GEOLOGICA CARPATHICA, AUGUST 2015, 66, 4, 311–329 doi: 10.1515/geoca-2015-0028

Assessing provenance of Upper Cretaceous siliciclastics using spectral γ-ray record

DANIEL ŠIMÍČEK[1]✉ and ONDŘEJ BÁBEK[1,2]

[1]Department of Geology, Palacký University, 17. listopadu 1192/12, 771 46 Olomouc, Czech Republic; ✉daniel.simicek@upol.cz
[2]Department of Geological Sciences, Masaryk University, Kotlářská 2, 611 37 Brno, Czech Republic; babek@prfnw.upol.cz

(Manuscript received January 13, 2015; accepted in revised form March 12, 2015)

Abstract: The relationship between contents of clay minerals/grain size and spectral γ-ray record (concentrations of K, U and Th) in sediments is used for interpretation of sedimentary facies in wire-line logs. However, this approach is often complicated by the multi-component nature of mineralogically immature siliciclastics. As mineralogy of the source material and grain-size sorting during transport both contribute to the detrital composition of the final sediment, a joint study of facies and outcrop γ-ray spectra can potentially make the latter an effective tool in provenance studies. This paper provides comparison of outcrop γ-ray data and detailed facies mapping with mineral and chemical composition of the rocks (modal composition; transparent heavy mineral assemblages; WDX SEM chemistry of minerals) and interprets them in terms of provenance changes. We studied the Upper Cretaceous, synorogenic siliciclastic sediments of the Mazák and Godula Formations (Silesian Unit of the Western Carpathians flysch belt). Decreasing mineral maturity of the studied sandstones is consistent with provenance change from craton interior — (Mazák Formation) to transitional continental and recycled orogen sources (Godula Formation). Two major phases of K, U and Th concentration shifts, which occurred close to the Mazák/Godula Formation and Middle/Upper Godula Members boundaries, are consistent with changes in main detrital modes. These trends indicate gradually accelerated influx of material derived from high-grade metamorphic and plutonic rocks during deposition of the Mazák and Godula Formations. These changes are interpreted as reflecting a gradual exhumation and erosion of deeper crustal levels of the source area, the Silesian ridge.

Key words: Western Carpathians, gravity-flow sediments, geochemistry, heavy minerals, γ-ray spectrometry, provenance.

Introduction

Gamma-ray (γ-ray) spectrometry (GRS) is a widely used petrophysical method, which is particularly useful for identification of facies, stratigraphic correlation and sequence-stratigraphic interpretations both below and on the surface (Van Wagoner et al. 1990; Slatt et al. 1992; Catuneanu 2006; Šimíček & Bábek 2015). The fundamental principle standing behind these applications is the effect of variable concentrations of K, Th (and sometimes U) bearing clay minerals and non-radioactive, sand-sized quartz grains in siliciclastic sediments, which render higher radioactivity to muddy sediments and low radioactivity to sands (Doveton 1994; Rider 1999). However, the multi-component nature of chemically poorly mature sediments complicates the use of GRS as a direct proxy of grain size, as numerous minerals typically contained in different grain size fractions can act as carriers of K, Th and U. As examples, we can mention silt-sized heavy minerals (zircon, apatite, monazite, thorite, etc.) as carriers of U and Th, clay minerals with highly variable contents of K (kaolinite, smectite, illite, etc.) and sand-sized K-feldspars, albite and micas with relatively high contents of K (Rider 1999; Svendsen & Hartley 2001; Šimíček et al. 2012). Mineral inclusions, organic matter and chemical precipitates further complicate the overall picture (cf. Rider 1999). Consequently, spectral γ-ray data can be interpreted in terms of sediment grain-size only with the knowledge of mineral composition and, the other way round, spectral γ-ray combined with grain-size data can be used as a proxy for changes in mineralogy and chemistry of detrital sediments (Šimíček et al. 2012). Deep-marine flysch siliciclastics of remnant and foreland basins, often characterized by low chemical maturity are particularly useful for studies of material provenance because their modal composition closely reflects the rock composition of their source area (Miall 1995; Critelli et al. 1997; Šimíček et al. 2012).

This paper is focused on the Upper Cretaceous deep-marine siliciclastics of the Mazák and Godula Formations, which represent well-exposed parts of the Silesian Unit, in the Outer Western Carpathians of Central Europe. The thrust sheets of the Silesian Unit are inverted relicts of sedimentary fill of the Silesian basin, which formed a part of the Outer Carpathians basin system. Synorogenic sedimentation in the Silesian basin commenced with the Late Cretaceous compressive tectonic regime in the Outer Carpathians (Golonka et al. 2000; Oszczypko 2004). The Silesian Unit is now preserved as a rootless nappe (Stráník et al. 1993; Picha et al. 2006). Due to the detachment of the inverted Silesian basin from its basement in the Miocene (Oszczypko 1999), any information about the source area in the basin are based entirely on studies of chemical, mineral and pebble composition (Książkiewicz 1962; Unrug 1968; Wieser 1985; Ślączka 1998; Golonka et al. 2000; Hanžl et al. 2000; Michalik et al. 2004; Poprawa et al. 2004). The sediment composition, facies and paleocurrent data indicate a predominant transport of material from the hypothetical Silesian ridge, which was situated on the southern margin of the basin (Książkiewicz 1962; Menčík et al. 1983; Oszczypko 2004; Słomka et al.

2004; Malata et al. 2006). Periods of increased sediment supply in the Outer Carpathians basins, which are indicated by the presence of clasts of "exotic rocks" (Unrug 1968; Grzebyk & Lesczyński 2006), always coincided with the main tectonic phases of the Carpathian orogeny (Książkiewicz 1977; Poprawa et al. 2006). The changes in depositional style of the siliciclastic turbidites therefore coincided with the source area changes, which in turn might have been recorded in the γ-ray characteristics of the sediments. A detailed spectral γ-ray logging and facies mapping can be used to trace the provenance changes related to important geotectonic events (cf. Šimíček et al. 2012). The aim of this paper is to achieve an insight into the relationship between facies, modal composition and γ-ray spectra in deep-marine siliciclastics and contribute to the discussion about the Late Cretaceous synorogenic evolution of the Outer Western Carpathians (cf. Menčík et al. 1983; Kováč et al. 1998; Golonka et al. 2000; Plašienka 2000; Haas & Csaba 2004).

Geological setting and stratigraphy

The Silesian Unit constitutes a part of the Menilite-Krosno group of superficial thrust sheets of the Western Carpathians flysch belt (Golonka et al. 2006). It plunges beneath the front of the Magura thrust sheets to the east and southeast and overlaps the flysch sediments of the Subsilesian nappe and Miocene deposits of the Carpathian Foredeep to the northwest (Lexa et al. 2000; Picha et al. 2006). Its sediment succession spanning Upper Jurassic to Lower Miocene, which is

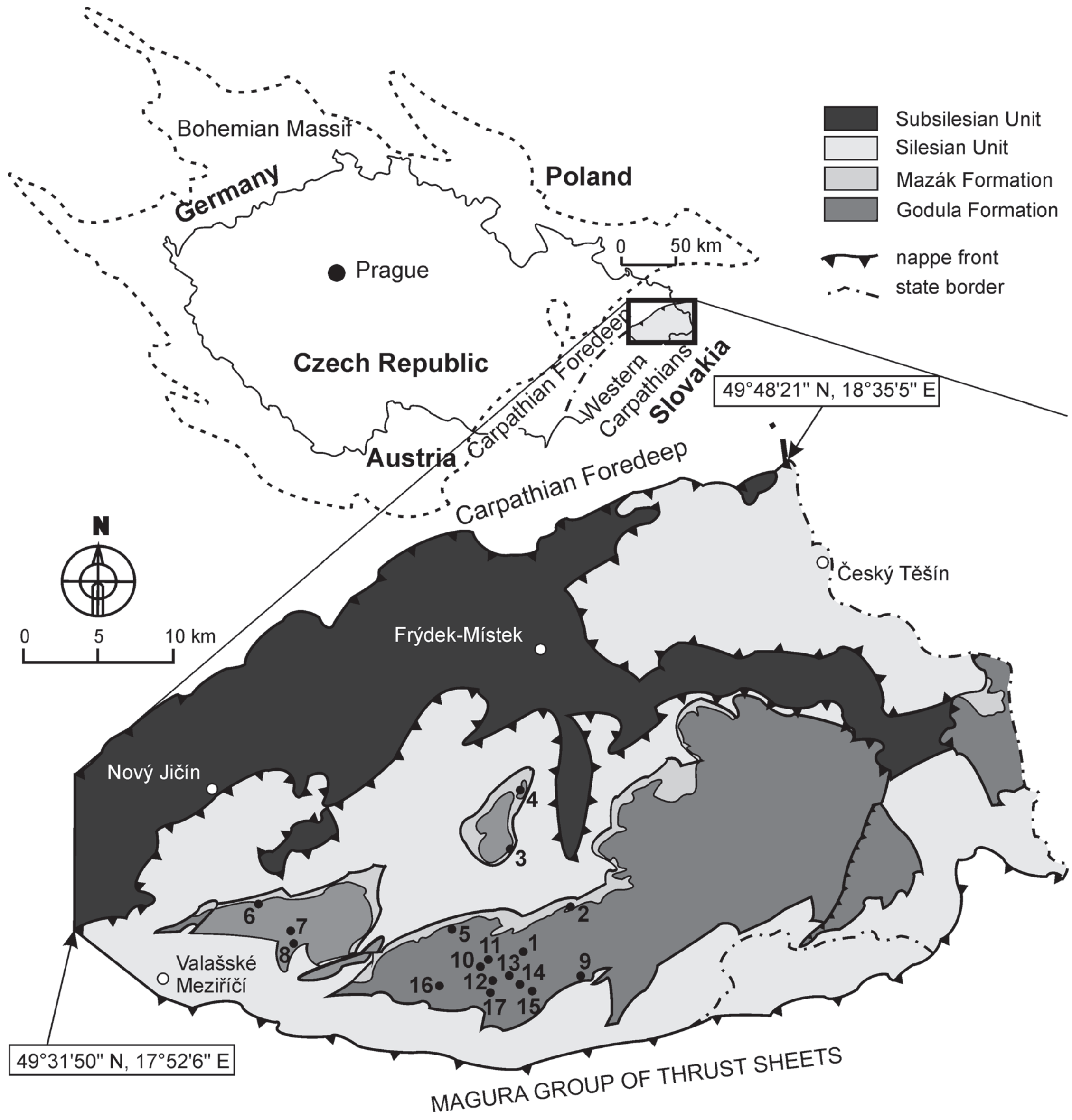
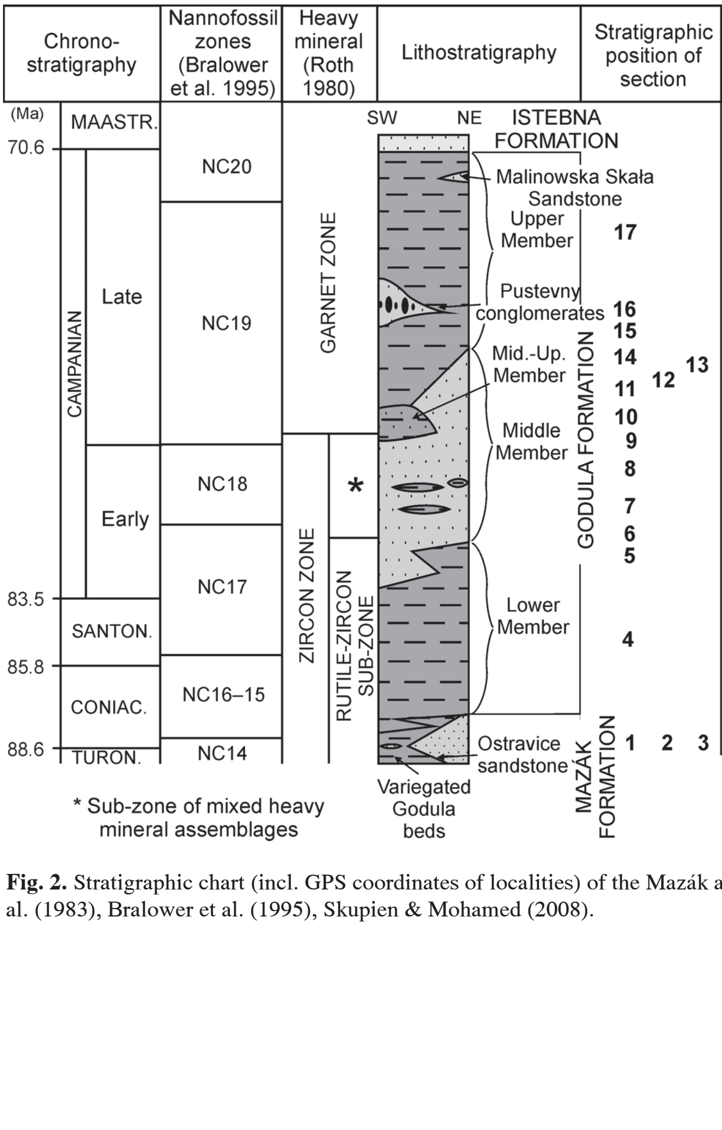


Fig. 1. Geological sketch of the Silesian Unit in the Czech Republic. **1** — Čeladná, **2** — Mazák Quarry, **3** — Ondřejník 2, **4** — Ondřejník 1, **5** — Trojanovice Quarry, **6** — Mořkov, **7** — Zubří 1, **8** — Zubří 2, **9** — Šance, **10** — Malinová, **11** — Pustevny 1, **12** — Pustevny 2, **13** — Magurka 2, **14** — Magurka 1, **15** — Čeladenka, **16** — Horní Rozpité, **17** — Pustevny 3.

up to 6 km thick, represents the most complete stratigraphic record in the Outer Carpathians Silesian basin. It was inverted and finally thrust over the Bohemian Massif foreland during the Miocene Styrian tectonic phase (Oszczypko 1999; Picha et al. 2006). We can distinguish three facies domains in the Silesian Unit: (I) Godula representing basin floor deposits, (II) Baška representing base of the slope sediments and (III) Kelč facies domain representing slope facies (Eliáš 1970; Menčík et al. 1983). The Godula facies domain, which is most extensive, is composed of the Vendryně, Hradiště, Veřovice, Lhota, Mazák, Godula, Istebna, Rožnov, Menilite and the youngest Krosno Formations (Matějka & Roth 1954; Eliáš 1970, 2002; Menčík et al. 1983; Eliáš et al. 2003). The Upper Cretaceous Mazák and Godula Formations (focus of this paper; Fig. 1) with maximum thickness of 3 km represent the best-exposed part of the Godula facies domain (Roth 1980).

The Mazák Formation (formerly "Variegated Godula Beds" — Menčík et al. 1983) of Middle Cenomanian–Coniacian (Skupien et al. 2009) is equivalent to the "Variegated shales" in Poland (Golonka 1981). The overlying Godula Formation is composed of Lower Godula (Santonian–Lower Campanian), Middle Godula (Lower–Upper Campanian) and Upper (Uppermost Campanian) Godula Members (Menčík & Pešl 1955; Skupien & Mohamed 2008). The biostratigraphy of both formations is based on agglutinated foraminifers and dinoflagellates (Olszewska 1984; Skupien & Mohamed 2008; Skupien & Smaržová 2009; Skupien et al. 2009). An alternative stratigraphic scheme is based on heavy-mineral zones (Roth 1980). The lower zircon zone, subdivided into rutile-zircon subzone and subzone of mixed assemblages, corresponds to the Mazák Formation, Lower Member and lower part of the Middle Member of the Godula Formation. Zircon, tourmaline and rutile predominate in the heavy minerals assemblages of this zone. The upper garnet zone with predominance of garnets, zircon and rutile comprises the upper part of the Middle and the Upper Members of the Godula Formation (Fig. 2).

The Mazák Formation is composed predominantly of red pelagic mudstones (oceanic red beds *sensu* Hu et al. 2005), which alternate with green-grey mudstones and thin layers of fine-grained turbidite sandstones (Golonka 1981; Menčík et al. 1983; Lemańska 2005; Jiang et al. 2009). This pelagic to hemipelagic sedimentary succession (Koszarski 1963) locally alternates with thick-bedded, medium- to coarse-grained sandstones and fine-grained conglomerates (Ostravice sandstone), which are interpreted as sediments of prograding submarine fans (Eliáš 1995; Picha et al. 2006). The Godula Formation comprises predominantly deep-marine gravity-flow deposits (Geroch & Koszarski 1988). The Lower and Upper Members are both characterized as predominantly thin-bedded successions of grey mudstones and fine- to medium-grained, glauconite-rich sandstones (Picha et al. 2006; Skupien & Smaržová 2009). They are interpreted as sediments of the middle zone (channel-to-lobe transition) and outer zone (lobe and wedge



Number and name of localities	GPS coordinates of localities
1 - Čeladná	49°30'32.624" N, 18°19'0.272" E
2 - Mazák Quarry	49°31'29.223" N, 18°24'22.749" E
3 - Ondřejník 2	49°33'27.010" N, 18°18'55.760" E
4 - Ondřejník 1	49°34'44.059" N, 18°18'55.954" E
5 - Trojanovice Quarry	49°30'10.263" N, 18°14'49.789" E
6 - Mořkov	49°31'8.278" N, 18°4'52.062" E
7 - Zubří 1	49°29'52.060" N, 18°6'27.920" E
8 - Zubří 2	49°29'58.958" N, 18°6'26.766" E
9 - Šance	49°29'49.889" N, 18°24'55.430" E
10 - Malinová	49°28'51.192" N, 18°17'1.858" E
11 - Pustevny 1	49°28'39.652" N, 18°17'9.039" E
12 - Pustevny 2	49°28'27.691" N, 18°16'59.093" E
13 - Magurka 2	49°28'11.870" N, 18°20'14.055" E
14 - Magurka 1	49°28'13.006" N, 18°19'3.160" E
15 - Čeladenka	49°27'14.387" N, 18°21'19.706" E
16 - Horní Rozpité	49°27'37.649" N, 18°13'44.833" E
17 - Pustevny 3	49°28'17.124" N, 18°16'54.807" E

Fig. 2. Stratigraphic chart (incl. GPS coordinates of localities) of the Mazák and Godula Formations modified after Roth (1980), Menčík et al. (1983), Bralower et al. (1995), Skupien & Mohamed (2008).

margin) of a sand-rich submarine fan (Menčík et al. 1983; Słomka 1995; Eliáš 2000). The Middle Member represents a medium- to thick-bedded sandstone succession with minor occurrence of heterolithic facies and mudstones, which was deposited by various gravity flows in the middle zone (distributory channels, lobes) of a sand-rich submarine fan system (Eliáš 1970; Menčík et al. 1983; Leszczynski 1989; Słomka 1995; Picha et al. 2006; Skupien et al. 2009).

It is assumed that the Silesian basin was isolated from other Outer Carpathians depocenters by two morphological elevations, the Baška ridge to north and the Silesian ridge to the south (Menčík et al. 1983). Paleocurrent data, facies distribution and clast composition data indicate that the sediments of the Godula facies domain were sourced from areas situated along the southern border of the basin (Unrug 1968; Golonka et al. 2000; Picha et al. 2006) and further dispersed by bottom currents in a basin-axis direction (Eliáš 1970; Oszczypko 2006). The Silesian ridge consisted of Cadomian to Variscan crystalline basement and its Upper Paleozoic to Mesozoic sedimentary cover (Unrug 1968; Słomka et al. 2004; Golonka et al. 2008). The intensity of tectonic activity in the Silesian ridge (Golonka & Krobicki 2006) along with the tectonic basin subsidence (Roth 1980; Menčík et al. 1983) and global sea-level changes (Słomka 1995; Oszczypko 2004) are considered as the major causes of switching between coarse-grained deposits of prograding and laterally migrating submarine fans and fine-grained background hemipelagic sedimentation (Eliáš 1970). Modal composition of sandstones of the Godula Formation indicates a gradual decrease in the input from sedimentary and meta-sedimentary sources and increased supply of high-grade metamorphic and plutonic detritus up-section (Unrug 1968; Krystek 1973; Žůrková 1975; Eliáš 2000; Grzebyk & Lesczyński 2006). This trend is explained by progressive exhumation of structurally deeper crustal parts of the Silesian ridge (Unrug 1968; Słomka et al. 2004).

Material and methods

Concentrations of K, U and Th were measured using an RS-230 Super Spec (GEORADIS s.r.o., Czech Republic) portable γ-ray spectrometer. The tool is equipped with 2×2″ (103 cm^3) BGO ($Bi_4Ge_3O_{12}$) scintillation detector. The counting time 240-s provides accuracy for K, U and Th concentrations assessment ±10 % in low to moderate radioactive sediments (cf. Svendsen & Hartley 2001). The combined error due to ambient conditions and the instrument for repeated measurements was estimated to be less than ±7.5 % for each element. Values of the standard γ-ray (SGR) were calculated using the Schlumberger NGT-A formula (see Rider 1999; Šimíček et al. 2012, p. 52). 525 outcrop GRS points were measured at 17 sections throughout the Mazák and Godula Formation (Figs. 1, 2). Outcrops were logged with 15-cm sampling interval, which corresponds to commonly used resolution of wire-line logs (Doveton 1994; Rider 1999). The measurement practice followed the recommended detector — outcrop geometry as described by Svendsen & Hartley (2001) and Šimíček & Bábek (2015). The GRS measurements were accompanied by detailed facies logging. The deep-water facies classification schemes of Lowe (1982), Pickering et al. (1986), Ghibaudo (1992), Mutti (1992) and Shanmugam (2006) were used in this paper. The stratigraphic position of sections within the Mazák and Godula Formations was inferred from the geological maps (Jurková & Roth 1967; Baldík et al. 2011) and previous investigations (Eliáš 1970, 1979; Roth 1980; Menčík et al. 1983).

The modal composition of sandstones was studied in 36 thin sections using the point-counting technique of Chayes (1956) with 300 points per thin section. The Gazzi-Dickinson method was used for statistical processing of the compositional data (Ingersoll et al. 1984; Dickinson & Lawton 2001; Caracciolo et al. 2012). Individual grains and mineral constituents of rock fragments in the grain size range 0.063–2 mm were described as particular mineral types. Lithic clasts consisting of grains smaller than 0.063 mm were described as rock types (Dickinson 1985). The following categories were used: monocrystalline (Qm) and polycrystalline (Qp) quartz, potassium feldspars (Fk), plagioclases (Fp), sedimentary and meta-sedimentary lithic clasts (Ls), plutonic and meta-plutonic lithic clasts (Lm), volcanic and meta-volcanic lithic clasts (Lv) and unidentified lithic clasts (Li). Modal compositional data were plotted in Q-F-L provenance ternary diagrams of Dickinson et al. (1983). Grain size distribution was measured from the a-axis of one hundred randomly selected grains per thin section (30 samples) using JMicroVision image analysis toolbox (Roduit 2008).

The provenance of clastic material of sandstones was studied from heavy mineral assemblages (50 samples). Heavy mineral concentrates were separated from the grain-size fraction 0.25–0.5 μm by separating in 1,1,2,2-tetrabromethane ($C_2H_2Br_4$) with density 2.96 g/cm^3. Specific minerals were determined using binocular microscope NIKON C-PS and polarizing microscope CarlZeiss Jena. Percentage shares were assessed semi-quantitatively by comparison with diagrams of percentage assessment of Tucker (2003).

The chemical composition of mineral carriers of radioactive K, U and Th was investigated using the electron microprobe Cameca SX100 at the Department of Geological Science, Masaryk University of Brno. Eight thin sections of sandstones coated with graphite were analysed in WDS and EDS mode. The voltage 15 keV, current 10 nA and diameter of beam 4 μm were used for analyses of framework grains. The parameters of analyses of accessory heavy minerals were 15 keV, 20–40 nA and 1–2 μm.

Phase analyses of 20 oriented samples of mudstones using X-ray diffractometer STOE Stadi P in transmission mode, with primary Ge (111) monochromator, linear PSD detector and Co anode were carried out at the Department of Geological Science, Masaryk University of Brno.

Results

Facies and facies stacking patterns

Nine sedimentary facies were identified in the Mazák and Godula Formations based on detailed lithological descriptions (cf. Pickering et al. 1986; Tucker 2003). They comprise the grain-size range from clast-supported conglomerates to mudstones (Table 1). Bed thicknesses vary from 3 m in some conglomerate and sandstone facies to several cm in the mudstones and fine-grained sandstones. Individual beds can sometimes have gradual contacts, usually on tops, but most frequently, they are sharp-based, flat or irregular with presence of various types of sole marks and bioturbation. The beds are usually massive or normally graded in the coarse-grained facies. Various types of internal stratification and lamination (planar, cross, convolute) are common particularly in the fine-grained facies (Table 1). The F1 facies are interpreted as deposits of hyperconcentrated gravity flows (or sandy debris flows *sensu* Shanmugam 2006) based on the presence of large floating clasts, inverse grading at bed bases and gravel- to coarse-sand lithology (Słomka 1995; Mulder & Alexander 2000; Gani 2004; Shanmugam 2006). The beds of facies F2 (and partly F3a), with predominantly coarse-sand lithology, massive bedding with graded top parts, presence of traction carpets near the bed bases, dish structures and frequent amalgamation, can be interpreted as deposits of high-density turbidity currents (Ghibaudo 1992; Mutti 1992). In particular, the basal parts of F3a beds correspond to Ta(b) divisions of Bouma sequence (Bouma 1962) or the S2 and S3 divisions of Lowe sequence (Lowe 1982). Beds interpreted as low-density turbidity current deposits (F4a–b, top parts of F3a) are characterized by sheet-like bed geometry, common presence of sole marks and T(b)cd divisions of the Bouma sequence (Mutti & Ricci Lucci 1972; Mutti 1977). Suspension fallout deposition is inferred for beds with predominant mudstone lithology (F6 facies), which can alternate with thin siltstone lenses and laminae representing distal, low-density turbidites or bottom-current reworked deposits (Bouma 1962; Stow 1979; Einsele 2000).

The proportion of individual facies (in thickness %) markedly differs across the stratigraphy (Table 2). Fine-grained and heterolithic facies (F4a alternating with F5 and F6) and local, thick wedge-shaped or lenticular sandstone-to-conglomerate bodies (facies F2 and F1a–b) of the Ostravice sandstone predominate in the Mazák Formation. A similar distribution of facies can be observed in the Lower and Upper Members of the Godula Formation, where thin-bedded, silty-to-clayey heterolithic facies (F6 and F4a–b) predominate over coarse-to-medium grained sandstone beds (facies F2 and F3a–b). Noteworthy features include the

Table 1: Facies classification of sediments of the Mazák and Godula Formations and interpretation of depositional processes.

Facies	Bed thickness	Grain size	Sedimentary structures	Bed contacts, bed geometry	Rip-up intraclasts	Depositional process
F1a	0.5–3 m	clast supported conglomerates	massive; occasional inverse grading at the bed base – d	occasional basal scours; irregular upper contact, bed geometry unknown	rare mudstone intraclasts	cohesive debris flows to hyperconcentrated flows transition
F1b	0.5–1.5 m	matrix supported conglomerates to pebbly sandstones	massive to faintly normally graded, sometimes parallel stratification; occurrence of clast supported conglomerates at the bed base; amalgamation	frequent basal scours; irregular upper contact, bed geometry unknown	rare mudstone intraclasts above bed bases (max. size 25 cm)	
F2	0.3–3 m	coarse-to-medium grained sandstones	massive, normally graded, rare parallel stratification near bed top, frequent dish structures, traction carpet at the bed base	frequent basal scours; flat upper contact; bed geometry unknown	occasional mudstone intraclasts above bed bases	high-density turbiditic flows
F3a	0.3–1.5 m	medium grained sandstones	normally graded; amalgamation, frequent traction carpet near bed base; Ta division of Bouma sequence	frequent basal scours; flat upper contact; sometimes laterally wedged		
F3b	10–30 cm	medium to fine grained sandstones	normally graded; parallel stratification; cross to convolute stratification at the bed top, Tb division of Bouma sequence, amalgamation	flat lower and upper contact; sheet-like geometry		
F4a	~ cm–10 cm	fine grained sandstones to siltstones	parallel lamination; sometimes normally graded; T(b)c,d division of Bouma sequence	various types of sole marks or bioglyphs at the bed base; flat upper contact; bed wedging		low-density turbiditic flows
F4b	~ cm–30 cm	siltstones with mudstone lamines	cross lamination; convolution, bioturbation, sometimes faintly parallel lamination	non-erosive contacts; flat base; undulated top		
F5	~ cm– ~ dm	mudstones with siltstone lens	parallel laminated	non-erosive contacts; flat base		deposition from suspension clouds of low-density turbiditic flows
F6	~ cm–20 cm	mudstones	faintly parallel laminated to massive			hemipelagic suspension deposits

Table 2: Average distribution (in thickness %) of facies (see Table 1) in the lithostratigraphic units of the Mazák and Godula Formations.

Stratigraphy	Mazák Formation		Godula Formation				
	Ostravice sandstone	Variegated Godula Beds	Lower Member	Middle Member	Middle to Upper Member	Upper Member	Upper Member
Facies							Pustevny sandstone
F1a	9.4	–	–	–	–	–	2.6
F1b	32.3	–	–	–	–	–	31.3
F2	55.2	–	18.9	28.7	7.6	11.7	25.3
F3a	3.1	–	2.4	8.3	11.1	15.4	23.2
F3b	–	–	8.5	7.2	2.8	15.4	–
F4a	–	41.7	28.1	27.2	32.5	16.6	2.6
F4b	–	–	15.9	10.7	3.0	5.9	–
F5	–	35.6	9.8	4.4	6.6	7.4	–
F6	–	22.7	19.5	13.6	36.3	27.6	15.0

local coarse-grained sand bodies of the Pustevny sandstone in the Upper Member, which share similar facies characteristics and distribution with the Ostravice sandstone. Compared to the other members of the Godula Formation, the Middle Member is characterized by moderate prevalence of the thick sandstones (facies F2 and F3a) over the heterolithic facies succession (F4a–b and F6).

γ-ray data from outcrops and the effect of facies on distribution of radioactive elements

The average total γ-ray values from the whole dataset are 86.3 nGy·h^{-1} (σ 26.09) dose rate and 116.3 API (σ 36.10) standard γ-rays (SGR). The average concentrations of radioactive elements are 2.8 % of K (σ 0.9), 2.8 ppm of U (σ 1.1) and 11.7 ppm of Th (σ 3.9). The contributions of Th, K and U to the total radioactivity (SGR) are roughly balanced, as indicated by statistically significant positive correlations (linear regression coefficient $R^2=0.94$ for SGR:Th; $R^2=0.86$ for SGR:K and $R^2=0.81$ for SGR:U). The radioactive elements K, Th and U show various degrees of mutual correlation in particular lithostratigraphic units: $R^2=0.5$–0.8 for Th:U, $R^2=0.3$–0.9 for Th:K, $R^2=0.1$–0.6 for K:U (Fig. 3). It is noteworthy that the K/Th and K/U ratios (indicated by linear regression line in Fig. 3) are markedly shifted in the Upper Godula Member as compared to the other stratigraphic levels. Th/U ratios range from 2 to 7 in most of samples, and the average value is 4.2. We observed only very low variation of Th/U values between particular facies at the same stratigraphic levels.

Siliciclastic sediments of the Mazák Formation usually have low radioactivity while the Godula Formation shows moderate to low radioactivity (Table 3, Fig. 4). Conglomerate and sandstone facies (F1a–b, F2, F3a, F3b) have average SGR values of 106.5 API and average concentrations of K: 2.6 %, U: 2.6 ppm and Th: 10.6 ppm. Heterolithic facies, siltstones and mudstones (F4a–b, F5, F6) usually show higher radioactivity than stratigraphically equivalent coarse-grained facies. Their average SGR values are 130.8 API and average concentrations of K: 3.1 %, U: 3.3 ppm and Th: 13.3 ppm. This facies dependence of the total as well as spectral radioactivity values is maintained in all the stratigraphic units (Table 3). The biggest disproportion between coarse- and fine-grained facies was observed in the Mazák Formation, where sandstone facies reach only 35 % of the values obtained in mudstones. In contrast, sandstone facies in the Upper Godula Member reach up to 85 % of the values from the mudstones.

No statistical correlation was found between the concentrations of K, Th and mean grain size of sandstones ($R^2=0.11$; $R^2=0.12$, respectively).

Stratigraphic distribution of outcrop spectral γ-ray data

The facies distribution of the γ-ray data reveals that the differences between mean SGR values and K, U and Th concentrations in the coarse- and fine-grained facies are lower at some stratigraphic levels than at others (Table 3). The highest contrast between facies and, therefore, the best

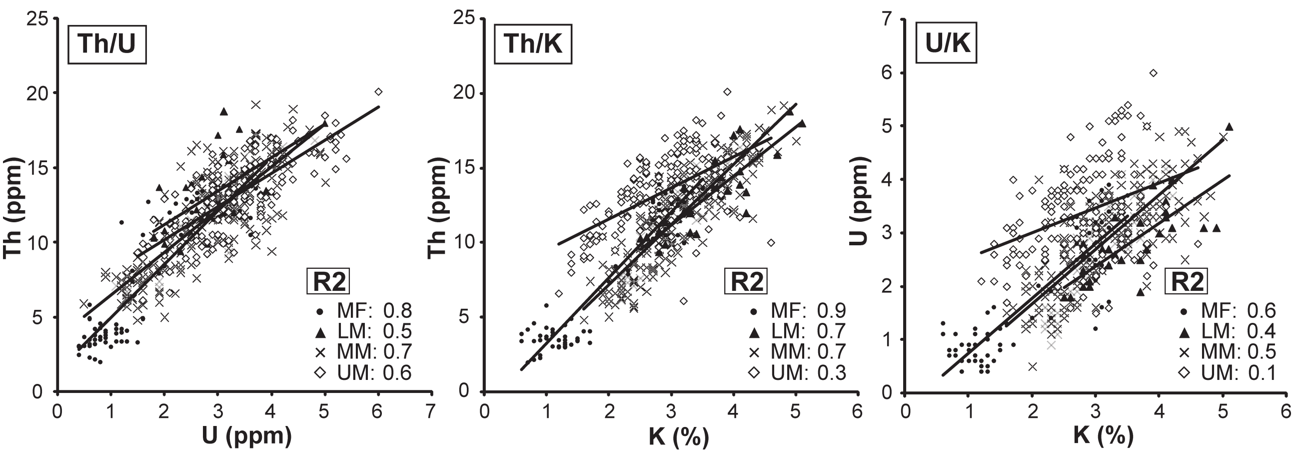


Fig. 3. Bivariate plots of K, U and Th concentrations in the Mazák Formation (MF) and the Lower Member (LM), Middle Member (MM) and Upper Member (UM) of the Godula Formation, including linear regression lines and values of R^2.

Table 3: Mean (bold), minimum and maximum values and standard deviation (italics) of outcrop GRS data and their facies and stratigraphic distribution. **SST+CG** — Sandstones and conglomerates (F1a–b, F2, F3a, F3b facies), **SLST+MDS** — Heterolithics, siltstones and mudstones (F4a–b, F5, F6 facies), **OS** — Ostravice sandstone, **VGB** — Variegated Godula Beds (oceanic red beds), **LM** — Lower Godula Member, **MM** — Middle Godula Member, **MM–UM** — Aggregated data from Middle and Upper Godula Member, **UM** — Upper Godula Member.

		Whole dataset		**LITHOSTRATIGRAPHY**										
				Mazák Formation		**Godula Formation**								
						LM		MM		MM–UM		UM		
		SST+CG	SLTS+MDS	SST+CG (OS)	SLST+MDS (VGB)	SST+CG	SLTS+MDS	SST+CG	SLTS+MDS	SST+CG	SLTS+MDS	SST+CG (PS)	SST+CG	SLTS+MDS
SGR (API)	**Mean**	**106.5**	**130.8**	**39.6**	**117.7**	**125.5**	**171.6**	**81.1**	**139.9**	**128.1**	**148.2**	**51.1**	**114.4**	**140.8**
	Min.	25.8	73.0	25.8	73.0	95.8	121.6	57.1	127.4	78.0	104.3	37.1	62.5	77.7
	Max.	176.2	194.4	55.0	140.5	163.6	194.4	138.3	166.7	176.2	186.5	137.0	169.9	191.2
	St. dev.	*36.23*	*23.26*	*6.78*	*13.47*	*20.37*	*27.12*	*19.64*	*12.03*	*21.74*	*20.86*	*26.42*	*22.96*	*20.72*
K (%)	**Mean**	**2.60**	**3.10**	**1.10**	**3.00**	**3.30**	**4.80**	**2.40**	**3.80**	**3.10**	**3.75**	**1.60**	**2.50**	**3.05**
	Min.	0.60	1.40	0.60	2.00	2.50	3.20	1.60	3.40	1.60	2.50	1.00	1.20	1.40
	Max.	4.60	5.10	1.70	3.50	4.20	5.10	3.80	4.70	4.60	5.00	4.60	3.50	4.10
	St. dev.	*0.89*	*0.66*	*0.29*	*0.28*	*0.54*	*0.75*	*0.43*	*0.40*	*0.65*	*0.59*	*0.86*	*0.59*	*0.58*
U (ppm)	**Mean**	**2.60**	**3.30**	**0.80**	**2.80**	**2.60**	**3.10**	**1.80**	**3.30**	**3.10**	**3.65**	**1.30**	**3.15**	**4.05**
	Min.	0.40	1.20	0.40	1.20	1.80	2.70	0.50	2.60	1.60	2.60	0.70	1.60	2.20
	Max.	5.30	6.00	1.60	3.90	4.00	5.00	3.90	4.00	5.00	4.80	4.60	5.30	6.00
	St. dev.	*1.05*	*0.90*	*0.29*	*0.66*	*0.62*	*0.90*	*0.68*	*0.39*	*0.63*	*0.64*	*0.61*	*0.88*	*0.84*
Th (ppm)	**Mean**	**10.60**	**13.30**	**3.55**	**12.00**	**12.00**	**16.95**	**7.70**	**13.80**	**12.90**	**14.80**	**4.65**	**12.15**	**15.25**
	Min.	2.00	7.40	2.00	7.40	9.40	12.10	4.80	11.50	7.60	10.80	2.60	6.60	8.60
	Max.	18.00	20.10	5.80	13.90	17.60	18.80	14.70	16.10	17.70	18.90	10.00	18.00	20.10
	St. dev.	*3.94*	*2.41*	*0.74*	*1.53*	*2.21*	*2.59*	*2.27*	*1.60*	*2.27*	*2.04*	*2.20*	*2.45*	*2.19*
Th/K	**Mean**	**3.78**	**4.12**	**2.89**	**4.03**	**3.70**	**3.66**	**3.35**	**3.56**	**4.02**	**3.96**	**2.85**	**5.29**	**5.22**
	Min.	1.91	2.70	1.94	3.13	2.86	3.38	2.33	2.70	2.85	3.36	1.91	3.47	2.85
	Max.	7.63	6.28	6.50	4.96	4.30	3.84	5.47	3.95	7.00	4.74	3.92	7.63	6.28
	St. dev.	*0.98*	*0.74*	*1.15*	*0.40*	*0.38*	*0.18*	*0.62*	*0.33*	*0.73*	*0.34*	*0.43*	*0.94*	*0.77*
Th/U	**Mean**	**4.20**	**4.20**	**4.40**	**4.30**	**4.80**	**5.13**	**4.54**	**4.54**	**4.00**	**4.23**	**3.83**	**3.88**	**3.73**
	Min.	2.00	2.80	2.20	2.90	3.40	3.60	2.50	2.90	2.40	3.03	2.00	3.04	2.84
	Max.	11.80	9.40	9.70	9.40	7.20	6.06	11.80	5.35	6.60	6.44	6.50	6.45	6.00
	St. dev.	*1.20*	*1.00*	*1.70*	*1.20*	*0.90*	*1.02*	*1.37*	*0.77*	*0.80*	*0.65*	*1.11*	*0.71*	*0.68*

sensitivity of γ-rays to facies is observed in the Mazák Formation, where mudstones (K: 2.0–3.5 %, U: 1.2–3.9 ppm, Th: 7.4–13.9 ppm) are significantly more radioactive than sandstones (K: 0.6–1.7 %, U: 0.4–1.6 ppm, Th: 2.0–5.8 ppm). We can see an equally good sensitivity to facies in the lower part of the Middle and partly in the Lower Godula Members (Fig. 4). On the other hand, there is a large overlap between GRS value ranges from sandstones + conglomerates and mudstones + heterolithics in the upper part of the Middle Member and the Upper Godula Member (Table 3), which indicates lower sensitivity of γ-ray spectra to facies and lower sediment maturity, as compared to the underlying strata.

Distribution of GRS data in the proximal-to-distal facies spectrum for different stratigraphic levels is shown in Fig. 5. The principal stratigraphic trends in GRS can be documented in the thick-bedded, coarse-grained sandstones of facies F2. In the Mazák Formation, the mean concentrations of U (1.5 ppm) and Th (3.3 ppm) in the facies F2 are low. In the same facies, the mean concentrations are much higher (U: 2.5 ppm, Th: 10.5 ppm) in the Lower Member and even higher (U: 4.5 ppm, Th: 12.5 ppm) in the Upper Godula Member (Fig. 5). The other facies also reveal this increase of U and Th concentrations in the stratigraphically younger strata, which probably indicates change in the detrital composition of sediments. On the other hand, a great variability of K distribution causes the poor facies-sensitivity of the Th/K ratio, which is considered a good proxy of lithology in many carbonate and siliciclastic systems (Fertl 1979; Kumpan et al. 2014; Šimíček & Bábek 2015). However, high Th/K ratios are effective in distinguishing between the Upper Godula Member and the underlying stratigraphic levels (Figs. 3, 5).

Stratigraphic distribution of GRS data in sandstones shows two prominent levels, which are marked by an increase in U and Th concentrations and Th/K ratios (Fig. 5). The former one is situated at the base of the Lower Godula Member while the latter is situated at the base of the Upper Godula Member. There is another significant increase of SGR and concentrations of U and Th within the Middle Member. The only exception is the Pustevny sandstone, where U and Th concentrations are markedly lower and K markedly higher than the mean values in the Upper Member (Table 3). These changes in γ-ray spectra are linked with variations in the modal composition of sandstones and they probably indicate significant provenance changes.

Modal composition of siliciclastics and stratigraphic trends

The majority of samples obtained from the Mazák and Godula Formations plot within the subarcose and feldspar-rich lithic arenite fields in Folk's classification diagrams

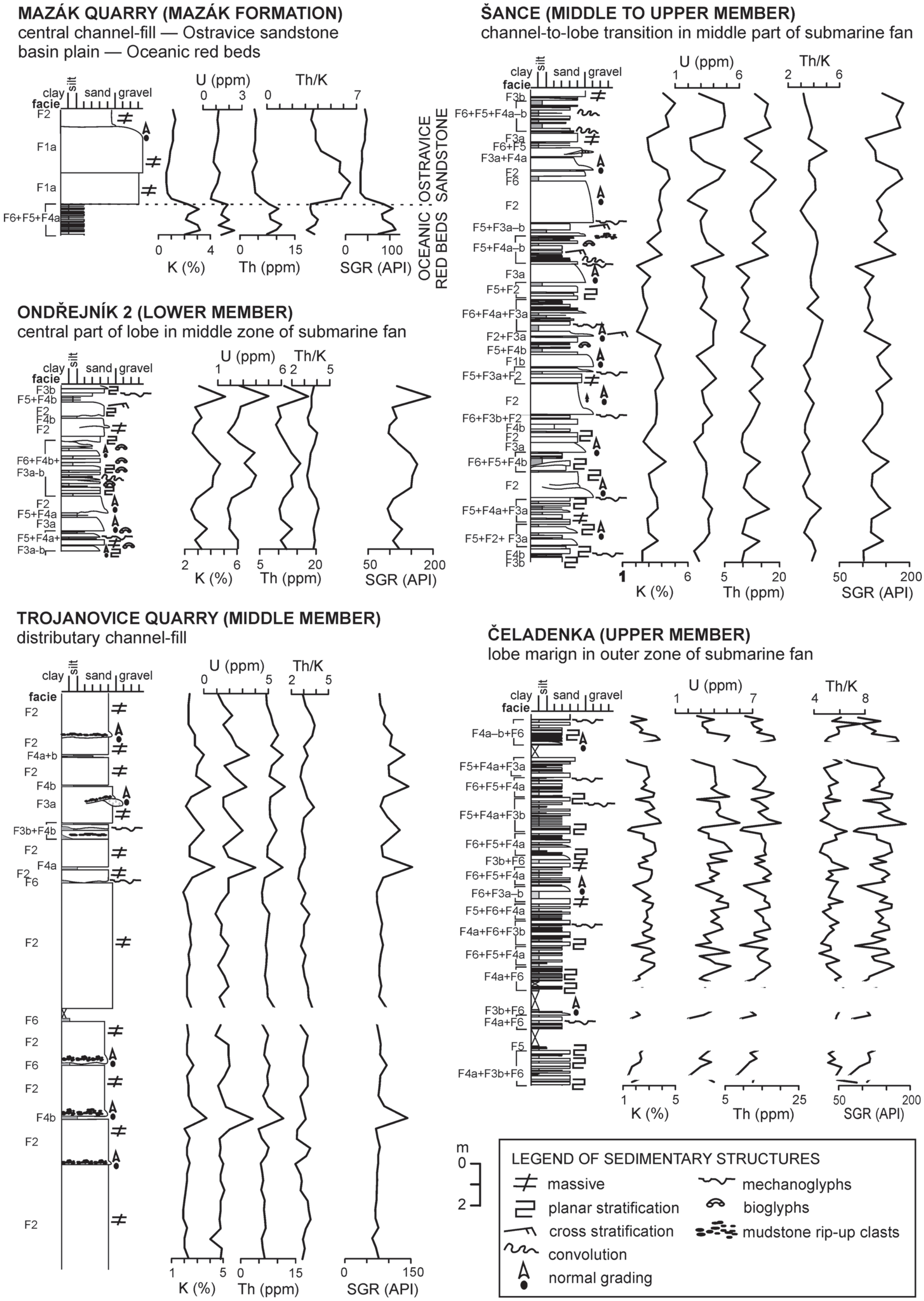


Fig. 4. Combined lithology (incl. facies and sedimentary structures) and GRS logs of selected sections, which represent the Mazák Formation and particular stratigraphic levels of the Godula Formation.

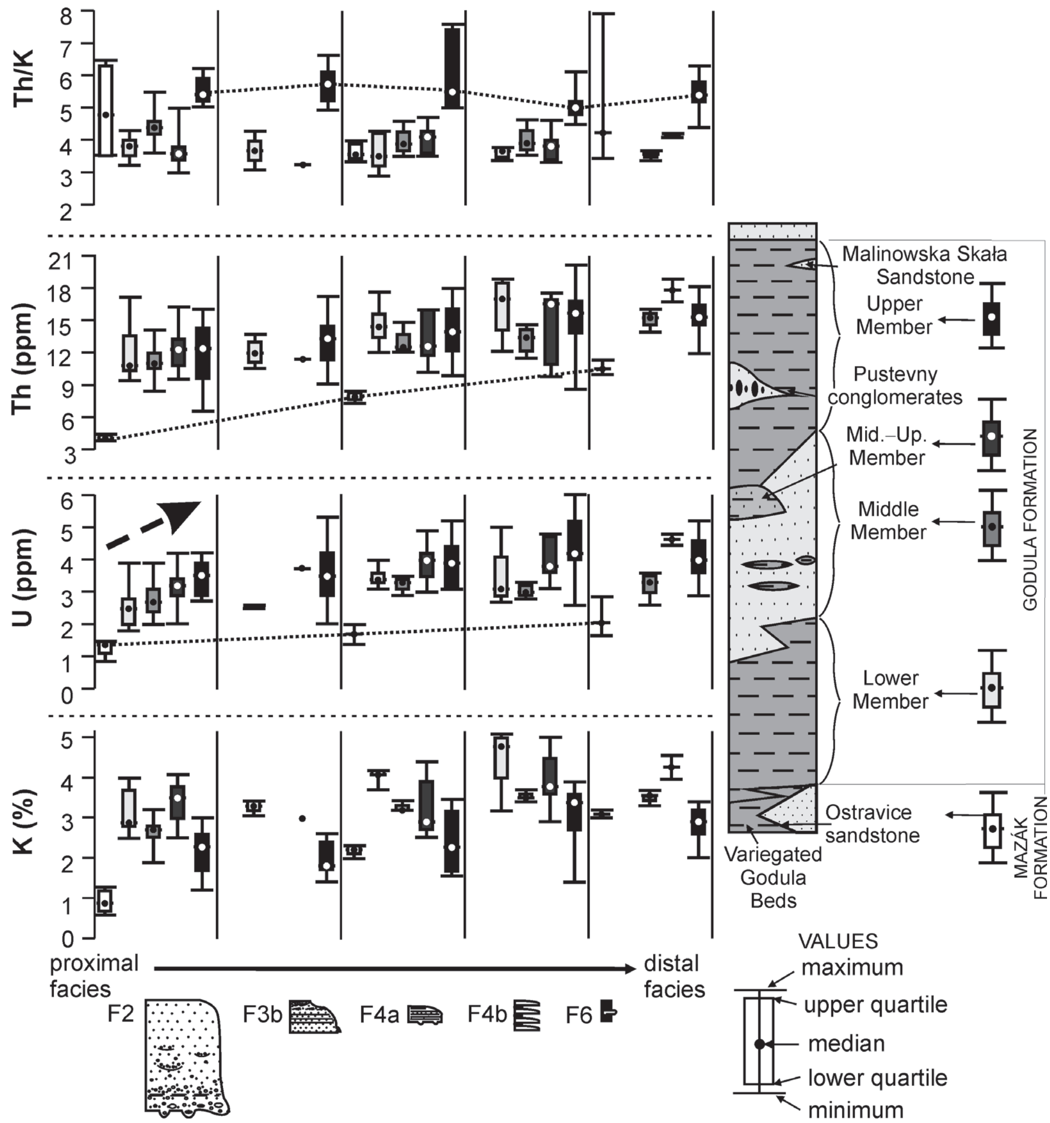


Fig. 5. Box-and-whisker plots of K, U and Th concentrations in the most abundant facies of the Mazák Formation and the Lower, Middle, Middle–Upper and Upper Members of the Godula Formation. Notice the progressively increasing U concentrations in facies F2 up-section (arrow), the markedly low U and Th concentrations in the Mazák Formation and high Th/K values in the Upper Member.

(Folk 1974). Samples of quartz arenite and arcose composition are less common. The average contents of the principal sandstone components are quartz 71.5 %, feldspars 23.5 % and lithic clasts 5 %.

In the Gazzi-Dickinson provenance ternary diagrams (Dickinson et al. 1983), one can see a clear shift from quartz-rich cratonic sources in the Mazák Formation to predominantly transitional continental crust sources in the Lower Member of the Godula Formation, followed by another shift towards recycled orogenic sources in the Middle and Upper Members (Fig. 6).

Quartz grains (total quartz, Qt) comprise both monocrystalline (Qm; mean 54.6 %) and polycrystalline (Qp; mean 16.9 %) grains. Qm grains with non-undulose extinction predominate over those with undulose extinction. About 60 % of Qp grains consist of more than three crystal individuals. Generally, the contents of Qt decrease throughout the Mazák and Godula Formations. The highest percentages of Qt (86.5–88.4 %) were found in the Ostravice sandstone of the Mazák Formation (see Table 4), in which Qp (24.7–33.8 %) prevail over Qm grains producing the lowest Qm/Qp ratios (1.2–1.9) from the whole study area. In the Godula Formation, the contents of Qt vary between 60.3 % (upper part of the Middle Member) and 81.6 % (Pustevny sandstone, Upper Member), Qm grains significantly prevail over Qp grains and Qm/Qp ratios generally increase from the Lower Member (mean 2.9) toward the Upper Member (mean 4.8).

The feldspar grains (F) comprise potassium feldspars (Fk; mean 12.5 %) and plagioclases (Fp; mean 10.7 %). K-feldspars are mostly represented by un-zoned to slightly zoned orthoclase grains without alteration or with weak signs of sericitization. Microcline with typical cross twinning is rare. Plagioclases (mostly albite) have typical lamellar twinning while most of the grains are altered by albitization. Perthitic

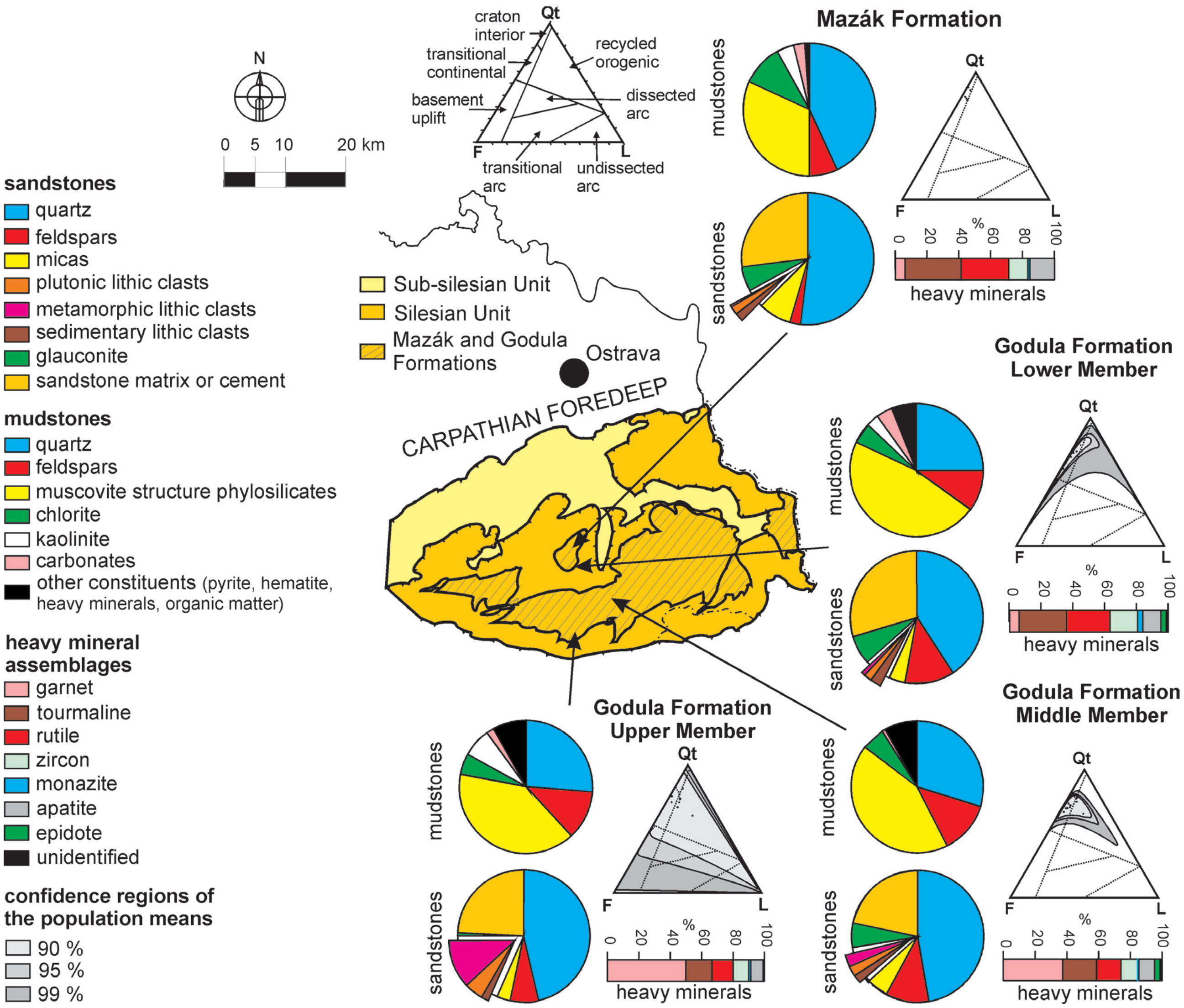


Fig. 6. Combined diagrams summarizing main provenance changes across the Mazák and Godula Formations (heavy mineral assemblages in bar plots, sandstone and mudstone components in pie plots and QFL data in Dickinson's et al. (1983) provenance ternary plots with logistic normal confidence regions of the population means and 90 %, 95 % and 99 % confidence limits, cf. Weltje 2002).

feldspars are also common. The lowest concentrations of feldspar grains (F) were found in the Mazák Formation in which Fp grains prevail over K-feldspars. Total feldspar concentrations are highest in the Lower Member and then both the Fp and Fk contents decrease towards stratigraphically younger members of the Godula Formation. Except for several samples from the Lower Godula Member, Fk prevails over Fp in most of the Godula Formation (see Table 4).

Lithic clasts (L) consist predominantly of plutonic and meta-plutonic (Lm) and sedimentary and meta-sedimentary (Ls) clasts. Volcanic clasts (Lv), represented by very rare andesite and rhyolite fragments, were only identified in several samples from the Mazák Formation (max. 0.7 %). The Lm group (0.3–13.7 %; mean 4.7 %) consists of fine-grained granites, gneisses and granulites. The Ls group (0.3–10.0 %; mean 1.1 %) comprises clasts of limestone, silicite, mudstone, siltstone, greenschist, micaschist and phyllite. The contents of Lm and Ls are relatively low and balanced in the Mazák Formation, the Lower Godula Member and lower part of the Middle Godula Member. From the upper part of the Middle Godula Member upwards, we can observe an increasing abundance of L clasts, in particular the Lm group (see Table 4). Plutonic rocks (mainly granitoids) predominate over high-grade metamorphics in the Mazák Formation and the Lower Godula Member. From the Middle Godula Member upwards, high-grade metamorphic rocks (mainly gneisses and granulites) start to predominate. The composition of Ls group also changes across the stratigraphic column, with limestone and chert clasts predominating in the Mazák Formation and pelites and metapelites being more abundant from the Lower Godula Member upwards.

In addition to Q, F and L clasts, mineral components such as micas, glauconite, heavy minerals and components contained in sandstone matrix can be important sources of K, U and Th radioactivity. Micas (mean 6.8 %) include muscovite and biotite, the latter usually being chloritized. Arenites of

Table 4: Recalculated sandstone compositional data. **OS** — Ostravice sandstone, **PS** — Pustevny sandstone, **VGB** — Variegated Godula Beds, **Qm** — Monocrystalline quartz, **Qp** — Polycrys talline quartz, **Fp** — Plagioclase feldspars, **Fk** — Potassium feldspars, **Lv** — Volcanic lithic clasts, **Lm** — Magmatic lithic clasts and their metamorphic equivalents, **Ls** — Sedimentary lithic clasts and their metamorphic equivalents, **Li** — Undetermined lithic clasts, **Qt** — Total quartz, **F** — Total feldspars, **L** — Total lithic clasts.

Lithostratigraphy		Locality	Sandstone compositional group										
			Qm (%)	Qp (%)	Fp (%)	Fk (%)	Lv (%)	Lm (%)	Ls (%)	Li (%)	Qt (%)	F (%)	L (%)
Godula Formation	PS	Pustevny 3	54.9	15.5	9.3	16.1	0	1.0	1.0	2.1	70.5	25.4	4.1
		Horní Rozpité	53.3	25.6	7.5	9.5	0.5	1.0	0.5	2.0	78.9	17.1	4.0
	Upper Mb	Čeladenka	56.3	25.3	6.8	6.3	0	2.6	1.1	1.6	81.6	13.2	5.3
			41.8	33.8	12.0	11.1	0	0.4	0	0.9	75.6	23.1	1.3
		Magurka 1	54.3	6.0	6.0	10.6	0.7	19.2	1.3	2.0	60.3	16.6	23.2
			47.4	19.6	9.3	13.9	0	8.2	1.0	0.5	67.0	23.2	9.8
	Middle–Upper Mb	Magurka 2	53.4	14.6	11.7	20.4	0	0	0	0	68.0	32.0	0
			63.0	8.2	14.7	9.2	0	1.6	2.2	1.1	71.2	23.9	4.9
		Pustevny 2	54.8	17.3	8.7	16.8	0.5	1.0	1.0	0	72.1	25.5	2.4
			55.4	15.5	9.3	16.6	0	1.0	1.0	1.0	71.0	25.9	3.1
		Malinová	52.6	19.9	6.4	9.0	0.6	10.3	1.3	0	72.4	15.4	12.2
			50.0	16.3	13.3	12.0	0	7.2	1.2	0	66.3	25.3	8.4
			62.9	10.3	11.9	11.9	0	1.0	1.0	1.0	73.2	23.7	3.1
		Šance	49.4	14.0	13.4	18.3	0	3.0	0.6	1.2	63.4	31.7	4.9
	Middle Member	Zubří 2	50.3	19.6	14.0	13.4	0.6	1.1	0.6	0.6	69.8	27.4	2.8
		Zubří 1	42.2	23.5	8.8	8.8	0	13.7	2.0	1.0	65.7	17.6	16.7
			56.5	15.1	10.2	17.2	0	0	0	1.1	71.5	27.4	1.1
		Mořkov	56.9	10.1	14.7	15.6	0	0	0.9	1.8	67.0	30.3	2.8
			53.2	21.6	5.8	11.7	0	1.2	3.5	2.9	74.9	17.5	7.6
		Trojanovice quarry	51.7	19.1	13.5	10.7	0.6	2.2	0.6	1.7	70.8	24.2	5.1
			55.7	16.5	10.3	12.9	0	1.0	1.5	2.1	72.2	23.2	4.6
			47.6	22.5	11.5	14.7	0.5	0.5	1.6	1.0	70.2	26.2	3.7
			56.6	21.5	7.8	8.3	0	2.0	0.5	3.4	78.0	16.1	5.9
			60.8	12.7	8.8	14.9	0	1.1	0.6	1.1	73.5	23.8	2.8
	Lower Mb	Ondřejník 1	60.2	12.4	11.2	13.7	0.6	0	0.6	1.2	72.7	24.8	2.5
			49.5	10.9	17.4	20.7	0	0	1.6	0	60.3	38.0	1.6
			51.1	18.6	11.7	16.5	0	1.1	0.5	0.5	69.7	28.2	2.1
			63.6	10.3	11.4	10.9	0	2.2	0.5	1.1	73.9	22.3	3.8
Mazák Formation	VGB	Ondřejník 2	54.9	14.9	14.4	14.9	0	0.5	0.5	0	69.7	29.2	1.0
		Mazák quarry	36.5	29.5	26.9	5.1	0	0.6	0.6	0.6	66.0	32.1	1.9
	OS	Čeladná	61.9	24.7	3.1	7.2	0	1.0	0.5	1.5	86.6	10.3	3.1
			57.4	31.0	3.7	6.9	0	0	0.5	0.5	88.4	10.6	0.9

the Lower and Middle Godula Members have the highest concentrations of micas. Mica grains sometimes contain inclusions of heavy minerals such as rutile, zircon or pyrite. Glauconite concentrations are highly variable in the studied sandstones (0–10 %; mean 3.2 %). They can rapidly fluctuate layer by layer at one locality. Heavy minerals occur in accessory amounts (0.2–1.9 %; mean 0.8 %) with tourmaline, rutile, zircon, garnets and apatite being the most abundant. Monazite and epidote are also present but their abundance in the heavy mineral assemblages is low (less than 2 %). Xenotime, barite, titanite and pyroxene are even less abundant (less than 1 %). SEM observation of thin sections indicate that some of these infrequent heavy minerals (monazite, xenotime) are relative abundant in the silt and clay size fractions of the sandstones. The composition of heavy mineral assemblages changes throughout the Mazák and Godula Formations. Most significant is the increase of garnet abundance at the expense of tourmaline, rutile, zircon and apatite in the Middle and Upper Godula Members. The sandstone matrix is primarily composed of a mixture of quartz, plagioclase, chlorite, sericite, illite and heavy minerals. However, the primary matrix is partly or fully replaced by calcite or quartz cement in most of the samples.

Chemical composition of the potential carriers of K, U and Th (K-feldspars, muscovite, biotite, glauconite, zircon, apatite, monazite) were studied in selected sandstone samples using the scanning electron microscope in the WDX mode. The results are shown in Table 5.

The mineral composition of mudstones was studied using X-ray powder diffraction. Phyllosilicates (micas, clay minerals with mica-structure, chlorite and kaolinite) and other detrital minerals in the clay fraction (quartz, feldspars and calcite) were identified in all samples. Other constituents such as pyrite, hematite and heavy minerals are present but rare. The clay minerals, quartz, feldspars and chlorite are the most important components of the mudstones. Illite and clay minerals with mica structure are most abundant in the Lower (mean 47 %) and Middle Godula Members and least abundant in the Mazák Formation (mean 32 %). In contrast, quartz grains are most abundant in the Mazák Formation (mean 43 %) and least abundant in the Godula Formation (mean values, Lower Member: 25 %; Middle Member: 30 % and Upper Member: 27 %). Contents of feldspars increase from the Mazák Formation (mean 7 %) towards the Godula Formation (mean values, Lower Member: 10 %; Middle and Upper Members: 12 %). The highest contents of chlorite were found in the Mazák Formation (mean 10 %) while they are less abundant in the Godula Formation (mean 6 %).

Discussion

Interpretation of depositional environment

A basic interpretation of the depositional setting was inferred from association of facies at five larger sections (10–40 m) (see Fig. 4). The record of sedimentary structures, facies (Table 1) and facies associations suggest that most of the deposition took place in a sand-rich submarine fan system in a deep-marine setting (cf. Mutti & Ricci Lucci 1972; Nor-

Table 5: Concentrations of K_2O, UO_2 and ThO_2 (SEM WDX) in the main mineral carriers of K, U and Th in sandstones of the Godula Formation. The mean abundance of minerals was calculated from point counting of the analysed thin sections. Most important sources of K, U and Th are indicated in bold letters.

Apatite	Zircon	Monazite	**Heavy minerals**	Glauconite	Albite	Biotite	Muscovite and sericite	K-feldspar	**Mineral/ group of minerals**	
13	**24**	**19**		**17**	**22**	**15**	**11**	**20**	**Number of sampled points**	
frequent	**frequent**	**frequent**	**0.8 %**	**3.2 %**	**10.7 %**	**4.4 %**	**2.4 %**	**12.5 %**	**Mean abundance**	
0.01	0	0		**8.0**	**0.2**	**3.3**	**10.4**	**15.3**	*mean*	**K_2O (wt. %)**
0	0	0		6.9	0	0.1	8.5	13.4	*min.*	
0.02	0	0		9.5	0.5	8.6	10.7	16.3	*max.*	
0.03	0.06	**0.35**		0	0	0	0	0	*mean*	**UO_2 (wt. %)**
0	0.01	0.13		0	0	0	0	0	*min.*	
0.16	1.79	1.03		0	0	0	0	0	*max.*	
0.01	0.01	**4.98**		0	0	0	0	0	*mean*	**ThO_2 (wt. %)**
0	0	1.04		0	0	0	0	0	*min.*	
0.19	0.13	19.59		0	0	0	0	0	*max.*	

mark 1978; Shanmugam & Moiola 1988; Reading & Richards 1994; Stow et al. 1996; Einsele 2000; Shanmugam 2006).

Specific elements of the submarine fan system can be seen in particular sections. Basal erosive contact of a submarine channel with underlying pelagic sediments is inferred from facies stacking patterns in the Mazák Quarry section (Fig. 4). Channel deposition of the Ostravice sandstone (cf. Eliáš 1970) is inferred from poor size sorting, predominance of coarse-grained facies (F1a and F2), frequent amalgamation of beds and sharp bed contacts (cf. Mutti & Ricci Lucci 1972; Shanmugam & Moiola 1988; Reading & Richards 1994; Słomka 1995). Although lenticular or wedge-shaped beds, which are characteristic for channel sediments, were not detected due to limited outcrop exposure, they were previously described from the Ostravice sandstones by Eliáš (1970). Basin plain sedimentation of the Mazák Formation is documented by thinly bedded red mudstones alternating with quartz-rich fine-grained turbidites (Menčík et al. 1983; Lemańska 2005). Sediments of distributory channels (Trojanovice Quarry section, Fig. 4) are interpreted on the basis of marked prevalence of coarse-grained, massive sandstone facies (F2, F3a–b) over thin bedded fine-grained facies (F4a–b, F6); poor sorting; amalgamation of beds, sharp bed contacts and abundance of mudstone rip-up clasts (Mutti & Ricci Lucci 1972; Shanmugam & Moiola 1988; Reading & Richards 1994; Bouma 2000). This section corresponds to Type II of the channel sediments classification of Słomka (1995). Sediments of the Ondřejník 2 and Šance sections are interpreted as sandstone lobe deposits, as indicated by irregular vertical CU-trends (compensation cycles), sheet-like bed geometry and predominance of coarser-grained sandstone facies. The occurrence of thick-bedded and amalgamated beds of coarse-to-medium grained sandstones with Ta-b divisions of Bouma sequence suggest deposition in proximal part of sandstone lobe near the mouth of distributory channel for the Šance section. Generally, thinner beds, presence of finer-grained facies with Tb-Tc Bouma division and abundant sole marks and trace fossils in bed bases at Ondřejník 2 section indicate deposition in the non-channelized part of a sandstone lobe (Mutti & Ricci Lucci 1972; Shanmugam & Moiola 1988; Reading & Richards 1994; Słomka 1995; Bouma 2000). The Čeladenka section, characterized by thin- to very thin bedding, equal proportion of sandstones and mudstones and predominance of facies with Tb-c-d-e Bouma divisions represents sediments of outer sandstone lobe (Mutti & Ricci Lucci 1972; Shanmugam & Moiola 1988; Reading & Richards 1994; Słomka 1995; Bouma 2000; Mattern 2002).

Sources of K, U and Th signal

The possible sources of γ-ray signal in the studied sediments were identified from the results of modal and chemical analyses of sandstones and mudstones. The contribution of K, U and Th to the overall natural radioactivity (SGR) is relatively well balanced, which is typical for immature siliciclastic sediments (IAEA 1990). Potassium is mainly contained in the minerals of sand grain-size such as K-feldspars, muscovite, biotite (incl. partly chloritized grains), which are abundant in the studied samples, and glauconite, which is only locally abundant (Table 5). Low K_2O concentrations were also detected in albite, which is an abundant framework grain in sandstones (Table 5), as well as in carbonate cement. The most important sources of U and Th are heavy minerals (cf. Howell & Aitken 1996) in particular monazite and zircon. Relatively low concentrations of UO_2 and ThO_2 were also detected in apatite grains (Table 5). A considerable part of the K, U and Th radioactive signal can also be linked with the presence of lithic clasts (IAEA 1990; Doveton 1994), in particular pelitic and meta-pelitic, plutonic and high-grade metamorphic rocks (cf. Šimíček et al. 2012). In addition, sandstone matrix can contain silt- and clay-sized radioactive grains such as sericite, K-feldspars, illite, kaolinite, chlorite and monazite, which are abundant in the sand fractions of the sandstones as well as in the mudstones.

The composition of mudstones (X-ray powder diffraction) is qualitatively similar to the composition of sandstones (Fig. 6). Illite and I/S-structure clays are most probably the main sources of potassium, followed by feldspars and sericite. The relatively high radioactivity of mudstones is traditionally attributed to the adsorption of Th and U on the surface of clay minerals and U on organic matter — an important constituent in some mudstones (IAEA 1990; Herron & Matteson 1993; Rider 1999). Part of the Th and U signal can be carried by silt- and clay-sized heavy minerals in the mudstones. We have not analysed organic matter, which can be a possible source of U, but the contents of total organic carbon are typically low in the sandstones (<1 %) as well as in the mudstones (< 0.5 % on average; maximum 1 %) of the Mazák and Godula Formations (Matýsek & Skupien 2005; Skupien & Smaržová 2009). We therefore do not regard organic matter as an important source

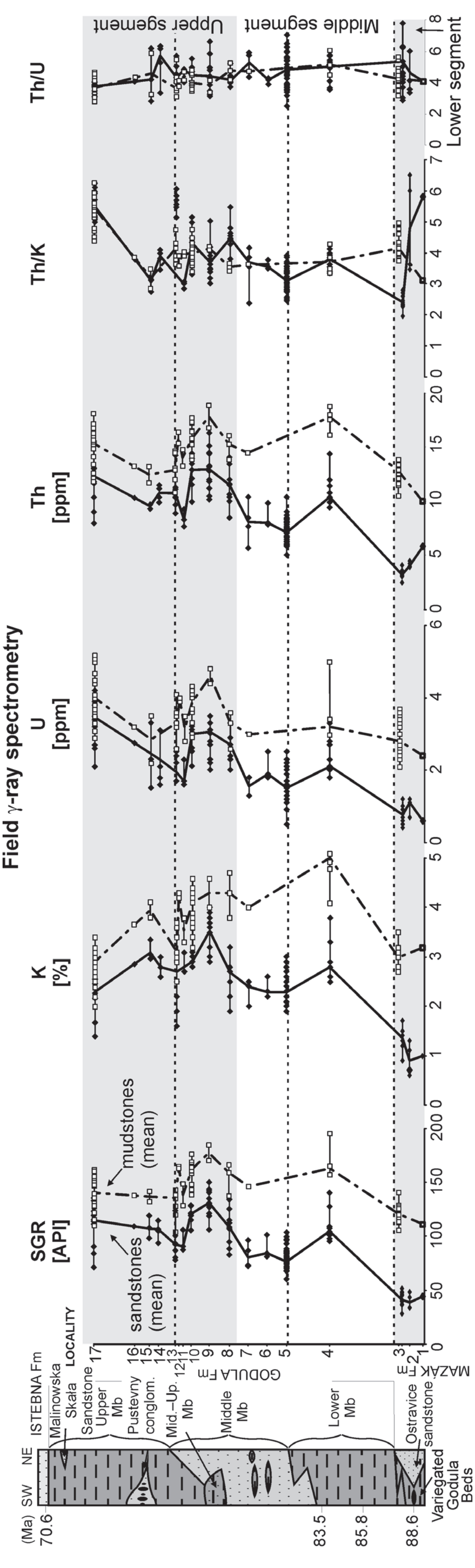


Fig. 7. Stratigraphic variation of K, U and Th concentrations, SGR and Th/K and Th/U ratios in sandstone and mudstone facies throughout the Mazák and Godula Formations. For numbers of localities see Fig. 1.

of radioactivity in the studied siliciclastics, which is in accordance with our previous results from the Carboniferous turbidites of the Moravo-Silesian Basin (Šimíček et al. 2012, p. 59).

In summary, the overall radioactivity of the Mazák and Godula Formation is not controlled by a prominent single mineral carrier but by a mix of different mineral sources of a very broad grain-size range.

Facies effects of γ-ray spectra

The total radioactivity and concentrations of K, U and Th in sandstones, in general, are lower than in mudstones due to the compositional contrasts between sand, silt and mud fractions. This difference, however, changes throughout the stratigraphy of the Mazák and Godula Formations (Figs. 7, 8). The highest contrasts were found in the Mazák Formation, in which maximum concentrations of K, U and Th in sandstones are lower than minimum values in mudstones (Figs. 4, 7). This sensitivity of the radioactive signal to facies/grain size is consistent with the primary interpretation of γ-ray logs as an indicator of "shaliness" (Doveton 1994; Rider 1999). Much of the compositional contrast is provided by the "dilution" effect of the non-radioactive quartz framework grains and calcite replacement cements in the quartzose sandstones. On the other hand, the higher radioactivity of the mudstones is driven by the contents of clay minerals and other phyllosilicates. The Lower Godula Member and lower part of the Middle Member also show relatively high facies sensitivity of the γ-ray logs. However, the facies sensitivity is rather reduced in several sections from more distal submarine fan settings (thin-bedded turbidites) (Ondřejník 2; Fig. 4). Frequent variations of thin (<20 cm), lithologically contrasted layers cause homogenization of the GRS due to mixing of γ-ray signals from multiple layers in the 2 π geometry (cf. Bristow & Williamson 1998; Svendsen & Hartley 2001).

Starting from the upper part of the Middle Godula Member, the contrast between the K, U and Th concentrations in sandstone- and mudstone+heterolithic facies tend to decrease upwards (Figs. 7, 8). The value of GRS signal as indicator of facies/grain size is therefore reduced. This trend is obviously related to the decreasing contents of quartz grains in sandstones at the expense of K-, U- and Th-bearing minerals including feldspars, micas, glauconite, lithic clasts and specific components of sandstone matrix (including heavy minerals at the Middle/Upper Godula Member boundary).

Interpretation of facies trends from the γ-ray record (Myers & Bristow 1989) is unreliable in mineralogically immature siliciclastic systems (Šimíček et al. 2012). This is caused by the highly variable modal composition of the same grain-size fractions in sandstones (quartzo-feldspathic and quartzo-feldspatholithic arenites) and by the low compositional contrast between sand-size fraction, sandstone matrix and mudstones.

The increasing rate of detrital influx (decreasing mineral maturity of sandstones) results in decreasing facies effect on the γ-ray spectra. Our results show, that detrital effect generally increased during deposition of the Mazák and Godula Formation. Comparing the GRS data from the same lithologies, we can interpret the changes in concentrations of K, U and Th in terms of provenance changes.

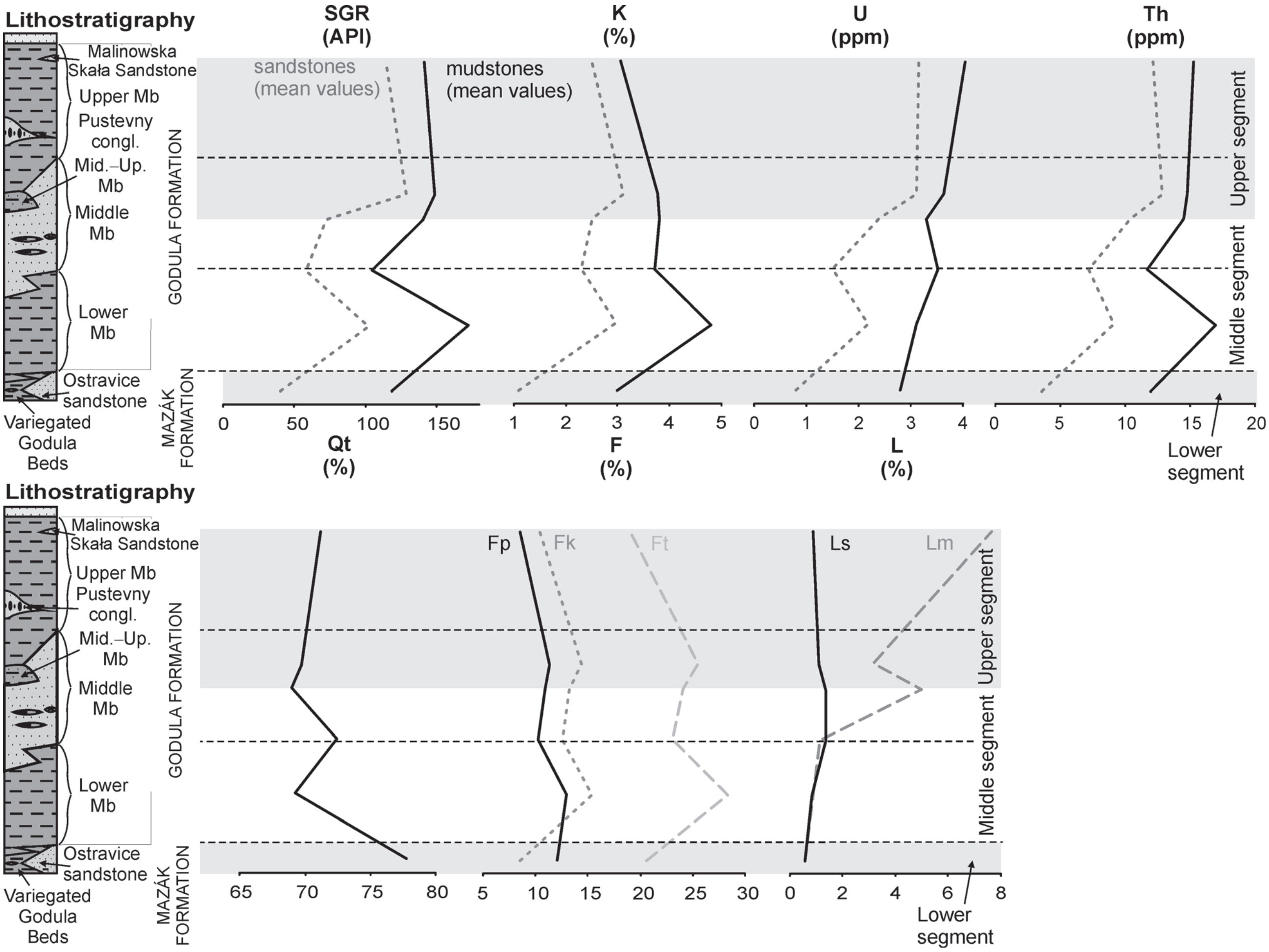


Fig. 8. Stratigraphic distribution of mean values of SGR, concentrations of K, U and Th and the main components of sandstones throughout the Mazák and Godula Formations. **Qt** — Total quartz, **F** — Feldspars, **Ft** — Total feldspars, **Fk** — Potassium feldspars, **Fp** — Plagioclases, **L** — Lithic clasts, **Lm** — Plutonic and meta-plutonic lithic clasts, **Ls** — Sedimentary and meta-sedimentary lithic clasts.

Sandstone provenance, GRS values and the tectonic evolution of the source area

The petrological study reveals that the dominant source of clastic material for the Mazák and Godula Formations was in plutonic and high-grade metamorphic rocks and, to a lesser extent, in sedimentary and low-grade metamorphic rocks. Volcanic material is rare (cf. Unrug 1968; Grzebyk & Leszczyński 2006). The plutonic and high-grade metamorphic sources are indicated by lithic clasts of granites, gneisses and granulites; slightly undulose Qm grains; Qp grain composed of more than 3 sub-grains (*sensu* Basu 1985); unzoned or slightly zoned potassium feldspars and high proportions of tourmaline, rutile, garnet and zircon in the heavy mineral assemblages (Blatt 1967; Basu 1985; Helmond 1985; Krainer & Spoil 1989; Morton 2003; Das 2008; Gallala et al. 2009). The sedimentary and meta-sedimentary sources are indicated by lithic clasts of limestones, cherts, mudstones, chlorite schists, mica-schists and by sparse rounded (re-sedimented) heavy-mineral grains (cf. Deer et al. 1992). Volcanic rocks including andesite and rhyolite clasts together with sparse Cr-spinels in the heavy mineral fraction document moderately active or distant island arc volcanism (*sensu* Čopjaková & Sulovský 2003; Grzebyk & Leszczyński 2006). This complies with the rock composition of the hypothetical intra-basin Silesian ridge source (Menčík et al. 1983; Picha et al. 2006). The Silesian ridge consisted of Proterozoic (Cadomian) granites, Variscan metamorphic rocks metamorphosed into eclogite and granulite facies and Variscan and post-Variscan low-grade metamorphic and sedimentary cover units (Michalik et al. 2004; Poprawa et al. 2004; Nejbert et al. 2005; Budzyń et al 2008).

The study of mineral and chemical composition of sandstones, combined with the GRS record reveals two major changes in modal composition during the deposition of the Mazák and Godula Formations. Consequently, we can divide the entire stratigraphic succession into three segments. The lower segment corresponds to the Mazák Formation (Turonian–Lower Coniacian) and the rutile-zircon heavy mineral sub-zone of Roth (1980). The main sources were identified in quartz-rich plutonic rocks, supplemented by limestone and chert clasts and rare pelite and meta-pelite rocks. The samples plot in the cratonic source field of the ternary diagrams of Dickinson et al. (1983). All this indicates relatively high mineral and textural maturity of the sandstones. The cratonic clastic sources document the initial phase of the Silesian ridge uplift due to the Upper Cretaceous compressional tectonic movements in the Central Carpathians (Menčík et al. 1983). The sand bodies of the coarse-grained Ostravice sandstone indicate the first pulses of synorogenic sedimentation in the Western Carpathians and a transition from passive to active continental margin regime (Picha et al. 2006). The high contents of non-radioactive quartz, low contents of less stable framework K-feldspars, plagioclases, micas and glauconite, the rutile- and tourmaline dominated heavy mineral assemblages, the low contents of lithic clasts and replacement of sandstone matrix by non-radioactive quartz or carbonate cement are responsible for very low spectral γ-ray signals in the sandstones of the Mazák Formation (Fig. 8). Consequently, in terms of the GRS signal, the Mazák Formation can be clearly distinguished from the overlying Godula Formation (Fig. 7).

The middle segment corresponds to the Lower Member and the lower part of the Middle Godula Member (Lower Coniacian–Lower Campanian). Its upper boundary coincides with the boundary between zircon and garnet heavy mineral zones of Roth (1980). As compared with the lower segment, the mineral and structural maturity of sandstones considerably decreased in the middle segment. The QFL data plot within the continental transition field in the ternary diagrams of Dickinson et al. (1983) (Fig. 6). Sandstones of the middle segment have significantly lower contents of quartz and higher contents of feldspars (especially K-feldspars), micas and glauconite. Among the lithic clasts, plutonic rocks still predominate but clasts of micaschists and gneisses are also abundant. The replacement of sandstone matrix by cement is not as common as in the lower segment. Rutile and tourmaline still predominate in the heavy mineral assemblages but their proportion slightly decreases at the expanse of zircon, monazite and locally garnets. The relative decrease of tourmaline, which indicate lithium-poor granites (*sensu* Grzebyk & Leszczyński 2006) and increase of zircons, which is also typical for granitoid rocks (Speer 1980) is probably related to a complex composition of the Cadomian crystalline basement of the Silesian ridge. This compositional complexity may parallel the Silesian ridge to the Brunovistulian terrane constituting the southeastern margin of the Bohemian Massif (Kalvoda et al. 2008). The change in modal composition of sandstones of the middle segment correlates with the onset of the Mediterranean (Subhercynian) orogenic phase in the Alpine–Carpathian system spanning the Coniacian to Maastrichtian times (Książkiewicz 1977; Menčík et al. 1983). Rapid uplift and erosion of the Silesian ridge combined with high topographic gradient, short transport and rapid deposition during the tectonic compression can be considered as causes of the lower sandstone maturity in the middle segment and the associated changes in the γ-ray spectra (Fig. 8).

The upper segment corresponds to the upper part of the Middle Godula Member and the Upper Godula Member (Upper Campanian) and the garnet heavy mineral zone of Roth (1980). The ternary diagrams of Dickinson et al. (1983) indicate increasing supply of material from recycled orogen (Fig. 6). This is consistent with the continuing decrease of mineral and textural maturity of sediments (the highest abundance of quartzo-feldspathic and quartzo-feldspatholithic arenites). This composition is characteristic for acid to intermediate magmatic sources and their high-grade metamorphic country rock (Zuffa 1985), which is the supposed composition of the Cadomian and Variscan crystalline basement of the Silesian ridge (Michalik et al. 2004; Poprawa et al. 2004; Nejbert et al. 2005; Budzyń et al 2008). An increased supply of such material was observed at the base of the upper segment. In contrast to the middle segment, high-grade metamorphic rocks such as gneisses and granulites predominate over magmatic rocks (cf. Unrug 1968) in the upper segment. This is consistent with the decreasing proportions of rutile and tourmaline at the expanse of garnets in the heavy mineral assemblages (Roth 1980). Composition of garnets changes from almandine-dominated assemblages in the upper part of the Middle Member to pyrope-dominated ones in the Upper Member (*sensu* Grzebyk & Leszczyński 2006). Clasts of pelites and meta-pelites are less abundant than in the middle segment, and limestones and silicites disappear altogether. The base of the upper segment is characterized by significant increase of the concentrations of K, U and Th and SGR values. The upper parts of the Middle Godula Member show the highest values of SGR and K concentrations. Consistently with the sandstone composition data, the γ-ray contrast between mudstones and sandstones is considerable reduced in the upper segment, indicating further reduction of mineral maturity and accelerated transport and deposition of eroded material.

Conclusions

The Mazák and Godula Formations of the Silesian Unit have moderate levels of natural radioactivity which is controlled by a roughly balanced contribution from K, U and Th. K-feldspars, micas, albite, glauconite and clay minerals (mainly illite and I/S clays) were identified as major sources of potassium; heavy minerals, in particular monazite, zircon and apatite, and possibly clay minerals as sources of U and Th (the latter due to adsorption of U and Th). Additional cryptic radioactive sources are probably associated with mudstones, lithic clasts of meta-pelites in sandstones and fine-grained matrix in sandstones. This multi-component nature of the radioactive signal provides an explanation for the lack of statistical correlation between concentrations of the main mineral constituents and concentrations of K, U and Th.

The radioactive signal is sensitive to facies/grain size. Mudstones and heterolithic deep-water facies are more radioactive than gravity-flow sandstones and conglomerates from the same stratigraphic levels. However, the facies contrast of the γ-ray signal strongly varies throughout the stratigraphic column. In the Mazák Formation at the base of the succession, the contrast is highest but it decreases in two steps towards the younger strata, first at the base of the Lower Godula Member and second at the base of the Upper Godula Member. The reduced facies contrast is caused by increasing radioactivity of sandstones due to their lower mineral maturity and lower compositional contrast between sand-, silt- and clay-size fractions at the top of the succession. With these facies effects taken into account, the spectral γ-ray signal can be used as an indicator of provenance changes.

The siliciclastics of the Mazák and Godula Formations are derived mainly from plutonic and high-grade metamorphic sources and to a lower extent from sedimentary and low-grade metamorphic rocks of the hypothetical intra-basin Silesian ridge. Based on modal analyses, mineral chemistry and the spectral γ-ray record, the whole succession can be subdivided into three stratigraphic segments with contrasting provenance. The lower segment (Mazák Formation) is predominantly supplied from a cratonic source (mainly quartz-rich plutonic rocks, limestone and chert clasts). The high contents of non-radioactive quartz and low contents of chemically unstable radioactive minerals are responsible for the very low radioactivity of the Mazák Formation and high γ-ray contrast between sandstone and mudstone facies. Sandstones of the Mazák Formation can be easily distinguished from sandstones of the Godula Formation based on their very low radioactivity. The middle segment (Lower and lower parts of the Middle Godula Member) is supplied from transitional continental sources, mainly plutonic rocks but also micaschists and gneisses. The middle segment correlates with the onset of the Mediterranean (Subhercynian) orogenic phase in the Alpine–Carpathian system. Intensive uplift and erosion in the source area caused significantly lower mineral maturity of the sandstones as compared with the lower segment. Increased influx of K-feldspars and micas and lower textural maturity of the sands are responsible for the higher concentrations of K, U and Th in the γ-ray spectra. The upper segment (upper parts of the Middle Member and the Upper Godula Member) is predominantly supplied from recycled orogen, which is consistent with the further decrease of mineral and textural maturity of the sandstones. Its sediments are derived from high-grade metamorphics such as gneisses and granulites predominating over magmatic rocks, which is consistent with the change from zircon-dominated to garnet-dominated heavy mineral zones of Roth (1980). The base of the upper segment is associated with a prominent increase in concentrations of U and Th and values of SGR.

The stratigraphic variation in K, U and Th concentrations and total radioactivity are consistent with the changes of main detrital modes in sandstones of the Mazák and Godula Formations. They jointly indicate a progressive uplift, increasing topographic gradient and enhanced erosion of crystalline basement of the Silesian ridge. Outcrop γ-ray spectrometry proves to be a useful tool in studies of siliciclastic provenance in geotectonic settings associated with plate convergence.

Acknowledgments: This research was funded by the Project POST-UP II (Nr. CZ.1.07/2.3.00/30.0041). We thank reviewers Prof. Jan Golonka (Katedra Geologii Ogólnej i Geoturystyki AGH, Kraków, Poland) and doc. Ing. Petr Skupien (Faculty of Mining and Geology VŠB-TUO, Ostrava, Czech Republic) for their helpful comments.

References

Baldík V., Novotný R., Gilíková H., Bubík M., Kryštofová E., Buriánek D., Krejčí O., Kolejka V., Nývlt D., Pecina V., Janderková J., Kociánová L., Müller P., Krumlová H., Geršlová E., Kašpárek M., Kašperáková D., Konečný F., Krejčí Z., Kunceová E., Havlín Nováková D., Otava J., Paleček M., Sedláček J. & Švábenická L. 2012: Explanation to basic Geological Map of the Czech Republic 1:25,000, 25–231 Rožnov pod Radhoštěm. *MS Archiv ČGS*, Praha, 1–169 (in Czech).

Basu A. 1985: Reading provenance from detrital quartz. In: Zuffa G.G. (Ed.): Provenance of arenites. *NATO ASI Series C 148, D. Reidel Publishing Company*, Dordrecht, 231–248.

Blatt H. 1967: Original characteristics of clastic quartz grains. *J. Sed. Petrology* 37, 401–424.

Bouma A.H. 1962: Sedimentology of some flysch deposits: a graphic approach to facies interpretation. *Published PhD Thesis, Elsevier Pub. Co.*, Amsterdam, 1–168.

Bouma A.H. 2000: Fine-grained, mud-rich turbidite systems: model and comparison with coarse-grained, sand-rich systems. In: Bouma A.H. & Stone C.G. (Eds.): Fine-grained turbidite systems. *Mem. Amer. Assoc. Petrol. Geol.,* 72; *Soc. Sed. Geol., Spec. Publ.* 68, 9–20.

Bralower T.J., Leckie R.M., Sliter W.V. & Thierstein H.R. 1995: An integrated Cretaceous microfossil biostratigraphy. In: Berggren W.A., Kent D., Aubry M.-P. & Hardenbol J. (Eds.): Geochronology, time scales and global stratigraphic correlation. *SEPM Spec. Publ.* 54, 65–79.

Bristow C.S. & Williamson B.J. 1998: Spectral gamma ray logs: core to log calibration, facies analysis and correlation problems in the Southern North Sea. In: Harvey P.K. & Lovell M.A. (Eds.): Core-Log integration. *Geol. Soc. London, Spec. Publ.* 136, 39–52.

Budzyń B., Hetherington C.J., Williams M.L., Jercinovic M.J., Dumond G. & Michalik M. 2008: Application of electron probe microanalysis Th–U-total Pb geochronology to provenance studies of sedimentary rocks: An example from the Carpathian flysch. *Chem. Geol.* 254, 148–163.

Caracciolo L., Von Eynatten H., Tolosana-Delgado R., Critelli S., Manetti P. & Marchev P. 2012: Petrological, geochemical, and statistical analysis of Eocene–Oligocene sandstones of the Western Thrace Basin, Greece and Bulgaria. *J. Sed. Res.* 82, 7, 482–498.

Catuneanu O. (Ed.) 2006: Principles of sequence stratigraphy. 1st ed. *Elsevier Pub. Co.*, New York, 1–386.

Chayes F. 1956: Petrographic modal analysis. *Wiley*, New York, 1–113.

Critelli S., Le Pera E. & Ingersoll R.V. 1997: The effects of source lithology, transport, deposition and sampling scale on the composition of southern California sand. *Sedimentology* 44, 653–671.

Čopjaková R. & Sulovský P. 2003: Detrital Cr-spinels in Culm sediments of the Drahany Upland and their tectonic significance. *Bulletin mineralogicko-petrologického oddělení Národního muzea v Praze* 11, 144–148 (in Czech).

Das P.K. 2008: Petrography of sandstones of Thekopili Formation, Jaintia Hills district, Meghalaya. *Bull. Pure Appl. Sci. F. Geol. Sci.* 27, 19–24.

Deer W.A., Howie R.A. & Zussman J. 1992: An introduction to the rock-forming minerals. 2nd ed. *Longman Sc. Technic.*, Harlow, 1–698.

Dickinson W.R. 1985: Interpreting provenance relation from detrital modes of sandstones. In: Zuffa G.G. (Ed.): Provenance of Arenites. *NATO ASI Series C* 148, *D. Reidel Publishing Company*, Dordrecht, 333–363.

Dickinson W.R. & Lawton T.F. 2001: Tectonic setting and sandstone petrofacies of the Bisbee basin (USA–Mexico). *J. S. Amer. Earth Sci.* 14, 475–504.

Dickinson W.R., Beard S., Brakenridge G.R., Erjavec J.L., Ferguson R.C., Inman K.F., Knepp R.A., Lindberg F.A. & Ryberg P.T. 1983: Provenance of North American Phanerozoic sandstones in relation to tectonic setting. *Geol. Soc. Amer. Bull.* 94, 2, 222–235.

Doveton J.H. 1994: Geologic Log Interpretation: Reading the rocks from wireline logs. *SEPM Short Course Notes* 29, 1–169.

Einsele G. 2000: Sedimentary basins: Evolution, facies and sediment budget. *Springer-Verlag*, Berlin, Heidelberg, 1–792.

Eliáš M. 1970: Lithology and sedimentology of the Silesian unit in the Moravo–Silesian Beskydy Mts. *Sbor. Geol. Věd, Geol.* 8, 7–99 (in Czech).

Eliáš M. 1979: Facies and paleogeography of the Silesian unit in the western part of the Czechoslovak Flysch Carpathians. *Věst. Ústř. Úst. Geol.* 54, 6, 327–339.

Eliáš M. 1995: Sedimentology of the Mazák beds and Ostravice sandstone (Godula formation). *Zpr. Geol. Výzk. v r. 1994*, 44–45 (in Czech).

Eliáš M. 2000: Relation between Pústevny Sandstone and Malinovská skála Sandstone (Godula Formation s.s.) in the Beskydy Mts. *Geol. Výzk. Mor. Slez. v r. 1999*, 64–66 (in Czech).

Eliáš M. 2002: The Rožnov Formation — new formal name for the Submenilitic Formation of the Godula development of the Silesian Unit (Moravskoslezské Beskydy Mts., Kelčská pahorkatina Upland, Outer West Carpathians). *Geol. Výzk. Mor. Slez. v r. 2000* 8, 27–28 (in Czech).

Eliáš M., Skupien P. & Vašíček Z. 2003: Proposal of the arrangement of the lithostratigraphy in the lower part of the Silesian unit on territory of Czech Republic (Outer Western Carpathians). *Sborník vědeckých prací Vysoké školy báňské — Technická univerzita Ostrava, Řada hornicko-geologická* 49, 8, 7–13 (in Czech).

Fertl W.H. 1979: Gamma Ray spectral data assist in complex formation evaluation. *Proc. 6th. European Symp., Pap. Q*, 1–54.

Folk R.L. 1974: Petrology of sedimentary rocks. 2nd ed. *Hemphill Press*, Austin, 1–182.

Gallala W., Gaied M.E. & Montacer M. 2009: Detrital mode, mineralogy and geochemistry of the Sidi Aich Formation (Early Cretaceous) in central and southwestern Tunisia: implications for provenance, tectonic setting and paleoenvironment. *J. Afr. Earth Sci.* 53, 159–170.

Gani M.R. 2004: From turbid to lucid: A straightforward approach to sediment gravity flows and their deposits. *Sed. Record* 2, 3, 4–8.

Geroch S. & Koszarski L. 1988: Agglutinated foraminiferal stratigraphy of the Silesian Flysch Trough. *Abh. Geol. Bundesanst.* 41, 73–79.

Ghibaudo G. 1992: Subaqueous sediment gravity flow deposits: practical criteria for their field description and classification. *Sedimentology* 39, 3, 423–454.

Golonka J. 1981: Geological map of Poland, explanations; Bielsko-Biała sheet. *Geol. Inst. Publ. House*, Warszaw, 1–63 (in Polish).

Golonka J. & Krobicki M. 2006: From Tethyan to Platform Facies. Outer Carpathians. In: Wierzbowski A., Aubrecht R., Golonka J., Gutowski J., Krobicki M., Matyja B.A., Pieńkowski G. & Uchman A. (Eds.): Jurrasic of Poland and adjacent Slovakian Carpathians. *Field trip guidebook of 7. Int. Congr. on the Jurrasic system*, Kraków, 11–15.

Golonka J., Oszczypko N. & Ślączka A. 2000: Late Carboniferous–Neogene geodynamic evolution and palaeogeography of the Circum-Carpathian region and adjacent areas. *Ann. Soc. Geol. Pol.* 70, 107–136.

Golonka J., Krobicki M., Waśkowska-Oliwa A., Vašíček Z. & Skupien P. 2008: Main paleogeographical elements of the West Outer Carpathians during Late Jurassic and Early Cretaceous times. *Kwart. AHG Geologia* 34, 3/1, 61–72 (in Polish).

Golonka J., Gahagan L., Krobicki M., Marko F., Oszczypko N. & Ślączka A. 2006: Plate tectonic evolution and paleogeography of the Circum-Carpathian region. In: Golonka J. & Picha F.J. (Eds.): The Carpathians and their foreland: Geology and hydrocarbon resources. *Mem. Amer. Assoc. Petrol. Geol.* 84, 11–46.

Grzebyk J. & Lesczyński S. 2006: New data on heavy minerals from the Upper Cretaceous–Paleogene flysch of the Beskid Slaski Mts. (Polish Carpathians). *Geol. Quart.* 50, 2, 265–280.

Haas J. & Csaba P. 2004: Mesozoic evolution of the Tisza Mega-unit. *Int. J. Earth Sci.* 93, 297–313.

Hanžl P., Finger F., Krejčí O., Schitter F., Buriánková K. & Stráník Z. 2000: Petrography, geochemistry and age of granitic pebbles from the Moravian part of the Carpathian Flysch. *Geol. Carpathica* 52, 2, 101–103.

Helmond K.P. 1985: Provenance of feldspatic sandstones — the effect of diagenesis on provenance interpretation. In: Zuffa G.G. (Ed.): Provenance of arenites. *NATO ASI Series C 148, D. Reidel Publishing Company*, Dordrecht, 139–163.

Herron M.M. & Matteson A. 1993: Elemental composition and nuclear parameters of some common sedimentary minerals. *Nucl. Geophys.* 7, 383–406.

Howell J.A. & Aitken J.F. 1996: High resolution sequence stratigraphy: innovations and applications. *Geol. Soc. London, Spec. Publ.* 104, 1–374.

Hu X., Jansa L., Wang C., Sarti M., Bąk K., Wagreich M., Michalík J. & Soták J. 2005: Upper Cretaceous oceanic red beds (CORBs) in the Tethys: occurrences, lithofacies, age and environments. *Cretaceous Research* 26, 3–20.

IAEA 1990: The use of gamma ray data to define the natural radiation environment. *IAEA-TECDOC-566*, Vienna, 1–173.

Ingersoll R.V., Bullard T.F., Ford R.L., Grimm J.P., Pickle J.D. & Sares S.W. 1984: The effect of grain size on detrital modes: a test of the Gazzi–Dickinson point counting method. *J. Sed. Petrology* 54, 1, 103–116.

Jiang S.Y., Jansa L., Skupien P., Yang J.H., Vašíček Z., Hu X. & Zhao K. 2009: Geochemistry of intercalated red and gray pelagic shales from the Mazak Formation of Cenomanian age in Gzech Republic. *Episodes* 32, 1, 3–12.

Jurková A. & Roth Z. 1967: Explanations to map-sheet 1:50,000 M-34-85 B Frýdlant nad Ostravicí. *Ústř. Úst. Geol.*, Praha, 1–110 (in Czech).

Kalvoda J., Bábek O., Fatka O., Leichmann J., Melichar R., Nehyba S. & Špaček P. 2008: Brunovistulian Terrane (Bohemian Massif, Central Europe) from Late Proterozoic to Late Palaeozoic: a review. *Int. J. Earth Sci.* 97, 3, 497–518.

Koszarski L. 1963: Types fondamentaux des depots et principales etapes de leur developpement dans de geosynclinal du flysch des Carpathes Septentrionales. *Assoc. Geol. Carpatho–Balkan., V. Congres, Comm. Scient. III/I 1*, Bucuresti, 253–267.

Kováč M., Nagymarosy A., Oszczypko N., Ślączka A., Csontos L., Marunteănu M., Matenco L. & Márton E. 1998: Palinspastic reconstruction of the Carpathian-Pannonian region during the Miocene. In: Rakús M. (Ed.): Geodynamic development of the Western Carpathians. *Slovak Geol. Surv.*, Bratislava, 189–217.

Krainer K. & Spoil C. 1989: Detrial and authigenic feldspars in Permain and Early Triassic Sandstones, Eastern Alps. *Sed. Geol.* 62, 57–77.

Krystek I. 1973: Subdivision of the Godula Group based on heavy mineral studies. *Věst. Ústř. Úst. Geol.* 48, 1, 47–50 (in Czech).

Książkiewicz M. (Ed.) 1962: Geological atlas of Poland. Stratigraphic and facial problems. Fasc. 13. Cretaceous and early Tertiary in Polish Carpathians. *Inst. Geol.*, Warszawa (in Polish).

Książkiewicz M. 1977: The tectonics of the Carpathians. In: Pozaryski W. (Ed.): Geology of Poland. IV. Tectonics. *Wyd. Geol.*, Warszawa, 476–669.

Kumpan T., Bábek O., Kalvoda J., Matys Grygar T. & Frýda J. 2014: Sea-level and environmental changes around the Devonian–Carboniferous boundary in the Namur–Dinant Basin (S Belgium, NE France): A multi-proxy stratigraphic analysis of carbonate ramp archives and its use in regional and interregional correlations. *Sed. Geol.* 311, 43–59.

Lemańska A. 2005: Comparison of deep-water agglutinated foraminifera from the hemipelagic variegated shales (Lower Turonian–Lower Santonian) and the turbiditic Godula beds (Upper Santonian–Campanian) in the Lanckorona–Wadowice area (Silesian Unit, Outer Carpathians, Poland). *Stud. Geol. Pol.* 124, 259–272.

Leszczyński S. 1989: Characteristics and origin of fluxoturbidites from the Carpathian flysch (Cretaceous–Paleogene), South Poland. *Ann. Soc. Geol. Pol.* 59, 3–4, 351–390.

Lexa J., Bezák V., Elečko M., Mello J., Polák M., Potfaj M. & Vozár J. 2000: Geological map of Western Carpathians and adjacent areas. *Št. Geol. Úst. D. Štúra*, Bratislava.

Lowe R.D. 1982: Sediment gravity flows. II. Depositional models with special reference to the deposits of high-density turbidity currents. *J. Sed. Petrology* 52, 279–297.

Małata T., Olszewska B., Poprawa P., Skulich J. & Tomaś A. 2006: The source areas of the Carpathian exotic rocks in the light of the petrological and paleontological studies. *Przegl. Geol.* 54, 10, 853–854 (in Polish).

Matějka A. & Roth Z. 1954: Report on geological maping between Ostravice and Stonávka (Moravo-Silesian Beskydy Mts.). *Zpr. Geol. Výzk. (Ústř. Úst. Geol.) v r. 1953*, 97–106 (in Czech).

Mattern F. 2002: Amalgamation surfaces, bed thicknesses, and dish structures in sand-rich submarine fans: numeric differences in channelized and unchannelized deposits and their diagnostic value. *Sed. Geol.* 150, 3–4, 203–228.

Matýsek D. & Skupien P. 2005: Phosphorite concretions in the Upper Cretaceous of the Silesian Unit. *Geol. výzk. Mor. Slez. v r. 2004*, 34–36 (in Czech).

Menčík E. & Pešl V. 1955: About lithological development of the Godula beds in the Ostravice river catchment area. *Sbor. Ústř. Úst. Geol.* 21, 1, 127–139 (in Czech).

Menčík E., Adamová M., Dvořák J., Dudek A., Jetel J., Jurková A., Hanzlíková E., Houša V., Peslová H., Rybářová L., Šmíd B., Šebesta J., Tyráček J. & Vašíček Z. 1983: Geology of the Moravskoslezské Beskydy Mts. and the Podbeskydská pahorkatina Upland. *Ústř. Úst. Geol.*, Praha, 1–304 (in Czech).

Miall A.D. 1995: Collision-related foreland basin. In: Busby C.J. & Ingersoll R.V. (Eds): Tectonics of sedimentary basis. *Blackwell Science Inc.*, Oxford, 393–424.

Michalik M., Gehrels G. & Budzyń B. 2004: Dating of gneissic clasts from Gródek on the Jezioro Rożnowskie Lake (Silesian unit). *VII. Ogólnopolska Sesja Naukowa „Datowanie minerałów i skał“*, Kraków, 101–106.

Morton A.C. 2003: Heavy minerals. In: Middleton M.J. (Ed.): Encyclopedia of sediments and sedimentary rocks. *Kluwer Acad. Publ.*, Dordrecht, 356–358.

Mulder T. & Alexander J. 2000: The physical character of subaqueous sedimentary density flows and their deposits. *Sedimentology* 48, 2, 269–299.

Mutti E. 1977: Distinctive thin-bedded turbidite facies and related depositional environments in the Eocene Hecho Group (South-central Pyrenees, Spain). *Sedimentology* 24, 1, 107–131.

Mutti E. 1992: Turbidite sandstones. *AGIP, Inst. Geol., Università di Parma*, Parma, 1–275.

Mutti E. & Ricci Lucchi F. 1972: Turbidites of the northern Apennines: Introduction to facies analysis (English translation by Nilson T.H. 1978). *Int. Geol. Rev.* 20, 125–166.

Myers K.J. & Bristow C.S. 1989: Detailed sedimentology and gamma-ray logging characteristic of a Namurian deltaic succession II: gamma-ray logging. In: Whateley M.K. & Pickering K.T. (Eds.): Deltas: sites and traps for fossil fuels. *Geol. Soc. London, Spec. Publ.* 41, 81–88.

Nejbert K., Skulich J., Poprawa P. & Małata T. 2005: Geochemistry of granitoid pebbles from the Western Outer Carpathians (Poland). *Pol. Tow. Mineral. Prace Spec.* 25, 341–344.

Normark W.R. 1978: Fan Valleys, channels, and depositional lobes on modern submarine fans; characters for recognition of sandy turbidite environments. *AAPG Bull.* 62, 912–931.

Olszewska B. 1984: A paleoecological interpretation of the Cretaceous and Paleogene foraminifers of the Polish Outer Carpathians. *Inst. Geol. Biul.* 346, 7–62 (in Polish).

Oszczypko N. 1999: From remnant oceanic basin to collision-related foreland basin — a tentative history of the Outer Western Carpathians. *Geol. Carpathica, Spec. Issue* 50, 161–163.

Oszczypko N. 2004: The structural position and tectonosedimentary evolution of the Polish Outer Carpathians. *Przegl. Geol.* 52, 8–2, 780–791.

Oszczypko N. 2006: Late Jurassic–Miocene evolution of the Outer Carpathian fold-and-thrust belt and its foredeep basin (Western Carpathians, Poland). *Geol. Quart.* 50, 1, 169–194.

Picha F.J., Stráník Z. & Krejčí O. 2006: Geology and hydrocarbon resources of the Outer Western Carpathians and their foreland, Czech Republic. In: Golonka J. & Picha F.J. (Eds.): The Carpathians and their foreland: Geology and hydrocarbon resources. *Mem. Amer. Assoc. Petrol. Geol.* 84, 49–175.

Pickering K.T., Stow D.A.V., Watson M. & Hiscott R.N. 1986: Deep-water facies, processes and models: a review and classification scheme for modern and ancient sediments. *Earth Sci. Rev.* 22, 75–174.

Plašienka D. 2000: Paleotectonic controls and tentative palinspastic restoration of the Carpathian realm during the Mesozoic. *Slovak Geol. Mag.* 6, 2–3, 200–204.

Poprawa P., Małata T., Oszczypko N., Słomka T., Golonka J. & Krobicki M. 2006: Tectonic activity of sediment source areas for the Western Outer Carpathian basins — constraints from analysis of sediment deposition rate. *Przegl. Geol.* 54, 10, 878–887 (in Polish).

Poprawa P., Małata T., Pécskay Z., Banaś M., Skulich J., Paszkowski M. & Kusiak M. 2004: Geochronology of crystalline basement of the Western Outer Carpathians sediment source areas — preliminary data. *Miner. Soc. Pol., Spec. Pap.* 24, 329–332.

Reading H.G. & Richards M. 1994: Turbidite systems in deep-water basin margins classified by grain size and feeder system. *AAPG Bull.* 78, 5, 792–822.

Rider M. 1999: The geological interpretation of well logs. 2nd ed. *Whittles Publishing Services*, Caithness, 1–288.

Roduit N. 2008: JMicroVision: Image analysis toolbox for measuring and quantifying components of high-definition images. Version 1.2.7. Software available for free download at: *http://www.jmicrovision.com*

Roth Z. 1980: The stratigraphy of the Godula Group of the Moravskoslezské Beskydy Mountains in the light of deep boreholes. *Věst. Ústř. Úst. Geol.* 55, 2, 75–83 (in Czech).

Shanmugam G. 2006: Deep-water processes and facies models: implications for sandstone petroleum reservoirs. *Handbook Petrol. Explor. Prod. 5, Elsevier*, Amsterdam, 1–496.

Shanmugam G. & Moiola R.J. 1988: Submarine fans: characteristics, models, classification, and reservoir potential. *Earth Sci. Rev.* 24, 6, 383–428.

Skupien P. & Mohamed O. 2008: Campanian to Maastrichtian palynofacies and dinoflagellate cysts of the Silesian Unit, Outer Western Carpathians, Czech Republic. *Bull. Geosci.* 83, 2, 207–224.

Skupien P. & Smaržová A. 2009: Carbon stable isotopes of the organic matter from the Upper Cretaceous of the Silesian Unit. *Geol. Výzk. Mor. Slez.*, 75–77 (in Czech).

Skupien P., Bubík M., Švábenická L., Mikuláš R., Vašíček Z. & Matýsek D. 2009: Cretaceous oceanic red beds in the Outer Western Carpathians of Czech Republic. In: Hu X., Wang Ch., Scott R.W., Wagreich M. & Jansa L. (Eds.): Cretaceous oceanic red beds: stratigraphy, composition, origins and paleoceanographic and paleoclimatic significance. *SEPM Soc. Sed., Spec. Publ.* 94, 99–109.

Slatt R.M., Jordan D.W., D'Agostino A. & Gillespie R.H. 1992: Outcrop gamma-ray logging to improve understanding of subsurface well log correlations. In: Hurst A., Griffiths C.M. & Worthington P.F. (Eds): Geological application of wireline logs. 2nd ed. *Geol. Soc. London, Spec. Publ.* 65, 3–19.

Słomka T. 1995: Deep-marine siliciclastic sedimentation of the Godula beds, Carpathians. *PAN, Prace Geol.* Kraków 139, 1–131 (in Polish).

Słomka T., Golonka J., Krobicki M., Matyszkiewicz J., Olszewska B., Oszczypko N. & Wieczorek J. 2004: Significance of exotics in reconstruction of geological structures of source areas and geotectonic evolution of the Carpathian basins. In: Krobicki M. (Ed.): Carpathian exotic pebbles— the importance in paleogeographic-tectonic reconstruction. [Egzotyki karpackie — znaczenie w rekonstrukcjach paleogeograficzno-tektonicznych.] *Ogólnopolskie seminarium*, Kraków, 5–8 (in Polish).

Speer J.A. 1980: Zircon. *Rev. Mineral. Geochem.* 5, 67–112.

Stow D.A.V. 1979: Distinguishing between fine-grained turbidites and contourites on the deep-water margin of Nova Scotia. *Sedimentology* 26, 371–387.

Stow D.A.V., Reading H.G. & Collinson J.D. 1996: Deep seas. In: Reading H.G. (Ed.): Sedimentary environments: processes, facies and stratigraphy. *Blackwell Science*, Oxford, 395–453.

Stráník Z., Menčík E., Eliáš M. & Adámek J. 1993: Flysch belt of Western Carpathians, autochthonous Mesozoic and Paleogene in Moravia and Silesia. [Flyšové pásmo Západních karpat, autochtonní mesozoikum a paleogén na Moravě a ve Slezsku.] In: Přichystal A., Obstová V. & Suk M. (Eds.): Geology of Moravia and Silesia. *Moravské zemské muzeum a Sekce Geol. Věd PřF MU*, Brno, 107–120 (in Czech).

Svendsen J.B. & Hartley N.R. 2001: Comparison between outcrop-spectral gamma ray logging and whole rock geochemistry: implications for quantitative reservoir characterization in continental sequences. *J. Mar. Petrol. Geol.* 18, 657–670.

Ślączka A. 1998: Age of the andesitic rocks in the Sub-Silesian unit (Outer Carpathians). *XVI Congr. CBGA, Abstracts*, Vienna, 559–559.

Šimíček D. & Bábek O. 2015: Spectral gamma-ray logging of the Grès d'Annot, SE France: An outcrop analogue to geophysical facies mapping and well-log correlation of sand-rich turbidite reservoirs. *Mar. Petrol. Geol.* 60, 1–17.

Šimíček D., Bábek O. & Leichmann J. 2012: Outcrop gamma-ray logging of siliciclastic turbidites: separating the detrital provenance signal from facies in the foreland-basin turbidites of the Moravo–Silesian basin, Czech Republic. *Sed. Geol.* 261–262, 50–64.

Tucker M.E. 2003: Sedimentary rocks in the field: the geological field guide series. 3rd ed. *John Wiley & Sons Ltd.*, New York, 1–234.

Unrug R. 1968: The Silesian cordillera as the source of clastic material of the flysch sandstone of the Beskid Śląski and Beskid Wyspowy ranges (Polish Western Carpathians). *Rocz. Pol. Tow. Geol.* 38, 81–164 (in Polish).

Van Wagoner J.C., Mitchum R.M., Campion K.M. & Rahmanian V.D. 1990: Siliciclastic sequence and stratigraphy in well logs, cores, and outcrops: concepts for highresolution correlation of time and facies. *AAPG Methods, Explor. Ser.* 7, 1–99.

Weltje G.J. 2002: Quantitative analysis of detrital modes: statistically rigorous konfidence regions in ternary diagrams and their use in sedimentary petrology. *Earth Sci. Rev.* 57, 211–253.

Wieser T. 1985: Some remarks on sedimentation, composition and provenance of exotics-bearing conglomerates in the Western Polish Carpathians flysch formations. *Guide to excursion 1. XIII Congr. CBGA*, Cracow, 57–68.

Zuffa G.G. (Ed.) 1985: Provenance of arenites. *NATO ASI Series C* 148, *D. Reidel Publishing Company*, Dordrecht, 139–163.

Žurková I. 1975: Heavy minerals in sandstones from the Godula formation in the western part of Jablunkov furrow. *Věst. Ústř. Úst. Geol.* 50, 5, 297–299 (in Czech).