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Geological evolution of the southwestern part of the Veporic Unit (Western Carpathians): based on fission track and morphotectonic data

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Abstract: Zircon and apatite fission track (FT) and morphotectonic analyses were applied in order to infer quantitative constraints on the Alpine morphotectonic evolution of the western part of the Southern Veporic Unit which is related to: (1) Eo-Alpine Cretaceous nappe stacking and metamorphism of the crystalline basement in the greenschist facies. (2) Exhumation phase due to underthrusting of the northerly located Tatric-Fatric basement (~ 90–80 Ma), followed by a passive en-block exhumation with cooling through ~320–200 °C during the Palaeocene (ZFT ages of ~61–55 Ma). (3) Slow Eocene cooling through ~245–90 °C, which most likely reflected erosion of the overlying cover nappes and the Gosau Group sediments. Cooling reached up to 60 °C till the Oligocene (AFT ages of ~37–22 Ma) in association with erosion of cover nappes. The efficient Eocene erosion led to the formation of the first Cenozoic planation surface with supergene kaolinization in many places. (4) The early Miocene erosion coincided with surface lowering and resulted in the second planation surface favourable for kaolinization. (5) In the middle Miocene, the study area was covered by the Poľana, Javorie, and Vepor stratovolcanoes. (6) The late Miocene stage was related to the erosion and formation of the third Cenozoic planation surface and the final shaping of the mountains was linked to a new accelerated uplift from the Pliocene.

Keywords: Western Carpathians, Veporic Unit, morphotectonic evolution, fission track analysis, planation surfaces, exhumation.

Introduction

Following the collision and nappe stacking processes during the Alpine orogeny, the study area underwent an episode of exhumation as a result of such factors as compressive tectonics, post-orogenetic unroofing, and isostatic readjustment. Modern measurement techniques, such as zircon and apatite fission track analyses, have helped to establish useful exhumation and denudation chronologies.

The Western Carpathians occupy the north-eastern part of the Alpine orogen of Europe. In the west, the Western Carpathians are connected with the Eastern Alps and share a similar Variscan and Alpine tectonic evolution. They are traditionally divided into three principal parts — External, Central, and Internal (e.g., Plašienka et al. 1997, 1999; Froitzheim et al. 2008), or two principal parts — Outer and Inner Western Carpathians (e.g., Mišík et al. 1985 pp. 304–344; Biely 1989; Bezák et al. 2004; Hók et al. 2014), depending on application of either Mesozoic or Cenozoic structure, respectively.

The Veporic Unit represents the middle of the thick-skinned thrust sheets (a.k.a. the Middle Group of Nappes — cf. Hók et al. 2014) incorporated into the Eo-Alpine structure of the Central Western Carpathians. It is overthrust by the Gemeric Unit (a.k.a. the Upper Group of Nappes) along the Lubeník–Margecany thrust and both override the Tatric sheet (a.k.a. the Lower Group of Nappes) in the north-west along the Čertovica thrust (Fig. 1). This Eo-Alpine nappe pile is tectonically overlain by the Jurassic Meliata subduction-accretionary complex (Kozur & Mock 1973, 1997; Faryad 1995; Faryad & Henjes-Kunst 1997; Lačný et al. 2016) and by the Silicic thin-skinned nappe system (e.g., Mello 1979) and is exposed from beneath the post-nappe Palaeogene and Neogene sedimentary formations and Neogene to Quaternary volcanites or volcano-sedimentary covers (Dublan et al. 1997a,b).

Much progress has been made in recent years towards understanding the processes of Alpine metamorphism during the nappe stacking (e.g., Vrána 1964; Janák et al. 2001; Finger et al. 2003; Lupták et al. 2004; Jeřábek et al. 2008a,b, 2012), but the Late Cretaceous to Cenozoic evolution is still not well understood. Therefore, the principal aim of this work is to apply new fission track data together with sedimentological, stratigraphical, structural, and morphological knowledge for the purpose of revealing quantitative constraints on the Mesozoic to Cenozoic morphotectonic evolution of the external part of the Southern Veporic Unit, immediately after the Eo-Alpine nappe stacking and metamorphism. This study addresses both the Early Cretaceous collisional thrusting

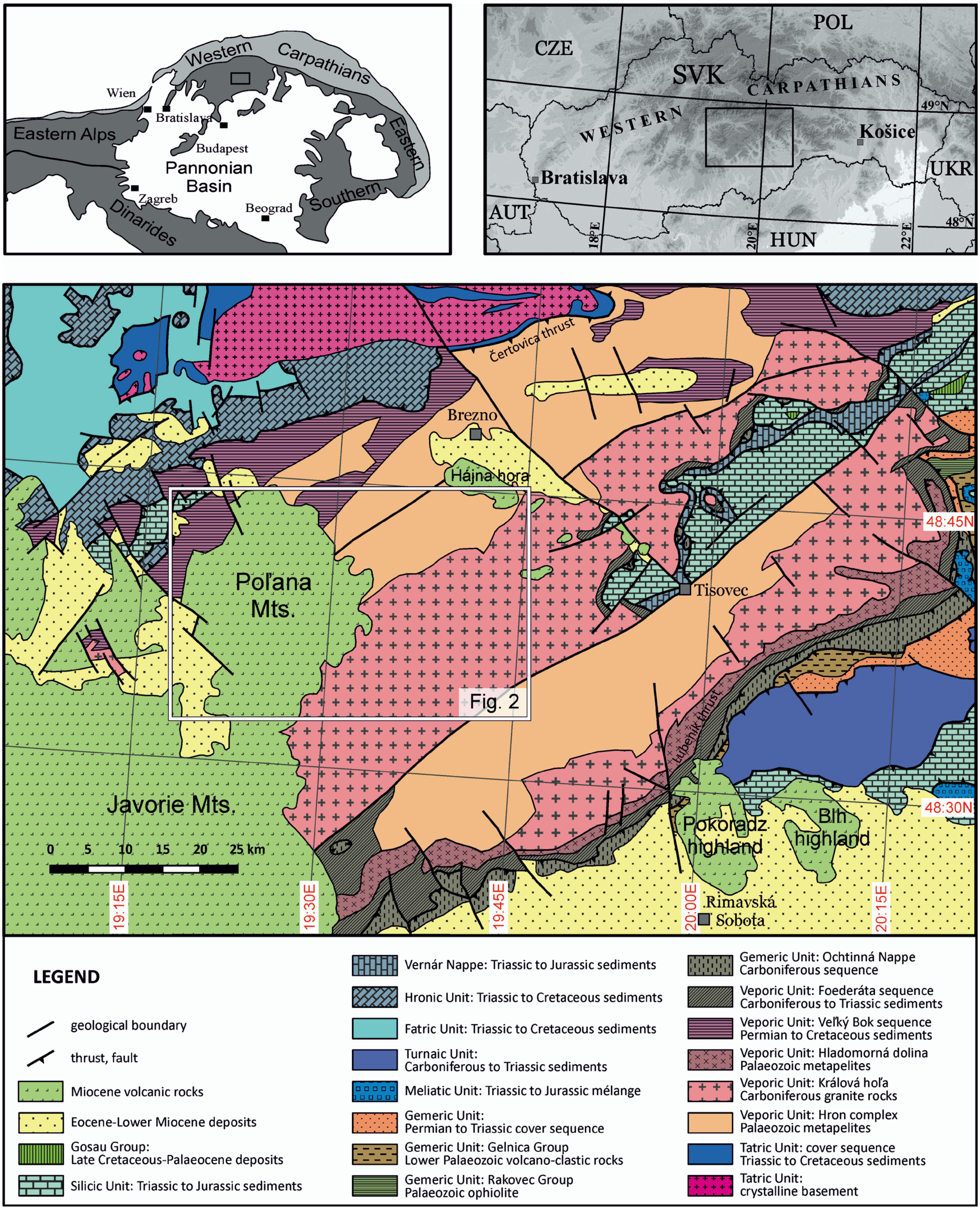


Fig. 1. Tectonic map of the Veporic Unit and surrounding area (according to Bezák et al. 2004; Jeřábek et al. 2012; Vojtko et al. 2016; modified).

phase followed by the Late Cretaceous/Palaeocene collapse and the Eocene–Quaternary post-collisional evolution of this part of the Western Carpathians.

Geological framework

Veporic Unit

The Veporic crystalline basement is composed of Palaeozoic volcano-sedimentary rocks characterised by medium-grade Variscan metamorphic overprint (Vrána 1964; Méres & Hovorka 1991; Kováčik et al. 1996; Putiš et al. 1997; Jeřábek et al. 2008a), which are located in the footwall of high-grade Variscan migmatites and Upper Devonian–Lower Carboniferous I- and S-type granite rocks (~370–350 Ma; Siman et al. 1996; Michalko et al. 1998; Broska et al. 2013). This nappe structure (Figs. 1 and 2), with the footwall metasediments and amphibolites (Hron Complex) and the hanging wall granite rocks (Kráľova Hoľa Complex), has been previously associated with Alpine thrusting (Klinec 1966, 1976; Bezák et al. 1997). During the Permian, the basement was intruded by several smaller A-type granitic bodies (e.g., Hrončok Granite; Bezák et al. 1999a; Finger et al. 2003) and was locally affected by the low-pressure/medium-temperature metamorphism (Finger et al. 2003; Jeřábek et al. 2008b). The Alpine tectono-metamorphic phase is characterised by Cretaceous amphibolite facies conditions in the structural footwall, which gradually decrease to greenschist facies conditions towards the structural hanging wall (Janák et al. 2001; Jeřábek et al. 2008a). The Cretaceous metamorphism was associated with the development of a subhorizontal mylonitic fabric, which developed during E–W orogen-parallel stretching induced by the northward overthrusting of the Gemeric

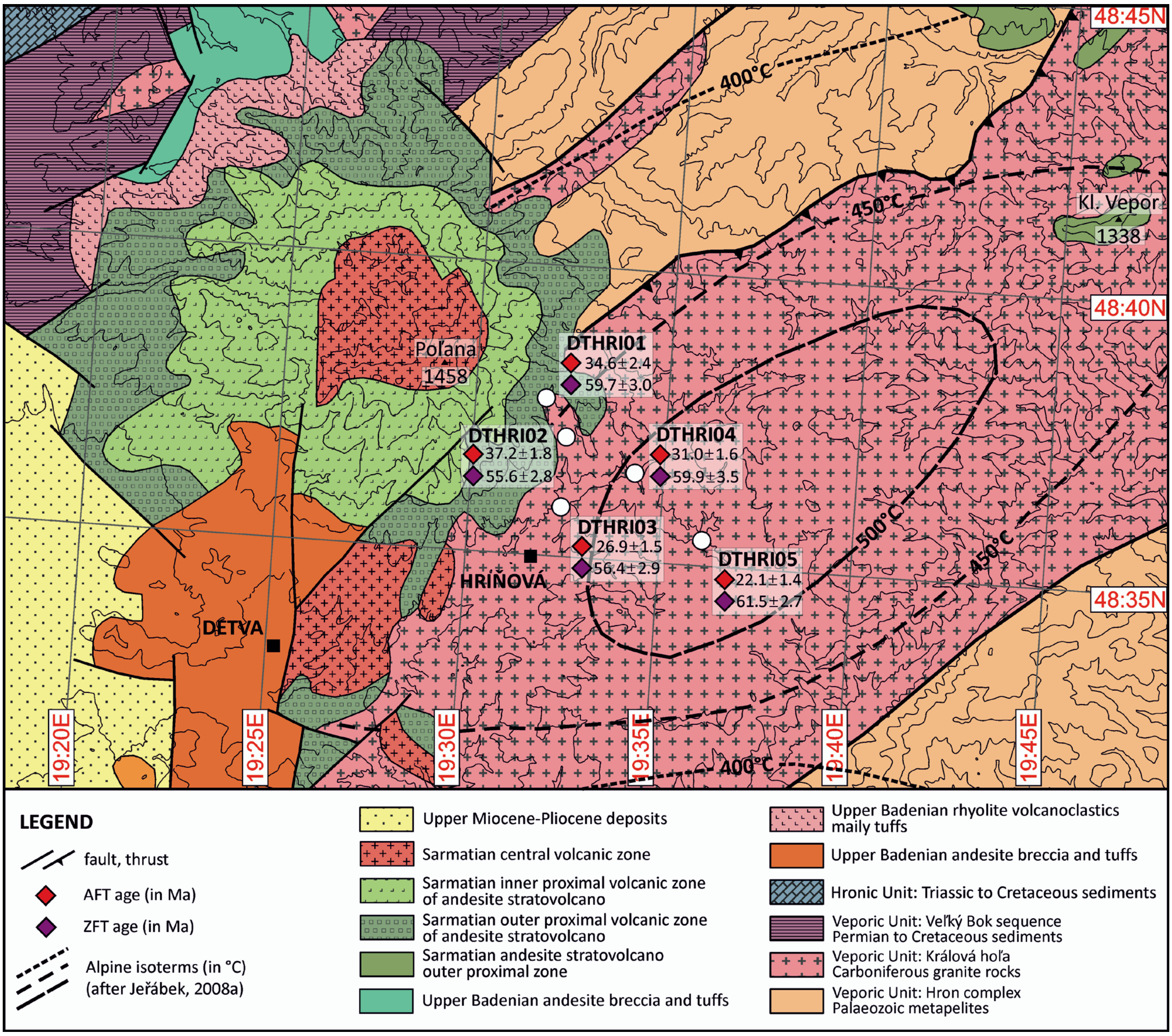


Fig. 2. Tectonic map of the study area showing new zircon and apatite fission track data (ZFT, AFT). Note: map modified according to Bezák et al. (2004).

Unit (Lexa et al. 2003; Jeřábek et al. 2007, 2012; Bukovská et al. 2013). With the ongoing collision, this fabric was folded. The southern boundary between the Veporic and Gemeric units was intruded by the subsurface Rochovce Granite of Cretaceous age (Hraško et al. 1999; Poller et al. 2001; Kohút et al. 2013). The Alpine exhumation in the central part of the Veporic Unit took place during the Late Cretaceous to Palaeogene (Králiková 2013; Vojtko et al. 2016).

The Veporic cover is characterised by the Foederata sequence overlying the Southern Veporic crystalline basement and by the Veľký Bok sequence which is confined to the Northern Veporic crystalline basement. The Foederata cover sequence forms an autochthonous or para-autochthonous sedimentary cover of the Variscan Southern Veporic crystalline basement. Its probable age ranges from the Late Carboniferous to Late Triassic. Jurassic rocks were inferred by several authors (e.g., Klinec 1976) but their presence was not proved yet (e.g., Plašienka 1993; Bezák et al. 1999a,b; Vojtko 2000; Vojtko et al. 2000, 2015). The cover sequence, together with its basement, is metamorphosed under the greenschist facies and intensively ductilely deformed (350–400 °C at 400–450 MPa; cf. Lupták et al. 2003; Jeřabek et al. 2008a). However, the south-eastern portion of the crystalline basement suffered 450–500 °C (Jeřábek et al. 2008a). The study area is located only in a part of the southern Veporic Unit where the cover has already been removed by erosion (Fig. 2).

Post-nappe sedimentary and volcanic formations

To understand the circumstances of the Veporic crystalline basement exhumation, we have to consider several depositional stages in the time span from Late Cretaceous to Cenozoic. The oldest sedimentary sequence of the Late Cretaceous to earliest Palaeocene age is represented by grey calcareous claystones which belong to the Gosau Group (Mišík 1978; Mišík & Sýkora 1980; Gašpariková 1986; Vass et al. 2001). The redeposited Upper Cretaceous fauna frequently occurs also in pre-transgressive to transgressive deposits of the Buda Basin.

After a long period of erosion in the whole Veporic and Gemeric area, a new sedimentary cycle started by deposits of the fore-arc type Central Carpathian Palaeogene Basin in the Priabonian. The sedimentary succession is mainly composed of deep marine, siliciclastic turbidites of the Eocene–Oligocene age. The termination of sedimentation can be indirectly dated to the Oligocene/Early Miocene boundary. The erosive remnants of these deposits occur in the Horehronie Valley — Ľubietová, Brezno, and Tisovec sites (Pulec 1966; Vojtko 2000; Plašienka & Soták 2001; Zlinská et al. 2001; Soták et al. 2005; Žecová et al. 2006; Vojtko et al. 2015; Fig. 1).

Beside this, the southern part of the Veporic area was covered by transgressive deposits of the Upper Rupelian to Lower Chattian Číž Formation (Fm.) (Vass & Elečko 1982) which in the lowermost part contains kaolin clays from weathered crusts located in the north (Kraus 1989). This formation belongs to the retro-arc type Buda Basin (Tari et al. 1993). The overlying Chattian to Aquitanian deposits are composed of basinal calcareous claystones and siltstones belonging to the Lučenec Fm. (Andrusov 1965; Vass & Elečko 1982; Vass et al. 2007). The evolution of the Buda Basin was terminated by the Aquitanian eastward extrusion of the ALCAPA Mega-unit from the Eastern Alpine-Adriatic collisional zone (cf. Csontos et al. 1992; Kováč et al. 2016). In the Late Aquitanian–Early Burdigalian (Eggenburgian), the Fiľakovo-Pétervására Basin developed. However, the spatial extent of deposits was less than the deposits of the Buda Basin (Sztanó 1994; Halásová et al. 1996; Kováč et al. 2016). During the Burdigalian, an activity of the Pannonian asthenolith resulted in uplift and marine regression accompanied by extensive acid volcanism in this area. However, in the deeper part of the depression continental sediments with layers of rhyodacite tuffs (in northern Hungary) were deposited (Vass 1995).

The overlying Salgótarján Fm. is represented by paralic sedimentation with coal seams typical for the Novohrad-Nógrad Basin. However, a gradual northward sea transgression led to subsequent marine sedimentation in southern Slovakia. Subsidence of the basin reached the maximum in the Late Burdigalian (Karpatian), which was immediately followed by rapid regression and erosion. The last transgression in the area of southern Slovakia occurred during the middle Langhian (Early Badenian), at this time the marine and deltaic sediments of the Vinica Fm. were deposited (Vass 1977, 2002).

Volcanic activity prevailed in the study area during the Langhian and Serravallian (Badenian–Sarmatian) when the Javorie, Poľana, and Vepor stratovolcanoes developed. These predominantly andesite stratovolcanoes have a complex, polygenetic structure, and polystage development (cf. Konečný et al. 1983, 1998a,b; Lexa et al. 1993; Dublan et al. 1997a,b; Lexa & Konečný 1998; Vojtko 2000; Konečný et al. 2015a,b).

Finally, river sediments (Poltár Fm.) with high contents of kaoline clays were deposited in the Southern Slovak Basin. These sediments are of the same age of ~6–7 Ma as the basaltic volcanism of the Podrečany Fm. (Balogh et al. 1981; Kantor & Wiegerová 1981; Vass & Kraus 1985; Konečný et al. 1995).

Methods

Fission track analysis

For geochronological study, five Upper Devonian to Lower Carboniferous granite rocks samples were collected from the western part of the Veporic crystalline basement close to the Poľana Stratovolcano. All the samples were taken from surface outcrops (Fig. 2).

Apatite and zircon fission track (AFT, ZFT) analyses were carried out at the Fission Track Laboratory of Isotope Geochemistry section, Vrije Universiteit Amsterdam. After the conventional mineral separation (crushing, sieving, magnetic, and heavy liquid separation), apatites were mounted in epoxy resin, while zircons were placed in PFA® Teflon sheets. Polished apatite mounts were etched in 7 % HNO_3 for 35 s at

20 °C (temperature controlled) in order to reveal spontaneous fission tracks. In the case of polished zircon mounts, the eutectic melt of NaOH-KOH was used at the temperature of 225 °C for 20 hours. The etched mounts were attached against an external detector and subsequently irradiated at the nuclear reactor in Munich. During irradiation, the neutron flux was monitored using CN5 dosimeter glass for the apatite mounts and CN1 dosimeter glass for the zircon mounts. After irradiation, induced fission tracks in external detector muscovites of the mineral mounts were etched in 48 % HF for 12 min at 21 °C; the external muscovites of the dosimeters were etched in 48 % HF for 16 min at 21 °C. Fission tracks were counted with 1250× magnification with a dry objective using a computer-controlled Zeiss Axioplan microscope equipped with an automated Dumitru stage. All the samples were analysed using the external detector method as described by Gleadow (1981). The zeta calibration approach (Hurford & Green 1983) was adopted to determine the central FT ages. Data processing was carried out using the TRACKKEY program, version 4.2.f (Dunkl 2002). The probability of grains counted in a sample belonging to a single population of ages was assessed by a $P(\chi^2)$ probability test (Galbraith 1981). Long axes of the FT etch-pits (D_{par} method; Donelick 1993; Burtner et al. 1994) were measured as a proxy for annealing properties. Track lengths were measured on horizontal confined tracks in c-axis parallel surfaces of apatites and were normalized for crystallographic angle using a c-axis projection (Donelick et al. 1999; Ketcham et al. 2007). Thermal histories of the samples were modelled using the HeFTy® programme (Ketcham 2005) and multi-kinetic annealing model of Ketcham et al. (2007). D_{par} values of apatites were included in the modelling as indicator for the chemical composition of the single grain ages.

Morphotectonic analysis

All the samples for thermochronological study were collected along a nearly horizontal profile with altitudinal difference between sampling sites of up to 200 m. Because the sampling sites are located in the area of relatively well preserved palaeosurfaces, including the largest one, the Sihlianska planina (plateau), altitudinal positions of sampling sites with respect to these palaeosurfaces were analysed. Although the structure and quantity of fission track data do not allow accurate modelling of coupled thermal and geomorphic history (e.g., Safran 2003; Valla et al. 2011), some valuable indications of former palaeorelief were obtained from the morphotectonic analysis.

The remnants of the palaeosurface have been delineated on a 10-m resolution DEM (DMR SR 3 provided by the Topographic Institute in Banská Bystrica) using highly automated DEM-based fuzzy-logic methodology developed by Haider et al (2015). As a first step, four basic raster images, namely slope, curvature, terrain roughness index (TRI), and relative high (RH) raster, were generated from DEM using standard tools integrated in the ArcGIS Info 10.2 (including 3D Analyst and Spatial Analyst extensions). For TRI calculation, the ArcGIS Toolbox for Surface Gradient and Geomorphometric Modeling, version 2.0-0 (Evans et al. 2014) was also used. To obtain more compact results, some degree of smoothing was applied to these basic raster images. Then, the fuzzy membership maps were generated using fuzzy logic criteria similar to those proposed by Haider et al. (2015). Accordingly, the maximum (membership degree of 100 %) and minimum (membership degree of 0 %) thresholds for slope membership raster were set to 10° and 30°. This means that flat surfaces with slopes of up to 10° are considered as hundred percent potential planation surfaces. Applying linear change of membership degree between thresholds and providing that the likelihood is >80 %, flat surfaces tilted more than 14° are not considered to be potential planation surfaces. Alternatively, the maximum and minimum threshold values for TRI membership raster were set to 80 and 100 m. For curvature membership raster we used Gaussian membership type with midpoint “0” and spread “1”. The criteria for construction of the RH surface were modified to match the characteristics of the relief in the study area. The elevation points used for interpolation of the local erosional base level map were acquired as intersections of 3rd and higher orders streams (Strahler ordering) and contour lines (50 m contour interval), excluding the few elevation points clearly on spread planation surfaces. Subsequently, the RH surface was obtained as a difference between the recent topography and this erosional base level surface. To exclude river terraces and young pediments, threshold values for fuzzy membership RH map were set to 100 and 50 m for maximum and minimum membership degrees, respectively. The final map of palaeosurfaces was obtained using fuzzy overlay (“and” type) of all four membership raster. To increase the reliability of the results, the likelihood threshold for the fuzzy overlay raster was set to 90 %. The focal statistics (floating window size 30 by 30 pixels) were used to obtain final raster image.

To take into consideration possible neotectonic differentiation of a previously uniform palaeosurface, the most distinctive morpholineaments were visually identified in the surrounding of outcrops used for fission track analysis. The depth of the sampling site below planation surface was computed by subtraction of its altitude from the maximum altitude of best preserved remnant of the planation surface from the surrounding area bounded by morpholineaments.

Fission track data

The locations and analytical results of the samples are presented in Figs. 2–4. All samples were taken from surface outcrops. The data are displayed in Table 1 as central ages (Galbraith & Laslett 1993) with errors quoted as ±1σ.

ZFT ages were determined for five samples (DTHRI01–DTHRI05 samples), yielding Palaeocene to Early Eocene central ages ranging from 61.5±2.7 to 55.6±2.8 Ma (Table 1). All ZFT ages passed the chi-squared probability test ($P(\chi^2)$ >5 %; Galbraith 1981), indicating that all grains in each sample belong to one homogeneous age population. The radial plots of single grain ZFT ages are shown in Fig. 3.

The AFT ages of five apatite samples (DTHRI01–DTHRI05 samples) yielded Late Eocene to earliest Miocene central ages varying from 37.2±1.8 to 22.1±1.4 Ma (Table 1). All AFT ages passed the chi-squared probability test ($P(\chi^2)$ >5 %; Galbraith 1981). The radial plots of single grain AFT ages are shown in Fig. 4.

In order to quantify fluorine and chlorine contents of the apatite specimens, the D_{par} values were measured as well. The samples displayed the same range of D_{par} values, between 1.6 and 2.5 µm (Table 1), indicating fairly similar chemical compositions and relatively fluorine rich apatites (Burtner et al. 1994) with a low resistance to annealing (Ketcham et al. 1999). Confined fission track length distributions were determined on apatite samples in order to obtain information about their thermal history. The track length distributions of confined horizontal tracks exhibit unimodal and negative skewedness with mainly broad standard deviation (SD >1.5 µm) and a relatively small range in mean track lengths between 12.3 and 13.3 µm (Table 1). Such track length distributions are indicative for basement rocks with slow cooling or prolonged residence in the apatite partial annealing zone (APAZ; ~60–120°C; e.g., Wagner & Van den haute 1992).

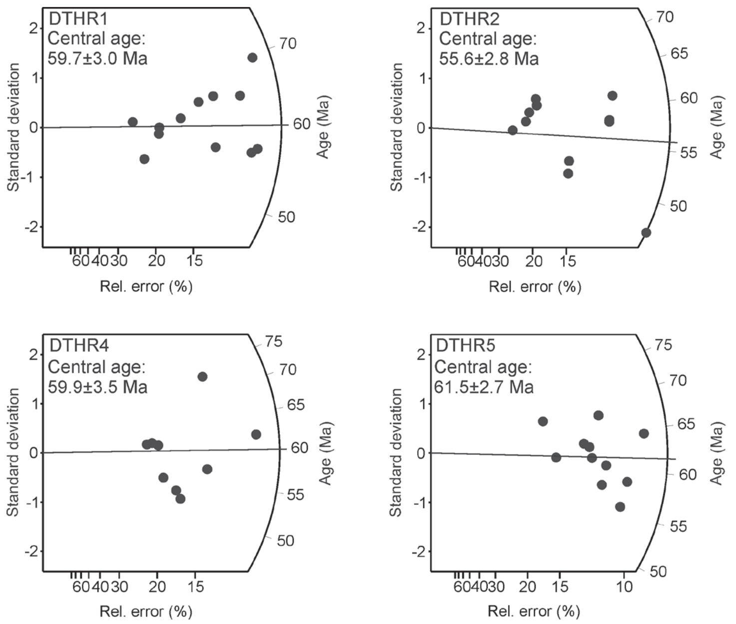


Fig. 3. Radial plots of single-grain zircon fission track (ZFT) age data.

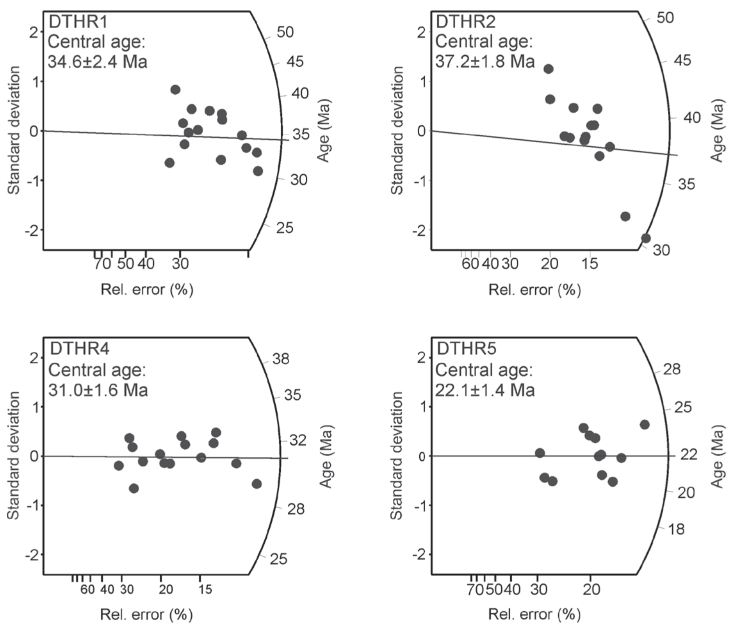


Fig. 4. Radial plots of single-grain apatite fission track (AFT) age data.

Table 1: ZFT and AFT data from the western part of the Veporic Unit, Western Carpathians. Zeta± *1σ* — FT ages were calculated using the zeta calibration method (Hurford & Green 1983) with error quoted as ± *1σ* (Green 1981). Zircon ages were calculated using dosimeter glass CN1 with a zeta value of 128±3 year/cm^2 (analyst: Paul Andriessen), apatite ages using dosimeter glass CN5 with a zeta value of 358±10 year/cm^2 (analyst: Paul Andriessen); N=number of counted grains per sample; ρ_s, (ρ_i)=density of spontaneous (induced) tracks (×10^6 tr/cm^2); Ns, (Ni)=number of counted spontaneous (induced) tracks; ρ_d = density of dosimeter tracks (×10^6 tr/cm^2); Nd=number of counted dosimeter tracks; P(χ^2)=probability of obtaining χ^2 values for n degrees of freedom where n= number of crystals–1; Central age (Ma) ± *1σ* error (Galbraith & Laslett 1993). D_{par}=average diameter of the fission track etch-pits parallel to the crystallographic c-axis (Donelick 1993); MTL=mean confined horizontal track length; SD=standard deviation of track lengths; N(L)=number of horizontal confined tracks measured.

Sample code	Latitude (WGS-84)	Longitude (WGS-84)	Altitude (m asl.)	PaleoAlti/Depth (m)	Petrography	Chronostratigraphy	N	ρ_s	Ns	ρ_i	Ni	ρ_d	Nd	P(χ^2) %	Central age (Ma) ± 1σ	Dpar µm	MTL µm	SD µm	N (L)
Zircon																			
DTHR1	48°37'58.34"N	19°31'42.19"E	780	880/100	tonalite	Late Devonian–Early Carboniferous	17	5.175	1709	23.983	792	4.346	8973	99.5	**59.7±3.0**				
DTHR2	48°37'18.65"N	19°32'02.58"E	680	880/200	tonalite	Late Devonian–Early Carboniferous	11	5.054	1669	25.194	832	4.346	8973	78.2	**55.6±2.8**				
DTHR3	48°35'50.68"N	19°32'17.55"E	580	770/190	tonalite	Late Devonian–Early Carboniferous	11	4.511	1507	22.150	740	4.346	8973	88.8	**56.4±2.9**				
DTHR4	48°36'30.59"N	19°34'09.78"E	590	770/180	tonalite	Late Devonian–Early Carboniferous	9	4.928	1126	22.759	520	4.346	8973	80.9	**59.9±3.5**				
DTHR5	48°35'25.61"N	19°36'00.37"E	710	820/110	granodiorite	Late Devonian–Early Carboniferous	11	7.399	2685	33.289	1208	4.346	8973	97.7	**61.5±2.7**				
Apatite																			
DTHR1	48°37'58.34"N	19°31'42.19"E	780	880/100	tonalite	Late Devonian–Early Carboniferous	15	2.563	309	13.344	1609	10.099	20850	99.9	**34.6±2.4**	1.64–2.34	12.95	1.39	32
DTHR2	48°37'18.65"N	19°32'02.58"E	680	880/200	tonalite	Late Devonian–Early Carboniferous	14	5.813	759	28.156	3676	10.099	20850	72.8	**37.2±1.8**	1.91–2.34	12.30	1.71	74
DTHR3	48°35'50.68"N	19°32'17.55"E	580	770/190	tonalite	Late Devonian–Early Carboniferous	15	4.053	554	27.146	3711	10.099	20850	99.9	**26.9±1.5**	1.79–2.34	13.08	1.48	102
DTHR4	48°36'30.59"N	19°34'09.78"E	590	770/180	tonalite	Late Devonian–Early Carboniferous	15	4.511	615	26.225	3575	10.099	20850	100.0	**31.0±1.6**	2.16–2.53	12.67	1.59	109
DTHR5	48°35'25.61"N	19°36'00.37"E	710	820/110	granodiorite	Late Devonian–Early Carboniferous	12	3.952	349	32.212	2845	10.099	20850	99.9	**22.1±1.4**	1.85–2.24	13.31	1.66	53

Results of morphotectonic analysis

The arrangement of potential planation surfaces (formation of the near-sea-level, low-relief erosional surface) is depicted in Fig. 5. The largest remnant westward of Hriňová town is the Sihlianska planina (plateau) that is generally considered as the Miocene planation surface (Lukniš 1964), formed before the origin of the Poľana Stratovolcano (Urbánek 2002). Clear altitudinal differences are visible along morpholineaments (palaeosurfaces reach up 900 m a.s.l. on the south, 1100 m a.s.l. on the west and 1000 m a.s.l. on the north) pointing to neotectonic differentiation of the palaeosurface (Lacika 1993). It is supported by directional coincidence of morpholineaments and main faults (cf. Fig. 2 and 4). The Slatina Valley is a local centre of relative subsidence, where the palaeosurface reaches only 770 m a.s.l. near DTHRI03 and DTHRI04.

There is no ZFT age-altitude relationships in the study area, but certain relation can be seen between the age and depth below the preserved planation surface (Table 1). It is known, that relatively deep ZFT closure isotherm is generally less sensitive to palaeotopography (Braun 2002; Glotzbach et al. 2015). Thus, the ZFT data usually do not provide valuable information about the character of contemporary palaeorelief. On the other hand, we can suppose a former altitudinal unity of till now preserved palaeosurface and its neotectonic differentiation. Considering nearly flat ZFT isotherm, a denudation rate less than 0.02 mm/year can be calculated from the relation between ZFT age and depth below the palaeosurface. This value is much lower than the modelled cooling rates (Figs. 6 and 7). Therefore, small differences in altitude of sampling sites or depth below the palaeosurface indicate that all samples are from the same ZPAZ zone.

The AFT ages are considerably younger than those obtained in other Veporic areas (cf. Vojtko et al. 2016). Moreover, the disturbed character of ages is obvious, but no significant relationships were detected between age and altitude of sampling site or depth below the planation surface. On the other hand, the elevation differences between sampling sites are very small, and if neglected (i.e. sampling profile will be considered as horizontal), such age perturbation can indicate pronounced topography during AFT system closure (Braun 2002; Glotzbach et al. 2015). In this case, the younging trend from the NW to SE indicates a palaeoslope inclined from SE to the NW.

Alpine tectonic evolution

Eo-Alpine Early Cretaceous nappe stacking

The Alpine shortening and burial history of the Veporic Unit began in the Early Cretaceous following overthrusting of the Jurassic Meliata subduction-accretionary complex onto the Gemeric Unit (Kozur & Mock 1973; Maluski et al. 1993; Dallmeyer et al. 1996; Faryad & Henjes-Kunst 1997; Árkai et

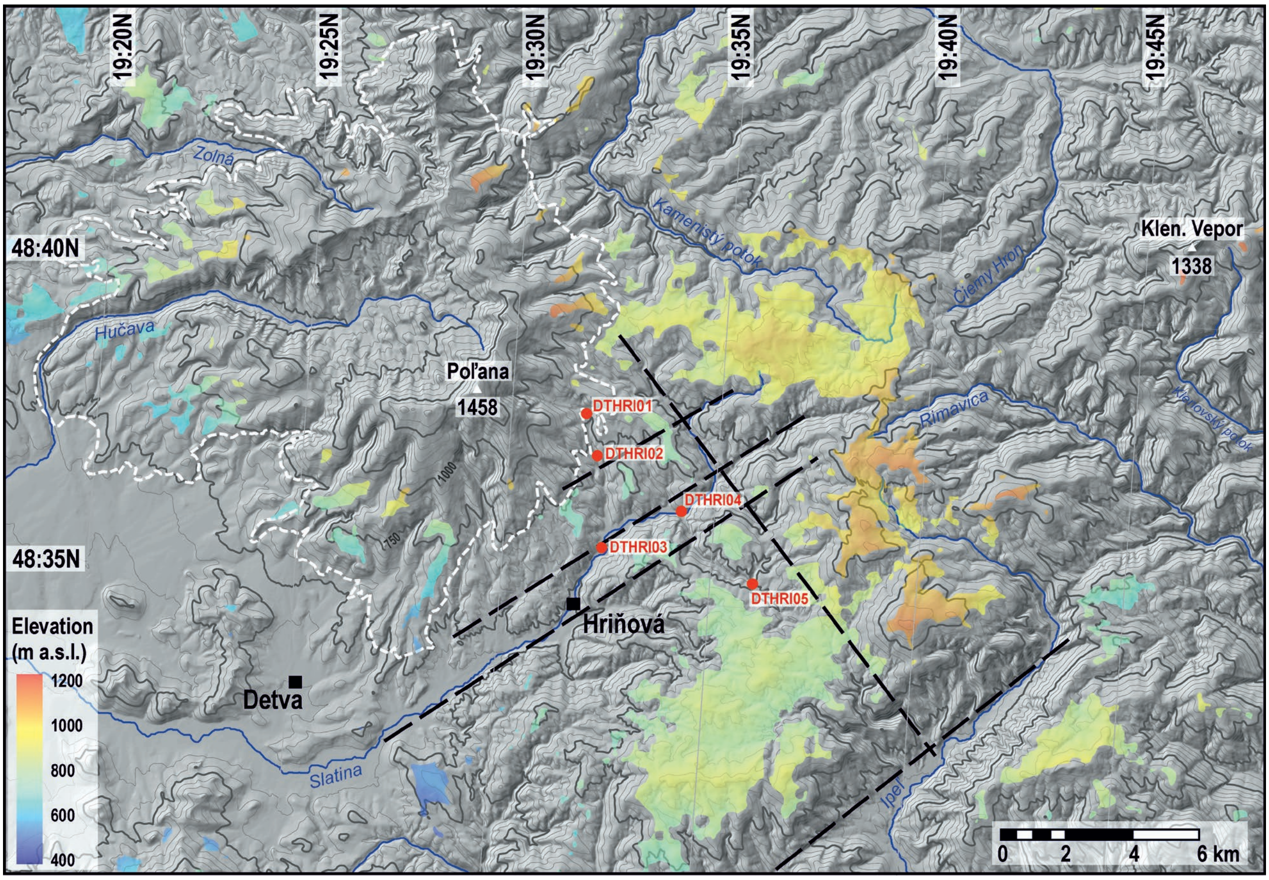


Fig. 5. Present-day topography of the study area with remnants (coloured) of oldest planation surfaces. The colours of surfaces represent their altitude. Black dashed lines represent the main morpholineaments related to the FT sampling sites. White dashed line delineates outer proximal zone of the Poľana Stratovolcano. Contour intervals — 250 m (heavy contours), 50 m (fine contours).

al. 2003; Putiš et al. 2014, 2015; Lačný et al. 2016). In these early convergent stages, the Veporic Unit suffered an internal shortening and thickening documented by upright folding. These early structures were mostly obliterated by the subsequent major deformation associated with the development of subhorizontal mylonitic foliation and the W–E orogen-parallel extension (Hók et al. 1993; Plašienka 1993; Janák et al. 2001; Jeřábek et al. 2007, 2008a). In the study area, the Alpine metamorphism shows the greenschist facies conditions of 450 °C at 400–450 MPa (Jeřábek et al. 2008a). The Eo-Alpine metamorphism may be as old as ~115 Ma, as was revealed by $^{40}Ar/^{39}Ar$ and K–Ar ages (Maluski et al. 1993; Kováčik et al. 1996, 1997) and Sm–Nd whole rock-garnet isochron (~109 Ma; Lupták et al. 2004). The orogen-parallel extension of the Veporic Unit finished at latest by ~97 Ma, as is suggested by the post-kinematic growth of monazite in the southern Foederata cover sequences, revealed by the laser ablation ICP-MS dating (Bukovská et al. 2013). At the same time, the northern part of the Veporic Unit still experienced thrusting and internal imbrications, related to the onset of underthrusting of the Fatric basement from the north, recorded by white mica $^{40}Ar/^{39}Ar$ ages of ~95–90 Ma from the lower-grade shear zones in the northern parts of the Veporic dome (Plašienka 2003; Putiš et al. 2009).

Cretaceous to Neogene exhumation/denudation

The continuing N–S convergence and initiation of underthrusting of the Tatric–Fatric basement southward switched the Gemeric-driven subvertical shortening in the Veporic Unit to the Tatric–Fatric-driven horizontal N–S shortening (Jeřábek et al. 1012). This process caused upright folding of the earlier subhorizontal fabric and the development of crustal-scale folds (Jeřábek et al. 2008a, 2012; Vojtko et al. 2016). This indicates that the major exhumation displacement and cooling from ~400 to 350 °C occurred before ~80 Ma (Figs. 6–8), most probably in association with the formation of large-scale cuspate antiforms (Vojtko et al. 2016). The upper part of the basement, most likely due to southward underthrusting of the Tatric–Fatric basement, were affected by an eastward unroofing of the overlying rock sequences (Fig. 9). Beside this, the schellite-molybdenite stockwork mineralization was emplaced as fine disseminations and veinlets. Genetically, the mineralization is confined to pre-existing subvertical E–W trending cleavage and is related to intrusion of the subsurface Upper Cretaceous Rochovce granite occurring in the close proximity to the Lubeník thrust zone. It is dated by zircon U–Pb isochrones revealing ages from ~76 to 82 Ma (Hraško et al. 1999; Poller et al. 2001; Kohút et al. 2013) and post-dates the

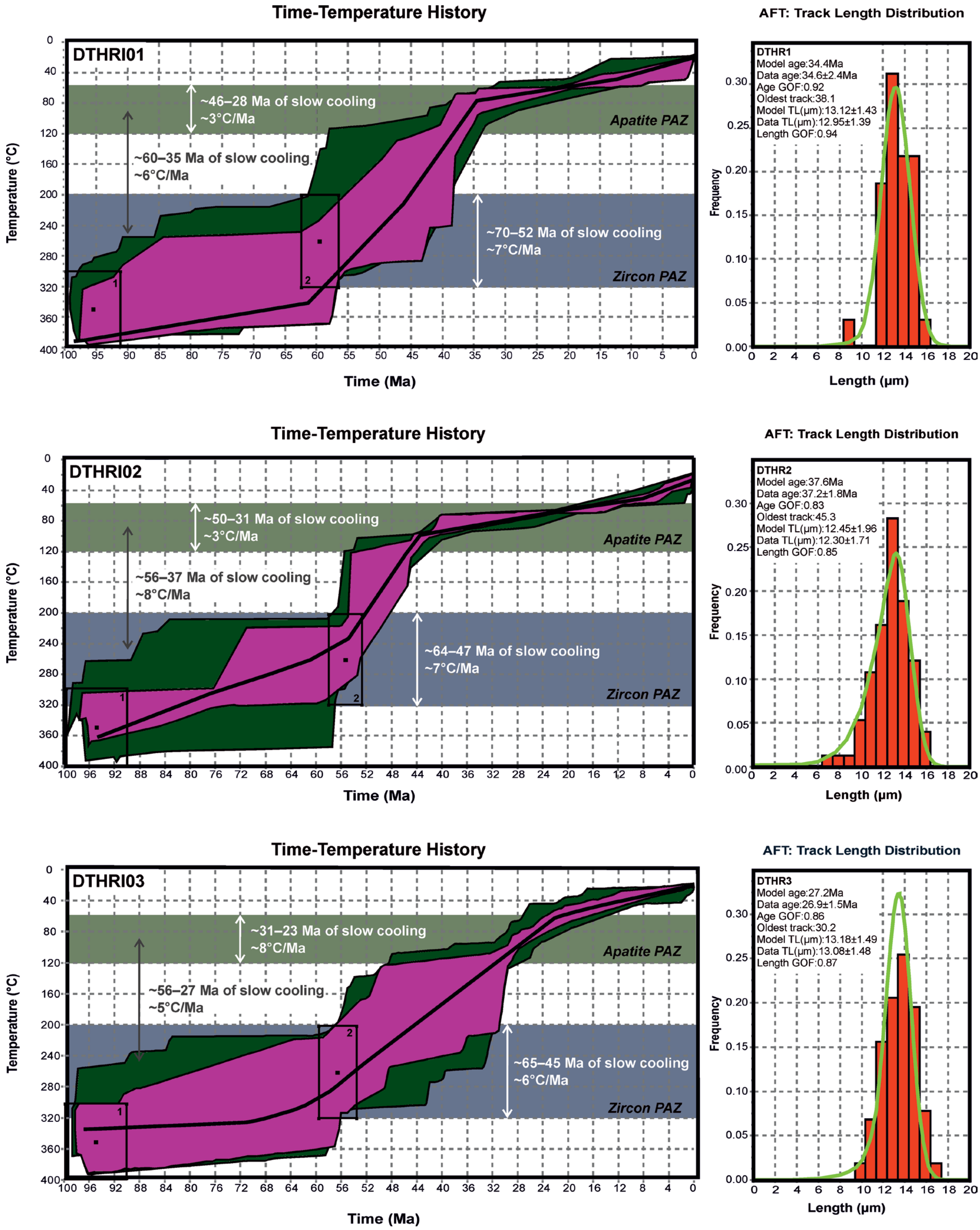


Fig. 6. Thermal modelling of AFT data: (from top to bottom: DTHR1, DTHR2, and DTHR3 samples) obtained by HeFTy program (Ketcham 2005). Results are displayed in time-temperature diagrams (left diagrams). Magenta envelope — good fit; green envelope — acceptable fit; black line — best fit; black box — fixed constraints defined according to independent geological and geochronological data (1 — burial beneath the Eo-Alpine nappe stack (e.g., Vojtko et al. 2016); 2 — ZFT age). Right diagrams: frequency distribution of measured confined track length data overlain by a calculated probability density function (best fit). Model age, Data age — model and data calculated age. Age GOF, Length GOF — goodness of fit (statistical comparison of the measured input data and modelled output data, where a "good" result corresponds to value of 0.5 or higher, an "acceptable" result corresponds to a value of 0.05, and "the best" result corresponds to a value of 1). Note that modelled *t-T* paths are valid only inside 120–60 °C (Apatite PAZ — partial annealing zone). Data outside this temperature range may not necessarily represent the real thermal trajectory of a sample, unless constrained by other data. Oldest track: the age of the oldest fission track that has not fully annealed. Model TL, Data TL — mean lengths of the model and data, and the standard deviations of length distributions.

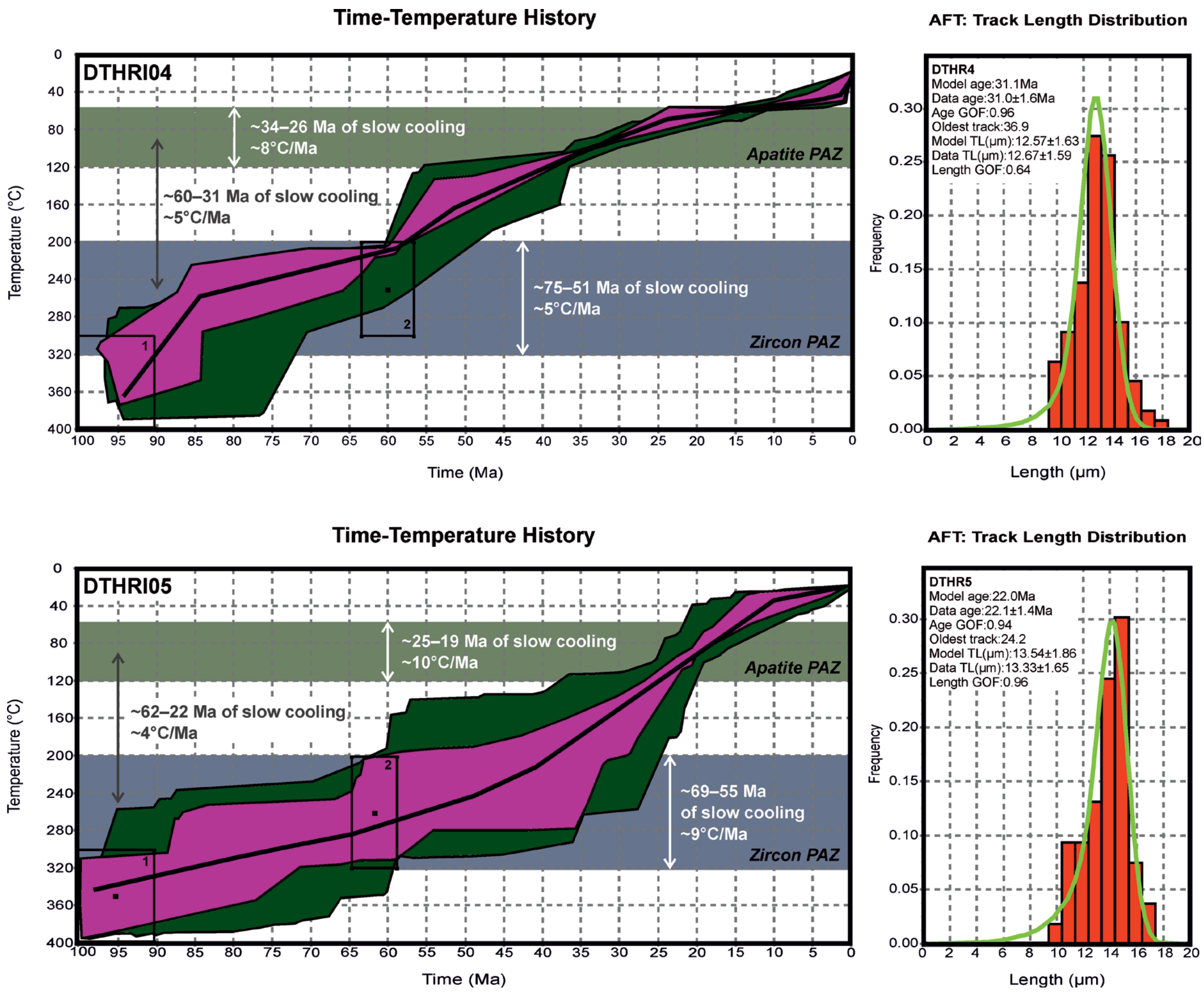


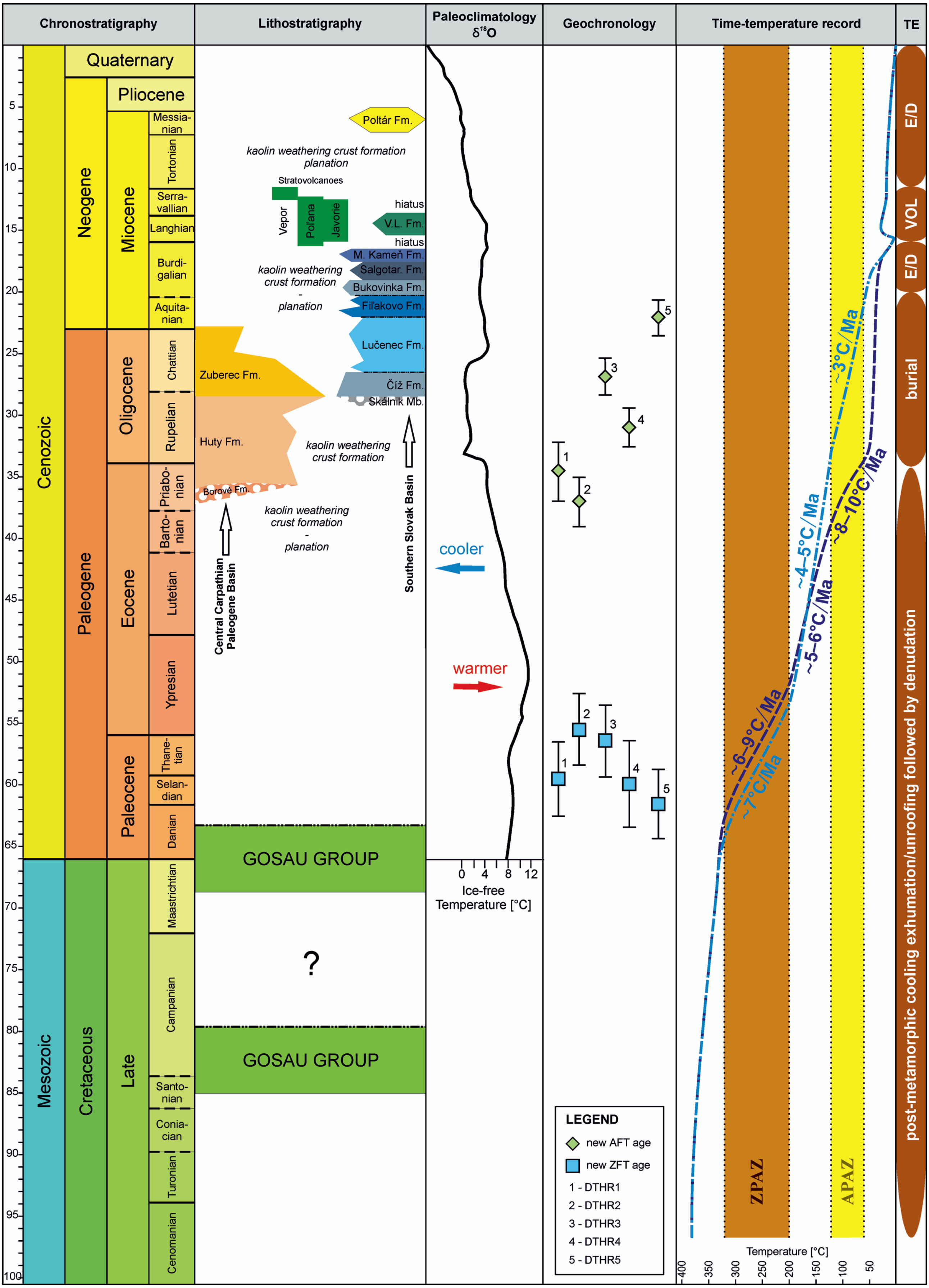
Fig. 7. Thermal modelling of AFT data: (from top to bottom: DTHR4 and DTHR5 samples) obtained by HeFTy program (Ketcham 2005). Results are displayed in time-temperature diagrams (left diagrams). For further explanation see Fig. 6.

unroofing of the Veporic Unit from underneath the overlying Ochtiná Nappe and Gemeric Unit along a low-angle detachment shear zones (Hók et al. 1993; Plašienka 1993; Madarás et al. 1996; Jeřábek et al. 2012; Bukovská et al. 2013; Novotná et al. 2015). Most probably, data obtained from newly formed phlogopite from the Muráň Nappe sole that yielded flat Ar–Ar spectra with plateau ages at 84 to 91 Ma (Milovský & Plašienka, 2007) could indicate unroofing of the Veporic Unit. The Drienok–Vernár–Muráň cover nappe system was detached and transported south-eastward together with the Gemeric Unit and the Meliata accretionary prism.

Since the Late Cretaceous to Palaeocene, a passive en-block exhumation of the already finalised internal structure of the Southern Veporic Unit was probably controlled by isostatic balancing of thickened crust and progressive erosion. In the study area, the cooling of the crystalline basement through ~320–200 °C (zircon partial annealing zone, ZPAZ; Tagami et al. 1998) with a slow cooling rate of ~6–9 °C/Ma (Figs. 6–8) was revealed by new ZFT data of 61.5±2.7 to 55.6±2.8 Ma (Figs. 3, 6–8). These Palaeocene ages can be explained by slower or delayed exhumation of the western portion of the Veporic metamorphic dome with respect to its central part (cf. Kráľ 1977; Plašienka et al. 2007; Vojtko et al. 2016).

Additionally, new AFT data of 37.2±1.8 to 22.1±1.4 Ma (Fig. 4) indicate that continuous slow cooling (~3–10 °C/Ma) progressed from ZPAZ to APAZ (temperature interval of ~245 °C to 90 °C; Figs. 6–8) during the Eocene. Such slow cooling likely reflects erosion-controlled exhumation of the

Fig. 8. Summary of litostratigraphy, palaeoclimatology, geochronology, and time-temperature record, indicating Mesozoic to Cenozoic geodynamic evolution of the Vepor domain. Explanations: *Lithostratigraphy* — **V.L. Fm.** – Vinica and Lysec formations; *Palaeoclimatology* — oxygen isotope curve ($\delta^{18}O$) for Cenozoic (modified according to Zachos et al. 2001); *Time-temperature record* — **ZPAZ, APAZ** – zircon and apatite partial annealing zones, dashed line represents idealized fit for the low-thermal evolution of the DTHRI01 and DTHRI02 samples and dot-dashed line represents idealized fit for the low-thermal evolution of the DTHRI03, DTHRI04, and DTHRI05 samples, computed values represent cooling rates in mm a year; **TE** – tectonic events, **E/D** – exhumation vs. denudation, **V** – volcanic activity.


Chronostratigraphy
Lithostratigraphy
Paleoclimatology δ18O
Geochronology
Time-temperature record
TE
Quaternary
Pliocene
Messinian
Tortonian
Serravallian
Langhian
Burdigalian
Aquitanian
Chattian
Rupelian
Priabonian
Bartonian
Lutetian
Ypresian
Thanetian
Selandian
Danian
Maastrichtian
Campanian
Santonian
Coniacian
Turonian
Cenomanian
Neogene
Miocene
Paleogene
Oligocene
Eocene
Paleocene
Cenozoic
Mesozoic
Cretaceous
Late
Poltár Fm.
kaolin weathering crust formation
planation
Stratovolcanoes
Vepor
Poľana
Javorie
hiatus
V.L. Fm.
hiatus
M. Kameň Fm.
Salgotar. Fm.
Bukovinka Fm.
Fiľakovo Fm.
kaolin weathering crust formation
planation
Lučenec Fm.
Zuberec Fm.
Číž Fm.
Skálnik Mb.
Huty Fm.
kaolin weathering crust formation
Borové Fm.
kaolin weathering crust formation
planation
Central Carpathian Paleogene Basin
Southern Slovak Basin
GOSAU GROUP
?
GOSAU GROUP
cooler
warmer
Ice-free Temperature [°C]
LEGEND
new AFT age
new ZFT age
1 - DTHR1
2 - DTHR2
3 - DTHR3
4 - DTHR4
5 - DTHR5
ZPAZ
APAZ
Temperature [°C]
~3°C/Ma
~8–10°C/Ma
~4–5°C/Ma
~5–6°C/Ma
~6–9°C/Ma
~7°C/Ma
E/D
VOL
E/D
burial
post-metamorphic cooling exhumation/unroofing followed by denudation

southern Veporic crystalline basement from beneath the overlying complexes formed by the thin pile of the Drienok–Vernár–Muráň cover nappe system and, probably, by sediments of the Gosau Group (Fig. 9). However, we also assume the "secondary burial" of the Vepor Unit beneath these complexes. Consequently, the emplacement of the thin Drienok–Vernár–Muráň cover nappes into its final position should have occurred after the final stages of the Veporic Unit exhumation (after 55 Ma), but definitively before the transgression of Central Carpathian Palaeogene Basin (before 35 Ma) or the Southern Slovak Basin sequences. This assumption is supported by an important difference between the Alpine temperatures determined for the uppermost level of the Veporic Unit (~350 °C, based on metamorphic mineral assemblage and ZFT data; Lupták et al. 2003; Jeřábek et al. 2008a; Vojtko et al. 2016) and for the lowermost level of the Drienok–Vernár–Muráň nappe system (~150 °C based on conodont colour alteration index; Havrila 2011), excluding their common metamorphic evolution. Moreover, occurrences of the cover nappes with the Gosau Group in the study area during the Eocene–Oligocene is also proved by variegated Oligocene to Early Miocene transgressive conglomerates of the Southern Slovak Basin, which are composed of pebbles from several Eo-Alpine units (e.g., Vass & Elečko 1982; Vass et al. 1989, 2007).

Additionally, the perturbed AFT ages (~37–22 Ma) point to the existence of pronounced palaeorelief during the AFT system closure. It is not definitely clear, however, whether this palaeorelief was related to the huge erosional remnants of the Silicic nappe pile in the study area, or to the Late Eocene to Oligocene extension-related deepening of Palaeogene basins (Soták et al. 2001; Kováč et al. 2016). However, the AFT ages point rather to the second option.

Modelling of the AFT parameters has provided a fairly clear picture about the low-temperature thermal evolution of the southern Veporic crystalline basement since the Late Eocene (Figs. 6 and 7). Based on the strongly reduced mean track lengths and scarcity of long tracks, the *t-T* paths exhibited two cooling groups with respect to their cooling rate. The first group (DTHRI01 and DTHRI02 samples) represents slow cooling or prolonged residence in the APAZ with cooling rate of ~3 °C/Ma. On contrary, the second group (DTHRI03, DTHRI04, and DTHRI05 samples) is characterised by slightly faster cooling in the APAZ with cooling rate of ~8–10 °C/Ma.

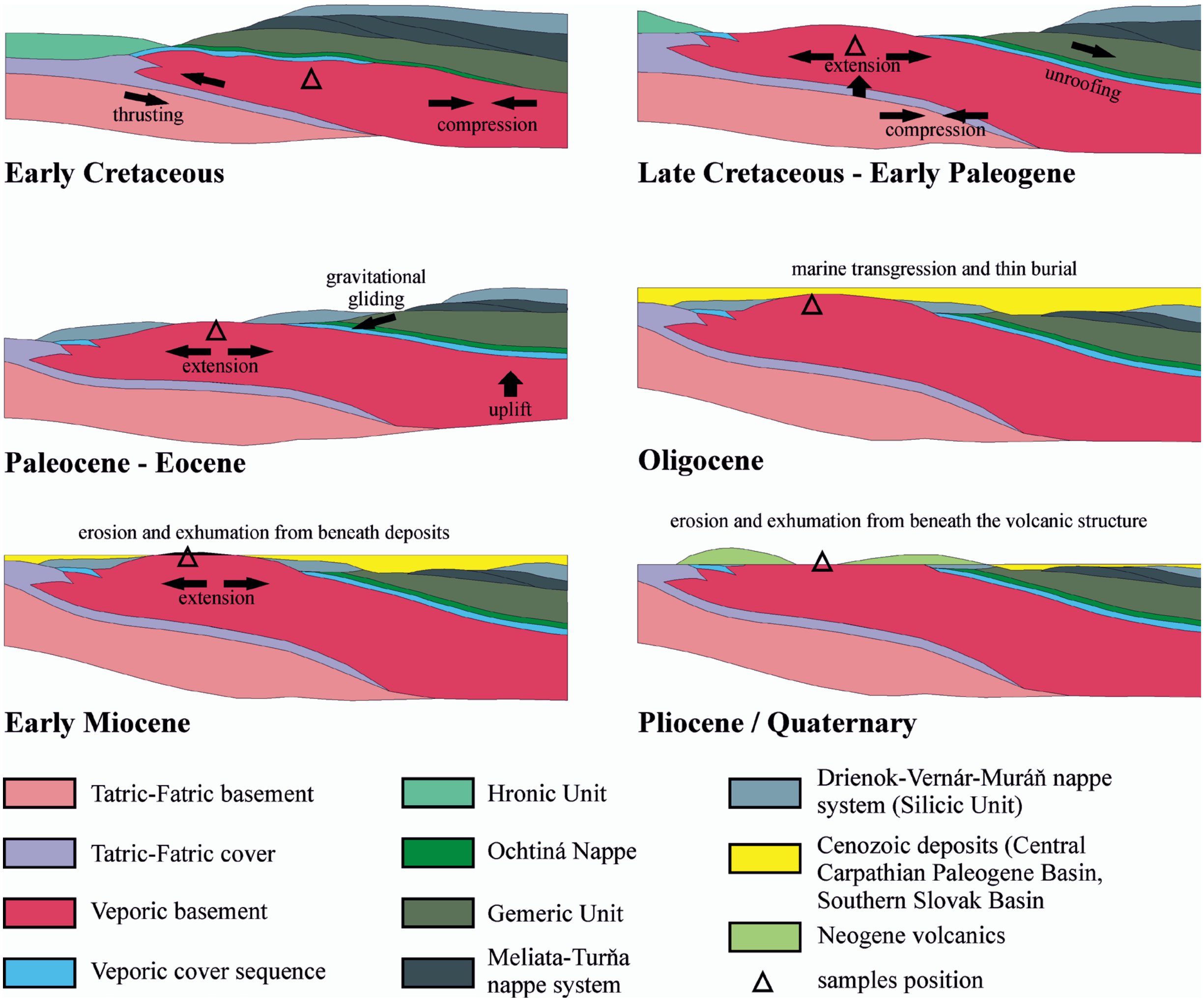


Fig. 9. Mesozoic to Cenozoic geodynamic evolution of the Vepor Mountains and their surroundings.

However, both groups indicate the Late Eocene to Oligocene continual cooling up to present-day temperatures (Figs. 6–8).

The final stage of the Late Eocene erosion of the main Vepor area can be related to the planation. This planation episode is generally related to in situ chemical weathering and formation of kaolin weathering crust in the Vepor area conditioned by high global annual temperature and humidity (Fig. 8). There is no direct evidence about this planation stage in the study area, but we suppose that the basal formation (Skálnik Member of the Číž Formation; Early/Late Oligocene boundary) of the Southern Slovak Basin (e.g., Vass et al. 1989, 2007) likely contains kaolin transported from the study area. Most probably, abrasion during slow Eocene transgression played an important role in the final stage of formation of the first Cenozoic planation surface. Nevertheless, in many places the Oligocene to Lower Miocene sea transgressed onto the Veporic and Gemeric crystalline basements again (Fig. 9; e.g., Soták et al. 2005; Vass et al. 2007; Kováč et al. 2016; Vojtko et al. 2016).

A maximum burial of the Veporic Unit beneath this post-nappe sedimentary sequence is considered to be in the Late Oligocene and coincides with deposition of the Huty Fm. (Soták et al. 2001) in the external part of the Central Western Carpathians and also with deposition of the Číž Fm. in the Southern Slovak Basin (Vass & Elečko 1982; Vass et al. 1989, 2007). In the Veporic Unit, the evidence for the Oligocene burial is found in the Brezno Depression where at least 800 m of the Upper Eocene to Oligocene strata is still preserved (Planderová 1966; Pulec 1966; Sitár 1966). Small erosive remnants have also been preserved in the vicinity of Tisovec town (cf. Klinec 1976; Vojtko 2000, 2003; Plašienka & Soták 2001; Soták et al. 2005). Nevertheless, the Oligocene deposits could not be thick enough because there are no indications of AFT system reheating (Figs. 8 and 9; based on Vojtko et al. 2016 and data therein).

After the deposition of the Oligocene to Lower Miocene sedimentary sequence, the prolonged erosional exhumation of the Vepor area can be assumed. The early Miocene erosion in a humid and warm climate nearly completely removed older Cenozoic sediments and remnants of superficial nappes. Thus, the Middle Miocene volcanism likely occurred here in a relatively flat landscape forming the second Cenozoic palaeosurface (Figs. 8 and 9) favourable for supergene kaolinization. Remnants of kaolin weathering crust indicate that the Veporic granitoid basement must have experienced a period of tectonic quiescence and was exposed to intensive chemical weathering (Kraus 1989). Consequently, the so-called mid-mountain level is probably the third Cenozoic palaeosurface, remnants of which are widespread in the modern relief of the Western Carpathians. Traditionally it is considered to be the Upper Miocene surface, but it can integrate also remnants of older palaeosurfaces (Lukniš 1964; Minár 2003) as in the eastern part of the Veporic massif where inheritance of Eocene planation surfaces (cf. Vojtko et al. 2016) and its rejuvenation during the early Miocene was documented. In contrast, the scattered AFT ages from ~37 Ma to 22 Ma from almost the same altitude level beneath the mid-mountain level indicate enhanced topography during AFT system closure and do not support preservation of the Eocene planation surface in this locality. However, uncovering and integration of the early Miocene planation surface into mid-mountain level cannot be excluded.

Formation and erosion of the Neogene stratovolcanoes

Tectonic quiescence period in the Early Miocene was replaced by lithospheric stretching, intramontane basins formation, and volcanism in the Central and Internal Western Carpathians. In the study area, the volcanic activity started in the south-western part by Langhian stratovolcanic suite (Konečný et al. 1998a,b) belonging to the Javorie Stratovolcano and Langhian to Serravallian Šútovka Statovolcano (Šimon et al. 2013). Later on, after the period of volcanic and tectonic quiescence (lasted about 1 million years), the volcanic activity was renewed and progressed towards the central part of the Veporic domain (Figs. 8 and 9). The Poľana Stratovolcano was formed during the Serravallian and its remnants are the best preserved in the recent relief. Most probably, the youngest stratovolcano was represented by the Vepor Stratovolcano (late Serravallian; Konečný et al. 2015a,b), but on the contrary it is nearly totally missing in the recent relief. The products of the Serravallian Poľana and Vepor stratovolcanoes probably completely covered and conserved the older planation surface. The total thickness does not exceed more than 1.5 km, because the AFT system was not reheated during the Neogene in the study area.

After the volcanic activity ceased, erosional processes removed almost the whole volcanic cone of the Vepor Stratovolcano, significantly destroyed the Serravallian Javorie stratovolcanic cone and slightly disrupted also the Poľana Stratovolcano from the Late Miocene. Isostatically counterbalanced uplift and erosion of the Veporic domain most probably quickly uncovered the Lower Miocene planation surface on the Veporic crystalline basement from beneath the volcanic structure (Figs. 8 and 9). After the intensive mechanical weathering, the period of tectonic quiescence dominated by chemical weathering occurred and caused the third phase of planation, as well as formation of small kaolin crusts that developed not only on the Veporic crystalline basement, but also on the volcanic and carbonate rocks (Fig. 8; e.g., Lukniš 1964; Kraus 1989; Gaál 2008). The existence of this phase of planation is supported by formation of planation surfaces on the Serravallian volcaniclastics rocks in the periphery of the Veporic area (e.g., Hájna Hora, Pokoradz, and Blh highlands; Fig. 1), as well as extensive remnants of planation surfaces inside the older neovolcanic mountains in the west (Kremnické vrchy, Ostrôžky). Superposition of the youngest Serravallian lava flows on truncated older volcanosedimentary formations points to an integration of older surfaces into the Late Miocene mid-mountain level in altitude of ~1000 m a.s.l. Exhumation of older planation surfaces and integration of extensive depositional volcaniclastic plains enabled formation of a stepped flat relief that occupied majority of the region, excluding the Poľana Stratovolcano. This type of planation surface was called a tectoplain due to a complex long-term development in prevailing extensional tectonic regime (Minár 2003; Minár et al. 2011). During the Pontian to Pliocene (~6–3 Ma), the tectonic quiescence period was replaced by exhumation which caused that the kaolin weathering crust to be washed away and deposited in the Poltár Fm. of the Lučenec Basin (Fig. 8). Transportation and deposition of this weathered crust from the crystalline basement was controlled by depositional environments. Basically, the quartzitic sand abundant in kaolin deposits derived from granitoids was transported to the alluvial–lacustrine environment where it was accumulated on northern slopes of the basin. Clays were ultimately eroded and washed out and deposited in the basin where they form a matrix of gravels and sands (Poltár Fm.).

The present-day morphology of the Vepor Mountains, characterised by sharply cut valleys to the crystalline basement (Upper Ipeľ river, Rimavica river or Kamenistý potok creek), points to an accelerated Pliocene and Quaternary uplift (Figs. 8 and 9), which was probably controlled by erosion-induced isostatic adjustment of the area after the removal of a considerable amount of the volcanic complexes from this region (especially the Vepor Stratovolcano).

Conclusions

In order to provide an insight into the morphotectonic evolution of the external zone of the southern Veporic Unit, an internal part of the Eo-Alpine (Cretaceous) orogenic wedge of the Central Western Carpathians, a combination of geochronology together with regional geological, sedimentological, and geomorphological investigations was used. Based on this research, several principal Alpine tectono-thermal stages of burial and exhumation processes can be defined.

During the Eo-Alpine Early Cretaceous nappe stacking, the Veporic crystalline basement was buried beneath the northward overthrusting Gemeric Unit and overlying Jurassic Meliata accretionary complexes. The crystalline basement was buried at least to the depth of ~15 km and suffered a greenschist facies metamorphic overprint. In this early convergent stage, the Veporic Unit underwent an internal thickening and shortening which led to the formation of a penetrative subhorizontal mylonitic fabric.

After the Early Cretaceous burial, a major exhumation phase started and most likely it was associated with two distinct cooling mechanisms related to underthrusting of the northerly Tatric–Fatric crust. The first exhumation and cooling from ~400 to 350 ºC, as a result of initial underthrusting and Veporic unroofing, took place before ~80 Ma.

Since the Late Cretaceous, a continual underthrusting led to a passive en-block exhumation of the already finalised internal structure of the Veporic Unit. In the western portion of the Southern Veporic Unit, the Palaeocene slow cooling through the temperature interval of ~320–200 °C was revealed, which is approx. 10 Ma later than in the central part of this unit. During the Early Eocene, a deceleration of exhumation rate in temperature conditions from ~250 to 90 °C (temperature interval between APAZ and ZPAZ medians) was computed, which most likely reflects burial of the Southern Veporic crystalline basement beneath the thin Silicic superficial nappe system and the Gosau Group strata.

The slow cooling continued up to the latest Eocene to Oligocene when the basement rocks reached the temperature zone of ~120–60 °C. This process can be related to subsequent erosion of the overlying strata. In many places the Eocene erosion was efficient enough, because the Oligocene to Lower Miocene strata were deposited directly onto the Veporic basement. At this time, the first Cenozoic planation surface with kaolin weathering crust was probably formed in the Vepor area. However, disturbed AFT ages indicate the existence of pronounced relief in the study site. Palaeotopography was related either to preservation of huge relics of the Drienok–Vernár–Muráň cover nappes in the southern part of the study area, or to the Late Eocene to Oligocene extension-related deepening trend of the Central Carpathian Palaeogene Basin.

The Early Miocene is characterised by a period of tectonic quiescence. Slow erosion in a humid and warm climate led to the formation of broad areas of subdued relief favourable for supergene kaolinization. Most likely, the second Cenozoic planation surface was formed in the Vepor domain in this period. The tectonic quiescence period in the Early Miocene was replaced by Middle Miocene volcanic activity. The volcanic products completely covered the early Miocene palaeosurface and were probably a crucial reason for its preservation in many places.

After the cessation of volcanic activity, a tectonic quiescence period prevailed during the Late Miocene. At this time, an isostatically compensated erosion most probably quickly removed a lot of the Middle Miocene volcanic structures and uncovered the second Cenozoic planation surface developed on the Veporic crystalline basement. During the tectonic quiescence period, this planation surface was remodelled and significantly lowered by intensive mechanical and chemical weathering and gradual denudation accompanied by formation of the third Cenozoic planation surface, as well as small amount of kaolin crust not only on the Veporic crystalline basement, but also on the flattened volcanic rocks. Thus the mid-mountain level, planation surface preserved till now in the central parts of the Vepor and surrounding mountains, has a polygenetic character of a tectoplain. It is a result of a complex history including repeated erosion, peneplanation, and exhumation during a mostly extensional tectonic regime.

Since the Pontian, erosional processes of the Veporic crystalline basement led to transportation and deposition of the weathering crust into the alluvial to lacustrine environments on the northern flanks of the Lučenec and Rimava basins (Poltár Fm.). The final shaping of the Vepor Mountains has been linked to a new accelerated tectonic activity since the Pliocene.

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