REVIEW PAPER

ENVIRONMENTAL VARIABILITY OF THE PONTO-CASPIAN AND MEDITERRANEAN BASINS DURING THE LAST CLIMATIC MACROCYCLE

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ABSTRACT. This paper reviews reconstructions of the the evolution of the Ponto-Caspian basin system to certain parts of the Pontian-Mediterranean system in order to analyze their correlation and response of the systems to the global climate change. The Ponto–Caspian and Mediterranean basins belong to different types of water basins and evolved differently in the Late Pleistocene responding in different ways to the global climate change. The paleogeographic reconstructions and correlation analysis of the Late Pleistocene events (within the last climatic macrocycle) made it possible to view the evolution of the basins as parts of a single system allowing to identify certain specific features and patterns in their functioning. The study is based on the analysis and integration of the data published by numerous researchers including the author of the paper and numerous colleagues from many countries who have been studying the paleogeography of the Ponto-Caspian and Mediterranean regions in the Late Pleistocene.

KEY WORDS: Caspian Sea, Sea of Azov, Black Sea, Marmara Sea, Eastern Mediterranean, ancient passages, late Pleistocene, sea level change, climate change, paleoenvironmental reconstruction, correlation

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INTRODUCTION

The Ponto-Caspian and Mediterranean basins represent a system of intracontinental water bodies, relicts of the Paratethys sea basin, different in their natural characteristics and paleogeographic evolution. The Ponto-Caspian part of the system includes an isolated basin of the Caspian Sea, the Azov–Black Sea basin, which at certain periods connects with the ocean, and the Manych Depression, which occasionally functions as a strait between the Caspian and the Pontian basins. The Mediterranean part of the system is composed of the eastern part of the Mediterranean Sea, permanently connected to the ocean; the Sea of Marmara, which forms a kind of a «gate» between the Black and Mediterranean seas and is at certain periods isolated from the adjacent sea basins, and the Bosporus and Dardanelles straits (Fig. 1).

Fig. 1. Ponto-Caspian and Mediterranean basins

The evolution of the described natural system is influenced by multiple factors. This review aims to reveal connections between the global and regional climate changes, as well as between sea level fluctuations in the Caspian Sea, Black and Azov seas and the basins of the Eastern Mediterranean and the evolution of their environments.

The paleogeographic analysis focuses on the last climatic macrocycle, which corresponds to the time interval from the last interglacial period (MIS 5e) to the present day (MIS 1), thus spanning the entire Late Pleistocene. The interval covers several global climatic events different in their magnitude and impact, including glacial and interglacial periods, their individual stages and phases in their development. The natural systems of the Caspian and Pontian sea basins and those of the Mediterranean significantly differ in their characteristics. As a result, their response to the changes in paleoclimate, correlation of their transgression and regression phases, periodicity of the connection between the basins, the exchange of water and the biological diversity development – all provide the historic basis for predictive estimates of the environmental conditions in the region under the climate change.

The history of the above listed basins and their environments in the Late Pleistocene has been studied for a long time (more than three centuries). The studies in the Caspian region were started by Pallas (1776), Eichwald (1824), continued by Mushketov (1895), Andrusov (1888, 1900), Bogachev (1903), Pravoslavlev (1908, 1926), and many others. Results of multidisciplinary investigations were summarized in a number of monographs (Fedorov 1957, 1978; Vasilyev 1961; Moskvitin 1962; Svitoch, Yanina 1997; Rychagov 1997; Yanina 2005, 2012; Svitoch 2014). The first paleogeographic reconstructions of the Black Sea basins were performed by Andrusov (1889, 1890, 1925, a.o.) and developed further by Pavlov (1925), Gubkin (1913), Arkhangelskiy and Strakhov (1938), and many others. The results of integrated studies are presented in monographs by Fedorov (1963), Nevesskaya (1965), Popov (1983), Mikhailesku (1990), Svitoch et al. (1998), Izmailov (2005). Many specialists studied the evolution of environments of the Mediterranean basins (Lamothe 1899; Gignoux 1913; Issel 1914; Blanc 1937; Shimkus 1981; Bruckner 1986; Keraudren and Sorel 1987; Castradori 1993; Cita et al. 1973; Çagatay et al. 2000, 2009; Kaminski et al. 2002; Mudie et al. 2002; Cecilia et al. 2008; Wegwertha et al. 2014; Büyükmeriç et al. 2016; Krijgsman et al. 2019; Casini et al. 2020 and many others).

Hundreds of papers concerning the aspects of the environmental development in individual basins and the region as a whole have been published by now. And yet most of the paleogeographic problems in every region are still debatable. Among them is the topic of correlation between the events within the Ponto-Caspian basin system and the individual parts of the Pontian-Mediterranean system, as well as the response of both systems to the changes in the global and regional climate. Also, the question regarding the correlation between the paleogeographic events within the region and on the adjacent territories has not been done answered yet. In the present paper, the author addresses the above-stated problems and proposes a possible answer to some of the discussed questions.

The study is based on the analysis and integration of the data published by numerous researchers including the author of the paper and numerous colleagues from many countries who have been studying the paleogeography of the Ponto-Caspian and Mediterranean regions in the Late Pleistocene. The recent decades are marked by a sharp increase in the amount of such research, which indicates a growing interest of the global scientific community in the history of those intracontinental basins. That interest may be attributed to a considerable role the basins have played in the past evolution of the continent environment and still play today.

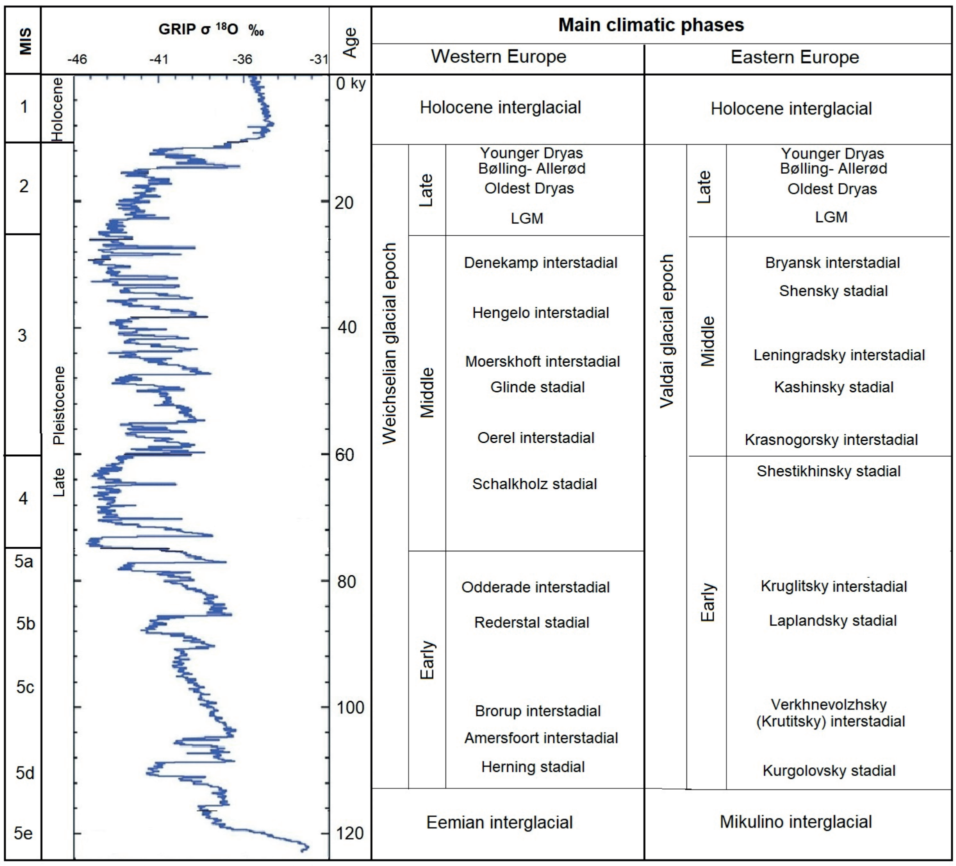

Fig. 2. The last climatic macrocycle and the climatostratigraphic scheme of Europe (Reprinted from Novenko 2016)

THE LAST CLIMATIC MACROCYCLE

Eemian (Mikulino) interglacial

The last interglacial period is thoroughly studied and described in scientific publications. In the climatostratigraphic scheme of European Russia it is denoted as Mikulino Interglacial (Fig. 2) (Dynamics… 2002; Shik 2014; Novenko 2016), and in the stratigraphic schemes of Western and Central Europe – as Eemian Interglacial (Kukla et al. 2002; Litt and Gibbard 2008; Brewer et al. 2008).

The position of the interglacial in the geochronological scheme and its time period is an important and still debatable question in the Pleistocene paleogeography. At present, the generally accepted approach to estimating the interglacial duration and its phases is based on the correlation of the oxygen isotope data in the deep-sea sediments and in the ice cores, which is an indication of the global climate changes (marine isotope stages – MIS) (Kukla et al. 2002; Berger et al. 1981; Imbrie et al. 1984). The majority of specialists agree that the Eemian (Mikulino) Interglacial corresponds to MIS 5e (Shackleton 1969). Its duration is estimated at 13 ka (128 to 115 ka BP) with the climatic optimum falling on ~125 ka BP. That geochronological position of the interglacial has been accepted by the International Commission on Stratigraphy (Litt and Gibbard 2008; Head 2020). Numerous dating results obtained using thorium and uranium technique (Kuznetsov et al. 2002; Gaigalas et al. 2005) and the OSL procedure (Degering and Krbetschek 2007; Boettger et al. 2009; Mania et al. 2010; Kurbanov et al. 2019) do not contradict the attribution of the last interglacial to that interval.

The ice cores obtained from deep drilling of the Greenland and Antarctic ice sheets present a natural archive, which can be used to study the natural environmental dynamics. As an example, the oxygen isotope data obtained from deep drilling in Greenland, conducted within the NGRIP (NorthGRIP Project members 2004) and NEEM (North Greenland Eemian Ice Drilling (NEEM) international project, 2007–2012) projects, provided a continuous record of the climate fluctuations over the last 123 and 130 ka, which allowed to estimate the chronological boundaries of the interglacial at 130-115 ka BP, with its thermal maximum falling on ~126 ka BP (NEEM Project members 2013; Turney et al. 2010).

Studies of marine deposits indicate a fast rise of the sea level at the beginning of the interglacial (Boreal transgression) related to the melting of the ice sheet (Zagwijn 1983, 1996; Forsström 2001; Kopp et al. 2010). The global sea level reached its current position around 127 ka BP, and exceeded it by 6-7.5 m (Kopp et al. 2010) or even by 7-9 m (Dutton and Lambeck 2012) at the peak of the transgression. Lowering of the sea level started around 116-118 ka BP (Kopp et al. 2013). The data on the dripstone calcite and stalagmites in the caves of southern Europe and the eastern Mediterranean indicate considerable warming that occurred around 129.7 – 125.8 ka BP (Drysdale et al. 2005; Wainer et al. 2011). The results of high-resolution analysis of the oxygen isotope composition and geochronological studies of Entrische Kirche cave (the Austrian Alps) allowed to date the upper boundary of the interglacial at approximately 118 ka BP (Meyer et al. 2008).

The age of the interglacial upper boundary (and, therefore, its duration) is still under discussion (Helmens 2014). Some authors include in the Eemian Interglacial not only MIS 5e but also 5d (Kukla et al. 2002; Brauer et al. 2007); others (Molodkov and Bolikhovskaya 2002, 2006) attribute the entire MIS 5 interval to the Mikulino Interglacial and distinguish periods with cold snaps within it. The interglacial duration is estimated at 75 ka. New data obtained every year is abundant, though not conclusive. Kukla et al. (1997) proposed to distinguish between the Eemian Interglacial *sensu stricto* (s.s.) – a warm period, identified in the deposits from Western Europe – and the Eemian Interglacial *sensu lato* (s.l.) – a period, when a thermophilic forest was present in southwestern and southern Europe.

Weichselian (Valday) Ice Age

Many specialists that consider the Eemian (Mikulino) Interglacial to correspond to MIS 5e attribute the next interval MIS 5d-a (complicated in structure regarding its climate) to the Vistulian (Valday) Ice Age. According to the stratigraphic schemes conventionally used for European Russia, the Late Pleistocene includes the Early Valday glaciation (MIS 5d-a and MIS 4), Middle Valday megainterstadial (MIS 3), and the Late Valday glaciation (MIS 2) (Fig. 2). MIS 5d-a interval includes the Kurgolovo cooling (5d stage), the Upper Volga (Krutitsa) interstadial (5c), Lapland cooling (5b), and Kruglitsa interstadial (5a) (Dynamics of... 2002; Paleoclimates… 2009). In Eastern and Central Europe several stages are identified within the Weichselian (Vistulian) Ice Age, namely, the early (MIS 5d-a), middle (MIS 4 – MIS 3) and late (MIS 2) stages (Mangerud 1989; Litt et al. 2007), while periods of warming and cooling are also distinguished within the stages.

The Late Pleistocene period corresponding to MIS 5 d-a lasted for 40 ka (115–75 ka BP). There is an opinion (Lavrushin et al. 2002; Shik 2014) that this interval should be considered as a separate stage under the name of, for example, Eovalday. In the author's opinion, it should be considered as a transitional (from interglacial to glacial) epoch. A deep and prolonged cooling corresponding to MIS 4 stage (75-60 ka BP) represents the Kalinin stage of the Valday glaciation. The Middle Valday epoch (MIS 3) is confined between 60 and 25 ka BP and includes a series of periods with relative cooling and warming; the period as a whole is characterized by a general decrease in the climate continentality (Paleoclimates… 2009). The late Valday (Ostashkov) glacial stage corresponds to MIS 2 (25-11.7 ka BP). The maximum cooling during the Valday glacial period is dated at 22 to 18 ka BP. At that time the entire boreal region of Europe not covered by the ice sheet represented a single hyperzone (Dynamics of… 2002) where the landscapes developed under a heavy influence of the cryogenesis. According to Rinterknecht et al. (2018), «local LGM» in the central part of the East European Plain is dated at 20 to 20.5 ka BP, while the deglaciation began around 17–15 ka BP.

The Late Glacial (14.7–11.7 ka BP) was characterized by short-term climate fluctuations. The time interval included distinct warming with two ineterstadials – Bølling (14.7–14.0 ka BP) and Allerød (13.6–12.9 ka BP), separated by a colder period known as Older Dryas (Walker et al. 2009; Merkt and Müller 1999; Dzieduszyńska and Forysiak 2019). A considerable cooling – Younger Dryas – occurred during 12.9–11.7 ka BP (Walker et al. 2009; NorthGRIP Project members 2004; EPICA community members 2004). The beginning of distinct warming immediately after the cold period is considered to be the lower boundary of the Holocene (Head 2019).

PONTO-CASPIAN AND MEDITERRANEAN SEA BASINS IN THE LATE PLEISTOCENE

Caspian Sea

The scheme of the Late Pleistocene events on the Caspian Sea includes the late Khazarian (late Khazarian and Hyrcanian transgressions) and Khvalynian (early and late Khvalynian) transgressive epochs separated by the

Atelian regression. Each of the above-named events was complicated by transgressive and regressive phases and oscillations, resulting from climate fluctuations and widely varying in their magnitude and direction.

Late Khazarian transgressive epoch

There are two transgressions (stages) distinguished within the late Khazarian epoch – the late Khazarian and the Hyrcanian. The level of late Khazarian transgressive basin reached about minus 10 meters at its maximum (Svitoch and Yanina 1997; Rychagov 1997; Yanina 2012). The sea expansion and its coast were described in a number of works (Leontyev et al. 1977; Rychagov 1977). The water was rather warm, as suggested by the composition of the mollusk assemblage dominated by crassoid Didacna (typically *D. nalivkini* and *D. surachanica*), which is characterized by large size and thick valves. The assumption is corroborated by the abundance of *Corbicula fluminalis* in the freshened water of the Northern Caspian at that time; currently, that species is encountered only in the south of the Caspian region (Yanina 2005; Bezrodnykh et al. 2015). The salinity was higher than at present – from 10–12‰ in the Northern Caspian to 14–15‰ in the Southern basin (Yanina 2012; Svitoch 2014). The pollen assemblages indicate a warm and dry climate (Abramova 1972; Yakhimovich et al. 1986). A distinctive feature of the sea was a predominance of depositional coasts (Leontyev et al. 1977) and a large volume of accumulative sand bodies, which might be an indicator of a prolonged stay of the sea at the same level, with only insignificant fluctuations.

The late Khazarian transgression was followed by sea level lowering. This is suggested by the gaps in marine sedimentation processes, that are distinguished in the sequences exposed in the coastal scarps (Fedorov 1957; Popov 1983; Rychagov 1997; Svitoch and Yanina 1997; Yanina 2005, 2012) and in the sedimentary series found in the Northern Caspian (Yanina et al. 2014; Bezrodnykh et al. 2015). At present, there is no data indicative of the scale of the sea level lowering.

The issue of the Hyrcanian transgressive sea basin has been a widely debated topic for many years (Yanina et al. 2014). The Hyrcanian transgressive stage was identified in the Caspian history by Goretsky (1957) and Popov (1967) based on the analysis of boreholes drilled in the northwest of the Caspian Lowland and the Vostochny (Eastern) Manych valley. Their position was subjected to a harsh criticism by many specialists (Vasilyev and Fedorov 1965; Fedorov 1978; Shkatova 2010; Svitoch et al. 1998), most of which rejected the idea of the Hyrcanian stage. Recently, however, the materials obtained from drilling in the Northern Caspian during the oil exploration allowed to return to that problem (Yanina et al. 2014; Sorokin et al. 2017). The drilling of the Upper Pleistocene series revealed the Caspian marine deposits corresponding to the Hyrcanian transgressive basin. The typical feature of its fauna is the joint occurrence of «Khvalynian-like» fauna (*Didacna subcatillus, D. cristata*) and rare late Khazarian mollusks. The sea basin was freshened and exceeded the late Khazarian basin in size. The pollen assemblages suggest a somewhat cooler and wetter climate (Yanina et al. 2014).

The late Khazarian transgressive epoch is attributed to the beginning of the Late Pleistocene. As has been shown by uranium series dating, the Late Khazarian transgressive stage corresponds to 127–122 ka BP (Shkatova 2010), while the entire Late Khazarian epoch is dated at 127–76 ka BP (Rychagov 1997; Shkatova 2010). Dating by the electron spin resonance technique (ESR) allowed to date the stage to the period from 140 to 85 ka BP (Molodkov and Bolikhovskaya 2006). The continental deposits exposed in the Srednyaya Akhtuba section in the lower reaches of the Volga correspond to the late Khazarian and Hyrcanian stages in the Caspian Sea evolution. Their age determined by the OSL (optically stimulated luminescence) technique corresponds to the entire MIS 5 stage (Yanina et al. 2017).

Atelian regression

The end of the Khazarian stage in the Caspian Sea evolution was marked by a deep regression. The estimates of the regression amplitude vary widely – from -43 m (Badyukova 2016), and -53 m (Leontyev et al. 1977), to -100 m (Bezrodnykh et al. 2017) and -140 м (Lokhin and Mayev 1990).

The estimates of the age and climate of the regressive stage also vary over a wide range. The deposits of the stage were first described by Pravoslavlev (1908, 1926), who attributed them to hot desert environments. A number of his colleagues – geologists Mazarovich (1927), Nikolayev (1941), Zhukov (1936) – shared the same opinion. Pavlov (1925) dated the Atelian continental deposits to the «beginning and culminating epoch of the Wurmian glaciation». That point of view was supported by Moskvitin (1962). Further on, the view of the Atelian epoch as a cold (glacial) interval was shared by the majority of researchers, though their opinions differed widely as to which particular cold interval it was. Vasilyev (1961) related the regression to the time of the Dnieper glacial period – Mikulino interglacial. In the opinion of Chepalyga (2004) and Lavrushin (Lavrushin et al. 2014) the regression coincided with the maximum of the last glaciation (Late Valday, Ostashkov) on the East European Plain. Some specialists (Fedorov 1978; Yakhimovich et al. 1986; Yanina 2012, 2014) correlate the Atelian regression with the Kalinin glacial epoch. There is also an opinion (Svitoch 2014; Svitoch et al… 1998) that the regression lasted longer and continued from the Kalinin glacial maximum to that of Ostashkov.

The Atelian continental deposits are widely spread in the north of the Caspian Lowland; quite often they form wedge-like structures penetrating deeply into underlying layers. Those wedges and frost fissures indicate severe climate conditions and occurrence of permafrost. Shells of freshwater and terrestrial mollusks found occasionally in the deposits are characterized by an oppressed appearance. Bone remains of mammals belonging to the «mammoth assemblage» indicate a cold climate of the Atelian time interval. That is supported by tundra-steppe pollen assemblages recovered from the Atelian deposits (Grichuk 1954; Moskvitin 1962).

In the depositional series of the Northern Caspian, the Atelian interval is represented by paleodepressions and erosional landforms, which are distinctly visible in the seismic stratigraphic profiles (Bezrodnykh et al. 2017). As appears from the composition of organic remains, the period was characterized by the presence of lakes with fresh or brackish water and wetlands. The pollen assemblages are indicative of rather diverse landscapes north of the Caspian Sea – from forests dominated by conifers to periglacial forest-steppe and tundra-forest-steppe (Bolikhovskaya et al. 2017).

The Atelian regression was dated by the TL method to the period 80–28 ka BP (Shakhovets and Shlyukov 1987). The age of its final phase was determined by the OSL method at 48.68±3.10 ka BP (Yanina et al. 2017); a few radiocarbon dating results fall into the interval of 44.40–41.80 ka BP, which corresponds to the first half of the intra-Valday interstadial warming. Well-developed cryogenic wedges found at the base of Atelian series deeply penetrating into the underlying deposits (dated to MIS 5) suggest that accumulation of the subaerial Atelian deposits began during the cold (glacial) epoch MIS 4 and the sea regression occurred in MIS 4 – the 1st half of MIS 3.

Khvalynian transgressive epoch

The Atelian regression was followed by a «great» Khvalynian transgression when the Caspian Sea reached its highest level over the entire Neopleistocene history. The Khvalynian basin left its traces on all the coasts and has been described in sufficient details (Leontyev et al. 1977; Rychagov 1997). Almost all the specialists agree that the Khvalynian transgression proceeded in two stages – the early and late Khvalynian, which were separated by Yenotayevka regression (Fedorov 1957, 1978; Rychagov 1977, and others). The sea level during the early Khvalynian transgressive stage reached 48–50 m a.s.l. at its maximum. The Lower Volga valley was enclosed by an extended estuary about 500 km long, the sediments of which are exposed in most of the sections (Fedorov 1957; Vasilyev 1961; Moskvitin 1962; Svitoch and Yanina 1997). One of the typical facies is the so-called «chocolate clays» – a distinctive kind of the Khvalynian basin deposits of the Caspian Sea (Makshaev and Svitoch 2016).

Traces of the ancient coastlines left by different stages of the early Khvalynian transgression are present in a form of erosional and depositional terraces on the coasts of Middle and Southern Caspian; in the Northern Caspian there are traces of incised deltas and other coastal landforms. The most distinct terraces are present at 34–36 m (Talginsksya terrace, Rychagov 1970), 20–22 m (Buynakskaya terrace, Fedorov 1956), 14–15 m (Turkmenian terrace, Fedorov 1957). The terrace development was probably related to the transgressive stages of the sea alternating with regressions (Rychagov 1970; Chepalyga 2006); others attribute the coastal landform development to temporary delays in the sea retreat during regressive phases (Fedorov 1957; Vasilyev 1961). A number of researchers (Britsyna 1954; Arkhipov 1958; Vasilyev 1961) studied the marine sediments in the Northern Caspian and came to a conclusion about two early Khvalynian transgressive phases separated with a regression (Eltonian regression, by Vasilyev (1961). This opinion is shared by Chepalyga (Chepalyga 2006).

The fauna inhabiting the sea basin was relatively poor and represented mostly by *Didacna parallella*, *D. protracta*, and *D. ebersini*, though crassoid didacnas were completely absent. The mollusks are characterized by thin shells, often small in size. The salinity is estimated at 11–12‰ at the main water body of the Early Khvalynian Caspian Sea and at 5–0.5‰ at the estuary (Kvasov 1975; Yanina 2012; Svitoch 2014). The small size of mollusks suggests rather low temperatures of water, which is confirmed by the pollen assemblages. For example, those recovered from the chocolate clays (in the lower reaches of the Volga) indicate the presence of periglacial landscapes (tundra-steppe, periglacial forest-steppe, periglacial open forests and parklands) (Bolikhovskaya and Makshayev 2020).

The end of the Early Khvalynian period was marked by the Yenotayevka regression with the sea level dropping, according to different estimates, to -43…-45 m (Rychagov 1997), -84 m (Varushchenko et al. 1987) or even to -110 m below sea level (Lokhin and Mayev 1990). Continental deposits occur in the Khvalynian sedimentary sequences (Brotskiy and Karandeeva 1953), though most often the regression may be traced by erosional landforms (Svitoch and Yanina 1997; Rychagov 1997). According to pollen data (Sorokin et al. 1983), the climate was dry and cool.

The sea level during the late Khvalynian transgression reached about 0 m at its maximum (Leontyev et al. 1977; Rychagov 1997). The Volga River divided into two channels near its mouth. The composition of late Khvalynian mollusk fauna was similar to that of the early Khvalynian; the only exception was the dominance of *D. praetrigonoides*, which occupied only insignificant biotopes in the early Khvalynian basin. The mollusk fauna composition suggests the salinity of about 10-12‰ (Yanina 2005). The abundance of mollusks with larger and more massive shells may be a result of a higher water temperature than in the early Khvalynian. Palynological data also indicates general warming in the region (Abramova 1972; Vronskiy 1976; Yakhimovich et al. 1986).

The lowering of the late Khvalynian sea level proceeded irregularly; periods of a slowdown are marked by the coastlines detected at altitudes of -10 to -12 m (Sartass stage) and -16 to -18 m (Dagestan stage) (Leontyev and Fedorov 1953). The time of the Dagestan stage proved to correspond to the Holocene epoch (New Caspian basin) (Rychagov 1997). New data obtained by Svitoch (2011) confirmed it to be an independent Holocene transgression, which was followed by regression as the climate became more arid (Abramova 1972).

The age of the Khvalynian transgression, as well as that of its individual stages, is still debatable. The early Khvalynian stage was dated by the TL to the period 70-40 ka BP, and the late Khvalynian – to 20-10 ka BP (Rychagov 1997). According to 14C and 230Th/234U data, the Khvalynian transgression may be dated to 19-8 ka (Kvasov 1975; Svitoch and Yanina 1997; Leonov et al. 2002; Tudryn et al. 2013; Arslanov et al. 2016). Values close to the radiocarbon (14C) ones were obtained using the OSL technique (Yanina et al. 2017).

Another transgressive basin has been recently uncovered by drilling in the Northern Caspian, which is the earliest Khvalynian basin (Bezrodnykh et al. 2014, 2015; Yanina et al. 2017; Sorokin et al. 2018), with mollusk fauna including *Didacna subcatillus, D. zhukovi, D. parallella*. Judging from the mollusk habitus, the sea was moderately warm. From numerous radiocarbon dates the period of its existence is estimated to span from 37 ka BP (the 2nd half of the MIS 3 interstadial) to LGM (MIS 2) (Bezrodnykh et al. 2014, 2015). That transgressive series is covered by a regressive layer dated by radiocarbon to 22–20 ka BP and corresponding to LGM. The overlying sedimentary series (repeatedly dated by radiocarbon) preserves traces of several transgressive and regressive phases of the Khvalynian basin evolution chronologically corresponding to the main phases of the degradation of the late Valday glaciation (Yanina et al. 2017).

It may be concluded from the above that the Late Pleistocene paleogeographic events in the Caspian region are closely related to the changes in the global climate (Fig. 3). The late Khazarian «little» transgression developed during the Mikulino (Eemian) Interglacial (MIS 5e). The Hyrcanian transgression that was characterized by a higher sea level developed during the transitional stage from the interglacial towards the Valday glaciation. As the maximum of the early Valday (Kalinin glaciation) approached, the Hyrcanian basin retreated under the conditions of cold and dry climate. The Atelian regression corresponded to the Kalinin glacial epoch and the first stage of the interstadial warming (MIS 3). At the 2nd half of the interstadial, the early stage of the Khvalynian transgression started. The transgressive changes of the sea level were interrupted by a regression during the period of the maximum cooling and drying of the late Valday (MIS 2, LGM).

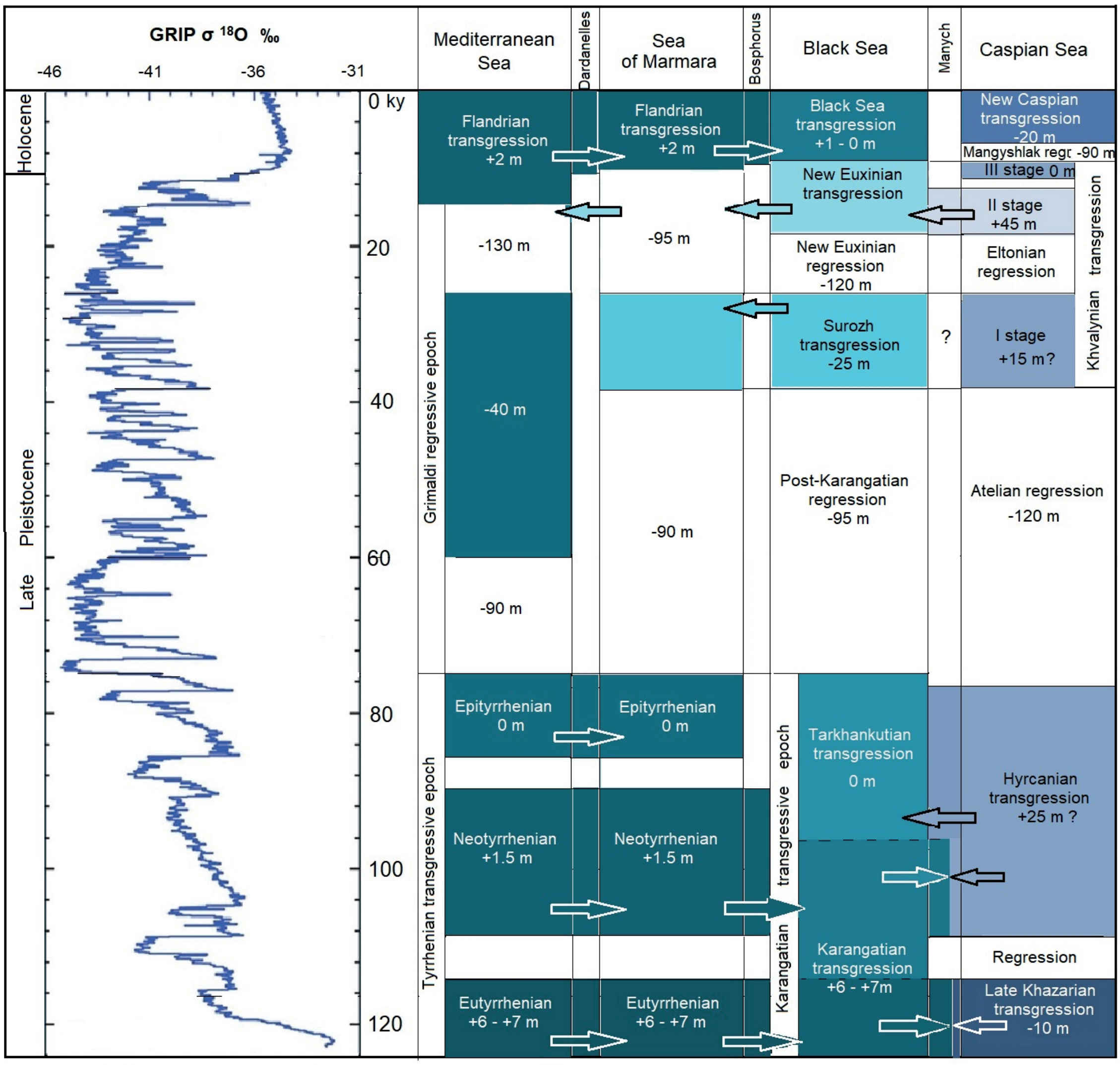

Fig. 3. Ponto-Caspian and Mediterranean basins during the late Pleistocene and their correlation More saturated shade of color shows higher salinity of the basin. Arrows indicate the water inflow and migration of fauna

The transgression resumed during the period of deglaciation. The most notable events of the Late Glacial – warming phases of Bølling and Allerød – activated the melting of the ice sheet and degradation of the permafrost within the drainage basin; those processes contributed to the further development of the Khvalynian basin transgression. That period was marked by the «chocolate clay» accumulation in the Volga estuary and in the pre-Khvalynian depressions in the Northern Caspian. The phases of considerable cooling known as Oldest Dryas, Older Dryas, and Younger Dryas resulted in a reduced runoff from the drainage area and regression of the Caspian Sea. The most significant regression corresponded to the Younger Dryas. The final phase of the Khvalynian transgression corresponds to the abrupt warming, which is considered as the Pleistocene / Holocene boundary. The decrease in sea level ended with the Mangyshlak regression, which developed under the conditions of a continental climate in the Boreal period of the Holocene.

Pont

The Pontian basin (including the Sea of Azov and the Black Sea) occupied an intermediate position between the Caspian and Mediterranean seas and therefore was influenced by both. Its Late Pleistocene history was marked by an alternation of marine and brackish-water basins. In the Late Pleistocene, the only marine basin was Karangatian which developed as a result of the Mediterranean water inflow. The next marine transgression occurred in the Holocene. The post-Karangatian, Surozh and New Euxinian basins were also characterized by brackish water.

Karangatian transgressive epoch

The fore-Karangatian drop of the sea level gave way to the Karangatian transgression at the beginning of the Late Pleistocene. Its deposits are widespread and the basin paleogeography has been studied in details (Andrusov 1904, 1925; Arkhangelskiy and Strakhov 1938; Nevesskaya 1965; Dimitrov and Govberg 1979; Božilova and Djankova 1976; Zubakov et al. 1982; Krystev et al. 1990; Markova and Mikhailesku 1990; Yanko et al. 1990; Nesmeyanov and Izmailov 1995; Svitoch et al. 1998; Dodonov et al. 2006; Sorokin 2011; Filipova-Marinova et al. 2012; Shumilovskikh et al. 2013a.b; Wegwerth et al. 2014; Krijgsman et al. 2019; Kurbanov et al. 2019).

The Karangatian period was marked by a large interglacial transgression that exceeded the present-day sea level by 6-7 m, while the water salinity reached up to 30 ‰. There are two stages distinguished in the transgression development – the Karangatian and Tarkhankutian, each of them characterized by faunal assemblages with different proportions of stenohaline and euryhaline groups of mollusks. Two phases are also noted in the Karangatian stage. The earlier – Tobechik phase (Nevesskaya 1965) was marked by a wide distribution of species typical of the sea up to the present days (*Cerastoderma glaucum, Abra ovata* and others). The sea level in the basin was below that of today. The second phase (Karangatian) was characterized by the dominance of the halophilic species including those that are currently absent from the basin (*Cardium tuberculatum,* et al.). High salinity was observed in the southern part of the Sea of Azov as well. Another distinctive feature of the transgression was a higher water temperature, which is suggested by the malacofauna composition as well as the presence of thermophilic subtropical species of diatoms and pollen assemblages (Zhuse et al. 1980; Vronskiy 1976).

A series of the U/Th dates obtained for the transgression fall within the period of 140–70 ka BP (Arslanov et al. 1975, 1983; Dynamics… 2002). The ESR (electronic spin resonance) dates fall into the period of 127-121 ka BP (Dynamics… 2002). According to the OSL data, the earlier stage of the transgression developed around 131–120 ka BP, and the later one – around 120–100 ka BP (Kurbanov et al. 2019).

The Tarkhankut stage deposits yielded faunal assemblage, that included impoverished Mediterranean mollusk fauna, barren of halophilic elements and dominated by *Cerastoderma glaucum* and *Abra ovata*. The entire basin was confined within the present-day outlines of the Black Sea coasts and the salinity did not exceed 14-15‰. There were some Caspian species – *Didacna cristata, D. subcatillus, D.* ex gr. *protracta*, in the Tarkhankut basin, but mostly confined to limited area sites (Yanina 2012; Sorokin et al. 2019).

Post-Karangatian regression

The sea level during the regression dropped to around -80 … -100 m (Fedorov 1978; History of the geological evolution … 1988). I.P. Balabanov and Ya.A. Izmailov (1988) recorded the presence of *Didacna* sp. shells in the deposits. Judging from the diatom species composition, the water in the basin was cold and characterized by low mineralization (History of the geological evolution… 1988). A considerable cooling of the climate is indicated by the pollen assemblages mostly corresponding to dry and cool steppes on the Black Sea coasts (Shcherbakov et al. 1979).

Surozh transgression

Not all the specialists studying the Late Pleistocene history of the Pontian basins accept the Surozh transgression. It was established by Popov (1955), later Goretskiy (1957) applied to it the term 'Alanian'. The highest level of the basin is estimated at -25 …-20 m abs. The deposits of that transgressive basin are found on the Black Sea shelf (History of the geological evolution… 1988; Kuprin and Sorokin 1982; Shcherbakov 1982). The pollen assemblages recovered from the cores suggest a climate warming (Shcherbakov et al. 1979) and the period, corresponding to the Surozh basin, is estimated at 40-25 ka BP (Shcherbakov 1982).

New Euxinian epoch of regression and transgression

The New Euxinian stage in the Pontian basin development began with a significant regression. The basin level at its minimum was estimated at -80 m (Shcherbakov et al. 1977); -90 m (Fedorov 1978), at -100 to -110 m (Ostrovskiy et al. 1977); and at about -140 to -150 м (Rayan 1997; Winguth et al. 2000). Most of the specialists believe that the New Euxinian basin represented a completely isolated lake. According to the data obtained by Sholten (1974), the bottom of the Bosporus Strait is located at a depth of 100 m, which suggests either a constant one-way discharge of water from the New Euxinian basin or isolation of the latter. The extent of this significantly freshened basin covered the area of deep water, continental slope, and the lower part of the shelf. It was inhabited by mollusk fauna of freshwater (*Viviparus duboisianus, Lithoglyphus naticoides, Valvata piscinalis* etc.) and freshened brackish-water (*Monodacna, Dreissena rostriformis, Dr. polymorpha*) species, dominated by dreissenas.

During the period of regression, a low coastal plain existed in place of the Azov Sea, with the Don River flowing across it (Kaplin and Shcherbakov 1986). The Don mouth was located ~50 km south of the Strait of Kerch. The mouths of the rivers Dnieper, Dniester, and Danube joined together and formed a great canyon and a joint delta. The diatom flora (Zabelina and Shcherbakov 1975) indicates a considerable cooling. The pollen assemblages also suggest a cold and dry climate (Vronskiy 1976; Mudie et al. 2007; Filipova-Marinova et al. 2012). The emerged shelf and low coastal plains were dominated by the landscape similar to periglacial ones. As follows from the available data (Shimkus et al. 1977; Degens and Ross 1972; Briceag et al. 2019), the cooling in the region reached its maximum at 22–23 ka BP.

The existence of the regressive basin is dated to 22-17(16) ka BP (Shcherbakov et al. 1977; Balabanov and Izmailov 1989); 25-22 ka BP (Degens and Ross 1972). Some other specialists dated the regression maximum to a later time – 14-12 ka BP (Ostrovskiy et al. 1977).

Many researchers noted complex transgressive and regressive patterns in the New Euxinian basin dynamics (Balabanov and Izmailov 1988; Balabanov 2006; Murdmaa et al. 2006). The sea level was rising from 16 to 12.5 ka BP (Balabanov and Izmailov 1989) and reached -45 m (Varushchenko 1975). The final transgressive phase of the New Euxinian stage when the sea level reached -25 m is dated to 9.8 ka BP (Kovalyukh et al. 1977; Balabanov 2006). In addition, Murdmaa and his colleagues (Murdmaa et al. 2006) identified another basin – Antian with the water level reaching -30 m around 13 ka BP.

The New Euxinian transgressive basin was inhabited by brackish-water fauna dominated by mollusks that prefer very low salinity and belong mostly to *Monodacna, Adacna* and *Dreissena* genera. Typically euryhaline Mediterranean species are completely absent (Nevesskaya 1965; Popov 1983; Fedorov 1978). Shells of an early Khvalynian *Didacna ebersini* (Fedorov 1978) are occasionally encountered and some Khvalynian ostracods (Popov and Suprunova 1977) have been identified. The inflow of Mediterranean water into the New Euxinian basin first occurred around 9.8-9.5 ka BP (Jones and Gagnon 1994) when the Holocene Black Sea transgression began in the Pontian region.

Caspian mollusks in the Late Pleistocene Black Sea

The presence of the Caspian Sea mollusks in the Azov – Black Sea basins is of primary importance for the correlation of events in the Pontian and Caspian basins as it proves the functioning of the paleo-straits between the basins. The Caspian mollusk fauna consists of species autochthonous for the Caspian Sea and endemic to the Pontian-Caspian basin. In the Neopleistocene the Caspian mollusks occasionally migrated through the Manych Strait into the Black Sea basin and evolved there.

As follows from the analysis of the Late Pleistocene malacofauna of the Pont (Yanina 2005), brackish-water didacnas persisted in isolated (freshened) water areas throughout the Karangatian epoch. Two groups of mollusks, different in origin, are identified: (1) Euxinian-Uzunlarian species (*Didacna pontocaspia, D. borisphenica*), that survived the period of increased salinity in the freshened parts of the basin; (2) Caspian species (*Didacna cristata, D. subprotracta, D. subcatillus*), which most likely penetrated into the retreating Karangatian basin with the Hyrcanian water and settled within a few limited areas of the Tarkhankut basin.

During the period of the New Euxinian regression, when the basin was noticeably freshened, all the didacnas became extinct. The New Euxinian transgressive basin was dominated by semifreshwater Caspian species (*Monodacna, Adacna, Hypanis, Dreissena*), with occasional *Didacna moribunda* (Andrusov 1926; Fedorov 1963; Semenenko and Sidenko 1979), identical to *Didacna ebersini* which is an index species of the Khvalynian fauna (early Khvalynian assemblage) of the Caspian Sea.

The presence of the Caspian assemblage members (Hyrcanian and early Khvalynian species) in the Tarkhankut and New Euxinian (Pontian) basins suggests the Caspian water inflow into the Black Sea during that period, most likely, via the Manych Strait.

Some researchers (Arkhangelskiy and Strakhov 1938; Dimitrov and Govberg 1979) believe that the faunal elements of the Old Euxinian basin could have survived during the marine Karangatian transgression in the freshened limans and then spread out over the New Euxinian basin. Others (Shnyukov et al. 1981) believe that the brackish-water New Euxinian species migrated from the Caspian Sea to the Surozh basin; they could have survived in the most freshened areas and then – at the New Euxinian time – expanded widely. As has been stated by zoologists (Mordukhay-Boltovskoy 1960) who have studied the Caspian fauna in the Azov-Black Sea basins, if the relicts of the Old Euxinian fauna persisted in the modern basin, the species composition of the Caspian fauna in the two basins (the Caspian and the Azov-Black Sea) would be quite different. The isolation is a powerful factor of species formation. The fact that the species of the two isolated basins are similar, means that the Caspian fauna existing now in the Azov-Black Sea basin persists seemingly since the end of the Pleistocene.

It can be concluded that the paleogeographic events that took place in the Pontian basin in the Late Pleistocene were closely connected with the global climate change (Fig. 3). The global warming at MIS 5e and the rise of the sea level forced the Mediterranean water to enter the Black Sea depression which resulted in the Karangatian transgression. The global cooling during the transition to the Valday glacial epoch initiated a drop of the Karangatian sea level that followed the global sea-level lowering. Separately from the Karangatian sea, the Hyrcanian basin of the Caspian Sea under similar conditions transgressed and a part of its water was transferred to the Tarkhankut basin (the 2nd stage of the Karangatian epoch). The early Valday ice age (Kalinin, MIS 4) was marked by the presence of the post-Karangatian regressive basin. Its level became somewhat higher (transgressed) during the interstadial warming (MIS 3), though still remained below zero mean sea level. The late Valday glacial epoch (Ostashkov glaciation, MIS 2) resulted in the most distinct sea level drop (at the LGM) and caused a deep New Euxinian regression of the Pontian basin. When the continental ice sheets and the permafrost degraded, the New Euxinian basin transgressed, though sea level was still negative. At that time water of the early Khvalynian transgression of the Caspian basin was partly discharged through the Manych into the Pontian basin. In the Holocene the global interglacial warming resulted in the inflow of the Mediterranean Sea water into the Black Sea basin and development of transgression.

Manych

The analysis of the Manych Strait functioning based on the Quaternary series studies in natural exposures and cores plays an important part in correlating the events and understanding the connection and interaction between the Caspian and Pontian basins in the Late Pleistocene.

Judging from the stratigraphic position and malacofauna recovered from the Manych valley deposits, there was an ingressive bay there at beginning of the Late Pleistocene (the Karangatian transgression maximum) which penetrated as far east as the Caspian – Black Sea water divide (Fedorov 1978; Popov 1983; Yanina 2014; Kurbanov et al. 2018). The presence of the Karangatian fauna in its deposits (*Cerastoderma glaucum, Chione gallina, Chlamys glabra, Ostrea edulis*) suggests a rather high salinity in the central part of the bay (~18–20‰). The head of the bay was close to the Kalaus River mouth. A wide distribution of *Cerastoderma glaucum* and disappearance of more halophilic species indicates considerably freshened water (up to 10‰) (Popov 1983).

Popov (1983) identified two stages in the Karangatian Sea ingression, the second marked by an increased ingression range. It can safely be assumed that the earlier stage corresponded chronologically to the development of an inlet of the late Khazarian basin with a lower water level (Yanina 2012). The 2nd stage of the ingression correlates with the Hyrcanian transgression with a bay deeply penetrating westward via the Eastern Manych valley. When the level of the Karangatian basin dropped and the ingressive inlet shrank, the Caspian (Hyrcanian) water penetrated into the strait bringing mollusks *Didacna cristata, D. parallella, D. subcatillus, Monodacna caspia, Dreissena polymorpha*. The salinity in the strait (judging from the malacofauna) was about 8–10‰, the water was notably freshened by the inflowing streams.

The Hyrcanian deposits in the central part of the Manych depression are dated using OSL at 107±7 ka BP (Kurbanov et al. 2018). It supports the earlier conclusion about the Karangatian sea level lowering (Tarkhankut stage) and the inflow of the Hyrcanian water during the cooling at the transition from the Milulino Interglacial to the Valday glaciation. At the end of the Hyrcanian a lake appeared in the Manych valley – Burtass lake, which, according to Goretskiy (1953) chronologically correlates with MIS 4 (Popov 1983; Kurbanov et al. 2018).

The next strait opening occurred during the Early Khvalynian epoch when the Caspian level reached about 50 m a.s.l. This follows from the geomorphological structure of the Manych depression and is substantiated by the paleontological findings recovered from its deposits (Popov 1983; Svitoch et al. 2010; Yanina 2012). The first stage in the strait development was marked by erosion processes. The early Khvalynian water reached the Manych threshold and flowed towards the Black Sea basin cutting through the Burtass lake sediments and subaerial deposits above them. That stage in the strait development is expressed in the linear hollows and ridges in the Manych depression (Svitoch et al. 2010) and Abeskun deposits that contain the early Khvalynian species *Didacna ebersini* (Goretskiy 1953; Popov 1983). That stage is dated using radiocarbon at 17–16 ka BP (Yanina 2012)

The next stage in the strait evolution was apparently depositional and was marked by the formation of fine deposits between the ridges and development of a ~22 m high terrace. Among the mollusks there were *Didacna ebersini, D. protracta, D. subcatillus, Monodacna caspia, Adacna laeviuscula, Hypanis plicatus, Dreissena polymorpha, Dreissena* *rostriformis distincta* present. The geological structure of the deposits suggests an ingressive type of the strait, with water penetrating the eroded valley and the stream increasing gradually in capacity (Svitoch et al. 2010). The strait of that type could have developed during the transgressive stage of the early Khvalynian basin when the sea level reached ~22 m. The sedimentary sequence and mollusk assemblages recovered from it indicate a unidirectional migration of the mollusk fauna from the Caspian Sea into the Pontian New Euxinian basin (Svitoch and Yanina 2001; Yanina 2005, 2006; Chepalyga 2004, 2006; Svitoch 2006, 2007). That stage, the last one in the strait development in the Pleistocene, is dated at 14.8–14.3 ka BP (Svitoch et al. 2010).

It can be concluded that during the Late Pleistocene the Manych Strait with a one-way flow of the Caspian water to the Pontian basin was open three times, namely once in the Hyrcanian and twice in the early Khvalynian intervals of the Caspian history. An ingressive bay that existed in the Manych valley during the Karangatian interglacial transgression of the Pont reached as far east as the Caspian–Black Sea water divide (Fig. 3).

Mediterranean Sea

The Late Pleistocene regime of the Mediterranean Sea depended on the global sea-level fluctuations due to continuous connection between the sea and the Atlantic Ocean through the Strait of Gibraltar.

Tyrrhenian transgressive epoch

The Tyrrhenian epoch represents a complicated and prolonged transgressive-regressive period in the Mediterranean Sea history. It was the most prominent interval in the Neopleistocene paleogeography, marked by a wide distribution of the tropical malacofauna of the Senegalese type, with its most characteristic type Strombus bubonius. It has been found that the penetration of tropical malacofauna elements started as early as the Middle Neopleistocene and persisted during at least a part of the Late Neopleistocene (Cita et al. 1973; Paskoff and Sanlaville 1980; Zazo and Goy 1984; Zubakov 1986; Svitoch et al. 1998; a.o.). There are as many as four marine terraces with Tyrrhenian malacofauna known on various coasts of the Mediterranean.

Three stages can be distinguished in the Tyrrhenian Sea level rise during the Late Pleistocene – Eutyrrhenian, Neotyrrhenian and Epityrrhenian, corresponding to climatic substages MIS 5e, 5c and 5a (Zubakov 1986; Svitoch et al. 1998). The main peak of transgression with the sea level ~4-6 m above present sea level corresponded to the climatic optimum of the Eemian Interglacial (MIS 5e), characterized by warmer and wetter climate compared to the current conditions. The 2nd transgressive rise of the sea level exceeded the present level by 1.5 m and occurred under the climate conditions similar to the present days. The 3rd sea level rise corresponding to MIS 5a was insignificant and did not rise above the modern level of the Mediterranean (Çağatay et al. 2009).

Grimaldi regressive epoch

The Tyrrhenian transgression was followed by a prolonged interval of decreased sea level (MIS 4-2). The process was rather irregular. During the glacial epoch (MIS 4) the Tyrrhenian sea level dropped by 60 - 90 m (Blanc 1937; Shimkus 1981; Svitoch et al. 1998; Çağatay et al. 2009). The deep-sea deposits attributed to that time are distinguished by alternating layers with warm-water and cold-water foraminifera indicative of insignificant warming and cooling of the climate. Pollen assemblages recovered from the deposits and dated to the interstadial warming (MIS 3) display also an alternation of the subtropical and sub-boreal (boreal) vegetation. The layers also differ in the proportion of the thermophilic planktonic foraminifera. At that time the level of the regressive post-Tyrrhenian basin increased up to (though no more than) -40 m abs. The Mediterranean deep-sea sediments dated to the last glacial epoch (MIS 2) are dominated by cold-tolerant foraminifera; the pollen assemblages abound in birch, pine, and Artemisia (Shimkus 1981). Numerous data sources indicate a sharp and deep drop of the sea level at the LGM with the estimates varying from 100 to 300 m (Keraudren and Sorel 1987; Zubakov 1986; Bruckner 1986; Svitoch et al. 1998; Çağatay et al. 2009; a.o.).

Flandrian transgression

In the Mediterranean region the post-glacial glacio-eustatic rise of the sea level is known as Versilian or Flandrian transgression. Its beginning is dated to the early post-glacial time (~17–15 ka BP), and further development has been thoroughly studied in various Mediterranean regions (Keraudren and Sorel 1987; Aksu et al. 1999; Badertscher et al. 2011; Cecilia et al. 2008; a.o.). The transgression began with a large volume inflow of the North-Atlantic water into the Mediterranean Sea and resulted in a wide distribution of the modern-type mollusk fauna of relatively thermophilic species represented by Mediterranean-Lusitanian and Mediterranean-Canarian forms (*Chlamys glabra, Mytilaster lineatus, Corbula mediterranea, Pitar rudis*, a.o.), moderately thermophilic Keltian (*Mytilus galloprovincialis, Cardium paucicostatum, Donax venustus*) and rather cold-loving Keltian forms (*Nucula nucleus, Ostrea edulis, Cerastoderma glaucum, Chione gallina, Solen vagina*). The amplitude of the sea level rise 10–9 ka BP is almost universally adopted to be up to -30 m abs.

Therefore, it may be safe to assume that the evolution of the Mediterranean basins through the Late Pleistocene was controlled by fluctuations of the World Ocean level, which in turn had been initiated by the global climate changes.

The Sea of Marmara

The Sea of Marmara, which is called the «gate» from the Mediterranean to the Black Sea, is a paleogeographically important element of the system under consideration that presents evidence of the interaction between the two sea basins in the past.

It has been established that the Late Pleistocene history of the Marmara Sea presented an alternation of marine and lacustrine stages. A marine basin existed in the Marmara Sea basin at the beginning of the Late Pleistocene (Eemian Interglacial, MIS 5e). The marine conditions developed as a result of the sea water invasion when the Mediterranean Sea level rose above the Dardanelles Strait threshold (at present its altitude -65 m). The sapropel layers rich in organic matter mark those events in the sedimentary record. The malacofauna was similar to that of Tyrrhenian age in composition: it included numerous Mediterranean species, mostly from euryhaline to moderately stenohaline; the stenohaline marine species are less common, which suggests the salinity about 28-30 ‰ (Çağatay et al. 2009; Büyükmeriç et al. 2016; Meriç et al. 2018; Krijgsman et al. 2019). Three marine stages are recognized in the Late Pleistocene history of the Marmara Sea – during the interglacial (MIS 5e) and during the warm intervals of the transitional period (MIS 5c and MIS 5a). The most significant influx of marine water occurred during the Eemian Interglacial period. The 4th marine stage began at the very end of the Pleistocene and continued during the Holocene Interglacial.

When the sea level dropped below the Dardanelles threshold, the marine environments were replaced with lacustrine ones. This was recorded in the Upper Pleistocene sedimentary series by the accelerated erosion, formation of rills, and the lacustrine deposition. The Sea of Marmara turned to a freshened brackish-water lake during the cooling phases of the transitional period (MIS 5d and MIS 5b), and in the glacial epochs (MIS 4 and MIS 2). The lacustrine environments persisted during the interstadial warming (MIS 3) when the ocean level rose. Evidently, the threshold in the Dardanelles was higher at that time and therefore prevented the marine water inflow into the Sea of Marmara basin. Some authors (Meriç et al. 2017) consider a possibility of the water discharge from the Surozh (Pontian) basin.

The lowest stand of the lake level (-85 to -95 m) was observed during the last glacial maximum (LGM) when it was completely isolated from the ocean (Çagatay et al. 2000; Algan et al. 2001; Aksu et al. 2002; Hiscott et al. 2002; Badertscher et al. 2011). The Marmara Sea turned to a freshened brackish-water lake of the «Caspian» type, with the salinity varying from 1 to 7‰. The seasonal contrasts were sharp. Such climatic and hydrological conditions persisted approximately till 20 cal ka BP. The first wave of the warming occurred between 20 and 18 cal ka BP; it was followed by a considerable increase of the melted water inflow. The lake was almost completely devoid of fauna except for occasional representatives of *Dreissena* and *Theodoxus* mollusk genera. Flora of brackish-water and freshwater diatoms was rather scarce. The coasts were mostly treeless (Filipova-Marinova et al. 2004; Mudie et al. 2001, 2007), which resulted in intensive erosion and transportation of the abundant clastic material into the lake.

An episode of a considerable inflow of the freshened water from the Black Sea was recorded in the interval between ~18 to 15 cal ka BP (Aksu et al. 1999, 2002; Çagatay et al. 2000; Bahr et al. 2005; Major et al. 2006; Herrle et al. 2018). Probably, it was a result of the Black Sea level rising above the Bosporus threshold at the time of the New Euxinian transgression in the Black Sea basin. The Bølling-Allerød interstadial brought warm conditions to the Marmara Lake, which is reflected in the sediments, fauna, and flora of the lacustrine facies. Brackish-water and freshwater diatoms and woody material became less abundant just before the marine incursion. Once having reached the Dardanelles threshold, the Mediterranean water began to gradually fill the Sea of Marmara depression.

A considerable cooling during the Younger Dryas period resulted in lowering of the Marmara Sea level and water inflow from the Black Sea, the intensity of the process is still open to discussion (Çagatay et al. 2000; Algan et al. 2001; Hiscott et al. 2002). Judging from the pollen assemblages, the Younger Dryas was the most arid interval of the last ice age in the Eastern Mediterranean (Filipova-Marinova et al. 2012; Mudie et al. 2002, 2007). The Mediterranean water inflow repeated in a rather short time, less than ~1.5 ka later as estimated by Reichel and Halbach (2007). The isotope analysis results revealed the following pattern: when the marine water enters the lake, it becomes saline in a rather short time (1 or 2 ka); when the marine water is outflowing through the Dardanelles Strait, its freshening proceeds slowly taking 3 to 5 ka (Reichel and Halbach 2007). The marine regime established in the Sea of Marmara basin by 9.3-9.0 ka BP (Büyükmeriç et al. 2016).

It can be safely concluded that the Marmara Sea development in the Late Pleistocene was mostly controlled by the level fluctuations in the Black Sea and the Mediterranean, which in turn resulted from the global changes of the climate. An interconnection between the basins depended also on the altitude of thresholds in the Bosporus and Dardanelles straits.

CORRELATION AND INTERRELATIONS OF THE EVENTS

The paleogeographic analysis of the basins forming the system of the Pontian–Caspian–Mediterranean seas (subsequently referred to as the System) allows to analyze the correlation between transgressive and regressive events in individual parts of the System and to reconstruct the functioning of the entire System under the conditions of global climate change (Fig. 4).

In the Caspian Sea, the beginning of Late Pleistocene was marked by a «little» late Khazarian transgression that occurred during the interglacial epoch (MIS 5e). The sea level rise resulted from the climate conditions characterized by a high humidity at the Mikulino (Eemian) Interglacial optimum. At the same time, in the context of the interglacial transgression of the global sea level that reached 6-7 m above the current level, all the sea basins connected to the ocean experienced a transgressive rise of the level: the Tyrrhenian transgression in the Mediterranean, and Karangatian in the Pontian basins. All the straits between the sea basins were functioning. The Karangatian water penetrated deeply into the Manych depression forming an ingression bay, though the Manych threshold prevented its discharge into the late Khazarian basin. The transgression of the marine basins reached its maximum during Neopleistocene, while the transgressive basin during the late Khazarian was isolated from the seas and its level stayed below present mean sea level (Fig. 4a).

The Tyrrhenian transgression proceeded in three stages, two of which were marked by the water penetrating into the Pontian basins and causing the two-stage Karangatian transgression. The 2nd Karangatian stage (Tarkhankut) developed together with the Hyrcanian basin of the Caspian Sea. Its water penetrated deeply into the Eastern Manych valley forming a bay. Another ingressive bay – Karangatian – existed from the Pontian side. The global cooling and the onset of the Valday (Weichselian) glaciation initiated a regressive trend in all the marine basins following the global sea level drop. The drop of the Karangatian sea level led to a gradual reduction of the Manych bay length and finally to its complete disappearance. The same climatic conditions favored the Hyrcanian lake basin transgression by increasing the positive constituent of its water balance. The basin level rose above the Manych threshold, and its water flowed towards the Pont, which resulted in a slight rise of the sea level and a decrease in salinity. Such correlation of the events is corroborated by the presence of Hyrcanian malacofauna in Manych and in the Tarkhankut Pontian basin (Fig. 4 b).

The long-lasting and structurally complicated Tyrrhenian transgression in the Mediterranean, as well as the two-stage Karangatian transgression in the Pontian basin and the Khazarian transgressive epoch, also two-stage, in the Caspian Sea – all of them developed within the period corresponding to MIS 5 – Eemian Interglacial *sensu lato* (s.l.), or Eemian Interglacial (MIS 5e), and the period transitional to the glacial period (MIS 5d-a). It should be noted that the transgressions developed in different basins for various reasons. The marine

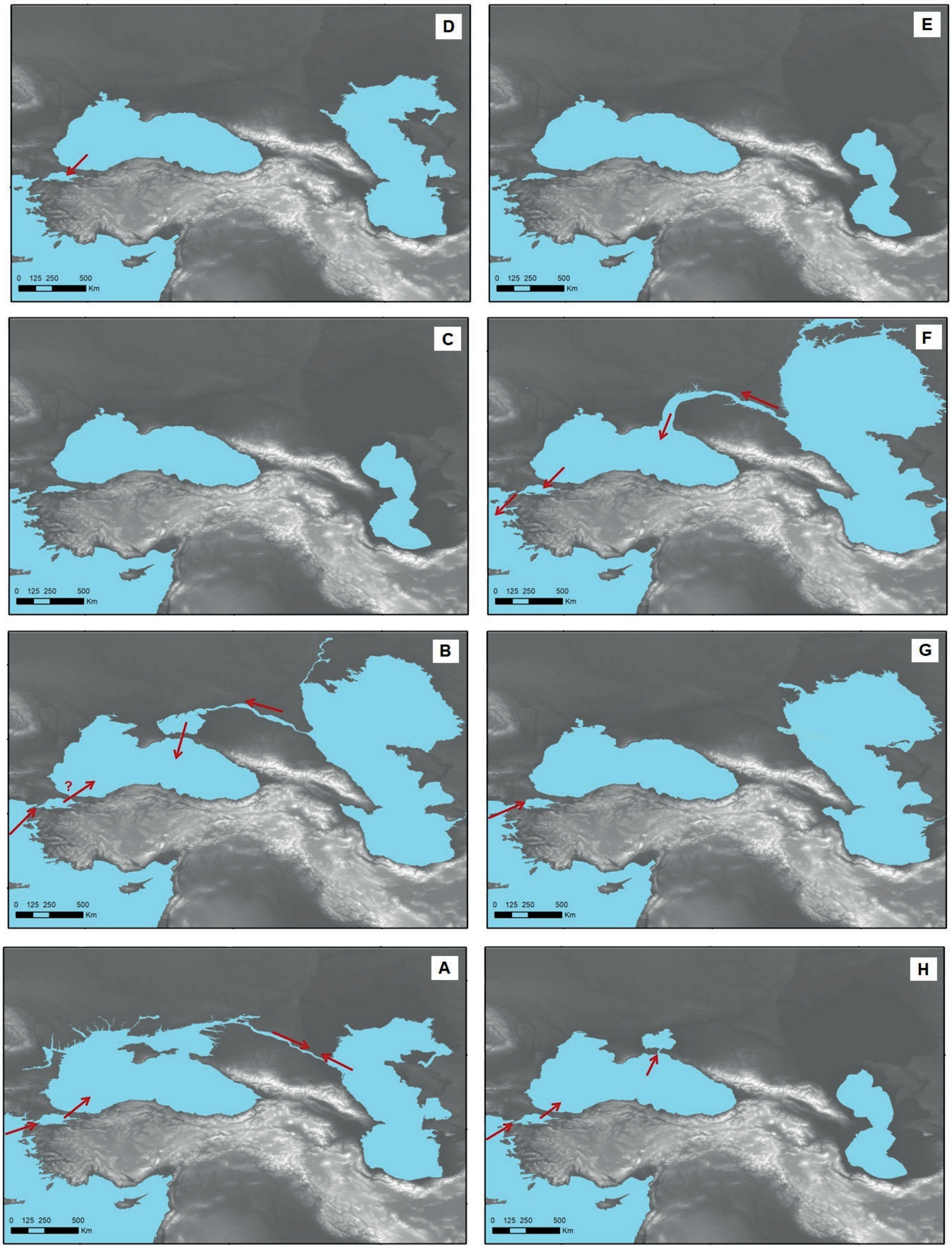

Fig. 4. The Pontian – Caspian – Mediterranean system functioning under conditions of the global climate changes

A - Mikulino - Eemian interglacial (MIS 5e): Tyrrhenian transgression in the Maditerranenan and Marmara seas, Krangatian marine transgression in the Pont (with ingression into the Manych valley) and the Late Khazarian lake transgression in the Caspian Sea;
B - Transition from interglacial to glacial (MIS 5 d-a): Tyrrhenian transgression in the Maditerranenan and Marmara seas, Tarkhankutian basin in the Pont and Hyrcanian transgression in the Caspian, Hyrcanian passage in the Manych Valley;
C - Early Valdai – Early Weichselian glaciation (MIS 4), Glacial maximum: Grimaldi regression in the Mediterranena and Marmara seas, Post-Karangatian regression in the Pont and Atelian regression in the Caspian;
D - Interstadial warming (MIS 3, second half): Inter-Grimaldi transgression in the Mediterranean Sea, lake transgression in the Sea of Marmara, Surozh transgression in the Pont and beginning of the Khvalynian transgression in the Caspian;
E - Late Valdai – Late Weichselian glaciation (MIS 2), Glacial maximum: the Grimaldi regression in the Mediterranean and Marmara seas, the Neoeuxinian regression in the Pont and regression (Eltonian?) in the Khvalynian (Caspian) basin;
F - Degradation of Glaciation (MIS 2): the Grimaldi basin in the Mediterranean and Marmara seas, the Neoeuxinian transgression in the Pont and maximum of the Khvalynian transgression in the Caspian; cascade of basins;
G - Glacial degradation (MIS 2): beginning of the Flandrian transgression in the Mediterranean and Marmara seas, the Neoeuxinian transgression in the Pont and Late Khvalynian transgression in the Caspian;
H – Holocene (MIS 1, beginning): the Flandrian transgression in the Mediterranean and Marmara seas, the Chernomorean (Black Sea) transgression in the Pont and the Mangyshlak regression in the Caspian Sea.
Arrows indicate the water inflow and migration of fauna.

transgressions resulted from the rise of the global sea level above the thresholds in straits connecting individual seas in the System. The Caspian transgressions resulted primarily from the positive water balance in the drainage basin. The discharge through the Manych and the strait functioning depended on the sea level rise above the Manych threshold.

During the glacial time (MIS 4) all the basins in the considered System were at a regressive state: the Roman regression in the Mediterranean, post-Karangatian regression in the Pontian basin and Atelian regression of the Caspian Sea. Lake basins existed in the depressions of the Marmara Sea, the Black Sea, and the Caspian Sea, a lacustrine plain of the Pre-Don occupied the area of the Sea of Azov, and none of the straits was functioning (Fig. 4c). Marine basins connected to the ocean followed its regression (the glacial regression). The Caspian regression resulted from the negative water balance under the conditions of glacial climate.

The rise of the Mediterranean Sea level began during the 1st half of the interstadial warming (MIS 3) following the interstadial rise of the global sea level. The straits were closed, and isolated lakes persisted in the depressions in places of the Marmara Sea, the Black Sea and the Caspian Sea.

The transgressive rise of the level in the Mediterranean Sea continued during the 2nd half of the interstadial warming (MIS 3). The Dardanelles Strait was closed. A lake regime persisted in the Marmara Sea depression, though the basin level has increased due to the increasing water inflow. The same reason has led to the post-Karangatian regression changing to a small-scale Surozh transgression (with sea level not reaching 0 m abs). It had been speculated that a part of its water was flowing into the Sea of Marmara. The Mediterranean water did not flow to the Marmara, or to the Pontian basins. The Manych Strait was closed (Fig. 4d). The Mediterranean Sea transgression was induced by the global sea level rise during the interstadial period. The sea level, however, was below the Dardanelles Strait threshold. The development of the lake basins in the Marmara Sea, Pontian basin, and the Caspian Sea depended on the relationship between the inflow and outflow constituents of the water balance in each of them.

During the glacial epoch (MIS 2) all the basins in the System have experienced regression. The Last Glacial Maximum (LGM) was marked by a deep regression of the global ocean which was followed by regression (Grimaldi) of the Mediterranean Sea. In the isolated basins of the Marmara Sea, Pontian and Caspian seas the LGM was marked by cold and dry conditions, which resulted in a considerable drop of the water level (Grimaldi, New Euxinian and Elton regressions) (Fig. 4e).

The deglaciation epoch led to the water level rise in all the basins under consideration. The Mediterranean Sea, closely related to the ocean in its evolution, transgressed gradually, though irregularly. The «Great» Khvalynian transgression developed in the Caspian Sea due to the sharp increase of the water inflow into that lake. One of the reasons for the high level of the Khvalynian basin was a high threshold of the Manych depression. After it was exceeded, the Khvalynian water started flowing into the New Euxinian (Pontian) lake basin. There were two stages distinguished in the Khvalynian water discharge corresponding to the transgression reaching 45-50 m a.s.l. (initial phase of the warming) and 22-20 m a.s.l. (Bølling warming). The Pontian basin and the Marmara Sea levels depended heavily on the levels of the Caspian and Mediterranean seas. The Mediterranian did not exert any influence onto the New Euxinian basin of the Pont till the Holocene. However, it was twice subjected to the impact of the Caspian due to the Manych Strait opening, which led to the increase of the New Euxinian basin level. Because of the low threshold in the Bosporus, the New Euxinian basin turned to a flow-through (drained) lake: it received the water from the Khvalynian basin of the Caspian, and when the lake level exceeded the Bosporus threshold it started flowing from the New Euxnian basin into the lake located in the Marmora Sea basin. Supposedly, the water could flow further, through the Dardanelles Strait into the eastern part of the Mediterranean (Grimaldi) Sea. So, a system of flow-through lakes developed as follows: Khvalynian basin of the Caspian Sea – New Euxinian (Pontian) basin – Grimaldi basin of Mediterranean (Fig. 4f).

Under the conditions of the ongoing deglaciation, the transgressing Mediterranean Sea reached the Dardanelles Strait threshold, the strait opened and marine water entered the Sea of Marmara, which gradually transformed into a marine basin. Another episode of the New Euxinian water outflow into the Marmara Sea occurred at the end of the Late Pleistocene, in the Younger Dryas. During the beginning of the Holocene, the marine regime stabilized. The sea level reached the Bosporus threshold, and the marine water started to fill the New Euxinian (Pontian) basin, which gradually turned into the modern Black Sea and the Sea of Azov (Fig. 4g). In the Caspian Sea, the cold and dry climate of the Younger Dryas led to a lowered level of the Khvalynian basin (Enotayevka regression). The subsequent sudden warming induced a transgressive rise of the sea level during the Late Khvalynian stage. The Khvalynian epoch in the Caspian Sea evolution ended with the Mangyshlak regression under conditions of a strongly continental climate in the Boreal period of the Holocene.

CONCLUSION. THE PONTIAN – CASPIAN – MEDITERRANEAN SYSTEM FUNCTIONING UNDER THE CONDITIONS OF THE GLOBAL CLIMATE CHANGE

The Pontian–Caspian and Mediterranean basins belonged to different types of water basins and evolved differently in the Late Pleistocene responding in different ways to the changes in global climate. Paleogeographic reconstructions and correlation analysis of the Late Pleistocene events (within the last climatic macrocycle) made it possible to view the evolution of the basins as parts of a single system allowing to identify certain specific features and patterns in their functioning.

The interglacial epoch (MIS 5e) was marked by transgression in all the basins in the System, which could be attributed to different reasons. Marine transgressions resulted from the rise of the global sea level and the opening of the straits (as the sea level exceeded the strait threshold) between the elements of the System. The Caspian lake transgression resulted from the positive water balance of the basin. The marine transgression reached its highest level, while the Caspian transgressive basin stayed below present mean sea level.

During the transition to the glacial period (MIS 5d-a), the Mediterranean Sea level was unstable: its development was interrupted twice (MIS 5d and 5b) by the level drop below the Dardanelles Strait threshold. Those events also affected the Marmara Sea, where marine transgression developed in two stages. During the first stage (MIS 5c) the level exceeded the Bosporus threshold, which led to the rise of the Black Sea level. In the Caspian Sea, the climatic conditions of the transitional period resulted in positive water balance, which caused transgressive evolution of the Caspian basin. The Caspian water flowing through the Manych into the Pontian basin and opening of the strait towards the Mediterranean resulted from the Caspian Sea level rising above the Manych threshold elevation.

During the peaks of the glacial epochs (MIS 4 and MIS 2) all the basins of the System were at regressive stages. None of the straits was functioning. The sea basins that have a connection with the ocean followed its regression; the Caspian Sea, however, regressed due to negative water balance under the conditions of the glacial climate. The colder and dryer conditions of the LGM resulted in even deeper regression.

The interstadial warming (MIS 3) was marked by a small increase in the sea level (not exceeding 0 m abs.). In the Mediterranean Sea, the transgressive level rise resulted from the global sea level increase. In contrast to that, the Marmara, Black and Caspian seas were isolated from the ocean during the glacial period, and their levels depended on the water balance conditions in their drainage basins.

The degradation of the last glaciation (MIS 2) and transition to the Holocene interglacial resulted in the level rise in all the basins (though different in magnitude), which was interrupted by the cold climate phases (Oldest, Older and Younger Dryas). The Mediterranean Sea was constantly connected to the ocean and transgressed accordingly. The highest level rise was observed in the Caspian Sea, which can be attributed to a considerable increase of the water input. The maximum sea level was controlled by the height of the Manych threshold, once that level was exceeded, the Caspian water started flowing into the Pontian basin, the level of which was then located below the present mean sea level. In total there were two instances of the outflow. The input of the Caspian water into the Pont led to the level rise and a discharge of water through the strait into the Sea of Marmara and further into the Mediterranean, thus forming a system of drainage lakes.

The marine regime in the Mediterranean – Pontian part of the system stabilized at the beginning of the Holocene, during the period of the interglacial transgression of the ocean. The Caspian Sea continued its development as a lake basin with high sensitivity to the climate fluctuations.

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