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1 **Long-lasting Cadomian magmatic activity along an active northern**
2 **Gondwana margin: U–Pb zircon and Sr–Nd isotopic evidence from**
3 **the Brunovistulian Domain, eastern Bohemian Massif**

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17 **margin;** U–Pb geochronology; Geochemistry

18

19 **Running header:** Long-lasting Cadomian magmatism in the eastern Bohemian Massif

20

21 **Abstract**

22 Cadomian magmatic complexes of the Brunovistulian Domain crop out in the eastern
23 termination of the Bohemian Massif. However, the age, nature and geotectonic affinity of
24 some of pre-Variscan (meta-)igneous rock complexes from this domain are still unknown.
25 Geochronological and geochemical study of the granitic rocks across the Brunovistulian
26 Domain reveals new information about the timing and nature of this magmatic activity
27 originally situated along the northern margin of Gondwana. Zircon U–Pb data (601 ± 3 Ma,
28 Brno Massif ; 634 ± 6 Ma, paraautochthonous core of the Svatka Dome; 568 ± 3 Ma, Bíteš
29 orthogneiss from the allochthonous Moravicum indicate the prolonged magmatic activity
30 within the Brunovistulian Domain during the Ediacaran. The major- and trace-element and
31 Sr–Nd isotopic signatures show heterogeneous geochemical characteristics of the granitic
32 rocks and suggest a magmatic-arc geotectonic setting. The two-stage Depleted Mantle Nd
33 model ages (*c.* 1.3–2.0 Ga) indicate derivation of the granitic rocks from a relatively primitive
34 crustal source, as well as from an ancient and evolved continental crust of the Brunovistulian
35 Domain.

36 These results constrain the magmatic-arc activity to *c.* 635–570 Ma and provide a further
37 evidence for a long-lived (at least *c.* 65 Myr) and likely episodic subduction-related
38 magmatism at the northern margin of Gondwana. The presence of granitic intrusions derived
39 from variously mature crustal sources at different times suggests heterogeneous crustal
40 segments to having been involved in the magmatic-arc system during its multi-stage
41 evolution.

42

43 **Introduction**

44 The Variscan orogenic belt in Europe incorporates many of Cadomian magmatic complexes
45 produced by the Late Proterozoic–Cambrian igneous activity. The widespread and
46 voluminous arc-related magmatism was originally situated along the Andean-type active
47 margin of the Gondwana supercontinent and represents an important episode of crustal
48 growth within Europe and Western Asia (Nance et al. 1991, 2002; Murphy et al. 2002; von
49 Raumer et al. 2002; Pereira et al. 2011). Continental segments of the Cadomian belt including
50 the magmatic arc (Nance et al. 1991, 2002; Murphy et al. 2004) rifted off the northern
51 Gondwana margin during the Early Palaeozoic (Nance et al. 1991, 2010; Kemnitz et al. 2002;
52 Linnemann et al. 2008) and subsequently accreted to Laurussia during the Variscan Orogeny
53 (Franke 2000; Winchester et al. 2002). The overall extent and duration of the Cadomian
54 magmatic-arc activity is constrained, from localities scattered through the Variscan belt, to be
55 Neoproterozoic–Early Cambrian (Nance et al. 1991; von Raumer et al. 2002; Murphy et al.
56 2004; Linnemann et al. 2008).

57 Arc-related magmatic suites have been extensively reported from the Cadomian basement
58 in the Teplá–Barrandian Unit (Mašek and Zoubek 1980; Zulauf et al. 1997; Dörr et al. 1998,
59 2002; Sláma et al. 2008a; Hajná et al. 2010, 2013; Drost et al. 2011) and the Saxothuringian
60 Domain (Linnemann and Romer 2002; Linnemann et al. 2014) of the Bohemian Massif,
61 Iberian Massif (Fernández-Suárez et al. 2000; Bandres et al. 2002; Albert et al. 2015a, b;
62 Rubio-Ordóñez et al. 2015), the Eastern Pyrenees (Castiñeiras et al. 2008; Casas et al. 2015),
63 the Armorican Massif (D’Lemos et al. 1990; Strachan et al. 1996; Chantaine et al. 2001;
64 Gerdes and Zeh 2006), the Alps (Schaltegger et al. 1997; Neubauer et al. 2002; Schulz et al.
65 2004), the Tauride–Anatolide Platform (Ustaömer et al. 2005; Gürsu and Conczoglu 2008;
66 Şahin et al. 2014) and the Central Iranian Block (Moghadam et al. 2015).

67 The Brunovistulian Domain of the Bohemian Massif is generally assumed to be a
68 continental segment derived from the northern Gondwana margin (Dudek 1980; Finger et al.

69 2000a; Kalvoda et al. 2008). From the Late Devonian to Carboniferous (*i.e.* Variscan), the
70 Brunovistulian Domain was incorporated into the collision zone between Laurussia and peri-
71 Gondwana microcontinents (Matte et al. 1990; Franke 2000; Winchester et al. 2002;
72 Schulmann et al. 2009). Despite relatively well-constrained Cadomian formation of the
73 Brunovistulian Domain, the age, nature and geotectonic affinity of some bodies of pre-
74 Variscan (meta-)igneous rocks are still unknown, because of the scarcity of modern
75 geochronological and isotopic data from this area.

76 The aim of this paper is to present new U–Pb zircon ages of (meta-)granitic rocks from
77 three key parts of the Brunovistulian Domain in the south-western Moravia. The results of the
78 geochronological study are combined with Sr–Nd isotopic signatures and whole-rock
79 geochemistry. The newly acquired results are compared with previously published
80 geochronological and geochemical data from the Brunovistulian Domain in order to provide
81 better insight into the pre-collisional (pre-Variscan) evolution of the crystalline basement in
82 the easternmost part of the Variscan orogenic belt.

83

84 **Geological setting**

85 The eastern part of the Bohemian Massif (Fig. 1a, b) is traditionally subdivided into two
86 domains. The medium-grade Brunovistulian Domain (Suess 1926; Dudek 1980; Schulmann et
87 al. 1991, 1994; Kalvoda et al. 2008) in the east was underthrust beneath the high-grade
88 Moldanubian Domain to the west (Suess 1912; Dallmeyer et al. 1995; Franke 2000;
89 Schulmann et al. 2009) (Fig. 1b, c, 2).

90 Derivation of the Brunovistulian Domain from the northern margin of Gondwana is
91 generally accepted (Matte et al. 1990; Finger et al. 1995, 2000a; Friedl et al. 2000), but its
92 exact provenance remains controversial. The Brunovistulian Domain is mostly considered as
93 a continental segment of Avalonian (South American) affinity (Moczydlowska 1997; Tait et
94 al. 1997; Finger et al. 2000a; Friedl et al. 2000; Mazur et al. 2010) merged together with the

95 Moldanubian Domain during the Variscan collision. In contrast, several authors rather
96 suggested that it had a peri-Baltic affinity (Belka et al. 2002; Vavrdová et al. 2003; Kalvoda et
97 al. 2002, 2008; Nawrocki et al. 2004). The thrust boundary between the Moldanubian and
98 Brunovistulian domains is mostly assumed to be a remnant of an Early Palaeozoic ocean
99 (Höck et al. 1997; Finger et al. 1998) representing a southward curved continuation of the
100 Rheic suture (Finger et al. 1998; Murphy et al. 2006; Linnemann et al. 2008; von Raumer and
101 Stampfli 2008; Nance et al. 2010; Mazur et al. 2012) that can be traced further to the west
102 along the European Variscan belt. Some studies suggested a subduction zone to have been
103 located between the Moldanubian and the Brunovistulian domains (Matte et al. 1990; Höck et
104 al. 1997; Franke 2000; Konopásek et al. 2002; Finger et al. 2007) at least during the early
105 Variscan evolution. However, Schulmann et al. (2009, 2014) and Košler et al. (2014)
106 proposed a different idea, namely that the Moldanubian Domain represents just a rifted and
107 thinned marginal part of the Brunovistulian continental segment. This concept assumes that
108 these units were never separated by a large-scale oceanic domain and that the formation of the
109 Moldanubian Domain took place during the Early Palaeozoic and subsequent Variscan
110 orogenic evolution.

111 The Brunovistulian Domain itself is further subdivided into the Brunovistulicum *sensu*
112 *stricto* (*s.s.*), and the Moravicum (continuing as a Silesicum to the north; Hanžl et al. 2007b
113 and references therein) (Fig. 1b, c, 2), which differ from each other mainly in the degree of
114 the Variscan reworking. The strong Variscan deformation largely overprinted the primary
115 depositional/emplacement structures in these units and their mutual pre-Variscan position
116 remains a matter of discussion. During the Variscan Orogeny, the Moravicum nappe system
117 was thrust over the Brunovistulicum *s.s.* from the west (Suess 1912, 1926; Dudek 1980;
118 Schulmann et al. 1991). Moreover, most of the Brunovistulicum *s.s.* is covered by the
119 Devonian sedimentary rocks, the Variscan flysch sequences and the Cretaceous–Paleogene

120 sedimentary rocks of the Outer Carpathians from the East (Dudek 1980; Jelínek and Dudek
121 1993; Hladil et al. 1999).

122 The exposed parts of the Brunovistulicum *s.s.* are dominantly represented by the magmatic
123 rocks of the Brno and the Thaya (Dyje) massifs (Suess 1912; Dudek 1980; Finger et al. 1995,
124 2000a; Leichmann and Höck 2008) and by paraautochthonous metagranite body located in the
125 footwall of the Moravicum (the core of the Svatka Dome) (Fig. 1c, 2). Relicts of their host
126 pre-Cadomian basement (Dudek et al. 1980; Fritz et al. 1996) are preserved as blocks of
127 gneisses and migmatites.

128 The Brno Massif has been interpreted as a Cadomian rock assemblage built by the Western
129 and Eastern granitoid complexes (*e.g.*, Finger and Pin 1997; Hanžl and Melichar 1997) and by
130 a relict of ocean domain (the Central Basic Belt) sandwiched in between them (Hanžl and
131 Melichar 1997; Finger et al. 2000a, b; Leichmann and Höck 2008) (Fig. 1c, 2). The Western
132 Granitoid Complex consists of granites, granodiorites, diorites and also abundant blocks of
133 thermally affected host-rocks, whereas the Eastern Granitoid Complex is built mainly by
134 granodiorites, tonalities and quartz diorites. The Central Basic Belt contains low-grade
135 metamorphosed mafic plutonic and volcanic rocks.

136 The granitoids of the Thaya Massif are assumed to be a south-western continuation of the
137 Western Granitoid Complex of the Brno Massif (Finger et al. 1995, 2000a; Leichmann and
138 Höck 2008) off-set by the marginal fault of the Permian Boskovice Basin.

139 The core of the Svatka Dome is made up by a paraautochthonous metagranite body
140 (Souček et al. 1992; Hanžl et al. 2007a). Low-grade metamorphosed Devonian siliciclastic
141 sediments and limestones (Hladil et al. 1999) are incorporated within the metagranites as
142 narrow tectonic slices close to the overthrust Moravian nappes.

143 The rock association of the Brunovistulicum *s.s.* is generally considered as a product of a
144 subduction-related magmatism (Jelínek and Dudek 1993; Finger and Pin 1997). The Eastern

145 Granitoid Complex has been interpreted as a result of a primitive arc-related magmatism,
146 whereas the Western Granitoid Complex was more likely produced by melting of a pre-
147 existing continental crust (Hanžl and Melichar 1997; Finger et al. 2000a).

148 For the Central Basic Belt, a minimum age of 725 ± 15 Ma was proposed based on the Pb–
149 Pb zircon evaporation data from associated rhyolites (Finger et al. 2000b).

150 The Neoproterozoic intrusion of the Brno Massif granitoids is documented by the U–Pb
151 zircon age of 584 ± 5 Ma from the Western Granitoid Complex diorite (van Breemen et al.
152 1982) as well as by the $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages of amphiboles (586.9 ± 0.5 Ma from Eastern
153 Granitoid Complex and 596.9 ± 2.1 Ma from the Western Granitoid Complex diorites) (Fritz
154 et al. 1996). The metagranite from the Thaya Massif provided a Late Neoproterozoic U–Pb
155 zircon age of 575 ± 2 Ma (Friedl et al. 2004) and Rb–Sr whole-rock age of 551 ± 6 Ma
156 (Scharbert and Batík 1980).

157 The Moravicum constitutes a north–south elongated belt (Fig. 1c, 2) of deformed and
158 metamorphosed rocks. It is represented by the Svatka Dome and the Thaya (Dyje) Dome
159 anticlinal structures, forming tectonic windows (Suess 1912), and the Letovice Complex
160 (Höck et al. 1997; Soejono et al. 2010). The Moravicum, separated from the Moldanubian
161 Domain by the Micaschist Zone (Suess 1912), has been traditionally considered as a
162 deformed margin of the Brunovistulicum (Dudek 1980; Schulmann et al. 1991). Other
163 workers (*e.g.*, Winchester et al. 2006) proposed that the Moravicum is an independent
164 fragment of the Avalonian crust sandwiched between the strongly deformed and
165 metamorphosed Moldanubian Domain in the west and relatively undeformed rocks of the
166 Brunovistulicum *s. s.* in the east. The Moravicum was affected by the Variscan Barrovian-
167 type metamorphism (Höck 1995; Štípská and Schulmann 1995; Štípská et al. 2015), where the
168 metamorphic inversion was caused by imbrication of crustal nappes (Suess 1912; Štípská and
169 Schulmann 1995).

170 The Svatka and Thaya domes are made up by an assemblage of orthogneiss (Bíteš
171 orthogneiss in the Svatka Dome, Bíteš and Weitersfeld orthogneisses in the Thaya Dome)
172 and metapelite nappes (Dudek 1980; Schulmann et al. 1991; Štípská and Schulmann 1995)
173 thrust over the Brunovistulian basement (Fig. 1c, 2). The U–Pb zircon ages of *c.* 580 Ma from
174 the Bíteš orthogneiss (Friedl et al. 2000, 2004) and matching $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende cooling
175 age of 575.6 ± 2.2 Ma from associated amphibolite (Fritz et al. 1996) confirm its Cadomian
176 protolith age. The Letovice Complex is formed mainly by amphibolites and metagabbros
177 interpreted as a relict of an Early Cambrian (*c.* 530 Ma) incipient oceanic basin located
178 between the Brunovistulian and the Moldanubian domains incorporated into the Variscan
179 thrust-nappe system of the Moravicum (Soejono et al. 2010) (Fig. 1c, 2).

180

181 **Sample descriptions**

182 The sampling was focused on regionally important (meta-)igneous bodies within the
183 Brunovistulian Domain in the south-western Moravia (Fig. 1c). Our aim was to constrain their
184 intrusive ages, ages of inherited zircons and whole-rock geochemical characteristics. Samples
185 UD 3 and UD 5 were collected from the Brunovistulicum *s.s.* and sample UD 2 from the
186 Moravicum.

187 Biotite–amphibole granodiorite UD 3, from the abandoned quarry Anenský mlýn (WGS84
188 coordinates: N 49° 08.103', E 16° 31.906'), is a characteristic rock of the Western Granitoid
189 Complex of the Brno Massif. The granodiorite encloses abundant fine-grained mafic enclaves
190 of variable size and shape that display mixing/mingling textures. The sample UD 3 is
191 medium-grained, weakly porphyritic and exhibits a random or locally weak magmatic fabric
192 defined by the preferred orientation of K-feldspar phenocrysts (Fig. 3a). The granodiorite UD
193 3 consists of the biotite–amphibole–plagioclase–K-feldspar–quartz mineral assemblage.
194 Biotite and amphibole are locally chloritized and plagioclase is partly sericitized (Fig. 3b).
195 Accessory zircon, apatite and opaque minerals are also present.

196 Strongly deformed metagranite UD 5 was collected from the core of the Svatka tectonic
197 window south of Dolní Loučky (WGS84 coordinates: N 49° 21.405', E 16° 21.795'), and
198 represents granitoids of the Brunovistulian basement in the tectonic footwall of the
199 Moravicum nappe system. This fine-grained metagranite has only locally preserved igneous
200 texture and generally shows well-developed anastomosing solid-state foliation (Fig. 3c). The
201 metagranite UD 5 consists of mostly recrystallized quartz, K-feldspar and plagioclase with
202 fine-grained muscovite, scarce biotite and chlorite (Fig. 3d). Accessory minerals are zircon,
203 apatite and carbonate.

204 Sample UD 2 of the Bíteš orthogneiss from the vicinity of Štěpánov nad Svatkou (WGS84
205 coordinates: N 49° 30.143', E 16° 21.611'), represents typical meta-granitic lithology of the
206 Moravicum. The orthogneiss UD 2 is made mainly of quartz accompanied by the mineral
207 assemblage plagioclase–K-feldspar–muscovite–clinozoisite and small amount of biotite (Fig.
208 3e, f). Opaque mineral, apatite and zircon are the main accessories. Quartz and feldspars are
209 often recrystallized and feldspars replaced by sericite. The sample UD 2 shows fine- to
210 medium-grained porphyroclastic texture (Fig. 3e) and is characterized by a subhorizontal
211 high-grade foliation defined by alternation of recrystallized polymineralic plagioclase–K-
212 feldspar and quartz domains separated by bands of muscovite and biotite (Fig. 3f).

213 **Analytical techniques**

214 *Whole-rock geochemistry*

215 The whole-rock major- and trace-element analyses of magmatic rocks were determined in the
216 Acme Analytical Laboratories Ltd., Vancouver, by Inductively-Coupled Plasma Mass
217 Spectrometry (ICP–MS). Total abundances of the major- and minor-element oxides ('Group
218 4A') were determined by ICP-Emission Spectrometry (ICP-OES) following a $\text{LiBO}_2/\text{Li}_2\text{B}_4\text{O}_7$
219 fusion and dilute nitric digestion. Loss on ignition (LOI) was obtained by weigh difference
220 after heating to 1000 °C. The detection limits are 0.01 wt. % for most of the oxides, except

221 Fe₂O₃ (0.04 %), P₂O₅ (0.001 %) and Cr₂O₃ (0.002 %). Rare earth and refractory elements
222 were determined by ICP-Mass Spectrometry (ICP-MS) following a LiBO₂/Li₂B₄O₇ fusion and
223 nitric acid digestion of a 0.2 g sample ('Group 4B'). In addition a separate 0.5 g split was
224 digested in Aqua Regia and analysed by ICP-MS to report the precious and base metals (Pb,
225 Ni, Zn and Cu, 'Group 1DX'). See <http://www.acmelab.com> for details of the analytical
226 procedure and respective detection limits. Data management, recalculation, and plotting of the
227 whole-rock geochemical data were facilitated using *GCDkit* (Janoušek et al. 2006).

228 *Strontium–neodymium isotopic compositions*

229 For the radiogenic isotope determinations, samples were dissolved using a combined HF–
230 HCl–HNO₃ digestion. Strontium and bulk REE were isolated by exchange chromatography
231 techniques following the procedure of Pin et al. (1994) (PP columns filled with Sr. spec and
232 TRU. spec Eichrom resins, respectively). The Nd was further separated from the REE fraction
233 on PP columns with Ln. spec Eichrom resin (Pin and Zalduegui 1997). Complete analytical
234 details were reported by Míková and Denková (2007).

235 Isotopic analyses were performed on a Finnigan MAT 262 thermal ionization mass
236 spectrometer housed at the Czech Geological Survey in dynamic mode using a single Re
237 filament with Ta addition for Sr measurement and double Re filament assembly for Nd. The
238 ¹⁴³Nd/¹⁴⁴Nd ratios were corrected for mass fractionation to ¹⁴⁶Nd/¹⁴⁴Nd = 0.7219 (Wasserburg
239 et al. 1981), ⁸⁷Sr/⁸⁶Sr ratios assuming ⁸⁶Sr/⁸⁸Sr = 0.1194. External reproducibility was
240 estimated from repeat analyses of the BCR-1 (¹⁴³Nd/¹⁴⁴Nd = 0.512621 ± 20 (2σ, n = 5) and
241 NBS 987 (⁸⁷Sr/⁸⁶Sr = 0.710248 ± 28 (2σ), n = 10) isotopic standards. The Rb, Sr, Sm and Nd
242 concentrations were obtained by ICP-MS in Acme Laboratories (see above).

243 The decay constants applied to age-correct the isotopic ratios are from Steiger and Jäger
244 (1977: Sr) and Lugmair and Marti (1978: Nd). The ϵ_{Nd}^i values were obtained using Bulk
245 Earth parameters of Jacobsen and Wasserburg (1980) (¹⁴⁷Sm/¹⁴⁴Nd_{CHUR} = 0.1967 and present-

246 day $^{143}\text{Nd}/^{144}\text{Nd}_{\text{CHUR}} = 0.512638$), the two-stage Depleted Mantle Nd model ages ($T_{\text{Nd}}^{\text{DM}}$) were
247 calculated after Liew and Hofmann (1988) ($^{147}\text{Sm}/^{144}\text{Nd}_{\text{DM}} = 0.219$, present-day
248 $^{143}\text{Nd}/^{144}\text{Nd}_{\text{DM}} = 0.513151$, average crustal $^{147}\text{Sm}/^{144}\text{Nd}_{\text{CC}} = 0.12$).

249 *Laser ablation ICP-MS U–Pb zircon dating*

250 About twenty kilograms of the fresh rock were crushed, sieved, and zircon grains were
251 separated from the samples using the Wilfley shaking table and heavy liquids, mounted in
252 epoxy-filled blocks and polished. Zoning patterns in individual grains were observed, and
253 presence of older inherited components checked, by cathodoluminescence detector mounted
254 on the electron microprobe at the Institute of Petrology and Structural Geology, Charles
255 University in Prague.

256 The U–Pb and Pb–Pb zircon ages were obtained using two different laser-ablation (LA)
257 ICP-MS analytical protocols at the University of Bergen, Norway:

258 a) Isotopic analysis of zircon by laser ablation ICP-MS followed the technique described in
259 Košler et al. (2002) and Košler and Sylvester (2003). A Thermo-Finnigan Element 2 sector
260 field ICP-MS coupled to a 213 nm solid state Nd-YAG laser (NewWave UP213) at Bergen
261 University, Norway, was used to measure Pb/U and Pb isotopic ratios in zircons. The sample
262 introduction system was modified to enable simultaneous nebulisation of a tracer solution and
263 laser ablation of the solid sample (Horn et al. 2000). Natural Tl ($^{205}\text{Tl}/^{203}\text{Tl} = 2.3871$; Dunstan
264 et al. 1980), ^{209}Bi and enriched ^{233}U and ^{237}Np (>99%) were used in the tracer solution, which
265 was aspirated to the plasma in an argon–helium carrier gas mixture through an Apex
266 desolvation nebuliser (Elemental Scientific) and a T-piece tube attached to the back end of the
267 plasma torch. A helium gas line carrying the sample from the laser cell to the plasma was also
268 attached to the T-piece tube. The laser was fired at a repetition rate of 5 Hz and energy of 80
269 mJ. Linear laser rasters (30–100 microns) were produced by repeated scanning of the laser
270 beam at a speed of 10 microns/second across the zircon sample surface. Typical acquisitions

271 consisted of 40 second measurement of blank followed by measurement of U and Pb signals
272 from the ablated zircon for another 110 seconds. The data were acquired in time resolved –
273 peak jumping – pulse counting mode with 1 point measured per peak for masses 202
274 (flyback), 203 (Tl), 204 (Pb), 205 (Tl), 206 and 207 (Pb), 209 (Bi), 233 (U), 237 (Np), 238
275 (U), 249 (^{233}U oxide), 253 (^{237}Np oxide) and 254 (^{238}U oxide). Raw data were corrected for
276 dead time of the electron multiplier and processed offline in a spreadsheet-based program
277 (Lamdate; Košler et al. 2002). Data reduction included correction for gas blank, laser-induced
278 elemental fractionation of Pb and U and instrument mass bias. Minor formation of oxides of
279 U and Np was corrected for by adding signal intensities at masses 249, 253 and 254 to the
280 intensities at masses 233, 237 and 238, respectively. No common Pb correction was applied to
281 the data but the low concentrations of common Pb were checked by observing $^{206}\text{Pb}/^{204}\text{Pb}$
282 ratio during measurements. Residual elemental fractionation and instrumental mass bias were
283 corrected by normalization to the natural zircon reference material GJ-1 (Jackson et al. 2004).
284 Zircon reference material 91500 (Wiedenbeck et al. 1995) was periodically analysed during
285 the measurement for quality control and the obtained mean value of 1065 ± 5 (2σ) Ma
286 corresponds with the published reference value of *c.* 1065 Ma (Wiedenbeck et al. 1995).

287 b) A Nu AttoM high resolution ICP-MS coupled to a 193 nm ArF excimer laser
288 (Resonetics RESolution M-50 LR) at Bergen University, Norway, was used to measure the
289 Pb/U and Pb isotopic ratios in zircons. The laser was fired at a repetition rate of 5 Hz and
290 energy of 80 mJ with 19 microns spot size. Typical acquisitions consisted of 15 second
291 measurement of blank followed by measurement of U and Pb signals from the ablated zircon
292 for another 30 seconds. The data were acquired in time resolved – peak jumping – pulse
293 counting mode with 1 point measured per peak for masses $^{204}\text{Pb} + \text{Hg}$, ^{206}Pb , ^{207}Pb , ^{208}Pb ,
294 ^{232}Th , ^{235}U , and ^{238}U . Due to a non-linear transition between the counting and attenuated (= analogue) acquisition modes of the ICP instruments, the raw data were pre-processed using a

296 purpose-made Excel macro. As a result, the intensities of ^{238}U are left unchanged if measured
297 in a counting mode and recalculated from ^{235}U intensities if the ^{238}U was acquired in an
298 attenuated mode. Data reduction was then carried out off-line using the Iolite data reduction
299 package version 3.0 with VizualAge utility (Petrus and Kamber 2012). Full details of the data
300 reduction methodology can be found in Paton et al. (2010). The data reduction included
301 correction for gas blank, laser-induced elemental fractionation of Pb and U and instrument
302 mass bias. For the data presented here, blank intensities and instrumental bias were
303 interpolated using an automatic spline function while down-hole inter-element fractionation
304 was corrected using an exponential function. No common Pb correction was applied to the
305 data but the low concentrations of common Pb were checked by observing $^{206}\text{Pb}/^{204}\text{Pb}$ ratio
306 during measurements. Residual elemental fractionation and instrumental mass bias were
307 corrected by normalization to the natural zircon reference material Plešovice (Sláma et al.
308 2008b). Zircon reference materials GJ-1 (Jackson et al. 2004) and 91500 (Wiedenbeck et al.,
309 1995) were periodically analysed during the measurement for quality control and the obtained
310 mean values of 599.9 ± 2.1 (2σ) Ma and 1063.0 ± 3.2 (2σ) Ma are accurate within the
311 published reference values (600.5 ± 0.4 Ma, Schaltegger et al. 2015; 1065 Ma, Wiedenbeck et
312 al. 1995, respectively). The zircon U–Pb ages are presented as concordia diagrams generated
313 with the ISOPLOT program v. 3.6 (Ludwig 2008).

314 **Results**

315 *Whole-rock geochemistry*

316 The geochemical data from the samples UD 3, UD 5 and UD 2 were compared with
317 previously published chemical analyses of granitic rocks from the Western and Eastern
318 granitoid complexes of the Brno Massif (Hanžl and Melichar 1997; Leichmann and Höck
319 2008), as well as metagranites and mylonites from the paraautochthonous basement of the

320 Svatka Dome and the Bíteš orthogneiss nappe (Moravicum) (Souček et al. 1992; Hanžl et al.
321 2007a).

322 *Major elements*

323 The three studied samples are subalkaline granites to granodiorites as demonstrated by the
324 multielement R_1 – R_2 plot (De La Roche et al. 1980) (Fig. 4a) as well as by the Total Alkalis–
325 Silica (TAS) diagram (Cox et al. 1979) (Fig. 4b). Both metagranite UD 5 and orthogneiss UD
326 2 are moderately peraluminous ($A/CNK = 1.12$, Table 1) in contrast to subaluminous (A/CNK
327 $= 1.04$) granodiorite from the Brno Massif. This is also documented by the multielement B–A
328 diagram (Debon and Le Fort 1983) modified by Villaseca et al. (1998) (Fig. 4c). In the binary
329 SiO_2 – K_2O plot (Peccerillo and Taylor 1976) (Fig. 4d), the samples UD 5 and UD 2 classify as
330 (normal-K) calc-alkaline to high-K calc-alkaline, while the Brno Massif granodiorite (UD 3)
331 is distinctly potassic. In all four classification diagrams the newly analyzed samples fall close
332 to the fields defined by the previously published compositions of the Western Granitoid
333 Complex and Moravicum orthogneisses, as appropriate. The only exception is the
334 granodiorite sample (UD 3) being enriched in K_2O in the SiO_2 – K_2O plot (Fig. 4d).

335 Major-element composition of all three new samples is silicic ($SiO_2 = 69.2$ – 71.6 wt. %)
336 (Fig. 5) with variable K_2O/Na_2O ratio ranging from 0.71 in orthogneiss (UD 2) through 1.33
337 of metagranite (UD 5) to 1.55 in relatively potassic granodiorite (UD 3). The major-element
338 compositions of all three new samples plot within the compositional range of the Brno Massif
339 granitoids, metagranites and mylonites of the Svatka Dome core and the Bíteš orthogneisses.
340 They often show a negative correlation of SiO_2 with major- and minor-element oxides (TiO_2 ,
341 Al_2O_3 , FeO , MgO and CaO).

342 *Trace elements*

343 The trace-element patterns, in spider plot normalized by average composition of the upper
344 continental crust (Taylor and McLennan 1995), are very similar to each other and show

345 mostly trends close to the upper crustal average, with distinct troughs in Th, U, Nb, and Ta and
346 perceptible depletion in HREE (Fig. 6a). All three samples are also depleted in P; the
347 orthogneiss UD 2 shows in addition spikes in Ba and Sr.

348 Chondrite-normalized (Boynnton 1984) REE patterns (Fig. 6b) are also very similar in all
349 three samples, featuring moderate enrichment in LREE ($La_N/Yb_N = 15.3\text{--}44.3$, $La_N/Sm_N =$
350 $4.5\text{--}10.4$; Table 2) with weak depletion in HREE. Typical of metagranite UD 5 and
351 orthogneiss UD 2 are weak negative Eu anomalies ($Eu/Eu^* = 0.59$ and 0.84 , respectively),
352 while the Brno Massif granodiorite UD 3 displays a distinctly positive one ($Eu/Eu^* = 1.26$),
353 perhaps reflecting feldspar(s) accumulation.

354 Both types of multielement patterns for the studied samples best fit within the variability of
355 Moravicum orthogneisses, but also within the Western Brno Massif granodiorites (Fig. 6).

356 In addition, the Zr concentrations in all three samples were used to determine zircon
357 saturation temperatures (Watson and Harrison 1983), which should provide a maximum
358 constraint upon the magma temperature. The calculated temperatures are $825\text{ }^\circ\text{C}$ for
359 granodiorite UD 3, $760\text{ }^\circ\text{C}$ for the protolith of the metagranite UD 5 and $749\text{ }^\circ\text{C}$ for the
360 protolith of the orthogneiss UD 2.

361 *Sr–Nd isotopic data*

362 The Sr–Nd isotopic compositions of the three samples, both raw and age-corrected to their
363 respective intrusive ages, are summarized in the Table 3. The Sr–Nd isotopic composition of
364 the Western Brno Massif granodiorite UD 3 is the most primitive, close to the Bulk Earth
365 ($^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7048$; $\varepsilon_{Nd}^i = -1.0$). The Bíteš orthogneiss UD 2 on the other hand reflects a
366 generation from a mature crustal source ($^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7101$; $\varepsilon_{Nd}^i = -10.0$); the
367 paraautochthonous metagranite UD 5 falls between these two extremes ($^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7052$;
368 $\varepsilon_{Nd}^i = -3.7$). The variability in the Nd isotopic compositions is directly reflected by the two-

369 stage Depleted Mantle Nd model ages (T_{Nd}^{DM}) (granodiorite UD 3: 1.33 Ga, metagranite UD 5:
370 1.57 Ga, Bíteš orthogneiss UD 2: 2.01 Ga).

371 *Laser ablation ICP-MS U–Pb zircon dating*

372 Three samples (UD 3, UD 5, UD 2) from the main parts of the Brunovistulian Domain were
373 dated using LA-ICP-MS U–Pb method on zircon. Isotopic data with corresponding ages are
374 given in Online Resource. The oscillatory zoning observed in cathodoluminescence (CL)
375 images of zircon grains from all samples (Fig. 7) corresponds to the crystallization from melt.

376 The zircon grains from the granodiorite of the Brno Massif (UD 3) are transparent, pale
377 brown to colourless and generally have long-prismatic habitus. In CL images, most grains are
378 euhedral and oscillatory zoned (Fig. 7a). A total of 44 analyses were performed in the sample
379 UD 3, of which 31 were used. Dating of sample UD 3 yielded a concordia age of 601 ± 3 Ma
380 (2σ , Fig. 8a), interpreted as the Late Proterozoic intrusive age of the granodiorite. No
381 inherited zircon cores were either observed in CL images or were detected by the LA-ICP-MS
382 analyses.

383 Zircon population from the metagranite UD 5 (the Brunovistulian basement in the core of
384 the Svatka Dome) is heterogeneous and consists of pale brown to light pink, euhedral and/or
385 subhedral grains. Cathodoluminescence images show mostly euhedral oscillatory growth
386 zoning and infrequent inherited cores (Fig. 7b). From sample UD 5, 19 analyses were
387 performed, of which only 11 were concordant. These concordant analyses combine into a
388 concordia age of 634 ± 6 Ma (2σ , Fig. 8b), interpreted as the magmatic crystallization age of
389 this metagranite. The single analysis of c. 1670 Ma is interpreted as a xenocrystic core while
390 the detection of one c. 400 Ma zircon most probably reflects Pb loss during metamorphism.

391 Zircon population of the Bíteš orthogneiss UD 2 (Moravicum's nappe) contains generally
392 prismatic, euhedral, colourless to pale pink grains, mostly with oscillatory zoning. Corroded
393 and rounded inherited cores are very common (Fig. 7c). CL-bright outer rims, possibly related

394 to a recrystallization or new zircon growth, were also found in some of the grains (Fig. 7c).
395 However, their rims were too thin to be dated by LA-ICP-MS. A total of 76 analyses were
396 performed in the sample UD 2, of which 74 were used. Analyses placed outside the cores
397 combine into a concordia age of 568 ± 3 Ma (2σ , Fig. 8c), interpreted as the Late Proterozoic
398 crystallization age of the Bíteš orthogneiss magmatic protolith. Dating of the frequent
399 inherited cores shows a range of ages between *c.* 1.1 and 2.1 Ga and two dates at *c.* 2.5 Ga
400 and *c.* 2.7 Ga (Fig. 9).

401 **Discussion**

402 *Age of Cadomian magmatism in the Brunovistulian Domain*

403 The new LA-ICP-MS U–Pb zircon age of 601 ± 3 Ma for the Western Granitoid Complex of
404 the Brno Massif (granodiorite UD 3) is significantly older than the conventional U–Pb zircon
405 age of 584 ± 5 Ma for a diorite of the same geological unit dated by van Breemen et al.
406 (1982). The Ar–Ar hornblende dating of the Eastern Granitoid Complex diorite (south of
407 Blansko) also gave a younger age of 586.9 ± 0.5 Ma (Fritz et al. 1996) as did the metagranite
408 from the Thaya Massif (575 ± 2 Ma: U–Pb zircon age of Friedl et al. 2004). On the other
409 hand, our age agrees remarkably well with the Ar–Ar hornblende age of 596.9 ± 2.1 Ma
410 obtained from a diorite, also from the UD 3 locality (Anenský mlýn quarry), by Fritz et al.
411 (1996), indicating a relatively rapid cooling.

412 The new 634 ± 6 Ma LA-ICP-MS U–Pb zircon age for the metagranite UD 5 represents
413 both the first available geochronological datum from the paraautochton in the core of the
414 Svratka Dome and also the oldest yet known intrusive age for granitoids of the
415 Brunovistulicum *s.s.*

416 The protolith of the Bíteš orthogneiss UD 2 from the allochthonous Moravicum has also
417 Late Proterozoic, albeit significantly younger, intrusive age of 568 ± 3 Ma that correlates well
418 with the U–Pb ages reported from the Moravicum and the Silesian Domain (van Breemen et

419 al. 1982; Friedl et al. 2000, 2004; Kröner et al. 2000; Oberc-Dziedzic et al. 2003; Mazur et al.
420 2010, 2012), the Teplá–Barrandian Unit (Dörr et al. 2002; Hajná et al. 2013) as well as from
421 the Saxothuringian Domain (Linnemann et al. 2014).

422 Taken together, our new *in situ* U–Pb zircon data confirm the generally held idea of
423 Cadomian origin of the Brunovistulian Domain (Dudek 1980; Finger et al. 2000a; Leichmann
424 and Höck 2008). In particular, they bring further evidence for a long-lived Late Proterozoic
425 (Ediacaran) arc-related magmatic activity within the Brunovistulian Domain (at least *c.* 635–
426 575 Ma). This is largely in line with the published 630–530 Ma K–Ar cooling ages (Dudek
427 and Melková 1975; recalculated to the new K decay constants of Steiger and Jäger 1977) of
428 the samples from deep boreholes drilled into the eastern part of the Brunovistulian Domain
429 concealed under the sediments of the Carpathian Foredeep in SE Moravia.

430 *Prospective sources of the Cadomian magmas*

431 The studied (meta-)granitic rocks, and literature data from the same units, show major- and
432 trace-element characteristics resembling calc-alkaline, continental magmatic arc-related
433 granites (Fig. 10). However, new whole-rock geochemical and Sr–Nd isotopic signatures
434 provide an evidence that the granodiorite UD 3 of the western part of the Brno Massif and the
435 metagranite UD 5 from the footwall of the Moravicum (Brunovistulicum *s.s.*) originated by
436 partial melting of a geochemically less evolved source, whereas the orthogneiss UD 2 of the
437 Moravicum was generated from an ancient, mature crustal segment.

438 By their chemistry, the samples from the Brunovistulicum *s.s.* (UD 3 and UD 5) resemble
439 I- or transitional I/S-type granite suites. Most typically, the sample UD 3 is subaluminous,
440 less siliceous (granodioritic) and thus falling into the field of low-peraluminous granites in
441 Fig. 4c. Moreover, it is depleted in Th, U, Nb and Ta if compared with typical upper crustal
442 compositions (Fig. 6a). Its Sr–Nd isotopic composition is the most primitive of the studied
443 samples, close to the Bulk Earth ($^{87}\text{Sr}/^{86}\text{Sr}_i = 0.705$; $\varepsilon_{Nd}^i = -1.0$). It falls just at the least

444 evolved limit of the Sr–Nd isotopic data ($^{87}\text{Sr}/^{86}\text{Sr}_i = 0.705\text{--}0.710$; $\varepsilon_{Nd}^i = -1.0$ to -7.0) from
445 the Western Granitoid Complex (aka Thaya Terrane) of Finger and Pin (1997). The
446 neodymium in the paraautochthonous metagranite UD 5 is somewhat less radiogenic
447 ($^{87}\text{Sr}/^{86}\text{Sr}_i = 0.705$; $\varepsilon_{Nd}^i = -3.7$).

448 Still, the most characteristic features of the Western Granitoid Complex are the Sr–Nd
449 isotopic compositions at the other end of the spectrum, resembling mature continental crust
450 (Finger et al. 2000a). Based on these observations, as well as on elevated silica and potassium
451 contents, the same authors assumed mostly metasedimentary source of the granitic magmas,
452 with only limited participation of the juvenile lower crustal lithologies or mantle-derived
453 magmas. The latter notion is also supported by field observations of mingling between mafic
454 and felsic magmas in the Anenský mlýn quarry (UD 3).

455 The rather evolved chemistry of the Western Granitoid Complex contrasts with the much
456 more primitive Sr–Nd isotopic signature of the Eastern Granitoid Complex (Slavkov Terrane):
457 $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.704\text{--}0.705$; $\varepsilon_{Nd}^i = -1.0$ to $+3.0$ (Finger et al. 2000a) implying a geochemically
458 little evolved source of the granitic magmas, perhaps young calc-alkaline rocks, and/or
459 significant mantle contribution. Worth noting in this context, however, is that the
460 metasedimentary lithologies in this unit also show a CHUR-like isotopic signature ($^{87}\text{Sr}/^{86}\text{Sr}_i$
461 $= 0.704\text{--}0.706$; $\varepsilon_{Nd}^i = -1.0$ to $+2.0$; unpublished data of Finger and Pin cited by Finger et al.
462 2000a).

463 On the other hand, the Bíteš gneiss UD 2 is a typical S-type granite (see *e.g.*, high SiO_2 ,
464 elevated A/CNK, thus falling into field of felsic peraluminous granites in Fig. 4c, as well as
465 abundance of inherited zircon cores). A viable genetic model is a partial melting of a mature,
466 ancient crustal source ($^{87}\text{Sr}/^{86}\text{Sr}_i = 0.7101$; $\varepsilon_{Nd}^i = -10.0$). Similar ε_{Nd}^i values of -10 to -11 ,

467 corresponding to two-stage Nd model ages (T_{Nd}^{DM}) of *c.* 2 Ga, were previously reported from
468 analogous orthogneiss samples by Liew and Hofmann (1988) and Finger et al. (2000a).

469 *Remarks on correlation of the Brunovistulicum s.s. and the Moravicum*

470 The granitic magmas of the Brunovistulicum *s.s.* most likely originated from relatively
471 immature crustal material and have little or no inherited zircon component. On the contrary,
472 the protolith of orthogneiss from the Moravicum was derived from a more evolved continental
473 crust rich in older detritus, especially of Mesoproterozoic and Palaeoproterozoic age. These
474 results cast doubts on the concept that the orthogneisses of the Moravicum represent just
475 deformed equivalents of the Brunovistulicum *s.s.* granitoid rocks (Dudek 1980; Schulmann et
476 al. 1991, 1994). Instead, detected differences suggest that the rocks of these two units
477 represent products of melting of distinct crustal sources with potentially different provenance.
478 In any case, the geochemical characteristics place the magma source of all studied lithologies
479 into an evolved continental arc setting.

480 However, a small number of studied samples does not allow proper description of
481 magmatic-arc evolution that resulted in the formation of the Brunovistulian Domain rock
482 assemblage. The age and geochemical span of the obtained data corresponds to an episodic
483 magmatic activity observed within both ancient and modern continental magmatic-arc
484 systems (Paterson and Ducea 2015) and to variation in chemical composition caused by
485 continental magmatic-arc dynamics (Ducea et al. 2015).

486 *Significance of inherited zircon age populations*

487 Almost no inherited zircon ages were detected in the samples from the Brunovistulicum *s.s.*
488 either due to lack of older zircons in their source(s) or the fact that they did not survive the
489 partial melting event. The spectrum of Mesoproterozoic, Palaeoproterozoic and Neoproterozoic
490 ages obtained from the zircon cores in the Bíteš orthogneiss sample UD 2 (Moravicum) could

491 be interpreted as a recycled population of zircons from the melted source, which has been
492 likely of sedimentary origin. The Nd model age of *c.* 2.0 Ga for the Bíteš orthogneiss UD 2
493 corresponds with the age spectrum obtained from inherited zircon cores of the same sample (*n*
494 = 6; Fig. 9) and indicates recycling of a mainly *c.* 2 Ga old crustal component, possibly
495 reflecting the Palaeoproterozoic (Eburnean orogenic phase of the West African Craton) event
496 commonly described within the Moldanubian Domain (Kröner et al. 1988; Wendt et al. 1993;
497 Friedl et al. 2004; Janoušek et al. 2010; Košler et al. 2014), its unmetamorphosed equivalent
498 the Teplá–Barrandian Unit (Strnad and Mihaljevič 2005; Drost et al. 2007) and the
499 Saxothuringian Domain (Linnemann and Romer 2002; Linnemann et al. 2014). This fact
500 would suggest that the Bíteš orthogneiss nappe has been derived from the Moldanubian
501 Domain. On the other hand, detailed provenance analysis of detrital zircons in
502 metasedimentary rocks from the Moldanubian Domain and the Moravicum (Košler et al.
503 2014) indicated that the protoliths of these rocks were deposited in separate basins, yet
504 spatially related prior to the Variscan Orogeny.

505 The absence of Tonian and Cryogenian ages (for a general review of zircon age spectra of
506 the Cadomian complexes see discussion in Dörr et al. 2015) could exclude Minoan and
507 Armorican terranes as a source area of the studied rock.

508 The Meso- and Palaeoproterozoic zircon cores age populations (well defined peaks
509 between *c.* 1.2 and 2.4 Ga) from some of the metaigneous complexes in NE Austria and SW
510 Poland (*e.g.*, Bíteš and Strzelin gneisses; Friedl et al. 2000, 2004; Oberc-Dziedzic et al. 2003;
511 Mazur et al. 2010, 2012) are similar to our detected inherited age spectrum in the Bíteš
512 orthogneiss (Fig. 9), and also to the previously published detrital zircon age data from the
513 Moravicum (Košler et al. 2014). These age populations can be correlated with the orogenic
514 events reported from the Amazonian cratonic province (Cardona et al. 2009; McLelland et al.
515 2010). In contrast, the abundance of gneisses derived from early Palaeozoic granitic protoliths

516 together with lack of Mesoproterozoic and Palaeoproterozoic inheritance are considered as
517 evidence of the North African affinity typical of the Armorican terranes (Linnemann et al.
518 2004, 2008; Samson et al. 2005). On this basis, the whole Brunovistulian Domain has been
519 correlated with the South American (Avalonian) part of Gondwana (Friedl et al. 2000, 2004;
520 Mazur et al. 2010).

521 However, broadly similar zircon populations from the Neoproterozoic sedimentary rocks
522 were found also in several parts of Baltica (*e.g.*, Kuznetsov et al. 2010; Bingen et al. 2011) as
523 well as within terranes belonging to the Trans European Suture Zone (Łysogóry and
524 Małopolska massifs) (Valverde-Vaquero et al. 2000; Nawrocki et al. 2007). Presence of key
525 Meso- and Palaeoproterozoic zircon ages within different continental segments (both
526 Avalonia and Baltica) indicates that only zircon age data themselves are not useful to
527 unequivocally distinguish the provenance of the Brunovistulian Domain. The original position
528 of the Brunovistulian Domain during Cadomian Orogeny still remains uncertain. Further
529 detailed geochronological studies of zircon inherited cores from magmatic rocks or detrital
530 zircons from sedimentary rocks could shed more light on this issue.

531 *Geotectonic implications*

532 The new age data combined with the geochemical signatures from all the studied parts of the
533 Brunovistulian Domain suggest their origin in the same continental arc setting.

534 The obtained time span of the protolith crystallization ages could mean that the continental
535 arc-related magmas were created in course of a long-lasting Late Neoproterozoic episodic
536 magmatic activity within the Brunovistulian Domain. Long duration of the magmatic system
537 could have enabled an involvement of heterogeneous crustal components – as indicated by
538 variable Nd model ages – that may have come from spatially distant domains. Such a
539 persistent Cadomian subduction zone activity has been proposed along the whole active

540 northern margin of the Gondwana supercontinent (period of main magmatism at *c.* 635–570
541 Ma; Murphy et al. 2004; Nance and Linnemann 2008).

542 The long-lasting widespread subduction-related magmatism has been over last 15 years
543 reported from many Cadomian basement complexes. The presence of arc-derived clastic
544 material in the Cadomian accretionary wedge-type sequences of the Teplá–Barrandian Unit
545 indicates voluminous arc-related magmatic activity at *c.* 610–560 Ma (Dörr et al. 2002; Sláma
546 et al. 2008a; Hajná et al. 2013). Detrital zircon age spectra from the Cadomian basement of
547 the Saxothuringian Domain show arc-type magmatism in the interval of *c.* 750–570 Ma
548 (Linnemann et al. 2014).

549 The long-lived Late Proterozoic magmatic arc along the northern Gondwana margin has
550 been inferred for the Iberian Massif (Pereira et al. 2011; Albert et al. 2015a; Rubio-Ordóñez et
551 al. 2015) and the Eastern Pyrenees (Casas et al. 2015). The Neoproterozoic arc-related
552 magmatic activity with the time span between *c.* 630 and 550 Ma is also well documented in
553 the orogenic belts of the West Gondwana not involved into younger orogens. These are for
554 example the Dom Feliciano–Kaoko Belt (see summary in Konopásek et al. 2016), Ribeira
555 Belt (Heilbron and Machado 2003) and the Araçuaí Belt (Tedeschi et al. 2016) today exposed
556 along the east coast of Brazil. Moreover, the similar time interval of arc-related magmatism
557 has been reported from the Timanides in the NE margin of Baltica (western continuation of
558 the Cadomian orogen in Neoproterozoic; *e.g.*, Pease et al. 2004; Kuznetsov et al. 2007). The
559 existence of Early Cambrian magmatic arc along the northern margin of East Gondwana is
560 documented in the eastern Mediterranean region (Romano et al. 2004; Dörr et al. 2015), the
561 Western Pontides (Şahin et al. 2014) and the Central Iranian Block (Moghadam et al. 2015).

562 The view that all three studied parts of the Brunovistulian Domain belonged to the same
563 magmatic arc is, however, challenged by the presence of Mesoproterozoic to
564 Palaeoproterozoic inherited zircon ages in the Bíteš orthogneiss, by the lack of inherited

565 zircon ages in the Brunovistulicum *s.s.* granitoids, and by different Depleted Mantle Nd
566 model ages of the whole-rock samples. Winchester et al. (2006) pointed out the fact that the
567 Mesoproterozoic zircon ages obtained from the orthogneisses belong to the Moravicum, but
568 not to the Brunovistulicum *s.s.* and suggested the possibility of former independence of the
569 Moravicum basement from the Brunovistulicum *s.s.* Nevertheless, the lack of old zircon cores
570 can be simply caused by rare occurrence of inheritance typical of relatively hot and less
571 siliceous granites with I-type or mixed I/S-type affinity (Miller et al. 2003; Janoušek 2006 and
572 references therein). This is in line with the particularly high zircon saturation temperature of
573 825 °C calculated for the granodiorite UD 3.

574 The scenario of two independent crustal segments showing continental arc magmatism
575 would indicate existence of an oceanic suture between the Moravicum and the
576 Brunovistulicum *s.s.* Several studies from this region nonetheless render the existence of such
577 a large ocean basin unlikely (absence of relicts of the ocean floor-related rocks and/or
578 evidence for HP–LT metamorphism) (Schulmann et al. 1991; Hanžl et al. 2007a).
579 Occurrences of the metamorphosed Devonian continental and marine sedimentary rocks
580 sandwiched between the Brunovistulicum *s.s.* and the Moravicum (Schulmann et al. 1991;
581 Hanžl et al. 2007a) has been interpreted as remnants of small basins rather than of a subducted
582 extensive ocean domain (Hladil et al. 1999). The Devonian lithospheric extension regime,
583 reported by Kalvoda et al. (2008) from the Brunovistulicum *s.s.*, led more likely to the
584 development of an attenuated continental crust with narrow segments of oceanic crust in the
585 marginal parts of the Brunovistulian Domain.

586 The concept that the Brunovistulicum *s.s.* and the Moravicum belonged to the same
587 continental segment seems most likely, even though the independent provenance of both units
588 cannot be completely excluded. Our data confirm that the Brunovistulian Domain was a part
589 of an Andean-type active margin formed along the northern border of the Gondwana

590 supercontinent (Fig. 11a) after its final amalgamation (Nance et al. 1991, 2002; Murphy et al.
591 2004). The northern Gondwana margin-derived continental fragments rifted off the
592 supercontinent mainland (Fig. 11b) during the Cambrian–Ordovician extensional event
593 (Murphy et al. 2006; Linnemann et al. 2008; von Raumer and Stampfli 2008; Žák et al. 2013).
594 Position and evolution of the Brunovistulian Domain during most of the Early Palaeozoic are
595 unclear until the regional Devonian extension (Hladil et al. 1999; Kalvoda et al. 2008) and
596 subsequent incorporation into the Variscan collision (Schulmann et al. 2009, 2014; Štípská et
597 al. 2015; see Fig. 11c for details).

598 Comparisons of age and duration of magmatic-arc activity and its possible sources in the
599 Brunovistulian Domain with both adjacent (Teplá–Barrandian Unit and Saxothuringian
600 Domain) and more distant (West Gondwana, NE margin of Baltica *etc.*) Cadomian basements
601 reveal significant similarities. These findings indicate their common evolution and probable
602 spatial relations during the Ediacaran. Moreover they underline a global importance of the
603 widespread Neoproterozoic–Cambrian arc-related magmatic activity throughout Peri-
604 Gondwana and Baltica terranes for a crustal growth. However, the questions of detailed
605 evolution and mutual configuration of different Cadomian basements remain open and thus
606 represent a challenge for future research.

607

608 **Conclusions**

609 The presence of granitic rocks in the Brunovistulian Domain with Late Proterozoic
610 (Ediacaran) crystallization ages provides a further evidence for a long-lived, voluminous and
611 widespread Cadomian magmatic activity at the northern margin of Gondwana. Geochemical
612 fingerprints show that the granitic rocks of the Brunovistulicum *s.s.* and the Moravicum were
613 formed in a continental magmatic-arc environment. The range of crystallization ages indicates
614 episodic magmatic activity within this arc, which has been active for at least 65 Myr during

615 Late Proterozoic (*c.* 635–570 Ma). The whole-rock geochemical character, Sr–Nd isotopic
616 signatures and, in the case of the Bíteš orthogneiss, abundance of zircon inheritance seem to
617 indicate distinct crustal sources and melting conditions for each of the studied intrusions.

618 The parental magmas were generated by partial melting of independent segments of a
619 continental crust characterized by variable two-stage Depleted Mantle Nd model ages (*c.* 1.3–
620 2.0 Ga). Variability of the sources was most likely caused by different and originally distant
621 portions of the continental crust involved into the long-lasting magmatic-arc system. Both
622 parts of the Brunovistulian Domain (Brunovistulicum *s.s.* and the Moravicum) together with
623 other temporally related complexes within the Variscan belt formed segments of the northern
624 Gondwana margin until the Early Palaeozoic times, when they were rifted off during the
625 supercontinent break-up. These Cadomian continental segments were finally amalgamated
626 during the Early Carboniferous Variscan collision.

627

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636

637 **Figure captions**

638 **Fig. 1** a – Position of the Bohemian Massif (BM) within Europe. b – Generalized geological
639 map of the Bohemian Massif (modified after Franke 2000). SX: Saxothuringian
640 Domain; TB: Teplá–Barrandian Unit; MD: Moldanubian Domain; BD: Brunovistulian
641 Domain; SZ: Silesicum. The solid rectangle represents the studied area. c – Simplified
642 geologic map of the Brunovistulian Domain (modified from the geologic map of the
643 Czech Republic, 1:500,000; Cháb et al. 2007).

644 **Fig. 2** Mutual position of individual units within the Brunovistulian Domain in the schematic
645 geological cross-section along the line A–B (location shown in Fig. 1). Stars indicate
646 approximate locations of the geochemical and geochronological samples.

647 **Fig. 3** Macro and microphotographs of studied samples from the Brunovistulian Domain. a, b
648 – Brno Massif granodiorite UD 3; c, d – Svatka Dome metagranite UD 5; e, f – Bíteš
649 orthogneiss UD 2.

650 **Fig. 4** Classification diagrams for the (meta-)igneous rocks of the Brunovistulian Domain
651 showing samples UD 2, UD 3, UD 5 and literature data (semi-transparent symbols;
652 Souček et al. 1992; Hanžl and Melichar 1997; Hanžl et al. 2007a; Leichmann and Höck
653 2008). Brno Massif data do not include the A-type Hlína suite (Hönig et al. 2014). a –
654 Multielement plot R_1 – R_2 (De La Roche et al. 1980). b – Modified SiO_2 – $\text{Na}_2\text{O} + \text{K}_2\text{O}$
655 (TAS; in wt. %) classification diagram proposed by Cox et al. (1979) for plutonic rocks.
656 The discrimination boundary between the subalkaline and alkaline domains is after
657 Irvine and Baragar (1971). c – Multielement B–A diagram (Debon and Le Fort 1983)
658 modified by Villaseca et al. (1998). l-P: low-peraluminous, m-P: moderately
659 peraluminous, h-P: highly peraluminous, f-P: felsic peraluminous. d – SiO_2 – K_2O plot

660 (wt. %) with the discrimination boundaries between the tholeiitic, calc-alkaline, high-K
661 calc-alkaline and shoshonitic rocks of Peccerillo and Taylor (1976).

662 **Fig. 5** Binary plots of silica vs. selected major- and minor-element oxides (wt. %). Data
663 designation as in the Figure 4.

664 **Fig. 6** Multielement diagrams for the samples of the granitic rocks from the Western
665 Granitoid Complex of the Brno Massif, Bíteš orthogneiss from the Moravicum, and
666 metagranites from the core of the Svatka Dome. The semi-transparent background
667 fields are defined by the literature data from the same units (Souček et al. 1992; Hanžl
668 and Melichar 1997; Hanžl et al. 2007a; Leichmann and Höck 2008). a – Average Upper
669 Continental Crust (Taylor and McLennan 1995) normalized spider plot. b – Chondrite-
670 normalized (Boynnton 1984) REE patterns.

671 **Fig. 7** Cathodoluminescence images of typical zircon grains extracted from the studied
672 samples. Laser spots and $^{206}\text{Pb}/^{238}\text{U}$ ages with 2σ uncertainties for UD 2, UD 3 and laser
673 rasters and $^{206}\text{Pb}/^{238}\text{U}$ ages with 2σ uncertainties for UD 5 are marked. Laser spot-size
674 was 19 μm for UD 2 and UD 3 and 14–24 μm for UD 5.

675 **Fig. 8** U–Pb concordia diagrams and calculated concordia ages (in blue) for magmatic zircons
676 (LA-ICP-MS data). **a** – Brno Massif granodiorite UD 3; **b** – Svatka Dome
677 metagranite UD 5; **c** – Bíteš orthogneiss UD 2. n – number of used analyses (more
678 than 90% concordance)/total number of analyses.

679 **Fig. 9** Probability density plot (bin width = 65 Ma) showing $^{206}\text{Pb}/^{238}\text{U}$ zircon ages (LA-ICP-
680 MS data) from the Bíteš orthogneiss UD 2 from the Moravicum (ISOPLOT; Ludwig
681 2008).

682 **Fig. 10** Geotectonic discrimination diagrams for the (meta-)igneous rocks of the
683 Brunovistulian Domain. Data designation as in the Figure 4. **a** – Multielement plot Al–
684 $\text{Fe}^{\text{T}} + \text{Ti} - \text{Mg}$ of Jensen (1976) showing the calc-alkaline (CA) character of the studied
685 rocks. TH: tholeiitic series. **b** – Th–Zr/117–Nb/16 ternary diagram (Wood 1980). N-
686 MORB: normal-type mid-oceanic ridge basalts; E-MORB (WPT): enriched mid-
687 oceanic ridge basalts (within-plate tholeiites); WPA: within-plate alkali basalts; CAB:
688 calc-alkaline basalts; IAT: island-arc tholeiites. **c** – Binary plot Y–Nb (Pearce et al.
689 1984). ORG: Ocean Ridge Granites, VAG: Volcanic Arc Granites, syn-COLG:
690 Collision Granites, WPG: Within Plate Granites. **d** – Binary plot Y + Nb–Rb (Pearce
691 et al. 1984) (the same abbreviations).

692 **Fig. 11** Schematic sketches demonstrating proposed tectonic evolution of the Brunovistulian
693 Domain. **a** – Cadomian magmatic-arc stage at the Late Proterozoic; **b** – Early
694 Cambrian initiation of the Gondwana margin break-up (“back-arc” position of the
695 Letovice Complex is assumed); **c** – Variscan continental collision at the Early
696 Carboniferous; **References:** (1) this study, (2) Kemnitz et al. (2002), (3) Linnemann et
697 al. 2008, (4) Sláma et al. 2008a, (5) Nance et al. (2010), (6) Finger et al. (1998), (7)
698 Mazur et al. (2012), (8) Soejono et al. (2010), (9, 10), Schulmann et al. (1991, 1994),
699 (11) Štípská and Schulmann (1995), (12) Štípská et al. (2015).

700 **Table captions**

701 **Table 1** Major-element whole-rock geochemical analyses (wt. %)

702 **Table 2** Trace-element whole-rock geochemical analyses (ppm)

703 **Table 3** Sr–Nd isotopic data

704

705 **Electronic supplementary material**

706 Laser ablation ICP-MS U–Pb zircon data

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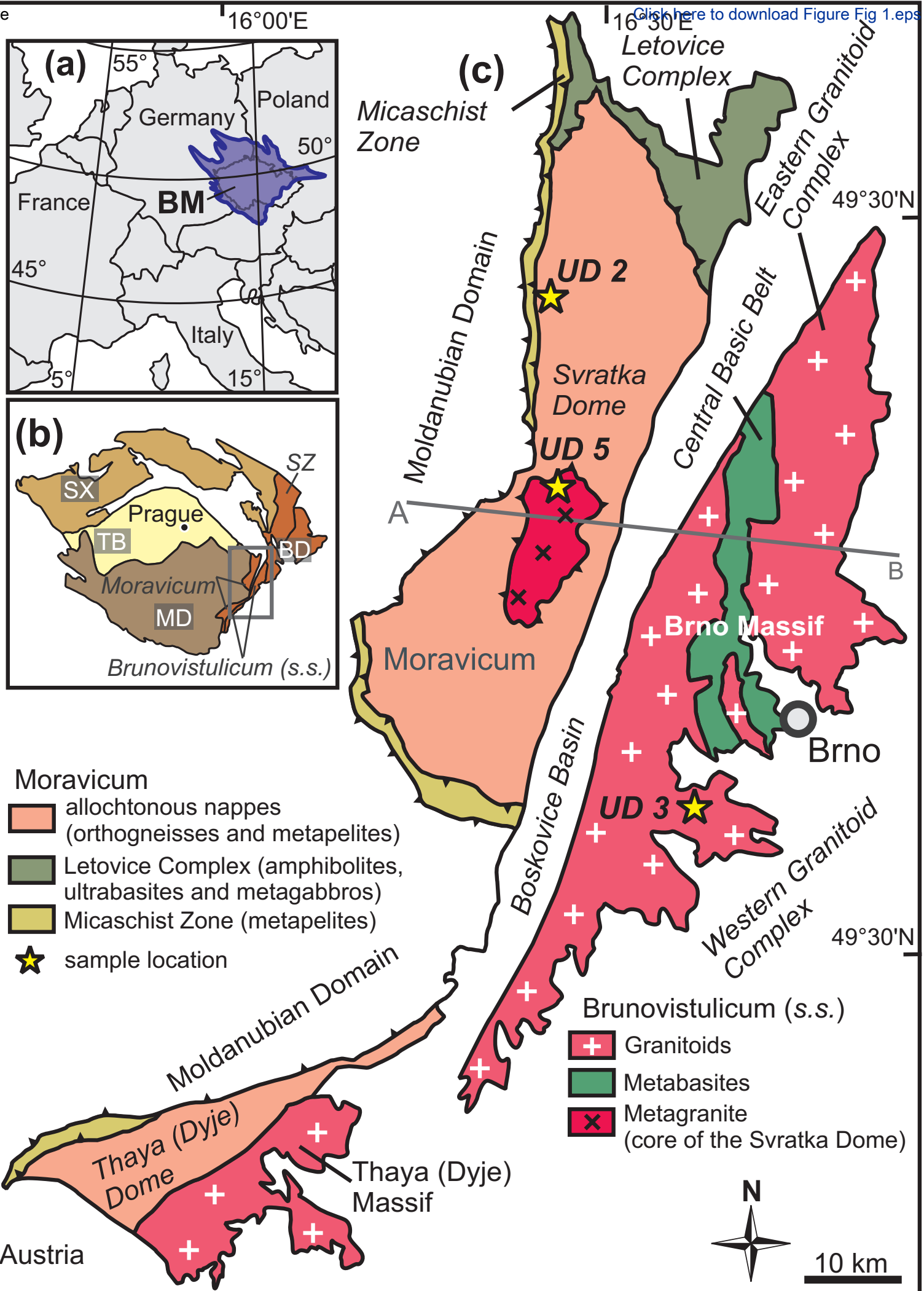
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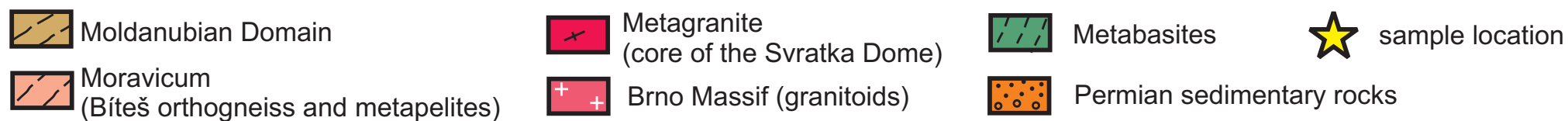
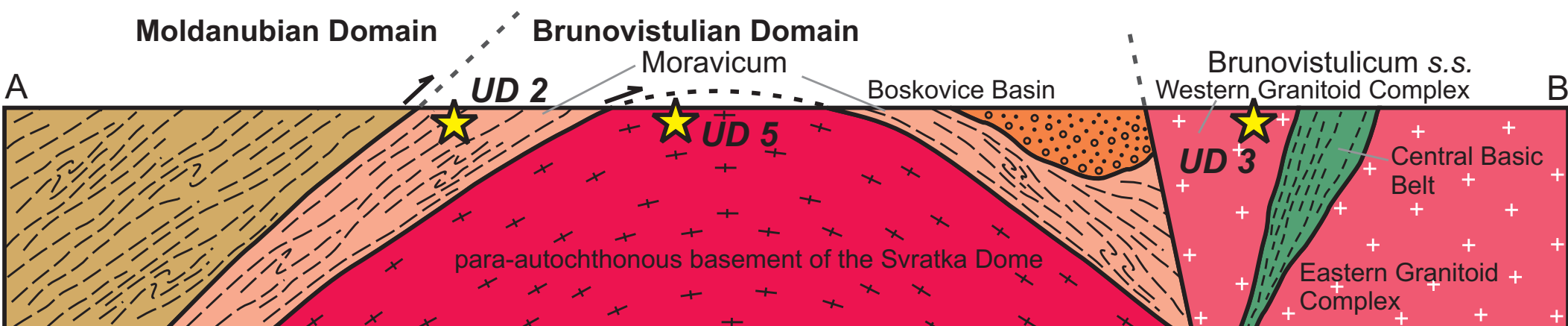
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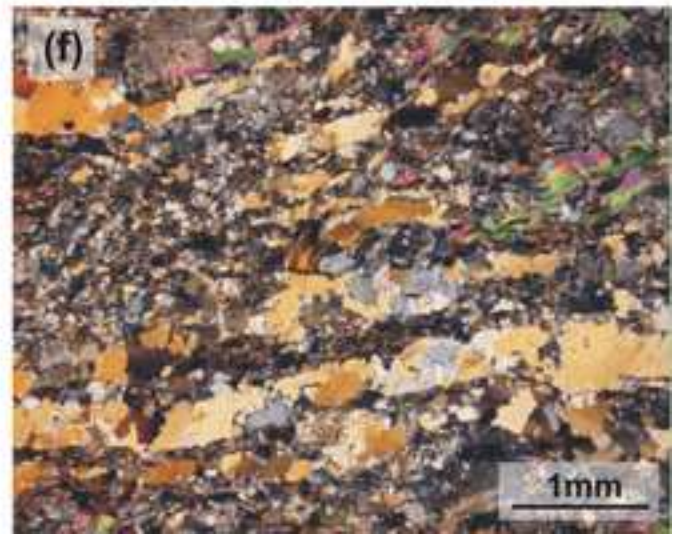
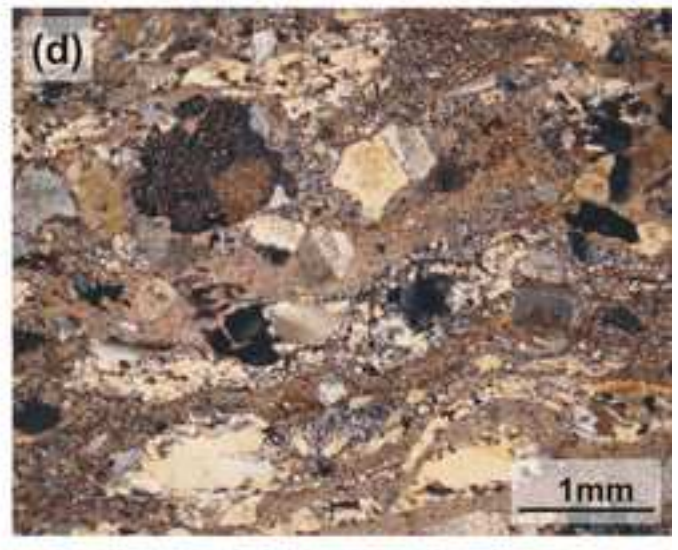
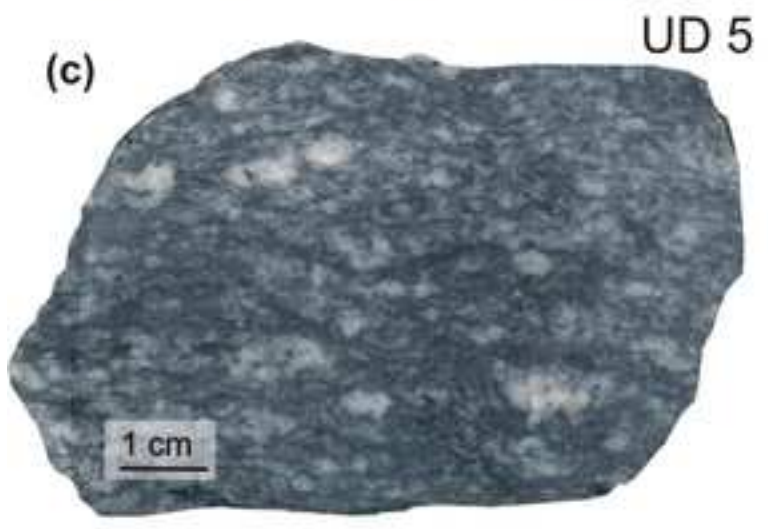
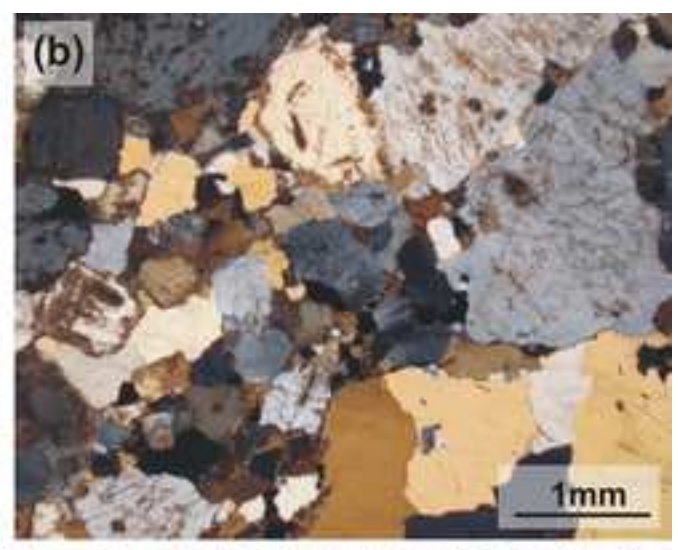
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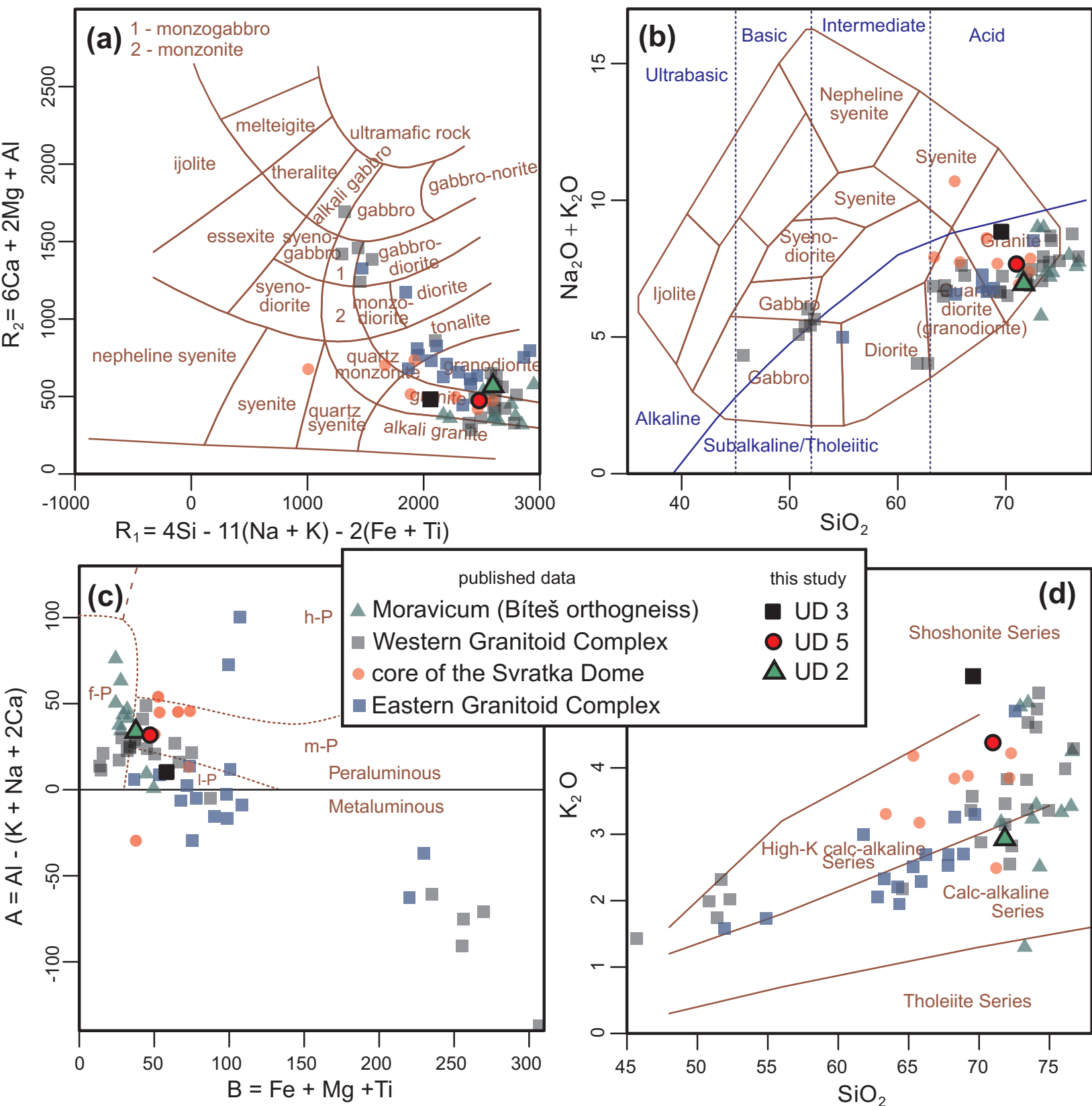
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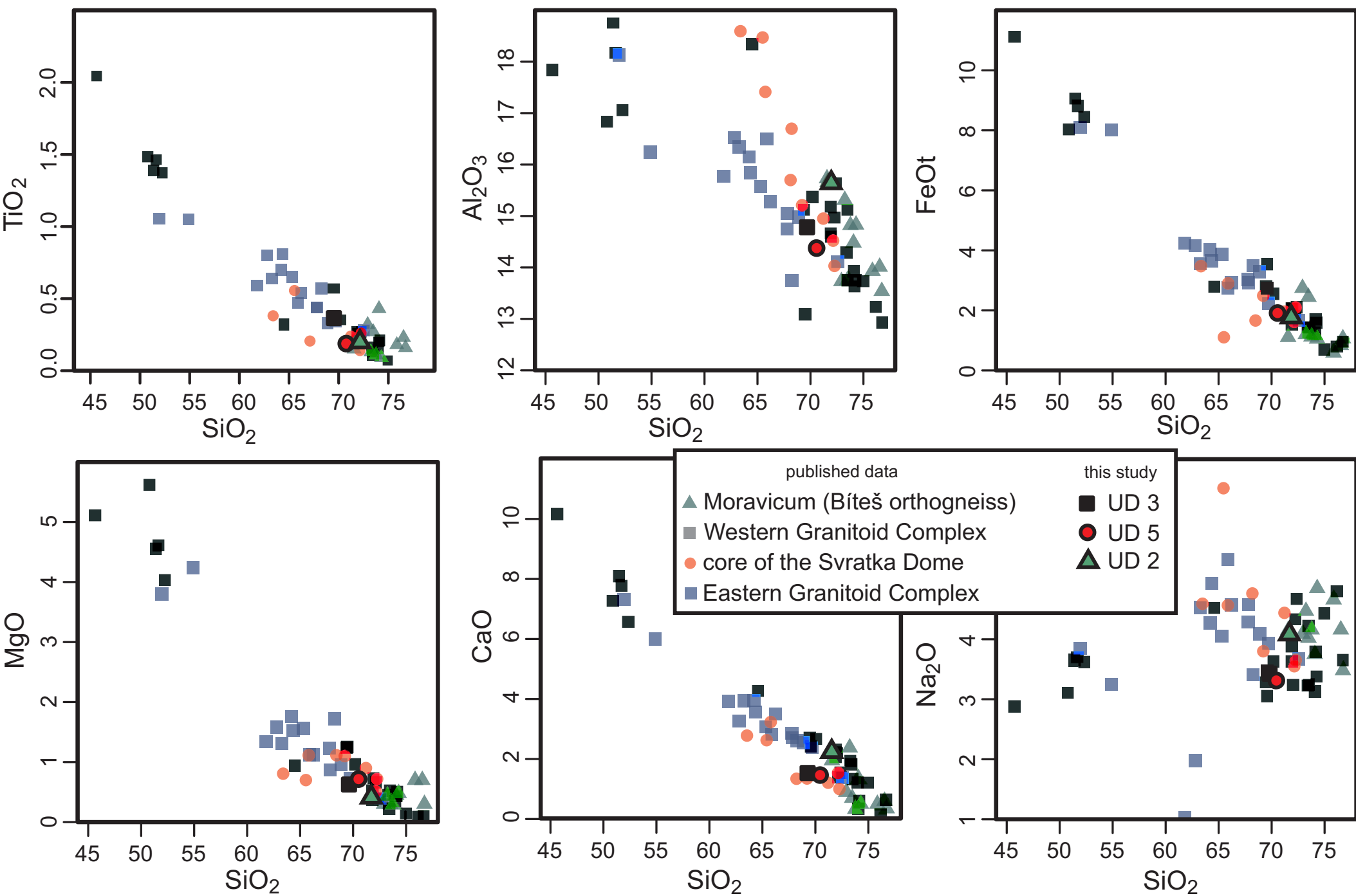
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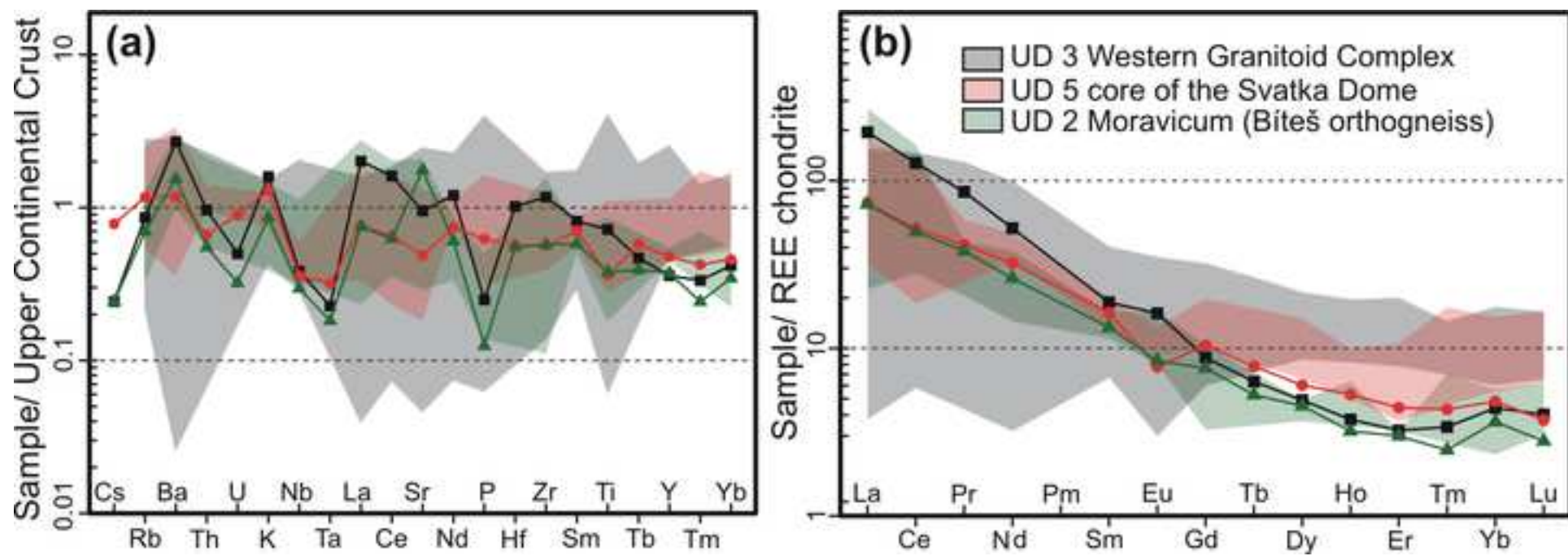


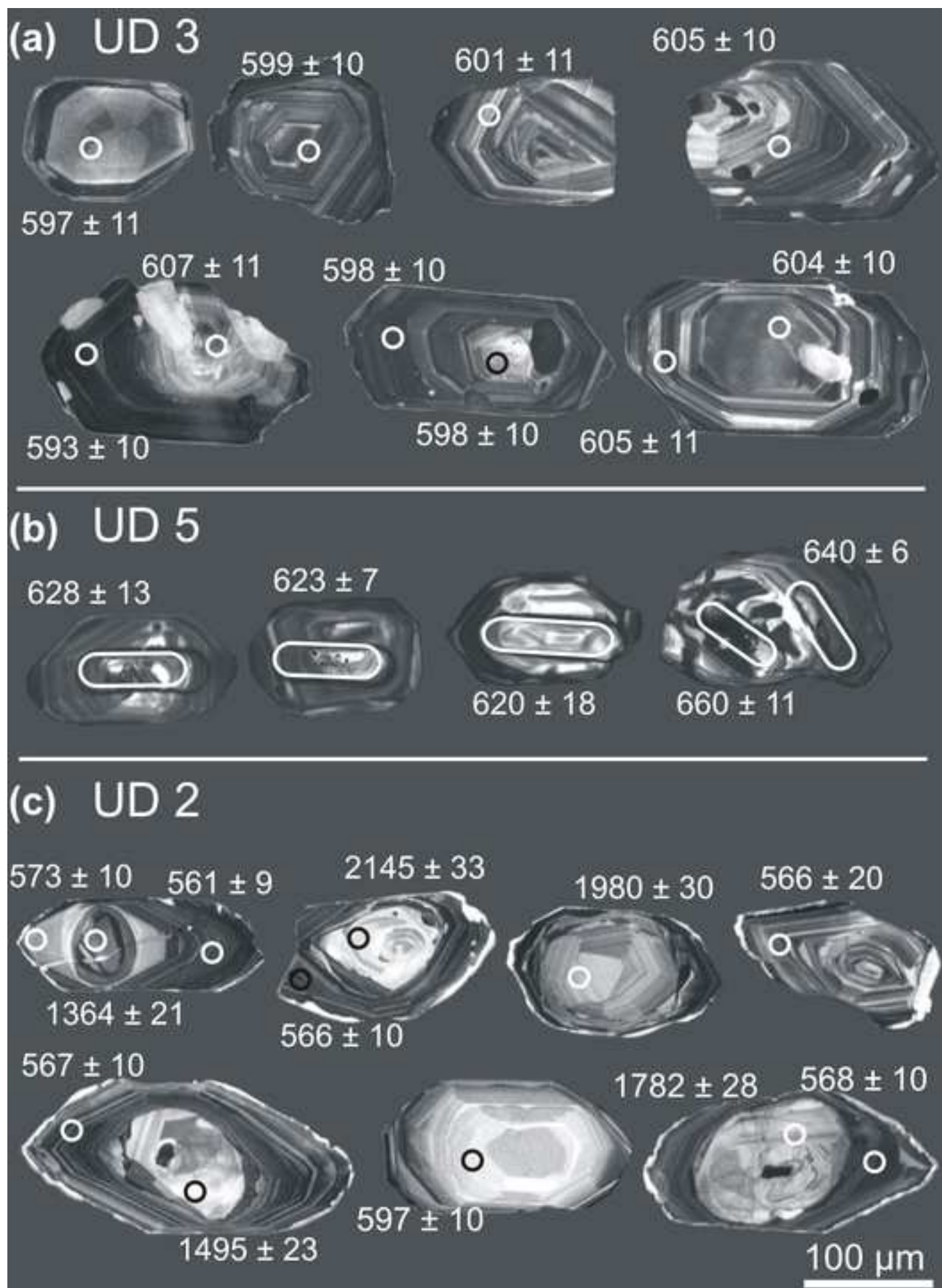


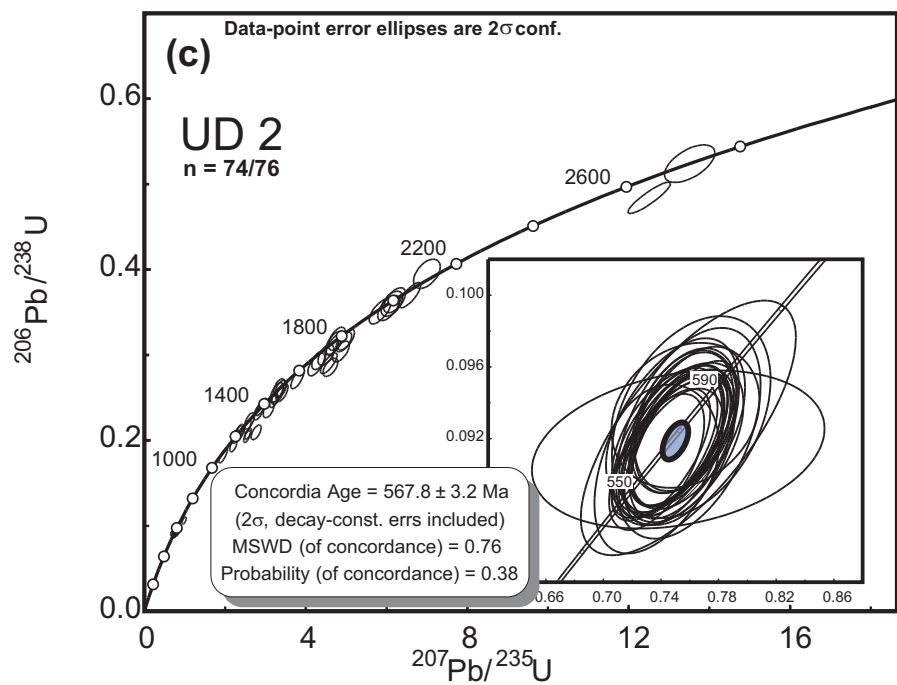
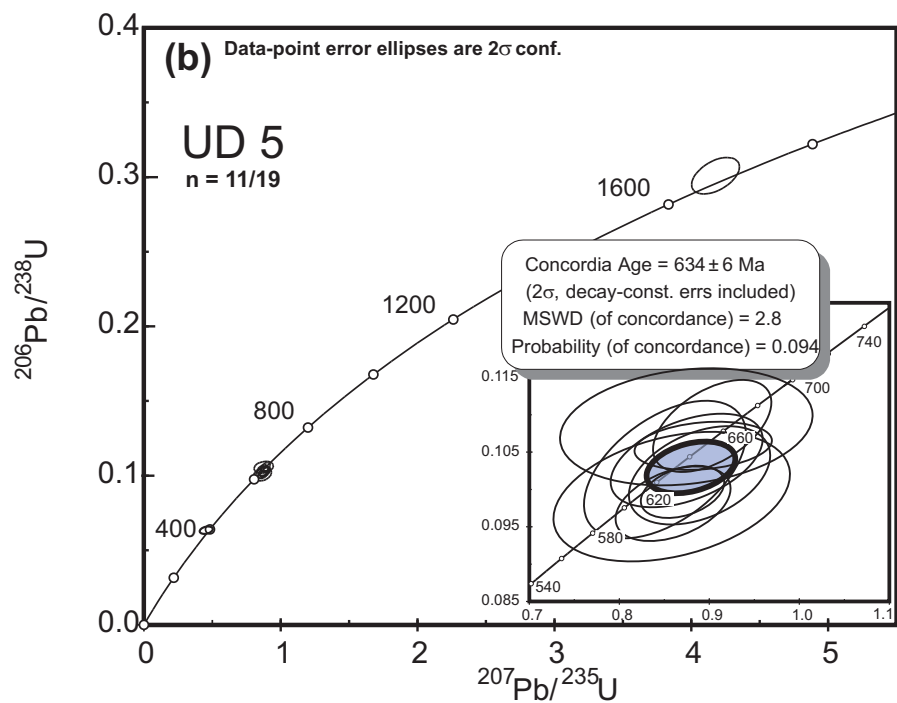
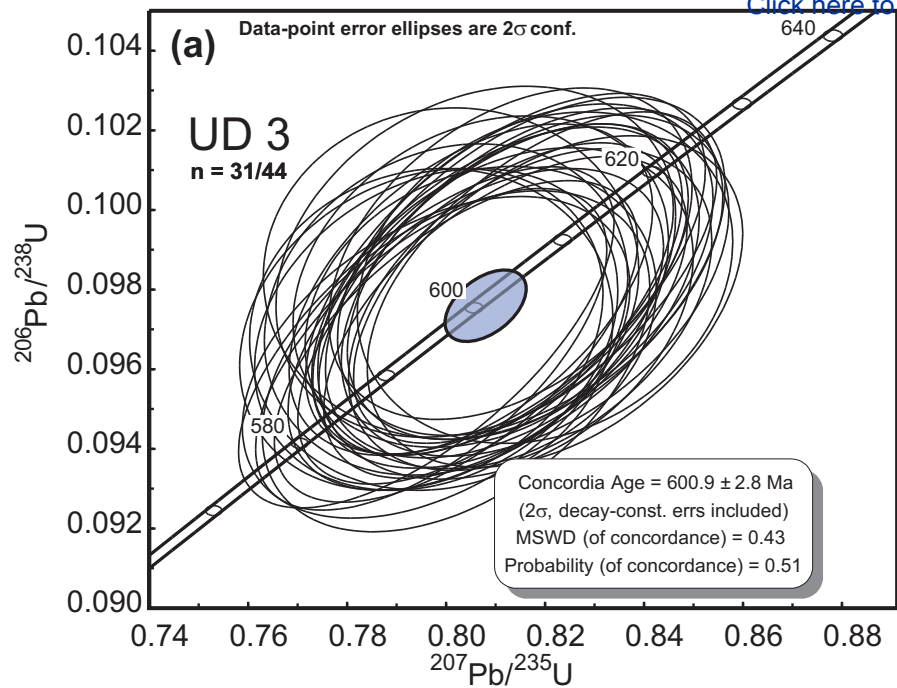


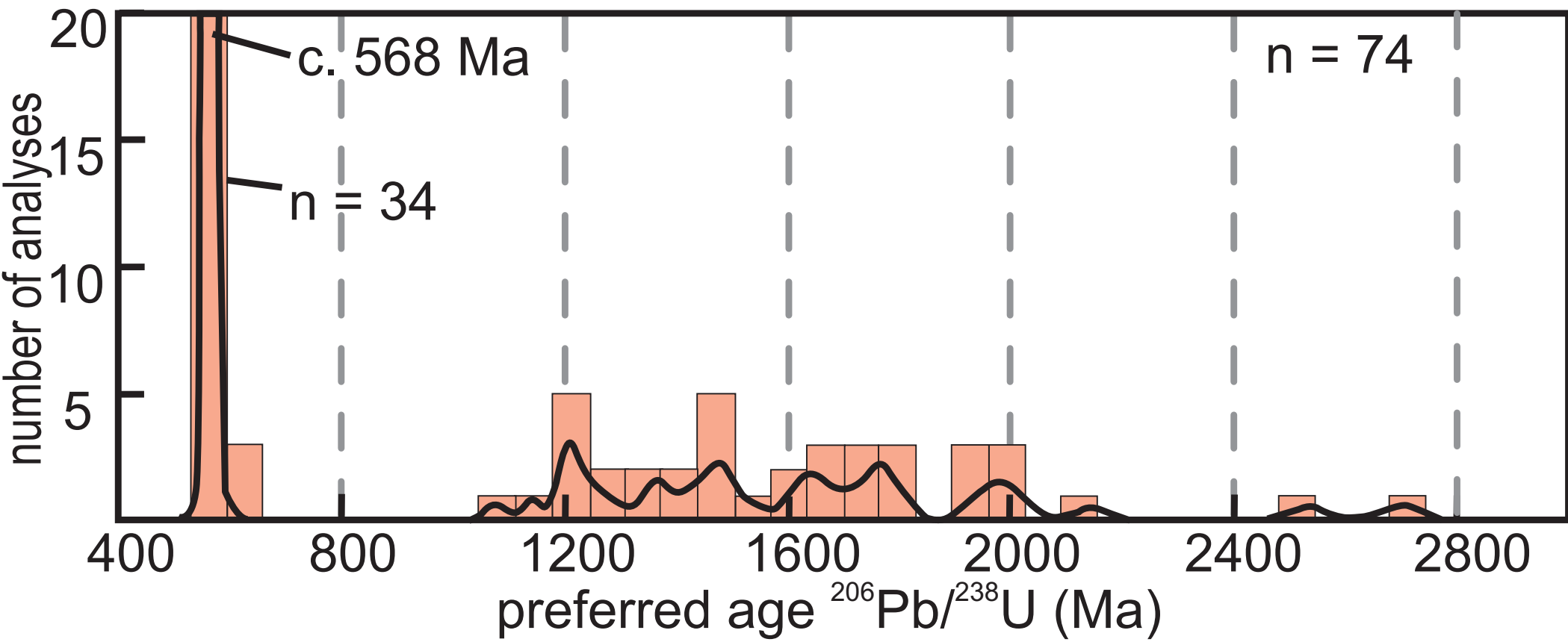


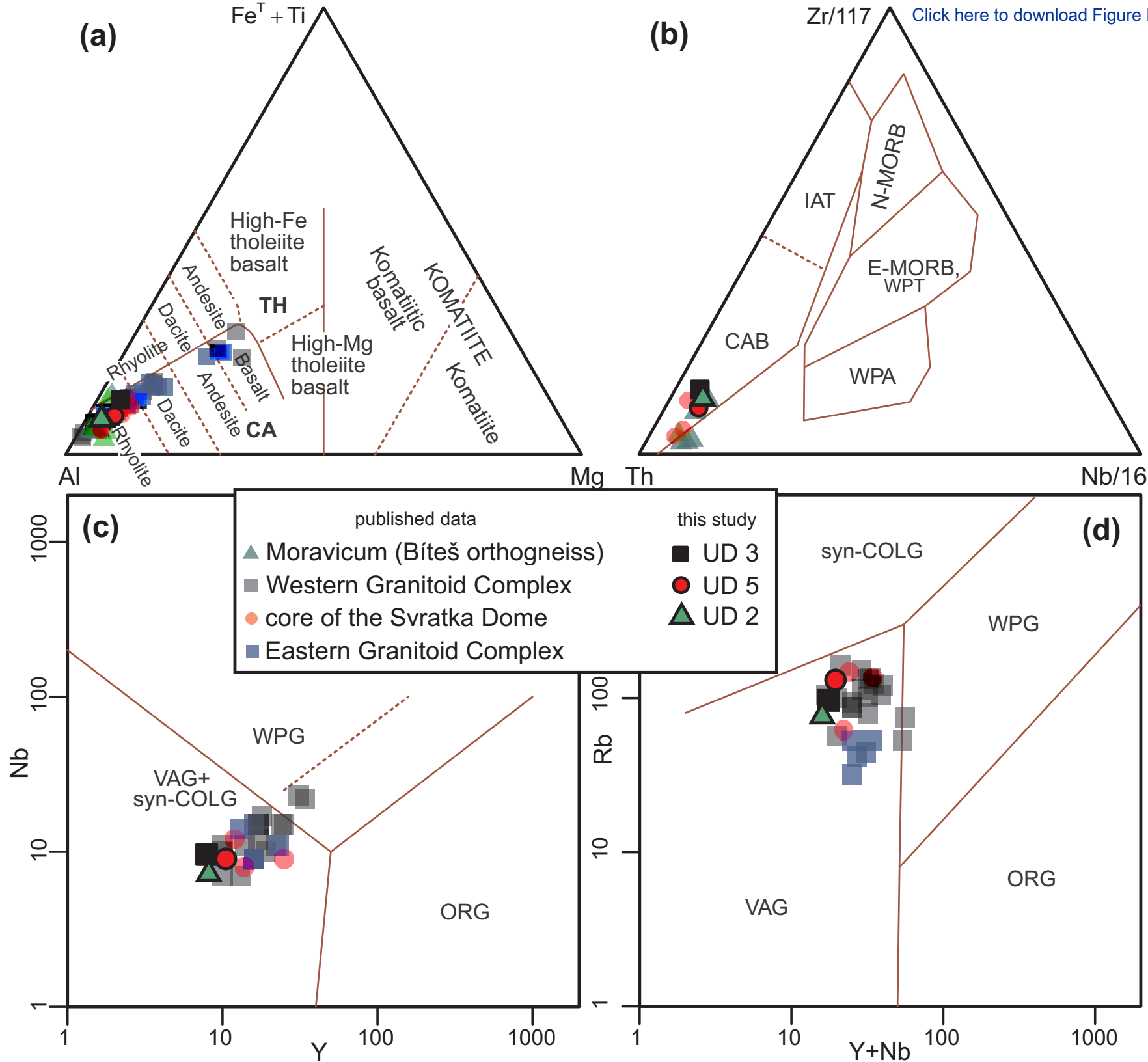


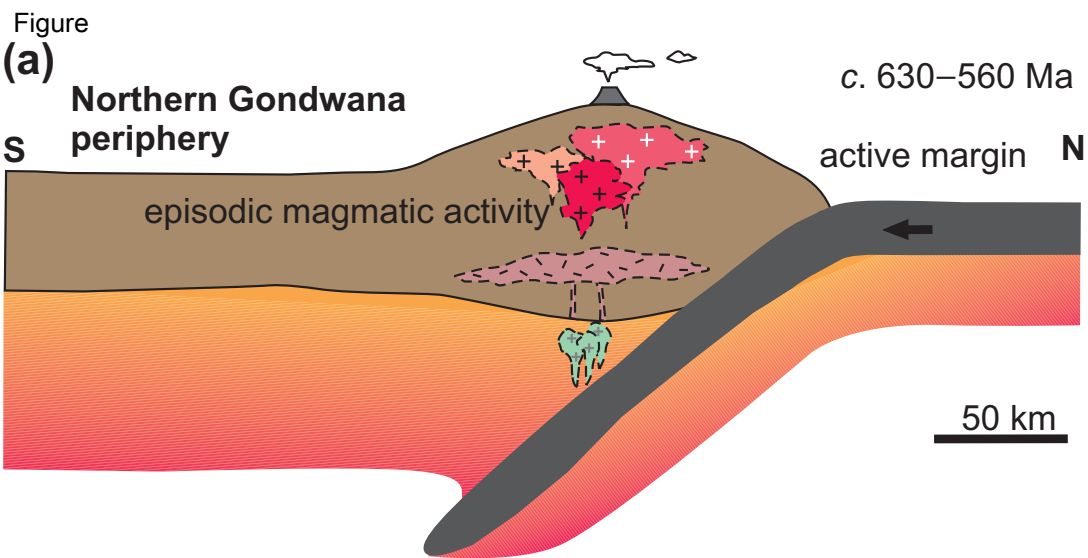






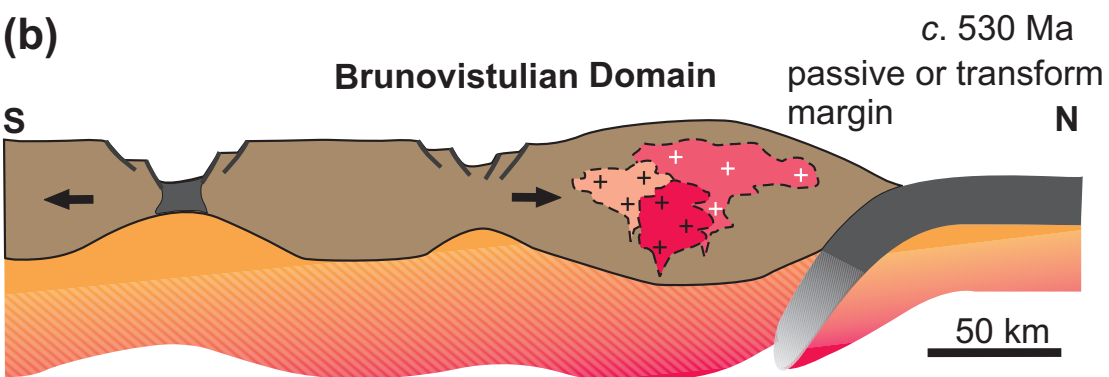






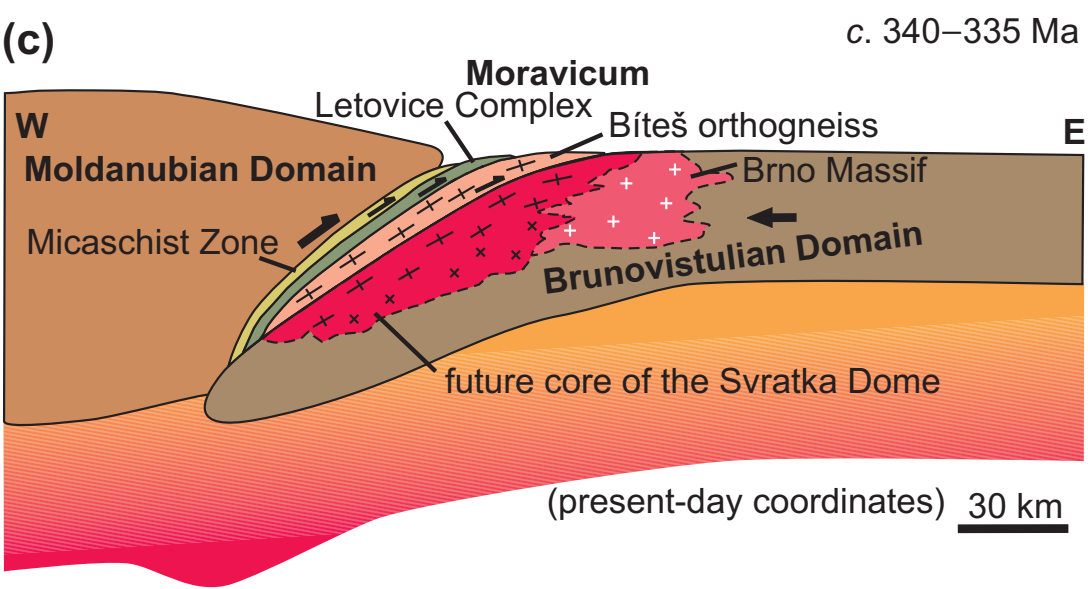
Long-lived subduction of oceanic lithosphere beneath the Gondwana continental margin during the Late Proterozoic and c. 65 Myr duration of multi-stage magmatic activity [1]

Episodic magmatic activity involving partial melting of various crustal sources within the magmatic-arc system [1]



Transition from active to passive/transform continental margin followed by initiation of separation from the Gondwana mainland during the Early Cambrian [2, 3, 4, 5]

Lithospheric thinning resulted in origin of the Rheic Ocean [6, 7] or more likely incipient oceanic basins [8]



Underthrusting of the Brunovistulian Domain beneath the thickened Moldanubian orogenic root, imbrication of the western margin of the Brunovistulian Domain (Moravicum) at c. 340 Ma [9, 10]

Exhumation of the Moravian nappe system [11, 12] and crustal-scale folding (development of the Svatka and Thaya domes) at c. 335 Ma [10]

Table 1 Major-element data (wt. %)

	UD 3	UD 5	UD 2
SiO₂	69.57	70.98	71.64
TiO₂	0.36	0.18	0.19
Al₂O₃	14.78	14.40	15.66
FeO_t	2.74	1.87	1.77
MnO	0.06	0.07	0.04
MgO	0.63	0.68	0.43
CaO	1.50	1.47	2.26
Na₂O	3.47	3.29	4.06
K₂O	5.38	4.38	2.90
P₂O₅	0.04	0.10	0.02
LOI	0.9	2.2	0.6
Σ	99.43	99.62	99.57
K₂O/Na₂O	1.55	1.33	0.71
A/CNK	1.04	1.12	1.12
mg	29.11	39.31	30.19

Table 2 Trace-element data (ppm)

	UD 3	UD 5	UD 2
Rb	96.5	131.0	78.1
Cs	0.9	2.9	0.9
Ba	1486.0	647.0	851.0
Sr	333.3	170.1	616.4
Th	10.3	7.1	5.9
U	1.4	2.5	0.9
Zr	222.8	107.0	108.8
Hf	5.9	3.3	3.2
Nb	9.6	9.0	7.4
Ta	0.5	0.7	0.4
Sc	3	3	2
Ni	5.2	2.3	1.8
Co	2.8	1.9	1.8
Pb	5.9	7.4	3.1
Zn	31	36	39
Cu	1.2	1.8	0.9
Y	7.9	10.5	8.2
La	60.4	22.9	22.5
Ce	103.2	41.9	40.1
Pr	10.47	5.07	4.65
Nd	31.3	19.3	15.7
Sm	3.67	3.18	2.60
Eu	1.19	0.57	0.62
Gd	2.28	2.70	1.97
Tb	0.30	0.37	0.25
Dy	1.58	1.94	1.46
Ho	0.27	0.38	0.23
Er	0.68	0.93	0.63
Tm	0.11	0.14	0.08
Yb	0.92	1.01	0.76
Lu	0.13	0.12	0.09
La_N/Yb_N	44.3	15.3	20.0
La_N/Sm_N	10.4	4.5	5.4
Eu/Eu*	1.26	0.59	0.84
Yb_N	4.4	4.8	3.6

Table 3 Sr–Nd isotopic data

Sample	Age	⁸⁷ Rb/ ⁸⁶ Sr	⁸⁷ Sr/ ⁸⁶ Sr	2s_Sr	⁸⁷ Sr/ ⁸⁶ Sr _i	¹⁴⁷ Sm/ ¹⁴⁴ Nd	¹⁴³ Nd/ ¹⁴⁴ Nd	2s_Nd	¹⁴³ Nd/ ¹⁴⁴ Nd _i	εNd _i	T _{NdDM} 2stg
UD 3	601	0.8383	0.711957	0.000006	0.704772	0.0709	0.512093	0.000009	0.511814	-1.0	1.33
UD 5	634	2.2329	0.725420	0.000014	0.705227	0.0996	0.512043	0.000011	0.511629	-3.7	1.57
UD 2	568	0.3669	0.713056	0.000011	0.710085	0.1001	0.511765	0.000012	0.511392	-10.0	2.01

¹ subscripts 1 indicate age-corrected isotopic ratios

² two-stage Nd model ages (Ga) (Liew and Hofmann 1988)