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Argon and fission track dating of Alpine metamorphism and basement exhumation in the Sopron Mts. (Eastern Alps, Hungary): thermochronology or mineral growth?

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Summary

The crystalline basement rocks of the Sopron Mountains are the easternmost and most isolated outcrops of the Austroalpine basement of the Eastern Alps. Ar/Ar and K/Ar dating of phengitic mica indicates that the Eoalpine high-pressure metamorphism of the area occurred between 76 and 71 Ma. Short-lived metamorphism is characterised by fluid-poor conditions. Fluid circulation was mostly restricted to shear zones, thus the degree of Alpine overprint has an extreme spatial variation. In several metamorphic slices Variscan mineral assemblages have been preserved and biotite yielded Variscan and Permo-Triassic Ar ages. Different mineral and isotope thermometers (literature data) yielded temperatures of 500–600 °C for the peak of Alpine metamorphism in the Sopron Mountains, but muscovite and biotite do not show complete argon resetting. Thus, we consider this crystalline area as a well constrained natural test site, which either indicates considerably high closure temperatures (around 550 °C) for Ar in muscovite and biotite in a dry metamorphic environment, or which is suitable for testing the widely applied methods of temperature estimations under disequilibrium conditions.

Apatite fission track results and their thermal modeling, together with structural, mineralogical and sedimentological observations, allows the identification of a post-metamorphic, Eocene hydrothermal event and Late Miocene-Pliocene sediment burial of the crystalline rocks of the Sopron Mountains.

Introduction

The Sopron Mountains are one of the easternmost outcrops of the 500 km long belt of Austroalpine basement of the Eastern Alps (Fig. 1a; *Tollmann*, 1977). Towards the east, several km thick Miocene clastic sequences cover the basement, and

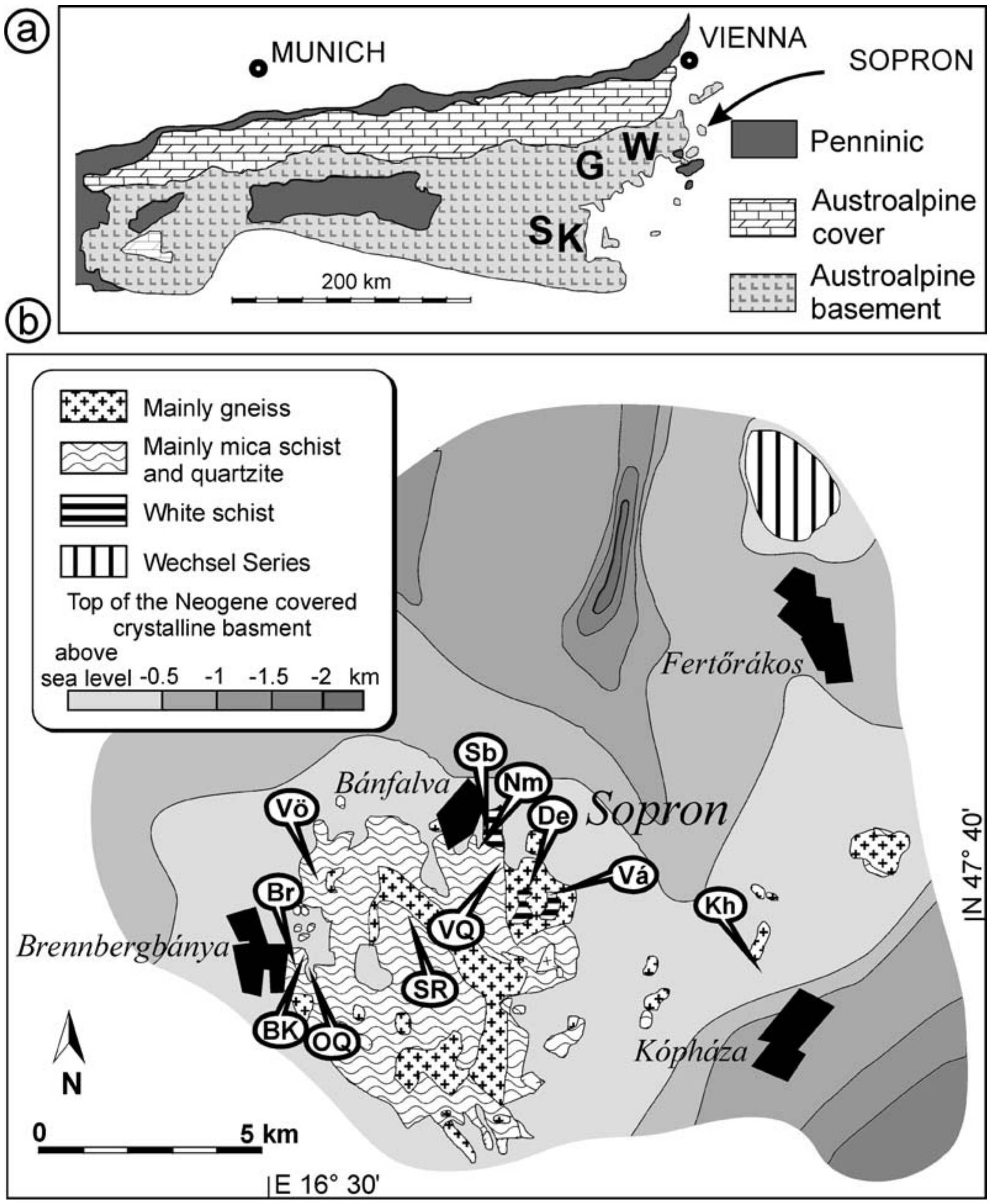

Fig. 1. **a** Geographic position within the Eastern Alps **b** Geological sketch map of the Sopron Mountains (after *Fülöp*, 1990). *Austroalpine units in the overview map: S* Saualpe; *K* Koralpe; *W* Wechsel; *G* Gleinalm; *localities in the Sopron Mountains: BK* Brennberg, Kovács árok; *OQ* Oromvég Quarry; *Br* Brennberg; *Vö* Vöröshid; *SR* SR-1 borehole; *Sb* Bánfalva; *Nm* Nándormagaslat; *De* Deákkút; *Vá* Váris; *Kh* Kópháza; *VQ* Vashegy Quarry

Variscan units, overprinted by Eoalpine (Late Cretaceous) greenschist-amphibolite facies metamorphism are exposed in the Western Carpathians (*Koroknai* et al., 2001). Towards the west the lithologic composition of the Austroalpine basement is rather variegated, and the distribution of the metamorphic grades shows a collage-like pattern (*Frey* et al., 1999; *Schuster* and *Thöni*, 2001) due to the Late Cretaceous-Tertiary displacement of blocks following the Eoalpine metamorphism (*Neubauer* and *Frisch*, 1993; *Frisch* et al., 2000). All these features can be observed on a smaller scale in the Sopron Mountains (Fig. 1b). The basement comprises slices of para- and orthometamorphic lithologies and these slices are separated by flat-lying ductile shear zones partly transformed to whiteschist (leucophyllite). The grade of the post-Variscan metamorphic overprint varies between and also within the tectonic blocks. In some micaschists mineral assemblages formed during Variscan and Permo-Triassic high-temperature (HT) metamorphism are preserved (*Schuster* et al., 2001), whereas in other parts of the same crystalline slab (sometimes in the same quarry) a completely Alpine mineral paragenesis was formed (*Török*, 1996). Geothermobarometry of Alpine assemblages indicates metamorphic conditions of up to 1300 MPa and 500–600 °C (see compilation below and in *Török*, 2001). The high variability in the mineralogy and the well-constrained P–T conditions of the metamorphic events make the Sopron Mountains a unique natural laboratory for an empirical study of argon resetting in white mica and biotite. In this study we aim to integrate the geochronological results into a coherent thermotectonic history and show the evolution in a new type of diagram which presents the datable events versus age and not simply the time/temperature path.

Geological setting

The Sopron Mountains form a group of isolated hills (up to 559 m a.s.l.), surrounded by Neogene molasse sediments, at the margin of the Pannonian Basin. The crystalline formations are mainly medium-grade mica schists and orthogneisses of very variable mineralogical composition. They belong to the pre-Mesozoic basement of the Eastern Alps and are related to the "Grobgneis" series of the Lower Austroalpine nappe complex. The crystalline schists are mostly made up of orthogneisses (Sopron Gneiss Formation, SGF) and mica schists (Sopron Micaschist Formation, SMF). *Draganits* (1998) has proposed another nomenclature; he distinguished the Sopron Series (orthogneisses and monotonous diaphtoritic mica schists) with Alpine mineral assemblages and the Óbrennberg-Kalter Bründl Series (mainly mica schist) with relatively well-preserved pre-Alpine mineralogy. Whereas the former was related to the Lower Austroalpine units, the latter one was assigned to the Middle Austroalpine (*Draganits*, 1998). In the following we will use the SMF-SGF subdivision and abbreviations according to *Kisházi* and *Ivancsics* (1985).

The SMF was affected by pre-Variscan – Variscan metamorphism producing andalusite–sillimanite–biotite schists (*Lelkes-Felvári* et al., 1984; *Kisházi* and *Ivancsics*, 1985a; *Draganits*, 1998) and it was overprinted by HT/LP Permo-Triassic metamorphism (*Schuster* et al., 2001). For the latter event the pressure–temperature conditions were estimated at 650 °C and 300–500 MPa by *Draganits* (1998). From the different mineral reactions *Török* (1999) concluded a sequence of metamorphic stages. The highest temperature assemblages reached the beginning of the granulite facies (650–700 °C, 240–380 MPa). Some parts of SMF also underwent Alpine HP metamorphism producing kyanite–chloritoid and chlorite-bearing muscovite schists. Garnet formed locally during Alpine metamorphism (*Draganits*, 1998). According to different authors the P–T conditions of this high-pressure amphibolite facies event were:

450–500 °C, 1200 MPa, *Török* (1996)
560 ± 30 °C, 1300 MPa, *Demény* et al. (1997)
550 ± 30 °C, 950 ± 150 MPa, *Draganits* (1998)
450–550 °C, 1300–1400 MPa, *Török* (1998).

These geothermobarometric data are based on both mineral equilibria and oxygen isotope fractionation data.

Whiteschists or leucophyllites are very characteristic rocks of the area. These Mg-rich, strongly foliated, sometimes kyanite-bearing leuchtenbergite–muscovite–quartz schists form few meters thick shear zones. Their genesis has been intensely debated; derivation from Mg-rich pelitic protoliths as well as a pure metasomatic origin was discussed (*Lelkes-Felvári* et al., 1982; *Kisházi* and *Ivancsics*, 1987). Metamorphic temperatures were estimated at 460–480 °C by *Prochaska* (1991). *Demény* et al. (1997) related these temperatures to Alpine metamorphic peak conditions, whereas *Török* (2001) related them to retrograde metamorphic processes.

Rather few geochronologic data are available from the Sopron Mountains proper. *Draganits* (1996) has presented mica Rb/Sr and Ar/Ar ages measured at the Laboratory of Geochronology, Vienna University. The Rb/Sr ages range from 214 to 102 Ma, the coarser fractions yielding older ages than the finer ones. The

Table 1. *K/Ar ages from the crystalline schists of the Sopron Mts. Abbreviations of the sample sites are shown in Fig. 1b*

Code	Lithology	Phase	Fraction [μm/A]	K [wt%]	^{40}Ar(rad) *	^{40}Ar(rad) [%]	Age ± 1s [Ma]
Nádormagaslat (Nm)							
Nm-1	gneiss	w. m.	315–630	8.65	5.45	95.3	155.0 ± 5.2
Nm-2	gneiss	w. m.	160–200	7.82	2.81	64.6	90.1 ± 3.6
	gneiss	w. m.	315–630	8.68	4.61	88.6	131.7 ± 5.0
Nm-3	gneiss	w. m.	200–315	7.97	3.24	74.8	101.7 ± 3.9
Nm-4	gneiss	w. m.	315–630	8.86	4.95	89.8	138.5 ± 5.3
Váris (Vá)							
Vá-1	gneiss	bt	80–160	7.17	2.5	52.5	87.5 ± 3.7
	gneiss	bt	315–630	7.40	2.23	52.8	79.1 ± 3.3
Vá-1	gneiss	w. m.	80–160	8.48	3.09	69.9	91.6 ± 3.6
	gneiss	w. m.	315–630	8.42	3.95	70.3	117.0 ± 4.5
Vá-2	gneiss	w. m.	315–630	8.10	5.25	89.3	160.0 ± 6.1
Vá-3	gneiss	w. m.	315–630	8.64	3.77	88.3	109.1 ± 4.2
Vá-4	gneiss	w. m.	315–630	8.50	4.96	77.5	144.2 ± 5.5

(*continued*)

Table 1 (*continued*)

Code	Lithology	Phase	Fraction [μm/A]	K [wt%]	^{40}Ar(rad) *	^{40}Ar(rad) [%]	Age ± 1s [Ma]
Vá-4	gneiss	bt	315–630	7.29	2.27	70.0	78.5 ± 3.3
Vá-5	gneiss	w. m.	315–630	8.43	3.97	93.7	117.4 ± 4.5
	gneiss	bt	63–315	7.25	2.98	74.5	102.8 ± 4.0
	gneiss	bt	315–630	7.15	2.9	79.1	101.6 ± 3.9
Vá-7	gneiss	bt	315–630	6.82	2.43	50.1	89.6 ± 3.8
Kópháza (Kh)							
Kh/7	gneiss	w. m.	80–160/1A	8.89	3.1	93.2	87.9 ± 3.3
Kh/8	gneiss	w. m.	80–160/1.1A	8.65	3.01	98.4	87.4 ± 3.3
Kh/9	gneiss	w. m.	80–160/1.2A	7.12	2.55	97.9	90.1 ± 3.4
Kh/6	gneiss	w. m.	70–180	6.02	2.03	90.0	84.7 ± 3.2
Vashegy quarry (VQ)							
3144	leucophyllite	w. m.	63–100	7.19	2.186	88.1	76.6 ± 2.9
		w. m.	500–630	7.67	4.39	94.4	137.4 ± 5.4
3146	gneiss	w. m.	200–630	6.92	2.287	93.6	83.1 ± 3.2
		w. m.	500–630	9.81	4.099	94.5	104.4 ± 4.0
Oromvég quarry (OQ)							
SO94-2	Ky–Cld–Ms	w. m.	56–100	4.19	1.977	50.8	117.5 ± 5.0
SO94-3	Ky–Cld–Ms	w. m.	>500	8.29	6.005	83.5	177.3 ± 6.7
SO94-4	Ky–Cld–Ms	bt	250–160	5.46	5.957	90.6	260.9 ± 9.9
	Ky–Cld–Ms	bt	100–160	5.06	4.952	89.1	235.7 ± 9.0
	Ky–Cld–Ms	w. m.	63–100	3.69	1.481	52.4	100.5 ± 4.2
	Ky–Cld–Ms	w. m.	160–250	3.30	1.499	27.2	113.2 ± 6.6
Brennberg, Kovács-trench (Br)							
2823	And–Sil–Bt	w. m.	70–315	4.68	2.61	85.4	138.4 ± 5.3
		bt	70–160	7.36	8.4	92.6	272.0 ± 10.0
		bt	160–315	7.44	8.55	89.0	274.0 ± 11.0
3147	And–Sil–Bt	bt	125–200	7.46	7.849	97.5	252.2 ± 9.6
		w. m.	125–200	5.74	2.221	84.0	96.9 ± 3.7
Brennberg, side valley of Kovács-trench (BK)							
3148	granite	w. m.	315–630	5.25	2.455	86.3	116.5 ± 4.4
		bt	63–200	5.67	4.667	86.0	200.2 ± 7.6
3149	Bt	bt	125–315	6.78	9.225	96.0	319.7 ± 12.1
SO94-1	Gar–Bt	bt	63–100	5.60	7.851	95.2	328.5 ± 12.5
		bt	160–250	6.47	8.79	93.0	319.5 ± 12.1
		w. m.	63–100	5.96	2.493	85.7	104.5 ± 4.0
		w. m.	250–630	6.78	2.589	25.5	95.6 ± 5.8
Sopronbánfalva (Sb)							
2532	leucophyllite	w. m.	70–160	6.69	2.16	93.6	81.1 ± 3.1

μm/A: Size fraction in micrometer and Ampere used at the isodynamic magnetic separation with side angle of 10°; *: ^{40}Ar(rad) in 10^{-5} cc STP/g. *bt*: biotite, *w. m.*: white mica, *Gar-, And-, Sil-, Bi*: garnet-, andalusite- sillimanite- and biotite-schists total gas ages of the micas from pegmatites range from 286 to 155 Ma. Monazite U–Th–Pb electron microprobe ages indicate Variscan HT metamorphism, and the diffuse lower intercept around 80 Ma points to some Alpine phosphate mineralisation (*Nagy* et al., 2002).

From the adjacent areas *Frank* et al. (1996) dated metamorphic rocks belonging to the Wechsel Series from the borehole Fertőrákos-1004, about 10 km NE of the Sopron Mts (Fig. 1b). Variscan to Late Cretaceous Ar/Ar ages were obtained; most data fall in the interval of 220–170 Ma. *Frank* et al. (1996) explained the considerable variation of the ages by high ^{40}Ar overpressure and incorporation of excess Ar into the dated minerals during nappe stacking. The zircon and apatite fission track (FT) ages of the crystalline formations of the Fertőrákos Hills are between 62 and 54 Ma and between 46 and 41 Ma, respectively.

Müller et al. (1999) reported radiometric ages from the Wechsel crystalline unit, situated west of the study area (Fig. 1a). They measured both Variscan and Alpine ages (375–270 Ma and 90–80 Ma, respectively) and concluded that during the Alpine, greenschist facies event the new minerals were formed only along distinct shear bands. In the surrounding, less deformed polymetamophic rocks the Variscan ages were not affected. *Schuster* et al. (2001) has pointed out the importance of the Permo-Triassic HT/LP metamorphic event, which has numerous manifestations also in the surroundings of the Sopron Mountains. The Permian and Triassic ages (270–240 Ma) of the high temperature mineral assemblages were determined by Sm/Nd, Rb/Sr and Ar/Ar methods.

A very detailed reconstruction of the timing of thrusting events during Late Cretaceous compression (100 to 70 Ma) was attempted by *Dallmeyer* et al. (1998). They investigated highly deformed thrust planes throughout the whole Austroalpine pile and concluded that the Alpine ages related to greenschist facies metamorphism are only recorded along the detachment zones.

Compilations of the radiometric ages of the entire Austroalpine realm and a more recent synthesis on its easternmost part can be found in *Frank* et al. (1987) and *Schuster* and *Thöni* (2001), respectively.

Argon chronology

52 K/Ar and 6 Ar/Ar age determinations were performed on white mica, biotite and feldspar. The feldspar ages were previously published in *Balogh* and *Dunkl* (2001); in this study we will therefore focus on the mica Ar ages. The analytical techniques of K/Ar and Ar/Ar dating (ATOMKI, Debrecen, Hungary) have been described by *Balogh* et al. (1999) and by *Balogh* and *Simonits* (1998). Sampling sites are shown in Fig. 1. The results are presented in Table 1.

Sopron Gneiss formation

Crystalline basement rocks of the Sopron Mts. contain white micas, the crystal size of which ranges from 20 μm to 5 mm. In the orthogneisses white mica occurs in two generations: the older, coarser-grained type is classified as muscovite, the younger, finer-grained one is phengite (Fig. 2). *Török* (1996) supposed that the

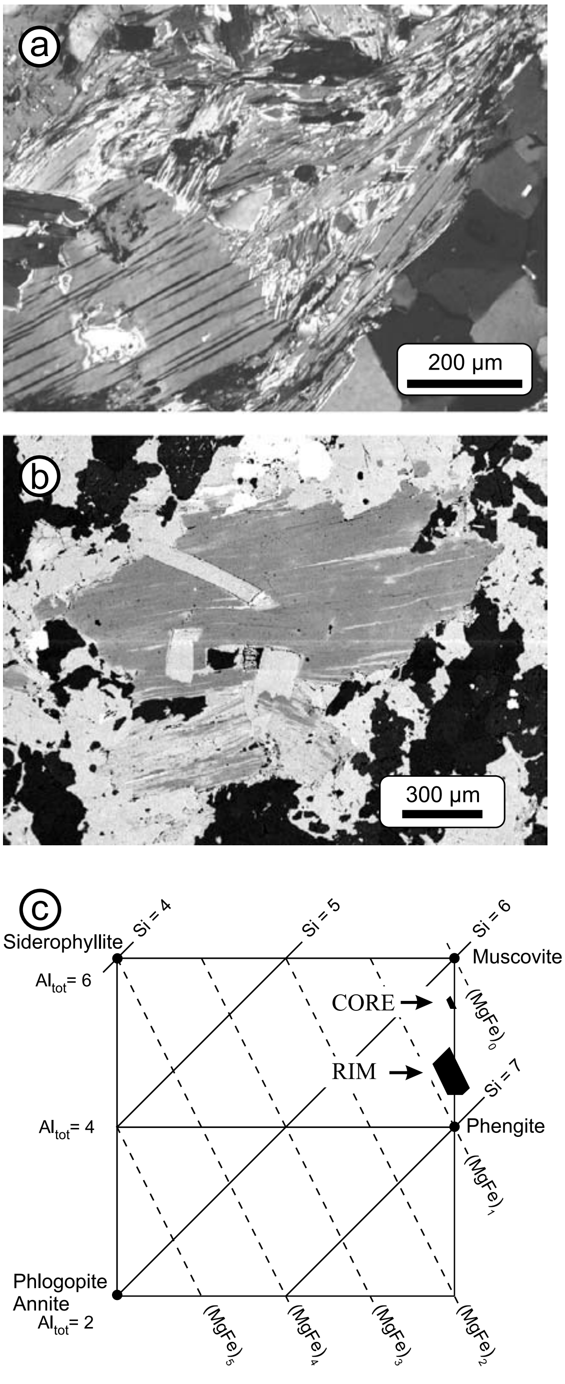

Fig. 2. **a** Photomicrograph showing two well distinguishable white mica generations: an old (Variscan) coarse-grained and a new (Alpine) finer-grained one. The latter is mainly arranged in shearbands (Vá-1 sample). **b** On the backscattered electron image the different composition of the older cores of muscovitic composition (gray) and the finer overgrowths of phengitic composition (light gray) is well visible. The two phases show tight intergrowths (Vá-4 sample). **c** Compositions of old (core) and new (rim) mica generations. The broader compositional range of the second phengitic mica generation is in harmony with its texture and its rapid growth under disequilibrium conditions

low-Si muscovite cores are of magmatic origin, whereas the phengite is thought to have formed during Alpine metamorphism.

The white mica K/Ar ages range from 160.0 Ma to 76.6 Ma in the SGF. The youngest age (76.6 ± 2.9 Ma) was measured on fine-grained white mica from leucophyllite while the youngest age measured on a gneiss sample is 84.6 ± 3.3 Ma. Size fractions of white mica are plotted against their K/Ar ages in Fig. 3. There is a

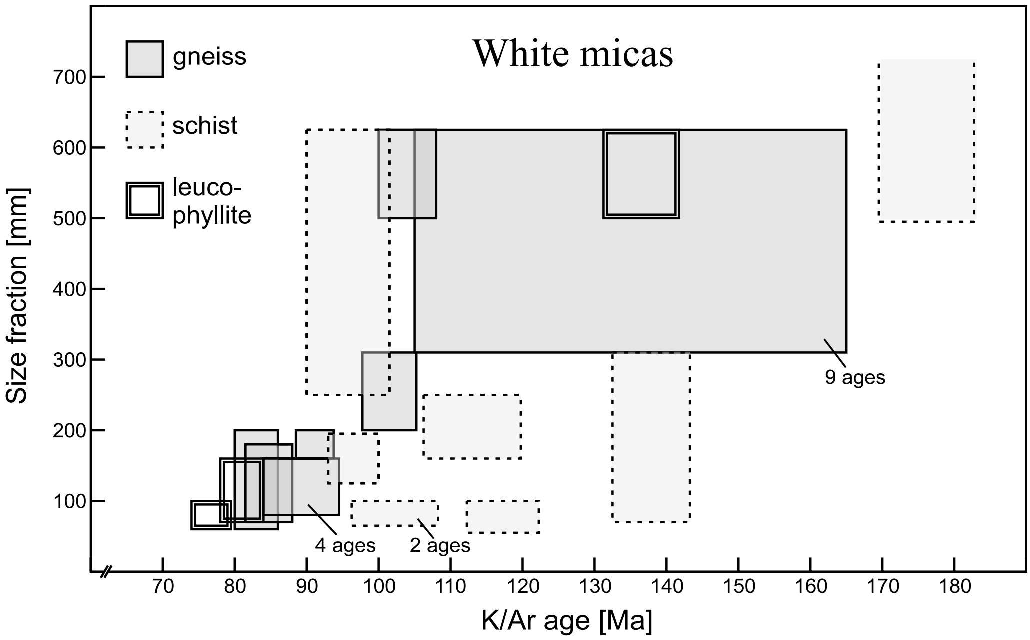

Fig. 3. Relation of white mica K/Ar ages and size fractions

tendency that finer-grained mica fractions dominated by phengitic compositions are younger than the coarse-grained micas. Their Late Cretaceous age supports the assumption that phengites were formed during the Eoalpine tectonothermal event, although K/Ar ages alone are insufficient proof for the timing of phengite formation. This is because: (i) even fine-grained phengite fractions may contain an older muscovite component; in this case the real phengite age is younger than the K/Ar datum and (ii) the closure temperature of phengite is a bit lower than that of muscovite (*Wijbrans* and *McDougall*, 1986). In the latter case the Alpine overprint would have caused greater Ar release from the phengite. The older white mica ages of coarser size fractions can be interpreted either in terms of a partial reset of K/Ar ages or incorporation of variable amounts of excess Ar during the Alpine metamorphism. The more reliable scenario is that the old (pre-Late Cretaceous) K/Ar ages of the coarse white mica fractions reflect partial Ar loss and thus these are Permo-Triassic – Alpine mixed ages.

Ar/Ar geochronology was performed in order to check the crystal-chemical homogeneity of the measured white mica fractions and to characterize the possible age components. In the coarse-grained muscovite No. 2509 (fom the Váris quarry) the oldest ages (193.6 ± 2.3 Ma and 204.1 ± 2.1 Ma) correspond to degassing steps at higher temperatures (Fig. 4a). This age range is close to that of HT Permo-Triassic metamorphism (*Schuster* et al., 2001). The complexity of the spectrum reflects the fact that the dated fraction is a mixture of crystals having different mineral chemistry. This coarse size fraction contains both the old big muscovite, and the later formed, finer phengite.

A finer-grained mica fraction (No. 2508, also from Váris quarry) is mostly composed of phengite of the second mica generation. The step-heating ages give

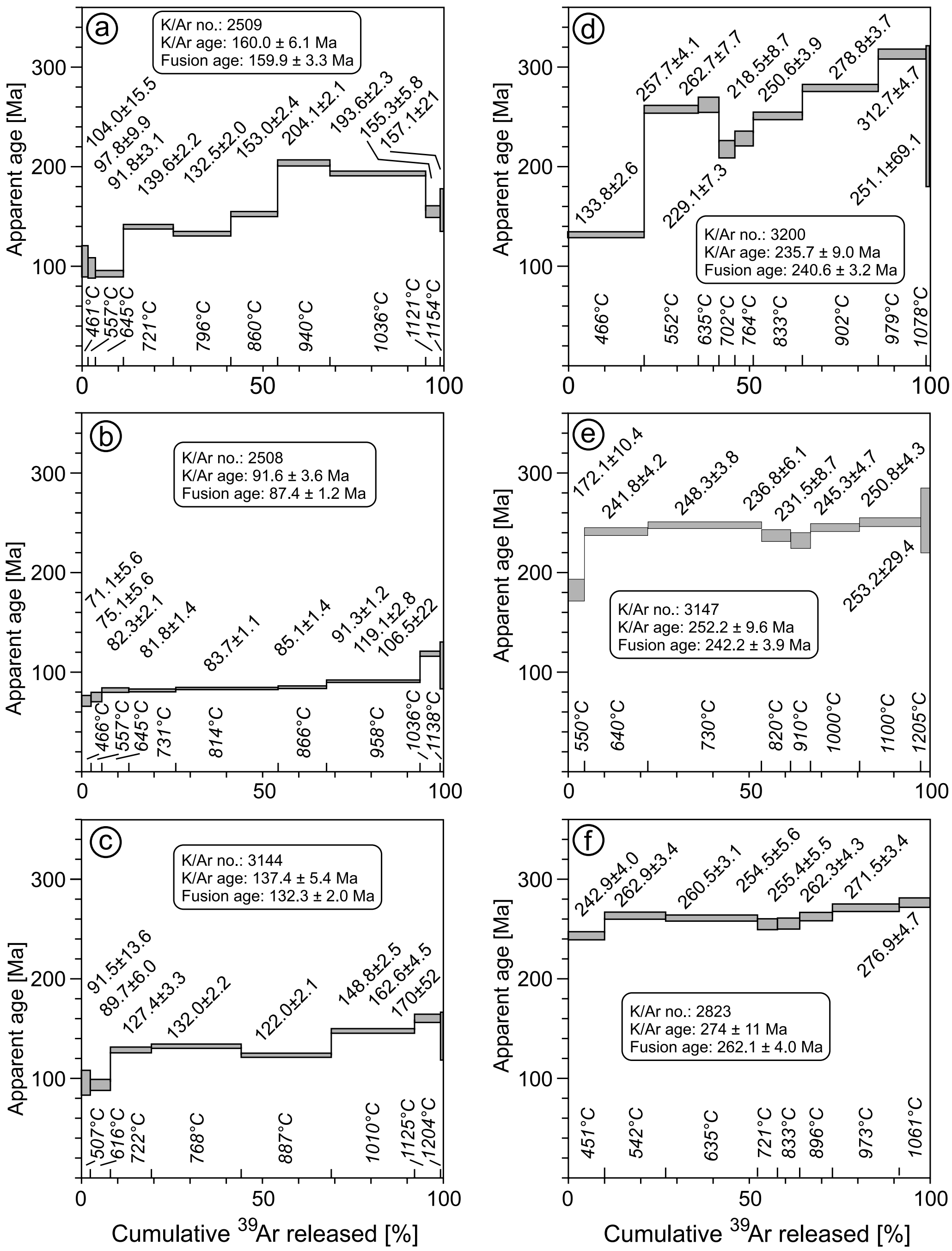

Fig. 4. $^{40}Ar/^{39}Ar$ age spectra of different size fractions of white micas: **a** Váris-2, orthogneiss, muscovite, 315–630 µm. **b** Váris-1 orthogneiss, phengite, 80–160 µm. **c** Vashegy, leucophyllite, 500–630 µm. $^{40}Ar/^{39}Ar$ age spectra of biotites from the Sopron Micaschist Formation: **d** Oromvég quarry; kyanite–chloritoid–muscovite–schist; 100–160 µm. **e** Brennberg, Kovács trench; andalusite–sillimanite–biotite–schist; 125–200 µm. **f** Brennberg, Kovács trench; andalusite–sillimanite–biotite–schist; 160–315 µm

a considerably narrower range than those of the coarse mica fraction (Fig. 4b). The formation of the main mass of the phengite occurred probably between 83 and 75 Ma. The fractions degassed at higher temperatures are probably mostly derived from relicts of the first mica generation. Figure 2b gives an illustration of the fine intergrowth of the two mica generations. Mica concentrates coarser than 10 μm always contain composite grains. The youngest white mica K/Ar age of 76.6 ± 2.9 Ma from the leucophyllite at Vashegy is close to the lowest values of the age spectrum on phengite No. 2508, showing that ages of fine-grained white micas in the leucophyllite mainly belong to the second generation and the dated fraction contains a minor amount of inherited old muscovite.

The third sample selected for Ar/Ar dating was the coarse-grained white mica No. 3144 from a leucophyllite at Vashegy. It shows a definite overprint during the Late Cretaceous event (first two steps), whereas at higher degassing temperatures the ages are above 160 Ma (Fig. 4c).

Biotites of the SGF (Váris quarry) yield ages from 102.8 Ma to 78.5 Ma (Table 1). The lack of correlation between the age and atmospheric Ar concentration suggests that partial resetting of ages was the dominant reason for the scatter of biotite ages. The similarity of the youngest white mica (76.6 ± 2.9 Ma) and biotite (78.5 ± 2.9 Ma) ages, the "isochron age" of the feldspars (74 ± 11 Ma, *Balogh* and *Dunkl*, 2001) and the ages at the lowest Ar/Ar degassing temperature steps (71.1 ± 5.6 Ma and 75.1 ± 4.7 Ma) on the fine-grained white mica from the gneiss of the Váris quarry show that all these ages were controlled by the Late Cretaceous metamorphism. Allowing for some excess Ar in the white micas and biotites, the period from 76 Ma to 71 Ma is regarded as the most likely age of this Alpine event.

Sopron Micaschist formation

The biotite K/Ar ages in the parametamorphic rocks are considerably older than in the orthogneisses. The dated metamorphic rocks have a rather variable mineralogical composition. The garnet–biotite schist, andalusite–sillimanite–biotite schist and kyanite–chloritoid–muscovite schist yielded biotite Ar ages from 328 to 200 Ma. The youngest age was measured on the biotite fraction of a granite dike. The ages show neither correlation with the K contents nor with the size fractions of the analysed phases (in contrast to the white micas; see Table 1).

The biotite K/Ar ages in the SMF are systematically older than the white mica ages. In the following we will investigate the possibility of the incorporation of excess Ar in the biotite. *Kulp* and *Engels* (1963) demonstrated that in metamorphic biotites the K/Ar system is more resistive to exchange processes than the Rb/Sr system. Hence, Ar ages older than Rb/Sr ages are insufficient to prove the presence of excess Ar. In the SMF the oldest biotite ages are close to the U–Th–Pb ages of 310 ± 34 Ma to 296 ± 41 Ma measured on monazite by *Nagy* et al. (2002), and to the Rb/Sr isochron age measured by *Frank* et al. (339 ± 12 Ma, 1996) on muscovite from a gneiss in the borehole Fertőrákos-1004. Thus, the oldest K/Ar ages are in accordance with the metamorphic history of the studied area. Ar is captured by the minerals from the intergranular fluid, which always contains an atmospheric component. If the $^{40}Ar/^{36}Ar$ ratio in the intergranular fluid was more or less uniform, a positive correlation can be expected between age and atmospheric Ar concentration. The lack of this correlation (Table 1) is an additional argument to regard the oldest biotite K/Ar ages (328.5–319.5 Ma) as a close, minimum approximation for the timing of Variscan metamorphism of the andalusite–sillimanite–biotite schist in the SMF.

Ar/Ar age spectra have been recorded on three biotite samples from the SMF (Fig. 4). An Alpine overprint of a post-Triassic age is recorded only in biotite from the kyanite–chloritoid–muscovite schist sampled at the Oromvég quarry (Fig. 4d). At the last degassing step this biotite yielded an age of 312.7 ± 4.7 Ma, which is slightly younger than the oldest K/Ar ages from the SMF. Biotites from the andalusite–sillimanite–biotite schist do not show any Alpine overprint (Fig. 4e, f). This is not a unique observation in the Eastern Alps. *Müller* et al. (1999) found that in the easternmost Lower Austroalpine nappes the Alpine overprint is detectable only in the $<63\,\mu m$ mineral fraction. The narrow scatter of the step ages points to better crystal chemical homogeneity than in the sample presented in Fig. 4d. However, all three spectra show younger ages at mid-temperature degassing steps. This phenomenon has been studied for biotite with chlorite contamination by *Lo* and *Onstott* (1989). They argued, that recoiled ^{39}Ar atoms are captured by the chlorite, thus producing a lower ^{40}Ar(rad)/^{39}Ar ratio. At 700–900 °C, when the chlorite decomposes, this excess ^{39}Ar will be liberated and can lower the age. This is also a convincing explanation for our data. Moreover we think that younger ages at mid-temperature steps argue against the presence of excess Ar in the biotite. Namely: concentration of excess Ar is independent from the K distribution, so the presence of excess Ar would cause an age increase in the low-K phase, compensating the younger ages caused by the recoiled ^{39}Ar atoms.

Disregarding the first degassing steps and the younger ages at medium temperatures, the two biotites from the andalusite–sillimanite–biotite schist gave plateau-like Permo-Triassic ages. This is best seen in biotite No. 3147 (Fig. 4e), for which ages range from 241.8 Ma to 250.8 Ma for 81% of ^{39}Ar released. Recent chronological data (e.g. *Berka* et al., 1998; *Schuster* and *Thöni*, 1996; *Schuster* et al., 1999) from the Austroalpine crystalline basement indicate widespread Permo-Triassic HT/LP metamorphism caused by extensional tectonics. The Permian and Triassic step ages in the Ar/Ar spectra are assumed to reflect the timing of this HT/LP, extension-related event in the SMF.

On closure temperatures

The ages presented above indicate a rather complex evolution of the crystalline formations of the Sopron Mountains. In addition to the thermal resetting of the Ar chronometer, mineral growth, phase transformation and deformation could have played a major role (*Villa*, 2001, 2003). In the following we shall point out the difficulties of assigning temperature values to the age data.

The closure temperature concept introduced by *Dodson* (1973) has been used successfully for several decades. *Purdy* and *Jäger* (1976) evaluated the following closure temperatures for micas in the Central Alps: Rb–Sr in muscovite and phengite 500 ± 50 °C, K/Ar in muscovite and phengite 350 ± 50 °C and Rb–Sr and K/Ar in biotie 300 ± 50 °C. *Harland* et al. (1990) listed the most commonly used closure temperatures for a set of minerals and methods.

They gave 350 °C, 280 ± 40 °C and 230 °C for the K/Ar system in muscovite, biotite and plagioclase, repectively, and 500–600 °C for the U–Th–Pb system in monazite. However, it became clear, that this concept and the widely accepted closure temperatures are unsuitable for the interpretation of radiometric data of certain metamorphic regions. It has been proven that isotope systems in minerals may survive temperatures well above their accepted ''closure'' temperatures. The recalculated closure temperatures published by *Villa* (1998) are systematically higher than the formerly used ones (e.g. 500 °C and 450 °C for the K/Ar system in muscovite and biotite and 770 °C for the U/Pb system in monazite) in cases, when stress and fluids did not promote isotopic exchange. He remarked ''most minerals can exchange isotopes even at the Earth's surface, if provided sufficient fluid''.

Circulating fluids and stress will reduce the recalibrated closure temperatures. It has been proven experimentally by *Kulp* and *Engels* (1963) that radioactive parent-daughter systems may exchange isotopes at temperatures below 100 °C. The decrease of closure temperature is an insufficiently studied phenomenon; it depends certainly on stress, fluid chemistry and fluid/rock ratio and the temperature of these processes. As a test of fluid effects, we calculated closure temperatures for phlogopite under hydrothermal conditions from the data by *Giletti* (1974) as cited by *McDougall* and *Harrison* (1988). Assuming cooling rates of 9.5×10^{-13}–9.5×10^{-14} K/s and effective grain sizes of 1–0.05 mm, closure temperatures from 488 °C to 358 °C were obtained, which are higher than those given by *Harland* et al. (1990, 280 ± 40 °C). This is in line with the recalibrated value of *Villa* (1998, 450 °C). Considering the results of *Kulp* and *Engels* (1963), the disappointing uncertainty of closure temperatures of minerals formed under hydrothermal conditions becomes clear. On the other hand, the success of the closure temperature concept of *Dodson* (1973) and the traditional closure temperature values listed by *Harland* et al. (1990) are promising and motivating for a better understanding of the influence of metamorphic conditions on the closure temperatures.

In the Sopron Mts. stress and fluid circulation during Alpine tectonics certainly reduced the closure temperatures along the shear zones. The differences of Ar release properties from hydrous minerals during vacuum and hydrothermal extraction has been studied and established by *Norwood* (1974) and *Gaber* et al. (1988): vacuum extraction results in unrealistically low activation energies and closure temperatures.

Due to these uncertainties, we refrain from assigning temperature values to the K/Ar ages. Instead, we only point to the relation of formation and closure temperatures of the dated minerals. It is well established that, when Ar release is controlled by temperature, muscovite retains Ar better than biotite; this can be observed in the SGF. In the SGF newly formed phengite is present together with old, only partially rejuvenated muscovite. *Wijbrans* and *McDougall* (1986) reported on coexisting phengite and muscovite, where the muscovite gave younger Ar ages. This implies that both muscovite and phengite can be formed below the closure temperature of the other mineral. In the SMF phengite ages are systematically younger than biotite ages. This shows, that phengite can be formed also below the ''dry closure temperature'' of biotite and that in the SMF the age relation of biotite and phengite is not controlled by temperature.

Geological factors controlling the Ar ages

Our K/Ar and Ar/Ar data show a very complex picture which is not free of apparent contradictions. The considerations listed in the previous section discussing "closure temperatures" indicate that the pure thermochronologic approach is not able to reveal the meaning of the observed Ar age pattern of the Sopron Mts. We consider two major factors controlling the arrangement of Ar ages of micas: i) The growth of sheet silicates along highly deformed zones, which were the major fluid transporting pathways, and minor mineral growth and only very minor Ar resetting within the blocks with less strain and dry metamorphic conditions. ii) The Late Cretaceous extensional tectonics responsible for displacements of metamorphic slabs having different grades of post-Variscan Ar resetting and Alpine mineral growth.

i) Mineral growth

Török (1996, 1998) recognised the different crystal chemistry of the texturally older muscovites and the younger phengitic overgrowths which was supposed to be Alpine. According to the Ar chronology it became evident that the phengitic mica formed in the latest phases of Eoalpine metamorphism. Eoalpine cooling in the Austroalpine basement usually started around 100 Ma and the youngest mica Ar ages are ca. 70 Ma (see compilations in *Frank* et al., 1987; *Stüwe*, 1998; *Thöni*, 1999). The Alpine eclogite facies metamorphism affected the basement of Koralpe-Saualpe around 100 ± 10 Ma (*Thöni* and *Jagoutz*, 1992; *Thöni* and *Miller*, 1996; Fig. 1a). The exhumation of the eclogitic rocks in the Koralpe-Saualpe started already about 90 ± 3 Ma (*Thöni*, 1999). The syn-tectonic phengite growth during high-P metamorphism of the Sopron basement (*Török*, 1996, 1998) occurred considerably later, i.e. around 76–71 Ma. The significant age difference of metamorphism of the Austroalpine subunits indicates the multiphase and migrating character of the Eoalpine metamorphism. The basement of the Sopron Mts. belongs to those crystalline slabs, which reached the HP metamorphic conditions in the latest phase of the Eoalpine cycle.

Several mica fractions yielded pre-Late Cretaceous Ar ages. Three biotite fractions of the BK (Brennberg, Kovách trench) samples (Fig. 1b and Table 1) have Variscan ages (328–319 Ma with 93 to 96% radiogenic Ar content). These samples contain well-preserved pre-Alpine mineral assemblages and textures (*Török*, 1999); thus these Ar ages can be interpreted as the pristine record of Variscan metamorphism. Another sample from the BK locality, three from the Br outcrops (Brennberg, see Table 1) and several Ar/Ar and Rb–Sr data in *Draganits* (1996), yielded Permo-Triassic K/Ar and Ar/Ar ages (Fig. 4d, f). These biotite ages (from 274 to 200 Ma) reflect the complete or partial Ar loss during the Permo-Triassic HT-LP metamorphism of the Eastern Alps (*Schuster* et al., 1999).

Most problematic is the interpretation of the ages between 200 and 110 Ma, namely between the Permo-Triassic and the Eoalpine thermotectonic events. Early Cretaceous ages are known from the Eastern Alps, but they occur mainly in the Upper Austroalpine nappe system in lower grade metamorphosed Palaeozoic sequences (*Frank* et al., 1987). These intermediate Ar ages can be created by 1) partial Ar loss from Variscan or Permian ages during Eoalpine overprint; 2) by the physical mixing of lamellae of early and late mica generations or 3) by a Mesozoic tectonothermal event. We suggest that mixing of phases is the main reason for these "intermediate" ages. For example, the Ar/Ar sample presented in Fig. 4a shows composite crystal chemistry (*Villa* et al., 1997), whereas the similarly coarse sample in Fig. 4c indicates better homogeneity. In the latter sample the partial reset of an old, pre-Alpine, coarse phase may have played some role. We cannot exclude the presence of a Jurassic-Early Cretaceous resetting period (*Vance*, 1999; *Dunkl* et al., 1999) but the study area is too complex to trace such a weak event.

– How have the pre-Alpine ages been preserved?

The Alpine high-pressure metamorphism created variable amounts of phengite, Ca–Fe–Mn garnet, biotite, K-feldspar, albite, kyanite and clinozoisite (*Lelkes-Fevári* et al., 1984; *Török*, 1998). However, in some crystalline blocks the Variscan mineral assemblages are still perfectly preserved (*Lelkes-Felvári* et al., 1984; *Kisházi* and *Ivancsics*, 1985; Török, 1999), and the intrusive textures of some granitoid bodies have also suffered only negligible deformation (*Kisházi* and *Ivancsics*, 1989; *Draganits*, 1998). The Alpine transformation is very variable, rarely complete. Even the leucophyllites contain inherited, big mica fish of muscovitic composition, although this highly ductile lithology was formed along shear zones with strong chemical and mineral transformations (*Demény* et al., 1997).

Török (1998) considered that "the high-pressure metamorphism occurred in a fluid-rich environment". We disagree and argue that the above outlined heterogeneity of Ar ages, the preservation of pre-Alpine mineral assemblages and the often observed disequilibrium phenomena can only be the consequence of a rapid metamorphic evolution. We conclude that the metamorphic fluids were able to trigger significant new mineral growth only along the major shear zones. Similar incomplete transformations are known from other HP terranes: *Ch. Miller* (pers. comm.) reported transition from gabbro to eclogite within 10 cm distance from the Koralpe (*Miller* and *Thöni*, 1997). From the Caledonides *Robinson* (1991) reported HT metamorphic assemblages not showing any mineral trasformations during subsequent HP overprint due to the lack of deformation and fluids.

The closeness of the average of zircon fission track ages (69 ± 6 Ma, see later) to the K/Ar and Ar/Ar phengite ages indicates that the post-HP thermal relaxation period was extremely short. Our interpretation of the observed features is summarized in Fig. 5. This time/event plot distinguishes fission track thermochronologic data, mineral growth and resetting through diffusion of Ar. The known thermotectonic events and the interpretation of the ages are noted below the plot. The Variscan mica formation is well documented. Later, during the Permo-Triassic HT metamorphism a few of the studied samples have lost the Carboniferous biotite Ar ages either by thermal diffusion or by new growth of biotite. The Alpine metamorphism is interpreted as a short-lived overprint and the HP event was followed by rapid cooling. According to our data a complete Ar reset due to thermal diffusion is neither detectable in the white micas nor in the biotite samples.

– What kind of methodical consequences can be drawn from this data set?

Different authors using basically different methods published similar temperature and pressure estimates for the Alpine metamorphism of the crystalline rocks in

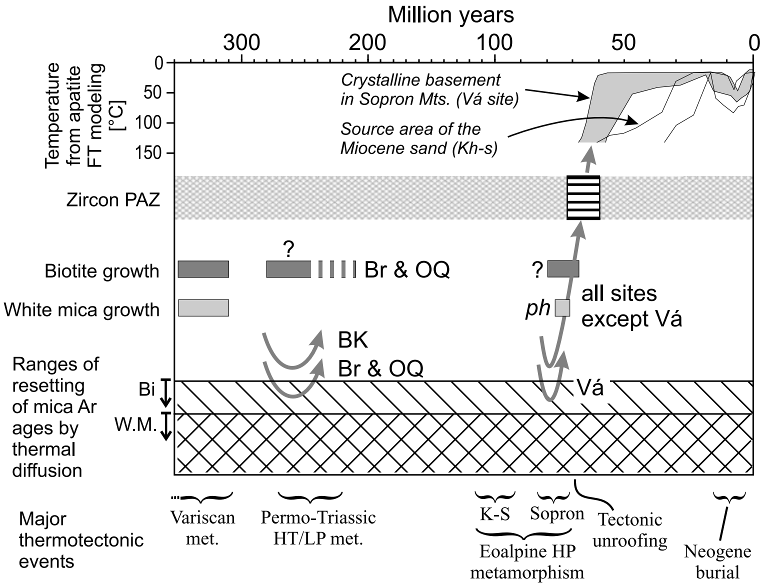

Fig. 5. Evolution of the basement of Sopron Mountains deduced from argon and fission track geochronology. The horizontal axis represents the time. Along the vertical axis four relevant factors are presented: (i) the temperature of the 0–130 °C interval deduced from apatite FT thermochronology, (ii) the zircon partial annealing zone (PAZ) and the time, when the basement of the Sopron Mountains passed this temperature range, (iii) growth of biotite and white mica and (iv) the range of thermal reset of mica Ar ages. It is not possible to present the complete path through the Variscan and Alpine evolution, only some segments of the history are known. Thus, the *darker and lighter gray blocks* represent mica formation in the different sites; the *gray arrows* approaching and entering the biotite reset temperature show the highest temperature periods reached by the different blocks and also the lack of thermal resetting of Ar ages during the metamorphic events in some sites; and the sub 130 °C *thermal paths* refer to the post-metamorphic thermal history in the shallow crust. *Bi* biotite; *W.M.* white mica; *ph* phengitic composition; *K-S* Koralpe-Saualpe; *Br* Brennberg, side valley; *BK* Brennberg, Kovách trench; *Kh-s* Kópháza sand; *OQ* Oromvég quarry; *Vá* Váris

the Sopron Mountains: *Török* (1996, 1998, 2001): 450–600 °C, 1200–1400 MPa; *Demény* et al. (1997): 560 ± 30 °C, ~1300 MPa; *Draganits* (1998): 550 ± 30 °C, 950 ± 150 MPa. These data and the incomplete Ar reset of Variscan muscovites even in the Alpine shear zones characterised by considerable bulk chemical changes would imply that the closure temperature given by *Purdy* and *Jäger* (1976) and adopted in the compilation of *Harland* et al. (1990) underestimate the Ar reset temperature in a dry metamorphic environment. Taking examples only from the neighbouring Alpine metamorphic formations, *Dallmeyer* et al. (1998) concluded that the greenschist facies metamorphism (300 to 450 °C) was "generally insufficient to reach white mica closure temperatures of K/Ar and Rb–Sr isotopic systems". *Müller* et al. (1999) observed similar, "paradox", unreset Variscan Ar ages in the Wechsel nappe which underwent Alpine greenschist facies metamorphism. They noted that in the less deformed polymetamophic rocks the "metamorphic overprint virtually caused no rejuvenation of Variscan mineral ages". In the Caledonides the metamorphic temperature estimated to be about 400 °C created a new biotite generation, but the K/Ar and Rb/Sr ages of the older

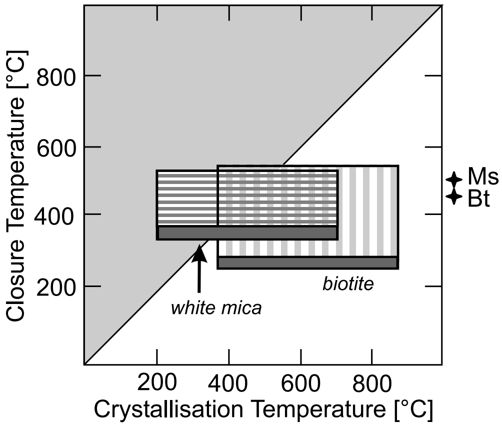

Fig. 6. Plot showing the relation of mineral growth and closure temperatures of white mica and biotite (after *Cliff*, 1993, with simplifications). The horizontal dark gray bars are taken from *Cliff* (1993) and represent the crystallisation and closure temperatures of micas according to *Harrison* et al. (1985) and *Purdy* and *Jäger* (1976) for biotite and white mica, respectively. The vertically and horizontally hatched rectangle areas were drawn according to the negligible and partial Ar reset for biotite and muscovite at ca. 550 °C metamorphic temperature recorded in our study. Stars at right indicate the closure temperature estimations of *Villa* (1998) for muscovite (Ms) and biotite (Bt)

biotites have not changed (*Verschure* et al., 1980). *Villa* (1998) argued also for the need of re-consideration of the widely used, too low closure temperatures. He placed the muscovite and biotite closure temperature to 500 and 450 °C, respectively (see discussion above). Still, we must not forget that the higher closure temperatures advocated by *Villa* (1998) are not yet supported by laboratory experiments and the coexistence of phengite and muscovite in the Sopron crystalline schists shows that equilibrium was not reached during the Alpine metamorphism.

Cliff (1993) designed an impressive plot showing the relation of mineral formation and closure temperatures. However, following the above mentioned considerations this plot must be modified (Fig. 6). The significantly higher closure temperature estimates would mean a striking limitation on the application of the biotite and white mica Ar chronometers as "thermochronometer of 350 °C".

ii) Structural control

We cannot neglect the fact that the Austroalpine crystalline basement in the eastern Eastern Alps and also in the Sopron Mts. is frequently crosscut by flat-lying ductile shear zones (*Kisházi* and *Ivancsics*, 1985; *E. Draganits* and *L. Csontos*, pers. comm.). The tectonic interpretation of the mylonitic and leucophyllitic shear zones is different from author to author. *Dallmeyer* et al. (1998) interpreted the Late Cretaceous Ar/Ar ages of the shear zones as the record of a sequence of thrusting, whereas *Ratschbacher* et al. (1989) related the stretching lineations to a late orogenic extensional event. We favour a Late Cretaceous (ca. 70–65 Ma) tectonic unroofing to explain the observed rapid cooling. On the other hand, the Sopron

Micaschist and Sopron Gneiss Formations record slightly different metamorphic histories. Both contain Alpine phengite, but the bulk transformation ratio is weaker in the schists than in the orthogneiss bodies, although the plasticity and permeability of the former lithologies are higher. The neighboring crystalline slices were displaced along the (partly leucophyllitic) mylonitic zones. This transport postdates the climax of Alpine metamorphism, because the leucophyllites were formed "after the peak of the Alpine high-pressure metamorphism" (*Török*, 2001). Thus, we conclude that the extensional phase controls the Ar ages in the shear zones. The whole pile of partly reset Austroalpine crystalline slices of the Sopron Mountains can be considered as an extensional allochton formed during Late Cretaceous time.

Fission track chronology – detection of post-metamorphic thermal history

The mica Ar geochronometers register only the mineral growth and the higher temperature interval of the thermal evolution of the crystalline basement. The post-Cretaceous exhumation history can be traced by zircon and apatite fission track thermochronology. The description of the experimental procedure can be found in *Dunkl* et al. (2001) and the results are listed in Table 2.

Zircon FT results

Only the orthogneiss samples (Vá and Nm) and the Kőbérc mica schist (OQ site) contain zircons in proper amounts. The zircon crystals have rather high spontaneous track densities; the majority of the samples was not measurable. The apparent ages have an average of 69 ± 6 Ma.

Apatite FT results

The apparent FT ages show a slight scatter; the average of the apparent ages in the crystalline samples is around 50 Ma (Table 2). The chemical composition of the apatite has an impact on the thermal stability of the fission tracks (*Green* et al., 1986). Thus we have determined the chlorine and fluorine contents of some representative samples by the electron microprobe of the Laboratory for Geochemical Research (Budapest). The apatites in the orthogneiss samples are nearly pure fluorapatite ($\sim$0.03 wt% Cl), but the substitution by chlorine in the Kőbérc micaschist ($\sim$0.3 wt%) has a definite effect on the track stability, and it was considered in the thermal modeling. The low relief of the area does not allow an age/elevation evaluation of apatite results (*Wagner* et al., 1977).

The uranium distribution within the crystals is very heterogeneous at the locality Nádormagaslat. In all the Nm samples the apatite grains have a low-uranium rim with <20 ppm U, whereas the cores (and also the bulk of apatites of the other gneiss samples) contain $\sim$70 ppm U. This overgrowth probably formed during a post-metamorphic hydrothermal stage, because the margins of the crystals are very rich in fluid inclusions (Fig. 7).

The youngest apatite FT age was measured in the detrital grains of a Miocene sandstone sample from Kópháza (Table 2). This fanglomeratic, unsorted, partly

Table 2. *Fission track results obtained on samples from Sopron Mts*

Locality	Code	Lithology	Cryst.	Spontaneous ρs (Ns)	Induced ρi (Ni)	Dosimeter ρi (Ni)	$P(\chi^2)$ [%]	FT age* [Ma ± 2s]	Track length (n)** [μm]
Apatite ages									
Város	Vá-1	gneiss	15	23.9 (880)	32.0 (1181)	4.197 (4254)	27	58 ± 3.1	14.1 ± 1.0 (50)
Város	Vá-2	gneiss	27	29.1 (2160)	49.6 (3673)	4.117 (4254)	<1	46 ± 2.3	
Város	Vá-3	gneiss	14	20.9 (720)	26.0 (896)	4.199 (4254)	61	62.6 ± 3.5	14.1 ± 1.0 (50)
Város	Vá-4	gneiss	15	17.8 (656)	23.0 (850)	4.201 (4254)	98	60.2 ± 3.5	14.0 ± 1.1 (50)
Város	Vá-5	gneiss	15	20.8 (766)	30.8 (1134)	4.186 (4254)	17	52.3 ± 3.1	13.9 ± 0.8 (50)
Város	Vá-6	gneiss	8	20.3 (504)	30.0 (744)	4.186 (4254)	6	53.4 ± 4.2	
Nádormagaslat	Nm-1	gneiss	30	4.68 (1011)	8.32 (1798)	4.188 (4254)	<1	43.9 ± 3	14.0 ± 0.8 (50)
Nádormagaslat	Nm-2	gneiss	23	7.92 (688)	10.5 (909)	4.19 (4254)	10	57.9 ± 3.7	13.9 ± 1.1 (50)
Nádormagaslat	Nm-3	gneiss	27	9.15 (954)	15.7 (1643)	4.192 (4254)	3	44.9 ± 2.6	13.5 ± 1.0 (50)
Nádormagaslat	Nm-4	gneiss	23	6.81 (813)	12.0 (1436)	4.195 (4254)	<1	42.1 ± 3.6	13.8 ± 1.2 (50)
Deákkút	De	gneiss	19	20.1 (1296)	25.1 (1616)	3.988 (9969)	1	59.6 ± 3.4	13.5 ± 1.0 (34)
SR-1 104.2m	SR	gneiss	25	9.46 (1205)	14.0 (1788)	3.988 (9969)	14	50.1 ± 2.4	13.7 ± 1.1 (45)
Kőbérc	OQ	gneiss	23	10.6 (597)	16.6 (937)	4.182 (4254)	17	49.7 ± 3.2	14.4 ± 1.0 (50)
Vöröshidi	Vö	micaschist	21	2.11 (507)	3.63 (874)	4.184 (4254)	19	45 ± 3.1	14.0 ± 1.1 (50)
Bánfalva	Sb	leucoph.	28	2.59 (615)	9.23 (2194)	7.89 (12961)	6	40.9 ± 2.5	14.3 ± 0.7 (9)
Kópháza	Kh-g	gneiss	25	8.06 (902)	29.0 (3251)	7.89 (12961)	<1	42.3 ± 2.6	13.9 ± 1.0 (24)
Kópháza	Kh-w	gneiss	9	5.16 (197)	16.4 (627)	7.89 (12961)	75	46.1 ± 3.9	
Kópháza	Kh-s	sand	60	4.94 (2015)	9.12 (3720)	3.41 (2189)	<1	33.6 ± 1.8	13.3 ± 1.5 (50)
Zircon ages									
Város	Vá-1 + 3	gneiss	13	73.7 (1406)	22.8 (434)	1.216 (1997)	1	72.4 ± 6.1	
Város	Vá-4	gneiss	17	60.5 (1374)	18.5 (421)	1.219 (1997)	3	71.1 ± 5.3	
Város	Vá-5	gneiss	20	63.2 (1572)	22.2 (552)	1.224 (1997)	48	63 ± 3.6	
Nádormagaslat	Nm-1	gneiss	16	70.0 (1668)	22.3 (524)	1.222 (1997)	22	70.5 ± 4.1	
Nádormagaslat	Nm-2 + 3	gneiss	10	69.9 (928)	26.7 (354)	1.215 (1997)	11	57.6 ± 4.8	
Nádormagaslat	Nm-4	gneiss	9	53.6 (611)	16.7 (190)	1.213 (1997)	59	70.5 ± 6.2	
Kőbérc	OQ	gneiss	17	89.7 (2126)	17.7 (419)	0.822 (1858)	<1	77.5 ± 8	

Cryst: number of dated zircon crystals
Track densities (ρ) are as measured ($\times 10^5$ tr/cm^2); number of tracks counted (N) shown in brackets
$P(\chi^2)$: probability obtaining Chi-square value for n degree of freedom (where n = no. crystals-1)
* Central ages calculated using dosimeter glass: NBS 962 with $\zeta_{zircon} = 363.5 \pm 6$ and $\zeta_{apatite} = 373 \pm 8$
** Track length distributions can be found: http://www.sediment.uni-goettingen.de/staff/dunkl/archive/sopron.html

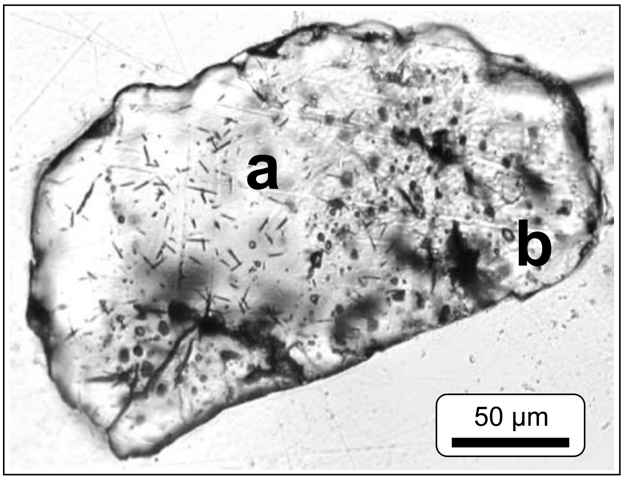

Fig. 7. In the Nádormagaslat gneiss samples the rim of the apatite crystals is poor in uranium, but rich in fluid inclusions. Photomicrograph showing this overgrowth which has probably a hydrothermal origin in a fragment of an apatite crystal; a) the clear central zone rich in spontaneous tracks (~67 ppm U), b) the marginal zone is poor in uranium (~14 ppm) and rich in fluid inclusions

very coarse, partly sandy sediment forms the immediate cover of the SGF. It contains numerous boulders and pebbles of lithologies, which are not known from the Sopron Mountains sensu stricto. Coarse garnet amphibolites and tonalitic orthogneiss can be found in the Wechsel series exposed to the north and west of Sopron (Fig. 1, *Frank* et al., 1996; *Müller* et al., 1999). The age distribution is rather compact, the majority of single grain ages are between 45 and 20 Ma. The 33 Ma average of single grain ages is considerably younger than the apparent apatite FT ages of the basement of the Sopron Mountains.

The confined track length data (Table 2) show slightly shortened track populations, but the spreads of the distributions are narrow (~1 μm).

Thermal modeling of apatite FT results

The evaluation of the confined track length distributions allows us to resolve the meaning of the apparent apatite FT ages (*Gleadow* et al., 1986). The narrow spread of track length distributions indicates rapid cooling. Computer modeling using the AftSolve software (*Ketcham* et al., 2000) supports this interpretation. For the modeling the following conditions were applied:

- *annealing model: Ketcham* et al. (2000),
- *chlorine content:* 0.03 wt% for all gneiss samples, 0.1% and 0.3% for the Vöröshid and Kőbérc mica schist samples, respectively,
- *initial track length:* 16.0 and 15.3 μm,
- *paths calculated:* up to 1 000 000.

In case of the Vá and Nm samples bulk averages of all data and amalgamated track lengths were applied as the samples were collected from small areas (see

Table 2). The modeling algorithm finds fit to only a very low proportion of the tested paths, when the input data are so well constrained (>4000 spontaneous and induced tracks; 250 and 200 track length data in Vá and Nm samples, respectively).

The initial time-temperature point for the modeling was chosen at 70 Ma and 250 °C – according to the mean of zircon FT ages. The fixed time-temperature bars for the modeling intervals were determined to give no limitation for the fitting. Beyond the "end-point" of the modeling (0 Ma, 10 °C) another "fix-point" was applied for the thermal modeling indicating that the crystalline formations were exhumed already to the surface in Miocene time (*Dunkl* and *Frisch*, 2002). The Austroalpine crystalline basement was covered during the Middle Miocene with fluvial and later with marine sediments (*Telegdi-Roth*, 1881). This erosional unconformity is well exposed around Fertörákos, Brennberg and also at Kópháza (where we sampled the basal beds of the Miocene sequence, see above). Furthermore we suppose that in the central part of the Sopron Mountains this paleo-erosional level was close to the recent surface of the crystalline formations. This reasoning provides the basis for insertion of a "fix-point" for the modeling at 17 Ma, 14 °C.

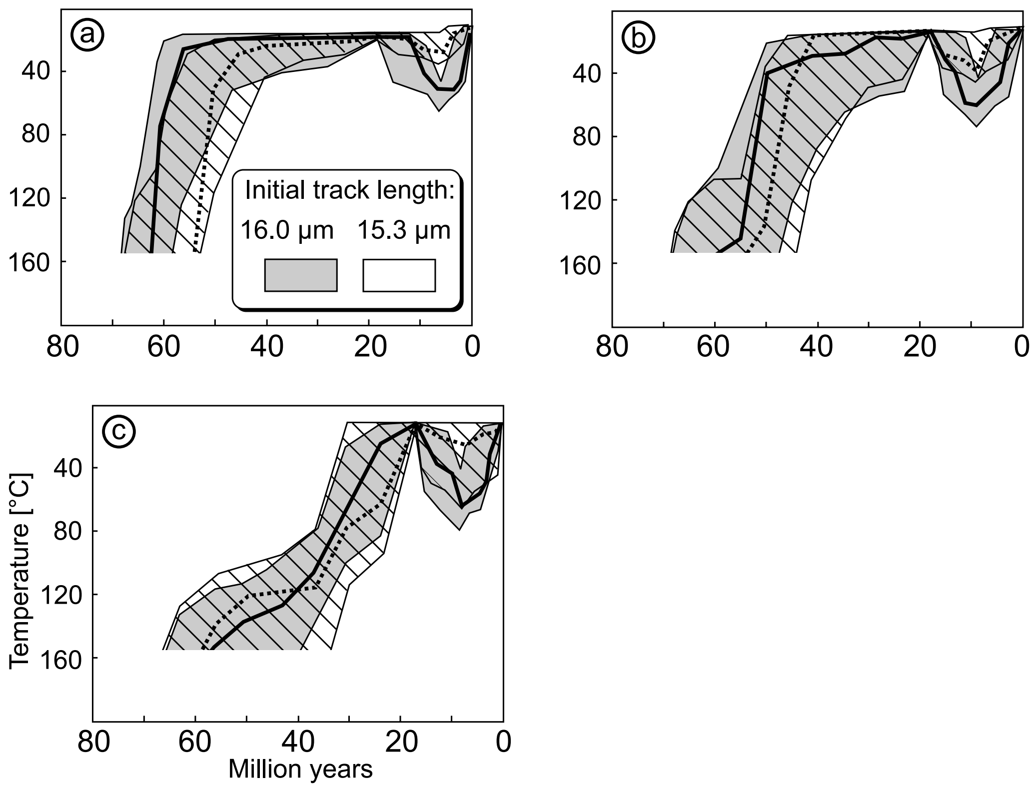

Fig. 8. Time-temperature plots showing the results of thermal modeling of apatite FT results done by AftSolve software (*Ketcham* et al., 2000). Solid and dotted lines are presenting the best-fit-runs supposing 16.0 and 15.3 μm initial track length, respectively. **a** Váris gneiss (mean age of 6 samples and all lengths together), **b** Nádormagaslat gneiss (mean age of 4 samples and all lengths together), **c** Kópháza sand (computing based on all grains except two grain ages older than 100 Ma)

This scenario assumes that the top of the crystalline basement was at surface temperature during the Middle Miocene.

For the amalgamated Vá sample the modeling yields a thermal path composed of stepwise cooling segments. The samples underwent very rapid cooling between 65 and 50 Ma (Fig. 8a). The modeling results indicate a low temperature period (near-surface stay) for the rest of the Palaeogene. It is difficult to give an exact temperature estimation for the post-Middle Miocene period. The modeling results for the period after the sediment burial depend on the assumption on the initial track length. The problem roots in the fact that the induced tracks created by neutron irradiations in apatite yield track lengths around 16 μm, whereas the thermally undisturbed volcanic formations have apatites with ca. 15 μm track length (*Gleadow* et al., 1986). This has an impact on the final part of modelled time-temperature paths. We performed the modelings using two values (16.0 and 15.3 μm) to study the effect of this factor. Supposing the greater figure the pooled Váris gneiss sample requires significant post-depositional burial heating, while using the initial value of 15.3 μm Late and post-Miocene overprint is smaller, but still exists (Fig. 8a).

To resolve this dilemma we have to consider the Miocene evolution of the region. The immediate surroundings of the Sopron Mts. is characterised by very complex basin morphology (Fig. 1b). Variable thickness of Neogene syn- and post-rift sequences are typical, and the crystalline areas are also crosscut by steep faults with over one hundred metres of vertical displacement (*Draganits*, 1996). We suppose that the major part of the basin relief is the result of late and post-Miocene fault tectonics and thick syn-rift sediments also covered the actually exposed crystalline areas. Thus, those modeling results, which indicate post-Middle Miocene burial heating, can be considered reliable.

The young fault tectonics can also have an effect on the different exhumation of the blocks within the mountains and probably resulted in the observed variation of apatite apparent ages. The modeling of the pooled results of the Nádormagaslat samples resulted in time-temperature paths basically similar to those of Váris (Fig. 8b). The good fits for the post-metamorphic cooling period give a slightly younger range and the burial-related thermal overprint is slightly more pronounced.

The basal sediments of the Miocene cover derived from an area, which had an Early Tertiary cooling history and which is different from that of the Sopron Mountains proper. This difference is in harmony with the observed exotic character of the boulders. The modeling indicates continuous cooling of the source area for the whole Tertiary period before its exhumation to the surface and sediment burial in Miocene time.

Evaluation of the fission track results

The average of the zircon FT ages (69 ± 6 Ma) can be related to the minimum age of the Eoalpine HP metamorphism (78–74 Ma). These data indicate rapid cooling during the exhumation of the Austroalpine basement following the Late Cretaceous metamorphism (Fig. 5). Similar cooling was found in the Austroalpine formations of the Fertörákos Hills, the Gleinalm dome and in the Wechsel area (*Neubauer* et al., 1995; *Frank* et al., 1996), and also the western blocks of the Austroalpine realm (*Fügenschuh* et al., 1997; *Elias*, 1998).

The Váris sample group gave the oldest apparent apatite FT ages while the Nádormagaslat samples and the leucophyllite from Bánfalva yield the youngest ones. Thermal modeling based on apatite FT results indicates a cooling history which is composed of three well distinguishable periods. The results from Váris indicate rapid cooling in the Early Palaeogene as a continuation of cooling from higher temperatures as detected by mica growth and zircon FT thermochronology. This can be related to the cessation of the exhumation by tectonic unroofing that terminated the Late Cretaceous orogeny.

The difference in the age of Palaeogene cooling between the Vá and Nm samples can be explained in three ways. i) The two sites belong to different structural blocks and had actually different cooling histories due to Late Cretaceous-Early Palaeogene extensional tectonics. ii) The difference is the result of Neogene normal faulting and thus reflects the differences in Late Miocene burial and exhumation histories. In this case the shift of Palaeogene cooling between Figs. 8a and b would be only an apparent phenomenon. iii) The Nádormagaslat and also the Bánfalva sample with the youngest apatite FT age (derived from the immediate vicinity of Nm, see Fig. 1b) was affected by post-metamorphic hydrothermal fluid activity. Although we can not exclude the first two cases, iii) is thought to be the major reason. The Nádormagaslat locality contains numerous manifestations of hydrothermal activity; the gneiss is densely penetrated by quartz veins, the biotite is fully chloritised, the accessoric apatite crystals have an overgrown rim with low uranium content, and this rim is extremely rich in fluid inclusions (Fig. 7). The multiphase hydrothermal activity of the Sopron area was described by *Török* (1996, 2001). He also observed growth of phosphate minerals in shear zones in relation with the last stage, high salinity fluids. This fluid activity was related to the retrograde phase and this interpretation fits well to the youngest apparent apatite FT age of Nádormagaslat.

Rb/Sr geochronology of the area indicates also some Eocene event. The biotite ages range from 55 to 41 Ma (*Kovách* and *Sudár-Svingor*, 1988) and *Draganits* (1996) reported a biotite Rb/Sr age of 53 Ma. There is further evidence for the hydrothermal activity in the Eastern Alps. In more western areas of the Austroalpine realm *Prochaska* et al. (1995) reported 46 Ma old K/Ar ages on $<2\,\mu m$ fractions of white mica from a hydrothermal ore deposit. It is noticeable that these data are younger than the biotite K/Ar and the zircon FT ages. The Rb/Sr results are in the range of apatite FT apparent ages, thus they can be interpreted neither as mineral growth, nor as cooling ages. Laboratory experiments show the high mobility of Rb and Sr in the crystal lattice of biotite even at low temperature in a hydrothermal environment (*Kulp* and *Engels*, 1963; *Bracke* et al., 1992).

The observed thermal stagnation during the Eocene and Oligocene agrees well with the considerations on the relative passive tectonic behaviour of the Eastern Alps during Palaeogene times (*Hejl*, 1997; *Dunkl* et al., 2001). The Late Neogene burial-related thermal overprint has already been discussed. It is rather difficult to quantify this event, but its presence indicates the latest milestone in the thermal evolution of the area. The Neogene syn- and post-rift sediments had a much broader extent and the Sopron Mountains were buried before the inversion of the western Pannonian Basin in Pliocene times (*Horváth* and *Cloething*, 1996).

Conclusions

1. Two generations of white micas occur in the gneisses of the Austroalpine basement of the Sopron Mountains. The coarser, older generation is muscovitic whereas the younger, finer-grained micas occur as overgrowth on muscovite and have phengitic composition. A characteristic Ar age difference can be observed between muscovite and phengite. The latter formed during Eoalpine metamorphism, the muscovite retained pre-Alpine Ar ages due to incomplete resetting.
2. The age of the Eoalpine high-P metamorphic overprint of the Sopron Gneiss Formation can be approximated by the lowest K/Ar ages on biotite (78.5 ± 2.9 Ma) and muscovite (76.6 ± 2.9 Ma), by the "isochron age" of feldspars (74 ± 11 Ma), and by Ar/Ar ages measured at the lowest temperature degassing steps on leucophyllite from Váris quarry (71.1 ± 5.6 Ma and 75.1 ± 4.7 Ma). The most likely time interval for the Alpine event is 76–71 Ma.
3. The high-P metamorphism of the Sopron Mts. is considerably younger than the eclogite facies metamorphism and the following cooling of the Koralpe-Saualpe further west. It throws light on the multiphase character and the temporal shift of Eoalpine metamorphism in the Austroalpine units.
4. Carboniferous biotite K/Ar ages (328.5–319.5 Ma) from the andalusite–sillimanite–biotite schist with well-preserved pre-Alpine mineral assemblages at Kovács trench, Óbrennberg, are regarded as a close approximation of the age of Variscan metamorphism. Another slab of this schist complex was affected by HT Permo-Triassic metamorphism; biotite fractions crystallised during this event yield Ar ages between 272 and 236 Ma.
5. In the kyanite–chloritoid–muscovite schist the white mica fractions gave Late Cretaceous Ar ages. During the formation of Alpine white mica the pre-Alpine biotites of the same rock bodies (mentioned in points 3 and 4) have suffered negligible Ar loss; they retained the Variscan and Permo-Triassic Ar ages.
6. The disequilibrium mineral assemblages and the parallel coexistence of pure Variscan (& Permo-Triassic) biotite ages and Alpine white mica ages in the same rock indicate that the Eoalpine metamorphic overprint was a short-lived event. The metamorphic fluids were channelled and generated predominant Alpine mineral assemblages only along shear zones. In the Sopron Micaschist Formation the manifestation of Alpine overprint is limited to the growth of white mica.
7. The published estimates on Alpine metamorphism indicate a temperature range from 450 to 600 °C using different methods/calibration systems. This temperature is higher than the highest estimation published for the closure temperature of muscovite (*Villa*, 1998) and differs from the observed partial retention of radiogenic Ar during the Alpine metamorphism.
8. Flat lying mylonitic shear zones formed after the climax of the HP metamorphism. They are interpreted as major detachment planes formed during Late Cretaceous extension. Zircon FT data (average: 69 ± 6 Ma) show that

post-metamorphic cooling was rapid; it was the consequence of tectonic unroofing. The structure of the Sopron basement is considered as an extensional allochton.

9. Thermal modeling based on apatite FT results indicates that the recently exposed level of crystalline formations reached a near-surface position already in Palaeogene times. There are traces of hydrothermal activity of probably Eocene age following the post-Eoalpine exhumation.
10. The Late Palaeogene period was dominated by stagnation. The thick Neogene sedimentary sequences, which are preserved in the basins around the Sopron Mountains, also covered the crystalline, presently exhumed mountainous areas. This sedimentary cover was removed probably only in Pliocene times.

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