environment of the Northern Black Sea area, and to assess their possible impact on human migrations and subsistence, with a special emphasis on the Mesolithic-Neolithic interface.

2. PRESENT-DAY ENVIRONMENT

The area under discussion includes the Black Sea Lowland. Geologically, it belongs to the North Pontic Depression, a large foredeep of the crystalline basement of the East European Plate, which lies at a depth in excess of 2000 m. Its surface forms an undulating loess-covered plain, tilted to the south, where it contacts the Black Sea coastal landforms. The plain is crossed by several rivers, both large (the Prut, Dniester, Dnieper, Southern Bug, and Don) and small (the Kogulnik, Sarata, and many others). The lower stretches of the larger rivers often form impressive estuaries (or *limans*).

The climate of the Ukrainian steppe is continental, with cold to moderately cold winters, usually with intensive snowfalls, as well as hot and very dry summers. Annual precipitation is less than 400 mm, which gradually diminishes in the eastern direction. Summer droughts occur regularly, usually every three years.

The predominant vegetation is steppe, or treeless grassland consisting of drought-resistant herbs. It gradually transforms into a forest-steppe to the north, with patches of deciduous forest within river valleys. The Black Sea steppe is part of the huge belt of grassland extending eastward from the mouth of the Danube River, along the northern shore of the Black Sea, across the lower Volga River, through Kazakhstan and Southern Siberia into Mongolia. Steppe-like landscapes have been present there since the middle to late Miocene (15–11 mya) as a result of global cooling of the climate (Zubakov and Borzenkova 1990). In the geobotanical sense, there are three varieties of steppe: hydrophytic (including plants adapted to a humid environment), mesophytic (communities thriving in moderate humidity), and xerophytic (species that tolerate dry and ultra-dry conditions).

The steppe biota combine biodiversity with frequent and often unpredictable variations in biomass. Studies in the steppe and semi-desert areas of

Central Asia (Rodin 1977) identified annual, seasonal, and long-term fluctuations in the population densities of both plants and animals, which were primarily controlled by precipitation. Notwithstanding the high agricultural potential of the chernozem soil, the regularly occurring droughts lead to frequent crop failure and make this area one of high-risk farming.

3. THE INITIAL SETTLEMENT

During the Last Interglacial and the early stages of the Last Glacial, the northern Black Sea area supported groups of Neandertals, who manufactured different Mousterian industries (Micoquian, Levallois-Mousterian, Blade Mousterian; see Chabai, this volume). The radiometric dates for Mousterian sites in Crimea provide the oldest age of 70 ky BP, with the main cluster dating to around 40–30 ky BP. The youngest Mousterian sites date to the 'Arcy Interstadial' (Marks and Chabai 1998; Chabai, this volume). The later episode, not clearly recognizable in the Black Sea loess stratigraphy, dates to 32–28 ky BP in Western Europe (Djindjian *et al.* 1999). Neandertal skeletal remains have been found in several cave sites in the Crimean Mountains (Bonch-Osmolovsky 1940; Kolosov 1979), and the ^{14}C ages obtained for the bones of a Neandertal child in Mezmaiskaya Cave in the northern Caucasus (Golovanova *et al.* 1999) indicate that Neandertals survived there until 29 k years ago. This dating agrees with the evidence from Crimea.

The period of intensive Neandertal presence in the Black Sea area corresponded to OIS 5 (b–d), and 4, i.e., the initial (Early Valdaian and Sherstikhinian) stages of the Last (Würm, Weichselian, or Valdai) Ice Age, with interruptions by mild episodes: the Brørup/Upper Volgian and Odderade. Pollen data suggest that during the colder episodes, rare pine and birch forests intermingled with widespread grassland on the East European Plain, whereas the mild episodes were marked by the expansion of pine and spruce forests, mostly along river valleys (Velichko *et al.* 2002).

During OIS 5 (b–d), the Black Sea remained connected with the world ocean through the Mediterranean, its sea level being close to or higher than that of the present, as witnessed by the Karangat shorelines dated to 125, 111, and 80 ky BP (Chepalyga 2002). This period was followed by a regression recorded by brackish water deposits with *Dreissena* (Chepalyga 2002), possibly reflecting a glacio-eustatic regression coinciding with the Sherstikhinian cooling of 72–58 ky BP (Arslanov 1992).

Anatomically modern humans (AMH), who eventually replaced the Neandertals, appeared on the East European Plain and in the Caucasus between 41 and 32 ky BP (Golovanova *et al.* 1999; Dolukhanov *et al.* 2001). DNA analysis demonstrates (Ovchinnikov *et al.* 2000), that the Neandertals both in Europe and Caucasus were genetically unrelated to the AMH, and hence, the expansion of modern humans was a result of new migrations from Africa, in which course, groups supposedly reached the Black Sea area via the Levant and southeastern Europe.

These migrations occurred during OIS 3, which included the prolonged 'Middle-Valdaian Megainterstadial'. It comprised several mild episodes (Hengelo, Arcy-Denekamp, Bryansk) with unstable climate and predominantly pine and birch forests spreading along the valleys (Velichko *et al*. 2002). In the Black Sea basin, this period supposedly coincided with the Tarkhankutian transgression dated to 62.7 ky (U-Th) and 31.3 ky (^{14}C). At that time, water level in the Black Sea reached –30 to –35 m, and remained connected with the Mediterranean Sea (Chepalyga 2002). As recorded by Early Upper Paleolithic sites, radiating groups of AMH advanced along the major rivers (the Dniester, Dnieper, and Don) with intensive settlement in the Carpathian area, the Dnieper and Don basins, Crimea, and northern and western Caucasus (Figure 1).

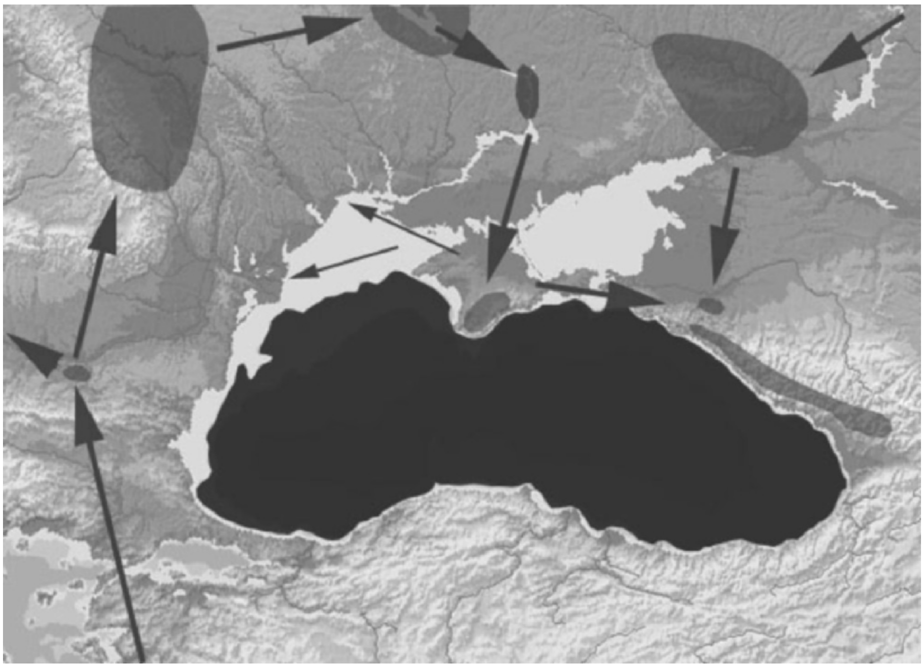

Figure 1. The Pontic basin at *ca.* 18 ky BP. Key: dark area = the Neoeuxinian basin; gray areas = concentrations of Upper Paleolithic sites; arrows = migratory routes.

A marked increase in the densities of AMH sites in the East European Plain coincided with OIS 2. This period included the LGM (20–18 ky BP), which brought maximum advance of the ice sheets in Northern Eurasia and an intensive accumulation of loess in the periglacial zone. In this climate of maximum cold and reduced precipitation, the entire East European Plain was covered by uniform periglacial-type landscapes combining tundra and steppe elements (Velichko *et al.* 2002).

This period also marked a maximum regression in the Neoeuxinian basin, when the brackish mega-lake that preceded the Black Sea stood at about –100 m below present sea level (Chepalyga 2002; Shmuratko, this volume). The greater part of the north Pontic shelf, including the Azov Sea, was exposed as a loess-covered, low-lying plain with lakes and meandering rivers. This period was followed by the culmination of the Khvalynian transgression of the Caspian Sea (17–14 ky BP), during which surplus water spilled into the Black Sea via the Manych Depression (Chepalyga 2002, this volume). The occurrence of Upper Paleolithic sites along the northern and eastern coastal areas of the Black Sea implies that the Manych Spillway did not impede the movements of Paleolithic groups.

The high frequencies of radiocarbon dated sites in the 'periglacial' zone of the East European Plain show a clear increase in population densities in the time ranges of 29–26 and 24–18 ky BP (Dolukhanov *et al.* 2001). Significantly, population densities in Central and Western Europe markedly decreased at that time, with some areas, such as southern Germany and Britain, being virtually depopulated (Houseley *et al.* 1997).

Groups of Paleolithic hunters moving into the Pontic Lowland from the north settled within the valleys of small rivers. A large site dated to 18 ky BP, Anetovka 2, belonged to specialized bison hunters and combined the functions of a tool-making workshop, butchering site, and a cult center (Stanko *et al.* 1989). Another site of approximately similar age, Amvrosievka, was a short-lived kill-site of bison hunters (Krotova 1999). It is very likely that similar settlements occurred on the exposed Pontic shelf.

During the final episodes of the Last Ice Age, the climate gradually became milder and wetter. Artyushenko (1970) identified pollen spectra comparable to those of the Alleröd in Western Europe: a warm interval between 11,800 and 11,000 years BP, revealing the dominance of oak, lime, elm, maple, and hazelnut. This period coincided with the glacio-eustatic rise of sea level in the Black Sea (the Neoeuxinian transgression). The rate of this transgression was controlled by the intensity of ice-sheet melting, and it increased during the mild interstadial episodes. An accelerated sea-level rise coincided with the Alleröd interstadial, when the Black Sea rose from –70 to –10 m in less than 1000 years (Shmuratko, this volume).

Vegetation changes in addition to excessive hunting, led to a nearly total extinction of bison. Large Paleolithic settlements disappeared and much smaller sites arose in the southern part of the Pontic Lowland (such as Bol'shaya

Akkarzha, near Odessa). The lithic inventory acquired a microlithic character with an acknowledgeable occurrence of elements stemming from the Caucasus (Stanko *et al.* 1989). There are no indications of any large-scale migrations or abrupt changes of subsistence that could reflect accelerated sea-level rises.

4. THE MESOLITHIC

With the beginning of the Holocene, *ca.* 10,500 calBC, summer temperatures in Europe rose by at least 6º C (Isarin and Bohnke 1999). During the Preboreal and Boreal periods of the Holocene (10,000–7000 calBC), dry steppe with *Carex*-Gramineae and *Artemisia*-Gramineae dominated the vegetation cover of the Pontic Lowland. Pollen data show a gradual spread of forest-like vegetation starting 8000 calBC. Forests were restricted to river bottomlands and foothills. Initially, they comprised pine and birch, then later included broad-leaved species such as oak, elm, and an understory of hazelnut. Steppe vegetation on the watersheds acquired a mesophytic character.

According to Balabanov (this volume) and Yanko-Hombach (this volume), Holocene changes in the level of the Black Sea had a fluctuating character. Balabanov's scheme shows a transgressive peak of –25 m reached at 9350–9200 calBC, and three transgressive phases taking place between 8550 and 6640 calBC, one of which (7330–7050 calBC) was linked to the breakthrough of Mediterranean water. Later sea-level transgressions occurred at 4600, 4200, and 3800 calBC, culminating at 3000 calBC. The amplitude of the greatest transgressions was on the order of 20 m, and the rate of sea-level rise did not exceed 6–7 cm per annum (Shmuratko, this volume).

An alternative scenario suggested by Ryan *et al.* (1997), Ryan and Pitman (1998), and Ryan *et al.* (2003) argues that, following the post-glacial rise in global sea level, the Mediterranean Sea catastrophically breached the Bosphorus Strait at *ca.* 7400 calBC, and rapidly refilled the Black Sea basin, flooding the shelf. This event allegedly led to drastic environmental changes and stimulated the expansion of farming in Europe.

4.1 Crimea

Mesolithic sites in Crimea are distinguishable by their topographic position (predominantly in mountainous valleys), subsistence (hunting of forest and forest-steppe animals, with an increased role of gathering and fishing), and lithic inventories. Stone tools retained their Upper Paleolithic character, the main distinctions being the refinement of core-reduction techniques, resulting in an increased production of regular prismatic blades and bladelets. Retouched tools were dominated by small end scrapers with varying frequencies of geometric microliths (Stanko 1982:99).

4.1.1 Settlement Pattern

Mesolithic levels are known in at least 20 caves and rockshelters found in the canyons of small rivers and streams crossing the northern range of the Crimean Mountains. At least several sites bear evidence of prolonged habitations (Bibikov *et al.* 1994).

4.1.2 Subsistence

The early Mesolithic saw the disappearance of mammoth, reindeer, cave bear, and cave hyena (Vekilova 1971). Mesolithic prey included grassland species (wild horse, saiga antelope, aurochs, wild sheep and goats) increasingly supplemented by those adapted to forests (red deer, roe deer, wild boar, and brown bear) (Bibikov *et al*. 1994). Mesolithic deposits contain numerous bones belonging to large fish, both freshwater (salmon, trout, and roach), and brackish-water species (pike-perch and catfish). At several sites (Shan-Koba, layer 3; Fatma-Koba; and Kurzak-Koba) were found large concentrations of the shells of edible common snails (*Helix aspera*). These snails make up the shell midden site of Laspi 7, located on the seashore in Sebastopol Bay. Earlier reports suggested the occurrence of domesticated pig at the Mesolithic site of Tash-Air (Krainov 1960), however, Tsalkin (1970:261–266) has convincingly demonstrated that the bones identified as pigs either penetrated from younger deposits, or belonged to wild boar.

4.1.3 Material Culture

Several archaeological cultures are distinguished in the Crimean Mesolithic (Telegin 1982, 1989). The Shan-Koba Culture, typified by several cave sites and the coastal site of Laspi 7, features industries manufactured on small prismatic blades, with a high proportion of backed blades, end scrapers, burins, and geometrics (lunates and trapezes). The Murza-Koba industries are found at Crimean cave sites above the Shan-Koba levels. They include notched and backed blades as well as end scrapers made on truncated blades. Geometrics are rare but include a diagnostic *fossil directeur*, the Murza-Koba trapeze. Bone and antler tools are not numerous but include characteristic small harpoons. The small rockshelter of Syuren' 2 in the middle chain of the Crimean Mountains occupies a special position. The inventory of the lower level includes tanged points with direct analogies in the Polish Swiderian.

4.1.4 Burials

Level 3 of Murzak-Koba rockshelter includes a double burial of a male and female. Partly preserved skeletons were found beneath stone slabs, side-by-side, in an extended posture, facing east. The age of the female is estimated as 20–25, that of the male as 40–50 (Bibikov *et al.*1994:105–108). A burial of an adult male, with an estimated age of about 40 years, was also found in the Fatma-Koba rockshelter. This skeleton lay in a contracted posture on the right side in a shallow grave.

4.1.5 Chronology

Using typological criteria and stratigraphic evidence, Telegin (1982:49) divided the Crimean Mesolithic into two stages: the older being the Shankobian, and the younger the Murzakobian. A series of radiocarbon dates for the Laspi 7 site gives an age range of 8750–6100 calBC.

4.2 The Pontic Lowland Steppe

4.2.1 Settlement Pattern

Numerous Mesolithic sites are found on the Danube-Dniester interfluve in the western part of the Pontic Lowland (Figure 2). They form a hierarchy of settlements linked to topographic position. The largest sites (Mirnoe 1 and Beloles'e) were located on the lower terraces of the limans, which at the time were marshy river bottomlands, rich in waterfowl. Allegedly, they were base-camps inhabited on a year-round basis. According to Stanko (1982), the Mirnoe community included twenty to twenty-five individuals, representing three or four nuclear families, each with four to six individuals. Smaller sites in a similar setting supposedly belong to task groups budding off from the main center. The location of several sites at higher elevations in the foothills implies seasonal transhumance.

Mesolithic settlements are also located in the terraced valleys of the major rivers (the Dnieper, Dniester, and Seversky Donets). A dense cluster of sites is found in the middle reaches of the Dnieper: Igren', Oskorivka, Vasilievka, Yamburg, and others.

4.2.2 Subsistence

The existing evidence, including the chemical analysis of bones and dental pathology, clearly indicates that a hunting and gathering subsistence

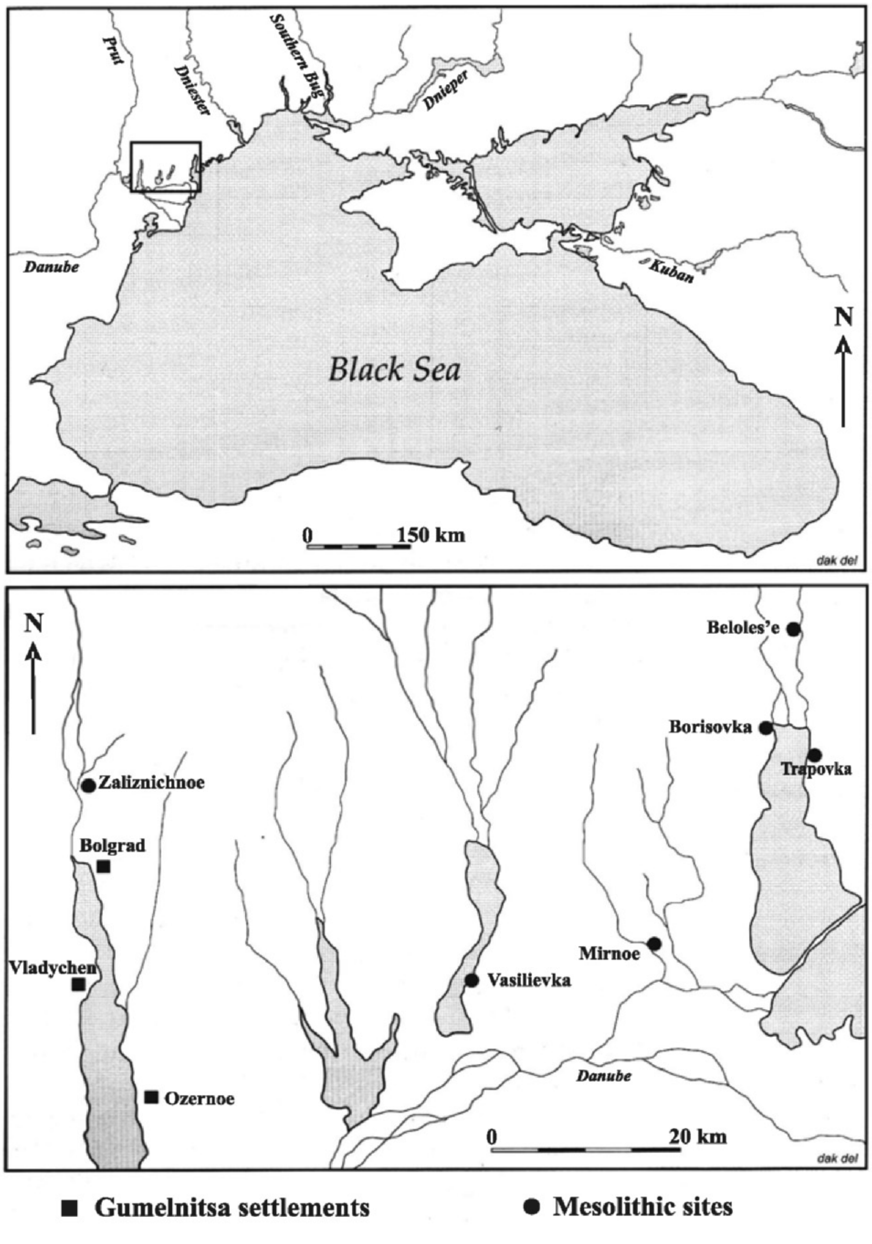

Figure 2. Prehistoric sites in the Danube-Dniester interfluve (Dolukhanov 2001).

economy persisted in southern Ukraine throughout the Mesolithic period (Lillie 1996). It is equally evident that the Mesolithic economies developed in an environment of increasing scarcity among hunting resources. Bison, the principal Paleolithic prey species, had been largely exterminated by that time. Mesolithic groups targeted the less numerous herds of wild horse (or tarpan) and saiga antelope, or more solitary animals, such as aurochs. Fishing and gathering increasingly supplemented the hunting of land mammals.

Chemical analysis of human bones from Mesolithic cemeteries on the Dnieper shows shifts in the concentration of barium and strontium and increased ^{13}C values, indicating a greater consumption of plant foods (Jacobs 1994). Korobkova (1993) has identified at Mirnoe a group of prismatic blades with microscopic traces of linear use wear, which she and Stanko (1982) view as evidence of "reaping knives" used for harvesting edible plants. Pashkevich (1982:136) has found the grains of several potentially edible plants in the deposits of Mirnoe: white goosefoot (*Chenopodium album*), black bindweed (*Polygonum convoluvulus*), vetch (*Vicia hirsuta*), and sorel (*Rumex acetosa*).

In all parts of the Pontic Lowland, Mesolithic food-acquisition strategies increasingly relied on the use of aquatic resources and harvesting of plant food. Sites were concentrated in landscapes with diverse and predictable wild resources, especially marine estuaries, lakes, and river valley floors. The Mesolithic lifestyle combined stability with seasonal transhumance. In several cases, large permanent base-camps were surrounded by smaller seasonal stations. Communal cemeteries strongly suggest increasing sedentism and territorial control.

4.2.3 Material Culture

The Kukrekian is typified by the "Kukrekian armature," a truncated, notched, and ventrally retouched blade, which was first recognized at the open-air site of Kukrek in the Crimean steppe. Implements of this type have been identified at other sites of a later date in Crimea, on the Dnieper River (Igren' 8), and in various areas of the Pontic Lowland. The Grebenikian, as exemplified at Mirnoe, Girzhevo, and other sites, is yet another culture group, predominantly in the Odessa District. The implement inventory includes end scrapers and microliths consisting exclusively of trapezes.

4.2.4 Burials

Large communal cemeteries, unknown in earlier periods, reflect substantial social changes that marked the transition to the Mesolithic. A group of large cemeteries, Vasilievka, Volos'ke, and others, lies near the Dnieper Rapids, south of Dnieperpetrovsk (Stolyar 1959; Telegin 1982). Individuals within the same cemeteries are associated with distinct burial rites, suggesting cultural homogeneity: the majority was buried in a contracted position, and some in an extended supine position.

Still more importantly, the skeletons belong to at least three distinct physical types (Gokhman 1966, 1986; Potiekhina 1999). The first group, consisting of individuals with broad and high-relief faces, is viewed by Gokhman (1966:187) as belonging to the autochthonous "Cro-Magnon" population of the Central and Eastern European Upper Paleolithic. The second group was found only at Volos'ke cemetery and includes individuals with very long, narrow faces typical of the "Mediterranean race." The third type, found at Vasilievka 1 and among the dead buried in extended supine positions at Vasilievka 3, features narrow faces and protruding jaws. At Volos'ke in addition to Vasilievka 1 and 3, there are numerous cases of ribs and vertebrae penetrated by flint arrowheads indicating a violent death.

4.2.5 Chronology

AMS measurements (Anthony 1994; Jacobs 1994; Lillie 1996, 1998) suggest the age of Vasilievka 3 cemetery as 10,400–9200 calBC. Using typological criteria, Stanko (1982) and Telegin (1982, 1989) distinguish an early stage in the Ukrainian steppe Mesolithic (as exemplified by Beloles'e) that reveals an essentially late Paleolithic technology, regular prismatic blades, and a limited number of geometrics. The later stage includes Grebenikian (Mirnoe, Girzhevo, and others) and also Kukrekian sites. Several radiocarbon dates are available for the eponymous site and Igren' 8 on the Dnieper. Existing radiocarbon dates for Igren' 8 span a wide range between 8800 and 6100 calBC. The oldest evidence of pottery is from the lower levels of the Rakushechnyi Yar site located on a small island in the lower stretches of the Don River, with radiocarbon dates indicating an age of 5846±128 calBC (Dolukhanov and Shukurov 2004).

4.3 Mesolithic: Summary

In all parts of the Pontic Lowland, Mesolithic strategies of food-collecting increasingly relied on the use of aquatic resources and harvesting of plant foods. Sites were concentrated within landscapes with diverse and predictable wildlife resources, especially, marine estuaries, lakes, and river floors. The Mesolithic lifestyle combined stability with seasonal transhumance. In several cases, large permanent base-camps were surrounded by smaller seasonal stations. The presence of communal cemeteries strongly indicates greater sedentism and increased territorial control.

Existing evidence firmly suggests that the Mesolithic communities in the northern Pontic area developed essentially on a local basis, without any significant influence from outside. At any rate, there is no discernible signal that could reflect any change in the subsistence or population dynamics at the time of the alleged 'Flood', at *ca.* 7400 calBC. Yet ,both the anthropological and archaeological data indicate minor infiltrations from the Mediterranean area that proceeded throughout the Mesolithic.

5. THE NEOLITHIZATION

The spread of the Neolithic and the adoption of farming in Western Asia and Europe largely coincided with the Climatic Optimum of the Holocene. In the greater Mediterranean area, 8 to 7 thousand years ago, this climate change took the form of increased precipitation while temperatures dropped by 1.5 to 2.0° C (Rohling and de Rijk 1999). The Black Sea Lowland witnessed the maximum spread of mixed coniferous-deciduous forests in the lower mountains and within valley bottomlands. Forest-steppe vegetation became dominant on the watersheds, bringing together mesophytic grassland and isolated stands of oak, elm, lime, and maple (Kremenetski 1991).

V. Gordon Childe (1925) first modeled the Neolithization as a major migration of farming populations from Western Asia. The spread of farming into Central Europe is associated with sites of the Linear Pottery Culture (LBK). Statistical analysis of LBK radiocarbon dates places its most probable age at 5227 calBC (Dolukhanov and Shukurov 2004). Farming in that area was spreading at a very rapid rate, *ca.* 4 km/year.

LBK sites appeared on the Dniester and smaller rivers of Moldavia and southwestern Ukraine during its middle stage, presumably originating from the Carpathian basin. The Bug-Dniestrian Culture on the Dniester and Southern Bug witnessed the direct interaction of local hunter-gatherers with these LBK farming communities.

A different model of Neolithization is proposed for the greater part of the East European Plain, where the Neolithic is associated with sedentary, pottery-making, hunting, and food-gathering communities. Analysis of radiocarbon dates for Neolithic cultures here shows that the tradition of pottery-making was spreading from the east (Dolukhanov and Shukurov 2004). The early pottery sites of the Yelshanian Culture on the Lower Volga yield an age of 6910±58 calBC. Rakushechnyi Yar on the Lower Don dates to 5846±128 calBC, and the Bug-Dniestrian dates to 6121±101 calBC.

The tradition of pottery-making in each case was accepted by the local Mesolithic groups and involved only limited human displacements. Thus, the dead buried at the Neolithic Dnieper-Donetsian cemeteries have similar cranial measurements as the individuals found at the Mesolithic site of Vasilievka 3 (Potiekhina 1999). On the other hand, pottery-making hunter-gatherers had contacts with farming communities, as is demonstrated by the presence of LBK pottery at Bug-Dniestrian sites.

Fully developed farming economies are attested in the forest-steppe of the Pontic Lowlands in sites of the early Tripolye (Cucuteni) Culture, *ca.* 4500–4350 calBC (Wechler 1994). Small settlements located in close proximity to river flood plains show the full range of cultivated plants: hulled wheats, naked six-row barley, and hulled barley, as well as animal remains of both wild (red deer, boar, roe deer, and elk) and domesticated species (cattle, pig, sheep/goat). The archaeological evidence and radiocarbon dates suggest that the Tripolye (Cucuteni) Culture arose from the interaction of eastward migrants from the Boian Culture area in Romania with autochtonous Bug-Dniestrian groups.

The spread of farming in the steppes of the Danube-Dniester interfluve is documented by sites of the Gumelnitsa Culture. These sites, located in Romania and Bulgaria and dated between 4900 and 3800 calBC, featured mounded, tell-type settlements, developed stock breeding and agriculture, and copper metallurgy. The Gumelnitsa sites in the Odessa Oblast' of Ukraine and in southern Moldavia have yielded radiocarbon ages that calibrate to 4900–4000 calBC. Large Gumelnitsa settlements (Bolgrad, Ozernoe, Nagornoe, and Novoselskoe) are located on the elevated terraces of the Yaplug liman, which was, at that time, ingressed by marine water in the course of the New Black Sea transgression.

Sharing basic common characteristics with their western counterparts, the sites in the Danube-Dniester interfluve reveal several peculiarities, such as the lack of *tells* and artificial fortifications, and differences in the architecture of dwellings, types of ceramic wares, and working tools. Cattle were shorter than those in Romania, emmer was a common staple food, and there was no evidence of hunting (Subbotin 1983).

All stages of the Gumelnitsa Culture were roughly contemporaneous with the Cucuteni-Tripolye Culture, and numerous examples of mutual interaction are apparent (Comşa 1987). Tsvek (1996) particularly stresses intensive contacts with Gumelnitsa that are recognizable in the sites of the Alexandrovka group of the eastern Tripolye area.

The Gumelnitsa Culture was the result of expansions of several related groups from a core area in the northern Balkans. These groups were bound together by multiple trade links, yet they maintained their cultural identity over a prolonged period of time. The expansion into the dry steppe of the Danube-Dniester interfluve (a high-risk agricultural area) became possible only with the increased precipitation of the Climatic Optimum.

Significantly, both scenarios of Neolithization (the spread of farming and the spread of pottery-making) occurred in the northern Pontic area significantly *later* than the time of the alleged 'Flood' (*ca.* 7400 calBC). These were complex processes involving both limited-scale migrations and diffusion of cultural and technological knowledge among local hunting and gathering communities. They were in no way controlled by environmental catastrophes. On the other hand, one may acknowledge in several cases that the periods of major extension of farming communities in the northwestern and northern Pontic area coincided with minor Black Sea transgressions. This may be due to the general increase of precipitation in the river catchment, which enhanced the agro-climatic potential of places within the steppe and forest-steppe (see Dergachev and Dolukhanov, this volume).

6. THE POST-OPTIMUM

Starting *ca.* 4800–3500 calBC, annual precipitation diminished by 50 mm, while mean monthly temperatures fell by 2º C for July and by 1º C for January compared to today. The combined effect of extensive farming and increased aridity led to a decline in the Mediterranean plant species, oak, lime, and elm (Kremenetski 1991).

After the peak of the New Black Sea transgression, sea level dropped over the course of the Phanagorian regression. Saltwater molluscs from sand deposits in the Lower City of Olbia at a depth of 0.1–0.3 m below present sea level yielded a radiocarbon age of 4200 calBC, suggesting that sea level at this time was at least 1.6 m below that of today (Shilik 1997).

Saltwater molluscs were collected on the bottom of the Bug liman: from a layer of quartz sand at a depth of 1.50–1.60 m that underlay archaeological deposits of the 5th–3rd centuries BC. The samples yielded an age of *ca.* 4000 calBC, implying that the coeval sea level was at least 3 m lower than at present (Shilik 1997:120). A later stage of this regression is attested in the Yargylach Cay, on the Tarkhankut Peninsula in western Crimea. Archaeological deposits bearing ceramics attributable to the Catacomb and Timber Grave Cultures were found on the bottom of the bay, suggesting a mid-second millennium BC lowering of sea level by at least 1 m compared to today (Shilik 1997:120).

Significant transformations can be seen within the Tripolye societies during their latest phase (3780–3320 calBC). One of the Late Tripolye groups, Usatovo, spread throughout the Prut-Dniester-Southern Bug interfluve and became increasingly dependent on stock breeding, especially of horses and sheep/goats. Defensive structures became more common, and profound changes in social structure appear conspicuously in burial rites, particularly at the eponymous site, Usatovo, northwest of Odessa (Patokova 1979). Three large kurgans stood out from the rest of the burials, apparently belonging to the local elite. They contained individual male burials with rich inventories of grave goods that included Aegean-type daggers fashioned from arsenical copper, and copper flat axes and chisels (Zbenovich 1974). Graves containing female burials also were accompanied by rich grave goods, including pottery and personal adornments. The nearby cemeteries consisted of clusters of flat graves, each cluster apparently belonging to kin relatives as evidenced by the occurrence of figurines of a similar type. Based upon this evidence, the Usatovo society may be viewed as a military-oriented chiefdom, with a strong local power base as shown by elite burials, and containing several lineages bearing particular symbols of kin identity.

Further socio-economic development resulted in the total collapse of the Late Tripolye agricultural societies and the gradual rise of the Pit-Grave, Catacomb, and Timber Grave Cultures. The Pit-Grave or Yámnaya Culture reached its peak during the third millennium calBC, when it encompassed the entire East European steppe from the Urals in the east to the lower Danube in the west.

In the traditional view, the Pit-Grave Culture consisted of pastoral stock breeding groups that kept cattle, sheep, goats, and horses. Archaeobotanical evidence has recently lent support to an earlier view that this region also included a sedentary agricultural component (such as Mikhailovka) in areas sufficiently rich in agricultural resources (Pashkevich 1997). The Mikhailovka site grew to the size of one and a half hectares in its final stage of development and was surrounded by fortifications with stone ramparts and ditches.

Merpert (1968) views the Pit-Grave Culture as a ‘culture-historical entity’ based on a common ideology (primarily the kurgan mortuary rite), which resulted from the integration of various local cultural traditions. Yet, the kurgan rite was not a complete novelty in the steppe area, as it was in use at Usatovo and earlier still during the middle stages of Tripolye. Pit-Grave burials may be viewed as symbols of the enhanced power of local chieftains. The prestige artifacts in the rich graves included imported items of gold and copper, mace heads (Nikol’ski and Mariupol groups), and wheels or even complete wheeled carts (e.g., at Storozhevaya Mogila, near Dnieperpetrovsk on the Dnieper River).

The replacement of the Late Tripolye by the Pit-Grave Culture was generally peaceful. There are few indications of hostilities and much evidence for both trade and social interaction, particularly in the diffusion of pottery. There is also evidence for direct reoccupation, as at Maidanetskoe, where Pit-Grave mounds were erected immediately atop the fortified Late Tripolye settlement (Videiko 1994). Direct military occupation of the entire area seems unlikely, as the Late Tripolye population far outnumbered the Pit-Grave groups according to Videiko (1994).

It has been noted (Renfrew 1979:142–143) that the “abrupt termination” of the “prosperous Chalcolithic cultures” occurred quasi-simultaneously in various parts of southeastern Europe. These changes can be viewed as adaptations to social crisis, triggered by adverse climatic conditions and the ensuing scarcity of food resources.

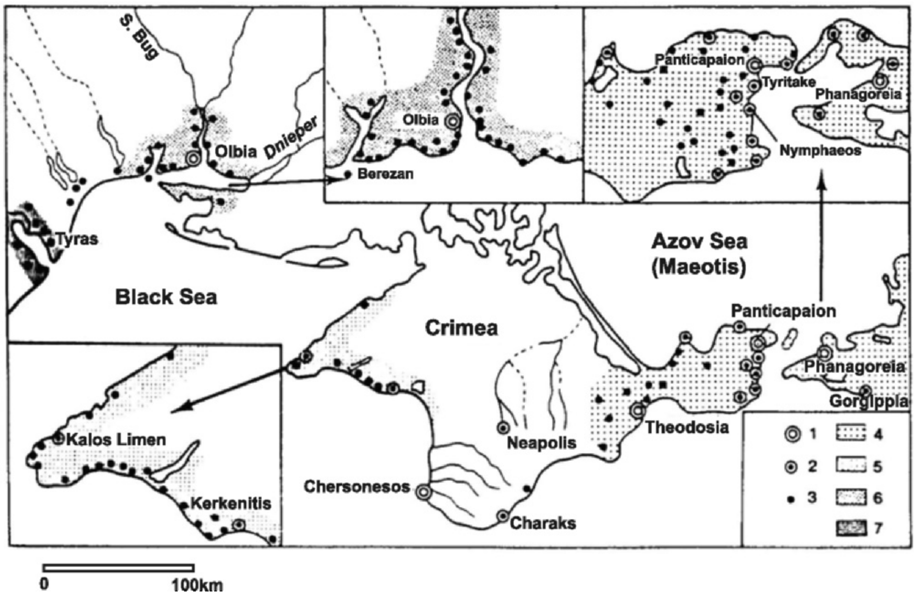

*Figure 3.*Greek colonization of the northwestern Black Sea area (after Kryzhitskii 1997).

7. THE GREEK COLONIZATION

Between the seventh and fifth centuries BC, a string of Greek colonies emerged along the northern Black Sea coast: Tyras in the Dniester Liman; Olbia in the Bug-Dnieper Liman; Chersonesos, Theodosia, and Panticapaeus in Crimea; Gorgippia on the Caucasian coast; and a great many minor ones (Figure 3). The initial Greek colonization proceeded against the backdrop of the Phanagorian regression, which attained its minimum in the fifth century BC (Fedorov 1978). Judging by the position of Greek ceramics found in subaerial deposits on the bottom of the Bug Liman, sea level at the time can be assessed at –11 m at the lowest (Shilik 1997:121).

One of the earliest Greek settlements, dated to the seventh century BC, was established on the island of Berezan, about 2 km off the coast of the Bug Liman. At the time of the Greek colonization, this island was part of the mainland, and it remained so until the 1st century AD (Nazarov 1997).

The subsequent sea-level rise, the Nymphaean transgression, started in the late 1st century AD. The chronology and magnitude of this transgression have been assessed by the position of fortifications in the port of Chersonessos (the present day Quarantine Bay of Sebastopol, in Crimea). Detailed survey of the Nymphaean terrace along the Bug Liman near Olbia shows that maximum sea level was similar to today and was attained in the late 7th–early 8th centuries AD (Shilik 1997:122–123).

The seaward displacement of Chersonesos's ramparts in the 13th–14th centuries suggests a new drop in sea level by about 3 m below the present stand (Figure 4). This "Korsunian regression" reached its lowest level of –3.0 to –3.6 m in the late 14th–early 15th centuries AD (Shilik 1997:125–126).

8. CONCLUSIONS

The expansion of Neandertals into the Black Sea area and Caucasus from the Levant and southeastern Europe occurred during the time of the Karangat Sea (OIS 5–4), when the Pontic basin was connected with the Mediterranean and the world ocean.

Early modern humans penetrated the Black Sea area as a result of a new migration from the Levant and southeastern Europe during one of the mild episodes of the Valdai Megainterstadial (OIS 3), when the Black Sea was connected with the Mediterranean at a level close to the present one.

Mesolithic groups emerged between *ca.* 10,500 and 6000 calBC, essentially supported on a local resource base in an environment of increased temperature, rainfall, and gradual rise in sea level. There are no indications of major population movements or changes in subsistence that could suggest adjustment to the alleged 'Flood' at *ca.* 7400 calBC.

The earliest manifestations of pottery-making in the southern part of the East European Plain, including the northern Pontic area, appeared around 7000–6000 calBC, while early farming cultures (affiliated with the LBK) emerged *ca.* 5200 calBC. Fully developed farming economies belonging to the Tripolye (Cucuteni) and Gumelnitsa Cultures became established in the Pontic Lowland during the fifth millennium BC (4900–4000 calBC). The process of Neolithization involved both limited-scale migrations and diffusion of cultural and technological knowledge among local hunter-gatherer communities, which occurred considerably later than the alleged 'Flood' (7400 calBC). In several cases, major extensions of farming communities into the northwestern and northern Pontic area coincide with minor Black Sea transgressions. This may be due to the general increase in precipitation, which triggered the influx of fresh water into the Black Sea and enhanced agro-climatic resources in the coastal areas.

Early farming cultures collapsed during the third millennium calBC under conditions of growing aridity and regression in the Black Sea. The Pontic Lowland was subsequently populated by nomadic pastoral groups.

The Greek colonies began spreading in the coastal Black Sea area during the seventh to fifth century BC, against the backdrop of a low sea level.

The Holocene transgression/regression cycles in the Black Sea occurred with the periodicity of 1250 years (Figure 4).

Existing archaeological data strongly support a scenario of gradual environmental changes in the northern Pontic area during the Late Pleistocene and Holocene. Reliable evidence suggests that the changes in subsistence and cultural dynamics resulted from a combination of socio-economic and environmental factors, fluctuations in precipitation being the most important among the latter.

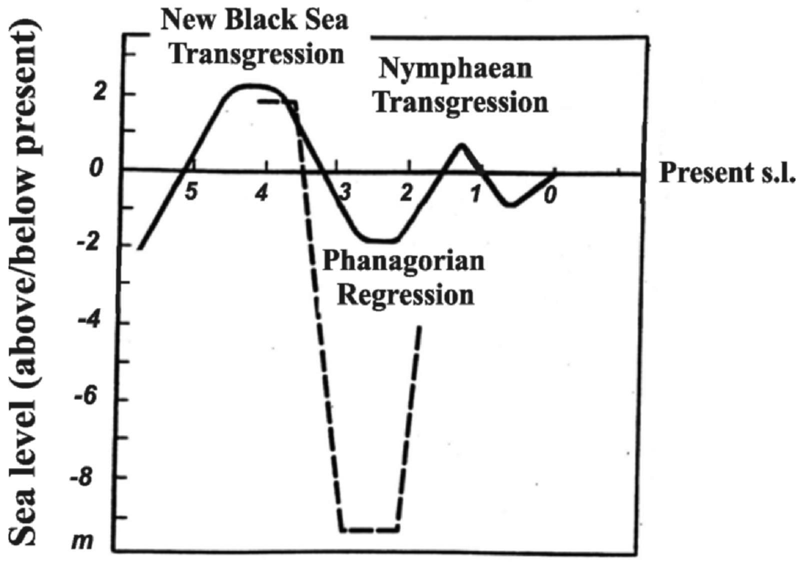

*Figure 4.*Fluctuations in the level of the Black Sea level during the middle to late Holocene; solid line (after Fedorov 1978), dotted line (after Shilik 1997).

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HOLOCENE MEDITERRANIZATION OF THE SOUTHERN CRIMEAN VEGETATION: PALEOECOLOGICAL RECORDS, REGIONAL CLIMATE CHANGE, AND POSSIBLE NON-CLIMATIC INFLUENCES

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Abstract: Despite their relatively high latitude and isolation from the Mediterranean biome, the southern coast of the Crimean Peninsula and the Novorossiysk-Tuapse region of the western Caucasus harbor the largest concentrations of Mediterranean flora in the northern Black Sea region. This paper examines previously published pollen and soil records covering the interval from 12,000 BP to the present, plant ecology research, and field observations in order to investigate the plants of Mediterranean origin in the modern communities of southwestern Crimea. The inquiry considers local Holocene vegetation change and soil development as a function of regional climate change. In addition, the paper discusses possible direct and indirect influences by human agency, sea currents, and sea level changes on the formation and modification of Mediterranean floristic elements in Crimea.

Keywords: Pollen, soils, climate, archaeology, sea currents, Mediterranean vegetation, Crimea, Black Sea, Holocene

1. INTRODUCTION

Despite its relatively high latitude (44º 25'–44º 40' N.) and complete isolation from the Mediterranean biome, the southern coast of the Crimean Peninsula and the Novorossiysk-Tuapse region of the western Caucasus contain the two largest concentrations of Mediterranean flora on the northern shores of the Black Sea (Figure 1). In these two areas, local climatic factors such as rela-

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*Figure 1.*The Black Sea region and the areas of Mediterranean vegetation .

tively warm temperatures and a short, but conspicuously dry, summer season favor the growth of subtropical and Mediterranean flora.

This paper reviews the post-glacial vegetation history of southwestern Crimea based on pollen and soil records from the Heraklean Peninsula and the adjacent Chyornaya Valley (Figure 2). It also discusses vegetation change in relation to climate and human intervention, as well as the possible influences of sea level migration and marine currents. Evidence consists of pollen and soil records published by Cordova and Lehman (2003, 2005), which provide proxy information allowing inferences about paleoc limatic change. Unfortunately, pollen data are lacking for the Novorossiysk-Tuapse region, and therefore the inquiry must be focused on Crimea alone.

Section 2 describes the biogeographical context of the study area and the main characteristics of climate, vegetation, and soils in southern Crimea. Section 3 reviews ideas from various scientific sources, mostly Russian, that concern the origin of Mediterranean floras in the northern Circum-Pontic region.

2. THE REGIONAL AND LOCAL CONTEXT OF CRIMEA'S MEDITERRANEAN VEGETATION

Between 46% and 52% of native plant species of Crimea's southern mountains are also found in the Eastern Mediterranean region (Didukh 1992).

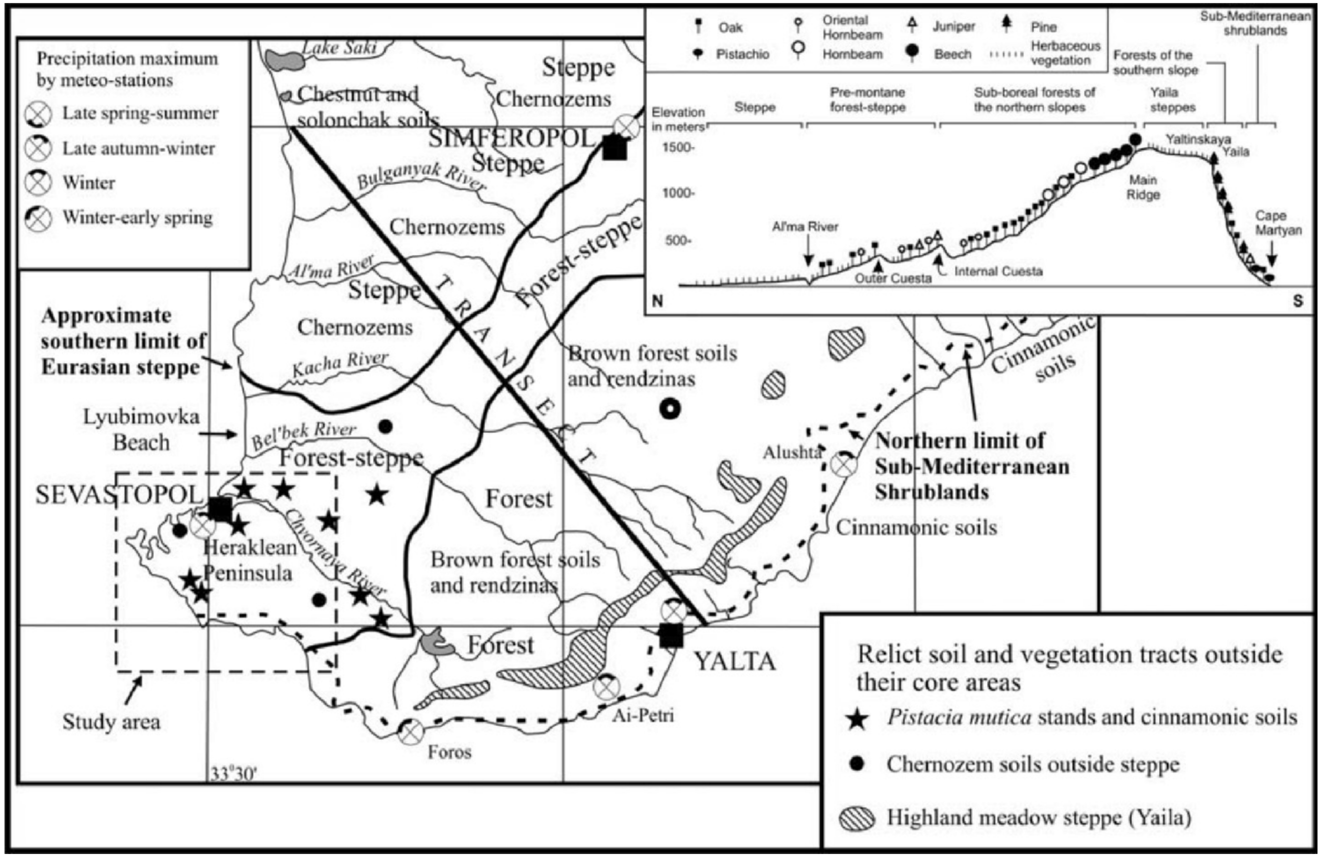

Figure 2. Southwestern Crimea and vegetation transect.

These species are concentrated mostly along a narrow strip of Crimea's southern coast as well as in other isolated spots (Figure 2, map and inset transect). The climate of the southern Crimean coast is considered of Mediterranean type because of its mild winter temperatures and summer drought (Walter 1943, 1974; Yena *et al.* 1996), but its temperatures are lower and summer drought periods shorter than most locations in the Eastern Mediterranean (Douguedroit and Zimina 1987). The climate of this part of Crimea is thus often referred to as Sub-Mediterranean (Grishankov 1976; Yena 1978; Vyed' 1983).

Sparse distribution of meteorological stations in Crimea makes it difficult to define the limits of Sub-Mediterranean climate. Some Mediterranean floristic elements, such as *Pistacia mutica*, suggest approximate limits, which correspond to the areas with relatively warm temperatures and summer drought (Figure 1). Therefore, the distribution of *P. mutica* is here provisionally taken as demarcating the modern extent of Crimea's Sub-Mediterranean climate. This area includes the southern and southwestern coast, and its climatic characteristics include precipitation maxima in winter, late fall-winter, or winter-early spring, as well as mean January temperatures that never drop below 0° C.

The Crimean Mountains create a protective barrier against the freezing effects of Arctic air blasts that sweep southward through the region in winter. Warm, westerly and southwesterly winds during the winter and a relatively warm sea keep temperatures high along the southern coast (Vodop'yanova 1986). The combination of these warm winds, warm sea waters, and cold air also produce storms that bring winter precipitation to southwestern Crimea (Vyed' 1983; Vodop'yanova 1986). Similar weather conditions often occur in the Novorossiysk-Tuapse region, where vegetation is remarkably similar to the southern Crimean coast (Didukh *et al.* 1990).

Despite its varied Mediterranean plant species, Crimea lacks some typical Mediterranean elements. For example, olive (*Olea europaea*) and all the evergreen oaks of the Mediterranean region are absent in Crimea's native flora. In recent centuries, these species have been introduced and adapted to the warmest spots of the southern Crimean coast through acclimatization research carried out under the aegis of the Nikitskiy Botanical Garden in Yalta (Wulff 1926; Molchanov and Rubtsov 1986).

There are three native species of oak in Crimea, *Quercus pubescens*, *Q. petraea*, and *Q. robur*, of which only *Q. pubescens* has strong Mediterranean ties. In Crimea, *Q. pubescens* occupies areas below 500 m mostly, though in the Mediterranean basin, it is usually found at higher elevations, e.g., in Turkey, where it grows mainly between 600 m and 1700 m (Davis 1982). At such elevations, temperatures are too low for other, typically Mediterranean oaks, and for this reason, Horvat *et al.* (1974) classify *Q. pubescens* as a Sub-Mediterranean element. Nevertheless, exceptions exist, as is indicated by the *Q. pubescens* occupying lowland areas in the Argive Plains of the Peloponnese

(Jahns 1993).

In southern Crimea, typical Mediterranean elements occupy elevations from 0 to 300 m, though they can be found higher depending upon local conditions. The majority of such elements are not found farther north, as is the case with *Pistacia mutica*, *Cistus tauricus*, *Ruscus ponticus*, *Arbutus unedo*, and *Pyracantha coccinea*, among others (Vodop'yanova 1986). Most of the native Mediterranean floristic elements in southern Crimea are components of *shiblyak* and *phrygana*, two plant communities often equated with the Mediterranean *maquis* and *garrigue*, respectively (Shelyag-Sosonko *et al.* 1982; Didukh 1992).

Shiblyak consists of associations of evergreen and deciduous shrubs and short trees. The term comes from the Slavic languages of former Yugoslavia, where a similar community exists in the mountains facing the Adriatic Sea. Although shiblyak is often equated with the Mediterranean maquis, Rikli (1943–48) suggests that the scarcity of evergreen shrubs in shiblyak communities differentiates it from maquis. Nonetheless, both communities share various ecological characteristics. Both are dominated by shrubs, and both are the result of the degradation of oak forests.

Phrygana is a community of aromatic, spiny herbs and scrub, which are unpalatable to livestock. The term derives from the Greek word for low scrub. Phrygana communities are often understood as the result of the degradation of shiblyak, intense grazing, and soil erosion, however, it is possible that the presence and distribution of phrygana scrub are also influenced by soil and lithology, as seems to be the case in areas of flysch and mudstone deposits in Crimea's south coast (Korzhenevskii and Klyukin 1990). There are several types of phrygana in Crimea, the most common being the *tomillares* type, which is dominated by plants of the Labiatae (mint) family (Didukh 1992). Less abundant than the Labiatae is the Cistaceae family, which includes the genera *Cistus* and *Helianthemum*. Other plants commonly found in phrygana are *Asphodeline taurica* and *A. lutea*. Underrepresented species include various members of the Rosaceae and Fabaceae families. In addition, a specific group of phrygana communities grows on the badly eroded schists and mudstones of Jurassic age. These communities include *Capparis herbaceae* and *Athraphaxis replicata*, among others (Korzhenevskii and Klyukin 1990). As in the case of the association between shiblyak and maquis, the Crimean phryganas are different from those of the Mediterranean region, although they are remarkably similar in various ecological aspects.

Although Mediterranean species form between 40 and 50% of the total flora of southern Crimea, a large number of them are considered rare and endangered (Cordova *et al.* 2001b). Communities of native Mediterranean vegetation are essentially restricted to those tracts of land within protected territories. Some of these species, however, are now outnumbered by species recently introduced from the Mediterranean region and naturalized in an environment relatively similar to their places of origin.

In addition to flora, soils also provide testimony of present and past Mediterranean climates. *Cinnamonic* soils, or the close equivalent of *terra rossa* and calfersic soils in the Russian and Soviet Classification, are widespread in both southern Crimea and the Novorossiysk-Tuapse region (Gerasimov 1954; Krupsky and Polupan 1979). Steppe areas contain variations of the southern *chernozem*, or black prairie soils, and the mountains reveal typical *rendzina* and mountain forest soils (Krupsky and Polupan 1979). These soil genetic types are useful stratigraphically because, in the sedimentary sequences, they seem to indicate climatic fluctuations between humid (forest), steppe (chernozem), and Sub-Mediterranean (cinnamonic).

3. THE ORIGIN OF MEDITERRANEAN FLORAS IN THE NORTHERN BLACK SEA REGION

The pollen records discussed in this paper reveal an expansion of Mediterranean and Sub-Mediterranean floras in southwestern Crimea during the Holocene. It should be acknowledged, however, that paleoecological studies of Miocene and Pliocene deposits have identified taxa of Mediterranean origin and paleoclimates similar to those of the Mediterranean region. Fossil records from different parts of Eurasia show that the Mediterranean biome originated in the warmer climates of the Miocene, but its further differentiation occurred during the variable cold and warm episodes of the Pliocene and Pleistocene (Pignatti 1978; Suc 1984). During the Miocene and Early Pliocene, the Mediterranean biotic realm extended to the north and east beyond its present distribution, but as the Sarmatic and Meiotic Seas shrunk in size and disappeared, a large number of species became isolated in numerous areas of the Middle East and Central Asia (Shchekina 1979; Didukh 1992).

Parishkura (1978) analyzed pollen from the Late Miocene and Pliocene beds exposed at Lyubimovka Beach, north of Sevastopol (Figure 1). Veklich (1982) discussed these pollen records in relation to soil chronosequences developed in the same deposits and came to the conclusion that subtropical and Mediterranean conditions prevailed during these periods. Shchekina (1979) correlated pollen data from the Lyubimovka beds with other sections in mainland Ukraine, where similar plant taxa existed, suggesting that warm climates extended as far north as Kiev. Pollen recovered by Parishkura (1978) at Lyubimovka included various taxa of subtropical origin, such as *Cornus*, *Rhamnus*, and *Rhus*. Many taxa found in these assemblages are not particularly associated with the modern Mediterranean realm, however. Such taxa include *Castanea*, *Carya*, *Pterocarya*, *Zelkova*, and *Juglans*, among others, whose relicts are today secluded in the Colchidian (Georgia), Pontic (northeastern Turkey) and Hyrcanian (northern Iran) forests, where climates are not Mediterranean (Tumajanov 1971; Zohary 1973; van Zeist and Bottema 1991; Didukh 1992). On the other hand, typical modern Mediterranean native plants growing in Crimea today, such as *Pistacia*, *Jasminum*, and *Ligustrum*, were not reported from any of these beds at Lyubimovka. Unfortunately, Parishkura (1978) reports most plant taxa at the family level only, which does not necessarily indicate the occurrence of specific taxa of Mediterranean origin. For example, the abundance of Ericaceae does not necessarily denote *Arbutus*, which is the only member of this family among the modern Mediterranean flora in Crimea and the Novorossiysk-Tuapse regions.

Another piece of evidence regarding Tertiary relict vegetation comes from southern Crimea and the Novorossiysk-Tuapse regions, where *Pinus pityusae* var. *stankeviczii*, a Tertiary Mediterranean relict now extinct in the Mediterranean region, still thrives (Didukh 1992). Phylogenetically, *P. pityusae* can be traced to *P. brutia*, which is found in the Eastern Mediterranean (Price *et al.* 1998), however, there are no continuous records showing the phylogenetic differentiation of *Pinus* in the northern Black Sea region.

The hypothesis regarding the Tertiary origin of Mediterranean floras in Crimea and Novorossiysk-Tuapse has been discussed by Vul'f (1943), Maleev (1946, 1948), Grosset (1979), Podgorodetsky (1988), Didukh *et al.*(1990), and Didukh (1992). Their discussions led to the hypothesis that most Tertiary Mediterranean floras in Crimea and the Novorossiysk-Tuapse region survived the cold phases of the Pleistocene in warm *refugia* (Didukh 1992). The discontinuous paleobiogeographical record from the Pliocene to the present leaves this hypothesis untested.

The data presented here do show how plant taxa of Mediterranean affiliation expanded in southwestern Crimea after the last glacial period. It is not clear whether this resulted from the expansion of thermophilous plants out of their glacial refugia or immigration of species from lower latitudes. Unfortunately, pollen analysis has failed to answer this question, yet DNA studies like those proposed by Comes and Kadereit (1998) could help track some Mediterranean and Sub-Mediterranean elements in the northern Black Sea to their interglacial or even Tertiary origins.

4. PALEOECOLOGICAL RESEARCH IN SOUTHWESTERN CRIMEA

The pollen and soil records discussed here are presented and described in detail by Cordova and Lehman (2003, 2005) and summarized in Figure 3.

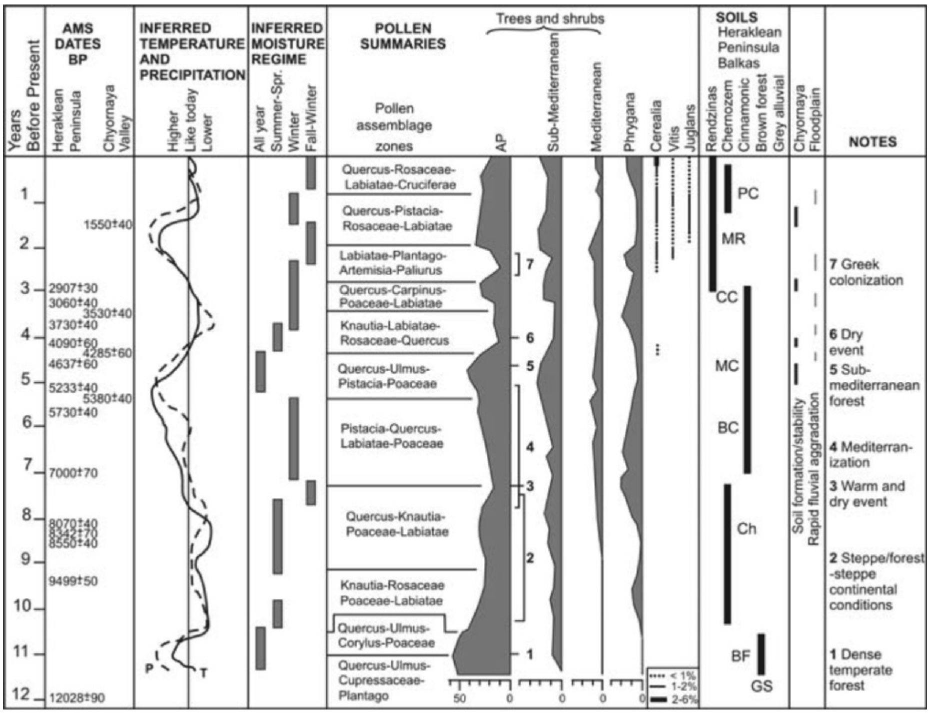

Figure 3. Vegetation change, soils, and inferred climates for the Holocene in SW Crimea.

They were obtained from seven stratigraphic soil-profile sections and four pollen sequences supported by a total of seventeen AMS radiocarbon dates (Table 1). In addition, archaeological occupation remains contributed to the chronological control of some sections.

In order to simplify the diagram of Figure 3, the substantial number of plant taxa identified in the pollen profiles was condensed into groups. Each group consists of a set of selected pollen types linked to plant taxa of a particular ecological community. These communities include AP (arboreal pollen), phrygana, Mediterranean, and Sub-Mediterranean taxa (Table 2). All members of the phrygana group are present in the Mediterranean group. This is due to the fact that membership in the Mediterranean group depends upon climatic associations, whereas the phrygana plant community comprises taxa of Mediterranean affiliation. A shiblyak group was not included in the curve because some taxa from this community are also present in dense forests, and because the pollen of some important shiblyak species is not morphologically distinctive. Nonetheless, the majority of pollen types of shiblyak plant taxa are distributed in the Mediterranean and Sub-Mediterranean group.

Pollen assemblage zones in the summarized diagram were qualitatively determined based on the four most common taxa present in a set of samples in

Table 1. AMS conventional and calibrated radiocarbon dates.

Lab number	Conventional ^{14}C Age (years BP)	Median Calibrated Age (years calBP)	Section*	Depth Below Surface (cm)	Dated material**
Beta-127553	1550±40	1446	NG2	68	char
T-16416A	2907±30	3041	AA	25–30	ash
Beta-127550	3060±40	3277	AA	75–90	bosed
Beta-137097	3530±40	3790	NG2	139–152	bosed
Beta-156480	3730±40	4076	BBBP-3	60–80	bosed
Beta-137095	4090±60	4607	EF	130–140	bosed
T-16426A	4258±60	4805	NG2	230–240	bosed
T-16420A	4637±60	5390	MM2	110–120	bosed
T-16417A	5233±40	5981	AA	170–180	bosed
Beta-127554	5380±40	6180	NG2	261–274	bosed
Beta-127552	5730±40	6525	MM2	245–270	bosed
Beta-137094	7000±70	7819	BB	187	char
Beta-127551	8070±40	9009	BBBP-2	90–100	bosed
T-16421A	8342±70	9353	MM2	340–350	bosed
Beta-156479	8550±40	9529	BBBP-2	90–135	bosed
T-16418A	9490±50	10,810	AA	310–320	bosed
T-16422R	12,028±90	14,048	BBBP-2	590–600	bosed

* See Cordova and Lehman (2003, 2005) for locations and descriptions.
** char = charcoal, bosed = bulk organic sediment

Table 2. Ecological groups of pollen taxa mentioned in text and Figure 3.

Group	Sum of the Following Pollen Types and Taxonomic Groups
Phrygana	Cistaceae, Labiatae, *Asphodeline*, *Sarcopoterium*, *Thymelea*
Mediterranean	*Pistacia*, Jasminum*, Thymelea hirsuta, Sarcopoterium, Asphodeline*, and Cistaceae
Sub-Mediterranean	*Quercus pubescens* type, Cotinus,* Cornus,* Paliurus,* and Verbascum*

* Taxa typical of shiblyak communities

the pollen sequences (*e.g. Quercus*-*Carpinus*-Poaceae-Labiatae). For details on the criteria involved in this determination, see Cordova and Lehman (2005). For details on pollen groups and problems concerning the determination of pollen of cultivated and wild varieties (e.g., *Vitis*, *Pistacia*, *Juglans*, Fabaceae, and Rosaceae) and their alternative interpretation in pollen diagrams, see Cordova and Lehman (2003).

AMS radiocarbon dates on the diagram of Figure 3 are uncalibrated, but a list of calibrated dates is presented in Table 1. Calibrations were carried out following the University of Washington Quaternary Isotope Lab Radiocarbon Calibration Program REV 4.3, which is based on the "intcal98.14c" curve by Stuiver *et al.*(1998). When reported, calibrated dates are presented as calBC, calAD, or calBP; uncalibrated dates are reported as BP.

Precipitation and temperature in the diagram of Figure 3 are inferred from taxa for which limits of moisture and temperature are known in the modern plant communities of Crimea (see Cordova and Lehman 2005). Thus, the curves are not meant to have precise numerical validity, but only relative parity with the pattern shown by the same plants in their contemporary climatic settings within the study region.

The diagram of Figure 3 also summarizes information on soil development (see section 4.1), periods of stability (soil formation), and rapid fluvial aggradation in the Chyornaya River floodplain. Notes one to seven on the far right column refer to the numbers indicated in the pollen summary.

4.1 Sediments and Soils

The oldest deposit described consists of a gray alluvial soil (GS) formed in alluvial silts dated 12,028±90 BP, which is buried by mud drapes and lenses of coarse sand and gravel, and a sequence of alluvial silts. A brown forest soil (BF) developed on top of the alluvial silt deposits. Subsequently, a meadow chernozem (Ch) developed on top of the BF soil. The smooth boundary between the BF and Ch soils contains lithics dated to the Late Mesolithic Murzak Koba period, according to the chronology by Bibikov *et al.*(1994). AMS radiocarbon dates for the Ch soil range between 9499±50 and 8070±40 BP.

Above the chernozem soil, a sequence of three cinnamonic soils developed on rapidly accumulating silt deposits. The first was a poorly developed brown cinnamonic (BC) that grades into a dark gray-brown meadow cinnamonic (MC), and this was in turn covered by silts on which a calcic cinnamonic (CC) developed. The BC soil formed under the warm conditions of a Mediterranean-like climate, similar to the modern cinnamonic soils of the southern Crimean coast. Although no dates were obtained from it, it is possible to place it between 7000±70 and 5233±90 BP. The MC soil is richer in organics than the BC, suggesting increased moisture although still under Mediterranean climatic conditions. Three dates (5730±40, 5233±40, and 4090±60 BP) put this moister climatic event somewhere between 5700 and 4100 BP. The CC soil is a cinnamonic with carbonate filaments, suggesting dry conditions. The top of the CC soil seems to be truncated by erosion, except for one section where part of an A horizon is still visible above the carbonate filament horizon. The CC soil is bracketed between 4090±60 and 3060±40 BP.

The meadow rendzina soil (MR) developed on gravelly-loam colluvium deposited on the erosional unconformity that truncates the top of the CC soil. AMS dates from the bottom of the MR soil at sections AA and BBBP-3 yielded 2907±30 and 3730±40 BP, respectively. Pottery of Bronze Age date is mixed in with the colluvium, however, suggesting that this soil may have accumulated between 3000 and 3500 BP.

A core from the Chyornaya River floodplain contains a continuous 2.8 m sequence of overbank and backswamp sediment accumulation. The bottom of the sequence is dated 5380±40 BP, and sedimentary and chemical data indicate five major depositional units separated by three buried soils (Cordova and Lehman 2003). The wetland nature of soils in the Chyornaya floodplain allows no direct correlation with soils in the Heraklean Peninsula.

4.2 Pollen Zones and Environmental Change

The *Quercus*-*Ulmus*-Cupressaceae-*Plantago* pollen assemblage zone was obtained from silt layers burying the gray alluvial soil (GS). Radiocarbon dates place this assemblage around 12,000 BP. The palynological content of this zone was 59% arboreal pollen (AP), of which 35% corresponded to broadleaf trees. The presence of boreal tree species (e.g., *Betula* and *Alnus*) indicates cool and moist conditions. Despite the increase in AP, the bottom of this zone contained high frequencies of Cupressaceae and *Plantago*, suggesting that despite forest dominance, unstable conditions prevailed. Thus, this assemblage marks a transition from cool and dry to cool and wet conditions.

The *Quercus*-*Ulmus*-*Corylus*-Poaceae pollen zone was associated with the brown forest (BF) soil. Although no absolute dates exist for this deposit, it is presumed to have formed between 12,000 and 11,000 BP. AP was 35%, of which 30% comprised broadleaf trees. The relatively high frequencies of *Quercus*, *Corylus*, and *Sambucus* suggest a dense forest cover. *Corylus* and *Sambucus* grow on forest edges, which implies the presence of open areas. Possibly, human activities account for the existence of such openings. In fact, lithics of the Crimean Mesolithic Murzak Koba industry appeared atop the deposit containing this pollen assemblage (Cordova and Lehman 2005). Alluvial brown forest soil (BF) formation shows that stable conditions were finally reached after continuous silt accumulation. Overall, the pollen assemblage of this zone suggests a temperate climate with an even annual rainfall distribution.

The *Knautia*-Rosaceae-Poaceae-Labiatae zone was characterized by a decrease in broadleaf trees and an increase in *Artemisia*, *Knautia*, and other steppe taxa, which suggests the predominance of a cool and dry climate. A typical southern chernozem soil (Ch) developed at this time, which correlates with the steppe-like vegetation suggested by the pollen assemblage. Thus, pollen and soil records indicate that, at this time, a continental, cool and dry climate with precipitation maximum in late spring and summer predominated.

The *Quercus-Knautia*-Poaceae-Labiatae zone demonstrated a dominance of steppe vegetation, although broadleaf-tree pollen remained relatively high (15.5 to 27.6%). Sub-Mediterranean arboreal species such as *Cornus*, *Paliurus*, and *Quercus pubescens* increased. The appearance of *Pistacia*, *Jasminum,* and Cistaceae pollen toward the top of the assemblage marks the transition to a warmer climate with a dry summer. *Mentha*-type pollen increased, suggesting not only a Mediterranean-like climate, but also intensification of pastoral activities. Overall, this pollen assemblage suggests a transition from steppe/forest-steppe vegetation to Sub-Mediterranean woodland similar to the modern southern coast of Crimea. It is probably at this time when shiblyak and phrygana communities began to form as a result of the combined effects of climatic change and pastoral activities by the Neolithic inhabitants of the Heraklean Peninsula.

The *Pistacia-Quercus*-Labiatae-Poaceae zone shows a rapid increase in Mediterranean and Sub-Mediterranean plant taxa. In particular, the frequencies of *Pistacia* (6.5–7.3%) and *Jasminum* (1.2–2.9%) show peaks unmatched in other periods. These percentages are even higher than the modern pollen spectra of areas where these two taxa grow (Cordova and Lehman 2003). Pollen of phryganoid plants such as *Asphodeline*, Cistaceae, and Labiatae continued to rise. Among the Sub-Mediterranean trees and shrubs, *Quercus pubescens*, *Cotinus*, *Paliurus*, and *Cornus* maintained a strong presence among the arboreal taxa. Although *Q. pubescens* thrives in areas of both summer and winter rain, it is well adapted to conditions of summer drought. Therefore, it remained the principal oak species as climate shifted to more Mediterranean-like conditions. The formation of a brown cinnamonic soil (BC) during this period suggests the establishment of a xeric moisture regime (prolonged summer drought).

The *Quercus-Ulmus-Pistacia*-Poaceae zone denotes an increase of broadleaf trees, especially *Quercus* and *Ulmus*, and marks a climatic amelioration. *Fagus* pollen, which is rare in all the pollen assemblages, appears for the first time in this zone, but its frequencies are below 0.5%. When compared with modern pollen rain (Cordova and Lehman 2003), this low percentage indicates long-distance transport by wind from the mountains. Meadow cinnamonic soils indicate increased moisture. It is possible that there were two precipitation maxima during the year, one in the fall-winter and the other in late spring or early summer.

The *Knautia*-Labiatae-Rosaceae-*Quercus* zone marks a decline in arboreal pollen and suggests a dry phase. The peak of *Knautia* may represent the return of steppe conditions, but *Artemisia* is inexplicably absent in the assemblage. The increase of Labiatae (mainly *Mentha*-type pollen) hints at an increase in pastoral activities, but some Labiatae are relatively drought-tolerant since they are commonly found in the steppes. This may suggest that pastoral activities increased under drying conditions. Though archaeological sites of the

Neolithic period in the Heraklean Peninsula are poorly known, there is substantial evidence for rapid 'Neolithization' in nearby regions (Cohen 1996).

The *Quercus*-*Carpinus*-Poaceae-Labiatae zone shows a resurgence of broadleaf trees. Although *Quercus* recovers slightly, *Carpinus* and *Ulmus* increase considerably during this phase. It is possible to infer a sub-humid climate with occasional droughts in the summer, probably similar to the modern forest areas immediately west of the Chyornaya Valley. In general, pollen assemblage and soil development at this time indicate climatic amelioration concomitant with increasing human disturbance. AMS dates show that this moisture improvement occurred shortly before the establishment of the Greek farming colony in the Heraklean Peninsula, and thus it is most likely that climatic amelioration played a role in attracting settlers to the region.

The Labiatae-*Plantago*-*Artemisia*-*Paliurus* zone shows that as broadleaf trees declined, shrubs such as *Paliurus* and those of the Rosaceae family increased. The abundance of Labiatae and other phryganoid herbs, as well as weeds and cultivated species such as *Vitis* and Cerealia, suggest a managed environment. Thus, vegetation change in this zone is marked mainly by disturbance created through the establishment of the Greek *chora* of Tauric-Chersonesos (about 420 to 100 calBC). Due to the noise imposed by the anthropic indicators in the pollen assemblage, it is difficult to infer climatic conditions, however, it could be assumed that the warm and moist conditions of the previous zone continue into this one.

The *Quercus*-*Pistacia*-Rosaceae-Labiatae zone marks an increase of arboreal vegetation and, at the same time, the development of a weed complex. The reduction of *Vitis* and Cerealia suggests a decline in Greek farming, however, *Juglans* (walnut) appears for the first time, although it does so later in the assemblage. The Labiatae and a series of weeds indicate sporadic farming and pastoral activities. Thus, pollen assemblage and soil development suggest an amelioration of climate together with agricultural decline. A meadow rendzina soil developed on the colluvial fills in the *balkas* (local Russian for a small ravine). The formation of the Post-Classical period chernozem (PC) in the uplands reflects an anthropogenic change rather than a climatic consequence.

The *Quercus*-Rosaceae-Labiatae-Cruciferae zone embraces combined assemblages from meadow rendzina soils (MR) in the Heraklean Peninsula and from the top alluvial unit in the Chyornaya floodplain. The palynological content of this zone suggests that *Quercus* and shrubs of the Rosaceae family became dominant in the wooded landscapes. Other broadleaf trees such as *Pistacia*, *Carpinus*, and *Ulmus* remained low. Among other trees, *Pinus* increased through reforestation. The increase in Labiatae pollen suggests phrygana expansion due to pastoralism. Members of the mustard family (Cruciferae) increased, henceforth becoming one of the most widespread weeds in the modern landscape (Cordova and Lehman 2003). Climatic conditions are difficult to determine due to the noise created by reforestation and introduction of foreign species, as well as agricultural and urban development.

4.3 Vegetation Change in Adjacent Regions

Very few dated pollen records currently exist in other regions of Crimea, however, these available data allow rough correlations with sequences obtained from the Heraklean Peninsula. Gerasimenko (published in Cohen *et al.* 1996) reconstructed the vegetation history around Skalistiy rock shelter on the northern slopes of the mountains. The DR-3 (Younger Dryas) phase is marked in Skalistiy by a return of steppe conditions, and such an event is clear in the data from this study. Subsequently, climatic conditions in Skalistiy improve, whereas in the Heraklean Peninsula, they remain relatively dry until around 8000 BP.

Gerasimenko also studied the late Holocene lacustrine deposits of Lake Saki in western Crimea (brief summary in Cordova *et al.* 2001a), where a dry phase around 4000–3500 BP is also recorded. In addition, the increase in cultivated plants during and after the Hellenistic period (*Vitis*, *Juglans*, and Cerealia) is recorded in the sediments of Lake Saki.

The Holocene paleoecological sequences from the lower Dnieper region (Kremenetski 1995, 1997) and from southeastern Ukraine (Gerasimenko 1995, 1997) show some parallels with those studied in southwestern Crimea. For instance, the AP increase recorded in these steppe locations during the Atlantic period occurs slightly later than in the Heraklean Peninsula. This amelioration corresponds to the *Quercus*-*Ulmus*-*Pistacia*-Poaceae pollen assemblage zone (~5700–4500 BP). Perhaps vegetation and soil in southwestern Crimea responded earlier to this change due to its southern location and proximity to the sea. The dry event between 4200 and 3300 BP (SB-2) recorded in the pollen and soils of the Heraklean Peninsula is represented in the pollen diagrams of Kardashinski Swamp and southeastern Ukraine. In addition, soils dated to this stage in the steppe present low organic content and carbonate development (Ivanov 1992; Gerasimenko 1995, 1997). Between 3000 and 2500 BP, the AP curve reflects an expansion of forested areas, most likely induced by precipitation increase. This forest expansion is curbed around 2500 BP by the Greek colonization (Figure 3), however, after the decline of farming in the chora of Chersonesos, AP pollen recovers. This event may be related not only to farming decline, but also to climatic amelioration (Cordova and Lehman 2005). The pollen records obtained by Gerasimenko (1995) in the southern Ukrainian steppes show a moisture increase during the early Sub-Atlantic phase (SA-1), dated roughly between 2500 and 1500 BP.

Correlation with palynological records from around the Black Sea is difficult because most pollen sequences in this region come from lake basins located at high elevations, where post-glacial forest recovery was delayed by low temperatures and dry continental conditions. Unfortunately, most of the pollen sequences from lowland areas do not cover the entire Holocene because these lakes were formed behind coastal sand bars and dunes and did not come into being until the sea reached a level similar to the present. Thus, pollen diagrams from Adatepe Akgöl, Kücük Akgöl, and Tatlı Gölü in northern Turkey (Bottema *et al.* 1993–94) or the lakes in eastern Bulgaria (Bozilova and Beug 1994; Lazarova 1995) cover only the Middle and Late Holocene, when vegetation change was largely influenced by local human activities.

Dated pollen sequences from sea bottom sediment cores exist for the western and southwestern areas of the Black Sea basin and the Sea of Azov, but some taxa identified in the southern Crimean profiles are not recorded in the sea sediment cores. For example, the Labiatae and Cistaceae, important families in the phrygana and the Mediterranean flora of Crimea, and overall in the Mediterranean region, are poorly represented or even absent in sea sediment cores of both the Mediterranean and Black Seas. This is partly the result of the poor pollen production and dispersal of some species. They can be recorded in terrestrial deposits near the actual location of plant communities, but not far away in the sea. In contrast, tree pollen is recorded both nearby and far away due to its high dispersing capabilities. Therefore, correlations between the terrestrial pollen sequences here discussed and diagrams from sea bottom cores should be based mainly on AP and individual tree genera.

Pollen sequences from continental shelf sediments off the Bulgarian coast prepared by Atanassova (1995) show a predominance of steppe vegetation before ~11,000 BP, a transition from xeric to humid conditions between ~11,000 and ~7,000 BP, and a subsequent increase in arboreal pollen. A similar trend is observed in a diagram from the Sea of Azov basin published by Vronsky (1988), but this diagram lacks radiocarbon dates for the Middle and Late Holocene.

Mudie *et al.* (2002) obtained pollen from sea bottom sediments north of the Bosphorus and from sediments of the Marmara Sea that show Mediterranean taxa appearing earlier. In core B-7, PZ-2b, in the southern Black Sea, *Pistacia* appears and peaks between 9000 and 6000 BP, which correlates with pollen records of the Marmara Sea (Mudie *et al.* 2002) and Eastern Mediterranean deep sea cores published by Rossignol-Strick (1995). *Pistacia* would be expected to appear much later in Crimea, which lies at a higher latitude and experiences relatively lower temperatures. Amazingly, *Pistacia* appears and peaks between 9529 calBP (8550 BP) and 5981 calBP (5233 BP) in the pollen diagrams of southwestern Crimea. Is this the result of a rapid warming that allowed Mediterranean trees to come out of their glacial refugia? Or is this a sign of rapid migration from southern latitudes? There is no answer to the first question due to the lack of paleobotanical data, but there are some clues to an answer to the second question when looking at marine changes during the Holocene.

5. POSSIBLE INFLUENCES BY PREHISTORIC AND HISTORIC CULTURES

Another factor to consider in the Holocene migration of flora is human agency, which may have occurred through movements of population or through local disturbance. Humans could have carried plants or seeds intentionally or accidentally. Plant introductions and adaptations are better known for the Greek, Roman, Byzantine, and later periods (Cordova and Lehman 2003), however, for previous centuries, there are no well-dated paleobotanical records that support early introduction of southern flora. Studies in the Sinop Peninsula in Turkey reveal pre-Greek trade routes in the Circum-Pontic area (Doonan, this volume). Such trade routes may have reached Crimea.

Another possibility is that human disturbance may have created conditions appropriate for the expansion of Mediterranean plants that were still secluded in refugia. This could be the case in the notable expansion of phrygana between 7000 and 5000 BP (Figure 3). The rise of phrygana scrub in Crimea and the Mediterranean region is often associated with grazing and general agricultural pressure. Although phryganoid taxa appear in the diagram before the Neolithic, it is not until *ca.* 8000 BP that they show a considerable increase. Any conclusive linking of early pastoral influences with the growth of phrygana will require further ecological, archaeobotanical, and palynological research.

The effects of Greek farming on deforestation (*ca.* 2500 BP) and phrygana expansion are more convincing, since they are better documented archaeologically and historically. *Vitis* and Cerealia-type pollen increase commensurate with an increase in pollen from phryganoid shrubs and a reduction in arboreal pollen. Additionally, the expansion of *Juglans* during the Roman and Byzantine periods (*ca.* 2000–1000 BP) is also recorded in the pollen diagrams (Cordova and Lehman 2003).

6. POSSIBLE INFLUENCES OF MARINE CHANGES ON THE HOLOCENE CLIMATES AND VEGETATION OF CRIMEA

6.1 Sea Level Changes and Their Possible Effects on Climatic Conditions

The continental shelf around Crimea is wide and shallow, and as a consequence, low sea levels exposed large tracts of land around the peninsula. The lowest shoreline proposed by Ballard *et al.*(2000) at –155 m for the lake occupying the Black Sea Basin in the early Holocene might have had an impact on Crimea's climate, as extensive areas around the modern shores of the peninsula became exposed (Figure 4).

The emergent lowland to the west, north, and east of Crimea may have diminished the climatic influence of the sea, reducing moisture and creating continental conditions similar to those of southern mainland Ukraine. Maximum and minimum temperatures may have been more extreme, leading to warmer summers and colder winters than at present. The northwesterly and westerly winds that today bring moisture to southwestern Crimea may have become drier, since the sea had regressed southward. Pollen and paleosol records suggest steppe conditions in the Heraklean Peninsula from 11,000 to 7500 BP. The decline of broadleaf trees, the increase of steppe taxa, and the development of a chernozem soil suggest cool, dry climate conditions, which may have resulted from increasingly continental conditions produced by a drop in sea level.

Although evidence exists for an early Holocene low shoreline, the timing and mechanisms of the post-glacial refilling of the Black Sea are still debated. While Ryan *et al.* (1997) and Ryan and Pitman (1998) propose catastrophic flooding from the Mediterranean Sea, Ballard *et al.* (2000) propose a sudden but not catastrophic sea-level change over several centuries. Although the original date for this event was centered around 7.15 ky BP (Ryan *et al.* 1997), analysis of sea floor morphology as well as chemical and biological data have prompted Ryan's team to propose that the saltwater flood from the Mediterranean into the Black Sea could have occurred as early as 8.4 ky BP (Ryan *et al.* 2003). Opposing the idea of a flood, Aksu *et al.* (2002), Sperling *et al.* (2003), and Mudie *et al.* (2004) propose a gradual reconnection between the Black and Mediterranean Seas over the course of several millennia.

Ballard *et al.* (2000) obtained dates from fresh- and saltwater molluscs, placing the Black Sea refilling event between 7460±55 and 6820±55 BP, which agrees with the original date of 7150±100 BP proposed by Ryan *et al.* (1997). Based on studies of sediments in the Marmara Sea, Aksu *et al.* (2002) and Sperling *et al.* (2003) have proposed earlier dates for the reconnection of the seas through the Bosphorus. In any scenario, the reconnection and refilling of the Black Sea was completed before 6500 BP, with minor fluctuations occurring later on (see Ballard *et al.* 2000). The importance of this date is its proximity to 7000 BP, when striking changes in vegetation and soil development occurred in the Heraklean Peninsula (Figure 3). There is, however, insufficient evidence to attribute climate changes to the drop and rise of sea level.

6.2 The Possible Effects of Sea Currents

If Holocene immigration of southern flora occurred, then sea currents represent potential carriers of floral pioneers. This supposition emerged after

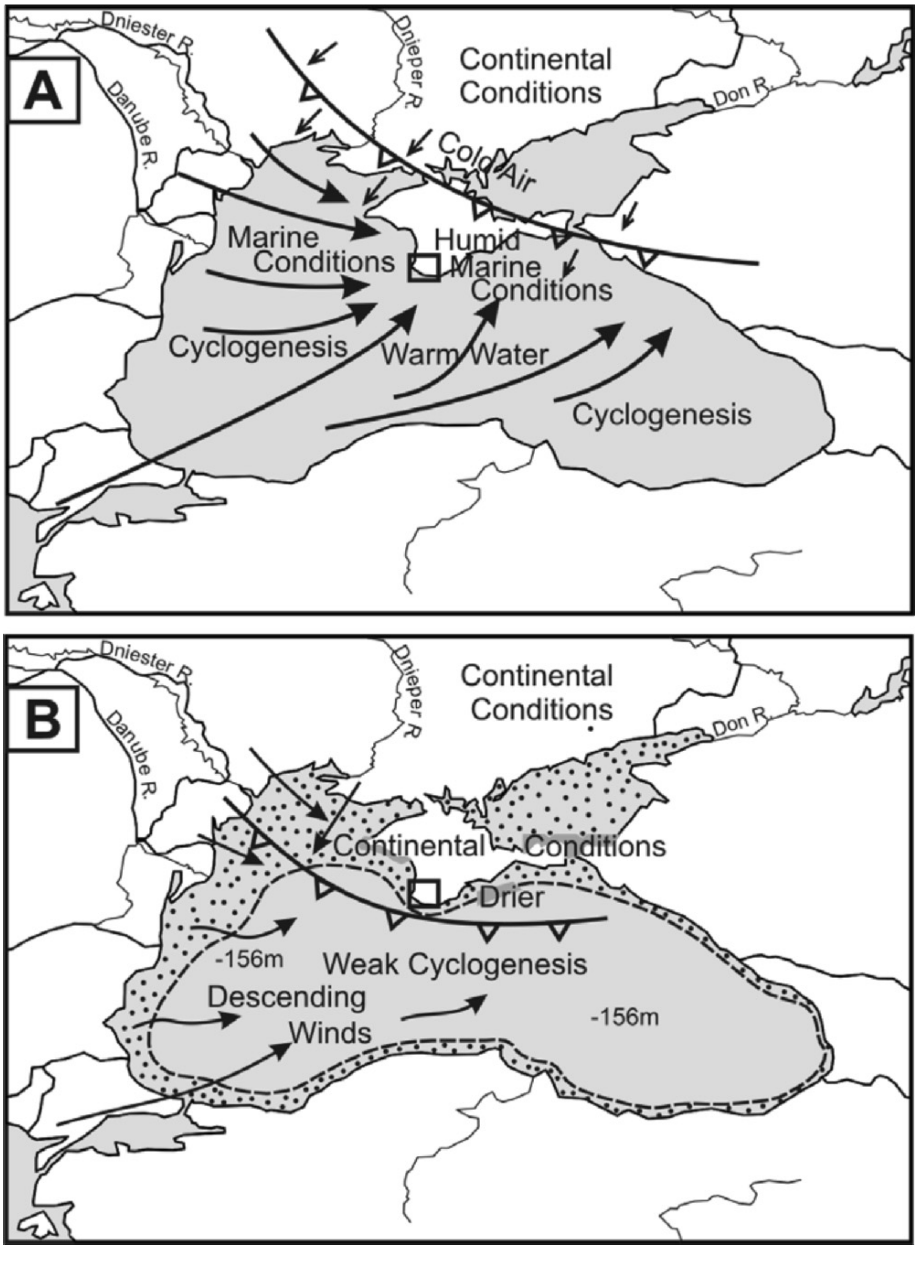

Figure 4. The Black Sea and its effects on Crimea's climatic conditions. A. Present predominance of marine climatic conditions; B. Hypothetical effects of low sea levels, in which continental conditions would prevail.

a series of observations on materials drifting along the southern Crimean coast. The drifting materials included both organic and non-biodegradable inorganic items, such as plastic bottles, spray cans, rubber sandals, etc. At the beach of Batyliman (see location on Figure 1), trash items with legible labels were sorted according to their country of manufacture (Table 3). The majority of floating items (58.6 %) were made in Turkey, suggesting that most items floated along the coast or were carried by currents all the way from Anatolia. Today, a sea current crosses the Black Sea northward, directly from the Sinop Peninsula in Turkey, and makes land at the southwestern coast of Crimea (Figure 5).

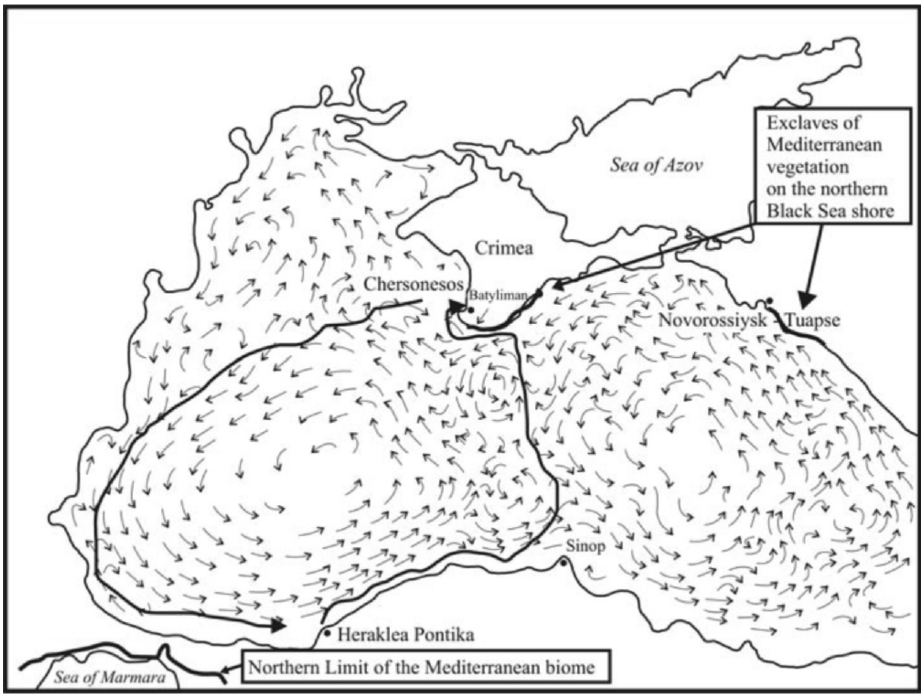

Figure 5. Black Sea currents and Strabo's trade and population migration route.

This is probably the same current mentioned by the Greek geographer Strabo in a ship itinerary between Heraklea Pontika and Chersonesos (Strabo 1924:7.4.3), which served as a guiding path for the main routes of trade and colonization of the western part of the Black Sea basin in Hellenistic times (Maximowa 1959; Gaydukevich 1969; Zolotarev 1979; Hind 1998). If the same circulation system existed in the Middle Holocene, then one could propose that the currents may have carried seeds and twigs of plants from lower latitudes to the southern Crimean shores. Recovered trash accumulations at Batyliman do not prove this assumption. Factors such as trading of these products around the Circum-Pontic area or the disposing of trash from ships must be considered.

Organic items in the trash at Batyliman include tree stumps, twigs, seeds, and other plant remains, which were not studied at this time. Very often, tree stumps drift along the southern coast of Crimea. On some occasions, their root systems carry lithologic materials not found in the geologic formations of Crimea, indicating long distance transport by means of sea currents (Alexander A. Klyukin, Tavriya National University, personal communication).

Table 3. Classification of trash found on the shore at Batyliman according to their place of manufacture.

Country of Manufacture	Number of Items
Circum-Pontic countries	**59**
Turkey (41), Ukraine (10), Russia (7), Bulgaria (1), Romania, Moldova, and Georgia (0)	
Other countries	**16**
Germany (3), Italy (3), Malaysia (2), Japan (2), France (1), Slovakia (1), Saudi Arabia (1), India (1), Namibia (1), and Brazil (1).	
Total	**75**
Distribution by type of material (N = 75) 69 floatable, non-degradable materials (e.g. plastic, rubber, aluminum) 6 perishable materials (e.g., cardboard, all of which were manufactured in Ukraine).	

The inorganic trash and tree stumps that wash up on Crimean beaches may not prove a direct connection from Turkey via currents, but they point to the possibility that sea currents may have played a role in carrying genetic plant material. Moreover, it is important to bear in mind that northward migration of vegetation via sea currents could have happened only after the Black Sea attained levels similar to the present. Coincidently, a similar set of currents flows toward the Novorossiysk-Tuapse region of the western Caucasus (Figure 5), and therefore, a scenario similar to the one hypothesized above can be applied here as well. Given the uncertainty that still haunts the timing of the refilling of the Black Sea, however, this scenario remains a hypothesis.

7. CONCLUSIONS

Pollen data from southwestern Crimea show a rapid change from forest to steppe around 11,000 BP and from steppe to Mediterranean shrubland and scrub during the period 7500 and 6000 BP. Transformations in the pollen assemblage are mirrored by changes in soil development, first from brown forest soils, to chernozem, and then to the Mediterranean cinnamonic soils. In part, the changes were due to rapid warming and altered circulation, which may have provided increased winter rains and a relatively long drought during the summer. After 4000 calBP, pollen of Mediterranean taxa decreases, suggesting that Mediterranean floras gradually retreated to their present distribution along the relatively warm and dry southern coast of Crimea.

The increased frequency of pollen from typical phrygana species suggests that humans may have also influenced the process of mediterranization of the Crimean vegetation. Unfortunately, there are no substantial local archaeological records to reconstruct the influence of humans on the process of vegetation transformation in pre-Hellenistic times.

Along the northern Black Sea coast, sea level changes brought about considerable shoreline migration due to the extensive area of continental shelf lying at shallow depths. A drop of less than 100 m in sea level compared to that of today may have affected Crimea more than any other region of the Black Sea coast. A drop of 150 m could easily transform Crimea from a peninsula to a highland set amidst a broad flat plain, and the elimination of widespread sea water surfaces around Crimea may have held tremendous consequences for precipitation and temperature in the study area. Although steppe conditions are reflected in the pollen diagram for the period between 11,000 and 7500 BP, it is not clear whether this vegetation emerged due to purely climatic influences or whether changing sea-land conditions should also be implicated as an additional factor affecting the climate.

Sea currents can also be viewed as a possible vehicle for plant migration from lower latitudes during the warming phase of the Middle Holocene. This assumption lacks adequate support, however, since evidence cannot be marshaled to reject the alternative assumption that Mediterranean plant elements may have existed in Crimea since the Tertiary Period, surviving the cold phases of the Pleistocene in refugia, and expanding during the interglacials. Unfortunately, the lack of continuous paleobotanical records for the Pleistocene leaves this question untestable for now.

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14

PONTIC-CASPIAN MESOLITHIC AND EARLY NEOLITHIC SOCIETIES AT THE TIME OF THE BLACK SEA FLOOD: A SMALL AUDIENCE AND SMALL EFFECTS

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Illustrations by Dorcas R. Brown

Abstract: Even if the Black Sea rose catastrophically and flooded the North Pontic plain about 7600–7300 calBC following Ryan and Pitman's revised chronology, only 100 to 150 individual foraging bands of 50–75 people each would have been forced out of the drowned North Pontic plains over a front more than 1000 km wide. We might be able to detect faint signs of their movements in some of the changes in distribution and types of flint tool kits in the Early Mesolithic, but there were no massive migrations outward from a flooded Black Sea basin. A review of Mesolithic and early Neolithic archaeological data in Ukraine provides little or no archaeological support for a sudden shift in human behavior at the time of the proposed flood.

Keywords: Mesolithic, forager populations, forager mortuary practices, subsistence change, ceramics, Neolithic, equid hunting, microliths, sea-level change

1. GEOGRAPHIC SETTING

The Black Sea sits within a deep geological trough with steep mountain walls on its southern, eastern, and southwestern sides. On its northern side, the slope of the land is gradual, and a broad region of arid plains borders the sea. This North Pontic steppe is surrounded by uplands to the north and west. These higher elevations catch most of the moisture carried in by the westerly Atlantic winds, leaving the lowland plain arid and treeless. To the east, the plain merges with the steppes and deserts north of the Caspian Sea, which, like the Black Sea, is surrounded by high, steeply sloping mountains on the south and by low-lying

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plains on the north.

During the Upper Paleolithic, between 32,000[1] and 10,000 years ago, a diverse array of hunter-gatherer societies occupied the Pontic-Caspian steppes (Soffer and Praslov 1993; Sinitsyn 2003). What is now the southern fringe of the Eurasian forest zone was then a periglacial steppe, the famous ‘mammoth steppe’, where Ice Age hunters built hide-covered huts over frameworks of mammoth tusks at places such as Mezirich and Mezin on the middle Dnieper drainage and at Kostenki II on the middle Don. Reindeer hunters lived in the Carpathian foothills at Molodova V (Grigor’eva 1980).

In what is now the steppe zone north of the Black and Caspian Seas, bison and horse hunters lived at sites such as Amrosievka and Anetovka II. The hunting patterns and tool kits of these hunters in the south were quite different from those of the colder mammoth steppe to the north. The area of the southern steppes was also much larger than it is now.

Ryan and Pitman (1998) suggested that a major Flood event filled the Black Sea trough catastrophically during the early Holocene, drowning a large portion of the North Pontic steppe and causing wide-ranging migrations among the humans who lived there, with profound effects on neighboring societies throughout eastern Europe. This paper reviews changes in economy and material culture among the Mesolithic societies in the North Pontic-Caspian steppe during the terminal Pleistocene and early Holocene. It focuses on the time frame of the Black Sea Flood in Ryan and Pitman’s proposal in order to look for significant changes in human settlement patterns or economic strategies that might have resulted from the event.

In addition, the dating of the Flood has been revised since Ryan and Pitman first suggested it, from about 6200–6000 calBC (7200 BP) in their original publication (Ryan *et al.* 1997) to about 7600–7300 calBC (8400 BP) in their modified chronology (Ryan *et al.* 2003). The paper will examine a range of human societies of both the 7th and 8th millennia BC in an effort to assess the likelihood of large-scale migrations occurring at the time originally proposed as well as the subsequent redating.

2. A BRIEF REVIEW OF THE FLOODS

At the height of the Last Glacial Maximum, *ca.* 20 ky BP, so much water was locked into the ice sheets and frozen into the earth in the form of permafrost that global sea level dropped 120 meters below that of today, while the northern shore of the Black Sea (certainly) and that of the Caspian Sea (possibly) retreated southward into their deepest basins. What is now the shallow northern shelf of the Black Sea was a grassy plain 100,000 km^2 in area, occupied by herds of bison (*Bison priscus*) and horses. Animals and the people who hunted them were able to move across the frozen Pontic-Caspian steppe with ease. But as the world's climate warmed during the period of deglaciation, the northern glaciers and permafrost melted, swelling the Pontic-Caspian rivers into impassable torrents. Vast lakes and swamps drained southward into the basins of the Black and Caspian Seas, and beginning *ca.* 16 ky BP, both seas grew to enormous size, drowning the plains to their north. The pattern of human settlement in the region was heavily dependent upon hydrological stability, and the geological and hydrological events of late glacial and early post-glacial times are disputed, as this volume makes clear. What follows is my understanding, based on recent literature as well as papers presented at the Columbia University conference,[2] of the hydro-geological history that provided the setting for the human occupation of the Pontic-Caspian steppes.

According to Chepalyga (this volume), during the 2000 years between about 16,000–14,000 BP, the Caspian stood at the highest levels recorded in recent geological history. The northern shoreline extended almost to Saratov on the Volga River and to just south of Orenburg on what is now the Ural River, blocking any east-west movement on the land surface south of the Ural Mountains. Russian geologists refer to this stage of Caspian evolution as the Khvalynian Sea. Under this enormous sheet of slightly brackish water (salinity of about 6–10 parts per thousand), a thick sediment of sand, clay, and silt accumulated that became home to a rich bottom community of shellfish (*Didacna*, *Dreissena*). The Caspian rose high enough to spill westward through the Manych Depression, now an arid zone of eroded karst limestone north of the North Caucasus Mountains, into the present Sea of Azov, and thence into the Black Sea. This Manych spillway poured enough water into the Black Sea basin to fill it to near its modern shoreline in perhaps 100 years, after which it broke through the Bosphorus Strait and flowed out into the Aegean, which finally drained the excess meltwater westward into the world ocean. Lake levels were briefly stabilized in both the Black and Caspian basins.

Continued warming led to the establishment of the belt of arid steppe north of the Black and Caspian Seas. Summer evaporation decreased flow within the feeder rivers, and both seas began to lose more water through evaporation than they gained through riverine inflows. The Khvalynian Sea retreated to near the modern Caspian shoreline, and the Manych channel became a series of shallow lakes in a dry, karst steppe. The Caspian would never again be connected to the Black Sea.

The Black Sea also contracted, though the pattern of its sea-level variations is not agreed upon by all researchers. Probably by about 13 ky BP, the link to the Aegean through the Bosphorus dried up. Ryan and Pitman propose a decrease in the Black Sea water level from 14.7–12 ky BP, then a rise from runoff, then another drawdown between 10–8.4 ky BP before their major flood at about 7.4–7.2 ky BP, when the level of the Aegean and Marmara Seas rose above the Bosphorus sill and poured into the Black Sea basin.

Aksu and Hiscott agree that the Black Sea declined after Chepalyga's Khvalynian flood, but in contrast, they propose that there was continuous outflow from the Black Sea to the Marmara after 11 ky BP (Aksu *et al.* 2002). Insufficient countercurrent inflow of marine water into the Black Sea kept the Pontic basin brackish, but aspects of marine conditions began to appear by 8.4 ky BP (about 7600–7300 calBC), and full colonization by Mediterranean marine molluscs occurred by 7.1 ky BP (about 6200–6000 calBC), accounting for the salinity changes at both of the chronological points that have been indicated by Ryan and Pitman as possible Flood events (see Hiscott *et al.*, this volume).

If Ryan is right, the 100,000 km^2 plain that represents the northern shelf of the Black Sea was exposed before about 8.4 ky BP. Herds of horses and saiga antelope moved across dunes and steppes that are today tens of meters beneath the waves. Submerged beach lines and coastal dunes have been identified by Ryan and others on the Black Sea shelf at depths from 150 to 50 m (Lericolais, this volume). These features are associated in their scenario with the drawdown after 10,000 BP and before their catastrophic flood about 8.4–7.2 ky BP.

If such an inundation happened—and it is still a matter of debate—the rising sea would have drowned the wide plain on the northern Black Sea shelf. By the end of the Late Mesolithic period in the North Pontic region (about 6000 calBC), the Black Sea had risen to the level of the world ocean. It has remained connected to the Aegean through the Bosphorus and Dardanelles Straits since then. Caspian Sea level has been more changeable because it is affected only by variation in river flows, rainfall, and evaporation rates in the surrounding watersheds, not by any linkage to the world ocean. Evaporation, which currently removes about 1 m of water per year over the entire Caspian Sea surface, and river discharge, to which the Volga is the greatest contributor, are both strongly affected by climate. Based on sedimentary evidence, Russian scientists have produced a complex chart of Caspian transgressions (rises) and regressions (falls) extending over the last 10,000 years (Matyushin 1986:136).

3. THE HUMAN CONSEQUENCES

Let us suppose that the flood happened as described by Ryan and Pitman. Are its effects visible archaeologically? Did any significant cultural or economic changes happen in the North Pontic or Caspian steppes about 7500–6000 calBC, a period embracing both the original and the revised chronologies for the catastrophic Flood event?

The societies of the Pontic-Caspian region at this time were foragers. There were no towns, no domesticated animals, and no farmers. The population of the plain north of the Black Sea, the largest land area affected by the proposed abrupt rise in sea level, probably consisted of small bands of mobile hunters and fishers, judging from sites investigated in the higher, dry parts of Ukraine. Ryan and Pitman (1998) initially speculated that the Black Sea flood of the early Holocene could have been the root cause of three migratory responses.

(1) It could have prompted a migration southward across Anatolia toward Mesopotamia, where the refugees' memory of a massive flood eventually became the inspiration for the Biblical flood of Noah.

(2) It could have pushed other populations westward, eventually bringing about the introduction of agricultural economies across central Europe by Linear Pottery farmers.

(3) It could have started a series of migrations north of the Black Sea that were responsible for the spread of the Indo-European languages.

The recent re-dating of the Flood (Ryan *et al.* 2003) has placed the event 1500–1000 years earlier than originally proposed, separating it chronologically from any of these events. As the data stand now, the first of these propositions is improbable and the second and third are impossible.

On the first proposition, the Biblical flood account is generally thought to have been inspired by a native Mesopotamian flood myth in which the role of Noah was played by the virtuous king Ziusudra (Kramer 1963:163–164, 224; Best 1999). There is no archaeological evidence for a migration from northern Anatolia into Mesopotamia around 7600–7300 calBC. The earliest Neolithic sites of northwestern Anatolia, near the Bosphorus, are dated about 6500 calBC, and their occupants migrated northward into that region from central Anatolia, i.e., they moved in the opposite direction, toward the Black Sea not away from it (Özdoğan 1999).

On the second proposition, the Linear Pottery expansion across central Europe began about 5500 calBC, so it could not have been caused by a Black Sea flood that occurred about 2000 years earlier and hundreds of kilometers distant from the nearest Linear Pottery site. Even if the events were closer chronologically, the Linear Pottery expansion had no connection with events near the Black Sea (Bogucki 1996).

Finally, the expansion of the Indo-European languages could not have begun before about 4000 calBC, since Proto-Indo-European contained vocabulary for wool and for wheeled vehicles, neither of which existed before 4000 calBC (Mallory 1989; Anthony 1995). Although Proto-Indo-European could well have been spoken in the steppes north of the Black Sea, it did not exist until 3000 years after the proposed early Holocene flood. This might be one reason why there is no hint of a world-destroying flood in the most archaic Indo-European myths.

Although these initial dramatic claims cannot be supported, it is still productive to consider the possible effects of the Black Sea floods on foraging societies of the late Upper Paleolithic and Mesolithic in the North Pontic steppes.

3.1 The Late Paleolithic and Mesolithic in the Caspian Depression

The Khvalynian-era Caspian Sea separated the western steppes–those west of the Ural Mountains–from the eastern steppes of Kazakhstan and Central Asia for about 2000 years, from 16,000 to 14,000 BP. East-west travel on foot was possible only across the Ural Mountains to the north. Earlier, during the cold Last Glacial Maximum, the Upper Paleolithic cultures of the Siberian steppes east of the Ural Mountains already were somewhat different from those west of the Urals (Boriskovsky 1993; Lisitsyn 1996), but during the Khvalynian flooding, the expanded sea cut off contact, isolating the cultures of the western steppes even more from those of Kazakhstan and Central Asia. As a result, the differences between them intensified. At the end of the Pleistocene, the spread of forests across the former mammoth steppe created an ecological contrast between the southern steppe cultures, which continued to hunt wild equids on the open plains, and the cultures of the north, which slowly adapted to life as forest-zone foragers.

After the Caspian retreated to approximately its modern basin (after about 12,000 BP), the land newly exposed by the Khvalynian regression, the North Caspian Depression, became a dry and challenging environment (Figure 1). A large sand desert called the Ryn Peski, or Red Sands (now the northern-most sand desert in the world), lies north of the Caspian and northeast of the Volga delta. During humid periods, the Ryn Peski was covered by steppe, and in dry periods, as today, it reverted to desert. The Caspian Depression elsewhere had a floor of variegated clay and sandy soils, dotted with brackish lakes and covered by dry steppe containing salt-tolerant stands of *Artemisia*. Occasionally, winds exposed fossil shoals of old *Dreissena* shells. Herds of saiga antelope (*Saiga tatarica*), onagers (*Equus hemionus*), and horses (*Equus caballus*) were hunted across these saline plains by small bands of post-glacial hunters. Their camps have been found among the dunes northeast of the Volga at places such as the Early Mesolithic site of Je-Kalgan (Dzhe-Kalgan) and the Late Mesolithic site of Suek-Te. Similar flint tool kits, containing geometric microliths in lunate and trapezoidal shapes as well as end-scrapers on small blades, were used in sites south of the Volga, as at Kharba. Igor Vasiliev (Vasiliev *et al.* 1996; Vasiliev 1998) has compared these tool traditions to those of the North Caucasus (at Mesolithic sites such as Satanai and Tomuxlovka) and has suggested that the Caspian steppes were re-occupied after the Khvalynian flood by forager groups from the south. It is probably unwise to rely solely on lithic tool kits to identify the ethnic and geographic origins of forager societies, but on the strength of the evidence, similar microlithic tool kits were used during the Mesolithic over a large region that included both the Caspian Depression and the North Caucasus.

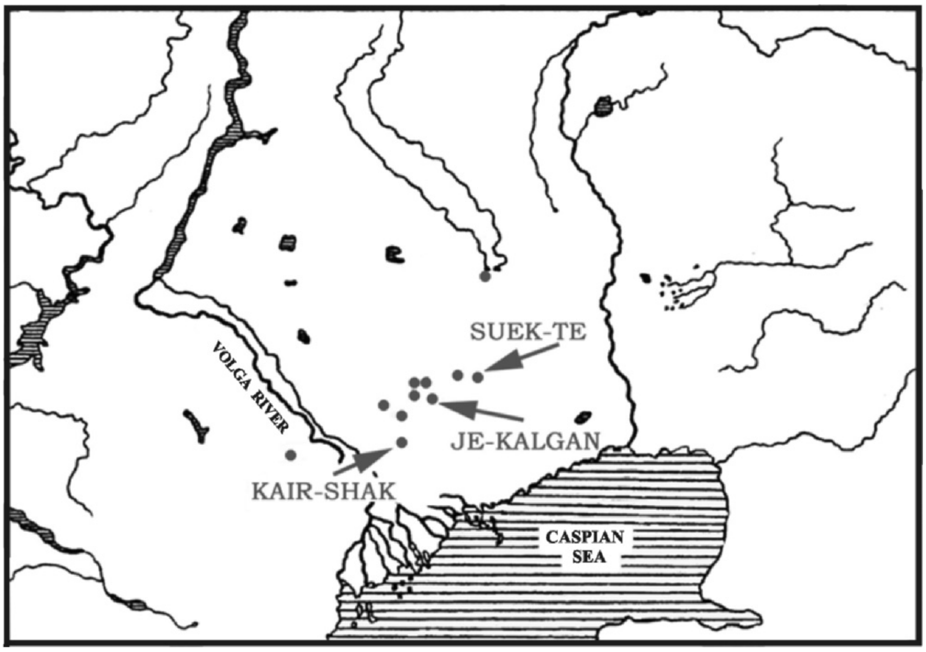


Figure 1. Map of Mesolithic sites in the North Caspian Depression (after Vasiliev *et al.* 1996).

3.1.1 Early Mesolithic in the North Caspian Depression: Je-Kalgan and Istai

Study of the Mesolithic in the North Caspian Depression was energized by a series of expeditions in the 1980s led by Igor Vasiliev of the Samara State Pedagogical Institute (Vasiliev *et al.* 1988; Ivanov and Vasiliev 1995; Vasiliev *et al.* 1996; Vasiliev 1998). At the Early Mesolithic camps of Burovaya 53 and Je-Kalgan in the North Caspian desert, Mesolithic garbage dumps contain almost exclusively the bones of onagers, *Equus hemionus* (Gorashchuk and Komarov 1998). In a few locations, the modern dunes have covered and protected ancient land surfaces that reveal an Early Mesolithic steppe environment. The floor of an Early Mesolithic dwelling about 3 m long and 2 m wide was preserved in such a context at Je-Kalgan.

Early Mesolithic forager camps of about 8000–7000 calBC differed from those that would follow in the Late Mesolithic. In the Early Mesolithic, the North Caspian climate was cooler and moister than it is today, and numerous inland lakes filled low places in what were then grasslands. Early Mesolithic campsites seem to have been quite small. The scatter of tools and animal bones at Je-Kalgan covered an area of just 10x10 m, and other sites were even smaller (Istai VI measured 4x4 m; Burovaya 53 measured 2x3 m). These Early Mesolithic sites seem to represent the camps of single families or small hunting parties. Most of them were situated within low swales or depressions rather than on hilltops. The flint tool inventories included many geometric microliths (principally long, broad parallelograms, and lunates with unifacial retouch on the arc), micro-chisels, and blunted blades, with few scrapers and no trapezoid forms (Figure 2).

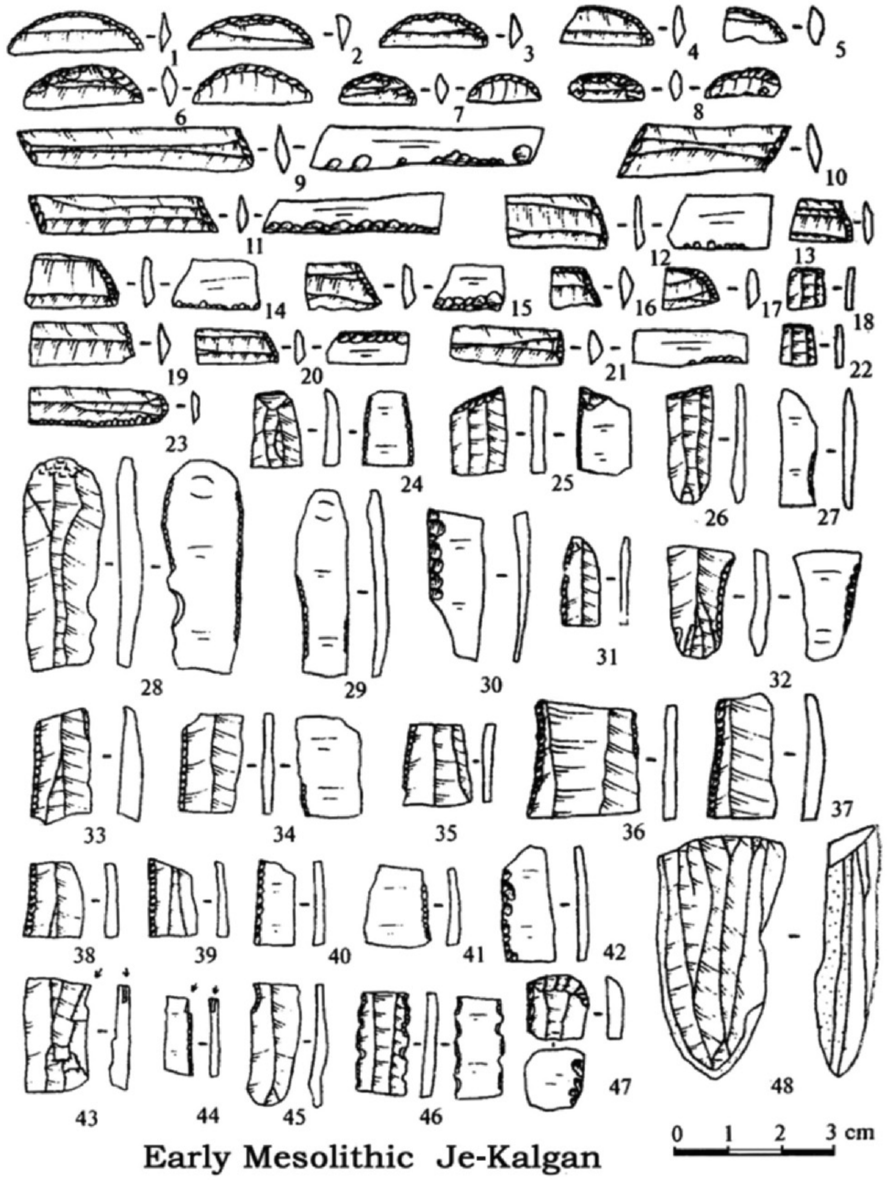


Figure 2. Flint tools of the North Caspian Mesolithic: Je-Kalgan (after Vasiliev *et al.* 1996).

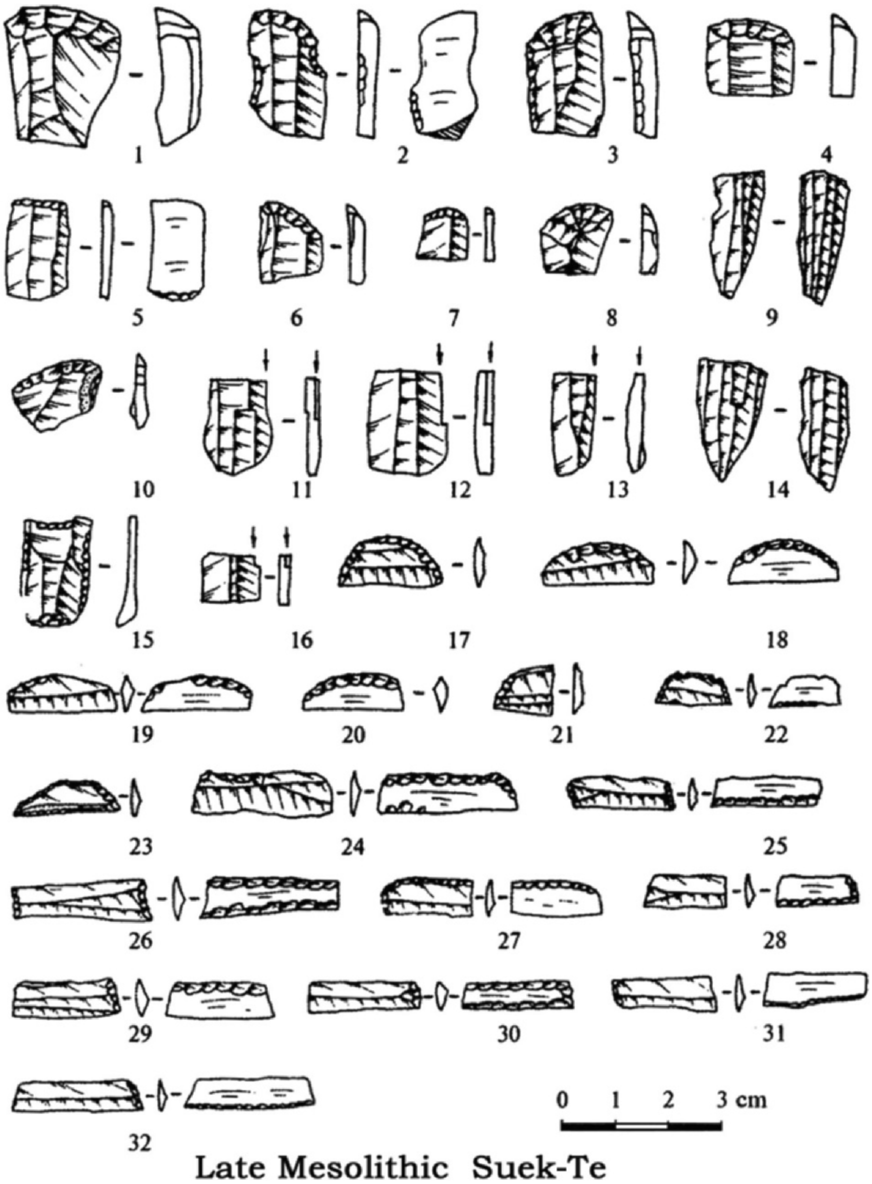


Figure 3. Flint tools of the North Caspian Mesolithic: Suek-Te (after Vasiliev *et al.* 1996).

3.1.2 Late Mesolithic in the Caspian Depression: Suek-Te and Istai IV

During the Late Mesolithic, about 7000–6000 calBC, the North Caspian plains became increasingly dry. Pollen evidence suggests that desert conditions spread at the time, with a peak dry episode about 6000 calBC, after which humid conditions returned. The dry episode was probably related to the very well-documented cold oscillation that occurred about 6200 calBC across the northern hemisphere (Perry and Hsu 2000). Late Mesolithic camps reached larger sizes; Suek-Te covered an area of 60x40 m. Onager hunting still seems to have supplied most of the meat in the diet (Kuzmina 1988). No screens or flotation methods were used in these excavations, however, and a more systematic attempt to use field methods that could recover fish and small animal bones might change the dietary picture. Late Mesolithic sites are more often found on ridge tops than in depressions, and the central concentration of sites moved 100 km to the north and east of the Early Mesolithic hunting grounds, away from the shrinking Caspian and its growing collar of desert.

The type site for the Late Mesolithic is Istai IV. Here, as at other Late Mesolithic sites, hunters used a grey flint different from the yellow-grey flint preferred during the Early Mesolithic. Tool kits included many end-scrapers on blades, finely retouched microblades 5–7 mm wide, oblique-ended microblades, short parallelograms, trapezoids, and short lunates with bifacial retouch on the arc (Figure 3). The percentage of geometric microliths was lower than in the Early Mesolithic.

Vasiliev and other specialists can see in these Late Mesolithic tool kits recognizable predecessors for the tools of the Neolithic cultures of the North Caspian region. On an even broader scale, they perceive important similarities among the Late Mesolithic toolmaking traditions of many lowland steppe foraging groups, including those of the North Caspian steppes (Istai IV type), the West Caspian steppes (Kharba type), and the Azov-Crimean steppes (Kukrekskaya type and Zimovniki type). Forager sites in all of these regions display similar lunates, parallelograms, and trapezoids, with a gradual decrease in the first two microlith types and an increase in trapezoids over time. When pottery-making emerged, the Early Neolithic cultures of the Azov-Caspian region (Surskii, Kair-Shak III) made pots with somewhat similar shapes and similar patterns of decoration. Vasiliev has advanced the idea that the dry Azov-Caspian steppes constituted a kind of 'cultural region' during the Late Mesolithic and Early Neolithic, a network of interacting forager bands.

To the north, the foragers of the Pontic-Caspian forest-steppe were not very different; many of the same microlithic traditions can be seen there. Unfortunately, no Late Mesolithic cemeteries are known in the Azov-Caspian steppes, so the physical type is unknown. In contrast, Mesolithic foragers living east of the Caspian Sea and east of the Urals made quite different kinds of tools and seem to have belonged to distinct social networks.

3.2 The Mesolithic in the North Pontic Steppes

Around the Black Sea, most of the forager camps of the Epipaleolithic and Early Mesolithic (about 9000–7000 calBC) were concentrated in locations well away from the modern shoreline, with the exception of the Danube-Dniester region (Figure 4) (Telegin 1982; Kol'tsov 1989). The principal Early Mesolithic settlements in the North Pontic region were situated in the same territories occupied during the Upper Paleolithic: (1) high in the Eastern Carpathians, where the reindeer hunters of Molodova V in the Dniester valley became the deer hunters and riverine fishers of the Early Mesolithic, and (2) in the steppes between the Dnieper and Donets river valleys, where the bison and horse hunters of the Late Paleolithic became the deer and horse hunters of the Early Mesolithic. Early Mesolithic occupation (at least 20 sites) is found in interior locations within the central part of Dobruja, the region of sandy hills and rock outcrops south of the Danube delta (Păunescu 1987). Early Mesolithic horse hunters also lived northwest of the Black Sea at Beloles'e, in the Dniester-Prut steppes. The Crimean Mountains sheltered a cluster of Early Mesolithic camps established within cathedral-like caves high up in the peaks, of which Shan Koba is the type site. The Late Mesolithic type site is Murzak-Koba.

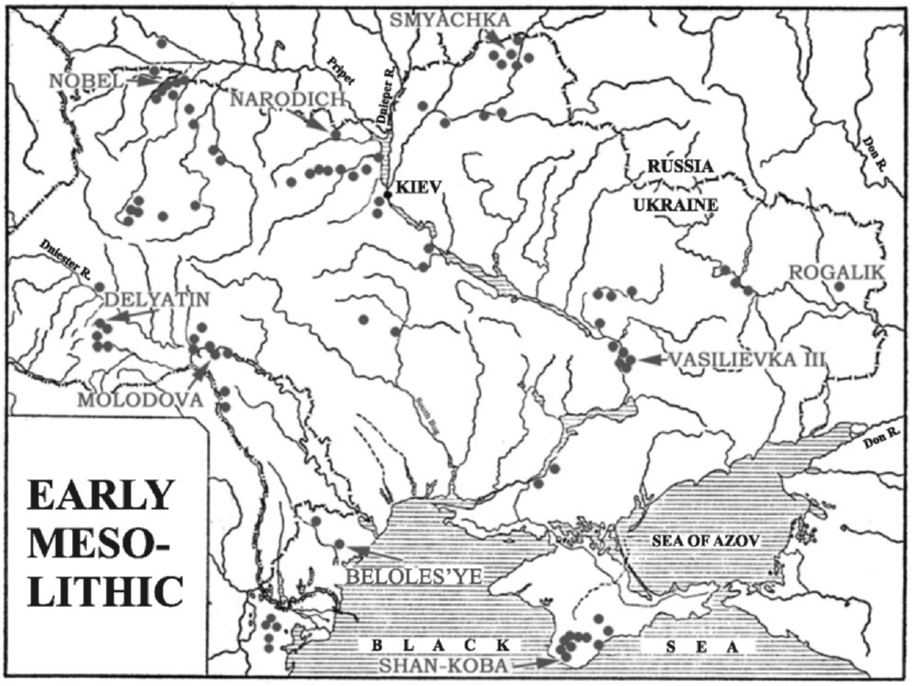


Figure 4. Map of Early Mesolithic sites in the North Pontic region (after Telegin 1982).

The Black Sea flood, if it occurred, would have converted the Crimean Peninsula from an isolated spur above a broad coastal plain to a peninsula almost completely surrounded by the sea at some time between about 7500–6000 calBC. If the Mesolithic forager population density on the pre-flood North Pontic plain was about 0.08 people/km^2, which is Dolukhanov's estimate for post-glacial forager population densities in the steppe zone (Dolukhanov 1979:23), up to 8000 people could have been displaced when the plain was flooded. They would have migrated to higher ground in the hills north and northwest of the Black Sea, and perhaps toward the new shoreline around the Crimean Peninsula. Archaeology does suggest that new sites appeared in the lowland steppe, near the modern coast, at the end of the Early Mesolithic and beginning of the Late Mesolithic, about 7000 calBC (Figure 5). One of these sites, Kukrek, provided the type assemblage for a new kind of flint tool kit (Kukrekskaya) that was widely used in the North Pontic steppes during the Late Mesolithic, 7000–6000 calBC. There was no significant change in settlement patterns, economy, or lithics that corresponded with the original date for the proposed flood at about 6000 calBC, but the shift from Early to Late Mesolithic tool types could possibly be correlated with the revised earlier date.

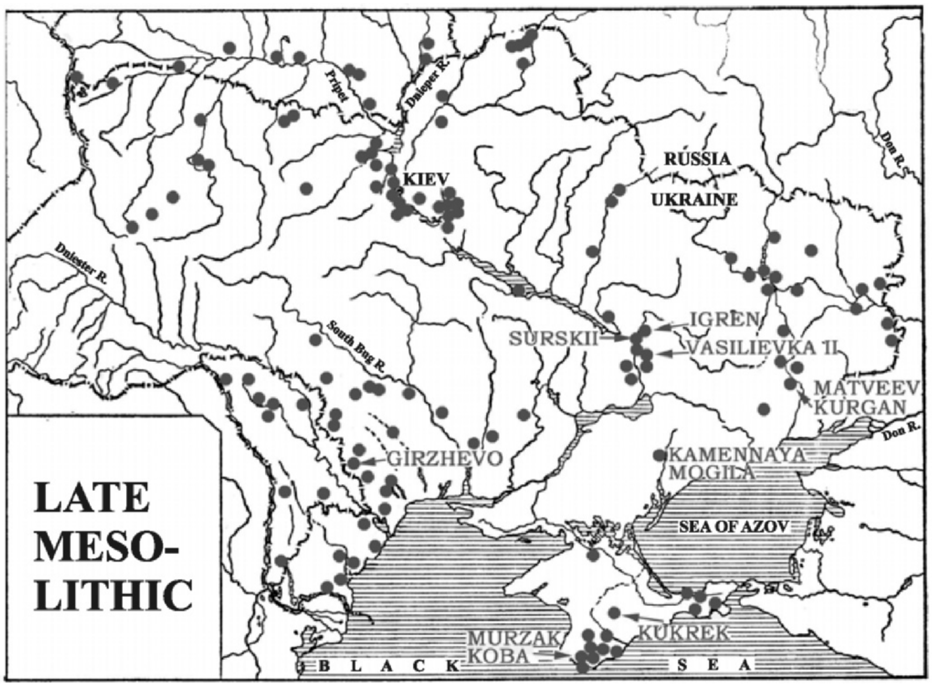


Figure 5. Map of Late Mesolithic sites in the North Pontic region (after Telegin 1982).

3.2.1 Mesolithic Settlements and Cemeteries at the Dnieper Rapids

The largest and most interesting cluster of Mesolithic sites in the North Pontic region was located near the Dnieper Rapids. The rapids started at modern Dnepropetrovsk and dropped 50 m in 10 cascades along a distance of 66 km.

Each cascade once had its own name and character. Fish migrating upstream, like the sudak (*Lucioperca*), could be taken in vast quantities at the rapids, and the shallow pools between the cascades were home to wels (*Silurus glanis*), a type of catfish that grows to 5 meters in length. Bones of both types of fish were found in Mesolithic camps such as Igren 8 near the rapids (Figure 6). At Igren 8, the deepest stratum, F, contained Late Mesolithic Kukrekskaya materials. Above this, strata E and E1 contained Surskii Early Neolithic, and the following stratum D1 held Dnieper-Donets I Middle Neolithic materials. At the end of the rapids, there was a ford at Kichkas where the river could be crossed relatively easily on foot. This ford was important because the high-water level in the Dnieper seems to have been higher during the Mesolithic than it is now. Mesolithic occupation sites located 3 m above high water levels of today were covered by water-deposited sediments laid down in the past by floods that must have post-dated the occupation of the sites.

The rapids and many of the archaeological sites associated with them were inundated by dams and reservoirs built between 1927 and 1958 (most of the data in this section is taken from Telegin 1982 and Kol'tsov 1989). Among the many sites discovered near the rapids during dam construction were eight Mesolithic cemeteries, most importantly Vasilievka I (24 graves), Vasilievka II (32 graves), Vasilievka III (45 graves), Marievka (15 graves), and Volos'ke (19 graves). Some cemeteries (Vasilievka I and III) that have been published as Late Mesolithic, broadly about 7000–6000 calBC, seem now to be very Early Mesolithic, closer to 8000 calBC, according to new radiocarbon dates (Jacobs 1993; Anthony 1994). Two cemeteries that were assigned to the Early Neolithic (Vasilievka II and Marievka) are now dated to the Late Mesolithic, 6500–6000 calBC. On close examination, these two cemeteries do not contain ceramics or any other Neolithic-era artifacts, so the new radiocarbon dates probably are accurate. Some changes in human skeletal morphology and diet that were once thought (Jacobs 1993) to have occurred between the Late Mesolithic (at Vasilievka III, for example) and the Neolithic (at Vasilievka II) now appear to have occurred between the Early and Late Mesolithic, so they cannot be associated with the beginning of herding and cereal cultivation. Nor can they be associated with a flood dated about 6000 calBC, at the end of the Late Mesolithic. But they could, in principle, be correlated with an event that happened just before 7000 calBC.

This writer does not follow Telegin *et al.* 2002 in extending the label 'Mariupol culture' back to 7000 calBC to include the Late Mesolithic cemeteries of Marievka and Vasilievka II. The term 'Mariupol culture' refers to the most distinctive regional aspect of the Eneolithic Dnieper-Donets culture, phase II (DDII). The appearance of deep ossuary pits filled with multiple layers of skeletons numbering in the dozens and accompanied by copper objects, ceramic vessels, polished axes, maces, and other unusual grave gifts happened only in the DDII period, after about 5400–5200 calBC, which also corresponded to the first

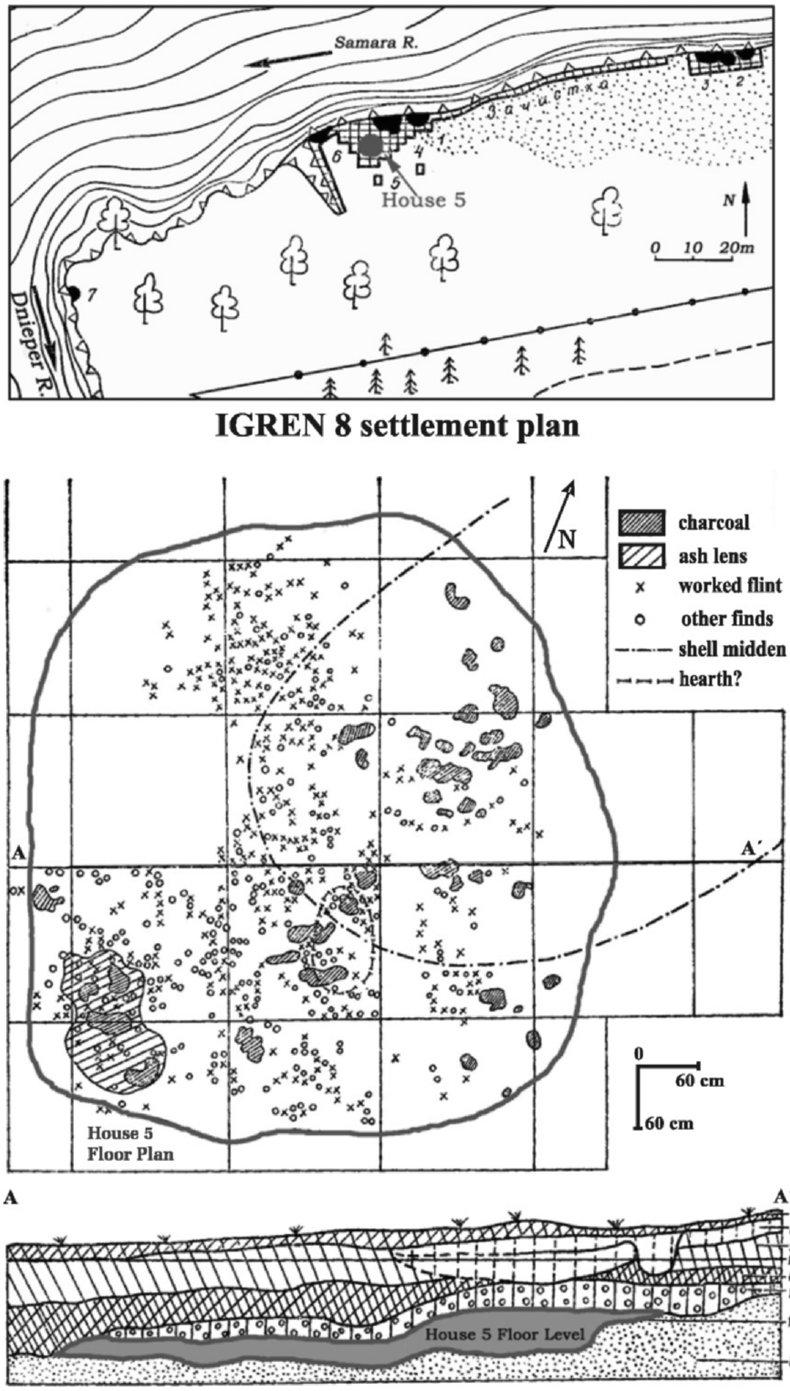


Figure 6. The Late Mesolithic settlement of Igren 8 (after Telegin 1982).

appearance of domesticated animals in the Dnieper valley. It is clear from the new dates that Vasilievka II and Marievka did not belong with this group of sites—they were mis-assigned to the Mariupol type. Vasilievka II and Marievka contain neither the kinds of artifacts nor the fully developed collective mortuary customs that were distinctive of the Mariupol-Nikol'skoe period. They should be called Late Mesolithic.

Whether during the Mesolithic or Neolithic, the Dnieper Rapids were a more stable natural feature than the Black Sea itself: they existed independent of the elevation of sea level. The stability and predictability of riverine fish resources and the existence of good fording places were the favorable factors that made the rapids a key territory throughout North Pontic prehistory, and no comparable cluster of Mesolithic cemeteries exists anywhere else in the Pontic-Caspian region. Elsewhere, the dead were buried rather casually in isolated graves or in very small groups of two or three in or near old campsites. The formal burial plots near the Dnieper Rapids probably functioned partly as visible and permanent claims to land rights, public statements that a particular cluster of families had traditionally lived and died in that territory. This implies a certain level of competition over the resources of the Dnieper Rapids. Several skeletons had pieces of microlithic flint blades embedded in their bones, the result of being shot or stabbed with arrows or spears armed with microliths. Two of the 19 individuals buried at Volos'ke and two (perhaps three) of the 45 at Vasilievka III were wounded by weapons tipped with Kukrekskaya-type microlithic blades. Both of these cemeteries were Early Mesolithic or even Epipaleolithic, dated about 8000 calBC, so competition and the creation of formal cemeteries around the rapids began long before 7500–6000 calBC.

3.2.2 Changes in Flint Tool Kits and Economy

Archaeologists have attempted to define regional culture groups in the Mesolithic of the North Pontic steppes based on the presence/absence or proportion of different kinds of flint tools. Telegin (1982) defined nine regional groups just in Ukraine during the Early Mesolithic and 11 regional groups for the Late Mesolithic. He then consolidated these into four Early Mesolithic and five Late Mesolithic culture 'types' or culture 'circles'. The principal difference between the Early and Late Mesolithic tool kits in the steppe zone was a decrease in the frequency of microlithic tools, from 20% in the Early Mesolithic to 8% in the Late Mesolithic (Telegin 1982: Table 7). But this change seems to have been a gradual trend, and it continued until microlithic tools disappeared from steppe tool kits in the Eneolithic. A hypothetical catastrophic Black Sea flood at either 7600–7300calBC (toward the end of the Early Mesolithic) or 6000 calBC (toward the end of the Late Mesolithic) had no visible effect on this continuous trend in lithics.

The steppe Late Mesolithic microlithic province is normally divided into two broad groups. One, Grebeniki, is found principally in the western Pontic steppes, centered on the lower Dniester and Bug River valleys, but extending southward to the lower Danube at, for example, Zaliznichnye and Mirnoe (Stanko and Subbotin 1979; Stanko 1986). Occasionally, Grebeniki-like tool assemblages are found farther east, as at Matveev Kurgan. Grebeniki assemblages contain many geometric microliths, about 15% of all tools, including many trapezoids, but no lunates and very few parallelograms. The other group, Kukrekskaya, was concentrated in the eastern part of the Pontic-Caspian steppes, centered in the Crimean and Azov steppes. It featured Mesolithic tool kits with a lower proportion of microliths (1–3%), almost all of which were simple blade inserts, not chipped into geometric shapes. Other steppe groups, like those of the Crimean Mountains (Murzak Koba type) and the dry desert-steppes of the Caspian Depression (Khalba and Istai IV), had many different geometric microliths, including lunates, trapezoids, and parallelograms. It is not at all clear what these differing proportions of flint tools meant in terms of culture or function, although some have speculated on the functions of different kinds of microliths (Sapozhnikova and Sapozhnikov 1986). The two main groups are not well separated geographically, since Kukrekskaya kits occur in sites northwest of the Black Sea (Girzhevo), and Grebeniki assemblages appear in the east (Matveev Kurgan).

Like the foragers of the Caspian Depression, the hunter-gatherers of the Pontic steppes depended economically on equid hunting (Bibikova 1978; Benecke 1997). In the Dniester-Prut steppes, horses were the primary game animals at both Early Mesolithic Beloles'e (75% of recovered fauna were horse) and Late Mesolithic Girzhevo (62% of recovered fauna were horse). Another equid, the extinct hydruntine (*Equus hydruntinus*), was also hunted in the northwestern Pontic steppes. On the northern and eastern part of the Crimean Peninsula, in both Early and Late Mesolithic levels at Frontovoe 1 and 3, horses and onagers were the principal game animals. Horses and onagers also were the principal game animals in the steppes north of the Sea of Azov, at Kammennaya Mogila (Benecke 1997) and at Matveev Kurgan (Krizhevskaya 1991). Matveev Kugan is called Early Neolithic (see below), but it was occupied at the Late Mesolithic/Early Pottery Neolithic boundary, about 6500–6000 calBC. In the upland forests and around the Dnieper Rapids, red deer, aurochs, and wild boar were the most important quarry. It is difficult to identify any change in hunting patterns in the coastal steppes corresponding with a proposed inundation of the northern plain either at 7600–7300 calBC or 6000 calBC. Equid hunting provided most of the meat in the diet both before and after these dates.

3.3 Early Pottery and the Beginning of the Neolithic

The adoption of ceramics, which made possible the first fire-resistant pots that could be left simmering on the fire, defines the beginning of the

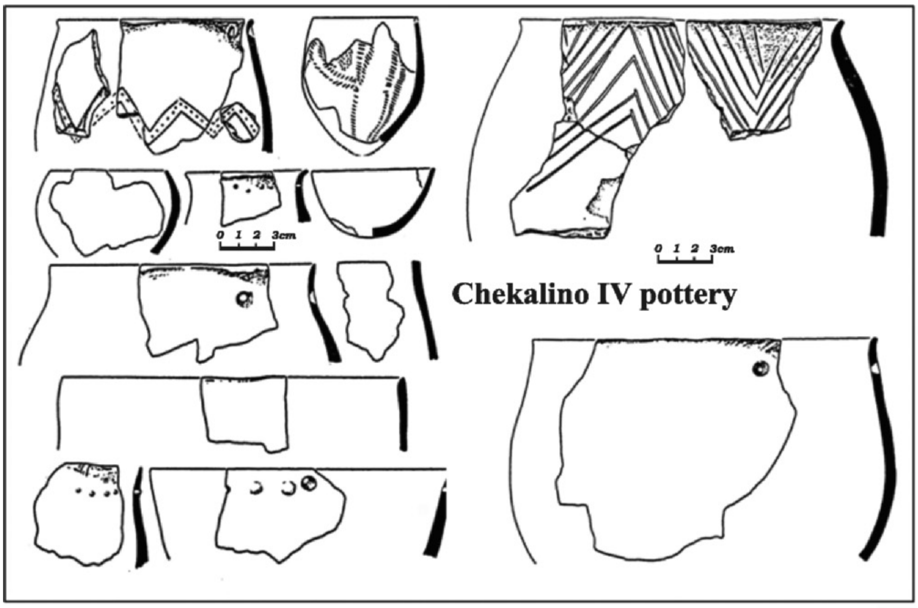
Neolithic in the terminology of Soviet and post-Soviet archaeology. It is odd that the old Soviet definition of 'Neolithic' societies depended on the presence of a new technology, pottery, and disregarded economy. In the Western terminology, 'Neolithic' refers to societies with a food-producing economy, regardless of ceramic skills. The Marxist definition ignored the mode of production and the Capitalist one depended entirely on it. Apparently, archaeologists were happy to ignore the political implications of archaeological terminology.

The first pottery appeared in the Pontic-Caspian region between 7000–6500 calBC. This was about the time originally proposed for the Black Sea Flood and the inundation of the North Pontic plain. Neolithic origins were associated with the Flood in Ryan and Pitman's book (1998). Radiocarbon dates reported in Telegin (1982), Telegin *et al.* (2002), and Timofeev and Zaitseva (1997) have had the effect of reassigning several 'Neolithic' cemeteries to the Mesolithic, blurring the line between the two periods. Many Westerners do not realize that the Soviet/post-Soviet definition of the term 'Neolithic' is quite different from the Western definition. So this seems like a good opportunity to review the Mesolithic/Neolithic transition.

3.3.1 The First Ceramics in the Pontic-Caspian Region

Pottery technology, the hallmark of the Neolithic period in Soviet terminology, has recently been shown to be surprisingly old in the middle Volga region. In fact, sites in the middle Volga region have the oldest pottery in all of Europe (Mamonov 1995; Timofeev and Zaitseva 1997; Bobrinsky and Vasilieva 1998). The middle Volga is 600 km north of the Caspian Sea and even farther from the Black Sea, so it is difficult to see how the appearance of ceramics in the middle Volga region could have been caused by sea-level changes in the Black Sea.

The first ceramics in the middle Volga region are called the Elshanka type. They were made of fine clayey silt collected from lake bottoms, coiled into vessels, and fired at 450–600° C. Elshanka vessels were round-bottomed or flat-based pots and open bowls decorated with incised zig-zag lines, rows of impressed pits, and fingernail impressions (Figure 7). They are dated to 7000–6500 calBC in the Sok and Samara valleys, east of the Volga. Chekalino IV, an Elshanka site on the Sok River, has 6 radiocarbon dates taken on shell. The two oldest dates seem anomalous: 8990±100 and 8680±100 BP. The other four fall between 8050±100 and 7940 ±100 BP, or about 7000–6700 calBC. Equally early dates for sites with Elshanka ceramics have come from Lebyazhinka IV on the Sok (8470±140 BP) and Ivanovskaya on the upper Samara River (8020±90 BP). Even earlier ceramics have been found much farther to the east, around Lake Baikal in Siberia, but it is uncertain if the middle Volga ceramics were inspired by these or not.



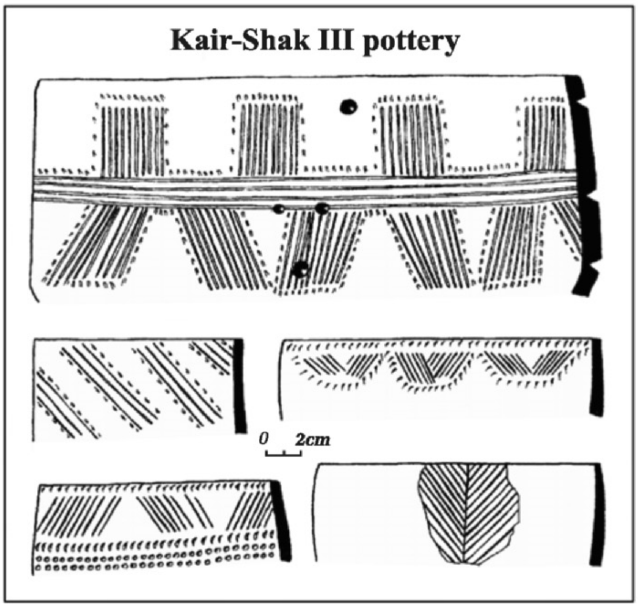


Figure 7. Elshanka Early Neolithic pottery vessels from Chekalino IV (after Mamonov 1995), and Late Neolithic Kair-Shak III pottery (after Barynkin and Kozin 1998).

The appearance of pottery in the middle Volga region marks the transition to the Early Neolithic. From this northeastern source, ceramic technology diffused southward—so influences actually moved from north to south, not from the Pontic basin to the north. Later but still very primitive ceramics were found at Matveev Kurgan in the steppes north of the Sea of Azov, dated 6400–6000 calBC. The two camps at Matveev Kurgan (1 and 2) have yielded a handful of very small pieces of straight-walled, undecorated, fired clay, perhaps sherds from straight-walled, flat-bottomed vessels, as the excavator has suggested, or perhaps griddles or liners for storage pits, or something else. No pot rims or bases were found.

Well-made pottery seems to have been adopted widely across the Pontic-Caspian steppes after about 6200–6000 calBC. Shell-tempered pottery decorated with incised geometric motifs is dated about 6200–5800 calBC at the Early Neolithic settlement on Surskii Island in the Dnieper Rapids (Telegin *et al.* 2003). A superficially similar kind of pottery, but assigned to a distinct archaeological culture, is dated about 5800 calBC in the North Caspian Depression at Kair-Shak III. Pottery like that at Kair-Shak III is stratified on top of even older pottery at the North Caspian settlement of Kugat. This lower Kugat pottery is undated but might be as old as Surskii (Barynkin and Kozin 1997). Both the Surskii pottery and the Kair-Shak III pottery featured S-profiled, shell-tempered, round-bottomed pots, decorated with incised and stamped geometric designs. Somewhat similar wares were found in the Azov steppes at Kammennaya Mogila. The settlement on Surskii Island contains, in addition to pottery, a bone net-making tool decorated with carved geometric motifs very similar to those carved on bone armrings in the nearby Late Mesolithic cemetery of Vasilievka II, dated about 7000–6500 calBC. Such a connection shows continuity in decorative design between the Late Mesolithic and Early Neolithic in the Dnieper Rapids region (Figure 8).

In the North Pontic forest-steppe uplands, the earliest radiocarbon-dated ceramics are early Bug-Dniester wares found at Bazkov Ostrov in the Southern Bug valley (perhaps about 6200–6000 calBC), followed by the Soroki camps in the middle Dniester valley (about 5900–5700 calBC). The Bug-Dniester sequence shows that pottery-making spread into the forest-steppe northwest of the Black Sea from two directions. First, the older native Pontic-Caspian pottery tradition spread from the east into the Southern Bug valley at about 6200–6000 calBC. Second, ceramics were adopted in the Dniester valley at about the same time that pottery-making pioneer farmers migrated into Moldavia from the west at about 5900–5700 calBC. These pioneer farmers of the Criş Culture also carried the Neolithic agricultural complex (in the Western sense of the Neolithic) into the North Pontic region.

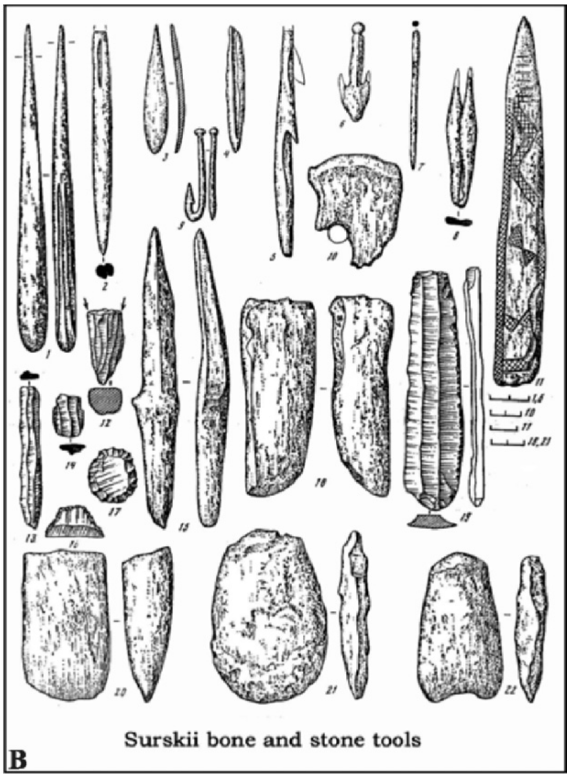


Figure 8. Decorative motifs on bone from (A) the Late Mesolithic Vasilievka II cemetery and (B) the Early Neolithic Surskii settlement (after Kol'tsov 1989).

3.3.2 The Local Domestication of Animals

Brief comment should be made on the persistent claims that economies based on the herding of domesticated animals had appeared in the Pontic steppes before 6000 calBC, that is, before domesticated animals were imported into the region by Criş Culture farmers from the Danube valley. The belief that herding economies appeared independently had three sources.

The first was confusion between the Soviet and the Western senses of the term 'Neolithic'. The *ceramic* Neolithic certainly did begin before 6000 calBC in the Pontic-Caspian region, but the economy remained one of hunting, fishing, fowling, and gathering.

The second source was V.I. Bibikova, the chief archaeozoologist at the Institute of Zoology in Kiev during the post-war blossoming of Soviet archaeology from the 1950s through the 1970s. Her argument for a local domestication of cattle and pigs was re-examined statistically by Benecke (1997) and was not supported. The occasional bones of domesticated pigs and cattle that Bibikova identified from Late Mesolithic ('aceramic Neolithic' in her terms) deposits at Surskii, Igren 8, and Soroki probably came from later levels at these complexly stratified sites. The sites were dug quickly, with few boxes, labels, and bags.

(The multiple individual bags and boxes of Western archaeology are the byproduct of a consumer culture in which every activity is commodified; Soviet archaeology pushed ahead with much less equipment.) Storage spaces were overcrowded. Even in Kiev, bones were kept in crumbling boxes in temporary outbuildings and in the unheated apse of a Medieval church. Given these field and lab conditions, a few animal bones could easily be placed in the wrong box.

The third source of the idea that animal domestication began independently in Ukraine is the archaeological site of Matveev Kurgan. The domesticated animals at this site included not just cattle and pigs, which had local wild counterparts, but also sheep (*Ovis orientalis*). There were no wild sheep in the Pontic-Caspian steppes; none appeared in Mesolithic bone assemblages. Matveev Kurgan site 1 (MK 1) seems to be dated to 6400–6000 calBC by radiocarbon dates: 7505±210 BP (GrN 7199) and 7180±70 BP (Le 1217). This is astonishing, because *Ovis orientalis* was native to the Near East, and at 6400–6000 calBC, there were no sheep-keeping cultures nearby from which the Matveev Kurgan people could have acquired Near Eastern sheep. The nearest shepherding cultures were still confined to northern Greece, Macedonia, southwestern Bulgaria, and Anatolia at this early date, all of which were at least 1300 km distant. The radiocarbon age of MK 1 predated the appearance of the Hassuna-related cultures of the Caucasus at Shulaveri, which, at 6000 calBC, were the earliest Caucasian groups to possess sheep. So how could sheep appear in the Pontic steppes so early? Where did they come from?

Matveev Kurgan is located in the Mius River valley north of the Sea of Azov. Two sites numbered 1 and 2, and located across a river from each other, were excavated here by L.Y. Krizhevskaya between 1968–1973 (Krizhevskaya 1991). MK 1 contained the remains of an oval dwelling and is thought to have been a residential camp. MK 2 was identified as a fishing and hunting station. Both contained Grebeniki-type microlithic flint tools, a typical Late Mesolithic complex, and are assumed by the excavator to have been occupied by the same people at the same time. The two radiocarbon dates from MK 1, averaging about 6400–6000 calBC, are frequently cited, but a piece of animal bone from the Grebeniki-type, Late Mesolithic deposit at MK 2 produced a third date, 5400±200 BP (Le 882), or about 4400–4000 calBC. This second dating shows that the Late Mesolithic deposit at MK 2 contained an unrecognized component from an era 2000 years later, when domesticated animals were common. Krizhevskaya combined the bones and artifacts from all levels into a single assemblage, arguing that flint tools of the same types were found in all levels. But at MK 1, the bulk of the flint tools and animal bones was found at 40–70 cm depth (Krizhevskaya 1991:8), while the dwelling floor and hearths were discovered deeper, at 80–110 cm (Krizhevskaya 1991:16). This vertical displacement indicates the presence of multiple occupation levels, with most of the animal bones coming from the upper level(s). The radiocarbon date from MK 2 shows that later archaeological deposits, unrecognized at the time, were mixed with the older materials. The mixing of older and younger deposits could explain the 21 bones of sheep/goat (4.6% of the 453 identified bones) found at MK 1 and four bones of sheep/goat (1% of those identified) at MK 2. Most of the animal bones were those of wild animals, principally horses, onagers, and wild pigs.

But Bibikova also identified many of the bones of horses, pigs, and cattle as those of domesticated varieties. All together, 24% of the bones from MK 1 and 2 were identified as those of domesticated animals. Bibikova's methods for identifying domesticated horses, which relied almost entirely on size, were not adequate; it remains a matter of debate whether early domesticated horses exhibited any systematic metric differences from their wild cousins (Anthony 1991; Anthony and Brown 2000; Benecke and von den Dreisch 2003). The horses at Matveev Kurgan probably were wild. Domesticated cattle and pigs are easier to identify, but wild varieties of both animals were also present at MK 1 and 2, so it is vital to know from which level the domesticated bones came. Unfortunately, we don't know this because all levels were combined in the analysis. Radiocarbon dating clearly demonstrates that later Eneolithic bones were mixed with the Mesolithic component at MK 2. As no other steppe site contains domesticated animals this old, Matveev Kurgan is not convincing as the earliest steppe site to reveal evidence of husbandry. It is, however, an important site for understanding the general outline of steppe forager economies at 6500–6000 calBC, the time of initial adoption of ceramics.

4. SUMMARY: GREAT SHOW, SMALL AUDIENCE

Human societies adapted successfully to the dramatic changes in climate and sea level that affected the Pontic-Caspian region as the last Ice Age waned. Although the changes in sea level and salinity were significant, whether sudden or gradual, their front-line effects on human societies were small and might be detected only by looking very closely at minor archaeological changes. Dolukhanov's (1979:23) population estimates suggest that, even if the Black Sea rose catastrophically and flooded the North Pontic plain following Ryan and Pitman's scenario, only 100 to 150 individual foraging bands of 50–75 people each were forced out of the drowned North Pontic plains over a front more than 1000 km wide. We can't really see the signs of such a sudden demographic movement in flint tool kits, subsistence economy, settlement types, or the creation of new technologies at any time between 7500 and 6000 calBC. No massive migrations moved outward from a flooded Black Sea basin. There was no horde of refugees carrying a legend of a great flood south to Mesopotamia. The floods in the Black Sea basin, whether they occurred suddenly or gradually, caused few long-term changes in human behavior.

ENDNOTES

1. The 32,000 BP date is from two sites at Kostenki on the Don River: Kostenki 1 layer III, and Kostenki 14 or Markina Gora. At the latter site, Sinitsyn has documented the oldest decorative beadwork in Europe in the context of an early Aurignacian assemblage of flint tools.

2. "The Black Sea Flood: Archaeological and Geological Evidence" held at Columbia University in New York City, October 18–20, 2003.

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15

FLUCTUATIONS IN THE LEVEL OF THE BLACK SEA AND MESOLITHIC SETTLEMENT OF THE NORTHERN PONTIC AREA

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Abstract: The ecological situation in late Paleolithic and Mesolithic times is often assessed on the basis of a dynamic equilibrium between the economic needs of humans and the ecosystem's natural capacity for resource renewal. This concept assumes a static notion of social structures and presumes not only stable subsistence systems, but also the persistence of a mode of production similar to those ethnographically recorded among hunter-gatherer groups on the periphery of developed civilizations. Based on paleoecological analyses of the North Pontic Steppe during the late Paleolithic and Mesolithic, we advance a different scenario: that human adaptation to environmental change was neither passive nor evolutionary, but both active and dynamic.

Keywords: adaptation, demography, catastrophe, population, paleoecology, ecological crisis

1. INTRODUCTION

Starting in the late Paleolithic, intensive hunting with the newly introduced bow and arrow led to the depletion of the large game animals that constituted the specialized prey species. This decline stimulated the quest for novel food resources through technical improvements in the tool-kit and the appropriation of additional foraging grounds. These improved tools and hunting techniques created an additional impact on the environment, interfering with natural processes. It has been proposed that rapid change in human society was enabled by such repeated disruptions in the equilibrium between environment and human needs. Chronic crises that affected prehistoric hunting and gathering communities on both local and regional scales finally resulted in the global crisis

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that struck hunting-based groups throughout the North Pontic Steppe in the late Mesolithic. Major changes took place within these communities and led to either demographic imbalance or an outflow of population (at least partially) from the settled area.

Fluctuations in sea level and the commensurate shrinking and expansion of littoral areas had considerable impact on the settlement pattern of the final Paleolithic and Mesolithic societies there. Two scenarios may be visualized. In the first one, rise in sea level had a catastrophic character and resulted in the total destruction of human settlements along the coast (Ryan and Pitman 1998; Book of Genesis 7:19–24). In the second one, the sea rose gradually at a rate of 1–9 m per millennium (Kaplin 1982), thereby allowing human groups to abandon their sites slowly. In the latter case, the rising water gradually inundated the sites without fully destroying them, and archaeological deposits would have been well preserved beneath the silty or peaty sediments. In the former case, no final Paleolithic or Mesolithic sites would have survived in the northwestern Pontic coastal territory.

Well-preserved final Paleolithic and Mesolithic sites on the Northwestern Pontic Lowland indicate a settlement hiatus of hunter-gatherer groups between 15.0 (Bol'shaya Akkarzha) and 6.5 ky BP (Igren' 8). These sites yield no evidence of an abrupt abandonment or disturbance that could have been caused by a catastrophic flood occurring at around 7.15 ky BP according to Ryan and Pitman (1998).

2. METHODS OF INVESTIGATION

In order to relate sea level migration in the Black Sea with prehistoric human settlement, one must assess first the paleoenvironmental changes and their potential impact on habitation sites, and second the spatial distribution of Paleolithic and Mesolithic sites in the coastal area, as well as their cultural and economic character. In dealing with this problem, we combined available geological and paleoenvironmental evidence pertaining to the mechanisms and character of sea level and climate changes with our observations on the stratigraphy, pollen, animal remains, and planigraphy of coastal Stone Age sites. We framed the problem theoretically within the concept of cultural adaptation, which views crises of the entire prehistoric subsistence system as stemming from reduced biomass and triggering either a readaptation to the changed environmental conditions or a migration in search of new resources.

Catastrophes, both global and local, have punctuated the human past causing innumerable losses and suffering, and often destroying all traces of human life in large areas. The consequences of these catastrophes cannot pass unnoticed, yet in each case, the evidence requires scrupulous and balanced assessment. It should be kept in mind that several catastrophes perceived by the ancients as global events (such as the Biblical Flood) might now be viewed as local ones with only limited impact.

3. PALEOENVIRONMENT

During the late Pleistocene and early Holocene (14–8 ky BP), the ecological setting in the North Pontic Steppe resulted largely from the interaction between changes in the physical environment with local human population dynamics and technical equipment, i.e., level of tool productivity.

Today, the North Pontic Steppe coincides mostly with the North Pontic Lowland, which stretches from the Lower Danube to the western ridges of the Donetsk Upland. The hydrological network, which was essentially created during the Pliocene-Pleistocene, established its main geomorphic features. The North Pontic Lowland can be divided into two regions according to the degree of surface erosion: one dissected by the ravines and gullies west of the Southern Bug River, and a relatively level plain east of that river. The most heavily dissected area is located within the Southern Bug-Danube interfluve, where the Black Sea cliffs are predominantly high and steep, and they often show evidence of marine abrasion. On the coast, the river valleys are mature, their width reaching 3–4 km, while the smoothed shorelines are often cut by ravines. This ravine-gully network becomes denser, and the valleys steeper, farther north.

In the early Holocene, the coastal area of the Pontic Lowland differed substantially from that of today, as it was much influenced by fluctuations in the level of the Black Sea (Figure 1). Based on the global eustatic rise of the Black Sea level and geological survey of the shelf, the following periods of stable shoreline are identifiable throughout the course of the Neoeuxinian and New Black Sea transgressions (Kaplin 1982): (1) late Pleistocene, at –55 to –65 m; (2) end of the early to beginning of the middle Holocene, at –20 to –30 m; (3) beginning of the late Holocene, at –10 to –15 m (Figures 1 and 2).

During the course of the Neoeuxinian regression, the present-day shelf was an exposed, loess-covered, alluvial plain connecting Crimea with Dobruja. Relief differed greatly from current conditions. The lower base level that existed at the time triggered erosive deepening of river valleys by an estimated 45–50 m (Ivanov and Ishchenko 1974). Hence, during the late Pleistocene and early Holocene, the western sector of the Pontic Lowland was available for human settlement and featured much deeper river valleys, ravines, and gullies, generally resembling the present-day setting to the north of the Lowland (Shcherbakov *et al.* 1976).

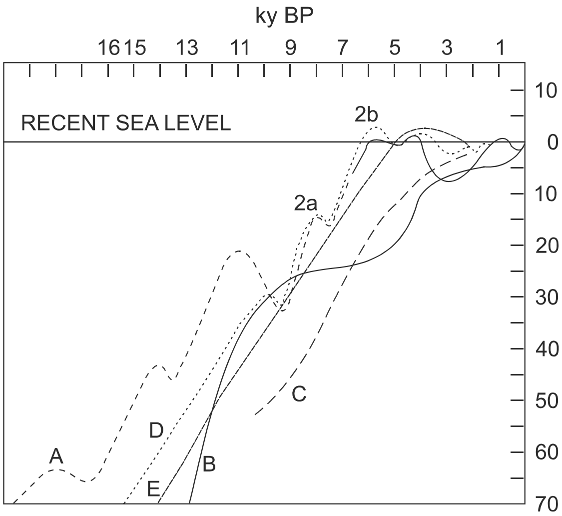


Figure 1. Schematic fluctuation curves for the level of the Black Sea in the late Pleistocene and Holocene. **Key:** A after Fedorov (1977); B after Shcherbakov *et al.* (1976); C after E.N. Nevessky (1967); D and E global late- and post-glacial transgression: D after Fairbridge (1962) and E after Müller-Beck (1967).

According to the above paleoenvironmental reconstruction, one can see that final Paleolithic and Mesolithic settlements (such as Bol'shaya Akkarzha, Kamenka, Beloles'e, Mirnoe, Trapovka, and others) were situated in the steppe zone at a great distance from the coeval shoreline: 100–170 km as the crow flies, or 200–250 km along the river. The low-elevation position of such sites as Bol'shaya Akkarzha, Beloles'e, and Mirnoe today is the result of sea-level rise and ingression into the lower reaches of the Dnieper, Southern Bug, and Dniester Rivers and their numerous tributaries (Stanko 1982, 1997a, b).

As a result, archaeological surveys targeted at final Paleolithic and Mesolithic sites on the northwestern Black Sea shelf and along the submerged river valleys might be deemed promising. Several flint tools retrieved from boreholes in various places on the shelf have already been identified by this writer. Such an archaeological survey of the northwestern Black Sea shelf might solve a number of major problems related to the character and chronology of the submergence, migrations, and the interrelationship between prehistoric groups of the Balkans, Central Europe, and Crimea.

The reconstructed environment of the Northwestern Pontic Lowland sheds new light on the possible settlement pattern of late Paleolithic and Mesolithic communities in Southeastern Europe, and in the Danube Basin in particular. Several writers have noted apparent similarities in the material culture of Late Paleolithic and Mesolithic sites in Bulgaria (Dzhambazov and Margos 1960), Romania (Boriskovsky 1964; Păunescu 1970), the northwestern Pontic area (Stanko 1982, 1985, 1997a), and Crimea (Danilenko 1960; Stanko 1997b). This similarity may be explained by direct contacts among human groups in all these areas (Stanko 1991), because in the late Pleistocene and Holocene, the northern Pontic area formed a huge steppe plain, twice as large as the present one, that was cut by numerous river valleys and ravines with forested slopes that bore coniferous cover during the colder stages, and deciduous cover during the warmer ones. It should be noted that broad-leaved forests were always present in that area (Artyushenko 1970; Paskevich 1981, 1982; Sirenko and Turlo 1986).

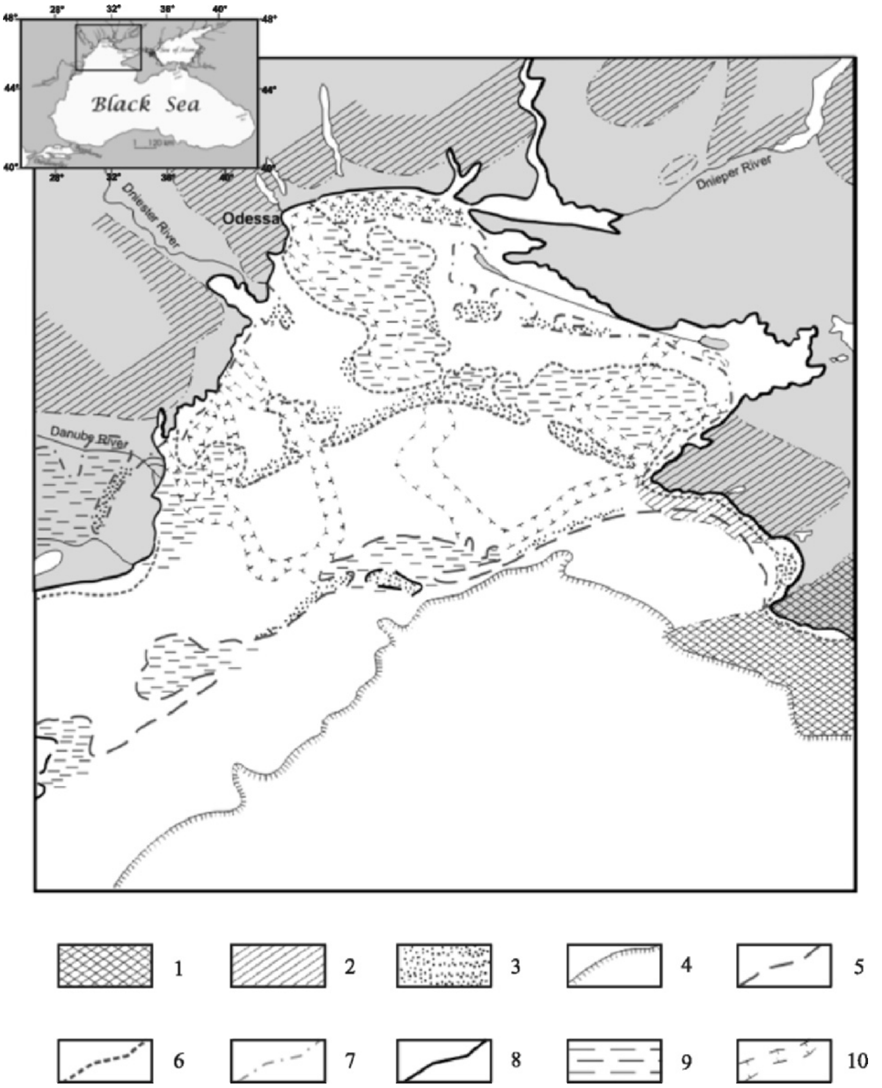


Figure 2. Paleogeography and Geomorphology of the Northwestern Pontic shelf. **Key:** 1 = Alpine-type thrust-and-fold Crimean Mountains; 2 = Post-Pontic plateau; 3 = Pliocene-Pleistocene alluvial plain; 4 = continental slope margin; 5 = initial Neoeuxinian coastline; 6 = final Neoeuxinian coastline; 7 = Old Black Sea coastline; 8 = present coastline; 9 = late Pleistocene limans; 10 = late Pleistocene river beds.

The animal world was rich, judging from the faunal remains found at late Paleolithic and Mesolithic sites. Faunal assemblages of prey species from the northern Pontic area were dominated by bison in the late Pleistocene, and by aurochs in the Holocene. Apart from these, species identified in late Pleistocene and early Holocene deposits include horse, woolly rhinoceros, reindeer, red deer, saiga antelope, and the now extinct wild equid *Equus hydruntinus* (Bibikova 1974, 1975, 1978, 1982; Bibikova and Belan 1979; Bibikova and Starkin 1989).

Lying beyond the glaciers during the ice ages, the steppe provided a refuge for both humans and animals. The present studies have made it possible to trace human population dynamics in connection with environmental changes since the Last Glacial Maximum. At that time, an intensive infiltration into the North Pontic area from the north resulted in the appearance of several large settlements, such as Sagaidak 1, Muralovka, Zolotovka, Anetovka 1, and others. The emergence of these sites coincided with the local extinction of the mammoth and woolly rhinoceros. It should be remarked that bones of both mammoth and woolly rhinoceros at late Paleolithic sites in the North Pontic area are extremely rare. Apparently, these animals survived Mousterian exploitation in only reduced numbers, and a few mammoth bones have been identified in the lower strata of only three late Paleolithic sites: Oskorivka, Vladimirovka, and Leski, in a specific industrial context deprived of microliths. Apart from Vladimirovka, bones of woolly rhinoceros were identified at Sagaidak 1, the only reliably radiocarbon dated site in this group (21,240±200 BP; LE-1602). Significantly, bones of neither mammoth or woolly rhinoceros have been found at sites of more recent age (20–15 ky BP). Beginning around 20 ky BP, bison became the principal hunted prey, and bones of this species are generally dominant in the faunal assemblages of late Paleolithic sites in the North Pontic Steppe (Boriskovsky 1961, 1964).

4. POPULATION DYNAMICS

Starting with the middle stage of the late Paleolithic, site inventories indicate the widespread use of microlithic blades and bone arrow- and spearheads. Backed microblades and micropoints, usually interpreted as inserts for complex arrow- and spearheads (Nuzhnyi 1992), became dominant among the secondarily retouched flint artifacts. Temporary stabilization of hunter-gatherer communities was achieved through an improved hunting technique using a bow and arrow. Several large, permanent settlements arose at the time with evidence of mass bison kills, including Amvrosievka (983 individuals) and Anetovka 2 (about 300 individuals). Judging from the number of butchered animals, one can hardly agree with Reeves (1983), who argued that the inhabitants of North

America's Great Plains *ca.* 6 ky BP were skillful bison hunters, whereas Mesolithic Europeans remained "primitive hunter-gatherers." Reeves obviously ignored the fact that the epoch of courageous and skilled bison hunters in the East European Steppe occurred at a much earlier age (20–14 ky BP), whereas by the 4th millennium BC, stock breeding and agriculture became firmly established in the area.

Influx from the Caucasus intensified with the beginning of the late glacial period (14–12 ky BP), resulting in a general increase in population density in the North Pontic Steppe. Judging from the limited number of animal remains at the sites, the density of bison conspicuously diminished. This period was marked by increased social mobility; in their quest for prey, humans settled the banks of smaller rivers and ravines.

At that time, the entire subsistence system was gradually restructured. The small herds of bison that survived the late glacial period practically disappeared by the end of the Pleistocene, and in the early Holocene, bison became extinct among the prey species of Mesolithic groups in the Northwestern Pontic Steppe. Mesolithic faunal finds reveal very few herd animals (such as wild ass and saiga antelope) that were common in the Pleistocene. These herd animals were replaced by solitary ones (such as aurochs, roe deer, and wild boar) that were better adapted to semi-closed biomes of forest-steppe type. Aurochs were the most commonly recovered within the Mesolithic Pontic Steppe sites (Bibikova 1975).

Modifications in animal distribution were related not only to climatic and landscape changes. Pollen evidence reveals (Figures 3 and 4) that vegetation shifts occurred as well (e.g., the spread of sylvatic coenoses on river bottomland and in the ravines as well as the widespread appearance of pasture meadows and hay lands in the watershed), but later, in the Atlantic period. The pollen record clearly shows that no drastic environmental changes took place during the Pre-Boreal and Boreal periods, i.e., during the Mesolithic (Pashkevich 1981, 1982).

At the beginning of the Mesolithic, the shoreline lay far from its present location. Yet, by the end of the period, the gradual sea-level rise had led to river mouth ingression. The archaeological layer at Mirnoe that is dated by ^{14}C to 7200±80 BP (close to the date of the alleged 'Flood') was stratified beneath peat-like deposits suggesting the occurrence of a small mire at the time (Stanko 1982). There is no evidence, however, of a precipitous abandonment of the coastal area, and Mesolithic sites existed in the area for yet another millennium. The occurrence of Mesolithic-type stone inventories within early Neolithic sites, which began spreading through the Pontic Lowland at 6500-6000 BP, suggests that small Mesolithic groups persisted all that time.

The demographic processes that occurred in the North Pontic area during the early Holocene (the Pre-Boreal) were complex. As the archaeological record shows, a small group of people moved into the Danube-Dniester interfluve

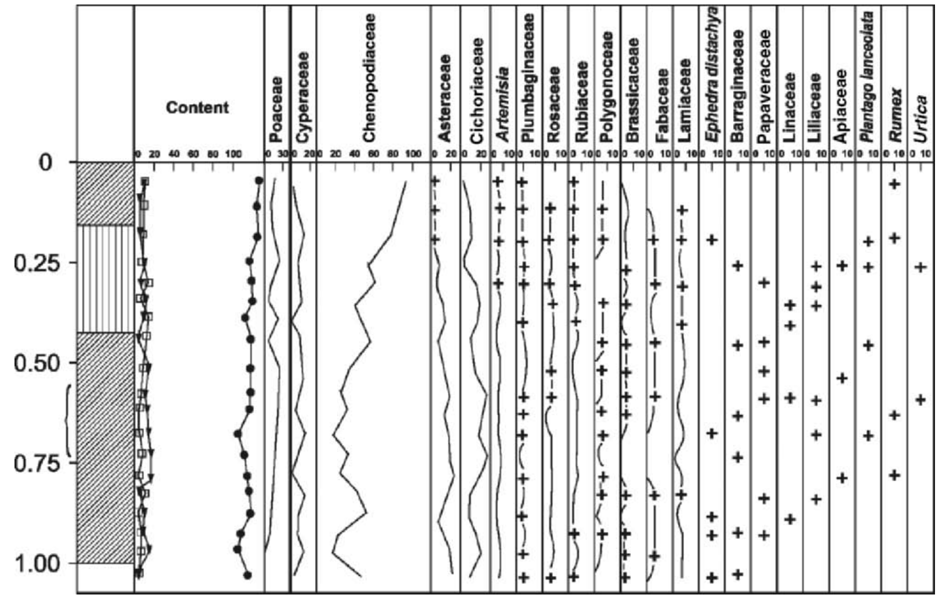


Figure 3. Pollen diagram from Beloles'e after Pashkevich (1981); + indicates less than 2%.

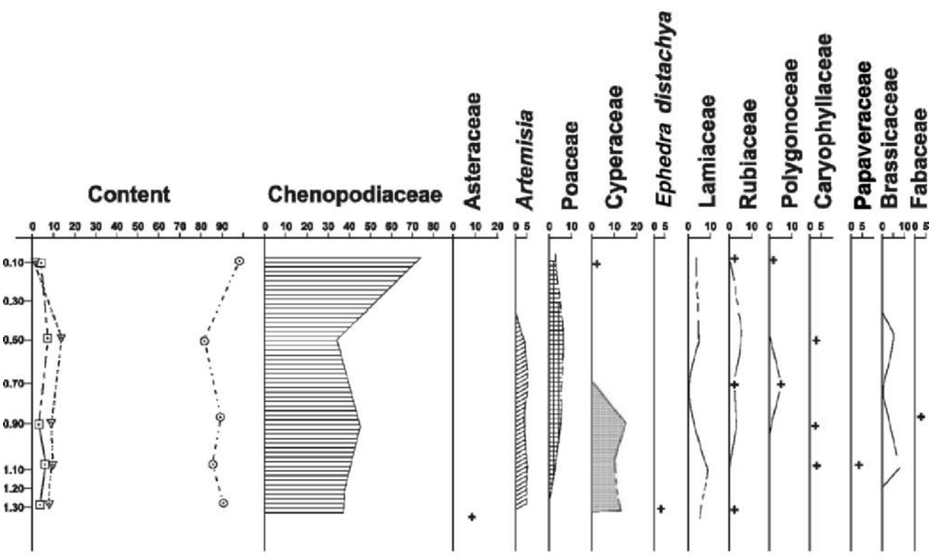


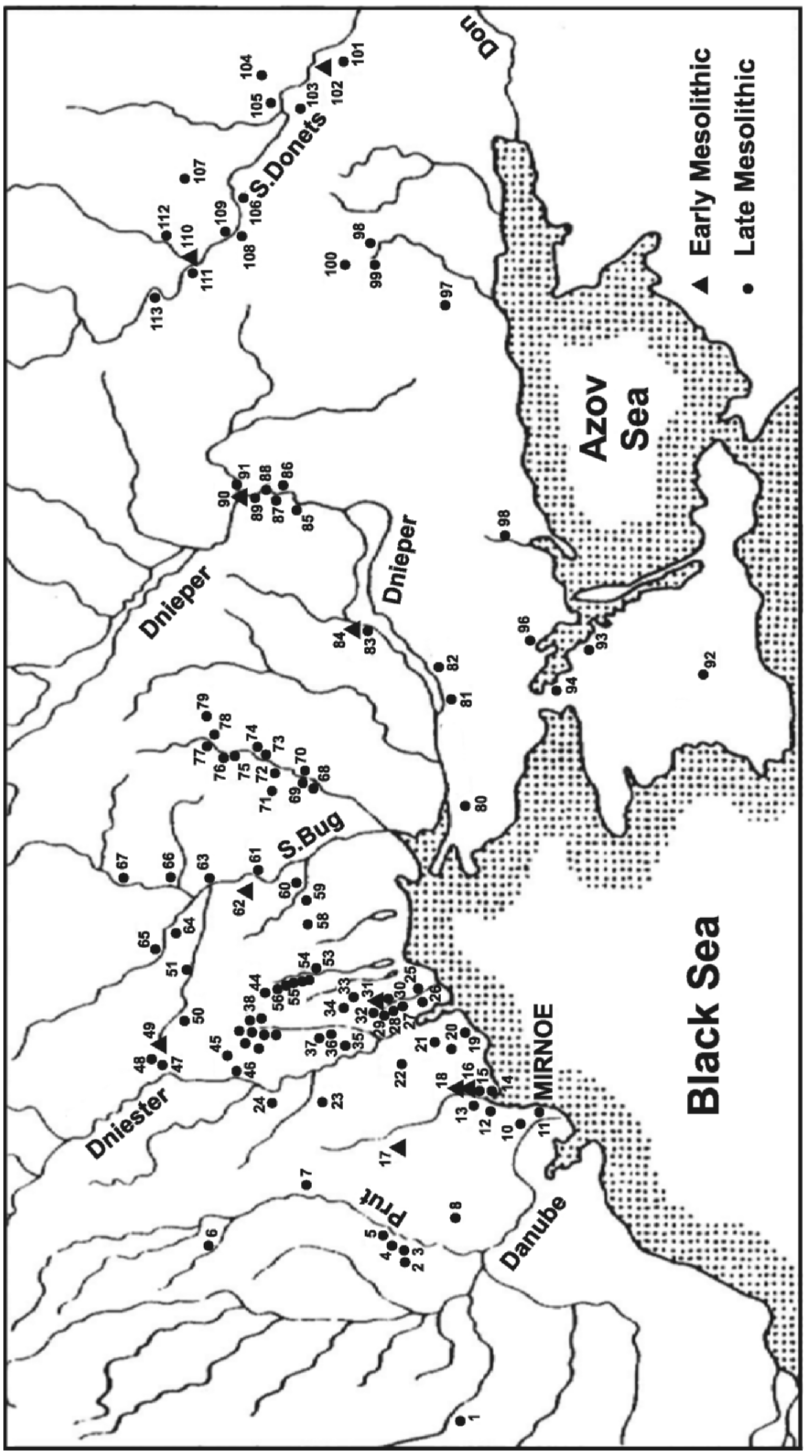
Figure 4. Pollen diagram from Mirnoe after Pashkevich (1982); + indicates less than 2%.

from the Balkan-Danube area (Stanko 1985, 1991, 1997a, b), and a restructuring of the hunting-gathering subsistence pattern followed (Tringham 1971, 1973). At the same time, other groups moved in from the Caucasus-Middle Asian area to the east (Efimenko 1953; Danilenko 1969; Stanko 1997a). Contacts with Gontsy-Borshevo communities in the north intensified, and under their influence, a new cultural entity, the Rogalik-Tsaritsian, arose in the steppe. In the late Mesolithic, its offshoots reached the Lower Danube (Dvorianinov 1976; Stanko 1986; Gorelik 1984, 2001). As this occurred, there was further development of the traditional cultures established in the steppe since the late Paleolithic (e.g., Anetovian, Kamennobalkian, and others).

No large settlements are recorded for this period. Investigated sites usually show very few fragments of animal remains, or none at all. Among the identifiable fauna, horse dominated both in the west (Beloles'e) and in the east (Rogalik-Yakimovska) (Bibikova 1978; Gorelik 2001). The bones of bison are practically absent, and those of aurochs are rare, particularly in the east. Apparently, aurochs had only started to penetrate the North Pontic area.

The emergence of farming was a gradual process deeply rooted in the local traditions, and in no way was it connected with a catastrophic flooding of the coastal area. The social and ecological situation that arose in the North Pontic area during the early Holocene does suggest the character of a local crisis, but it stemmed from the increased population density due to an influx of new groups (Figure 5) combined with the decreased biological productivity of the steppe due to the decline in ungulate herds. Bison, the principal prey species during the late Paleolithic, had been reduced to near extinction from overkill, and the rarefied herds of wild horse moved elsewhere in quest of better pastures. New species of solitary animals, mainly the aurochs, could not fully satisfy human needs, and the adoption of new hunting techniques necessitated a restructuring of the whole hunting-gathering subsistence pattern.

Geometric microliths, elements of a new type of inserted hunting armament, became widespread in Mesolithic stone inventories (as exemplified by Tsarinka, Leont'evka, Sorskoi 5, Rogalik-Yakimovska, and other sites). Micropoints and backed blades, in use since the late Paleolithic as inserts for arrow- and spearheads, were further improved. Grooved bone points were also widely employed. Food gathering became more complex. Potentially edible plants–lamb's quarters (*Chenopodium album*), black bindweed (*Polygonum convolvulus*), hairy vetch (*Vicia hirsuta*), and common sorel (*Rumex acetosa*)–have been identified in the archaeological deposits of Mirnoe. The broad use of plant foods by Mesolithic people and their cooking in ceramic vessels has been discussed by Danilenko (1969) and Bibikov (1977). Yet, hunting remained the principal means of subsistence for Mesolithic groups in the Pontic steppe, even as resources became increasingly depleted. Further intensification of hunting was no longer possible; it could only aggravate ecological imbalances. The relatively high population density, both in the valleys and on the watersheds, left virtually no vacant foraging territories. Stock rearing was the only way to increase productivity (Krizhevskaya 1974, 1992; Stanko 2003).



Early Mesolithic
Late Mesolithic
Black Sea
Azov Sea
Danube
Prut
Dniester
S.Bug
Dnieper
Dnieper
S.Donets
Don
MIRNOE

Figure 5. (Facing page) Mesolithic sites in the North Pontic area. **Key:** 1 - Laposh; 2 - Yerbichen' (Spinoasa); 3 - Beneasa II; 4 - Beresti; 5 - Malusteani; 6 - Ripiceani; 7 - Sarateni; 8 - Kiryatibya; 9 - Zaliznichnoe; 10 - Vasil'evka (Kitai-ozero); 11 - Mirnoe I, II, III; 12 - Borisovka; 13 - Tatrabunaty; 14 - Trapovka; 15 - Novoselidy I; 16 - Beloles'e IV; 17 - Beloles'e I; 18 - Kantemir; 19 - Shirokoe; 20 - Diviziya; 21- Marzaleevka; 22 - Tsarichanka; 23 - Zaim; 24 - Cobusca-Veche; 25 - Baraboi I, IV, V; 26 - Mar'yanovka; 27 - Iosipovka; 28 - Mirnoe-na-Baraboe; 29 - Dobrozheani; 30 - Krasnogorka I, II, III; 31 - Vasil'evka; 32 - Shcherbanka; 33 - Karpovo; 34 - Budyachki; 35 - Grebeniki; 36 - Girzhevo; 37 - Trostyanec; 38 - Karabanovo; 39 - Nikolaevka; 40 - Priimovo; 41 - Balashevo; 42 - Balashevo I; 43 - Malye Khatki; 44 - Frunzovka; 45 - Dovzhanka; 46 - Orlovka; 47 - Kodyma; 48 - Kodyma V; 49 - Tsarinka; 50 - Poznanka; 51 - Elenonka; 52 - Stepanovka; 53 - Prichepovka; 54 - Tsybulevka I-IV; 55 - Tsebrikovka; 56 - Katarzhino; 57 - Kominternovskoe; 58 - Skosarevka; 59 - Vasil'evka-na-Chichiklee; 60 - Slobodka; 61 - Arbuzova Balka; 62 - Anetovka; 63 - Konecpol'; 64 - Zaval'e; 65 - Gvozdev; 66 - Sinjukhin Brod; 67 - Novo-Arhangel'sk; 68 - Tyrnovo; 69 - Annovka; 70 - Novo-Petrovskoe; 71 - Lenino II, III; 72 - Bubinka; 73 - Glubokaya; 74 - Balakha; 75 - Sofievka I; 76 - Kunova Balka; 77 - Sagaidak; 78 - Varvarovka; 79 - Kazanka; 80 - Lower Dnepr Sands (Kuchugury); 81 - Kakhovka; 82 - Kairy; 83 - Anastas'evka; 84 - Leont'evka; 85 - Gadjucha Balka; 86 - Vasil'evka I; 87 - Vovnichi; 88 - Kizlevyii; 89 - Lokhansty; 90 - Surskoi Ostrov; 91 - Igren'; 92 - Kukrek; 93 - Dolinka; 94 - Vishnevka; 95 - Sergeevka; 96 - Kamennaya Mogila; 97 - Listvyanka; 98 - Mospino; 99 - Kremennaya Gora (Aleksandrovka); 100 - Plishcheevka; 101 - Murzina Balka; 102 - Zimovniki; 103 - Melovoe; 104 - Teplaya; 105 - Petrovka; 106 - Raigorodok; 107 - Pelrovo-Orlovskoe; 108 - Dronovka; 109 - Drobyshevo; 110 - Rubcy; 111 - Grushevakha; 112 - Prishib; 113 - Peirovskoe.

It took two more millennia until farming became established in the Pontic steppe. No tools suitable for tilling hard local soils are evident at Mesolithic sites. Neither have potentially domesticable wild plants ever been identified in the steppe biomes. By contrast, aspects of stock breeding were deeply rooted in the Mesolithic economies. Mesolithic hunters not only exterminated but also effectively controlled the wild herds. Analysis of mortality curves in the Mirnoe fauna shows that hunting focused on adult and semi-adult individuals only, while the young ones were spared and possibly additionally fed. The latter suggestion is substantiated by the finds of blades with sickle sheen identified by Korobkova (1993); supposedly, these blades were used for harvesting native grasses.

The available records of early Neolithic sites in the steppe, Matveev Kurgan 1 and 2 (Krizhevskaya 1974, 1992), show subsistence to have been of a complex character, based on hunting, fishing, and incipient cattle breeding, with no evidence of agriculture.

5. CONCLUSION

Early forms of stock breeding arose in the Northern Pontic coastal area because of a combination of factors, both anthropogenic (increased population density and improved hunting techniques) and environmental (changes in climate, modifications in the ungulate population, restricted feeding grounds, and sea-level migration), that gradually took hold there beginning with the early Holocene. Available evidence pertinent to both paleoenvironment and population dynamics reveals no catastrophes for the time span between 14 and 6 ky BP.

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16

THE NORTHWESTERN BLACK SEA: CLIMATIC AND SEA-LEVEL CHANGES IN THE LATE QUATERNARY

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Abstract: This paper summarizes the environmental changes, including water-level fluctuations, in the Black Sea during the Late Pleistocene and Holocene based upon data drawn from the literature as well as the authors' preliminary research results in the western Black Sea and Danube Delta. Evidence from the Danube Delta has been used to construct a curve of relative water-level changes in the Black Sea over the last 100,000 years. Arguments for and against the catastrophic flood hypothesis are discussed, but questions involving the rate of sea-level change and related coastline migrations during the Holocene cannot be answered confidently without more extensive study of the problem.

Keywords: Sea level, glacial, lowstand, interglacial, highstand, flooding, Black Sea, Danube Delta, Bosphorus sill, Mediterranean Sea

1. GENERAL SETTING

The Black Sea is one of the largest enclosed seas in the world, covering an area of about 4.2 x 10^5 km^2 and possessing a maximum depth of 2,212 m. Its total volume is 534,000 km^3, but most of this water (the 423,000 km^3 that lies below a depth of 150–200 m) is anoxic and contaminated with H_2S.

The Bosphorus and Dardanelles Straits provide the sole connection between the Black and Mediterranean Seas. The Bosphorus is narrow (0.76–3.6 km) and shallow (presently 32–34 m at the sill). It restricts the two-way water exchange between the very saline Mediterranean Sea (with a salinity of 38–39‰) and the more brackish Black Sea (about 17‰ at the surface and 22‰ at

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the bottom). The surface discharge of Black Sea water has been estimated at about 600 km^3/yr (~20,000 m^3/s), while the heavier Mediterranean undercurrent streams into the Black Sea, resupplying it with about half its outflow, 300 km^3/yr (~10,000 m^3/s) (Özsoy *et al*. 1995).

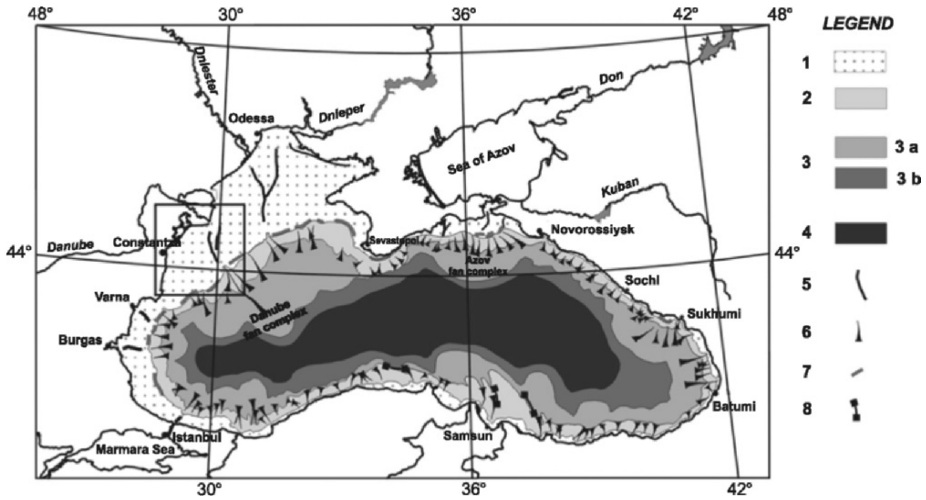

Figure 1. The Black Sea: main geomorphologic and sedimentogenetic provinces (after Panin *et al.* 1997): (1) Continental shelf; (2) Continental slope; (3) Continental apron: (3a) superior, (3b) inferior; (4) Deep zone; (5) Limits among provinces; (6) Main canyons; (7) Paleo-valleys; (8) Structural elements.

The northwestern corner of the Black Sea is especially suitable for a study of sedimentation and coastline migration during the Late Quaternary (Figure 1). Here, the continental shelf widens dramatically and encompasses about 25% of the total area of the sea. Here also, two of the largest rivers in Europe deliver their water and sediment load into the Black Sea–the Danube, with a water discharge of about 200 km^3/yr, and the Dnieper, which contributes 54 km^3/yr. The discharges of smaller rivers, such as the Dniester (310 m^3/sec) and Southern Bug (82 m^3/sec), add a little more, bringing the total inflow into the northwestern Black Sea to about 255 km^3/yr.

2. ENVIRONMENTAL CHANGES

Throughout its geologic history, the Black Sea experienced major sea-level changes accompanied by severe environmental modifications, including geomorphological reshaping. The most spectacular changes were driven by the Quaternary glaciations and deglaciations that reflect responses to Milankovitch cycles of 100 and 20 ky periodicity. These sea-level fluctuations endowed the restricted channels of the Bosphorus and Dardanelles with the power to control the Black Sea's connection to the Mediterranean. When global sea level dropped below the Bosphorus sill, water level in the Black Sea varied according to regional conditions independent of those in the world ocean. Perhaps the most important consequence of these lowstands was the interruption of Mediterranean inflow into the Black Sea, which, lacking any saline contribution, gradually became a giant brackish to freshwater lake.

Quaternary environmental changes in the Black Sea have been studied in detail from different perspectives, including geomorphology (of the coastal, shelf, slope, and bathyal zones), sedimentology (sedimentary environment, sediment architecture, and geochemistry), faunal and floral assemblages, paleomagnetism, and absolute age dating. Attention has focused largely on the Late Quaternary because these recent events can be characterized with much greater accuracy. Nevertheless, studies over the course of a century have not eliminated all the controversies surrounding what happened and when.

Although the interpretation of glacial and interglacial effects differs to some extent from author to author, the principal Pleistocene glacial periods in Europe (Danube, Günz, Mindel, Riss, and Würm) produced regressive phases in the Black Sea. Connections with the Caspian Sea were sometimes maintained via the Manych Depression, but the Pontic basin became isolated from the Mediterranean. Faunal assemblages during highstands were typically marine with Mediterranean influences, but during the lowstands, they were transformed, becoming representative of freshwater with a pronounced Caspian character. Features generated during the lowstands, such as paleo-cliffs, old submerged beach ridges, and sand dunes, are numerous on the shelf and shelf break, but they are difficult to date, while the amplitudes of sea-level change cannot readily be ascertained because their indicators are often masked by uplift or subsidence within the coastal and shelf zones.

Shcherbakov *et al.* (1978, 1979) suggested that environmental changes in the Black Sea can be recognized and characterized through the sometimes contradictory features found in three main geomorphologic and sedimentary zones: (1) the coastal zone, (2) the continental shelf, including the shelf break, and (3) the bathyal zone.

The following tables summarize the evolutionary phases of the Black Sea and provide correlations with the Mediterranean and Central Alpine Europe. Table 1 (A and B), reproduced from Fedorov (1978) with a few modifications, involves the Black Sea coastal and shelf zones. Table 2 (A and B), slightly modified from Shcherbakov *et al.* (1979), provides stratigraphy and correlations between the Black Sea's inner and outer shelves and the bathyal zone. All tables reflect a "classic" view of Black Sea development.

Table 1A. Stratigraphy and correlations of the Upper Quaternary phases for Europe, European Russia, and the Black Sea Region; after Fedorov (1978) with minor modifications.

General scale		Europe		European Russia	Black Sea Region	
					General stratigraphic scale	
Holocene		Flandrian		Holocene	Black Sea Horizon	Nymphaean
						Phanagorian
						New Black Sea
						Old Black Sea
Pleistocene	Upper	Grimaldian - Würm (regression to –100/–130 m)		Ostashkovian	Neoeuxinian	Late Neoeuxinian
				Mologo-Sheksnian		Early Neoeuxinian (Post-Karangatian)
				Kalinian		
		Neotyrrhenian (terrace at 2–8 m asl)		Mykulinian	Karangatian	Upper Karangatian
						Lower Karangatian
	Middle	Regression (Riss II ?) Deepening of Bosphorus to 100 m		Moskovian	Upper Euxinian-Uzunlarian	Regression
		Eutyrrhenian (Tyrrhenian Ib) (terrace at 10–20 m)		Odyntzovian		Uzunlarian
		Regression (Riss I ?)		Dneprian		Late Paleoeuxinian
					Lower Euxinian-Uzunlarian	Regression
		Paleotyrrhenian (Tyrrhenian Ia) (terrace at 18–30 m)		Lykhvinian		Paleouzunlarian
						Early Paleoeuxinian
	Lower	Mindel (Roman Regression)		Okan	Regression	
		Cromerian	Sicilian 2 (terrace at 60 m)	Dnestrian	Chaudinian	Upper Chaudinian
			Sicilian 1 (terrace at 100 m)			Lower Chaudinian
					Gurian-Chaudinian	
		Günz (regression)				Regression
Eo-Pleistocene		Emilian-Calabrian		Morozovian-Nogayskian	Gurian	

Table 1B. Stratigraphy and correlations of Upper Quaternary phases for the coastal and inner shelf zones of the Black Sea; after Fedorov (1978) with minor modifications.

General stratigraphic scale		W and NW Black Sea	N Black Sea, Crimea, Kerch, Taman	E Black Sea, Caucasus
Black Sea Horizon	**Nymphaean**	Terrace at 2 m;sands with *Cardium edule* L., etc.	Terrace at 2 m; sands with *Cardium edule* L., etc.	Terrace at 2 m;sands with *Cardium edule* L., etc.
	Phanagorian	Regression to –6/–8 m; archaeological layers V-I c. BC	Regression to –6/–8 m; archaeological layers V-I c. BC	Regression to –6/–8 m; archaeological layers V-I c. BC
	New Black Sea	Terrace at 4–5 m; sands and shells with *Cardium edule* L., *Chlamys*, *Ostrea*, *Mytilus*	Terrace at 4–5 m; sands and shells with *Cardium edule* L., *Chlamys*, *Ostrea*, *Mytilus*	Terrace at 4–5 m; sands and shells with *Cardium edule* L., *Chlamys*, *Ostrea*, *Mytilus*
	Old Black Sea	Clayey sands with *Cardium edule* L., etc., at 10–20 m water depth on shelf	Clayey sands with *Cardium edule* L., etc., at 10–20 m water depth on shelf	Clayey sands with *Cardium edule* L., etc., at 10–20 m water depth on shelf
Neo-euxinian	**Late Neoeuxinian**	Würmian loess; clays with *Monodacna caspia* Eichw., *Dreissena polymorpha* Pall., at 20–30 m water depth on shelf	Clays with *Monodacna caspia* Eichw., *Dreissena polymorpha* Pall., at 20–30 m water depth on shelf	Clays with *Monodacna caspia* Eichw., *Dreissena polymorpha* Pall., at 20–30 m water depth on shelf
	Early Neoeuxinian (Post-Karangatian)	Regression to –60/–80 (–130) m; Würmian loess; deepening of the valley incisions	Loess-like deposits; alluvial-deltaic sands; deepening of Kerch Strait	Regression; deepening of the valley incisions to –60/–80 m
Karangatian	**Upper Karangatian**	Terrace at 15–16 m Shells and sands with *Cardium tuberculatum* L., *Paphia senescens* Coc., etc.	Terrace at 8–12 m (4–8 m Taman); shells and clays with *Cardium tuberculatum* L., *Paphia senescens* Coc., *Aporrhais pespelicani* L., etc.; at the base, clays with *P. senescens* Coc., *Certhium vulgatum* Burg.; terrace at 12–15 m (Pshady valley), 25–30 m (Sochi region)	Shells with *Cardium tuberculatum* L., *Paphia senescens* Coc., *Aporrhais pespelicani* L., *Cerithium vulgatum* Burg., etc.
	Lower Karangatian			
Upper Euxinian-Uzunlarian	**Regression**	Regression; clayey loess-like deposits	Clayey deposits with *Limnea*, *Planorbis*; pebbles with *Viviparus*	Regression; alluvial pebbles, terminal moraine at Amtkheli
	Uzunlarian	Terrace at 35–40 m (Bulgaria); upper Babel layers, sands with *Didacna nalivkini* Wass., etc.; uppermost lagoonal clays	Clayey sands with *Cardium edule* L., *Didacna nalivkini* Wass, etc.	Terrace at 25–30 m (Pshady) and 35–37 m (Pshady valley); pebbles, sands with *Cardium edule* L., *Mactra stultorum* L., *Scrobicularia*
	Late Paleoeuxinian		Sands and clays with *Didacna nalivkini* Wass., *D. pontocaspia* Pavl., *Viviparus*	Terrace at 40–42 m (Pshady valley); sands, conglom., limestones with *D. nalivkini* Wass., *D. subpyramidata* Prav., at the base *Balanus*
Lower Euxinian-Uzunlarian	**Regression**	Regression	Regression	Regression; diluvium
	Paleouzunlarian	Sands, clays with *Didacna pallasi* Prav., *D. nalivkini* Wass.; Lower Babel layers; lagoonal clays with *Didacna pseudocrassa* Pavl., etc.	Continental deposits with the Mandzhil terrace	Terrace at 45–50 m (at Ashe, Makopse, Magri); pebbles with *C. edule*, *Paphia* sp., *Chione gallina*
	Early Paleoeuxinian			Terrace at 60–65 m (Dzhubgy); sands, pebbles with *Didacna baericrassa* Pavl., *D. pallasi*, *C. edule*
Regression		Alluvial sands with *Viviparus* and Tyraspol complex of mammalians	Top deposits with *Archidiscodon* sp.	Regression

Table 2A. Stratigraphy and correlations of Upper Quaternary phases for the continental shelf of the Black Sea; after Shcherbakov *et al.* (1979) with minor modifications.

Northern Europe					BLACK SEA				
Stratigraphic subdivisions					Bathymetric zone 0–50 m		Bathymetric zone 50–200 m		
					Layers	Molluscs	Horizon	Molluscs	Diatomaea
Holocene		Upper		Subatlantic - 2800	**Dzhemetinian**	*Divaricella divaricata*, *Gafrarium minimum*, *Pitar rudis*, *Cardium papillosum*	**Phaseolinus muds**	*Modiolus phaseolinus*	*Coscinodiscus radiatus*, *Thalassiosira excentrica*, *Actinocyclus ehrenbergii*, *Cyclotella kutzingiana*, *Cyclotella aceolata*
		Middle		Sub-boreal - 4800 Atlantic	**Kalamitian**	*Chione gallina*, *Spisula subtruncata*, *Mytilus galloprovincialis*	**Mytilus muds**	*Mytilus galloprovincialis*, *Cardium edule*	*Coscinodiscus radiatus*, *Thalassiosira excentrica*, *Asteromphalus robustus*
				- 7800	**Bugazian-Viteazian**	*Cardium edule*, *Abra ovata*, *Corbula mediterranea*, *Mytilaster lineatus*, *Monodacna caspia*, *Dreissena polymorpha*	**6800±140**	*Mytilus galloprovincialis*, *Cardium edule*, *Monodacna caspia*, *Dreissena polymorpha*	*Thalassiosira excentrica*, *Stephanodiscus astraea*, *Synedra buculus*, *Navicula palpebralis* var. *semiplena*
		Lower		Boreal - 9400 Pre-boreal	**Neoeuxinian**	*Monodacna caspia*, *Dreissena polymorpha*	**8550±130**		
Upper Pleistocene	**Würm (Valdai)**	Upper	Ostashkovian g.	- 10,200 Younger Dryas Allerød Lower Dryas Bölling		*Dreissena polymorpha*, *Viviparus fasciatus*, *Unio* sp.	**13,500±1500**		
				Gothiglacial Pomeranian	**Karkinitian**	*Dreissena polymorpha*, *Cardium edule*	**17,760±200**	*Dreissena rostriformis distincta*	
				Frankfurtian Brandenburgian - 25,000	**Tarkhankutian**	*Cardium edule*, *Abra ovata*, *Dreissena polymorpha*	**~ 22,000** **~ 25,000**		
		Middle	M-S. i.g.	Paudorf Arcy Gottweig - 40,000	**Surozhian**				
		Lower	K.g.		**Regression**				
				- ~ 65,000	**Post-Karangatian**				
	Riss-Würm	Mikulinian interglacial		Eemian - ~ 125,000	**Karangatian**				

Abbreviations:
g. = glacial
i.g. = interglacial
M-S. i.g. = Mologo-Sheksnian interglacial
K.g. = Kalinian glacial

Table 2B. Stratigraphy and correlations of Upper Quaternary phases for the bathyal zone of the Black Sea; after Shcherbakov *et al.* (1979) with minor modifications.

Northern Europe					BLACK SEA				
Stratigraphic subdivisions					Bathyal zone – northern part		Bathyal zone – southern part		
					Horizon	**Diatoms, Molluscs**	**Horizon**	**Nannoplankton, Dinoflagellates**	**Age**
Holocene		Upper		Subatlantic - 2800 Sub-boreal	**Coccolith ooze** *(Dzhemetinian)*	*Coscinodiscus radiatus*, *Endictia oceanica*, *Thalassiosira excentrica*, *Asteromphalus robustus*, *Rhizosolenia calcar-avis*	**Coccolith ooze Unit 1**	*Emiliania huxleyi*, *Lingulodinium* sp., *Peridinium* sp.	
		Middle		- 4800 Atlantic	**Sapropel-like muds** *(Kalamitian)*		**Sapropel muds Unit 2**	*Braarudo-sphera bigelovi* *Peridinium trochoidem*	
				- 7800	**Terrigenous-biogenic muds**		**Terrigenous-biogenic muds Unit 3**		7090±180
		Lower		Boreal - 9400 Pre-boreal - 10,200	**Hydrotroilitic muds** *(Neoeuxinian)*	*Stephanodiscus astraea*	**Nanno-fossil-rich terrigenous mud**	Reworked Cretaceous, Paleogene, Neogene coccoliths *Tectatodinium spirifirites*	8600±200
Upper Pleistocene	**Würm (Valdai)**	Upper	Ostashkovian glacial	Younger Dryas Allerød Lower Dryas Bölling Gothiglacial	**Terrigenous brown oxydated muds** **Clayey muds**	Fragments and young forms of *Dreissena rostriformis*, *Monodacna caspia*			13,850±200 16,900±270
				Pomeranian Frankfurtian Brandenburgian	*(Karkinitian)*	*Micromelania caspia*	**Lacustrine phase**		22,000
				- 25,000	*(Tarkhankutian)*	*Cardium edule*	**Marine phase**		25,000
		Middle	M-S. i.g.	Paudorf Arcy Gottweig - 40,000	*(Surozhian)*				40,000
		Lower	K.g.					**Abbreviations**: g. = glacial i.g. = interglacial M-S. i.g. = Mologo-Sheksnian interglacial K.g. = Kalinian glacial *(Between parentheses in italics)* = layers from the inner shelf	
				- ~ 65,000					
	Riss-Würm	Mikulinian interglacial		Eemian - ~ 125,000	*(Karangatian)*				

3. LATE QUATERNARY ENVIRONMENTAL CHANGES IN THE BLACK SEA

The following overview of Late Pleistocene and Holocene environmental changes in the Black Sea is based on sources in the literature as well as the authors' preliminary research results from the western Black Sea shelf and coastal zone, in addition to the Danube Delta.

3.1 Riss II glacial (Lower to Upper Euxinian-Uzunlarian)

The Riss II glacial (Moscovian) resulted in a regressive phase in the Black Sea, which dropped to ~ –100 m (Fedorov 1978) and accumulated freshwater Caspian fauna. During this phase, the Bosphorus Strait is thought to have been exposed to about 100 m.

3.2 Riss-Würm interglacial (Karangatian)

During the Riss-Würm interglacial (Mikulinian), eustatic rise in sea level allowed saline Mediterranean water to penetrate the Pontic basin. This Karangatian phase introduced marine conditions that lasted about 60 ky (from 125 to ~65 ky BP). The Karangatian sea level exceeded that of the present day by 8 to 12 m (Fedorov 1978; Ostrovsky *et al.* 1977; Chepalyga 1984). Connections were established with the Mediterranean Sea through the Bosphorus and Dardanelles, and most likely with the Caspian Sea through the Manych Depression as well. Marine terraces of this age have been found around the Black Sea basin at elevations as high as 4–8 m on the Taman Peninsula and 30–35 m along the Caucasus coast because of neotectonic uplift (Fedorov 1978; Ostrovsky *et al.* 1977). Ingressions into lowland areas, such as the Kolkhida Depression, the Manych Depression, and the Danube Delta, have been reported. Salinity during the Karangatian phase must have been high–ranging from 30 to 37‰ (Nevesskaya 1970)–and comparable to that of the Mediterranean, as Karangatian deposits in the coastal and shelf zones contain marine stenohaline and euryhaline fauna of Mediterranean type. Anoxic deep water with a high hydrogen sulphide content also formed during the Karangatian, together with sapropel deposits in the bathyal zone (Neprochnov 1980; Chepalyga 1984).

3.3 Lower Würm glacial (Post-Karangatian)

The Karangatian highstand was succeeded by the Lower Würm glaciation (Early Valdai, Kalinian), which lowered sea level significantly to between

–100 and –110 m (Ostrovsky *et al.* 1977; Chepalyga 1984). This Post-Karangatian phase–Early Neoeuxinian according to Fedorov (1978)–interrupted the Mediterranean connection again and transformed the sea back to a brackish, semi-fresh condition, completely oxygenated, and with a salinity of 5–10‰. The characteristic fauna were once more of Caspian type.

3.4 Middle Würm interstade (Surozhian)

In the Middle Würm (Middle Valdai, Mologo-Sheksnian interstade, Bryansk interstade), sea level rose very close to that of the present day (between –10 and 0 m). This Surozhian phase occurred between *ca.* 40–25 ky BP (Popov 1983), bringing saline water into the Pontic basin accompanied by marine endemic fauna with Mediterranean associations. According to Ostrovsky *et al.* (1977), salinity was comparable to that of the present Black Sea as well as the previous Karangatian highstand during its maximum transgression. Neprochnov (1980) suggests that the deep waters of the Surozhian basin achieved a high hydrogen sulphide concentration, with sapropel and pyrite formation.

Two subphases of the Surozhian transgression have been reported by Ostrovsky *et al.*(1977). During the earlier subphase, sea level held at almost –10 m, and in the later subphase, it rose to that of the present or slightly higher. Deposits with Surozhian fauna can be found in marine terraces along the Caucasian coast at +15 to +20 m. Elsewhere, they occur at lower elevations: the Kerch Strait, Karkinit Bay and peninsula, offshore on the Gallitzin Rise, in the Shagan lagoon (Shnyukov and Trashchuk 1976; Traschuk and Boltivets 1978), and in the Danube Delta (Panin *et al.* 1983).

On the southern Romanian coast, a Surozhian wave-cut cliff or bench was identified at 12–38 m depth (Figures 2 and 3). Locally, four different terraces appear within this depth interval at about –14, –22, –28, and –38 m (Figure 3A). These erosional, wave-cut features are overlain by several prograding wedges that developed during the sea-level fall after the Surozhian highstand (Figure 3B).

3.5 Upper Würm glacial (Neoeuxinian)

The beginning of the subsequent regression at about 25 ky BP is represented by the Tarkhankut layers, which still contain marine fauna, and the Karkinit horizon, which reveals brackish fauna and few late marine remnants.

The Upper Würm glaciation (Late Valdai, Ostashkovian) corresponds to the Neoeuxinian phase of the Black Sea. This extreme lowstand witnessed a dramatic water-level drawdown to depths variously estimated by researchers: –110 to –130 m (Ostrovsky *et al.* 1977), –90 to –110 m (Chepalyga 1984),

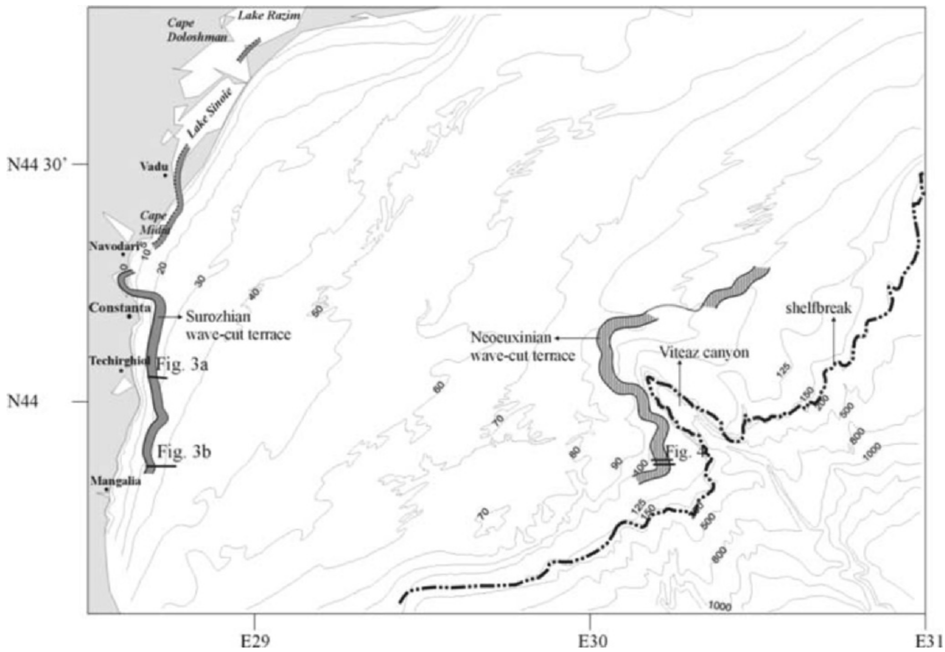

Figure 2. Map of the western continental shelf of the Black Sea (Romanian section) showing bathymetry and location of the shelf break and shorelines during the Neoeuxinian and Surozhian.

around –140 m (Ryan *et al.* 1997), and –100 to –110 m (Demirbağ *et al.*1999; and Görür *et al.* 2001). The shoreline receded far from its present position, especially in the northwest, where much of the wide continental shelf was exposed. Tributary rivers, especially the Paleo-Danube and the Paleo-Dnieper, incised sharply into the newly exposed areas, cutting to a depth of 90 m on the outer shelf.

During the Last Glacial Maximum (~19 to ~16 ky BP), the Neoeuxinian basin was probably completely isolated from the Mediterranean Sea. Water became brackish, then fresh–3 to 7‰ salinity or less according to Nevesskaya (1965)–well-oxygenated, and free of H_2S. The faunal record contains species of brackish to freshwater type, and in the bathyal zone, the microflora consisted of cold-water diatoms dominated by a depleted freshwater complex of *Stephanodiscus astraea* (Zhuze and Mukhina 1980).

At about 16–15 ky BP, post-glacial warming and ice cap melting began. The supply of meltwater from the glaciers to the Pontic basin was direct (via the Dnieper, Dniester, and Danube Rivers) and plentiful, and as a consequence, the Neoeuxinian water level rose very quickly, reaching and surpassing the Bosphorus sill level by ~12 ky BP. It is now generally believed that, during this phase, there was substantial freshwater outflow from the Neoeuxinian lake into

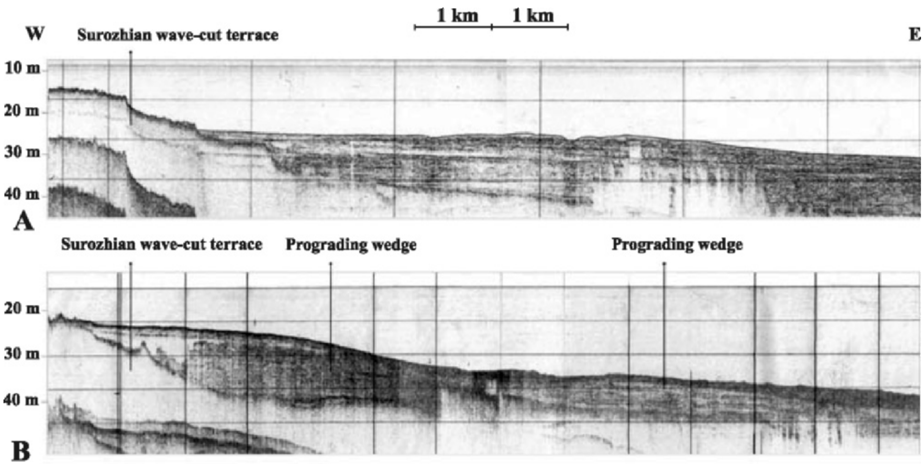

Figure3. Sub-bottom profiles (3.5 kHz) on the proximal shelf showing the Surozhian wave-cut terrace indicated in Figure 2. A (above): four wave-cut Surozhian terraces; B (below): overlying post-Surozhian sediments.

the Aegean Sea through the Bosphorus and Dardanelles Straits. This fresh water discharge has been estimated at about 190 km^3/year (Kvasov 1975).

The Neoeuxinian lowstand is marked on the distal edge of the Romanian shelf by a wave-cut terrace that can be followed for about 100 km along the shelf edge at depths between 98 and 115 m (Figures 2 and 4). This terrace was cut into prograding deposits that are interpreted as a shelf-perched lowstand wedge

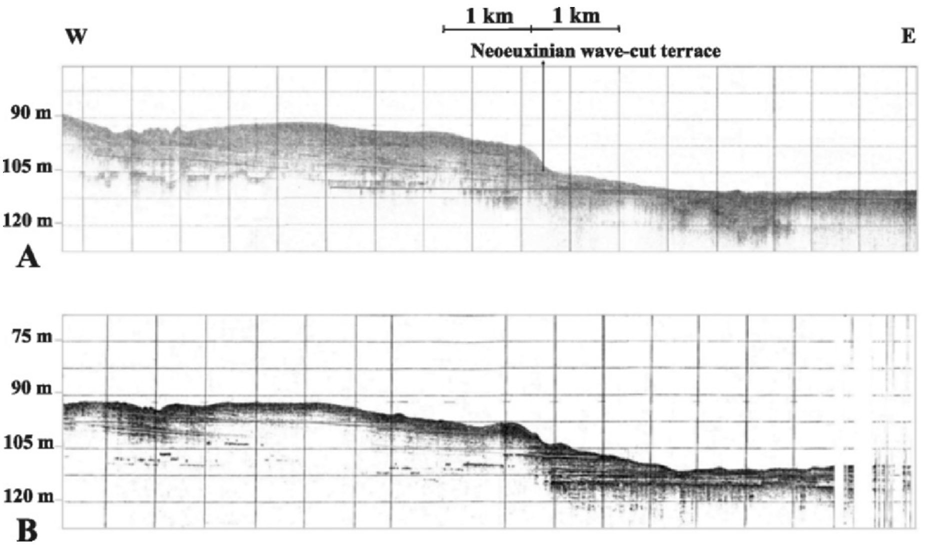

Figure 4. Sub-bottom profiles (3.5 kHz) on the distal shelf showing the Neoeuxinian wave-cut terrace indicated in Figure 2.

formed during a forced regression (Posamentier *et al.* 1992; Berné *et al.* 1998). Our preliminary results indicate that the Neoeuxinian shoreline formed a gulf landward of the Viteaz Canyon. Fluvial incisions on the continental shelf suggest that a large amount of sediment was transported into this gulf, probably controlling the development of the Viteaz Canyon (Winguth *et al.*2000; Popescu *et al.*2001, 2004).

3.6 Holocene

When the Mediterranean and Black Seas reached the same level (close to the present-day situation) between 9 and 7.5 ky BP, a two-way water exchange was established, and the transformation of the Neoeuxinian lake back into an anoxic brackish sea began. The maximum rise in the Black Sea (3–5 m above current level) occurred 4–3.5 ky ago during the Sub-boreal (Fedorov 1978). At this time, the so-called "Old Black Sea" terrace was formed. A rapid lowering of the water level to –5 to –8 m followed, and this phase corresponds to Fedorov's "Phanagorian regression," coeval with the first Greek colonization of the Black Sea coast. A new, short-lived ingression of the sea to a stand of +1 to +3 m then occurred; this transgression has been called the "Nymphaean" by Fedorov (1978), the "Istrian" by Bleahu (1963), and the "Dzhemetinian" by Nevesskaya (1965). By about the 10th century AD, the level of the Black Sea experienced a decline of 1–2 m, then a slow rise, which continues today.

Based on studies of the Caucasian coastal zone, Ostrovsky *et al.*(1977) reconstructed a much more complicated pattern of water-level fluctuations for the Black Sea. In our opinion, however, data from other areas of the Black Sea are not consistent with their results.

4. THE FLOOD HYPOTHESIS

In the late 1990s, a new hypothesis emerged. It proposed that a rapid and catastrophic marine flooding of the Black Sea by Mediterranean water took place at about 7.5 ky BP (Ryan *et al.*1997a, b). According to the authors, the level of the Black Sea was high enough during initial deglaciation for some fresh Pontic water to enter the Aegean Sea, but by about 12 ky BP, retreat of the ice sheet led to a temporary redirecting of meltwater into the North Sea. Deprived of this incoming meltwater during the cool Younger Dryas beginning ~11 ky BP, and under the influences of a drier and windier climate that lasted until 9 ky BP, the Black Sea experienced a new regression to –156 m. At the same time, Mediterranean sea level continued to rise up with the global ocean. This progressive increase finally reached the height of the Bosphorus sill by 7.15 ky BP and broke through, generating a massive torrent of salt water into the Black Sea basin. In the opinion of Ryan and Pitman (1998:234), the input rate was 200 times greater than the falls at Niagara and produced a surge in the level of the Black Sea approaching 30 to 60 cm per day that filled up the basin in a few years. A deeper Bosphorus sill (~ –85 m), however, might have led to an earlier reconnection with the Mediterranean and a different scenario of Black and Mediterranean Sea mixing (Major *et al.* 2002; Ryan *et al.* 2003).

New data collected during two cruises aboard the French *R/V Le Suroît* by the French-Romanian BLASON project seem to support, at least in part, a reconnection of the two seas after 9 ky BP.

This hypothesis is still debated, however. Recent studies conducted on the southern coast of the Black Sea, in the Bosphorus Strait (Algan *et al.* 2001), and in the Marmara Sea have led several scientists to propose that the Black Sea was flowing into the Marmara Sea between 9.5 and 7.2 ky BP. According to Aksu *et al.* (1999, 2002a, b), evidence for this outflow into the Marmara includes the uniform westward direction of cross-stratification seen in sand-prone deposits near the Dardanelles indicating strong westward flow toward the Aegean Sea (Hiscott *et al.*, this volume). It is also seen in the presence of a sapropel layer forming in the Marmara by 7 ky BP, which suggests a clear density stratification in the water column with less saline and less dense Pontic outflow at the surface. Studies of coastal sediments near the Sakarya River sustain a dating of 7.2 ky BP for a relatively high water level in the Black Sea: about –18 m, which was below present sea level yet higher than the Bosphorus sill (Görür *et al.* 2001). The flood hypothesis proposes a –156 m coastline for the Black Sea at this time, immediately prior to the proposed flood.

The sedimentary evolution of the Danube Delta gives strong evidence that argues against both a –156 m lowstand in the Black Sea during the early Holocene and a subsequent catastrophic flooding of the Pontic basin. The main stages of deltaic growth during the Holocene have been identified and dated by corroborating studies in geomorphology, geochemistry, mineralogy, structural and textural analysis, faunal analysis, and ^{14}C dating (Panin *et al.* 1983; Panin 1983, 1989, 1997, 1999). The phases are identified geographically in Figure 5.

If the ^{14}C dating is correct, this model suggests a highstand (very close to present day) by 11.7 ky BP, when the deltaic coastline was represented by the "Letea-Caraorman spit," now located about 25–30 km west of the present shore (Figure 6). From this point onward, no catastrophic event, including a sea-level drop to –156 m, can be recognized within the delta's sedimentary record. The subsequent phases of deltaic growth are fully continuous, and no gaps have been found between the successive stages of lobe progradation.

The main conclusion of the present review is that, in our opinion, more extensive studies are needed to fill the remaining gaps in our knowledge before the uncertainties about Late Pleistocene and Holocene sea-level changes and

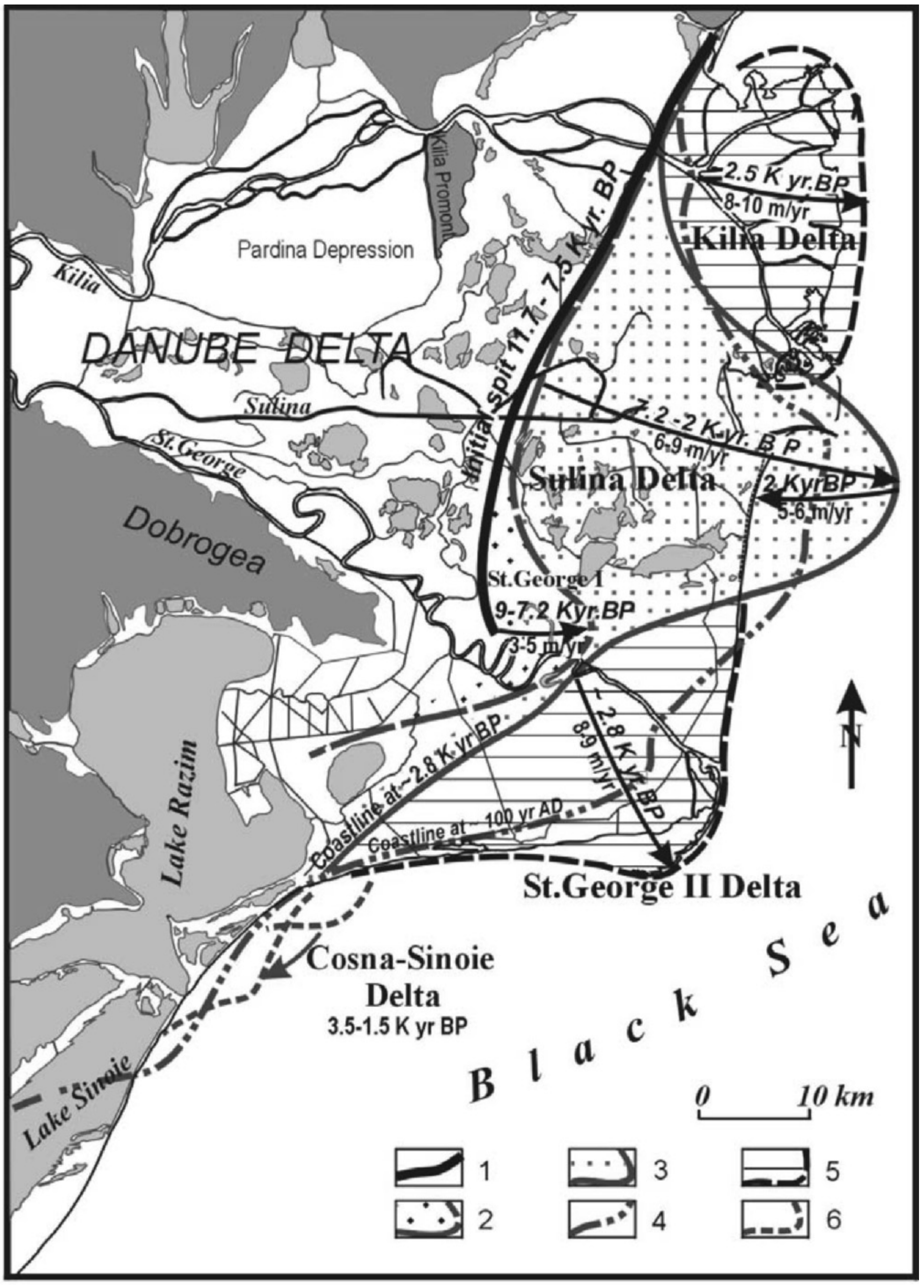

Figure 5. Evolution of the Danube Delta during the Holocene and corresponding coastline changes (after Panin 1997). (1) initial formation of the Letea-Caraorman spit at 11.7–7.5 ky BP; (2) St. George I Delta, 9.0–7.2 ky BP; (3) Sulina Delta, 7.2–2.0 ky BP; (4) coastline at 100 AD; (5) St. George II and Kilia Deltas, ~2.8 ky BP to the present; and (6) Cosna-Sinoie Delta, 3.5–1.5 ky BP.

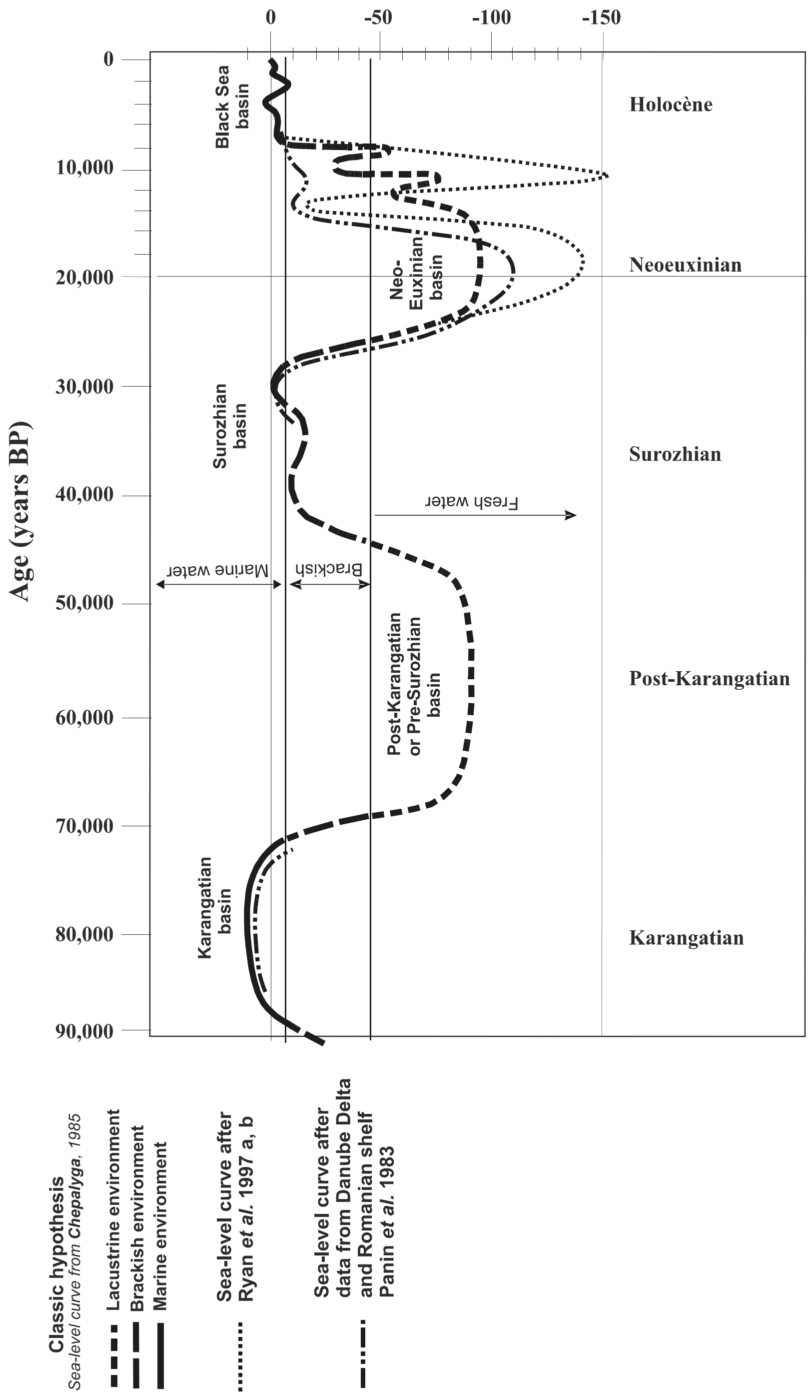

Figure 6. Black Sea level variation according to different authors.

and related coastline migration in the Black Sea can be eliminated and the ideas contained within the existing hypotheses can be decisively evaluated.

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17

SEA-LEVEL FLUCTUATIONS AND COASTLINE MIGRATION IN THE NORTHWESTERN BLACK SEA AREA OVER THE LAST 18 KY BASED ON HIGH-RESOLUTION LITHOLOGICAL-GENETIC ANALYSIS OF SEDIMENT ARCHITECTURE

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Abstract: This paper reconstructs sea-level fluctuations and coastline migration in the northwestern part of the Black Sea for the last 18 ky using high-resolution lithological-genetic analysis of shelf sediments. The analysis includes study of geomechanical and geochemical properties (textures, grain size, C_{org}, $CaCO_3$, pore water salinity and chemical composition, density, bulk density, water content, liquid limit, void ratio, and shear strength). The author's technique allows the rhythmic variations of geomechanical and geochemical properties to be used as a basis for reconstructing oscillating sea-level changes and coastline migration, which is demonstrated in three exemplary cores recovered in Karkinitsky Bay and the Dnieper Paleovalley, both on the Black Sea's northwestern shelf. Sedimentary episodes attributable to Shnitnikov's climatic (humidity) cycles have been revealed during the intervals between calendar years 1415, 551, 336, 286, 214, 169, and 136. Based on the length of these cycles, the mean sedimentation rate in Karkinitsky Bay is estimated as 0.34–0.51 mm/y, and in the Dnieper Paleovalley, 0.29–0.36 mm/y. The coastline migrated from ~ –90 m (*ca.* 18 ky BP) to ~ –30 m (*ca.* 12 ky BP), and then to ~ –20 m (9.5–9.2 ky BP), when the first penetration of Mediterranean water and its biota occurred. The Holocene transgression occurred in an oscillating manner with 'centennial' regression-transgression cycles, each lasting 1600–2000 years. The suggested technique of lithological-genetic analysis and the model of sedimentogenesis provide a basis for forecasting future sea-level changes, coastline migration, and sedimentogenesis on the northwestern shelf.

Keywords: Black Sea, Upper Pleistocene, Holocene, paleogeography, sedimentation cycles

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1. INTRODUCTION

At present, three scenarios have been proposed for the development of the Black Sea since the Last Glacial Maximum (LGM): (1) gradual (Arkhangel'sky and Strakhov 1938; Nevesskaya 1965; Il'ina 1966; Aksu *et al.* 2002); (2) catastrophic (Ryan *et al.* 1997, 2003, 2004; Major *et al.* 2002; Algan *et al.* 2002; Lericolais *et al.* 2004); and (3) oscillating, or non-catastrophically fluctuating (Ostrovsky *et al.* 1977; Fedorov 1978, 1982; Voskoboinikov *et al.* 1983; Tchepalyga 1984; Shnyukov 1985; Balabanov *et al.* 1982; Yanko 1990; Yanko and Gramova 1990; Selivanov 1996a, b; Shilik 1997; Svitoch *et al.* 1998; Chepalyga 2002; Balabanov, this volume; and many others).

Whatever scenario is correct, only shelf sediment architecture can provide insight into sea-level changes and coastline migration caused by low amplitude oscillations. After 10 ky BP, the level of the Black Sea never dropped below the isobath at ~ –40 m, nor did it exhibit a maximum amplitude of fluctuation greater than ~20 m (Yanko-Hombach, this volume).

Sedimentation on continental shelves, which receive globally no less than 50–70% of the total marine sediments, is universally controlled by depth, elevation of sea level, and hydrodynamic regime of the basin, and it varies through geologic time (Nevessky 1967; Shepard 1973; Gershanovich 1979; Sherbakov *et al.* 1979; Shnyukov 1984, 1985). The sedimentary facies structure and the character of deposition are largely defined by the "facies rule" of Valter-Golovkinsky (Lisitsin 1984).

The identification of different sedimentary facies and their depositional environment is based on studies of the most stable indices, which usually include sedimentary composition, structure, texture, and included fossils (Porter and Pettijohn 1963; Shepard 1973; Reineck *et al.* 1981; Selley 1981). These indices, however, are not always sufficient to define the facies environment, and this is the case with the Upper Pleistocene and Holocene muds widely distributed on the northwestern Black Sea shelf (Voskoboinikov *et al.* 1985).

Additional information can be obtained through the analysis of sequential changes in the geomechanical and geochemical properties of sediments (e.g., Manheim and Chan 1974; Bischoff and Sayles 1972; Shishkina 1972; Einsele *et al.* 1974; Levy 1974; Babynets' *et al.* 1981; Konikov 1993; Konikov *et al.* 1999). Previous studies have revealed 'centennial' regression-transgression cycles with a duration of 1600–2000 years that allegedly occurred against the background of the general Holocene sea-level transgression (Shnitnikov 1969; Ivanov and Shmuratko 1983; and others). These cycles are recorded in both sedimentary series architecture and spatial location of ancient coastlines in the Black Sea (Nevessky 1967; Shcherbakov *et al.* 1978; Fedorov 1982; Shnyukov 1982; Voskoboinikov *et al.* 1982a).

The present study focuses on the periodicity and frequencies of geomechanical and geochemical properties within the shelf sediments, which have permitted the identification of various depositional periods and enabled the development of a high-resolution sedimentation model for the northwestern Black Sea shelf. This opens the possibility of establishing a high frequency model for Black Sea level oscillations in the future.

2. MATERIAL

The study area includes the northwestern shelf of the Black Sea, ranging from the Ukrainian part of the Danube Delta to the northwestern coast of Crimea. This area was studied over the course of a large-scale geological survey by various Soviet and ex-USSR geological organizations since the early 1970s. The evidence includes about 200 boreholes, each 20–80 m long with an internal diameter of 127–169 mm; more than 2000 vibrocores up to 4.5 m long with an internal diameter of 80–100 mm; and more than 150 handset sampling sites. The sampling network is shown in Shnyukov (1984, 1985). Main results of the study are included in the database of the Laboratory for Engineering Geology of Sea Shores, Reservoirs, and Mountain Slopes of Odessa National University.

From the total number of cores, uninterrupted sediment columns have been chosen from 16 key sites. In the present work, these sites are exemplified by three cores located in diverse environmental settings (Figure 1):

(1) core 1136 (northwestern part of Karkinitsky Bay): 32°25'07" E, 45°35'02" N; water depth –30.5 m, length 377.5 m,

(2) core 2-82 (Dnieper Paleovalley): 30°47'20" E, 46°21'55" N; water depth –24.5 m; length 310 m, and

(3) core 37A-82 (Dnieper Paleovalley): 31°21'45" E, 46°17'30" N; water depth –19.6 m; length 307 m).

3. METHODS

Sampling was performed as follows. Geomechanical and geochemical properties of the sediments were studied continuously in each 2–2.5 cm of the core. Sediment structures were studied in samples that were 10 cm in length along the core column and 1 cm in width.

Stratigraphy of the sediment columns was based on taxonomic analysis, vertical and spatial distribution of molluscs (Nevesskaya 1965), and sometimes foraminifera (Voskoboinikov *et al.*1985; Yanko-Hombach, this volume), all in conjunction with a lithological-genetic analysis of the sediments. The stratigraphic scheme of Arkhangel'sky and Strakhov (1938) and Fedorov (1982) was applied.

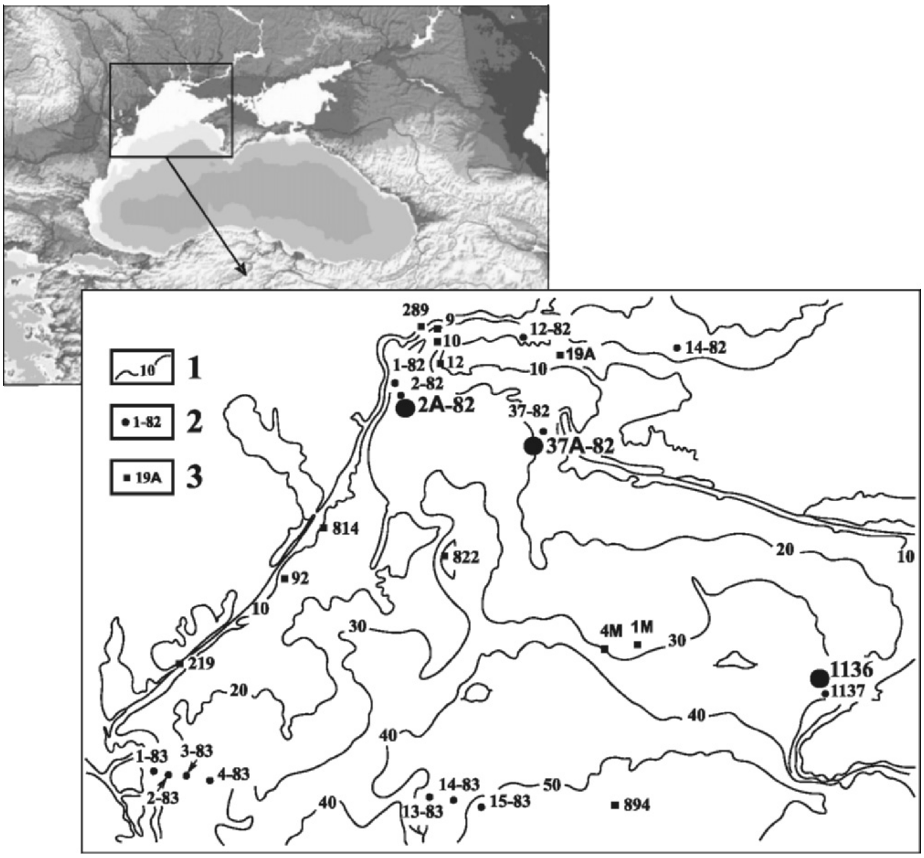

Figure 1. Study area on the northwestern shelf of the Black Sea. **Key:** 1 – isobath, 2 – vibrocores, 3 – boreholes dated by conventional radiocarbon method.

Grain-size study of the sediments was performed by wet sieving and pipette analysis as described in Inman (1952), Folk (1954), and Rukhin (1969).

Organic content (C_{org}) was identified by gas chromatography (Degens and Reuter 1964; Artem'ev 1976).

Clay minerals (<0.005 mm fraction) were identified by diffractometers URS-50IM and DRON-1.5, as described in Biscay (1964) and Butuzov *et al.* (1975) with no preliminary removal of organic matter and carbonates.

Carbonate content ($CaCO_3$) was identified through hydrochloric acid treatment of 1 g samples (Babynets' *et al.* 1981).

Pore water salinity and chemical composition were studied by chemical-analytical methods as described in Shishkina (1972) and Manheim *et al.* (1972).

Geomechanical properties, including density, bulk density, water content or moisture, liquid limit, void ratio, and shear strength (for slightly larger samples) were identified both at the ship's and stationary laboratories using

using techniques described by Akal (1972), A.W. Skempton (1953, 1970), and Babynets' *et al.* (1981).

Sediment texture was studied by X-ray analysis (Hamblin 1962, 1965; Voskoboinikov *et al.* 1983; Shnyukov 1984, 1985). In the Russian literature, sediment *texture* refers to the spatial distribution of constituent particles and their arrangement relative to each other (e.g., Rukhin 1969). In the English literature, the Russian definition of texture corresponds largely to what is meant by the term *structure* (e.g., Leet 1982). Textures identified on X-ray images were digitized (Figure 2) for further statistical treatment using Markov chains, a sequence of random values whose probabilities at a time interval depend upon the value of the number during the previous time (Krumbein and Graybill 1965).

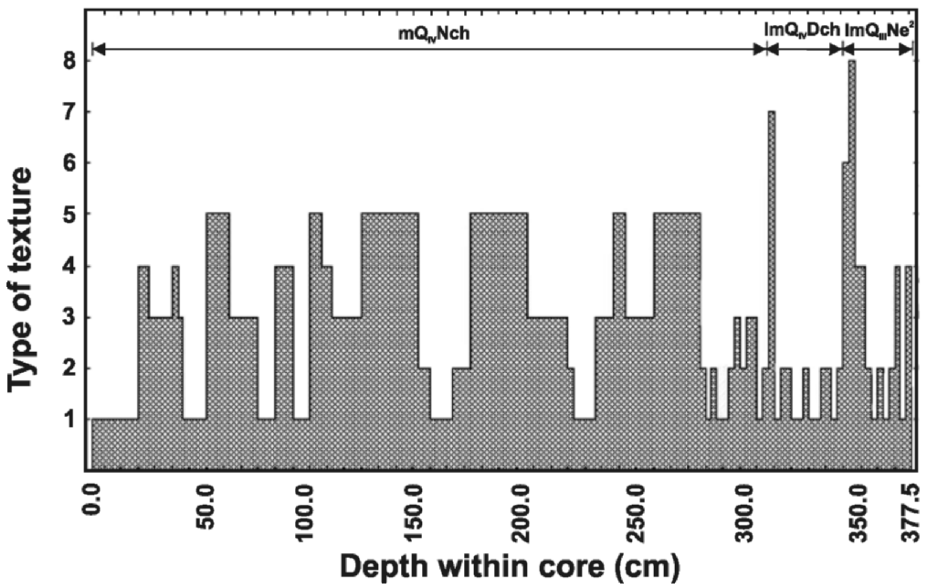

Figure 2. Types of textures (scaled numerically) in core 1136. **Key:** 1 – massive; 2 – bioclastic; 3 – horizontally-bedded; 4 – obliquely-bedded; 5 – sinuously-bedded; 6 – bioturbational; 7 – erosional, and 8 – peat; Ne – Neoeuxinian; Dch – Drevnechernomorian (Old Black Sea); Nch – Novochernomorian (New Black Sea); m – marine; lm – liman.

Statistical processing of the recovered data included trend, time-series, and spectral analysis (Box and Jenkins 1970; Kominz *et al.* 1979; Pisias *et al.* 1984), correlation, and regression (from a "Statistics" package). Statistically homogenous entities within the digital sequences have been identified using Rodionov's criterion (Rodionov 1968) and Markov chains (Krumbein and Graybill 1965). Using software developed in the Laboratory for Engineering Geology of Sea Shores, Reservoirs, and Mountain Slopes of Odessa National

University (Konikov *et al.*2002), matrices of transitive probabilities have been created, enabling the discovery of cyclical variations in the studied characteristics along the length of the cores.

4. RESULTS

4.1 Stratigraphy and Lithology

The sediment columns were divided into Upper Pleistocene and Holocene sections. The former represents Neoeuxinian beds, while the latter contains Drevnechernomorian (Old Black Sea) and Novochernomorian (New Black Sea) beds (Figures 3, 4, and 5).

No radiocarbon dates were obtained for these particular cores, however, ^{14}C records are quite abundant for many other cores in the area (e.g., Balabanov's Appendix 1, this volume). Upper Neoeuxinian beds on the Golitsyn Uplift inside the Dnieper Paleovalley (core 12, water depth –32 m; core 822, water depth –34.8 m; Figure 1) have yielded an uncorrected age of 12.7–10.2 ky BP (Balandin and Mel'nik 1987). The radiocarbon age of the Drevnechernomorian beds ranges from 9.8 ky BP (Grigor'ev *et al.*1984) to 8.9 ky BP (Balandin and Mel'nik 1987; Shnyukov 1984, 1985).

The Upper Neoeuxinian beds contain stiff clays (core 1136, Figure 3) or light gray silty muds (core 2-82, Figure 5), as well as thin layers of coquina consisting of entire shells and/or fragments of Caspian brackish water molluscs *Dreissena polymorpha* (Pallas), *D. rostriformis* (Desh.), and rare *Micromelania caspia lincta*Milashewitsch. The Upper Neoeuxinian deposits cover most of the northwestern Black Sea shelf up to the –20 m isobath, reflecting a highstand of the Neoeuxinian lake at around 10 ky BP (Figure 6).

Upper Neoeuxinian sediments were deposited during a warmer climate than that of the LGM as indicated by the shift from pine to broad-leaved forests (Komarov *et al.* 1979; Kvavadze and Dzheiranshvili 1989). Their thickness varies up to 25 m. In some places (e.g., the western part of the Golitsin Uplift, located at the mouth of Karkinitsky Bay), they are exposed on the sea floor (Ischenko 1974). Lithologically, they are rather monotonous and fill pre-Neoeuxinian depressions and paleoriver valleys (Shnyukov 1985).

In many cores recovered above the –90 m isobath, the Upper Neoeuxinian beds overlap unconformably the subaerial loams and aquatic sediments containing freshwater *Viviparus* sp., ostracoda *Candona* and *Candoniella*, and foraminifera *Ammonia novoeuxinica* (Yanko-Hombach, this volume) indicating a transformation of the bottom from an erosional to a subaquatic accumulative phase (Gozhik 1984; Shnyukov 1985). In many cores, the Upper Neoeuxinian beds are overlapped by peats and/or very coarse sedi-

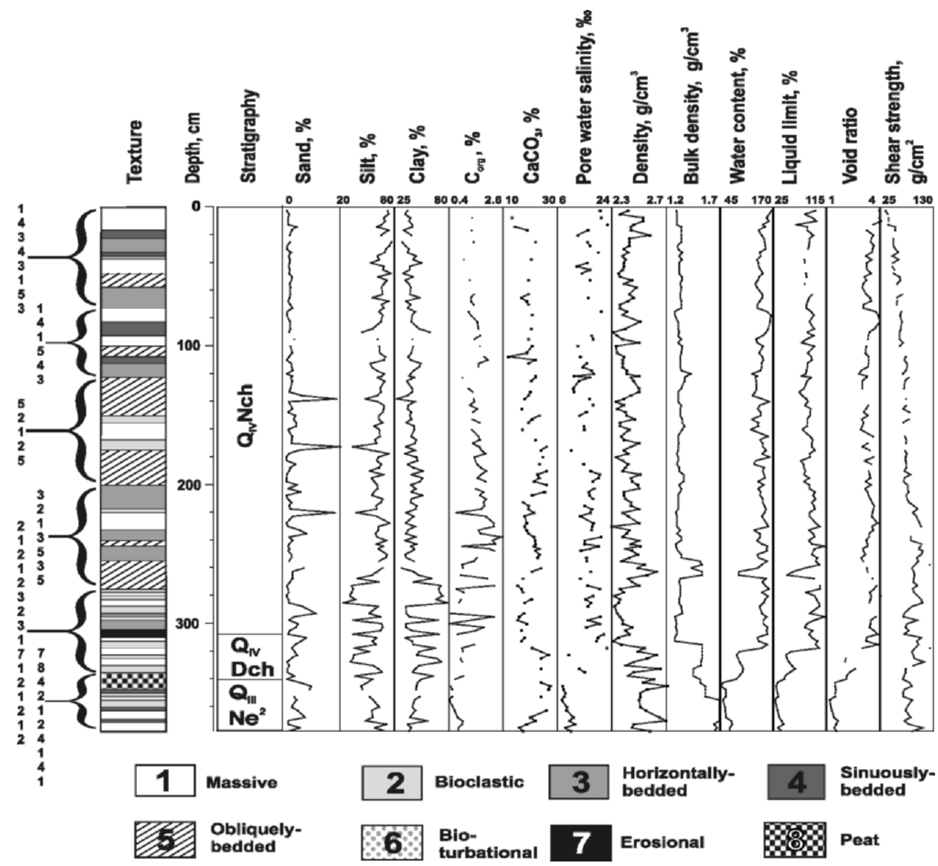

Figure 3. Geomechanical and geochemical sediment properties in core 1136. **Key:** Ne – Neoeuxinian; Dch – Drevnechernomorian (Old Black Sea); Nch – Novochernomorian (New Black Sea).

ments formed at the end of the Younger Dryas (*ca.* 10.2 ky BP) when the level of the lake dropped to about –55 m (Yanko-Hombach, this volume).

The Drevnechernomorian beds contain dark gray and greenish-gray silty and clayey muds with thin (2–7 mm) laminae of sand as well as thicker (0.5–15 cm) layers of mollusc shell coquina containing a mixture of Caspian brackish water species–*D. polymorpha*, *Hydrobia ventrosa* (Montagu), *Monodacna caspia* (Eichwald) (in core 2-82), *Abra ovata* (Philippi)–and Mediterranean euryhaline ones–*Cardium edule* Linné.

The Novochernomorian beds reveal dark gray mud with thin (3–7 mm) sandy and silty beds alternating with coquina layers. The latter are enriched with either intact or fragmented shells of Mediterranean molluscs *C. edule* Linné, *Mytilus galloprovincialis* (Lamarck), *Nassarius reticulatus* (Linné), *Bittium reticulatum* (Costa), and *Chione gallina* (Linné).

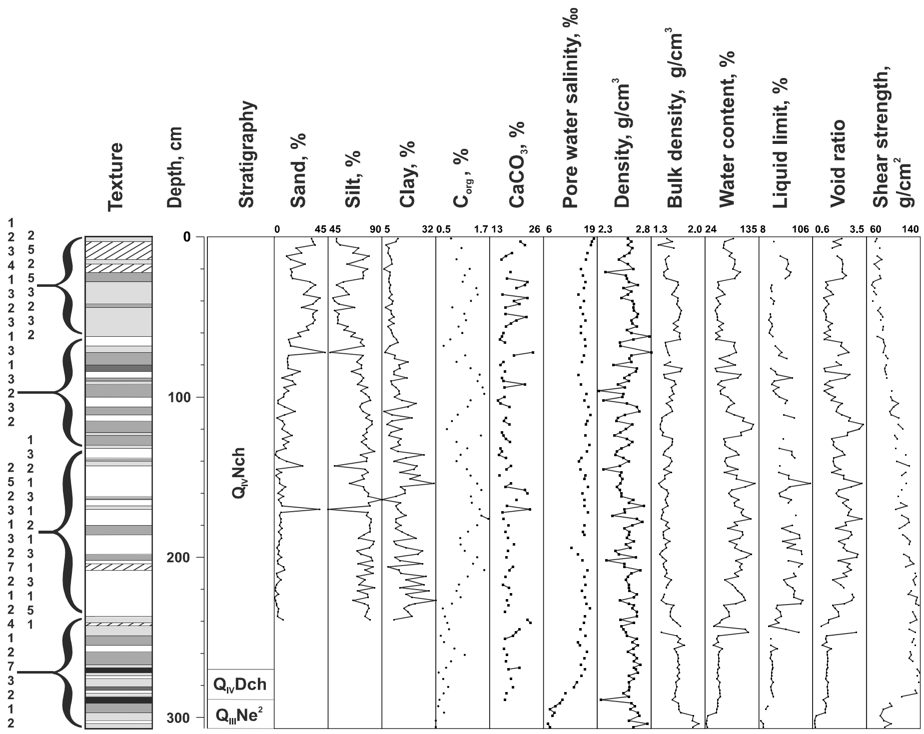

Figure 4. Geomechanical and geochemical sediment properties in core 37A-82. For key, see Figure 3.

The Upper Neoeuxinian-Drevnechernomorian boundary is fixed by the first appearance of *C. edule*, which marks the initial Holocene connection with the Sea of Marmara around 9.5 ky BP and recolonization of the Neoeuxinian lake by Mediterranean organisms. This boundary is present at a depth of 340 cm (from the bottom) in core 1136 (Figure 3), 295 cm in core 2-82 (Figure 4), and 298 cm in core 37A-82 (Figure 5).

The boundary between the Drevnechernomorian and Novochernomorian beds is fixed by the appearance of a wide distribution of Mediterranean molluscs that continues upwards. It lies at a depth of 310 cm in core 1136 (Figure 3), 270 cm in core 37A-82 (Figure 4), and 250 cm in core 2-82 (Figure 5).

The sediments in most of the cores are dominated by a <0.01 mm fraction (Figure 3 and 5), however, in some cores, the sandy fraction becomes more abundant (e.g., core 37A-82, Figure 4).

4.2 Geomechanical and Geochemical Characteristics

Massive, horizontally-bedded, sinuously-bedded, obliquely-bedded, bioturbational, and erosional textures were observed. The thickness of the vari-

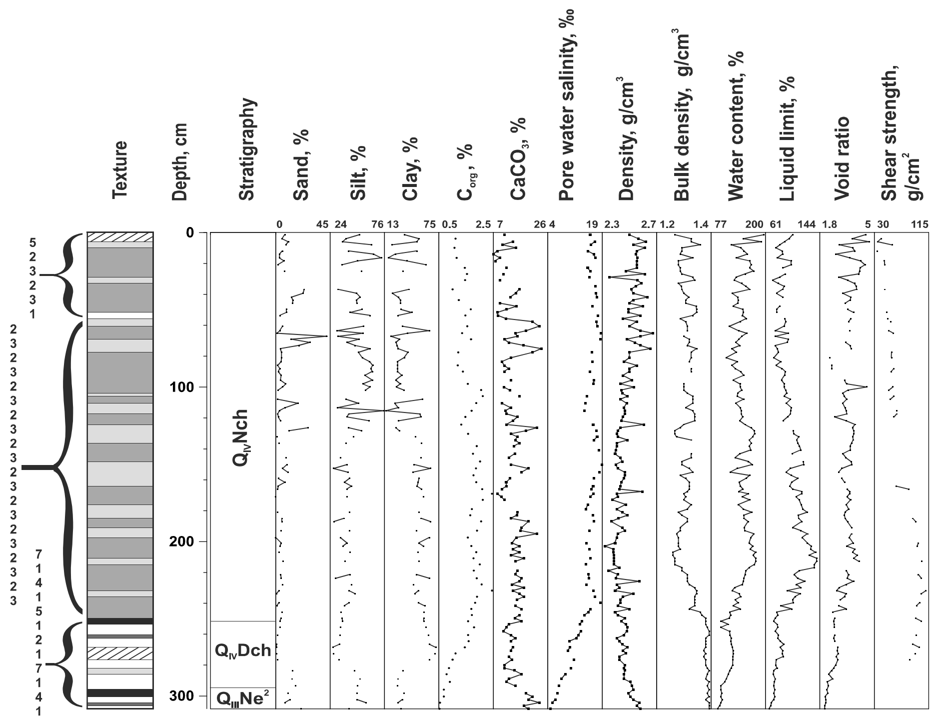

Figure 5. Geomechanical and geochemical sediment properties in core 2-82. For key, see Figure 3.

ously textured layers varied from 2–5 to 20–35 cm (Figures 3, 4, and 5).

The dominant textures found within Upper Neoeuxinian deposits were massive, horizontally-bedded, and bioclastic. An erosional texture was present at the Upper Neoeuxinian-Drevnechernomorian and the Drevnechernomorian-Novochernomorian boundaries. Drevnechernomorian beds were characterized by a mixture of texture types. Novochernomorian beds showed predominantly obliquely-bedded and sinuously-bedded textures (Figures 3, 4, and 5).

Each of the cores described here contained predominantly silty (30–70%) and clayey (10–40%) grain-size fractions. In cores 1136 and 2-82, the sandy fraction did not exceed 2–5%, with the exception of several intervals where it reached 20% (Figure 3) or even 45% (Figure 5). Core 37A-82 contained an elevated sandy fraction due to its location in shallower water and within closer proximity to the Tendra sand-spit. As a rule, the percentage of the sandy fraction increases in the sandy-coquina layers (Figures 3, 4, and 5).

C_{org} values did not exceed 0.45% in the Upper Neoeuxinian sediments. It was significantly higher (up to 2.7%) at the top of the Drevnechernomorian beds (Figure 3).

Mineralogical composition of the clayey fraction from deposits in Karkinitsky Bay and the Dnieper Paleovalley was fairly similar, with a domi-

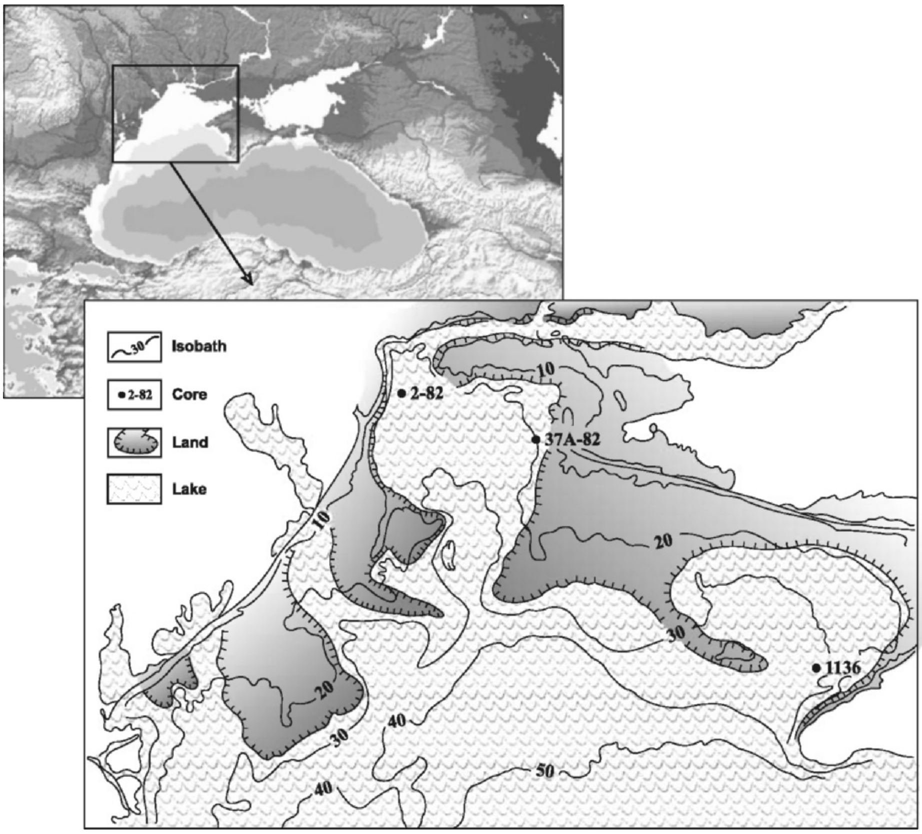

Figure 6. The northwestern shelf during the Late Neoeuxinian transgression (~13 ky BP).

nance of mixed-layer clay (up to 30%) and hydro-micaceous minerals (up to 70%), and the occurrence of kaolinite (up to 5–10%), chlorite (up to 5–15%), and (rarely) montmorillonite (up to 5%) (Voskoboinikov *et al.* 1983; Shnyukov 1985).

$CaCO_3$ content varied through a wide range between 5% and 40%. The higher $CaCO_3$ values, reaching 20–40%, usually correlated with increased shell content in the coarser fractions and were typical for Drevnechernomorian sediments (26–40%).

Pore water salinity varied from 4.6 to 23.4‰. In the Upper Neoeuxinian beds, it ranged between 4.6 and 11.5‰; in the Drevnechernomorian beds, it varied from 9 to 15‰; and in the Novochernomorian beds, it revealed values between 11 and 23.4‰.

The lowest water content (moisture) was found in Upper Neoeuxinian sediments. It dramatically increased upward in the Drevnechernomorian, and especially in the Novochernomorian beds (Figures 3, 4, and 5).

The highest sediment density was obtained for the Upper Neoeuxinian and the bottom part of the Drevnechernomorian beds. It decreased sharply in the Novochernomorian beds (Figure 7).

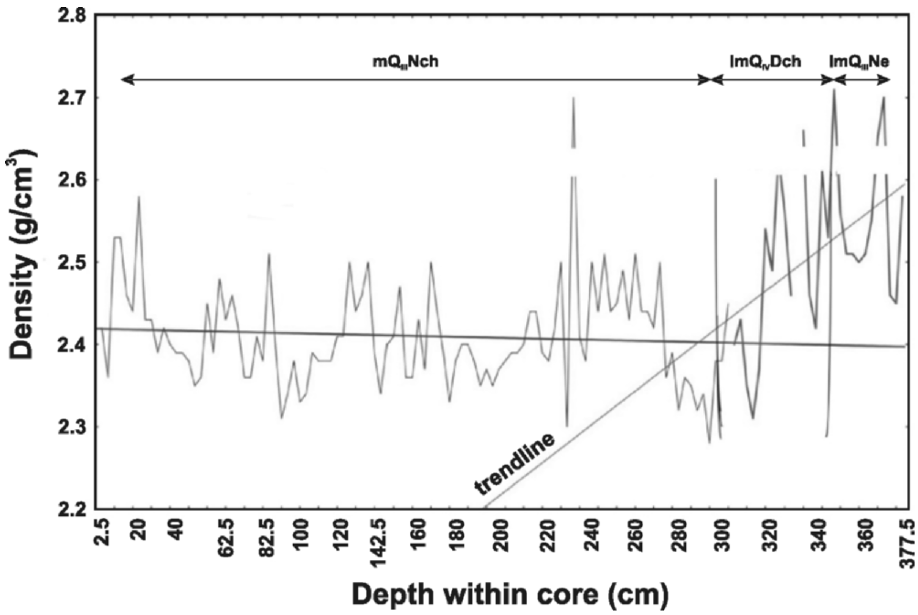

Figure 7. Sharp decrease of bulk density upwards within core 1136.

The liquid limit and void ratio showed similar changes with core depth. They increased significantly from the Upper Neoeuxinian-Drevnechernomorian boundary upwards (Figures 3, 4, and 5).

Shear strength values showed a gradual decrease upwards within the cores to a low value in the Novochernomorian beds.

4.3 Statistical Processing

Correlation analysis (Tables 1, 2, and 3) of recovered data allowed the discovery of regular trends (positive or negative) and irregular patterns in the sequential distribution of geomechanical and geochemical properties along the length of the cores. For example, the density of the Novochernomorian deposits reveals no distinct pattern (i.e., the absence of any discernible trend), while both the Drevnechernomorian and Upper Neoeuxinian sediments show a distinct trend, which is particularly apparent in core 1136 (Figure 7).

Similar patterns were discovered using other characteristics, e.g., water content, shear strength, etc. (Tables 1, 2, and 3). Parameters having the highest correlation coefficient were used as the most informative indices.

Table 1. Correlation matrix of geomechanical and geochemical parameters in core 1136.

Parameters	Depth in core	Texture	Density	Bulk density	Water content	Liquid limit	Shear strength	C_{org}	$CaCO_3$	Sand	Silt	Clay	Void ratio	Porosity	Pore water salinity
Depth in core	1.00	0.04	0.40*	0.62*	–0.59*	–0.49*	0.55*	–0.36*	0.18	0.20	–0.37*	0.26	–0.64*	–0.65*	–0.42*
Texture		1.00	–0.15	–0.09	0.14	0.07	0.10	0.20	0.01	–0.21	0.02	0.09	0.10	0.11	0.13
Density			1.00	0.69*	–0.68*	–0.65*	–0.03	–0.38*	0.26	0.16	0.01	–0.09	–0.64*	–0.66*	–0.44*
Bulk density				1.00	–0.92*	–0.88*	–0.01	–0.58*	0.21	0.21	–0.08	–0.03	–0.94*	–0.95*	–0.66*
Water content					1.00	0.96*	0.12	0.59*	–0.32	–0.34	0.03	0.14	0.96*	0.97*	0.78*
Liquid limit						1.00	0.21	0.57*	–0.32	–0.35*	–0.02	0.20	0.92*	0.93*	0.78*
Shear strength							1.00	0.09	–0.04	–0.03	–0.33*	0.35*	0.00	0.02	0.06
C_{org}								1.00	–0.13	–0.42*	0.31*	–0.10	0.59*	0.60*	0.35*
$CaCO_3$									1.00	0.31*	–0.01	–0.15	–0.35*	–0.31*	–0.28
Sand										1.00	–0.24	–0.26	–0.29*	–0.30*	–0.32*
Silt											1.00	–0.87*	0.12	0.12	0.11
Clay												1.00	0.02	0.04	0.05
Void ratio													1.00	0.99*	0.78*
Porosity														1.00	0.78*
Pore water salinity															1.00

* - statistically significant

Table 2. Correlation matrix of geomechanical and geochemical parameters in core 2-82.

Parameters	Depth in core	Texture	Density	Bulk density	Water content	Liquid limit	Shear strength	C_{org}	$CaCO_3$	Sand	Silt	Clay	Void ratio	Porosity	Pore water salinity
Depth in core	1.00	–0.11	0.12	0.73*	–0.65	–0.66	0.67*	–0.46	0.28	–0.37	–0.74*	0.77*	–0.76*	–0.75*	–0.58
Types of texture		1.00	–0.72	–0.47*	0.38	0.34	–0.05	0.13	0.12	0.35	–0.07	0.03	0.38	0.39	0.44
Density			1.00	0.6	–0.47	–0.62	–0.35	–0.56	–0.56	–0.73*	0.29	–0.22	–0.39	–0.38	–0.78*
Bulk density				1.00	–0.87*	–0.95*	0.18	–0.66	–0.14	–0.65	–0.2	0.26	–0.92*	–0.91*	–0.81*
Water content					1.00	0.93*	–0.37	0.4	0.05	0.42	0.11	–0.15	0.95*	0.96*	0.55
Liquid limit						1.00	–0.13	0.65	0.23	0.63	0.09	–0.15	0.91*	0.90*	0.76*
Shear strength							1.00	0.29	0.56	0.15	–0.71*	0.69*	–0.45	–0.45	0.11
C_{org}								1.00	0.07	0.44	0.12	–0.16	0.38	0.37	0.88*
$CaCO_3$									1.00	0.72*	–0.54	0.46	0.01	–0.01	0.2
Sand										1.00	0.02	–0.12	0.48	0.45	0.70*
Silt											1.00	–0.99*	0.23	0.23	0.15
Clay												1.00	–0.28	–0.28	–0.22
Void ratio													1.00	0.99*	0.56
Porosity														1.00	0.54
Pore water salinity															1.00

* - statistically significant

Table 3. Correlation matrix of geomechanical and geochemical parameters in core 37A-82.

Parameters	Depth in core	Texture	Density	Bulk density	Water content	Liquid limit	C_{org}	$CaCO_3$	Sand	Silt	Clay	Void ratio	Pore water salinity
Depth in core	1.00	–0.55*	–0.04	–0.69*	0.74*	0.78*	0.18	–0.35	–0.80*	0.56*	0.86*	0.76*	0.23
Texture		1.00	0.13	–0.00	–0.05	–0.23	–0.43	0.21	0.28	–0.02	–0.58*	–0.00	–0.10
Density			1.00	0.33	–0.36	–0.41	–0.19	0.48	0.46	–0.49	–0.24	–0.15	–0.46
Bulk density				1.00	–0.98*	–0.94*	–0.08	0.39	0.86*	–0.78*	–0.67*	–0.97*	–0.51
Water content					1.00	0.96*	0.11	–0.47	–0.90*	0.83*	0.67*	0.97*	0.44
Liquid limit						1.00	0.15	–0.52	–0.90*	0.78*	0.75*	0.92*	0.47
C_{org}							1.00	0.34	–0.05	0.03	0.06	0.08	0.15
$CaCO_3$								1.00	0.64*	–0.60*	–0.47	–0.35	–0.01
Sand									1.00	–0.91*	–0.76*	–0.83*	–0.31
Silt										1.00	0.43	0.75*	0.16
Clay											1.00	0.65*	0.43
Void ratio												1.00	0.38
Pore water salinity													1.00

* - statistically significant

Based on spectral analysis, we constructed spectrograms indicating the statistically valid periods for each characteristic. In the present study, we focused on the cyclicity of sedimentation, choosing to illustrate sedimentation rhythms using pairs of characteristics (Figures 8 and 9).

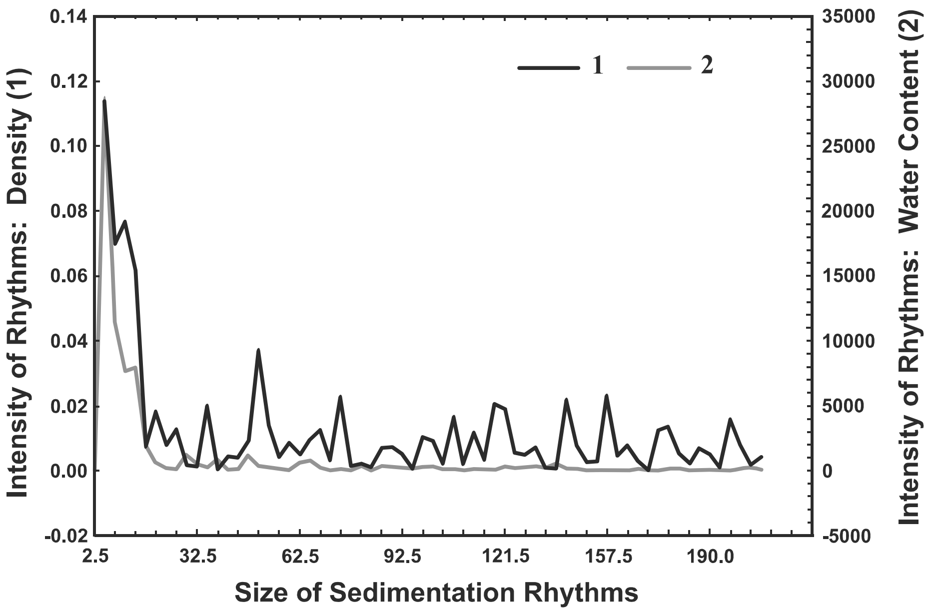

Figure 8. Periodograms of density (1) and water content (2) variability in core 1136.

Based on conjugate analysis, statistically significant periods (rhythms) were identified for all studied characteristics (Konikov *et al.* 2002). Taking into account the sampling intervals, the following depositional segments along the core lengths have been identified:

core 1136	core 37A-82	core 2-82
8–10 cm	5–7 cm 9–11 cm	6–8 cm
13–15 cm	14–16 cm	14–16 cm
22–25 cm		22–23 cm
51–54 cm	49–51 cm	51–59 cm
81–83 cm		85–88 cm

Metrical variation from core to core is due to differences in the thickness of the Novochernomorian sediments in the three sample locations. Statistically, this neither jeopardizes estimation of the magnitude of the cycles and their plausibility, nor prevents their correlation (Box and Jenkins 1970).

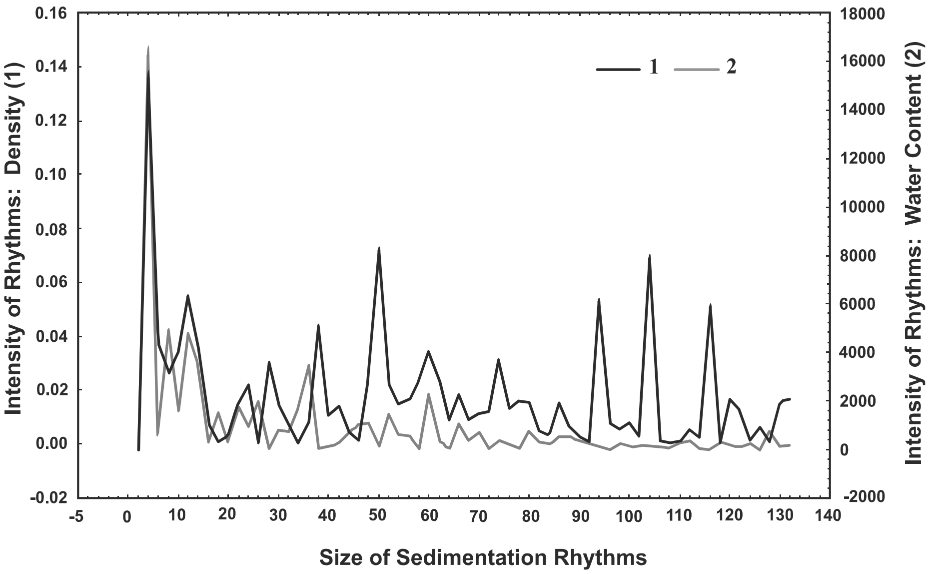

Figure 9. Periodograms of density (1) and water content (2) variability in core 37A-82.

For a visual presentation of rhythmicity, we carried out a frequency smoothing for density values using the BARTLETT function. The resulting graphs (Figures 10 and 11) clearly demonstrate the interrelation of rhythms of various frequencies.

Using Rodionov's criterion, the statistically homogenous elements were identified and their numerical characteristics established (Konikov 1993, 1995). The cyclical occurrence of homogeneous elements and the structure of cycles were estimated based on analysis of the matrices of transitional probabilities (Krumbein and Graybill 1965).

5. DISCUSSION

5.1 Sedimentation Model

Sediments with massive textures are characterized by high dispersibility and poor sorting. They are usually formed under moderate hydrodynamic activity and remoteness from sources of clastic material (Reineck *et al.* 1981; Selley 1981; Voskoboinikov *et al.* 1983).

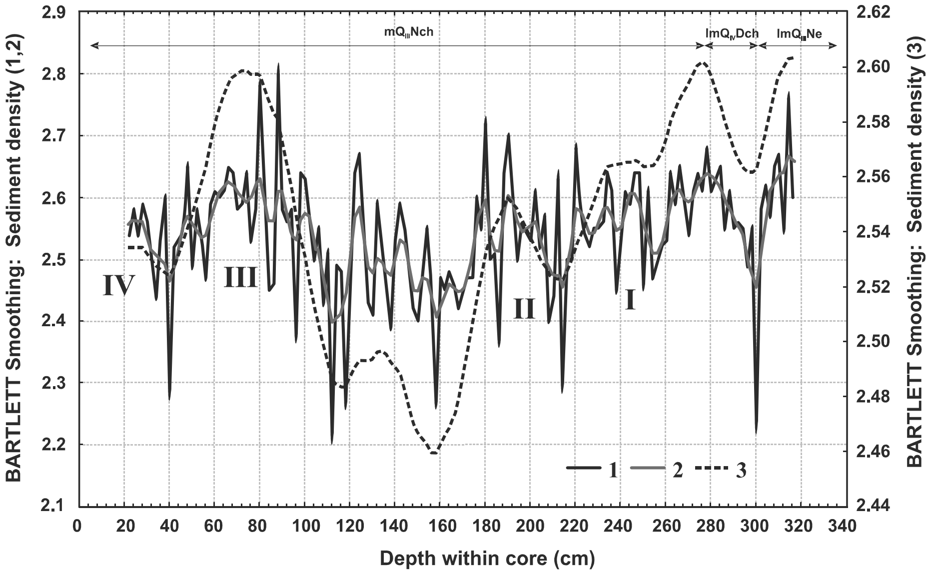

Figure 10. BARTLETT smoothed curve of sediment density in core 1136. Smoothing: 1 – based on 5 quantities (rhythm size is 13 cm); 2 – based on 3 quantities (rhythm size is 8 cm); 3 – based on 9 quantities (rhythm size is 22.5 cm); I–IV = transgressive-regressive cycles.

Bedded textures are characterized by the alternation of clay (about 0.5–3 mm thick) and siltstone-sand (about 2–7 mm thick) microlayers, and they reflect a sedimentation of pulsating character. As a rule, they are formed in rather shallow basins under frequent sea-level fluctuations of small amplitude.

Oblique bedded textures are usually formed by variations in hydrodynamic regime (e.g., currents). Sinuous textures are formed as a result of wave influence on the bottom.

Bioclastic textures are largely represented by coquina with oozy filler. The resting position of shells enables one to reconstruct the paleohydrodynamic conditions in the basin. In cores 1136 and 2-82, the horizontal orientation of the shells reflects a quiet sedimentological regime. In core 37A-82, the bioclastic material in some streaks is represented mainly by the debris of bowls with traces of wave processing. Apparently, these streaks were accumulated during shallow sea conditions near the coastline.

Within the cores, strata of coquina are often superimposed upon sediments of massive texture, obviously testifying to a deepening of water depth. Such a combination of textures occurs principally within Novochernomorian and Drevnechernomorian sediments and is typical of sedimentation under shallow

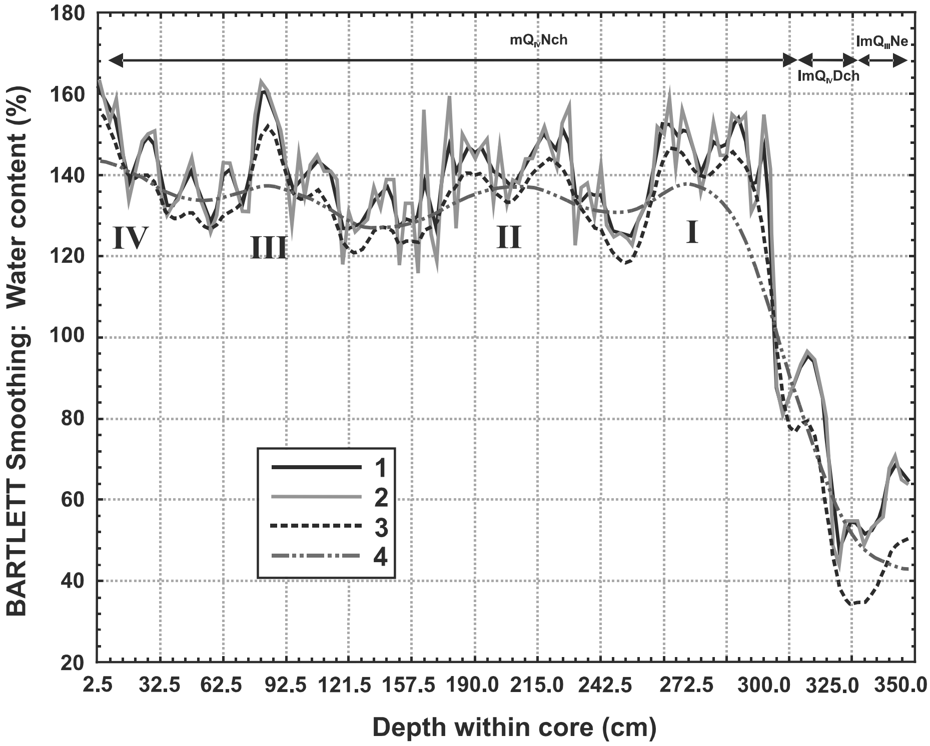

Figure 11. BARTLETT smoothed curve of water content (moisture) in core 1136. Smoothing: 1 – based on 5 quantities (rhythm size is 13 cm); 2 – based on 3 quantities (rhythm size is 8 cm); 3 – based on 9 quantities (rhythm size is 22.5 cm); 4 – based on 33 quantities (rhythm size is 82.5 cm); I–IV = transgressive-regressive cycles.

lagoon conditions (Reineck *et al.* 1981; Selley 1981; Voskoboinikov *et al.* 1983; Shnyukov 1985).

In average Novochernomorian sections, bedded textures prevail, testifying to increased hydrodynamic activity within the basin. Such conditions date to a drop in the Black Sea that we have named the Khadzhibeian regression (Voskoboinikov *et al.* 1982b; Shnyukov 1985).

Each of the selected texture types is repeated many times in the columns, reflecting a cyclical character. This is demonstrated by the results of Markov analysis on these sequences (Konikov 1993, 1995), which indicate that there are two types of two-element cycles: massive-bioclastic and horizontal stratified-oblique stratified. In core 1136, the first cycle repeats itself seven times, and second cycle repeats itself 16 times.

This finding confirms the rhythmic character of the sedimentation process on the northwestern shelf, exclusive of wave erosion and biomorphic texture (streaks of peat). Traces of wave erosion on beds at the Drevnechernomorian-Novochernomorian boundary indicate that sea level was lower then by

28–30 m than that of today. The depth of Karkinitsky lagoon at that time was probably about 3–5 m.

The stratum of peat in the thicker Upper Neoeuxinian deposits at a depth of 370 cm from the bottom testifies to the stagnation of sea level and the likely formation of deposits under marsh conditions. Karkinitsky Bay then represented a shallow, half-closed, low-salinity lagoon, and this corresponds with Nevessky's data (1967). It is important that the shallow depths of the Upper Neoeuxinian and Drevnechernomorian basins did not interfere with the accumulation of oozy deposits in Karkinitsky Bay and the Dnieper paleolagoon, as well as other shelf sites. At the present time, the formation of muds occurs in depths less than 20 m on the shelf, in limans, and in lagoons.

The salinity and chemical composition of sediment pore water reflect the immediate hydrochemical conditions of sedimentation, that is, they characterize the mineralization and chemical composition of water within the sedimentation basin (Bruevich 1952; Valyashko 1971; Shishkina 1972; Konikov 1993). However, in the course of diagenetic transformation of the sediments, the salinity of pore waters can be slightly increased as a result of pore water interacting with the deposits. The chemical composition can change, too, but only in a certain direction (Bischoff and Sayles 1972; Shishkina 1972; Strakhov 1972; Levy 1974; Konikov 1993).

At the end of the Late Neoeuxinian in Karkinitsky and the Dnieper paleolagoons (Figure 1), sedimentation proceeded under brackish water conditions. The beginning of the Drevnechernomorian is connected with a sharp increase in water salinity up to 15–20‰, which is obviously related to intensive influx of saltwater from the Sea of Marmara (Konikov 1993).

For the last 7500 years, water salinity in lagoons and limans has changed significantly (from 15 up to 26‰). We associate increased salinity with the Holocene regressions. Declining water level and temporary isolation of lagoons and limans resulted in increasing water salinity, which changed its composition. Such a process is observed in some modern limans of the northern Black Sea and northern Crimea (Voskoboinikov *et al.* 1982a; Shnyukov 1984; Konikov 1993).

The appearance of a trend in the distribution of magnitudes of geomechanical and geochemical properties characterizes the process of diagenetic modifications of deposits as a whole. The sharp changes in shear resistance and other properties at the Drevnechernomorian-Novochernomorian boundary can be interpreted as indirect indications of stratigraphic subdivision.

Pulsating variations in grain-size composition and other characteristics, seen against the background trend, testify to the pulsating nature of sedimentation on the shelf.

Simple calculations permit one to determine the duration of the various depositional cycles that have been identified. For instance, the length of each sedimentation cycle in the column can be determined by dividing the full potential of the column by the number of sediment cycles. Then, knowing the duration of the continuous sedimentation process, it is possible to determine their duration in years.

The analysis of the structure of the depositional series in Karkinitsky Bay and the Dnieper Paleovalley convinces the author that in the Novochernomorian, sediment accumulation happened without interruptions. It is therefore possible to determine the duration of the sedimentation periods for this stage of geological history. The author accepts the point of view of P.V. Fedorov (1982), who determined the maximum duration of the Novochernomorian stage as 7500 years. Now it is not difficult to calculate the duration of each of the selected periods (rhythms). For the investigated columns, the following periods (in years) were obtained: 1974, 1415, 551, 336, 286, 214, 169 and 136. The periods of such duration are well known by ^{18}O and ^{14}C content and the data of the astrophysical research (Khlystov *et al.* 1992; Shmuratko 2001).

A cycle the size of 1974–1415 is proportional to Shnitnikov's cycle (millennium cycle). It is repeated within the Novochernomorian deposits 3.5 times (Figures 9, 10, and 11). The last semi-cycle corresponds to the transgressive phase after the Phanagorian regression of the Black Sea. Higher frequency cycles, which are reflected in the cyclic geological and climatic processes, are stipulated by the astronomical events in the solar system. For example, the periodicity of functions of the barycentric movement of the Sun is as follows (in years): 169.3; 380–392.6; 649.5–760 (Khlystov *et al.*1992). On this scale, which is restored using the concentration of atmospheric carbon (^{14}C), the spectrum of solar activity for the last 7500–8000 years is characterized by the following periods at about: 29, 60, 80–90, 200, 370, 500, 700, 1000, and 2300 years (Khlystov *et al.*1992; Shmuratko 2001). The given information convinces us of the reality of the obtained frequencies of sedimentation rhythms on the northwestern shelf of the Black Sea.

5.2 Paleogeography

In accordance with our previous estimates (Voskoboinikov *et al.*1983) and in the opinion of other researchers (e.g., Shcherbakov *et al.*1978; Fedorov 1982; Shnyukov 1985), the level of the Neoeuxinian lake at the LGM (20 ky BP) was about –90 to –100 m. Numerous cores have recovered deltaic deposits of alluvial-marine genesis, sediments of coastal wave-cut facies, and riverine erosional downcuttings filled in by Neoeuxinian alluvium up to the –75 to –95 m isobath (Konikov *et al.*1999). The northwestern Black Sea shelf was a dry terrain dissected by river valleys (Figure 12). If the level of the lake had been –140 m as has been suggested by Ryan *et al.*(1997), river erosion would have been much deeper.

The Neoeuxinian basin transgressed during the interval of 18–12.7 ky BP, reaching –32 to –35 m below present at 12.7–12.5 ky BP (Shnyukov 1985;

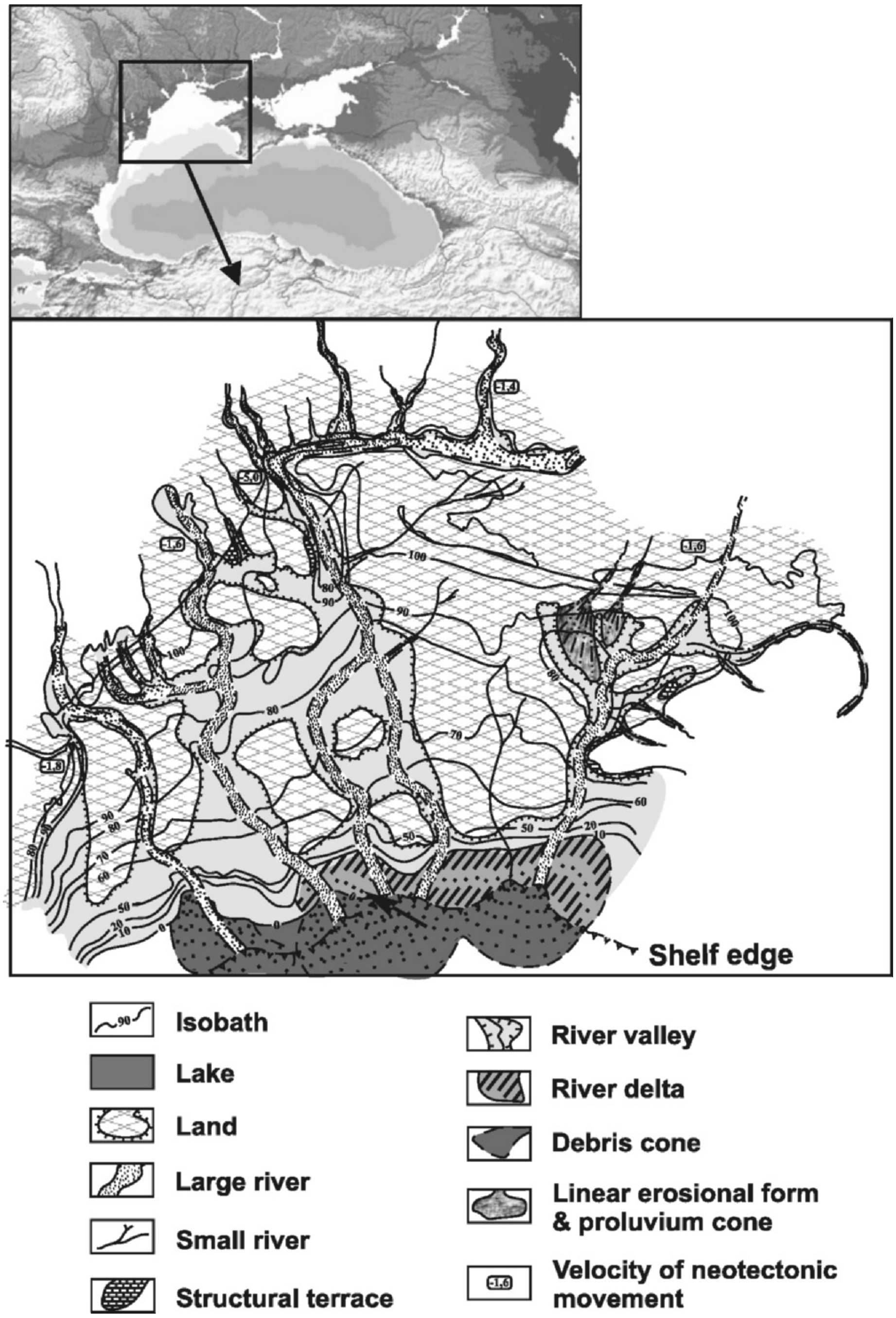

Figure 12. Northwestern shelf during the Early Neoeuxinian regression (~20 ky BP).

Balandin and Mel'nik 1987) (Figure 6).

For a prolonged period, until 9.8–9.5 ky BP, the basin stalled at this level. Possibly, in the final stage of its transgression, the Late Neoeuxinian basin rose to a higher level, reaching –18 to –20 m, as dated Late Neoeuxinian deposits were encountered at such depth in the North Pontic limans (Shnyukov 1985). The present writers (Shnyukov 1985) do not exclude the probability of minor regression at about 11.8–10 ky BP.

The basin ingressed into the river valleys to a distance of 20–30 km seaward from the present shoreline position. In the course of this ingression, a large Dnieper paleolagoon, as well as most of the Black Sea limans were formed, including the Dniestrovsky, Khadzhibeysky, Kuyalnitsky, Tiligulsky, Berezansky, and others (Voskoboinikov *et al.* 1982b; Shnyukov *et al.* 1982, 1984; and others). This is confirmed by the occurrence of clay and clayey muds within the limans at –18 to –26 m and dating to 13.1–9.2 ky BP (Shnyukov 1984).

According to the writers' evidence, a large semi-isolated shallow lagoon was formed in the area of present-day Karkinitsky Bay at 9.5–9.2 ky BP. Its depth is estimated to have been about 14 m, while the sea level stood at –18 to –20 m (Figure 2).

The attribution of both Neoeuxinian shorelines, as well as the older basins, the Karangat and Ancient Euxinian, to the levels of –30 to –70 m (Konikov *et al.* 1999) is by no means accidental. Seismic reflection profiles (Tugolesov 1985) and geological sequences show three structural-tectonic steps on the northwestern Black Sea shelf: the first one stretching from the present-day coastline seaward to the –30/–35 depth contour; the second one lying between the –30/–35 and –65/–70 depth contours; and the third one from –65/–70 to the shelf edge. This configuration apparently reflects the impact of recent tectonic movements on the paleogeography of the shelf area.

During the Holocene, the sedimentary environment of Karkinitsky Bay underwent repeated changes due to sea-level oscillations. It alternatively became a bay open to the sea or an isolated shallow lagoon depending upon shoreline position (Nevessky 1967; Shcherbakov *et al.* 1978).

Yet, due to the effect of the Pleistocene-Holocene tectonic submergence (Shcherbakov *et al.* 1978) and the low magnitude (5–15 m) of Holocene sea-level oscillations (Fedorov1982; Voskoboinikov *et al.* 1982a; Zubakov 1986; Ivanov and Shmuratko 1983), the conspicuous indices of wave reworking are apparent only in the Drevnechernomorian deposits.

According to the same writers' evidence, the Drevnechernomorian sea level was ~20 m below its present position. During the Novochernomorian, transgressive-regressive sea-level changes on the outer shelf were reflected only in the composition and character of bottom deposits (Konikov 1992, 1993). Closer to the present-day shoreline (with a water depth of less than 20 m) and in the limans, the transgressions and regressions are firmly recorded in the ancient shorelines, identifiable by beach facies (Leonov 1982; Voskoboinikov *et al.*

1982b; Ivanov and Shmuratko 1983; Shnyukov 1984).

The sequence of Late Pleistocene and Holocene events suggested here is essentially in accord with that shared by several leading researchers (Degens and Ross 1972; Fedorov 1982; Ivanov and Shmuratko 1983), and it provides a more detailed account of the sedimentary environments.

6. CONCLUSIONS

It has become apparent that during the course of the Neoeuxinian transgression, from 18 to 12 ky BP, the level of the Black Sea rose from –90/–100 to –30 m. By 9.5–9.2 ky BP, sea level stalled at an elevation of 18–20 m below its present elevation. During the Holocene, repeated transgressive-regressive oscillations of varying duration and magnitude occurred against the background of a general sea-level rise. As follows from the paleogeographic and paleo-hydrochemical reconstructions, various sedimentary environments resulting from the sea-level oscillations repeatedly replaced each other across the Black Sea shelf. Sedimentation on the northwestern shelf was dominated by transgressive-regressive sea-level oscillations and other hydrodynamic activity. This conclusion follows numerous paleogeographic and paleo-hydrochemical reconstructions (Bruevich 1952; Nevessky 1967; Degens and Ross 1972; Shcherbakov *et al.* 1979; Fedorov 1982; Voskoboinikov *et al.* 1982a; Ivanov and Shmuratko 1983; Shnyukov 1985; Konikov 1993; and others).

A new curve of sea-level changes in the Black Sea over the past 18,000 years has been plotted based on the evidence of lithogenetic analysis and absolute radiometric dates (Figure 13).

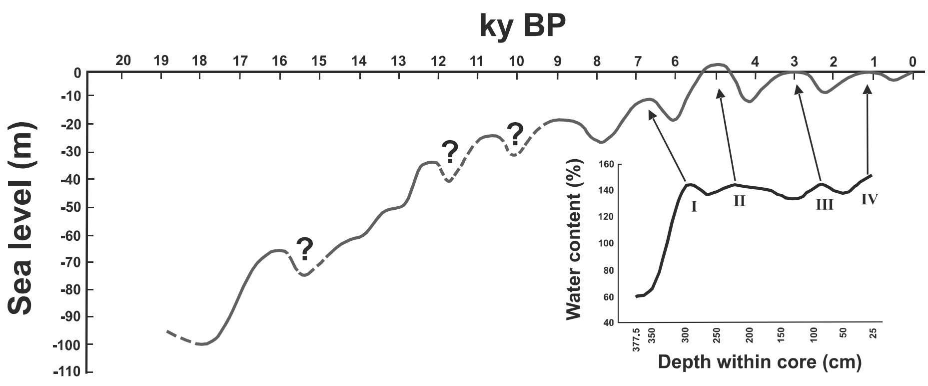

Figure 13. Reconstruction of sea-level changes in the Black Sea for the last 20 ky. The curve (smoothing based on 33 quantities) represents water content of sediments in core 1136; I–IV = transgressive-regressive cycles.

In summary, it can be demonstrated that during the late Pleistocene and Holocene, sedimentation on the northwestern shelf of the Black Sea resulted from transgression-regression oscillations of sea level and other hydrodynamic activity. The heterogeneity of sedimentation is reflected in the sequential changes of sediment properties. Sedimentation proceeded in statistically discernible cycles. Based on statistical processing of the available data, the duration was estimated in calendar years as: 1974–1415, 551, 336–286, 214, 169, and 136.

The Black Sea transgression-regression phases are controlled by Shnitnikov's climatic (humidity) cycles. The following regressive phases are identifiable in the architecture of Holocene sediments on the northwestern Black Sea shelf and its limans: Phanagorian, 2800–2200 BP (Fedorov 1982); Khadzhibeian, 4800–4200 BP (Voskoboinikov *et al.* 1982a; Shnyukov 1985; Konikov 1993); Tiraian, 6500–5800 BP (Ivanov and Shmuratko 1983); and Drevnechernomorian, 8500–7500 BP (Fedorov 1982). The amplitude of these oscillations is assessed as 5–15 m. Cycles with a duration of 700–500 years do not always yield clear signals in the geologic sequences, but the plainly recognizable Medieval regression of the 13th–18th century (the Little Ice Age) is a notable exception (Shnitnikov 1969).

High frequency sedimentation rhythms have been deduced from sequential patterning of the physical and mechanical properties of strata in the northwestern Black Sea shelf based on sediment structure, composition, and other variables. Based on the length of the sedimentary cycles, the mean sedimentation rate in Karkinitsky Bay has been estimated as 0.34–0.51 mm/y, and in the Dnieper Paleovalley, 0.29–0.36 mm/y.

Hence, using the proposed methodology, it has been possible to identify periods and cycles of various durations that illuminate paleoclimatic changes in the study area. Apart from that, the sedimentary texture, lithology, and physical properties (such as salinity and chemical composition of pore water) have been deemed reliable proxy evidence of the sedimentary environment on the shelf.

The development of a high-resolution model of Late Pleistocene-Holocene sedimentation for the northwestern Black Sea shelf constitutes a basis for forecasting future sea-level changes.

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18

WATER-LEVEL FLUCTUATIONS IN THE BLACK SEA SINCE THE LAST GLACIAL MAXIMUM

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Abstract: Two IFREMER oceanographic surveys carried out in the northwestern Black Sea in 1998 and 2002 complement previous seabed mapping and subsurface sampling by various international expeditions. They show that the lake level rose on the continental shelf to at least the –40 to –30 m isobath based on the landward limit of a *Dreissena* layer representative of very low salinity conditions (< 5‰). The Black Sea then shows clear evidence for an onset of marine conditions at 7150 BP. From these observations, Ryan *et al.* (1997) concluded that the Black Sea could have filled abruptly with saltwater cascading in from the Mediterranean. Despite critical discussions of this interpretation, recent IFREMER discoveries of well preserved drowned beaches, sand dunes, and soils appear to lend support to the flood hypothesis. This new evidence includes (1) multibeam echo-sounding and seismic reflection profiles that reveal wave-cut terraces at about –100 m, (2) Romanian shelf cores that show an erosion surface indicating subaerial exposure well below the sill of the modern Bosphorus, (3) ^{14}C ages documenting a colonization of the former terrestrial shelf surface by marine molluscs at 7150 BP, (4) evidence of sea water penetration into the Black Sea in the form of recent canyon heads at the Bosphorus outlet, and (5) palynological analysis and

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dinocyst studies that pinpoint the arrival of freshwater during the Younger Dryas and, later, the rapid replacement of Black Sea dinocysts by a Mediterranean population.

Keywords : Rapid transgression, Younger Dryas, Seismic stratigraphy, multibeam geomorphology, Bosphorus outlet, Black Sea continental shelf

1. INTRODUCTION

In 1997, Ryan and Pitman (Ryan *et al.* 1997) presented astonishing evidence supporting a catastrophic flooding of the Black Sea basin about 7500 years ago and speculated on the role it may have played in the spread of early farming into Europe and much of Asia. Their book, *Noah's Flood: The New Scientific Discoveries About The Event That Changed History* (Ryan and Pitman 1998), proposed that the Black Sea flood could have cast such a long shadow over succeeding cultures that it inspired the deluge account in the Babylonian Epic of Gilgamesh and, in turn, the story of Noah in the Book of Genesis.

This hypothesis resulted from a joint Russian-American expedition conducted in 1993 on the continental shelf south of the Kerch Strait (Major 1994; Ryan *et al.* 1997). High-resolution seismic reflection profiles, cores precisely targeted on these profiles, and ^{14}C accelerator mass spectrometry (AMS) dating of recovered fauna permitted several conclusions. The survey revealed a buried erosional surface strewn with shell gravel that extended across the broad continental margin of the northern Black Sea and beyond the shelf break (Evsylekov and Shimkus 1995; Major 1994). The cores revealed evidence of subaerial mud cracks at 99 m beneath the sea surface, algal remains at 110 m, and *in situ* roots of shrubs in desiccated mud at ~123 m. Each site lay well below the 70 m depth of the Bosphorus bedrock sill (Algan *et al.* 2001; Gökaşan *et al.* 1997).

The combined evidence suggested to the research team that a drowning event in the Pontic basin may have resulted from a marine transgression into a vastly shrunken lake. It appeared that this inundation subsequently deposited a uniform drape of marine mud upon the former terrestrial surface, creating a sapropel layer equally thick in depressions as on crests of dunes, and with no sign of landward-directed onlap of the sedimentary layers in the drape (Ryan *et al.* 2003). The ^{14}C age determinations documented a simultaneous subaqueous colonization of the terrestrial surface by marine molluscs at 7100 BP,[1] and this date was assigned to the flooding event.

The previous shrunken lake stage and subsequent flooding preclude the possibility of outflow to the Sea of Marmara at that time. Recently, however, Aksu *et al.* (2002c) presented arguments for persistent Holocene outflow from the Black Sea to the eastern Mediterranean and for non-catastrophic variations in the level of the Black Sea during the last 10,000 years (Aksu *et al.* 2002a).

These findings are in direct contradiction to the flood hypothesis.

In 1998 and 2002, two Black Sea projects, one European and the other more broadly international, enabled IFREMER to conduct two oceanographic surveys that afforded the opportunity to complete work previously begun on seabed mapping and subsurface sampling. The main objectives of these cruises were to prepare for the European ASSEMBLAGE Project, which would assess Black Sea sedimentation since the Last Glacial Maximum (LGM), quantify the impacts of climate change, and evaluate the sensitivity of the Black Sea system to external forcing. Progress in resolving these major issues can be achieved only through examination of geomorphology and stratigraphy from the shelf to the abyssal deep in the northwestern corner of the Black Sea.

These surveys revealed that lake level in the Pontic basin rose on the shelf to at least the –40 to –30 m isobath, which corresponds to the landward limit of a layer containing the shelly debris of *Dreissena rostriformis distincta*, a mollusc indicative of freshwater conditions. This rise in freshwater suggests that the lake which preceded the formation of the Black Sea served as an important catchment for meltwater draining out of the Fennoscandian ice cap: Melt Water Pulse 1A during the Bölling-Allerød interval (Bard *et al.* 1990). Possibly, the lake level at this time broke over the sill of its outlet and spilled into the Mediterranean through the Marmara Sea, however, the onset of salt water conditions creating the Black Sea during the mid-Holocene (7150 BP) has been clearly demonstrated. Although opposing hypotheses have been proposed (Aksu *et al.* 1999b, 2002a, b, c), recent discove ries of well preserved drowned beaches, sand dunes, and soils provide new evidence supporting the Ryan and Pitman flood hypothesis.

2. GEOLOGICAL BACKGROUND

The Black Sea is a 2.2 km deep basin with a wide continental shelf in its northwestern corner (Figure 1). External connection to the Mediterranean Sea is over a sill within the Bosphorus Strait. The role played by this strait in controlling the salinity and stratification of both seas as well as the production of anoxia in the Black Sea has been often discussed (Arkhangel'sky and Strakhov 1938; Scholten 1974; Muramoto *et al.* 1991; Rohling 1994; Lane-Serff *et al.* 1997; Abrajano *et al.* 2002). The general view is that during periods of low global sea level, the connection between the Black Sea and the ocean was severed. The Black Sea then freshened as a result of continued river discharge, expanding into a vast, well ventilated lake with a shoreline established at the level of its outlet (Chepalyga 1984; Hodder 1990), from which excess water was exported to the Mediterranean.

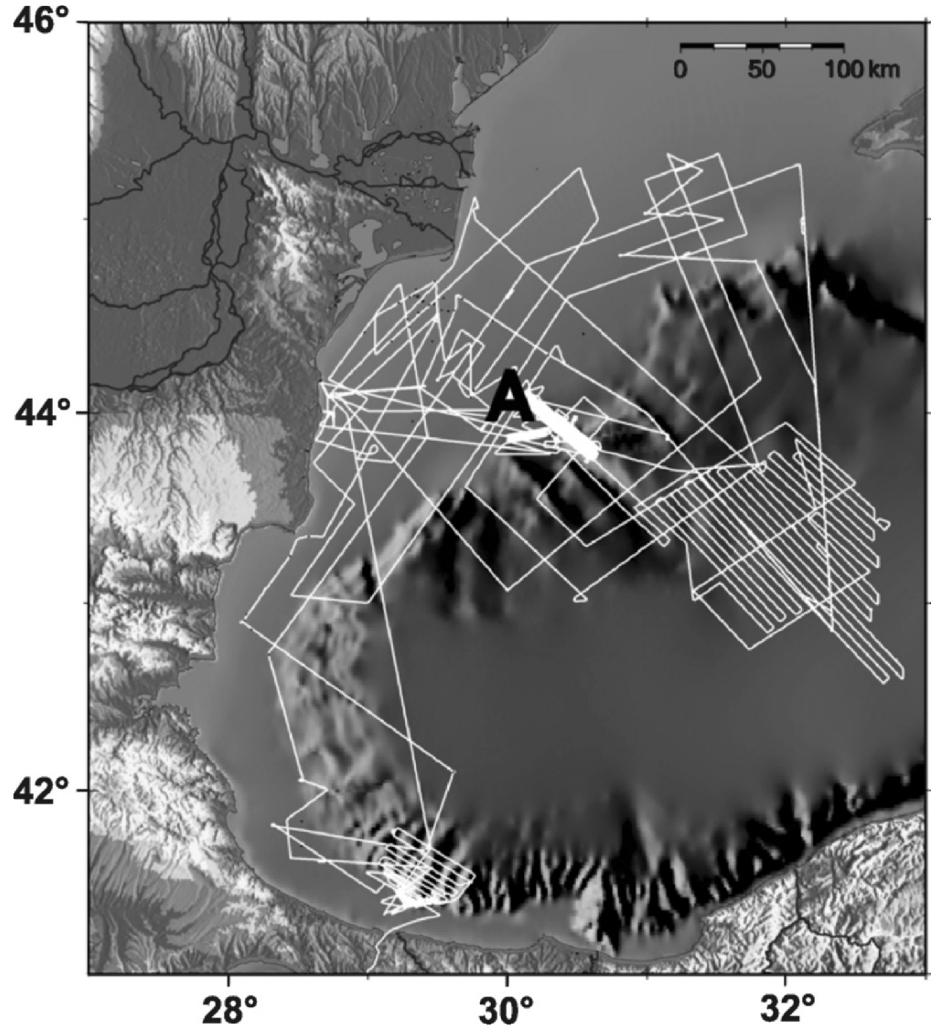


Figure 1. Bathymetry of the semi-enclosed Black Sea basin and IFREMER survey route locations. "A" represents location depicted in Figure 3.

The largest European rivers, the Danube, Dnieper, and Dniester, deliver their water and deposit their sediment loads into the northwestern Black Sea. The Danube drainage basin, 817,000 km^2 in areal extent, charges the Black Sea at an estimated rate of 6047 m^3/s (almost 190 km^3/yr), while the river's sediment discharge at its mouth was about 51.7 million tons per year (t/yr) before the river was dammed in 1970 and 1983 (Bondar 1998). Estimates of total sediment discharge subsequent to damming average less than 30–35 million t/yr, of which only 4–6 million t/yr is sandy material (Panin 1997). During glacial lowstands, and especially at the beginning of interglacials, sediment discharge from these rivers was probably much higher.

It has long been recognized that the Black Sea became isolated from the Marmara and Mediterranean Seas during glacial intervals when the world ocean level fell below the Bosphorus sill depth (–35 m). It has also been postulated that water level in the Black Sea began to rise in step with that of the Marmara and Mediterranean Seas during times when the level of the world ocean had risen above the Bosphorus sill depth (Degens and Ross 1974). Recent sediment analyses along the margins of the Black Sea suggest, however, that water-level fluctuations in the Pontic basin were somewhat more complex, with high lake levels occurring during the wet intervals of late glacial stages and low lake levels occurring during the drier intervals of early interglacials (Chepalyga 1984).

Another widely accepted hypothesis postulates that the Black Sea had always maintained a continuous outflow through the Bosphorus and Dardanelles Straits, even during highly arid glacial intervals (Chepalyga 1984; Kvasov and Blazhchishin 1978). This perspective essentially assumes that precipitation and river input into the Pontic basin always exceeded loss from local evaporation. Indeed, meltwater from the former ice sheets in Fennoscandia, northern Asia (Grosswald 1980), and the central Alps transformed the Black Sea into a giant freshwater lake a number of times during the past (Federov 1971; Ross *et al.* 1970), most recently during the Late Pleistocene Neoeuxinian stage of the LGM, between 25–18 ky BP (Arkhangel'sky and Strakhov 1938; Nevesskaya and Nevessky 1961; Nevesskaya 1965; Ross *et al.* 1970; Fedorov 1971).

Ryan *et al.* (1997) published evidence suggesting that the Black Sea became a giant freshwater lake during the LGM. This evidence included new AMS ^{14}C dates, abrupt changes in the organic carbon content, water content, and $\delta^{18}O$ from core material dated approximately 7150 BP, as well as the occurrence of a widespread unconformity interpreted as an erosional surface subaerially exposed during the last glacial. The distribution of recorded depths for this unconformity implies that the surface of the freshwater lake must have fallen to levels greater than 100 m below its outlet. Ryan *et al.* (1997) inferred from this that, by about 7150 BP, the sill depth of the Bosphorus was breached by the rising world ocean, and a catastrophic flooding of the continental shelf of the Black Sea occurred (Figure 2).

Evidence that does not support the catastrophic flood hypothesis includes variously aged sapropels sampled from the eastern Mediterranean and the Black Sea. Sapropel S1 in the Aegean is generally thought to have been deposited between about 8000 and 5500 BP (Aksu *et al.* 1999a, b; Fontugne *et al.* 1994), although deposition may have lasted until 5300 BP (Rohling and de Rijk 1999). Aksu *et al.* (1999a) suggest that during this time, nutrient-rich freshwater from the Black Sea reduced the surface salinity of the eastern Mediterranean, thereby increasing the stability between the surface and deep waters and decreasing deep circulation. As a result, high surface productivity and restricted

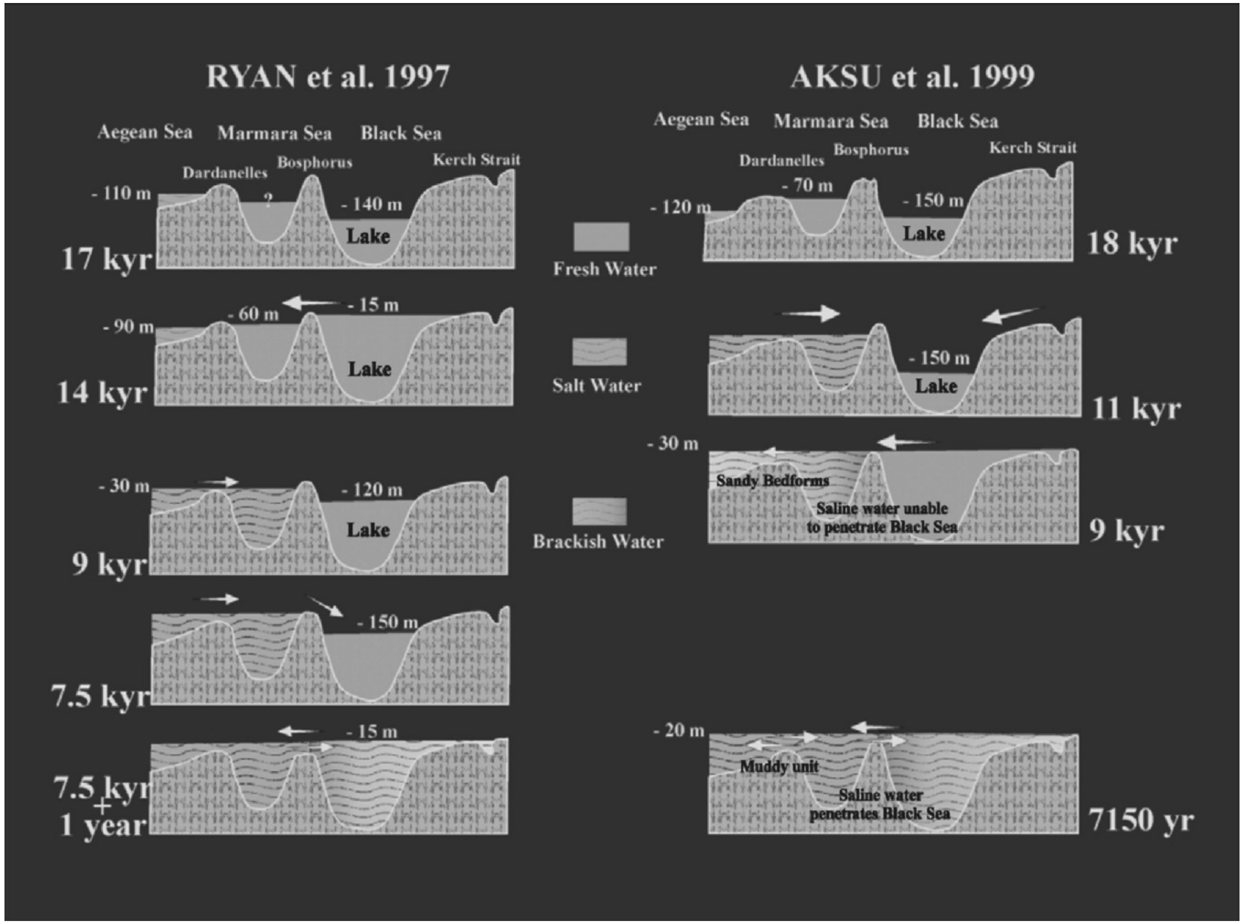


Figure 2. Alternative schemes of reconnection between the Black and Mediterranean Seas.

circulation provided conditions favorable for sapropel formation. Rohling (1994), however, suggests that sapropel formation in the Black Sea started about 550 years later than it did in the eastern Mediterranean, when denser water from the Mediterranean, upon entering the Pontic basin, displaced the nutrient-rich waters in the Black Sea upward toward the surface (Calvert 1990; Calvert and Fontugne 1987). This lag is probably too large to be accounted for by the catastrophic flooding hypothesis.

3. OBSERVATIONS

3.1 Acquisition and Processing of Data

During the two campaigns on board the *R/V Le Suroît* a DGPS system afforded precise positioning. In some areas (for example, the dune field of Figure 3), real-time navigation processing with an accuracy of a few meters made it possible to follow parallel track lines spaced 200 m apart at a speed of 4 knots. The vessel was equipped with a Simrad EM 1000 swath bathymetry system, and very high resolution seismic lines were shot simultaneously using a mud-penetrator and Chirp sonar system. All data acquisition was synchronized and digitally recorded.

An EM 1000 multibeam echo-sounder provided mapped bathymetry by processing the returned echo of each sonic transmission through a selection of 68 pre-formed beams trained across a strip of sea floor effectively four times as wide as the water depth. Each beam gives a depth resolution on the order of 10 centimeters over a five by two meter footprint along the surveyed swath, and all obtained at the survey speed of 4 knots. An image of backscatter reflectance energy was generated along with the digital elevations. Navigation, bathymetry, and image data were processed and synthesized in order to plot positions automatically, and thereby produce bathymetric maps and an image mosaic.

The very high resolution seismic reflection sources were a single channel mud-penetrator with a central frequency around 2500 Hz and an XSTAR Chirp sonar, sweeping between 4 and 16 kHz. The digital data acquisition was done in real time on the Triton-Elics Delph PC-based system.

Coring was achieved using a Kullenberg piston core and, in some cases, a vibrocorer.

3.2 Topography

Bathymetry data provided by the multibeam echo-sounder are illustrated in Figure 4. Prominent in the northern half of the survey area are linear ridges

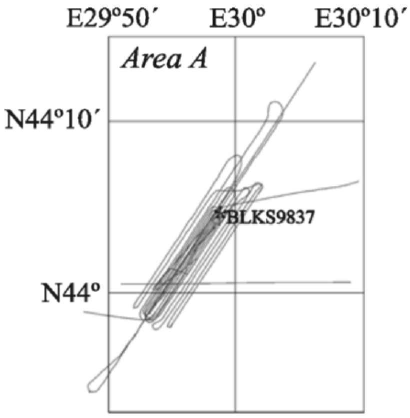


Figure 3. Map showing the location of profiles acquired on the dune field of Area A (indicated on Figure 1).

four to five meters in relief with an average spacing of 750 m. The ridges strike almost uniformly at an azimuth of 75±10°, they typically reveal asymmetrical cross-sections with the steeper sides facing to the southeast, and they possess a length to width ratio exceeding four. Numerous depressions 100 to 1800 m in diameter and 3 to 9 m deep also populate the southern half of the corridor. Depths of individual depressions are greatest at the base of their northeast walls, and they shoal to the southwest. In the center of the surveyed corridor, some depressions align within the troughs between the linear ridges.

3.3 Subsurface Structure

The ridges and depressions can be viewed in cross-section using high resolution seismic reflection profiles. The sub-bottom profiler and Chirp sonar provide seismic penetration to tens of meters and define layering to the sub-meter scale. The profiles indicate that the ridges in the north are asymmetrical, and with one exception, their steeper side faces to the southeast. These ridges are superimposed on a reverberant "bottomset" reflector that is sometimes conformable with subjacent strata, but in many cases truncates them (Figure 5). Although the high ground in the south has a mound-like appearance, seismic surveys show an asymmetrical cross-section with the crest and steeper slope predominantly located on the south side. The interiors of the ridges and mounds contain steeply-dipping "foreset" clinoforms with the same asymmetry and orientation as the cross-section topographic profiles.

The ridges, mounds, and depressions of the mid and outer shelf are everywhere draped by a thin layer of sediment of remarkably uniform thickness, no more than a meter (Figure 5). The linear ridges surveyed in this study are aligned somewhat obliquely to the regional bathymetric contour and to the paleo-shoreline outlined by wave-cut terraces (Figure 6).

3.4 Sediment Cores

Sediments obtained by coring provide ground truth for the reflection profiles. Sampling into the interior of a ridge on the dune field (core BLKS9837

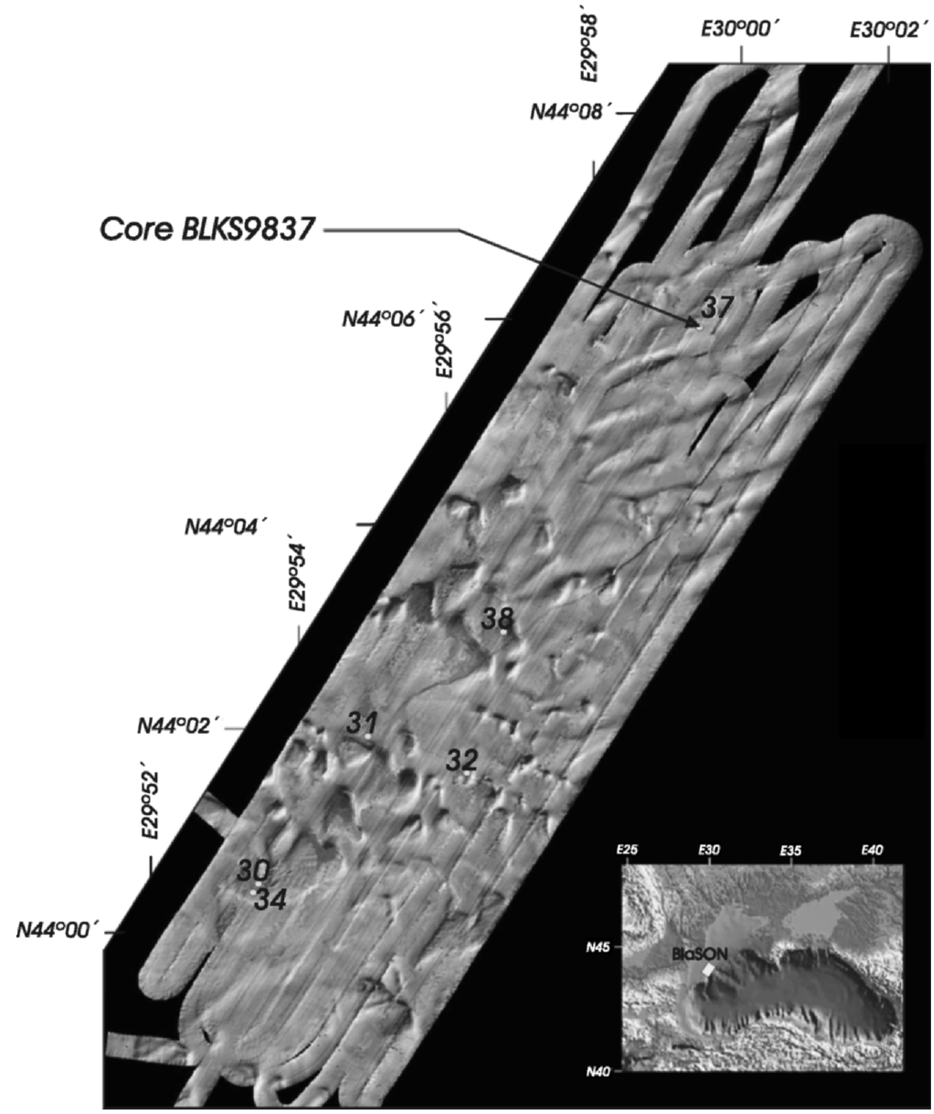


Figure 4. Multibeam bathymetry of the sand dune field and the location of core BLKS9837.

taken in a water depth of 68 m, core length 190 cm, position N44°00.54'-E029°58.87') recovered dark sand rich in opaque heavy minerals and shell fragments (Figure 7). Sampled minerals include quartz, garnet, and ilmenite. Shell fragments belong to freshwater mussels of *Dreissena rostriformis distincta*. Coring of the bedded sediments underlying the dunes and upon which they formed extracted silty red and brownish clay with thin lenses containing fresh to slightly brackish water molluscs (*Dreissena* and *Monodacna* spp., respectively).

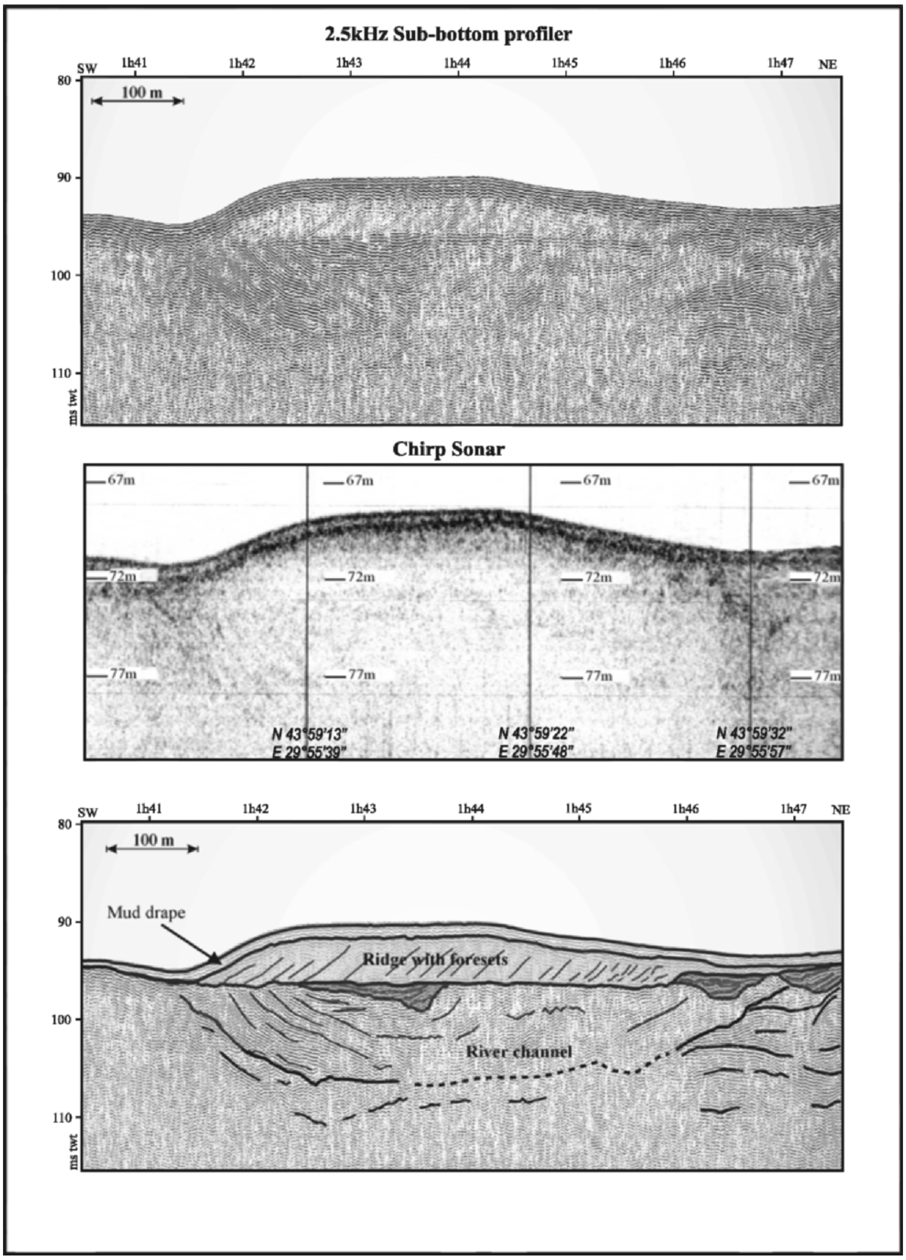
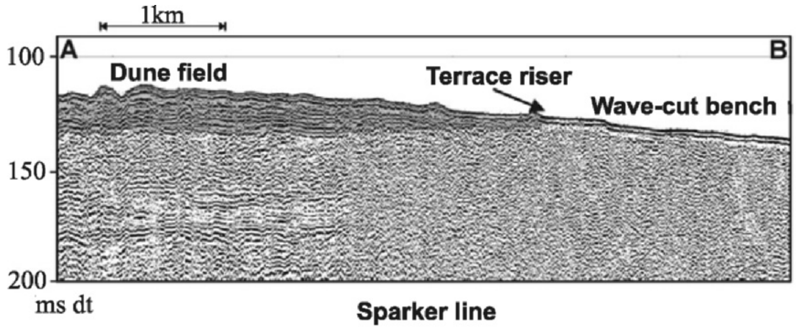


Figure 5. Seismic profiles across a sand dune.

These molluscan specimens return AMS radiocarbon dates spanning the interval 8585 to 10,160±90 BP (without reservoir and dendro-chronologic calibration).



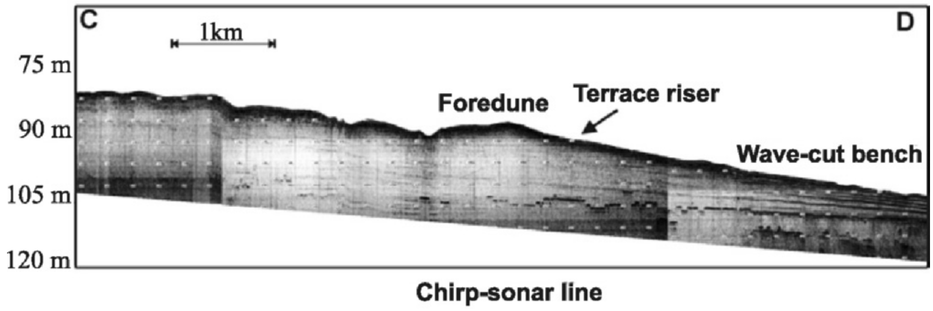


Figure 6. Seismic profiles across the wave-cut terrace.

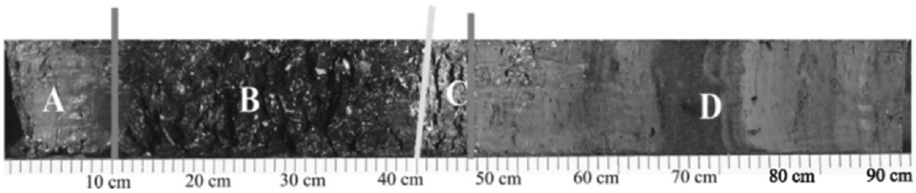


Figure 7. Core BLKS9837 (N44°00.54'-E29°58.87') recovered from dune field; A = layer with Modulus dated from 3000 BP to present; B = layer with *Mytilus edulis* dating between 6900 and 3000 BP; C = layer with *Dreissena* sp., the youngest dating 7150 BP.

Molluscs recovered in the cores from within the uniform surface drape reveal exclusively saltwater species, such as *Mytilus edulis* (also known as *Mytilaster*) and *Cerastoderma edule*. Those sampled near the base of the drape date between 6590 and 7770±80 BP.

Cores that penetrated though the drape reveal a sharp basal contact between the overlying organic mud with its marine molluscs and an underlying sand of variable thickness, rich in *Dreissena* sp. detritus. Recent palynological and dinocyst analyses by Speranta-Maria Popescu (Popescu *et al.* nd) on samples from within these cores indicate an abrupt freshwater transition during the

Younger Dryas followed by an abrupt replacement of the Black Sea dinocysts by a Mediterranean population at 7150 BP.

4. DISCUSSION

The sand dune fields and associated wave-cut terrace are interpreted as coastal zone relics that have persisted since 9680 to 8360 BP. At that time, a wave-cut terrace presently at –100 m with dunes and pans between –80 and –65 m would have lain well below the level of the external ocean (Fairbanks 1989). A lake could have existed below global sea level only if the Bosphorus barrier was higher than the external ocean, thereby preventing outflow. Burial of the dunes and pans by a drape of mud is not sufficient to imply a sudden infilling of the depression once the Bosphorus barrier was breached, but considering the evidence of (1) an abrupt transition from shell hash (a very condensed layer with brackish fauna) to mud, and (2) the impressive preservation of dunes and pans with no preferential infilling of the depression, a rapid terminal transgression of the Black Sea seems a compelling interpretation. The Caspian Sea, which appears to have experienced similar phenomena (with the exception of the last reconnection), reveals in the Iranian coastal regions of Mazandaran and Gilan comparable sand dunes on the terrestrial surface as high as 20 m that lie parallel to the seashore and which contain sandy particles and fragments of shells. The relic Black Sea dunes are occasionally cut across by aeolian deflation, which testifies to a previous lowstand.

Water-level fluctuations in the Black Sea appear to be directly linked to climate variability (Kvasov 1975; Svitoch *et al.* 2000). The Caspian Sea, another enclosed basin, reacted similarly. When not connected to the Mediterranean via the Black Sea, it reached highstands and achieved outflow in cold periods and drew down to lowstands through evaporation in warm periods. When the Mediterranean Sea penetrated the Dardanelles Strait at *ca.* 12,000 BP, the level of the Marmara was below its outlet (Aksu *et al.* 1999b, 2002b).

Whether catastrophic flooding is hydraulically possible remains open to question. Lane-Serff *et al.* (1997) modeled the events following the connection of the Black Sea to the Mediterranean hydraulically and assumed that the sill cross-section has remained more-or-less unchanged. They showed that significant freshwater outflow from the Black Sea occurred only 500–1000 years after sea level in the Mediterranean reached the Bosphorus sill depth. A two-layer exchange between the two seas was initiated when an upper limit to the water flux was able to pass over the sill. This delay corresponds in order of magnitude to the lag between the onset of sapropel deposition in the Mediterranean and that in the Black Sea. Furthermore, they also demonstrated that it took 2500–3500 years for the bulk of the freshwater in the Black Sea to be replaced by salty

Mediterranean water and for euryhaline-marine conditions to be established. This period corresponds to the time of eastern Mediterranean sapropel deposition.

Debate continues about whether a catastrophic flood in the Black Sea significantly impacted habitation, migration, and cultural practices of European Neolithic populations. The archaeological interpretations presented by Ryan and Pitman (1998) depend upon the principal archaeological assumption that after the onset of agriculture, the ancient Near East suffered a drought that forced the first farmers to find refuge in a more hospitable climate, that of the pre-flood Black Sea coast. Kalis *et al.*(2003) suggest that a more realistic picture of the Neolithic diaspora into Europe consists of various waves, currents, and eddies of people, bringing animals and plants. In some areas, an influx of migrating farmers replaced sparse local foraging populations with Neolithic agricultural communities. Elsewhere, early farming communities appeared as isolated pioneer outposts within a multicultural landscape of foragers and farmers.

5. CONCLUSION

Recent surveys carried out on the northwestern continental shelf of the Black Sea imply that the early Holocene lake level rose on the shelf to at least the –40 to –30 m isobath as evidenced by the landward limit of the *Dreissena* layer, which is characteristic of freshwater conditions. This rise in freshwater level is consistent with interpretations of the Pontic basin as an important catchment for meltwater drained from ice caps during Melt Water Pulse 1. It is possible that the lake, having filled with freshwater at this time, rose to the level of its outlet and spilled into the Mediterranean. However, in the mid-Holocene, at 7150 BP, the onset of salt water conditions is clearly evidenced in the Black Sea. From these observations, Ryan *et al.*(1997) came to the conclusion that the Black Sea could have been filled by saltwater cascading from the Mediterranean.

Despite discussions to the contrary, recent IFREMER seismic surveys indicate the presence of well preserved drowned beaches, sand dunes, and soils supporting the interpretations of Ryan and Pitman that the Black Sea was rapidly filled with marine water from the Mediterranean during the mid-Holocene.

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ENDNOTES

1. The abbreviation BP means radiocarbon years before present (1950) with neither correction for reservoir age nor calibration to calendar years. In Ryan *et al.* (1997) and Ryan and Pitman (1998), ages were expressed in calendar years, with 7500 calBP (= 5550 BC) equivalent to 7100 BP.

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19

ARCHAEOLOGICAL AND PALEONTOLOGICAL EVIDENCE OF CLIMATE DYNAMICS, SEA-LEVEL CHANGE, AND COASTLINE MIGRATION IN THE BULGARIAN SECTOR OF THE CIRCUM-PONTIC REGION

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Abstract: Evidence for Black Sea level changes during the Holocene are contained in the biostratigraphic, geomorphological, geological, and archaeological records. Based on 30 existing radiocarbon dates for the Bulgarian coastline, a general sea level curve for the Holocene is here suggested, details of which can be filled in as new data become available. The Holocene transgression reveals a gradual and cyclical character. From 1924 to 1998, average eustatic rise has been 2.5– 2.7 mm/y, but during the last 6000 years, sea level has exceeded that of the present on a few occasions, as indicated by marine terraces along the coast at elevations of 2–5 m. The observed differences between coastlines of varying age are due to their heterogeneous structure, neotectonic occurrences, and the lithologic and physiographic characteristics of the individual areas. The beginning and end of the transgressions can be traced in the sediments of the flooded estuaries and firths. A stratigraphic subdivision of these sediments is presented here, together with a discussion of the climate and vegetation changes reconstructed from the stratigraphic sequence.

Keywords: biostratigraphy, Late Pleistocene, Holocene, pollen, dinoflagellates, molluscs, paleoenvironmental reconstruction, Black Sea transgressions and regressions

1. INTRODUCTION

Although investigations have been ongoing for the past three decades, the Quaternary history of the Black Sea shelf is still not fully understood. Much

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of the evidence has been removed by repeated displacement in the high energy wave zone, and the complex, heterogeneous conditions under which shoreline sediments were laid down have created incomplete cross-sections and frequent lithological transitions. Despite these challenges, stratigraphic investigation of Late Pleistocene and Holocene sediments in the western Black Sea area is important not only for reconstructing paleoenvironmental changes but also for predicting and preventing unfavorable natural phenomena that could affect coastal and marine resources in the future.

Past biostratigraphic research on the Bulgarian Black Sea shelf has focused mostly on the reconstruction of paleoecological conditions along the coastline and within the shelf zone (Bozilova *et al.* 1979; Komarov *et al.* 1979; Filipova-Marinova 1986; Filipova and Dimitrov 1987; Bozilova *et al.* 1992; Filipova-Marinova *et al.* 2002). The data obtained have helped to:

(1) describe past vegetation and climatic changes and outline marine sediment stratigraphy (Govberg 1978; Filipova *et al.* 1983; Filipova *et al.* 1989; Atanassova 1990, 1995; Stoyanova 1990; Shopov 1991; Shopov *et al.* 1992),

(2) inform about sea-level fluctuations (Filipova-Marinova and Christova 2001), and

(3) locate submerged prehistoric settlements (Toncheva and Margos 1959; Filipova *et al.* 1993; Bozilova and Beug 1994; Draganov 1995; Lazarov 1996; Bozilova and Tonkov 1998; Filipova-Marinova and Bozilova 2003; Angelova and Draganov nd).

In order to correlate available data more precisely with the current European stratigraphic scale and to elucidate the mechanisms and causes for ecological changes (climate, human impact, tectonic movements, etc.), absolute age data have also been taken into consideration (Kuptzov *et al.* 1979; Nikolaev *et al.* 1980; Dimitrov 1982; Krastev and Michova 1990).

In this paper the latest paleoecological and archaeological data relating to vegetation, climate, and sea-level changes during the Late Pleistocene and Holocene are presented.

2. THE STUDY AREA

2.1 Geomorphology of the Black Sea Shelf

On the basis of the relief, shape, time of formation, character, and speed of sedimentological processes, three geomorphologic zones can be outlined in the western Black Sea shelf: littoral or inner, central, and peripheral or outer (Dimitrov 1979).

The littoral zone is considered to be of Holocene age (Dimitrov 1979).

It begins at the coast and continues to a depth of 50 m. Active wave impact, erosion, and accumulation are characteristic (Khrischev 1984). The littoral zone is separated from the central zone by a depression 17–20 m deep on the northern Black Sea shelf and 65–70 m deep on the southern Bulgarian Black Sea shelf.

The central zone lies between 50 and 72 m in depth. Within it, three subzones run parallel to the coast: an inner depression, a depositional bar, and a depositional plain. The area experiences a high sedimentation rate, typically about 2.5 m/kyr (Khrischev 1984).

The peripheral zone extends to depths between 90 and 120 m. It is subdivided into an outer depression and a barrier bar, and its low sedimentation rate is due to highly active bottom currents (Dimitrov 1979).

2.2 Climate

According to Velev (2002), the Bulgarian Black Sea coast exists within the Continental-Mediterranean climatic area and is influenced by the proximity of the sea. The northern and the southern parts of the coast differ. Climate in the north is affected by strong continental influences. Prevailing winds are north-easterly, and annual precipitation is about 450–500 mm, with a maximum in June and a minimum in February. Mean January temperature is around 0° C, dropping to –2° C inland. In the south, climate is transitional Mediterranean. Mean annual precipitation is estimated at about 500–600 mm, with rainfall mostly in the autumn-winter seasons. Mean January temperature is 2–3° C, and in July, it is 22° C. The dry summer period lasts from July to September. Winds blow mostly from the southeast and rarely from the northeast.

2.3 Modern Vegetation

According to Bondev (1991), the study area falls within the Black Sea region of the Euxinian province of the European deciduous forest. Today, much of the coastal area is used for arable farming. Natural forests are dominated by *Quercus cerris* L. and *Q. frainetto* Ten., but forests of *Carpinus betulus* L. with *Acer campestre* L., *Q. cerris*, and *Tilia tomentosa* Moench also occur. Only in the northern coastal area, in the South Dobruja region, there is steppe vegetation, which is dominated by *Agropyron brandzae* Pantu et Solac., *Koeleria brevis* Stev., *Stipa lessingiana* Trin. et Rupr., *Artemisia lerchiana* Weber, *Adonis vernalis* L., *Paeonia tenuifolia* L., and others.

Forests containing Mediterranean elements, such as *Carpinus orientalis* Mill., *Phyllirea latifolia* L., *Fraxinus ornus* L., *Quercus pubescens* Willd., and *Celtis australis* L., are distributed along the southern Black Sea coast. Forests with more continental species grow in lowland sites and on hilltops, as well as on the Strandzha Mountains. They comprise mainly *Quercus cerris*, *Q. frainetto*, *Q. polycarpa* Schur., and *Carpinus betulus*. Forests of *Fagus orientalis* Lipsky, with an undergrowth of evergreen shrubs (*Rhododendron ponticum* L., *Ilex aquifolium* L., and *Daphne pontica* L.) cluster within the more humid ravines of the Strandzha Mountains.

Riparian forests lining rivers and lakes are periodically inundated. Their components are mainly *Fraxinus oxycarpa* Willd., *Ulmus minor* Mill., *Carpinus betulus*, *Quercus pedunculiflora* C. Koch, and *Alnus glutinosa* (L.) Gaerth. The riparian forests are particularly rich in lianas: *Hedera helix* L., *Periploca graeca* L., *Smilax excelsa* L., *Vitis vinifera* L., and *Clematis vitalba* L.

Along the periphery of coastal lakes, reed formations are dominated by *Phragmites australis* (Cav.) Trin. ex Steud., *Typha latifolia* L., *T. angustifolia* L. and *Schoenoplectus lacustris* (L.) Palla (Kochev and Jordanov 1981).

Psammophytic herb communities, mostly of *Leymus racemosus* (Lam.) Tzvel. ssp. *sabulosus* (Bieb.) Tzvel., *Ammophilla arenaria* (L.) Link, *Centaurea arenaria* Bieb. ex Willd., *Galilea mucronata* (L.) Parl, and shrub communities with *Cionura erecta* (L.) Grsb., etc., grow on the sandy beaches and dunes.

2.4 Archaeological Background

Underwater archaeological investigations along the Bulgarian Black Sea coast over the last ten years have enabled the collection of detailed information on the prehistoric settlements submerged at depths of 4 to 8 m (Lazarov 1996). These settlements were occupied during two principal chronological periods.

Some of the submerged sites in the Varna Lake and Bay of Sozopol area date to the late stages of the Eneolithic and the initial stages of the Transitional Period (Final Eneolithic). The Late Eneolithic in Bulgaria ranges from 4450 to 4200 calBC. As for the western Black Sea coastline, a series of radiocarbon dates yielding corrected dates between 6420 and 6290 calBP was cited by Boyadziev (1995, 1998) and two additional dates, 5300±45 and 5310±40 BP/6250 calBP, have been reported by Preisinger *et al.* (2004).

From the perspective of cultural development along the Black Sea coast, the Eneolithic settlement in the harbor of Sozopol may be placed between the end of the Varna culture (phase III) and the beginning of the Chernavoda I culture (Boyadziev 1998). Along with the site at Cape Atia (near Sozopol), it probably represented a southern version of the Varna culture, with regional features still lingering behind in the south after the culture had vanished to the north of Cape Emine (Draganov 1995). In cultural and economic respects, the Late Eneolithic culture in that area could be regarded as a more progressive Black Sea version of the Kodzhadermen-Gumelnitza-Karanovo VI culture, which was widespread in northeastern and southern Bulgaria (Todorova 1985).

After the Eneolithic, the settlements were flooded and destroyed by intermittent oscillations in the level of the Black Sea or by downward movement of the land south of Burgas Lake. This break corresponds to the occupation hiatus before the next appearance of settlements after the Transitional Period (4200– 3200 calBC). At this time, the sea began to withdraw, exposing the Bay of Sozopol for human habitation once more (Draganov 1995).

Some Early Bronze Age settlements were built atop the remains of previous Late Eneolithic ones (Sozopol and Varna), while others arose in entirely new locations–e.g., Atia, Ropotamo (south of Sozopol), Ourdoviza, and Akhtopol (near the Veleka River, southern Bulgarian coast). These settlements were occupied mainly between 2900 and 2700 calBC, i.e., the second phase of the Early Bronze Age (Kuniholm *et al.* 1998). Radiocarbon determinations yielding dates from 4810 to 4517 calBP are cited by Boyadziev (1995, 1998).

3. MATERIALS AND METHODS

This paper summarizes data obtained during the investigation of 17 cores from shelf and deep-water zones of the Black Sea, together with 8 cores from coastal lakes and river estuaries (Figure 1). Reconstructing paleoclimatic dynamics and coastline migrations for the Black Sea requires a range of research methods, including analysis of molluscan fauna, spores and pollen, dinoflagellates, lithology, geomorphology, and seismostratigraphy, together with archaeology and radiocarbon dating. The ^{14}C dates (Table 1) have been calibrated according to Stuiver and Reimer (1993). Data from surveys conducted between 1980 and 1999 are summarized in Table 2 and set against the stratigraphic scheme (Shopov 1991) for Black Sea Quaternary sediments.

Sea-level changes during the Holocene have been delimited by analyzing terrace complexes along the seashore, locating marine phases that denote sea-lake oscillations, and comparing them with available radiocarbon dates. A curve of sea-level fluctuations during the last 11,000 years has been constructed (Figure 2). The curve represents a suggestion only; tectonic factors, landslides, erosion, and some differences in the dating make absolute precision impossible.

4. RESULTS AND DISCUSSION

4.1 Vegetation History and Climate Dynamics

Upper Neoeuxinian marine deposits are found below the 30 m isobath in almost all cores on the shelf. In the peripheral shelf zone, these deposits are represented by shell detritus as well as aleuritic and clayey silts between several

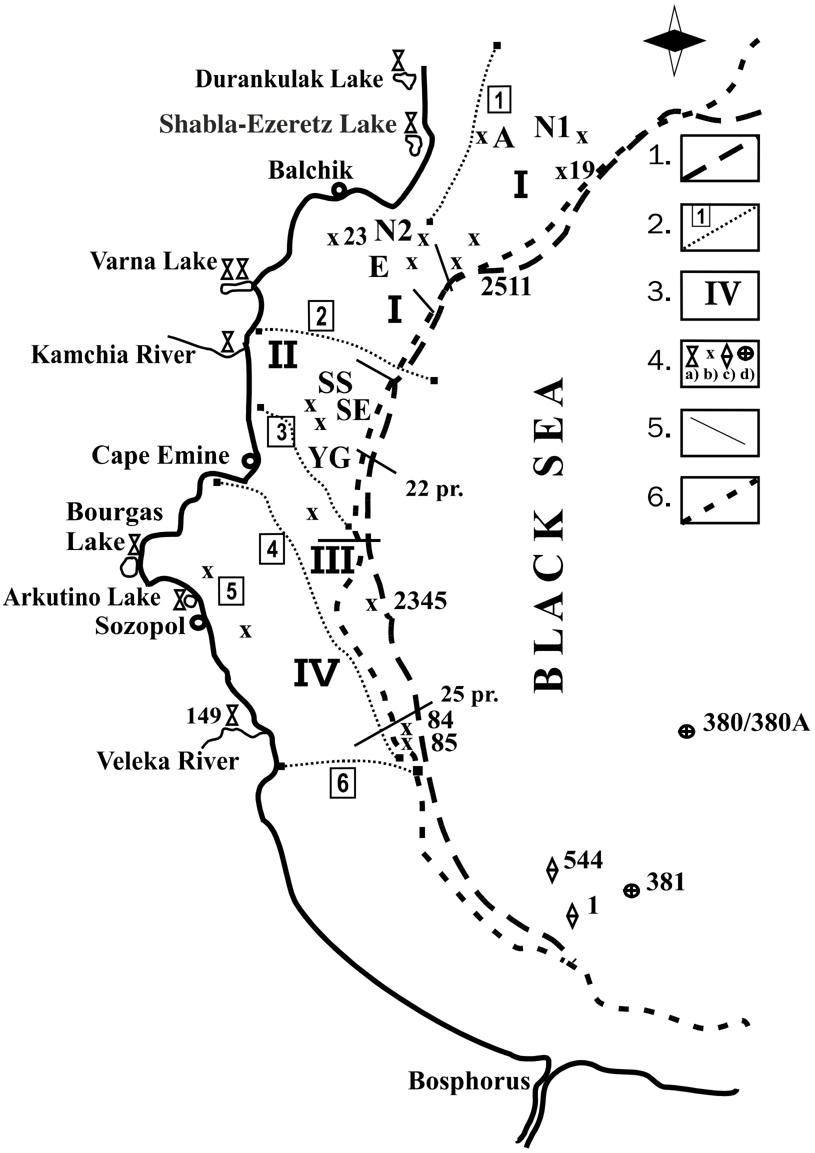

Figure 1. Scheme of the Bulgarian Black Sea shelf showing the location of investigated cores: **1.** boundary between the shelf and the continental slope; **2.** basic fault zones reflected in the bottom relief: (1) Eastern Tyulenovo fault, (2) Southern Moesian fault, (3) Pre-Balkan fault, (4) Hind Balkan fault, (5) Sozopol fault, (6) Eastern Rezovo fault; **3.** basic tectonic zones: (I) Moesian Platform, (II) Lower Kamchia Depression, (III) Eastern Balkan Range, (IV) Burgas Depression; **4.** location of cores: (a) cores from lake and river terrace sediments, (b) cores from shelf sediments, (c) cores from deep-water sediments, (d) deep-water sounding 380/380A and 381 conducted on *R/V Glomar Challenger*; **5.** continuous seismoacoustic profiles (CSP); **6.** 100 m isobaths; Abbreviation of structures: N1–Northern, A–Aprilska, N2–Nanevska, E–Elizavetinska, SS–Samotino Sea, SE–Samotino East, YG–Yuriiy Gogin.

Table 1. Radiocarbon dates from the Western Black Sea coast.

Lab. N°	Material Dated	Sample Depth (cm)	^{14}C Age (Uncal. BP)	± Errors	Calibrated Years (BC)	Calibrated Years (BP)
MSU-1182	molluscan shells	+110	290	90	AD 1640	310
GIF-5651	peat	–180	1570	90	AD 530	1420
Lu-2382	molluscan shells	–200	2290	50	381	2330
GIF-5652	peat	–230	2440	90	510	2460
MSU-2294	molluscan shells	–450	2670	220	810	2760
MSU-1171	molluscan shells	+300	3120	50	1400	3350
MSU-1292	molluscan shells	+400	3210	220	1490	3440
MSU-1293	molluscan shells	+250	3500	150	1780	3730
GIF-6542	peat	–280	3570	60	1890	3840
GIF-5654	peat	–380	3700	105	2040	3990
Hv-12038	calcareous mud	–535	3760	56	2140	4090
Bln-4114	wood	–400	3980	60	2470	4420
Lu-2241	gyttja	–340	4020	60	2497	4446
Lu-2242	calcareous mud	–360	4080	60	2586	4535
Bln-4111	wood	–400	4170	50	2755	4704
GIF-6534	peat	–360	4330	60	2920	4870
GIF-5655	peat	–470	4400	100	2930	4880
Hv-11488	mud and shell detritus	–400	4730	70	3390	5340
MSU-1236	molluscan shells	+200	4960	100	3720	5670
MSU-1181	molluscan shells	+500	5140	270	3960	5910
GIF-6034	peat	–580	5650	100	4470	6420
GIF-6035	peat	–695	5990	100	4850	6800
Lu-2381	molluscan shells	–420	6170	150	5070	7020
Hv-12839	calcareous mud	–530	6636	163	5520	7470
GIF-6036	peat	–785	6800	110	5630	7580
Hv-11490	gyttja	–1110	7495	95	6360	8310
Hv-19926	peat	–2290	8355	75	7430	9380
Hv-19404	peat	–2930	8580	70	7570	9520
Hv-19403	peat	–3700	9190	95	8130	10,080
Hv-19927	peat	–4210	9945	160	9090	11,040

centimeters to one meter thick, which form clearly defined depositional bodies of coastal or barrier type at a depth of 100–120 m (Khrischev and Shopov 1978). These sediments mark the lowest level of the Neoeuxinian Basin.

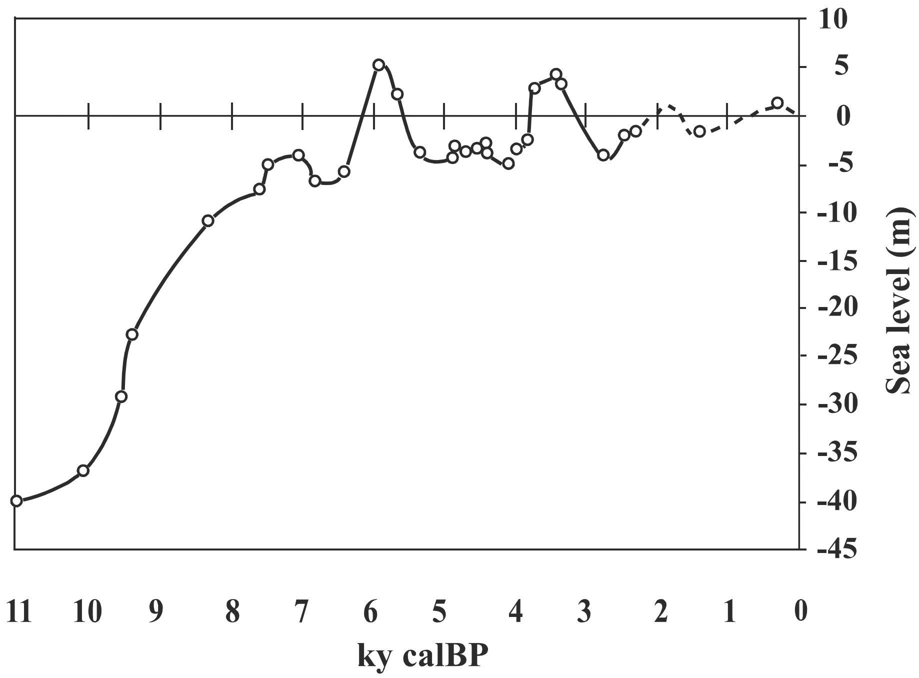

Figure 2. Paleoeustatic curve of Black Sea level oscillations for the last 11,000 years calBP.

In the central shelf zone, Neoeuxinian deposits differ in both thickness and composition. According to CSP data, their thickness reaches 31 m within the Aprilska Structure area, while in other locations, it ranges between 4–12 m. Inward and outward from the central shelf, thickness decreases, eventually wedging out completely. On seismograms, Neoeuxinian layers are distinguished by poor undulatory stratification characteristic of continental deposits and near-shore marine sediments. Their extent is delimited by the 30 m isobath.

In Core C-3 of the Aprilska Structure, Upper Neoeuxinian deposits cover Usunlarian sediments transgressively. They are represented by aleuritic and pellitic silt, uniform, with obscure stratification, and enriched with detritus, which appears as an important component of their lithification. In Core C-2 of the Northern Structure and Core C-3 of the Samotino Sea Structure, the sediments are clayey and sandy. The Upper Neoeuxinian sediments are referable to the Late Pleniglacial (22,000–18,000 calBP), the Late Glacial (18,000–11,500 calBP), and the Early Holocene (11,500–8800 calBP). On the Bulgarian coast, the Late Glacial sediments have been dated to 14,610±200 BP/17,490 calBP and 11,430±330 BP/13,340 calBP (Dimitrov 1982).

Most characteristic of the Upper Neoeuxinian faunal complex are the freshwater Caspian molluscan species *Dreissena rostriformis distincta* Andrus and *Monodacna caspia* Eichwald. The complex also comprises *Dreissena* *polymorpha regularis* Andrus and *Clessiniola variabilis* Eichwald. The presence of the euryhaline Mediterranean mollusc *Cardium edule* Lamarcki Reeve, and *Paphia rugata* (B., D., et D.) should also be noted, because it marks the deposition of sediments at the end of the Neoeuxinian transgression.

Pollen spectra show the domination of herb communities over arboreal vegetation. The Pleniglacial and the stadials of the Late Glacial are clearly represented by the spread of cold steppe, with species of *Artemisia*, Poaceae, Chenopodiaceae, and many other taxa of Asteraceae predominating. Stands of *Pinus*, *Betula*, *Quercus*, and *Corylus* were present among the herb communities. Stands of *Pinus* probably occupied the higher parts of the coastal plateaus, where conditions were more favorable for the growth of trees because of increased atmospheric humidity. According to Niklewski and van Zeist (1970), not just low temperatures but primarily low humidity was a limiting factor in the development of arboreal vegetation during glacial stages. The composition of the pollen spectra suggests a very cold, dry climate during the Younger Dryas stadial of the Late Glacial (13,000–11,500 calBP), recognized everywhere in Europe as an episode of pronounced cooling (Berglund *et al.* 1994; Walker *et al.* 1994). As a result of prevailing cold and dry conditions during the Late Glacial, *Artemisia* and Chenopodiaceae appear abundantly in the terrestrial and lacustrine deposits of southwestern and northern Turkey (Bottema *et al.* 1993–94) as well as northwestern Greece and Central Europe (van Zeist *et al.* 1975). Palynological records from the deep basins of the Marmara Sea also suggest an abundance of *Artemisia* and Chenopodiaceae, indicative of a cold and dry climate during the Late Glacial (Caner and Algan 2002).

Two maxima of *Pinus diploxylon*-type pollen in the Neoeuxinian sediments are probably connected with the spread of *Pinus nigra* forests into lower terrain together with the deciduous taxa *Quercus*, *Carpinus betulus*, *Ulmus*, *Corylus*, *Tilia*, *Betula*, etc., during the Bölling (15,500–14,000 calBP) and Allerød (14,000–13,000 calBP) interstadials of the Late Glacial. Some authors (van Zeist *et al.* 1975) consider that the spread of pine forests during the Late Glacial was probably stimulated by such temporary climatic improvement and especially by the rise in humidity.

Neoeuxinian sediments are also characterized by a unique dinoflagellate assemblage consisting of two species: *Spiniferites cruciformis* Wall and Dale, and *Tectatodinium psilatum* Wall and Dale. This assemblage is similar to the fresh- to brackish water New Euxinic Stage assemblage (Unit 3) of Wall and Dale (1974) and to dinocyst assemblage zones B2 and B1c (Mudie *et al.* 2004). The low diversity is comparable to that described by Wall *et al.* (1973) and Mudie *et al.* (2001, 2002). The assemblage is associated with sea surface salinities less than 7‰ (Deuser 1972; Wall and Dale 1974; Mudie *et al.* 2001), while Chepalyga (2002) indicates a lower salinity (less than 5‰). According to Wall *et al.* (1973), *S. cruciformis* and *T. psilatum* are cool water, low salinity steno-

Table 2. Biostratigraphy and paleoenvironmental reconstruction of Late Pleistocene and Holocene sediments of the Bulgarian Black Sea area.

Chronological scale ky calBP	Section		Alpine Stratigraphic scale	Archaeological chronology (Todorova 1979)	Transgressions and hydrological regime (Fedorov 1982): horizons	Chronological scheme of the Black Sea Shelf (Shopov 1991): regional layers	Chronological scheme of the Black Sea Shelf (Shopov 1991): sublayer	Molluscan fauna	Pollen assemblages	Coastal vegetation	Dinocyst assemblages	Paleoecological reconstructions
	Holocene		Subatlantic	Iron Age	Nymphaean Transgression; Phanagorian Regression	Black Sea layer	New Black Sea sublayer	*Modiolus phaseolinus* *Spisula subtruncata triangulata*	*Quercus, Ulmus, Carpinus, Fagus, Alnus, Salix*	Mixed oak forests and formations of flooded forests along the river valleys	*Lingulodinium machaerophorum, Spiniferites ramosus, Peridinium* sp.	Increase in humidity and decrease in temperature; salinity similar to the present
2.6	Holocene		Subboreal	Bronze Age	New Black Sea Transgression	Black Sea layer	Old Black Sea sublayer	*Mytilus galloprovincialis* *Cardium papillosum* *Cardium exiguum* *Cardium edule* *Hydrobia ventrosa* *Monodacna caspia caspia*	*Quercus, Carpinus betulus, Corylus, Ulmus*	Mixed oak forests; increase of *Carpinus betulus*	*Lingulodinium machaerophorum, Cymatiosphaera globulosa,* maximum of *Lingulodinium machaerophorum*	Slight drying up of the climate
5.8	Holocene		Atlantic	Transitional period; Eneolithic	New Black Sea Transgression; Increase in salinity	Black Sea layer	Old Black Sea sublayer		Maximum of *Quercus, carpinus betulus, Corylus, Ulmus,Tilia*	Balanced mixed oak forests		Climatic optimum: maximum of temperature and humidity; increase in salinity
6.8 7.9 8.8	Holocene		Atlantic	Neo-lithic	First invasion of Mediterranean water	Neoeuxinian	Upper	*Monodacna caspia caspia* *Dreissena rostriformis distincta* *Dreissena polymorpha regularis* *Clessiniola variabilis*				
	Holocene		Boreal		First invasion of Mediterranean water	Neoeuxinian	Upper		*Quercus, Ulmus, Corylus*	Open oak forests	*Tectatodinium psilatum, Spiniferites cruciformis*	Amelioration of the climate
10.0	Holocene		Preboreal		First invasion of Mediterranean water; Regression	Neoeuxinian	Upper		*Quercus, Betula, Corylus, Artemisia,* Chenopodiaceae	Xerophytic herb communities, stands of trees		Amelioration of the climate
11.5	Upper Pleistocene (Würm)	Late Glacial	Younger Dryas		Neoeuxinian Transgression Outflow to the Mediterranean	Neoeuxinian	Upper		*Artemisia,* Chenopodiaceae	Steppe		Cold and dry climate; slight freshening
13.0	Upper Pleistocene (Würm)	Late Glacial	Allerød			Neoeuxinian	Upper		*Pinus, Artemisia,* Chenopodiaceae	Pine forests and steppe		Cool climate, fresh to brackish water
14.0	Upper Pleistocene (Würm)	Late Glacial	Older Dryas			Neoeuxinian	Upper		*Artemisia,* Chenopodiaceae, Poaceae	Steppe		Cold and dry climate; fresh to brackish water
	Upper Pleistocene (Würm)	Late Glacial	Bölling			Neoeuxinian	Upper		*Pinus diploxylon, Artemisia*	Pine forests and steppe		Cool climate, fresh to brackish water
15.5	Upper Pleistocene (Würm)	Pleniglacial				Neoeuxinian	Upper; Lower	Continental sediments	*Artemisia,* Chenopodiaceae	Steppe		Cold and dry climate, fresh to brackish water

haline species that were common in Late Glacial to Early Holocene sediments. *S. cruciformis* also occurred in Late Glacial sediments from Lake Kastoria, northern Greece (Kouli *et al.* 2001) and in the brackish Caspian and Aral Seas (Fabienne *et al.* 2004). The presence of these species is probably connected with the influx of glacial meltwater from the Russian Platform and the Carpathian Mountains (Stanley and Blanpied 1980; Aksu *et al.* 1999; Chepalyga 2002). This melting ice also led to rising global sea level, which has stabilized since 8218 calBP (Davis *et al.* 2003).

The disappearance of cysts of *S. cruciformis* and *T. psilatum* is dated to 6880±240 BP/7650 calBP in western Black Sea shelf sediments (Filipova-Marinova 2003a) and to 6135±75 BP/7010 calBP in sediments from the continental slope (Atanassova 1995). Mudie *et al.* (2001) consider that the virtual disappearance of *S. cruciformis* and *T. psilatum* suggests the inability of these stenohaline taxa to survive an apparently abrupt salinity change to values of 10–12‰ (Deuser 1972) or 18‰ (Wall and Dale 1974). Their demise could also be explained by low temperature tolerance. Though such a depositional pattern could result from an abrupt climatic change, it could also be the consequence of an unconformity in which sediment erosion leaves a hiatus. The latest ^{14}C date for Neoeuxinian sediments is 10,670 BP/12,601 calBP while the earliest date for the Early Holocene is 8080±20 BP/8984 calBP (Dimitrov 1982). According to Khrischev and Georgiev (1991), regional erosion mainly affected sediments deposited near the Pleistocene/Holocene boundary, corresponding to the drastic change in hydrological regime during the rapid sea level rise and establishment of the Mediterranean connection. This hiatus shows that the terrestrial sediments of the shelf were exposed and eroded at a time of low sea level, and that they were then flooded by sea water some time before 6880±240 BP/7650 calBP.

First, single cysts of the euryhaline dinoflagellate *Lingulodinium machaerophorum* (Deflandre and Cookson) Wall and the acritarch *Cymatiosphaera globulosa* Takahashi appeared at about 9630±540 BP/10,890 calBP in deep-water sediments (Filipova *et al.* 1989), and about 9000 BP/10,000 calBP (interpolated date) in the estuary of the Veleka River (Filipova-Marinova 2003b). In shelf sediments, they appeared at about 6800±240 BP/7650 calBP (Filipova-Marinova 2003b). Wall and Dale (1974) consider that outflow from the Black Sea to the Marmara Sea was sufficient to prevent marine organisms from entering in large numbers until about 7000 years ago, when the influence of meltwater had declined substantially. According to Chepalyga (2002), post-glacial water level increase in the Black Sea continued until 9000–8000 BP/ 10,000– 8800 calBP, when the rising world ocean began to impede freshwater outflow from the Black Sea. In some Black Sea cores, oxygen isotopic measurements on calcium carbonate have suggested that Black Sea salinity was about 5–7‰ during the Early Holocene, while ion diffusion studies (cited by

Deuser 1972) have indicated a salinity of ~3.5‰ from 20,000 to 8000 BP/22,000 to 8800 calBP. Most likely, an accidental ingress of Mediterranean sea water through the Bosphorus also occurred, proved by the finds of single specimens of the euryhaline gastropod *Hydrobia ventrosa* (Mtg.) together with the molluscs *Cardium edule* and *Mytilaster lineatus* (Gm. in L.), single cysts of the dinoflagellate *Lingulodinium machaerophorum*, and the acritarch *Cymatiosphaera globulosa* at levels between 9000 BP/10,000 calBP and 8355±75 BP/9380 calBP in the core of the Veleka River (Filipova-Marinova 2003b). The appearance of *Cardium edule* larvae and some other marine species along the Caucasian shelf during the Early Holocene has also been recorded (Chepalyga 2002). This is the difference between the Neoeuxinian sediments related to the Pleniglacial and those related to the Early Holocene. Surface-water circulation dominated by southward export of Black Sea surface waters with little northward penetration of Mediterranean surface water across the Bosphorus since ~10,500 to 6000 BP/12,400 to 6800 calBP is cited by Aksu *et al.* (2002).

At the Early/Middle Holocene boundary, an influx of Mediterranean saltwater and an abrupt change in biota occurred. Old Black Sea sediments are characterized by a heterogeneous, mechanically-formed thanatocoenosis with redeposited Neoeuxinian Caspian molluscs (*Dreissena rostriformis distincta* and *Monodacna caspia pontica*), Mediterranean immigrants (*Cardium edule* and *Hydrobia ventrosa*), marine euryhaline species inhabiting waters with a salinity close to that of today (*Mytilus galloprovincialis* Lam., *Cardium exiguum* Gmel., and *C. papillosum* Poli), and contemporary stenohaline species (*Spisula subtruncata triangulata* Ren., *Pitar rudis* Poli, *Chione gallina* Linné, *Nassa reticulata* Linné, and *Scala communis* Lam.).

Dinocyst assemblages experience a change after 7600 BP/8368 calBP. An abundance of the euryhaline species *Lingulodinium machaerophorum* and the acritarch *Cymatiosphaera globulosa* shows that seawater salinity was higher than it is today. According to some authors (Degens and Hecky 1973; Wall and Dale 1974), this can be attributed to an influx of saline water, a slowing down in the rate of sea level rise, a slow overflow of the shelf areas, and considerable improvement in the climate. Based on evidence, Ryan *et al.* (1997, 2003) proposed that Black Sea salinification did not start until 7150±40 BP/7921 calBP and that post-glacial connection between the Mediterranean and the Black Sea occurred as a catastrophic flood of saline water into the Black Sea that rapidly inundated the basin.

Palynological data allow more detailed paleoecological reconstructions, but such evidence reflecting vegetation dynamics during the Early Holocene are scarce along the western Black Sea shelf because of the unconformity and sediment erosion at the Pleistocene/Holocene boundary. Useful data have come mainly from investigated cores on the continental slope, abyssal plain, and in the estuary of the Veleka River. The Preboreal and Boreal chronozones (11,500–

8800 calBP) are characterized by the spread of xerophytic herb communities and the rapid migration of arboreal taxa (*Pinus*, *Betula*, *Quercus*, and *Ulmus*) that survived the severe conditions of the Late Glacial in nearby refugia in the Strandzha Mountains (Atanassova and Bozilova 1992; Shopov *et al.* 1992; Filipova-Marinova 2003b). Mountains of the Balkan Peninsula are considered to have been favorable sites for tree survival by many authors (Beug 1975; Huntley and Birks 1983; Bennett *et al.* 1991; Birks and Line 1993). The first slight increase in deciduous arboreal pollen is registered in two places: sapropel muds from the deep-water part of the western Black Sea dating to 9630±520 BP/10,890 calBP (Filipova *et al.* 1989) and in the estuary of the Veleka River and dating to 9945±160 BP/11,040 calBP. Climate warming and increase of arboreal taxa from 10,737±315 BP/12,670 calBP is also suggested by Shimkus *et al.* (1977) and Komarov (1983). During the Preboreal, the distribution of arboreal taxa was probably stimulated by temperature rise and persistence of the continental climate. Scattered stands of *Quercus*, *Ulmus*, *Tilia*, *Fraxinus*, and *Acer* occurred. Climatic improvement during the Boreal caused a gradual replacement of the xerophytic herb communities by trees and formation of open oak forests. According to Kutzbach and Webb (1993), winter insolation and changes in mean winter temperature of the coldest month have an effect on the decline in residual LGM ice cover occurring up to 7500 BP/8218 calBP.

During the Atlantic chronozone (8800–5800 calBP), optimum climatic conditions (high temperature and humidity) favored extensive spreading of mixed oak forests, the main components of which were *Quercus*, *Ulmus*, *Tilia*, *Fraxinus*, and *Acer*. The most characteristic feature of the coastal zone is the increase of *Carpinus betulus* at 5770±105 BP/6560 calBP in the area of Arkutino Lake (Bozilova and Beug 1992) and at 5650±100 BP/6420 calBP in the area of Shabla-Ezeretz Lake (Filipova 1985). In addition to being part of the mixed oak forests, *Carpinus betulus* also formed separate communities at higher altitudes and on northern slopes. In the eastern Stara Planina Mountains, hornbeam formed a separate belt (Filipovitch 1987). The areas occupied by *Fagus* were also enlarged. During that period, steppe vegetation was preserved only in the coastal area of South Dobruja due to the insignificant increase in humidity, which could not compensate for the rise in temperature and the desiccating effect of the wind. Even during the climatic optimum, only steppe forest vegetation existed in this region, while in the nearby area of Srebarna Lake, there were mixed oak forests (Lazarova and Bozilova 2001).

According to the archaeological chronology (Todorova 1979), the Late Atlantic could be referred to the Eneolithic (4450–4200 calBC). Human impact was significant and influenced the natural forest vegetation along the western Black Sea coast (Bozilova and Filipova 1991; Bozilova and Beug 1994; Filipova-Marinova and Bozilova 2003). This activity is reflected in the pollen diagrams of the coastal lakes as a decrease in arboreal taxa and the presence of cultivated cereals, such as *Triticum* and *Hordeum*, and of some weeds and ruderal plants, such as *Plantago lanceolata*, *Polygonum aviculare*, *Urtica*, and *Centaurea cyanus*. There is no doubt that the degradation of the mixed oak forests at this time is connected with agriculture as well as population increase.

Changes in forest composition occur during the Subboreal chronozone (5800–2600 calBP). The most characteristic feature is the significant presence of *Carpinus betulus* and the decrease of *Ulmus*. *Carpinus orientalis* increases, marking the beginning of destructive human activities between 3200 and 2500 calBC during the Early Bronze Age (Bozilova and Beug 1994; Bozilova and Tonkov 1998; Filipova-Marinova and Bozilova 2003).

A humidity increase is observed at the beginning of the Subatlantic chronozone (2600–0 calBP). This is seen in the increase of *Alnus*, *Salix*, and *Fraxinus excelsior* pollen and the flooding of forests along the mouths of rivers running into the Black Sea, dated to 3185±100 BP/3380 calBP (Bozilova and Beug 1992). Beech (*Fagus*) forests spread into the higher regions and into wet ravines. This last stage of vegetation development is connected with the formation of modern plant communities along the coastal zone.

4.2 Sea Level Change and Coastal Migration

Chepalyga (1984) dates the beginning of the Holocene by ^{14}C from 10,700 to 9700 BP/12,630 to 10,950 calBP. Early in the Holocene, the Black Sea transgression marked a new stage in the geological history of the Pontic basin, transforming its paleoenvironment. The inflow of Mediterranean water raised Black Sea salinity. Many authors identified the beginning of the Holocene by the first appearance of euryhaline Mediterranean molluscan and dinoflagellate species. According to Degens and Hecky (1973), salinity began rising in the Black Sea before 9300 BP/8275 calBP, and the influx of Mediterranean water dated to 7300 BP/8000 calBP.

Ryan *et al.* (1997, 2003) suggest that the marine invasion occurred as an abrupt drowning, submerging the exposed continental shelf to a depth of more than 100 m by 7150±40 BP/7925 calBP. Ballard *et al.* (2000) revealed a paleoshoreline at a depth of 155 m on the southern Black Sea shelf, indicating the occurrence of a flood in the interval 7460–6820 BP/8183–7616 calBP. Evidence of such a rapid transgression has also been found using high-resolution profiles from the Bosphorus Strait and its exit to the Black Sea shelf at a depth of –105 m (Demirbağ *et al.* 1999). Submergence of the southern shelf area beneath the rising sea level is indicated between 8590±145 and 7415±115 BP/9500–8160 calBP (Algan *et al.* 2002). This discrepancy implies that the beginning of the Holocene in the Black Sea is several thousand years younger than the conventional Late Pleistocene/Holocene boundary of 10,500 BP/12,400 calBP (INQUA decision of 1969). Most biostratigraphers dealing with continental deposits in Europe keep to this boundary. It is also accepted for the territory of Bulgaria and the Black Sea coast (Bozilova 1982). In accordance with the radiocarbon dating of the Black Sea shelf sediments, this boundary ranges from 8620±70 to 7070±90 BP/9530–7840 calBP. During the Neoeuxinian, the Black Sea was half isolated from the system of the world ocean, and the effects of global climatic events took place with a certain delay (Khrischev and Shopov 1978; Nikolaev *et al.* 1980; Dimitrov 1982; Shopov 1991). But Shimkus *et al.* (1977) resolve this discrepancy by establishing the Black Sea Holocene as synchronous with the global chronology (Ryan *et al.* 2003). The absolute dating of the base of the Early Holocene deep-water sediments is 10,737±315 BP/12,670 calBP and the upper part of these sediments is dated at 7480±540 BP/8250 calBP. This freshwater Early Holocene Stage (HL_I) corresponds to the Upper Neoeuxinian Stage (Shimkus *et al.* 1977). A change of molluscan and dinocyst assemblages in the western Black Sea shelf occurred after 7600 BP/8370 calBP.

At the end of the Late Glacial and beginning of the Holocene, sea level in the world ocean was rising continuously, eventually impeding freshwater outflow from the Black Sea. Yet, Mediterranean water was unable to penetrate substantially into the Black Sea. With the post-glacial transgression, the Black Sea basin simultaneously expanded to the 30–34 m isobath on the western shelf, in the process flooding riverine lowlands (Aksu *et al.* 2002; Chepalyga 2002). This is attested at depths between 40.1 and 24.9 m in a core from the firth-estuary of the Veleka River. Lagoon-estuarine sediments characterized by low organic content and the sporadic occurrence of molluscan fauna and dinoflagellate cysts were deposited between 9945±160 and 8355±75 BP/11,040–9380 calBP. The composition of the molluscan fauna, with a predominance of the brackish water species *Dreissena polymorpha* (Andr.) Pal. and the presence of the stenohaline dinoflagellate *Spiniferites cruciformis*, suggests comparatively low salinity, confirming an Early Holocene age of the sediments. During that time, an accidental ingress of the sea into the firth occurred, as indicated by the finding of single specimens of the euryhaline gastropod species *Hydrobia ventrosa* together with the molluscan species *Cardium edule*, *Mytilaster lineatus*, as well as euryhaline dinoflagellate cysts of *Lingulodinium machaerophorum* and the acritarch *Cymatiosphaera globulosa* at some levels (Petzold 1995; Filipova-Marinova 2003a).

After 8355±75 BP/9380 calBP and until about 7000 BP/7790 calBP (interpolated date), sediments were rich in organic material and very high in carbonate content. The latter originates from the high concentration of molluscs. These sediments contain a mixture of euryhaline Mediterranean molluscs *Hydrobia ventrosa* and the relict Caspian fresh- to brackish water *Monodacna caspia caspia* Eichwald. The composition of this molluscan assemblage reflects a gradual salinization that would allow the joint existence of Caspian and euryhaline Mediterranean species. Against this background, the fluctuations of freshwater with predominance of *Dreissena polymorpha* are explicable. The presence of the gastropod species *Valvata* sp. indet. emphasizes again the freshwater character of some sediments. In other layers of the sequence, the euryhaline Mediterranean species *Abra alba* Wood. and *Cardium edule* predominate. Typical marine euryhaline dinoflagellate cysts of *Lingulodinium machaerophorum* and the acritarch *Cymatiosphaera globulosa* are not identified in all studied layers, thereby giving reason to suspect an oscillating connection of the firth with the sea.

The rise in sea level before 8130±170 BP/8990 calBP (Svitoch *et al.* 1997) was particularly sharp. Radiocarbon dating was performed on shells of *Chione gallina*, *Donax truculus* (L.), and *Paphia rugata* from lagoon-estuarine type sediments at the base of a section in a sand-strip situated 5 km to the north of the town of Nessebar.

A steady mutual water-mass exchange between the firth and the sea was established after 6880±240 BP/7650 calBP. Sandy deposits with typical molluscan fauna of *Mytilus galloprovincialis* and *Hydrobia ventrosa* were found. In addition, *Mytilaster lineatus*, *Corbula maeotica* (Mil.), and *Cardium edule* were present. The composition and the considerably more widespread occurrence of this assemblage, compared to the others, reflects conditions of an expanding transgression and salinization of the sea in the mid-Holocene, which eventually brings about the final disappearance of Caspian species as the euryhaline marine species rise to dominance. The typical euryhaline dinoflagellate cysts *Lingulodinium machaerophorum* and acritarch *Cymatiosphaera globulosa* are identified in all layers. Most probably, this connection is a result of the influx of Mediterranean waters and the rise of sea level after 7000 BP/7790 calBP (Wall and Dale 1974; Chepalyga 1984; Ryan *et al.* 1997, 2003).

Data from pollen and mollusc analysis show that coastal lake formation is connected with the Holocene transgression, which is dated at Lake Varna before 7495±95 BP/8310 calBP, at Lake Durankulak at about 6170±150 BP/7020 calBP (Bozilova and Tonkov 1998), and at Lake Shabla-Ezeretz at 6800±110 BP/7580 calBP (Filipova 1985).

The presence of peat layers and changes in gastropod composition in these lake deposits provide evidence for periodic isolation and reconnection of the lakes with the sea (Shopov and Yankova 1987). With a generally rising sea level during the Atlantic, certain relative falls can also be observed at 5990±100 BP/6800 calBP in Lake Shabla-Ezeretz. Pollen evidence shows a considerable percentage increase in the typically marshy coenosis of *Phragmites australis* Grav. and *Typha latifolia* L., together with an increase in cultural indicators, such as cereals, weeds, and ruderals (Filipova 1985; Filipova-Marinova and Bozilova 1990). A cultural layer revealed at Lake Varna at 4.3–5 m in depth and dated to 5390±65 BP/6190 calBP also confirms this lowering of the water level

(Bozilova 1986). Evidently, settlements of the Late Eneolithic existed at the sea side. Some remains were revealed at a depth of 5–7 m in the Varna-Beloslav estuary, north of Cape Atia (near Sozopol), in the harbor of Sozopol, and Kiten (former Ourdoviza, near Sozopol) (Bozilova and Ivanov 1985; Draganov 1995; Lazarov 1996; Todorova 2002; Filipova-Marinova and Bozilova 2003). A series of radiocarbon dates between 5595±100 and 5475±50 BP (6400–6287 calBP/ 4840–4370 calBC) are in conformity with the radiocarbon chronology for Bulgarian prehistory (Görsdorf and Boyadziev 1997).

The rise in sea level at about 5140±270 BP/5910 calBP was particularly sharp. The date indicates that an enlargement of coastal lakes probably took place after the Eneolithic. More support for this suggestion is in the Eneolithic burial place around Lake Durankulak, where some of the graves are today underwater (Dimov 1990; Marinova 2003). The present estuarine-bay of Varna-Beloslav turned into a large embayment of the sea, reaching the mouth of the rivers Provadiyska and Devnya. Fragments of a sea terrace of this period survive at an elevation of 5 m (Krastev *et al.* 1990). Some fossils have been revealed in a terrace close to the town of Beloslav. Biostratigraphic analyses show that at the beginning and end of that event, the natural reservoir was considerably fresher, with an abundance of *Theodoxus pallasi* (Lindh.) and *Dreissena polymorpha*, but during the maximum of the transgression, its salinity was close to that of the sea, as is shown by the presence of halophilous and thermophilous Mediterranean molluscan species such as *Ostrea edulis* (L.), *Mytilaster lineatus*, and others.

Sea level rise from 5140±270 to 4960±100 BP/5910–5670 calBP is also marked in the coastal terraces at 2–5 m between the town of Kavarna and Lake Tuzlata (Krastev *et al.* 1990) and the mid-section of the sand-strip south of Nessebar, where numerous shells of Mediterranean molluscs were found. Late Eneolithic settlements were flooded and destroyed as a result of this transgression or a sinking of the land south of Burgas Lake. A 700–800 year hiatus between the Eneolithic and Early Bronze Age settlements can be seen in numerous cores (Lazarov 1996; Boyadziev 1998). There are almost no anthropogenic indicators apart from single pollen grains of *Cerealia*-type, *Plantago lanceolata*, and *Polygonum aviculare*. This interval could be correlated with the Transitional Period (4200–3200 calBC) and connected with a decline in human occupation (Todorova 1979). Hay (1988) suggested that the highest sea level after 5100 BP/5910 calBP was a result of the melting of the ice-caps, and according to Badyukov and Kaplin (1979), sea level exceeded today's level by 2–3 m.

A temporary lowering of sea level occurred between 4400±100 and 3760±56 BP/4880–4090 calBP. The connection between some coastal lakes and the sea was broken, most probably as a result of increasing drought (Bozilova and Filipova 1986). The complete drainage of Lake Durankulak is shown by the presence of a peat layer and the absence of diatoms, which has been ^{14}C dated to 4080±60 and 4020±60 BP/4540–4450 calBP (Bozilova and Tonkov 1998). Similarly dry conditions have been detected for the period 4200–3700 BP/4720–3990 calBP in the northern Black Sea area (Kremenetski *et al.* 1999), and archaeological remains excavated in the western coastal area also provide evidence for a lowering of sea level during that period.

Human activity along the western Black Sea coast resumed during the Early Bronze Age, as the sea withdrew and thus made the Bay of Sozopol habitable again (Filipova-Marinova and Bozilova 2003). During the later Early Bronze Age, some settlements were established above remains of Late Eneolithic settlements (Sozopol and Varna), while others arose in new locations (Atia, Ropotamo, Kiten, and Akhtopol). Archaeological finds recovered at a distance of 800–900 m from the present coastline in the ancient valley of the river Karaagatch at a depth of 2.5 m have been dated by ^{14}C to 4160–4000 BP/4620–4410 calBP, i.e., the second phase of the Early Bronze Age (Porozhanov 1991; Draganov 1995). The settlement of Kiten provides the longest Early Bronze Age oak chronology from the Balkans and the Aegean. Eighty-three cross-sections of oak pilings collected from the submerged Early Bronze Age sites of Kiten, Burgas, Primorsko, and Sozopol on the western Black Sea coast of Bulgaria were dated. The radiocarbon ages range from 4246±19 to 4122±23 BP/4832–4605 calBP (Kuniholm *et al.* 1998, this volume). According to Kuniholm *et al.* (1998), a combination of sea-level change and tectonic subsidence is thought to explain why apparent habitation sites are now 5–10 m underwater. The increase in anthropogenic impact occurs simultaneously with an observed decline in forests.

The next rise in sea level is shown by both the abandonment of the prehistoric Early Bronze Age settlements and the considerable presence of cysts of the typical association of the euryhaline dinoflagellate *Lingulodinium machaerophorum* and the acritarch *Cymathiosphaera globulosa* in marine sediments after 3570±60 BP/3840 calBP. This rise in sea level is marked by the deposition of peat with molluscan detritus in the coastal lakes. Terraces at an elevation of 3 m were formed east of the town of Balchik at Lake Tuzlata, where shells of the marine mollusc *Ostrea edulis* have been dated to 3120±50 BP/3350 calBP. Similar marine deposits can be observed south of Cape Shabla on an abrasion terrace 4 m high, as well as in cross-sections through the estuaries of the rivers Kamchia, Fandakliyska, and Veleka, where they are dated to 3500±150 and 3120±50 BP/3840 and 3350 calBP (Krastev *et al.* 1990).

The beginning of the Subatlantic, 2500 BP/2600 calBP, was marked by colder global temperatures (Hay *et al.* 1991). The low level of the Black Sea between 3200 and 1950 calBP may have been a result of lower rates of freshwater inflow during the Phanagorian Regression from 3000 to 2200 BP/3200–2300 calBP (Chepalyga 1984). The drop in Black Sea level varies in different reports from 5 to 15 m (Ostrovsky *et al.* 1977; Shilik 1977; Fedorov 1978, 1982; Chepalyga 1984; Balabanov and Izmailov 1988), while coeval sea level in the world ocean ranges from 0 to 2 m below present (Mörner 1971). Along the Bulgarian Black Sea shelf, the effects of the regression are reflected in the re-exposure of the existing firths 1.5 km into the sea and a 10 m decrease in sea level (Khrischev *et al.* 1980). A marked lowering of the water level in coastal lakes and the accumulation of a thick peat layer also occurred. Lithified sediments found 5–10 m below present sea level in the old sand spits of Alepu and Chimovo (near the Nessebar) further indicate the low position of sea level (Khrischev *et al.* 1980). From this period as well, submerged ancient settlements are found, everywhere at a depth of 4–6 m, with remnants of harbor basins at depths of 8–12 m (Lazarov 1975). Historical documentation indicates that the peak of the Phanagorian Regression was reached around the middle of the fifth century BC/2500 BP (Orachev 1990). Deposits of the time found in cross-sections and cores are characterized by a mixed lithological composition of aleurites, sands, and gravel of continental alluvial and lake phases. They occur on top of the Holocene sediments and contain rare shells of terrestrial molluscs and plant remains.

High precipitation levels in Europe generally coincided with the Nymphaean Transgession. Archaeological evidence suggests higher rainfall and comparatively moist conditions in Europe between 2000 and 1400 BP/1950–1300 calBP (Bouzek 1983; Hay *et al.* 1991). The transgression flooded ports and ancient towns along the coast and exceeded recent sea level by 1–2 m. The firth of Mandra was open for watercraft, and the ports of Skafidah and Deultum were founded along its shore. Nymphaean Sea sediments are represented by well sorted sands built by high coastal swells, sand spits, and the rear parts of the beaches. The Nymphaean Sea terrace on the Bulgarian coast is clearly marked by the beach containing New Black Sea and modern deposits (Fedorov 1978).

After 800 BP, European climate gradually became cooler and drier. In the Alps, glaciers expanded at 220–120 BP (Hay *et al.* 1991). During the Middle Ages, sea level fell. According to Herman Shkorpil, during the battle between Vladislav Varnentchik and the Turks in 1444, the land between Varna and the residential district of Asparukhovo was marshy and difficult to pass. This confirms the occurrence of the Korsunian Regression. It was followed by the Lasian Transgression, which was still remembered by people of the middle of the last century. The medieval settlement of Skafidah was flooded in the course of that transgression. Its remnants have been observed under the waters of Lake Mandra, near the present village of Skeff (Lazarov 1975). Formation of the 2 m high sand bar at the mouth of the Batova River is dated by the complex of molluscs *Donax trunculus* and *Chione gallina* to 290±90 BP/310 calBP (Krastev *et al.* 1990).

According to the records of oceanographers in Varna and Burgas, the average eustatic rise along the Bulgarian coast of the Black Sea for the period

1928 to 1998 is about 2.5–2.7 mm per year (Markov *et al.*1991; Vesselinov and Mungov 1998).

5. CONCLUSIONS

All paleontological data indicate arid climatic conditions with steppe vegetation during the Pleniglacial and Late Glacial stadials. A slight increase in moisture during the interstadials of the Late Glacial is reflected in the spread of pine forests. Increase in temperature favored the expansion of deciduous trees during the Preboreal and the formation of open oak forest during the Boreal. At 8800 calBP, optimum climatic conditions (increase in temperature and higher moisture) led to the maximum extent of balanced, mixed oak forests. Increased humidity and decreased temperature are reflected in the Subatlantic by the formation of flooded forests along the river valleys.

Based on existing radiocarbon dates for the Bulgarian coast, a sea-level curve for the Holocene is here suggested. The Holocene transgression reveals a gradual and cyclical character. On three occasions, the Black Sea exceeded present sea level: up to 5 m at 5910 calBP, up to 4 m at 3730 calBP, and up to 1 m at 310 calBP.

The first appearance of single euryhaline specimens of the gastropod *Hydrobia ventrosa*, the molluscan species *Cardium edule* and *Mytilaster lineatus*, and the dinoflagellate cysts of *Lingulodinium machaerophorum* with the acritarch *Cymatiosphaera globulosa* is established after 11,040 calBP in sediments of the Veleka River estuary on the southeastern Black Sea coast.

A steady water-mass exchange between the Black Sea and the Mediterranean dates to 7650 calBP, as reflected by the domination of the typical euryhaline molluscan fauna of *Mytilus galloprovincialis* and *Hydrobia ventrosa* together with dinoflagellate cysts of *Lingulodinium machaerophorum* and the acritarch *Cymatiosphaera globulosa*. A modern transgression is currently being observed, with sea level rising slowly at 2.5 to 2.7 mm/yr.

Eustatic sea level fluctuation has been a controlling factor in the evolution of the Bulgarian Black Sea coast and shelf. The effect of tectonic movements is seen at the regional scale.

The western Black Sea shelf and continental slope suffered a regional stratigraphic hiatus at the Pleistocene/Holocene boundary due to an unconformity and sediment erosion from 11,590±240 to 6880±260 BP/13,520–7650 calBP.

The submergence of Late Eneolithic settlements (4450–4200 calBC) and Early Bronze Age settlements (3200–2500 calBC) as well as the interruption of human activity during the transgressive stages after 5140 and 3730 calBP have been traced.

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20

DENDROCHRONOLOGY OF SUBMERGED BULGARIAN SITES

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Abstract: During the late 1980s, 83 timber pilings off the Black Sea coast of Bulgaria at Kiten were sampled for dendrochronological research. According to the excavators, the pieces derived from house foundations associated with a now submerged Early Bronze Age habitation site. The wood, mostly oak, formed a 285-year tree-ring sequence, currently the longest Early Bronze Age oak chronology from the Balkans, with at least four, possibly five, building phases identified. The four major phases are represented by at least 10 specimens each, and all construction episodes span a 64-year period. A limited number of specimens exhibited dates falling between these phases, suggesting that maintenance activities involving wood replacement were ongoing between phases. Nine 15-year-long ring sequences were wiggle-matched to the decadal radiocarbon calibration curve of Stuiver and Becker. With an error of ±10 years, the resulting dates place the cutting of the trees for phase 1 at *ca.* 2778 BC and for the possible phase 5 at *ca.* 2715 BC. Kiten is only one of several submerged sites discovered by Bulgarian underwater archaeologists. In a discussion of flooding in the Black Sea basin, the submergence of a site occupied during the early 3rd millennium BC requires some explanation, considering that the date proposed by Ryan and Pitman for their inundation lies in the 6th millennium. An even earlier set of submerged timbers from Eneolithic Sozopol has now been dated and is here reported.

Keywords: dendrochronology, sea-level change, Black Sea, Bulgaria, tree-ring dating

1. INTRODUCTION

Tree-ring dating is deceptively simple. Some species of trees add annual growth increments in recognizable sequences. When trees in a specific

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climatic region such as the Black Sea basin are similarly affected by yearly changes in the climate, their rings can be matched or 'crossdated' with one another so that a given ring can be assigned to a specific calendar year (see Kuniholm and Striker 1987; Kuniholm 1996a). Sometimes a felling time within a year can be identified (Kuniholm 2001). Dendrochronology is the only archaeometric dating technique that furnishes this kind of annual or sub-annual resolution. The method works only with species having clear, annual growth rings, but it works with wood that may be dry, wet (bog sites or shipwrecks), or burned (charcoal).

Crossdating is the fundamental principle on which all dendrochronology is based, and therefore, it deserves further explanation. The researcher has to be assured that rings from two or more specimens were formed in the same year. Neither simple ring-counts nor a single pattern of co-variation in ring-width, i.e., a 'signature', are sufficient (see Kuniholm 2001 for illustrations). In order to avoid the possibility of accidental, spurious 'matches', dendrochronologists prefer to compare specimens possessing at least 100 rings and multiple signatures rather than shorter-lived ones, which may not preserve enough signatures to guarantee the fit. These ring-patterns may be generated by a wide variety of causes, some of which yield a true signal, others merely noise.

The ring-patterns that are most usually crossdatable are the trees' synchronized response to some mutually-experienced climatic stimulus (Hughes *et al.*2001). In some regions, it is principally rainfall or lack of it; in others, it is temperature; in yet others, it is some combination of the two. Around the Black Sea, precipitation from April to June seems to be the limiting factor (Hughes *et al.* 2001). This stimulus-and-response is therefore specific to a climatic region, and in this paper, the region in question runs from Georgia in the east along the northern Turkish coast to Greece and the former Yugoslavia in the west. We do not yet know whether we will obtain crossdating some day with trees in Romania and Ukraine. The climatic boundaries for dendrochronological crossdating have been best determined, in practice, by trial and error.

When continuous tree-ring chronologies do not exist from the present to antiquity, alternative methods of placement are needed. One of the best methods is radiocarbon wiggle-matching, which will be elaborated on below.

1.1 Preliminary Digression: Çatak in Southeastern Turkey and the Tigris Trout of James Prosek

Since students of the Black Sea may not be regular readers of the sports section in the *New York Times*, we draw the reader's attention to an article by James Prosek entitled "Seeking God's fish, a/k/a Tigris trout," which appeared on Sunday, March 23, 2003. Prosek caught a trout–an image of which he kindly provided for the Columbia University conference–in the headwaters of the Tigris

River at Çatak, about 30 km east of Lake Van. The genetic ancestry of this trout traces back to the end of the last glaciation, or about 13,000 years ago, when the (then) freshwater lakes we now know as the Caspian and the Black Seas were connected. It is quite unthinkable that the trout, and its cousins in the headwaters of the Euphrates, migrated upriver from the Persian Gulf.

Prosek has now published *Trout of the World* (2003) in which this trout and its relatives are illustrated and discussed. The very simple point to be made here is that we all need to remember that modern coastlines and landforms are not necessarily those of remote antiquity. Thus, it is not at all outrageous to think of Early Bronze Age habitations ten meters under today's sea level.

1.2 Secondary Digression: Troy of Early Bronze Age I and Wiggle-matched Dates from the Schliemann Trench

This information was published in *Studia Troica III* (Korfmann and Kromer 1993) and again in *Acta Archaeologica* (Kuniholm 1996b). Our second point is to remind readers, for whom "wiggle-matching" may be a novel concept, of a remarkable V-shaped radiocarbon anomaly in Anatolian pines in the 29th century BC, which makes the placement of the Troy I chronology, with its radiocarbon dates forming a coeval and similarly remarkable V-shaped anomaly, a decade and a half before 2700 BC all the more secure.

2. SUBMERGED COASTAL SITES IN BULGARIA—EARLY BRONZE AGE KITEN

In 1998, we published the results of measurements from 85 oak and boxwood pilings collected in 1988 and 1989 from Burgas, Sozopol, Primorsko, and Kiten, about an hour's drive north of the Turkish frontier. The geologists at this meeting showed images of the relatively shallow shelf which runs along the western edge of the Black Sea and on which the Kiten site and other similar sites were built. We were told at the time of collection by a team of Bulgarian archaeologists and divers that the pilings came from Early Bronze Age house foundations, now submerged about 8–10 meters beneath the surface of the Black Sea. The tree-rings formed a 285-year sequence encompassing four, possibly five, building phases. Wiggle-matching at the Heidelberg radiocarbon laboratory of nine 15-year-long ring sequences produced the results in Table 1, which were published in the James Harvey Gaul memorial volume (Kuniholm *et al.* 1998).

We emphasize that the accuracy of this wiggle-matched placement in time is provided by the same 29th century BC V-shaped anomaly in the radiocarbon calibration curve that has already been noticed in Trojan pines. Here, it is repeated in the Bulgarian oaks–see Kuniholm (1996b:332, Figure 8). The archaeological context at Kiten was described to us by Professor Henrieta Todorova as "the forming phases of the Černa Voda-Ezerovo culture / Ezero VI-IX = 2900–2700 Cal. BC," so the dates are in accord with the material assemblage, except that here we are able to be much more specific.

Table 1. Phases and dates for Kiten pilings (Kuniholm *et al.* 1998).

PHASE	KITEN RELATIVE DATE	RADIOCARBON DATE*
1	trees cut *ca.* 1073	= 2778 BC±10 (bark present)
	7 year interval	
2	trees cut *ca.* 1080	= 2771 BC±10 (very few exterior rings missing)
	20 year interval	
3	trees cut *ca.* 1100	= 2751 BC±10 (waney edge)
	14 year interval	
4	trees cut *ca.* 1114	= 2737 BC±10 (waney edge)
	22 year interval	
5?	trees cut *ca.* 1136	= 2715 BC±10 (very few exterior rings missing)

*NOTE: These radiocarbon dates were based on the 1993 calibration curve (Stuiver and Becker 1993). The now current 1998 calibration curve (Stuiver *et al.* 1998; the 2004 version is not yet available) yields a date one year earlier (i.e., 2716 BC) for the possible Phase 5.

Each of the dendrochronological phases at Kiten consists of at least ten trees, all ending in the same terminal year. Other specimens yielded dates between these phases, which suggests that some maintenance activities involving wood replacement were continuous. We were promised plans of the site by the excavator Kalin Porozhanov but as yet have not seen any. It would be instructive to study the spatial relationship to determine whether people were building and rebuilding as the water level rose, or whether the recovered wood assemblage is simply a series of pilings from four separate and approximately contemporary buildings. The small building model shown to us at Kiten does not make sense for a marshy environment because the house foundation appeared to be dug into the soil, making for a rather wet cellar. The function of the pilings on either side of the foundation is unclear to us.

Our dendrochronological work at Kiten does not exist in isolation. The Kiten tree-ring chronology crossdates with the Early Bronze Age site of Demircihüyük, with the latter ending at Kiten Relative Date 1146, or, using the wiggle-matched dates, 2705 BC±10 (Kuniholm 1987; Kuniholm *et al.* 1998).

The message from Kiten, then, is that throughout the 28th century BC, and as late as 2715 BC, people were able to build and live on the western shore of the Black Sea, even if somewhat damply. This land surface is currently ten meters below the surface of the Black Sea. We need the help of the geologists to explain why there was such a change: how much due to tectonic subsidence, how much due to rising sea level.

3. SUBMERGED COASTAL SITES IN BULGARIA—ENEOLITHIC SOZOPOL

We have a second collection of timber pilings, mostly oak, retrieved in 1988 by Aleksandar Durman and Hristina Angelova from an Eneolithic site (Quadrant D) about ten meters under water in Sozopol harbor. Again, we have not seen copies of the plans. The Sozopol oak forms a 224-year tree-ring chronology, and radiocarbon determinations made at the AMS facility at the University of Arizona place the end-date at 4140 BC±19 (calibrated). There is no phasing as we saw at Kiten. But again, it is worth noting that as of the mid-42nd century BC—or 14 centuries before Kiten—people were building and living next to a Black Sea that was much lower than it is today.

Our intuitive feeling is that although we are now looking at a salty replacement for what was once a freshwater lake (the kind of lake in which the Tigris trout's ancestors might have swum), the change did not come as a single enormous splash, thereby filling the Black Sea with salt water up to its present brim. The last ten meters of sea-level change took place some time between 2700 BC and today. We leave to the geologists the explanation and estimation of what part of this is due to absolute sea-level change and what part to land movement.

4. FUTURE WORK

At the time of our visits to Bulgaria in 1988 and 1989, we were told by the divers of an off-shore "forest," from which it was alleged that timbers could be produced for dendrochronological analysis. Perhaps inadequate knowledge of Russian made us misunderstand what was being said, and perhaps Dr. Angelova might be able to shed further light on this. Certainly, if any such material could be recovered and analyzed, this would be additional useful information for conditions along the Black Sea coastline at specific times in antiquity. The same applies to anything formerly terrestrial recovered by Dr. Ballard's submersibles. Shipwrecks, of course, do not provide this kind of information other than that the wreck site was at one time navigable water.

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THE NEOLITHIZATION OF THE NORTH PONTIC AREA AND THE BALKANS IN THE CONTEXT OF THE BLACK SEA FLOODS

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Abstract: The neolithization of Southeastern Europe, including the Balkans and the North Pontic area, took place in an environmental context. This paper examines specifically its connection to the effect of sea-level changes in the Black Sea. There is no evidence of catastrophic flooding that might have caused large-scale migrations of early farming populations. At the same time, it is apparent that the major stages in the spread of farming groups coincided with sea-level rises that might have been caused by increased precipitation in the river basins.

Keywords: Southeastern Europe, Holocene, neolithization, climate change, Black Sea level fluctuations, cultural diffusion, migrations

1. INTRODUCTION

In the context of European prehistory, the transition from the Mesolithic to the Neolithic, or from hunting and gathering to agriculture, is viewed as one of the most important episodes in the process that paved the European path to civilization.

The mechanism by which the Neolithic spread in Europe is represented by two basic models: migration and cultural diffusion. The neolithization model based on direct migrations is omnipresent in the work of Childe, who proposed that the European Neolithic was created by "farmers spreading from the southern cradle of cereals" (Childe 1958:110). More recently, this idea has taken the form of demic expansion, i.e., a 'wave of advance' (Ammerman and Cavalli-Sforza

V. Yanko-Hombach et al. (eds.), The Black Sea Flood Question, 489-514.

1973). This model was further substantiated by genetic markers (Menozzi *et al.* 1978; Cavalli-Sforza *et al.* 1994) that have been interpreted as indicating the movement of farming populations from Anatolia into Europe. Renfrew (1987, 1996) linked the dispersal of farming with the spread of Indo-European speech.

According to the migrationist concept, populations from Anatolia moved into the western and northern areas of the Aegean littoral (Nea Nikomedeia, Sesklo, and other sites), settled Thrace, and spread farther north into the Middle Danube and Central Transylvania, importing the first domesticated plants and animals, as well as agricultural skills (Demoule 1993; Srejovič 1993; Demoule 1994; Parzinger and Özdoğan 1995; van Andel and Runnels 1995; Lazarovici 1996; Titov 1996; Ursulescu 2000; and others).

There are several variations on the migrationist model that range from a direct colonization of hitherto unpopulated areas, to the annihilation of previous Mesolithic groups (Childe 1958; Ammerman and Cavalli-Sforza 1973) and élite dominance (Renfrew 1987). Zilhão (1993, 2001) views neolithization as a 'leapfrogging colonization' by small seafaring groups along the Mediterranean coast.

An alternative approach views neolithization as an adoption of agriculture by indigenous hunter-gatherers through the diffusion of cultural and economic novelties by means of intermarriages, assimilation, and borrowing (Tilley 1994; Whittle 1996; Thomas 1996, 2003).

A unifying position advocated by Zvelebil (1986, 1996) distinguishes three phases in the transition to agriculture: availability, substitution, and colonization. Each phase operated in the broader context of an 'agricultural frontier' (see also Zvelebil and Lillie 2000). The 'individual frontier mobility' concept relates neolithization to 'small-scale' contacts between hunter-gatherers and farmers at the level of individuals and small groups linked by kinship. Several writers (Gronenborn 1999; Price *et al.* 2001) argue that neolithization involved small groups of immigrant farmers who came into contact with 'local forager-herder/horticulturalists.'

The advent of radiocarbon dating provided a new instrument for testing the various models of neolithization. The first radiocarbon measurements seemed to confirm the Childean concept of *Ex Oriente lux*, indicating that the "Neolithic way of life penetrated Europe from the southeast spreading from Greece and the south Balkans…" (Clark 1965:67). Later publications based on comprehensive radiocarbon data for Neolithic sites suggested a more balanced view. Tringham (2000:216–217) discussed the spread of new techniques resulting from an expansion of population and their adoption (or rejection) by local groups.

A recent analysis of a large dataset of Neolithic radiocarbon measurements (Gkiasta *et al.* 2003) has basically confirmed the earlier results (Clark 1965; Ammerman and Cavalli-Sforza 1973) in showing a correlation between the earliest occurrence of the Neolithic with the distance from its assumed source in the Near East.

The neolithization of Southeastern Europe proceeded against the background of substantial environmental changes, which included modifications in climate, vegetation, and soils. No less important were the changes that affected the Pontic area, especially those related to fluctuations in the level of the Black Sea and associated phenomena in the littoral zone. The aim of the present article consists in assessing the impact these changes had on the neolithization of Southeastern Europe and the Balkans.

2. METHODS

It should be remarked that the establishment of a direct causal relationship between environmental changes and the social processes that occurred in connection with neolithization is a very complicated task, particularly so for the southern areas of Eastern Europe. The main reason is the insufficient evidence available based on multidisciplinary research. There have been very few examples of such investigations (Ivanov and Vasil'ev 1995; Shilik 1997; Dolukhanov 1979, 2001; Levkovskaya *et al.* 2003).

The second reason is connected with the fact that neolithization in Southeastern Europe proceeded mainly in the forest-steppe zone, where natural conditions were more favorable for the complex development of farming and livestock breeding. In contrast, the Azov-Black Sea steppe zone, where this process was originally aimed predominantly at the development of a livestock economy, became intensively inhabited much later, only from the beginning of the Chalcolithic. Due to these circumstances, Neolithic and Chalcolithic sites are extremely rare in the north Black Sea-Azov Sea coastal area, with few multi-level stratified sites containing long sequences of archaeological deposits.

Taking into consideration the above, the present writers have chosen the following methods to tackle the problem:

(1) assessment of existing evidence pertinent to environmental changes, including fluctuations of Black Sea level;

(2) comparative analysis of the data from the main, clearly-stratified, multi-level Mesolithic to Neolithic sites, their geomorphic setting, and their subsistence base;

(3) analysis of the spatial distribution of Mesolithic, Neolithic, and Chalcolithic cultural entities in the northwestern and northern Black Sea-Azov Sea coastal areas; and

(4) assessment of the possible relationships between archaeologically-established cultural changes and independently-established environmental changes.

3. THE ENVIRONMENT

3.1 Climate, Vegetation, Soils

According to the pollen data, at the beginning of the Holocene, 10,500 BC, summer temperatures in Europe rose by at least 6º C (Isarin and Bohnke 1999). The North Pontic area pollen record (Kremenetsky 1991) also indicates that, during the Preboreal and Boreal periods (10,000–7000 BC), dry steppe with *Carex*-Gramineae and *Artemisia*-Gramineae dominated the vegetation cover of the Pontic Lowland. Rare forests began spreading along the river valleys from *ca.* 7000 BC onward. Initially, they consisted of pine and birch, and later included the broad-leaved species oak and elm, with an understory of hazelnut, all restricted to bottomlands and terraced slopes. The steppe vegetation on the watersheds gradually acquired a mesophytic character. Precipitation markedly increased 8 to 7 ky ago in the entire Mediterranean region (Rohling and de Rijk 1999). In the Pontic area, this period coincided with the maximum spread of mixed coniferous-deciduous forests in the uplands and valley bottoms. The silvo-steppe, with isolated stands of oak, elm, lime, and maple, extended over the watershed plain (Kremenetsky 1991).

3.2 The Black Sea

According to the 'classical' scenario (Fedorov 1978) and its latest modifications (Shmuratko, this volume), during the Allerød interstadial—11,800 to 11,000 years BP—the level of the Black Sea rose from –70 to –10 m (part of the Neoeuxinian transgression). When sea level reached –20 m, i.e., above the Bosporus sill, the low salinity water from the Black Sea started flowing southward through the strait into the Marmara Sea, and from there into the Aegean Sea. Further rise in the level of the Mediterranean Sea slowed the brackish water discharge from the Black Sea, and permitted the penetration of saltwater, thus establishing the present two-flow regime in the Bosporus and Dardanelles Straits. As a result, marine molluscs colonized the Black Sea shelves. The ensuing glacio-eustatic New Black Sea transgression resulted in a gradual rise of sea level, which reached its maximum of 2.0–2.5 m at 3800–3600 BC. At this stage, seawater ingressed into the river valleys and formed the estuaries which eventually developed into the present-day *limans*.

An alternative scenario was suggested by Ryan *et al.*(1997) and Ryan and Pitman (1998). According to this hypothesis, following the post-glacial rise, the Mediterranean Sea catastrophically breached the Bosporus Strait at *ca.* 6100 BC (reported as 7150 BP) and rapidly refilled the Black Sea basin, flooding the shelf. In a later publication (Ryan *et al.*2003), the flood date was pushed still earlier to 8400 BP, or 7500 BC. This alleged inundation led to drastic environmental changes, and stimulated the transition to farming in Europe. Current geologic investigations carried out in the Marmara Sea and on various shelf areas of the Black Sea (Aksu *et al*. 2002, and also Hiscott *et al*, Yanko-Hombach, Balabanov, and others in this volume) proved the robustness of the 'classical' scenario. They provided additional evidence proving that, in the early Holocene, it was the Black Sea that first breached the Bosporus, overflowing into the Marmara and Aegean Seas.

According to the multi-disciplinary data obtained over the past decades by Russian scientists, and based on multiple radiocarbon-dated cores and seismic profiles (Balabanov, Yanko-Hombach, this volume), it has become apparent that Holocene changes in the level of the Black Sea possessed a fluctuating character. At the beginning of the Holocene, 10,500–10,000 years ago, the level of the Black Sea was 40–50 m below that of the present. The Neoeuxinian transgression finally brought sea level up to –25 m by 9350–9200 BC. The end of the Neoeuxinian stage (8650–8300 BC) was marked by a substantial regression to –41.1 to –42.9 m. Then three transgressive phases are identifiable between 8550–6640 BC. The maximum sea-level rise at this stage was linked to the breakthrough of Mediterranean water into the Black Sea basin, signaled by the appearance of euryhaline (saline) Mediterranean fauna. The transgressive Late Bugazian phase, the maximum of which is estimated at 7330–7050 BC, corresponded with a sea-level rise up to –16 or –17 m. During the Vityazevian stage, 7050–5400 BC, sea level rose to –9 or –10 m. A regression occurred at 6.9–7.1 ky BP (6000–5600 BC), during which sea level fell to –20 m, marking the end of the Vityazevian stage. The Dzhemetinian stage (5000–3000 BC) consisted of three transgressive phases at 4600, 4200, and 3800 BC. The next substage culminated at 3000 BC with the maximum sea-level rise. These data, which are deemed as the most reliable, are used further in this text.

3.3 Stratified Mesolithic-Neolithic Sites

In view of the scarcity of multi-level sites and their often inadequate documentary records, only a few sites are described below. Yet, the key sites, including deposits of the Late Mesolithic and Middle Neolithic, are sufficiently dated by radiocarbon determinations.

3.3.1 Soroki I, Soroki II

Two multi-level sites of the Early Neolithic Bug-Dniestrian Culture are located south of the town of Soroki, Moldova, at a distance of 600 m from each other (Figure 1, *28*). They lie on a narrow, low floodplain of the Dniester River (6–8 meters above the river, 65–70 m above sea level), at the bottom of the steep slope of the upper terrace. These sites were excavated by V.I. Markevich in the late 1950s to early 1960s (Markevich 1974).

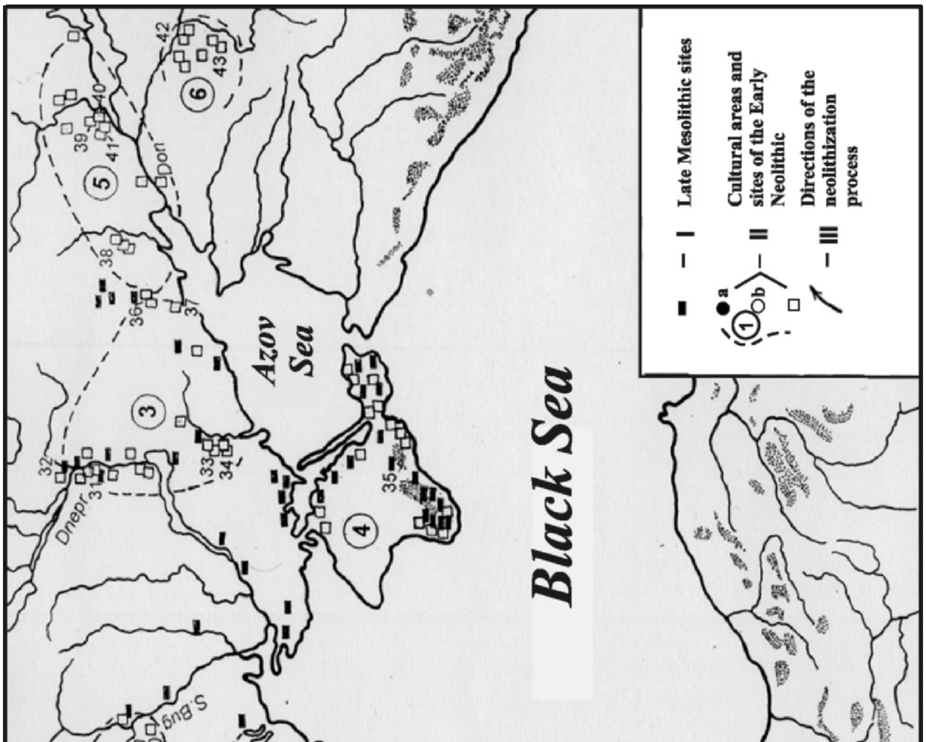
Late Mesolithic sites
Cultural areas and sites of the Early Neolithic
Directions of the neolithization process
Black Sea
Azov Sea
Dnepr
Don
S. Bug
Dnestr
Prut
Donau
Theiss
Morava
Jiu
Iskar
Struma
Vardar
Marica

Figure 1. (Facing page) Mesolithic to Early Neolithic cultures and key sites in the Balkan-Danubia and North Pontic areas.
Cultures: 1 = Starčevo-Criş (a = early, b = late); 2 = Bug-Dniestrian; 3 = Surian; 4 = Crimean; 5 = Rakushechnyi-Yarian; 6 = Kremenian.
Sites: 1 – Sesklo; 2 – Nea Nikomedeia; 3 – Hoca Çeşme; 4 – Anzabegovo; 5 – Kremikovtsy; 6 – Divostin; 7 – Cârcea; 8 – Starčevo; 9 – Gura Baciului; 10 – Szarvas; 11 – Lepenski Vir; 12 – Padina; 13 – Alibeg; 14 – Ogrădeana; 15 – Schela Cladovei; 16 – Ostrovul Mare; 17 – Koprivets; 18 – Ovčarovo; 19 – Suceava; 20 – Trestiana; 21 – Sacarovca; 22 – Erbiceni; 23 – Bereşti; 24 – Garvăn; 25 – Mirnoe; 26 – Grebeniki; 27 – Girzhevo; 28 – Soroki; 29 – Baz'kov; 30 – Pugach; 31 – Surskiy; 32 – Igren'; 33 – Kamennaya Moghila; 34 – Semenovka; 35 – Kukrek; 36 – Razdol'noe; 37 – Mariupol'; 38 – Matveev Kurgan; 39 – Samsonovskoe; 40 – Rakushechnyi Yar; 41 – Razdorskoe; 42 – Rassypnaya; 43 – Kremennaya.

Each of the sites includes three clear archaeological levels. The lower level (2) of the Soroki I site and two lower levels (3 and 2) of the Soroki II site (Figure 2B) are characterized by the absence of pottery, but they contain the remains of domestic animals (*Bos taurus* L., *Sus scrofa domestica* L., *Canis familiaris* L.), with an obvious dominance of wild species. Two upper levels (1a and 1b) of the Soroki I site and the upper level (1) of the Soroki II site include pottery together with remains of the same animals. On this basis, these sites are viewed as reflecting the transition from the Mesolithic to the Neolithic. An alternative view (Dolukhanov 1979) is that these were basically hunter-gatherer communities involved in an exchange network with farming groups.

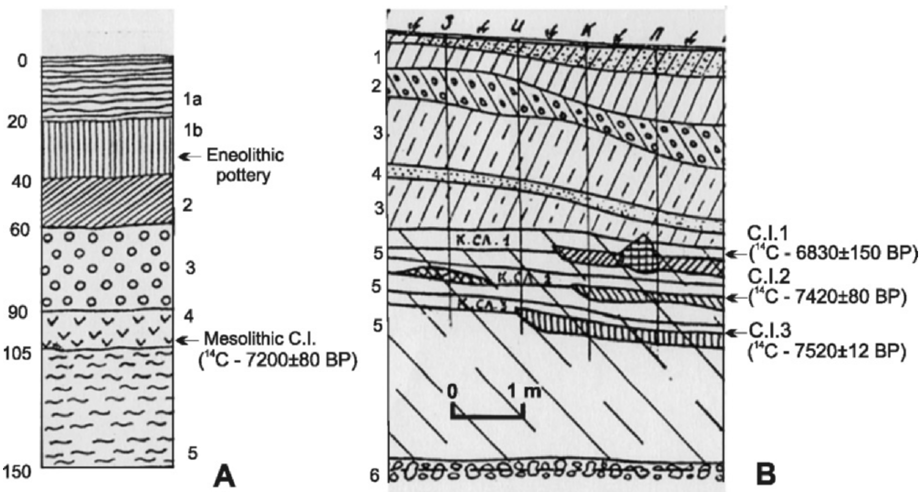

Figure 2. Stratigraphic sequences of Mirnoe (A) and Soroki II (B) sites. **Key:** A: 1a = turf; 1b = clayey loam, the Chalcolithic level; 2 = light gray, sandy, clay loam with plant detritus; 3 = clayey loam; 4 = yellowish clayey loam, the Mesolithic level, with carbonate inclusions; 5 = light brown, heavy clayey loam (after V. N. Stanko 1982). B: 1 = gray forest soil; 2 = clayey loam with calcareous rubble; 3 = heavy, clayey loam; 4 = sandy, clay loam; 5 = light loess-like clay loam, archaeological level; 6 = pebbles (after V.I. Markevich 1974).

There are several radiocarbon dates for all three cultural levels of Soroki II: 6500–6210 BC (Bln-588) for the lower level (3), 6430–6080 BC (Bln-587) for level 2 (without pottery), and 6000–5450 BC (Bln-586) for the pottery-bearing level 1. Recently, seven radiocarbon dates with an average age of 6121±143 BC were obtained for Bug-Dniestrian sites on the Pivdennyi Bug; all satisfy the statistical test for contemporaneity (Dolukhanov and Shukurov 2004).

In the process of excavation, it was stratigraphically established that the cultural levels of the two sites were associated with the alluvial clayey loam (up to 2.0 m thick) that lies directly on riverine pebbles. At Soroki I, the lower cultural level (2) was separated by a thick layer (up to 1.0 m) of limestone gravel, probably eroded out from an adjacent ancient ravine. Alongside, two lenses of alluvial sediments can be observed, allegedly corresponding to separate periods of abrupt rise in the level of the Dniester River.

Riverine flooding is clearly visible in the sequence at Soroki II (Figure 2B), where all three cultural levels are separated by sterile, and almost equally thick, beds of alluvial sediment. Ultimately, the site was buried under thick deposits of sandy, clayey loam and limestone gravel caused by a major landslide.

3.3.2 Kamennaya Moghila 1

This important multi-level site lies on a hilltop in the floodplain of the right bank of the Molochnaya River's ox-bow, within the village of Terpeniye, Melitopol region, Zaporozhe Province, Ukraine (Figure 1, *33*). It comprises several archaeological levels. The Late Mesolithic deposit has been radiocarbon-dated to 6430–6150 BC; the Early Neolithic levels are attributable to the Surian Culture with three ^{14}C dates: 6230–6020 BC (Ki-4022), 6220–5900 BC (Ki-4226), and 6030–5780 BC (Ki-7667); and the Late Neolithic levels of the Azov-Dnieprian Culture have yielded three ^{14}C dates: 5480–5250 BC (Ki-4025), 4890–4850 BC (Ki-4024), and 5290–4800 BC (Ki-4023) (Kotova 2003).

3.3.3 Semenovka 1

The site, in the village of Semenovka, Melitopol region, Zaporozhe Province, Ukraine (Figure 1, *34*), lies on the first terrace of the right bank of the Molochnaya River. The Neolithic deposits of this multi-layer site are found in different parts of the terrace, dissected by a brook. In one area, deposits attributed to the Surian Culture were identified at a depth of 1.4 to 1.8 m, overlying the sterile, clayey loam. Available ^{14}C dates for this level are: 6260–6310, 6090–6140, 6080–6140, 5930–5720, and 5890–5620 BC. The second level, consisting of "light-gray humified loamy sand," is associated with the Azov-Dnieprian Culture of the developed Neolithic. It was found on a sandy elevation and dated by radiocarbon to 5480–5200 BC (Kotova and Tuboltsev 1996; Kotova 2003).

Pollen analysis performed by Pashkevich (Dolukhanov 1979) reveals a vegetation of bunchgrass steppe with rare trees (birch, alder, elm, and hornbeam). Nearly 90% of the animal remains from the Neolithic layers belong to domesticates–predominantly cattle, followed by horse, sheep, goat, and pig. The wild species include red deer, boar, and bison, with numerous remains of fish and birds.

3.3.4 Rakushechnyi Yar

Rakushechnyi Yar (Figure 1, *40*) is a clearly stratified Neolithic settlement located on a small island in the lower stretches of the Don River, about 100 km upstream from the city of Rostov. The site includes 23 archaeological layers (Belanovskaya 1995), of which the deepest (23–26) belong to the Early Neolithic. The layers are 5–15 cm thick and interbedded with sterile sand or silt. Archaeological deposits, which are not identical in each layer, allegedly resulted from seasonal occupations. Fireplaces and the remains of surface dwelling structures were recognized in several strata. Animal remains comprise both wild (red deer, roe deer, fox, hare, and numerous birds) and domesticated species (sheep, goat, cattle, dog, and horse, the last either wild or domestic). Numerous shells of edible molluscs (mostly *Viviparus*) indicate the importance of food gathering. The flint industry includes end scrapers made from blades and flakes, retouched blades, and borers. Arrowheads and geometrics (symmetrical trapezes) occur only in the upper levels. The pottery is often tempered with organic matter and includes both flat- and pointed-bottom varieties. Their ornaments are usually restricted to the upper part of the vessel and consist of triangular notches forming horizontal rows, small pits, and incised lines. The developed character of the material culture and the apparent absence of Mesolithic elements imply that Rakushechnyi Yar was not the oldest Neolithic site in the area, but its preceding stage remains to be found.

3.3.5 Matveev Kurgan 1 and 2

Two Early Neolithic sites, Matveev Kurgan 1 and 2, are located in the valley of the Miuss River, on the littoral of the Azov Sea (Figure 1, *38*) (Krizhevskaya 1991). Site 1 includes the remains of a surface dwelling with hearths and post-holes, as well as an open, allegedly ritual fireplace. At Site 2, open fireplaces and large stone and clay inlays were found. The animal remains of the both sites are dominated by wild species: aurochs, red deer, roe deer, beaver, wolf, wild boar, and two equids—the kulan (*Equus hemionus*) and the extinct hydruntine (*E. hydruntinus*), the latter two species being more typical of the Mesolithic age. The domesticates, which made up 18–20% of the total assemblage, include horse, cattle, sheep/goat, pig, and dog.

Both sites contained rich lithic industries, with no less than 600 cores (both single- and double-platformed). Elongated broad blades and less numerous flakes dominated the assemblage. End scrapers made from large flakes and retouched blades were found together with various tools made from blades. There were about 90 geometric microliths, mostly trapezes, in both symmetric and asymmetric varieties. Several 'bifacial' flint axes were reported, yet the number of slate polished axes was much greater. A diverse bone-and-antler industry found at both sites included spear- and arrowheads, awls, and their fragments. Both sites have yielded slate sinkers for fishing nets. Only a handful of pottery items were found at either site: 6 fragments at Site 1, and 21 at Site 2. The pottery fragments were unornamented and manufactured of silty clay without any apparent artificial tempering.

Six out of 10 dates from the lower layers of both Rakushechnyi Yar and Matveev Kurgan satisfy the criterion for contemporaneity, yielding an average age of 5863±130 BC. The remaining dates include one younger (5000 BC) and three older (6550–6850 BC) (Dolukhanov and Shukurov 2004).

3.3.5 Razdorskoe 1

Another site, Razdorskoe 1 (Figure 1, *41*), lies 0.5 km downstream on the 4–8 m high terrace of the right bank of the Don River, on an alluvial fan formed by a brook flowing within a large gully. The site has 23 cultural layers with a total thickness of 5.7 m. The lowermost layer (Belanovskaya 1995) is referred to the Early Neolithic Rakushechnyi Yar Culture; the only ^{14}C date for this level, 9470±310 BP (IGAN-722), is apparently too old, probably due to contamination. The second and third lowest levels (or 22nd and 21st) are associated with various horizons of the Lower Don (or Mariupol') Culture of the developed Neolithic-Early Chalcolithic. The site is still under investigation, and only preliminary data are available (Kiyashko 1987). The pollen (Kremenetsky 1991) from the lowest levels accumulated in an early Holocene environment dominated by *Carex*-Gramineae steppe and pine forests spreading along sandy terraces (Kremenetsky 1991:123).

3.3.6 Mirnoe

Mirnoe (Kilia region, Odessa Province, Ukraine) is one of the few fully investigated Late Mesolithic sites in the northwestern Black Sea coastal area (Figure 1, *25*). The site lies on the floodplain of the Dracula River, 800 m southwest from the southern limit of the village of Mirnoe. The site was discovered in 1963, and systematic excavations were carried out in 1969–1976 by V.N. Stanko, resulting in the exposure of 1807 m^2 of archaeological deposits (Stanko 1982). There is only one radiocarbon date for the cultural level: 6230–5890 BC (Le-1647). A 'hearth area' identified in the central part of the settlement, included several hearths and so-called 'backers' pits' (circular pits with traces of fire, supposedly intended for cooking) with a high concentration of animal bones and flint pieces. As suggested by use-wear analysis, lithic tools were used predominantly for butchering and skin dressing. Retouched blades are also numerous, while cores were few in number. Eighteen clusters of implements were identified in the peripheral part the site, their boundaries not always being clear. Among 214 animal bones found at the site were identified: horse (minimum number of individuals = 7); aurochs (3), saiga (3), and fox (1). The finds of elongated prismatic blades with characteristic sickle sheen were particularly significant; based on the use-wear analysis, they were identified as 'reaping knives' intended for harvesting edible plants. Several potentially edible plants were identified by Pashkevich (1982) in the cultural deposits: white goosefoot (*Chenopodium album*), black bindweed (*Polygonum convolvules*), hairy vetch (*Vicia hirsuta*), and sorel (*Rumex acetosa*).

The Mesolithic layer (from 0.15 to 0.30 m thick) was found in a yellowish sandy clay overlying a light yellow-brown heavy clayey loam of apparently lacustrine origin. The archaeological layer was covered by a light brownish clayey loam deposited in an environment of rising lake level (Figure 2A). Pollen analysis (Pashkevich 1982) indicates that the Mesolithic settlement occurred in an environment of forb-bunchgrass steppe. The pollen spectra of the brownish clayey loam show a vegetation cover of mesophytic steppe with rare broad-leaved trees and shrubs, apparently spread along the river banks.

3.4 The Distribution Areas of Mesolithic-Neolithic Cultural Entities

We proceed from the well-known postulate that every significant cultural phenomenon of the past (from a site to an archaeological culture) covered a distinct territorial space. The confines of the space around a site (often referred to as *site exploitation territory*, or SET) indicates the area habitually used by site occupants as the source of their livelihood and an arena of various activities. Its concrete location was dictated by numerous factors, which included prevailing subsistence (hunting-gathering, broad-spectrum agriculture, or stock breeding), type of site (base camp, permanent settlement, seasonal encampment, etc.), and general lifestyle (peripatetic hunting, sedentary, seasonal transhumance). No less important were the availability of sustainable and predictable natural resources within the *site catchment areas* (SCA). In the case of hunter-gatherer subsistence, these resources include the landscapes suitable for hunting, fishing, collecting of edible molluscs and other wild foods, as well as sources of raw materials. In the case of early agriculture, they include arable land and the combination of climatic factors referred to as agro-climatic potential. The absence of archaeological sites in any landscape area may be seen as indicating adverse environmental conditions impeding human settlement.

Our analysis is based on the distribution of archeological sites of the four main cultural periods: the Late Mesolithic, the Early and Late Neolithic, and the Early Chalcolithic. This sequence of cultural development falls within the very Early to Middle Holocene, in the range between *ca.* 10,000 and 4000 BC.

Intensive settlement of the northern Black Sea steppe area starts with the Mesolithic. This will be apparent from a comparison of the distribution of Late Paleolithic (Figure 3) and Mesolithic sites (Figure 4). These maps are based on Smyntyna (1999:240) with additions for Moldova and the eastern provinces of Romania. The comparison clearly demonstrates that, in the Late Paleolithic period, the majority of the sites concentrated in the present-day forest-steppe zone and to the north of it, whereas in the Mesolithic period, the sites shifted southward into the present-day littoral zone. It is significant that many of them are in the immediate vicinity of the contemporary coast, occupying the lowland plain locations of the littoral. It is clearly observed for the Crimean steppe, the northern Black Sea littoral, and the Azov Sea littoral as well as for the northwestern Black Sea littoral, including Dobruja.

The peculiarities in the spatial distribution of the Late Paleolithic vs. Mesolithic sites clearly stem from their environmental settings. Under the conditions of the Last Glacial Maximum, population densities of Late Paleolithic groups markedly increased in the periglacial zone of Eastern Europe, where the conditions for hunting big game were optimal. These groups penetrated the Pontic Lowland only occasionally, and settled in the valleys of small rivers (Dolukhanov, this volume). In contrast, during the early Holocene, when big game became extinct, Mesolithic groups settled predominantly in the Pontic Lowland, where deep river valleys and low mountain slopes offered greater and more predictable resources for hunting, fishing, and food-gathering. As follows from the spatial distribution of Mesolithic sites in the Danube-Dniester interfluve in southern Bessarabia (Stanko *et al.* 1999; Dolukhanov 2001; Dolukhanov *et al.* 2004), some Mesolithic sites occupy rather high places (Figure 5, altitudinal levels IV-III). But several sites (Mirnoe 1, 2, 3, Vasilievka, Divizia, Noveselitsa 1, Tatarbunary 1, and others, Figure 5, *1–9*) are found on the lowest (I-II) horizons, at elevations from 10–15 to 40–50 m above sea level. Significantly, the low position of Mesolithic sites is equally typical of Romanian Dobruja. It is particularly the case of the Garvăn site (Figure 5, *13*), situated on the lowest sandy terrace, flooded currently by the Danube (Păunescu 1987a:3; 1987b:12). One possible explanation for this is seasonal transhumance, with the largest sites of sedentary type (Mirnoe 1, Beloles'e) being located on the lowermost levels, close to the river floors, and the smaller sites found in the foothills, resulting from the resettlement of smaller groups during the summers.

Considering the ^{14}C dates for Mirnoe and other Late Mesolithic sites (Erbiceni, Kamennaya Moghila I, Marievka, and Vasilievka), as well as those for the Early Neolithic sites discussed below, the Late Mesolithic of the North Pontic Plain may be dated between *ca.* 7000 and 6000 BC (8100–7200 BP).

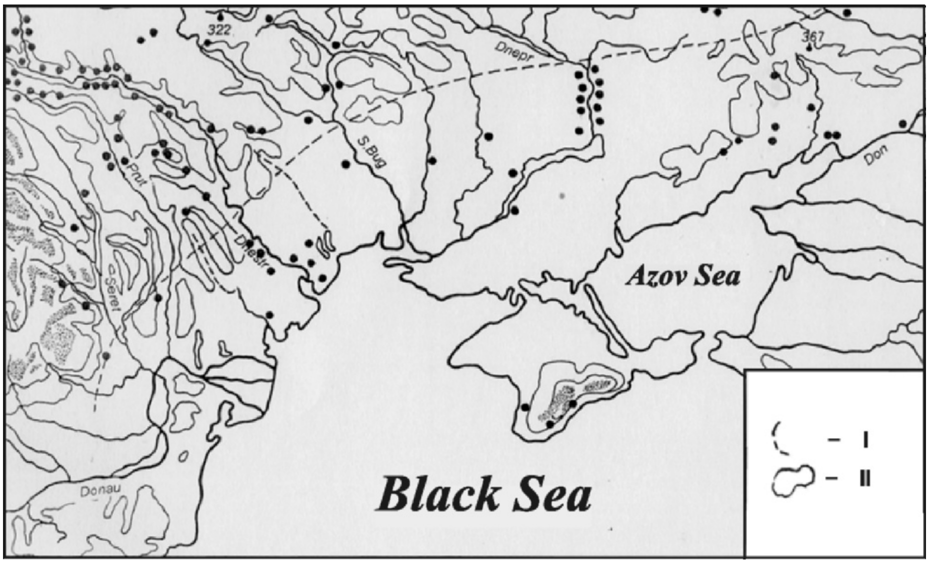

Figure 3. Late Paleolithic sites in the North Pontic area; after E.V. Smyntyna (1999) with additions. **Key:** I = Present steppe/forest-steppe limit; II = 200 m contour.

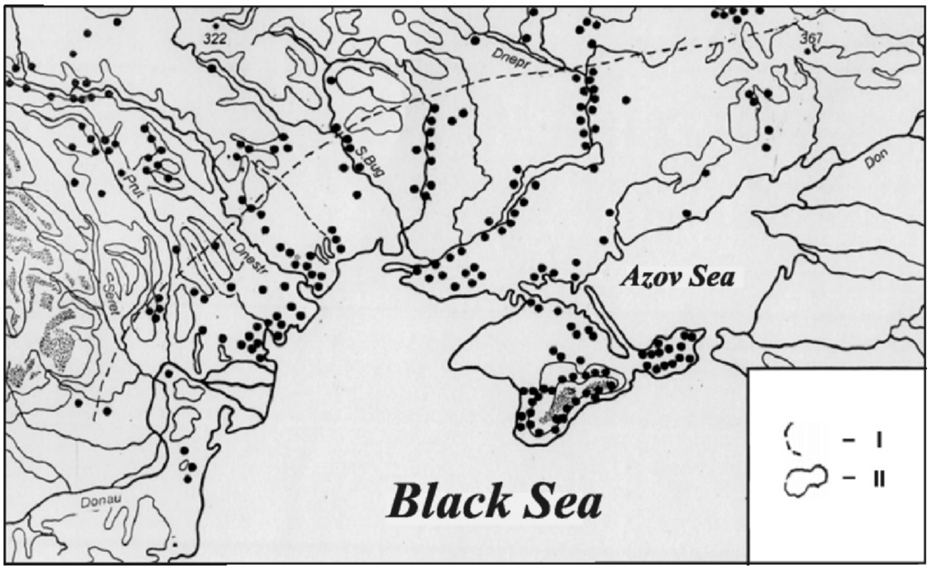

Figure 4. Late Mesolithic sites in the North Pontic area; after E.V. Smyntyna (1999) with additions. **Key:** I = Present steppe/forest-steppe limit; II = 200 m contour.

The Neolithic period in the discussed area is subdivided into two main subperiods. The earlier subperiod in the Balkan-Danubian area is connected to the development of the Protosesklo-Karanovo I-II-Starčevo-Criş cultural community, which spread eastward from the Carpathians almost up to the Dniester

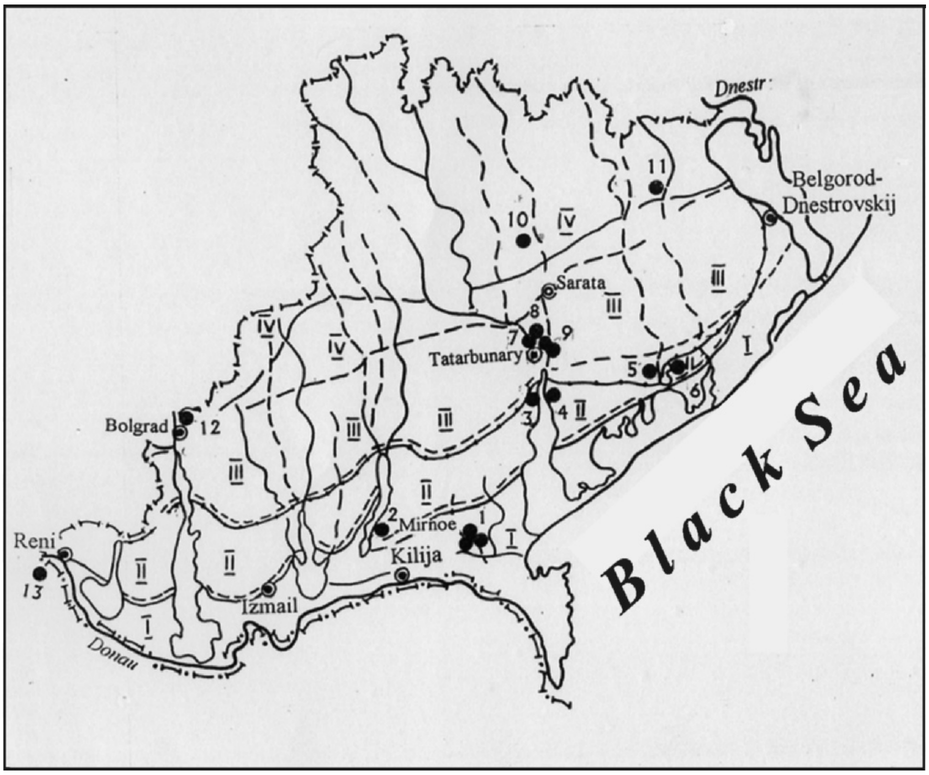

Figure 5. Late Mesolithic sites in the Danube-Dniester interfluve. **Altitudinal levels:** I = 10–20 m; II = 20–50 m; III = 50–85 m; IV = 85–140 m. **Sites:** 1 – Mirnoe 1–3; 2 – Vasilievka; 3 – Borisovka; 4 – Trapovka; 5 – Divizia; 6 – Novoselitsa 1; 7 – Tatarbunary 1; 8 – Kogylnik; 9 – Beloles'e 1, 4; 10 – Cantemir; 11 – Tsarichanka; 12 – Zaliznichnoe; 13 – Garvăn (Romania).

River in its later stages. The development of the Early Bug-Dniestrian, Surian, Rakushechnyi-Yarian, the early stages of Azov-Dnieprian, and the Kremennaya II type sites in the Pontic Lowland fall roughly within the same subperiod. The dating, without taking into consideration the aceramic period of Thessaly, is approximately 6500 to 5500 BC.

The later Neolithic subperiod coincides with the occurrence of Vinča type cultures (typified by Vinča-Turdaş, Banat, Vădrasta, Dudeşti), the Hamangian, and the Boian (early stages) in the Balkan-Danubian area, and east of the Carpathians, by the development of the Linear Pottery Culture (in the Carpathian-Dniester area), late Bug-Dniestrian, late Azov-Dnieprian (Mariupol'), and the Lower Don Culture. This subperiod dates between 5500 and 5000 BC.

The earlier and later sites of the Protosesklo-Karanovo I-II-Starčevo-Criş culture encompass vast areas of the Balkan interior, while it was totally absent from the coastal zone of the western Black Sea littoral. Significantly, the eastern limit of Karanovo I-II settlements runs along the low and middle courses of the Marica River (Hoca Çeşme, Aşaği Pinar), the upper course of the Kamčiya River

(Ovčarovo, and others), and the Lom River (Koprivets, and others), which flow into the Danube. No Early Neolithic sites have ever been found in Bulgaria's Black Sea coastal area, despite intensive surveys. Bulgarian experts have repeatedly underestimated this evidence (Todorova 1986; Popov 1996; Mateva and Skakun 2003). Figure 6 shows the main Early Neolithic sites in the discussed area with the omission of the Crimean steppe, which is insufficiently studied and has no radiocarbon dates (Kolosov 1985).

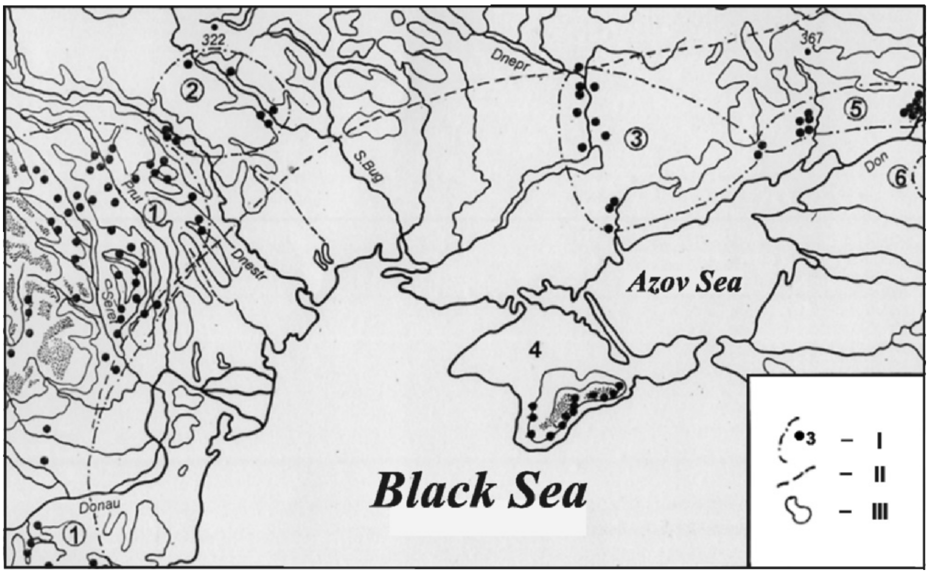

Figure 6. Early Neolithic sites and cultures in the North Pontic area. **Key:** I = Sites and cultural areas; II = Present steppe/forest-steppe limit; III = 200 m contour. **Cultures:** 1 = Starčevo-Criş-Karanovo I-II (Koprivets 1–2); 2 = Bug-Dniestrian (early); 3 = Surian; 4 = Crimean; 5 = Rakushechnyi-Yarian; 6 = Kremenian.

To interpret this phenomenon, one should remember that the subsistence base of all the abovementioned Early Neolithic Balkan sites was agriculture and stock rearing. After their first appearance in Greece at about 7000 BC, village-type settlements expanded rapidly in the Balkans with economies based on the exploitation of domesticated animals and plants of Near Eastern origin: sheep, goats, pigs, and cattle for animals, and wheats (*Triticum dicoccum*, *T. monococcum*, *T. aestivum*), barley (*Hordeum vulgare* ssp. *distichon*), pulses (*Lens culinaris*, *Vicia ervilia*, *Pisum sativum*) for plants (Perlès 2003). Correspondingly, these settlements were located in areas with high agro-climatic potential: terraced intermountain depressions sufficiently rich in water resources with light, easily arable soils (Dolukhanov 1979). In Romania, the Starčevo-Criş sites are found in the Lower Danube Plain and Moldova, usually in the lower river valleys (Maricescu-Bîlcu 1993). Obviously, the present-day western Black

Sea littoral with its poor soils and dry, unpredictable climate offered much less opportunity for farming economies in their early stages.

The location of Early Neolithic sites in lowlying geomorphological settings is particularly typical of the Eastern steppe area, as exemplified by Semenovka 1 (lower level), in the vicinity of Melitopol (Figure 1, *34*). Farther west, in the basin of the Lom River (Figure 6, lower left corner), several early Neolithic sites, classified by V. Popov as belonging to the Koprivets group, are found on the lower floodplain. But, as the author notes, such a location is typical only of sites in the lower Koprivets horizons (levels 1 and 2). All later sites shifted abruptly southward and were sited, as a rule, on higher ground, at 150–250 m above sea level (Popov 1996:30, 33). The lowest Koprivets horizon includes monochrome pottery, and the second one can reliably be synchronized with Karanovo I (Popov 1996:89). There are no ^{14}C dates available for Koprivets, but based on Karanovo I parallels, its age may be estimated at about 6000–6300 BC. Three of the five dates known for the settlement of Semenovka 1 (see above) point to the same interval. The low position of these sites may be due to the fact that densely forested floodplains with abundant wildlife resources were more attractive for the area's Neolithic settlers, for whom hunting-gathering activities still played a significant role in subsistence.

The cultural situation in the entire North Pontic area changed abruptly with the transition from Early to Late Neolithic. At that time, the forest-steppe of the Eastern Carpathian area became densely populated by settlements of the agricultural Linear Pottery Culture (Figure 7, *4*). The later stages of the Bug-Dniestrian (Figure 7, *5*) continued their development further east. As previously, the steppe zone remained sparsely populated. Apparently, there are no sites in the basins of the Lower Dniester, the Southern Bug, and much of the Dnieper. The Azov-Dnieprian (Mariupol'), the Lower Don sites, and those of Crimea's 'Steppe Neolithic' (Figure 7, *6–8*) are relatively rare. But it is notable that, despite their scarcity, several sites are found on the Azov Sea littoral, as well as in northern and eastern Crimea (Dolinskoe and Frontovoe, respectively) and on the Lower Don (Liventsovka, Botai, Zalivnoye). During this period, a new settlement appears at Semenovka (level 3).

Yet, there is indirect evidence suggesting that the scarcity of sites does not adequately reflect population density. This may be illustrated by the position of the Mariupol' cemetery, located at an elevation of 15–20 m above sea level on a residual island in the mouth of the Kalmius River on the Azov Sea coast (Makarenko 1933).

The most significant changes occur in the Lower Danube area and Dobruja, a lowland with numerous lagoons, marshes, and densely forested residual islands (Munteanu 1996). In the previous period, the Early Neolithic, this entire area was virtually empty of settlement. By contrast, in the Late Neolithic period, it became immensely rich in archeological sites belonging to various cultures, sites of which were often found on low-lying and presently

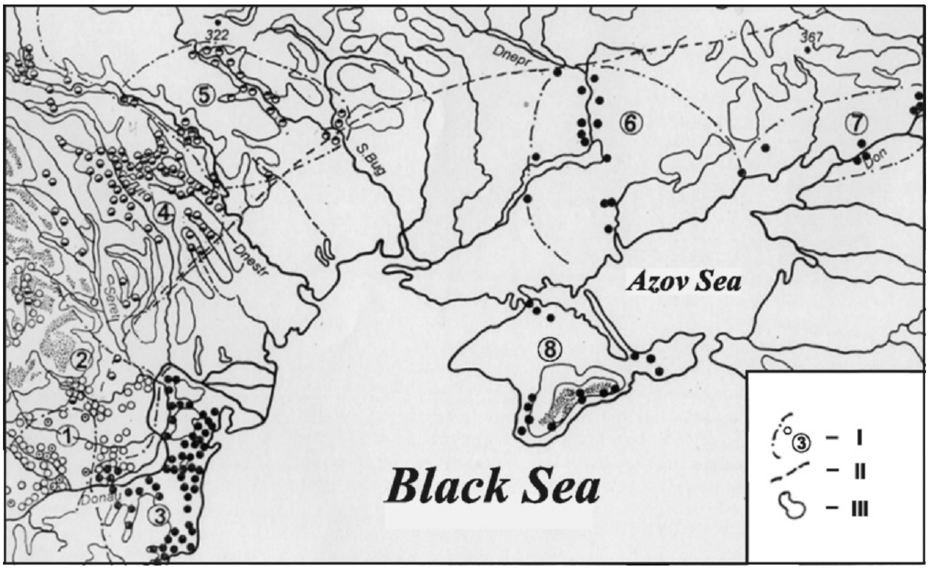

Figure 7. Late Neolithic sites and cultures of the North Pontic area. **Key:** I = Sites and cultural areas; II = Present steppe/forest-steppe limit; III = 200 m contour. **Cultures:** 1 = Dudeşti; 2 = Boian (early); 3 = Hamangia; 4 = Linear Pottery; 5 = Bug-Dniestrian (late); 6 = Azov-Dnieper (Mariupol'); 7 = Lower Don; 8 = Crimean.

flooded levels. On the left bank of the Danube, these are the sites of Dudeşti (Comşa 1971) and those representing the early stages of the successive Boian Cultures (Comşa 1974) (Figure 7, *1–2*). In Dobruja, these are the sites of the Hamangia Culture (Haşotti 1997) (Figure 7, *3*). So far, there are but few radiocarbon dates for the Dudeşti and Hamangia sites, and they reflect only their later stages. However, it is generally accepted that the Dudeşti Culture is synchronous with Hamangia's early phases. On the other hand, the Dudeşti Culture is usually attributed to the Vinča cultural assemblage. The emergence of these two cultures in the Balkans dates to about 5500 BC and corresponds to the early stages of the Vinča-Turdaş and Banat Cultures. Therefore, at this time, the Lower Danube and Dobruja became available for large-scale settlement of culturally variegated groups, for which subsistence was based largely on grain agriculture and stock rearing, with lesser importance placed on hunting and fishing. This period of intensive lowland settlement continued throughout the existence of the Dudeşti Culture and the early stages of the Boian and Hamangia Cultures until about about 5000 BC, when the number of sites belonging to the final stages the Boian and the Hamangia Cultures in the Lower Danube (Figure 8, *1*) and Dobruja (Figure 8, *2*) declined considerably compared with the previous period (cf. Figure 7, *2–3*).

Tell-type sites of the Early Chalcolithic Gumelniţa A-Aldeni II Culture in Bulgaria and Romania appear from *ca.* 4900 BC, clustering usually on elevated locations in the Lower Siret and Lower Prut areas (Figure 8, *3a*)

(Dragomir 1983:17).

The sites of Gumelniţa-Bolgrad type were the first agricultural settlements to appear on the steppe littoral of southern Bessarabia (Subbotin 1983). Sites in the Danube-Dniester interfluve were usually located on uplands: the terraced banks of freshwater lakes and *limans*, the banks of major rivers (the Danube and Prut), and in a few cases the banks of smaller rivers and streams. These naturally fortified settlements had easy access to arable soils: the carbonate chernozem that developed atop the loess-covered terraces. The largest settlements were close to the river estuaries transgressed by the sea, 2–3 m above present-day sea level (Fedorov 1978). Two radiocarbon dates obtained for the site of Vulcaneşti III in southern Moldavia suggest an age of 4896–4006 BC.

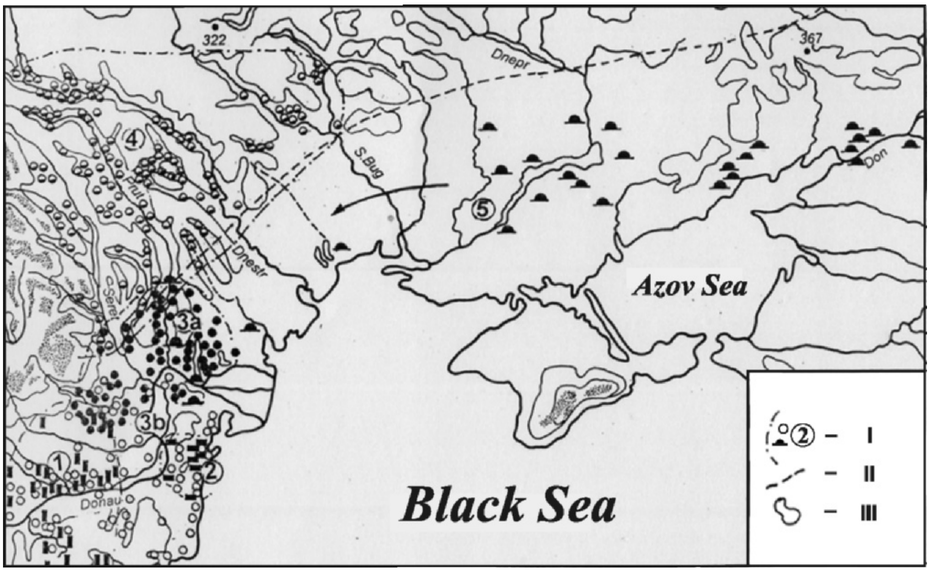

Figure 8. Early to Middle Chalcolithic sites and cultures of the North Pontic area. **Key:** I = Sites and culture areas; II = Present steppe/forest-steppe limit; III = 200 m contour. **Cultures:** 1 = Boian (late); 2 = Hamangia (late); 3a = Bolgrad-Aldei II; 3b = Gumelniţa -Karanovo VI; 4 = Precucuteni-Tripolye A; 5 = Novo-Danilovka type (the arrow shows their westward movement).

Subsequently, the Black Sea littoral was densely occupied by sites of Gumelniţa-Karanovo VI-Varna type (Figure 8, *3b*). Notwithstanding repeated culture changes, the sites in the littoral zone grew continuously in number until the Middle Chalcolithic, which was marked by the emergence of the Cernavoda I-Pevets Culture at *ca.* 4200–4000 BC (Todorova 1986; Bojadjiev *et al.* 1993).

Sites of the Precucuteni II-III-Tripolye A Culture represented the large-scale spread of agricultural communities into the forest-steppe zone east of the Carpathians at about 4500–4350 BC (Figure 8, *4*). The situation in the steppe zone remains obscure at this time. Ground burials and the first barrow complexes of Novo-Danilovka type began spreading widely in the steppe at the end of the

Precucuteni-Tripolye A transitional and the initial Cucuteni A-Tripolye B1 periods (Figure 8, *5*). These burial sites, apparently genetically related to the Khvalynsk Culture of the Middle Volga, are deeply wedged into the area of farming cultures in the west, including the Tisza Plain and Thrace, and correspond to the first wave of Indo-European dispersal, according to the hypothesis of Gimbutas (1979).

4. DISCUSSION

As follows from the previous review, considerable changes in spatial distribution occurred among prehistoric cultures within the Balkan-northern Black Sea area during the Early and Middle Holocene. These changes were accompanied by modifications in subsistence, settlement pattern, and cultural affiliations of local communities. Changes that could suggest human migrations of varying scale, cultural diffusion, and interactions or combinations of these processes, proceeded against a background of significant environmental shifts, including changes in climate, vegetation, and soil types, as well as the Black Sea shoreline. Shoreline migration became the focus of both scholarly and public interest after the series of publications by Ryan, Pitman, and their colleagues (Ryan and Pitman 1998; Ryan *et al.*1997, 2003), who suggested the occurrence at *ca.* 8400 BP (*ca.* 7200 BC)–revised from their earlier dating of 7150 BP (*ca.* 6000 BC)–of a catastrophic breach of Mediterranean saltwater through the Bosporus Strait resulting in an abrupt drowning of more than 100,000 km^2 of previously exposed Black Sea continental shelf. This, they propose, led to a rapid spread of farming from Greece, Bulgaria, Romania and the Marmara coast inland along the major river valleys of Southeastern Europe (Ryan and Pitman 1998).

Multiple investigations conducted by marine geologists in the Black Sea area and supported by recent multidisciplinary research (Yanko-Hombach, Balabanov, Konikov, Chepalyga, this volume) clearly demonstrate that the Ryan-Pitman 'Flood' scenario is based on a limited scope of incorrectly interpreted evidence. Therefore, it can no longer be seriously considered. Nonetheless, recent investigations do confirm the occurrence of multiple fluctuations and smaller-scale migrations of the Black Sea coastline over the course of the Holocene. The question is whether such changes in the coastal environment had a discernible impact on the subsistence and movements of prehistoric groups.

It is quite obvious that these shoreline migrations proceeded at a slow rate (2.5–5 cm/year) and were restricted nearly exclusively to the present-day continental shelf. Thus, they could not have had any *direct* effect on prehistoric groups beyond the littoral zone. The maximum rise of sea level that resulted in the formation of the transgressive shoreline at an elevation of 2.0–2.5 m above present sea level, occurred 4.6–4.4 ky BP or 2800–3000 BC (Balabanov, this volume). The effect of this and similar sea-level rises was restricted mainly to the river estuaries and *limans*, which were repeatedly ingressed by the sea.

The general trend of the Holocene transgression in the Black Sea was controlled by glacio-eustasy: the rise of water volume in the global ocean due to the post-glacial ice sheet melting. At the same time, the secondary and tertiary fluctuations observable in the Black Sea might have been caused by local factors, such as tectonics. No less important a factor was change in the Black Sea water balance, which depends greatly on the freshwater river discharge flowing into the Black Sea (presently estimated at 310 km^3/yr). Therefore, the observable sea-level rises correspond to intervals of increased volume of river discharge, which, in its, turn, should reflect an increase in precipitation within the catchment areas of these rivers. Hence, the observed transgressions in the level of the Black Sea may be considered as indicative of episodes of increased precipitation in the riverine drainage basins. This might have had a considerable effect on the environment, as well as human movements in the steppe and forest-steppe area.

The curve of Holocene sea-level changes (Balabanov, this volume) indicates the occurrence of low-frequency cycles with periods of 1800–2000 years. It should be noted that evidence of a 1470-year cycle was reported (Grootes and Stuiver 1997; Schulz *et al.* 1999) from the glacial-age Greenland ice record. It has been suggested that similar cycles are acknowledgeable throughout the entire Holocene as well (Bond *et al.* 2001). Cycles of similar magnitude were previously identified by Shnitnikov from Russia's Holocene records (Shnitnikov 1973:16).

Based on the abovementioned evidence, one can assess the possible effect of sea-level changes on the neolithization of the Balkans and North Pontic area over the course of the main established chronological units.

4.1 Mesolithic (10,000–6500 BC or 10,250–7700 BP)

This was a period of generally low sea level, ranging between –55 and –20 m below present. A considerable part of the continental shelf was exposed, and the shoreline was located in places 80–100 km seaward of its present position. Larger sites, some of which were base camps inhabited on a year-round basis, were located on the lower terraces of the limans, at that time marshy river bottomland. Smaller sites at higher levels were formed by groups budding off from the main centers during seasonal transhumance.

4.2 Early Neolithic (7,000/6500–5500 BC or 8100/7700–6600 BP)

This period corresponds to the occurrence of Protosesklo-Karanovo I-II Cultures in the Balkan interior, and Early Bug-Dniestrian, Surian, Rakushechnyi-

Yarian, the early stages of Azov-Dnieprian, and the Kremennaya II type sites in the North Pontic area. During this period, there occurred several considerable sea-level rises, which never exceeded a level of –9 m below present. The Late Bugazian phase reached a level of –16 to –17 m at 7330–7050 BC. During the Vityazevian phase (7040–5740 BC), sea level rose to –9 to –10 m. Subsequently, the emergence of early farming settlements in Greece and their large-scale spread in the Balkans coincided with several transgressive phases, which are indicative of a general increase in rainfall.

4.3 Later Neolithic (5500–4000 BC or 6600–5200 BP)

This period corresponds to the large-scale spread of Linear Pottery Culture in the western Balkans and Moldavia; the later stages of the Bug-Dnestrian, the Azov-Dnieprian (Mariupol'), the Lower Don sites, and those of Crimea's 'Steppe Neolithic' farther east. Considerable changes occur in the previously unoccupied Lower Danube and Dobruja, where settlements of the Dudeşti, Boian, and Hamangia Cultures appear. Sites of Gumelniţa-Bolgrad type were the first agricultural settlements to appear on the steppe littoral of southern Bessarabia. This period includes several significant sea-level rises: the late Vityazevian transgression at *ca.* 6800–5800 BC reaching –8 m, and also the Early Dzhemetinian culminating close to the present sea level at 4600 and 4200 BC. This means that the expansion of farming settlements covering areas of the northwestern coast proceeded within an environment of increased humidity and repeated sea-level rises.

4.4 Chalcolithic (4500–3200 BC or 5650–4500 BP)

This period of the Cucuteni-Tripolye Culture marks an intensive expansion of early farming settlements in the northern Pontic forest-steppe. Significantly, this period roughly coincides with the Dzhemetinian transgression (5000–3000 BC), during which peaks at 4600, 4200, 3800, and 2800–3000 BC brought sea level close to or slightly above that of the present.

5. CONCLUSIONS

(1) Following from the above-cited records, no evidence is available to substantiate catastrophic sea-level rises that can be linked with large-scale migrations of early farming groups.

(2) The alleged sea-level rise at *ca.* 8400 BP (*ca.* 7400 BC) in fact preceded the emergence of early farming communities in the Black Sea area. The Late Mesolithic groups that existed at the time were basically stable with limited

evidence suggesting seasonal transhumance.

(3) According to reliable evidence, at the time of the alleged ‘Flood’ at 7150 BP (*ca.* 6000 BC), sea level rose from –32 to –8 m and in no way affected the agricultural settlements which were usually located within intermontane depressions at a great distance from the littoral. Expansion of settlements recognized at that time in Greece and the western Balkans is due to a general increase of precipitation that created the conditions favorable for early farming.

(4) Significantly, periods of major extension of farming communities in the northwestern and northern Pontic area coincide with the minor Black Sea transgressions, possibly due to higher precipitation in the river catchments, which enhanced the agro-climatic potential in the steppe and forest-steppe.

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22

HOLOCENE CHANGES IN THE LEVEL OF THE BLACK SEA: CONSEQUENCES AT A HUMAN SCALE

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Abstract: Recent research on Holocene changes in the Black Sea has provided an extraordinary data set for archaeologists studying the region. Most immediate discussion has focused on population movements, the location of submarine sites, or the origins of myth and legend, but proposed links between a dramatic rise in the Black Sea and the spread of agriculture across Europe, the floods of the Bible, or the Gilgamesh epic are unhelpful. They spring from long-abandoned approaches to human behavior in which causation is assumed to be simple and direct, of environmental origin, and, more often than not, grounded in the movement of populations (i.e., cultures). From the perspective of a modern, critical archaeology, the Holocene changes in the Black Sea have important consequences, yet their importance lies not in how they may have immediately or directly changed life in the Circum-Pontic region but in how an increasingly refined and broadened documentation of these changes can radically refigure our understanding of a critical period in European pre-history.

Keywords: Mesolithic, Neolithic, southeastern Europe, Iron Gates Gorges sites, scales of analysis

1. INTRODUCTION

This paper investigates the evidence for possible impacts of a Black Sea transgression in the seventh millennium calBC on the ways people lived along the Black Sea's western coast. Chronologically, the closest significant change in human behavior occurred at the start of the Neolithic, *ca.* 6500 calBC in the southern Balkans (i.e., Greece) and at 6000 calBC further north and west (in

V. Yanko-Hombach et al. (eds.), The Black Sea Flood Question, 515-536.

northern Bulgaria, southern Romania, Serbia, and beyond). The record of human activity before and after the sea-level rise in the latter, northern region, will be examined, and possible connections between coastal change and archaeologically documented patterns in human behavior will be discussed.

It is this writer's intention to question widely held (yet unsupportable) generalizations about (1) the causes of change in prehistoric societies, (2) the coherence of the Neolithic as an archaeological reality, and (3) attempts to marry significant geological events (such as the Black Sea rise) with changes in the patterns of past human activities. It is suggested in conclusion that most of the current debate over the Black Sea flood is misdirected. The significance of any transgression is not to be found in arguments about direct causes for major changes in human behavior, e.g., the emergence or spread of agriculture. The true significance is that the rise in sea level preserved a unique archaeological record of early Holocene southeastern Europe, and its documentation presents us with one of the most outstanding archaeological discoveries of the past half-century.

2. BACKGROUND TO THE FLOOD

Recent research into the paleohistory of the Black Sea has revolutionized our understanding of the marine geology of the region. Work by Popp (1969), Panin (1974, 1997), Skiba *et al.* (1976), Shcherbakov (1979), Shcherbakov and Babak (1979), and Popov (1983) established stratigraphies for shelf sediments and documented changes from brackish to marine fauna. The lack of reliable dating, however, prevented correlation of these changes with patterns of human behavior. The opening of the Black Sea to western researchers in the late 1960s beginning with Ross, Degens, and others (Ross *et al.* 1970; Degens and Ross 1972, 1974; Ross and Degens 1974) initiated a new era of scientific cooperation.

By the end of the 1990s, Ryan, Pitman, and others argued for a dramatic rise in sea level at 7150 BP. Their writings, which made popular and scientific headlines, suggested links between a flood and two important major events in world (pre)history and folklore: the introduction of settled farmers into central Europe during the Neolithic and the genealogy of the Flood Myth that appears in the Bible (Ryan and Pitman 1998; Ryan *et al.*1997a, 1997b, and Ryan *et al*, this volume). Relying on a change in strontium 87/86 ratios in Black Sea molluscs, Ryan's research team has recently re-dated the rise in sea level to 8350±70 BP uncorrected and uncalibrated (Ryan *et al.* 2003). This date calibrates to 7500 calBC, however, using a marine curve and local ΔR offset for the Black Sea of 67±26 years (Siani *et al.* 2000), the date calibrates to

7000–6640 calBC, rounded to the nearest decade (see Stuiver and Reimer 1993; Stuiver *et al.* 1998a, 1998b).

The important consequences of the model proposed by Ryan and his colleagues are that the Black Sea shoreline (as it existed immediately prior to the rise in the first quarter of the seventh millennium calBC) is currently ~ –80 m below the surface of the Black Sea, and that when the sea rose, it did so quickly without reworking and eroding the paleoshoreline. The archaeological significance is that the transgression conserved an early Holocene landscape on the large and flat (or slightly undulating) coastal shelf of Bulgaria, Romania, and Ukraine, which preserves clearly visible river channels that once flowed across the shelf towards the ancient shore.

In separate research, Robert Ballard's team has shown that the ancient beaches sloping gently toward the coastline contain fragile but well-preserved features (e.g., concreting pavement), thus demonstrating that when the sea level rose, it did so quickly without disturbing shoreline structures (Ballard *et al.* 2000). These flooded shelf areas should therefore harbor a superb archaeological resource from the early Holocene.

Ryan's hypothesis has been challenged, however. Proponents of an alternative model argue for a longer, more gradual rise in the Black Sea, starting much earlier in the post-glacial (Görür *et al.* 2001; Aksu *et al.* 1999, 2002a, 2002b; Hiscott and Aksu 2002; Hiscott *et al.* 2002; Kaminski *et al.* 2002; Major *et al.* nd; and see other papers in this volume, such as Hiscott *et al.*, Yanko-Hombach, Panin and Popescu, Balabanov, Chepalyga, Kuprin and Sorokin, and Shmuratko). From an archaeological perspective, supporters of both models have missed the boat. Unquestionably, changes in the character and level of the Black Sea are important, yet their value is not as stimuli for human action (at least not at the scale that has been proposed to date). Rather, the continually refined documentation of the Black Sea rise provides an extraordinary archaeological data set for understanding the critical changes that took place in southeastern Europe at the end of the seventh millennium calBC.

Assessing whether any connection exists between the Black Sea transgression and changes in the archaeological record requires an understanding of (1) how and where people were living before 7000–6640 calBC in this region, and (2) the principal changes in human behavior that occurred at the end of the seventh millennium calBC (i.e., from the start of the Neolithic). Furthermore, once the similarities and differences in lifeways before and after the rise in sea level have been clarified, it will be necessary to recognize that changes in ancient human behavior occurred over a variety of scales (e.g., spatial, chronological, perceptual) and thus must be understood over a similar range of scales.

3. HUMAN BEHAVIOR BEFORE AND AFTER 7000–6640 calBC

What was life like along the western coast of the Black Sea at 7000–6640 calBC when the sea level rose and inundated the wide coastal plain of Bulgaria and Romania? The responsible, but unsatisfying, answer is that we do not know. Despite more than a century of dry land excavation and field survey along the western coast of the Black Sea, very little is known about life in the period that preceded the appearance of the first communities of pottery-making, settled farmers, who define the Neolithic and who have been firmly dated no earlier than 6000 calBC in this part of southeastern Europe (see Whittle 1996; Bailey 2000).

Whereas research in other parts of Europe to the north and west has documented a dynamic late Pleistocene and early Holocene way of life (i.e., a Mesolithic), excavation and analysis in southeastern Europe has preferred to investigate the Neolithic and subsequent proto-historic and historic periods, neglecting the mysteries surrounding the activities of early Holocene societies (see discussions on archaeological research agendas in Velkov 1993; Bailey 1998; Kotsakis 1998). While detailed work has focused on the first appearance of Anatomically Modern Humans in the Balkans at 50–60 ky BP (Kozłowski 1982; Pawlikowski *et al.* 1990; Koz łowski *et al.* 1994; Ivanova and Sirakova 1995), we still have a poor understanding of the Upper Paleolithic (45–10 ky BP), and especially the period starting with the Last Glacial Maximum at 18 ky BP.[1]

Most frustrating is the major gap in the archaeological sequence between *ca.* 10,000 and 6000 years calBC: the Mesolithic record from southeastern Europe is patchy at best (a full discussion appears in Bailey 2000: 32–36). Based on the better studied Mesolithic record in other parts of Europe, one can predict that traces of corresponding activity in southeastern Europe will be ephemeral (i.e., short-term camps, hunting stations, and activity places). Such sites are most likely to be found along rivers, lakes, and seas, under significant accumulations of alluvium. Alluvial landscapes can experience high levels of change over relatively short periods of time (cf. Howard and Macklin 1999, Macklin 1999, and Bailey *et al.* 2002). There are also several examples of deeply buried Mesolithic occupations in the flat Pannonian alluvium of Serbia and Hungary. A few sites from such contexts are known from Romania and Bulgaria. The 100,000 sq km western Black Sea coastal shelf that was lost at the end of this poorly understood (and previously thought of as poorly populated) period suggests that perhaps the Mesolithic is not missing at all in southeastern Europe. On one hand, existing evidence argues that much of it lies beneath deep alluvium along inland river courses or around ancient, extinct lakes, buried to a depth and in positions that make their discovery by standard field survey almost impossible. (For a good example from another part of Europe, see Louwe Kooijmans 2001). On the other hand, perhaps the missing Mesolithic is better preserved and more accessible (i.e., without the levels of alluvium that have built up along the dry land rivers) on the surface and along the river valleys that can be seen on the flooded western coastal plain of the Black Sea.

3.1 Pre-7000 calBC Human Activities: the Missing Mesolithic

Based on the few well studied dry land sites from the period before the 7000–6640 calBC Black Sea rise, it is clear that the landscapes of southern Romania and northern Bulgaria were inhabited by highly mobile groups that hunted, fished, and gathered what they needed to live.[2] High biomass zones like those found along coasts, rivers, and other lowland water bodies (e.g., marshes, lagoons) would have provided a relatively abundant and easily accessible supply of food. Knowledge of the natural environment, its rhythms and schedules, and an understanding of how best to exploit (but also manage and conserve) available wild resources were the essential tools and technologies of people in this region at this time. Group mobility (as well as group fissioning and regrouping) was a necessary and good thing, providing flexible solutions to social and political, as well as the inevitable economic problems and crises.

Absent from this mobile existence was the greater level of attachment and commitment to particular parts of the landscape that came to characterize the relationships communities had with their natural and social environments in later Neolithic times. However, even before the Neolithic, it is clear that people identified particular places in specific landscapes with distinct meanings and uses. One such place is Pobiti Kamuni in northeastern Bulgaria (Dzhambazov and Margos 1960; Gatsov 1984a, 1984b; Bailey 2000:32–33, 128–129). Studies of Pobiti Kamuni reveal a long period of intermittent use stretching back before the Last Glacial Maximum, but importantly, also containing at least one period of activity dating to between the ninth and the seventh millennia calBC (Gatsov 1995:74). In truth, Pobiti Kamuni is not one isolated site; it is a large area (up to 50 sq km) consisting of individual surface scatters of flint tools and flint working debris. Analysis of the thousands of cores, tools, blades, and bladelets from the area suggests that while these were places to which people came to make flint tools, they were not the places at which people used those tools (Gatsov 1995). Pobiti Kamuni was but one of many places to and from which people moved within a larger mosaic of social and economic landscapes.

Different from Pobiti Kamuni are the better known semi-sedentary sites found during rescue excavations along the Danube in Romania and Serbia, where the river runs through the Iron Gates Gorges. At the Cuina Turcului rock shelter, for example, several hearths and assemblages of flint scrapers together with the debris from flint working document early Holocene activity areas that date as early as the tenth millennium calBC (Păunescu 1978; Radovanović 1996a:319). At other early sites in the Gorges, similar evidence for communal activities has been found. Some of the earliest sites containing durable architecture were concentrated in this stretch of the river between the eighth and the sixth millennia calBC. Small trapezoidal buildings have been recovered with semi-permanent superstructures (made from saplings, tree branches, and trunks) that sat on stone foundations arranged around carefully prepared plastered floors. With their longest side facing the river, these structures occupied narrow terraces squeezed between the river and the steep forested slopes leading to higher terraces. Faunal evidence indicates that significant quantities of large river fish were caught and consumed, that pigs, dogs, and red deer were also important food sources, and that food storage also took place (i.e., smoking fish). By the seventh millennium calBC, pottery and domesticated food stuffs are also present.[3]

The Gorges sites are dramatic. Not only do their buildings represent early appearances of oddly shaped (later Neolithic buildings are oval or recti-linear) durable, semi-permanent architecture, but they also contain significant concentrations of human remains (nine sites have produced bones from over 600 individuals) as well as strikingly decorated material culture (e.g., boulders pecked with human and animal appearances).

The Gorges sites are good examples of the kind of human record that is missing from the rest of the region.[4] The location of the sites is significant: tightly packed on thin strips of terrace land between the river and the slopes. Though the use of any one site or building was probably not continuous or permanent (perhaps seasonal use is the most accurate reconstruction), it is striking that people chose to build durable structures and to bury their dead in these places. In doing so, they were marking out particular locations along certain stretches of the river. The combination of access to large river fish on one side and red deer and pig from the wooded slopes behind would have made these places extremely popular. Indeed, these locations would have taken on particular significance and meaning to the people who used the strange little buildings, caught and ate fish, and buried their dead under the floors and in the spaces between structures (see especially Chapman 1992). These were special places, perhaps best understood by the juxtaposition of rich wild resources (river fish and deer) and the unique social, economic, and spiritual atmosphere created by the co-existence of the living and dead.

The greatest lesson to take from the Gorges sites may not be in the presence of burials here, or the early durable architecture, or even the exotically decorated boulders. Perhaps their greatest importance lies in the way that these sites were discovered and in their locations along the high-biomass strips of riverbank. If it was not for rescue excavations in advance of hydroelectric dam construction, there is little chance that the sites would have ever come to light, and without the Gorges sites, our understanding of the pre-Neolithic would be even poorer than it is. More importantly, the fact that the Gorges sites sit low down along the alluviated banks of a major river suggests that these are the types of places in which archaeologists should be looking for more evidence of this period (i.e., deeply alluviated locations along other watercourses). In this sense, the apparent emptiness of the pre-Neolithic landscape may be more a factor of heavy alluviation along many river valleys than it is a consequence of an actual absence of people and sites.

Equally important for understanding the apparent absence of the Mesolithic in the Balkans may be the changes in the Black Sea that occurred at 7000–6640 calBC. If the rise in the Black Sea removed from the region's landscapes a large coastal plain (cut by significant river systems, coastal lagoons and marshes), then that flooded plain must contain much of the missing pre-Neolithic record, perhaps in localized concentrations similar to what was found in the Gorges. At present, in addition to the Danube Gorges sites and the lithic resource sites such as Pobiti Kamuni, there is little else to help fill the otherwise empty landscape of this region throughout the period before the Black Sea transgression and prior to the start of the Neolithic. What is clear, however, is that the landscape (both the dry land and the currently submerged coastal plain) was not empty of people.

3.2 Post-6000 calBC Human Activities: the Neolithic

If the archaeological record of the pre-Neolithic in the northern Balkans is thin, the evidence after 6000 calBC is abundant. From this time, people started living in new ways, adopting new technologies to exploit novel species of plants and animals, using ceramic pyrotechnology to produce a wide range of vessels and other objects, and building new, durable places in which to live. Pigs, dogs, and cattle, which had been hunted for millennia, were supplemented by new species, such as sheep and goat, and all were managed as domesticates, though hunting remained a significant social and economic activity.

Following the introduction of simple architectural constructions, new forms of organization arose that linked social groups together in particular places, and in many cases, for long periods of time. People developed new perceptions of the landscape (of its products and of the rights of access to those products) and new conceptions of how individual people (and groups of people) should be associated or differentiated. Social and economic life focused on houses and households within bounded villages, though at the beginning of this period, organization of people across social and natural spaces was more fluid and open. In many places, less permanent camps emerged before longer lasting villages. The firing of clay to make pots, tools, and figurines was a novelty that had fundamental consequences. In addition to the importance of pottery vessels as a new container technology, fired clay became a major medium with which (and literally upon which) occurred an explosion of symbolic expression. Whittle (1996), Bailey (2000), Chapman (2000), and Tringham (2000) provide a more complete discussion of the Neolithic in this region.

3.2.1 Explaining the Neolithic Transformation

The differences in lifeways that distinguish the post-6000 calBC Neolithic from what came before (even based on the very thin record that we have for that earlier period) are clear and fundamental. Early explanations of the changes that occurred at 6000 calBC were principally economic, e.g., Childe's (1936) powerful model for a shift from a food gathering to a food producing system. They assumed that these changes were absolute, temporally abrupt, had the same character and cause as those that marked the appearance of the Neolithic in the southern Balkans at 6500 calBC (i.e., Greek Thessaly), and were the result of singular events, such as migrations of people or changes in climate.

Until very recently, it has been acceptable to speak of the early European Neolithic as a single way of life. Even as mid-twentieth century applications of nuclear physics to absolute dating made it clear that the Neolithic appeared in different parts of Europe at different times (Renfrew 1973), it was still assumed that the repertory of activities and technologies that made up the Neolithic was the same wherever and whenever it appeared. More sophisticated work has broken down the assumed homogeneity of the Neolithic (Whittle 1996; Thomas 1999). It is now clear that even within a single region, e.g., Thessalian Greece, there was significant variation in how people lived, even in how the same people lived at different times of the year or during the same season at a single site (Whittle 1996; Kotsakis 1999, 2005; Bailey 2000; Souvatzi 2000, nd; Halstead 2005). The same is true in the Northern Balkans (Greenfield 1993, 2000, nd a, nd b; Greenfield and Jongsma nd a, nd b; Greenfield *et al.* nd). The recognition of such variation within a single landscape makes any attempt to generalize across larger regions and between parts of Europe foolhardy at best.

Significant progress in research on the varied components of the Neolithic package (i.e., sedentism, ceramic pyrotechnology, animal and plant domestication) has opened up our understanding of the adaptation in new and increasingly complex ways. For example, it is no longer accurate to speak simply about the domestication of plants. There are many different scales of relationship between people and plants, and large-scale, field-based, crop cultivation of highly productive and robust species such as wheat and barley was a relatively late development in European prehistory. It is clearly not evident in the Early Neolithic of the northern Balkans (Greenfield and Jongsma nd a, nd b;

Greenfield *et al.* nd). Evidence for the wide-spread clearance of land for planting fields of crops does not appear in the archaeological record for many regions until the late Bronze or early Iron Ages, *ca.* 2500 calBC (Willis 1994, 1995). It is much more likely that the early selection and exploitation of particular plants were more heterogeneous processes that entailed small-scale use of both wild and managed species, which functioned through a combination of garden-sized plantings with sophisticated understanding of local wild resources.

In similar ways, research into human-animal interactions has ranged well beyond simplistic ideas about corrals or the farmyard and has exploded assumptions about the ways early Europeans exploited animals (Higgs 1972; Sherratt 1981; Ingold 1980; Halstead 1998). A general claim for the economic importance of domesticating animals has been replaced by subtle understanding not only of differing scales of animal exploitation (e.g., for primary and secondary products, via herding and grazing or hunting and managing wild stock), but also for different scales of consumption for the products of different sized animals (Greenfield 1988, 1991, 1993, nd a, nd b; Russell 1998).

Even in these well argued fragmentations of the long accepted but overly simplified perceptions of plants and animals in the Neolithic, most current explanations retain a level of generalization that smooths the data in an unrealistic way and presents the non-specialist reader with a charade that proposes that there was a particular Neolithic way of living that can be clearly and cleanly documented by the presence of domesticated plants and animals. Indeed, as debate continues to pick apart the increasingly fuzzy entity that archaeologists have called the Neolithic, it has become progressively clear that none of the constituents of the original package occur without significant variation across the regions (indeed even within a single region) and through the several millennia that make up the period.

Other recent arguments have (1) acknowledged the differences between the northern and southern Balkans (Halstead 1989; Greenfield 1993; Jongsma and Greenfield 2001; Greenfield and Jongsma nd a, nd b), (2) given more credit to the choices made by indigenous local pre-Neolithic inhabitants in adapting, adopting, and rejecting particular elements from the Neolithic package of technologies, plants, animals and social organs (Zvelebil and Lillie 2000), and (3) argued the probability of a less exact, less complete, and less absolute transition to the Neolithic way of living (Zvelebil 1986, 1994; Greenfield and Jongsma nd a, nd b; Greenfield *et al.* nd; and papers in Bailey *et al.* 2005). The concept of sedentism has been subject to similar assaults and critical redefinitions; it is no longer acceptable to assume that permanent buildings document year-round sedentism, that the inspiration for the construction of early architecture was simply the provision of shelter, or even that the same group of people could not exploit two apparently contradictory types of settlement systems, e.g., complementary villages of permanent houses and more mobile camps of temporary pit-features (see papers in Bailey *et al.*2005).

A major result of these refinements and redefinitions of the Neolithic and its constituent parts is that we can no longer speak of one Neolithic. There were many Neolithics that appeared and disappeared at different times and places. The distinction between what was Neolithic and what was not has been irrevocably blurred both in terms of chronological sequence (i.e., the permeability of boundaries with pre- and post-Neolithic phenomena) and in terms of an individual definition of typical Neolithic behavior (i.e., there are no universal activities represented by the terms animal and plant domestication or permanent sedentism; see papers in Bailey *et al.* 2005). A fundamental consequence of breaking down the Neolithic as an archaeological construct is the devaluing of the earlier, easy explanations for the origins of the Neolithic, which proposed a clearly defined package of goods, techniques, and knowledge that could have been brought into southeastern Europe by migrating groups from the Near East. There was no one origin to the Neolithic lifestyle in southeastern Europe, nor even a set of easily identifiable events that caused people to change their lives in ways that might appear to us (looking back 8000 years) as dramatic and radical. It is much more likely that the patterns of behavior that eventually accumulated and that archaeologists uncover today, are the result of very gradual alterations, testings, adaptations, rejections, re-alignments, regressions, and adoptions of a host of alternative components of living. Change was slow.

4. MULTIPLE SCALES OF HUMAN BEHAVIOR AND GEOLOGICAL EVENTS

If we accept that the transition to the Neolithic is most accurately understood as a gradual process, then what role might changes in the Black Sea occurring at 7000–6640 calBC have had within such an extended transition? Again, the answer is neither simple nor straightforward but lies in a rigorous examination of the relationship between geological events and changing patterns of human behavior. One of the greatest obstacles to an accurate understanding of the rise in the Black Sea is the still current assumption that geological events, physical (and cultural) topographies, and human communities are homogenous, i.e., that they are easily modeled entities that react in predictable ways making them understandable at a single general level. In reality, we can approach both geological and human events on many different scales.

For example, when discussing the effects of the Black Sea rise, geologists often make broad assumptions and conclusions that incorporate broadly what, in reality, archaeologists hold to be heterogeneous spatial scales. In this sense, this writer has a range of different (but not mutually exclusive) perceptions of the Black Sea and its western coast. Such perceptions include:

(1) *the largest scale*, upon which the Black Sea itself is an entity compared to the Aegean, the Caspian, or the Marmara,

(2) *the medium scale*, where the Turkish coast can be distinguished from the northwestern coast along the littoral of Romania and Ukraine,

(3) *the small scale*, in which one section of a regional coast is compared to another section of that same coast,

(4) *the local scale*, where a particular inlet or river mouth is distinct from an inlet or river mouth a few kilometers along the coast, and finally

(5) *the intimate human scale*, upon which the specific reaction of one individual is separated from the reaction of a second individual standing in the same place on that coast at the same time.

It is important to note that each of these spatial scales requires a separate detail of enquiry, a different set of information, and a different focus of research. Most critically, each will create a different image of the same geological event. Furthermore, each scale will provoke a different type of explanation and will provide a different level of understanding of both people and environment. In the context of any potential effects of a Black Sea transgression at 7000–6640 calBC therefore, we cannot simultaneously generalize across all of these different scales. Each scale of spatial dimension will entail a different understanding of the geological events and their potential human consequences.

In a similar way, one could investigate the potential effects of the rise in the Black Sea along a range of temporal scales. One could examine the immediate effects of the marine transgression. Indeed, this is the scale at which popular interest in the research of the Ryan team has focused: the BBC *Horizon* and *Ancient Voices* programmes reconstructed dramatic scenes of people thrashing about in the crashing waves of a flood. However, there is variation and a range of understanding even at this single temporal scale: how are we to define immediate? The span of time that the sea took to rise to its new level? The time it took for its new composition to stabilize? As a series of individual moments when each significant landform of the coastal shelf was submerged?

At the other end of the range of temporal scales (i.e., the longest term), one could examine the effects of all of the rises in sea level and all the changes of water from brackish to marine in the Black Sea within its complete paleohistory (i.e., from its origin as a geological entity to the present). Furthermore, between these temporal extremes, there are many other scales that would repay investigation, whether they are measured in subjective analytic terms (e.g., millennial, century, decade, annual, seasonal, monthly, daily), whether they follow climatic divisions (e.g., Bölling, Older Dryas, Allerød,Younger Dryas, PreBoreal, Boreal, and Atlantic), or whether they refer to a human life as the scale (e.g., the effects of changes over the life of an individual, 25–50 years, or over any number of generations of ancestors or descendants living on the coast).

The recognition that such ranges of scale exist can begin to clarify the causes of conflict over the different reactions (i.e., archaeological versus geological) to the proposed consequences of the rise in the Black Sea. Perhaps the grandest suggestion, that the transgression is the origin for the Biblical or Gilgamesh flood epics, sits comfortably at the highest, most general, spatial and temporal level. It applies across the Black Sea without reservation; it fits very badly, however, with the particular contexts of individual archaeological sites in their specific topographies and cultural locations, and it does not fit at all at the most intimate, human scale, outlined above.

Tremendous power dwells at the most general scale. It is the place of myth and legend, and it relies on trust and faith in the unmeasurable. It is untraceable, unquantifiable, and literally (and to its benefit) unscientific. However, it is also of least value for understanding the prehistoric past at the human scale. Explanations and proposals that thrive at the most general scales have extraordinary influence over our understanding of the past precisely because they are untethered to reality by the specificity of facts; they are beyond both proof and disproof. They rely on faith. Explanations developed at this level have particularly dangerous consequences, as they often invoke otherwise invisible mechanisms of migration, invasion, and culture change that have little if any palpable or documentable traces in the archaeological record. Worse, they are often deployed in racial and nationalist claims for rights to residence. To write about migrations in the prehistoric past is to construct generalized, mobile human communities for which there is not significant archaeological evidence.

4.1 The Human Scale

So how are we to understand the human consequences of the 7000–6640 calBC changes in the Black Sea? This writer suggests focusing on the human scale, acknowledging that there would have been many different effects and reactions to the changes in the Black Sea, in many different places (within the soon-to-be submerged coast and plain) and at many different times (during the inundation and in the immediate aftermath of the flooding), and especially, in the period of geological, environmental, and social stabilization that must have followed the loss of the coastal plain. Perhaps it is this latter period which is the most interesting for our understanding of consequences of the changes. If there were movements (or even displacements) of people, then they would have occurred during this period: re-organizations of previously mobile communities into new landscapes (perhaps similar to previous landscapes, though just as likely dramatically different). How long might this period have lasted? It is difficult to know. Was it even a coherent period that had a recoverable beginning or end? Did it even have one set of unshifting geographic foci?

Would there have been an instantaneous shift in lifestyle and a ready adaptation to new landscapes, new resources, and new special places? This is highly unlikely. Much more probable would have been a substantial period of stabilization, both for the composition and level of the Black Sea as well as for the dynamics and tensions between groups of people across the terrestrial landscape. Could this period have been the 650–1000 years that separate the inundation of the coastal plain and the first archaeological appearance of the new way of living that has come to be called the Neolithic? Perhaps. Would such a period of the gradual settling of water, land, resources, and people which fills the half millennium that leads up to 6000 calBC help to explain the exceptional coherence and longevity of the Neolithic way of life in this region for the next 2500 years? Though impossible to assess based on current understanding of the pre-6000 human landscape of Romania and Bulgaria, it is a highly attractive and stimulating possibility that should be addressed using new research strategies.

5. MOVING RESEARCH FORWARD

To conclude, if we want rigorously to investigate the effects of the Black Sea changes of 7000–6640 calBC, then we need a much better understanding of the record of human behavior along the ancient western coasts of the Black Sea as it existed previously. This record can be produced only by collaboration between dry land archaeologists and marine geologists. The aim should be to reconstruct the complete pre-7000–6640 calBC landscape, including the vast area of the currently submerged coastal plain. We need a programme of research that will take us past the debate of the rate and date of rise in sea level. Indeed, the research proposed here will resolve, once and for all, the arguments over the date of the Black Sea rise. We need a project to create a detailed map of the coastal plain (using the data gathered to date by all of the teams which have been working in the region) and which can coordinate a large-scale, detailed coring and submarine excavation programme. Key project goals would be to use the mapping data to identify those places on the coastal plain that have the highest probability of containing pre-7000–6640 calBC sites, and to core and excavate these locations in order to recover cultural, environmental, and datable material. The result would be the recovery of a long lost Holocene landscape, which this writer believes will contain the missing Mesolithic of this part of southeastern Europe, and which would provide a scientifically supportable answer to the question, what were the human consequences of the 7000–6640 calBC changes in the Black Sea?

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ENDNOTES

1. Of the more than 150 sites that Alexandru Păunescu includes in his recent corpus of Mesolithic and Paleolithic sites in southern Romania (i.e., between the Danube and the Carpathians), only 12 date to the end of the Upper Paleolithic (i.e., Epipaleolithic or Tardigravettian) or the Mesolithic (i.e., the Tardenoisian or the "Schela Cladovei Culture type") (Păunescu 2000:Table 1).

2. See examples at Franchthi Cave (van Andel and Vitaliano 1987; Perlès 1987, 2004; Jacobsen and Farrand 1988; Shackleton 1988; Wilkinson and Duhon 1990; Hansen 1991; Vitelli, 1993; Farrand and Jacobsen 1999; Vitelli and Dengate 1999), Grotta dell' Uzzo (Bietti 1981; Piperao 1981), and the discussion below about the Danube Gorges sites.

3. For a full discussion of the Danube Gorges, see Srejović (1967, 1969, 1972), Boroneanţ (1970, 1982, 1989, 2001); Prinz (1987); Voytek and Tringham (1989); Chapman (1989, 1992, 2000); Radovanović (1996a, 1996b); Whittle (1996); Bonsall *et al.* (1997, 2000); Radovanović and Voytek (1997); Borić (1999, 2002a, 2002b, 2003, 2005); Bailey (2000:62–71); and Borić *et al.* (2004). There are also more recent excavations at Schela Cladovei; their full publication is eagerly anticipated (see Boroneanţ 1973; Bonsall *et al.* 1997).

4. Of the 12 Mesolithic sites listed from Romania, all but three are from the Danube Gorges. The exceptions are Largu Cornul Malului and Largu Le Calentir (both in Buzau County), and Lapos Poina Roman (in Prahova County). For Largu Cornul Malului, see Păunescu (1979:517–518, 2000:114), Dragomir (1957:300–301, 1959:475–476). For Largu Le Calentir, see Păunescu (1979:518, 2000:114–115). For Lapos Poina Roman, see Păunescu (2000:118–130), Mogoşanu and Bitiri (1961:215–226), Mogoşanu (1960:127–128, 1962:145–151, 1964:337–350, 1978:349), Cârciumaru and Beldiman (1994:380), and Cârciumaru (1996:425).

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23

MORPHOTECTONIC DEVELOPMENT OF THE SOUTHERN BLACK SEA REGION AND THE BOSPHORUS CHANNEL

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Abstract: The southern coast of the Black Sea is bounded by the east-west trending Pontides and Strandjas mountain ranges, and between them flows the Bosphorus, a narrow channel representing the only seaway connecting the Black Sea with the Aegean-Mediterranean system. All of these morphological entities are young, formed during the Plio-Quaternary period. A major force in the formation of the Pontides has been north-south compression, generated in eastern Anatolia by the progressive northward advance of the Arabian Plate. At its western end, the Pontides terminate in the Marmara region, where north-south extension has been deforming the land since the Late Miocene, producing a number of grabens and horsts aligned east to west. The Strandjas represent one of these horsts that extends westward along the trend of the Pontides to form the Black Sea's southwestern margin. Located between the Pontides and the Strandjas are the lowlands around Istanbul, where deep erosion has produced a flat-lying surface. Beginning in the Pliocene, when the North Anatolian Fault (NAF) extended into the Marmara Sea basin, the Istanbul horst began to be deformed under a dextral shear stress regime. This regime generated conjugated pairs of oblique faults, which caused the Bosphorus to take shape as a zigzagging channel.

Keywords: Bosphorus, Pontides, North Anatolian Fault, horst and graben

1. INTRODUCTION

This paper describes the major morphological characteristics of the land south of the Black Sea and discusses their tectonic implications. A brief history of the development of the Bosphorus will also be presented.

V. Yanko-Hombach et al. (eds.), The Black Sea Flood Question, 537-569.

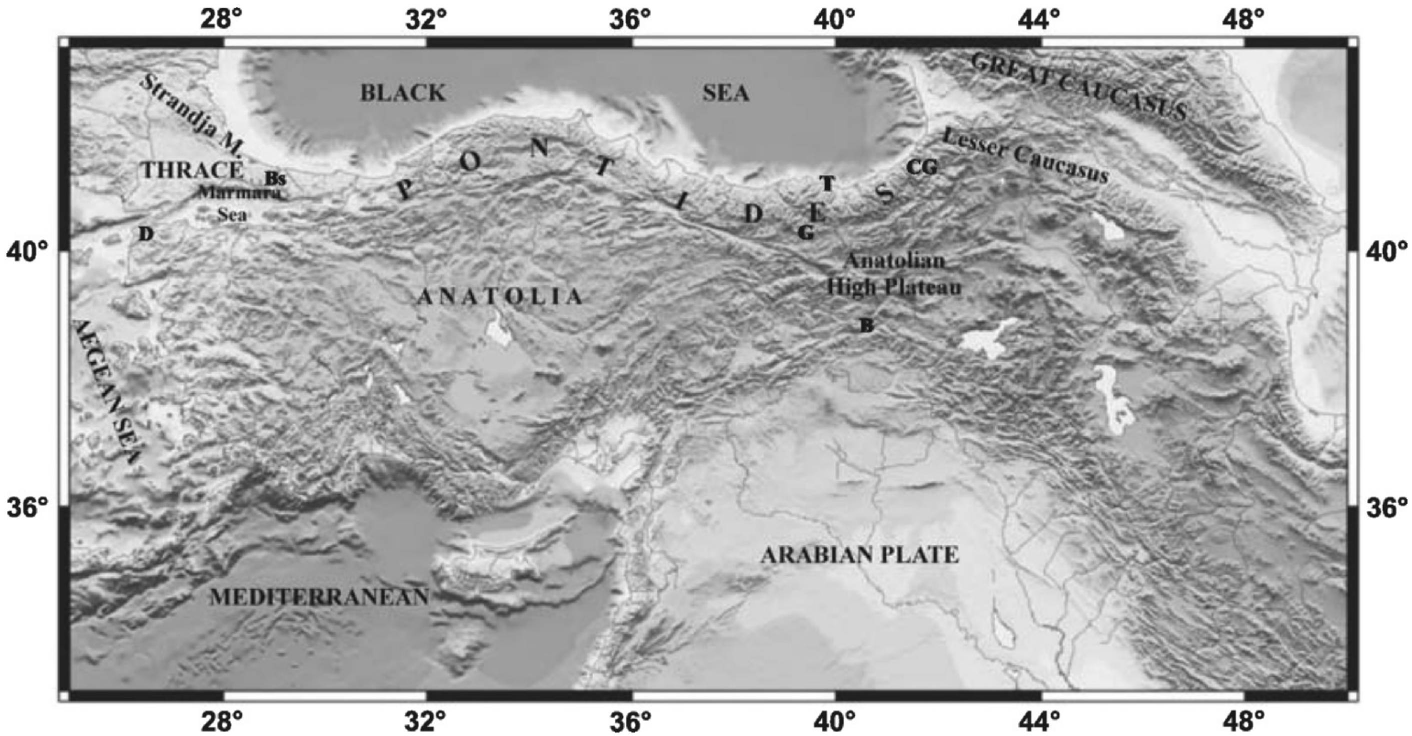


Figure 1. Topographic map of Turkey and surrounding regions. N-S compression in eastern Anatolia has created a high plateau, while N-S extension in western Anatolia has produced many E-W trending horsts and grabens. The Pontides range borders the Black Sea, and the NAF zone runs parallel just to its south. The Black Sea shelf is narrower in the south than it is in the west or north. Abbreviations: (D) Dardanelles, (Bs) Bosphorus, (T) Trabzon, (G) Gümüşhane, (CG) Çoruh River gorge, (B) Bingöl. (The map was derived from USGS-GT0P030.)

1.1 Geographic Setting

The Black Sea is bounded on the south by a long mountain range known as the Pontides (Figure 1), which run parallel to the Black Sea coast and form steep slopes. The Black Sea shelf, lying in front of the range, is relatively narrow, varying from 1 to 10 km in width. Beyond the shelf is a steep continental slope leading to a deep basin of >2000 m.

The Pontides rise to over 3500 m in the east, but they gradually diminish toward the west, terminating abruptly in the Marmara region. From here to Thrace, there are no high mountains. The Strandja mountain range then rises to the north of Thrace, following the line of the Pontides and forming a >2000 m high barrier in the southwestern corner of the Black Sea.

To the south of the Pontides lies the North Anatolian Transform Fault, or NAF, which spans the entire length of Anatolia. In the west, the NAF is buried beneath the Marmara Sea, but it reappears in the Gelibolu (Gallipoli) peninsula before entering the northern Aegean (Figure 1).

The Marmara Sea covers a young and deep basin connecting the Black Sea with the Aegean-Mediterranean system through two straits, the Dardanelles in the southwest (D in Figure 1) and the Bosphorus in the northeast (Bs in Figure 1). Both straits have bedrock sills covered by recent sediments, and current minimum depths are 70 m in the Dardanelles and 32 m in the Bosphorus. Water connections between the Mediterranean system and the Black Sea have been determined in the past by two major factors: (1) global sea level changes associated with climatic variations, and (2) rapid morphological development of the region, which appears to be closely connected with strong tectonic forces.

1.2 Tectonic Evolution

The rich geological data of the Pontides and the southern Black Sea coastal zone have been documented in an extensive literature, however, morphotectonic descriptions are meager and restricted to a few sentences. Hamilton (1842) was one of the first earth scientists to visit the region and report his observations. Oswald (1912) was the first to give an accurate account of the regional morphology. He noted the major longitudinal fault steps that run parallel to the coast and descend the steep mountain flanks toward the Black Sea. He also described minor faults cutting these fault blocks transversely. Stratil-Sauer (1927) produced a few geological maps and recorded some geological and geographical descriptions. A number of local studies were carried out in the mountains, mainly in the 1950s and 60s, among which may be mentioned the works of Erinç (1953), Inandık (1955), Ardel (1968), and Gattinger *et al.*(1962).

Important contributions to knowledge of the morphotectonics of the Eastern Pontides were made by a group of German geologists, including Schultze-Westrum and Zankl (Maucher *et al.* 1962). They studied the Giresun region and made observations similar to those of Oswald, but they reached somewhat different conclusions. They asserted that, from the Mesozoic to the present, the Pontides developed as Hertz-type broken mountains, in which vertical tectonic events were dominant. Their explanation was applicable only to the latest stage of uplifting in the development of the Pontides, however, as it ignored the fact that the Pontides are located in the middle of the Alpine-Himalayan Mountain belt. Their conclusions have been proven wrong by most subsequent studies (Şengör *et al.* 1980; Yılmaz and Tüysüz 1984; Robinson *et al.* 1995; Yılmaz *et al.* 1997).

The Pontides are, in fact, young mountains, formed after the Miocene Epoch during the Neotectonic period (Şengör 1980). Their appearance followed a chain of events associated with the gradual disappearance of the Tethyan Ocean (Şengör and Yılmaz 1981).

The basin that contains the present Black Sea emerged earlier, during the Paleotectonic period (Şengör 1980), and its evolution may be summarized as follows. Basin formation was initiated early in the Cretaceous Period (Görür 1988). Possibly, it occurred during the collapse of the previous orogen that had resulted from the complete elimination during the Dogger Epoch (Middle Jurassic, 180–159 my BP) of the PaleoTethys Ocean from what is now eastern Europe, the Black Sea, and surrounding regions (Şengör *et al.* 1980). The subsequent convergence of colliding continental plates caused crustal thickening during the ensuing Malm Epoch (Late Jurassic, 159–144 my BP). With the collapse of this orogen, rifting began during the Berriasian Age (144–137 my BP) of the Early Cretaceous (Yiğitbaş *et al.* 1999), which led to drifting during the Aptian-Albian Ages (121–99 my BP) (Yılmaz *et al.* 1997).

The basin became oceanic through the progressive drifting of a back arc basin behind a volcanic island arc that grew along the line of the present Pontide chain, due possibly to the southward migration of the arc front. As a result of this, arc region cooled and collapsed during the Campanian Age (83–71 my BP) of the Upper Cretaceous. During the Lutetian Age (49–41 my BP) of the Middle Eocene, the NeoTethys Ocean swelled to cover a wider area that included Anatolia (Yılmaz *et al.* 1997). A gentle bulge along the axis of the Pontides appeared in connection with the elimination of the NeoTethys Sea during the late Eocene and Oligocene Epochs (Yılmaz *et al.* 1997). The sea of that period, delimited by the Pontide bulge in the south, is commonly referred to as the ParaTethys (Jones and Simmons 1997).

By the Early Miocene (23–16 my BP), the Tethyan Ocean extended from the Indian Ocean to the Eastern Mediterranean, but it had already receded along the Bitlis-Zagros suture of southern Anatolia. A sea still covered eastern and southwestern Anatolia after the Tethys Ocean disappeared (Yılmaz 1993). The steady northward advance of the African Plate against the resistant oceanic plate located beneath the Paratethys Sea began to squeeze all of eastern Anatolia (Şengör and Kidd 1979). Initially, the entire eastern part of Anatolia was elevated as a block during the Langhian Age of the early Middle Miocene (16–15 my BP). During the Late Miocene (11–5 my BP), most of eastern Anatolia was occupied by interconnected lakes, lying just above sea level, and within which sediments of low energy environments, primarily fine detritals and carbonates, were deposited (Şaroğlu and Yılmaz 1986). This evidence suggests that the Pontides had not yet risen and, therefore, had not yet begun to act as a source of clastic material.

Toward the end of the Late Miocene, the region passed through a denudational phase that created a region-wide erosional surface. In eastern Anatolia, extensive remains of this surface have been recognized and used as a time reference with respect to subsequent events. Any major morphological entity laid down above it may be regarded as a younger feature (Yılmaz *et al.* 1997; Yılmaz 2002). Starting from the Late Miocene, sporadic development of volcanoes occurred across this flat erosional surface, and extruded lava flowed over large expanses of it (Yılmaz *et al.* 1987, 1998). The rise of eastern Anatolia, particularly the Pontides and, along its periphery, the Bitlis-Zagros system, postdates the development of the erosional surface. Evidence for this can be seen in the initiation of increasingly coarser and more voluminous clastic materials, derived from these highlands, being transported into the surrounding basins, mainly the Black Sea basin. The first clastic sediments are not older than the early Pliocene (Robinson *et al.* 1997).

The ongoing north-south compression affecting eastern Anatolia produced east-west trending folds, reverse faults, and conjugated pairs of northeast and northwest trending strike-slip faults (Şengör and Kidd 1979; Yılmaz *et al.* 1987), and as a consequence, the crust has become progressively shortened and thickened. The NAF and the East Anatolian Transform Fault (EAF) finally developed (Figure 2) when crustal thickness exceeded 46 km (Zor *et al.* 2002), beginning in the Pliocene (Şaroğlu and Yılmaz 1991; Yılmaz 2002).

The continental wedge delimited by the EAF and the NAF represents a discrete tectonic entity called the Anatolian Plate. This Anatolian Plate began to transfer the energy stored by the over-thickened crust of the constricted eastern Anatolian region by protruding westward from the point of intersection of the two transform faults (McKenzie 1972, 1976; Şengör 1980).

The north-south compression created by the northward movement of the Arabian Plate (18–20 mm/y according to Reilinger *et al.* 1997) was blocked by the old and resistant, trapped oceanic lithosphere underlying the Tethys basin (Zonenshain and Le Pichon 1986; Banks and Robinson 1997). The two peripheral belts–the Eastern Pontides-Lesser Caucasus belt in the north and the

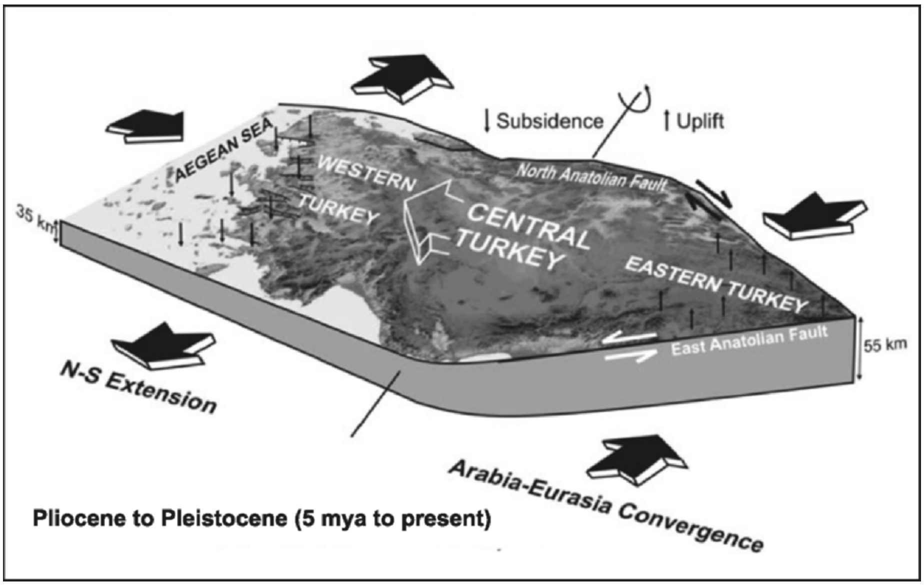


Figure 2. Block diagram displaying the present deformation pattern of Anatolia. Compression in eastern Anatolia produced two transform faults, the NAF and the EAF, with the central part forming a discrete Anatolian Plate, which is forced westward to generate N-S extension in western Anatolia and the Aegean; modified after Şengör (1980).

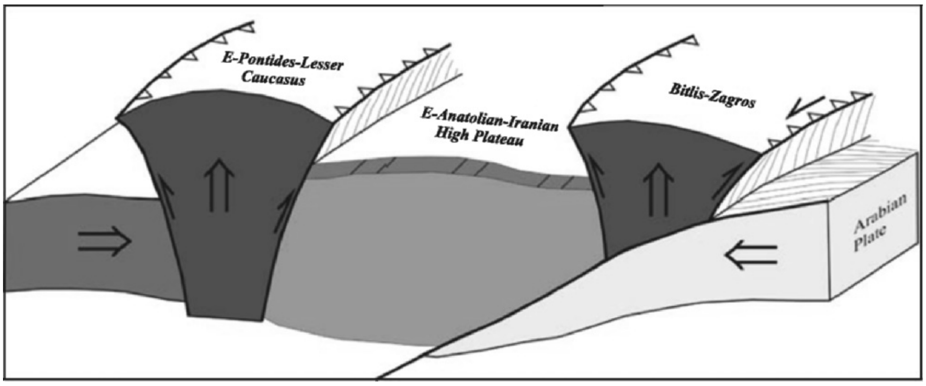


Figure 3. Block diagram displaying the major morphotectonic characteristics of eastern Anatolia and surrounding regions. Arrows indicate relative motions. Note the ramp basin characteristics of eastern Anatolia; its central region is over-thickened and slightly bulged. Accordingly, major rivers run sub-parallel to the flanks of the mountain belts for hundreds of kilometers.

Bitlis-Zagros belt in the south–also had beneath them rigid high-grade metamorphic rocks (Yılmaz 1993; Yılmaz *et al.* 1997). In contrast, the Eastern Anatolian–Iranian high plateau overlies the Eastern Anatolian Accretionary

Complex (EAAC), which formed due to the elimination of the Tethys Ocean (Şengör and Yılmaz, 1981) and acted like a cushion between the Pontides and the Bitlis belt (Şengör and Yılmaz 1981). As a result, when squeezed by compressive forces, these peripheral belts were thrust to the north (Robinson *et al.* 1997; Gökaşan 1996) as well as to the south (Şaroğlu and Yılmaz 1986), becoming elevated at a higher rate (1.1 mm/y) than the central eastern Anatolian region (0.6 mm/y). Thus, a giant ramp basin has been formed (Figure 3).

2. MORPHOTECTONICS

2.1 The Pontides and Its Vicinity

The northern edge of the narrow Black Sea continental shelf of Anatolia gives way to a steep continental slope leading to a deep remnant oceanic basin (Neprochnov *et al.*1974; Zonenshain and Le Pichon 1986; Banks and Robinson 1997). Along its southern edge, the entire shelf is delimited by the 1500 km long Pontides mountain chain that extends from the Lesser Caucasus to the Marmara.

On the map (Figure 1), the mountain chain displays a sinusoidal pattern, arcing northwards at both ends and southward at the center. These curves along the axis of the mountain range are apparently related to regional tectonics. During the late Miocene, the north-south compression produced an extremely constricted and thickened region around the city of Bingöl in central eastern Anatolia (B in Figure 1). Axes of major folds and faults follow the virgation pattern of a locally squeezed morphology and fan out from this area of the Anatolian high plateau somewhat like a sheaf of wheat tied in the center. The Pontides conform closely to the pattern. The chain trends to the northwest on the western side of the Gümüşhane-Trabzon line (G to T on the map), and it trends to the northeast on the eastern side.

Along the northern edge of the eastern Anatolian high plateau, the Pontides rise steeply like a wall, separating the plateau from the Black Sea. As a result, the area's seaward drainage has been diverted. The major rivers flow along the valley axes, sub-parallel to the flank of the mountain chain, for hundreds of kilometers before cutting across the mountain toward the sea.

The present mountain chain is made up of long, thin, normal fault-bounded blocks (Figures 4 and 5) that define the morphotectonic character. The width of the fault blocks varies greatly, from a few hundred meters to a few kilometers, but these faulted blocks form discrete steps that steeply descend the flank of the mountains (Figure 5).

The connection between the closed drainage system south of the Pontides (within the eastern Anatolian ramp) and the Black Sea appears to have been

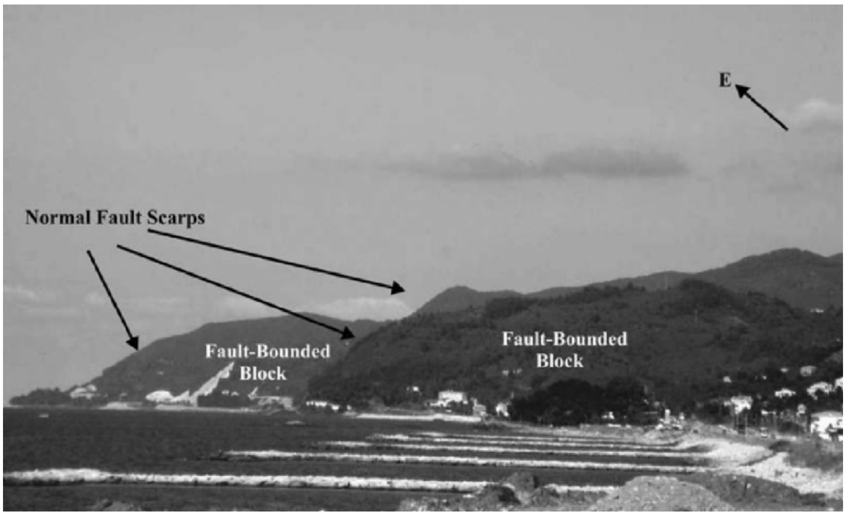


Figure 4. A view from the eastern Pontides range. Steep slopes correspond to normal faults which divide the flank of the mountains into long and thin, fault-bounded blocks.

Figure 5. A closer look at fault-bounded rectangular blocks along the northern flank of the eastern Pontides. Inward dipping of the top surfaces of the fault-blocks indicates the listric nature of the normal faults along which the flank of the mountain descends seaward in abrupt steps.

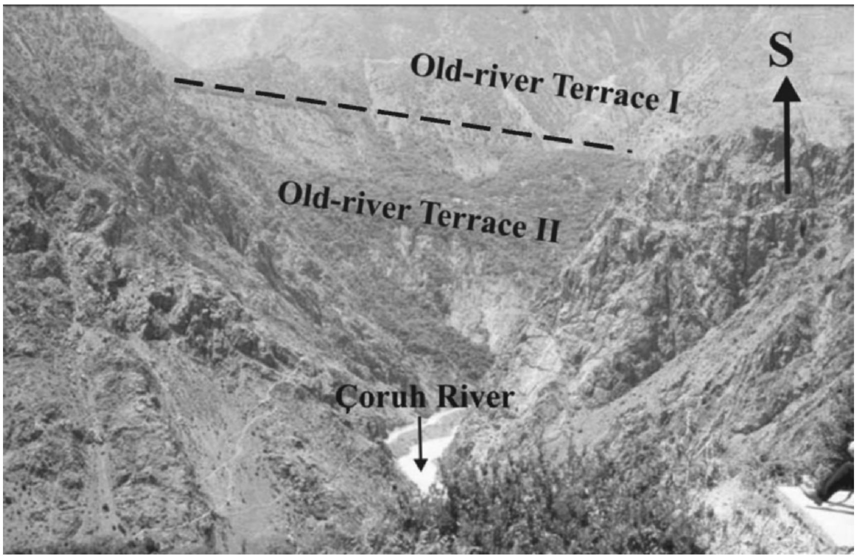


Figure 6. A view from Çoruh Gorge in which old hanging river terraces may be identified at different elevations.

established recently. Unconsolidated fluvial sediments, possibly late Quaternary in age, have been deposited along the Çoruh Gorge (CG in Figure 1), which contains the only river to cut through the Pontides in the eastern part of the range. No older sediments are identified within the Gorge. The profile of the Çoruh River valley is broad at the higher topographic levels, and becomes narrower, deeper, and steeper downward. The changes from broad to narrow profile are sharp and display evidence of at least three episodes of mountain uplift, including a number of elevated river terraces at varied heights along the Gorge (Figure 6). The top of the mountains, which were elevated to over 3000 m, preserve remnants of the flat-lying erosional surface (Figure 7).

On the northern and southern flanks of the Pontides, the planes of the normal faults are slightly listric (curved) in nature, as reflected by back-tilting of the fault blocks (Figure 5). Long and thin major fault blocks are terminated at some distances by secondary faults that commonly trend at right angles to the major faults. They also commonly display normal fault characteristics with slight oblique-slip components. In Figure 8, coast-parallel major faults are clearly visible in the background, and in the foreground, the secondary faults can be seen cutting the continuity of the long blocks and dividing them into smaller blocks. Together, the longitudinal and transverse faults divide the mountains into fault blocks of various sizes, and as a result, a complicated domino-fault pattern has been produced. This pattern determines the irregularly formed morphological features of the present coast, including discontinuous erosional terraces observed

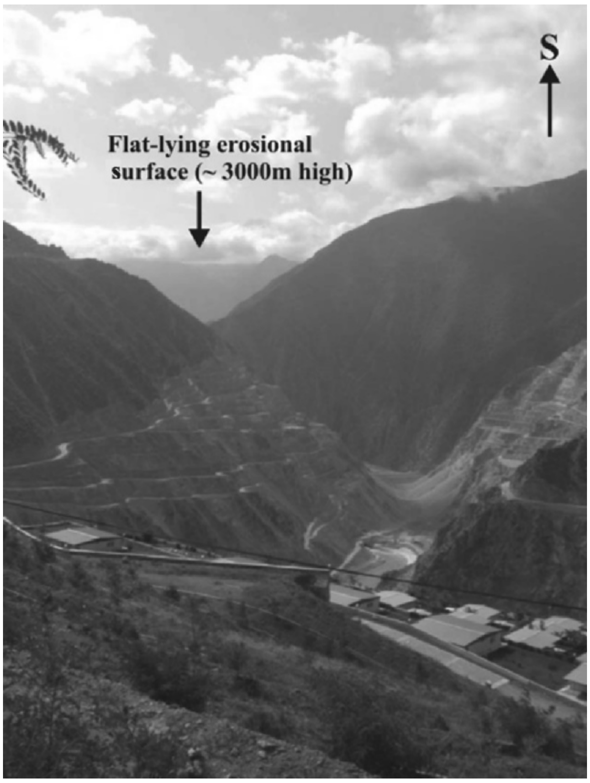


Figure 7. A southerly view of Çoruh Gorge. Slopes become narrower and steeper downward, indicating rapid uplift and severe downward incision, and a flat-lying erosional surface can be seen at the horizon.

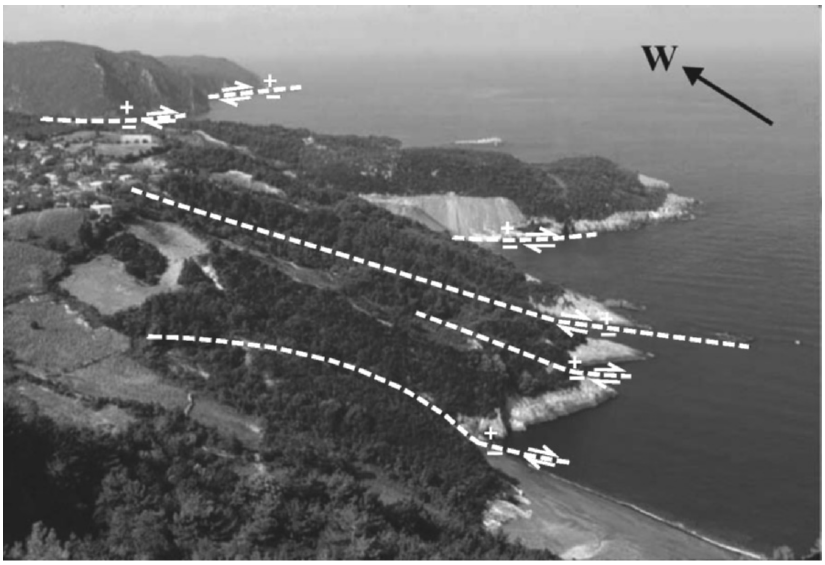


Figure 8. A view of the irregular morphological features within the Black Sea coastal region.

at varying heights up to 30m, and beaches that have been elevated as well as isolated in narrow pockets.

Making stratigraphic correlations from one beach or terrace to the next is difficult, because each one has a slightly different developmental history as indicated by the different rates of uplift measured in the coastal region. The uplift rates vary from about 0.5 to 3 mm/y (Tchepalyga *et al.*1997; Aytaç 2003). Such diversity appears due to the varying rates of descent of the normal fault blocks, though the rate of mountain uplift has been rather uniform overall.

2.2 The Istanbul Region

The Pontides gradually lose their morphological identity toward the west and disappear completely in the Marmara region of western Anatolia (Figure 1), where the forces of north-south compression that affect eastern Anatolia are replaced by those of north-south extension (Figure 2). These drastic changes in the tectonic regime are reflected in the morphology of the two regions. The extensional province to the west is represented by a number of east-west trending grabens and horsts. Western Anatolia and the Aegean have been undergoing north-south extension at the rate of more than 2.5 cm/y since the late Miocene (Yılmaz *et al.*2000).

In the Marmara region, the Pontides terminate abruptly along an approximately north-south trending strike-slip fault zone, the Adapazarı-Karasu Fault (Figure 9), associated with the NAF (Yiğitbaş *et al.*2004). This branch of the NAF possesses a westerly-directed thrust component, which sustains the height

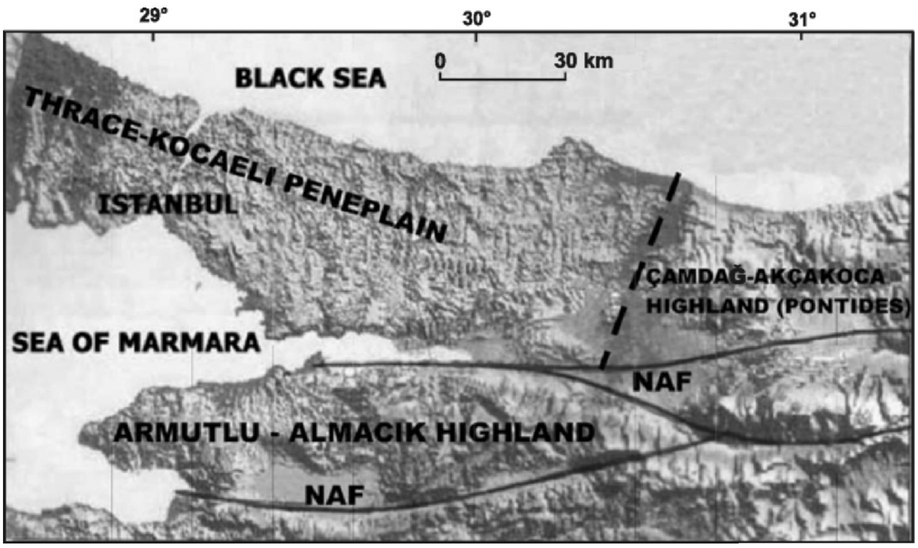


Figure 9. Topographic map of northwestern Anatolia. The Pontides end sharply along a N-S trending fault zone boundary. Along the straight coastline of the Black Sea is the Thrace-Kocaeli Peneplain. Branches of the NAF are buried under the Marmara Sea.

of the mountain in the extensional province and maintains a sharp contrast in morphology across the fault. The fault separates the mountainous region in the east from the low-lying flat land with a dendritic drainage pattern in the west. This flat plain extending along the Black Sea coast represents deeply-eroded Paleozoic rocks on which an erosional surface known as the Thrace-Kocaeli Peneplain (Pamir 1938) was formed during the Late Miocene (Emre *et al.* 1998; Yılmaz *et al.* 2000).

The flat land of the Istanbul region is located on a horst block (Figure 10A and B) bound in the south by a fault zone forming the steep northern slope of the deep basin occupied by the Marmara Sea. The fault is considered to be the western extension of the NAF (Demirbağ *et al.* 2003), however, it is thought to have formed initially in the early Pliocene as a normal fault during the opening of the Marmara Basin, and only later was it captured by the NAF system, possibly during the latest Pliocene-Pleistocene (Yılmaz 1997; Emre *et al.* 1998; Yaltırak 2002; Gazioğlu *et al.* 2002; Gökaşan *et al.* 2003).

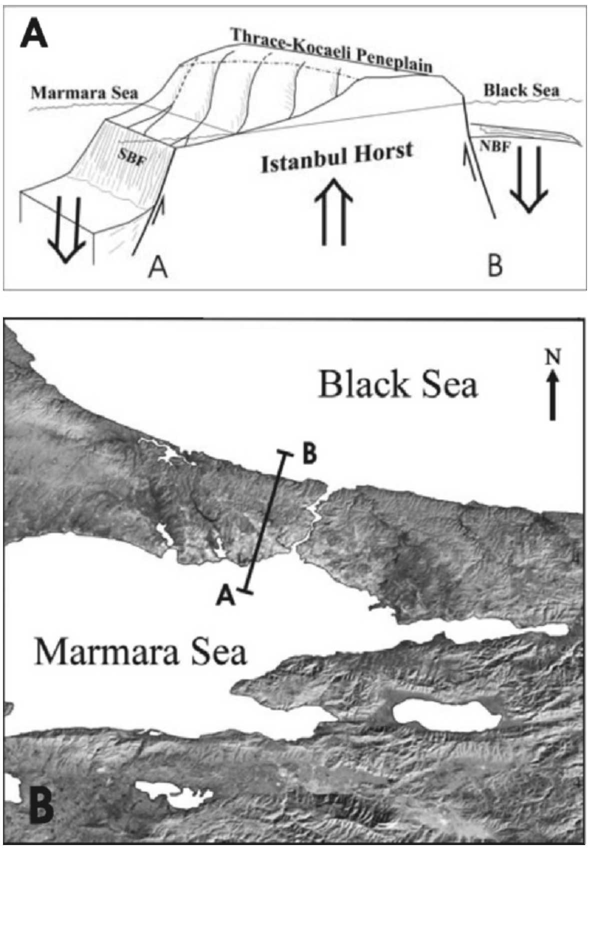
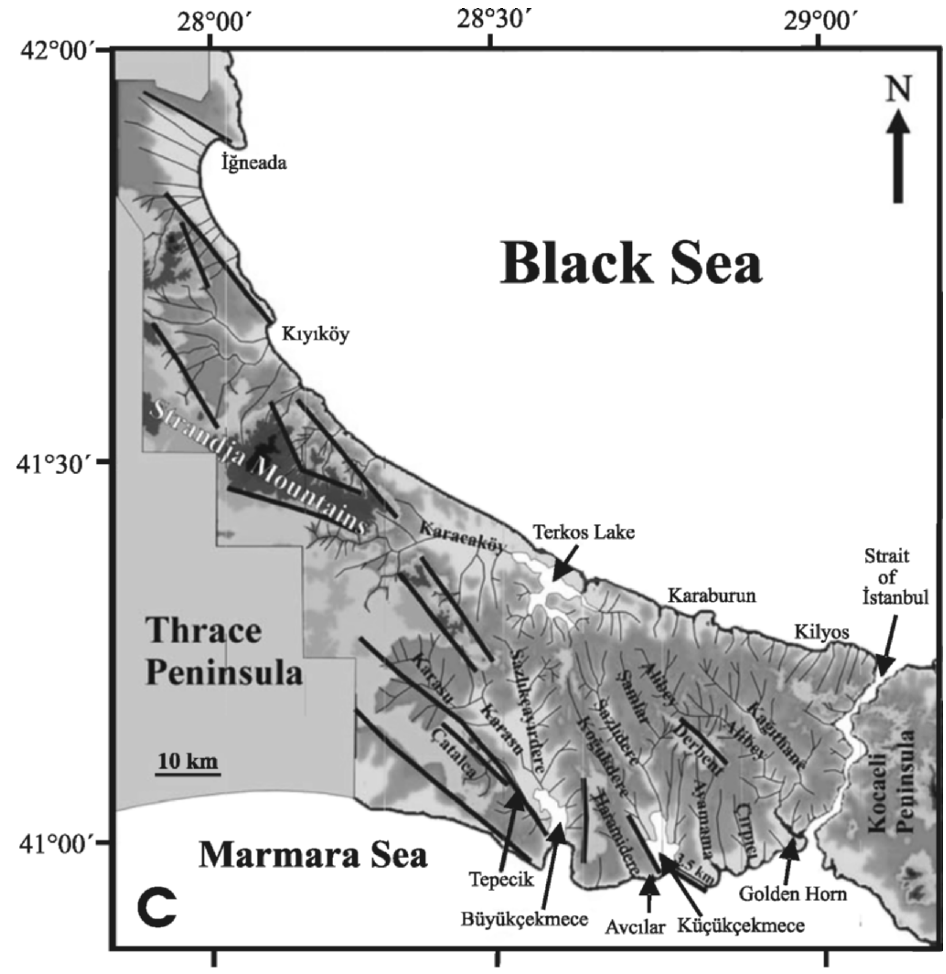


Figure 10. **A.** Schematic diagram displaying the horst block formed between two major faults. The NBF extends along the linear Black Sea coast, while the southern fault (SBF) corresponds to the northern shelf break of the Marmara Sea, which runs along the extension of the NAF. The flat surface at the top of the horst represents remnants of the Thrace-Kocaeli Peneplain. Diagonal lines represent the major trend of the drainage system, which is determined by the structural elements. **B.** Topographic map of the land surrounding the eastern part of the Marmara Sea. The NW and NE trending major streams may be noted on both sides of the Istanbul region. A-B: approximate direction of the cross section in Figure 10A. **C.** (Facing page) Map showing major lineaments and some morphological features of Istanbul and surrounding areas. Parallel streams, scarps, and displacements on the shoreline can be seen (after Gökaşan *et al.* 2002).



The Istanbul horst possesses a northern edge separating it from the Black Sea. It is a steeply dipping oblique fault zone, the Northern Boundary Fault, or NBF, which corresponds approximately to the shelf break (Figure 10A). The Bosphorus, the narrow strait connecting the Black Sea with the Aegean-Mediterranean system, is located on the Istanbul horst. The strait is a 0.7 to 3.5 km wide, zigzagging channel with an average depth of about 50 m and a maximum depth of 110 m. Along the channel, there are two bedrock sills, each located close to its two entrances. The sill near the northern entrance of the Bosphorus lies at –60 m and is situated in the submarine valley that is considered to represent a pre-existing channel extending into the Black Sea (Oğuz *et al.* 1990; DiIorio and Yüce 1999). The southern sill lies at –32 m and also extends along a midchannel that deepens toward the Marmara Sea (Alavi *et al.* 1989). The two sills are thought to contain Upper Quaternary sediments (Gökaşan *et al.* 1997), but sediment infill along the entire Bosphorus can reach locally to thicknesses of more than 130 m.

Remnants of the erosional surface may be observed on both sides of the Bosphorus at different elevations. On the eastern (Asiatic) side, dome-shaped monadnocks of resistant Lower Paleozoic quartzites rise up to 300 m above the peneplain. On the western (European) side, the erosional surface is better preserved and thus more recognizable. Seismic data reveal the extension of this surface northward where it lies 200 m beneath the sea level of the Black Sea (Demirbağ *et al.* 1999), having descended on the down-thrown block of the NBF. The dip-slip displacement along the fault is estimated to be greater than 400 m.

Close to the southern entrance of the Bosphorus is a little bay or inlet known as the Golden Horn (Figures 11 and 12). It is in reality an estuary formed at the mouth of the Kağıthane-Alibeyköy River, which drowned about 7400 years ago (Meriç 1990). Towards the plateau in the west, the river displays a clear meandering profile above the peneplain surface (Figure 13). This erosionally flattened plateau and the river above it have been elevated more than 200 m with respect to present sea level.

Figure 11. Aerial photo of the Golden Horn and its entry to the Bosphorus.

Figure 12. Aerial photo of the meandering profile of the Kağıthane-Alibeyköy River and its entrance to the Golden Horn. The Marmara Sea is visible in the background.

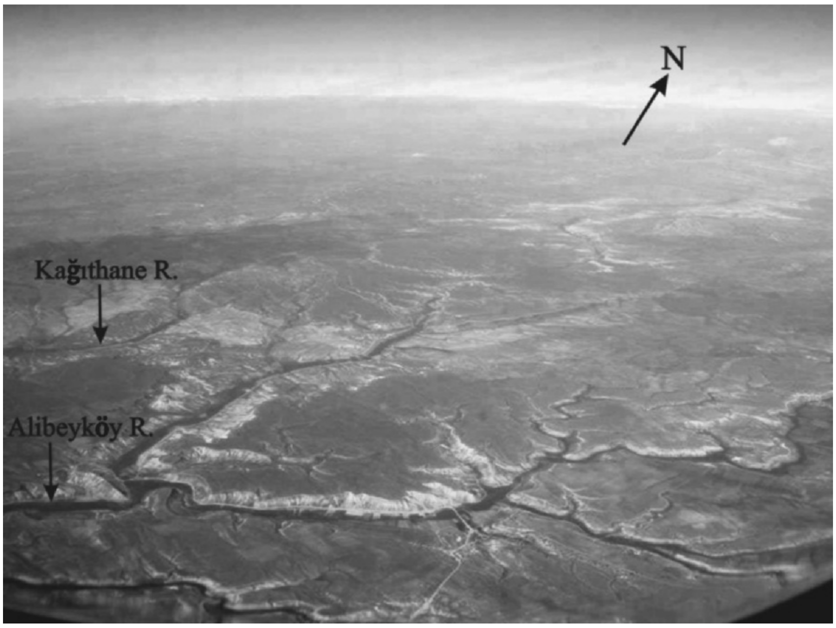


Figure 13. Aerial photo of the dendritic and meandering profiles of the Kağıthane-Alibeyköy River above the Thrace-Kocaeli Peneplain.

Figure 14. A view of the eastern side of the Bosphorus. Note the flat-lying erosional surface at the horizon and the gentle slopes toward the Bosphorus.

This rise has created a broad valley profile with slopes gently descending to the Bosphorus from the two surrounding plateaus (Figure 14). Such morphology may indicate that a river valley existed previously in place of the Bosphorus. This view is supported by data from seismic reflection profiles across and along the Bosphorus, which reveal a buried channel in the middle of the strait (Figures 15A, B) that may be interpreted as a paleo-river valley, the floor of which lies about 50 meters below present sea level. An approximately east-west trending drainage divide is also detectable (Figure 16) that appears to have been coeval with the channel. Prior to the opening of the Bosphorus, the drainage divide separated the two main rivers, one flowing north to the Black Sea and the other south to the Marmara Sea. The watershed crossed the Bosphorus between Anadolu Kavağı and Rumeli Kavağı. It runs very close to the Black Sea in the western plateau but comes closer to the Marmara Sea on the eastern plateau.

The paleo-river valley under the Bosphorus was cut into Paleozoic bedrock in the north and Upper Cretaceous bedrock in the south, but the overlying valley fill is primarily Upper Quaternary sediments, judging from recovered core samples (Meriç 1995; Algan *et al.* 2001).

A precise dating for the development of the Bosphorus is unknown, however, the following data may offer a relative age:

(1) Close to the eastern flank of the Bosphorus are outcrops of Early Miocene lacustrine sediments (Figure 17) and Late Miocene Paratethyan

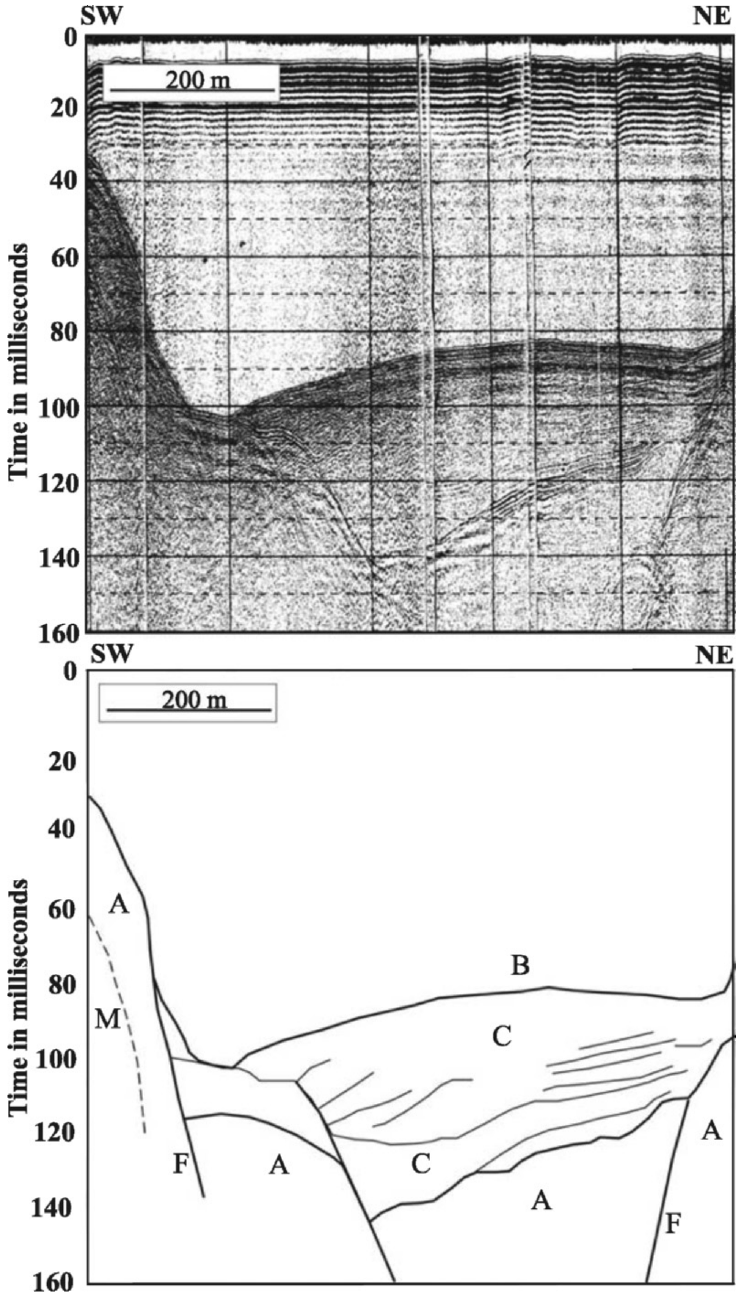
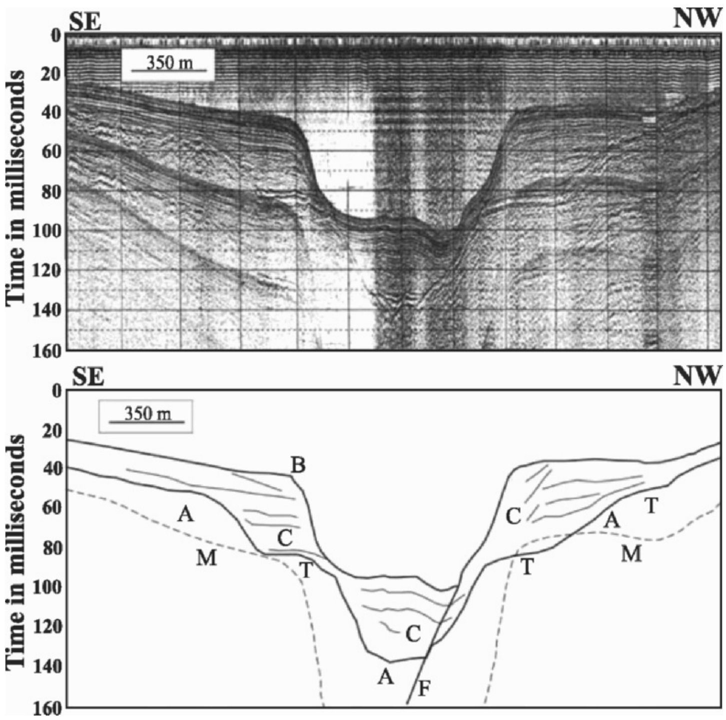


Figure 15. **A.** Seismic section across the Bosphorus showing the paleostream valley in the center. Note the fault-bound edges of the Bosphorus (after Gökaşan *et al.* 1997). Letters A to M correspond to stratigraphic layers: (A and M) basement, (B and C) young, soft cover sediments, (F) fault. **B.** (Top of following page) A seismic section across the Bosphorus showing a buried paleostream valley in the central part (after Gökaşan *et al.* 1997).



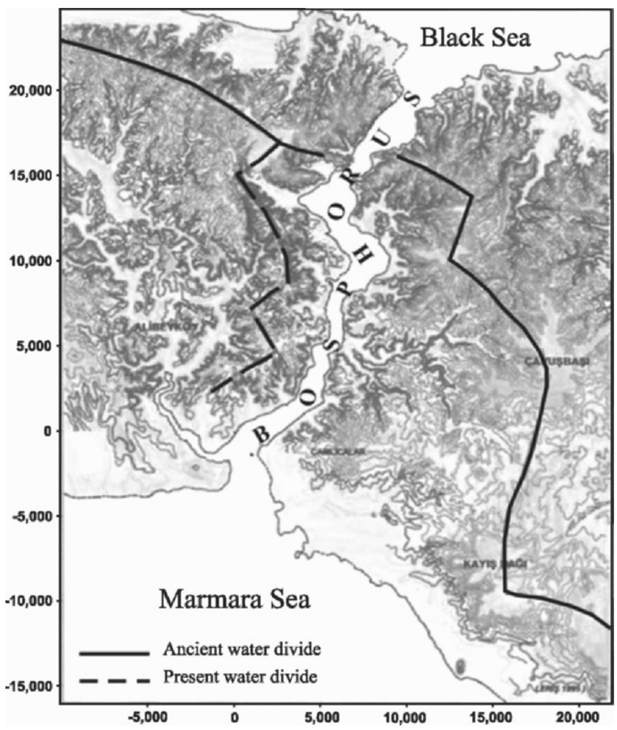


Figure 16. Topographic map of the Istanbul region. The continuous line is the paleo-water divide interrupted by the Bosphorus. The meandering characteristics of the streams above the peneplain surface are noticeable.

sediments (Sayar 1988; Steininger *et al.* 1985), which are now observed at 250 m above sea level (Oktay and Eren 1995).

(2) Adjacent to the Bosphorus on both sides, and located atop hills that have been elevated over 200 m, are outcrops of loosely cemented or unconsolidated fluvial sediments of the Pliocene (Figure 17).

(3) Lacustrine sediments around Karaburun on the Black Sea coast indicate an uplift of about 75 m along the northwestern coast of the Istanbul peninsula during the late Quaternary (Oktay *et al.* 2002).

According to this evidence, the opening of the Bosphorus postdates the Pliocene. Fluvial sediments similar to the Pliocene unit have been distinguished by seismic data from the central part of the Bosphorus, lending further support for a post-Pliocene development (Gökaşan *et al.* 1997).

Along the flanks of the Bosphorus, three different slope angles are readily distinguished from top to bottom (Figure 18). The slopes are gentle above, but they become steeper toward the bottom, with each change occurring at a sharp and angular inflection, indicating a morphological unconformity. These features document a rapid and episodic uplift of the Istanbul horst, while the stream valleys and tributaries adjoining the Bosphorus on both flanks display hanging valley characteristics.

Along these valleys, more than three distinctly developed elevated terraces may be distinguished. The steep lower slopes are commonly planar. Shallow seismic studies along and across the Bosphorus reveal that some of these planar surfaces correspond with the faults (Figures 15A and 15B). They cut the sea floor and determine the present bathymetry of the Bosphorus (Yılmaz and Sakınç 1990; Gökaşan *et al.* 1997; Oktay *et al.* 2002). Gökaşan *et al.* (1997) describe a number of northwest-southeast oriented right-lateral strike-slip faults, which appear to have played an active role in the development of the present morphology and drainage system (Figures 10B, C). They also identified faults on their seismic profiles displaying right-lateral displacement of the shoreline. These data also indicate an active tectonic role in the development of the Bosphorus. Such a view is further supported by numerous consistent lineaments that determine present parallel drainage and linear slopes on the plateaus surrounding the Bosphorus (Figure 10c). According to Gökaşan *et al.* (1997, 2003), some units, distinguished seismically, are syn-tectonic sediments deposited during the Pleistocene Epoch (Figure 15b), and some are ruptures formed during the Würm glaciation.

Paleozoic rocks are exposed on both sides of the Bosphorus, within which a number of sets of faults and lineaments have been identified and mapped by previous research (Alavi *et al.* 1989; Ketin and Kıran 1989; Gökaşan *et al.* 1997; Oktay *et al.* 2002). The faults are difficult to date. They may have formed at any time during the long period of orogenic evolution of the region, however, some of them extend into the Bosphorus and form its slopes (Figure 19).

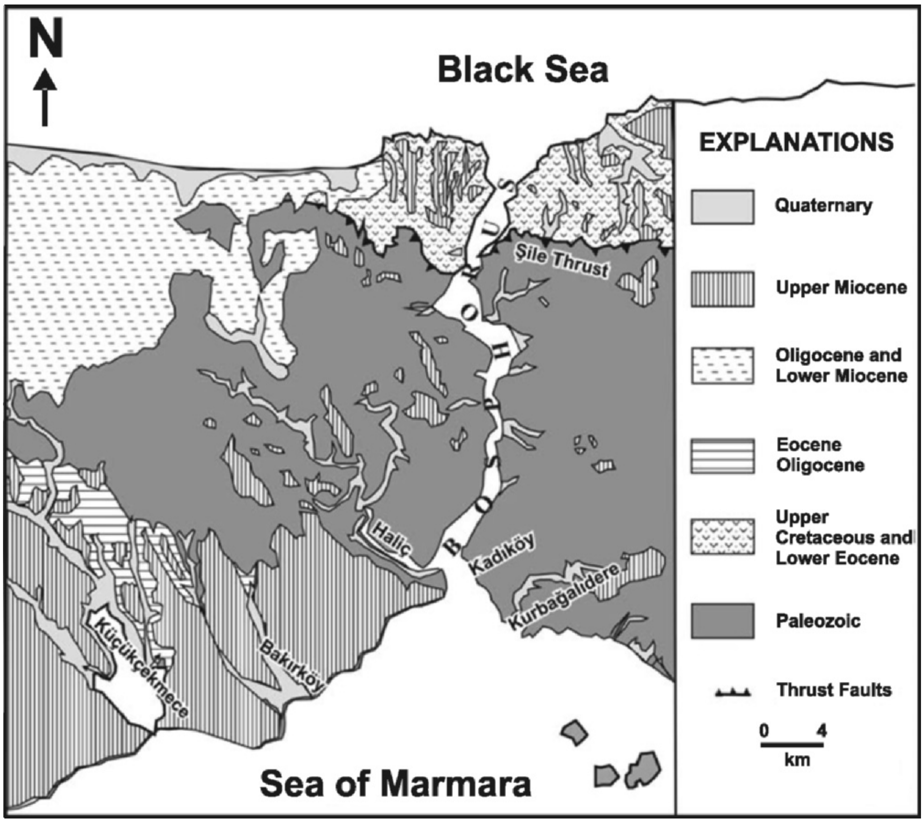


Figure 17. Geological map of northwestern Anatolia.

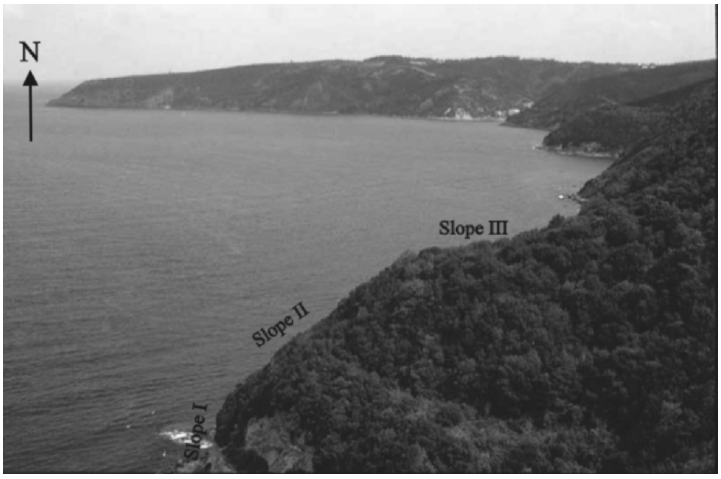


Figure 18. A northerly view of the flank of the Bosphorus near the Black Sea. Note the sharp changes in slope angles from the flat-lying top to the steep lower inclines that make three different slopes readily distinguishable.

In fact, on the map, most of the present boundaries of the Bosphorus are perfectly linear, which may reveal something of their structural origin (Figures 19 and 10B).

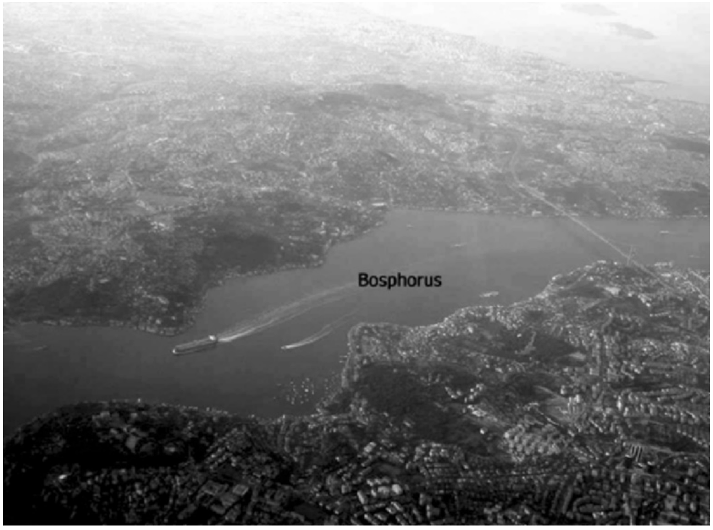


Figure 19. Zigzagging path of the Bosphorus. Note the straight shorelines, the continuity of the straight coastline into the land at the far side of the photograph, and the similarity of coastlines on both shores.

The zigzagging boundaries on both sides of the Bosphorus are similar, matching in size, shape, and pattern, so that if the two sides were merged, they would fit each other closely. Therefore, the faults on both flanks collectively may represent conjugated pairs of oblique faults, displaying strike-slip and dip-slip components. Such a fault pattern suggests generation by dextral shear stress.

3. DISCUSSION: THE OPENING OF THE BOSPHORUS

One of the earliest explanations for the opening of the Bosphorus was proposed by Strato Physicus (Straton Lamprakenos, d. *ca.* 270 BC) (from Rodier 1890). Starting in the 19th century, the subject was taken up by scientists such as Von Hoff (1822), Hochstatter (1870), Philipson (1898), and Andrussov (1900). Pamir (1938) tackled the problem more extensively, and his research was followed by many others, including Erinç (1940), Yalçınlar (1947), Scholten (1974), Stanley and Blanpied (1980), Yılmaz and Sakınç (1990), and Oktay and Sakınç (1993). Most of these studies were based on limited field and seismic data, however.

According to Ryan *et al.* (2003), the Black Sea was a lake throughout most of the past three million years. This freshwater basin was invaded by seawater during a few brief intervals. The episodes of freshwater incursion were associated with the overflowing of Caspian and Central Asian lake water into the Black Sea (Zubakov 1988; Chepalyga, personal communication, this volume). During Quaternary glacial stages, lake water sometimes rose considerably and spilled into the Marmara Sea through the Bosphorus (Stanley and Blanpied 1980; Lane-Serff *et al.* 1997). During the Neoeuxinian phase (22–9 ky BP), the Black Sea was also isolated from the Mediterranean Sea and remained a freshwater lake (Fedorov 1971; Ross and Degens 1974) because global sea level had fallen below the sill of the Bosphorus.

During this period, the shelf areas of the Black Sea were exposed to subaerial erosion and excavated by rivers until about 10–9 ky BP, when glacial meltwater filled the Black Sea lake (the Neoeuxinian Lake) and inundated the shelf areas. Offshore seismic data show that deposits of the Flandrian transgression lapped onto the older eroded units (Demirbağ *et al.* 1999).

A more complex and progressive reconnection that spanned the past 12,000 years has been suggested by Aksu *et al.* (1999, 2002). In their view, persistent Holocene outflow from the Black Sea into the eastern Mediterranean occurred. Sedimentological and chronostratigraphical studies of coastal plain sediments from all around the Black Sea indicate that the latest flooding by Mediterranean waters occurred sometime between 9 (Yanko-Hombach 2003, and personal communication) and 7.5 ky BP (Ryan *et al.* 1997, 2003).

In light of the accumulated data, the following history may be envisaged for the development of the Bosphorus.

A peneplain was formed in the Istanbul region (Figure 20A) during the late Miocene (Emre *et al.* 1998; Le Pichon *et al.* 2001; Oktay *et al.* 2002; Yaltırak 2002). The Istanbul horst rose (Figure 20B) under the north-south extensional regime (Gökaşan *et al.* 2003) during the late Miocene-early Pliocene, prior to the entry of the NAF into the Marmara region during the Pliocene-Pleistocene. In this new tectonic regime, the horst located between the NAF and the NBF, which runs above the Black Sea coast (Demirbağ *et al.* 1999; Oktay *et al.* 2002), began to be deformed under a dextral shear stress (Figure 20C). The submarine river valley extending along the Bosphorus supports the view of the right-lateral displacement along the NBF because it curves to the west and terminates abruptly by a steep scarp leading to the deep basin. There is no canyon lying in front of this submarine paleo-river valley. However, a few kilometers east is a submarine canyon curving in the opposite direction with respect to the river valley (Lericolais, personal communication). Similarly, the steep walls of the canyon head may also be related to fault zones which have caused lateral displacement along the canyon axis and truncated the canyon head. The canyon system cuts through all the seismic units and erosional surfaces, which indicates that its formation is not yet complete. Its shelfward erosion is in an incipient stage (Demirbağ *et al.* 1999; Algan *et al.* 2002; and Lericolais, personal communication).

The major driving force of the dextral shear stress regime affecting the Istanbul horst may be attributed to the NAF (Oktay *et al.* 2002), along which a slip rate of about 2 cm/y is presently measured by Global Positioning System (GPS) (Reilenger *et al.* 1997); this rate is estimated to have been relatively constant during the last million years (Barka and Kadinsky-Cade 1988).

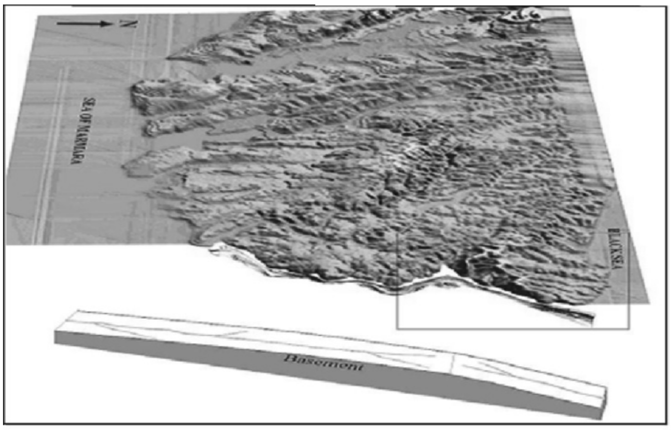
The NBF, although locally displaying oblique displacement with right-lateral slip and dip-slip components (Demirbağ *et al.* 1999), appears to represent the resistant side, or the less active side, of the Istanbul horst with respect to the NAF. Under the dextral shear, conjugated pairs of faults and joints were initially generated. During the following stages, when the horst was forced to rotate clockwise, the blocks bounded by the faults (synthetic and antithetic shears) caused offset of the basin boundary (Gökaşan *et al.* 1997, 2002), and thus the fault-bounded blocks began to move away from one another.

Gökaşan *et al.* (1997) and Oktay *et al.* (2002) have recently provided field and seismic data from the Bosphorus and surrounding areas while stressing the role of tectonic forces in the development of the Bosphorus. In fact, they followed the two-stage hypothesis initially proposed by Yılmaz and Sakınç (1990). According to this idea, a river valley formed first, then faulting enlarged the valley and brought it to its present form.

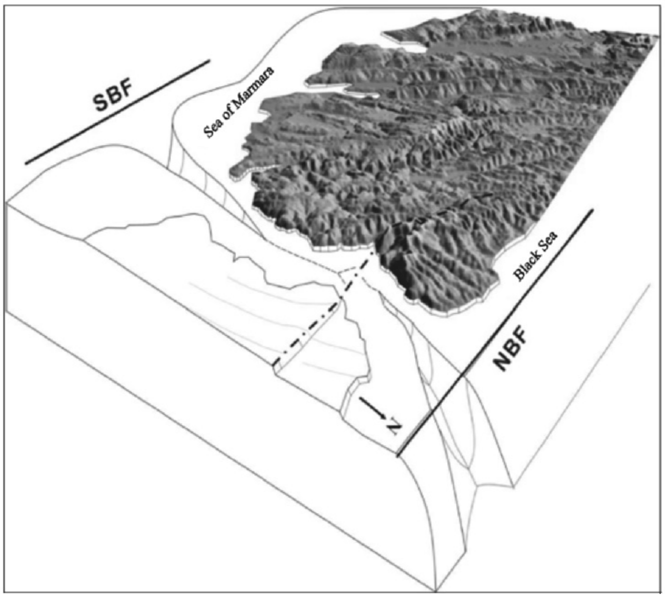
To explain the tectonic stage of Bosphorus development, Oktay *et al.* (2002) proposed a block rotation model analogous to the Dibblee model (1977) that was applied to the San Andreas fault and neighboring areas by Nur *et al.* (1989) and Peacock *et al.* (1998). According to this hypothesis, both sides of the Bosphorus rotated clockwise independently between the NAF and NBF. The major difference between the Oktay *et al.* (2002) model and the one proposed in this paper is that a counter-clockwise rotation seems to be more suitable to the dextral shear stress field that was generated in the region than the clockwise rotation of the blocks suggested by Oktay *et al.* (2002). Oktay *et al.* suggest independent rotations of the two flanks of the Bosphorus, while it is here suggested that reverse rotation operated on both sides and produced the same deformation. Moreover, Oktay *et al.* attribute an insignificant role to erosional processes, however, this paper proposes that, in addition to tectonics, deep incision and erosion contributed substantially to the development of the Bosphorus.

Coeval with the clockwise rotation, the Istanbul horst began uplifting (Demirbağ *et al.* 1999). A clear indication of this uplift is the presence of elevated, meandering river valleys, hanging river valleys located on both sides of the Bosphorus, and elevated river terraces. Hanging terraces along the Black Sea coast near Şile offer further support for the uplifting (Erinç 1953; Erol 1979; Erinç 1982; Ertek 1995).

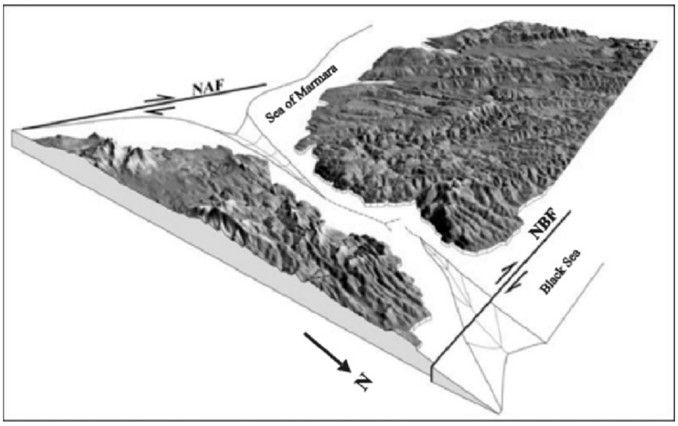
Figure 20. **A.-E.** Block diagrams showing sequential stages of the geological development of the Bosphorus.



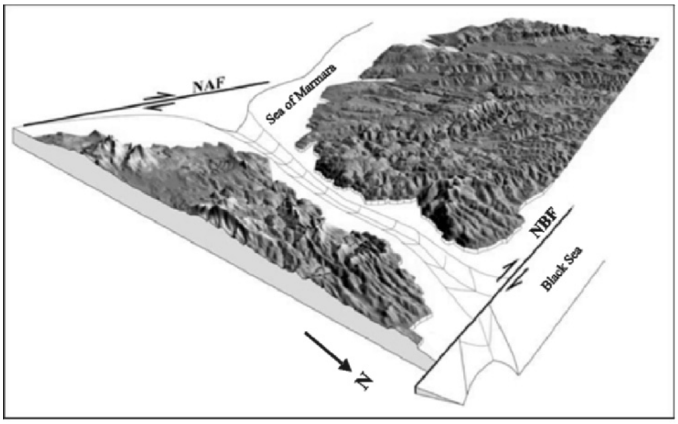
A. A flat-lying erosional surface known as the Thrace-Kocaeli Peneplain developed during the Late Miocene.



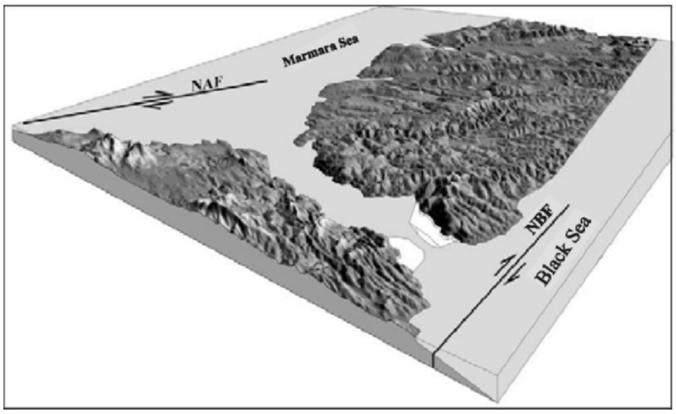
B. Under an extensional regime, a horst bounded by two fault zones was formed. The Southern Boundary Fault (SBF) corresponds to the shelf break of the Marmara Sea basin, and the Northern Boundary Fault (NBF) runs parallel to the Black Sea coast. On the horst, two streams formed, separated by a water divide. The horst rose, and downward incision increased progressively, reactivating some of the previously formed faults.



C. The NAF extended to the Marmara region, possibly during the Pliocene-Pleistocene interval, and captured the SBF. Under the influence of the NAF, the Istanbul block began to be deformed by dextral shear. During this period, the NBF behaved as a passive, or less active, site. The dextral shear generated conjugated pairs of oblique faults along which the fault blocks began to move away from each other; this led to the opening of the Bosphorus.



D. At the river mouths, small inlets or bays developed. Headward erosion along the two streams advanced, leading to the formation of one continuous valley. During this stage, the presence of two basement sills prevented water circulation along the valley.



E. Sea water circulation began when global sea level rose above the basement sills of the Bosphorus. Available data on the timing of this event are not conclusive, and thus, the topic is widely debated.

Judging from the amount of displacement across the faults bounding the Istanbul horst (Demirbağ *et al.* 1999; Oktay *et al.* 2002) and the present elevation of the Miocene, Pliocene, and Quaternary sediments (Oktay *et al.* 2002), the uplift rate for the area from Istanbul to Thrace (Yaltırak *et al.* 2002) may be estimated at 0.3–0.5 mm/y. Upward movement of the Istanbul horst accelerated downward incision along the valleys as well as headward erosion along the two major rivers that occupied the place of the Bosphorus today (Figure 20C). The erosion became severe, and when it reached an advanced stage, the two valleys merged to form one continuous valley (Figure 20D).

Water circulation within this valley did not begin as soon as the strait developed. In its initial stages, the valley floor stood at a much higher elevation than the water levels of the Marmara and Black Seas. Small bays or inlets (Figure 20d) developed at the mouths of the valleys (Gökaşan *et al.* 1997), where core data reveal 17.5 meters of Paleoeuxinian (*ca.* 300,000 year old) lagoon sediments (Meriç *et al.* 2000). Demirbağ *et al.* (1999) also documented seismic and surface evidence supporting the existence of a lagoon or bay environment at this time.

In subsequent stages, deep carving along the valley floor advanced, and the present steeply sloped flanks began to form (Figure 20C). Water incursion onto the valley floor began from two ends, based on evidence from cores taken of sediments that were deposited in brackish water ponds (Meriç *et al.* 2000; Algan *et al.* 2002) about 32–22 ky BP (Çağatay *et al.* 2000).

According to Ryan *et al.* (2003), the Black Sea dropped 105 to 140 m after the close of the Pleistocene. Such a depth would be far below the basement sills of the Bosphorus. According to Yanko-Hombach (2003), however, the lake level rose rapidly from –150 to –50 m during the 3000 years between 16 and 13 ky BP. During this interval, the depressions in the central Bosphorus became the site of an isolated Neoeuxinian lake in which lacustrine sediments were deposited until about 12 ky BP (Algan *et al.* 2001). Global sea level was lower than the bedrock sill depth of the Dardanelles Strait, and as a result, Mediterranean water could not enter the Marmara Sea to reach the Bosphorus (Stanley and Blanpied 1980; Çağatay *et al.* 2000).

Indications of the onset of water circulation along the Bosphorus are meager and restricted to a few cored sediment samples. Possibly, circulation began during the post-glacial Neoeuxinian transgression (Shimkus *et al.* 1978; Ryan *et al.* 2003), when the freshwater lake covering the Black Sea basin rose considerably (Figure 20E) (Chepalyga, personal communication). According to Algan *et al.* (2001), the Black Sea was receiving glacial meltwater, and its water level was rising after the Younger Dryas (Stanley and Blanpied 1980). This rise, and consequent outflow from the Black Sea, is supported by the formation of sapropels in the Marmara Sea at 10–6.5 ky BP (Algan *et al.* 1999, 2002).

With the continuous rising of global sea level, the Marmara Sea was flooded with Mediterranean water at about 12 ky BP. Ryan *et al.* (1997) and Ryan and Pitman (1998) suggest an abrupt Mediterranean-Black Sea connection possibly between 9 and 7 ky BP. This matter is still controversial and extensive discussions appear in recent and current literature. The reader is referred to those works (Görür *et al.* 2001; Aksu *et al.* 2002; and Ryan *et al.* 2003) as well as other papers in this volume for review of the details.

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SEA-LEVEL CHANGES MODIFIED THE QUATERNARY COASTLINES IN THE MARMARA REGION, NORTHWESTERN TURKEY: WHAT ABOUT TECTONIC MOVEMENTS?

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Abstract: Geological and paleoceanographic studies in the Marmara and Black Seas indicate that climate changes during the Quaternary modified coastlines and influenced prehistoric human life there. The effect of active tectonics on coastal modification is poorly known, however, and much less appreciated. This paper summarizes field evidence and relevant literature for the Marmara region in an effort to explore the modifying role of active tectonics on the area's coastlines. The Marmara region is a tectonically very active belt in northwestern Turkey with an extensive paleogeographical evolution related to escape tectonics and development of the strike-slip North Anatolian Fault (NAF). Here, east-west and northeast-southwest oriented depressions have created lakes, marshes, and fluvial systems since the Late Pliocene and Pleistocene along splays of the NAF concurrent with Plio-Pleistocene uplift and denudation. Post-Pleistocene uplift and denudation, accompanied by seismic movements during historical times, have formed wide alluvial plains suitable for human settlement. The distribution of Paleolithic sites exclusively on high terraces, and the total absence of pre-Bronze Age (5000 to 3000 BP) sites, are particularly noteworthy in the area, which includes the Bosphorus, Izmit Bay, Sapanca Lake, and the Sakarya River. Active tectonics continue to reshape the region, as is plainly demonstrated by the 1999 Marmara earthquake sequence.

Keywords: Active tectonics, Marmara region, northwestern Turkey, North Anatolian Fault, seismic movement, uplift, absence of pre-Bronze Age sites, coastal modification

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1. INTRODUCTION

Turkey lies in the tectonically active Alpine-Himalayan mountain system, which is the Earth's youngest orogenic belt. Convergence of the Arabian and African plates with the Anatolian and Eurasian plates (Figure 1) has led to the creation of

(1) mountain ranges, such as the Zagros, the Caucasus, and the Pontides to the northeast and east of the Anatolian block (Dewey *et al.* 1973; McKenzie 1978; Şengor and Yılmaz 1981),

(2) a 1500-km long right-lateral strike-slip fault zone that accommodates the westward escape of the Anatolian block to the north (Şengör 1979; Şengör and Canitez 1982; Şengör *et al.*1985; Westaway 1994), and

(3) a subduction zone along the southwestern boundary of the Anatolian block (Le Pichon and Angelier 1979, 1981).

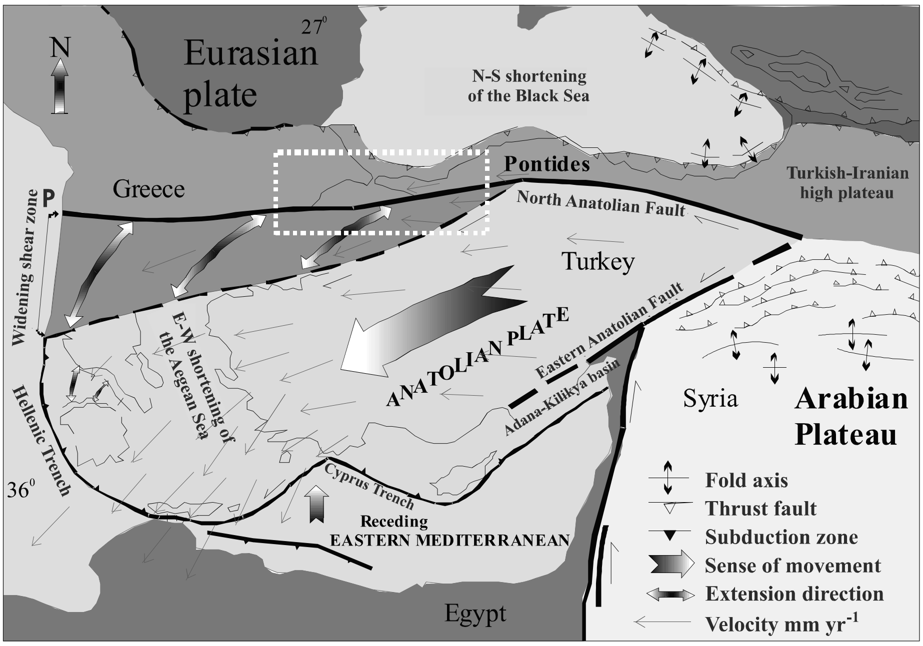


Figure 1. The principal active tectonic features of northwestern Turkey superimposed on the plate tectonic setting of the eastern Mediterranean (after McKenzie 1978; Le Pichon and Angelier 1979). GPS velocity field relative to a Europe-fixed reference as in Straub and Kahle (1994), Reilinger *et al.* (1997), and Kahle *et al.* (1998) is indicated by arrows. The small rectangle with white outline indicates the region of study.

In this orogenic setting, the Black Sea is a post-collisional basin, developed after closure of the northern branch of the Neo-Tethys Ocean (Yılmaz *et al.* 1995). The evolution of the Marmara Sea, Izmit Bay, and Saroz Bay was influenced by the North Anatolian Fault (NAF) (e.g., Barka and Kadinsky-Cade 1988; Koral and Eryılmaz 1995; Le Pichon *et al.* 2001; Armijo *et al.* 2002), and the Aegean Sea formed as a back-arc basin to the north of the Cretean arc (McKenzie 1972; Le Pichon and Angelier 1979, 1981). Western Anatolia is one of the most rapidly extending regions in the world (Figure 1).

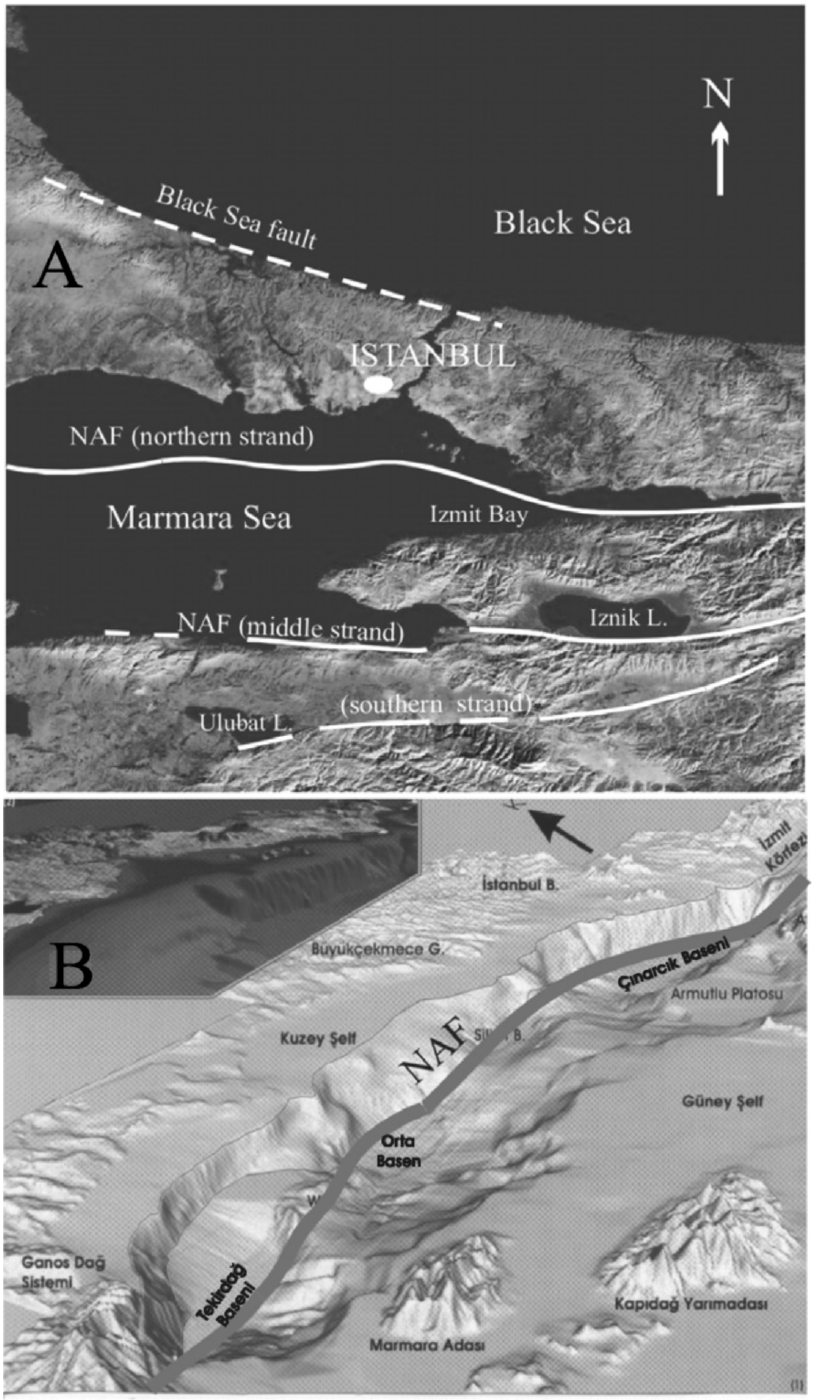
The regional landscape has been continuously and intensively modified since the Tertiary. The most recent of the tectonic activities that have affected coastal geomorphology (e.g., Yılmaz 2003) are the least known due to lack of uncovered signatures. The evidence of very active tectonics implies, however, that prehistoric human life in the Black Sea-Marmara Sea-Aegean Sea region must have been influenced more often than is suggested by archaeological evidence. This paper reviews geological, seismological, and archaeological data that emphasize the role of active tectonics. In this way, it differs from most studies that have primarily focused on the modification of Quaternary coastlines by climatic changes (e.g., Ryan *et al.* 1997; Ryan and Pitman 1998). This approach is necessary in order to distinguish between the effects of active tectonics and climate on the coastal zones of the region collectively called the western Pontics.

2. GEOLOGICAL BACKGROUND

Northwestern Turkey and the western Pontics comprise prominent tectonic features: overthrusts, listric faults, and the NAF (Figures 1 and 2). Overthrusts occur within the mountain ranges; they are related to continental convergence in Eocene-Oligocene times (Yılmaz *et al.* 1995). Listric faults occur along the coastal margins of the mountain range. The seismically active NAF transects the region in an east-west orientation (Figure 2C).

The NAF is the most important geotectonic feature of the region. It is part of the boundary between the Eurasian Plate to the north and the Anatolian Plate to the south, extending from Eastern Anatolia, through the Marmara region, and into the Aegean Sea (Ketin 1969; Şengör 1979; Şengör *et al.* 1985; Barka and Kadinsky-Cade 1988; Barka 1997). Previous studies have described it either as a single shear zone (Şengör *et al.* 1985) or as en-echelon strike-slip faults along which depressions were formed as pull-apart basins (Barka and Kadinsky-Cade 1988; Barka 1992; Koral and Eryılmaz 1995; Armijo *et al.* 2002).

Just east of the Marmara region, the NAF splays into three major strands (Figure 2). The northern strand extends through depressions such as the Bolu and Adapazarı basins, Sapanca Lake, and Izmit Bay, and then traverses the Marmara Sea. The middle strand continues through depressions such as Iznik Lake and Gemlik Bay, forming a transpressional segment with the northern strand, and


N
Black Sea fault
Black Sea
A
ISTANBUL
NAF (northern strand)
Marmara Sea
Izmit Bay
NAF (middle strand)
Iznik L.
(southern strand)
Ulubat L.
B
İstanbul B.
Büyükçekmece G.
Çınarcık Baseni
Armutlu Platosu
Kuzey Şelf
NAF
Güney Şelf
Orta Basen
Ganos Dağ Sistemi
Tekirdağ Baseni
Marmara Adası
Kapıdağ Yarımadası

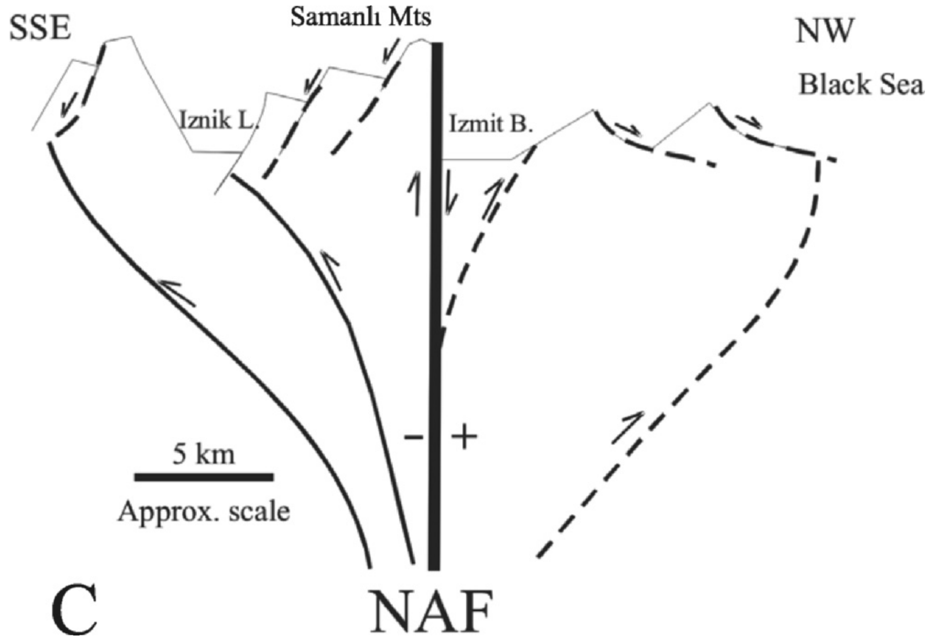


Figure 2. **A.** Morphology and principal tectonic features of the Marmara region; relief map is from NASA's Visible Earth web site. **B.** Bathymetry of the Marmara Sea with its three deep basins from east to west the Çınarcık, central, and Tekirdağ basins. Bathymetry map is after Gökaşan *et al.* (2001). Each basin is separated from each other by ridges. **C.** Schematic section showing active tectonic structures of the region (+ sign indicates the approaching fault block).

frames the southern shelf of the Marmara Sea with a southward bend. The southern strand is morphologically less obvious and appears as a splay of the middle strand (Crampin and Evans 1986). It extends through depressions such as the Yenişehir basin, the Uluabat Lake, and the Kuş (Apolyont) Lake.

The NAF caused a series of modifications in the region, though there is no general agreement on their age and displacement (e.g., Şengör 1979; Şengör *et al.* 1985; Barka and Kadinsky-Cade 1988; Barka 1992; Armijo *et al.* 1999; Bozkurt 2001; Hubert-Ferrari *et al.* 2002). Several suggestions have been offered as to its age: (1) that initiation began in the late Middle Miocene (Şengör 1979, 1980), (2) that it is earliest Pliocene, about ~5 my BP (Barka and Kadinsky-Cade 1988), or (3) that it is even as young as 1.8 my BP. Estimates of total offset along the NAF in the west range from 85 km (Armijo *et al.* 1999) to the commonly accepted 20–25 km (Seymen 1975; Şengör 1979; Şengör *et al.* 1985; Barka and Gülen 1988; Arpat and Şenturk 2000; Barka 1992; Herece and Akay 2003). These offsets indicate a 5–10 mm (Barka 1992) or 17±2 mm (Hubert-Ferrari *et al.* 2002) yearly rate of motion.

Different tectonic models have been proposed on the nature and length of the faults along the NAF in Izmit Bay and the Marmara Sea (e.g., Şengör 1979; Barka and Kadinsky-Cade 1988; Suzanne *et al.* 1990; Barka 1992; Ergün and Özel 1995; Koral and Öncel 1995; Wong *et al.* 1995; Okay *et al.* 1999, 2000; Siyako and Şaroğlu 2000; Yaltırak *et al.* 2000; Gökaşan *et al.* 2001; Le Pichon *et al.* 2001; Armijo *et al.* 2002). Recent marine seismic reflection profiles obtained, for example, by Smith *et al.* (1995), Okay *et al.* (1999, 2000), and Le Pichon *et al.* (2001) reveal that the Marmara Sea comprises several deep marine asymmetric strike-slip basins (i.e., the Çınarcık, Central Marmara, and Tekirdağ basins) with Plio-Quaternary fill reaching to thicknesses of over 3 km, separated by northeast-trending submarine ridges that rise several hundred meters above the sea floor (Figure 2).

Two distinct, steep, continuous bathymetric escarpments bound the Marmara Sea on the north and south, marking the location of major active faults (Okay *et al.* 2000). These faults enter the Marmara Sea through Izmit Bay in the east and emerge in Thrace in the west (Tüysüz *et al.* 1998; Yaltırak *et al.* 1998; Armijo *et al.* 1999), bounding the northern margin of Gallipoli Peninsula (Dardanelles), and extending into the Aegean Sea (Figure 1). The westerly extension of the fault zone into the Aegean Sea bounds the Northern Aegean Trough (Lyberis 1984; Roussos and Triantafyllos 1991; Taymaz *et al.* 1991; Barka 1992; Saatçılar *et al.* 1999).

These tectonic features occur in a geological setting of sedimentary units ranging from Paleozoic to the present, including metamorphics/metaophiolites. Upper Cretaceous to Miocene marine sediments occur unconformably over these basement rocks, and they are in turn overlain by Ponsien-Pliocene sediments containing weakly cemented clastics representing a fluvial and lacustrine depositional environment (Akartuna 1968; Bargu and Sakınç 1989; Koral and Şen 1994; Sakınç *et al.* 1999).

The Tertiary sedimentary strata of the region with syn-tectonic features indicate continuous tectonic uplift in northwestern Turkey since the Late Eocene-Oligocene (Onal 1984; Siyako *et al.* 1989; Sen *et al.* 1996; Görür *et al.* 1997; Turgut and Eseller 2000). Uplift was synchronous with the growth of the Pontides (Yılmaz *et al.* 1995) and the Carpathians (Muratov *et al.* 1978; Ross 1978; Kojumdgieva 1983), and it was accompanied by widespread deposition of fluvial strata in the Black Sea basin through the Late Miocene to Pliocene during which brackish seas formed (Robinson *et al.* 1996; Nikishin *et al.* 2003).

During the late Quaternary, the Black Sea was repeatedly isolated from the Marmara and Mediterranean Seas due to varying climatic conditions (Ross *et al.* 1970; Degens and Ross 1972; Deuser 1972; Ross and Degens 1974; Stanley and Blanpied 1980; Yanko 1990; Tchepalyga 1995). For instance, the Black Sea became a giant freshwater lake during the last glacial maximum (18,000 BP), with the water level at about –100 m (Ryan *et al.* 1997; Ryan and Pitman 1998). During the post-glacial sea-level rise at about 7.15 ky BP, Mediterranean water reached and refilled the Black Sea through the Istanbul strait. This refilling was proposed as an abrupt, catastrophic inundation, giving rise to a new interpretation of the Biblical/mythological Noah's Flood (Ryan *et al.* 1997; Ryan and Pitman 1998). A number of other papers have been published in favor of such a Black Sea flood hypothesis (Demirbağ *et al.*1999; Ballard *et al.* 2000; Winguth *et al.* 2000), yet others that are critical due to contradictory evidence have also appeared (Aksu *et al.*1999; Ça ğatay *et al.*2000; Aksu *et al.* 2002a, b; Kerey *et al.* 2004; Yanko-Hombach 2004).

3. EVIDENCE FOR AN ACTIVE TECTONIC SETTING

Present and historical seismic activity provides ample evidence for tectonic activity in the region (Figure 3). Seismic activity ranges from micro-seismicity to major events, and it occurs largely along the NAF and its splays. Seismic events are prominently strike slip, though there are normal and reverse-slip events as well (e.g., Alptekin *et al.* 1986).

Major seismic events are almost always accompanied by prominent fault rupture, which produces subsidence, uplift, and lateral displacement of landforms (Barka 1996; Koral *et al.* 2001b) (Figure 4). For instance, the August 1999 earthquake in the eastern Marmara caused, along its 100 km long rupture path, right lateral displacement up to 5 m in places such as Arifiye and Akyazı (Barka 1999; Herece 1999; Koçyiğit *et al.*1999; NEIC 1999; Toksöz *et al.*1999; Akyüz *et al.* 2000; Emre *et al.*2000) (Figure 4A). These rupture segments displayed a dip-slip component in wide Quaternary alluvium and formed sag ponds. When they occurred along shorelines of marine or lacustrine environments, as in the coastline of Gölcük, Değirmendere (Izmit Bay), and Sapanca Lake (Figure 4B, C, D), coastal modifications developed (Öztürk *et al.* 2000) (Figure 5). The rupture transected the bank of the Sakarya River, where a sharp clockwise turn exists in its course, overlapping the surface rupture of the previous 1967 Mudurnu earthquake (Ms=7.1) in the Adapazarı depression (Akyazı). The rupture was either localized or occurred in a broad zone (up to several tens of meters). In places, it splayed into branches and showed discontinuities.

Moreover, major destructive earthquakes have repeatedly modified the region throughout the historical past (Ambraseys and Finkel 1987, 1991; Eyidoğan *et al.*1991; Ambraseys and Finkel 1995). For example, the August and November 1999 events changed the morphology near the epicenters of the previous earthquakes of 967, 1719 (Io=V), 1754, and 1878. Major earthquakes of 1343, 1509, 1766a, 1766b and 1894 (Io=VIII) affected the Marmara Sea and its vicinity (Ambraseys and Finkel 1995), and the earthquakes of 69, 121, 269, 362, 478, 554 and 740 impacted the historical town of Izmit (Nicomedia) and

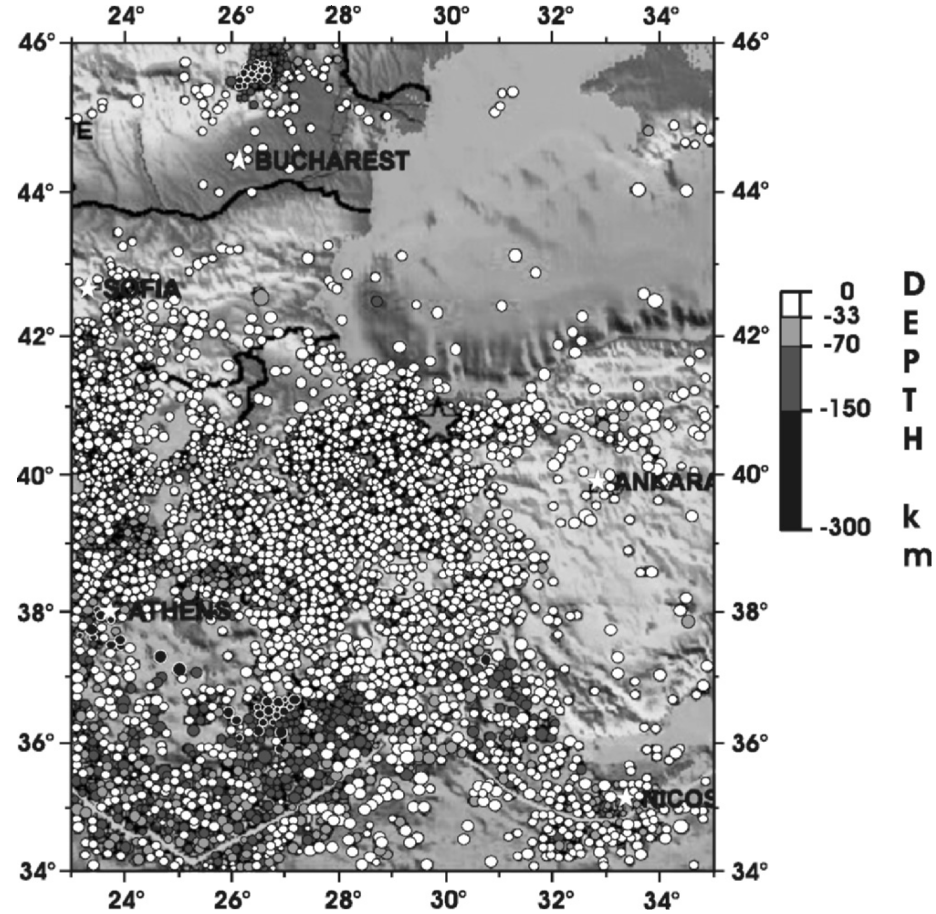
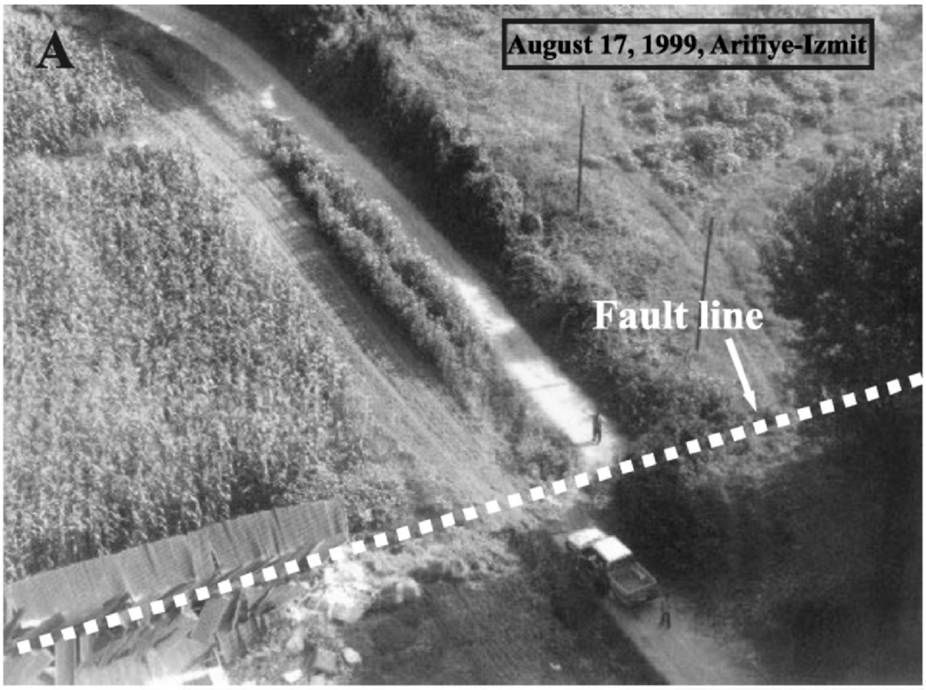


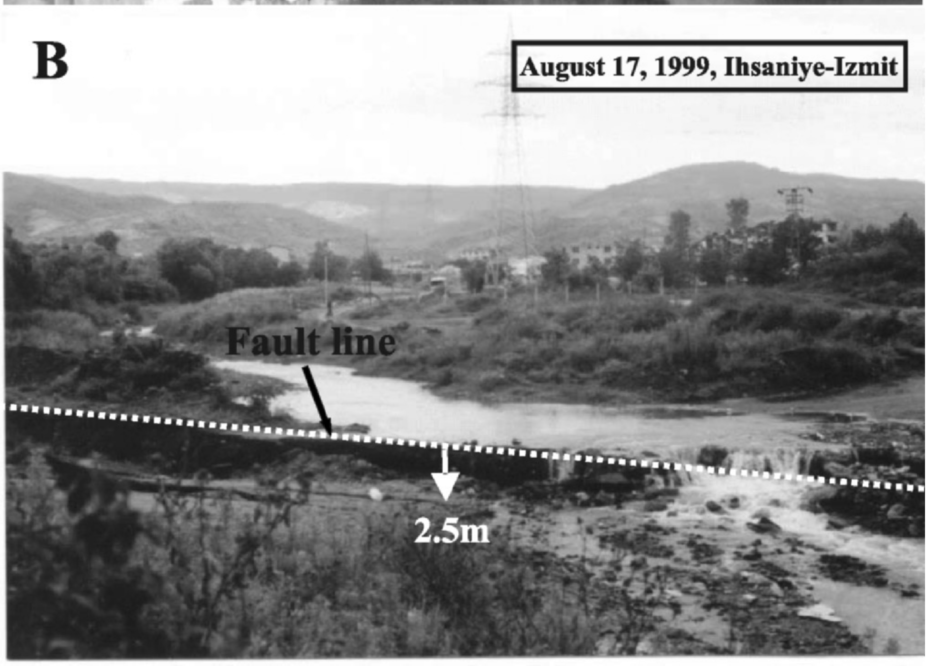
Figure 3. Seismicity of western Turkey between 1990–1999 (NEIC 1999). This map shows that northwestern Turkey is one of the most seismically active areas in the eastern Mediterranean. Seismicity along the NAF is characterized by fewer, but larger, events. Zigzag line at bottom represents plate boundary.

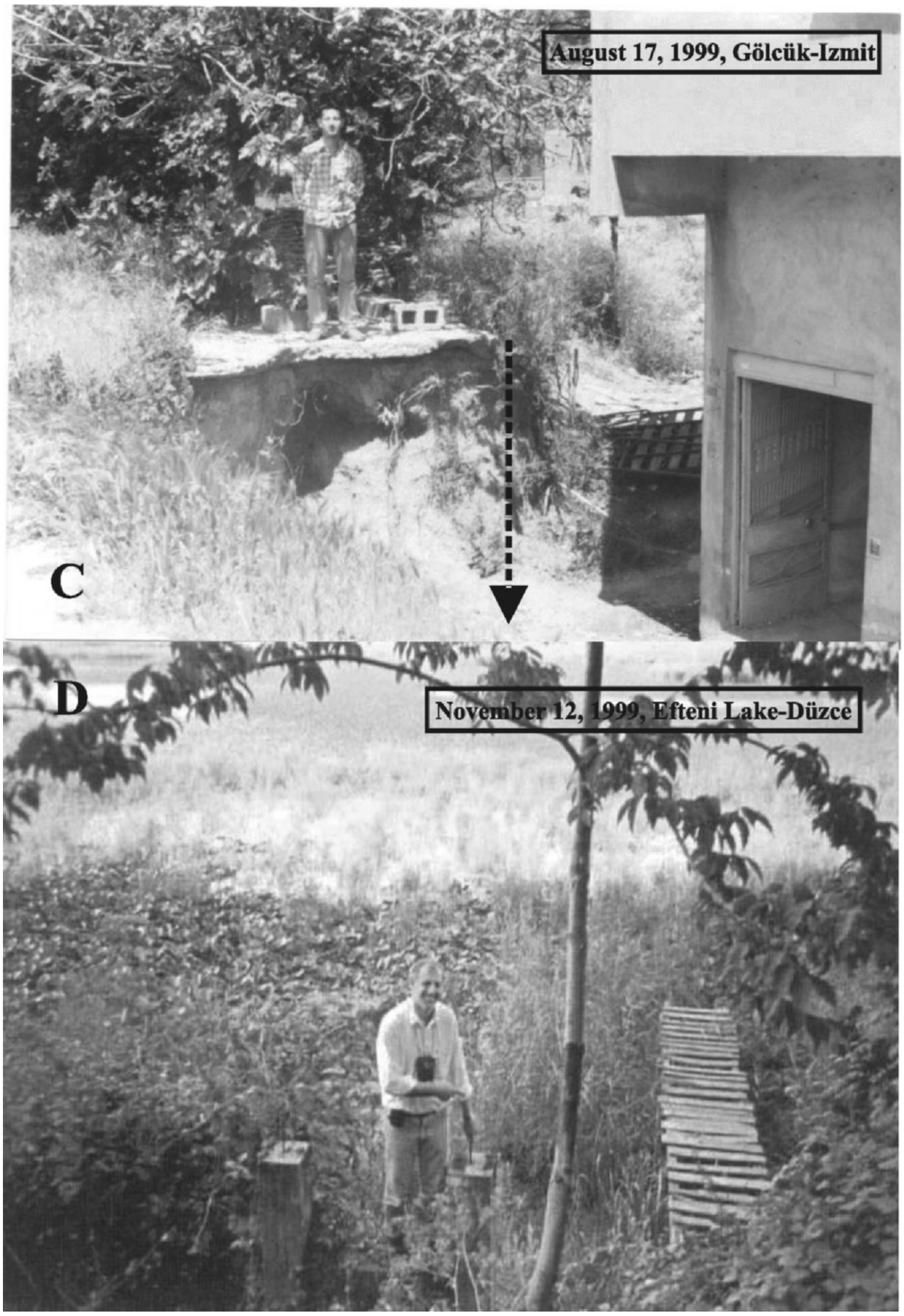
Lake Iznik (Nicea) that lie on the middle strand of the NAF. Many other destructive earthquakes were identified by trench studies in Saroz and Izmit bays (Rockwell *et al.* 1998).

GPS readings from networks in the Marmara Sea, Aegean Sea, and Anatolia reveal further evidence of tectonic movements. The network in the

Figure 4. Facing and following pages: **A.** Surface rupture resulting from the August 17, 1999 earthquake. Lateral displacements up to 5 m oriented east-west were observed along the 120 km long affected area (from KOERI website). **B.** and **C.** Vertical displacements due to faulting or seismically induced gravitational movements were on the order of a few meters in Gölcük. **D.** Coastal modifications in Efteni Lake following the Nov. 12, 1999 Düzce earthquake. The stairs leading to the lake were displaced laterally and vertically more than 2 m.


A
August 17, 1999, Arifiye-Izmit
Fault line

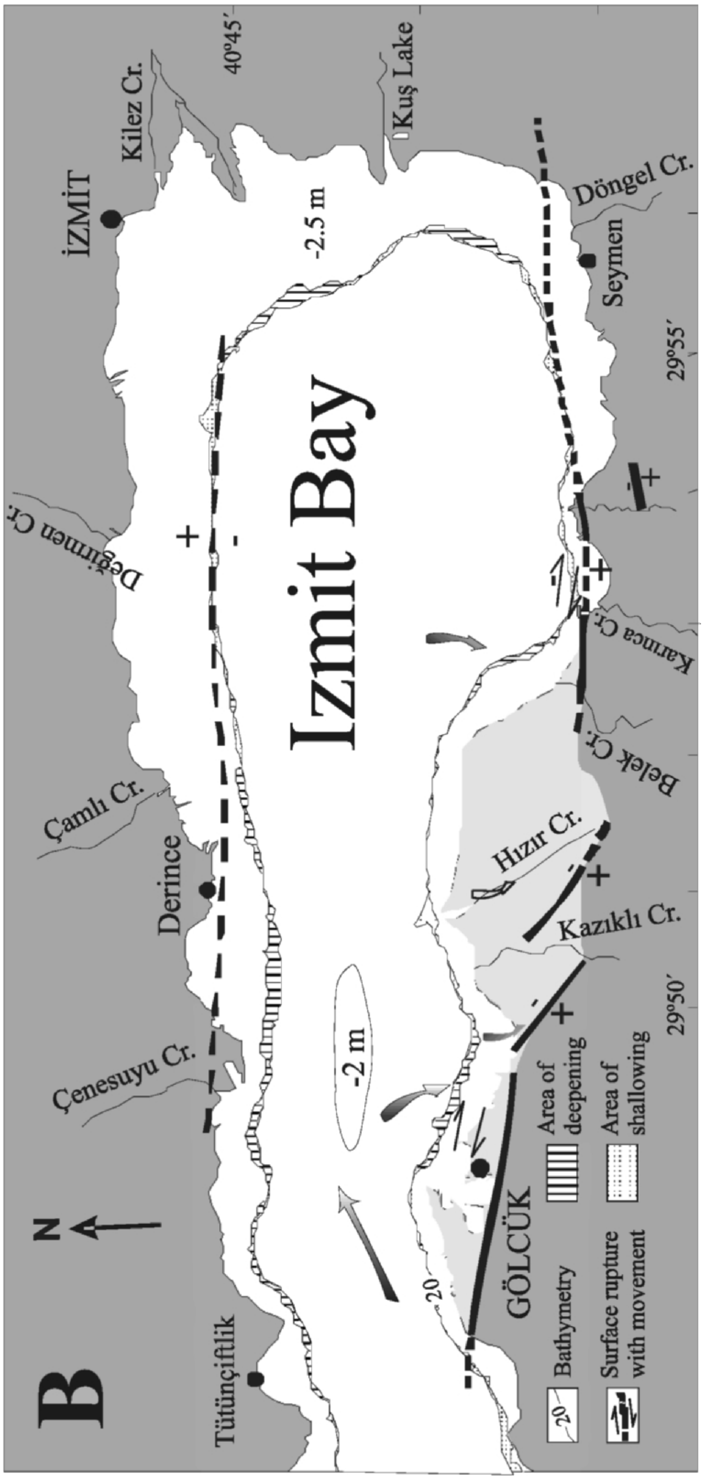

B
August 17, 1999, Ihsaniye-Izmit
Fault line
2.5m

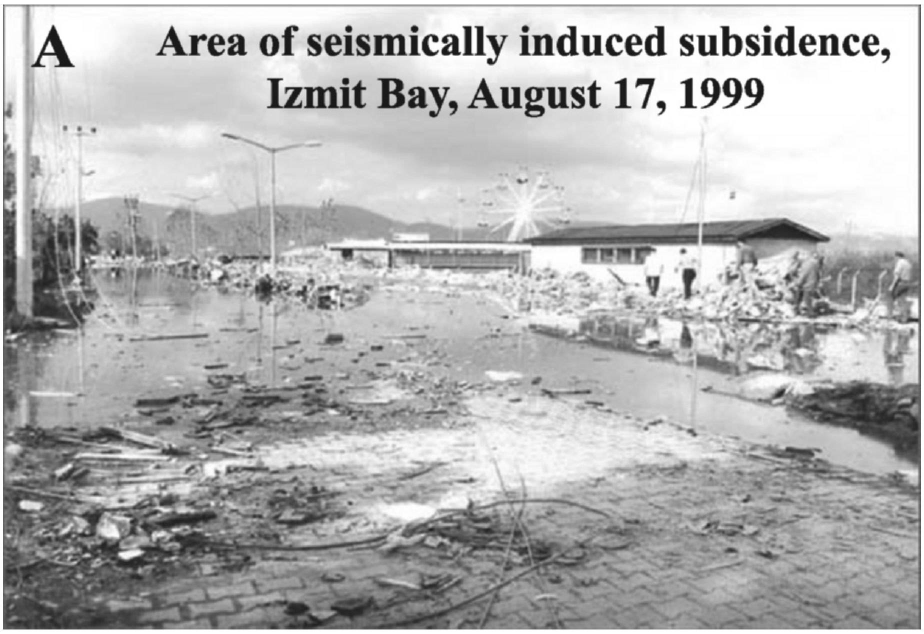

August 17, 1999, Gölcük-Izmit
C
D
November 12, 1999, Efteni Lake-Düzce

Marmara region, which surveys the northwestern part of the boundary between the Eurasian and Anatolian plates, indicates an average rate of 22 mm yr^{-1} for the westward movement of the Anatolian Plate along the NAF relative to the Eurasian Plate (Straub and Kahle 1994; Oral *et al.* 1995; Le Pichon *et al.* 1995; Straub and Kahle 1995; Reilenger *et al.* 1997; Straub *et al.* 1997; Kahle *et al.*1998; McClusky *et al.* 2000). Stations in the Caucasus Mountains indicate a relative shortening on the order of 10 mm yr^{-1} across the Caucasus (Kahle *et al.* 1998). The network in the southern part of the Aegean indicates essentially a homogeneous, southwest-oriented motion. In Greece, rates reach up to 35 mm yr^{-1} relative to the Eurasian Plate. The motion of Anatolia is considered to be a coherent, counterclockwise rotation about a pole located near the northern Sinai Peninsula (Rotstein 1984; Reilinger *et al.* 1997; Kahle *et al.* 1998).

Tectonic movements are also reflected in the drainage system and morphotectonic features of northwestern Turkey (Figure 6). Major rivers of the region indicate sharp diversions in their courses as they traverse a splay of the NAF. The Sakarya, Simav, Mustafa Kemalpaşa, and Gönen rivers are examples, and the Sakarya River, for instance, exhibits a 22 km dextral diversion in the Adapazarı basin (Figure 6B).

The principal morphotectonic units in northwestern Anatolia (the Kocaeli Peneplain, the Samanlıdağ and Bilecik highs, and the NAF depressions) contain tilted Pliocene erosional surfaces and sedimentary units with two different drainage systems and hanging valleys, all indicative of recent tectonic activity. The Kocaeli Peneplain, with an average elevation of 150–250 m, is tilted to the north with its water division line lying proximal to the NAF. There, north-trending streams drain into the Marmara Sea via hanging valleys (Emre *et al.* 1998). The Samanlıdağ and the Bilecik highs, with an average elevation of 700– 1000 m, are transected by the northern, middle, and southern strands of the NAF. They are overlain by late Miocene-Pliocene fluvial and lacustrine sedimentary units, which were deformed and elevated to the height of mountain ranges (Akartuna 1968; Genç 1986; Bargu and Sakınç 1989; Erendil *et al.* 1991). The Samanlıdağ plateau exhibits a drainage system of wide-based, low-energy, north-trending streams. Near the NAF, this drainage system connected to another one consisting of short streams with wild morphology via hanging valleys (Emre *et al.* 1998). NAF-related depressions, such as the Adapazarı basin, Sapanca Lake, Izmit Bay, and the Marmara Sea, that occur along the northern strand of the NAF contain late Pliocene-Holocene sediments and form a boundary between the Kocaeli Peneplain and the Samanlıdağ unit (Emre *et al.* 1998) (Figure 6). In the Adapazarı depression, the late Pliocene-Pleistocene age sedimentary unit inclines 20–30º to the south. In the Sapanca Lake and Izmit Bay depressions, Pleistocene deposits occur and exhibit lateral displacements up to 10 km (Emre *et al.* 1998). The Geyve and Pamukova basins, Iznik Lake, and the Gemlik and Bandırma Bays along the middle strand of the NAF, and the Yenişehir basin and Uluabat (Apolyont) and Kuş (Manyas) Lakes along the


B
N
Tütünçiftlik
Çenesuyu Cr.
Derince
Çamlı Cr.
Değirmen Cr.
İZMİT
Kilez Cr.
40°45′
Kuş Lake
-2.5 m
Izmit Bay
-2 m
Döngel Cr.
Seymen
29°55′
Karınca Cr.
Belek Cr.
Hızır Cr.
Kazıklı Cr.
29°50′
GÖLCÜK
20
Bathymetry
Surface rupture with movement
Area of deepening
Area of shallowing



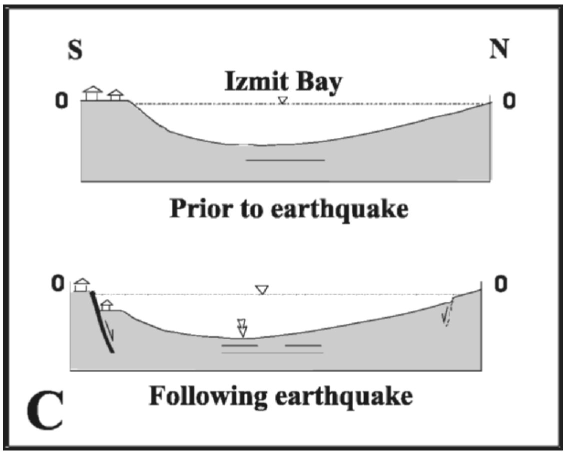


Figure 5.
A. Coastal modifications along Izmit Bay after the earthquake of August 17, 1999.
B. (facing page) Bathymetric changes occurred both in Izmit Bay and Sapanca Lake (after Özturk *et al.* 2000). Light colored area indicates the location where coastal modification occurred.
C. Suggested mechanism for coastal changes.

southern strand of the NAF are underlain by Upper Miocene and Pliocene fluvial, then lacustrine, sediments (Genç 1986). In the Pamukova basin, there are Pleistocene sediments. All these depressions are overlain by Quaternary alluvium, indicating a recent formation age.

Uplifted beach-dunes, wave-cut notches and Pleistocene-Holocene terraces also indicate active tectonics in the Pontic region. Uplifted terraces occur at a height of 30–170 m on the southern coasts of the Black and Marmara Seas (Andrusov 1900; Erentöz 1953; Erinç and İnandık 1955; Erinç 1956; Chaput 1957; Göney 1964; Akartuna 1968; Ketin and Abdüsselamoğlu 1970; Erol and Nuttal 1973; Erol and Inal 1980; Bargu and Sakınç 1989); Ardos 1992;

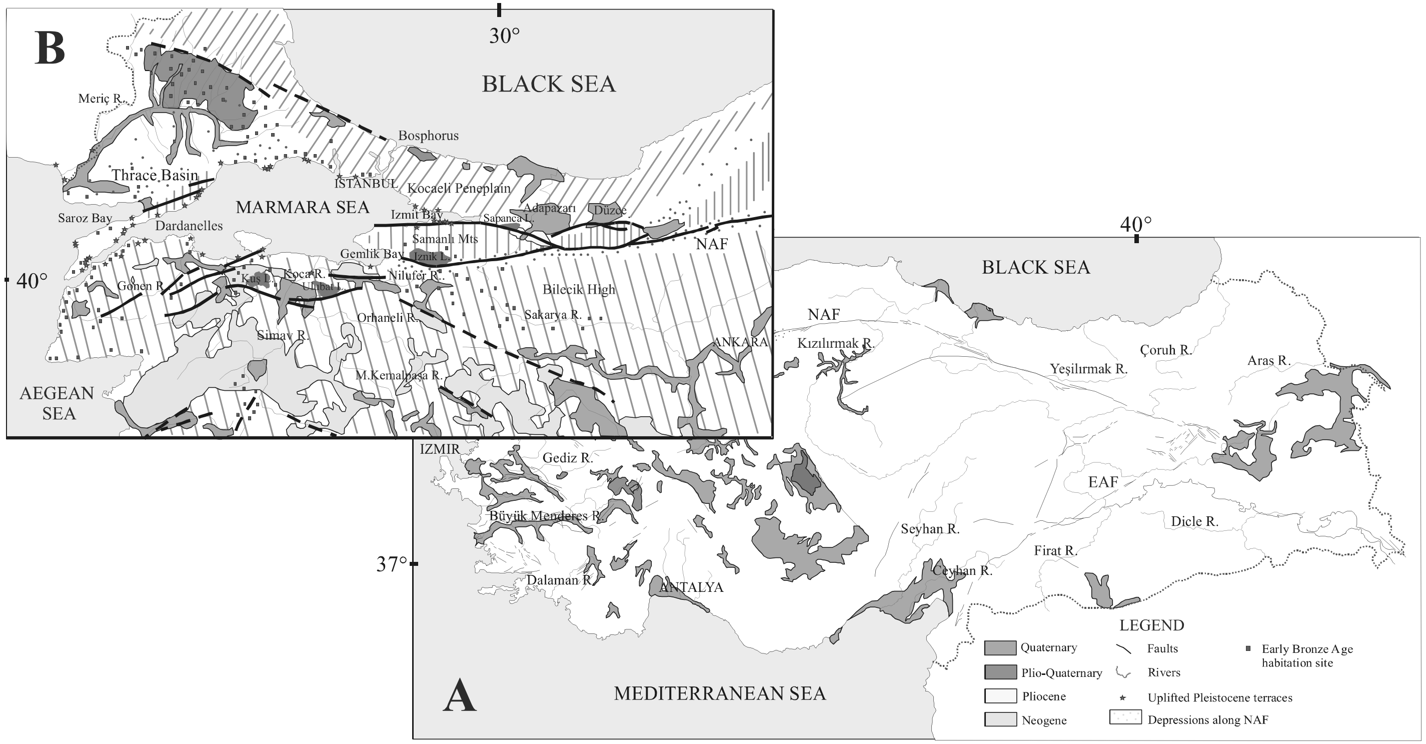


Figure 6. **A.** Quaternary outcrops seen in relation to active tectonic features of northwestern Turkey (after MTA map and Şaroğlu *et al.* 1992). **B.** Pliocene-Quaternary outcrops and active faults of northwestern Turkey. Major river systems reflect the influence of active tectonics. The Sakarya, Kızılırmak, Simav, and Mustafakemalpaşa Rivers exhibit the prominent role played by active tectonics on geomorphology. Uplifted Pleistocene outcrops are indicated by stars, morphotectonic units by cross-hatching and dots, and Early Bronze Age sites by small squares (after Özdoğan 2003).

Bargu 1993; Barka and Sütçü 1993; Erol and Çetin 1995) (Figures 6B and 7). A well-developed beachrock occurs along the northern shores of the Çanakkale Strait (the Dardanelles) at İyisu, situated at an elevation of 16.6 m. U/Th measurements on an *Ostrea edulis* shell extracted from this beachrock revealed an age of 205 ky BP, which yields an average uplift rate of ~0.26 mm yr^{-1} (Yaltırak *et al.*2002). Uplifted marine terraces of Pleistocene age occur at three different morphologic levels in the vicinity of Izmit Bay (Bargu and Sakınç 1989). The oldest (yielding an age of 260 ky BP) is situated southwest of Karamürsel on Upper Miocene sediments at levels 60–90 m above sea level (Paluska *et al.* 1989). It contains a brackish fauna indicated by *Mytilus galloprovincialis* Lamarck, *Venerupis calverti* Newton, *Ostrea lamellosa* Brocchi, *Chlamys opercularis* Linné, *Cerastoderma edule* Linné, suggesting a minimum uplift rate of 0.50 mm yr^{-1}. The other terrace occurs east of Karamürsel at 20–25 m on Eocene sediments and contains *Ostrea* shells representing Mediterranean fauna that yield U/Th ages of 130 ky BP (Paluska *et al.* 1989; Sakınç and Bargu 1989). This terrace suggests a minimum uplift rate of 0.25 mm yr^{-1}. There are no marine terraces along Gemlik Bay. Conversely, while there are no terraces along Sapanca Lake, there are some along Iznik Lake (Ardel 1956; Akartuna 1968; Bilgin 1967; Ikeda *et al.* 1991).

Active fault scarps, displaced underwater channels, and recent fault ruptures indicate the role of active tectonics in the Marmara Sea (Figure 2). Seismic stratigraphic sections from this marine environment include many faults and onlaps, demonstrating tectonic activity and uplift. For instance, the Tekirdağ basin–an active, rhomb-shaped depression developed on a negative flower structure along the NAF–has a basin floor at a water depth of –1150 m. The basin lies along a releasing bend of the strike-slip fault and is bounded on the northern side by the NAF and on the southern side by a steep submarine slope defined by a normal fault (Okay *et al.*1999). The basin is strongly asymmetrical in cross-section, with the thickness of the syn-transform Pliocene-Quaternary strata increasing from a few tens of meters on the submarine slope to over 2.5 km adjacent to the North Anatolian Fault (Okay *et al.*1999). The Tekirda ğ is flanked by a constraining bend, along which the syn-transform strata are being underthrust. This thrusting seems to be responsible for the uplift of the submarine slope to a height of 924 m. (Okay *et al.* 1999).

4. DISCUSSION: CAUSES OF COASTAL TRANSFORMATIONS

Northwestern Turkey has undergone extensive coastal and ecological transformations since the early Mesozoic Era, ~200 my BP. Throughout the Mesozoic, it lay on the southern margin of the Neo-Tethys Ocean that linked the

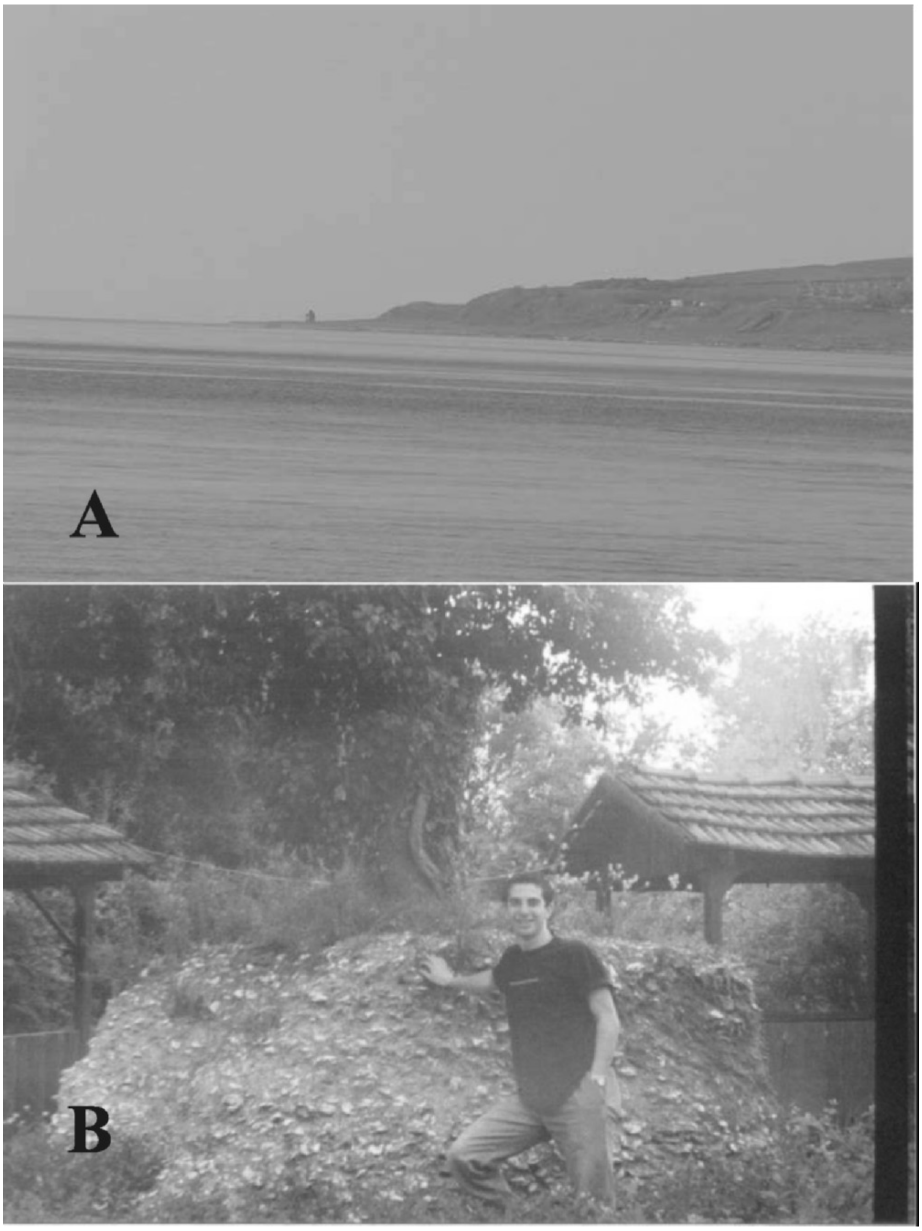
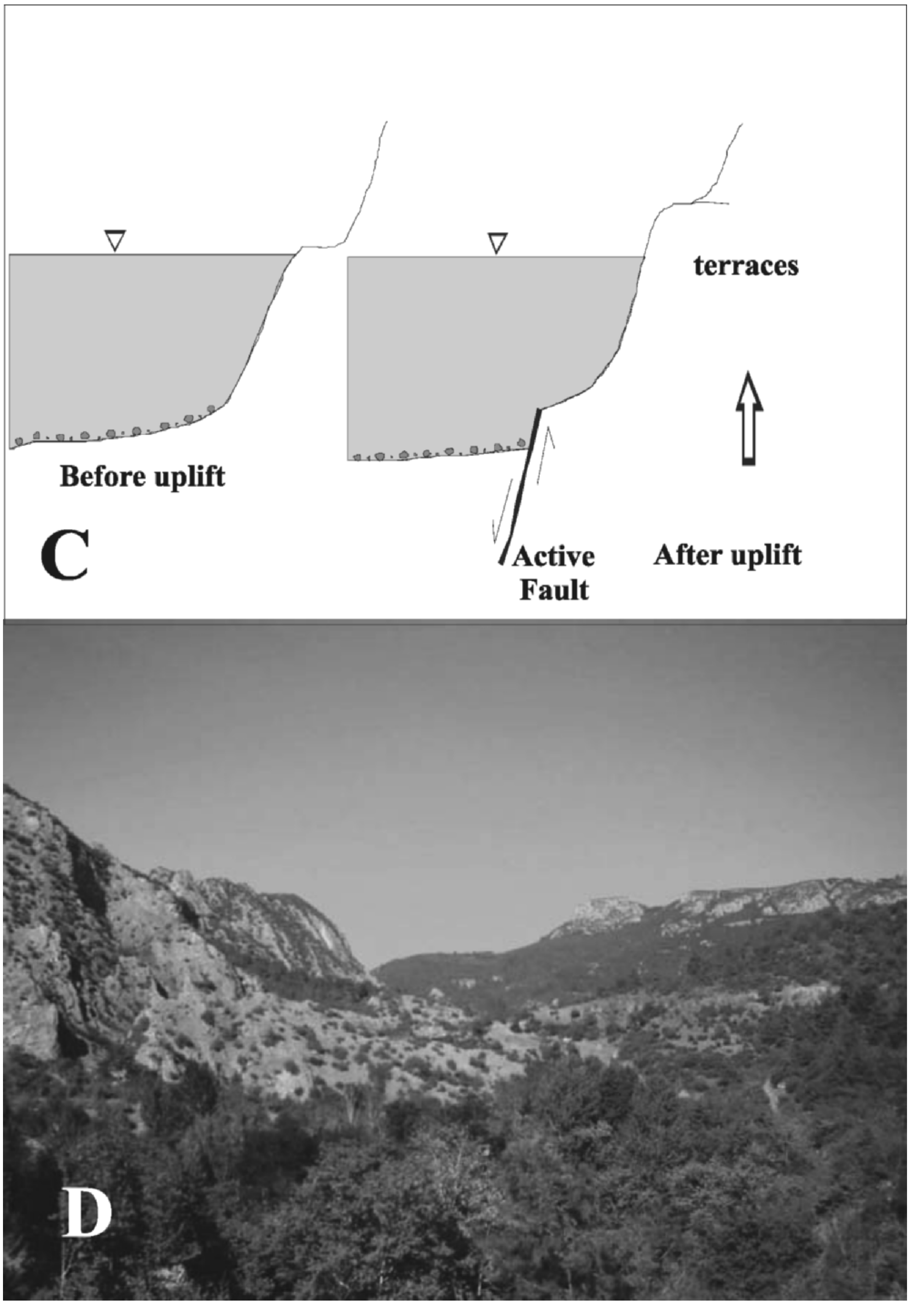


Figure 7. **A.** Example of uplifted coastal notches near Büyük Çekmece along the northern Marmara coast. **B.** Uplifted Pleistocene terrace along the coast of Izmit Bay. **C.** Suggested mechanism of uplift for terraces and coastal notches. **D.** A deeply-cut valley indicating rapid uplift east of Yenişehir, eastern Marmara region.



basins of the Atlantic and Pacific Oceans (e.g., Okay *et al.* 1994; Yılmaz *et al.* 1995). It then existed within the semi-isolated Ponto-Caspian basins as the African continent collided with Eurasia, and the mountain chains of the

Carpathians, the Pontides, and the Caucasus gradually rose in the mid-Tertiary. During the Upper Miocene-Pliocene, it existed as the Euxinian basin (Black Sea) (Reid and Orlova 2002). It then became the Chauda Lake Sea after the late Mindel (Elsterian) Glaciation between 500,000–400,000 BP (Yanko and Troitskaya 1987; Tchepalyga 1995). During the Riss-Würm Interglacial (150,000–100,000 BP), the Euxinian depression became connected to the Mediterranean basin, but with the Würm Glaciation, the connection with the outside ocean was again broken. A connection to the Mediterranean Sea was established once more through the Bosphorus and the Dardanelles about 9400 BP (Yanko-Hombach 2004) or 7500 BP (Ryan *et al.* 1997).

These extensive geographic transformations in the western Pontics were attributed primarily to oceanographic conditions. In an orogenic zone, however, one would expect the paleogeographical evolution to have been heavily influenced by tectonic movements. The uplift evident in the regressive sedimentary sequence of the Pontic region is characterized by a transition from shallow marine conditions (Middle Miocene) to lacustine and fluvial conditions (Upper Miocene to Pliocene) (Siyako *et al.* 1989; Sen *et al.* 1996; Turgut and Esseller 2000), and this is further implied by the unconformities/discontinuities that have been recognized in seismic profiles as well as Pliocene and Pleistocene age deposits (Çetin *et al.* 1995; Meriç *et al.* 1995; Özhan and Bayrak 1998). Sustained tectonic uplift since the late Pleistocene is apparent in elevated Pleistocene terraces, shorelines, and morphological features such as wave-cut notches and V-shape valleys (Figure 7). Uplift has thus been a prominent component of the tectonic activity in northwestern Turkey.

How much uplift has occurred in the region since the Pleistocene is problematic. Tectonic uplift rates obtained from raised terraces in the western Marmara region are quite variable. Some rates were derived under the assumption that the terrace was deposited at or just below the coeval global sea level. For instance, deposition of the İyisu section (Çanakkale, the Dardanelles) occurred about 200 ky BP at a sea level that was –36 m below that of the present (Chappell and Shackleton 1986), which suggests a 53 m post-depositional uplift of the beachrock at an average rate of ~0.26 mm yr^{-1} (Yaltırak *et al.* 2002). Similar calculations for the dated successions along the Dardanelles reveal uplift rates ranging from ~0.27 to ~0.73 mm yr^{-1} (Yaltırak *et al.* 2002). The variations are interpreted as differential uplift rates among individual fault blocks within the North Anatolian Fault beneath the Marmara Sea floor (Aksu *et al.* 2000). These uplift rates exceed those observed in adjacent regions of western Turkey. Other rates obtained from the staircase of four high terraces of cemented fluvial gravel at ~360, ~330, ~255, and ~225 m above river level within the Büyük Menderes River drainage catchment are interpreted as having developed largely from the Middle Pleistocene onwards. A terrace ~190 m above river level with a lower-level staircase capped by basalt K/Ar dated to ~1.2 my BP has been interpreted as indicating uplift rates of ~0.1 mm yr^{-1} or more during the latest Pliocene but increasing to near ~0.2 mm yr^{-1} in the Middle and Late Pleistocene for reasons not directly related to active normal faulting. Here, local isostatic consequences of the rate change were superimposed over the regional surface uplift (Westaway *et al.* 2003).

Repeated transitions from high-energy to low-energy environments are apparent in the alternating gravel-silt-clay layers and unconformities of the late Quaternary sedimentary sequence in Izmit Bay (Çetin *et al.* 1995; Ediger and Ergin 1995; Özhan and Bayrak 1998). These indications may represent episodes of rapid tectonic uplifting that occurred during the region's recent geological history. This suggests that uplift history may be more complicated than has been envisaged, and that the reported uplift rates could be very conservative estimates due to exclusion of localized subsidence, such as occurred during the August 17, 1999 earthquake (Figure 5) (Rebaï *et al.* 1993; Öztürk *et al.* 2000). When geomorphological-bathymetrical features of the region are considered, uplift rates appear on the order of, minimally, 1 mm yr^{-1}. This amount is much less than the Holocene uplift that has been suggested for a similar setting within Iskenderun Bay, in the northeastern corner of the Mediterranean (Koral *et al.* 2001a). It is, however, partially consistent with that of a 600 m raised snow-border of the last glacial period in the eastern Black Sea mountains (Erinç 1984) as well as GPS readings (1–10 mm) along the Black Sea (Kahle *et al.* 1998; Tari *et al.* 2000).

Even considering the suggested rates of uplift, the Marmara region must have experienced significant differences from reported paleogeographic reconstruction models during the Quaternary glacial periods. It has been assumed that the Marmara Sea—which forms an oceanographic gateway between the world's largest permanently anoxic basin, the Black Sea, and the Mediterranean Sea—was isolated from the Aegean Sea during the glacial periods, becoming a lake. It has also been assumed that it reconnected through the Strait of Çanakkale (the Dardanelles) during interglacial highstands, which allowed Mediterranean water to penetrate northward into the Marmara Sea and, eventually, the Black Sea (e.g. Aksu *et al.* 1999). When considered together with detailed topographic and bathymetric maps of the western Marmara region, the average rate of tectonic uplift reveals that a degree of water exchange existed between the Aegean Sea and the Marmara Sea even during the peak of the Middle Pleistocene glacial period (e.g., when global sea level was situated at approximately –100/ –80 m). This is supported by the dominance of *Mytilus edulis* as the most common brackish-bivalve in the present-day Marmara raised coastal terraces (Yaltırak *et al.* 2002).

The reported uplift rates are very small, however, compared to the westerly lateral displacement of the Anatolian Plate along the NAF, which GPS measurements and geological offsets indicate moves an average of 20 m every thousand years. The NAF splits into three branches in the eastern Marmara region to accommodate this displacement relative to the stable Eurasian Plate.

The rate is estimated at 16–25 mm yr^{-1} based on about 25 km offsets of geological features and late Miocene outcrops in the eastern Marmara (Barka 1992; Hubert-Ferrari *et al.* 2002), or an average of 22 mm yr^{-1} based on GPS readings (Straub and Kahle 1995; Kahle *et al.* 1998). Such rates indicate lateral morphological change on the order of 15,000–20,000 m over 1 my and 30,000–40,000 m over 2 my, which is evident from the diverted river courses in the region (Figure 6). An inevitable geological consequence of the high rate of lateral displacement along the NAF is the very young age, perhaps Pleistocene, of many morpho-tectonic features, such as Izmit and Saroz Bays, the Bosphorus, and perhaps the Samanlı and Gaziköy mountains.

The spatial distribution of sedimentary units in the geological map of the region indicates that the NAF has been active from as early as the Upper Miocene (Şengör *et al.* 1985) or as recent as the Pliocene (~5 my BP). There is, however, no sound evidence that the fault system was the same then as it is today. It has been suggested that in the Upper Miocene-Pliocene, northwest-oriented faults existed in the Thrace region (Perinçek 1987, 1991; Koral and Şen 1995; Koral 1998). These faults had a very different geometry from that of the current NAF and brought about the formation of rectilinear basins in the region at that time (Koral and Şen 1995; Sen *et al.* 1996; Koral 1998). The post-Pliocene uplift (and denudation), caused in part by rotating fault blocks (Rotstein 1984; Westaway 1990), coincided with a new generation of east-west and northeast-southwest oriented, tectonically-controlled depressions along the NAF, creating lakes, marshes, and fluvial systems of Plio-Quaternary and Pleistocene age, as is indicated by Pleistocene and Holocene outcrops (Figure 6B). Post-Pleistocene and Holocene uplift and denudation, accompanied by seismic movements on the NAF, caused the filling of these basins to form wide alluvial plains suitable for human settlement.

Geological data on late Quaternary tectonics is limited. Archaeological evidence, however, is totally absent for the interval between 5000 and 3000 BP in this tectonically-active environment. Pre-Bronze Age habitation (younger than 5000 BP) is totally absent within the territory between Iznik Lake and the Black Sea, including the rich alluvial plains around Izmit Bay, Sapanca Lake, and the Sakarya River (Figure 6). It is hard to conceive of a cultural discontinuity for the entire span of the Bronze Age (5–3 ky BP) in the eastern Marmara region, as the area constituted a cultural bridge between Anatolia and the Balkans throughout the Neolithic and Chalcolithic periods (10–5 ky BP). An extensive maritime trade network was established all around the Aegean to the Çanakkale strait, and well-documented Early Bronze Age sites, such as Troia, Demirci Höyük, Beşiktepe, Kumtepe, Hacılartepe, and Orhangazi, are present on the coastal strip from the entrance of the Dardanelles to Iznik Bay (Özdoğan 2003). Furthermore, although there are several large Bronze Age mounds around Iznik Lake, no sites occur around Sapanca Lake despite its very similar setting and distance of only

50 km. This is strange, but it is consistent with the absence of terraces around Sapanca Lake.

This archaeological abnormality coincides geographically with the last high sea-level stand of the Riss-Würm interglacial period. It also coincides with the tectonically very active region of the NAF and its transpressional segments. Multidisiplinary research is needed to discover the causes of these coincidences, however, it is easy to envisage lateral displacement on the order of 2 km and vertical displacement of more than 100 m over the course of 100,000 years. These values would no doubt have affected not only the nature of water connections through the straits (Koral 1993; Oktay *et al.* 2002; Yaltırak *et al.* 2002) but also the coastal areas within the region.

6. CONCLUSIONS

Neotectonic effects in northwestern Turkey are evident in uplifted terraces/beach sands, tilted deposits/erosional surfaces, diverted creeks, deeply-cut valleys, truncated ridges, and sag ponds. Such features indicate the importance of tectonic uplift and lateral movements in the geomorphological/morphotectonic development of the region. There are two questions of principal interest: (1) How recently did these features develop? and (2) How far did they influence the paleogeography of the recent geological past? Though very limited at present, data indicate the possibility that many geological/morphological features may be representative of the latest period of neotectonics: late Pleistocene-Holocene. Within the last 100,000 years, lateral displacements on the order of 2 km, and vertical displacements minimally on the order of 100 m must have occurred in northwestern Turkey. Even at the presently accepted rates, the connections between adjacent basins and coastal configurations would have been seriously modified.

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25

SEA-LEVEL CHANGES DURING THE LATE PLEISTOCENE-HOLOCENE ON THE SOUTHERN SHELVES OF THE BLACK SEA

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Abstract: Seismic profiles and gravity cores collected from shelf areas off the Sakarya Delta and Thrace Peninsula, together with available previous studies, were evaluated in this research. A major shelf-crossing erosional unconformity was observed, lying as deep as –120 m along the southern margin of the Black Sea. This erosional surface indicates a lowstand of the Black Sea's lake. The present shelf edge was therefore a former paleoshore environment of the pre-existing Neoeuxinian basin. Unit 1 overlies the erosional surface and includes a widespread, thin mud drape, the Sakarya Delta, and various irregular depositional features off the Thrace Peninsula. Unit 2 was deposited at the shelf edge and consists of a lowstand sedimentary wedge (off the Sakarya Delta) and seaward prograding clinoforms (off the Thrace Peninsula), all indicative of deposition during the lowstand. Radiometric dates from sedimentary cores collected above the erosional unconformity vary between 11.8 and 8.6 ky BP. Further studies will be required to obtain a better understanding of the timing and intensity of these sea-level changes.

Keywords: Black Sea, sea level, seismic stratigraphy, radiocarbon ages

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1. INTRODUCTION

The Black Sea is presently the largest inland anoxic body of water. Its connection to the world ocean is provided by three straits (the Bosphorus, Dardanelles, and Gibraltar) and two linking inland seas (the Marmara and Mediterranean). During the Late Quaternary, the Black Sea was subjected to dramatic geological and oceanographic changes, and as a result of its confined geographical situation, its sea-level changes during glacial and interglacial periods have been more complicated than those of the global oceans.

Based on the investigations conducted by scientists of the former Soviet Union, sea level in the Black Sea varied in a fluctuating manner during the Late Pleistocene-Holocene. Rises and falls occurred at scales as small as 10 m and less, and they took place somewhat independently of sea-level changes in the world ocean. Controversy was introduced into this matter in the late 1990s when Ryan *et al.* (1997) proposed a hypothesis that contradicted the fluctuating scenario of the Russian model. They suggested instead a rapid and abrupt transgression of Mediterranean water, which catastrophically invaded the subaerially exposed shelf of the Black Sea. The force and volume of this inundation were compared to the biblical flood of Genesis, and the authors even suggested that the Black Sea event might have given rise to the later Noachian Flood narrative (Ryan and Pitman 1998). Their proposed great flood in the Pontic basin was dated at about 7150 BP in Ryan *et al.* (1997) but revised to 8400 BP in Ryan *et al.* (2003).

Criticizing this flood view, Aksu *et al.* (2002a, b), Hiscott and Aksu (2002), Hiscott *et al.* (2002), and Kaminski *et al.* (2002) advocated for a persistent and strong outflow of the Black Sea during the entire period of the late glacial to Holocene, and they presented multidisciplinary data collected from the Marmara Sea and from the southwestern shelf of the Black Sea in support.

The aim of this paper is to summarize previous evidence from the southern (Turkish) shelf of the Black Sea together with new data in order to (1) contribute further information regarding the mode and intensity of sea-level changes in the Black Sea and (2) test each of the opposing views.

1.1 Fluctuating Sea Level During the Late Pleistocene to Holocene

The Pleistocene epoch in the Black Sea was characterized by alternations between a flow-type and a semiclosed water body. In a broad sense, marine transgressions occurred during interglacials, whereas during glacial periods, the basin became a flow-through brackish water body (Fedorov 1988; Svitoch *et al.* 2000). Such flow-through regimes should have been interrupted, however, when the maximum lowstand dropped below the sill depth of the Bosphorus. When the level of Caspian Sea was higher than that of the Black Sea, a flow-through system channeled via the Manych-Kuma area decreased the elevational differences between the two basins. Only after Mediterranean water began to spill over the Bosphorus sill in the early Holocene was a stable connection established between the Mediterranean and Black Seas. Sea-level fluctuations in the Black Sea did not occur in synchrony with those in the Mediterranean or Caspian Seas, as they were largely controlled by the depth of the sills in the Manych Depression and at the Bosphorus and Dardanelles Straits (Pirazzoli and Pluet 1991:99–105). Estimates of the ages and amplitudes of these fluctuations vary among the investigators who have studied them.

Fedorov (1988) distinguished a two-phase transgression in the Black Sea (Neoeuxinian and Holocene) with the waning of glaciation. The first phase resulted from increasing river runoff and produced a slightly brackish flow-through system discharging via the Bosphorus. It came to an end at about 9–10 ky BP with no entry of salty Mediterranean water. Although some investigators (Nevesskaya 1965; Kvasov 1975; Fedorov 1978; Shcherbakov *et al.* 1978) felt that the Neoeuxinian basin had always been flow-through in character, evidence of a closed lake receiving Caspian water prior to the establishment of a flow-through system has been demonstrated during the Neoeuxinian phase using isotopic analysis (Nikolaev 1995; Svitoch *et al.* 2000).

The second phase of the transgression was marine; this is evident from the boundary between the brackish Neoeuxinian deposits and those of the overlying Old Black Sea that contain Holocene transgression sediments with predominantly Mediterranean shells. The fauna of these deposits indicates that the brackish flow-through basin was replaced immediately, or after a short interval, by a weakly saline basin not exceeding 10‰ from 9–8 ky BP (Svitoch *et al.* 2000), and then by a saline basin reaching 19 or 20‰ (Fedorov 1988). The transgression maximum was 1.5–2 m higher than present sea level. This Holocene transgression triggered by the influx of Mediterranean water through the Bosphorus was dated to about 8–7 ky BP (Kvasov 1975; Fedorov 1978; Shcherbakov 1982).

Different investigators have proposed different numbers of transgressive and regressive cycles during the Holocene. Fedorov (1988) distinguished two phases of Holocene transgression, whereas Arslanov *et al.* (1983) and Balandin and Trashchuk (1982) found five transgressive-regressive phases based on archaeological materials. Yanko (1990) recognized six transgressive-regressive cycles during the Holocene using biostratigraphic, paleoecologic, and paleogeographic analysis of foraminifera. At the beginning of the Holocene transgression, the Black Sea rose from –65 m to –35 m between 9.4 and 8.0 ky BP according to Yanko-Hombach *et al.* (2002). During the following transgression-regression cycles between 8 and 5 ky BP, sea level oscillated from –55 to

–15 m and rose in absolute terms to several meters higher than the present level at about 5 ky BP (Nevessky 1967; Svitoch *et al.* 2000; Yanko-Hombach *et al.* 2002). In the late Holocene, however, only small sea-level fluctuations occurred: a fall of 6–8 m during the Phanagorian regression between 3–2 ky BP and a subsequent rise of 2–3 m during the Nymphaean transgression of about 2–1 ky BP (Fedorov 1978).

1.2 Abrupt Drowning of the Shelf in the Holocene

Ryan *et al.* (1997) proposed a catastrophic drowning of the exposed Black Sea shelf by an invasion of Mediterranean water at 7.15 ky BP and later revised the dating to 8.4 ky BP (Ryan *et al.* 2003). Based upon climatic and hydrographical features of both the Caspian and the Black Seas, together with the evidence of calcium carbonate content, isotopes of oxygen and carbon (Major *et al.* 2002), coastal dunes, and erosional surfaces, Ryan *et al.* (2003) also suggested that the Black Sea (Neoeuxinian basin) had been an inland lake whose surface had been drawn down beyond its shelf break to a lowstand level of –105 m by evaporation and reduced river input at about 11 ky BP. During the cool Younger Dryas interval (11–10 ky BP), the changing water balance of the Black and Caspian Seas caused a sea-level rise to the –30 m isobath, at which point Black Sea water spilled through the Bosphorus into the Sea of Marmara. As the climate became warmer, however, the Black Sea shoreline fell to the level of the outer shelf or, occasionally, beyond the shelf edge due to aridity; pollen spectra in deep-sea cores indicate extremely dry conditions (Ryan *et al.* 2003). At 8.4 ky BP, a second transgression was produced by Mediterranean water, which had by then penetrated the Bosphorus Strait, risen to the level of its sill, broken through, and abruptly refilled the lake. In this way, a basin that had dropped to –95 m was catastrophically inundated, submerging the exposed continental shelf to the –30 m level once again (Ryan *et al.* 2003).

Lending support to this low-level lake scenario, Ballard *et al.* (2000) found a paleoshoreline at a depth of –155 m on the southern Black Sea shelf off the Sinop Promontory. The paleoshoreline was characterized by a typical beach profile with a beach berm, low tide terrace, a longshore trough exposing the surface of the ancient lake bottom, and a longshore sand bar. Molluscs recovered from this paleoshoreline fall into distinct freshwater and marine assemblages. The oldest and youngest freshwater molluscs from the dredged sample gave radiocarbon ages of 15.5 and 7.4 ky BP, respectively, while the oldest marine molluscs dated to 6820 BP. High resolution seismic studies showed evidence of a rapid transgression over an erosional surface; this surface was found at a water depth of –105 m at the Bosphorus and its exit to the Black Sea shelf (Demirbağ *et al.* 1999) and in the area of the Sakarya Delta (Algan *et al.* 2002). Winguth *et al.* (2000) found the lowstand level dating to the Last Glacial Maximum at

–151 m on the Romanian shelf.

Sperling *et al.* (2003) showed indirect evidence of a Mediterranean origin for this Black Sea transgression by explaining the source of the sapropelic layers in the Marmara Sea. Heavy isotopes and high sea surface water salinities indicate that the initiation of sapropel deposition was associated with the increasing primary production brought on by Holocene global warming.

Decades before these investigations, Ivanov and Shmuratko (1983) had already compared continental moisture variations with the level of oceans and lakes, and explained the sea-level variations within the Black Sea in a way consistent with that proposed by Ryan *et al.*(2003). They suggested that a one-way flow of Black Sea water into the Mediterranean occurred during the Würm glacial maximum because the Azov-Black Sea basin had a positive balance. During the post-glacial warming, however, the level of the Black Sea dropped, according to the closed seas and lakes rule, until it was compensated by inflow of Mediterranean water through the Bosphorus.

1.3 Non-catastrophic Connection and Persistent Outflow Since ~11 ky BP

Post-glacial inflow of Mediterranean water to the Black Sea began as occasional spilling over the Bosphorus (~9 to 7 ky BP), based on multi-disciplinary data from deep basin Black Sea sediments and, indirectly, from the global sea-level curve (Degens 1971; Deuser 1974; Maynard 1974; Scholten 1974; Ross and Degens 1974).

Stanley and Blanpied (1980) modeled the development of the two-way exchange of water masses through the straits during the last 12 ky. Their model postulated that prevailing anaerobic lacustrine conditions in the Marmara Sea gradually shifted to partially anoxic conditions as stratification was established by excess freshwater overflow from the Black Sea between 12.0 and 9.5 ky BP. At that time, the Black Sea was receiving large annual meltwater discharges from the major tributary rivers. The two-way flow regime was established with the Marmara Sea between 9.5 and 7.0 ky BP as a result of the confluence of both Mediterranean and Black Sea overflows. Today's oceanographic conditions began ~3 ky BP with the reduction of excess freshwater loss from the Black Sea.

Evidence of continuous outflow from the Black Sea during the last glacial-Holocene period has also been shown by studies carried out in the Aegean and Marmara Seas, and on the southern shelf of the Black Sea. Strong outflow of the fresher Black Sea water has been implicated as a very important factor in the formation of the most recent Mediterranean sapropel (Ryan 1972; Stanley 1978; Thunell and Lohman 1979). The most recent sapropel S1 in the Aegean Sea was deposited between 9.6 and 6.4 ky BP in association with low salinity surface water originating from the Black Sea, which prevented vertical mixing and deep water ventilation (Aksu *et al.* 1995a, b). Using high resolution seismic data, Aksu *et al.* (1999) showed that a vigorous outflow from the Black Sea between about 9.5 and 7 ky BP formed west-directed sandy bedforms in the western Marmara Sea, and they also correlated the sill depths of the Dardanelles and Bosphorus straits with the global sea-level curve. They proposed that the formation of most of the Mediterranean sapropels was tied to periods of outflow from the Black Sea during the Late Quaternary.

Çağatay *et al.* (2000) pointed out the significance of a strong, nutrient-rich, freshwater, Black Sea outflow for the formation of sapropelic layers in the Marmara Sea between 10.6 and 6.4 ky BP. Further, a great majority of the articles in the Special Issue of *Marine Geology*, volume 190, favor a strong and persistent outflow from the Black Sea to the Mediterranean at about 11–9 ky BP based on seismic, geochemical, isotopic, and micropaleontological data.

Aksu *et al.* (2002a) showed that the uppermost unit (Unit 1) on the southwestern Black Sea shelf, which consists of a lowstand systems tract, a transgressive systems tract, and a highstand systems tract, was deposited during the last glacial lowstand and Holocene. Linear extrapolated ages from the sedimentary cores indicated that the transgression started at about 11–10.5 ky BP (Aksu *et al.* 2002a). Two sapropel layers, M1 and M2, were defined in the Marmara Sea, having been deposited in the intervals 29.5–23.5 ky and 10.5–6.0 ky BP (Aksu *et al.* 2002b). Multi-proxy paleoenvironmental data from dated cores demonstrated that their formation was closely related to the existing communication with the Black Sea (Aksu *et al.* 2002b; Abrajano *et al.* 2002). Benthic foraminifera in the Marmara Sea sediments indicated that saline Mediterranean water could not penetrate into the Black Sea until after 9.1 ky BP due to the strong outflow of brackish-fresh Black Sea water (Kaminski *et al.* 2002). A subaqueous delta at the southern exit of the Bosphorus was also shown to be built by the strong and persistent Black Sea outflow from the strait at about 10 ky BP (Aksu *et al.* 2002c; Hiscott *et al.* 2002).

2. PHYSICAL CHARACTERISTICS OF THE COASTAL ZONES AND SHELVES ALONG THE SOUTHERN BLACK SEA MARGIN

The unique relief of the Black Sea is one of the main factors to be considered in studying the sedimentological processes associated with sea-level migration (Figure 1). The southern margin of the Black Sea is both morphologically and geologically different from the northern margin. Extensive plains and large rivers occur in the north, whereas the high and steep North Anatolian

Mountains (Pontides) and narrow shelf zone of the southern Black Sea littoral (see Yılmaz, this volume) create very different hydrodynamic and sedimentation regimes in the shallow water of the coastal zone and in the shelf area. The physical characteristics of this southern margin will be summarized briefly below, proceeding from east to west.

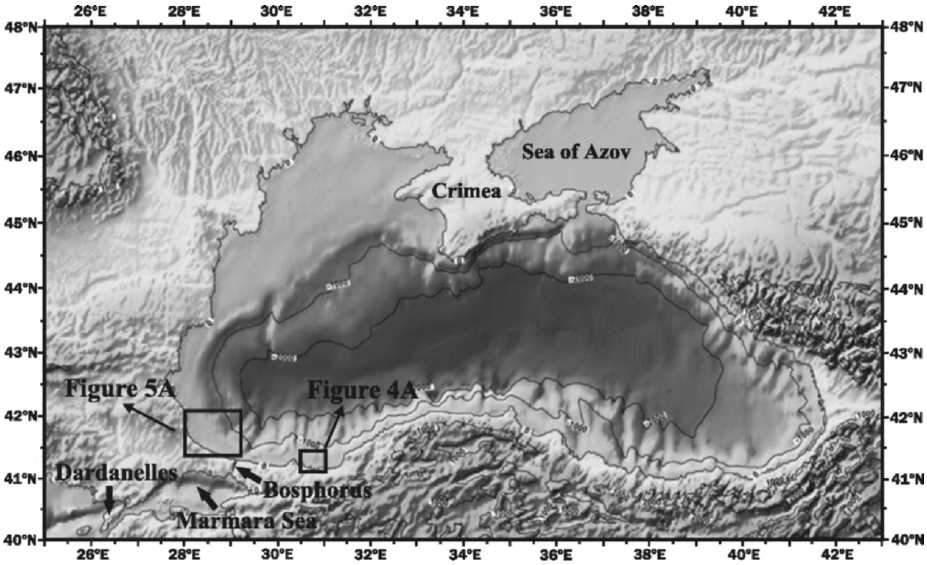


Figure 1. Physiographic features of the Black Sea and its surroundings.

The Turkish Black Sea coastline (Figure 2) is about 1625 km long (Darkot *et al.* 1975). Its topographic character is mainly erosional with 20–30 m high cliffs, however, depositional coasts with relatively low relief exist in the western and central sectors, where the Kızılırmak and Yeşilırmak Rivers have built large deltas (İzbırak 1972; Atalay 1994). Relief varies from east to west. The high Pontide Mountains extend almost parallel to the coast from the eastern boundary of Turkey to the Sakarya River in the west, forming a series of cliffs or rocky promontories that alternate with bay-head beaches. These mountainous coasts include high terraces with a fluvial cover at 60–70 m and erosional depositional surfaces of Late Pliocene-Early Pleistocene age up to 200–250 m. Middle and Upper Pleistocene terraces are evident at 5, 10–15, and 20–24 m (Erol 1985). The coastline west of the Sinop Promontory is extremely steep, with cliffs and the continuation of a fairly wide continental shelf. Relief decreases west of the Sakarya River, however, sedimentary rock formations of Paleozoic to Tertiary age associated with volcanics form high cliffs with small bay-head shingle beaches in this part of the coast (İnandık 1963).

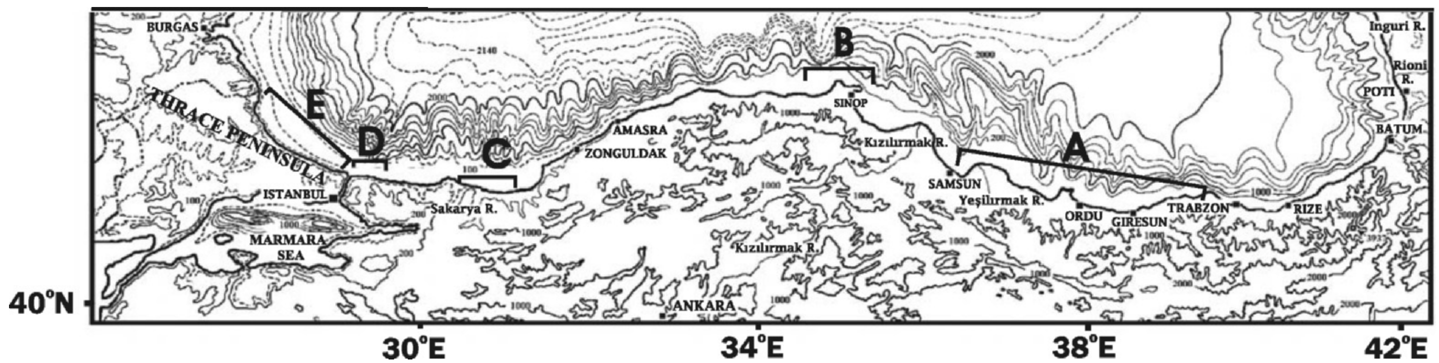


Figure 2. Physiographic features of the Turkish coast and shelf zone (from Ross and Degens 1974). Capital letters indicate the shelf areas studied by various investigators: A Trabzon to Samsun; B Sinop Promontory; C Sakarya Delta; D Northern exit of the Bosphorus; and E the Thrace Peninsula.

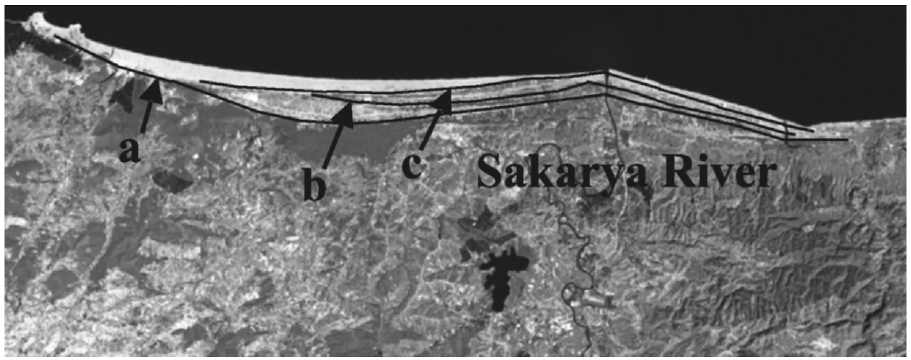


Figure 3. Prograding shorelines in the Sakarya Delta.

Old shorelines in the Sakarya Delta were distinguished based upon onshore studies, aerial photographs, and Landsat imagery (Figure 3). Shoreline (a) is the oldest and coincides with inactive (old) cliffs. The boundary between the mobile coastal dunes and the stationary ones landward of them delineates shoreline (b). The present shoreline (c) is a narrow sandy beach zone. Seaward progradation of the shorelines is related to the growth of the delta (Algan *et al.* 2002). The coastal zone is dominated by wide beaches and extensive dune fields to the west of the Bosphorus and becomes extremely high farther west in the Thrace Peninsula where the Strandja Mountain chain forms a cliff coast (Erol 1985).

The Turkish shelf zone is narrow compared to the northwestern part of the Black Sea. It is situated in front of the Pontide Mountains and is incised by numerous submarine canyons. Hence, the shelf sediments are not preserved extensively but are immediately transported via these canyons beyond the shelf edge at –100 to –110 m and into the deep Black Sea basin. In the southwest, the bottom morphology of the shelf is locally smooth with particularly irregular characters (described below in section 3.3).

The width of the shelf varies between 10 and 15 km from the east to the Bosphorus in the west. Local widening of the shelf results from sedimentation by rivers, such as the Kızılırmak and Yeşilırmak. The shelf area narrows to less than 2 km where the Sakarya River forms its delta. The Sakarya Delta lobe is deeply dissected by a submarine canyon to a water depth of about –50 m a short distance (1.5 km) from the river mouth. This deep channel comprises two discrete heads that coalesce at –800 m water depth and then continue to –2000 m as one large canyon. At the northern exit of the Bosphorus, the shelf is cut by an 80–90 m deep channel representing the northward extension of the strait (Di Iorio and Yüce 1999). This channel disappears at the outer shelf, however, a deep canyon is located at the shelf break aligned with its course. The widest part of the shelf (maximum 25 km) occurs along the Thrace Peninsula in the west, but it narrows again to the east (minimum 5 km).

3. SEISMIC STRATIGRAPHY OF THE SOUTHERN SHELVES

Previous investigations on the Turkish shelf have been few (Norman and Atabey 1987; Okyar *et al.*1994; Okyar and Ediger 1999; Demirbağ *et al.*1999; Görür *et al.* 2001; Aksu *et al.* 2002a; Algan *et al.* 2002) compared to the extensive studies conducted in the northwestern shelf area. A brief summary of these previous studies (between Trabzon and the Bosphorus in Figure 2) is given below together with new data from the southwestern (Thrace Peninsula) shelf.

3.1 The Shelf Area Between Trabzon and Sinop

The shelf area between Trabzon and Samsun was studied by Okyar *et al.* (1994) and Okyar and Ediger (1999) using high-resolution seismic reflection profiles (Figure 2). They defined two seismic units (A and B) separated by an irregular reflector (R). This reflector represents an erosional unconformity, developed at the falling of sea level during the Last Glacial Maximum (Okyar *et al.* 1994). The upper unit (A), observed in the nearshore zone, has a thickness between 5 and 20 m as well as variable internal configurations. Prograding sigmoid reflectors are interpreted as representing shallow water deposits related to Holocene sedimentation from the coastal rivers, whereas parallel reflectors are formed under uniform rates of accumulation (Okyar *et al.* 1994). Acoustic turbidity caused by enhanced gas content (mainly methane of biogenic origin) in the sediments prevented the researchers from observing further details of the Unit A sequence (Okyar and Ediger 1999). Locally, on a sill-like structure with a relief of 15 m located in 30–45 m water depths, the Unit A sequence displays onlaps indicative of rising sea level. The lower Unit B sequence is characterized by parallel and chaotic reflections. Based on onshore borehole data, it represents both consolidated alluvial sequences and Cretaceous volcanic formations. Locally, it displays toplap terminations at 91 m water depth to its upper boundary, which is the erosional surface (R) (Okyar *et al.* 1994).

Farther west, near Sinop (Inceburun Peninsula) Promontory (Figure 2), Norman and Atabey (1987) recognized rocky surfaces, widespread landslides, and four platforms at varied depths using high-resolution seismic profiles. The platform levels at –15, –35, –50, and –65 m display lateral discontinuities over the shelf break at –100 m. Some platforms contain sediment deposits up to 20 m thick. They also found two possible lowstand terraces at –15 and –50 m partly eroded during subaerial exposure. North of the promontory, the platforms at –50 and –65 m were completely eroded as indicated by the exposed and inclined rocky surface. Without any age determination, Norman and Atabey (1987) postulated that the –65 and –50 m platforms correspond to the lowstands at 12 and 10 ky BP by comparison with the dated platforms in Crimea (Balabanov *et al.* 1981). Using the same correlation approach, the –35 m platform was associated with 8.3 ky BP. Regressions, wave erosion, tectonic activities, and landslides were offered as the reasons for discontinuity between these platforms.

3.2 The Shelf Area Between the Sakarya Delta and the Bosphorus

High-resolution seismic reflection profiles collected from the southern shelf where the Sakarya River constructed its delta (Figure 2) indicate three distinctive seismic units (Algan *et al.* 2002). The most prominent feature is a widespread erosional unconformity that truncated the lower units. It separates Unit 1 from Units 2 and 3 below, and it is located at depths of –92 to –105 m, which was calculated using an average seismic velocity of 1500 m/s from 122–140 msec twtt (milliseconds of two-way travel time) (Figure 4A–F). This erosional unconformity between Units 3 and 1 on the shelf extends to the shelf edge where it overlies Unit 2. It is nearly flat in the middle and outer shelf areas, and becomes relatively shallower in the landward direction. This widespread erosional surface indicates a subaerial exposure of the shelf during the last lowstand of the Black Sea.

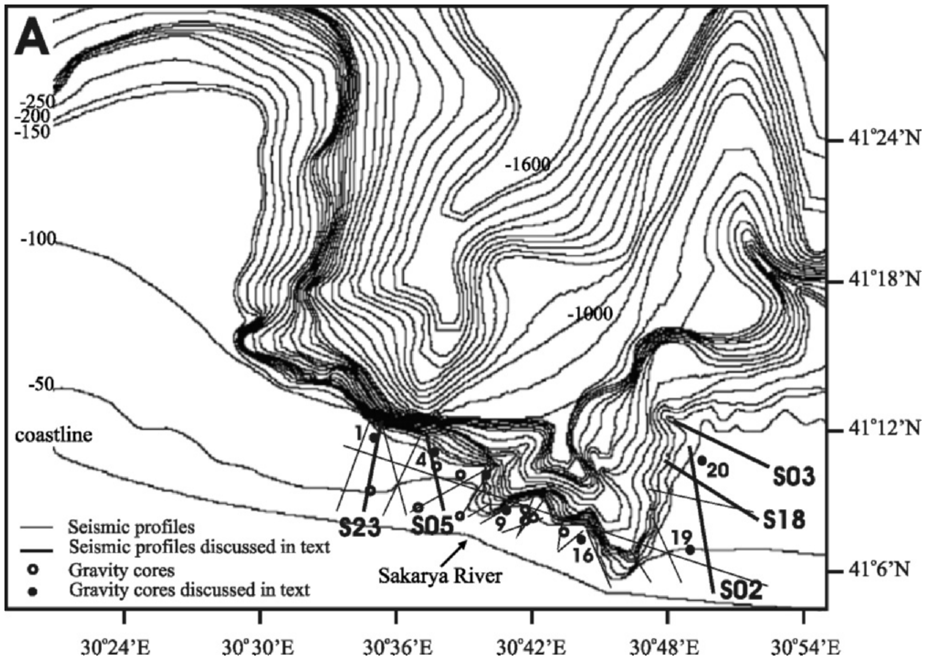
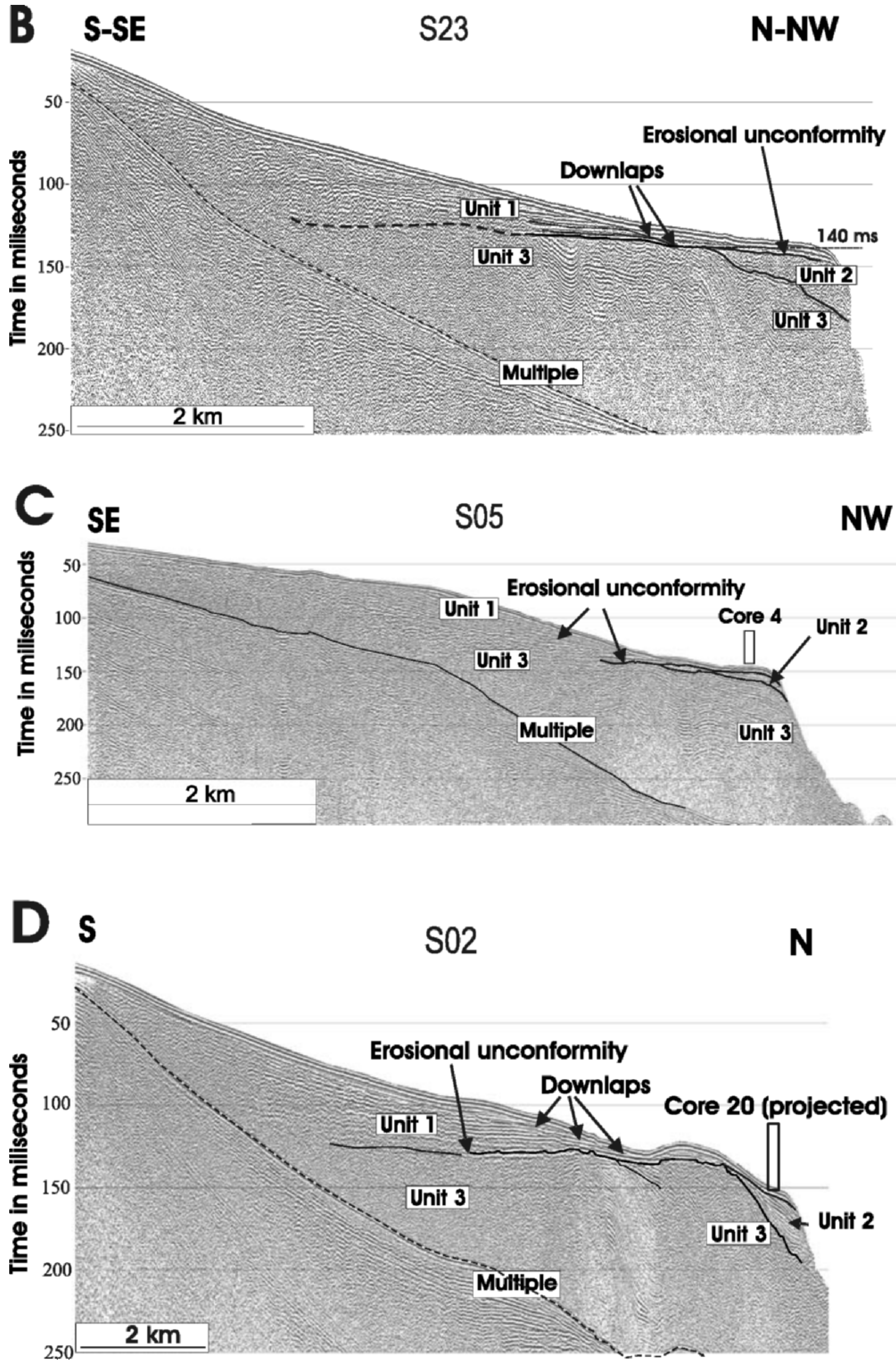
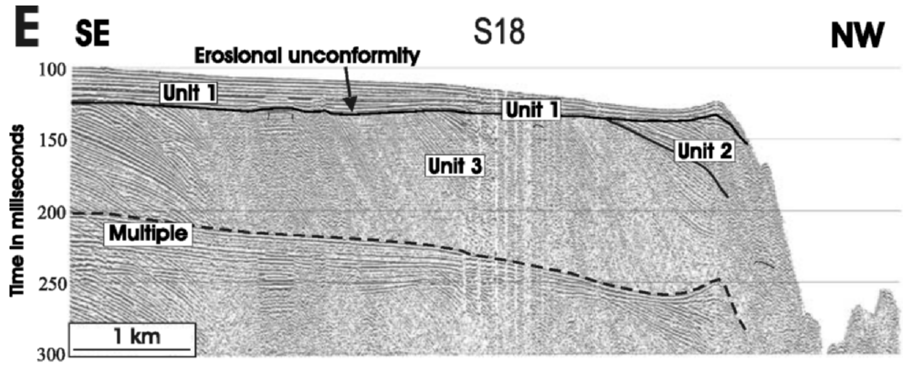


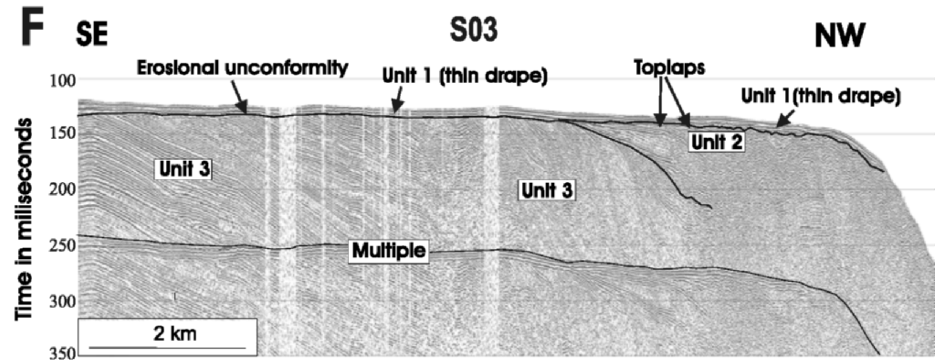
Figure 4. **A** Bathymetric map and location of seismic profiles and gravity cores collected from the shelf area off the Sakarya River with depths in meters (adapted from Algan *et al.* 2002). **B–F** (following pages) Single channel seismic reflection profiles collected in May 1999 on board *R/V Arar* showing seismic stratigraphic Units 1–3 (B, C, D, and F are modified from Algan *et al.* 2002).

The uppermost Unit 1 reveals a high-angle oblique progradational pattern of seaward-dipping reflectors in areas proximal to the Sakarya River mouth (Figure 4B–D), whereas it shows a parallel aggradational character in more distal areas (Figure 4E). The downlap terminations occur at a water depth of about –90 to –96 m (120–128 msec twtt). The thickness of Unit 1 increases landward up to ~30 to 50 msec, it becomes thinner seaward as well as laterally,



B
S-SE
S23
N-NW
Time in miliseconds
50
100
150
200
250
Erosional unconformity
Downlaps
Unit 1
140 ms
Unit 3
Unit 2
Unit 3
Multiple
2 km
C
SE
S05
NW
Time in miliseconds
50
100
150
200
250
Erosional unconformity
Unit 1
Core 4
Unit 2
Unit 3
Multiple
Unit 3
2 km
D
S
S02
N
Time in miliseconds
50
100
150
200
250
Erosional unconformity
Downlaps
Core 20 (projected)
Unit 1
Unit 3
Unit 2
Unit 3
Multiple
2 km





and it almost disappears east of the river mouth (Figure 4F). East of the river mouth, where the shelf is smooth, flat, and lacking deltaic deposition, a thin mud layer overlies the erosional surface.

Unit 2 reveals an oblique-parallel internal configuration and an onlap relationship with the next lower unit, while erosional and toplap terminations are observed at its upper contact. It is wedge-shaped and exists only at the shelf edge; its thickness is limited where the canyon incision is deep. Unit 2 represents lowstand deposits laid down at a time of lower sea level (about –105 m, or ~140 msec).

Beneath the major erosional unconformity lies the lowermost Unit 3, which has been interpreted as the basement. Its upper boundary delineates the shelf area and is covered by Unit 2 at the shelf edge. The internal reflectors of this unit are inclined parallel, locally chaotic, and deformed by faults that affect the folded strata and filled channel. Unit 3 is possibly an extension of Upper Cretaceous limestone, volcanogenic sedimentary rocks, and also Oligocene-Lower Miocene sedimentary sequences that are exposed onshore (Oktay *et al.* 1992; Yılmaz *et al.* 1997).

The lateral continuation to the west of this erosional unconformity has been recognized at the same water depth on the shelf near the northern exit of the Bosphorus channel (Figure 2) by Demirbağ *et al.* (1999). Their fourth unit is defined as depositional, covering older units with onlaps at the shelf edge in a manner similar to Unit 2 on the shelf beyond the Sakarya Delta. They also observed a thin veneer of mud overlying the erosional unconformity and defined it as Unit E.

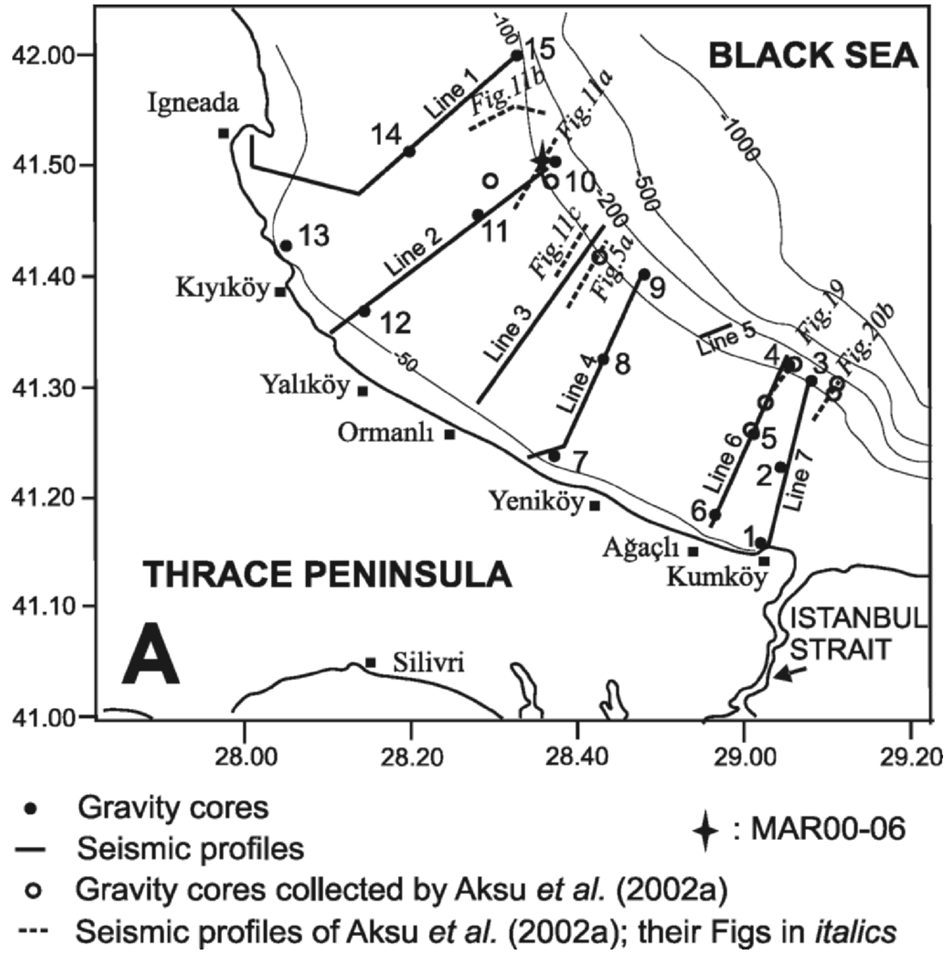
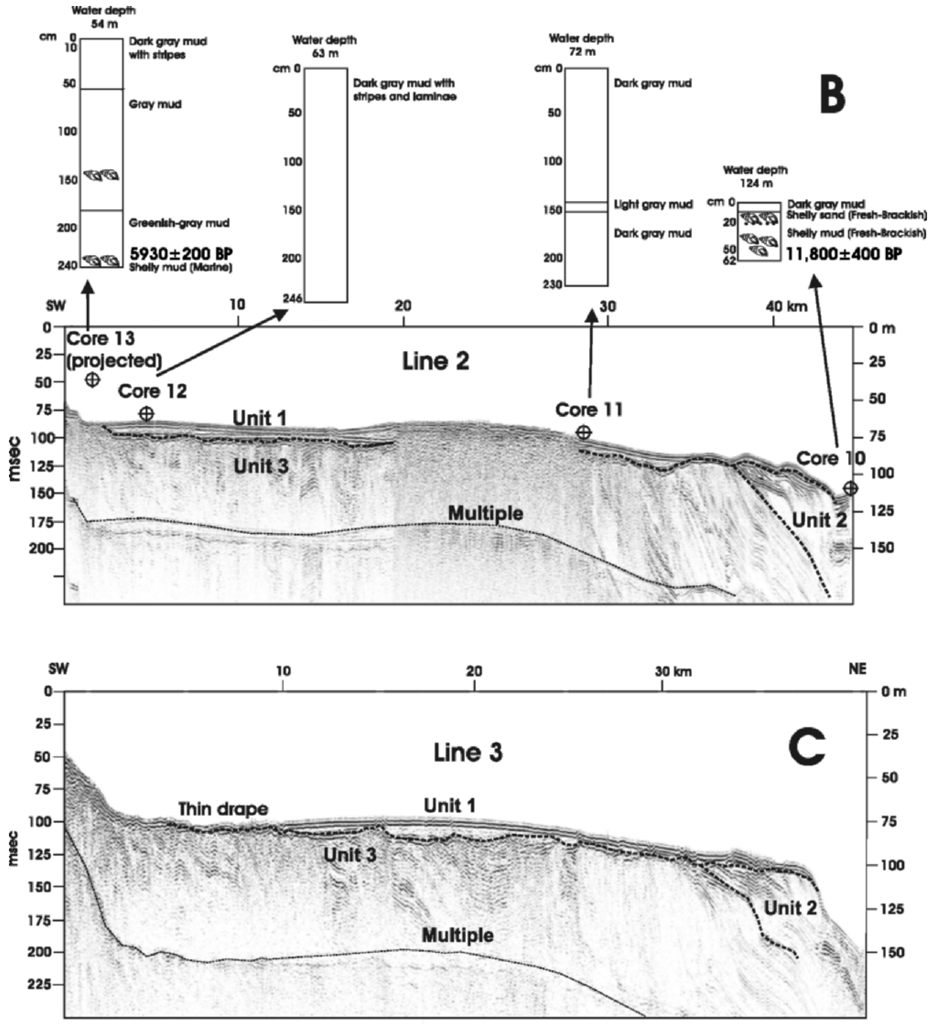
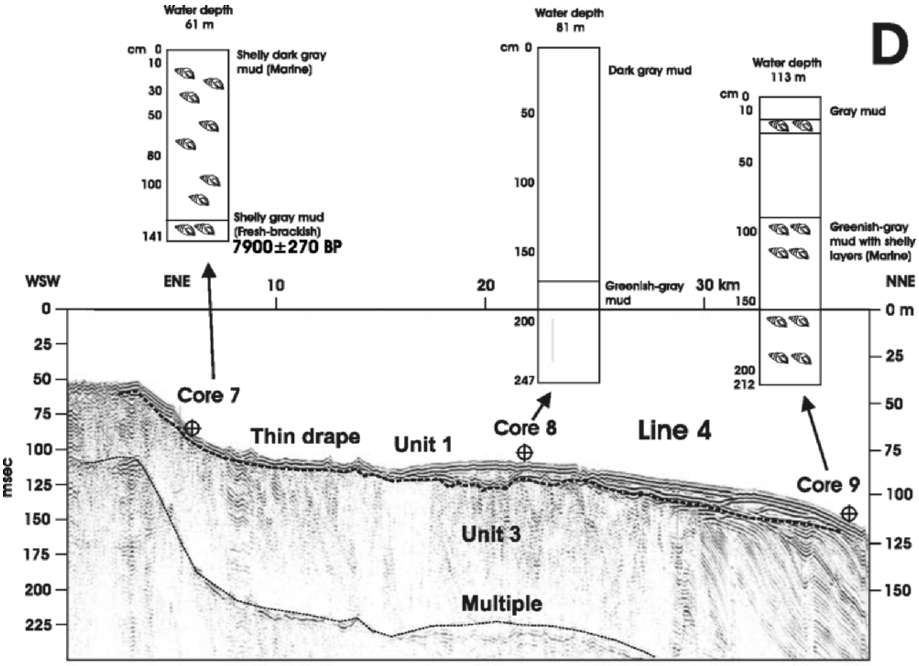


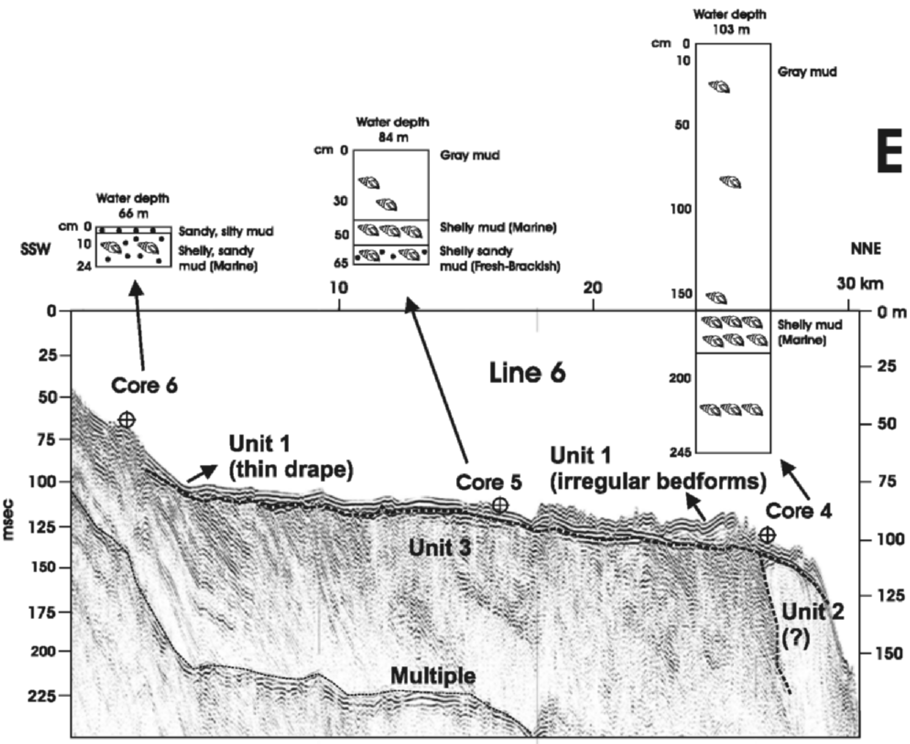
Figure 5. **A** Location of seismic profiles and gravity cores collected from the southwestern shelf area of the Black Sea (from Ergin *et al.* 2003). **B–F** (following pages) Single channel seismic reflection profiles collected during August 1999 on board *R/V Arar*, showing the seismic stratigraphic Units 1–3. Generalized lithological characteristics of the gravity cores are shown in related profiles; core 13 shown in Figure 3C was collected ~15 km northwest of core 12.

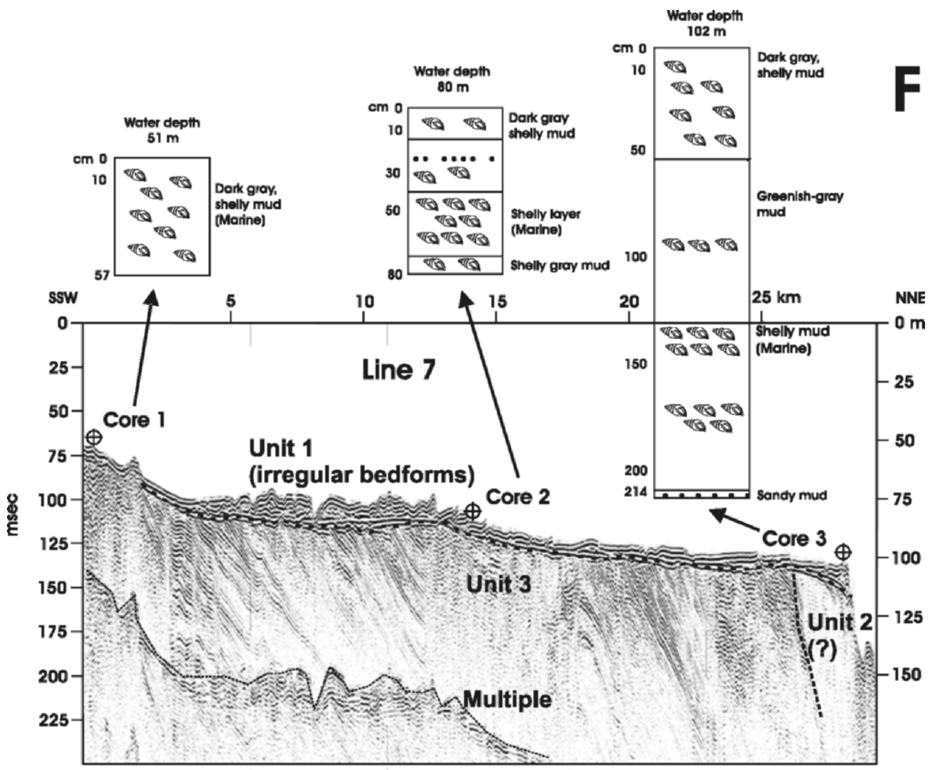
3.3 The Shelf Area Off the Thrace Peninsula

In August 1999, seven high-resolution seismic profiles were collected from the southwestern shelf of the Black Sea using a sparker system on board the *R/V Arar* of Istanbul University (Figure 5A). In the widest part of the shelf along the Thrace Peninsula (Figure 5B–F), three distinctive seismic units and a widespread erosional surface were recognized that appear similar to the units on the shelf off the Sakarya Delta. Unit 1 unconformably overlies the older units and consists of various depositional features. In the westernmost part of the area (Figure 5B, C, D), it is represented by a drape with parallel and weak reflectors, reaching a thickness of 10–20 msec. This Unit 1 drape is widespread over most of the inner and outer shelf with parallel internal configuration (Figure 5B, C).




D
Water depth
61 m
cm 0
10
30
50
80
100
141
Shelly dark gray mud (Marine)
Shelly gray mud (Fresh-brackish)
7900±270 BP
Water depth
81 m
cm 0
50
100
150
200
247
Dark gray mud
Greenish-gray mud
Water depth
113 m
cm 0
10
50
100
150
200
212
Gray mud
Greenish-gray mud with shelly layers (Marine)
WSW
ENE
10
20
30 km
NNE
0 m
msec
Core 7
Thin drape
Unit 1
Core 8
Line 4
Core 9
Unit 3
Multiple


E
Water depth
66 m
cm 0
10
24
Sandy, silty mud
Shelly, sandy mud (Marine)
Water depth
84 m
cm 0
30
50
65
Gray mud
Shelly mud (Marine)
Shelly sandy mud (Fresh-Brackish)
Water depth
103 m
cm 0
10
50
100
150
200
245
Gray mud
Shelly mud (Marine)
SSW
NNE
10
20
30 km
0 m
msec
Core 6
Line 6
Unit 1 (thin drape)
Core 5
Unit 1 (irregular bedforms)
Core 4
Unit 3
Unit 2 (?)
Multiple



At the shelf edge and above the erosional unconformity, this unit displays gentle, shingled, lenticular reflectors (Figure 5D). Unit 1 also displays depositional bodies having wavy morphologies and/or irregular bedforms on seismic profile lines 6 and 7 (Figure 5E, F). These distinctive features are found close to the channel north of the Bosphorus at water depths of –80 to –100 m but grading shoreward to a thin drape. Their distribution is not continuous, and though mostly parallel, their internal configuration does not show continuity either.

Higher resolution data indicated that the seismic unit above the erosional unconformity (α in Aksu *et al.* 2002a) consists of four subunits separated by two strong reflectors (α1 and α2). Subunits 1B and 1C were interpreted as barrier islands/beaches and sediment ridges, sediment waves, and current-generated marine bars on the basis of their geometry and seismic facies attributes (Aksu *et al.* 2002a). The locations of our seismic lines 6 and 7 (Figure 5E, F) are very close to those of Aksu *et al.* (2002a), so it is likely that the same features were observed in our respective profiles (see Figure 5A compared to their Figures 19 and 20). This deposition appears to have occurred during a relatively rapid but progressive sea-level rise (Aksu *et al.* 2002a). Similar depositional features were observed extending from the Ukrainian margin onto and across the Romanian

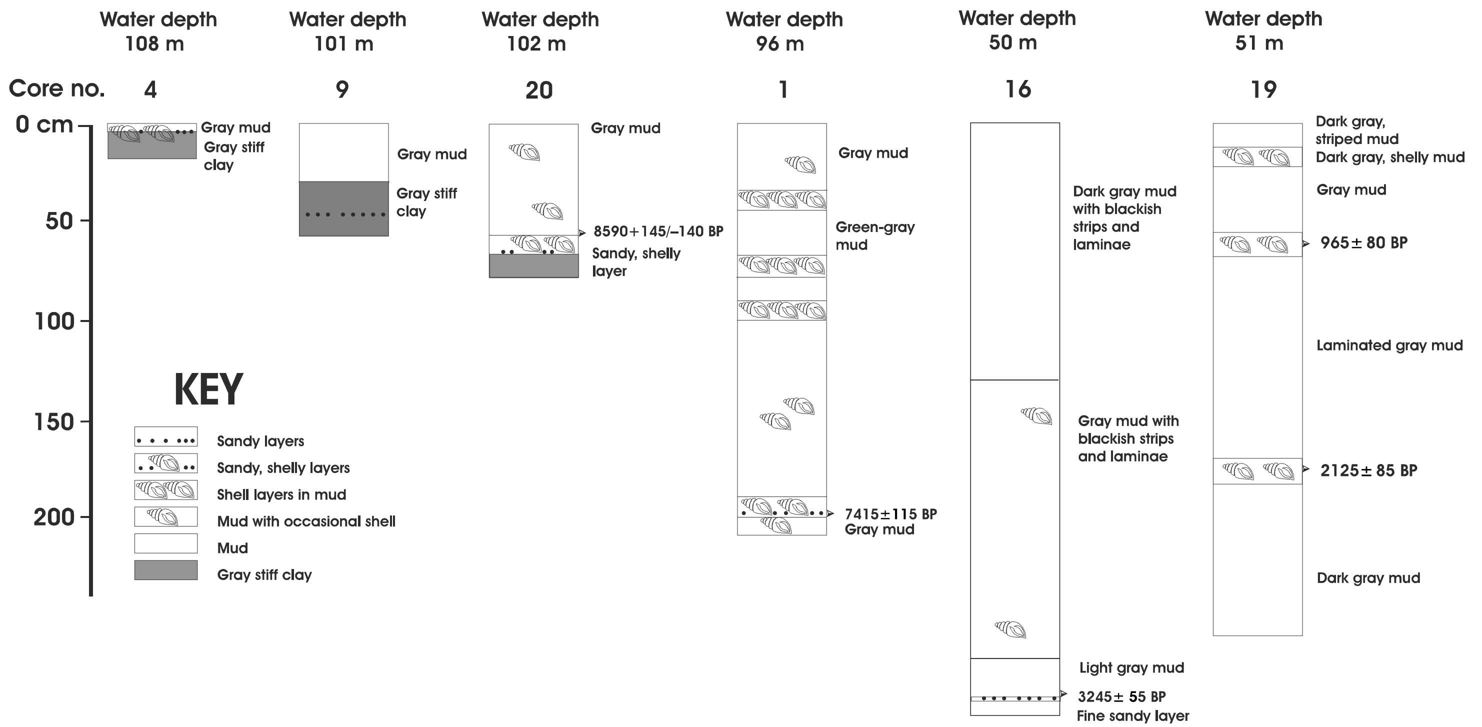


Figure 6. Generalized lithological characteristics of the gravity cores collected from the Sakarya Delta (see Figure 4A for locations).

margin, but these have been interpreted as drowned beaches, sand dunes, and soils from a coastal/littoral environment that surrounded a shrunken lake (i.e., the drawn down early Holocene Black Sea) prior to the marine inunda-tion (Ryan *et al.* 2003; Lericolais *et al.* 2003). Nevertheless, both interpretations can be reconciled by accepting that an initial coastal or littoral deposition took place that was later drowned by rising sea level.

Unit 2 occurs only at the shelf edge and can be observed in Figure 5B and C. It is represented by seaward-prograding clinoforms, indicating various lowstands and subsequent rises of sea level. These lowstand shelf-edge delta lobes have been observed by Aksu *et al.*(2002a) and defined as a subunit of Unit 1 (Unit 1A), underlying unconformity α (see Figure 5A to compare the seismic lines). The oldest sequence (Unit 3) is represented by acoustically strong reflectors and characterized by faulted and folded strata. It is draped by a thin veneer in the landward direction (Figure 5C–F). The erosional unconformity truncates Units 2 and 3 and is located at a depth of ~ –90 to –120 m on the outer shelf.

4. SEDIMENTATION ON THE SHELF

The gravity cores collected from the Sakarya Delta (Figure 4A) contain both deltaic sediments and the thin mud drape that overlie the erosional unconformity. The lithology of the cores collected in the vicinity of the river mouth (deltaic sediments) consists of light and dark gray, fine-grained mud, including laminated layers (Figure 6). These laminated layers comprise mainly silt and/or dark colored, fine sands, and some include shell fragments. Discrete shelly mud or shelly sand layers also occur locally in the eastern (core 19) and western edges (core 1) of the deltaic sediments. *Ammonia* (*A*. *tepida*, *A*. *parkinsoniana*, and *A*. sp.) is the dominant benthic foraminiferal genus in all the cores, as it is widespread in the Black Sea (Yanko 1990). *Elphidium* sp., *Quinqueloculina* sp., and *Porosononion* sp. are very scarce. In the deltaic sediments, the molluscs include *Plagiocardium papillosum*, *Acanthocardia paucicostata*, *Modiolus adriaticus*, *Chamelea gallina*, *Spisula subtruncata*, *Gouldia minima*, *Mactra* sp., *Hydrobia* sp., and *Rissoina* sp. Radiocarbon age determination of *Modiolus*, *Plagicardium*, and *Mactra* shells yielded ages of 965±80 BP at 60–65 cm and 2125±85 BP at 175–177 cm. At the bottom of core 16, which was collected near core 19, a coarse-grained, shelly layer gave an age of 3245±55 BP. Age determinations indicate the high sedimentation rate (80–90 cm/1000 yr) that is normally expected in deltaic environments.

Cores 4, 9, and 20 were collected from the shelf break (Figure 4A) at about –100 m, where the deltaic sediments thin out. These cores penetrated deposits that were laid down over the erosional surface, as can be seen in the seismic profiles S05 and S02 (Figure 4C, D). Comparing the core location with the seismic profiles suggests that the hard substratum beneath this layer should be interpreted as the upper part of the erosional surface. The thin mud drape is located at the top, its thickness varying between 2 and 60 cm (Figure 6). Below this mud, a relatively coarse-grained, shelly layer occurs. There is a marked contact between this shelly layer and the stiff clay deposit with low water content at the base of these cores.

The uppermost 60 cm of core 20 contain greenish-gray mud, including occasional shells, while below this depth, a 2-cm-thick, shell-enriched layer overlies fine, sandy, stiff clay at the base of the core. The upper 30 cm of the core contain *Plagiocardium papillosum* and *Modiolus* sp., however, below 40 cm, *Dreissena polymorpha* and *D. rostriformis pontocaspia* are abundant, together with a few *Acanthocardia* sp., *Corbula gibba*, and *Limaria* sp. *Ammonia* is the most dominant foraminiferal genus, together with a minor presence of *Quinqueloculina* sp. only in the upper 25 cm of the core. The ^{14}C dating of a *Dreissena* shell from above the stiff clay gave an age of 8590 +145/–140 BP. Lithological characteristics of this core indicate that deposition began under high-energy conditions over the stiff, eroded substrate that lies at about –100 m, and it continued under low-energy conditions, thereby suggesting a rapid deepening of an initially shallow environment. The ^{14}C age from above this substrate suggests that the submersion occurred at about 8.6 ky BP.

Core 1 was recovered where the deltaic sediments thin out toward the shelf edge at a water depth of –96 m (Figure 4A). The lithology of this core consists of greenish-gray mud with discrete shelly layers 4–6 cm thick (Figure 6). Between 157 and 192 cm from the top of the core, the greenish-gray mud undergoes light and dark alternations, then below a sharp contact lies an 8-cm-thick, sandy, shelly layer. Molluscan fauna in core 1 are represented by *Fusus* sp., *Corbula* sp., and *Cardium* sp. in the upper part of the core, while below 180 cm, both brackish (*Dreissena polymorpha*, *D. rostriformis pontocaspia*, *Monodacna caspia*) and marine types (*Plagicardium papillosum*, *Acanthocardia* sp., *Mytilus* sp., *Corbula gibba*, *Crenalla* sp., *Natica* sp., *Hydrobia* sp., and *Turritella* sp.) coexist. *Mytilus* shells obtained between 192 and 200 cm within the core yielded a ^{14}C age of 7415±115 BP. The appearance of Mediterranean fauna in this layer reflects changing conditions in the Black Sea. The timing of the Holocene entry of Mediterranean water into the Black Sea varies between 9.5 to 7 ky BP depending upon the investigator. This age is within the limits of previous findings by Ross and Degens (1974), and it is close to those of Kvasov (1975), Fedorov (1978), Shcherbakov (1982), Ryan *et al.* (1997), Aksu *et al.* (2002c), and Ryan *et al.* (2003), however, it is slightly later than that of Kaminski *et al.* (2002). Nevertheless, our dating does not imply a precise timing for the changing conditions but might mark the beginning of colonization by euryhaline molluscs after connection with the Mediterranean.

A total of 15 gravity cores were collected from the southwestern Black Sea shelf along the Thrace Peninsula. Sediments deposited in the inner and mid-shelf zones comprise predominantly fine-grained, locally-laminated, dark and light grayish-green mud with distinct shelly layers (Figure 5B–F). However, from mid-shelf to the shelf edge, the coarse-grained fraction and shell contribution increase. The dominant molluscan compositions of the sediments are indicated in parentheses in Figure 5B–F (Marine or Fresh-Brackish). Core 13 was recovered from the inner shelf (water depth of 54 m) and consists of almost homogeneous mud, but it includes some shell fragments at its base (Figure 5C). These basal shells (210–213 cm within the core) are predominantly *Mytilus*, dated to 5930±200 BP, with few *Dreissena* sp. Core 12 was collected from a water depth of 63 m and represents Unit 1 of the inner to mid-shelf environment (Figure 5C). It consists almost entirely of mud with tiny black laminae and rare shell fragments. Core 11 was obtained from a mid-shelf area and is similar to Core 12 in texture and composition. Core 10 was recovered from the shelf edge at a water depth of 124 m; it could not penetrate farther than 62 cm due to the hardness of the bottom. The lower 42 cm consist of hard, shell-rich mud, which underlies 10 cm of fragmented shells with some silt and fine sand layers. The top 10 cm of the core consist of dark gray mud. Mollusc shells from this core are entirely *Dreissena* sp. *Dreissena* shells from the base of core 10 (55–60 cm) provide an age of 11,800±430 BP (Figure 5C). Aksu *et al.* (2002a) recovered their core MAR00-06 from a location very close to that of core 10 (Figure 5A), and it provided an age of 7.8 ky BP from a Mediterranean mollusc at almost 40 cm above the α unconformity. Subunits 1B, 1C, and 1D were covered by core MAR00-06. Considering the location of the core within our seismic profile (Figure 5B), the hardness of the substrate, and the proximity to the topset-foreset transition of the shelf-edge delta (Aksu *et al.* 2002a), it is likely that this core reflects sedimentation above the erosional surface.

Core 7 is 140 cm long and was obtained from a water depth of 61 m on the inner shelf (Figure 5D). It contains greenish-gray mud with abundant shell fragments. Down to a core depth of 110 cm, the molluscs are of Mediterranean affinity (*Mytilus*, *Cardium*), whereas shells and fragments of *Dreissena* and *Monodacna* prevail toward the base. Basal *Dreissena* shells have given an age of 7900±270 BP. The presence of brackish fauna has been reported on the inner shelf of the northern side of the Black Sea (Shcherbakov *et al.* 1978). Core 8 was collected from the mid-shelf at a water depth of 81 m, where Unit 1 becomes thicker. It is 247 cm long and contains dark greenish-gray mud with rare shell fragments down to 170 cm; the color of the mud changes to light gray toward the base of the core. The changing lithology at the shelf edge is evident from core 9 (Figure 5D), within which shells are present in several discrete layers throughout the 212 cm length of the core. The shells, however, are most abundant below a core depth of 100 cm as discrete layers between 3 and 7 cm thick within a matrix of alternating dark and light greenish-gray mud.

Core 6 was very short (24 cm). It was recovered from the inner shelf where the erosional surface may have exposed the sea floor. Because of the hard bottom condition, the gravity corer did not penetrate deeply. The lithology consists of abundant mollusc shells in a sandy matrix (Figure 5E). The molluscs are mostly *Mytilus*. Core 5 was collected from the mid-shelf at a water depth of 80 m. Similarly, this core could not penetrate much deeper than 60 cm (Figure 5E). Sediments in the upper 40 cm of the core are greenish-gray mud with locally abundant shells, whereas below this depth, and down to 58 cm, sediments gradually become coarser-grained due to the presence of fragmented shells. Between 58 and 65 cm depth, a fine, sandy layer occurs with abundant and dominant *Dreissena* sp., as in core 7. Core 4 captured a 245 cm sedimentary record and was obtained from a water depth of 103 m at the shelf edge (Figure 5E). Sediments from this core were mud, however, abundant shelly layers occurred at various depths.

Core 1 was raised from a water depth of 51 m and represents 57 cm of the inner shelf environment (Figure 5F). Sediments from this core comprise abundant shells in sandy mud, similar to core 6. Core 2 was recovered from a water depth of 80 m, in an area where the sea floor is marked by irregular bedforms. The top 20 cm can be identified as shelly mud, and below this, to a depth of 27 cm, a layer of broken shells and fine sand occurs. From 27 to 71 cm, both complete and broken shells are dominant, and a shelly mud appears at the base. Core 3 was collected from the shelf edge at a water depth of 102 m. It consists mainly of mud, often interlayered with shelly materials. At the base of the core (210–214 cm) was noted a layer of fine sand.

Along the shelf off the Thrace Peninsula, sedimentation gives rise mainly to fine-grained, siliciclastic mud, but in both shelf edge and inner shelf areas, a coarse-grained contribution by biogenic materials appears to be important. Dark gray mud containing Mediterranean mollusc species reflects the highstand drape, however, cores 13, 10, 7, and 5 penetrated to layers containing brackish fauna at their base, suggesting that the marine drape overlies Neoeuxinian sediments.

5. DISCUSSION AND CONCLUSIONS

The major shelf-crossing erosional unconformity observed along the southern margin of the Black Sea occurs at a water depth of about –100 m, however, its actual depth varies from as shallow as –91 m in the east near Trabzon-Samsun to as deep as –120 m on the southwestern shelf. The presence of submarine terraces or platforms in the vicinity of the Sinop Promontory (Norman and Atabey 1987) seems to be of local importance, since such terraces are not found (or preserved) along the rest of the southern shelf. Moreover, no age determination is available from the sediments deposited atop these features. A paleoshoreline at –155 m (Ballard *et al.* 2000) in the same area has been dated to 15.5 ky BP, suggesting submersion after that time.

In this study, the seismic units above the erosional unconformity were defined as Unit 1 and consisted of a widespread thin mud drape, deltaic sediments of the Sakarya Delta, and irregular bedforms off the Thrace Peninsula. Other erosional gaps defined by Aksu *et al.* (2002a) and Ryan *et al.* (2003) could not be observed due to lack of required resolution in our data for these sediments. The mud drape was extensively observed on the southern shelf. It lies just above the erosional unconformity between the Sakarya Delta and the Bosphorus (Demirbağ *et al.* 1999) and covers the irregular bedforms of the outer shelf off the Thrace Peninsula. Sediments just above the unconformity off the Sakarya Delta were dated to ~8.6 ky BP in core 20, suggesting a sea-level rise from about –100 m and an immediate submergence of the shelf edge. At about 7.4 ky BP, the fresh/brackish water basin reveals evidence of salinization via an influx of Mediterranean water.

Alternatively, another age determination obtained above the erosional unconformity may indicate an earlier lacustrine lowstand well after the glacial maximum but before 11.8 ky BP. The molluscan fauna from core 10 were distinct from all other cores, as they were entirely dominated by brackish types. If the transgression began at about 11.8 ky BP from –124 m and rose continuously to –100 m by about 8.6 ky BP, then there could not have been any continuous outflow from the Black to the Marmara Sea before 8.6 ky BP due to the higher sill depth of the Bosphorus Strait. Moreover, global sea level does not support such a scenario in that time span.

Instead, two separate transgressions, as proposed by Major *et al.* (2002) and Ryan *et al.* (2003), might provide a better explanation. The richness of the *Dreissena* shells above the erosional unconformity at the shelf edge off the Thrace Peninsula indicates a submergence by fresh or brackish water. Brackish fauna are also observed at the base of cores 13, 7, and 5, which were collected from the inner part of the southwestern shelf. As was previously pointed out (Kuprin *et al.* 1974; Shimkus *et al.* 1978; Major *et al.* 2002), this evidence indicates a local rise of the Black Sea covering the shelf, followed by another drop prior to the onset of marine conditions. This local rise was shown to have occurred during the Younger Dryas cool interval (11–10 ky BP) as a result of an increase in precipitation versus evaporation over the Black and Caspian Seas (Ryan *et al.* 2003). Although our age of 11.8 ky BP is slightly older than the time of the Younger Dryas, the location of the sediments and their constituent fauna support the conclusion that deposition over the erosional surface around

11 ky BP was due to the Neoeuxinian lake. After the falling sea level of the subsequent warm period (Major *et al.* 2002), the last transgression must have submerged the shelf area at about 8.6 ky BP, reaching about –20 m as indicated by a borehole on the deltaic plain. Above the freshwater sediments at –18 m, Mediterranean fauna dominate. Görür *et al.* (2001) dated the boundary between the lake sediments comprising the fresh-brackish molluscan fauna and the alluvial sediments of Sakarya River to 8090 BP.

The data presented in this study were collected from the southern margin of the Black Sea and appear to be compatible with the model proposed by Major *et al.* (2002) and Ryan *et al.* (2003). Indications of the smaller sea-level fluctuations that occurred during the Late Holocene (Ostrovsky *et al.* 1977; Chepalyga 1984) would not have been detected given the limits of this study. In view of the transgressive-regressive cycles scheme for the last 10–5 ky, Holocene regression is most likely the main reason for the gap between 10 and 8.4 ky BP (Yanko-Hombach *et al.* 2002). Distinct depositional bodies (irregular bedforms) observed on the outer part of the relatively wide shelf off the Thrace Peninsula occur at –80 to –100 m depths, but since our seismic data lack the necessary resolution, we cannot discuss whether their origin should be considered marine as Aksu *et al.* (2002a) propose or terrestrial as Ryan *et al.* (2003) suggest. The littoral depositional setting, however, suggests that sea level at the time stood at about –80 m, lower than both the sediment (–32 m) and basement (–70 m) sill depths of the Bosphorus (Gökaşan *et al.* 1997). The Black Sea, therefore, could not outflow through the strait when they were being deposited. Although the deposition of sapropel has been linked to a large outpouring of nutrient-rich, less saline Black Sea water, which brought about density stratification in the water column of the Marmara Sea (Çağatay *et al.* 2000; Aksu *et al.* 2002b; Abrajano *et al.* 2002) and in the Aegean Sea (Aksu *et al.* 1995a, b) at about 10.6 to 6.4 ky BP, new findings on the stable oxygen isotope ratios of foraminifera contradict the presence of low salinity surface water during the deposition of sapropel S1 (Sperling *et al.* 2003).

Further studies along the southern margin of the Black Sea are necessary in order to establish lateral correlation and integration of the data on a basin-wide scale.

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THE FROZEN BOSPHORUS AND ITS PALEOCLIMATIC IMPLICATIONS BASED ON A SUMMARY OF THE HISTORICAL DATA

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Abstract: Historically, the first evidence of a frozen Bosphorus was noted during the time of Herodotus. Analysis of the historical data about freezing events in the Bosphorus (at Istanbul) reveals the existence of four main cold periods since 1 AD. The first occurred around the 1st century. Although the temperature was close to, or perhaps a little lower than, that of the present, three successive freezing events are indicated between 7–17 AD. The second cold period was in the 4th century, when another freezing event was reported in 401 AD. After a slight temperature increase up to the beginning of the 8th century, the third cold period extended from the mid-8th to the 13th century, during which the Bosphorus and even parts of the Black Sea were repeatedly frozen, and floating ice masses entered the Sea of Marmara. Winters were markedly mild for 400 years starting from the mid-13th century. The fourth cold period began early in the mid-17th century and has lasted to the present day; it has been characterized by severe winters, however, the intensity of the winter cold has gradually diminished during this interval. Our aim is to evaluate existing historical data on these modern cold periods, to analyze instrumental meteorological data, and to provide suitable data for future correlations with the amplitude and frequency of paleoglacier advances both in the Alps and in Anatolia. These four periods are more or less coeval with the phases of glacial advance in the Northern Hemisphere. As the accuracy of the historical data increases with time, evidence is more detailed for the fourth period. During this interlude, which coincides with the Little Ice Age, freezing events were not all coeval with reliable central European evidence. This can be explained by the low index of the North Atlantic Oscillation that resulted in higher precipitation ratios.

Keywords: climate change, Istanbul, Little Ice Age, Anatolia, paleoclimate, Bosphorus freezing

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1. INTRODUCTION

Istanbul is a huge metropolis connecting continents, cultures, and religions, a home to approximately 15 million people (Figure 1). Its history begins with the foundation of Byzantion in Hellenistic times and continues through the Roman, Byzantine, and Ottoman periods to today. Most of the accounts dealing with the establishment of Istanbul focus on the year 2700 BP, though it is evident that people had already settled around Istanbul before this date. In the region, archaeological surveys have revealed the presence of continuous settlement, the earliest sites of which appeared 300,000 years ago.

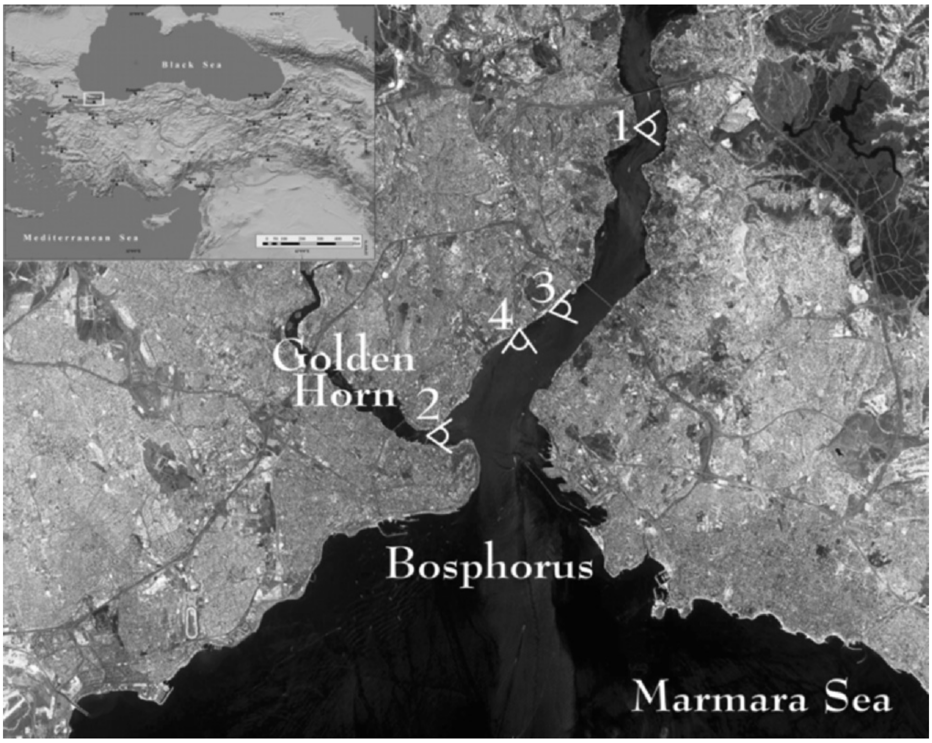

Figure 1. Satellite image of Istanbul with approximate orientations of the photographs in the following figures: Figure 2 = 1; Figure 3 = 2; Figure 4 = 4; Figure 5 = 3.

At around the Last Glacial Maximum, Anatolia (including Istanbul) experienced a substantial expansion of cold steppe vegetation at the expense of forest and woodland. Closed lakes were far more extensive than at present, which is best explained by a reduction in evaporation together with evapotranspiration losses, accompanied by higher catchment runoff coefficients (Kuzucuoğlu and Roberts 1997). Meanwhile, the Black Sea was a large lake to the north, and Marmara Sea was a lake without any connection to the Aegean and Black Seas. During the Late Glacial period, this region experienced several climatic oscillations until the present-day environment was established, approximately 3000 years ago (Özdoğan 1997).

Herodotus, the Greek historian of the 5th century BC, supplied important information about Istanbul, notably that there were originally two cities (2002). The former was Kalkhedon, situated on the Asian coast of the Bosphorus. The latter was Byzantion, located on the "Historical Peninsula" on the European side. Also, there was a third city, Selymbria, but this was situated on the northern Marmara coast, far away from the two cities that formed the nucleus of today's Istanbul. Herodotus stated that Kalkhedon was founded 17 years before Byzantion. The Greek geographer Strabo (1st century BC to 1st century AD) recorded similar data in his famous work, *Geography* (2000). Byzantion, a well-protected harbor city, was established in 663 BC by colonists from cities in the Peleponnese. Due to its position dominating the Bosphorus Strait, it grew rich in little time, becoming the key point for trade between the Black and Marmara Seas. It then started to control both the Bosphorus traffic and passages between Asia and Europe.

In this study, the aim is to evaluate existing historical data on the recorded cold periods, to analyze instrumental meteorological data, and to provide a suitable synthesis for further studies.

2. FREEZING EVENTS IN THE BOSPHORUS AND BLACK SEA REGION IN HISTORICAL RECORDS

Herodotus provided the first information about freezing in the Bosphorus. In addition to this event, which occurred during the Cimmerian historic time, Herodotus reported another freezing occurrence in the Kerch Strait of the Azov Sea. Subsequently, there is no indication of a freezing event in the Bosphorus until the time of Augustus. In *Epistulae Ex Ponto* (Letters from Black Sea) and *Tristia* (Songs of Madness), Ovid (1977) mentioned three successive freezing episodes in the Black Sea region between 7 and 17 AD.

Between the time of Ovid until the 4th century AD, there is no information on freezing of surface waters in the Bosphorus and Black Sea region, although freezing of several rivers, especially the Danube, was reported. The Roman historian, Ammianus Marcellinus (1808), reported not only the freezing of rivers that flow into Black Sea but also that parts of Black Sea open to northerly winds were covered by thick ice.

Scaliger, in his quotation from *Chronicon Pachale* (Paschal Chronicle), mentioned that the Black Sea was completely frozen in 401 AD, at the time of

Emperor Arcadius (395–408 AD) (Chikhachef 1864). Freezing was so intense that ice masses floated into the Marmara Sea continuously even a month into spring (March?). According to another source, winter in Istanbul was so severe in the same year that the Bosphorus and Golden Horn were completely frozen. The inhabitants of Istanbul believed that this was the "Wrath of God" (Pamukciyan 2002).

Around 600 AD, Jordanes (1882) mentioned that the Azov Sea was frozen but reported that the Don River was not affected by the cold event. After the cold spell documented by Scaliger, no freezing in the Bosphorus or Black Sea basin was reported for 300–350 years. On the other hand, the winters of 401, 418, and 441 AD (Schnurrer 1823–25) were harsh in Western Europe. Also, the winters of 660 AD–data from Theophanes (1997)–and 760 AD–data from Nicephorus, Patriarch of Constantinople (1990)–in Istanbul were dominated by extensive, abnormal cold, and the region was under snow for several months.

The Bosphorus was reportedly frozen again in 739 AD (Hammer-Purgstall 1827–35). Byzantine Emperor Leo III (717–741 AD) was against praying in front of icons, and ordered that the fresco of Jesus be removed from the gate of the Palace. During his reign, the winter of 739 AD was so violent that the Bosphorus was completely frozen. Inhabitants and priests related this extraordinary phenomenon to the struggle against the saints (Pamukciyan 2002).

At the time of the Emperor Konstantinos V Kopronymos, freezing events with harmful consequences were reported throughout his reign (741–775 AD). The first one occurred in 753 AD, during the winter of which, ice masses were dragged by the force of the elements from the Black Sea to the Marmara Sea (Pamukciyan 2002). This occurred again in 756 and 763 AD. Byzantine chronographers recorded these events in detail, and several historians made different observations, descriptions, and interpretations relating to this decade (753–763 AD). Later, at the beginning of 12th century, the historian Zanoras described the 8th century:

> In this interim (Emperor Kopronymos), there was a harsh winter. Not only rivers, also the Black Sea is frozen partially. Because of the frozen Bosphorus, sea passage from Byzantion to Chrysopolis (Üsküdar) on the Asian side is hardened. The inhabitants who could walk to Üsküdar, and loads could be carried with animals and two-wheeled ox-carts. It is known that similar phenomena occurred in other seas. With spring warming, the ice was broken into pieces and these pieces started to flow and were dragged by strong winds. Many domestic and wild animals were killed by the collision of island like ice masses. Some of these masses were dragged quickly and hit the city walls, with the intensity of the strike not only the walls but also buildings nearby were damaged (Indjidjian 2000).

The Byzantine chronographer Glykas (1836) described the same event. The freezing that occurred several times when Emperor Kopronymos was in power was also described by Theophanes the Confessor (1997), who lived in this time. Theophanes mentioned the freezing of a broad stretch of 100 miles along the coast of the Black Sea, first to a depth of 30 and later to 50 fathoms, and the occupation of the Bosphorus and Hellespontus by a thick ice cover:

> In the same year (762 AD), starting in early October, there was very bitter cold, not only in our land, but even more so to the east, the north, and the west, so that on the north coast of the Pontos to a distance of 100 miles the sea froze from the cold to a depth of thirty cubits. The same happened from Zigchia to the Danube, including the river Kouphis, the Danastris, the Danapris, and Nekropelai, and the rest of the coast as far as Mesembria and Medeia. All this ice was snowed upon and grew by another twenty cubits, so that the sea became indistinguishable from land: upon this ice wild men and tame animals could walk from the direction of Chazaria, Bulgaria, and other adjoining countries. In the month of February (763 AD) of the same 2nd indiction this ice was, by God's command, split up into many different mountain-like sections which were carried down by the force of the winds to Daphnousia and Hieron and, by way of the straits, reached the City and filled the whole coast as far as the Propontis, the islands, and Abydos. Of this I was myself an eyewitness, for I climbed on one of those [icebergs] and played on it together with some thirty boys of the same age. Some of my wild and tame animals also died. Anyone who so wished could walk without hindrance as on dry land from Sophianai to the City and from Chrysopolis to St Mamas and to Galata. One of the icebergs struck the jetty of the Acropolis and crushed it. Another huge one struck the wall and shook it greatly so that the houses on the inside partook of the quake. It then broke into three pieces and ringed the City from the Mangana to the Bosphorus, rising in height above the walls. All the inhabitants of the City, men, women, and children, ceaselessly watched these things and would return home with lamentation and tears, not knowing what to say. (Theophanes the Confessor 1997)

According to Byzantine historians, this cold period was one of the most significant events in history. Patriarch Nicephorus supplied important information on winter 763 AD in Istanbul:

> In the beginning of autumn (762 AD) winter has come with abnormal colds; all, also saline waters are frozen which affected inhabitants of the city severely. 100 miles stretch of the sea [1 Roman mile is approximately 1.5 km] is covered by ice like in the regions north of Black Sea. Ice invaded most of the rivers; the coasts of Mesembria and Medeia were a solid mass as ice was 30 coudée thick [old unit of length, being the distance from elbow to fingertips; 30 coudee = about 13–14 m]. Also snowfall was so heavy that this ice is enclosed by 20 coudée [about 9–10 m] of snow and all morphological differences between sea and coast disappeared. Now a white cover unified sea and land. All parts of Black Sea facing north were solidified [= frozen]. Especially the areas of Hazars and around the Scythian's Lands were inaccessible and unsuitable for human and animal life. After a while (according to Georgios Kedrenos in February) this significant crystal crust broke into several pieces and these were uplifted in the middle of the sea like Pyramids. Most of them, dragged by winds, were smashed and sunk in the opening of the Bosphorus to the Black Sea near Daphnusia, which was a powerful castle. Most of them entered into the Bosphorus. They filled up all the curls of the water way and connected Asia and Europe. They

> formed a land bridge between two continents and it was easier to pass the strait by walking instead of using boats. Accumulated ice masses in the Bosphorus without any delay were dragged into Propontis [Marmara Sea] and even reached Abydos. There they accumulated again in a perfect way to form a structure like a monolith and Propontis lost its sea characteristics. One of these huge icebergs was grounded in the bottom of Constantinopolis Fortress, and shook the city walls so that inhabitants were excited. Icebergs accumulated in front of the Fortress, then invaded all waterways. They accumulated to the same height as the city walls. As a result inhabitants of the city were able to go out of the city from the harbor by crossing these icebergs and they can walk to the Galata Castle on the other side from Constantinopolis Fortress (Chikhachef 1864).

There is no evidence of cold winters in Western Europe during this decade (753–763 AD), however, a period of intense cold after December, 770 AD, was mentioned. This cold period was extensively felt in almost all of Europe (Schnurrer 1823–25). In 787 AD, European authors reported an abnormal cold pulse in Italy (Schnurrer 1823–25) that went unmentioned in Byzance.

Freezing in the Black Sea occurred again at the start of 9th century. In 800 AD, it was frozen to a considerable depth (Schnurrer 1823–25). The winter of 801/802 was also noted in Europe for cold temperatures. As a whole, the 9th century is accepted as one of the coldest periods in Europe. In 820 AD, the Danube, Elbe, and Rhine Rivers were frozen for a month. Furthermore, the winters of 832, 855, 859, 864, 874, and 880 AD were so harsh, especially in 859, that the Adriatic Sea was frozen, and Italy was under snow for 100 days (Schnurrer 1823–25). Despite these climatic events, Byzantine chronographers documented no severe winters or freezing events for the Bosphorus.

Glykas and Symeon Logotheta revealed four freezing events in the following century. At the time of Emperor Romanus I Lecapenus (919–944 AD), winters were cold in 928 and 934 (Hammer-Purgstall 1827–35). In the time of Emperor Nicephorus II Focas (963–969 AD), temperatures fell again (data from Glykas). Also the winter of 993 AD, at the time of Basileus II (976–1025 AD), was reported as severe by Byzantine chronographers (data from Glykas). Leo Grammaticus, who lived in the time of Emperor Romanus, reported severe cold and stated that one of these periods lasted for 120 days, with severe frost occurring (Indjidjian 1794).

Europe experienced harsh winters in the 10th century. They occurred either before the two freezing events in the time of Emperor Romanus or between the reigns of Emperors Romanus and Nicephorus Focas, in the years of 912, 927, 940, and 943 AD. Freezing of the Bosphorus and the Black Sea was reported by several authors in 1011 during the period of Basileus II (Schnurrer 1823–25). A severe winter was also reported in Europe. Ice was floating on the Nile River (Michaud 1825–29).

After the severe winter 1011, there is no evidence for freezing of coastal waters in the Black Sea region for 210 years, with the exception of the freezing of the Kerch Strait that connects the Azov and the Black Seas. This period is known as the "Little Climatic Optimum," or the Medieval Warm Period (Telelis 2000). In 1221, during the time of Emperor Robert I (1219–1228), a very severe winter struck Byzance such that the Bosphorus and Golden Horn were completely frozen (Pamukciyan 2002). This recurred in 1232 (Hammer-Purgstall 1827–35).

Both Europe and the East witnessed several harsh winters, whereas in Istanbul and in the Black Sea, a period of 221 years passed without any freezing. The summer of 1043 was recognized in Europe as "A year with snowfall, which covered both the German farmers and their yield" (Chikhachef 1864). In the winter of 1076, Rennes was frozen, and in 1113, following a severe winter in France, there was a very hot and dry summer when dead trees caught fire under "the burning sun" (Chikhachef 1864). As a consequence of heavy snowfall in Baghdad in 1117, snow cover reached the height of a man and remained for 14 days (Hammer-Purgstall 1842–43). Several locations in Europe had to experience very low temperature during the winters of 1124, 1126, 1127, 1129, 1179, and 1210 (data from Chikhachef 1864).

In winter of 1232, after a very hot summer in 1231, the Bosphorus was frozen again, and this event does not coincide with severe winters in Europe. In 1234, the Adriatic Sea was also frozen as a result of a severe winter in Italy, and it was possible to walk across (Indjidjian 1794). During a 385-year period, the Adriatic Sea was frozen only once, yet this event was not duplicated in the Black Sea region. Similarly, the 1232 freezing event in the Bosphorus was not felt in the Adriatic Sea.

After the 1232 Bosphorus freezing, a 250-year warm interval ensued. Petrus Gyllius (1489–1555) cited (1561, 1562) two severe winters during his time, writing that even the sea surface from Kağıthane to Galata, on the harbor side of Istanbul, was frozen, so that boats could advance only by breaking the ice. Unfortunately, the author did not provide the years. These events probably took place between 1520 and 1550 (Hammer-Purgstall 1827–35).

According to Hammer-Purgstall (1827–35), thick ice covered the Bosphorus and connected the two continents in 1620, at the time of Sultan Osman II (1618–1622). Demetrius Cantemir, in his *History of the Ottoman Empire*, reported frost and freezing of the Bosphorus due to severe cold, and that the inhabitants of Istanbul were walking from Istanbul to Üsküdar in the winter of 1621 (Pamukciyan 2002). This event was mentioned by Hammer-Purgstall and Cantemir with one year difference (but we conclude that it is the same event). According to Nicephorus (1829–55), there was first a heavy snowfall and then dry cold so that the Bosphorus was frozen. People were able to walk from Galata to Üsküdar and from Üsküdar to Istanbul in January of 1621.

During the 339 years from 1232 to 1621, there were several harsh winters in Europe. Those of 1292, 1322, 1323, 1341, 1342, 1358, 1363, 1399, 1042, 1407, 1408, 1421, 1433, 1434, 1457, 1491, 1506, 1513, 1514, 1534, and 1607 were recognized especially as "difficult years with severe winters"

(Chikhachef 1864). Such severe winters were not experienced in the Bosphorus and Black Sea region. Only the winters of 1341, 1342, and 1343 brought unusual cold to Istanbul (Schnurrer 1823–25). In 1513, the Rhine and Danube Rivers were frozen when cold conditions dominated Europe, but Ottoman chronographers recorded no freezing events in the Bosphorus (Pamukciyan 2002).

But in 1669, the Bosphorus was again frozen, at least partially, with floating ice masses on the waterway (Schnurrer 1823–25). Between 1621 and 1669, severe cold periods from Europe and even from America are known. The cold winters of 1642, 1658, 1667, and 1669 were especially significant. Hailstones of 1 kg reportedly fell in Egypt on July 28, 1667. Until 1755, cold winters were more common in Europe, Asia, and America. The winters of 1670, 1673, 1695, 1697, 1705, 1709, and 1716 witnessed partial freezing along the coasts of Denmark. During the winter of 1729, the Danube froze over three times. The winter of 1740 affected all of Europe, while the winters of 1741, 1744, and 1750 were extreme in America (Hammer-Purgstall 1827–35). Even though there was no freezing reported for the Bosphorus, the 1750 winter was difficult in Istanbul.

In 1755, during the time of Sultan Osman III (1754–1757), when severe winters were absent, Turkish chronographers noticed an ice cover on the Bosphorus that enabled one to walk from Ortaköy to Sütlüce on February 16th (Pamukciyan 2002). In January or February of 1779, the Bosphorus was once again frozen due to cold weather conditions (Indjidjian 1794).

During the 68-year interval from 1779 to 1823, central and northern Europe experienced a series of long and very cold winters. Considering their continuity, duration, and intensity, it is very difficult to find a cold period of equal severity. Europe endured 20 harsh winters between 1668 and 1816, i.e., the winters of 1768, 1775, 1776, 1778, 1779, 1780, 1784, 1785, 1786, 1789, 1793, 1796, 1799, 1800, 1802, 1808, 1810, 1812, 1813, and 1816 (data from Chikhachef 1864). This represented an almost continuously cold period, with only two-year breaks. This period marks the maximum of the Little Ice Age in Europe (Grove 1990). Was the Little Ice Age a global or a regional European phenomenon? In this context, data from the Bosphorus region are crucial. Some of the significant parameters of this cold period are outlined below.

On January 27, 1776, temperature was measured at –28.7° C in Leipzig, –22.5° C in Montdidier, –21.5° C in Nancy, –20° C in Paris, and –21.5° C in Vienna (Schnurrer 1823–25). Nearly all the rivers of Europe were frozen from the beginning of December, 1784, to the end of February, 1785. On February 27, 1785, temperature was –29.2° C in Waldheim, Germany. Cold weather reached abnormally low levels in Paris, Naples, and several cities of Spain in the same month of that year (Schnurrer 1823–25). In most of Europe, temperatures were lower than –25° C in the winter of 1789 (Schnurrer 1823–25). From the end of summer, 1801, harsh, cold conditions dominated all of Europe. Temperatures were measured at –48° C in Bialystok, Poland, on January 1, 1802, –27.3° C in

St. Petersburg on March 12, –24° C in Yekaterinburg on May 11. The sea in Livorna was frozen the same year (Schnurrer 1823–25). England and Germany were under heavy snowfall in 1802; temperature was measured at –29.2° C in Stuttgart. After the severe winter of 1802, tropical weather conditions dominated the whole summer, and temperatures as high as 37.5° C were recorded in Vienna, for example (Schnurrer 1823–25). In 1808, a severe winter dominated southern Europe, especially Italy. On February 27, Naples was under heavy snowfall, and the Rhone River in southern France was frozen (Schnurrer 1823–25). The 1809 winter followed the same pattern, with temperatures of –9.7° C in Naples, –35° C in Lithuania, –42.2° C in St. Petersburg, and the mercury froze in Moscow (Schnurrer 1823–25). The winter of 1812 was the last in the series of severe winter weather, but with an "inverse" temperature pattern, e.g., the thermometer registered –6.2° C in Stuttgart and –11.2° C in Naples (Schnurrer 1823–25). Although the severe winters in Europe had an important geographical extension and were felt in Hungary and/or Thrace, which are relatively closer to the Black Sea basin, there is no evidence of freezing in the Bosphorus and Black Sea region in historical records for that time period.

The Black Sea and Bosphorus witnessed another severe winter in 1823, however. The northern coasts of the Black Sea froze completely, and ice masses were dragged into the Bosphorus. Also, the Golden Horn was partially frozen (Pamukciyan 2002). For this year, the cold event recorded in the Bosphorus and Black Sea seems to correlate with difficult conditions dominating Europe. For instance, temperatures were measured at –30° C in Hamburg between February 21 and 26. Mean temperature in Berlin was around –23° C (peak –35° C), in Bucharest –23° C to –25° C, –37° C in Sweden, and –11° C in St. Petersburg. Meanwhile, Spain and Portugal were under snow (Schnurrer 1823–25).

The Golden Horn froze on February 6, 1849 (Chikhachef 1864). From January 9 to 10, 1862, the Bosphorus was again frozen (Pamukciyan 2002). During this winter, temperatures dropped abruptly north of the Black Sea (–30° C in Moscow on December 9, 1862), whereas this winter was almost unfelt in France, England, and even Vienna.

When the Golden Horn froze in 1857, inhabitants could drive from Halıcıoğlu to Eyüp (from the Archives of the Kandilli Observatory). The Golden Horn and Bosphorus froze again in 1878 during the Russian-Ottoman Wars (Pamukciyan 2002). The same event occurred again in 1893 (from the Archives of the Kandilli Observatory).

The most severe winter after that time occurred in 1928. The Golden Horn was partially frozen, and icebergs were observed in the Bosphorus. Again in 1929, a part of the sea froze, and ice masses were dragged from the Black Sea and accumulated in the Bosphorus (Figures 1, 2, and 3).

The last freezing event in the Bosphorus was during the second harsh winter of the young Turkish Republic in 1954. Ice masses accumulated in the Bosphorus and prevented marine traffic for several days (Figures 1, 4, 5, and 6).

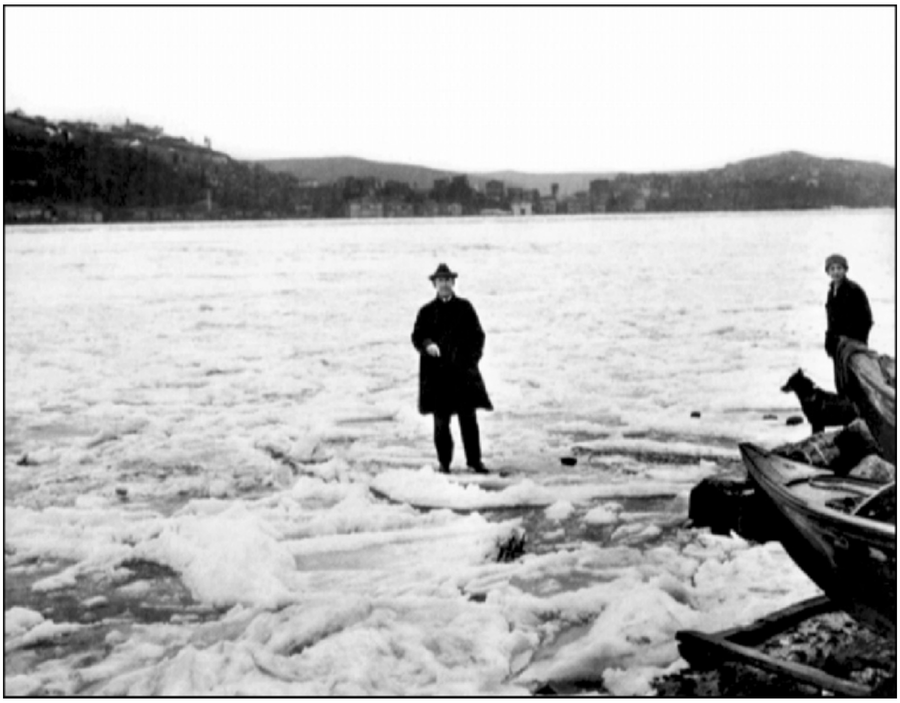

Figure 2. Partially frozen Bosphorus in the winter of 1929; position 1 in Figure 1 (from Üster 2000).

Figure 3. Frozen Golden Horn and Bosphorus in the winter of 1929; position 2 in Figure 1 (from Üster 2000).

Figure 4. Ice masses in the Bosphorus in the winter of 1954; position 4 in Figure 1 (from Üster 2000).

Figure 5. Ice masses accumulated in the Bosphorus in the winter of 1954; position 3 in Figure 1 (from Üster 2000).

3. DISCUSSION

According to the historical records, over the last 2000 years, there have been 40 severe winters in Istanbul accompanied by (1) freezing of the Bosphorus and the Golden Horn and (2) ice masses dragged in from the Black Sea. During these two millennia, 14 freezing events have been reported in the Black Sea only. The Bosphorus has been totally frozen 12 times and partially frozen 6 times. The Golden Horn has been totally frozen 15 times and partially frozen 6 times.

Most descriptions in the historical records show that freezing occurs, and

Figure 6. Iceberg floating on the Bosphorus in early spring of 1954 (from Üster 2000).

ice masses are dragged in from the Black Sea, due to strong easterly winds that blow during severe winters. For instance, the descriptions of Zanoras, Theophanes, and Nicephorus for the winters of 753, 756, and 763 mention cold and strong winds. Historians also recorded that dry, cold conditions subsequent to heavy snowfall brought about freezing in the Bosphorus during the 1621 winter. In order to freeze saline water and transport ice masses from the Black Sea, through the Bosphorus, and across the Marmara Sea as far as the Dardanelles, strong winds and cold, dry air masses are required. Within the general atmospheric circulation patterns affecting the Eastern Mediterranean region (Figure 7), such winds are driven by the strong thermal high pressure system that covers a large part of the Asian continent and affects the region substantially in winter. This Siberian High Pressure System causes dry and cold continental polar air masses to move in a southwesterly direction. It is logical to conclude that extension and intensification of the Siberian High occurred during winters when the Bosphorus froze.

The effect on these events of the North Atlantic Oscillation (NAO), which dominates atmospheric variability in the subtropical northern hemisphere during winter, is unclear (Figure 8). The longest precipitation record for Turkey is from Istanbul (Erinç and Bener 1961), but the relationship between total winter precipitation and freezing events remains, unfortunately, poorly understood. Causes of this inconsistency could be the absence of long-term meteorological measurements and inconsistencies in the historical data for older events.

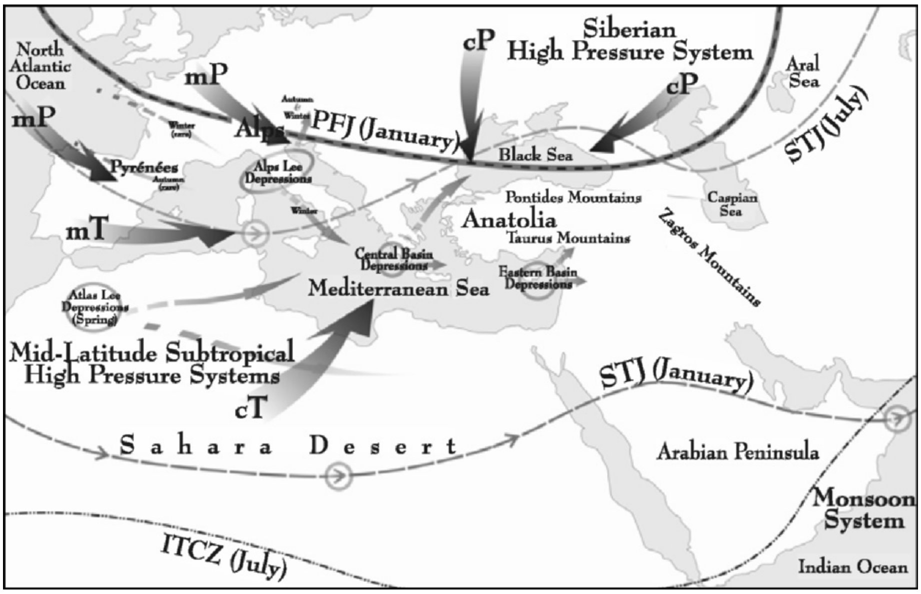

Figure 7. Mean positions of the Polar Front Jet (PFJ), Subtropical Jet (STJ), and Intertropical Convergence Zone (ITCZ) in winter and summer in the Mediterranean Region, and schematically located low pressure and high pressure systems that influence the climate of the Eastern Mediterranean: cP = Continental Polar Air Mass; mP = Marine Polar Air Mass; cT = Continental Tropical Air Mass; mT = Marine Tropical Air Mass (from Akçar and Schlüchter 2005).

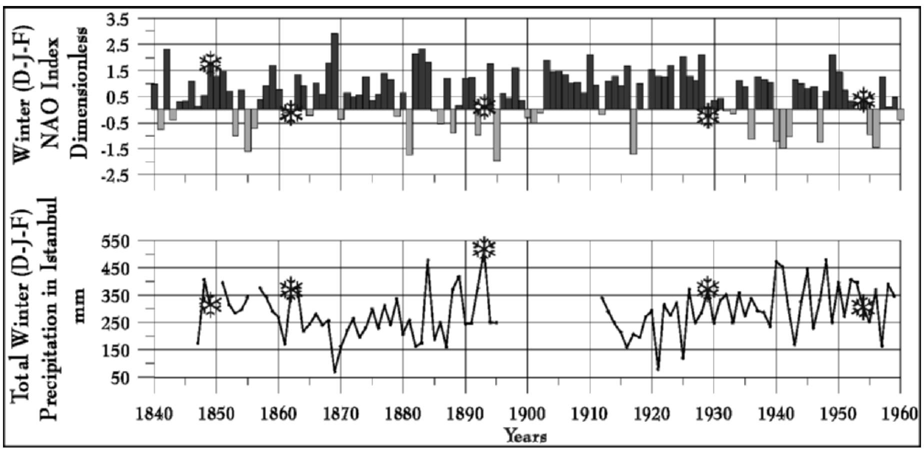

Figure 8. Total winter precipitation in Istanbul (data from Erinç and Bener 1961); Winter North Atlantic Oscillation Index (http://www.cru.uea.ac.uk/cru/data/nao.htm) and Freezing Events in Bosphorus between 1840 and 1960 (modified from Yavuz *et al.* 2003).

The historical data on freezing events in the Black Sea and Bosphorus show four main cold periods since 1 AD (Figure 9). The first period was around the 1st century. Although the temperature was close to or perhaps a little lower than at present, three successive freezing events were indicated on the Black Sea coast between 7–17 AD. The second cold period occurred in the 4th century, when parts of the Black Sea were frozen, and floating icebergs reached the Marmara Sea in 401 AD. The third cold period extended from the mid-8th to the 13th century, when the Bosphorus and parts of the Black Sea were repeatedly frozen, and floating icebergs were present in the Marmara Sea in the winters of 739, 753, 756, 763, 928, 934, 1011, 1221, and 1232. The fourth cold period began in the early mid-17th century and lasted to the present day. It has been characterized by severe winters (with freezing of the Bosphorus, the Golden Horn, and parts of the Black Sea in 1621, 1669, 1755, 1779, 1823, 1849, 1862, 1857, 1878, 1893, 1928, 1929, and 1954), however, the intensity of the winter cold has gradually diminished during this interval.

The four episodes are more or less contemporaneous with the phases of glacial advance in the Northern Hemisphere (Grove 1990). As the accuracy of the historical data increases with time, evidences for events are more frequent and more reliable for the fourth period. During this time interval that coincides with the Little Ice Age, freezing events are not always contemporaneous with the central European evidence.

5. CONCLUSIONS

Over the last 2000 years, 40 severe winters occurred, which were accompanied by freezing of the Bosphorus and the Golden Horn and by the dragging of ice masses from the Black Sea. Fourteen freezing events are reported for the Black Sea. The Bosphorus has been totally frozen 12 times and partially 6 times. The Golden Horn has been totally frozen over 15 times and partially 6 times (Figure 9).

The effect of the NAO Index on the precipitation pattern and on the freezing events in the Bosphorus is not clear. The climate of the region is strongly influenced by the Siberian High Pressure System during winter. The frozen Bosphorus during the most recent cold periods (from 1621 to 1954) can be explained by the interaction of the North Atlantic Oscillation and the Siberian High Pressure Systems (Figures 7 and 8).

The four cold episodes are more or less coeval with phases of glacial advance in the Northern Hemisphere (Grove 1990). As the accuracy of the historical data increases with time, evidence for freezing events is more frequent and reliable in the fourth period. At this time, which coincides with the Little Ice Age, freezing events were not always coeval with the central European evidence.

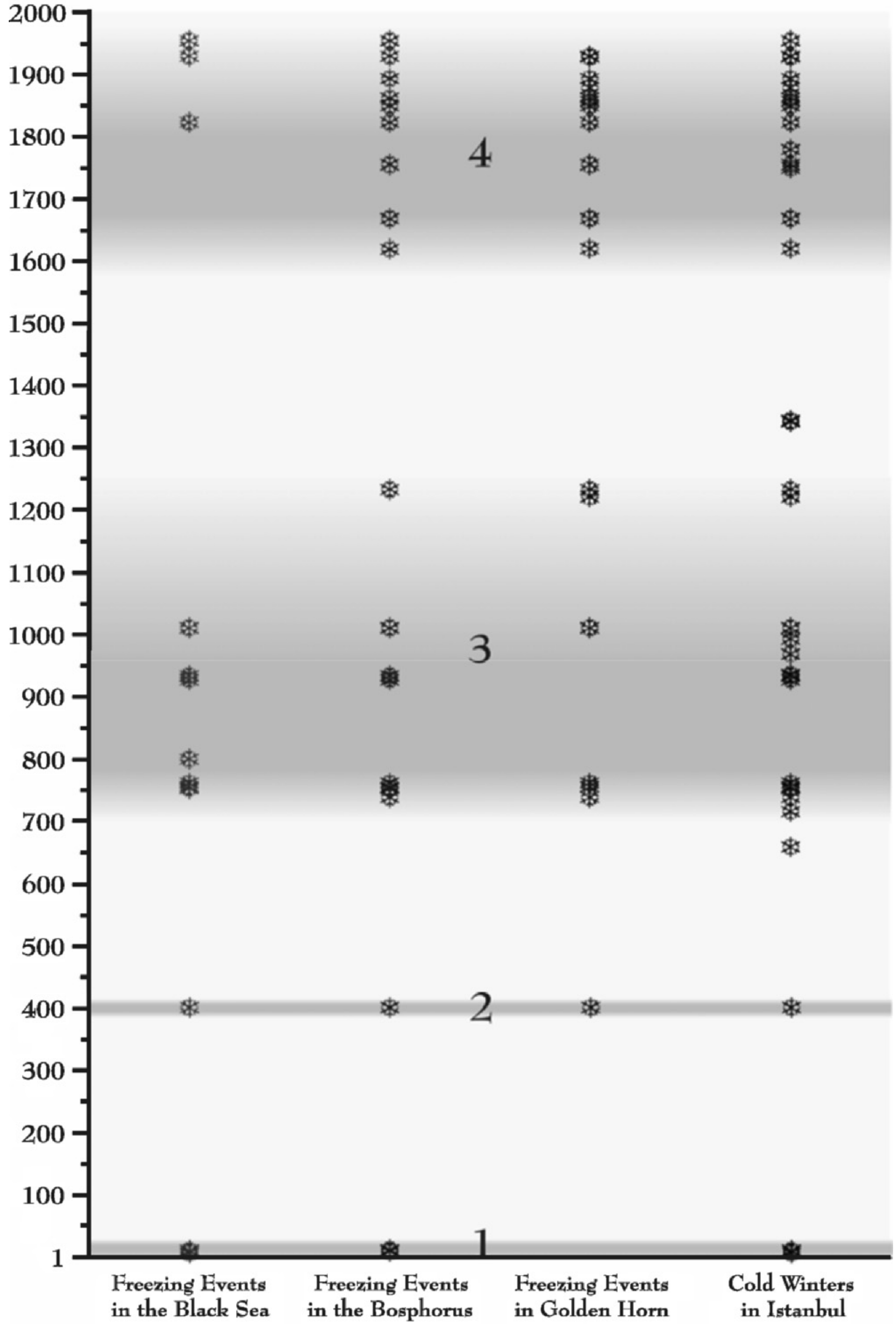

Figure 9. Distribution of historically recorded cold events since 1 AD with four main cold phases.

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COASTAL CHANGES OF THE BLACK SEA AND SEA OF MARMARA IN ARCHAEOLOGICAL PERSPECTIVE

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Abstract: In light of the controversial Black Sea flood hypothesis, this paper surveys the settlement history of Turkish coastal areas along the Black and Marmara Seas to determine if evidence exists for a disruption that might have been caused by catastrophic environmental factors. Paleolithic and Mesolithic sites are common along most of this coastline, but later sites of Neolithic through Bronze Age date are mysteriously lacking. No cultural changes can yet be ascribed to an abrupt inundation, but the evidence is difficult to interpret. Unfortunately, the archaeological record of the entire Circum-Pontic region is highly uneven, and existing data from different sectors are also incompatible. Each part of the Black Sea basin has its own distinctive cultural sequence, and each has evolved its own schools of archaeology that employ not only different methods and terminologies but also different ways of thinking. Despite the need, there have been no serious attempts to link these cultural areas together to form a regional synthesis. Thus, a comprehensive archaeological assessment embracing the entire Black Sea region stands out as a challenging but necessary task.

Keywords: Prehistoric archaeology, geoarchaeology, Turkey, Black Sea, Marmara Sea, cultural interaction.

1. INTRODUCTION

The cultural setting of the Circum-Pontic region attracted almost no interest until about a decade ago, when attention was suddenly drawn to it by the disclosure of a new hypothesis suggesting that a cataclysmic flood had filled the Black Sea basin relatively recently (*ca.* 8400 BP). This suggestion, developed

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by natural scientists based principally on marine geological evidence, was published together with speculations on the possible implications of such an inundation for cultural history. These implications included sensational issues, such as Noah's Flood, the origin of the Sumerians, and the beginnings of agriculture and civilization. This Flood Hypothesis was widely publicized and discussed in the general media, but ironically, the essence of the proposal, i.e., its geological aspect, was almost forgotten. The focus of reporting was skewed toward the presumed cultural consequences of the catastrophe, and, again ironically, archaeologists tended to ignore these scenarios as fabrications deserving no serious consideration because of their spectacular nature.

The Black Sea has nevertheless continued to fascinate the public, but on a scholarly level, the intrigue it is generating has provided a unique opportunity to bring together a broad spectrum of specialists, not only archaeologists but also natural scientists, whose curiosity about the region has been piqued. This paper is part of that multidisciplinary effort. It will survey the archaeological evidence along the southern coast of the Black Sea in search of cultural discontinuities that might have been caused by abrupt environmental changes.

1.1 A Prelude To Reaching A Mutual Understanding

Archaeology is a social science, and like other social sciences, there is little uniformity in its axioms and principles. Practices and theoretical concepts vary considerably according to different schools of thinking, which have their roots in the historical development of the science. The subject matter of archaeology is the cultural formations established by human beings in the past, however, as the components that make up a culture are extremely variable and in some cases unpredictable, so it is that universal rules or definitions are extraordinarily difficult to formulate. Thus, archaeologists think in a mode different from that of natural scientists, who are more accustomed to directly testable hypotheses and greater regularity in the behavior of their subject matter.

Nevertheless, archaeologists and natural scientists have frequently conducted cooperative research, and by the second half of the 20th century, collaboration became more institutionalized, yielding a new generation of disciplines under the overall label of "archaeological science." An array of specializations has since emerged in which experts, originating either from archaeology or natural science, have developed sufficient interdisciplinary knowledge to apply a wide range of investigative techniques to the solution of archaeological problems. The number of scientists who have devoted time and energy to broadening their expertise across disciplines has not been great, however, and there has been a tendency toward overspecialization.

Geoarchaeology is an apt example. The facile symbiosis between archaeology and geography that characterized the early investigations of Pumpelly

(Pumpelly 1908; Champlin 1994) in Central Asia at the beginning of the 20th century had already become complexly interdisciplinary with the work of Butzer in the late 1950s (Butzer 1958). Over the past few decades, however, geoarchaeology has been so overspecialized that natural scientists who moved in geoarchaeological directions grew somewhat isolated from their original discipline. These specialists have made significant contributions to archaeology by elucidating the relationships between ancient culture and habitat, but diminishing dialogue between natural scientists and geoarchaeologists has regrettably been a negative consequence. Due to this widening gap, new and indispensable discoveries in, for example, geomorphology, paleoclimatology, and marine geology, tend to go unnoticed by archaeologists, while research conducted by geoarchaeologists remains largely inaccessible to natural scientists because it is usually published in archaeological journals and reports.

The new hypothesis of an abrupt transgression of the Black Sea has opened up the possibility for greater collaborative research across disciplines. William Ryan and Walter Pitman (1998), the natural scientists who proposed the hypothesis, reached beyond their expertise to consider the remarkable consequences such an event would have imposed on ancient cultures living along the paleocoastline. With little detailed knowledge of the archaeological record, they conjectured that a disaster of this magnitude might be tied to the legend of Noah's Flood. They further suggested that it might have led to massive population displacement, ultimately giving way to the beginning of civilization. The historical part of this hypothesis appeared so unfounded to the archaeological and geoarchaeological communities, however, that it was dismissed by most. Only a single archaeological team that had already become involved in the matter continued to entertain such possibilities. Media coverage of the hypothesis soon led to neglect of the geological research in favor of the spectacular cultural repercussions, luring the popular imagination toward issues such as the biblical flood story, the impact of cataclysmic events on collective memory, the beginnings and spread of agriculture, and massive migrations.

This surge of publicity had some favorable results because it eventually convinced many scientists to look at the Black Sea basin more closely. Even at this early stage, the stimulus has produced major contributions to our knowledge together with a wider audience for news of such discoveries.

The marine geological findings are still poorly understood by specialists outside the earth sciences, namely archaeologists and historians. Accordingly, this new initiative to bring together experts from distant fields of natural and social science to share their evidence on the Black Sea is most welcome. As few archaeologists have taken an active part in this debate, and their expertise is usually geographically confined, this paper will begin with a summary of archaeological research in the Circum-Pontus and a short overview of the fundamental character of each area for the non-specialists.

2. CULTURAL SETTING OF THE BLACK SEA BASIN

The Black Sea is one of the largest land-locked marine bodies in the world. It constitutes a single and fairly uniform geographical entity, but from a human perspective, it is an extensive open space that has segregated, or even isolated, territories from one another. Each of the surrounding areas has a distinctive environment and cultural tradition, so much so that it is not possible to speak of a collective "Black Sea Culture." There are, in fact, five separate sectors with three buffer zones lying between them (Figure 1).

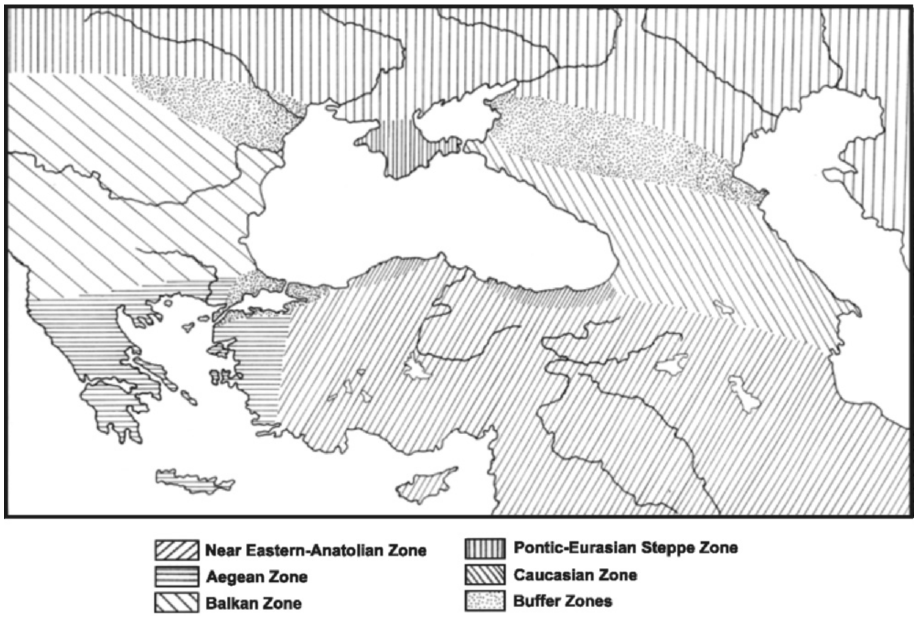

Figure 1. Cultural areas surrounding the Black Sea basin.

The sectors are discrete not only from the cultural and ecological perspective, but each local archaeological community has developed its own approach to describing its antiquity. Distinct schools of archaeological thinking have developed over the last two centuries, fostering a great deal of area specialization and little, if any, interest in other parts of the region. Isolated from each other, the individual sectors have characterized their ancient cultures in their own way, and though the terminology is everywhere the same, the terms have different implications from area to area. Correlating different cultures has become tricky business, making regional interpretation of cultural change very challenging.

The temporal and cultural differences between areas and the scope of the terminological problem can be appreciated by a simple comparison (Table 1). In the Near East and Anatolia, the Upper Paleolithic period (the latest subdivision of the Old Stone Age) was followed by two transitional stages known as the Epipaleolithic and Proto-Neolithic, while the subsequent Neolithic period comprises two distinct cultural stages labeled the Pre-Pottery Neolithic and the Pottery Neolithic. In the Near East, the transition from mobile hunting and gathering to sedentary village life dependent on farming took place during the late Epipaleolithic, and by the Pre-Pottery Neolithic, the appearance of large villages based on mixed farming and herding demonstrates a significant level of cultural complexity. In the Eurasian steppes, however, it is the Upper Paleolithic period that marks a high level of cultural complexity, as the entire territory was occupied by semi-sedentary hunters and fishers at the time. A very long-lived Mesolithic interval follows, while the subsequent Neolithic period is represented by a mobile, nomadic lifestyle dependent on the herding of animals. The terms used to describe the sequential stages of cultural development have a totally different meaning depending on the area.

The body of archaeological knowledge in any of these areas has grown to such a level that it would be very difficult–if not impossible–to master the cultural history everywhere. In earlier years, when far fewer data had been assembled within a far less voluminous archaeological literature, regional overviews were possible. At present, it would be hard to envision any single archaeologist attempting such a task.

Archaeological evidence is surely pertinent to the solution of the problems raised by the proposed early Holocene flood scenario in the Black Sea basin. Archaeological findings are usually viewed as the consequence of adaptation to environmental conditions, but here, they also represent a potential testing ground that can help verify terrestrial effects of the presumed marine events. In other words, correlating the archaeological evidence with the results of natural science investigations may play a decisive role in confirming hypotheses. Before attempting such a step, however, archaeology must resolve its own problems by synchronizing the cultural sequences that have been established independently around the Pontic basin.

The following sections provide brief outlines of the major cultural zones that surround the Black Sea.

2.1 The Eurasian or Pontic Steppe

The steppes, one of the world's largest mono-ecological zones, extends from Mongolia in northeast Asia to the Carpathian basin in Central Europe. It is characterized by a severe continental climate. The Ural Mountains form the most prominent feature in the generally monotonous topography of undulating lowlands, broken occasionally by large south-flowing rivers. The homogeneity of this natural setting is reflected in its cultural history. Socioeconomic systems and other aspects of historically-known lifeways, as well as the general character of the archaeological remains, are uniform throughout the Eurasian steppes.

In most places, the Upper Paleolithic and Mesolithic cultures are much better documented and understood than those of later periods. The Neolithic began considerably later than its appearance in the Near East. Recognized by simple, coarse pottery and domestic sheep and goats, it must have been introduced from elsewhere, as there is no evidence for indigenous development. The “Neolithic way of life” in the Eurasian steppes differed from that of other parts of the Old World. It refers to a nomadic routine dependent on herding, an economic mainstay that survived into the Medieval period. Thus, there is no association of the term with the presence of permanent settlements. Most of the archaeological evidence has been derived from burial sites, commonly referred to as *kurgans*. Even in cases when settlements have been recovered, they are usually one-period occupations. As the pace of cultural change was very slow, such single component sites are not very helpful in establishing a cultural sequence. Thus, chronological reconstruction depends heavily on “horizontal stratigraphy”[1] and radiometric dating.

Through most of prehistory and to some degree during historic times, interaction between the steppes and other areas was extremely limited. It was usually related to the trading of precious metals and metal artifacts. Only at times of crisis, usually triggered by climatic changes, was there movement of people out of this region, e.g., the “steppe invasions” or “Kurgan migrations.”

Within this vast territory, the Crimean Peninsula stands out as a distinct micro-region, occasionally displaying cultural manifestations related to the rest of the steppe regions but still somewhat different from them (Figure 1, denser hatching).

2.2 Moldavia

The northwestern corner of the Black Sea served as a cultural buffer zone between the Balkans and the steppes. Centered on the catchment of the Dniester River, the Moldavian area extends outward, encompassing the territory between the Bug and Danube Rivers. Depending upon the strength of influences from east or west, it was incorporated within either of the adjacent regions as a contact zone for the transfer of commodities and technologies. At times of crisis in the steppes, it became the Balkan frontier and received the primary impact of the steppe invasions. Settlement sites are well evidenced, at least for certain periods, however, such sites rarely yield a continuous sequence of occupations.

The chronological and cultural terminology employed for Moldavia are mostly in concordance with those of the Balkans.

Table 1: Simplified chart of the chronological terminology used in different cultural areas around the Black Sea basin.

years calBC	Near East-Anatolia	Thrace-Marmara	The Aegean	The Balkans	Moldavia	The Steppes	Caucasus
0	Iron Age	Iron Age	Iron Age	Iron Age	Iron Age	Iron Age	Iron Age Late Bronze Age Middle Bronze
1000	Late Bronze Age Middle Bronze Age	Late Bronze Age	Late Bronze Age Middle Bronze Age	Late Bronze Age	Bronze Age	Bronze Age	Early Bronze Age
2000	Early Bronze Age	Early Bronze Age	Early Bronze Age	Early Bronze Age			
3000	Late Chalcolithic Middle Chalcolithic	? Late Neolithic	Chalcolithic	Eneolithic Late Neolithic	Eneolithic	Eneolithic	Chalcolithic
4000		Middle Neolithic Early Neolithic	Late Neolithic Middle Neolithic	Middle Neolithic Early Neolithic	Late Neolithic		
5000	Early Chalcolithic Late Pottery Neolithic		Early Neolithic		Early Neolithic	Neolithic	Neolithic
6000	Early Pottery Neolithic	Aceramic ?	Aceramic ?	Mesolithic	Mesolithic	Mesolithic	Mesolithic
7000	PPNC PPNB	Mesolithic	?				
8000	PPNA						
9000							
10,000	Proto-Neolithic		Mesolithic				
11,000							
12,000	Epipaleolithic	?	Upper Paleolithic	Upper Paleolithic?	Upper Paleolithic	Upper Paleolithic	Upper Paleolithic

2.3 The Balkans

The large peninsula known as the Balkans, or at times as Southeastern Europe, comprises different geographical units, yet it still displays significant cultural uniformity. As the Balkan peninsula is presently divided into numerous national states, the cultural uniformity of the past is not reflected in the archaeological terminology. Often, the same culture appears under different names in each of the Balkan countries, leading to considerable confusion.

Ancient events in the Balkans were influenced by their proximity to the Aegean and Anatolia, two major centers of cultural activity. The close interaction is most evident during the Neolithic, which is often referred to as the "Anatolian-Balkan Culture Complex." In the Balkans, however, chronological implications of terms such as Neolithic or Chalcolithic are not the same as they are in the Aegean or in Anatolia, and these differences in the naming of cultural stages have a historical basis. The Early Neolithic of the Balkans, also known as the Karanovo I–Starčevo–Criş stage, represents the beginning of sedentary life in the Balkans. Following Near Eastern practice, the first settled villages were designated Early Neolithic, but only subsequently was it discovered that sedentarization emerged later in the Balkans than it did in the Near East. The first Balkan villages of the Early Neolithic turned out to be coeval with the Early Chalcolithic in Anatolia.

Numerous multi-layered mound sites exist in the Balkans, and therefore, the relative sequence of cultures is securely established. There are, nevertheless, lacunae in the stratigraphic record. The later stages of the Upper Paleolithic and subsequent Mesolithic period are poorly understood, while archaeological evidence from the 2nd millennium BC is likewise problematic.

2.4 The Marmara

The Marmara area is located between the Black Sea, the Balkans, the Aegean, and Anatolia. Its two coastal belts–one on the European side and the other on the Asian side–are separated by the Sea of Marmara and the two straits, namely the Bosphorus at the north giving access to the Black Sea and the Dardanelles to the west opening to the Aegean. Although it is small and looks very much like a buffer zone, the Marmara is of prime importance in understanding the interaction between Europe and Anatolia. It is also located strategically at the bottleneck in the main maritime route connecting the Aegean with the Circum-Pontic sphere. Any changes in marine conditions, either in the Black or Aegean Seas, have consequences in the Sea of Marmara, and thus, archaeological evidence between the two straits may be of significant value in verifying interpretations of findings beyond.

It is relatively easy to cross the Sea of Marmara and the straits, even by primitive seafaring; these water bodies do not constitute real barriers to transit between Europe and Asia. Any break in relations between the northern and southern sides of the Marmara must, it would seem, implicate to some extent cultural events that involve its role in regional communication. Whether the Marmara served as cultural bridge or barrier, the more insurmountable obstacles were probably not its natural ones.

The cultural terminology in current use in the Marmara area is confusing, as there is no consensus on a uniform standard. Depending upon the research objective, the terms applied will follow either Balkan, Anatolian, or Aegean usage. For example, a site dating to 5500 BC in Eastern Thrace will likely be categorized as Early Neolithic according to the cultural sequence established for the Balkans, but a coeval context on the Anatolian side would be called Late Neolithic/Early Chalcolithic, while along the Aegean littoral, a similar site might be assigned to the Middle Neolithic.

2.5 The Aegean

Even though the Aegean area is beyond the Black Sea basin, it must still be considered an integral part due to its important maritime connections. The archaeological sequence of the Aegean, which includes mainland Greece, the Aegean islands, and western Anatolia, is well-attested, at least in its basics, through numerous multi-layered habitation sites. Moreover, there are strong parallels between the Aegean and the Near East, providing links with the traditional, historical chronology.

Any change in the sea level of the Black Sea should be reflected in the Aegean, on condition that the connecting straits of the Dardanelles and Bosphorus are open. Numerous paleoenvironmental studies have been conducted in the Aegean, yielding ample evidence for Holocene sea-level changes and climatic fluctuations.

2.6 Anatolia and the Near East

The Anatolian peninsula is an extension of the Near East. The connection is not only a geographic one, but through a considerable part of cultural history, both regions have developed together. Likewise, Anatolian archaeology has evolved under the strong influence of the Near East, and in contrast to the confusion that always exists with cultural designations of the Aegean and Southeastern Europe, similar terminology is used in both Anatolia and the Near East.

The Black Sea lies just to the north of Anatolia, the long coastline dominated topographically by high mountain ranges that run parallel to the shoreline, leaving no definable coastal plain (Figure 1, denser hatching). This rough terrain is broken by the estuaries of the Kızılırmak and Yeşilırmak Rivers, which terminate in small deltaic plains. With the exception of these plains, almost nothing is known of the prehistoric and protohistoric cultures of the entire Black Sea littoral.

2.7 The Caucasus

The territory known as Caucasia comprises two discrete geographical units: Greater Caucasus in the north and Transcaucasia in the south. The distinction is not only geographic, but also cultural, as the archaeological evidence varies considerably from one part to the other.

Transcaucasia lies at the northern periphery of the Near East and is closely related to both the East Anatolian and Iranian plateaus. The area forms a buffer zone between the steppe cultures to the north and the Near Eastern civilizations to the south. Transcaucasia, and especially the Kura-Araxes River drainage, contains numerous settlement sites, but none preserves a lengthy occupation record, making chronological control and the establishment of cultural sequences difficult. Cross-cultural seriation is possible only when imports from the Near East are found. Transcaucasia and the Anatolian highlands merge together with no distinct boundary. On occasion, steppe cultures have crossed the Caucasus and expanded seamlessly into the Anatolian plateau via this route.

The northern foothills of the Greater Caucasus form an ecological microniche along the Kuban River valley. The area seems to have been a refuge for groups from the steppes. Most of the archaeological evidence derives from burials, not settlement sites, and thus chronological control is weak and problematic. A relative chronology has been constructed on the basis of imported high-status objects recovered in kurgan graves.

3. ARCHAEOLOGICAL EVIDENCE: A CONSPECTUS OF THE DATA FROM THE TURKISH COASTAL STRIP

Considering the wealth of evidence, the diversity of cultural groups, and the long time span, this paper cannot present an archaeological overview of the vast Anatolian territory adjacent to the Black Sea. Such an effort would yield only a simplistic generalization of a complex story.[2] It must suffice to underscore some of the critical points relevant to the main theme of this volume.

Working out cultural sequences for the Circum-Pontic sphere has not been easy. Due to problems rooted in the history of research in each area, such as lack of study or biases in designing research strategies, the available data do not allow one to draw a comprehensible picture. The Turkish sector of the Black Sea coastal strip is no exception, but, here, the lacunae in knowledge are due mainly to the lack of coastal sites, not the deficiency of research. The data are so limited and scattered–both geographically and chronologically–that how they should be interpreted seems an issue of secondary importance.

Despite the paucity of evidence from the littoral areas, the interior of Anatolia is a rich source of information. A general understanding of what was happening can be obtained because the Anatolian peninsula was closely tied to the traditional centers of the Near East, and numerous stratified mound sites yielding concrete archaeological sequences as well as occasional finds of ancient written documents have left Anatolian culture history more firmly established than in any of the other sectors around the Black Sea.

In examining the Anatolian evidence, the essential first question to ask is why there are so few coastal sites.

3.1 The State of the Evidence: Paucity of Coastal Sites

The pattern of archaeological resources along the northern Turkish coast is a strange one: while earlier prehistoric sites of the Paleolithic and Mesolithic periods are relatively well represented throughout most of the coastal strip, the only attested sites of the Neolithic, Chalcolithic, and Bronze Ages occur on the deltaic plains of the Kızılırmak and Yeşilırmak Rivers. This odd distribution is the opposite of that found in most other areas, where Paleolithic and Mesolithic sites are usually rare or absent while Chalcolithic and especially Bronze Age sites are extremely common. Farther inland, both in Central Anatolia and Thrace, hundreds of Chalcolithic and Bronze Age sites can be found, but finds datable to the Paleolithic or Mesolithic periods are few in number and in many places, they have never been identified.

A reasonable explanation is still elusive for the lack of settlement sites along the Black Sea coastline of Turkey between the Mesolithic and Greek Colonization periods. Despite the presence of local museums in the area and the fruitless attempts of many archaeologists, nothing has turned up that can be firmly dated to the Neolithic or to the Bronze Age.[3] The paucity of discovered sites cannot be blamed on a paucity of research, and neither can the mountainous coastline and other topographic obstacles be held responsible, as numerous Greek colonial sites are located throughout the coastal strip.

A more detailed examination of the archaeological evidence from the various periods is presented in the following sections. A map showing the location of sites mentioned in the text appears in Figure 2.

3.2 The Evidence of Early Prehistory (Paleolithic to Mesolithic Periods)

Wherever intensive surface surveys have been conducted along the Black Sea coast, abundant Paleolithic and Mesolithic evidence has been discovered. Lithic scatters from these cultural horizons are found on the first coastal terraces, and they extend inland within the valley systems. None of these sites has been excavated, so it is not possible to say whether any represent *in situ* campsites or whether they are all secondarily deposited. It is undeniable that the frequency of discovery indicates a high concentration of activity.

Figure 2. Archaeological sites mentioned in the text.

The area around the Bosphorus has been studied most intensively (Gatsov and Özdoğan 1994; Runnels and Özdoğan 2003). Lithic scatters or debris concentrations have been recovered most often within the fossilized coastal dunes, but they are not found here exclusively. Locations containing lithic assemblages typologically datable to the Early and Middle Paleolithic periods are known throughout the coastal area. The geological setting of these sites has not yet been worked out, however, and therefore, they do not yet provide a reliable basis for discussion here. On the other hand, some early Upper Paleolithic sites with Early Aurignacian assemblages have been discovered, the most prolific of which is situated within the coastal dunes of Ağaçlı, about 30 km west of the Bosphorus. There is no clear evidence of sites inhabited during the later stages of the Upper Paleolithic, but the dunes were re-occupied subsequently by Mesolithic or Epipaleolithic communities.

The fact that such assemblages extend along the coastal strip on sand dunes implies that the level of the Black Sea must have been close to its present state during both the Early Aurignacian as well as the Mesolithic periods. Due to the lack of excavated sites, no absolute dates have been obtained for these assemblages. Typological cross-dating with other regions suggests that the Early Aurignacian must be earlier than 25,000 BP, and that the Mesolithic probably spans the time between 14,000 and 9000 BP. It is not possible to say without further research whether the lack of late Upper Paleolithic sites is due to cultural reasons or to changes in the level of the Black Sea, but it can be concluded that during the late Mesolithic (10th to 9th millennia BP), the coastline must have been similar to that of the present, and the Black Sea must have provided a favorable littoral environment attractive to communities dependent on fishing and mollusc collecting.

3.3 The Evidence of Later Prehistory (Neolithic to Bronze Ages)

To date, only the fringes of the two deltaic plains of the Kızılırmak and Yeşilırmak Rivers have revealed settlement sites occupied continuously from late prehistoric times to the Greek Colonization period. Archaeological material found in these sites clearly indicates intensive interaction with the Anatolian interior. Considering that Anatolians understood and employed seafaring since the early Neolithic, the absence of human habitation for several millennia along most of the Black Sea coast is puzzling.

There are two possible explanations: (1) the prehistoric communities of the Anatolian plateau had no interest in maritime opportunities within the Black Sea region and therefore did not settle the littoral areas, or (2) there are, indeed, coastal sites, but, for some reason, they have not yet been found. The first explanation seems unlikely, as there are numerous elements in the cultural assemblages of Central Anatolia that possess common features with those of the Balkans.[4] This view is supported by the excavations at İkıztepe, Dündartepe, and Kocagöz Höyük, sites located on the deltaic plains noted above. For an overall assessment of these sites, with extensive bibliographic references, see Yakar (1991).

Archaeological evidence around the Sea of Marmara (Figure 3) provides further insight into the problems of coastal site distribution. Detailed assessments of the area have already been extensively published (Özdoğan 1997, 1999, 2003), and some of the critical issues are discussed here, organized for clarity according to the principal geographicalnits: the Dardanelles, the Sea of Marmara, and the Bosphorus.

3.3.1 The Dardanelles

Ample evidence of settlement sites dating from the Neolithic (7th millennium BC) to the present is found along the coasts of the Dardanelles. There is no indication in the archaeological record of an occupation break or irregularity that would imply disruption by external factors.

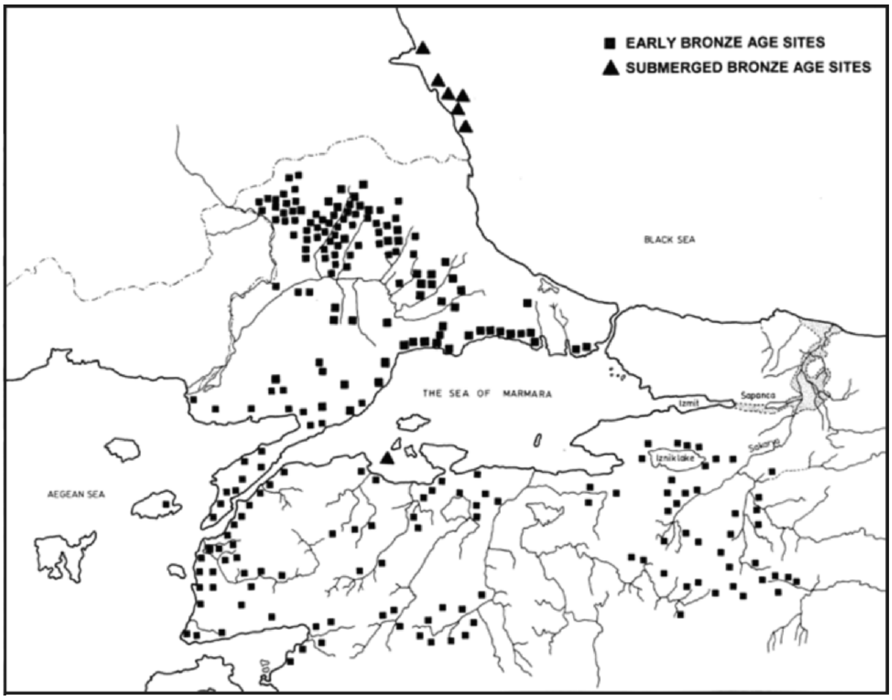

Figure 3. Distribution of Early Bronze Age sites in the Marmara area.

3.3.2 The Sea of Marmara

Along the northern, European coast of the Sea of Marmara, archaeological site distribution matches that of the Dardanelles. Beginning at the strait, the archaeological record is more or less complete all the way up to the Bosphorus at Istanbul. It is a different story in the southeastern section, that is, along the Anatolian coastal strip running from the Kapıdağ peninsula to the entrance of the Bosphorus. Here, no coastal sites dating between the 6th millennium BC and the 8th century BC have been found. Considering the concentration of ancient settlements, especially of the 3rd and 2nd millennia BC, situated only a little farther inland, the absence of littoral habitation is again very puzzling. The fact that this sector overlies the tectonically active "North Anatolian Fault" suggests that coastal sites might have been submerged by land movement and not a relative rise in sea level. The presence of a submerged mound near the island of Avşa (Özdoğan 2003) seems to support this view.

3.3.3 The Bosphorus

Throughout the Bosphorus, there are no sites earlier than the 8th century BC. It might be argued that the presence of a large metropolis like Istanbul would have wiped out prehistoric sites, yet, in spite of all efforts, not even a single artifact predating the 8th century has been recovered. Reason demands that there should be early historic occupation here. During the 2nd millennium BC, in particular, the considerable rivalry between the Hittites and Mycenaeans, who sought control of maritime routes, is plainly demonstrated at Troy on the Dardanelles. It is almost unthinkable that neither the Hittites nor the Mycenaeans had installed strategic strongholds near the mouth of the Black Sea. It is also unlikely that all sites dating to the periods in question have been eroded because the Bosphorus terraces have revealed well preserved Paleolithic surfaces. Whether the absence of later sites is due to tectonics or other natural phenomena cannot be determined at this time, but these site distributions, like others of coastal Turkey, still need an explanation.

Returning to the Black Sea, the evidence along the Bulgarian littoral also reveals no prehistoric sites located directly on the coastal strip. There is, however, substantial evidence of submerged sites in the Black Sea and under the lagoons (Draganov 1995), and all are datable to the time periods that are missing along the Anatolian coast and in the Bosphorus. One is led to surmise that there might have been appreciable change in the early Holocene coastline of the Black Sea. Possibly, the Black Sea was lower than the present day, at least between the 4th and late 2nd millennia BC. If this assessment were correct, then one would expect to see similar conditions in the Marmara and Aegean Seas, with the implication that the straits were open, but according to present knowledge, there is no such indication, neither in the Marmara nor in the Aegean. Presenting further hypothetical resolutions of this problem exceeds the expertise of an archaeologist. Somehow, archaeology and natural science must find a way to settle the matter.

4. CONTINGENCY OR CONTINUITY IN CULTURE

In this section, an effort will be made to sort out the indications of “irregularities” in the archaeological record of the southern Black Sea littoral–if there truly are any aside from the long habitation breaks presented above–that suggest an abrupt interruption in the occupation sequence that would be difficult to justify solely in terms of culture change.

4.1 The Neolithic Dispersal

Incipient food production is evident by 13,000 BP in the core area of "Neolithization" within the Near East. This beginning entailed more than a simple shift in subsistence patterns but also involved an almost complete transformation of late glacial hunting and collecting practices. Reflections of this new way of life can be seen in the introduction of new tools and technologies, which grew slowly through several formative millennia until reaching maturity at about 9000 BP, at which time it rapidly expanded outside its core region. Though many aspects of this complex picture are still debated, the basic characteristics and routes of expansion have been clarified through decades of excavation and interpretation of recovered evidence from hundreds of sites. There is no space left for other speculative scenarios.

Some suggestions have linked the spread of Neolithic economies with the proposed abrupt changes in sea level associated with the flood hypothesis. From the archaeological point of view, this is not possible. The littoral areas of the Black Sea never played an active role in the transmission of food production from the central Anatolian plateau to the Balkans. On the contrary, the initial movement of farming populations from Anatolia westward to the Balkans went through the Aegean, carrying with it the characteristic elements of the "Neolithic package": polished axes and adzes, pressure flaked projectile points, stone bracelets, ear studs, steatopygous figurines, stone and pottery vessels, as well as domestic sheep, goats, cattle, wheat, barley, and legumes. These introduced elements have no predecessors in the Balkans, and thus, it is a simple matter to trace their origins. The Black Sea littoral could not have participated in the movement of Neolithic culture out of Anatolia because settled village farms appeared in the Aegean and the inner basins of the Balkans, such as Macedonia, the Sofia plain, and even Hungary, considerably earlier than their first emergence in the coastal areas of the Black Sea.

4.2 The End of the Sixth Millennium BC

At the end of the 6th millennium, roughly 5300 to 4800 BC, a significant cultural change took place both in the Balkans and in western Anatolia. In the Balkan stratigraphy, this change corresponds to the transition from Early to Middle Neolithic, marking the end of the Karanovo I–Starčevo assemblages and the appearance of Vinča elements. Population increased as pottery and tools that had not developed out of preceding assemblages were introduced. Some have called this a process of "Vinčaization," but there is substantial debate among archaeologists about the mode and mechanism of the transformation (Özdoğan 1993; Efe 1990, 1995).

In western Anatolia at this time, a total cultural displacement occurred. It was first noted by James Mellaart (1960), who proposed that the cause was a migration from the Balkans, but it later became clear that the same phenomenon was also happening there. No concrete evidence suggests a source for the new cultural elements, but it is not impossible that the influence originated and spread from the Black Sea littoral, the only area that is not verifiably incorrect because data are completely absent. Such an assumption is based on negative evidence and must be considered too speculative to advance with confidence. Whether or not these cultural changes have anything to do with variations in marine conditions is an issue in need of further examination.

4.3 The Bronze Age

By the early stages of the Bronze Age, there is considerable evidence for population movements, both in the Caucasus as well as in Anatolia and the Balkans. It is almost certain that these changes were due to cultural turbulence, which might have been triggered by climatic oscillations. For an overview of the Bronze Age in the Marmara area, see Özdoğan (2002).

5. FINAL WORDS

This paper attempted to present an overview of the current archaeological evidence along the southern Black Sea coast. It should provide a foundation for natural scientists unfamiliar with cultural evolution in the area by introducing the enigmatic site distributions and the debates that surround these findings. Though it ends without reaching a conclusion, its success will be measured by how much it fosters greater interdisciplinary collaboration.

ENDNOTES

1. Most archaeological sites contain superimposed deposits, which serve to establish both stratigraphic sequencing and relative dating of their respective assemblages. Sites occupied for only a single period, however, provide no such temporal sequencing of cultures, and their chronological relationship with other sites must be worked out by comparing changes in typology or style of incorporated artifacts. This dating process is often called "horizontal chronology," or seriation.

2. There are several well written volumes on the later prehistory of Turkey, each with a different scope and coverage. Among these, the writer would suggest Karul (2002), Karul (2003), Özdoğan and Başgelen (1999), and Yakar (1991).

3. Two hoard finds, one from Ordu and the other from Trabzon, might be the only exceptions. The former is of the 2nd millennium BC (Bittel and Schneider 1944) and the latter, Late Chalcolithic of the Varna type (Rudolph 1977). Both hoards, however, are of unknown provenience.

4. It is not possible to present details for such parallelism here. For specific cases, see Özdoğan (1996), Summers (1993), and Thissen (1992). Nevertheless, it is of interest to note that the prehistoric assemblages of Central Anatolia have more parallels in the western Pontic area, such as the Varna region, than in the Marmara region. This strongly implies that the connection between Central Anatolia and the Balkans was by sea, not by land.

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SUBMERGED PALEOSHORELINES IN THE SOUTHERN AND WESTERN BLACK SEA — IMPLICATIONS FOR INUNDATED PREHISTORIC ARCHAEOLOGICAL SITES

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Abstract: During the course of six field seasons, from 1998 to 2003, groups from the Institute for Exploration (IFE) and, recently, the Institute for Archaeological Oceanography (IAO) at the University of Rhode Island Graduate School of Oceanography (URI-GSO) have conducted expeditions to the Black Sea for marine geological and archaeological research. These expeditions were conducted in partnership with groups from other institutions, including the University of Pennsylvania, Florida State University, Massachusetts Institute of Technology, the Institute for Nautical Archaeology at Texas A&M University, and Woods Hole Oceanographic Institution. Oceanographically, the Black Sea is a distinctive body of water with highly preservative anoxic bottom water and a submerged coastal plain that was exposed and possibly habitable for periods of time in the recent geologic past. Archaeologically, the Black Sea basin has a rich maritime-oriented history with a number of coastal and near-coastal sites around the present-day perimeter that contain important information about human history from Paleolithic times to the present. Using a suite of deep submergence technologies, a deep-towed side-scan sonar and subbottom profiling system, an optical imaging towsled, and remotely operated vehicle (ROV) systems, we have discovered, surveyed, imaged, and sampled several interesting geological and archaeological sites that shed more light on the ancient natural and human history of the region. These include several well-preserved prehistoric shorelines presently submerged about 150 m below present-day sea level, a very well preserved wooden ship dating to Byzantine times about 1500 years ago, an amphora-laden trading vessel dating to the Hellenistic period about 2400 years ago, and a possible site of human habitation along one of the ancient shorelines that was inundated during the Neolithic period more than 8000 years ago.

Keywords: underwater archaeological sites and survey, submerged paleoshorelines, sea-level migration, Holocene

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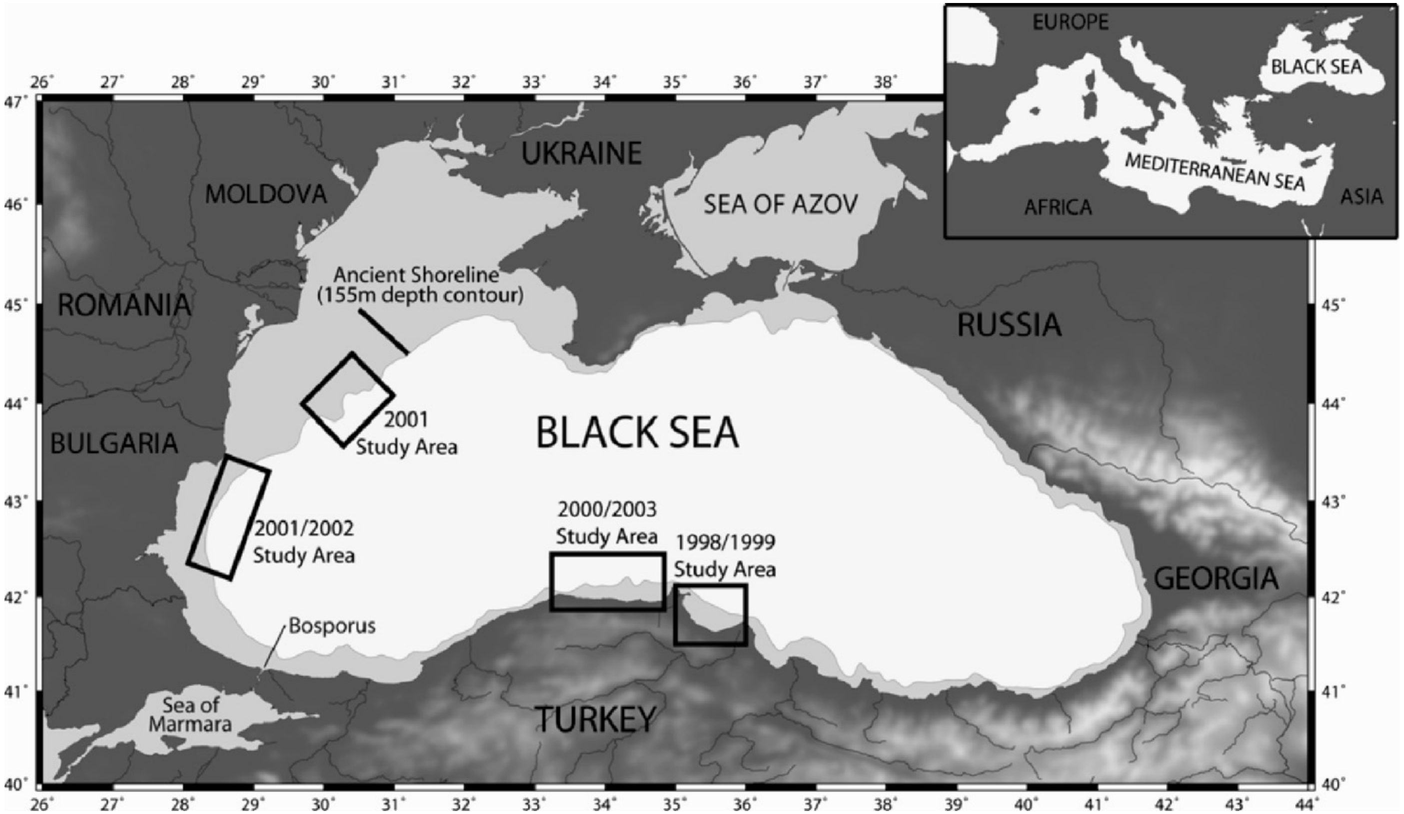


Figure 1. Map of the Black Sea region. Black boxes mark the general locations of 1998–2003 study areas. Outline of the 155 m isobath is indicated, representing the approximate location of the pre-flood shoreline.

1. INTRODUCTION

The Black Sea (Figure 1) is an interesting body of water from both oceanographic and archaeological points of view. During the early Holocene, the water of the Black Sea was significantly fresher than today, possibly contained within a lake-like or estuarine environment (Ross *et al.* 1970; Ryan *et al.* 1997). The lake-like environment then changed, perhaps catastrophically, to a marine environment as the global rise in sea level caused saline waters to breach the sill of the Bosporus Strait and spill into the Black Sea basin, which at that time was at a lower elevation (–25 to –40 m according to Aksu *et al.* 2002 and Görür *et al.* 2001, but perhaps as low as –150 m according to Ryan and Pitman 1998). An important feature of the Black Sea is that below an average depth of 150 meters, the water is anoxic. The lack of oxygen prohibits benthic and mid-water biota from existing below this depth. The bottom waters became anoxic around the same time that they became saline, directly following the aforementioned marine inundation (Deuser 1974; Wilkin *et al.* 1997).

An additional unusual feature is that the surface currents of the Black Sea circulate in two large cyclonic (counterclockwise-rotating) gyres, one occupying the eastern half of the basin, the other occupying the western half (Oguz *et al.* 1993). The two gyres meet in the central portion of the basin where the eastern gyre flows southward and the western gyre flows northward. Non-coincidentally, this unique flow regime lies along a suspected ancient north-south trade route that crossed the Black Sea from the Sinop region of northern Turkey to Crimea in the southernmost Ukraine (Hiebert 2001).

Researchers associated with the Institute for Exploration conducted expeditions to the Black Sea during six field seasons, from 1998 through 2003 (Figure 1). During the first five seasons, the objectives were to map the submerged landscape that was inundated by rising sea level (including former river valleys), identify ancient shoreline features, search landward of these paleo-shorelines for evidence of drowned Neolithic human settlements, and explore the sea floor for well-preserved ships of antiquity that sank along suspected trade routes. From 1998 to 2000, the region of interest was the northern coast of Turkey, in the southern Black Sea. From 2001 to 2002, the focus shifted to the western Black Sea coast and continental shelf of Bulgaria and Romania. During the summer of 2003, we returned to the southern Black Sea to continue our explorations of the pre-flood landscape and ancient shipwrecks, revisiting the sites discovered during the 2000 season and testing a new remotely operated vehicle system designed for archaeological excavations in the deep sea.

Many of the operational details and scientific results from these expeditions have been published elsewhere. Several of these publications deal primarily with the geology and oceanography (Ballard *et al.* 2000; Coleman 2003); others deal primarily with the archaeology (Ballard *et al.* 2001; Hiebert

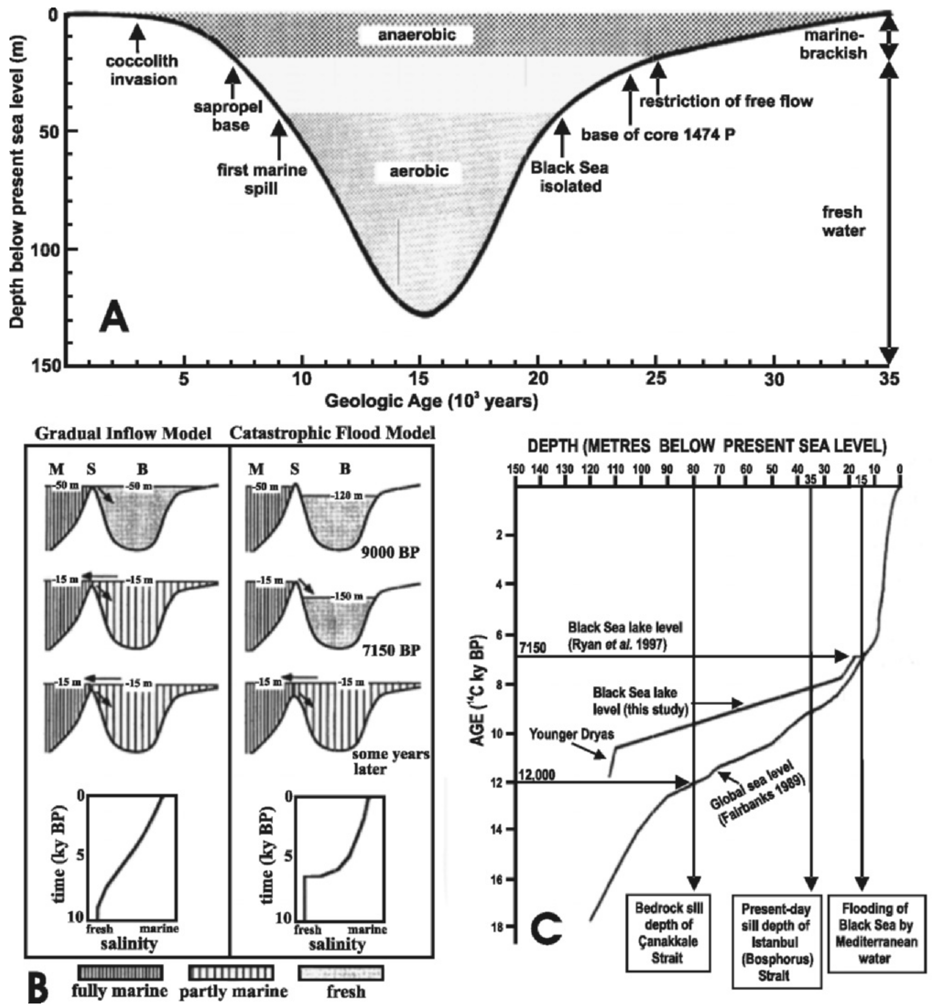


Figure 2. Models representing sea-level rise histories for the Black Sea basin. Upper figure (A) follows Ross and Degens (1974). The global sea-level rise curve for the last 35,000 years, after Milliman and Emery (1968), shows that once sea level dropped below the sill depth of the Bosporus, the Black Sea became isolated, and its waters freshened. The lake remained isolated until the first marine spillover, about 9000 BP, followed by deposition of the marine sapropel unit, and followed most recently by deposition of the coccolithic ooze layer. Lower left figure (B), after the model of Ryan *et al.* (1997), proposes a catastrophic flood of the basin *ca.* 7150 BP that would have rapidly changed the lake into a sea, in contrast to the gradual inflow model. The Ryan model is also depicted in the adjacent, lower right figure (C), presented as a sea-level rise curve, which shows a competing model from Görür *et al.* (2001). Their sea-level rise curve (middle) suggests that the marine invasion of the Black Sea was more gradual.

2001; Hiebert *et al.* 2002); and others deal primarily with the technology (Coleman *et al.* 2000, 2003; Coleman and Ballard 2004). This paper extracts information from these publications but places the main focus on the submerged geological evidence in support of a rapid flooding event and on the potential for discovering presently submerged sites of human occupation within the inundated landscape of the Black Sea shelf.

1.1 Holocene Oceanography

There has recently been considerable renewed interest in the Holocene evolution of the Black Sea region following the work of several research groups and their contrasting hypotheses about the latest connection between the Black Sea and the global ocean through the Sea of Marmara and Mediterranean Sea (i.e., Aksu *et al.* 1995; Ryan *et al.* 1997). This interest escalated following the publication of William Ryan and Walter Pitman's book, *Noah's Flood* (1998), in which they proposed a catastrophic influx of marine water around 7150 BP, which flooded about 150,000 square kilometers of the coastal landscape surrounding the "New Euxine Lake." This transition from a lake to a sea was first discovered and documented in the late 1960s by geologists who analyzed many sediment cores collected in the Black Sea (Ross *et al.* 1970; Degens and Ross 1972; Ross and Degens 1974).

Several groups of marine geologists have challenged the work of Ryan and Pitman and have suggested that (1) the inundation was much more gradual (Aksu *et al.* 1999; Görür *et al.* 2001), and (2) that it may have, in fact, progressed in the opposite direction, from the Black Sea into the Sea of Marmara and the Aegean Sea (Aksu *et al.* 1995; Aksu *et al.* 1999; Aksu *et al.* 2002; Hiscott *et al.* 2002). However, the discovery of a submerged paleoshoreline off northern Turkey (Ballard *et al.* 2000) and the mapping of ancient river channels that once flowed across a terrestrial landscape, such as those off the northern (Ryan *et al.* 1997) and western (Coleman 2003) Black Sea coast, indicate that the water level of the Black Sea was up to 155 meters lower than the present-day level prior to the marine inundation. More recent work suggests the inundation occurred about 8400 BP (Major *et al.* 2002) and that the level of the ancient lake was higher.

The Holocene evolution of the Black Sea is illustrated schematically in Figure 2, with models proposed by Ross and Degens (1974), Ryan *et al.* (1997), and Görür *et al.* (2001) based on their respective data and interpretations. Other competing models are not depicted here but appear in various papers within this volume (i.e., Algan *et al*; Balabanov; Chepalyga; Glebov and Shel'ting; Hiscott *et al*; Konikov; Kuprin and Sorokin; Lericolais *et al*; Shmuratko; Shuisky; and Yanko-Hombach).

Ross *et al.* (1970) published the first western reports on the recent geology of the basin based on the analysis of deep-water sediment cores. In nearly every core, these authors identified three basic lithologies: a modern uppermost unit of coccolithic ooze (Unit I), or deep-sea mud containing stratified layers of plankton-rich carbonaceous clay, underlain by a thick sequence of sapropelic mud (Unit II) with very high organic content, underlain by a sharp transition to light-colored lacustrine clay (Unit III) that persisted to the base of most cores. This same sedimentary sequence has been reported by a number of other authors who examined different cores throughout the Black Sea (Hay *et al.* 1991; Çağatay 1999). According to Ross *et al.* (1970), the marine transition started gradually about 9000 BP, prior to deposition of Unit II.

The model first put forward by Ryan *et al.*(1997) appears in Figure 2B (lower left). It suggests that a catastrophic flood occurred about 7150 BP, abruptly transforming the lake to a sea. Just prior to the flood, the lake was interpreted to have been about 150 meters lower in elevation than today, although other data show a shallower lowering of about 75 meters (Görür *et al.* 2001; Hiscott *et al.* 2002). Görür *et al.* (2001) presented new data from a shallow-water survey near the Sakarya River in northern Turkey suggesting that the marine transition occurred at 7150 BP, but the inflow was more gradual, and the elevation of the lake was much higher than Ryan *et al.*(1997) had reported (Figure 2C, lower right). Although the curves look different, all of the models presented in Figure 2 indicate that the proto-Black Sea water level 12,000 years ago was more than 100 m lower than the present-day level. Therefore, despite the rapid influx of marine waters (or glacial meltwater), a major portion of the Black Sea shelf was inundated since that time, and that fact has significant archaeological implications.

1.2 Prehistoric Archaeology

During the peak of the Würm glaciation, the late Pleistocene climate was harsh in northern Europe and Russia. Nevertheless, humans migrated and spread throughout Europe and its environs (van Andel and Tzedakis 1996). The cold, dry climate interval (Younger Dryas) following this last Ice Age could have made the Black Sea basin a desirable place to live. It is quite conceivable that human populations settled near-shore coastal environments surrounding the lake-like oasis. In fact, many prehistoric archaeological sites have been discovered throughout parts of interior and coastal Turkey, Bulgaria, and Romania.

Contradictory data exist, however, and have prompted interpretations that argue against the habitability of the proto-Black Sea coast (i.e., Mudie *et al.* 2004). The Late Paleolithic, Mesolithic, and Early Neolithic periods are represented by a scattering of archaeological sites, but data are sparse. The only inhabitants of the region during these periods were probably small bands of indigenous hunters and gatherers (Ryan and Pitman 1998). Özdoğan (2003) showed that no archaeological sites of Neolithic to Bronze Age date exist near the present-day Black Sea shoreline, but some sites cluster in areas away from the coast.

In Eurasia and the Near East, by middle to late Neolithic times, human populations flourished, small settlements coalesced into larger villages, and stone technology advanced. Two noteworthy sites representative of advanced Neolithic cultures in the region are Lepenski Vir (Srejovic 1972) along the banks of the Danube, and Çatal Hüyük (Mellaart 1967) in south-central Turkey. Although both of these sites are far from the Black Sea coast, and on opposite sides of the Sea, they provide strong evidence of regional advanced cultures with significant direct parallels represented in painting and sculpture (Ryan and Pitman 1998). Though they remain controversial, agricultural village sites dating to at least the seventh millennium BP have been discovered along the western and southern Black Sea coasts according to Hiebert *et al.* (2002) and Ballard *et al.* (2001), and contra Özdoğan (2003). Subsequently, evidence for ceramic and metal technology appears along these same coasts (Ballard *et al.* 2001).

In coastal Bulgaria (Figure 1), a significant prehistoric archaeological discovery was made – the earliest evidence of gold metallurgy anywhere in the world and perhaps the earliest civilization in Europe, dating to 6600 calBP (Florov and Florov 2001). The site in question, a necropolis in the Varna Lakes region of Bulgaria in the present-day Provadiyska River valley, contained many graves, some of which included copper tools along with jewelry and ornaments made from processed gold. The people who occupied the Varna Lakes region probably developed copper and gold metallurgy themselves, independently inventing it in Europe far from the influence of earlier Near Eastern metallurgists (Renfrew 1980). Another Bulgarian archaeological site from a similar time period has been excavated at Karanovo. The Neolithic is represented here, dating to about 8000 BP and perhaps earlier, transitioning to the Copper Age (early Chalcolithic) at about 7000 BP and later (Renfrew 1980). Bulgaria's central location has made it a connection between a diverse array of Near Eastern, Mediterranean, Russian steppe, and central European cultures, and for this reason the region possesses some of the most significant archaeological resources in all of Europe (Bailey and Panayotov 1995; Florov and Florov 2001).

In 1985, divers in a Russian submersible found and recovered a large ceramic plate (Dimitrov 1999) in about 90 m of water off the Bulgarian coast. This artifact was retrieved from the presently submerged ancient Provadiyska River valley, near the paleoshoreline, about 50 km east of Varna. This plate was not found in association with any other artifacts, and its depositional context has not been precisely determined, therefore more work needs to be done to interpret its significance as a Neolithic artifact.

Elsewhere in the Black Sea region, archaeological sites of the late Neolithic have been discovered on the southern tip of the Crimean peninsula in Ukraine; they contain extensive evidence of shellfish gathering for use as a food resource (Burov 1995). Other evidence exists on rock carvings for whale hunting, and dolphin bones are present at a few sites. These finds indicate that the early inhabitants of the Black Sea coast exploited littoral resources following the marine inundation. This could suggest that earlier hunter-gatherers exploited the coastal regions of the Neoeuxinian lake for potential food resources. Some researchers have argued that the Black Sea coast would not have been habitable during the Neolithic for a variety of reasons, including saline soils and brackish water preventing agricultural activities (i.e., Mudie *et al.*2002).

2. METHODOLOGY

We set out to explore and document submerged prehistoric and historic archaeological sites in the region armed with three ideas (Ballard *et al.*2001): (1) that much evidence existed for diverse Mesolithic and Neolithic populations in inland areas outside the coastal perimeter of the Black Sea, (2) that two prominent marine geologists had recently proposed their hypothesis of a major catastrophic flood in the Black Sea basin perhaps linked to the biblical legend of Noah (Ryan and Pitman 1998), and (3) that another prominent marine geologist had suggested that the best preserved ships of antiquity should be found in the Black Sea (Bascom 1976). Initial survey sites were selected in the southern and western Black Sea to carry out the first phases of archaeological exploration.

The primary methodologies employed during the 1998–2002 expeditions were: remote sensing (primarily geophysical prospecting) using side-scan sonar and subbottom profiler, and ground truthing the remotely sensed targets using geological sampling equipment, remotely operated vehicle (ROV) systems, and a submersible. In each case, survey tracklines were chosen to cross the predicted location of submerged shorelines near suspected paleo-river channels, searching for potential inundated terrestrial archaeological sites, and to cross into deeper, anoxic water to search for well-preserved ancient shipwrecks. The primary methodology during our return visit to the Turkish sites in 2003 was subsurface exploration of selected submerged archaeological sites, limited artifact collection, and advanced site mapping using a new ROV system. Many of the methodologies discussed here have been published in greater detail elsewhere (i.e., Coleman 2002; Coleman 2003; Coleman and Ballard 2004).

During 1998 and 1999, a number of fishing vessels were chartered out of Sinop, Turkey, and used to collect the remote sensing data and geological samples (Ballard *et al.* 2000). The survey location was east of the Sinop Peninsula. The primary tools used to map the submerged surficial geology and explore potential archaeological targets were a Marine Sonics side-scan sonar towfish and an Edgetech DF-1000 side-scan sonar towfish. During the 1999 survey, acoustic targets and geological features identified during both years were investigated with a small ROV. The ROV was equipped with scanning sonar and video equipment that aided in locating and identifying the targets. In addition to the collection of sonar and video data, geological samples were collected during the 1999 survey using a small frame dredge. The shell samples collected by the dredge along the Sinop paleoshoreline were sent to Dr. Gary Rosenberg of the Academy of Natural Sciences for species identification, and selected specimens were submitted for radiocarbon dating to the National Ocean Sciences Accelerator Mass Spectrometer (NOSAMS) facility at the Woods Hole Oceanographic Institution (WHOI).

During the 2000 expedition, we chartered the British motor vessel *Northern Horizon*, which could be driven by a dynamic positioning system, to operate WHOI's deep-towed side-scan sonar *DSL-120* and IFE's newly developed imaging sled *Argus* and ROV *Little Hercules* (Coleman *et al.* 2000). In general, the same methodologies were employed as during the previous field seasons, but this time we used a larger ship and larger survey equipment capable of reaching deeper water depths, mapping wider swaths, and operating 24-hours per day. A new survey region to the west of the Sinop Peninsula was explored during the 2000 expedition that we identified as a potential high probability area for locating submerged archaeological sites due to its wide and relatively flat shelf, and because the region to the west also lay along a suspected ancient trade route between Sinop and the Bosporus. The first phase of this survey was a side-scan sonar survey, followed by ROV dives on selected acoustic targets. The ROV was used for close-up video inspection of the acoustic targets, and to collect a limited amount of geological and archaeological samples. Selected archaeological samples were submitted to Beta Analytic's radiocarbon laboratory for analysis.

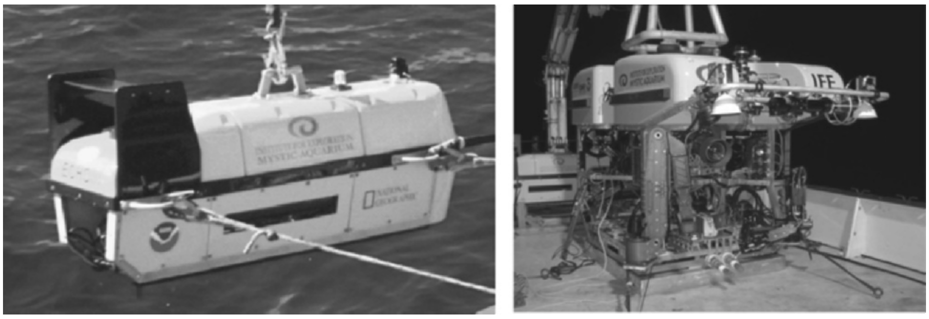
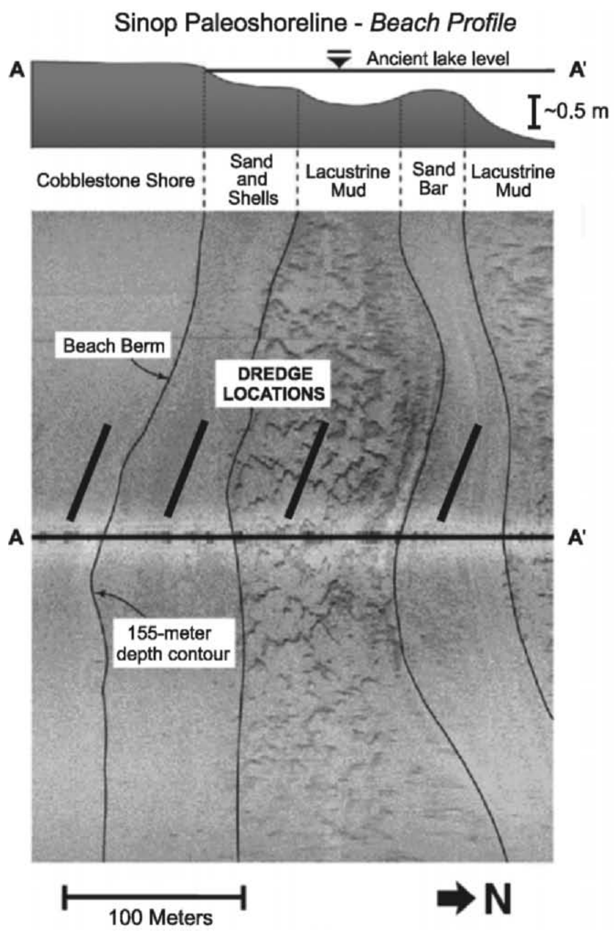


Figure 3. Two of the seafloor mapping and imaging systems used during several of our Black Sea expeditions. On the left is IFE's deep-towed side-scan sonar and subbottom profiler, *Echo*, and on the right is *Hercules*, a large, remotely operated vehicle (ROV) capable of underwater archaeological excavation.

During the 2001 survey, we used the research vessel *Akademik* operated by the Bulgarian Academy of Sciences Institute of Oceanology (BAS-IO) out of Varna, Bulgaria, to tow a newly developed side-scanning and subbottom profiling sonar named *Echo* (Coleman 2003) (Figure 3, left). A number of locations off the Bulgarian coast were surveyed, primarily along the suspected paleo-shorelines and paleo-river channels, and also where the Holocene sediment thickness was at a minimum. Another region was surveyed far off the Romanian


Sinop Paleoshoreline - *Beach Profile*
A
A'
Ancient lake level
~0.5 m
Cobblestone Shore
Sand and Shells
Lacustrine Mud
Sand Bar
Lacustrine Mud
Beach Berm
DREDGE LOCATIONS
155-meter depth contour
100 Meters
N

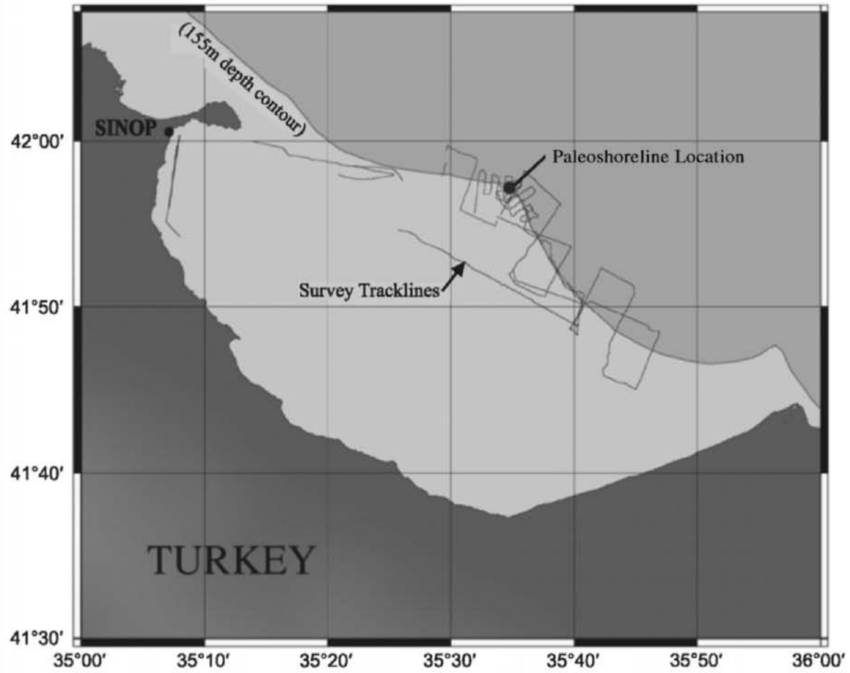

(155m depth contour)
SINOP
Paleoshoreline Location
Survey Tracklines
TURKEY
42°00′
41°50′
41°40′
41°30′
35°00′
35°10′
35°20′
35°30′
35°40′
35°50′
36°00′

Figure 4. (Facing page) Side-scan sonar image collected across the 155-meter depth contour, traveling south to north about 30 km east of Sinop, Turkey (left). Upper portion of figure is a vertical section across the paleoshoreline from A – A' revealing a "typical" beach profile (after Ballard *et al.* 2000). Dredge locations for cobble and shell samples are depicted as lines crossing the submerged beach parallel to the contours. Sinop paleoshoreline location is depicted on the map (right).

coast, surrounding and within the Viteaz Canyon, near the distal reaches of the Danube fan. This acoustic survey successfully identified a number of interesting acoustic targets and seafloor geological features. In addition to mapping during the 2001 survey, we collected sediment cores with a gravity corer. The cores were split, photographed, analyzed, and described onboard the ship.

During 2002, we used the BAS-IO research submersible *PC-8B*, equipped with scanning sonar and video equipment, for 8 dives on selected acoustic targets identified from the 2001 survey. We collected some geological and archaeological samples using the submersible's manipulator. Some of the archaeological samples were submitted to the NOSAMS facility at WHOI for radiocarbon dating.

In 2003 on the *R/V Knorr* and with the newly developed ROV *Hercules*, we returned to selected sites previously discovered off the Turkish coast in 2000 (Coleman *et al.*2003) (Figure 3, right). This ROV was specifically designed to conduct limited robotic archaeological excavations on the seafloor using a sediment and water suction system, water jets, and excavation tools to carefully and effectively remove sediment from around artifacts, and to collect geological and archaeological samples (Webster 2004). Shipwreck studies included advanced mapping with acoustic and optical imaging sensors and limited subsurface exploration and sampling (Coleman *et al.*2003).

3. RESULTS AND INTERPRETATION

3.1 Offshore North-Central Turkey

The 1998 survey off Sinop, Turkey, was very successful in establishing the partnerships and laying the groundwork for future archaeological survey research in the region. In addition to the marine survey, a terrestrial survey commenced to investigate the maritime archaeological landscape of the Sinop region (Doonan 2004, this volume). The 1999 survey successfully imaged and sampled submerged geological features off the Sinop coast. Tens of square kilometers of side-scan sonar data were collected that revealed the nature of the seafloor geology and a number of other acoustic targets. One particular location showed evidence indicative of a relic shoreline (Figure 4, left). Located about 30 kilometers due east of Sinop (Figure 4, right), this shoreline immediately stood out as a unique feature on the side-scan sonar record. The acoustic character of the seafloor geology changed as our survey progressed from south to north across the feature. In addition, the relief of the seabed changed subtly. The dives undertaken by the ROV along the same transect revealed a gently dipping coastal plain with lots of cobbles and well-rounded stones that gradually were replaced by muddy sediment. The topographic profile along this transect appears identical to that of a modern beach, with a berm and offshore sand bar (Figure 4, top left). The dredge samples collected a number of diverse shoreline materials, including smooth cobbles, rounded stones and pebbles, and a large quantity of shells.

Table 1. Taxonomy and radiocarbon dating results for molluscs collected from the Sinop paleoshoreline during the 1999 expedition; dates reported by the National Ocean Sciences Accelerator Mass Spectrometer (NOSAMS) facility, Woods Hole, MA (after Ballard *et al.* 2000). Dates are in radiocarbon years before present. Shell types are indicated by the letters F and S: F = freshwater, S = saltwater. Refer to Figure 5 for a plot of age ranges for the different shell types.

Accession #	Species of Mollusc	Type	Age (years BP)	± Error
OS-21644	*Papillicardium papillosum*	S	3200	70
OS-21645	*Modiolula phaseolina*	S	2810	40
OS-21646	*Dreissena rostriformis* var. *distincta*	F	15,550	120
OS-21647	*Dreissena rostriformis* var. *distincta*	F	10,800	75
OS-21648	*Dreissena rostriformis* var. *distincta*	F	10,450	70
OS-21649	*Dreissena rostriformis* var. *distincta*	F	10,450	70
OS-21650	*Dreissena rostriformis* var. *distincta*	F	10,400	70
OS-21651	*Dreissena rostriformis* var. *distincta*	F	10,500	70
OS-21652	*Trophonopsis breviata*	S	2800	45
OS-21653	*Bittium reticulatum*	S	3860	50
OS-21654	*Abra alba*	S	4130	45
OS-21655	*Mytilus galloprovincialis*	S	6820	55
OS-21656	*Parvicardium exiguum*	S	3190	50
OS-21657	*Turricaspia caspia lincta*	F	12,100	85
OS-21658	*Turricaspia caspia lincta*	F	9310	65
OS-21659	*Turricaspia caspia lincta*	F	7460	55
OS-21660	*Turricaspia caspia lincta*	F	7590	55
OS-21661	*Turricaspia caspia lincta*	F	7480	55
OS-22454	*Mytilus galloprovincialis*	S	4830	45
OS-22455	*Mytilus galloprovincialis*	S	6470	45
OS-22456	*Mytilus galloprovincialis*	S	6020	45
OS-22457	*Mytilus galloprovincialis*	S	4000	45

A complete representative set of mollusc shells from the dredge hauls was identified by species and submitted for radiocarbon analysis. The results from the radiocarbon dating are presented in Table 1 and illustrated graphically in Figure 5 (top). The two most common shell types from this paleoshoreline were those of *Dreissena rostriformis*, an extinct mollusc with a close affinity to those from the Caspian Sea (Ballard *et al.* 2000) and *Mytilus galloprovincialis*, a salt water mollusc that is prevalent in the modern Black Sea. Another mollusc found among the shoreline samples was *Turricaspia caspia lincta*. These molluscs can survive in fresh to brackish (low salinity) water but are referred to hereafter as “freshwater” species to distinguish them from the higher salinity molluscs. Most of the shells from the dredge hauls were well-articulated, many even with both valves intact, a fact that is attributed to their *in situ* nature. A significant observation concerning the freshwater shells is that they were all completely bleached white in color (Figure 5, top). This qualitative assessment indicates that the shells were likely exposed to the sun and air at some time during their existence. As indicated in Table 1, the dates of all the freshwater shells were older than 7400 BP, while those of all the saltwater shells were younger than 6800 BP, indicating that the transition from the Neoeuxinian lake

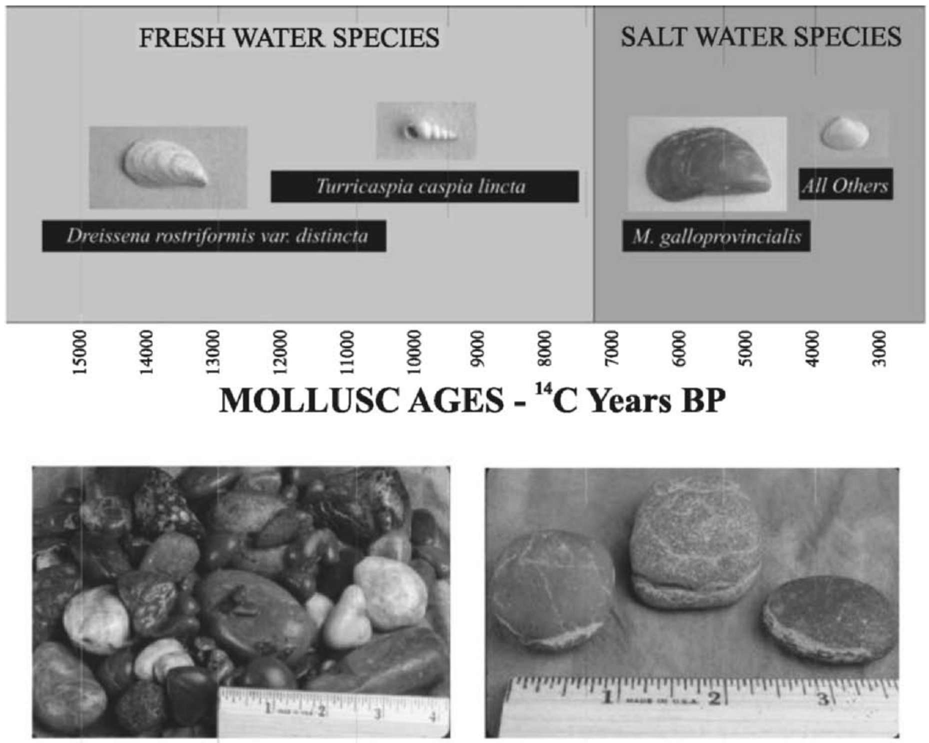


Figure 5. Samples collected in the dredge during the 1999 survey off Sinop. Age range (upper figure, in radiocarbon years BP) of selected mollusc shells collected in a dredge haul along the strike of the Sinop paleoshoreline, seaward of the 155-m depth contour. Total mollusc age ranges from the data in Table 1 are represented by the length of each black bar. Freshwater molluscs plot older than 7460 BP, saltwater molluscs plot younger than 6820 BP (Ballard *et al.* 2000). Lower left figure shows rounded, poorly-sorted cobbles collected in a dredge haul near the Sinop paleoshoreline, landward of the 155-m depth contour near the beach berm, as indicated in Figure 4. On the lower right is a subset of these cobbles that show a concreting ring around the flat circumference of each, indicating they were once welded together (Ballard *et al.* 2000).

to the Black Sea occurred within that interval. Because *T. caspia lincta* can survive in brackish water, it is likely that the transition started to take place prior to 7400 BP, as suggested by Major *et al.* (2002).

In addition to the mollusc shells, a large collection of rock samples, mostly cobblestones, was collected in the dredge hauls (Figure 5, bottom left). All of the collected stones were well-rounded, indicating they had existed in a high-energy geologic setting. Since this region was never covered by glaciers, the only explanation is that these stones were worked by moving water, either in a riverine, or near-shore marine or lacustrine environment, such as along an exposed shoreline. A number of the cobblestones were flat and revealed a ring of fine-grained, concreted mud around their circumference (Figure 5, lower right). These samples likely came from the top of a relict backshore portion of an exposed beach, possibly well-worn by surf for some time during the marine transgression. Following the inundation, fine-grained sapropelic mud draped a thin layer on top of this ancient shoreline (Ballard *et al.* 2000).

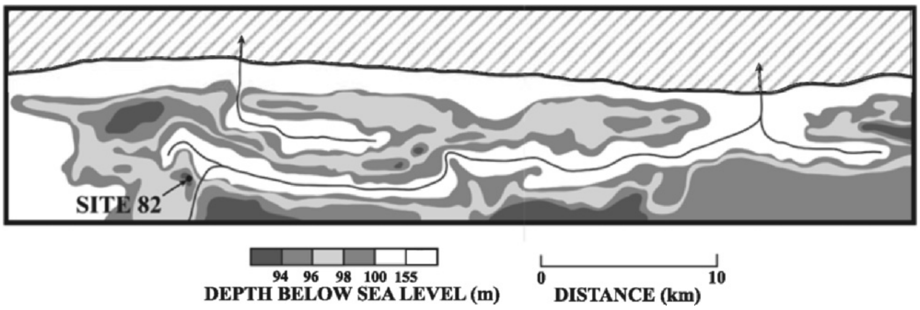


Figure 6. Bathymetric contour map for the 2000/2003 survey areas indicated in Figure 1. The undersea topography presents a slightly undulating landscape cut by lower-elevation paleo-river valleys that drained into the ancient lake. Location of side-scan sonar target #82 is indicated at the lower left near the confluence of two valleys (refer to Figure 7 for more detail).

During the 2000 expedition, hundreds of square kilometers of the continental shelf and upper slope west of Sinop were surveyed with side-scan sonar and a variety of bathymetric mapping techniques. The results of the bathymetric mapping effort are depicted in Figure 6. The submerged landscape is characterized by a rolling topography with small-elevation peaks and valleys, and a prominent paleoshoreline. This environment is interpreted as consisting of small hills between small drainage valleys (Ballard *et al.* 2001). Remnants of ancient river channels that flowed through these valleys could be seen in the side-scan records. Numerous acoustic targets littered this submerged landscape, the majority of which were large bedrock outcrops and boulders. At about 100 m depth on the continental shelf, three ancient shipwrecks were discovered that proved to be from the Roman-Byzantine periods, based on the typology and provenience of the cargo of amphorae (Ballard *et al.* 2001).

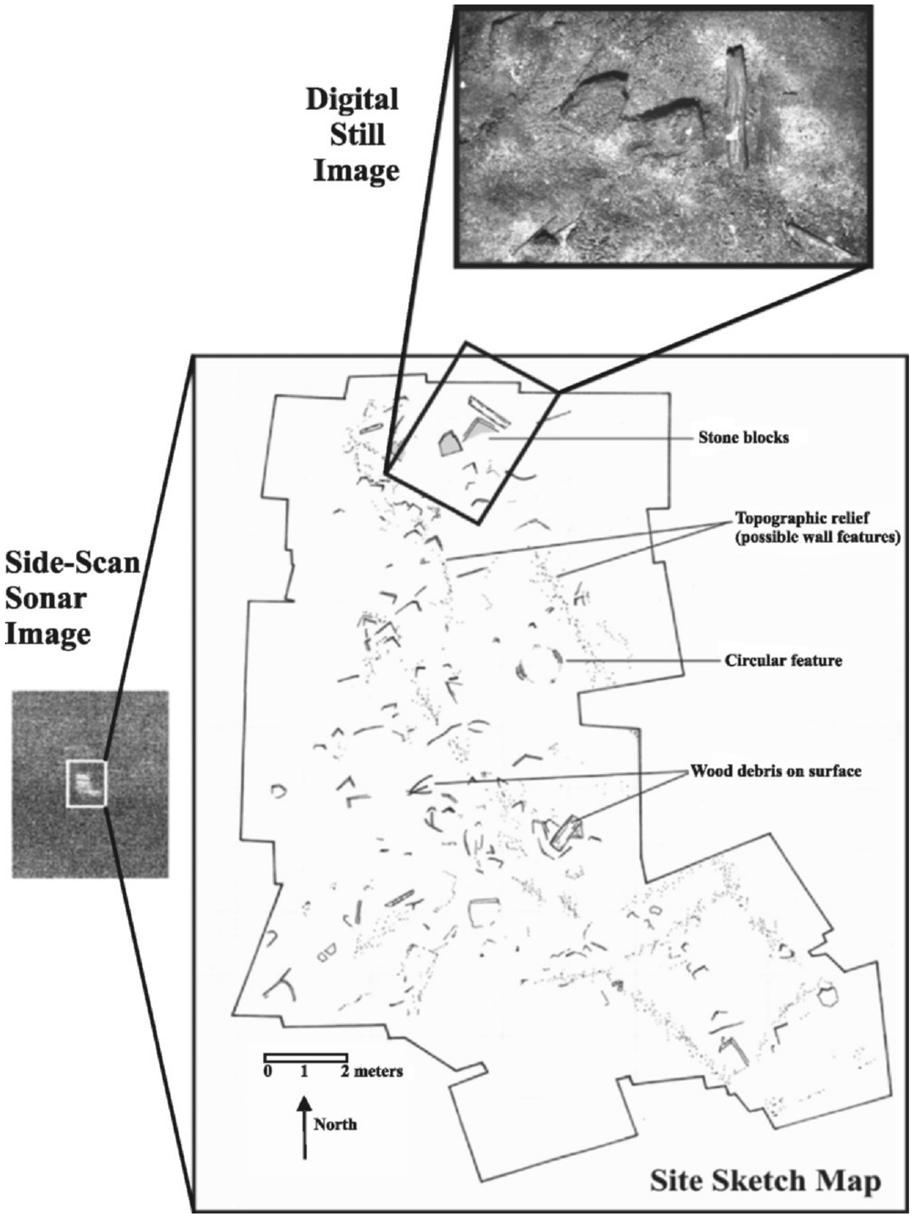


Figure 7. Site sketch map (center) of Site 82 revealing the spatial distribution of surficial objects and features, including many pieces of hewn wood, stone blocks with a worked appearance, and possible subsurface man-made structures (after Ballard *et al.* 2001 and Hiebert *et al.* 2002). The original side-scan sonar target is pictured at left in reversed gray-scale, making strong light-colored reflections. Approximate dimensions of the site are 10 by 15 m. A photograph (top) looking vertically down on the site, collected by an underwater electronic still camera, reveals some of the wood and stone features.

In addition, a beautifully preserved ancient wooden shipwreck was found at a depth of about 350 m on the continental slope west of the other shipwrecks. A small wood sample collected from this wreck, proved it was the same age as the others, dating to about 1500 radiocarbon years BP (Ballard *et al.* 2001).

While collecting side-scan sonar data, an isolated acoustic target stood out prominently on one of the records and was immediately recorded as a point of interest (Figure 7, left). It was chosen as the first dive site in 2003 using the ROV *Little Hercules*. The site, which became known as Site 82 (it was the 82nd sonar target we collected) appeared to have non-natural looking stone blocks and wood that, because of their shape and orientation, gave the appearance of having been worked and modified by humans. A high resolution bathymetric survey of the site using the *DSL-120* proved that it was perched at an elevation several meters higher than the immediate surroundings and near the confluence of two submerged river channels (Figure 6) (Ballard *et al.* 2001). This relief and location relative to rivers is significant in that it is common for Anatolian Bronze Age settlements to be sited in a such a fashion (Özdoğan 2003).

We used the *Argus* towsled to collect a series of electronic still camera (ESC) images (Figure 7, top). A number of adjacent ESC images were spliced together to create a composite mosaic image, from which we compiled an interpretive archaeological site map (Figure 7, center) (Ballard *et al.* 2001; Hiebert *et al.* 2002). A large quantity of wooden de bris was scattered around the site, mixed in with a number of oddly arranged stone blocks. Due to the degree of seismic activity along the Anatolian coast, it is possible that earthquakes were responsible for this odd arrangement. We collected a limited number of wooden artifacts from the site; eventually, these yielded young dates in the range of 100–250 years BP (Ballard *et al.* 2001), and we interpret this material to be recent objects that floated into the site. None of the other wooden samples were dated. This site and the general vicinity need to be investigated further in order to confirm or reject the hypothesis of a Neolithic origin for the objects.

3.2 Offshore Bulgaria and Romania

During the 2001 expedition to the western Black Sea, we again mapped hundreds of square kilometers of the continental shelf and upper slope off the coasts of Bulgaria and Romania using a side-scan sonar and subbottom profiler. Several survey regions were mapped and sampled by collecting gravity cores. In total, three regions on the Bulgarian continental shelf and slope and two regions on the Romanian shelf and slope, including the Viteaz Canyon were surveyed (Figure 8). A total of 18 gravity cores were collected, 10 off the coast of Bulgaria and 8 off the coast of Romania (Figure 8; Table 2), in water depths ranging from about 80 to more than 1600 m. The amount of sediment recovered in each core ranged from 30 to 355 cm (Table 2), depending primarily on the sediment type. The side-scan sonar survey located dozens of interesting acoustic targets and the

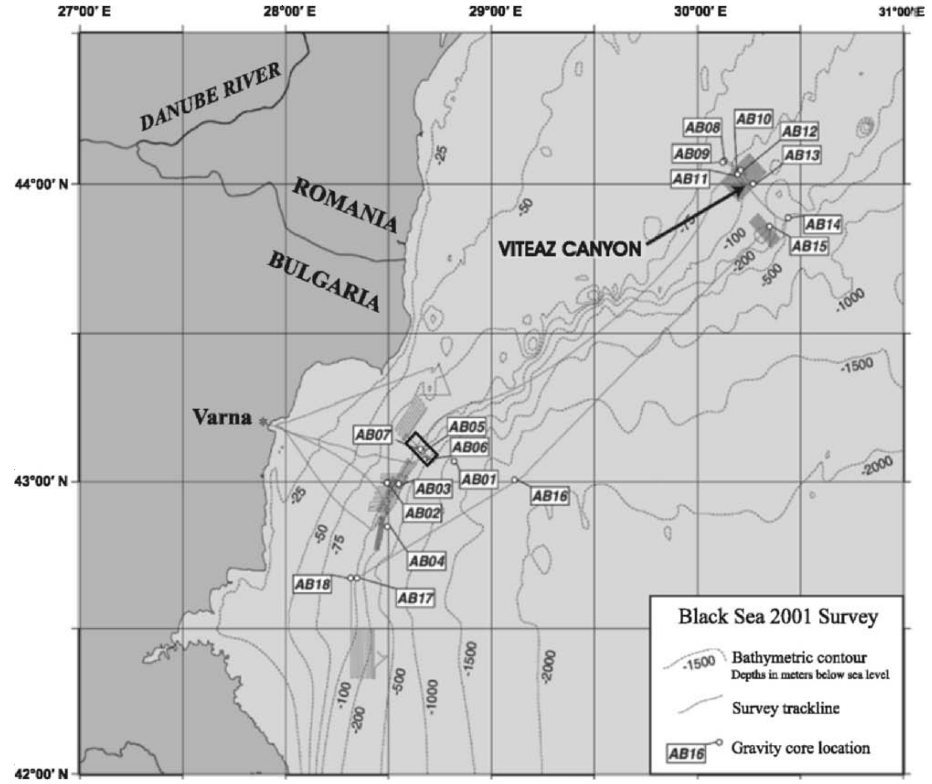


Figure 8. Survey regions in 2001, off Bulgaria and Romania. Using the *R/V Akademik* out of Varna, Bulgaria, five survey regions (three off Bulgaria and two off Romania) were chosen near locations where paleo-rivers met paleoshorelines and where Holocene sediment cover was thin. Side-scan sonar and subbottom profiler survey tracklines are depicted, as are positions of gravity cores (listed in Table 2), which were taken where subbottom profiler data had been collected. Small black box indicates location of Figure 9.

subbottom profiles helped us interpret the sub-seafloor geology, together with the gravity core data.

A focus of this expedition was to survey the shelf region off Varna, Bulgaria (Figure 8), near the location where the Paleo-river Provodiyska flowed across the presently submerged shelf. This ancient river valley is the location where an ancient plate was found previously (Dimitrov 1999). The survey region depicted in Figure 9 represents the intersection of the paleo-river with the paleo-shoreline (Dimitrov *et al.* 2003). A subbottom profile that crossed the delta of the Paleo-river Provodiyska parallel to the paleoshoreline (Figure 9) shows a number of fine sedimentary layers (Figure 10), and cores from the location of this profile indicate that the sediments were deposited from the late Pleistocene to early Holocene. This interpretation is based on a qualitative interpretation of the split cores, including fossil molluscs identified as stratigraphic markers, but these are

Table 2. Gravity core locations and details from the 2001 Bulgaria/Romania survey on *R/V Akademik.* Refer to Figure 9 for a plot of core locations.

Core ID	Date	Latitude	Longitude	Water Depth (m)	Recovered Length (cm)
AB1	8/16/01	43°04.2000' N	028°49.2200' E	1111	110
AB2	8/18/01	42°59.7700' N	028°29.7500' E	84	30
AB3	8/18/01	42°59.5400' N	028°33.0700' E	145	133
AB4	8/19/01	42°52.1000' N	028°29.6000' E	115	153
AB5	8/22/01	43°06.2600' N	028°39.8900' E	500	130
AB6	8/22/01	43°05.0636' N	028°41.1673' E	570	162
AB7	8/22/01	43°06.5900' N	028°39.3600' E	140	160
AB8	8/25/01	44°04.6472' N	030°07.7733' E	90	33
AB9	8/25/01	44°04.3733' N	030°07.3609' E	85	80
AB10	8/25/01	44°02.5900' N	030°12.1206' E	140	74
AB11	8/25/01	44°02.0320' N	030°11.6894' E	120	45
AB12	8/25/01	44°02.8300' N	030°12.5800' E	115	32
AB13	8/25/01	44°00.0987' N	030°16.1837' E	350	350
AB14	8/25/01	43°53.2900' N	030°26.4200' E	784	355
AB15	8/26/01	43°51.5329' N	030°20.9826' E	117	240
AB16	8/27/01	43°00.8700' N	029°08.6700' E	1629	235
AB17	8/27/01	42°40.3513' N	028°20.9182' E	120	270
AB18	8/27/01	42°40.3596' N	028°19.1361' E	90	100

not dated. Cut into the stratigraphic section are a number of smaller tributary-like features, or small river channels. This geologic structure resembles a cross-section through a modern day river delta and the presence of gas vents also suggests this interpretation. The gas is identified by small reflections in the water column above the diffractive acoustic lenses in the sediment profile. This apparent paleo-delta is presently submerged about 150 m below present-day sea level and may represent a relic river delta on the Neoeuxinian lake margin, which lay just seaward of the paleoshoreline at a depth of 150 m. These gas-charged sedimentary structures were commonly found at similar water depths throughout the survey regions.

Another prominent geologic feature evident in several subbottom profiles was the position of a pinched-out acoustic reflector in ~145 m of water depth (Figure 11). This reflector is traceable from areas of deeper water in the sections, where a thicker layer of sediment overlies the bedrock, and it pinches out at shallower depths, where it is overlain unconformably by modern (Holocene) sediment. This unconformity represents a break in the geologic record, which translates to a break in the continuity of geologic time as a result of a change from a depositional to erosional environment. In the case of a coastal shoreline setting, which this region represents, the erosional regime could have been above water, and the depositional regime could have been below water—although this is a matter of debate because, in high-energy beach environments,

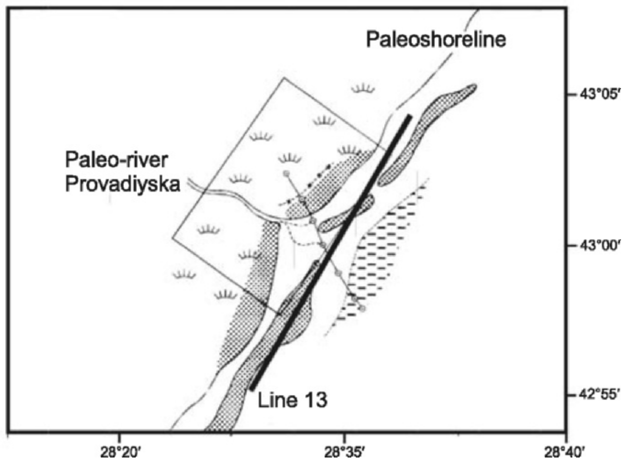


Figure 9. Interpretive sketch of the Paleo-river Provadiyska where it crossed the shelf and met the paleoshoreline of the Black Sea, about 35 km east of Varna, Bulgaria (after Dimitrov *et al.* 2003). The heavy black line represents the location of the subbottom profiles for Line 13, as shown in Figure 10.

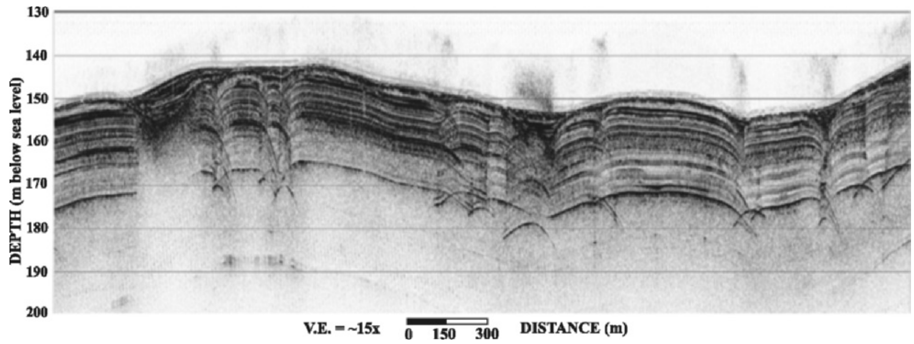


Figure 10. Subbottom profile of Line 13 collected parallel to and seaward of the paleoshoreline, roughly perpendicular to the axis of the Paleo-river Provadiyska. Individual strata are revealed across the length of the profile, which reveals the hummocky deltaic topography now submerged beneath about 150 m of water. Strong reflections in the water column and diffractive lenses within the strata indicate the presence of a large amount of gas beneath the surface, which is being expelled into the water. Profile location is shown in Figure 9. Vertical exaggeration is about 15x.

erosion from wave action could extend down tens of meters below the surface. We have interpreted the location where the unconformity pinches out as the location of the paleoshoreline.

The interpretation of the gravity cores reveals more evidence to support a rapid transition from a lacustrine environment to a marine environment (Major *et al.* 2002). A number of the cores that were collected from deep water on the continental slope show a marked change in sediment type from light gray clay to dark gray mud (Figure 12). This sedimentary horizon in the deep cores is extremely sharp and conformable, suggesting a rapid transition in energy regimes and water chemistry. In the gravity cores from the shallow continental shelf, there was no visible evidence of this rapid transition, either because of erosion

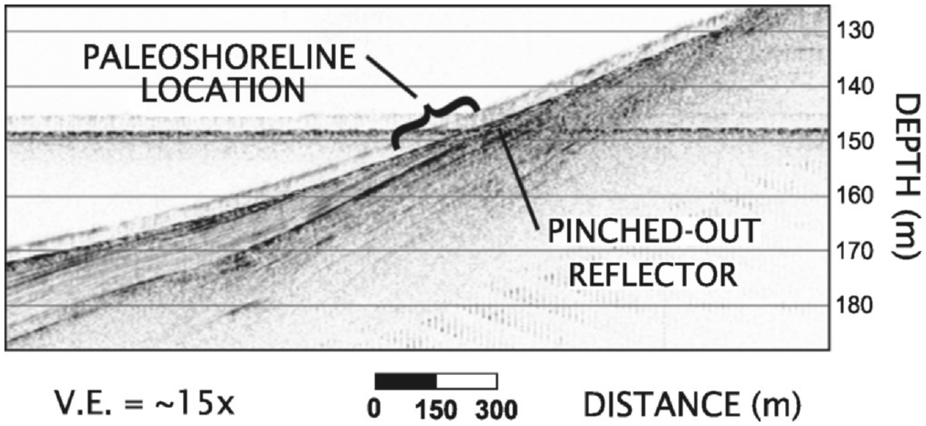


Figure 11. Subbottom profile crossing a western Black Sea paleoshoreline on the present-day continental slope. The acoustic unit between the lake floor and pinched-out reflector represents pre-flood Neoeuxinian sediment, which is conformably overlain by Holocene marine mud.

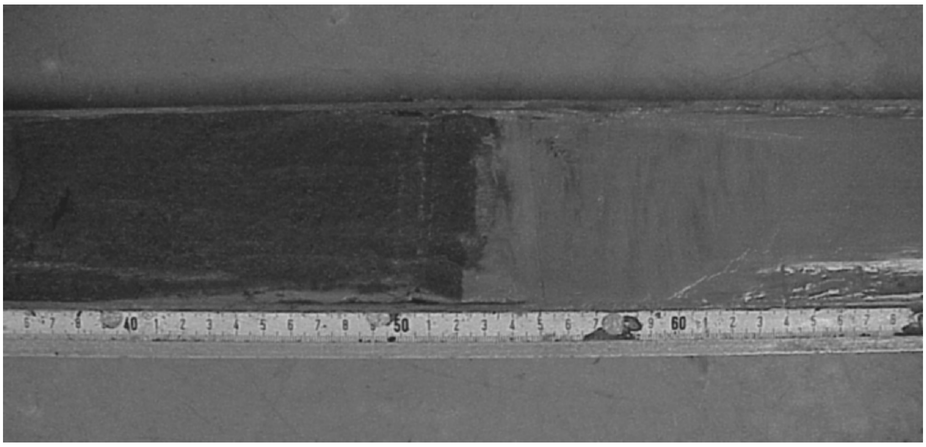


Figure 12. Gravity core AB-6 collected during the 2001 expedition off the Bulgarian shelf in 570 m water depth. Photograph of the middle portion of the split core (35 to 68 cm from surface) taken soon after it was collected. Top of the core is to the left, bottom to the right. Left side represents darker colored marine mud, right side represents lighter colored lacustrine clay. Core location is indicated in Figure 8 and Table 2.

or a rapid chemical change. The deep water cores (e.g., AB-6, Table 2 and Figure 8) record this rapid transition, but cannot indicate details about the level of the Black Sea prior to the transition, or the direction of flow during the transition period—see Major *et al.* (2002) for additional details.

During the 2002 expedition off the Bulgarian coast using the submersible *PC-8B*, we dove on some targets identified during the 2001 sonar mapping expedition and were successful in locating and investigating four shipwrecks. Three shipwrecks were estimated to be only several hundred years old (one was radiocarbon dated to be about 400 years BP). However, one of the ships contained a cargo of ancient amphorae. Using the submersible's manipulator, we sampled one of the clay jars and recovered fish bones with prominent cut marks. Radiocarbon dating of the bone material indicated the shipwreck was about 2300 years old (Coleman *et al.* 2003). In the western Black Sea, the rocky outcrops that were ubiquitous in the southern Black Sea were lacking. Nearly every acoustic target we dove on with the submersible in 2002 turned out to be a man-made object such as modern trash or shipwrecks.

Finally, an interesting observation was made while traversing the seafloor in the submersible at a depth of about 160 m in the southern survey region. Here we imaged what appeared to be an extensive field of relic mollusc shells, contained *in situ* on the seabed (Figure 13). Although we were not able to collect any of the shell samples, we recorded video footage of this apparent bed of fossil mussels. This was not confirmed by sampling, but the bed was expansive and consistent for several tens of meters along the slope contour. All the shells were observed to be aligned in the same direction, perpendicular to the nearby paleoshoreline. This relic feature is interpreted as being indicative of an ancient near-shore shallow water environment, presently submerged to a depth far greater than ones at which large mussel beds are known to exist.

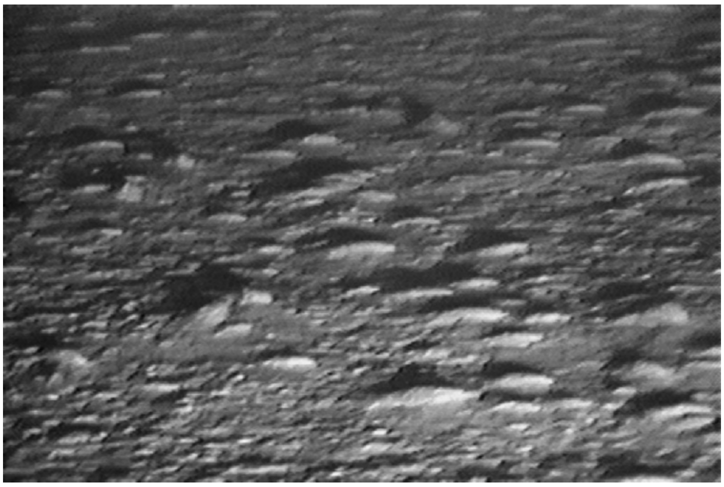


Figure 13. Paleoshoreline features of ancient mollusc shells located *in situ* on the seabed as they were in the Early Holocene, but now lying at a depth of 160 m. No samples were collected, but it is hypothesized that these were ancient mussels living in a near-shore environment prior to the inundation of the Black Sea. The still photograph comes from video collected by the submersible during the 2002 expedition. Dive location is within the southernmost survey region as shown in Figure 8. Although the photo is of poor quality, it represents the best documentation other than direct visual observation from the submersible.

4. DISCUSSION AND CONCLUSION

One of the primary goals we wished to achieve during the first years of archaeological and oceanographic surveying in the Black Sea was to map the submerged landscape on the continental shelf off the coasts of north-central Turkey, Bulgaria, and Romania in hopes of locating a datable human settlement along the ancient shores of the inundated Neoeuxinian lake. Unfortunately, that goal was not fulfilled completely; however, we did achieve an approximate delineation of the submerged pre-flood paleoshoreline in all of our survey boxes. Landward of the paleoshorelines, inundated landscapes, consisting primarily of flooded coastal geological formations, were acoustically and visually imaged and sampled, resulting in the mapping and identification of a number of paleo-river channels and other relic geological features. One unique site was located off the coast of Turkey and was mapped and imaged as representing a possible pre-flood site of human occupation. In addition to the geological and archaeological surveying, several shipwrecks were discovered, one beautifully preserved in the deep anoxic waters off the Turkish coast, another from the Hellenistic period of about 300 BC off the Bulgarian coast, the oldest ship found to date in the Black Sea. The mapping revealed a drowned landscape from a time when the region was exposed subaerially, prior to the Holocene flood. Throughout each survey box, geophysical data was collected and ground truthed by geological sampling and ROV and submersible investigations, revealing a number of interesting features associated with the submerged paleoshorelines of the Black Sea.

A significant result from our archaeological and oceanographic surveying in the Black Sea is a delineation of the submerged paleoshoreline at 155 m water depth off the coast of north-central Turkey and off the coasts of Bulgaria and Romania. Landward of the paleoshoreline, in all survey regions, portions of the inundated landscape were mapped, characterized, and sampled. In addition to the shoreline features, a number of submerged river channels and deltaic features were also delineated. Regardless of the rapidity of the marine inundation into the basin, which flooded this shoreline and surrounding environments, we conclude that the coastal landscape surrounding the shores of the Neoeuxinian lake were exposed and habitable during the recent geologic past. Whether or not the inundation was catastrophic, rapid, or more gradual does not alter the fact that many coastal archaeological sites may now lie submerged, exposed or buried on the drowned Black Sea continental shelf. Therefore, this region needs to be explored and mapped in greater detail to learn more about the ancient history of the Black Sea basin.

The discovery of Site 82 in 95 m of water about 20 km off the coast of northern Turkey in 2000 stimulated discussions by archaeologists that deep-water terrestrial sites could be preserved and found on the Black Sea shelf. More work on this site and elsewhere on the submerged shelf is needed in order to obtain conclusive results. Results of the Bulgarian survey indicate that four of the eight side-scan sonar targets were shipwrecks, and several other targets were man-made objects including what appeared to be fishing gear and trash (for example, abandoned oil drums and filled plastic bags). The lack of rocky outcrops on the Bulgarian shelf helps the preliminary identification of targets, and future archaeological oceanographic survey work will undoubtedly result in the discovery of new man-made objects on the seafloor. We conclude that the western shelf of the Black Sea holds great promise for the application of geophysical surveying and visual inspection techniques to the prospection for submerged archaeological sites. One of the problems could be high near-shore sedimentation rates potentially burying these sites, making them more difficult to locate and investigate with side-scan sonar.

The discrepancy between the depths of the paleoshorelines in the southern and western Black Sea basin (140 m vs. 155 m) can be accounted for by a number of factors. One mechanism might be differential isostatic and/or tectonic adjustments on either side of the basin. The dynamics of the earth's crust and upper mantle in northern Turkey are dominated by the Anatolian Fault, a major plate boundary that is tectonically very active. Earthquakes and faulting cause localized uplift and downwarping of the surface, which could easily account for the difference in paleoshoreline elevation. In addition, the marine inundation of the Black Sea loaded an enormous amount of water onto the exposed shelf regions, causing them to isostatically adjust by subsidence. The amount of subsidence is a function of a number of factors including the crustal thickness and amount of impacted terrain. These differences could also account for differential elevation adjustments of the paleoshoreline. We also observed a large quantity of submerged landslides off the Turkish coast that could also account for differential isostatic adjustments.

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29

NEW EVIDENCE FOR THE EMERGENCE OF A MARITIME BLACK SEA ECONOMY

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Abstract: This article argues that changes in settlement pattern from the Bronze to early Iron Age suggest a new maritime orientation in the economy of the Sinop Promontory. Bronze Age evidence (3rd–2nd millennia BC) suggests a primarily land-based economy supplemented by maritime resources, while Iron Age evidence (early 1st millennium BC) indicates a stronger maritime orientation. Long-distance overseas interaction from as early as the 5th millennium is suggested by stylistic parallels between local finds and other Black Sea coasts. Early ceramic parallels probably signal incidental contact resulting from fishing or other maritime activities. Iron Age maritime intensification encouraged coastal settlement with clearer parallels between Sinopean and overseas ceramic and architectural assemblages. Exchange and fishing are likely to have been key components of the Iron Age economy in advance of the Circum-Pontic economic expansion brought about by Greek colonization in the second half of the 7th century BC.

Keywords: Sinop, Cimmerians, exchange, Black Sea economy, Bronze Age, Iron Age, settlement patterns

1. INTRODUCTION

In their book entitled *Noah's Flood*, Ryan and Pitman (1998:165ff.) suggest that the attitudes of local populations toward the Black Sea may have been profoundly affected by the trauma of the flood. The process, timing, and cultural context of the post-glacial infilling of the Black Sea (ancient "Pontus") to its current level may yet be incompletely known, but understanding the diverse and dynamic relationships of local populations with the sea is essential to constructing a history of cultural and economic development in the region. The Sinop Regional Archaeological Project (SRAP) is an ongoing inter-

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disciplinary case study of the relationship between local communities and the varied inland, coastal, and maritime environments on the south coast of the Black Sea (Figure 1).

To its south, the Sinop Promontory is cut off from mainland Anatolia by the forbidding Pontic Mountains, and because of the difficulty of passage through these mountains, inhabitants of the promontory have remained either isolated or oriented toward the sea. The distinctive ecology and rugged physical geography of the southern and eastern coasts of the Black Sea, where high rainfall and mild temperatures prevail, contrast sharply with the dry, lowland steppes of the northern side (Figure 2). These contrasting environmental conditions have given rise to the production of different resources, and under favorable economic circumstances, this has in turn stimulated intensive trade (in particular, during the Hellenistic, late Roman, and Ottoman periods).

Figure 1. The Black Sea, indicating places named in the text.

The diversity of cultures and available resources along the coasts of the Black Sea provided strong incentive for exchange. The Romanian historian Bratianu called the Black Sea a "plaque tournant," a turntable around which the great cultures of ancient and modern times revolved (Bratianu 1969:43). Economic and political patterns within the region have persisted over millennia between the great trading and nomadic world of the Eurasian steppe, the powerful military empires of Mesopotamia and Anatolia, and the rich mercantile societies of the Mediterranean. Although these cultural adaptations have been stable throughout antiquity, interactions between the individual cultures have been more variable and dynamic. Trade among Pontic communities has generally flourished in the context of *heterarchical* political economies (Crumley 1994), when political and economic power are dispersed, and trade is managed by specialized networks of agents acting as a trade diaspora (Stein 1999). From Scythians to Russians, Hittites to Ottomans, the major political axis of the sea has been oriented north-south.

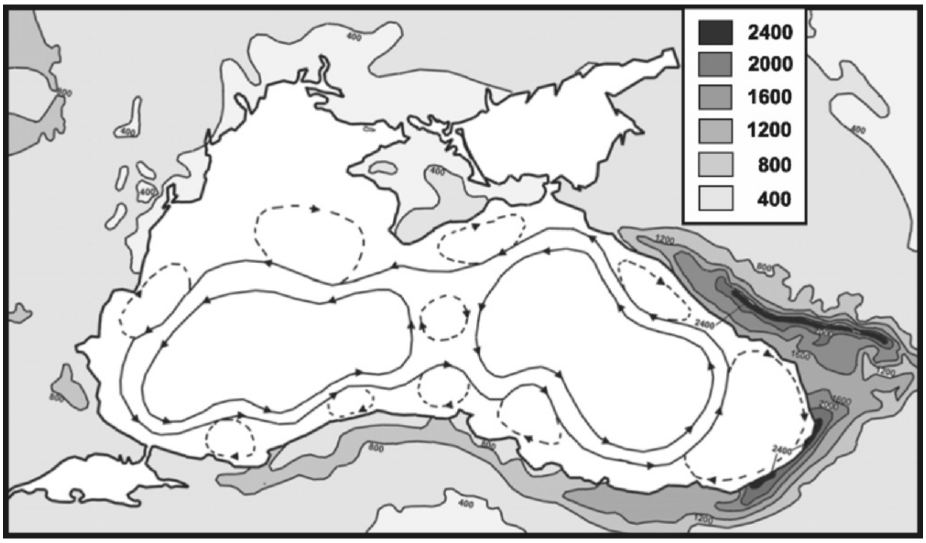

Figure 2. Rainfall and surface current circulation of the Black Sea region. Rainfall distribution is indicated in mm/yr at 400 mm intervals varying from light gray (low rainfall) to dark gray (high rainfall); data from Steinhauser (1970). Primary surface currents are indicated in solid lines, and secondary currents in dashed lines.

Since the Greeks of Miletus colonized a necklace of colonies around the shores of Pontus in the 7th century BC, Mediterranean cultures have exerted a powerful economic pull on the region. Mediterranean agents (Greeks, Romans, Genovese, Venetians) have often established closely networked enclaves of traders in major Pontic ports, in which linguistic, cultural, and familial bonds enabled stable exchange relationships to emerge. In the few brief episodes in which a single political power dominated north and south coasts (Roman and Ottoman), the connection with the Mediterranean through the Thracian Bosphorus was critical. Exercising political dominance over the region took extraordinary efforts, and even under Roman domination, the east was not completely under imperial control (Braund 1994).

Several themes wind through the long history of maritime economy in

Sinop (Doonan 2004a).

(1) There is little evidence that trade sustained local economies before the 1st millennium BC.

(2) Local engagement with the sea over the long term tends to ebb and flow in cycles rather than evolve over time.

(3) Exchange was incidental to regional interaction and not primarily responsible for it before the establishment of a Greek colonial trade diaspora in the 7th century BC.

2. THE BLACK SEA TRADE PROJECT

In 1996, we initiated the Black Sea Trade Project (BSTP) to study the communities on the Sinop Promontory and their integration into the dynamic Pontic political economy over the past 5000 years (Hiebert *et al.* 1998). We combine the study of terrestrial cultures, their production and demography, with an investigation of submerged landscapes and seascapes in a unique multi-disciplinary project (see Ballard *et al.*2001; Coleman and Ballard, this volume, on the underwater component of the project). We chose the Sinop Promontory as the focus of our case study because Sinop holds a unique strategic, economic, and cultural position in the Black Sea. It is situated at the point where two opposing gyres in the surface currents flow north and south, assisting sailors on the difficult and dangerous deep-water crossing. The multidisciplinary Sinop Regional Archaeological Project (SRAP) provides historical, cultural, and demographic context for the underwater research.

A systematic archaeological survey of the hinterland around Sinop ties together the various components of SRAP. The Sinop Promontory, like the majority of the south coast, poses many challenges to systematic archaeological survey. Total areal coverage has been considered by many archaeological surveys to be the preferred approach to regional study (Fish and Kowalewski 1990), but because the Sinop Promontory is so large and diverse, we have employed a sampling program (Doonan 2004b). To date, we have conducted systematic surveys in the Demirci and Karasu valleys, Boztepe, and the upland Sorkum area, recording almost 180 archaeological loci in almost 400 field tracts (Figure 3). In the field, we line up about a dozen fieldwalkers at ten meter intervals; they proceed along a transect and collect all of the archaeological evidence they find in one-meter swaths. This evidence is carefully recorded, including counting, weighing, and photographing of all ceramic finds. Evidence for the date of occupation is provided by study of ceramic types. We are currently developing a program to date our field ceramics directly by thermo-luminescence and to establish chronology and provenience of local ware-types based on physical characteristics (Bauer 2002; Doonan and Bauer 2005). This will make even the humblest scraps of pottery useful to our assessment of production and exchange throughout the Sinop hinterland. Doonan (2004b) contains further discussion of the methods employed by the survey, including excavations as well as geomorphological and remote sensing studies.

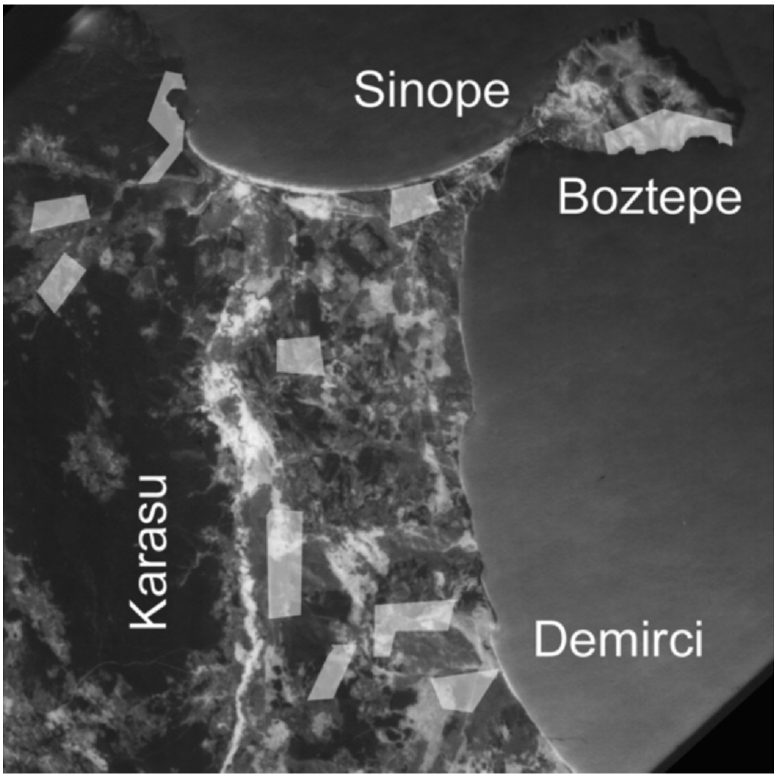

Figure 3. CORONA photograph of the Sinop Promontory with sample quadrats in the Demirci and Karasu valleys and Boztepe highlighted. Dark areas are heavily forested offering almost no visibility.

2.1 Emergence of the Black Sea Economy: the View from Sinop

SRAP has documented a sequence of occupation at Sinop going back more than 10,000 years. Over the millennia, communities chose different settlement locations based on the availability of resources. This discussion examines three settlement categories, based on the relationship of a settlement's catchment area (the space surrounding the settlement that is likely to have been most important for the gathering of resources critical to the function of the economy) to the sea:

(1) inland (> 2 km from the sea), with full terrestrial catchment areas in a 2 km radius,

(2) near-coastal (< 2 km of the sea), with potential access to maritime and terrestrial catchment areas, and

(3) coastal (right on the coast), with significant limitations of access to terrestrial resources.

This paper focuses on the early formation of a Black Sea interaction system, and so the evidence under consideration will be limited to the Paleolithic (pre-10,000 BC) through late Classical (4th century BC) times (Table 1). The patterns of interaction that emerged in these early phases have continued to play an influential role in subsequent developments, which are discussed elsewhere (Doonan 2004a).

Table 1. Major periods and cultural trends on the Sinop Promontory, Neolithic through Greek Classical periods.

Period	Dates (BC)	Cultural Trends and Exemplary Loci
Paleolithic-Chalcolithic	>10,000 – 2700	Some presence noted (Inceburun, Mezarliktepe, Maltepe) Possible evidence of Neolithic contact with west Pontus (Ilıpınar), Karanovo VI (Bulgaria) Data too scarce to establish clear cultural and settlement patterns
Early Bronze Age	2700 – 2000	Widespread settlement inland (Gulluavlu) Some settlement in locations with potential for terrestrial-maritime exploitation (Kocagoz) Ceramics parallel Ikiztepe (Samsun-Turkey), Troy II (NW Turkey), Catacomb Culture (Ukraine)
Middle-Late Bronze	2000 – 1000	Settlement similar to Early Bronze Expanding contacts with central Turkey seen in ceramics and Hittite references to Kashka tribes Possible early establishment of coastal settlements (Kösk Höyük-Gerze).
Iron Age	1000 – 650	Expansion of coastal settlements (Sinop Kale NW, Fener-Akliman, Kösk Höyük-Gerze) Clear connections between Sinop Kale NW and northern Black Sea (Berezan)
Greek Archaic	late 7th – early 5th century	Foundation of Sinope and network of colonial dependents Very limited relationship between Sinope and hinterland
Greek Classical	early 5th – late 4th century	Athenian cleruchy founded at Sinope Secondary port (Harmene) Datames's siege of Sinope, loss of eastern colonial network Mass production of Sinope amphoras, export to northern Black Sea

Evidence of human presence on the Sinop Promontory during the Paleolithic has been documented at Inceburun, the northernmost point on the Anatolian landmass (Isin 1998; Hiebert 2001). Inceburun seems misleadingly "coastal" today, yet it was an upland site at the time of its occupation (Hiebert 2001). Any evidence of pre-Holocene maritime or coastal settlement must be found through underwater research, as discussed in Ballard *et al.* 2001. Very little information can be gleaned from Inceburun because of the conditions of the remains; finds were recovered from severely wind-eroded sandy deposits (Akkan 1975). The paucity of similar finds from other parts of the Sinop Promontory affords little additional evidence.

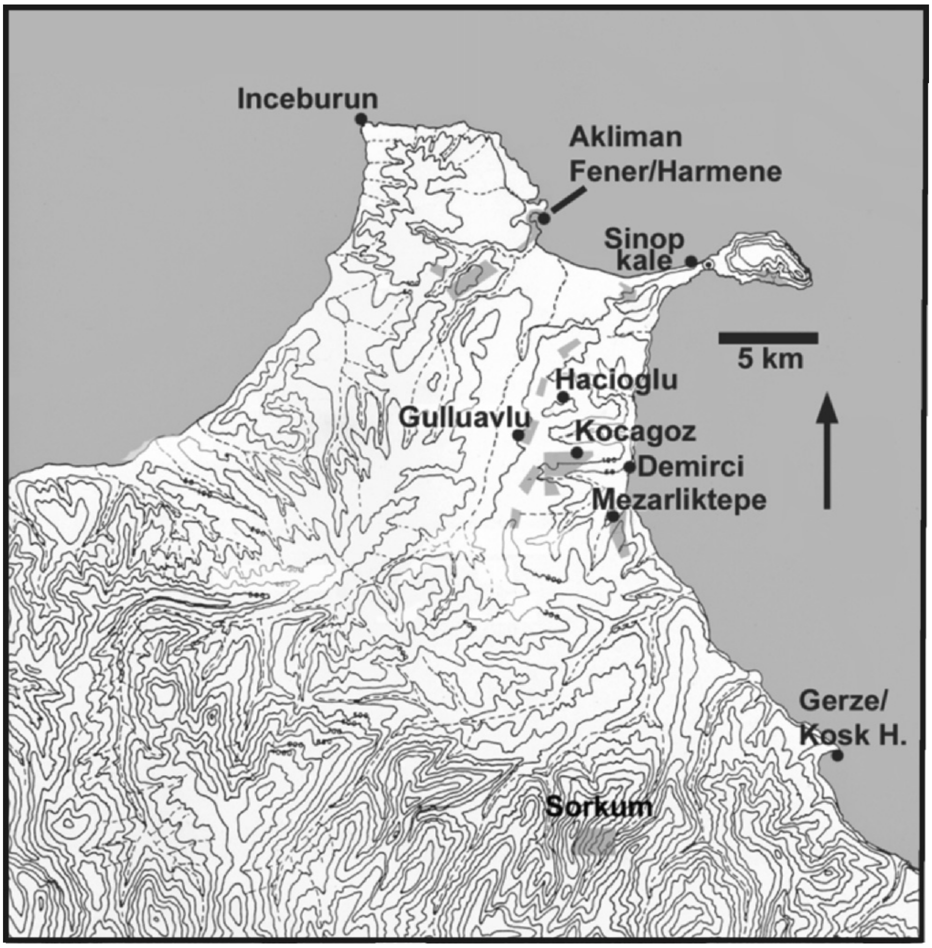

Figure 4. The Sinop Promontory indicating places named in the text.

Neolithic and Chalcolithic settlement data suggest some interest in maritime resources, but the evidence is insufficient. The site of Mezarliktepe near the coast of the Demirci Valley was sited to take advantage of a likely estuary, although it is sheltered from the coast by the Keçioğlu ridge. This kind of siting, near but not directly on the coast, is similar to pre-Bronze Age settlements in the Sea of Marmara, such as 6th millennium Ilıpınar (Roodenberg

2001), and Ikiztepe in the Kızılırmak River delta east of Sinop (Alkım *et al.* 1988). The ceramic assemblage at Mezarliktepe also reveals similarities to wares of the Marmara and Pontic regions (Doonan *et al.* 1999). Incised wares with pierced lug handles resemble those at Ilıpınar (Roodenberg 1995) and early Ikiztepe (Bilgi personal communication; see Alkım *et al.* 2003). Patterns of early settlement and contact will be better understood through the research currently undertaken by Alex Gantos.

Settlement appears to have become widespread in the Early Bronze Age, i.e., from the mid 3rd millennium onward. Settlement remains sparse on the coast during the EBA, but it expanded in areas within sight of the coast and inland. Alex Bauer is studying the transfers of technology between these settlements and those on other coasts of the Black Sea. Sites are typically mounds (but not necessarily stratified höyüks) placed in positions overlooking transportation routes, like Kocagoz and Gulluavlu (Figure 4). Assemblages show connections to other areas around the Black Sea. The assemblage at Kocagoz, for example, shows connections particularly with settlements to the east and south, like Ikiztepe, while the Gulluavlu assemblage points south, west, and north, towards central Anatolia, Troy, and the northern Pontic coast. Both Kocagoz and Gulluavlu were situated far enough from the coast that they had access to ample terrestrial catchment areas, yet the extensive parallels of their ceramic assemblages suggest diverse connections. Although both are inland sites, they strongly suggest extensive interaction around the Black Sea. This does not necessarily imply a maritime-oriented economy, however.

Just about the turn of the 1st millennium, a new kind of settlement started to appear on the Sinop Promontory. These settlements were located right on the coast with little or no clear terrestrial catchment. Just beneath the north-west corner of the Sinop Kale, a multi-phase locus was observed eroding from an exposed scarp during the 1997 season. Hiebert directed an excavation of the scarp in 2000. The ceramics recorded in the survey at Sinop Kale NW show remarkable similarities to sites in Crimea and Ukraine, notably Berezan (Solovyov 1999; Doonan nd). The stone architecture at Sinop Kale also shows close similarities to that of pit houses from the north coast (Solovyov 1999; Kuznetzov 1999; Tsetskhladze 2004). During the 2003 season, SRAP documented similar architecture at Kösk Höyük in Gerze, which we hope to investigate further in coming seasons (Doonan and Bauer 2005).

A famous passage in Herodotus mentions that there was a north Pontic "Cimmerian" foundation at the site of Sinop:

> The Cimmerians in their flight from the Scythians into Asia also made a colony on the peninsula where the Greek city of Sinope has since been founded. (IV.12.2)

The Cimmerians are a shadowy and ill-defined ethnic group from the early 1st millennium, who are best known for their invasions of Anatolia (Ivantchik 2001). There is nothing to suggest that they were a maritime power or that these horsemen loaded their animals and weapons onto ships to sail across the Black Sea. No more likely is the possibility that they marched over the forbidding Pontic mountains to found a colony at the site of Sinop. A more plausible explanation might be as follows. Greek colonists arriving at Sinope in the late 7th century BC would have observed ruins of a village with the same kind of pottery and architecture that their contemporaries were seeing on the north coast and explained these remains as belonging to Cimmerians (already shadowy and mythical in the north as well as the south). Further confirmation for this theory would have been available from locals, to whom the inhabitants of the coastal village would have been outsiders (the material culture associated with this site is distinct from local traditions). This myth was eventually passed down to Herodotus who included it in his history of the place.

The coastal settlements at Sinop Kale and Kösk Höyük highlight a remarkable development in the history of the Sinop Promontory. These are the first settlements located directly on the coast in places that would have necessitated a maritime-oriented economy. Fishing and trade are the most significant economic activities at the water's edge in Sinop. It may well be that both of these drew the interest of settlers from the north coast. It may even be that the first crossings were made by northern fishermen taking advantage of the seasonal circulation of fish stocks around the Black Sea. Bonito and sardines are most prominent among a host of valuable fish species that migrate around the Black Sea each year in a predictable pattern (Doonan 2002). If these coastal settlements were seasonal fishing camps, these more adventurous patterns of fishing may lie behind the pattern of intensified contacts in the western half of the Black Sea at the end of the 2nd and beginning of the 1st millennium BC (Vanchugov 2001; Pienazek-Sikora 2003). In a recent review of evidence for Greek colonization in the Black Sea, Tsetskhladze (1994) rightly emphasizes the western Black Sea as critical to understanding Cimmerian settlement and the foundation of Sinope, following the Roman geographer Strabo I.3.21.

The founding of the colony at Sinope by Greeks from Miletus was highly significant in the development of the integrated Black Sea economy, although perhaps not the watershed in Black Sea history it has normally been understood to be (Doonan 2004a, nd). The existence of widespread maritime connections among Black Sea communities represented by the coastal sites in Sinop province implies that the Greeks moved into a region where long-distance interactions were already developing. The Milesians established a colony at Sinope near the site of the modern port around 630 BC (Akurgal 1956; Boysal 1959). According to Greek sources, the foundation of Sinope was followed quickly by a succession of colonial foundations connecting a path to the fabled metal resources of the eastern Black Sea (Graham 1982). In myth, the nymph Sinope resisted the persistent advances of Apollo, the Greek god associated with colonization, and Zeus, king of the gods (Ps-Skymnus 986-87, Apollonius Rhodicus 944ff.; Ivantchik 1998). Sinope's brave resistance to the gods' overtures may mythologize a pattern observed by SRAP: almost no settlements in the hinterland show evidence of Greek interaction before the fourth century BC (Doonan 2004a). The hinterland was clearly not connected to the port for over two centuries, when amphorae produced near the port and imported ceramics begin to be evident along the coasts and in highland areas of the promontory (Doonan 2004a). This engagement is almost certainly linked to the disruption of colonial trade networks in particular by the satrap (Persian governor) Datames, who was carving out a local territory for himself at the expense of his Persian masters during the second quarter of the fourth century (Doonan 2004a). Sinope resisted Datames's incursions, but the port's relationship to its colonies in the eastern Black Sea was severed, prompting an expansion of production in the hinterland on the promontory. For the next millennium, the hinterland on the promontory was closely connected to the Black Sea regional economy.

2.2 Settlement Trends in the Sinop Promontory

Considering the broader patterns that have emerged so far in our systematic survey, we can see a general predominance of settlements away from the coast before the Iron Age (Figure 5). The extensive Bronze Age settlement in near-coastal and inland areas must represent economies largely oriented towards terrestrial resources. The tantalizing possibilities of long-distance interaction represented by similarities in ceramic style and technology are as yet difficult to explain, given the lack of evidence for coastal settlement. It is possible that a certain amount of exchange was incidentally related to fishing or other maritime activities. Current evidence for Iron Age settlement on the coast suggests a dispersed pattern, although we have only a limited sample of systematically investigated coastal areas. Also, the distinctive appearance of settlements in port locations outside the systematic survey areas (for example, Kösk Höyük and Sinop Kale NW) together with the strong trans-Pontic influences evident at these sites suggests the emergence of a significant maritime component in the Sinop economy on the eve of Greek colonization.

The Greeks from Miletus established a colony at Sinop in the late 7th century BC, probably not long after the abandonment of the Sinop Kale NW site. They named the town Sinope, and their colony was the jumping-off point for an extensive colonization of the region. Our investigations at Sinop Kale NW suggest that the Greek colonial process did not suddenly bring long-distance communications to the Black Sea, it expanded pre-existing communications networks in the Black Sea. The network of colonies and the colonial diaspora established by the Greeks structured subsequent interactions around the Black Sea for more than a millennium.

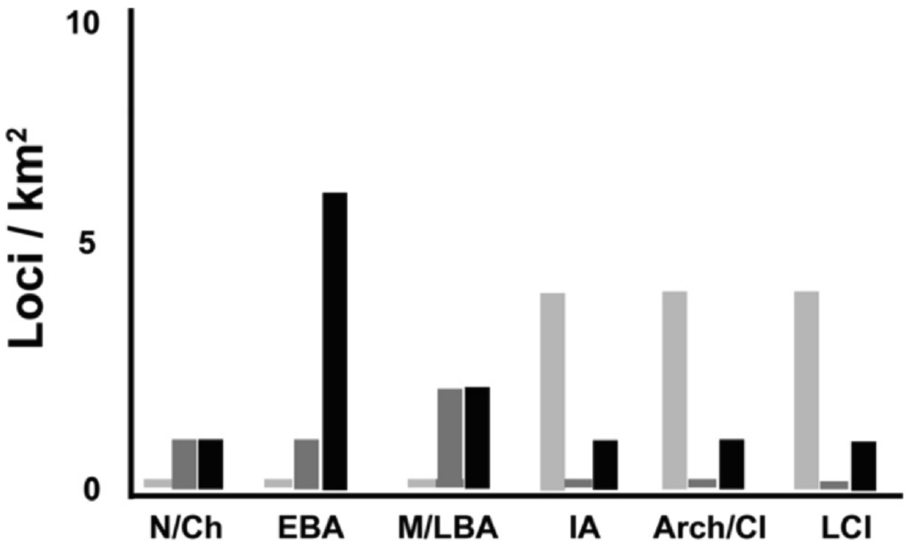

Figure 5. Frequencies of archaeological loci per square kilometer in coastal (light gray), near-coastal (darker gray) and inland (black) catchment areas. Only areas investigated using a systematic research design are included.

ACKNOWLEDGMENTS

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30

HOLOCENE SEA-LEVEL CHANGES OF THE BLACK SEA

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Abstract: This article summarizes the changes in sea level and salinity of the Black Sea during the Holocene epoch. Paleogeographical reconstructions are based on detailed geological surveys of the shelf and Holocene terraces in the Kerch-Taman region and the Caucasus littoral, as well as the northwestern Pontic area. The chronological framework derives from the comprehensive database of radiocarbon dates obtained from Black Sea sediments in the USSR and Russia. These data are presented systematically here for the first time in Appendix I at the back of the volume.

Keywords: Black Sea, Caucasus Black Sea coastal area, northwestern Pontic area, Holocene, paleogeographical reconstructions, radiocarbon dates

1. INTRODUCTION

Sea-level change in the Pontic basin over the past 20,000 years has been discussed in many publications (Nevessky 1961; Nevesskaya 1965; Ostrovsky 1967; Fedorov 1977; Shcherbakov *et al.* 1977; Shilik 1977; Ostrovsky *et al.* 1977; Shnyukov *et al.* 1981; Dzhandzhgava 1979; Shnyukov *et al.* 1981; Dzhanelidze 1980; Balabanov *et al.* 1981; Balandin *et al.* 1982; Dzhandzhgava *et al.* 1982; Izmailov 1982; Palatnaya 1982; Voskoboinikov *et al.* 1982; Molodykh *et al.* 1984; Balabanov and Izmailov 1988). Much consensus has recently been reached among Russian scholars regarding general trends in the evolution of the basin, and most share the view that the large-scale regressive-transgressive cycle was linked to Late Pleistocene (Late Valdai) climatic cooling and Holocene warming. There are no major differences in the assessment of the timing and scale of the maximum sea-level drop, as well as the character and rate of the Holocene sea-level rise. Differing opinions exist, however, over the time and scale of the smaller fluctuations superimposed upon the Holocene trans-

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gression, the salinity of the basin, the sequence of orictocenoses (localized fossil assemblages), and the first appearance of Mediterranean fauna after the brackish Neoeuxinian stage.

2. MATERIALS AND METHODS

Abundant evidence relating to the Late Pleistocene and Holocene history of Black Sea landforms has emerged from a large-scale geological survey conducted in the Caucasus littoral, the Kerch-Manych region, and the north-western Pontic area, most notably in the areas of Kobuleti, Sukhumi, Novyi Afon, Gudauta, Pitsunda, Gagra, Adler, Lazarevskoye, the Kerch Strait, the Black Sea, the Kuban River Deltas on the Azov Sea, and the northwestern Pontic limans/lagoons (Figure 1). Descriptions of the methods and most of the results have appeared in several earlier publications (Ostrovsky *et al.* 1977; Balabanov and Gey 1981; Balabanov *et al.* 1981; Arslanov *et al.* 1982; Izmailov 1982;

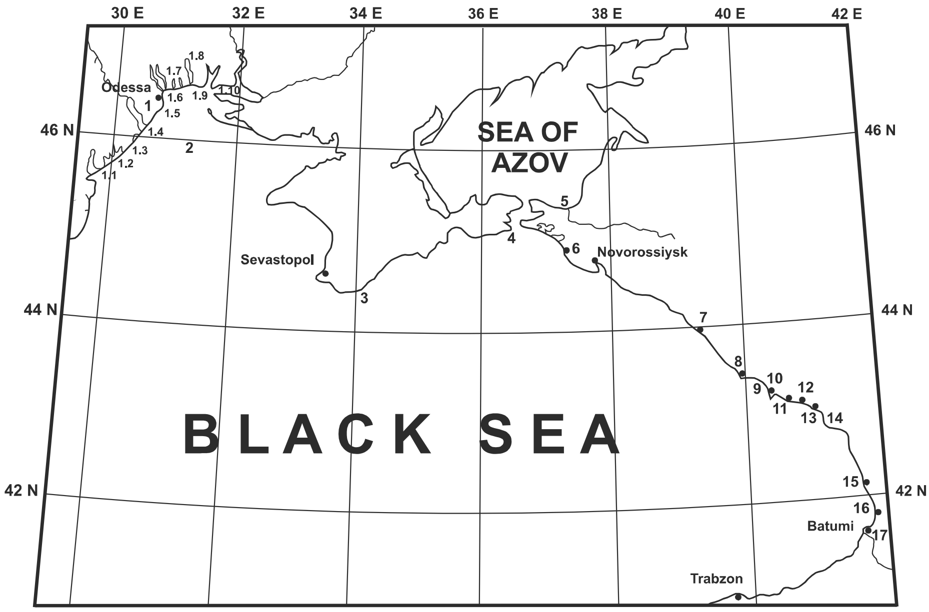


Figure 1. Study area. **Key:** 1 = Limans of the northwestern Black Sea coastal zone: 1.1 – Shagany, 1.2 – Alibey, 1.3 – Budakskii, 1.4 – Dniestrovskii, 1.5 – Khadzhibeyskii, 1.6 – Kuyalnitskii, 1.7 – Bolshoi Adzalykskii, 1.8 – Tiligulskii, 1.9 – Berezanskii, 1.10 – Dniepro-Bugian; 2 = northwestern Black Sea shelf; 3 = southern Crimean shelf; 4 = depth of Kuban River; 5 = Kerch Strait; 6 = Anapa shore; 7 = Lazarevakoe shore; 8 = Adler shore and shelf; 9 = Gagra shore and shelf; 10 = Pitsunda shore and shelf; 11 = Gudauta shore and shelf; 12 = Novyi Afon shore and shelf; 13 = Sukhumi shore and shelf; 14 = Kodori River Delta; 15 = seaside of Colchis Plain; 16 = Kobuleti shore and shelf; 17 = Batumi shore and shelf.

Nesmeyanov *et al.* 1987; Balabanov and Izmailov 1988; Izmailov *et al.* 1989).

The survey incorporated intensive geologic studies of the Holocene terraces and off-shore shelf of the Black Sea, including: (1) identification of stratigraphic units and genesis of the types of Holocene sediments, together with their lithological composition and facies variability in stratigraphic sequences as well as along the strata; (2) determination of the thickness of stratigraphic-lithological units; and (3) paleogeographical reconstructions of the marine depositional landforms, identification of their main features, and assessment of shoreline migration rate. Surveyors also carried out a high-precision correlation of ancient shorelines with the various stages of Holocene transgression. Special attention was paid to paleogeographic reconstructions, vertical and horizontal positions of ancient shorelines, as well as their instrumental locations.

The examination focused on the precise correspondence of ancient shorelines within tectonically distinct areas in order to assess the altitudinal impact of recent tectonic movements. Correlations were obtained through exact dating of peats formed in coastal lagoons and fossil mollusc assemblages. which are diagnostic features for shoreline migration.

The project was carried out with: (1) topographic survey, (2) coring and sondaging, (3) bathymetry of the shelf, (4) geophysical survey of the shelf using multibeam sonar and seismic-acoustic profiles, and (5) sampling for macro- and microfauna as well as pollen analysis, and the plotting of respective diagrams.

In the key coastal areas, 10–12 geological cross-sections were obtained, each comprising 6–10 cores into the sea floor, 3 or 4 being extended to depths of 30–40 m. Inshore cores normally exposed the totality of Holocene sediments. Cores on the sea floor that penetrated 25 to 30 m deep usually captured the total Holocene sequence, or, in rare cases, 60–70% of their total thickness.

The recovered cores were sampled at intervals of 0.25–0.5 m (rarely 1.0 m) for grain-size study as well as macro- and micropaleontological analysis. Chronological control was obtained via radiocarbon dating of either fossil molluscs (from marine sediments) or peat samples (from terrestrial sediments). In some instances, radiocarbon dating was performed on wood samples associated with the enclosing sediments.

Radiocarbon dating was conducted in various laboratories and both intra- and extra-laboratory cross checking was carried out whenever possible to minimize instrumental error. Dates are listed in Appendix I at the back of the volume, with the ^{14}C labs indicated in an associated table endnote.

3. OBJECT OF STUDIES

The sediments of the Chernomorian horizon (Q_{III}^{4}-Q_{IV}), formed during the Holocene transgression, are widespread on the Black Sea shelf and in the lower reaches of the larger rivers. Their complete 80–90 m thickness was studied by the writer in the Pitsunda-Gagra area.

The stratigraphic complexity of the Chernomorian horizon on the terraces and shelf stems from:

(1) The polygenetic character of the sediments, which include lithologically distinct marine, lagoonal, riverine, and paludine facies, the contours of which were subject to repeated modification by periodic transgressive and regressive episodes, all against the background of a major transgression.

(2) A high sedimentation rate, averaging 1–2 m per century during the last 12–13 thousand years, and reaching 15–20 m per century in areas of intensive deposition. This has resulted in the metachronous character of lithological contacts and facies zones, which have experienced repeated spatial migration (the "facies rule" of Valter-Golovkinsky, see Lisitsin 1984).

(3) An irregular distribution of molluscan and foraminiferal assemblages in the sediments as well as spatial displacement of the orictocenoses due to variations in depth and salinity of the Late Pleistocene-Holocene basin (Nevesskaya 1965). This makes the stratigraphic use of such assemblages difficult. Different sedimentary facies may include similar orictocenoses, and similar orictocenoses may often be encountered in sediments of various ages, both in sequence and in space. These occurrences make the Holocene biostratigraphy rather complicated. At the same time, a proper study of orictocenoses is fundamental for paleoenvironmental reconstructions.

To date, geological investigation of the Holocene terraces along the Caucasian littoral has failed to provide universally acceptable criteria for a detailed stratigraphic subdivision of Late Pleistocene and Holocene sequences. In addition, there is no generally accepted methodology for such investigation.

Stratigraphic subdivision of the Chernomorian horizon followed the paleogeographical approach of Zhizhchenko (1959), which was applied to Holocene sediments in the Black Sea by Balabanov *et al.* (1981). This approach involves analysis of the sequence of sedimentogenesis and paleoecology within the basin in relation to general physical and climatic changes and is based on high-resolution facies-lithological and paleontological study. The bio-facies regularities found for various paleogeographical units were interpreted in rhythmic-stratigraphic terms, and paleogeographic episodes were dated using ^{14}C measurements and evidence from archaeological or historical sources.

In view of the complexities of both facies zonality and paleontological distributions in the New Black Sea sediments, stratification and correlation were carried out for four principal sedimentation types: marine, riverine, subaerial, and lacustrine-paludine. Interpretation of Holocene geological evidence in the coastal areas of the Caucasus ultimately permitted the development of spatial-temporal models, including detailed isochronous surface projections for the transgression maxima and regression minima (Figure 2).

4. RESULTS

The survey compared Holocene shoreline elevations from various structural-tectonic units of the eastern Black Sea and Azov Sea coastal areas, ranging from the stable and submerging West Kuban trough to the equally stable but uplifting Caucasus littoral. It found that the elevations of synchronous shorelines have remained basically constant. In many cases, the magnitude of tectonic deformation remained within the bounds of acceptable error given the lithological-facies and geomorphic criteria used to estimate shoreline elevation. Hence, the rate of recent tectonic deformation in the investigated area during the Late Pleistocene and Holocene corresponded to the average rate determined from the deformation of marine terraces, that is, less than a few meters over 10,000 years.

The investigations failed to support the view of Serebryanny (1978, 1982) and other scholars, who have argued that, due to local tectonic movements, the synchronous shorelines in various Black Sea coastal areas vary considerably in elevation. One may with justification plot a regional curve of Black Sea water level fluctuations based on the geologic spatial-temporal models developed for key areas of the Caucasus littoral: Kobuleti, Sukhumi, Novyi Afon, Gudauta, Pitsunda, Gagra, Adler, Lazarevskoye, and Kerch-Taman. In each of these areas, reliable geochronological data (Table 1) were used, while local peculiarities were taken into account. Much evidence was also obtained from prolonged investigations of the northwestern Pontic limans and shelf (Figure 3).

Based on the combined evidence, three types of rhythmic fluctuations are identifiable in the paleogeographic record of the Black Sea. The first type includes cycles related to major periodic climate change on the order of Ice Age glacials and interglacials. The present article deals with the Holocene section of this series only. Shorter fluctuations within these intervals correspond to eustatic cycles; those lasting a few thousand years are labeled *stages*, and those with a duration of a few centuries are called *phases*. The regression minima are considered temporal boundaries separating the stages and phases.

The first of the two Holocene stages corresponds to the initial transgression (11.0–6.0 ky BP), which was marked by a sustained rise in sea level related to the wave of Holocene warming. The sea rose to nearly its present level (–3.5 to –4.0 m), and salinity increased gradually (from 1.5–2.0 to 15–16‰). The final phase of this stage witnessed maximal ingression of the sea into the valleys of major and minor rivers throughout the Black Sea littoral.

The second Holocene stage corresponds to the transgression maximum and covers the last 6000 years. During the initial phase of this stage, sea level reached its present-day position, and according to majority opinion, it rose yet higher, to +1 or +2 m, during a later phase, 3–5 ky BP. Also at that time, Black Sea salinity reached its current values, fluctuating between 18 and 22‰.

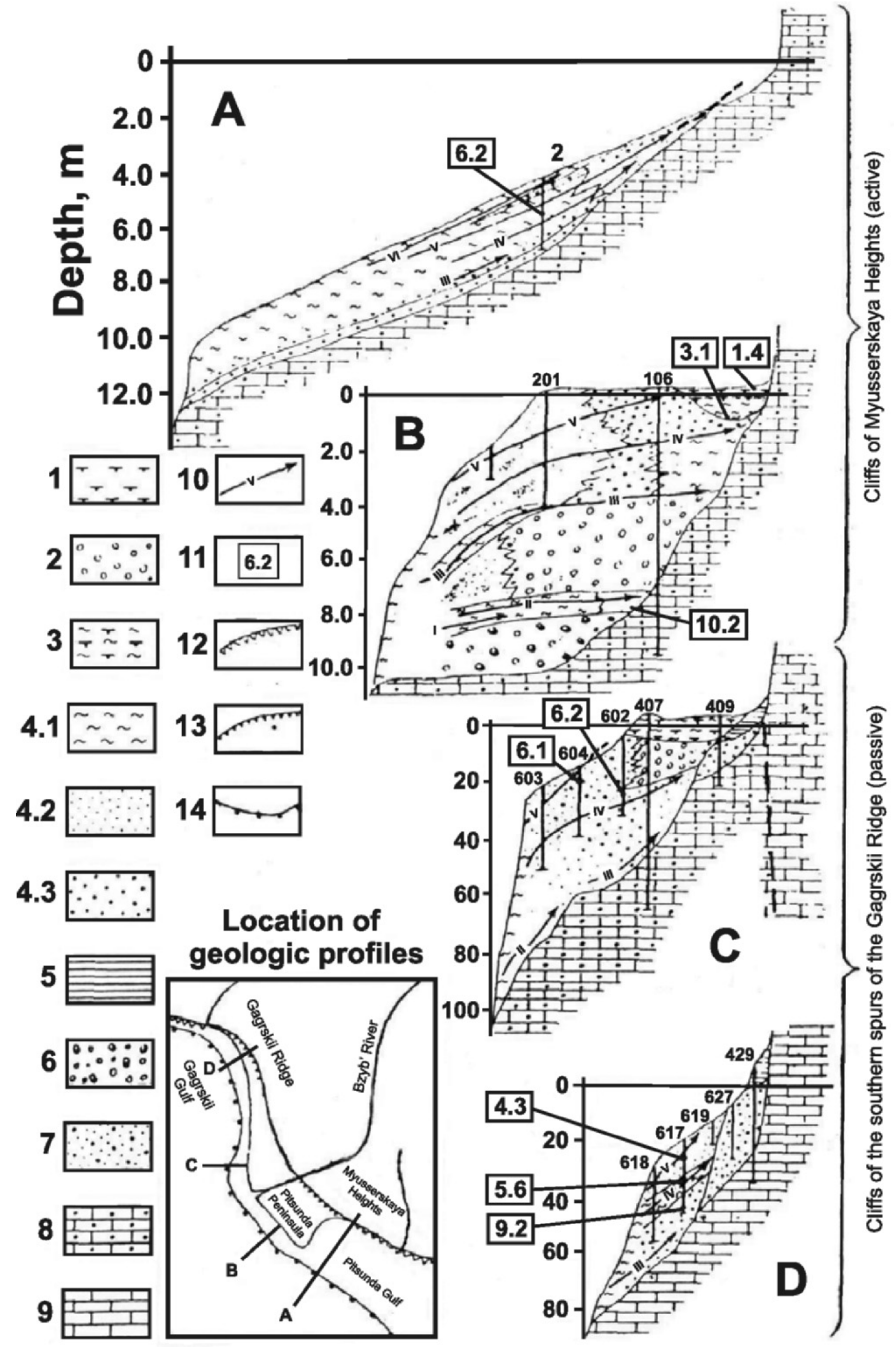

Depth, m
A
B
C
D
Cliffs of Myusserskaya Heights (active)
Cliffs of the southern spurs of the Gagrskii Ridge (passive)
Location of geologic profiles
Gagrskii Ridge
Gagrskii Gulf
Bzyb' River
Pitsunda Peninsula
Myusserskaya Heights
Pitsunda Gulf

Figure 2. (Facing page) Geological sections and scheme of Pitsunda-Gagra shelf (Dzhandzhgava *et al.* 1982). **Key:** 1–4 = Holocene (Q_{IV}): 1 – biogenic sediments and peat, 2 – alluvium, pebbles, gravel, boulders, 3 – lacustrine alluvium, muddy clay often with peat, 4 – marine sediments (4.1 – muddy clay, 4.2 – sand, 4.3 – gravel, gravel-pebbles, pebbles); 5 = Upper Pleistocene-Holocene (Q_{III-IV}): deluvial loams, loams; 6–7 = Upper Pleistocene (Q_{III}): 6 – alluvium, pebbles, boulders, 7 – marine sediments (sand, gravel, and pebbles); 8 = Neogene (N): conglomerates, sandstone, clay; 9 = Lower Cretaceous (K_1): limestone; 10 = Stratoisochrone: I – missing?, II – 10.2–9.8, 8.3–8.2, 16.5 ky BP, III – missing?, VI – 1.6–1.5 ky BP (dotted line shows part of isochrone eroded by the following transgression), V– 4.0–3.9 ky BP; 11 = Radiocarbon age (uncorrected); 12 = passive cliff; 13 = active cliff; 14 = shelf edge (isobath 100–110 m).

4.1 Holocene Stage 1

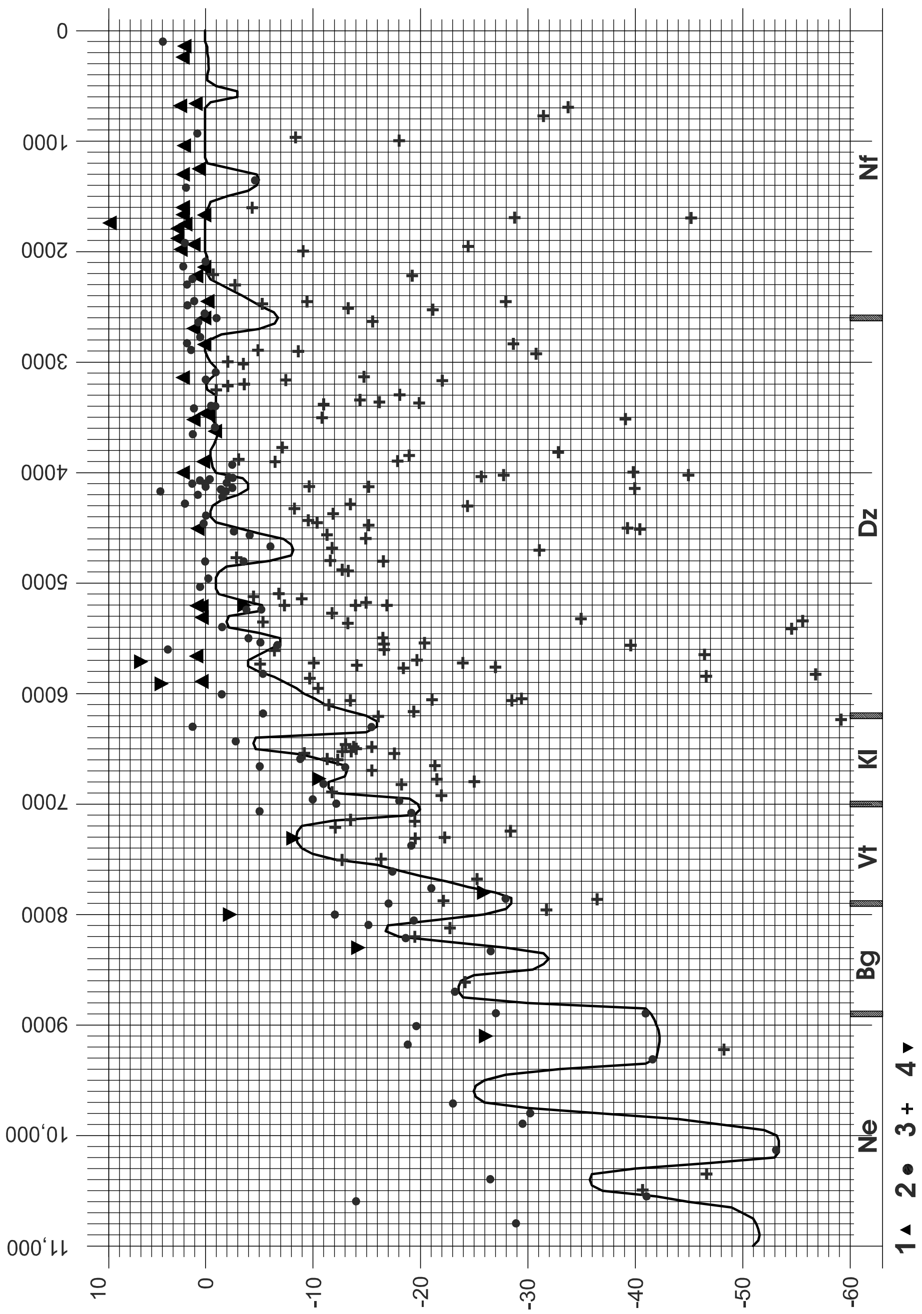
The first Holocene stage is considered a transition from the regressive to the transgressive state. Based on foraminiferal evidence and essentially following L.A. Nevesskaya's (1965) scheme, it is subdivided into four phases: Neoeuxinian (Ne), Bugazian (Bg), Vityazevian (Vt), and Kalamitian (Kl) (Ostrovsky *et al.* 1980; Balabanov *et al.* 1981; Balabanov and Izmailov 1988).

4.1.1 The Neoeuxinian Phase

The Neoeuxinian phase (ending at 9.6 ky BP) featured the proliferation of typical Neoeuxinian fauna (*Monodacna caspia*, *Dreissena polymorpha*, *Micromelania caspia*, etc.) in the basin's peripheral areas. This phase started with a deep 'Post-Yenikalian' regression and shows a complex pattern of rhythmic sea-level fluctuations.

Our investigations have clarified the shoreline position during this phase. Complete sequences of Neoeuxinian sediments with typical fauna outcrop in the Kerch-Taman area and along the Caucasus littoral. They have also been exposed in cores at depths of –20 to –25 m. Sediments belonging to earlier Neoeuxinian oscillations were examined in great detail in numerous northwestern Pontic limans (Molodykh *et al.* 1984). The importance of these studies is that dated Neoeuxinian sediments lay in close proximity to the shoreline. A series of measurements was also obtained in the area of the Kerch Strait (Kovalyukh *et al.* 1977). These studies clearly demonstrate that in the time span of 21.5–10.7 ky BP, sea level rose at least three times to –21 to –25 m, the most probable ages of the transgressive peaks being 12.2, 11.6, and 10.9 ky BP. Each peak was separated by shallow regressions not exceeding a few meters in magnitude (Balabanov and Izmailov 1988).

Following an insignificant sea-level rise, indications of a deep regression become apparent in early Holocene deposits of 10.7–10.0 ky BP. At the mouth of the Khobi River in the western Colchis lowland, plant detritus exposed in a core at a depth of –41 m was dated by radiocarbon to 10,550±200 BP (Dzhanelidze 1980). A radiocarbon age of 10,130±180 BP was obtained for a sample of buried peat extracted from a core at a depth of –52.6 to –53.0 m on the Pitsunda Peninsula. In


0
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Nf
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Figure 3. (Facing page) Schematic curve of sea-level change in the Black Sea based on radiocarbon dating: 1 = ^{14}C of marine bars, spits, and terraces using mollusc shells and wood; 2 = ^{14}C of marine sediments using mollusc shells; 3 = ^{14}C of coastal-lagoon peats on peat and wood; 4 = ^{14}C of alluvial-lacustrine sediments using mollusc shells.

both cases, one cannot exclude the possibility of tectonic submergence. In other parts of the Caucasus littoral, deposits of a similar age were found below –40 m. On the Adler shelf, Neoeuxinian sediments were radiocarbon-dated to 10,350±270 BP at a depth of –80 m. In the Kerch Strait, a date of 10,530±190 BP was obtained from a sample recovered from a depth of –42 m, and on the Pitsunda Peninsula, a date of 10,220±140 BP was obtained from a sample at –75.4 m.

During the final transgression of the Neoeuxinian phase at *ca.* 9.8 ky BP, sea level rose to –25 m. In the traditional scheme of the Black Sea, Neoeuxinian sediments are identified as reflecting the growth of low salinity, brackish water fauna. Marine clay with mollusc shells of *Dreissena polymorpha*, *Monodacna caspia*, *Micromelania elegantula*, *Clesseniola variabilis*, and *Theodoxus pallasi* were found on the Sukhumi shelf at an elevation of –28 m. Littoral marine Neoeuxinian sediments found in the Dnieper-Bug liman at –20 to –21 m yielded a radiocarbon date of *ca.* 9.8 ky BP (Kovalyukh *et al.* 1977).

The end of the Neoeuxinian phase was marked by a considerable regression, indices for which were established on the Sukhumi shelf: a core exposed a layer of buried peat at a depth of –41.1 to –42.9 m, which gave a radiocarbon age of 9310±80 BP. Similar deposits on the northwestern Pontic shelf were dated to 8900 BP (Inozemtsev *et al.* 1984).

4.1.2 The Bugazian Phase

The Bugazian phase (9.2–7.9 ky BP) included clear signals for two transgressions separated by regressions dated 8.3–8.4 ky BP. Evolution of the first phase brought complex fluctuations as evidenced by buried peat layers exposed within cores on the northwestern Pontic shelf at Grigor'evka at an elevation of about –15 m and with a radiocarbon age of 8880±290 BP (Parunin *et al.* 1974). A similar date was obtained on the Gudauta shelf: 8890±150 BP.

The maximum rise of sea level at this time is linked to the breakthrough of Mediterranean water into the Pontic Basin. This event is indicated by the appearance of euryhaline Mediterranean fauna (Yanko-Hombach, this volume), often in the form of rare larvae and immature species of *Cardium edule*, *Abra ovata*, and, infrequently, *Chione gallina* and *Spisula subtruncata*. The appearance of Mediterranean foraminifera was also noted in sediments of the same age on the Sukhumi shelf (Yanko and Troitskaya 1987).

The first regression (between the first and second Bugazian phases) cut the influx of Mediterranean water and led to the extinction of the euryhaline micro- and macrofauna, at least within the Caucasus shelf. This regression is reliably dated by samples of buried peat exposed between two layers of lagoon clay within cores from the Adler littoral. Peat removed from a depth of –18.15 to –18.65 m gave a date of 8210±130 BP, and another sample recovered at a depth of –26.4 to –26.5 m yielded 8330±60 BP. The bottom peat covering the paleolagoon floor at a depth of –32 m may be accepted as a reliable marker for the maximum regression at this stage (Izmailov 1982).

The Late Bugazian transgression that reached a maximum at an estimated 8.15 ky BP corresponded with a water level rise to –16 or –17 m, as evidenced by the littoral-marine facies in the Kerch Strait, on the Sukhumi shelf, backwater and lagoon sediments of the Imeretia lowland, and similar deposits elsewhere. This transgression saw a new and more pronounced step in the evolution of Bugazian-type fauna, with a clear dominance of Mediterranean species over Neoeuxinian ones at the transgression maximum. The Bugazian fauna, which include *Cardium edule*, *Abra ovata*, *Hydrobia ventrosa*, *Mytilaster lineatus*, but abundant *Monodacna caspia*, *Dreissena polymorpha*, and *Theodoxus pallasi*, *Micromelania elegantula*, and *Clessiniola variabilis* were identified on the Sukhumi shelf at a depth of –27.8 m and yielded an age of 8260±300 BP.

The Bugazian phase ended with a regression dated to 7.9 ky BP, based upon buried peat layers in the area of the Imeretia lowland near Adler. Buried peat exposed at a depth of –27 to –28 m in a core south of the mouth of the Mzymta River yielded dates of 7850±120 and 7760±130 BP (Arslanov *et al.* 1975). A thick layer of peat exposed on the floodplain of the Mzymta River at a depth of –16.4 to –19.4 m was dated in the range of 7610±70 to 8050±70 BP. Similar dates were obtained at the Colchis coast for buried peat layers at an elevation of –17 m in the mouth of the Rioni River dating to 7900±60 BP (Dzhanelidze 1980) and at an elevation of –12 m in the Supsa River dating to 8000±940 BP (Kvavadze 1978). Regressive marine pebbles and gravel with euryhaline fauna in the area of the Sukhumi Peninsula have been radiocarbon-dated to 7860±70 BP. It should be noted that, starting from the Late Bugazian phase, the regressive sea-level decreases were not connected to a desalinization of the Pontic basin, although in several cases a depletion of the molluscan and foraminiferal assemblages has been acknowledged.

4.1.3 The Vityazevian Phase

During the Vityazevian phase (7.9–7.0 ky BP), sea level rose to as high as –9 to –10 m. Based on geological and geomorphological evidence, the shoreline corresponding to the transgression maximum was identified at this elevation with only small variations (±1 to 2 m) in all coring locations, from the Azov Sea and Kuban River Delta to the Colchis area. Regional distinctions are evident in the collected molluscs. Orictocenoses of this phase on the Caucasus shelf may be characterized as Early Kalamitian. They reveal a prevalence of *Chione*, *Corbula*, and *Cardium*, with an admixture of *Spisula*, *Nassarius*, and *Bittium*. Depleted yet typical Vityazevian orictocenoses that, for the most part, lack *Chione*, *Spisula*, and several other species, prevail in the area of the Kerch Strait and the Kuban River Delta. More than ten radiocarbon dates have been obtained using samples from these marine sediments, collected at depths of –13 to –15 m. Molluscs from the Sukhumi shelf yielded dates in the range of 7.2–7.6 ky BP, and similar dates were obtained for Vityazevian sediments in the northwestern Pontic limans (Molodykh *et al.* 1984).

The end of the Vityazevian phase, at 6.9–7.1 ky BP, was marked by a regression of sea level to –20 m. Buried peat yielding dates of that order was exposed within five cores from the interfluvial area between the Mzymta and Psou Rivers at depths of –12.1 to –19.1 m. The exact determinations were: 6970±120, 7000±140, and 7090±90 BP (Izmailov 1982).

4.1.4 The Kalamitian Phase

The Kalamitian phase (7.0–5.9 ky BP), which terminates the first stage of the Holocene transgression, included at least two oscillations. As sea level rose, inshore lagoonal peat was formed at increasingly higher levels. Hence, in the Mzymta-Psou interfluvial area, a buried peat dating to 6890±120 BP was exposed in the rear part of the Imeretia lowland at –8.7 m. The radiocarbon age of buried peat recovered at –8.7 m on the Sukhumi Peninsula was 6590±70 BP. The Nabadian peat north of Poti at about –5 m yielded an age of 6660±60 BP (Cherdyntsev *et al.* 1966; Sluka 1969), and at the village of Chaladidi (between the Rioni River and Paleostomi Lake) at an elevation of –13 m, an age of 6600±100 BP (Burchuladze *et al.* 1975).

The transgressive shoreline has been well established geologically at a depth of –3 to –4 m and dated to *ca.* 6.5 ky BP. To date, over 20 radiocarbon measurements have been taken on Kalamitian marine sediments in various parts of the Black Sea basin, nearly all of them using Kalamitian-type molluscs. In many areas, the Kalamitian transgression maximum was marked by deep ingression of the sea into the lower reaches of river valleys. The extent of this ingression was not exceeded during later phases of the Holocene transgression, notwithstanding the further rise in sea level. Similar phenomena have been observed in the areas of Sukhumi, Pitsunda (Balabanov *et al.* 1981), Adler, and also in the Kuban River Delta (Izmailov *et al.* 1989), where the submerged Kalamitian shorelines (the shell walls) were exposed by cores more than 30 km off the present-day shoreline.

The subsequent regression, which probably occurred at 6.2–6.0 ky BP, is identifiable in the peats exposed at –5.5 to –8.0 m in the Imnati bog (Neishtadt *et al.* 1965) and at –15.4 m in the Dnieper-Bug liman (Gozhik and Novosel'sky 1989). Existing geological evidence does not exclude a deeper level for the regression (possibly reaching –20 m or more), but the survey has so far been unable to obtain radiocarbon dates for continental sediments at this depth.

4.2 Holocene Stage 2

The second transgressive stage of the Holocene in the Black Sea encompasses the last 6000 years. During this interval, the pattern of sea-level fluctuation remained unchanged. Yet, in contrast to the previous stage, the transgression peaks corresponded to actual sea level. Two main transgressive phases, the Dzhemetinian (Dz) and Nymphaean (Nf), were separated by the Phanagorian regression. Dzhemetinian-type mollusc assemblages, corresponding to a salinity similar to today, remained dominant throughout the whole Black Sea perimeter (Nevesskaya 1965).

4.2.1 The Dzhemetinian Phase

The Dzhemetinian phase (6.0–2.6 ky BP) is distinctive not only in its sediments, but also in the relief features of the present-day coast and, most important, several generations of coastal bars (Balabanov *et al.* 1981; Arslanov *et al.* 1982). Three main transgressional subphases were separated by two regressions dated to 4.8–4.6 and 4.2–4.0 ky BP, judging from buried peat layers in various parts of the Caucasus littoral.

Three transgressions occurred during the first subphase at 5.7, 5.4–5.3, and 5.1–4.9 ky BP, respectively. The second subphase culminated at 4.6–4.4 ky BP, bringing with it maximum sea-level rise. The third subphase consisted of four short-lived transgressions, reflected in their corresponding coastal bars.

All transgressive subphases were separated by regressions of up to 6 m deep, as evidenced by the dates for buried peats on the Sukhumi littoral (5680 BP from a depth of –6.6 m), on the right bank of the Bzyb' River (5240±90 BP from –5.0 to –5.2 m), on the Colchis littoral (5240±60 BP from –3.75 m), and in the Gagra Bay coastal area (4800±90 BP from –5.0 m), among other places.

The main transgression maximum at 4.6–4.4 ky BP corresponds to coastal bars established at various points along the Black Sea littoral. On the Pitsunda Peninsula, it created a coastal bar at 2.0–2.5 m, which surrounded an ancient lagoonal depression in the peninsula's rear (Balabanov *et al.* 1981). The occurrence of transgressive shorelines of that age was confirmed in other areas as well. For example, an age of 4.5 ky BP has been reported for the terraces deposited along the shores of several limans in the northwestern Pontic area (Molodykh *et al.* 1984), and at 0.8 m high and dated to 4500±60 BP developed in the Zhesterov ridge of the Kuban River Delta (Izmailov *et al.* 1989).

During the Dzhemetinian phase, the most reliably established regression is the one dated 4.2–4.0 ky BP. Radiocarbon dates have been obtained from about ten buried peat layers of that age exposed in the mouth of the Khobi, Inguri, Bzyb', and other rivers, as well as the Early Bronze Age archaeological site within the mouth of Dzhikumur River near the town of Ochamchira (Voronov 1969). All these sites are 5 m below present sea level, and remarkably, all the dates cluster within the limited span of 4.25–4.05 ky BP.

Following this regression, during the second half of the Dzhemetinian phase, three small-scale regressions (less than 1.5–2.5 m) have been identified. The first regression occurred at ~3.9–3.8 ky BP based on peat from the Churia River in the Colchis littoral (3930±200 BP) and from Zmeiny Island (3860±90 BP). The next regression (–0.55 m) occurred at 3.4–3.3 ky BP based on peat from the Gagra (3400±80 BP) and Kobuleti littoral (3420 ±90 BP). The regression at 3.1–3.0 ky BP is evidenced in peat exposed within the Paleostomi Lake sequence on the Colchis littoral (3150+90 BP) and in the mouth of the Inguri River (3090±100 BP).

4.2.2 The Phanagorian Regression

The extent of the Phanagorian regression (~2.7–2.4 ky BP) is still under discussion. Both geological (Ostrovsky *et al.* 1977) and archaeological (Shilik 1972, 1977) evidence suggests that sea level fell by at least 7–8 m. This drop receives further confirmation from the remains of city walls attributable to the ancient Greek city of Dioskoria that are found at a depth of 6–10 m on the bottom of Sukhumi Bay (Balabanov and Gaprindashvili 1987). Radiocarbon dates of that age have come from peat layers at levels reaching –2.5 m at several sites in the urban areas of Gagra, Sukhumi, Kobuleti, and Poti (Rukhadze and Solov'ev 1964; Dzhanelidze 1980).

4.2.3 The Nymphaean Phase

During the Nymphaean phase (from 2.6 ky BP until the present time, or beginning with the 2nd century BC), sea level consistently remained close to its present position. This phase may be subdivided into two or three transgressions separated by small-scale regressions. The transgressions are marked by their respective generations of coastal bars laid down adjacent to each other unconformably in the greater part of the surveyed areas, particularly in the Kuban River Delta, Imeretia lowland, and Pitsunda Peninsula. Details of the paleogeographic setting during this phase were discussed by Balabanov (1984).

It should be noted that the occurrence of episodes of low sea level (the Medieval regression phases) was also confirmed by radiocarbon-dated buried peat and archaeological evidence. Thus, a date of 1350±60 BP was obtained for a peat layer at –3.6 m on the Pitsunda Peninsula, and the remains of a tower of the 5th to the 7th centuries AD were found on the floor of Sukhumi Bay at a depth of 2.8 m (Voronov 1969). Within the Nymphaean, a recent sea-level rise can be identified, starting after the late Medieval regression caused by the Little Ice Age.

4.3 Summary of Results

The evidence and analysis presented above permit the reconstruction of a high-resolution history of Holocene sea-level and salinity changes in the Black Sea with an accuracy of a few hundred years (Figure 3).

A vast number of radiocarbon dates (Table 1 in Appendix I), of which 150 records have been obtained by the author in the course of his studies, as well as substantial data from archaeological sites of the Bronze Age, Classical antiquity, and the Middle Ages that are located on the Holocene terraces and off-shore shelf, have contributed to three distinct lines of evidence. The first one includes the dates obtained from mollusc shells deposited either below or close to sea level; the second consists of archaeological sites or radiocarbon-dated in-shore-lagoonal peat layers formed slightly above sea level; and the third involves samples providing the chronological and altitudinal control for the surface of ancient coastal bars and terraces belonging to various phases of the final stage of the Holocene transgression. All points on the curve (Figure 3) correspond to the elevation of the dated samples. It should be noted that some previously published geochronological and archaeological materials were not accompanied by precise locations or elevations within their source geological and archaeological sections.

5. CONCLUSIONS

Analysis and synthesis of evidence accumulated by the large-scale geological survey of Black Sea terraces and adjoining off-shore shelf areas has enabled a qualitatively new outline of the Holocene history of the Pontic Basin, and in the process, it has considerably enhanced the accuracy of the eustatic curve (Figure 1). A solid geochronological foundation has been provided for several key stages in Black Sea evolution, especially those linked to the succession of mollusc assemblages. In this writer's opinion, the discovery of peat layers exposed by cores below present sea level was particularly important, as it refutes the previously held opinion about an interrupted development of the recent Black Sea transgression without intermediate decreases in sea level. In view of the abundant geological and chronological evidence, this writer is convinced that the occurrence of these paleogeographical episodes is beyond any doubt, yet the magnitude of the regressions needs further specification. One may state with confidence that a significant sea-level rise (in excess of 0.5–1.0 m) coinciding with the Holocene climatic optimum has not been confirmed by the existing evidence.

The paleoceanographic reconstructions outlined above show no evidence for catastrophic flooding of the Black Sea at 7150 BP or at 8400 BP as proposed by Ryan *et al.* (1997, 2003) and Ryan and Pitman (1998). At that time, sea level rose from –28 m to –8 m (Vityazevian transgression phase) over the course of 400–500 years with an average rate of 5 m $100y^{-1}$. Also by that time, the Black Sea had already acquired its two-way connection with the Mediterranean Sea through the connecting straits of the Dardanelles and Bosphorus, and it had been re-colonized by Mediterranean organisms (the Vityazevian molluscan complex of Nevesskaya 1965).

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31

SEA-LEVEL CHANGES AND COASTLINE MIGRATIONS IN THE RUSSIAN SECTOR OF THE BLACK SEA: APPLICATION TO THE NOAH'S FLOOD HYPOTHESIS

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Abstract: This paper discusses the stratigraphy and Quaternary history of the Russian sector of the Black Sea shelf. The key element of this history was eustatic sea-level rise that proceeded against a background of less pronounced neo-tectonic movements. In the light of existing evidence, the alleged catastrophic flooding of the Black Sea at 8.4 ky BP (Ryan *et al.* 1997, 2003) seems improbable.

Keywords: Pleistocene, Holocene, sea-level changes, seismic reflector, transgression, unconformity

1. INTRODUCTION

Since 1949, Yuzhmorgeologiia Enterprises (YMG) has been carrying out systematic geologic prospecting in areas of the Aral, Caspian, Azov, Black, and Mediterranean Seas, and since 1987, this institution has been actively involved in geoecological monitoring of the Russian sector of the Azov, Black, and Caspian Sea basins. Over this time, a substantial database has been gathered that includes information from gravitation-magnetic prospecting, various types of seismic-reflection profiling, sonar imaging, core sampling, drilling, under-water photography, scuba diving, as well as geochemical and other specialized analyses. Investigations of the upper portion of the Black Sea shelf sequence

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were based chiefly on single-channel seismic reflection profiling (SPARKER), multibeam echo-sounding, sonar imaging, bottom sediment sampling, and drilling.

The principal concepts regarding the geology and Quaternary history of the Black Sea that were proposed about twenty years ago based on evidence at that time have now been further developed in the light of new evidence (Shnyukov *et al.* 1979; Korsakov *et al.* 1982; Glebov and Skryabina 1984; Glebov and Sosnovsky 1984; Esin *et al.* 1986; Glebov *et al.* 1986; Glebov 1987; Glebov *et al.* 1996a, b, 1999; Glebov 1999). These concepts stem from previously formulated theories (Arkhangel'sky and Strakhov 1938; Degens and Ross 1972; Fedorov 1977; Ostrovsky *et al.* 1977a, b; Shcherbakov *et al.* 1977a, b; Fedorov 1978a, b; Shcherbakov and Babak 1979; Shnyukov *et al.* 1979; Shnyukov and Makarenko 1981; Balabanov *et al.* 1981; Shcherbakov 1982a, b; Fedorov 1984; and others) and in no way contradict the newly available evidence (Aksu *et al.* 1999, 2002a, b; Yanko-Hombach *et al.* 2002; and others).

Due to insufficient core drilling, seismic stratigraphic analysis remains the main instrument for studying the Quaternary geology and paleogeography of the Russian Black Sea shelf.

The Upper Pleistocene/Holocene transgressive-regressive cycle (the Neoeuxinian-Old/New Black Sea period, from 15–14 ky BP until the present) had an oscillatory character and proceeded against a background of general sea-level rise in the absence of catastrophic 'floods'. Available evidence makes it possible to suggest a quantitative forecast of future sea-level changes in the Black Sea and the development of its coastal areas.

The goal of this paper is to analyze the Late Quaternary history of the Black Sea in an attempt to discover whether there are any indications of the early Holocene catastrophic flooding proposed by Ryan *et al.* (1997, 2003) and widely discussed in the west. The paper concludes that, based upon available evidence, the alleged abrupt inundation of the Black Sea basin at 7.2 or 8.4 ky BP (Ryan *et al.* 1997, 2003) seems improbable.

2. MATERIALS AND METHODS

The present paper is based predominantly on results from geological survey of the upper portion of the sedimentary sequence on the Russian Black Sea shelf (Figure 1).

Geological survey and other investigations carried out by the YMG used multibeam echo-sounding (SIMRAD), high-resolution seismic and acoustic profiling (SPARKER, BOOMER, Sub-Bottom Profiler), submarine-towed Side-Scan Sonar, bottom sediment sampling, drilling, seafloor photography, scuba diving, and the study of onshore exposures. Density of seismic profiling and

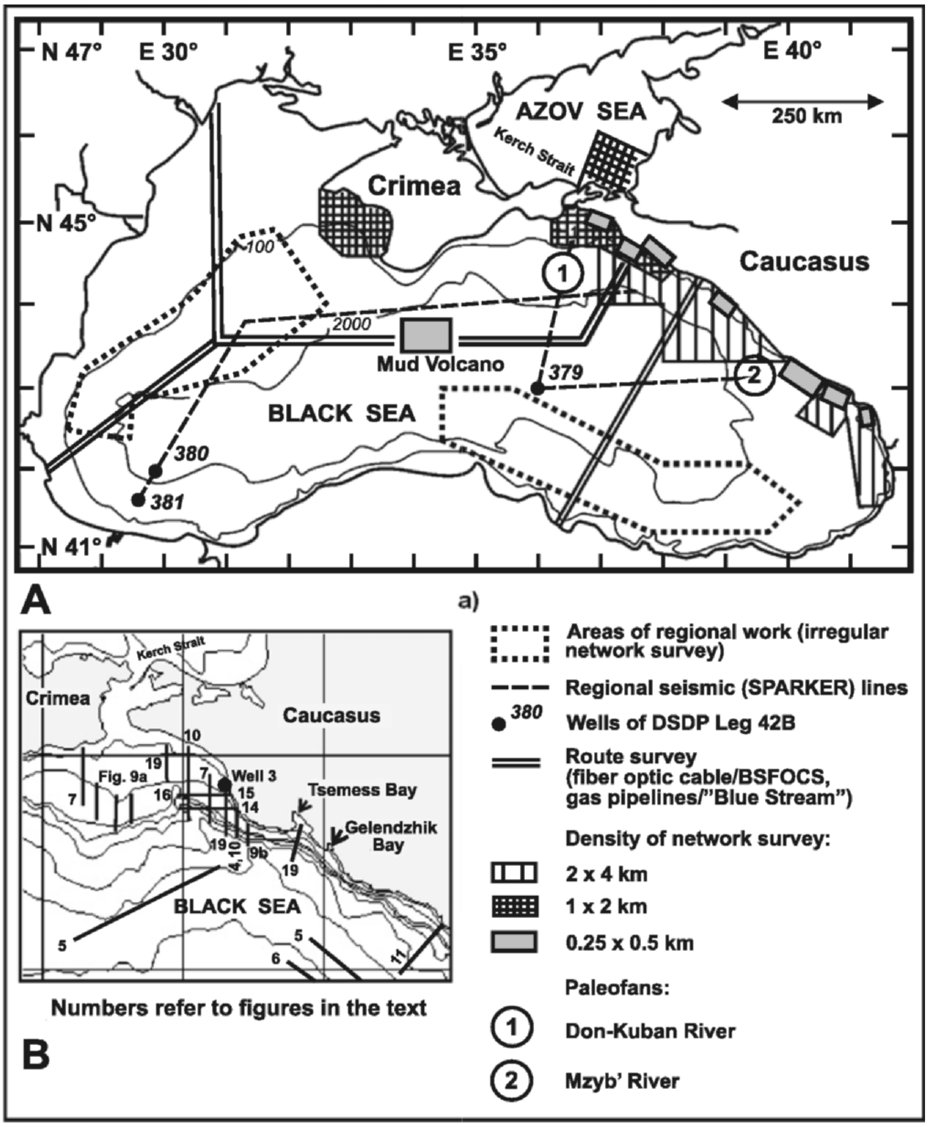


Figure 1. A. Locations of the geological and geophysical investigations in the upper portion of the sedimentary sequence of the Russian Black Sea shelf. B. Some of the seismic profiles and cores discussed in the text.

core sampling varied from 0.25 by 0.25 km to 2 by 4 km and yielded a substantial database containing thousands of kilometers of seismic profiles and sampling sites, as well as hundreds of meters of drill cores. YMG has at its disposal rich biostratigraphic materials that are not discussed here because they have been the subject of a special publication (Yanko-Hombach *et al.* 2002).

Nonetheless, these data have been used to interpret the seismic evidence.

In recent years, YMG has obtained a unique picture of the bottom relief of the Russian Black Sea sector (Figure 2) using a multidirectional eco-sounder (SIMRAD EM-12) in the course of detailed survey.

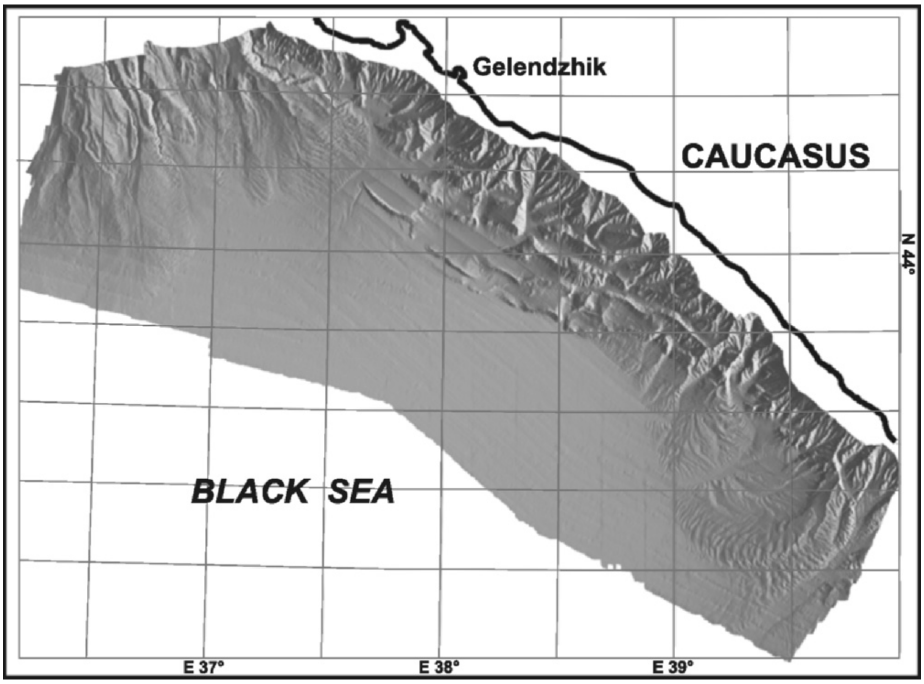


Figure 2. Shadow image (SIMRAD EM-12) of the Black Sea bottom along the Russian sector.

As mentioned above, seismic profiling (Payton 1977) is currently the only instrument providing sufficient data to establish a valid record of Quaternary stratigraphy and paleogeography. The senior author has dealt with these problems over many years (Shnyukov *et al.* 1979; Korsakov *et al.* 1982; Glebov and Skryabina 1984; Glebov and Sosnovsky 1984; Glebov *et al.* 1986; Esin *et al.* 1986; Glebov 1987; Glebov *et al.* 1996a, b, 1999; Glebov and Shilyaev 1999; Glebov 1999; and see YMG Reports 8,9,11,12,14–18,26,28,39, 40). The seismic stratigraphy discussed below derives from the interpretation of single-channel seismic-reflection profiling with a resolution from 1 to 4–5 m. The construction of average velocity plots for the entire Black Sea area was based on the velocity analyses of onshore cores and experimental marine seismic soundings (YMG Reports 1–12,14–17,19–29,34–39), as well as special tests performed on core #380 DSDP (Gorshkov *et al.* 1993).

Mean sound velocity values (recalculated twtt in seconds converted to depth below sea level) were estimated as 1500 m/sec in water and 1600 m/sec in Quaternary sediments. Data from shelf coring and sampling as well as onshore geological databases (Figures 3 and 4) contributed to the interpretation of seismic evidence.

Stratigraphic subdivision of reflectors and seismic units was achieved using data pertinent to the deeper Black Sea areas, and particularly the Don-Kuban alluvial fan. In the latter case, the regressive units (clinoforms) were clearly distinguishable in seismic profiles from the interbedded transgressive bodies of predominantly aggradational (nepheloid) sedimentation. Four successive seismic units identified in the records of the Don-Kuban alluvial fan and tentatively classified as Pre- Old Black Sea, Post-Karangatian, Post-Chaudinian, and Middle Chaudinian, indicate a prograding environment (Gorshkov *et al.* 1993; Andreev *et al.* 1999, Glebov *et al.* 1999; Prutsky 2000; YMG Reports 30–32,37,38). Recently, the accuracy of their stratigraphic position has been improved (Figure 5, top; Figure 8).

The architecture of the Don-Kuban alluvial fan units is viewed as an ideal set of seismic markers clearly distinguishable over long distances (Gorshkov *et al.* 1993, Glebov *et al.* 1999; YMG Reports 28,29). Their stratigraphic attribution is based on climatic stratigraphic evidence and may be considered a seismic 'stratotype' for Quaternary sediments in the entire north-eastern Black Sea region.

The investigation also focused on the alluvial fan stratigraphy of the Mzyb' and Shakhe Rivers (Figure 5, bottom) and the sedimentary apron of the Black Sea continental slope (Figure 6), where the indices of at least three Pleistocene-Holocene transgression-regression cycles have been recognized (Andreev *et al.* 1999; Prutsky 2000; YMG Reports 30–34,36,37).

In the present paper, 'seismic stratigraphic units' have been identified by the seismic reflectors that delimit them, above and below (Figure 7). The correlation of seismic stratigraphic units for Black Sea Quaternary deposits is based on (1) the analysis of unconformity surfaces and related reflectors and seismic facies, (2) typical antecedent and present-day relief, and (3) biostratigraphic data (Figure 8).

2. RESULTS

2.1 Geomorphology

The present-day relief of the Black Sea shelf was formed during the Pleistocene-Holocene in an environment of uninterrupted Caucasus uplift and sea-level oscillations. The Caucasian shelf is of variable width, and the greater part of its edge is marked by a scarp of presumably tectonic origin (Figure 9).

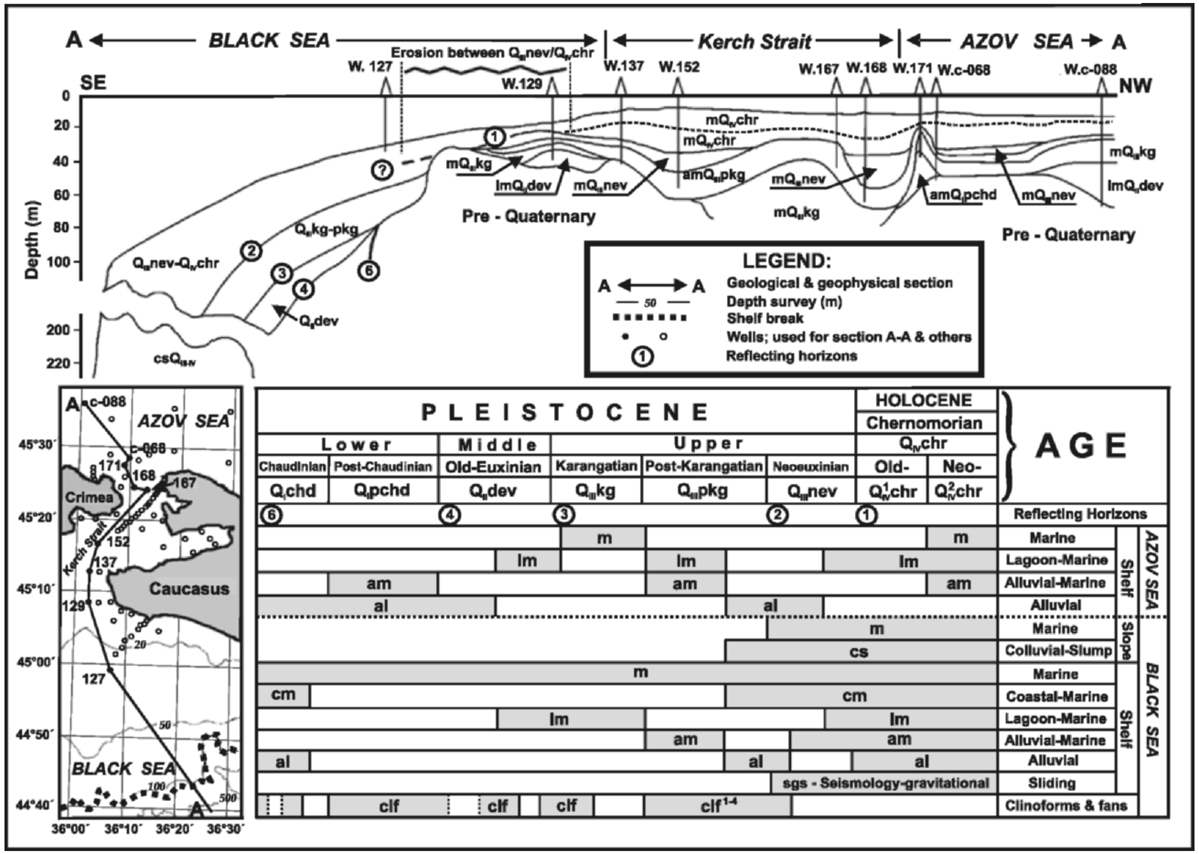


Figure 3. Example of geological data. The table includes acronyms and principal reflectors used in figure captions and in the text.

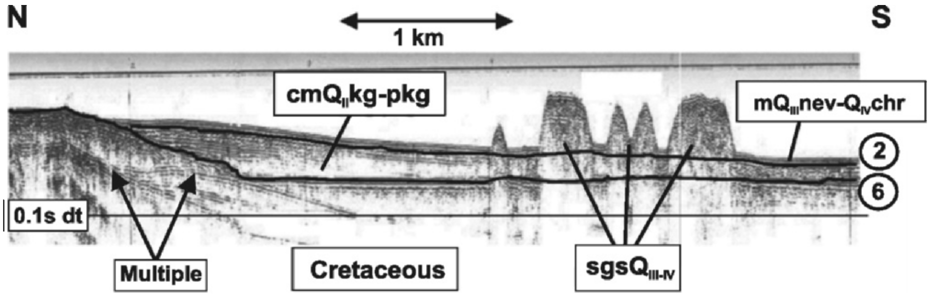


Figure 4. Example of reflector 2 stratification, based on seismic-gravitational landslips of known ages (Fragment of Profile 298131).

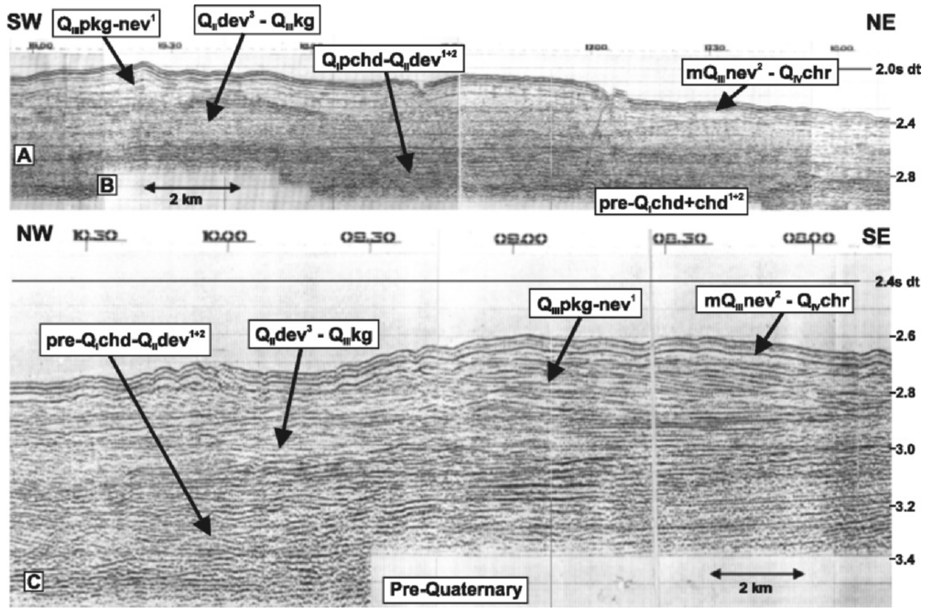


Figure 5. Alluvial fan structures of the Don-Kuban (top = fragment of Profile 49117) and Shakhe River (bottom = fragment of Profile 49102). A, B, and C = CDP (common depth point) reflectors; they are labeled A,Б, and В in the Russian reports.

Depth at the shelf edge varies between 85 and 115 m, representing the impact of different tectonic movements (Shnyukov *et al.* 1979; Shnyukov and Makarenko 1981; Glebov 1987; Glebov *et al.* 1996a; Prutsky 2000; YMG Reports 3,12,15,16,25,33,35). In the Kerch and Georgian sectors, one notes a gently-stepped transition to the continental slope, evidence for a continuing steady subsidence of the shelf edge (YMG Reports 9,15,21,22,35).

At several locations, the shelf edge is marked by an offshore bar or series of bars (Figure 10). On the Caucasian and Bulgarian shelf, these bars tend to

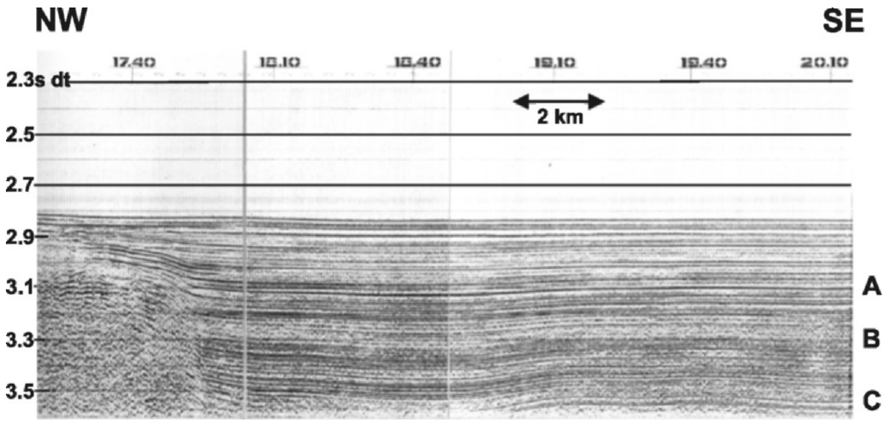


Figure 6. Structure of Quaternary sedimentary deposits at the bottom of the continental slope (fragment of Profile 49101); A, B, and C = CDP reflectors.

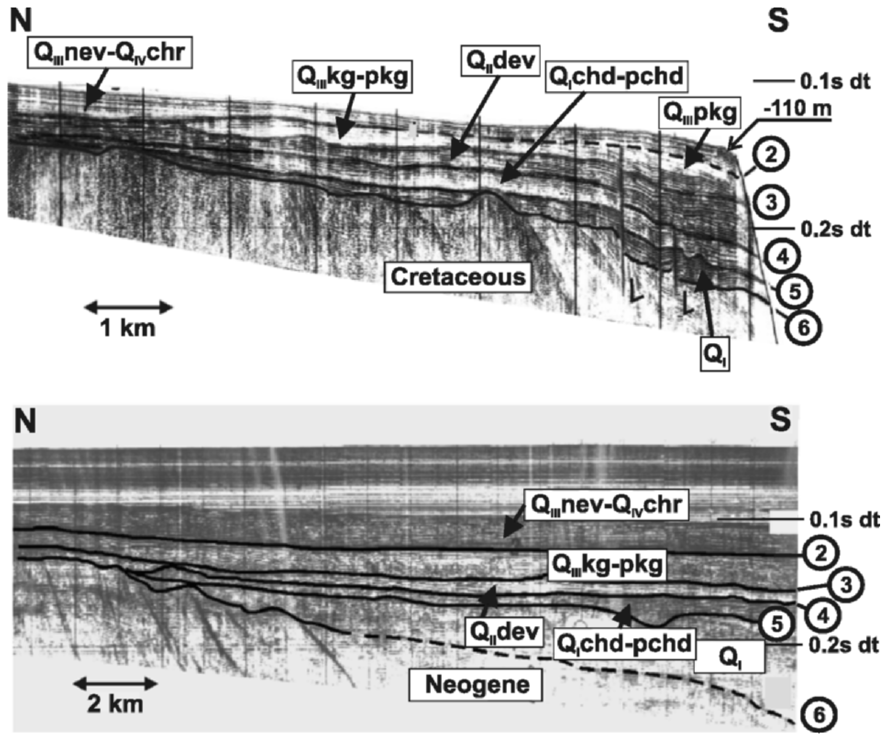


Figure 7. Key reflectors of the Caucasus (top = fragment of Profile 298116) and Kerch-Taman shelf areas (bottom = fragment of Profile 237652).

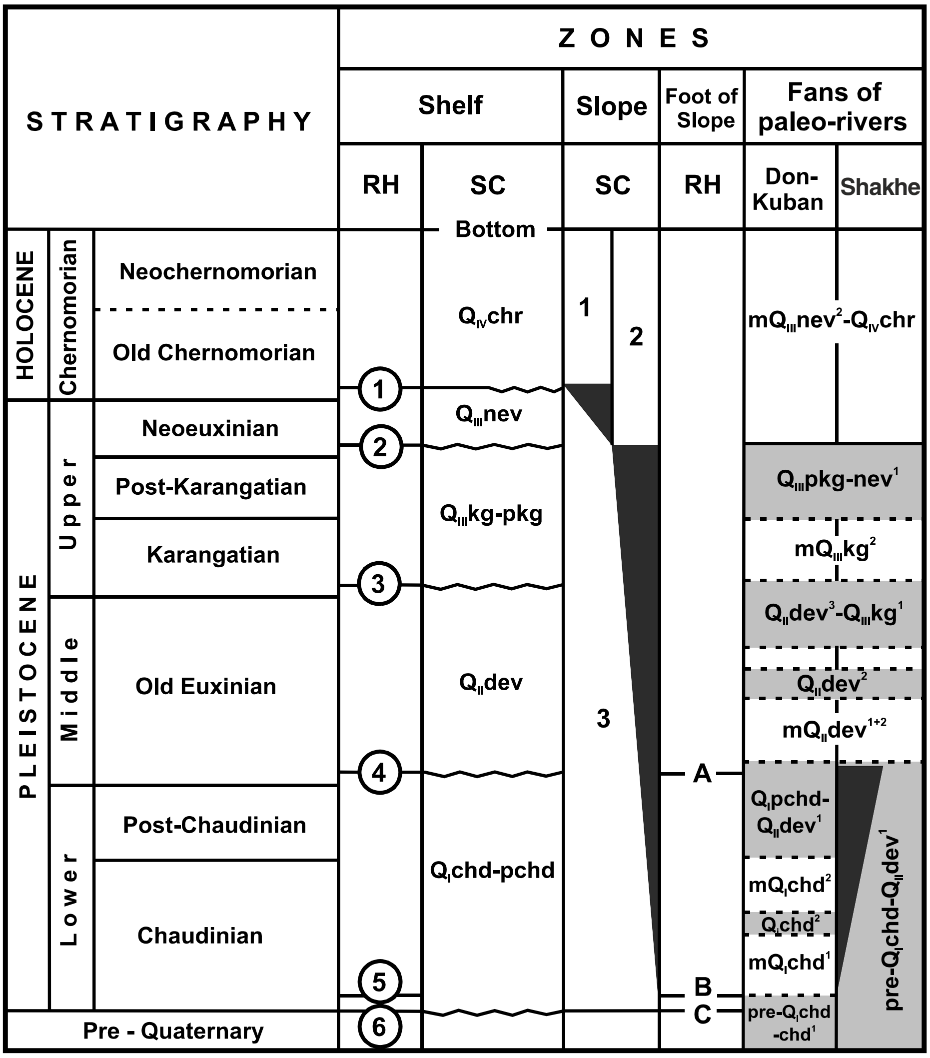


Figure 8. Tentative seismic stratigraphy for Quaternary deposits of the shelf, continental slope, and alluvial fans of the northeastern Black Sea area. Key: RH = reflecting horizons (1–6 = seismic-acoustic profiles, A-C = CDP reflectors); SC = Seismic complexes.

occur at water depths of 70–90 and 86–96 m, respectively (Esin *et al.* 1980; Goncharov and Evsyukov 1985; Glebov 1987; Glebov *et al.* 1996a; YMG Reports 15,33,36,38). An off-shore bar in the Kerch Strait was found at a depth of –143 to –147 m, which likely resulted from tectonic subsidence (Glebov 1987; Glebov *et al.* 1996a; YMG Report 15).

On the shelf surface, one can observe steps of submerged terraces (Figure 11) and off-shore bars clustering at certain bathymetric levels: cliffs and

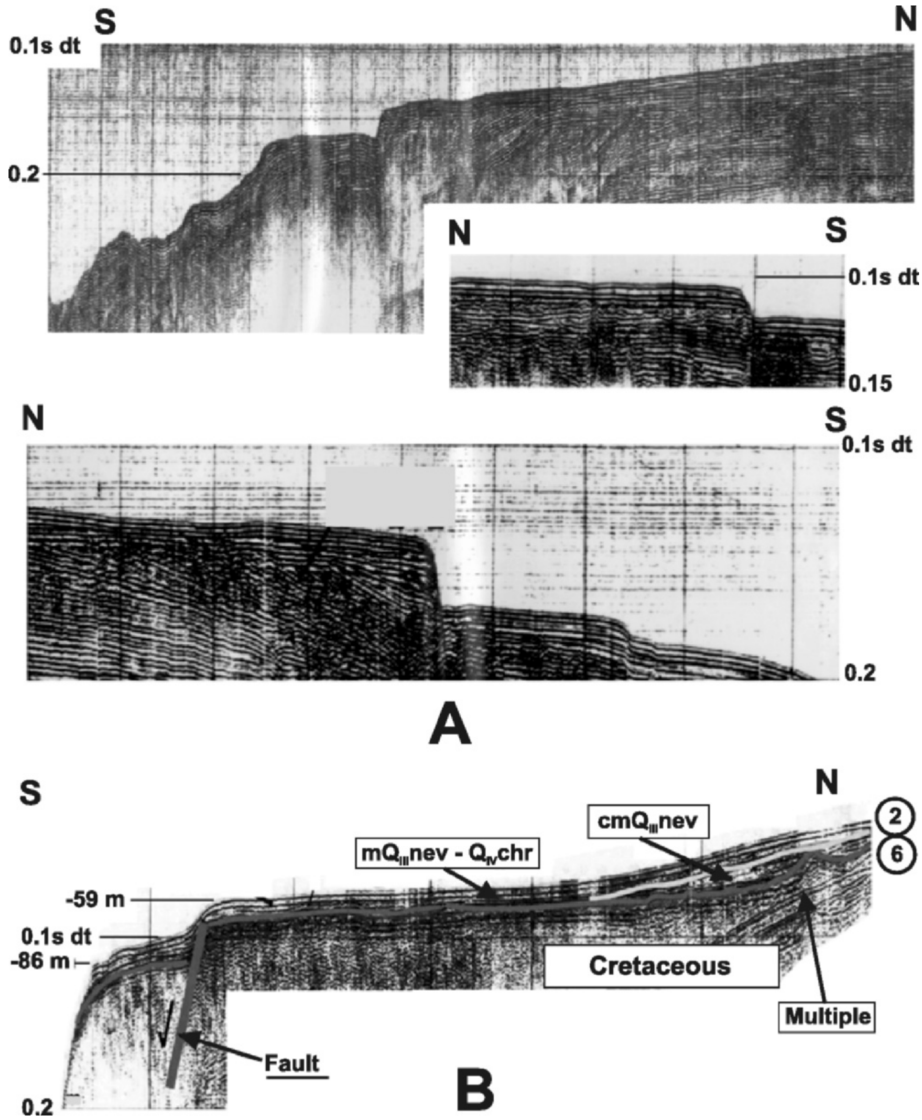


Figure 9. Shelf-edge scarps at Kerch-Taman (a = fragments of Profiles 237604,20,30) and the Caucasus (b = fragment of Profile 298134).

terraces (at –8 to –12 m, –15 to –20 m, –35/–40 to –45/–50 m, –53 to –71 m, –60 to –70 m, –68 to –80 m, –74 to –78 m, –82 to –93 m, –83 to –88 m, –86 to –115 m, –97 to –107 m, and –104 to –115 m), and smoothed bars (at –43 to –54 m, –55 m, –65 m, and –80 m). The most commonly encountered terrace, at –35 to –40 m, corresponds to the Bosphorus sill. Several terrace levels are reliably correlated with those established on the shelf of southern and western Crimea,

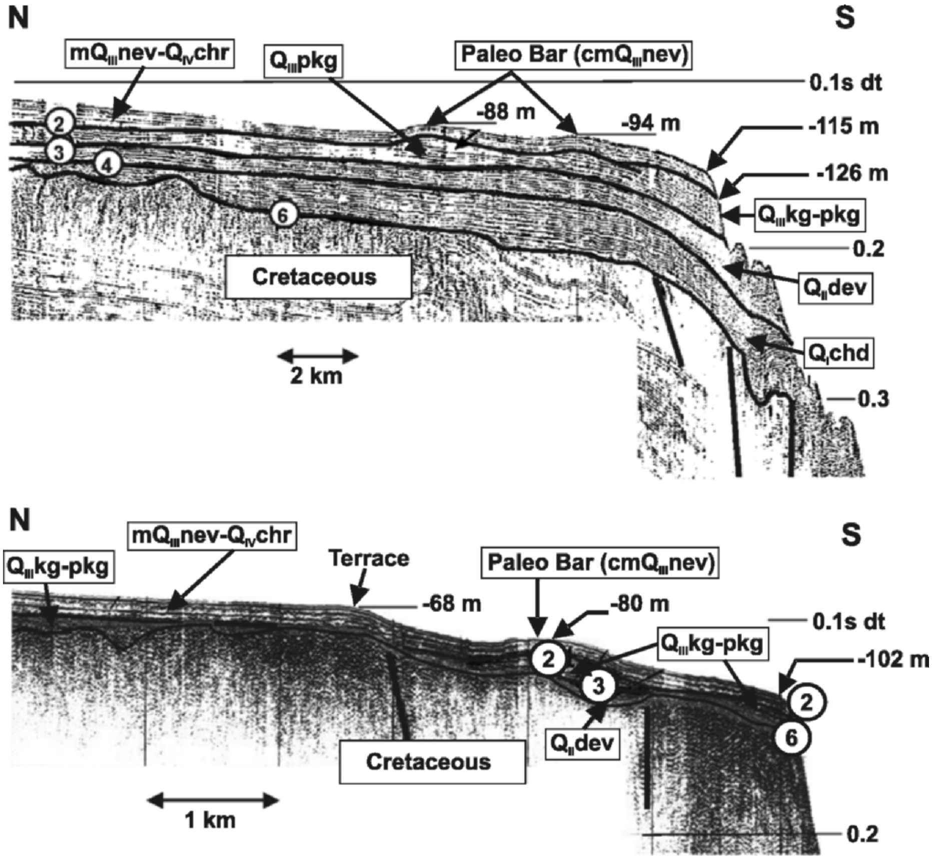


Figure 10. Off-shore bars (top = fragment of Profile 298112; bottom = fragment of Profile 298131).

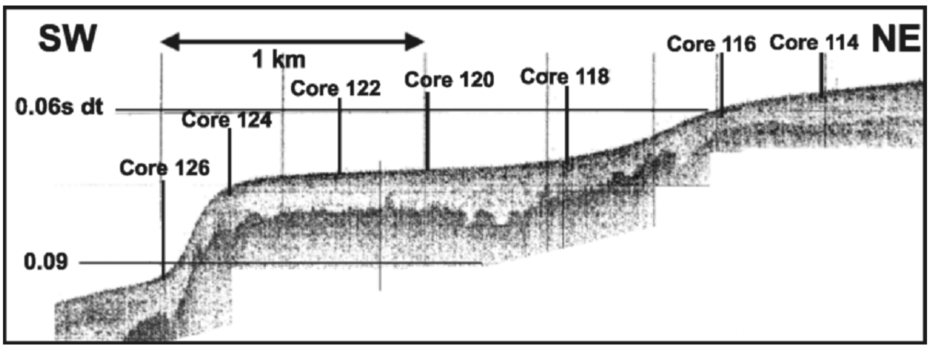


Figure 11. Example of Pleistocene-Holocene terraces on the shelf (fragment of Profile 3192119).

the Danube (Shimkus *et al.* 1980), and Bulgaria (Půrlichev and Nikolov 1977). These terraces are relics of ancient shorelines, and the noted differences in elevation are obviously due to recent tectonic movements. Their elevation reaches 10–27 m (Shimkus *et al.* 1980; Esin *et al.* 1986; Glebov 1987; Glebov *et al.* 1996a).

2.2 Quaternary History

Eustatic sea-level oscillations occurring against a background of large-scale neo-tectonic movements constitute the key elements of Black Sea Quaternary history. The main milestones may be summarized as follows. The total isolation of the Pontic-Caspian basin, which existed since the Miocene, was disrupted in the Pleistocene, when limited linkage with the Mediterranean Sea and the world ocean was re-established. Since that time, the alternating glacial and interglacial stages controlled sea-level oscillations in the Pontic and Caspian basins. Their cycles never coincided (Milanovsky 1968; Kvasov 1977). Black Sea transgressions occurred with the eustatic rises in ocean level during the late glacial and interglacial periods. In contrast, transgressions in the land-locked Caspian Sea, located in the arid zone, were controlled by the massive influx of melting water from the ice sheets during the latter half of the ice ages, while its regressions corresponded to the intensive surface evaporation and decreased river discharge of the interglacials and earlier half of the ice ages.

Calculations (Esin *et al.* 1986) based on a mathematical model of marine abrasion that takes into account tectonic movements and sea-level oscillations demonstrate clearly that a shelf 1–2 km wide could have been formed during the past 1.6 my. In reality, the shelf area is several times wider, and thus its formation must have begun in the Pliocene.

It should be noted that throughout the Quaternary, shelf sedimentation occurred on a heterogenic subaerial- and marine-eroded surface created during regression-transgression cycles ranging in age from the Lower Pleistocene to the Upper Pleistocene/Holocene. Thus, this surface presents a heterogeneous bottom for the Quaternary series and does not coincide with the lower boundary of the period in the stratigraphic sense.

This surface is clearly manifested in the seismic wave field in the form of a regional unconformity (reflector C). Plainly visible on this surface are terraces and ancient offshore bars at an elevation of less than 20 m (clustering at depths of –20, –30, –40, –50 to –60, –85 to –95, –105 to –110, –120 m, and more), as well as ancient continental slopes (Shnyukov *et al.* 1979; Glebov 1987; Glebov *et al.* 1996a; Glebov 1999; YMG Reports 5,8,9,11,12,14–17, 21,22,25, 33,34,36–39,40) (Figure 12). These landforms are also apparent in the thicknesses of the overlying younger sediments.

It should also be noted that recent tectonic movements have substantially

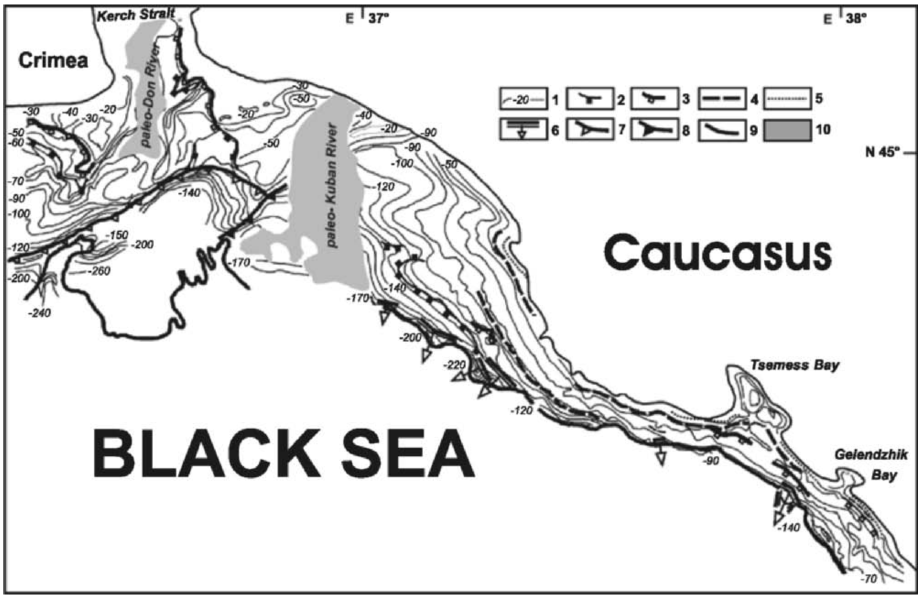


Figure 12. Structural map of the surface of Pre-Quaternary deposits. **Key:** 1 – isohyps (m), abrasion paleo-cliffs: 2 – Early Pleistocene; 3 – Middle-Late Pleistocene; 4 – Late Pleistocene; 5 – Holocene; 6 – faults, paleo-slopes: 7 – Late Pontian; 8 – Kimmerian; 9 – Early Pleistocene; 10 – zones without reflection correlation.

affected the elevation of both present and ancient landforms, in addition to the unconformity levels (principal seismic reflectors) (Ostrovsky *et al.* 1977a; Fedorov 1978a; Shnyukov *et al.* 1979; Shnyukov and Makarenko 1981; Glebov 1987; Nesmeyanov and Izmailov 1995; YMG Reports 8,9,12,13,15,36).

The present concept of Quaternary history is based on the history of Quaternary Black Sea level oscillations (Glebov 1987; Figure 13, top), with adjustments based on new seismic evidence pertinent to the Late Pleistocene-Holocene interval (Figure 13, bottom). The age of ~7.2 ky BP has been accepted as the Neoeuxinian/Old Black Sea boundary (Nevesskaya 1965; Degens and Ross 1972; Shcherbakov *et al.* 1977a, b; Shimkus *et al.* 1978; Shcherbakov and Babak 1979; Dimitrov 1982; Shcherbakov 1982b). For comparison and further discussion, we reproduced the curves suggested by Balabanov *et al.* (1981) based on ^{14}C AMS dates, as well as that of Ryan *et al.* (2003).

Based on existing data and conceptual understanding (Arkhangel'sky and Strakhov 1938; Fedorov 1977; Ostrovsky *et al.* 1977a, b; Fedorov 1978a, b; Shnyukov and Makarenko 1981; Fedorov 1984; Glebov *et al.* 1986; Esin *et al.* 1986; Glebov 1987; Glebov *et al.* 1996a; Glebov 1999), several stages in Black Sea Quaternary history are recognized:

Table 1. Large-scale regressive stages and their maxima.

Pre-Chaudinian	regression maximum of > –180 m
Mid-Chaudinian	up to –40/–50 m
Post-Chaudinian	> –160 m
Mid-Old Euxinian	up to –30/–40(?) m
Old Euxinian/Karangatian boundary	up to –130 m
Post-Karangatian/Early Neoeuxinian	up to –120/–140 (?)m
Short-lived Pre Old Black Sea	–34 (possibly –45 m)

Apart from these stages, the position of oscillating transgression levels has been estimated. Evidently, the Chaudinian and Old Euxinian basins were fairly similar to the contours of the present Black Sea; their levels slightly exceeded the present one. The Neoeuxinian transgression did not reach current sea level, however, having been lower by –8.5 m.

2.2.1 The Pre-Late Pleistocene

The regression depth at the **Gurian-Chaudinian boundary** was estimated at > –100 m (Fedorov 1978a, 1984). Later studies (Glebov 1987; Glebov *et al.* 1996b; YMG Report 15) focused on the unconformity at the shelf edge that was unaffected by tectonic subsistence, and its position was estimated at > –180 m. Such a depth is likely related to the deeply buried incisions of the Don, Kuban, Psou, and other paleo-rivers, as well as the canyons on the continental slope (Shnyukov *et al.* 1978, 1979; Korsakov *et al.* 1982; Glebov 1987; Gorshkov *et al.* 1993; Glebov *et al.* 1999; YMG Reports 15,30–32,34, 36,37).

The **Chaudinian transgression** (reflector 5) is recognized at a depth of –178 to –180 m (Figure 7, top), yet its elevation could be affected by an 8–10 m deep fault. This transgression truncated the underlying Gurian sediments and left discernible fragments of an abrasion cliff off the Caucasian coast at an elevation of –120 to –130 m. On the Kerch-Taman shelf, fragmented abrasion cliffs of that period are detectable at an elevation of –60 to –50 m. Supposedly, the abrasion presently visible in the bedrock relief at –30 to –40 m was related to the two phases of the Chaudinian transgression (Glebov 1987; YMG Reports 8,9,15).

Indices of the deep **Post-Chaudinian regression** and subsequent **Old Euxinian transgression** (reflector 4) are reliably fixed to an elevation of at least –160 m (with an adjustment for tectonic subsidence) (Figure 7, top). The Old Euxinian transgression triggered the subparallel-bedded, low amplitude (‘clay-ey’) seismic facies.

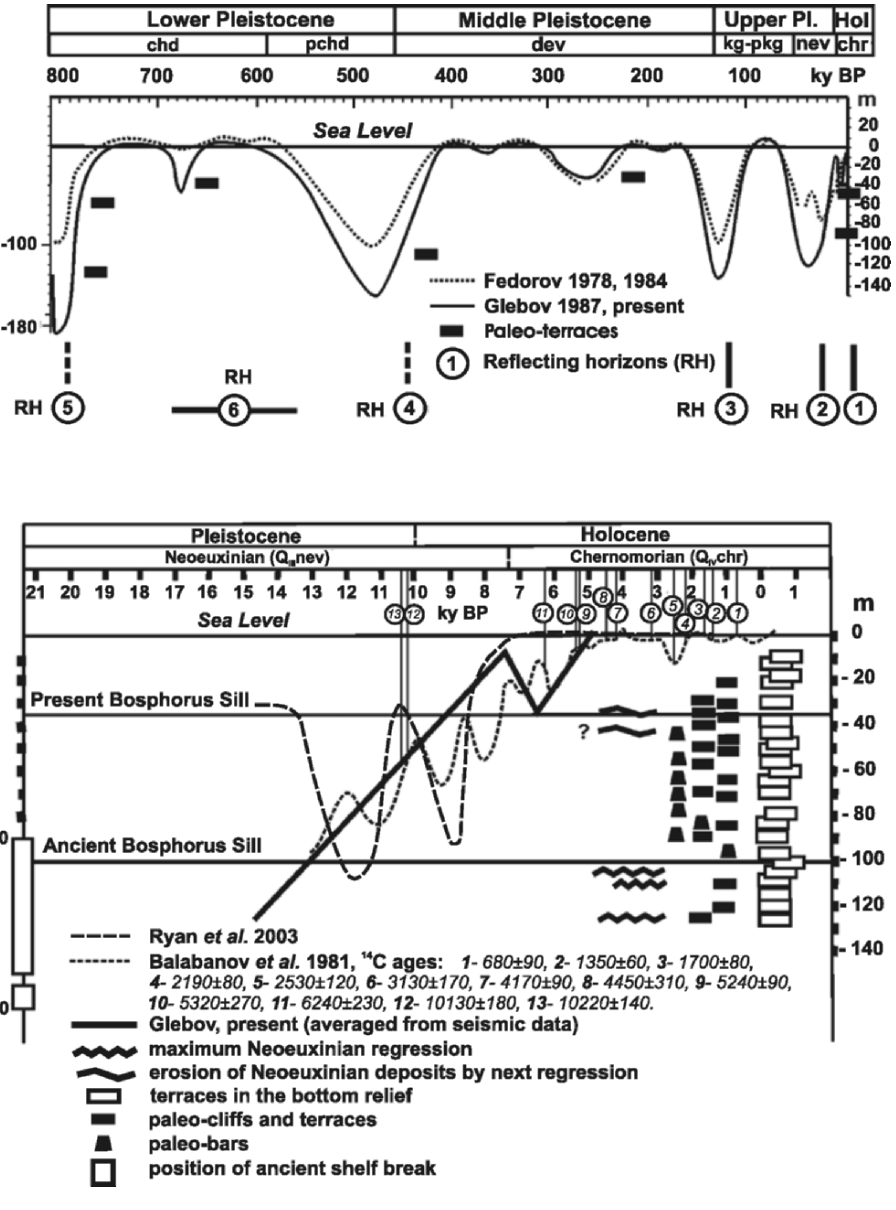


Figure 13. Quaternary sea-level changes in the Black Sea (top); Late Pleistocene-Holocene oscillations in the level of the Black Sea (bottom).

The **regression at the Old Euxinian/Karangatian boundary** was in excess of –130 m, and possibly during its course, the Bosphorus was deepened from –80 to –100 m (Scholten 1974; Fedorov 1977, 1978a). According to an alternative view, the Bosphorus did not exist prior to 100 ky BP (Aksu *et al.* (2002a).

The **Karangatian transgression** (reflector 3) was large in scale, yet its traces were practically obliterated by the subsequent Post-Karangatian/Early Neoeuxinian erosion (Shnyukov and Makarenko 1981; Glebov 1987, 1999). Taking into account a possible deepening of the Bosphorus, a linkage with the Mediterranean Sea may have been re-established at this time. Pre-Neoeuxinian sediments that had accumulated at the edge of the Caucasus started burying the tectonic scarps (Figure 7 top; Figure 14, top). An accumulative outstretching of the continental shelf proceeded actively in the Kerch-Taman area, where it advanced 4–10 km by the beginning of the Neoeuxinian period and acquired a smoothed character due to the infilling of canyons (Shnyukov *et al.* 1979; Shnyukov and Makarenko 1981; Glebov 1987; Gorshkov *et al.* 1993; Glebov 1999; YMG Reports 9,15). A similar 'avalanche-like' sedimentation was also noted on the Caucasian shelf (Figure14, bottom; YMG Reports 21,22).

2.2.2 Late Pleistocene-Holocene

The **Pre-Neoeuxinian regression**, which started during the Karangatian period, is most clearly evident in the Quaternary sequence (Figure 15).

Relics of the Post-Karangatian regression are detectable in several areas (Figures 7 and 10). The elevation of the Pre-Neoeuxinian shoreline on the Caucasian shelf has been identified at –126 m (Glebov1987; YMG Reports 9,15).

Like other areas on the periphery of the Black Sea basin, the Kerch-Taman zone (Arkhangel'sky and Strakhov 1938) experienced intensive block uplifting during the early Upper (Late) Pleistocene (the Post-Karangatian period) that had an amplitude of 60–80 m and an apparent slip of 20–40 m, and faded out toward the middle Upper Pleistocene (Shcherbakov *et al.* 1977b; Shnyukov and Makarenko 1981). This precludes its use for estimating sea-level changes.

It is generally accepted that prior to 15 ky BP, the surface of the Neoeuxinian 'sea-lake' lay at a depth of –140 m (Yanko-Hombach *et al.* 2002). Simultaneously, the Khvalynian Sea rose in the present Caspian basin. A short-lived outflow via the Manych Depression may have existed (Fedorov 1978a). The Black Sea was isolated from the Caspian basin between 18 and 15 ky BP, at a time of maximum drop in sea level (Shcherbakov *et al.* 1977b).

The beginning of the **Neoeuxinian transgression** (reflector 2) occurred at 15–14 ky BP (Shcherbakov 1982b; Yanko-Hombach *et al.* 2002).

During the period of 15–10 ky BP, the rate of sea-level rise was concomitant with that of the world ocean. Initially reaching 20 mm/year during the initial stage of the transgression, it decreased to 9–10 mm/yr during the final stage (Gozhik 2001; Yanko-Hombach *et al.* 2002).

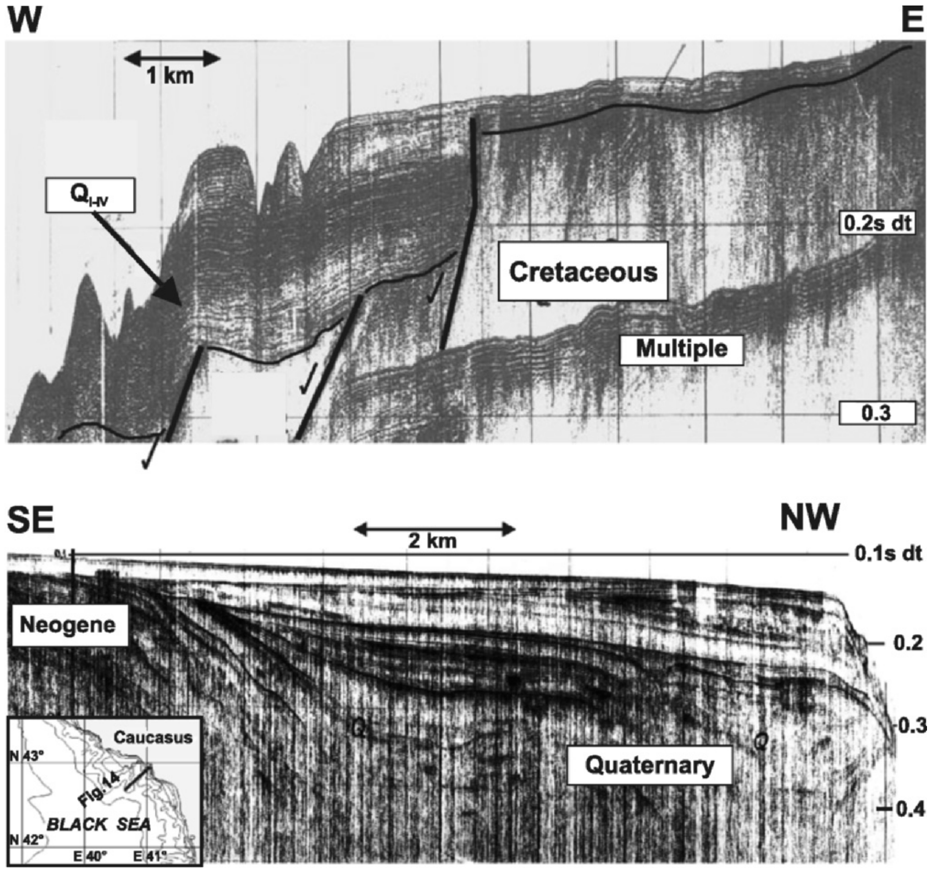


Figure 14. Buried evidence of tectonic subsidence at the edge of the Caucasian shelf (top = a fragment of Profile # 2981114) and avalanche-like sedimentation at the edge of the Georgian shelf (bottom = a fragment of Profile # 308850).

Offshore bars were formed at the shelf edge in the initial stage of the transgression at a depth of –80 to –94 m, and a cliff emerged at –68 m (Figure 10). Cores extracted along the shelf edge (at –80 to –90 m) show marine-littoral facies, including beach gravel (Shcherbakov *et al.* 1977a; Shcherbakov 1982b).

A similar offshore bar, in existence since the Chaudinian period, has been noted on the Bulgarian shelf at an elevation of –80 to –100 m, in a context of littoral facies (Dimitrov 1978). Remains of an ancient (Neoeuxinian?) cliff were recognized on the northwestern shelf at ~ –100 m, along with areas of littoral bars at –30 to –40 and –50 to –80 m (Panin *et al.* 1977).

Neoeuxinian-related cliffs and offshore bars in addition to the accumulation of marine-littoral sediments (Glebov 1987; YMG Reports 8,9,15) suggest a slow sea-level rise with stillstands and minor regressions.

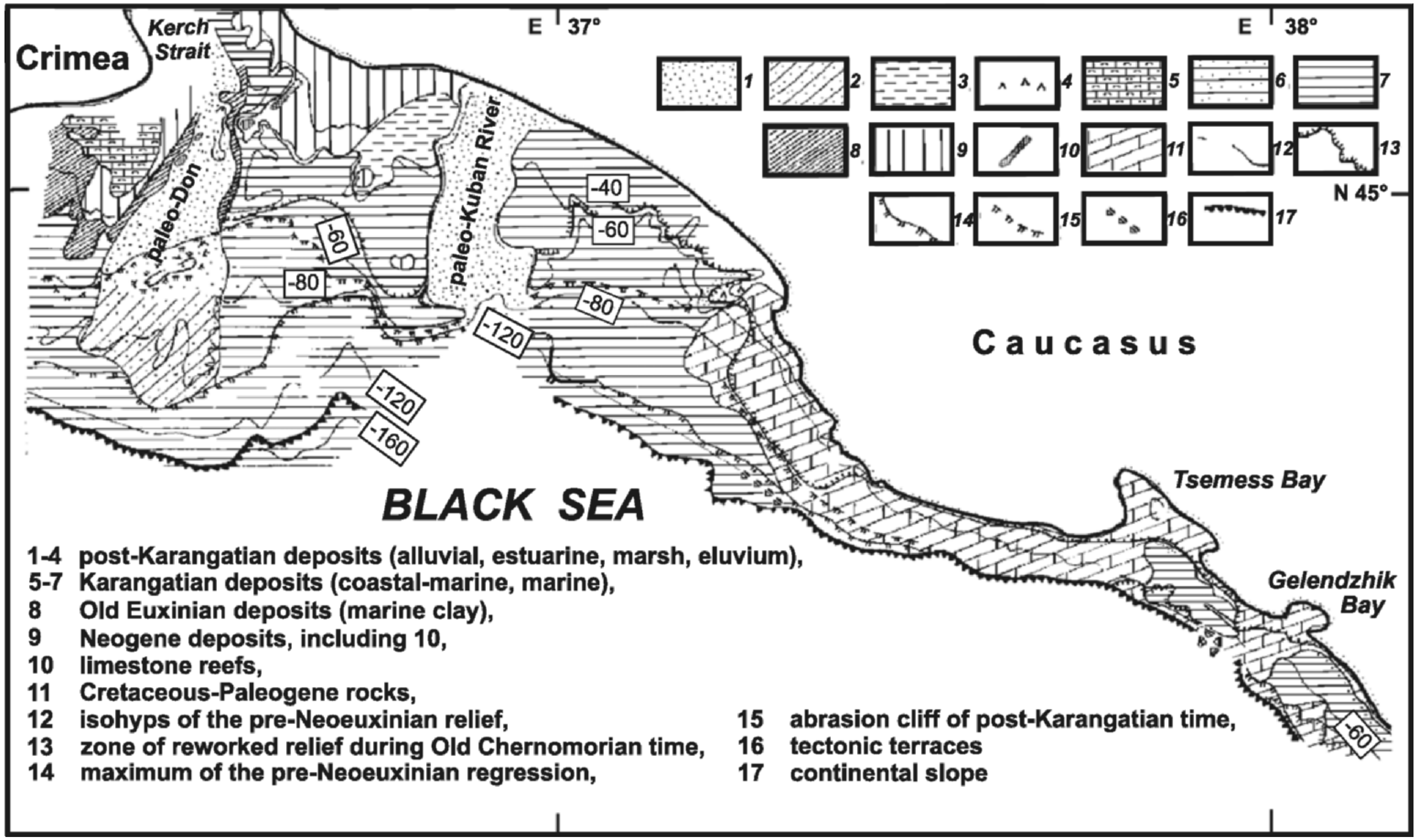


Figure 15. Paleogeographical setting of the early Neoeuxinian period.

Based on the interrelation of the Don and Kuban River prodeltas with the Post-Karangatian and Early Neoeuxinian abrasion cliffs, one can conclude that the infilling of the Don and Kuban valley floors began after the Post-Karangatian regression maximum, resulting from the damming of the Early Neoeuxinian transgressing sea. Alluvial (deltaic?) sediments of that age are reflected in the interrupted, hummocky, and chaotic seismic signals from the infilling of erosional river channels (Figure 16, top). This profile intersects the submerged Kuban paleo-valley described in previous publications (Shnyukov *et al.* 1978; Shnyukov and Makarenko 1981; Glebov *et al.* 1986; Glebov1987; YMG Report 15). At the same time, a slow accretion of clinoforms proceeded along the marginal part of the Kerch-Taman shelf (Figure 16, bottom).

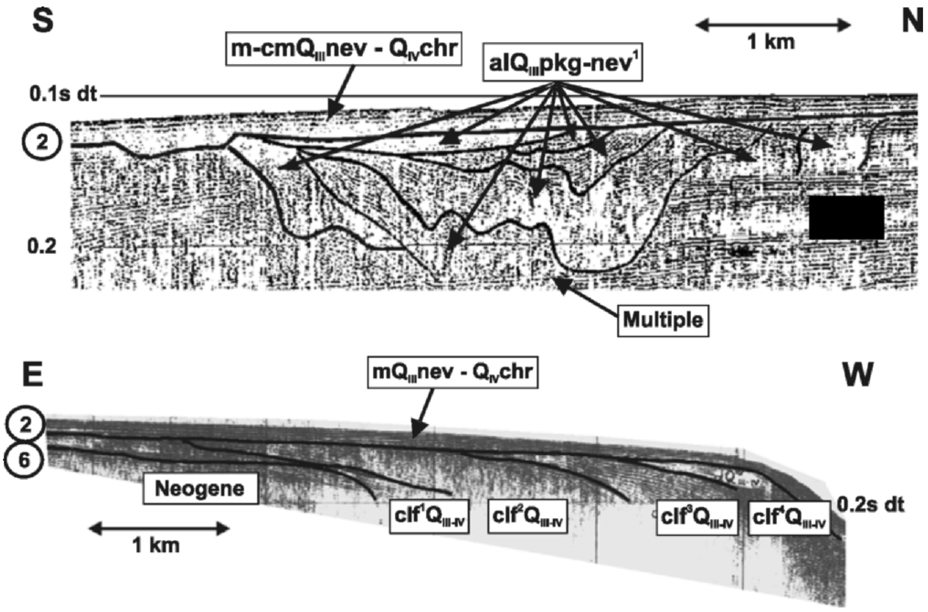


Figure 16. Infilling of the Paleo-Kuban valley (top = a fragment of Profile # 298102) and clinoforms of the lateral shelf accretion (bottom = a fragment of Profile # 2981117).

The subsequent development of the Neoeuxinian transgression was marked by the accumulation of marine sediments on the Black Sea shelf. They are generally represented on the seismic records in the form of bedded signals with prevailing low-amplitude seismic facies. In the cores, these sediments consist of clayey silt. Only in the Azov Sea did the Neoeuxinian sediments comprise subaerial, alluvial-lacustrine formations. The Neoeuxinian littoral-marine deposits reach essentially up to the present-day shoreline.

The infilling of overdeepened river floors on the inner shelf by alluvial and estuarine-marine sediments likely commenced in the final Neoeuxinian period (Ostrovsky *et al.* 1977b; Figure 17). The lower part of the sequence as recovered in cores has been ^{14}C AMS-dated to 10.53±1.90, 10.35, and 8.9–7.3 ky BP (Prutsky 2000; YMG Reports 32,34).

The sea ingressed into the overdeepened depressions of Tsemess and Gelendzhik Bays during the maximum stage of the Neoeuxinian transgression (Glebov 1987; YMG Report 15). In several cores (## 5G, 6G, 8G, 9G, and 18G), Neoeuxinian estuarine-marine deposits 13 m thick ($lmQ_{III}nev$) consisting of

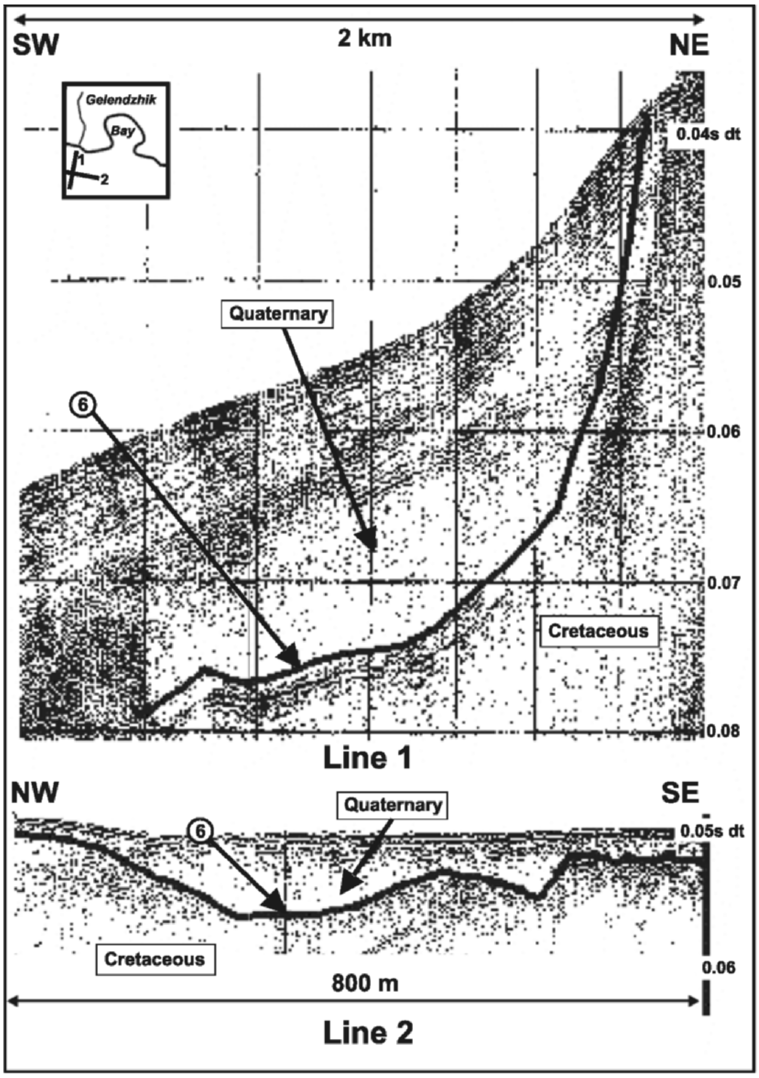


Figure 17. Infilling of the overdeepened floor of the Ashamba River (off Gelendzhik).

pellitic silt and clay with numerous plant detrita and peat horizons have been recovered beneath 10 m of Old and New Black Sea silts.

The liman (estuarine) Neoeuxinian sediments are identifiable as a transparent seismic section. Reflector 1, found at its surface (Figure 18, top), can be traced to a depth of –8.5 m (Figures 18 and 17, bottom) and likely corresponds to the maximum of the Neoeuxinian transgression. A similar pattern occurs in Tsemess Bay. According to the evidence from coring (Shnyukov and Makarenko 1981), Neoeuxinian deposits were encountered in the Kerch Strait at a depth of –10.9 m.

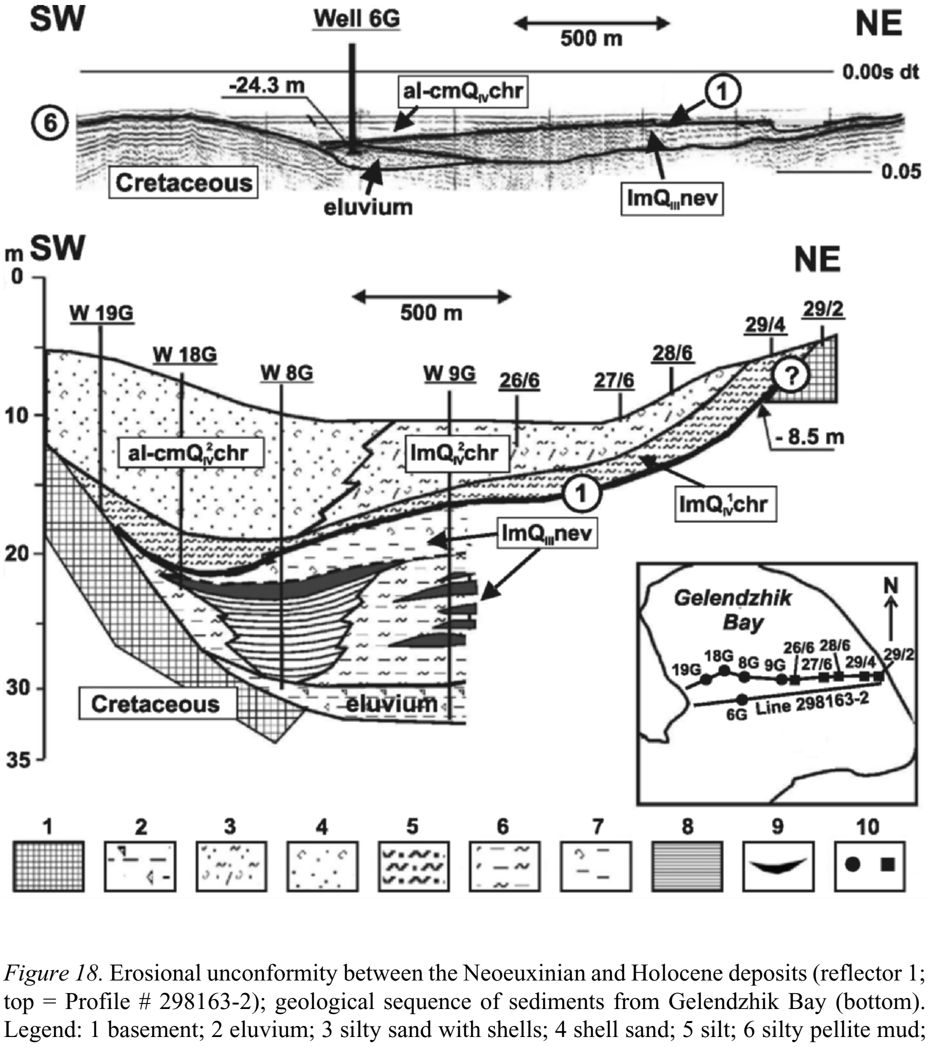


Figure 18. Erosional unconformity between the Neoeuxinian and Holocene deposits (reflector 1; top = Profile # 298163-2); geological sequence of sediments from Gelendzhik Bay (bottom). Legend: 1 basement; 2 eluvium; 3 silty sand with shells; 4 shell sand; 5 silt; 6 silty pellite mud; 7 pellite mud with shells; 8 clay; 9 peat; 10 wells and cores.

The final Neoeuxinian stage was marked by a small-scale regression. Based on the occurrence of the subsequent marine erosion of the Old Black Sea transgression (reflector 1), its elevation is estimated as –34 m. Reflector 1 is identified in seismic records as the shallowest unconformity starting from the sea bottom (Figures 18 and 19).

Based on the evidence obtained by coring, this boundary is viewed as the Pre-Holocene one (Glebov 1987; YMG Report 15). Its deeper position, in excess of –45 m (Figure 19, bottom), is probably due to a recent tectonic subsidence of the Kerch-Anapa area (Shnyukov *et al.* 1979; Shnyukov and Makarenko 1981; Glebov 1987).

In the greater part of the Black Sea basin, Neoeuxinian deposits have been completely eroded in the active wave impact zone (inner shelf) and in areas of active incision by submarine canyons to a depth of 2000 m. Evidence for a regression in the final Neoeuxinian stage has been found by coring in the area of the Kerch Strait (Shnyukov and Makarenko 1981).

In the Azov Sea, estuarine-marine deposits of Old Azov age (Bugazian-Vityazevian stage, according to the loess stratigraphy) fill depressions in the pre-Holocene relief, unconformably overlying Neoeuxinian, Post-Karangatian (Surozhian), and Karangatian sediments, and (in Kazantip Bay) Old Euxinian deposits.

In cores from the Georgian shelf, the bottom of the Holocene deposits is nearly always marked by an erosion surface in the form of a basal horizon. A depositional gap is indicated by the occurrence of redeposited ostracoda at the bottom of the sequence (Yanko and Troitskaya 1987). A small-scale regression prior to the start of the Holocene Black Sea transgression is also evidenced by peat horizons at the Neoeuxinian-Old Black Sea transgression boundary in Colchis. This regression-transgression cycle is linked to the level of erosion of Neoeuxinian deposits (at –30 to –20? m); previously, it had been interpreted as the maximum transgression of the Neoeuxinian Sea.

The subsequent Black Sea transgression proceeded simultaneously with the Flandrian transgression in the Mediterranean basin. At that time, the lower Bosphorus current allegedly eroded the sill in the Istanbul area, deepening it and forming a valley, the floor of which now drops from –35/–40 m to –80/–90 m at the entrance to the Black Sea (Fedorov 1978a). The increased influx of Mediterranean water led to stagnation (a deoxidizing medium) inside the Black Sea (Shcherbakov *et al.* 1977b; Shcherbakov 1982a, b).

In a geochemical sense, the influx of Mediterranean water and hydrosulfuric poisoning are evidenced by a marked change in molybdenum accumulation during the Old Black Sea period (Mitropol'sky *et al.* 1982). Along the shelf and down the continental slope, the Black Sea sediments formed during the most recent transgression nearly universally cover the underlying deposits.

The Black Sea Holocene transgression was marked by the formation of the present-day bench. An estuarine-marine environment was re-established in

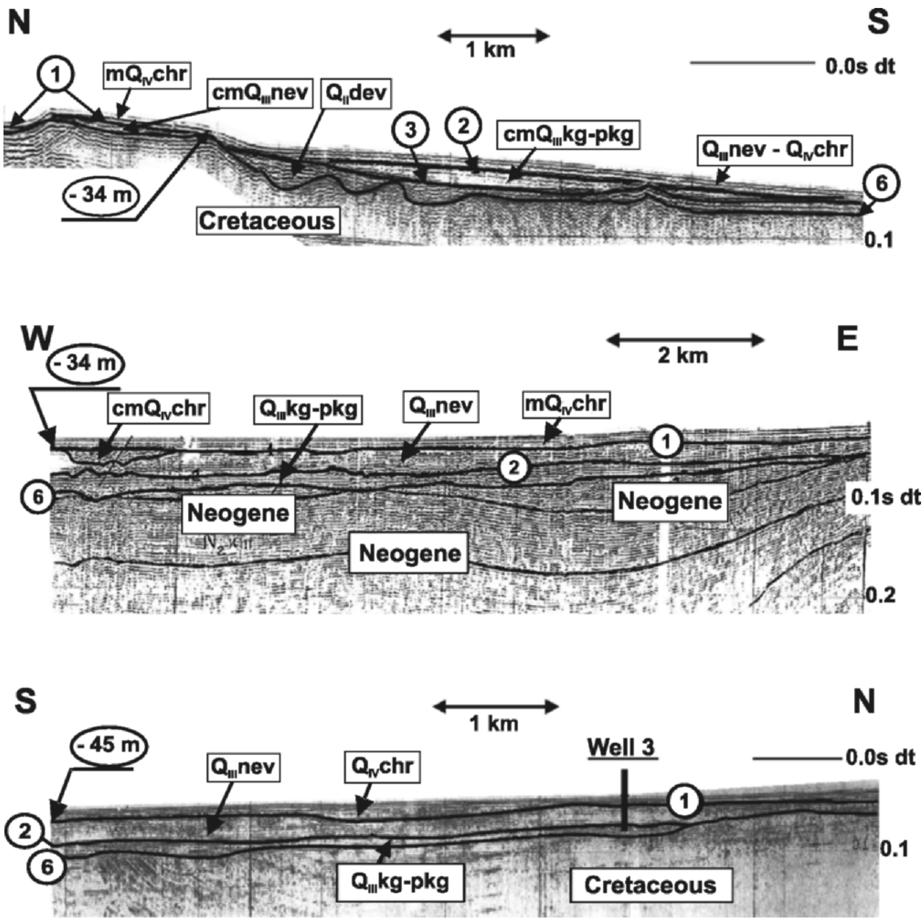


Figure 19. Possible regression maximum detectable between the Neoeuxinian and Holocene deposits (top = a fragment of Profile # 298169; middle = a fragment of Profile # 2981200). Recent tectonic submergence of the Taman shelf (bottom = a fragment of Profile # 298121).

Tsemess and Gelendzhik Bays in Old Black Sea time.

The transition from the Old to New Black Sea deposits was gradual. The Black Sea transgression had an oscillatory character, with sea level exceeding its present elevation by 2–2.5 m in the New Black Sea phase. It is generally accepted that sea level reached its maximum during the Climatic Optimum, which occurred between 3–5 ky BP (Fedorov 1977; Ostrovsky *et al.* 1977b; Balabanov *et al.* 1981; Yanko and Troitkaya 1987; Gozhik 2001). The Phanagorian regression, during which sea level fell to between –5 and –13 m,is substantiated by archaeological evidence dated to *ca.* 2.7 ky BP (Shilik 1977).

Numerous archaeological sites of Greek and Roman antiquity are known from the lower Holocene marine terrace. They include Classical and Iron Age sites, as well as those of later ages (Hermonasse, Phanagoreia, Tmutarakan, and sites on the northern shore of Taman Bay).

The beginning of the **contemporary transgression** is estimated to have taken place between the 2nd/3rd centuries and 500 AD. Pollen analyses of marine sediments indicate considerable climate change during the Holocene, including the last three millennia: an alternation of warm, cool, wet, and dry climatic phases of varying duration. The paleoclimate of the Black Sea area featured the occurrence of periodic pluvial phases with a substantial increase in water drainage within the entire catchment. These phases corresponded with a marked increase in bio-productivity in the Black Sea basin and the formation of sapropelic ooze.

2.3 Other Black Sea Areas (With Special Reference to the West Crimean Shelf)

Characteristics of the Quaternary sequence in the Russian Black Sea sector outlined above in no way contradict evidence from other areas (Korsakov *et al.* 1982; Glebov and Skryabina 1984; Glebov and Sosnovsky 1984; Glebov *et al.* 1986; YMG Reports 3,10–12,14,19–24,26–29). The investigations carried out in 1978–1981 by YMG on the West Crimean shelf (YMG Reports 11,14) revealed similar morphological-structural elements, reflectors, and seismic-stratigraphic units (Figure 20), as well as deposits representing the main stages of geological history.

The Cimmerian-Kuyalnikian paleo-shoreline is the deepest (–110 to –150 m), and the shoreline at –65 to –75 m lies at an intermediate position within the Neoeuxinian transgression. The Neoeuxinian period corresponds to the progradation of the shelf, and the maximum sea-level rise at the end of the Neoeuxinian period, as evidenced in the Black Sea limans, reached an elevation of –18 m. These sequences also indicate regressive erosion that occurred between the deposition of Neoeuxinian and Bugazian sediments (Granova 2001).

The levels from –35 to –45 m correspond to the pre-Old Black Sea regression maxima. These levels are distinguished by cliff formation in the pre-Karangatian period and detectable in the present-day relief. Traces of the Holocene Black Sea transgression are also evident at a depth of –22 m.

Transgression phases were accompanied by the flooding of drainage basins. Wave-induced transport of sediments formed sand-spits in the shallows of coastal bays with the emergence of shallow limans and lagoons. These formations appeared at three levels during the course of the contemporary transgression: at –55 to –60, –30 to –35 and, –20 to –25 m, which confirms the slow rise of sea level with oscillations at this stage. Limans inherited from previous stages are currently being formed.

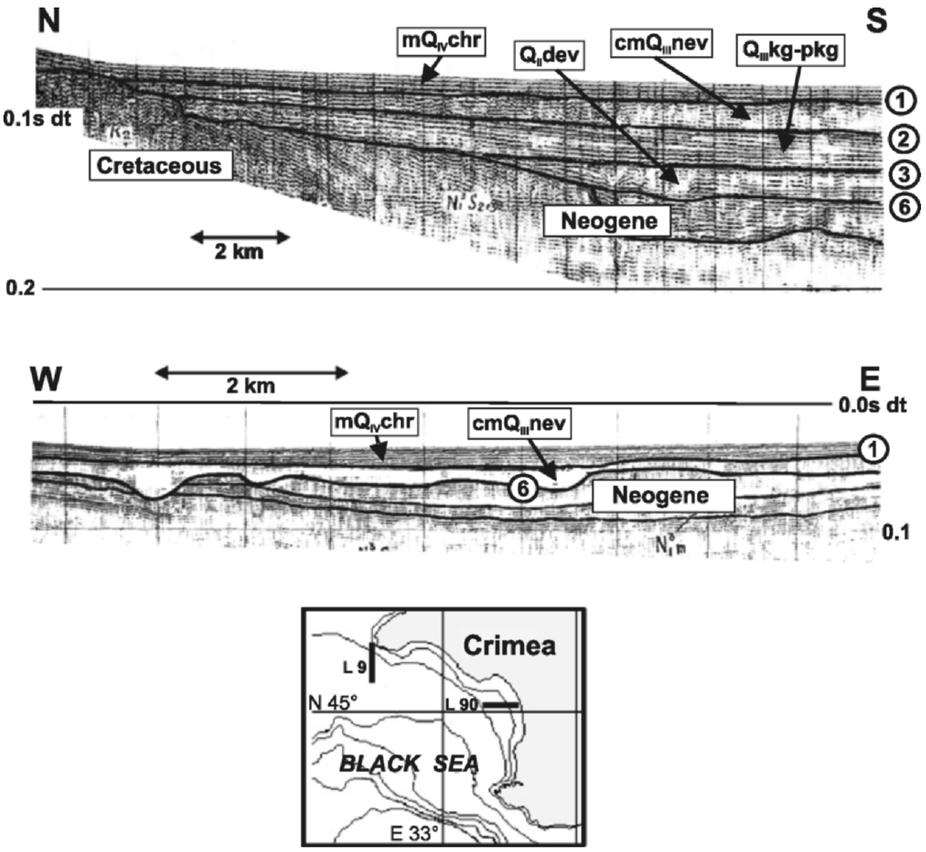


Figure 20. Principal reflectors and seismic units of the West Crimean shelf (top = a fragment of Profile # 09; bottom = Profile # 90).

3. DISCUSSION

The geology and Quaternary history of the Black Sea shelf outlined above are largely based on the existing conceptual framework (Arkhangel'sky and Strakhov 1938; Degens and Ross 1972; Fedorov 1977; Ostrovsky *et al.* 1977a, b; Shcherbakov *et al.* 1977a, b; Fedorov 1978a, b; Shcherbakov and Babak 1979; Shnyukov *et al.* 1979; Shnyukov and Makarenko 1981; Balabanov *et al.* 1981; Shcherbakov 1982a, b; Fedorov 1984; and others) and are consistent with evidence discovered more recently (Aksu *et al.* 1999, 2002a, b; Yanko-Hombach *et al.* 2002; and others). The currently advanced hypothesis of a catastrophic flooding of the Black Sea (Ryan *et al.* 1997, 2003) needs a separate discussion.

This hypothesis may be criticized from several perspectives: paleoclimatology, paleoceanology, Black Sea drainage balance, position of the Bosphorus sill, composition of Late Pleistocene and Holocene Black and Marmara Sea sediments, paleomorphology and seismic stratigraphy of the upper part of the geological sequence, as well as many other factors. All these criticisms were convincingly formulated by Aksu *et al.* (1999, 2002a, b). Their critical arguments are well in accord with the evidence and concepts advanced more than twenty years ago by Soviet scholars (including the present authors). Regrettably, these classical studies are not always sufficiently known to western scholars.

In the wake of mounting criticism, the proponents of the 'Flood scenario' have modified their initial views, taking into account new evidence, mainly from Russian publications. Ryan *et al.* (2003) now acknowledge two transgressive stages in the Black Sea: during the Younger Dryas (10–11 ky BP) from –105 to about –30 m, and a practically instantaneous one (at 8.4 ky BP) from –95 again to –30 m (sic!). Significantly, Major *et al.* (2002) discuss various possible positions of the Bosphorus sill and the related interaction of the Black Sea and Mediterranean basins.

In view of the ongoing discussion, the present authors make the following observations.

First, it should be stressed that the materials obtained during a single mission in 1993 are not sufficiently exhaustive for far-reaching conclusions. This mission included studies of only two areas that were limited in size and included the Kerch Strait, which is strongly affected by active litho-dynamic processes. The scope of analytical investigations was obviously insufficient for the revolutionary conclusions put forward in the initial publication (Ryan *et al.* 1997). Reliability of samples taken at depths ranging from –123 to –140 m should be treated with caution, as this area is located at the shelf edge and upper portion of the continental slope, places that are also strongly affected by active litho-dynamic processes, such as sediment redeposition (Glebov 1987). Detailed analysis of much more extensive and more representative materials from the southwestern Black Sea area (Aksu *et al.* 2002a) leads to opposite conclusions.

Second, the paper by Ryan *et al.* (2003) includes several inconsistencies. The quoted seismic sequences for the northern Black Sea area (Ryan *et al.* 2003:528, their Figure 2) were obtained by YMG and were interpreted by Glebov in 1981 (YMG Archive, File 15). Yet, these sequences have never been published in the form reproduced by Ryan *et al.* (2003), neither in the cited paper (Esin *et al.* 1986) nor in any other. This initial scheme was subsequently modified based on a more complex interpretation of large amounts of new evidence; it resulted in a new stratigraphic scheme of seismic reflectors (Figure 21), as well as a new understanding of their interrelations and geologic history.

Ryan *et al.* (2003) failed to notice that their initial interpretation of reflector stratification contradicts their fundamental idea about the 2-meter-thick drape of Neoeuxinian-Black Sea sediments being indicative of a rapid shelf submergence.

Considering the first unconformity (reflector 1) in the deeper part (Figure 21, top and bottom) as the Neoeuxinian/Holocene boundary, they had to enlarge the thickness of Old and New Black Sea deposits to 14–15 m. This would imply the occurrence of a relatively short (between 8–7 and 5 ky BP) yet large-scale regression at the Neoeuxinian-Holocene boundary with a drop in sea level in excess of 110 m. Such thicknesses of Black Sea deposits have never been found under normal conditions on the shelf. As stated above, the marine erosion between the Neoeuxinian and Holocene sediments has been found on the coastal part of the shelf to a depth of –34 m (Figure 21, middle). The maximum value of –45 m is a possible result of a recent tectonic subsidence.

Third, according to the initial hypothesis of Ryan *et al.* (1997), the Neoeuxinian Black Sea was catastrophically flooded by an influx of Mediterranean water over the Bosphorus sill more powerful than Niagara Falls. It filled the Pontic basin from a level of –150 m to –15 m in less than two years. This estimate lies within the margin of error for the absolute dating. The modified hypothesis (Ryan *et al.* 2003) suggests a catastrophic flooding at 8.4 ky BP with a practically instantaneous sea-level rise of 100 m. In view of the existing evidence, this is quite improbable, as by 9–7 ky BP, sea level had already reached –20 to –30 m (Shcherbakov *et al.* 1977b; Fedorov 1978a, b; Balabanov *et al.* 1981; Shcherbakov 1982a, b; Kuprin and Sorokin 1982; Fedorov 1984; Yanko-Hombach *et al.* 2002; and many others). This estimate is consistent with the present authors' findings (Figure 13, bottom).

According to the more probable scenario, Late Pleistocene sea-level rise in the Black Sea was due to a sharply increased influx of glacial meltwater rather than an abrupt penetration by the Mediterranean Sea. The period of 8–7 ky BP was marked by a hydrological balance shift in favor of the Mediterranean Sea, and not a catastrophic breakthrough of the latter into the former (Shcherbakov *et al.* 1977b).

The Neoeuxinian transgression had an oscillatory pattern, with oscillations of various orders proceeding against a backdrop of general sea-level rise (Fedorov 1977; Ostrovsky *et al.* 1977b; Balabanov *et al.* 1981; and many others). Yanko-Hombach *et al.* (2002) have convincingly proven the oscillatory pattern of the Late Pleistocene-Holocene transgression using biostratigraphic evidence. We confirm this based on the evidence of marine-littoral landforms (cliffs and offshore bars), Neoeuxinian seismic facies (indicative of deceleration, stillstands, and minor regressions), as well as good preservation of the antecedent topography (indicative of the accelerated, flood-like sea-level rise). The latter evidence is well known in other parts of the Black Sea and remains the subject of debates (Aksu *et al.* 2002a; Ryan *et al.* 1997, 2003).

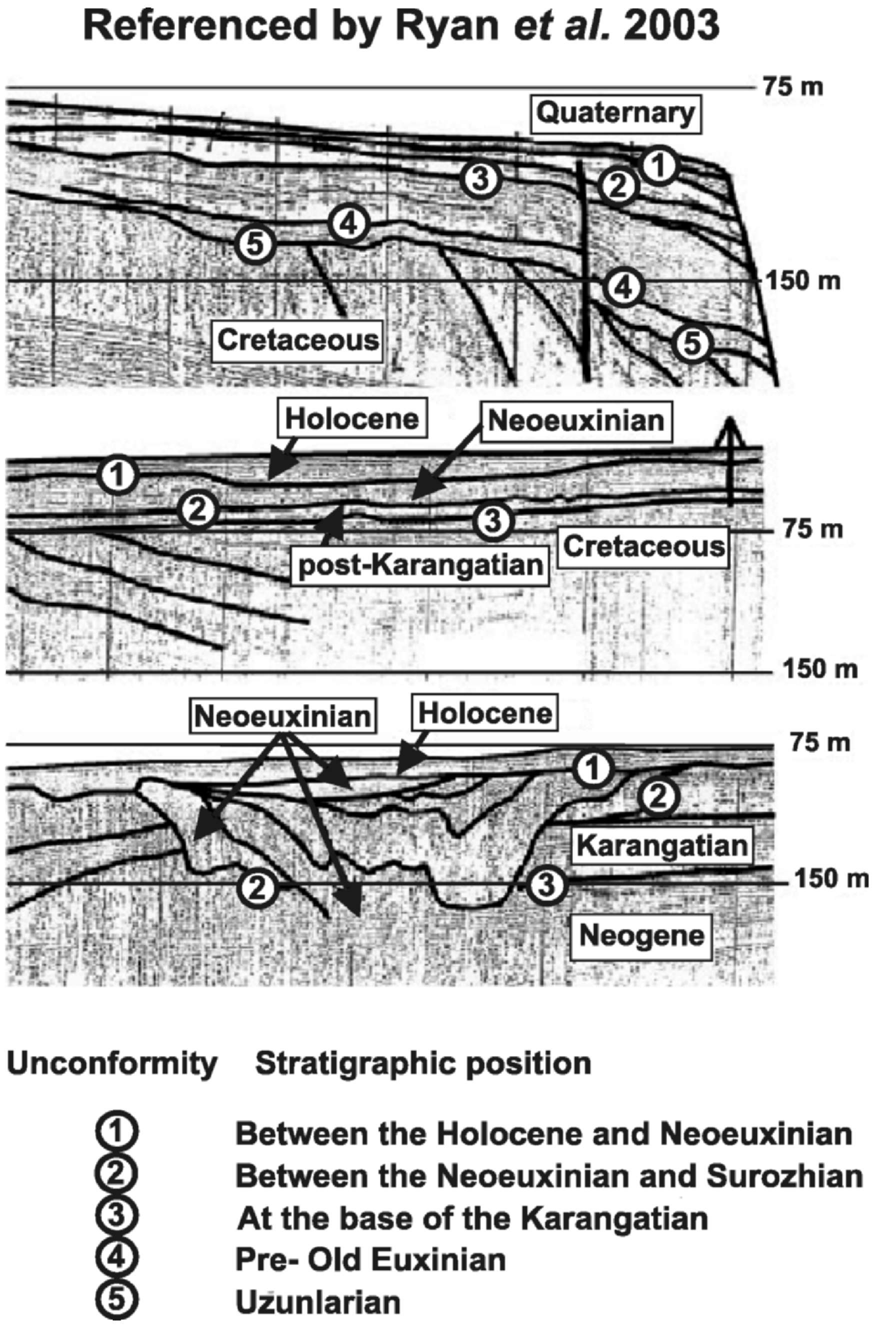
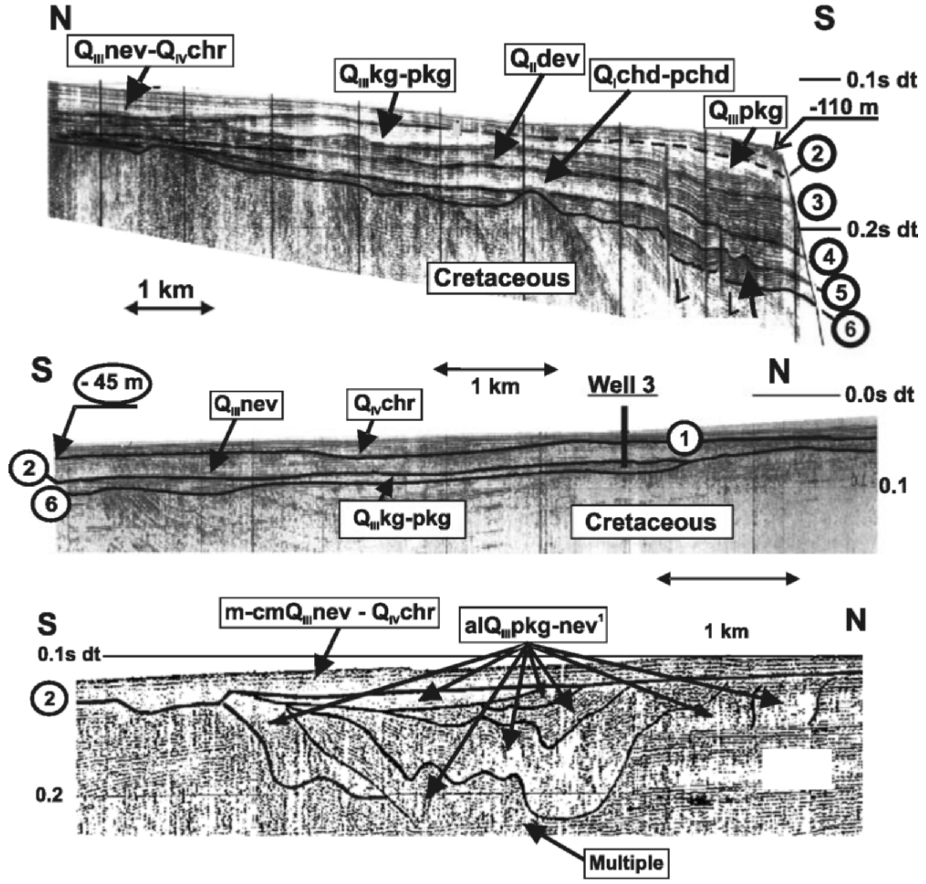


Figure 21. Comparison of fragments of seismic profiles (this page and facing).

Present paper
(Figures 7 top, 19 bottom, 16 top)



Unconformity	Stratigraphic position
1	Between Holocene and Neoeuxinian
2	At the base of the Neoeuxinian
3	At the base of the Karangatian
4	At the base of the Old Euxinian
5	Pre-Chaudinian
6	Bedrock (pre-Quaternary)

Like any other hypothesis, that of the Black Sea 'Flood' has a right to exist until either finally confirmed or rejected. Yet, this hypothesis is in no respect preferable to the existing ones. Simple calculations show that present-day river discharge alone (without the influx of glacial meltwater and spilling through the Bosphorus) could produce a sea-level rise of 100 m in the Black Sea basin in 80–100 years. This scenario implies a rapid rise (up to 1–1.5 m per year), yet not a catastrophic flooding. However, the bulk of the existing evidence argues against even such a scenario.

4. SCENARIO FOR THE FUTURE

Analysis of the Late Pleistocene-Holocene history of sea-level changes clearly indicates a global pattern involving transgression-regression cycles of diminishing amplitude from the Early Pleistocene to the present–that is, a curve of 'fading' or decreasing maxima and minima (Figure 13, bottom). This offers an optimistic scenario for the future, with low probability of global catastrophic floods. Recorded increases in oscillation frequency in the later Holocene are most likely due to greater resolution of biostratigraphic evidence from coring and sampling. In its entirety, the duration of the Holocene has been no more than one third that of the preceding Late Pleistocene (Riss-Würm) Interglacial; it seems likely that similar, yet undetected, oscillations occurred during its course. By extrapolation, one might anticipate future cyclical climate-controlled changes in salinity and sea level with a duration varying between a few hundred and a few thousand years. Strong and prolonged cooling, accompanied by a deep Black Sea regression, similar to those of the Upper Pleistocene Ice Age, may be expected no sooner than the next 20–30 thousand years.

Presently, we are witnessing a Black Sea transgression, and its immediate consequences should be discussed in more detail. A relatively rapid sea-level rise in the world ocean (at an estimated rate of up to 1 cm/year) resulting from the greenhouse effect, and the related problems of coastal protection against increased abrasion and inundation are being actively debated. It has been suggested (Gozhik 2001) that, with an increase of mean annual temperature by 3° C, sea level could rise 75–115 cm by the mid-21st century.

The present authors have shown (Esin *et al.* 1986) that the Black Sea shelf surface was formed during the Pleistocene-Holocene by a transgressing sea rising at the rate of *ca.* 1 mm/year (Figure 22A, B). According to our estimate,at 1.5 ky BP, the shoreline was located 8–13.4 m seaward from its present position. With the current abrasion rate, the shelf zone will be widened by 10–16 m at the expense of dry land in about 2000 years. According to the Odessa tide-gauge readings, the present rise in sea level varies from 1 mm/yr (1880–1920) to 2.5–3 mm/yr (1920–1960) (Gozhik 2001), which is similar to eustatic values.

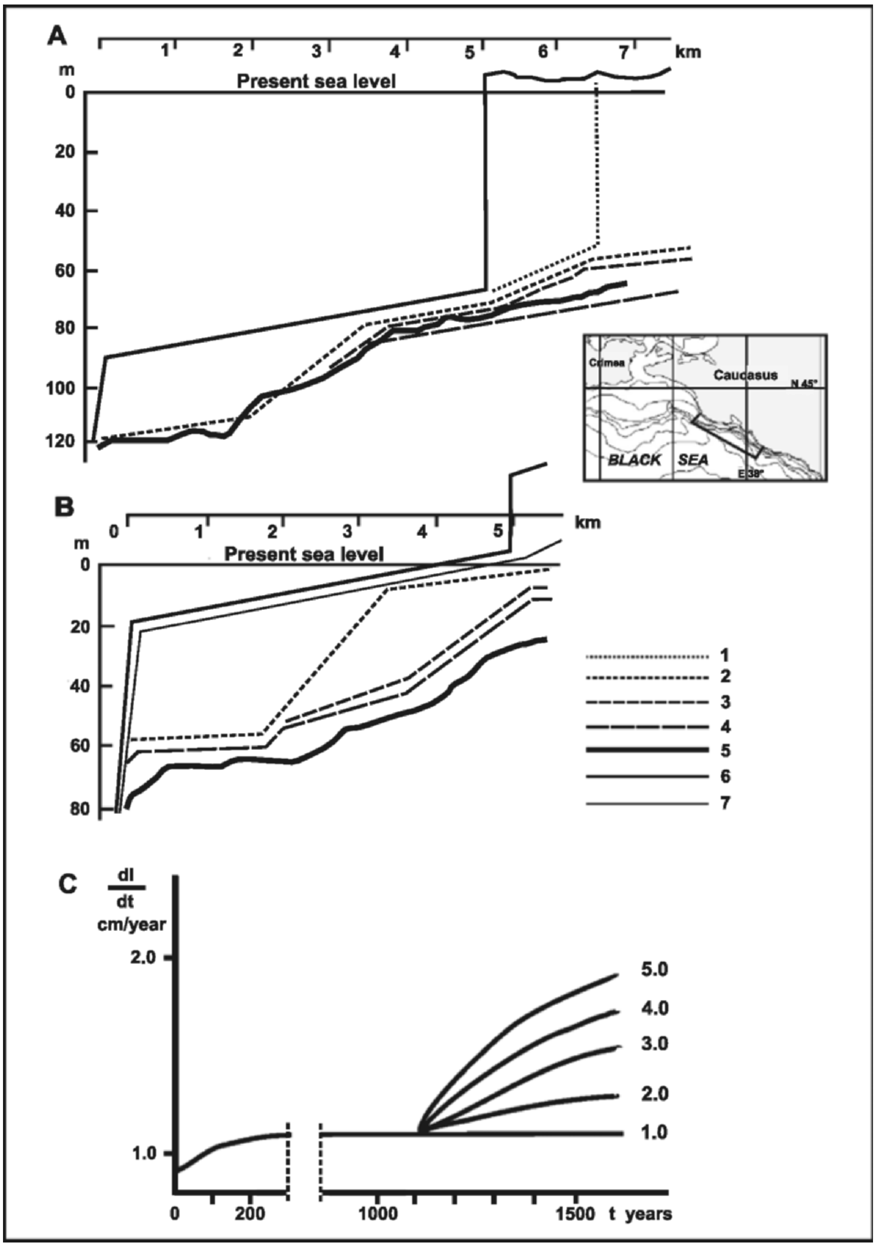


Figure 22. Scenarios (A,B) of shelf modification. Estimated shelf profiles by the end of transgressions: 1 = Chaudinian; 2 = Early Old Euxinian; 3 = Late Old Euxinian; 4 = Karangatian; 5 = Neoeuxinian; 6 = present profile of Pre-Quaternary shelf surface; 7 = initial, Pre-Chaudinian shelf profile; C = changes in the cliff recession rate at various rates of sea-level rise (from 1 to 5 mm/year).

It should be remembered that the predominantly erosional Russian Black Sea coast is currently receding due to a reduction in sediment supply. The coastline recession rate, depending on the rock composition, varies between 1–2 cm/year and 0.6–12 m/year. Esin *et al.* (1993) suggested a mathematical model that enables one to obtain quantitative characteristics of the abrasion as a function of sea-level rise (Figure 22C). According to this estimate, the abrasion rate will accelerate, ultimately increasing twofold. Similar processes are expected to occur across the Circum-Pontic area.

Sea-level oscillations, marine abrasion, and flooding of coastal areas cannot be discussed without considering recent tectonic movements. A common trend recognized for the Caucasian shelf has been occurring throughout the Quaternary against a background of Caucasian orogenic growth on one hand, and the depression of the Black Sea basin on the other: the shelf becomes narrower and turns into an actively fractured flexure, reworked by sea abrasion. This conclusion is confirmed by young faults detectable along the shelf edge (Figures 9 and 10, top). The observable shelf displacement toward dry land is likely due to expansion of the Black Sea basin. The tectonic-erosional destruc-tion of the shelf has proceeded at a variable pace, and it has been suggested that, in the future, the folded structures will be transformed into blocks. Contraction of platform-type shelves, the emergence of block structures, and the activation of flexures and disjunctions are also envisaged (Nesmeyanov and Izmailov 1995). Multidirectional block movements occurring along with stable, recent tectonic uplift are a characteristic feature of the recent epoch.

The oscillatory character of recent tectonic movements is indicated by the elevational differences between coeval terraces along the Black Sea littoral zone (Ostrovsky *et al.* 1977a; Fedorov 1978a; Nesmeyanov and Izmailov 1995). Analysis of Pleistocene shoreline elevations shows that only the positive recent tectonic movements exceed the amplitude of eustatic Black Sea level oscillations. Exclusively negative movements with an amplitude of –0.5 to 0.8 mm/year are characteristic of the graben-gulf bottoms of Tsemess and Gelendzhik Bays, as well as the areas of Anapa, Tuapse, Sochi, and Sukhumi (Blagovolin and Pobedonostsev 1973; Nesmeyanov and Izmailov 1995).

5. CONCLUSIONS

Seismic stratigraphy reliably reflects the main features in the geological structure of the Black Sea shelf and its Quaternary history. The foundation of the suggested model is likely to remain basically intact even in light of new evidence and a more accurate assessment of reflector ages; the key levels and their relations will not change.

This model reveals an oscillatory character for the Pleistocene-Holocene transgression. Its main elements were suggested 20 years ago (Shnyukov *et al.* 1979; Korsakov *et al.* 1982; Glebov and Skryabina 1984; Glebov and Sosnovsky 1984; Esin *et al.* 1986; Glebov *et al.* 1986; Glebov 1987; Glebov *et al.* 1996a, b, 1999; Glebov 1999), and its evidence is consistent with the most widely-accepted conceptual framework as well as newly gathered data.

The proposal by Ryan *et al.* (1997, 2003) of a catastrophic flooding of the Black Sea at 8.4 ky BP that includes a practically instantaneous 100 m rise in sea level is deemed improbable.

This paper includes evidence, previously unpublished in English, about the Neoeuxinian transgression maximum reaching an elevation of –8.5 m. The previously described level of –20 to –30 m likely corresponds to the marine erosion between the Neoeuxinian and Old Black Sea sediments.

The estimated fading of transgression-regression cycle amplitudes from the Late Pleistocene to the present, as well as the estimated rate of the present transgression, clearly signal the absence of global-scale catastrophes in the geologic future.

Based on existing forecasts of sea-level rise and the related shoreline destruction by abrasion, protective measures can be taken in a timely manner. Yet, the development of more accurate and high-precision models of future changes in the Circum-Pontic area necessitates detailed, complex investigation of the less studied areas as well as the key ones, focused especially on the Late Pleistocene and Holocene deposits and involving the detection and interpretation of seismic units of various orders. As a final test of the model, additional drilling and sampling in key areas are also needed. The establishment of an international body would be highly instrumental in the co-ordinated monitoring of the hydrological regime and the ecological situation in coastal areas of the entire Circum-Pontic region.

Apart from detailed studies of regression-transgression cycles in the Black Sea, other issues should be scrutinized, including the dynamics of hydrogen sulphide poisoning, regional seismicity, current seismic-gravitational processes, the modern manifestations of mud volcanism, hydrocarbon seepage on the sea bottom (Glebov *et al.* 2001), and the ecological setting in general.

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LANGUAGE DISPERSAL FROM THE BLACK SEA REGION

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Abstract: There are no known linguistic facts that would be best explained by a catastrophic flood or other major event involving the Black Sea in Neolithic times. This lack of evidence, however, could be due as much to our limited knowledge, both theoretical and factual, of linguistic paleogeography in western Eurasia as to the absence of such an event in antiquity. This paper will review what is known about the ages, origins, means, and causes of spread for all the Eurasian and Near Eastern language groups west of India (including now-extinct ancient languages), in an attempt to find any possible signs that the Neolithic linguistic geography of the area might have been affected by a Black Sea exodus or culture crash. The null hypothesis is that nothing out of the ordinary happened to languages in the vicinity of the Black Sea anywhere between the origin of food production in Southwestern Asia about 10,000 years ago and the dawn of history in Mesopotamia about 5000 years ago.

Keywords: language family, language spread, Indo-European, Afroasiatic, linguistic diversity, Caucasus, language extinction

1. INTRODUCTION

As basic background information, two important facts about the historical and modern distribution of language families in western Eurasia should be noted. First, the recent and modern diversity of language families there is notably low (Austerlitz 1980; Nichols 1992:232ff., 1997a). Second, the extent and frequency of large language family dispersals and language spreads there is notably high (Nichols 1992:123ff., 1997a, b, 1998).

Languages of the area in antiquity were not those of today but their

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ancestors (together with other languages now extinct). Therefore, the present focus will be on language families rather than individual modern languages, and specifically, it will concern the kind of family that can be called a *stock*, which is the oldest clade that is both demonstrably a clade and reconstructable by standard linguistic comparative method (Nichols 1992:24–25, 1997a:363, 2003a). A stock is demonstrable if it exhibits (1) shared paradigms, especially shared irregular paradigms (Greenberg 1960; Nichols 1996; Campbell 2003), or similar shared closed sets of forms, (2) identical or near-identical three-phoneme sequences in its basic vocabulary with significantly greater than chance frequency (Bender 1969), and/or (3) evidently cognate vocabulary with recurrent sound correspondences having significantly greater than chance frequency (Oswalt 1991; Ringe 1992; Nichols 1996; Kessler 2001). A stock is reconstructable if its basic vocabulary displays regular sound correspondences and if cognates can generally be distinguished from non-cognates. Examples of stocks include Indo-European, Semitic, Kartvelian, and Uralic. Isolates such as Basque, Korean, and Zuni are one-language stocks. Three languages of the ancient Near East–Sumerian, Hurrian-Urartian, and Hattic–are often classified as isolates because they have no attested sister languages, though, of course, in their own time they may have had sister languages that happen not to have been written. Afroasiatic, the ancestor to Semitic and its sister stocks, is demonstrably a family (Greenberg 1960; Newman 1980) but has not so far been shown to exhibit regular correspondences. Daniels (2004) gives a summary table of the main evidence for this, while Appleyard (1999) and Hetzron (1990a) cite some lexical evidence. At present, Afroasiatic is the only clear example of a demonstrable but not reconstructable family.[1]

A stock can have many major branches (as Indo-European does), only a few (as Uralic does, with a primary bifurcation into Finno-Ugric and Samoyedic), or only one (as isolates like Basque do). Young families lacking external kin, like Chumashan or Japanese-Ryukyan, are also one-branch stocks with recent divergence. A stock can have many daughter languages (as Austronesian does), only a few (as Kartvelian and West Caucasian do), or only one (as isolates do).

Because language is transmitted by learning and can be adjusted to the social and cultural context, all languages change constantly. As a result, the evidence crucial to demonstrating relatedness eventually drops out of usage or changes beyond recognizability. Evidence sufficient to prove relatedness rarely survives more than several millennia. Most of the oldest firm stocks are around 6000 years old (good examples are Indo-European, Austronesian, Uto-Aztecan, and the Semitic and Chadic branches of Afroasiatic). Under the right circumstances–chiefly, when a family is well attested in a number of daughter languages and happens to have a durable and distinctive grammatical "signature"–somewhat older families can be detected. Two such families fall into the greater Black Sea area, as will be discussed below. As a rule of thumb, then, language families can rarely be traced back more than about 6000 years, and probably never more than about 10,000 years, at least not by currently available methods. (These figures are very approximate; they are empirically based on estimates of language family ages and are therefore susceptible to change.)

Of course, genealogical descent lines continue back to the origin of language (whenever, wherever, and whatever that was), but most of this prehistory lies beyond the reach of scientific knowledge. Fortunately, the time frame of the proposed Black Sea flood is within the window of visibility, albeit barely.

Language families can be dated by various means, most of them quite approximate. It is common to reckon a family's age by comparing its internal lexical and grammatical diversity with that of well-known benchmark families such as Romance, Slavic, or Polynesian (1500–2000 years); Iranian, Mayan, or Oceanic (around 4000 years); or Indo-European or Austronesian (around 6000 years).

Glottochronology is a dating technique that assumes a constant rate of vocabulary loss or change and computes an age from the percentage of words, taken from a standard list of glosses, that are cognate and have the same meaning. Embleton (2000) provides a recent overview, and Trask (1996) gives a textbook presentation. In reality, of course, rates of change are not constant. Compare Icelandic, which has almost no loan vocabulary, with English, which possesses an everyday vocabulary that is about half Romance and a glottochronological wordlist with several Romance and Norse items. Both English and Icelandic are equally removed in time from ancestral Proto-Germanic. Conservatism vs. innovation are generally evident in linguistic data, however, and in the case of English vs. Icelandic, a cursory inspection of either the historical or the linguistic facts would show that English has had an accelerated rate of change due to intensive contacts, while Icelandic remains conservative due to isolation and deliberate choice of an archaizing literary standard. Glottochronology is a binary procedure, and its accuracy improves with the number of daughter languages that can be compared pairwise. English stands out in a glottochronological survey of Germanic languages just as it stands out in a typological survey. When the facts are critically assessed and the outliers (like English) weighted appropriately, glottochronology can be useful though approximate. Alpher and Nash (1999) use the same wordlist and assumptions for reasonably successful subgrouping of Australian languages.

The foregoing describes what may be called raw or uncalibrated glottochronology. Embleton (1986, 1991) proposes a calibration based on typology, linguistic geography, and areal interaction which gives remarkably accurate results but requires labor-intensive data collection (including typological and cultural data), coding, and computation. (Below, calibrated dates are estimated based on her findings.) Lohr (1999) and Dolgopol'sky (1964) propose shorter gloss lists with supposedly less variation in retention rates. Gray and Atkinson (2004) use a Bayesian method which allows for variable retention rates, but it still badly needs calibration (as shown by the fact that their application to the independently very well dated Indo-European stock yielded absurd results).[2] Nichols (nd) shows that the rate of loss of vocabulary items in particular semantic fields varies with language type.

When an ancestral language can be identified with a well-dated archaeological culture, dating becomes more confident. Well-known successes include the identification of Proto-Oceanic with the Lapita Culture of coastal Melanesia (Kirch 1997) and the identification of Proto-Indo-European with the first use of wheeled transport by the Pit-Grave Culture (Anthony 1995; Darden 2001). In the latter case, it is Proto-Indo-European in the narrow sense, excluding its Anatolian branch, that has a native vocabulary for wheeled transport and is firmly datable.

2. LINGUISTIC DIVERSITY AND SPREADS IN THE BLACK SEA AREA

In Neolithic Mesopotamia and Anatolia, the diversity of language families must have been considerable. This is probable because, in the third to second millennia BC, the density of families as calculated from only the attested languages is high (Nichols 1992:232ff.), and the attested languages were likely but a fraction of the languages and families that existed in the area. Now, the diversity in this area is low, as languages have been absorbed in several post-Neolithic spreads of state and imperial languages. In third millennium BC Mesopotamia, Akkadian absorbed Sumerian and no doubt several other languages. In the first millennium BC, Aramaic absorbed Akkadian and perhaps many others. Finally, beginning in the seventh century AD, Arabic absorbed much of Aramaic.

In Anatolia, the Anatolian branch of Indo-European must have absorbed several earlier languages (Hittite is known to have absorbed Hattic); its daughters and the rest of the indigenous linguistic diversity were effaced by the later spreads of Greek (an Indo-European branch) and then Iranian (also Indo-European) and Turkic languages. West of the Black Sea, the pre-Indo-European languages were displaced or absorbed by Indo-European spreads, and the early Indo-European Balkan languages were absorbed in later spreads, most recently that of Slavic in the 5th–6th centuries. North of the Black Sea there has been a series of large-scale spreads on the steppe: early Indo-European, then the Iranian branch of Indo-European, then Turkic languages, then (incipiently) Mongolian, and finally Russian and Ukrainian.

These various spreads have all involved language families that were not in the area in Neolithic times but have entered later either from the east (Indo-European, Turkic) or south (Semitic). Absorbing prior languages by language shift, they have removed most of the evidence we might have had on the linguistic consequences of the Black Sea flood. Only to the east of the Black Sea, in the Caucasus, do we have pre-Indo-European survivors: the three indigenous language families of the Caucasus–West Caucasian, East Caucasian, and Kartvelian–all greatly different from each other and the other languages of western Eurasia. They can tell us something about the pre-Indo-European linguistic character of the Black Sea area and are the first place to look for evidence of a Black Sea flood.

3. THE CANDIDATES FOR FLOOD-TRIGGERED DISPERSALS

Could a disastrous Black Sea flood account for the spread of the language of an early farming society? This attractive hypothesis assumes that the flood scattered populations at the northwestern frontier of farming. The flood refugees moved outwards, bringing agriculture to their northern, western, and northeastern neighbors, who were still hunter-gatherers. (The hypothesis also requires that the refugees were able to save their seed stock and livestock and resettle in new territory as farmers; this will be assessed below.) This paper will now review what is known about the age, likely origin, and early history of the language families attested in the greater Black Sea area. A language family is a candidate for a flood-triggered dispersal if it began to diverge around 8000 years ago, originated near the Black Sea, and experienced a significant early dispersal. Families will be reviewed in order of their approximate ages.

3.1 Afroasiatic

The world's oldest known language family (but not a stock; see above) has, by now, spread all across northern Africa and the Near East. Its daughter branches are Semitic (e.g., Arabic), Egyptian/Coptic (now extinct), Berber (e.g., Tamazight), Chadic (e.g., Hausa), Omotic (e.g., Dizi), and Cushitic (e.g., Somali). (Whether Cushitic is a single branch, two or three branches, or a residual category is debated.[3])

The age of Afroasiatic is unknown, but its major branches are of at least Indo-European-like age, so its dispersal must have been at least as early as the Black Sea flood and quite possibly earlier (Hetzron 1990a:649; Appleyard

1999). Newman (1980) mentions the possibility of 10,000 years, and Militarev (2000) calculates a glottochronological date on this order. Since we do not yet have firm sound correspondences for Afroasiatic, glottochronology, which requires precise identification of cognates, is not applicable here.

Afroasiatic is on the whole strongly African in its grammatical and phonological typology, and its origin is generally taken to be northeastern Africa, where its greatest present genealogical diversity is found. A minority opinion sees Proto-Afroasiatic as spoken by one of the first farming communities of Mesopotamia, from which it dispersed with the spread of farming (e.g., Militarev 2000; for discussion, see Appleyard 1999; Voigt 1999; and especially Ehret *et al.* 2004). Even if a Mesopotamian origin is assumed, however, Afroasiatic is a poor candidate for a Black Sea dispersal, as Afroasiatic languages have never been close to the Black Sea and certainly have never spread north of it. In protohistorical and historical times, languages of the Semitic branch have spread northward into Mesopotamia, not southward from there.

3.2 East Caucasian (a.k.a. Nakh-Daghestanian or Northeast Caucasian)

At about 8000 years (an estimate of a calibrated glottochronological age), East Caucasian is the oldest datable language stock (Nichols 2003b, and work in progress). Though it is the right age, it originated in the wrong place: in the eastern Caucasus, probably in today's northern Azerbaijan and close to the Caspian Sea, a location over 300 km (200 mi) from the Black Sea. Though well diversified with over 30 daughter languages, it has spread very little and has never been attested beyond the Caucasus highlands, foothills, and (at least in recent times) foothill/lowland interface. Its reconstructable vocabulary includes terms for the earliest domesticates but not later ones, and cultural continuity to historical times makes it likely that an East Caucasian language was spoken by the ancient occupants of the archaeological site of Chokh in southern Daghestan, one of the earliest farming communities outside of Mesopotamia (Zohary and Hopf 2000:221; for cultural and likely linguistic continuity in the area, Gadzhiev 1991:16–19). The East Caucasian language Avar is now spoken in the areas around Chokh. Well ensconced at 1887 meters (6194 feet) on an upper tributary of the Sulak River, which empties into the Caspian, and with archaeological ties to earlier sites nearby, Chokh is unlikely to represent a fresh recent arrival from the Black Sea area. Settlement and resettlement of the Caucasus highlands, and spreads of dialects, languages, and cultural elements in the Caucasus seem always to have moved up river valleys, so the Chokh environs are likely to have been settled from the east.

Another obstacle to assuming a Black Sea origin for this culture is the considerable ecological diversity along the southern flanks of the Caucasus; some time lag would be expected due to local adaptation along the way, and there is no beeline route from the Black Sea to Chokh. As of about 8000 years ago, domestication had spread radially from Mesopotamia and is attested to the north and northeast in both the eastern and western parts of the south Caspian region. The most parsimonious view of the origin of Chokh is that it is part of this general diffusion. Thus, Proto-East Caucasian appears to have diversified and taken root in the eastern Caucasus foothills and highlands as an early consequence of the initial spread of agriculture from Mesopotamia.

3.3 Indo-European

This ultimate ancestral stock of English had probably the largest ancient inland range of any language family, and in most of its range, its daughters were the actual majority spoken languages on the ground and not just an imperial or trade superstratum.

On several kinds of linguistic evidence, Indo-European is about 6000 years old. Its earliest attested daughters–Hittite, Sanskrit, and Mycenaean Greek–are transparently similar to one another and unlikely to have diverged for more than two or three millennia before their attestations: Hittite is attested from the 17th century BC, Mycenaean Greek from the 13th, and Vedic Sanskrit from the 6th, though it reflects the codified and carefully transmitted language of about 1000–1500 BC. The reconstructable Proto-Indo-European vocabulary includes words for ‘wool’, ‘yoke’, and other aspects of ox traction, and the large subclade that excludes the Anatolian branch also has a series of words for wheeled transport and a word for ‘horse’ (Anthony 1995; Darden 2001). The family must have originated in an area where wool and ox traction were known before wheels and horses, and where wheeled transport was known early, all of which points clearly to the Pit-Grave Culture of the Black Sea Steppe, where wheeled transport first appeared about 5500 years ago (Mallory 1989; Anthony 1995; Darden 2001) and wool somewhat before that.

Indo-European underwent a wide spread, but this was about 2000 years too late for our Black Sea flood. It is not even very likely that the distant ancestors of the Indo-Europeans came from the Black Sea area. The cultural antecedents of the Pit-Grave Culture appear to lie to the east, in the Khvalynsk Culture of the Caspian steppe (Mallory 1989:206ff.; Darden 2001 citing Gimbutas 1997:56). The best candidate for a first-order sister to Indo-European is Uralic (Oswalt 1991; Ringe 1998; but see Kessler and Lehtonen 2006), which originated east of the Urals. Janhunen (2000, 2001:63), however, notes the great typological distance between Proto-Uralic and Proto-Indo-European and does not advocate genetic relatedness.

There is also evidence of diffusion, not necessarily direct, between early Indo-European and early Uralic (e.g., Campbell 1990, 1998 and references therein; but again, see Janhunen 2001). Indo-European shares the root consonants of its personal pronouns not only with Uralic but also with Kartvelian of the southern Caucasus as well as several language stocks of Siberia (Nichols 2001; Nichols and Peterson 2005). The personal pronoun consonantism pattern is unique to Eurasia, and therefore it is a very strong indicator of provenance. It involves *m* as first consonant in a first person singular form and an apical obstruent in the second person singular:

Indo-European:	English	I	me	thee
	Latin	*ego*	*me*	*tu*
Kartvelian:	Georgian	*me*	*me*	*shen*
Finno-Ugric:	Finnish	*minä*		*sinä*
Eastern Siberian Isolate:	Yukagir	*met*		*tet*

For all of these reasons, the ancestor of Indo-European is likely to have arrived in the Black Sea steppe area from the east.[4]

3.4 Semitic

This family, well over 5000 years old, originated in the Arabian Peninsula. The inscriptional evidence of Akkadian and Eblaite, distinct East Semitic languages of the mid-third millennium BC, shows that the initial branching into East and West Semitic and the differentiation of East Semitic had already occurred by that time. Akkadian is attested in southern Mesopotamia before 2500 BC. It gradually absorbed Sumerian and spread through most of Mesopotamia beginning about 1900 BC as the Babylonian and then Assyrian imperial language. In the mid-first millennium BC, Aramaic, another Semitic language, replaced Akkadian throughout Mesopotamia, and in the first millennium AD, the spread of Arabic reduced Aramaic to a series of enclaves.

Semitic is too young to have been dispersed by the Black Sea flood, and it originated very far from the Black Sea (see Hetzron 1990b, 1997; Faber 1997).

3.5 Kartvelian

This is the family of Georgian and its three sisters, indigenous to the southern Caucasus and distinctive in morphological type. It is probably somewhat over 5000 years old, an estimate of a calibrated age. (The raw glottochronological age is at least 4000 years: Gamkrelidze and Ivanov 1997:777fn. 19, citing Gamkrelidze and Machavariani 1965:17.)

The Kartvelian family has spread only modestly from its area of origin in the central Georgian foothills. (For Kartvelian, see Klimov 1998; Harris 1991; Gamkrelidze and Ivanov 1995:777; descriptive grammars include Hewitt 1995; Tuite 1997; Cherchi 1999.)

3.6 West Caucasian (a.k.a. Northwest Caucasian and Abkhaz-Adyghe)

This is another small but fairly old family with five attested daughter languages, of which non-linguists may have heard of Abkhaz and Circassian. West Caucasian probably originated near the northeastern Black Sea coast (its place of greatest diversity) and spread very little until Circassian and Kabardian displaced or succeeded Alanic (Iranian) on the north Caucasus plain in the Middle Ages. The family is phonologically and morphologically highly distinctive in Eurasia and in general, with a very large consonant system, very few vowels, very complex consonant clusters, multiple prefixation, head-marking morphosyntax, and a morpheme canon consisting of little more than an onset (often complex) and a gloss (see Kuipers 1975; Hewitt 1989, Hewitt and Khiba 1989; Colarusso 1992; Smeets 1984; Chirikba 1996; Jaimoukha 2001). The type is so unusual that it is tempting to see it as a sole survivor of something now utterly extinct in Eurasia, but it could as well have originated in a Minoan or Trojan trade colony or Cucuteni-Tripolye remnant as in a flood survivor. Gensler (1993) finds some possibly diagnostic structural resemblances between West Caucasian and the grammatical type now found in far western Europe and North Africa, which would also be consistent with this survivor scenario (as is discussed further below) but does not point specifically to a Black Sea source.

To summarize, no attested language family of western Eurasia originated in the right time and place to have been dispersed by the Black Sea flood. No language family has even two of the three criteria of right time, right place, and significant early dispersal. The families with the widest dispersals are Indo-European and Afroasiatic, both of which originated far from the Black Sea. The Indo-European dispersal is 2000 years too late for the flood, and that of Afroasiatic may have been too early.

The fact that major linguistic spread zones extend along the Black Sea coast to both north and south, however, means that most of the early Circum-Pontic languages have gone extinct. For all we know, at around 6000 years ago, when the major language spreads were beginning, the Black Sea may have been ringed with the daughters of a then-2000-year-old family which dispersed because of the flood. Their other sisters may have spread far into eastern and central Europe, only to be absorbed later by Indo-European. Except for the

Caucasus, whatever languages were spoken around the Black Sea in pre-Indo-European times have long since gone extinct without written attestation, and we will never know what stock or stocks they belonged to.

4. THE SOCIOLINGUISTICS OF LANGUAGE SPREADS AND DISPERSALS

It is a truism that during language shift people gradually abandon a local, less prestigious, or economically less useful language for an inter-ethnic, prestigious, or economically useful language. That is, they abandon their language for that of a larger or more privileged community. Assuming that there was a Black Sea flood, let us hypothetically trace the likely fate of the people it displaced. Farming communities on what is now submerged shelf may have been wealthy and powerful, and their language sociolinguistically dominant, but as farmers, their wealth was in their land, and they lost that when they were flooded out.

The inundation probably occurred before the development of animal traction (Darden 2001:192–193), so the refugees brought with them only what they could carry and what livestock they could keep together. Just how much they could have brought depends on how fast the water rose. Whether they could have saved any seed stock depends on the time of the year. Those affected first were farthest from the present coast and probably had no economic or kinship ties to communities living above the present water level, so they are unlikely to have prospered in their places of refuge. Those with such ties were closer to the present coast, but by the time they had to leave, the societies into which they moved were already stressed by the presence of those who had fled earlier.

Thus, the flood refugees faced economic hardship or even ruin, negotiating for the right to settle on the lands of others, their population probably scattered. They are likely to have reverted to hunting and gathering at least for a time. As disadvantaged refugees, they may well have contributed to the spread of genes and some cultural elements, but their language was not socio-linguistically advantaged and would probably have died out within a few generations. A relevant comparison is the eruption of Thera around 1645 BC, which triggered a tsunami that destroyed the Minoan fleet and ash clouds that caused crop failures over a larger area. Far from triggering a spread of Minoan civilization and its language, this disaster seems to have weakened the civilization and contributed to the extinction of the language, which was replaced by Mycenaean Greek from the mainland. Cultural absorption and language shift were the fate of the Minoans and very likely also of any Black Sea populations impacted by the flood.

5. THE LINGUISTIC SIGNATURE OF EARLY DOMESTICATION ZONES

There seems to be a consistent linguistic profile in the regions of first independent domestication. In and around the domestication zone, one finds several well-entrenched old language stocks, testifying to early population growth and sociolinguistic strength as a result of domestication (Diamond and Bellwood 2003). This can be called a *stock cluster*. These stocks, and often other nearby languages, exhibit strong linguistic areality (grammatical and lexical convergence due to long contact) that appears to be ancient. Often, we can identify what will be called a *catalyst stock*, sometimes but not necessarily part of the stock cluster, which seems to have spread far beyond the domestication zone as part of the initial spread of domestication.

The best-studied case is Mesoamerica, where several old stocks (Mayan, Otomanguean, Totonacan, Mixe-Zoque) and several isolates (Huave, Cuitlatec, Tarascan, Tequistlatec, Xincan, Nahuan) display pronounced and distinctive areality (Campbell *et al.*1986). The Uto-Aztecan stock, which does not display these areal features and may or may not have been part of the original stock cluster, has spread far to the north, apparently with the spread of maize agriculture (Hill 2001), and can be taken as the catalyst stock.

In Southeast Asia, the center of rice domestication, the very old Austroasiatic language family is an evident catalyst stock (Blust 1994). The stock cluster has largely been overrun by the expansion of Chinese but apparently includes the Daic and Hmong-Mien stocks as remnants, and possibly also Pre-Proto-Austronesian. The question of areality is not a straightforward one. Austroasiatic and Austronesian, and probably also early Sino-Tibetan, have important structural features in common (simple syllable structure, limited inflectional morphology marked mostly by clitics or isolating formatives, minimal or no number inflection). Perhaps secondarily, a distinctive morpho-syntactic type has centered on Southeast Asia and the stock cluster: a radically isolating type with minimal inflection, monosyllabic roots, and tones. This type now affects all of Daic and Hmong-Mien, the Chinese branch and some parts of the Tibeto-Burman branch of Sino-Tibetan, and Vietnamese (Austroasiatic).

In northern China, the center of millet domestication, the catalyst stock in received view is Sino-Tibetan. The stock cluster has been mostly overrun by the Chinese spread, but it may well include Japanese and Korean. The areality is now obscured by the later shift of Chinese to the Southeast Asian type, but the indigenous ancient areality would appear to be the mildly agglutinating, strictly head-final type exhibited by Japanese and Korean, and by languages of traditionally non-agricultural peoples of Manchuria and Siberia (the Tungusic family, whose daughters include Manchurian, is an example).

In New Guinea, where horticulture developed independently in the highlands over 10,000 years ago, linguists regard the Trans-New Guinea group, whose membership and description are still being put together, as a possible catalyst group (it cannot yet be labeled a stock) responsible for the spread of horticulture through the highlands (Ross 1995, 2005; Pawley 1998, 2001). There is strong areality in the highlands, involving case marking, ergativity, clause chaining with switch reference, and two-tone systems.

In Africa, various plants were independently domesticated in the western reaches of the savannah zone that lies between the rain forest and the Sahel. Here, the catalyst family would appear to be Niger-Congo in the narrow sense, comprising a set of families with resemblant gender systems that must therefore be related: Benue-Congo, Kwa, Gur, Adamawa-Ubangi, Kru. The stock cluster includes several West African stocks that are often classified in Niger-Congo, though relatedness has not been demonstrated: Dogon, Ijoid, Mande, and possibly others. Areal features include simple to isolating morphology, tone systems with downstep, and verb serialization. These are found in the stock cluster and in some of the Niger-Congo languages.

In North Africa, where cattle were domesticated early and perhaps independently of domestication in the Near East, the catalyst family is Afroasiatic and the stock cluster has largely gone extinct in the region of early livestock use. As a result of post-Neolithic desiccation, the domestication center–today's Sahara Desert–emptied out linguistically and ethnically and is now a major spread zone. A number of stocks and probable stocks now ring the southern periphery of the Sahara, and some of these probably belong to the stock cluster: e.g., (from west to east) Niger-Congo, its putative western affiliates, Songhay, Saharan, Central Sudanic, the isolate Fur, Kadugli-Krongo, Kordofanian, East Sudanic. (A number of these are often grouped together as Nilo-Saharan, a grouping never demonstrated. For the evidence *pro*, see Bender 1989, 1991, 1997.) It is difficult to factor areal features out of these stocks from larger areal features of Africa, but they probably include fairly complex inflection, including internal change as a common marking type, complex plural marking on nouns, and verb-initial word order.

Domestication of reindeer occurred about 4000 years ago in southeastern Siberia to northwestern Manchuria, probably stimulated by knowledge of horse domestication to the west. The large spread of the Tungusic language family is a likely result. This family belongs to a distinctive and strong areal cluster including most language families of Siberia and especially the Turkic and Mongolian families. The resemblance among Tungusic, Turkic, and Mongolian is strong enough to have persuaded many investigators of their genealogical relatedness (under the rubric of Altaic), though when proof of the relatedness is sought, it proves difficult to exclude the less similar isolates Japanese and

Korean from the grouping (which is not demonstrated in any case). The bibliography on Altaic is large; for recent literature, see discussions *pro* in Vovin (1999) and *con* in Janhunen (2001) . The areality also affects the Uralic stock and Yukagir, an isolate or small family, groups for which an Altaic classification is generally not proposed.

The situation in Mesopotamia, however, is different from all of these, at least on the surface. All three diagnostic elements seem to be lacking. There is no diversity of well-entrenched ancient stocks surviving now in Mesopotamia, no clear catalyst family, no evident areality linking the four lineages attested in and near ancient Mesopotamia (Sumerian, Elamite, Semitic, Hurrian/Urartian), and no deep-seated areality surviving today. The chief reason for this is that subsequent spreads of Semitic, Iranian, and Turkic languages have completely effaced the original indigenous languages of Mesopotamia. The location of Mesopotamia in close proximity to a desert, the most drastic type of spread zone, and the long history of empire in the area are responsible for these secondary spreads. If the stock cluster and catalyst stock had survived, it would be relatively easy to identify likely survivors from the Black Sea flood zone.

Piperno *et al.* (2004) find that sedentary foraging with grinding and baking of wild grain was already evident by about 20,000 years ago in the Fertile Crescent, long before domestication. This intensification probably entailed population growth, so that the Fertile Crescent could have been a long-standing center with a slow outward spread of people, cultures, and languages beginning long before domestication. This makes the general lack of areality among the ancient Near Eastern languages even more puzzling. Perhaps the pattern evident in historical times, of language spread into rather than out of Mesopotamia, began much earlier.[5]

6. THE CAUCASUS AS AN ENCLAVE

The Caucasus is one of two linguistic enclaves in Eurasia that preserve the grammatical types that must have dominated Eurasia (or at least mid-latitude Eurasia) before the spread of Indo-European and its successor spreads along the greater Silk Road (Tuite 1996; Bickel and Nichols 2003). The great age of this early grammatical profile is shown by the fact that it appears to have survived in the Pacific Rim linguistic population (Tuite 1996; Bickel and Nichols 2003, 2005), which separated from mainland Eurasia before the Neolithic. Since agriculture spread to the Caucasus from Mesopotamia very early (Zohary and Hopf 2000: 221, 246), it is at least possible that part of the Mesopotamian stock cluster and even the catalyst stock survive in the Caucasus enclave.

Despite the conspicuous widespread presence in the Caucasus of a few structural properties unusual in Eurasia, such as moderately to very complex consonant inventories, greater than average inflectional complexity, and ergativity (of quite different kinds), there is really no Caucasian linguistic areality (Tuite 1999; Nichols 1999) that can be taken as the criterial ancient Mesopotamian areality. Rather, we may tentatively identify two different subzones in the Caucasus. The eastern one possesses phrasal predicates, simple syllable structure, complex noun inflection, minimal or no person marking, noun classification, clause chaining, and tones and phonation types. Some of these, notably the phrasal predicates and simple syllable structure, seem to be more generally ancient Southwest Asian features. In the western one, we find very complex consonant systems, complex syllable onsets, minimal or simple noun declension, multiple role marking and complex verb inflection, prefixal or proclitic person marking, and finite subordination with conjunctions. Some of the western features are very loosely echoed in the Balkan peninsula: multiple role marking, proclitic role marking, minimal noun declension, finite subordination. It is the western type that is strikingly exotic in western Eurasia and rare worldwide.

The East Caucasian language stock mostly reflects the eastern type, West Caucasian the western type, and Kartvelian a mix. This typological east/west division coincides with an archaeological distinction between the eastern and western Caucasus that begins before the Neolithic and continues to the early Middle Ages (Gadzhiev 1991:17). It is possible, but not at all certain, that one of the two types may reflect ancient Mesopotamian areality. It is even possible that the division into eastern vs. western types goes back to early Mesopotamia. (The Fertile Crescent plus Mesopotamia is a sizable area that must have harbored considerable diversity of language families before the rise of empires.) There is similarly no way to determine which if any of the indigenous Caucasian stocks is the catalyst stock, though East Caucasian, with its likely connection to the early agricultural site of Chokh, is a good candidate.

Mesopotamian domestication spread outward in four directions: southward to North Africa, westward to Anatolia and thence to Europe, northward to the Caucasus and thence to the western steppe, and eastward to Central Asia and thence to India. The Black Sea splits the westward and northward trajectories, and if the Black Sea shelf was inhabited by early farmers, these two trajectories may have been one. Gensler (1993), surveying 20 structural properties, finds strong and unique typological affinities between insular Celtic languages and northern African languages, chiefly Afroasiatic, and a small but striking cluster of affinities between these two and West Caucasian. The effects in Celtic, a structurally deviant Indo-European branch, are not inherited but apparently represent substratal or diffusional effects. There would seem to be three possible explanations for the Celtic-Afroasiatic affinities: (1) spread of agriculture up the Atlantic coast of Europe from North Africa, (2) later contacts between an

Afroasiatic-speaking Mediterranean seafaring culture (e.g., Phoenician) and western Europe–this scenario has been argued by Vennemann in several papers (e.g., 2001a, b)–and (3) survival of the western Mesopotamian grammatical type only at the far western frontiers of its spread, namely in North Africa and Atlantic Europe. (Gensler's pattern is in itself evidence for typo-logical affinity between ancient Mesopotamia and the ancient grassland that is now the Sahara, though it does not tell us whether that affinity was caused by the spread of agriculture or antedated it.)

On any of these three interpretations, evidence of the putative western Mesopotamian/western Caucasian structural type is surprisingly localized and weak in Europe. Granted that most of the Indo-European languages of today's Europe were not the first Indo-European languages in their present territories but have overrun and absorbed the Indo-European frontier languages. For a likely early core lexical substratum in the then-frontier Indo-European languages of central Europe see Schrijver (1997); for a detailed consideration of evidence of pre-Baltic and pre-Slavic Indo-European substrata in Baltic and Slavic, see Andersen (2003). Admittedly, the cultural, and therefore linguistic, impact of earliest Indo-European on pre-Indo-European farmers in Europe was unusually strong (Darden 2001:216ff.). Even so, one might expect more pronounced substratal effects of pre-Indo-European languages on Indo-European languages, especially if the two were of very different structural types.[6] Yet even Basque, a pre-Indo-European language of far western Europe, does not pattern with Celtic in Gensler's survey but is if anything reminiscent of the eastern Caucasian and Southwest Asian type. What is known about Etruscan seems to point in the same direction. Could it be that a Black Sea flood cut off or at least constricted the western Mesopotamian pipeline to Europe, and flow from the eastern pipeline filled the gap? This very speculative possibility seems to be the strongest linguistic evidence to be found for a catastrophe in the vicinity of the Black Sea at the time of the early expansion of domestication.

7. CONCLUSIONS

A Black Sea flood, if it occurred, is unlikely to have triggered linguistic and cultural dispersals of those flooded out. It is more likely to have contributed to the extinction of their language or languages. Nothing in the linguistics of western Eurasia demands a flood or other catastrophe at the time of early agriculture as its explanation, and no language family attested in Eurasia has spread from the vicinity of the Black Sea in the time frame of the early spread of farming. The near-absence of the putative western Mesopotamian and western Caucasian structural type in Europe may or may not require explanation, but if it does, extinction or weakening by an event or situation in the Black Sea region about 8000 years ago is one speculative interpretation that may be borne by the facts but certainly is not demanded by them. The location of the Black Sea, sandwiched as it is between the Anatolian and steppe spread zones, and the location of Mesopotamia next to the Saharan spread zone, have caused much possible evidence to be lost due to successive language spreading. What linguistic evidence can be recovered is not very conclusive on the question of whether the Black Sea flood occurred. On the other hand, a firm consensus among earth scientists as to the recent prehistory of the Black Sea would give linguists welcome information that we could apply in interpreting the linguistic facts.[7]

ENDNOTES

1. A group that is reconstructable but not demonstrable cannot be called a family or stock and is not necessarily a clade. With enough ingenuity, recurrent correspondences can be found between any group of languages; any number of speculative super-ancient "families" have been "found" by assuming (rather than demonstrating) relatedness and seeking recurrent correspondences. Two mutually incompatible reconstructions have been worked out for Afroasiatic: that of Orel and Stolbova (1995) and that of Ehret (1995). Since they are incompatible, at least one of them must be random, which proves that one can find recurrent correspondences that are random (Ratcliffe 2003; see also Appleyard 1999). Therefore, recurrent sound correspondences do not in themselves prove genealogical relatedness of languages.

2. Their Indo-European dating is 8700 years for its initial branching (the separation of the Anatolian branch from the rest), 7000 years for the separations of Tocharian and Greek-Armenian, and all major branching complete by 5000 years ago. But, as discussed again below, Proto-Indo-European had terms for wool and wheeled transport, things unknown until after about 6000 years ago. The Indo-European dissolution is so well dated that it should have been used to calibrate Gray and Atkinson's method.

Similar early chronologies for Indo-European have been advocated by archaeologists wishing to see the Indo-European distribution as a great singularity caused by the spread of farming from the Fertile Crescent (Renfrew 1987; main ideas reviewed in Darden 2001). In fact, it is no singularity but a good example of language spreading from the steppe, a regular phenomenon in Eurasia. The chronology of the Indo-European dispersal and of several of its main branches is very solid. Culture elements are firmly reconstructed and clearly belong to the late Neolithic steppe and not to early Mesopotamia, and the words for wool, transport, etc. clearly reconstruct to the protolanguage as lexical items in those meanings. The chronology of Indo-European is as clear as the chronology of early farming sites in Mesopotamia, and redating Indo-European so as to locate it in the early Fertile Crescent is scientifically as unsound as, for example, redating major early Neolithic archaeological sites such as Hacılar or Jarmo to 3500 BC just in order to link them to Indo-European.

3. For the Afroasiatic family tree, see the Linguist language resources area website: http://linguistlist.org/langres/index.html

4. Colarusso (e.g., 1997, 2003) argues for a Pontic origin of Indo-European and a deep genetic

connection to West Caucasian. The small vowel inventory of Proto-Indo-European and certain of its morphophonemic alternations are indeed reminiscent of West Caucasian (see Kuipers 1960, 1976; Gamkrelidze and Ivanov 1995:115ff.), but the morpheme structure canons and overall morphosyntactic types of the two stocks are very different. The mythological parallels adduced by Colarusso are convincing but hard to date, and may not require more than the already well established Indo-European presence in the western steppe at some date after the Indo-European breakup.

5. Akkadian is a good example of this process: an immigrant to Mesopotamia from the Semitic homeland to the south, it absorbed Sumerian and was then itself later absorbed by Aramaic. A similar pattern whereby languages of rural and nomadic peoples displace urban languages is also typical of Central Asia and may in fact be the general rule for oasis civilizations.

6. Work in progress by Peter Schrijver and Willem Vermeer finds evidence that Hattic, Minoan (the pre-Mycenaean language of Crete, attested in Linear A inscriptions), and the autochthonous substratal language to several central European Indo-European branches share typological features such as extensive verbal prefixation and head-initial word order, echoed in part in West Caucasian and possibly representing the language type of the first Neolithic farmers. Even this distribution is less, in geographic range and substratal impact, than might have been expected for the first expansion of farming.

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33

TIMING OF THE LAST MEDITERRANEAN SEA–BLACK SEA CONNECTION FROM ISOSTATIC MODELS AND REGIONAL SEA-LEVEL DATA

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Abstract: Water levels in the Mediterranean and Black Seas since the Last Glacial Maximum have varied substantially across the region because of the influence of the melting of the last great ice sheets in redistributing ice and water over the Earth's surface. This spatial variability is significant for discussions of the timing of water exchange between the Aegean and Black Sea, which reached *ca.* –10 m relative to present sea level at 12 ky calBP. In the absence of direct observational evidence, sea-level change at sill locations is predicted here using isostatic models that have been calibrated against observational data from other Mediterranean localities. If one assumes a depth of ~ –32 m for the Bosphorus sill, the Black Sea is predicted to have been reached by rising Mediterranean water between about 10.3 and 9.5 ky calBP. Alternatively, if the Bosphorus bedrock gorge at ~ –100 m depth is adopted as the sill, then the first Mediterranean influx over the shallower Dardanelles sill at ~ –80 m is predicted to have occurred between 15 and 13.7 ky calBP.

Keywords: sea-level prediction, glacio-hydroisostasy, sill height, Black Sea, Marmara Sea

1. INTRODUCTION

The connection between the Black Sea and the Mediterranean Sea via the Sea of Marmara remains a controversial issue. Was there a catastrophic flooding of the Black Sea when a rising Mediterranean Sea breached the Bosphorus sill as Ryan *et al.* (1997, 2003), Ryan a nd Pitman (1998), Major

V. Yanko-Hombach et al. (eds.), The Black Sea Flood Question, 797-808.

et al. (2002), and Ryan (this volume) have proposed, or was the connection one of gradual water exchange as described by Aksu *et al.* (2002), Hiscott and Aksu (2002), and Hiscott *et al.* (this volume)? And when did this event occur? Some researchers place the first post-LGM (Last Glacial Maximum) incursion of Mediterranean water into the Black Sea as early as 12,800 BP (one of the two scenarios presented by Major *et al.* 2002), while others date it later. It took place about 11,000–10,500 BP according to Aksu *et al.* (2002) and Hiscott and Aksu (2002), but it was later still at 9100 BP according to Kaminski *et al.* (2002). A date of 8400 BP has been proposed in Ryan *et al.* (2003) and in the second scenario of Major *et al.* (2002), and it was placed as recently as 7100 BP in Ryan *et al.* (1997).

If sea-level change is known or can be predicted, and the depths of the topographic barriers separating the Aegean and Black Seas are also known, then it becomes possible to predict the timing of the first contacts between the Aegean, Marmara, and Black Seas. The principal sources of uncertainty would be (1) the accuracy of the local sea-level curve, and (2) whether there has been significant modification of the sills by erosion, sedimentation, or local tectonics.

This paper employs a predictive approach to estimating the timing of inter-basin connection based upon isostatic models of sea-level change that have been calibrated against regional Mediterranean sea-level data. Previous studies have used the Barbados sea-level curve of Fairbanks (1989)–e.g., Aksu *et al.* (2002)–or earlier models by Milliman and Emery (1968)–e.g., Ryan *et al.* (1997)–or Chappell and Shackleton (1986)–e.g., Aksu *et al.* (2002)–to estimate the timing of the post-LGM flooding of the Black Sea by the Mediterranean without first testing whether these local sea-level functions are appropriate for the Black Sea sill location. In fact, as demonstrated by Lambeck (1995, 1996), sea levels from the Aegean to the Black Sea vary substantially because of the combined effects of glacio- and hydroisostasy, even in the absence of tectonics. Furthermore, the results can be expected to differ from observed values at Barbados or similar sites because Mediterranean and Black Sea levels were strongly influenced by the nearness of the former European ice sheet and by the coastline geometry of the ocean basins into which the glacial meltwater drained (Lambeck and Purcell 2005).

There appear to be no sea-level observations from areas near the Bosphorus and Dardanelles sills that controlled the exchange of water between the Aegean and the Black Sea, so instead we use a regional model for the Mediterranean that has been compared to sea-level data from tectonically stable regions. The calibrated model is then used to predict sea-level change at the sills based on the assumptions that (1) there was no tectonic contribution to the sea-level signal or that this has been corrected in some way, and (2) lateral variation in the effective parameters that define the Earth's response to changing water and ice loads on time scales of 10^4 years is unimportant.

The first assumption has been tested against sites where the position of the last interglacial shoreline is known. The second assumption must remain an article of faith until full three-dimensional mantle response models have been developed and tested against observational data. Until this is achieved, the following predictions must be based on a range of effective earth model parameters that yield predictions consistent with field data that have been obtained across the region. The data used include information from Israel (Sivan *et al.* 2001, 2004), from locations in Greece where vertical tectonic displacements are believed to be small (van Andel and Shackleton 1982; Lambeck 1996), from Italy (Lambeck *et al.* 2004), and from the French Mediterranean (Lambeck and Bard 2000). These data span the critical period during which the sills guarding the Black Sea were likely to have been breached by a post-LGM Mediterranean transgression.

2. PREDICTION OF MEDITERRANEAN SEA LEVEL

The model predicts sea-level change due to the growth and decay of the ice sheets of the last glacial cycle, i.e., from the last interglacial to the present. It incorporates the planet's deformation as well as the variations in gravitational field caused by shifting ice and water loads and their redistribution over the Earth's surface. Meltwater returns to the oceans, and as it does, the configuration of the ocean basins and the shape of their margins become time dependent, but throughout the process, the ocean surface remains equipotential. Ice loads are constrained by examining the response of the crust (mainly reflected by sea-level change) in the formerly glaciated areas, while the time history of the total ice volume is inferred from sea-level data far from the glaciated regions. The most recent discussion of the theory used here is in Lambeck and Johnston (1998) and Lambeck *et al.* (2003). The theory is consistent with that of, for example, Mitrovica and Milne (2003). Model parameters for the Mediterranean appear in Lambeck and Purcell (nd), and the ice-volume equivalent sea level used is that of Lambeck and Chappell (2001).

Figure 1 illustrates the predicted sea level (relative to today's values) in the Mediterranean and Black Sea for *ca.* 12,000 BP. It clearly shows the spatial variability in sea-level response that can be expected across the region. The dominant high frequency spatial signal is due to the changes in sea-floor loading by meltwater added since the onset of post-LGM deglaciation. This is seen primarily as a subsidence within the basins, such as the Black Sea and the Western Mediterranean, and a relative uplift of adjacent land bodies. Superimposed on this is a longer wavelength variation, mainly with a north-south trend, that is the planet's sea-level response to changes in ice load over northern Europe and North America, details of which may be found in Lambeck and Purcell (nd). Thus, the Black Sea will have a different sea-level response than, say, the Levantine coast, which lies farther from the former ice loads.

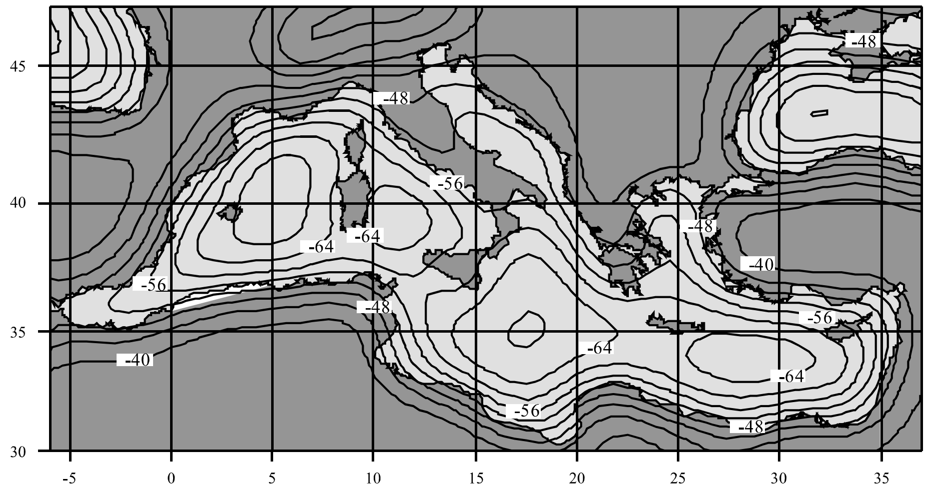

Figure 1. Predicted sea level across the Mediterranean at 12,000 calBP. Predictions assume tectonic stability of the crust but include glacio-hydro-isostatic effects based on realistic ice and earth models that have been calibrated against observational data from across the region (from Lambeck and Purcell 2005).

This spatial variability is further illustrated in Figure 2 as a time series at the four Mediterranean sites discussed below. In this case, predictions for Villefranche (on the French Mediterranean coast) and Tirins (Peloponnese, Greece) are very similar, but they differ from the Levantine coast result by ~10 m at 10,000 years BP. As a consequence, one should not use a sea-level curve from any other location in the Mediterranean as a proxy for change in the Bosphorus without first correcting for differential isostatic effects. Likewise, one cannot use the Barbados sea-level function unless it can be demonstrated that the isostatic effects are fortuitously the same for these sites as they are for the Black Sea sills.

In this paper, we calibrate the isostatic rebound model at locations in the Mediterranean where vertical tectonic displacements are small or known (from the MIS 5.5 shoreline elevation), we estimate effective earth and ice sheet parameters that describe the observed response, and then we interpolate to locations of particular interest, in this case the connection between the Mediterranean and the Black Sea. Figure 1 shows one result of such an interpolation for the Mediterranean and Black Sea basins as a whole.

In the next section, predictions are based on the Lambeck and Purcell (2005) model, and accuracy estimates depend upon uncertainties in both the earth- and ice-model parameters, as well as the accuracy of the global ice volumes for the post-LGM period, as discussed therein. The model predictions have been successfully tested against observations from outside the Mediterranean, includ-ing the Barbados data (see Lambeck *et al.*2002).

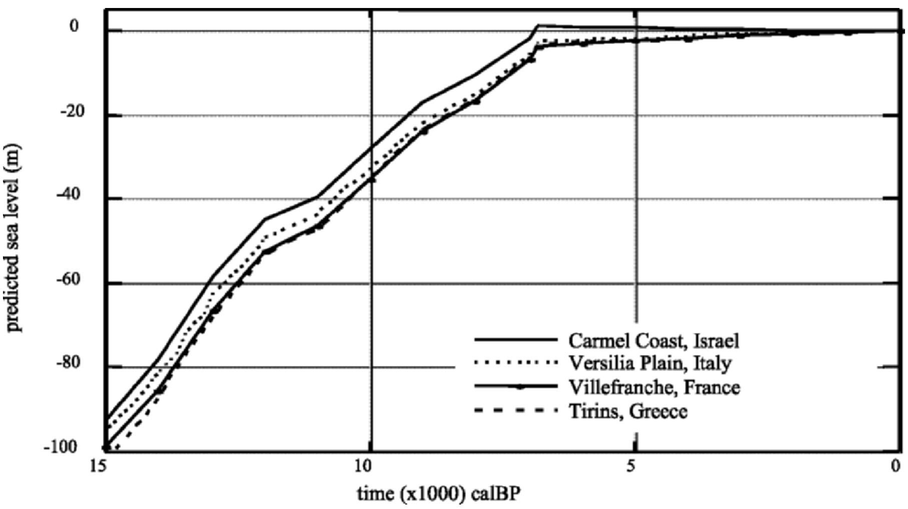

Figure 2. Predicted sea levels at four Mediterranean locations based on the model in Figure 1.

3. CALIBRATION OF THE SEA-LEVEL MODEL

Figure 3 illustrates predicted sea levels at several localities and compares them to observed values for the same locations. Figure 3A describes the Carmel coast of Israel. Here, underwater and coastal archaeological research has provided constraints on the local sea-level curve from ~9500 calBP to the present. The offshore evidence includes submerged prehistoric settlements, water wells, and shipwrecks whose ages can be established by ^{14}C dating (e.g., Galili and Weinstein-Evron 1985; Galili *et al.*1988; Galili and Nir 1993). The onshore evidence includes remnants of anchorages, slipways, piscinas, quarried ponds, and water wells (e.g., Raban 1981, 1983; Raban and Galili 1985; Galili and Sharvit 1999; Sivan and Galili 1999), but we have used only those indicators that can be related to sea level and whose age determinations are believed to be reliable. Generally, the predictions are consistent with the observational data, although the latter are limited to times after ~9 ky BP.

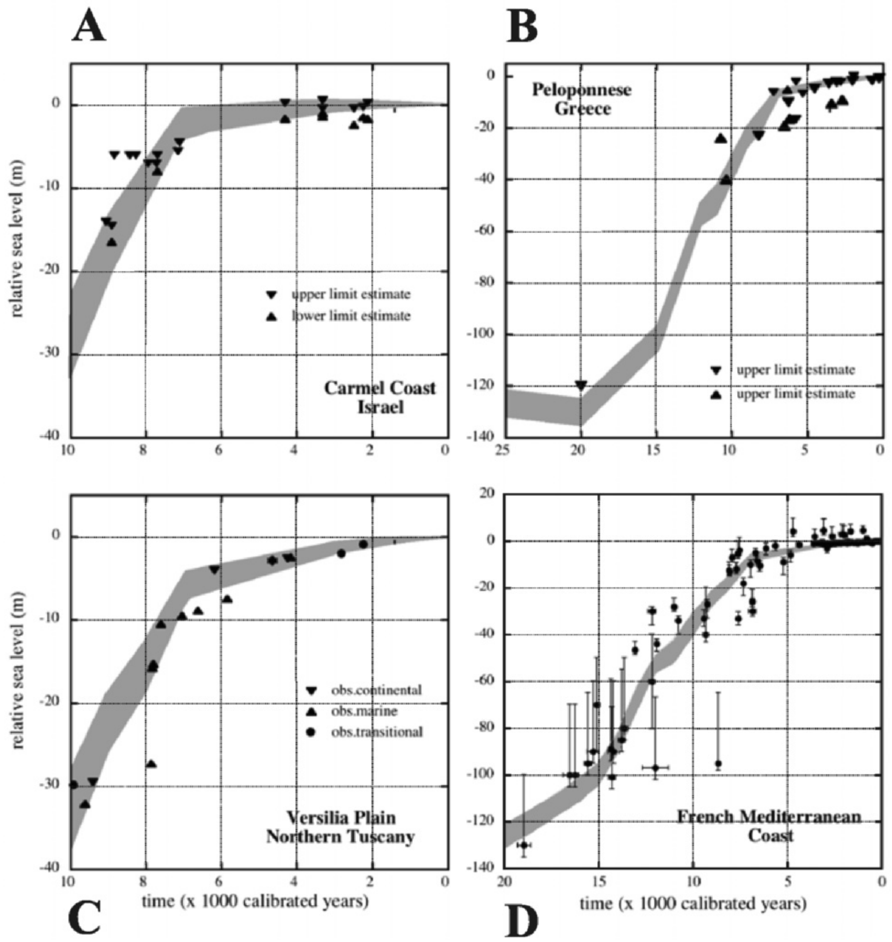

Figure 3. Comparison of predicted and observed relative sea levels at four Mediterranean localities. Upper and lower limits for predicted values (the grey zone) are based on uncertainties in the model parameters; for details, see Lambeck and Purcell (2005). Observational accuracies or upper and lower limits are shown where appropriate. Data sources are discussed in the text.

Figure 3B compares several sites in the Greek Peloponnese, where tectonic stability is suggested by the occurrence of the Last Interglacial sea levels at a few meters above present sea level (Kelletat *et al.* 1976; van Andel 1987). Data from several sites have been combined here: from the heads of the Argolikós, Messiniakós, and Lakonikós Gulfs, from Navarinou Bay (Kraft *et al.* 1975, 1977, 1980), and a LGM sea-level estimate from the Argolikós (van Andel and Lianos 1984). Some spatial variability between these sites is predicted, but this is less than the observational accuracy. The results have been combined here into a single sea-level curve. Predicted values are for the mean of the sites, and the associated accuracy includes the contribution from the expected spatial variability. Much of the evidence is again of a limiting nature; upper limits are provided by terrestrial material and lower limits are provided by marine material. Agreement of the predicted sea levels with this evidence is again satisfactory, although the data points before ~ 7 ky BP are few.

A more detailed Holocene record with data extending back 10,000 years is available from the Versilia Plain of northern Tuscany (Antonioli *et al.*2000). The continental sediments generally lie above, and the marine materials lie below, predicted values (Figure 3C). Data from other Italian localities that are either tectonically stable, or where tectonic corrections can be made, are also consistent with the model back to about 11,000 BP (e.g., from Cape Palinura, Calabria, and the North Adriatic; Lambeck *et al.*2004). The fourth locality is on the French Mediterranean coast where sea-level indicators of variable quality extend back to the LGM (Lambeck and Bard 2000). For convenience, and because the observational accuracies are generally larger than the predicted spatial variability for the observation sites, the field data have here been projected onto a single sea-level curve, which is compared with the predicted values in Figure 3D. Here also, the agreement is satisfactory, and from this brief comparison, as well as from comparisons with other localities beyond the Mediterranean (Lambeck *et al.*2002), we conclude that the model predictions provide a satisfactory interpolation device for calculating sea-level change in unsurveyed areas, in this case for the Aegean–Marmara–Black Sea connection.

4. PREDICTED SEA LEVEL AT THE MARMARA AND BLACK SEA SILLS

Two sill locations are considered, one at the southern end of the Bosphorus (Figure 4A) and the other within the Dardanelles near Cape Nara Burun (Figure 4B). In addition, sea-level profiles have been predicted along a section from the Aegean to the Black Sea that runs through the Dardanelles, the Sea of Marmara, and the Bosphorus (Figure 4C). One immediate observation is that the predicted sea levels for the two sill locations are substantially different: ~5 m at 10,000 BP and >9 m after 14,000 BP. Also notable is that the Black Sea prediction lies well below that for the localities illustrated in Figure 3 and that observations from none of these sites would form a good proxy for sea level in the upper Bosphorus.

We therefore use the model predictions, calibrated against observed data from the other localities, to estimate the times at which the sills would likely have been breached. For example, if the present Bosphorus sill depth of ~ –32 m (Major *et al.*2002) is used, then the Mediterra nean water in the Marmara basin would have reached this elevation between about 10,300 and 9500 calBP,

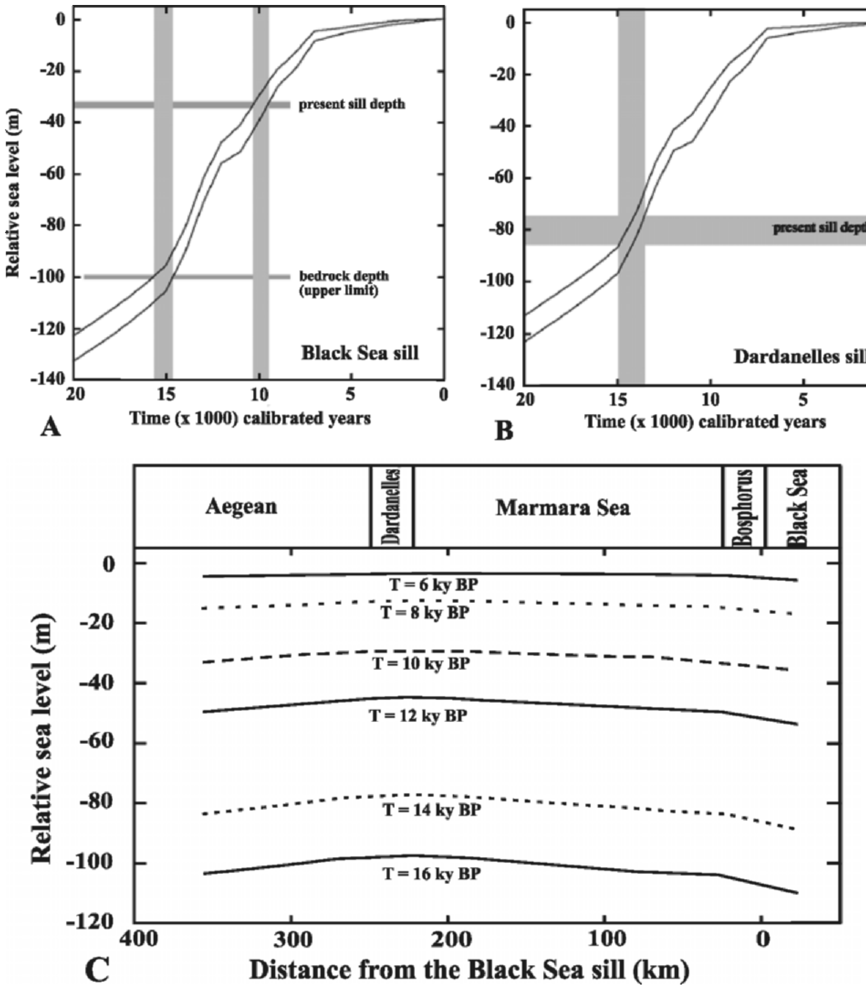

Figure 4. Predicted sea levels at two sill locations between the Aegean and Black Seas; present sill depths and predicted times of transgression are indicated. **A**. The Black Sea sill at the southern end of the Bosphorus, with two possible sill depths (its present depth at ~ –32 m and bedrock depth at ~ –100 m). **B**. The Dardanelles sill near Cape Nara Burun at a depth of ~ –80 m. **C**. Variation in predicted sea level for selected intervals along an Aegean to Black Sea transect. Variation within the Dardanelles is generally small, and the precise location of this sill is not important.

the spread reflecting the uncertainty in the model predictions. This compares with the limits of 9400 calBP and 8000 calBP proposed by Major *et al.* (2002) and Ryan *et al.* (2003), respectively. Alternatively, if the ~ –100 m bedrock depth is adopted as the top of the ancient barrier, then the first post-LGM Mediterranean inflow is predicted to have occurred between 15,700 and 14,600 calBP, with the proviso that there was no higher barrier in the Dardanelles at that time. If the Dardanelles sill was at –80±5 m (Major *et al.* 2002), then this elevation would have been breached between 15,000 and 13,700 calBP, the uncertainty reflecting both the model's limitations and the uncertainty in the reported sill depth. This compares with the dating of 15,000 to 12,800 calBP proposed by Major *et al.* (2002). Note that at these early times, circulation between the northern Aegean and the rest of the Mediterranean was much more restricted than it is today (Lambeck 1996).

4. CONCLUSION

In the absence of direct observational evidence for relative sea-level change within the Aegean-Black Sea corridor, glacio-hydro-isostatic models are used here as interpolation devices to predict changes at sill locations that are understood to have controlled the exchange of water between the Aegean, Marmara, and Black Seas. The absolute accuracy of the model predictions rests on the ability to separate the parameters that describe the Earth's response to loading and those that define the surface ice loads. When the model is used for interpolation purposes only, full decoupling of the parameters is not critical, provided that the model parameters are calibrated against regional sea-level data. For the examples used here, agreement is satisfactory and within the accuracies of both the observational evidence and the model predictions.

Based on these results, and in the absence of local tectonic movements, we estimate that, assuming that the Bosphorus sill elevation has remained unchanged at ~ –32 m, the Mediterranean reached this height for the first time since the LGM at between 10,300 and 9500 calBP, and the Dardanelles sill, at ~ –82 m, was reached between 15,000 and 13,700 calBP. Being higher than the low-level bedrock Black Sea sill, this also establishes the alternative timing for the Black Sea flooding by the Mediterranean.

Accuracy of the observational evidence is generally not high, however, and improved results are desirable from the region if these times of transgression are to be refined. Obtaining improved estimates will require additional high-accuracy information on local sea-level change in the region. In this regard, data from the Black Sea coast itself may be useful, particularly away from tectonically active zones, because of the variation predicted to occur here (Figure 1).

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34

CLIMATIC CHANGES IN THE EASTERN MEDITERRANEAN FROM THE LAST GLACIAL MAXIMUM TO THE LATE HOLOCENE

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Abstract: Time series of proxy data allow reconstruction of paleoenvironmental change in the Eastern Mediterranean region since the Last Glacial Maximum (LGM) to the late Holocene. During the LGM, climate was mostly humid, and in the subsequent Post Glacial, it became drier, except for a humid break represented by the Younger Dryas. Holocene climate in the region underwent a series of changes, some of which were severe. During most of the cold periods, the sea regressed, and precipitation rates were high, causing lake levels and ground water tables to rise. Warm periods witnessed marine transgression, climatic drying, and dropping levels of surface and subsurface water bodies, while sands, brought from the Nile, were deposited along the shoreline. The warm-dry periods brought about the collapse of many socio-economic systems, especially those along the desert fringe, while during the cold-humid periods, the desert belt flourished.

Keywords: climate change, Eastern Mediterranean, Last Glacial Maximum (LGM), Holocene.

1. EASTERN MEDITERRANEAN CLIMATE DURING THE LAST GLACIAL MAXIMUM

During most of the Last Glacial Maximum (LGM), 23 to 18 ky BP, the Near East enjoyed abundant precipitation. Lakes and marshes were widespread, and the Dead Sea reached its highest stand between about 26 and 18 ky BP (Neev and Emery 1967; Neev and Hall 1977; Begin *et al.* 1985; Bartov *et al.*

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2002). At approximately the same time, shallow lakes or marshes existed in Wadi Feiran in southern Sinai (Issar and Eckstein 1969), which date between 24 and 20 ky BP according to ^{14}C age determinations conducted by A. Kaufman of the Weizman Institute of Science.

The humid climate of the LGM recharged the Nubian sandstone aquifers underlying the Arabian-Saharan deserts, including the Sinai and Negev, and the age of this water beneath Sinai has been dated between 30 and 20 ky BP (Issar *et al.* 1972; Gat and Issar 1974). Evidence of a humid period during the LGM can also be derived from the frequency curve of radiocarbon ages obtained from groundwater in the confined aquifers of Syria, Jordan, Iraq, and northern Saudi Arabia (Geyh *et al.* 1986; Geyh 1994). Furthermore, Upper Paleolithic flint tools have been found in cultural deposits throughout the Negev and Sinai deserts, demonstrating that these arid regions were hospitable to groups of hunters and gatherers during most of the LGM (Goring-Morris 1993).

Also during this time, heavy dust storms followed by rainstorms deposited thick loess layers in the deserts as well as along the plains bordering them (Issar and Bruins 1983). Fossil water under the Sinai and Negev deserts is rich in sulfate, and the overlying loess is gypsiferous (Issar *et al.* 1987; Issar *et al.* 1988), facts that suggest water and dust from these storms contained abundant $CaSO_4$. This may be explained by the special southern and southwestern trajectories of the cyclonic lows driving the storms (Leguy *et al.* 1983).

Alternatively, Bar-Matthews *et al.* (1997) have suggested that the isotopic abundance of ^{18}O (–2.3 to –4 ‰) and ^{13}C (–7.5 to –11 ‰) from the speleothems of Soreq Cave in the northwestern Judean Hills reflects an arid climate during the period of 25 to 17 ky BP. According to these indicators, deposition within the cave should have occurred in the temperature range of 12° to 16° C, and with an annual rainfall of 300–450 mm. This conclusion, however, contrasts with other observations discussed by the same authors in the same paper, namely that during this time a substantial growth of speleothems occurred in Soreq Cave. This growth is indicated by large Low Magnesium Calcite (LMC) crystals relatively free of detritus but with the highest $\delta^{18}O$ and $\delta^{13}C$ ‰ (Bar-Matthews *et al.* 1997:162). This finding also does not correlate with the rate of growth of stalagmites in the caves of northern Israel, which was elevated despite the heavier isotopic ratio. This evidence, as well as that already discussed, prompted Geyh to conclude that, indeed, cold and humid periods occurred at 30, 25.5, 20.5, and 18 ky BP (Geyh 1994). Still, the heavy composition of environmental isotopes in the stalagmites has yet to be explained.

Geyh (1994) has suggested that although the ^{18}O and ^{13}C isotopic ratios of the stalagmites reflect climatic changes, it is still not possible to deduce from such evidence temperature values due to the influence of kinetic isotope fractionation, which affects the isotopic composition of the stalagmites. Geyh maintains (personal communication, November 1999) that a different precipitation source yielded the isotopically enriched water found in Levantine cave stalagmite formed during the LGM. He suggests that during glacial maxima, high pressure over the cold northern continent delivered the driving force for dominant storm trajectories from the still hot and humid regions of Sudan and Ethiopia. Such isotopically heavy precipitation with a deuterium excess of about +10 (Global precipitation line) occurs from time to time in the area of Palmyra today (Geyh personal communication).

The opinion of the present author is that the heavy composition of the isotopes of oxygen and carbon in the stalagmites is due to autumn, spring, and even summer rains. Precipitation during the relatively warmer seasons likely accounted for higher evaporation rates that in turn produced the heavier isotopic composition of the water which infiltrated the underground.

2. CLIMATE DURING THE TRANSITION FROM THE LGM TO THE HOLOCENE

After *ca.* 20 ky BP, heavy isotope ratios of oxygen and carbon in the Soreq Cave speleothems trend downward, gradually becoming similar to those of today. This evidence demonstrates that climate began moving toward a Mediterranean type–in the contemporary sense of the word–with warm, dry summers and cold, wet winters. There were, of course, fluctuations in temperature and, as a result, in the amount of precipitation on millennial, century, and decadal scales, as well as from year to year.

A more detailed look at the transition from the LGM to the Holocene shows, in addition to a sharp decrease in both $\delta^{18}O$ and $\delta^{13}C$, very abrupt jumps to higher values (Bar-Matthews *et al.* 1998). One of these jumps occurred at *ca.* 19 ky BP, which may be interpreted as a warm and dry spell. This evidence correlates well with the low level of the Sea of Galilee at this period. A submerged archaeological site containing the remains of a hunter-gatherer camp of the early Kebaran culture (Ohalo II) was found on the former banks of the lake. The site's discovery in 1989 was due to a dramatic decline in lake level that left it sufficiently exposed to excavate on dry land, and its presumably continuous underwater burial environment yielded remarkable preservation. Wooden building remnants were recovered as well as plentiful dietary evidence, including wild varieties of Mediterranean grains, nuts, fruits, game, and fish. Twisted fibers used for fishing nets were also found (Nadel *et al.* 1994).

From 17 to 15 ky BP, speleothems developed irregular, alternating thick and thin laminae with variable colors (from white to dark red) and higher detrital components (Bar-Matthews *et al.* 1998) implying climatic oscillations from cold and humid to warm and dry. Early Epipaleolithic sites occupied at this time are generally small and temporary, indicative of mobile foraging groups of limited membership, perhaps nuclear families (Goring-Morris 1993:1120–21). Also, in the more humid parts of the Levant, camp sites during this transitional period are in the vicinity of perennial springs or localities with a high ground water table near marshlands.

After *ca.* 15 ky BP, the isotopic composition of stalagmites within Soreq Cave assumed the general character of that of the Holocene, suggesting that the climate became Mediterranean (cold, humid winters and warm, dry summers). More frequent winter and spring storms may have marked the interval from 14.5 to 13 ky BP due to a southerly shift in the westerlies, and this change may have had an effect on the arid parts of the Levant. Indeed, Bar-Yosef reports on the appearance of many small Geometric Kebaran sites throughout the Sinai and Negev at just this time (Bar-Yosef 1993).

An abrupt rise in the percentage of heavy oxygen and carbon isotopes in the Soreq Cave stalagmites marks the interval between 13.5 and *ca.* 11.5 ky BP, corresponding to the Bölling/Allerød warming. The ratio between the two isotopes is unlike that of the Holocene but is similar to that which characterized the Post Glacial period. These two millennia may represent the final return of a glacial climatic regime, namely one with summer precipitation.

At the mound site of Jericho, in the Jordan rift valley near the perennial ‘Ein el Sultan spring, the lowermost layer contained Epipaleolithic Natufian implements dated 11,166±107 BP, 11,090±90 BP, and 10,800±180 BP (Burleigh 1981). In the same layer was found a rectangular enclosure with sockets for non-structural uprights that surrounded a clay platform bearing small saucer-like basins on its upper surface. It has been assumed that this installation served as a shrine established by the hunter-gatherer inhabitants near the spring (Kenyon 1979:24, 1993:675).

In general, the Natufian culture began the adaptation to sedentarism. This was probably due to the necessity of surviving a period of abrupt climatic fluctuations from pluvio-glacial (the Younger Dryas phase of *ca.* 10–11 ky BP) to Mediterranean climate (Holocene) and their attendant ecological effects. One can assume that hunting and fishing groups foraged over an extensive hinterland surrounding their permanent settlement as game dwindled due to over-hunting. This factor may also have increased the vegetal component of the diet, which might be surmised from the assemblage of tools for harvesting and processing grain and other seeds or nuts. The settling down probably required the invention of new food production technologies, such as storage, especially during the dry seasons and years of drought. It is difficult to imagine that, with their development of specialized tools, objects of art, and social and religious customs, Natufians failed to see that the sowing of seed during the dry season would guarantee the yield of enough grain during the humid season to feed their community. Although no concentrated effort appears to have been invested in the selection of grains from the best yielding plants, the re-seeding nevertheless ushered in agriculture. In those localities where Natufian groups adopted a sedentary way of life, selection of seeds that emphasized the most favorable properties to satisfy human needs, slowly gained dominance (see Henry 1989; Bar-Yosef and Valla 1991; Moore *et al.* 2000).

3. CLIMATE DURING THE HOLOCENE

The early Holocene in Israel, which corresponds archaeologically to the Pre-Pottery Neolithic A, or PPNA (*ca.* 10,500–9500 BP), is considered to be a warm and less humid interval (even though the Eastern Mediterranean as a whole was less warm and more humid at the time). This dry phase quickly became warm and more humid during the Pre-Pottery Neolithic B (PPNB, *ca.* 9500–7500 BP), and later, a colder and wetter period arose during the Pottery Neolithic A, or Yarmukian culture (*ca.* 7500–7000 BP). According to the isotope data (Bar-Matthews *et al.* 1998), the climatic trend during the Pre-Pottery and Pottery Neolithic periods seems to be generally humid but drying out toward the end.

Beginning *ca.* 6500 BP, the Chalcolithic period ushered in a new era of artistic tradition and technical knowledge that lasted for about thirteen centuries. Oxygen and carbon isotope ratios in cave and lake deposits show that, around 6500 BP, the climate became colder and thus more humid throughout the Fertile Crescent (Issar 2003). Agricultural settlements appeared in areas that had not been settled during earlier millennia. In the eastern part of the Fertile Crescent, people descended from the highlands into the lower Tigris and Euphrates valleys and began to build the first urban centers in this region. On the western side of the Fertile Crescent, terraces for controlling flash floods were built in the Negev Desert (Avner 1998).

The Chalcolithic cultures of the Levant vanished toward the end of the sixth millennium BP, and archaeologists explain their disappearance by positing a combination of factors such as armed conflict, political collapse, commercial competition with Egypt, or climatic changes (Levy 1998). Archaeological remains have not yet revealed any conclusive causes, but examination of the isotopic and paleo-environmental data shows that the region underwent a severe drying phase at the time that lasted for 200 to 300 years (Bar-Matthews *et al.* 1998:211).

During the ensuing Early Bronze Age, a relatively short warm and dry period was followed by a colder and wetter climate that lasted about 500 years. A variety of worldwide evidence shows that toward the end of the third millennium BC, corresponding to the end of the Early Bronze III, climate became warmer on a global scale (Bar-Matthews *et al.* 1998; Issar 2003). Again, it brought about glacial melting and a rise in the level of the world ocean. In the Near East, by contrast, lake levels dropped and deserts expanded. Mesopotamia experienced a drying trend, and its soils became increasingly saline. Jacobsen and Adams (1958) correlated the historical documents of the time (clay tablets excavated from the ruined city mounds) with the settlement patterns of ancient towns and villages within the Tigris and Euphrates River valleys. They found that from about 2300 BC, there were many reports of salinization, and that the percentage of barley in relation to wheat in the yearly harvests increased over the succeeding centuries. As barley is more salt-tolerant than wheat, their conclusion was that humans, by their failure to irrigate wisely and drain the valley properly, brought too much water to the area, thereby causing the groundwater table to rise and salts to infiltrate the upper soil layers by capillary movement. The new paleo-climatological data here discussed shifts the blame from the shoulders of the ancient Mesopotomians to global climate change.

The drying up of the semi-arid regions triggered the movement of its inhabitants towards the remaining greener areas. Most mound sites suffered abandonment, while the extensive agricultural regimes of urban society gave way to a small-scale mixed farming and herding economy in Early Bronze IV in the late third millennium BC (Dever 1998).

The warm, dry phase extended into the Middle Bronze Age, as seen in increasingly heavier isotopic ratios of ^{18}O in Galilean cave speleothems and Lake Kinneret sediments (Issar 1998). Inhabitants of the new Middle Bronze Age cities of Canaan developed subsurface water systems to cope with the scarcity of moisture and to provide a secure source to withstand sieges in their increasingly fortified strongholds. In Mesopotamia and Egypt, water projects were initiated on a regional basis.

Eventually, climate ameliorated, yielding a more humid phase by about 1500 BC, which was followed by another short but marked warm oscillation around 1300 BC. A global cold spell marked the start of the first millennium BC, when the level of the Dead Sea rose by about 50 meters (its total rise during the LGM was 200 m) (Frumkin *et al.* 1991). The desert of the Levant flourished, allowing the build-up of an extended trade network. After a short warm, dry break from 500–400 BC (Bookman *et al.* 2004), climate became humid again during the third century BC (Bar-Matthews *et al.* 1998; Issar 2003; Issar and Zohar 2004), enabling the Nabataeans to build their desert cities and develop agriculture in the surrounding valleys using a sophisticated system of flood harvesting. The Roman Empire then extended into the desert belt of the Levant.

A series of invasions by groups from the cold steppes of central and northeastern Asia due to the shortening of the warm and dry season, encroached again and again on the frontiers of empires in the west and east. The Chinese were threatened by the nomadic Xiongnu from the third century BC, while the borders of the Roman world were attacked by Germanic peoples during the third century AD and by the Huns during the fifth century AD. To protect their empire, the Romans built a line of fortifications, the *limes*, while the Chinese constructed the Great Wall. At the same time, hunger and strife due to the failure of rice crops characterized the Kofun Period of royal mound graves in Japan (Sakaguchi 1982).

At *ca.* AD 600, the climate turned warm again. Sea level began to rise and gradually inundated Roman port installations and other buildings along the former shorelines from Caesarea in Palestine to Cadiz in Spain. On the other hand, the level of the Dead Sea fell below that of the present, and its shallow southern basin dried up (Frumkin *et al.* 1991).

Also at this time, a stronger monsoon regime caused a rise in the level of the Nile, bringing increased amounts of silt and sand down to the Mediterranean Sea from the Nubian Desert (Stanley *et al.* 2001). This water-borne sediment was carried eastward by the littoral sea currents and, combined with wind-blown materials from the coast of northern Africa, it covered most of the coastal plain of Palestine. Once flourishing cities along the desert fringe became deserted, stony ghost towns. Northern and central Arabia dried out as the moisture-bearing westerlies weakened. Desertion of the *limes* settlements of the Byzantine Empire and the collapse of Sasanian Persia gave added momentum to the Arab invasion under the banner of Islam, enabling the conquest of the entire Fertile Crescent, Persia, northern Africa, and the Iberian Peninsula.

By *ca.* AD 900, climatic warming reached its climax. Europe enjoyed optimal conditions, causing an abrupt increase in population and an escalating demand for cultivatable land. Around AD 1000, a cooler phase began (Bar-Matthews *et al.* 1998; Issar 2003; Issar and Zohar 2004) as the Little Ice Age gradually took hold.

The proxy-data time series and tree ring data available from various parts of the Near East demonstrate that a phase of warming, and thus drying, of the Levant began in the first half of the 17th century AD (Issar and Zohar 2004). This turn of fate led to generalized village desertion and agricultural decline in the central and southern parts of the coastal plain of Palestine. Possibly, the interval may also have provided the opportunity for the massive sand dunes of the southern Levant to complete their encroachment onto the coastal plain.

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35

CLIMATE, SEA LEVEL, AND CULTURE IN THE EASTERN MEDITERRANEAN 20 KY TO THE PRESENT

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Abstract: Between the Last Glacial Maximum (*ca.* 18,000 years ago) and about 5,000 years ago, global sea level fluctuated between –130 m and 0 m, but in the Eastern Mediterranean, these fluctuations are not known with any precision. Similarly, the climate of the very late Pleistocene and early Holocene changed rapidly, but in the Eastern Mediterranean, these changes are hitherto not well documented nor securely dated. While the coastal sediments that accumulated during this period are well studied and precisely dated, their correlation with both sea level and climatic factors is still debated. Consequently, the possible impact of climatic events on the move from hunting and gathering to food production is also a matter of debate.

Keywords: Eastern Mediterranean, shoreline, coastal stratigraphy, Pleistocene, Holocene, agricultural revolution

1. INTRODUCTION

In this paper, the sediments of the last 20 ky on the Israeli coastal plain will be described together with their cultural contents. These sediments are composed mainly of wind-blown quartz sand, either consolidated by calcareous cementation (Issar 1968; Gavish and Friedmann 1969) or unconsolidated (Neber *et al.* 2002). Quartz sand is carried into the Mediterranean by the River Nile (Pomerancblum 1966; Rossignol 1969) and transported along the Egyptian and Israeli continental shelf by Mediterranean waves and currents (Emery and Neev 1960; Carmel *et al.* 1984). Subsequently, the sand is blown inland by the prevailing east-bound (Yaalon and Laronne 1971; Goldsmith *et al.* 1990), or sometimes north-northeast bound (Neber 2002) winds.

V. Yanko-Hombach et al. (eds.), The Black Sea Flood Question, 819-832.

The sand beds are sometimes separated by soils, which are typically a reddish sandy loam (Hamra soil, Rhodoxeralf) or a brown Regosol (Vertisol) on the coastal plain (Avnimelech 1962; Issar 1968; Dan and Yaalon 1971). The previously estimated minimum of 10 superimposed paleosols in the Israeli coastal plain (Ronen 1975) has recently been doubled (Neber 2002). Some of the red loams contain archaeological remains, and these range in age from the Lower Paleolithic, *ca.* 1 million years ago (Ron *et al.* 2001; Laukhin *et al.* 2001) to the terminal Paleolithic (Ronen *et al.* 1975; Godfrey-Smith *et al.* 2003). As presently known, then, the sand sediments on the coastal plain have accumulated over approximately the last 1 my. It should be noted that cultural remains on the coastal plain have been encountered only in the red loams. None have ever been found in the sand layers or in an incipient Regosol.

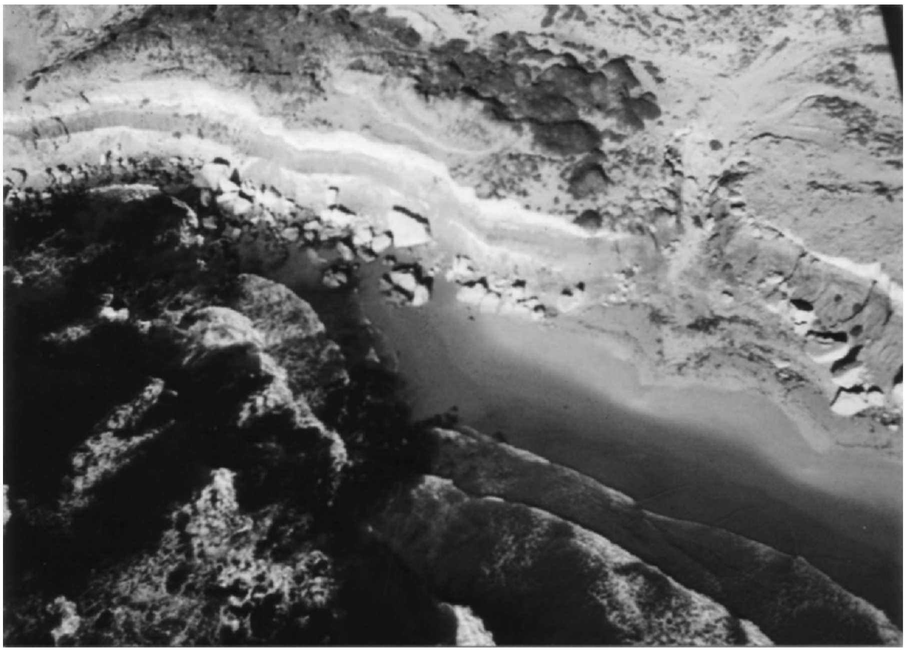

Figure 1. The eastern Mediterranean shoreline cliff near Netanya, aerial view looking east.

2. THE COASTAL SEDIMENTS

The following sediments have been identified on the Eastern Mediterranean shore (see idealized section in Figure 8). They were dated by infrared optically stimulated luminescence (IRSL) and by thermoluminescence (TL). The layers are, from the top:

2.1 Unit 1

Surficial sediments (Figures 1 and 2) consist of loose aeolian sand about 5 m thick that penetrated inland from the coast from 5.3±0.7 ky BP to the present (Neber *et al.* 2002). These are the "Hadera Sands" and underlying "Taarucha Sands" (Horowitz 1979; Gvirtzman *et al.*1998). The low-Mg calcite, aragonite, and quartz sand (Neber 2002) of Unit 1 is interstratified with four weakly developed paleosols (Vertisols) very similar to the most recent surface Regosol (Figure 3). The paleosols, which contain abundant land snails composed of aragonite, date to 5.0, 3.6, 1.6 and 0.8 ky BP, respectively (Neber 2002). Good correlation exists between the three latest soils and the three late Holocene humid phases identified by oxygen isotopic compositions ($\delta^{18}O$) and dated to 3.2, 1.3 and 0.7 ky BP, respectively (Schilman *et al.* 2002).

*Figure 2.*The eastern Mediterranean shoreline c liff near Netanya, layers 1 through 4.

2.2 Unit 2

Below Unit 1 lies a calcareous sandstone ca. 2 m thick (Figure 4). This "Tel-Aviv Kurkar" (Horowitz 1979) has recently been dated between 6.2±0.7 and 4.1±0.3 ky BP (Frechen *et al.* 2001).

Figure 3. Present day incipient soil (Layer 1) covers the "Tel Aviv Kurkar" (Layer 2) which overlies the "Netanya Hamra" (Layer 3).

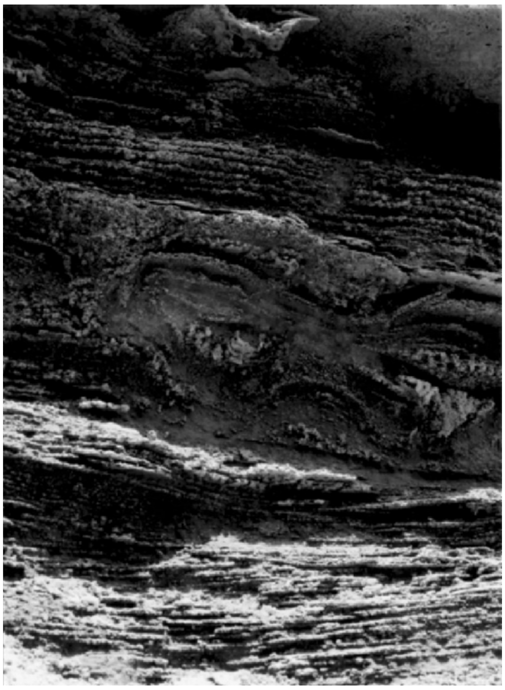

Figure 4. "Tel-Aviv Kurkar" aeolianite (Layer 2) with wind-induced undulations.

2.3 Unit 3

The Epipaleolithic red loam, or the "Netanya Hamra" (Figure 5), is 0.5–2.0 m thick (Horowitz 1979; Gvirtzman *et al.* 1984) and likely formed between 20.3±5.2 and 11.2±1.4 ky BP (Frechen *et al.*2002). The span of 57.0±5.1 to 5.3±0.3 attributed to it by Porat *et al.*(2004) probably considers two superposed soils. The Netanya Hamra normally overlies Unit 3a, the "Dor Kurkar" aeolianite (Figure 6), which is not present in the section of Figure 8.

*Figure 5.*The Epipaleolithic red loam (Layer 3, “Netanya Hamra”) near Netanya.

*Figure 6.*The incipient soil (Layer 4, “Nahsholim Sand”) *ca.* 35 ky BP covered by layer 3a, the “Dor Kurkar” aeolianite (30–20 ky BP, eroded away in Figure 8 sequence).

2.4 Unit 4

Unit 4 is an incipient soil (Vertisol) up to 2 m thick, labeled “Café au Lait” (Avnimelech 1962) and “Nahsholim Sands” (Gvirtzman *et al.* 1984). Dated between 50–30 ky BP (Frechen *et al.* 2002), it underlies the “Dor Kurkar” and overlies the “Ramat Gan Kurkar.”

2.5 Unit 5

The “Ramat Gan Kurkar” (Gvirtzman *et al.*1984) is a weakly consolidated, *ca.* 30 m thick aeolianite exposed at the base of the coastal cliff and dated 65–50 ky BP (Engelmann *et al.*2001; Frechen *et al.* 2002) (Figure 8).

3. CULTURAL EVOLUTION

Around 20,000 years ago during the Last Glacial Maximum, the Upper Paleolithic period came to an end in the Levant (Gilead 1991), and the Kebaran culture of the Epipaleolithic period began (Godfrey-Smith *et al.* 2003). Nomadism in the Kebaran was to some extent hampered by the heavy grinding equipment in the tool kit (Ronen *et al.* 2003). The necessity of leaving immovable implements in place may be seen as an initial step in the transition from hunting and gathering to sedentism and food production. This cultural transformation, known as the "Agricultural Revolution," was one of the most important changes in human history, and it was accomplished gradually, through the following phases (see Table 1).

3.2 The Kebaran Culture

The Epipaleolithic Kebaran Culture (20–13 ky BP) is characterized by microlithic tools composed of backed bladelets, points, and geometric shapes. The growing importance of vegetal foods in the Kebaran diet is attested by the manufacture of specialized equipment: grinding stones (Arensberg and Bar-Yosef 1973; Ronen *et al.* 1975), the oldest "food processor."

Grinding tools were frequently made of basalt, a hard volcanic stone, difficult to work and of limited geographical distribution (Ronen *et al.* 2003). The use of basalt 20 ky ago followed an extremely long period during which this raw material was ignored; basalt use has been attested at only two Lower Paleolithic sites in the Jordan Valley (Ronen *et al.* 2003). Hence, the renewed use of this exotic stone in food preparation may have been an elite symbol, perhaps indicating the existence at this early date of competitive feasts and the complex social order implied (Hayden 1995).

3.2 The Natufian Culture

Following the Kebaran, the Natufians (13,000–11,000 BP) produced a tool kit that was similarly characterized by microlithic tools, but now governed by lunates. They appear to have established the first sedentary settlements, though the reasons for this change are not yet fully understood (Kaufman 1992; Valla 1995). The shift radically altered living practices, and one consequence must have been that land acquired a new significance. Territorial concerns over year-round possessions are now clearly attested: stone-built circular houses of simple plan replaced the "invisible architecture" of former times. Substantially-built permanent houses mark the decline of human mobility and declare the group's attachment to a prescribed territory. A growing number of inhumations in Natufian residential sites (Valla 1995) and the domestication of dogs (Davis and Valla 1978) further affirm newly-emerged territorial claims (Ronen 2004).

Table 1. Major phases of the "Agricultural Revolution."

CULTURAL PHASE	DATE (calBP)	SOCIAL CHANGE	SUBSISTENCE	TECHNOLOGY
Pottery Neolithic	8500		Olives Fruits Wine?	Pottery
Late PPNB	9500	Temples Mega-sites	Fishing village Pastoral societies	
Early PPNB	10,500	Mother-Goddesses Cattle worshipped Treated skulls Rectangular houses	Animals domesticated	Lime plaster Projectiles
PPNA	11,500	Sacred places Numerous burials		
Natufian **Late Epipaleolithic**	13,500	Communal house Initial hierarchy Round houses Dog	Cereals domesticated Animals controlled Cereals cultivated	Sickles
S E D E N T A R I Z A T I O N				
Early Epipaleolithic	20,000	Burials rare	Hunter-gatherers	Mortars Arch
Upper Paleolithic	40,000	"Egalitarian" societies		

Natufians invented the sickle to improve plant collecting efficiency, especially with wild cereals (domesticated grains appear in the archaeological record only later). Social hierarchy is indicated (Fellner 1995) by several individuals who were buried with elaborate jewelry, a possible sign of ranking. Natufian sites contain a communal house where social events apparently took place. This house was far larger (9 m in diameter) than ordinary living structures (2–3 m in diameter), was architecturally distinct, and contained objects of symbolic nature–e.g., a cache of pestles and a dog's head (Valla 1995). The communal house and signs of social hierarchy followed so closely upon the emergence of sedentism that prior existence among late glacial nomads of the cultural practices they entailed, albeit unrecorded, may be suspected.

3.3 Neolithic Culture

A permanent settlement increases pressure on local resources, making a future food crisis inevitable (Henry 1989; Ronen and Winter 1998), climatic changes notwithstanding. It is unclear exactly when humans started to cultivate plants (Willcox 2000) and control animals (Kolska-Horwitz *et al.* 1999). At any rate, the earliest colonists in Cyprus had already brought with them managed plants and animals (Vigne 2001), roughly at the transition from the Pre-Pottery Neolithic A (PPNA, 11,000–10,500 BP) to the Pre-Pottery Neolithic B (PPNB 10,500–8500 BP). Hence, initial domestication must have been practiced in the Levant earlier still. Domestication opened the way to amass resources. An apparently more "privatized" socio-economic regime emerged in the PPNB with family-owned storage facilities and rectangular houses (Flannery 2002).

In the PPNB, inequality and social tensions within the community probably grew. At this time, the first arrowpoints appeared with perfect aerodynamic properties (Gopher 1994). Since domestication had rendered hunting a secondary strategy of food procurement, arrows may have had a social significance beyond hunting. Large communal houses exist in the PPNA (Stordeur and Abbès 2002), semi-subterranean, with a circular plan. In the PPNB, "temples" appear with outstanding features including orthostats, platforms and combustion zones (Rollefson *et al.* 1992). A mother goddess and her partner, the bull, were worshipped across the Levant (Cauvin 2000). Burials now included elaborately treated skulls taken from a small fraction of the community's deceased (Kenyon 1957).

Toward the end of PPNB, subsistence strategies diversified with the appearance of fishing villages (Galili *et al.* 2002) and pastoral societies. In the following phase, the Pottery Neolithic (8500–6500 BP), the agricultural revolution was completed with fruit domestication, leading to the manufacture of olive oil and wine (Zohary and Hopf 1988). With the emergence of metallurgy (*ca.* 6000 BP), humanity was propelled into modern technology, while the rising post-glacial sea reached its present level. The oldest port cities in the Eastern Mediterranean were established around 5000 years ago, and thereafter, history is recorded in written sources.

4. DISCUSSION

Various interpretations have been offered in an attempt to correlate observed coastal sediments (aeolianites, red loam, and incipient soils) with climatic events and oscillating sea level (Dan and Yaalon 1971; Ronen 1975, 1983; Brunnacker *et al.* 1982; Wieder and Gvirtzman 1999). Sea level in the

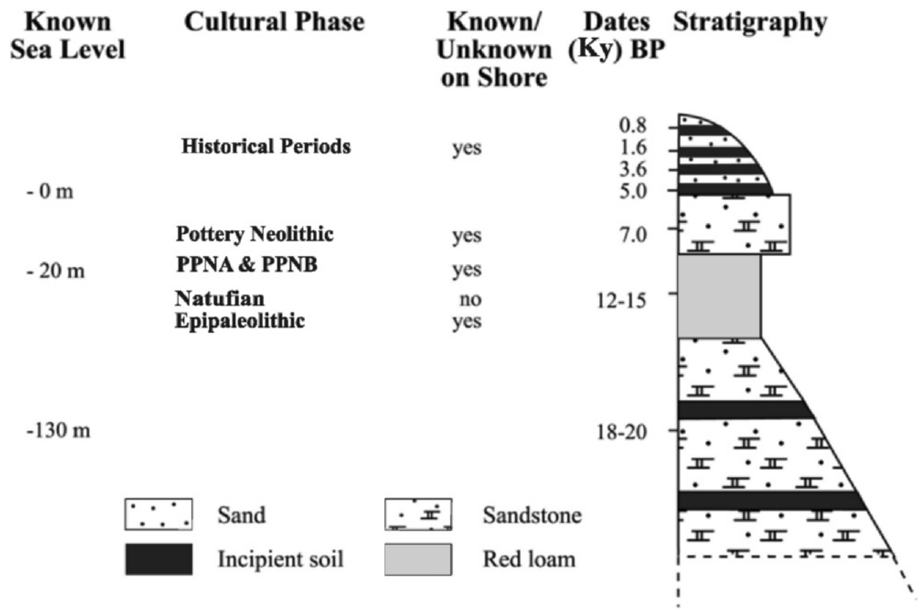

Figure 7. Schematic section in the coastal cliff approximately correlated with cultural phases.

Eastern Mediterranean is known with confidence at only three points along the timeline discussed here. It dropped to about –130 m during the Last Glacial Maximum, around 20 ky BP, it rose to about –20 m by 8 ky BP (Galili *et al.* 2002), and it attained its present elevation (0 m) at around 5 ky BP (Figure 7).

How sea level change affected humans is far from clear. The relatively rapid sea rise of 130 m substantially reduced living space on the coastal plain, probably forcing a concomitant rise in population density which must have had important social and economical consequences: "…the loss of coastal plains is arguably the most important postglacial environmental event" (van Andel and Hansen 1987:62). There is, however, no consensus on the role this event played in the Agricultural Revolution.

The three uppermost incipient soils of Unit 1 in the Sharon shoreline cliff coincide with three humid episodes identified by oxygen isotopic compositions ($\delta^{18}O$) (Schilman *et al.* 2002). The aeolianite of Unit 2 appears to be in phase with Mediterranean sapropel S1 (Neber 2002), which is dated about 7.8±4.0 ky BP (Cheddadi and Rossignol-Strick 1995) and about 8.5 ky BP (Lourens *et al.* 1996). The Younger Dryas (12–11.5 ky BP) is not recorded in the section discussed here, coming at the upper end of Unit 3, the Epipaleolithic soil or Netanya Hamra.

Severe erosion apparently occurred during the low sea stand of the Last Glacial Maximum, *ca.* 18 ky BP, as indicated by the absence of Dor Kurkar in the section described here (Figure 8) and by the hiatus of some 20,000 years

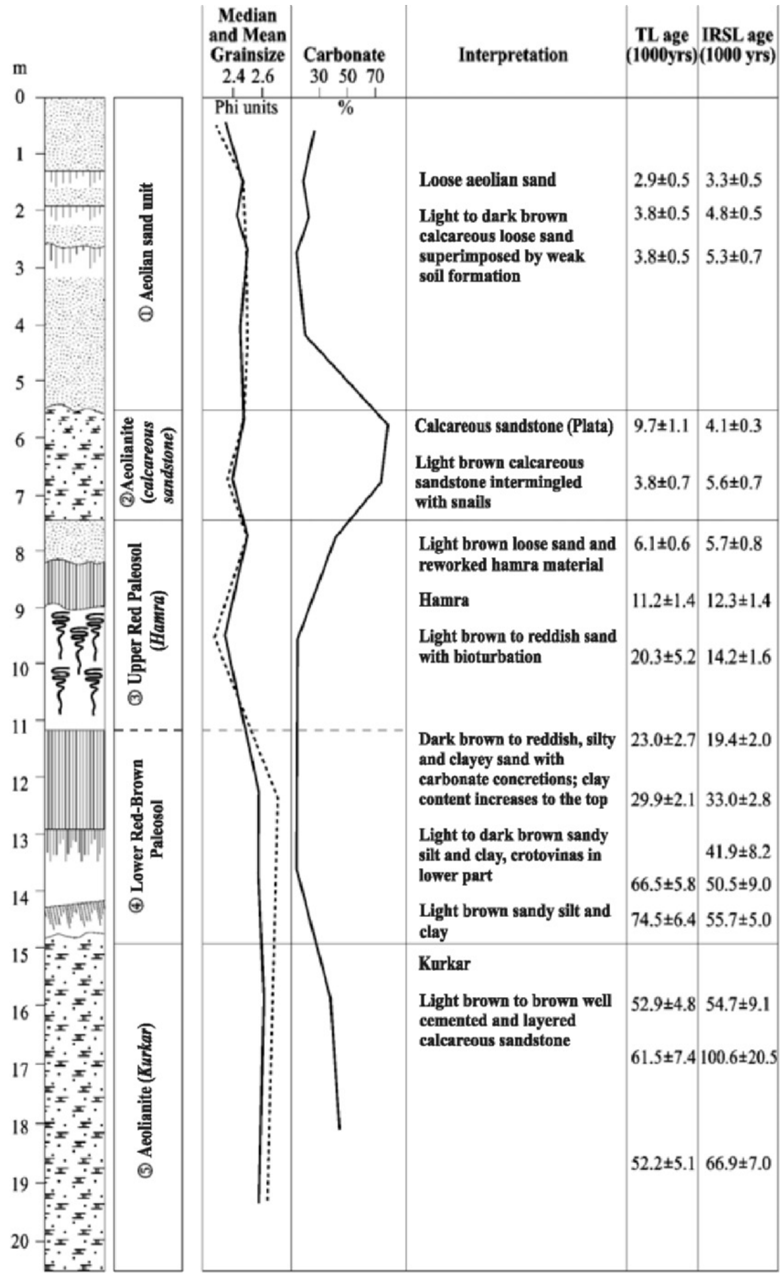

Figure 8. Idealized section in the coastal cliff near Netanya with IRSL and TL dates (from Frechen *et al.* 2001, 2002).

between the Netanya Hamra and the underlying aeolianite elsewhere in the shore cliff (Engelmann *et al.* 2001).

Wieder and Gvirtzman (1999:234) attribute sand accretion to interglacial periods and pedogenic episodes to glacial periods, but such a rigid attribution does not fit the facts presented here. Several pedogenic events, including the growth of an important red loam and a few Vertisols, occurred during the last glacial and the Holocene (Figure 8). Nor does the model fit recent data on the variation of oxygen ($\delta^{18}O$) and carbon ($\delta^{13}C$) isotopic composition during the Pleistocene (Vaks *et al.* 2003) and Holocene (Schilman *et al.* 2002).

A generalized model may be suggested (Ronen 1975). During low sea stands (= cold period), sand on the shallow continental shelf was exposed to the wind, and large scale movement inland led to sand accumulation on the coast. Pedogenesis would require, on the contrary, the cessation of sand deposition, thereby implying a high sea stand (= warm period). Rapid sand accretion prevents pedogenesis (Porat *et al.* 2004), while weak sand accretion coupled with adequate humidity may allow for a weak pedogenesis, such as the brown sandy Vertisols. That the weak Holocene Vertisols formed in humid phases was amply shown recently (Schilman *et al.* 2002). A complete halt to sand accretion and a wet Mediterranean climate are required for the formation of red loam (Rhodoxeralf) (Ronen 1975; Brunnacker *et al.* 1982; Wieder and Gvirtzman 1999). During periods of sand accretion, humans avoided the coastal plain. Only when sand accretion ceased did humans use the coastal environment.

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APPENDICES

Appendix 1: Radiocarbon dates for Late Pleistocene and Holocene sediments of the Black Sea coast and shelf from former USSR sources, arranged from youngest to oldest.
I.P. Balabanov, compiler

Appendix 2: Radiocarbon dates for Late Pleistocene and Holocene sediments of the Black Sea and Sea of Azov coasts and shelf from former USSR sources not included in Appendix 1 and from western sources, arranged from youngest to oldest.
Valentina V. Yanko-Hombach, compiler

Appendix 3: Programs of October-November, 2003, conferences.

Appendix 1. Radiocarbon dates for Late Pleistocene and Holocene sediments of the Black Sea coast and shelf from former USSR sources, arranged from youngest to oldest. Compiled by I.P. Balabanov.

Geographic Area*	Locality	Mode of Extraction	Elevation/ Water Depth (m)	Depth Within Core/ Outcrop (m)	Lithologic Unit/ Description	Laboratory Number**	Uncali-brated Age	± Error	References***
Colchis coast	Imnatskoe bog (southern coast of Paleostomi Lake)	Exploratory hole	6	2	Sphagnous peat	Mo-249	100		Vinogradova *et al.* 1963; Neishtadt *et al.* 1965
Anapa coast	Bamlyuk village, marine terrace	Exploratory hole	2	0.2	Mollusc shells with 30% removal of the outer layer	ЛГ-89А	140	110	Badinova *et al.* 1976
Anapa coast	Bamlyuk village, marine terrace	Exploratory hole	2	0.2	Mollusc shells with 60% removal of the outer layer	ЛГ-89В	240	90	Badinova *et al.* 1976
Pitsunda Peninsula	Nymphaean coastal bar	Exploratory hole 001/2	2.5	2.5	Mollusc shells dated after selection: *Chione gallina* (КИГН-437) and *Ostrea edulis* (КИГН-442)	КИГН-441	270	50	
Kuban River Delta	Dziginskaya system of coastal bars, 6 km from Kiziltash liman	Core Д-1	1	0.2	*Cardium edule*	ЛУ-1877	660	90	Izmailov *et al.* 1989
Pitsunda Peninsula	Medieval coastal bar, western part of Pitsunda	Exploratory hole 2647	3.1	0.85	Wood in buried soil	ЛУ-524а	680	90	Balabanov *et al.* 1981
Ochamchiri shelf	Coastal shelf	Core 6/88	–29.2	4.6	Mollusc shells	МГУ-1185	690	100	
Pitsunda shelf	Pitsunda Bay, 0.6 km from the mouth of Tsanigvarta River, 0.8 km from the shore	Core 9	–28	3.5	Mollusc shells	ГПИ-104	770	35	Buachidze *et al.* 1975; Balabanov *et al.* 1981
Colchis coast	Left bank of Churiya River valley, 6 km from the river mouth	Core 369		0.5	Peat	ГИН	930	50	Timofeev and Bogolyubova 1998
Northwestern shelf	Dniepro-Bugian liman	Core 481	–5.7	2.6	Mollusc shells	КИ-1570	970	40	Gozhik and Novosel'sky 1989

Geographic Area*	Locality	Mode of Extraction	Elevation/ Water Depth (m)	Depth Within Core/ Outcrop (m)	Lithologic Unit/ Description	Laboratory Number**	Uncali-brated Age	± Error	References***
Novyi Afon		Core 501	–15.2	2.8	Mollusc shells	КИГН-402	990	70	
Kerch Strait	Pesochnoe village, marine terrace (3-3.5 m)	Core 1017	3.4	1.5	Mollusc shells	ЛУ-110	1040	80	Badinova *et al.* 1976
Kuban River Delta	Markitanskaya spit	+2	2	1.5	Mollusc shells	МГУ-1549	1250	50	
Pitsunda Peninsula	Ancient boggy lagoon	Core 304	–0.7	3.9	Peat	ЛУ-1031	1350	60	Balabanov *et al.* 1981
Kobuleti coast		Core 380	2	0.2	Wood in peat	ГИН-860а	1420	80	Kind *et al.* 1975, 1976
Kuban River Delta	Northern bank of Kazachya ridge	Outcrop	2	0	Mollusc shells	КИ	1600	50	Artyukhin *et al.* 1989
Northwestern shelf	Budakskii liman	Core 221	–2.4	2	Mollusc shells		1600	70	Palatnaya 1982; Molodykh *et al.* 1984
Pitsunda Peninsula	Nymphaean coastal bar	Exploratory hole 10	2	0	Mollusc shells	ВСЕГИНГЕО II - 35	1665	86	Kuptsov *et al.* 1975
Pitsunda Peninsula	Nymphaean coastal bar	Exploratory hole 001/2	2.5	2.5	*Chione gallina* control of samples КИГН-441, КИГН-442	КИГН-437	1670	70	
Pitsunda shelf	Berth of Pitsunda resort, 0.08 km from the coast	Core 4	–15.8	12.9	*Ostrea edulis*	ГПИ-70	1690	240	Buachidze *et al.*1975; Balabanov *et al.* 1981
Pitsunda shelf	Pitsunda Bay, 0.6 km from the mouth of Tsanigvarta River, 2.6 km from shore	Core 10	–38.25	7	Mollusc shells	ГПИ-105	1700	160	Buachidze *et al.*1975
Kobuleti coast	First terrace of Natanebi River	Outcrop 13	10	1.2	Wood	ТБ-396	1740	50	Apakidze *et al.* 1987
Kuban River Delta	Kazachya ridge (~2 m), 1.4 km from Mostovyanskyi village	Outcrop	2	1	*Cardium edule* control of samples ТБ-398	ЛУ-1884	1740	50	Izmailov *et al.* 1989

Colchis coast	Ureki village, first coastal bar, to the south from the mouth of Supsa River	Exploratory hole	1.75	0	Mollusc shells		1750	240	Dzhandzhgava *et al.* 1979
Kobuleti coast	High flood-land of Natanebi River	Exploratory hole 344	13	2	Wood in alluvium	ТБ-490	1760	50	Apakidze *et al.* 1987
Kobuleti coast		Core 84	2.5	0.6	Peat	ТБ-486	1760	45	
Pitsunda Peninsula	Nymphaean coastal bar	Grubbing 1505	9	6.5	Mollusc shells	ВСЕГИНГЕО II - 29	1790	170	Kuptsov *et al.* 1975; Balabanov *et al.* 1981
Pitsunda Peninsula	Nymphaean coastal bar	Grubbing 1505	9	6.5	Mollusc shells (control of sample ВСЕГИНГЕО II-29)	ЛУ-548	1880	100	
Gagra coast	Beach cliff, river terrace of interfluve between Byzb' and Ol'ginka Rivers	Outcrop without number	3	1	Peat	ТБ-350	1930	45	Apakidze *et al.* 1987
Kuban River Delta	Kazachya ridge, 1.4 km from Mostovyanskyi village	Outcrop without number	1	0	*Cardium edule* (control of sample ЛУ-1884)	ТБ-398	1935	40	
Gagra shelf	Lower stretches of Zhvava-Kvara River	Core 607	–20	4.5	Mollusc shells	ЛУ-724	1950	110	
Pitsunda Peninsula	Nymphaean coastal bar	Grubbing 134	2.75	0.55	Mollusc shells	ВСЕГИНГЕО II - 30	1980	180	Kuptsov *et al.* 1975
Tuapse coast	Near Tuapse town	Core 14	2	11	*Chione gallina*	ЛУ-308В	2000		Tertychny 1974
Colchis coast	Imnatskoe bog (southern shore of Paleostomi Lake)	Exploratory hole	6	6	Sphagnous peat	Мо-251	2100	150	Vinogradova *et al.* 1963; Neishtadt *et al.* 1965
Gagra coast	Beach cliff, interfluve between Bzub' and Ol'ginka Rivers	Outcrop without number	3	0.9	Peat	ТБ-348	2130	45	Apakidze *et al.* 1987
Pitsunda Peninsula		Exploratory hole 001/2	2.5	2.5	*Ostrea edulis*	КИГН-442	2140	180	
Pitsunda Peninsula	Ancient boggy lagoon	Grubbing 1420a	0.3	0.8	Mollusc shells	ЛУ-542	2190	80	Balabanov *et al.* 1981

Geographic Area*	**Locality**	**Mode of Extraction**	**Elevation/ Water Depth (m)**	**Depth Within Core/ Outcrop (m)**	**Lithologic Unit/ Description**	**Laboratory Number****	**Uncali-brated Age**	**± Error**	**References*****
Gudauta shelf	Cape Souk-Su	Core 520	–6.8	12.4	Mollusc shells	КИГН-435	2220	120	
Kuban River Delta	Slobodkinsky Ridge (~1.5 km), 2.5 km from Slobodka village	Outcrop	1.5	0.75	*Cardium edule*	ЛУ-1885	2220	40	Izmailov *et al.* 1989
Kobuleti coast		Core 380	2	0.8	Wood in peat	ГИН-860в	2240	150	Kind *et al.* 1975
Gagra coast	Beach cliff, Alachadze village	Outcrop	3	1.25	Peat	ТБ-351	2300	45	Apakidze *et al.* 1987
Kuban River Delta	Dziginskay system of coastal bars, 6 km from Kiziltash liman	Core 6-99	1.5	4.3	Mollusc shells	МГУ-1528	2300	110	
Gagra shelf	Lower stretches of Zhvava-Kvara River	Core 607	–20	8	Mollusc shells	ЛУ-725	2450	150	
Gagra coast	Coastal part of Ol’ginka River valley	Exploratory hole 5039a	2.2	1.1	Wood, peat	ЛУ-649	2450	80	
Kuban River Delta	Markitanskaya spit	Core M-2	2	2.3	Mollusc shells	МГУ-1552	2450	70	
Northwestern shelf	Berezanskii liman	Core 645	–7	2.5	Mollusc shells		2450	40	Molodykh *et al.* 1984
Pitsunda Peninsula	Ancient boggy lagoon	Core 190	–0.7	4.5	Mollusc shells	ВСЕГИНГЕО II - 12	2470	380	Balabanov *et al.* 1981
Kobuleti coast	Kobuleti peat	Exploratory hole	2.5	0.9	Peat		2480	200	Dzhandzhgava *et al.* 1979
Sukhumi shelf	Western part of Sukhumi Bay	Core 718	–9.3	3.9	Mollusc shells	ТБ-384	2510	50	Apakidze *et al.* 1987
Pitsunda Peninsula	Western side of Pitsunda Peninsula	Core 231	2.8	24	*Ostrea edulis*, *Chlamys glabra*, and others	ВСЕГИНГЕО II - 120	2530	120	Balabanov *et al.* 1981
Northwestern shelf	Dniepro-Bugian liman	Core 846	–5	4.6	Mollusc shells	КИ-2181	2560	30	Gozhik and Novosel’sky 1989
Colchis coast	Gagidskoe bog	Core 57	1	1	Peat	ГИН-647	2580	50	Kind *et al.* 1975

Colchis coast	Sakorkio village	Exploratory hole			Charcoal from 4th c. AD archaeological site	ТБ-6	2600	145	Apakidze *et al.* 1987
Colchis coast	Moltavskoe bog, suburbs of Poti town	Core 44	1	2	Peat	ГИН-645	2600	500	Kind *et al.* 1975
Kerch Strait	Chushka spit	Core 259	0.4	16	Mollusc shells	ЛУ-414	2630	160	
Kobuleti coast	Kobuleti peat	Core ?	2.5	1.8	Peat		2640	300	Dzhandzhgava *et al.* 1979
Novyi Afon coast	Marine slope of the first terrace, 0.1 km from the shoreline, in vicinity of the railway station Gvandra	Core 418	7.8	6.8	Mollusc shells	КИГН-313	2690	90	
Kobuleti coast		Core 380	2	1.5	Wood in peat	ГИН-860г	2770	100	Kind *et al.* 1975
Kobuleti coast		Core 387	7.5	5.8	Wood in peat	ГИН-862а	2830	100	Kind *et al.* 1975
Pitsunda shelf	Berth of Pitsunda resort, 0.08 km from the coast	Core 4	–15.8	12.9	Mollusc shells	ВСЕГЕНГЕО - II - 33	2830	150	Kuptsov *et al.* 1975
Batumi coast	Batumski (Burun-Tabiya) Cape, stadium in Batumi	Outcrop	2	2	Mollusc shells in pebbles		2840	180	Dzhandzhgava *et al.* 1979
Gagra coast	Beach, 0.7-0.8 km north of Colchis channel mouth, near Gnilushka River	Exploratory hole 008	1.8	0.4	Wood in peat	КИГН-586	2880	40	
Kuban River Delta	Dziginskaya system of coastal bars, 6 kilometers from Kiziltash liman	Core 6-99	1.5	6.3	Mollusc shells	МГУ-1529	2890	150	
Northwestern shelf	Dniepro-Bugian liman	Core 621	–5.8	2.75	Mollusc shells	КИ-1574	2900	40	Gozhik and Novosel'sky 1989
Pitsunda shelf	Pitsunda Bay, in front of the mouth of Tsanigvarta River, 2.08 km from shore	Core 3	–24.8	6.05	Wood	ВСЕГЕНГЕО - II - 9	2920	340	Kuptsov *et al.* 1975
Northwestern shelf	Tiligulskii liman	Core 67		2	Mollusc shells		3000	90	Molodykh *et al.* 1984
Northwestern shelf	Shagany liman	Core 170	–2.24	1.3	Mollusc shells		3020	110	Molodykh *et al.* 1984

Geographic Area*	Locality	Mode of Extraction	Elevation/ Water Depth (m)	Depth Within Core/ Outcrop (m)	Lithologic Unit/ Description	Laboratory Number**	Uncali-brated Age	± Error	References***
Colchis coast	Mouth of Inguri River	Exploratory hole	1.25	2.2	Alder(-tree) peat	ТА-1303	3090	100	Serebryanny *et al.* 1984
Pitsunda Peninsula	Ancient boggy lagoon	Core 473	0	14.8	*Divaricella divaricata*, *Pitar rudis*, *Spisula subtruncata*, *Ostrea edulis*, *Chione gallina*, *Bittium reticulatum*, *Cardium edule*	ЛУ-1033	3130	170	Balabanov *et al.* 1981
Colchis coast	Grigoleti village, second coastal bar, mouth of Supsa River	Exploratory hole	5	3	Mollusc shells		3140	280	Dzhandzhgava *et al.* 1979
Colchis coast	Chaladidi village, interfluve between Rioni River and Paleostomi Lake	Exploratory hole	6	6	Mollusc shells	ТБ-56	3150	90	Burchuladze *et al.* 1975
Northwestern shelf	Dniepro-Bugian liman	Core 332	–4.4	3	Mollusc shells		3160	80	Molodykh *et al.* 1984
Kobuleti shelf	Coastal shelf	Core m20	–20	2	Mollusc shells	ТБ-480	3170	50	Apakidze *et al.* 1987
Northwestern coast	Alibey liman spit	Core 178	1.6	5.2	Mollusc shells		3200	100	Molodykh *et al.* 1984
Northwestern shelf	Dniepro-Bugian liman	Outcrop without number	–1.8	0.3	Mollusc shells under cultural layer at archaeological site of Olbia	ЛЕ-1171	3210	50	Shilik 1972
Pitsunda Peninsula	Novochernomorskaya system of coastal bars	Exploratory hole 003	1.5	2.5	*Chione gallina*	КИГН-440	3250	160	
Gudauta shelf	Cape Souk-Su	Core 522	–16.5	1.5	Mollusc shells from clay sand	КИГН-577	3300	80	
Sukhumi shelf	2 km NNW from the mouth of Gumista River	Core 723	–9.8	4.5	Mollusc shells	ТБ-361	3340	50	Apakidze *et al.* 1987
Sukhumi shelf	Western part of Sukhumi Bay	Core 702	–15.4	0.7	Mollusc shells	ТБ-382	3360	50	Apakidze *et al.* 1987

Gudauta shelf	Near Cape Souk-Su	Core 520	–6.8	13.1	Mollusc shells	КИГН-436	3370	90	
Gudauta shelf	Left bank of Khipsta River	Core 535	–8.7	2.35	Mollusc shells in fine sand	КИГН-578	3380	80	
Gagra coast	North part of coastal plain	Core 459	2.7	3.25	Peat	ЛУ-651	3400	80	
Colchis coast	Moltavskoe bog, suburbs of Poti town	Core 44	1	2.75	Peat	ГИН-646	3400	200	Kind *et al.* 1975
Kobuleti coast	?	Core 84	2.5	1.4	Peat	ТБ-487	3420	90	
Colchis coast	Chaladidi village, "Zugra" hill, suburbs of Poti town	Exploratory hole			Charcoal; Late Bronze archaeological site XIV-X c. BC	ТБ-5	3470	190	Apakidze *et al.* 1987
Northwestern coast	Dniepro-Bugian liman, terrace +2 m	Outcrop without number	2	2.4	Mollusc shells	Mo-500	3480	60	Shilik 1972; Devirts *et al.* 1972
Northwestern coast	Odessa Bay port elevator	Core 31	1.2	12.1	Mollusc shells	ЛУ-1102	3500	170	
Pitsunda Peninsula	Beach, eastern side of Pitsunda Cape	Core 230	1	40	Mollusc shells	ВСЕГИНГЕО II - 141	3520	130	Kuptsov *et al.* 1975
Pitsunda Peninsula	Coastal bar	Grubbing 60	2.5	1.5	Mollusc shells	ВСЕГИНГЕО II - 28	3520	320	Kuptsov *et al.* 1975
Pitsunda Peninsula	SE of Lake Zmeinoe	Exploratory hole 006	0	0.94	Peat	КИГН-582	3550	50	
Pitsunda Peninsula	SE of Lake Zmeinoe	Exploratory hole 006	0	0.94	Peat	КИГН-583	3600	50	
Pitsunda Peninsula	Coastal bar	Exploratory hole 002	1.5	2.5	Mollusc shells	КИГН-438	3620	140	
Gagra coast	Beach, 0.7-0.8 km north of Colchis channel mouth, near Gnilushka River	Exploratory hole 008	1.8	0.6	Peat	КИГН-587	3650	190	
Gagra coast	1.8-1.9 m above sea level south of Colchis channel mouth (Ol'ginka River)	Outcrop	1.9	0	Peat	МГУ-296	3690	120	Parunin *et al.* 1974; Kvavadze 1974
Kuban River Delta	Dziginskaya system of coastal bars, 6 km from Kiziltash liman	Core 6-99	1.5	8.7	Mollusc shells	МГУ-1526	3770	60	

Geographic Area*	Locality	Mode of Extraction	Elevation/ Water Depth (m)	Depth Within Core/ Outcrop (m)	Lithologic Unit/ Description	Laboratory Number**	Uncali-brated Age	± Error	References***
Novyi Afon shelf		Core 1a/88	–30.4	2.45	Mollusc shells	МГУ-1190	3820	280	
Pitsunda Peninsula	Neochernomorian terrace, NW of Lake Inkit	Exploratory hole 007	0	0.36	Peat (control of sample КИГН-584)	КИГН-585	3850	60	
Sukhumi shelf	2 km NNW of the Gumista River mouth	Core 724	–14	5	Mollusc shells	ТБ-388	3850	50	Apakidze *et al.* 1987
Northwestern coast	Zmeiny Island	Core 862		2.4	Peat	КИ-690	3860	90	Kovalyukh *et al.* 1977
Kobuleti coast		Core 12		3	Mollusc shells	ТБ-393	3890	50	Apakidze *et al.* 1987
Novyi Afon shelf		Core 507	–14.9	2.9	Mollusc shells	КИГН-404	3900	70	
Northwestern coast	Marine terrace 1 km from the mouth of Berezanskii liman	Outcrop without number	1	0.9	Mollusc shells		3900	100	Molodykh *et al.* 1984
Northwestern shelf	Budakskii liman	Core 221	–2.4	4	Mollusc shells		3900	100	Palatnaya 1982
Colchis coast	Left bank of Churiya River, 6 km from its mouth	Core 369		3.75	Peat	ГИН	3930	200	Timofeev and Bogolyubova 1998
Gagra coast	Lower stretches of Zhvava-Kvara River	Core 609	–17.8	22	Mollusc shells	ЛУ-726	4000	140	
Novyi Afon coast	Psyrcha village, marine terrace 2.5 m above sea level	Exploratory hole	2.5	0.5	Mollusc shells overlying sand layer containing ceramics of the Kobanian culture (~4 ky BC)	ЛГ-95	4000	180	Badinova *et al.* 1976
Gagra shelf	Lower stretches of Zhvava-Kvara River	Core 607	–20	25	Mollusc shells	ЛУ-727	4020	230	
Gudauta shelf	Left bank of Khipsta River	Core 536	–20.5	7.3	Mollusc shells in fine sand	КИГН-576	4020	80	
Northwestern shelf	Khadzhibeyskii liman	Core 342		4.6			4020	50	Molodykh *et al.* 1984

Sukhumi shelf	Interfluve betw. Kelasuri and Madzharka Rivers	Core 717	–22.5	3.1	Mollusc shells	ТБ-353	4040	50	Apakidze *et al.* 1987
Colchis coast	Anakliyski peat, south of the Inguri River mouth		1.25	3.5	Peat		4050	50	Burchuladze *et al.* 1975
Colchis coast	Churiyski peat			3.5	Peat		4050	60	Burchuladze *et al.* 1975
Pitsunda Peninsula	Neochernomorian terrace, NW of Lake Inkit	Exploratory hole 007	0	0.36	Peat; control of sample КИГН-585	КИГН-584	4060	100	
Kobuleti coast	?	Core 385	3.5	3	Wood in peat	ГИН-861в	4070	120	Kind *et al.* 1975
Colchis coast	Anakliyski peat in the Inguri River mouth		1.25	3.2	Sedge-reed peat	ТА-1301	4090	90	Serebryanny *et al.* 1984
Kobuleti coast	?	Core 385	3.5	2.25	Wood in peat	ГИН-861б	4100	300	Kind *et al.* 1975
Colchis coast	Poti			3.5	Peat	ГИН-103	4100	50	Zubakov 1974
Northwestern coast	Zmeiny Island	Core 862		2.6	Peat	КИ-691	4120	60	Kovalyukh *et al.* 1977
Colchis coast	Imnatskoe bog (southern shore of Paleostomi Lake)		6	8.5	Peat	Мо-253	4130	190	Vinogradova *et al.* 1963; Neishtadt *et al.* 1965
Novyi Afon shelf	Coastal shelf	Core 517	–6.4	3.3	Mollusc shells	КИГН-414	4130	90	
Northwestern coast	Odessa Bay, port elevator	Core 8	1.45	16.6	Mollusc shells	ЛУ-1099	4130	150	
Gagra shelf	Lower Zhvava-Kvara R.	Core 607	–20	20	Mollusc shells	ЛУ-721	4140	160	
Colchis coast	Nabadskoe bog, right bank of Rioni River	Core 43	1	3.5	Sedge-reed peat	ГИН-108	4140	50	Cherdyntsev *et al.* 1966; Sluka 1969
Colchis coast	Gagidskoe bog		1	2.5	Peat		4150	40	Sluka 1969
Gagra coast	Interfluve between Bzyb' and Ol'ginka Rivers	Core 316	2	3.9	Peat	ЛУ-507	4170	90	Balabanov *et al.* 1981
Colchis coast	20 km from Khobi River mouth, Horcha village (Khobski county); 1.3 m thick Pogrebennyi peat	Outcrop	4.5	0.3	Peat	ТБ-70	4170	50	Burchuladze *et al.* 1975

Geographic Area*	Locality	Mode of Extraction	Elevation/ Water Depth (m)	Depth Within Core/ Outcrop (m)	Lithologic Unit/ Description	Laboratory Number**	Uncali-brated Age	± Error	References***
Gagra coast	Beach, 0.7-0.8 km north of Colchis channel mouth, near Gnilushka River	Exploratory hole 008	1.8	1.1	Peat	КИГН-588	4200	60	
Gagra coast	Northern part of the coastal plain	Core 459	2.7	4.25	Peat	ЛУ-650	4220	70	
Gagra coast	Underwater slope south of Colchis channel (Ol'ginka River), 1.8-1.9 m above sea level	Outcrop without number	1.9	0	Peat	ТБ-43	4280	60	Burchuladze *et al.* 1975
Kobuleti shelf	?	Core M24	–7.1	6.3	Mollusc shells	ТБ-481	4290	50	Apakidze *et al.* 1987
Gagra shelf	Interfluve between Ol'ginka and Tskherva Rivers	Core 617	–15.5	8.8	Mollusc shells	ЛУ-1065	4300	80	Dzhandzhgava *et al.* 1982
Kuban River Delta	Dziginskay system of coastal bars, 6 km from Kiziltash liman	Core 6-99	1.5	9.7	Mollusc shells	МГУ-1527	4330	90	
Sukhumi shelf	Interfluve between Kelasuri and Madzharka Rivers	Core 716	–10.4	1.4	Mollusc shells	ТБ-357	4370	60	Apakidze *et al.* 1987
Colchis coast	Right bank of Supsa River, 3.5 km from mouth	Core 212	2	2	Peat	ГИН	4390	80	Timofeev and Bogolyubova 1998
Lazarevskoye coast	Left bank of Psezuapse River (close to the mouth)	Core 441	2.3	12	Mollusc shells	ЛУ-189	4440	100	Arslanov *et al.*1975
Pitsunda Peninsula	Ancient boggy lagoon	Core 110	0	10.3	*Cardium edule*, *Ostrea edulis*, *Mytilus galloprovincialis*, *Paphia discrepans*	ВСЕГИНГЕО II - 24	4450	310	Balabanov *et al.* 1981; Kuptsov *et al.* 1975
Gagra coast	Coastal plain, mouth of Ol'ginka River	Exploratory hole without number	2	1.8	Peat	ТБ-42	4460	150	Burchuladze *et al.* 1975

Pitsunda Peninsula	Cape Inkit	Core 202	2.9	18	*Mytilus galloprovincialis*, *Divaricella divaricata*, *Spisula subtruncata*, *Chione gallina*, *Corbula mediterranea*, *Cardium edule*	ЛУ-541	4480	60	Balabanov *et al.* 1981
Kuban River Delta	Zesterovataya ridge, 3 km from Prigibsky farmstead	Outcrop	0.8	0.15	*Cardium edule*	ЛУ-1880	4500	60	Izmailov *et al.* 1989
Ochamchira shelf	Open sea	Core 6/89	–29.2	10	Mollusc shells	МГУ-1186	4500	120	
Northwestern coast	Marine terrace 1 km from the mouth of Berezanskii liman	Outcrop without number	1.75	1.45	*Cardium edule*, *Mytilus galloprovincialis*, *Hydrobia ventrosa*		4500	120	Molodykh *et al.* 1984
Pitsunda shelf	Pitsunda Bay in front of the mouth of Tsanigvarta River, 0.8 km from the shore	Core 1	–32.5	7.85	Mollusc shells	ГПИ-103-86	4510	200	Buachidze *et al.* 1975
Colchis coast	Anakliyski peat, south of the Inguri River mouth		1.25	3.85	Sedge-reed peat	ТА-1300	4530	70	Serebryanny *et al.*1984
Lazarevskoye coast	Right bank of Psezuapse River, close to the mouth	Core 401	1.77	13	Mollusc shells	ЛУ-199	4560	160	Arslanov *et al.* 1975
Colchis coast	Anakliyski peat, south of the Inguri River mouth		1.25	5.35	Peat in boggy clays	ТА-1299	4570	90	Serebryanny *et al.*1984
Pitsunda shelf	Western part of Pitsunda Bay, berth of Pitsunda resort, 0.25 km from shore	Core 6	–40.3	26.6	Mollusc shells	ГПИ-97	4580	170	Buachidze *et al.* 1975
Northwestern coast	Odessa Bay, port elevator	Core 31	1.2	16.1	Mollusc shells	ЛУ-1100	4600	680	
Sukhumi coast	Distal part of the Sebastopolisskyi Cape	Core 100	2.5	8.5	Peat	ТБ-371	4670	60	Apakidze *et al.* 1987
Novyi Afon shelf	?	Core 505	–8.4	3.3	Mollusc shells	КИГН-408	4680	80	
Kodora shelf	Kodorski Cape	Core 4/88	–17.3	13.7	Mollusc shells	МГУ-1187	4700	150	

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Kobuleti coast	?	Core 88	2.1	5	Mollusc shells	ТБ-484	4770	60	Apakidze *et al.* 1987
Gagra shelf	Underwater slope south of Colchis channel (Ol'ginka River)	Outcrop	–3.5	7.0	Peat	МГУ-	4800	90	Kvavadze 1974
Kobuleti coast	?	Core 385	3.5	3.45	Wood in peat	ГИН-861г	4800	100	Kind *et al.* 1975
Northwestern shelf	Dniepro-Bugian liman	Core 481	–5.7	10.9	Mollusc shells	КИ-1571	4800	100	
Northwestern shelf	Dniepro-Bugian liman	Core 493	–6.6	5	Mollusc shells		4800	100	Gozhik and Novosel'sky 1989
Novyi Afon shelf	?	Core 514	–8.9	4	Mollusc shells	КИГН-416	4890	95	
Adler coast	Interfluve betw. Mzymta and Psou Rivers, 3.4 km south of Mzymta River	Core A-11	3.8	17	Mollusc shells	ЛУ-702	4900	60	
Pitsunda Peninsula	In vicinity of poultry farm	Exploratory hole 005	0	0.3	Peat	КИГН-581	4960	270	
Kuban River Delta	Sadkovskaya system of coastal bars (1 m length), 2 km SE of Sadki village	Outcrop without number	1.0	0.6	*Cardium edule*, control of sample ЛУ-1879	ТБ-397	5025	50	
Northwestern shelf	Khadzhibeyskii liman	Core 342		6.8			5100	70	Molodykh *et al.* 1984
Kobuleti coast	?	Core 88	2.1	6.5	Mollusc shells	ТБ-485	5120	60	Apakidze *et al.* 1987
Lazarevskoye coast	Right bank of Psezuapse River (close to the mouth)	Core Л-9	4	13	Mollusc shells	ЛУ-703	5140	90	
Sukhumi coast	Central part of Sukhumi Cape	Core 42	1	16	Mollusc shells	ТБ-341	5180	60	Apakidze *et al.* 1987
Kobuleti coast	?	Core 71	4.8	8.3	Wood in lake-alluvial sediments	ТБ-488	5200	55	

Novyi Afon coast	Coastal part of the Aapsta River valley	Core 55	1.62	15.5	Wood in marine pebbles	КИГН-207	5200	80	
Northwestern shelf	Dniepro-Bugian liman	Core 846	–5	8.9	Mollusc shells	КИ 2182	5200	200	Gozhik and Novosel'sky 1989
Northwestern shelf	Dniepro-Bugian liman	Core 846	–5	11.8	Mollusc shells	КИ 2183	5200	90	Gozhik and Novosel'sky 1989
Northwestern shelf	Budakskii liman	Core 221	–2.4	5	Mollusc shells		5200	60	Palatnaya 1982; Molodykh *et al.* 1984
Kuban River Delta	Sadkovskaya system of coastal bars (1 m length), 2 km SE of Sadki village	Outcrop without number	1.0	0.6	*Cardium edule*, control of sample ТВ-397	ЛУ-1879	5210	50	Izmailov *et al.* 1989
Kuban River Delta	Derevyankovskaya system of coastal bars, 2 km south of Derevyankovskaya village	Outcrop without number	1.0	0.5	*Cardium edule*	ЛУ-1883	5210	60	Izmailov *et al.* 1989
Gagra coast	Coastal plain between Bzyb' and Ol'ginka Rivers	Core 16	2	7.2	Wood, peat	ЛУ-508	5240	90	Balabanov *et al.* 1981
Colchis coast	Left bank of Churiya River, 6 km from mouth	Core 369		5	Peat	ГИН	5240	60	Timofeev and Bogolyubova 1998
Kobuleti shelf	?	Core m14	–6	5.7	Mollusc shells	ТБ-483	5270	60	Apakidze *et al.* 1987
Pitsunda shelf	Western part of Pitsunda Bay (berth of Pitsunda resort), 0.25 km off shore	Core 6	–40.3	28.6	*Mytilus galloprovincialis*	ГПИ-78	5270	180	Burchuladze *et al.* 1975; Balabanov *et al.* 1981
Kuban River Delta	Derevyankovskaya system of coastal bars west of Derevyankovskaya village	Outcrop without number	1.5	1.0	*Cardium edule*, control of sample ЛУ-1881	ТБ-399	5310	55	
Pitsunda shelf	Erosion bench in the upper reach of "Akula" canyon	Outcrop without number	–35	0	Wood	МГУ-	5320	170	Balabanov *et al.* 1981
Novyi Afon shelf	?	Core 1/88	–53	2.5	Mollusc shells	МГУ-1184	5340	120	
Kobuleti coast	?	Core 7	5.8	11.2	Mollusc shells	ТБ-392	5350	55	

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Northwestern shelf	Tiligulskii liman	Core 9	–9	4.3	Mollusc shells		5360	170	Molodykh *et al.* 1984
Colchis coast	5-6 km from the mouth of Supsa River, Supsa village		9	10.5	Peat	МГУ; ТЛ-61	5400	700	Kvavadze 1978
Gagra shelf	Lower stretches of Zhvava-Kvara River	Core 607	–20	34.6	Mollusc shells	ЛУ-720	5410	320	
Colchis coast	Churiiski peat, Khobi River mouth	?	1	5	Peat		5500	80	Dzhanelidze 1980
Northwestern shelf	Dniepro-Bugian liman	Core 493	–6.6	9.9	Mollusc shells	КИ-1572	5500	80	Gozhik and Novosel'sky 1989
Colchis coast	Mouth of Inguri River		1.25	6.35	Peat	ТА-1298	5540	100	Serebryanny *et al.* 1984
Sukhumi shelf	2 km NNW from the mouth of Gumista River	Core 723	–9.8	10.5	*Spisula subtruncata*, *Turbonilla delicata*, *Chione gallina*, *Bittium reticulatum*, *Retusa umbilicata*, *Mytilaster lineatus*, *Cardium edule*, *Hydrobia ventrosa*	ТБ-362	5540	60	Apakidze *et al.* 1987
Tuapse coast	Lazarevskoe village, marine terrace	Core 1209	1.4	18	Mollusc shells	ЛУ-195	5550	380	Arslanov *et al.* 1975
Gagra shelf	Interfluve betw. Ol'ginka and Tskherva Rivers	Core 617	–15.5	24	Mollusc shells on terrestrial loams	ЛУ-1064	5560	70	Dzhandzhgava *et al.*1982
Sukhumi coast	Rear part of peninsula, 0.8 km west of Adzapsh River	Core 125	3	9.6	Peat	ТБ-391	5570	50	Apakidze *et al.* 1987
Colchis coast	20 km from the Khobi River mouth, Horcha village (Khobski county); 1.2 m buried peat	Outcrop without number	4.5	1	Peat	ТБ-60	5600	50	Burchuladze *et al.* 1975
Northwestern coast	Alibey liman	Core 178	1.6	8	Mollusc shells		5600	170	Molodykh *et al.* 1984

Northwestern shelf	Dniepro-Bugian liman	Core 621	–5.8	10.75	Mollusc shells	КИ-1575	5600	70	Gozhik and Novosel'sky 1989
Pitsunda Peninsula	SW coastal part of the peninsula	Core 114	3.6	50	Mollusc shells	ВСЕГИНГЕО II - 8	5650	95	Kuptsov *et al.* 1975
Kuban River Delta	Sadkovskaya system of coastal bars (1 m length), 10.4 km S of Sadki village	Outcrop	1.0	0.25	*Cardium edule*	ЛУ-1878	5660	60	Izmailov *et al.* 1989
Northwestern shelf	Dniepro-Bugian liman	Core 495	–5	16.1	Mollusc shells	КИ-1573	5700	90	Gozhik and Novosel'sky 1989
Northwestern shelf	Dniepro-Bugian liman	Core 493	–6.6	13	Mollusc shells		5700	90	Gozhik and Novosel'sky 1989
Kobuleti coast	?	Outcrop 10	12	5.9	Wood	ТБ-394	5710	60	
Adler coast	Interfluve betw. Mzymta and Psou Rivers, 3.4 km S of the Mzymta River	Core 1204y	2	12	Wood in marine sediments	ЛУ-187	5720	120	Arslanov *et al.* 1975
Sukhumi shelf	2 km NNW from the mouth of Gumista River	Core 724	–14	10	*Spisula subtruncata*, *Rissoa splendida*, *Chione gallina*, *Cardium exiguum*, *Bittium reticulatum*, *Nassarius reticulatus, Retusa truncatula, Retusa umbilicata, Pachia discrepans, Cardium edule*, *Hydrobia ventrosa*, rarely: *Ostrea edulis*, *Modiolus adriaticus*, *Gastrana fragilis*	ТБ-389	5720	60	Apakidze *et al.* 1987
Novyi Afon coast	Right bank of Tsevekvara River valley	Core 414	3	8	Mollusc shells	КИГН-122	5730	120	
Adler coast	Marine terrace on the interfluve between the Kudepsta and Mzymta Rivers	Core A-18	5.87	20	Mollusc shells	ЛУ-753	5740	80	

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Sukhumi shelf	Near the Gumista River mouth	Core 725	–7.1	11.3	Mollusc shells	ТБ-390	5760	60	Apakidze *et al.* 1987
Gagra coast	Northern part of the coastal plain	Core 407	3	30	Mollusc shells	ЛУ-730	5760	160	
Colchis coast	Imnatskoe bog (southern shore of Paleostomi Lake)		6	11.3	Sphagnous peat	Мо-254	5820	210	Vinogradova *et al.* 1963; Neishtadt *et al.* 1965
Novyi Afon shelf	?	Core 1/88	–53	3.7	Mollusc shells	МГУ-1191	5830	840	
Pitsunda shelf	Pitsunda Bay, in front of the Tsanigvarta River mouth, 0.8 km from the shore	Core 1	–32.5	14.15	Mollusc shells	ВСЕГИНГЕО - II - 43	5840	410	Kuptsov *et al.* 1975
Adler coast	Marine terrace on the interfluve betw. Kudepsta and Mzymta Rivers	Core A-17	5.33	15	Mollusc shells	ЛУ-754	5860	70	
Kuban River Delta	Derevyankovskaya system of coastal bars, W of Derevyankovskaya station	Outcrop without number	1.5	1.0	*Cardium edule*, control of sample ТБ-399	ЛУ-1881	5890	50	Izmailov *et al.* 1989
Kobuleti coast	?	Outcrop 10	12	7.8	Wood	ТБ-395	5910	60	Apakidze *et al.* 1987
Kobuleti coast	?	Core 70	2.7	13.1	Mollusc shells	ТБ-482	5950	55	
Kobuleti coast	?	Core 387	7.5	9	Wood in peat	ГИН-862в	6000	500	Kind *et al.* 1975
Sukhumi shelf	Interfluve between Kelasuri and Madzharka Rivers	Core 717	–22.5	6.9	Mollusc shells	ТБ-355	6050	60	Apakidze *et al.* 1987
Sukhumi shelf	Interfluve between Kelasuri and Madzharka Rivers	Core 716	–10.4	3	Mollusc shells	ТБ-358	6060	60	Apakidze *et al.* 1987
Gagra shelf	Lower stretches of Ol'ginka River	Core 604	–14	7	Mollusc shells	ЛУ-677	6060	180	Dzhandzhgava *et al.* 1982

Sukhumi shelf	Interfluve between Kelasuri and Madzharka Rivers	Core 717	–22.5	6	Mollusc shells	ТБ-354	6060	60	Apakidze *et al.* 1987
Pitsunda Peninsula	SW coastal part of the peninsula	Core 214	5.2	30.0	Mollusc shells	ВСЕГИНГЕО II - 41	6070	130	Kuptsov *et al.* 1975
Northwestern shelf	Budakskii liman	Core 221	–2.4	9	Mollusc shells		6100	80	Palatnaya 1982
Gagra shelf	Lower stretches of Ol'ginka River	Core 602	–7.1	12.2	Mollusc shells	ЛУ-669	6160	160	Dzhandzhgava *et al.* 1982
Adler coast	Interfluve between Mzymta and Psou Rivers, rear part of the terrace	Core A-9	1.62	7	Wood in peat	ЛУ-700	6180	70	
Adler shelf	Canyon on Cape Konstantinovski	Outcrop without number	–16	0	Mollusc shells	ТБ-347	6210	60	
Pitsunda shelf	Within Pitsunda bayhead, opposite Tsanigvarta River estuary, 0.05 km from shore		–42	17.2	*Mytilus galloprovincialis*		6240	230	Burchuladze *et al.* 1975; Balabanov *et al.* 1981
Kobuleti coast		Core 387	7.5	6.25	Wood in peat	ГИН-862б	6300	2000	Kind *et al.* 1975
Northwestern shelf	Dniepro-Bugian liman	Core 855	–8.5	6.9	Peat	КИ-2241	6300	300	Gozhik and Novosel'sky 1989
Gudauta coast		Mine 1001/1	2	18	Wood	КИГН-230	6430	200	
Sukhumi coast	Rear part of the peninsula, 0.4 km W of Adzapsh River	Core 36	5.2	8	Peat	ТБ-352	6430	60	Apakidze *et al.* 1987
Novyi Afon shelf	Coastal shelf	Core 502	–8.5	4.5	Mollusc shells	КИГН-407	6460	190	
Sukhumi shelf	Interfluve between Kelasuri and Madzharka Rivers	Core 716	–10.4	5	Mollusc shells	ТБ-359	6480	60	Apakidze *et al.* 1987
Novyi Afon shelf	Coastal shelf	Core 503	–8.7	5	Mollusc shells	КИГН-400	6480	180	

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Kerch Strait		Core 14	–6.5	7.5	*Dreissena rostriformis*	МГУ-400	6500	250	Shcherbakov *et al.* 1977
Novyi Afon shelf	Coastal shelf	Core 510	–7.9	4.8	Mollusc shells	КИГН-410	6520	100	
Sukhumi coast	Right bank of Gumista River valley, 2 km NNW of mouth, on the sea shore	Core 50	4.5	18	Mollusc shells	ТБ-369	6520	70	Apakidze *et al.* 1987
Sukhumi shelf	2 km NNW of the mouth of the Gumista River	Core 722	–6	3.2	Mollusc shells	ТБ-385	6540	60	Apakidze *et al.* 1987
Sukhumi shelf	Interfluve betw. Kelasuri and Madzharka Rivers	Core 716	–10.4	7.1	Mollusc shells	ТБ-360	6540	60	Apakidze *et al.* 1987
Sukhumi coast	Distal part of Sebastopolisskyi Cape	Core 100	2.5	11.3	Peat	ТБ-372	6590	65	Apakidze *et al.* 1987
Novyi Afon shelf	Coastal shelf	Core 516	–5.3	6	Mollusc shells	КИГН-415	6590	110	
Lazarevskoe coast	Left bank of Psezuapse River (close to the mouth)	Core Л-7	0.82	13	Mollusc shells	ЛУ-701	6600	120	
Novyi Afon shelf	Coastal shelf	Core 501	–15.2	6.2	Mollusc shells	КИГН-403	6650	130	
Colchis coast	Chaladidi village, "Simagre" locality between Rioni River and Paleostomi Lake		6	19	Peat	ТБ-55	6660	100	Vinogradova *et al.* 1963; Neishtadt *et al.* 1965
Colchis coast	Right bank of Rioni River, center of Nabadskoe bog	Core 43	1	6	Peat	ГИН-127	6660	60	Cherdyntsev *et al.* 1966; Sluka 1969
Sukhumi coast	Right bank of Gumista River, 2 km NNW from mouth, on the sea shore	Core 50	4.5	20	Mollusc shells	ТБ-376	6690	70	Apakidze *et al.* 1987
Novyi Afon coast	Coastal part of the Aapsta River valley	Core 58	6.53	17	Wood in liman clay	КИГН-204	6770	130	

Novyi Afon coast	Coastal part of the Aapsta river valley	Core 55	1.62	23.2	Wood in marine clay	КИГН-205	6780	120	
Northwestern shelf	Odessa-Dniester seaside	Core 189	–16.6	8.4	*Dreissena polymorpha*, *Viviparus viviparus*, *Lithoglyphus naticoides*, *Theodoxus pallasi*		6800	90	Inozemtsev *et al.* 1984
Adler coast	Interfluve between Mzymta and Psou Rivers, 3.4 km S of the Mzymta River	Core A-12	2	13	Peat	ЛУ-699	6810	120	
Lazarevskoye coast	Right bank of Psezuapse River (close to the mouth)	Core 401	1.77	20	Mollusc shells	ЛУ-509	6820	210	
Novyi Afon shelf	Coastal shelf	Core 503	–8.7	3	Mollusc shells	КИГН-399	6890	120	
Sukhumi coast	Left bank of Gumista River, near the mouth	Core 63	2	24	Mollusc shells	ТБ-378	6920	70	Apakidze *et al.* 1987
Gudauta coast	?	Core 1	2	12	Peat	ТБ-400	6960	60	Apakidze *et al.* 1987
Adler coast	Interfluve betw. Mzymta and Psou Rivers, 3.5 km S of the Mzymta River	Core 1203y	2	20	Peat	ЛУ-184	6970	120	Arslanov *et al.* 1975
Adler coast	Interfluve betw. Mzymta and Psou Rivers, rear part of marine terrace	Core A-1	8.85	21	Peat	ЛУ-635	7000	140	
Gagra shelf	Underwater slope, S of Colchis channel (Ol'ginka River)	Outcrop without number	0	5	Peat	ТБ-47	7060	100	Burchuladze *et al.* 1975
Adler coast	Interfluve betw. Mzymta and Psou Rivers, middle part of marine terrace	Core A-2	1.17	20.3	Peat	ЛУ-638	7090	90	
Sukhumi coast	W part of Sukhumi Cape	Core 49	2.5	16	Mollusc shells	ТБ-377	7140	70	Apakidze *et al.* 1987
Northwestern shelf	Tiligulskii liman	Core 9	–9	10.5	Mollusc shells		7150	190	Molodykh *et al.* 1984

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Sukhumi coast	Left bank of Gumista River, 1.3 km SSE, sea shore	Core 93	3	15	Mollusc shells	ТБ-379	7210	70	Apakidze *et al.* 1987
Northwestern coast	2.5 km from Primorskoe village	Core 219	1.5	29.8	Plant remains in Bugazian overlapping Neoeuxinian sediments	ЛГ-165	7250	190	Badinova *et al.* 1978
Northwestern shelf	Dniepro-Bugian liman	Core 621	–5.8	16.5	Mollusc shells in layer of peat between Bugazian and Vityazevian sediments	КИ-1576	7300	120	
Sukhumi shelf	Interfluve between Kelasuri and Madzharka Rivers	Core 717	–22.5	9.2	*Cardium edule*, *Corbula mediterranea*, *Hydrobia ventrosa*	ТБ-356	7300	70	Apakidze *et al.* 1987
Adler coast	High floodland, 4.2 km from mouth of Psou River	Core 545б	16.93	25	Wood in delta-liman sediments	ЛУ-687	7310	120	
Sukhumi coast	W part of Sukhumi Cape	Core 49	2.5	22	Mollusc shells	ТБ-375	7310	70	Apakidze *et al.* 1987
Kuban River Delta	Small Dziginka village	Core Д-2	1	20.15	Peat	ЛУ-1882	7380	80	Izmailov *et al.* 1989
Northwestern shelf	Dniepro-Bugian liman	Core 858	–2.2	8.75	Peat	КИ-1583	7500	110	Gozhik and Novosel'sky 1989
Sukhumi shelf	2 km NNW from mouth of the Gumista River	Core 722	–6	10.3	Mollusc shells	ТБ-387	7500	70	Apakidze *et al.* 1987
Sukhumi coast	W part of Sukhumi Cape	Core 41	1.3	14	Mollusc shells	ТБ-373	7500	70	Apakidze *et al.* 1987
Northwestern shelf	Dniepro-Bugian liman	332	4.4	13.0	Mollusc shells		7520	120	Molodykh *et al.* 1984
Crimea shelf	W of Sevastopol and S of Balaklava	Core 101	–97	1	*Dreissena rostriformis*	МГУ-401	7560	140	Parunin *et al.* 1974
Adler coast	Right bank of Mzymta River	Core A-14	2.58	20	Wood in peat	ЛУ-748	7610	70	

Sukhumi shelf	2 km NNW of the mouth of the Gumista River	Core 723	–9.8	13	*Corbula mediterranea*, *Mytilaster lineatus*, *Cardium edule*, *Abra ovata*, *Hydrobia ventrosa*, rarely *Monodacna caspia*	ТБ-363	7630	250	
Adler coast	Interfluve betw. Mzymta and Psou Rivers, 3.4 km S of the Mzymta River	Core 1204y	2	23	Peat	ЛУ-185	7760	130	Arslanov *et al.* 1975
Northwestern shelf	Dniester coastal zone	Core 183	–9.9	12.3	*Abra ovata*, *Dreissena polymorpha*, *Monodacna caspia pontica*		7800	90	Inozemtsev *et al.* 1984
Northwestern shelf	Odessa sand-bank	Core 67	–24	1.8	Mollusc shells in alluvium: *Viviparus viviparus*, *Clessiniola variabilis*, *Lithglyphus naticoides*		7800	90	Inozemtsev *et al.* 1984
Adler coast	Interfluve betw. Mzymta and Psou Rivers, 3.5 km S of the Mzymta River	Core 1203y	2	30	Wood in peat	ЛУ-183	7850	120	Arslanov *et al.* 1975
Sukhumi coast	Distal part of Sukhumi Cape	Core 48	1.5	38	Mollusc shells	ТБ-374	7860	70	Apakidze *et al.* 1987
Colchis coast	Right bank of the mouth of Rioni River, near Poti		2	19	Peat above shallow marine sediments	ТБ-86	7900	60	Dzhanelidze 1980
Adler coast	High floodland of Psou River, 4.2 km from mouth	Core 544	19.85	22	Wood in delta alluvium	ЛУ-686	8000	150	
Colchis coast	Valley of Supsa River, 5-6 km from the mouth		9	21	Peat	МГУ; ТЛ-11	8000	940	Kvavadze 1978
Northwestern shelf	Odessa sand-bank	60	–7.5	6.6	*Dreissena polymorpha*, *Monodacna caspia pontica*, *Clessiniola variabilis*		8030	80	Inozemtsev *et al.* 1984
Adler coast	Right bank of Mzymta River	Core 1-14	2.58	22	Wood in peat	ЛУ-752	8050	70	
Adler coast	High floodland of Psou River, 4.2 km from mouth	Core 544	19.85	35	Peat in delta-liman sediments	ЛУ-689	8090	50	

Geographic Area*	Locality	Mode of Extraction	Elevation/ Water Depth (m)	Depth Within Core/ Outcrop (m)	Lithologic Unit/ Description	Laboratory Number**	Uncali-brated Age	± Error	References***
Northwestern shelf	Odessa sand-bank	Core 14/16		0.5	Peat, bark		8130	70	Inozemtsev *et al.* 1984
Northwestern shelf	Budakskii liman	Core 221	–2.4	17	Bugazian-like molluscs in marine sediments atop continental sediments		8200	105	Palatnaya 1982
Adler coast	Interfluve between Mzymta and Psou Rivers, rear part of marine terrace	Core A-1	8.85	27.5	Sphagnous peat	ЛУ-636	8210	130	
Sukhumi shelf	2 km NNW from the mouth of Gumista River	Core 723	–9.8	13	*Mytilaster lineatus* (young specimen), *Cardium edule*, *Abra ovata*, *Hydrobia ventrosa*, *Dreissena polymorpha* (mass form), *Monodacna caspia*, *Theodoxus pallasi*, *Micromelania elegantula*, *Clessiniola variabilis*	ТБ-364	8260	300	
Adler coast	Interfluve betw. Mzymta and Psou Rivers, middle part of marine terrace	Core A-2	1.17	27.7	Peat	ЛУ-637	8330	60	
Northwestern shelf	Odessa sand-bank	Core 1		12.2	Silty peat		8400	90	Inozemtsev *et al.* 1984
Crimean shelf	W of Sevastopol and S of Balaklava; S shore Crimea	Core 101	–97	2.5	*Dreissena rostriformis*	МГУ-319	8550	130	Parunin *et al.* 1979; Shcherbakov *et al.* 1977
Northwestern coast	Adzalykskii liman, Bolshoy sand spit, Novaya Dophnovka village	Core 289	1.5	25.7	Marine molluscs in abundant plant debris	ЛГ-161	8610	170	Badinova *et al.* 1978
Northwestern coast	Dniester liman	Core 157	2	25.2			8700	150	Inozemtsev *et al.* 1984

Pitsunda Peninsula	Central part of peninsula	Core 125	1.6	77	*Dreissena polymorpha*, *Monodacna caspia*, *Theodoxus pallasi*, *Micromelania caspia*, control of samples ЛГ-168 and ЛУ-365	ВСЕГИНГЕО II - 14	8850	160	Kuptsov *et al.* 1975; Balabanov *et al.* 1981
Northwestern shelf	Grigor'evka village, 60 km E of Odessa, near the shore	Core 10		10	Peat	МГУ-48	8880	290	Parunin *et al.* 1979; Kaplin *et al.* 1977
Gudauta shelf	Cape Souk-Su	Core 522	–16.5	10.55	Peat	КИГН-443	8890	150	
Northwestern shelf	Central part of Golitsin Uplift	Core 151	–38	3	Peat		8900	130	Inozemtsev *et al.* 1984
Northwestern shelf	Odessa sand-bank	Core 1/16		3.6	Peat		9000	100	Inozemtsev *et al.* 1984
Northwestern shelf	Dniester coastal zone	Core 182	–12.9	6.7	Silt with plant debris		9000	100	Inozemtsev *et al.* 1984
Northwestern shelf	Odessa-Dniester seaside	Core 166	–23.5	2.5	Mollusc shells in alluvium (?)		9100	120	Inozemtsev *et al.* 1984
Gudauta shelf	Left bank of Khipsta River	Core 535	–8.7	10.1	Peat	КИГН-446	9180	380	
Gagra shelf	Interfluve betw. Ol'ginka and Tskherva Rivers	Core 617	–15.5	32.8	Mollusc shells	ЛУ-1060	9220	1770	Dzhandzhgava and Balabanov 1982
Northwestern shelf	Grigor'evka village, 60 km E of Odessa, near the shore			16.3	Peat	МГУ-47	9240	380	Parunin *et al.* 1974; Kaplin *et al.* 1977
Sukhumi shelf	Sukhumi Bay, 0.8 km SE of the Basla River mouth	Core 721	–14.9	26.7	Peat	ТБ-346	9310	80	Apakidze *et al.* 1987; Yanko and Troitskaya 1987
Novyi Afon shelf	Coastal shelf	Core 503	–8.7	10.4	Mollusc shells	КИГН-401	9350	350	
Northwestern shelf	Open sea	Core 187	–19.6	3.3	Mollusc shells in alluvium: *Dreissena polymorpha*, *Clessiniola variabilis*, *Lithoglyphus naticoides*, *Viviparus viviparus*		9400	180	Inozemtsev *et al.* 1984

Geographic Area*	Locality	Mode of Extraction	Elevation/ Water Depth (m)	Depth Within Core/ Outcrop (m)	Lithologic Unit/ Description	Laboratory Number**	Uncali-brated Age	± Error	References***
Northwestern shelf	Odessa Bay	Core 162	–10.1	18.2	*Dreissena polymorpha*, *Dreissena rostriformis*, *Lithoglyphus naticoides*, *Viviparus viviparus*		9500	150	Inozemtsev *et al.* 1984
Northwestern shelf	Odessa sand-bank	Core 10	–12.9	14.9	Mollusc shells		9600	120	Inozemtsev *et al.* 1984
Northwestern shelf	Open sea	Core 185	–17.7	5.3	Peat		9700	150	Inozemtsev *et al.* 1984
Northwestern coast	Sand spit of Kuyalnitsky lagoon	Core 153	2	32.2	Peat		9800	400	Inozemtsev *et al.* 1984
Northwestern shelf	Odessa Bay	Core 162	–10.1	19.4	Peat		9900	110	Inozemtsev *et al.* 1984
Pitsunda Peninsula	Central part of peninsula	Core 125	1.6	77	*Dreissena polymorpha*, *Monodacna caspia*, *Theodoxus pallasi*, *Micromelania caspia*, control of samples ВСЕГИНГЕО II - 14 and ЛУ-365	ЛГ-168	9960	340	Balabanov *et al.* 1981
Pitsunda Peninsula	Ancient lagoon in the rear part of the peninsula	Core 138	0.3	53.4	Wood in peat above marine sediments with *Dreissena polymorpha*, *Monodacna caspia*, rarely *Clessiniola variabilis*	ВСЕГИНГЕО II - 124	10,130	180	Balabanov *et al.* 1981
Northwestern shelf	Golitsin Uplift, Core 4M, 1971y.	Core 1		2	*Viviparus* in alluvium-lacustrine sediments (?)		10,200	450	Kovalyukh *et al.* 1977
Pitsunda Peninsula	Central part of peninsula	Core 125	1.6	77	*Dreissena polymorpha*, *Monodacna caspia*, *Theodoxus*, *Micromelania*, control of samples ВСЕГИН ГЕО II -14 and ЛГ-168	ЛУ- 365	10,220	140	Balabanov *et al.* 1981

Sukhumi shelf	2 km NNW of the Gumista River mouth	Core 724	–14	32.6	*Dreissena polymorpha*, *Theodoxus pallasi*, rarely *Monodacna caspia*, *Clessiniola variabilis*	ТБ-370	10,350	100	
Adler coast	Interfluve between Mzymta and Psou Rivers, 2.5 km S of the Mzymta River, central part of marine terrace	Core 1206y	3	83	Shell rock with fine sand: *Dreissena polymorpha*, *Theodoxus pallasi*, rarely *Monodacna caspia*, *Micromelania caspia*, *Clessiniola variabilis*	ЛУ-350	10,350	270	Arslanov *et al.* 1975
Northwestern shelf	Odessa sand-bank	Core 10	–12.9	13.6	Mollusc shells in alluvium (?) *Dreissena polymorpha*		10,400	180	Inozemtsev *et al.* 1984
Northwestern shelf	Odessa-Dniester seaside	Core 168	–27	7.4	Mollusc shells in alluvium: *Dreissena polymorpha*, *Viviparus viviparus*, *Micro-melania lincta*, *Unio* sp.		10,500	150	Inozemtsev *et al.* 1984
Kerch Strait	Chushka Spit	Core 259	0.4	41.2	Mollusc shells	ЛУ-366	10,530	190	
Colchis coast	Coastal peat, mouth of Khobi River, Kulevi village	Core 47	1	42	Plant detritus	ГИН-632	10,550	200	Kind *et al.* 1975
Northwestern shelf	Odessa-Dniester seaside	Core 101		11.6	Peat		10,600	300	Inozemtsev *et al.* 1984
Northwestern shelf	Odessa sand-bank	Core 26/18	13	1	Peat		10,600	160	Inozemtsev *et al.* 1984
Kerch Strait	?	Core 17	–6.5	12.5	*Dreissena rostriformis*	МГУ-405	10,800	200	Shcherbakov *et al.* 1977
Northwestern shelf	Odessa sand-bank	Core 10	–12.9	16	Peat		10,800	600	Inozemtsev *et al.* 1984

* Geographic Regions:
Northwestern: Ukraine
Kerch-Taman region: Kuban River Delta, Kerch Strait
Caucasus Black Sea coast (from N to S): Anapa, Tuapse, Lazarovkoe, Adler, Gagra, Pitsunda, Gudauta, Novyi Afon, Sukhumi, Kodori, Ochamchiri, Colchis, Kobuleti, Batumi

** Laboratory Index of Organizations Performing Radiocarbon analyses:

Мо	Institute of Geochemistry and Analytical Chemistry of the Academy of Sciences of the USSR
ГИН	Geological Institute of the Academy of Sciences of the USSR
МГУ	Moscow State University
ВСЕГИНГЕО	All-Union Scientific Research Institute of Hydrogeology and Engineering Geology, Ministry of Geology of the USSR
ЛЕ	Leningrad Branch of the Institute of Archaeology of the Academy of Sciences of the USSR
ЛГ	All-Union Research Geological Institute of the Academy of Sciences of the USSR
ЛУ	Research Geographical-Economic Institute of Leningrad State University
КИ	Institute of Geochemistry and Physics of Minerals of the Academy of Sciences of the Ukrainian Soviet Socialist Republic
КИГН	Institute of Geological Sciences of the Academy of Sciences of the Ukrainian Soviet Socialist Republic
ТБ	Tbilisi State University
ТА	Institute of Zoology and Botany of the Academy of Sciences of the Estonian Soviet Socialist Republic

*** When a primary source is not specified, the date is being published here for the first time. All sources appear in the reference list for the paper by Balabanov in this volume, pp. 725-730.

Appendix 2. Radiocarbon dates for Late Pleistocene and Holocene sediments of the Black Sea and Sea of Azov coasts and shelf from former USSR sources not included in Appendix 1 and from western sources, arranged from youngest to oldest. Compiled by V.V. Yanko-Hombach.

Geographic Area*	Locality	Mode of Extraction	Elevation/ Water Depth (m)	Depth Within Core/ Outcrop (m)	Lithological Unit/Description of Dated Material/Age	Laboratory Number**	Uncali-brated Age	± Error	References***
Caucasian shelf		Core 36	–215	0.0–0.08		КИ-669	100		Mitropol'sky, pers. comm., 12.07.02
Northwestern shelf		Core 3	53	0.05	Mollusc shells in silty mud	КИ-682	100		Balandin and Mel'nik 1987
Northwestern shelf	Profile Tarkhankut-Cape Zmeinyi	Core 12	–90	0.0–0.1	Shell-bearing silty mud	КИ-659	120	70	Mitropol'sky, pers. comm., 12.07.02
Northwestern shelf		Core 12	–90	0.1	Mollusc shells in silty mud	КИ-46659	120	70	Balandin and Mel'nik 1987
Crimea shelf		Core 1306	–130	0.0–0.03	Shell-bearing silty mud	КИ-146	130	50	Mitropol'sky, pers. comm., 12.07.02
Northwestern shelf		Core 30	–90	0.1	Mollusc shells in calcareous silty mud	КИ-665	160	60	Balandin and Mel'nik 1987
Caucasian shelf		Core 38	–192	0.0–0.15	Mud	КИ-678	240	50	Mitropol'sky, pers. comm., 12.07.02
Crimean shelf		Core 3	–80	0.0–0.12	Mud	КИ-667	270	70	Mitropol'sky, pers. comm., 12.07.02
Crimean shelf		Core 26	–80	0.0–0.12	Silty mud	КИ-667	270	150	Mitropol'sky, pers. comm., 12.07.02
Northwestern shelf		Core 4	53	0.1	Mollusc shells in clayey mud	КИ-680	270	70	Balandin and Mel'nik 1987
Northwestern shelf	Profile Tarkhankut-Cape Zmeinyi	Core 7	–85	0.0–0.1	Shell-bearing silty mud	КИ-661	350	80	Mitropol'sky, pers. comm., 12.07.02
Crimean shelf		Core 1306	–130	0.03–0.12	Clayey mud with *Modiolus phaseolinus*	КИ-156	620	100	Mitropol'sky, pers. comm., 12.07.02
Caucasian shelf		Core 39					670	90	Mitropol'sky, pers. comm., 12.07.02

Geographic Area*	Locality	Mode of Extraction	Elevation/ Water Depth (m)	Depth Within Core/ Outcrop (m)	Lithological Unit/Description of Dated Material/Age	Laboratory Number**	Uncali-brated Age	± Error	References***
Northwestern shelf		Core 29	–84	0.4	Mollusc shells in silty mud	КИ-666	990	10	Balandin and Mel'nik 1987
Crimea shelf		Core 1306	–130	0.12–0.24	Clayey mud with *M. phaseolinus*	КИ-148	1400	120	Mitropol'sky, pers. comm., 12.07.02
Northwestern shelf	Left bank of Kuyalnitskii liman	Core 51	0.5	1	Kalamitian mollusc shells		1440	40	Gozhik *et al.* 1987
Northwestern shelf		Core 7	–70	0.2–0.3	Calcareous silty mud	КИ-664	1520	190	Mitropol'sky, pers. comm., 12.07.02
Northwestern shelf	Profile Cape Khersones-Romanian coast	Core 19	–190	0.2–0.26	Calcareous silty mud	КИ-662	1620	170	Mitropol'sky, pers. comm., 12.07.02
Northwestern shelf	Yagorlytskii Bay	Core 4/163	–5.2	0.0–1.2	Dzhemetinian mollusc shells		1700	90	Gozhik *et al.* 1987
Caucasian shelf		Core 38	–192	1.3–1.42	Mud	КИ-679	1740	100	Mitropol'sky, pers. comm., 12.07.02
Crimean shelf		Core 3	–53	0.8–0.9		КИ-683	1750	100	Mitropol'sky, pers. comm., 12.07.02
Northwestern shelf		Core 27	–88	0.88	Mollusc shells in silty mud	КИ-672	1800	100	Balandin and Mel'nik 1987
Northwestern shelf	Central trench	Core 38p	–19	0.4–0.45	Dzhemetinian mollusc shells		1830	60	Gozhik *et al.* 1987
Northwestern shelf	Central trench	Core 3p	–19	0.1–0.13	Dzhemetinian mollusc shells		1880	70	Gozhik *et al.* 1987
Northwestern shelf		Core 4	–53	0.83	Mollusc shells in clayey mud	КИ-681	1930	100	Balandin and Mel'nik 1987
Northwestern coast	Dniestovskii liman	Core 6 ИГН АН УССР	–60	60.4	Calcareous mud	КИ-655	1950	200	Balandin and Mel'nik 1987
Northwestern shelf		Core 1	–60	0.3–0.4	Calcareous silty mud	КИ-655	1950	200	Mitropol'sky, pers. comm., 12.07.02
Northwestern shelf	Central trench	Core 38p	–19	0.65–0.7	Dzhemetinian mollusc shells		2110	90	Gozhik *et al.* 1987

Southwestern shelf	41°50.38' N x 28°37.54' E	Core MAR00-06	–127	0.45	*Mytilus* spp.	TO-9138	2160	60	Aksu *et al.* 2002c
Northwestern shelf	Tendrovskii Gulf	Core 35/134	–15.6	1.5–1.6	Dzhemetinian mollusc shells		2300	80	Gozhik *et al.* 1987
Northwestern shelf	Danube	Core 17					2350	130	Mitropol'sky, pers. comm., 12.07.02
Northwestern shelf		Core 17	–150	0.4	Mollusc shells in silty mud	КИ-642	2350	130	Balandin and Mel'nik 1987
Northwestern shelf	Tendrovskii Gulf	Core 35/134	–15.6	1.8–2.0	Dzhemetinian mollusc shells		2450	80	Gozhik *et al.* 1987
Northwestern shelf	Dniester seashore	Core 181	–12.9	1.2–2.0	Dzhemetinian mollusc shells		2500	50	Gozhik *et al.* 1987
Northwestern shelf	Profile Tarkhankut-Cape Zmeinyi	Core 5	–52	0.65–0.75	Shell-bearing silty mud	КИ-660	2540	100	Mitropol'sky, pers. comm., 12.07.02
Northwestern shelf	Tendrovskii Gulf	Core 22/138	–12.7	0.0–0.45	Mollusc shells		2600	100	Gozhik *et al.* 1987
Northwestern shelf	Tendrovskii Gulf	Core 22/138	–12.7	0.45–0.85	Dzhemetinian mollusc shells		2640	90	Gozhik *et al.* 1987
Northwestern shelf	Danube	Core 16					2700	105	Mitropol'sky, pers. comm., 12.07.02
Northwestern shelf		Core 16	–140	0.6	Mollusc shells in silty mud	КИ-673	2700	105	Balandin and Mel'nik 1987
Western shelf		Core 2362	–102	2.5 –2.65	Silty mud with Neoeuxinian fauna	ИОАН-155	2740		Malovitsky *et al.* 1979
Turkish shelf	Near Sinop	Frame dredge	–140 to –170		*Trophonopsis breviata*	OS-21652	2800	45	Ballard *et al.* 2000
Turkish coast	Coastal plain of Sakarya River	Core KSK-20		8.5	Wood		2810	125	Görür *et al.* 2001
Turkish shelf	Near Sinop	Frame dredge	–140 to –170		*Modiolula phaseoline* (correct Latin name is *Modiolus phaseolinus*)	OS-21645	2810	40	Ballard *et al.* 2000
Northwestern shelf	Tendrovskii Gulf	Core 20/134	–13	1.1–1.45	Dzhemetinian mollusc shells		2880	100	Gozhik *et al.* 1987
Northwestern shelf	Tendrovskii Gulf	Core 20/134	–13	1.45–1.65	Dzhemetinian mollusc shells		2930	70	Gozhik *et al.* 1987

Geographic Area*	Locality	Mode of Extraction	Elevation/ Water Depth (m)	Depth Within Core/ Outcrop (m)	Lithological Unit/Description of Dated Material/Age	Laboratory Number**	Uncali-brated Age	± Error	References***
Northwestern shelf	Profile Cape Tarkhankut-Romanian coast	Core 18	–180	0.5–0.6	Calcareous silty mud	КИ-663	2960	180	
Northwestern shelf	Tendrovskii Gulf	Core 35/134	–15.6	2.35	Dzhemetinian mollusc shells		2970	80	Gozhik *et al.* 1987
Northwestern shelf	Profile Tarkhankut-Cape Zmeinyi	Core 2	–50	0.9–1.0	Calcareous silty mud	КИ-656	2980	170	Mitropol'sky, pers. comm., 12.07.02
Northwestern shelf	Dnieper-Bugskii liman	Core 5/144	–4.8	1.8–1.9	Dzhemetinian mollusc shells		3010	90	Gozhik *et al.* 1987
Sea of Azov	Belosaraiskaia spit	Core 36	–4		Novoazovian *Mytilus galloprovincialis*	КИ-332	3100	170	Semenenko and Kovalyukh 1973
Turkish shelf	Near Sinop	Frame dredge	–140 to –170		*Parvicardium exiguum*	OS-21656	3190	50	Ballard *et al.* 2000
Turkish shelf	Near Sinop	Frame dredge	–140 to –170		*Papillocardium papillosum*	OS-21644	3200	70	Ballard *et al.* 2000
Western shelf		Core 2362	–102	0.1–0.2	Mud with *Phaseolinus*	Woods Hole	3400		Malovitsky *et al.* 1979
Western shelf		Core 2362	–102	0.1–0.2	Mud with *Phaseolinus*	Woods Hole	3400		Malovitsky *et al.* 1979
Sea of Azov	Near Belosaraiskaia spit	Core 36	–8		Dzhemetinian *Cardium edule*	КИ-306	3450		Semenenko and Kovalyukh 1973
Northwestern shelf		Core 711	–39.4	0	Dzhemetinian mollusc shells		3480	45	Gozhik *et al.* 1987
Northwestern shelf	Dnieper-Bugskii liman	Core 3a/158	–3.2	1.7–1.9	Dzhemetinian mollusc shells		3500	55	Gozhik *et al.* 1987
Western shelf		Core 2345	–122	0.15–0.25	Mud with *Phaseolinus*	Woods Hole	3780		Malovitsky *et al.* 1979
Turkish shelf	Near Sinop	Frame dredge	–140 to –170		*Bittium reticulatum*	OS-21653	3860	50	Ballard *et al.* 2000
Northwestern shelf	Central trench	Core 187	–19.6	0.7–1.0	Dzhemetinian mollusc shells		4000	80	Gozhik *et al.* 1987
Western shelf		Core 2360	–91	0.3–0.4	Mud with *Modiolus*	Woods Hole	4020		Malovitsky *et al.* 1979

Northwestern shelf	Central trench	Core 187	–19.6	0.0–0.7	Dzhemetinian mollusc shells		4100	70	Gozhik *et al.* 1987
Turkish shelf	Near Sinop	Frame dredge	–140 to –170		*Abra ovata*	OS-21654	4130	45	Ballard *et al.* 2000
Northwestern shelf	Central part, profile Cape Khersones-Romanian coast	Core 21	–64	0.44–0.52	Calcareous silty mud	КИ-643	4170	310	Mitropol'sky, pers. comm., 12.07.02
Northwestern shelf	Central trench	Core 39p	–19	2.6–2.65	Dzhemetinian mollusc shells		4330	100	Gozhik *et al.* 1987
Northwestern shelf	Dnieper-Bugskii liman	Core 3a/158	–3.2	2.0–2.15	Dzhemetinian mollusc shells		4400	100	Gozhik *et al.* 1987
Turkish coast	Coastal plain of Sakarya River	Core KSK-18	2	8.5	Shell		4900	180	Görür *et al.* 2001
Northwestern shelf	Tendrovskii Gulf	Core 38/140	–13	0.5–0.65	Kalamitian mollusc shells		4930	80	Gozhik *et al.* 1987
Northwestern shelf	Dnieper 44° 57.6' N x 32°5.5' E	Core BS1	–68	0.64	L1/Marine *Mytilus galloprovincialis*		5080	115	Major 2002
Northwestern shelf	Tendrovskii Gulf	Core 30/129	–17.5	1.9–2.1	Kalamitian mollusc shells		5160	90	Gozhik *et al.* 1987
Northwestern shelf	Tendrovskii Gulf	Core 38/140	–13	1.15–1.25	Kalamitian mollusc shells		5300	125	Gozhik *et al.* 1987
Southwestern shelf	41°49.01' N x 28°30.68' E	Core MAR00-05	–83	0.6	*Mytilus* spp.	TO-9137	5460	70	Aksu *et al.* 2002b
Northwestern shelf	Tendrovskii Gulf	Core 22/138	–12.7	1.25–1.3	Kalamitian mollusc shells		5530	100	Gozhik *et al.* 1987
Northwestern shelf	Dniestrovskii trench	Core 493	–6.5	9.9–10.6	Kalamitian mollusc shells		5550	80	Gozhik *et al.* 1987
Northwestern shelf	Central trench	Core 38p	–19	2.7–2.73	Kalamitian mollusc shells		5560	90	Gozhik *et al.* 1987
Southwestern shelf	41°27.26' N x 29°16.01' E	Core MAR98-04	–112	0.24	*Mytilus*	TO-7782	5680	60	Aksu *et al.* 2002b
Northwestern shelf	Dnieper-Bugskii liman	Core 493	–6.5	13.0–13.1	Kalamitian mollusc shells		5700	90	Gozhik *et al.* 1987
Northwestern shelf	Yagorlytskii Bay	Core OGU	1	0.4–0.5	Kalamitian mollusc shells		5770	80	Gozhik *et al.* 1987

Geographic Area*	Locality	Mode of Extraction	Elevation/ Water Depth (m)	Depth Within Core/ Outcrop (m)	Lithological Unit/Description of Dated Material/Age	Laboratory Number**	Uncali- brated Age	± Error	References***
Sea of Azov	Central part	Core 39		4.6	Drevneazovian *Chione gallina*	КИ-307	5770		Semenenko and Kovalyukh 1973
Southwestern shelf	41°42.16' N x 28°43.32' E	Core MAR00-08	–96	0.54	*Mytilus* spp.	TO-9139	5780	70	Aksu *et al.* 2002b
Southwestern shelf	41°27.26' N x 29°16.01' E	Core MAR98-04	–112	1.04	*Mytilus*	TO-7783	5780	60	Aksu *et al.* 2002b
Northwestern shelf	Tendrovskii Gulf	Core 38/140	–13	0.75–0.95	Kalamitian mollusc shells		5800	100	Gozhik *et al.* 1987
Turkish coast	Coastal plain of Sakarya River	Core KSK-4	5	9.3	Shell		5810	160	Görür *et al.* 2001
Northwestern shelf	Central trench	Core 189	–16.6	8.4	Kalamitian mollusc shells		6000	90	Gozhik *et al.* 1987
Northwestern shelf	Profile Cape Khersones-Romanian coast	Core 20	–181	0.7–0.9	Calcareous silty mud	КИ-641	6110	240	Mitropol'sky, pers. comm., 12.07.02
Southwestern shelf	41°42.38' N x 29°06.31' E	Core MAR00-09	–115	1.19	*Mytilus* spp.	TO-9525	6132	139	Aksu *et al.* 2002b
Sea of Azov	Central part	Core 39			Drevneazovian *Chione gallina*	КИ-308	6200		Semenenko and Kovalyukh 1973
Caucasian coast	Pitsunda Bay	Core 2					6240	230	Balabanov *et al.* 1981
Northwestern shelf	Yagorlytskii Bay	Core 4/163	–5.2	1.2–1.75	Kalamitian mollusc shells		6280	75	Gozhik *et al.* 1987
Northwestern shelf	Central trench	Core 38p	–19	2.75–2.85	Kalamitian mollusc shells		6320	120	Gozhik *et al.* 1987
Northwestern shelf	Tendrovskii Gulf	Core 38/140	–13	1.0–1.15	Kalamitian mollusc shells		6400	105	Gozhik *et al.* 1987
Northwestern shelf		Core 817	–20	2.5	Peat with plant remains	КИ-65	6400	300	Balandin and Mel'nik 1987
Northwestern shelf		Core 1517	–130	0.4	Kalamitian mollusc shells		6430	100	Gozhik *et al.* 1987
Northwestern shelf	Dnieper 44°57.6' N x 32°5.5' E	Core BS1	–68	1.21	L1/Marine *Mytilus*		6470	30	Major 2002

Northwestern shelf	Tendrovskii Gulf	Core 20/134	–13	1.75–1.95	Kalamitian mollusc shells		6540	120	Gozhik *et al.* 1987
Southwestern shelf	41°42.16' N x 28°43.32' E	Core MAR00-08	–96	1.16	*Mytilus* spp.	TO-9090	6590	70	Aksu *et al.* 2002b
Northwestern shelf	Dnestrovskii trench	Core 92	–7.2	1.9–2.1	Kalamitian mollusc shells		6600	100	Gozhik *et al.* 1987
Southwestern shelf	41°49.01' N x 28°30.68' E	Core MAR00-05	–83	1.67	*Cardium* ssp.	TO-9088	6660	70	Aksu *et al.* 2002b
Northwestern shelf	Danube 44°15.2' N x 30°24.68' E	Core BLKS9801	–92	0.33	L1/Bugazian *Mytilus galloprovincialis*		6750	70	Major 2002
Southwestern shelf	41°19.82' N x 29°45.53' E	Core MAR00-23	–98	1.7	*Mytilus* spp.	TO-9526	6760	60	Aksu *et al.* 2002b
Western shelf		Core 62	–93	0.55–0.65	Sapropel and *Mytilus* mud	Woods Hole	6800		Malovitsky *et al.* 1979
Turkish shelf	Near Sinop	Frame dredge	–140 to –170		*Mytilus galloprovincialis*	OS-21655	6820	55	Ballard *et al.* 2000
Northwestern shelf	Dnieper 44°57.6' N x 32°5.5' E	Core BS1	–68	1.34	L1/Bugazian *Mytilus galloprovincialis*		6860	35	Major 2002
Northwestern shelf	Dnieper 44° 53.4' N x 32°9.3' E	Core BS7/AK93-7	–108	0.92	L1/Bugazian *Mytilus galloprovincialis*	OS-2357	6860	35	Major 2002; http://imager.ldeo.columbia.edu/BlackSeaShelf/Table.html
Western shelf		Core 2362	–102	0.6–0.7	Shell hash with Vityazevian fauna	ИОАН-157	6890	630	Malovitsky *et al.* 1979
Northwestern shelf	Dnieper 44°57.6' N x 32°5.5' E	Core BS1	–68	1.3	L1/Bugazian *Cardium exiguum*		6910	45	Major 2002
Northwestern shelf		Core 817	–20	2.6	Peat with plant remains	КИ-66	6920	120	Balandin and Mel'nik 1987
Caucasian coast	2 km NNW of Gumista River mouth	Core 722	–6	5.4–6.1	Mollusc shells	ТБ-386	7040	70	Balabanov, pers. comm., 09.07.03
Northwestern shelf		Core 6	–6.8	4	Peat with plant remains	КИ-51	7040	170	Balandin and Mel'nik 1987
Northwestern shelf	Tiligulskii liman	Core 67	Not known	5.5–7.2	Kalamitian mollusc shells		7080	140	Gozhik *et al.* 1987

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Northwestern shelf		Core 817	–20	2.7	Peat with plant remains	КИ-69	7100	105	Balandin and Mel'nik 1987
Northwestern shelf		Core 711	–39.4	0.25	Kalamitian mollusc shells		7110	70	Gozhik *et al.* 1987
Northwestern shelf	Dnieper 45°21.9' N x 31°49.7' E	Core BS3-2	–49	0.49	*Monodacna caspia*	OS-2321	7130	40	Major 2002; http://imager.ldeo.columbia.edu/BlackSeaShelf/Table.html
Northwestern shelf	44°58.8' N x 32°11.1' E	Core BS12/AK93-12	–78	1.44	L1/Bugazian *Cardium edule*	OS-2325	7140	45	Major 2002; http://imager.ldeo.columbia.edu/BlackSeaShelf/Table.html
Northwestern shelf	Dnieper 44°54.4' N x 32°8.5' E	Core BS8	–99	1	L1/Bugazian *Cardium exiguum*		7140	45	Major 2002
Northwestern shelf	Dnieper 44°57.6' N x 32°5.5' E	Core BS1/AK93-1	–68	1.34	L1/Bugazian *Monodacna caspia*	OS-2357	7220	40	Major 2002; http://imager.ldeo.columbia.edu/BlackSeaShelf/Table.html
Turkish shelf	Near Sinop	Frame dredge	–140 to –170		*Turricaspia caspia lincta*	OS-21659	7460	55	Ballard *et al.* 2000
Western shelf		Core 2362	–102	0.7–0.9	Shell hash with mixed Bugazian and Vityazevian fauna	ИОАН-153	7480	540	Malovitsky *et al.* 1979
Turkish shelf	Near Sinop	Frame dredge	–140 to –170		*Turricaspia caspia lincta*	OS-21661	7480	55	Ballard *et al.* 2000
Northwestern coast	Dniestrovskii liman	Core 3	–6.6	13.9	Peat	КИ-50	7500	200	Balandin and Mel'nik 1987
Northwestern coast	Dnieper-Bugskii liman, Ochakovskii profile 1-1	Core 303a		12	Mud	КИ-733	7520	120	Mitropol'sky, pers. comm., 12.07.02

Northwestern coast	Dnieper-Bugskii liman, Ochakovskii profile I-I	Core 303a		12		КИ-733	7520	120	Mitropol'sky, pers. comm., 12.07.02
Northwestern shelf	Tendrovskii Gulf	Core 22/136	–15.7	3.2–3.5	Bugazian-Vityazevian mollusc shells		7590	110	Gozhik *et al.* 1987
Turkish shelf	Near Sinop	Frame dredge	–140 to –170		*Turricaspia caspia lincta*	OS-21660	7590	55	Ballard *et al.* 2000
Northwestern shelf	Danube 44°8.13' N x 29°21.85' E	Core BLKS9815	–93	0.55	Bugazian *Monodacna caspia*		7655	75	Major 2002
Northwestern shelf		Core 775	–24.5	0.45–0.6	Bugazian-Vityazevian mollusc shells		7670	90	Gozhik *et al.* 1987
Southwestern shelf	41°50.38' N x 28°37.54' E	Core MAR00-06	–127	1.24	*Mytilus* spp.	TO-9089	7770	70	Aksu *et al.* 2002b
Northwestern shelf	Tendrovskii Gulf	Core 22/138	–12.7	2.0–2.6	Bugazian-Vityazevian mollusc shells		7790	110	Gozhik *et al.* 1987
Northwestern shelf	Danube 44°0.64' N x 29°53.72' E	Core BLKS9831	–75	1.1	*Monodacna caspia* L2/Shell hash		7865	70	Major 2002
Northwestern shelf	Danube 44°15.2' N x 30°24.68' E	Core BLKS9801	–92	0.33	L1/Bugazian *Dreissena* sp.		7940	75	Major 2002
Northwestern shelf		Core 775	–24.5	0.7–0.85	Bugazian-Vityazevian mollusc shells		8000	110	Gozhik *et al.* 1987
Turkish coast	Coastal plain of Sakarya River	Core KSK-20	2	28	Peat		8090	120	Görür *et al.* 2001
Northwestern shelf	Dnieper 45°21.9' N x 31°49.7' E	Core BS3-2	–49	0.52	*Dreissena rostriformis* Shell hash		8185	90	Major 2002
Northwestern shelf	44°53.4' N x 32° 9.3' E	Core BS7	–108	1	*Dreissena rostriformis* L2/Shell hash		8250	35	Major 2002
Northwestern shelf	Tendrovskii Gulf	Core 33/129	–16.7	1.25–1.55	Bugazian-Vityazevian mollusc shells		8260	115	Gozhik *et al.* 1987
Northwestern shelf	Danube 44°2.86' N x 29°56.79' E	Core BLKS9838	–77	0.02	*Dreissena* sp. L2/Shell hash		8275	70	Major 2002
Northwestern shelf	Danube 44°0.66' N x 29°53.71' E	Core BLKS9834	–76	0.41	*Dreissena* sp. L2/Shell hash		8300	70	Major 2002
Northwestern shelf	Danube 44°0.64' N x 29°53.72' E	Core BLKS9830	–57	0.22	*Dreissena* sp. In sand		8305	80	Major 2002
Northwestern shelf	Danube 44°1.9' N x 29°54.84' E	Core BLKS9831	–75	1.61	*Dreissena* sp. L2/Shell hash		8345	70	Major 2002

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Northwestern shelf	Danube 44°1.9' N x 29°54.84' E	Core BLKS9831	–75	1.76	*Dreissena* sp. L2/Shell hash		8360	70	
Northwestern shelf	Danube 44°0.54' N x 29°58.87' E	Core BLKS9837	–68	0.39	*Dreissena* sp. L2/Shell-bearing mud		8360	75	Major 2002
Northwestern shelf	Danube	Core 12	–9.5	5.6	Peat with plant remains	КИ-120Б	8400	300	Balandin and Mel'nik 1987
Northwestern shelf		Core 1	–6.6	6.2	Peat with plant remains	КИ-119	8400	350	Balandin and Mel'nik 1987
Northwestern shelf	Dnieper 44°55.2' N x 32°8.0' E	Core BS11	–91	1.55	*Dreissena rostriformis* L2/Shell hash		8415	70	Major 2002
Northwestern shelf		Core 19a	–6	8.3	Peat with plant remains	КИ-118	8500	200	Balandin and Mel'nik 1987
Northwestern shelf	Central trench	Core 805	–21	1.25–1.3	Bugazian-Vityazevian mollusc shells		8510	95	Gozhik *et al.* 1987
Northwestern shelf	Kerch 44°40.0' N x 36°34.5' E	Core BS24	–110	1.64	*Dreissena polymorpha*		8550	80	Major 2002
Northwestern shelf		Core 778	–22.8	0.7–0.85	Bugazian-Vityazevian mollusc shells		8635	90	Gozhik *et al.* 1987
Northwestern shelf	Danube 44°15.2' N x 30°24.68' E	Core BLKS9801	–92	0.34	L1/Bugazian *Dreissena* sp. in shell-bearing sand		8660	75	Major 2002
Northwestern shelf		Core 711	–39.4	0.55	Bugazian-Vityazevian mollusc shells		8850	95	Gozhik *et al.* 1987
Northwestern shelf	Tiligulskii liman	Core 38	–11	12.6–15.0	Bugazian-Vityazevian mollusc shells		8900	60	Gozhik *et al.* 1987
Northwestern shelf		Core 711	–39.4	0.5	Neoeuxinian mollusc shells		8900	150	Gozhik *et al.* 1987
Sea of Azov	Belosaraiskii Bay	Core 26		18	Bugazian *Monodacna caspia*	КИ-262	8940	200	Semenenko and Kovalyukh 1973
Northwestern shelf	Kerch 44°39.2' N x 36°35.2' E	BS22	–129	1.1	*Dreissena polymorpha* L2/Shell hash		8990	75	Major 2002
Northwestern shelf	Central trench	Core 14/40	–23.6	3.15–3.3	Neoeuxinian mollusc shells		9100	150	Gozhik *et al.* 1987

Northwestern shelf	Kerch 44°39.2' N x 36°35.2' E	Core BS22	–129	1.1	*Dreissena rostriformis* L2/Shell hash		9235	75	Major 2002
Northwestern coast	Dnieper-Bugskii liman, Ochakovskii profile 1-1	Core 304a		12	Mud	КИ-734	9280	160	Mitropol'sky, pers. comm., 12.07.02
Sea of Azov	Belosaraiskii Bay	Core 26		18	Bugazian *Cardium edule*	КИ-261	9280	200	Semenenko and Kovalyukh 1973
Turkish shelf	Near Sinop	Frame dredge	–140 to –170		*Turricaspia caspia lincta*	OS-21658	9310	65	Ballard *et al.* 2000
Northwestern shelf	Central trench	Core 187	–19.6	3.3–4.1	Neoeuxinian mollusc shells		9400	180	Gozhik *et al.* 1987
Northwestern shelf	Tiligulskii liman	Core 67	–16	9.5–10.1	Neoeuxinian mollusc shells		9500	190	Gozhik *et al.* 1987
Northwestern shelf	Central trench	Core 162	–10.1	18.0–18.4	Neoeuxinian mollusc shells		9500	150	Gozhik *et al.* 1987
Northwestern shelf	Danube 44°7.99' N x 29°19.53' E	Core BLKS9814	–55	1.3	*Cardium exiguum* in shell-bearing, silty mud of saline pond		9580	80	Major 2002
Northwestern shelf	Tendrovskii Gulf	Core 4/60	–16	2.4–3.3	Neoeuxinian mollusc shells		9700	100	Gozhik *et al.* 1987
Northwestern coast	Dnieper-Bugskii liman	Core 329	–3	18	Mollusc shells	КИ-735	9800	180	Kovalyukh *et al.* 1977
Sea of Azov	Belosaraiskii Bay	Core 27		15.0–20.0	Neoeuxinian *Viviparus fasciatus*	КИ-282	9820		Semenenko and Kovalyukh 1973
Northwestern shelf	Dnieper 44°51.6' N x 32°21.3' E	Core BS14	–140	0.2	*Dreissena rostriformis* in L4/*Dreissena* coquina		10,000	50	Major 2002
Northwestern shelf	Danube 44°8.95' N x 30°39.24' E	Core BLKP9805	–131	0.14	*Dreissena* sp. in L4/*Dreissena* coquina		10,090	130	Major 2002
Northwestern shelf	Golitsin Uplift	Core 1(4M)		2	*Viviparus fasciatus*	КИ-286	10,200		Semenenko and Kovalyukh 1973
Northwestern shelf	Danube 44°15.2' N x 30°24.68' E	Core BLKS9801	–92	0.45	*Dreissena* sp. in L1/ Bugazian shell-bearing sand		10,250	90	Major 2002
Northwestern shelf	Danube 44°7.38' N x 30°42.98' E	Core BLKS9806	–135	0.11	*Dreissena* in L4/*Dreissena* coquina		10,260	95	Major 2002
Northwestern shelf	Golitsin Uplift	Core 2 (1m)		2.3	*Viviparus viviparus*		10,360	560	Shnyukov, ed., 1985

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Northwestern shelf	44°53.4' N x 32°9.3' E	Core BS7	–108	0.95	*Dreissena rostriformis* in L2/Shell hash		10,400	50	Major 2002
Turkish shelf	Near Sinop	Frame dredge	–140 to –170		*Dreissena rostriformis var. distincta*	OS-21650	10,400	70	Ballard *et al.* 2000
Northwestern shelf	Danube 44°5.76' N x 30°46.80' E	Core BLKS9807	–163	0.03	*Dreissena* sp. in L4/*Dreissena* coquina		10,430	80	Major 2002
Turkish shelf	Near Sinop	Frame dredge	–140 to –170		*Dreissena rostriformis var. distincta*	OS-21648	10,450	70	Ballard *et al.* 2000
Turkish shelf	Near Sinop	Frame dredge	–140 to –170		*Dreissena rostriformis var. distincta*	OS-21649	10,450	70	Ballard *et al.* 2000
Turkish shelf	Near Sinop	Frame dredge	–140 to –170		*Dreissena rostriformis var. distincta*	OS-21651	10,500	70	Ballard *et al.* 2000
Northwestern shelf	Danube 44°12.0' N x 30°32.21' E	Core BLKS9804	–101	0.14	*Dreissena* sp. in L4/*Dreissena* coquina		10,560	75	Major 2002
Northwestern shelf	Danube 44°4.04' N x 30°50.68' E	Core BLKS9810	–378	0.94	*Turricaspia caspia* in L5/Shell-bearing calcareous mud	ETH-23298	10,640	80	Major 2002;Major *et al.* 2002
Western shelf		Core 2361	–88	0.7–0.85	Neoeuxinian shell hash	Woods Hole	10,670		Malovitsky *et al.* 1979
Turkish shelf	Near Sinop	Frame dredge	–140 to –170		*Dreissena rostriformis var. distincta*	OS-21647	10,800	75	Ballard *et al.* 2000
Northwestern shelf		Core 19a	–6	10	Peat with plant remains	КИ-117	10,800	300	Balandin and Mel'nik 1987
Northwestern shelf	Dnieper 44°53.4' N x 32°9.3' E	Core BS7	–108	1.4	*Dreissena* in L4/*Dreissena* coquina		10,800	65	Major 2002
Sea of Azov	Belosaraiskii Bay	Core 27		19.2	Neoeuxinian *Viviparus fasciatus*	КИ-264	10,900	490	Semenenko and Kovalyukh 1973
Northwestern shelf		Core 1506	–200	1.1–1.4	Neoeuxinian mollusc shells		10,980	390	Gozhik *et al.* 1987
Northwestern shelf	Dnieper 44°54.4' N x 32°8.5' E	BS8	–99	1.3	*Dreissena rostriformis* in L4/Sandy *Dreissena* coquina		11,190	80	Major 2002

Northwestern shelf	Kerch 44°39.2' N x 36°35.2' E	Core BS22	–129	1.15	Unidentified gastropod in L4/Shelly mud		11,260	90	Major 2002
Northwestern shelf	Dnieper 44°53.5' N x 32°8.8' E	Core BS10/AK93-10	–106	1.45	*Dreissena rostriformis* in L4/Sandy *Dreissena* coquina	OS-2324	11,350	35	Major 2002; http://imager.ldeo.columbia.edu/BlackSeaShelf/Table.html
Northwestern shelf	Danube 44°4.04' N x 30°50.68' E	Core BLKS9810	–378	1.18	*Dreissena rostriformis* in L5/Shelly calcareous mud	ETH-23299	11,410	110	Major 2002; Major *et al.* 2002
Western shelf		Core 62	–93	0.9–1.1	Shell hash bearing silty sand with Neoeuxinian fauna	ИОАН-787	11,430	330	Malovitsky *et al.* 1979
Sea of Azov	Azov Sea	Core 40			Peat with *Planorbis* sp.	КИ-328	11,480	?	Semenenko and Kovalyukh 1973
Sea of Azov	Belosaraiskii Bay	Core 27		20–21.7	Neoeuxinian *Dreissena polymorpha*	КИ-284	11,480		Semenenko and Kovalyukh 1973
Northwestern shelf	Dnieper 44°53.0' N x 32°9.2' E	Core BS9	–123	1.15	*Turricaspia caspia* in L4/Sandy *Dreissena* coquina		11,540	85	Major 2002
Northwestern shelf	Budakskii liman	Core 221	–22	22.0–23.0	Neoeuxinian mollusc shells		11,570	140	Gozhik *et al.* 1987
Western shelf		Core 2345	–122	1.1–1.25	Shell hash with Neoeuxinian fauna	ИОАН-791	11,590	240	Malovitsky *et al.* 1979
Caucasian coast	Gudauta coast Cape Souk-Su	Core 120 M	2	16.3–16.9	Mollusc shells	КИГН-221	11,680	230	Balabanov, pers. comm., 07.16.02
Northwestern coast	Berezanskii liman	Core 657		13.8-13.9	Mollusc shells		11,700	180	Molodykh *et al.* 1984
Northwestern shelf	Central trench	Core 189	–16.6	11.3	Neoeuxinian mollusc shells		11,700	150	Gozhik *et al.* 1987
Western shelf		Core 2345	–122	1.4–1.5	Shell hash with Neoeuxinian fauna	ИОАН-161	11,710	1100	Malovitsky *et al.* 1979
Crimean shelf	Yalta	Core 165	–112	1.5	Mollusc shells		11,900	180	Gozhik *et al.* 1987
Northwestern shelf	Golitsin Uplift	Core 2 (1m)		2.3	Neoeuxinian *Viviparus fasciatus*, *V. contectus*	КИ-260	12,050		Semenenko and Kovalyukh 1973
Turkish shelf	Near Sinop	Frame Dredge	–140 to –170		*Turricaspia caspia lincta*	OS-21657	12,100	85	Ballard *et al.* 2000

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Northwestern coast	Dnieper-Bugskii liman	Core 854	–5.4	15.9–16.2	Mollusc shells	КИ-1582	12,200	180	Molodykh *et al.* 1984
Sea of Azov	Belosaraiskii Bay	Core 27		20–21.7	Neoeuxinian *Viviparus fasciatus*	КИ-265	12,250	400	Semenenko and Kovalyukh 1973
Northwestern shelf	Danube 44°5.23' N x 30°47.98' E	Core BLKS9809	–240	0.15	*Dreissena rostriformis* in L5/Shelly calcareous mud	ETH-22156	12,310	95	Major *et al.* 2002
Northwestern shelf	Dnieper 44°53.5' N x 32°8.8' E	Core BS10	–106	1.46	Wood in L5/Shell-bearing mud/silt		12,330	90	Major 2002
Northwestern shelf	Golitsin Uplift	Core 2 (1m)		2.3	Neoeuxinian *Dreissena polymorpha*	КИ-283	12,360	500	Semenenko and Kovalyukh 1973
Northwestern shelf		Core 894	–63	1.3	*Dreissena polymorpha* in shell-bearing sand	КИ-383	12,400	500	Balandin and Mel'nik 1987
Northwestern shelf	Dnieper-Bugskii liman	Core 1407	–7.4	15.8–16.0	Neoeuxinian mollusc shells		12,520	180	Gozhik *et al.* 1987
Northwestern shelf		Core 822	–31	3.8	Mollusc shells in loamy sand	КИ-811	12,700	500	Balandin and Mel'nik 1987
Northwestern shelf	Danube 44°4.04' N x 30°50.68' E	Core BLKS9810	–378	1.54	*Dreissena rostriformis* in L5/Shell-bearing mud	ETH-23300	12,790	110	Major 2002;Major *et al.* 2002
Northwestern shelf	Danube 44°4.04' N x 30°50.68' E	Core BLKS9810	–378	1.86	*Dreissena rostriformis* in L5/Shell-bearing mud	ETH-23301	12,920	100	Major *et al.* 2002
Sea of Azov	Belosaraiskii Bay	Core 26		19.2	Neoeuxinian *Monodacna caspia*	КИ-263	13,100	800	Semenenko and Kovalyukh 1973
Northwestern shelf	Dnieper 44°55.3' N x 32°16.5' E	Core BS13	–165	1.07	*Dreissena rostriformis* in L5/Shell-bearing mud/silt		13,250	90	Major 2002
Crimean shelf	South of Balaklava	Core 81	–100	2.9	Mollusc shells		13,500	500	Gozhik *et al.* 1987
Caucasian shelf	Pitsunda Peninsula, ancient swampy lagoon	Core 138	1	73.5–77.0	Mollusc shells	ВСЕГИН-ГЕО II-140	13,680	610	Kuptsov and Nechaev 1975
Northwestern shelf	Central trench	Core 67	–24	8.0–10.2	Neoeuxinian mollusc shells		13,800	110	Gozhik *et al.* 1987
Northwestern shelf	Danube 44°5.23' N x 30°47.98' E	Core BLKS9809	–240	1.15	*Dreissena rostriformis* in L5/Shell-bearing mud	ETH-22157	14,010	100	Major *et al.* 2002

Western shelf		Core 2378	–96	1.0–1.2	Terrigenous shell-bearing sediments with Neoeuxinian fauna	ИОАН-802	14,610	210	Malovitsky *et al.* 1979
Western shelf		Core 2378	–96	1.0–1.2	Mollusc shells in terrigenous shell-bearing sediments with Neoeuxinian fauna	ИОАН-802	14,610	200	Malovitsky *et al.* 1979
Northwestern shelf	Dnieper 44°51.6' N x 32°21.3' E	Core BS14/AK93-14	–140	2.15	*Dreissena rostriformis* in L5/*Dreissena*-bearing mud	OS-2360	14,700	65	Major 2002; http://imager.ldeo.columbia.edu/BlackSeaShelf/Table.html
Northwestern shelf	Danube 44°5.23' N x 30°47.98' E	Core BLKS9809	–240	2.15	*Dreissena rostriformis* in L5/Shell-bearing mud	ETH-21127	14,950	100	Major *et al.* 2002
Northwestern shelf		Core 814	–10	2.3		КИ-814	15,100	370	Balandin and Mel'nik 1987
Caucasian coast	North part of Gagra Bay	Core 432	–3.5	50.0–50.1	Mollusc shells	ЛУ-728	15,230	220	Balabanov, pers. comm., 2003
Turkish shelf	Near Sinop	Frame dredge	–140 to –170		*Dreissena rostriformis var. distincta*	OS-21646	15,550	120	Ballard *et al.* 2000
Northwestern shelf		Core 711	–39.4	0.8	Neoeuxinian mollusc shells		16,720	170	Gozhik *et al.* 1987
Western shelf		Core 2360	–91	1.0–1.25	Neoeuxinian shell- bearing silty mud	ИОАН-150	17,080	510	Malovitsky *et al.* 1979
Western shelf		Core 62	–93	1.4–1.5	Shell hash bearing silty sand with Neoeuxinian fauna	ИОАН-788	17,180	300	Malovitsky *et al.* 1979
Northwestern shelf		Core 1513	–79	1.5–1.7	Neoeuxinian mollusc shells		17,400	500	Gozhik *et al.* 1987
Northwestern shelf	Danube 44°4.04' N x 30°50.68' E	Core BLKS9810	–378	7.04	*Dreissena rostriformis* in L5/Shell-bearing mud	ETH-23302	17,760	130	Major 2002;Major *et al.* 2002
Crimean shelf	South of Balaklava	Core 81	–100	1	Mollusc shells		17,780	200	Gozhik *et al.* 1987
Caucasian coast	Coastal plain of Gagra Bay	Core 459	2.7	16	Peat	ЛУ-652	19,460	580	Balabanov, pers. comm., 07.16.02
Northwestern shelf	Danube 44°7.38' N x 30°42.98' E	Core BLKS9806	–135	0.38	*Dreissena rostriformis* in L5/Shell-bearing mud/silt		20,490	150	Major 2002

Geographic Area*	Locality	Mode of Extraction	Elevation/ Water Depth (m)	Depth Within Core/ Outcrop (m)	Lithological Unit/Description of Dated Material/Age	Laboratory Number**	Uncali-brated Age	± Error	References***
Northwestern shelf	Danube 44°5.23' N x 30°47.98' E	Core BLKS9809	–240	8.4	*Dreissena rostriformis* in L5/Shell-bearing mud	ETH-21128	20,580	150	Major *et al.* 2002
Northwestern shelf	Dnieper 44°55.2' N x 32°8.0' E	Core BS11	–91	1.75	Wood in L6/Stiff deltaic mud		22,750	200	Major 2002
Caucasian coast	Gudauta shelf, left bank of Khipsa River	Core 535	–8.7	11.3–11.5	Peat		23,900	680	Balabanov, pers. comm., 07.16.02
Northwestern shelf	Danube 44°5.47' N x 30°47.21' E	Core BLKS9808	–186	1.27	*Dreissena rostriformis* in L5/Shell-bearing mud/silt		24,160	190	Major 2002
Northwestern shelf	Danube 44°0.66' N x 29°53.71' E	Core BLKS9834	–76	0.6	Bulk organic C in L6/Stiff deltaic mud		24,980	200	Major 2002
Northwestern shelf	Danube 44°0.66' N x 29°53.71' E	Core BLKS9834	–76	2.35	Bulk organic C in L6/Stiff deltaic mud		26,630	230	Major 2002
Western shelf		Core 2345	–122	2.75–2.8	Silty mud with Neoeuxinian fauna	ИОАН-162	26,950		Malovitsky *et al.* 1979
Western shelf		Core 2362	–102	4.75–5.0	Silty mud with Neoeuxinian fauna	ИОАН-154	27,295	1120	Malovitsky *et al.* 1979
Western shelf		Core 2345	–122	4.1 –4.2	Clayey mud with Neoeuxinian fauna	ИОАН-164	27,780		Malovitsky *et al.* 1979
Western shelf		Core 2345	–122	4.3–4.4	Clayey mud with Neoeuxinian fauna	ИОАН-165	28,100		Malovitsky *et al.* 1979
Western shelf		Core 2345	–122	2.25–2.4	Silty mud with pebbles and Neoeuxinian fauna	ИОАН-790	29,650		Malovitsky *et al.* 1979
Western shelf		Core 2378	–96	2.3–2.4	Silty mud with Neoeuxinian fauna	ИОАН-792	30,170		Malovitsky *et al.* 1979
Western shelf		Core 2378	–96	1.7–2.0	Terrigenous shelly sediment with Neoeuxinian fauna	ИОАН-794	31,420		Malovitsky *et al.* 1979
Southwestern shelf	41°27.26' N x 29°16.01' E	Core MAR98-04	–112	1.18	White mussel	TO-7784	33,550	330	Aksu *et al.* 2002b
Northwestern shelf		Core 3	–10	3	Peat in loess-like loam	КИ-812	33,600	1500	Balandin and Mel'nik 1987
Northwestern shelf		Core 53	–10	5	Peat in loess-like loam	КИ-813	45,000		Balandin and Mel'nik 1987

Western shelf		Core 2361	–88	0.85–1.0	Shell hash with Chaudinian fauna	Woods Hole	50,000		Malovitsky *et al.* 1979
Western shelf		Core 2360	–91	1.25–1.4	Shell hash with Chaudinian fauna	Woods Hole	50,000		Malovitsky *et al.* 1979

* Geographic Regions (clockwise):

- Turkish coast, Turkish shelf
- Southwestern shelf
- Western shelf
- Northwestern coast, Northwestern shelf
- Crimean shelf
- Sea of Azov
- Caucasian coast, Caucasian shelf

** Laboratory Index of Organizations Performing Radiocarbon analyses:

КИ	Institute of Geochemistry and Physics of Minerals of the Academy of Sciences of the Ukrainian Soviet Socialist Republic
ИОАН	Institute of Oceanology, Academy of Sciences of the USSR
ETH	Hoenggerberg AMS facility (Eidgenössische Technische Hochschule, Zürich)
ЛУ	Research Geographical-Economic Institute of Leningrad State University
ВСЕГИНГЕО	All-Union Scientific Research Institute of Hydrogeology and Engineering Geology, Ministry of Geology of the USSR
КИГН	Institute of Geological Sciences of the Academy of Sciences of the Ukrainian Soviet Socialist Republic
ТБ	Tbilisi State University
TO	IsoTrace Radiocarbon Laboratory, Accelerator Mass Spectrometry Facility, University of Toronto (Errors attached to the calibrated calender years represent 95% confidence limits, but do not account for analytical errors of the ^{14}C dates.
OS	National Ocean Sciences AMS Facility, Department of Geology and Geophysics, Woods Hole Oceanographic Institution
Woods Hole	Woods Hole Oceanographic Institution (no specimen number provided)

*** All sources appear in the reference list for the paper by Yanko-Hombach in this volume, pp. 183-199.

Appendix 3. Programs of October-November, 2003, conferences.

NATO Advanced Research Workshop (ARW)
"CLIMATE CHANGE AND COASTLINE MIGRATION"
GeoEcoMar, National Institute of Marine Geology and Geo-ecology, Bucharest, Romania
October 1-5, 2003

October 1, 2003 Arrival

October 2, 2003

Morning Session

8:00 Objectives of the NATO ARW
N. Panin, Romania; and **V. Yanko-Hombach**, Canada.

PANEL 1: LATE PLEISTOCENE-HOLOCENE (LAST 20,000 YEARS) MIGRATION OF THE COASTLINE IN THE CIRCUMPONTIC REGION; GEOLOGICAL AND GEOPHYSICAL EVIDENCE

8:30 PLENARY: FLOOD HYPOTHESIS IN THE BLACK SEA: AN OVERVIEW.
W.B.F. Ryan, USA; and **G. Lericolais**, France.

9:15 PLENARY: CLIMATE CHANGE AND COASTLINE MIGRATION IN THE CIRCUMPONTIC REGION: GEOLOGICAL AND PALEONTOLOGICAL RECORDS.
A. Tchepalyga, Russia; and **V. Yanko-Hombach**, Canada.

10:00 PLENARY: NEOTECTONIC DEVELOPMENT OF THE CIRCUMPONTIC REGION: AN OVERVIEW.
Y. Yılmaz, Turkey.

10:45 Questions and discussion

11:15 CLIMATE DYNAMICS, SEA-LEVEL CHANGE, AND COASTLINE MIGRATION IN THE BULGARIAN SECTOR OF THE CIRCUMPONTIC REGION.
M. Georgiev, Bulgaria.

12:00 CLIMATE DYNAMICS, SEA-LEVEL CHANGE, AND COASTLINE MIGRATION IN THE ROMANIAN SECTOR OF THE CIRCUMPONTIC REGION.
N. Panin, Romania.

12:45 Questions and discussion

Afternoon Session

2:00 CLIMATE DYNAMICS, SEA-LEVEL CHANGE, AND COASTLINE MIGRATION IN THE UKRAINIAN SECTOR OF THE CIRCUMPONTIC REGION.
Yu. Shuisky, Ukraine.
2:30 CLIMATE DYNAMICS, SEA-LEVEL CHANGE, AND COASTLINE MIGRATION IN THE RUSSIAN SECTOR OF THE CIRCUMPONTIC REGION.
A. Glebov, Russia.
3:00 Questions and discussion
3:15 MOLLUSCAN ECOLOGY/PALEOECOLOGY IN THE RECONSTRUCTION OF COASTAL CHANGES.
D. Basso, Italy.
3:45 CLIMATE DYNAMICS, SEA-LEVEL CHANGE, AND COASTLINE MIGRATION IN THE TURKISH SECTOR OF THE CIRCUMPONTIC REGION.
O. Algan and **M. Ergin**, Turkey.
4:15 Questions and discussion
5:15 CLIMATE DYNAMICS, SEA-LEVEL CHANGE, AND COASTLINE MIGRATION IN THE GREEK SECTOR (AEGEAN SEA) OF THE CIRCUMPONTIC REGION.
N. Conispoliatis, Greece.
5:45 CLIMATE DYNAMICS, SEA-LEVEL CHANGE, AND COASTLINE MIGRATION IN THE EASTERN MEDITERRANEAN SECTOR OF THE CIRCUMPONTIC REGION.
A. Issar, Israel.
6:15 Questions and discussion
6:30 *Ad Hoc* Committee: Splinter Group: Geology, Neotectonics, and Geophysics of the Circumpontic Region

October 3, 2003

PANEL 2. LATE PLEISTOCENE-HOLOCENE MIGRATION OF THE COASTLINE IN THE CIRCUMPONTIC REGION: PALEONTOLOGICAL AND ARCHEOLOGICAL EVIDENCE

Morning Session

8:00 PLENARY: CLIMATE DYNAMICS AND MIGRATION OF THE COASTLINE AS FACTORS FOR HUMAN DEVELOPMENT IN

THE CIRCUMPONTIC REGION: AN OVERVIEW.
M. Özdoğan, Turkey.

8:45 PLENARY: SUBMERGED SETTLEMENTS IN THE CIRCUMPONTIC REGION: AN OVERVIEW.
H. Angelova, Bulgaria.

9:30 ARCHAEOLOGICAL AND PALEONTOLOGICAL EVIDENCE OF CLIMATE DYNAMICS, SEA-LEVEL CHANGE, AND COASTLINE MIGRATION IN THE BULGARIAN SECTOR OF THE CIRCUMPONTIC REGION.
M. Filipova-Marinova, Bulgaria.

10:00 Questions and discussion

10:30 ARCHAEOLOGICAL AND PALEONTOLOGICAL EVIDENCE OF CLIMATE DYNAMICS, SEA-LEVEL CHANGE, AND COASTLINE MIGRATION IN THE ROMANIAN SECTOR OF THE CIRCUMPONTIC REGION.
P. Alexandrescu, Romania.

11:00 ARCHAEOLOGICAL AND PALEONTOLOGICAL EVIDENCE OF CLIMATE DYNAMICS, SEA-LEVEL CHANGE, AND COASTLINE MIGRATION IN THE UKRANIAN SECTOR OF THE CIRCUMPONTIC REGION.
V. Chabai, Ukraine.

11:30 Questions and discussion

Afternoon Session

1:00 ARCHAEOLOGICAL AND PALEONTOLOGICAL EVIDENCE OF CLIMATE DYNAMICS, SEA-LEVEL CHANGE, AND COASTLINE MIGRATION IN THE RUSSIAN SECTOR OF THE CIRCUMPONTIC REGION.
N. Leonova and **A. Tchepalyga**, Russia.

1:30 ARCHAEOLOGICAL AND PALEONTOLOGICAL EVIDENCE OF CLIMATE DYNAMICS, SEA-LEVEL CHANGE, AND COASTLINE MIGRATION IN THE CAUCASIAN SECTOR OF THE CIRCUMPONTIC REGION.
V. Kuznetsov, Russia.

2:00 Questions and discussion

2:30 ARCHAEOLOGICAL AND PALEONTOLOGICAL EVIDENCE OF CLIMATE DYNAMICS, SEA-LEVEL CHANGE, AND COASTLINE MIGRATION IN THE TURKISH SECTOR (SOUTHERN SHELF OF THE BLACK SEA, THE SEA OF MARMARA, THE BOSPORUS, AND THE DARDANELLES) OF THE CIRCUMPONTIC REGION.

M. Özdoğan and **E. Meriç**, Turkey.
3:00 ARCHAEOLOGICAL AND PALEONTOLOGICAL EVIDENCE OF SEA-LEVEL CHANGE AND COASTLINE MIGRATION IN THE GREEK (AEGEAN SEA) SECTOR OF THE CIRCUMPONTIC REGION.
A. Andrinopoulos, Greece
3:30 Questions and discussion
4:00 ARCHAEOLOGICAL AND PALEONTOLOGICAL EVIDENCE OF SEA-LEVEL CHANGE AND COASTLINE MIGRATION IN THE EASTERN MEDITERRANEAN SECTOR OF THE CIRCUMPONTIC REGION.
A. Ronen, Israel.
4:30 Questions and discussion
4:45 *Ad Hoc* Committee: Splinter Group: Archaeology and Paleontology

October 4, 2003

PANEL 3. MODELING OF CLIMATE DYNAMICS, COASTLINE MIGRATION, AND GEOMORPHOLOGY IN THE CIRCUMPONTIC REGION

Morning Session

8:00 PLENARY: MODELING OF CLIMATE DYNAMICS: AN OVERVIEW.
A. Kislov, Russia.
8:45 PLENARY: MODELING OF COASTLINE MIGRATION AND GEOMORPHOLOGY IN THE CIRCUMPONTIC REGION WITH RESPECT TO CLIMATE DYNAMICS: AN OVERVIEW.
A. Selivanov, Russia.
9:30 Questions and discussion
10:00 POSTERS

1. PALEONTOLOGICAL AND GEOCHEMICAL PROXIES IN RECONSTRUCTION OF PLEISTOCENE-HOLOCENE CLIMATE DYNAMICS.
S. Spezzaferri, Switzerland.
2. ACTIVE TECTONICS IN THE MODIFICATION OF THE QUATERNARY COASTLINES IN THE MARMARA REGION, WESTERN PONTICS.
H. Koral, Turkey.
3. HIGH-RESOLUTION MODEL OF SEDIMENTATION ON THE NORTHWESTERN SHELF OF THE BLACK SEA:

APPLICATION FOR CLIMATE DYNAMICS, COAST-LINE MIGRATION, AND SEA-LEVEL CHANGES.
E. Konikov, Ukraine.

10:30 Round Table: The ARW Project overview (Project co-ordinator V. Yanko-Hombach) and general discussion (all participants).

Afternoon Session

3:00 Field trip to the deltaic sites.

October 5, 2003 Departure

International Conference on "THE BLACK SEA FLOOD: ARCHAEOLOGICAL AND GEOLOGICAL EVIDENCE"
Columbia University Seminar on the Ancient Near East, New York NY
October 17–20, 2003

October 17, 2003 Arrival

October 18, 2003

Morning Session

8:45 INTRODUCTION AND WELCOME
Allan S. Gilbert, Fordham University, Bronx, NY, and Columbia University Seminar on the Ancient Near East

PANEL 1: THE FLOOD; DISCOVERY, DATING, AND ALTERNATIVE PERSPECTIVES

9:00 DISCOVERY OF THE FLOOD.
Walter C. Pitman, III, Lamont-Doherty Earth Observatory, Columbia University, Palisades, NY

9:30 NEW DEVELOPMENTS FROM CONTINUED EXPLORATIONS.
William B.F. Ryan, Lamont-Doherty Earth Observatory, Columbia University, Palisades, NY

10:00 SEISMIC, STRATIGRAPHIC, AND STABLE ISOTOPE EVIDENCE FROM THE BLACK, MARMARA, AND NORTHEASTERN AEGEAN SEAS: ALTERNATIVE INTERPRETATIONS.

Ali E. Aksu, Memorial University, St. John's, Newfoundland, Canada

10:30 GLACIAL TO HOLOCENE BENTHIC FORAMINIFERA IN THE BLACK, MARMARA, AND NORTHEASTERN AEGEAN SEAS.
Michael A. Kaminski, University College, London

11:00 Discussion

Afternoon Session

PANEL 2: ANATOLIAN AND BALKAN COASTAL SITES AND THEIR INTERPRETATION

2:00 THE BLACK SEA FLOOD IN ARCHAEOLOGICAL PERSPECTIVE: PROSPECTS AND QUESTIONS.
Mehmet Özdoğan, Istanbul University

2:30 BLACK SEA COASTAL CHANGES AND UNDERWATER EXCAVATIONS OF THE BULGARIAN LITTORAL.
Hristina Angelova, Center of Underwater Archaeology, Sozopol, Bulgaria

3:00 DENDROCHRONOLOGY OF SUBMERGED BULGARIAN SITES.
Peter I. Kuniholm, Cornell University, Ithaca, NY

3:30 NEW EVIDENCE FOR THE EMERGENCE OF A MARITIME BLACK SEA ECONOMY.
Owen P. Doonan, California State University, Northridge

4:00 HOLOCENE CHANGES IN THE LEVEL OF THE BLACK SEA: CONSEQUENCES AT A HUMAN SCALE.
Douglass W. Bailey, Cardiff University, Wales

4:30 Discussion

October 19, 2003

Morning Session

PANEL 3: BLACK SEA BASIN, HYDROLOGY, AND PROSPECTS FOR UNDERSEA DISCOVERY

9:00 THE LATE QUATERNARY HISTORY OF THE BLACK SEA AND ITS ADJACENT BASINS: A CRITICAL OVERVIEW OF THE FLOOD HYPOTHESES.
Andrei L. Tchepalyga, Russian Academy of Sciences, Moscow; and **Valentina Yanko-Hombach**, Avalon Institute for Applied Science,

Winnipeg, Canada
9:30 MORPHOTECTONIC EVOLUTION OF THE SOUTHERN BLACK SEA REGIONS.
Yücel Yılmaz, Kadir Has University, Istanbul
10:00 THE BLACK SEA: OXIC, SUBOXIC, AND ANOXIC LAYERS.
James W. Murray, University of Washington, Seattle
10:30 FROM MOUNTAIN-TOP TO OCEAN BOTTOM: COMPREHENSIVE ARCHAEOLOGICAL RESEARCH ALONG THE BLACK SEA COAST.
Fredrik T. Hiebert, University of Pennsylvania, Philadelphia
11:00 Discussion

Afternoon Session

PANEL 4: UNDERSEA DISCOVERIES AND CULTURAL DISTRIBUTIONS TO THE NORTH AND EAST

2:00 RECENT DEEPSEA SURVEY RESULTS IN THE BLACK SEA.
Robert D. Ballard and **Dwight F. Coleman**, Institute for Exploration, Mystic, CT, and University of Rhode Island
2:30 STEPPE POPULATIONS NORTH AND EAST OF THE BLACK SEA BETWEEN 6200-5500 calBC: THE FLOOD AND THE TRANSITION TO FOOD PRODUCTION.
David W. Anthony, Hartwick College, Oneonta, NY
3:00 THE EARLY NEOLITHIC CULTURES OF EASTERN EUROPE.
Valentin Dergachev, Moldavian Academy of Sciences, Chinisau
3:30 THE MEDITERRANIZATION OF THE SOUTHERN CRIMEAN FLORA DURING THE HOLOCENE.
Carlos E. Cordova, Oklahoma State University, Stillwater
4:00 ORIGINS OF EURASIAN LANGUAGE FAMILIES IN RELATION TO THE BLACK SEA FLOODING.
Johanna Nichols, University of California, Berkeley
4:30 Discussion

October 20, 2003

Morning Session

9:00 CONFERENCE SUMMARY AND OVERVIEW.
Valentina Yanko-Hombach, Avalon Institute for Applied Science, Winnipeg, Canada

Geological Society of America Annual Meeting
T104. “NOAH'S FLOOD” AND THE LATE QUATERNARY GEOLOGICAL AND ARCHAEOLOGICAL HISTORY OF THE BLACK SEA AND ADJACENT BASINS
Washington State Convention and Trade Center, Seattle
November 2-5, 2003

November 4, 2003

Session No. 178 Valentina Yanko-Hombach and Jim Teller, Presiding

1:30 EVIDENCE FOR A BLACK SEA FLOODING EVENT.
William B.F. Ryan, Lamont Doherty Earth Observatory, Columbia University, Palisades, NY; **Namik Çağatay**, Geology Department, ITU Maden Fakultesi, Ayazaga, Istanbul, Turkey; **Candace O. Major**, Laboratoire des Sciences du Climat et de l'Environnement, Gif-sur-Yvette, France; and **Gilles Lericolais**, DRO-Geosciences Marines, IFREMER, Laboratoire Environnements Sedimentaires, Plouzané, France.

1:45 LATE GLACIAL GREAT FLOOD IN THE BLACK SEA AND CASPIAN SEA.
Andrei Chepalyga, Institute of Geography, Russian Academy of Science, Moscow, Russia.

2:00 CONFLUENCE OF CLIMATE CHANGE AND CULTURAL COMPLEXITY IN SOUTHERN MESOPOTAMIA: IMPLICATIONS FOR BIBLICAL FLOOD MYTHOLOGY.
James P. Kennett, Department of Geological Sciences and Marine Science Institute, Univ of California Santa Barbara, Santa Barbara, CA; and **Douglas J. Kennett**, Department of Anthropology, Univ of Oregon, Eugene, OR.

2:15 MORPHOTECTONIC DEVELOPMENT OF THE SOUTHERN BLACK SEA REGION AND THE SURROUNDINGS.
Yücel Yılmaz, Kadir Has Univ, Cibali, Istanbul, Turkey.

2:30 "NOAH'S FLOOD" AND THE LATE QUATERNARY HISTORY OF THE BLACK SEA AND ITS ADJACENT BASINS: A CRITICAL OVERVIEW OF THE FLOOD HYPOTHESES.
Valentina Yanko-Hombach, Avalon Institute of Applied Science, Winnipeg, Canada.

2:45 Paper Withdrawn

3:00 EARLY HOLOCENE SEA LEVEL CURVE OF THE COAST OF ISRAEL, EAST MEDITERRANEAN.
Dorit Sivan, Institute for Maritime Studies, Univ of Haifa, Israel;

and **Kurt Lambeck**, Research School of Earth Sciences, The Australian National Univ, Canberra, Australia.

3:15 THE CONNECTIONS BETWEEN THE BLACK SEA AND MEDITERRANEAN DURING THE LAST 30 KY.
Oya Algan, Institute of Marine Sciences and Mgmt, Vefa, Istanbul, Turkey.

3:30 THE BLACK SEA DURING THE LAST 20.000 YEARS: SEA LEVEL, SALINITY AND CLIMATE.
Anton, Preisinger and Selma Aslanian, Mineralogy, Technical Univ of Vienna, Vienna, Austria.

3:45 CLIMATIC CHANGES MODIFIED THE QUATERNARY COASTAL LINES IN THE MARMARA REGION, WESTERN PONTICS: WHAT ABOUT ACTIVE TECTONICS?
Hayrettin Koral, Department of Geological Engineering, Istanbul University, Avcılar, Istanbul, Turkey.

4:00 THE BLACK SEA FLOOD: ARCHAEOLOGICAL & GEOLOGICAL EVIDENCE; A SUMMARY OF THE COLUMBIA UNIVERSITY CONFERENCE.
Allan S. Gilbert, Department of Sociology & Anthropology, Fordham University, Bronx, NY.

4:15 WATER EXCHANGE BETWEEN MEDITERRANEAN AND BLACK SEAS DURING LATE GLACIAL-HOLOCENE: EVIDENCE FROM MARMARA AND BLACK SEAS.
M. Namik Çağatay and **Naci Görür**, Geological Engineering Department, Faculty of Mining and Eurasian Institute of Earth Sciences, Ayazaga, Istanbul, Turkey.

4:30 HOLOCENE BLACK SEA ENVIRONMENTS ACCORDING TO PALYNOLOGY.
Sr Speranta-Maria Popescu, PaléoEnvironnements et PaléobioSphère, Univ Claude Bernard-Lyon 1, Villeurbanne, France; **Martin V. Head**, Geography, Univ of Cambridge, Cambridge, UK; and **Gilles Lericolais**, DRO Géosciences Marines, IFREMER, Plouzané, France.

4:45 WAS THE LAST RAPID SEA CHANGE IN THE BLACK SEA LINKED TO A CATASTROPHIC EVENT RECORDED BY MANKIND ?
Gilles Lericolais, DRO-Geosciences Marines, IFREMER, Laboratoire Environnements Sedimentaires, Plouzané, France; **Irina Popescu**, RCMG, Univ of Gent, Gent, Belgium; **Nicolae Panin**, GEOECOMAR, Bucharest, Romania; **François Guichard**, LSCE, CNRS-CEA, Gif-sur-Yvette, France; and **Sr Speranta-Maria**

Popescu, Centre de Paléontologie stratigraphique et Paléoécologie, Université Claude Bernard-Lyon 1, Villeurbanne, France.

5:00 STABLE AND RADIOGENIC ISOTOPE CONSTRAINTS ON THE DEGLACIAL HISTORY OF THE BLACK AND MARMARA SEAS.

Candace O. Major,[1], **William B.F. Ryan**,[2], **Steven L. Goldstein**,[2], **Laurent Labeyrie**,[1], and **Namik Çağatay**,[3] (1) Laboratoire des Sciences du Climat et de l'Environnement, CNRS, Gif-sur-Yvette, France; (2) Lamont Doherty Earth Observatory, Palisades, NY (3) Geology Department, ITU Maden Fakültesi, Ayazaga, Istanbul.

5:15 LATE GLACIAL FRESHENING OF THE LEVANTINE BASIN: ISOTOPIC EVIDENCE FROM ISKENDERUN BAY.

Silvia Spezzaferri, Geology and Paleontology, Univ of Fribourg, Fribourg, Switzerland; **Valentina Yanko-Hombach**, Avalon Institute of Applied Science, Winnipeg, Canada; and **Sebastien Bruchez**, Geology and Paleontology, Univ of Lausanne, Lausanne, Switzerland.

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