

SOURCE ROCKS FOR HEAVY MINERALS IN LOWER PART OF MENILITE FORMATION OF SKOLE NAPPE (POLISH FLYSCH CARPATHIANS), BASED ON STUDY OF DETRITAL GARNET AND TOURMALINE

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Abstract: The study focused on the chemistry of detrital garnet and tourmaline from sediments of the Boryslav and Kliva Sandstone types in the Oligocene part of the Menilite Formation of the Skole Nappe (Western Outer Carpathians, Poland), with regard to provenance. Almandine and almandine-pyrope compositional varieties are the most common garnets, with minor almandine-pyrope-grossular garnet. Scarce garnet grains, with grossular and spessartine as the dominant end-members, are also present. The tourmaline belongs to the alkali tourmaline principal group and represents the schörl-dravite series. The detrital garnet and tourmaline display strong, compositional similarities to minerals, occurring in igneous and metamorphic rocks of the Bohemian Massif, as well as to detrital grains, deposited within the internal basins of the massif. This suggests that the primary rocks for the garnet and tourmaline may be crystalline complexes of the Bohemian Massif. However, other uplifted areas, similar to the complexes of the Bohemian Massif, cannot be ruled out. Such hypothetical areas could be located in the northern foreland of the Carpathian basins. Euhedral tourmaline and other minerals, occurring in the heavy-mineral assemblages studied, most probably were derived from eroded and presently not exposed, crystalline complexes, originally situated in the Skole Basin foreland or within the basin.

Key words: Detrital tourmaline, detrital garnet, provenance, flysch, Menilite Formation, Oligocene, Skole Nappe.

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INTRODUCTION

The analysis of heavy minerals is a widely used tool in provenance reconstruction for ancient and modern, clastic, sedimentary rocks. The process of provenance determination starts with conventional analyses, which yield basic information about the frequency and varieties of the heavy minerals. Nowadays, this step is followed by detailed studies of the chemical composition of minerals, which permits the identification of specific characteristics of the primary rocks in probable source areas.

Garnet and tourmaline are two of the most commonly studied, heavy minerals, in terms of chemical composition for the determination of provenance. Both garnet and tourmaline are very common in ancient sediments, since they are stable and resistant during transport, weathering and diagenesis (e.g., Morton and Hallsworth, 1999, 2007). Because they are sensitive to P-T conditions, oxygen fugacity and the chemical conditions controlling their growth in the parent rocks, their composition is of great importance to the determination of the petrographic features of the primary source rocks (see Mange and Wright, 2007).

The heavy-mineral associations of the Oligocene sandstones of the Menilite Formation (northern Polish part of the Skole Nappe) were described briefly by Tokarski (1947) and Szczurowska (1970, 1971, 1973) and recently in detail by Salata and Uchman (2012). The recent, conventional analyses of heavy minerals revealed that the immediate source area for the Oligocene sandstones probably was composed mainly of sedimentary rocks, such as sandstones and conglomerates. The most likely source rocks appear to have been the Carboniferous or younger clastics that form the Małopolska and the Upper Silesia blocks, although the possibility of metasediments cannot be excluded (Salata and Uchman, 2012). Such a hypothesis is supported by comparison with the heavy-mineral data, obtained for Carboniferous sandstones of the Małopolska and Upper Silesia blocks (Turnau-Morawska and Łydka, 1954; Łydka, 1955; Kryszowska *et al.*, 1960; Siedlecka and Krysowska, 1962; Zachuń, 1996), and the presence of Carboniferous coal fragments in sandstones of Kliva type in the Menilite Formation (Kotlarczyk and Śliwowa, 1963).

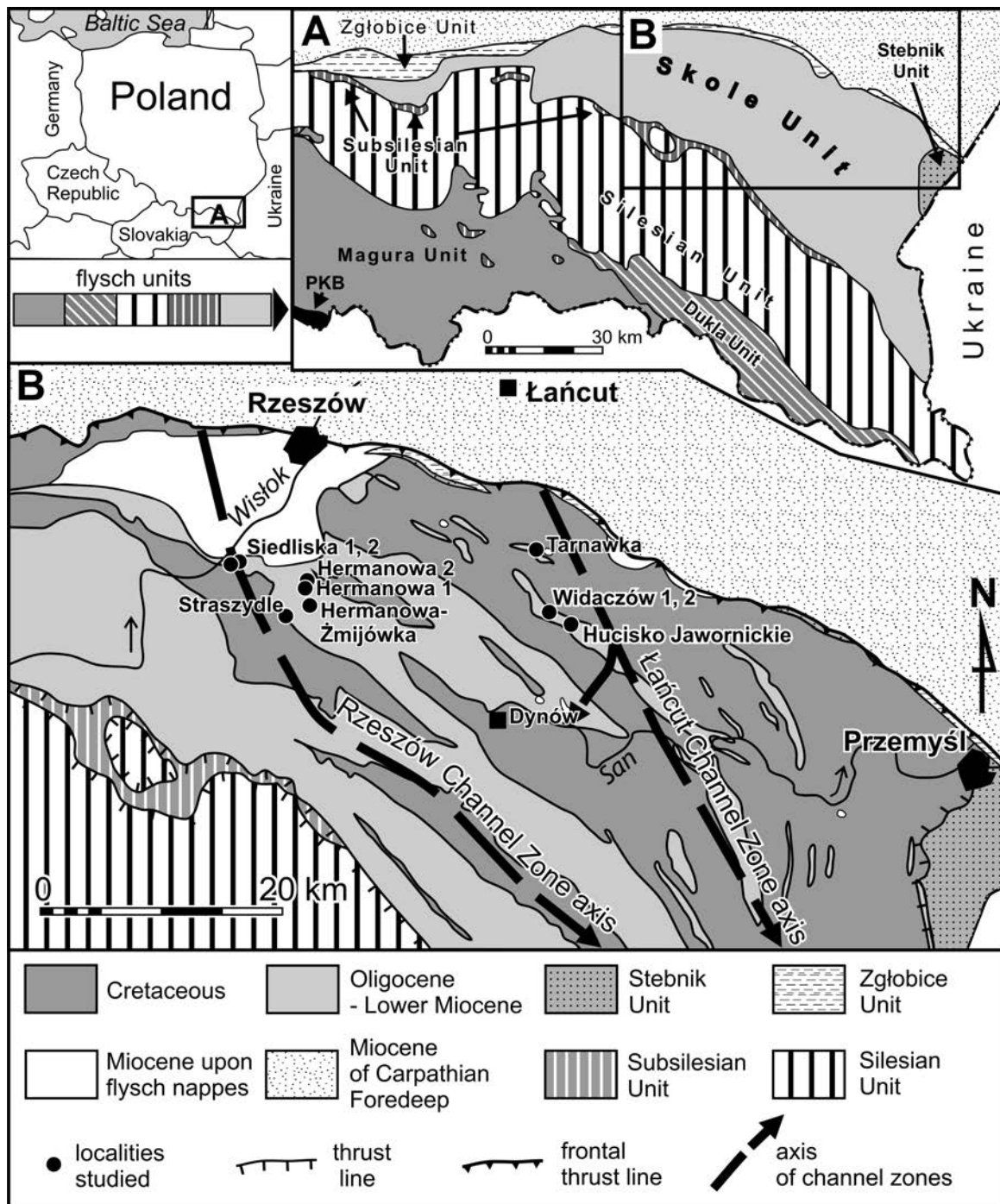


Fig. 1. A – Main structural units in eastern part of Polish Flysch Carpathians, with location of study area (based on Żelaźniewicz *et al.*, 2011), PKB – Pieniny Klippen Belt; B – Geological sketch map of study area, with sampling locations (modified from Salata and Uchman, 2012, based on Kotlarczyk and Leśniak, 1990)

The Upper Silesia and Małopolska blocks may be the most recent, but not necessarily the only source areas for the Oligocene of the Skole Basin. Thus, the determination of the source rocks of the heavy minerals studied and their original location is problematic. The mineral composition of the heavy-mineral assemblages, especially the presence of garnet, kyanite, and staurolite and the frequencies of these minerals (Salata and Uchman, 2012), indicate that they originated from medium- to high-grade, metamorphic rocks. However, because the garnet and tourmaline also could

have originated from igneous rocks, detailed chemical analyses of the minerals are required for a more specific provenance determination.

This work follows and builds upon the recent study on heavy minerals (Salata and Uchman, 2012), and includes chemical analyses of garnet and tourmaline. On the basis of the chemical compositions of these two mineral groups, the present work provides more specific information about the petrographic types of source rocks. Comparison of the analytical results with the available, published data permits fur-

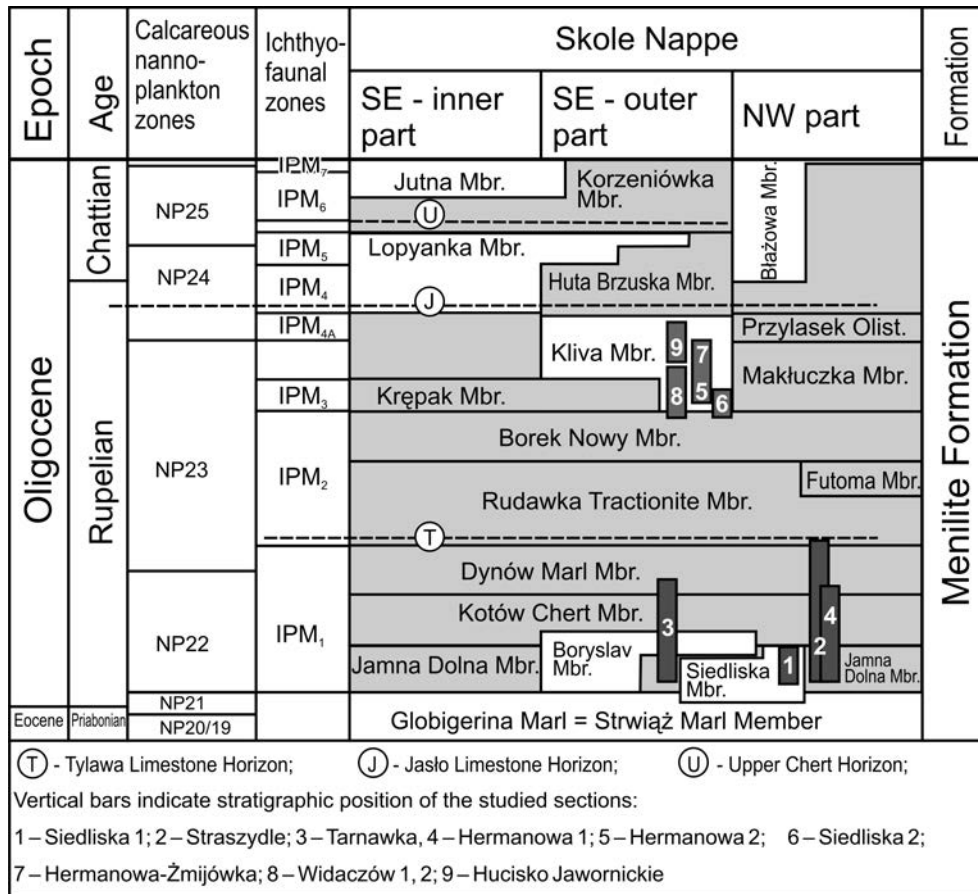


Fig. 2. Stratigraphic scheme of Menilite Formation, Skole Nappe, in Oligocene (modified from Salata and Uchman 2012, based on Kotlarczyk *et al.*, 2006), with stratigraphic positions of sections studied

ther constraints on the designation of possible locations of one or more primary source areas for the garnet and tourmaline studied.

GEOLOGICAL BACKGROUND AND SAMPLE SETTING

The Skole Nappe is the most northerly and also the lowest, structural unit of the Middle Group (Żelaźniewicz *et al.*, 2011 and references therein) in the Polish part of the Western Outer Carpathians (Fig. 1A). The nappe extends from the Polish-Ukrainian border to the Brzesko area. The northern margin of the Skole Basin, initially at least 150 km wide (Gałała *et al.*, 2012), was also the southern edge of the European Platform. Since the Outer Carpathian nappes are overthrust over the Carpathian Foredeep, the edge of the platform currently is not exposed, thus preventing access for direct investigation.

The palaeotransport directions indicate that the northern source area for sediments, up to lower parts of Lower Miocene, is generally related to the sedimentary cover of the Northern European Platform (e.g., Książkiewicz, 1962; Kotlarczyk, 1966, 1976; Ślęczka and Unrug, 1966; Malata and Poprawa, 2006). However, the position of the main channel zones, particularly in the study area (Fig. 1B), indicates transport towards the S and SE (Kotlarczyk and Leś-

niak, 1990). The possible, initial location of the Skole Basin (e.g., Golonka *et al.*, 2006), situated far to the S of the present-day position of the frontal Carpathian overthrust, and the palaeogeography before the Carpathian rotation, suggest that the most probable source areas were the southern margins of the Upper Silesia and Małopolska blocks. According to Kotlarczyk and Śliwowa (1963), the foreland scarps evolved from initially steep to gentle during the Oligocene.

Sediments, representing the time interval from Lower Cretaceous to Lower Miocene, make up the flysch sequence of the Skole Nappe (Figs. 1B, 2). The Oligocene part is composed of the Menilite Formation (Fig. 2). The formation is the most characteristic succession in the Skole Nappe, consisting of dark-brown shales with hornstones, deposited on the periodically anoxic floor of the deep-sea flysch basin (Kotlarczyk and Uchman, 2012). According to Malata and Poprawa (2006), during the accumulation of the Menilite-Krosno series, the sedimentation rate increased from initially about 25–35 m/Ma to over 100 m/Ma, resulting in a thickness of a few hundred metres for the Menilite Formation in the Skole Nappe. Clastic material, transported to northern parts of the Skole Basin, is represented mainly by the Boryslav Sandstone (Tołwiński, 1917), the Kliva Sandstone and the Błażowa Sandstone members; these are the lower, middle and upper parts of the Menilite Formation, respectively (Fig. 2). Kliva-type sandstones are present also in the Siedliska Member (Fig. 2). The Siedliska Conglomerate

(Kropaczek, 1917a, b) is limited mainly to an area to the S and SE of Rzeszów (Kotlarczyk and Leśniak, 1990). The sediments of the Boryslav Sandstone and Kliva Sandstone members accumulated mainly from gravitational flows in channel zones SE of Rzeszów and Łańcut (see Fig. 1). Nevertheless, sediments of the Kliva Sandstone Member also may be found outside the main channel zones (Kotlarczyk and Leśniak, 1990).

The Siedliska Sandstone Member is up to 70 m thick, and composed mainly of thick and medium beds of grey conglomerates, sandstones, and calcareous mudstones, rich in organic material, interbedded with brown mudstones (Kropaczek, 1917a, b; Blaicher and Nowak, 1963; Kotlarczyk, 1985, 1991). The Boryslav Sandstone Member, coeval with the Siedliska Sandstone Member in its lower part, is represented mainly by 20–30 m of poorly cemented quartz (locally glauconitic) sandstones, with locally occurring beds of silicified mudstones. The sandstones are typically brown, light grey or yellowish, thick-bedded and fine- to medium-grained (Totwiński, 1917; Świdziński, 1947; Kotlarczyk and Leśniak, 1990). The Kliva Sandstone Member is similar to the Boryslav Sandstone Member in lithology. It is dominated by white or yellowish, poorly cemented, well-sorted to conglomeratic, thick and very thick-bedded quartz arenites. They are mostly fine- to medium-grained sandstones, containing quartz, feldspar, muscovite, glauconite, lithoclasts and pebbles of coal, black cherts and quartzitic sandstones (Żgiet, 1963; Kotlarczyk, 1966, 1976; Ślącza and Unrug, 1966).

Localities, sampled for the analysis of heavy-mineral frequencies, are situated SE of Rzeszów and Łańcut, and distributed within the two main channel zones (Fig. 1). The sandstone samples represent the Siedliska Sandstone Member, the Boryslav Sandstone Member and the Kliva Sandstone Member. Additionally, some samples were collected from two sandstone beds, close to the top of the Kotów Chert Member and at the base of the Dynów Marl Member, as well as from sandstone beds, interbedded with the shales of the Rudawka Tractionite Member and in the lower part of the Menilite Formation in the Hermanowa 1 section (Figs 1, 2; for details of sample locations, see Salata and Uchman, 2012). The mineral frequencies were investigated in 47 samples.

ANALYTICAL PROCEDURE

The sandstone samples represent weakly or very weakly consolidated rocks. Accordingly, the procedures of heavy-mineral separation included gentle crushing (when needed) and rinsing with water to clean off the clay fraction, followed by sieving to obtain the 63–250 μm fraction. The heavy minerals were separated, using sodium polytungstate with a density of 2.9 g/cm^3 . The heavy-mineral assemblages recovered were mounted in Canada balsam, and described and counted, using a polarising optical microscope. In each sample, 200 to 300 grains of transparent, non-micaceous minerals were counted, according to the ribbon method (Galehouse, 1971). Mineral frequencies were calculated as volume percentages (Salata and Uchman, 2012).

Observations in the microfield and chemical analyses of mineral inclusions in garnet and tourmaline grains were performed, using a HITACHI S-4700 Field Emission Scanning Electron Microscope in the Laboratory of Field Emission Scanning Electron Microscopy and Microanalysis, at the Institute of Geological Sciences, Jagiellonian University.

The chemical composition of the garnets and tourmalines in carbon-coated, polished thin sections was determined, using a Cameca SX-100 electron microprobe (EMP), operated in a wavelength dispersion (WDS) mode, at the Joint-Institute Analytical Complex for Minerals and Synthetic Substances of Warsaw University. The WDS analytical conditions were as follows: 15 kV accelerating voltage, 20 nA beam current and a focused beam. The following, synthetic and natural mineral standards were used for calibration: Si (wollastonite for tourmaline and diopside for garnet analyses), Al and K (orthoclase), Cr (Cr_2O_3), Ti (rutile), Mg (diopside), Fe (Fe_2O_3), Mn (rhodonite), Ca (wollastonite), Na (albite), F (phlogopite).

Garnet and tourmaline grains were chosen randomly for electron microprobe analyses. Single-spot analyses were performed in 80 garnet grains, 1 spot per grain. Twenty grains, representing different compositions, were analysed in traverses to investigate garnet chemical zonation. Twenty grains of tourmaline were analysed in traverses, revealing internal, chemical heterogeneity, but only in grains that displayed dark and light zones in the high-contrast back-scattered electrons (BSE) imaging. Consequently, the homogeneous grains in BSE imaging were analysed in the mode of two spots per grain. A total of 91 tourmaline grains was analysed.

Boron was not measured in the tourmalines, but three atoms of boron were assumed to be present in the tourmaline formula. Therefore, the weight percentages of B_2O_3 , necessary to produce 3 boron atoms, were calculated, assuming $\text{OH} + \text{F} = 4 \text{ apfu}$ (atoms per formula unit), and that all iron is Fe^{2+} . The calculations were normalised to 31 anions. Although the exact contents of B, Li and H could not be measured and consequently the calculated cation amounts are not precise, the results are sufficiently accurate for basic provenance interpretations (e.g., Henry and Guidotti, 1985; Henry *et al.*, 2011).

RESULTS

Mineral composition

Zircon, tourmaline, rutile, staurolite, garnet and Al_2SiO_5 polymorphs are the main constituents of the heavy-mineral assemblages studied. Zircon comprises 6–33% (% = frequency percents; mean value, $\text{mv} = 18\%$), tourmaline 8–35% ($\text{mv} = 19\%$), rutile 5–47% ($\text{mv} = 20\%$), staurolite 4–27% ($\text{mv} = 17\%$), garnet 0–25% and Al_2SiO_5 polymorphs 5–27% ($\text{mv} = 15\%$). The latter are represented mainly by kyanite (4–27%; $\text{mv} = 14\%$), which was identified on the basis of its platy habit and perfect cleavage. Pleochroic andalusite and fibrous sillimanite are scarce. Additionally, single grains of apatite, epidote, brookite, chrome spinel and very rare monazite also occur in some samples. Frequencies of heavy minerals fluctuate between samples and channel

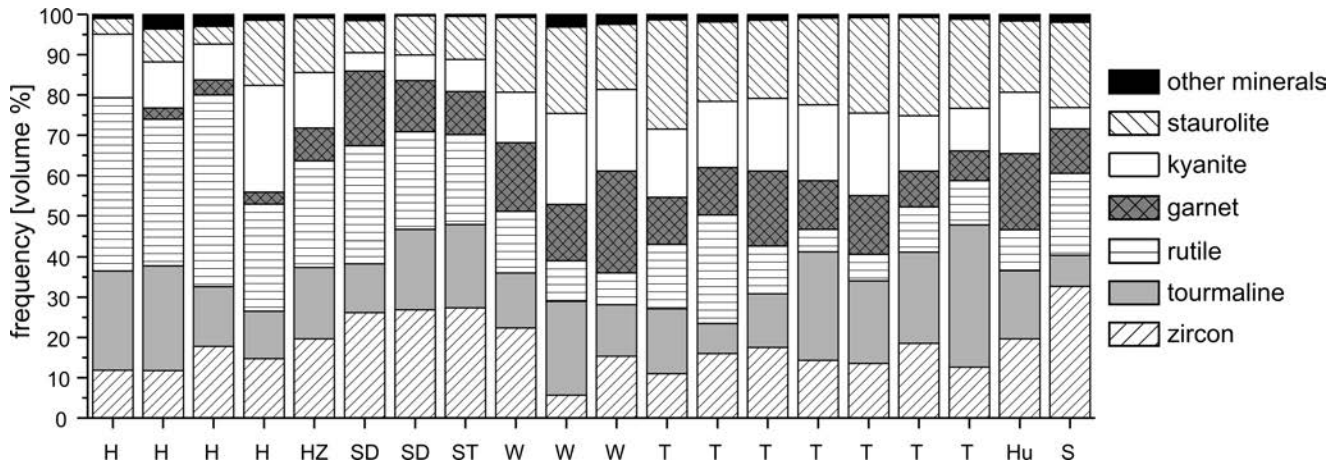


Fig. 3. Frequencies of heavy minerals in representative samples of Oligocene sandstones of Menilite Formation in Skole Nappe. Other mineral field includes andalusite, epidote, apatite, chrome, spinel, sillimanite, brookite and monazite (Salata and Uchman, 2012). Abbreviations for sample locations: H – Hermanowa; HŻ – Hermanowa Żmijówka; SD – Siedliska 2; W – Widaczów; T – Tarnawka; Hu – Hucisko Jawornickie; S – Siedliska

zones and there is no obvious trend. Small fluctuations in mineral frequencies are not regular and related to the sandstone fractions and hydraulic behaviour of heavy minerals, rather than to any variation in source-rock composition (for mineral frequencies in representative samples see Fig. 3; for details see Salata and Uchman, 2012). The ZTR (zircon + tourmaline + rutile) index of the assemblages varies in a very broad range of 36–80% ($mv = 57$).

Garnet

The smallest amounts of garnet were noted in the very fine-grained sandstones of the Hermanowa 1 section, where

its low frequency is most probably related to the sand fraction. Garnet occurs mainly as grains with sharp boundaries, probably a result of crushing during transportation, but well rounded grains are also common. For the most part, they have been affected by dissolution processes. Euhedral, not fractured garnet crystals are very rare. Garnet grains, mounted in Canada balsam, are predominantly colourless in transmitted light, but some of them are pinkish, yellowish or salmon-pink in colour. Scarce green grains also occur (Salata and Uchman, 2012).

Three main compositional varieties of the garnet were distinguished (Fig. 4; Tab. 1):

- 1) almandine-dominated $Prp_{4-19}Alm_{58-87}Sps_{0-18}Grs_{0-18}$

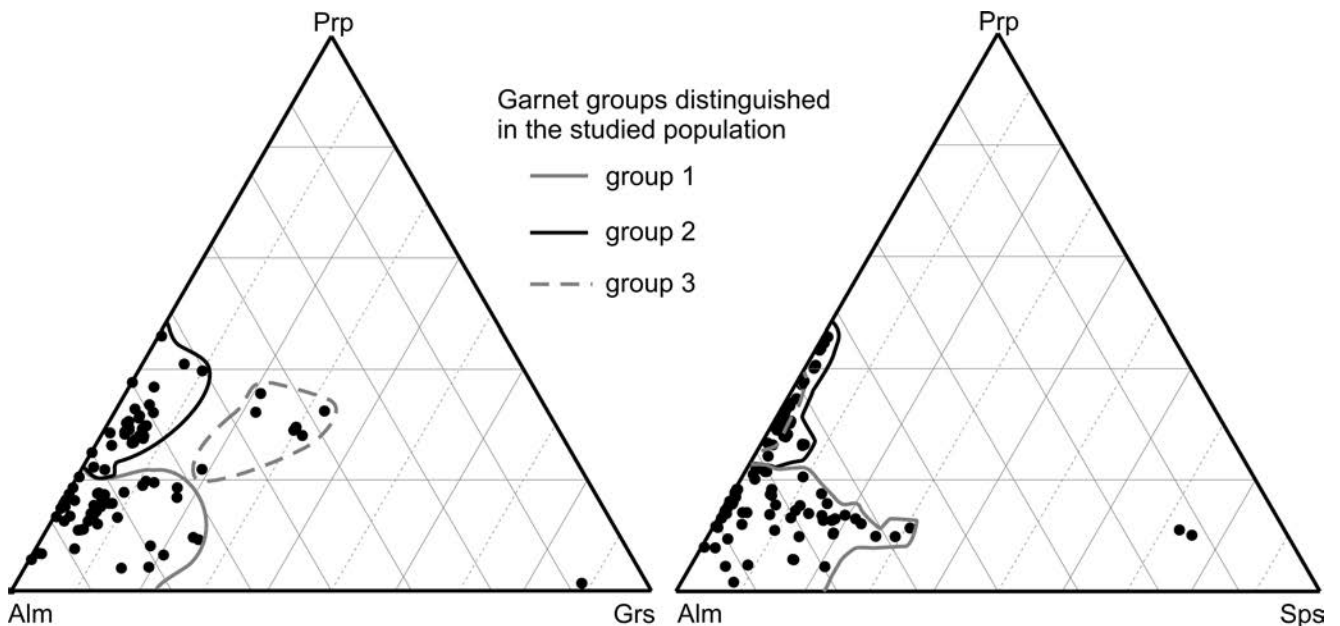


Fig. 4. Composition of detrital garnets ($n = 100$) from Oligocene sandstones of Menilite Formation, Skole Nappe. Abbreviations: Alm – almandine; Grs – grossular; Prp – pyrope; Sps – spessartine

Table 1

Representative analyses of detrital garnets from Oligocene sandstones of Menilite Formation in Skole Nappe.
Oxides in [wt%]

| Analysis | M_22 | M_6 | M_G6 p.8 | M_G6 p.15 | M_24 | M_G3 p.2 | M_G3 p.8 | M_6 | M_G13 p.5 | M_G13 p.9 | M_26 | M_60 | M_41 | M_21 | M_10 | M_32 | M_26 |
|--|--------|--------|-------------|--------------|--------|-------------|-------------|--------|--------------|--------------|--------|-------|--------|-------|--------|--------|--------|
| SiO ₂ | 36.72 | 37.80 | 37.24 | 37.34 | 36.76 | 37.13 | 36.76 | 37.65 | 39.21 | 39.42 | 37.48 | 38.32 | 39.77 | 39.31 | 38.72 | 37.03 | 33.86 |
| TiO ₂ | 0.02 | 0.12 | 0.01 | b.d.l. | 0.07 | 0.01 | 0.03 | 0.09 | 0.05 | 0.03 | 0.12 | 0.17 | 0.07 | 0.20 | 0.33 | 0.22 | b.d.l. |
| Al ₂ O ₃ | 21.04 | 21.77 | 20.31 | 20.47 | 21.07 | 20.20 | 20.28 | 21.61 | 21.51 | 21.52 | 21.35 | 20.52 | 22.67 | 21.36 | 19.90 | 20.62 | 19.68 |
| Cr ₂ O ₃ | b.d.l. | 0.08 | b.d.l. | 0.02 | 0.01 | b.d.l. | b.d.l. | 0.02 | b.d.l. | 0.06 | 0.04 | 0.01 | 0.07 | 0.01 | b.d.l. | 0.03 | 4.83 |
| Fe ₂ O ₃ | 0.00 | 0.00 | 1.30 | 1.00 | 0.00 | 1.26 | 0.98 | 0.00 | 0.99 | 1.12 | 0.08 | 1.62 | 0.00 | 1.13 | 3.07 | 1.01 | 0.00 |
| FeO | 38.28 | 28.14 | 32.99 | 32.46 | 31.90 | 27.85 | 27.87 | 29.85 | 24.39 | 24.30 | 33.52 | 25.61 | 20.32 | 18.28 | 3.74 | 5.77 | 31.55 |
| MgO | 1.83 | 3.19 | 3.70 | 4.08 | 0.95 | 2.33 | 2.38 | 2.37 | 11.28 | 11.61 | 5.46 | 5.27 | 9.30 | 7.00 | 0.03 | 2.15 | 3.04 |
| CaO | 2.00 | 2.09 | 1.50 | 1.54 | 6.20 | 0.89 | 0.89 | 8.11 | 1.54 | 1.36 | 1.38 | 7.99 | 7.73 | 12.16 | 33.46 | 5.39 | 0.83 |
| MnO | 0.12 | 7.98 | 3.22 | 2.86 | 3.16 | 11.15 | 11.11 | 1.45 | 0.40 | 0.36 | 1.03 | 0.34 | 0.27 | 0.43 | 0.34 | 29.32 | 3.09 |
| Total | 100.01 | 101.16 | 100.26 | 99.76 | 100.13 | 100.81 | 100.29 | 101.14 | 99.37 | 99.79 | 100.46 | 99.84 | 100.20 | 99.88 | 99.59 | 101.53 | 96.87 |
| Number of cations calculated on the basis of 12 oxygen atoms | | | | | | | | | | | | | | | | | |
| Si | 2.984 | 2.991 | 2.993 | 3.001 | 2.975 | 2.999 | 2.986 | 2.971 | 3.000 | 3.000 | 2.968 | 3.002 | 2.998 | 3.003 | 2.981 | 2.949 | 2.848 |
| Ti | 0.002 | 0.007 | 0.001 | 0.000 | 0.005 | 0.001 | 0.002 | 0.005 | 0.003 | 0.002 | 0.007 | 0.010 | 0.004 | 0.011 | 0.019 | 0.013 | 0.001 |
| Al | 2.016 | 2.030 | 1.923 | 1.939 | 2.009 | 1.923 | 1.941 | 2.010 | 1.940 | 1.930 | 1.992 | 1.894 | 2.014 | 1.923 | 1.806 | 1.935 | 1.951 |
| Cr | 0.000 | 0.005 | 0.000 | 0.001 | 0.001 | 0.000 | 0.000 | 0.001 | 0.000 | 0.004 | 0.003 | 0.001 | 0.004 | 0.001 | 0.000 | 0.002 | 0.321 |
| Fe ³⁺ | 0.000 | 0.000 | 0.079 | 0.061 | 0.000 | 0.077 | 0.060 | 0.000 | 0.057 | 0.064 | 0.005 | 0.096 | 0.000 | 0.065 | 0.178 | 0.061 | 0.000 |
| Fe ²⁺ | 2.602 | 1.863 | 2.217 | 2.182 | 2.159 | 1.882 | 1.894 | 1.970 | 1.561 | 1.547 | 2.220 | 1.678 | 1.281 | 1.167 | 0.241 | 0.385 | 2.219 |
| Mg | 0.221 | 0.377 | 0.443 | 0.488 | 0.115 | 0.280 | 0.288 | 0.279 | 1.287 | 1.318 | 0.645 | 0.615 | 1.046 | 0.798 | 0.004 | 0.256 | 0.381 |
| Ca | 0.175 | 0.177 | 0.129 | 0.133 | 0.538 | 0.077 | 0.077 | 0.686 | 0.126 | 0.111 | 0.118 | 0.671 | 0.625 | 0.996 | 2.760 | 0.460 | 0.075 |
| Mn | 0.009 | 0.535 | 0.219 | 0.195 | 0.217 | 0.763 | 0.764 | 0.097 | 0.026 | 0.023 | 0.069 | 0.023 | 0.017 | 0.028 | 0.022 | 1.978 | 0.220 |
| Total | 8.007 | 7.984 | 8.002 | 7.998 | 8.016 | 8.001 | 8.011 | 8.019 | 7.999 | 7.998 | 8.026 | 7.987 | 7.988 | 7.990 | 8.009 | 8.037 | 8.016 |
| End-members [mol%] | | | | | | | | | | | | | | | | | |
| prp | 7.4 | 12.8 | 14.7 | 16.3 | 3.8 | 9.2 | 9.5 | 9.2 | 42.9 | 43.9 | 21.1 | 20.6 | 35.2 | 26.7 | 0.1 | 8.3 | 11.7 |
| alm | 86.6 | 63.1 | 73.7 | 72.8 | 71.3 | 61.9 | 62.4 | 65.0 | 52.0 | 51.6 | 72.8 | 56.2 | 43.2 | 39.1 | 7.9 | 12.5 | 68.1 |
| sps | 0.3 | 18.1 | 7.3 | 6.5 | 7.2 | 25.1 | 25.2 | 3.2 | 0.9 | 0.8 | 2.3 | 0.8 | 0.6 | 0.9 | 0.7 | 64.3 | 6.8 |
| grs | 5.8 | 5.8 | 0.3 | 1.3 | 17.7 | 0.0 | 0.0 | 22.6 | 1.3 | 0.3 | 3.5 | 17.6 | 20.8 | 30.0 | 82.2 | 11.8 | 0.0 |
| adr | 0.0 | 0.0 | 3.9 | 3.0 | 0.0 | 3.8 | 3.0 | 0.0 | 2.9 | 3.2 | 0.3 | 4.8 | 0.0 | 3.2 | 9.0 | 3.1 | 0.0 |
| uv | 0.0 | 0.2 | 0.0 | 0.1 | 0.1 | 0.0 | 0.0 | 0.1 | 0.0 | 0.2 | 0.1 | 0.1 | 0.2 | 0.1 | 0.0 | 0.1 | 13.4 |
| Grt gr. ¹ | 1 | 1 | 1 | 1 | 1 | 1 | 1 | 1 | 2 | 2 | 2 | 3 | 3 | 3 | - | - | - |

¹ Grt gr. – main garnet groups distinguished in the studied garnet population (see text); b.d.l. – below detection limit; prp – pyrope; alm – almandine; sps – spessartine; grs – grossular; adr – andradite; uv – uvarovite. Structural formula based on 12 oxygen atoms and 8 cations. The Fe²⁺ and Fe³⁺ contents determined assuming garnet stoichiometry.

comprising 60% of the analysed population. Also assigned to this group is garnet, enriched in spessartine or grossular end-members, with the composition Prp₄₋₁₂Alm₅₆₋₇₁Sps₂₀₋₂₉Grs₀₋₁ or Prp₆₋₉Alm₆₁₋₇₂Sps₂₋₈Grs₂₀₋₂₃ respectively;

2) pyrope-enriched almandine Prp₂₁₋₄₄Alm₅₀₋₇₃Sps₁₋₆Grs₀₋₁₀ (33% of the population);

3) pyrope-almandine-grossular showing composition of Prp₂₇₋₃₅Alm₃₉₋₄₄Sps₁Grs₁₈₋₃₀ (4% of the population).

In addition, single grains of grossular (Prp₀Alm₈Sps₁Grs₈₂), spessartine (Prp₈₋₉Alm₁₃₋₁₄Sps₆₃₋₆₄Grs₁₀₋₁₂) (Fig. 4; Tab. 1) and uvarovite-enriched almandine (Prp₁₂Alm₆₈Sps₇Grs₀UV₁₃) have been also found (Tab. 1). The garnet grains are chemically homogenous (Fig. 5A–C; Tab. 1), but it should be noted that the detrital garnet population represents a mixture of small, single crystals, as well as pieces of larger ones, which in fact may represent fragments of initially zoned, large garnet crystals.

Tourmaline

Tourmaline is present commonly as rounded grains, with mainly olive-brown pleochroic colours, while euhedral tourmaline crystals are less frequent. Occasionally, irregularly coloured, rounded pinkish and blue tourmalines were found, but they, especially the two latter, are very rare and occur as single grains only in some samples (Salata and Uchman, 2012).

The X-site in the tourmaline structure is dominantly occupied by Na, in most cases exceeding 0.6 *apfu*, while the Ca and K contents are commonly below 0.1 and 0.02 *apfu*, respectively. The X-site vacancy in the largest part of the tourmaline population is below 0.5, although it exceeds 0.5 in two analyses (Tab. 2). The chemistry indicates that the tourmalines belong to the alkali group; only the two grains with X-site vacancies higher than 0.5 belong to the X-va-

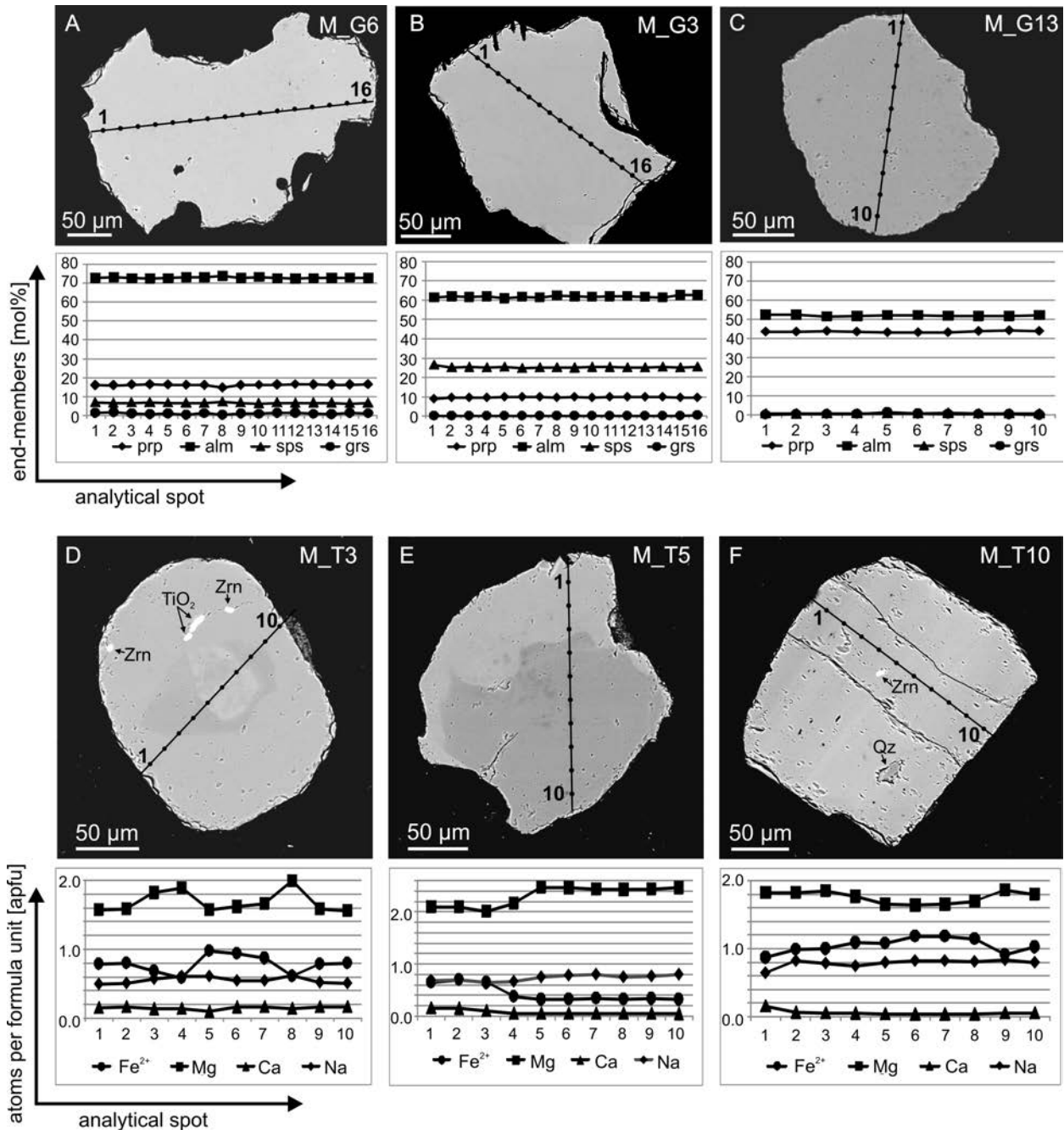


Fig. 5. Representative BSE images and compositional profiles in garnet (A–C) and tourmaline (D–F). Representative chemical analyses of these minerals are presented in Tables 1 and 2. Abbreviations: Qz – quartz; TiO₂ – TiO₂ polymorph; Zrn – zircon

cant group (see Henry *et al.*, 2011) (Fig. 6A). Because the dominating divalent cations are Fe and Mg, the tourmaline displays the composition of schörl or dravite, respectively. Sporadic, Fe-rich tourmaline, characterized by an X-site vacancy of over 0.5, represents the foititic tourmaline species (Tab. 2; Fig. 6B). The Mg/(Fe+Mg) ratio varies considerably in a broad range from 0.03 to 0.78 in homogenous tourmalines and reaches 0.88 in zonal ones (Tab. 2). All the tourmaline grains display a Si content close to the stoichiometric value of 6 *apfu*, and high Al amounts, exceeding 6 *apfu* in almost all cases (Tab. 2). The remaining components are present in minor amounts: the Ti content is mostly lower than 0.1 *apfu*, increasing up to 0.2 *apfu* only in scarce

grains, while Mn and Cr occur mostly in trace amounts of < 0.03 *apfu* (Tab. 2).

The tourmaline is mostly homogenous, although chemical zonation, related to varying Fe and Mg contents, was observed. Zones, visible as bright in BSE imaging, are enriched in Fe (and in Ca in some cases) and depleted in Mg, while in the dark zones, the relationship of these elements is reversed (see Fig. 5D–F). Such chemical variations usually do not influence affiliation to a mineral species (Tab. 2). The tourmalines typically contain quartz inclusions. Dravitic tourmaline, beside quartz inclusions, contains also the TiO₂ polymorph + zircon (Fig. 5D), zircon (Fig. 5F) and the TiO₂ polymorph + monazite or pyrite.

Table 2

Representative analyses of compositionally homogenous and heterogeneous, detrital tourmalines from Oligocene sandstones of Menilite Formation in Skole Nappe. Oxides in [wt%]

| Analysis | M_29 | M_6 | M_19 | M_16 | M_15 | M_25 | M_1 | M_3 | M_27 | M_2 | M_26 | M_170 |
|---|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|
| SiO ₂ | 37.08 | 34.38 | 34.91 | 37.32 | 36.27 | 37.20 | 36.06 | 36.75 | 35.65 | 36.21 | 36.89 | 36.05 |
| TiO ₂ | 0.39 | 0.30 | 0.54 | 0.82 | 0.20 | 0.82 | 0.37 | 0.82 | 1.29 | 0.43 | 1.79 | 0.23 |
| B ₂ O ₃ ¹ | 10.75 | 10.34 | 10.37 | 10.91 | 10.68 | 11.02 | 10.66 | 10.88 | 10.28 | 10.66 | 10.68 | 10.73 |
| Al ₂ O ₃ | 31.71 | 33.61 | 33.80 | 32.17 | 35.97 | 35.23 | 32.40 | 34.26 | 26.60 | 31.81 | 30.71 | 36.35 |
| Cr ₂ O ₃ | 0.02 | 0.02 | 0.13 | 0.22 | 0.11 | 0.12 | 0.57 | 0.13 | 0.03 | 0.06 | b.d.l. | b.d.l. |
| MgO | 8.92 | 0.28 | 0.70 | 8.92 | 1.43 | 6.27 | 7.43 | 6.68 | 6.35 | 6.63 | 5.96 | 2.78 |
| CaO | 0.81 | 0.22 | 0.02 | 0.59 | 0.07 | 0.28 | 1.09 | 0.35 | 1.57 | 1.28 | 0.39 | 0.17 |
| MnO | 0.01 | 0.22 | 0.12 | 0.02 | 0.09 | b.d.l. | b.d.l. | 0.01 | 0.13 | 0.03 | 0.05 | 0.04 |
| FeO _{tot} | 4.68 | 16.54 | 14.34 | 4.50 | 11.85 | 6.06 | 6.06 | 5.86 | 13.57 | 9.01 | 9.13 | 9.47 |
| Na ₂ O | 2.32 | 2.18 | 1.93 | 2.60 | 1.29 | 2.02 | 1.88 | 2.16 | 2.07 | 1.86 | 2.25 | 1.60 |
| K ₂ O | 0.01 | 0.07 | 0.03 | 0.03 | b.d.l. | 0.01 | 0.07 | 0.01 | 0.02 | 0.03 | b.d.l. | 0.01 |
| F | 0.03 | 0.12 | 0.34 | 0.05 | 0.19 | 0.12 | 0.10 | 0.10 | 0.10 | 0.08 | 0.16 | 0.05 |
| Total | 96.73 | 98.26 | 97.22 | 98.14 | 98.23 | 99.14 | 96.67 | 98.01 | 97.63 | 98.08 | 98.00 | 97.46 |
| O = F | 0.01 | 0.05 | 0.14 | 0.02 | 0.08 | 0.05 | 0.04 | 0.04 | 0.04 | 0.03 | 0.07 | 0.02 |
| TOTAL | 96.71 | 98.21 | 97.07 | 98.12 | 98.15 | 99.09 | 96.63 | 97.96 | 97.59 | 98.04 | 97.94 | 97.43 |
| Numbers of ions based on 31 anions (O, OH, F) | | | | | | | | | | | | |
| Si | 5.993 | 5.779 | 5.853 | 5.946 | 5.901 | 5.866 | 5.881 | 5.870 | 6.028 | 5.906 | 6.004 | 5.839 |
| Ti | 0.048 | 0.038 | 0.069 | 0.098 | 0.024 | 0.097 | 0.045 | 0.099 | 0.164 | 0.052 | 0.219 | 0.028 |
| 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 |
| Al | 6.041 | 6.658 | 6.680 | 6.040 | 6.898 | 6.548 | 6.228 | 6.449 | 5.300 | 6.113 | 5.891 | 6.939 |
| Cr | 0.002 | 0.003 | 0.017 | 0.027 | 0.014 | 0.015 | 0.074 | 0.016 | 0.004 | 0.008 | 0.000 | 0.000 |
| Mg | 2.150 | 0.069 | 0.174 | 2.119 | 0.346 | 1.474 | 1.805 | 1.592 | 1.600 | 1.612 | 1.447 | 0.671 |
| Ca | 0.140 | 0.039 | 0.003 | 0.101 | 0.013 | 0.047 | 0.190 | 0.061 | 0.284 | 0.223 | 0.068 | 0.029 |
| Mn | 0.002 | 0.031 | 0.017 | 0.002 | 0.012 | 0.000 | 0.000 | 0.002 | 0.018 | 0.004 | 0.007 | 0.005 |
| Fe ²⁺ | 0.632 | 2.325 | 2.011 | 0.600 | 1.612 | 0.799 | 0.826 | 0.782 | 1.918 | 1.229 | 1.242 | 1.283 |
| Na | 0.727 | 0.710 | 0.627 | 0.802 | 0.407 | 0.616 | 0.593 | 0.668 | 0.677 | 0.587 | 0.709 | 0.502 |
| K | 0.002 | 0.014 | 0.007 | 0.007 | 0.000 | 0.001 | 0.014 | 0.001 | 0.003 | 0.006 | 0.001 | 0.002 |
| Total | 18.736 | 18.666 | 18.457 | 18.743 | 18.227 | 18.464 | 18.657 | 18.538 | 18.997 | 18.740 | 18.590 | 18.297 |
| X _{vac} ² | 0.13 | 0.24 | 0.36 | 0.09 | 0.58 | 0.34 | 0.20 | 0.27 | 0.04 | 0.18 | 0.22 | 0.47 |
| X _{Mg} ³ | 0.77 | 0.03 | 0.08 | 0.78 | 0.18 | 0.65 | 0.69 | 0.67 | 0.45 | 0.57 | 0.54 | 0.34 |
| M.N. ⁴ | D | S | S | D | F | D | D | D | S | D | D | S |

DISCUSSION

Constraints on type of source rock

Garnet is an accessory phase in various, metamorphic and igneous rocks. The most common garnet hosts are metapelites and metapsammites, formed under conditions of medium- to high-grade metamorphism (e.g., Miyashiro, 1975; Yardley, 1989; Deer *et al.*, 1997). The almandine-rich garnets from group 1 are most typical for various rocks of amphibolite facies conditions (Figs 7, 8). However, almandine and spessartine-rich almandine also may be derived from granitic to pegmatitic rocks, as well as rhyolites to andesites (e.g., Deer *et al.*, 1997; Fig. 7). Almandine garnet, enriched in the pyrope end-member, representing group 2, may originate from metamorphic rocks of granulite to eclogite facies (Figs 7, 8), but also from intermediate, igneous rocks, derived from deep parts of the crust (Fig. 7).

Pyrope-almandine-grossular of group 3 are common in high-grade metabasites, such as eclogites, and high-pressure granulites (Figs 7, 8). The source rock of one grain of grossular was a skarn, a very low-grade metabasic rock or an ultrahigh-temperature calc-silicate granulite (Fig. 7).

Tourmaline can crystallize during igneous, metamorphic, and, more rarely, diagenetic processes (e.g., Henry and Guidotti, 1985; Henry and Dutrow, 1996; Dutrow and Henry, 2011 and references therein; van Hinsberg *et al.*, 2011a, b). Most of the tourmalines studied display proportions of Al, Fe_{tot} and Mg, typical for crystals, forming in Al-rich and Al-poor metapelites and metapsammites (Henry and Guidotti, 1985; Fig. 9A). The prevailing part of the tourmaline grains analysed display moderate to high Mg content, with a Mg/(Mg+Fe) ratio range of 0.4–0.7 (Tab. 2), which is typical for metapelites (Henry and Dutrow, 1996). Additionally, tourmaline, forming in Al-rich metapelites

Table 2 continued

| Tourmaline grain | M_T3 (Fig. 5D) | | | | M_T5 (Fig. 5E) | | | | M_T10 (Fig. 5F) | | | |
|---|----------------|--------|--------|--------|----------------|--------|--------|--------|-----------------|--------|--------|--------|
| analytical spot | 1 | 4 | 6 | 8 | 1 | 3 | 5 | 8 | 1 | 3 | 6 | 9 |
| SiO ₂ | 36.73 | 37.66 | 37.03 | 37.52 | 37.39 | 37.77 | 38.19 | 38.02 | 36.66 | 37.49 | 36.93 | 39.74 |
| TiO ₂ | 0.83 | 0.64 | 0.88 | 0.67 | 0.80 | 0.82 | 0.24 | 0.28 | 0.73 | 0.92 | 1.03 | 0.94 |
| B ₂ O ₃ ¹ | 10.83 | 10.98 | 10.73 | 10.94 | 10.80 | 10.91 | 10.90 | 10.92 | 10.70 | 10.79 | 10.65 | 11.46 |
| Al ₂ O ₃ | 33.99 | 33.58 | 32.09 | 32.96 | 31.17 | 32.20 | 31.57 | 32.06 | 31.80 | 30.94 | 30.62 | 33.00 |
| Cr ₂ O ₃ | 0.06 | 0.06 | 0.05 | 0.09 | 0.05 | b.d.l. | b.d.l. | 0.05 | 0.08 | 0.07 | 0.02 | 0.05 |
| MgO | 6.57 | 8.02 | 6.70 | 8.41 | 8.76 | 8.50 | 10.40 | 10.25 | 7.54 | 7.71 | 6.75 | 8.26 |
| CaO | 0.89 | 0.80 | 0.91 | 0.81 | 0.97 | 0.69 | 0.39 | 0.35 | 0.94 | 0.33 | 0.24 | 0.35 |
| MnO | b.d.l. | 0.03 | b.d.l. | b.d.l. | b.d.l. | b.d.l. | 0.01 | b.d.l. | 0.07 | b.d.l. | b.d.l. | 0.06 |
| FeO | 5.86 | 4.38 | 6.96 | 4.57 | 4.85 | 4.87 | 2.43 | 2.49 | 6.42 | 7.43 | 8.69 | 7.27 |
| Na ₂ O | 1.57 | 1.98 | 1.74 | 2.03 | 2.22 | 2.10 | 2.42 | 2.44 | 2.04 | 2.54 | 2.59 | 2.87 |
| K ₂ O | 0.01 | b.d.l. | b.d.l. | b.d.l. | 0.05 | 0.02 | b.d.l. | 0.07 | b.d.l. | b.d.l. | b.d.l. | b.d.l. |
| F | b.d.l. | 0.05 | 0.02 | 0.05 | 0.15 | 0.22 | 0.22 | 0.22 | 0.10 | 0.02 | 0.05 | 0.12 |
| Total | 97.33 | 98.17 | 97.11 | 98.04 | 97.19 | 98.10 | 96.75 | 97.14 | 97.06 | 98.24 | 97.58 | 104.13 |
| O = F | 0.00 | 0.02 | 0.01 | 0.02 | 0.06 | 0.09 | 0.09 | 0.09 | 0.04 | 0.01 | 0.02 | 0.05 |
| TOTAL | 97.33 | 98.15 | 97.10 | 98.02 | 97.13 | 98.00 | 96.66 | 97.05 | 97.02 | 98.23 | 97.56 | 104.08 |
| Numbers of ions based on 31 anions (O, OH, F) | | | | | | | | | | | | |
| Si | 5.892 | 5.959 | 5.998 | 5.964 | 6.020 | 6.016 | 6.092 | 6.052 | 5.955 | 6.037 | 6.026 | 6.028 |
| Ti | 0.100 | 0.076 | 0.107 | 0.080 | 0.096 | 0.099 | 0.029 | 0.034 | 0.089 | 0.111 | 0.127 | 0.108 |
| 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 |
| Al | 6.427 | 6.263 | 6.125 | 6.173 | 5.913 | 6.043 | 5.934 | 6.015 | 6.087 | 5.871 | 5.889 | 5.899 |
| Cr | 0.007 | 0.007 | 0.007 | 0.011 | 0.006 | 0.000 | 0.000 | 0.007 | 0.010 | 0.009 | 0.003 | 0.006 |
| Mg | 1.572 | 1.891 | 1.619 | 1.992 | 2.102 | 2.018 | 2.472 | 2.432 | 1.825 | 1.850 | 1.642 | 1.868 |
| Ca | 0.153 | 0.135 | 0.157 | 0.138 | 0.167 | 0.117 | 0.066 | 0.060 | 0.163 | 0.057 | 0.042 | 0.057 |
| Mn | 0.000 | 0.005 | 0.000 | 0.000 | 0.000 | 0.000 | 0.001 | 0.000 | 0.009 | 0.000 | 0.000 | 0.008 |
| Fe ²⁺ | 0.786 | 0.580 | 0.942 | 0.607 | 0.653 | 0.649 | 0.324 | 0.331 | 0.872 | 1.001 | 1.186 | 0.922 |
| Na | 0.489 | 0.607 | 0.546 | 0.625 | 0.691 | 0.648 | 0.749 | 0.752 | 0.642 | 0.793 | 0.820 | 0.843 |
| K | 0.001 | 0.000 | 0.000 | 0.001 | 0.011 | 0.003 | 0.000 | 0.015 | 0.000 | 0.001 | 0.000 | 0.000 |
| Total | 18.428 | 18.524 | 18.501 | 18.591 | 18.661 | 18.594 | 18.668 | 18.696 | 18.651 | 18.730 | 18.734 | 18.739 |
| X _{vac} ² | 0.36 | 0.26 | 0.30 | 0.24 | 0.13 | 0.23 | 0.18 | 0.17 | 0.20 | 0.15 | 0.14 | 0.10 |
| X _{Mg} ³ | 0.67 | 0.77 | 0.63 | 0.77 | 0.76 | 0.76 | 0.88 | 0.88 | 0.68 | 0.65 | 0.58 | 0.67 |
| M.N. ⁴ | D | D | D | D | D | D | D | D | D | D | D | D |

¹ calculated on stoichiometry

² X_{vac} = X-site vacancy

³ Mg/(Mg+Fe)

⁴ Mineral name: D – dravite; S – schörl; F – foitite; b.d.l. – below detection limit

that display X-site vacancies mostly in the range of 0.5–0.2, may be interpreted as having been formed in temperatures of 200–650°C, while tourmaline with X-site vacancy values of < 0.2 crystallized in temperatures above 750°C (Henry and Dutrow, 1996). Since most of the tourmalines from the Menilite Formation show an X-site vacancy > 0.2, it may be assumed that they developed under medium-grade metamorphic conditions. The low Ca content in the tourmaline population, mostly not exceeding 0.2 *apfu*, provides additional information on the Ca-poor type of source metasedimentary rocks (Fig. 9B). Additionally, several dozens of analyses, plotted in the provenance discrimination diagrams, indicate provenance, related to a Li-poor granitoids, pegmatites and aplites (Fig. 9A, B), and only one analysis is

located in the field of Fe³⁺-rich, quartz-tourmaline rocks, calc-silicates and metapelites (Fig. 9A). Tourmaline, formed in Li-poor granitic rocks, generally have a schörlitic composition, with very low Mg/(Fe+Mg) ratios in the range of 0.0–0.4. Nevertheless, tourmaline from metamorphosed, granitic rocks may have similar compositions (Henry and Dutrow, 1996). To conclude, judging on the “environmental” diagrams, the two main, parent rocks for the tourmalines studied are Al-enriched or depleted, Ca-poor metasediments and granitic rocks, depleted in Li. The pink or blue colour of some tourmalines, found among the grains, mounted in Canada balsam suggests an elbaite composition. However, such tourmaline composition was not determined in the population studied.

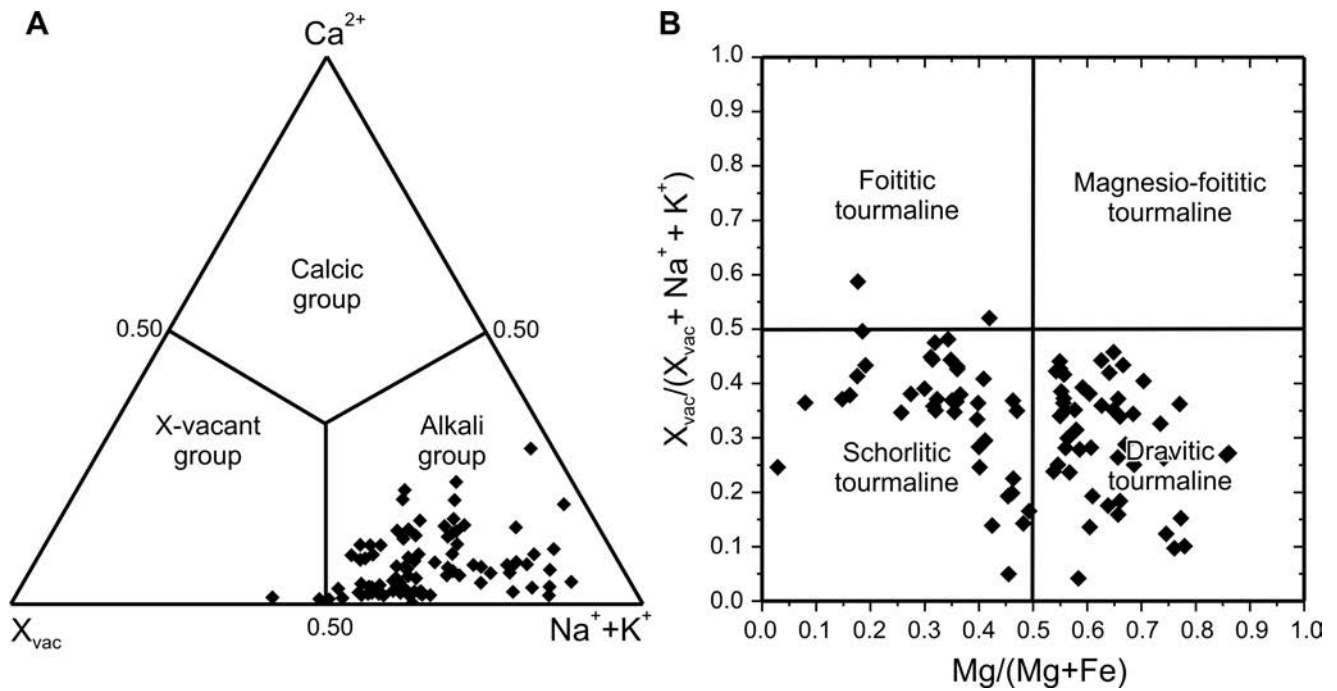


Fig. 6. Classification of detrital tourmalines ($n = 91$) from Oligocene sandstones of Menilite Formation, Skole Nappe: **A** – primary tourmaline groups, based on dominant occupancy of X-site; **B** – generalized tourmaline species, based on $Mg/(Mg + Fe)$ versus $X_{vac}/(X_{vac} + Na + K)$ ratios (Henry *et al.*, 2011)

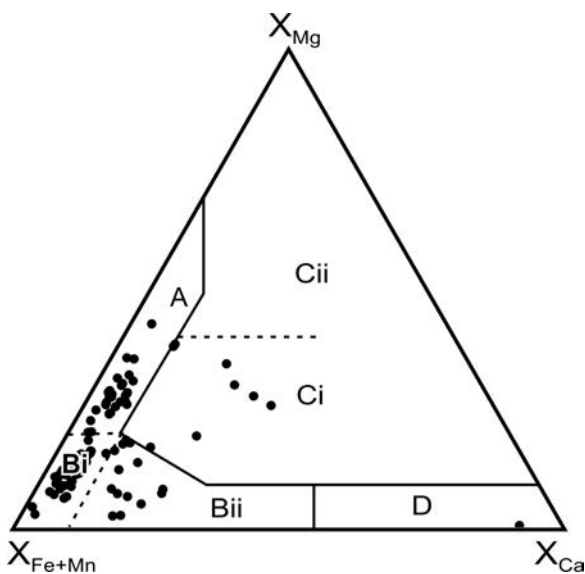


Fig. 7. Composition of garnets of Oligocene sandstones of Menilite Formation in ternary plot, referring to natural, sedimentary garnet concentrations (Mange and Morton, 2007): A – garnets mainly from high-grade granulite-facies metasediments or charnockites, but also from intermediate-acidic, igneous rocks, sourced from deep in crust; Bi and Bii – garnets, derived in general from amphibolite-facies metasediments. However, garnet populations that plot exclusively in Bi field may also be derived from intermediate-acidic, igneous rocks; Ci and Cii – field of garnets mainly from high-grade metabasic rocks, where Cii imply sourcing from ultramafic rocks, such as pyroxenites and peridotites; D – garnets, generally derived from metasomatic rocks, very low-grade metabasic rocks or ultrahigh-temperature metamorphosed calc-silicate granulites. X_{Fe} , X_{Mg} , X_{Ca} , X_{Mn} denote ionic contents of Fe, Mg, Ca and Mn, normalized according to Droop and Harte (1995)

Potential source areas of detrital material

Salata and Uchman (2012), on the basis of data, obtained from conventional heavy-mineral studies, concluded that the immediate source of heavy minerals in the Menilite Formation could be clastic rocks in the sedimentary cover of the Małopolska and Silesian blocks. This idea was supported by the rounded morphology of the dominant part of heavy minerals, suggesting their multicycle history, although their transportation from a distant source or provenance from metasediments also were taken into account. The presence of Carboniferous coal fragments and the similarity of heavy-mineral assemblages led to the conclusion that Carboniferous, clastic rocks of the Małopolska and Silesian blocks are the most likely sources. However, younger, clastic rocks of the blocks, as well as older flysch strata were not excluded (Salata and Uchman, 2012 and references therein). Nonetheless, the low or moderate ZTR index values (around 30–50%) of the heavy-mineral assemblages studied seem to contradict this idea, since the index values of multicycle sediments are much higher, as much as 90 or 100% (Hubert, 1962). The high index value may be modified, if simultaneously with sedimentary rocks, some crystalline massifs were eroded or pebbles of crystalline rocks were disintegrated, yielding minerals, less stable than zircon, tourmaline and rutile. Such a situation might have existed during the Oligocene, as suggested by the presence of sharp, euhedral grains of zircon, tourmaline, platy kyanite and euhedral garnet in the heavy-mineral assemblages of the sandstones studied. Moreover, long transportation of sediments from a source area would not necessary affect first-cycle mineral assemblages. Mechanical abrasion seems to have much less influence on the stability of heavy

Table 3

Literature data of garnet composition from crystalline rocks of Sudetes, clastic rocks, filling internal, sedimentary basins in Bohemian Massif and from sedimentary cover of Upper Silesian Block

| Garnet composition | Host-rock | Locality | References |
|--|--|----------------------------|--------------------------------|
| Prp<52Alm70-75Sps<52Grs4-6 | Upper Carboniferous sandstones | Upper Silesia Coal Basin | Kusiak <i>et al.</i> (1999) |
| Prp10-73Alm14-73Sps1-28Grs3-31 | Middle Jurassic sands and sandstones | Kraków–Wieluń Upland | Méres <i>et al.</i> (2012) |
| Prp1-3Alm63-72Sps1-22Grs6-34 | leucocratic two-mica gneiss | Kamieniec Metamorphic Belt | Puziewicz and Rudolf (1998) |
| Prp5-22Alm69-82Sps1-4Grs5-23 Prp4-8Alm62-77Sps6-18Grs9-18 | coarse-blastic mica-schist fine-blastic mica-schist | | Józefiak (1998) |
| Prp7-11Alm46-79Sps1-19Grs9-39 | two-mica gneiss | | Puziewicz <i>et al.</i> (1999) |
| Prp7-32Alm39-68Sps0-5Grs16-38 | eclogites | | Bakun-Czubarow (1998) |
| Prp6-9Alm44-67Sps23-46Grs1-3 | metamorphic pegmatite | | Pieczka <i>et al.</i> (1997) |
| Prp8-17Alm61-76Sps5-22Grs3-10 Prp31-39Alm49-55Sps2Grs8-12 | gneiss, metapegmatite granulite | Góry Sowie Massif | Budzyń <i>et al.</i> (2004) |
| Prp38Alm40Sps1Grs21 | HP granulite | | O'Brien <i>et al.</i> (1997) |
| Prp17-43Alm37-62Sps1-2Grs8-23 | metabasites | | Dziedzic (1996) |
| Prp17-23Alm44-59Sps1-4Grs19-30 | metagabbro | | Kryza and Pin (2002) |
| Prp0-1Alm39-40Sps57-59Grs1 | pegmatite | | Strzegom–Sobótka Massif |
| Prp1-7Alm45-67Sps1-15Grs19-44 | granite and granito-gneiss | Karkonosze–Iżera Massif | Oberc-Dziedzic (1991) |
| Prp10-71Alm9-58Sps1-60Grs2-24 | Late Cretaceous sandstones | North Sudetic Synclinorium | Biernacka and Józefiak (2009) |
| Prp3-45Alm40-63Sps1-33Grs2-27 | Permian sandstones | Intra-Sudetic Basin | Biernacka (2012a) |
| Prp1-29Alm2-77Sps0-41Grs3-77 | Carboniferous sediments | | Felicka (2000) |
| Prp41-50Alm46-54Sps1-2Grs0-6 | Granulite clasts in Carboniferous sediments | Bohemian Massif | Kotková <i>et al.</i> (2007) |
| Prp5-47Alm50-88Sps0-4Grs0-9 | Carboniferous Culm sediments | | Čopjaková <i>et al.</i> (2005) |

minerals than diagenesis (Morton and Hallsworth, 1999 and references therein).

The provenance information on the petrographic types of source rocks, obtained from chemical analyses of detrital garnet and tourmaline from the Menilite Formation, corresponds well with the petrographic types of pebbles, occurring in the Carboniferous clastic rocks of the Upper Silesia Block. Metamorphic schists, granulites, gneisses and various, granitic rocks, as well as pegmatites, were reported from the Carboniferous sandstones and conglomerates, filling the Upper Silesia Coal Basin (e.g., Turnau-Morawska and Łydko, 1954; Paszkowski *et al.*, 1995 and references therein). Moreover, the chemical composition of the detrital garnet analysed is compatible with the composition of detrital garnet, occurring in the Carboniferous sediments of the Upper Silesia Coal Basin (Kusiak *et al.*, 1999) (Tab. 3). Therefore, the data presented on garnets to some extent support the assumption (Salata and Uchman, 2012) that the heavy-mineral assemblages from the Kliva Sandstone types in the Menilite Formation were derived, *inter alia*, from the Carboniferous cover of the Skole Basin foreland. Additional information is in the recently published data on the chemical composition of detrital garnets from the Middle Jurassic, clastic rocks of the Kraków–Wieluń Upland (Méres *et al.*, 2012). Compositional garnet varieties, determined in the study cited, are to large extent comparable to those, occurring in the Menilite Formation, although the garnet population from the Middle Jurassic rocks is richer in high-pyrope varieties. The similarity of garnet populations indicates that

not only Jurassic clastics may be taken into account as possible source rocks, but also that the garnets of the Menilite Formation may have come from the same source areas.

Considering all of the aspects, mentioned above, the following question arises: where were the crystalline-massif sources of the minerals located? Some clues are provided by palaeotransport directions, measured in the Carboniferous sediments of the Upper Silesia Block (e.g., Gradziński *et al.*, 1961; Doktor and Gradziński, 1998). These suggest areas, located on the SW, NW, N (Gradziński *et al.*, 1961) and S (Doktor and Gradziński, 1998) margins of the Upper Silesia Coal Basin. Nevertheless, according to Gradziński *et al.* (2005 and references therein) the palaeotransport directions measured by Gradziński *et al.* (1961) are unreliable, owing to the small amount of data. However, the types of pebbles (Paszkowski *et al.*, 1995), the petrographic study of the sandstones (Świerczewska, 1995) and the chemical dating of detrital monazite (Kusiak *et al.*, 2006) provide data that indicate igneous and metamorphic rocks of the Bohemian Massif, as dominating sources for the Carboniferous of the Upper Silesia Coal Basin. The massif also was considered as a source area for detrital, pyrope-rich garnet from the Middle Jurassic sediments of the Kraków–Wieluń Upland (Méres *et al.*, 2012), although the authors proposed another source, located generally to the S of their area of investigation, in a more proximal position than the Bohemian Massif, with respect to the Kraków–Wieluń Upland. This hypothetical land may be covered at present by the Carpathian overthrust. However, its existence has not been con-

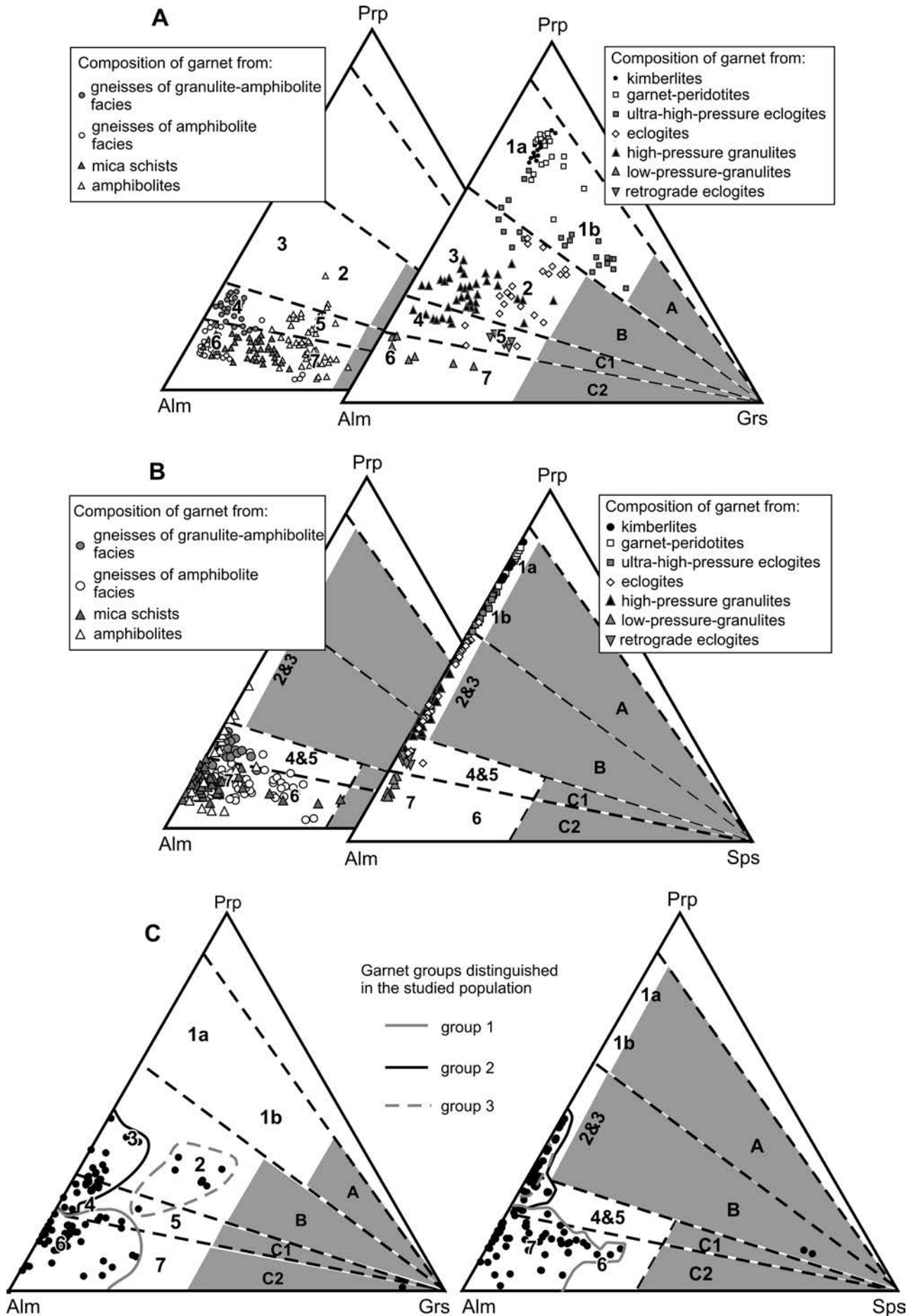


Fig. 8. Composition of garnets of Oligocene sandstones of Menilite Formation, compared to garnet composition from various metamorphic and igneous rocks. **A, B:** Composition of garnets from UHP/HP metamorphic conditions in classification diagrams “pyrope-almandine-grossular” (according to Aubrecht *et al.*, 2009 and references therein; Méres *et al.*, 2012): field A – garnet compositions from HP/UHP conditions; field B – garnet compositions from granulite and eclogite facies conditions; field C1 – transitional field of garnet compositions from high amphibolite to granulite facies conditions; field C2 – garnet compositions from amphibolite facies conditions (this field includes also garnet from blue schists, skarns, serpentinites, igneous rocks, etc.). 1a – garnet from UHP eclogites, garnet – peridotites and kimberlites; 1b – garnet from UHP eclogites; 2 – garnet from HP eclogites and HP mafic granulites; 3 – garnet from HP felsic and intermediate granulites; 4 – garnet from gneisses metamorphosed under P-T conditions transitional to granulite and amphibolite facies; 5 – garnet from amphibolites metamorphosed under P-T conditions transitional to granulite and amphibolite facies; 6 – garnet from gneisses metamorphosed under amphibolite facies conditions; 7 – garnet from amphibolites metamorphosed under amphibolite facies conditions.

Source rocks of individual garnets: *Right diagram:* garnet from HP granulites in Góry Sowie Mts (Polish Sudetes), garnet from peridotites, eclogites and granulites from Bohemian Massif, garnet from HP and UHP eclogites and garnet peridotites from Western Gneiss Region (WGR, Norway), garnet from kimberlites, garnet from eclogites with inclusions of diamond, garnet from HP granulites, UHP eclogites with inclusions of coesite and garnet peridotites from Saxonian Erzgebirge and Granulitgebirge. *Left diagram:* garnet from mica schists, gneisses and amphibolites and amphibolized eclogites occurrence in pre-Alpine basement rocks of Western Carpathians Mts (see Aubrecht *et al.*, 2009 and references therein). **C:** composition of garnet population studied from lower part of Menilite Formation in classification diagrams. Abbreviations as in Fig. 4

firmed by direct investigations. Taking into consideration the compositional similarities of the detrital garnets from the Menilite Formation and the Middle Jurassic clastics (Méres *et al.*, 2012), the source proposed by the authors mentioned also may be considered in provenance studies of the heavy-mineral assemblages.

The presently exposed, igneous and metamorphic bodies, available for direct investigations and comparative studies, are widespread within the Bohemian Massif (see e.g., Dallmeyer *et al.*, 1995; Mazur *et al.*, 2006; Żelaźniewicz *et al.*, 2011 and references therein). These include crystalline rocks, containing garnet and/or tourmaline, similar in com-

position to the detrital grains from the Oligocene part of the Menilite Formation.

The more proximal area is in the Western Sudetes. Garnet varieties, corresponding well in composition with the garnets studied, were described from various rocks in the Sowie Góry Massif, the Strzegom–Sobótka Granite Pluton and the Orlica–Śnieżnik and Karkonosze–Izera massifs, as well as from rocks of the Kamieniec Metamorphic Belt (Tab. 3). The composition of the detrital garnets studied is also consistent with the chemistry of the detrital garnets from the Carboniferous, Permian and Turonian clastics, filling sedimentary basins in the Sudetes (Tab. 3), for which

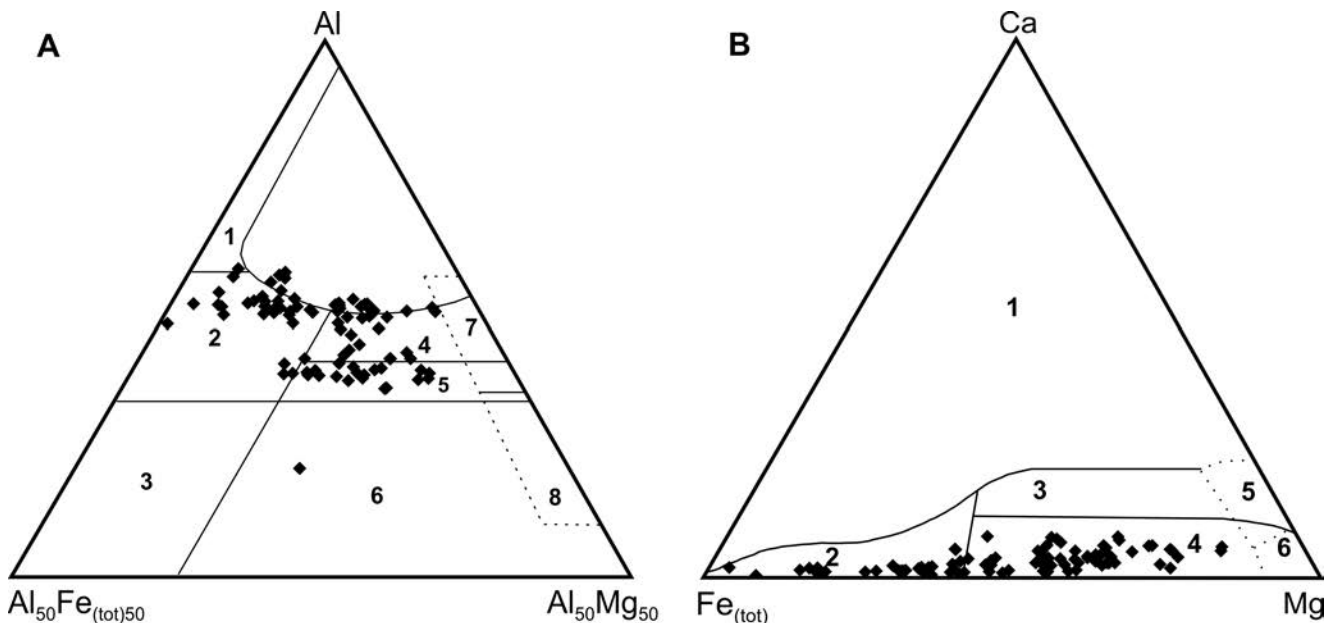


Fig. 9. Provenance of tourmalines studied in “environmental” diagrams of Henry and Guidotti (1985). **A:** 1 – Li-rich granitoids, pegmatites and aplites; 2 – Li-poor granitoids, pegmatites and aplites; 3 – hydrothermally altered, granitic rocks; 4 – Al-rich metapelites and metapsammites; 5 – Al-poor metapelites and metapsammites; 6 – Fe³⁺-rich quartz-tourmaline rocks, calcsilicates and metapelites; 7 – Ca-poor ultramafites; 8 – metacarbonates and metapyroxenites. **B:** 1 – Li-rich granitoids, pegmatites and aplites; 2 – Li-poor granitoids, pegmatites and aplites; 3 – Ca-rich metapsammites, metapelites and calcsilicates; 4 – Ca-poor metapsammites, metapelites and quartz-tourmaline rocks; 5 – metacarbonates; 6 – metapyroxenites

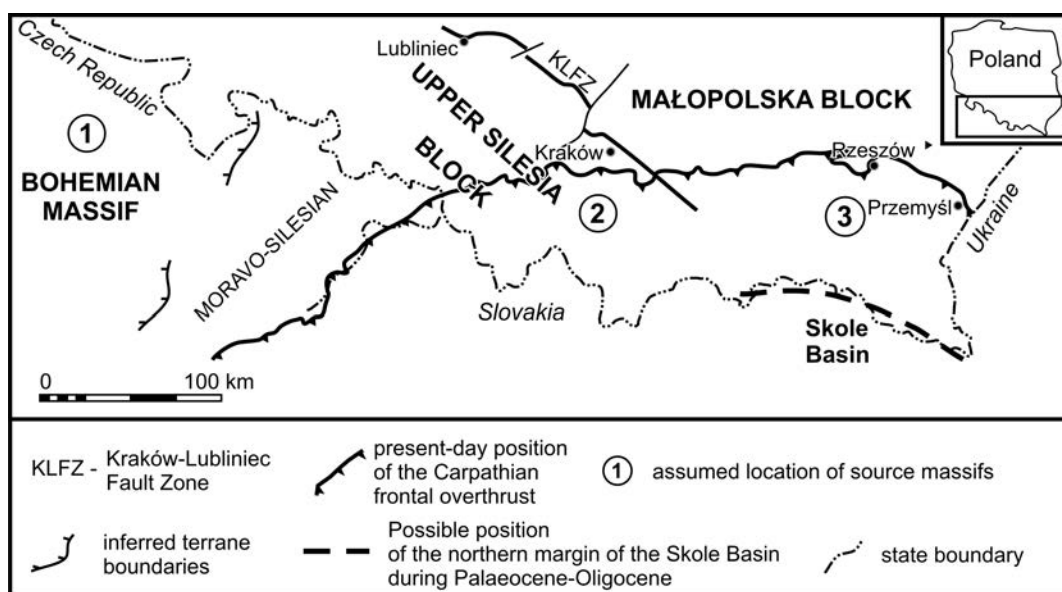


Fig. 10. Sketch-map (based on Żelaźniewicz *et al.*, 2011), showing assumed location of crystalline massifs, which could be primary source areas of heavy minerals from Oligocene sandstones in Skole Nappe: 1 – crystalline complexes of Bohemian Massif; 2 – hypothetical area, proposed as source area for Middle Jurassic clastics of Kraków–Wieluń Upland (see Méres *et al.*, 2012); 3 – hypothetical massif, located directly in Skole Basin foreland.

some of the massifs, mentioned above, were determined as source rocks (Felicka, 2000; Biernacka and Józefiak, 2009; Biernacka, 2012a).

Additionally, pyrope-almandine garnet with an elevated Ca-content may be found in granulitic and eclogitic bodies of parts of the Bohemian Massif (Kotková, 2007 and references therein; see also Fig. 8) that are more internal, in relation to the Sudetes. Moreover, several dozen grains of the detrital garnets analysed, especially those with low amounts of grossular and spessartine end-members, correspond well in composition with detrital garnets from the Lower Carboniferous clastics of the Culm Basin, in the eastern part of the Bohemian Massif (Tab. 3). According to Copjaková *et al.* (2005) and Kotková *et al.* (2007), this kind of garnet should have come from the eroded part of the massif, as garnet of such a composition is not present at the present-day erosion level. A similar composition to the detrital garnets studied also is characteristic for detrital garnets from the Permian sandstones and conglomerates of the Krkonoše (Karkonosze) Piedmont Basin (Martínek and Štolfová, 2009), which originated in the Moldanubian crystalline complexes.

Like the garnets, the tourmaline data provide analogous clues about provenance. The detrital tourmaline shows similarity in composition to tourmaline, occurring in several areas in the Sudetes. These include the Karkonosze–Izera Massif, with its metamorphic envelope, the Strzegom Granitoid Massif, the Góry Sowie Block, granitic intrusions within Jordanów Śląski and the Szklary massifs (Janeczek, 1985; Pieczka and Kraczk, 1988; Pieczka, 1996; Słaby and Kozłowski, 2005). Tourmaline of schörl-dravite composition, comparable to the detrital grains studied, is also quite abundant in the Variscan, granitic rocks of Moldanubicum, in the Czech part of the Bohemian Massif (e.g., Novák *et*

al., 2004; Buriánek and Novák, 2007 and references therein). Tourmaline from these localities belongs to the alkali group and displays similar amounts of the main cations and values of Mg/(Mg+Fe) index, as well as the X-site vacancy, to that measured in the detrital tourmaline analysed. Moreover, it occurs as similar colour varieties.

Also comparable in composition is the detrital tourmaline, occurring in the Upper Cretaceous sandstones in the North Sudetic Synclinorium (Biernacka and Józefiak, 2009; Biernacka, 2012b), which represent recycled, clastic material, transported from the eroded Bohemian Massif (Biernacka, 2012b). A considerable part of the tourmalines from the North Sudetic Synclinorium resemble in composition the tourmalines from pegmatites, occurring in the granulitic bodies of the Bohemian Massif, and from the melasyenites of the Trebic Pluton from the Moldanubian Zone of the massif (Biernacka, 2012b). Therefore, the detrital tourmaline from the Menilite Formation may originate from the same petrographic source types as the detrital tourmaline of the North Sudetic Synclinorium.

Taking into account all the similarities discussed above, several areas can be considered as primary source areas for the garnet and tourmaline populations studied. These include the Bohemian Massif, a hypothetical area, located S of the Kraków–Wieluń Upland or metamorphic and igneous bodies, located possibly in the foreland of the Skole Basin (Fig. 10). Some parts of the Bohemian Massif, such as the Góry Sowie Massif, appear to be too small, judging by their present dimensions. However, the Góry Sowie Massif in particular, according to Biernacka and Józefiak (2009) may be underestimated in terms of its size, since a large part of it may be covered.

The problem of provenance of the majority of tourmaline and garnet grains, occurring in the heavy mineral as-

semblages studied seems to be explained to some extent. However, the origins of the relatively scarce euhedral, well-preserved grains, including tourmaline, zircon, kyanite and garnet (Salata and Uchman, 2012), merit brief discussion. If crystalline source massifs are to be considered as possible sources of the minerals, they must have been located in a more proximal position than the Bohemian Massif, in relation to the Skole Basin, during the Oligocene. Such massifs, eroded or overthrust by the Carpathian nappes, could have existed in the foreland or within the Skole Basin. However, this hypothesis is not confirmed by direct studies. Another possibility for the provenance of the euhedral grains is through release during the disintegration of crystalline rock pebbles, present in the source area. Nevertheless, both hypotheses, explaining the provenance of euhedral grains, are contradicted by the scarcity and low diversity of the pebbles, found in the sandstones of the Kliva type, among which mainly sedimentary rocks and coal particles were found (see Kotlarczyk and Śliwowa, 1963; Kotlarczyk, 1979).

CONCLUSIONS

The garnets studied display a chemical composition, similar to that of the detrital garnets from the Carboniferous and Middle Jurassic clastics of the Upper Silesia Block sedimentary cover. This implies that these sediments could be considered as source rocks for the clastic material in the Skole Basin, during the Oligocene.

The chemical composition of the garnets indicates that they were formed in various metamorphic rocks under conditions, ranging from amphibolite to eclogite facies. However some of them also may have originated from igneous bodies. The composition of tourmalines, like that of the garnets, indicates a provenance from metasediments and igneous rocks, mainly of granitic composition. The detrital garnets and tourmalines display a resemblance to garnet and tourmaline crystals, occurring in metamorphic and igneous rocks of the Bohemian Massif as well as to detrital grains, occurring in clastic sediments, which accumulated in its internal basins and foreland. This suggests that the parent rocks of the minerals studied could be located within the massif. However, other hypothetical, uplifted areas, composed of metamorphic and igneous rocks, petrographically similar to those of the Bohemian Massif and located in the northern foreland of the Carpathian sedimentary basins, cannot be excluded as possible source areas.

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