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Heavy minerals and exotic pebbles from the Eocene flysch deposits of the Magura Nappe (Outer Western Carpathians, eastern Slovakia): their composition and implications on the provenance

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Abstract: The study aims to reconstruct the crystalline parent rock assemblages of the Eocene Strihovce Formation (Krynica Unit) and Mrázovce Member (Rača Unit) deposits, based on the heavy mineral suites, their corrosive features, geochemistry of garnet and tourmaline, zircon cathodoluminescence (CL) images, and exotic pebble composition. Both units are an integral part of the Magura Nappe belonging to the Flysch Belt (Outer Western Carpathians). Corrosion signs observable on heavy minerals point to different burial conditions and/or diverse sources. The compositions of the detrital garnets and tourmalines as well as the CL study of zircons indicate their origin in gneisses, mica schists, amphibolites, and granites in the source area. According to observed petrographic and mineralogical characteristics, palaeoflow data and palaeogeographical situation during the Eocene may show that the Tisza Mega-Unit crystalline complexes including a segment of the flysch substratum could represent the lateral (southern) input of detritus for the Krynica Unit. The Rača Unit might have been fed from the northern source formed by the unpreserved Silesian Ridge. The Marmarosh Massif (coupled with the Fore-Marmarosh Suture Zone) is promoted to be a longitudinal source.

Key words: Eocene, Outer Western Carpathians, Magura Basin, exotic pebbles, heavy minerals, geochemistry, provenance

1. Introduction

Heavy-mineral assemblages in the sediments can provide valuable information and thus serve as indicators of the palaeogeographic connections between individual palaeogeographical domains (Michalík, 1993) on provenance reconstruction of ancient and modern clastic sedimentary rocks (e.g., Morton, 1987; Morton and Hallsworth, 1999; Morton et al., 2004, 2005; Čopjaková et al., 2005; Oszczypko and Salata, 2005; Mange and Morton, 2007). Chemical composition of heavy minerals is dependent on the parent rock composition and P/T conditions under which they originated (crystallisation, postmagmatic fluid attack, metamorphism). Some of them are resistant to weathering, mechanical effects of transport, and burial diagenesis in connection with intrastratal dissolution. Therefore, heavy minerals are usually excellent provenance indicators, ideally in combination with palaeoflow analysis and investigation of exotic pebbles (pebbles or fragments of rock, preserved in sandstones and conglomerates, comprising various rocks derived from the hypothetical or destroyed source area).

Previous provenance studies on the Palaeogene deposits from the eastern part of the Magura Nappe (Flysch Belt, Outer Western Carpathians) were focused on either petrography of major framework grains (Ďurkovič, 1960, 1961, 1962) or on exotic pebble composition (Leško and Matějka, 1953; Wieser, 1967; Nemčok et al., 1968; Marschalko, 1975; Oszczypko, 1975; Marschalko et al., 1976; Mišík et al., 1991a; Oszczypko et al., 2006, 2016; Olszewska and Oszczypko, 2010). Based on heavy mineral suites, the provenance has been also investigated (Ďurkovič, 1960, 1965; Starobová, 1962), and recently more detailed results were reported from electron microprobe analyses (e.g., Salata, 2004; Oszczypko and Salata, 2004, 2005; Bónová et al., 2016, 2017).

New information on exotic pebbles, morphological features of heavy minerals, garnet and tourmaline geochemistry, and zircon cathodoluminescence analysis obtained from the Eocene clastic deposits of the Mrázovce Member belonging to the Rača Unit (RU) and of the Strihovce Formation belonging to the Krynica Unit (KU) are presented in this study. This is further supported by

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the palaeoflow analysis, and the possible source material of deposits is discussed. Our new data from the Strihovce Fm. are interpreted in the context of previous studies on palaeoflow directions (Koráb et al., 1962; Nemčok et al., 1968; Oszczypko, 1975; Kováčik et al., 2012) and exotic pebble compositions (Marschalko et al., 1976; Mišík et al., 1991a; Oszczypko et al., 2006). The aim of this paper is to review and reevaluate the published data, as well as to interpret the new results from petrographic and mineralogical study of the Eocene deposits from the Krynica and Rača units cropping out in the eastern part of the Magura Nappe.

2. Geological background and potential source areas of Eocene deposits

The Magura Nappe is the innermost tectonic unit of the Flysch Belt (Outer Western Carpathians, OWC). It is subdivided (from the south to north) into three principal tectono-lithofacies units: the Krynica, Bystrica, and Rača units (Figures 1a and 1b). These units consist of deep-sea, mostly siliciclastic deposits of Late Cretaceous to Oligocene age. In the south, the Magura Nappe is tectonically bounded by the Klippen Belt, while in the north-east it is in tectonic contact with the Dukla Unit belonging to the Fore-Magura group of nappes (e.g., Lexa et al., 2000).

The Rača Unit represents the northernmost tectono-lithofacies unit of the Magura Nappe. Based on lithofacies differences in its northern and southern parts, two zones are distinguished (Figure 1b, Kováčik et al., 2011, 2012): the Outer Rača Unit (Siary Unit in the Polish OWC) and the Inner Rača Unit (Rača Unit s.s. in the Polish OWC). The Outer Rača Unit consists of the Beloveža and Zlín formations. The Beloveža Fm. (Early Eocene – Middle Eocene) is formed by thin-bedded flysch and variegated claystones. The lower part of the Zlín Fm. (Middle Eocene – Early Oligocene) is composed of the

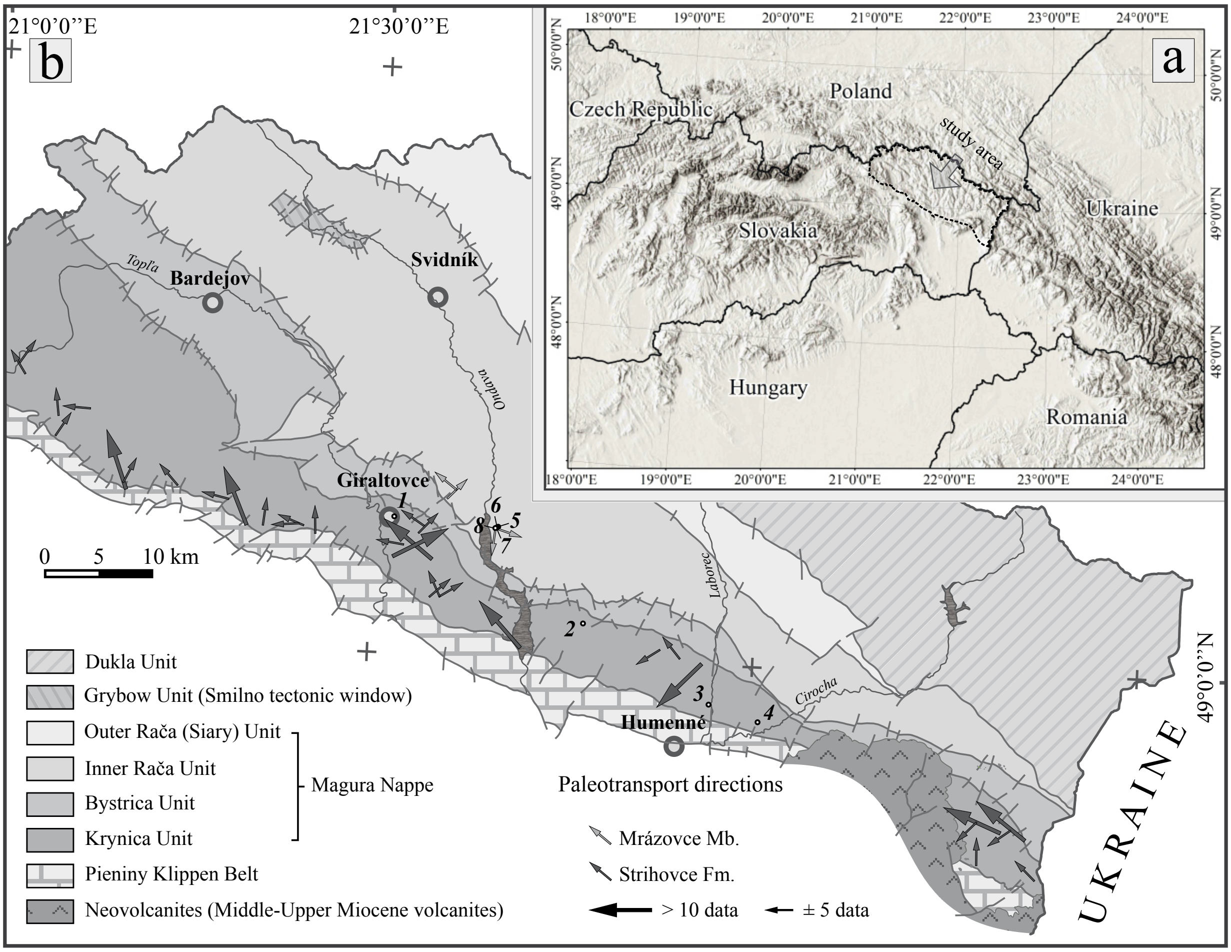

Figure 1. a) DTM map showing the position of the studied area in Central Europe; b) simplified and partly modified structural sketch map of the NE part of the Slovak Flysch Carpathians (according to Stránik, 1965; Koráb, 1983; Nemčok, 1990; Žec et al., 2006; Kováčik et al., 2011; Bónová et al., 2017; http://mapserver.geology.sk/gm50js) with sampling locations (1 – GIR-1; 2 – KOS-1; 3 – UD-1; 4 – KNC-1, KNC-4; 5 – MRA-1, 6 – MRA-2, 7 – MRA-3, 8 – MRA-4).

glauconite-sandstone facies, whereas the upper part is usually formed by the claystone facies. The total thickness of the formation is reaching 1500–2500 m. The Inner Rača Unit superficially covers a considerably larger area. It has more variegated facies content than the Outer Rača Unit. It is built of the following formations: Kurimka Fm. (sensu Samuel, 1990); Beloveža, Zlín, and Malcov fms (Kováčik et al., 2011, 2012). The underlier of the Kurimka Fm. (Late Cretaceous – Early Eocene) is not known, towards the overlier it gradually evolves into the Beloveža Fm. The formation is divided into flysch and sandstone facies. The Beloveža Fm. (Palaeocene – Middle Eocene) crops out in the frontal parts of particular slices (or in cores of anticlinal structures) of the Inner Rača Unit. The lower part of the formation is formed by the Mrázovce Member, whereas the upper part is formed by thin-bedded flysch with the intercalations of variegated claystones. The thickness of the Beloveža Fm. commonly reaches 200–250 m, with maximum up to 2000 m (Nemčok et al., 1990). The lowermost part of the Beloveža Fm. – Mrázovce Member (sensu Kováčik et al., 2012) has a character of the upward-fining and upward-thinning flysch succession (channel-levee complex) with palaeoflow direction prevailingly from NW to SE (Kováčik and Bóna, 2005). In the group of crystalline exotic pebbles within the Mrázovce Mb. were found muscovite-biotite quartzite, quartzitic paragneiss, quartzitic micaschist, granodiorite, and ultrabasic? rock. Limestones, sandstone, and chert were also described (Kováčik et al., 2012). The overlier of the Beloveža Fm. is formed by the Zlín Fm. (Middle Eocene – Early Oligocene). The formation is composed of several facies (or lower lithostratigraphic units): Makovica sandstones with local layers of conglomerate, glauconite-sandstone facies, coarse-grained sandstones and conglomerates, claystone facies, and dark-grey and olive-green calcareous claystones with quartzose-carbonate and glauconitic sandstones. The transition into the overlying Malcov Formation (Late Eocene – ?Late Oligocene) is gradual at numerous places and a common occurrence of the Malcov and Zlín lithotypes is expressed by the defining of the Zlín-Malcov facies (calcareous claystones, quartzose-carbonate, and glauconitic sandstones).

The Bystrica Unit is overthrusted on the Inner Rača Unit in the north-eastern side and in the south it is in tectonic contact with the Krynica Unit. The oldest lithostratigraphic unit is the Beloveža Fm. (Palaeocene – Middle Eocene) consisting of the sandstone facies (locally with conglomerates) and the thin-bedded flysch. The Zlín Fm. (Middle Eocene – Late Eocene) is formed prevailingly by the sandstone facies and claystone facies.

The Krynica Unit is the southernmost tectono-lithofacies unit of the Magura Nappe. It consists of the Proč, Čergov, Strihovce, and Malcov formations. The Proč Fm. is commonly regarded as a part of the Pieniny Klippen Belt (e.g., Nemčok, 1990; Lexa et al., 2000). Latter research in the this area proved the facies transition (Jasenovce Mb.) between the Proč and Strihovce fms and so both formations constitute an integral part of the Krynica Unit (Potfaj in Žec et al., 2006). The Strihovce Fm. (Early Eocene – Late Eocene) dominates in the eastern part of Flysch Belt (Žec et al., 2006; Kováčik et al., 2012) and represents several 100-m-thick bed successions of quartzose-greywacke (Strihovce) sandstones with intercalations of conglomerates. A significant facies is represented by the polymictic conglomerates with exotic pebbles (Marschalko et al., 1976; Mišík et al., 1991a): granite, orthogneiss, micaschist, metalydite, migmatite, quartz porphyry, rhyolite, and basic volcanics. Arkose, arkosic quartzite, Triassic limestones containing ostracods and foraminifers, Jurassic siliceous limestones with chert, radiolarian siliceous limestones, dark flecked marl limestones ("fleckenmergel"), Dogger-Malm biomicrites, Kimmeridgian-Tithonian shallow-water and pelagic limestones, Late Jurassic-Early Cretaceous limestones with calpionels, and Cretaceous, Palaeocene to Middle Eocene limestones and sandstones with foraminifers were also identified (Mišík et al., 1991a). Significant for the Strihovce conglomerates are red orthogneisses (Marschalko et al., 1976). In the Eocene deposits of an equivalent formation (the Piwniczna Sandstone Member of the Magura Formation and Tylicz/Krynica facies, Olszewska and Oszczypko, 2010) in Poland were found granitoids, gneisses, mica schists, phyllites, quartzites, and a small amount of basic volcanic rocks and Mesozoic carbonates (Oszczypko, 1975; Oszczypko et al., 2006, 2016). Analyses of heavy mineral suites from the Strihovce Fm. showed garnet dominance over zircon, rutile, tourmaline, and staurolite (Ďurkovič, 1960; Starobová, 1962; Bónová et al., 2010). High Cr-spinel content was also noted (Starobová, 1962; Winkler and Ślączka, 1992; Bónová et al., 2017). Maťašovský (1999) described the garnet, ilmenite, rutile, zircon, leucoxene, epidote, tourmaline, apatite, pyroxene, and gold. The sandy claystones are developed in the overlier of these polymictic conglomerates. The flysch facies is locally presented with intercalations of variegated claystones. The Malcov Fm. (Late Eocene – ?Late Oligocene) is the youngest formation of the KU in the region. For the KU, sedimentary gravity flows brought clastic material mostly from S, SE, and E to the N, NW, and W (longitudinal filling, Koráb et al., 1962). Several data point to the directions from SW to NE. It was supposed that the lateral filling longitudinally turned to the axis of the basin (l. c.).

During the Late Cretaceous to Palaeogene the Magura Basin was supplied with clastic material from source areas situated on the northern and southern margins of the basin.

The northern source area is traditionally associated with the Silesian Ridge/Cordillera (e.g., Książkiewicz, 1962; Eliaš, 1963; Krystek, 1965; Soták, 1986; 1990; 1992;

Grzebyk and Leszczyński, 2006), but other sources like the Bohemian Massif (Nemčok et al., 2000) and European Platform (Golonka et al., 2000, 2003; Golanka, 2011) were also proposed. The Silesian Ridge/Cordillera was an elevated area, consisting of the pre-Albian formations of the Magura substratum and tectonically annexed parts of the Brunovistulicum (Soták, 1990, 1992), or it was originally part of the North European Platform (Golonka et al., 2014). It is known only from exotics and olistoliths occurring within the various units of the Outer Western Carpathians (l. c.). The Silesian Ridge was uplifted in the Late Cretaceous to Palaeocene (Poprawa and Malata, 2006), to Middle Eocene (Kováč et al., 2016) or up to the Oligocene (Książkiewicz, 1962; Golonka et al., 2006). Golonka et al. (2006) and Waśkowska et al. (2009) suggested an existence additional intrabasinal ridge, the Fore-Magura Ridge, which supplied the Magura basin during the Palaeocene from the North. According to Mišík et al. (1991a) the Silesian Cordillera had no equivalent in the eastern-Slovakian zone of the Flysch Belt.

The southern source area is not still unambiguously determined. Leško (1960) and Leško and Samuel (1968) proposed the Marmarosh Cordillera (partially identified with the present development of the Marmarosh Massif), which detached the Magura and Klippen Belt spaces until the Late Lutetian in the east. On the other hand, the Marmarosh Ridge is considered an extension of the Silesian Ridge (Bąk and Wolska, 2005) and could feed the Magura Basin from the north-eastern side (e.g., Oszczypko et al., 2005, 2015). The presence of the intrabasinal Marmarosh Ridge between the Magura and Dukla basins was also suggested (Leszczyński and Malata, 2002; Ślączka et al., 2006; Gągała et al., 2012). It uplifted during the Late Eocene and drowned in the Early Oligocene due to tectonic loading (Gągała et al., 2012). Koráb and Ďurkovič (1973, 1978) demonstrated the existence of a mutual sedimentary basin for the Magura and Dukla units during the Middle Cretaceous to Early Oligocene in eastern Slovakia, i.e. these units were sedimented in a basin that was not divided by a cordillera. Ślączka and Wieser (1962) and Ślączka (1963) proposed small islands of the Marmarosh and Rachov massifs situated between the Dukla and Silesian (northern) subbasins. Nemčok et al. (1968), Nemčok (1970), and Samuel (1973) also envisaged an exotic cordillera that had been fed to the Magura Basin from the south. For the KU (Strihovce Fm.), Marschalko et al. (1976) and Mišík et al. (1991a) devised the South-Magura Cordillera (Magura Cordillera sensu Rakús et al., 1990). This cordillera was active predominantly during the Eocene and was constituted from the substratum of the Magura Basin (l. c.). Marschalko et al. (1976) suggested the consuming of the South-Magura Cordillera during the Oligocene. According to Potfaj (1998), this cordillera existed only until the Middle Eocene. Based on the study of exotic crystalline pebbles, Oszczypko et al. (2006), Salata and Oszczypko (2010), and Olszevska and Oszczypko (2010) devised the Eocene exhumation of the Magura basement in the KU. The siliciclastic material could also be supplied from a SE source area (Dacia and Tisza Mega-Units) and carbonate material from the ALCAPA Mega-Unit: Central Carpathian Block and Pieniny Klippen Belt (l. c.). This interpretation of carbonate source could be excluded because of the different biofacies of the Mesozoic sequences (Mišík et al., 1991a). Palaeogeographic reconstructions based on the heavy mineral composition of the Eocene-Oligocene deposits and Cr-spinel geochemistry supported by the palaeoflow data suggest that during the Eocene to Lower Oligocene the source area for the eastern part of the Magura Basin was located in the Fore-Marmarosh suture zone (Eastern Carpathians; Bónová et al., 2017). Late Eocene to Late? Oligocene deposits mainly in the RU could be derived from the Marmarosh Massif and also the Fore-Marmarosh Suture. For the KU, a significant contribution of detrital material from medium- to high-grade metamorphic complexes of the Villáni-Bihor and Békés-Codru zones (crystalline basement of the Tisza Mega-Unit) was proposed by Bónová et al. (2016). Part of the clastic material could be redeposited from older flysch formations (l. c.).

3. Sampling and methods

Quantitative exotic pebble analysis (130 pebbles with parameters up to 11 cm) was performed for several localities within the Mrázovce Mb. deposits. The pebble material was obtained from an exposure in the Mrázovce stream (GPS: N 49°06.446, E 21°39.385) and from debris of the conglomerate occurrences (GPS: N 49°06.727, E 21°39.611, Figures 1a and 1b). The thin sections were prepared from 25 samples and were examined under a polarising microscope. Published data were used for the Strihovce Fm. (Oszczypko, 1975; Marschalko et al., 1976; Mišík et al., 1991a). Sandstone samples were selected for optical heavy mineral analysis covering the Strihovce Fm. from the Krynica unit (KU) and the Mrázovce Mb. from the Rača unit (RU).

For the KU, heavy minerals were separated from the sandstone-conglomerate facies (Strihovce Sandstones s. s.) of the Kamenica n/Cirochou and Košarovce localities (KNC-1, KNC-4, and KOS-1 samples), from the flysch facies of the Giraltovce locality (GIR-1 sample), and from the matrix of polymictic conglomerates of the Udavské locality (UD-1 sample).

For the RU, heavy minerals were recovered from the MRA-1, MRA-2, MRA-3, and MRA-4 samples of the Mrázovce locality (Figure 1b).

The weight of the samples was about 3–5 kg. To separate the heavy minerals, the samples were crushed, sieved, and gently washed by water across a Wilfley vibrating table. In this study, the total heavy mineral concentrates were obtained from the grain-size fraction of 0.01–0.63 mm through the standard separation method using tribromomethane with a specific gravity of 2.89 g/cm^3. Approximately 350 translucent heavy minerals were counted in randomly selected traverses for each sample. Detrital minerals (garnets, tourmalines, and zircons) were embedded in epoxy resin and polished. Minerals were analysed in polished thin sections using an electron microanalyser (CAMECA SX 100, State Geological Institute of Dionýz Štúr, Bratislava, Slovak Republic) with the WDS method at accelerating voltages of 15 kV, beam current of 20 nA, and electron beam diameter of 5 µm. To measure concentrations of various elements the following natural and synthetic standards were used: orthoclase (Si *Kα*), TiO_2 (Ti *Kα*), Al_2O_3 (Al *Kα*), Cr (Cr *Kα*), fayalite (Fe *Kα*), rhodonite (Mn *Kα*), forsterite (Mg *Kα*), wollastonite (Ca *Kα*), NiO (Ni *Kα*), willemite (Zn *Kα*), and V_2O_5 (V *Kα*). The crystallochemical formula of garnet was normalised to 12 oxygens and conversion of iron valence (Fe^{3+} and Fe^{2+}) according to ideal stoichiometry. Analysed points for tourmaline were located in the centre, on the core-rim and on the rim of the grains. Tourmaline structural formula was calculated on the basis of 31 oxygens, (OH + F) = 4 *a.p.f.u.*, B = 3 *a.p.f.u.* Cathodoluminescence was used for the observation of the zircon zoning. It was carried out with the same instrument at an accelerating voltage of 8 kV and beam current of 1×10^{-3} nA. Silicates in pebble exotics were studied by electron microprobe JEOL JXA 8530FE at the Earth Sciences Institute in Banská Bystrica (Slovak Republic) under the following conditions: accelerating voltage 15 kV, probe current 20 nA, beam diameter 2–5 µm, ZAF correction, counting time 10 s on peak, 5 s on background. Used standards, X-ray lines, and D.L. (in ppm) are: Ca(Kα, 25) – diopside, K (Kα, 44) – orthoclase, F (Kα, 167) – fluorite, Na (Kα, 43) – albite, Mg (Kα, 41) – olivine, Al (Kα, 42) – albite, Si (Kα, 63) – quartz, Fe (Kα, 52) – hematite, Cr (Kα, 113) – Cr_2O_3, Mn (Kα, 59) – rhodonite, V (Kα, 117) – $ScVO_4$, Ti (Kα, 130) – rutile, Cl (Kα, 12) – tugtupite. Their structural formulas were calculated as previously described.

Selected heavy minerals were analysed via scanning electron microscopy (SEM) using a TESCAN VEGA-3 XMU (operating at 20 kV) equipped with an EDX energy dispersive spectrometer for their surface characterisation (Department of Condensed Matter Physics, Pavol Jozef Šafárik University in Košice, Slovak Republic). The mineral samples were fixed on a carbon sticker and covered by Au.

4. Results

4.1. Exotic pebble analysis

Krynica Unit. Composition of pebbles considered in the discussion was excerpted from the published data (Oszczypko, 1975; Marschalko et al., 1976; Mišík et al., 1991a).

Rača Unit. About 23% of the pebbles analysed are represented by phyllite, garnet micaschist, and gneisses (Figures 2a and 2c), 6% of them are formed by tourmaline-bearing pale granite (Figure 2b), and 3% of the exotics belong to cataclastic granite. About 38% of pebbles appertain to subarkose, quartz arenite, and quartzite, following organogenic limestone, limestone (10%), and dark siliceous rocks (19%). Some limestone pebbles show signs of a syngenetic splitting connected with the matrix penetrating them. Rounded quartz is the most abundant (it is not counted in the statistics considering its high concentration).

Petrographic characteristics of pebbles. **Phyllite** is fine-grained rock composed mainly of undulose quartz, biotite, white mica, and plagioclase feldspar, rarely graphite. Secondary minerals are represented by calcite and hematite (after opaque minerals). In some samples the biotite is baueritised or intensively chloritised. **Garnet micaschist** is formed by undulose quartz and feldspar containing the anhedral crystals of garnet. The subhedral garnet porphyroblasts show signs of local chloritisation. They are often surrounded by quartz and white mica, more sporadically by chloritised biotite. Garnet porphyroblasts represent grossular-almandine with a spessartine component, the content of which decreases slightly toward the rim ($Alm_{76-78}Grs_{12-14}Prp_{7-8}Sps_{1-5}$). Zircon, tourmaline, and opaque minerals are in accessory amounts. Subhedral zoned dravitic tourmaline [Mg/(Mg + Fe) = 0.6-0.72] is subrounded by mica and quartz. Quartz and chlorite penetrate the tourmaline grain and form its microboundinage, signalising the brittle deformation behaviour of minerals (Figure 2d). **Gneiss** shows usually a banded texture. The first type of gneisses consists of the K-feldspar and plagioclase, which form the porphyroblasts in the quartz-muscovite matrix. Zircon, staurolite, and kyanite (?) rarely occur. In the second type of gneisses, the porphyroblasts are represented by a destroyed (retrograde) garnet (Figure 2a) coupled with K-feldspar, chloritised biotite, and quartz in the quartz-muscovite matrix. The chemical composition of garnet corresponds to almandine with variable content of grossular and spessartine molecules ($Alm_{73-79}Prp_{5-8}Grs_{9-12}Sps_{3-10}$). The rock foliation is surmounted by graphite. Ore minerals and zircon rarely occur. The porphyroblasts in the third type of gneisses are composed of the sigmoidal garnets enclosed in TiO_2 polymorphs, zircon, and apatite and also of the sericitised K-feldspars. The geochemistry

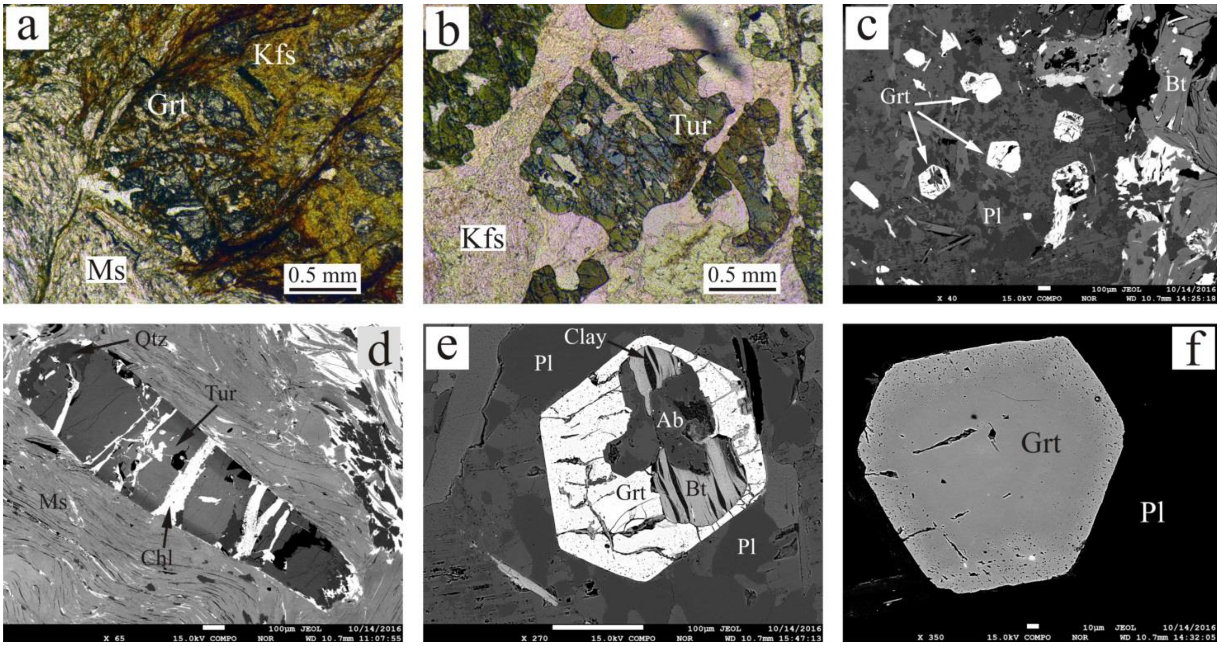

Figure 2. Microphotographs in plane polarised light (a, b) and backscattered electron images (c–f) of exotic pebbles from the Mrázovce Mb. deposits: a) retrograde garnet coupled with K-feldspar in gneiss pebble; b) pleochroic tourmaline in granite pebble; c) euhedral (prograde) garnets enclosed in plagioclase in gneiss pebble; d) tourmaline from micaschist pebble with fractures filled by quartz and chlorite; e, f) c in detail.

of garnet indicates uniform composition as in a previous type ($Alm_{78-82}Prp_{4-8}Grs_{8-11}Sps_{2-7}$). A groundmass consists mainly of muscovite with biotite. Adjacent to the garnets there is a slightly higher proportion of quartz and feldspar than in the micaceous part of the groundmass. This type of gneisses is characterised by the highest quartz content. Euhedral small garnets enclosed in plagioclase are characteristic for the fourth type of gneisses (Figures 2c, 2e, and 2f). EMP analyses revealed their zoned character. Garnets show grossular-almandine composition with an increase of the pyrope component at the expense of the grossular toward the rim, signalising the prograde metamorphism ($Alm_{63-68}Prp_{5-9}Grs_{20-27}Sps_{1-5}$). Biotite, muscovite, quartz, zircon, rutile, and ore minerals are also present. **Cataclastic granite** consists mainly of K-feldspar, plagioclase, undulose and partially recrystallised quartz, rare muscovite, and pseudomorphosis after pyrite. Some quartz crystals seem to be distinctly elongated. The fractures in feldspars are filled by quartz. **Granite** consists of quartz, orthoclase, microcline showing evident cross-hatched twinning, plagioclase with lamellar twining, and tourmaline showing very distinct pleochroism (Figure 2b). Zoned tourmaline shows schorlitic-dravitic composition (molar X_{Mg} = [Mg/(Mg + Fe)] varies from 0.45 to 0.56). The alkali feldspar is present in much higher proportions than the plagioclase. The zircon and white mica are accessory minerals. **Subarkose** is composed mainly of quartz, K-feldspar, and plagioclase. Detrital zircon, muscovite, chloritised biotite, and epidote are present in accessory amounts. The matrix contains opaque minerals, probably iron oxides. The quartz is the main component of the **quartz arenite**. The altered feldspars, platy white mica, detrital zircon, tourmaline, and hematite (after opaque minerals) are scarce. This rock is cemented by calcite cement. Another type of quartz arenite shows the corrosive structure; the original shape of quartz grains is intensively destroyed by a corrosive influence of the hematite cement. Quartz is the dominant grain type in **quartzite**. Biotite and muscovite slices, sericitised and partially deformed feldspar with kink bands, zircon, rutile, and apatite are an unsubstantial. Some quartzite pebbles are cut by calcite veins. The recrystallized quartz and bands of graphite are the main component of **graphitic quartzite**. **Limestone** pebbles are represented either by clustered ones (calcite mass with unsharp restricted clusters of calcite mud) or organogenic limestones with dispersed microfossils (foraminifers).

4.2. Heavy minerals

Heavy mineral assemblages (HMAs) of the Strihovce Fm. (KU) consist of high proportions of garnet, zircon, rutile, and apatite. Subordinate amounts were obtained for tourmaline, epidote, staurolite, and Cr-spinel. Pyroxene, amphibole, glauconite, kyanite, monazite, and titanite rarely occur. The HMA of the Mrázovce Mb. (RU) is comparable to that of the KU (Figure 3) but certain differences are a mildly higher tourmaline concentration than in the Strihovce Fm. and the occurrence of barite.

4.2.1. Corrosion features

Surface textures of detrital minerals usually range from incipient corrosion to deep etching, reflecting a progressively increasing degree of weathering. Some ultrastable to stable grains are unweathered. Surface textures of the selected minerals are documented in Figure 4.

Krynica Unit. According to classification of Andò et al. (2012), a few detrital garnets represent almost unweathered euhedral grains (Figure 4a), but nevertheless the bulk of isometric grains are slightly rounded. Some garnets show a slight to advanced degree of corrosion. The textures caused by both weathering/dissolution and abrasion are observed on the same grain (Figure 4b). The mass of grains commonly show corroded outlines and large-scale facets (Figure 4c), and less frequently etch pits (Figure 4d). Among stable minerals, tourmaline is usually angular and unweathered, sometimes subrounded with an initial to slight degree of corrosion, while corroded rutile locally occurs (Figure 4e). Zircon is mildly rounded or euhedral and usually unweathered (Figure 4f).

Rača Unit. Contrary to the Strihovce Fm. deposits, detrital garnets from the Mrázovce Mb. show deeply etched to faceted grain surfaces. Weathering intensity of garnets is diverse (Figures 4g and 4h); grains with large-scale facets broadly prevail (Figure 4g). Stable minerals such as zircon, tourmaline, rutile, and apatite also show signs of corrosion. Zircon occasionally displays corrosion, preferentially metamictic grains. Some have euhedral shape (Figure 4i). Apatite and tourmaline usually show subhedral outlines and incipient corrosion (Figure 4j). Other tourmalines are completely transformed by corrosion to rounded grains with significant etch pits (Figure 4k). Subrounded to rounded (recycled) rutile grains reveal an initial to slight degree of corrosion (Figure 4l).

4.2.2. Heavy mineral ratios

The relative abundance of heavy minerals is reflected by the mineral indexes of garnet/zircon (GZi), chromian spinel/zircon (CZi), and apatite/tourmaline (ATi) (Morton and Hallsworth, 1994, 1999; Morton et al., 2005).

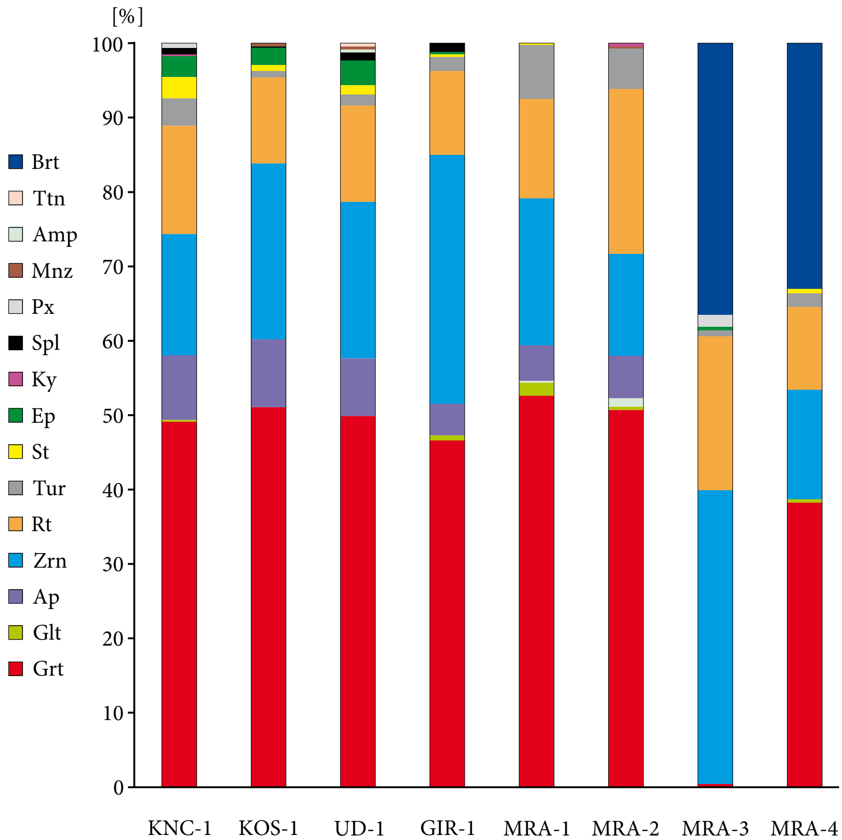

Figure 3. Heavy minerals in samples (%) from deposits of the formations investigated. Grt – Garnet, Glt – glauconite, Ap – apatite, Zrn – zircon, Rt – rutile, Tur – tourmaline, Sta – staurolite, Ep – epidote, Ky – kyanite, Spl – spinel, Px – pyroxene, Mnz – monazite, Amp – amphibole, Ttn – titanite, Brt – barite.

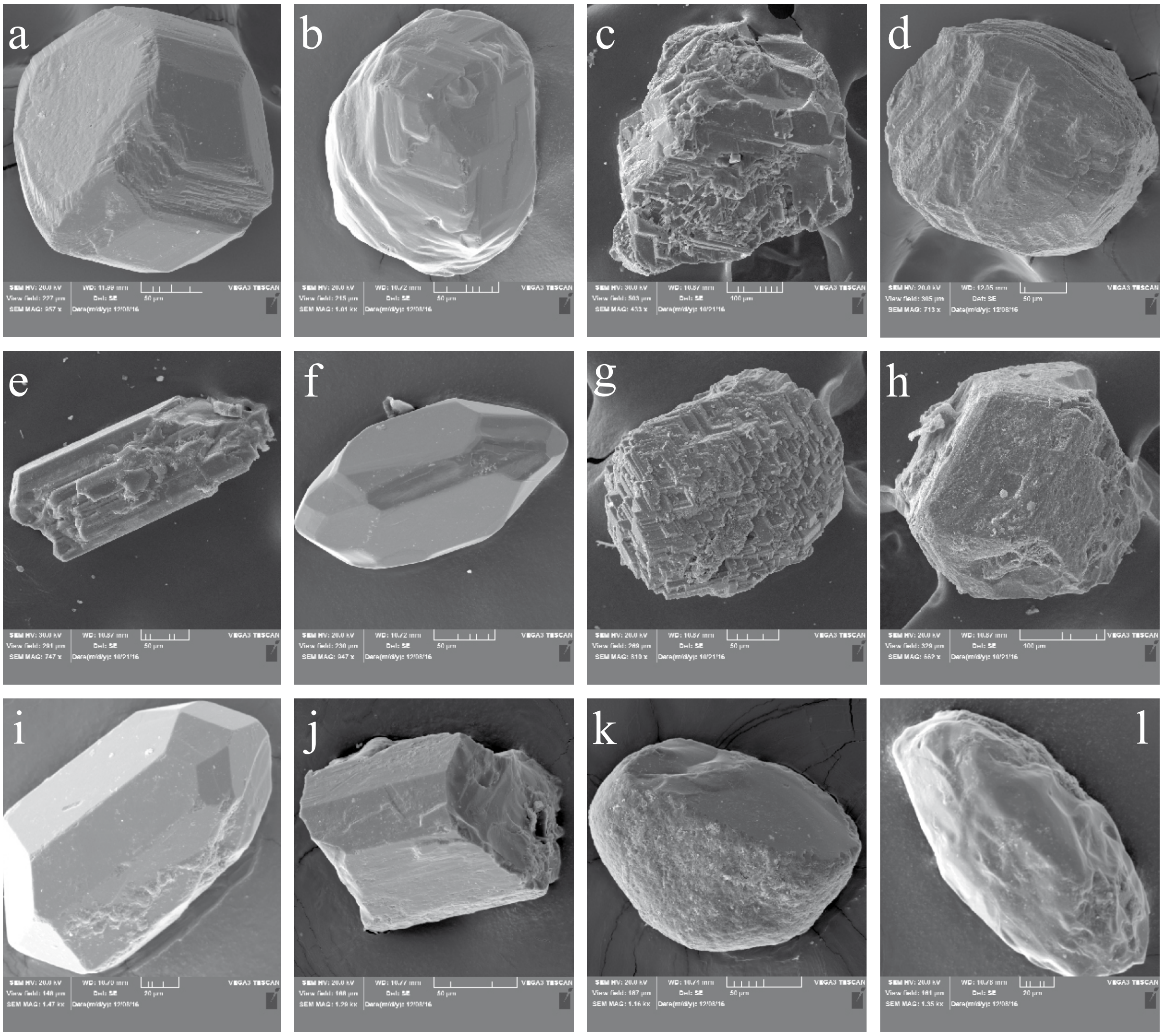

Figure 4. Scanning electron microscope images of detrital minerals point to their corrosion features from the Strihovce Fm. (a–f) and Mrázovce Mb. (g–l) deposits, respectively. For detailed description see the text.

The garnet versus zircon ratio, which is used to detect increasing chemical modification with sediment burial, ranges from 58 to 75 for the Strihovce Fm. and from 72 to 79 for the Mrázovce Mb. deposits. The apatite/tourmaline index, which is best suited for unravelling chemical alteration at the source and/or transport, is consistently high in all samples from the Strihovce Fm. (70–91), while lower values (40–51) are common for the Mrázovce Mb. deposits. Interestingly, apatite is completely lacking in the MRA-4 sample. The chromian-spinel/zircon index, which varies from 3.4 to 5 in the Strihovce Fm., provides a good reflection of source area characteristics because these minerals are comparatively immune to alteration during the sedimentary cycle. This index could be used to directly match sediments with source materials, even for suites of first-cycle origin (Morton and Hallsworth, 1994). Its rather high value indicates that a positive proportion of ophiolite detritus was chiefly supplied for the KU. On the other hand, the CZi values are negligible in the Mrázovce Mb. deposits. The ZTR index (percentage of the combined zircon, tourmaline, and rutile grains among the transparent, nonmicaceous, detrital heavy minerals, sensu Hubert, 1962), which reflects the sediment maturity, is within the range of 34%–36% (sporadically 46%) for the Strihovce Fm. and from 28% to 41% for the Mrázovce Mb.

4.2.3. Heavy mineral geochemistry

Heavy mineral analyses were performed aiming at identifying possible differences in heavy mineral compositions that can be accounted to the sediment provenance of each formation. This study is focused on garnet (Table 1) and tourmaline (Table 2). These mineral groups show some chemical variations. Results are shown in Figure 5.

Garnet. Detrital garnets from the KU form either irregular sharp fragments or isometric subrounded grains. Contrary to it, garnets from the RU are predominantly represented by subangular and subrounded fragments with apparent corrosion-induced marks (above-mentioned). Garnets in both units are pink and pale orange, usually free from inclusions, or colourless with dark dusty inclusions. The composition of garnets studied is illustrated in the ternary classification diagram of Morton et al. (2004) using almandine + spessartine, pyrope, and grossular as poles and the discrimination fields A, B I, B II, and C (Figure 5a).

Krynica Unit. Garnets from the *sandstone-conglomerate facies* (KNC-1, KNC-4, KOS-1 samples) are represented by the pyrope-almandines ($Alm_{73\text{-}83}Prp_{10\text{-}20}Grs_{2\text{-}4}Sps_{3\text{-}7}$),

Table 1. Representative microprobe analyses of detrital garnets from the Strihovce Fm. (KU) and the Mrázovce Mb. (RU) deposits. Oxides are in wt.%.

Mineral	Garnet													
Unit	Krynica Unit								Rača Unit					
Sample	UD-1		GIR-1		KNC-1		KOS-1		MRA-1		MRA-2		MRA-3	
Point	c	r	c	r	c	r	c	r	c	r	c	r	c	r
SiO_2	38.92	38.39	36.70	36.71	36.17	36.51	36.81	36.79	37.16	37.89	38.06	37.54	37.40	38.40
TiO_2	0.51	0.10	0.00	0.00	0.00	0.15	0.01	0.01	0.17	0.05	0.02	0.00	0.07	0.11
Al_2O_3	21.38	21.40	20.86	21.18	21.37	21.13	20.58	21.15	20.62	20.67	20.97	20.82	21.07	21.05
Cr_2O_3	0.00	0.00	0.03	0.03	0.02	0.00	0.02	0.04	0.00	0.00	0.01	0.01	0.00	0.01
Fe_2O_3*	0.00	0.00	2.48	1.58	2.62	2.13	2.71	1.52	0.00	0.00	0.00	0.00	0.00	0.00
FeO	26.34	25.57	27.09	27.54	30.71	13.95	30.11	30.95	18.46	19.48	31.49	31.62	25.54	23.49
MnO	0.22	0.19	8.47	8.36	7.76	18.88	5.74	5.72	17.34	13.34	2.69	3.48	5.32	2.06
MgO	6.67	5.82	3.49	3.28	1.89	0.19	3.40	2.96	1.10	1.80	5.27	4.33	1.02	0.77
CaO	5.59	7.38	1.55	1.60	1.02	8.10	1.59	1.54	5.14	6.68	1.01	1.02	9.39	14.21
Total	99.62	98.86	100.7	100.3	101.5	101.1	101.0	100.7	100.0	99.91	99.52	98.82	99.80	100.1
Si	3.033	3.016	2.940	2.949	2.908	2.928	2.946	2.952	3.007	3.039	3.034	3.033	3.000	3.040
Ti	0.030	0.006	0.000	0.000	0.000	0.009	0.001	0.000	0.010	0.003	0.001	0.000	0.004	0.007
Al	1.964	1.982	1.969	2.005	2.025	1.997	1.942	2.000	1.967	1.955	1.970	1.982	1.992	1.964
Cr	0.000	0.000	0.002	0.002	0.001	0.000	0.001	0.003	0.000	0.000	0.001	0.001	0.000	0.000
Fe^{3+}	0.000	0.000	0.150	0.096	0.159	0.129	0.163	0.092	0.000	0.000	0.000	0.000	0.000	0.000
Fe^{2+}	1.717	1.680	1.815	1.849	2.065	0.936	2.015	2.077	1.249	1.307	2.100	2.136	1.713	1.555
Mn	0.014	0.012	0.575	0.569	0.529	1.282	0.389	0.389	1.189	0.907	0.182	0.238	0.362	0.138
Mg	0.775	0.682	0.417	0.392	0.227	0.023	0.406	0.355	0.133	0.215	0.626	0.522	0.122	0.091
Ca	0.467	0.622	0.133	0.138	0.088	0.696	0.136	0.133	0.446	0.574	0.086	0.088	0.807	1.205
Total	8	8	8	8	8	8	8	8	8	8	8	8	8	8
Alm	57.75	56.07	61.73	62.72	71.01	31.86	68.39	70.34	41.40	43.53	70.13	71.58	57.04	52.01
Prp	26.07	22.76	14.19	13.31	7.79	0.78	13.78	12.01	4.42	7.16	20.92	17.49	4.05	3.05
Grs	15.46	20.69	4.20	4.46	2.80	22.17	4.26	4.29	14.70	19.09	2.87	2.96	26.82	40.17
Sps	0.48	0.42	19.55	19.29	18.18	43.66	13.21	13.16	39.41	30.19	6.08	7.98	12.04	4.63
Uv	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.01
Adr	0.00	0.00	0.32	0.21	0.22	1.43	0.36	0.20	0.00	0.00	0.00	0.00	0.00	0.00
Ca-Ti Grt	0.24	0.06	0.00	0.00	0.00	0.10	0.00	0.00	0.08	0.03	0.00	0.00	0.05	0.14

Fe_2O_3* – calculated; c – core, r – rim.

Table 2. Representative microprobe analyses of detrital tourmalines from the Strihovce Fm. (KU) and the Mrázovce Mb. (RU) deposits. Oxides are in wt.%.

Mineral	Tourmaline																	
Unit	Krynica Unit									Rača Unit								
Sample	UD-1		GIR-1		KNC-1		KOS-1			MRA-1			MRA-2		MRA-3		MRA-4		
Point	c	r	c	r	c	r	c	c/r	r	c	c/r	r	c	r	c	r	c	c/r	r
SiO_2	37.21	37.60	36.55	37.34	36.95	36.98	37.12	36.72	36.57	36.83	36.83	36.56	35.10	36.58	36.84	36.92	37.14	36.79	37.01
TiO_2	0.77	0.67	1.04	0.71	0.43	1.08	2.64	0.53	0.88	0.81	0.28	0.65	0.11	0.85	0.42	0.59	0.29	1.18	0.75
B_2O_3*	10.63	10.79	10.66	10.82	10.80	10.59	10.78	10.45	10.59	10.65	10.52	10.49	10.23	10.48	10.57	10.51	10.56	10.51	10.64
Al_2O_3	30.98	32.13	32.73	33.77	34.91	30.50	29.04	30.91	32.21	29.64	31.31	31.05	33.44	30.51	31.25	30.57	31.07	29.98	31.38
Cr_2O_3	0.00	0.09	0.19	0.06	0.05	0.05	0.26	0.03	0.16	0.06	0.05	0.08	0.04	0.04	0.05	0.07	0.00	0.06	0.00
MgO	6.67	8.31	7.61	7.89	5.63	6.86	11.72	5.14	6.19	10.47	6.34	6.07	0.57	6.44	5.81	5.79	5.59	5.53	6.01
CaO	0.30	0.40	0.93	0.59	0.54	0.51	2.56	0.10	0.48	2.63	0.07	0.55	0.17	0.62	0.08	0.09	0.10	0.20	0.36
MnO	0.00	0.00	0.05	0.05	0.04	0.02	0.00	0.02	0.00	0.02	0.03	0.01	0.09	0.04	0.07	0.05	0.03	0.06	0.01
FeO_{tot}	8.23	4.65	4.04	3.06	6.25	7.89	0.53	10.10	7.15	3.52	8.00	8.46	14.37	8.09	9.59	9.51	9.97	10.43	9.06
Na_2O	2.44	2.42	1.79	1.96	1.73	2.32	1.46	2.24	2.03	1.43	2.42	2.03	1.57	2.20	2.68	2.66	2.54	2.51	2.46
K_2O	0.02	0.02	0.06	0.04	0.03	0.03	0.05	0.01	0.01	0.02	0.02	0.01	0.03	0.01	0.01	0.01	0.00	0.01	0.01
NiO	0.00	0.00	0.00	0.02	0.00	0.00	0.00	0.02	0.01	0.02	0.16	0.03	0.00	0.00	0.01	0.00	0.01	0.01	0.00
F	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.08	0.00	0.00	0.05	0.00	0.00	0.00	0.00	0.00
Cl	0.01	0.01	0.02	0.01	0.01	0.00	0.02	0.01	0.02	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.01	0.01
H_2O*	3.66	3.71	3.67	3.72	3.72	3.64	3.71	3.60	3.64	3.67	3.62	3.61	3.52	3.61	3.64	3.62	3.64	3.62	3.66
O=F	–0.01	–0.01	–0.01	–0.01	–0.01	–0.01	–0.01	–0.01	–0.01	0.00	0.03	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	100.9	100.8	99.33	100.1	101.1	100.5	99.90	99.88	99.94	99.78	99.77	99.61	99.24	99.51	101.02	100.39	100.94	100.90	101.36
Si	6.083	6.054	5.958	6.000	5.946	6.071	5.985	6.106	6.002	6.010	6.078	6.060	5.963	6.065	6.056	6.105	6.110	6.085	6.047
AlT	0.000	0.000	0.042	0.000	0.054	0.000	0.015	0.000	0.000	0.000	0.000	0.000	0.037	0.000	0.000	0.000	0.000	0.000	0.000
$T_{tot.}$	6.083	6.054	6.000	6.000	6.000	6.071	6.000	6.106	6.002	6.010	6.078	6.060	6.000	6.065	6.056	6.105	6.110	6.085	6.047
B	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000
Cr	0.001	0.011	0.025	0.008	0.006	0.007	0.033	0.004	0.021	0.007	0.007	0.011	0.005	0.005	0.006	0.010	0.000	0.008	0.000
AlY+Z	5.969	6.099	6.247	6.396	6.568	5.902	5.503	6.058	6.229	5.700	6.090	6.066	6.660	5.961	6.055	5.958	6.025	5.845	6.044
Ti	0.095	0.081	0.127	0.086	0.052	0.133	0.320	0.066	0.108	0.099	0.035	0.082	0.014	0.106	0.052	0.073	0.036	0.147	0.092
Fe	1.125	0.626	0.550	0.411	0.841	1.083	0.072	1.405	0.981	0.481	1.104	1.173	2.042	1.122	1.319	1.316	1.372	1.443	1.238
Mn	0.000	0.000	0.007	0.007	0.005	0.003	0.000	0.003	0.000	0.003	0.005	0.001	0.013	0.005	0.009	0.007	0.004	0.009	0.002
Mg	1.625	1.994	1.850	1.890	1.350	1.679	2.816	1.275	1.514	2.547	1.561	1.500	0.145	1.591	1.423	1.427	1.371	1.363	1.463
Ni	0.000	0.000	0.000	0.002	0.000	0.001	0.000	0.003	0.001	0.002	0.021	0.004	0.000	0.000	0.001	0.000	0.001	0.001	0.001
$Y+Z_{tot.}$	8.814	8.810	8.806	8.800	8.822	8.806	8.744	8.813	8.855	8.839	8.821	8.836	8.879	8.791	8.865	8.790	8.809	8.816	8.839
Ca	0.052	0.070	0.163	0.102	0.093	0.090	0.442	0.017	0.085	0.459	0.012	0.097	0.031	0.110	0.015	0.016	0.018	0.036	0.062
Na	0.772	0.754	0.567	0.611	0.539	0.738	0.457	0.723	0.646	0.454	0.775	0.651	0.517	0.707	0.855	0.853	0.810	0.806	0.779
K	0.005	0.004	0.012	0.008	0.007	0.005	0.011	0.002	0.002	0.005	0.003	0.003	0.006	0.002	0.001	0.002	0.001	0.002	0.002
$X_{tot.}$	0.830	0.828	0.741	0.721	0.639	0.834	0.910	0.742	0.733	0.918	0.790	0.751	0.554	0.819	0.871	0.871	0.829	0.843	0.844
$X_{vac.}$	0.170	0.172	0.259	0.279	0.361	0.166	0.090	0.258	0.267	0.082	0.210	0.249	0.446	0.181	0.129	0.129	0.171	0.157	0.156
F	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.042	0.000	0.000	0.026	0.000	0.000	0.000	0.000	0.000
Cl	0.003	0.001	0.004	0.002	0.002	0.000	0.006	0.002	0.007	0.001	0.000	0.000	0.000	0.000	0.001	0.002	0.000	0.003	0.001
Mg#	0.59	0.76	0.77	0.82	0.62	0.61	0.98	0.48	0.61	0.84	0.59	0.56	0.07	0.59	0.52	0.52	0.50	0.49	0.54

B_2O_3*, H_2O* – calculated; vac. – vacancy; c – core, c/r – core/rim, r – rim; Mg# – Mg/(Mg+Fe).

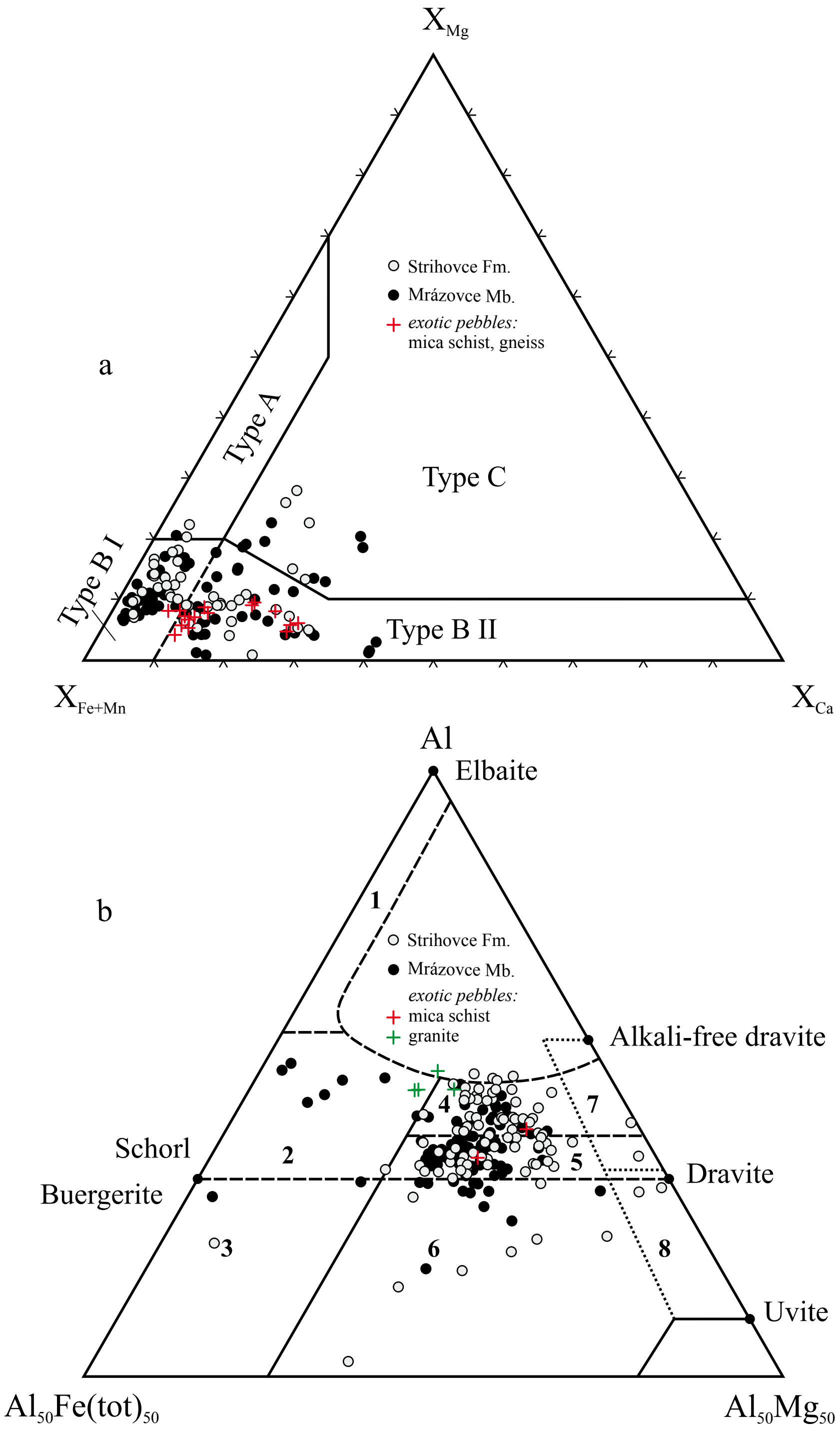

Figure 5. a) Composition of detrital garnets from the siliciclastics studied and exotic pebbles in a Fe + Mn-Mg-Ca ternary diagram (Morton et al., 2004): type A – Grt from granulites; type BI – Grt from intermediate to acid igneous rocks; type B II – Grt from metasedimentary rocks of amphibolite facies; type C – Grt from metabasic rocks. b) Al-Fe-Mg diagram for tourmalines (Henry and Guidotti, 1985). (1) Li-rich granites; (2) Li-poor granites and aplites; (3, 6) Fe^{3+}-rich quartz-tourmaline rocks; (4) metapelites and metapsammites coexisting with Al-rich phases; (5) metapelites and metapsammites not coexisting with Al-rich phases; (7) low-Ca metaultramafic rocks, Cr- and V- rich metasedimentary rocks; (8) metacarbonates and metapyroxenites.

grossular-pyrope-almandines ($Alm_{55}Prp_{28}Grs_{15}Sps_1$), almandines (87 mol% Alm), and unzoned grossular-almandines ($Alm_{78}Grs_{11}Prp_6Sps_4Adr_1$). Zoned garnets, in which almost all end-member species vary, specifically from ($Alm_{71}Sps_{18}Prp_8Grs_3$) to ($Alm_{32}Prp_1Grs_{22}Sps_{44}$), or from ($Alm_{48}Prp_4Grs_{18}Sps_{29}$) to ($Alm_{68}Prp_{11}Grs_{17}Sps_3$), are also found. Quartz, tourmaline, and biotite represent the inclusions. For the *matrix of polymictic conglomerates* (UD-1 sample), grs-alm garnets ($Alm_{60-78}Grs_{12-29}Prp_{5-15}$) with variable prp content are typical. The prp-alm garnets with grossular ($Alm_{56-58}Prp_{26}Grs_{15-21}$) and prp-alm ones ($Alm_{77-81}Prp_{14-17}Grs_{2-7}Sps_{1-7}$) were distinguished. Garnets contain infrequent inclusions such as rutile, quartz, and apatite. For *flysch facies* (GIR-1 sample), prp-sps-alm garnets ($Alm_{59-63}Sps_{20-23}Prp_{10-14}Grs_{4-7}$) and zoned grossular-almandines ($Alm_{65-73}Grs_{14-24}Prp_{8-10}Sps_{2-3}$), typical of increasing almandine at the expense of the grossular component toward the rim, are common. Pyrope-almandines ($Alm_{71}Prp_{23}Grs_3Sps_2$) are scarce.

Rača Unit. There are unzoned pyrope-almandines ($Alm_{74-84}Prp_{12-17}$), grossular-pyrope-almandines ($Alm_{61-70}Prp_{19-23}Grs_{10-16}$), grossular-almandines ($Alm_{62-80}Grs_{13-30}$), and grossular-almandines with pyrope ($Alm_{48}Grs_{30}Prp_{20}$) or spessartine ($Alm_{40}Grs_{40}Sps_{20}$), along with spessartine-almandines with pyrope ($Alm_{57-70}Sps_{15-31}Prp_{8-10}$) or grossular ($Alm_{50-70}Sps_{11-30}Grs_{11-17}$). Zoned grossular-almandines with variation in pyrope and/or grossular components (from $Alm_{61}Grs_{23-28}Sps_{10}Prp_{4-6}$ to $Alm_{58-72}Grs_{12-21}Prp_{9-12}Sps_7$ and from $Alm_{57}Grs_{27}Prp_4Sps_{12}$ to $Alm_{52}Grs_{40}Prp_3Sps_5$), respectively, were also found. In these garnets, the Ti amount correlates positively with grossular content. They usually constitute inclusions such as ilmenite, zircon, allanite, and quartz. White mica, chlorite, and plagioclase appear together within sps-grs almandine.

Tourmaline. Tourmaline occurs usually as short and abrupt prismatic grain, usually of brown to dark brown colour. Rounded and subrounded tourmalines with the same colour are scarcer. Sharp-edged splinters were also found. All forms noted above were found in both the Krynica and Rača units. Some tourmalines are inclusion-rich: quartz and zircon occurred in the RU, while quartz, albite, rutile, ilmenite, apatite, zircon, and titanite were found in the KU.

Krynica Unit. The EMP analyses show that the detrital tourmalines belong to the alkali-tourmaline primary group, in which Na^+ predominates (0.53-0.89 apfu) over Ca^{2+} (0.01–0.38 apfu) and K^+ (<0.01 apfu). Only one grain (inherited core) represents the calcic-tourmaline primary group, with Ca^{2+} at 0.44 apfu (KOS-1 sample). Generally, the Y-site position is dominated by Mg^{2+} (1.12-2.82 apfu) and Fe^{2+} (up to 1.72 apfu) with subordinate content of Mn^{2+} (0.0-0.02 apfu). Molar X_{Mg} = [Mg/(Mg + Fe)] values vary in the wide range of 0.37 to 0.99. Tourmalines from the *sandstone-conglomerate facies* (KNC-1, KNC-4, and KOS-1 samples) could be divided into three categories: zoned grains with an inherited core (Figures 6a and 6b), a developed inner rim, and overgrowth marginal zone; zoned grains with no inherited core; and unzoned grains. Zoned tourmalines display a shift from schorlitic-dravitic inherited core to overgrowth showing dravitic composition in the rim (sensu Henry et al., 2011). Some inherited cores show pure dravite composition (up to 11.72 wt.% MgO) with high Ti (up to 2.64 wt.% TiO_2) and eventually Cr (0.26 wt.% Cr_2O_3) contents; one detritic core belongs to schorlitic tourmaline (21 wt.% FeO). Tourmalines from the *matrix of polymictic conglomerate* (UD-1 sample) as well as from the *flysch facies* (GIR-1 sample) show identical characteristics. They are zoned (Figure 6c), with or without an inherited core, and point to a dravitic composition (Henry et al., 2011). Their molar X_{Mg} = [Mg/(Mg + Fe)] value is in the range of 0.50 to 0.96.

Rača Unit. Detrital tourmalines belong to the alkali-tourmaline primary group, in which Na^+ predominates (0.41–0.87 apfu) over Ca^{2+} (0.0–0.26 apfu). Some inherited cores represent the calcic-tourmaline primary group, with Ca^{2+} from 0.42 to 0.46 apfu (MRA-1, MRA-2 samples). Based on the dominant divalent cations in the Y-site position, which are also Fe and Mg, tourmalines belong to dravitic ones (Henry et al., 2011). Molar X_{Mg} = [Mg/(Mg + Fe)] values in tourmalines range from 0.46 to 0.84. Some inherited cores show a schorlitic composition (MRA-2 sample; Figure 5b). Molar X_{Mg} = [Mg/(Mg + Fe)] values in these cores range from 0.05 to 0.35.

According to the diagram indicating the environment of tourmaline origin (Henry and Guidotti, 1985), the grains were derived from metapelites that coexisted or did not coexist with Al-rich phases, sporadically from quartz-tourmaline rocks (Figure 5b). The grains coexisting with the Al-rich phase show low to medium content of Ca and Ti. Schorlitic inherited cores found mainly in the Mrázovce Mb. deposits originated from Li-poor granitoids, while dravitic cores identified just in the Strihovce Fm. originated in metacarbonates and metapyroxenites or Ca-poor ultramafites (Figure 5b). Unzoned tourmalines, typical of the Strihovce Fm., indicate origin in Al-rich metapelites (Henry and Guidotti, 1985).

4.2.4. Zircon internal structure

In both units, zircon forms either euhedral to subhedral short-prismatic dipyramidal grains or long-prismatic (to acicular) ones without signs of corrosion (Figures 4f, 4i, and 6d–6i). Both groups are colourless, pale yellow, or pink. Rounded zircon shapes are also present. Their colour is the same – colourless, pink, and yellowish. Rounded zircons are more common in the Mrázovce Mb. deposits.

Krynica Unit. Following the CL images, a couple groups could be distinguished. For *sandstone-conglomerates*

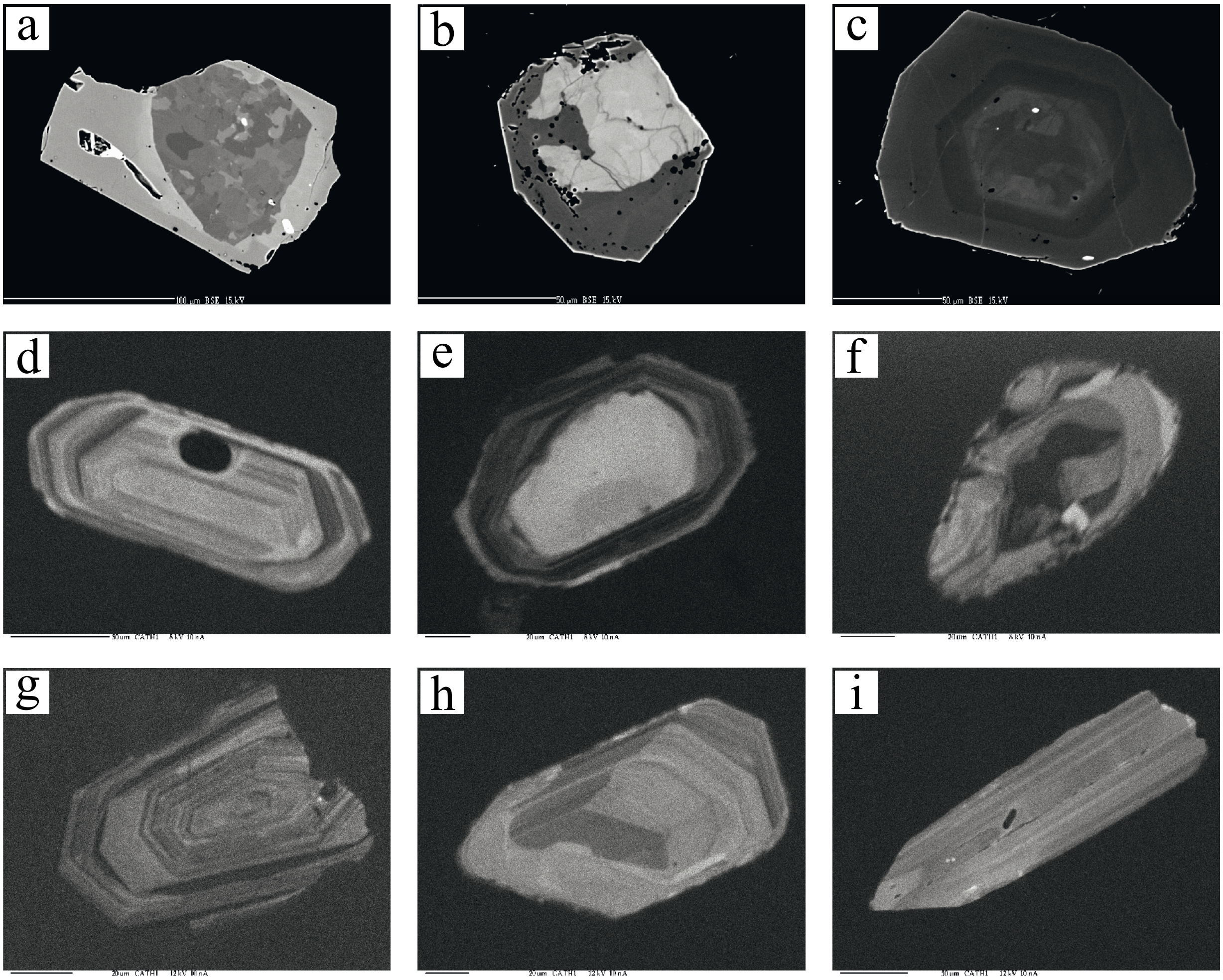

Figure 6. Representative BSE images of detrital tourmaline (a–c) and CL images of zircon (d–i) from the siliciclastics investigated: a–c) tourmaline with inherited cores from the Strihovce Fm. (KOS-1, KNC-1, UD-1 samples); d–f) internal structure of zircons from the Strihovce Fm. (KNC-1, KNC-4, GIR-1 samples); g–i) zircons from the Mrázovce Mb. deposits (MRA-2, MRA-1 samples). For detailed description see the text.

facies (KNC-1, KNC-4, KOS-1 samples), the euhedral to subhedral short-prismatic zircons with a concentric oscillatory zoning and local marginal resorption are characteristic (Figure 6d). Compared to the remaining groups, they are scarcer. The second group is represented by zircons with unzoned inherited partially resorbed cores with a new zircon phase, "embayments". This phase is resorbed in the final zircon growth stage by a zone with irregular oscillatory zoning (Figure 6e). Zircons with recrystallised inherited cores, on which a new concentric to oscillatory zoned phase has grown, represent the third group. Zircons from the *matrix of polymictic conglomerates* (UD-1 sample) show regular oscillatory zoning without local recrystallisation patterns. Some zircons exhibit a sector zoning. Unzoned zircon grains indicating the first-stage rapid crystallisation represent an individual group. A few grains show broadly reworked internal texture. In the *flysch facies* (GIR-1 sample), there are zircons with convolute zoning gradually undergoing to chaotic internal texture. It is developed in the whole grain's profile (Figure 6f). Zircons showing concentric to regular oscillatory zoning are also present.

Rača Unit. A few types of zircons could also be distinguished in the Mrázovce Mb. deposits: euhedral grains with inclusions of ilmenite, quartz, titanite, and feldspar; drum-like zircons with acute pyramids and rounded grains. Cross-sections through zircons point to wide inner variability: zircons without or with inherited cores showing regular oscillatory zoning (Figure 6g); zircons with sector zoning; rounded zircons with patchy to chaotic inner texture; euhedral zircons showing oscillatory zoning, which is sharply abrupt, by the growing of the newly formed rehomogenised phase (Figure 6h); and finally dendritic zircons (Figure 6i).

5. Discussion

5.1. Provenance of the heavy minerals

Corrosion features of heavy minerals. The relative roughness of the exterior of a grain or the facets developed on its surfaces can provide information about the mechanical and corrosion history that the detrital grain has experienced. Faceted garnets in heavy mineral suites usually denote the dissolution after deposition (e.g., Hansley, 1987; Morton et al., 1989; Salvino and Velbel, 1989). In this case, some heavy minerals can fully disappear.

In the HMA from the Strihovce Fm., grains almost without corrosion signs coupled with those showing the corrosion formed during weathering (indicative of a palaeoclimatic setting) or diagenetic processes are common. Existence of weathering and unweathering garnet surfaces simultaneously denotes their erosion before deposition. Faceted grains occur in sandstones with calcite cement. The abundance of Ca, which can act as a buffer to acid pore solutions, could control the process of etching (Borg, 1986). An additional alternative is a resedimentation from older formations, or it may reflect provenance from multiple and differently weathered sources, including bedrock of various types and maturity.

For the Mrázovce Mb., the shape of garnets seems to be a result of diagenesis. Postdepositional dissolution is indicated by the absence of any evidence of abrasion on their surfaces (Figure 4g). Furthermore, the bulk grains in the HMA show different degrees of diagenetic dissolution. Indeed, the polymictic character of detrital garnet associations and Ca-rich garnets in the HMA indicate that associations were not substantially modified via diagenetic dissolution. Therefore, they can be considered as the original and corresponding to the source rocks. The scarcity of minerals such as kyanite and staurolite suggests that they were sporadically present in the source rocks. This is consistent with the tourmaline geochemistry in the HMA. Rare euhedral tourmalines without any corrosion marks appear to be first-cycle origin from a proximate source (MRA-2 sample), though the bulk tourmaline grains show recycled origin. A similar situation relates the zircons. Barite derivation is debatable. Based on crystal-morphology characterisation, we speculate that abundant barite is diagenetic. Diagenetic barite formed in sediments is large (20–700 µm) and typically consists of flat, tabular-shaped crystals or nodules in sedimentary layers or mounds of porous crystals, which exhibit a layered appearance of platy crystals in diamond-shaped clusters (Paytan et al., 2002; Griffith and Paytan, 2012). Barite from the Mrázovce Mb. deposits usually shows tabular-shaped crystals. Diagenetic barite is known from the Sub-Silesian Unit in the flysch Carpathians (Leśniak et al., 1999). The sedimentary origin of barite distributed in the Rača deposits (obtained from the panned concentrates) was referred by Bačo et al. (2015). Summarising, there are differences in the corrosion features of HMAs between deposits studied, indicating with all likelihood the deeper burial of the Mrázovce Mb. deposits counter to the Strihovce Fm.

Heavy mineral ratios. The HMAs show the rather immature character of deposits studied via their low values of the ZTR index. The variability of garnet and zircon surfaces denotes that the deposited heavy minerals were supplied from primary source rocks (metamorphic and igneous origin), which were subjected to weathering during the Eocene, and also from the recycled source rocks assumed mainly for the Mrázovce Mb. deposits.

According to provenance-sensitive heavy mineral ratios proposed by Morton and Hallsworth (1994, 1999), both the Krynica and Rača subbasins were supplied by sources of well-stocked garnet (high GZi). Despite a significant degree of corrosion on garnet surfaces, typical mainly of the Mrázovce Mb., the GZi index shows no loss of garnet due to burial. The presence of Ca-rich garnets, for instance, which are less stable than Ca-poor ones during diagenesis (Morton, 1987; Morton and Hallsworth, 2007), also confirms this presumption. Moreover, the HMAs consist of ultrastable and less stable heavy minerals, which excludes the possibility that the components more prone to disintegration underwent significant dissolution at the final deposition site or during the burial.

Potential sources for detrital garnets. The elevated Alm content in combination with the lower Prp suggests amphibolite-facies conditions (Deer et al., 1992, 1997), partially leucosomes in migmatites (Suggate and Hall, 2014). The low grossular values (<10 mol%) in the almandines are more typical of felsic rocks (gneisses and felsic granulites), whereas a high grossular content occurs in garnets from micaschists and rocks of mafic composition (amphibolites, mafic granulites). The high spessartine content is an indicator of igneous source rocks (granitoids and/or pegmatites and volcanic rocks; Deer et al., 1997 and references therein), as well as of low-P/T metamorphic rocks, especially those in thermal aureoles (Miyashiro, 1955; Deer et al., 1982; Morton et al., 2004).

Heterogeneous association of detrital garnets indicates their origin in different types of medium-grade metapelites such as garnet micaschists, gneisses, and amphibolites (Figure 5a). Scarce pyrope-rich almandine garnet (Prp ~28 mol%) with grossular could have originated in mafic granulites. Almandines with the Sps component may have been derived from granites or pegmatites. Analogous composition of detritic garnets was found in the Upper Cretaceous-Palaeocene fms of the Krynica Unit in Poland (Salata, 2004). The most iron-rich almandine (87 mol% Alm) could be related to metamorphic pelitic parent rocks (Deer et al., 1997). Grossular-rich almandine with pyrope (the Mrázovce Mb.) may occur in amphibolites (l. c.). Garnets showing chemical zoning, where the different distribution of Fe, Mg, Ca, and Mn via individual zones of almandines was observed, could also be formed in the low- to medium-grade regionally metamorphosed pelitic rocks. These garnets were found in both formations studied. Garnets showing continuously decreasing X_{Grs} and X_{Sps} and increasing X_{Prp} and X_{Alm} toward the rim suggest increasing temperature during the crystallisation of their rims (Spear, 1993). There are also zoned garnets with Ca-rich rims (Grs up to 40 mol%). The simplest interpretation for this chemical zonation of garnet is that the sharp increases of grossular content and Ti may reflect successive Ca-metasomatic events, i.e. infiltration by high-Ca fluids (Stowell et al., 1996). Another explanation of the growth of grossular rims on pre-Alpine(?) garnet cores is their Alpine metamorphic overprint. This phenomenon, characteristic for metapelites and metagranitoids, is well known from the crystalline basement of the Central Western Carpathians (Korikovsky et al., 1990; Méres and Hovorka, 1991) and the Tisza Mega-Unit (Horváth and Árkai, 2002), though the Grs component reached up to 30 mol% (l. c.). Extensive solid solution between grossular and almandine has also been found for garnets from low-grade metamorphic rocks (Deer et al., 1997).

Potential sources for detrital tourmalines. The majority of grains display a metamorphic origin. The rather low proportion of Ca and X-site vacancy in the tourmalines studied suggests a medium grade of metamorphism according to Henry and Dutrow (1996). This is also supported by the relatively high Mg/Fe ratio (average 1.96 and 1.66 for the KU and RU, respectively) in the bulk of tourmalines, typical of the medium grade (Henry and Guidotti 1985).

Zoned tourmalines (with inherited cores) indicating a polycyclic genesis, at which the rims are enriched by dravitic component, denote the activity of metamorphic fluids derived from the regional metamorphosis of metapelite. The inherited cores that originated in metaultramafites, metacarbonates, or Cr-rich metasediments found in the Strihovce Fm. have not been identified yet in the western part (the Biele Karpaty Unit) of the Flysch Belt (Bačo et al., 2004). This unit is considered to be a palaeogeographic and tectonic equivalent of the Strihovce Nappe (Eastern flysch zone; Potfaj in Bezák et al. 2004). Tourmalines that originated in igneous-hydrothermal fluids derived from granites were not found in the Strihovce Fm. Contrary to it, more frequent inherited cores of schorlitic composition, which could be related to the Li-poor granitic genetic environment (Henry and Guidotti, 1985), were identified in the Mrázovce Mb. deposits. The schorlitic-dravitic tourmaline from granite pebbles shows a similar chemical composition as the hydrothermal tourmaline from the Gemeric granite (Central Western Carpathians, Broska et al., 2012). It suggests that some detrital schorlitic-dravitic tourmaline may have been derived from granite, whereas fluids required for its formation come from the regional metamorphosis of metapelites (Jiang et al., 2008).

Potential sources for detrital zircon. Variations in zircon colour, crystal shape, and internal zoning suggest a variety of source rocks. The population of elongate, euhedral zircons with oscillatory zoning points to significant igneous input (Vavra, 1993, 1994) into the Magura Basin, suggesting that deposition mainly of the Strihovce Fm. deposits was relatively close to an igneous source. Some zircons occur as euhedral drum-like grains, sometimes with sector, convolute, or chaotic internal structure indicating a metamorphic origin (Corfu et al., 2003; Hoskin and Schaltegger, 2003). Complicated zoned patterns such as truncated zoning, which is following by new zircon overgrowths, are indicative for recrystallization in the metamorphic conditions. These zircons are typical for the Mrázovce Mb. deposits.

Provenance remarks. The HMAs and especially the geochemical data of garnets well reflect the geological situation in the source areas. Chiefly garnets in both formations investigated show generally similar chemical compositions, suggesting a provenance from lithologically resembling source rocks. We note that the garnets with rather high grossular content (up to 40 mol%) in the Mrázovce Mb. deposits were not found in the Strihovce Fm. Insufficient staurolite in the Mrázovce Mb. deposits may suggest the contribution from the staurolite-free parent rocks (staurolite was found only in one gneiss pebble). The HMAs point to a provenance dominated by low- to medium-grade (rarely high-grade) metamorphic terranes including rocks of igneous origin for both units. Although it may seem that tourmalines actually show metasedimentary origin, their initial source rocks were diverse – granite (RU) versus ultramafic rock (KU). High concentrations of zircon are attributed to widespread low-grade metamorphic (phyllite, micaschist) and igneous rocks. Their euhedral shape, typical of the KU, suggests the granitoid origin. The occurrence of Cr-spinels clearly points out a ultramafic (ophiolitic) source that had been feeding mainly the southern part of Magura basin (Krynica subbasin). The Cr-spinel amount in the Mrázovce Mb. deposits would rather suggest that the ophiolites did not deliver detritus in large quantities.

5.2. Exotic pebble origin

The diversified petrographic inventory of exotics in both units includes the crystalline and sedimentary rocks. The pebbles known from the KU (Stráník, 1965; Wieser, 1967; Nemčok et al., 1968; Marschalko et al., 1976; Mišík et al., 1991a; Oszczypko et al., 2006; Salata and Oszczypko, 2010) signalise a significant lithological variability compared to those from the RU. The determining factor for the source area may be an absence of Silurian graptolite shales, Culm sediments, Devonian and Dinantian limestones, and Carboniferous coal (Mišík et al., 1991a), which are known as clasts in the external parts of the Flysch Belt (e.g., Soták, 1985).

The rounded quartz, dark-grey siliceous rocks, quartzites, and granites observed within the RU exotics may denote their resedimentation. Poorly rounded pebbles of metamorphic rocks such as phyllite and micaschist, representing the rocks with rather little stability, point to transport from a relatively proximate source or existence of a primary (nonrecycled) source. Eventually, resedimentation of the rounded material from older formations and their mixture with local, short-distance transported material appears to be a viable explanation. Among crystalline pebbles found in the Mrázovce Mb. deposits, there were not discovered any red orthogneiss, limburgite and kersantite, red granite, or metalydite, which are specific to the Strihovce Fm. (Marschalko et al., 1976); beyond, granite with nongranitic origin of tourmaline is a significant mark for the Mrázovce Mb. deposits.

5.3. Palaeogeographic notices – implications for the Eocene provenance

The heavy mineral analysis unambiguously confirmed a terrigenous material in the source area. It may be speculated about the Tisza Mega-Unit, which is made up of Variscan crystalline complexes and post-Variscan – Alpine overstep sequences. In the pre-Alpine (Variscan) basement of the Tisza Mega-Unit the prevailing rock association consists of gneiss, micaschist, amphibolites, and granitoids (Szederkényi et al., 2012; Haas and Buday, 2014). Based on the palaeogeographic position of the Tisza Mega-Unit during the Eocene (Csontos, 1995; Csontos and Vörös, 2004; Handy et al., 2014; Kováč et al., 2016), palaeoflow analysis indicating the SE source (Koráb et al., 1962; Kováčik et al., 2011, our data), and the data dealing with its exhumation (e.g., Merten et al., 2011; until the Middle Eocene according to Kounov and Schmid, 2013), this unit could be a possible source area. Chemical analyses of garnets studied show similar composition to those from the Tisza Mega-Unit crystalline basement (cf. Horváth and Árkai, 2002). Finally, monazite chemical dating from exotic gneiss pebble (Poprawa et al., 2006) and exotic metapelitic rocks derived from the "southern source" (Oszczypko et al., 2016) signalise Variscan ages. On the other hand, there are serpentinite and eclogite (Baksa Complex) broadly identical to those from the Moldanubicum of the Bohemian Massif (Horváth et al., 2003). Following the petrological features of ultrabasic and granitoid rocks, age, and fossil content (e.g., Silurian black shales), for the SW part of the Tisza crystalline basement and the Bohemian Massif, a similar geological development was suggested (Klötzli et al., 2004; Kovács et al., 2016 and references therein). Detritus from these types of rocks has not been found in the Strihovce flysch siliciclastics. Detrital garnets studied show different compositions beside those from the Bohemian Massif deposits and juxtaposed units (Biernacka and Józefiak, 2009; Kowal-Linka and Stawikowski, 2013 and references therein). They are commonly enriched in pyrope molecule (Čopjaková et al., 2005), typical of eclogites and/or granulites (Méres, 2008). Certain differences are also found in the tourmaline composition (Biernacka, 2012 and references therein). In consequence, the Bohemian Massif as well as the units or terranes with a similar development may be excluded as a potential source.

At first sight, there are no considerable differences between the Mesozoic (Triassic-Jurassic) sequences of the Tisza Mega-Unit (cf. Véber, 1990; Hass and Peró, 2004; Vozár et al., 2010; Kovács et al., 2011) and the Strihovce exotic pebbles of the Mesozoic age (Mišík et al., 1991a). Nevertheless, first, a low-grade regional metamorphism connected with the Cretaceous nappe stacking within the Tisza microplate (Árkai, 2001) was not recognised in the Strihovce pebbles (Mišík et al., l. c.), albeit the Mesozoic formations of the Tisza Unit were locally metamorphosed only in the vicinity of the Alpine overthrusts (Árkai, 2001). In the second place, there are no conodonts in the Triassic exotic pebbles (Mišík et al., 1991a), while for the Tisza pelagic limestone facies (appearance in the Mecsek and Villáni units) they are known (Kovács et al., 2005, Kovács and Rálisch-Felgenhauer, 2005). For the Békes Unit, conodonts were not identified. It should also be mentioned that the Cretaceous sequences cannot be fully paralleled. Although the Early Cretaceous fossil genera from the Tisza Mega-Unit carbonates (Császár and Turnšek, 1996) were also described in the Strihovce exotic pebbles, the bulk of them indicate a specific sedimentary environment unknown so far from the KU detritus. Senonian limestones identified just from boreholes are sporadically preserved in the Tisza Mega-Unit (Császár and Haas, 1984), so they could not have supplied the Krynica subbasin, which is filled by Senonian limestone-bearing pebbles (Mišík et al., 1991a). Despite this, we assume that the Krynica subbasin might have been fed namely by crystalline detritus, mostly from the NE part (today's coordinates) of the Tisza Mega-Unit.

Rare and exotic rocks such as red orthogneiss, kersantites, and limburgites represent the particularity of the KU pebbles (Marschalko et al., 1976). They were interpreted as material that originated in the substratum of the Magura Basin, concretely from the North European Platform (NEP, Marschalko et al., 1976). Platform kersantites (NW Poland) show different mineral compositions (Pendias and Ryka, 1974) compared to those from the KU. The red orthogneisses often associated to anorthosites are known from the NEP (e.g., Cymerman, 2007), while anorthosite pebbles were not found in the exotic material. "Granitic gneisses with red feldspars" were identified from a drill hole within the Rzeszotary Terrane in Poland (Konior, 1974). The present basement of the KU could be formed by the Upper Silesian Unit (USU) and/or Małopolska Terrane (MPK, Rylko and Tomaś, 2005). The USU along with Brunovistulicum may continue to the Moesia Terrane (Seghedi et al., 2005; Oczlon et al., 2007; Kalvoda et al., 2008). The USU, MPK, and West Moesia represent a piece of Baltica-derived crust (Oczlon et al., 2007), notwithstanding that the Małopolska Terrane partially differs from the USU (Malinowski et al., 2005). This may indicate Late Neoproterozoic modification of the USU crust or accretion of exotic crust, now underlying the USU, to the southern margin of Baltica (Oczlon et al., 2007). The crystalline basement of the MPK is unknown, while the USU constitutes the northern part of the Brunovistulicum, the Precambrian basement of which includes granitoids and amphibolite facies metamorphic rocks (e.g., Franke et al., 1995). On the other hand, the Rzeszotary Terrane cropping out southwards of Krakow (Poland), entrained between the southern USU and MPK, may represent a displaced fragment of East Avalonian crust and not form the basement to any of those (Oczlon et al., 2007). The same type of crust is cropping out in Dobrogea. Metamorphic formations vary in metamorphic grade from middle-upper amphibolite facies to lower greenschist facies (Seghedi, 2012 and references therein). In the South Dobrogea, the cratonic basement of the Moesian Platform is comparable to that of the Ukrainian Shield, containing Archean gneisses (Giuşcă et al., 1967; Seghedi et al., 2005). This proximal Baltican terrane was displaced here along the Trans-European Suture Zone (Oczlon et al., 2007).

Taking into account a complicated evolution and the structure of the flysch substratum, about which we have only poor information from the exotics and perhaps from heavy minerals, the source area for the lateral (southern) input of detritic material for the Krynica subbasin remains enigmatic and speculative. We consider that it could be a crustal segment with basement rocks (Proterozoic and Palaeozoic fragments) from Caledonian terranes on the southwestern margin of the East European craton. Since the Permian time, the Outer Carpathian domain belonging to the eastern European continental border between the Bohemian Massif and the Ukrainian Shield was peneplenised. The Triassic sedimentary record in this area is mostly unknown (Michalík, 2011).

Summarising, the source area for lateral input referred to as the South-Magura Cordillera (sensu Marschalko et al., 1976 and Mišík et al., 1991a) could be a crustal fragment from aforementioned units incorporated into the ravel of the Tisza crystalline block. This block could have evolved independently during the Senonian. We assume that its position was not subparallel to the Pieniny Klippen Belt (e.g., Mišík et al., 1991a) but was rather subparallel to the Tisza Mega-Unit realm (Figure 7).

The Marmarosh Massif, located roughly SE (palaeocoordinates) from the Magura Basin during the Early to Middle Eocene, seems to be a potential (longitudinal) source for clastics of crystalline origin. It consists of volcanites, metasediments metamorphosed in a green-schist facies, gneisses, amphibolites, and granites of Proterozoic to Palaeozoic age (Zlatogurskaja et al., 1976; Khain, 1984; Grinchenko et al., 2005). Metasedimentary rocks commonly contain the tourmalines of schorlitic-dravitic series (Szakáll et al., 2002; Matkovskyi et al., 2011) and their garnets show similar geochemistry (Matkovskyi, 2009; Matkovskyi et al., 2011) as the detrital ones from the Strihovce Fm. deposits. According to Bónová et al. (2017), the geochemistry of volcanic chromian spinels from the Strihovce Fm. indicates that they may have been derived from the Fore-Marmarosh Suture Zone (sensu Hnylko, 2011b; Hnylko O et al., 2015). Trough Triassic deposits with radiolarites that occurred in the Rachov Massif (Kazintsova and Lozynyak, 1985) reveal their distinct character compared to Triassic exotic pebbles from the Strihovce Fm. (Mišík et al., 1991a) per quod it does not promote the near position but does not exclude the origin of psammitic-pelitic detritus from this source area. Such detrital material could be transported via turbidite flows from several hundred kilometres in distance.

Although it may seem that the Bohemian Massif was a possible source for the Rača Unit, because it was proposed for more external zones of the Flysch Belt, much like for the western sector of the Magura Nappe (e.g., Krystek, 1965; Otava et al., 1997, 1998; Salata, 2014), this idea has no basis for the Mrázovce Mb. siliciclastics. Deficiency of rutile and abundant epidote in the Culm horizonts of Drahany Upland (Čopjaková et al., 2001) constituting the mixture of rocks from the Bohemian Massif (Hartley and Otava, 2001) are an important signal for diverse sources. The amount of these minerals in the Mrázovce Fm. deposits is reversed (Figure 3). The depth of burial could have been an argument for epidote lack but rutile is an ultrastable mineral. It cannot be ascribed to resedimentation of material from older strata, which could change the ratio between referenced minerals, because the tourmaline is significantly lower in the Mrázovce Mb. deposits than in the Late Cretaceous to Palaeocene fms (cf. Salata, 2004). High-grade metamorphic detritus known from the Culm Basin (Hartley and Otava, 2001; Čopjaková et al., 2005), perhaps even from Cretaceous and Palaeogene flysch deposits of the Ukrainian Carpathians (Silesian and Skole

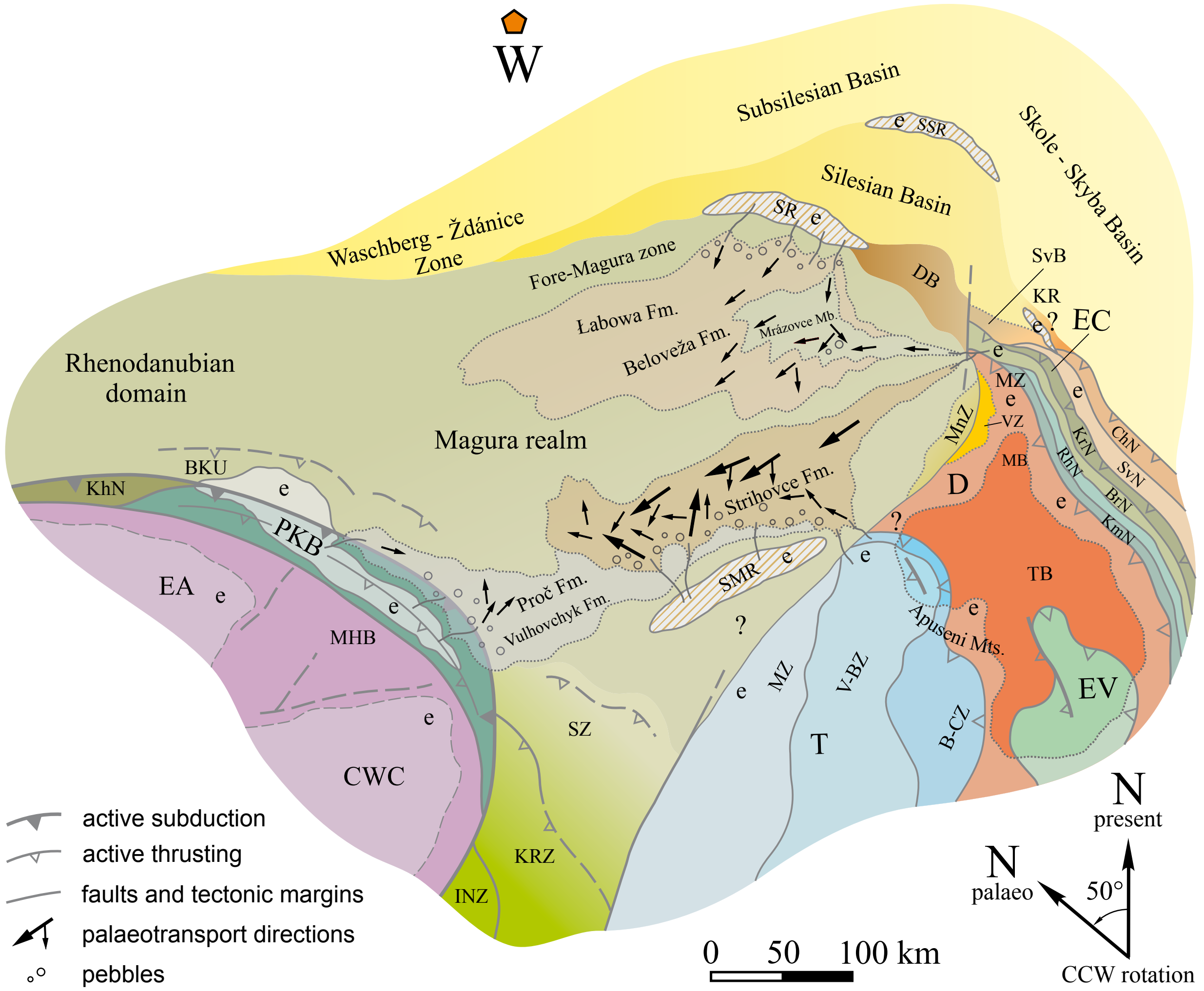

Figure 7. Schematic palaeogeographic situation of the Magura Basin and adjacent tectonic units during the Late Ypresian (created on the basis of own investigations and research of Koráb et al., 1962; Stránik, 1965; Contescu et al., 1966; Nemčok et al., 1968; Marschalko, 1975; Marschalko et al., 1976; Soták, 1990; Mišík et al., 1991a, 1991b; Oszczypko and Oszczypko-Clowes, 2006, 2009; Márton et al., 2007, 2013; Schmid et al., 2008; Hnylko, 2011a; Merten, 2011; Merten et al., 2011; Kováčik et al., 2011, 2012; Plašienka, 2012; Handy et al., 2014; Hnylko and Generalova, 2014; Plašienka and Soták, 2015; Hnylko O et al., 2015; Hnylko OM et al., 2015; Hnylko and Hnylko, 2016; Bónová et al., 2016, 2017; Kováč et al., 2016). MnZ – Monastyrets Zone; DB – Dukla basin, SR – Silesian Ridge, SSR – Sub-Silesian Ridge, KR – Kumane Ridge, SMR – South Magura Ridge (Cordillera); CWC – Central Western Carpathians, MHB – Myjava-Hričov Basin, EA – Eastern Alps; KhN – Kahlenberg Nappe; PKB – Pieniny Klippen Belt; BKU – Biele Karpaty Unit; INZ – Inacovce Zone; KRZ – Kricevo Zone; SZ – Szolnok Zone; T – Tisza Mega-Unit; MZ – Mecsek Zone, V-BZ – Villány-Bihor Zone, B-CZ – Békés-Codru Zone; D – Dacia Mega-Unit; TB – Transylvanian Basin (land and epicontinental area), MB – Maramures Basin (trough), VZ – Vezhany Zone; EV – Eastern Vardar ophiolitic unit; EC – Eastern Carpathians; MR – Marmarosh (Rakhov) Zone (elevation), ChN – Chornohora Nappe; SvN – Svydovets Nappe, SvB – Svydovets Basin (dipping part); KrN – Krasnoshora Nappe, Fore-Marmarosh Suture (Ceahlau): BrN – Burkut Nappe; RhN – Rakhiv Nappe, KmN – Kamyanyi Potik Nappe, W – Wien, e – emerged land.

units), where it has been interpreted as a material from the Bohemian Massif (Tsymbal and Tsymbal, 2014), was not found in the Mrázovce Mb. deposits.

In terms of palaeogeography, there could be mentioned the fauna of "Frydek type" biofacies discovered in the Mrázovce Mb. (Kováčik et al., 2006). This fauna is known from northern flysch nappes – Sub-Silesian, Silesian, and Skole units (e.g., Liszkowska and Morgiel, 2013). It suggests along with palaeoflow directions (NW) the Fore-Magura provenance – most probably the Silesian Ridge. The geological composition of the Silesian Ridge varied from north to south (Soták, 1992; Budzyń et al., 2008, 2011) and did not relate to the Bohemian Massif (Soták, 1990).

6. Conclusions

Based on heavy mineral assemblages obtained here, the terrigenous crystalline material dominates in both investigated formations. Although the heavy mineral compositions indicate that the major crystalline sources for the Eocene siliciclastic formations of the Rača and Krynica units were greenschist to amphibolite facies of the metamorphic rocks and granitoids, some dissimilarities occurred in these suites as well as exotics suggesting the different sources. While for the Krynica Unit we consider that the Tisza Mega-Unit crystalline complexes included a segment of the flysch substratum that could represent the lateral input, the Rača Unit might have been fed from the northern source formed by the Silesian Ridge. The Marmarosh Massif (coupled with the Fore-Marmarosh Suture Zone) is promoted to be a longitudinal source.

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