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Quaternary time scales for the Pontocaspian domain: Interbasinal connectivity and faunal evolution

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A B S T R A C T

The Pontocaspian (Black Sea - Caspian Sea) region has a very dynamic history of basin development and biotic evolution. The region is the remnant of a once vast Paratethys Sea. It contains some of the best Eurasian geological records of tectonic, climatic and paleoenvironmental change. The Pliocene-Quaternary co-evolution of the Black Sea-Caspian Sea is dominated by major changes in water (lake and sea) levels resulting in a pulsating system of connected and isolated basins. Understanding the history of the region, including the drivers of lake level and faunal evolution, is hampered by indistinct stratigraphic nomenclature and contradicting time constraints for regional sedimentary successions. In this paper we review and update the late Pliocene to Quaternary stratigraphic framework of the Pontocaspian domain, focusing on the Black Sea Basin, Caspian Basin, Marmara Sea and the terrestrial environments surrounding these large, mostly endorheic lake-sea systems.

1. Introduction

The Black Sea and Caspian Sea basins are the present-day remnants of the ancient Paratethys Sea (Laskarev, 1924), an epicontinental water-mass that developed since the earliest Oligocene in central Eurasia as the northern branch of the Tethys Ocean. It was separated from the southern, Mediterranean branch by the Alpine-Caucasus-Himalayan orogenic belt that progressively formed by ongoing tectonic collision of the Eurasian plate with the African-Arabian and Indian plates (Rögl, 1999; Popov et al., 2006). In Oligo-Miocene times, the Paratethys Sea covered large parts of Europe and Asia, stretching from southern Germany in the west to western China in the east. A complex combination of Mio-Pliocene tectonic uplift, glacio-eustatic sea level fluctuations, and sedimentation by major deltaic systems, progressively filled the

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Fig. 1. Ancient map of the Pontocaspian region after Kircher (1678), who in his "Mundus Subterraneus" already envisaged that the Caspian Basin must have been connected to the open ocean to explain its relatively high salinity (> 10 ‰) today. The Caspian Sea is in fact an isolated long-lived lake since at least 2.6 Ma. Kircher considered a subterraneous channel to the Persian Gulf for the marine connection. The location of this marine connection is still enigmatic today.

marginal sedimentary basins in the west and east. Consequently, the Paratethys sea drastically retreated, which influenced the regional climate of Eurasia (Ramstein et al., 1997) and facilitated mammal (including hominid) migration between Africa, Asia and Europe (Bar-Yosef and Belmaker, 2011).

Throughout its entire history, Paratethys formed a series of restricted basins, separated by shallow, tectonically active, gateways, that were extremely sensitive to small climatic and tectonic variations (e.g., Popov et al., 2006; Palcu et al., 2017). The semi-enclosed basin configuration resulted in extreme palaeoenvironmental dynamics including anoxic, hypersaline, brackish to fresh water conditions. Some salinity regimes in the basin, that were supposed to require connectivity to the open ocean, have been a major scientific puzzle for many centuries (Fig. 1: Kircher, 1678). The long-lived isolated position of the basins, combined with the exceptional palaeoenvironmental conditions, created faunal communities that are endemic to the Paratethys region, and that waxed and waned through geological history (e.g., Harzhauser et al., 2002). They obtained maximum extension during the latest Messinian (Lago-mare) times ~5.5 Ma, when Paratethyan faunas occupied the entire Mediterranean Basin as well (e.g., Guerra-Merchán et al., 2010; Stoica et al., 2016). Today the Paratethyan faunas are at their minimum: the so-called Pontocaspian communities are now almost entirely restricted to small enclaves in the major deltaic and estuarine systems of the northern Black Sea as well as the Caspian Sea (Grigorovich et al., 2003; Yanina, 2012a). The origin, evolution and migration of these characteristic Pontocaspian faunal elements is still not fully understood (e.g., Wesselingh et al., 2008).

The Pliocene-Quaternary co-evolution of the Black Sea-Caspian Sea is dominated by major changes in sea/lake levels. These may have been driven by external components, resulting from opening and closing gateways to the Mediterranean or Arctic ocean, as well as by internal components where hydrological and climatic changes induced by glacial-interglacial cycles may have created periods of intermittent interbasinal connectivity (e.g., Badertscher et al., 2011; Yanina, 2014). A simplified scenario is that the two basins were isolated during lowstands, when individual water levels, environmental conditions, and faunal composition were largely determined by the local hydrological budgets. The two basins became connected during highstands of the Caspian Sea. During such periods, overflow of the Caspian Sea through the Manych low-land connection north of the Greater Caucasus enabled faunal exchange (Fig. 2). Environmental conditions became similar in both basins by mixing of the water masses and consequently migration and blending of the Pontocaspian fauna took place. Additionally, the Black Sea became connected to marine waters of the Mediterranean during interglacial highstands and the Pontocaspian biota were marginalized. These marine transgressions did not reach the Caspian Basin. Sea/lake level changes in the Pontocaspian region have been gigantic. Over 1000 m lake level rise has been proposed for the late Pliocene Productive Series-Akchagylian transition in the South Caspian Basin (see van Baak et al., 2017 and references therein). Also on short time scales, lake level changes were very significant as shown by Caspian Holocene variations of ~100 m affecting historic settlements on the Caspian coast (Kroonenberg et al., 2007). The Quaternary sea/lake level history of the Black Sea – Caspian Sea domain is still enigmatic

Fig. 2. Present-day drainage area of the Pontocaspian domain. Yellow circles denote the locations of the stratotype sections of the main Quaternary stages of the Caspian Basin and Black Sea Basin: 1) Akchaghylian on Krasnovodsk peninsula (Turkmenistan), 2) Apsheronian on Apsheron Peninsula (Azerbaijan), 3) Bakunian in Baku (Azerbaijan), 4) Kuyalnikian (Ukraine), 5) Gurian (Georgia), 6) Chaudian on Cape Chauda (Crimea) and 7) Uzunlarian (Crimea). White circles denote the locations of key sections: 1) Pyrnuar (N38.93, E56.26), 2) Malyi Balkhan (N39.27, E54.97), 3) Yuzhny Urundzhik (N39.27, E54.50), 4) Ushak (N40.45, E53.37), 5) Lokbatan (N40.33, E49.75) and Jeirankechmez (N40.24, E47.09), 6) Duzdag (N40.70, E46.92) and Bozdag (N40.80, E46.84), 7) Pantashara (N41.23, E46.36) and 8) Kvabebi (N41.48, E45.68) and Kushkuna (N41.25, E45.44).

and awaits consistent interpretations, although significant progress has been made recently by geochemical proxies (e.g. strontium ratios, oxygen isotopes) that may clarify the timing and amount of connectivity (Major et al., 2006; Badertscher et al., 2011).

Understanding the fundamental mechanisms and processes that influenced both the geological and historical changes in sea level, connectivity, climate, and environment (e.g. salinity, anoxia, etc.) is also crucial for a coherent understanding of the future economical and sustainable developments in the region. The immensely rich hydrocarbon fields of both the Black Sea and especially the South Caspian Basin are the product of changing interbasinal connectivity, that generated the anoxic source rocks of the Oligocene Maikop Series, the deltaic reservoir rocks of the Pliocene "Productive Series" and the brackish water cap rock of the Plio-Pleistocene Akchagylian clays (e.g., Hinds et al., 2004; Vincent et al., 2010). Paleoenvironmental changes in the region are causing the biodiversity crisis that the Pontocaspian fauna is experiencing today (e.g., (Grigorovich et al., 2003; Popa et al., 2009).

One of the key problems to understand the complex and intertwined geological history of the Pontocaspian region is the absence of reliable stratigraphic correlations between the Black Sea and Caspian Sea and between the lake/marine and continental domains. The lack of open marine faunal assemblages in the Paratethys generally hampers a straightforward correlation to the standard geological time scale, and the presence of mainly endemic faunas resulted in regional time scales for the different Paratethyan subbasins (e.g., Hilgen et al., 2012; Nevesskaya et al., 2003). For the Quaternary, individual time scales have been developed for the Caspian and the Black Sea region, both mainly based on their own characteristic faunal (usually mollusc) assemblages from rich, but local sites (e.g., Yanina, 2014). Cross-correlation has mainly been done on biostratigraphic arguments, because radiometric, magnetostratigraphic and astronomical data are scarce and/or their results controversial. Carbon dating has been thoroughly applied to Holocene rocks (e.g., Yanina, 2014 and references therein), but radiometric datings (K/Ar or Ar/Ar) of older Quaternary rocks are very rare, because of the common lack of intercalated volcanic ashes (e.g., Chumakov and Byzova, 1992). Magnetostratigraphic correlations to the geomagnetic polarity time scale have been widely produced for the lower Quaternary successions (e.g., Molostovsky, 1997). They are more problematic for rocks younger than 780 kyr (last full reversal of the magnetic field), because there this technique only works if it is possible to sample with a resolution that is high enough (< 1 ka) to pick up the reversal excursions of the Brunhes chron (Laj and Channell, 2007; Singer et al., 2014). Consequently, many of the regional stage boundaries are still poorly dated and serious age uncertainties exist between the current geological time scales for the region.

Here, we present a comprehensive overview of the existing stratigraphic and geochronologic data for the Caspian Sea, the Black Sea and adjacent continental domains. The main result will be an update of the Quaternary geological time scale for the Pontocaspian region which will allow better understanding of the succession of geological events, especially dealing with the major changes in interbasinal connectivity and faunal evolution. We will furthermore discuss the state-of-the-art on the existing techniques and mechanisms that allow a better understanding of Quaternary paleoenvironments, interbasinal connectivity, fauna migration patterns and sea level change in these long-lived anomalohaline lake systems, and the corresponding environmental changes in the terrestrial counterparts.

2. Quaternary time scales of the Pontocaspian domain

2.1. The Caspian Sea region

The Caspian Sea is a lake: it is the world's largest endorheic water body (Fig. 2) extending over 1200 km in latitude (36°-47°N), and 195-435 km in longitude (46°-56°E). The surface area and water volume of the Caspian Basin critically depend on the regional hydrological balance. The Caspian Sea is divided into three subbasins of roughly similar surface area, but widely differing in depth and volume. The North Caspian Basin (< 15m deep, 1% volume) is separated by the Mangyshlak sill from the Middle Caspian Basin (< 800m, ~33%) which is in turn separated by the Apsheron threshold from the South Caspian Basin (< 1025m, ~66%) (Panin et al., 2005; Zonn et al., 2010). At present, Caspian water level is ~27 m below global sea level, which gives a surface area of ~371,000 km^2 and volume of ~78,200 km^3. The Caspian Basin is a huge reservoir of anomalohaline (often referred to as brackish) water. It is highly sensitive to climatic changes in its catchment area (3.5 million km^2), which extends far northward to the central part of the East European Plain (Panin et al., 2005; Zonn et al., 2010). During the Quaternary, the catchment extended to include almost entire Central Asia. Today, the Caspian catchment contains forests and steppes in the Volga and Ural valleys and mountainous forests and arid regions in the Caucasus and Transcaspian areas. The salinity of the present-day Caspian Sea changes from 1‰ near the Volga delta in the north to 13.5 ‰ in the south (Dumont, 1998). The Volga discharge provides 85-90% of the total fresh water influx and forms the main element in the hydrological budget (Agapova and Kulakova, 1973; Mamedov, 1997; Zonn et al., 2010).

In late Miocene (Pontian) times the Caspian Basin was still connected to the Black Sea, forming the final phase of the ancient Paratethys Sea (Popov et al., 2004, 2006; Krijgsman et al., 2010). The Caspian Basin became isolated from the Black Sea in the earliest Pliocene (Van Baak et al., 2016a), when a major drop in its water level resulted in far southward retreat of lake environments and an associated progradation of the Volga's fluvio-deltaic deposits that reached the South Caspian Basin (Fig. 3a). This so-called Productive Series is the main South Caspian hydrocarbon reservoir unit (e.g., Hinds et al., 2004). The nature and extent of lake conditions in the southern basin at the time are unknown as these deposits are often several km below surface. Since its isolation from the Black Sea and open ocean, the Caspian Basin has experienced numerous transgressions and regressions with water level fluctuations of several tens to hundreds of meters resulting in enormous changes of its shoreline, especially in the flat northern part (Varuschenko et al., 1987; Svitoch, 2010a; Yanina, 2014). Our review of the Caspian stratigraphy starts with the late Pliocene Akchagylian transgression that resulted in the first of several lake phases that extend all the way to the modern Caspian Sea. Here we review the definitions and commonly used Plio-Quaternary stratigraphic subdivisions in the Pontocaspian domain, including their geochronological constraints, which allow a detailed correlation of the Caspian Sea with the Black Sea and the terrestrial and global ocean records.

2.1.1. Akchagylian

2.1.1.1. Description. During the Akchagylian age (late Pliocene-earliest Pleistocene) the largest Caspian transgression occurred, with shores extending well into the middle Volga and southern Urals to the north as well as the Sea of Azov in the west and the Aral Sea in the east (Fig. 3b). The Caspian Basin was a saline lake with major endemic faunal radiations but also events occurred that saw the introduction of marine foraminifera. The widespread fine grained intervals of the Akchagylian form the hydrocarbon cap rocks in many places in the Caspian Basin.

The Akchagylian stage was defined by Andrusov (1912), who described these deposits from the Krasnovodsk Peninsula on the eastern side of the Caspian Sea (Andrusov, 1889, 1902, 1906) and named this stage after the Akchagylian district (Fig. 2). Andrusov did not designate a stratotype, and later the Ushak well was proposed as lectostratotype (a stratotype used to describe a stratigraphic unit that does not have a sufficient holostratotype). The stratigraphic profile of the Ushak well was thoroughly studied by many authors; the Akchagylian deposits are around 100 m thick but the base is not exposed (Dvali et al., 1932; Ali-Zade, 1961; Cheltsov, 1965; Nevesskaya, 1975a; Danukalova, 1996). In the stratotype area, Akchagylian deposits transgressively overlie Neogene red-colored rocks and are in turn overlain by Apsheronian limestones (Ali-Zade, 1961). In the Ushak profile, Ali-Zade (1961) and Cheltsov (1965) determined: 1) a 21 m lower unit consisting of alternations of calcareous clays, shelly limestones, calcareous sandstones and two thin layers of volcanic ashes, 2) a 25 m middle unit comprising grey sandstones with ferruginous concretions, and 3) a 59 m upper unit consisting of an alternation of clays, siltstones, sandstones and shelly limestones. Because the Akchagylian in the Ushak well section has no clear boundaries with underlying and overlying units, Danukalova (1996) suggested the Malyi Balkhan section in western Turkmenistan as neostratotype (a stratotype designated after the holo/lectostratotype as a replacement for it) and the Pyrnuar section as hypostratotype (an additional stratotype, in different geographic context). The Pantashara and the Bozdag/Duzdag sections in Transcaucasia (Azerbaijan) were also proposed as neostratotypes (Danukalova, 1996).

The Akchagylian deposits are widespread in the Caspian area and their distribution shows the maximum extent of the Akchagylian transgression (Fig. 3b). The maximum thickness of the Akchagylian (up to 750 m) is reached in the deep Caspian basins whereas it is generally much thinner (0.5-10 m) in the periphery of the basin. In several sections of Azerbaijan (Lokbatan, Jeirankechmez) the Akchagylian overlies the deltaic deposits of the Productive Series of the South Caspian Basin (Van Baak et al., 2013; Van Baak, 2015). In eastern Georgia, (e.g., Kvabebi section), the Akchagylian is deposited above the so-called pre-Akchagylian unconformity (Buleishvili, 1960), indicating syntectonic deformation in the western Kura Basin (see Adamia et al., 2017). It is angularly unconformable and paraconformable at some points. Akchagylian deposits are overlain by conglomerate and sandy units there. Palynological data of the northern regions show dominating taiga forests (Kuznetsova, 1966) whereas the southwestern regions contain pollen that indicate open steppe landscapes (Kovalenko, 1971; Naidina and Richards, 2016).

The Akchagylian mollusc faunas are best characterised by the high number of endemic mactrid and cardiid bivalve species (Danukalova, 1996). The Akchagylian is often subdivided into three substages, based on its mollusc assemblages (Golubyatnikov, 1904, 1908; Kolesnikov, 1940; Ali-Zade, 1954; Yakchemovich et al., 1970; Paramonova, 1994). The lower substage is marked by low variety of genera and species containing *Aktschagylia subcaspia, A. karabugasica, A. inostrantzevi, Cerastoderma dombra dombra*. The middle substage is characterized by high species numbers within the genera *Aktschagylia, Andrussovicardium, Miricardium* and *Avicardium*. The upper substage is characterized by low numbers of mollusc species including *Aktschagylia subcaspia, A. ossoskovi, Cerastoderma dombra dombra* and species of *Dreissena* and *Theodoxus*. The gastropods *Pirenella caspia* s.l., *"Clessiniola" intermedia, "C." utvensis,* and *"C." vexatilis* occur in all substages. The usage of this threefold scheme may lead to an arbitrary allocation of units, especially as it is only based on high or low mollusc species richness. For example, other investigations demonstrated that middle Akchagylian molluscs appeared at different stratigraphical levels, which led to an alternative subdivision of the Akchagylian in two substages (Popov, 1969; Trubikhin, 1977; Danukalova, 1996; Nevesskaya et al., 2003).

The lower part of the Akchagylian is marked by a sudden occurrence of euryhaline marine foraminifera like *Cassidulina crassa, C. prima,* possibly *C. reniforme,* and *C. obtusa, Ammonia beccarii, Cibicides lobatulus, Spirillina sp., Discorbis multicameratus, Miliolina aksaica*

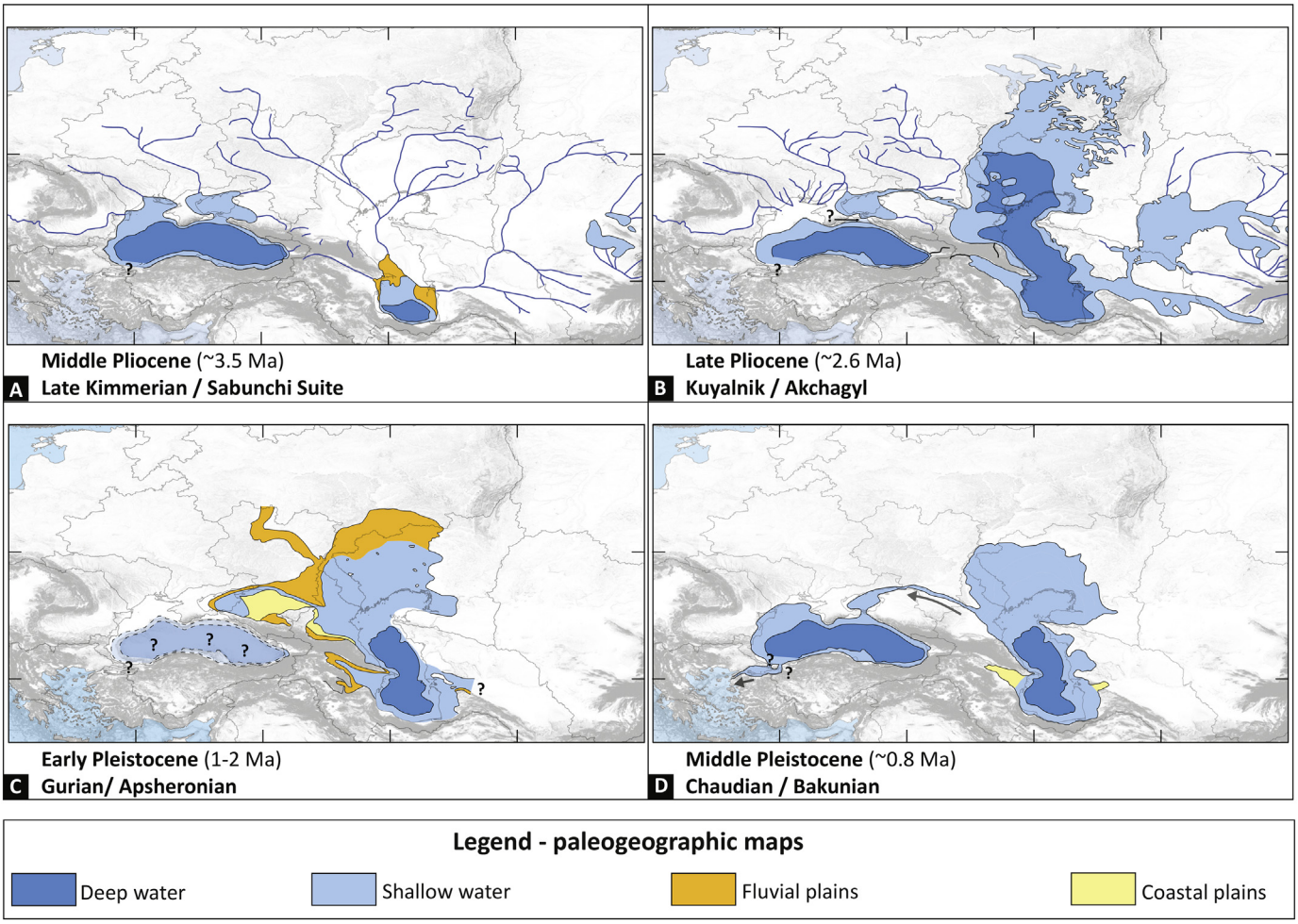

Fig. 3. Paleogeographic maps for the Plio-Pleistocene Pontocaspian region. A) Middle Pliocene; B) Late Pliocene; C) Early Pleistocene; D) Middle Pleistocene. Based on Vinogradov, 1961, 1969 and Abdurakhmanov et al. (2002).

(Agalarova, 1976; Richards et al., 2018) as well as small sized biserial Bolivinidae. The latter are nowadays considered to belong to the planktonic genus *Streptochilus* (Smart and Thomas, 2006, 2007). During this marine influx, the freshwater dominated Pliocene ostracod fauna became replaced by an anomalohaline ("brackish") water assemblage consisting of *Limnocythere alveolata*, *L. luculenta*, *L. tschaplyinae*, *Typhlocypris gracilis, Loxoconcha eichwaldi, Candona candida*, and *C. combibo* (Van Baak et al., 2013; Fig. 4). The upper part of the Akchagylian is characterized by the presence of *Eucythere naphtatscholana, Amnicythere andrussovi, A. nata, A. multituberculata*, *A. cymbula, Leptocythere gubkini, Euxinocythere praebosqueti, Loxoconcha babazananica,* and *Camptocypria acronasuta*.

2.1.1.2. Correlation. The age of the Akchagylian is mainly based on paleomagnetic investigations and correlation to the Geomagnetic Polarity Time Scale (GPTS) in combination with radio-isotopic dating. The interpreted ages of the boundaries are subject of debate. One of the first paleomagnetic results was obtained from western Turkmenistan and indicated that the onset of the Akchagylian in the South Caspian Basin started slightly below a normal-reversed boundary (Khramov, 1960, 1963). We consider this now the Gauss-Matuyama boundary, which corresponds to an age of 2.58 Ma. Note here that the first geomagnetic polarity time scale ever published, appeared later (Cox et al., 1963). Magnetostratigraphic investigations of numerous sections in Turkmenistan and Azerbaijan confirmed this N-R polarity pattern straddling the base of the Akchagylian (Trubikhin, 1977). The Pyrnuar section in western Turkmenistan, however, showed two small reversed zones in the lower Akchagylian, which were correlated with chrons C2An.2r (Mammoth) and C2An.1r (Kaena) by Trubikhin (1977). Consequently, the base of the Akchagylian was matched to the base of the Gauss chron (C2An) at an age of 3.6 Ma, a correlation that has been accepted in numerous publications (Trubikhin, 1977; Semenenko and Pevzner, 1979; Sidnev, 1985; Molostovsky, 1997). Recently, Gurarii (2015) re-investigated the polarity pattern of the Pyrnuar section and concluded that the two reversed zones are too small to be interpreted as Mammoth and Kaena and that the base of the Akchagylian correlates better with the upper part of the Gauss chron (between 3.0 and 2.6 Ma).

In the northern Pre-Caspian region, the base of the Akchagylian Stage has been placed at the lower Nunivak chron (C3n.2n) at an age of ~4.5 Ma, but the main Akchagylian transgression event there was also dated at the Gauss-Matuyama boundary (Yakhemovich et al., 1981, 2000). Hence, it was suggested to correlate the lower/upper substage boundary with the Gauss/Matuyama reversal (Nevesskaya et al., 2005). The Akchagylian deposits of the Azov region and the Northern Greater Caucasus only show a single normal to reverse polarity change, interpreted as the Gauss-Matuyama boundary, although the base Akchagylian is tentatively placed at the Gilbert-Gauss reversal at 3.6 Ma (Naidina and Richards, 2016).

The ash layers from the lowermost Akchagylian of Azerbaijan have been dated by Chumakov (Chumakov et al., 1988) using a fission track method at 3.34 ± 0.5 Ma, although they later reported that these ages were not very accurate and had to be confirmed by new data (Chumakov et al., 1992). Magnetostratigraphic data from the Lokbatan

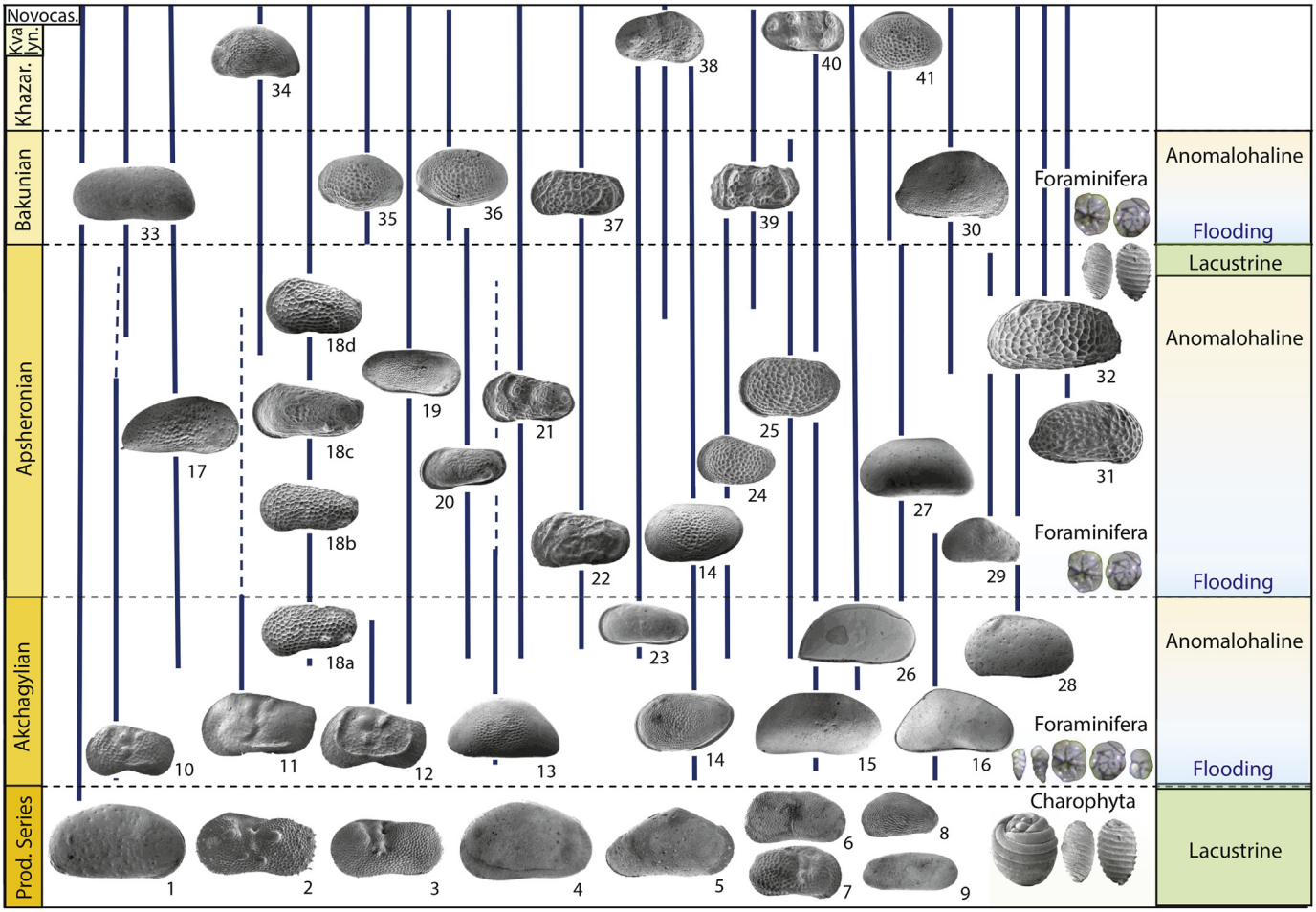

Fig. 4. Distribution of the main ostracod species, foraminifera and charophyta in the Caspian Basin during the Plio-Pleistocene, based on literature data. 1) *Cyprideis torosa*; 2) *Ilyocypris gibba*; 3) *I. bradyi*; 4) *Cyprinotus salinus*; 5) *Eucypris* sp.; 6) *Pseudocandona compressa* (juvenile); 7) *Limnocythere aralensis*; 8) *Zonocypris membranae*; 9) *Darwinula stevensoni*;10) *Limnocythere alveolata*; 11) *L. luculenta*; 12) *L. tschaplyinae*; 13) *Typhlocypris gracilis*; 14) *Loxoconcha eichwaldi* (different stages of evolution); 15) *Candona candida*; 16) *C. combibo*; 17) *Eucythere naphtatscholana*; 18a) *Amnicythere andrussovi;* 18b) *A. palimpsesta*; 18c) *A. normalis*; 18d) *A. saljanica*; 19) *Amnicythere nata*; 20) *Leptocythere gubkini*; 21) *A.multituberculata*; 22) *Euxinocythere praebosqueti*; 23) *A. cymbula*; 24) *Loxoconcha babazananica*; 25) *L. petasa*; 26) *Camptocypria acronasuta*; 27) *Caspiocypris filona*; 28) *Tyrrhenocythere bailovi*; 29) *Xestoleberis chanakovi*; 30) *T. amnicola donetziensis*; 31) *T. azerbaidjanica*; 32) *T. papillosa*; 33) *Camptocypria acronasuta*; 34) *Bacunella dorsoarcuata*; 35) *Loxoconcha gibboides*; 36) *L. lepida*; 37) *Euxinocythere bosqueti*; 38) *Cytherissa bogatschovi*; 39) *Euxinocythere bacuana*; 40) *A. quinquetuberculata*; 41) *Loxoconcha endocarpus*

section were not straightforward but led to the conclusion that the lower Akchagylian boundary should be around ~3.2 Ma (Van Baak et al., 2013). This was in good agreement with results from the Kvabebi section in eastern Georgia where the uppermost two normal subchrons of Gauss were attributed to the Akchagylian (Agustí et al., 2009). In the Kushkuna section of Azerbaijan, however, the lowermost Akchagylian only contains one normal polarity zone, correlative to the top Gauss (Trubikhin, 1977). Recently, the Akchagylian ash layers of the Lokbatan and Jeirankechmez sections of Azerbaijan were radio-isotopically dated with the $^{40}Ar/^{39}Ar$ method, which resulted in a revised age of 2.7 Ma for the base of the Akchagylian in Azerbaijan (Van Baak, 2015). This is in excellent agreement with the magnetic polarity pattern of the Jeirankechmez section that shows the two reversed subchrons of the Mammoth and Kaena in the upper part of the Productive Series (Khramov, 1963; Van Baak, 2015). The earlier magnetostratigraphic correlation of Lokbatan was thereby rejected (Van Baak, 2015).

In conclusion, most evidence and interpretations converge towards two different ages for the base of the Akchagylian (Fig. 5). The “classic option” mainly relies on the magnetostratigraphic correlation of the Pyrnuar section and dates the base of the Akchagylian to the base of the Gauss at an age of 3.6 Ma. This is the officially accepted age in Russian stratigraphy (Provisions, 2003; Nevesskaya et al., 2005). The “young option” mainly depends on sections in Azerbaijan, where the integration of $^{40}Ar/^{39}Ar$ dating and magnetostratigraphy indicates an age of 2.7 Ma for the base of the Akchagylian, in the uppermost normal part of the Gauss chron (Khramov, 1960; Van Baak, 2015).

2.1.2. Apsheronian

2.1.2.1. Description. Caspian Sea conditions in the Early Pleistocene Apsheronian were similar to today. The Caspian Basin was occupied by an anomalohaline lake. A major salinity decrease during the late Akchagylian corresponds with the almost complete extinction of characteristic Akchagylian bivalve faunas and subsequently the typical Apsheronian fauna evolved. This endemic fauna became increasingly dominated by extant endemic Caspian groups. During the Apsheronian, the Caspian Basin was mostly an isolated basin that may have had rare, short-lived connections only to the Black Sea via the Manych Strait (Fig. 3c).

The Apsheronian stage was established based on outcrops on the Apsheronian peninsula (Fig. 2) near Baku (Azerbaijan), and the Bailov Cap district was proposed as lectostratotype locality (Andrusov, 1923; Nevesskaya, 1975b). There, the Apsheronian deposits conformably overlie the Akchagylian deposits and are overlain, with an angular unconformity, by Bakunian strata (Nevesskaya, 1975b; Stratigraphy,

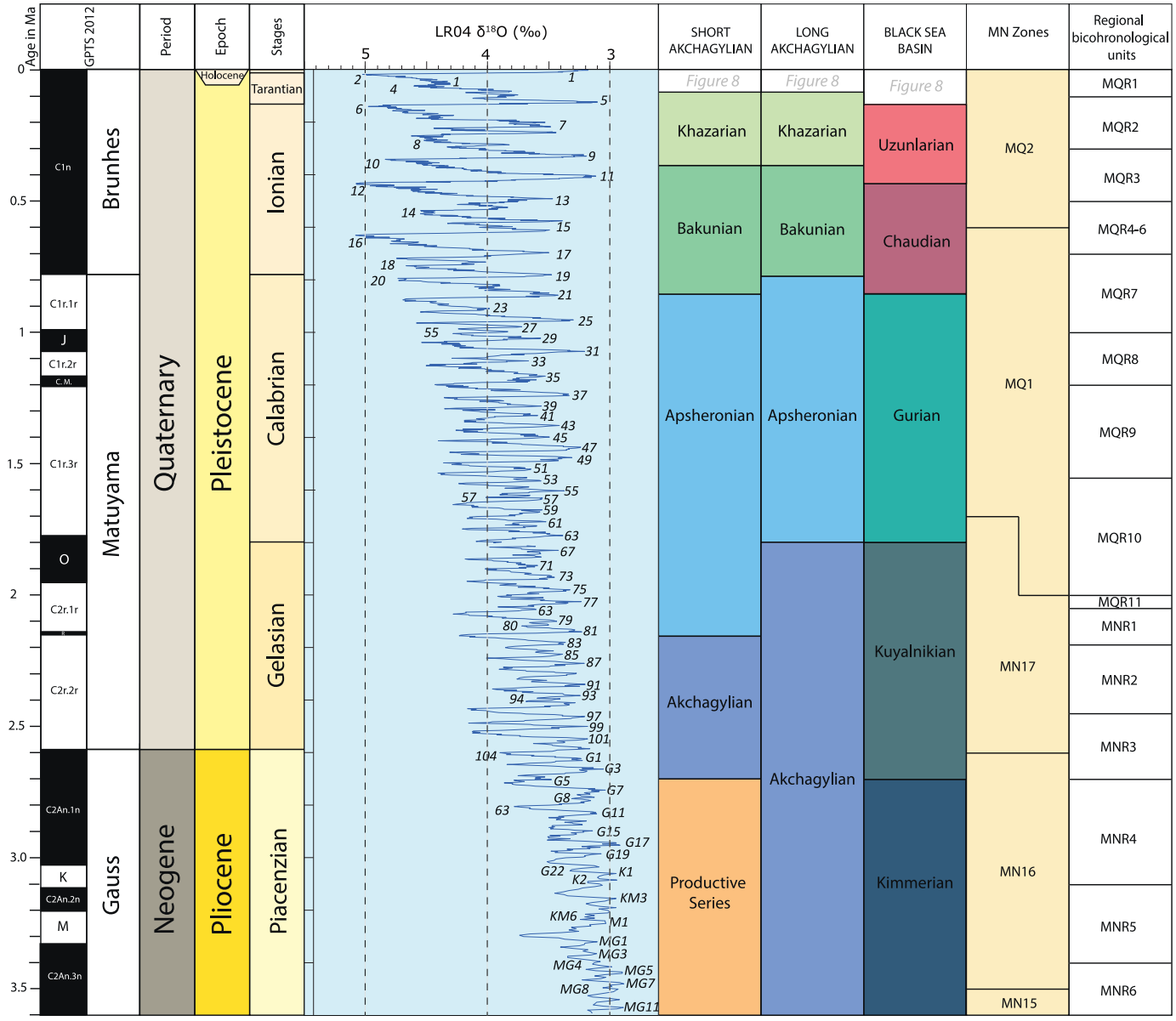

Fig. 5. Plio-Quaternary time scales for the Pontocaspian domain. GPTS with Systems and Stages are after Hilgen et al. (2012), oxygen isotope curve with numbered Marine Isotope Stages (MIS) is after Lisiecki and Raymo (2005). M = Mammoth (C2An.2r), K = Kaena (C2An.1r), R = Reunion (C2r.1n), O = Olduvai (C2n), C.M. = Cobb Mountain (C1r.2n), J = Jaramillo (C1r.1n). Ages of the Cobb Mountain Subchron are after Channell (2017). On the right side are the time scales for the Caspian Basin, Black Sea Basin and the terrestrial domain (mammal (MN/MQ) zonation and regional (MNR/MQR) biochronological units).

1982).

The size of the Caspian Sea slightly decreased compared to the Akchagylian, so Apsheronian deposits are, almost everywhere, found conformably overlying Akchagylian strata (Stratigraphy, 1982). Palynological data mainly reveal treeless landscapes, indicating an intensification of the continental climate and an increasing aridity when compared to the Akchagylian (Naidina and Richards, 2016).

The Apsheronian Stage is subdivided into three substages, based on changes in the composition of the mollusc fauna (Andrusov, 1923; Kolesnikov, 1940; Zhidovinov et al., 2000; Sidnev, 1985). The lower substage is characterized by a species-poor assemblage of low saline to fresh water bivalves (*Dreissena, Corbicula, Apsheronia*) and gastropods (*Lymnaea, Streptocerella, Turricaspia, Theodoxus*). The middle substage is marked by the first occurrence of the mollusc genera *Parapsheronia* and *Didacna*. The upper zone is characterized by a general depletion of the saline fauna and the disappearance of the ribbed apsheronids of the middle interval. This upper (Tyurkyanian) interval is poorly recognizable in many parts of the Caspian Basin and in some regions only a twofold division has been used (Stratigraphy, 1982).

Euryhaline foraminifera of the genus *Ammonia* spp. have been observed at the base of the Apsheronian, suggesting a brief increase in salinity. Most of the ostracod species from the Akchagylian continued to evolve within the Apsheronian stage but became more frequent. The Apsheronian comprises the ostracod genera *Leptocythere* and *Caspiolla*; the middle Apsheronian is marked by a widespread radiation within the genus *Leptocythere* (Fig. 4). *Tyrrhenocythere azerbaidjanica*, *T. papillosa*, *T. bailovi*, *Cyprideis torosa* and *Cytherissa bogatschovi* are also commonly observed (Fig. 4).

The Tyurkyanian stage was defined by Khain (1950) for the continental deposits (up to 100 m) between the Apsheronian and Bakunian successions (Stratigraphy, 1984). These deposits contain fresh water molluscs (*Viviparus diluvianus*, *Valvata piscinalis*, *V. antiqua, Bithynia* sp., *Lithoglyphus naticoides*, *Pisidium amnicum*, *P.* cf. *supinum*, *P.* cf. *subtruncatum, Sphaerium rivicola*, *Unio* sp., *Dreissena* sp.) and correspond to

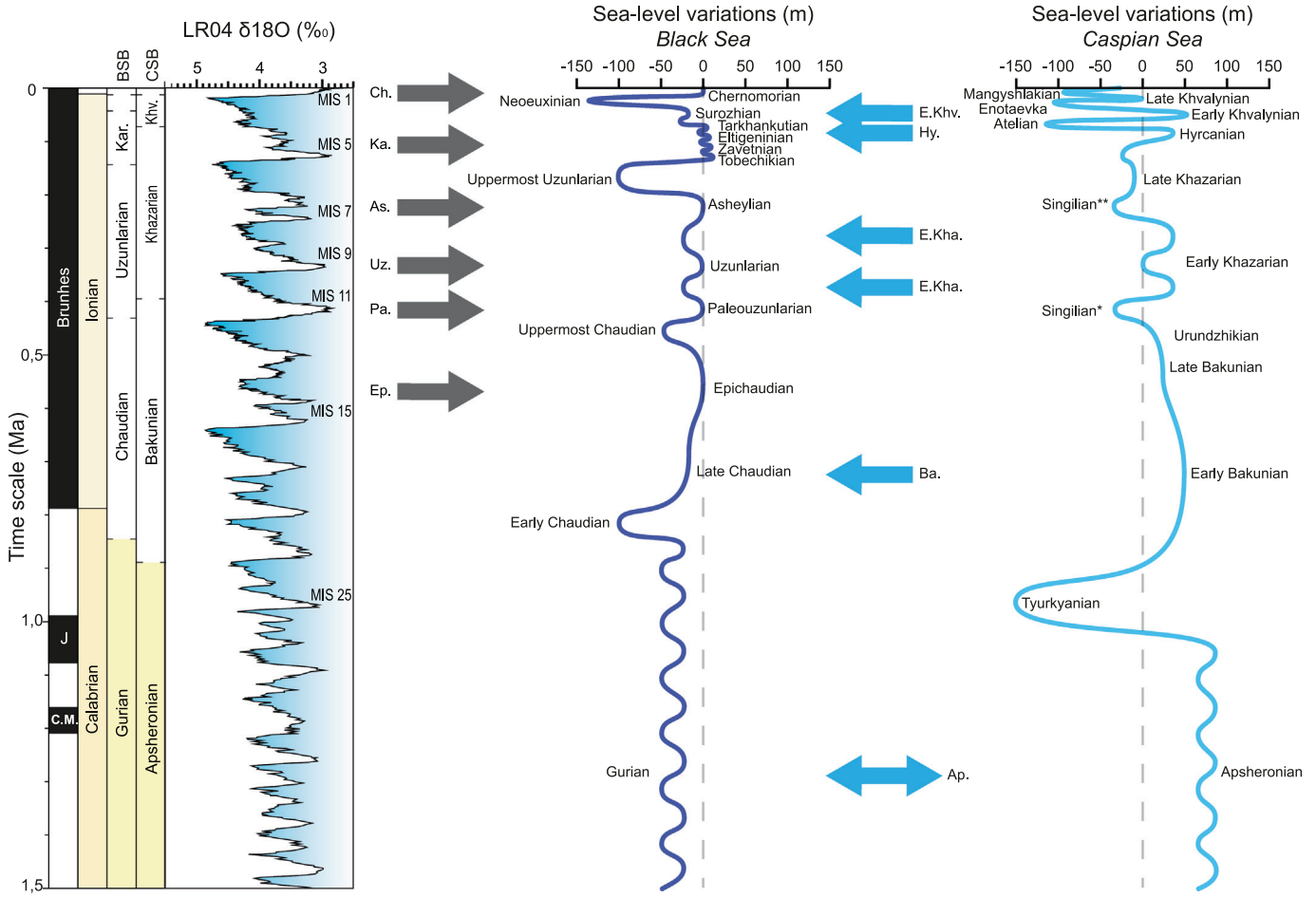

Fig. 6. Schematic reconstruction of the Black Sea and Caspian Sea water-level curves in comparison to global oxygen isotopes records of Lisiecki and Raymo (2005) during the Pleistocene to Holocene. Associated interbasinal water exchanges marked by arrows pointing to the right for Mediterranean waters flooding into the Black Sea, arrows pointing to the left for Caspian Sea waters flooding into the Black Sea and double arrows for bidirectional water exchange between Black Sea and Caspian Sea. N.B. Two options exist in literature regarding the position of the Singilian: *Svitoch (2013b and references therein) and **Zastrozhnov et al. (2018 and references therein).

a strong (~150 m; Fig. 6) regressional phase (Danukalova et al., 2016; Svitoch, 2013a; Svitoch, 2013b; Yanina, 2012a; Yanina, 2012b; Zastrozhnov et al., 2018). In the fluvio-lacustrine Tyurkyanian deposits, ostracods typical of shallow fresh water environments are dominant (e.g., *Ilyocypris bradyi, I. gibba, Pseudocandona compressa, Candona neglecta, Cypris subglobosa, Darwinula stevensoni, Eucypris* sp.). The Tyurkyanian deposits are well-expressed in the eastern margin of the South Caspian Basin and are locally present in the North Caspian Basin where they are known from drill cores and have a maximum thickness of 38 m (Danukalova et al., 2016). Here, we consider the Tyurkyanian as the upper part of the Apsheronian.

2.1.2.2. Correlation. An intensive paleomagnetic campaign to study the Apsheronian deposits in the Pontocaspian region took place in between 1960 and 1980 (Khramov, 1960, 1963; Trubikhin, 1977; Kochegura and Zubakov, 1978; Semenenko and Pevzner, 1979; Yakhemovich et al., 1981). The magnetostratigraphic correlations of sections in Azerbaijan and Turkmenistan were straightforward and have placed the Akchagylian/Apsheronian boundary at the base of a normal polarity interval, that was assumed to correspond to the Olduvai subchron (C2n), corresponding to an age of 1.95 Ma (Kochegura and Zubakov, 1978; Semenenko and Pevzner, 1979; Sidnev, 1985). Recently, multidisciplinary investigations based on integration of paleomagnetic, Ar/Ar dating, ostracod faunal analysis and lithological changes have been conducted again in the Kura Basin of Azerbaijan and found the base of the Apsheronian close to, but slightly below the base of the Olduvai (Van Baak et al., 2013). Paleomagnetic data from the Duzdag section in Azerbaijan confirm that the base of the Apsheronian corresponds to a normal polarity interval (Pevzner, 1986). The magnetic polarity pattern at Duzdag, however, indicates that this normal polarity interval correlates to the Reunion subchron rather than to the Olduvai, giving a slightly older date of ~2.1 Ma for the basal Apsheronian.

Further revisions of the boundary have been proposed based on the new definition of the Neogene-Quaternary boundary. In 1998, the Interdepartmental Stratigraphic Committee of Russia (ISC) established the Neogene-Quaternary boundary on Russian territory at the top of the Olduvai subchron corresponding to an age of 1.8 Ma (Provisions, 1998). With an aim of correlating the Russian national stratigraphic chart to the global Geological Time Scale, the Akchagylian/Apsheronian boundary was made equal to the Neogene-Quaternary boundary, and therefore arrived at an age of 1.8 Ma. Since this time, many time scales in the Pontocaspian region place the base of the Apsheronian at the top of the Olduvai subchron (Nevesskaya et al., 2005).

The Tyurkyanian deposits are suggested to correspond to the lowermost part of the Middle Pleistocene (Semenenko and Pevzner, 1979; Trubikhin, 1977). According to fission track data, however, the age of the Tyurkyanian is 1050 to 950 ka (Ganzey, 1984).

In conclusion, two different ages exist for the top of the Akchagylian/base of the Apsheronian (Fig. 5). The official geological

time scale of Russia dates the Akchagylian/Apsheronian boundary at 1.8 Ma, resulting in a "long Akchagylian" of 1.8 Myr. The alternative correlation, based on the Duzdag section of Azerbaijan, dates this boundary at > 2.1 Ma, suggesting a "short Akchagylian" of 0.6 Myr if the age of 2.7 for the base Akchagylian in Azerbaijan is used, or an "intermediate Akchagylian" of 1.5 Myr if the age of 3.6 Ma for the base of the Akchagylian is used (Fig. 5).

2.1.3. Bakunian

2.1.3.1. Description. Conditions in the Middle Pleistocene Bakunian stage resembled the modern Caspian Sea in size, fauna and salinity regimes (Fig. 3d). During highstands punctuated overflows towards the Black Sea existed. Widespread carbonate rocks formed during warmer phases of the Bakunian in the South Caspian Basin; these are the main source for building material there.

The Bakunian Stage was defined by Sjögren (1891) and comprises the sedimentary strata that conformably overlie the Apsheronian deposits in the eastern part of the Apsheron peninsula of Azerbaijan (Fig. 2). Sjögren (1891) did not propose any stratotype and Golubyatnikov (1904) suggested the section Gora Bakinskogo Yarusa on the Apsheron Peninsula, which was later thoroughly studied by other researchers (Nalivkin, 1914; Fedorov, 1957; Mamedov and Aleskerov, 1988; Svitoch et al., 1992; Svitoch and Yanina, 1997; Yanina, 2005). The Neftyanaya Balka section was suggested as an additional reference section (Yanina, 2012b, 2013). Neftyanaya Balka is situated in the Kura Basin and has the advantage that it contains well defined stratigraphic boundaries for the Bakunian stage. The mollusc fauna of the Bakunian stratotype is characterised by a number of bivalve species including *Didacna parvula, D. catillus, D. rudis, D. carditoides, D. eulachia, D. mingetschaurica, D. pravoslavlevi, D. lindleyi*. The taxa *D. parvula* and *D. catillus* are index-species, while *D. rudis* and *D. carditoides* are considered characteristic species (Bogachev, 1932a, 1932b; Fedorov, 1957, 1978a; Vekilov, 1969; Svitoch et al., 1992; Yanina, 2005, 2013; Nevesskaya, 2007; Svitoch and Yanina, 2007).

Sediments of the Bakunian are widespread on most Caspian coasts. In tectonic depressions, Bakunian deposits are deeply buried, and are only exposed in local uplifted structures. Sediments represent a range of facies and their thickness varies between a few meters on high terraces, to hundreds of meters (maximum thickness ~500 m) in the depressions. Spores and pollen indicate a relatively cold and humid climate in the early Bakunian, with forest formation (birch, alder, oak, maple, elm) on the western coast (Abramova, 1974; Filippova, 1997) and forest-steppe assemblages (pine, birch, alder, elm) in the lower Volga region (Moskvitin, 1962). The late Bakunian was marked by moderately warm and humid climate, as expressed by forest-steppes (oak, caracas, hophornbeam) in the Kura Basin (Svitoch et al., 1998), widespread evergreens in Dagestan (Abramova, 1974), and steppes in the valleys of the Ural river (Yakhemovich et al., 1986).

The base of the Bakunian is determined by a significant transgression that reached its maximum extent in the first half of the Middle Pleistocene (=lower Neopleistocene), but was much smaller than the Akchagylian transgression (Fig. 3c). The presence of Bakunian *Didacna parvula, D. rudis, D. carditoides, D. catillus* in the Black Sea (Chaudian Stage) and in the Manych Depression (Fedorov, 1978a; Popov, 1983; Svitoch et al., 1998, 2010; Yanina, 2005, 2006) represents Caspian overflows into the Black Sea Basin through the Manych Strait.

The Bakunian Stage is commonly subdivided into two substages based on different mollusc fauna, in particular the bivalve genus *Didacna* (Fedorov, 1957, 1978a; Moskvitin, 1962; Vekilov, 1969; Popov, 1983; Mamedov and Aleskerov, 1988; Yanina, 2005, 2012b, 2013; Nevesskaya, 2007). Three morphological groups of the *Didacna* genus exist: the *crassoidal, catilloidal* and *trigonoides* groups (Svitoch, 1967; Yanina, 2005, 2013; Nevesskaya, 2007). The faunas of the lower Bakunian are dominated by the first two groups and furthermore includes taxa like *Didacna parvula, D. catillus* and *D. fedorovi*. In addition, *Dreissena rostriformis* is widely distributed. The most characteristic species are *Didacna parvula* and *D. catillus*. The upper Bakunian predominantly contains didacnas of the transitional crassoidal-catilloidal group (*D. rudis, D. carditoides*) and crassoidal group (*D. eulachia, D. mingetchaurica, D. pravoslavlevi, D. bacuana*) and representatives of the catilloidal and trigonoidal groups are rare. The most characteristic species are *Didacna rudis, D. carditoides* and *D. eulachia* (Yanina, 2005, 2013; Nevesskaya, 2007).

The transgressive base of the Bakunian is marked by a level rich in euryhaline foraminifera, represented mainly by *Ammonia* spp., *Nonion* sp. and *Cibicides lobatulus*. The Bakunian ostracod community is dominated by endemic species adapted to unusual salinity settings such as *Eucythere naphtatscholana, Loxoconcha eichwaldi, L. petasa, L. babazananica, Tyrrhenocythere azerbaidjanica, T. papillosa, T. amnicola donetziensis, Cytherissa bogatschovi* and *Bacunella dorsoarcuata* (Fig. 4). Furthermore, many leptocytherid and candonid species occur. During the Bakunian, a morphological transformation of the carapace features took place, especially in loxoconchids. Several early stage morphotypes evolved into new species that characterize the present-day Caspian Sea ostracod fauna, including *Loxoconcha endocarpus*, *L. lepida* and *L. gibboides*.

The Urundzhikian stage has been defined by Fedorov (1946) as an independent stratigraphic unit corresponding to the final stage of the Bakunian transgressive cycle (Fedorov, 1993, 1999). The stratotype of the Urundzhikian is the section Yuzhny Urundzhik in Western Turkmenistan (Fedorov, 1946). As an additional reference section Neftyanaya Balka in Azerbaijan was suggested, which contains well defined stratigraphic boundaries (Svitoch and Yanina, 2007). Urundzhikian deposits are known from the southern Caspian Basin only (Kura Basin, Apsheron peninsula and southwest Turkmenistan). They are characterized by numerous *Didacna* species: *D. eulachia, D. mingetschaurica, D. pravoslavlevi, D. colossea, D. shirvanica, D. bergi, D. karelini, D. porsugelica, D. čelekenica, D. rudis,* and *D. carditoides*. Trigonoidal and catilloidal didacnas are rare. Characteristic species are *D. eulachia, D. pravoslavlevi* and *D. kovalevskii*. The size of the endorheic Urundzhikian lake phase slightly exceeded the area of the modern Caspian Sea. Here, we consider the Urundzhikian as the upper part of the Bakunian.

2.1.3.2. Correlation. The beginning of the Bakunian transgression is generally considered to correlate with the lower part of the Middle Pleistocene as defined on the International Quaternary Chart (Fedorov, 1978a; Rychagov, 1997; Yanina, 2012a, 2013; Svitoch, 2013a; Zastrozhnov et al., 2018). The age of Bakunian deposits is ~600 ka according to fission track analysis, and 378-480 ka according to thermoluminescent dating (Lavrishchev et al., 2011a, 2011b). Based on palynological data, the lower Bakunian is correlated to isotope stages MIS 18-16 (750-625 Ka) and the upper Bakunian to MIS 15-13 (625-475 ka). The large size and thickness of *Didacna* shells and the high carbonate content of Urundzhikian sediments suggest warm climatic conditions. The Urundzhikian transgression is consequently correlated to the interglacial stage MIS 11 (Svitoch and Yanina, 2007; Yanina, 2012a).

Magnetostratigraphic data indicate that the Bakunian predominantly correlates with the normal Brunhes Chron, which provides a maximum age of 780 ka (Asadullayev and Pevzner, 1973). A recent study of the Xocashen section in western Azerbaijan places the Apsheronian-Bakunian boundary between the Jaramillo and Bruhnes chrons, in a reversed subchron correlative to chron C1r.1r with an age of 0.88-0.85 Ma (Van Baak et al., 2013), but these authors did not distinguish the Tyurkyanian.

In conclusion, the age of the lower Bakunian boundary is constrained to the age interval between 0.88 and 0.75 Ma (Fig. 5). The official geological time scale of Russia places the Apsheronian/Bakunian boundary at the Brunhes/Matuyama reversal at an age of 0.78 Ma.

2.1.4. Khazarian

2.1.4.1. Description. During the Middle-Late Pleistocene Khazarian and

Hyrcanian phases the Caspian Sea experienced large sea level changes and episodic overflow into the Black Sea occurred (Fig. 6). In between these high-stands, deep regressions took place and essentially the modern Caspian system evolved.

The Khazarian Stage has been defined by Andrusov (in Pravoslavlev, 1913) and comprises the sedimentary strata that conformably overlie Bakunian deposits in the North Caspian Basin and the Manych Strait. Fedorov (1953) suggested a twofold subdivision into a lower and upper Khazarian, which is officially accepted in Russian literature. Lower Khazarian deposits in boreholes on the Apsheron peninsula were also called "Gyurgyanian" (Dashevskaya, 1936, 1940), a name that can be found in some stratigraphical schemes (e.g., Nevesskaya, 2007). Many researchers have considered another unit in the northern Pre-Caspian area, the Singilian, either at the base of the lower Khazarian (Pravoslavlev, 1913, 1932; Fedorov, 1957, 1978a; Moskvitin, 1962; Goretskiy, 1966; Rychagov, 1997), or as an independent unit between Bakunian and lower Khazarian intervals (Sedaikin, 1988; Svitoch and Yanina, 1997; Yanina, 2012b; Svitoch, 2013b). In these scenarios the Singilian is considered time-equivalent with the Urundzhikian (Fig. 6). Popov (1983), however, considered the Singilian as the regressive phase of the lower Khazarian. More recently, the regressional Singilian phase was placed between the early and late Khazarian transgressions (Zastrozhnov et al., 2018).

The early Khazarian transgression corresponds to a sea level high stand of +15 m (Fedorov, 1957; Vasiliev, 1961; Rychagov, 1997) or +20-25 m (Svitoch and Yanina, 2007). The lower Khazarian deposits are widespread along the Caspian coast, penetrating far inland along paleodepressions. The sediment thickness ranges from a few meters to thick sequences in the depressions. The maximum thickness is recorded in the Kura Basin, where it reaches 600 m (Vekilov, 1969). Spore and pollen data indicate a three-stage change in vegetation: from goosefoot-sagebrush steppe at the beginning of early Khazarian time, to taiga type forests in its middle and periglacial steppe at the end (Zastrozhnov et al., 2018).

The lower Khazarian is characterized by the broad development of the trigonoidal group of didacnas (*Didacna subpyramidata, D. paleotrigonoides, D. gurganica, D. mishovdagica, D. trigonula, D. trigonoides chazarica*), representatives of the crassoidal group (*Didacna praevoslavlevi, D. nalivkini, D. delenda, D. apscheronica, D. ovatocrassa, D. subcrassa, D. pontocaspia tanaitica*) and rare species of the catilloidal group (*Didacna dilatata, D. subcatillus, D. vulgaris, D. lindleyi, D. adacnoides*). Characteristic species for the lower Khazarian are *Didacna subpyramidata* and *D. paleotrigonoides*. Numerous other anomalohaline molluscs are present as well (*Monodacna caspia, Hypanis plicatus, Adacna vitrea*, etc.). Typical ostracods are *Caspiolla gracilis, Candoniella subellipsoida, Leptocythere arevina, L. martha, L. quinquetuberculata, L. gibboida, Bacunella dorsoarcuata, Loxoconcha endocarpa, Cyprideis littoralis* and *Cytherissa cascusa* (Ushko and Shneider, 1960; Sedaikin, 1988; Svitoch and Yanina, 1997). Among foraminifera, *Ammonia caspica* and *Mayerella brotzkajae* occur (Yanko, 1989).

The early Khazarian spans several interglacials and glacials. Khazarian deposits of the Manych Depression include two overflow phases (Fedorov, 1978a; Svitoch and Yanina, 1997; Yanina, 2005, 2012a). The early Khazarian transgressions mostly developed under relatively cold climate conditions but some of the transgressions developed during interglacial intervals. Traces of permafrost are found in deposits of the Northern Pre-Caspian region, which represented a periglacial interval during glacial intervals (Grichuk, 1954; Moskvitin, 1962; Zhidovinov et al., 1984). Deciduous woods developed in Dagestan (Abramova, 1974). Glaciers existed in the Caucasus mountains and cold and humid conditions existed in Azerbaijan during these glacials (Milanovsky, 1966; Dumitrashko et al., 1977; Aleskerov, 1990). Lower Khazarian spores and pollen from boreholes in the Southern Caspian are dominated by grassy plants and trees (pine, fir, birch, alder, willow) (Vronsky, 1976).

The late Khazarian transgression (Fig. 7a) corresponds to a sea level high stand of -10 m (Kaplin et al., 1977a; Popov, 1983; Svitoch and Yanina, 1997), but lacked any connection with the Black Sea Basin (Fig. 6). Common gigantism of shells, high carbonate content in the sediment, and the presence of oolites indicates warm climate conditions (Yanina, 2014). Salinities ranged between 10 to 12‰ in the northern part and up to 14-15‰ in the southern part of the Caspian Basin, i.e. higher than today (Yanina, 2014). Pollen assemblages testify to moderately warm interglacial climates (Abramova, 1974).

The upper Khazarian is marked by widespread species of the crassoidal *Didacna* group (*Didacna surachanica, D. subcrassa, D. hyrcana, D. nalivkini, D. delenda, D. ovalis, D. karabugasica, D. subovalis, D. ovatocrassa, D. schuraosenica*). Trigonoidal and cattiloidal forms are rare. The index species of the upper Khazarian is *Didacna surachanica* (Fedorov, 1957; Yanina, 2005; Nevesskaya, 2007). In more oligohaline and freshwater deposits *Corbicula fluminalis* is common.

Foraminifera are represented by *Ammonia novoeuxinica, A. caspica, Elphidium capsicum, E. shochinae, Frorilus trochospiralis, Majerella brotzkajae, Cornuspira minusculla* and *Milioniella risilla* (Yanko, 1989). The Khazarian ostracod fauna is a continuation of the Bakunian fauna (Fig. 4). The most common species are *Candona schweyeri*, *Fabaeformiscandona* sp., *Eucythere naphtatscholana*, *Cyprideis torosa*, *Loxoconcha eichwaldi*, *L. petasa, L. lepida, L. babazananica, Tyrrhenocythere azerbaidjanica, T. papillosa, T. amnicola donetziensis, Euxinocythere bacuana, Amnicythere andrussovi* (and its associated morphotypes), *A. striatocostata, Bacunella dorsoarcuata* and *Cytherissa bogatschovi* (Sedaikin, 1988).

After the main late Khazarian transgression, but before the Khvalynian transgression, another transgression has been described from boreholes in the North Caspian Basin; the Hyrcanian (or Girkanian) transgression (Fig. 6; Popov, 1955, 1967; Goretskiy, 1957; Yanina, 2013; Sorokin et al., 2018). Hyrcanian deposits contain "Khvalynian-like" fauna of *Didacna subcatillus*, *D. cristata, D. pallasi, D. subcrassa*, but also the mainly freshwater *Corbicula fluminalis*. The widespread occurrence of *C. fluminalis* is indicative of the warm water character of the basin. The Hyrcanian mollusc fauna in the Manych Depression shows that Caspian waters were draining to the Black Sea (coevally with the last phase of the Karangatian transgression in that basin) through the Manych Strait (Popov, 1983; Yanina, 2014).

The Khazarian sediments (including the Hyrcanian in their upper part) are generally separated from the lower Khvalynian deposits by the Atelian regression (Fig. 6), when the Caspian Sea level was significantly lowered (Fig. 7b). Based on seismic-acoustic profiling, the maximum lowstand during the Atelian is estimated at -120 to -140 m (Lokhin and Maev, 1990; Maev, 1994). Vast areas of the Caspian shelf were exposed and river incisions were deep (Fedorov, 1978a; Rychagov, 1997). Atelian deposits contain mammal remains of the Upper Paleolithic faunal complex, including mammoth, horse, reindeer, etc., indicative of tundra-steppe and cold arid continental climate. Towards the end of the Atelian epoch, the climate became even warmer. In the vegetation, the share of arboreal pollen (birch, pine and spruce trees) increased while elm, oak and linden re-appeared. The importance of xerophytes decreased, while grasses and herbaceous vegetation expanded. Steppe and forest-steppe environments became dominant (Grichuk, 1954; Chiguryaeva and Khvalina, 1961; Moskvitin, 1962; Bolikhovskaya et al., 2017).

2.1.4.2. Correlation. The lower Khazarian is generally considered to be equal to the upper part of the Middle Pleistocene, while the upper Khazarian is correlated with the lower part of the Upper Pleistocene. According to Th-U, TL and electron spin resonance data, lower Khazarian deposits are dated at ages of > 300, > 250, 148-177, and 142-108 ka. They have a normal polarity (Brunhes) and one or two subzones of reverse polarity (Shkatova, 2010). The Khazarian comprises three transgressive stages which correspond to MIS 10, 8 and 6 (Fig. 6; Yanina, 2014).

Upper Khazarian deposits are dated by Th-U and TL methods at 127-

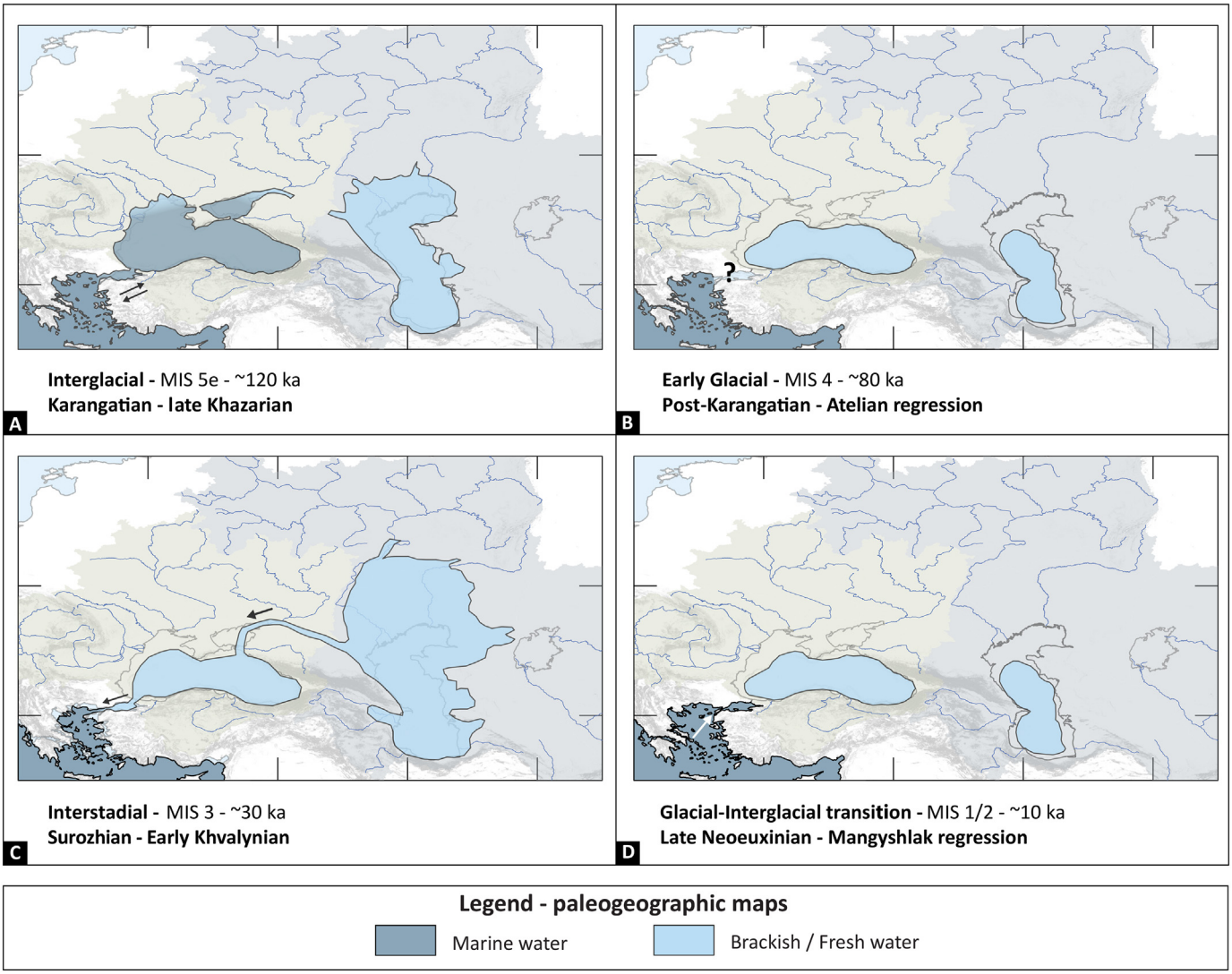

Fig. 7. Paleogeographic maps for the late Pleistocene Pontocaspian region. Arrows indicate the water flow direction in the gateway regions. All maps are based on Yanina (2014).

87 and 122-(130)-84 ka (Shkatova, 2010), 144-90 ka (*Geochronology…*, 1974), 130-91 ka (Leontiev et al., 1975), and 122-106 ka (Shakhovets, 1987; Shkatova, 2010). According to Uranium-Ionium dates, the age of the late Khazarian transgression was estimated at between 114 and 75 ka (Leontiev et al., 1975; Rychagov, 1997), and between 115 and 81 ka (Arslanov et al., 1978). Results of electronic paramagnetic (spin) resonance (ESR) revealed ages between 140 and 85 ka (Bolikhovskaya and Molodkov, 1999, 2008; Molodkov and Bolikhovskaya, 2009) and 108-85 ka (Shkatova, 2010). These results led to the conclusion that the late Khazarian transgression phase took place between 127 and 122 ka, the initial regression phase at 117–114 ka BP, and the full regression phase at 114–85 ka, coeval with MIS 5e, 5e-d, and 5d-a, respectively (Shkatova et al., 2010). Late Khazarian and Hyrcanian deposits of the Srednyaya Akhtuba section (lower Volga area) were OSL dated at 112 ± 5 ka, 102 ± 5, 87 ± 4 ka indicating they also correspond to MIS 5 (Yanina et al., 2017a, 2017b). Paleomagnetic measurements of upper Khazarian sediments confirmed the occurrence of a reversal excursion in five sections. Its age (from Th–U, TL and ESR measurements) is estimated between 117 ± 7.5 and 89 ± 11 ka (Shkatova, 2010), and thus is most likely to correspond to the reversed polarity Blake event (~120 ka). It is generally accepted that the late Khazarian transgression corresponds to the Eemian Interglacial.

Most age constraints indicate that the upper Khazarian deposits have an age between 125-85 ka, corresponding to MIS 5 (Fig. 8). The Hyrcanian transgression corresponds to the final part of the interglacial interval MIS 5a (85 ka). The Atelian regression then most likely peaked at an age of 85-75 ka. The final phase of regression is dated by OSL at 48 ± 3 (Yanina et al., 2017b) and by ^{14}C at 45–41 ka (Bezrodnykh et al., 2017) suggesting it corresponds to the first half of the interstadial warming of MIS 3.

2.1.5. Khvalynian

2.1.5.1. Description. The Khvalynian stage developed in the Late Pleistocene glacial period. Very high transgressions with an overflow event towards the Black Sea interchanged with very deep regressions. Salinities were somewhat depressed compared to today and a unique landscape (Baer knolls) developed in the North Caspian plains.

The Khvalynian Stage has been defined by Andrusov (in Pravoslavlev, 1913) and comprises the sedimentary strata that transgressively overlie the Atelian and Khazarian deposits in the North Caspian Basin and Manych Strait. The Khvalynian transgression is by far the most extensive sea-level rise in the Late Pleistocene history of the Caspian Sea (e.g., Yanina, 2014) (Fig. 7c). Khvalynian deposits are generally subdivided into lower and upper Khvalynian highstand intervals that are separated by the Enotaevka regression (Fig. 6).

The well preserved paleocoastlines of the early Khvalynian Sea

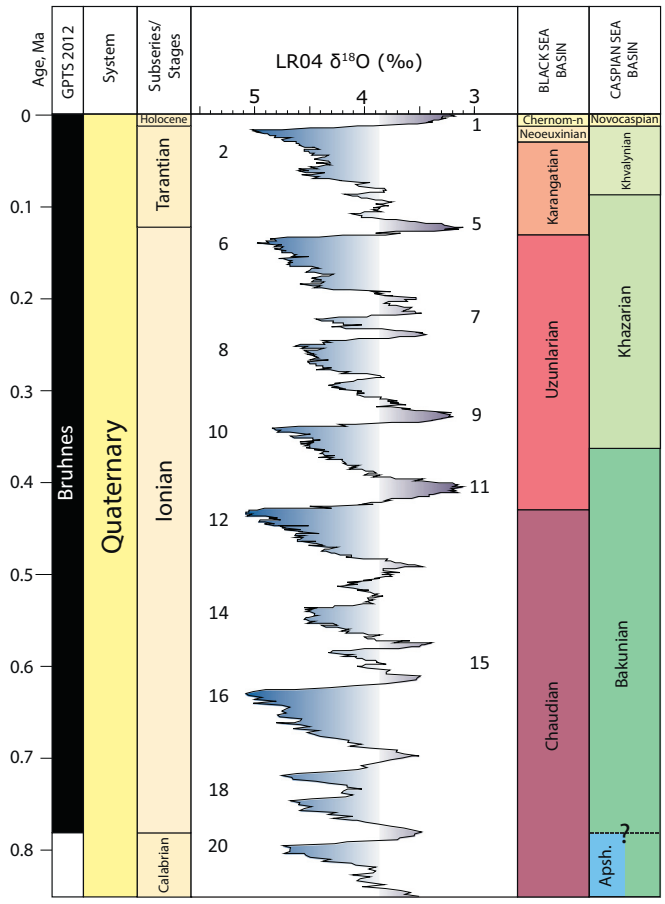

Fig. 8. Quaternary time scales for the Pontocaspian domain. GPTS with Systems and Stages is after Hilgen et al. (2012), oxygen isotope curve with numbered Marine Isotope Stages (MIS) is after Lisiecki and Raymo (2005). On the right side are the time scales for the Caspian Basin and the Black Sea Basin.

allow precise determination of facies distributions. Its sedimentary thickness is usually only a few meters, but can reach over 100 m in the Kura and Western Turkmenistan basins. The lower Khvalynian deposits comprise widely different sediments, including the characteristic liman-type "chocolate clays" of the lower Volga area that probably formed by accumulation of fine brown sediments that derived from the periglacial landscapes of the hinterland (Moskvitin, 1962; Goretskiy, 1966; Makshaev and Svitoch, 2016; Tudryn et al., 2016). Palynological data confirm a cold climate (Abramova, 1974; Yakhemovich et al., 1986). The early Khvalynian water level reached +50m, and Caspian Sea water spilled over to the Black Sea via the Manych Strait (e.g., Popov, 1983; Varuschenko et al., 1987; Yanina, 2014). During this period the North Caspian Basin was much larger than today (Fig. 7c).

The lower Khvalynian and upper Khvalynian are separated by the Enotaevka regression (Brotsky and Karandeeva, 1953), which reached a maximum lowstand at -105 m (Maev, 1994). Terrestrial Enotaevka deposits and numerous unconformities are described from the coastal regions (Leontiev and Fedorov, 1953; Vasiliev, 1961; Rychagov, 1974; Fedorov, 1978a; Svitoch and Yanina, 1997). The regressive Enotaevka unit in the North Caspian Basin consists of deltaic deposits with marked cross-bedding (Roslyakov et al., 2007). It is prominent in the deeper parts and pinches out towards the shore where it is characterized by an unconformity (e.g. Svitoch and Yanina, 1997). According to pollen data, arid cool climate conditions existed (Sorokin et al., 1983). During the late Khvalynian transgressive stage, sea level reached about 0 m (which is 27 m above todays Caspian levels: (Fedorov, 1957; Varuschenko et al., 1987; Rychagov, 1997). No overflow to the Black Sea existed. Upper Khvalynian deposits are found on all Caspian coasts in the interval of -20 to 0 m, but their thickness does not exceed a few meters.

In the lower Khvalynian *Didacna ebersini, D. parallella, D. protracta* are abundant, while *D. cristata, D. subcatillus, D. praetrigonoides, D. delenda and D. zhukovi* are rare. Characteristic species are *D. parallella* and *D. protracta*. On the eastern coast, these species are replaced by *D. cristata* and *D. zhukovi*. The thin shells of these molluscs indicate relatively low water temperatures compared to the present-day Caspian Sea. The structure of mollusc fauna is indicative of relatively low salinities, although regional variations existed. Salinities in the North Caspian Basin (3-4‰) were slightly higher than today, whereas salinities in the southern basin (~11 ‰) were slightly lower (Yanina, 2014). The upper Khvalynian deposits comprise clays, silts and sands with *Monodacna caspia*, *Dreissena polymorpha*, *D. grimmi*, and gastropods. Multiple *Didacna* species include *D. praetrigonoides*, *D. parallella*, *D. protracta*, and more rarely *D. subcatillus*. *Didacna praetrigonoides*, a rare species in the early Khvalynian, became dominant. The salinity of the main water body of the late Khvalynian basin was very similar to the early Khvalynian (Yanina, 2014). The relative abundance and thick shells of the molluscs indicate warmer conditions in late Khvalynian times then during early Khvalynian. Palynological data confirm a general warming of the Caspian region (Grichuk, 1954; Abramova, 1974; Vronskiy, 1974; Vronsky, 1976; Sorokin et al., 1983; Yakhemovich et al., 1986).

Ostracods of the Khvalynian are mainly represented by *Loxoconcha gibboides, L. endocarpa, Caspiolla gracilis, Leptocythere propinqua, L. martha, Paracyprideis enucleate, Cyprideis torosa* (Sedaikin, 1988). Foraminifera are dominantly brackish water species (Yanko, 1989).

2.1.5.2. Correlation. The ages of the Khvalynian transgressions are rather ambiguous and remain subject of discussion (Kaplin et al., 1972, 1977a, 1977b; Kvasov, 1975; Leontiev et al., 1975; Arslanov et al., 1978, 2016; Svitoch and Yanina, 1997; Rychagov, 1997; Bezrodnykh et al., 2004; Badyukova, 2007; Yanina, 2014). Based mostly on (now considered outdated) TL dates, the age of the lower Khvalynian deposits was estimated at 70-40 ka, and the age of the upper Khvalynian deposits at 20-10 ka (Leontiev et al., 1975; Rychagov, 1997). According to ^{14}C and $^{230}Th/^{234}U$ data the age of the Khvalynian is 19-8 ka (Kvasov, 1975; Svitoch and Yanina, 1997; Leonov et al., 2002; Chepalyga et al., 2008; Svitoch et al., 2008; Tudryn et al., 2013; Arslanov et al., 2016). Based on the radiocarbon dates from drilling the North Caspian Basin, lower Khvalynian ages are 30-21 ka and upper Khvalynian ages 19-12 ka (Bezrodnykh et al., 2004, 2015). The OSL dating of the chocolate clay in the lower Volga area resulted in ages of 15 ± 1 and 13 ± 0.5 ka (Yanina et al., 2017b). Paleomagnetic studies of the lower and upper Khvalynian sediments revealed the presence of two magnetic excursions, identified at several sections (Seroglazka, Lenino, and Yenotaevka), which are likely to correspond to the Laschamps (41 ka) or Mono Lake (32 ka) events (Laj and Channell, 2007). The discrepancies between these results from different methods, which may even lead to inversions in the chronology, show that the ages for these transgressions are not yet well established (Tudryn et al., 2013).

According to Sorokin et al. (2018), the lower Khvalynian transgression correlates with the global interstadial warming of the younger half of MIS 3 at an age of 30-21 ka, and is caused by an increase in the surface runoff from the catchment area. The Khvalynian sea level rise was interrupted at the LGM (time of maximum cooling and aridization, MIS 2) and resumed when the ice sheet was decaying. The warm phases of Bølling and Allerød promoted ice sheet melting along with thawing of permafrost, the latter having been widespread in the Volga drainage basin (Sorokin et al., 2018). The Khvalynian came to its end at the first sharp warming that resulted in the rise of the Caspian level and is generally taken as marking the Pleistocene/Holocene boundary.

In conclusion, it seems that most age constraints for the base of the Khvalynian converge on an age of 35 ka (young option), while the top is

estimated at ~10 ka (Fig. 8). The age of the maximum Enotaevka regression separating the lower and upper Khvalynian is around ~15 ka.

2.1.6. Novocaspian

2.1.6.1. Description. The Holocene Novocaspian stage represents the modern Caspian Sea settings and faunas.

The Novocaspian (Holocene) Stage, defined by Bogachev (1903), comprises the sedimentary strata that transgressively overlie the Khvalynian deposits in the North Caspian Basin. Novocaspian deposits are developed on all Caspian coasts below -19 (-20) m. In the North Caspian Basin, they are represented by shallow water sands with numerous fresh-and anomahaline molluscs species. The Novocaspian is separated from the upper Khvalynian by the regressive Mangyshlakian facies (defined by Zhukov, 1945), that represents the deltaic progradation of the Volga and Ural rivers during a major sea level fall up to – 80 m or even – 113 m (Figs. 6, 7d) (Varuschenko et al., 1987; Bezrodnykh et al., 2004, 2015). The Mangyshlakian deposits contain peats and sands with plant detritus and species poor assemblages of fresh water and oligohaline molluscs (*Dreissena polymorpha, Lymnaea, Unio*) but lack *Didacna* species (Sorokin, 2011).

The Novocaspian mollusc fauna is marked by various *Didacna* species of the crassoidal and trigonoidal groups: *Didacna crassa, D. baeri, D. trigonoides, D. pyramidata, D. longipes, D. barbotdemarnii*. In addition, numerous other anomalohaline molluscs are present like *Monodacna caspia, Hypanis plicatus, Adacna vitrea*, etc. Characteristic for the Novocaspian are the species *D. baeri* and *D. trigonoides,* plus the entry of *Cerastoderma glaucum* (Fedorov, 1953, 1957). The uppermost beds contain *Mytilaster minimus* and *Abra segmentum,* which were anthropogenically introduced into the Caspian from the Azov/Black Sea during the 20th century. Spores and pollen spectra contain up to 25% of tree pollen, mainly pine and birch, indicative of the relatively humid climate in the Holocene (Sorokin, 2011).

2.1.6.2. Correlation. Judging from the absolute age determinations, the Mangyshlakian regression peaked between 10 and 8 ka, and the Novocaspian deposits are all younger than 7 ka (Fig. 8). The AMS ^{14}C data from boreholes in the Volga delta suggest a lowstand around 8000 BP. Volga delta data indicate a continuously rising sea level between 5000 and 3000 BP until a highstand was reached at -25 m around 2600 BP (Overeem et al., 2003; Kroonenberg et al., 2008). During the historically well-known mediaeval Derbent regression (Rychagov, 1997) Caspian Sea levels dropped to -34 m, and possibly even -45 m as recorded in the deeper parts of the offshore Kura delta in Azerbaijan. A second highstand around 300 BP is documented in an outermost barrier in Dagestan (Kroonenberg et al., 2007). The two highstands appear to coincide with two well-known periods of increased precipitation in Eurasia, the 2600 BP event (Van Geel and Renssen, 1998) and the Little Ice Age, whereas the Derbent regression seems to be coeval with the Warm Mediaeval Period (Kroonenberg et al., 2007).

2.2. The Black Sea region

The Black Sea is today a marginal sea of the Mediterranean (Fig. 2) and has a surface area of 436,400 km^2 (excluding the Sea of Azov), a maximum depth of 2,212 m, and a volume of 547,000 km^3. Its E-W extent is about 1175 km (27°27'-41°42') and it stretches ~800 km N-S (46°33'- 40°56'). At present, it is the world's largest meromictic water body; deep waters do not mix with the upper water layers that receive oxygen from the atmosphere. As a result, over 90% of the deep Black Sea volume is anoxic. Circulation patterns are primarily controlled by basin topography and fluvial inputs, which result in a strongly stratified vertical structure. The Black Sea has a positive freshwater balance: it receives more fresh water from the rivers and rainfall than it loses from evaporation. The Black Sea consequently experiences an estuarine type of water transfer with the Mediterranean Sea via a shallow threshold (35-40 m) at the Bosphorus Strait, with bottom inflow of dense Mediterranean water below a surface outflow of fresh Black Sea water into the Marmara Sea. The salty Mediterranean inflow mixes with the basin's fresher waters, which results in an average salinity of 18 – 22‰ for the Black Sea surface waters, i.e. much lower than the Mediterranean (37 – 38‰).

During Quaternary to Recent times, periods of isolation and episodic connection with the Mediterranean Sea (through the Marmara and Aegean seas) largely controlled the paleoenvironmental conditions in the Black Sea (Zubakov, 1988; Badertscher et al., 2011; Van Baak et al., 2016b). The ancient Bosphorus Strait may have been slightly deeper than today, as the sill depth in the Paleozoic bed rock is estimated at ~85 m in the Dardanelles Strait (Algan et al., 2001). However, the Bosphorus gateway itself evolved only in the Middle Pleistocene (McHugh et al., 2008). When global/Mediterranean Sea levels were above the Bosphorus sill, marine water connections existed all the way to the Black Sea. During such connection phases the Black Sea level tracked that of global sea levels. When the Mediterranean levels were below the sill, the Black Sea turned into an isolated, saline lake basin. Major rivers like the Danube, Dniester, Dnieper, and Don (*via* the Sea of Azov), supply fresh water to the Black Sea; together they drain a large part of continental Europe (Fig. 2). During intervals with a positive water balance the Black Sea level remained at the sill height and one-directional flow towards the Mediterranean Sea occurred. In times of negative water budgets, lake levels dropped until the total inflow (precipitation and river influx) equalled the evaporation in the Black Sea Basin (e.g., de la Vara et al., 2016). Salinities during these lake phases were typically in the oligohaline-mesohaline ranges, very similar to today's Caspian Sea.

In the northeast, the Black Sea is connected to the Sea of Azov through the Kerch Strait (Fig. 2). The Sea of Azov (45°12′-47°17′N, 33°38′-39°18′ E) has a surface area of 39,100 km^2 and is 13 m deep in its central part. Two large rivers, the Don and Kuban, flow into the Sea of Azov. Annually 49.2 km^3 waters flows to the Black Sea, while 33.8 km^3 returns, resulting in an average salinity of ~11 ‰. Recently, anthropogenic reduction of river drainage strengthened the inflow of the Black Sea waters and increased the average salinity up to 13.8 ‰. In the geological history, the Sea of Azov frequently desiccated during glacio-eustatic lowstands and the Don and Kuban rivers directly drained into the Black Sea, south of the modern Kerch Strait (Yanina, 2012a).

During the Pontian (6.1-5.6 Ma), the Black Sea was connected to the Mediterranean in the south, the Dacian Basin of Romania in the west and to the Caspian Basin in the east (Popov et al., 2004, 2006; Krijgsman et al., 2010; Van Baak et al., 2015b, 2016a, 2017). During early Pliocene times (regional Kimmerian age) the Black Sea Basin became isolated and transformed into a long-lived lake (Fig. 3a). During the late Pliocene-early Quaternary (Kuyalnikian), modern Pontocaspian faunal elements appeared in the Black Sea Basin, where a succession of saline lake stages developed, increasingly punctuated (from the Middle Pleistocene onwards) by short marine connectivity phases similar as today. Here we review the stratigraphic development of the Black Sea domain during the Plio-Quaternary.

2.2.1. Kuyalnikian

2.2.1.1. Description. At the onset of the Pleistocene, the Pontocaspian domain consisted of two autonomous provinces; the mesohaline-polyhaline Akchagylian basin of the Caspian region and the oligohaline-mesohaline Kuyalnikian basin of the Black Sea region (Fig. 3b). The late Pliocene-early Pleistocene Kuyalnikian Beds (Sintsov, 1875) are composed of shallow-water sands, sandstones, and clays with marly interbeds and represent a stratigraphic analogue of the Akchagylian sediments. They represent an anomalohaline, mostly isolated long-lived lake in the Black Sea Basin.

The Kuyalnikian (or Kujalnician) is named after the "Kuyalnik Estuary" north of Odessa in the Ukraine (Fig. 2) and inherited its brackish-water regime from the preceding Pontian and Kimmerian water bodies of the Black Sea, after which the basin gradually

desalinated. Salinity in the Kuyalnikian was relatively low, based on the abundance of fresh water mollusc and ostracod species. The configuration of the Kuyalnikian basin was relatively similar to the present-day Black Sea Basin (Fig. 3b), and included the south Ukrainian shelf and the Azov-Kuban and Rioni gulfs (Popov et al., 2006). The Kuyalnikian basin had no direct connection with the open ocean, but an ephemeral connection probably existed with the Akchagylian basin of the Caspian region as Akchagylian molluscs like *Aktschagylia subcaspia* and *Cerastoderma dombra* are known from Kuyalnikian sequences (Taman beds) in the Azov area (Nevesskaya et al., 1986). However, there is no other evidence that Akchagylian faunas penetrated the Black Sea Basin.

Palynological data from Western Georgia (Shatilova et al., 1998) and Eastern Ukraine (Sirenko, 2000) show frequent changes of vegetation, temperatures and humidity, which reflect the generally increasing climatic cyclicity of the Kuyalnikian compared to Pliocene. In Western Georgia, warm and humid phases were characterized by mixed conifer-broadleaved forests, whereas cool and less humid phases saw extensive expansions of conifer forests (Shatilova et al., 1998, 2011). North of the Black Sea, the cooler and drier phases supported increasingly expanding steppe environments. Correspondingly, broadleaved forests became more restricted to warm and humid phases, less diverse and limited to riparian habitats (Sireko, 2000).

In the Azov-Kuban region, the Kuyalnikian generally comprises a ~12-16 m thick unit of bluish-grey clays and sands; its most substantial thickness of ~80 m is reported from drillings in the northern Azov region. The Kuyalnikian stage is commonly represented by a two-fold subdivision with a lower Kuyalnikian containing bivalves like *Dreissena, Pseudocatillus, Prosodacna* and *Lymnocardium* and an upper Kuyalnikian ("Taman Beds" of Eberzin, 1935) that comprises Akchagylian-type molluscs like *Avimactra subcaspia* and *Cerastoderma dombra* (Nevesskaya et al., 1986; Popov et al., 2006). Kuyalnikian deposits are widely distributed in the south-western part of Georgia (Taktakishvili, 1984; Molostovsky, 1997; Kirscher et al., 2017). Characteristic mollusc fossils there include *Pontalmyra medeae, Pseudocatillus pleonexia* and *Pseudocatillus postdonacoides* (Nevesskaya et al., 2003).

Stratigraphically and/or paleoenvironmentally significant ostracods include brackish water species like *Bakunella dorsoarcuata*, *Caspiolla acronasula*, *Loxoconcha eichwaldi* and *L. petosa* and fresh water species like *Caspiocypris labiate* and *Cypria arma*. Small-sized euryhaline benthic foraminifera like *Ammonia beccari* and *Elphidium incertum* have only been reported from the northern Black Sea region (Karmishina, 1975). The lowermost Kuyalnikian and the Taman Beds contain rare intercalations of nannofossils with zonal species such as *Discoaster brouweri* and *D. pentaradiatus* correlative to zones NN17-18 (Semenenko and Lulieva, 1978; Semenenko and Pevzner, 1979).

In the central Black Sea cores of DSDP Leg 42B, the Kuyalnikian arguably corresponds to the Units IVa and III (Van Baak et al., 2016b). The lower Kuyalnikian may correlate to subunit IVa (644.6–718 mbsf) which is composed of sideritic and diatomaceous sediments. The upper Kuyalnikian may correspond to the lower part of Unit III (446.5–644.5 mbsf) which is of a freshwater nature, characterized by lacustrine chalk (Seekreide) and mud (Ross and Neprochnov, 1978).

2.2.1.2. Correlation. Paleomagnetic data from the Azov-Kuban region indicate that the upper Kimmerian and lower Kuyalnikian deposits have normal polarity whereas the "Taman Beds" and overlying deposits show reversed polarity. This pattern has been revealed in numerous sections and cores; Chegerchi (Semenenko and Pevzner, 1979), Tobechik and Arshintsevo (Molostovsky, 1997), Krasny Kut, Babach-Tarama, Kryjanovka, and Liventsovka (Tretyak and Volok, 1976), Taman (Zubakov and Kochegura, 1976; Zubakov, 1990). The normal to reversed polarity transition in the mid-Kuyalnikian is generally correlated with the Gauss/Matuyama boundary at 2.58 Ma (Zubakov, 1990).

Paleomagnetic results from western Georgia also indicate a lower Kuyalnikian of normal polarity and an upper Kuyalnikian of reversed polarity; this is expressed in outcrops along the Skurdubi and Tsina-Gele rivers, in the Kvarbeti II section and in some scattered sites (Molostovsky, 1997). The normal polarity zone of the lower Kuyalnikian here is also correlated with a part of the Gauss chron (Zubakov, 1990), while the upper Kuyalnikian is correlated to the lowermost Matuyama chron (Molostovsky, 1997).

In conclusion, most paleomagnetic data indicate that the lower boundary of the Kuyalnikian is located in the upper part of the Gauss chron. Consequently, we place the boundary here at a tentative age of 2.7 Ma, which makes it similar to the base of the "short" Akchagylian in the Caspian region (Fig. 5).

2.2.2. Gurian

2.2.2.1. Description. During the Early Pleistocene, the Black Sea Basin was occupied by the anomalohaline Guria Lake. In this lake an endemic Pontocaspian fauna evolved. No connections existed with the marine realm. Possibly only ephemeral connections existed with the Caspian Basin through the Manych Strait north of the Caucasus.

The Gurian is named after Guria, western Georgia (Fig. 2). According to most literature, the Gurian beds in the Black Sea Basin are mostly the equivalents of the Apsheronian of the Caspian region. The Gurian basin was significantly smaller than the modern Black Sea Basin (Fig. 3c). Consequently, Gurian exposures on land are relatively scarce and occur mainly in western Georgia and in the northwestern part of the Kerch Peninsula.

The Gurian strata of Georgia were first described by Ilyjn (1929, 1930) and are subdivided into two lithostratigraphic units: the Khvarbetian and Naderbazetian. The two units have been divided on the basis of their cardiid assemblages (Kvaliashvili, 1976). The Khvarbetian is only found in southern Georgia and comprises shallow "brackish" lacustrine greyish-bluish marls with dreissenids in its lower part and small gastropods including *Laevicaspia* and *Turricaspia* in the top. The Naderbazetian is more widespread and dominated by bioclastic calcarenites (Ilyjn, 1930) that contain shells of *Digressodacna* (Taktakishvili, 1984).

The Gurian climate in Western Georgia was quite similar to the Kuyalikian one, with climatic changes mainly affecting temperatures, while humidity remained high and relatively stable. The taxonomic composition of the vegetation also remained similar and continued the trend of stepwise reduction of thermophilous elements; especially at the end of the Gurian a large number of Pliocene relicts became locally extinct in Western Georgia. In parallel, the structure of vegetation changed into a more pronounced altitudinal zonation of forests. The lowlands and higher mountain zones remained covered by poly-dominant broadleaved and conifer forests, respectively. But for the first time in this region, beech forests formed a separate community in the middle altitudes of the increasingly uplifting Caucasus mountain ranges (Kogoshvili, 1977; Milanovsky, 1977; Tsagareli, 1980; Shatilova et al., 2011, 2014)

2.2.2.2. Correlation. The age and stratigraphic position of the Kuyalnikian/Gurian boundary is not straightforward and has been correlated differently by various authors; with the reversed interval above the Olduvai (~ 1.7 Ma; Pevzner (1986), with the upper part of the Reunion subchron (~2.1 Ma; Molostovsky, 1997), with the middle part of the Olduvai (~1.9 Ma; Alexeeva et al., 1981) and with the top of the Olduvai (1.8 Ma; Popov et al., 2006). Zubakov and Borzenkova (1990) correlate the main part of the Gurian with the reversed Matuyama chron. Recent magnetostratigraphic investigations of the Rioni Basin of Georgia also revealed a long reversed interval in the Gurian Khvarbeti section that is correlated to C1r.2r (Kirscher et al., 2017). Based on the magnetostratigraphic pattern of the Khvarbeti section, the base of the Gurian was correlated with the top of the Olduvai at an age of 1.8 Ma, although it must be noted here that the Kuyalnikian-Gurian boundary is not present in the section itself and

relied on correlation with nearby sections (Kirscher et al., 2017).

In summary, although there are many uncertainties in the published magnetostratigraphic results, we tentatively place the base of the Gurian at the top of the Olduvai subchron at an age of 1.8 Ma (Fig. 5). This correlation makes the Gurian roughly coeval with the Apsheronian of the Caspian region and synchronous with the Calabrian stage of the Mediterranean and standard GTS (Kirscher et al., 2017).

2.2.3. Chaudian

2.2.3.1. Description. During the Middle Pleistocene Chaudian, the Black Sea Basin was dominated by anomalohaline lake conditions (Fig. 3d). Several episodes of connection with the Caspian Basin occurred. Also several overflow events into the Marmara Basin occurred where Chaudian faunas were established temporarily. At the same time the Black Sea first experienced marine incursions from the Mediterranean Sea during interglacial highstands.

The Chaudian takes its name from the promontory Cape Chauda on the Kerch Peninsula of Crimea (Fig. 2), where also the stratotype section is located (Andrusov, 1889; Fedorov, 1963; Nevesskaya, 1963). The Chaudian basin was characterized by multiple changes in water level ranging from –140 to –30m (Sorokin and Babak, 2011). Differently oriented tectonic movements favored different preservation of Chaudian sediments. In the eastern part of the Black Sea, only coastal transgressive facies are preserved on uplifted terraces (Sorokin and Babak, 2011). In the deeper Black Sea Basin, the Chaudian deposits are represented by uniform clays and silts with sapropelitic interlayers in the upper part. These sediments contain assemblages of fresh-water and brackish-water cryophilic diatoms (Ross, 1974; Svitoch, 2008).

The Chaudian has been subdivided into a lower Chaudian and an upper Chaudian, separated by a break in sedimentation (Fedorov, 1963). The lower Chaudian is represented by sediments and fossils indicating a freshened basin with an average salinity of 8-10 ‰ (Yanina, 2015). Lower Chaudian deposits are characterized by the molluscs *Tschaudia tschaudae*, *T. guriana*, *Didacna guriensis*, *D. supsae*, *D. pseudocrassa* and *D. baericrassa* (Yanina, 2014). They also contain brackish water foraminifera (Chepalyga et al., 1989). According to diatoms, spores and pollen, cold climatic conditions with periglacial vegetation prevailed (Chernyshova, 1980; Zhuze et al., 1980) and most elements of thermophilic flora had disappeared (Shatilova, 1974). In Georgia, the Chaudian climate was temperate in the upper and middle forest belts, and warm-temperate in the lower ones (Shatilova et al., 2011). Humidity was generally high. The early Chaudian climate was warmer and altitudinal zonation more distinct than in the late Chaudian, which appears to have been temperate at most elevations (Shatilova et al., 2011). Major glaciers developed in the Caucasus, while cold pluvial conditions marked the lower areas (Dumitrashko et al., 1977).

The upper Chaudian strata are represented by shallow-water carbonates with abundant fauna consisting of *Didacna pseudocrassa, D. baericrassa, D. olla, D. tamanica*, *Monodacna subcolorata, M. cazecae, Hypanis plicatus, Dreissena chaudae*, *D. pontocaspia* (Fedorov, 1963; Nevesskaya, 1963; Svitoch et al., 1998; Yanina, 2012b). The salinity of the basin was on average 12–13 ‰ (Yanina, 2005). In addition, upper Chaudian strata also contain *Didacna* species characteristic of the upper Bakunian in the Caspian Sea (*Didacna rudis, D. carditoides, D. catillus, D. eulachia*), leading to the term Chauda-Bakunian (Fedorov, 1963; Popov, 1983; Yanina, 2005). This suggests one or several connections through the Manych Strait existed with drainage of Caspian waters into the Azov region (Fedorov, 1963; Popov, 1983; Yanina, 2005). On the periphery of the late Chaudian basin and near the mouths of large rivers, coastal lagoons with anomalohaline water fauna formed. Forest-steppe communities with prevalence of broad-leaved forest vegetation dominated the surrounding areas (Arap et al., 1990).

The upper part of the Chaudian deposits contains shells of Mediterranean euryhaline molluscs such as *Cerastoderma glaucum*, *Mytilus galloprovincialis* and *Scrobicularia plana* (Andrusov, 1896; Keraudren, 1970; Fedorov, 1978a) and planktonic foraminifera such as *Globigerina bulloides, Globorotalia hirsuta* and *G. crassaformis* (Yanko, 1989), suggesting episodic marine incursions. On the Caucasian coast, a terrace with a species poor Mediterranean mollusc fauna has been determined as "pre-Uzunlarian" (Solov'ev, 1977), expressing the first Pleistocene influx of Mediterranean waters into the Black Sea Basin. This semi-marine basin is also referred to as Epichaudian (Fedorov, 1978a), Patrayian (Zubakov, 1988) or Karadenizian (Chepalyga, 1997).

The uppermost part of the Chaudian is marked by a 40-60 m regression (Fig. 6), evidenced by deep erosion of river valleys and estuaries that transitioned into shallow lakes (Mikhaylesku, 1990). On the northern coasts of the Black Sea accumulation of thick loess successions took place.

2.2.3.2. Correlation. In several geological time scales of the Black Sea region (e.g., Shatilova et al., 2009), the lower boundary of the Chaudian is indicated as being coeval with the Brunhes-Matuyama boundary, which has an age of 780 ka (Hilgen et al., 2012). In other time scales, however, the Brunhes-Matuyama boundary is placed between lower and upper Chaudian (Zubakov, 1988; Nevesskaya et al., 2005). Recently, the Gurian-Chaudian transition was magnetostratigraphically determined in the Khvarbeti section (at level 157m) of Georgia, where it occurs in a reversed interval interpreted to correspond to chron C1r.1r and an age of ~0.85 Ma (Kirscher et al., 2017) (Fig. 5). The marine transgression that caused the first penetration of Mediterranean molluscs into the late Chaudian is correlated with MIS 15 or 13 (Yanina, 2012a).

2.2.4. Uzunlarian

2.2.4.1. Description. The Middle Pleistocene Uzunlarian is a continuation of the conditions of the Chaudian in the Black Sea Basin. Pontocaspian faunas were widespread but less diverse and several marine connections occurred during interglacial highstands.

The Uzunlarian (Arkhangelsky and Strakhov, 1938) is a transgressive stage, that follows the regression at the end of the Chaudian. The stratotype section of the Uzunlarian is the Uzunlar section on the Kerch Peninsula (Fig. 2: Arkhangelsky and Strakhov, 1938; Fedorov, 1963; Chepalyga et al., 1989). In the lower, sandy part of the section molluscs adapted to unusual salinities occur such as *Didacna pontocaspia, D. pallasi, D. subpyramidata, D. baericrassa, Monodacna, Hypanis* and *Dreissena* spp. In the upper part, Pontocaspian species were replaced by euryhaline Mediterranean molluscs such as *Cerastoderma glaucum*, *Abra segmentum* and *Chamelea gallina*. The Uzunlarian represents an interval of repeated marine conditions in the Black Sea Basin (Fedorov, 1963, 1978a; Zubakov, 1988). The connection with the Mediterranean Sea was re-established via the Bosphorus Strait, salinity increased from 10–13‰ to 15–17‰, and euryhaline species resistant to salinity variations (e.g. *Cerastoderma glaucum*) penetrated the Black Sea (Fedorov, 1978a). Two different levels with lower Khazarian molluscs in the sedimentary successions of the Manych Depression and the Azov region prove that Caspian waters drained at least twice through the Manych Strait into the Black Sea Basin (Fedorov, 1978a; Popov, 1983; Yanina, 2012a).

The Uzunlarian deposits comprise evidence of both warm and cold periods. Marine transgressions of Mediterranean waters took place in interglacial periods, while ingressions of cold and brackish Caspian waters occurred during glacials. During interglacial intervals the Black Sea Basin became populated by euryhaline Mediterranean molluscs like *Paphia*, *Scrobicularia*, *Cerastoderma*, *Abra, Chione, Mytilus* and *Ecrobia* spp. (Svitoch et al., 1998). Most of them co-occured with *Didacna* species. Stenohaline ostracods and planktonic foraminifera (*Neogloboquadrina pachyderma, Globigerina bulloides* and *Globigerinoides ruber*) are also documented. Sea level was close to zero and salinity reached 15-17‰. During glacial episodes the Black Sea Basin became isolated from the Mediterranean and became populated with *Didacna baericrassa, D. pontocaspia, D. borisphenica, D. raricostata, D. akschaena*, and the Caspian (early Khazarian) species *Didacna pallasi, D. subpyramidata* and *D.*

pallasi. The water level was well below today's and salinity decreased to ~7 ‰ near the river mouths and 11–12 ‰ in the open basin (Nevesskaya, 1963, 2007; Svitoch et al., 1998; Yanina, 2005, 2012a).

The exact number of marine and brackish water intervals in the Uzunlarian is contentious: one alternation of Caspian and marine type deposits (Old Euxinian-Uzunlarian) has been recognized by Arkhangelsky and Strakhov (1938) and Nevesskaya (2007); two alternations were identified by Fedorov (1978) and Chepalyga (1997); and three (named Paleouzunlarian, Uzunlarian and Asheylian) were proposed by others (Fig. 6; Ostrovsky et al., 1977; Zubakov, 1988; Nesmeyanov and Izmailov, 1995; Yanina, 2012a). A major regression phase separates the Uzunlarian from the overlying Karangatian (Fedorov, 1978a; Popov, 1983). This regression resulted in deep erosion of paleo-valleys; e.g., the Bosphorus incised to -100 m (Fedorov, 1978a). On the western (Bulgarian) margin of the Black Sea, however, no sedimentation break has been observed between the Uzunlarian and Karangatian (Hrischev and Shopov, 1979; Grigoriev et al., 1985; Svitoch et al., 1998). In the lakes of the Danube Delta cold-resistant fresh water stagnophile mollusc species of *Planorbis, Sphaerium, Pisidium*, etc. settled, while pine, birch, alder, and fir-tree dominated the vegetation (Mikhaylesku, 1990).

2.2.4.2. Correlation. Accurate absolute age determinations of Uzunlarian deposits are very rare. According to Zubakov (1988), the Uzunlarian is dated between 580 and 300 ka and corresponds to MIS 15. Later, it was established that the Uzunlarian transgression took place in the second half of the Middle Pleistocene at an age of ~250 ka (e.g., Svitoch, 2003). Based on OSL dating, the upper part of the Uzunlarian was dated at ~227,5 ka (Sychev, 2017). According to paleogeographical reconstructions based on *Didacna* species, Yanina (2012b) correlated the three marine transgressive phases (Fig. 6) with MIS 11 (Paleouzunlarian), MIS 9 (Uzunlarian) and MIS 7 (Asheylian). In this scenario, the age of the lower Uzunlarian boundary is located below MIS 11 at around ~400 ka (Figs. 5 and 8).

2.2.5. Karangatian

2.2.5.1. Description. The Karangatian represents the most widespread marine conditions in the Black Sea Basin established during the penultimate Eemian (MIS 5e) interglacial and its transition into the last glacial (Fig. 7a). During MIS 5, basin settings were very similar as today, but sea level and salinity were higher. Towards MIS4 and MIS 3 the Black Sea Basin became isolated.

The deposits overlying the Uzunlarian were initially attributed to the Mediterranean Tyrrhenian stage near Sudak village in Crimea (Andrusov, 1905), but later renamed as Karangatian after the sedimentary successions on Cape Karangat (Arkhangelsky and Strakhov, 1938). The stratotype of the Karangatian is the Eltigenian terrace, located on the Kerch peninsula (Andrusov, 1905; Fedorov, 1963; Nevesskaya, 1965; Yanko et al., 1990; Dodonov et al., 2006). There, the Karangatian deposits are represented by shell beds and shelly sandstones with numerous *Ostrea edulis, Flexopecten glabra, Mimachlamys varius, Mytilus galloprovincialis, Chamelea gallina, Acanthocardia tuberculatum, Cerastoderma glaucum, Aporrhais pespelecani, Cerithium vulgatum,* etc. They overlie a 60m thick clay sequence of Uzunlarian age with freshwater molluscs (*Lymnea* and *Planorbis*) and rare *Didacna* (Zubakov, 1988). Deep water sediments of the Karangatian in the Black Sea Basin are similar in structure and composition to the Holocene sapropel and clay-coccolith oozes.

Based on mollusc fauna, the Karangatian was subdivided in three intervals that show first a gradual increase of salinity and then a decrease (Nevesskaya, 1965). The Karangatian stage includes four to five minor transgressive intervals (Fig. 6). The first highstand (Tobechikian) consists of two subcycles which are characterized by an impoverished Mediterranean fauna with *Cerastoderma glaucum*, *Abra segmentum, Mytilaster lineatus, Paphia senescens, Chamelea gallina,* and *Ecrobia* cf. *maritima* (Nevesskaya, 1965). In between and on top of these transgressive strata, hydromorphic soils are developed indicative of exposure during lowstands. In the Tobechikian the foraminifera *Elphidium caspicum azovicum* and *Ammonia parasovica* were found (Yanko et al., 1990). The second (Zavetnian) highstand is also represented by two subcycles with *Ostrea edulis, Donax trunculus, Cerithium vulgatum, Solen vagina* and *Pholas dactylus* (Zubakov, 1988). In between and on top are dune sand intervals. A third highstand (Eltigenian), overlying the previous units with an erosional unconformity, is composed of sands with stenohaline faunal elements including *Acanthocardia tuberculata, Dosinia lupinus*, *D. exoleta*, *Ensis ensis*, *Barbatia barbata*, *Chama gryphoides, Mimachlamys varia*, *Scrobicularia plana*, *Jagonia reticulata*, *Donax variegatus, Hiatella arctica*, *Anomia ephippium* and *Gari depressa* (Nevesskaya, 1965; Zubakov, 1988). The Eltigenian includes numerous foraminifera like *Massilina inaequalis, Quinqueloculina laevigata, Q. bicornis,* and *Porosononion martcobi ponticus* (Yanko et al., 1990). The fourth (Tarkhankutian) stage is represented by a species-poor Mediterranean mollusc assemblage lacking halophilic elements. The dominant species are *Cerastoderma glaucum* and *Abra segmentum,* which indicate that salinity decreased during the latest Karangatian sea level fall (Nevesskaya, 1965; Kuprin and Sorokin, 1982; Chepalyga et al., 1989). The Tarkhankutian contains the foraminifera species *Ammonia compacta, A. parasovica, Elphidium ponticum, Porosononion subgranosus mediterranicus,* and *Haynesina anglica.* In the Kerch-Taman area, the Caspian (Hyrcanian) species *Didacna subprotracta*, *D. subcatillus, D. cristata* are found together with euryhaline marine species. They got across the Manych in times of Karangatian sea regression (Popov, 1983; Yanina, 2014; Yanina et al., 2017a; Kurbanov et al., 2018). Some authors consider a fifth, post-Karangatian (Surozhian) highstand, which contains the Khvalynian index species *Didacna ebersini* along with marine species (Popov, 1955, 1983). Others, however, consider the marine fauna as redeposited (e.g., Fedorov, 1978a; Fedorov, 1978b; Nevesskaya, 1965; Yanina, 2012b; Yanina, 2006).

The Karangatian transgression was the largest of the entire Quaternary history of the Black Sea, with a maximum water level approximately 6-8 m higher than today. Warm saline Mediterranean waters entered the Black Sea Basin through the Bosphorus Strait and a one-way migration of euryhaline and stenohaline Mediterranean fauna took place. At the maximum of the transgression, the salinity of the Karangatian sea was about 30‰ (Nevesskaya, 1965). High salinities were also characteristic for the southern part of the Sea of Azov, the Manych and the Don estuary. The Karangatian transgression extended all the way into the Manych Depression but marine mollusc species never entered the Caspian Basin. The Karangatian sea was further marked by relatively warm waters, as shown by subtropical diatom species (Zhuze et al., 1980). Palynological data also indicate warm and moderately dry conditions (Artyushenko et al., 1972, 1973; Vronsky, 1976; Koreneva, 1980). Coastal lowlands were occupied by steppe and forest-steppe landscapes (Sorokin, 2011). During the second half of the Karangatian, the climate cooled and coastal lagoons transformed into lakes inhabited by fresh-water molluscs (*Planorbis, Lymnea*, and others). In Western Georgia, during the warm early Karangatian lowlands were covered by swamp forests and mixed broadleaved communities; middle mountain belts carried beech forests, and conifer forests were restricted to the upper mountain belt. With the progressive cooling during the late Karangatian, conifer forests began to migrate to lower elevations (Shatilova et al., 2011).

2.2.5.2. Correlation. A variety of absolute age estimates exist for the Karangatian/Surozhian, based on radiocarbon, uranium-ionium, and thermoluminescence methods. According to isotopic and geochronological data, the Karangatian transgression took place between 129 and 70–80 ka ago (Arslanov et al., 1993). Uranium dates for the Eltigenian transgression range from 129-88 ka (Arslanov et al., 1983) or 127-120 ka (Zubakov, 1988). ESR age estimates are constrained between 127–121 ka (Velichko, 2002). The Surozh highstand is dated at 62-47 ka (Arslanov et al., 1983) or 40-25.

Paleomagnetic inclination data of the Karangatian Tuzla section on the Black Sea coast of the Taman Peninsula reveal a reversal excursion dated at ~110 ka that coincides with a similar anomalous direction in the Eltigen section on the Kerch peninsula (Dodonov et al., 2006). These reversal excursions probably correlate with the Blake reversed event that has an age of 100 ka (Singer et al., 2014). At Cape Tuzla, the Surozhian beds are also found on top of a reversal excursion that was assumed to correspond to the Laschamp reversed event at 43ka (Zubakov, 1988).

On the basis of these dates, we estimate that the Karangatian represents an interval of approximately 50–60 thousand years with its base at the beginning of MIS 5e at an age of ~130 ka and the maximum of the Karangatian transgression roughly corresponding to the global sea level highstand during the Eemian during MIS 5e (Fig. 8) (Chepalyga, 1997; Velichko, 2002; Yanina, 2014). The presence of Hyrcanian (Caspian) mollusc species in the Tarkhankutian beds (Popov, 1983; Yanina, 2014) suggests that this stage corresponds to the terminal transition from the Eemian interglacial (MIS5a) to the MIS4 glacial epoch. The Surozhian highstand (up to -25 m) took place during MIS 3 (Yanina, 2014).

2.2.6. Neoeuxinian

2.2.6.1. Description. The Late Pleistocene Neoeuxinian forms the final Pontocaspian lake phase in the Black Sea Basin. The Neoeuxinian lake had low salinities and lake levels were predominantly very low. However, at least one overflow event from the Caspian occurred and at least one relative highstand resulted in a Black Sea-Marmara Sea connection judging from Neoeuxinian communities there (Fig. 7c). The Neoeuxinian phase was terminated with early Holocene marine ingresssions which restricted Pontocaspian biota to the modern coastal habitats along the northwestern Black Sea coast and the marginal areas of the Azov Sea.

Neoeuxinian sediments were first recognized by Andrusov (1918), while Arkhangelsky and Strakhov (1938) introduced the term Neoeuxinian. These sediments developed on the Black Sea floor at depths below 20 m and are represented by diverse sandy-clayey deposits with slightly anomalohaline and freshwater fauna (Yanchilina et al., 2017). The Neoeuxinian includes two intervals. The lower Neoeuxinian corresponds to a deep regression with mostly fresh water molluscs like *Dreissena, Viviparus, Valvata* and *Unio*. The connection with the Sea of Marmara was disrupted during this regressive phase and the Black Sea transformed into a freshwater to oligohaline basin with salinities not exceeding 1.5–3.0‰. At this time, the Black Sea reached a minimum level of -110 m (Ostrovsky et al., 1977; Balabanov and Izmaylov, 1989) to -150 m (Ryan et al., 1997). The Black Sea shelf and the Kerch Strait were exposed, and the Sea of Azov was a flat lowland crossed by the channel of the Don river. The deltas of the Danube and Dnieper rivers were located 200 km offshore from their present-day position. Diatom flora provides evidence for cold environments (Zabelina and Scherbakov, 1975) and a cold and arid climate is also reconstructed from spores and pollen data (Artyushenko et al., 1972). Most of the Pontocaspian region was covered by cold steppe vegetation, but temperatures and humidity remained higher in Georgia. There, the colder climate is expressed in the expansion of conifer forests, while broadleaved plants retreated and survived in the refuge of warm gorges (Shatilova et al., 2011).

The upper interval is dominated by Pontocaspian mollusc species such as *Dreissena* cf. *bugensis, Monodacna caspia, M. colorata, Adacna vitrea* and *Hypanis plicatus*. Caspian (Khvalynian) shells of *Didacna ebersini* (Andrusov, 1918; Fedorov, 1963; Semenenko and Sidenko, 1979) and Khvalynian ostracod species (Popov and Suprunova, 1977) and rare foraminifera (Yanko, 1989) occurred as well. This interval corresponds to a rise in water level to -20 m, and to Caspian waters flowing through the Manych Strait (Svitoch and Yanina, 2007). At the same time, the presence of Neoeuxinian faunas in the Marmara Basin and northern Aegean shows Black Sea overflow (Islamoğlu and Chepalyga, 1998; Taviani et al., 2014; Büyükmeriç, 2016).

2.2.6.2. Correlation. The age of the Neoeuxinian regression is estimated at 22–16 ka (Scherbakov et al., 1977; Balabanov and Izmaylov, 1989) or at 25–22 ka (Degens and Ross, 1972). From 16 to 12.5 ka, the water level is thought to have increased because of the influx of Caspian (Khvalynian) waters overspilling into the Black Sea as a result of intensive thawing after the last glacial maximum (LGM) (Balabanov and Izmaylov, 1989; Balabanov, 2006; Murdmaa et al., 2006; Svitoch and Yanina, 2007; Yanchilina et al., 2017). The age of the Khvalynian molluscs from the Manych Strait is 14.7-12.6 ka (Svitoch et al., 2010; Yanina, 2014). Hence, we conclude that the Neoeuxinian basin developed during MIS 2 with the deep regression coinciding with the LGM and the late Neoeuxinian transgression with the deglaciation phase (Fig. 8).

2.2.7. Chernomorian

2.2.7.1. Description. The Chernomorian represents the modern Holocene conditions in the Black Sea.

The Neoeuxinian/Chernomorian boundary is based on the first emergence of the Mediterranean molluscs *Cerastoderma glaucum* and *Abra segmentum.* The Chernomorian sediments are deposited all over the Black Sea, with clays rich in hydrogen sulfide in deep settings and lagoonal silts and shelly limestone terraces at the margins. The Chernomorian is subdivided in two subunits, the Old Chernomorian and Neochernomorian, based on different mollusc assemblages (Fedorov, 1978a). Old Chernomorian sediments are dominated by brackish-water species and the emergence of euryhaline Mediterranean elements. The Neochernomorian contains a more diverse assemblage of Mediterranean molluscs. Nevesskaya (1963, 1965) described four different stages in Chernomorian mollusc assemblages (Bugazian, Vityazevian, Kalamitian, and Dzhemetian) and concluded that both sea level and salinity gradually increased until the present. According to Fedorov (1978a, 1982), the maximum Neochernomorian transgression reached 2–2.5 m above modern sea level and corresponds to the highest salinities of the Holocene Black Sea. During the Chernomorian, Pontocaspian habitats became suppressed by the Black Sea ingression into deltaic and coastal lakes and lagoons, as shown by Vespremeanu-Stroe et al. (2017) for the Danube delta area.

Numerous publications have discussed the problem of reconstructing the exact mode of sea level rise in the Holocene Black Sea Basin. According to Ryan et al. (1997), Mediterranean waters filled the Black Sea Basin with an enormous speed, flooding the entire region within a couple of years. These authors furthermore link this catastrophic deluge to the Biblical Flood of Noah (Ryan and Pitman, 1999). In contrast, many others showed that sea level increased gradually in a reciprocating mode, with various speed and depending on different mechanisms (Balabanov and Izmailov, 1988; Aksu et al., 2002; Balabanov, 2006; Taviani et al., 2014; Yanchilina et al., 2017).

2.2.7.2. Correlation. The Holocene transgression of the Black Sea, caused by Mediterranean inflow, is dated at 11 ka (Balabanov, 2006), at 10 ka (Ostrovsky et al., 1977), 9 ka (Degens and Ross, 1972), 8-9 ka (Vinogradov et al., 1969; Kuprin and Sorokin, 1982) and 7-8 ka (Kvasov, 1975; Fedorov, 1978a). According to Ryan et al. (1997), Noah's flood happened ~7.2 kyr ago. The age of the Neochernomorian phase is determined by radio-carbon dating at 5-3 ka (Svitoch et al., 1994, 1995, 1998); at 5.5–5 ka and 4.5-4 ka (Fedorov, 1982; Selivanov, 1996), and at 4.2-3.8 ka (Arslanov et al., 1988).

2.3. The Marmara Sea region: gateway between Mediterranean and Pontocaspian

The Marmara Basin (Fig. 9) is a complex, deep basin located between the Black Sea in the north and the Aegean/Mediterranean Sea to the south. Together with the adjacent Iznik Lake Basin it contains a

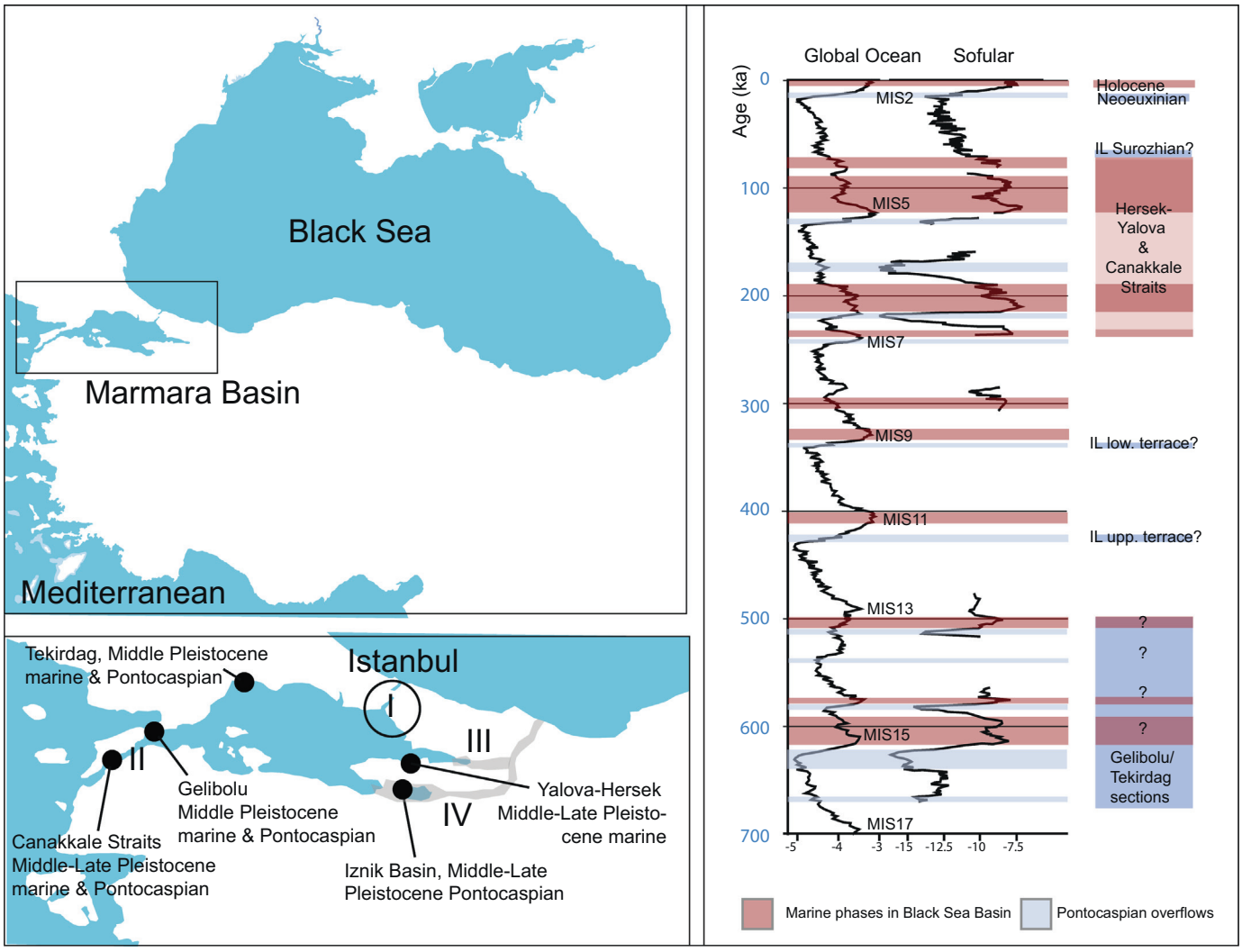

Fig. 9. Stratigraphic framework for middle-late Quaternary marine and Pontocaspian intervals in the Marmara Basin and surroundings. The Sofular isotope record is from Badertscher et al. (2011). Blue: Pontocaspian intervals, Red: marine intervals. IL = Iznik Lake. Gateways: (I) Bosphorus, (II) Dardanelles, (III) Sapanca corridor and (IV) Iznik corridor.

Middle-Late Pleistocene succession of Pontocaspian and marine Mediterranean units that testifies of a dynamic history (Ergun and Ozel, 1995; Parke et al., 1999; Okay et al., 2000; Imren et al., 2001; Le Pichon et al., 2001). Currently it is connected through two gateways, the Bosphorus in the north with a sill depth of -35 m (McHugh et al., 2008) and the Dardanelles to the south with a sill depth of - 70 m b.s. (Aksu et al., 1999; Hiscott et al., 2002). The Marmara Basin and its gateways have an active tectonic history including regional uplift of marginal areas, strong subsidence within the basin and locally, very strong uplift along the branches of the active North Anatolian Fault System (NAFS: Aksu et al., 2000; Le Pichon et al., 2016; Tari and Tüysüz, 2016). The current Bosphorus probably developed only in the Middle Pleistocene and Middle Pleistocene connections between the Marmara Sea and Black Sea Basin may have run through an Iznik or a Sapanca corridor (Le Pichon et al., 2016; Fig. 9).

The oldest Pontocaspian deposits in the Marmara Basin comprise a series of mostly gravelly deposits along the NW shores of the Marmara Basin that have been assigned a Chaudian age based on correlation with Chaudian *Didacna* species known from the Black Sea Basin (Andrussov, 1900; Taner, 1983). Outcrops are known from the surroundings of Tekirdag and especially from Gelibolu and Çanakkale. The Gelibolu outcrops contain a 25 m thick succession that may contain several overflow phases and the presence of *Lentidium mediterraneum* (reported as *Corbula mediterraneum* in Taner, 1983) implies possible intermittent connection with marine habitats as well.

Late Middle Pleistocene marine deposits and faunas are known from marine terraces in the Yalova area (Büyükmeriç et al., 2016) and the shores of the Dardanelles (Erol, 1965, 1985, 1992; Erol and Nuttal, 1972; Erol and İnal, 1980). Late Middle Pleistocene Pontocaspian deposits with *Didacna* are known from a series of terraces on the northern shores of Lake Iznik (Islamoğlu, 2009; Büyükmeriç, 2016), while *Monodacna* faunas were found in boreholes in Lake Iznik (Meriç et al., 2017). Using tectonic uplift calculations provided by Yaltırak et al. (2002), the ages for the terrace levels are between 322 and 575 ka, corresponding to Chaudian/Old Euxinian intervals of the Black Sea Basin. These authors, as well as Meriç et al. (1995), argued for a paleoconnection with the Marmara Basin, because strong uplift has been detected along the southern branch of the NAFZ.

Early Late Pleistocene (MIS 5) marine deposits and faunas are found in various areas in and around the Marmara Basin (Meriç et al., 1995, 1999; Büyükmeriç, 2016; Büyükmeriç et al., 2016). These deposits are rich in mollusc faunas with a composition almost identical to the Uzunlarian and Karangatian faunas of the Black Sea Basin. They also lack several of the stenohaline species characteristic of coeval Mediterranean faunas (Büyükmeriç et al., 2016). This assemblage suggests slightly reduced salinities (mesohaline-polyhaline) compared with the

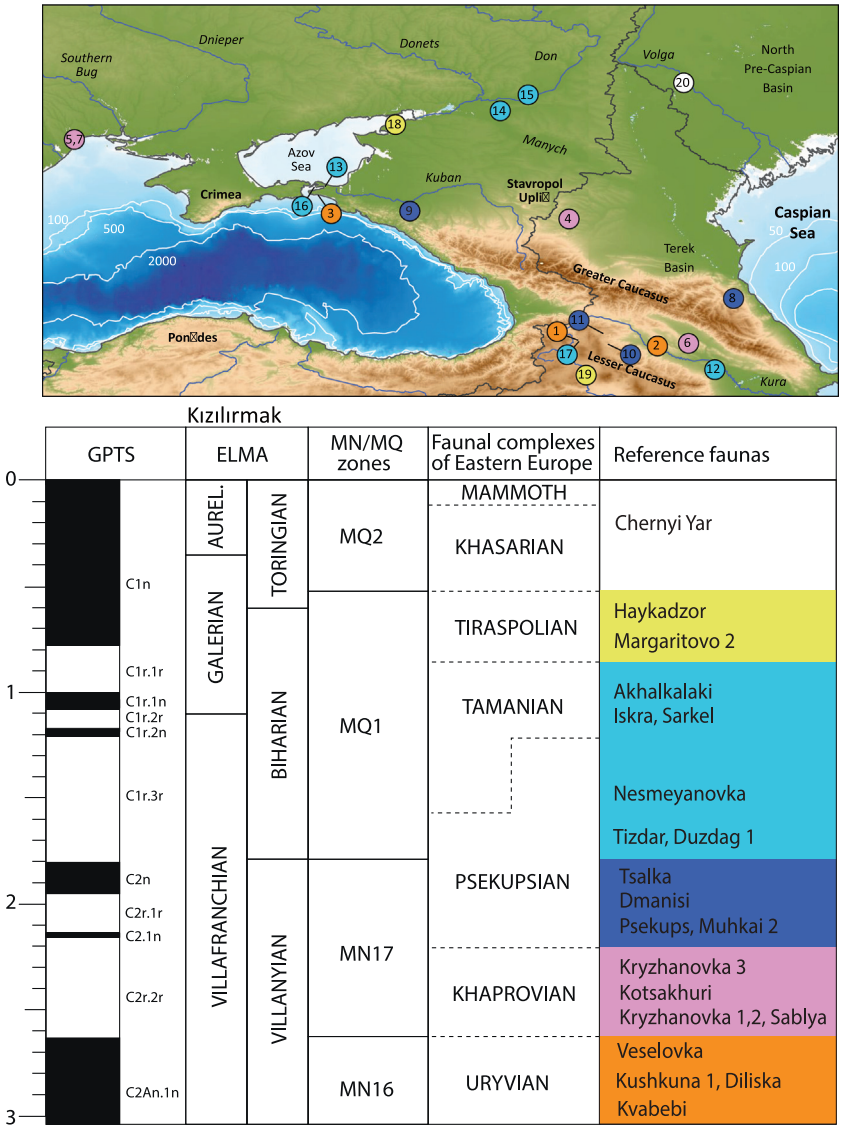

Fig. 10. Left: Correlation of late Pliocene - Pleistocene continental stages and reference faunas of the Pontocaspian domain with standard chronostratigraphy and mammal-based biochronological units. Right: Map of mammal reference localities. MN 16: Early Villafranchian/Early Villanyian 1) Diliska (N41.43, E43.46), 2) Kushkuna 1 (N41.28, E45.45), 3) Veselovka (N45.13, E36.92). MN 17: Middle Villafranchian/Late Villanyian 4) Sablya (N44.53, E43.23¬), 5) Kryzhanovka 1,2 (N46.56, E30.8), 6) Kotsakhuri (N41.35, E46.23), 7) Kryzhanovka 3 (N46.56, E30.8). MN 17-MQ1: Late Villafranchian/Late Villanyian-Early Biharian 8) Muhkai 2 (N42.25, E47.36), 9) Psekups (N44.73, E39.23), 10) Dmanisi (N41.33, E44.34), 11) Tsalka (N41.6, E44.09). Late Villafranchian – Early Galerian / Early Biharian 12) Duzdag 1 (N40.68, E46.89), 13) Tizdar (N45.36, E37.1), 14) Nesmeyanovka (N47.29, E41.5), 15) Sarkel (N47.7, E42.2), 16) Iskra (N45.35, E36.83), 17) Akhalkalaki (N41.4, E43.48). Galerian / Late Biharian 18) Margaritovo 2 (N46.94, E38.87), 19) Haykadzor (N40.54, E43.65). Aurelian / Toringian 20) Chernyi Yar (N48.01, E46.11).

Mediterranean conditions at the time.

Meriç et al. (2017) reported Pontocaspian faunas of Late Pleistocene age from Lake Iznik boreholes and made a tentative correlation with Surozhian intervals of the Black Sea Basin. The faunas imply another more recent connection between Lake Iznik and the Pontocaspian basins, although a paleogeographic explanation for such young connections is still uncertain. The final Pontocaspian phase in the Marmara Basin is marked by submerged Neoeuxinian deposits formed during MIS2. Around 15 ka ago, the presence of fully developed Neoeuxinian faunas found in cores and box cores in a number of sea floor localities in the southern half of the Marmara Sea overflows indicates that overflows from the Black Sea Basin must have existed (Islamoğlu and Chepalyga, 1998; Taviani et al., 2014; Büyükmeriç, 2016). The early Holocene flooding of the Dardanelles threshold at ca 9.3 ka brought the current marine conditions back to the Marmara Basin (Büyükmeriç et al., 2016).

3. Constraining stratigraphic ages of terrestrial units alongside the Caspian and Black Sea Basins

Present-day Pontocaspian species live well into the lower courses of adjacent rivers. Understanding the age of these extended Pontocaspian habitats helps our understanding of the evolution in the lakes themselves. Furthermore, the terrestrial biotas can provide additional time constraints on the lake successions as well as insights into major regional climate and environmental disruption episodes, important to understand evolution within the lake basins. Finally, episodes of lake basin isolation should correspond to terrestrial migration corridors whereas periods of lake interconnection should correspond to terrestrial habitat fragmentation. Characterising faunal distributions and improving age constraints is therefore a major contribution in understanding the evolution of the Pontocaspian region for both lacustrine and terrestrial biotas. Here we focus on the terrestrial sequences of the key regions between the Caspian and Black Sea basins in the Azov-Crimea region and their importance to understand Pontocaspian lake basin evolution.

3.1. The fresh water mollusc record

The fresh water mollusc record of lowland units that are coeval with the Pontocaspian lake phases can be subdivided in a late Pliocene-early Pleistocene phase (the so-called Levantine phase) and a Pleistocene phase. Before the Levantine phase mollusc faunas were characterized by the occurrence of long-lasting subtropical lineages with few boreal species (Stratigraphy…, 1982). In the Dacian Basin of Romania, the local term “Romanian” is commonly used for the late Pliocene-early

Pleistocene fresh water environments (see Papaianopol et al., 2003; Van Baak et al., 2015a and references therein).

3.1.1. Late Pliocene-Early Pleistocene

The increased climatic fluctuations at the Pliocene-Pleistocene transition correspond with the appearance of boreal taxa that formed the Levantine fauna association. The Levantine fauna is subdivided into three groups: early (corresponding to late Kimmerian - early Akchagylian), middle (middle - late Akchagylian) and late Levantine (Apsheronian) (Stratigraphy…, 1982).

The early Levantine is characterized by the ascent of peculiar sculpted subtropical types of Unionidae and Viviparidae. Pearly fresh water mussel groups such as the (sub-) genera *Rugunio, Cuneopsidea, Ritia* and *Cyclopotomida* emerged while the genus *Plicatibaphia* went extinct during this stage (Stratigraphy…, 1982).

During the middle Levantine, the pearly freshwater mussel genera *Cuneopsidea, Rugunio, Ritia, Pristinunio* became extinct, while the genus *Bogatschevia* appeared. Boreal species appeared for the first time and a transition from thermophilous to more cold-resitant communities occurred. The middle Levantine is well known from the Veselovka and Tizdar mammal sites (Fig. 10: Taman Peninsula, Russia). The Veselovka fauna is dominated by *Dreissena theodori kubanica* and sculpted *Unionidae* species, such as *Rugunio lenticularis* in the lower part, and *Avimactra* species in the upper part. The highly ornamented unionid groups probably represent stable warm temperate to subtropical conditions, comparable to modern Mediterranean climates, but with plenty of precipitation in the Black Sea Basin (Chepalyga, 1985).

The lower part of the Veselovka section, with both freshwater and brackish-water molluscs, is correlated to the Kuyalnikian. The upper part with *Avimactra* shells of the Akchagylian on Taman (Eberzin, 1940). The presence of *Dreissena theodori kubanica*, which went extinct at the end of the Kuyalnikian (Nevesskaya et al, 1997), determines the fauna as pre-Gurian. According to Kochegura and Zubakov (1978), the Gauss/Matuyama boundary (2.6 Ma) is present in Veselovka. The site has a small mammal fauna corresponding to zone MN16(b) (Tesakov, 2004).

The late Levantine faunas are characterized by the gradual disappearance of subtropical elements and increasing prominence of boreal species. The unionoid *Pseudosturia* appeared at this time. *Bogatschevia* species, representing subtropical conditions, declined. Key late Levantine sites are the Iskra and Maly Kut localities (Fig. 10) on the Taman Peninsula that yield fresh water faunas with a variety of pearly fresh water mussel species as well as some Pontocaspian species (*Dreissena polymorpha, D.* cf. *eichwaldi, Cardiidae* indet., *Clessiniola* sp.) indicating the presence of lowland Pontocaspian-fluvial habitats. The occurence of the bivalve *Pseudosturia* indicates colder climatic conditions than during the middle Levantine, although the climate was still moderately warm. In addition to the freshwater fauna, the Iskra locality produced shells of the Apsheronian *Dreissena* cf. *eichwaldi*. Younger late Levantine faunas are known from the Margaritovo 1, 2, Semibalki 1 and Port-Katon 3 locations in the Northeast Azov Sea region. These faunas are characterized by a large number of fresh water species which are close to modern relatives. Thermophile taxa such as *Borysthenia naticina, Parafossarulus* sp., *Fagotia* spp., *Lithoglyphus* pp. and *Pisidium sulcatum* indicate warm climatic conditions. Such conditions are also suggested by thick-walled shells of Unionidae and Viviparidae in the Semibalki 1 site. The presence of cardiids in the Semibalki fauna suggests proximity of anomalohaline Pontocaspian lake habitats.

The late Levantine localities yield small mammal associations attributed to regional MQR7 and MQR8 zones (Tesakov et al., 2007a). Late Levantine deposits are reversely magnetized corresponding to the reverse polarity interval between the Jaramillo subchron and the Brunhes-Matuyama reversal. The Nesmeyanovka section in the Rostov region contains the pearly fresh water mussel species *Bogatschevia scutum, Pseudosturia caudata, P. brusinaiformis, Potamoscapha tanaica* (Stratigraphy…, 1982) and a small mammal fauna corresponding to the upper part of the MQR9 regional zone (Fig. 5).

Well-dated late Pliocene-early Pleistocene fresh water faunas are also known from the Dacian Basin (Papaianopol et al., 2003). They developed from 4.15 Ma onward, when a paleoenvironmental change from anomalohaline to fluvio-lacustrine environments took place related to progressive basin infill (Van Baak et al., 2015a; Jorissen et al., 2018).

3.1.2. Pleistocene phase

The composition of the fresh water mollusc fauna acquired a near-modern appearance in the early Middle Pleistocene. Various localities in the Azov region such as Platovo, Zeleny, Semibalki 2, Port-Katon 4 yield rich early Middle Pleistocene molluscan faunas. The mollusc association is represented by 46 taxa, almost all extant. The predominance of *Lithoglyphus* represents river habitats. In addition, limnophiles and stagnophiles (e.g. *Valvata, Planorbis, Gyraulus* spp.) are very common. The few cardiid occurrences point to the proximity of Pontocaspian estuarine/saline habitats. Most of the early Middle Pleistocene species are temperate species, but there also some warm temperate species such as *Lithoglyphus* spp., *Fagotia esperi*, and *Borysthenia naticina* (Glöer, 2002) indicating relatively warm (interglacial) conditions. The mollusc fauna from the Port-Katon 4 locality is younger and contains a suite of modern boreal freshwater species and a single extinct species (*Parafossarulus crassitesta*). Thermophilic elements are lacking in Port-Katon 4, which indicates a colder climate in the younger part of early Middle Pleistocene.

During the late Middle Pleistocene to Late Pleistocene, the onset of extensive continental glaciations on the Russian Plain increasingly affected the Pontocaspian lowlands and had a strong influence on the development and distribution of freshwater molluscs (Gozhik and Dacenko, 2007). During interglacials, warm-water migrants from the south appeared in the molluscan assemblages. Only boreal taxa remained during glacial intervals. The mollusc assemblages of this period are characterized by modern species. During the warm interglacials MIS 9-11 warm-temperate assemblages with common thermophilous *Corbicula* were present. Faunas from cold (glacial) intervals are impoverished and dominated by limnophilous and stagnophilous molluscs (Stratigraphy 1982). The beginning of the late Middle Pleistocene is characterised by the fauna from the Mikhailovskoe location with *Viviparus* sp., *Lithoglyphus pyramidatus, Valvata* spp., *Anisus spirorbis*, and the bivalve *Corbicula fluminalis*. This association includes three thermophilic species: *Lithoglyphus pyramidatus, Borysthenia naticina, Corbicula fluminalis*. Today, *Corbicula* lives in regions with mean annual temperatures (MAT) between 13 and 17 °C (compared to a MAT of c 8 °C for the Rostov region today (Volkova, 1966)). Late Middle Pleistocene faunas from the Beglitsa section are younger and comprise an admixture of fresh water (e.g. *Valvata pulchella, Bithynia leachi, Anisus strauchianus, Vallonia pulchella, V. costata,* Unionidae indet., *Dreissena polymorpha)* and Pontocaspian species *(Clessiniola* sp., *Didacna* cf. *pallasi, D.* cf. *borisphenica* and *Monodacna caspia*). The association represents both lowland rivers and their transition towards Pontocaspian lake environments.

The early Middle Pleistocene faunas correspond to the so-called Cromerian faunas of western Europe (Preece, 2001). The interval contains the last occurrence of *Borysthenia goldfussiana*, which went extinct at the end of the early Middle Pleistocene (Gittenberger and Janssen, 1998; Frolov and Kurshakov, 2015). The small mammal associations are typical for the Tiraspol faunal complex and belong to the regional zone MQR 5 with an estimated age range of ~550-700 ka (Vangengeim and Tesakov, 2008). Only normal polarities were found (in e.g. Platovo, Port-Katon, Kagalnik) that were attributed to the Brunhes chron (Agadjanian, 1976; Tesakov et al., 2007a). The small mammal fauna in the Port-Katon 4 locality belongs to the later part of the Tiraspol faunal complex and correlates with the regional zone MQR4 (Tesakov et al., 2007a). *Didacna pallasi* penetrated across the Manych Strait into the Black Sea in the early Middle Pleistocene and survived almost to the

end of the Middle Pleistocene (Yanina, 2012a). The youngest Pleistocene fauna is correlated with the post Eemian part of the Late Pleistocene.

3.2. The terrestrial mammal record

Mammals provide a reliable means of stratigraphic control in terrestrial domains due to their fast morphological evolution and high origination/extinction rates, combined with regional and continent-wide dispersals and their sensitivity for climatic and environmental change. A European biochronological zonation was elaborated by Mein (1999, 1990, 1975), who included characteristic taxa, typical associations, first and last appearance datums of the genus rank taxa, and reference localities/fauna for his MN zonation scheme (Fig. 10). The MN scheme contains 17 units covering the timespan from Aquitanian to Gelasian (Mein, 1976; Steininger, 1999; Hilgen et al., 2012). Several extensions were proposed to cover the late Pliocene and Pleistocene, e.g. “MNQ zones” (Guérin, 1982; Agustí, 1986; Agustí et al., 1987; Nomade et al., 2014) that are relevant for the Pontocaspian region. Here, the system of Fejfar and Heinrich (Fejfar and Heinrich, 1981, 1990; Fejfar et al., 1998) is favoured, since the other schemes rely mainly on western European data (Spain, Italy and France).

A parallel and integrated system of European Land Mammal “ages”/”stages” is being developed and widely used (Mein, 1990; de Bruijn et al., 1992; Hilgen et al., 2012). The time span of Pleistocene regional stages of Eastern Paratethys roughly corresponds to three “ages”/”stages” including four MN zones of the European biochronological system: Villanyian (= early and middle Villafranchian; MN16, MN17, 3.5-1.8 Ma), Biharian (= late Villafranchian to middle Galerian; MQ1, 1.8 -0.6 Ma), and Toringian (late Galerian to Aurelian; MQ2, 0.6-Recent) (see Fig. 10) (Fejfar and Heinrich, 1981; Fejfar et al., 1998). Furthermore, several regional biochronological zonations exist. The classical zonation scheme is a system of Eastern European “faunal complexes” by Gromov (1948) based on large mammals, and essentially based on the evolutionary stages of mammuthoid elephants. The Pleistocene in the Gromov scheme is divided into six units, Khaprovian, Psekupsian, Tamanian, Tiraspolian, Khazarian, and Late Paleolithic or Mammoth complexes (Fig. 10). The parallel MQR/MNR regional biochronology of eastern Europe is based on concurrent ranges of small mammals, mostly on rapidly evolving arvicoline rodents with 14 units covering the Pleistocene (Vangengeim et al., 2001; Tesakov, 2004; Tesakov et al., 2007b).

In general, the succession of Pleistocene mammalian faunas of the Pontocaspian region is correlating well to coeval faunas elsewhere in Europe and western Asia. Certain regional differences (especially in small mammals) do not hamper direct correlations using mammalian species with vast geographic ranges. The fossil record of mammals in the Pontocaspian area is patchy and combines some better studied regions and areas with insufficient data. In the Pontocaspian region, a number of fossil mammal localities occur within anomalohaline lake and shallow marine deposits and contribute to stratigraphic bracketing of regional lacustrine zonations.

3.2.1. MN 16: Early Villafranchian/Early Villanyian

Small mammal criteria: Large European *Mimomys* with external cement (*M. hajnackensis/hassiacus* evolving into *M. polonicus*). Two subzones are defined: MN16a (*Mimomys hajnackensis*), MN16b (*M. polonicus*)(Fejfar et al., 1998).

Additional criteria: Smaller *Mimomys* of the *stehlini-hintoni* grade, *Kislangia cappettai*, *K. ischus*, *Borsodia novoasovica*.

Large mammal criteria: In addition to typical Pliocene forms like *Anancus arvernensis, Hipparion rocinantis/crusafonti*, the end of this period documents the appearance of basal *Archidiskodon meridionalis rumanus*/cf. *gromovi* (MN16b), the relatively large deer *Arvernoceros ardei* and first *Equus, Gazellospira* and *Protoryx*. These faunas were also characterised by the presence of *Dinofelis, Vulpes, Homotherium* and *Lynx issiodorensis*. In the Caucasus new members of the Bovinae subfamily appeared, which are not recorded in the European early Villafranchian (*Protobison kushkunensis* and *Ioribos aceros*) (Vekua, 1972; Lebedeva, 1978; Sotnikova, 1989; Vislobokova et al., 2001).

Ages of the boundaries: Lower boundary is close to the Gilbert/Gauss boundary (Opdyke et al., 1997; Vangengeim et al., 2005) or located in the middle Gauss chron (Agustí et al., 2001). Upper boundary is close to 2.6 Ma.

3.2.1.1. Eastern European Faunal complex: Uryvian/Skortselian. Characteristic faunas in the region: Volna, Veselovka on Taman Peninsula, Russia; Kushkuna 1 in Azerbaijan; Simbugino, south Urals, Russia (Tesakov, 2004); Kvabebi and Diliska (MN 16b) in Georgia (Agustí et al., 2009).

Correlation with lacustrine stages: Correlates with lower Akchagylian and lower Kuyalnikian. Fauna of Volna is found on top of upper Kimmerian deposits; faunas of Kvabebi, Kushkuna 1 and Simbugino (MN16b) are found in shallow marine deposits of the lower Akchagylian. The Veselovka fauna is confined to the Kuyalnikian corresponding to the very end of the Gauss chron (Tesakov, 2004).

3.2.2. MN 17: Middle Villafranchian/Late Villanyian

Small mammal criteria: Large European vole *Mimomys pliocaenicus* (including its more primitive stage *M. praepliocaenicus*).

Additional criteria: Smaller *Mimomys* gr. *hintoni livenzovicus – reidi*; *Pitymimomys* gr. *stenokorys-pitymyoides;* rhizodont lagurines *Borsodia praehungarica.*

Large mammal criteria: Common occurrence of archaic southern elephant *Archidiskodon meridionalis gromovi*, horse *Equus* (*Allohippus*), comb-antlered deer *Eucladoceros,* smaller mastodont *Anancus arvernensis alexeevae* and camel *Paracamelus alutensis.* The characteristic assemblage of this time includes *Pliocrocuta perrieri, Elasmotherium, Cervalces* (*Libralces*), *Leptobos,* large *Canis,* badgers *Meles* and large felid *Homotherium crenatidens.* The regional record includes *Adjiderebos cantabilis* (Alexeeva, 1977; Dubrovo and Burchak-Abramovich, 1986; Titov, 2008; Sotnikova and Titov, 2009; Kahlke et al., 2011).

Ages of the boundaries: Lower boundary near 2.6 Ma, upper boundary estimated from ca. 2.0 to 1.8 Ma.

3.2.2.1. Eastern European Faunal complex: Khaprovian. Characteristic faunas in the region: Sablya, Kryzhanovka 1, 2, 3, Kushkuna (2, 3 and 4 bone bearing horizons), Kotsakhuri.

Correlation with lake stages: Correlates with middle-late Akchagylian and upper Kuyalnikian. Sablya is confined to upper Akchagylian deposits; Kryzhanovka 1, 2: upper Kuyalnikian deposits, reversely magnetized (lower Matuyama Chron), Kryzhanovka 3: fresh water deposits of upper Kuyalnikian, lower Matuyama Chron.

3.2.3. MN 17-MQ1: Late Villafranchian/Late Villanyian-Early Biharian

Small mammal criteria: *M. pliocaenicus – M. ostramosensis.*

Additional criteria: *Mimomys* gr. *reidi-pusillus*, *Kislangia rex, Borsodia newtoni-arankoides,* first *Allophaiomys deucalion*.

Large mammal criteria: *Archidiskodon meridionalis meridionalis, Stephanorhinus etruscus, Eucladoceros orientalis orientalis,* peculiar *Megaloceros stavropolensis, Pachycrocuta brevirostris, Panthera gombaszoegensis.* Last occurrences of *Nyctereutes megamastoides, Gazellospira torticornis. Palaeotragus, Gallogoral, Vulpes alopecoides, Lynx issiodorensis.* First occurrence of *Bison, Capra, Praeovibos, Soergelia*, *Pontoceros* (Lordkipanidze et al., 2007; Kahlke et al., 2011; Tesakov et al., 2013).

Ages of the boundaries: Lower boundary estimated from ca. 2.0 to 1.8 Ma, upper boundary near 1.6 Ma.

3.2.3.1. Eastern European Faunal complex: Psekupsian. Characteristic faunas in the region: Psekups, Georgievsk, Muhkai (Northern Caucasus, Russia), Palan-Tukan (Azerbaijan), Dmanisi (Figs. 10, 11), Tsalka (Georgia).

Fig. 11. Dmanisi fauna, selected specimens. Top row from left to right: *Homotherium crenatidens*, skull, Dm.5/154.A1/87; *Megantereon whitei*, skull, D1341; *Panthera onca georgica*, mandible dex., D2027, holotype; *Lynx issiodorensis*, skull, D3497; *Pachycrocuta brevirostris*, juvenile mandible dex., Dm.65/63.B1x.153; *Canis etruscus*, skull with mandible, D3420. Second row from left to right: *Mammuthus meridionalis taribanensis*, articulated hind leg fragment, Dm.64/68.B1x.338; *Stephanorhinus etruscus*, skull, D3270; *Equus stenonis*, skull, D353. Third row from left to right: *Palaeotragus priasovicus*, metatarsus fragment dex., D1757; *Cervalces* cf. *gallicus*, antler fragment dex., D1949; *Arvernoceros insolitus*, antler dex., D2747, holotype; *Praemegaceros obscurus*, antler fragment dex., D430; *Pseudodama nestii*, antler sin., D1495. Fourth row from left to right: *Bison (Eobison) georgicus*, skull fragment, D354, holotype; *Gallogoral meneghini sickenbergii*, skull, Dm.4/154.B1.79; *Capra dalii*, horn core dex., D75, holotype; *Pontoceros surprine*, skull fragment D5552, holotype.

Correlation with lake stages: Correlates with upper Akchagylian and upper Kuyalnikian. Georgievsk, fluviatile deposits are located directly above the upper Akchagylian.

3.2.4. Late Villafranchian – Early Galerian / Early Biharian

Small mammal criteria: The range of *Mimomys intermedius* (= *M. savini*).

Additional criteria: Appearance and diversifications of rootless voles of the primitive *Allophaiomys* grade (up to 1 Ma), early *Microtus* (ca. 1 – 0.6 Ma). Decline of diverse mimomyine voles. Appearance and diversification of rootless lagurine voles, genera *Prolagurus, Lagurodon, Kalymnomys, Eolagurus*.

Large mammal criteria: Characteristic late southern elephants *Archidiskodon meridionalis meridionalis* grading into *A. m. tamanensis*, which coexisted with *Equus* (*Allohippus*) cf. *major*, *Elasmotherium caucasicum*, *Eucladoceros orientalis pliotarandoides, Pontoceros ambiguus, Bison (Eobison) tamanensis*, *Homotherium latidens, Pachycrocuta brevirostris*, *Panthera gombaszoegensis* and *Canis tamanensis* (Vereshchagin, 1957; Forstén, 1999; Sotnikova and Titov, 2009; Kahlke et al., 2011; Bukhsianidze et al., 2014).

Ages of the boundaries: Lower boundary estimated near 1.6 Ma. Upper boundary, ca. 0.8 Ma.

3.2.4.1. Eastern European Faunal complex: Tamanian. Characteristic faunas in the region: Tizdar, Iskra (Taman Peninsula, Russia), Margaritovo 1, Semibalki 1, Port-Katon (North-Eastern Sea of Azov Region), Duzdag 1 (Azerbaijan), Akhalkalaki (Georgia).

Correlation with lake stages: Correlates with Apsheronian and Gurian. Uppermost Kujalnikian deposits of Tizdar yielded one of the oldest small mammal faunas with basal *Allophaiomys pliocaenicus*. The fauna of Duzdag 1 with *Archidiskodon meridionalis* is in lower Apsheronian deposits. The fauna of Iskra with *Allophaiomys pliocaneicus* co-occurs with Apsheronian type molluscan fauna.

3.2.5. Galerian / Late Biharian

Small mammal criteria: Appearance of modern species/genera of rootless arvicolines of the *Microtus* group, in co-occurence with late rhizodont water voles *Mimomys intermedius* (=*Mimomys savini*).

Large mammal criteria: The abundance of large herbivorous *Mammuthus trogontherii*, *Bison schoetensacki, Palaeoloxodon antiquus, Stephanorhinus kirchbergensis, Cervalces latifrons, Equus süssenbornensis.* Among carnivores, *Panthera spelaea fossilis* and *Ursus deningeri* are present (Gromov, 1948; Nikiforova et al., 1971; Forstén, 1999).

Ages of the boundaries: Lower boundary ca. 1.2-1.0 Ma, upper boundary 05-0.6 Ma.

3.2.5.1. Eastern European Faunal complex: Tiraspolian. Characteristic faunas in the region: Haykadzor (Armenia), Tiraspol (Transdniestria), Margaritovo 2, Semibalki 2, Kagalnik, Port-Katon 4 (North-Eastern Sea of Azov Region)

Correlation with lake stages: Correlates with Bakunian and Chaudian.

3.2.6. Aurelian/Toringian

Small mammal criteria: The range of water vole *Arvicola*.

Additional criteria: Essentially modern mammalian fauna, with a number of sequentially disappearing taxa (e.g. megamammals of the "Mammoth fauna").

Large mammal criteria: *Mammuthus trogontherii chosaricus, Mammuthus intermedius, Palaeoloxodon antiquus, Elasmotherium sibiricum, Stephanorhinus kirchbergensis, Equus hydruntinus, E. latipes, Camelus knoblochi, Cervus* cf. *elaphus, Megaloceros giganteus, Bison priscus, Saiga tatarica*

Ages of the boundaries: Lower boundary is 0.36 Ma. Upper boundary – 70 ka.

Characteristic faunas in the region: Raigorod, Chernyi Yar, Nikolskoe (lower Volga), Veshenskaya, Beglitsa (lower Don area).

3.2.6.1. Eastern European Faunal complex: Khazarian. Correlation with lake stages: Khazarian (Caspian Basin), early Euxinian to Chernomorian (Black Sea Basin) regional stages. Lower Volga localities Raigorod, Chernyi Yar, Nikolskoe, etc. are situated directly in the upper Khazarian fluviatile deposits.

3.2.7. Late Aurelian/Late Toringian

Small mammal criteria: *Microtus oeconomus, Arvicola terrestris*, *Eolagurus luteus, Lagurus lagurus, Cricetus cricetus.*

Large mammal criteria: Typical representatives of mammoth fauna, most part of which are common for recent North Eurasian associations: *Mammuthus primigenius, Cervus elaphus, Alces alces, Bison priscus, Saiga tatarica, Canis lupus*, *Alopex lagopus, Vulpes vulpes*.

Ages of the boundaries: Lower boundary is ca. 70 ka. Upper boundary – 11 ka.

3.2.7.1. Eastern European Faunal complex: Mammoth. Characteristic faunas in the region: Siniy Yar, Biryuchya Balka, Muralovka, Kamennaya Balka (Lower Don Basin), Nechereziy (Kuban river Basin, North Caucasus)

Correlation with lake stages: Khvalynian

3.3. The hominin record

In terms of hominin evolution, the Pontocaspian domain plays an important, yet not completely understood, role. The region is located on cross-roads between Africa, Asia and Europe and provided migration pathways for early hominins. The Pontocaspian lakes as well as some of the high mountain ranges, semi-desert and dense forest belts may have provided obstacles to migration, yet the presence of a variety of landscapes may have facilitated dispersal as well as settlement. A complete review, including a relationship between the extent of the Pontocaspian climatic and lake evolution on one side, and the hominin evolution itself on the other side, is beyond the scope of this paper. There is still a huge debate on subjects such as migration path, explanation for gaps in the hominin record, and driving factors. Because the Pontocaspian region is regarded as a highly important region for hominin migration, including the major sites of Dmanisi and Ubeidiya, together with important late Pleistocene hominin sites, we will give a short summary of the most important hominin sites and their implications in and around this region (Fig. 12, Table 1). This might serve as a basis for future studies on the relationship of the Pontocaspian climate, the related extent of the lakes, and hominin migration. In order to understand the potential role of the Pontocaspian region for hominin evolution we extend our review to the wider region.

The oldest calibrated evidence for the presence of early hominins is recorded in the Southern Caucasus at the site of Dmanisi, in Georgia, where Oldowan lithic tools have been described from bed A, associated with mammalian remains (Ferring et al., 2011). This level is placed in a normal geomagnetic interval, identified as Olduvai, following magnetobiostratigraphic and radiometric criteria (Fig. 12). Above bed A, bed B of the Dmanisi section shows a reversed polarity and was Ar^{40}/Ar^{39} dated to ca. 1.77 Ma (Ferring et al., 2011). It is this bed B that contains the oldest hominin remains in Eurasia, represented by 5 skulls and associated 4 mandibles, as well as post-cranial skeletal elements belonging to at least 3 individuals (Vekua et al., 2002; Lordkipanidze et al., 2005, 2007). The hominins from Dmanisi are referred to an early population of *Homo erectus*, bearing many archaic traits (Lordkipanidze et al., 2013). A rich vertebrate assemblage is associated with the hominin remains, including *Mammuthus meridionalis, Equus stenonis, Stephanorhinus etruscus, Eucladoceros* aff. *tegulensis, Bison* (*Eobison*) *georgicus, Pontoceros* sp., *Canis etruscus, Pliocrocuta perrieri, Megantereon megantereon* and *Homotherium crenatidens* (Fig. 11). The small mammal association is dominated by gerbils (*Parameriones* aff. *obeidiensis*) and hamsters (*Cricetulus* n. sp.).

One of the few additional pre 1 Ma actual hominin remains was described from Ubeidiya, Jordan Valley, Israel, which has been biochronologically dated to ~1.5 Ma (Belmaker et al., 2002). This site also represents the first appearance of early Acheulian technology in the Levant corridor (Doronichev and Golovanova, 2010; Bar-Yosef and Belmaker, 2011). A somewhat younger record (1.17-1.24 Ma) of hominin presence is recorded in Anatolia at Gediz river, in the Izmir province, where a quartzitic artefact (possible a hard hammer product flake) was reported by Maddy et al. (2015). Also in Anatolia, a more abundant assemblage comes from the site of Dursunlu (Konya province, Central Anatolia), which includes lithic artifacts of Oldowan mode, mainly quartz flakes and flake fragments, associated with a mammal assemblage dominated by *Equus altidens* and *E. mosbachensis* (Güleç et al., 2009). Paleomagnetic analysis of the section reveals a dominant reverse interval punctuated by three normal episodes, which suggests a late Early Pleistocene age, covering Jaramillo and the Matuyama-Brunhes boundary (MBB) (between 0.99 and 0.78 Ma). The site of Kalepe Deresi 3, in Central Anatolia, dated to a maximum age of 1 Ma, yielded around 4000 Acheulian tools including obsidian bifaces, rhyolite pebble tools, flakes and cleavers (Slimak et al., 2008). Also in Anatolia a partial skull assigned to *Homo erectus* was reported from the site of Kocabaş (Denizli Basin), dated between 1.2 and 1.6 Ma (Lebatard et al., 2014). It includes a large fragment of right parietal, a fragment of right frontal preserving part of the supraorbital torus and a fronto-parietal fragment (Kappelman et al., 2008; Vialet et al., 2012, 2014; Aytek and Harvati, 2016). The morphometric study of this specimen suggests a link with the African fossils, *Homo ergaster* and early *Homo erectus*, and distinguishes it from specimens of Dmanisi and Asian *Homo erectus* (Vialet et al., 2018). The Kocabaş finding is associated with large mammals (*Archidiskodon meridionalis*, *Palaeotragus*, *Equus* cf. *altidens*, *Bos* sp., *Dama* sp., *Stephanorhinus etruscus*, *Bison* sp., *Leptobos, Cervalces*; Boulbes et al., 2014). The presence of early Palaeolithic material is also reported in the area, such as an artefact of a hard-hammer flake (Maddy et al., 2015) and clactonian flakes (Aytek et al., in press).

Another example that highlights the Levant corridor as one preferred path of hominin migration between Africa and Eurasia is located in Gesher Bernot Ya'akov in Israel (Fig. 12). It has been dated using the presence of the MBB at 780 ka (Goren-Inbar et al., 2000). Its Acheulian assemblage is dominated by cleavers and bifaces and shows strong affinities to Africa (Bar-Yosef and Belmaker, 2011). It has been interpreted as a climatically driven migration wave originating from Africa and ceasing in the Levant (Bar-Yosef and Belmaker, 2011). These developed cleaver rich assemblages are rare north of the Levant corridor, Kaletepe Deresi 3 being one of the exceptions (Doronichev and Golovanova, 2010).

The Treugol'naya Cave in the northern Caucasus, is an eastern European site with Lower Paleolithic, core-flake tools, which has been dated by electron spin resonance (ESR) to between 364 ± 11 and 583 ± 25 ka (Doronichev and Golovanova, 2010). Interestingly, the Treugol'naya Cave assemblage lacks Acheulian bifaces.

In the Southern Caucasus, the Acheulian is well represented in a number of excavated caves: Azokh, Koudaro and Tsona (Liubin and

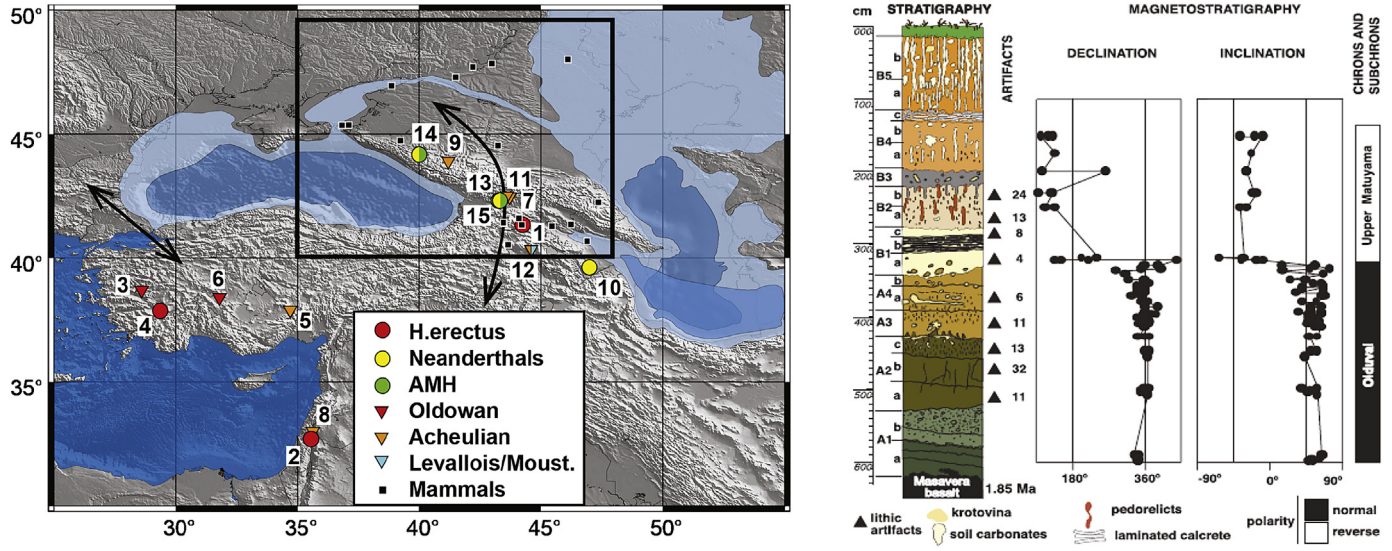

Fig. 12. Left: Paleogeographic map with the location of hominin-related sites in the study area. Numbers correspond to site description in Table 1. Arrows show migration pathways out of Africa and into Europe. Right: Stratigraphy, archaeology and paleomagnetic dating of the oldest hominin findings in Dmanisi, Georgia (from Ferring et al., 2011).

Bosinski, 1995; Liubin, 2002; Bar-Yosef and Belmaker, 2011; Fernández-Jalvo et al., 2016). At Azokh, a number of levels record the evolution from the early to the late Acheulian, starting with few core-choppers. The richest layers (VI and V) present a late Acheulian industry including bifaces performed in a Levallois way. Level V also yielded a fragment of *Homo* mandible (Gadzhiev and Gusseinov, 1970; Liubin and Bosinski, 1995; Liubin, 2002).

At the Koudaro cave, the layer 1 of Koudaro I is characterized by an abundant late Acheulian industry, including side scrapers, core-choppers, elongated bifaces and flake cleavers. Also three human teeth appeared associated with this industry. TL datings provided consistent ages of 360 ± 90 ka and 350 ± 70 ka. However, a reverse interval has been identified at level 6, immediately below the Acheulian layers, where an early Middle Pleistocene (Galerian) mammalian fauna was recognized (Liubin and Bosinski, 1995). This suggests that the Acheulian layers are older than the Matuyama-Brunhes boundary (> 780 ka). At Kudaro III, layers 6-8 delivered a more restricted Acheulian assemblage, including bifaces and flake tools. TL dating of level 5 provided a and age of 252 ± 51 ka and 245 ± 49 ka, while dating of level 8a resulted in 560 ± 112 ka (Liubin, 2002).

The large Tsona cave, located in South Ossetia (Georgia) at an altitude of 2150 m, includes an Acheulian succession, starting with a lower level (layer 7) that delivered a small early Acheulian assemblage. Overlying this level, level 6 produced a rich late Acheulian sample of about 100 artifacts, including 29 bifaces. The chronology of these deposits is not well established. The scarce rodent content suggests at least an early Middle Pleistocene or a late Early Pleistocene age for some of the lower levels of this cave. The Lower to Middle Paleolithic transition has been identified in Nor Geghi 1, Armenia (Adler et al., 2014). The Nor Geghi 1 tools are thought to show the synchronous use of Acheulian bifaces and more developed Levallois technology for the first time outside Africa (Adler et al., 2014).

The Neanderthal record is by far too sparse in this region to make any detailed statements. However, Ortvale Klde, Georgia, represents an exceptional site to study Neanderthal-modern human interactions at the Middle to Upper Paleolithic transition (Adler et al., 2008). Rare findings of Neanderthals in the Caucasus region (Mezmaiskaya Cave) have been interpreted to represent a refugia role (Stewart and Stringer, 2012). Additionally, comparison of the lithic assemblages from the northern and southern slopes of the Caucasus (Ortvale Klde versus Mezmaiskaya) yield a transition from dissimilar to similar going from Middle to Upper Paleolithic, which suggests that the Caucasus represented a biogeographical and social barrier for the Neanderthals, whereas the modern humans spread across it much more easily (Bar-Yosef et al., 2006).

Ultimately, the Pontocaspian region might not represent just a segment along the migration path between Africa and Eurasia, but rather be a favourable place, where hominins lived repeatedly for several periods during most of the Pleistocene.

4. Caspian Sea and Black Sea coevolution; open questions and new directions

The Pontocaspian domain experienced numerous lake/sea-level variations and prolonged periods of isolation and short intervals of connection during the Plio-Pleistocene and Holocene due to its sensitivity to internal forcing factors (e.g., Bezrodnykh et al., 2004; Rychagov, 1997; Svitoch, 2013b). In these semi-isolated basins, sea-level changes were often of larger magnitude and higher frequency compared to open marine settings (Kroonenberg et al., 1997, 2000). Here we examine these sea-level changes in a framework of Caspian Sea and Black Sea coevolution and interbasinal connectivity. We furthermore discuss several open questions regarding the paleogeographic evolution of the Pontocaspian region and describe the present-day and future perspectives of the potential environmental hazards the region may experience.

4.1. *The Akchagylian flooding*

During the Pliocene-Quaternary, major sea-level variations of several hundreds of meters have been inferred for the Caspian Basin. After a long period of low sea-level during the Pliocene Productive series, there have been at least six major transgressive stages distinguished (Kroonenberg et al., 1997; Yanina, 2012a), some of which resulted in overflows towards the Black Sea Basin. The Akchagylian flooding is one of the most extreme events of sea level rise in the Pontocaspian domain, but still subject of considerable debate.

Despite inconsistencies in the definition of the Akchagylian there are several common points which are more or less universally agreed upon: 1) the Akchagylian was marked by a large scale transgressive event, 2) there was an increase in the salinity of the water during parts of the Akchagylian compared with the water in which the underlying and overlying strata were deposited, and 3) a switch in environmental

Table 1
Key sites and locations in the Pontocaspian (and surroundings) with hominin findings.

Site	Age	Type			Tools				Location		Position rel. Great Caucasus		Reference
		Hominins							Lat N	Long E			
	[ka]	H.erectus	H.neander	AMH	Oldowan(I)	Acheulian(II)	Levallois	Mousterian(III)	[°]	[°]	south	north	
1. Dmanisi	1800	x			x				41,34	44,24	x		Ferring et al. (2011)
2. Ubeidiya	1400-1200	x				x	x		32,70	35,56	x		Belmaker et al. (2002)
3. Gediz River	1170-1240				x				38,70	28,60		x	Maddy et al. (2015)
4. Kocabas	> 1100	x							37,87	29,36		x	Lebatard et al. (2014)
5. Kalepe Deresi 3	1000					x			37,90	34,70		x	Slimak et al. (2008)
6. Dursunlu	990-780				x				38,40	31,80		x	Güleç et al. (2009)
7. Tsona	~780					x			42,46	43,72	x		Mgeladze & Moncel (2016)
8. Gesher Benot Ya'akov	780 ka					x			33,01	35,63	x		Goren-Inbar et al. (2000)
9. Treugol'naya	600-350					x (no bifaces)			43,90	41,20		x	Doronichev and Golovanova (2010)
10. Azokh/Azych	300??-100		x			x			39,62	46,99	x		Fernández-Jalvo et al. (2016)
11. Koudaro	250-560					x			42,50	43,60	x		Mgeladze & Moncel (2016)
12. Nor Geghi1	200-400					x	x		40,34	44,60	x		Adler et al. (2014)
13. Ortvale Klde	33-50(38)		x	x				x	42,31	43,26	x		Adler et al. (2008)
14. Mezmaiskaya Cave	33-44(32)		x	x				x	44,17	40,00		x	Golovanova et al. (1999)
15. Dzudzuana	27-24		x	x				x	42,31	43,32	x		Bar-Yosef et al. (2006)

conditions caused a observed change in faunal assemblages, primarily based on malacofauna (Andrusov, 1902; Danukalova, 1996), although zoo- and phytoplankton also reacted to the event (Nevesskaya et al., 2003; Van Baak, 2015). There are several well studied sections of the Akchagylian in the Caspian catchment (Yurevich, 1966; Danukalova, 1996; Danukalova et al., 1998; Van Baak et al., 2013; Forte et al., 2015), but only a few coeval exposures are known from the Black Sea region, making correlation between the two basins problematic.

The Akchagylian lake expansion covered vast areas of southern Russia, Kazakhstan, Turkmenistan, Iran, Azerbaijan and Georgia (Popov et al., 2006) and is in many places characterised by extensive shale deposits. A transgression on the scale of the Akchagylian would have required that vast amounts of water entered the basin. There are two ways this could have happened: a climatically controlled switch to a more positive hydrological budget or connection with overspill from an adjacent waterbody (Fig. 13).

A transgression during the Akchagylian could be achieved by an increase in catchment precipitation (either directly through rainfall or *via* snowfall, glacier formation and melt), possibly combined with a reduction in precipitation over the lake surface. Given its vast northern drainage, a positive water budget in Eastern Paratethys is a perfectly plausible phenomenon that has previously been cited as a driver of inter-basin connection dynamics (Marzocchi et al., 2016). However, only fresh water can be delivered in this fashion and therefore it cannot account for the apparent high salinities or transfer of fauna and plankton. Although a positive water budget could have played a role in the transgression, it is not possible to increase lake volume and salinity simultaneously through this mechanism. Salt dissolution from diapiric exposures in the northern Caspian lowlands could potentially have contributed to the salinity increase but would probably only do so to a minimal degree. A connection with a more saline water body is therefore necessary to explain the Akchagylian phenomenon in full.

There are three potential directions of marine connection between the Caspian Sea and the global oceans (Fig. 13); from the south (Indian Ocean), from the west (*via* the Mediterranean and Black Sea) and from the north (Arctic Ocean). A connection between the Caspian Sea and the Indian Ocean has previously been considered as a possibility during the Pontian, either through the region of modern day Pakistan, Afghanistan and Turkmenistan, or through modern day Iran *via* the Persian Gulf (Van Baak et al., 2016a). However, it is probable that the mountains of the Alpide Belt (specifically the Talysh, Alborz, Zagros and Hindu Kush) had been uplifted sufficiently to prevent this connection by the time of the Akchagylian transgression (Axen and Lam, 2001; Ballato et al., 2011). Although this route is not ruled out completely, a huge rise in global sea level would have been required in order to breach this barrier and there is no evidence for a sea level rise of this magnitude at a global scale (Lisiecki and Raymo, 2005; Miller et al., 2005; Rohling et al., 2014). Paleontological indicators to support a southern connection are lacking in the Caspian Basin. It is therefore quite unlikely compared with the remaining options.

The Black Sea is a potential pathway for connection with the Caspian given that such a connection occurred during the Pontian (Van Baak et al., 2016a; Grothe et al., 2018), and a number of times after the Akchagylian (e.g. Yanina, 2014). This mechanism could allow transgression, salinity increase and faunal/plankton transfer. However, connection to the Black Sea (with a contribution from the Black Sea rather than purely overspill from the Caspian) would require global eustatic sea-level to rise above the minimum elevation of the gateway region (probably *via* the Kuma-Manych spillway, present day height 27 m above global sea level (Svitoch, 2010a)). Since this is a tectonically active region, sill height was likely different in the past (Svitoch, 2010b; Svitoch and Makshaev, 2011). Nevertheless, a Black Sea to Caspian connection could only occur during a global sea-level high-stand, i.e. during an interglacial episode or during a high stand in the Black Sea (depending on the connectivity of the Black Sea with the Mediterranean). Furthermore, there are not many accessible Akchagylian

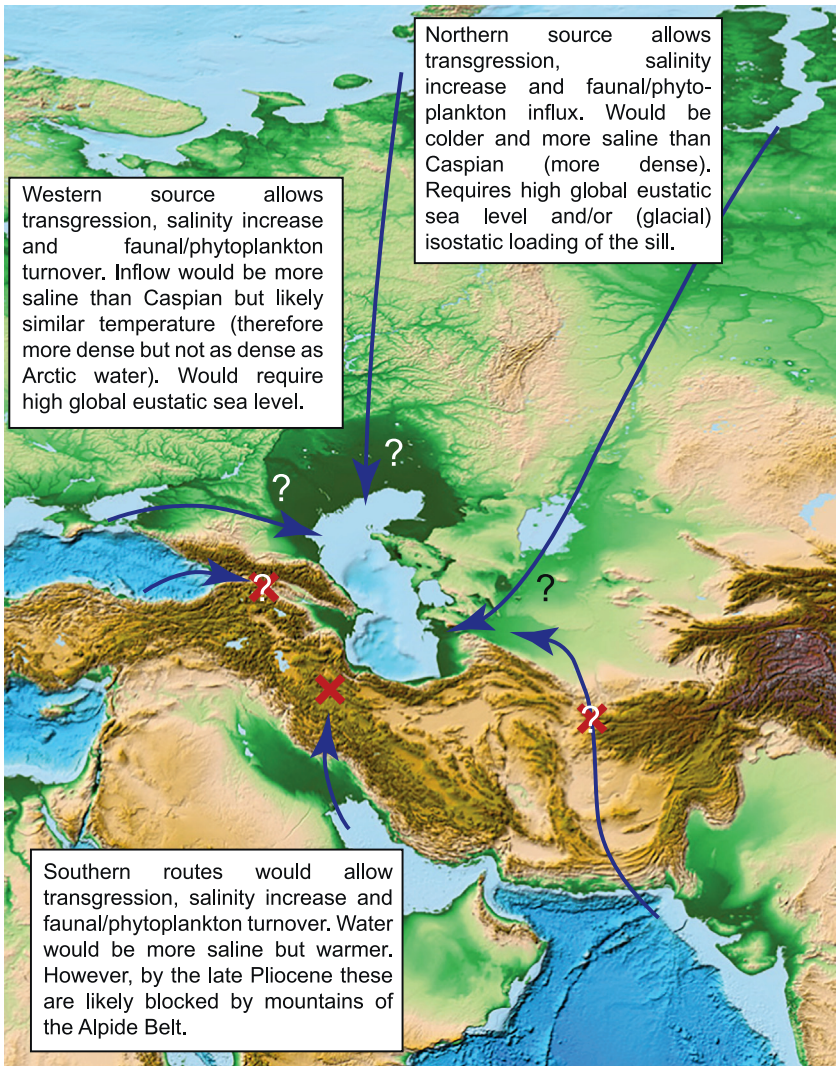

Fig. 13. Map showing potential Akchagylian transgression mechanisms and possible connection points with brief outline of the requirements for each and the observed phenomena each would be able to explain. The main mechanisms discussed are climate driven base level rise and connection with an adjacent basin. Three potential adjacent water bodies are considered as the source of incoming water (Arctic Ocean, Indian Ocean and the Black Sea). For each of the three two potential gateways are suggested.

equivalent outcrops in the Azov/Black Sea region making it difficult to draw a correlation between the contemporary faunas of the Black Sea and Caspian basins. Marine conditions were lacking in the Black Sea Basin during the Late Pliocene. The fact that the Akchagylian mollusc fauna only extends westward as far as the Sea of Azov (Danukalova, 1996) and not, as far as we know, into the Black Sea might cast doubt on whether there was a true connection between the Mediterranean-Black Sea system and Caspian Sea or whether the inferred connection with Azov was caused purely by overspill from the Caspian. It is therefore unlikely that the very high lake levels and relative high salinities of the Akchagylian can be explained by a transgression through the Black Sea Basin.

The final option would be a connection to the Arctic Ocean, either west of the Urals, from the Barents Sea and over the Russian Platform or east of the Urals from the Kara Sea via the area that is now occupied by the Aral Sea (Fig. 13). In common with the western connection, a northern source is consistent with the observed transgression, a number of faunal indicators and salinity increase (although we are not currently aware of any contemporary time-equivalent sections in the Arctic region that could be used to test this hypothesis). Presence of fauna that are genetically related to Arctic species appears to support a connection with the Arctic during the Late Pliocene. Within the Akchagylian, levels with boreal-arctic foraminifera were found that match a proposed northern marine connection (Agalarova, 1976; Van Baak, 2015; Richards et al., 2018). The modern day threshold between the Caspian and the Arctic is at 125 m above global sea level (Esin et al., 2016), so high eustatic sea level being the main driver of such a connection is impossible, even during full interglacials. However, isostatic lowering of the sill height due to loading by northern hemisphere ice sheets may help to explain a potential northerly connection (Van Baak, 2015; Richards et al., 2018).

4.2. Origin of the Pontocaspian faunas

Modern Pontocaspian faunas (e.g. faunas dominated by extant Pontocaspian genera) assembled in the Black Sea Basin and Caspian Basin during the Early Pleistocene. Individual Pontocaspian taxa have a history well before the onset of the Quaternary. Some major diversifications of Pontocaspian taxa occurred already in the late Miocene, as is shown for Pontocaspian gobies (Schwarzhans et al., 2017).

Most Pontocaspian lineages appear to derive from central Paratethys and possibly Anatolian ancestors. With the onset of the Pannonian age (c. 11.6 Ma; ter Borgh et al. (2013)) in Hungary and adjacent territories, a continuous series of anomalohaline long-lived

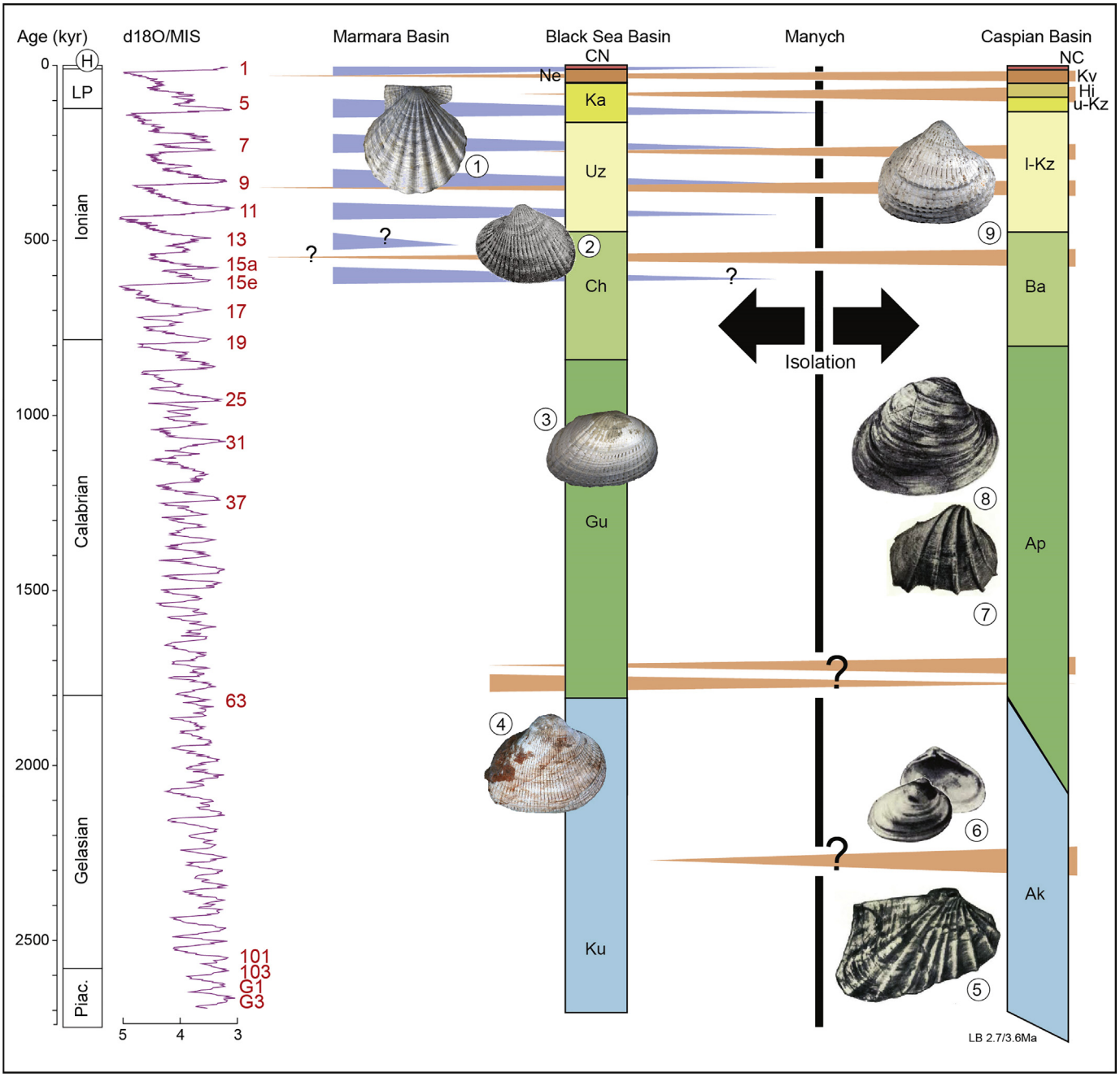

Fig. 14. The successive mollusc faunas in the Pontocaspian region provide clear insights into basin connection and isolation. Age and MIS zonation from Cohen et al. (2013). H = Holocene; LP = Late Pleistocene. Regional phases: CN = Chernomorian; Ne = Neoeuxinian; Ka = Karangatian; Uz = Uzunlarian; Ch= Chaudian; Gu = Gurian; Ku = Kujalnikian; NC = Novocaspian; Kv = Khvalynian; Hi = Hircanian; u-Kz = upper or late Khazarian; l-Kz = lower or early Khazarian; Ba = Bakunian; Ap = Apsheronian; Ak = Akshagylian. Representative bivalve species: (1) *Flexopecten glaber*; (2) *Didacna pseudocrassa*; (3) *Didacnocatilus guriensis*; (4) *Submonodacna pleistopleura*; (5) *Avicardium rariiferum*; (6) *Avimactra subcaspia*; (7) *Apsheronia raricostata* s.l.; (8) *Apsheronia propinqua* s.l.; (9) *Didacna subpyramidata* (1) from Büyükmeriç et al., 2016; (2) from Nevesskaya, 2007; (5-8) from Kolesnikov, 1940.

lakes developed in central-eastern Europe whose faunas are dominated by lymnocardiine radiations (Magyar et al., 1999; Popov et al., 2006; Harzhauser and Piller, 2007). The Pontocaspian lake system forms the final phase of this Paratethys system.

Some Pontocaspian groups have an arctic origin (e.g., Caspian seals, gammariid crustaceans: Kosarev and Yablonskaya, 1994; Palo and Väinölä, 2006). The introduction of the arctic elements in the Pontocaspian region likely occurred during the Akchagylian as shown by the occurrence of arctic-boreal foraminifera (Agalarova et al., 1940; Van Baak, 2015; Richards et al., 2018). This age fits very well with molecular clock estimates of the divergence of the Caspian Seal from its arctic sister species (Palo and Väinölä, 2006) of 2-3 Ma.

In the Caspian Sea Basin, the Akchagylian-Apsheronian boundary forms the base of the Pontocaspian interval (Fig. 14). The Akchagylian was dominated by an excessive radiation of mactrid and cardiid bivalves, the latter being unrelated to modern Pontocaspian genera (Danukalova, 1996). These were entirely replaced by Apsheronian faunas dominated by modern Pontocaspian bivalve genera such as *Didacna*, *Monodacna*, *Adacna* and *Hypanis* (Kolesnikov, 1940). At this stage it is unknown why such a drastic and apparently sudden reorganization of biota occurred in the Caspian Sea Basin at the Akchagylian-Apsheronian boundary, yet a major environmental upheaval

must have been involved.

In the Black Sea Basin, the onset of the Pontocaspian phase is gradual and spans the Kimmerian-Kuyalnikian-Gurian transition. Kuyalnikian bivalve faunas are dominated by *Didacna* and *Dreissena* whereas Kimmerian faunas are dominated by ancestral Paratethyan bivalve genera such as *Congeria*, *Lymnocardium*, *Prosodacna* and *Paradacna*. The Early Pleistocene Gurian fauna has a Pontocaspian character.

The origin of several dominant Pontocaspian genera is uncertain as broad agreement on their taxonomic delimitation is lacking. Miocene and Pliocene records of *Didacna* and *Monodacna* exist for the Pannonian, Dacian and Pontian Basins (Wesselingh et al., 2008), all or part of these probably concern different genera and are in need of revision. The oldest records of *Didacna* and *Monodacna* appear to be of Pliocene age from two adjacent basins in SW Anatolia: Baklan Basin (*Monodacna*: Wesselingh and Alçiçek, (2010)) and Denizli Basin (*Didacna*: Wesselingh et al., 2008), who at the time estimated the stratigraphic age as Late Miocene). The role of these Anatolian basins in the origin of modern Pontocaspian biota remains to be established still (Wesselingh and Alçiçek, 2010).

In summary, Pontocaspian species are species that evolved within the Black Sea – Caspian Sea basins in the past few million years under anomalohaline to fresh water conditions. Many of the species form lineages that typically go back into the Miocene Paratethyan Basins and these are termed Pontocaspian lineages. However, modern Pontocaspian communities that are dominated by the species of the genera *Didacna, Monodacna, Laevicaspia* and *Turricaspia* only emerged in the Early Pleistocene of the Caspian and Black Sea Basins.

4.3. Faunal indicators of basin isolation and connectivity

Isolation and connection of basins can be very well determined by the amount of species bei99ng shared between adjacent basins (Wesselingh, 2007; and references therein). Two neighbouring basins with similar ecological conditions will develop endemic faunas when separated long enough and will experience faunal exchange and homogenization when connected. Molluscs, ostracods and foraminifera with their good preservation potential are very well suited to assess the degree of isolation and interbasin connectivity of Pontocaspian basins.

Extensive bivalve and gastropod radiations occurred in the Caspian and Black Sea basins in the past 3 million years resulting in faunas rich in endemic species (Fig. 14). Turnover rates in several groups have been excessively high. For example, the Middle-Late Pleistocene contains at least seven well recognizable cardiid faunas in both the Black Sea Basin and Caspian Basin as a result of high origination and evolution rates. The gastropod evolution is less well-constrained due to a lack of comprehensive taxonomic works on the Quaternary faunas. Both groups do suffer from a taxonomic impediment and there is a long tradition of change and conflicting opinions of species boundaries. Resolving this taxonomic impediment is of utmost importance to increase their biostratigraphic value and to understand the isolation and connectivity phases.

Ostracod occurrences in the Caspian and Black Sea basins also reflect the evolution of the region since the latest Pliocene. High speciation rates were the result of variable and high environmental pressure. This resulted in species rich and morphologically diverse faunas reflecting environmental dynamics of each of the Pontocaspian basins. The main issue emerging from studying Neogene and Quaternary ostracod faunas from the Pontocaspian realm is the presence of conflicting taxonomic definitions. They originate mainly from ambiguous descriptions of ostracod species. The quality of hand- drawings, which were predominant in publications at the beginning of the 20th century, varies drastically and often the described species are beyond recognition. Due to confusing approaches in the course of the 20th century new species were created, often based on poorly recognizable morphological features that might represent ecological variation (e.g. salinities) rather than species characters per se. The great number of Pontocaspian taxa described therefore might be the result of a number of created morphotypes, that do not represent the true ostracod diversity. A taxonomic harmonization including re-evaluation of originally described species, based on type collections and sections is urgently required. Recently ostracodologists characterizing species of Pontocaspian origin have published a series of papers using modern and standardized imaging approaches such as morphometric analysis and/or high resolution SEM pictures (Boomer et al., 1996, 2010; Rostovtseva and Tesakov, 2009; Van Baak et al., 2013).

4.3.1. Isolation during the Productive Series-Kimmerian

During the deposition of the Productive Series in the Caspian Basin the dominant facies association was fluvial - lacustrine. Accordingly, freshwater taxa capable of surviving in low energy environments like deltaic areas or lakes replaced the previous brackish Pontian fauna. Ostracod assemblages have been recovered from the lacustrine intercalations at the top part of the Productive Series and they mainly consist of freshwater species as well as the euryhaline *Cyprideis torosa*. Besides ostracods, charophytic algae have been frequently observed. Virtually nothing is known about potential deep lake conditions in the South Caspian Basin at the time as these deposits are deeply buried. While the Productive Series in the Caspian Sea was deposited under fluvio-lacustrine conditions, the Black Sea Basin was an anomalohaline lake throughout the Kimmerian. There, the mollusc and ostracod fauna was dominated by “Pontian” type taxa and their direct descendants. No exchange of faunas existed between Caspian and Black Sea basins indicating prolonged isolation of the two basins.

4.3.2. Connectivity during the Akchagylian-Kuyalnikian

After the Akchagylian flooding event, the former freshwater dominated ostracod fauna of the Productive Series became replaced by an anomalohaline (“brackish”) water assemblage. These ostracod faunas are poorly preserved, especially in the basinal part, as a result of alternating dysoxic-and hypoxic conditions. Weakly calcified shells of candonids and limnocytherids dominate. Towards the top of the Akchagylian, dysoxic intervals become less frequent and the ostracod community becomes more diverse, with stronger calcified and better preserved valves. The Akchagylian ostracod fauna is dominated by a high number of endemic species that evolved in situ in the Caspian Basin.

During the Akchagylian a very diverse mollusc fauna with flocks of bivalve species of *Akchagylia* and *Akchagylocardium/Avicardium* evolved in the Caspian Basin (Danukalova, 1996). These faunas reached into the Azov region and Kerch region but have not been reported from elsewhere in the Black Sea Basin. At the same time the Black Sea Basin contained its own endemic Kuyalnikian mollusc faunas (Taktakishvili, 1987). Akchagylian salinity was higher (upper mesohaline to polyhaline) then the Kuyalnikian one (oligohaline to lower mesohaline), so intermittent connections may have failed to establish Akchagylian species into the lower salinity Kuyalnikian lake. The presence of two completely contemporaneous endemic bivalve faunas in the Akchagylian of the Caspian-Azov and Kuyalnikian of the Black Sea Basin strongly suggests prolonged isolation by a terrestrial corridor separating the Caspian from the Black Sea. The corridor must have been located in the Kerch-Crimea region. Almost the entire endemic Akchagylian cardiid and mactrid bivalve fauna went extinct at the end of the Akchagylian and became replaced by other genera in the subsequent Apsheronian in the Caspian Basin.

4.3.3. Isolation and connectivity during the Apsheronian-Gurian

Most of the Early Pleistocene mollusc faunas in both Black Sea and Caspian basins were completely dissimilar with only very limited exchange of species (for example the bivalve genus *Didacna* migrated in the Caspian Sea from the Black Sea; Nevesskaya (2007)). Gurian faunas of the Black Sea Basin have almost nothing in common with the time

equivalent Apsheronian of the Caspian Basin. Both faunas were rich in species and endemic to their respective basins. The continued presence of almost totally different and endemic faunas again shows the prolonged isolated nature of the basins during the Early Pleistocene. The first occurrence of typical Pontocaspian genera such as *Turricaspia*, *Laevicaspia, Didacna, Monodacna, Adacna* and *Hypanis* in the Apsheronian is not well understood and subject to further studies. Some of the genera, such as *Monodacna*, have records extending back in the Pliocene of southwestern Turkey (Wesselingh and Alçiçek, 2010). Remarkably, no apparent spatio-temporal populations are known from the Black Sea Basin before the introduction of this species in the Caspian Basin.

During the Apsheronian stage, saline conditions prevailed in the Caspian Basin with fresher conditions in marginal deltaic and coastal areas. The base of the Apsheronian is characterized by the appearance of euryhaline foraminifer *Ammonia* spp. suggesting the existence of another but minor marine connection event. Additionally, species of *Cibicides* and *Cassidulina* were observed in lower abundances. Ostracod species diversified during the Apsheronian. Many Akchagylian species further evolved into Apsheronian endemics and became more abundant.

4.3.4. Isolation and connectivity during the Middle-Late Pleistocene

This situation dramatically changed during the Middle and Late Pleistocene when Caspian overflows repeatedly connected the Caspian and Black Sea basins and faunas (Svitoch et al., 2008; Badertscher et al., 2011). Caspian species groups (like *Didacna* species) extended from time to time their ranges into the Black Sea Basin and at times even into the Marmara Basin and its satellite basins (Islamoğlu, 2009). However, the number of suggested overflows from the Caspian to Black Sea by Baderstcher et al., 2011, has not been observed in the Black Sea strata that suggest fewer overflow events (Yanina, 2006, 2012a). The lower Bakunian transgression in the Caspian Basin coincides with abundant occurrences of euryhaline foraminifera such as *Ammonia* spp., *Nonion* sp. and *Cibicides lobatulus*. The ostracod community was diverse, dominated by anomalohaline endemic species. Several Bakunian ostracod species from the Caspian Basin also are common in Chaudian and Uzunlarian deposits of the Black Sea indicating faunal interchange. In the Black Sea Basin these species persisted up to Neoeuxinian times. The Bakunian ostracod fauna starts to resemble the fauna from the Late Pleistocene - Recent Caspian realm and the Late Pleistocene ostracod assemblages from the Black Sea region. Their presence in both realms is indicative of repeated overflows from the Caspian Sea. During interglacial intervals the Black Sea reconnected to marine settings of the Mediterranean and Pontocaspian species became suppressed in marginal habitats along the north coast.

Marine waters reached the Black Sea Basin during interglacial highstands since MIS13 or MIS15 (Middle Pleistocene). They reached well into the Manych Strait during MIS5e but never crossed the threshold to the Caspian Basin (Fedorov, 1978b; Svitoch et al., 1998, 2010). Caspian overflow events during the Middle-Late Pleistocene time interval were of short duration and most of the time the Black Sea Basin was isolated. In the Middle-Late Pleistocene *Didacna* species did establish in the Black Sea Basin, but usually did not persist long after cessation of connection events. The last Caspian overflow dates to c 15 ka (Khvalynian).

4.4. Geochemical proxies of interbasinal connectivity

Periods of basin isolation and/or connection can in theory also be resolved by studying the geochemical record, for instance by the strontium ($^{87}Sr/^{86}Sr$) and oxygen ($\delta^{18}O$) isotope ratios, providing sections are well dated. The Black Sea and the Caspian Sea basins have experienced numerous hydrological and environmental changes resulting from phases of isolation and connection both to each other and to the open ocean. Here we discuss the strontium and oxygen isotopic records of the Black Sea and the Caspian Sea as an alternative approach to reconstructing the connectivity history of these basins. The changing pattern of connectivity between individual basins and the open ocean, is key to reconstructing the past environmental and hydrological dynamic of the region and for understanding the natural variation in biodiversity.

4.4.1. Strontium isotopes

Strontium (Sr) isotopic ratios ($^{87}Sr/^{86}Sr$) in an open marine setting have been used as a stratigraphic tool (Hess et al., 1986; Crame et al., 1999; McArthur et al., 2012) because although this ratio in ocean water changes through geological time, at any given time, the $^{87}Sr/^{86}Sr$ is the same everywhere in the global ocean (Elderfield, 1986). This is because the residence time of the Sr isotopes in the global ocean is longer than the ocean mixing time. Sr isotopic ratio as a dating tool, however, becomes problematic in marginal marine or lacustrine basins as the strontium isotopic composition of the basin may be dominated by non-oceanic sources like river and ground water leading to an $^{87}Sr/^{86}Sr$ that differs from the global ocean (e.g., Ingram and Sloan, 1992; Vonhof et al., 1998; Flecker et al., 2002). The Sr concentration and isotope ratios of river and ground water depends on its catchment geology and this can vary substantially between different drainage basins (Palmer and Edmond, 1992; Stein et al., 1997; Krom et al., 1999; Major et al., 2006). For marginal and isolated basins then, $^{87}Sr/^{86}Sr$ of a basin is controlled by the input and mixing of different water masses that have distinct isotopic signatures and Sr concentrations, i.e., freshwater and ocean water (Albarède and Michard, 1987; Vonhof et al., 1998; Major et al., 2006; Kober et al., 2007; Holmes et al., 2010; Vasiliev et al., 2010). The $^{87}Sr/^{86}Sr$ ratio of the water is incorporated and preserved by organisms that precipitate carbonate shells, e.g, foraminifera, ostracods and molluscs. Critically, unlike the better-known stable oxygen isotopes, Sr isotopes do not appear to be fractionated by biological processes such as vital effect (Chivas et al., 1985; Reinhardt et al., 1998a) or by water phase changes (Åberg, 1995; Capo et al., 1998). Consequently, $^{87}Sr/^{86}Sr$ measured in well preserved fossil carbonate can be used to evaluate quantitatively the relative input of different source water into the basin (Ingram and Sloan, 1992; Reinhardt et al., 1998b). In addition, the $^{87}Sr/^{86}Sr$ record through time of two (or more) adjacent sub-basins can be used to reconstruct the connectivity history between them. At its simplest, the $^{87}Sr/^{86}Sr$ record of two basins will show similar values during periods of connection but will diverge from each other during periods of isolation if the water inputs to each basin are different.

4.4.2. Oxygen isotopes

The oxygen isotopic composition ($\delta^{18}O$) of a marginal marine or lacustrine basin depends on several factors such as the isotopic composition of precipitation and evaporated water, as well as river runoff, groundwater and oceanic input (Craig et al., 1963; Dansgaard, 1964; Stuiver, 1970). Like Sr isotopes, organisms living in the water preserve the $\delta^{18}O$ of the water during carbonate precipitation. However, unlike Sr, oxygen isotopes are affected by species specific vital effects, i.e different species fractionate the two main stable isotopes of oxygen (^{18}O and ^{16}O) differently. Combination of these factors affect the $\delta^{18}O$ preserved in fossil carbonates, and therefore, teasing apart the individual signals that impact the $\delta^{18}O$ of a basin is quantitatively complex and open to interpretation. This is particularly problematic when trying to reconstruct past connectivity in basins where fossil fauna is extinct, forcing reliance on the assumption that nearest living relatives will have the same vital effect fractionation. However, the use of $\delta^{18}O$ of various source waters into the basin as a baseline can provide qualitative insight into the hydrological changes of the basin. In principle, higher $\delta^{18}O$ of the basin water indicates input of and mixing with isotopically heavy marine water (enriched in $\delta^{18}O$) or strong evaporation in an isolated setting. Lower $\delta^{18}O$ of the basin water indicates increased input of isotopically light freshwater (depleted in $\delta^{18}O$) such as precipitation derived from high latitudes and high altitudes, glacial and

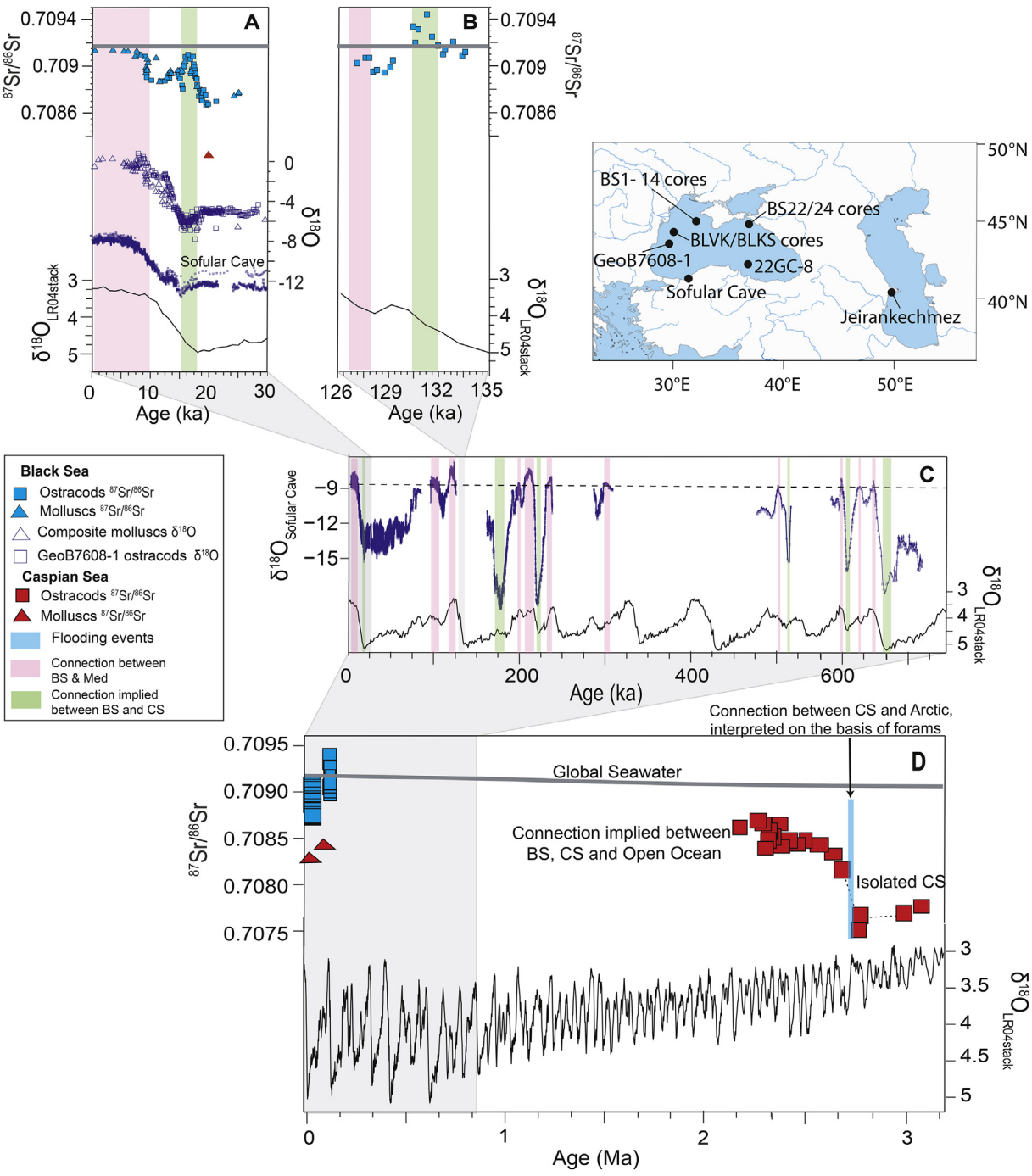

Fig. 15. The $^{87}Sr/^{86}Sr$ and $\delta^{18}O$ data from the Black Sea (blue) and Caspian Sea (red) for different time slices (A - D). Different symbols represent different archives; ostracods (boxes), mollusc (triangles), speleothem (circles). The $^{87}Sr/^{86}Sr$ of the global ocean is obtained from McArthur et al. (2012), for the Black Sea is obtained from Major et al. (2006) and Wegwerth et al. (2014), for the Caspian Sea obtained from Page (2004) and Van Baak (2015). The global benthic $\delta^{18}O$ is obtained from Lisiecki and Raymo (2005), for the Black Sea is obtained from Bahr et al. (2006), Major et al. (2006) and the $\delta^{18}O$ of the Sofular Cave is obtained from Badertscher et al. (2011).

inland river runoffs, reflecting the broad latitudinal change in $\delta^{18}O$ driven by the global hydrological cycle (Dansgaard, 1964).

4.4.3. Interbasinal connectivity

The geochemical evidence of the connectivity between the Black Sea, the Caspian Sea and the Open Ocean between 3 Ma to 700 ka is limited but subject of ongoing studies. There are no oxygen isotopic records and very few strontium isotopic data available for this period. Van Baak (2015) conducted strontium measurements for the Pliocene-Pleistocene transitional samples (3.2 - 2 Ma) in the Caspian Basin (Fig. 15). There are no data from the Black Sea Basin during this period. Van Baak (2015) suggested that the Caspian Basin was isolated until about 2.7 Ma and at least locally dominated by rivers with low $^{87}Sr/^{86}Sr$. At about 2.7 Ma, the Sr isotopic ratios rise substantially before stabilizing at values that are consistent with Black Sea ratios measured on samples from more recent periods when the Black Sea was isolated from Mediterranean Sea (e.g., during LGM). Van Baak (2015) suggested that the Caspian Sea was progressively connected to the Black Sea during this period with additional source with high $^{87}Sr/^{86}Sr$, likely from the ocean. The likely route for the ocean water into the

Caspian Sea at about 2.7 Ma, is discussed in Section 4.1. No faunal evidence exists for extensive connections between the Caspian and Black Sea Basin at the time though.

Sparse data are available for both strontium and oxygen isotopes for the last 700 ka. However, the data is exclusively from the Black Sea and the surrounding area and heavily dominated by studies that focus on the last 30 ka (Aksu et al., 2002; Bahr et al., 2005, 2006, 2008; Major et al., 2006; Nicholas et al., 2011; Piper and Calvert, 2011; Mudie et al., 2013; Constantinescu et al., 2015; Yanchilina et al., 2017). Currently, the Black Sea is connected to the Mediterranean Sea via the Marmara Sea and the strontium and oxygen isotopic values of the present day Black Sea are 0.709133 ± 0.000015 and -2.4 ‰, respectively (Fig. 15a; Major et al., 2006). The strontium isotopic record for the last 30 ka shows that the Black Sea was an isolated basin during the last glacial maximum (Major et al., 2006). This is consistent with the oxygen isotopic record from the western Black Sea (Bahr et al., 2006, 2008; Badertscher et al., 2011). Increased $^{87}Sr/^{86}Sr$ along with the depleted (negative) $\delta^{18}O$ during the termination of the LGM (18 – 16 Ka) has been suggested to result from increased freshwater input into the Black Sea, either from the northern Black Sea rivers or Caspian Sea spill over caused by high melt water delivered by the Volga (Major et al., 2006; Bahr et al., 2008). The final shift of the $^{87}Sr/^{86}Sr$ and $\delta^{18}O$ in the Black Sea toward the present day value at 9.4 ka marks the Holocene connection of the Black Sea to the open ocean. However, there are studies conducted in the Black Sea that provide slightly different timing for the last Black Sea-Mediterranean connection (see Section 2.2.7).

A multi proxy study (ice-rafted debris, organic and inorganic sediment geochemistry, $TEX_{86}$and $U^{K'}_{37}$ derived water temperatures) of the Black Sea during the penultimate glacial and the Eemian interglacial (133.5 – 122.5 ka) shows that the Black Sea experienced two meltwater pulses during the Termination II and was connected to the Mediterranean Sea during the Eemian (Fig. 14b; Wegwerth et al., 2014). Increased $^{87}Sr/^{86}Sr$ during the second melt water pulse (131.5–130.5 ka) was interpreted as resulting from the increased influx from the Amu Darya river in the Caspian Basin and subsequent overflow towards the Black Sea Basin during the deglaciation (Wegwerth et al., 2014). The authors suggest that the water from the Amu Darya, which is fed by highly radiogenic Sr from western Himalayas (Henderson et al. (1994) and references therein), entered the Caspian Sea and finally connected the Black Sea with the Caspian Sea via the Manych Depression. This assumption is supported by studies suggesting that the Amu Darya river directly entered the Caspian Sea at the beginning of the Holocene (Boomer et al., 2000). However, no time equivalent record for the Caspian Sea exists to prove this hypothesis.

The only long-term (0-700 ka) oxygen isotopic record (Fig. 15c) from the Black Sea comes from well-dated speleothems collected from Sofular Cave, located 10 km from its southern coast in northern Turkey (Badertscher et al., 2011)). The $\delta^{18}O$ record of the Sofular Cave shows twelve episodes when the values surpassed $\delta^{18}O$ of -8.5 ± 1‰. The authors correlated the evolution of the $\delta^{18}O$ of the Sofular Cave with the $\delta^{18}O$ of the Black Sea and suggest that -8.5 ‰ is the characteristic $\delta^{18}O$ value for the Black Sea when it is connected to the Mediterranean Sea. The $\delta^{18}O$ record thus implies twelve connectivity periods between the Black Sea and the Mediterranean Sea during this time. On the other hand, the authors interpret the excursions towards isotopically depleted $\delta^{18}O$ values as reflecting inflow from the Caspian Sea.

Overall, where both strontium and oxygen isotopic records exist for the same time period, they are in accord with each other and show that the Black Sea and Caspian Sea are dynamic basins, with multiple periods of isolation and connection. The Sr isotopic data are more sensitive to the connectivity between the basins. On the other hand, the oxygen isotopic data depend on several factors including regional climate and ice volume, thus making the interpretation of the data and robust conclusions about the connectivity between the basins very difficult. The combination of $^{87}Sr/^{86}Sr$ and the $\delta^{18}O$, however, provides an untapped potential to tease apart the relationship between the connectivity and its two major drivers, i.e., climate and tectonics. Currently new isotope time series are being developed for both basins that will further improve our understanding of the connectivity history.

4.5. Drivers of sea level change in the Pontocaspian domain

The lake/sea level evolution of the Black Sea Basin and Caspian Basin is very complex resulting from different driving processes and settings. Disentangling the contribution of the various drivers of lake/sea level evolution through time is an ongoing effort that combines climate modelling and a variety of geological and geomorphological approaches.

The two main Pontocaspian gateway regions (Bosphorus and Manych) are affected by active tectonics (Ryan et al., 2003; Le Pichon et al., 2016). Furthermore deposition and erosion shaped these gateways too (Belousov and Enman, 1999; Svitoch and Makshaev, 2011). Tectonic activity may affect vertical land movement and force basin shape and bathymetric changes. However, such tectonic movements are generally slow (< 1 mm/year) (Rodionov, 1994; Elguindi and Giorgi, 2005; Giorgi, 2006; Svitoch, 2013a). Thus, their impact on short-term lake level changes is very small compared to climate and sea level change.

The Caspian Sea has experienced large variations in water level throughout its history (Fig. 16). In the 20th century alone a 3m drop and subsequent rise in lake level occurred in times of decades (Kroonenberg et al., 2007, 2008; Chen et al., 2017). On millennial time scales regressions occurred of over 100 m deep. During most of the Pliocene-Quaternary history the Black Sea Basin was disconnected from the open ocean, hence its lake levels were driven by global sea levels only during short highstand episodes (Badertscher et al., 2011; Yanina, 2012a). In intervening periods similar processes as those in the Caspian drove lake level variations.

Lake level variations in the Pontocaspian domain mainly thus depend on hydro-climatological processes resulting in changes of the evaporation/precipitation balance of the basins and their catchments (Rodionov, 1994; Kroonenberg et al., 2005; Esin et al., 2010; Yanina, 2014; Yanina et al., 2017a). Apart from rain and evaporation, significant contributions to the hydrologic budget of these basins occur in periods of glacial and permafrost development and demise (Rodionov, 1994; Kroonenberg et al., 2008; Yanina, 2012a)). Increased lake areas did increase the potential for evaporation.

Today, water supply is mostly controlled by runoff contributions from Eurasian rivers (e.g., Volga, Danube, Don, Dniester and Dnieper) with some additional direct contribution of rain over the lakes. The river runoff is determined by regional climate systems and responds to global climate change. On multi-millennial timescales there has been increased river discharge during warm interglacial periods. Further contributions occurred due to melting of the ice sheets during periods of deglaciation. Furthermore, “extinct” river systems contributed significantly to lakes in the past. Central Asian rivers like Amu Darya and Syr Darya were very prominent during the Quaternary when they flowed into the South Caspian Basin depositing a substantial delta (Kroonenberg et al., 2005; Torres, 2007).

Previous studies suggested that climate change was the dominant driver of Quaternary sea level variations in the Pontocaspian domain, but there are conflicting ideas concerning the relative importance of different potential drivers as well as the mechanisms/routes by which they impacted basin connectivity and sea/lake level. For example, different opinions exist on the importance of particular drivers of lake level change in the Caspian Basin during the last interglacial. Wegwerth et al. (2014) suggested that the two meltwater pulses occurred during the MIS6-MIS5 transition (133.5 – 122.5 ka) resulted in an initial overflow of the Black Sea into Mediterranean Sea (before marine flooding of the Bosphorus of MIS5e). These overflow events were possibly driven by increased influx from the Amu Darya catchment into the Caspian Sea. In contrast, Badertscher et al. (2011) indicated that these

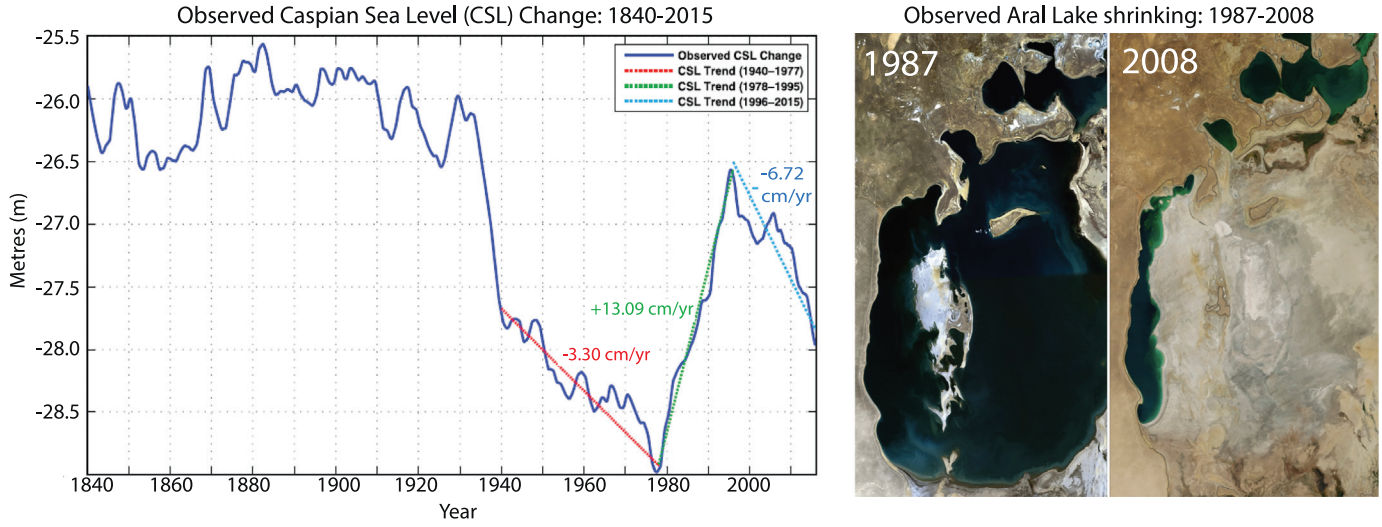

Fig. 16. Dramatic water level changes in the Pontocaspian domain. Left: Monthly mean Caspian Sea level (CSL) changes observed by tide gauges (1940–1997) and satellite altimetry (1997 to 2015) from Chen et al. (2017). Right: Recent shrinking of the Aral Lake, comparison of satallite images from 1987 (USGS) and 2008 (NASA).

particular meltwater pulses could have joined the Black Sea via the Caspian Sea from the Fennoscandian ice sheet. The main water sources of highstands during the late Pleistocene (early: ~35–25 ka BP and late Khvalynian: ~17–12 ka BP) are also debated (Yanko-Hombach and Kislov, 2017). For example, melting of ice sheets in the Russian Plain or overflow of ice-dammed pro-glacial lakes south of the Siberian ice sheets were proposed as major water sources for these Khvalynian transgressions (for details see review by Svitoch, (2009)). Tudryn et al. (2016) indicated that during the late Pleistocene both the Caspian and Black Sea basins were a trap for south-eastern Scandinavian Ice-Sheet meltwater pulses. The snow/ice melt flux joined the Pontocaspian region via the Dniepr and Volga rivers. These pulses were dated for the Black Sea between 17.2 and 15.7 cal kyr BP and the Caspian Basin from LGM until ~13.8 cal kyr BP respectively (Soulet et al., 2013; Tudryn et al., 2016).

Modelling efforts (Kislov and Toropov, 2007a, 2007b; Renssen et al., 2007; Esin et al., 2010) have also indicated how the Pontocaspian sea level was affected by changes in climate, especially during the Late Pleistocene to present day. Based on climate model outputs from PMIP (Paleoclimate modelling Inter-comparison Project) it was suggested that the primary control on lake level during the mid-Holocene (6 ka BP) and LGM (21 ka BP) were changes of river runoff rather than precipitation and evaporation over the sea (Kislov and Toropov, 2007a).

4.6. The Pontocaspian in the Anthropocene

Pontocaspian lake systems and biota have experienced rapid and profound changes throughout its geological history. Yet a variety of human activities has accelerated changes and are affecting the region and its unique biota even further. Planned large scale engineering works in the Caspian Sea and Black Sea Basin together with climate change may pose even further challenges for the Pontocaspian ecosystems.

Pontocaspian ecosystems in the Black Sea Basin are currently restricted to the coastal zones (deltas, estuaries/limans, lagoons) and adjacent lower river basins of the northwestern shelf and the Sea of Azov. The most important anthropogenic impacts in these ecosystems are (1) increased sediment load associated with forest clearing since the middle ages (Giosan et al., 2012), (2) damming of rivers modifying flow regimes and restricting mobility of faunas (Semenchenko et al., 2015) effectively providing sharp upper boundaries of Pontocaspian habitats (that normally grade into fluvial habitats), (3) strong nutrient increase due to waste water and use of agricultural fertilizers and pollution (e.g. (Kosarev and Yablonskaya, 1994), (4) conversion of Pontocaspian wetlands into pasture (Popa et al., 2009), (5) shipping lane dredging affecting major salinity gradients in the Pontocaspian habitats (Zhulidov et al., 2018) and (6) the introduction of alien species strongly affecting the native biota (e.g. Grigorovich et al., 2002; Kosarev and Yablonskaya, 1994). The current conservation status of Pontocaspian biota in the Black Sea Basin is not well understood. Strong reductions in distribution areas (Popa et al., 2009) and reduced abundances of Pontocaspian species (Abdurakhmanov et al., 2002; Yanina, 2015) have been reported. For several species no recent records of living populations have been obtained suggesting strong decline.

The oldest human imprints in the Caspian Basin have been proposed as the introduction of the brackish water cockle *Cerastoderma glaucum* with Paleolithic settlements across the Black Sea – Caspian Sea drainage divide c 8000 yr ago (Fedorov, 1957; Yanina, 2012b, 2015). Coastal areas and river basins were impacted by human activities in similar ways as the Black Sea Pontocaspian habitats (Kosarev and Yablonskaya, 1994; Grigorovich et al., 2002). Invasive species established in the Caspian Basin through accidental introduction with ships and a range of exotic species were deliberately introduced in the 20th century to increase food production for local fish. Invasions are ongoing: the latest invasive species is *Mytilopsis leucophaeta*, a mussel native to the Caribbean, that in the past few years has invaded many of the ports (Heiler et al., 2010). The invasions have resulted in very strong decline of Pontocaspian communities within decennia (examples in Kosarev and Yablonskaya, 1994). Several formerly widespread species, such as *Dreissena caspia* and *D. elata* appear to have gone extinct and for many more, including hydrobiid species groups, no recent records exist of living populations. The Caspian biodiversity crisis has been described as ecocide by Dumont (1995). However, the true scale of the biodiversity crisis is uncertain as there is a paucity of recent data. Rapid sea level changes in the Caspian Sea (up to a few metres/decade during the later 20th century) apparently are natural variation (Fig. 16).

The situation in Lake Aral is even more dramatic (Fig. 16): the lake itself is a precarious balance of inflowing Central Asian rivers and evaporation and in the Holocene Lake Aral experienced draughts (Filippov and Riedel, 2009). Yet the almost entire disappearance of the lake between 1957 and 2015 has been driven by extraction of river waters for agriculture. As a result, most of the Pontocaspian species in Lake Aral have vanished (Andreeva, 1989; Boomer et al., 2000).

In general, the continuity of many Pontocaspian lineages, especially those of Paratethyan origin, is in peril. The center of Pontocaspian diversity is located in the Caspian Sea and the demise of the native faunas is of major concern. Several planned and unplanned activities pose further threats to Pontocaspian ecosystems. Higher global sea levels associated with projected global warming will affect coastal Pontocaspian habitats of the Black Sea. The proposed Istanbul shipping lane may have impacts on the marine properties of the Black Sea Basin and provide another gateway for alien species. Climate change also may strongly impact the precipitation/evaporation balances of the Caspian and Aral catchments. These precarious balances drove lake level variations in the past. Plans for massive water extraction from the Caspian Sea to alleviate water shortages on the Iranian plateau may tip the balance.

The Pontocaspian region has a very long and dynamic history of lake basin evolution and fauna evolution resulting in a unique fauna adapted to the unusual salinity settings and partial isolation of the basins. Human activities are driving unprecedented change in many parts of the region, changing the nature of this system profoundly and permanently.

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