Earth-Science Reviews 246 (2023) 104594

Contents lists available at ScienceDirect

Earth-Science Reviews

journal homepage: www.elsevier.com/locate/earscirev

The legacy of the Tethys Ocean: Anoxic seas, evaporitic basins, and megalakes in the Cenozoic of Central Europe

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ARTICLE INFO

Keywords:
Carbon sink
Paleogeography
Stratigraphy
Endemism
Strait

ABSTRACT

At the end of the Eocene, the demise of the Tethys Ocean led to the formation of one of the largest anoxic seas in the last 50 million years of Earth history. This long-lived anoxic water body, named *Paratethys*, covered large parts of central Eurasia and functioned as a major carbon sink for 15-20 million years, characterised by the deposition of cherts, anoxic turbidites and black shales. The anoxic episode was followed by a phase of instable connectivity where full marine episodes alternated with evaporitic crises and lacustrine episodes, resulting in the deposition of marine molasses, evaporites and continental-lacustrine sediments. Finally, Paratethys transformed into a megalake that progressively filled with clastic sediments from the neighbouring mountain ranges.

Paratethys was tectonically fragmented in numerous sub-basins that spread W-E from the Alpine and Carpathian orogens to the East European Platform. Most Paratethyan stratigraphic records from Central and Eastern European tectonically-active regions are not complete and thus hamper paleogeographic and paleoenvironmental reconstructions. The only exception is the *Outer Carpathian Basin*, located in the external part of the Carpathian arc in Central Europe, that preserved a complete record of Tethys demise and the rise and fall of Paratethys. The Outer Carpathians sedimentary successions show various lithologies that reflect an interplay of interbasinal connectivity and water exchange with the global ocean.

Here we review the stratigraphic schemes of the different tectonic domains of the Outer Carpathians and describe the most complete records to produce a Carpathian-wide framework for the Eocene to Miocene evolution of Paratethys, *the lost sea of Eurasia*. Finally, we focus on the paleogeographic reconstructions of the interbasinal Paratethys connections and discuss how marine connectivity influenced anoxia and hypersalinity and impacted the Cenozoic depositional environments in central Europe.

1. Introduction

The western branch of the *Tethys Ocean* ceased to exist during the Cenozoic, related to Africa-Arabia-Eurasia continental collision (Jovane et al., 2009; Khain et al., 2010; Rögl, 1998). Combined with a global sea-level drop that occurred during the Eocene-Oligocene transition, the *Western Tethys realm* fragmented into two large epicontinental seas (Popov et al., 2004; Rasser et al., 2008) – Paratethys and Mediterranean (Fig. 1). In the north, Paratethys stretched from Germany to China and was separated from the Mediterranean Sea in the south by the Alpine-Carpathian-Himalayan continental collision zone, a tectonically active region characterised by subduction, obduction, back-arc rifting, and shear zone processes (Schmid et al., 2019). The Paratethys realm consisted of a tectonically stable region, the *Eastern Paratethys* (Harzhauser et al., 2002; Jones and Simmons, 1997; Popov et al., 2019; Steininger and Godfrid, 1999), centred around the Black Sea – Caspian basins, and

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https://doi.org/10.1016/j.earscirev.2023.104594
Received 16 December 2022; Received in revised form 3 October 2023; Accepted 9 October 2023
Available online 13 October 2023

a tectonically active region with smaller and shorter-lived marine basins, the *Western Paratethys* and *Central Paratethys* (Kováč et al., 2017b; Popov et al., 2004; Sant et al., 2017). Paratethyan connections with the global ocean, via the Mediterranean and North Sea, controlled circulation patterns and shaped the particular Paratethys environments (Fig. 2; Palcu and Krijgsman, 2021).

1.1. The changing paleogeography of Paratethys: anoxic sea, salt giant and megalake configurations

The exceptionally well-preserved and diverse Paratethyan sedimentary records represent some of the most complete archives documenting the closing of an ocean realm (Fig. 2). These records provide temporal and spatial stratigraphic challenges due to poor connectivity with the global ocean, that greatly reduced the diversity of oceanic taxa and hampers correlations with the standard geological time scale. Consequently, regional stratigraphic schemes have been developed, using cosmopolitan and endemic species and local faunal assemblages as chronostratigraphic markers (Piller et al., 2007b; Rasser et al., 2008, 2016; Raffi et al., 2020). The rise and fall of endemic species were often not centred in one unified basin, so various regional stratigraphic schemes were developed in parallel (Fig. 1) for the major Paratethyan sub-domains.

The Paratethys realm was subdivided in three or two major sub-domains (Fig. 1). Some favour two subdomains: Central Paratethys - covering west and central Europe; and Eastern Paratethys - covering Eastern Europe and Asia, (Papp et al., 1974; Piller et al., 2007a; Popov et al., 2004). Others use three sub-domains (West, Central and East) different in size and life-span: *Western Paratethys* (Fig. 1) roughly matched the North-Alpine Foredeep; *Central Paratethys* (Fig. 1) enclosed (from W to E) the Vienna Basin, Styrian Basin, the Pannonian Basin and the Transylvanian Basin; and *Eastern Paratethys* (Fig. 1), comprising the Scythian Basin, Indol-Kuban Basin, Terek- Mangyshlak Basin, Caspian Basin and the Black Sea Basin (Nevesskaya et al., 1984; Rögl, 1999; Steininger and Rögl, 1984; Vetö, 1987). The Outer Carpathian Basin (OCB), situated between the Central and Eastern Paratethys, initially belonged to Central Paratethys until the Middle Miocene and afterwards became part of the Eastern Paratethys.

The *Outer Carpathian Basin (OCB)* was the largest and longest-lasting tectonically active basin, containing crucial information about connectivity and environmental change during and after the Tethys demise. The sediments of the OCB are intrinsically connected with Carpathian tectonics, older in the Polish sector (Western Carpathians) where thrusting and infilling with sediments was completed by the Miocene,

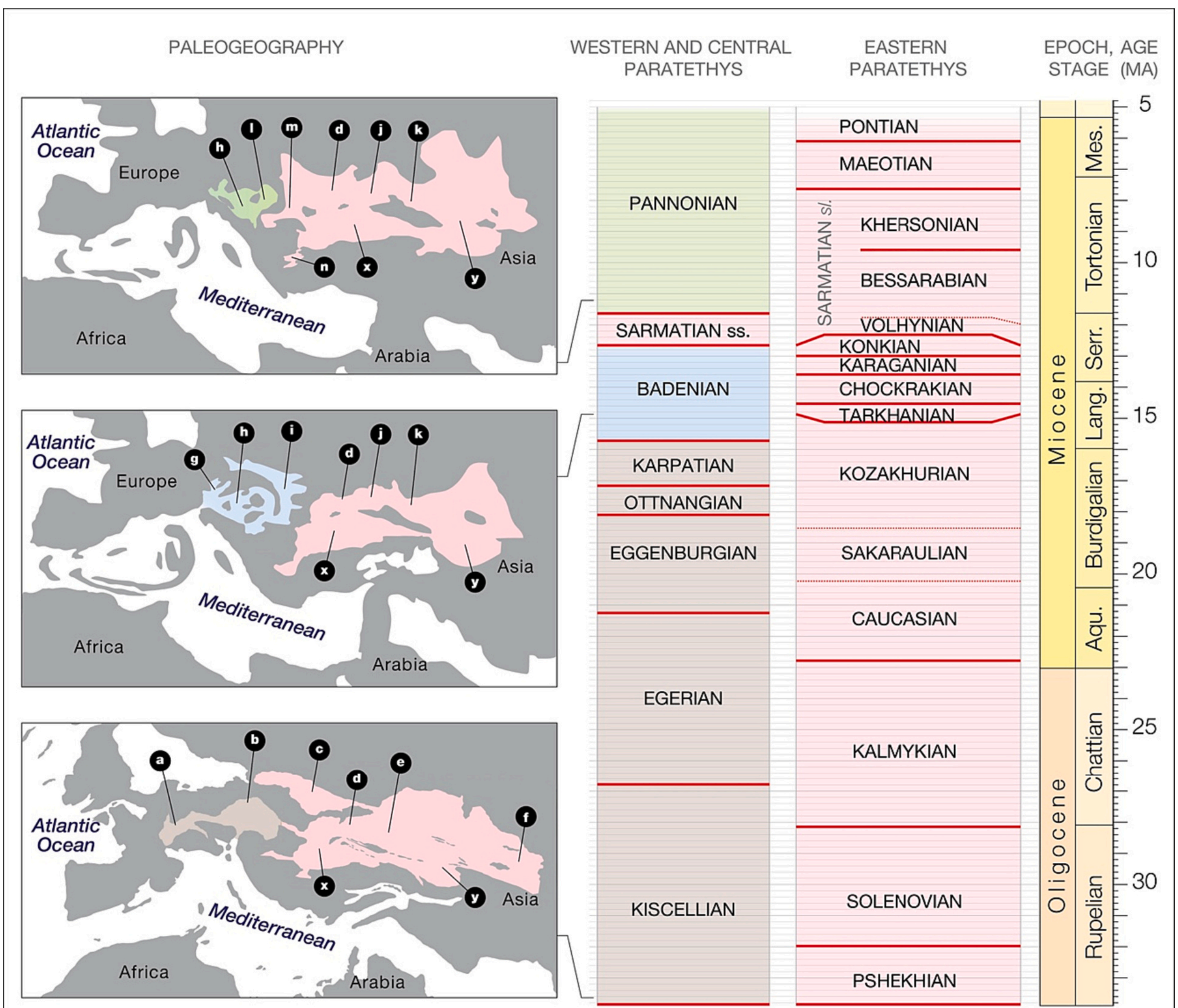

Fig. 1. Paleogeography and stratigraphy of the Paratethyan Realm (Popov et al., 2004; Raffi et al., 2020). The sub-basins of Paratethys during the Paleogene and Neogene: a. North-Alpine Foredeep, b. Outer Carpathian, c. Pripyat, d. Scythian, e. Fore-Caucasus, f. Turan, g. Vienna, h. Pannonian, i. Carpathian Foredeep, j. Indol-Kuban, k. Terek-Mangyshlak, l. Transylvanian, m. Dacian, n. North-Aegean; x. The Black Sea, y. Caspian.

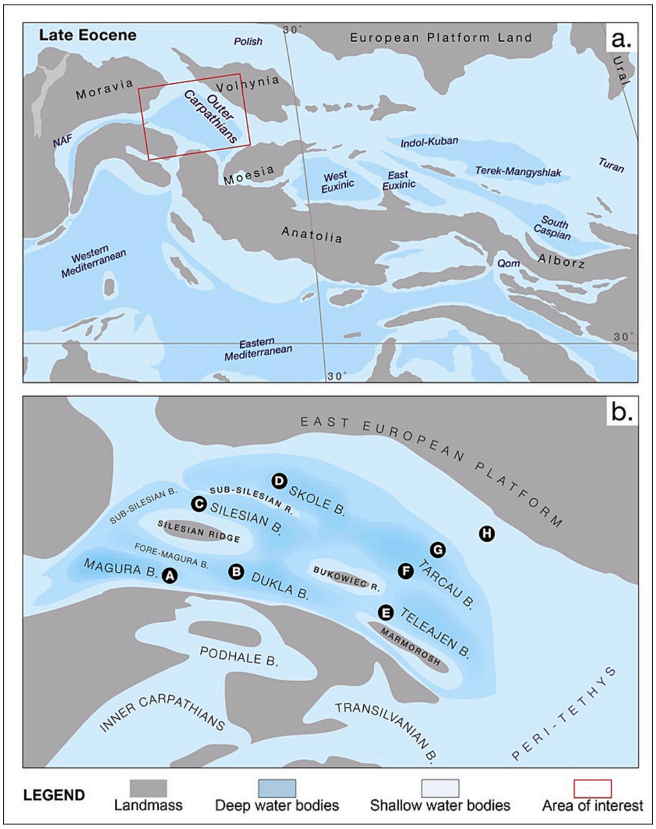

Fig. 2. The northern Tethys Realm sub-basins and ocean connections before the formation of Paratethys The sediment infill history of each of these basins is influenced by basin connectivity and reflects sediment sources from the European Foreland. nearby landmasses (after Ciurej and Haczewski, 2016; Dziadzio et al., 2016; Garecka, 2012; Golonka, 2011; Golonka et al., 2007; Kotlarczyk and Uchman, 2012; Kováč et al., 2016; Książkiewicz, 1977; Leszczyński and Malata, 2002; Oszczypko-Clowes and Zydek, 2012; Palcu and Krijgsman, 2021; Ślączka et al., 2006; Starzec et al., 2018; Teťák et al., 2019). (a.) General map of the Tethys realm during the Late Eocene, highlighting the fragmented character of the northern region of Tethys (Peri-Tethys) Note the weak connections with the global ocean *via* restricted seaways. (b.) The Outer Carpathian basin is characterised by fragmented sub-basins, shaped by regional tectonics: A-Magura b., B-Dukla b., -Silesian b., D-Skole b., E-Teleajen b., F-Tarcău b., G-Vrancea b., H-East

and younger in the Ukrainian and Romanian sectors (Eastern and South-Eastern Carpathians), where the thrust front continued propagating eastwards marking the post collisional stage in the Outer Carpathians (Golonka, 2011). The youngest tectonics occurred in the East Carpathians, namely the Wallachian phase, between Pliocene and Quaternary times (Săndulescu, 1984). Later stratigraphic dating of structures, and thereby paleostresses, argued that the Wallachian phase was not a brief event at the Pliocene-Quaternary boundary, but lasted from middle Tortonian to early Pleistocene (Hippolyte and Sandulescu, 1996).

The Romanian sector of the OCB (South-Eastern Carpathians) contains one of the most complete sedimentary records of Paratethys history, revealing its paleogeographic and paleoenvironmental evolution from the Eocene till Recent. A large part of its literature, however, is found in local Romanian publications, poorly accessible and lacking stratigraphic and geochronological updates. Here, we review the Eocene, Oligocene, and Miocene stratigraphic subdivisions of the Romanian Sector of the OCB. We create a detailed timeline for the lithological and environmental changes and discuss how these reflect the restructuring of the marine Paratethys gateways. This results in revised stratigraphic and paleogeographic reconstructions and reveals how the formation of the main Paratethyan rock units are related to variations in external and internal basin connectivity.

2. The Outer Carpathian Basin

The Outer Carpathian Basin evolution is related to the opening of the Alpine Tethys, in Jurassic - Early Cretaceous times. During the northward movement of the Inner Carpathian ALCAPA block toward the North European plate, an accretionary wedge, marked by the development of thick flysch sequences with olistostromes, triggered the partial closure of the Outer Carpathian basins (Schulz et al., 2005). At the beginning of the Oligocene the Outer Carpathian Basin became stratified and developed anoxic environments, due to poor connectivity with the global ocean. It was affected by changes in ocean connectivity as well as connectivity with the neighbouring *Eastern Paratethys* subbasin (Palcu et al., 2019b). The OCB basin gradually filled with sediments, that are outcropping mainly in Poland, Ukraine, and Romania and to a lesser extent in Slovakia, Czechia.

2.1. Tectonic background

The Carpathians are a range of mountains that stretch in an ~1,500 km long arc across Central and Eastern Europe. Mountain formation was linked to the emplacement of continental blocks during Africa Eurasia collision (e.g., Jolivet and Faccenna, 2000). *The Alpine-Carpathian-Pannonian mega-unit (ALCAPA)* and the *Tisza–Dacia blocks* moved north-eastwards and eastwards and occupied a concave embayment on the *Eastern European-Scythian-Moesian Platforms* of the European foreland (Schmid et al., 2008; Tischler et al., 2008). The resulting accretionary prism comprises various turbiditic successions of Late Jurassic to Miocene age. These accumulated on a thin-stretched continental crust of the European Platform's original passive margin in an array of narrow deep-water basins separated by sub-aqueous to sub-aerial ridges, referred to as "cordilleras" (Fig 1b).

From a geotectonic perspective, the Carpathians are traditionally subdivided into an older range, known as the Inner Carpathians, and a younger range known as the Outer Carpathians (Ślączka et al., 2006). The Outer Carpathians are made up of a stack of nappes, each associated with a sub-basin. Due to intense folding and overthrusting, only sediments from the central parts of these sub-basins are generally preserved (Ślączka et al., 2006). Three main depositional stages are observed: an early anoxic stage characterised by the development of black shales (Jurassic to Albian); a second oxic stage marked by the formation of red and variegated shales (Cenomanian–Eocene) and a third and final stage characterised by brown bituminous shales (Oligocene–Early Miocene) (Băncilă, 1958; Dumitrescu, 1952a; Dziadzio et al., 2016; Leszczyński, 1997; Oszczypko-Clowes and Zydek, 2012; Rauball, 2020; Rauball et al., 2019; Săndulescu, 1984; Ştefănescu, 1978). The thin-skinned nappes and intra-nappe imbricate thrust-sheets of the OCB were tectonically stacked in the Late Oligocene to Early Miocene and piled up northwards and eastwards onto the Miocene foreland basin at the flexural margin of the European Platform (Leszczyński et al., 2015; Leszczyński and Malata, 2002).

In the western *Outer Carpathians* (Poland and Czechia), these nappes consist of the Magura Nappe, the Fore-Magura Group of nappes, and the Dukla, Silesian, Subsilesian, and Skole nappes (Fig. 2).

In the eastern *Outer Carpathians,* the nappes are known as Skole, Skyba, and Boryslav–Pokuttya nappes in the Ukrainian sector (Ślączka, 1996; Ślączka et al., 2006) and correspond (Fig. 3) to the Tarcău, Vrancea and the Subcarpathian nappes in the Romanian sector (Băncilă, 1958; Dumitrescu, 1952b; Mrazec and Popescu-Voiteşti, 1914; Săndulescu, 1984, 1994). The Romanian terminology employs the Dacides and Moldavides names to describe tectonic units of the Eastern Carpathians. The *Dacides* represent nappes and thrusts of Cretaceous and Miocene ages, while the *Moldavides* correspond to Cenozoic nappes and thrusts (Săndulescu, 1984). The Carpathian Foredeep represents the youngest depocenter where molasse deposits accumulated. It consists of

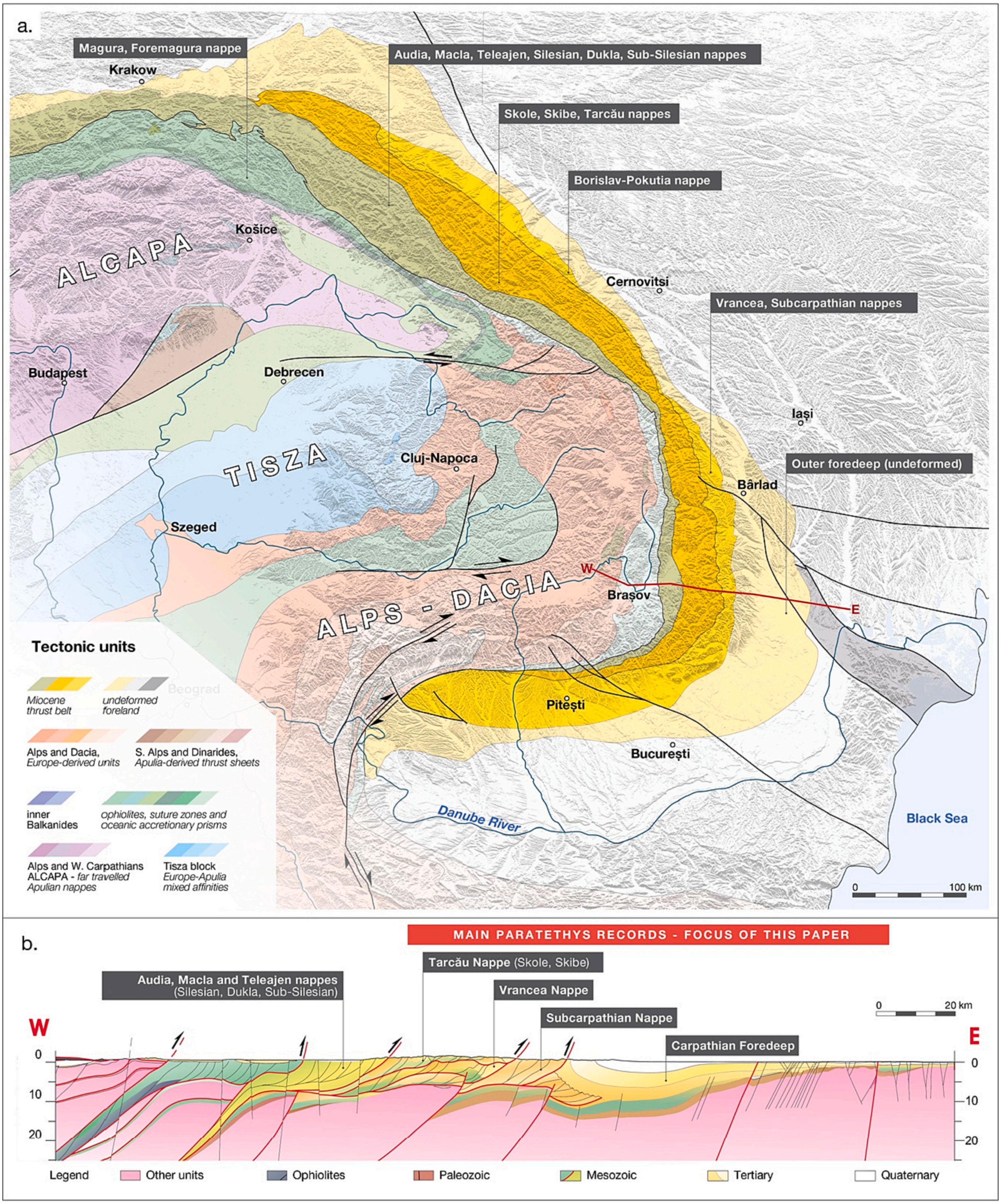

Fig. 3. The Carpathian structural domains linked with Paratethys (after Matenco et al., 2010) a. The Outer Carpathian nappes, b. The cross-section in the Carpathian bend region (no vertical exaggeration) highlights the tertiary deposits of interest.

two distinct zones; a folded internal zone of the Carpathian Foredeep, in contact with the Carpathian nappes (e.g., the Diapir Folds zone) and an external zone characterised by undeformed foreland deposits.

The complete Paratethys record of the OCB, covering the Oligocene – Pliocene interval, is present in the Eastern Carpathians of Romania, preserved in several successions exposing sediments of the Tarcău, Vrancea and Subcarpathian nappes and the Carpathian Foredeep.

The Tarcău Nappe (Săndulescu, 1994) is a polyfacial unit with sedimentary formations of the Lower Cretaceous to the Lower Miocene. During the Paleogene, the geometry of the Tarcău basin was extremely complicated, with many supply sources (internal and external) being active at different degrees, at different times (Bădescu, 2005). This complexity was gradually reduced throughout the Eocene. Starting with the Oligocene, the sedimentary facies became largely uniform, supplied by external sedimentary sources from the platform regions in the east and northeast and internal sources from the Carpathian orogen (Roban et al., 2022). In Miocene times, the Tarcău nappe is significantly reduced due to post-Miocene erosion and non-deposition (Bădescu, 2005). Complete Miocene successions are found only in the Romanian Carpathian Bend, where evaporitic sedimentation (gypsum of *the Cornu unit, along with conglomerates of the Brebu Fm.*) occurred in the Early Miocene, followed by molasse deposits (*Doftana Fm.* – Lower Miocene) the first molasse appearance in the Tarcău basin. Salt, stromatolites, tuffs, and breccia levels also occur (Ştefănescu et al., 1981).

The Vrancea Nappe (Marginal Folds Nappe) (Dumitrescu, 1952a; Ionesi, 1971; Săndulescu, 1994) emerges in a series of tectonic semi-windows (Vrancea, Slănic-Oituz, Bistriţa, Putna) or tectonic windows (Dumesnic, Mitocu lui Bălan) and has been intercepted under the Tarcău nappe by a series of drillings (Bădescu, 2005). This unit is present in the Polish Ukrainian sector, known as *Boryslav-Pokuttya Nappe*. Lower Cretaceous shales represent the main tectonic decollement of the nappe that consists of sediments of Lower Cretaceous-Miocene age. Like the Tarcău basin, the Vrancea Nappe basin is characterised by diverse sedimentary facies, with limited flysch deposition, occurring mainly in the Paleocene- Eocene interval.

The Subcarpathian Nappe is the outermost unit of the Moldavides (Săndulescu, 1984). It was initially separated (Mrazec and Popescu-Voiteşti, 1914) as an incipient nappe, overthrust above the salt formation, and later confirmed by drilling and also named *Pericarpathian unit* (Băncilă, 1958). The Subcarpathian nappe contains Upper Eocene and Oligocene deposits and Lower Miocene molasse within the Carpathian bend zone.

The Carpathian Foredeep consists of a folded internal flank and an external unfolded flank, separated by a flexure. The sedimentary successions are predominantly represented by Upper Miocene-Pliocene molasses with components exclusively of Carpathian origin. The internal flank of the Foredeep, also known as *Diapir Folds Zone* represents the youngest Alpine structure in the foredeep area of the Eastern Carpathians (Ştefănescu et al., 1981). This zone also comprises the Wallachian folds that affected the Carpathian foreland along the *Intramoesic* and *Peceneaga-Camena* faults (Hippolyte et al., 1999).

2.2. The Stratigraphy of the Romanian sector of the Outer Carpathian Basin (OCB)

A long history of research (initiated in the late 19th Century), different schools studying Paratethys geology (e.g., Moscow, Vienna), and the 20th Century geopolitical isolation during the cold war led to the development of parallel terminologies and the attribution of different meanings to the same rock units. In the southern sector of the OCB, Romanian-specific stratigraphic and lithologic terminologies have been employed. Here we provide a simplified framework to describe key paleoenvironmental and palaeogeographic changes in the southern sector of OCB.

2.2.1. Eocene Tethys: turbidite deposits

The Eocene OCB deposits mainly represent expressions of the Tethys realm, with complex geometry, characterised by flysch deposition in well-ventilated environments connected with the global ocean (Tab. 1). During the Eocene, the *Tarcău-Krosno basin* (Fig. 2b) was characterised by three lithofacies areas with different sediment supply and environmental characteristics. In the western (inner) part of the Tarcău basin,

Table 1
Eocene – lowermost Oligocene Tethyan flysch and hemipelagic formations in the SE Outer Carpathian Basin.

TARCĂU NAPPE
TARCĂU LITHOFACIES. Internal (western facies) supplied with sandy, mica-rich sediments from the emerging ridges and the submerged cordilleras of the Carpathian orogen. Main sub-units:
Tarcău Fm. Sandstones with small intercalations of red or green clays (~2000 m thick) contain bioturbations at their base (10 m thick) and frequent intercalations of clays, marls, limestones, and conglomerates with schist clasts. Sediments with rich ichnofaunas indicative for an oxic depositional regime (Brustur, 1997). The oxic regime is indicated also by the presence of variegated (mainly red) shales (Giurgiu-Ghelinţa beds – Săndulescu, 1984). *Age: Middle-Upper Eocene.*
Podu Secu Fm. Flysch composed of a rhythmic alternation of calcareous sandstones with clays rich in bioglyphs, informally named 'hieroglyphic beds' (Waśkowska, 2015). Towards the top, variegated shales (mainly red) occur. The top of this unit is marked by a dm-thick level rich in globigerina, named the "globigerina marl". *Age: Upper Eocene*
DOAMNA LITHOFACIES. External (eastern facies) supplied with clay, marl, and limestone material from the platform units in the eastern part of the basin. Main lithological units:
Straja Fm. Silica-rich unit consisting of clays and silts with glauconitic quartz-rich sandstones, limestones, and silicolitic sponge-rich beds. Coloured in brick-red with thin green and striped intervals, with interbeds of glauconitic sandstones and silicolites (gaize, spongolites, radiolarites). Locally, contains grey to blackish shales; Dysoxic.
Suceviţa Fm. A rhythmic alternation of green or grey calcareous sandstones, variegated shales, in cm-dm thick banks and greenish-grey clays, locally grading into white quartz sandstones (Scorbura sandstone). It is the lateral equivalent of the Leşunt fm. of the Vrancea nappe. Oxic depositional regime.
Doamna limestone Fm. A mainly pelagic unit - fine-grained limestones, white-yellowish, greenish or brown, with siliceous concretions (Bădescu, 2005); Oxic regime.
Bisericani Fm. A succession of clays, marls, grey or green siltites with occasional sideritic concretions. The beds also contain lenticular intercalations of breccias and conglomerates with green schists (sourced from the foreland) in a clayey matrix. Oxic depositional regime based on ichnofaunas (Brustur, 1997).
Lucăceşti Fm. Variable thickness (5-30 m) composed of intercalations of conglomerates, sandstones, and silt-clay layers. It is synchronous with the Ardeluţa Fm. and similar to the Oligocene Kliwa Fm., but unlike the Kliwa Fm., the Lucăceşti is associated with oxic environments (Amadori et al., 2012, Belayouni et al., 2009).
TAZLĂU LITHOFACIES. Intermediate facies in the Tarcău basin, an interference of the western, mica-rich Carpathian orogen and eastern, marl and limestone-rich platform sediment sources. Main sub-units:
Straja Fm. Alternating clays and silts with glauconitic quartzite sandstones, limestones, and red-brick coloured sponge-rich beds. Predominantly arenitic towards western areas. Oxic (Amadori et al., 2012)
Tazlău Fm. Flysch-type sandstones and variegated shales (hieroglyphic beds) rich in micaceous content and greenschist detritus with very rare intercalations of Tarcău sandstone. Typically developed in the central areas of the Eastern Carpathians, it correlates laterally with the Ciunget fm. and the Leşunţ fm. To the south, it merges with the Colţi-Valea Rea Fm. (Micu, 1982).
Plopu Fm. (synchronous with Podu Secu and Bisericani fm.) Consist of clay-sandstone flysch similar to the variegated flysch, but more clayey, with thinner sandstones and intercalations of red clays in the lower part (Bartonian). It appears to have anoxic levels in the top part (Belayouni et al., 2007).
Lupoaia Fm. (synchronous and similar to Ardeluţa Fm. and the Lucăceşti Fm.).
VRANCEA NAPPE
Lithological units similar to Tarcău nappe. These units occur in tectonic windows and half-windows with deposits of Eocene and younger ages in the north and Paleocene and younger ages in the south. Main sub-units:
Piatra Uscată Fm. (= Straja fm.) located in the Vrancea half-window.
Jgheabu Mare Fm. (= Suceviţa fm.) outcropping in the Bistriţa half-window; Oxic (Amadori et al., 2012).
Greşu Fm. and Buciaş Fm. (= Suceviţa Fm.), located in the Vrancea half-window; The upper third of the formation contains a series of intercalations of limestone beds with siliceous accidents equivalent to the Doamna limestones (Săndulescu, 1990).
Bisericani Fm. and the Lucăceşti Fm., described above.

the lithofacies consists of deposits rich in sands with high content of micas. This was related to an active supply from the emerging ridges of the Carpathian orogen or emerged cordilleras (Fig. 2b). To the east, the *Doamna lithofacies* consists of very diverse deposits, rich in limestones and marls supplied from the foreland. A third area, the *Tazlău lithofacies* formed in between the Tarcău and Doamna basins (Ştefănescu et al., 1981), represents a transition with alternations of western and eastern supply. The Vrancea Nappe contains deposits similar to the Tarcău basin (Table 1).

The Early to Middle Eocene (Ypresian-Lutetian) stratigraphic record

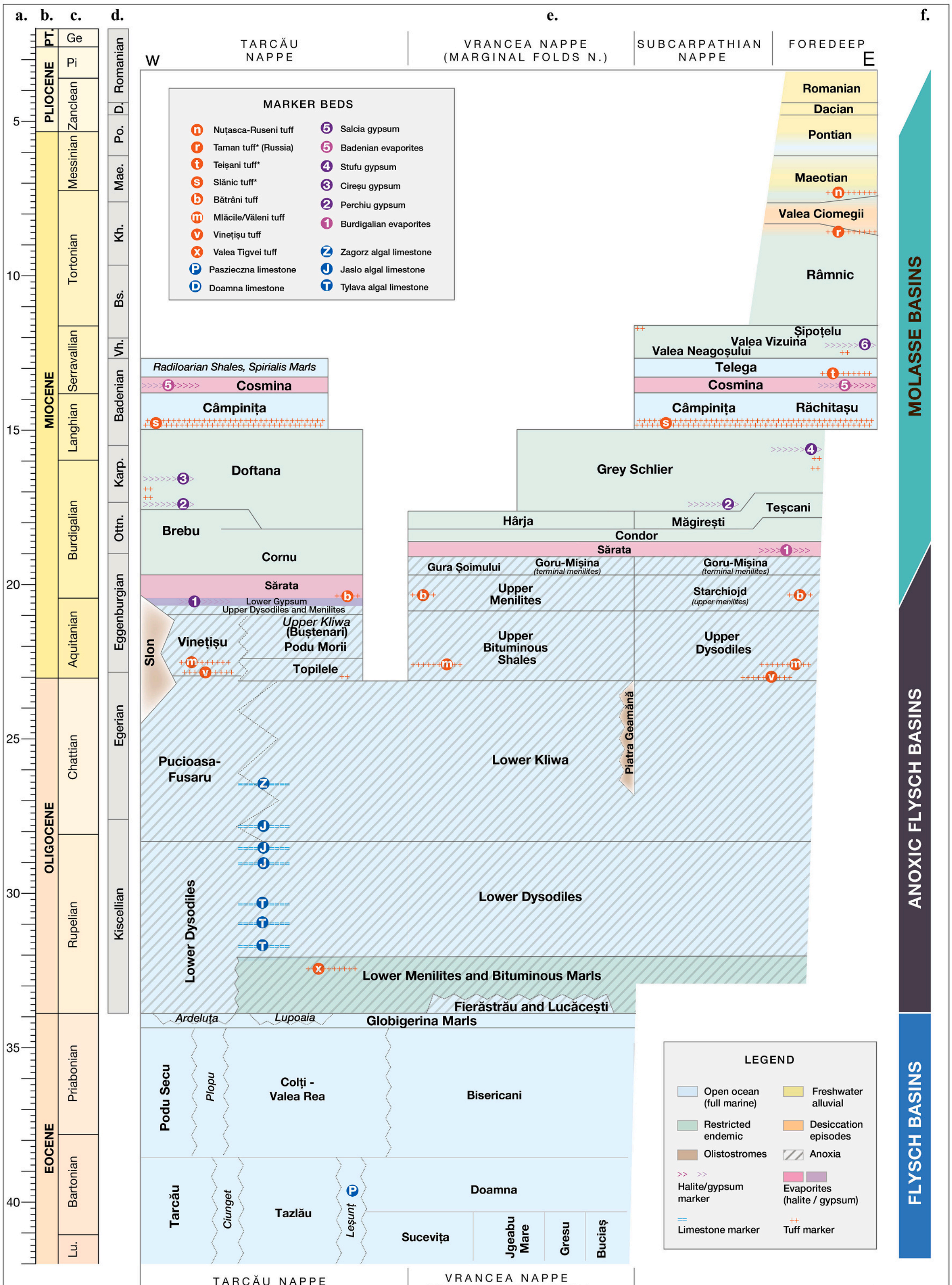

Fig. 4. Stratigraphic chart (simplified) of the Eocene-Pliocene formations from the Romanian Carpathians (references in the text and tables) **:** a. ages; b, c. global epochs and stages; d. regional stratigraphic units; e. key stratigraphic units and paleo-salinity estimations; f. simplified depositional environments. Acronyms: Lu – Lutetian, Pi – Piacenzian, Ge – Gelasian, Vh – Volhynian, Bs – Bessarabian, Kh – Khersonian, Mae – Maeotian, Po – Pontian, D – Dacian.

of the OCB is characterised by numerous sedimentary formations (Fig. 3). From the inner to the outer parts of the basin, the main lithofacial units and/or formations are: *Tarcău, Straja, Ciunget, Tazlău, Sucevița, Leşunt, Jgheabu Mare, Colţi, Greşu and Buciaș* (Fig. 3; Table 1). Most facies comprise massive flysch-type sandstone that contains thin intercalations of red or green clay, variegated shale occasionally rich in siderites, conglomeratic lenses, limestone intercalations and siliceous concretions (Fig. 4).

The most common Middle-Upper Eocene (Bartonian-Priabonian) deposits of the OCB are termed *Podu Secu, Plopu, Paszieczna, Doamna, Lucăceşti, Colți, Bisericani* facies/formation (Fig. 5; Table 1). In the Tarcău domain, they mainly comprise clay-sandstone flysch or variegated flysch composed of a rhythmic alternation of calcareous sandstone and red and/or green marls (Fig. 5).

In the Vrancea and Marginal Folds domain, monotonous successions of clays, clayey marls, grey or green silts are present in which, locally sideritic concretions appear. All flysch units are generally overlain by an uppermost Eocene marker bed, a calcareous unit known as *"globigerina marl"* that frequently contains sandstone intercalations. In some cases, it is even completely replaced by thick sandstone. The marls develop when a pelagic sedimentation regime is installed and deposits rich in planktonic foraminifera, especially *Globigerina* sp., accumulated (Książkiewicz, 1975).

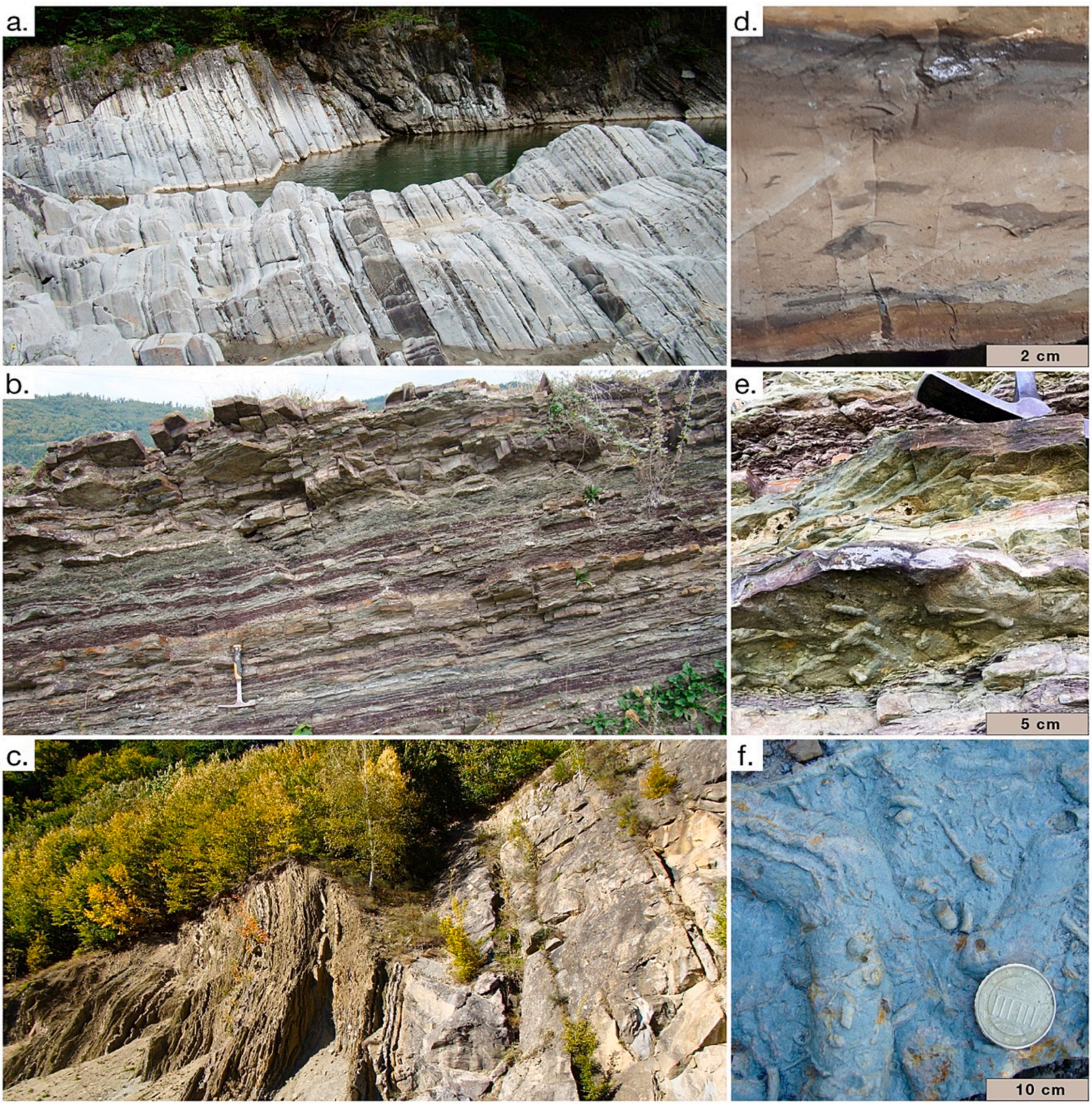

Fig. 5. Eocene facies of the Tethyan flysch formations. b. Doamna limestone Fm., b. Podu Secu Fm., c. Plopu Fm, d. Doamna Fm. - limestones with siliceous nodules *(detail)*, e. The variegated aspect of the flysch from Podu Secu Fm. *(detail)*, f. Bioturbations are very common in the flysch of the Plopu Fm *(detail)*. (Photos: A. De Leeuw, I. Maris)

2.2.2. The Oligocene – Early Miocene Paratethys: bituminous flysch formations

After the Eocene-Oligocene transition, the Paratethyan environments show large-scale *anoxic conditions*. The Oligocene and earliest Miocene deposits represent a dysfunctional marine realm, characterised by flysch deposition in stratified and poorly ventilated environments (Fig. 6). As a result, successive sequences of *bituminous clays* (dysodilic shale - *see supplementary materials*) and *bituminous cherts* (*menilites – see supplementary materials*) developed in the external part of the OCB (Pătruț, 1955; Popescu, 1952). The lower menilites and dysodiles units represent a peak in anoxia during the Rupelian. These anoxic deposits are overlain by the massive Kliwa and Fusaru sandstone units (Săndulescu, 2011; Pătruț, 1955; Popescu, 1952), the main reservoir rocks of the OCB (see section 2.2.2). The main stratigraphic units of the bituminous formation are concislely presented in Table 2.

The overall aspect of bituminous turbidites is the occurrence of fine-grained anoxic sediments, interrupted from time to time (in different areas) by masses of clastic sediments. These clastics have two source areas: an internal source provided Carpathian-type material for the Fusaru clastic units, and an external source provided the Kliwa-type orthoquartzite material from the foreland edge (currently subducted under the Moldavide nappes) (Frunzescu, 2013).

Unlike the Eastern Paratethys anoxia, where micropaleontological markers discontinuous or absent (Sachsenhofer et al., 2018), in the Central Paratethys, i.e., the OCB deposits preserved in the Oligocene foraminifers (mainly benthonic) and nannofossils. The identification of calcareous nannofossil zones provided age estimates in most formations and, together with carbonatic and tuff marker beds, have resulted in a geochronologic reference frame (Supp. fig. 1). The intercalated carbonatic marker beds (e.g., *Tylawa, Jaslo* and *Zagórz* - *see supplementary materials for more information*) are thin layers (mm to dm-thick) of pelagic coccolith limestone (Alexandrescu and Brustur, 1985; Ciurej and Haczewski, 2012, 2016; Grasu et al., 2018; Haczewski, 1984; Ionesi, 1986; Melinte-Dobrinescu and Brustur, 2008; Melinte, 2005; Nowak, 1965; Stefanescu et al., 1993). The *Tylawa, Jaslo* and *Zagórz* limestones have been correlated across major tectonic units and facies zones of the Outer Carpathians over distances up to 550 km and have been accepted as chronohorizons (Bojanowski et al., 2018; Racki and Narkiewicz,

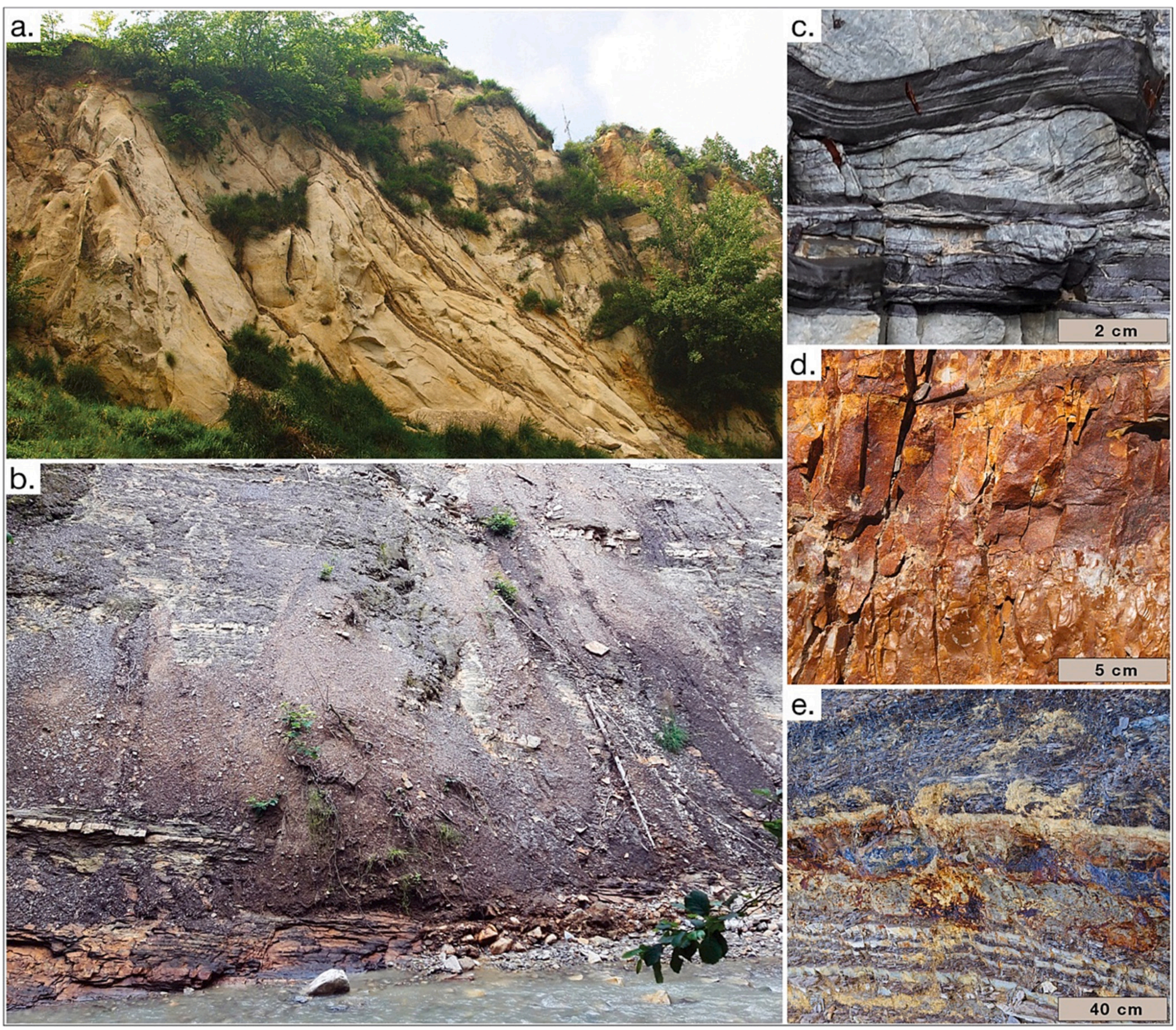

Fig. 6. Oligocene facies of Paratethyan anoxic flysch formations: a. Upper Kliwa Fm.; b. Lower Dysodiles, with concretion levels, c. Bituminous Marls – alternations of carbonatic (white) and organic rich (black) levels *(detail)*; Menilite surface with glassy appearance and an iron oxidation crust *(detail)*, e. Dysodiles with sulphur-rich levels (jarosite) and oxidation crusts *(detail)*.

Table 2
Bituminous flysch formations (Lower Oligocene – Lower Miocene) in the SE Outer Carpathian Basin of Paratethys.

FUSARU-PUCIOASA FACIES. Internal (Carpathian) bituminous lithofacies
Bituminous marls with a specific bluish-white alteration. Age: earliest Rupelian, upper part of NP21 nannofossil zone (Melinte, 2005).
Lower Menilites are organic-rich sediments rich in silica; Age: early Rupelian, lower part nannofossil zones NP22-NP23 (Melinte, 2005).
Lower dysodiles Fm. The sequence begins with black to brownish menilites; cherts or dark red siliceous rocks (widespread marker horizon), interbedded with anoxic shales. They frequently contain fish scales. Peak Anoxia (Amadori et al., 2012; Belayouni et al., 2009).
Pucioasa Fusaru Fm. Part of the Krosno-Fusaru lithofacies, similar to the Eocene Tarcău sandstone, a mica-rich greywacke (Săndulescu et al., 1995), characteristic of proximal flysch with occurrences of conglomerates containing mainly rolled clasts of Carpathian origin. The external occurrences of the Fusaru sandstone contain levels of Kliwa sandstone, leading to an alternation of sandstones with Carpathian clasts and Kliwa-type quartz sandstones; Age: late Oligocene (Chattian) - early Miocene (Aquitanian) interval (Roban et al., 2022).
Vinețișu Fm. (Grigoraș, 1955) Izvoarele beds (Pătruț, 1955). Thick "hieroglyphic" turbidites (1000–2000m) consisting of alternating grey sandstones and marls (Rusu et al., 1996). The lower part of the formation has two tuff marker beds (Stefanescu et al., 1993): the Vinețișu tuff (green-blackish with biotite, NN1 biozone) and the Mlăcile tuff (white, strongly bentonised, NN2 biozone).
Starchiojd Fm. (Popescu and Popescu, 2002); Upper dysodiles and menilites (Coquand, 1867; Dumitrescu, 1948). Black-grey dysodilic shales alternating with cherts (menilites), brown diatomites (white when altered), and rare poorly-cemented Kliwa type-sandstone. The base of the formation contains tuffs (Bătrâni tuff). Thick diatomite packages (20–45m) occur in the Buzău River area; Age: late Aquitanian-early Burdigalian, nannofossil zone NN2 (Melinte, 2005).
Slon Fm. (Săndulescu et al., 1995). Lateral facies characterised by olistostromes with wildflysch character (olistoplaques) occurring in the innermost parts of the basin at various levels (particularly common in the interval corresponding to Vinețișu Fm.).
Gura Șoimului Fm. (Stoica, 1953); Goru-Mișina beds (Dumitrescu, 1952a, 1952b). Grey marls alternating with sandstones containing large olistolith bodies from older anoxic formations. Facies restricted to the Vrancea Nappe (corresponding to the Vinețișu Fm.). Sub-Oxic (Amadori et al., 2012).
KLIWA SANDSTONE FACIES. External (foreland) bituminous lithofacies
Lower Menilites with bituminous marls Fm. (Săndulescu et al., 1995). Similar to the external facies (see above).
Lower Dysodiles Fm. Bituminous clay rocks extremely poor in microfauna but relatively rich in fossil fish species (Rupelian). Intermittently interbedded with frequent sideritic limestone levels. Peak anoxia (Amadori et al., 2012; Belayouni et al., 2009). AGE: upper Rupelian-lower Chattian, NP23 and lower part of NP24 nannofossil biozone (Ştefănescu and Melinte, 1994). Contains the Tylawa coccolithic limestones; Age: early Rupelian, NP23 biozone (Melinte, 2005).
Kliwa Sandstone Fm. (Lower Kliwa) (Pătruț, 1955; Popescu, 1952). Sandstone rich in reworked quartz sands and greenschist elements from the foreland. Clayey intercalations, with variable thickness (centimetres to metrics), are represented by bituminous clays with sulphur efflorescences such as dysodilic shales, anoxic. AGE: late Oligocene - earliest Miocene, NP24-NN1 biozones (Melinte, 2005; Roban et al., 2022).
Podu Morii Fm. (Teisseyre, 1911). Alternating clays, thin arenites associated with Jaslo limestones. The lower and middle parts resemble the Pucioasa Beds, the upper part (Izvoarele Beds) contains an alternation of grey, grey-greenish clays, and convolute calcareous sandstones exhibiting biogenic sedimentary structures (e.g., Sabularia & Mammilichriis) at the base as well as pyroclastic interbeds (Văleni tuff); Jaslo coccolithic limestones age: Chattian, NP24 biozone (Melinte, 2005).
Topilele Fm. (Stefanescu et al., 1993). Sandy flysch consists of alternating dysodiles and grey sandstones. Locally present in the Carpathian bend area as a transition between the Lower Kliwa and Podu Morii units; Age: upper Chattian– Aquitanian, the Oligocene-Miocene boundary in the lower part of the formation.
Buștenari Fm. (partly equivalent with Upper Kliwa Sandstone) (Stefanescu et al., 1993). An alternation of sands (Kliwa sandstone and sands, carbonatic sandstone with mica and convolute structures), silts (silty marls and coal-rich silts) and clays (marls, marls with siderites, dysodiles and bentonitic clays). The lower third is dominated by clays, while the remainder is sandy (coarsening-upwards trends, Frunzescu, 1989). Microbreccias with greenschist elements, volcanic tuffs and diatomites can occur in thin layers; Age: early Miocene (Burdigalian), NN2 biozone (Melinte, 2005; Roban et al., 2022).
Starchiojd Fm. (Popescu and Popescu, 2002); Upper Menilites (Stoica, 1944), similar to the internal facies (see above); Age: late Aquitanian - early Burdigalian, nannofossil zone NN2 (Melinte, 2005).

2006).

The anoxic sediments of the OCB are generally characterised by lack or scarcity of limestones and dolomites; thus, the exceptional occurrence of widespread limestone marker beds is very important for stratigraphic correlations. In most sections, the limestones are split by clastic deposits that are usually much thicker than the limestones. Investigations of the *Jaslo Limestone* has shown that the clastic rocks represent event deposits (Haczewski, 1989; Korab and Kotlarczyk, 1977) and the laminated limestone represents the bulk of the geological time (Ciurej and Haczewski, 2012; Jucha and Kotlarczyk, 1959; Jucha and Krach, 1962; Melinte-Dobrinescu and Brustur, 2008). Several tuff beds have been identified throughout the south-eastern OCB (e.g., *Văleni, Bătrâni, Vinețișu* and *Mlăcile* tuffs – (Alexandrescu, 1994; Alexandrescu et al., 1981, 1991, 1992; Alexandrescu and Brustur, 1985; Ştefănescu et al., 1989). They represent good stratigraphic markers, mostly regional, but their ages remain poorly constrained, based only on biostratigraphic data.

In the innermost zone of the *Tarcău nappe*, the *Slon Formation* is interfingering with the *Fusaru Sandstone* and part of the *Vinețișu Formation* , being Upper Oligocene-lowermost Miocene (lower Burdigalian) in age (Melinte and Băceanu, 1996). It contains olistoliths of internal origin, mostly Upper Cretaceous variegated shales. The origin of *Slon Fm.* was likely related to the rising of the neighbouring Inner Moldavides nappes (*Audia and Macla*) during the *Savian tectogenesis* in the Late Oligocene (Stefanescu et al., 2007).

2.2.3. Lower and Middle Miocene Carpathian foredeep: evaporites and molasse formations

The Lower-Middle Miocene deposits of the OCB represent expressions of a molasse basin (Fig. 7), a trapped sea characterised by clastic deposition in anomalohaline environments having fluctuating, generally poor connections with the global ocean (Table 3). During the Early Miocene, tectonic activity triggered significant changes in the depositional regime of the OCB sub-basins (Săndulescu, 1984; Săndulescu et al., 1995) and anoxic turbidites are progressively replaced by molasse deposits and evaporites.

In the Burdigalian, a first stage of folding and detachment occurred associated with NE-ward thrusting (Maţenco et al., 1997; Matenco and Bertotti, 2000; Ştefănescu et al., 1980), known in the classic Paratethys literature as the *Old Styrian tectonic phases* (e.g., Stefanescu et al., 2007). During this tectonic episode, the Tarcău Nappe was largely emerged, and the sedimentary basins receded to the southeast, where molassic and evaporitic formations of late Burdigalian to Langhian age were deposited

Molasse deposition in shallow facies and under limited/absent marine connectivity continued throughout the Early Miocene, punctuated by several local gypsum evaporitic episodes. The oldest (Burdigalian) evaporitic episode, characterised by salt and gypsum precipitation, is recorded in the OCB basins at ~19 Ma (Roban et al., 2022), although other studies place it at around 17 Ma (Mărunţeanu, 1999).

The Lower Miocene anoxic flysch is replaced by predominantly evaporitic formations and molasses (*Brebu*, *Doftana*, *Grey Schlier)* throughout the entire OCB domain. The molasse deposits are supplied from two clastic sources: an external source with greenschist clasts from the foreland (in the Subcarpathian and Vrancea nappes) and an internal source supplied by the Carpathians (Tarcău Nappe). The OCB molasse predominantly consists of rhythmic alternations of reddish conglomerate (gravel and sand) at the base and greyish silt and clay at the top (Mărunţeanu, 1999; Săndulescu et al., 1995).

The Burdigalian succession contains several lithostratigraphic markers (Grasu et al., 1999; Olteanu, 1953): the *Perchiu Gypsum* and the *Valea Calului Marls* (represented by reddish clays) are found at the base, while the *Stufu Gypsum* is located in the upper part (Fig. 4; Table 3). The interval between the *Perchiu Gypsum* and the *Valea Calului marls* consists of grey marls and sandstones, called the "*Grey Formation*" (Grasu et al., 1999).

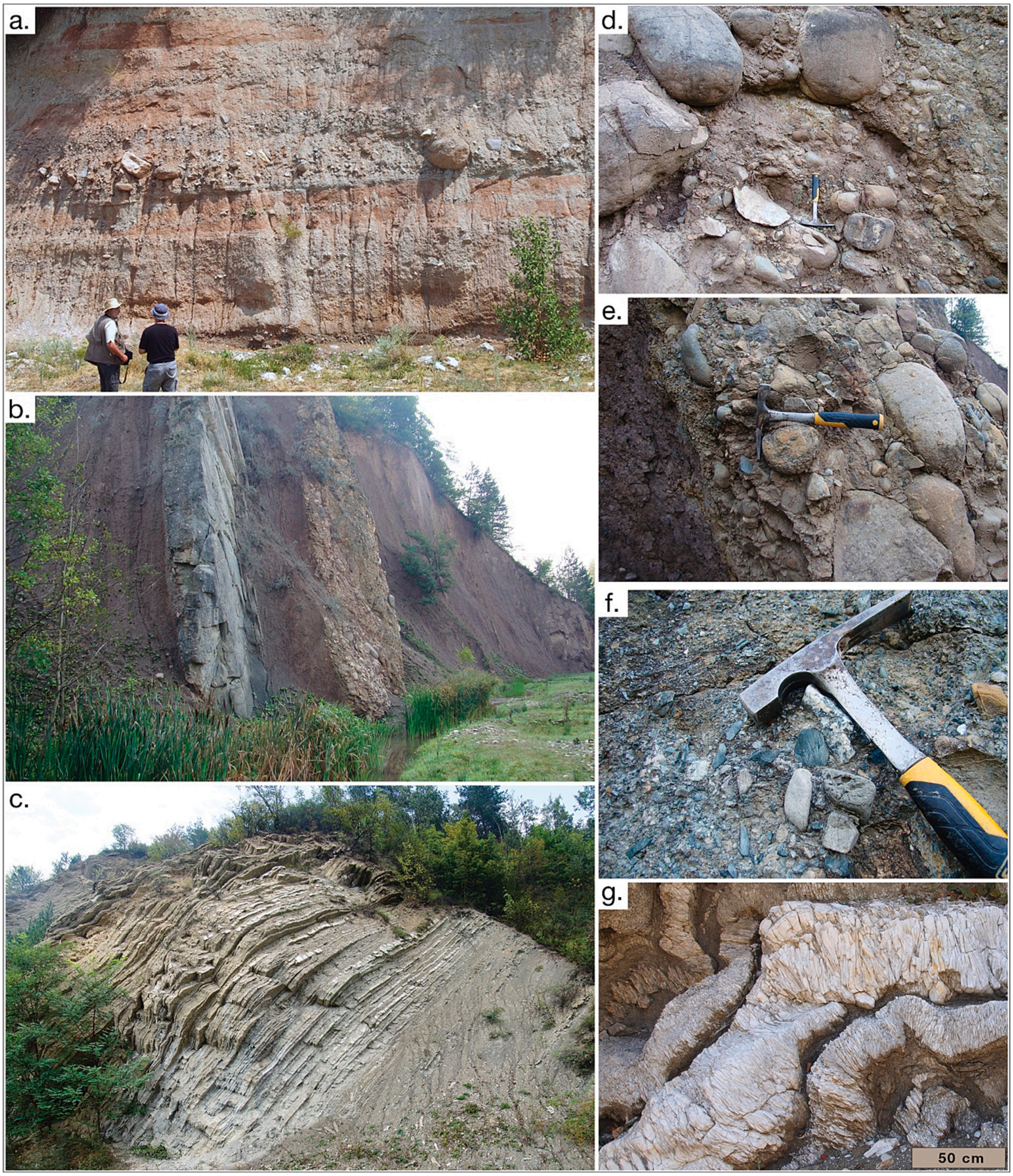

Fig. 7. Key Lower Miocene molasse formations: a, b. Brebu conglomerates (*red horizons*); c. Gypsum layers from the Grey Schlier unit Fm.; d-e Carpathian-type (internal) conglomerates *(detail)*, f. Dobruja-type, greenschist-rich (foreland-derived) conglomerates *(detail)*, g. Gypsum fold *(detail)* in Sărata Fm.

Middle Miocene tectonic activity (Bartol et al., 2012; Matenco et al., 2003; Matenco and Bertotti, 2000; Tărăpoancă et al., 2003), known in the classic literature as the *Young Styrian tectogenesis* (Săndulescu, 1988) and a profound marine transgression (Ćorić et al., 2009; Harzhauser and Piller, 2004; Hohenegger et al., 2009; Holcová et al., 2018; Kováč et al., 2004; Rögl, 1999; Rögl et al., 2002; Rundic et al., 2014; Sant et al., 2019a; Schreilechner and Sachsenhofer, 2007) increased the size and depth-domain of the Carpathian Foredeep, which became the main depositional area in the Outer Carpathian region. The Middle Miocene deposits of the Carpathian Foredeep consist of fine-grained sediments rich in marine fauna (Fig. 8), linked to the mid Langhian or Badenian transgression, which is recognized throughout the entire Carpathian region (Mandic et al., 2019; Sant et al., 2019b; Studencka, 1999), at ~15 Ma (Sant et al., 2019a). The marine episode was followed by a second evaporitic succession linked to the so-called *Badenian Salinity Crisis* of the Central Paratethys, which spanned between 13.8 and 13.4 Ma (de Leeuw et al., 2010, 2013, 2018; Peryt, 2006). In the OCB, the evaporitic succession (Cosmina Fm.) generally started with a relatively monotonous salt breccia with clay matrix and many clasts, both in terms of size and petrography (Crihan, 1999a; Krézsek et al., 2023; Olteanu, 1951a; Popescu, 1951; Săndulescu et al., 1995; Schleder et al., 2019; Tamaş et al., 2018)

Table 3
Early-Middle Miocene Paratethys molasse formations in the Carpathian Foredeep with nannoplankton age estimations from (Mărunţeanu, 1999).

CENTRAL PARATETHYS. CARPATHIAN FOREDEEP. Middle Miocene molasse deposits
Sărata Fm. Rhythmic alternation of layers of gypsum and clayey – silty granofacies rocks (Frunzescu, 1996; Frunzescu and Anastasiu, 1995; Frunzescu et al., 1995). The clastic material consists of black shaley clays, grading into grey-whitish clays with high bituminous content, similar to dysodiles, and grey-purple marls, located at the top of the evaporitic sequence. In the Subcarpathian and Vrancea nappes, the Salt Formation developed as discontinuous salt bodies and intercalations of sedimentary breccias, conglomerates, clays, or saliferous marls and gypsum. Maximum thickness reaches 350-400 m, but its geographical distribution is discontinuous (Frunzescu, 2013). Evaporitic basin (Mărunţeanu, 1999) of Burdigalian age (nannofossil zone NN2b).
Condor Fm. A sequence containing marls, clays, sandstones, and coarse granular arkose sandstone. Shallow basin (Frunzescu, 2000), Burdigalian age (nannofossil zone NN2b).
Cornu Fm. Breccia with clay matrix representing the post evaporitic phase. It is unstratified and with heterogeneous clastic compositions. The upper part contains a grey-yellowish gypsum level (0,50–4m) and a rhythmic flysch sequence with red marls representing the subaerial exposure of older deposits—shallow basin (Filipescu et al., 2020) of Burdigalian age (nannofossil zone NN3).
Brebu Fm. A rhythmic alternation of conglomerate and clayey sandstones with sandy marls, fining upwards. The sandstones are characterised by vertical grain sorting structures, oblique lamination, current structures and ichnologic casts. Shallow basin (Frunzescu, 2000), Burdigalian age (nannofossil zones NN3-NN4a).
Măgireşti Fm. and Pietricica Hârja, Pleşu, Bârseşti conglomerates. Massive conglomerates generally rich in "greenschist" elements and diverse clastic material of Jurassic and Eocene limestones, Permo-Triassic sandstones and Jurassic marls (Săndulescu et al., 1995) located in the external part of Subcarpathian Nappe, on the basin margins. Shallow basin margin (Frunzescu, 2000), Burdigalian age (nannofossil zone).
Măgireşti Fm. (Hârja Fm.) Sandy-marly molasse (Săndulescu et al., 1995) corresponding to the Early Miocene conglomerates and the sandy or sandy-marly molasses that overlay the conglomerates. Shallow basin (Frunzescu, 2000), Burdigalian age (nannofossil zone NN3).
Teşcani Fm. A marly-sandy molasse, rich in "greenschist" clastic material (Mărunţeanu, 1999). Shallow Marginal basin margin (Frunzescu, 2000) of Burdigalian age (nannofossil zone NN4a).
Grey Schlier unit. Marly-sandy molasse, rich in "greenschist" clastic material (Mărunţeanu, 1999). Shallow basin (Filipescu et al., 2020; Ştefănescu, 1984; Ş tefănescu et al., 1980) of Burdigalian/Langhian age (nannofossil zone NN4).
Doftana Fm. Molasse consisting of silty clay layers, often red or grey, containing current structures at the bottom. In the terminal part, yellow-rusty bituminous calcareous shales (stromatolites) occur, followed by brown-rusty, slightly bituminous, compact gypsum. At the top, widespread ichnologic markers indicate intertidal conditions of deposition. Gypsum appears as thin and rare intercalations or as the Perchiu and Cireşu gypsum, two thick, widespread gypsum marker beds. Shallow basin (Filipescu et al., 2020; Ştefănescu, 1984; Ştefănescu et al., 1980) of Burdigalian/Langhian age (nannozone NN4).
Câmpiniţa Fm. (Slănic tuff). Marine succession starts with *"globigerina horizon"* (Popescu, 1951), a marly unit (Crihan, 1999a) marked by major intercalations of volcanic deposits called Slănic tuff in the South Carpathians (Murgeanu et al., 1968), similar to the Dej tuff in Transylvania (Meruţiu, 1912; Popescu-Voiteşti, 1915). The external part of Subcarpathian Nappe is characterised by calcareous sandstones (Răchitaşu Fm). Middle Miocene (15 – 13.8 Ma) age (Crihan, 1999b; Sant et al., 2019a, 2019b).
Cosmina Fm. (Slănic evaporites). Evaporitic succession, linked to the so-called Badenian Salinity Crisis (de Leeuw et al., 2010, 2013, 2018; Peryt, 2006). It generally consists of a monotonous salt breccia with clay matrix and a large variety of clasts, both in size and petrography (Crihan, 1999b; Olteanu, 1951a; Popescu, 1951;Săndulescu et al., 1995). The Badenian evaporitic formation is represented by gypsum and/or halite and can be distinguished from the Burdigalian salt formations by the different detrital material of the matrix (the Badenian evaporites lack "greenschist" components). Selenite gypsum occurs locally (Micu, 1982), similar to the selenites of the external flank of the Foredeep in south Poland and the Sub-Carpathian Ukraine (Frunzescu, 2000). Middle Miocene age (13.8 – 13.4 Ma)
Telega Fm. (Radiolarian shales & Spirialis marls) (Olteanu, 1951b; Popescu, 1951; Crihan, 1999a, 1999b). Deposits with normal marine character (de Leeuw et al., 2018; Popescu, 1995). Depositional environments of the radiolarian shale are favourable to the accumulation of clayey deposits rich in siliceous organisms, radiolarian, ebriids, dinoflagellates, and rarely, foraminifera and nannoplankton. The *Spirialis* (planktonic gastropods) marls consist of clayey granofacies represented by compact grey calcareous limestones, with rare intercalations of sand/sandstone. Middle Miocene age (13.4 – 12.8 Ma).

The Badenian evaporitic formation (early Serravallian) is represented by gypsum and/or halite (Fig. 8) and can be distinguished from the evaporitic formations of the *Early Burdigalian Salinity Crisis* by the different detrital material of the matrix (the Badenian evaporites lack "greenschist" components). Wide crystallized "selenite" gypsum occurs locally (e.g., *Valea Rea,* Micu, 1982) and is similar to the selenite gypsum of the external flank of the foredeep in south Poland and the Sub-Carpathian Ukraine (Frunzescu, 2000).

The Middle Miocene evaporitic succession is overlain by deposits with a normal marine character correlated throughout the basin (Popescu, 1995). These marine environments, such as *Radiolarian shales* and *Spirialis marls* units (Crihan, 1999a, 1999b; de Leeuw et al., 2018; Olteanu, 1951b; Popescu, 1951) are characterised by radiolarian-rich clays, favourable to the accumulation of siliceous organisms, radiolarian, ebriids, dinoflagellates, and rarely, ostracods, foraminifers and nannofossils, along with marls, rich in planktonic gastropods (*Spirialis* sp.), probably indicating a stratified sea. After the Badenian stage, the deposition shifted from open marine to restricted brackish environments (Liu et al., 2017). This change triggered the *Badenian-Sarmatian Extinction Event* (BSEE), dated at 12.65 Ma (Palcu et al., 2015), when most marine fauna of the OCB basin became extinct (Crihan and Маrunteanu, 2006; Filipescu and Silye, 2008; Harzhauser and Piller, 2007; Studencka, 2016; Studencka and Jasionowski, 2011), a feature also observed as well at the Konkian-Volhynian boundary in the Eastern Paratethys (Palcu et al., 2017).

2.2.4. Late Miocene Dacian basin: lacustrine deposits of the Paratethys megalake and Pliocene-Quaternary basin fill

The deposits of the early *Sarmatian regional stage* indicate a joint Eastern and Central Paratethys domain. During the middle Sarmatian, the Tarcău, Vrancea and Subcarpathian nappes were thrusted over the foreland units. The Vrancea nappe became covered by the Tarcău nappe (Krézsek et al., 2023; Stefanescu et al., 2007), except for several tectonic windows (Dumesnic and Mitocul lui Bălan) and half-windows (Vrancea, Oituz, Bistriţa, and Putna). At the beginning of the *Sarmatian regional stage*, the Outer Carpathian region became incorporated in the Eastern Paratethys (Jipa and Olariu, 2009). To distinguish it from the active foredeep, part of the Central Paratethys those times, the name *Dacian basin* is used for the Sarmatian and post-Sarmatian basin. The sediments of the Dacian basin represent the filling of this last relict of Outer Carpathian basins (See Table 4).

The *Sarmatian* succession consists in the monotonous clay-silt formations corresponding to the *Volhynian and lower Bessarabian substages*, often rich in pyrite and pyritized fossils (e.g., insects) – indicating anoxic/suboxic depositional environments (Munteanu, 1998). The clays are covered by intervals richer in sands, which increase in thickness and eventually grade into the sandy sediments, corresponding to the *upper Bessarabian substage* characterised by the predominance of coarse siliciclastic sequences (sands, sandstones, silts, conglomerates) of early Tortonian age (Fig. 9). These sandy units are the expression of extensive alluvial systems that discharged in the Dacian Basin. The largest of these deltaic systems, dubbed Balta Formation (de Leeuw et al., 2020; Matoshko et al., 2016) formed the southern limit of the axial drainage system of the Carpathian foreland basin and developed in vast areas of eastern Romania, western Ukraine and most of the area of the Republic of Moldova.

During the *Khersonian substage* (corresponding to the late Tortonian Stage, global chronostratigraphy), the lithology becomes predominantly characterised by continental deposits alternating with lacustrine/brackish sequences (Fig. 9). In the southeast, the Khersonian succession covers a reduced area and corresponds to a large erosional gap in the former marginal areas of the basin (Munteanu, 1998).

The deposits of the Maeotian regional stage (lower Messinian) (Fig. 9) lay either concordantly over Khersonian limestones and continental deposits or transgressively with frequent angular unconformities over older deposits (Lazarev et al., 2020; Palcu et al., 2019b). The

Fig. 8. Middle Miocene molasse formations. a. Thick green tuffs (Slănic tuff) from the Câmpiniţa Fm., b. Salt diapir from Cosmina Fm., c. Gypsum from the Cosmina Fm. overlain by clays of the Telega Fm.; d. White tuff in the Răchitaşu Fm., e. Slănic Prahova Salt Mine in the Cosmina Fm., f. Algal gypsum from Cosmina Fm., g. Gypsum nodules in Cosmina Fm.

deposits of the Pontian, Dacian and the Romanian regional stages (latest Miocene-Pliocene) have a widespread development and represent the final stage of deposition in aquatic environments, followed by coarse alluvial deposits, terraces and landslides of Quaternary ages (Andreescu et al., 2011). The present-day structural configuration was finished during the Lower Pleistocene tectonic movements (*Wallachian tectogenesis*), when the youngest deposits were also involved in folding and faulting processes (Dicea, 1995; Stefanescu et al., 2007).

3. Paleogeographic and paleoenvironmental reconstructions of the Outer Carpathian Basin

Tectonic rearrangements of the internal seaways and straits, as well as sediment infill have produced many configurations of the Paratethyan realm (Allen and Armstrong, 2008; Dercourt, 2000; Haq, 1981; Harzhauser et al., 2002; Jovane et al., 2009; Oberhänsli and Hsü, 1986), but the evolution of Paratethys after the separation from the Tethys realm (Fig. 1) can be summarized in three key paleogeographic snapshots: ***"Anoxic Giant***", *(Oligocene - early Miocene)* was a phase when large

Table 4
Middle Miocene Paratethys molasse formations in the Carpathian Foredeep.

PARATETHYS, DACIAN BASIN. Miocene (Middle to Upper) to Pliocene molasse deposits
Valea Neagoșului Fm. Clays and blue-grey silty clays, tuffs, sandstone concretions, and conglomerates (Papaianopol, 1992; Munteanu, 1998). Brackish anoxic/dysoxic facies of Middle Miocene age (12.8 – 12.3 Ma; Crihan, 1999a, 1999b; Liu et al., 2017).
Valea Vizuinei Fm. (Munteanu, 1998), also known as Dara beds (Papaianopol, 1992). Clayey formation with gypsum intercalations, deposited between two regressive phases, characterised by coarse clastics and gypsum precipitation (end Kossovian and lower Volhynian). The upper gypsum level (Salcia gypsum) is the most developed. Brackish anoxic/dysoxic facies of Middle Miocene age (12.3 – 11.6 Ma; Crihan, 1999a).
Șipoțelu Fm. Marls and silty clays with levels of fossil-rich sandstones (Andreescu, 1972). It marks a broad development of the Paratethyan endemic faunas of the Sarmatian and corresponds to the local Volhynian and Bessarabian regional stages (Papaianopol, 1992; Munteanu, 1998). Brackish anoxic/dysoxic deposits, Late Miocene in age (11.6 – 9.7 Ma; Crihan, 1999a, 1999b).
Râmnic Fm. Sands with sandstone concretions, sandstones, and clays (Andreescu, 1972), with oolitic intervals, partly corresponding to the "marly-sandy-oolitic levels" (Iorgulescu, 1953), lower part of the Istrița limestone (Papaianopol, 1992), Milcov Fm (= beds, Motaș et al., 1976). A brackish shallowing oxic basin of Late Miocene age (9.7 – 7.65 Ma; Crihan, 1999a, 1999b; Palcu et al., 2019b).
Valea Ciomegii Fm (=beds, Munteanu, 1998). Continental deposits alternating with freshwater and brackish levels known as "variegated complex," "striped complex," "package of the purple-green clays," "purple-green marl series," "brackish-freshwater complex" (Andreescu, 1972; Ciocârdel, 1952; Ciocârdel and Patrulius, 1950; Macarovici et al., 1967; Motaș et al., 1976; Munteanu, 1998; Pana, 1966). Partially desiccated basin with freshwater and continental deposits of Late Miocene age (8.7 – 7.65 Ma; Crihan, 1999a; Lazarev et al., 2020; Palcu et al., 2021).
Maeotian regional stage deposits. Consist in a lower Maeotian depositional interval, with clays and marls, sandy in the peripheric zones and an upper Maeotian depositional interval dominated by sandy marls, micaceous sands, and diverse sandstones. At the top of the lower Maeotian, the sediments are disturbed by slumps and erosional episodes, followed by the redeposition of reworked lower Maeotian sediments and fauna (Pana, 1966). Brackish early Maeotian and freshwater environments with marine influxes during late Maeotian; Late Miocene age (7.65 – 6.1Ma; Lazarev et al., 2020; Palcu et al., 2019b; Stoica et al., 2013; Vasiliev et al., 2021).
Pontian regional stage deposits. Consist of a lower Pontian clayey granofacies, a middle Pontian of predominantly sandy granofacies and an upper Pontian marly-sandy facies. The base of the Pontian contains a short influx of mesohaline to polyhaline microfauna, including benthic and planktonic foraminifera from the Aegean, defined as the "Pontian Salinity Incursion." Restricted marine - freshening (Krijgsman et al., 2010; Stoica et al., 2007; Vasiliev et al., 2010) Late Miocene-Pliocene age, i.e., 6.1 – 4.8 Ma, Jorissen et al., 2018).
Dacian regional stage deposits. An alternation of yellowish-white, fine to medium-grained sands, with intercalations of grey-purple-brown marls, silty marls, coarse granular rusty sandstones, and small gravels in thin layers. To the top, the fine-grained facies is dominant in the form of marls and blackish-brown clays and layers of coal. Brackish-freshwater deposits of Pliocene age (4.8 – 4.4 Ma; Jorissen et al., 2018)
Romanian regional stage deposits. Consist of a lower clayey-sandy subdivision (including brown compact sandy clays, alternating with grey-yellowish sands with low cross-stratification, fine-medium or coarse granular, with grey, brown marls, sometimes with calcareous concretions) and a higher subdivision (mainly gravely, including medium and fine granular gravels with polygenic subrounded and rounded clasts. Freshwater deposits of Pliocene-Pleistocene age (4.4 – 2.2 Ma; Van Baak et al., 2015).
Quaternary deposits. Alluvial deposits, terraces and landslides (Andreescu et al., 2011; Necea et al., 2005). Pleistocene gravels frequently occur deformed, highlighting the recent appearance of the Wallachian tectonic phase (Hippolyte and Sandulescu, 1996).

anoxic seas experiencing various degrees of anoxia, developed throughout Paratethys; ***"Changing Seas"*** *(early and middle Miocene)* was an episode when the sub-basins of Paratethys experienced alternations between fully marine and anomalohaline environments (hypersaline or hyposaline); ***"Megalakes"*** *(Late Miocene)* was the last stage of Paratethys, characterised by *poor connectivity with-* or *complete disconnections from-* the global ocean and brief incursions of marine water masses from the Mediterranean Sea. The lithological successions from the southern sector of the Outer Carpathian basin represent the only complete record of these paleogeographic and paleoenvironmental changes from the tectonically active region of Paratethys.

3.1. Eocene open ocean environments with short-term local anoxia

The Eocene stratigraphic units of Central Europe mostly reflect deep, tectonically-dynamic, open-ocean environments. Paleogeographic reconstructions (Fig. 10a) point to a particular morphology, characterised by deep narrow basins, and abundant clastic fill. The clastic Kliwa, and Fusaru (Krosno) sandstones reflect the proximity to *cordilleras* that were emerging and undergoing erosional processes, but their occurrences are mainly related to the regional tectonic setting and are not linked with paleoenvironmental changes. Throughout the Eocene, multiple gateways existed in the western Tethys region, allowing a good connection between the Tethyan sub-basins and between Tethys and the global ocean (Palcu and Krijgsman, 2021). The Eocene is generally perceived as the time when the northern Tethys region was characterised by well-connected and well-oxygenated basins, with short-lived local occurrences of anoxic sediments described in the OCB: e.g., the "black shale" of the Silesian (Tarcău) Nappe of the Ukrainian Carpathians (Hnylko and Hnylko, 2019), the "black Eocene" of the Polish Carpathians (Waśkowska et al., 2015). In the Eocene of the Eastern Paratethys a period of more significant restriction occurred, leading to the Kuma anoxic event (Beniamovski et al., 2003; Cramwinckel et al., 2022; van der Boon et al., 2019; Zastrozhnov et al., 2019). Limited age control exists and palaeogeographical constraints are too poor to establish if and how these restrictions are related to the Middle Eocene Climatic Optimum or its aftermath (Bohaty et al., 2009; Cramwinckel et al., 2022; Giorgioni et al., 2019; Sluijs et al., 2013; van der Boon et al., 2021; Zhang et al., 2013). While the geography and chronology of Eocene anoxia remains poorly constrained, it is important to note that open marine oxic environments prevailed in most of the Central Paratethys basins up to the Eocene-Oligocene Transition.

The Eocene paleogeographic episode was furthermore dominated by the deposition of *marine turbidites* in the central European basins. The OCB domain was characterised by good connections with the oceans, albeit a decline during the Middle Eocene (Bartonian) took place, when a deficiency in connectivity led to the formation of deep-water stratification and anoxia. This restriction may be linked with connectivity changes in the Eastern Paratethys region. After the Middle Eocene connectivity restriction, the basins recover their ventilation until the Early Oligocene. The late Eocene OCB turbidites are generally replaced by *"globigerina marl"* at the Eocene-Oligocene Transition. These marls appear to develop throughout Paratethys during the uppermost Eocene and may represent foraminiferal blooms due to increased nutrient-rich superficial waters flowing out from the anoxic basins.

3.2. Oligocene restriction and anoxia

During the latest Eocene (Terminal Eocene Cooling Event) and throughout the Eocene-Oligocene transition (Priabonian – early Rupelian), the OCB transformed from a carbonate factory into a poorly ventilated basin (bituminous marls) and ultimately switched into an anoxic basin (Fig. 10b). The *de facto* transition to anoxia occurred in the Oligocene – in the early Rupelian – corresponding to the calcareous nannofossil zone NP21, upper part (Melinte-Dobrinescu and Brustur, 2008), when the basin experienced severe restrictions. This Early Oligocene restriction correlates with the Solenovian "isolation" of the Eastern Paratethys (Rusu, 1977, 1999; Báldi and Báldi-Beke, 1985; Merklin, 1974;; Sachsenhofer et al., 2018). Similar restrictions have been documented in events throughout the West-Central Paratethys region (Soták, 2010). Restriction of the connections to the proto-Mediterranean region *via* the Slovenian basin was particularly important for the internal connections in Paratethys, given that the faunal assemblages in the OCB became dominated by boreal fauna elements (Soták, 2010), an event described as the abrupt and synchronous occurrence of the *Spiratella Sea* (Báldi, 1984).

Fig. 9. Upper Miocene facies of the foredeep fill and molasse formations. A. Sands and sandstones of the Șipoțelu fm.; b. Continental sands and paleosols of the Râmnic fm.; c. Maeotian lacustrine sediments; d. Selenite from Valea Vizuinei fm.; e. Sandstone concretions in the Râmnic fm.; f. Istrița limestone beds with limestones and oolitic sands; g. Intra-Maeotian slump with clay debris fill, capped by coarse bioclastic debris.

The restrictions in circulation and the increased stratification in the OCB were not constant throughout the Oligocene (Table 2). The initial restriction comprises three episodes, all of them in the Rupelian, with deposition of typical lithologies: (1) *Bituminous Marls* corresponding to the onset of anoxia, followed by (2) *Lower Menilites* – peaking in anoxia (Amadori et al., 2012; Belayouni et al., 2009; Puglisi et al., 2006), and (3) *Lower Dysodiles,* when the basin connectivity improved while the basin remained poorly ventilated. These anoxic deposits formed in the OCB consecutive with a freshening in Eastern Paratethys suggesting they might represent lithological expressions of exchange between the two basins in a context of limited connections with the ocean. As marine connectivity slightly improved, anoxic turbidites became the dominant lithology in the OCB. The Central European basins remained poorly ventilated until the Early Miocene (Fig. 11a), when sub-oxic episodes such as those found in the *Gura Șoimului Formation* (Amadori et al., 2012) indicate a new phase of vertical basin circulation.

3.3. Miocene restricted anomalohaline environments and evaporites

Miocene paleogeographic reorganizations created diverse

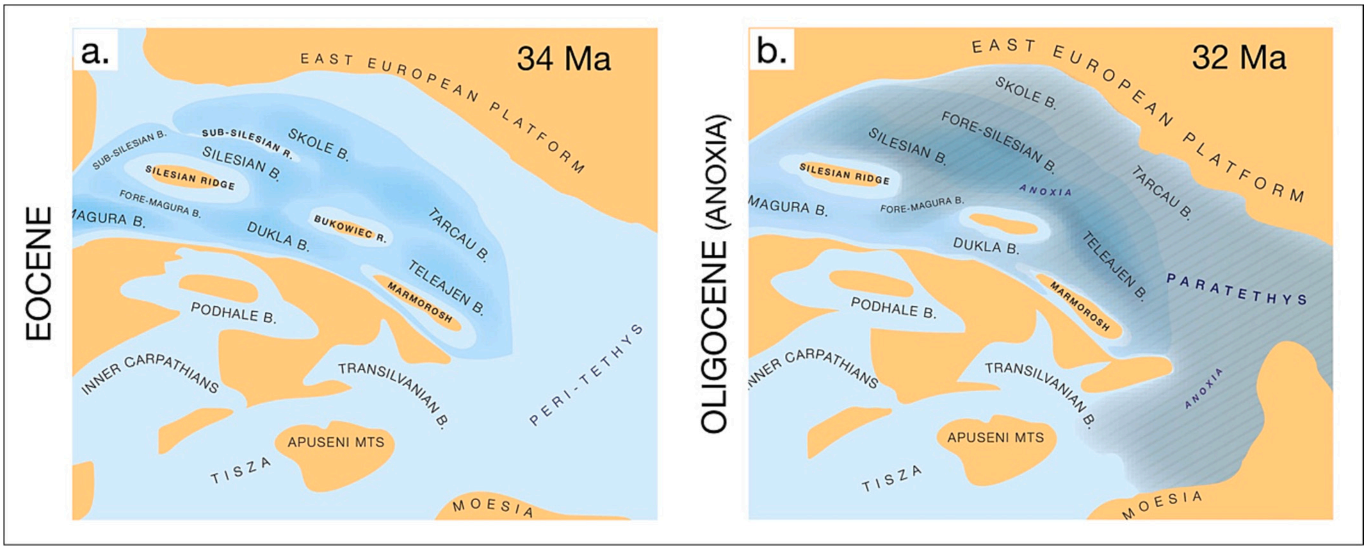

Fig. 10. Simplified paleogeographic sketches highlighting the evolution of the Outer Carpathian Basin during the Late Eocene (a) and Early Oligocene (b) (after Golonka, 2011; Golonka et al., 2007; Palcu and Krijgsman, 2021). Reconstructions indicating the main sub-basins and connections with nearby basins. Orange areas represent landmasses, blue areas represent marine basins (darker blue – deep basins). The brown shades indicate anoxic domains.

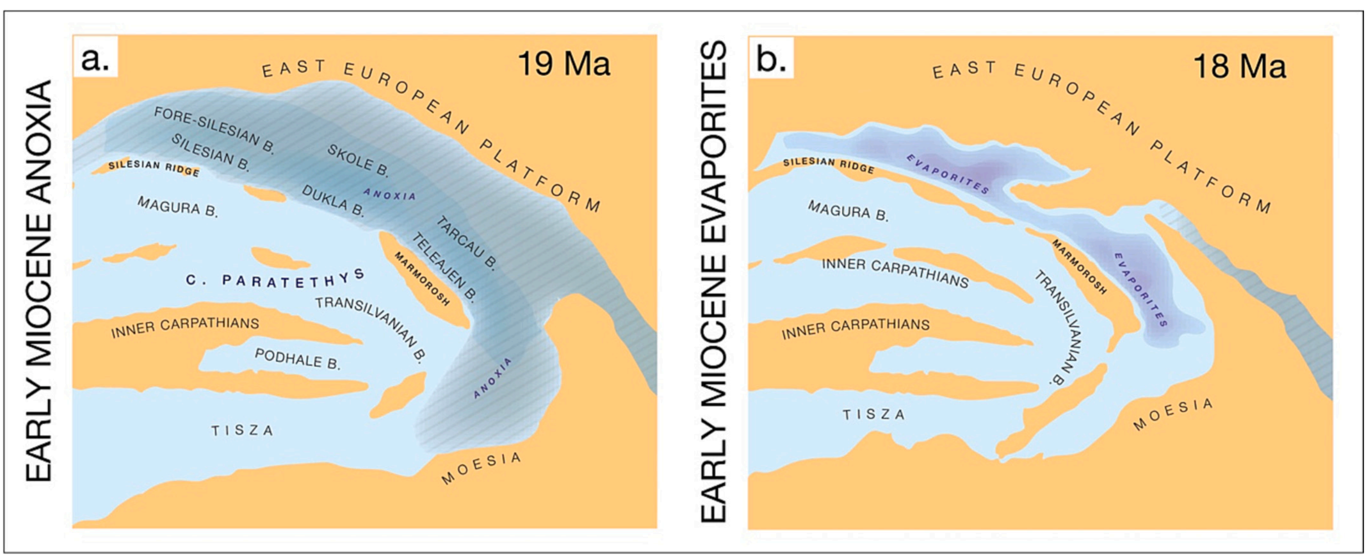

Fig. 11. Simplified paleogeographic sketches highlighting the evolution of the Outer Carpathian Basin during the Early Miocene (after Golonka, 2011; Golonka et al., 2007; Palcu and Krijgsman, 2021). Reconstructions indicating the main sub-basins and connections with nearby basins. Orange areas represent landmasses, blue areas represent marine basins. The brown shades indicate anoxic domains, and the purple shade shows the evaporitic domains.

environments dominated by *brackish and marine molasse, evaporites and carbonatic sediments*. Molasse deposits formed in the OCB at the beginning of the Early Miocene when marine connectivity deteriorated further. During the Burdigalian stage, the connectivity within the OCB was affected by the Old Styrian tectonic phase (Săndulescu, 1988). Intense eastward thrusting of the Outer Carpathian nappes led to uplift and the formation of thick olistoplaques at the front of the nappe thrusts, which likely affected intra-Paratethyan connectivity in the Polish-Ukrainian region. This tectonic episode led to restrictions in marine connectivity and the formation of evaporites in the Ukrainian and Romanian sectors of the OCB (Fig. 11b). Halite and gypsum precipitated in restricted pools during the earliest Burdigalian and late Burdigalian, as the basin became further isolated. During and after the evaporitic episodes, the western connections with the global ocean re-opened and a relatively well-distributed presence of marine faunistic and floristic elements was found and described from most of the Lower Miocene sedimentary record (Mărunţeanu, 1999).

Following the Burdigalian evaporitic episodes and the Old Styrian tectonic phase, the OCB was greatly reduced in size (Cheremisska and Cheremissky, 2019; Kováč et al., 2017a; Popov et al., 2006; Saulea et al., 1969) and characterised by shallow aquatic environments. In the Sub-carpathian area, sediments indicate a major continental contribution, while scarcity in fossils suggests that the environment was unfavourable to marine life (Mărunţeanu, 1999). Regional gypsum deposits indicate partial desiccation episodes. Frequent ichnologic evidence, traces of land mammals and birds (Brustur, 1997), suggest that large basin areas were covered by shallow to tidal aquatic environments.

Middle Miocene Tectonic activity in Central Europe (Fodor et al., 1999; Huismans et al., 2001; Kováč et al., 2004; Vakarcs et al., 1998), also known as the *Young Styrian tectonic phase in* the classical literature (e.g., Săndulescu, 1988), combined with a global sea-level high-stand led to the transgression of marine environments (Fig. 12a) on large

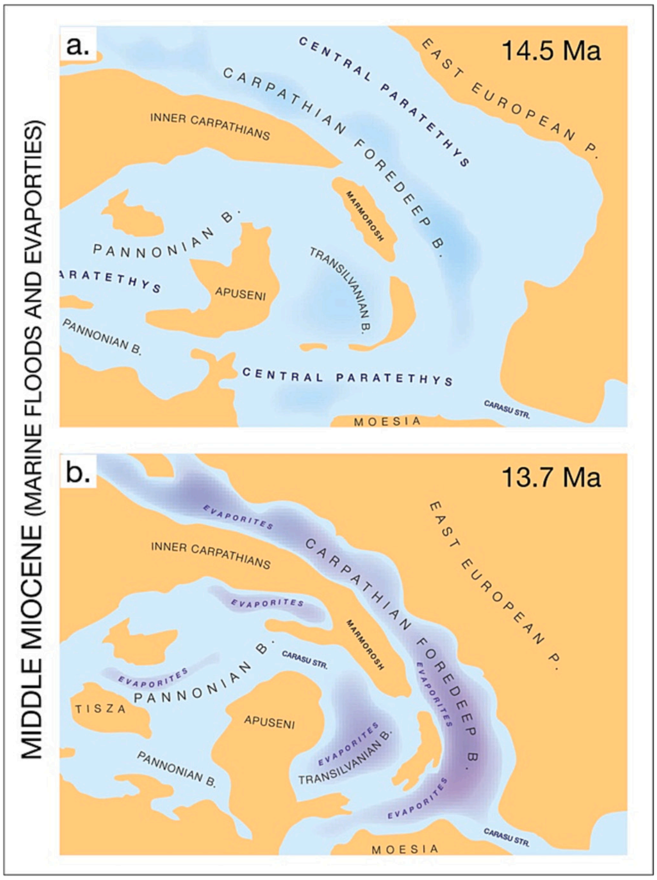

Fig. 12. Simplified paleogeographic sketches highlighting the evolution of the Outer Carpathian Basin during the middle Miocene (after Golonka, 2011; Golonka et al., 2007; Palcu and Krijgsman, 2021). Reconstructions indicating the main sub-basins and connections with nearby basins. Orange areas represent landmasses, blue areas represent marine basins and purple shades show the evaporitic domains.

swathes of Central Europe. This drastic paleoenvironmental change, that occurred at ~15 Ma (Sant et al., 2019b, 2020), during the Langhian (Badenian regional stage) is known as the Badenian flooding event (Harzhauser and Piller, 2007; Holcová et al., 2018; Kováč et al., 2007; Piller et al., 2007b) and has been documented throughout Central Paratethys, in Poland (Oszczypko and Oszczypko-Clowes, 2012), Ukraine (Nevesskaja et al., 2003), Czechia (Brzobohatý and Stráník, 2012; Hladilová et al., 2014; Rybár et al., 2016; Švábenická, 2002), Slovakia (Csibri et al., 2022), Hungary (Báldi et al., 2002), Romania (Beldean et al., 2013; Chira et al., 2000; Filipescu and Silye, 2008; Sant et al., 2019b, 2020), Croatia (Brlek et al., 2016; Ćorić et al., 2009; Sremac et al., 2016), Bosnia-Hercegovina (Mandic et al., 2019), Serbia (Rundic et al., 2014) and Bulgaria (Popov et al., 2006).

Before the transgression, a limited area of Central Europe was covered by the Karpatian Sea, a basin characterised by brackish-marine environments, spanning from Slovakia, Czechia, and Austria to Slovenia via rifts in the central Hungarian region (Kováč et al., 2016; Piller et al., 2007b; Sant et al., 2017). Thus, during the high stands that accompanied the Middle Miocene climatic optimum, the basins of Central Europe became better connected with each other and with the global ocean.

In the OCB, this transgressive event led to the expansion of the basin and increases in bathymetric regimes (Beldean et al., 2013). Former tidal zones turned to deep offshore areas (Bicchi et al., 2003; Sztanó, 1995) and the new marine realm transgressed on nearby lands (Holcová et al., 2018; Sant et al., 2019a). After the Middle Miocene transgressive and tectonic episode, the basin depocenters shifted eastwards (Popov et al., 2004) but also expanded on top of older nappes that were previously emerged (Bădescu, 2005). That is why Badenian and post-Badenian deposits are often referred to as *"post-tectogenetic cover"* (Haas, 2012). The Badenian flooding opened connections between the OCB and the Vienna Basin via a Polish-Czech strait (Palcu and Krijgsman, 2021), but the OCB was also connected with the Pannonian and Transylvanian basins via a system of straits fragmenting the Eastern Carpathians (Kováč et al., 2017a). A key connection to the east (Carasu Strait) allowed passage of marine waters to Eastern Paratethys. This triggered water exchanges between the "open marine" Central Paratethys and the "anoxic, anomalohaline" Eastern Paratethys, followed by the shutdown of the anoxic condition in the Eastern Paratethys sea (Palcu et al., 2019b).

Global eustatic changes led to a deterioration of marine conditions at the beginning of the Serravallian at ~13.8 Ma (de Leeuw et al., 2010), when the OCB experienced a second evaporitic crisis (Fig. 12b). This Middle Miocene event, known as the "Badenian Salinity Crisis" (de Leeuw et al., 2013; Harzhauser et al., 2018; Peryt, 2006) is spread throughout the OCB and beyond in the nearby Transylvanian, Maramureș and Pannonian basins to the west. The evaporitic facies did not spread in the regions west than of the Pannonian basin, e.g., Vienna basin (Harzhauser et al., 2018), suggesting that the chokepoint in marine connectivity must have been in the Pannonian region. Tectonic changes in the Pannonian basin ended the connectivity restrictions and terminated the evaporitic phase at ~13.4 Ma (de Leeuw et al., 2018; Peryt, 2006). After the evaporitic phase ceased, the basin recovered, becoming initially stratified and then fully marine during the middle Serravallian (Melinte-Dobrinescu and Stoica, 2014). During the Middle Miocene, the Eastern Paratethys region alternated between a sea poorly connected with the ocean via the OCB and a lake that discharged waters to the OCB (Palcu et al., 2017; Simon et al., 2019).

3.4. Late Miocene megalake environments

At 12.65 Ma the paleogeography of the entire Paratethys realm changed again due to reorganizations in the internal connectivity that allowed a mixing of the Eastern and Central Paratethys water masses while the connectivity with the global ocean decreased (Palcu et al., 2015, 2017). The improved connectivity between the Central and Eastern Paratethys was triggered by progressive down-flexing of the lithosphere in the area of southern Moldavia and SW Ukraine due to eastward subduction roll-back and coincident advance of the Carpathian wedge, which brought this formerly exposed area below sea-level (de Leeuw et al., 2020). This "unification" of the Paratethyan basins led to two simultaneous faunal crises in the region: (1) the Badenian-Sarmatian Extinction (Harzhauser and Piller, 2007; Palcu et al., 2015) event corresponding to the extinction of marine fauna in Central Paratethys while (2) the Konkian-Volhynian event (Palcu et al., 2017) led to significant faunal changes in Eastern Paratethys. In the aftermath of these two events, the unification of the Paratethys realm marks the beginning of the Sarmatian regional stage (Piller et al., 2007b), characterised by a thriving endemic fauna.

After almost a million years of effective internal connectivity (Fig. 13a), at the Serravallian-Tortonian boundary (11.6 Ma), Paratethys broke into a collage of basins again (ter Borgh et al., 2013) along the alignment of the East and South Carpathian Mountains. The smaller lake Pannon broke apart from Paratethys in the west of the Carpathians (Magyar et al., 2013; 2021) developing a rich "Pannonian" endemic fauna that evolved from the Sarmatian endemic faunas of Paratethys (Fig. 13b). After the fragmentation, Eastern Paratethys continued to function as a large unified aquatic realm characterised by endemic Sarmatian environments and faunas (Popov et al., 2004). Eastern Paratethys stretched from the Carpathian Mountains in the west to modern Kazakhstan in the east, incorporating the OCB (Popov et al., 2006, 2010). Both the *Pannonian* and *Eastern Paratethys* basins were deprived of significant connections with the global ocean, although short-lived, weak marine connections cannot be completely excluded. The rise of

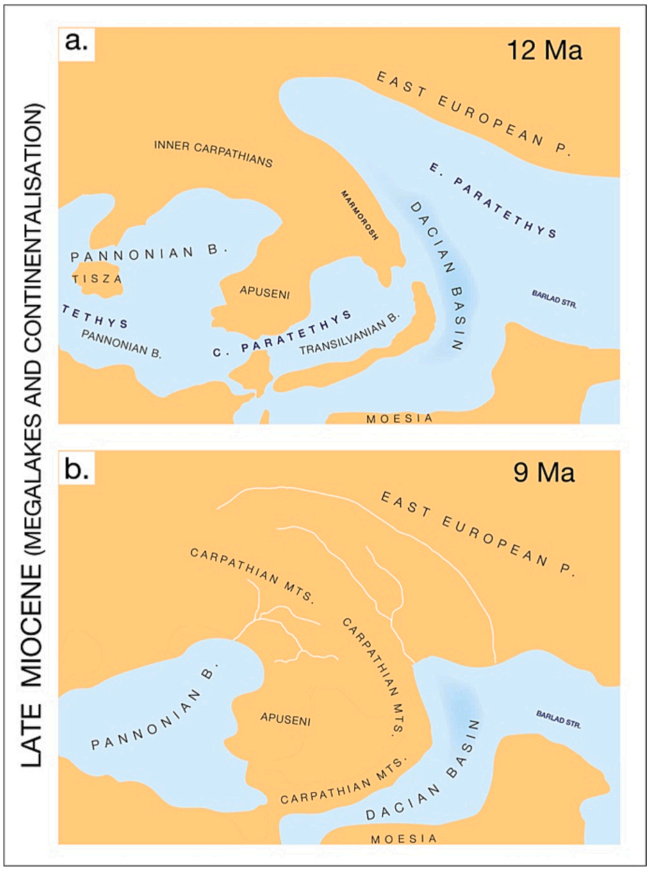

Fig. 13. Simplified paleogeographic sketches highlighting the evolution of the Outer Carpathian Basin during the late Miocene (after Golonka, 2011; Golonka et al., 2007; Palcu and Krijgsman, 2021). Reconstructions indicating the main sub-basins and connections with nearby basins. Orange areas represent landmasses, blue areas represent marine basins.

the Carpathian thrust-fold belt not only affected the western connection of the OCB with the Transylvanian and Pannonian basins but also provided large amounts of sediments that began to fill the depressions (Jipa, 2018; Matoshko et al., 2016). Gradually the sea retreated, initially from the Polish-Ukrainian sector in the NW and then along the Carpathians towards the SSE (de Leeuw et al., 2020).

Brackish and freshwater molasse developed in the OCB during the Late Miocene, when the western connections faded, and the Dacian Basin became a marginal basin of Eastern Paratethys. During the Tortonian-early Messinian, the remaining Paratethys realm experienced significant base-level fluctuations, expressing the climatic forcing of their watershed. In the second half of the Tortonian, it suffered several desiccation crises (Butiseacă et al., 2021). In the maximum desiccation episodes, such as the Great Khersonian Drying (Lazarev et al., 2020; Palcu et al., 2019b), the connectivity between Eastern Paratethys and the OCB was greatly reduced, and the latter experienced episodes of freshening possibly as a result of restriction of its connection with the Eastern Paratethys by the Balta Delta (Matoshko et al., 2016). This regressive trend was reversed in the late Tortonian due to a global cooling episode that changed the water balance of Paratethys to positive values. This change, in turn, triggered a gradual refill of Eastern Paratethys (7.65-7.4 Ma), known as the Maeotian transgression (Lazarev et al., 2020; Palcu et al., 2019b). During the early Messinian (6.9-6.7 Ma, Golovina et al., 2019), a discharge towards the nearby Mediterranean basin reinstalled a new connection with the global ocean through the Aegean region (Krijgsman et al., 2020). The Black Sea basin became linked with the Mediterranean realm, although the connection remained weak (van Baak et al., 2017; Floroiu et al., 2011; Grothe et al., 2016; Krijgsman et al., 2010; Stoica et al., 2013; Vasiliev et al., 2005, 2021). In Messinian and later stages, the Dacian Basin continued filling up with sediments while gradually losing its marine and brackish character (Van Baak et al., 2015; Olariu et al., 2018; Vasiliev et al., 2021).

4. Outlook, future perspectives & concluding remarks

This review shows that a restricted aquatic basin can develop heterogenous aquatic environments due to changes in marine connectivity. This raises the exciting opportunity of a combined approach to investigate these anomalohaline waterbodies in more detail and to better understand their past, present and future behaviour. Despite their broad occurrence (Fig. 14), bituminous marls are still poorly understood and often confused with anoxic, carbon-free black shales (bituminous shales). These marls can be markers of a particular connectivity window between marine basins and their anoxic neighbours. The mechanisms of marine connectivity associated with the formation of carbonate factories remain poorly understood but might be similar to changes in the connectivity of the modern Black Sea with the Mediterranean Sea via the Sea of Marmara. The Black Sea can become supplier of nutrient-rich waters to the nearby marine basins due to the onset of natural density pumps as a result of marine connectivity increases due to sea-level rise and anthropogenic activities. Similar density pumps are considered responsible for major paleoenvironmental cataclysms in the Paratethys realm such as the Badenian-Sarmatian Extinction Event (Palcu et al., 2015) or the demise of the Paratethys Anoxic Giant (Palcu et al., 2019a).

Large anoxic seas are important natural carbon sinks and in the case of long-lived anoxic giants the amount of carbon trapped in the anoxic sediments can be considerable. However, the impact of anoxic carbon sinks on global climate is still poorly understood. It is estimated that the anoxic sediments of the Eurasian Epicontinental Sea of Tethys captured between 0.72×10^{12} T and 1.3×10^{12} T (720-1300 GT) of organic sedimentary carbon released during the Paleocene-Eocene Thermal Maximum (PETM) (Kaya et al., 2022). A significantly larger amount of organic sedimentary carbon, around 60×10^{12} T (60,000 GT), is estimated to have been captured by the Paratethys anoxic giant during the latest Eocene-Oligocene, at a rate of $\sim6\times10^{12}$ T (6,000 GT) per Ma (Allen and Armstrong, 2008), that would correspond to ~12% of the estimated global organic carbon flux in the late Paleogene (Raymo, 1994). This flux, even though just a crude estimation, shows that the overall effect of the carbon drawdown of Paratethys would have suppressed atmospheric CO_2 levels during the EOT and in the Oligocene (Fig. 14). Further investigations are required to confirm these estimations and clarify the role of Paratethys in the Eocene-Oligocene Transition (EOT) greenhouse-icehouse transition.

The sediments of the Paratethys anoxic giant represent an important hydrocarbon source rock associated with some of the largest natural gas and oil reservoirs on the planet. The sea bottom of these anoxic seas can be rich reservoirs of methane, trapped under the seafloor in solid form as methane clathrates (gas hydrates). The modern Black Sea is the largest anoxic sea on Earth and contains a large reservoir of gas hydrates (Kruglyakova et al., 2004; Riboulot et al., 2018; Vassilev and Dimitrov, 2002) that might be at risk of being released from the sea-bottom due to global warming, sea-level rise and improved connectivity. Remarkably, the mechanism of this natural process remains poorly understood and awaits future modeling studies.

In the case of Paratethys evaporites, its Lower Miocene gypsum and halite deposits are relatively poorly studied and depositional models for the Middle Miocene evaporites differs from the well-known Mediterranean Messinian models (Andreetto et al., 2021; Cita, 1982; Flecker et al., 2015; Hernández-Molina et al., 2014; Hsü et al., 1973; Krijgsman and Meijer, 2008; Lugli et al., 2010; Müller and Mueller, 1991; Roveri et al., 2014, 2016; Selli, 1954). More studies are required to clarify and improve the understanding of Paratethyan evaporite deposits and their chronological models to allow direct comparison with the other evaporitic basins of the world.

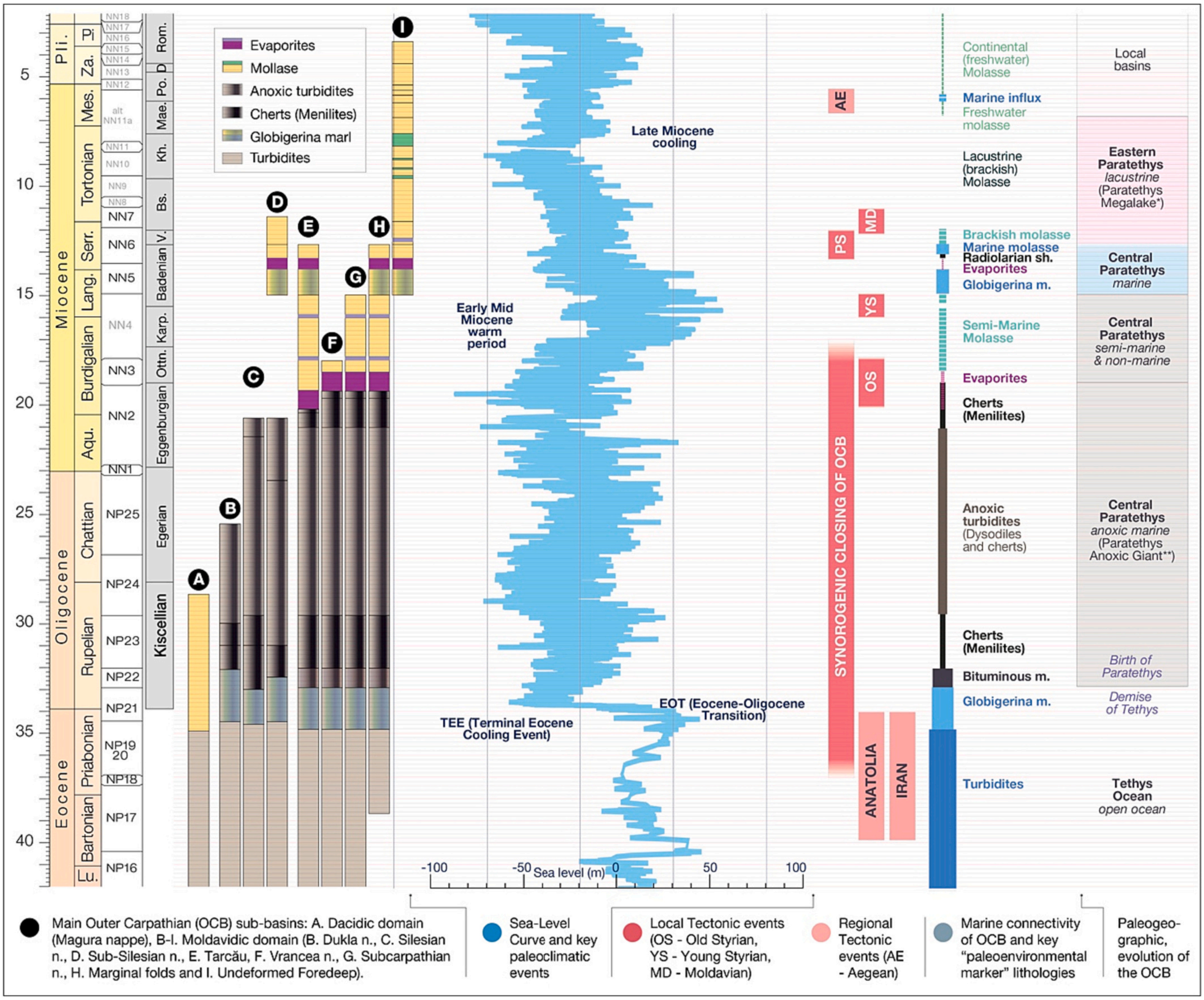

Fig. 14. Sedimentary records of the OCB sub-domains correlated with the eustatic curve (Miller et al., 2020), tectonic events, specific lithological features and qualitative estimations of marine connectivity trends based on the biostratigraphic data reviewed in this study. The paleogeographic evolution of the OCB can be resumed to several episodes, determined by the connectivity of the basin with the global ocean.

The paleoenvironmental change towards the endorheic Paratethys megalake resulted in the development of rich endemic floras and faunas, such as nannoplankton, malacofauna, microfauna (e.g., ostracods), vertebrate fauna (e.g., fish or marine mammals) (Čorič, 2005; Lukeneder et al., 2011). Part of this fauna survived and currently thrives in the Caspian Sea, the largest modern endorheic megalake (e.g., Krijgsman et al., 2019). The special faunas of these megalakes have been studied over more than a century but the exact mechanisms and the required conditions for the development and spread of the endemic species between isolated basins or after the basin isolation ceases.

Our review of Cenozoic paleoenvironmental change in the Central European basins highlights several key research directions and research questions to better understand what happens during and after the demise of oceans and seas. After the demise of the Tethys Ocean, long transition phases, in the order of tens of millions of years (more than 30 million years in the case of Paratethys), are required to shift the once marine realms to continental environments. These paleoenvironmental changes were mainly governed by the degree of marine connectivity with the global ocean. During episodes of increased marine connectivity, the basins were influenced by the global ocean reflecting global climate trends. As connectivity with the ocean dwindled, however, the environments became more influenced by the surrounding continental landmasses reflecting local climate trends. The transition from ocean to continental environments was characterised by a dominant regressive trend in the marine connectivity: alternating between gradual and sharp connectivity decrease episodes and short-lived connectivity increase trends. These changes were in turn governed by the interplay between tectonic processes and eustatic fluctuations.

Tectonic processes provided a direct influence on the basins by means of geometric deformations (subduction, nappe thrusting and back-arc extension) but also contributed indirectly to the increase of sediment supply that filled the basins. The main impact of tectonics, however, relates to marine connectivity, as uplift and sediment supplied by nearby orogens effectively choked the sea-straits. Tectonic activity also triggered "gateway opening" episodes, associated with shear zones, back-arc extension, or the flexure of foreland basin. Tectonic activity also triggered episodes of basin expansion, due to basin subsidence processes, opening of piggy-back basins and back-arc basins. Also, it influenced the formation and relocation of fore-bulge, back-bulge and foredeep zones affecting internal connectivity and inducing more complexity in reconstructing the evolution of former oceanic basins.

Detailed stratigraphic correlations show a resonance relation between eustatic fluctuations and tectonic events: sea-level drops occurring during the tectonic closure of basins and of marine gateways were associated with regional evaporitic crises while global sea-level rise episodes that correlated with the opening of back-arc basins and gateways triggered large “marine flood” events. During and after the demise of the Tethys Ocean, its former oceanic basins underwent exotic paleoenvironmental configurations that reflect specific degrees of connectivity with the ocean such as the carbonate factories, anoxic seas, evaporite basins and megalakes.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Data availability

No data was used for the research described in the article.

Acknowledgements

Funding: This publication is part of the project “Impact of sea-level rise on anoxic basins: Paratethys vs. Black Sea” [grant Veni.212.136] of the research Talent programmed – Veni which is financed by the Dutch Research Council (NWO) of DVP. DVP also acknowledges the Fundação de Amparo à Pesquisa do Estado de São Paulo (FAPESP) for financial support through grant 2018/20733-6. DVP, MMD, EA and MS also acknowledge Project PNRR C9 - I8 "Multiproxy reconstruction of Eurasian Megalakes, connectivity and isolation patterns during Neogene-Quaternary times", code 97/15.11.2022, Contract No. 760115/23.05.2023. WK acknowledges the Dutch Research Council (NWO) [grant 865.10.011].

Photo credits to *Arjan de Leeuw* (Figs.: 1a, 1e, 6a, 7b, 7c , 7d, 7e, 7f, Supp. 1a, Supp. 1b, Supp. 1c, Supp. 1e, 2 Supp. 1f,); *Isabela Maris* (Figs.: 1d, 6b, 6c, 9e, Supp. 1d,); *Dan V. Palcu* (Figs.: 1b, 1c, 1f, 6d, 6e, 7a, 7g, 8a, 8b, 8c, 8d, 8f, 8g, 9a, 9b, 9c, 9d, 9f, 9g); Wikimedia commons (Fig.: 8e).

Appendix A. Supplementary data

Supplementary data to this article can be found online at https://doi.org/10.1016/j.earscirev.2023.104594.

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