*GEOLOGICA CARPATHICA*, AUGUST 2014, 65, 4, 307–326 doi: 10.2478/geoca-2014-0021

# Cretaceous–Quaternary tectonic evolution of the Tatra Mts (Western Carpathians): constraints from structural, sedimentary, geomorphological, and fission track data

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*(Manuscript received July 17, 2013; accepted in revised form June 5, 2014)*

**Abstract:** The Tatra Mts area, located in the northernmost part of Central Western Carpathians on the border between Slovakia and Poland, underwent a complex Alpine tectonic evolution. This study integrates structural, sedimentary, and geomorphological data combined with fission track data from the Variscan granite rocks to discuss the Cretaceous to Quaternary tectonic and landscape evolution of the Tatra Mts. The presented data can be correlated with five principal tectonic stages (TS), including neotectonics. TS-1 (~95–80 Ma) is related to mid-Cretaceous nappe stacking when the Tatric Unit was overlain by Mesozoic sequences of the Fatric and Hronic Nappes. After nappe stacking the Tatric crystalline basement was exhumed (and cooled) in response to the Late Cretaceous/Paleogene orogenic collapse followed by orogen-parallel extension. This is supported by 70 to 60 Ma old zircon fission track ages. Extensional tectonics were replaced by transpression to transtension during the Late Paleocene to Eocene (TS-2; ~80–45 Ma). TS-3 (~45–20 Ma) is documented by thick Oligocene–lowermost Miocene sediments of the Central Carpathian Paleogene Basin which kept the underlying Tatric crystalline basement at elevated temperatures (ca. >120 °C and <200 °C). The TS-4 (~20–7 Ma) is linked to slow Miocene exhumation rate of the Tatric crystalline basement, as it is indicated by apatite fission track data of 9–12 Ma. The final shaping of the Tatra Mts has been linked to accelerated tectonic activity since the Pliocene (TS-5; ~7–0 Ma).

**Key words:** Western Carpathians, Tatra Mts, tectonics, sedimentology, geomorphology, fission track analysis.

## Introduction

The Western Carpathians are the north-easternmost extension of the European Alps. They form a northward-convex E–W trending mountain arc approximately 500 km long and 300 km wide (Fig. 1) belonging to the Alpine orogenic belt in Central Europe. In the west, the Western Carpathians are linked to the Eastern Alps and in the east to the Eastern Carpathians. The Western Carpathians are divided into three principal zones — the External Western Carpathians (EWC), Central Western Carpathians (CWC), and Internal Western Carpathians (IWC) (e.g. Andrusov et al. 1973; Maheľ 1986; Kozur & Mock 1996, 1997; Plašienka et al. 1997; Plašienka 1999; Froitzheim et al. 2008).

The Tatra Mts cover a relatively small area of 785 km$^2$ in the northern portion of the CWC →

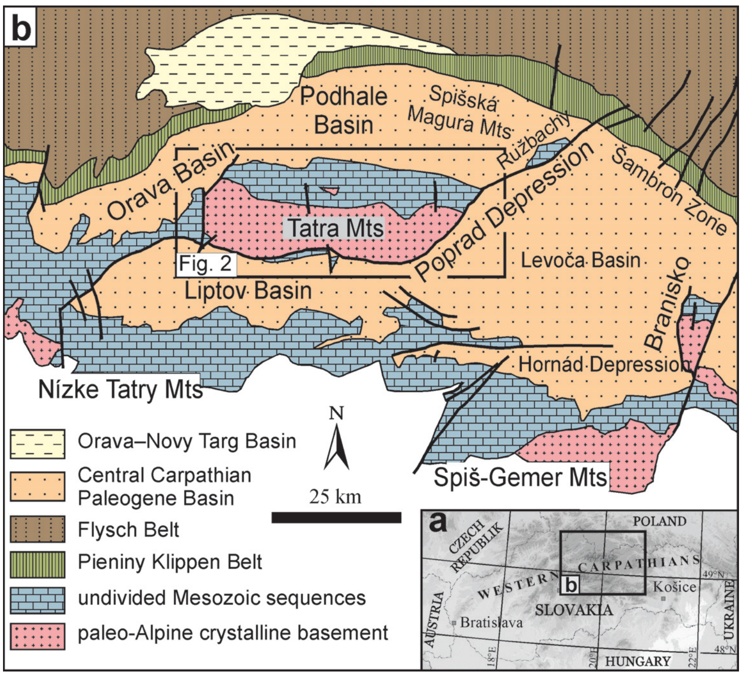


**Fig. 1. a** — Location of the Tatra Mts, **b** — Simplified tectonic map of the Tatra Mts broader area (modified according to Bezák et al. 2004).

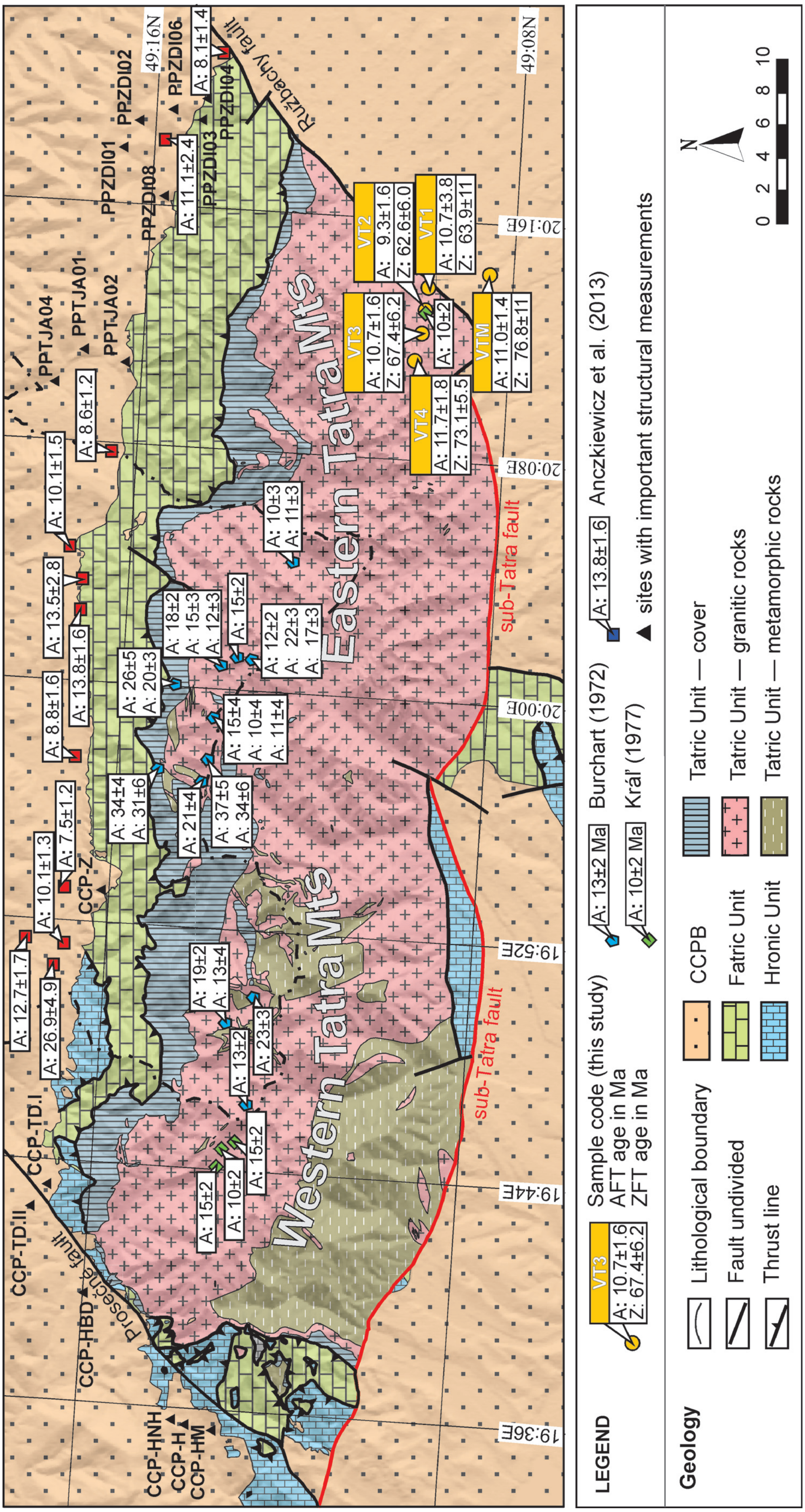


**Fig. 2.** Tectonic map of the Tatra Mts depicting new ZFT and AFT data as well as published FT data (modified according to Nemčok et al. 1994). **Note:** the grey line at the southern foot of the mountains represents the strike of the sub-Tatra fault.

along the state border of Slovakia and Poland (Fig. 1). Their impressive landform is highlighted by Gerlachovský štít (peak) (2655 m a.s.l.), the highest peak in the whole Carpathian arc. Geomorphologically, the mountains are divided into the Eastern (High Tatra Mts and Belianske Tatry Mts) and Western Tatras. The Tatra Mts form an asymmetrical horst-like W-E trending megaanticline. The southern boundary against the Liptov and Poprad subbasins belonging to the Central Carpathian Paleogene Basin (CCPB) is marked by the sub-Tatra fault (Figs. 1 and 2 — Mahel' et al. 1967, 1986). As a result of their unique morphometric character, they were recently defined as markedly the smallest basic morphostructural region of the Western Carpathians (Minár et al. 2011).

In spite of numerous publications on the structural geology of the Tatra Mts area (e.g. Nemčok et al. 1993; Janák et al. 2001; Sperner et al. 2002; Jurewicz 2005; Vojtko et al. 2010), their Alpine geodynamic evolution is still not precisely revealed. Moreover, the great majority of published studies focus only on partial time and thematic aspects. Since the Alpine post-collisional tectonic evolution of the Tatra Mts is still not fully understood, a comprehensive synthetic model using a multidisciplinary approach is needed. For this purpose the structural, sedimentary, and geomorphic data combined with zircon and apatite fission track analysis (ZFT and AFT) were applied.

To achieve a comprehensive model of the tectonic evolution of the Tatra Mts during the Cretaceous to Quaternary, we have focused on the following problems: (i) even though common features of the evolution of the Tatra Mts and the other CWC mountains exist, the recent unique character of the mountains is conditioned by specific features of Cretaceous to Quaternary development, revealed from structural, sedimentary, geomorphological, and geochronological data; (ii) on the other side, relationships between the extreme relief and kinematics of the sub-Tatra fault system and paleostress changes probably exist. They can be revealed by comparison of paleostress variations and development of asymmetry on the basis of lithological, structural, and geomorphological data; (iii) the time of the neotectonic morphological individualization of the Tatra Mts that should be in line with geomorphologic evidence can be estimated from the sedimentary, structural, and geochronological records.

## Regional geological and geomorphological settings

The Tatra Mts are bounded by the denudation remnants of the CCPB (Figs. 1 and 2). The region, located near the boundary zone between the CWC and IWC (boundary: Pieniny Klippen Belt) was affected by strong Miocene deformation (e.g. Ratschbacher et al. 1993; Nemčok & Nemčok 1994; Kováč & Hók 1996; Plašienka et al. 1997; Pešková et al. 2009; Vojtko et al. 2010).

The core of the Tatra Mts is formed by the Tatric crystalline basement (Fig. 2) which is composed of two thick-skinned Variscan tectonic units (Kahan 1969; Janák 1994). Both tectonic units differ in metamorphic grade and lithology. The Jalovec Unit (lower one) occurs only in a tectonic inlayer in the south-western portion of the mountains (Western Tatra Mts). This unit is built up by a monotonous complex of schists and gneisses with intercalations of quartzite. The Baranec Unit (upper one) has a more variegated composition and consists of two complexes. Metamorphic rocks are predominantly composed of migmatite, orthogneiss, paragneiss, and occasionally amphibolite with relics of high pressure metamorphic rocks. The main mass of the upper unit comprises different types of Variscan granites (Kahan 1969; Nemčok et al. 1993, 1994; Janák 1994; Janák et al. 1996, 1999).

The Tatric sedimentary cover (Tomanová cover sequence) is largely of Late Permian to Cretaceous age (Permian–Early Turonian). The autochthonous cover sequence contains the Javorinská Široká, Tomanová, and Osobitá successions. The allochthonous cover sequences are located predominantly in the central part of the mountains and include the Červené vrchy, Giewont, and Široká partial nappes. Overall the Tatric cover sequences display a monoclinal structure with a moderate northward inclination (Fig. 2). The thickness ranges from several hundred meters to a maximum of ~2000 m in the Kominy Tyłkowe area (for further information see Nemčok et al. 1993).

The tectonically overlying the Tatric Unit is a nappe derived from the area between the Tatric and Veporic realms (Biely & Fusán 1967). It can be subdivided into several smaller nappes (predominantly the Bobrovec, Suchy Wierch, Havran, and Bujačí partial nappes) or duplexes that differ mainly by lithostratigraphic content, position, and regional distribution (Fig. 2). The stratigraphic range of the Fatric Unit is Early Triassic–Early Cretaceous. The age of thrusting is constrained by the deposition of a synorogenic flysch — the Poruba Formation (Albian to Early Turonian) in the Fatric and Tatric units (e.g. Andrusov et al. 1973; Plašienka 2003).

The Hronic Unit represents structurally the highest nappe system. It appears only in the western portion of the Tatra Mountains (Fig. 2). This nappe contains sedimentary strata, from Anisian to Toarcian in age. It is composed of two partial nappes, the Siwa Woda (lower) and the Furkaska-Koryčiská (upper) (Nemčok et al. 1993). The lithological composition of these partial nappes refers to the Ludrová succession which is interpreted as a slope between the Reifling Basin and the Mojtín-Harmanec carbonate platform (cf. Havrila 2011).

These nappes form the substratum of the Eocene to earliest Miocene CCPB sedimentary succession (Podtatranská skupina Group; cf. Gross et al. 1984, 1993; Sliva 2005 — Fig. 2) which is predominantly composed of deep-marine siliciclastics up to 4 km in thickness (Soták et al. 2001). The Podtatranská skupina Group is divided into four formations (Golab 1959; Roniewicz 1969; Gross et al. 1984) — the Borové Formation a.k.a. "Numulitic Eocene", Huty Formation a.k.a. Zakopane Member, Zuberec Formation a.k.a. Chocholów Formation, and Biely potok Formation a.k.a. Ostrysz Formation (Fig. 3 column "Lithostratigraphy") with stratigraphic span from Middle Eocene to earliest Miocene (Olzsewska & Wieczorek 1998; Soták 1998, 2010; Gedl 2000; Garecka 2005). Traditionally, the CCPB is interpreted as a fore-arc basin located behind the Carpathian accretionary wedge (e.g. Royden & Báldi 1988; Tari et al. 1993; Kázmér et al. 2003) and now surrounding the Tatra Mts.

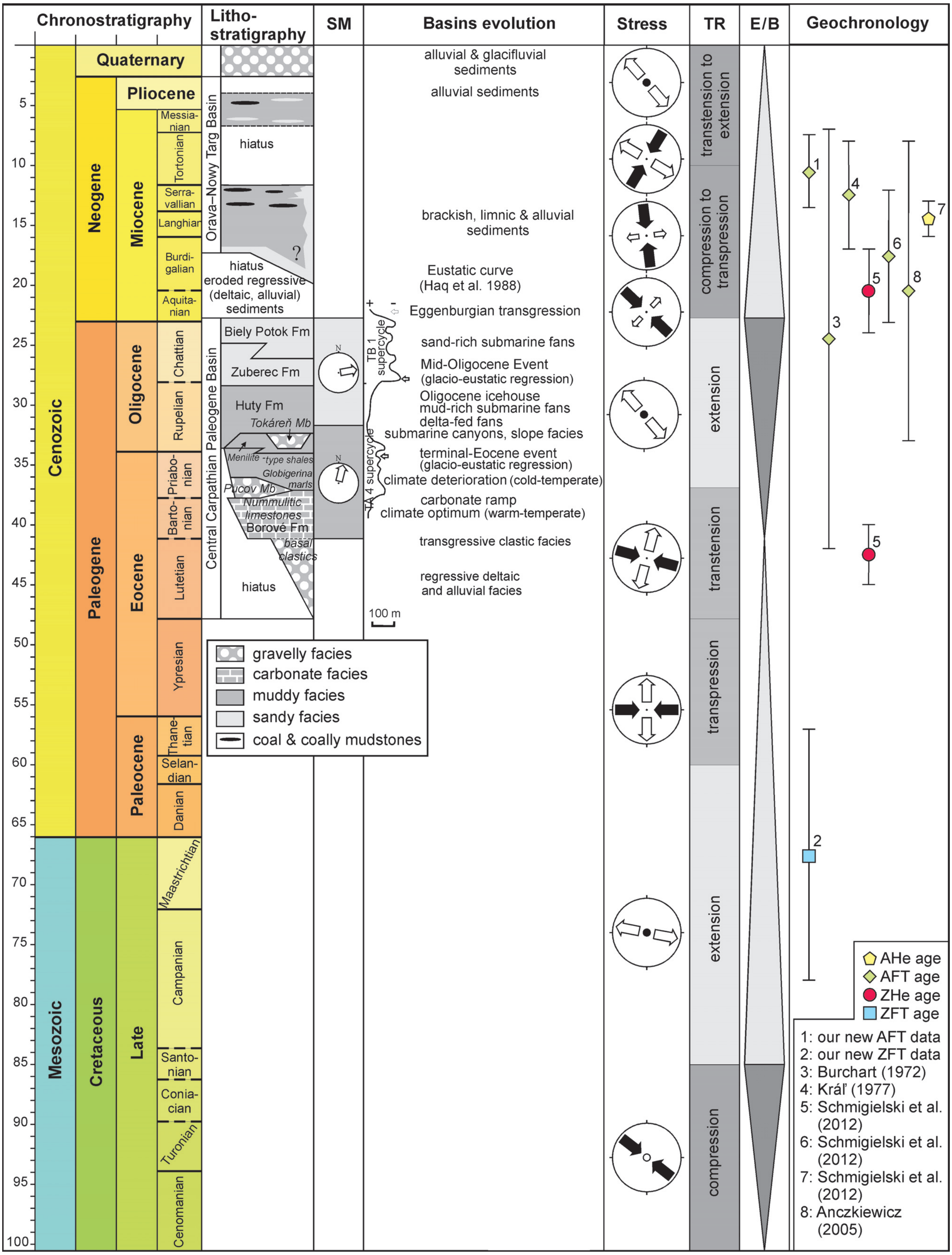


**Fig. 3.** Summary of litostratigraphy, paleotransports, paleostress fields, tectonic regimes, exhumation/burial, and geochronological data, indicating Mesozoic to Cenozoic geodynamic development of the Tatra Mts and its surroundings. **SM** — sedimentary marks show principal direction of paleocurrents, **TR** — tectonic regime, **E/B** — exhumation (denudation) vs. burial (accumulation).

The Miocene and Pliocene sediments are not preserved in the Tatra Mts and they are not confirmed from its immediate surrounding as well. However, the Middle Miocene to Pliocene terrestrial and fresh water sediments, up to 1300 m thick form the fill of the Orava–Nowy Targ Basin (Fig. 3 column "Lithostratigraphy") located several kilometers north (Watycha 1976; Gross et al. 1993). Their original extents probably reached further southward and covered a large part of the adjacent sediments of the CCPB (Wagner 2011).

Significant amounts of Quaternary sediments can be found in the southern foothills of the High Tatra Mountains where massive Pleistocene moraine and fluvioglacial sediments are more than 400 m thick. These sediments are thought to have been deposited in a graben which was related to normal faulting along the Ružbachy and sub-Tatra faults (e.g. Nemčok et al. 1993, 1994).

The Tatra Mts form an asymmetrical horst structure associated with the Miocene exhumation, which occurred along the sub-Tatra fault system as was confirmed using several geological methods (e.g. Burchart 1972; Andrusov et al. 1973; Kráľ 1977; Maheľ 1986; Kováč et al. 1994; Plašienka et al. 1997; Janák et al. 2001; Baumgart-Kotarba & Kráľ 2002; Struzik et al. 2002; Anczkiewicz 2005; Anczkiewicz et al. 2005, 2013; Śmigielski et al. 2012). The exhumation continued during the neotectonic period (Pliocene–Quaternary) as is indicated by the large amount, several hundred meters in thickness, of the Quaternary glaciofluvial sediments distributed predominantly at the southern foot of the Tatra Mts (e.g. Nemčok et al. 1993).

The Tatra Mts differ from the rest of the Western Carpathians by their exceptional geomorphological character. The mountains are characterized by dominant glacial relief formed during the Pleistocene (Lukniš 1973) and the most intensive periglacial processes in the recent (Midriak 1983). Absence of significant remains of planation surfaces is another specific geomorphic feature of the Tatra Mts. The variation of altitude (750–2650 m a.s.l.) is bigger than anywhere else in the Western Carpathians. In addition, the mountains are characterized by values of mean slopes (~25°) and available relief (~700 m), which are much higher than in the other mountains of the CWC (~5° vs. ~150 m).

Distinctive W–E as well as N–S geomorphological differentiation is typical for the Tatra Mts. Altitudinal difference between the Eastern and Western Tatra Mts reach about 400 m. However, the Belianske Tatry Mts (north-eastern part of the Eastern Tatra Mts) are approximately 500 m lower than the rest of the Eastern Tatra Mts (High Tatra Mts). N–S altitudinal asymmetry is also clearly visible in the High Tatra Mts alone — the difference between average altitudes of S and N ridges reaches about 100 m here (Holle 1909 ex Lukniš 1973).

## Methods

### *Structural analysis*

Standard procedures for brittle fault-slip analysis and paleostress reconstruction are now well established (Etchecopar et al. 1981; Angelier 1990, 1994). Paleostress axes characterized by the brittle structures were computed by the WinTensor software (Delvaux & Sperner 2003) using the method of Angelier (1984). Fault data were inverted to obtain the four parameters of the reduced stress tensor: $\sigma_1$ (maximum principal stress axis), $\sigma_2$ (intermediate principal stress axis), $\sigma_3$ (least principal stress axis), and also the $\Phi$ ratio of principal stress differences. The stress regime was expressed numerically from the $\Phi$ ratio using an index $\Phi'$ (according to Delvaux et al. 1997).

### *Sedimentological analysis*

The sedimentary and geodynamic development of the Tatra Mts and their surroundings have been studied also using sedimentological analysis applied to the CCPB strata, especially focusing on the Spišská Magura Mts and eastern part of the Podhale region. Detailed sedimentary logs, available from well-preserved outcrops, have been constructed and correlated to obtain information on sedimentary environments. Paleotransport reconstruction was based on measurement of existing marks on bedding planes, cross bedding, lamination, and also on measurement of channel axis and synsedimentary folds as well. Provenance analyses of coarse clastics were based on macroscopic and thin-section observations.

### *Geomorphological analysis*

Analysis was performed by use of digital elevation models derived from the basic topographic maps of the Slovak Republic 1:50,000, geological maps 1:50,000 (Nemčok et al. 1994) and geomorphological map 1:50,000 (Lukniš 1968). ArcGIS 9.3 software was used for processing.

The paragenetic system of the Studený potok stream (including not only recent but also all Pleistocene accumulations of the stream) was the subject of analysis. It includes the area of Slavkovský štít (peak) (2452 m a.s.l.) where fission track analysis was performed. The denudation segment — DS (valley in the Tatra Mts) and accumulation segment — AS (glacial and glaciofluvial sediments on the foothills) were distinguished. On the basis of outcrops of pre-Quaternary rocks (Paleogene deposits and crystalline rocks) recorded in geological and geomorphological maps, a few boreholes and geophysical profiles, a first approximation of the Quaternary base surface of the AS was created (using Topo to Raster interpolation from the selected points). The final approximation was performed by adding new points of the Quaternary base (–3 m below recent surface of Quaternary sheet) in the places were the first approximation rose up to the recent surface.

Envelope surfaces were created for both, the AS and DS from the highest ridge line points. Control by recent surface was performed too (envelope surface cannot reach below recent surface). The volume of preserved accumulation as well as all retention space in the AS was calculated by subtraction of the Quaternary base surface from the recent and envelope surfaces. The volume of the DS was determined by analogy.

### *FT analysis*

Sample preparation and FT analysis were carried out at the Fission Track Laboratory of the Institute of Geology, University of Innsbruck. Apatite and zircon separation was done with standard heavy mineral separation techniques using magnetic and heavy liquid separation. The apatites were mounted in epoxy resin and zircons in PFA® Teflon, then ground and polished. Spontaneous fission tracks were revealed by etching in 6.5% $HNO_3$ for 40 s at 20 °C. The zircons were etched in a NaOH–KOH eutectic melt for 4–8 h at 235 °C. Irradiation of both apatite and zircon was carried out at the FRM II research reactor in Garching, Germany. Neutron flux was monitored using CN5 (for apatite) and CN1 (for zircon) dosimeter glasses. After irradiation, the induced fission tracks in external detector muscovites were etched in 40% HF for 45 min at 20 °C. The fission tracks were counted with 1250× magnification with a dry objective using a computer-controlled Zeiss Axioplan microscope equipped with an automated AUTOSCAN stage. The samples were analysed using the external detector method (EDM) as described by Gleadow (1981). All ages are 'central ages' (Galbraith & Laslett 1993) with errors quoted as $\pm 1\sigma$. The central ages were calculated using the zeta calibration method (Hurford & Green 1983) with a zeta factor of $372 \pm 35.8$ year/cm$^2$ (apatite, CN5 glass) and $185.8 \pm 11$ year/cm$^2$ (zircon, CN1 glass) (analyst: S. Králiková). Data processing was carried out using the TRACKKEY program, version 4.2.f (Dunkl 2002). Only grains with their prism planes parallel to the polished surface were used for age dating. 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 on polished apatite surfaces (Dpar method — Donelick 1993) were measured as a proxy for annealing properties. Horizontal confined fission track lengths were measured in apatite samples and mean horizontal confined track lengths (Mean HCTL) with one-sigma standard deviation ($\sigma$) were determined.

## Data used

### *Structural data*

Structural data obtained from the Tatra Mts area were used to reconstruct the Late Cretaceous to Quaternary tectonic evolution of the study area. Four main deformation phases related to principal burial vs. exhumation/denudation processes were identified in the paleo-Alpine nappe units and one in the CCPB deposits. The last deformation phase affected both the paleo-Alpine nappe units as well as the CCPB deposits. The Cenozoic paleostress fields were calculated in detail and published in the works of Pešková et al. (2009), Vojtko et al. (2010), and Sůkalová et al. (2012). Analysis of structural measurements, as well as a geological and structural study of the Tatra region show relative clockwise rotation of the paleostress field during the Late Cretaceous to Cenozoic (Figs. 3, 4 and Table 1).

Late Cretaceous–Paleocene orogenic collapse, exhumation/denudation, post dating the nappe stacking, is documented within the whole CWC territory especially in the Veporic and Gemeric belt (e.g. Hók et al. 1993; Plašienka 1993, 1999; Jeřábek et al. 2012). The compressional tectonic regime was followed by extensional tectonics (Fig. 3) which occurred predominantly on low angle normal faults.

From the Paleocene to Eocene a change from transpression to transtension together with a shift from WSW–ENE to ENE–WNW compression can be observed (Fig. 3). The timing brackets for this stage are based on stratigraphic arguments with the upper limit given by undeformed Upper Eocene to Oligocene formations.

The Early–Middle Miocene was a time characterized by reverse faulting and, to a lesser extent, by strike-slip faulting with the maximum stress axis ($\sigma_1$) oriented NW–SE (Fig. 3). Especially, intense folding on all scales can be observed throughout and the reverse sub-Tatra fault allowed for uplift and exhumation of the Tatra Mts in the northern part of the working area.

During the late Middle and Late Miocene a gradual change in the paleostress orientation was observed. The paleostress axis ($\sigma_1$) rotated to the NE–SW position in the Late Miocene (Fig. 3). Additionally, the tectonic regime passed from transpression through transtension to tension. The Middle to Late Miocene was characterized by a generally N–S-trending compression in a strike-slip and compressional tectonic regime.

The youngest tectonic regime (neotectonics) is characterized by the E–W extension, which has been documented in the Orava–Nowy Targ Basin (Pešková et al. 2009). The Quaternary tension is parallel to the Western Carpathian arc ($S_h$), and the $S_H$ of the stress field is generally N–S (Fig. 3). However, some disturbances can occur, for example in the Kozie Chrbty Mts and the Hornád Depression in the south of the study area (Sůkalová et al. 2012).

### *Sedimentological data*

The aim of the sedimentological study was to better understand the burial vs. exhumation story of the Tatra Mts region during the Cenozoic. The main goal was to complement all relevant previous sedimentary studies of the CCPB and Orava–Nowy Targ Basin in combination with our data and to provide a contemporary view of the basins' development. Sedimentological data played a principal role in solving the Cenozoic evolution of the Tatra region.

Formation of the CCPB started with the deposition of transgressive facies from the Middle to Late Eocene (Lutetian–Priabonian) while the Oligocene to earliest Early Miocene is characterized by a thick complex of deep-sea fan sediments (Fig. 3). Basin evolution was controlled by tectonics, as documented by syn-sedimentary deformation (Starek 2001; Pešková et al. 2009; Vojtko et al. 2010), by climatic and by sea-level changes (Janočko & Jacko 1998; Soták 1998, 2010; Soták & Starek 2000; Soták et al. 2001).

Sedimentation of the Middle–Upper Eocene Borové Formation started with thick alluvial, deltaic and shallow marine clastics overlain by shallow marine and organodetrital limestones. The uppermost part of the formation is composed of bryozoan and Globigerina marls (Soták 2010). The overall transgressive character of the Borové Formation was interrupted in the Priabonian by a short regression (Fig. 3), namely the Pucov Member (cf. Starek et al. 2012).

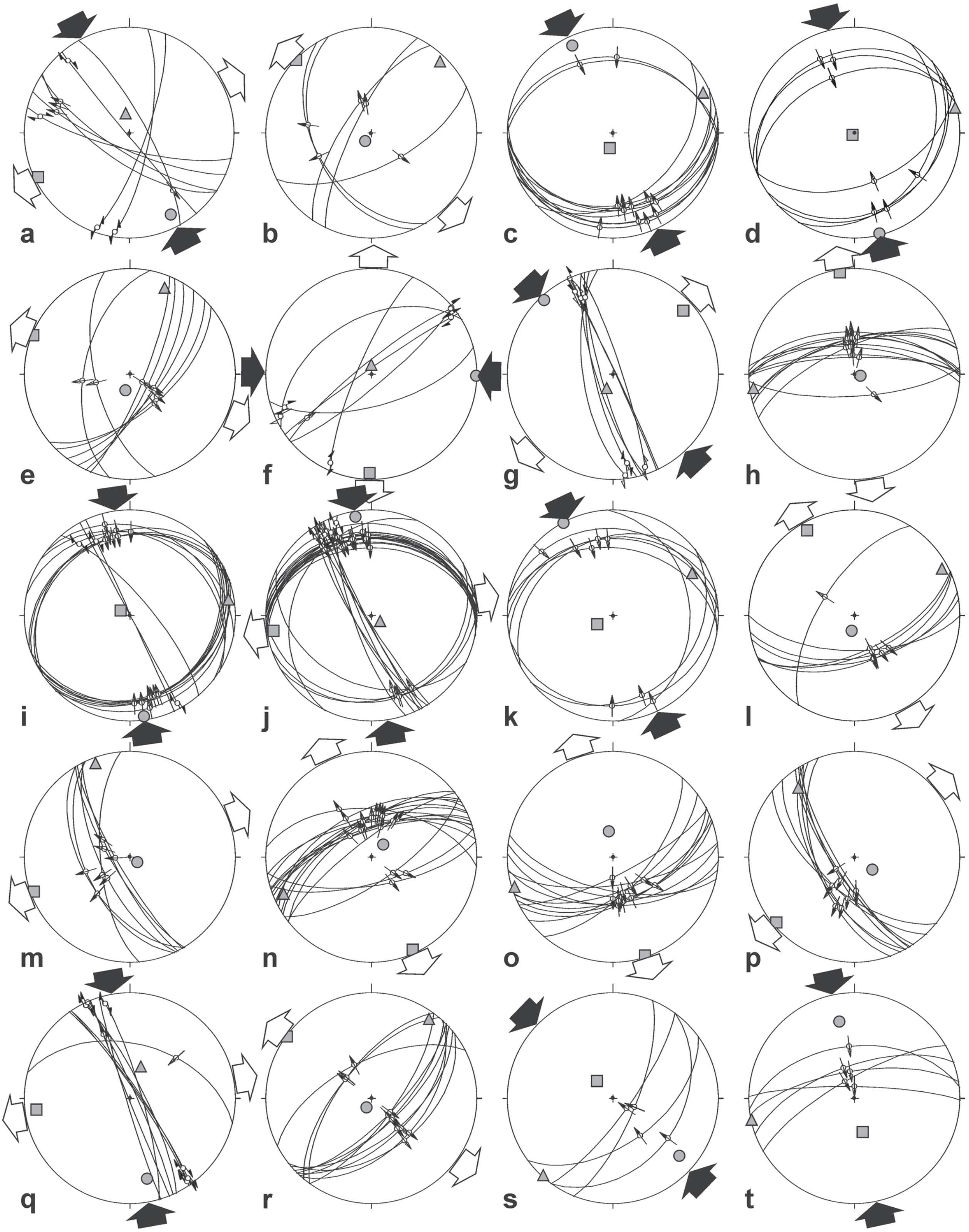


**Fig. 4.** Examples of paleostress reconstructions for the Tatra Mts region. **Explanation:** Stereogram (Lambert's net, lower hemisphere) with traces of fault planes, observed slip lines and slip senses and principal paleostress axes (circle = $\sigma_1$, triangle = $\sigma_2$ and square = $\sigma_3$). **Note:** For further information on paleostress tensors see Table 1 and for evolution see Fig. 3 column "Stress". **a** — CCP-HM1, **b** — PPTJA04A, **c** — PPTJA04B, **d** — PPTJA04C, **e** — PPTJA04D, **f** — PPTJA01A, **g** — PPZDI01A, **h** — PPZDI01B, **i** — PPZDI01D, **j** — PPZDI01E, **k** — PPZDI01F, **l** — PPZDI01G, **m** — PPZDI01H, **n** — PPZDI02A, **o** — PPZDI02E, **p** — PPZDI02F, **q** — PPZDI08A, **r** — CCP-TDII.1, **s** — CCP-TDI.2, **t** — CCP-Z4.

**Table 1:** Paleostress tensors from fault slip data. **Explanations:** Site — code of locality, n — number of fault used for stress tensor determination, $n_T$ — total number of fault data measured, $\sigma_1$, $\sigma_2$, and $\sigma_3$ — azimuth and plunge of principal stress axes, $\Phi$ — stress ratio ($\sigma_2$–$\sigma_3$/$\sigma_1$–$\sigma_3$), Method — direct inversion (DI), numerical dynamic analysis (NDA), Event — tectonic regime (flattening, strike-slip, extension).

| Site | Formation | n | $n_T$ | $\sigma_1$ | $\sigma_2$ | $\sigma_3$ | $\Phi$ | Method | Event |
|---|---|---|---|---|---|---|---|---|---|
| CCP-H1 | Huty Fm | 7 | 26 | 139/07 | 230/09 | 011/79 | 0.38 | DI | flattening |
| CCP-H2 | Huty Fm | 6 | 26 | 096/71 | 226/13 | 319/14 | 0.48 | DI | extension |
| CCP-H3 | Huty Fm | 4 | 26 | 306/83 | 120/07 | 210/01 | 0.5 | DI | extension |
| CCP-H4 | Huty Fm | 3 | 26 | 195/22 | 298/28 | 072/53 | – | NDA | flattening |
| CCP-H5 | Huty Fm | 5 | 26 | 205/03 | 296/05 | 078/84 | 0.5 | DI | flattening |
| CCP-HBD1 | Huty Fm | 5 | 6 | 308/15 | 217/04 | 113/74 | 0.44 | DI | flattening |
| CCP-HBD2 | Huty Fm | 1 | 6 | 063/22 | 153/01 | 244/68 | – | NDA | flattening |
| CCP-HBD3 | Huty Fm | 2 | 6 | 215/76 | 010/12 | 101/06 | – | NDA | extension |
| CCP-HM1 | Borové Fm | 9 | 46 | 153/14 | 350/75 | 244/04 | 0.61 | DI | strike-slip |
| CCP-HM2 | Borové Fm | 6 | 46 | 210/20 | 065/65 | 305/13 | 0.38 | DI | strike-slip |
| CCP-HM3 | Borové Fm | 15 | 46 | 237/87 | 001/02 | 091/02 | 0.29 | DI | extension |
| CCP-HM4 | Borové Fm | 4 | 46 | 300/33 | 035/07 | 135/56 | 0.25 | DI | flattening |
| CCP-HM5 | Borové Fm | 4 | 46 | 116/21 | 300/68 | 207/02 | 0.47 | DI | strike-slip |
| CCP-HNH1 | Huty Fm | 3 | 4 | 270/06 | 167/64 | 003/25 | – | NDA | strike-slip |
| CCP-TDI.1 | Zuberec Fm | 3 | 11 | 163/62 | 063/05 | 330/27 | – | NDA | extension |
| CCP-TDI.2 | Zuberec Fm | 4 | 11 | 299/11 | 029/01 | 127/79 | 0.5 | DI | flattening |
| CCP-TDI.3 | Zuberec Fm | 2 | 11 | 359/25 | 190/65 | 097/01 | – | NDA | strike-slip |
| CCP-TDII.1 | Zuberec Fm | 12 | 33 | 208/82 | 036/08 | 306/01 | 0.5 | DI | extension |
| CCP-TDII.2 | Zuberec Fm | 4 | 33 | 131/18 | 222/02 | 317/72 | 0.61 | DI | flattening |
| CCP-TDII.4 | Zuberec Fm | 2 | 33 | 218/74 | 309/01 | 039/16 | – | NDA | extension |
| CCP-Z1 | Huty Fm | 5 | 25 | 255/30 | 032/52 | 152/22 | 0.5 | DI | strike-slip |
| CCP-Z2 | Huty Fm | 4 | 21 | 163/21 | 026/63 | 260/17 | 0.46 | DI | strike-slip |
| CCP-Z3 | Huty Fm | 3 | 25 | 157/59 | 055/07 | 312/30 | – | NDA | extension |
| CCP-Z4 | Huty Fm | 5 | 25 | 349/27 | 258/01 | 166/63 | 0.5 | DI | flattening |
| CCP-Z5 | Huty Fm | 5 | 25 | 136/62 | 315/28 | 226/00 | 0.5 | DI | extension |
| PPTJA01A | Zuberec Fm | 1 | 2 | 105/70 | 272/20 | 3/4 | – | NDA | extension |
| PPTJA01B | Zuberec Fm | 1 | 2 | 4/16 | 273/2 | 176/74 | – | NDA | flattening |
| PPTJA02A | Huty Fm | 2 | 3 | 109/38 | 206/8 | 306/51 | – | NDA | extension |
| PPTJA02B | Huty Fm | 1 | 3 | 160/52 | 206/8 | 356/37 | – | NDA | extension |
| PPTJA04A | Zuberec Fm | 5 | 38 | 216/82 | 44/7 | 314/1 | 0.43 | DI | extension |
| PPTJA04B | Zuberec Fm | 12 | 38 | 335/10 | 66/7 | 191/78 | 0.40 | DI | flattening |
| PPTJA04C | Zuberec Fm | 7 | 38 | 166/2 | 76/2 | 218/88 | 0.72 | DI | flattening |
| PPTJA04D | Zuberec Fm | 8 | 38 | 195/77 | 22/13 | 292/2 | 0.21 | DI | extension |
| PPVTA01A | Gutenstein Lm. | 7 | 15 | 91/0 | 1/83 | 181/7 | 0.20 | DI | strike-slip |
| PPVTA01B | Gutenstein Lm. | 2 | 15 | 140/7 | 231/10 | 15/78 | – | NDA | flattening |
| PPZDI01A | Huty Fm | 7 | 107 | 317/5 | 204/78 | 48/11 | 0.14 | DI | strike-slip |
| PPZDI01B | Huty Fm | 13 | 107 | 106/85 | 262/4 | 352/3 | 0.36 | DI | extension |
| PPZDI01C | Huty Fm | 4 | 107 | 98/74 | 226/9 | 318/12 | 0.51 | DI | extension |
| PPZDI01D | Huty Fm | 18 | 107 | 172/4 | 81/6 | 298/82 | 0.39 | DI | flattening |
| PPZDI01E | Huty Fm | 32 | 107 | 351/6 | 122/82 | 261/7 | 0.12 | DI | strike-slip |
| PPZDI01F | Huty Fm | 8 | 107 | 332/0 | 62/16 | 241/7 | 0.25 | DI | flattening |
| PPZDI01G | Huty Fm | 8 | 107 | 190/78 | 62/7 | 331/9 | 0.38 | DI | extension |
| PPZDI01H | Huty Fm | 7 | 107 | 126/83 | 2340/6 | 250/4 | 0.41 | DI | extension |
| PPZDI02A | Huty Fm | 16 | 58 | 44/77 | 247/11 | 156/4 | 0.31 | DI | extension |
| PPZDI02B | Huty Fm | 7 | 58 | 278/74 | 145/11 | 53/11 | 0.58 | DI | extension |
| PPZDI02C | Huty Fm | 3 | 58 | 61/74 | 314/5 | 222/16 | – | NDA | extension |
| PPZDI02D | Huty Fm | 1 | 58 | 257/25 | 153/27 | 9/58 | – | NDA | strike-slip |
| PPZDI02E | Huty Fm | 12 | 58 | 351/70 | 253/3 | 162/20 | 0.52 | DI | extension |
| PPZDI02F | Huty Fm | 9 | 58 | 126/73 | 321/17 | 230/4 | 0.49 | DI | extension |
| PPZDI03A | Carpathian Keuper | 7 | 37 | 284/38 | 107/52 | 15/1 | 0.62 | DI | strike-slip |
| PPZDI03B | Carpathian Keuper | 4 | 37 | 246/74 | 118/10 | 26/13 | 0.41 | DI | extension |
| PPZDI03C | Carpathian Keuper | 12 | 37 | 265/88 | 21/1 | 111/2 | 0.37 | DI | extension |
| PPZDI03D | Carpathian Keuper | 5 | 37 | 181/20 | 294/47 | 76/36 | 0.36 | DI | extension |
| PPZDI03E | Carpathian Keuper | 4 | 37 | 1/22 | 96/12 | 212/65 | 0.35 | DI | flattening |
| PPZDI03F | Carpathian Keuper | 5 | 37 | 203/17 | 324/60 | 105/25 | 0.22 | DI | strike-slip |
| PPZDI04A | Carpathian Keuper | 5 | 16 | 329/11 | 236/13 | 98/73 | 0.47 | DI | flattening |
| PPZDI04B | Carpathian Keuper | 11 | 16 | 31/85 | 206/5 | 296/0 | 0.50 | DI | extension |
| PPZDI06A | Borové and Huty Fm | 14 | 14 | 150/6 | 241/11 | 29/77 | 0.47 | DI | flattening |
| PPZDI08A | Tokáreň Mb | 9 | 12 | 168/23 | 19/64 | 263/12 | 0.51 | DI | strike-slip |
| PPZDI08B | Tokáreň Mb | 3 | 12 | 160/78 | 349/12 | 259/2 | – | NDA | extension |

The thick Lower Oligocene Huty Formation comprises mostly fine-grained deep water deposits (Fig. 3). Its lower regressive part is formed by various types of dark, organic-rich shales and thick intraformational conglomerates of the Tokáreň Member (Sliva 2005). In the eastern part of the study area, sedimentary facies suggest a rather proximal position (Marschalko & Radomski 1970; Janočko & Jacko 1998; Janočko et al. 2000; Sliva 2005). The N to NE directed paleocurrents (Fig. 3) together with clasts (CWC source) from fossil submarine canyons (Tokáreň Member), outcropping on the northern slopes of the Tatra Mts, are clear evidence for Gemeric and Veporic sources and non-existence of the Tatra Mts. Although in the Podhale and Orava parts of the CCPB the paleotransport is not unequivocally determined, meter-sized Mesozoic boulders in thick unsorted breccias and conglomerates favour short transport and proximity to the basin margin and the CWC source.

The Lower Oligocene transgressive part of the Huty Formation is formed by typical mud-rich deep marine fan deposits without thick sandstones and conglomerates (Soták 1998; Soták et al. 2001). Sediments of the Huty Formation gradually change into regressive facies of the Zuberec and Biely potok Formations — typical deep marine flysch deposits (Fig. 3). Paleocurrent analysis showed reorientation from marginal to basin axis position, indicating an increase in tectonic activity and change of sedimentary sources (Sliva 2005). In the Orava–Spišská Magura Mts region, eastward prograding fans were deposited, while in the Šarišská vrchovina–Levočské vrchy Mts fans prograded westwards (e.g. Marschalko & Radomski 1960; Krysiak 1976; Starek 2001; Sliva 2005).

The youngest age (NP25–NN1 nannoplankton zone) of the CCPB sediments was reported from the Biely Potok Formation in many sites (Gedl 2000; Soták & Starek 2000; Starek et al. 2000; Soták et al. 2001; Starek 2001; Garecka 2005; Sliva 2005). Sedimentological research in these deposits confirmed their submarine fan origin (Janočko & Jacko 1998; Soták 1998; Janočko et al. 2000; Soták & Starek 2000; Starek et al. 2000; Soták et al. 2001; Starek 2001; Sliva 2005). Gradual shallowing and decrease of salinity is indicated by microfossil associations (Starek et al. 2000; Soták et al. 2001). Regressive, deltaic, and fluvial facies have not been identified due to Miocene basin inversion followed by denudation. The upper limit for the regressive deposits could be determined by the NN2 Zone because a new transgressive sedimentary cycle started at the NN2/NN3 boundary in the East Slovakian Basin (e.g. Soták et al. 2001). Our estimation points to several hundred meters of missing regressive sediments (deltaic and alluvial) which most probably have been redeposited in newly formed Neogene basins in the hinterland.

The Neogene sedimentary record is preserved in the Orava–Nowy Targ Basin, which originated during the Middle Miocene (Watycha 1976; Nagy et al. 1996). The basin was previously considered to be a retro-arc basin (Roth (Ed.) 1963). However, because of its position near the Periklippen shear zone, some authors considered the basin to be a pull-apart type (Pospíšil 1990; Pomianowski 2003). The basin infill is composed of Burdigalian, Langhian, and Serravalian coarse-grained sandstone, claystone, and intercalations of lignitic claystone to lignite. The sediments were deposited in brackish, limnic, and alluvial environments. The basin also contains Upper Miocene to Lower Pliocene greenish-grey claystone and siltstone with intercalations of sandstone lying erosively on the Middle Miocene strata (Fig. 3).

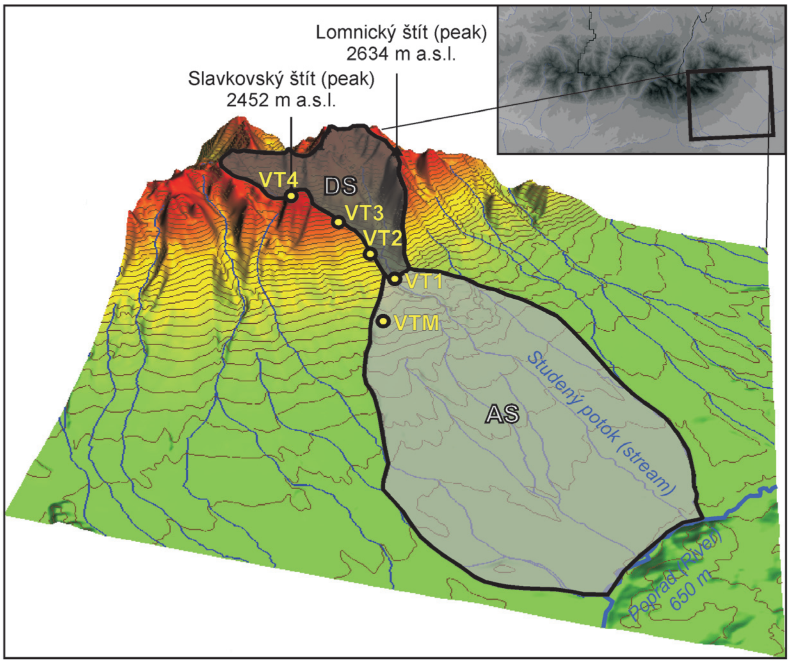


**Fig. 5.** Paragenetic system of the Studený potok stream (Slavkovský štít and Lomnický štít peak area). **DS** — denudation part of the system (Malá and Veľká studená dolina — valleys in the High Tatras), **AS** — accumulation part of the system (Podtatranská kotlina — basin). **VT-4–VT-1, VTM** — sample code locations. Contour interval 10 m.

**Table 2:** Denudation rates (**DR**) estimated from various possible ages of oldest Quaternary sediments, **DH** — mean denudation levelling of the denudational segment.

| | Area [km$^2$] | Volume [km$^3$] | DH [m] | DR 500 kyr [mm/a] | DR 750 kyr [mm/a] | DR 1000 kyr [mm/a] |
|---|---|---|---|---|---|---|
| **Denudation segment** | 14.05 | 2.07 | | | | |
| **Accumulation segment** | 69.51 | | | | | |
| **Preserved sediments** | | 0.83 | 59 | 0.12 | 0.08 | 0.06 |
| **Minimum (recent) storage space** | | 3.68 | 262 | 0.52 | 0.35 | 0.26 |
| **Construed maximum storage space (rising 0.03 mm/a)** | | 3.96<br>4.10<br>4.24 | 282<br>292<br>302 | 0.56 | 0.39 | 0.32 |

### *Geomorphological data*

The main goal of geomorphological analysis was to estimate a minimum Quaternary denudation in the Studený potok paragenetic system (Fig. 5). The estimation was limited by the following vagueness: (i) storage sedimentary space on the mountain foothill is open so only a fragment of former accumulation is preserved; (ii) an unknown amount of sediments was not deposited on the foothills (suspended, bed and dissolved load of the Studený potok (stream) shifted directly to the Poprad River); (iii) the glacial and glaciofluvial sediments, preserved on the highest watersheds of the accumulation segment (AS), were never numerically dated and their age is only roughly estimated as Early Pleistocene to Mindel (Elsterian) (Lukniš 1968; Nemčok et al. 1993); (iv) storage space computed from the altitude of the highest recent remnants of Quaternary sediments is underestimated because of their denudation; (v) glacial cycles could infill and empty some parts of the storage space several times, which again leads to underestimation of the resulting denudation rates.

Denudation rates from in situ produced cosmogenic nuclides from regolith profiles (Cockburn & Summerfield 2004) as well as denudation rates computed for small flat river catchments from terrace deposits (e.g. Schaller et al. 2002; Hidy et al. 2014) generally vary about 0.01–0.1 mm/a. A set of estimations of minimum denudation rates was done with respect to the age of the oldest sediments (500–1000 kyr) as well as the maximum denudation rate of their remnants (0.03 mm/a — see Table 2).

Beside this, additional relevant morphometric data were obtained from the morphometric analysis: (i) difference between mean altitude of denudation segment (1880 m a.s.l.) and accumulation segment (779 m a.s.l.) reached 1100 and 1200 m a.s.l. in medians, respectively; (ii) depths of valley bottoms below dividing ridges predominantly vary from 300 to 600 m; (iii) volume of denudation segment — valley is approximately half the volume of the accumulation segment — storage space (denudation/accumulation volume ratio is 0.56 for minimum and 0.49 for maximum storage space). It means that about half the accumulation has come from the space above the present valley and denudation of ridges is about half the general denudation.

### *FT data*

The sampling strategy was to quantify amplitude and timing of vertical movements in the Tatra Mts, using the example of Slavkovský štít (peak). In order to determine exhumation rate, an elevation profile in the facetted fault slope of Slavkovský štít (peak) was sampled every 400 m of altitude. Because of the possibility to record rocks from above the recent peaks in the mountains and to specify information about the denudation rate in the Tatra Mts by this way, the oldest moraine material below Slavkovský štít (peak) was studied as well. Four samples from the Carboniferous granitic rocks of Slavkovský štít (peak) together with one sample from the granite boulders of Middle Pleistocene moraine in the foothills were dated by the FT method. All samples were taken from surface outcrops (Figs. 5 and 6; Table 3).

Zircon central ages were obtained from five samples (VT-1, VT-2, VT-3, VT-4, and VTM) and range from 62.6 ± 6.0 to 76.8 ± 11 Ma (Table 3, Fig. 2). All analyses passed the $\chi^2$-squared test ($P(\chi^2)$ > 5 %; Galbraith 1981), indicating that all single grain ages statistically belong to the same population. Unfortunately samples VT-1 and VTM yielded a too low number of grains for further meaningful treatment although the ages are in line with the other samples.

Apatite central ages are significantly younger than the aforementioned zircon central ages, ranging between 9.3 ± 1.6 and 11.7 ± 1.8 Ma (Table 3,

**Table 3:** ZFT and AFT data of this study. **N** — number of counted grains per sample, **$\rho_s$, ($\rho_i$)** — density of spontaneous (induced) tracks ($\times 10^5$ tr/cm$^2$), **Ns, (Ni)** — number of counted spontaneous (induced) tracks, $\rho_d$ — density of dosimeter tracks ($\times 10^5$ tr/cm$^2$), **Nd** — number of counted dosimeter tracks, **$P(\chi^2)$** — probability of obtaining $\chi^2$ values for n degrees of freedom where n = number of crystals-1; central age (Ma) ± 1 σ error (Galbraith & Laslett 1993), **Dpar** — measured long axis of etch-pits which are parallel to the crystallographic c-axis (Donelick 1993), **L** — mean horizontal confined track lengths, **SD** — 1σ standard deviation of the sample, **N (L)** — number of horizontal confined tracks measured, **SE** — standard error of the mean.

| Sample | Latitude | Longitude | Altitude | Petrography | Chronostratigraphy | N | $\rho_s$ | Ns | $\rho_i$ | Ni | $\rho_d$ | Nd | $P(\chi^2)$ | Central age | Dpar | L | SD | N (L) | SE |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Code | WGS-84 | | (m) | | | | | | | | | | % | (Ma) ± 1σ | (μm) | (μm) | (μm) | | (μm) |
| *Zircon* | | | | | | | | | | | | | | | | | | | |
| VT1 | 49°09'47.9"N | 20°13'39.7"E | 1180 | granodiorite | Carboniferous | 3 | 52.87 | 92 | 40.81 | 72 | 5.33 | 3078 | 70 | 63.9 ± 11 | | | | | |
| VT2 | 49°09'49.9"N | 20°12'54.4"E | 1664 | granodiorite | Carboniferous | 10 | 53.31 | 441 | 40.98 | 339 | 5.2 | 3078 | 98 | 62.6 ± 6.0 | | | | | |
| VT3 | 49°09'53.2"N | 20°12'07.3"E | 2033 | granodiorite | Carboniferous | 14 | 59.77 | 528 | 43.36 | 383 | 5.290 | 3078 | 88 | 67.4 ± 6.2 | | | | | |
| VT4 | 49°10'00.4"N | 20°12'07.3"E | 2456 | granodiorite | Carboniferous | 26 | 59.64 | 1442 | 40.08 | 969 | 5.32 | 3078 | 100 | 73.1 ± 5.5 | | | | | |
| VTM | 49°08'27.9"N | 20°14'11.9"E | 986 | moraine | Middle Pleistocene | 3 | 63.15 | 157 | 39.82 | 99 | 5.25 | 3078 | 89 | 76.8 ± 11 | | | | | |
| *Apatite* | | | | | | | | | | | | | | | | | | | |
| VT1 | 49°09'47.9"N | 20°13'39.7"E | 1180 | granodiorite | Carboniferous | 4 | 1.45 | 9 | 26.48 | 164 | 10.46 | 3000 | 93 | 10.7 ± 3.8 | 1.1 | | | | |
| VT2 | 49°09'49.9"N | 20°12'54.4"E | 1664 | granodiorite | Carboniferous | 22 | 1.640 | 53 | 34.94 | 1129 | 10.70 | 3000 | 100 | 9.3 ± 1.6 | 1.3 | 12.1 | 1.7 | 32 | 0.3 |
| VT3 | 49°09'53.2"N | 20°12'07.3"E | 2033 | granodiorite | Carboniferous | 20 | 2.36 | 84 | 37.32 | 1326 | 9.060 | 3000 | 100 | 10.7 ± 1.6 | 1.3 | 12.7 | 1.5 | 53 | 0.2 |
| VT4 | 49°10'00.4"N | 20°12'07.3"E | 2456 | granodiorite | Carboniferous | 21 | 2.14 | 75 | 29.94 | 1051 | 8.82 | 3000 | 100 | 11.7 ± 1.8 | 1.6 | 12.7 | 1.6 | 84 | 0.2 |
| VTM | 49°08'27.9"N | 20°14'11.9"E | 986 | moraine | Middle Pleistocene | 50 | 1.72 | 151 | 29.48 | 2593 | 10.156 | 3000 | 100 | 11.0 ± 1.4 | 1.4 | 13.4 | 1.4 | 48 | 0.2 |

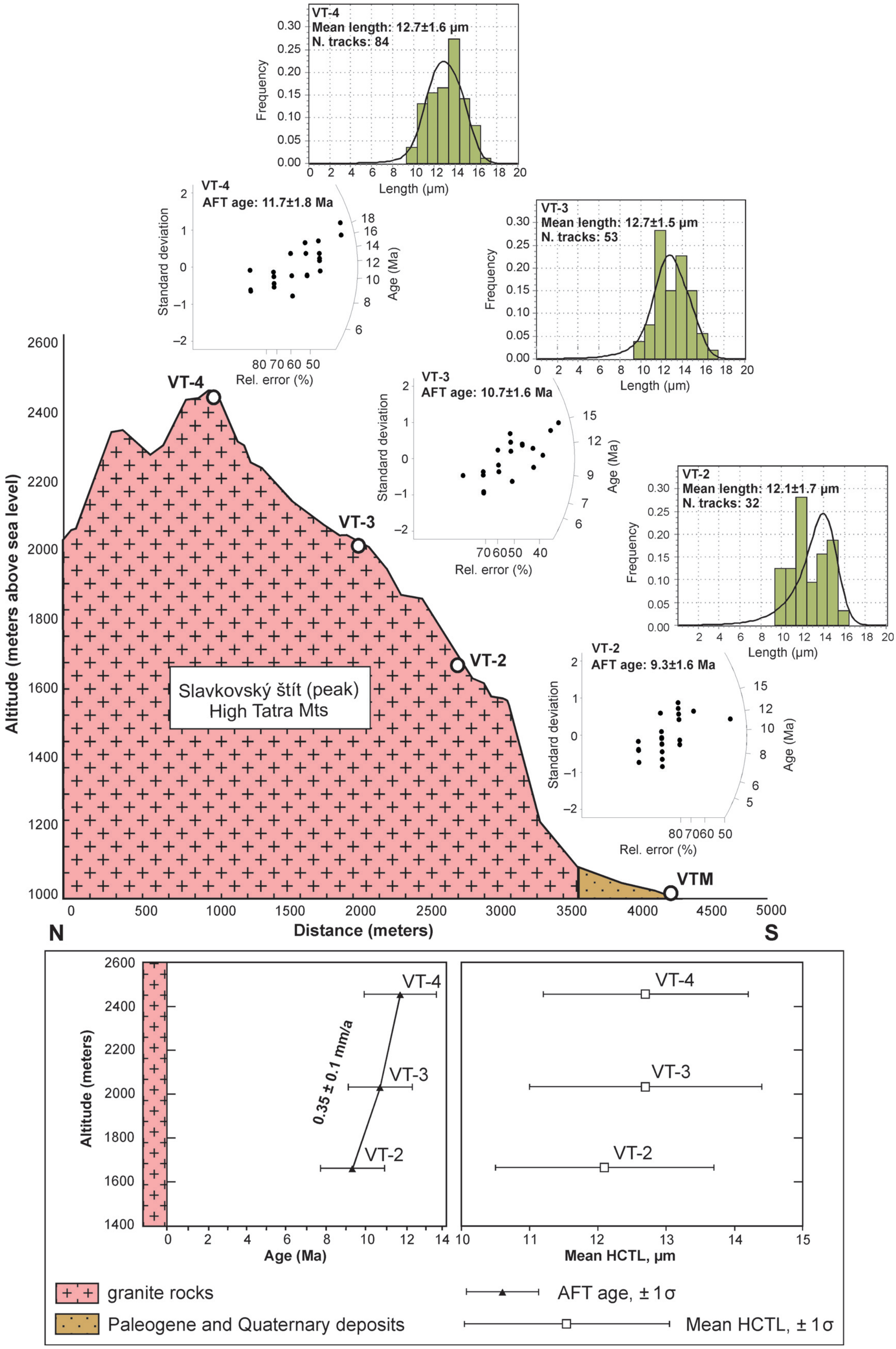


**Fig. 6.** Schematic elevation profile of Slavkovský štít (peak), High Tatra Mts, showing samples location, lithology, radial plots of single-grain AFT age data with ± 1 σ errors, horizontal lengths diagrams (N. tracks — number of measured tracks), central AFT ages with estimated exhumation rate, and mean horizontal confined track lengths (Mean HCTL) with 1 σ standard deviation.

Fig. 2). All apatite samples (VT-1, VT-2, VT-3, VT-4, and VTM) passed the $\chi^2$-squared test ($P(\chi^2)$ >5 %; Galbraith 1981). Only sample VT-1 provided too few grains for meaningful interpretation. Despite the fact that not too many fission track lengths could be measured, the track length distributions (TLD) of horizontally confined fission tracks are unimodal, negatively skewed, with fairly broad standard deviations [SD between 1.4 (VTM sample) and 1.7 µm (VT-2 sample) Table 3] and short mean horizontally confined track lengths [Mean HCTL between 13.4 (VTM sample) and 12.1 µm (VT-2 sample) Table 3]. Such TLD's are common in basement rocks with slow protracted cooling through the apatite partial annealing zone (APAZ; ~60–120 °C; e.g. Wagner & Van den Haute 1992). In all of the analysed apatites, Dpar values do not vary much, ranging between 1.1 and 1.6 µm (Table 3), indicating fairly similar chemical compositions and relatively fluorine rich apatites (Burtner et al. 1994) with a low resistance to annealing (Ketcham et al. 1999).

## Interpretation and discussion

### *The first tectonic stage (TS-1; ~95–80 Ma)*

TS-1 reflects mid-Cretaceous nappe stacking originated during the northwards progressing paleo-Alpine Cretaceous convergence (Fig. 7). Within the working area detached cover nappes formed in the northern frontal part of the accreted Carpathian system (e.g. Plašienka et al. 1997; Jurewicz 2005). Generally, NW directed thrusting propagated from the inner CWC outwards into the foreland lower plate (Plašienka 2003; Prokešová et al. 2012). Deformation occurred under brittle-ductile to brittle conditions with consistent top-to-the-northwest shear (Fig. 8a,b). During this stage the Tatric Unit took a lower position in the nappe stack and was metamorphosed under anchizonal and/or lower greenschist facies conditions (Fig. 8b). The fully annealed zircon samples indicate that the metamorphic temperature was in excess of ~320 °C (proposed upper temperature limit of the zircon partial annealing zone — ZPAZ; Tagami et al. 1998). The peak metamorphic temperature had to be less than ~350 °C within the Tatric crystalline basement, because the $^{40}Ar/^{39}Ar$ and Rb/Sr on muscovite and biotite methods yielded isotopic age values for the granitic rocks in the range of 330–280 Ma. These ages most probably record magmatic cooling followed by exhumation of the granite pluton during the Late Variscan orogenesis and they are not affected by the Alpine tectogenesis (Janák & Onstott 1993; Maluski et al. 1993; Janák 1994).

### *The second tectonic stage (TS-2; ~80–45 Ma)*

After the tectonic burial beneath the paleo-Alpine nappe pile, an exhumation by unroofing of the Tatric Unit occurred during the Late Cretaceous–Paleocene (Fig. 7). Exhumation processes occurred predominantly on the low angle normal faults where the principal tension axis operated in the generally W–E direction with general top-to-the-east shear (see Fig. 3). In the western part of the mountains, tectonic contact of the Hronic and extremely reduced Fatric units vs. the Tatric crystalline basement along the low-angle faults is considered to be a consequence of this exhumation process but the kinematics are almost opposite.

The Late Cretaceous–Early Paleocene cooling of the Tatric basement from the depth of ca. ~10 km was also revealed by the presented ZFT data with the ages scattered between ~60 and 70 Ma. Due to the late-Eocene transgression it is obvious that at least the western Tatric basement cooled from Cretaceous $T_{max}$ to surface temperatures. Only because of reheating in the course of formation of the CCPB, the AFT samples do not record this early phase of cooling (e.g. Kráľ 1977; Anczkiewicz 2005 and data herein) except at least two detrital samples in the Podhale Basin (5/02 and 22/02) with AFT peak ages of 55–60 Ma (cf. Anczkiewicz 2005). Formation of the CCPB caused substantial burial of the Tatric basement and total resetting of the AFT system in the east. However,

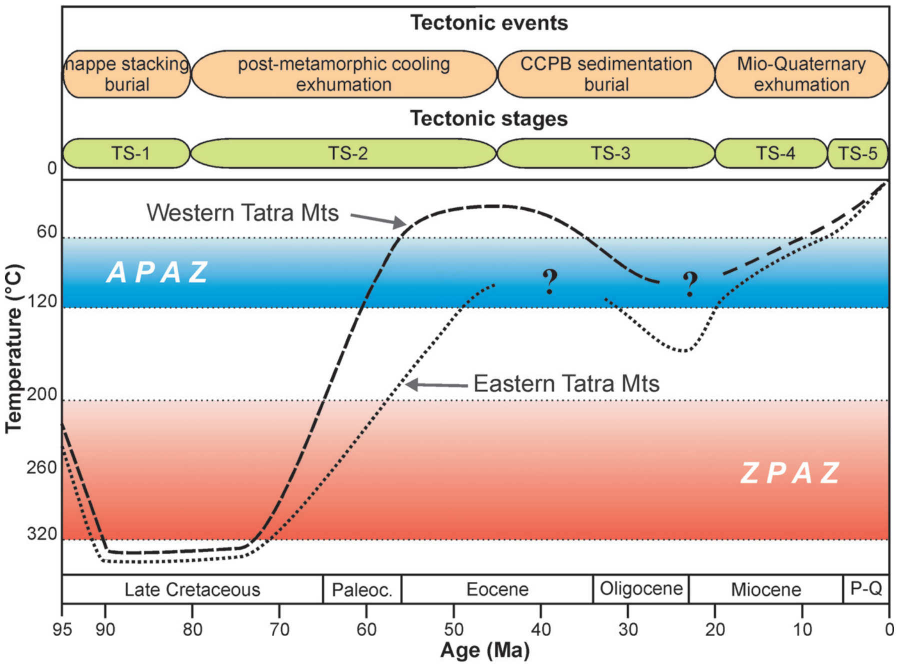


**Fig. 7.** Summarizing time-temperature evolution of the Tatra Mts crystalline basement drawn from regional geological considerations (main tectonic events and tectonic stages recorded in the study area are depicted at the top of the diagram), from results of ZFT and AFT data, and additional published thermochronological data. **ZPAZ** — zircon partial annealing zone, **APAZ** — apatite partial annealing zone; double dot-dashed line represents idealized fit for the tectono-thermal evolution of the Tatric crystalline basement.

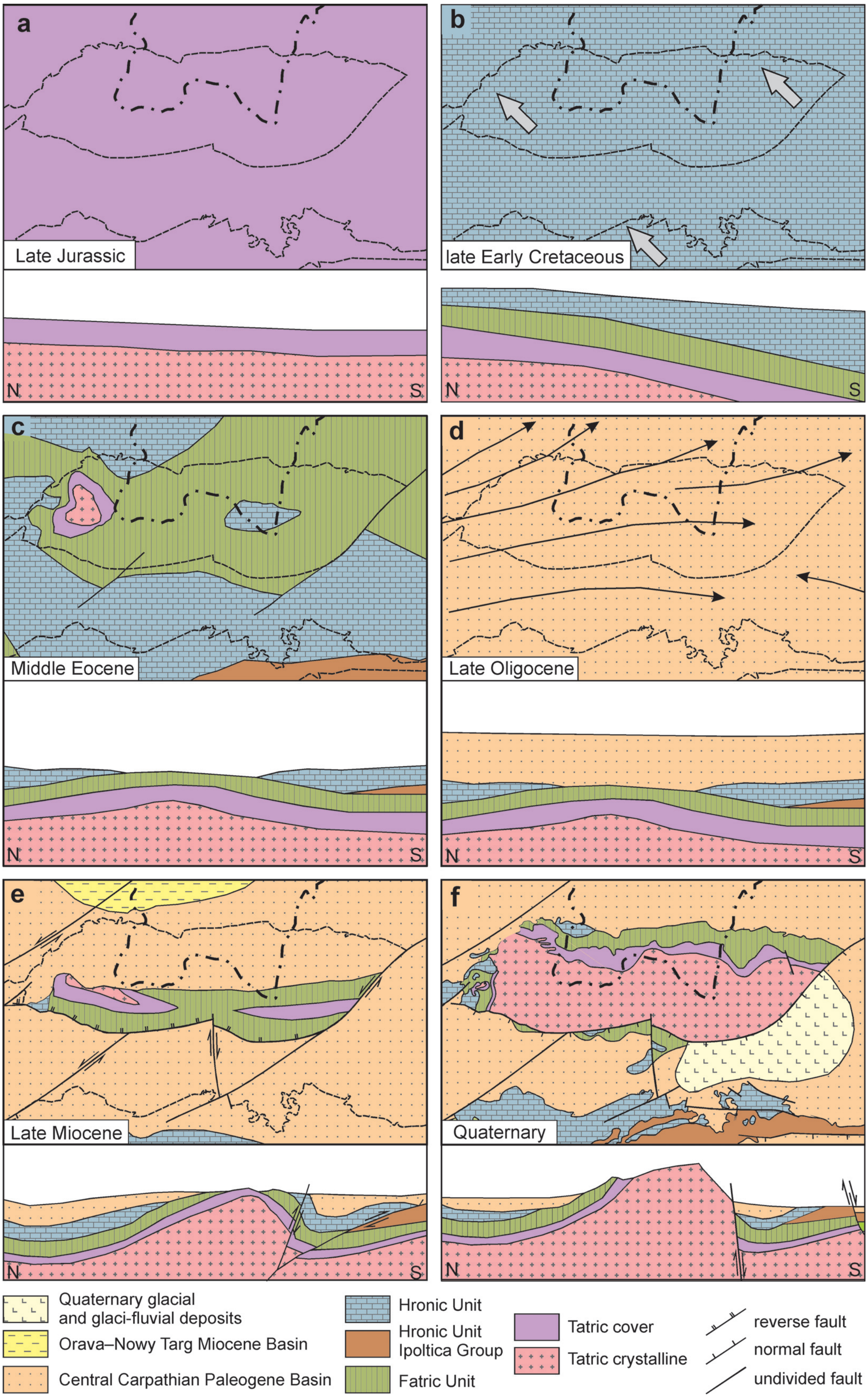


**Fig. 8.** Schematic sketch of the geodynamic evolution of the Tatra Mts during the Early Cretaceous to Quaternary depicted in tectonic maps and simplified geological cross sections striking approximately along the middle part of the maps. **Note:** grey arrows in 8b represent general direction of thrusting.

partial resetting and thus less overburden may probably be indicated by older (mixed?) AFT ages in the Western Tatra Mts (Burchart 1972; Struzik et al. 2002 and Anckiewicz 2005 — see samples T3 and 9/01).

Basically, these exhumation and denudation processes were recorded in the whole of the Tatra Belt as well as in the Veporic and Gemeric units. Published ZFT data from the Tatra Belt yielded ages of ~70 Ma in the Považský Inovec Mts (Kováč et al. 1994), ~53 Ma in the Tribeč Mts (Kováč et al. 1994), and ~45 to 70 Ma in the Malá Fatra Mts and the basement of the Turiec Basin (Danišík et al. 2010; Králiková 2013; Králiková et al. 2014). In the southern zone of the CWC, the Veporic Unit yielded ZFT age of ~65–75 Ma (Plašienka et al. 2007; Králiková 2013; Vojtko et al. 2013) and the Gemeric Unit provided ZFT ages between ~62 and 88 Ma (Plašienka et al. 2007). According to the ZFT dating results, the first conspicuous exhumation processes after the Alpine nappe stacking definitely took place mainly during the Late Cretaceous–Paleocene. It is evident, that this exhumation, well-known in the Veporic Unit (e.g. Hók et al. 1993; Plašienka 1993, 1999; Jeřábek et al. 2007, 2008, 2012), prograded to the external, Tatra Unit.

After the exhumation the hanging wall leftovers were markedly eroded and a last remnant of the Hronic Unit is preserved in the north-western edge of the Tatra Mts. In the Western Tatra Mts (Košariská and Mačacie diery sites), erosion was even more efficient and Upper Eocene strata transgressed onto the pre-Mesozoic Tatric basement (cf. Passendorfer 1958; Andrusov 1959; Gorek & Scheibner 1966; Nemčok et al. 1994). It should be stressed that some authors interpret the contact between Upper Eocene strata and granite in the Košariská site as tectonic origin (e.g. Uhlig 1897; Lugeon 1903). However, this contact lacks any tectonic reworking. Scarce paleo-current markers (towards west to north-west), the Borové Formation clasts and granite pebbles in the stratotype site of the Pucov Member conglomerates (Priabonian) point to a source area in the Western Tatra Mts during the Priabonian. In the central and eastern portion of the Tatra Mts such direct evidence of the Tatric crystalline basement erosional level has never been recorded (Fig. 8c). Moreover, there are no indications of Tatric crystalline pebbles in the Borové Formation and Tokáreň Member so far (e.g. Janočko et al. 2000; Sliva 2005).

### *The third tectonic stage (TS-3; ~45–20 Ma)*

TS-3 is linked to the formation of the CCPB. It shows a fore-arc basin position developed on the destructive plate-margin of the ALCAPA microplate and behind the EWC accretionary wedge (Tari et al. 1993; Soták et al. 2001; Kázmer et al. 2003). During the sedimentation of the CCPB sequences, the former area of the Tatra Mts was buried under the relatively thick Middle Eocene–lowermost Miocene strata (Figs. 7 and 8d). The known thickness of the CCPB sediments is ~3.5 km in the Levočské vrchy Mts (Gross et al. 1999), ~2.25 km in the Spišská Magura Mts (Janočko et al. 2000), ~3.0 km in the Podhale Basin (Kępińska 1997), ~1.5 km in the Liptov Basin (e.g. Gross et al. 1980), and ~2.5 km in the Orava region (Fusán et al. 1987). At present, the Liptov Basin has the minimum thickness of the CCPB strata among the portions of the CCPB. This is apparently caused by its more intense uplift followed by erosion during the Neogene and Quaternary. In addition, the known paleotransports documented in the Oligocene sequences (in the Orava, Liptov, Podhale, and Spišská Magura Mts) point to eastward transport of sedimentary material fed by gravity flows (Sliva 2005). From the sedimentological standpoint, the Oligocene sequences of the Orava region, Liptov and Podhale basins, and Spišská Magura Mts represent one submarine fan system (Fig. 8d). It means that the Tatra region was completely overburdened by the CCPB sediment pile. The inferred thickness of the eroded and preserved section varies from <3.1–3.9 km in the Orava region to >5.5–6.9 km in the Spišská Magura Mts close to the Ružbachy fault, calculated for 20 and 25 °C/km paleogradients (e.g. Środoń et al. 2006). On the basis of these data, the total thickness of the CCPB sedimentary sequences in the Tatra region, eroded and preserved, can be estimated as ~4.0–5.0 km. Thus, a combined tectonic and sedimentary total overburden of the Tatric basement on the order of ~4.0 km in the west and up to 7.0 km in the eastern portion of the study area can be assumed after the CCPB formation, as was depicted also by K-Ar dating on bentonite at ~16–19 Ma (Środoń et al. 2006). With an assumed geothermal gradient of ~25 °C/km this leads to metamorphic temperatures in the Tatric basement (except the west with slightly lower temperatures) between 160 and 220 °C. Published zircon U-Th/He (ZHe) data (Śmigielski et al. 2012) and the presented AFT data are well in line with this estimate. Predominantly there are two preferred models: (i) on the basis of X-ray diffraction study of illite combined with K-Ar dating of bentonites, Środoń et al. (2006) postulate burial beneath variable thickness of Cenozoic sediment pile (~4 km in the west to ~7 km in the east of the study area) at the stable geothermal gradient of 21 ± 2 °C/km as the only heat source. The authors evaluation of the burial depth is independently supported by fluid-inclusion data published by Hurai et al. (2002) that refer to burial depth of 5.3–6.5 km (130–205 °C) close to the Ružbachy fault, and 2–3 km (95–100 °C) in the Orava territory. Besides, Środoń et al. (2006) published additional evidence of burial as the only heat source that was revealed by measurements of different grain densities and porosities in the Bukovina and Chocholow wells in the Podhale Basin. According to the authors, these differences respond primarily to the lithostatic load and not to the temperature; (ii) according to AFT dating study from the Podhale–Spišská Magura Basin, Anczkiewicz et al. (2013) assume both heating associated with mid-Miocene volcanism and variable thickness of Oligocene- and potentially also Miocene sediments as the responsible heat source. Their interpretation shifts the extent of the mid-Miocene thermal episode to the northern edge of the Pannonian block and thus provides an extra heat source. According to the authors, the additional heat source could be linked with the elevated paleogeothermal gradient (>25 °C/km) estimated by Marynowski & Gaweda (2005) and Kepińska (2006), on the basis of vitrinite reflectance and illite-smectite data.

On the basis of our research together with known geological knowledge we inclined to the first model presented by Środoń et al. (2006). For many reasons, it is disputable to consider the interpretation of ‘the mid-Miocene thermal event as the extra heat source that prograded to the northern edge of the Pannonian block' proposed by Anczkiewicz et al. (2013). One of the arguments is the lack of evidence that this thermal event affected the Tatra Mts crystalline basement. This is documented by a big scatter in AFT ages that were not completely reset in the Middle Miocene, or even later. Another important fact is that in the areas close to the huge volcanic edifices, such as the Tribeč, Žiar, and Veporské vrchy Mts, the AFT system was not influenced by this thermal event, providing the ages older than Miocene or Oligocene (see AFT data in the works of Danišík et al. 2004, 2008; Plašienka et al. 2007; Králiková 2013; Vojtko et al. 2013). Therefore, it is very unlikely that the mid-Miocene thermal event would affect the very remote and relatively cold periphery of the CWC without influencing immediately surrounding areas with the main volcanic activity (Central Slovak Neovolcanic Field region). Additionally, our new ZFT data of Late Cretaceous to Paleocene age, that were not influenced by an elevated paleogeothermal gradient (>25 °C/km) in the Middle Miocene, may be regarded as further circumstantial evidence. If we assume ~7 km of total overburden of the Tatra Mts paleo-Alpine basement after the formation of the CCPB in the Early Miocene and the Mesozoic strata both nappe and cover sequences, then a paleogeothermal gradient of 30 °C/km, or even more, should at least partially or completely reset the ZFT system in the eastern part. However, we missed this evidence. The same argument applies to the ZHe data published by Śmigielski et al. (2012). On the basis of these arguments it is necessary to look for another interpretation because the paleogeothermal gradient had to be in the range of 20–25 °C/km. For these reasons, the most likely interpretation appears to be the combination of burial beneath the CCPB strata and subsequent exhumation in the Miocene as was described in detail in the work of Králiková et al. (2014).

### *The fourth tectonic stage (TS-4; ~20–7 Ma)*

TS-4 is characterized by exhumation of the Tatric crystalline basement after the deposition of the Oligocene–lowermost Miocene strata followed by a quiet period predominantly during the Late Miocene to Early Pliocene (Fig. 7). The measured brittle structures in the CCPB sediments indicate that the exhumation process was controlled by a compressional tectonic regime (Early Miocene), followed by a compressional to strike-slip tectonic regime (Middle to Late Miocene) and finally by an extensional tectonic regime (Pliocene to Quaternary). The paleostress field progressively changed, where the $S_{Hmax}$ relatively rotated from NW–SE through N–S to NE–SW directions (Fig. 3; for further information and detailed paleostress data see Pešková et al. 2009; Vojtko et al. 2010; Sůkalová et al. 2012). The obtained data helped us to understand the tectonic processes during the exhumation, denudation, and formation of the modern Tatra Mts relief. The Tatra Mts are an asymmetrical horst where the Mesozoic sedimentary cover and nappe units occur on the northern slopes with the moderate to steep inclination of bedding planes northward (Fig. 8e). The principal role during the exhumation of the mountains is ascribed to the W–E striking sub-Tatra fault with its several cross-cutting and arching fault segments such as the Ružbachy fault on the east and Prosečné (a.k.a. Choč-Prosečné-Krowiarski) fault on the west. The results of fault-slip analysis and paleostress reconstruction make it possible to interpret the kinematics of the main fault zones participating in the Tatra Mts structure.

Previous models of the Tatra Mts exhumation/uplift were predominantly based on map-scale geological data. Generally, five models were used in the past: (i) exhumation happened on the W–E sub-Tatra reverse fault with north inclination as was accepted predominantly during the first half on the 20$^{th}$ century (e.g. Sokolowski 1948; Gorek 1956; Andrusov 1958, 1968; Maheľ et al. 1967; Biely & Fusán 1967; Piotrowski 1978); (ii) the exhumation occurred on the sub-Tatra normal fault. This interpretation was supported by technical and geophysical works (Gross 1973; Gross et al. 1980) where the fault was interpreted as subvertical, or even a steeply south-dipping fault plane. However, this fault plane orientation was recognized only on its accompanying subsidiary faults. This model was widely accepted especially in the second half of the 20$^{th}$ century (e.g. Maheľ 1986; Nemčok et al. 1993); (iii) a new conceptual and kinematically consistent model was proposed by Sperner (1996) and Sperner et al. (2002). The model was based on measurement of brittle deformation in the broader area of the Tatra Mts. The main displacement on the sub-Tatra fault was considered to be reverse southward movement with strike-slip combination on the Ružbachy and Prosečné faults and the Tatra massif was uplifted as a compressional horst structure with a combination of strike-slip duplex (Ružbachy horst); (iv) the sub-Tatra fault is a part of planar normal faults which were produced by tilt and spin rotations of domino-type prismatic upper crustal blocks that formed due to horizontal top-to-the north simple shear of the CWC crust triggered by underthrusting of the Northern European plate (Grecula & Roth 1978; Marko 1995); (v) the last proposed model of the Tatra massif exhumation as a compressive structure involving the basement was published by Janák et al. (2001) according to whom the sub-Tatra fault operated as a frontal subvertical fault and the fault bend syncline was located in/along the northern foot of the Tatra Mts.

Interpretation of fault-slip analysis and paleostress reconstruction can help us to understand the kinematics of main fault shear zones in the Tatra region (Figs. 3 and 8). The Early Miocene paleostress was caused by a compressional tectonic regime with the NW–SE principal compression, and exhumation of the Tatra Mts started most probably along the reverse Ružbachy fault and transpressive to oblique reverse sub-Tatra fault. The uplift started as a compressive structure involving the basement as was proposed by Janák et al. (2001). During the Middle Miocene, the compressional axis rotated to the generally N-S direction and the Ružbachy fault operated as a sinistral strike-slip fault (transform fault) and the sub-Tatra fault as a reverse fault. Exhumation was carried out by a combination of strike-slip movement and reverse faulting (Sperner 1996; Sperner et al. 2002). The Late Miocene paleostress field can be characterized by the NE-SW oriented principal compressional axis (e.g. Sperner 1996; Pešková et al. 2009; Vojtko et al. 2010; Sůkalová et al. 2012) and the kinematics of the Ružbachy and Prosečné faults successively changed from transpression via transtension up to normal faulting. Along the Ružbachy fault, the hanging wall was located in the SE side with down-throw movement and along the Prosečné fault, the hanging wall down-throw north-westward. The Tatra Mountains started to be an individualized horst structure in the northern portion of the CWC. During this stage, the reverse movement of the sub-Tatra fault changed to oblique transpressive sinistral strike-slip up to transtensive sinistral strike-slip at the end of the Late Miocene. The Pliocene to Quaternary period is characterized by an extensional tectonic regime with general principal tension in the NW-SE direction. The orientation of the paleostress field caused predominant normal faulting along the sub-Tatra and Ružbachy faults where the south block subsided. Most authors agree that the Prosečné, sub-Tatra, and Ružbachy faults, which restricted the asymmetric horst-like structure of the Tatra Mts, operated as normal faults during the neotectonic phase (e.g. Maheľ 1986; Nemčok et al. 1993).

The timing of the Middle to early Late Miocene exhumation processes of the Tatra Mts crystalline basement was revealed by two independent thermochronological methods, using AFT and apatite U-Th/He (AHe) analysis (Burchart 1972; Kráľ 1977; Struzik et al. 2002; Anczkiewicz 2005; Anczkiewicz et al. 2005; Śmigielski et al. 2012 — Fig. 2). The presented new AFT ages in the range of $9.3 \pm 1.6$ and $11.7 \pm 1.8$ Ma (Figs. 2 and 6) from the Slavkovský štít (peak) area are in good agreement with the previous published AFT data from the Tatra crystalline basement (Kráľ 1977; Anczkiewicz et al. 2005), revealing the late Middle to early Late Miocene exhumation from the depth of at least 5 km (recalculated by the geothermal gradient at 20–25 °C/km). The elevation profile in Slavkovský štít (peak) allowed us to determine an approximate exhumation rate of ca. $0.35 \pm 0.1$ mm/a. The result is consistent with obtained track length measurements of identical short Mean HCTL and fairly broad SD (VT-2–VT-4 samples), indicating simple continuous moderate cooling of the Tatric crystalline rocks through the APAZ at $9.3 \pm 1.6$ to $11.7 \pm 1.8$ Ma (Fig. 6). It is clear that no particular event occurred in this time span. A similar exhumation rate ($0.2 \pm 0.1$ mm/a) was determined in the Mount Rysy, High Tatra Mts, during the Miocene (Anckiewicz et al. 2005). The revealed early Late Miocene slow exhumation rate can be related to the initial planation surface formation of the modern relief (intramountain level) in the Western Carpathians (e.g. Mazúr 1965; Starkel 1969; Minár et al. 2011; Zuchiewicz 2011). This period was replaced by accelerated uplift during the latest Late Miocene to Pleistocene (Fig. 8f). According to Baumgart-Kotarba & Kráľ (2002), the uplift from the ~2 km depth (~60°) most probably started in the time span of ~7–2 Ma. Moreover, the authors postulate an uplift rate of up to 1.0 mm/a for this time range because of a short time to uncover the Tatric crystalline basement of the High Tatra Mts. Additionally, sedimentological data confirm the outlined scenario. After Tokarski et al. (2012) all the Neogene gravels of the Miocene–Pliocene sediments of the Orawa–Nowy Targ Basin are devoid of clasts derived from the Tatra Mts where no prominent relief existed at that time. To the contrary, the Pieniny Klippen Belt was subject to denudation and had more considerable relief then now. The Late Miocene relatively quiet tectonic period and low relief in the Tatra Mts is documented by the oldest fine-grained cave sediments in the Belianska cave, dated to ~6.15–4.18 Ma (Bella et al. 2011). Similar results were obtained from speleothem dating in the Polish Tatra caves, indicating their oldest denudation surface as latest Miocene or younger in age (Głazek 1996). The upper limit of denudation surface formation (> 1.2 Ma) was gained by study of the evolution of the phreatic stage in the Tatra caves (Gradziński et al. 2009).

### *The neotectonic stage (~7–0 Ma)*

The impressive morphology of the Tatra relief points to a new acceleration in tectonic activity during the latest Late Miocene to Pleistocene. Recent research on the sedimentological record from the Orava–Nowy Targ Intramontane Basin supports this view. The research showed a considerable difference between the Neogene (fine-grained) and the Quaternary (coarse-grained) deposits (Tokarski et al. 2012). Neotectonic fault activity is also proved by significant evolution of calcareous tufa and travertine mounds in the Polish (e.g. Mastella & Rybak-Ostrowska 2012) and also in the Slovak (e.g. Sůkalová et al. 2012) parts of the Tatra Mts area. The youngest story was indirectly documented by granite boulders of Middle Pleistocene moraine located at the foot of the High Tatra Mts. The granite boulders revealed significantly older AFT age (~$11.0 \pm 1.4$ Ma, VTM sample; Table 3) than the age of deposition. This AFT age refers to the simple moderate cooling event of Slavkovský štít (peak) during the latest Middle to early Late Miocene, documented by the same age and track length measurement records as in the VT-2 to VT-4 samples. In addition, the lag time between the AFT age ($11.0 \pm 1.4$ Ma) inherited from the Tatra massif and the time of deposition of the VTM sample (Middle Pleistocene) indicates an acceleration of exhumation rate during this time span. This premise indicates that an accelerated exhumation rate of the Tatra massif must have occurred ~6.5–1.0 Ma. Beside this, slip rate analysis of the Vikartovce fault which is parallel to the sub-Tatra fault close to the Tatra Mts provided dip-slip movement along the fault ~1 mm/a during the Late Pleistocene (Vojtko et al. 2011).

Holocene (postglacial) denudation of the High Tatra Mts ridges was estimated from the Holocene sedimentary infill volume of 17 valleys. Statistically consistent data lead to a mean lowering of the ridges by ~5 m for the Studený potok valley (Lukniš 1968). It is equivalent to a mean denudation rate of ~0.5 mm/a. From the denudation/accumulation volume ratio (0.56 — see chapter 4.3) of the Studený Potok paragenetic system and considering that a significant part of the removed rocks is not reflected in the volume of accumulation storage space, an estimate of the Holocene denudation rate of ~1 mm/a can be assumed.

U-series dating of cave speleothems (Gradziński et al. 2009) points to a mean rate of valley deepening of 0.2–0.3 mm/a during the last 200 kyr in the northern part of the Tatra Mts. However, this significant discordance can be eliminated by several factors: (i) N–S asymmetry of the Tatra Mts uplift could lead to smaller deepening of the valleys located on the north side; (ii) a fundamental part of the denuded ridge material remained in the valleys during the interglacial times, but was later removed by glaciers during the next glacial); (iii) the intensity of interglacial ridge denudation was increased by the effect of rapid denudation in the paraglacial stage; (iv) more intensive widening of valleys in comparison with their deepening in the last 200 kyr could play a significant role. On the basis of the aforementioned data, a minimum denudation rate ~0.5 mm/a is considered to be most presumable since the Middle Pleistocene.

## Conclusions

The Tatra Mts and their surroundings underwent a complex Alpine tectonic evolution, which can be divided into several tectonic stages, based on structural, sedimentary, geomorphic, and FT data:

♦ A first tectonic stage (*TS-1; ~95–80 Ma*) can be dated back to the late Early Cretaceous and is coeval with nappe stacking (Fig. 7). At this time, the Tatric crystalline basement was buried below the paleo-Alpine — Fatric and Hronic units (Fig. 8b). The fully annealed zircon samples suggest that the metamorphic temperature was in excess of ~320 °C (Fig. 7);

♦ The principal compressional stage of the Alpine orogene was replaced by the Late Cretaceous to Paleogene orogene collapse followed by an orogen-parallel extension (Fig. 8c), revealed by structural and ZFT data of ~60–70 Ma (Figs. 2 and 7). The extensional tectonics were replaced by transpression to transtension (Fig. 3) during the Late Paleocene to Eocene period (*TS-2; ~80–45 Ma*);

♦ The Late Eocene to Earliest Miocene can be characterized by formation of a marginal sea of the Peri-Tethyan Basin represented by the CCPB (Fig. 8d). It shows a fore-arc basin position developed on the destructive plate-margin and behind the EWC accretionary wedge. During the sedimentation of the CCPB sequences, the crystalline basement of the Tatric Unit was buried in between the APAZ and ZPAZ (*TS-3; ~45–20 Ma*; Fig. 7);

♦ The final cooling of the Tatra massif was a result of asymmetric neotectonic exhumation that could be linked with the sub-Tatra faulting in the southern edge of the massif during the Neogene (*TS-4; ~20–7 Ma*; Figs. 6, 7 and 8e). The exhumation processes of the Tatric crystalline basement from the depth of ~5 km were dated to the Middle/early Late Miocene, according to AFT and AHe data (Burchart 1972; Kráľ 1977; Struzik et al. 2002; Anczkiewicz 2005; Anczkiewicz et al. 2005; Śmigielski et al. 2012; and data herein). Additionally, the early Late Miocene evolution was characterized by formation of the basic Western Carpathian planation surface — intramountain level;

♦ It can be stated that the final appearance of the mountains range in morphology above the surrounding foreland was a consequence of a new acceleration in tectonic activity during the latest Late Miocene to Pleistocene (*neotectonic stage — TS-5; ~7–0 Ma*; Figs. 7 and 8f). On the basis of the geomorphological data, a denudation rate of ~1 mm can be considered for the whole Quaternary, whereas a minimum denudation rate of ~0.5 mm/a can be assumed since the Middle Pleistocene (Table 2).

**Acknowledgment:** This publication is the result of the project implementation: Comenius University in the Bratislava Science Park supported by the Research and Development Operational Programme funded by the ERDF Grant No.: ITMS 26240220086. The work was financially supported by the Slovak Research and Development Agency under the contracts Nos. APVV-0625-11, APVV-0099-11, APVV-0315-12, APVV ESFEC-0006-07, by the Grant Agency of the Czech Republic No. GAČR:13-15123S and by the VEGA agency under contracts No. 1/0193/13. B. Fügenschuh acknowledges financial support by the Austrian Science Fund (FWF) Grant No. I 138-N19. Thanks are due to reviewers for their accurate and constructive reviews.

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