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Astronomical forcing of Carboniferous paralic sedimentary cycles in the Upper Silesian Basin, Czech Republic (Serpukhovian, latest Mississippian): New radiometric ages afford an astronomical age model for European biozonations and substages

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ABSTRACT

The ~3-km-thick Ostrava Formation in the foreland Upper Silesian Basin, which is located along the Czech-Polish border, provides the thickest and most complete record of mixed shallow-marine to continental sediments of Serpukhovian (Carboniferous) age in the former equatorial Pangea. The coal-bearing strata of the formation show a prominent rhythmic architecture with ~200 faunal bands, of which approximately 80 contain marine fauna. Three new and three existing U-Pb zircon ages of intercalated volcanic tuffs measured using chemical abrasion isotope dilution thermal ionization mass spectrometry (CA-IDTIMS) allow for the calibration of rhythmic patterns of several orders. The genetic sequences show an average duration of 100 kyr, in line with orbital forcing by short eccentricity. This allows the astronomical tuning of these cycles in the Ostrava Formation to a time axis with a resolution of ~100 kyr, providing temporal constraints on lithostratigraphic units, floral and ammonoid biozones, and West European substages. The Pendleian/Arnsbergian boundary defined at the base of the E2a ammonoid zone is now constrained to 325.9 Ma and the base of the *Lyginopteris stangeri* Zone at the base of the Ostrava Formation to ~329.2 Ma. The boundary of this zone with the subsequent *L. larischii* Zone corresponds to the base of the Jaklovec Member at ~325.8 Ma. The correlation of the major marine bands with equivalent bands in the Pennine Basin (England) and the Midland Valley of Scotland based on ammonoid faunas is suggested.

1. Introduction

The Upper Silesian Basin is currently the most important hard coalfield of Carboniferous age in Europe. Over three-fourths of its ~7400 km^2 area is located in Poland, and the remaining area is located in the Czech Republic (Zdanowski and Żakowa, 1995; Dopita et al., 1997). Its coal-bearing strata evolved from the marine deposition of siliciclastic flysch in the foreland of the Variscan orogenic belt following marine regression near the Viséan/Serpukhovian boundary (Kumpera and Martinec, 1995). The Serpukhovian coal-bearing deposition of the Ostrava Formation (or the Paralic Series in the Polish part of the basin) was interrupted by frequent marine transgressions (Havlena, 1964). Up to 3200 m of the sediments of the Ostrava Formation thus represent one of the thickest and most complete coal-bearing paralic sedimentary records of the Serpukhovian worldwide (Gastaldo et al., 2009b). It contains over 200 coal seams, of which ~90 have, at least locally, workable thickness and quality (Havlena, 1964; Žídková et al., 1997). Rich floral assemblages enable its subdivision into biozones (Purkyňová, 1970, 1977, 1997). Approximately 200 individual invertebrate macrofaunal horizons were recognised and merged into 27 groups for correlation (Řehoř, 1962; Řehoř and Řehořová, 1972). The marine faunas of the basin allow for high-resolution stratigraphy at the basin scale (e.g., Vašíček and Růžička, 1957; Řehoř and Řehořová, 1972; Vašíček, 1982, 1986) and their correlation with the major marine bands of Western Europe (Řehoř, 1970).

Although the first study addressing the stratigraphy of the Upper

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Silesian Basin appeared as early as the 18th century (Gerhard, 1781), the issue of its litho- and bio-stratigraphy was extensively addressed towards the end of the 19th century and during the 20th century. During this time, cycles of several orders of magnitude were distinguished, and approximately 45 stratigraphically important volcaniclastic beds, intercalated in coals (tonsteins) or in clastic sediments (whetstones), were detected in the formation (Dopita and Králík, 1977; Łapot, 1994; Martinec, 1997). Faunal horizons and embedded volcaniclastic beds served as tools for building the detailed internal stratigraphy of the formation, as required by the coal-mining industry.

Three high-precision CA-IDTIMS U-Pb ages of the volcaniclastic horizons in the lower part of the Ostrava Formation indicate that transgressive-regressive coal-bearing cyclothems (or genetic cycles) could be driven by 100-kyr eccentricity cycles (Gastaldo et al., 2009b; Jirásek et al., 2013a). New ages for three other newly dated tonsteins from the upper part of the formation provide a unique opportunity to verify the orbital-forcing hypothesis on the origin of the genetic cycles and major cyclothems and potentially tune these to a time domain at a ~100-kyr resolution. This would allow for the calibration of the Ostrava Formation, its stratigraphic subdivisions and early Namurian floral and faunal biozones to the International Chronostratigraphic Chart. Thus, the Ostrava Formation has the potential to serve as one of the most complete reference sections of paralic Serpukhovian strata for marine – non-marine biostratigraphic correlations in Europe. This study aims to contribute to this topic.

2. Geology and stratigraphy of the Upper Silesian Basin

2.1. Geological setting

The Upper Silesian Basin formed during the final stages of the evolution of the larger Moravo-Silesian Paleozoic Basin (defined by Unrug, 1966; see review by Kalvoda et al., 2008), in the eastern domain of the Central European Variscides, as a part of its outer zones, the Rhenohercynicum and Subvariscicum (Kalvoda et al., 2008; Kalvoda and Bábek, 2010; Buła et al., 2015 - Fig. 1B). The evolution of the Moravo-Silesian Basin, which is located on the Cambrian and Ordovician sedimentary cover of the Brunovistulicum basement in the foreland of the Variscan Orogen, began during the Late Devonian (Kumpera and Martinec, 1995; Narkiewicz, 2007). Carboniferous sedimentation began with preflysch carbonates and continued through the marine siliciclastic flysch (Culm) facies during the late Tournaisian and Viséan to the coal-bearing paralic and finally continental molasse facies (Kumpera, 1990). This Serpukhovian to Moscovian coal-bearing sequence of the larger synorogenic Moravo-Silesian Basin is referred to as the Upper Silesian Basin. The current post-erosional remnant of the basin has a roughly triangular shape that extends southward from Poland into the Czech Republic (Fig. 1C). The basin fill is overlain by Permian, Mesozoic, and Cenozoic strata in Poland, whereas Outer Carpathian nappes (Jurassic to Paleogene) and Neogene deposits of the Carpathian Foredeep cover the Carboniferous sediments in the southern part of the basin in the Czech Republic (Dopita et al., 1997).

Based on paleomagnetic data, the position of the Upper Silesian Basin was confined to close to the paleoequator. Krs et al. (1993) suggested the mean paleomagnetic directions and virtual paleomagnetic pole position to be $\Phi_p = 29.90°$ N, $\lambda_p = 167.74°$ E, dp = 5.52° and dm = 10.81° during the early stages of coal-bearing deposition.

2.2. Basin stratigraphy

The sedimentary fill of the Upper Silesian Basin consists of two major coal-bearing sedimentary units, i.e., the paralic and continental units. The stratigraphically earlier paralic deposition of the Ostrava Formation, or the Paralic Series in Poland, includes a wide spectrum of sedimentary environments (Kędzior et al., 2007). It is characterised by the seemingly cyclic alternation of marine and continental siliciclastics with coal seams, and frequent but thin intercalations of volcaniclastic material (Dopita and Králík, 1977). The Ostrava Formation is subdivided in stratigraphic order into the Petřkovice, Hrušov, Jaklovec and Poruba members (Fig. 2), based on the presence of spatially continuous megafaunal horizons (Kotas, 1995; Řehoř, 1997). Strata following the Ostrava Formation were deposited after a hiatus (Gothan, 1913b; Havlena, 1982; Purkyňová, 1977, 1997) and consist of continental sediments with freshwater faunal bands and coals, the latter being up to ~16 m thick (Dopita and Kumpera, 1993). These terrestrial strata are assigned to the Karviná Formation (Bashkirian) in the Czech part of the basin and are further subdivided into the Saddle, Suchá, and Doubrava members. This subdivision differs from that in the Polish part of the basin, where sedimentation continued up to the early Gzhelian (Fig. 2).

The most important marker horizons used for stratigraphic identification during the exploration stages and correlation within the basin are volcanoclastic rocks, marine horizons or their groups, and large lithosomes of coarse-grained sediments, as well as some prominent coal seams. The Ostrava Formation contains at least 17 tonsteins (argillitized tuffs in coal seams), ~32 tuffites and rocks with volcanic admixtures (so-called "whetstones") (Martinec, 1997; Horák et al., 1992). The thicknesses of the tonsteins range from millimetres up to 30 cm, while those of some whetstones can reach up to 15 m (Jirásek et al., 2013a). They serve as important isochrons for the basin stratigraphy, and some have great spatial extent. Also valuable for correlation are some thick and spatially widespread horizons of coarse clastics (coarse-grained sandstones to conglomerates); for example, the Castle Conglomerate Unit. Jirásek et al. (2013b) suggested that these coarse deposits are most likely a result of a significant event reflecting global climate change, namely, the onset of glacial interval C2 of Fielding et al. (2008a).

The paleogeographic configuration of the Ostrava Formation, as well as the distribution of its major sedimentary environments, were significantly affected by the geotectonic position of the Upper Silesian Basin. It was situated between the Variscan orogenic belt to the west and the low-relief epi-Cadomian Brunovistulian platform along its eastern margin (Dopita and Kumpera, 1993). The depocentre of the Ostrava Formation was a flat coastal area open to the north from which marine transgressions occurred (Havlena, 1964; Pešek et al., 1998). The synorogenic deposition of the Ostrava Formation is manifested by the gradual migration of the depocentre to the east and by the erosion of previously deposited sediments on the western flank of the basin during its simultaneous uplift and folding (e.g., Kandarachevová, 2011; Hýlová et al., 2013; Petrušková, 2013). Due to high sedimentary input into the basin, nearly all marine bands are siliciclastic in nature. The maximum subsidence and, in turn, greatest thickness of up to 3200 m is concentrated in a NNE-SSW striking zone along the western edge of the basin in the foreland of the prograding Moravian nappes (Žídková et al., 1997). Further east, on the stable Brunovistulicum basement, however, the thickness of the unit decreases to a tenth of that along the western margin. In the same direction, the number and thickness of coal seams and of invertebrate macrofaunal horizons decrease, while the sand content of the formation increases (e.g., Hýlová et al., 2013; Hýlová et al., 2016). Differences in the thickness of the basin fill and the absence of lithostratigraphic and/or biostratigraphic correlation markers lead to uncertainty in the exact correlation between the "classical" western part of the basin and the condensed "peripheral" development along its eastern margin. Therefore, in eastern Poland, Czarnocki (1907) and Doktorowicz-Hrebnicki (1935) divided the sedimentary sequence, which lacks most of the correlation markers, into the Sarnow, Flora, and Grodziec members (Fig. 2).

Although the subdivision of the Ostrava Formation and its equivalent strata (Paralic Series) in the Polish part of the basin is similar, there is disagreement in the definition of the boundaries of its members (e.g., the boundaries of the Petřkovice/Hrušov and Jaklovec/Poruba members - Fig. 2). In this study, we use the Czech definitions, as most data regarding the Ostrava Formation are from the Czech part of the basin

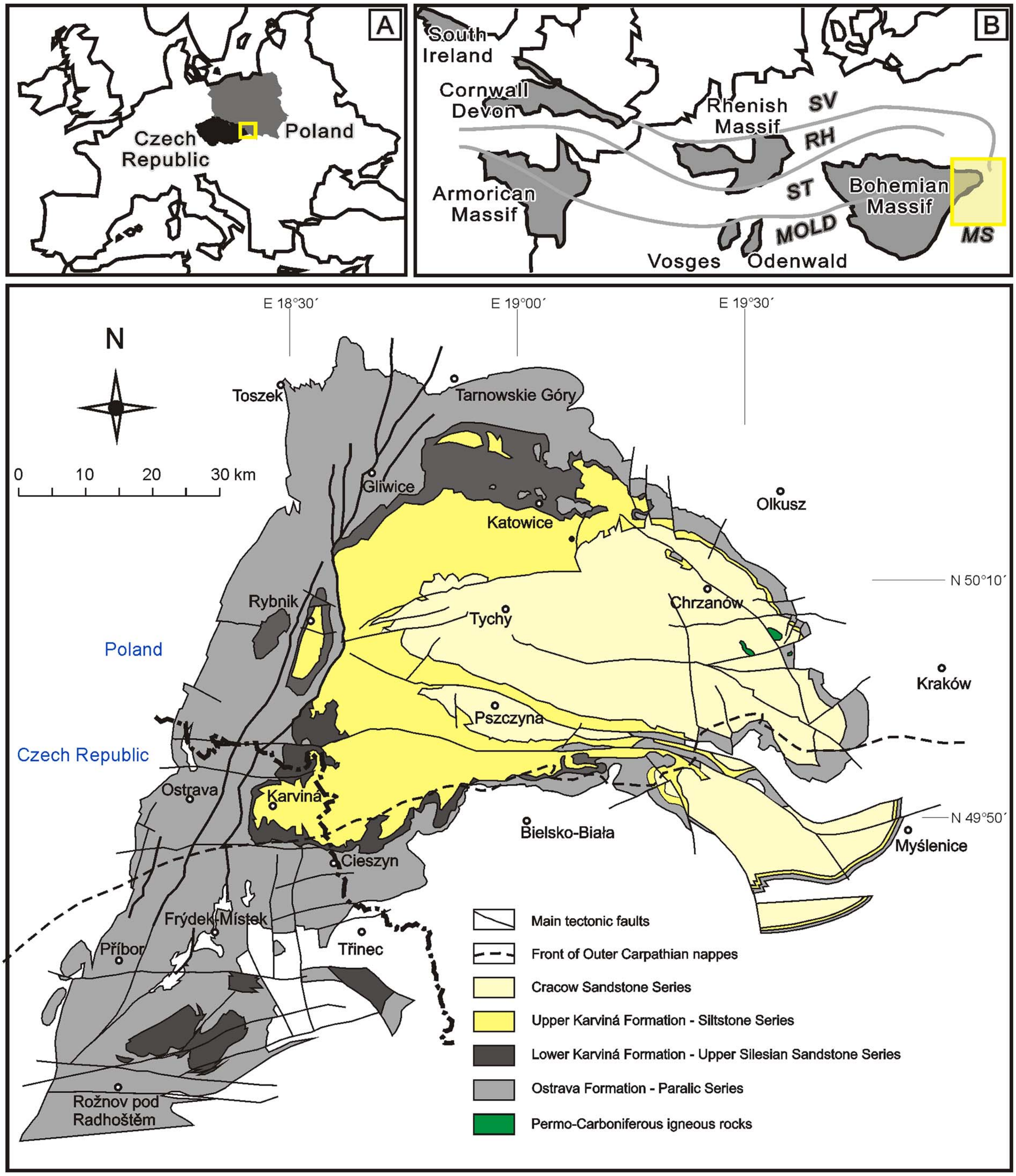

Fig. 1. A: Location of the Czech Republic and Poland. B: Location of the Bohemian Massif and the study area within the European Variscides. Major tectonic elements: SV – Subvariscan Zone, RH – Rheno-Hercynian Zone, ST – Saxo-Thuringian Zone, MOLD – Moldanubian Zone, MS – Moravo-Silesian Zone. C: Detailed map of the areal extent of the Ostrava Formation strata and location of the exploratory boreholes used in this study.
(Modified after Jirásek et al., 2013b, 2017).

(see Hýlová et al., 2013). For purposes of comparison, Fig. 2 provides an overview of the terminology used in the Czech and Polish parts of the basin.

2.3. Biostratigraphic zonation of the Ostrava Formation

The beginning of biostratigraphic research in the Upper Silesian Basin dates back to the second half of the 19th century, as coal-mining activity provided access to the fossil record. Both fauna and flora attracted the attention of geologists and became useful tools for basinal correlation purposes (Dopita et al., 1997).

The invertebrate fauna of the Ostrava Formation are concentrated into laterally widespread horizons of marine, brackish, or freshwater origin. Approximately 200 faunal horizons have been identified, of which less than half show marine or brackish faunas (Řehoř and Řehořová, 1972). Individual horizons are difficult to identify; therefore, Řehoř (1962) defined 27 groups of faunal bands that can be distinguished more easily. The most outstanding marine transgressions of basin-wide scale are concentrated into the Štúr, Františka, Enna, Barbora, and Gaebler groups and are used to divide the Ostrava Formation

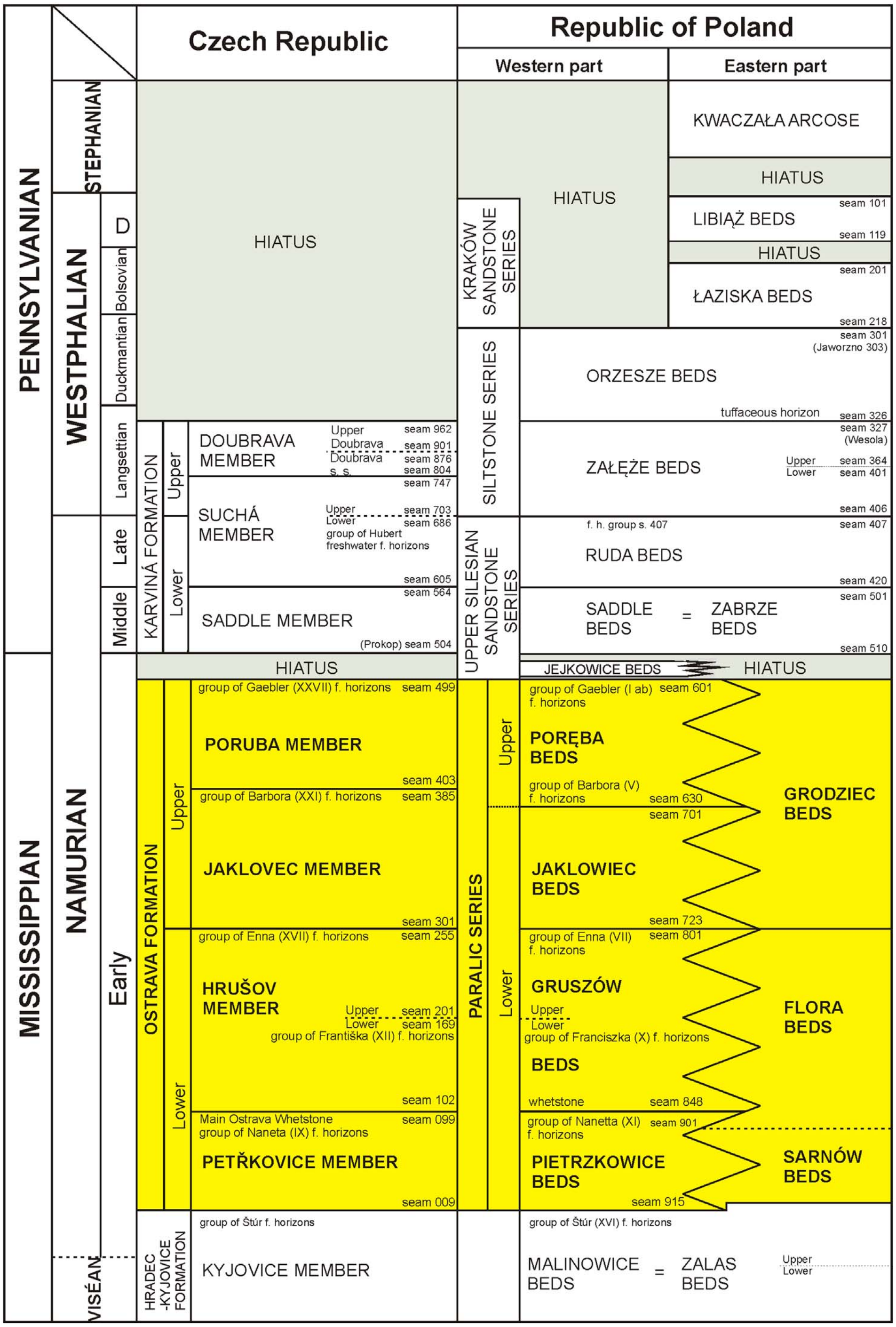

Fig. 2. Lithostratigraphic division of the Czech and Polish parts of the Upper Silesian Basin.
(According to Hýlová et al., 2013, modified).

into individual members (Fig. 2).

Using biological criteria to interpret paleosalinity, Řehoř (1966) distinguished assemblages of freshwater, brackish, euryhaline, and stenohaline faunas in the Ostrava Formation (Fig. 3). A certain degree of taxonomical overlap exists between adjacent zones. However, no mixing of freshwater and marine faunas occurs, and the mixing of brackish and marine faunas is very rare (Řehoř and Řehořová, 1972). The freshwater assemblages are characterised by the predominance of bivalves of the genera *Carbonicola*, *Naiadites*, *Curvirimula*, and *Porubites*. Brackish environments are represented by the substantially variable genus *Lingula*. Euryhaline faunas mostly comprise benthic bivalves of the families *Malletiidae*, *Nuculidae*, and *Nuculanidae*, which are representatives of the genus *Janeia*, and most members of the superfamily *Pholadomyaceia*. Euryhaline brachiopods are represented by the genera *Pleuropugnoides* and *Orbiculoidea*, whereas *Euphemites* is an important gastropod genus. Stenohaline faunas are typified by the presence of bivalve genera *Posidonia*, *Selenimyalina*, *Streblochondria*, and *Dunbarella* and gastropods of the genera *Bellerophon*, *Retispira*, *Cymatospira*, *Glabrocingulum*, and *Straparolus*. Brachiopods are represented by representatives of the order *Productida*. Less common, but indicative of fully marine conditions, are the remains of goniatites, nautiloids, trilobites, and corals (Fig. 3).

The use of invertebrate macrofauna, especially goniatites (Roemer, 1870; Patteisky, 1933; Wirth, 1933, 1935), initially appeared to be the most promising tool for the biostratigraphic subdivision of the formation. Unfortunately, most macrofauna are represented by

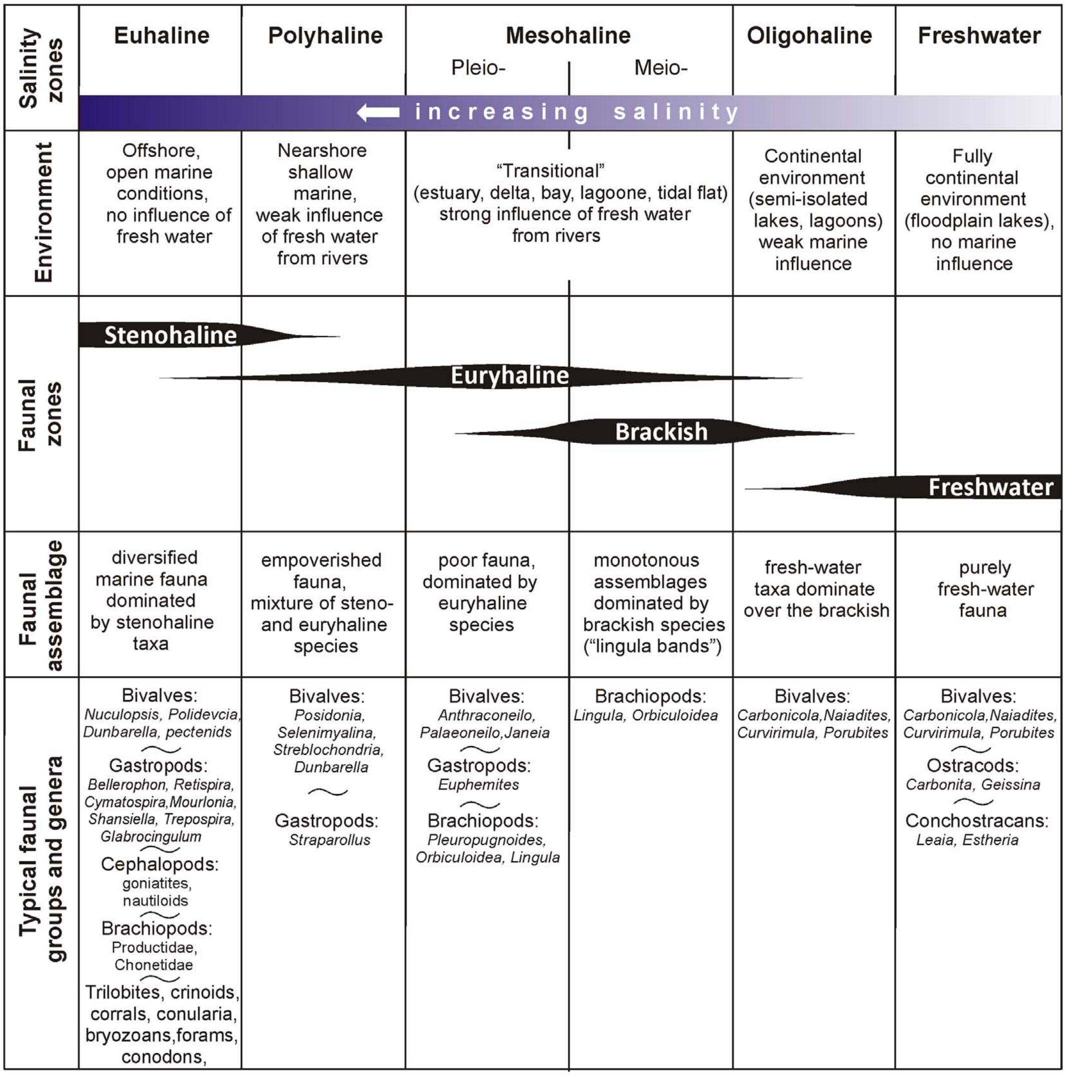

Fig. 3. Salinity-based distribution and composition of faunal assemblages in the Ostrava Formation.
Compiled from data in Řehoř and Řehořová (1972) and Řehoř (1997).

stratigraphically less important groups, such as pelecypods (41%), arthropods (16%), brachiopods (14%), and gastropods (12%), and only 11% of cephalopods, of which approximately half are goniatites (Řehoř and Řehořová, 1972; Bojkowski, 1967). Therefore, the correlation of marine horizons to existing faunal biozones remains problematic in some cases (Fig. 4). Nevertheless, most authors agree on assigning the goniatite Zone E (*Eumorphoceras*) to the entire Ostrava Formation. The definitions of the subzones E_1 (*Eumorphoceras pseudobilingue*) and E_2 (*Eumorphoceras bisulcatum*) are subject to different interpretations (Table 1). This is because there is an absence of some key taxa and, as such, other species typical of an association of a particular subzone are used for local biostratigraphic subdivision (Dopita et al., 1997). For example, only sporadic remains of the biozone taxon *Eumorphoceras bisulcatum leitrimense* were identified in the Gaebler group at the top of the Poruba Member (Řehoř and Řehořová, 1972). In addition, the index species *Cravenoceratoides nititoides*, which is typical of the goniatite subzone E_{2b}, was identified in the uppermost part of the same horizon (Vašíček, 1983).

The microfauna of the Ostrava Formation has not been systematically studied in detail. Conodonts of the Poruba Member were studied by Vašíček (1982, 1986, 1987), who found the biozone taxa *Gnathodus bilineatus bollandensis* and *Cavusgnathus naviculus*, thus confirming the affiliation of the Gaebler marine band below the top of the member to the goniatite subzone E_2. Vašíček and Růžička (1957) described foraminifera from the Ostrava Formation, but these were mostly new species that were unimportant to global stratigraphy. Two horizons were defined in the upper Ostrava Formation based on ostracods; one occurs in the Barbora group of faunal horizons, and the second occurs in the Gaebler group of faunal horizons (Přibyl, 1957). Similarly, newly identified species were found to be of little significance to global stratigraphy.

Another suitable and widely applied biostratigraphic approach is based on abundant macroflora, which together with Scottish and Belgium contemporaneous flora are among the richest and best-studied Serpukhovian floras of Europe and North America. The flora of the Ostrava Formation has been studied since the second half of the 19th century (e.g., Helmhacker, 1874; Stur, 1877). However, its first systematic use in biostratigraphy dates to the beginning of the 20th century (Gothan, 1913a, 1913b; Šusta, 1928; Patteisky, 1933). This flora is characterised by a high diversity of lyginopterids and medulosallean pteridosperms. Overall, approximately 235 whole plant species have been identified in the Ostrava Formation (Gastaldo et al., 2009b). Biostratigraphic investigations have progressed substantially since the 1950s, when extensive borehole exploration provided a unique opportunity for collecting stratigraphically located plant fossils. Several thousands of boreholes have been drilled in the Czech and Polish parts of the basin, and most of them have been investigated for fossils. Although paleobotanical research has been performed independently in

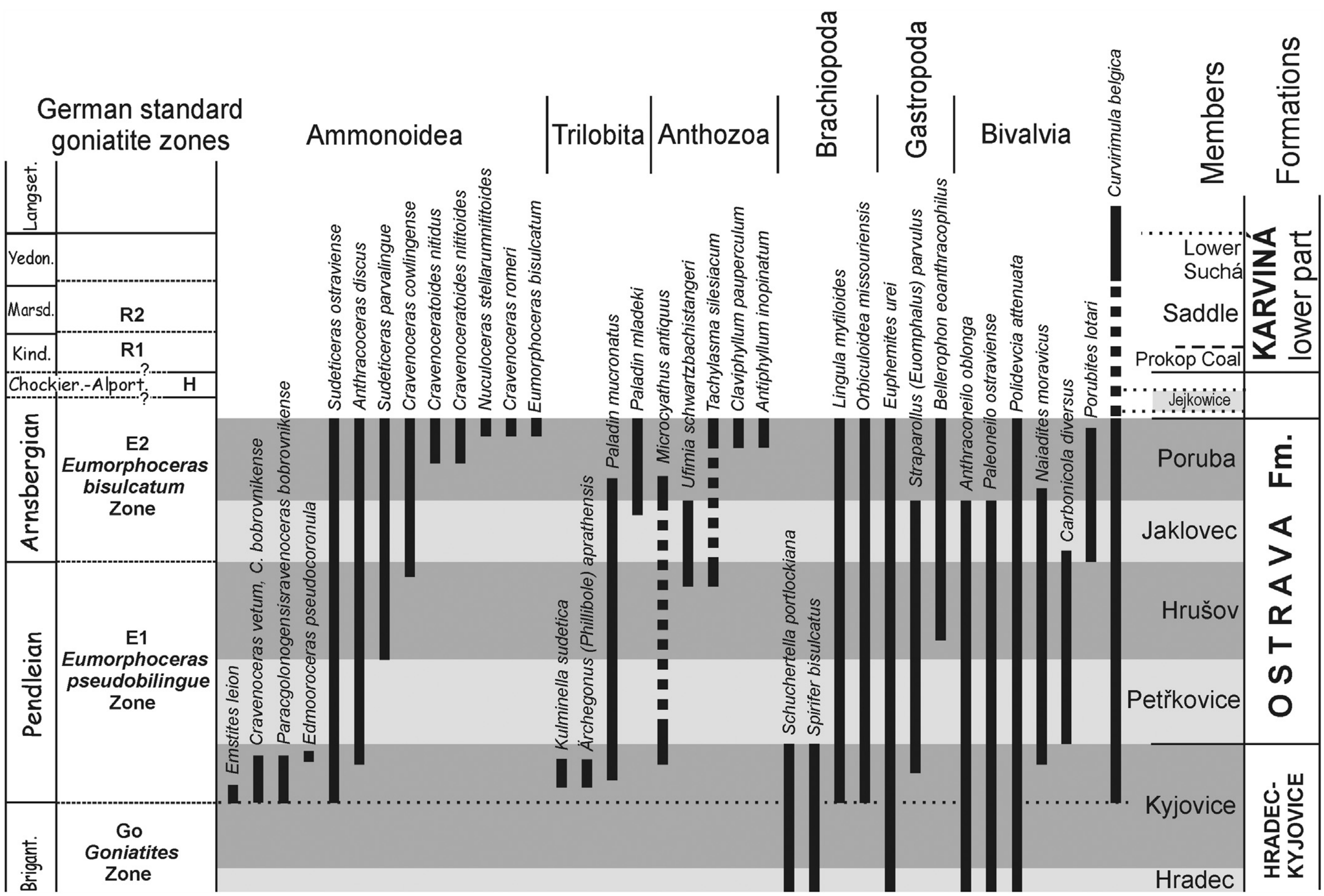

Fig. 4. Stratigraphic ranges of key Serpukhovian macrofaunal taxa in the Ostrava Formation.
Compiled after Řehoř and Řehořová (1972), Kumpera (1983), Musiał et al. (1995), and Řehoř (1997). Resources for German standard goniatite zones are summarised in Amler and Gereke (2002).

Table 1
Goniatite and floral zones and subzones of the Ostrava Formation defined by various authors. Notes: (1) In all cases, the goniatite zones shown in the table terminate at the roof of the Poruba Member, in which the last marine horizon of the Upper Silesian Basin occurs. (2) All floral zones also terminate at the roof of the Poruba Member, where a significant hiatus occurs. (3) The upper part of the goniatite zone Go is equivalent to the *Emstites novalis* Zone established by Korn (1994).

		GONIATITE ZONATIONS			FLORAL ZONATIONS			
		Řehoř 1970	Havlena 1982, Musiał et al. 1995	Dopita et al. 1997	Jachowicz 1972	Purkyňová 1970, 1996	Kotasowa, Migier 1995	Gastaldo et al. 2009b
OSTRAVA FORMATION	Puruba Member	E_2		E_2	N_{6-7}	NA_3	IV	4
	Jaklovec Member				N_{4-5}			
	Hrušov Member	E_1	E_2	E_1	N_3	NA_2	III	3
	Petřkovice Member				N_{1-2}			2 1
Kyjovice Member		Go	E_1	Go		NA_1	II	

both parts of the basin, it resulted in similar floral zonations (Table 1). In Poland, this research was conducted by Stopa (1957, 1967); however, Kotasowa and Migier (1995) established the currently used floral zonation based on extensive data from boreholes and coal mines (Table 1). In the Czech part, the flora was systematically studied by Havlena (1960, 1977) and Purkyňová (1962, 1963, 1970, 1977, 1997), and the zonation suggested by the latter author is now in use (Fig. 5). She defined two floral zones in the Ostrava Formation. The first occurrence of *Lyginopteris stangeri* (Stur) was recorded at the base of the first zone, while a more complex floral change marks its upper boundary at the base of the Enna group of faunal horizons at the boundary of the Hrušov and Jaklovec members (Fig. 5). This significant floral turnover was described by Gastaldo et al. (2009a). The same authors tried to create a more detailed stratigraphic division using

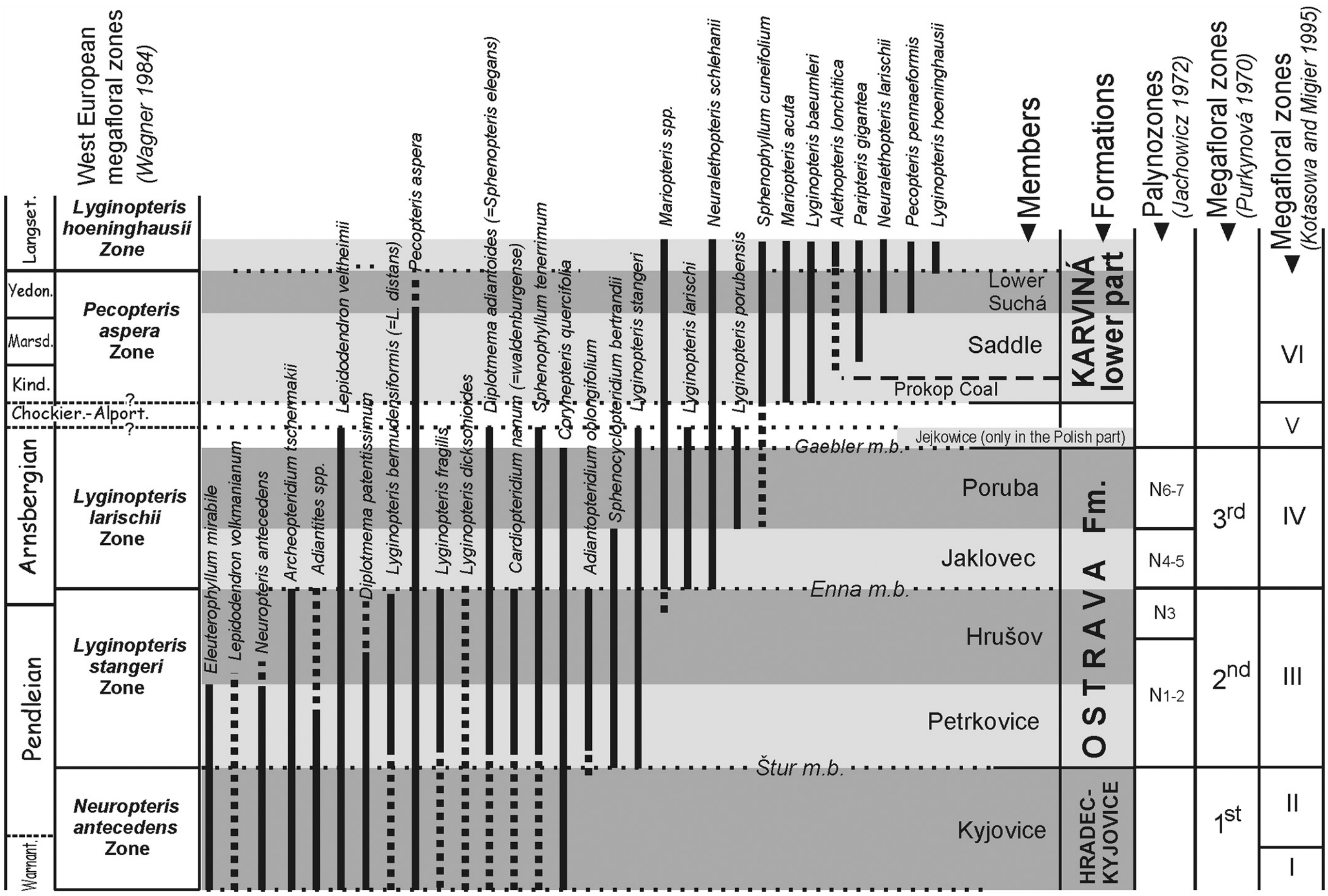

Fig. 5. Stratigraphic ranges of key Serpukhovian macrofloral taxa found in the Ostrava Formation and plotted against the Wagner (1984) biozonal scheme.

cluster analysis and Bayesian statistical methodology and subsequently defined 4 zones of floral associations in the Ostrava Formation, which were further subdivided into 7 subzones (Table 1).

Four palynological zones were distinguished in the Ostrava Formation in Poland by Jachowicz (1972). These include associations with a predominance of the genera *Punctatisporites* ($N_{1–2}$), *Lycospora* (N_3), *Densosporites a Lycospora* ($N_{4–5}$), and *Lycospora* ($N_{6–7}$). However, their ranges differ from those of the macrofloral zonations (Table 1). Indeed, a more recent study by Bek (2008) highlighted the problematic issues contained in the aforementioned publication. That author found it difficult to establish whether the differences between Czech and Polish studies (Table 1) are due to real differences between the two regions of the same basin or due to different approaches to nomenclature and taxonomy and differences in preservation.

2.4. Sequential architecture of the Ostrava Formation

The Ostrava Formation was deposited in a very flat coastal lowland characterised by a series of environments ranging from fluvial to offshore (Fig. 6). Their spatiotemporal distribution varied as the coastline transgressed and regressed landward and seaward due to relative sea-level changes (Jansa, 1967; Skoček, 1991; Doktor and Gradziński, 1999; Kędzior et al., 2007; Gastaldo et al., 2009a, 2009b). In addition, a prominent decrease in tectonic subsidence to the east together with autogenic processes resulted in a complex vertical repetition of lithologies showing seemingly cyclic patterns of strata at several time scales and magnitudes.

The first references to the cyclic pattern of sediments in the Upper Silesian Basin appeared in approximately the mid-20th century (e.g., Born, 1936; Přibyl, 1954; Zeman, 1960); however, this mostly comprised only a vague definition of individual sequences. In 1960, Jansa and Tomšík, based on detailed sedimentological studies of boreholes, distinguished six 350- to 850-m-thick megasequences (megacycles) of basin-wide extent separated by the thickest and most widespread marine bands (Fig. 7).

Individual megacycles consist of lowstand, continental, and highstand and inundated parts, in which the smaller-scale cyclothems (cycles of lower order) have different characteristics. The regressive interval of the megacycles corresponds to the onset of continental sedimentation with a higher proportion of coarse fluvial clastics and thick coal (Havlena, 1964). It comprises upward-fining cyclothems, which start with erosively based, thick, amalgamated sandstones followed by rooted soils with thick-banded, collovitrinite-rich coals. A fully marine fauna in mudstone is very rarely found above coal in this regressive part of the megacycles. During the subsequent transgressive part of the megacycles, the coal-forming peat swamps were flooded and changed into lakes and, subsequently, to lagoons with a brackish (linguloid) fauna. Siltstones prevail over fine-grained sandstones in this part of the megacycle. The “inundation” (~highstand) part of the megacycle represents maximum relative sea levels. This interval mostly consists of thick shallow-marine sandstones and thin, only locally mineable coals. The shales overlying the coals usually contain brackish and marine faunas.

Each megacycle consists of several mesocycles showing indices of transgressive and regressive development, although these are less pronounced compared to the megacycles (Jansa and Tomšík, 1960). The mesocycles are composed of several “basic” cycles. However, these have not been studied in detail over the whole Upper Silesian Basin.

Havlena (1964, 1970, 1982) defined ~10-m-thick, upward-fining cyclothems starting with an erosional surface followed by sandstones

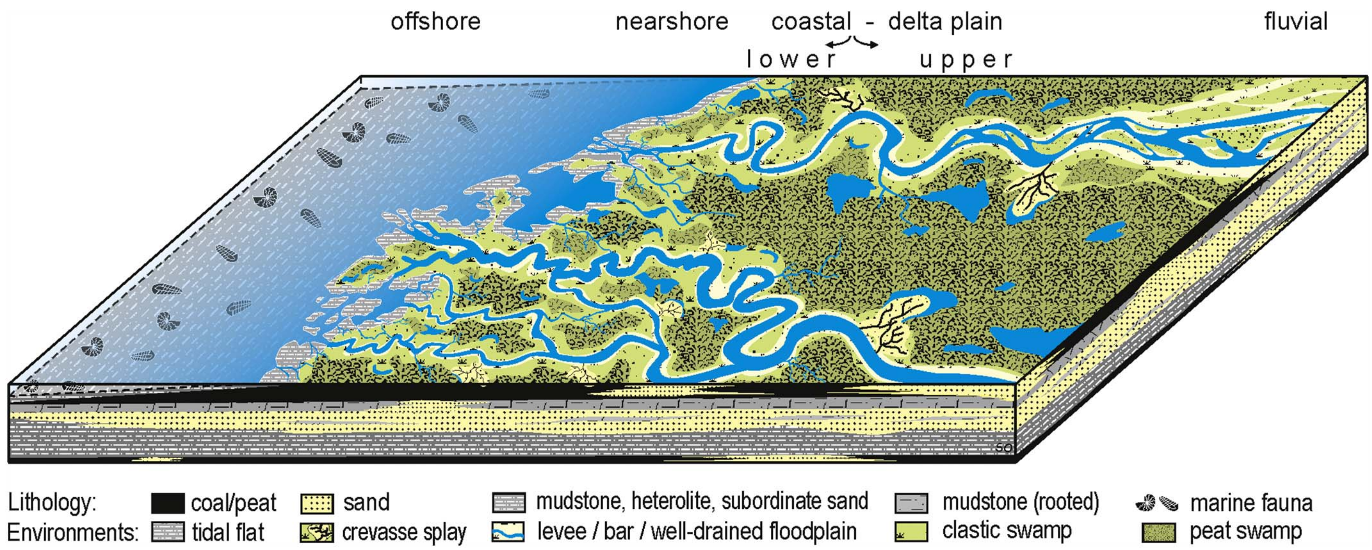

Fig. 6. Idealised model of the depositional system of the Ostrava Formation. The block diagram shows the early transgressive stage when delta plains become flooded and river mouths gradually transformed into estuaries.

that grade upwards to siltstones with abundant *Stigmaria* remains and coal seams followed by a thin mudstone layer containing lacustrine, brackish or marine fauna and drifted plant fragments.

Another cyclical pattern was suggested by Durčáková (1969), who divided the Ostrava Formation in the Czech part of the basin into 27 "partial lithostratigraphical units" defined as transgressive – regressive cycles, that mostly correspond to the groups of faunal bands defined by Řehoř (1962) (Fig. 7). This concept was later modified by Fialová et al. (1978), who reduced the number of "partial lithostratigraphical units" of the Poruba Member from 6 to 5. However, Žídková et al. (1997) questioned the role of using "partial lithostratigraphical units" for lithostratigraphy and renamed them as "partial lithological units" (Fig. 7). Lithological units further consist of basic cycles. Žídková et al. (1997) distinguished approximately 220 basic cycles throughout the Ostrava Formation that are mostly between 5 and 15 m in thickness. These basic cycles consist of coarse-grained sandstones deposited on an erosional surface, fining up into rooted siltstones and coal seams, and followed by siltstones or mudstones, often with fauna. However, Žídková et al. (1997) also mentioned "incomplete" cycles lacking coal seams and/or faunal bands. Some erosional surfaces at the bases of some cycles may reveal topographic differences of up to several metres. However, no significant depositional break disrupting floristic and faunistic biozones has been identified. The only biostratigraphically identified hiatus is between the Ostrava and Karviná formations (or Upper Silesian Sandstone Series in Poland) and is marked by an interruption of floral continuity (Gothan, 1913b) by ganister marking a prominent paleosol horizon at the top of the Ostrava Formation (Dopita and Králík, 1971).

Skoček (1991) was the first to suggest the effect of global glacio-eustatic sea-level change as a driving force controlling repetitive sedimentation patterns in the Upper Silesian Basin. Today, this concept is generally accepted and supported by more recent studies. Using this approach, Gastaldo et al. (2009b) subdivided the Ostrava Formation into 54 glacio-eustatically driven genetic cycles sensu Galloway (1989) (Fig. 7) based on a study of several hundred borehole successions. Individual genetic cycles comprise one or more discrete coal and marine bands, and their thicknesses vary across the coalfield. The base of each cycle is marked by a transgressive erosional (ravinement) surface above continental deposits (either coal or roof mudstone), as indicated by the appearance of estuarine or marine fauna and/or by high gamma ray values.

The average duration of 11 of these genetic cycles is 0.83 ± 0.24 Ma between the U-Pb dated Ludmila and Karel tonsteins in the lower part of the formation. This overlaps with the short-period (100 kyr) eccentricity cycle at the 95% confidence interval. Gastaldo et al. (2009b) also extrapolated downward sedimentation rates based on the average cycle duration and estimated the base of the Serpukhovian stage (~base of the formation) to have occurred at 329.7 Ma.

The brief overview of the previous studies of the cyclic architecture of the Ostrava Formation shows that various approaches have been used and that different results have been achieved. The largest differences exist on the finest scale of basic or genetic cycles, the thickness and number of which often vary, both vertically and laterally (Fig. 8). The thicknesses of the basic cycles in different studies vary between 2 and ~40 m, but the most common values range from 5 to 13 m (for an overview, see Žídková et al., 1997). Although sections of basic cycles usually include faunal (often marine) bands above the coal, the authors agree that many cycles are incomplete and lack a faunal band, coal or other parts of the cycle. None of these authors has attempted to use the basic cycles for correlation purposes. This is because the number of the basic cycles laterally varies mostly (but not always) in response to changes in the thickness of the formation controlled by tectonic subsidence, which decreased strongly to the east away from the orogenic foreland (Fig. 8). However, even within a small area where the thickness of the formation (members) and the number of cycles are similar, the correlation of the basic cycles is often difficult even between neighbouring boreholes at a scale of hundreds of metres. This could imply that the basic cycles are mostly of autogenic origin.

In contrast, the partial lithostratigraphic (Durčáková, 1969) and lithologic (Žídková et al., 1997) units generally represent much thicker internal units of the Ostrava Formation. The stratigraphic ranges of these mostly copy the Řehoř (1962) subdivision of the formation into 27 groups of faunal bands (Fig. 7). The average thickness of individual units varies between ~50 and 170 m, and is approximately 110 m on average. Their thickness also varies laterally within individual units, mostly reflecting changes in subsidence, and some neighbouring units may be difficult to separate (Žídková et al., 1997). The number of faunal bands varies both between individual lithostratigraphic/lithologic units (3–15) and within them. Despite local problems in the definition of these lithological units, it is assumed that they represent allocycles controlled by base-level changes driven climatically by orbital forcing. These units thus potentially may represent an object suitable for testing their consistency in the duration of their deposition, together with the genetic cycles of Gastaldo et al. (2009b) that are also defined as allocycles of glacioeustatic origin.

Megacycles are several hundred metres thick, and tend to become

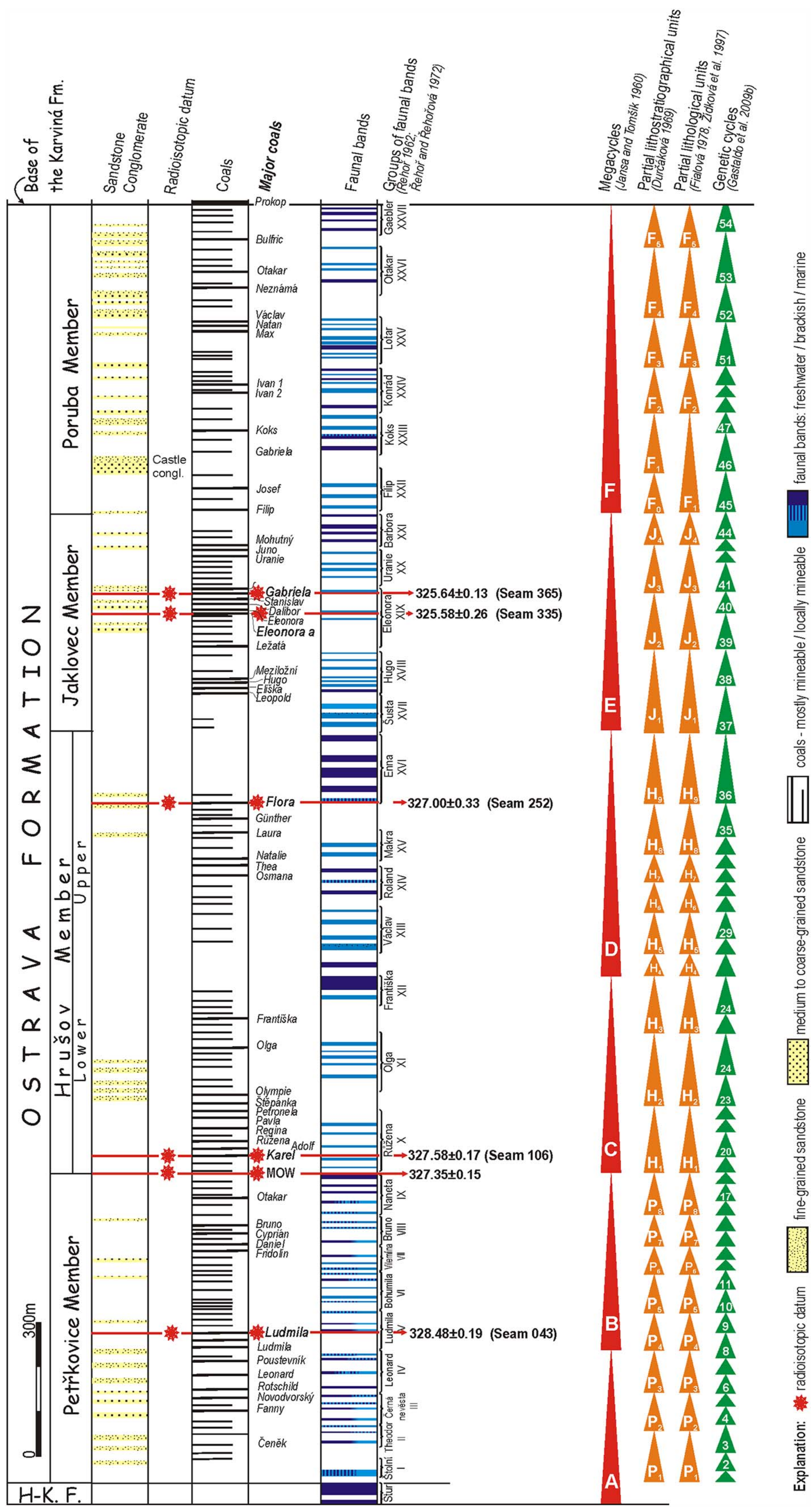

Fig. 7. Cyclic patterns of the Ostrava Formation as interpreted by different authors. Major sandstone bodies, coal seams and their names, and faunal bands are indicated in separate columns. Major sequences and different classifications of genetic or basic sequences are given to the right.

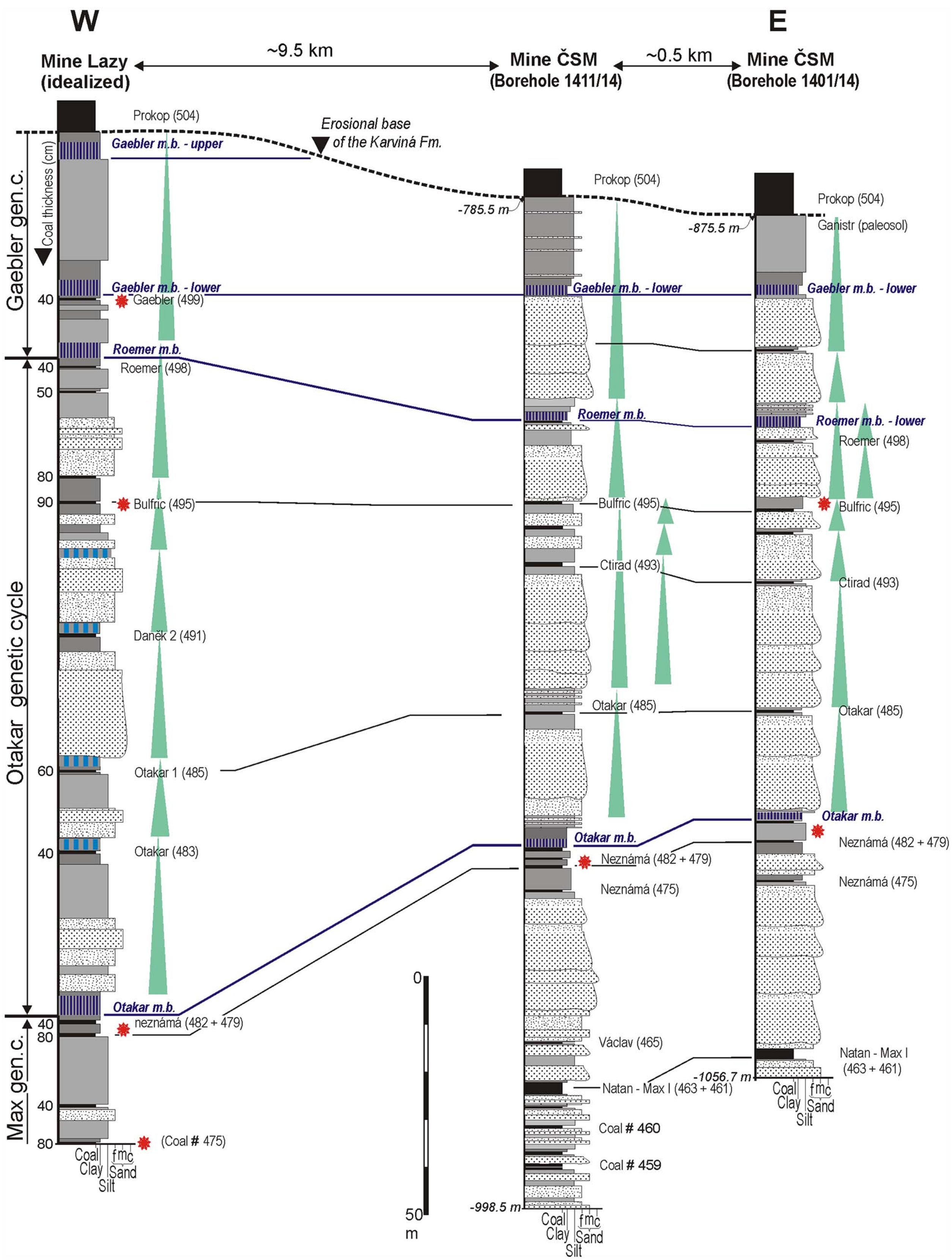

Fig. 8. Lateral variability in the architecture of the Otakar and Gaebler genetic cycles on top of the Ostrava Formation. Examples of basic cycles are indicated by triangles. Red stars indicate presence of tonsteins. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Table 2
An overview of published chronostratigraphic data from the Ostrava Formation of the Upper Silesian Basin. MOW = the Main Ostrava Whetstone horizon (tuffite).

	horizon	Hess and Lippolt 1986	Banaś 1999	Gastaldo et al. 2009b	Jirásek et al. 2013a
Poruba Member	479	319.0 ± 1.7 Ma 319.9 ± 1.6 Ma	312.2 ± 6.7 Ma 321.6 ± 6.6 Ma 323.9 ± 6.6 Ma	323.9 ± 4.0 Ma	
Jaklovec Member	365	324.8 ± 1.2 Ma		329.3 ± 4.1 Ma	
	335	319.0 ± 1.7 Ma 319.9 ± 1.6 Ma		328.8 ± 4.3 Ma	
Hrušov Member	106			328.01 ± 0.08 Ma	327.58 ± 0.17 Ma
	MOW				327.35 ± 0.15 Ma
Petřkovice Member	043			328.84 ± 0.16 Ma	328.48 ± 0.19 Ma

thicker upwards in the formation (Fig. 7). They are bounded by the most widespread and important marine bands and can be correlated not only across the Czech part of the basin but also across the entire Upper Silesian Basin (Havlena, 1982; Hýlová et al., 2013). They are considered here as informal lithostratigraphic units that could be of glacioeustatic origin.

2.5. Geochronology

The radioisotopic dating of volcaniclastic horizons embedded in the succession of the Ostrava Formation has been applied over the past three decades (Table 2). Lippolt et al. (1984) and Hess and Lippolt (1986), using $^{40}Ar/^{39}Ar$ total gas and plateau ages, dated sanidine from the three coal tonsteins of the Jaklovec and Poruba members. The tonstein of seam 335 (Eleonora) yielded an age of 326.5 ± 2.8 Ma as an average of two measured samples, the tonstein of seam 365 (Gabriela) yielded an age of 324.5 ± 1.2 Ma, and the tonstein of seam 479 (Max) yielded an age of 315.0 ± 1.9 Ma as the average of two measured samples.

Banaś (1999), using the $^{40}K/^{40}Ar$ method, dated sanidine, biotite, and illite/smectite in bentonite from seam 610 in the Poruba Member from the Szombierki and Grodziec mines in Poland, probably corresponding to the tonstein from seam 479 of the Poruba Member in the Czech part of the basin. However, usable dates were provided by sanidines (323.9 ± 6.6, 321.6 ± 6.6 Ma and 312.2 ± 6.7 Ma), whereas biotite and illite/smectite were subject to strong diagenetic changes. Their radioisotopic ages were considered by the author to be unreliable for estimating the timing of deposition.

High-resolution CA-IDTIMS U-Pb dating in the Upper Silesian Basin was first applied by Gastaldo et al. (2009b). They dated zircons from tonsteins of seam 043 (Ludmila) in the Petřkovice Member and seam 106 (Karel) in the lower Hrušov Member, which provided ages of 328.84 ± 0.16 Ma and 328.01 ± 0.08 Ma, respectively. These authors also recalculated earlier data published by Hess and Lippolt (1986) using new standards and a new ^{40}K decay constant and obtained ages of 323.9 ± 4.0 Ma for the tonstein of seam 479 (Max), 329.3 ± 4.1 Ma for the tonstein of seam 365 (Gabriela), and 328.8 ± 4.3 Ma for the tonstein of seam 335 (Eleonora). Based on their genetic cycles, Gastaldo et al. (2009b) also estimated the time of deposition of the Ostrava Formation to be 550 m/Ma and extrapolated the ages of the base and top of the formation to be 329.7 Ma and ca. 323.9 Ma, respectively. Jirásek et al. (2013a), using the CA-IDTIMS U-Pb method, dated zircons from the Main Ostrava Whetstone horizon on the boundary between the Petřkovice and Hrušov members and obtained an age of 327.35 ± 0.15 Ma. Using the recalibration of the isotope dilution tracer, these authors also recalculated the ages of the tonsteins of seams 043 (Ludmila) and 106 (Karel) previously published by Gastaldo et al. (2009b). These recalculated ages are given in Table 2.

3. Methods

Field data consists of the examination of sedimentary patterns in the cores of eight underground boreholes owned by the OKD company, available in the Green Gas DPB company. These boreholes were between 150 and 800 m long, and those examined for the purpose of this study stratigraphically covered the upper half of the Petřkovice and the base of the Hrušov members in the lower part and the Poruba Member and the top of the Jaklovec Member in the upper part of the Ostrava Formation. The sedimentological characterisation of these boreholes helped us to better understand the facies in the context of sea-level changes and to check for the delineation of genetic cycles in available stratigraphic intervals.

Because of the decline in coal mining activity in the Czech part of the basin over the last two decades, the larger part of the Ostrava Formation section is currently accessible neither in coal mines nor in boreholes. Instead, only written and graphic documentation of a few hundred boreholes was available to the authors. A larger set of boreholes and data from coal mines, now mostly abandoned, were used by Dopita et al. (1997) to compose a generalised section of the Ostrava Formation; we adopted these here as the most representative section of the Ostrava Formation to demonstrate its cyclic patterns recognised by various authors (Fig. 7).

To test the cyclic patterns of the Ostrava Formation and its calibration, existing radioisotopic dating was extended by three new samples of tuffs (tonsteins) from the Czech part of the Upper Silesian Basin. These are (in stratigraphic order) as follows: sample GP4 - tonstein of the Flora Coal (Seam No. 252, Upper Hrušov Member) from the Zárubek Mine; sample GP2 - tonstein of the Eleonora Coal (Seam No. 335, Jaklovec Member) from the Julius Fučík Mine; and sample GP1 – tonstein of the Gabriela Coal (Seam No. 365 = 15th Petřvald Coal, Jaklovec Member) from the Julius Fučík Mine. All samples are from the collection of the Geological Pavillion of Professor František Pošepný at VŠB – Technical University of Ostrava.

The U-Pb geochronology methods used for isotope dilution thermal

Table 3
New CA-IDTIMS U-Th-Pb isotopic data for zircons from the Ostrava Formation.

							Radiogenic Isotopic Ratios								Radiogenic Isotopic Dates					
Grain	$\frac{Th}{U}$	^{206}Pb* x10^{-13} mol	mol% ^{206}Pb*	$\frac{Pb^*}{Pbc}$	Pbc (pg)	$\frac{^{206}Pb}{^{204}Pb}$	$\frac{^{208}Pb}{^{206}Pb}$	$\frac{^{207}Pb}{^{206}Pb}$	% err	$\frac{^{207}Pb}{^{235}U}$	% err	$\frac{^{206}Pb}{^{238}U}$	% err	corr. coef.	$\frac{^{207}Pb}{^{206}Pb}$	±	$\frac{^{207}Pb}{^{235}U}$	±	$\frac{^{206}Pb}{^{238}U}$	±
(a)	(b)	(c)	(c)	(c)	(c)	(d)	(e)	(e)	(f)	(e)	(f)	(e)	(f)		(g)	(f)	(g)	(f)	(g)	(f)
sample GP-1 (Coal 365, Gabriela)																				
z6	0.691	0.3460	99.09%	34	0.27	1973	0.216	0.059257	0.255	0.734754	0.310	0.089929	0.103	0.661	576.6	5.5	559.34	1.34	555.118	0.55
z5	**0.075**	**0.1760**	**98.14%**	**14**	**0.28**	**972**	**0.024**	**0.053072**	**0.645**	**0.379506**	**0.730**	**0.051862**	**0.195**	**0.928**	**331.9**	**14.6**	**326.67**	**2.04**	**325.945**	**0.62**
z2	**0.131**	**0.1351**	**97.70%**	**12**	**0.26**	**784**	**0.041**	**0.053389**	**0.661**	**0.381644**	**0.736**	**0.051845**	**0.141**	**0.720**	**345.4**	**15.0**	**328.25**	**2.06**	**325.839**	**0.45**
z8	**0.033**	**0.5920**	**99.35%**	**40**	**0.32**	**2776**	**0.011**	**0.053153**	**0.262**	**0.379858**	**0.299**	**0.051831**	**0.083**	**0.747**	**335.4**	**5.9**	**326.93**	**0.84**	**325.752**	**0.26**
z1	**0.202**	**1.2968**	**99.50%**	**55**	**0.54**	**3601**	**0.064**	**0.053062**	**0.133**	**0.378983**	**0.182**	**0.051801**	**0.068**	**0.741**	**331.4**	**3.0**	**326.29**	**0.51**	**325.567**	**0.21**
z3	**0.075**	**0.8166**	**99.26%**	**36**	**0.51**	**2430**	**0.024**	**0.053067**	**0.203**	**0.378999**	**0.248**	**0.051798**	**0.072**	**0.794**	**331.7**	**4.6**	**326.30**	**0.69**	**325.548**	**0.23**
z7	0.372	1.3718	98.41%	18	1.84	1132	0.118	0.052984	0.220	0.377482	0.267	0.051672	0.071	0.713	328.1	5.0	325.18	0.74	324.777	0.24
z4	0.100	1.1366	94.91%	5	5.06	355	0.032	0.053319	0.437	0.377467	0.489	0.051345	0.092	0.713	342.2	9.9	325.17	1.36	322.772	0.29
sample GP-2 (Coal 335, Eleonora)																				
z6	**0.452**	**0.5817**	**99.30%**	**42**	**0.34**	**2590**	**0.143**	**0.053201**	**0.242**	**0.380353**	**0.285**	**0.051852**	**0.076**	**0.661**	**337.4**	**5.5**	**327.30**	**0.80**	**325.881**	**0.24**
z2	**0.286**	**2.6312**	**99.80%**	**141**	**0.44**	**8990**	**0.090**	**0.052983**	**0.084**	**0.378381**	**0.141**	**0.051796**	**0.066**	**0.928**	**328.0**	**1.9**	**325.85**	**0.39**	**325.538**	**0.21**
z3	**0.429**	**1.4142**	**99.45%**	**54**	**0.64**	**3302**	**0.135**	**0.053027**	**0.164**	**0.378661**	**0.207**	**0.051791**	**0.070**	**0.720**	**329.9**	**3.7**	**326.05**	**0.58**	**325.507**	**0.22**
z5	**0.102**	**0.9124**	**99.45%**	**49**	**0.42**	**3274**	**0.032**	**0.052907**	**0.169**	**0.377717**	**0.216**	**0.051778**	**0.071**	**0.747**	**324.8**	**3.8**	**325.36**	**0.60**	**325.431**	**0.23**
z4	0.342	1.5228	97.75%	13	2.91	802	0.108	0.053158	0.235	0.379037	0.283	0.051715	0.071	0.741	335.5	5.3	326.33	0.79	325.040	0.23
z1	0.307	1.8816	99.31%	41	1.08	2619	0.097	0.053057	0.151	0.378286	0.198	0.051710	0.067	0.794	331.3	3.4	325.78	0.55	325.010	0.21
z7	0.334	1.1426	99.10%	32	0.86	2002	0.106	0.053046	0.226	0.377112	0.269	0.051561	0.068	0.713	330.8	5.1	324.91	0.75	324.095	0.22
sample GP-4 (Coal 252, Flora)																				
z2	**0.286**	**0.2206**	**97.79%**	**13**	**0.41**	**818**	**0.090**	**0.053646**	**0.825**	**0.385316**	**0.895**	**0.052093**	**0.127**	**0.606**	**356.2**	**18.6**	**330.94**	**2.53**	**327.356**	**0.40**
z4	**0.499**	**0.1547**	**97.39%**	**11**	**0.34**	**691**	**0.158**	**0.054170**	**0.907**	**0.388998**	**0.982**	**0.052081**	**0.158**	**0.542**	**378.1**	**20.4**	**333.64**	**2.79**	**327.288**	**0.50**
z5	**0.264**	**0.6118**	**99.40%**	**46**	**0.31**	**2984**	**0.084**	**0.052997**	**0.154**	**0.380160**	**0.211**	**0.052025**	**0.078**	**0.816**	**328.7**	**3.5**	**327.16**	**0.59**	**326.942**	**0.25**
z1	**0.256**	**0.3866**	**99.06%**	**30**	**0.30**	**1920**	**0.081**	**0.052876**	**0.270**	**0.379129**	**0.318**	**0.052003**	**0.093**	**0.627**	**323.5**	**6.1**	**326.40**	**0.89**	**326.806**	**0.30**
z6	0.318	0.7791	99.40%	47	0.39	3002	0.100	0.053055	0.195	0.379935	0.239	0.051938	0.073	0.698	331.1	4.4	326.99	0.67	326.407	0.23
z8	0.408	1.1253	99.64%	82	0.34	5030	0.129	0.052956	0.153	0.378441	0.196	0.051830	0.070	0.722	326.9	3.5	325.89	0.55	325.747	0.22
z7	0.375	0.3560	97.51%	11	0.76	724	0.118	0.053084	0.210	0.39561	0.264	0.053767	0.104	0.664	344.3	4.7	338.46	0.76	337.61	0.34

Notes:
(a) z1, z2, etc. are single zircon grains or fragments annealed and chemically abraded after Mattinson (2005).
(b) Model Th/U ratio iteratively calculated from the radiogenic $^{208}Pb/^{206}Pb$ ratio and apparent $^{206}Pb/^{238}U$ date.
(c) Pb* and Pbc represent radiogenic and common Pb, respectively; mol % $^{206}Pb^*$ with respect to radiogenic, blank and initial common Pb.
(d) Measured ratio corrected for spike and fractionation only. Fractionation is estimated at 0.16 ± 0.03%/a.m.u. based on analysis of NBS-981 and NBS-982.
(e) Corrected for fractionation, spike, and common Pb; up to 1 pg of common Pb was assumed to be procedural blank: $^{206}Pb/^{204}Pb$ = 18.042 ± 0.61%; $^{207}Pb/^{204}Pb$ = 15.537 ± 0.52%; $^{208}Pb/^{204}Pb$ = 37.686 ± 0.63% (all uncertainties 1-sigma). Excess over blank was assigned to initial common Pb, using the two-stage Pb isotope evolution model of Stacey and Kramers (1975) at the nominal sample age.
(f) Errors are 2-sigma and propagated using the algorithms of Schmitz and Schoene (2007).
(g) Calculations based on the decay constants of Jaffey et al. (1971). $^{206}Pb/^{238}U$ and $^{207}Pb/^{206}Pb$ ratios and dates corrected for initial disequilibrium in $^{230}Th/^{238}U$ using Th/U [magma] = 3.
Bold is used for the zircon grains which were used for the age calculations and for Concordia diagrams (Fig. 9).

ionization mass spectrometry follow those previously published by Davydov et al. (2010). Zircon crystals were subjected to a modified version of the chemical abrasion method of Mattinson (2005), reflecting a preference to prepare and analyse carefully selected single-crystal fragments. Many analyses were undertaken on crystals previously mounted, polished, and imaged by cathodoluminescence (CL), and selected on the basis of zoning patterns. The U-Pb ages and uncertainties for each analysis were calculated using the algorithms of Schmitz and Schoene (2007) and the U decay constants of Jaffey et al. (1971). Other details of analytical parameters can be found in the notes to Table 3. Uncertainties are based upon non-systematic analytical errors, including counting statistics, instrumental fractionation, tracer subtraction, and blank subtraction. These error estimates should be considered when comparing our $^{206}Pb/^{238}U$ dates with those from other laboratories that used tracer solutions calibrated against the EARTHTIME gravimetric standards. When comparing our dates to those derived from other decay schemes (e.g., $^{40}Ar/^{39}Ar$, $^{187}Re/^{187}Os$), the uncertainties in the tracer calibration (0.05%; Condon et al., 2007) and U decay constants (0.108%; Jaffey et al., 1971) should be added to the internal error in quadrature, which increases the total error to ± 0.38 Ma for each of these samples.

The currently available radioisotopic ages of volcaniclastic rocks embedded in the Ostrava Formation (Gastaldo et al., 2009b and Jirásek et al., 2013a), together with new data from the tonsteins of the Flora (252), Eleonora (335) and Gabriela (365) coals, are used here to generate a calibrated age model of the formation following the approach used by Davydov et al. (2010). This model is based on the genetic cycles defined by Gastaldo et al. (2009b). In the first step, we tested the duration of the cycles between the dated intervals. Subsequently, we plotted the genetic cycles against time. The resulting time-calibrated model of the Ostrava Formation allows for the interpolation and extrapolation of the ages of individual members, as well as those of the base and top of the formation, with a resolution of approximately 100 kyr.

Regional chronostratigraphic units were used in the text, in addition to the international nomenclature of the Late Carboniferous subdivision of western and central Europe (cf. Menning et al., 2006; Davydov et al., 2012).

4. Results

4.1. Radioisotopic ages of dated tonsteins

The heavy minerals separated from sample GP1 (Jaklovec Member, Seam 365 Gabriela) yielded an abundant, homogeneous population of elongate prismatic zircons that are readily identified as volcanic in origin. The cathodoluminescence imaging of zircon crystals from the GP1 tuff reveals predominantly non-luminescent crystals comprising slightly brighter homogeneous cores mantled by darker oscillatory zoned overgrowths. A lesser number of crystals are either uniformly luminescent or have irregular luminescent cores. The latter minority of grains are interpreted as xenocrystic, while the dominant non-luminescent crystals were confirmed as juvenile and U-rich during screening via LA-ICPMS, yielding ages of ca. 310 to 340 Ma and uranium concentrations of 500 to 600 ppm. Eight grains were selected for further CA-IDTIMS analysis on the basis of their CL patterns. Chemical abrasion in concentrated HF at 180 °C for 12 h resulted in the significant dissolution of the zircon crystals, as anticipated from their high U concentrations and somewhat translucent characteristics. One grain (z6) yielded discordant isotopic ratios and a significantly older $^{206}Pb/^{238}U$ date of 555 Ma, which is interpreted as inherited or containing inherited components. The majority of the cluster analyses (n = 5) produced a weighted mean $^{206}Pb/^{238}U$ date of 325.64 ± 0.13(0.20)[0.40] Ma (MSWD = 0.90). *Quoted errors for individual analyses are thus of the form* ± $X(Y)[Z]$, *where* X *is solely the analytical uncertainty,* Y *is the combined analytical and tracer uncertainties, and* Z *is the combined*

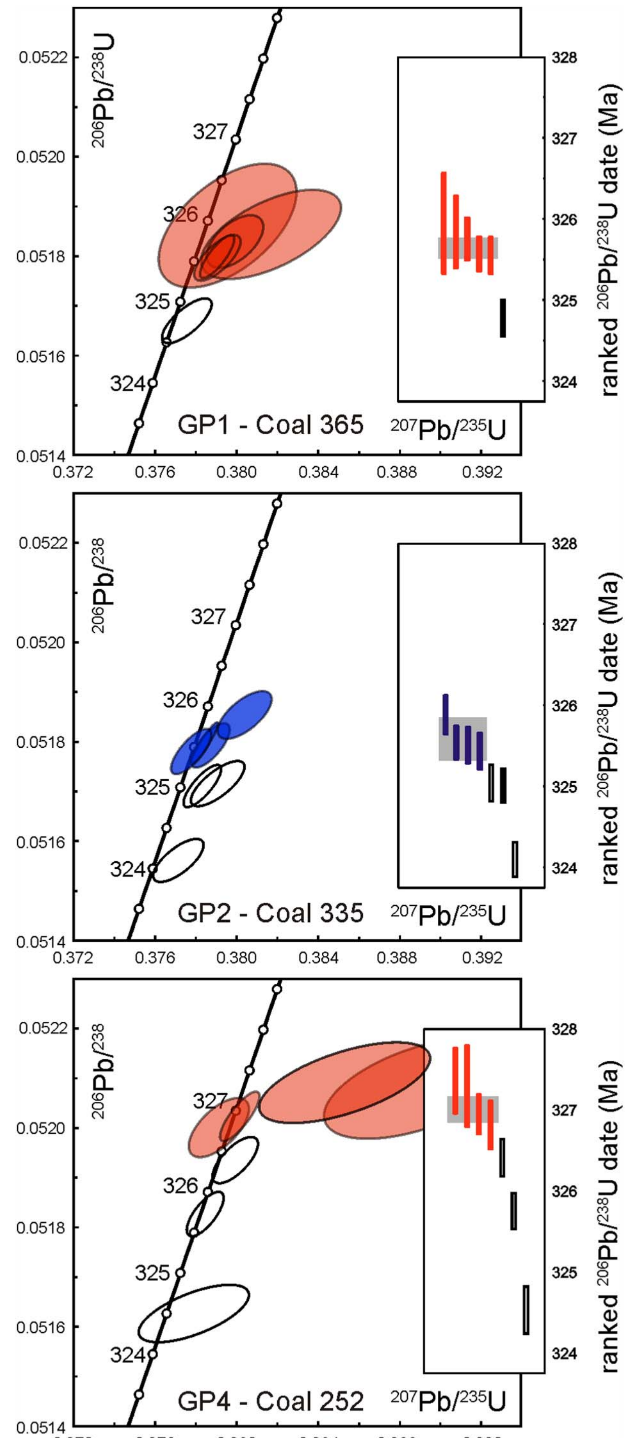

Fig. 9. U-Pb concordia diagrams and age plots for all new single zircon analyses from the Ostrava Formation tonsteins (tuffs).

analytical, tracer and ^{238}U *decay constant uncertainties*. It is interpreted as dating the eruption and deposition of this tuff. Two other analyses yielded anomalous dates approximately 1 to 3 Ma younger than the majority cluster and are interpreted to have experienced residual Pb-loss unmitigated by chemical abrasion (Fig. 9).

Zircons were less abundant in sample GP2 (Jaklovec Member, Seam 335 Eleonora), and comprised a bimodal population of both elongate and equant, prismatic, fine- to medium-sand-sized crystals, with indications of rounding on some grain edges. In CL images, the more equant grains were more likely to be highly luminescent and to contain irregularly resorbed cores. The more elongate grains ranged from more luminescent cores to darker oscillatory-zoned mantles; the relative proportions of bright and dark domains varied from grain to grain. LA-ICPMS confirmed relatively low U concentrations and anomalously old ages (ca. 470 to 650 Ma) for the luminescent cores of the equant population of zircon grains. Conversely, the elongate zircon crystals yielded high U concentrations ranging from 200 to 700 ppm, with ages of ca. 330 to 300 Ma; these grains were targeted for CA-IDTIMS analysis. Eight single zircon grains were selected on the basis of CL images and LA-ICPMS spot ages; of the eight, one crystal completely dissolved under chemical abrasion, while the remaining were significantly corroded. The four oldest grains yielded statistically similar isotopic ratios, with a weighted mean $^{206}Pb/^{238}U$ date of 325.58 ± 0.26(0.31)[0.46] Ma (MSWD = 2.87), which is conservatively interpreted as the eruption and depositional age of the tuff (Fig. 9). As with sample GP1, several anomalously younger analyses are interpreted as having experienced residual Pb-loss unmitigated by the chemical abrasion of these high-U crystals.

Similar to sample GP1, the heavy minerals separated from sample GP4 (Upper Hrušov Member, Seam 252 Flora) yielded an abundant, homogeneous population of elongate prismatic zircons that are readily identified as volcanic in origin. The CL imaging of zircon crystals from the GP1 tuff reveals predominantly non-luminescent crystals comprising slightly brighter homogeneous cores mantled by darker oscillatory zoned overgrowths. Truncated, luminescent, and likely inherited cores are observed in only a few grains. LA-ICPMS spot analyses confirmed the very high U contents of these crystals (> 1000 ppm) and juvenile ages of ca. 310 to 335 Ma. Unfortunately, eight single zircon grains selected on the basis of CL images were significantly dissolved during chemical abrasion, visually estimated as from 80 to 100% by volume. Four crystals yielded clustered U-Pb ages with a weighted mean $^{206}Pb/^{238}U$ date of 327.00 ± 0.33(0.36)[0.50] Ma (MSWD = 2.10), while three analyses scattered to younger ages (Fig. 9).

We obtained four dated stratigraphic levels covering the majority of the Ostrava Formation (Fig. 7) to calculate the duration of the genetic cyclicity in three discrete intervals: (1) Ludmila – MOW/Karel, (2) MOW/Karel – Flora, and (3) Flora – Eleonora/Gabriela. The results summarised in Fig. 10 show that the duration of the genetic cycles in these defined discrete intervals varies significantly. The genetic cycles approach 100 kyr between the Ludmila and MOW/Karel, as was found by Gastaldo et al. (2009b). The duration of the genetic cycles, where the Flora tonstein is involved, exhibits an extremely short (29 kyr) and extremely long (347 kyr) periodicity for intervals 2 and 3, respectively (Fig. 10).

This suggests that the age of the Flora Coal tonstein may have been affected by inheritance, as indicated by its age, which is very close to that of the Karel tonstein in the lower part of the Hrušov Member, although the stratigraphic position of the Flora Coal is approximately 800 m higher near the top of the Hrušov Member. The challenges of extracting a robust depositional age from the high-uranium, radiation-damaged zircon crystals of the Flora Coal tonstein have been noted before. It is certainly possible that the oldest cluster of zircons comprised low-U inherited cores that remained after chemical abrasion. For this reason, we have excluded the age of the Flora tonstein from our calculations.

4.2. Genetic cycles of the Ostrava Formation, their sedimentary environment, and palaeogeographical setting

As was previously suggested, among the existing cyclic patterns defined for the Ostrava Formation, only the partial lithological units of Žídková et al. (1997) and the genetic cycles of Gastaldo et al. (2009b) are defined throughout the entire formation and are related to basin-scale, base-level changes and thus have the potential to be glacioeustatically (i.e., climatically) driven. The 26 partial lithological units

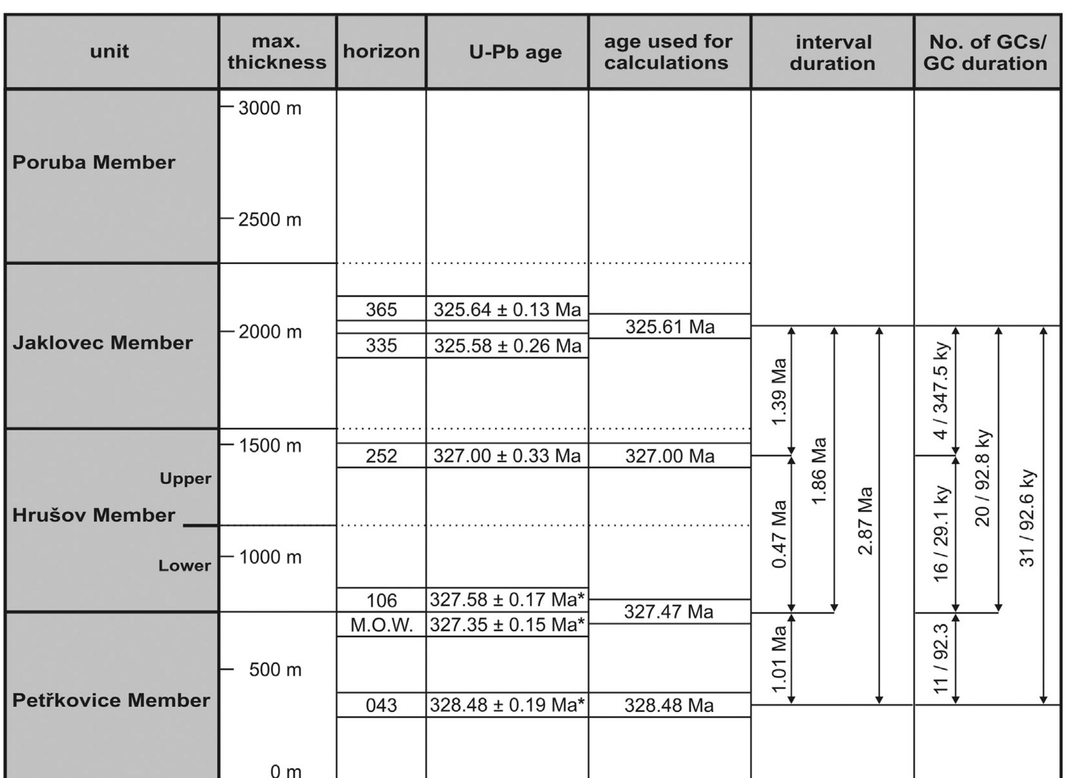

Fig. 10. New chronostratigraphic data from the Ostrava Formation and the calculated duration of the genetic cycles of Gastaldo et al. (2009b) as the arithmetic mean of genetic cycle length within radioisotopically constrained intervals. Notes: U-Pb data marked with asterisks were published or recalibrated by Jirásek et al. (2013a; see 2.6). M.O.W. = the Main Ostrava Whetstone horizon (tuffite).

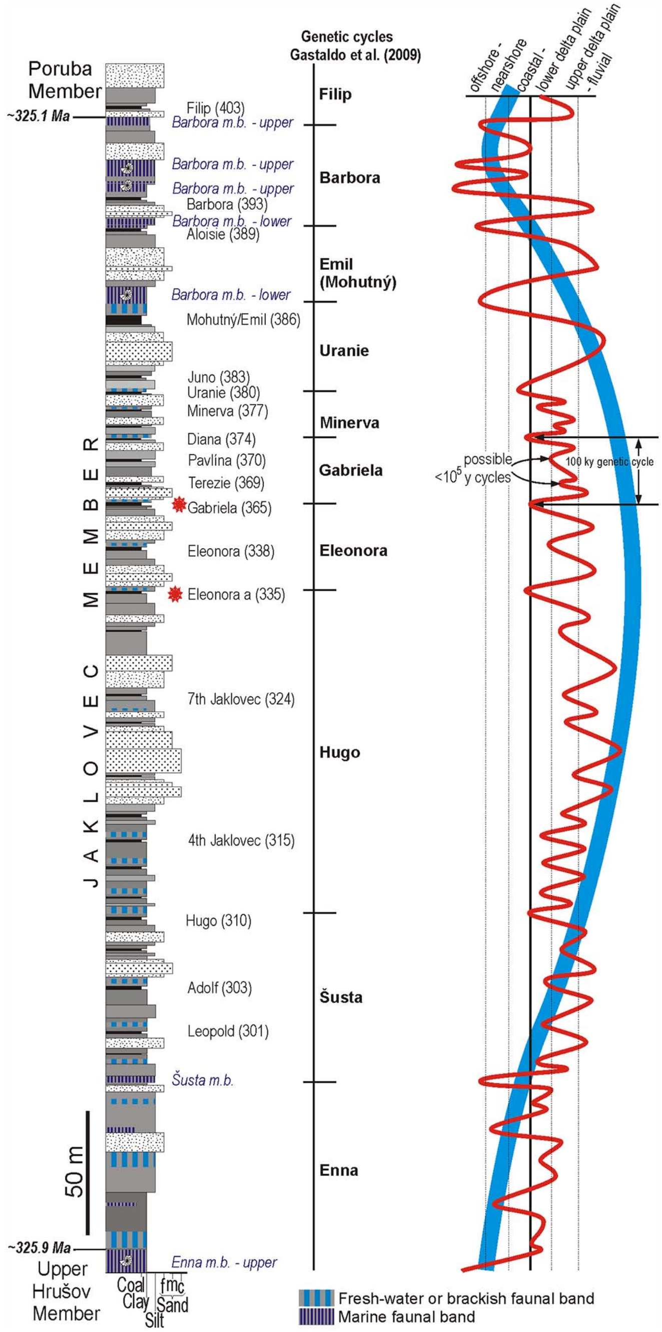

Fig. 11. Genetic cycles of the Jaklovec Member and suggested onlap-offlap curve indicating the cyclicity of the three superimposed orders. The onlap-offlap curve is based on the characteristics of the facies and the macrofauna of the invertebrate faunal bands.

basically correspond to the group of marine bands defined by Řehoř (1962), who grouped marine horizons bearing stratigraphically similar or identical faunas. Using this biostratigraphic approach yielded units that were determinable across most of the basin; however, this approach did not consider sedimentological criteria. The presence of several faunal bands up to several tens of metres apart suggests that more than one transgressive-regressive cycle is included within any partial lithostratigraphic unit. This is indicated by the arithmetic mean of their duration, which, for dated intervals, varies between 185 and 200 kyr, which are values that fall between short and long eccentricity cycles. In contrast, the 54 genetic cycles of Gastaldo et al. (2009b) are defined based on the sequence stratigraphic principles established earlier in the coeval Black Warrior Basin (Alabama), which has a similar geotectonic position (Gastaldo et al., 1993). The identification of the

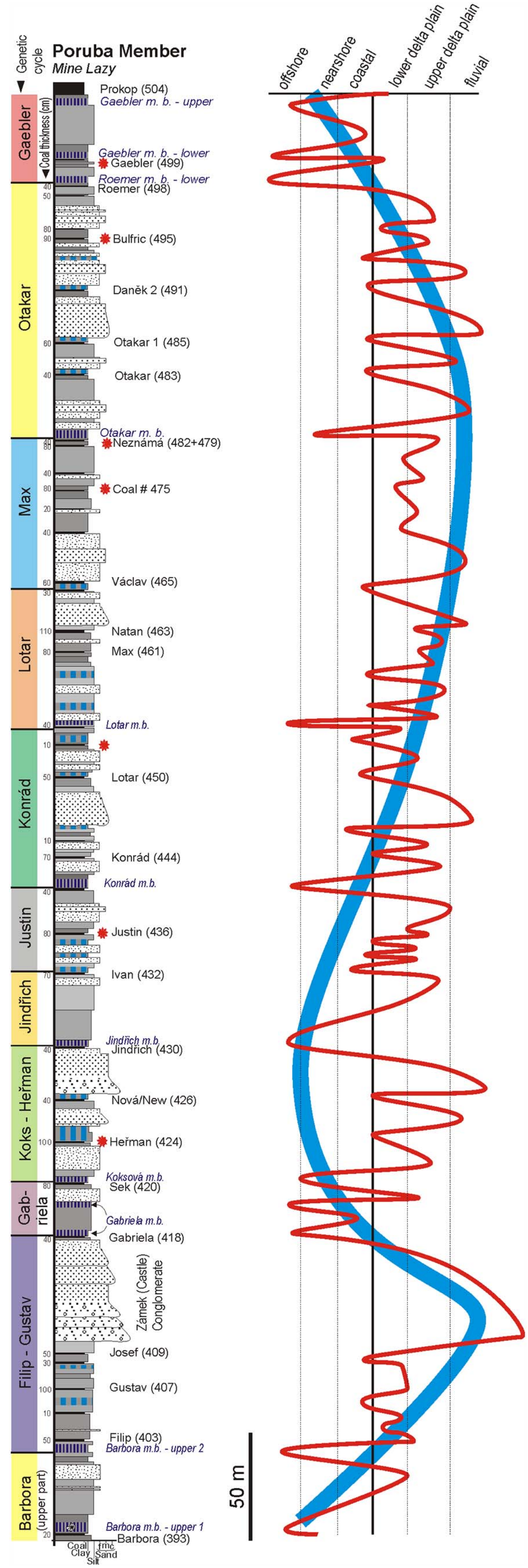

Fig. 12. Poruba Member and its genetic cycles in a mining field of the Lazy Mine in the central part of the Ostrava – Karviná coalfield. The onlap-offlap curve is based on the characteristics of facies in cored material and those of the macrofauna of the invertebrate faunal bands. The generalised section for the mining field of the Lazy Mine in the central part of the Ostrava-Karviná coalfield is provided by the DPB Paskov company.

genetic cycles in the Ostrava Formation benefited mainly from hundreds of graphic logs of boreholes and electric log curves of these boreholes where faunal bands could be stratigraphically identified by spikes in gamma curves even in the absence of preserved invertebrate faunal elements.

The examination of available boreholes in combination with published data allowed us to further investigate the original description and definition of the genetic cycles of Gastaldo et al. (2009b). Here, the adopted genetic cycles of Gastaldo et al. (2009a, 2009b) are bounded by regional transgressive erosional surfaces that overlie continental deposits (Figs. 11, 12). These transgressive surfaces are usually the contact between coal or roof shale and overlying estuarine to marine deposits with megafaunal remains (Gastaldo et al., 2009b).

The faunal bands are typically dark grey claystone to mudstone with siderite nodules (Havlena, 1964). They represent the zone of maximum flooding usually achieved in nearshore to offshore settings depending on the magnitude of relative sea-level rise or sediment input indicated by the characteristics and diversity of macrofauna (Řehoř and Řehořová, 1972). Fully marine conditions with diversified marine fauna were, however, reached only during major marine transgressions, mostly those that separated individual megacycles (Dopita et al., 1997). There is no overlap between freshwater and marine faunas, and only in rare instances (0.2% of approximately 30,000 tested associations) did Řehoř and Řehořová (1972) observe a mix of linguloid and marine faunas. Individual taxa occur in characteristic and regionally stable associations that, according to Řehoř (1966, 1977, 1997) and Řehoř and Řehořová (1972), represent original autochthonous faunal assemblages occupying ecologically different habitats. Their alternation throughout the band succession reflects environmental changes related to salinity and possibly bathymetry (Řehoř and Řehořová, 1972; Paproth et al., 1994; Gastaldo et al., 1993; Heckel, 1995, 2002; Dusar et al., 2000; Krzeszowska, 2015).

The interval above the faunal bands consists of a coarsening-upward sequence of mudstone grading to interlaminated mudstone and fine-grained sandstone, the latter increasing in proportion upwards in the Ostrava Formation. Sand laminae in some genetic cycles are regular or of rhythmically variable thickness on millimetre to centimetre or, less commonly, decimetre scales (Figs. 13, 14). Shallow, decimetre- to ~2-metre-thick channel-form sandstone bodies with intercalated mud laminae locally appear in the upper part of the coarsening-upward sequence. The described lithologies and their geometries and primary textures indicate deposition in a tidal regime (Gastaldo et al., 1993; Demko and Gastaldo, 1996) on a tidal flat with associated tidal creeks and channels that were deposited around the high and initiation of lowering relative sea level.

The characteristics of the following part of the succession depend on the magnitude of relative sea-level change and strongly vary among individual genetic cycles (Figs. 11, 12). Tabular sandstone bodies, of which a few are up to 10 m thick, are typical and interpreted as deltaic mouth bars (Fig. 13) or sand barriers (Havlena, 1964; Jansa, 1967; Doktor and Gradziński, 1999; Kędzior et al., 2007). Purely fluvial sandstones are rare (Havlena, 1964). They form linear bodies that are a few to > 20-m-thick, with an erosional base and internal erosional surfaces, coalified tree trunks; they lack a tidal bundle signature (Havlena, 1964; Jansa, 1967). They were deposited when sea level was approximately at its lowest point and the area was drained by fluvial systems, some of them braided (Jirásek et al., 2013b). They are concentrated in the most "continental" middle parts of the Lower Hrušov and Jaklovec members. However, the best example is the Castle Conglomerate Unit in the Gabriela cycle, which is located near the base of the Poruba Member (Jirásek et al., 2013b).

Where fluvial sandstones are absent, the genetic cycle consists of sandstone and mudstones alternating in centimetre-scale thin beds or laminae. These were deposited in a coastal setting established during lowstand and initial sea-level rise. These facies include mosaics of delta plain, tidal flat, and coastal lake environments (Fig. 13). Mudstones are rooted by stigmarian rhizophores with appendices and are typically weakly pedogenically altered. On top occur ~1-m-thick (max. 3-m-thick) and laterally widespread coalbeds and/or a few coals that are usually thin (Fig. 13). Coals are locally split by adjacent fluvial channels filled by sandstone, indicating that the peat swamps were deposited on coastal or delta plains separated by distributary channels (Filák, 2006). Sediment patterns found in the roof shales indicate peat drowning. They are usually grey mudstone with drifted remains of the aerial parts of plants, but may be thin bands of sapropelic coal or mudstone with a freshwater fauna (Havlena, 1964). Drowning due to initial sea-level rise, marked by a transgressive surface and followed by dark-grey mudstone or claystone with siderite nodules and brackish to marine fauna, indicate the base of the subsequent genetic cycle.

5. Discussion

5.1. Astronomical forcing of the depositional cycles in the Ostrava Formation

Gastaldo et al. (2009b) were only able to calibrate their genetic only in the lower part of the Ostrava Formation between the Ludmila and Karel coals (Fig. 7) based on high-precision data available at that time. The average duration (arithmetic mean) of the genetic cycles in this narrow stratigraphic interval is approximately ~100 kyr and was interpreted as a record of orbital forcing related to short-eccentricity. Two newly available dated stratigraphic intervals provide an excellent opportunity to test the consistency of the duration of the genetic cycles of Gastaldo et al. (2009b) throughout the remainder of the formation.

Of the six U-Pb CA-IDTIMS dated tonsteins or tuffite beds, two couples are close to each other, representing neighbouring genetic cycles (Figs. 7 and 15). Assuming that the duration of the cycle approaches 100 kyr, their ages are within the interval of uncertainty of the method employed in the current study. This is evident from the ages obtained from the Gabriela (365) – Eleonora (335) and Karel (106) – Main Ostrava Whetstone (MOW) (which yield differences of 0.06 and 0.20 Ma, respectively). Therefore, we calculated the arithmetic mean of both ages as a representative age of the stratigraphic level of each tonstein set (Fig. 10).

We subsequently tested the average duration of the genetic cycles in two other intervals: (4) Ludmila – Eleonora/Gabriela and (5) MOW/Karel – Eleonora/Gabriela. The durations of the genetic cycles in both of these intervals provide very similar results that are compatible with those of the Ludmila – Karel interval of Gastaldo et al. (2009b). The durations of cycles in intervals 1, 4, and 5 (Fig. 10) vary between 92.3 and 92.8 kyr. This calculated genetic cycle duration of 92.5 ka is consistent throughout most of the Ostrava Formation and probably represents a record of orbital-forcing related to short eccentricity cyclicity.

The re-evaluation of the cycle hierarchy recognition was beyond the scope of the current manuscript; also, it has been done multiple times. Here, we followed the genetic cycle recognition of Gastaldo et al. (2009b). One may argue that occasionally fifth and fourth order may be mixed in the genetic cycle recognition. In this, fifth order is close to or similar to the basic cycles, and fourth order represents the genetic cycles. For example, in Fig. 12, the sea level reconstruction shows three fourth-order sea-level variations from the base of the Gabriela to the base of the Lotar. However, there are five genetic cycles recognised in this interval. By doing this, one would underestimate the average duration of the fourth-order cyclicity, possibly making it closer to the expected value for short eccentricity. The opposite may, however, also have occurred with an extraordinary thick genetic cycle potentially comprising two less well-defined genetic cycles (Fig. 7). Therefore, we argue that our estimate of the average duration of the genetic fourth-order sequences in the Ostrava formation is likely fairly accurate.

Short eccentricity forcing can only occur via precession forcing, as eccentricity is the amplitude modulator of precession and has a

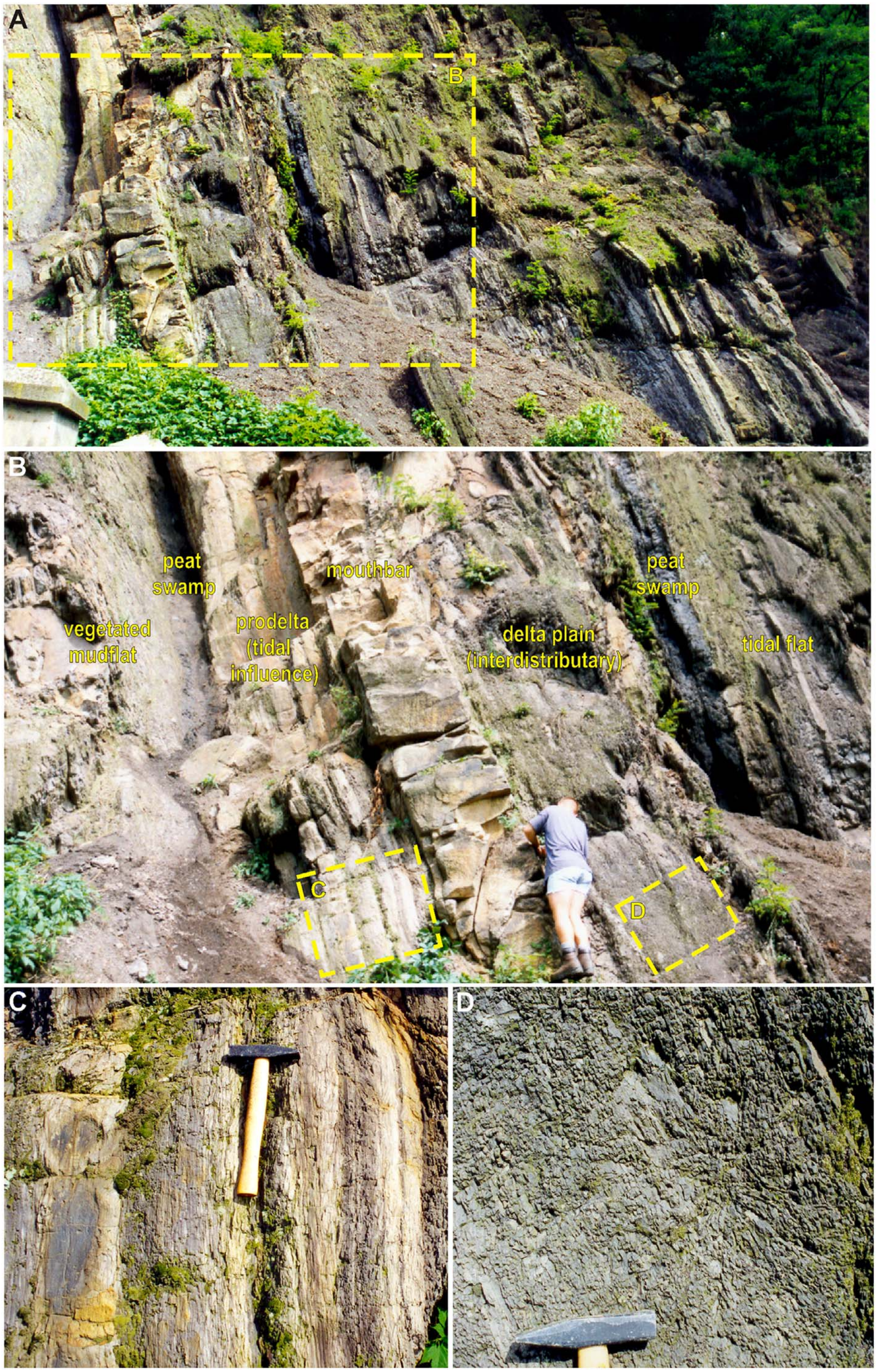

Fig. 13. Upper part of the Petřkovice Member around the Naneta marine band exposed along the Odra River on the Landek Hill in Ostrava. A – Overall view of the sedimentary succession exposed next to a gate at the Mine Museum (formerly the Anselm Mine) showing the rapid alternation of lithologies deposited in a coastal area; B – Detail of coastal succession with interpreted environments; C – Detail of heterolithic prodeltaic sediments with a possible tidal influence; D – Close-up view of mudstone with angular blocky structure that is densely rooted by *Stigmaria* directly below the coal seam.

negligible effect on Earth's insolation on itself. If only short eccentricity is observed in a record, it must originate from non-linear feedback mechanisms to the forcing. In the Ostrava Formation, the genetic cycles are one order higher than the basic cycles (Fig. 8). Basic sea-level cycle counting between the Gabriela base and Lotar base yield 15 sea-level cycles in 3 fourth-order, genetic cycles, if indeed the larger scale sea-level variations are traced as previously discussed. That would yield five fifth-order cycles per fourth-order cycle. In the Carboniferous, one would expect approximately six precession cycles per short eccentricity cycle. We recognised that five fifth-order per fourth-order cycles is in fairly good agreement with respective precession and short eccentricity forcing, as precession cycles are hardly developed during eccentricity minima and thus cycle counts are usually underestimations. This hypothesis must be further evaluated by detailed analysis of the basic cycles, including their lateral variability.

Evaluating the presence of long eccentricity forcing of 400-kyr length is more difficult with the current data. The megacycles and partial lithostratigraphic units seem to occur at this third-order time scale. More independent criteria have to be developed to trace this order of alternations, though fourth- and fifth-order cycle recognition would also greatly benefit from such criteria. There are seven to eight sequences in the Ostrava Formation that consist of coal-rich and coal-poor intervals (Fig. 7). These could well correspond to the long eccentricity forcing, though such a conclusion would be very premature.

The mechanism by which orbital forcing has driven these sedimentary sequences has been the subject of extensive discussion

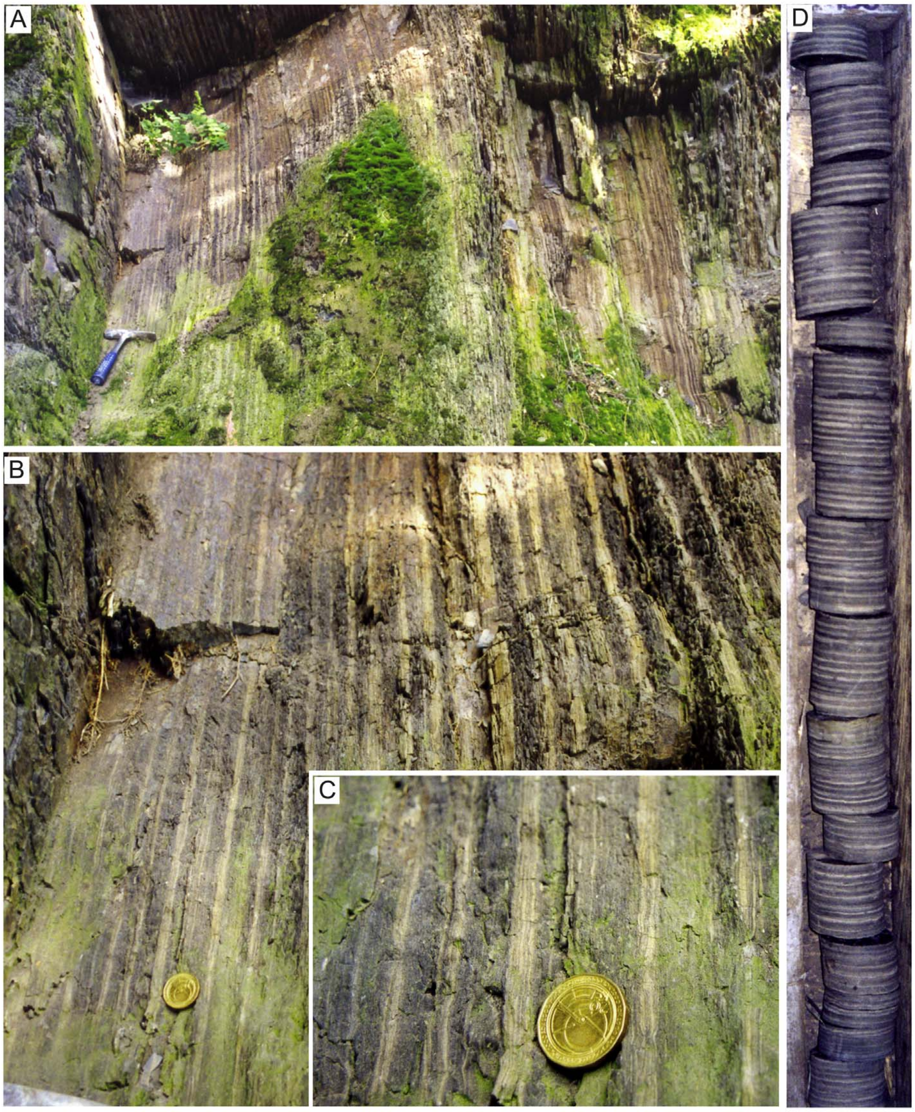

Fig. 14. Rhythmites of the Naneta marine band exposed along the Odra River on Landek Hill in Ostrava (A–C) and in a Paskov Mine borehole (D). A – View of the exposed rhythmites showing alternation of sandy and muddy laminae of centimetre-scale thicknesses, B + C – Detail of the rhythmites showing systematic thinning and thickening of mudstone/sandstone laminae, suggesting tidal origin related to "neap-spring" tidal cyclicity. Coin diameter is 26 mm. D – Sandstone – mudstone couplets showing rhythmic thickening and thinning of sandstone beds. The succession is 1-m-long, and the core diameter is 45 mm.

elsewhere and is beyond the scope and goal of the current study. Cycles were likely driven via glacioeustasy, which is the interpretation made for clastic Pennsylvanian successions throughout various basins in Europe and North America (Heckel, 1986; Izart et al., 2003; Davydov et al., 2010; Schmitz and Davydov, 2012; Pointon et al., 2012). Upstream controls via variable sediment input causing relative sea level changes should, however, not be ignored for the Ostrava Formation, especially because of the likely presence of precession forcing. Ice volume changes causing sea-level variability occurring at high latitudes are known to be mostly driven by obliquity and eccentricity, such as those during the Pleistocene. Little precession control is present in the Pleistocene sea-level variability, but a dominant precession control on the fifth-order, basic cycles in the Ostrava Formation would more likely connect these to upstream variability.

5.2. *Relative onlap-offlap changes constrained from faunas and facies*

Macrofaunal invertebrate remains in the Ostrava Formation are concentrated in thin bands interpreted as primary autochthonous biogenic accumulations (Řehoř and Řehořová, 1972). This, however, implies that spatial changes in faunal assemblages in and among individual bands of the Ostrava Formation (Řehoř, 1962, 1977, 1997; Řehoř and Řehořová, 1972) are due to changes in salinity and possibly also bathymetry. This interpretation is consistent with similar observations in other Carboniferous basins elsewhere in Europe and North America (Calver, 1968, 1969; Liu and Gastaldo, 1992; Paproth et al., 1983, 1994; Krzeszowska, 2015). Following these studies, Řehoř and Řehořová (1972) distinguished the relative salinity of the faunal bands based on the presence/absence of salinity-diagnostic species. The euhaline zone of "normal" salinity passes, as salinity decreases, to polyhaline, pleiohaline, meiomesohaline, and oligohaline to freshwater zones. Each zone is characterised by the presence of particular taxa or their combination (Fig. 3). A stenohaline fauna represents the fully marine conditions of the euhaline zone, whereas freshwater faunas are typical in freshwater to oligohaline zones. Brackish fauna, represented by the genera *Lingula* and *Orbiculoidea*, is typical for the meiomesohaline zone and adjacent parts of neighbouring salinity intervals (Řehoř and Řehořová, 1972; Gibson and Gastaldo, 1987). The euryhaline fauna includes taxa tolerant to changes in salinity (Fig. 3) that are dependent upon the intensity of marine transgressions but, in detail, are influenced by the morphology of coast and freshwater pulses, where rivers enter the sea (Demko and Gastaldo, 1996). In the case of the robust dataset obtained from ~1000 boreholes and a number of coal mines from the entire Czech part of the basin, these local anomalies in salinity are mapped (e.g., Řehoř and Řehořová, 1972). Only the maximum salinity observed in any individual faunal band is taken into account when reconstructing the magnitude of marine transgressions and the maximum landward shift of the coastline. In contrast, the maximum intensity of sea-level decrease in the intervals between faunal bands is interpreted based on the presence and characteristics of "terrestrial" lithofacies (e.g., rooted mudstones, coal or typical fluvial sediments lacking a tidal signature).

This approach allows us to use faunal bands and facies as proxies for the reconstruction of a relative onlap-offlap curve and, in turn, to interpret the magnitude of sea-level change. These are dominantly driven by eustasy; however, upstream climate change, tectonics and autocyclic mechanisms, including the compaction of sediments and mouth bars or distributary channel switching, are responsible for spatiotemporal variations in genetic cycles (Čepek, 1980; Gastaldo et al., 1993). Tectonic activity was the major cause of basin subsidence and, in turn, of accommodation space and the long-term preservation of sediments. The intensity of this subsidence decreased significantly from the western margin of the "synorogenic" basin to its eastern part (Dopita et al., 1997; Kędzior et al., 2007; Hýlová et al., 2013). Subsidence can be expressed as average depositional rates (Table 5) at the scale of the Ostrava Formation or its members over a temporal scale of thousands and millions of years. However, it has been suggested that repetitive tectonic (e.g., coseismic) pulses could generate base-level changes, and, in turn, accommodate sediments over a large area in a geological instant (Plafker and Savage, 1970; Gastaldo et al., 2004). For example, Phillips et al. (1994) and Phillips and Bustin (1996) reported the transgression of a Holocene coastal mangrove peat as the result of earthquake-induced subsidence. This, in turn, raises the question of whether tectonically induced subsidence could not serve as an alternative allocyclic mechanism generating or overprinting the genetic cycles of the Ostrava Formation. We, however, believe that the fairly consistent duration of genetic cycles over the tested intervals relates the cycles to astronomical cycles as the major control on their formation. It is, however, assumed that spatially and temporally variable tectonic subsidence is responsible for lateral and vertical changes in the thicknesses of individual genetic cycles.

The resulting onlap-offlap curve shows shifts in the coastline of various magnitudes and time scales. The shortest cycles are < 10^5 kyr because there are usually two, and locally up to four, individual cycles within one genetic cycle of Gastaldo et al. (2009b). Approximately 125 cycles are indicated in the onlap-offlap curve (Fig. 15), whereas Žídková et al. (1997) interpreted 269 basic cycles in the Ostrava Formation. The average duration of onlap-offlap cycles is 40 kyr and that of the basic cycles is ~18.5 kyr, based on the extrapolated ages of the Ostrava Formation presented here. The duration of these cycles approximates obliquity and precessional orbital frequencies and, thus, suggests possible allocyclic control. Autocyclic mechanisms (e.g., compaction) documented from similar sedimentary basins elsewhere in North America and Europe (Gastaldo et al., 1993; Demko and Gastaldo, 1996; Read, 1994; Read and Forsyth, 1989) certainly acted when generating the cyclic pattern of the Ostrava Formation at the finest scale. However, to distinguish between allocycles and autocycles at this scale is beyond the scope of this study.

The short eccentricity genetic cycles of Gastaldo et al. (2009b) are usually marked by more prominent peaks compared to the sub-eccentricity cycles they bracket. Both cyclicity scales are superimposed on a longer cyclic trend that overlaps with the megacycles of Jansa and Tomšík (1960) and Havlena (1964). Similar to the megacycles, individual sub-eccentricity cycles on the onlap-offlap curve follow long-term transgressive and regressive trends, separated by shorter intervals of unusually high ("super-highstand") or low ("super-lowstand") sea level. These higher-order highstands correspond with the major groups of marine bands of basin-wide extent separating individual members. These include the Štúr/Štolní (329.2 Ma), Naneta (327.6 Ma), Františka (326.7 Ma), Enna (325.9 Ma), Barbora (325.2 Ma), and Roemer/Gaebler (324.2 Ma). These bands range between 0.7 and 1 Ma apart, and we suggest that each likely represents the glacioeustatic expression of lower frequency eccentricity modulation in the Upper Silesian Basin (Olsen and Kent, 1999). They each preserve a fully marine and high-diversity autochthonous fauna including bivalves, gastropods, brachiopods, trilobites, and corals (Fig. 15), indicating normal salinities and relatively stable substrates with a restricted influx of suspended sediment at some distance away from coastal discharge. Such conditions could exist distal to the shoreline below the wave base (Demko and Gastaldo, 1996). The basic cycles in these intervals are lithologically less contrasting, having been deposited between offshore and inshore settings and lacking terrestrial facies.

Higher-order lowstands correspond to the most "terrestrial" part of the megacycles. The most prominent are found in the Lower Hrušov and Jaklovec members, where they span relatively long intervals of 327.3–326.8 Ma and 325.6–325.2 Ma, respectively. Faunal bands in these intervals are either absent or are of freshwater or oligohaline character (Fig. 15). The most prominent lowstand interval is, however, the Castle Conglomerate Unit in the lower part of the Poruba Member at ~324.95 ± 0.05 Ma. This interval is up to 115 m thick and was deposited in a braided fluvial system probably linked to a significant base-level decrease. The onset of the C2 glaciation interval in Australia, or a local tectonic event, was suggested as the driving force by Jirásek et al. (2013b).

5.3. Radioisotopically calibrated and astronomically tuned model of the Ostrava Formation

The consistency of the genetic-cycle duration throughout most of the Ostrava Formation allows for subsequent chronostratigraphic tuning on a linear time axis. Applying a 100-kyr periodicity provides a calibrated age model of the Ostrava Formation and its biostratigraphic record with a resolution of 10^5 years (Fig. 16). The extrapolated ages of the beginning and end of the Ostrava Formation are now constrained to fall within the interval 329.2–324.2 Ma. This suggests the deposition of this ~3-km-thick paralic succession over 5 Ma. This results in an average depositional rate of 63 cm/kyr for the most subsiding western part of the basin. The calibrated and astronomically tuned model of the Ostrava Formation also allows us to extrapolate the ages of individual members of the formation, their duration and, in turn, the depositional rates throughout the formation (Tables 4 and 5).

The Ostrava Formation comprises six megacycles (Jansa and Tomšík, 1960; Fig. 7), which, when the average sedimentation rate is applied, results in an average megacycle duration of 0.83 Ma. If we consider 269 basic cycles in the Ostrava Formation, as recognised by Žídková et al. (1997), each would be equivalent to 18.6 kyr, which approaches a precessional overprint with a length of 18.6 kyr (?) in the Carboniferous (Berger et al., 1992). Because our contribution is focused on genetic cycles, we leave this issue for future work. It is important to note that the tuning methodology of assigning 100 kyr to subsequent genetic cycles likely results in minimal estimates, as cycles may not be developed in intervals of low eccentricity related to longer periodicities.

5.4. Serpukhovian biozonal schemes - temporal constraints

The absolute time-calibrated and astronomically tuned age model of the Ostrava Formation enables us to apply numerical time constraints on faunal and floral biozones and regional chronostratigraphic substages.

5.4.1. Macrofloral zones

Detailed data on the stratigraphic ranges of plant species in the Ostrava Formation (e.g., Purkyňová, 1970, 1996, 1997) allow us to constrain the widely used floral zones of Wagner (1984) and Wagner and Alvarez-Vázquez (2010). Although already noted by Wagner (1984), they have been neither precisely constrained nor correlated with the local biozones of Purkyňová (1977) and Kotasowa and Migier (1995).

Plotting the stratigraphic ranges of key taxa against the lithostratigraphic subdivision of the formation and their comparison with Wagner

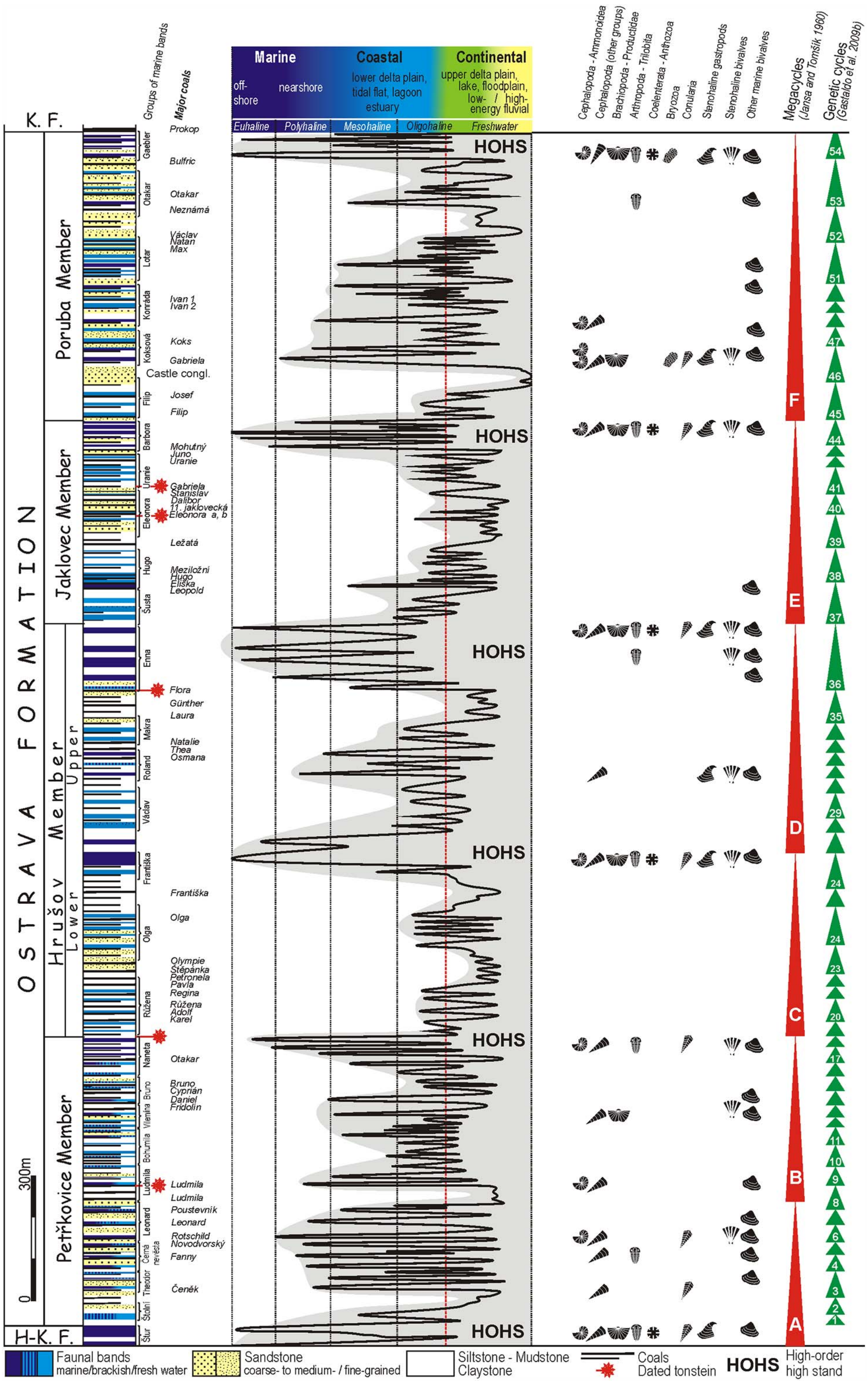

Fig. 15. Interpreted onlap-offlap curve for the Ostrava Formation derived from the characteristics of faunal assemblages and interpretation of the sedimentary environment, based on published data and the author's own sedimentological data.
Distribution of macrofauna compiled from the data of Řehoř and Řehořová (1972).

(1984) indicates that the macrofloras correspond to the *Lyginopteris stangeri* and *Lyginopteris larischii* zones (Figs. 5, 17). The base of the *L. stangeri* Zone is marked by the onset of this taxon and approximately coincides with the top of the Štúr group of marine bands and, thus, with the base of the Ostrava Formation. This position is in agreement with the sporadic presence of *Neuropteris antecedens* in the lower part of the Petřkovice Member, overlapping from the preceding zone of the same name. The Petřkovice and following Hrušov members contain species typical of the *Lyginopteris stangeri* Zone and are characterised by the presence of "Viséan" genera *Archeopteridium*, *Adiantites*, *Cardiopteridium*, *Adiantopteridium*, or *Eleuterophyllum* (Fig. 5). The base of the overlying *Lyginopteris larischii* Zone is marked by disappearance of these taxa as well as by the onset of *Lyginopteris larischii* and *Neuralethopteris schlehanii*, which are both typical of the *L. larischii* Zone (Fig. 5). This major floral overturn occurred at the base of the Jaklovec Member (Purkyňová, 1970, 1977, 1997) and is associated with the prominent marine transgression of the Enna group (Gastaldo et al., 2009a, 2009b). The top of the *Lyginopteris larischii* Zone, however, falls within the prominent hiatus between the Ostrava and Karviná formations, spanning approximately from the late Arnsbergian to the end of

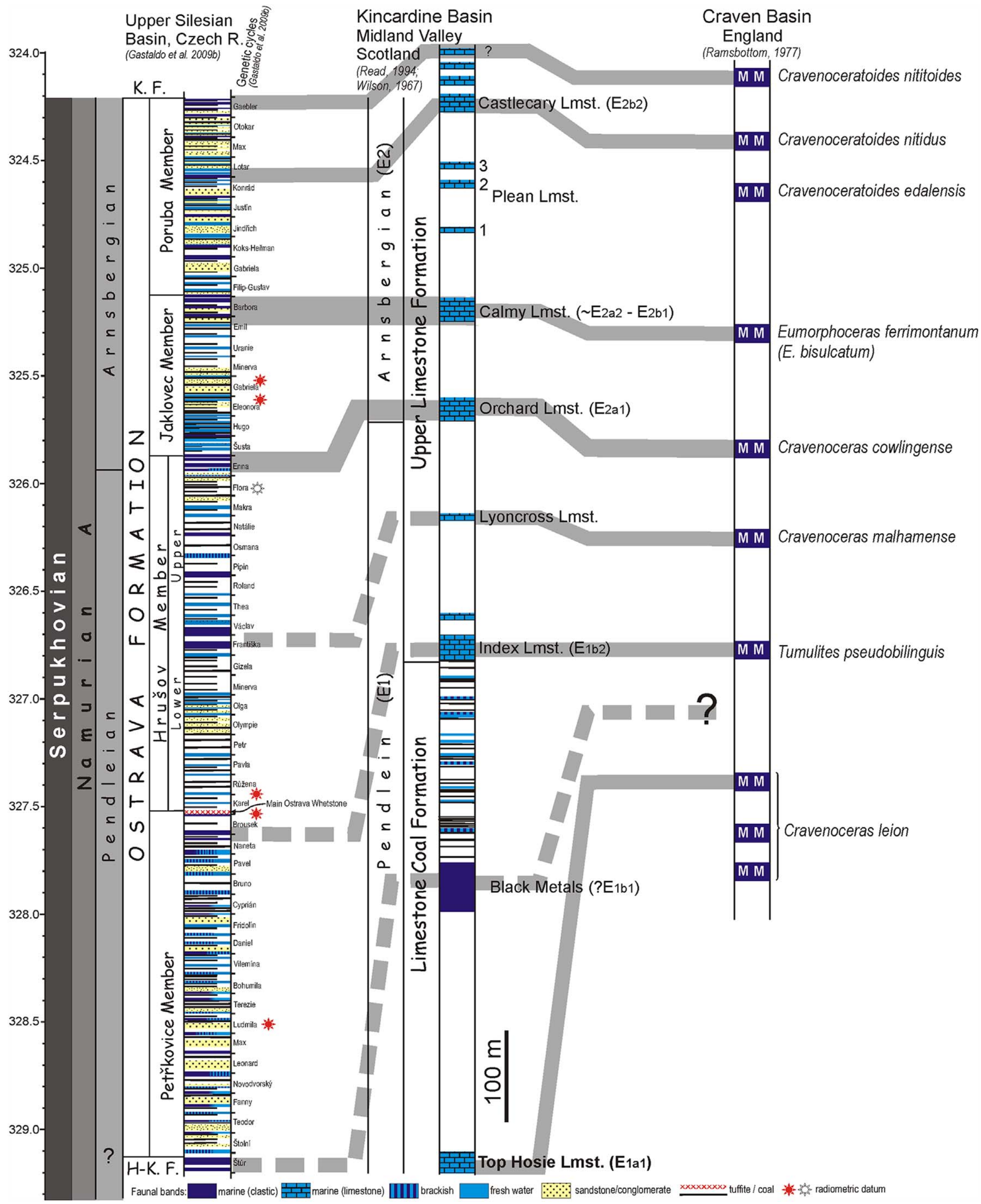

Fig. 16. Chronostratigraphically calibrated age model of the Ostrava Formation based on U-Pb radioisotopically interpreted durations of genetic cycles. The correlation of major marine bands of the Ostrava Formation with those on the British Isles is suggested based upon goniatite faunas.

the Alportian (Zdanowski and Żakowa, 1995). This gap is marked by a floral break identified by Gothan (1913b). The base of the Karviná Formation is already indicative of the "Namurian B" *Pecopteris aspera* Zone. Species typical of the Ostrava Formation, including *Sphenophyllum tenerrimum*, *Diplotmema adiantoides* (= *Sphenopteris elegans*), *Lyginopteris stangeri*, *L. bartonecii*, and *Lepidodendron veltheim* in the late Arnsbergian Jejkowice Member, occur within the gap between the formations in localised areas in the Polish sector of the basin. This indicates that the *L. larischii* Zone spans a significant part of the gap between the formations. The *L. stangerii* and *L. larischii* zones, as constrained here, correspond to the 2nd and 3rd floral zones of Purkyňová (1977, 1997) in the Czech part of the basin and to the IIIrd and IVth zones of Kotasowa and Migier (1995) in Poland (Fig. 17).

The fairly precisely identified stratigraphic position of the *Lyginopteris stangeri* and *L. larischii* zones within the radioisotopically calibrated Ostrava Formation allows us to calibrate their ages. The base of the *L. stangeri* Zone coincides with the base of the Ostrava Formation, which is now constrained to ~329.2 ± 0.2 Ma (2σ). The base of the following *L. larischii* zone corresponds with the boundary between the Hrušov and Jaklovec members with an extrapolated age of ~325.8 ± 0.2 Ma. The top of the zone, however, falls within the hiatus and is not recorded in the Czech part of the basin. In Poland, the locally preserved Jejkowice Member indicates the continuation of the zone during a time when no deposition or erosion prevailed over a

Table 4
Extrapolated ages of lithostratigraphic boundaries and marine bands of the Ostrava Formation.

Stratigraphy	No. of GC	Calculated age (Ma)
Top of the Ostrava Fm.	15.5	324.2
Base of the Poruba M.	5.5	325.1
Base of the Jaklovec M.	2.5	325.8
Františka marine band	12.5	326.8
Base of the Hrušov M.	20.5	327.5
Leonard marine band	33.5	328.7
Base of the Ostrava Fm.	38.5	329.2
Duration of the Ostrava Formation deposition		**5.00**

Fm. – formation, M. – member, GC – genetic cycle. Calculations related to the 335/365.

Table 5
Duration of deposition of members within the Ostrava Formation and depositional rates.

Member/ formation	Max. thickness (m)	No. of cycles	Duration of deposition (Ma)	Sedimentation rate (cm/kyr)
Poruba M.	900	10	0,93	97
Jaklovec M.	400	8	0,74	54
Hrušov M.	1100	18	1,67	66
Petřkovice M.	750	18	1,67	45
Ostrava Fm.	3150	54	5.00	58

major part of the basin. However, this member is not radioisotopically calibrated; therefore, constraining the exact top of the zone is currently not possible.

5.4.2. Marine invertebrate zones

Although over 200 species of Carboniferous invertebrates have been identified in the Czech part of the Upper Silesian Basin, only a few are of stratigraphic importance (Řehoř, 1977), with the most useful group being the goniatites. Those indicative of the E_1 Zone (Holdsworth and Collinson, 1988; Riley et al., 1995) have not been found in the Ostrava Formation. However, *Eumorphoceras pseudobilingue*, which is typical of the E_{1b2} marine band, occurs below the Štúr marine band in the upper part of the dominantly marine strata underlying the Paralic Series (the equivalent of the Ostrava Formation) in Poland (Zdanowski and Żakowa, 1995). This report is consistent with the presence of *Cravenoceras* (*Emstites*) *leion* and *Edmooroceras pseudocoronula* in the upper part of the stratigraphically equivalent Kyjovice Member in the Czech part of the basin (Fig. 2). The former is an index taxon for the base of the Pendleian (Namurian or E_{1a1} marine band) on the British Isles (Riley et al., 1995), and the latter is the key species for the same stratigraphic level in Germany (Korn, 1993, 1994). These occurrences suggest that the base of the Namurian occurs somewhere between the Kyjovice and the Bobrovníky horizons of the Kyjovice Member. This position also approaches the Viséan/Serpukhovian boundary (Fig. 17), which is currently located at 330.9 Ma (Schmitz and Davydov, 2012; Davydov et al., 2012), although its precise age cannot be estimated from our data.

The sporadic occurrence of goniatites indicative of the E_2 Zone, including *Cravenoceras cowlingense*, starts in the Enna Marine Band in the Polish part of the basin (Zdanowski and Żakowa, 1995). In the Czech part, this index taxon, which is used to identify the base of the Arnsbergian Substage, was reported by Řehoř (1977) from the base of the Poruba Member. A more diversified assemblage of goniatites of the E_{2b} Zone (*Cravenoceratoides edalensis*), however, has been found in the Gaebler group of marine bands at the top of the Ostrava Formation,

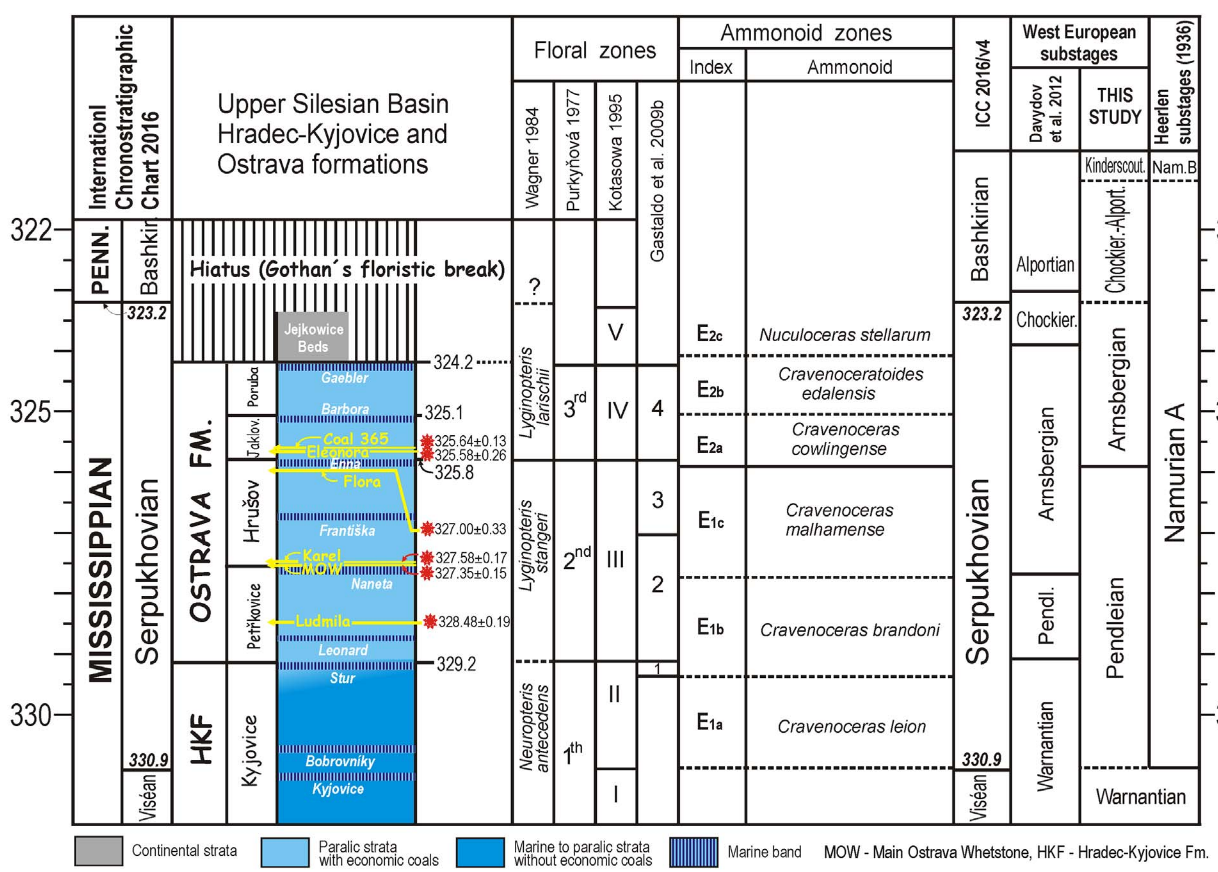

Fig. 17. Stratigraphic distribution of macrofloral species within the Ostrava Formation and ranges of macrofloral and ammonoid biozones. Ranges of regional and global stages are suggested.

constrained to ~324.2 Ma. The presence of *Cravenoceratoides nitidus*, *C. nititoides*, and *Eumorphoceras leitrimense* indicates a position in the E_{2b2} to E_{2b3} subzones (Vašíček, 1983; Řehoř, 1997). This implies that the highest Arnsbergian subzone E_{2c} probably falls within the gap between the Ostrava and Karviná formations, although *Nuculoceras stellarum* at the bottom of the E_{2c} Subzone was reported from the Gaebler group of marine bands by Zdanowski and Żakowa (1995) in Poland. These goniatite associations led Řehoř (1997) to the conclusion that the Pendleian/Arnsbergian boundary is most likely at the base of the Jaklovec Member, which is here calibrated to have an age of 325.8 Ma. However, we prefer to place it at the base of the Enna group of marine bands in the upper part of the Hrušov Member (~325.9 Ma), which is in agreement with macrofauna from Poland (Fig. 17). This age, however, does not correspond with the ~327.7 Ma position of this boundary, as interpreted in the Donets Basin (Davydov et al., 2010) and adopted by Davydov et al. (2012). The top of the Arnsbergian Substage can only be approximately constrained from our data because the youngest E_{2c} Zone is missing and the extrapolated age of the top of the Ostrava Formation of ~324.2 Ma corresponds to the E_{2b3} marine band. This age is very close to the 324.54 ± 0.26 Ma age of the B9 bentonite from the E_{2b2} marine band in the Pennine Basin (Pointon et al., 2012).

5.5. Correlation of the Ostrava Formation with contemporaneous successions in the UK

It is assumed that significant changes in sea level in the Ostrava Formation are of eustatic origin. If so, similar eustatic oscillations should be recorded in contemporaneous basins in Europe. To test this assumption, we attempt to correlate major eustatic events on our onlap-offlap curve (Fig. 15) with similar sea-level changes recorded in basins of the British Isles. The Pennine Basin is a classic area of high-resolution ammonoid zonation (Ramsbottom, 1974, 1977a, 1977b, 1979; Ramsbottom et al., 1978; Holdsworth and Collinson, 1988; Riley et al., 1995); Scotland is of a similar depositional environment that generated paralic coal-bearing cyclothems (Read, 1988, 1994; Read and Forsyth, 1989). Both areas show cyclicity of several orders considered by many authors to be driven by eustatic sea-level fluctuations (e.g., Holdsworth and Collinson, 1988; Maynard and Leeder, 1992; Martinsen et al., 1995; Hampson et al., 1997; Waters and Davies, 2006), representing a far-field response to climatically driven changes in a polar ice-sheet in southern Gondwana (Veevers and Powell, 1987; Isbell et al., 2003; Fielding et al., 2008a, 2008b; Gulbranson et al., 2010). These areas are an appropriate target for comparison to the Ostrava Formation to test eustatic control and to improve correlation among these basins.

In the Pennines, early Namurian (~Serpukhovian) strata are represented by the lower part of the Millstone Grit, which is a complex of marine to non-marine sandstones and shales showing cyclic repetition of several orders. Invertebrate faunas are concentrated into marine bands interpreted as transgressive systems tracts and maximum flooding surfaces (Holdsworth and Collinson, 1988; Waters and Condon, 2012). Many of these marine bands have been identified in other basins in Scotland, Ireland, Northern France, Belgium, the Netherlands, and Germany (Ramsbottom, 1979), indicating that eustasy was the primary control of formation (Waters and Condon, 2012). The estimation of their periodicities depends on the precision of existing radioisotopic data and varies between 185 ka (Holdsworth and Collinson, 1988) and 65 ka based on SHRIMP U-Pb zircon dates (Riley et al., 1995). The recent application of high-precision U-Pb LA/CA-ICP-MS dating (Pointon et al., 2012; Waters and Condon, 2012) is principally in agreement with those previously presented and suggests short eccentricity control. Pointon et al. (2012) calculated the average periodicity of Arnsbergian to early Langsettian sedimentary cycles to range between 89 and 124 ka. Similarly, Waters and Condon (2012) constrained the average durations of the Pendleian–early Arnsbergian and Chokierian–Bolsovian cycles to be ~111 and ~150 ka, respectively. The existence of a superimposed sub-100 ka obliquity or precession cyclicity has been proposed within the Pennine Basin by Waters et al. (2008) and Tucker et al. (2009) and is supported by the presence of multiple flooding surfaces with the same ammonoid assemblages. In addition, long-period cycle units, separated by major eustatic events and so-called mesothems, were identified by Ramsbottom (1977a, 1979), who estimated their average duration to be between 1.1 and 1.35 Ma.

The ~800-m-thick Limestone Coal (Pendleian) and Upper Limestone (Pendleian – Arnsbergian) formations in the Midland Valley of Scotland (Fig. 16) record widespread eustatic sea-level oscillations with two or more periodicities (Read and Forsyth, 1989; Read, 1988, 1994). The longer periodicity resulted in six major marine transgressions with an interval of ~1.1 Ma and included the marine bands of Hosie, Black Metals, Index, Orchard, Calmy and Castlecarry. Each of these bands comprises thick and widespread limestone beds, except the Black Metal, which is a hemipelagic shale (Fig. 16). The facies between these major transgressions include fluviodeltaic environments generally controlled by high-frequency allocycles, related to the higher-frequency Milankovitch orbital parameters (Read, 1994). As a result, an apparently cyclic succession of laterally persistent fluvio-deltaic cyclothems, with economical coals (~1 m) and fresh- to brackish-water faunal bands in their roof shales in areas far from strong fluvial influence, developed. Ramsbottom (1977a, 1977b) correlated the major marine transgressions in Scotland with the "black shales" of the southern Pennines, which contain thick-shelled ammonoids.

The correlation of the Ostrava Formation with the Millstone Grit in the Pennine Basin is based on the presence of the key taxa *Cravenoceratoides nitidus*, *C. nititoides*, and *Eumorphoceras leitrimense* in the Gaebler group of marine bands. Their occurrence suggests a correlation with the E_{2b2} to E_{2b3} subzones (Vašíček, 1983; Řehoř, 1997). A correlation with the E_{2b3} is preferred because of the presence of *C. nititoides*, which is the key taxon of the marine band. This interpretation also better fits the extrapolated age of the Gaebler group of marine bands (324.2 Ma) when compared to the 324.54 ± 0.26 Ma age of the B9 bentonite from the underlying E_{2b2} marine band in the Pennine Basin (Pointon et al., 2012). Thus, the E_{2b2} *Cravenoceratoides nitidus* marine band correlates with some of the marine bands in the middle part of the Poruba Member (e.g., Lotar group of marine bands). In contrast, the basal Arnsbergian E_{2a1} band with *Cravenoceras cowlingense* corresponds to the Enna group of marine bands (~325.9 Ma) on top of the Hrušov Member (Figs. 16, 17), based on findings of this taxon in the Polish part of the basin.

The correlation of major marine bands in the upper part of the Ostrava Formation with those in the Midland Valley of Scotland is based on the correlation of Ramsbottom (1977b) of the latter area with the Pennines. This suggests that the Castlecarry Limestone in Scotland may correspond to the Lotar group of marine bands in the Czech Republic and Poland, which is the major transgression in the middle part of the Poruba Member (Fig. 16), whereas the bands above the Castlecarry (~in stratigraphic position of the *C. nititoides*) probably correspond to the Roemer and Gaebler marine bands at the top of the Ostrava Formation. Another major group of marine bands lower in the section of the Ostrava Formation is the Barbora. Its stratigraphic position and invertebrate fauna best correlate with the Calmy Limestone (E_{2a2}–E_{2b1}) in the Midland Valley of Scotland. Similarly, ammonoid faunas of the Enna marine bands at the top of the Hrušov Member, marking the base of the Arnsbergian in the Upper Silesian Basin, correlate with the Orchard Limestone in Scotland.

Due to the absence of key ammonoid taxa, the correlation of the lower part of the Ostrava Formation is less clear. We suggest the correlation of the Františka group of marine bands in the Hrušov Member (extrapolated age ~326.7 Ma) with the Lyoncross Limestone of Scotland and the *Cravenoceras malhamense* band (E_{1c1}) of the Pennine. Similarly, the group of Naneta marine bands on top of the Petřkovice Member probably correlates with the Index Limestone of Scotland and the *Tumulites pseudobilinguis* band (E_{1b2}) of the Pennines, whereas the

Štúr group of marine bands is located at a stratigraphic position that most likely corresponds to the Black Metals marine band in Scotland. The basal Pendleian marine band bearing *Cravenoceras leion* (E_{1a1}) in the Pennines and its equivalent Top Hosie Limestone in Scotland correlate with the Bobrovníky Faunal Horizon in the Kyjovice Member (Kumpera, 1983). Its depth below the base of the Ostrava Formation, is, however, difficult to estimate because of the absence of reliable borehole data in this non-coal-bearing succession. This correlation is supported by the presence of the key taxon in all three areas.

6. Conclusions

The Ostrava Formation in the Czech part of the Upper Silesian Basin (= Paralic Series in Poland) represents the most complete coal-bearing Serpukhovian succession of equatorial Pangea. This ~3-km-thick unit exhibits a cyclic pattern of several superimposed orders which are interpreted as a record of glacioeustatic sea-level fluctuation of a far-field response to high-latitude glaciation. This study provides three new high-precision U/Pb ages of 327.00 ± 0.33 Ma, 325.58 ± 0.26 Ma, and 325.64 ± 0.13 Ma from volcanic ash beds (tonsteins) intercalated in coals. These new radioisotopic ages, together with existing ones with the same precision, allow us to test the length of genetic cycles recognised by Gastaldo et al. (2009a). The test confirmed a consistent ~100-kyr duration of the genetic cycles throughout the calibrated part of the formation. Thus, the genetic cycles represent a record of short-eccentricity orbital forcing and allow the tuning of these cycles and, in turn, the Ostrava Formation, on a linear time axis. The resulting calibrated and tuned model has a profound impact on the temporal constraints of this unit and its depositional rates, as well as on biozones and regional substages and their correlation to global stages. Particular conclusions are as follows:

The Ostrava Formation was deposited during a 5-My interval, with extrapolated Serpukhovian ages of its onset and termination of 329.2 and 324.2 ± 0.50 Ma, respectively. This interval is shorter than the previous estimate of Jirásek et al. (2013b) and suggests an average depositional rate of the formation of ~0.6 mm/y. However, the depositional rates of individual members within the formation varied between 0.45 and 0.97 mm/y, with the highest average sedimentation rate identified in the youngest Poruba Member.

At least three superimposed orders of cyclicity have been identified. Genetic cycles are units recording consistent periodicity at ~100 kyr throughout the calibrated part of the formation. Each genetic cycle comprises two, or mostly more, sub-eccentricity cycles, suggesting obliquity or precession cyclicity control. The major cycles (megacycles) marked by principal marine bands vary between 0.7 and 1 Ma, which may represent long-period eccentricity modulation.

1) The radioisotopically calibrated model of the Ostrava Formation allows us to constrain the Pendleian/Arnsbergian boundary to approximately 325.9 Ma.
2) The rich macroflora of the Ostrava Formation enables us to define floral biozones used for Carboniferous strata in other parts of Europe. The base of the *Lyginopteris stangeri* Zone coincides with the base of the Ostrava Formation and is constrained to ~329.2 Ma. The top of the zone corresponds with the base of the Jaklovec Member and is constrained to ~325.8 Ma, which is also the age of the base of the stratigraphically higher *L. larischii* Zone. Its top lies in the hiatus following the Ostrava Formation and can be only roughly constrained to approximately 323.0 Ma.
3) The presence of key taxa in some marine bands of the Ostrava and adjacent parts of the underlying Hradec-Kyjovice Formation allows us to approximate the stratigraphic position of the ammonoid zones in these units. The correlation of the major Serpukhovian marine bands of the Upper Silesian basin with those in the Pennine Basin and the Midland Valley of Scotland is suggested.
4) Based on ammonoid fauna and macroflora, the Pendleian/Arnsbergian boundary is estimated to be at approximately 326 Ma, which contrasts with the age of ~328 Ma presented by Davydov et al. (2012).

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Appendix A. Supplementary data

Supplementary data associated with this article can be found in the online version, at https://doi.org/10.1016/j.earscirev.2017.12.005. These data include the Google maps of the most important areas described in this article.

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