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Oxfordian neptunian dykes with brachiopods from the southern part of the Kraków-Częstochowa Upland (southern Poland) and their links to hydrothermal vents

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Abstract Neptunian dykes with abundant brachiopods, cf. Lacunosella sp. and fragments of echinoderms, occur in Oxfordian limestones in the southern part of the Kraków-Częstochowa Upland. The dykes fill fissures that have opened in the massive limestones due to local extension of the sedimentary basin located along the northern, passive margin of the Tethys Ocean. These fissures transmitted warm hydrothermal solutions that controlled the mass growth of free-living bacteria and microbial mats feeding the fauna, mostly brachiopods and echinoderms, settling the seafloor around the fissures. For some time, the fissures remained empty and their vertical walls were settled by stromatolites. Infilling of the fissures was an abrupt event related to faulting in the Oxfordian and to the rejuvenation of dislocations cutting through the Paleozoic basement.

Then, in the Cretaceous, and primarily in the Cenozoic, tectonic discontinuities filled with neptunian dykes were penetrated by karst waters and by hydrothermal solutions, which partly silicified the carbonate material infilling the dykes. The formation of dykes is genetically related to the Late Jurassic, Pan-European stress-field reorganization caused by the opening of the Northern Atlantic and Tethys Oceans.

Keywords Neptunian dykes · Hydrothermal vents · Brachiopods · Upper Jurassic

Introduction

Neptunian dykes are defined as fissures within rocks exposed on the sea bottom that have been filled with submarine sediments (Flügel 2004). The formation of such fissures is interpreted as an effect of: (1) depositional processes, (2) synsedimentary features, (3) diagenetic transformations of sediments, and, above all, (4) extensional tectonics in the sedimentary basin (see e.g., Land and Goreau 1970; Wendt 1971; Smart et al. 1988; Martire 1996; Winterer et al. 1991; Santantonio 1993; Winterer and Sarti 1994; Molina et al. 1995; Wieczorek and Olszewska 2001; Łuczyński 2001; Flügel 2004; Wall and Jenkyns 2004; Črne et al. 2007; Montenat et al. 2007; Kandemir and Yilmaz 2009; Kołodziej et al. 2010; Reolid and Molina 2010; Reolid et al. 2010; Nieto et al. 2012; Barski 2012).

The Upper Jurassic neptunian dykes hosted in carbonate succession from the southern part of the Kraków-Częstochowa Upland (KCU) have already been described in several short announcements (Hoffmann and Matyszskiewicz 1989; Wieczorek and Krobicki 1994; Krajewski 2004; Krajewski and Matyszskiewicz 2005; Krajewski 2006).
2004; Jędrys and Krajewski 2007), but their development has never been studied in detail. Neptunian dykes cutting through the Oxfordian microbial-sponge carbonate buildups were encountered in quarries in Młynka and Czajowice as well as in the Prądnik River and Będkowice Valleys. Among them were dykes filled with brachiopods *Rhynchonella* sp. (*Wieczorek and Krobicki 1994; Krajewski 2004*) or with carbonate detrital material (*Koszarski 1995; Krajewski 2004; Krajewski and Matyszkiewicz 2004; Jędrys and Krajewski 2007*).

Although brachiopods are common in Oxfordian sediments of the KCU, their accumulations are rather unique. In the literature, several species were identified: *Lacinussella cracoviensis* (*Quenstedt*) and rare *L. arolica* (*Oppel*), *L. trilobata formis* Wiśniewska, *Septaliphoria asteriana* (d’Orbigny), *S. moravica* (Uhlig), *Sellithyris engeli* (Rollier), *Zelleria delmontana* (*Oppel*) and *Terebratulina substriata* (Schlotheim) (*Różyczyk 1948; Wierzbowski 1970; Wiśniewska-Żelichowska 1971; Heliasz and Racki 1980*). In Oxfordian microbial-sponge buildups, the most common are *L. cracoviensis*, *S. asteriana* and *T. substriata* (*Różyczyk 1948; Wiśniewska-Żelichowska 1971; Heliasz and Racki 1980*), whereas in inter-buildup sediments, *S. engeli*, *Z. delmontana*, and *S. moravica* were found (*Heliasz and Racki 1980*). In the southern part of KCU, the dominant form is *L. cracoviensis*, characterized by local morphotypes (*Różyczyk 1948*) of mostly symmetric shells and average sizes significantly larger than those observed in the northern KCU, where asymmetric forms prevail (*Różyczyk 1948; Wierzbowski 1970; Heliasz and Racki 1980*). Presumably, this species predominates in sediments filling the studied neptunian dykes.

In recent years, exploration in the Big (Polish: Duża) Cave located in the slope of the Kluczwoda Valley discovered a neptunian dyke filled with numerous brachiopods (*Nowak 2014*). This work presents the development of unique dykes filled with brachiopods discovered in only three localities: (1) in the operating Młynka Quarry, (2) in the Big Cave in the Kluczwoda Valley and (3) in the Grodzisko Rock in the Prądnik River Valley (Fig. 1). Field observations, combined with the results of microfacies and geochemical analyses and cathodoluminescence, isotopic and fluid inclusion studies, enabled us to reconstruct the formation of these neptunian dykes, and to discuss the regional context of their occurrence.

**Geological setting**

The KCU is located in southern Poland and belongs to the regional tectonic unit known as the Silesian-Kraków Homocline (SKH), built of Triassic, Jurassic, and Cretaceous sediments (Fig. 1). Here, Mesozoic strata unconformably cover Precambrian and Paleozoic formations cut by the Kraków-Lubliniec Fault Zone (KLFZ) and the somewhat younger Krzeszowice-Charsznica Fault, which displaces it (*Habryn et al. 2014*). The KLFZ, active since the Early Paleozoic (*Morawska 1997; Zaba 1999*), is accompanied by manifestation of Paleozoic intrusions. Some of the intrusions became erosively exposed in the Oxfordian forming topographic highs upon the sea bottom. The areas underlain by those intrusions whose top parts were not eroded revealed less intense subsidence as compared to the neighboring areas, thereby creating structural highs upon the sea bottom, which became colonized by benthic organisms (*Matyszkiewicz et al. 2006a*).

Upper Jurassic carbonates were deposited on the passive shelf of the northern margin of the Tethys Ocean, to which the KCU also belongs. They represent a transgressive–regressive, Callovian-Lower Kimmeridgian tectono-stratigraphic unit (*Kutek 1994*) and are developed as microbial-sponge megafacies, which comprises bedded and massive facies as well as submarine gravity flows (Fig. 2; *Matyszkiewicz et al. 2015b*).

**Bedded facies**

The Upper Jurassic succession (Fig. 2) begins with a series of thin-bedded marls and alternating limestones with abundant benthic and nektonic fauna of the Early and early Middle Oxfordian ages. These strata grade upwards into a strongly differentiated limestone complex. Sediments of the Middle Oxfordian bedded facies represent thin- to medium-bedded pelitic limestones locally intercalated with marls and thin calciturbidites. The upper Middle Oxfordian and Upper Oxfordian successions are dominated by thick-bedded limestones with flints. The topmost part of the Upper Jurassic section of the Lower Kimmeridgian age is composed of bedded limestone-marl strata, known mostly from well logs (*Burzewski 1969*).

**Massive facies**

The massive facies comprise various types of carbonate buildups (terminology after *Riding 2002*), as well as olistoliths separated from massive limestone complexes and embedded within debris flow deposits. Carbonate buildups started to develop at the end of the Early and beginning of the Middle Oxfordian (Fig. 2) as small, sponge-microbial, low-relief carbonate mud mounds without rigid frameworks. These mounds then evolved over time into: (1) microbial-sponge segment reefs with laminar frameworks; (2) filled frame reefs with initial, reticulate rigid frameworks, which, in turn, were later replaced by (3) open frame reefs with well-developed reticulate rigid frameworks (*Matyszkiewicz et al. 2012*). Development
of carbonate buildups attained its climax at the end of the Middle and beginning of the Late Oxfordian. Carbonate buildups began to disappear in the Late Oxfordian and sea bottom relief became partly leveled.

Sediment gravity flows

Sediment gravity flows are common throughout the Upper Jurassic succession, being particularly spectacular at the break between the Oxfordian and the Lower Kimmeridgian (Fig. 2). These are mostly debris flow sediments and calciturbidites. Their occurrence is associated with the temporary deterioration of conditions controlling the growth of carbonate buildups, coeval, active, syngentic tectonics (Matyszkwicz 1996, 1997), and sea-level fluctuations. Debris flow sediments were observed particularly along the margins of the tectonic horsts (Matyszkwicz 1996; Ziółkowski 2007; Matyszkwicz et al. 2012), where the best conditions for observation are.

Materials and methods

Two neptunian dykes with abundant brachiopods were exposed in the 1980s and 1990s at the operating Młynka Quarry (MQ). Unfortunately, these dykes have not survived; however, the authors possess about 20 samples collected from their infillings.
The Big Cave (BC) site in the Maączna Rock was discovered in the years 2013–2014 in the course of the exploration of a cave located in the eastern slope of the Kluczwoda Valley (Nowak 2014; Wrzak 2014). This is the only available neptunian dyke with brachiopods from which 25 samples were collected from the infilling material and the host rock.

The last-studied neptunian dyke was exposed for a short time in 2001 in the Grodzisko Rock (GR), located in the northern slope of the Prądnik River Valley, in an excavation dug for the construction of a house (Krajewski 2004). At this site, several samples were collected of the brachiopod-rudstone and the host rock.

From samples collected at the MQ and BC sites, 25 thin-sections and 15 polished sections, which served as the basic materials for microfacies analysis, were prepared. Thin-sections from both the MQ and BC sites were subjected to CL examinations; in those from the BC site, additional measurements of FI homogenization temperatures were made in quartz crystals and microprobe analyses of cements were carried out. Material was collected from both the MQ and BC samples for geochemical and Nd isotope analyses.

The principal geochemical analyses were carried out at Activation Laboratories Ltd. (ACTLABS) in Ancaster (Canada) using fusion-inductively coupled plasma (FUS-ICP) for ten samples from the BC and two others from the MQ sites.

The CL analyses were carried out on polished, uncovered thin-sections with an 8300 Mk III cold cathode instrument at the Institute of Geological Sciences, Polish Academy of Sciences, in Kraków.

Measurements of FI homogenization temperatures ($T_h$) in quartz were carried out at the immersion heating stage (Kozłowski et al. 1979). In total, 52 inclusions were measured, following the principles established by Roedder (1984). The freezing runs yielded data enabling calculation of the concentrations of salts in the fluids (Crawford 1981; Kozłowski 1984) and identification of the presence of carbon dioxide and methane in fluid inclusions. Each heating and cooling run was repeated three times; for interpretation, only those runs were used for which the obtained values were satisfactorily consistent (cf. Roedder 1984).

Microprobe analyses of carbonate cements were completed at the Laboratory of Critical Elements, Faculty of Geology, Geophysics and Environment Protection at the
AGH University of Science and Technology in Kraków using a JEOL Super Probe JXA-8230.

Nd isotope studies were carried out at the Isotope Laboratory of the Adam Mickiewicz University in Poznań (Poland) with a Finnigan MAT 261 multi-collector thermal ionization mass spectrometer. Nine samples collected from the neptunian dyke at the BC site and two others taken from the MQ site were analyzed.

The Młynka Quarry

The Młynka Quarry is located at the southern margin of the Tenczynek Horst, which borders the Krzeszowice Graben to the south (Fig. 1). The exposed Lower and Middle Oxfordian succession includes zones from the Cordatum to the Transversarium (Głowniak 2006; Jurkowska and Kołodziej 2013). The Upper Jurassic sequence begins with marls with sponges and ammonites which grade upward into thin-bedded, pelitic limestones with marl intercalations (Fig. 2). A horizon of glauconitic marls occurs in the upper part of the pelitic limestones (Jurkowska and Kołodziej 2013), followed by massive limestones representing the Plicatilis-Transversarium zones and developed as microbial-sponge facies with siliceous sponges, bivalves, brachiopods, ammonites, bryozoans, and fragments of echinoderms, sometimes accompanied by numerous specimens of Crescentiella sp. (Hoffmann and Matyszkiwicz 1989). In the top portion of the quarry walls, debris flow with clasts of massive limestones is locally present.

Two neptunian dykes with attitudes N20W 90 and N10W 90 were exposed in the upper bench of the quarry. Both dykes were filled with densely packed brachiopods embedded within a light-beige matrix. The dykes cut through massive limestones in which ammonites were found, indicative of the Transversarium Zone (Głowniak 2006). The maximum thickness of the dykes was about 0.5 m. The dykes covered the exposed surface of massive limestone, which was over a dozen meters long and up to 10 m high (Wieczorek and Krobicki 1994).
The infilling material in dykes (Figs. 3, 4) exhibits both vertical and lateral variability. The walls of the dykes were covered by vertical stromatolite of about 1.5 cm in thickness (Figs. 3, 5a), which, along a 10-cm stretch, contact floatstone, which in turn fills the marginal parts of the dyke and changes into rudstone towards its centre. Stromatolite on the wall of massive limestone is composed of peloidal and micropeloidal lamina, 0.5 to 3 mm thick. The thicker lamina are composed of peloids and thinner of micropeloids (Fig. 5a). The lamina are parallel, but sometimes wavy also. Within lamina composed of micropeloids are occasionally observed oval single unidentified coprolites up to 0.8 mm in diameter.

In the floatstone, brachiopods found at contact with the massive limestone are greatly flattened and elongated parallel to the walls of the dyke (Fig. 3). Deformation of shells decreases towards the centre of the dyke. In the rudstone, brachiopod shells contact each other and preserve their oval shapes, but are sometimes crushed. Locally, in the central part of the dyke we observed angular clasts, up to 20 cm in diameter, developed as wackestone–packstone and boundstone with calcified siliceous sponges (Fig. 3).

In the matrix we identified *Crescentiella morronensis*, oncoids up to 0.5 mm in diameter, peloids, single bioclasts, unidentified coprolites, benthic foraminifera, and tuberoids (Fig. 5b–e). Locally, accumulations of peloids are present. The internal sediments filling the brachiopod shells show development similar to that of the enclosing matrix (Fig. 5c, d). In the latter, we identified additionally fragments of echinoderms up to several millimeters in diameter.

![Fig. 4](image-url) Floatstone from the infilling of the MQ dyke. Brachiopod shells show varying degrees of infilling. Some are filled with coarse crystalline quartz (arrows). The matrix is wackestone-packstone or grainstone with echinoderm detritus
(Fig. 5b) with thin microbial envelopes. The internal sediments, along with carbonate cement, fill the shells entirely or geopetally.

In flattened brachiopod shells located close to the walls of the dyke, the boundary between internal sediment and cements is vertical (Fig. 3). The orientation of the geopetal infillings of shells found in the central parts of the dykes is chaotic (Fig. 4). Some shells are empty; others are filled with carbonate cement or quartz.

In the largest shells, we sometimes observed juvenile forms of brachiopods up to several millimeters across (Fig. 5d). The surfaces and interiors of brachiopods shells contain diversified cements. The most common is radiaxial fibrous cement, which locally coats the shells from outside (Fig. 5d) but sometimes also covers the interiors, particularly if the shells are only partly filled with the internal sediment (Fig. 5e, f). In such cases, the radiaxial fibrous cement is usually covered with dogtooth cement. The internal surfaces of shells can also be covered with two other cements: (1) bladed and granular, sometimes found on the surfaces of internal sediment in partly filled shells (Fig. 5c) and (2) granular, observed in shells initially filled with internal sediment which, however, contacts the shell walls at present, whereas the granular cement fills the centers of the shells (Fig. 5c). Occasionally, brachiopods shells are filled with granular quartz (Fig. 5e, f), which hosts only scarce fluid inclusions. Sporadically, the internal sediment infillings contain fine (up to 1 mm across) aggregates of microflamboyant quartz (Fig. 5c). The internal surfaces of cavity walls are rarely laminated crystal silt with single bioclasts, mostly microoncoids up to 1 mm across, with unidentified fragments of bioclasts in the nuclei, are common. The microoncoids form local accumulations together with Crescentiella morronensis, single peloids, and small bioclasts up to 1 mm across. Bioclast surfaces are usually covered with films of isopachous granular cement, which grade towards the centers of the bioclasts into dogtooth cement contacting granular and blocky calcite cements (Fig. 10a, b). Locally, the fine-grained matrix of floatstones is replaced by microcrystalline granular quartz. We also observed aggregates of microflamboyant quartz, up to several millimeters across, which replaced echinoderm plates and cements (Fig. 8a).

In the matrix, cavities 1–2 cm long and 0.5 cm high are common, filled with internal sediment developed as irregularly laminated crystal silt with single bioclasts, mostly echinoderm plates embedded within syntaxial cement (Figs. 8a, 10c). The internal surfaces of cavity walls are occasionally covered with dogtooth cement (Fig. 8b–d). Sometimes the larger (several centimeters long) cavities are filled with coarse crystalline, granular quartz with the relics of blocky calcite cements. Two generations of this quartz occur: (1) granular quartz with abundant fluid inclusions at the walls of cavities and (2) inclusion-free, granular quartz (Fig. 10c, d).

The shells are: (1) empty; (2) geopetally filled with chaotic orientation; or (3) entirely filled with internal sediment, cement, or even coarse crystalline quartz (Fig. 7a–c). The interiors of brachiopod shells are entirely or partly filled with internal sediments and diversified carbonate cements (Fig. 10f) or by coarse crystalline, granular quartz (Fig. 10e). In the parts of shells filled with cements, over the internal sediments we observed granular and blocky calcite cements along with bladed calcite cement, the latter growing from the upper surface of the shells towards their centers (Fig. 10f). In geopetally filled shells, the internal sediments are mudstones in the lower parts of the shells (in situ) and packstones deposited immediately over the sediment, at the point of contact with the cement. Locally,
aggregates of microflamboyant quartz can be found in brachiopod shells (Fig. 10f).

Some brachiopod shells are crushed and displaced along with infilling internal sediment (Fig. 10f). The cracks along which displacements proceeded are filled with blocky calcite cement. Displacement was also observed in an echinoderm spine whose fragments were moved against each other along a narrow crack filled with granular cement and quartz with relics of blocky calcite cement (Fig. 9d).

The silicification process observed within of the brachiopod shells (Fig. 10e) produced three types of infillings: (1) granular quartz crystals of diameters up to about 2 mm, entirely filling the shell; these quartz crystals host rare fluid inclusions; (2) large (up to 5 mm across) quartz crystals with numerous fluid inclusions; and (3) irregularly distributed aggregates of microflamboyant quartz encountered in shells entirely filled with internal sediment. The internal walls of shells filled with granular quartz are locally covered by dogtooth cement (Fig. 10e).

The cracks, several millimeters wide, which cut through the dyke are filled with mudstone/wackestone or by internal sediment with Fe-oxides, sometimes rich in crushed calcite crystals (Fig. 9a–c). The walls of these cracks are covered with blocky calcite cement or dogtooth cement locally overgrown by syntaxial calcite cement with increased Fe content (Figs. 9a–c, 11; Table 1).

The Grodzisko Rock

The Grodzisko Rock, located in the northern slope of the Prądnik River Valley, is composed of microbial-sponge and microbial-Crescentiella massive limestones. In adjacent outcrops, massive limestones contain ammonites indicative of the Bifurcatus zone (Ziółkowski 2007), whereas in the overlying, pelitic, and detrital limestones (wackestones–packstones), fossils typical of the Bimammatum Zone were found (Ziółkowski 2007).

The contact of the neptunian dyke with the massive limestones is sharp. The brownish material infilling the dyke is greatly karstified and disintegrated. The matrix around the brachiopod shells has barely survived. Brachiopods form the rudstone, composed of chaotically arranged but perfectly preserved and usually empty shells (Fig. 7d). The outer surfaces of shells locally show disarticulation (cf. Olóriz et al. 2002). The degree of fragmentation is very low. Sporadically, we observed crushed shells or their fragments. Some shells are filled with coarse crystalline quartz.

Brachiopods from the neptunian dykes

The brachiopods from MQ, BC, and GR could not be precisely identified due to the exclusively relic-grade preservation of brachial apparatuses. In brachiopod accumulations, shells were occasionally crushed, but most were perfectly preserved. In any case, our brachiopods belong to a monospecific assemblage of rhynchoellids, cf. Lacunosella sp., although forms with smoother shell surfaces are also present. The latter resemble terebratulids but presumably, based on the presence of subtle ridges, are only variants of rhynchoellids. The shells are articulated. Both juvenile and adult forms are present. The abundance in MQ reaches 33 specimens, in the BC 14 specimens, and in GR reaches 18 specimens per 25 cm². The width of the shells from MQ varies from 5.2 to 30 mm, the length from 5.7 to 27.2 mm, with average values of 15 and 13 mm, respectively; from BC the width varies from 6 to 30.1 mm, the length from 5.9 to 27.2 mm, with average values of 25 mm for both, and from GR the width varies from 6 to 30.1 mm, the length from 6.2 to 27.2 mm, with average values of 25 mm for both.

Analytical results

General geochemistry

The floatstone infilling the BC neptunian dyke is locally highly silicified (SiO₂ content up to 68.5 %, Table 2). In samples rich in SiO₂, we observed increased Ba content (63–110 ppm) whereas silica-free samples contained distinctly less Ba (9–38 ppm). In silicified samples, Sr content is lower (from 92 to 128 ppm) but varies in silica-free
samples from 161 to 181 ppm. The floatstone from the MQ neptunian dyke show a chemical composition similar to silica-free fragments of the BC dyke (Table 2).

**Cement geochemistry**

The results of microprobe point analyses of dogtooth calcite cements and syntaxial calcite overgrowth cements collected from fractures cutting through the BC dyke are listed in Table 1 and shown in Fig. 11. The syntaxial cements indicate increased Fe content (up to 2.090 ppm), whereas in the dogtooth cements Fe contents do not exceed 80 ppm and are usually below the MPA detection limit. The remaining analyzed elements show no distinct correlations with these two types of cements.

**Nd isotope geochemistry**

The results of Nd isotope studies of the matrix of floatstones from the MQ and BC neptunian dykes show some differences (Table 3). For 9 BC samples, the $\varepsilon_{\text{T}}(T = 160 \text{ Ma})$ values change from $-8.1$ to $-6.3$ (average value: $-7.2$),

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**Fig. 6** Fragment of a neptunian dyke in a corridor wall of the Big Cave in the Kluczwoda Valley. The contact zone of the dyke (on the right) and massive limestone (on the left) is covered with brownish karst sediment (black arrow). To the left of the coin, breccia of massive limestone (b) is exposed.

**Fig. 7** Infilling of the BC (a–c) and GR (d) neptunian dykes: a contact of the dyke (right) and massive limestone (left). Visible in the contact zone are fragments of vertical laminated stromatolites (white arrows), which were separated from the wall of massive limestone along a short distance due to fault displacement. Brachiopod shells located close to the wall are crushed, whereas those located in the centre are fully preserved. Some shells are geopetally filled with grainstone (black arrows) with numerous echinoderm plates. The top is indicated by big white arrow; b floatstone with brachiopods and fragments of echinoderms in the matrix. On the right: a fully preserved echinoid test (arrow). The rock is fractured and the fractures are impregnated with black Mn dendrites and brownish Fe oxides; c floatstone: locally, rudstone with brachiopods. In the matrix, fragments of echinoid tests and spines are visible (arrows). In the lower part of the photograph, brachiopod shells, the cavity between them, and the adjacent matrix are all filled with coarse crystalline quartz; d empty shells contact each other. The matrix was removed by karstic processes.
whereas for two MQ samples the $e_{(T = 160 \text{ Ma})}$ values are $-6.9$ and $-6.8$.

The $T_{2DM}$ age for BC samples varies from 1.46 to 1.61 Ga (average value: 1.54 Ga), for MQ samples from 1.50 to 1.51 Ga (Table 3). Differences are minor but relevant to the stratigraphic positions of both dykes.

Fluid inclusions

The fluid inclusions found in quartz crystals are rare and small (<10 µm) and filled with the liquid phase, which is an aqueous solution with small gas bubbles. Some small (<5 µm long) inclusions developed only after cooling, i.e. after breakdown of the metastable homogeneous state. Since the formation pressure of quartz crystals should be low and probably close to that of a rather thin seawater column, the actual formation temperature of inclusions does not differ significantly from that determined using homogenization temperatures.

Our inclusions can be divided into two groups. The first includes the earliest inclusions present in the central parts of quartz crystals, which homogenized at $T_h = 89$–$54 \degree$C. These are relatively large (from ca 5 to nearly 10 µm), primary inclusions, i.e. syngenetic with the appropriate zones of studied quartz crystals. Their homogenization temperatures are quite high and variable. Moreover, the same parts of quartz crystals host fluid inclusions of the second group, which are smaller (<5 µm) and yield homogenization temperatures $T_h$ of 52–51 °C. In two cases, two or three first-group inclusions shared the same growth zone with the single inclusion of the second group, and each inclusion provided a different homogenization temperature. This would suggest the heterogeneous trapping of the fluid portion, i.e. at least the first group of inclusions were formed through overgrowth of a portion of liquid with a small gas bubble. The gas bubbles may have formed due to a very local pressure decrease or release of gas from the wall rock sediment.

The outer parts of quartz crystals contain fluid inclusions of the second group, usually smaller than those from the central parts and sometimes not as clearly syngenetic with quartz crystallization. Nevertheless, the majority are primary, or most probably primary. Moreover, they display almost constant $T_h$ values, from 52 to 49 °C. Unquestionable evidence was found in the growth zone in the part of a quartz crystals where heterogeneous trapping occurred together with homogeneous: two inclusions of $T_h = 51 \degree$C were identified on two sides of an inclusion which homogenized at 72 °C (Fig. 12; Table 4). Thus, the two inclusions originated from homogeneous trapping; their $T_h$ values indicate the formation temperature of quartz, whereas the third inclusion formed by heterogeneous trapping homogenized at an incidental temperature.

All the inclusions were filled with diluted solutions (4.9–3.3 wt% of total salt concentration) with NaCl and CaCl$_2$ as the main components. Moreover, inclusions of the first group may have included very low admixtures of KCl, too low for accurate quantitative determinations. As well, traces of carbon dioxide or methane were detected in the gas phase of some inclusions. However, it should be remembered that low-concentration components may have been omitted due to the small sizes of the studied inclusions.

Discussion

Position in the ecosystem and trophic relationships

In Upper Jurassic microbial-sponge megafacies from the northern margin of the Tethys, both terebratulids and rhynchonellides are more common in Upper Jurassic carbonate buildups than in inter-buildup facies (cf. Wagenplast 1972; Nitzopoulos 1974; Gwinner 1976; Brugger 1999; Helm and Schülke 1999; Helm 2005; Olóriz et al. 2006; Courville et al. 2007; Lazăr et al. 2011). Moreover, brachiopods from carbonate buildups exhibit significant diversity of morphotypes (Helm 2005). In dense populations, decreasing size, asymmetry and even deformations of shells have been observed (Asgaard 1968; Boullier 1993; Brugger 1999; Lazăr et al. 2011).

Lacunosella sp. range from the Middle Jurassic to the Early Cretaceous and are cited as forms typical of the
northern margin of the Tethys (Sandy 1988). Accumulations of *L. cracoviensis* in Oxfordian microbial-sponge buildups were described from the Dobrogea region in southeastern Romania (Grădinaru and Bărăbulescu 1994). However, in that region, *L. cracoviensis* occurs rather in larger microbial-sponge and microbial buildups, whereas
*L. arolica* is typical of small buildups (Herrmann 1996; Lazár et al. 2011). According to Lazár et al. (2011), the environments particularly preferred by accumulations of *L. cracoviensis* are the slopes of microbial-sponge and microbial buildups, where some brachiopods might have grown in situ and others might have been relocated down the slope to those sites. *L. cracoviensis* is regarded as a shallow-marine form and its shells are settled by epifauna, which is related to the presence of distinct ridges (Krawczyński 2008). The discovery of juvenile brachiopods in the shells of adult animals found in sediments filling neptunian dykes indicates that juveniles may have lived inside the living adult brachiopods (cf. Lazár et al. 2011), that is, may have applied an endosymbiotic life strategy.

Brachiopods have a low metabolic rate, low nutrient demands (Gahr 2005; Abdelhady and Fürsich 2014) and are able to assimilate dissolved substances during times of low influx of particulate food (Steele-Petrović 1976, 1979). Brachiopod dominance may be enhanced by reduced oxygen levels and a low nutrient supply (Gahr 2005; Tomašových 2006). However, defining the environment in this way does not explain why nearly monospecific accumulations of brachiopod assemblages occur in the studied neptunian dykes.

Although brachiopods are almost unknown from recent active vent sites (Tunnicliffe 1992a, b; Lee et al. 2008), we are of the opinion that the domination of brachiopods in the environment around the neptunian dykes may have resulted from the action of shallow-water hydrothermal vents. Similarly to hydrocarbon seeps, such sites support mass growth of specific fauna, sometimes including brachiopods (Sandy and Campbell 1994; Campbell and Bottjer 1995; Little et al. 1997, 1999, 2004; Van Dover 2000; Peckmann et al. 2007, 2011; Gischler et al. 2003; Lee et al. 2008; Sandy 2010; Bujtor 2011; Sandy et al. 2014; Kiel et al. 2014; Hryniewicz et al. 2015).

The depositional environment of Upper Jurassic microbial-sponge facies in the northern margin of the Tethys occupied variable depths. The upper parts of carbonate buildups grew close to the storm wave base, whereas the inter-buildup zones were located up to 100 m deeper (Matyszkiewicz 1999; cf. Olóriz et al. 2006). Hence, this environment can be linked to shallow-water hydrothermal vents (Dando 2010). Contrary to their deep-sea counterparts, such vents are characterized by the lack of vent-specific taxa (Tarasov et al. 2005) or by the presence of only one or two non-vent-obligate genera or higher taxa (Dando 2010). In the vicinity of shallow-water hydrothermal vents, not only is seawater temperature locally higher, but venting fluids may also reduce salinity due to the input of ascending meteoric water (Stüben and Glasby 1999; Dando et al. 2000), which implies the development of specific biocones. The zones of shallow-water hydrothermal seepage are well known for particularly intensive growth of bacteria and microbial mats (cf. Stanier and Cohen-Bazire 1977; Tunnicliffe 1988, 1991, 1992a, b; Taviani 1994; Mounji et al. 1998; Dando et al. 2000; Tarasov et al. 2005; Reolid and Abad 2014), which may have provided nourishment for brachiopods feeding on phytoplankton, bacteria, colloidal and dissolved nutrients (Steele-Petrović 1976, 1979). According to Little et al. (2004), some Mesozoic brachiopod taxa may have had chemosymbionts as well (cf. Sandy 1995; Campbell and Bottjer 1995; Sandy and Campbell 1994). Moreover, Campbell (2006) states that brachiopods appear to be attracted to vents or seeps, because such sites provide access to high concentrations of bacterioplankton to be filtered from the water column. Brachiopods from neptunian dykes, like all modern brachiopods (cf. Little et al. 2004), were certainly filter feeders, and the water energy must have been sufficient to keep food particles in suspension (cf. Ayoub-Hanna et al. 2014).

Another, much less common group of organisms found in sediments filling the neptunian dykes is regular and irregular echinoderms. We found fragments of echinoid plates as well as fully preserved specimens (Fig. 7b). In hydrothermal vent or seep environments, echinoderms are rather rare, or absent (Grasse 1985; Desbruyères et al. 2006). In the central part of KCU, echinoids were observed in Upper Oxfordian carbonate buildups, in microbial mats with numerous coproliths, whose origin can also be related to warm hydrothermal seeps (Matyszkiewicz et al. 2006b; Kochman and Matyszkiewicz 2012). The source of nourishment for echinoids may also have been free-living bacteria settled in the neighborhood of fractures or microbial mats flourishing close to seep deposits during the activity of hydrothermal vents or during the final phase of chemohermas in the basin (cf. Gaillard et al. 2011).

**The rationale for the occurrence of hydrothermal vents**

The composition of Late Jurassic epicontinental sea waters occupying the southern part of the KCU corresponds to that of recent oceanic waters (Olivier and Boyet 2006; Matyszkiewicz et al. 2012. The isotopic composition of Nd documents global changes in the supply of more radiogenic waters related to hydrothermal activity over a large area (Stille and Fischer 1990; Stille et al. 1996; Frank 2002; Dera et al. 2015) but does not reflect the local, particularly low-temperature action of hydrothermal fluids due to their extremely short residence time (<1 year) in the vicinity of hydrothermal vents (Halliday et al. 1992). The primary Nd-isotope composition of hydrothermal vents is not derived directly from the water column, but is representative of shallow burial conditions (cf. Jakubowicz et al. 2015).

In a sample of unaltered Upper Oxfordian limestone from the northern part of KCU, the $^{143}Nd/^{144}Nd$ value is
Fig. 10 Internal sediments, cements and silicification in the BC neptunian dyke: a infilling of the upper part of cavity with floatstone. Syntaxial calcite cement is observed on echinoid plates (centre right and upper right), grading with optical continuity towards the cavity centre into blocky calcite cement with relics of granular cement (arrow). Surfaces of tuberoids (upper right and lower right) are locally covered with isopachous granular cement, which grades towards the cavity centers into dogtooth cement, followed by granular cement. The box marks the area shown in b. Crossed nicols; b CL image of several generations of cements infilling a cavity. Early diagenetic, isopachous granular, syntaxial and dogtooth cements show dark-orange luminescence. Granular cement, which overgrows the others, shows dark-red luminescence. Blocky calcite cement of distinct zonal luminescence represents at least two generations of cements: the first developed on syntaxial cement (arrows), whereas the second filled the cavity centre; c floatstone with brachiopod shells. The spaces between shells are filled with laminated internal sediment, carbonate cements, and at least two generations of quartz. Upper left coarse crystalline quartz with numerous fluid inclusions replaces granular and blocky calcite cements, with inclusion-free quartz in optical continuity with the coarse crystalline quartz, occupying the centre of the cavity. The fragment marked with an arrow is shown in d. Crossed nicols; d CL image of two generations of quartz filling the spaces between brachiopod shells. The older generation (left and centre), which grows directly on cements covering the surface of the shells, contains numerous fluid inclusions. The younger quartz generation (on the right), which fills the cavity centre, comprises inclusion-free crystals with relics of blocky calcite cement in bright luminescent colors (arrows); e floatstone with brachiopod shells, the interiors of which are filled with granular quartz with rare inclusions. The inner walls of the shells are covered with dogtooth cement. Intraclasts, tuberoids, microoncoids, Crescentiella morronensis and echinoderm plates with syntaxial cement are observed in the matrix enclosing the shells. On the outer surface of shell occur gravi tative fibrous cement (arrow). Crossed nicols; f infilling of a brachiopod shell. The lower part of the shell is filled with internal sediment developed as mudstone and, somewhat higher, as micropeloidal packstone. Blocky calcite cement grows immediately over the surface of the internal sediment. The inner surface of the upper part of the shell is covered with bladed calcite cement. Locally (upper right), shell fragments are silicified. The lower part of the shell is broken and displaced together with internal sediment along a fracture (arrow) filled with blocky calcite cement. Crossed nicols.

–6.6 (Matyszkiewicz et al. 2015a). This value seems to correspond to the paleogeographic position of the studied fragment of shelf from the northern margin of the Tethys (cf. Sidorczuk et al. 2009). However, the \( \varepsilon_{\text{Nd}} \) values found in sediments filling both the neptunian dykes (particularly the BC) are less radiogenic and the range of \( \varepsilon_{\text{Nd}} \) values is 1.8 (Table 3), which cannot be explained by simple acquisition of Nd from the water column (cf. Jakubowicz et al. 2015). Hence, we are of the opinion that the diversity of \( \varepsilon_{\text{Nd}} \) values in the BC dyke representing sediment of the same age must be related to the influence of hydrothermal solutions sourced in the continental crust.

Finally, we relate the increased contents of Ba in sediments filling the neptunian dykes and silicification processes to the activity of hydrothermal solutions (Table 1) presumably caused by the rejuvenation of fault zones in the Cretaceous and, primarily, in the Cenozoic.

Fig. 11 Cross-plot of Mg + Sr and Fe + Mn values in dogtooth (green squares) and syntaxial calcite (blue triangles) cements from a tectonic fracture cutting the infilling of the BC dyke. The points correspond to the values listed in Table 1. Dogtooth and syntaxial calcite cements cluster in two separate fields, indicating two different precipitation environments: meteoric-phreatic and freshwater superficial, respectively.

Table 1 Contents of selected elements (in ppm) in dogtooth cements (dtc) and overgrowing syntaxial calcite cements (scc) from the infilling of the fracture cutting the BC dyke (cf. Figs. 9a–c, 11)

|       | Ba  | Si  | Fe  | Sr  | Mn  | Mg  |
|-------|-----|-----|-----|-----|-----|-----|
| dtc   | 1080| 210 | 0   | 780 | 210 | 60  |
| scc   | 0   | 4960| 1600| 0   | 290 | 1660|
| dtc   | 1320| 870 | 0   | 220 | 20  | 1330|
| scc   | 370 | 4660| 1360| 350 | 460 | 1280|
| dtc   | 0   | 2160| 1540| 970 | 0   | 940 |
| scc   | 0   | 1390| 0   | 330 | 80  | 1150|
| dtc   | 0   | 1080| 830 | 0   | 1250| 810 |
| scc   | 810 | 1070| 0   | 910 | 470 | 1050|
| dtc   | 0   | 6980| 2090| 760 | 0   | 1150|
| scc   | 0   | 880 | 80  | 0   | 170 | 930 |
| dtc   | 1050| 1110| 1430| 0   | 780 | 1150|
| scc   | 0   | 670 | 30  | 90  | 460 | 1120|
| dtc   | 0   | 1280| 720 | 270 | 740 | 410 |

The boxes with bold outlines mark results obtained for cements in direct contact.

Conclusions

The mechanism of formation and infilling of fractures

Brachiopod-bearing levels appear sometimes in transgressive phases greatly influenced by synsedimentary tectonics (Fürsich et al. 2005; Seyed-Emami et al. 2008; Baeza-Carratalá and Sephihrinasab 2014). Despite some differences in development, it seems that all of the studied neptunian dykes were formed under similar conditions, in the seepage...
Table 2  Results of geochemical analyses of infillings of the MQ and BC dykes

| Sample | SiO₂ (%) | Al₂O₃ (%) | Fe₂O₃ (%) | MnO (%) | MgO (%) | CaO (%) | Na₂O (%) | K₂O (%) | TiO₂ (%) | P₂O₅ (%) | L.O.I (%) | Total (%) | Ba (ppm) | Sr (ppm) | Lithology   |
|--------|-----------|-----------|-----------|---------|---------|---------|----------|---------|----------|----------|----------|-----------|----------|----------|-------------|
| Młynka Quarry |
| M6-2  | 13.31     | 0.05      | 0.10      | 0.056   | 0.32    | 47.31   | 0.03     | <0.01   | 0.001    | 0.02     | 37.75    | 98.95     | 9        | 160       | Floatstone  |
| M7-1  | 22.42     | 0.07      | 0.13      | 0.101   | 0.26    | 42.26   | 0.04     | <0.01   | 0.001    | 0.01     | 33.00    | 98.30     | 13       | 149       | Floatstone  |
| Kluczwoda Valley |
| K1-1  | 68.50     | 0.15      | 0.33      | 0.014   | 0.10    | 17.62   | 0.03     | 0.02    | 0.005    | <0.01    | 12.37    | 99.15     | 113      | 92        | Silicified floatstone |
| K1-2  | 22.68     | 0.12      | 0.15      | 0.18    | 0.24    | 42.58   | 0.04     | 0.02    | 0.003    | 0.01     | 32.54    | 98.40     | 38       | 164       | Floatstone  |
| K2-1  | 13.91     | 0.13      | 0.11      | 0.022   | 0.28    | 47.73   | 0.03     | 0.02    | 0.006    | 0.03     | 37.42    | 99.69     | 21       | 171       | Floatstone  |
| K2-2  | 61.23     | 0.09      | 0.37      | 0.011   | 0.12    | 21.63   | 0.03     | 0.01    | 0.002    | <0.01    | 16.96    | 100.40    | 63       | 105       | Silicified floatstone |
| K3-1  | 4.41      | 0.10      | 0.14      | 0.019   | 0.35    | 53.11   | 0.03     | 0.01    | 0.003    | 0.05     | 41.78    | 99.99     | 9        | 179       | Floatstone  |
| K3-2  | 11.79     | 0.22      | 0.39      | 0.018   | 0.29    | 47.91   | 0.03     | 0.03    | 0.008    | 0.03     | 37.02    | 99.85     | 21       | 181       | Floatstone  |
| K4-1  | 15.42     | 0.18      | 0.38      | 0.017   | 0.26    | 46.67   | 0.03     | 0.03    | 0.003    | 0.01     | 37.42    | 99.99     | 21       | 171       | Floatstone  |
| K4-2  | 52.05     | 0.11      | 0.32      | 0.013   | 0.14    | 26.11   | 0.04     | 0.02    | 0.003    | <0.01    | 20.69    | 99.41     | 81       | 118       | Silicified floatstone |
| K5-2  | 54.19     | 0.15      | 0.29      | 0.17    | 0.14    | 25.65   | 0.04     | 0.02    | 0.004    | 0.01     | 20.07    | 100.60    | 87       | 128       | Silicified floatstone |
| K5-1  | 65.25     | 0.12      | 0.32      | 0.11    | 0.09    | 19.19   | 0.04     | 0.02    | 0.002    | <0.01    | 14.74    | 99.81     | 87       | 113       | Silicified floatstone |

Table 3  Results of Nd isotope analyses for the MQ and BC dykes

| Sample | Weight (mg) | Nd (ppm) | Sm (ppm) | Sm/Nd | εNd (T = 0) | εNd (T = 160) | T<sub>2DM</sub> (Ga) |
|--------|-------------|----------|----------|--------|-------------|---------------|------------------|
| Młynka Quarry |
| M6-2  | 307.50      | 1.00     | 0.18     | 0.18   | 0.1095      | 512192 ± 17    | −8.7            | −6.9 | 1.51 |
| M7-1  | 286.27      | 0.95     | 0.17     | 0.18   | 0.1078      | 512196 ± 13    | −8.6            | −6.8 | 1.50 |
| Kluczwoda Valley |
| K1-1  | 128.88      | 3.30     | 0.70     | 0.21   | 0.1288      | 512151 ± 10    | −9.5            | −8.1 | 1.61 |
| K1-2  | 268.25      | 3.48     | 0.69     | 0.20   | 0.1204      | 512182 ± 09    | −8.9            | −7.3 | 1.55 |
| K2-1  | 291.82      | 2.07     | 0.38     | 0.18   | 0.1109      | 512201 ± 09    | −8.5            | −6.8 | 1.50 |
| K2-2  | 247.44      | 3.09     | 0.61     | 0.20   | 0.1186      | 512156 ± 10    | −9.4            | −7.8 | 1.58 |
| K3-1  | 298.90      | 1.90     | 0.36     | 0.19   | 0.1133      | 512194 ± 10    | −8.7            | −7.0 | 1.52 |
| K3-2  | 278.30      | 2.17     | 0.40     | 0.19   | 0.1126      | 512228 ± 12    | −8.0            | −6.3 | 1.46 |
| K4-1  | 283.37      | 3.33     | 0.63     | 0.19   | 0.1143      | 512188 ± 10    | −8.8            | −7.1 | 1.53 |
| K4-2  | 174.00      | 2.99     | 0.57     | 0.19   | 0.1160      | 512166 ± 10    | −9.2            | −7.6 | 1.57 |
| K5-2  | 103.07      | 2.36     | 0.46     | 0.20   | 0.1190      | 512186 ± 16    | −8.8            | −7.2 | 1.54 |
zones of hydrothermal solutions located along the tectonic discontinuities active in the Oxfordian (Fig. 13a–c). The infilling mechanism of both the MQ and the BC dykes was also similar. Unfortunately, the insufficient number of samples from the GR dyke precludes more precise reconstruction of its infilling process, but because the degree of fragmentation of brachiopod shells is very low, this association records probably an autochthonous to parautochthonous community relict (cf. Kidwell 1991).

Stromatolites, which cover the vertical wall of the MQ dyke (Figs. 3, 5a), indicate a temporal hiatus between its opening and infilling (Fig. 13c, d). The lack of cavities filled with internal sediments, along with the similar development of the matrix enclosing the brachiopod shells and the internal sediments filling the shells, indicates that the sediment which filled the fractures was unlithified. Moreover, the presence of echinoderm plates with microbial envelopes (Fig. 5b, e) indicates that their growth upon echinoderm fragments took place under surface conditions, in unconsolidated sediment laid down upon the sea bottom prior to its redeposition into open fractures (Fig. 14). The coexistence of juvenile and adult brachiopods suggests favorable living conditions around the fractures controlled by the abundance of nutrients. In this depositional environment, some brachiopod shells were filled in situ with internal sediments after the death of the organisms and decomposition of their soft tissues. These internal sediments are identical with the deposits enclosing the shells. Moreover, the internal upper surfaces of the shells were covered with early diagenetic, radiaxial fibrous and bladed calcite cements (Fig. 5; cf. Kendall and Tucker 1973; Reinhold and Kaufmann 2010; Richter et al. 2011). Simultaneously, the crusts of isopachous granular cements grew on some tuberoids and bioclasts embedded within the sediment (Fig. 14).

The varying orientations of the geopetal infillings of brachiopod shells indicate their redeposition into open fissures together with unconsolidated sediment covering the sea bottom. Infilling of fractures was triggered by tectonic activity and submarine gravity flow (Fig. 13d), as revealed by the presence of angular clasts of massive limestones in the infilling material (Fig. 3). Deformations of brachiopod shells, i.e. their ‘stretching’ along the fracture walls, document the stage of vertical displacement of tectonic blocks along fractures filled with still-unlithified sediment (Fig. 13e).

The infilling of the BC neptunian dykes, similar to that of the MQ, was preceded by a temporal hiatus. The preserved fragments of stromatolites growing on the vertical wall of the fracture cutting the massive limestone (Fig. 7a) constitute evidence that the fracture remained empty for some time. However, the infilling of the BC dyke differs from that of the MQ dyke in terms of: (1) the widespread development of syntaxial calcite cement on echinoderm plates (Figs. 8a, e, 9d, 10e) and the simultaneous absence of microbial envelopes; (2) the sporadic development of typical early diagenetic radiaxial fibrous and bladed calcite cements within both the brachiopod shells and the enclosing sediment; and (3) the presence of numerous cavities filled with internal sediment (Figs. 8a, c, f, 10c). These differences seem to reflect a shorter period of influence of seawater and microorganisms living on the sea bottom on sediment and the partial lithification of the sediment prior to redeposition into the fractures.
Table 4  Fluid inclusion data

| Lp. | $T_h$ (°C) | S (wt%) | NaCl (wt%) | KCl (wt%) | CaCl$_2$ (wt%) | CO$_2$ or CH$_4$ |
|-----|----------|--------|-----------|--------|-------------|------------|
| Preparation K2-2 |
| 1   | 87       | 4.7    | 2.0       | ~0.3   | 2.4         | –          |
| 2   | 82       | 4.9    | 2.2       | ~0.5   | 2.2         | tr. CO$_2$ |
| 3   | 74       | 3.7    | 3.0       | –      | 0.7         | Low CO$_2$ |
| 4   | 70       | 3.7    | 2.8       | –      | 0.9         | Low CO$_2$ |
| 5   | 58       | 4.1    | 3.1       | –      | 1.0         | Low CH$_4$ |
| 6   | 52       | 4.1    | 3.1       | –      | 1.0         | Low CH$_4$ |
| 7   | 52       | 4.0    | 3.0       | –      | 1.0         | –          |
| 8   | 51       | 4.1    | 3.1       | –      | 1.0         | Low CH$_4$ |
| 9   | 51       | 4.3    | 3.0       | –      | 1.3         | –          |
| 10  | 51       | 4.5    | 3.1       | –      | 1.4         | –          |
| 11  | 49       | 4.0    | 3.0       | –      | 1.0         | Low CH$_4$ |
| 12  | 49       | 4.5    | 3.2       | –      | 1.3         | –          |
| 13  | 49       | 4.2    | 3.2       | –      | 1.0         | –          |
| 14  | 49       | 4.2    | 3.1       | –      | 1.1         | –          |
| Preparation K4-2 |
| 15  | 89       | 4.9    | 2.1       | ~0.3   | 2.3         | tr. CO$_2$ |
| 16  | 88       | 4.9    | 2.4       | ~0.5   | 2.2         | –          |
| 17  | 85       | 4.5    | 2.5       | ~0.3   | 2.0         | –          |
| 18  | 83       | 4.6    | 2.4       | ~0.3   | 1.9         | tr. CO$_2$ |
| 19  | 77       | 3.3    | 2.5       | –      | 0.8         | Low CO$_2$ |
| 20  | 76       | 3.3    | 2.3       | –      | 1.0         | Low CO$_2$ |
| 21  | 72       | 3.6    | 2.5       | –      | 1.1         | tr. CO$_2$ |
| 22  | 54       | 4.0    | 3.6       | –      | 1.2         | Low CH$_4$ |
| 23  | 51       | 4.2    | 3.2       | –      | 1.0         | Low CH$_4$ |
| 24  | 51       | 4.1    | 3.0       | –      | 1.1         | –          |
| 25  | 51       | 3.9    | 2.9       | –      | 1.0         | –          |
| 26  | 51       | 4.3    | 3.1       | –      | 1.2         | –          |
| 27  | 51       | 3.9    | 2.0       | –      | 1.0         | Low CH$_4$ |
| 28  | 50       | 3.9    | 3.0       | –      | 0.9         | –          |
| 29  | 50       | 4.0    | 3.1       | –      | 0.9         | –          |
| 30  | 50       | 3.8    | 2.8       | –      | 1.0         | –          |
| 31  | 49       | 4.2    | 3.0       | –      | 1.2         | –          |
| 32  | 49       | 3.9    | 3.0       | –      | 0.9         | –          |
| 33  | 49       | 4.0    | 2.8       | –      | 1.2         | –          |
| 34  | 49       | 3.8    | 3.0       | –      | 0.8         | –          |
| Preparation K5-2 |
| 51  | 84       | 5.0    | 2.2       | ~0.5   | 2.3         | –          |
| 56  | 81       | 4.8    | 2.0       | ~0.5   | 2.3         | tr. CO$_2$ |
| 37  | 75       | 3.6    | 2.9       | –      | 0.7         | Low CO$_2$ |
| 38  | 72       | 3.5    | 2.5       | –      | 1.0         | Low CO$_2$ |
| 39  | 71       | 3.6    | 2.5       | –      | 1.1         | Low CO$_2$ |
| 40  | 52       | 4.4    | 3.3       | –      | 1.1         | Low CH$_4$ |
| 41  | 52       | 4.2    | 3.0       | –      | 1.2         | –          |
| 42  | 51       | 4.4    | 3.3       | –      | 1.1         | Low CH$_4$ |
| 43  | 51       | 4.0    | 2.8       | –      | 1.2         | –          |
| 44  | 51       | 4.2    | 3.0       | –      | 1.2         | –          |
| 45  | 50       | 4.1    | 3.0       | –      | 1.1         | Low CH$_4$ |
| 46  | 50       | 4.0    | 3.0       | –      | 1.0         | –          |
| 47  | 50       | 3.8    | 2.9       | –      | 0.9         | –          |
The partial lithification of the sediment is also indicated by the widespread development of syntaxial calcite cements on fragments of echinoderms as well as by the presence of numerous cavities entirely or partly filled with internal sediment (Fig. 14). Crystallization of syntaxial calcite cement was possible due to: (1) the lack of significant activity by microorganisms which did not produce microbial envelopes over the fragments of echinoderms contained in sediment redeposited into the fractures; (2) the presence of empty spaces in the sediment redeposited into the fractures (cf. Longman 1980); and (3) the influx of ascending warm fresh waters (cf. Waldken and Berry 1984; James and Choquette 1984). These cements entirely filled or covered the walls of fine fractures from the deeper part of the fractures.

After redepocition of sediment into the fractures, the successive stages of diagenesis proceeded, as revealed by the crystallization of dogtooth cement (Fig. 8e, f) in a marine phreatic environment (cf. Reinhold 1999). Numerous cavities present in the partly lithified sediment which abruptly invaded the fractures were gradually infilled with internal sediment that trickled down into the fissure system (cf. Wall and Jenkyns 2004). The infilling of fractures was presumably driven by slow currents, which are indicated by locally preserved, horizontal lamination of internal sediment infilling the cavities (Fig. 10c; cf. Winterer et al. 1991; Winterer and Sarti 1994; Łuczynski 2001; Wall and Jenkyns 2004). Such lamination forms during the horizontal flow of water currents in the fissure and the simultaneous influx of material from above. The deposition rate of these sediments was low, as indicated by the dogtooth cement covering the edges of small cavities (Figs. 8a–d, 9a–c; cf. Wall and Jenkyns 2004). This is the second generation of cements growing locally on the walls of small cavities and fractures within the sediment and on radiaxial fibrous and bladed cements sporadically covering the internal surfaces of brachiopod shells (Fig. 14).

At the end of the Late Jurassic took place a subaerial exposure of the deposits. In meteoric vadose environment developed gravitative fibrous cements. Subsequent Early Cretaceous tectonic activity (cf. Matyszkiewicz et al. 2015a) was related to the action of hydrothermal solutions, which migrated along the rejuvenated fault zone (cf. Matyszkiewicz et al. 2015a). It was connected with crystallization of dogtooth cement in meteoric-phreatic environment (cf. Reinhold 1999) and probably with local silicification of sediment infilling the fractures (Figs. 8e, f, 14). In the first silicification stage, quartz with copious fluid inclusion crystallized in fine fractures and on the surfaces of cavity walls, replacing carbonate cements (Fig. 10c, d).

During the burial diagenesis, both the granular and the blocky calcite cements crystallized in some still-empty brachiopod shells (cf. James and Choquette 1984). These cements entirely filled or covered the walls of fine fractures resulting from tectonic displacements within the lithified sediments (Fig. 14).

In the Cenozoic, the sediments infilling the neptunian dykes became conduits for the circulation of karst waters migrating along the rejuvenated fault zones. Cracks, small cavities and fractures developed in the matrix; along these, minor displacements occurred. In our opinion, karst processes are responsible for the crystallization of Fe-enriched syntaxial calcite cement, which overgrew the dogtooth cement covering the surfaces of fractures (Table 1; Figs. 9a–c, 11, 14). This cement presumably precipitated from interstitial superficial waters (cf. Jiménez de Cisneros et al. 1990, 1991). Tectonic fractures in which cements did not crystallize were then filled with ferruginous internal sediment (Fig. 9a).

The crystallization of quartz in cavities and brachiopod shells which were not entirely filled with internal sediments proceeded in the first stage of silicification. We interpret second stage of silicification as an effect of the invasion of hydrothermal solutions triggered by episodes of tectonic activity in the Cenozoic (Fig. 14), during which silica-enriched hydrothermal solutions migrated along the reactivated faults (Gołubowska et al. 2010; Kochman and Matyszskiewicz 2012; Matyszskiewicz et al. 2015a). In the second stage, inclusion-free quartz crystallized in the still-empty centers of cavities and in the interiors of brachiopod shells (Figs. 8f, 10c–e). Unfortunately, it is not clear which silicification stage is responsible for the crystallization of microgranular and microflamboyant quartz (Fig. 8a).

Poor preservation of the GR neptunian dyke precludes more extensive interpretation of its infilling mechanism.

Table 4 continued

| Lp. | $T_h$ (°C) | S (wt%) | NaCl (wt%) | KCl (wt%) | CaCl$_2$ (wt%) | $CO_2$ or CH$_4$ |
|-----|-----------|---------|------------|----------|---------------|---------------|
| 48  | 49        | 4.1     | 3.3        | –        | 0.8           | Low CH$_4$    |
| 49  | 49        | 3.9     | 3.0        | –        | 0.9           | –             |
| 50  | 49        | 4.1     | 3.1        | –        | 1.0           | –             |
| 51  | 49        | 4.2     | 3.1        | –        | 1.1           | –             |
| 52  | 49        | 3.9     | 3.0        | –        | 0.9           | –             |

Dashes indicate content below detection limits

$T_h$ homogenization temperature, $S$ total salinity, tr. trace
Fig. 13 Reconstruction of formation conditions of Oxfordian neptunian dykes in the southern part of the Kraków-Częstochowa Upland (explanations in the text). Drawing by P. Klapyta
However, the complete state of the brachiopod shells and the absence of sorting suggest (Fig. 7d) that their relocation into open fractures was abrupt, perhaps involving living animals. The presence of coarse crystalline quartz indicates that fractures filled with brachiopods operated as conduits not only for karst waters but also for hydrothermal solutions.

### Structural setting and paleogeographical context

Undoubtedly, processes active in the Late Jurassic, such as the opening of fractures on the sea bottom and the faulting and redeposition of sediments by submarine gravity flows, were related to tectonic activity in an extensional stress field generated by transregional factors. The opening of both the Northern Atlantic and Tethys Oceans resulted in the Late Jurassic Pan-European stress-field reorganization, which also included the passive, northern margin of the Tethys (Ziegler 1990; Allenbach 2001, 2002; Nieto et al. 2012) together with the recent KCU. In the same period, Paleozoic faults were also rejuvenated, which gave rise to pulses of subsidence and localization of Late Jurassic depocenters over the Late Paleozoic tectonic grabens (Rioult et al. 1991; Færseth 1996; Pittet and Strasser 1998; de Wet 1998; Allenbach 2001, 2002; Wetzel et al. 2003; Krajewski et al. 2014, 2016). These processes continued in the Cretaceous and Neogene.

The recent geological structure of the southern part of KCU is mostly the effect of Neogene tectonic deformations that occurred when the KCU area was cut by numerous faults, generating a horst and graben system due to the load from overthrusting Carpathian nappes (Dźulynski 1953; Gradziński 1972). The presence of Oxfordian neptunian dykes along with gravity-flow deposits and synsedimentary faults indicates that at least a portion of these horsts and grabens already existed in the Late Jurassic (Matyszkwicz 1996; Matyszkwicz et al. 2007; Matyja and Ziolkowski 2014) and were subsequently subject to multiple reactivation, most recently during the Alpine movements. The strike of at least a portion of the faults bordering the Krzeszowice Graben follows the main fault zones in the Paleozoic basement. This is particularly clear in the eastern part of the Krzeszowice Graben, where its northern border corresponds to the course of the KLFZ (Fig. 1; cf. Buła and Habryn 2011; Habryn et al. 2014). It is interesting that the BC neptunian dyke is located over this zone and is characterized by a roughly similar azimuth. The Late Jurassic activity of faults framing the Krzeszowice Graben is indicated by: (1) the gravity flow deposits described from its southern and northern margins (Matyszkwicz 2014, 2016).

### Table

| sea floor | infilling of neptunian dykes |
|-----------|-----------------------------|
| marine phreatic | meteoric vadose |
| meteoric phreatic (?)hydrothermal | burial diagenesis |
| extension and faulting | rejuvenation of faults |

| Oxfordian | Late Jurassic–Early Cretaceous | Late Cretaceous | Cenozoic |
|-----------|-----------------------------|----------------|---------|
| microbial envelopes on echinoderms | | | |
| syntaxial overgrowth on echinoderms | | | |
| isopachous granular cement | | | |
| radiolarian fibrous cement | | | |
| bladed cement | | | |
| dogtooth cement | | | |
| gravitative fibrous cement | | | |
| granular cement | | | |
| blocky calcite cement | | | |
| quartz silification | | | |
| Fe-enriched syntaxial cement | | | |

### Fig. 14

Relative timing of diagenetic and tectonic events
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