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U–Pb and $^{40}Ar/^{39}Ar$ dating of the Miocene fossil track site at Ipolytarnóc (Hungary) and its implications

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Abstract

Abundant Early Miocene mammal and bird tracks and a rich plant assemblage is preserved by the emplacement of an ignimbrite sheet of the Gyulakeszi Rhyolite Tuff Formation (GRTF) near Ipolytarnóc in northern Hungary. The tuff that overlies the track-bearing sandstone yielded a single-crystal zircon U–Pb total isochron age of 17.42 ± 0.04 Ma and a single-crystal laser-fusion plagioclase $^{40}Ar/^{39}Ar$ age of 17.02 ± 0.14 Ma (uncertainties are quoted at the 2σ level). An additional $^{40}Ar/^{39}Ar$ age of 16.99 ± 0.16 Ma was obtained from the equivalent rhyolite tuff near Nemti, where the underlying terrestrial clay yielded early proboscidean remains assigned to the MN4 mammal zone. The new, high-precision dates allow revision of the numeric age and correlation of the Ipolytarnóc fossil site and the GRTF, previously based on an average K–Ar age of 19.6 ± 1.4 Ma. The difference of 0.40 ± 0.15 Ma between the U–Pb and $^{40}Ar/^{39}Ar$ ages support the growing evidence for a systematic bias between the two isotopic systems due to the inaccurately known ^{40}K decay constant but likely also includes an undetermined pre-eruptive residence time of zircon. Published biostratigraphic data from under- and overlying marine strata establish correlation with the NN3 nannoplankton zone and, together with the new radioisotopic ages, suggest assignment of the fossils and the tuff to the Ottnangian regional stage of the Central Paratethys. A global correlation lends support to the recently suggested astronomical calibration of the Early Miocene time scale that revised the previous scales towards younger ages. The $^{40}Ar/^{39}Ar$ age from Nemti provides a reliable correlation of the MN4 mammal zone in Central Europe with the numeric time scale and places a minimum constraint on the age of the regional Proboscidean Datum, the migration event of proboscideans from Africa to Europe through the emerging "*Gomphotherium* landbridge". Contrary to suggestions for a significantly earlier European Proboscidean Datum, it appears that the originally suggested age of c. 17.5 Ma is realistic but it is significantly younger than the South Asian Proboscidean Datum.

Keywords: Miocene, U–Pb dating; $^{40}Ar/^{39}Ar$ dating; Proboscidean Datum; Central Paratethys regional stages

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1. Introduction

Abundant and well-preserved fossil vertebrate tracks are exposed on the topmost bedding planes of a Miocene sandstone near Ipolytarnóc, NE Hungary. The site is considered the prime fossil locality of Hungary and it has been recently nominated to the UNESCO list of World Heritage Sites. Since 1944, it has been protected as a Nature Conservation Area in and around Borókás-árok (Borókás Gully), ~2 km from the village of Ipolytarnóc near the Hungarian–Slovak border (Fig. 1). The mammal and bird tracks were discovered in 1900, and the most recent monographic study [1] documented 1644 footprints that belong to 11 species. After the latest excavations in 1993, nearly 3000 footprints are exposed on several hundreds of m^2 of the sandstone bedding plane.

Preservation of the tracks on an ancient riverbank has been attributed to a volcanic eruption that instantly covered the sand by a several meters thick sheet of rhyolitic ignimbrite [2]. The rhyolite tuff itself is also fossiliferous and contains abundant plant remains in its basal 20–40 cm. A recent paleobotanical study identified 64 taxa among the large collection of macrofloral remains (nearly 10 000 leaves) [3]. The assemblage is dominated by laurophyllous plants, indicative of a vegetation in warm and humid, subtropical climate [3]. The first fossil discovery at Ipolytarnóc, made in 1837, was that of a giant silicified tree trunk embedded at the sandstone—tuff transition. The conifer was described as *Pinuxylon tarnocziense* (Tuzson) Greguss, its original length exceeded 46 m [4], and parts of it are still preserved on site.

Despite the interest in this remarkable fossil site, the age of the fossiliferous sandstone and the rhyolitic tuff has not been satisfactorily determined. By convention rather than on the basis of tight biostratigraphic constraints, the tuff has been regarded to mark the base of the Ottnangian [5], a regional stage in the Lower Miocene of the Central Paratethys. Published K–Ar ages are scattered and carry large errors. Commonly, an average age

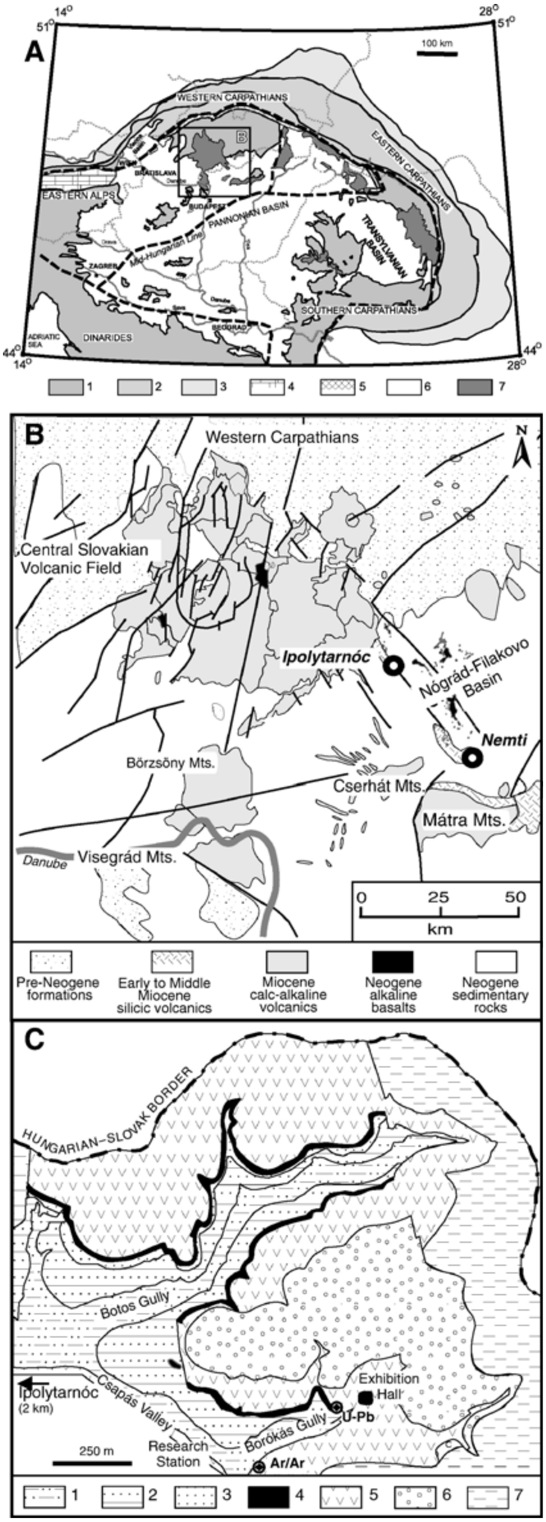

Fig. 1. A—Geotectonic setting of the study area within the Alpine–Carpathian–Pannonian region. Legend: 1—Inner Alpine and Carpathian belt, Dinarides, 2—Alpine–Carpathian flysch belt, 3—Carpathian molasse belt, 4—Northern Calcareous Alps, 5—Pienyny Klippen Belt, 6—Neogene sedimentary cover, 7—Neogene calc-alkaline volcanics. B—Generalized geology of the Nógrád–Filakovo basin and surrounding area, showing the location of Ipolytarnóc and Nemti. C—Geological map of the area around the Ipolytarnóc fossil site (after [2]), showing the location of dated samples. Legend: 1—Szécsény Schlier Formation, 2—Pétervására Sandstone Formation, 3—pebble conglomerate (Zagyvapálfalva Formation), 4—Track-bearing sandstone (Zagyvapálfalva Formation), 5—Gyulakeszi Rhyolite Tuff Formation, 6—Redesposited tuff and sandstone (Salgótarján Formation), 7—Variegated clay (Salgótarján Formation).

of the regionally extensive ignimbrite, 19.6±1.4 Ma, is also regarded as the age of Ipolytarnóc tuff [6]. Herein we report a new, high-precision single-crystal zircon U–Pb age and two plagioclase single-crystal laser fusion $^{40}Ar/^{39}Ar$ ages (one from Ipolytarnóc and another one from Nemti, located 50 km SE from Ipolytarnóc in the Nógrád Basin). Together, these dates provide a definitive age for the Ipolytarnóc fossils and a secure basis for their temporal correlation with other events. We discuss the implications of these ages (1) for comparison of the two radio-isotopic dating methods; (2) for timing of Miocene volcanism of the inner Carpathian arc; (3) for correlation of regional Paratethys stages and the Miocene time scale; and (4) for better resolving mammalian evolutionary and migration history (notably the age of regional Proboscidean Datum) and terrestrial-marine correlation.

2. Geological setting

Ipolytarnóc and the Nógrád Basin lie on the inner side of the Western Carpathian arc, near the northern margin of the Pannonian Basin (Fig. 1). The Alpine–Carpathian–Pannonian region is characterized by a complex Tertiary tectonic evolution [7]. Orogenic uplift of the Alpine range led to the formation of Paratethys, semi-isolated marine epicontinental basins with variable degree of connection to the Mediterranean, Atlantic, and Indopacific during the Paleogene and early Neogene [8]. The sedimentary basin evolution of the north Pannonian Basin is linked with the history of Central Paratethys and controlled by the interplay of eustasy and regional tectonics [9]. As part of the Central Paratethys, the Hungarian Paleogene Basin is interpreted as a retroarc flexural foredeep where molasse-type sedimentation prevailed until the Early Miocene [10]. Miocene successor basins, including the Nógrád–Filakovo Basin, record a change from compressional to extensional tectonic setting [10].

The Lower Miocene stratigraphy is well established locally for the vicinity of Ipolytarnóc (Fig. 2) [2], more broadly for the Nógrád Basin [5], and regionally for Hungary [11]. The chronostratigraphic framework uses the Central Paratethys regional stages of Eggenburgian, Ottnangian and Karpatian, correlated with the standard Burdigalian Stage [12]. Much of the upper Eggenburgian is comprised of a transgressive–regressive sedimentary cycle. At Ipolytarnóc, a borehole penetrated nearly 200 m of basinal, deep-water siltstone (Szécsény Schlier Formation). Its uppermost part is exposed on the surface and overlain by up to 50 m of locally glauconitic sandstone of nearshore facies (Pétervására Sandstone Formation). This unit is transitional to the Budafok Sand

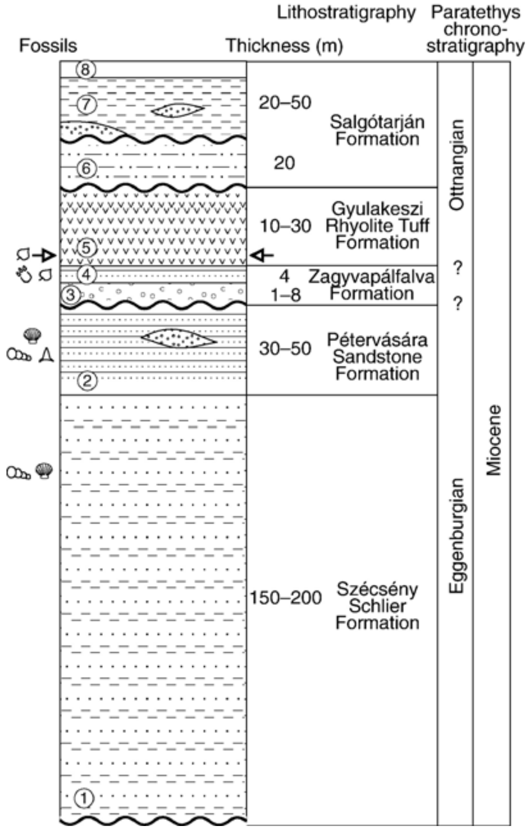

Fig. 2. Generalized Miocene stratigraphy of the Ipolytarnóc area (after [2]). The position of samples for radio-isotopic dating is marked by arrowheads, the track-bearing 'Ipolytarnóc beds' are marked by an asterisk. Lithology: 1—silty and clayey sandstone; 2—glauconitic sandstone with conglomerate lenses; 3—pebble conglomerate; 4—track-bearing sandstone; 5—rhyolitic ignimbrite; 6—redesposited tuff and sandstone; 7—variegated clay; 8—sand.

Formation, exposed farther to the west. At certain levels the sandstone contains an abundant marine Eggenburgian mollusk fauna [13] and shark teeth [14], and is in turn overlain by terrestrial strata of the Zagyvapálfalva Formation. An unconformity between the two formations is indicated by an irregular erosion surface [2]. Whereas elsewhere the Zagyvapálfalva Formation is dominated by variegated clays of continental to lagoonal facies, at Ipolytarnóc it contains 1–6 m of fluvial conglomerate overlain by 2 m of the track-bearing sandstone (informally called the Ipolytarnóc beds). The sequence is capped by a 10–30 m thick ignimbrite sheet, the subject of radio-isotopic dating reported here. The dated tuff is traditionally assigned to the Gyulakeszi Rhyolite Tuff Formation (GRTF), previously called the 'Lower Rhyolite Tuff'. This unit is the oldest product of the explosive silicic volcanism that is widespread in the

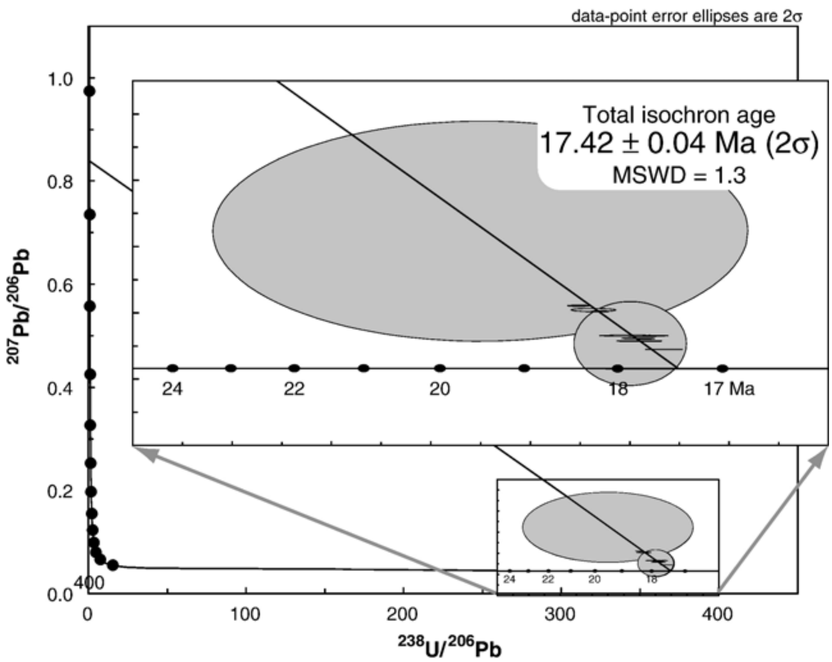

Fig. 3. Total U–Pb isochron diagram [20] of the Ipolytarnóc tuff sample. $^{204}Pb/^{206}Pb$ is perpendicular to the $^{207}Pb/^{206}Pb$–$^{238}U/^{206}Pb$ plane. The isochron is constrained by common Pb with $^{207}Pb/^{206}Pb = 0.84$ and $^{204}Pb/^{206} = 0.05464$.

Miocene of the Pannonian Basin. It is widely held that the silicic pyroclastics occur in three distinct horizons, although recently reported geochemical and petrologic evidence suggests the possibility of a more continuous temporal distribution of multiple eruptions [15]. The age of GRTF is known from numerous K–Ar dates and stratigraphically it is conventionally regarded to mark the base of the Ottnangian [6], even though the biostratigraphic constraints are loose. The ignimbrite grades upwards into reworked pyroclastics, overlain by terrestrial

Table 1
Zircon analytical data and calculated U–Pb ages

Sample	μg [a] zirc	ppm U	cm. Pb [b] (pg)	Th/U [c]	Total U/Pb isochron [d, e] $^{238}U/^{206}Pb$	%er	$^{207}Pb/^{206}Pb$	%er	$^{204}Pb/^{206}Pb$	%er	U/Pb concordia [f] $^{207}Pb/^{235}U$	%er	$^{206}Pb/^{238}U$	%er	ρ [g]	Age (Ma) $^{206}Pb/^{238}U$
IT.Z02	4.6	765	0.9	.55	360.9	1.6	.0667	0.4	.00139	2.1	.0172	1.7	.00270	1.2	.74	17.4±0.2
IT.Z03	7.4	797	1.1	.50	360.1	2.6	.0619	33.4	.00107	11.7	.0173	32.6	.00272	2.0	.07	17.5±0.3
IT.Z04	4.5	1686	1.1	.55	366.9	0.8	.0584	0.2	.00084	4.0	.0170	1.2	.00268	0.6	.60	17.3±0.1
IT.Z05	6.0	928	1.1	.54	362.0	1.0	.0634	0.2	.00116	1.1	.0173	1.2	.00270	0.8	.73	17.4±0.1
IT.Z06	3.2	2995	2.1	.64	361.7	0.7	.0650	0.2	.00124	1.0	.0174	1.1	.00270	0.6	.63	17.4±0.1
IT.Z07	3.9	629	1.2	.66	349.7	0.5	.0847	0.2	.00260	1.0	.0175	1.9	.00272	0.4	.54	17.5±0.1
IT.Z08	4.1	734	1.3	.56	352.7	1.0	.0821	1.1	.00243	3.9	.0173	3.4	.00271	0.8	.44	17.4±0.1
IT.Z09	7.9	619	5.0	.51	329.7	13.3	.1297	41.8	.00532	7.8	.0196	77.8	.00273	10.3	.14	17.6±1.8
IT.Z12	4.3	909	1.6	.59	353.9	0.4	.0813	0.4	.00233	4.1	.0175	3.0	.00270	0.4	.60	17.4±0.4

Ratios involving ^{206}Pb are corrected for initial disequilibrium in $^{230}Th/^{238}U$ adopting Th/U = 4 for the crystallization environment resulting in +90 ka for $^{206}Pb/^{238}U$ ages.
Uncertainties of individual ratios and ages are given at the 2Σ level and do not include decay constant errors.
[a] Sample weight is calculated from crystal dimensions and is associated with as much as 50% uncertainty (estimated).
[b] Total common Pb including analytical blank (analytical Pb blank is 0.8 ± 0.3 pg per analysis).
[c] Present day Th/U ratio calculated from radiogenic $^{208}Pb/^{206}Pb$ and age.
[d] Corrected for tracer contribution and mass fractionation (0.15 ± 0.09%/amu).
[e] 3D-isochron is constrained by an estimated $^{238}U/^{206}Pb = 0$, $^{207}Pb/^{206}Pb = 0.84$ (±1.7%) and $^{204}Pb/^{206}Pb = 0.05464$ (±5.5%).
[f] Ratios of radiogenic Pb versus U; data corrected for mass fractionation, tracer contibution and common Pb contribution.
[g] Correlation coefficient of radiogenic $^{207}Pb/^{235}U$ versus $^{206}Pb/^{238}U$.

variegated clay, siltstone, and sand of the Salgótarján Formation. Elsewhere in the Nógrád Basin, this formation contains economically important coal measures of paralic facies.

The other sampled site between Nemti and Bátonyterenye is located in the southern part of the abandoned Salgótarján mining district. The site is within the type area of the GRTF where the ignimbrite is underlain by variegated clay of the Zagyvapálfalva Formation. The sample site is only c. 1.5 km NNE of the Nemti clay pit, where the GRTF caps the terrigenous clay deposits that yielded proboscidean remains. In this area the GRTF is directly overlain by the coal-bearing Salgótarján Formation.

A new transgressive sedimentary cycle starts in the latest Ottnangian–Karpatian. The marine sandstone of the Kazár Member (Egyházasgerge Formation) contains a mollusk fauna dominated by the endemic Paratethys genus *Rzehakia,* and nannoplankton of the NN4 zone, thereby allowing biostratigraphic correlation both within the Central Paratethys and to the global standard [16,17].

As the Nógrád–Filakovo Basin straddles the Hungarian–Slovak state border, a partly different lithostratigraphic nomenclature is used in the Slovak literature [18]. Of interest are the Lipovany Sandstone Formation, equivalent of the Pétervására/Budafok formations, the Bukovinka Formation, equivalent of the Zagyvapálfalva Formation, and the "*Rzehakia* beds", referred to as the Medokys Member of Modry Kamen Formation in Slovakia and the Kazár Member of Egyházasgerge Formation in Hungary.

3. Radio-isotopic dating

3.1. U–Pb dating

A tuff sample was taken at Borókás-árok, from an outcrop beside the trail leading to the Exhibition Hall erected above the exposed track-bearing sandstone (Fig. 1). Standard mineral separation techniques were employed to extract zircons. The sample yielded abundant, colorless, clear zircons of elongated prismatic or needle-like morphology. The crystals were carefully examined in transmitted light using a petrographic microscope. In order to avoid averaging effects caused by older inheritance, xenocrystic contamination and Pb loss (or a combination of all), zircons were analyzed individually by low blank, isotope dilution thermal ionization mass spectrometry (IDTIMS) analytical techniques. Individual crystal weights range from 3 to 8 μg and have a median U concentration of c. 800 ppm. Analytical protocols follow those described in [19]. The total U–Pb isochron approach of Ludwig [20] was employed to extract the age of the sample which tests for both closed-system behaviour (i.e. concordance) and the assumption of an invariant common Pb and yields the smallest justifiable age-error of any possible U–Pb or Pb/Pb isochron.

A total isochron constrained by 9 analyses on individual zircons yields an age of 17.42 ± 0.04 Ma (2σ) with an MSWD of 1.3 (Fig. 3). One analysis showed an elevated common Pb concentration and was excluded (although including it does not change the result). The isochron is constrained by a conservative estimate of the common Pb with $^{207}Pb/^{206}Pb = 0.84$ (±1.68%) and $^{204}Pb/^{206}Pb = 0.05464$ (±5.5%). Rejecting the analyses with the largest uncertainties (Z03 and Z09) results in a slightly elevated MSWD (1.7) but essentially the same age and uncertainty. Alternatively, a weighted mean $^{206}Pb/^{238}U$ age (corrected for common Pb, see Table 1) yields the same age of 17.42 ± 0.06 Ma (2σ), also with a slightly elevated MSWD of 2.1. Scatter in excess of the analytical uncertainty may be an indication for minute amounts of Pb loss, may reflect pre-eruptive residence time of the zircons, or a combination of both [21]. Individual isotopic ratios including ^{206}Pb are corrected for excess ^{230}Th during the crystallization of the zircons (with an assumed Th/U of 4 in the host rock) that results in an age bias of 90 ka [22]. Varying the Th/U (e.g. to a range of 2 to 6) has only a minor effect on the age bias and thus the final age (20 ka at maximum), which is small compared to the possible presence of pre-eruptive residence time (see below for discussion).

3.2. $^{40}Ar/^{39}Ar$ dating

Tuff samples were collected for $^{40}Ar/^{39}Ar$ dating from Ipolytarnóc and Nemti. The GRTF overlying the track-bearing sandstone at Ipolytarnóc was sampled in an outcrop near the Research Station at the trail entrance leading to the Exhibition Hall (Fig. 1). The U–Pb and $^{40}Ar/^{39}Ar$ samples were taken from stratigraphically equivalent horizons at sites only 250 m apart. An additional sample was collected near Nemti, at another important fossil locality of similar age. The sampled outcrop of GRTF is a roadcut near the Rákóczi-telep storage facility of the Hungarian Geological Institute.

Both tuff samples (from Ipolytarnóc and Nemti) selected for $^{40}Ar/^{39}Ar$ dating are crystal-rich rhyolites with phenocrysts mainly of quartz, biotite, and plagioclase. Biotite in both samples shows substantial alteration and was hence considered unsuitable for dating. The plagioclase in both samples is optically clear and free of alteration, hence was selected for analysis.

Table 2
$^{40}Ar/^{39}Ar$ analytical data

Run ID	^{40}Ar (moles)	^{40}Ar (nA)	$\pm\sigma$ (nA)	^{39}Ar (nA)	$\pm\sigma$ (nA)	^{38}Ar (nA)	$\pm\sigma$ (nA)	^{37}Ar (nA)	$\pm\sigma$ (nA)	^{36}Ar (nA)	$\pm\sigma$ (nA)	$^{40}Ar^*/^{39}Ar_k$	$\pm\sigma$	$\%^{40}Ar^*$	Age (Ma)	$\pm\sigma$ (Ma)
Nemti																
6711-01A	3.97E−14	8.5669	0.0064	0.01256	0.00012	0.00563	0.00004	0.0454	0.0001	0.02832	0.00005	14.911	2.233	2.2	19.11	2.85
6711-01B	1.14E−14	2.3253	0.0024	0.09954	0.00014	0.00191	0.00003	0.7112	0.0023	0.00366	0.00002	13.081	0.087	55.7	16.77	0.11
6711-02	1.21E−13	25.2952	0.0170	0.22773	0.00049	0.01721	0.00006	1.5376	0.0031	0.07564	0.00008	13.335	0.313	11.9	17.10	0.40
6711-03A	8.71E−14	18.1902	0.0120	0.02196	0.00018	0.01176	0.00006	0.0970	0.0002	0.06056	0.00013	12.353	2.942	1.5	15.84	3.76
6711-03B	5.33E−15	1.1454	0.0008	0.07383	0.00013	0.00097	0.00002	0.3737	0.0008	0.00061	0.00002	13.504	0.102	86.8	17.31	0.13
6711-04A	6.98E−14	14.9915	0.0100	0.01067	0.00013	0.00966	0.00005	0.0456	0.0003	0.05021	0.00010	12.610	4.892	0.9	16.17	6.25
6711-04B	1.26E−14	2.5831	0.0028	0.08661	0.00019	0.00209	0.00003	0.7723	0.0019	0.00510	0.00003	13.146	0.115	43.8	16.85	0.15
6711-05A	2.05E−14	4.2323	0.0037	0.00546	0.00011	0.00274	0.00004	0.0271	0.0002	0.01410	0.00004	11.437	3.027	1.5	14.67	3.87
6711-05B	3.03E−16	0.0660	0.0006	0.00222	0.00007	−0.00003	0.00002	0.0165	0.0001	0.00012	0.00002	14.839	2.938	49.7	19.01	3.75
6711-05C	3.54E−15	0.5850	0.0014	0.00607	0.00007	0.00036	0.00002	0.0350	0.0002	0.00169	0.00002	14.655	1.271	15.1	18.78	1.62
6711-06A	2.60E−14	5.3446	0.0049	0.00940	0.00011	0.00342	0.00003	0.0384	0.0001	0.01773	0.00004	10.518	2.137	1.8	13.50	2.73
6711-06B	2.32E−16	0.0697	0.0005	0.00475	0.00006	0.00006	0.00002	0.0297	0.0001	0.00003	0.00001	13.470	0.940	91.5	17.27	1.20
6711-06C	5.41E−16	0.1041	0.0004	0.00612	0.00007	0.00009	0.00002	0.0372	0.0002	0.00012	0.00002	11.536	0.856	67.5	14.80	1.09
6711-07A	2.35E−14	5.0791	0.0042	0.00335	0.00009	0.00328	0.00003	0.0193	0.0001	0.01698	0.00004	16.694	5.508	1.1	21.38	7.01
6711-07B	1.11E−16	0.0317	0.0005	0.00213	0.00006	0.00006	0.00002	0.0201	0.0001	0.00003	0.00002	11.914	2.551	79.6	15.28	3.26
6711-07C	3.72E−14	5.5403	0.0170	0.10169	0.00027	0.00400	0.00005	0.8404	0.0069	0.01407	0.00026	14.241	0.792	26.0	18.25	1.01
6711-08	2.50E−15	0.5142	0.0012	0.01544	0.00009	0.00044	0.00002	0.0714	0.0002	0.00106	0.00003	13.448	0.646	40.3	17.24	0.82
6711-09	2.87E−14	6.1060	0.0110	0.07065	0.00013	0.00425	0.00003	0.3104	0.0006	0.01762	0.00004	12.965	0.314	15.0	16.62	0.40
6711-10	1.40E−14	3.0078	0.0027	0.00823	0.00009	0.00198	0.00003	0.0434	0.0002	0.00982	0.00003	12.556	1.610	3.4	16.10	2.06
6711-11	1.50E−15	0.2553	0.0006	0.01069	0.00009	0.00021	0.00002	0.0686	0.0002	0.00039	0.00002	13.612	0.588	56.8	17.45	0.75
Ipolytarnóc																
6712-01	1.24E−14	2.5504	0.0022	0.08519	0.00020	0.00200	0.00003	0.5179	0.0011	0.00490	0.00002	13.435	0.108	44.7	17.22	0.14
6712-02	1.81E−14	3.5883	0.0058	0.01817	0.00012	0.00244	0.00004	0.0828	0.0004	0.01130	0.00003	13.749	0.840	6.9	17.62	1.07
6712-02B	5.88E−16	0.2199	0.0009	0.02457	0.00014	0.00053	0.00002	0.0037	0.0001	0.00057	0.00002	2.035	0.268	22.8	2.62	0.35
6712-03	7.83E−14	16.4613	0.0140	0.08701	0.00030	0.01095	0.00005	0.7433	0.0019	0.05163	0.00007	14.289	0.578	7.5	18.31	0.74
6712-04	8.39E−15	1.7692	0.0016	0.00632	0.00008	0.00117	0.00003	0.0505	0.0003	0.00572	0.00003	12.521	1.474	4.4	16.06	1.88
6712-04B	3.19E−15	0.6115	0.0021	0.04152	0.00039	0.00062	0.00002	0.3216	0.0021	0.00032	0.00002	13.085	0.182	88.4	16.78	0.23
6712-05	6.74E−15	1.3812	0.0015	0.03855	0.00016	0.00104	0.00002	0.3626	0.0010	0.00306	0.00002	13.163	0.196	36.5	16.88	0.25
6712-06	4.72E−14	9.8979	0.0065	0.07204	0.00020	0.00674	0.00005	0.6697	0.0018	0.03032	0.00005	13.637	0.419	9.9	17.48	0.53
6712-07	7.06E−15	1.4872	0.0015	0.01390	0.00010	0.00098	0.00002	0.0550	0.0002	0.00448	0.00002	11.839	0.581	11.0	15.19	0.74
6712-08	9.91E−15	2.0441	0.0021	0.00475	0.00008	0.00137	0.00003	0.0329	0.0002	0.00675	0.00003	10.516	2.198	2.4	13.50	2.81
6712-09	3.02E−15	0.6667	0.0017	0.04769	0.00016	0.00068	0.00002	0.4154	0.0010	0.00023	0.00002	13.330	0.117	94.8	17.09	0.15
6712-10	5.15E−15	1.0436	0.0022	0.03626	0.00013	0.00085	0.00002	0.1800	0.0006	0.00195	0.00003	13.259	0.264	45.9	17.00	0.34
6712-11	4.77E−15	1.0840	0.0021	0.05115	0.00013	0.00098	0.00002	0.4188	0.0007	0.00148	0.00002	13.359	0.128	62.7	17.13	0.16
6712-12	1.84E−15	0.4076	0.0012	0.00840	0.00008	0.00031	0.00002	0.0291	0.0003	0.00102	0.00002	12.863	0.619	26.4	16.49	0.79
6712-12B	2.28E−15	0.5060	0.0010	0.03613	0.00013	0.00047	0.00002	0.1923	0.0004	0.00015	0.00002	13.261	0.137	94.4	17.00	0.17
6712-13	7.94E−15	1.6154	0.0015	0.04498	0.00013	0.00137	0.00003	0.2987	0.0006	0.00358	0.00002	12.945	0.174	35.9	16.60	0.22
6712-14	2.91E−15	0.6588	0.0010	0.04841	0.00016	0.00063	0.00002	0.2756	0.0006	0.00013	0.00002	13.293	0.159	97.4	17.04	0.20

Plagioclase was separated by conventional methods and crystals from the 710 to 1700 μm size fraction were irradiated in the Omega West reactor at Los Alamos National Laboratory for 3 h along with the Fish Canyon sanidine (FCs; 28.02 Ma; [23]) as a neutron fluence monitor. The samples were irradiated in wells in an Al disc similar to those figured in [23]. The two samples were irradiated in adjacent wells bracketed by the FCs standard. Identical J-values of 0.0007142±0.0000016 and 0.0007144±0.0000014 (2σ) were determined from the weighted mean of data from 7 crystals of the sanidine from each standard position.

The plagioclase crystals were degassed individually by fusion with an argon-ion laser, and after gettering the evolved gas, the relative abundances of argon isotopes were measured on an MAP 215C gas source magnetic sector mass spectrometer. Mass discrimination (1.0082±0.0014 per amu, 2σ) was monitored by analysis of 16 air pipette aliquots interspersed with the samples. Several crystals from each sample were degassed in 2–3 steps by incrementally increasing the laser power. Ar isotope data are shown in Table 2. Age calculations are based on 28.02 Ma [23] for the standard and the decay constants and isotope abundances of [24]. Crystals from the two tuffs are compositionally similar, with Ca/K ranging from 6–20 as deduced from the ^{37}Ar/^{39}Ar data. Isotope correlation diagrams (Fig. 4A,B) including all data from each sample yield robust isochrons [25] with ages of 17.02±0.14 and 16.99±0.16 Ma (2σ) for the Ipolytarnóc and Nemti tuffs, respectively, and atmospheric trapped components for both. Both isochrons have acceptable probabilities of fit (15% and 24%). The Ipolytarnóc and Nemti tuffs are indistinguishable in age and based on their similar petrography and Ca/K data for plagioclase, plus the probable wide distribution of such units, could represent a single ignimbrite eruption. However, correlation of the two tuffs is apparently at odds with paleomagnetic data indicating that they record opposite polarity (see below). It should be noted that our preliminary results were cited inaccurately and without permission by Goldsmith et al. [26]. The ages cited therein should be ignored as they are superceded by the present data.

4. Discussion

4.1. Difference between U–Pb and ^{40}Ar/^{39}Ar results

The U–Pb age (17.42±0.04 Ma) obtained for the Ipolytarnóc tuff is significantly older than the ^{40}Ar/^{39}Ar age (17.02±0.14 Ma). The 2.3 % difference of 0.40±0.15 Ma between these two ages is comparable in sense to that increasingly observed between U–Pb ages of zircon and ^{40}Ar/^{39}Ar ages for various minerals in volcanic rocks (e.g. as noted by [27,28]). It was shown that such a difference is wholly consistent with existing uncertainties in the ^{40}K decay constants and calibration data for standards, but also noted that, particularly for relatively young rocks such as in this study, a significant but unknown proportion of the bias could reflect pre-eruptive residence time of zircons in the magma.

Because pre-eruptive residence time of several hundred ka is well documented for zircons in silicic magma systems [21,29,30], we cannot confidently ascribe superior accuracy *a priori* to the U–Pb age compared with the ^{40}Ar/^{39}Ar age reported herein for the Ipolytarnóc tuff. In light of these considerations, we conclude that the U–Pb zircon age represents a maximum, and the ^{40}Ar/^{39}Ar age a minimum, age for the eruption. For comparison with Neogene time scales, which are mainly calibrated with ^{40}Ar/^{39}Ar data, our ^{40}Ar/^{39}Ar ages are more appropriate provided that comparisons are based on the same standard age basis. It should be noted, however, that the observed bias between the U–Pb and ^{40}Ar/^{39}Ar ages (and thus the

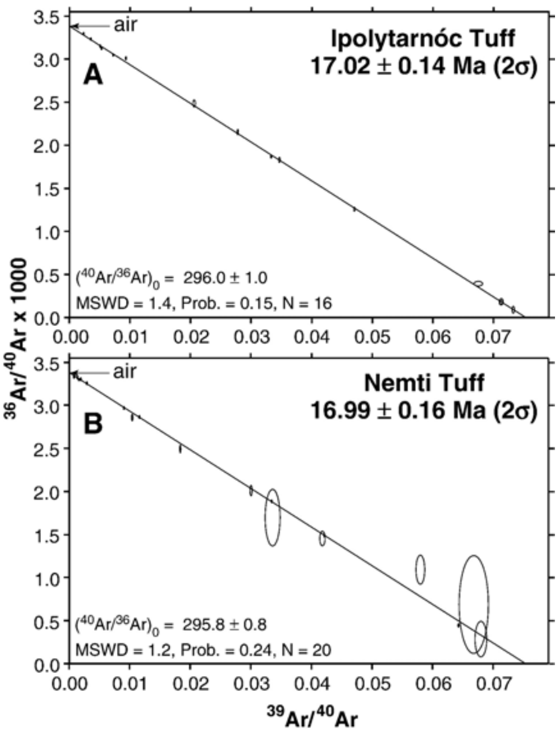

Fig. 4. Isotope correlation (inverse isochron) diagrams for plagioclase from the Ipolytarnóc (A) and Nemti (B) tuff samples. Error ellipses show 1σ errors. The ^{40}Ar/^{36}Ar of atmospheric argon (air) is shown by arrows. Diagrams include both stepwise and total fusion degassing data.

miscalibration of ages based on the K/Ar system) exceeds the uncertainties reported here. We anticipate that future studies will help to constrain the magnitude of this bias and, in the meantime, caution against the implicit use of uncorrected $^{40}Ar/^{39}Ar$ ages.

4.2. Implications for timing the Miocene volcanism of the inner Carpathian arc

Neogene volcanism in the Pannonian Basin started with voluminous, explosive eruptions that produced areally extensive silicic ignimbrite sheets [15,31]. It is widely held that the ignimbrites, of which the Gyulakeszi Rhyolite Tuff Formation (GRTF) is the oldest, were formed during three distinct volcanic episodes. The previously established numeric age of the tuff at Ipolytarnóc relies on the assumptions that it belongs to the GRTF which is in turn a product of a single eruption or a short volcanic episode. The GRTF has long been regarded as a stratigraphic marker horizon [32]. The average age of numerous K–Ar ages on GRTF, 19.6 ± 1.4 Ma, was therefore inferred as the age of the Ipolytarnóc tuff [6]. Published K–Ar dating of samples collected from Ipolytarnóc include 20.0 ± 2.0 Ma on biotite (average of two analyses) and 19.8 ± 3.0 Ma on plagioclase (errors are 1σ) [6]. Similarly, a sample from Nemti was dated as 20.9 ± 1.8 Ma (average of two analyses on biotite) [6]. However, another K–Ar age of 17.6 ± 0.8 Ma was also reported from this locality [33]. Unpublished data from the same laboratory show a significant scatter towards younger ages: the K–Ar ages of 6 additional samples from Ipolytarnóc vary between 5.7 ± 4.0 and 16.3 ± 1.6 Ma [34]. From nearby Lipovany in Southern Slovakia, fission track ages of 20.6 ± 0.5 and 20.1 ± 0.3 Ma were determined on rhyolite tuffs from the Lipovany Sandstone and the Bukovinka Formation, respectively [35].

Our new U–Pb and $^{40}Ar/^{39}Ar$ dates of ~17 Ma raise questions about the accuracy and reliability of previous K–Ar results, and/or the true synchrony of all GRTF tuff and the assumed short duration of the first volcanic episode producing the GRTF. Published K–Ar age distribution histograms of the large dataset on all pyroclastics from northern Hungary reveal a continuum of numeric ages between the lower GRTF and the middle and upper tuff horizons [6,36]. Considering the clear differences in their stratigraphic relationships, it may be taken as evidence of commonly disturbed isotopic systems and resulting dating inaccuracies that seriously limit the resolving power of the K–Ar method. This is further illustrated by the only $^{40}Ar/^{39}Ar$ age spectrum previously reported on biotite from Ipolytarnóc, which evidently lacks a plateau and yields a total fusion age of 19.0 ± 1.4 Ma [37].

Does it follow that our new, high-precision dates on the Ipolytarnóc and Nemti tuffs could be taken as the age of the GRTF? Caution is required as various lines of evidences were proposed recently which indicate paleomagnetic, geochemical, and petrologic differences between tuffs previously assigned to GRTF. Paleomagnetic studies established temporally well constrained block rotations during the tectonic evolution of the Carpathian–Pannonian region [33,36]. The oldest tuffs, assigned to GRTF, are of generally reverse polarity and show evidence of 70–90° westerly declination rotation. Younger, upper Ottnangian to Karpatian sedimentary rocks and pyroclastics of the "Middle Rhyolite Tuff" are characterized by only 30° rotation in the same sense. The anomalous behaviour of samples from Ipolytarnóc was noted as three ignimbrite samples display normal polarity and only 30° rotation [33]. Normal polarity, but a 90° rotation history, was determined from the underlying glauconitic sandstone [33]. Further studies are in progress that also involve samples from the track-bearing sandstone (E. Márton, pers. comm.). Ignimbrite and variegated clay samples from Nemti show reverse polarity and evidence of large rotation [33]. Thus paleomagnetic data may indicate that the age of tuff at Ipolytarnóc is different from the GRTF elsewhere. If the aforementioned paleomagnetic polarity data are valid, then the ages of the Nemti and Ipolytarnóc tuffs must be sufficiently different as to encompass a geomagnetic polarity reversal. An age difference of 0.03 ± 0.21 Ma is permitted by our data, which easily accommodates this possibility. However, according to most modern polarity time scales [38] both of these ages are expected to fall in the middle of subchron C5Cr, a reversed polarity interval of about 0.6 Ma duration (Fig. 5). Resolution of these possible inconsistencies is beyond the scope of the present paper, but should be addressed in future studies.

Instead of the three distinct volcanic episodes, nearly continuous eruptions from multiple volcanic centres were suggested as an alternative model for the Neogene silicic volcanism in Hungary [15]. Differences in rare earth element abundance patterns and zircon morphology point to different parent magmas, whereas physical volcanology indicates a variety of genetic modes for pyroclastics, all suggesting a composite eruption history for the GRTF [39]. A pyroclastic complex of several ignimbrite sheets, Plinian and phreatomagmatic eruption products, was also inferred from observations on the GRTF from the Bükk Foreland [40].

The new radio-isotopic dates suggest that at least parts of the GRTF are significantly younger than previously thought. If some of the reported K–Ar ages are accurate, then a complex and prolonged eruption history can be hypothesized. Confirmation will require a more comprehensive set of new U–Pb or $^{40}Ar/^{39}Ar$ dates from different outcrop areas of the GRTF.

4.3. Implications for the Miocene time scale and correlation of depositional sequences and Paratethys stages

Although the dated pyroclastics were deposited in a terrestrial environment, they are bracketed by marine strata amenable to biostratigraphic dating. The new dates are thus useful for calibrating the Paratethys chronostratigraphic scale and also, to a lesser extent, the standard geological time scale.

The Pétervására Sandstone underlying the track–bearing sansdstone yielded a diverse mollusk fauna of "Loibersdorf-type" that allows confident correlation with the Eggenburgian at its type area in the Austrian Molasse basin [13]. The youngest marine nannoplankton, derived from only a few meters below the footprint-bearing sansdstone, is assigned to the NN3 zone [41]. Similarly, the uppermost part of the Lipovany Formation at a nearby Slovak locality contained *Sphenolithus belemnos*, whose first appearance defines the base of NN3 zone [42].

Because strata of the younger marine sedimentary cycle are not preserved at Ipolytarnóc, an upper biostratigraphic bracket can only be inferred from elsewhere in the region. In Hungary, the oldest marine fossils above the coal-bearing Salgótarján Formation occur in the *Rzehakia* beds (Kazár Member of Egyházasgerge Formation). Apart from the endemic bivalves, nannoplankton including *Sphenolithus heteromorphus* (but not *S. belemnos*) also occur, indicating the NN4 zone [16]. The same conclusion was reached from a study of the equivalent strata in Slovakia (Medokys Member) [42]. The *Rzehakia* horizon is widespread in the Central Paratethys and is thought to represent the latest Ottnangian [17]. In Slovakia, marine ingressions occurring within the Salgótarján Formation (Plachtince Member) yielded *S. heteromorphus* together with *S. belemnos*, indicating the top of the NN3 nannoplankton zone [42]. The dated tuff therefore appears correlative with the NN3 nannoplankton zone, and permits correlation with both the standard and the regional time scale. However, this correlation remains at odds with the

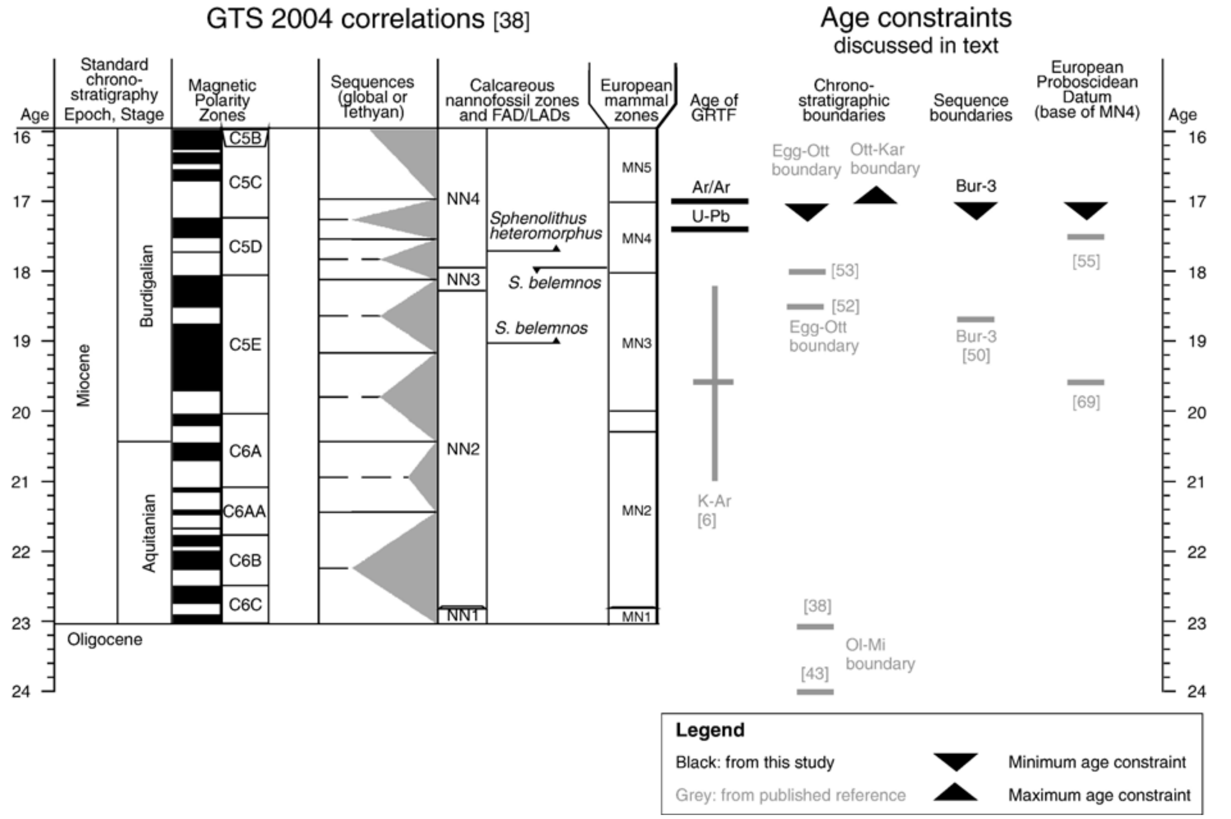

Fig. 5. The radio-isotopic ages of the GRTF in the framework of chronostratigraphic correlation of the Early Miocene (after [38]). Implied correlations with Paratethys chronostratigraphy, regional sequence stratigraphy, and mammalian biochronology are also shown. See text for discussion.

numeric age of the tuff (Fig. 5). In the biochronological framework of Central Paratethys stages, the age of the tuff cannot be older than late Eggenburgian or younger than late Ottnangian.

The numeric calibration of Early Miocene standard time scale, and the age of the Oligocene/Miocene boundary in particular, has been the subject of recent debate. The previously most widely used time scale (based on the FCs standard at 27.84 Ma) quotes 23.8 Ma for this epoch boundary [43]. This age recalculates to 24.0 Ma for comparison with the FCs age of 28.02 Ma [23] employed herein. A proposed astronomical calibration suggests a −0.9 Myr revision to 22.9±0.1 Ma [44]. The most recent geological time scale provides an age of 23.03 Ma for this boundary [38]. Support for both the older [45] and the younger [46] boundary age has been presented. Our radioisotopic dates with nannoplankton biostratigraphic constraints may help resolve the controversy because the discrepancy, albeit somewhat dampened, propagates to the early Miocene part of the time scale. To the oldest relevant nannoplankton datum, the first occurrence (FO) of *S. belemnos,* an astronomically tuned age of 18.9 Ma [47] was assigned versus 19.2 Ma in the earlier time scale [43]. For the last occurrence (LO) of this species, the same scales list 17.94 versus 18.3 Ma, whereas the FO of *S. heteromorphus* is determined as 17.70 versus 18.2 Ma. Clearly, our data compare more favourably with the younger ages of the astronomically calibrated scale. The apparent miscorrelation that arises from numeric calibration of the NN3 nannozone [38] and the radio-isotopic ages of GRTF (Fig. 5) remains to be resolved.

The new dates also provide a calibration point for regional sequence stratigraphic schemes. Several attempts have been made to interpret the depositional sequences of basins in the Carpathian–Pannonian area [48–50] and to correlate them with the global sequence charts [51]. The Miocene 3rd order depositional cycles resulted from an interplay of regional tectonic and global eustatic forcing. Although the Eggenburgian–Ottnangian cycle is broadly correlated to the global TB2.1 cycle, the reconstructed Late Eggenburgian sea level drop is a local phenomenon that records the overprint of regional tectonics [49]. Thus, in the North Pannonian Basin a regional unconformity (Bur-3) is recognized at the Eggerburgian–Ottnangian boundary, which is within the falling stage of the TB2.1 cycle and does not appear to correspond to any globally observed sequence boundary [50].

At this time in the Central Paratethys basin, a short interval of isolation was also recognized [42,50] which corresponded to increased continental communication (the establishment of the "*Gomphotherium* landbridge" [52]) that allowed rapid westward migration of mammals of African origin into Europe (see Section 4.4).

Locally at Ipolytarnóc, we equate the Bur-3 sequence boundary with the erosional base of the basal conglomerate of the Zagyvapálfalva Formation, immediately below the track-bearing sandstone. Therefore, the 17.0 Ma $^{40}Ar/^{39}Ar$ age of the GRTF provides a close minimum estimate for the age of this unconformity that may locally represent a hiatus of significant duration. (Note that the age of the Bur-3 was previously estimated at 18.7 Ma [50]).

Following deposition of the largely continental Salgótarján Formation, the latest Ottnangian–Karpatian marine transgression is correlated to the global TB2.2 sequence whose lower boundary was dated as 17.5 Ma [51]. Accepting this correlation would require a revision of the age of this global sequence boundary towards younger ages, in the same sense as suggested by the most recent global chronostratigraphic scale [38].

Lastly, the obtained radioisotopic dates are useful for calibrating the Eggenburgian–Ottnangian stage boundary within the Central Paratethys chronostratigraphy. In Hungary the GRTF has been conventionally regarded as the base of Ottnangian [6]. Commonly accepted age estimates for this stage boundary are 18.5 Ma [52] or 18.0 Ma [53]. It appears that the true age of this boundary is somewhat younger and falls between 17.4 and 17.0 Ma.

4.4. Implications for mammalian evolution and migration: the age of the Proboscidean Datum and correlation of MN zones

The Early Miocene is a critical interval for dispersal of mammals of African origin in Europe, following the initial establishment of a landbridge between the Afro-Arabian and Eurasian plates [54]. The most remarkable event is the dispersal of proboscideans recognized as the Proboscidean Datum [55]. This term was originally coined on the basis of a claim that the extension of this African group into Eurasia was a synchronous event. Multiple lineages of proboscideans extended their ranges into Europe at or near the Proboscidean Datum, including the mammutid *Zygolophodon*, the mastodont *Gomphotherium* and the deinothere *Deinotherium* [56]. Based on the first such finds from Portugal, the Proboscidean Datum was originally dated as 17.5 Ma [55], but later studies proposed a more prolonged migration event rather than a sharp datum [54]. In the Neogene mammal biochronologic scheme for Europe, the first occurrence of proboscideans was originally used to define the base of the MN4 zone [57,58]. Alternative interpretations identified it as the base of subzone MN3b [59], or distinguished the FOD of *Gomphotherium* marking the base of MN3 zone from the FOD of *Prodeinotherium* indicating MN4 [60]. Most recently, the European Proboscidean Datum was correlated with MN4 zone [58], which in turn was assigned a numeric age between 18.0 and 17.0 Ma, correlative of the Central Paratethyan Ottnangian and lowermost Karpathian stages [61]. Our results confirm correlation of the European Proboscidean Datum with the MN4 zone.

The validity of a single Old World Proboscidean Datum has been called into question by a number of authors based on the assumption that a landbridge that facilitated proboscidean extension from Africa to South Asia likely preceded the "*Gomphotherium* landbridge" that accommodated their extension into Europe by as much as several million years [54,62–64]. New evidence on an earlier Proboscidean Datum in South Asia includes the first occurrence of proboscideans in the late Oligocene at Dera Bugti in Baluchistan, western Pakistan [65]. The first occurrence of proboscideans in the Zinda Pir Dome of the Sulaiman Range, also in Pakistan, was recently dated as either latest Oligocene (correlative with Chron 8n, >26 Ma), or earliest Miocene (correlative with Chron 6Bn, <23 Ma) [66]. Mammalian biochronologic correlation between the Zinda Pir Dome and Dera Bugti faunas support the older, late Oligocene age of the Baluchistan faunas. As a result, the Proboscidea have been demonstrated to first occur much earlier in South Asia than in Europe. Recent discovery of a diverse proboscidean assemblage at Chilga, in northwestern Ethiopia, provides important new data on both the geographic source and chronology of ancestors of first occurring Eurasian proboscideans. The Chilga section yielded an $^{40}Ar/^{39}Ar$ age of 27.36± 0.11 Ma and is magnetostratigraphically correlated with Chron C9n (27.972–27.027 Ma) [67,68]. The proboscidean fauna is diverse and includes the most primitive known members of the Deinotheridae (*Chilgatherium harrisi*), Mammutidae (*Palaeomastodon* A and B), and Gomphotheriidae (*Gomphotherium* sp.) In addition, the Chilga faunas lack any evidence of Eurasian immigrants suggesting that a Late Oligocene–Early Miocene faunal exchange had not yet occurred. The South Asian data from Pakistan together with the Chilga fauna support a South Asian Proboscidean Datum within the Late Oligocene between 27 and 23 Ma.

Although Ipolytarnóc and the 19.6 Ma average age of the GRTF was previously used in arguments for an older date of the European Proboscidean Datum [54], the existence of proboscidean tracks were disputed [69]. Nevertheless, body fossils are known from the Zagyvapálfalva Formation, below the GRTF, from two localities. Teeth and tusks of *Gomphotherium angustidens* were collected at Nemti and *Prodeinotherium hungaricum* occurs near Salgótarján [70]. The fossils at Nemti were found below the dated tuff, hence our $^{40}Ar/^{39}Ar$ age provides a minimum age constraint for the Proboscidean Datum and the lower limit of the MN4 zone. As the $^{40}Ar/^{39}Ar$ ages from the two localities (Nemti and Ipolytarnóc) are indistinguishable, the likelihood that the track-bearing sandstone at Ipolytarnóc is marginally older, and hence predates the European Proboscidean Datum, remains small. Alternatively, it is more probable that proboscideans may have already lived in the area but did not frequent this site on the ancient riverbank. Similarly, a modern study in Africa found elephant tracks extremely rare at certain sites near bodies of water [71].

Previously only single European localities were radio-isotopically dated from both the MN3 and MN4 zones [53]. At Beaulieu (France), basalts $^{40}Ar/^{39}Ar$ dated as 17.5±0.3 Ma are contemporaneous with fossiliferous sediments that yielded the youngest assemblage where proboscideans and modern cricetids are still absent, assigned to the top of the MN3 zone [72]. (Note that comparison of our results with the date for Beaulieu is somewhat tenuous, as the latter is based on the Caplongue Hb standard (H. Maluski, written communication, 2004) which has not been intercalibrated with FCs standard to our knowledge.) In the coal mine at Belchatów (Poland), lacustrine sediments contain a rich small mammal assemblage together with *G. angustidens*, indicating the MN4 zone [73]. The horizon is bracketed by tuff layers that yielded average fission track ages of 17.3±0.4 and 17.0±0.7 Ma, respectively [74]. The dates presented here are in agreement with the Beaulieu and Belchatów dates and help establish a more tightly constrained lower age limit of the MN4 zone at 17.4 Ma. Previous correlation schemes suggested a somewhat older age of 18.0 Ma [53,75] or a younger age of 16.6 Ma [60].

The frequently cited argument that Ipolytarnóc yields the oldest European proboscidean record should be abandoned, as it was based on the now disputed presence of proboscidean tracks and an inferred average age of the tuff that is too old by more than 2 Ma. Yet the date from Nemti is useful for correlating early proboscidean faunas of the MN4 zone in Europe and assist in reaffirming the significant diachroneity of the European and South Asian Proboscidean Datums. Our data remains compatible with the originally proposed ~17.5 Ma age of the European Proboscidean Datum

[55] and suggest that proboscideans probably migrated from Africa into Europe significantly later than into South Asia [54,76]. The Hungarian faunas are now the most precisely and accurately dated among the European faunas that contain early occurring proboscideans correlated with the MN4 zone.

5. Conclusions

(1) We obtained a single-crystal zircon U–Pb age of 17.42±0.04 Ma and a laser-fusion plagioclase $^{40}Ar/^{39}Ar$ age of 17.02±0.14 Ma from the Gyulakeszi Rhyolite Tuff Formation (GRTF). These dates are regarded as the age of an exceptional Early Miocene fossil track site and plant assemblage that was preserved by the emplacement of the ignimbrite. Our results demonstrate that single-crystal U–Pb dating has advanced to become a powerful tool for high-precision age determination of zircons as young as Neogene.

(2) We also obtained an $^{40}Ar/^{39}Ar$ age of 16.99±0.16 Ma from an equivalent rhyolite tuff near Nemti, where the underlying variegated clay yielded early proboscidean remains assigned to the MN4 mammal zone.

(3) The difference of 0.40±0.15 Ma between the Ipolytarnóc U–Pb and $^{40}Ar/^{39}Ar$ ages is explained by the combined effects of possible pre-eruptive residence time of zircon and the inaccurately known ^{40}K decay constant that causes a 1–2% bias when $^{40}Ar/^{39}Ar$ and U–Pb ages are compared.

(4) The previously accepted 19.6±1.4 Ma age for the Ipolytarnóc tuff, based on an average K–Ar age for the GRTF, is significantly revised and superseded by our dating results.

(5) Further dating is needed to ascertain whether the ages reported here are representative for the entire GRTF or they correspond to younger eruptions of a volcanic episode that was more prolonged than previously thought.

(6) Published marine biostratigraphic data allow correlation of the Ipolytarnóc track-bearing sandstone with the upper part of NN3 nannoplantkton zone. Together with the radio-isotopic ages, they suggest correlation with the Ottnangian. Conventional correlation with the base of Ottnangian therefore needs reconsideration.

(7) The regional Bur-3 unconformity below the track bearing sandstone is likely to represent a hiatus of significant duration, whereas deposition of the continental Salgótarján Formation that overlies the GRTF was rapid.

(8) Our data support the recently suggested astronomical calibration of the Early Miocene time scale that revised the previous scales towards younger ages.

(9) The 16.99±0.16 Ma $^{40}Ar/^{39}Ar$ age of GRTF at Nemti provides a reliable correlation of MN4 mammal zone with the numeric time scale, as the dated tuff overlies terrestrial deposits that yielded *G. angustidens*. This date provides a key minimum constraint for the age of the European Proboscidean Datum, the migration event of proboscideans from Africa to Europe through the emerging "*Gomphotherium* landbridge". Contrary to suggestions for a significantly earlier European datum, it appears that the originally suggested age of c. 17.5 Ma is realistic.

(10) Our age for the European Proboscidean Datum reaffirms that it is substantially younger than the South Asian Proboscidean Datum, and that the previously implied Eurasian synchroneity of the exit of proboscideans from Africa should be set aside.

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