Origin and provenance of 2 Ma – 2 Ga zircons ejected by phreatomagmatic eruptions of Pliocene basalts in southern Slovakia

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Abstract: U-Pb ages of zircons recovered from Pliocene pyroclastic deposits in northern part of the Cenozoic intra-Carpathian back-arc basin (Pannonian Basin) span the interval from Pliocene (2.2 Ma) to Paleoproterozoic (Orosirian-Rhyacian, 1850-2115 Ma). The scattered U-Pb ages reflect eruption ages of the host basaltic volcanic centres, two episodes of post-Eocene magmatic crustal growth, and the possible tectonic affiliation, provenance and age of the subjacent basement or the sedimentary basin detritus sampled by the basaltic magma.

The youngest zircons define maximum ages of phreatomagmatic eruptions during the Late Miocene-Pliocene extension. These zircons are distinguished from older zircons by Zr/Hf (40-90) and Th/U ratios (0.5-4.5) as well as super-chondritic $\varepsilon\text{Hf}(t)$ values ranging from +7 to +14, indicating mantle-derived parental magmas. The locally increased Th/U ratios (up to 8) accompanied by Zr/Hf >60 are diagnostic of evolved phonolite parental melt. Hence, the youngest zircons can be interpreted as antecrysts, originating from evolved melts cogenetic

with the host alkali basalts. In contrast, older zircons represent xenocrysts scavenged by the uprising basalt from surrounding rocks.

Subordinate Eocene-Early Oligocene (29-38 Ma) sub-group of zircon xenocrysts is coincidental with the magmatism and volcanism along the Periadriatic lineament and the middle-Hungarian zone. The Early Miocene (18 Ma) cluster is coeval with the deposition of the Bükk Mountains felsic ignimbrite correlated with the onset of the back-arc extension that triggered Miocene sedimentation within the Pannonian Basin. The Eocene-Early Oligocene zircons have been likely scavenged from pyroclastic and ash-fall deposits of the Palaeogene retroarc basin subjacent to the Miocene basin infilling. Sub-chondritic $\epsilon\text{Hf}(t)$ values between -2.5 and -8 in the Eocene-Early Miocene zircons attest their crystallization from subduction-related felsic-to-intermediate melts containing large amounts of recycled crustal material. Palaeozoic-Proterozoic zircons create a heterogeneous population with variable trace element abundances and $\epsilon\text{Hf}(t)$ values. The determined age clusters are reminiscent of some basement units cropping out recently in Central Western Carpathians. Zircon Hf isotope data indicate recycling of up to 3.4 Ga old mafic crust and also the presence of 2 Ga old juvenile mafic crust. These units had either underlain the northern part of the Pannonian Basin during Pliocene, or had been exposed during the deposition of Miocene clastic sediments. The absence of Mesoproterozoic, Grenvillian zircons (0.9-1.8 Ga) in the pre-Cenozoic population of zircon xenocrysts is provisionally interpreted as indicating the evolution of the zircon source area within the west-African Craton.

Keywords: basalt, zircon, U-Pb-Hf isotopes, Pannonian Basin, Western Carpathians

1. Introduction

Zircon megacrysts, several mm to cm in diameter, often occur in Zr-undersaturated mafic magmas related to continental and oceanic intra-plate settings (Hoskin and Schaltegger 2003). The megacrystic zircons provide invaluable information addressing eruption ages of host basalts (Siebel et al. 2009, Hurai et al. 2013), and oxygen isotope composition, temperature and Nd-Hf systematics of subcontinental mantle (Valley et al. 1998, Paquette and Mergoïl-Daniel 2009, Siebel et al. 2009, Schaltegger et al. 2015). There is a consensus that the megacrystic zircons in mafic igneous rocks are in fact antecrysts crystallized from mantle-derived alkalic melts differentiated at various depths as felsic-to-intermediate intrusive bodies and transported to surface in younger basalt portions (e.g. Aspen et al. 1990; Guo et al. 1996; Upton et al. 2009; Paquette and Mergoïl-Daniel 2009; Yu et al. 2010; Sutherland et al. 2016; Schmitt et al. 2017). The term antecryst refers to zircon crystals crystallized from an earlier pulse of magma incorporated in a later pulse (Miller et al. 2007). In contrast, old zircons with different geochemical characteristics incorporated from the surrounding host rocks or those unrelated to the magmatic system are named xenocrysts (Johnson 1989).

Megacrystic zircons have also been detected in products of phreato-magmatic eruptions of the south-Slovakian Volcanic Field (SSVF) extending over northernmost promontories of the intra-Carpathian back-arc basin in the Lučenec Basin and the Cerova Highlands. Hurai et al. (2012, 2013) showed that the zircon megacrysts typologically and chemically resemble those present in syenite xenoliths from the same pyroclastic deposits (Huraiová et al. 1996; Hurai et al. 1998). Homogeneous U-Pb ages of zircons from maar sediments showed uninterrupted growth in sub-volcanic reservoirs within ± 100 ka. Only one of the studied zircon megacrysts showed heterogeneous zoning created by a 2.55 ± 0.14 Ma old core and 1.91 ± 0.07 rim, which reflected a $\sim 650 \pm 200$ ka long crystallization in a magmatic reservoir. Older inherited cores have not been observed (Hurai et al. 2010, 2012). Apatite and zircon crystals recovered from the same pyroclastic deposits returned ages almost equivalent (within uncertainty limits) closure ages of the U-Pb system at magmatic temperature (> 900 °C) and the low-temperature (~ 240 -

60 °C) post-eruptive cooling recorded by the (U-Th)/He system (Hurái et al. 2013), thus attesting the antecrystic character of the studied zircon megacrysts.

U-Pb isotopes have been hitherto determined in zircons from a total of five maars in the SSVF (Hurái et al. 2010, 2012, 2013). This study has been extended to additional maar-diatreme structures with the aim to better characterize not only megacrystic, but also sub-megacrystic zircons smaller than 200 µm. We provide new U-Pb and Lu-Hf isotope data from a total of 44 zircons recovered from maar-diatreme structures from SE part of the SSVF. Compared to previous results, the expanded dataset paints more complex picture of scattered ages, involving zircon antecrysts and older zircons xenocrysts containing also inherited cores. This paper is aimed to set the new U-Pb and Lu-Hf isotope data into geological context.

In other parts of the Carpathian back-arc basin, U-Pb, Lu-Hf, (U-Th)/He isotope or fission track zircon data are only limited to Middle Eocene-Early Oligocene tuffs in western Pannonian Basin (Danišík et al. 2015) and Early-to-Middle Miocene ignimbrites in Bükk Mts. in northern Pannonian Basin (Lukács et al. 2015, 2018). U-Pb ages of zircons in a metasedimentary granulite xenolith isolated from the basaltic tuff of Balaton Highlands in the central part of the Carpathian back-arc basin (Downes et al. 2015) recorded a complex thermal history correlated with late Mesozoic (90-100 Ma) and Palaeogene (40-65 Ma) metamorphism in the adjacent orogenic collisional belts, the increased heat flow at 15-16 Ma coincidental with the Miocene-Pliocene overheating and the final eruption of the host basalt at 3.7 Ma in Pliocene.

2. Geological setting

The south-Slovakian Volcanic Field (SSVF) is located within the Alpine-Carpathian-Pannonian (ALCAPA) mega-unit (Fig. 1a) composed of the Alpine-Carpathian orogenic belt and the superimposed intra-Carpathian back-arc basin (Pannonian Basin). The ALCAPA mega-unit is separated from the neighbouring Tisza-Dacia mega-unit to the south by the middle-

Hungarian zone, which represents the continuation of the Periadriatic lineament. Both mega-units were juxtaposed during Late Oligocene-Early Miocene lateral extrusion of the ALCAPA from Alpine collision zone (Kázmér and Kovács 1985, Csontos 1995). During Paleogene, the area experienced subduction accompanied by sedimentary basin formation in foreland, forearc (Kázmér et al. 2003) and retroarc (Tari et al. 1993) of the Alpine-Carpathian orogenic belt. Superimposed back-arc extension during Early Miocene controlled by the subduction of oceanic lithosphere underlying former flysch basins, and the ongoing subduction in Eastern Carpathians (Nemcok et al. 1998; Konečný et al. 2002) triggered the back-arc sedimentation of the Pannonian Basin starting at ~19 Ma (Royden et al. 1983), accompanied by intense volcanic activity.

Subduction with roll-back, slab-break-off, delamination, strike-slip tectonics and block rotation in the ALCAPA region has been accompanied by several episodes of syn- and post-collisional magmatic activity (Nemcok et al. 1998; Konečný et al. 2002; Seghedi et al. 2004; Harangi et al. 2006; Pécskay et al. 2006; Harangi and Lenkey 2007; Kovács et al. 2007; Lexa et al. 2010): 1) Eocene-Early Miocene volcanism activated by dextral strike-slip faulting along the Periadriatic lineament, subdivided into subordinate Eocene (60-37 Ma) and more extensive Late Eocene-Early Oligocene (37-25 Ma) and Late Oligocene-Early Miocene (25-17.5 Ma) pulses; 2) Early Miocene (21-11 Ma) calc-alkaline rhyolite-dacite volcanism coeval with early extensional phases; 3) areal- and linear-type middle-Miocene calc-alkaline andesite to bimodal andesite-rhyolite volcanism associated with the partial melting of mantle wedge above subducting slab (Szabó et al. 1992) and/or to late-stage back-arc extension (Lexa and Konečný 1974); and 4) disseminated Upper Miocene-Pleistocene Na-alkaline basaltic and basanitic volcanism (11.5-0.1 Ma) coincidental with the partial melting of mantle triggered by the asthenosphere bulging and crustal thinning (Embey-Isztin and Dobosi 1995). The thinned crust and lithospheric mantle in the study region is recently underlain in the depth of ~70 km by an

asthenospheric bulge (Tašárová et al. 2009; Bielik et al. 2010) uplifted to a depth of ~22 km in the central part of the Pannonian Basin (Horváth 1993).

The SSVF covers an area of about 500 km² along the Hungary-Slovakia state boundary (Fig. 1b). The SSVF is the classical region, where xenoliths of mantle peridotites (Hovorka and Fejdi 1980; Szabó and Taylor 1994; Huraiová and Konečný 1994; Konečný et al. 1995a; Downes and Vaselli 1995, Liptai et al. 2017) and granulites (Kovács and Szabó 2005) have been studied to decipher the composition and properties of lower crust and mantle, as well as processes associated with the mantle metasomatism and melting. Evidences for basaltic magma underplating, fractionation within mantle and crustal reservoirs, and generation of anorogenic magmas have been obtained in this area from the study of clinopyroxene and amphibole-bearing mafic cumulates (e.g. Huraiová et al. 1996; Zajacz and Szabó 2003; Kovács et al. 2004; Hurai et al. 2007; Zajacz et al. 2007) as well as syenites and A₁-type intra-plate granites (Huraiová et al. 1996, 2005, 2017a, b, 2019; Hurai et al. 1998, 2010).

The SSVF basalts span compositions from basanite through basaltic trachyandesite to phonotephrite. Trace element distribution patterns and radiogenic isotopes in the basalts (Embey-Isztin and Dobosi 1995) combined with those in mantle xenoliths (Downes and Vaselli 1995; Liptai et al. 2017) reflect a 7-25 % partial melting of spinel-bearing mantle source with basalt compositions locally affected by the enriched (metasomatized) mantle.

The SSVF basalts erupted within the interval from ~7 Ma to ~200 ka during several successive phases (Konečný and Lexa 2007). The initial Late Miocene phase was identified in NW part of the SSVF (Balogh et al. 1981). Superimposed Pliocene-Quaternary basalts erupted in SE part of the SSVF during 2nd to 6th consecutive phases (Konečný et al. 1995b). Phreatic and phreato-magmatic eruptions triggered and accelerated by the interaction of basalt with aquifers gave rise to maar-diatreme structures formed by Surtseyan, Strombolian and Hawaiian eruptions (Lexa et al. 2010). Subsequent vertical tectonic movements along NW–SE- and NE–

SW-trending faults exposed diatremes – feeder conduits of maars destructed by erosion (Konečný et al. 1999).

U-Pb zircon dating (Hurái et al. 2010, 2012) complemented with low-temperature (U-Th)/He geochronometry applied to zircon and apatite (Hurái et al. 2013) proved an Upper Miocene (~5.5 Ma) eruption age of maars near Pinciná and Fiľakovo–Castle Hill, and Pliocene eruption ages of maars near Hodejov (2.9 Ma), Hajnáčka (3.1 Ma) and Gemerské Dechtáre (3.2 Ma). Zircons extracted from a granite xenolith in the lava flow near Čamovce village (Huráiová et al. 2017b) returned an U-Pb age of 5.18 ± 0.02 Ma similar to the 5.2 ± 0.9 Ma whole-rock K/Ar age of the host basalt (Konečný 2007), thus attesting the capability of the U-Pb zircon geochronometer to constrain the basalt eruption ages.

3. Samples

Zircons have been recovered from two diatremes (Šurice, Tachty) and one maar (Fiľakovo–Red Hill). An isolated cliff in the southern margin of Šurice village named Owl’s Castle ($48^{\circ}13'34.04''\text{N}$, $19^{\circ}54'44.94''\text{E}$, Figs. 2a, b) is an exposed feeding channel of deeply eroded maar containing a block of subsided maar sediments in the upper part. The early infilling of the diatreme consists of palagonite tuffs and breccias with larger blocks and fragments of Early Miocene sandstones. The subsided block of stratified maar deposits is composed of tuffaceous siltstones and sandstones grading in the uppermost part into palagonite/tuff breccias with agglutinated scoria fragments and bombs. The sedimentary succession reflects the transition from early Surtseyan wet-surge to late Strombolian eruptions. The diatreme infilling as well as the subsided maar sediments are cut by basalt dykes, representing feeder channels of Hawaiian eruptions (lava fountains) and possibly of an eroded crater lava lake. The K/Ar whole rock dating of the basalt dyke returned an age of 4.0 ± 0.3 Ma (Konečný et al. 2003), although the diatreme itself has been affiliated with the 3rd volcanic phase lasting from 2.45 to 2.75 Ma

(Konečný 2007). The volcanoclastic material from the Šurice diatreme exhibits regular distribution of grain size fractions and significant volume of coarser grain-size fractions, indicative of sorting in a maar lake.

The diatreme in western part of Tachty village (48°9'21.07"N, 19°56'42.40"E, Figs. 2c, d) is filled with well-stratified, partly welded basaltic tuff containing sandstone fragments and blocks of the early Miocene Fil'akovo Formation (Burdigalian) subjacent to the tuff cone. The pyroclastic succession is dominated by lapilli tuffs with glassy lapilli embedded in siliciclastic matrix. Variable amounts of sideromelane and tachylite reflect a complex magma fragmentation history, recycling in the feeder conduit and interactions with water (Németh et al. 2011). The sharp sub-vertical contact of the diatreme wall with the caustically modified Burdigalian sandstone is preserved along the eastern side of the outcrop. The Tachty diatreme has also been affiliated with 3rd volcanic phase (Konečný 2007). Another small outcrop exposed in southern part of the Tachty village, 200 m SW from the previous sampling site, contains pyroclastic material mixed with larger sandstone fragments with green glauconite grains diagnostic of an Early Miocene sandstone/arcose. The sample studied is an unsorted mixture of volcanoclastic and sedimentary material ejected during a phreato-magmatic eruption from the proximal volcanic centre. Two zircon grains were recovered and studied from this site.

The Red Hill maar (48°16'43.27"N, 19°49'24.76"E) is situated in northern part of the Fil'akovo town. Similar to the proximal maar located in the centre of Fil'akovo (Castle Hill), it is built of a relict tuff ring of coarse-grained, unsorted wet-surge strata deposited onto Early Miocene sandstones of the Fil'akovo Formation. Superimposed is the succession with well-stratified fine- to coarse-grained palagonitized lapilli tuffs with variable proportions of vesicular basalt and occasional fragments of Miocene sediments. The alternation of finely laminated, massive, unsorted and sorted coarse strata indicates dry-surge, wet-surge and pyroclastic fall depositions. Both maars of Fil'akovo have been thought to be coeval with the Pleistocene Günz/Mindel boundary (Konečný et al. 1999). However, U-Pb zircon combined with (U-

Th)/He zircon and apatite data ([Hurai et al. 2013](#)) returned an Upper Miocene-Lower Pliocene age of the Castle Hill maar (5.62 ± 0.04 Ma, 5.5 ± 0.6 Ma and 5.4 ± 0.7 Ma, respectively). A sample from the Red Hill maar investigated in this study has been excavated from the eastern outer slope of the tuff ring of incoherent volcanoclastic material intercalated within siltstone (Figs. 2e, f).

Heavy mineral assemblage of the volcanoclastic sediments under study also involves pyroxene, amphibole, garnet, titanite, olivine, apatite, rutile, epidote, sillimanite and andalusite. Garnet, amphibole and pyroxene dominate in the Fil'akovo–Red Hill maar, whereas sediments from Tachty and Šurice diatremes are enriched in pyroxene and olivine ([Nemec and Huraiová 2018](#)).

4. Methods

Heavy mineral fractions hydro-dynamically separated from 10-20 kg of maar sediments were further processed by sieving, gravitational separation in heavy liquids and final electromagnetic purification. Zircons were hand-picked from diamagnetic fraction under binocular microscope, mounted in epoxy resin, ground and polished to expose crystal interiors.

Zircon chemical compositions supplemented by cathodoluminescence and back-scattered electron imaging were measured using a Cameca SX-100 electron microprobe at the State Geological Institute in Bratislava (Slovakia), using 15 keV accelerating voltage, 20 nA beam current, counting times between 5 and 25 s and 2-5 μm beam diameters. The instrument was calibrated with natural and pure synthetic compounds as described in [Hurai et al. \(2012\)](#). Matrix effects were resolved using the X-PHI correction method ([Merlet 1992](#)). Interferences were eliminated using empirical correction factors ([Amlı and Griffin 1975](#); [Konečný et al. 2004](#)). Detection limits were between 0.01 and 0.05 wt. %. Uncertainties varied from 1 % (major elements) to 10 % (trace elements).

U-Th-Pb isotope data on zircons were obtained *in situ* by laser ablation inductively coupled plasma spectrometry (LA-ICP-MS) at the Laboratoire Magmas & Volcans (LMV, Clermont-Ferrand, France). The analyses involved the ablation of minerals with a Resonetics Resolution M-50 Excimer laser system operating at a wavelength of 193 nm with a repetition rate of 3 Hz and a laser fluency of 2.5 J/cm². The laser spot diameter was set at 27-40 µm (Appendix S1.1). The ablated material was carried by helium and mixed with nitrogen (Paquette et al. 2014) and argon before injection into the plasma source of a Thermo Element XR sector-field HR-ICP-MS. The analytical protocol was similar to that reported by Mullen et al. (2018).

Data were corrected for U-Pb fractionation during laser sampling, and for instrumental mass discrimination (mass bias) by standard bracketing with repeated measurements of GJ-1 zircon (Jackson et al. 2004). The 91500 zircon reference material (Wiedenbeck et al. 1995) was also repeatedly analyzed as an unknown to independently monitor the reproducibility and accuracy of the corrections. Data reduction was carried out with the GLITTER® software package from Macquarie Research Ltd (van Achterbergh et al. 2001). For zircons younger than 10 Ma, ²⁰⁶Pb/²³⁸U ratios were corrected for the ²³⁰Th disequilibrium (Schärer 1984), assuming a Th/U ratio for the whole rock corresponding to 3±1 (Paquette and Mergoill-Daniel 2009). Data were not corrected for common Pb. Calculated ratios were exported, and concordia ages and diagrams were generated using the Isoplot/Ex v. 2.49 software package (Ludwig 2001).

Hafnium isotopes were measured with a Thermo Scientific Neptune Plus multi-collector ICP-MS coupled to a Resonetics M50E 193 nm excimer laser system at the LMV, Clermont-Ferrand, France. Data were collected in static mode (¹⁷¹Yb, ¹⁷³Yb, ¹⁷⁴Hf, ¹⁷⁵Lu, ¹⁷⁶Hf-Yb-Lu, ¹⁷⁷Hf, ¹⁷⁸Hf, ¹⁷⁹Hf). The operating conditions and instrument settings are described in Moyen et al. (2017), Paquette et al. (2017a) and reported in the Appendix S1.2. Mass bias effects on Hf were corrected using an exponential law and true value for ¹⁷⁹Hf/¹⁷⁷Hf = 0.7325 (Patchett and Tatsumoto 1980; Patchett et al. 1981). Data reduction was based upon user-selected background and sample integration intervals. Outlier Hf isotopic ratios for each analysis were

rejected using a two-standard deviation criterion. The $\varepsilon\text{Hf}(t)$ value is defined as the deviation of the sample $^{176}\text{Hf}/^{177}\text{Hf}$ from the CHondritic Uniform Reservoir (CHUR) in parts per 10^4 calculated according to the $\lambda^{176}\text{Lu}$ value = $1.865 \times 10^{-11} \text{ yr}^{-1}$ (Scherer et al. 2001) and the CHUR parameters ($^{176}\text{Hf}/^{177}\text{Hf} = 0.282793$; $^{176}\text{Lu}/^{177}\text{Hf} = 0.0338$) from Iizuka et al. (2015).

5. Results

5.1. Zircon ages

Several zircon groups have been distinguished according to their U-Pb ages (Appendix S2). Late Pliocene zircons represent the most abundant population (almost 30 % of all zircons) in Tachty and Šurice diatremes, but missing in the Red Hill maar. The Pliocene zircons form rounded and isometric grains, up to $1400 \times 1700 \text{ }\mu\text{m}$ in size, which often represent fragments of even larger crystals (Fig. 3). Corrosion was observed on the surface of smallest crystals. Dense oscillatory or domain zoning of Pliocene zircons are discerned in cathodoluminescence (CL), while inherited cores have not been observed. Crystal faces preserved in one grain (TA-2-1-2) were consistent with a simple combination of (101) pyramid and (100) prism. The zircon growth graded from the homogeneous central part to the rim crowded with innumerable thorite inclusions. The texture, reminiscent of that observed in zircon cores from Nb, REE-rich calc-alkalic granite xenoliths (Huraiová et al. 2017a), indicates the breakdown of $\text{ZrSiO}_4\text{-ThSiO}_4$ solid solution. The sharp boundary between the two zones rules out a post-growth, fluid-induced dissolution-precipitation. Similar syn-growth textures have been observed in zircon megacrysts from the Fil'akovo–Castle Hill maar (Hurai et al. 2013). The Late Pliocene zircons yielded ^{230}Th -corrected U-Pb ages ranging from 3.0 to 2.2 Ma (Fig. 4). Zircons from Tachty can be subdivided into three sub-groups clustering at $2.21 \pm 0.03 \text{ Ma}$ (24 analyses, 2 grains), $2.37 \pm 0.02 \text{ Ma}$ (32 analyses, 3 grains), and $2.50 \pm 0.04 \text{ Ma}$ (12 analyses, 2 grains). Zircons

from Šurice cluster at 3.00 ± 0.07 Ma (5 analyses, 1 grain), 2.81 ± 0.06 Ma (6 analyses, 1 grain) and 2.44 ± 0.02 Ma (37 analyses, 4 grains).

14 % of all investigated zircons returned Eocene-Miocene ages. One to three zircons of this group are present in each locality. Compared to the previous group, Eocene-Miocene zircons are smaller (typically 200-300 μm , occasionally up to 500 μm), almost euhedral elongated crystals with fully or partially preserved, locally corroded faces. Typical dense oscillatory and sector zonings are visible in CL. Some zircons contain rounded resorbed core (CV-N2-1-5). The group of Eocene-Miocene zircons yielded ages clustering at 35.1 ± 0.6 Ma, with additional individual data at about 38 and 29 Ma. One 35.1 Ma old zircon from Tachty contains an inherited 306 ± 9 Ma core. Three zircon grains from Šurice and Fil'akovo–Red Hill are concordant at 18.0 ± 0.3 Ma (Fig. 5a).

Palaeozoic zircons represent 25 % of the analyzed crystals. The Palaeozoic zircons are more frequent in the Fil'akovo-Red Hill maar than in the Šurice diatreme. The group of Palaeozoic zircons is texturally and morphologically rather heterogeneous, with smaller crystal sizes, typically 200-250 μm . They form elongated crystals and fragments showing local magmatic resorption and epiclastic corrosion. Dense oscillatory as well as patchy zoning of cores grading into marginal oscillatory growth zones are discernible in CL images. All zircons show discordant to concordant ages scattered between 300 and 510 Ma (Fig. 5b). A small group of two discordant 300 Ma zircons from Fil'akovo-Red Hill is coeval with the inherited core of the Eocene zircon from Tachty. Concordant ages span intervals from 504 ± 11 Ma (Early Cambrian) to 373 ± 10 Ma (Late Devonian). In spite of the distinct zoning and clear core-rim textural relationships, age contrasts have not been recorded within individual crystals.

Ages of Neoproterozoic zircons (23 % of investigated grains, Fig. 6a) span the interval from 570 Ma (Ediacaran) to 850 Ma (Cryogenian). CL textures range from weak domain (700-790 Ma) to dense oscillatory zonings (590-630 Ma). External morphology is isometric to elongated, with signs of magmatic and chemical corrosions. An 854 Ma old zircon shows

preserved crystal faces. Most investigated grains are doubly terminated crystals. Some zircons with sub-parallel parting and cataclasis have been excluded from isotopic study.

Rare Paleoproterozoic zircons (4 grains, 9 % of the investigated zircons) have been found only in the Fil'akovo–Red Hill maar (Fig. 6b). Internal textures range from dense oscillatory (e.g. FCV-Z1-3) to irregular domain zonings (CV-N2-3-1, CV-N2-1-6). Age of this zircon group covers the interval between 1850 and 2115 Ma (Orosirian-Rhyacian). The oldest zircon (FCV-Z1-5) exhibits a bright core-dark rim relationship expressed by two different ages, 2115 ± 18 Ma and 2065 ± 8 Ma, respectively. In spite of the complex internal zoning, ages of other zircons were homogeneous and concordant. Multiple analyses of the FCV-Z1-3 zircon plot along a discordia defining an upper intercept age of 1996 ± 14 Ma.

5.2. Zircon compositions and Lu-Hf isotopes

Pliocene zircons can be distinguished from older zircons in most discrimination diagrams based on relative and absolute U, Th, Y, Yb, Zr, Hf abundances (Figs. 7, 8 and Appendix S4). Zr/Hf and Th/U ratios are scattered within typical intervals of 40-90 and 0.5-3.5, respectively (Appendices S2 and S4). High Th/U ratios, between 4 and 8, in zircons from Tachty (TA-8-1, TA-1-1, TA-2-1-2, TA-Z1-2) are exceptional among all hitherto investigated SSVF zircons. Zr/Hf and Th/U ratios in Eocene-Oligocene zircons are considerably lower, ranging at 32-33 and 0.1-0.6, respectively. Pre-Cenozoic zircons exhibit relatively low Zr/Hf ratios, ranging from 30 to 50; whereas Th/U ratios are mostly significantly lower (0.1-0.8).

According to U-Pb and Lu-Hf isotope data summarized in Appendices S2 and S3, respectively, zircons from basalt maars of SSVF create clusters corresponding to several episodes of crustal growth (Fig. 9). Pliocene zircons differ from older zircons by unique super-chondritic $\varepsilon\text{Hf}(t)$ values, from 7 to 14, contrasting with sub-chondritic Eocene-Oligocene zircons, with $\varepsilon\text{Hf}(t)$ values ranging between -2.5 and -8. The pre-Tertiary zircon group forms

two discrete Palaeozoic-Neoproterozoic and Paleoproterozoic arrays, covering the interval of $\epsilon\text{Hf}(t)$ values between -11.5 and +13.

6. Discussion

6.1. Possible zircon sources

Pyroclastic deposits of maar structures are composed of a mixture of volcanoclastic material and disrupted rock fragments scavenged from feeder conduits. As such, they may theoretically contain zircons from several sources: 1) oceanic crust and associated marine sediments accreted to lithosphere during Miocene subduction (e.g. [Dobosi et al. 2003](#)), 2) pre-Cenozoic basement rocks within Early Oligocene transgressive conglomerates, gravels and sandstones ([Vass and Elečko 1992](#)), 3) Cenozoic sandstones of the Pannonian Basin ([Vass et al. 2007](#)), 4) magmatic reservoirs occurring at various crustal levels (e.g. [Huraiová et al. 1996, 2017a](#)).

Detrital zircons involved in the Cenozoic sedimentary infilling of the Pannonian Basin may also be provided by several distant sources: 1) uplifted Central Carpathian basement drained by rivers, 2) ash-fall deposits from Eocene-Early Miocene volcanic centres. Areal Middle Miocene andesite volcanism must have ejected a large amount of material distributed throughout the whole Pannonian basin. In contrast, pyroclastic ejecta from small Late Miocene to Pleistocene monogenetic volcanoes are probably negligible zircon suppliers owing to a smaller volume of the volcanic material distributed over substantially shorter distances.

The proximal and distal zircon sources can be discriminated according to external morphology of zircon crystals. Sharp crystal edges rule out longer sub-aquatic transport and mechanical abrasion (epiclastic corrosion), thus excluding the Cenozoic sedimentary detritus as the possible zircon source. Chemical corrosion of zircon crystals indicates the interaction with hot volatile magma and the absence of sub-aquatic abrasion. Decompression in rapidly

ascending basalt may result in a specific parallel parting of zircon crystals (e.g. [Rudnick and Williams 1987](#)). Rounded crystals indicate either magmatic resorption or the sub-aquatic abrasion. However, zircons armoured in xenoliths may retain their external crystal morphology until the ultimate phreato-magmatic explosion despite the longer transport in basaltic magma.

6.2. Origin of Cenozoic zircons

Cenozoic zircons from SSVF maars cover the interval from Late Eocene (35-38 Ma) to Pliocene (2.2 Ma). Their chemistry, internal zoning and morphology are unambiguously magmatic in origin. Pliocene zircons from Tachty and Šurice diatremes create coherent age clusters, which rule out their source in distal volcanic centres. The youngest zircon ages, i.e., 2.2 Ma in Tachty and 2.4 Ma in Šurice, define lower limits for eruption ages. The Šurice diatreme can thus be affiliated with the 3rd (2.45-2.75 Ma) volcanic phase, whereas the Tachty diatreme belongs to the 4th (2.25-1.6 Ma) volcanic phase according to the scheme proposed by [Konečný et al. \(1999\)](#) and [Konečný \(2007\)](#). The published 4.0 ± 0.3 Ma K/Ar whole-rock age of basalt dyke intersecting the Šurice diatreme ([Konečný et al. 2003](#)) should be thus regarded as erratic. The interval of 0.3-0.6 Ma between the oldest and youngest Pliocene zircons from individual diatremes is consistent with previous studies ([Hurai et al. 2010, 2013](#)) and may reflect the residence time in sub-volcanic reservoirs. The absence of post-Early Miocene zircons younger than the surrounding basin deposits precluded the estimate of the eruption age of the Fil'akovo–Red Hill maar.

Locally >1 mm megacrystic size, sector or oscillatory zonings and rounded shapes attributed to magmatic resorption are typical diagnostic of the Pliocene zircons. Frequent chemical corrosion reflects the interaction with volatile magma. Abundant crystal fragmentation can be ascribed to the mechanical instability and destruction during phreato-magmatic eruptions.

Zr/Hf ratios in the Pliocene zircons are scattered within the 40-90 interval, which is consistent with megacrystic zircons from other maar structures (Hurái et al. 2013). The locally increased Zr/Hf ratios (>60), significantly higher than those in zircons crystallizing from calc-alkaline granites and pegmatites (Wang et al. 2010), may reflect a late Zr-saturation in more alkaline parental melts (Boehnke et al. 2013). Th/U ratios in the Pliocene zircons range between 0.5 and 8.1, and overlap those of megacrystic zircons (0.2-4.6) from other maar structures (Hurái et al. 2013). The high Th/U ratios coupled with increased Zr/Hf >60 are consistent with an evolved alkaline syenite or phonolite parental melt (Belousova et al. 2002; Schmitt et al. 2010, 2017). The increased Zr/Hf and Th/U ratios in zircons from Tachty may reflect an alkaline sub-volcanic reservoir (probably phonotephrite), which could fractionate to phonolite in certain crustal levels at decreasing temperature.

Uranium abundances in Pliocene zircons mostly plot between the fields of zircons from highly evolved, Nb-REE-rich A₁-type calc-alkaline granite xenoliths (Zr/Hf = 11-21, >1200 ppm U) and relatively less-evolved alkali syenite xenoliths (Zr/Hf = 38-74, 2-85 ppm U). The decreased and strictly limited Hf contents (5,000-10,000 ppm), together with widely fluctuating Y contents (380-11,680 ppm) are typical of felsic differentiation products from mantle-derived mafic magmas (Pupin 2000). This zircon origin is also confirmed by the super-chondritic $\varepsilon\text{Hf}(t)$ values (+7 to +14), diagnostic of the magmas derived by a low-degree partial melting of lithospheric mantle and only weakly contaminated by an older crustal material. The $\varepsilon\text{Hf}(t)$ values in the Pliocene zircons are in the lower range of ~5.5 Ma zircons (Fig. 9, inset) recovered from A₁-type granites (Huraiová et al. 2017a). The close genetic relationship of megacrystic zircons and alkaline syenites is also supported by the occurrence of yttrialite inclusions in zircons from maar sediments and syenite xenoliths (Hurái et al. 2010, 2013, and unpublished data).

The Late Eocene-Early Miocene zircons with U-Pb ages scattered between 38 and 18 Ma form only a subordinate, but geologically interesting population of 1-3 grains in each locality

studied. Doubly terminated zircons show central domain encompassed by the zone with dense oscillatory zoning. Dominant are (110) prisms combined with variable proportions of (211) and (101) pyramids. The growth of (211) pyramid at the expense of the (101) pyramid was observed in some crystals (e.g., TA-Z1-1, 38 Ma). The opposite trend was observed in 18 Ma old zircons (e.g., CV-N2-1-5). The inherited 306 ± 9 Ma core in the Late Eocene, 35 Ma old zircon from Tachty (TA-Z1-1) indicates recycled Carboniferous crust.

Dominantly euhedral shapes without signs of magmatic resorption and only moderate chemical corrosion rule out the interaction of zircons with hot basalt as well as their mechanical abrasive treatment in fluvial environment. Hence, the zircons must have been either armoured in xenoliths during transport in magma, or scavenged from the feeder conduit of the maar structure in uppermost crustal levels. Considering their scarcity, inconsistent ages, lacking epiclastic as well as crystal-magma interaction features, the incoherent pyroclastic fall deposits subjacent to the Miocene basin infilling are the most probable sources for these zircons.

Zr/Hf values in Eocene-Miocene zircons between 31 and 33 are consistent with a late-magmatic origin (Wang et al. 2010). The dominantly anatectic parental magmas are also corroborated by sub-chondritic $\varepsilon\text{Hf}(t)$ values between -2.5 and -8. The chemical and radiogenic isotope fingerprints thus correspond to an orogenic, acidic-to-intermediate, subduction-related magmatism. The 35-38 Ma old zircons are coeval with the volcanic centers along the Periadriatic fault system and the middle-Hungarian zone interpreted as a subduction-related arc volcanism (Kovács et al. 2007; Danišik et al. 2015). The obtained ages overlap the oldest andesites from Velence and Récsk Mts. dated at 37 Ma (Benedek 2002).

The Early Miocene cluster at 18 Ma corresponds to the earliest syn-extensional eruptions within the Pannonian Basin marked by silicic ignimbrite deposits of the Bükkalja volcanic field south of Bükk Mountains. The ignimbrites are thought to reflect initial stages of the extension accompanied by an increased heat flow from upwelled asthenosphere. Sub-chondritic $\varepsilon\text{Hf}(t)$ values in Bükkalja tuff zircons overlap those from SSVF maars and indicate the prevalence of

the crustal component in the parental granitic magmas, thus indicating initially thick continental crust progressively thinned during the subsequent stages of silicic magma generation between 18 and 14 Ma (Lukács et al. 2015, 2018).

6.3. Origin and provenance of Palaeozoic-Proterozoic zircons

Dense oscillatory and sector zoning similar to that in the magmatic Eocene-Oligocene zircons is also typical for the Cambrian-Silurian zircons. The dominant (110) prisms associated with (110) and (211) pyramids (e.g., SE-SP-1-1, SE-VR-2-4, CV-N2-2-1), together with Th/U ratios between 0.2 and 0.6, and $\varepsilon_{\text{Hf}}(t)$ values between -2.2 and +2.2 favour the magmatic origin and hybrid felsic parental melts.

Nearly all Ediacaran zircons exhibit the dense oscillatory zoning associated with high Th/U ratios (up to 1.8). However, combinations of strongly negative and positive $\varepsilon_{\text{Hf}}(t)$ values (-8.4 in CV-N2-2-2 *versus* +12.8 in CV-N2-3-4) indicate synchronous occurrence of recycled crustal protolith and juvenile parental melt.

Cryogenian zircons exhibit diverse zonings. Nevertheless, their high Th/U ratios (0.2-1.1) combined with the super-chondritic $\varepsilon_{\text{Hf}}(t)$ values between +6.3 and +9.7 attest their close genetic relationship with juvenile magmas.

Th/U ratios of four Early Proterozoic zircons range between 0.4 and 0.6, thus attesting their magmatic origin. A typical “metamorphic” Th/U ratio of 0.01 (Rubatto 2002; Paquette et al. 2017b) was observed only in the 2115 ± 18 Ma core of the FCV-Z1-5 zircon, associated with a super-chondritic $\varepsilon_{\text{Hf}}(t)$ value projecting close to the depleted mantle array. This probably indicates the juvenile magmatic origin of the zircon grain overprinted by a solid-state metamorphic recrystallization. Other Paleoproterozoic zircons project below the CHUR, thus indicating the recycling of up to 3.4 Ga old mafic crust.

Assuming an evolution line of $^{176}\text{Lu}/^{177}\text{Hf} = 0.020$ characteristic of mafic crust (O'Neil et al. 2013), most $\varepsilon\text{Hf}(t)$ values obtained from Oligocene to Neoproterozoic zircons can be inferred from the continual recycling of ~1.0-2.0 Ga old crust mixed with a juvenile mantle-derived component, probably during the Gondwana assembly (Fig. 9). The Paleoproterozoic zircons indicate the Paleoproterozoic-to-Late Archean crust recycling during the assembly of Columbia and its later incorporation in the peri-Gondwanan Cadomian belt.

Owing to large variations in trace element composition, Lu/Hf isotopes, and internal zoning, the most probable sources of Palaeozoic to Proterozoic zircons must be sought in the subjacent middle to upper crustal basement units (proximal source) and/or those cropping out in Central Western Carpathians during the deposition of clastic fluvial sediments of the Cenozoic basins (distal source). This assumption is also supported by the study of mineral assemblages from the Red Hill maar, Tachty and Šurice diatremes, containing metamorphic almandine garnets and rutiles derived from metasedimentary rather than metavolcanic amphibolite-facies rocks (Nemec and Huraiová 2018).

U-Pb ages of pre-Cenozoic SSVF zircons overlap the Variscan, Caledonian, Cadomian and Eburnean orogenic growth cycles recorded also in neighbouring tectonic units of Central Western Carpathians. Rather surprising is, however, the very weak geochronological record of large Upper Carboniferous granitoid plutons exposed in Central Western Carpathians, as well as the lacking zircons coincidental with the acidic Permian-Triassic volcanism and the late-Mesozoic and Early Paleogene metamorphic overprint of crystalline basements of the Alps and Carpathians as well as the mafic granulite crust of the ALCAPA block. The gap between 850 Ma and 1.8 Ga indicates an interrupted crustal growth in the zircon source region, but this behaviour may also be random, caused by the limited number of investigated zircons.

Palaeozoic zircons from SSVF with mixed super- and sub-chondritic $\varepsilon\text{Hf}(t)$ values indicate a magmatic activity evolving from the Ordovician rifting to the Variscan collision. Relatively abundant Neoproterozoic zircons indicate the evolution within the Cadomian orogenic arc

(Neubauer 2002; Linnemann et al. 2008), also documented by an inherited zircon core (562 ± 43 Ma) in the Miocene zircon (13.2 ± 0.2 Ma) in the andesite laccolith near Fil'akovo (Bouloton and Paquette 2014). The population of pre-Cenozoic SSVF zircons can be thus correlated with an Armorican/Cadomian fragment of European Variscides (Putiš et al. 2008). Absence of Mesoproterozoic ages might indicate the evolution of the source area of SSVF zircons within the west-African Craton (e.g. Fernández-Suárez et al. 2004; Linneman et al. 2014; Henderson et al. 2016).

Proximal and distal zircon sources cannot be unambiguously discriminated within pyroclastic deposits. The absence of lithic fragments does not favour a proximal zircon provenance in the subjacent crust sampled by the ascending magma along feeding channels. However, alkalic syenite xenoliths representing the ultimate source of Pliocene zircons are missing as well, in spite of the maximum abundance of these zircons in all investigated maar-diatreme structures. Late Eocene-Early Miocene zircons derived from ash-fall deposits within the basin infilling also favour the source of the pre-Cenozoic zircons in clastic sediments, because blocks of the Miocene basin sediments are relatively abundant in the maar deposits. On the other hand, sharp crystal faces of most pre-Cenozoic zircons rule out longer sub-aquatic transport and mechanical abrasion, thus favouring a proximal (subjacent crust) rather than a distal (uplifted Central Carpathian basement) provenance. The proximal source in the subjacent crust should also be admitted for some Precambrian zircons showing the parallel parting deformation textures reminiscent of zircons from rapidly decompressed crustal xenoliths transported by ascending basalts (Rudnick and Williams 1987). However, the lower crustal granulite protolith of the ALCAPA mega-unit can definitely be ruled out as the source of pre-Cenozoic zircons, owing to the lacking high-temperature overprint and disturbance of the U-Pb system. At any rate, the question of proximal or distal provenance of pre-Cenozoic zircon xenocrysts in Pliocene maar structures remains open to further research.

7. Conclusions

Zircons recovered from pyroclastic deposits ejected by phreato-magmatic eruptions in northern part of the Pannonian Basin are shown to provide unconventional information pertaining to eruption ages of volcanoes and the nature of rocks scavenged from walls of feeding channels of maar-diatreme structures. Most zircons have been derived from intermediate to felsic differentiation products of intra-plate Pliocene alkali basalts. Owing to minimum residence times (less than 300 ka) in sub-volcanic reservoirs, the Pliocene zircons also define minimum eruption ages of the host maar-diatreme structures. Positive $\varepsilon_{\text{Hf}}(t)$ values indicate a mantle-derived magmatic source uncontaminated by the crustal material. The Upper Miocene-Pliocene magmatic zircons can safely be discriminated from older zircons by their increased U/Th and Zr/Hf ratios.

Cenozoic zircons come from Late Eocene-Early Oligocene volcanic centres along the middle-Hungarian zone, and the Miocene areal- and linear-type felsic-to-intermediate volcanism. These zircons have been scavenged by the ascending basaltic magma from ash-fall deposits from the base of the sedimentary basin infilling. Contrary to mantle-derived Pliocene zircon antecrysts, their negative $\varepsilon_{\text{Hf}}(t)$ values indicate extensive crustal contamination or recycling in the parental acidic-to-intermediate magma.

Pre-Cenozoic zircons returned ages scattered within the interval from Late Carboniferous (~300 Ma) to Paleoproterozoic (Orosirian-Rhyacian, 1.8-2.1 Ga). Their variable internal textures, heterogeneous sub- and super-chondritic $\varepsilon_{\text{Hf}}(t)$ values and trace element signatures are diagnostic of the multiple recycling of continental crustal source. It remains unclear, if the pre-Cenozoic zircons have been scavenged by the uprising basalt from the basinal detritus or from the subjacent basement. Their source region could be correlated with an Armorican/Cadomian fragment of European Variscides derived from the peri-Gondwanan Cadomian arc and the west-African Craton.

541

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546 **Supporting Information**

547

548 Additional Supporting Information may be found in the online version of this article:

549 **Appendix S1.1** Details of analytical conditions for U-Pb isotope analysis

550 **Appendix S1.2** Details of analytical conditions for Lu-Hf isotope analysis

551 **Appendix S2** Zircon U-Pb isotope data

552 **Appendix S3** Zircon Lu-Hf isotope data

553

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841 **Figure captions**

842

843 **Fig. 1. a)** Schematic sketch of the Carpathian arc and the Carpathian back-arc basin (modified
 844 after Pécskay et al. 2006; Kováč 2000): 1 – European platform, 2 – foredeep, 3 – Tertiary
 845 accretionary wedge, 4 – Klippen belt, 5 – back-arc basin, 6 – Variscan crystalline basement

with pre-Tertiary sedimentary cover and Mesozoic nappe units (Cretaceous accretionary wedge), 7 – Neogene andesite and rhyolite, 8 – Upper Miocene-Quaternary alkali basalt volcanism, 9 – major normal, strike-slip and thrust faults. Rectangle marks the south-Slovakian Volcanic Field (SSVF). **b)** Enlarged view of the SSVF with volcanic phases derived from whole-rock K/Ar data on basalts (Konečný et al. 2003; Konečný 2007), U/Pb zircon, and (U-Th)/He apatite dating of maars and diatremes (Hurai et al. 2013).

Fig. 2. Photo-documentation and geological cross-sections (Konečný and Lexa 2007) of sampling sites. **a-b)** The Šurice diatreme with subsided block of the maar infilling in the upper part. Explanations: 1 – palagonitized tuff breccia with fragments of Early Miocene sedimentary rocks, 2 – subsided block of a maar complex: a – ash tuff, b – fine tuff, c – coarse tuff, d – palagonitized tuff breccia, e – lapilli tuff with scoria fragments, f – agglutinated bombs in scoria matrix, 3 – dykes and short sills, 4 – sandstone of the Fil’akovo Formation, 5 – scree. **c-d)** Subsided maar deposits within the Tachty diatreme. Explanations: 6 – deformed bedded fine palagonite tuff, 7 – palagonite tuff: a – fine, b – coarse, 8 – palagonite tuff breccia, 9 – slumped palagonite tuff blocks in sandy sedimentary matrix, 10 – breccia composed of blocks of sedimentary rocks, 11 – Early Miocene sediments. **e-f)** Areal view (GoogleEarth, Google Inc.) and photo-documentation of the sampling site on the eastern flank of the Fil’akovo – Red Hill maar.

Fig. 3. Morphological and textural types of Pliocene zircons from Tachty (TA) and Šurice (SE). Scale bars (1000 μm) pertain to all images. BSE and CL refer to back-scattered electron and cathodoluminescence images, respectively.

Fig. 4. Tera-Wasserburg plot of Pliocene zircons from **a)** Tachty diatreme (TA), Tachty - playground (TA-Z) and **b)** Šurice diatreme (SE). Error ellipses display 2σ uncertainties.

872

873 **Fig. 5. a)** Tera-Wasserburg plot of Eocene-Miocene zircons from Tachty (TA), Šurice (SE) and
874 Red Hill (CV) with CL images of analysed zircon grains. **b)** Concordia plot of Palaeozoic
875 zircons from Fil'akovo–Red Hill maar (CV, FCV) and Šurice diatreme (SE) with CL images of
876 selected grains. Scale bars correspond to 100 μm . Error ellipses display 2σ uncertainties.

877

878 **Fig. 6. a)** Concordia plot of Neoproterozoic zircons from Šurice (SE) and Fil'akovo – Red Hill
879 (CV, FCV) with CL images of selected grains. **b)** Concordia plot of Paleoproterozoic zircons
880 from Fil'akovo–Red Hill maar with CL images of analysed grains. Scale bars correspond to 100
881 μm . Error ellipses display 2σ uncertainties.

882

883 **Fig. 7.** U (ppm) versus Th/U plot from LA-ICP-MS analyses ($n = 116$) of **(a)** Tertiary and **(b)**
884 pre-Tertiary zircons complemented with data ($n = 553$) from Hodejov, Fil'akovo–Castle Hill,
885 Gemerské Dechtáre ([Hurái et al. 2013](#)) and Hajnáčka–Bone Gorge maars ([Hurái et al. 2012](#)).
886 Zircon analyses from syenite ($n = 85$) and granite ($n = 29$) xenoliths ([Hurái et al. 2010](#); [Huraiová](#)
887 [et al. 2017a](#)) are also shown for comparison.

888

889 **Fig. 8.** Discrimination plots constructed from EPMA analyses of Tertiary **(a, c)** and pre-
890 Tertiary **(b, d)** zircons complemented with the analyses of Late Miocene-Pleistocene zircons
891 from Hodejov, Fil'akovo–Castle Hill, and Gemerské Dechtáre maars ([Hurái et al. 2013](#)), and
892 zircon data from syenite and granite xenoliths from Pinciná ([Hurái et al. 2010](#)) and Čamovce
893 ([Huraiová et al. 2017a](#)).

894

895 **Fig. 9.** Hf-isotope evolution diagrams for SSVF zircons. Dotted lines in **a)** show the Hf-isotope
896 growths of 1, 2.2 and 3.4 Ga old mafic crusts with the $^{176}\text{Lu}/^{177}\text{Hf} = 0.02$. Depleted mantle curve
897 was constructed using the present-day depleted mantle ϵ_{Hf} value = 16.9 ([Griffin et al. 2002](#)).

898 The detail in **b)** shows data on Tertiary detrital zircons from maars and diatremes, and those
899 from the Late Miocene (5.5 Ma) A₁-type granite from Čamovce ([Huraiová et al. 2017a](#)).