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Late Miocene to Pliocene palaeogeography of the Paratethys and its relation to the Mediterranean

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Abstract

The palaeogeographic evolution of the Paratethyan and Mediterranean realms are reconstructed with three maps ranging from the Late Miocene to the Middle–Late Pliocene. The maps are based on the facial analysis of selected deposits, clastic input, slope-slide process analyses as well as biogeographic data of the planktic, benthic and terrestrial biota. Characteristic fossil assemblages are used for palaeobathymetric and palaeohydrologic interpretations, restoration of palaeogeographic connections. The palaeogeographic reconstructions are palinspastically restored (after [Dercourt, J., Ricou, L.-E., Vrielynck, B. (Eds.), 1993. Atlas Tethys Palaeoenvironmental Maps. Gauthier-Villars, Paris, pp. 1–307, 14 maps; The Paleogeographic Atlas of Northern Eurasia, 1997. Inst. Tectonics Lithospheric Plates. Moscow. 26 maps], with modifications). Maps have been prepared for the terminal Tortonian/early Messinian–Late Pannonian/early Maeotian, Late Messinian–Pontian (salinity crisis time) and Piacenzian/Gelasian–Akchagilian. They illustrate the Neogene palaeogeographic evolution during and after the Attic orogenesis. Though the emerging mountain system of the Alpine foldbelt increasingly separated the Paratethys from the Mediterranean, Tethys–Paratethys connections remained extant and sufficiently effective for limited communication between both basins. They governed many features of the cyclic depositional history and biogeographic evolution of the Eastern Paratethys.

Keywords: Neogene; Paratethys; Pannonian; Euxinian–Caspian basins; Palaeogeography

1. Introduction

The Late Miocene–Early Pliocene orogenic phases significantly re-shaped the palaeogeography of areas adjacent to the Alpine foldbelt. A large-scale tectonic and sedimentary-palaeogeographic turnover affected the Late Neogene Mediterranean and Paratethyan realms particularly during the Late Tortonian (about 8 Ma ago; Meulenkamp et al., 2000) as a consequence of the Attic orogenic phase. Lateral tectonic movements in the western Mediterranean, uplifting in the Betic and Rifean regions, opening of the Tyrrhenian and Dead Sea basins, inception of stable marine deposition in the Aegean domain, and southward directed depocenter migration in the fore-Apennine basin determined the main features of the Messinian palaeogeographic, palaeobiogeographic

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and sedimentological history of the circum-Mediterranean realm.

Major geological features along the southern margin of the European platform include the end of sedimentation in the fore-Alpine and fore-Carpathian molasse basins and further palaeogeographic differentiation between the domains of the Central and Eastern Paratethys. The intra-Carpathian (Pannonian) basin became closed and modified into a long-lived brackish lake-sea ("Lake Pannon"), which received pronouncedly prograding deltaic clastics from the uplifted Carpathians. Major deformation in the Eastern Paratethys area and its surrounding mountain ranges (Pontides, Lesser Caucasus and Taurus) occurred in the Early Pliocene (Late Kimmerian, see Fig. 1), following earlier compressional block movements during the Maeotian and Early Pontian. As a result of these orogenic developments, the Black Sea depression obtained its modern outline (Tugolesov et al., 1985; Shcherba, 1993). Deposition in the Eastern Paratethys was affected by uplift of the Greater Caucasus, faulting in the Lesser Caucasus area, and a general regression of the marine realm. Marine deposition was terminated in large parts of Transcaucasia, on the Turanian Plate, and in the fore-Kopet Dagh Depression. During the Maeotian, the Eastern Paratethys gradually evolved into an intracontinental lake-sea with brackish water. The Dacic Basin, which had formerly been part of the Central Paratethys, from the Sarmatian belonged to the Eastern Paratethyan biogeographic domain. However, the record of the polyhaline plankton assemblages testifies passageways between the Paratethyan and Mediterranean marine realms, which were episodically opener than previously proposed.

1.1. Method of study

The region under consideration constituted a principal part of the Alpine foldbelt that was tectonically active during the entire Neogene. During the Late Neogene, the most significant lateral foldbelt movements, and associated block rotations, occurred in the western Apenninic zone of the Mediterranean (Boccaletti et al., 1990). For our palaeogeographic reconstructions, we used the restored palinspastic geographic base of Dercourt et al. (1993; with modifications) and "The Paleogeographic Atlas …" (1997), which take into account plate tectonic and palaeomagnetic data. The resultant palaeogeographic hypotheses concerning the reconstruction of straits, continental bridges, and water circulation patterns were tested by means of biogeographic data derived from planktic and benthic fossil associations and evidence of terrestrial invasions of faunas. In particular, the appearance of marine polyhaline migrants is used as an indicator for the existence of a connection with the open world ocean. However, benthos versus zoo- and phytoplankton data are often conflicting: deposits with brackish mollusks, ostracodes, and bryozoans may include polyhaline plankton associations. When interpreting these data, it is necessary to take into consideration that these benthos groups are characterized by longer life cycle and relatively heavy skeleton. Therefore autochthonous benthos remains more really reflect average environments in the burial place. Allochtonous burial, when shelly material was reworked by streams or redeposited to deeper bathymetric zones as a result of slope-slide processes, can be easily recognized from taphonomic data.

In contrast, the specific gravity of micro- and nannoplankton is more or less similar to that of seawater, so they can easily be transported by streams months after death. For this reason the presence of polyhaline planktic associations may be assumed to testify for influx of oceanic water, though it does not necessarily indicate fully marine environmental conditions at the location where the fossils were eventually deposited and buried. However, if marine microfossils occur very rarely, great caution is needed with respect to their palaeo-environmental interpretation, because wind can carry microfossils over considerable distances. (Illustratively for instance, M. Baldi-Beke (personal communication) observed marine nannoplankton remains in a dust sample collected in her garden near Budapest, more than 400 km away from the nearest sea).

1.2. Stratigraphic correlations

The stratigraphic correlation of the Pontian and Maeotian stages with those of the Mediterranean realm is presently the most troublesome issue in Paratethyan stratigraphy. Mostly, the Maeotian is correlated with the Middle–upper Tortonian, while the Pontian is assumed to be equivalent to the entire Messinian (Iljina and Nevesskaja, 1979; Neogenovaja sistema, 1986; Rögl, 1998). This correlation, however, disagrees with palaeomagnetic data. Based on primary reversed polarity of the Pontian, Trubikhin (1989) correlated it with the lower part of Chron 4 (=C3r) and with the upper Messinian. This correlation is supported by calcareous nannofossil data (NN11 in the *Dosinia* Beds of lower Maeotian and in lower Pontian and NN12 in Upper Pontian—Marunteanu and Papaianopol, 1998) from the Dacic Basin of South Romania and

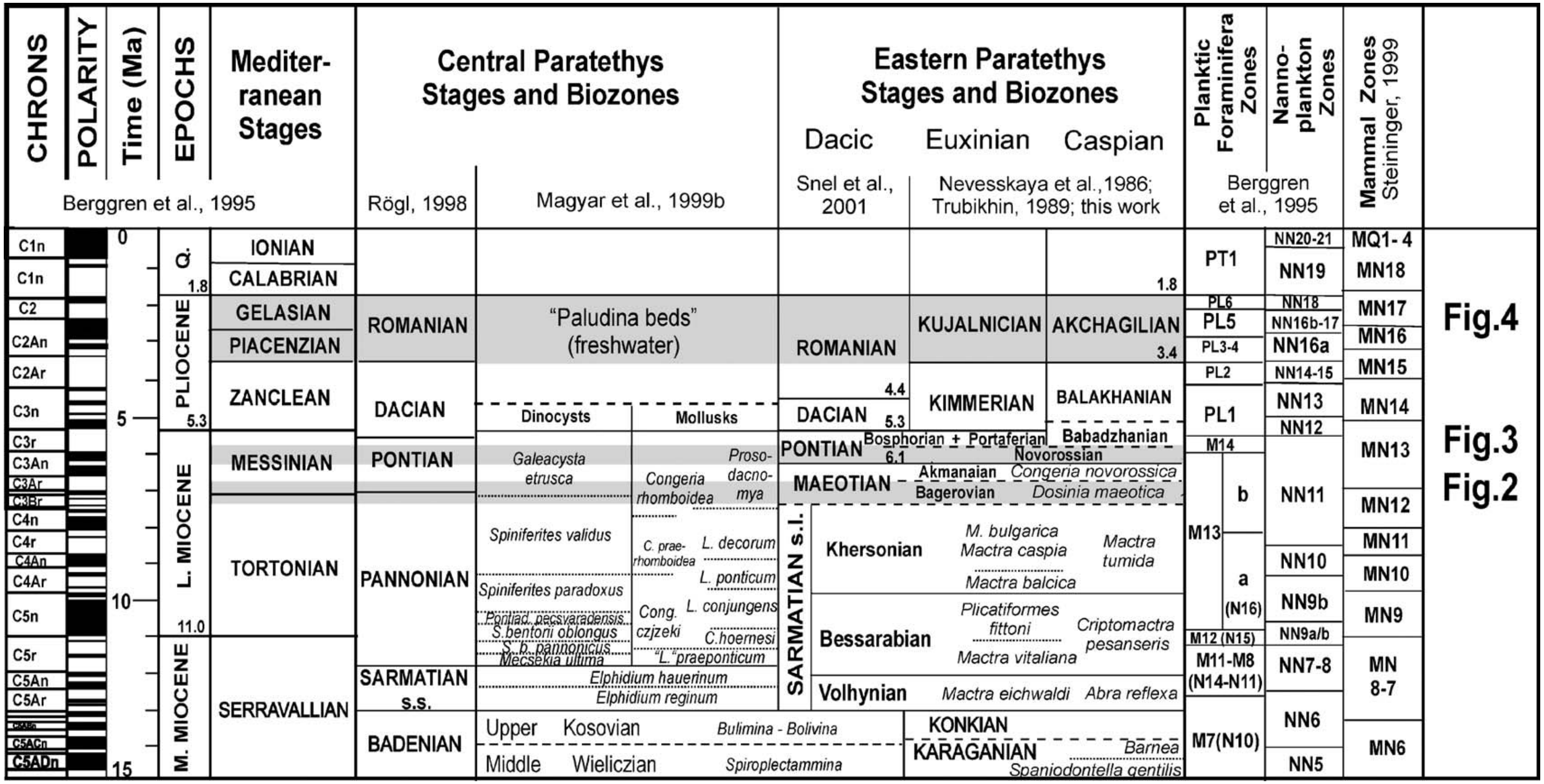

Fig. 1. Stratigraphic scheme of the Late Cenozoic Mediterranean (according to Berggren et al., 1995), Central (Rögl, 1998; Magyar et al., 1999b), and Eastern Paratethys (Dacic Basin according to Snel et al., 2001, Euxinian and Caspian basins — by Nevesskaya et al., 1986; Trubikhin, 1989). Mammal zones — according to Steininger (1999). Time intervals of the palaeogeographic maps (Figs. 2–4) indicated in right column.

correspondence with the current Astronomical Polarity Time Scale (Snel et al., 2001, Vasiliev et al., 2004, 2005) as well as biogeographic data concerning the origin of the Pontian fauna (Popov and Nevesskaya, 2000). In this case the Maeotian has to correlate with the uppermost Tortonian–Lower Messinian (Fig. 1). Though not all authors of this paper agree with this correlation (Iljina and Nevesskaja, 1979), this point of view is taken as a working hypothesis for this paper.

Due to the non-marine character of many younger Pliocene Paratethyan sequences, it is often not possible to subdivide biostratigraphically the Middle–Late Pliocene interval. As a consequence, the Pliocene map reflects a broad time interval from about 3.4 to 1.8 Ma. Altogether, it displays palaeogeographic configurations and environments that occurred after the large-scale tectonic reorganization of the late Early Pliocene. In terms of regional Paratethys stages and horizons, it corresponds to the Romanian of the Dacic Basin, the Kujalnician of the Euxinian Basin, and the Akchagilian of the Caspian Basin. These intervals can be relatively well delimited by both biostratigraphy and magnetostratigraphy.

1.3. General palaeogeography

1.3.1. Terminal Tortonian–Early Messinian: before salinity crisis (around 7 Ma)

1.3.1.1. Mediterranean. The Attic orogenesis and the Late Tortonian eustatic lowering of sea level initiated large-scale palaeogeographic and sedimentological changes in the Mediterranean. In this region these widespread events led to the shoaling and/or closing of the Betic Corridor (7.5–7.6 Ma years ago), the somewhat later shallowing of the Rifean Corridor (7.1–7.3 Ma), and palaeoceanographic modification of the Mediterranean Basin from an anti-estuarine into an estuarine circulation system. Moreover, it initiated extinction of deep-water benthic associations and accumulation of evaporites from the Late Tortonian onwards (Benson, 1976; Kouwenhoven et al., 1999; "Late Miocene …", 2001). Planktic and shallow-water benthic associations of Late Tortonian–Early Messinian age were rich and relatively diversified indicating subtropical environments of deposition with normal marine salinity (Benson, 1976). However, taxonomic composition of the most stenobiontic groups were less diversified than those found at the Atlantic coast of Iberia. The taxonomic diversity of the scleractinian coral associations was low (up to 6 species), and only the most eurybiontic genera–*Porites* and *Tarbellastraea*–were important reef builders, together with coralline algae (Esteban et al., 1996). The most probable limiting factor was fluctuation of salinity. Most diverse associations inhabited the southern and southeastern parts of the western Mediterranean, which probably indicates direction of water circulation. The northern carbonate platforms were built mainly by rhodolits.

1.3.1.2. Pannonian Basin: early Congeria rhomboidea, early Galeacysta etrusca biochrons. Lake Pannon, the large brackish lake that filled the Pannonian basin during the Late Miocene, flooded more and more dry land within the basin as the post-rift thermal subsidence of the basin floor continued. The water depth in some subbasins may have reached 1000 m. Condensed sequences of lacustrine carbonates were deposited in regions which did not receive significant clastic sediment supply. The uplift of the Alps and the Carpathians, however, provided increasing amount of clastic sediments; the rivers dumping the sediment load into the basin completely filled up the northwestern and northeastern regions, turning them into flat alluvial plains (Juhasz, 1991). The Transylvanian Basin also dried up, so Lake Pannon was restricted to the southern part of the basin. The southern shoreline, running parallel to the Sava river along the northern foots of the Dinarides, changed very little during the lifetime of Lake Pannon (Magyar et al., 1999a; Fig. 2).

The rich mollusk and ostracod fauna of Lake Pannon was almost fully endemic, and included lineages of both marine and freshwater origin (Müller et al., 1999). Adaptive radiation of cardiids, dreissenids, hydrobiids, and pulmonate gastropods was especially spectacular. Many of the diatom, dinoflagellate, nannoplankton, and fish species are also recognized as endemic. Whether the lacustrine basin experienced marine water incursions is ambiguous. Rare and sporadic findings of polyhaline nannofossils (Kollányi, 2000) and presence of supposedly marine elements in dinocyst associations (Suto, 1995) seem to suggest that marine connections were not fully severed. Altogether, the benthic fauna of mollusks and ostracodes was almost fully endemic, reflecting the persistent occurrence of a brackish lacustrine environment (Müller et al., 1999).

The mostly intermediate calc-alkaline arc-type magmatism, which was related to the Miocene to Pliocene subduction beneath the Carpathian arc, was active at this time in the Calimani and Gurghiu Mts. of the East Carpathians (Pécskay et al., 1995). The alkali basaltic volcanism of the basin proper had also started by this time. The volcanic eruptions were either subaerial (like in the Transdanubian Central Range of Hungary) or took

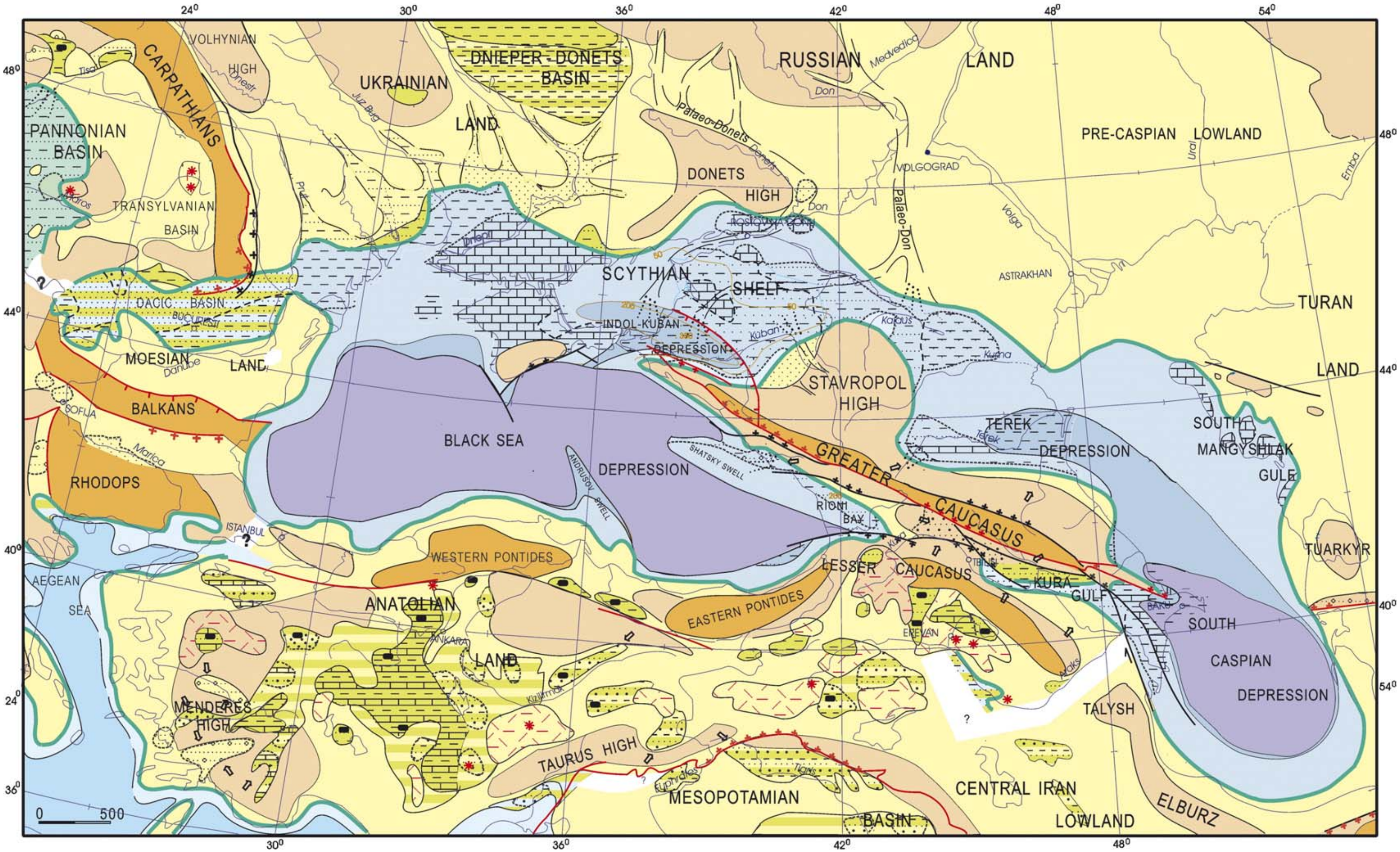

Fig. 2. Palaeogeographic map of the Eastern Paratethys for the terminal Tortonian–early Messinian (early Maeotian) on recent geographic base. Compiled by L.B. Ilyina, I.G. Shcherba and S.O. Khondkarian.

place under the water of Lake Pannon (like the "Kecel Basalt Formation" of the Hungarian Plain; Balázs and Nusszer, 1987).

1.3.1.3. Euxinian–Caspian Basin during the Maeotian. The short-term drop in sea level at end of the Sarmatian (sensu Barbot-de-Marny) followed by the Maeotian transgression has been estimated to be about 300 m in the Black Sea depression (Tugolesov et al., 1985; Robinson, 1997). Unconformities between the brackish (or hypersaline) upper Sarmatian and semi-marine lower Maeotian deposits are very common, suggesting drying up of large areas. The maximum of the transgression took place in the second part of the Early Maeotian (Stevanovich and Ilyina, 1982; Nevesskaya et al., 1986; Popov et al., 2004), when the basin system extended from the northern margin of the Moesian Plate (Dacic Basin) to southern Mangyshlak in the east (Fig. 2). However, its size was reduced relative to that during the Sarmatian.

The mid-Sarmatian orogenic movements in the central part of the basin resulted in strong northward progradation of the mountainous Caucasian Peninsula joined with the Iran-Lesser Caucasus land. The northern part of the Lower Kura depression accumulated relatively deep-water diatom oozes, whereas the peripheral part of the bay were marked by an accumulation of oolitic and detrital limestones that were replaced by terrigenous sediments in the southern direction. By the end of the Maeotian, the relict deep-water depression of the southeastern Caucasus opened toward the South Caspian Basin was compensated by sediments.

The Early Maeotian Basin was inhabited by mainly endemic species and subspecies of euryhaline Mediterranean genera, which evolved in the gulfs and lagoons of the Late Miocene sea. Some of them are known from the Aegean area (Stevanovich and Ilyina, 1982; Popov and Nevesskaya, 2000). Salinity of the early Maeotian Sea has been estimated to be generally 13–14 ppt, up to 17–18 ppt and reached maximum values (at least 25 ppt) in the second half of the early Maeotian in the Rioni Bay. This bay was invaded by true polyhaline Mediterranean mollusks *Alvania montagui* (Payr.), *Rissoa veittricosa* Desm., *Gibberula philippii* (Monterest.), *Ruditapes decussatus* (L.), and *Spiratella* sp., as well as by foraminifers, ostracodes, and nannoplankton association with *Catinaster coalitus* (Semenenko et al., 1995), which colonized the basin owing to the cyclonic current flowing along the southern coast of the Early Maeotian sea and favorable environments in this part of the sea (Ilyina et al., 1976). Ephemeral marine ingressions took place in other parts of the basin system too, as evidenced by the presence of polyhaline nannoplankton associations in lower Maeotian horizons in the Kerch–Taman area (Lyul'eva in Semenenko, 1987) and in *Dosinia* beds of Dacic Basin (Marunteanu and Papaianopol, 1998). The lowest salinity was characteristic of the western border of the basin. In the Serbian part of Dacian Gulf the marine regime was established only at the end of the Early Maeotian and was marked by the appearance and distribution of *Dosinia* and *Ervilia* (Stevanovich and Ilyina, 1982).

Shallow-water environments dominated the Eastern Paratethys. Deep water ones (more than 1000 m) existed only in the Black Sea and South Caspian depressions. The Indol-Kubanian and Terek-Caspian foredeeps were most deep shelfal depressions of the basin system. In much of the Eurasian shelf shallow limestones, clays, and sands accumulated. Oolitic and detrital-shelly limestones developed in the southern Ukraine, Crimea, Azov area and in South Mangyshlak. Alluvial-deltaic deposition prevailed around growing orogenic structures of the Greater Caucasus and along the northern margin of the Euxinian sea (Shcherba et al., 2001).

Thick sequences of molasse deposits accumulated in the continental-coastal areas of Transcaucasia and in the Kura and Rioni depressions (Shcherba, 1993; Shcherba et al., 2001). Conglomerates, clays, and sands of Dushet Formation (Maeotian–Pontian, up to 2000 m) were deposited in fluvial environments, containing remains of terrestrial and freshwater mollusks, mammals, and plants. The rivers that transported this material discharged into a lake in the mid-Kura depression, where low-carbonate sandy clays (up to 2500 m) of the Maeotian–Pontian Shirak Formation were deposited. In the South Azerbaijan–North Iran territory marine deposits with Early Maeotian foraminifers (*Quinqueloculina seminulum maeotica*) are known from the Nakhichevan depression (Neogenovaja sistema, 1986). In this Middle Araks–Nakhichevan area occurred a gulf of the Paratethys with a freshwater fauna including characteristic Eastern Paratethyan endemics since the Karaganian (Middle part of the Middle Miocene).

By the end of the Early Maeotian, the reactivation of orogenic movements resulted in general coarsening of sediments and angular unconformity at the Maeotian–Pontian boundary. As a consequence, the connection of the Paratethys with the Mediterranean gradually weakened, and its hydrological regime changed. The practically closed and freshened Late Maeotian basin widened. This process was accompanied by the flooding of river valleys in its northwestern margin and intense abrasion of coasts in the Rioni and South Mangyshlak bays. The latter bay reached the western Ustyurt area and its southern shoreline, which is locally marked by pebble beds, bounded the Karabogaz arch.

During the Late Maeotian, semi-marine fauna was replaced by brackish mollusk associations with *Congeria*, *Theodoxus*, *Pseudamnicola* and *Turricaspia*, brackish and freshwater diatoms and dinocysts. The salinity of the Late Maeotian sea probably lowered and its ionic composition also changed to Caspian-type one. However, Upper Maeotian sediments of the Rioni Bay (Galidzga River), Kerch and Taman area as well as the Dacic Basin (Marunteanu and Papaianopol, 1998) enclose elements of the Mediterranean fauna (beds with *Mactra superstes*, *Cerastoderma*, *Sphaeronassa mutabilis andrussovi*, *Bittium*, *Mohrensternia*—Ilyina et al., 1976; Popov's collection; intercalations with polyhaline nannoplankton—Lyul'eva in Semenenko, 1987; and marine diatoms with *Coscinodiscus*, *Thalasiosira decipiens*, *Nitzschia praereinholdi*—E.P. Radionova, unpublished data).

Connections with the Mediterranean probably occurred in the Aegean region, where lagoonal mollusk associations of the Dafni Formation (northern Greece, Serres) are very similar to those of the lower Maeotian (Stevanovich and Ilyina, 1982; Popov and Nevesskaya, 2000), and possibly in Transcaucasia or/and in the Taurus–Zagros area (Chepalyga, 1995), where the Upper Miocene–Pliocene sequences are poorly investigated.

The East European platform represented a lowland with the Mid Russian High in its center, and the Volhynian, Ukrainian and Donets highs in its southern parts. The Turan plate corresponded to a lowland largely without deposition. Orogenic movements in the Kopet Dagh and Pamir-Tien Shan zone led to the appearance of a wide molasse belt along their foothills that was composed of coarsely-grained, red-colored sequences of clastics.

1.3.2. Late Messinian: salinity crisis (6.1–5.3 Ma)

1.3.2.1. Mediterranean. The main palaeogeographic and bathymetric features of the Late Messinian basin system were similar to those of the Early Messinian. The salinity crisis markedly changed the sedimentological processes and facies of the basin. The pre-existing bathymetric zones became more distinct. Evaporites of different composition were deposited in different depth zones.

We adopted for this study the double-phase model of the Messinian crisis as developed by Clauzon et al. (2001). According to this hypothesis, the Mediterranean sea level was lowered in two phases separated by a flooding event. The first phase (5.8 Ma) affected only the Mediterranean basin margins and had a moderate amplitude (less than 100 m). It led to the accumulation of the lower gypsiferous evaporites. The second phase corresponded to a drastic drop in sea level that exceeded 1500 m and affected the whole basin (5.6 Ma). As a consequence, deep canyons were incised along the basin margins (palaeo-Nile, palaeo-Rhone, palaeo-Ebro, etc.) while a thick evaporitic series (up to 2 km according to seismic data) accumulated in the bathyal plains. Central parts of the basin deepest depressions are typified by the occurrence of soluble halite chloride salts (Chumakov in Malovitskiy et al., 1982). Between the two evaporitic phases brackish deposition took place in the "Lago Mare" facies. This member (so-called Terminal Complex) includes many stromatolithes, as well as lagoonal and shallow water deposits with brackish fauna of the Pontian type and continental intervals.

We tried to show this complex history on a single map (Fig. 3): bathyal environments with halite deposition in the Albouran, western and eastern Mediterranean depressions, surrounding inner shelf environments with marginal gypsum deposition that change into dominantly brackish environments in the Tyrrhenian, Adriatic, and Aegean regions prevailing marine facies with *Porites* reef-bodies in shallower coastal zones where the evaporites and "Lago Mare" facies are absent as these parts of the basin emerged during regression. In addition, some shallow depressions existed which were probably isolated from the open sea during the salinity crisis, and where marine and brackish sedimentation occurred without formation of evaporites (in Iberia, Rhone basin, Aegean area).

The brackish fauna and microflora of the "Lago Mare" included numerous genera and species in common with the Paratethys: endemic limnocardiines, *Congeria* and *Melanopsis* among the mollusks, *Loxoconcha djaffarovi* and *Cyprideis pannonica* ostracod associations, and *G. etrusca* among the dinocysts. Moreover, the "Lago Mare" brackish fauna contained Late Messinian endemics and was enriched by euryhaline marine faunal elements, which were absent in the Paratethys (*Cerastoderma*, *Mactra* among the mollusks). Probably, its origin is related to the oldest Late Miocene–Pannonian Paratethyan biota, but their subsequent development occurred independently. Some components of this specific biota, together with some rare Pannonian species, were ancestral to the younger Pontian fauna of the Euxinian–Caspian Paratethys. They migrated from the Mediterranean to the Eastern Paratethys through intermediate basins (??Aegean and

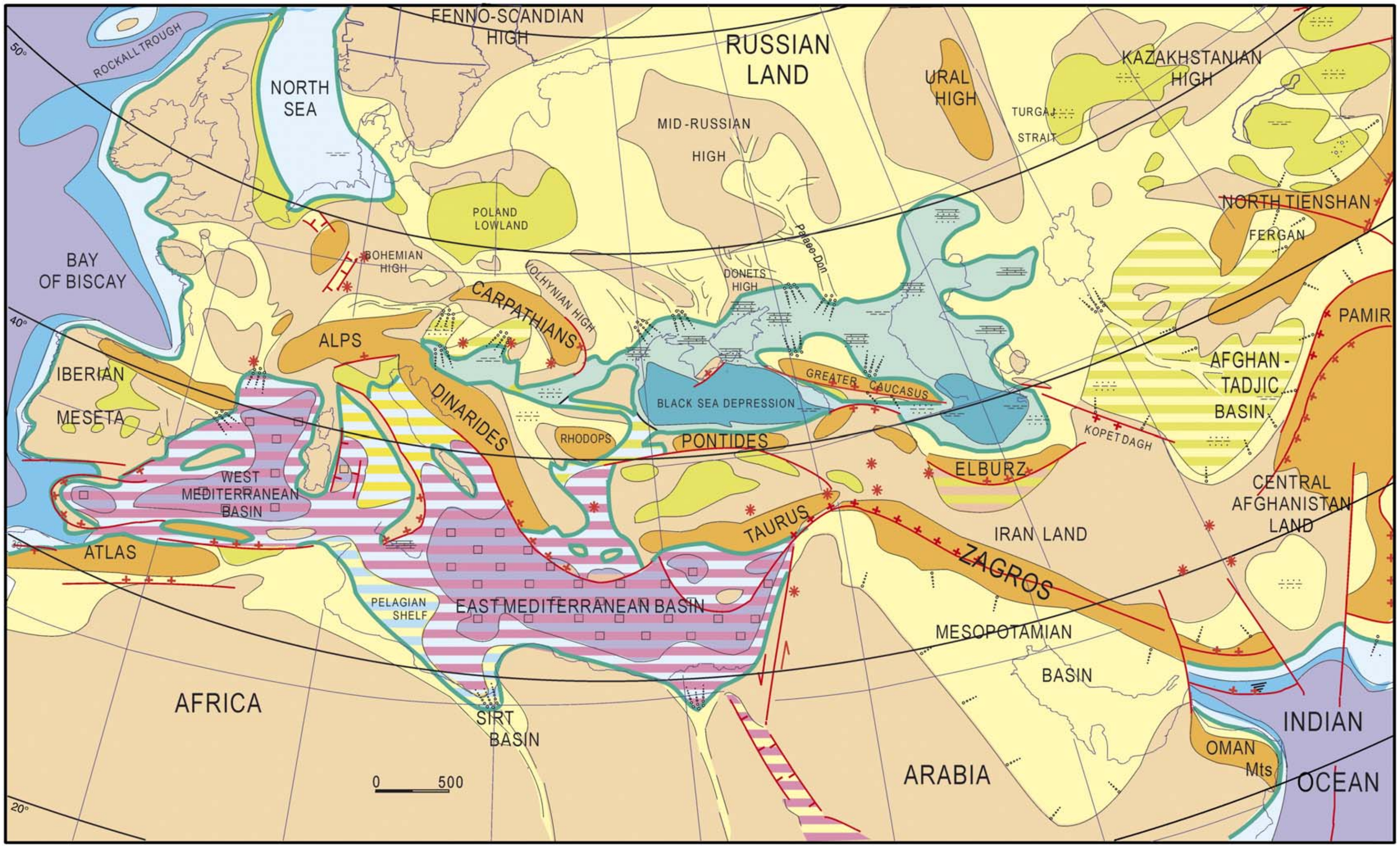

Fig. 3. Palinspastic reconstruction for the late Miocene, Late Messinian, salinity crisis (late Pannonian, early Pontian) (Popov et al., 2004).

Dacic basins; Eberzin, 1949; Nevesskaya et al., 1986; Popov and Nevesskaya, 2000).

1.3.2.2. Pannonian Basin: late C. rhomboidea, late G. etrusca biochrons. By this time, delta progradation from the northwest almost completely filled up the western part of the basin. Only the Drava depression, Slavonia, and the Sava depression remained subaqueous in the western Pannonian basin. As subsidence continued, the accommodation space in the dry land was filled with fluvial, terrestrial deposits. Delta lobes in the central part of the basin approached the lower part of the Tisza river from the northwest. The eastern part of the lacustrine basin received progradation from the northeast. The deep subbasins became gradually shallower as turbidites were deposited in the central parts of the depressions. The deep lacustrine environment was restricted to what is now southeast Hungary, northeast Croatia and north Serbia. The endemic mollusk and ostracod faunas of Lake Pannon flourished, and migrated probably episodically into the Eastern Paratethys during the Pontian via an outflow of the lake (Müller et al., 1999).

The calc-alkaline volcanism also lost area in the East Carpathians; now it was restricted to the North Harghita Mts. (Pécskay et al., 1995).

1.3.2.3. Euxinian–Caspian Basin during the Pontian. The Eastern Paratethys as shown in Fig. 3 corresponds to the maximum spreading of the sea in the Early Pontian (Novorossian). The Early Pontian basin was strongly enlarged, especially by transgression along its northern and eastern margins (Popov et al., 2004). In the north, the Early Pontian sea covered the southern half of the Donets High, which had been dry land and a source of clastic material since the Palaeogene, and reached the Volga-Don and Ergeni areas. In the northeast, the Early Pontian sea formed the large North-Pre-Caspian, North-Ustjurtian and South Mangyshlakian gulfs. Lower Pontian sediments are known from the Turkish coast of the Black Sea (Oszayar, 1977). The Dacic Basin belonged biogeographically to the Eastern Paratethys.

Sandy deposition prevailed in shallow coastal areas. Muds accumulated in deeper-water environments. Calcareous, shelly-detrital facies were widely distributed on the northern inner shelf of the Euxinian–Caspian Basin. Towards the Caucasus, Kopet Dagh and the western clastic sources the facies changed to muddy and sandy deposits. Deep-water environments existed only in the Black Sea and South Caspian depressions.

Based on the prevailing brackish fauna, the salinity of the basin was low, but it did not fall under 5–8 ppt. The Dacian part, where *Parvivenus widhalmi*–the single bivalve species of marine origin in Eastern Paratethyan Pontian–is lacking the salinity was probably lowest. At the same time the presence of nannoplankton (Marunteanu and Papaianopol, 1998) in the middle (NN11) and upper (NN12) Pontian testifies for episodic ingression of marine water. Marine communication between the Paratethys and the Mediterranean probably occurred via the Aegean Gulf (Stevanovic et al., 1989; Popov and Nevesskaya, 2000).

A pronounced regression at the beginning of the Late Pontian (Portaferrian) led to the emergence of the northern outer shelf of the Euxinian Basin. The Ciscaucasian Strait was closed, separating the Caspian Basin from the Euxinian Basin. The eastern lake-sea became restricted to the recent Middle and South Caspian depressions, including the Kura Gulf. This fall in sea level approximately corresponded in time with a drastic sea level drop in the Mediterranean (5.7 Ma according to initial Trubikhin in Stevanovic et al. (1989) and 5.6 Ma after Meulenkamp and collaborators (Snel et al., 2001), respectively). According to biogeographic data, however, the Caspian–Euxinian connection existed during the entire Pontian: particular Late Pontian species common in the Euxinian and Dacian basins are known from the Upper Pontian of Azerbaijan (Nevesskaya et al., 1986). New sharp regression took place at end of the Pontian when all recent shelf of the Black and Azov seas was emerged (Semenenko, 1987).

1.3.3. Middle–Late Pliocene: Piacenzian + Gelasian–Kujalnician–Akchagilian (3.5–1.8 Ma)

1.3.3.1. Mediterranean. Major features of the circum-Mediterranean palaeogeography after the Early Pliocene orogenesis were increasing rates of subsidence of depressions (Tyrrhenian and southern Aegean back-arc basins), and uplift of the surrounding mountain chains (Fig. 4). Pliocene vertical movements initiated the development of the present-day palaeogeographic features and river system (Meulenkamp et al., 2000; Popov et al., 2004).

In the western Mediterranean domain, the Strait of Gibraltar originated at the beginning of the Pliocene, after the Betic and Rifean corridors were closed in respectively the Tortonian and Messinian. Shallow marine environments persisted throughout the Pliocene all along the northern coast of the African continent. All Iberian domains were emerged in the Pliocene, except for some marginal basins. Continental facies (shallow-lake carbonates and fluvio-lacustrine clastics) accumulated in relatively small, intramontane basins.

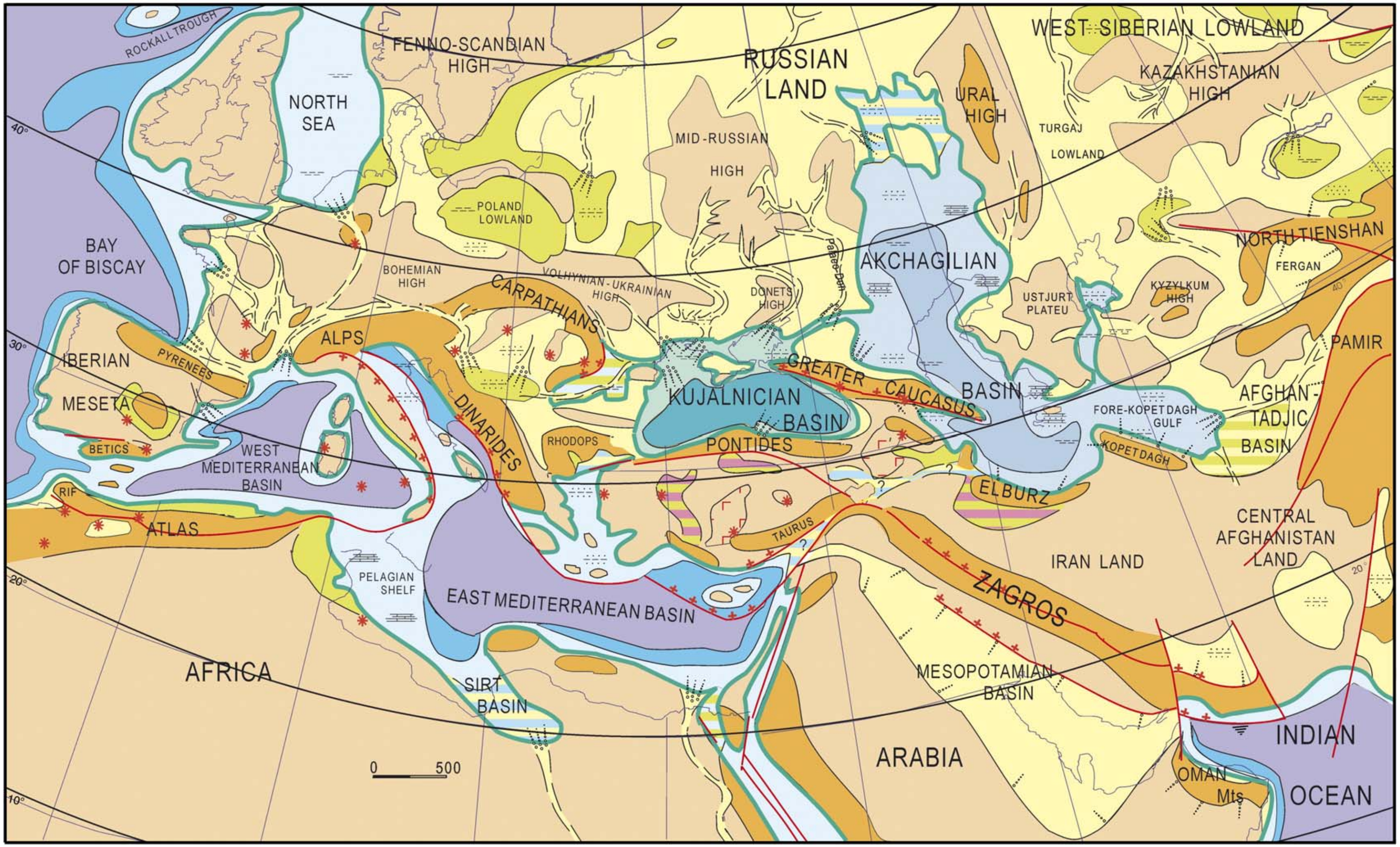

Fig. 4. Palinspastic reconstruction for the middle–late Pliocene, Piacenzian–Gelasian (“Paludina Beds”, late Romanian, Kujalnician, Akchagilian) (Popov et al., 2004).

1.3.3.2. Carpatho-Pannonian region. During the Late Pliocene, major folding and thrusting in the Outer Carpathians came to an end. The intermediate calc-alkaline volcanic activity along the intra-Carpathian arc was restricted to the Southern Harghita Mts. (Pécskay et al., 1995). Pliocene alkali-basaltic volcanism played an important role in the intra-arc domains. It occurred in the Styrian and Danube basins, the Great Hungarian Lowland and the Transylvania and Slovakian-Hungarian volcanic domains (Szabo et al., 1992; Martin and Nemeth, 2004). The general Late Miocene thermal subsidence of the Pannonian basin was replaced by a late-stage tectonic inversion (Horváth, 1995; Horváth and Tari, 1999). Uplift and subaerial erosion of the Neogene strata started in the western and northern parts of the basin, while some subbasins underwent continued subsidence. In these subbasins accumulation of continental clastics took place during the Pliocene. Pliocene deposits in the Drava and Sava depressions reached thicknesses up to about 1000 m (Meulenkamp et al., 1996).

Brackish and fluvio-lacustrine foredeep sedimentation continued in the Dacic Basin (Fig. 4), where the Pliocene (Dacian–Romanian) sequences reached thickness of thousands of meters. The mapped interval corresponds here to the Romanian, when extinction of limnocardiines and the appearance of unionids, viviparids and melanopsids portray the further decrease of the salinity of the basin. Ephemeral brackish-marine ingressions from the adjacent Euxinian domain reached the eastern and central parts of the overall fluvio-lacustrine Dacic basin. Brackish-water mollusks (*Euxinicardium*) and zonal calcareous nannoplankton associations (NN15–NN16—Marunteanu and Papaianopol, 1998) are known from these layers.

1.3.3.3. Euxinian Basin during the Kujalnician. Since Latest Miocene (Late Pontian) time, the Eastern Paratethys was subdivided into two major basins, the Dacian–Euxinian basin system and the Caspian basin. The Pliocene faunas of the former were inherited from the Pontian. Areally, the brackish basin corresponded approximately to the recent Black Sea, including the south Ukrainian shelf and the Azov-Kuban and Rionic gulfs (Fig. 4). The Kujalnician inland sea had no direct connection with the ocean, but there was an ephemeral corridor connecting it with the Akchagilian basin of the Caspian region: layers with *Aktschagilia subcaspia* and *Cerastoderma dombra* are known from the Kujalnician sequences of the Azov area. Shallow-water, terrigenous clastics with Kujalnician faunas are known from the North Black Sea Depression, eastern Crimea, Kerch and Taman peninsulas, Kuban and Rioni depressions. In all these regions, the Kujalnician deposits are characterized by similar fossils of brackish ostracodes and mollusks, such as *Prosodacna*, *Pachydacna*, *Chartoconcha*, *Pontalmyra*, *Oraphocardium*, *Euxinicardium*, and *Viviparus*.

The major part of the western East European platform, occupied by the Volhyn-Ukrainian and mid-Russian highlands, was intersected by branched river systems. Seismic and sedimentary data indicate the intensive growth of prodeltas at the mouths of platform rivers (palaeo-Danube, palaeo-Prut, palaeo-Dnester, palaeo-Dnieper in the northwestern part of the Black Sea, and the palaeo-Donets in the Azov-Kubanian regions), as well as uplifting of the Greater Caucasus (palaeo-Kuban, rivers of the western Ciscaucasia and Rioni Gulf regions), and Pontides.

1.3.3.4. Caspian Basin during the Akchagilian. During the Early Pliocene, the Caspian basin experienced a major regression, which was accompanied by a reduction of salinity. The pre-Akchagilian Balakhanian freshwater basin was restricted to the South Caspian Depression and the Kura Gulf. The sea level lowering resulted in deeply incised valleys (up to a few hundred meters) of the palaeo-Volga, palaeo-Kura, and palaeo-Amudarja rivers. The resultant rugged relief was invaded by the Middle–Late Pliocene Akchagilian sea, which was characterized by hemi-marine regime with reduced salinities and euryhaline biota of marine origin. Numerous endemic taxa of mollusks, ostracodes, and diatoms evolved from lagoonal Mediterranean ancestors. Relatively deep-water, clayey deposits (up to 350–500 m) accumulated in the entire central part of the basin extending from the pre-Caspian area to the South Caspian Depression. Similar deep-water muds (50–100 m) were deposited in the central part of the Lower Kura Depression. In the relatively shallow-water environments of the southern part of the basin, i.e. in eastern Georgia, western Azerbaijan, and in the fore-Kopet Dagh Gulf, saline clastic and calcareous sediments with rich endemic faunas (*Avicardium*, *Miricardium*, *Andrussella*, *Avimactra*—Paramonova, 1994) were deposited in mid-Akchagilian time. Marine terrigenous-clastic and evaporitic sediments were deposited contemporaneously in the southern Aral Gulf, which probably connected with the Kopet Dagh Gulf.

Northwards, the Akchagilian sea ingressed eventually the Volga and Kama basins. Poor and uniform, euryhaline faunas in these parts of the basin system reflect reduced salinities as compared to those of the main basin (Neogenovaja sistema, 1986). During the mid-Akchagilian, marine ingressions reached as far as the palaeo-Murgab, the Tedzhen and Amudarja valleys. The slopes of the Ural and Mugodzary included incised river valleys.

A new phase of orogenic movements in the Greater Caucasus resulted in a pronounced increase of clastic supply from the evolving chain towards its bordering plains. Thick successions of coarse clastics were laid down in the Kura intramontane depression (up to 900 m of continental conglomerates, sands and aleurites of the Alazan Series). To the east, these successions pass into shallow-marine sediments. Thick sandy sediments of the palaeo-Terek delta filled up the western part of the Terek-Caspian depression (Kunin et al., 1990). The coarsest molasse-type clastics developed around the Eastern Kopet Dagh, Pamir and Tien Shan orogenic belts (Polisak Suite of the Tadjic and Fergana depressions). The northern Karakum and central Kyzylkum areas were a highland where thick eolian successions were deposited. The palaeo-Zeravshan and palaeo-Sarysu rivers transported the clastics, which eventually accumulated in the lacustrine Aryskum Depression and along the northern margin of the Karakum. Sandy to clayey lacustrine sediments were also deposited in the Teniz Depression. Coarse alluvial material accumulated along the mountain chains. Farther to the east, fluvio-lacustrine, silty and clayey sediments were deposited in the Turgaj area and in western Siberia (Neogenovaja sistema, 1986).

Strong andezitic-basaltic volcanism took place in the orogenic belt along the southern border of the Paratethys (in Anatolia, Adzharo-Trialeti, Lesser Caucasus, Elburz) as well as in the Greater Caucasus, where eruptions in the Kazbek area of the Central Caucasus produced abundant volcaniclastics (Ruchs-Dzuar Suite in the palaeo-Terek delta—Neogenovaja sistema, 1986).

2. Discussion

Reconstruction of the Late Neogene history of the Paratethys, including its connections to the Mediterranean, still contains a lot of problematic issues. Divergent views on specific problems occur even among the authors of this paper. These debatable issues include (1) stratigraphic and geochronological correlation between individual basins; (2) origin of endemic faunas that seem to appear suddenly in a basin; (3) nature and location of some interbasinal palaeogeographic connections; and (4) the effect of the Messinian salinity crisis on the palaeogeography of the Paratethys.

(1) Correlation between the Late Miocene Eastern Paratethyan (Maeotian and Pontian regional stages), Central Paratethyan (Pannonian regional stage), and standard Mediterranean (Tortonian and Messinian stages) stratigraphic scales is the most sharply disputable issue in the Late Neogene Paratethyan stratigraphy. The most usual correlation proposed correspondence of the Maeotian with the Middle–Upper Tortonian, and Pontian with the entire Messinian (Iljina and Nevesskaja, 1979; Neogenovaja sistema, 1986; Rögl, 1998). This correlation was supported by absolute age data (Chumakov, 1993): 9.3 Ma for the lower boundary of the Maeotian, 8.0–8.4 Ma for the Lower/Upper Maeotian boundary, 7.1 Ma for the Maeotian–Pontian boundary, and 5.2–5.3 Ma for upper boundary of the Pontian. Zonal nannoplankton determinations of Lyul'eva (NN9–NN10 in base of the Maeotian—Semenenko, 1987) do not contradict to this correlation. This long duration of the Pontian, however, conflicts with the palaeomagnetic data. Based on the primarily reversed polarity of the Pontian, it is possible to correlate it with Chron 6 of the Harland scale and with the Upper Tortonian–Lower Messinian, as proposed by Pevzner, or alternatively with the lower part of Chron 4 (=C3r) and thus with the Upper Messinian (Trubikhin, 1989; Vasiliev et al., 2004, 2005). According to recent calcareous nannofossil data of Marunteanu (NN11 in the *Dosinia* Beds of the Lower Maeotian and in the Lower Pontian and NN12 in the Upper Pontian—Marunteanu and Papaianopol, 1998) from the Dacic Basin of South Romania, and correlation of magnetostratigraphic results with the current Astronomical Polarity Time Scale (Snel et al., 2001; Vasiliev et al., 2004, 2005), as well as biogeographic data on the origin of the Pontian fauna (Popov and Nevesskaya, 2000), the Pontian corresponds to the Upper Messinian only, as it was proposed by Stevanovic et al. (1989, p. 351), Esu and Taviani (1989), Trubikhin (1989), and others. In this case the Maeotian has to correlate with the uppermost Tortonian–Lower Messinian (Fig. 1).

As to the Pannonian basin, the confusion caused by the introduction of an Eastern Paratethyan stage, the Pontian, has been recently reviewed by Sacchi et al. (1997, 1999). Here we use the Upper Miocene biozones (Magyar et al., 1999b) rather than the regional stages (see Fig. 1).

(2) One of the most challenging migration event is the sudden appearance of typical brackish endemics at the beginning of the Pontian in the Eastern Paratethys. A few species of highly specific cardiid bivalves (*Pseudocatillus pseudocatillus*, *Paradacna abichi*) are common between the Pannonian basin and the Eastern Paratethys. However, six cardiid genera (*Euxinicardium*, *Pontalmyra*, *Pseudocatillus*, *Paradacna*, *Eupatorina*

and *Prosodacnomya*), 3–4 species (*P. pseudocatillus*, *Eupatorina littoralis*, *P. abichi* and, probably, *Prosodacnomya rostrata*) and a non-cardiid, *P. widhalmi*, are common between the Eastern Paratethys and the Upper Messinian of the Aegean basin, Greece (Popov and Nevesskaya, 2000). It seems obvious that the origin of some brackish endemic genera and species among bivalves, ostracodes, and dinoflagellates goes back to Lake Pannon where they first appeared and evolved; but the further direction of migrations is debatable. If they moved from the Aegean to the Eastern Paratethys, as supposed by Eberzin (1949), Stevanovic et al. (1989), Nevesskaya (in Nevesskaya et al., 1986, 2001), there must have been an earlier event when these specific biota went from Lake Pannon to the Aegean basin. An alternative scenario would be direct migration from Lake Pannon to the Eastern Paratethys (Dacian and/or Euxinian basins), and then to the Aegean. The first variation is strongly doubted by Lake Pannon workers, whereas the latter one is rejected by Eastern Paratethyan researchers. To compound the problems, preservation of the "Lago-Mare" fossils is not well enough and a lot of species endemic to the Mediterranean (or Aegean) are known from this time interval.

(3a) Connections between the Pannonian and Dacian basins during the Pontian have long been considered to facilitate mutual exchange of the aquatic fauna. Now it seems, however, that this connection worked as a biogeographic filter which allowed primarily a one-way traffic out of Lake Pannon towards the Eastern Paratethys (Müller et al., 1999).

(3b) Based on biogeographical data (full sets of the Early Maeotian mollusk, foraminifer, ostracod, and diatom species and Late Maeotian–Pontian mollusks and ostracodes in the Azerbaijan and South Mangyshlak associations), we proposed that at least temporal communications existed between the Euxinian and Caspian basins in the Transcaucasia before the late Pontian (Nevesskaya et al., 1986). The connection could have been located between the southern border of the Dzirul Massif and the Adzharo-Trialeti fault system, which occupied its recent position in the Late Kimmerian only.

(3c) The semi-marine Akchagilian biota of the Caspian basin Pliocene probably had Mediterranean origin. Way via the Euxinian basin was closed due to real brackish character of the contemporaneous Kujalnician basin. Marine elements among mollusks, ostracodes, foraminifers, polyhaline nannoplankton associations known from Azerbaijan (Golovina, oral communication) show possible corridor from the eastern Mediterranean to eastern Transcaucasia or South Caspian depression. However, the place of this passage (for instance, upper Euphrates valley—Chepalyga, 1995) remains highly speculative.

(4) The effect of the Messinian salinity crisis on the Paratethyan basins is difficult to show. The reasons of the early Pontian transgression and the late Pontian regression of the Eastern Paratethys were probably tectonical. An explanation of the regression as a consequences of an outflow of the Late Pontian waters in the Messinian sea (Semenenko and Teslenko, 1994) is insolvent, because in this case the Bosphorian fauna should appear in the Messinian sea, but the "Lago-Mare" fauna does not contain any specific Bosphorian species.

When seismic sequence stratigraphic studies of the basin fill started in the Pannonian basin, several authors expressed the view that the level of Lake Pannon rose and fell in phase with global eustacy (e.g. Pogácsás et al., 1988, 1992; Vakarcs et al., 1994). This approach already implied what Csató (1993) explicitly claimed: that the geometry of the intra-Carpathian basin fill does reflect the Messinian salinity crisis. This statement is as difficult to prove as to falsify, because the chronostratigraphic resolution of the Pannonian basin deposits is too poor in this time interval for such extrabasinal correlation (Magyar et al., 1999b).

3. Conclusions

The Attic and Rhodanian orogenic phases dramatically changed the framework of deposition and palaeogeography in the Mediterranean basin, and initiated large-scale tectonic and sedimentary-palaeogeographic turnover in the Late Neogene Paratethyan Realm (Steininger et al., 1985; Popov et al., 2004). Sedimentation in the fore-Alpine depression finished. The intra-Carpathian basin became a closed brackish lake-sea, which was gradually filled by lacustrine, deltaic, and fluvial deposits. Lake Pannon continuously shrank as delta progradation proceeded from the northwest and northeast, from the uplifted Alps and Carpathians. Sometimes during the Pliocene the thermal subsidence of the basin was inverted, and uplift and erosion of the Neogene sediments started in large portions of the basin, leading to the present-day topography.

Main orogenesis in the Greater Caucasus and in the mountain chains bordering the Eastern Paratethys from the south (Pontides, Lesser Caucasus, and Taurus) as well as volcanic activity took place in the Pliocene (Late Kimmerian), following earlier compressional movements during the Late Sarmatian and the Maeotian. The beginning of uplift in the Late Sarmatian dramatically changed the axial structure of the Euxinian–Caspian basin and resulted in the growth of the pre-Caucasus structure, surrounded by coastal sediments. The uplifting was most extensive in the central Caucasian traverse. Deposition in the Eastern Paratethys was affected by the uplift of the Greater Caucasus, faulting in the Lesser Caucasus area, and a general regression. As a result of these orogeneses, the Black Sea and South Caspian depressions obtained their modern outline (Tugolesov et al., 1985) and the Eastern Paratethys was subdivided into the Euxinian and Caspian basins from the second part of the Pontian.

Sedimentation in the Eastern Paratethys was dominated by shallow water deposits. Detrital-shelly facies developed on the Eurasian shelf during the Maeotian and Pontian, and alluvial-deltaic clastic deposition prevailed around growing orogens and big river deltas. Marine depositions of large areas in Transcaucasia, in the Turanian plate, and in the fore-Kopet Dagh depression were severed. A spacious coarse-clastic molasse belt developed along the foothills of the Lesser and Greater Caucasus, Kopet Dagh, and Tien Shan. During the Maeotian, the Eastern Paratethys gradually turned into an intracontinental brackish basin. The two Euxinian and Caspian basins existed as long-lived lake-seas until the Middle Pliocene (Akchagilian–Caspian part) and the mid-Pleistocene (Karangatian–Euxinian part), respectively. Nevertheless, biogeographic data testify closer connections between the Paratethyan and Mediterranean sea bodies than it was proposed before. Short-term marine ingressions into the Eastern Paratethys took place during the entire Maeotian, Pontian, Kimmerian, and Akchagilian, as evidenced by presence of Mediterranean benthos (in Maeotian and Akchagilian) and phytoplankton findings.

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