# A record of Eocene subaqueous large-scale mass movements in the external Tethys Ocean (Skyba Nappe, Outer Carpathians, Ukraine)

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***Abstract:*** Studies were carried out on the Popeli Beds, which were mainly deposited by mass movements along the northern slope of the Tethys Ocean and form an olistostrome sequence occurring within Paleogene deposits. The analysis focuses on the deposits exposed in the northern limb of the Pobuk Syncline – the structural element of the Oriv Skyba – in the external zones of the Skyba (Skole) Nappe. Structural, lithological, sedimentological and paleontological studies allowed for the identification and characterisation of the olistostrome sequence and revealed its structure. Six individual olistostrome complexes were distinguished and described in detail. Five of these are gravelly mudstones with oversized clasts, overlain by medium-bedded mudstones and sandstones, the last one consists of deformed packages of thin bedded mudstones and sandstones. The olistostrome sequence is composed of a wide range of lithological and stratigraphic components, including Tithonian – Berriasian shallow-water limestones (so-called Štramberk-type limestones), Upper Cretaceous limestones, sandstones, and siliceous rocks, as well as Paleogene mudstones, sandstones, limestones, and marls. Biostratigraphic analysis of foraminifera indicates the Bartonian as the time of displacement. The olistostrome niche migrated downslope from a lower shelf-upper slope position characterized by marl sedimentation, to slope areas dominated by turbidite sedimentation.

***Keywords:*** Outer Carpathians, Skyba Nappe, mass movements, olistostrome, foraminifera, exotic components

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## INTRODUCTION

Subaqueous mass movements play an important role in the sedimentation history of the Carpathian orogen. They were commonly formed in the Carpathian Tethys, especially during periods of tectonic distress associated with geodynamic stages of basin development (Cieszkowski et al. 2012, Golonka et al. 2015, 2022). These deposits are the result of slope instability, gravitational transport, and deposition (e.g., Abbate et al. 1970, Lucente & Pini 2003, Neuendorf et al. 2005, Cieszkowski et al. 2012, Festa et al. 2012, Golonka et al. 2015). It produces sedimentary bodies of different sizes from a few to tens of metres, typically characterised by chaotic structures called sedimentary mélanges (= olistostromes). Olistostromes occur in various structural units and within deposits of different ages in the Outer Carpathians (e.g., Książkiewicz 1958, Unrug 1968, Szymakowska 1976, Kulchytskyi 1977, Malik & Olszewska 1984, Słomka 1986, 2001, Cieszkowski 1992, Bromowicz 1998, Olszak 2006, Cieszkowski et al. 2009, 2012, 2017, Ślączka et al. 2012, Hnylko 2014, Waśkowska & Cieszkowski, 2014, Golonka et al. 2015, 2022, 2025, Jankowski 2015, Hnylko & Hnylko 2016, 2019, Strzeboński et al. 2017, Heneralova & Khomyak 2018, Bubík et al. 2020, Siemińska et al. 2020, Hnylko et al. 2021, Woyda et al. 2023, Waśkowska & Golonka 2025, Waśkowska et al. 2025), among which the Popeli olistostromes occupy a special position. They have considerable thickness and a widespread distribution. In the lithostratigraphic scheme it has been distinguished as an independent unit, informally referred to as the Popeli Beds, which are part of the Cretaceous – Paleogene sedimentary succession of the marginal part of the Outer Carpathians (Kotlarczyk 1985, Ponomareva 1988, Ryłko 2004) (Figs. 1, 2).

The Popeli Beds were described for the first time as the Popiele Series by Kropaczek in Grzybowski (1919) from the Borysław area. The Popeli Beds are distinguished from the surrounding deposits by the presence of dark sandy clays (also described as highly calcareous mudstones and marls) containing large blocks of shallow-water, uppermost Jurassic – lowermost Cretaceous limestones (the so-called Štramberk-type limestones), dolomites, quartzites, and phyllites (Grzybowski 1919, Rogala 1925, Świdziński 1948, Czaplicka et al. 1968, Gucik et al. 1973, Kulchytskyi 1977), and large blocks of marls and shales (Nowak 1927, Dżułyński & Kotlarczyk 1965, Dżułyński et al. 1979, Kotlarczyk 1988, Rajchel 1990). This lithostratigraphic division is also characterised by numerous fossil remains, with particular attention given to gastropods and foraminifera (Wiśniowski 1908, Grzybowski 1919, Rogala 1925, 1941, Syniewska 1937, Mjatliuk 1950, Krajewski & Urbaniak 1964, Watycha 1964, Morgiel & Szymakowska 1978, Krach 1985, Kotlarczyk 1988). The olistostromal origin of the Popeli Beds was described by Dżułyński and Kotlarczyk (1965).

In the Skole region (western Ukraine), the Popeli Beds are widely distributed in the marginal slice (thrust-sheet = "skyba" or "skiba") of the Outer Carpathians, and also in the more internal thrust-sheet, the so-called Oriv Skyba, where some of the most distal intra-basin olistostrome facies are directly accessible (Waśkowska et al. 2023). The main research challenge is that the age, internal structure, and sedimentological conditions of the formation of the Popeli olistostrome sequence remain debatable. The aim of the study is to determine sedimentological conditions of the olistostrome sequence and the age of the Popeli Beds based on comprehensive structure, lithological, and micropaleontological analysis.

## GEOLOGICAL OUTLINE AND THE STUDIED SECTION

The study was conducted in the Outer Carpathians, which are part of the Carpathian orogen formed during the Early Miocene as a result of the closure of the western part of the Tethys Ocean (Książkiewicz 1977, Oszczypko 2004, 2006, Golonka et al. 2019a, 2019b, 2021 and references therein). The Outer Carpathians are filled with Cretaceous – Miocene deposits, mainly clastic sediments (flysch), with a total thickness of up to several kilometres, which are detached from its sedimentary base and thrust over the Carpathian Foredeep. The Carpathian deposits are dominated by products of turbidite and, to a lesser extent, products of other sediment density flows which transported terrigenous material into the deep-sea parts of the basin (e.g., Dżułyński et al. 1979, Golonka & Picha 2006, Oszczypko et al. 2006, Golonka et al. 2019a). The Outer Carpathians are the externides with a nappe structure, where individual nappes are imbricated in a stack-like manner thrusted towards the NE (up to 3 km and more in the area studied, see Dosin 1963). The Boryslav-Pokuttya and Skyba nappes are the northernmost and structurally lowest units of the Outer Carpathians (Jankowski et al. 2004, Hnylko 2014) (Fig. 1). The Boryslav-Pokuttya Nappe is almost completely buried at depth and only Neogene sediments are exposed. The Skyba Nappe is called the Skole Nappe in the Polish tectonic nomenclature, after its *locus typicus* near the Ukrainian town of Skole (Nowak 1914). The term "Skyba" reflects the structural composition of this unit, which consists of several slices, locally known as "skybas" (or "skiby" in Polish). Along the intersection through the town of Skole, the Skyba Nappe consists of seven thrust-sheets known as (from outer to inner ones) Berehova, Oriv, Skole, Parashka, Zelemyanka, Rozhanka and Slavska skybas (Vialov et al. 1981).

In the vicinity of Skole and Verhnye Synyovydnye (see Fig. 1), the stratigraphic sequence of the Skyba Nappe is represented by the Senonian – Paleocene medium- and thick-bedded sandstones with thin intercalations of grey mudstones (Stryi Fm.), Paleocene thick-bedded sandstones (Yamna Fm.); Eocene variegated and green shales with thin- to medium-bedded sandstone intercalations (Manyava Fm.); medium- and thick-bedded sandstones (Vyhoda Fm.); thin- to medium-bedded sandstones and mudstones (Bystrytsia Fm.); olistostrome deposits (Popeli Beds), marls, sandstones and mudstones; Oligocene black shales and cherts (Menilite Fm.); and grey mudstones and sandstones (Lopyanets Beds) (Fig. 2).

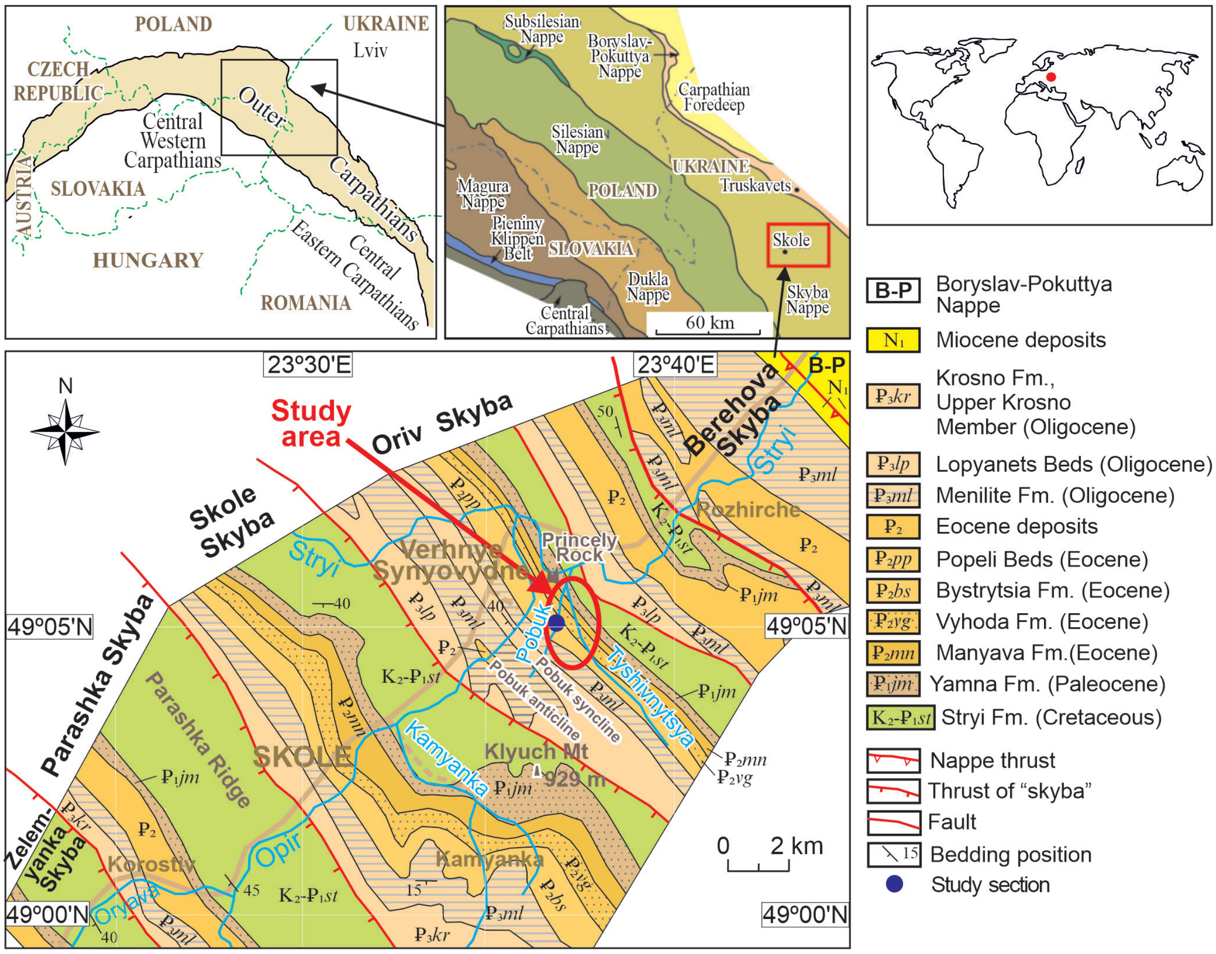


***Fig. 1.*** *Location of the study section. The geologic map modified after Dosin (1963), Jankowski et al. (2004), and Hnylko et al. (2022)*

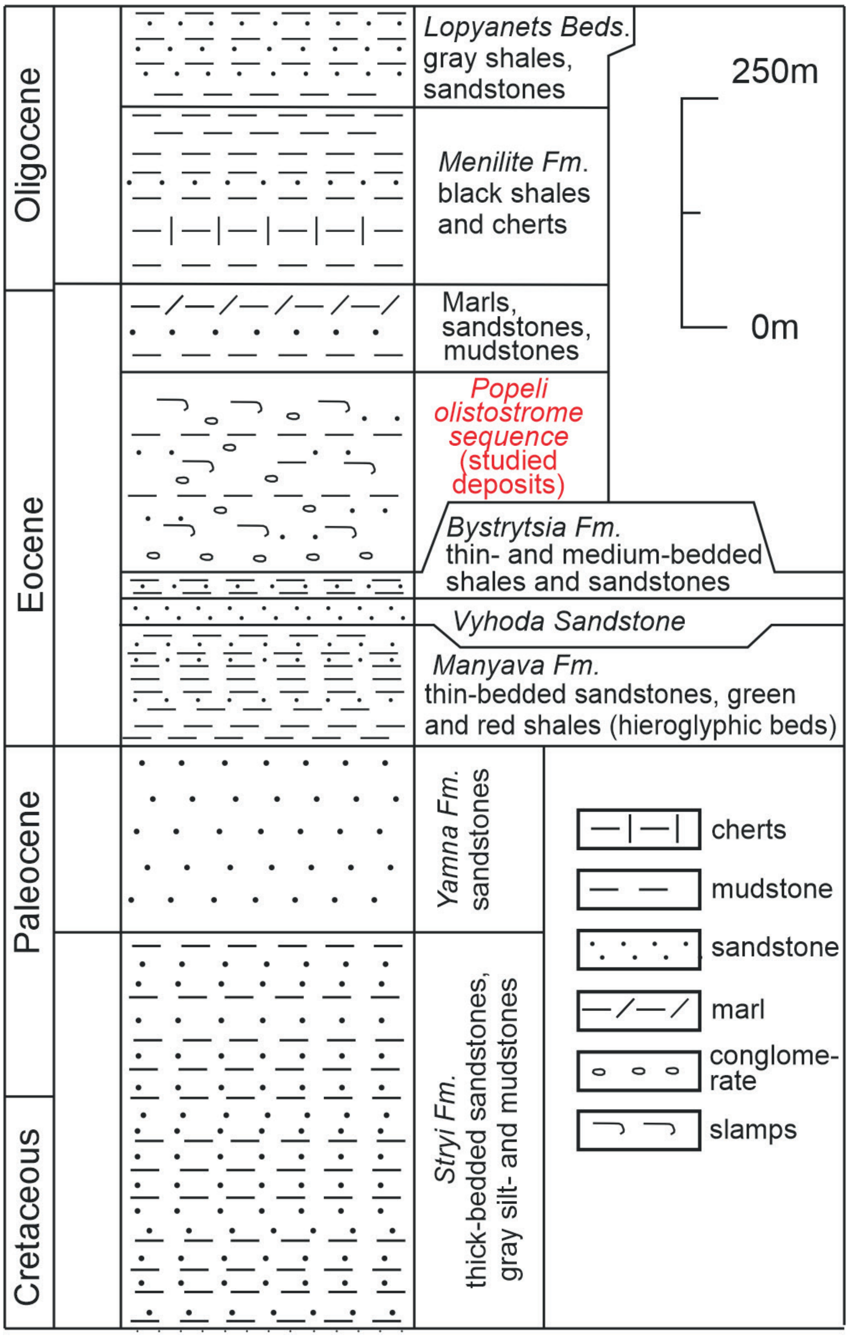


***Fig. 2.*** *Stratigraphic column of the Oriv thrust-sheet (= "skyba") of the Skyba Nappe in the study area in the basins of the Pobuk and Tyshivnytsia rivers (see also Fig. 1). The column is compiled based on our own research and using the state geologic map (Dosin 1963)*

In the study area, the Popeli Beds are part of the Berehova and Oriv skybas of the Skyba Nappe (see Fig. 1) and they are also known from the Boryslav-Pokuttya Nappe (Vialov et al. 1981). These beds occur within Paleogene deposits (see Figs. 1, 2).

The analysis of the Popeli Beds was carried out on the southern section, located on the northeastern limb of the Pobuk Syncline, about 8 km northeast of the town of Skole, situated in the Skole Beskyds of the Eastern Carpathians. This syncline is the structural element of the Oriv Skyba (Fig. 1).

The studied section is located in the village of Pobuk (start: 49,0853635°N 23,6139646°E and end: 49,0819310°N 23,6154362°E). The Popeli Beds are exposed along a series of outcrops in the Pobuk Stream, a short left-bank tributary of the Tyshivnytsya River, which flows into the Stryi River (Figs. 1, 3, 4). The Pobuk stream cuts across the structure of the syncline, providing accessible exposures for study.

## MATERIAL AND METHODS

### Fieldwork and sampling

To investigate the structure and lithological inventory of the Popeli olistostrome sequence, fieldwork was carried out in 2018, 2019 and 2024. This included classical geological mapping, logging with structural and textural identification of olistostrome deposits, an assessment of the lithological inventory, and sampling.

### Thin section analysis

Thin sections were prepared from selected clasts at the AGH University of Krakow, Faculty of Geology, Geophysics and Environmental Protection. They were microscopically analysed at the Department of General Geology and Geotourism (DGGG) AGH using a Nikon Eclipse LV100 POL optical microscope. Petrographic analyses were conducted, and where applicable, micropaleontological and biostratigraphical analyses were carried out.

### Foraminifera analysis

Soft rocks such as marls and mudstones were sampled for biostratigraphic analysis of foraminifera. They were subjected to standard micropaleontological preparation in the AGH Laboratory of Micropaleontology and Dendrochronology. Preparation consisted of the fragmentation of approximately 0.5 kg of sample (for samples with the prefix PA) or 200 g (prefix UK) in an aqueous sodium sulphate solution. Fragmentation involved the mechanical disintegration of the rock by repeated cycles of sodium sulphate crystallisation achieved by alternately cooling and heating the mixture. Subsequent stages included fractional separation, removal of particles smaller than 0.063 mm, and removal of salt by washing the residue through sieves. The residue was then dried in an oven at 80°C, next all microfossils were picked out. These foraminiferal tests were subjected to stratigraphic analysis and documented by imaging. Foraminiferal studies were carried out using a Nikon stereoscopic microscope at the DGGG AGH (samples with the prefix PA). Microfauna selection and determination from samples with the prefix UK were carried out at the Institute of Geology and Geochemistry of Fossil Minerals of the National Academy of Sciences of Ukraine, using MBS-9 and MPSU-1 microscopes. Imaging was performed with a scanning electron microscope at the Faculty Laboratory of Phase, Structural, Textural and Geochemical Research (AGH).

The thin sections are housed at DGGG AGH in the author's collection (Anna Waśkowska). The micropaleontological samples have been deposited at the European Foraminiferal Reference Centre of the Micropress Europe Micropaleontological Foundation (al. A. Mickiewicza 30, Krakow, Poland).

## RESULTS

To determine sedimentological conditions of the olistostrome sequence and the age of the Popeli Beds, comprehensive structure, lithological, and micropaleontological analyses were conducted.

### Lithological development

The olistostrome sequence could be subdivided into the six complexes. The deposits underlying and overlying the olistostrome sequence are also shortly described.

#### *Deposits underlying the olistostrome sequence*

The Popeli olistostrome sequence overlies a package of thin- to medium-bedded sandstones and mudstones of the Bystrytsia Formation that form a thin cover on the Vyhoda Sandstones (Fig. 3A). The Vyhoda Sandstones are fine- and medium-grained, medium-bedded. Within them, a few metres thick complex of thick-bedded, coarse- to very coarse-grained sandstones and conglomeratic sandstones occurs.

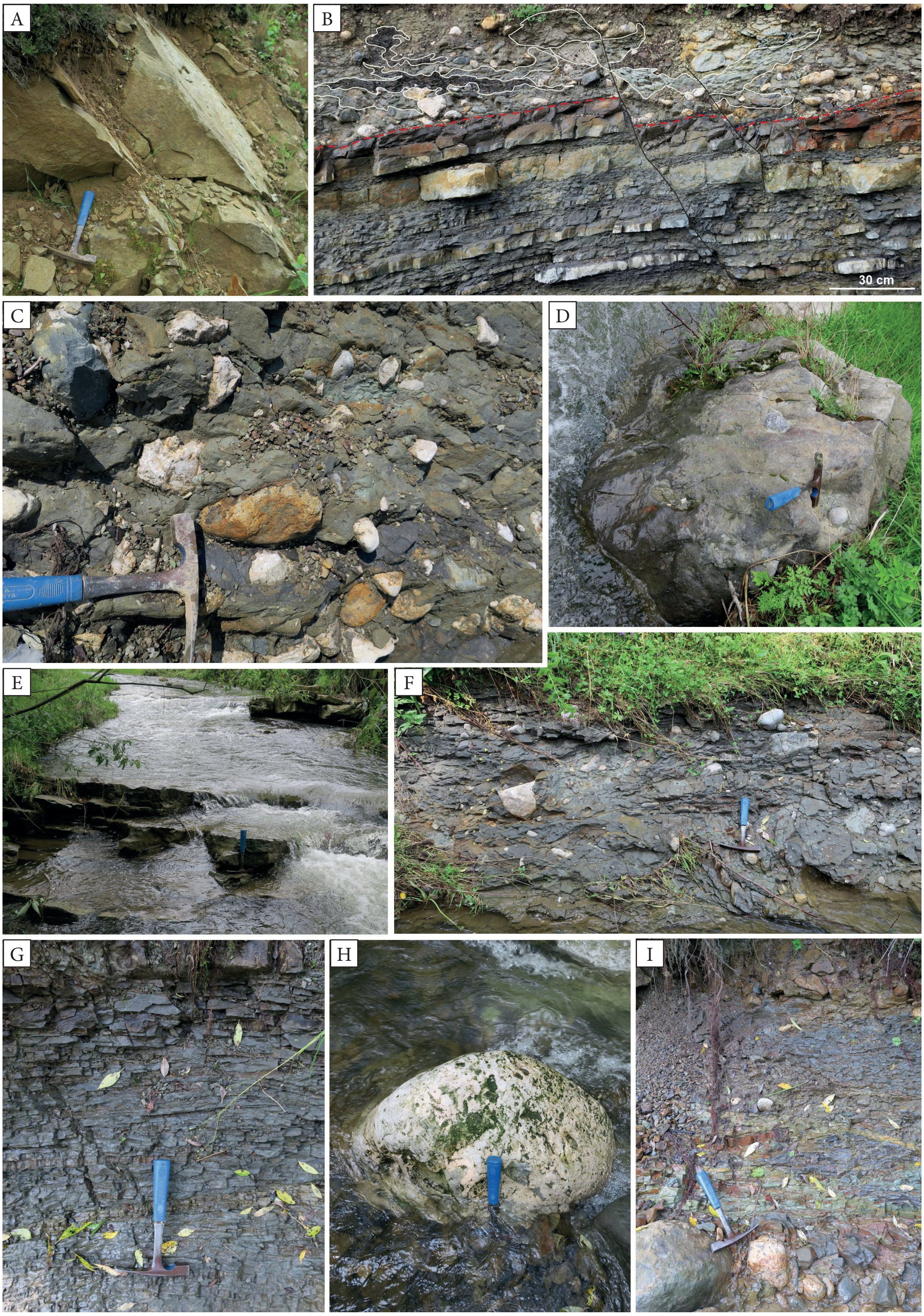


***Fig. 3.*** *Development of the studied deposits: A) the uppermost part of the Vyhoda sandstones; B) the lowermost part of the olistostrome complex; the basal erosional surface is marked by a red line, below the erosional surface are the underlying thin-bedded turbidites of the Bystrica-type, above the erosional surface is the lowermost part of the olistostrome gravelly mudstone (bigger marly and shaly clasts are marked with white line, the faults are marked with a black line); C) the lower part of the gravelly mudstones in Complex A; D) the limestone block with gravels in the surface caverns; E) sandy complex overlying the Complex A; F) the gravelly mudstone in Complex B; G) fragment of large mudstone clast in Complex E; H) the limestone block in Complex C; I) mudstone above the Complex D. The length of the hammer is 33 cm*

Directly beneath the olistostrome mudstones, from a few to over 40 cm in thickness, are dark grey, locally almost black, calcareous, with numerous visible trace fossils highlighted by dark patches. The sandstones, 10–20 cm in thickness, occasionally up to 30 cm, are grey, fine-grained, quarzitic with a glauconite admixture, weakly calcareous or non-calcareous with parallel and cross-lamination. The medium-bedded sandstones are massive at their bases, with cross and parallel laminations in the upper parts of beds.

### *Olistostrome sequence*

The olistostrome sequence is separated from the underlying deposits by an erosional sharp and irregular surface, that discordantly cuts into underlying thin-bedded turbidites (Fig. 3B). The sequence of olistostromes comprises at least six individual complexes (A–F), forming sheet-like bodies, some of which are separated by undisturbed deep-water deposits (Fig. 5).

***Complex A.*** The lowest olistostrome complex is 50 m thick. It begins with a 50 cm thick coarse-grained gravelly marl with numerous oversized gravelly clasts, including pebbles, cobbles, and occasionally boulders of various sedimentary rocks, up to 35 cm in diameter (Fig. 3C).

Well or moderately rounded clasts of grey- and cream-coloured limestones and cherty limestones are most common within the very large clasts. Subrounded clasts of grey and greyish green, relatively soft massive marls or marls with poorly defined parallel lamination are also present. Less common are clasts of grey sandstones and dark siliceous rocks, with occasional plastically deformed elongated green marls, up to 60 cm long. These very large clasts show a directional orientation parallel following the erosion surface. The greatest concentration of clasts occurs near the base. The matrix consists of grey and grey and grey-green gravelly marl or sandy marl. The vertical concentration of gravel changes irregularly as does the distribution of the green marl. Generally, the abundance and size of the gravel and content of the grey marly matrix decreases in an upwards direction.

After 50 cm, the deposits grade into a thick complex of coarse and medium gravelly marl with granules and pebbles, mainly well-rounded limestones, and rarely coarser clasts. The number of psephitic grains varies vertically and shows an irregular distribution. In the upper part of the sequence, elongated, plastically deformed clasts of green marl, as well as streaks and smears of green marl, occur within the matrix. Their orientation is consistent with the gravel clast distribution at the base of Complex A. At the top of the complex lies a well-rounded limestone block over 1 m in diameter (Fig. 3D). On its surface, pockets occur filled with green marl containing well rounded limestone pebbles and cobbles up to 10 cm in size, along with occasional phyllite clasts.

The gravelly marl is overlain by a 3 m thick package of glauconitic sandstone, beginning with fine-grained, parallel- and cross-laminated, medium-bedded sandstones and grading upward to medium-grained, very thick-bedded sandstones (Fig. 3E). Locally, these sandstones are coarse-grained or conglomeratic with scattered angular limestone granules. Trace fossils occur along the bedding planes. Above the sandstone package is a 10-metre-thick package of grey mudstones with bioturbations, with the alternation of very thin-bedded, parallel laminated grey sandstones in the upper part.

***Complex B.*** This complex begins with an erosion surface and documents the occurrence of another package of coarse-grained gravelly marl with very large gravelly clasts of pebbles and cobbles, occasionally boulders, with a total thickness of about 55 m (Fig. 3F). The concentration of pebbles and boulders is highest in the lower 30 cm of the package, with the coarser components becoming more dispersed higher up. This material is embedded in matrix consisting of dark grey, coarse-grained gravelly marlstone and sandy marlstone. In matrix the marl content decreases gradually upwards.

The very large clasts are dominated by well and very well-rounded limestones, ranging in size from 1 to 20 cm, occasionally up to 45 cm. In addition, rare poorly rounded or angular clasts of calcareous sandstones and, less frequently, conglomerates are present, together with sporadic clasts of greenish parallel laminated pelitic limestones a few centimetres in size. Small clasts of light green marl (macroscopically similar to the marl in Complex A) are also observed.

Upwards in the section, the number and size of very large gravel clasts varies. Approximately 5 m above the base, several centimetre-sized clasts of grey calcareous mudstones and, less frequently, light green marls appear in the matrix, along with limestone pebbles and angular sandstones. These are plastically deformed, stretched, or smeared within the matrix. The very large gravelly clasts show a linear orientation consistent with the stretching direction of the marl smears more or less parallel to the bedding. Approximately 10 m from the base, within the coarse-grained gravelly marl, isolated packages of grey calcareous mudstones up to 120 cm thick are observed. Scattered thin and medium layers (up to 30 cm thick) of mudstones with thin- and very thin-bedded fine-grained sandstones are present. In the uppermost part there are single large (50–120 cm), very well-rounded limestone blocks with burrow structures filled with grey marl.

An approximately 7-metre-thick undeformed package of grey mudstone and marl deposits overlies the coarse-grained gravelly marl (Fig. 3G). These mudstones are macroscopically similar to mudstones lying below Complex B.

***Complex C.*** The base of the next complex is marked by an erosion surface that is capped with a 10-metre-thick gravelly mudstone with pebble- and cobble-sized clasts. The very large gravelly clasts are dominated by angular or poorly rounded glauconitic and quarzitic sandstone clasts, ranging in size from a few centimetres to tens of centimetres. There are fewer rounded limestone pebbles, cobbles and blocks (Fig. 3H), and unrounded black cherts occur sporadically. The distribution of gravels in the mudstone changes irregularly upwards, with smaller clasts generally occuring in the upper part. A 3-metre-thick package of grey mudstones with bioturbations and rare thin- and very thin-bedded very fine-grained sandstones overlies the olistostrome Complex C.

***Complex D.*** This complex consists of 8 m of gravelly mudstone with very large gravelly clasts of pebbles and cobbles. Its structural and textural characteristic is very similar to analogous deposits in Complex C.

A nearly 20-metre-thick package of greyish-green, non-calcareous mudstones was found above Complex D (Fig. 3I). In its upper part there are rare very thin beds of silty mudstone and very fine-grained sandstone, showing flow-like plastic deformations.

***Complex E.*** The complex begins with an erosion surface overlain by 12 m of matrix supported conglomerate, which laterally changes at a distance of 8 m to gravelly mudstone with very large gravelly clasts of pebbles and cobbles (Fig. 4A). The well-rounded limestones dominate, occasionally reaching sizes of several tens of centimetres (Fig. 4B), together with angular and subangular clasts of massive grey mudstones. Angular clasts of sandstones occur in a smaller quantity. Within this complex a 150-centimetre-thick clast of grey mudstone extending over several metres, was documented. In the upper part of the complex, the matrix contains clasts of grey and grey-green sandy mudstones. Upwards in the Complex E, the quantity and size of pebbles and cobbles decreases, and the deposits change to gravelly mudstone. In the uppermost part, folded mudstone clasts up to 1.5 m thick are found, as well as deformed mudstone packages with thin sandstone beds forming plane deformation sheets. There are also 1.5-metre-long blocks of medium- to thick-bedded sandstone (occasionally occurring in packages of 2–3 beds), smaller amounts of plastically deformed green soft marls, and individual limestone blocks measuring 0.7 to 1.7 m in diameter.

***Complex F.*** Overlying the gravelly mudstone of Complex E is a 20-metre-thick sequence composed of deformed sheets of bedded deposits. They contain grey calcareous and non-calcareous mudstones ranging in thickness from a few centimetres to 40 cm. Some mudstone beds contain intercalations of gravelly mudstones or gravelly sandstones, which can be several centimetres thick. Very large gravelly clasts of limestones and sandstones are relatively rare. Mudstones are interbedded with rare thin- to very thin beds of grey, parallel and cross-laminated quarzitic sandstones.


***Fig. 4.*** *Development of the studied deposits: A) gravelly mudstones with numerous sandstone clast in Complex E; B) the limestone blocks from Complex E; C), D) deformed turbidites in Complex F; E) the thin-bedded turbidites above the Complex F; F) the mudstones with crushed mollusc shells. The length of the hammer is 33 cm*

The mudstone-sandstone packages occur in several-metre-thick sheets that are deformed (Fig. 4 C, D). Between the mudstone-sandstone bedded packages, there is a minor presence of mudstone matrix, as well as small sandstone clasts, and very rare, very large clasts of well-rounded limestones.

## *Deposits overlying the olistostrome sequence*

Above the olistostrome sequence, there is a sequence of thin- and medium-bedded mudstones and sandstones. It consists of grey and beige calcareous mudstones, ranging in thickness from a few to 30 cm, interbedded with thin- and medium-bedded, parallel and cross-laminated quarzitic sandstones. Rare, isolated gravel-sized clasts of limestone and sandstone occur in the mudstones. These deposits grade into non-calcareous mudstones with a few 1–2–centimetre-thick layers of parallel laminated quarzitic sandstones. Within these deposits, there is a several-metre-thick complex of massive dark grey mudstone containing randomly scattered fragments of thin-shelled molluscs up to 0.5 cm in size (Fig. 4F).

Higher up, the mudstones with sandstone layers grade into grey and beige marls with occasional very thin layers of calcareous sandstone. The Menilite Formation occurs above this.

## Biostratigraphic analysis of foraminifera

For biostratigraphic analysis, 26 samples of soft mudstones and marls were washed to obtain foraminifera tests. A total of 5 samples were taken from the deposits below the olistostrome sequence, 18 samples were taken from the olistostrome deposits, including matrix, cobbles/boulders and deposits between the olistostrome complexes, and 4 samples were collected from deposits overlying the displaced complex (Fig. 5, Table 1).

The samples varied considerably in both abundance and taxonomic diversity. Most of the assemblages included biostratigraphically significant forms (Fig. 6, Table 1 and Suppl. 1 – available online as a supplementary file to the article). Some samples contained forms representing a variety of age groups, so the youngest forms were used to estimate ages. The general state of preservation of foraminifera is good. In most specimens, the taxonomic features were sufficiently preserved to allow identification at the species or at least genus level. Among the foraminiferal assemblages, redeposited Cretaceous planktonic foraminifera stood out, showing clear signs of abrasion.

### *Deposits underlying the olistostrome sequence*

The upper parts of the Vyhoda Beds (sample PA 1) and the top 1.5-metre-thick part of the thin-bedded mudstone and sandstone complex immediately below the olistostrome (samples PA 6–9) were sampled (Fig. 5, Table 1). A sample taken from the Vyhoda Beds contains a deep-water agglutinated assemblage. That type of assemblage is typical for lower slope and abyssal plain environments and is common in Outer Carpathian deposits (Olszewska 1980). Its early Bartonian age was based on the presence of *Ammodiscus latus* (Grzybowski) and acmes of *Reticulophragmium amplectens* (Grzybowski) as well as acme S*piroplectammina spectabilis* (Grzybowski) (Morgiel & Olszewska 1981, Olszewska 1997, Kaminski & Gradstein 2005, Waśkowska 2021) (Table 1, Suppl. 1).

Thin-bedded deposits between Vyhoda Beds and the base of the olistostrome sequence contain a very poor paleontological record. Only a few planktonic foraminifera were found, except for sample PA 8 in which a poor deep-water assemblage of agglutinated foraminifera was recovered. The age was based on planktonic foraminifera. The sample PA7 provided the most precise data. The presence of *Turborotalia cerroazulensis* (Cole), *Globigerinatheka index* (Finlay), *Subbotina angiporoides* (Hornibrook), *Subbotina corpulenta* (Subbotina) indicates the late Lutetian – early Priabonian interval (Berggren & Pearson 2005, Pearson et al. 2006, Premoli Silva et al. 2008, Rögl & Egger 2012, BouDagher-Fadel 2015) (Table 1, Suppl. 1).

### *Olistostrome sequence*

***Matrix.*** Seven samples were collected from the olistostrome matrix within complexes A, B, C, E, and F (Fig. 5). They contain foraminiferal assemblages dominated by benthic foraminifera, both agglutinated and calcareous with a co-occurrence of planktonic foraminifera (Table 1). Biostratigraphic interpretations were based on the presence of planktonic foraminifera, combined with key agglutinated taxa.

The most precise biostratigraphic information was obtained from sample PA 11, collected from Complex A. It contained *Subbotina gortanii* (Borsetti), a species known from the fossil record since the early Bartonian (Pearson et al. 2006), occurring together with an assemblage of agglutinated foraminifera dominated by the *Reticulophragmium amplectens* (Grzybowski) (Table 1, Suppl. 1). In the Outer Carpathians, the acme of this species persists up to the early Bartonian (Waśkowska 2021). This coincidence strongly supports an early Bartonian age for the matrix in the lowermost part of the olistostrome sequence.

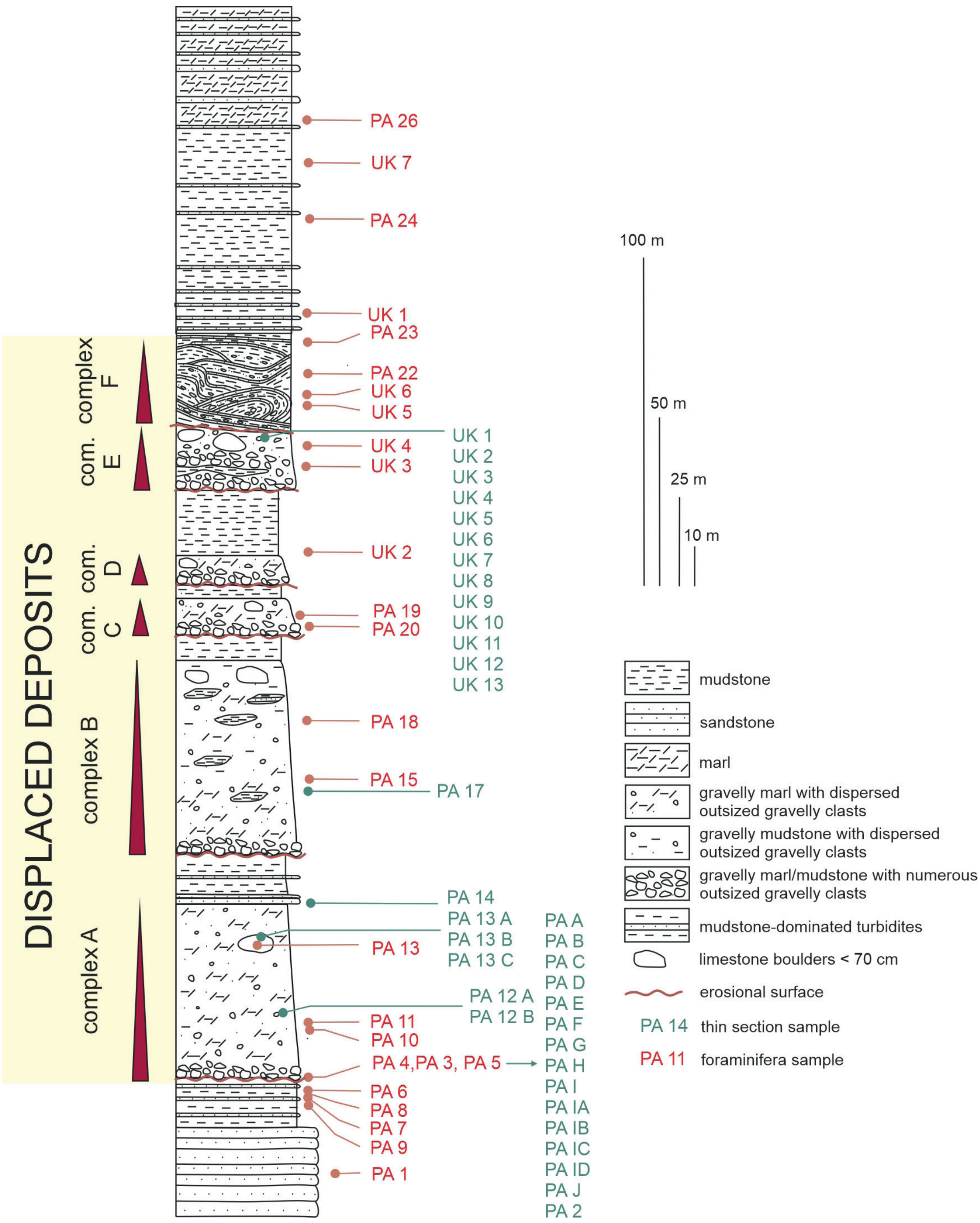


***Fig. 5.*** *Lithological log of the studied section with location of the analysed samples*

***Table 1***
*Stratigraphy and foraminiferal age indicators of olistostrome sequence (see also Figs. 5, 6)*

| Sample position | Sample number | Deposits | Assemblage | Biostratigraphic indicators | Age* |
|---|---|---|---|---|---|
| Vyhoda Beds | PA 1 | shale intercalation beetwen sandstones | agglutinated assemblage | *Ammodiscus latus* (Grzybowski), *Reticulophragmium amplectens* (Grzybowski) – acme, *Haplophragmoides walteri* (Grzybowski) – acme | early Bartonian |
| Bystrytsia Formation | PA 6 | thin-bedded deposits below olistostrome | poor planktic assemblage | *Dipsidripella danvillensis* (Howe et Wallace), *Subbotina corpulenta* (Subbotina) | E9–O4 (late Lutetian – Oligocene) |
| | PA 7 | thin-bedded deposits below olistostrome | planktic assemblage | *Turborotalia cerroazulensis* (Cole), *Globigerinatheka index* (Finlay), *Subbotina angiporoides* (Hornibrook), *Subbotina corpulenta* (Subbotina) | E11–E14 (late Lutetian – early Priabonian) |
| | PA 8 | thin-bedded deposits below olistostrome | poor agglutinated assemblage | *Ammomarginulina aubertae* Gradstein et Kaminski | Maastrichtian – Miocene |
| | PA 9 | thin-bedded deposits below olistostrome | poor assemblage | *Parasubbotina varianta* (Subbotina) | P1–O1 (Paleocene – Oligocene) |
| Complex A | PA 10 | light green gravelly marl (matrix) | diverse calcareous benthic and planktic assemblage | *Turborotalia increbescens* (Bandy) | E13–O1 (Bartonian – early Oligocene) |
| | PA 11 | greyish-green gravelly marl (matrix) | diverse calcareous benthic and planktic assemblage | *Subbotina gortanii* (Borsetti), *Reticulophragmium amplectens* (Grzybowski) – acme | early Bartonian |
| | PA 13 | light green marl (pocket fillings in 1-metre block) | diverse foraminiferal assemblage | *Globigerina officinalis* Subbotina, *Globoturborotalita* cf. *martini* (Blow et Banner), *Spiroplectammina spectabilis* (Grzybowski) – acme | E10 (late Lutetian – early Priabonian) |
| | PA 3 | light green marl (clast) | diverse planktic assemblage with calcareous benthos | *Acarinina aspensis* (Colom) | E7 (late Ypresian – early Lutetian) |
| | PA 4 | greenish-blue marly mudstone (clast) | diverse foraminiferal assemblage | *Acarinina cf. boudreauxi* Fleisher, *Acarinina soldadoensis* (Brönnimann), *Acarinina bullbrooki* (Bolli), *Glomospira* – acme, *Saccamminoides carpathicus* Geroch | E7 (late Ypresian – early Lutetian) |

***Table 1** cont.*

| | | | | | |
|---|---|---|---|---|---|
| Complex A | PA 5 | beige-green marl (clast) | diverse calcareous assemblage with agglutinated foraminifera | *Acarinina interposita* Subbotina, *Acarinina angulosa* (Bolli), *Acarinina esnehensis* (Nakkady), *Acarinina soldadoensis* (Brönnimann), *Acarinina pseudotopilensis* Subbotina, *Acarinina pentacamerata* (Subbotina), *Acarinina cuneicamerata* (Blow), *Acarinina bullbrooki* (Bolli), *Acarinina collactea* (Finlay), *Subbotina corpulenta* (Subbotina), *Subbotina eocaena* (Guembeli), *Subbotina senni* (Beckmann), *Parasubbotina hagni* (Gohrbandt) | E6–E7 (late Ypresian – early Lutetian) |
| Complex B | PA 15 | sandy mudstone (matrix) | poor assemblage of benthic foraminifera | *Spiroplectammina spectabilis* (Grzybowski) – acme | middle Lutetian – early Priabonian |
| | PA 18 | grey marl (clast) | calcareous assemblage dominated by benthic forms | *Acarinina bullbrooki* (Bolli), *Globigerinatheka subconglobata* (Shutskaya), *Globigerina officinalis* Subbotina, *Acarinina medizzai* (Toumarkine et Bolli) | late Lutetian – early Bartonian (E9–E11) |
| Complex C | PA 19 | gravelly mudstone (matrix) | agglutinated assemblages with calcareous benthos and rare plankton | *Globigerinatheka mexicana* (Cushman), *Globigerinatheka index* (Finlay), *Globigerina officinalis* Subbotina, *Dipsidripella danvillensis* (Howe et Wallace), *Ammodiscus latus* Grzybowski forma *ovoidalis* Waśkowska et Kaminski | E9–E14 (late Lutetian – early Priabonian) |
| | PA 20 | dark grey mudstone (clast) | agglutinated assemblage with calcareous benthos and rare plankton | *Dipsidripella danvillensis* (Howe et Wallace), *Spiroplectammina spectabilis* (Grzybowski) – acme, *Reticulophragmium amplectens* (Grzybowski) | late Lutetian (E9) – early Bartonian |
| Complex D/E | UK 2 | greyish-green mudstone (beetwen D and E complexes) | poor agglutinated assemblage | *Reticulophragmium amplectens* (Grzybowski) – acme | Lutetian – middle Bartonian |
| Complex E | UK 3 | greyish-green gravelly marly mudstone (matrix) | diverse benthic calcareous assemblage with planktic foraminifera | *Dipsidripella danvillensis* (Howe et Wallace), *Acarinina bullbrooki* (Bolli) | E9–E11 (late Lutetian – early Bartonian) |
| | UK 4 | greyish-greengravelly mudstone (matrix) | poor calcareous assemblage | *Dipsidripella danvillensis* (Howe et Wallace) | E9–O3 (late Lutetian – Oligocene) |

***Table 1** cont.*

| Sample position | Sample number | Deposits | Assemblage | Biostratigraphic indicators | Age* |
|---|---|---|---|---|---|
| Complex F | UK 5 | thin-bedded deposits (clast) | diverse agglutinated assemblage with calcareous benthos and rare plankton | *Glomospira* div. sp. – acme, *Recurvoides–Thalmannammina* – acme, *Reticulophragmium garcilassoi* (Frizzel) – acme, *Hormosina velascoensis* (Cushman), *Remesella varians* (Glaessner), *Anomalinoides danicus* (Brotzen) | late Paleocene |
| | UK 6 | thin-bedded deposits (clast) | poor agglutinated assemblage with rare plankton | *Globigerina officinalis* Subbotina | E10–M3 (late Lutetian – Miocene) |
| | PA 22 | thin-bedded deposits (clast) | diverse calcareous benthic and planktic assemblage with sparse agglutinated foraminifera | *Subbotina senni* (Beckmann), *Pseudohastigerina sharkriverensis* Berggren et Olsson, *Subbotina angiporoides* (Hornibrook), *Turborotalia cerroazulensis* (Cole), *Acarinina bullbrooki* (Bolli), *Reticulophragmium amplectens* (Grzybowski), *Spiroplectammina spectabilis* (Grzybowski) – acme | E11–E13 (late Lutetian – Bartonian) |
| | PA 23 | sandy mudstone (matrix) | poor agglutinated assemblage with rare calcareous foraminifera | *Subbotina eocaena* (Guembel) | E6– O7 (late Ypresian – Oligocene) |
| Deposits over olistostrome | UK 7 | thin-bedded mudstones and sandstones | poor planktic assemblage | *Pseudohastigerina naguevichiensis* (Mjatliuk) | E15–O1 (Priabonian – early Rupelian) |
| | UK 1 | thin-bedded mudstones and sandstones | barren | – | – |
| | PA 24 | thin-bedded mudstones and sandstones | barren | – | – |
| | PA 26 | mudstones with gastropods | barren | – | – |

*Age after Morgiel & Olszewska (1980), Geroch & Nowak (1984), Olszewska (1997), Berggren & Pearson (2005), Kaminski & Gradstein (2005), Olszewska et al. (2006), Pearson et al. (2006), Premoli Silva et al. (2008), Rögl & Egger (2012), BouDagher-Fadel (2015), Young et al. (2017), Waśkowska (2021).


***Fig. 6.*** *Selected foraminifera of the studied samples: 1 – Ammodiscus latus Grzybowski (sample (s.) PA 1); 2 – Reticulophragmium amplectens (Grzybowski) (s. PA 1); 3 – Haplophragmoides sp. (s. PA 4); 4 – Saccamminoides carpathicus Geroch (s. PA 4); 5 – Ammomarginulina aubertae Gradstein et Kaminski (s. PA 8); 6a–c – Subbotina gortanii (Borsetti) (s. PA 5); 7a–c – Subbotina corpulenta (Subbotina) (s. PA 3); 8a–c – Parasubbotina eoclava Coxall, Huber et Pearson (s. PA 4); 9a–c – Turborotalia increbescens (Bandy) (s. PA 10); 10a–c – Acarinina boudreauxi Fleisher (s. PA 3); 11a–c – Acarinina pseudotopilensis Subbotina (s. PA 3); 12a–c – Acarinina interposita Subbotina (s. PA 3); 13a–b – Acarinina aspensis (Colom) (s. PA 3)*

Sample UK 3 from Complex E contained a planktonic foraminiferal assemblage with *Dipsidripella danvillensis* (Howe et Wallace) – a species documented in the fossil record from the late Lutetian (Pearson et al. 2006) – as the youngest component (Table 1, Suppl. 1). In this sample, *Dipsidripella danvillensis* (Howe et Wallace) co-occurs with *Acarinina bullbrooki* (Bolli), a species known from the Lutetian to the early Bartonian (Berggren & Pearson 2005, BouDagher-Fadel 2015, Pearson et al. 2006, Premoli Silva et al. 2008) (Table 1, Suppl. 1). Assuming that both taxa are autochthonous (age-equivalent to the matrix), their presence suggests the early Bartonian as an upper age boundary for the matrix. It is consistent with age determinations from the lower and upper parts of the studied section.

Other samples from the matrix yielded less precise age determinations, indicating broader time intervals. However, all results include the early Bartonian within their range. In the samples from E and F complexes the deep-water Late Cretaceous – Paleocene forms occur as a redeposited component, while in the samples PA 10, PA 11, UK 3, UK 4 reposited Late Cretaceous planktonic foraminifera appear (Suppl. 1).

***Clasts.*** In total, nine samples were collected from marl and mudstone clasts derived from cobbles and boulders, as well as from deformed packages of thin-bedded mudstones and sandstones at various levels within the olistostrome sequence (Fig. 5, Table 1). The marly clasts contained mixed calcareous-agglutinated foraminiferal assemblages, dominated either by calcareous benthic or by planktonic foraminifera (Table 1). These assemblages were characterised by very high taxonomic diversity. Both their composition and the structure of the benthic foraminiferal assemblages are indicative of outer shelf to upper slope environments (e.g., Olszewska 1984, Kaminski & Gradstein 2005). Assemblages typical of deeper-water settings occurred in mudstone clasts, showing a dominance of either calcareous or agglutinated benthic foraminifera, with variable proportions of planktonic forms. The age of the samples was determined based on the occurrence of planktonic foraminifera, supported by diagnostic agglutinated taxa.

Clasts from Complex A (samples PA 3, PA 4, PA 5) contained foraminiferal assemblages indicative of a late Ypresian to early Lutetian age (E6 or E6–E7). The age of PA 3 and PA 4 samples was based on the occurrence of *Acarinina* species, including: *Acarinina aspensis* (Colom), *Acarinina* cf. *boudreauxi* Fleisher, *Acarinina soldadoensis* (Brönnimann), *Acarinina bullbrooki* (Bolli), and *Saccamminoides carpathicus* Geroch (Geroch & Nowak 1984, Pearson et al. 2006, Waśkowska 2011). Sample PA 5 yielded a more diverse assemblage of *Acarinina* and *Subbotina* species, including taxa important for age estimations: *Acarinina interposita* Subbotina, *Acarinina angulosa* (Bolli), *Acarinina esnehensis* (Nakkady), *Acarinina soldadoensis* (Brönnimann), *Acarinina pseudotopilensis* Subbotina, *Acarinina pentacamerata* (Subbotina), *Acarinina cuneicamerata* (Blow), *Acarinina bullbrooki* (Bolli), *Acarinina collactea* (Finlay), *Subbotina corpulenta* (Subbotina), *Subbotina eocaena* (Guembel), *Subbotina senni* (Beckmann), and *Parasubbotina hagni* (Gohrbandt) (Table 1). All of these samples also contain redeposited Late Cretaceous planktonic foraminifera in addition to Paleogene planktonic assemblages (Suppl. 1).

The infill of a cavity on the surface of a limestone block from the upper part of Complex A, marly clasts from Complex B, and silty clasts from Complex C contain younger foraminiferal assemblages. In sample PA 18, the co-occurrence of *Acarinina bullbrooki* (Bolli), *Globigerinatheka subconglobata* (Shutskaya), *Globigerina officinalis* Subbotina, and *Acarinina medizzai* (Toumarkine et Bolli) (Table 1, Suppl. 1) indicates an age ranging from the late Lutetian to early Bartonian (Pearson et al. 2006). In sample PA 13, the presence of *Globigerina officinalis* Subbotina, *Globoturborotalita* cf. martini (Blow et Banner), and the *Spiroplectammina spectabilis* (Grzybowski) acme (Table 1, Suppl. 1) points to an age interval of late Lutetian to early Priabonian (Pearson et al. 2006, Waśkowska 2021). In sample PA 20, the assemblage is characterised by the acmes of *Spiroplectammina spectabilis* (Grzybowski) and *Reticulophragmium amplectens* (Grzybowski), along with *Dipsidripella danvillensis* (Howe et Wallace) (Table 1, Suppl. 1), which together indicate a late Lutetian to early Bartonian age (Bieda et al. 1967, Olszewska 1980, 1997, Pearson et al. 2006, Waśkowska 2014, 2021).

Three samples were collected from a package of mudstones intercalated by thin-bedded sandstones of Complex F (Fig. 5, Table 1). Sample PA 22 contained *Subbotina senni* (Beckmann), *Pseudohastigerina sharkriverensis* Berggren et Olsson, *Subbotina angiporoides* (Hornibrook), *Turborotalia cerroazulensis* (Cole), and *Acarinina bullbrooki* (Bolli), along with relatively abundant *Reticulophragmium amplectens* (Grzybowski) and *Spiroplectammina spectabilis* (Grzybowski) (Table 1, Suppl. 1), which are interpreted as occurring within their acme. This assemblage suggests the late Lutetian to Bartonian age (Pearson et al. 2006, Geroch & Nowak 1984, Olszewska 1997, Waśkowska 2015, 2021). Sample UK 6 contained *Globigerina officinalis* Subbotina (Table 1, Suppl. 1), a species known from the late Lutetian to Miocene (Pearson et al. 2006), indicating that the sample is not older than the late Lutetian. In contrast, sample UK 5 yielded a deep-water benthic foraminiferal assemblage typical of the Outer Carpathian late Paleocene, including key indicator species such as *Reticulophragmium garcilassoi* (Frizzell) – acme, *Hormosina velascoensis* (Cushman), *Remesella varians* (Glaessner), *Anomalinoides danicus* (Brotzen), and *Glomospira* div. sp. – acme, as well as a characteristic *Recurvoides–Thalmannammina* acme (Table 1, Suppl. 1).

Sample UK 2 was collected from mudstone occurring between the olistostrome complexes D and E (Fig. 5, Table 1). It contained a deep-water foraminiferal assemblage with the *Reticulophragmium amplectens* (Grzybowski) acme, which in the Outer Carpathians is typical of Lutetian to middle Bartonian deposits (e.g., Morgiel & Olszewska 1981, Olszewska et al. 1996, Kaminski & Gradstein 2005, Waśkowska 2015, 2021) (Table 1, Suppl. 1).

### *Deposits overlying the olistostrome complex*

Four samples were taken above of the olistostrome sequence (Fig. 5). Sample UK 7 contains sparse Eocene – Oligocene planktonic foraminifera, including *Chiloguembelina cubensis* (Palmer) and *Pseudohastigerina naguevichiensis* (Mjatliuk) (Table 1, Suppl. 1). *Pseudohastigerina naguevichiensis* (Mjatliuk) appears in the Priabonian, within Zone E15 and lasted to early Rupelian Zone O1 (Pearson et al. 2006), and its presence suggests that the sample is not older than the Priabonian and not younger than early Oligocene.

## Microscopic analysis of clasts and matrix samples

A total of 35 thin sections were prepared and examined, within which 33 samples were from very large gravelly clasts of gravelly marl from complexes A and E (Suppl. 2 – available online as a supplementary file to the article, Fig. 7). Additionally, one sample from matrix of Complex B (PA 17), and one sample from sandstone overlying the Complex A were analysed (PA 14). The study included 19 limestone clasts, 12 sandstone clasts, and 4 other rocks (Suppl. 2).

Several groups of clasts can be distinguished (from the most to the less abundant):

1) Clasts of the Štramberk-type limestones (samples PA A, PA F, PA H, PA I, PA ID, PA 12B, PA 13B, PA 13C, UK 3, UK 4, UK 6, UK 11) are common thorough the whole section.
   They represent the most common exotic limestones in the Outer Carpathians, Tithonian to Berriasian in age (e.g., Hoffmann et al. 2021). Their characteristic microfacies can be easily identified even in small clasts. In the studied samples, they are represented by typical microfacies as coral boundstone, microbial-sponge boundstone or intraclastic-bioclastic grainstone/rudstone (reef talus) (Fig. 7A, B, Suppl. 2). Fossils and microfossils typical of the sha low-water, carbonate platform and its slope are numerous (Table 2). Foraminifera are represented by calcareous benthic forms, and agglutinated forms that built tests of calcareous material. Calcareous dinocysts typical of this time interval were also observed. It was possible to determine the age of some samples more precisely, mainly based on chitinoidellids of the subfamily Bonetinae (indicating the late Tithonian Boneti Subzone of the Chitinoidella Zone according to the calpionellid zonation) and calpionellids (especially *Crassicollaria intermedia* (Durand Delga) typical of the late Tithonian Crassicollaria Zone). The presence of the planktic crinoid *Saccocoma* sp. in some samples suggests an age no younger than the Tithonian. Occasionally, voids filled by younger deposits with planktic foraminifera can be observed in these limestones.

2) Clasts of the quartz-glauconitic calcareous sandstones (PA 2, PA 12A, UK 2, UK 5, UK 7, UK 10, UK 12, UK 13) (Fig. 7C) are more abundant in the upper part of the section. They include lithoclasts, mainly of Štramberk-type limestones. The sandstones are relatively rich in bioclasts, yielding benthic calcareous foraminifera, echinoderm elements, calcareous dinocysts, ostracod carapaces, fragments of bivalve shells, sponge spicules. Assemblages of planktic foraminifera, including *Subbotina* sp., indicate their Paleogene age. A similar lithology is represented by clast PA C, but the assemblage of the planktic foraminifera rather suggests Late Cretaceous age. The sandstones are very similar to samples PA 14 and PA 17, which represent the sandstones occurring above the Complex A, and the matrix of Complex B.
3) Clasts of more or less silicified rocks, rich in sponge spicules, occur in the Complex A (PA B) and E (UK 8, UK 9):
   – laminated foraminiferal-radiolarian wackestone (PA B) (Fig. 7A, D), with sponge spicules and other bioclasts; late Early Cretaceous – early Late Cretaceous in age, based on planktic foraminifera;
   – calcareous gaize (UK 8), rich in sponge spicules and less abundant other poorly preserved bioclasts; age indeterminate;
   – sandy spiculite packstone (UK 9) (Fig. 7E), with calcareous benthic and planktic foraminifers and other bioclasts; the presence of biserial planktic foraminifers suggest age no older than late Early Cretaceous.
4) Lithologies observed only in the individual clasts, in the lower part of the section, are represented by:
   – bioclastic wackestone/packstone with nummulites (PA E from the Complex A) (Fig. 7F), other bioclasts: smaller benthic and planktic foraminifera, fragments of red coralline algae and bryozoan colonies, echinoderm elements, serpulid tubes, ostracod carapaces; facies typical of the Paleogene (e.g., Bieda 1968);
   – sandy, partly silicified crinoid limestone (PA 13A from the Complex A), with fragments of bryozoan colonies, calcareous benthic foraminifera (including miliolids, neotrocholinids), siliciclastic grains and small clast with *Crassicollaria* sp.; based on the age of the clast with *Crassicollaria* (latest Tithonian – early Berriasian) and foraminifera, the age of the rock can be estimated as Early Cretaceous.
5) Rocks of indeterminate age, mostly barren of fossils, occur mainly in the Complex A: microsparitic limestone (PA D), fine-grained sandstone with siliceous cement (PA G), laminated siliceous, muddy marl (PA IA), laminated mudstone with ferruginous cement (PA IB), siltstone with siliceous cement (PA IC), fine-grained quartz sandstone (PA J, UK 1).

***Table 2***
*The most important fossils of the analysed clasts of the uppermost Jurassic – lowermost Cretaceous Štramberk-type limestones*

| Group | The most important fossils |
|---|---|
| Foraminifera | *Crescentiella morronensis* (Crescenti), *Mohlerina basiliensis* (Mohler), *Uvigerinammina uvigeriniformis* (Seibold et Seibold), *Neotrocholina valdensis* Reichel, *Neotrocholina molesta* (Gorbatchik), *Protopeneroplis striata* Weynschenk, *Protopeneroplis ultragranulata* (Gorbatchik), *Protomarssonella kummi* (Zedler), *Quinqueloculina multicostata primitiva* (Neagu), miliolids, spirillinids, trocholinids, neotrocholinids, and lagenids |
| Calcareous dinocysts | *Cadosina fusca fusca* Wanner, *Crustocadosina semiradiata semiradiata* (Wanner), *Colomisphaera tenuis* (Nagy), *Colomisphaera lapidosa* (Vogler), *Colomisphaera minutissima* sensu Borza, *Committosphaera pulla* (Borza), *Colomisphaera carpathica* (Borza) |
| Tintinnids | chitinoidellids of the subfamily Bonetinae, calpionellids (*Crassicollaria intermedia* (Durand Delga) |
| Microproblematica | *Thaumathoporella parvovesiculifera* (Raineri), *Labes atramentosa* Eliášová, and *Koskinobullina socialis* Cherchi et Schroeder |
| Other fossils | corals, calcified sponges, shells of bivalves, brachiopods and gastropods, echinoderm elements (including *Saccocoma* sp.), and bryozoan colonies, microencrusters, "Lithocodium" and Bacinella structures, calcimicrobes, dasycladacean green algae, serpulid tubes (including *Mercierella*? *dacica* Dragastan, *Terebella lapilloides* Münster), decapod *Carpathocancer triangulatus* (Misík, Soták et Ziegler), zoospores *Globochaete alpina* Lombard |

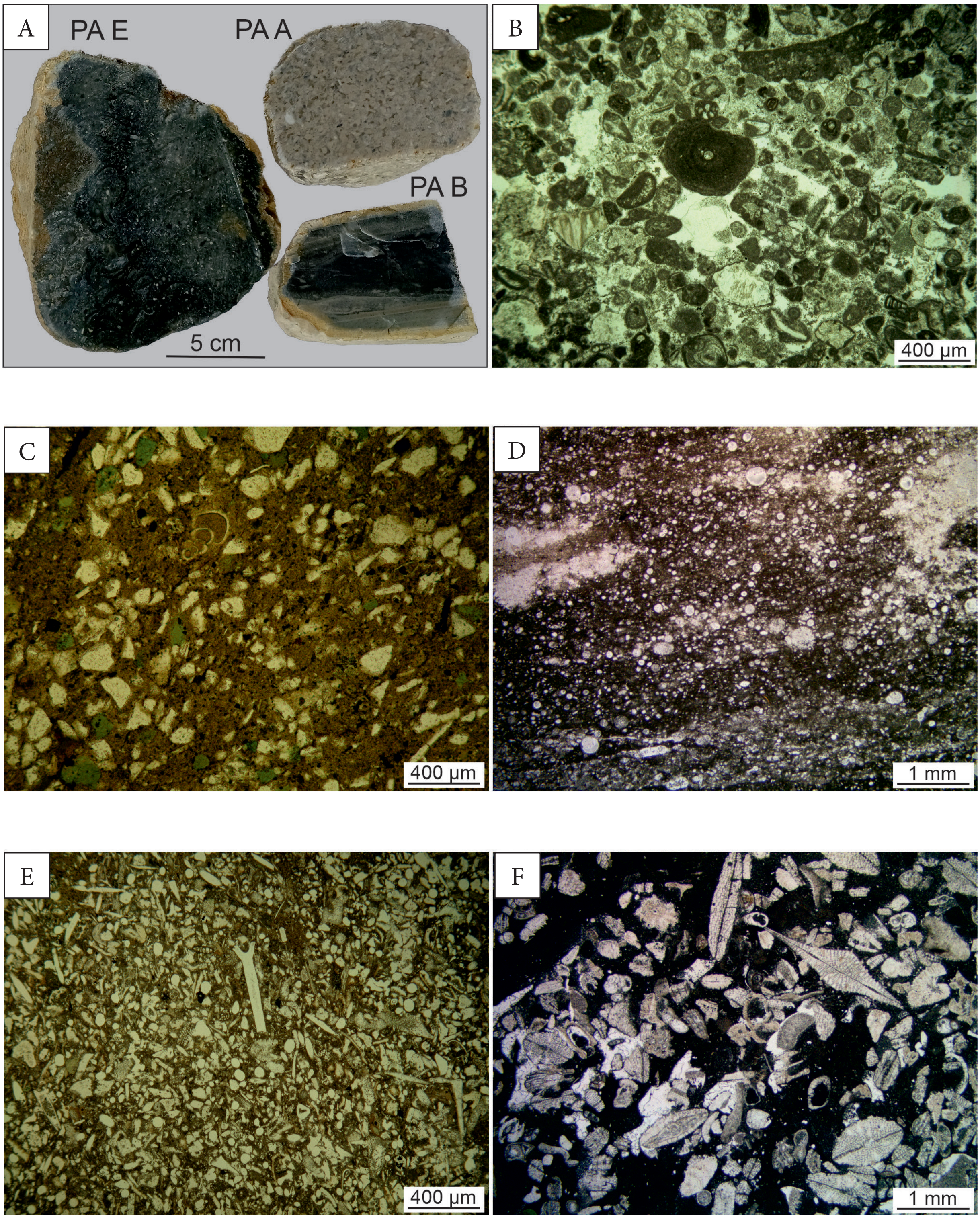


***Fig. 7.*** *Microscopic images of the chosen studied clasts: A) examples of exotic clasts: PA A – Štramberk-type limestone (intraclastic-coated rudstone), PA B – laminated foraminiferal-radiolarian wackestone, partly silicified, PA E – bioclastic wackestone/packstone with nummulites; B) Štramberk-type limestone (poorly washed bioclastic-coated grainstone), upper Tithonian (s. UK 6); C) fine-grained sandstone with calcareous-clay cement, Paleogene (s. UK 2); D) laminated radiolarian-foraminiferal wackestone, partly silicified, Upper Cretaceous (s. PA B); E) sandy spiculite packstone, not older than late Early Cretaceous (s. UK 9); F) bioclastic wackestone/packstone with nummulites, Paleogene (s. PA E)*

## INTERPRETATION

The Popeli olistostrome deposits in the Pobuk Syncline have a complex structure and consist of several separate complexes formed during individual depositional pulses, resulting in an olistostrome package with a total thickness of 220 m. The individual packages are separated by erosional surfaces. Most of them are also overlain by sedimentary successions formed as a result of turbiditic-type sedimentation, as indicated by the normal grading and the occurrence of sequences of sedimentary structures with planar- and cross-lamination (divisions of Bouma sequence). Within the olistostrome sequence, six displaced complexes of thicknesses varying from 8 to 50 m were identified (Fig. 5). These complexes are composed of gravelly marl or mudstone and contain varying amounts of gravel clasts (pebble- to boulder-sized) of different sizes redeposited as extra- and intrabasinal material. The exception is the last complex, which is dominated by deformation slabs composed of deep-marine mudstone-sandstone formations.

In the lower parts of the olistostrome sequence, the gravelly mudstones consist of deposits originating from the olistostromal niche located in lower shelf/uppermost slope position (Fig. 8). A marl matrix is present and soft rock clasts such as marls are very common. In the upper parts of the succession, the amount of material eroded from deeper parts of the basin along the displacement path increases progressively. The marl matrix gradually changes to a mud matrix, and the amount of mudstone clasts increases.

Observations of the lithological development indicate that each olistostrome complex has different characteristics, although a general decrease in contribution of limestone and marlstone clasts as well as marly matrix upwards the sequence are noted, together with a change in the lithological composition of the gravel clasts. An increase in the amount of sandstone and mudstone clasts, as well as sequences of alternating mudstones and sandstones, with a simultaneous reduction of marl clasts are observed. There is also a general decrease in the diversity of redeposited gravel further up the olistostrome sequence.

These gravels, particularly very large gravel clasts, are an important component of the olistostrome. They are both petrographically and stratigraphically diverse, including several groups of age-differentiated clasts:

1) Paleogene clasts, including fragments of sandstones, mudstones with deep-water foraminifera, upper Ypresian – lower Lutetian marls, upper Paleocene turbiditic sequences, and foraminiferal-algal limestones; these clasts are poorly rounded or angular;
2) uppermost Lower and Upper Cretaceous clasts, such as limestones and silicified limestones, and sandstones; the presence of sponge spicules is typical; these clasts are well to very well rounded;
3) clasts older than the late Early Cretaceous, represented almost exclusively by the *Štramberk*-type limestones; these clasts are very well rounded.

The material was delivered from outer zones of the Skyba Basin. The second and third groups represent older material that got into the source area through basement erosion or originated from outside the basin. They consist exclusively of hard rocks with a high degree of rounding, significantly exceeding that of the sandstones in the first group.

Changes in the composition of the olistostrome sequence, related to the disappearance of marl in the matrix and marly clasts, as well as the reduction in the amount of exotic material, are the result of a shift in the position of the niche. Firstly, it delivered mainly marly material to the inner part of the Skyba Basin from lower shelf/uppermost slope positions (Complex A). The upper complexes (B–C) contain a progressively increasing amount of deep-water material. It may be the result of the downslope migration of the landslide niche, from areas dominated by carbonate sedimentation to slope areas dominated by turbidite deposition (Fig. 8). Olistostrome complexes C–F are composed mostly of deep-water deposits, which suggests a slope position of the niche and probably its stabilisation. In the upper part of the olistostrome sequence, a stratigraphic inversion is preserved – displaced Eocene sequences are overlain by Paleocene material (Complex F). That reflects the progressive downcutting erosion, which subsequently involves older deposits.

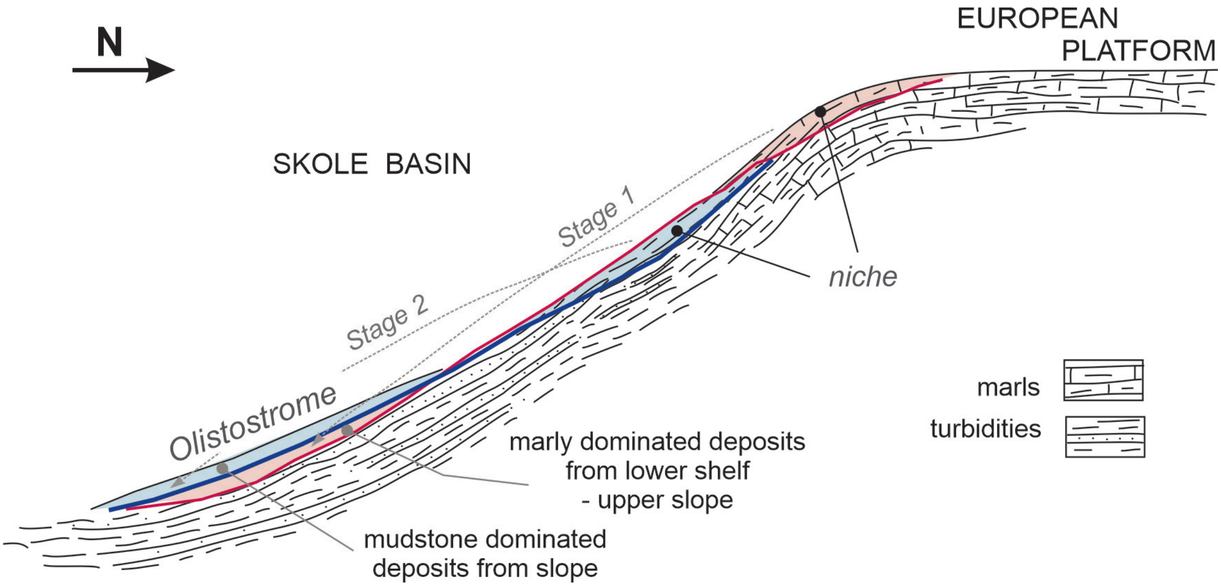


***Fig. 8.*** *Interpretation of the origin of the analysed Popeli Beds*

Shallow-water limestone pebbles are a constant element of the displaced complexes and occur as an individual component. In Complex F, two generations of limestone pebbles are present: large pebbles found in the mudstone matrix, which were incorporated during the formation of Complex F, and fine gravel-sized pebbles that are an integral component of mudstone-sandstone bedded deposits within the displaced complexes. The presence of carbonate grains within the turbidites indicates input from shallow-water areas during deposition and is not related to the genesis of the olistostrome. The paleotransport directions in the olistostrome and the overlying deposits are similar and indicate the delivery of material from N and NE.

The age of the displaced material was determined on the basis of foraminiferal biostratigraphy. Several factors were considered, including the age of the underlying and overlying formations, the matrix, and deposits occurring between the olistostrome complexes. Foraminiferal assemblages in the Vyhoda Sandstone, that is underlying the olistostrome, are of lower Bartonian age. Biostratigraphical analysis of the Bystrytsia Formation (directly below the olistostrome) did not allow for precise dating. Considering the biostratigraphic results from Bystrytsia Formation and data from the mudstone complex dividing C and D olistostrome complexes, it can be concluded that both were deposited in the lower – middle Bartonian interval. Dating the deposits above the olistostrome was challenging due to a very poor paleontological record. Most of the collected samples yielded no fossils. One sample taken 50 m above the top of the olistostrome complex contained an assemblage suggesting that this formation is no older than Priabonian.

## DISCUSSION

The deposits of the Outer Carpathians are characterised as being mostly turbidites formed by deposition from gravity currents (e.g., Dżułyński et al. 1959, Golonka & Picha 2006, Oszczypko et al. 2006, Golonka et al. 2019a). Within these formations, other genetic types of deposits are also present, including those resulting from a large-scale gravitational mass transport, known as olistostromes. Recent analyses indicate that such deposits are an important and widespread component of the sedimentary cover of the Outer Carpathians (Cieszkowski et al. 2012, Golonka et al. 2015, 2024 and references therein).

Olistostrome deposits are particularly abundant in the Silesian Domain of the Carpathians. This domain included the evolutionarily connected area of the Protosilesian, later Silesian, and Skyba basins, separated by the Subsilesian sedimentary zone, as well as the Menilite-Krosno Basin (Golonka et al. 2019b). Olistostrome deposits are of different ages and were formed during all developmental stages of the basins, from the initial to the terminal stages (Książkiewicz 1958, Dżułyński & Kotlarczyk 1965, Unrug 1968, Kulchytskyi 1977, Dżułyński et al. 1979, Malik & Olszewska 1984, Słomka 1986, 2001, Cieszkowski 1992, Cieszkowski et al. 2009, 2012, Ślączka et al. 2012, Waśkowska & Cieszkowski 2014, Jankowski 2015, Hnylko & Hnylko 2016, 2019, Strzeboński et al. 2017, Heneralova & Khomyak 2018, Bubík et al. 2020, Siemińska et al. 2020, Hnylko et al. 2021, Golonka et al. 2022, Woyda et al. 2023).

The Popeli Beds represent the most extensive olistostromes known from the Silesian Domain of the Tethys and also from the entire Outer Carpathian basinal area. It is the largest coherent complex of deposits formed as a result of mass displacement, with an estimated lateral extent of at least 200 km. The outcrops extend from the area south of Przemyśl in Poland to the Prut Valley in the Eastern Carpathians in Ukraine (Dżułyński & Kotlarczyk 1965). It forms a significant part of the Eocene profile of the Skyba Nappe, and its deposits – distinct in character and cartographic appearance – were assigned the rank of a lithostratigraphic member within the Hieroglyphic Formation by Rajchel (1990). However, the corresponding unit in Ukraine remains to be clarified.

In the study area, the olistostrome sequence has a thickness of about 220 m and form only part of the Popeli Beds, which extend to the base of the Menilite Formation. The olistostrome is overlain by a complex of thin-bedded turbidites with a marl package containing bivalve shells, reaching a thickness of about 100 m. Rajchel (1990) points out dynamic changes in the lateral development of the thickness of the Popeli Beds. In eastern Poland, it has been estimated at 130–100 m (Watycha 1964, Dżułyński & Kotlarczyk 1965, Gucik et al. 1973, Morgiel & Szymakowska 1978). Dżułyński et al. (1979), Kotlarczyk (1979, 1988), and Rajchel (1990) report up to 200 m in the Popeli region (*locus typicus*), while Grzybowski (1919) reported only 120 m. In the Pokuttya Carpathians, Tołwiński (1950) observed 50–100 m of Popeli Beds with gradual lateral transitions to green shales.

Not only is the vertical extent of the Popeli Beds considerable, but its lateral extent is also quite significant. Usually, the Popeli Beds are one of the standard lithostratigraphic units in the outermost areas of the Skyba Nappe and in the Boryslav-Pokuttya Nappe (Vialov et al. 1981) near the European Platform. In the area of Skole, the Popeli olistostrome occurs in the Oriv Skyba, which represents marginal but more internal basinal environments. Similarly, the presence of the Popeli Beds in the inner "skybas," correlated with the position of the Oriv Skyba, were noted by Rajchel (1990). This position indicates the range of distal mass movements into the Skyba Basin. The olistostrome sequence has a stratified structure, consisting of complexes that formed in a cycle of at least several separate large-scale mass movements. Dżułyński and Kotlarczyk (1965), Kotlarczyk (1988), and Rajchel (1990) stated that the Popeli Beds represent "a huge submarine avalanche of the sediment-flow type, or by a number of successive avalanches." The individual displaced packages are separated by stratified mudstones or mudstones with rare sandstone layers, representing a type of sedimentation completely different from the mass transport. The occurrence of deposits separating olistostrome complexes has already been noted in the geological literature for their lithological characteristics. Apart from the matrix and exotic components, quartz sandstone interbeds are also mentioned as constituents of the Popeli Beds (Czaplicka et al. 1968). In future studies, the age should be discussed to see if it is consistent with other research, and the composition of the exotic components should be re-examined.

Previously, it was assumed that material was displaced from the slope zones and supplied from turbidite-sedimented areas, from the Hieroglyphic Beds and from the Menilite Beds (Dżułyński & Kotlarczyk 1965). However, the assumption that the Popeli Beds were supplied with material from the Oligocene Menilite Beds, including the Kliwa Sandstones, which occur in the overburden and are younger, contradicts stratigraphic principles. Kotlarczyk (1988) emphasises the occurrence of shallow-water material – marls, mudstones with neritic fauna – displaced from the neritic zone, from platform areas located to the north, in accordance with paleotransport directions from the north or north-west (Rajchel 1990). In the studied section, redeposited material of platform and slope origin were recorded. It was transported from the northern and northeastern side (Dżułyński & Kotlarczyk 1965, Kulchytskyi 1977, Kotlarczyk 1988, Heneralova & Khomyak 2018), from the shelf-slope zones in the outer part of the Skyba/Boryslav-Pokuttya Basin, which also represents the marginal zones of the Carpathian Tethys.

The deposition of the olistostrome sequence took place in the intra-basin zones covered by deep-water sedimentation. In the area of the Boryslav, the Popeli Beds lie on the shales and fine-grained turbidites of the lower – middle Eocene Vytvytsa Formation (Ponomareva 1988, Vialov 1988a). They occur on the Vyhoda Sandstones or Vyhoda-Pasieczniański-type deposits (Tołwiński 1950, Czaplicka et al. 1968). According to our own geological observation (see Figs. 1, 2), the Popeli olistostrome in the Pobuk Syncline occurs above the fine-grained turbidites overlying the Vyhoda Sandstone, whose structure and age correspond to the Bystrytsia Formation. This formation is widely developed in Ukraine on the territory of the Skyba and Boryslav-Pokuttya nappes and corresponds to the part of the Vytvytsa Formation. Its thickness reaches 100–350 m (Vialov et al. 1981, Vialov 1988b). The small degree of thickness of the Bystrytsia Formation in the study area is due to its erosion caused by the olistostrome flow. During the deposition of the Vyhoda Sandstones in this part of the Skyba Basin, large-scale gravitational movements of sediments began, characterising the environment in which the first mass-displacement deposits were formed.

The Popeli Beds are overlain by the Menilite Formation. In the upper part they border the Sub-Menilitic Cherts or on a complex of variegated shales, which occur below the Menilite Cherts (Tołwiński 1950, Dżułyński & Kotlarczyk 1965), or on the Borysław Sandstone, which lies between the Popeli and the Menilite Beds (Świdziński 1948). In areas where the Popeli Beds are reduced, they laterally change into green shales, overlain by variegated mudstones and Menilite Beds (Tołwiński 1950). They are covered by the Borysław Sandstone (Czaplicka et al. 1968) – probably part of the Menilite Beds to the east.

Biostratigraphic analyses indicate that the formation of the olistostrome exposed in the Pobuk area occurred during the Bartonian. In this area, the Popeli Beds were considered as the middle – upper Eocene in age (Mjatliuk 1950). Near the city of Boryslav, the Popeli Beds were assigned a late Eocene age (Ponomareva 1988, Andreeva-Grigorovich 1999).

In past, the Popeli Beds were generally considered to be of late Eocene age (Nowak 1927, Świdziński 1948, Tołwiński 1950, Dżułyński & Kotlarczyk 1965, Czaplicka et al. 1968, Morgiel & Szymakowska 1978, Kotlarczyk 1988, Rajchel 1990, Ryłko 2004, Gedl 2013), or upper and middle Eocene age (Krajewski & Urbaniak 1964, Watycha 1964). This age is mainly based on biostratigraphic studies of molluscs and organic-walled dinocysts from the Przemyśl region. The analysis of molluscs by Rogala (1925) suggested an "Upper Eocene (Priabonian or Bartonian) age." This age was confirmed by Krach (1985), who revisited the mollusc collection from the Popeli Beds, originally studied by Rogala (1925) and Wiśniowski (1908). The molluscs were collected from grey, marly or sandy clays and shales found under the Menilite Cherts. In the analysed section of the Pobuk Syncline, above the main packages of displaced rocks, a package of massive marls (Fig. 4F), containing small, fragmented molluscs shells, was documented. This package occurs within Priabonian turbiditic deposits. The mollusc-bearing mass flow deposits are similar in age and stratigraphic position to those described by Wiśniowski (1908) and Rogala (1925). However, the preservation of the fauna is different, as the fauna in the marls is mostly found in the form of thin, fragmented shells.

Further biostratigraphic data for the Popeli Beds are provided by Garecka et al. (2008), who suggests an Eocene/Oligocene age for the boundary region between Poland and Ukraine. Age discrepancies may result from the development of the Popeli Beds and their origin related to the mass gravitational movements of sediment, which was a cyclic and time-extended phenomenon. The Popeli Beds have a complex structure consisting of individual, overlapping olistostrome complexes. Their lateral development is also not uniform, and during the Eocene, olistostrome activity alternately diminished or was initiated. Grzybowski (1919), in his description of this formation, points to lateral lithological changes and transitions into Hieroglyphic-type rocks. According to the literature (Nowak 1927, Świdziński 1948, Tołwiński 1950, Dżułyński & Kotlarczyk 1965, Czaplicka et al. 1968, Morgiel & Szymakowska 1978, Kotlarczyk 1988, Rajchel 1990, Ryłko 2004, Gedl 2013), this activity in the Przemyśl region mainly occurred in the upper Eocene, while data from the Pobuk area suggest that it began in the Bartonian. Therefore, the base is diachronous, while the top of the Popeli olistostrome is more or less isochronous.

The Bartonian is a special interval in the context of sedimentary development in the Outer Carpathian basins. In both the Magura and Silesian domains, large olistostrome deposits have been recorded as a result of large-scale sediment displacements. In the Silesian domain, the olistostrome of Lake Rożnów is known, which descended during the deposition of the Hieroglyphic Beds along the northern slope of the Silesian Ridge (Cieszkowski 1992, Waśkowska & Cieszkowski 2014). Meanwhile, the Osielec olistostrome from the Magura domain, deposited in Beloveza Formation, moved down along the southern slope of the basin, and the Kamienica olistostrome, associated with the migration of the accretion prism from the south, formed on the southern slopes of the former Magura Basin (Cieszkowski 1992, Bromowicz 1998, Olszak 2006, Waśkowska & Cieszkowski 2014, Cieszkowski et al. 2017, Waśkowska et al. 2024a, 2025). The Bartonian is an interval in which olistostrome deposits are a significant component of the sedimentary sections of various structural units, indicating tectonic instability in the Carpathian basins that led to the regional disruption of slope stability (Waśkowska et al. 2024b).

## CONCLUSIONS

The Oriv Skyba sequence of the Skyba Nappe is a distinct olistostrome sequence with a total thickness of 220 m. It overlies the Bystrytsia Formation and passes upward into a complex of thin-bedded mudstones alternating with sandstones with a marl package containing bivalve shells. The olistostrome sequence consists of at least six individual olistostrome complexes, ranging in thicknesses from a few metres to over 50 m. The first five olistostrome complexes are composed of gravelly mudstones with very large gravelly clasts, including middle Eocene turbiditic mudstone-sandstone packages, turbiditic sandstones, mudstones, upper Ypresian – lower Lutetian marls, Paleogene foraminiferal-algal limestones, upper Lower and Upper Cretaceous limestones, silicified limestones, and siliceous rocks of the same age, Upper Cretaceous calcareous sandstones, and the uppermost Jurassic – lowermost Cretaceous Štramberk-type limestones. The last complex consists of deformed upper Paleocene thin-bedded turbidite deposits.

During the Bartonian, large-scale mass movements of material occurred, resulting in the formation of an olistostrome sequence. Displacements took place along the northern slope of the Skyba-Boryslav-Pokuttya Basin from platform and slope areas. The main landslide niche migrated down the slope as the olistostrome sedimentation developed and continued. Initially, the niche was located in the upper part of the Tethys shelf or on the highest part of the basin slope and supplied significant amounts of marly material mixed with older, rounded clasts of hard sedimentary rocks that were present in this environment. Subsequently, the niche migrated downslope and supplied mudstones and sandstones.

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***Supplement 1.*** *List of stratigraphically important taxa in foraminiferal samples*

***Supplement 2.*** *Description of the studied thin sections of clasts and matrix samples*