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EARLY- vs. LATE OROGENIC GRANITOIDS RELATIONSHIPS IN THE VARISCAN BASEMENT OF THE WESTERN CARPATHIANS

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Abstract: The Variscan high-grade crystalline basement of the Western Carpathians contains granitic to granodioritic bodies transformed to various degrees into orthogneisses. The orthogneisses resemble the structures of the host regional-metamorphic rocks, indicating their syn-collisional evolution, predating the intrusion of the granitoids around 360–340 Ma. The Nízke Tatry Mountains orthogneiss, dated at ~381 Ma (NTJ-1 sample), emplaced during early partial melting stage deformation regime. Syn-tectonic magmatic flow and emplacement is suggested by the relics of magmatic foliations and lineations. Post-intrusion evolution comprises metamorphic/ductile medium-temperature fabrics of quartz, feldspars and micas, as well as metamorphic garnet growth. The granite-tonalite plutons emplaced between ~353 Ma (Malá Fatra Mountains) and ~343 Ma (Nízke Tatry Mountains, DUM-1 sample) during transtensive (?) deformation of the host metamorphic complex, show distinct magmatic fabric anisotropy, especially in their marginal parts, coeval with ductile fabrics of the orthogneisses. Petrologically the orthogneisses point to S-type granite or granodiorite, while the tonalites of large granitoid plutons seem to be of igneous I-type origin.

Key words: Variscan, Western Carpathians, U-Pb zircon age, orthogneiss, granitoid, microfabric.

Introduction

Temporal and spatial links between granite emplacement and active deformation sites wthin crustal-scale shear zones, have been demonstrated by many authors (e.g., Guineberteau et al. 1987; Paterson et al. 1989; Paterson & Tobisch 1992; D'Lemos et al. 1992, 1997; Miller & Paterson 1994; Bouchez & Gleizes 1995; Gleizes et al. 1997; Schofield & D'Lemos 1998; Brown & Solar 1998; Yenes et al. 1999). These authors found that syn-tectonic granites often record a whole range of thermal conditions during straining, from magmatic flow with primary magmatic crystallization texture (Bouchez et al. 1981; Bryon et al. 1994; Yenes et al. 1999), sub-magmatic or "sub-solidus" flow, where enough melt remains allowing limited crystal-plastic slip (Bouchez et al. 1992; Miller & Paterson 1994), to solidus deformations where crystal plastic deformation dominates (Gapais 1989; Paterson et al. 1989).

Orthogneisses from the supracrustal Jarabá Complex (Maheľ et al. 1968) of the Tatric basement (Central Western Carpathians, Figs. 1, 2) yield U-Pb zircon ages spanning from ~410 to 380 Ma. They include K-feldspar and plagioclase-bearing orthogneisses from the northern Tribeč Mountains (~410 Ma; Krist et al. 1992), Nízke Tatry Mountains (~390 Ma; Adamija et al. 1992) and from the Western Tatra Mountains (~405 Ma; Poller et al. 2000). Trondhjemitic orthogneisses from the leptynite-amphibolite complex (LAC; Hovorka & Méres 1993) (Fig. 2) of the Veporic basement (Central Western Carpathians) yield U-Pb zircon ages ~500 and ~350 Ma (the upper- and lower-intercepts of the discordia with the concordia; Putiš et al. 2001). The oldest meta-magmatics so far dated in the Western Carpathians include the Muráň granitic orthogneisses (~500 Ma upper intercept; Putiš et al. in prep.) and metagranitoids (474±14 Ma) from the north Veporic Unit dated by monazite chemical dating (Janák et al. 2002), both indicating a Late Cambrian–Ordovician magmatic event. They contrast with some dioritic orthogneisses, dated at ~346 Ma, which are interpreted as a remelting product of the LAC (Putiš et al. 1996, 1997, 2001; Filová et al. 2002). The main phase of (meso-Variscan) granitoid plutonism of the Tatric basement occurred at ~360–340 Ma (Petrík et al. 1994; Petrík & Kohút 1997).

In this paper, we present structural data from orthogneisses of the Tatric crystalline basement in the area of the Nízke Tatry and Malá Fatra Mountains (Figs. 1, 2), in relation to a large-scale thrust-fault system (Fig. 2), which affects stacked Variscan basement nappes (Putiš 1992, 1994). Our microfabric study examines whether an interaction exists between the gneissified granites to granodiorites (orthogneisses) and host regional-metamorphic rocks, as well as the emplacement of the younger plutonic granitoids.

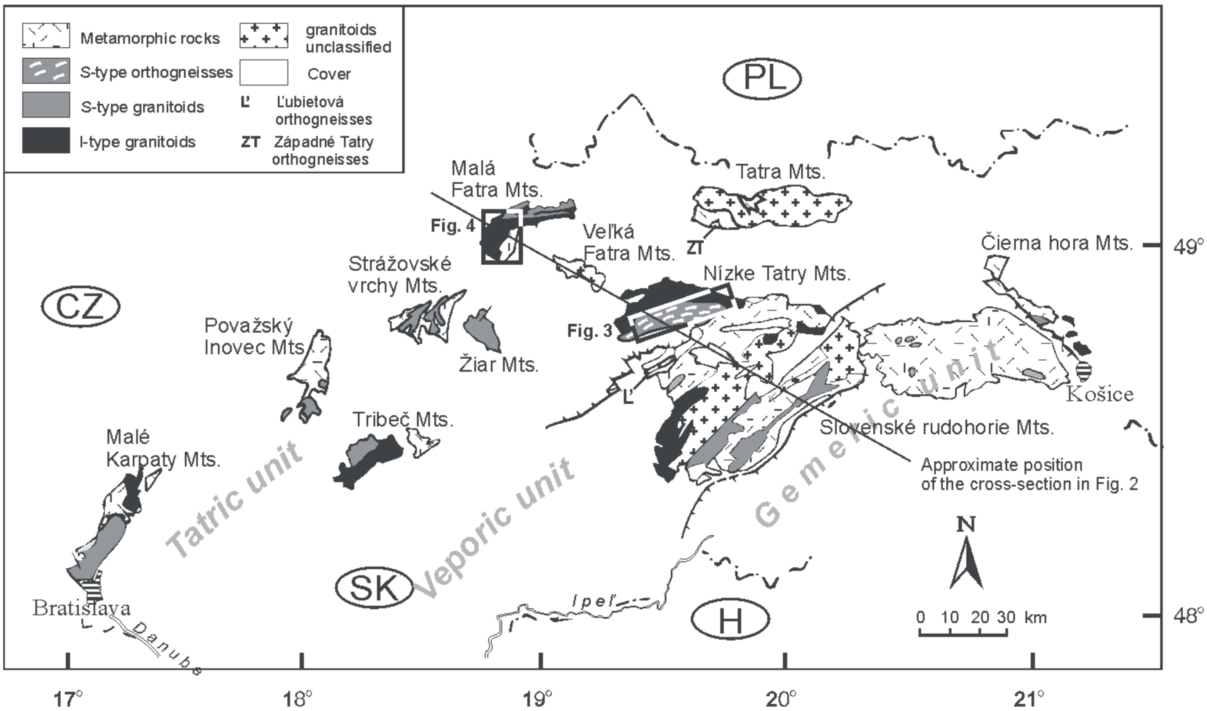

Fig. 1. Sketch-map of distribution of the Variscan West-Carpathian granitoid complexes. The study areas are indicated by frames.

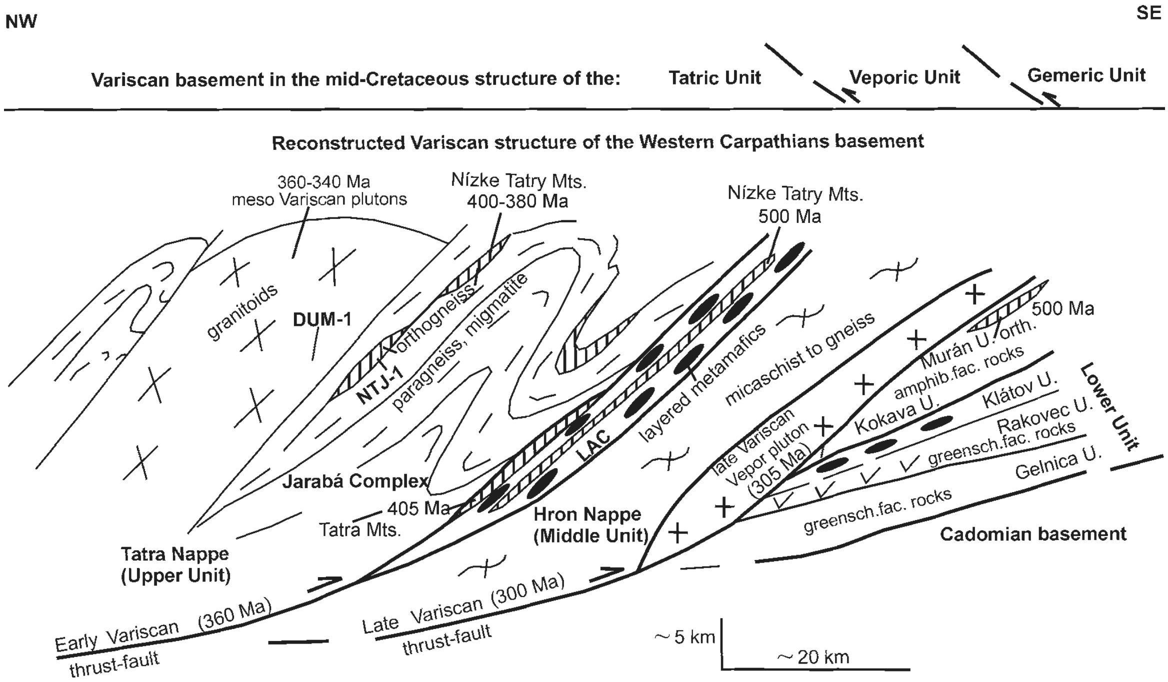

Fig. 2. Tectonic sketch-profile (from the Malá Fatra to Slovenské rudohorie Mts, see Fig. 1) of the Variscan structure of the Western Carpathians basement complexes, with location of dated orthogneiss (**NTJ-1**) and tonalite (**DUM-1**) samples. Geographical location of dated samples: NTJ-1 orthogneiss — taken from large outcrop in the Jasenie Valley, Struhár area, about 30 m east of the small dam; DUM-1 tonalite — taken from large outcrop along the main road above the Nižná Boca village, north of the Čertovica pass.

Structural setting and microfabrics of the orthogneisses and the granitoids

The Variscan high-grade crystalline basement outcrops in the mid-Cretaceous Tatric and Veporic Units (Fig. 1). The whole series includes, in addition to the basement and its Upper Carboniferous to mid-Cretaceous sedimentary cover, Mesozoic nappes and Paleogene sediments. These form the so-called core-mountains in the Tatric Zone, which represent Neogene–Miocene mega-horsts separated by graben sediments (Plašienka et al. 1997).

The Upper Variscan Tatra Nappe, a high-grade supracrustal metamorphic complex, is made of paragneisses, migmatites, amphibolites of the supra-crustal Jarabá Complex and granitoid plutons. Its lower-crustal sole is predominantly composed of metamafic rocks: amphibolites, inferred as retrograde eclogites (Janák et al. 1996), serpentinites and metagabbros of the LAC. This nappe overlies a medium-grade supracrustal complex (Fig. 2) devoid of granitoids and made of micaschists, gneisses and sporadic amphibolites, called the Middle Variscan Hron Nappe (Putiš 1992; Plašienka et al. 1997 — Fig. 6).

The Nízke Tatry orthogneiss

The K-feldspar orthogneiss (Koutek 1931; Zoubek 1951; Biely et al. 1992; Adamija et al. 1992; Putiš & Madarás 1993; Petrík et al. 1998; Madarás et al. 1999 — Fig. 1) forms several kilometres long and hundreds of metres wide lense-shaped bodies enclosed within the paragneiss-migmatites near the base of the Jarabá structural complex.

The orthogneisses are proto- to ultramylonitic (Fig. 3a). The protomylonitic type shows locally well preserved magmatic fabrics which are defined by oriented magmatic minerals (feldspars, micas) forming domains almost free from a ductile overprint (Fig. 3b). Originally, the rock was a synkinematic porphyric granite or granodiorite showing aligned prismatic K-feldspars and plagioclase crystals, surrounded by oriented muscovite and biotite flaky crystals, giving evidence of *syntectonic magmatic flow* (e.g., Tullis & Yund 1985; Gapais & Barbarin 1986; Aranguren & Tubía 1992) acquired at hypersolidus to solidus temperatures (~700–600 °C). In other domains, larger quartz crystals are weakly elongate, parallel-to-foliation ribbons formed during a *subsolidus deformation* coeval with the high-T deformation below granite solidus. The remaining quartz grains keep their interstitial location. At such subsolidus temperatures, around 600 °C, abundant strain-induced myrmekites (Simpson 1985) are developed parallel to the foliation along the K-feldspars megacrysts (Fig. 3b). Large plastically strained domains have macroscopically recognizable σ- and δ-type rotated feldspar porphyroclasts enclosed within dynamically recrystallized quartz-feldspar-mica mylonitic matrix, where asymmetric mica-fish result from basal-slip in biotite (Fig. 3c). Such microstructures may have formed at temperatures of 600–450 °C. Ductile flow at still lower temperatures (~300 °C) is marked by dynamic recrystallization of quartz ribbons into polygonal aggregates of equant grains (Fig. 3d). U-stage textural measurements of quartz in the ribbons suggest dominant prism <a> and basal <a> slips (Fig. 6a–d).

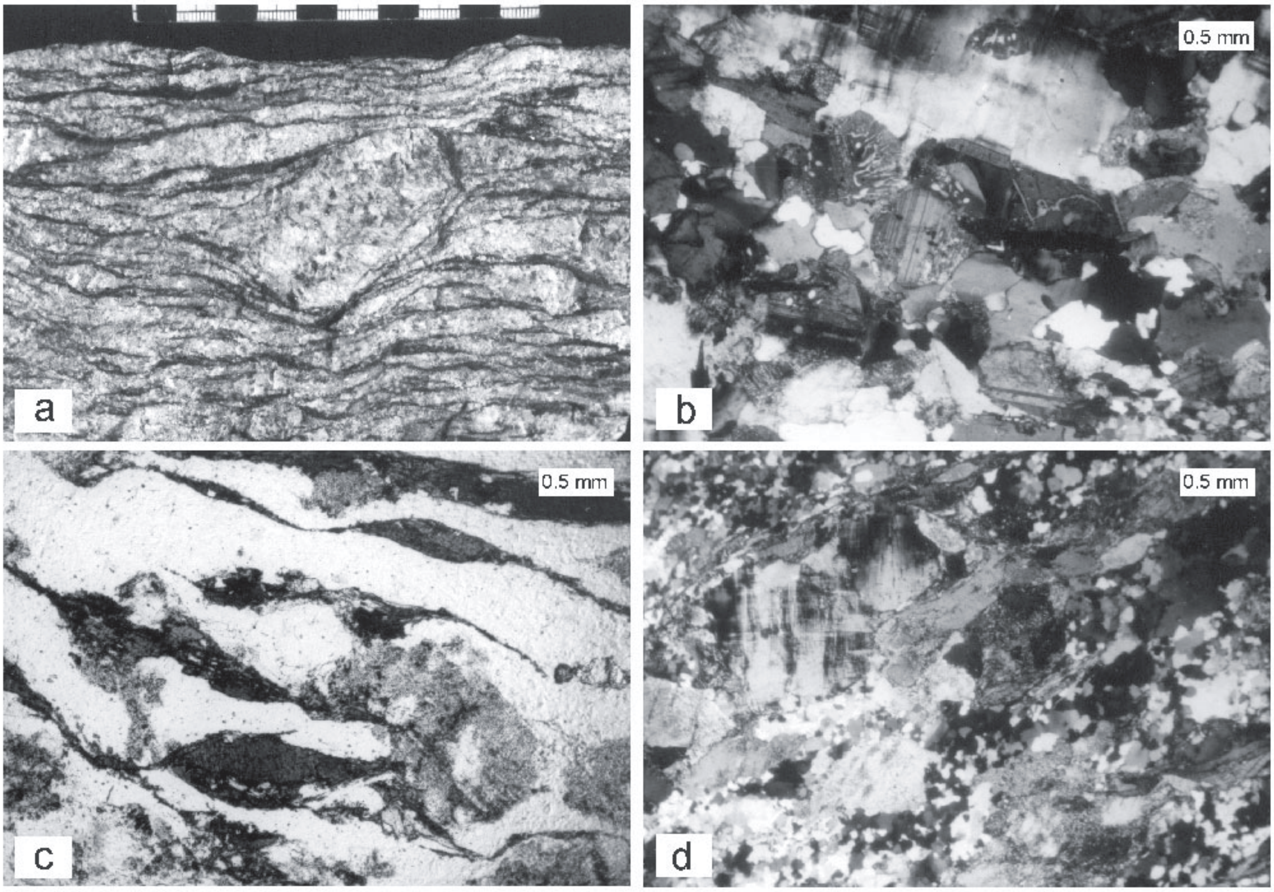

Fig. 3. Meso- and microfabrics of the Nízke Tatry orthogneiss (Jasenie valley-type, NTJ-1). **a** — typical augen-banded structure: 3 cm large K-feldspar porphyroclast with tails of plastically deformed and dynamically recrystallized margins within bands of stretched or dynamically recrystallized feldspar grains; **b** — parallel oriented growth of feldspars with subsolidus formation of strain myrmekite along the K-feldspar margins; **c** — biotite mica-fish undergoing slip along cleavage; **d** — low-temperature dynamic rotation recrystallization of quartz among undulose strained feldspar porphyroclasts. Scale bar: 1 cm (a), 0.5 mm (b–d).

Because the magmatic foliations and lineations are observed to be parallel to the late-metamorphic planar and linear fabrics (Fig. 7a–d) the deformation history recorded by the magmatic rocks is considered to be similar to that recorded by the host metamorphic rocks. Shape asymmetries around porphyroclasts point to a top-to-the-SE ductile thrusting of the meso-Variscan Upper Tatra Nappe over the Middle Hron Nappe, in present-day geographical coordinates.

The Malá Fatra orthogneiss

Orthogneisses in the Malá Fatra Mts (Figs. 4, 5a) were mentioned by Kamenický & Macek (1984) and Lupták (1996). They represent tens of metres thick, steeply inclined and kilometer-long lenses within the gneiss-migmatite-amphibolite metamorphic complex.

The highest-temperature *subsolidus deformation* features include narrow stripes of exsolved perthite followed by microclinization forming wide hatches organized into deformation bands, and twinned plagioclase crystals (Fig. 6g–h) due to plastic-slip (Jensen & Starkey 1985; Ji & Mainprice 1990). Large plastic deformation at high-temperature (>500 °C) produced quartz layers (Wilson 1975; Culshaw & Fyson 1984) that are mechanically separated from feldspar layers (Fig. 5b). These monomineralic ribbons probably formed by a dynamic migration recrystallization (Hirth & Tullis 1992). Their lattice fabric (Fig. 6e–f), measured using the U-stage, however shows a distinct (Y) maximum of c-axes, indicating that the prism <a> slip occurred (Hobbs 1981; Takeshita & Wenk 1988). Newly formed internal sub-ribbons, parallel to older ones, seem to result from dislocation flow and recovery process. A few ribbons, showing basal subboundaries indicative of a prism <c> slip, may have formed at higher subsolidus temperatures (~600 °C) (Blumenfeld et al. 1986; Gapais & Barbarin 1986; Mainprice et al. 1986).

The feldspar layers consist of dynamically recrystallized equant plagioclase and K-feldspar grains with slightly arcuate to straight boundaries meeting at ~120° triple junctions (Fig. 5c), suggesting a late kinematic annealing. Some lobate or sutured grain boundaries may indicate that the strain-induced grain boundary bulging, followed by subgrain rotation, has not been completely annealed. The presence of plagioclases either recrystallized at their periphery, or entirely dynamically recrystallized (Fig. 5b), suggests that a dislocation creep was an active deformation mechanism (Tullis & Yund 1985; Hacker & Christie 1990), and that subgrain rotation was a principal recrystallization process (Poirier & Guillopé 1978; Jensen & Starkey 1985; Ji & Mainprice 1990; Trimby et al. 1998). That the dynamic recrystallization (Urai et al. 1986; Drury & Urai 1990) took place at a high temperature is also documented by the higher albite content of the newly-formed K-feldspars (ca. 5–8 %) and by the growth of new garnet grains. Many of the plagioclase grains have their twins subparallel to the foliation plane. The (010) or (001) twins were measured with the U-stage, including small recrystallized plagioclase grains that originated by the dynamic recrystallization of original plagioclases in the mylonitic bands. Their $N_{Y(\beta)}$ (Fig. 6g) and $N_{Z(\gamma)}$ (Fig. 6h) optical directions point to a

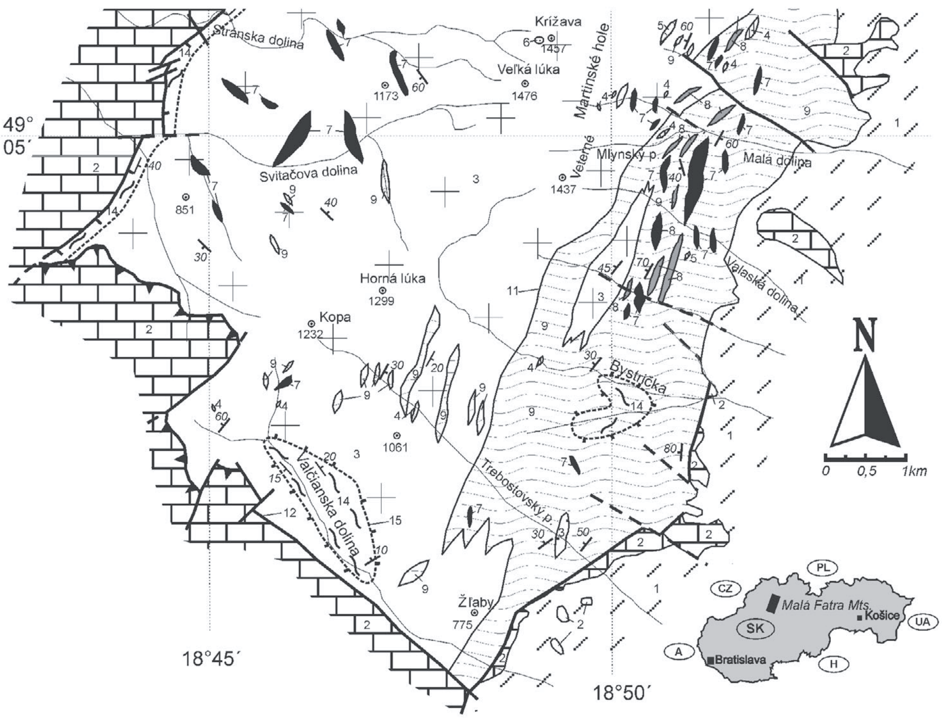

Fig. 4. Geological map of southern part of the Lúčanská Malá Fatra Mts (Rakús et al. 1988, modified by the authors). **1** — Quaternary and Neogene sediments; **2** — Mesozoic cover and nappe complexes; **3** — granodiorite to tonalite; **4** — pegmatite and aplite veins; **5** — lamprophyres; **6, 7** — amphibolite; **8** — orthogneiss; **9, 10** — paragneiss to migmatite; **11** — primary geological boundary; **12** — fault: observed; assumed; **13** — thrust plane of Mesozoic nappes; **14** — Alpine blastomylonites at the base of allochthonous crystalline complex; **15** — thrust plane of crystalline complex.

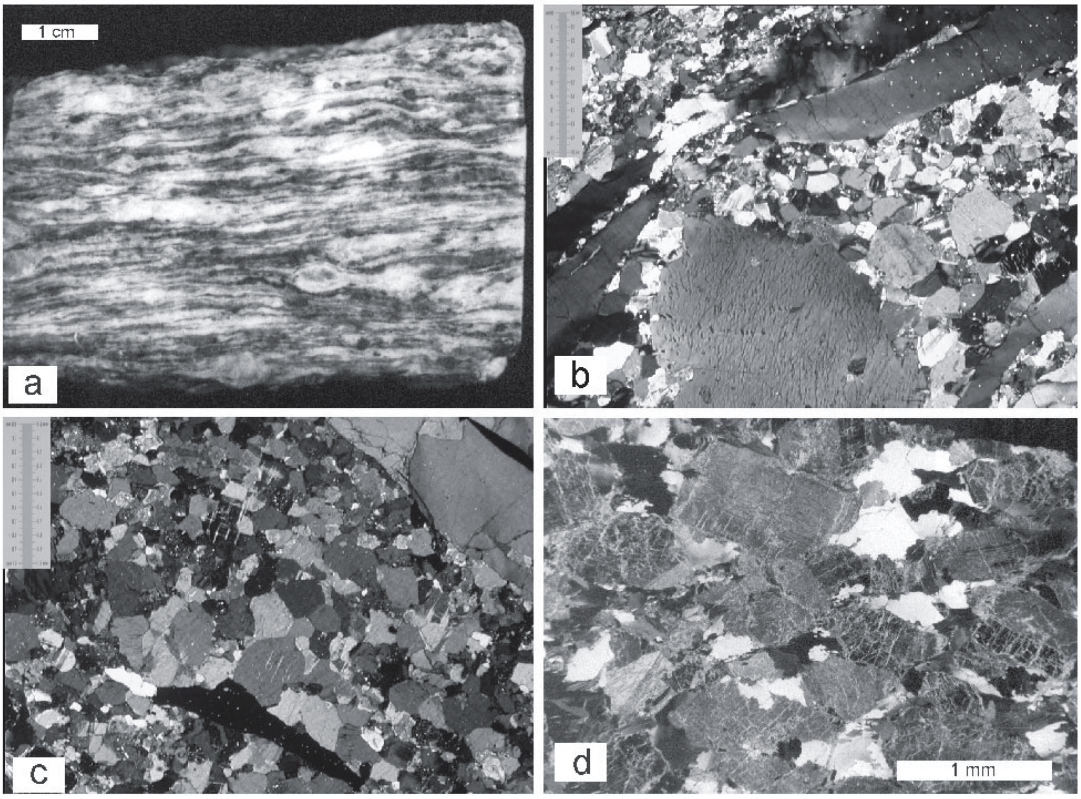

Fig. 5. Meso- and microfabrics of the Malá Fatra Mts orthogneiss (a–c) and the Ďumbier tonalite (d, DUM-1). **a** — mylonitic orthogneiss fabrics in XZ section. **b** — quartz ribbons alternating with layers of dynamically recrystallized feldspar grains and preserved feldspar core-mantle structures. **c** — recrystallized feldspars with their characteristic grain-boundary tripple junctions. **d** — magmatic foliation in Ďumbier tonalite formed by lath-shaped cumulated plagioclase, biotite and large quartz grains. Scale bar = 1 cm (a), 1 mm (b–d).

slip on {010} planes close to the foliation, in the direction $N_{Y(\beta)}$ = [001] subparallel to the lineation. This slip system also characterizes high-grade mylonites (Olsen & Kohlstedt 1985; Egydio-Silva & Mainprice 1999). The porphyroclasts of K-feldspars and plagioclases in this mylonitized granitoid were rotated during their dynamic recrystallization. New recrystallized grains, formed at the periphery of the porphyroclasts, constitute σ- or δ-type features tailing the porphyroclasts and pointing to a top-to-the-NW sense of shear. In host gneiss-migmatite rocks, characterized by a sillimanite lineation (Fig. 6j), similar microstructures and senses of shear are also observed, for example, asymmetric biotite-enriched pressure shadows tailing garnets and plagioclase porphyroblasts, asymmetric girdle patterns of biotite (Fig. 6i), and quartz (Fig. 6k) c-axes. The metamorphic, mylonitic and magmatic foliations (Fig. 7e–g), showing parallelism, reflect a complex evolutionary history similar to that described in the previous paragraph.

The Nízke Tatry granitoids

The granitoid cores of the Nízke Tatry and Malá Fatra Mountains (Fig. 1) represent late-orogenic plutonic slices emplaced within the metamorphic sequence (Fig. 2). They are structurally more or less parallel to their host metamorphic rocks having an attitude suggesting that a transpression-transtension regime was acting during their emplacement. Their contact with the metamorphic pile is often sharp, but in places it is represented by nebulitic migmatites grading into inhomogeneous, schlieren-bearing granites. Close to contacts, the strong magmatic fabrics (Siegl 1970, 1976), defined by lath-shaped feldspars and flaky micas (Fig. 5d) is observable. At some contacts, metre-scale xenoliths of para- and orthogneisses indicate that "magmatic stoping" acted along the walls of these plutonic slices, providing the space for these magmas to emplace within the Jarabá Complex of the Upper Tatra Nappe. Although no substancial superimposed ductile fabrics have been observed in these plutonic bodies, the magmatic structure is concluded to be coeval with the dominating mylonitic foliations of the hosting para- and orthogneisses (Fig. 7a–d).

The Malá Fatra granitoids

Zircons from plutonic granitoids show the age ~353 Ma (U-Pb, Shcherbak et al. 1990). We have studied the margins of the Malá Fatra (Veľká Lúka) pluton (Fig. 4) which display well-defined magmatic foliations defined by the preferred orientations of micas and feldspars (Fig. 7g). Their orientations are very constant with ~ENE-WSW strikes and steep dips towards the NNW, conformably to the host metamorphic rocks. Two differently oriented magmatic lineations are recorded. Some of them have direction of dip 325°–330° and plunge 70°–75° onto 335–350/70–80 foliation planes. These lineations probably represent the subvertical magmatic emplacement of the plutonic body. Most of the lineations, however,

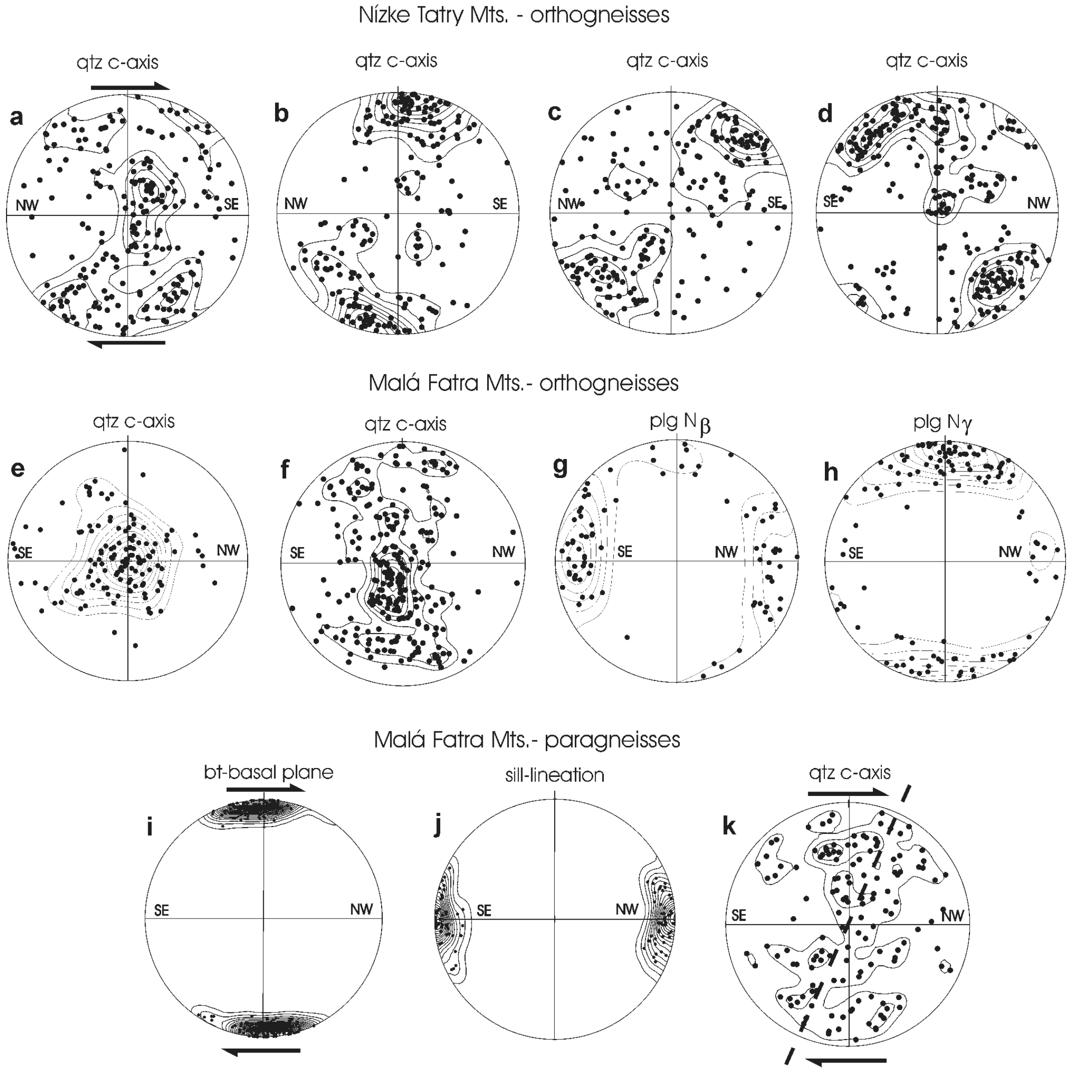

Fig. 6. Microtextural patterns of quartz, plagioclase, biotite and sillimanite. (a–d) quartz ribbon c-axis patterns (Nízke Tatry Mts, NT, orthogneiss): **a** — density levels of contours (d.l.c.): 8–6–5–3.5 [n = 215]; **b** — d.l.c.: 16–14–13–11.5–10–9–7–6–4.5 [n = 198]; **c** — d.l.c.: 9–8–6.5–5–4 [n = 372]; **d** — d.l.c.: 10–8.5–7–6–4.5 [n = 200]. (e–f) quartz ribbon c-axis patterns (Malá Fatra Mts., MF, orthogneiss): **e** — d.l.c.: 20–18.5–17–15.5–13–11.5–10–8.5–6–4.5 [n = 149]; **f** — d.l.c.: 24–20.5–17–13.5–10–6.5-3 [n = 191]. **g** — plagioclase N_β optical direction pattern (MF orthogneiss) d.l.c.: 11–10–8.5–7–6–4.5 [n = 83]. **h** — plagioclase N_γ optical direction pattern (MF orthogneiss) d.l.c.: 15–14–12.5–11–10–8.5–7–5.5 [n = 115]. **i** — biotite basal plane pattern (MF paragneiss) [n = 150], maximum = 23.2. **j** — sillimanite lineation pattern (MF paragneiss) [n = 80], maximum = 16.3. **k** — quartz c-axis patterns (MF paragneiss) 7.5–6–4 [n = 150].

are oriented at 240–260/15–30 onto 320–350/50–75 foliation planes. These orientations are ascribed to the ultimate stages of magma emplacement within a lateral strike-slip shear zone, affecting the surrounding metamorphic rooms with the ductilely deformed orthogneisses. Top-to the NW shear in ortogneisses indicates transtensional (?) deformation zone.

Petrological and geochemical comparison of the orthogneisses and granitoids in the Nízke Tatry Mountains

Because comparable data from the orthogneisses and granitoids of the Malá Fatra Mountains are not available, this paragraph concerns only the Nízke Tatry Mountains.

The Nízke Tatry orthogneisses

These orthogneisses usually have the composition of two mica (S-type) granites or granodiorites. Dominant minerals are subhedral An_{25-30} plagioclases, biotite and muscovite. Sub- to euhedral variably sericitized K-feldspars usually form large megacrysts 1–10 cm in size but also occur as an interstitial phase. In the mylonitic facies, the original magmatic phenocrysts are overgrown by albite, quartz and secondary K-feldspar to form augens. Petrík et al. (1998) showed that the K-feldspar megacrysts have a bell-shaped distribution of Ba concentrations that reach up to 4400 ppm in crystal centres. Such distributions were successfully modelled by *in situ* fractional crystallization, which supports the magmatic origin of these K-feldspar phenocrysts. Ba distributions also show that

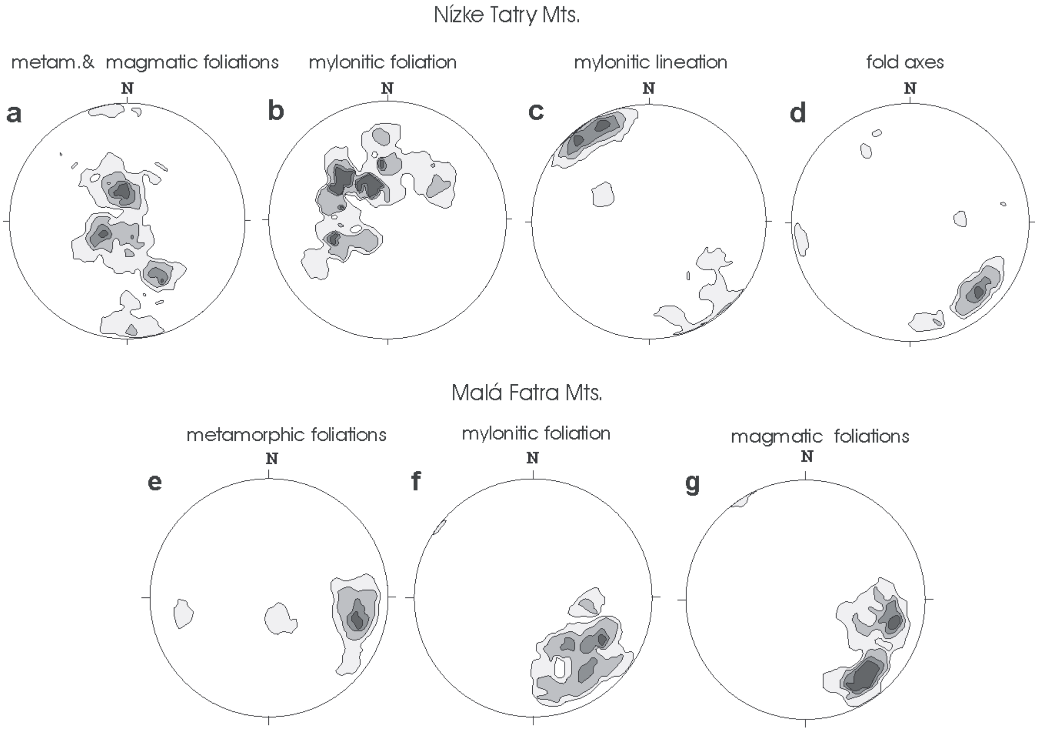

Fig. 7. Orientation diagrams of mesostructures from the Nízke Tatry (a–d) and Malá Fatra (e–g) Mts. (a–b) and (e–g) contour pole diagrams (planar data); (c–d) contour orientation diagrams (linear data). **a** — early-Variscan syn-tectonic metamorphic and/or magmatic foliations (69 data from orthogneisses and metamorphic rocks), density levels of contours (d.l.c.): 20–14.5–10–5. **b** — meso-Variscan mylonitic planes (27 data from orthogneisses), d.l.c.: 19–14–9.5–5. **c** — meso-Variscan mylonitic lineations (61 data from orthogneisses), d.l.c.: 26–19–13–6.5. **d** — metamorphic/ductile mesofold axes (35 data), d.l.c.: 19–14.5–10–5. **e** — early-Variscan metamorphic foliations (59 data from metamorphic mantle rocks), d.l.c.: 19–14–9.5–5. **f** — meso-Variscan mylonitic foliations (36 data from orthogneisses), d.l.c.: 17–13–8.5–4. **g** — Variscan foliation planes of xenolithic bodies and magmatic foliations in marginal domain of the granitoid pluton (17 data), d.l.c.: 25–19–12.5–6.

the megacrysts have formed from several individual cores, each having its own Ba profile. Biotite, that defines the foliation of orthogneisses, is dark-brown to pale yellow, fresh or variably chloritized. It is relatively rich in iron [Fe/(Fe+Mg) = 0.6] and reduced (Fe^{3+}/Fe_{tot} = 0.03), a feature which is typical of other S-type granitoids in the Western Carpathians (Petrík et al. 1994). Muscovite may be abundant, up to 10 vol. %, and in some mylonitic varieties, it encloses sillimanite indicating that former products of muscovite dehydration melting were subsequently re-hydrated. Accessory minerals are represented by abundant euhedral apatite, zircon and monazite, which are mainly enclosed in biotite. Ore minerals are represented by common pyrrhotite.

The orthogneisses are moderately acidic, peraluminous rocks, varying in composition from granodiorite (augen varieties) to tonalite (banded varieties) with SiO_2 varying between 67–76 wt. % and alumina saturation index (ASI) varying between 1.04 and 1.3. According to the limited trace element data set (Petrík et al. 1998), they have moderate contents in Rb (100–175 ppm), Ba (1100–316 ppm), Sr (234–120 ppm) and Y (11–28 ppm). Their REE concentrations are moderate to low, weakly fractionated, with distinct negative Eu anomaly and low La/Yb ratio (Fig. 9). These REE patterns are consistent with those from other Tatric orthogneissic cores, such as the Tribeč, Považský Inovec and Western Tatra Mountains (Méres & Hovorka 1992; Petrík 2001; Janák et al. 2001) and indicate a sedimentary protolith. In conclusion, major and trace elements favour an S-type granite precursor for these orthogneisses (Petrík et al. 1998).

The Nízke Tatry Ďumbier, Prašivá and Králička granitoids

Defined by Koutek (1931) as petrographically contrasted rocks, the dated (DUM-1) *Ďumbier granodiorites–tonalites and Prašivá granites* represent classical Variscan granitoid types of the Western Carpathians. The Ďumbier and Prašivá granitoids have been dated by the Rb/Sr method (Bagdasaryan et al. 1985) yielding a poorly constrained age at 362±21 Ma (I_{Sr} = 0.7079±0.0002). The silica content of the Ďumbier and Prašivá granitoids shows an unusually wide range from 60 to 72 %. The “mafic” subtypes, with 60–65 % SiO_2 contain 20–25 vol. % of biotite (Cesnak 1985) and are meta- to subaluminous. The more acidic subtypes have a peraluminous nature (A/CNK = 1–1.3) due to common sericite and late muscovite. The Prašivá type shows consistently higher K_2O contents 3–4 %, compared to 2–3 % in the Ďumbier type. Similarly, both types have common trace element trends: Ba fractionates from 2000 down to 300 ppm, Sr from 960 to 200 ppm, Zr from 250 to 50 ppm, values within the ranges of other Carboniferous I- and S-type granitoids (Petrík et al. 1994). In contrast with orthogneisses, rare earth elements show steep normalized patterns and no Eu anomaly (Petrík et al. 1994; Broska & Uher

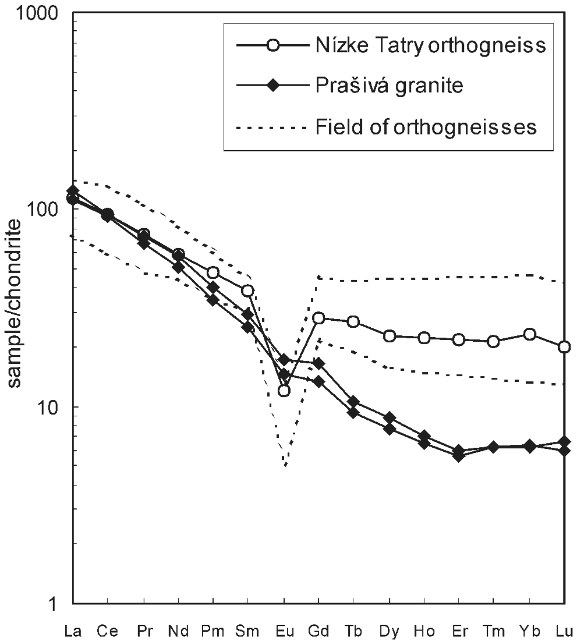

Fig. 8. Rare earth element patterns of orthogneisses and granites. The field of the West-Carpathian orthogneisses is based on the data of Méres & Hovorka (1993), Janák et al. (2001).

2001). They are apparently controlled by allanite or monazite (Fig. 8). The above-mentioned relatively high initial Sr ratio $I_{Sr} = {}^{87}Sr/{}^{86}Sr_{(350)} = 0.7079$, not consistent with the otherwise I-type petrographical character is noteworthy. This ratio may result from the mixing of a mafic (dioritic) and a mature supracrustal end-member (Petrík 2001). The latter end-member may be represented by another characteristic Nízke Tatry granite, the peraluminous *Králička* type characterized by high ratios of ${}^{87}Sr/{}^{86}Sr_{(365)} = 0.71601$ and ${}^{144}Nd/{}^{143}Nd_{(350)} = 0.511834$. The Rb/Sr age is 365±40 (Rb/Sr, Bagdasaryan et al. 1985; 2σ recalculated by Petrík 2000). The Králička granite occurs only within orthogneisses. The strontium isotopic composition, one of the highest among the Tatric granitoids, strongly indicates a mature, crustal protolith consistent with the S-type nature of the Nízke Tatry orthogneiss. The Králička granite also shows $\varepsilon Nd_{(350)} = -6.89$ (Kohút et al. 1999), the most negative value among all analysed West-Carpathian granitoids. The composition of the Králička granite was used by Petrík (2000) as a supracrustal felsic end-member in a mixing model of the West-Carpathian granitoids.

Geochronology

Isotopic-geochronology was performed at the Institute of Precambrian Geology and Geochronology (Russian Academy of Sciences, St. Petersburg) on a Finnigan MAT 261 8-collector mass-spectrometer in static mode. Zircons were extracted from crushed rock samples with heavy liquid and magnetic separation techniques. Hand-picked aliquots of zircon were analysed following the method of Krogh (1973). The total blanks were 0.05–0.1 ng Pb and 0.005 ng U. An air-abrasion treatment of the zircon was performed by the Krogh's (1982) technique. The PbDat and ISOPLOT programs by Ludwig (1991a,b) were used for uncertainties and correlations of U/Pb. On the basis of reproducibility of standard zircon analyses the uncertainties in U/Pb is defined at 0.5 %. Ages were determined using the decay constants given by Steiger & Jäger (1977). All errors are reported at the 2σ level. Corrections for common Pb were made using values of Stacey & Kramers (1975).

The Nízke Tatry NTJ-1 orthogneiss

The zircon population from sample NTJ-1 consists of idiomorphic and subhedral translucent, transparent, rarely nebulous prismatic and pyramidal lilac-brown crystals having regular magmatic zonation in cathodo-luminescence (CL). However some zircons contain cores or relicts of metamict cores traced by numerous tiny opaque inclusions, visible in CL images (Fig. 9a,b). The zircons have a length/width ratio of 1.5–3.0 and crystal sizes 30–150 μm. They appear to be of primary, igneous origin, with no evidence of metamorphic reworking.

Three sieve fractions (<60 μm, 70–60 μm and >100 μm; # 1–3 in Table 1) and two abraded zircon fractions (60–50 μm and 100–70 μm; # 4, 5 in Table 1) consisting of mostly idiomorphic and transparent zircons were analysed. On a concordia plot all the data points are discordant (Fig. 10a) and do not belong to a common regression line. Analyses of the smallest zircon fraction and both abraded zircon fractions define a discordia intersecting the concordia at 381.3±5.7 and 1232±31 Ma respectively (MSWD = 1.9). The data points for these zircons cluster near the lower intercept of the discordia. Displacement of two unabraded zircon fractions (# 2 and 3 in Table 1) from this discordia could be explained either by recent Pb loss in these zircons or by the presence of different age inherited components of radiogenic Pb. Taking into account the igne-

Table 1: U-Pb isotope data for the zircon from the orthogneiss sample NTJ-1.

№	Sieve fraction μm	Fract. weight mg	Concentr. ppm		${}^{206}Pb/{}^{204}Pb^a$	Isotopic ratios corrected for blank and common Pb^b				Rho^c	Age, Ma		
			Pb	U		${}^{207}Pb/{}^{206}Pb$	${}^{208}Pb/{}^{206}Pb$	${}^{207}Pb/{}^{235}U$	${}^{206}Pb/{}^{238}U$		${}^{207}Pb/{}^{235}U$	${}^{206}Pb/{}^{238}U$	${}^{207}Pb/{}^{206}Pb$
1	<60	0.70	56.8	705	363	0.05965±7	0.0581±1	0.5826±20	0.0708±2	0.93	466.2±1.6	441.2±1.4	590.9±2.6
2	–70+60	1.54	45.4	604	893	0.06165±3	0.0568±1	0.6196±19	0.0729±2	0.99	489.6±1.5	453.6±1.4	661.8±1.1
3	>100	1.23	40.5	523	1392	0.06372±4	0.0623±1	0.6729±22	0.0766±2	0.98	522.5±1.7	475.8±1.5	732.3±1.5
4	–60+50 A 50%	0.44	36.4	428	565	0.06271±6	0.0749±1	0.6791±22	0.0785±2	0.95	526.2±1.7	487.4±1.5	698.5±2.2
5	–100+70 A 30%	0.25	76.0	861	1175	0.06517±8	0.0858±1	0.7646±25	0.0851±3	0.93	576.6±1.9	526.4±1.6	779.8±2.5

Notes: ^{a} — measured ratio; ^{b} — uncertainties (95 % confidence level) refer to last digits of corresponding ratios; ^{c} – correlation coefficients of ${}^{207}Pb/{}^{235}U$ vs. ${}^{206}Pb/{}^{238}U$ ratios; 50 % of zircon removed during of the air-abrasion.

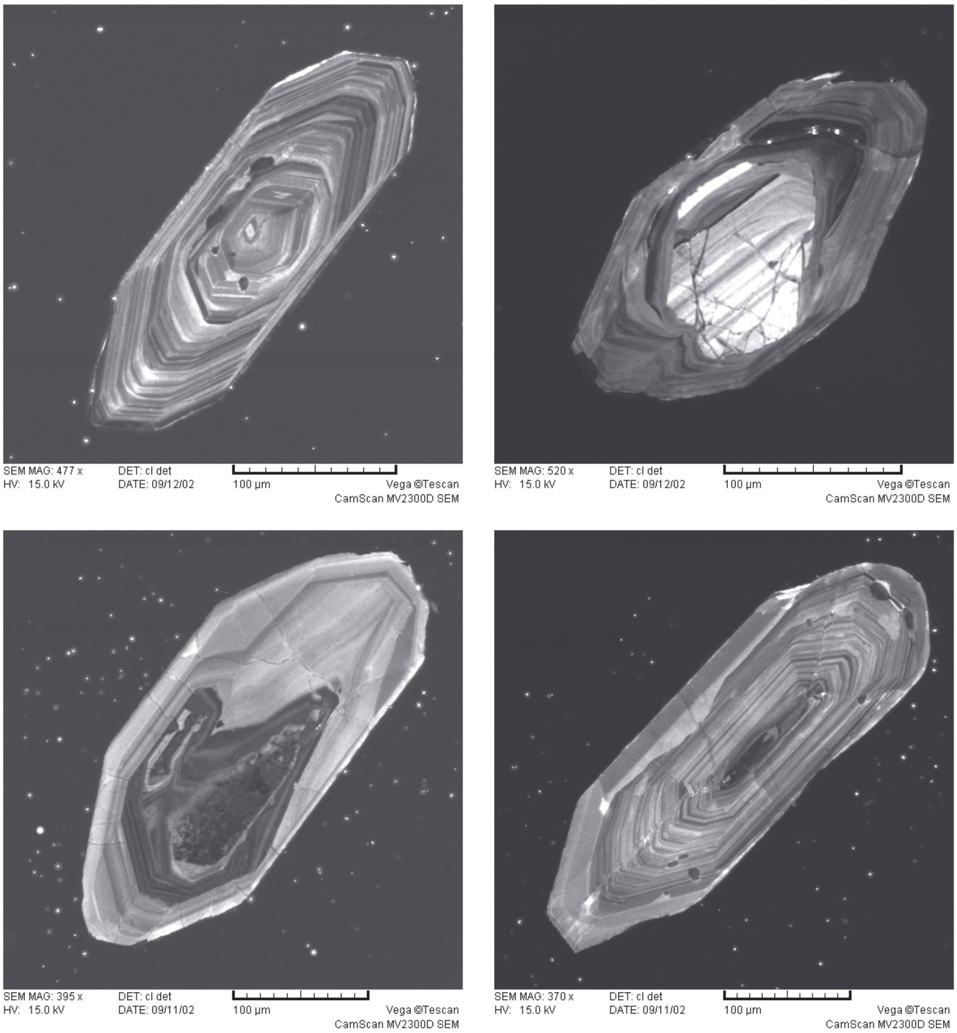

Fig. 9. CL images showing internal structure of zircons from samples NTJ-1 (**a**, **b**) and DUM-1 (**c**, **d**).

ous origin of the zircons, the lower intercept of the calculated discordia is interpreted as the primary emplacement age of the granitoids.

The Nízke Tatry DUM-1 tonalite

The zircon population from the Ďumbier tonalites (sample DUM-1) consists of idiomorphic and subhedral transparent (about 70 % of population) and cloudy, prismatic and pyramidal pale pink crystals showing a magmatic zonation in CL. The zircons often reveal oscillatory zonation and cores relics visible under CL (Fig. 9c,d). The range of crystal sizes is 40–250 μm. Zircons have a length/width ratio of 2.0–4.0, and appear to be of primary igneous origin.

Three sieve fractions (<60 μm, 80–60 μm and >100 μm; # 1–3 in Table 2) consisting of mostly idiomorphic and transparent zircon were analysed. A zircon from the smallest fraction (<60 μm; # 3, Table 2) was subjected to air-abrasion whereby about 40 % of its material was removed. Data points define a discordia intersecting the concordia at 343±3 and

Table 2: U-Pb isotope data for the zircon from the tonalite sample DUM-1.

№	Sieve fraction μm	Fraction weight mg	Concentr. ppm			Isotopic ratios corrected for blank and common Pb[b]				Rho[c]	Age, Ma		
			Pb	U	$^{206}Pb/^{204}Pb$[a]	$^{207}Pb/^{206}Pb$	$^{208}Pb/^{206}Pb$	$^{207}Pb/^{235}U$	$^{206}Pb/^{238}U$		$^{207}Pb/^{235}U$	$^{206}Pb/^{238}U$	$^{207}Pb/^{206}Pb$
1	>100	0.41	21.9	365	1660	0.0571±1	0.1192±1	0.4573±17	0.0581±2	0.73	382.4±1.4	363.8±1.1	496.1±5.6
2	–80+60	0.97	22.0	380	2636	0.0549±1	0.1291±1	0.4232±8	0.0559±2	0.86	358.4±0.7	350.7±0.7	408.4±1.6
3	<60 A 40%	0.76	10.7	175	790	0.0542±1	0.1585±1	0.4148±12	0.0555±2	0.53	352.3±1.1	348.2±0.7	379.1±4.8

Notes: [a] — measured ratio; [b] — uncertainties (95% confidence level) refer to last digits of corresponding ratios; [c] — correlation coefficients of $^{207}Pb/^{235}U$ vs. $^{206}Pb/^{238}U$ ratios; 40 % of zircon removed during of the air-abrasion.

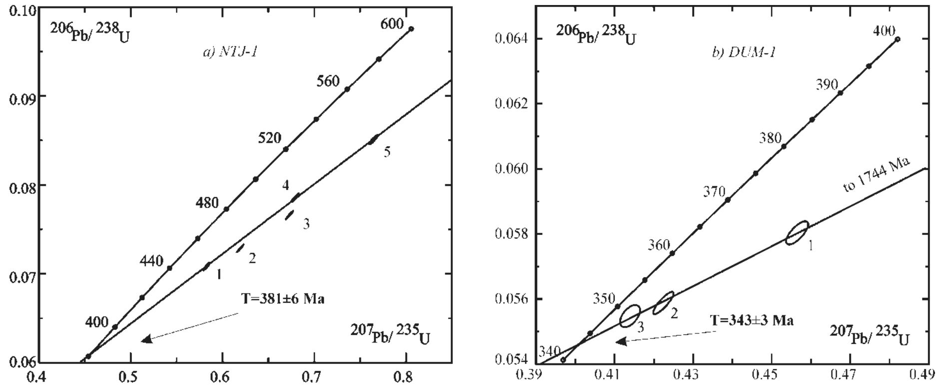

Fig. 10. Concordia diagrams for zircons from the Nízke Tatry Mts.: (**a**) orthogneiss NTJ-1 sample; (**b**) tonalite DUM-1 sample.

1744±205 Ma respectively, MSWD = 0.68 (Fig. 10b). The data points for this zircon cluster near the lower intercept of the discordia, ascribed to the presence of inherited radiogenic Pb component, mostly revealed in zircon from fraction >100 μm. Taking into account the igneous origin of the study zircon, the lower intercept age (343±3 Ma) is interpreted as the primary emplacement age of the tonalites. In conclusion, the Ďumbier tonalites were emplaced in the Variscan cycle, but later than the magmatic protoliths of the NTJ-1 orthogneiss.

Discussion

In agreement with Krist et al. (1992), our structural and geochronological results (although conventional multi grain zircon ages) strongly suggest that K-feldspar-bearing orthogneisses within the Jarabá structural complex (Fig. 2), including the dated sample NTJ-1 at ~380 Ma, originate from early Variscan syncollisional magmatism and metamorphism. This is consistent with the fabric similarities that are observed between both the "older" granites and orthogneisses, and the host metamorphics. The ongoing collisional thickening of the Upper Tatra Nappe raised the temperature of its middle- and lower-crustal complexes, and decompression along the thrust-fault, shear zones triggered partial melting and generation of large meso-Variscan (~360–340 Ma) mainly S-type granitoid plutons that emplaced together with older I-type granitoids (including the dated Ďumbier tonalite at 343±3 Ma) at mid-crustal level. Magmatic emplacement accompanied tectonic erosion of the wall mostly ductiley deformed rocks due to active subvertical probably transtension zones. Thus, a special "wall-type" magmatic stoping occurred in marginal zones of these granitoid plutons, characteristic of large xenoliths of the metamorphic mantle rocks, include the orthogneisses.

Having characterized the evolution of orthogneisses, three stages of nappe formation can be reconstructed from the structural and age relationships between the orthogneisses and the granitoid plutons within the Western Carpathians basement:

(1) *An early Variscan stage at* ~405–360 Ma took place, when the southeast-vergent mid-crustal Jarabá and lower crustal leptynite-amphibolite complexes were collisionally juxtaposed into the composite Upper Tatra Nappe. The orthogneisses, now dated at ca. 380 Ma, instead of ~405–380 Ma (Krist et al. 1992; Adamija et al. 1992; Poller et al. 2000) were magmatically emplaced at the base of the Jarabá Complex (Fig. 2). Therefore the magmatic age of K-feldspar-bearing orthogneisses is ascribed to the metamorphic peak of the collision regional-metamorphism within the Jarabá Complex (405–380 Ma), following an older (430–410 Ma) subduction/obduction metamorphic event recognized in the Variscan belt (von Quadt & Gebauer 1988; Matte 1991). The composite Upper Tatra Nappe is inferred to have overriden the Middle Hron Nappe at 380–360 Ma. The plagioclase orthogneiss (meta-trondhjemite) magmatically emplaced within the LAC (Fig. 2) at ~500 Ma (Putiš et al. 2001).

(2) *A meso-Variscan event took place at* ~360–340 Ma, when the Upper Tatra Nappe was intruded by large granitoid plutons emplaced into initially E–W to NE–SW (?) striking transtensional (?) shear zones within collapsing thickened crust. The mylonitic structures in these orthogneisses likely developed during this event. Part of them show SE-vergent thrusting close the base of the Jarabá Complex (in the Nízke Tatry Mts), or a transtensional top-to the NW shearing (in the Malá Fatra Mts).

(3) *A late-Variscan event took place at* ~340–300 Ma as the consequence of an oblique collision of the mostly low-grade basement complexes of the Lower Nappe (basement of the mid-Cretaceous Gemeric Unit, Fig. 2) with the early-orogenic Variscan basement complexes (basement of the mid-Cretaceous Tatric and Veporic Units). The youngest Variscan granitoids of the Upper Tatra Nappe, of mostly I-types (Sihla and Modra tonalites for example) fall into this time interval. They indicate a thermal event accompanying the late Variscan post-collisional collapse and formation of whole-crustal extensional faults.

Poller et al. (2000) have attributed the origin of the Western Tatra Mountains granitic orthogneisses (at ~405 Ma) to the subduction of an oceanic slab and generation of "older" granites in the upper plate active continental margin.

Conclusions

1. Our structural analysis of the orthogneisses from the Tatric basement Jarabá Complex (Nízke Tatry and Malá Fatra

Mountains) indicates that an early Variscan thrust-fault zone became active in-between the tectonically juxtaposed mid-crustal Jarabá Complex and lower-crustal leptynite-amphibolite complex that build most of the Tatric and Veporic crystalline basement of the Western Carpathians.

2. Magmatic ages and structures point to a regional-metamorphic, syn-collisional environment that rapidly transformed the oldest granites into orthogneisses.

3. We have distinguished the principal stages of microstructural evolution of the orthogneisses (magmatic, submagmatic and solidus) each having its characteristic deformation micro-mechanisms, textural patterns and mineral changes. Strong microstructural gradients point to higher strain rates during the former magmatic emplacement of the orthogneiss, in agreement with points 1 and 2, as well as during their final exhumation.

4. Our U-Pb dating of the Nízke Tatry (Jasenie, Struhár) orthogneiss yield an upper intercept at 1232±31 Ma, and a lower intercept at 381±6 Ma. The latter age is interpreted as the crystallization age of the original granite-granodiorite that was subsequently transformed into orthogneiss. The medium- to low-grade mylonitic fabrics of the orthogneisses reflect a continuous straining of the shear zone into which were emplaced large granitoid plutons. This continuum is in agreement with the obtained age of 343±3 Ma for the Ďumbier tonalite of the Upper Tatra Nappe in the Nízke Tatry Mountains.

5. Petrological-geochemical data indicate a sedimentary-metamorphic S-type protolith of the dated NTJ-1 orthogneiss. By contrast, the dated DUM-1 tonalite shows an igneous I-type origin.

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