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Tectonic-sedimentary evolution of the Dukla Nappe, Ukrainian Carpathians

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The knowledge of the stratigraphy, tectonics and evolution of the Dukla Flysch Nappe of the Ukrainian Carpathians has been supplemented on the basis of geomapping works, structural, sedimentological and micropaleontological studies. Four foraminiferal assemblages are distinguished in the studied sediments. The assemblages (1) and (2) containing Deep-Water Agglutinated Foraminifera suggested bathyal-abyssal depths. Assemblages (3) and (4) are characterized by planktonic and calcareous benthic foraminifera. The assemblage (3) indicates bathyal depths, and the assemblage (4) – upper bathyal–littoral depths. The following formations as indicators of the accretionary prism evolution were identified.

Pre-orogenic formation of the remnant flysch Carpathian Basin contains the sediments that accumulated between the active margins of the microcontinental terrane and the passive margin of Eurasia. It includes the Cretaceous–Eocene flysch containing foraminiferal assemblages (1) and (2). The pre-orogenic formation contains conglomerates with exotic clasts suggesting the uplifts interpreted as fore-bulges migrated towards foreland. Synorogenic formation is represented by latest Eocene–Oligocene deposits belonging to 'piggy-back' and trench-like facies. The 'Globigerina Marl' (Eocene/Oligocene transition) contain assemblages (3), and the Oligocene sediments (Turytsa and Dusyno subnappes of the Dukla Nappe) contain assemblages (4). The change from bathyal-abyssal depths to upper bathyal–lower littoral depths, which began at the turn of the Eocene and Oligocene, could be caused by syn-sedimentary tectonic movements. The diachronic 'younging' of the trench-like coarse-clastic facies suggests the migration of the trench. Post-orogenic formation was accumulated in the wedge-top basins on the Dukla Nappe. The identified formations indicate the accretionary prism growing.

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Introduction

A necessary part of identifying the oil and gas prospects in the region is the prediction of the distribution of potentially oil and gas bearing and generating strata. This necessitates the analysis of their geological/tectonic position, stratigraphy, lithofacies and sedimentary environments. In the Ukrainian Carpathians, potentially oil and gas-generating sediments, including black shales enriched in organic matter, are developed, mainly among the Lower Cretaceous (Spas and Shypot formations) and Oligocene (Menilite Formation) strata. The characteristics of this strata, as possible sources of hydrocarbons, are given in the works of (Kotarba et al., 2009; Sachsenhofer, Koltun, 2012; Krupskyi et al., 2014).

However in the Carpathians, black shales are known not only in the Oligocene and Lower Cretaceous, they are also present in the Upper Cretaceous and Paleocene-Eocene deposits in some structural-facies units (Vialov, 1981). In particular, in the Dukla Nappe, the black shales are developed both in the Lower and Upper Cretaceous formations, as well as in the Oligocene strata (Vialov, 1981; Vialov et al., 1988). Therefore, the analysis of the structure and development of the Dukla Nappe is of practical importance.

The Dukla Nappe was studied in detail by (Danysh, 1973), who developed the foundations of lithostratigraphy and deciphered the main features of its geological structure and evolution. This researcher discovered an interesting regularity – the migration to the northeast of the «cordilleras» (intra-basin ridges) growing, which occurred during the Cretaceous–Paleogene evolution of the Dukla basin. Later, stratigraphic schemes were proposed (Andreeva-Gryhorovych et al., 1985, Vialov et al., 1989) and stratotype sections of the Ukrainian Carpathian sediments were characterized (Vialov et al., 1988), which remain relevant even today.

In recent years, the evolution of the western segment of the Ukrainian Carpathians including

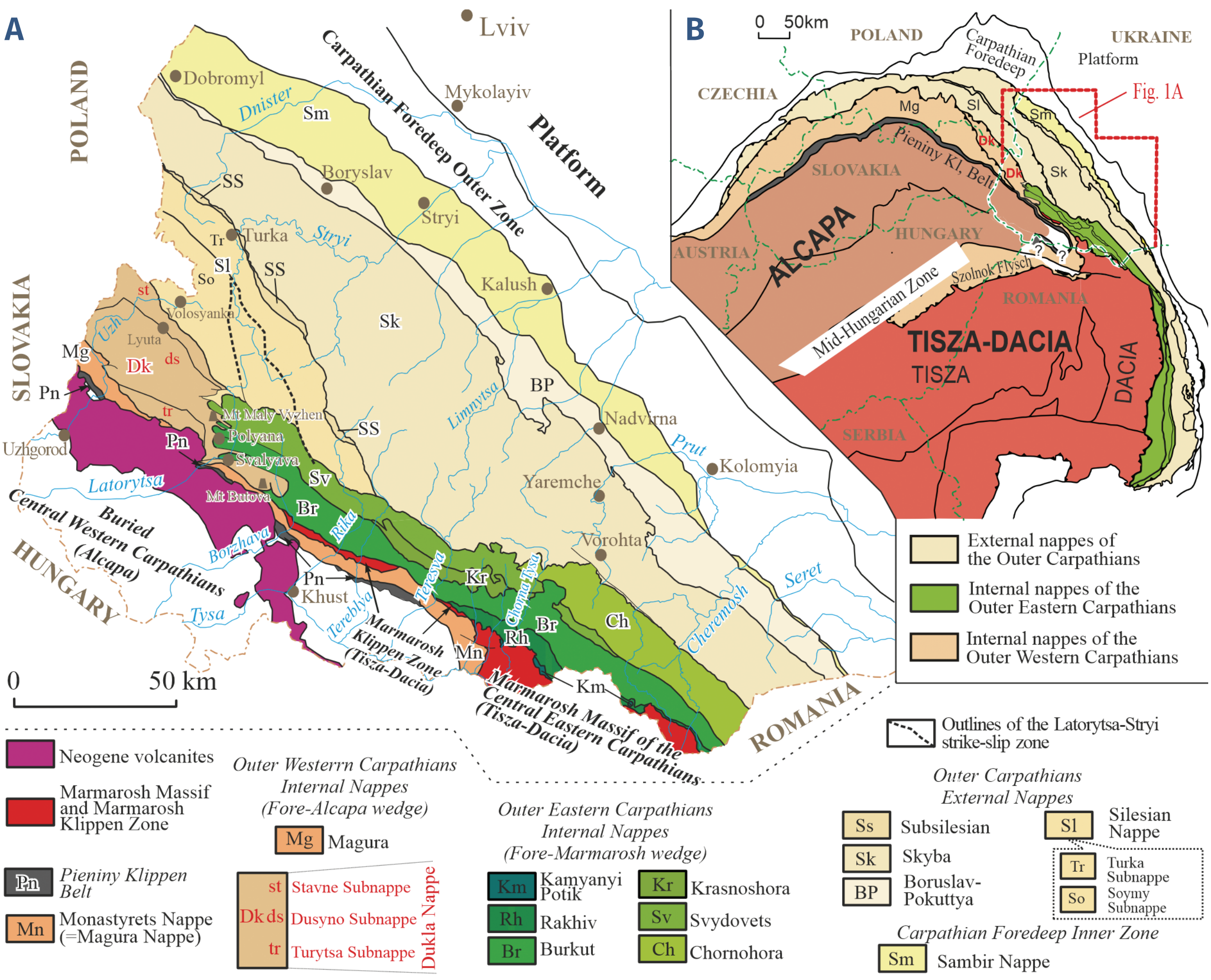

Fig. 1. Tectonic scheme *(A)* and geological position *(B)* of the Ukrainian Carpathians (after Hnylko et al., 2021, 2023 modified)

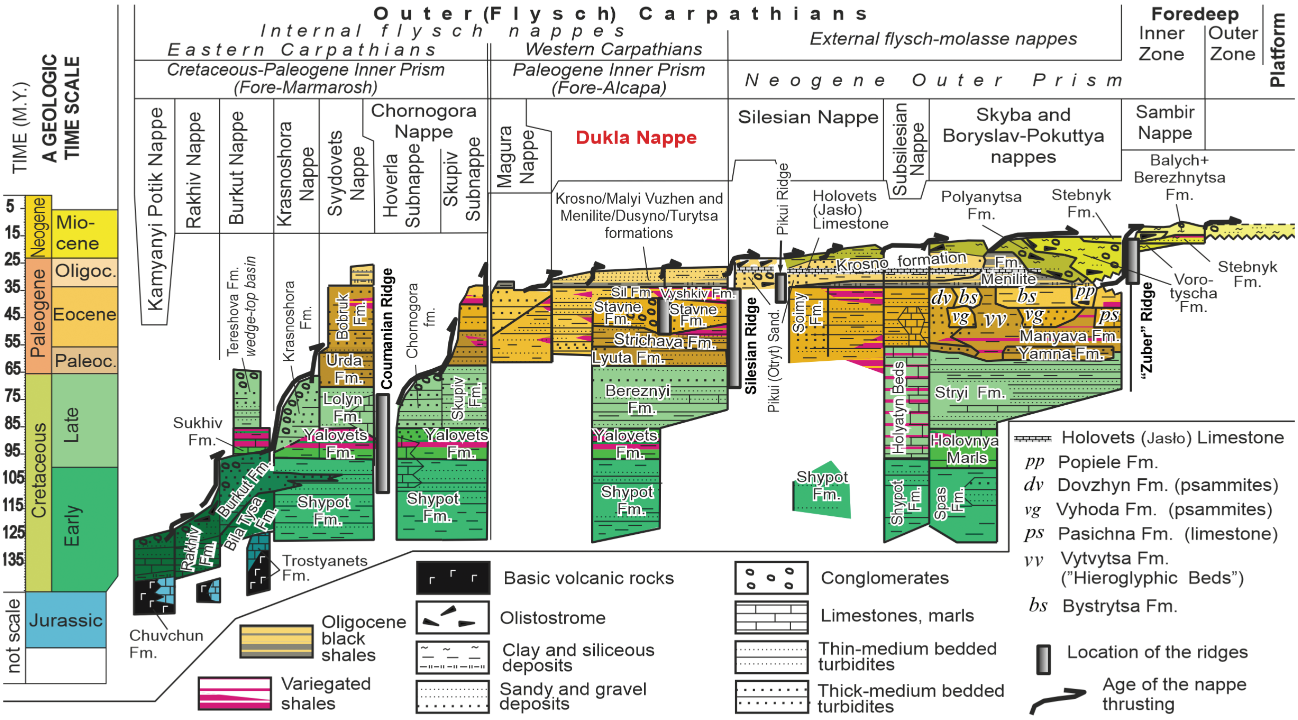

Fig. 2. Lithostratigraphic scheme of the Meso-Cenozoic of the Ukrainian Outer Carpathians (after Hnylko et al., 2023 modified)

Dukla Nappe has been considered, in terms of the accretionary prism development (Hnylko et al., 2015). The stratigraphy of the Dukla sedimentary succession was generalized, and some sedimentary processes and paleoenvironments were reconstructed on the base of the geological mapping, stratigraphic, sedimentological and structural investigations, as well foraminiferal analysis (Hnylko et al., 2015, 2020 and references therein) (Figs. 1, 2). The stratigraphy and sedimentary paleoenvironments of the black-shale formations located in the Ukrainian Carpathians, including the Dukla Nappe, were also supplemented (Hnylko et al., 2020, 2021 and references therein).

However, the creation of a complete and detailed picture of the tectonic-sedimentary evolution of the Dukla Nappe, the restoration of the sedimentary basin depths and the position of sedimentary complexes in the ancient accretionary prism, which will make it possible to predict the distribution of the potentially oil-and-gas-bearing complexes, remains an actual problem.

The purpose of the presented work is to supplement the knowledge of stratigraphy, geological structure and evolution of the Dukla Nappe in the Ukrainian Carpathians, and to reconstruct the sedimentary basin depths and the position of the Dukla sedimentary complexes in the ancient accretionary prism.

Method

Biostratigraphic, sedimentological and structural studies, as well as foraminiferal analysis were used. The material is both own field/laboratory data and literary sources. Lithostratigraphic subdivisions (formations, horizons) are represented mainly in accordance with stratigraphic schemes (Danysh, 1973; Andreeva-Grihorovych et al., 1985; Vialov et al., 1989; Gozhyk, 2013) and descriptions of stratotype sections (Vialov et al., 1988). Biostratigraphic and paleoecological studies are based mainly on the fauna of small foraminifera. The age of the formation is chiefly according to (Maslakova, 1955, 1965, 1966; Byzova et al., 1966; Glushko, Kruglov, 1971; Rozumeyko, 1980; Vialov 1981; Vialov et al., 1988, 1989; Andreeva-Grihorovych et al., 1985; Gozhyk, 2013; Hnylko et al., 2020, 2023). In the present work, only some of the most important species for determining age are given. For paleobathymetric reconstructions, the distribution of foraminifera in (hemi)pelagic sediments was analyzed based on both literature and own data. The bathymetry reconstructions of sedimentary paleobasins on the basis of small foraminifera was carried out in accordance with methods (Krasheninnikov, 1974; Murrey, 1976, 1991; Ivanik, Maslun, 1977; Gradstein, Berggren, 1981; Kaminski & Gradstein, 2005; Oszczypko et al., 2006).

Detailed geological mapping was carried out to identify the structure of some key areas including olistostrome/melange zones. The method of reconstruction of accretionary prism was used based to identify formations as indicators of the stages of its development, in particular the pre-orogenic, synorogenic and post-orogenic formations (Einsele, 1992; Plasienka, 2019) including wedge-top sediments (Artoni, 2013 and references therein).

Geological position

The Dukla Nappe is a large tectonic unit of the Outer (Flysch) Carpathians. The Outer Carpathians are made up of a several nappes thrusted over each other towards the north-east and over the Miocene Carpathian Foredeep. These nappes consist of allochthonous Late Jurassic–Neogene, mainly flysch sediments uprooted from their original substratum (Vialov, 1981; Kruglov et al., 1985, Kruglov, 1986; Golonka et al., 2006, 2019; Oszczypko, 2006; Hnylko et al., 2015, 2021 and references therein). The Jurassic-Early Cretaceous basic volcanic rocks of both oceanic and continental origin (Lyashkevich et al., 1995) are locally presended here in the base of some flysch successions as a remnants of the original substratum (see Fig. 2) (Hnylko, Heneralova, 2014; Krobicki et al., 2014 and references therein). The Outer Carpathian nappes were formed as the composed accretionary prism in the front of the both Alcapa and Tisza-Dacia (micro)continental terraines (e.g. Golonka et al., 2006, 2019; Oszczypko, 2006; Csontos, Vörös, 2004; Schmid et al., 2008, 2020; Hnylko et al., 2015, 2021; Kováč et al., 2016, 2017 and references therein).

The Outer Carpathian composed accretionary prism can be subdivided into three nappe systems (accretionary palaeoprisms). The first of them (Fore-Alcapa prism) was built at the front of the Alcapa Terrane. It includes such typical Western Carpathian units as the Magura and Dukla nappes. The second one (Fore-Tisza-Dacia prism) was formed at the front of the Tisza-Dacia Terrane. It contains the Kamyanyi Potik, Rakhiv, Burkut, Krasnoshora, Svydovets and Chornohora nappes located only in the Eastern Carpathians. Therefore from the geological points of view, the boundary between the Eastern and Western Outer Carpathians are located between the first and second nappe systems (Horvath, Galacz, 2006; Hnylko et al., 2015, 2021) (see Fig. 1). The third system with Silesian, Sybsilesian, Skole-Skyba, Boryslav-Pokuttya and Sambir nappes in Poland and Ukraine, and their prolongations in Romania were formed ahead of both already collided Alcapa Terrane (+Fore-Alcapa prism) and Tisza-Dacia Terrane (+Fore-Tisza-Dacia prism) (Csontos and Vörös, 2004; Hnylko et al., 2015, 2021; Kováč et al., 2016 and references therein). The Silesian, Syb-Silesian, Skole-Skyba nappe of the third system are located both in the Western and Eastern Outer Carpathians (see Fig. 1).

The Dukla Nappe (Unit) belongs to the Outer Western Carpathians. It is developed within the borders of Ukraine, Poland and Slovakia and located between such large units of the Outer Western Carpathians as the Magura and Silesian nappes. Dukla Nappe borders with the units of the Outer Eastern Carpathians such as the Burkut and Svydovets nappes in the Borzhava and Latorytsa river basins (see Fig. 1). In the Ukrainian Carpathians, the Dukla Nappe is divided (from south to north) into the Turytsa, Dusyno and Stavne subnappes (Vialov, 1981), which are somewhat different from each other in their sedimentary filling, especially in Paleogene flysch lithofacies.

In the Cretaceous–Paleogene time, the Dukla sedimentary basin was part of the Outer Carpathian remnant flysch basin located between the passive margin of Eurasia (the Eastern and Western European platform) and the active margins of the microcontinents situated in the Tethys Ocean. These microcontinents are known as the Alcapa and Tisza-Dacia terranes (now exposed as the crystalline massifs including the Central Western Carpathians and the Central Eastern Carpathians accordingly). Forming the Dukla Nappe was related to gradual closuring of the Outer Carpathian remnant flysch basin as a result of subduction of this basin (sub)oceanic basement beneath the Alcapa Terrane and the growing an accretionary prism in the front of this terrane. The accretionary prism was initially formed by the Pieniny Klippen Belt in the Late Cretaceous/Paleocene. Subsequently, the Magura and Dukla nappes were added to the prism in the Paleogene. Finally, the Silesian, Subsilesian, Skyba, Boryslav-Pokuttya and Sambir nappes were attached to the prism in the Early Miocene. The nappe pile was thrusted onto the Carpathian Foredeep in the Miocene (Golonka et al., 2006, 2019; Oszczypko, 2006; Hnylko et al., 2015, 2021; Kováč et al., 2016, 2017 and references therein) (see also Figs. 1 and 2).

Stratigraphic and sedimentological features of the Dukla Nappe sediments

In the Ukrainian Carpathians, the Dukla Nappe is filled with the Lower Cretaceous-Oligocene flysch deposits, which create a continuous (without angular and stratigraphic unconformities) sedimentary succession tectonically detached from its sedimentary basement (Fig. 2, 3). The age of the formations belonging to succession is given according to (Byzova et al., 1966; Maslakova, 1965, 1966; Rozumeyko, 1980; Andreeva-Gryhorovych et al., 1985; Vialov et al., 1988, 1989; Danysh, Ponomaryova, 1989; Romaniv, 1991; Gozhyk, 2013; Hnylko et al., 2020, 2023).

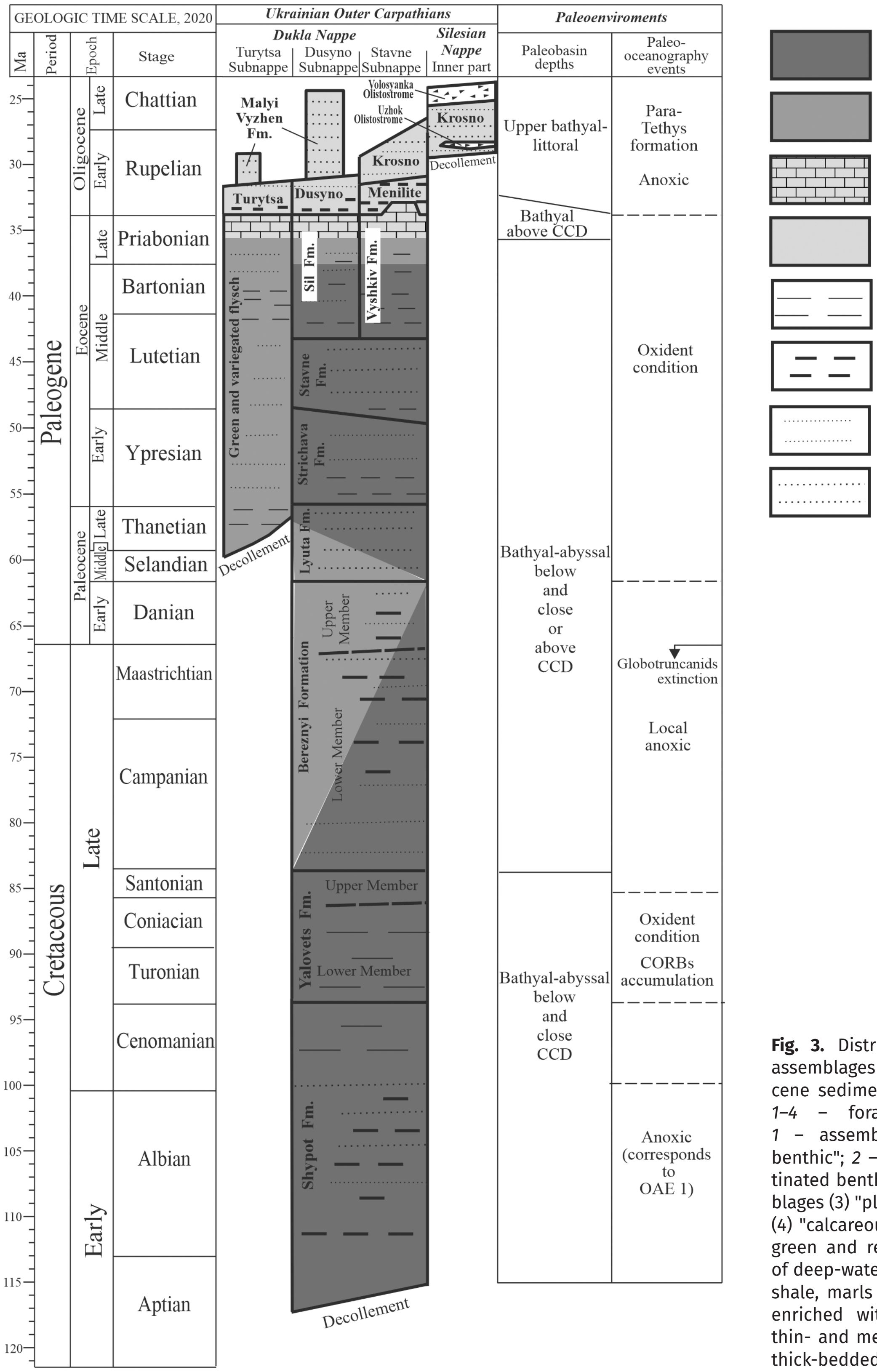

Fig. 3. Distribution of foraminiferal assemblages in the Cretaceous–Oligocene sediments of the Dukla Nappe. *1–4* – foraminiferal assemblages: *1* – assemblages (1) "agglutinated benthic"; *2* – assemblages (2) "agglutinated benthic-plankton"; *3* – assemblages (3) "plankton"; *4* – assemblages (4) "calcareous benthic-plankton"; *5* – green and red mudstone (analogues of deep-water pelagic clays); *6* – black shale, marls and siliceous sediments enriched with organic carbon; *7* – thin- and medium bedded flysch, *8* – thick-bedded flysch and sandstone

The Aptian–Cenomanian (Shypot Formation) is extended in the most tectonic units of the Ukrainian Carpathians (see Fig. 2). In the Dukla Nappe, it is represented by thin- and medium bedded rhythmic flysch composed mainly of quartz 'glassy' fine-grained sandstones and siltstones interlayered with the black shales enriched in organic matter, less with the grey-greenish shales. They are characterized by Bauma's turbidite intervals. Green mudstones of hemipelagic origin (3.5 m thick) lie in the top of the Shypot Formation. The total thickness of the Shypot Formation is 100 m in the Dukla Nappe. The age of the Shypot Formation is substantiated in detail for the deposits of the Chornohora Nappe by Aptian, Albian, and Cenomanian foraminifera (Byzova et al., 1966; Dabagian, 1969; Vialov et al., 1988).

Early Cretaceous foraminifera *Pseudobolivina variabilis* Vasicek, *Verneuilina neocomiensis* Mjatliuk as well as characteristic for the Albian species *Thalmanninella ticinensis* (Gandolfi), for Albian–Senomanian – *Plectorecurvoides alternans* Noth, *Trochammina normalis* Tairov, *Hedbergella planispira* (Tappan), and for Senomanian – *Bigenerina elongata* Tairov, *Spiroplectammina ammovitrea* Tappan were found in the Shypot Formation located in the Dukla Nappe (Maslakova, 1966; Rozumeyko, 1980). Along the Pestsiy Stream (left tributarity of the Lyuta River in the Uzh River Basin near the village of Lyuta, Transcarpathian Region, Dukla Nappe (see Fig. 1), green mudstone in the top of the Shypot Formation contain the foraminiferal assemblage, the Cenomanian age of which was identified by the joint occurrence of species *Plectorecurvoides alternans* Noth and *Dorothia crassa* (Marsson) (Vialov et al., 1988).

The Turonian–Santonian (Yalovets Formation) is developed in the most tectonic units of the Ukrainian Carpathians (see Fig. 2) and expressed by mainly red and variegated mudstone shales of (hemi) pelagic origin, which filling the lower member (up to 60 m) of the Yalovets Formation as well as the greenish-gray mudstones, siltstones and sandstones filling the upper member (60–70 m) of the Yalovets Formation. The lower member is characterized by a foraminiferal assemblage with the characteristic species Uvigerinammina jankoi Majzon, the age of the lower member is the Turonian–Coniacian (Maslakova, 1966; Hnylko et al., 2023). The first appearance of *Rzehakina epigona* (Rzehak) and *Caudammina gigantea* (Geroch) was noted in the upper member of the Yalovets Formation (Dukla Nappe, Lyuta and Zhdenivka river basins, Transcarpathian region), indicating the Santonian age of the sediments (Maslakova, 1966).

The Campanian–Lower Paleocene (Bereznyi Formation) distributed in the middle (Dusyno Subnappe) and outer (Stavne Subnappe) parts of the Dukla Nappe. It is composed of a typical flysch – an alternation of sandy turbidites and clay-marl hemipelagites. In the lower part of the Bereznyi Formation, the flysch is thin- to medium bedde; and in its upper part, it is thick-bedded. The presence of dark gray to black mudstones, shales and marls enriched with organic matter is characteristic. The thickmess of the Bereznyi Formation reaches 1000 m.

The characteristic Campanian–Maastrichtian *Globotruncana arca* (Cushman), *Globotruncana linneana* (Orbigny), *Rugoglobigerina kelleri* Subbotina occurred in the lower part of the Bereznyi Formation (Lyuta and Zhdenivka river basins, Dukla Nappe, Transcarpathian region) (Maslakova, 1966). Early Paleocene *Parasubbotina pseudobulloides* (Plummer), *Globanomalina compressa* (Plummer), *Praemurica inconstans* (Subbotina), *Globoconusa* cf. *daubjergensis* (Bronnimann) were identified in the upper part of the Bereznyi Formation (along the Uzh River near the village of Kostryno, Dukla Nappe, Transcarpathian region) (Vialov et al., 1988).

The Paleocene (Lyuta Formation) is extended in the middle (Dusyno Subnappe) and outer (Stavne Subnappe) parts of the Dukla Nappe. It is composed of massive and thick-layered polymictic gray sandstones (thickness 100–400 m). This is the deposits of high-density turbidite and grain flows. In this deposits, Paleocene foraminifers were identified, including *Nummulites frassi* Harpe (determination of Ya.V. Sovchyk) (see Danysh, 1973) as well as *Morozovella angulata* (White), *Globanomalina compressa* (Plummer), *Subbotina triloculinoides* (Plummer), *Anomalinoides danicus* (Brotzen), *Cibicidoides padellus* (Jennings), *Stensioina caucasica* (Subbotina), *Haplophragmoides mjatliukae* Maslakova (Vialov et al., 1988; Danysh, Ponomaryova, 1989).

The Paleocene–Eocene deposits distributed in the inner part of the Dukla Nappe (Turytsa Subnappe) are represented by green and variegated flysch, up to 750 m thick. This flysch is more sandy in the lower part of the sedimentary succession. It contains horizons of the red mudstones at different levels (Vialov, 1981). In general, these deposits are products of turbidite flows activity and hemipelagic clay sedimentation. They are characterized by the Late Zelandian–Thanetian *Globanomalina pseudomenardii* (Bolli), Ypresian *Morozovella mar-*

ginodentata (Subbotina) and Middle–Late Eocene *Acarinina bullbrooki* (Bolli), *Reticulophragmium amplectens* (Grzybowski) and *Ammodiscus latus* (Grzybowski) (Danysh, Ponomaryova, 1989).

The Lower Eocene (Strichava Formation) extended in the middle (Dusyno Subnappe) and outer (Stavne Subnappe) parts of the Dukla Nappe is expressed by medium-bedded greenish-gray flysch (thickness 350 m) – mainly medium-grained turbidites. Early Eocene foraminifera *Nummulites planulatus* Lamark, *Nummulites burdigalensis* Harpe (Golev, 1982), *Morozovella aragonensis* (Nuttall) and *Recurvoides smugarensis* Mjatliuk (Danysh, Ponomaryova, 1989) were found in the Strichava Formation.

The top of the Lower Eocene – Middle Eocene, Lutetian (Stavne Formation) is distributed in the middle (Dusyno Subnappe) and outer (Stavne Subnappe) parts of the Dukla Nappe and represented by mainly sandy flysch, sandstones, and coarse-clastic rocks, including gravelites and conglomerates in some places. The thickness of the Stavne Formation is up to 400 m. The flysch contains mainly turbidites. Coarse-clastic deposits consist of chiefly debris-flow products with exotic clasts including metamorphic rocks tending to the boundary between the Dusyno and Stavne Subnappes and, obviously, suggest the source area (paleoridge) buried between these subnappes. Characteristic foraminifera of the Late Ypresian–Lutetian *Morozovella aragonensis* (Nuttall), *Acarinina bullbrooki* (Bolli), *Turborotalia frontosa* (Subbotina), *Acarinina pentacamerata* (Subbotina) were found in the Stavne Formation (Danysh, Ponomaryova, 1989). Middle Eocene *Nummulites gallensis* Heim was identified in the middle part of the Stavne Formation (Vialov et al., 1988).

The Middle–Upper Eocene is composed of both thin-bedded green flysch of the **Sil Formation** (thickness 200–300 m) developed within the Dusyno Subnappe and thin-bedded variegated flysch, in places red shales of the **Vyshkiv Formation** (thickness up to 400 m) widespread within the Stavne Subnappe. The flysch is mainly composed of fine- and medium-grained turbidites. Middle–Late Eocene foraminifera *Reticulophragmium amplectens* (Grzybowski), *Ammodiscus latus* (Grzybowski), *Karreriella bartonica* Finlay characterize the Vyshkiv and Sil formations (Vialov et al., 1988). The Globigerinatheka tropicalis Zone (Lower Priabonian) was identified in the upper part of the Vyshkiv Formation in the sedimentary succession along the Vyshka River – the left tributary of the Uzh River (Dukla Nappe, Transcarpathian region) (Andreeva-Grygorovych et al., 1987).

The transitional Eocene/Oligocene deposits are represented by the **Sheshor horizon,** composed predominantly of gray hemipelagic marls (up to 10–20 m thick) enriched with planktonic foraminifera. These deposits known as **'Globigerina Marl'** are regionally distributed in the Carpathian region. Foraminifera of the Subbotina corpulenta Zone and nanoplankton of the Coccolithus subdistichus Zone and Helicosphaera reticulata Zones (upper Priabonian–Lower Rupelian) were identified in the "Globigerina Marl" in the sedimentary succession along the Vyshka River – the left tributary of the Uzh River (Dukla Nappe, Transcarpathian region) (Andreeva-Gryhorovych, 1987; Andreeva-Grygorovych et al., 1987).

The Oligocene in the lower part of its sedimentary succession in the Dukla Nappe is represented (see Fig. 2) either by black shales with sandstone intercalations (**Menilite Formation,** 20–80 m thick), or by dark to black marls with sandstone interbeds (**Dusyno Formation**, up to 300–350 m thick), or interlayering of quartzite-like fine-grained sandstones and dark gray and black mudstones and marls (**Turytsa Formation**, up to 700 m thick). The upper part of the Oligocene succession is composed of either gray flysch (**Krosno Formation**, up to 1000 m thick), or thick-bedded sandstones with clasts (up to 1 cm) of black mudstones (**Malyi Vyzhen Formation**, up to 300 m thick). Oligocene sediments complete the stratigraphic succession of the Dukla Nappe. Early Rüppelian foraminifera *Globigerina officinalis* Subbotina, *Subbotina vialovi* Mjatliuk, *Cibicidoides lopjanicus* Mjatliuk, *Bolivina aenariensiformis* Mjatliuk, *Bulimina elongata* Orbigny (Vialov et al., 1988; Danysh, Ponomaryova, 1989) characterize the lower part of the Dukla Oligocene succession. Rüpelian nanoplankton of the NP22-Helicosphaera reticulate Zone was found in the Dusyno Formation near the village of Dusyno and in the Uklinskyi Stream (Dukla Nappe, Transcarpathian Region) (Romaniv, 1991). Chattian nanoplankton of the NP25-Sphenolithus ciperoensis Zone was identified in the Malyi Vyzhen Formation in the Uklinskyi Stream (Transcarpathian Region) (Romaniv, 1991).

The Oligocene deposits completely make up the most inner **Volosyanka thrust-sheet of the Silesian Nappe,** which developed ahead of the Dukla Nappe front (Hnylko et al., 2021). The sedimentation of these deposits is related to the thrusting processes of the Dukla Nappe, therefore we briefly consider their stratigraphy. The deposits of the Volosyanka thrust-sheet is represented by the Oligocene Krosno Formation and Volosyanka Olistostrome. This olistostrome completes the stratigraphic succession and is overlapped by the Dukla Nappe (Hnylko et al., 2021).

The Krosno Formation is subdivided into three members here. The lower member composed of thin- to medium bedded gray flysch (with a thickness of 400 m). The middle member is expressed by the Pikui (Otryt on the neighboring Polish territory) Sandstone (up to 1000 m thick), at the base of which the Uzhok Olistostrome (up to 60 m thick) lies. The Uzhok Olistostrome contains clasts of exotic metamorphic rocks and shallow bioclastic limestones. The upper member of the Krosno Formation is represented by a thin- to medium bedded gray flysch (thickness up to 700 m).

The age of the Uzhok Olistostrome matrix is compared with the interval of standard foraminiferal O3–O5 zones of the upper Rüpelian–lower Chattian (middle part of the Oligocene) (Hnylko et al., 2021). The upper member of the Krosno Formation contains Upper Oligocene thin-laminated limestone Holovets Horizon (Jaslo Limestone in Poland).

The Krosno Formation is overlain by a thick (up to 1000 m) Volosyanka Olistostrome with flysch olistoliths probably slided from the uplifted front of the moving Dukla Nappe (Hnylko et al., 2021). In addition, the olistolith of red marls with Paleocene planktonic foraminifera of the Globanomalina pseudomenardii Zone was found in the Volosyanka Olistostrome (Hnylko et al., 2021 and references therein). This olistolith was apparently formed during the denudation of the thrust-sheet (with red marls) tectonically uprooted from the paleouplift between the Dukla and Silesian units during the thrusting of the Dukla Nappe onto the Silesian basin.

It should be noted that in the described Oligocene sediments, clearly expressed Bouma turbidite intervals were practically not observed by us (in contrast to typical turbidite textures in the Cretaceous-Eocene flysch), which may indicate the cessation of typical turbidite sedimentation in the Oligocene. Such change in sedimentation processes may indicate shallowing of the water basin (Poprawa et al., 2002; Oszczypko, 2006; Oszczypko et al., 2006; Dziadzio et al., 2019; Dziadzio & Matyasik, 2021).

Foraminiferal assemblages

Paleobathymetry of the Cretaceous-Paleogene basin of the Carpathians is largely based on foraminiferal microfauna. The Cretaceous-Eocene sediments of the Carpathians are generally dominated by siliceous Deep-Water Agglutinated Benthic Foraminifera (DWAF), which are similar to the Cretaceous-Paleogene microfauna of the deep-water (bathyal–abyssal) regions of the Atlantic, Pacific and Indian oceans according to their taxonomic composition and morphological features (Krasheninnikov, 1974; Ivanik, Maslun, 1977 and references therein; Kaminski, Gradstein, 2005 and references therein; Oszczypko et al., 2006). The Cretaceous-Paleogene deposits of the Carpathians contain also planktonic foraminifera, which suggests the marine normal saline conditions, as well as calcareous benthic foraminifera (Ivanik, Maslun, 1977; Oszczypko et al., 2006 and references therein). The paleobathymetry of the sediments filling the Dukla Nappe was previously discussed for both the Polish (Oszczypko et al., 2006 and references therein) and Ukrainian (Hnylko et al., 2020 and references therein) Carpathians. In the presented work, the foraminifera assemblages as indicators of paleobathymetry are distinguished (see Fig. 3) according to the methods outlined in the works (Murrey, 1976, 1991; Ivanik, Maslun, 1977; Kaminski, Gradstein, 2005 and references therein; Oszczypko et al., 2006).

Assemblages (1) 'agglutinated benthic' is defined in mainly non-calcareous (hemi)pelagic Cretaceous–Eocene deposits and composed of DWAF (see Fig. 3). Using the materials published in the works (Maslakova, 1966; Rozumeyko, 1980; Vialov et al., 1988; Hnylko et al., 2020 and references therein), such DWAF genera are characteristic for study deposits: *Glomospirella, Reophax, Thalmannammina, Recurvoides, Plectorecurvoides, Haplophragmoides, Trochammina* for the Shypot Formation (Aptian–Cenomanian); *Recurvoides, Plectorecurvoides, Haplophragmoides, Trochammina, Gerochammina, Uvigerinammina* for the lower member of the Yalovets Formation (Turonian–Coniacian); *Silicobathysiphon, Nothia, Rhabdammina, Hyperammina, Ammodiscus, Glomospira, Rzehakina, Reophax, Subreophax, Hormosina, Caudammina, Haplophragmoides, Recurvoides, Trochamminoides, Paratrochamminoides, Reticulophragmium, Spiroplectammina, Karrerulina* for the Senonian–Eocene deposits. The assemblages from red and green mudstones of the Yalovets Formation (Turonian–Coniacian) correspond to the Oceanic biofacies below the Calcite Compensatios Depth (CCD) (sensu Krasheninnikov, 1974, Gradstein, Berggren, 1981; Kaminski, Gradstein, 2005) according to the morphological features (small-sized test with fine-grained wall). The assemblages from Senonian-Eocene sediments correspond to the Flysch-type biofacies (sensy Gradstein, Berggren, 1981; Kaminski, Gradstein, 2005 and references therein) due to dominance fairly large and coarse-grained specimens and suggest bathyal–abyssal below or near CCD paleodepth.

Assemblages (2) 'agglutinated benthic–plankton' determined in Senonian–Eocene calcareous flysch (see Fig. 3) are composed of both DWAF and planktonic foraminifera. DWAF (the same as in assemblages (1) according their genera composition) generally prevail here. The planktonic foraminifera in Senonian sediments belong to the genera *Globotruncana* and *Rugoglobigerina,* and in Paleocene–Eocene sediments belong to the genera *Globigerina, Globigerinatheka, Parasubbotina, Subbotina, Globanomalina, Praemurica, Globoconusa, Morozovella* (using the data from the works (Maslakova, 1966; Andreeva-Grihorovych et al., 1987; Vialov et al., 1988; Hnylko et al., 2020). This difference in the genera composition of planktonic foraminifera reflects the event of the extinction of globotruncanids at the Cretaceous–Paleogene boundary. Calcareous benthic foraminifera *(Eponides, Anomalinoides, Cibicidoides, Stensioina)* are rarely found in the Senonian–Paleocene sediments. According to (Ivanik, Maslun, 1977; Oszczypko et al., 2006), the foraminifera of the assemblages (2) indicate bathyal–abyssal above CCD paleodepth.

Assemblages (3) 'plankton' determined in the Globigerina Marl (Sheshor horizon) are enriched with planktonic foraminifera with small admixtures of calcareous benthic foraminifera (see Fig. 3). Using the data from the works (Andreeva-Grihorovych et al., 1987; Dabagian, 1987; Ponomaryova, 1987; Vialov et al., 1988), planktonic foraminifera belong to genera *Catapsydrax, Globigerina, Subbotina, Dentoglobigerina, Tenuitella*, and benthic belong to genera *Nodosaria, Oridorsalis, Gyroidina, Heterolepa*. This microfauna is well preserved and according to (Murrey, 1976; Oszczypko et al., 2006) the foraminifera of the assemblages (3) indicate bathyal depths above CCD and foraminiferal lysocline.

Assemblages (4) 'calcareous benthic–plankton' determined in Oligocene sediments (Turytsa and Dusyno subnappes) are characterized by calcareous benthic foraminifera generally prevail here and planktonic foraminifera (see Fig. 3). Using the data from the works (Vialov et al., 1988 and references therein; Hnylko et al., 2020 and references therein), benthic foraminifera belong to genera *Robulus, Cibicides, Planulina, Asterigerina, Pararotalia, Bulimina, Neobulimina, Bolivina, Uvigerina* and planktonic belong to genera *Globigerina, Paragloborotalia, Globoturborotalita, Subbotina, Turborotalia, Pseudohastegirina, Tenuitella, Chiloguembelina*. According to (Murrey, 1976, 1991; Oszczypko et al., 2006), foraminifera of the assemblages (4) suggest the upper bathyal–littoral depths.

Some tectonic features

The Dukla Nappe, like other nappes of the Outer Carpathians, is characterized by a thrust-fold structure, i.e., the large Dukla Nappe is composed of smaller thrust-sheets thrusted over each other towards the north-east with a vergence that coincides to the general north-eastern vergence of the Ukrainian Outer Carpathians. This structure fully corresponds to the internal structure of accretionary prisms.

In addition, the Cretaceous-Eocene lower part of the Dukla sedimentary succession is more strongly deformed, while the Oligocene upper part of this succession are less intensively deformed. Oligocene sediments sometimes form gentle synclines, for example, the Maly Vyzhen Mountain brachysyncline (8 km north of Polyana, see Matskiv et al., 2003 and Fig. 1 for location) and Butova Mountain brachysyncline (southeast of Svalyava, see Matskiv, 2009 and Fig. 1 for location), while the underlied Cretaceous-Eocene formations are strongly deformed up to mélange. The cores of these gentle synclines are filled with Oligocene sediments (including coarse-grained Malyi Vyzhen Formation).

The Volosyanka thrust-sheet located ahead of the Dukla Nappe contains a synclinal fold. The syncline limbs are filled with the Oligocene Krosno Formation, and the syncline core is composed of the Volosyanka Olistostrome (Hnylko et al., 2021).

Formations – indicators of geodynamic development

The formations as indicators of the tectonic-sedimentary/geodynamic evolution were distinguished according to the existing theoretical models of tectono-sedimentary development of accretionary prisms/orogens (Einsele, 1992; DeCelles, Giles, 1996; Mutti et al., 2003) and the application of these models in the Carpathian region (Golonka et al., 2006, 2019; Oszczypko, 2006; Kováč et al., 2016, 2017; Plasienka, 2019; Schmid et al., 2020).

The pre-orogenic formation of the Carpathian remnant flysch basin is represented by the sediments that accumulated between the active margins of the microcontinental terrains and the passive margin of Eurasia. It includes the Cretaceous and Paleocene-Eocene flysch of the Dukla Nappe (Fig. 4). This flysch is composed mainly of the turbidite deposits and other gravity flow (e.g., debris-flow; grain-flow) deposits alternating with (hemi)pelagic sediments that contain DWAF (assemblages (1)) or both DWAF and planktonic ones (assemblages (2)). Here, the assemblages (1) (Aptian–Eocene) suggest to bathyal–abyssal depth below or close CCD and assemblages (2) (Senonian–Eocene) suggest to bathyal–abyssal depth above CCD.

Such sedimentation processes and deep-water environments are characteristic of the continental rise. In our case, this area was part of the Eurasian continental margin, bounded from the southwest by the active margin of the Alcapa microcontinental terrane (Hnylko et al., 2015; Kováč et al., 2016, 2017).

An important feature of the remnant Carpathian megabasin was the presence of intra-basin uplifts (ridges) within it: so-called ‘cordilleras’, indicated by the presence of exotic material in the flysch deposits such as clasts of metamorphic rocks, bioclastic shallow-water limestones, etc. Exotic material was derived from the basin margins and intrabasinal uplifts (ridges) located in the Outer Carpathian sedimentary realm. Now these paleoridges are buried under a Carpathian nappes (e.g., Vialov, 1981; Kruglov et al., 1985; Golonka et al., 2006, 2019; Oszczypko, 2006; Kováč et al., 2016; Cieszkowski et al., 2009, 2012; Gawêda et al., 2019, 2021; Hnylko et al., 2021; Kowal-Kasprzyk et al., 2021 and references therein).

The Aptian–Cenomanian Shypot Formation and the Turonian–Santonian Yalovets Formation are widespread in the Outer Carpathians (see Fig. 1), which indicates uniform sedimentation over a large part of the Carpathian Basin in this time. Moreover, in the Early Cretaceous (Aptian–Albian), the (hemi)pelagic deposits are enriched with black shales, which indicate anoxic conditions and poor oceanic circulations, which correspond in age to OAE1 (Oceanic Anoxis Event 1, see Gradstein et al., 2020). Yalovets

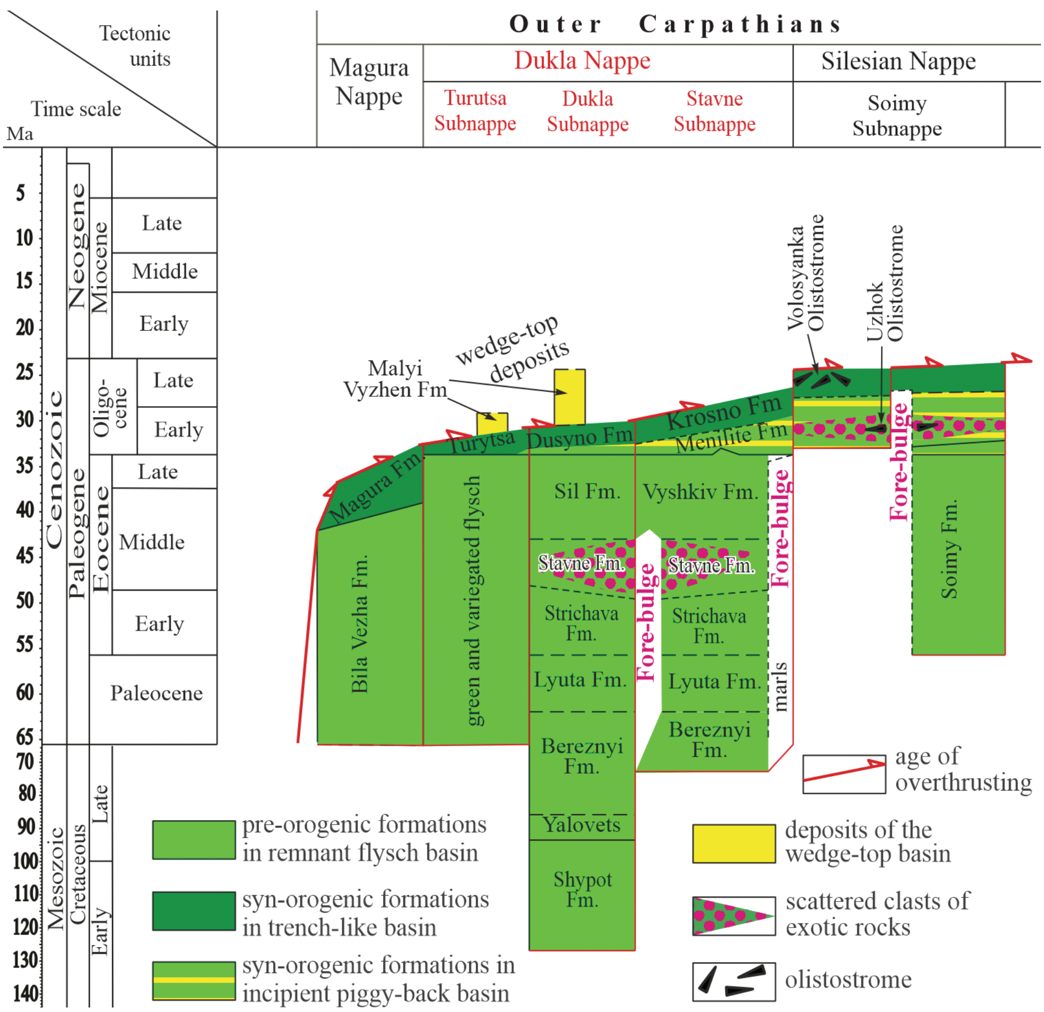

Fig. 4. Stratigraphic position of the sedimentary formations as indicators of the tectonic/geodynamic environments

Formation contains red shales (CORBs: Cretaceous Oceanic Red Beds, see Hu et al., 2012; Hnylko et al., 2023) suggested well oceanic circulation caused the activity of bottom currents and good aeration of the basin floor (Hnylko et al., 2023).

The Campanian–Lower Paleocene deposits in the Dukla Unit (Bereznyi Formation) are somewhat different from the deposits of other tectonic units of the Outer Carpathians (see Fig. 1) suggesting non-uniform sedimentation in Outer Carpathian Basin in this time. It is believed to be caused by the activation of intra-basin uplifts (ridges or 'cordilleras') including so called Fore-Magura and Silesian ridges, which limited the Dukla basin (Golonka, 2019) (Fig. 5). These paleoridges were reconstructed in the Polish Carpathians (Oszczypko 2006; Golonka, 2019 and references therein). In the Ukrainian Carpathians, remnants of the Silesian Ridge have also been recorded (see below), but no traces of the Fore-Magura Ridge have been found. The ridges have caused differences in sedimentation in the various basins between them (Danysh, 1973; Vialov, 1981; Kruglov et al., 1985; Golonka et al., 2006, 2019). In addition, these ridges could have caused the restriction of the circulation of oxygen-enriched bottom currents and local deposition of black mud sediments. As a result, the black mudstones, shales and marls enriched with organic matter of the Bereznyi Formation could be formed.

The Middle Paleocene-Eocene formations again contain the red shales suggesting restoration of well oceanic circulation of oxygen-enriched bottom currents.

In the Paleogene sediments of the Dukla Nappe and its foreland, the following indicators of the paleoridges are, firstly: exotic clasts of metamorphic rocks in the Eocene conglomerates of the Stavne Formation, which tend to the boundary between the Dusyno and Stavne subnappes of the Dukla Nappe; and secondly: the presence of fragments of exotic metamorphic rocks and shallow-water bioclastic limestones in the upper Rupelian–lower Chattian Uzhok Olistostrome and Late Oligocene Pikui Sandstone located in the Dukla Nappe foreland (Volosyanka thrust-sheet located in the inner part of the Silesian Nappe) (see Fig. 4). For the first case, the source area is the Eocene uplift ('Middle Cordillera' after Danysh, 1973) on the border of the Dusyno

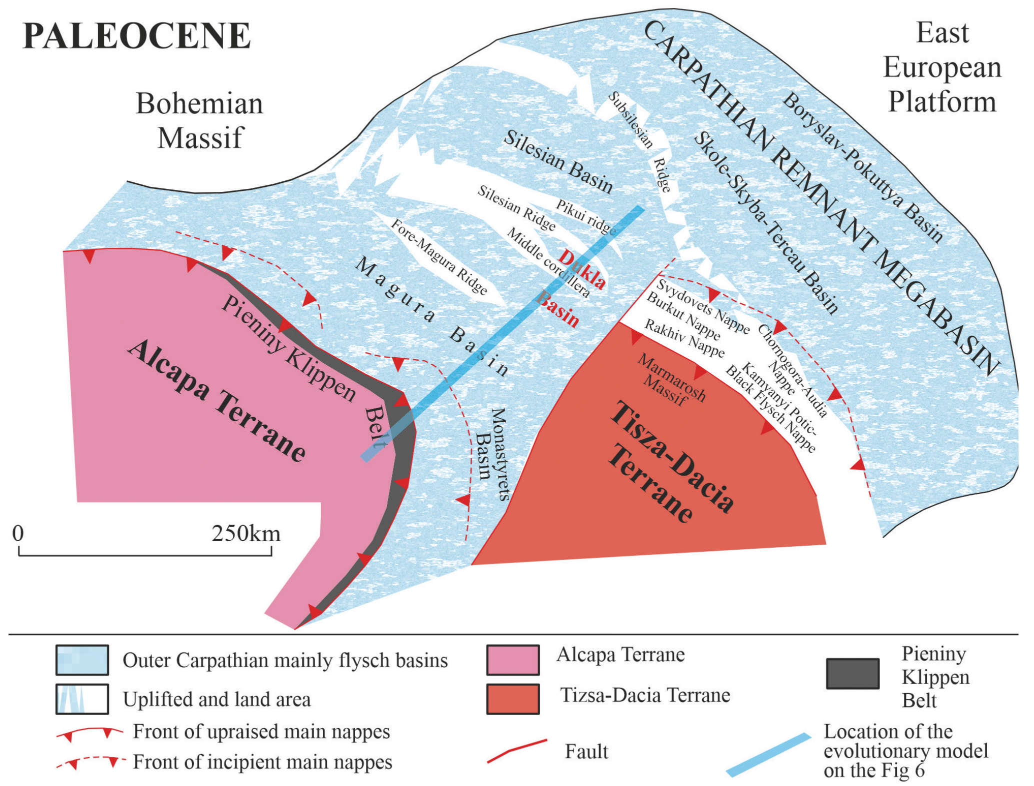

Fig. 5. Paleogeographic situation of the Carpathian realm during the Paleocene. Compiled using works of (Danysh, 1973; Vialov, 1981; Kruglov et al., 1985; Sandulescu, 1988; Csontos, Vörös, 2004; Golonka et al., 2006, 2019; Oszczypko, 2006; Schmid et al., 2008, 2020; Cieszkowski et al., 2009, 2012; Merten et al., 2010; Hnylko, Heneralova, 2014; Hnylko et al., 2015, 2021; Kováč et al., 2016; Gawêda et al., 2019, 2021; Kowal-Kasprzyk et al., 2021)

and Stavne subbasins, and for the second one, the source area is the middle/late Oligocene Pikui Ridge located in the foreland of the Dukla Nappe in the Silesian paleobasin (Hnylko et al., 2021). It is noted, the clasts of metamorphic rocks and bioclastic limestones are placed in the Oligocene synorogenic (see below) formation.

Another, third uplift, which was located between the Dukla and Silesian paleobasins, is suggested by red and green marls, which now make up the olistolith in the Volosyanka Olistostrome (Hnylko et al., 2021 and references therein). The red marls of this olistolith are enriched with late Zealandian–Thanetian (Globanomalina pseudomenardii Zone) planktonic foraminifera of good preservation, which according to (Oszczypko et al., 2006) indicates slowly (hemi)pelagic sedimentation at bathyal depths above both the CCD and the foraminiferal lysocline located much higher than the CCD. The ridge between the Dukla and Silesian paleobasins could have been a source area for the olistolith filled with these red shallow-water Paleocene marls. However, the age of this uplift can also cover a wider age interval up to the Oligocene, since the unification of sedimentation between the Dukla (Stavne subunit) and Silesian basins occured only in the Oligocene, when the Menilite and Krosno formation began to accumulate in both of them. Therefore, we assume that this third uplift, which could be a branch of the Silesian Ridge known in the Western Carpathians (Oszczypko, 2006; Golonka et al., 2019; Hnylko et al., 2021), existed until the Oligocene. In this case, we observe a gradual temporal migration of uplifts in the flysch basin in the foreland direction: from older uplifts in the southwest to younger ones in the northeast, namely: from the Eocene ridge into the Dukla basin (ridge between the Dusyno and Stavne subbasins, known as the 'Middle Cordillera' after (Danysh, 1973), to the pre-Oligocene ridge between the Dukla and Silesian basins (Silesian Ridge), and to the middle-late Oligocene ridge located in the foreland of the Dukla Nappe in the Silesian basin (Pikui Ridge) (Danysh, 1973; Hnylko et al., 2021) (see Figs. 2 and 4). We interprete these uplifts as the fore-bulges and associate their diachronic 'younging' towards the northeast with the processes of subduction and accretionary prism progradation (Fig. 6).

The synorogenic formation is represented by latest Eocene–Oligocene deposits belonging to both trench-like and 'piggy-back' facies. It includes sediments accumulated both ahead of the Magura/Dukla accretionary prism in the trench-like basin and in the areas further to the NE from the front of the Magura/Dukla accretionary prism outside the trench-like basin. However, the deposits of these areas were detached from the basement, uplifted and thrusted towards the platform starting from the Eocene-Oligocene boundary. In fact, incipient 'piggy-back basins' were formed in such areas (see Figs. 4, 6).

The transitional Eocene/Oligocene deposits are represented by the Globigerina Marl (Sheshor horizon) enriched with planktonic foraminifera with admixtures of calcareous benthic of assemblages (3) which indicate bathyal depths above CCD and foraminiferal lysocline. Change in the composition of benthic foraminifera – from DWAF in assemblages (1) and (2) to calcareous benthic in the assemblages (3) suggests a shallowing the Dukla sedimentary basin. The Oligocene sediments (Turytsa and Dusyno subnappes) are characterized by planktonic and calcareous benthic foraminifera of assemblages (4), which suggest the upper bathyal–littoral depths. The change from bathyal–abyssal to upper bathyal–littoral depths, which began at the turn of the Eocene and Oligocene, could be caused by syn-sedimentary tectonic movements such as the detachment of flysch masses from the sedimentary basement, their thrusting and uplifting (see Fig. 6).

In other words, at the turn of the Eocene and Oligocene, the flysch deposits of the Outer Carpathian megabasin, including the Dukla and Silesian basins, were involved into very gentle meganappe, on the top of which, the orogene formations (including Globigerina Marl, lower parts of the Turytsa, Dusyno, Menilite and Krosno formations) began to accumulate in the incipient piggy-back basin. Submarine slide/slamp complexes are widely developed in these formations.

Subsequently, the sandy lithofacies (the upper part of the Turytsa, Dusyno and Krosno formations) were accumulated in trench-like basin ahead of the Magura Nappe, and subsequently the Volosyanka Olistostrome was deposited in trench-like basin ahead of the Dukla Nappe front.

Therefore, we consider the Oligocene deposits completed the stratigraphic succession of the Dukla Nappe (the upper part of the Turytsa and Dusyno formation, and Krosno Formation), as well as the Oligocene Volosyanka Olistostrome developed ahead of he Dukla Nappe, to be a synorogenic trench formation. The diachronic 'younging' of trench lithofacies in the northeastern direction (from the upper part of the Turytsa to the Krosno lithofacies and to the Volosyanka Olistostrome

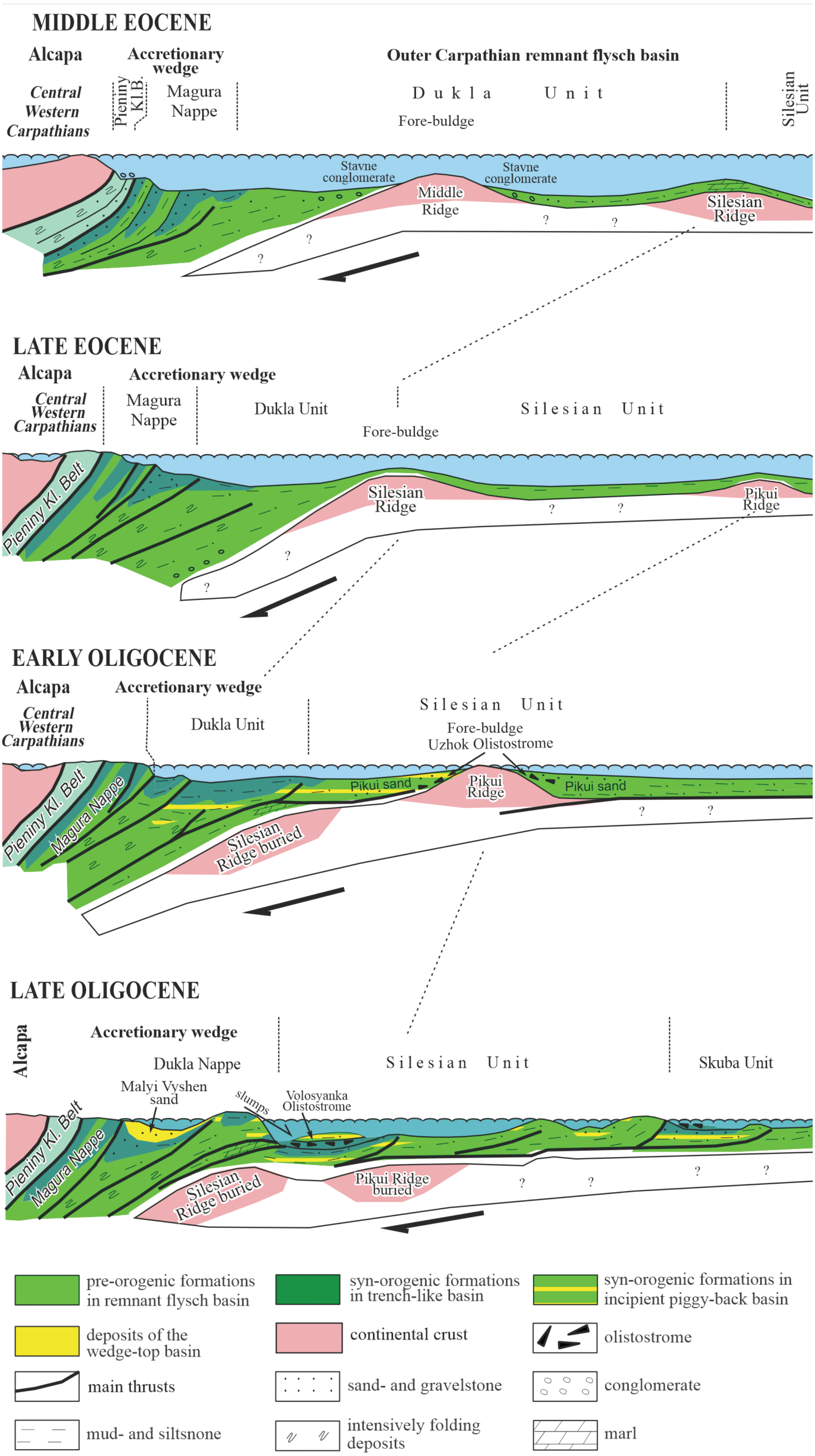

Fig. 6. Palinspastic reconstruction of the Dukla Unit tectonic-sedimentary evolution

(see Figs. 4, 6) is explained by the migration of the trench and the accretionary prism progradation in the same direction. Thus, before the Oligocene, the frontal unit of the prism was the Magura Nappe. In the Oligocene, the Turytsa Subnappe was attached to the prism and became frontal unit. Subsequently, the Dusyno and Stavne subnappes step by step were added to the prism. As a result, the Dukla Nappe became the frontal unit of the accretionary prism and supplied detrital material to the trench, where the Volosyanka Olistostroma accumulated (see Fig. 6). A peculiarity of the development of the Magura/Dukla prism was its thrusting onto the incipient ‘piggy-back basins’ probably located in the hinterland of the Outer Carpathian very gentle emerging meganappe (see Figs. 4, 6).

The post-orogenic formation was accumulated in the ‘wedge-top basin’ (Mutti et al., 2003; Artoni, 2013) after forming of the accretionary prism. It includes the sandy deposits of the Malyi Vyzhen Formation, which complete the stratigraphic succession in the back part of the Dukla Nappe (Dusyno Subnappe). Nannoplankton of the NP25 zone (Late Oligocene) were found in the upper part of the Malyi Vyzhen Formation (Romaniv, 1991). Thus, this part of the Malyi Vyzhen Formation is the same age as the synorogenic Volosyanka Olistostrome accumulated in the trench basin ahead of the movable Dukla Nappe front (see Figs. 4, 6). That is, the Malyi Vyzhen sandy sediments were accumulated in the wedge-top basin ‘on the body’ of the Dukla Nappe, which had been a frontal nappe of the accretionary prism at the same time and supplied the clastic material to the trench-like Volosyanka Olistostrome basin.

It should be noted, the Malyi Vyzhen Formation fills the cores of a very gentle synclines including both brachysyncline of the Maly Vyzhen mountain (8 km north of the city of Polyana) (Matskiv et al., 2003) and brachysyncline of the Butova mountain (southeast of the city of Svalyava) (Matskiv, 2009) (see Fig. 1 for location). The Oligocene sediments in these structures are almost undeformed, while the Paleocene-Eocene and Cretaceous strata underlying the Oligocene are highly deformed, intensively folded, and somewhere transformed into tectonic breccia and mélange. This suggests that the deformation of the Dukla unit starts in the pre-Oligocene time. Subsequently, the Dukla Nappe was formed as the frontal structure of the accretionary prism. Malyi Vyzhen sediments were accumulated on already deformed structures of the accretionary prism. The angular unconformity between the Eocene and Oligocene was not directly observed by geologists in the outcrops, that suggests the syn-sedimentary nature of the deformations.

Evolutionary features

Presented materials and available paleotectonic reconstructions (Golonka et al., 2006, 2019; Oszczypko, 2006; Hnylko et al., 2015, 2020, 2021; Kováč et al., 2016, 2017; Plasienka, 2019; Schmid et al., 2020) allow to detail the main features of evolution of the Ukrainian Carpathian Dukla Nappe.

The Carpathian basin was the part of the Alpine Tethys formed in the Jurassic and ealiest Cretaceous time during the rifting/spreading the southern margin of Eurasia. In the Cretaceous, it was located between the Eurasian passive margin and the active margins of the Alcapa and Tisza-Dacia microcontinents located in the Tethys Ocean. The subduction of the (sub)oceanic lithosphere (the basement of the flysch basin) beneath active microcontinental margins led to the deformation and detachment of flysch sediments from their substrate, and to growing the accretionary prism ahead of these terranes (Golonka et al., 2006, 2019; Oszczypko, 2006; Hnylko et al., 2015; Schmid et al., 2020 and references therein).

In the Cretaceous, an accretionary prism was formed ahead of the Alcapa active margin, which was “step by step” built up by various facies of the Peniny Klippen Belt (Oszczypko, 2006; Kováč et al., 2016; Plasienka, 2019). In the Aptian–Santonian time uniform sedimentation over a large part of the Outer Carpathian Basin. Here in the Aptian-Albian, anoxic conditions corresponding in age to OAE1 were developed (Shypot Formation with black shales); and in the Turonian–Santonian, well oxidant conditions and aeration of the basin floor existed (Yalovets Formation with CORBs). In the Campanian–Lower Paleocene, activation of intra-basin uplifts (ridges or ‘cordilleras’) in the Outer Carpathian Basin occurred (see Fig. 5). It could have caused the restriction of the circulation of oxygen-enriched bottom currents and local deposition of black mud sediments (Bereznyi Formation with black shales). In the Paleocene–Eocene, the restoration of well oceanic circulation of oxygen-enriched bottom currents in the Dukla basin occurred (Paleocene–Eocene Dukla succession with red shales).

In the Paleocene–Eocene, in front of fore-Alcapa prism composed of the Pieniny Klippen Belt formations, a deep-sea trench was formed in the Magura basin and filled with thick psammites of the Magu-

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ra Formation. The trench sediments were detached from the basement and added to the prism as the Magura Nappe. Subsequantly in the Oligocene, the sedimentary depocenter (trench) gradually shifted to the northwest into the Dukla basin (Hnylko et al., 2015) (see Figs. 4, 6).

In the Paleocene–Eocene, in the Dukla basin, the pre-orogenic formations were accumulated mainly by turbidite and other gravity flows, and by (hemi) pelagic sedimentation. The mainly bathyal–abyssal depths near or below the CCD were in this basin, as evidenced by both the sedimentary features and the (hemi)pelagic sediments contained DWAF. The intra-basinal uplift or 'Middle Cordillera' is recorded in the middle part of the Dukla basin, which supplied exotic clasts to the conglomerates of the Stavne Formation in the middle Eocene (see Figs. 2, 4). We consider this uplift to be a fore-bulge ahead of the Magura accretionary prism foreland (Figs. 6, 7). Such 'fore-bulge' (or 'outer rise) can be formed between a trench and an open basin due to the bending of a lithosphere plate (oceanic or continental) before its dipping into a subduction zone (DeCelles, Giles, 1996; Contreras-Reyes, Garay, 2018).

The forebulge gradually migrated in the foreland direction from the Eocene Middle Cordillera to the pre-Oligocene uplift between the Dukla and Silesian basins (Silesian Ridge), and subsequently to the Oligocene Pikuy Ridge ahead of the Dukla Nappe (see Figs. 2, 4, 6–8).

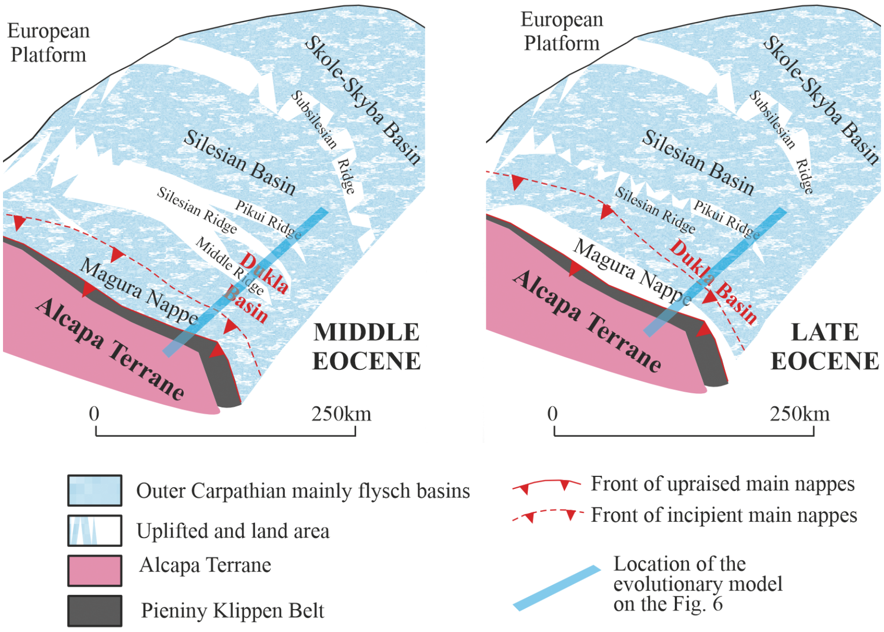

Fig. 7. Paleogeographic situation of the Dukla Basin and adjacent tectonic units during Middle-Late Eocene. Compiled using works of (Danysh, 1973; Vialov, 1981; Kruglov et al., 1985; Csontos, Vörös, 2004; Golonka et al., 2006, 2019; Oszczypko, 2006; Schmid et al., 2008, 2020; Cieszkowski et al., 2009, 2012; Hnylko et al., 2015, 2021; Kováč et al., 2016; Gawêda et al., 2019, 2021; Kowal-Kasprzyk et al., 2021)

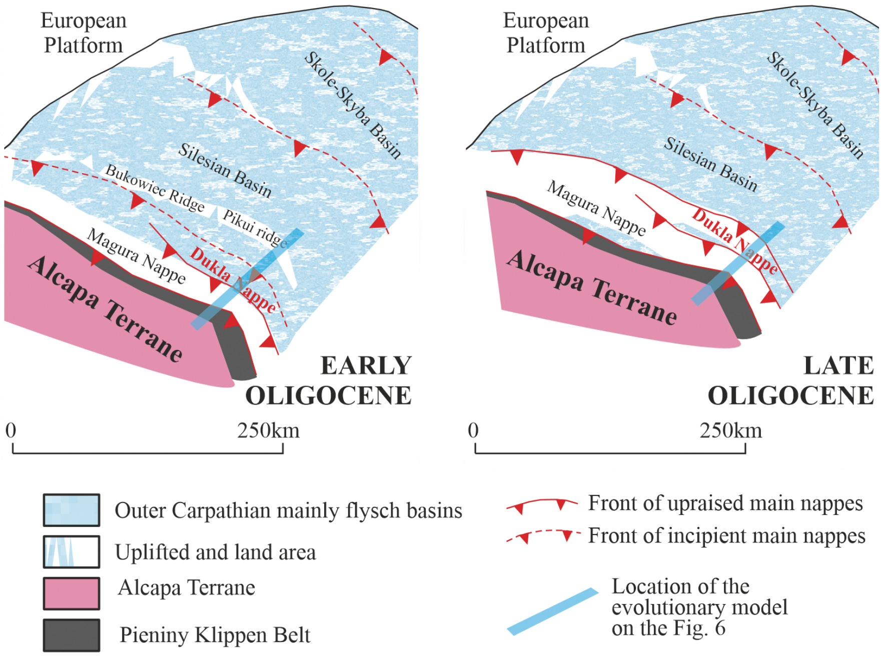

Fig. 8. Paleogeographic situation of the Dukla Basin and adjacent tectonic units during Oligocene. Compiled using works of (Danysh, 1973; Vialov, 1981; Kruglov et al., 1985; Csontos, Vörös, 2004; Golonka et al., 2006, 2019; Oszczypko, 2006; Schmid et al., 2008, 2020; Cieszkowski et al., 2009; Hnylko et al., 2015, 2021; Kováč et al., 2016)

During the Oligocene, the migration of the trench basin also occured in the same NE direction, which is reflected by diachronous 'younging' of synorogenic trench lithofacies (from the upper part of the Turytsa Formation up to the Krosno Formation and Volosyanka Olistostrome) (see Figs. 4, 6). The forebulge/trench migration was related to the growing of the accretionary prism (see Figs. 6–8).

In the late Oligocene, a post-orogenic formation including psammites of the Malyi Vyzhen Formation was accumulated on the top of the accretionary prism composed of the back part of the Dukla Nappe. At the same time, the synogrogenic Volosyanka Olistostrome was accumulated ahead of the prism front, i.e. ahead of the Dukla Nappe in the Silesian Basin. This shows that orogenic and post-orogenic processes occurred simultaneously in different parts of the accretionary prism (see Fig. 6).

It should be noted that Carpathian orogeny became especially active starting from the turn of the Eocene and Oligocene, when the depth of the flysch basin changed from bathyal-abyssal to upper-bathyal-sublittoral, which could be caused by the detachment of flysch masses from the sedimentary base, their thrusting and uplift, and subsequent attachment to the prism (Hnylko et al., 2020).

In addition, at the boundary between the Eocene and Oligocene, another important event such as closing deep-water connection link between the World Ocean and the Carpathian Basin occurred due to the formation of the Alps (Palcu et al., 2023). The Carpathian Basin was transformed into the segment of the Paratethys: systems of isolated and semi-isolated basins (Kováč et al., 2017 and references therein). As a result, the rich in organic matter Oligocene sediments with black shales including Menilite/Dusyno formations were formed.

Completion of forming the Dukla Nappe was accompanied by sedimentary basin uplift and termination of sedimentation in the Dukla unit and its foreland in post-Oligocene time.

Conclusion

Deposits of the Dukla Nappe and its foreland are divided into next formations.

The pre-orogenic formation of the Carpathian remnant flysch basin is represented by the Cretaceous–Eocene flysch composed mainly of products of catastrophic turbidite and other gravity flows, and (hemi)pelagic sedimentation. Here, non-calcareous (hemi)pelagic sediments (Aptian-Eocene) contain Deep-water Agglutinated Foraminifera (DWAF), and calcareous ones contain DWAF and planktonic foraminifera. Both sedimentary features and foraminiferal assemblages suggest to bathyal–abyssal sedimentary basins near and below or above the Calcite Compensatios Depth (CCD).

Synorogenic formation is represented by latest Eocene–Oligocene deposits belonging to both 'piggy-back' and trench-like facies. The first of them (Globigerina Marl, lower parts of the Turytsa, Dusyno, Menilite and Krosno formations) were accumulated on the top of the emerging Outer Carpathian gentle meganappe in the incipient 'piggy-back' basins; and the second one (sandy lithofacies of the upper part of the Turytsa, Dusyno and Krosno formations) as well as the Volosyanka Olistostrome were accumulated in the trench basin ahead of the Magura/Dukla accretionary prism. A peculiarity of the development of the Magura/Dukla prism was its thrusting onto the incipient 'piggy-back basins' probably located in the hinterland of the Outer Carpathian meganappe (see Figs. 4, 5). The Globigerina Marl and marl deposits of the Turytsa and Dusyno formations contain planktonic and calcareous benthic foraminifera which suggest shallowing of the syn-orogenic sedimentation from bathyal depths above CCD and foraminiferal lysocline during latest Eocene–earliest Oligocene (Globigerina Marl) up to upper bathyal–littoral depths during Oligocene.

Post-orogenic formation was accumulated in the wedge-top basins after forming of the accretionary prism. It includes the Late Oligocene sandy deposits of the Malyi Vyzhen Formation, which complete the stratigraphic succession in the back part of the Dukla Nappe (Dusyno Subnappe). Wedge-top Malyi Vyzhen Formation in the hinterland of the Dukla Nappe was deposited in the same time with the trench-like Volosyanka Olistostrome located in the foreland of the Dukla Nappe.

The identified formations indicate the main stages of the Dukla unit evolution.

In the Cretaceous–Eocene, the Dukla basin was located ahead of the Alcapa active (micro)continental margin and belonged to the Outer Carpathian deep-water remnant-type basin. Into this basin the forebulges were formed and migrated from the Eocene uplift into the Dukla basin (Middle Cordillera) to the pre-Oligocene uplift between the Dukla and Silesian basins (Silisian Ridge), and subsequently to the Oligocene Pikui Ridge ahead of the Dukla Nappe. During the Oligocene, the migration of the trench basin also occured in the same NE direction, which is reflected by diachronous 'younging' of synorogenic trench lithofacies (from the upper part of the Turytsa Formation up to the Krosno Formation). The fore-bulge/trench migration was related to the growing of the accretionary prism.

In the late Oligocene, a post-orogenic psammites of the Malyi Vyzhen Formation was accumulated on the top of the accretionary prism composed of the back part of the Dukla Nappe.

It should be noted that Carpathian orogeny became especially active starting from the turn of the Eocene and Oligocene, when the depth of the flysch basin changed from bathyal-abyssal to upper-bathyal-sublittoral, which could be caused by the detachment of flysch masses from the sedimentary base, their thrusting and uplifting. Completion of forming the Dukla Nappe was accompanied by sedimentary basin uplift and termination of sedimentation in the Dukla unit and its foreland in the post-Oligocene time.

The analysis of the tectonic-sedimentary evolution of the Dukla unit suggests that the potentially oil and gas-generating sediments, including black shales enriched in organic matter, could have formed in connection: (1) with global oceanic anoxic events (OAE1: Early Cretaceous black shales); (2) with regional events (Paratethys forming: Oligocene Menilite-like shales); and (3) with local tectonically-induced events including the growth of intra-basin uplifts/ridges. These ridges could have caused the restriction of the circulation of oxygen-enriched bottom currents and local deposition of black mud sediments (Campanian–Lower Paleocene Bereznyi Formation with black shales).

Доповнено уявлення про стратиграфію, тектоніку й еволюцію Дуклянського покриву Українських Карпат на основі геокартувальних робіт, структурних, седиментологічних і мікропалеонтологічних досліджень. Визначено чотири основні форамініферові угруповання, які вказують на глибини палеобасейнів: (1) «аглютинований бентос»; (2) «аглютинований бентос – планктон» (3) «планктон»; (4) «вапнистий бентос-планктон». Перші два містять глибоководні аглютиновані форамініфери (*англ.* Deep-Water Agglutinated Foraminifera – DWAF), які вказують на батіально-абісальні глибини як нижче, так і вище рівня компенсації кальциту (*англ.* Calcite Compensation Depth – CCD). Угруповання (3) містять головно планктонні форамініфери та вказують на батіальні глибини вище CCD і форамініферової лізокліни. Угруповання (4) містять вапнисті бентосні і планктонні форамініфери та вказують на глибини верхньої батіалі–субліторалі.

Були виділені такі осадові формації – індикатори розвитку акреційної палеопризми Західних Флішевих Карпат. Доорогенна формація залишково флішевого Карпатського басейну – це відклади, що накопичувались між активними окраїнами мікроконтинентальних терейнів та пасивною окраїною Євразії. Сюди відносимо крейдовий і палеоцен-еоценовий фліш Дуклянського покриву. Формація складена переважно продуктами діяльності турбідитних та інших гравітаційних потоків, а також фонової геміпелагічної седименації. Фонові осади містять форамініферові угруповання (1, 2). Тут глибоководні аглютиновані форамініфери належать здебільшого до родів *Glomospirella, Reophax, Thalmannammina, Recurvoides, Plectorecurvoides, Haplophragmoides, Trochammina* в шипотській світі (альб-сеноманський інтервал); *Recurvoides, Plectorecurvoides, Haplophragmoides, Trochammina, Gerochammina, Uvigerinammina* в нижній частині яловецької світи (турон–коньяк) та *Silicobathysiphon, Nothia, Rhabdammina, Hyperammina, Ammodiscus, Glomospira, Rzehakina, Reohax, Subreophax, Hormosina, Caudammina, Haplophragmoides, Recurvoides, Trochamminoides, Paratrochamminoides, Reticulophragmium, Karrerulina* в сенон-еоценових відкладах. Доорогенна формація та частково синорогенна містять відклади-індикатори внутрішньобасейнових піднять, зокрема конгломерати з уламками екзотики та мілководних мергелів і вапняків. Встановлено часову міграцію цих піднять у напрямку форланду орогену: від еоценового підняття всередині Дуклянського басейну (Серединна кордільєра) до доолігоценового – між Дуклянським і Сілезьким басейнами (Сілезька кордільєра) і до ранньоолігоценового – розвиненого у форланді Дуклянського покриву в Сілезькому басейні (Пікуйська кордільєра). Синорогенна формація – це малопотужні «глобігеринові мергелі» межі еоцену й олігоцену та відклади олігоцену. «Глобігеринові мергелі» збагачені планктонними форамініферами угруповання (3), а олігоценові відклади турицької і дусинської світ (Турицький і Дусинський субпокриви) містять вапнисті бентосні і планктонні форамініфери угруповання (4). Планктонні форамініфери угруповань (3, 4) належать до родів *Catapsydrax, Globigerina, Paragloborotalia, Globoturborotalita, Subbotina, Dentoglobigerina, Turborotalia, Pseudohastegirina, Tenuitella, Chiloguembelina,* а бентосні – до родів *Nodosaria, Robulus, Oridorsalis, Cibicides, Planulina, Asterigerina, Pararotalia, Bulimina, Neobulimina, Bolivina, Gyroidina, Heterolepa, Uvigerina*. Зміна батіальних–абісальних глибин на верхньобатіальні–субліторальні, яка розпочалась на рубежі еоцену й олігоцену, могла бути зумовлена зривом з основи флішевих мас, їх підняттям та загальним конседиментаційним насуванням Зовнішньокарпатського мегапокриву. Олігоценова седиментація відбувалась «на тілі» цього мегапокриву, а в тильній частині мегапокриву перед фронтом послідовно зростаючих Магурського та Дуклянського покривів формувався жолоб. До літофацій жолобу відносимо грубокластичні відклади, розвинені у верхах страграфічних розрізів олігоценових відкладів Дуклянського покриву, а також потужну Волосянківську олістострому, поширену перед фронтом Дуклянського покриву. Діахронне «омолодження» цих літофацій у північно-східному напрямку пояснюється міграцією жолобу та нарощуванням структур акреційної призми в цьому ж напрямку. Посторогенна формація накопичена після формування структури акреційної призми (*англ.* «wedge-top basins»). До неї відносимо піскуваті породи маловиженської світи, які заповнюють ядра пологих синкліналей і завершують стратиграфічний розріз тильної частини Дуклянського покриву. Виділені формації вказують на основні події розвитку Дуклянської одиниці. На активній окраїні Алькапи зростала акреційна призма, перед її фронтом накопичувались осади жолобу, дещо далі від призми розвивалось передове підняття (*англ.* «fore-bulge»), які мігрували в напрямку форланду. Міграція жолобу та передового підняття відображає процеси росту акреційної призми, її насування до північного сходу.

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