

Time calibration of a P–T path from a Variscan high-temperature low-pressure metamorphic complex (Bayerische Wald, Germany), and the detection of inherited monazite

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Abstract A temperature–time path was constructed for high-temperature low-pressure (HT–LP) migmatites of the Bayerische Wald, internal zone of the Variscan belt, Germany. The migmatites are characterised by prograde biotite dehydration melting, peak metamorphic conditions of approximately 850 °C and 0.5–0.7 GPa and retrograde melt crystallisation at 800 °C. The time-calibration of the pressure–temperature path is based on U–Pb dating of single zircon and monazite grains and titanite separates, on $^{40}\text{Ar}/^{39}\text{Ar}$ ages obtained by incremental heating experiments on hornblende separates, single grains of biotite and K-feldspar, and on $^{40}\text{Ar}/^{39}\text{Ar}$ spot fusion ages of biotite determined in situ from sample sections. Additionally, crude estimates of the duration of peak metamorphism were derived from garnet zoning patterns, suggesting that peak temperatures of 850 °C cannot have prevailed much longer than 2.5 Ma. The temperature–time paths obtained for two areas approximately 30 km apart do not differ from each other considerably. U–Pb zircon ages reflect crystallisation from melt at 850–800 °C at 323 Ma (southeastern area) and 326 Ma (northwestern area). The U–Pb ages of monazite mainly coincide with those from zircon but are complicated by variable degrees of inheritance. The preservation of inherited monazite and

the presence of excess ^{206}Pb resulting from the incorporation of excess ^{230}Th in monazite formed during HT–LP metamorphism suggest that monazite ages in the migmatites of the Bayerische Wald reflect crystallisation from melt at 850–800 °C and persistence of older grains at these temperatures during a comparatively short thermal peak. The U–Pb ages of titanite (321 Ma) and $^{40}\text{Ar}/^{39}\text{Ar}$ ages of hornblende (322–316 Ma) and biotite (313–309 Ma) reflect cooling through the respective closure temperatures of approximately 700, 570–500 and 345–310 °C published in the literature. Most of the feldspars' ages (305–296 Ma) probably record cooling below 150–300 °C, while two grains most likely have higher closure temperatures. The temperature–time paths are characterised by a short thermal peak, by moderate average cooling rates and by a decrease in cooling rates from 100 °C/my at temperatures between 850–800 and 700 °C to 11–16 °C/my at temperatures down to 345–310 °C. Further cooling to feldspar closure for Ar was probably even slower. The lack of decompressional features, the moderate average cooling rates and the decline of cooling rates with time are not easily reconciled with a model of asthenospheric heating, rapid uplift and extension due to lithospheric delamination as proposed elsewhere. Instead, the high peak temperatures at comparatively shallow crustal levels along with the short thermal peak require external advective heating by hot mafic or ultramafic material.

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Introduction

Granulite-facies metamorphism and migmatite formation by partial melting in mid to upper crustal levels previously thickened by continental collision have been described from quite a few orogenic belts (e.g. De Yoreo et al. 1991 and references therein). These findings seem in apparent conflict with most commonly accepted thermal models for lithosphere thickened by collision. These models predict high-temperature metamorphism

and partial melting due to crustal stacking, increased radiogenic heat production and thermal relaxation only in lower crustal levels (Thompson and Connolly 1995 and references therein). Heating of the mid and upper crust to the temperatures observed in high-temperature low-pressure (HT–LP) orogenic belts would take some tens of millions of years. In collisional orogens where this time span is not available, additional heat is required to trigger HT–LP metamorphism and melting (De Yoreo et al. 1991; Thompson and Connolly 1995; Zen 1995). In most common models the asthenosphere is invoked as an additional heat source, brought up to the base of the crust either by convective removal of the lithospheric mantle (e.g. England and Houseman 1989), delamination of the lithosphere (Bird 1979) or by slab breakoff (Davies and von Blanckenburg 1995). Other models consider advective heating by rising magmas or pre-collisional lithospheric extension (De Yoreo et al. 1991 and references therein).

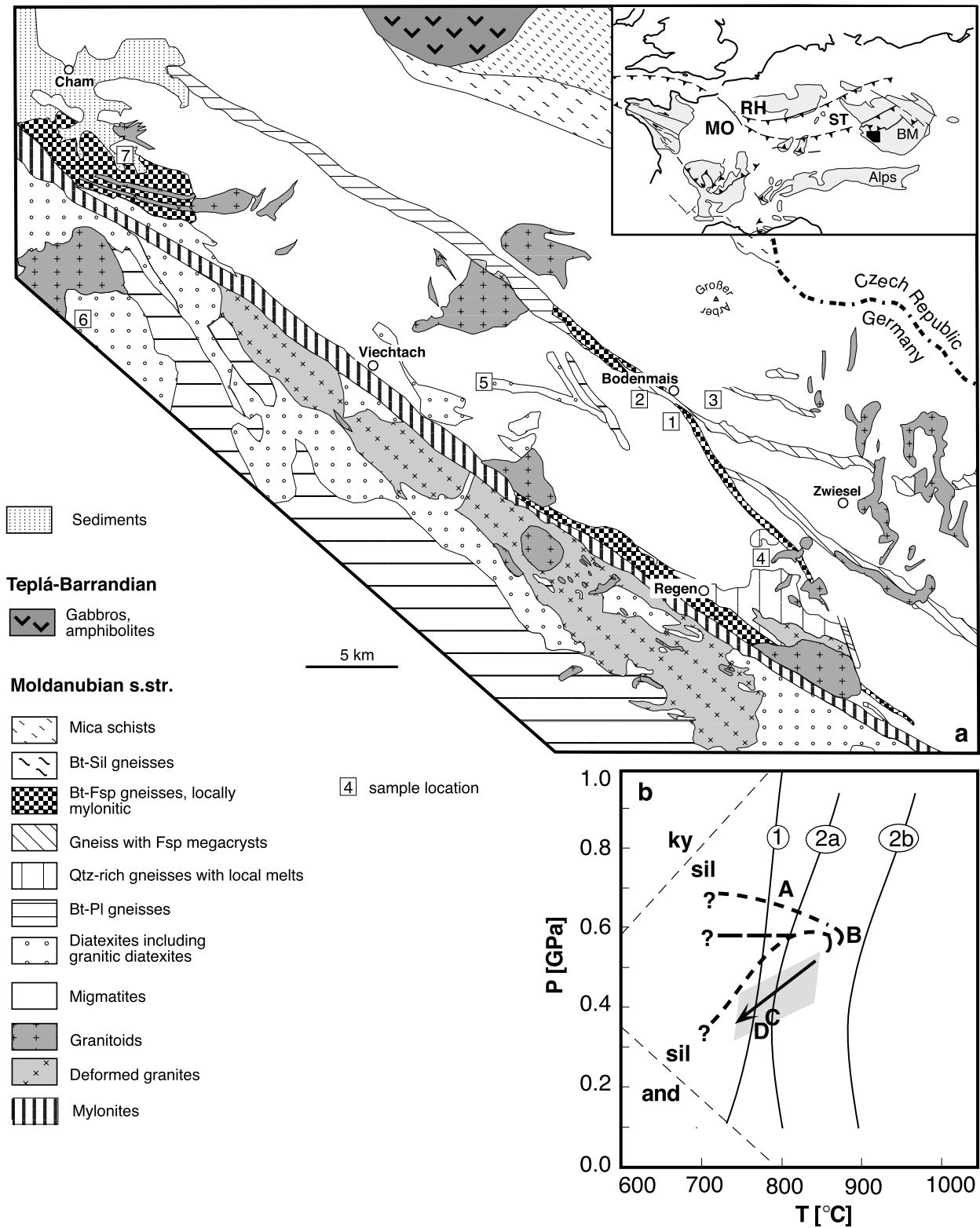
Determining the pressure–temperature–time (P–T–t) paths of metamorphic rocks in orogenic belts can be of considerable help in distinguishing the possible causes of HT–LP metamorphism. Heating by burial and subsequent thermal relaxation and exhumation of the crust result in comparatively steep, sometimes counterclockwise P–T paths of time spans of tens of millions of years. On the other hand, lithospheric or slab removal/detachment induces rapid uplift and extension of the previously thickened crust as recognised from quite a few mountain belts (e.g. Dewey 1988; Malavieille 1993; Ruppel 1995). In these cases, decompression (exhumation) of metamorphic rocks is accompanied by heating (Platt and England 1994). Heating during decompression and rapid exhumation of metamorphic rocks with cooling rates of > 100 to 500 °C/my have been described from a number of extensional settings (e.g. Kohn et al. 1993; van der Wal and Vissers 1993; García-Casco 1996; Soto and Platt 1999; Zeck and Whitehouse 1999). An important boundary condition for this mechanism is a considerable time lapse (at least 10 my; Platt et al. 1998) between lithospheric thickening and the onset of extension. Heating by underplating or rise of magmas induces P–T paths comparable to those typical of contact metamorphism, characterised by an almost isobaric temperature increase and slow isobaric cooling.

Within the Variscan belt of Europe, remnants of an HT metamorphic belt can be traced from the Bohemian Massif in the east via Schwarzwald, Vosges, Massif Central and the Armorican Massif to the Iberian Massif in the southwest (e.g. Matte 1986). Most of the belt represents fairly shallow crustal levels consisting of gneisses, migmatites and amphibolites. Granitoid plutons are abundant. The P–T conditions recorded by these rocks require an external heat source for HT–LP metamorphism (e.g. Le Métour 1978; Latouche et al. 1992; Brown and Dallmeyer 1996; Kalt et al. 1999). Late orogenic extension has been recognised from large-scale structures within the Variscan belt (e.g. Ménard and Molnar 1988; Costa and Rey 1995). However, within

Fig. 1a,b Variscan basement outcrops in central Europe, simplified geological map of the Bayerische Wald with sample locations, and P–T path of the migmatites. **a** Simplified geological map of the Bayerische Wald, modified after Kalt et al. (1999). Sample locations of migmatites, gneisses and amphibolites are indicated by numbers from 1–7 (compare Table 1). The *inset* shows Variscan basement outcrops in central Europe. *BM* Bohemian Massif; *MO* Moldanubian zone; *ST* Saxothuringian zone; *RH* Rhenohercynian zone. The *MO* and *ST* defined by Kossmat (1927) represent the internal part of the Variscan belt, characterised by high-grade metamorphism and plutonism. The *black rectangle* indicates the area shown in the map. **b** P–T path for the migmatites of the Bayerische Wald, modified after Kalt et al. (1999). *1* Experimentally determined minimum temperatures and pressures for dehydration melting of metapelites in the absence of aqueous fluids by the reaction biotite + sillimanite + plagioclase + quartz = garnet + K-feldspar + melt (Le Breton and Thompson 1988). *2a* Experimentally determined minimum temperatures and pressures for dehydration melting of metagreywackes in the absence of aqueous fluids by the reaction biotite + plagioclase + quartz = garnet + orthopyroxene + K-feldspar + melt (Vielzeuf and Montel 1994); *2b* biotite-out curve at higher temperatures in the same experiments. *2a* is a minimum temperature for migmatites of the Bayerische Wald. The biotite-out curve marks maximum temperatures for the migmatites as textures and biotite compositions indicate that biotite was not exhausted during partial melting. *A*, *B*, *C* and *D* refer to metamorphic stages as described in the section on geological setting, petrological context and geochronological background and in Kalt et al. (1999). The *shaded area* marks the results of thermobarometry, indicating equilibration down to stage *D*.

some of the HT–LP metamorphic areas, compressional structures often dominate (e.g. Tanner and Behrmann 1995; Behrmann and Tanner 1997) and evidence for considerable decompression from phase assemblages and thermobarometry is often lacking (e.g. Latouche et al. 1992; Kalt et al. 1999). HT–LP metamorphism has been dated at approximately 330–310 Ma B.P., depending on the location within the elongate belt (e.g. Grauert et al. 1974; Pin and Peucat 1986; Kalt et al. 1994a; Brown and Dallmeyer 1996). Eclogites contained in the HT–LP rocks occur as isolated exotic bodies of variable size. In the Schwarzwald, they record a high-pressure event at approximately 345–332 Ma (Kalt et al. 1994b; Kalt et al. 1997), suggesting that the time lapse between lithospheric thickening and the onset of heating may be small. Hence, at least for some parts of the Variscan HT–LP metamorphic belt, heating may have occurred during crustal thickening and may not have been accompanied by extension.

The Bayerische Wald in the Bohemian Massif (Fig. 1) forms part of the Variscan HT–LP metamorphic belt. The P–T path of the migmatites is characterised by prograde dehydration melting, by granulite-facies peak metamorphic conditions and by the absence of significant decompression during heating (Kalt et al. 1999; Fig. 1). The purpose of this investigation is to time-calibrate the P–T path of the migmatites in order to constrain boundary conditions for, and to distinguish between, various causes for HT–LP metamorphism. Due to the high peak temperatures, information on the prograde part of the P–T path is not to be expected. In



order to constrain the cooling history of the migmatites from peak metamorphism to low temperatures, the following dating techniques were applied: U-Pb dating of zircon, monazite and titanite, $^{40}\text{Ar}/^{39}\text{Ar}$ dating of hornblende, biotite and K-feldspar. In order to obtain

high-precision data and to be able to link them to petrological information, single grains were dated wherever possible. In order to test possible regional variations in the cooling path, various localities several kilometres apart were sampled.

Geological setting, petrological context and geochronological background

Most of the Bohemian Massif forms part of the internal Variscan HT belt, the Moldanubian and Saxothuringian zones according to the subdivision by Kossmat (1927; Fig. 1). Apart from the Teplá-Barandian unit, the Moldanubian zone is characterised by high-grade metamorphism and widespread granitoid intrusions (Fig. 1). It has a complex structure with nappes of HT–HP granulites, peridotites and eclogites thrust over HT–LP gneisses and migmatites. These gneisses and migmatites proper are devoid of MP or HP relics and only contain very rare exotic lenses of eclogite.

The Bayerische Wald (Fig. 1), located at the southwestern margin of the Bohemian Massif, comprises only the lower unit of HT–LP migmatites and gneisses, intruded by granitoid plutons of variable size. Most migmatites are of semipelitic to pelitic bulk composition and are characterised by the assemblages biotite + quartz + plagioclase + K-feldspar + cordierite + ilmenite. Additional garnet, spinel, orthopyroxene and sillimanite occur depending on bulk rock composition. The P–T path is most likely clockwise (Fig. 1) and can be divided into four stages. Apart from a first stage A only preserved in garnets of a few samples, the further P–T path of all migmatite samples is characterised by melt-producing biotite dehydration reactions in the absence of an aqueous fluid phase, peak conditions of approximately 850 °C/0.5–0.7 GPa and cooling (stage B) until the partial melts formed at stage B crystallised (stage C). Subsequent stage D, during which mineral equilibria were frozen (770–846 °C/0.44–0.51 GPa), is characterised by decompression and cooling. The amphibolites are included as bodies of variable size in the migmatites and show relic assemblages of either plagioclase + clinopyroxene + titanite or plagioclase + clinopyroxene + orthopyroxene + quartz + ilmenite, recording a granulite-facies stage with the same peak temperatures as those estimated for the migmatites.

Previous geochronological studies have shown HT–LP metamorphism to be of Carboniferous age. U–Pb ages of monazite grain-size fractions from migmatites of the Bayerische Wald range from 317 to 321 Ma (Grauert et al. 1974), similar to those obtained on metamorphic rocks of the Moldanubian zone in the adjacent Oberpfälzer Wald (317 ± 3 – 323 ± 3 Ma; Teufel 1988). K–Ar ages of biotite grain-size fractions from various metamorphic rocks of the Bayerische Wald scatter between 325 ± 5 and 309 ± 5 Ma (Kreuzer et al. 1989). For the granites intruding the migmatites, Rb–Sr mineral ages range from 302 ± 7 to 322 ± 5 Ma, and K–Ar mineral ages scatter between 296 ± 3 and 320 ± 8 Ma (compilation in Siebel 1998).

Sample selection

The aim of the study was to establish temperature–time paths for HT–LP migmatites and gneisses of the Bayerische Wald by U–Pb dating of zircon, monazite and titanite and $^{40}\text{Ar}/^{39}\text{Ar}$ dating of amphibole, biotite and K-feldspar. A temperature–time path ideally requires that all chronometers used should be applied to one outcrop or restricted area. This approach was limited in the case of the Bayerische Wald by the scarce occurrence of amphibolites and amphibole-bearing migmatites and by the fact that most of the latter were found to be hydrothermally overprinted. Therefore, amphibole dating could be performed only at a few outcrops (Table 1, Fig. 1). However, a larger number of localities (Table 1, Fig. 1) had to be selected for dating in order to detect possible regional variations in the cooling path. The sampling sites can be roughly grouped into a southeastern area around Bodenmais (Fig. 1), comprising locations 1–5 and a northwestern area around Cham (Fig. 1), comprising location 7 (and 6).

Two rock types were sampled for dating: (1) migmatites and one gneiss for U–Pb dating of zircon and monazite and $^{40}\text{Ar}/^{39}\text{Ar}$ dating of biotite and K-feldspar, and (2) amphibolites for U–Pb dating of titanite and $^{40}\text{Ar}/^{39}\text{Ar}$ dating of amphibole and in one case also $^{40}\text{Ar}/^{39}\text{Ar}$ dating of biotite. Phase assemblages, melting reactions, melt fractions and microstructures of the migmatites are described in detail in Kalt et al. (1999) and Berger and Kalt (1999). The amphibolites are fine- to medium-grained rocks with gneissose textures. Foliation is well pronounced and defined by amphibole and biotite (where present). Samples BW-112 and BW-132 contain quartz, clinopyroxene, orthopyroxene, biotite and accessory ilmenite and apatite in addition to amphibole and plagioclase. Pyroxene inclusions in some amphiboles indicate that amphibole grew at the expense of pyroxenes. However, the coexistence of unreacted pyroxenes with amphibole, both displaying stable grain boundaries, indicates equilibrium between the minerals. Sample BW-134 shows compositional banding parallel to foliation. Amphibole- and plagioclase-rich layers alternate with clinopyroxene- and plagioclase-rich layers. Sample BW-118 contains amphibole, plagioclase and quartz in textural equilibrium. Samples BW-134 and BW-118 contain accessory titanite.

Characteristics of the phases and sample preparation

The compositions of biotite, K-feldspar and amphibole from the samples used here for dating are listed in Table 2 (for analytical procedure see Kalt et al. 1999).

Zircon is present as accessory phase in the migmatites. It occurs interstitially in mesosomes and leucosomes and may also be enclosed in garnet and biotite. In thin section, zircons display anhedral to euhedral shapes and in cases zoning under crossed polars.

Table 1 Sample characteristics

Location ^a	Sample	Lithology	Phase assemblage ^b	Dating technique ^c
1	BW-22	Migmatite	crd + grt + bt + pl + qtz	⁴⁰ Ar/ ³⁹ Ar IH biotite
	BW-28	Migmatite	crd + grt + bt + Kfs + pl + qtz + sil + spl	⁴⁰ Ar/ ³⁹ Ar SF biotite
	BW-70L	Leucosome	Kfs + pl + qtz + bt	U-Pb zircon, U-Pb monazite
	BW-70M	Mesosome	crd + grt + bt + Kfs + pl + qtz + sil + spl	U-Pb monazite
2	BW-30	Migmatite	crd + grt + bt + pl + qtz	⁴⁰ Ar/ ³⁹ Ar IH biotite
	BW-32	Migmatite	crd + grt + bt + Kfs + pl + qtz	⁴⁰ Ar/ ³⁹ Ar SF biotite
	BW-67	Migmatite	crd + grt + bt + Kfs + pl + qtz + sil + spl	⁴⁰ Ar/ ³⁹ Ar SF biotite, ⁴⁰ Ar/ ³⁹ Ar IH K-feldspar
3	BW-86246	Gneiss	crd + grt + bt + pl + qtz + mt + spl	⁴⁰ Ar/ ³⁹ Ar IH biotite
4	BW-132	Amphibolite	hbl + cpx + opx + pl + qtz + bt + ilm	⁴⁰ Ar/ ³⁹ Ar IH hornblende, ⁴⁰ Ar/ ³⁹ Ar IH biotite
	BW-134	Amphibolite	hbl + cpx + pl + ttn	U-Pb titanite, ⁴⁰ Ar/ ³⁹ Ar IH hornblende
5	BW-118	Amphibolite	hbl + pl + qtz + ttn	U-Pb titanite, ⁴⁰ Ar/ ³⁹ Ar IH hornblende
	BW-120	Migmatite	crd + bt + Kfs + pl + qtz	U-Pb monazite, ⁴⁰ Ar/ ³⁹ Ar IH biotite, ⁴⁰ Ar/ ³⁹ Ar IH K-feldspar
6	BW-112	Amphibolite	hbl + cpx + opx + pl + qtz + bt + ilm	⁴⁰ Ar/ ³⁹ Ar IH hornblende
	BW-116	Migmatite	crd + bt + Kfs + pl + qtz + sil	U-Pb monazite, ⁴⁰ Ar/ ³⁹ Ar IH biotite, ⁴⁰ Ar/ ³⁹ Ar IH K-feldspar
7	BW-44	Migmatite	crd + grt + bt + Kfs + pl + qtz + sil	⁴⁰ Ar/ ³⁹ Ar IH biotite, ⁴⁰ Ar/ ³⁹ Ar SF biotite, ⁴⁰ Ar/ ³⁹ Ar IH K-feldspar
	BW-46	Migmatite	crd + grt + bt + Kfs + pl + qtz + spl + sil	⁴⁰ Ar/ ³⁹ Ar IH biotite, ⁴⁰ Ar/ ³⁹ Ar IH K-feldspar
	BW-90L	Leucosome	Kfs + pl + qtz + bt	U-Pb zircon, U-Pb monazite
	BW-90M	Mesosome	crd + grt + bt + Kfs + pl + qtz + spl + sil	U-Pb monazite, ⁴⁰ Ar/ ³⁹ Ar SF biotite

^a Location numbers correspond to those given in Fig. 1

^b Mineral abbreviations according to Kretz (1982)

^c *IH* incremental heating experiments on single minerals or mineral separates with a laser, *SF* spot fusion experiments on minerals in

thin sections with a laser. In addition to the minerals indicated, migmatites may contain accessory zircon, monazite, apatite, ilmenite, graphite, pyrrhotite and pyrite. Amphibolites may bear additional zircon, ilmenite and apatite

For the present study, only zircons formed during HT–LP metamorphism were of interest. Hence, zircons were only selected from leucosomes where the chance of being newly crystallised from melt during HT–LP metamorphism is high, whereas in mesosomes the probability of zircons containing inherited radiogenic lead or older zircon domains is much larger due to the restitic character of some mesosomes (Kalt et al. 1999). Under the binocular microscope, zircons partly showed core-rim structures. In order to avoid inheritance problems the analyses were carried out only on rim components (prisms and tips).

As is the case with zircon, monazite is an accessory phase in migmatites where it occurs interstitially and forms inclusions in garnet and biotite. Monazites are anhedral to platy crystals that partly display colour zoning under crossed polars. Though much less often than zircons, monazites may contain inherited radiogenic lead (e.g. Copeland et al. 1988), possibly in the form of an inherited monazite phase (Harrison et al. 1995). The case was described by Teufel (1988) for gneisses of the Oberpfälzer Wald. Therefore, in migmatites with clear leucosome–mesosome boundaries (BW-70, BW-90, Table 1), mesosomes and leucosomes were carefully separated before crushing. In one mesosome (BW-70M), monazite inclusions in garnet could be

separated. Under the binocular microscope, monazites displayed clear core-rim structures only in one case (BW-90M) where fragments were broken (Table 3).

Titanite is restricted to amphibolites. It displays irregular, contorted shapes where grown interstitially (BW-118) or forms flat sphenoidal crystals where included in amphibole and plagioclase (BW-134). No colour zoning was detected under the polarising microscope.

Amphibole in amphibolites grows mainly at the expense of granulite-stage pyroxenes, whereby it may completely replace pyroxenes in distinct layers of the rocks. X_{Mg} , Al_2O_3 and K_2O contents vary considerably among the samples, with K_2O being generally low (0.27–0.82 wt%, Table 2). The compositions correspond to magnesio-hornblende (BW-112), edenitic hornblende (BW-134) and actinolitic hornblende (BW-118, BW-132) according to the nomenclature of Leake (1978).

Biotite is one of the major phases in the migmatites. It occurs mainly in mesosomes and melanosomes and only rarely in leucosomes. Biotite is partly consumed by dehydration melting reactions. Due to the divariant character of the latter reactions, a large percentage of the biotite remains stable during partial melting. During subsequent cooling, new biotite grows and all biotite compositions are re-equilibrated on the retrograde path

Table 3 Results of U–Pb dating

Loca- tion	Sample, analysis no. and mineral properties ^a	Weight [μg] ^b	U [ppm] ^b	Th/U ^c	Pb _{com} [pg] ^d	$^{206}\text{Pb}/^{204}\text{Pb}$	$^{206}\text{Pb}/^{238}\text{U}$ f.g.	$^{207}\text{Pb}/^{235}\text{U}$ f.g.	$^{207}\text{Pb}/^{206}\text{Pb}$ f.g.	$^{206}\text{Pb}/^{238}\text{U}$ [Ma] ^f	+ / – [2σ] ^g	$^{207}\text{Pb}/^{235}\text{U}$ [Ma] ^f	+ / – [2σ] ^g	$^{207}\text{Pb}/^{206}\text{Pb}$ [Ma] ^f	+ / – [2σ] ^g
1	BW-70 M 1, m eu incl in grt [1] 2, m eu [1] 3, m eu [1] 4, m sb [1] 5, m eu incl A [1]	1 1 1 3 20	1918 2053 558 4932 3079	16.00 13.84 16.71 9.19 17.43	12.0 1.2 1.2 8.7 76.9	701 5547 1529 5375 2548	0.06842 ± 58 0.05162 ± 25 0.05136 ± 26 0.05143 ± 28 0.05045 ± 25	0.4867 ± 60 0.3764 ± 21 0.3745 ± 31 0.3738 ± 23 0.3667 ± 22	0.05159 ± 44 0.05288 ± 11 0.05261 ± 32 0.05271 ± 10 0.05271 ± 12	426.6 324.5 324.5 323.3 317.3	3.6 1.6 1.6 1.8 1.6	402.6 324.4 323.0 322.4 317.2	4.2 1.6 2.4 1.6 1.6	267 323.6 312 316.3 316.3	20 4.9 14 4.3 5.2
	BW-70L 6, m eu [1] 7, m eu [1] 8, m eu [1] 9, m eu [1] 10, m sb A [1] 11, z eu tips A [2]	3 10 4 1 1 2	5785 6278 7205 3095 7466 145	7.68 7.64 2.50 5.85 7.11 0.06	8.7 5.6 1.9 1.3 1.2 2.0	6458 35862 49912 7743 20059 489	0.05144 ± 27 0.05148 ± 33 0.05153 ± 31 0.05128 ± 24 0.05103 ± 29 0.05145 ± 26	0.3740 ± 22 0.3736 ± 25 0.3731 ± 24 0.3729 ± 20 0.3710 ± 23 0.3746 ± 65	0.05273 ± 10 0.05263 ± 08 0.05250 ± 08 0.05274 ± 10 0.05273 ± 10 0.05281 ± 82	323.4 323.6 323.9 322.4 320.9 323.4	1.6 2.0 2.0 1.4 1.8 1.6	322.6 322.3 321.9 321.8 320.4 323.1	1.6 1.8 1.8 1.4 1.8 4.8	317.0 312.7 307.3 317.6 317.2 321	4.3 3.3 3.6 4.3 4.6 35
4	BW-134 31, t fr lbr A [> 30] 32, t eu-sb f lbr [> 30]	333 170	21 16	1.47 1.79	405 178	73 68	0.05108 ± 44 0.05107 ± 44	0.383 ± 23 0.376 ± 23	0.0543 ± 32 0.0534 ± 33	321.2 321.1	2.6 2.8	329 324	17 17	384 347	129 136
5	BW-118 33, t an-fr lbr A [> 30] BW-120 23, m eu eq [1] 24, m eu eq [1] 25, m eu eq [1] 26, m eu eq incl [1]	173 2 1 2 2	110 1701 15364 2263 6956	0.64 8.02 5.63 9.28 14.74	132.2 3.3 3.2 5.7 12.0	477 3332 15535 2590 3836	0.05099 ± 26 0.05116 ± 27 0.05122 ± 26 0.05145 ± 24 0.05142 ± 25	0.3729 ± 36 0.3723 ± 23 0.3722 ± 21 0.3737 ± 21 0.3740 ± 21	0.05303 ± 39 0.05277 ± 16 0.05270 ± 08 0.05268 ± 12 0.05275 ± 10	320.6 321.7 322.0 323.4 323.2	1.6 1.6 1.6 1.4 1.6	321.8 321.3 321.3 322.4 322.6	2.6 1.8 1.6 1.6 1.6	330.2 318.9 315.9 314.9 318.0	16.5 7.0 3.6 5.3 4.5
6	BW-116 27, m eu-sb eq [1] 28, m eu-sb tip [1] 29, m eu eq [1] 30, m eu eq incl [1]	1 3 1 3	519 1979 1780 4569	10.95 7.59 8.31 11.13	2.6 2.3 6.7 33	697 8527 883 1360	0.05377 ± 29 0.05203 ± 25 0.05188 ± 24 0.05141 ± 28	0.3936 ± 44 0.3828 ± 22 0.3789 ± 27 0.3744 ± 25	0.05309 ± 49 0.05337 ± 13 0.05296 ± 23 0.05283 ± 17	337.6 327.0 326.1 323.2	1.8 1.6 1.6 1.8	337.0 329.1 326.2 322.9	3.2 1.6 2.0 1.8	333 344.5 327 321.3	21 5.4 10 7.4
7	BW-90M 12, m eu sp [1] 13, m sb A [1] 14, m eu [1] 15, m an core [1] 16, m an rim [1] 17, m sb A [1]	2 1 3 11 2 4	318 8915 5287 933 7234 4030	6.86 10.11 6.75 7.57 7.01 7.18	3.0 2.0 4.1 2.4 6.3 2.2	734 14368 12613 13806 7408 23541	0.05421 ± 27 0.05216 ± 28 0.05200 ± 30 0.05168 ± 25 0.05164 ± 29 0.05156 ± 26	0.3950 ± 37 0.3809 ± 23 0.3792 ± 24 0.3768 ± 21 0.3764 ± 23 0.3753 ± 21	0.05285 ± 39 0.05295 ± 10 0.05289 ± 08 0.05288 ± 09 0.05287 ± 09 0.05280 ± 08	340.3 327.8 326.8 324.8 324.6 324.1	1.6 1.8 1.8 1.6 1.8 1.6	338.0 327.7 326.4 324.7 324.4 323.1	2.8 1.8 1.8 1.6 1.8 1.6	322 326.7 323.9 323.6 323.1 320.0	17 4.3 3.6 3.7 3.7 3.5
	BW-90L 18, m eu [1] 19, m eu A [1] 20, m eu A [1] 21, m eu [1] 22, z prism [1]	1 2 4 1 1	12730 1558 10310 2632 241	2.90 2.74 2.78 7.03 0.34	3.1 3.0 8.9 0.9 0.8	13420 3415 15019 9312 952	0.05222 ± 32 0.05197 ± 26 0.05186 ± 31 0.05173 ± 24 0.05186 ± 26	0.3805 ± 25 0.3797 ± 22 0.3780 ± 25 0.3768 ± 20 0.3795 ± 40	0.05284 ± 09 0.05298 ± 13 0.05286 ± 09 0.05283 ± 11 0.05307 ± 46	328.2 326.6 326.0 325.2 325.9	2.0 1.6 2.0 1.4 1.6	327.4 326.8 325.6 324.7 326.7	1.8 1.6 1.8 1.4 3.0	321.8 328.1 322.7 321.4 332	3.7 5.7 3.9 47 20

^a m Monazite, z zircon, t titanite, eu euhedral, sb subhedral, an anhydral, eq equant, f flat, fr fragments, incl inclusion, lbr light-brown, A abraded, [1] number of grains analysed^b Weights known to better than 10% when over 10 μg , and to about 50% when less than 2 μg ; accuracy of U and Th concentrations is roughly proportional to uncertainty of sample weight^c Model Th/U ratio estimated from $^{206}\text{Pb}/^{206}\text{Pb}$ ratio and age of sample^d Total common Pb in sample, includes initial and blank Pb^e Measured ratio, corrected for fractionation and spike contribution^f Corrected for spike, fractionation, blank and initial common Pb (Stacey and Kramers 1975)^g 2σ uncertainty calculated by error propagation procedure that takes into account internal measurement statistics and external reproducibility as well as uncertainties in the blank and common Pb correction; errors are given as the last decimal places of the respective values

at very high temperatures (Kalt et al. 1999). Biotite inclusions in garnet may potentially record older stages of metamorphism or earlier metamorphic events with the $^{40}\text{Ar}/^{39}\text{Ar}$ chronometer (Kelley et al. 1997) through being armoured from Ar diffusion. Therefore, apart from single grains very thin, polished rock sections were prepared for dating biotite inclusions in garnet and biotite grains in the matrix. Biotites from migmatite samples have rather uniform compositions characterised by fairly low X_{Mg} values and K_2O contents >9 wt%. The largest variation is found in TiO_2 (3.82–5.12 wt%, Table 2). Biotite in gneiss sample BW-132 is in equilibrium with amphibole, plagioclase and quartz. It displays lower TiO_2 contents and higher X_{Mg} values compared to biotite from migmatites (Table 2).

K-feldspar occurs in mesosomes and in leucosomes of the migmatites. K-feldspars in leucosomes clearly crystallised from a melt during stage C. Conversely, their role in mesosomes is not clear as in general K-feldspar may be either consumed or formed by dehydration melting reactions depending on the $\text{K}_2\text{O}/\text{H}_2\text{O}$ ratio of the melt (Carrington and Watt 1995). However, K-feldspars in all samples are perthitic and thus record equilibration above approximately 700 °C. Above this temperature and at the relevant pressures, K-feldspar is generally homogeneous monoclinic, whereas below approximately 700 °C, spinodal decomposition of domains to a triclinic albite phase starts (Parsons and Brown 1991 and references therein). The exsolution lamellae observed in K-feldspars from migmatites of the Bayerische Wald have several scales. With a polarising microscope, parallel or subparallel lamellae of 50–250 μm widths can be observed. Back-scattered electron microprobe images reveal further exsolution lamellae of 1–5 μm widths. In thin sections, beginning formation of braid perthite can be recognised, resulting from rotation, broadening and coalescence of the albite lamellae. These transformations occur at and below 400 °C (Parsons and Brown 1991 and references therein) and are generally ascribed to slow cooling of the feldspars. While K-feldspars of samples BW-67, BW-120 and BW-116 only have minor inclusions, those in samples BW-44 and BW-46 contain numerous inclusions of plagioclase. All K-feldspar grains selected for dating come from leucosomes.

U–Pb dating

Analytical techniques

The U–Pb isotopic measurements were conducted at the Jack Satterly Geochronological Laboratory, Department of Earth Sciences, Royal Ontario Museum in Toronto. Minerals were separated by crushing with a jaw-crusher, pulverisation with a disk mill, heavy mineral enrichment on a Wilfley table, and subsequent magnetic separation on a Frantz isodynamic separator and density separation using heavy liquids. The minerals

to be analysed were separated under a binocular microscope and in part air-abraded (Table 3) following the technique of Krogh (1982). Single grains or single fragments of grains were dated except for zircon analysis of sample BW-70L (two grain fragments, Table 3) and titanite analyses (>30 grains, Table 3).

After a final selection, the minerals were washed in ca. 4 N HNO_3 on a hotplate and rinsed with H_2O and acetone. A mixed $^{205}\text{Pb}/^{235}\text{U}$ spike was used for U–Pb analyses of zircon, titanite and monazite. The spike was added to the sample after weighing and transfer to the dissolution vessel. Zircon was dissolved in HF (+ HNO_3) in Teflon mini-bombs at ca. 190 °C, monazite was dissolved in 6 N HCl in Savillex vials on a hotplate, and titanite was dissolved in HF (+ HNO_3) using Savillex vials on a hotplate. The solutions were subsequently evaporated, redissolved in 3.1 N HCl and passed through anion exchange resin in minicolumns in HCl medium to purify U and Pb (zircon and monazite). For titanite, a more complex HCl–HBr– HNO_3 procedure was necessary to purify U and Pb. Blanks were less than 2 pg Pb and 0.1 pg U for zircon and monazite and 10 pg Pb and 0.5 pg U for titanite.

Pb and U were collected together from the columns, loaded on outgassed Re-filaments together with H_3PO_4 and Si-gel, and run on a VG354 mass spectrometer using a Daly detector. Daly–Faraday conversion was 0.04%/a.m.u. Fractionation factors for U and Pb correspond to 0.1%/a.m.u.

General features of the monazite and zircon data

The results for 28 monazite, 2 zircon and 3 titanite analyses are given in Table 3 and presented in Concordia diagrams (Figs. 2–4). The data patterns for the monazites reveal various degrees of complexity and repeated analyses were carried out to verify the reproducibility of the ages and explore possible causes for the deviation of some of the samples.

All the monazite data plot on or slightly above the concordia curve, a common observation in monazite, reflecting on the one hand the resistance of monazite to lead loss, and on the other hand the presence of excess ^{206}Pb resulting from the incorporation of excess ^{230}Th at the time of formation (Schärer 1984; Parrish 1990). Although excess ^{230}Th mainly affects the U–Pb systematics of very young samples it may as well lead to several percent of excess ^{206}Pb in older samples provided the fractionation factor ($f = (\text{Th}/\text{U})_{\text{mineral}}/(\text{Th}/\text{U})_{\text{reservoir}}$) is large enough (Schärer 1984). The excess ^{206}Pb results in $^{206}\text{Pb}/^{238}\text{U}$ ages that are too high and $^{207}\text{Pb}/^{206}\text{Pb}$ ratios that are too low. A correction for the initial disequilibrium is generally possible but was not applied in this case as the Th/U ratio of the initial reservoir open to U, Th and Pb exchange with monazite is not known. Without correction, the most accurate estimate of the ages can be obtained from $^{207}\text{Pb}/^{235}\text{U}$ ratios which are not affected by the initial disequilibrium. Hence, all the monazite

ages reported in Figs. 2–4 and referred to in the text are based on weighted averages of $^{207}\text{Pb}/^{235}\text{U}$.

Zircon commonly incorporates very little Th at the time of formation and hence remains essentially unaffected by disequilibrium. In order to date only zircon parts grown during HT–LP metamorphism, very small zircon fragments had to be used that contained very little U and Pb, affecting the precision of the $^{207}\text{Pb}/^{235}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ ages through the common lead correction. Therefore, the $^{206}\text{Pb}/^{238}\text{U}$ ages are reported for zircon analyses. Titanite also incorporates only very little Th and is characterised by low U and Pb abundances. Therefore, $^{206}\text{Pb}/^{238}\text{U}$ ages are also indicated for titanite.

Ages of monazite and zircon from migmatites and gneisses

For sample BW-70, five analyses of monazite from the leucosome cluster tightly on or slightly above Concordia, yielding an average $^{207}\text{Pb}/^{235}\text{U}$ age of

321.8 ± 0.7 Ma (Fig. 2a). Analysis of two abraded zircon tips yields a concordant data point and an overlapping $^{206}\text{Pb}/^{238}\text{U}$ age of 323.4 ± 1.6 Ma (Fig. 2a). More complex relations are observed for the mesosome (Fig. 2b). Three of the monazite analyses yield overlapping results at a mean $^{207}\text{Pb}/^{235}\text{U}$ age of 323.3 ± 1.0 Ma. One analysis (no. 5, Table 3) yields a younger age of 317.2 ± 1.6 Ma (Fig. 2b). This grain contained several inclusions that probably contributed to the elevated initial common Pb (Table 3), but their effect on the age is not known. Another analysis (no. 1, Table 3) provides a reversely discordant but much older age (Fig. 2b). This grain was enclosed in garnet.

For sample BW-90, analyses of four monazite grains from the leucosome yield a weighted average $^{207}\text{Pb}/^{235}\text{U}$ age of 326.1 ± 1.7 Ma (Fig. 3a). The scatter of the data, however, exceeds analytical uncertainty. Analysis of a zircon prism from the leucosome yields a concordant data point and a $^{206}\text{Pb}/^{238}\text{U}$ age of 325.9 ± 1.6 Ma (Fig. 3a), identical to the monazite age. Five analyses of monazite from the mesosome are clustered tightly on or

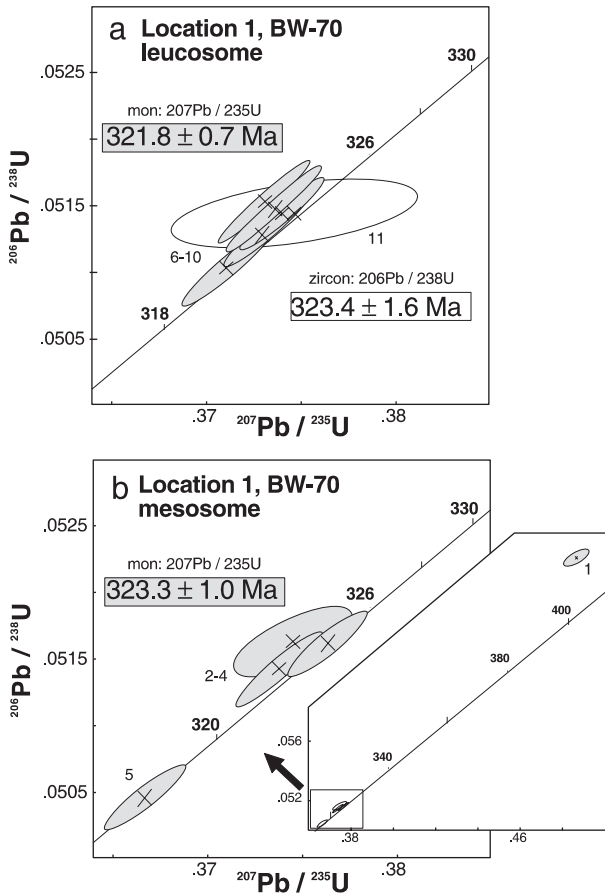


Fig. 2a, b U–Pb concordia diagram showing the results of U–Pb dating of monazite and zircon from migmatite sample BW-70 of location 1. **a** Leucosome; **b** mesosome. Errors are given at the 2σ level. For further explanation see section on general features of the zircon and monazite data and section on ages of monazite and zircon from migmatites and gneisses

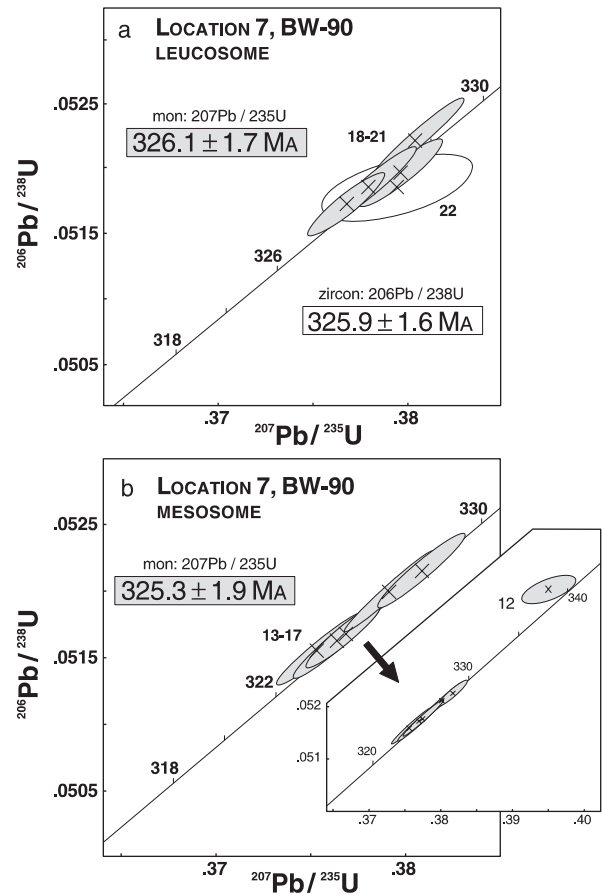


Fig. 3a, b U–Pb concordia diagram showing the results of U–Pb dating of monazite and zircon from migmatite sample BW-90 of location 7. **a** Leucosome; **b** mesosome. Errors are given at the 2σ level. For further explanation see section on general features of the zircon and monazite data and section on ages of monazite and zircon from migmatites and gneisses

slightly above the Concordia curve (Fig. 3b). The average weighted $^{207}\text{Pb}/^{235}\text{U}$ age is 325.3 ± 1.9 Ma, but the scatter of the data is beyond analytical uncertainty. A sixth analysis (no. 12, Table 3) of a simple prismatic crystal gave a significantly older age of 340 Ma. Some of the monazite grains of BW-90 mesosome showed apparent cores and overgrowths. One of these grains was broken into core and rim fragments (nos 15–16, Table 3) to test for a possible inherited origin of the core. However, both parts of the grain yielded identical ages (Table 3).

In migmatites BW-120 and BW-116, mesosome and leucosome could not be separated. Four monazite grains from sample BW-120 yield data points that overlap on or slightly above the Concordia curve with a mean $^{207}\text{Pb}/^{235}\text{U}$ age of 321.9 ± 0.8 Ma (Fig. 4a). The results for four monazite analyses of sample BW-116 are more complex (Fig. 4b). Three of the analyses are concordant but have $^{207}\text{Pb}/^{235}\text{U}$ ages ranging from 337 to 323 Ma. Another analysis (no. 28, Table 3) is slightly discordant (Fig. 4b). There is no obvious morphological or colour difference between the four monazite grains. The

$^{207}\text{Pb}/^{235}\text{U}$ age of the youngest monazite is 322.9 ± 1.8 Ma.

Ages of titanite from amphibolites

Two analyses were performed on titanite of sample BW-134, one of them using fragments that were abraded, the other using flat spheroidal crystals. Both show low U contents of 16–21 ppm and provide identical $^{206}\text{Pb}/^{238}\text{U}$ age values of 321.2 ± 2.7 Ma. Although the low $^{206}\text{Pb}/^{204}\text{Pb}$ ratios of about 70 make the age calculation quite sensitive to the choice of initial Pb, the variation introduced by this correction on the $^{206}\text{Pb}/^{238}\text{U}$ age is minimal. For example, the use of Stacey and Kramers (1975) model Pb compositions, calculated for ages spanning almost the entire Paleozoic and Mesozoic, yields a range of $^{206}\text{Pb}/^{238}\text{U}$ values covered almost entirely by the quoted analytical uncertainty. The titanite population of sample BW-118 consists of anhedral grains with irregular, contorted shapes. As the U content and the proportion of radiogenic Pb are much higher than in sample BW-134 (Table 3), the $^{206}\text{Pb}/^{238}\text{U}$ age of 320.6 ± 1.6 Ma is much more precise, and much less dependent on the choice of initial Pb. In this case a Paleozoic–Mesozoic range of model corrections produces a variation of less than half the analytical error.

$^{40}\text{Ar}/^{39}\text{Ar}$ dating

Analytical techniques

The $^{40}\text{Ar}/^{39}\text{Ar}$ isotopic measurements were conducted using the VULKAAN argon laserprobe at the Faculty of Earth Sciences, Vrije Universiteit Amsterdam. For incremental heating experiments, mineral separates were obtained by carefully crushing finger- to hand-sized samples, and sieving through 500 μm , 250 μm and 125 μm mesh fractions. The mineral separates were washed with deionised water, separated magnetically and handpicked. For an incremental heating experiment, 1–3 grains of biotite, 60 grains of hornblende and 1 grain of K-feldspar were used, respectively. The separates were put into Al tablets containing 20, 2-mm holes. On each tablet, 16 positions were loaded with samples and 4 with a flux monitor of known age (standard DRA-1, 24.99 Ma). For spot fusion experiments on biotite and K-feldspar, polished sample sections of 2×2 cm, 90 μm thick, were prepared. Their corners and edges were then broken to fit the sections onto an Al tablet. Additional tablets containing only the standard were also prepared. All tablets were wrapped in Al foil and loaded into cylindrical Al containers, with sample sections and standard tablets alternating. The containers were then irradiated in the Triga Reactor at the Oregon State University for 24 (first sample charge) and 18 h (second sample charge).

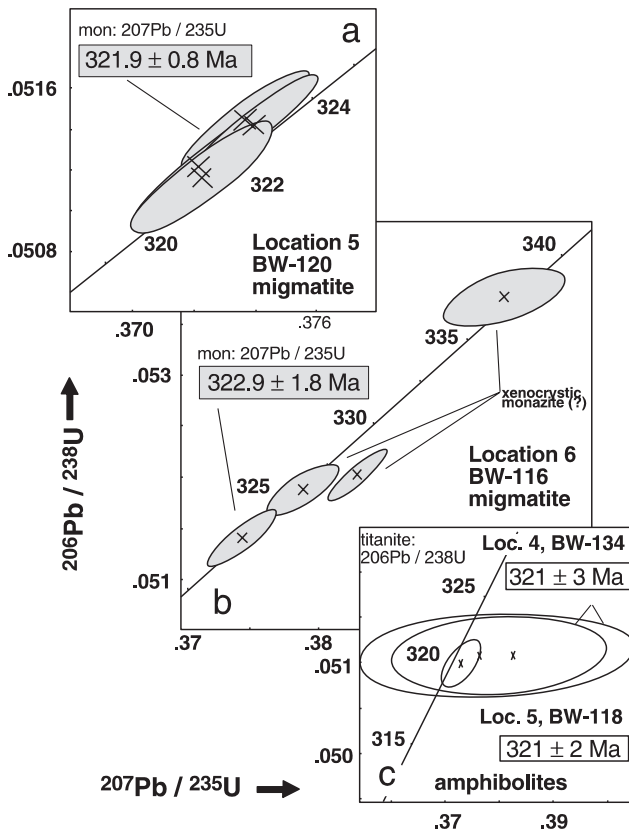


Fig. 4a–c U–Pb concordia diagrams showing the results of U–Pb dating of monazite and titanite. **a** Monazite from migmatite sample BW-120 (location 5); **b** monazite from migmatite sample BW-116 (location 6); **c** titanite from amphibolite samples BW-134 (location 4) and BW-118 (location 5). Errors are given at the 2σ level. For further explanation see section on general features of the zircon and monazite data and section on ages of monazite and zircon from migmatites and gneisses

The VULKAAN argon laserprobe consists of a 24 W argon ion laser (visible light, 488–524 nm), beam optics, a low volume UHV gas inlet system, and a MAP 215-50 noble gas mass spectrometer. Full technical and analytical laserprobe dating procedures were described by Wijbrans et al. (1995). The isotopic composition of Ar was measured in static mode. Intensities were recorded with a secondary electron multiplier detector using switchable pre-amplifier resistor settings (10, 100, 1000 M Ω) and peak jumping at half mass intervals from mass 40 down to mass 35.5. All measured values were corrected for mass discrimination, ^{37}Ar and ^{39}Ar decay, neutron-induced interferences from Ca and K, and procedural blanks. Correction factors were as quoted in Wijbrans et al. (1995). The J values (irradiation parameter) were calculated by measuring replicate standards from different locations within the tablets and the containers and by fitting a function through the data points, respectively. Ages were calculated using the decay constants recommended by Steiger and Jäger (1977). The quoted errors on the ages ($\pm 2\sigma$) include the uncertainty on the irradiation parameter J , the uncertainty on the mass spectrometric analyses of the Ar intensities, errors on the blanks and the mass interferences of radiogenic Ar with Ar derived from K, Ca and Cl. The uncertainty of a plateau age is calculated from the uncertainties of the individual steps, using the inverse variance of the $^{40}\text{Ar}/^{39}\text{Ar}$ ratios as a weighting factor.

Incremental heating experiments were performed by progressively degassing the sample grains using a defocused laser for 60 s at successively greater laser power increments. System blanks were measured before the first experiment and every fifth analysis thereafter. The blank correction applied to each sample used the blank integrated over the time of sample analyses. Typical system blanks were $3.92\text{--}6.63 \times 10^{-17}$, $5.18\text{--}8.91 \times 10^{-19}$, $8.91\text{--}33.7 \times 10^{-20}$, $2.61\text{--}3.23 \times 10^{-18}$ and $5.87\text{--}12.6 \times 10^{-19}$ mol for masses 40, 39, 38, 37 and 36, respectively. Spot fusion experiments on biotite and K-feldspar in sample sections were performed using a 2 W laser beam focused at 20 μm for approximately 2 s in 0.5 s pulses. System blanks were measured before each sample and every third analysis thereafter. Typical system blanks were $0.93\text{--}2.21 \times 10^{-16}$, $1.03\text{--}1.60 \times 10^{-18}$, $7.59\text{--}8.02 \times 10^{-19}$, $1.78\text{--}3.13 \times 10^{-18}$ and $1.61\text{--}1.78 \times 10^{-18}$ mol for masses 40, 39, 38, 37 and 36, respectively.

$^{40}\text{Ar}/^{39}\text{Ar}$ ages

The results of incremental heating and spot fusion experiments are given in Table 4. Representative age spectra obtained by incremental heating on hornblende, biotite and K-feldspar are shown in Figs 5, 6 and 7, respectively. Age plateaus are defined by at least three consecutive steps that overlap at the 95% confidence level (exclusive of uncertainty in J) and together contain at least 50% of the $^{39}\text{Ar}_K$ released by the sample. All

samples were additionally examined on inverse isochron diagrams ($^{36}\text{Ar}/^{40}\text{Ar}$ vs $^{39}\text{Ar}/^{40}\text{Ar}$), in which the age is given by the intercept on the $^{39}\text{Ar}/^{40}\text{Ar}$ axis and the composition of any trapped non-radiogenic argon by the intercept on the $^{36}\text{Ar}/^{40}\text{Ar}$ axis. In principle, the inverse isochron approach allows to determine whether the trapped argon is air-derived and/or excess argon. However, because the samples analysed here are fairly old and potassium-rich (except for hornblende), most of the data plot on or near the $^{39}\text{Ar}/^{40}\text{Ar}$ axis. Therefore, the $^{36}\text{Ar}/^{40}\text{Ar}$ intercepts have large uncertainties and give little information on initial ratios. The ages given by the $^{39}\text{Ar}/^{40}\text{Ar}$ intercepts generally coincide with the plateau and total fusion ages (Table 4).

Hornblende from three different locations (Fig. 1, Table 4) was analysed by incremental heating experiments. The separates yield plateau ages between 321.9 ± 3.3 and 313.7 ± 3.7 Ma (Table 4). Within error limits, there is good agreement between the plateau ages and the integrated total fusion ages (Table 4). The different plateau ages are not related to different sample locations as almost the entire range of ages is displayed by hornblende from location 4 (315.6 ± 3.5 to 321.9 ± 3.3 Ma, Fig. 5). A common characteristic of the hornblende age spectra is that the first, usually very small, heating step released argon with apparent ages that are considerably higher or lower than the plateau ages. The contribution of air argon in these first steps of incremental heating is comparatively large and the deviation of the apparent ages is most likely induced by the release of argon trapped on cracks or cleavage planes.

Biotite from all locations was analysed by incremental heating experiments on one or two grains, respectively. The obtained plateau ages range between 309.6 ± 3.5 and 313.2 ± 3.2 Ma (Fig. 6) and coincide with the integrated total fusion ages within error limits. Most spectra are characterised by very even plateaus with minor age deviations only in the first heating steps. As for hornblende, there are no significant age differences between the locations as the values overlap within error limits. Additionally, total fusion ages of single biotite grains from sample sections were obtained for some of the locations (Table 4). Garnet zoning patterns revealed that the inner cores of large garnets record the oldest metamorphic stage (Kalt et al. 1999). Therefore, biotite inclusions in garnet were analysed by spot fusion experiments and their mean ages were calculated separately from those obtained on matrix biotites (Table 4). However, no significant age difference between biotite from these two different textural positions is recognisable in any sample, and the mean spot fusion ages coincide with those obtained by incremental heating in every sample (Table 4). The spread in ages obtained by spot fusion experiments on single grains from sample sections is in most cases large, expressed by large errors on the mean ages (Table 4). This is probably because biotite grains were often rather small and some of the gas may have come from adjacent minerals or grain boundaries.

Table 4 Results of $^{40}\text{Ar}/^{39}\text{Ar}$ dating

Location	Sample	Mineral properties ^a	Incremental heating			Inverse isochron			Spot fusion			<i>J</i>				
			<i>n</i> ^b	MSWD	³⁹ Ar [%] ^c	PL ^d age [Ma]	+/- [2σ] ^e	TF ^f age [Ma]	+/- [2σ] ^e	<i>n</i> ^b	MSWD		⁴⁰ Ar/ ³⁶ Ar intercept [2σ] ^e	Age [Ma]	+/- [2σ] ^g	
1	BW-22	bt, 250–500 μm [2]	10	1.8	100.0	311.7	3.2	311.0	5.4	10	0.7	170.7	48.9	313.0	3.0	0.006161
	BW-28	bt, in matrix														0.005569
2	BW-30	bt, 250–500 μm [2]	11	2.4	96.8	311.6	3.5	310.3	5.7	11	2.4	173.0	174.5	312.4	5.5	0.006160
	BW-32	bt, incl in grt														0.005427
	BW-32	bt, in matrix														0.005427
	BW-67	bt, incl in grt														0.005259
	BW-67	bt, incl in grt														0.005259
	BW-67	Kfs, 250–500 μm [1]	10	109.2	100.0	299.3 ^j	2.9	292.2	2.9	10	109.2	–1099.5	1785.5	302.6	10.25	0.005259
3	BW-67	Kfs, 250–500 μm [1]	10	203.2	299.9	300.0 ^j	2.9	291.7	2.9	9	271.6	949.4 ^k	285.9	224.91	3.2	0.005259
	86246	bt, 250–500 μm [1]	5	0.5	99.7	312.0	3.1	312.1	3.9	5	0.4	330.0	145.6	311.9	3.2	0.006097
4	86246	bt, 250–500 μm [2]	7	0.3	96.0	312.6	6.2	308.2	24.8	11	0.5	181.3	346.1	313.7	4.9	0.006160
	BW-132	bt, 250–500 μm [3]	8	1.3	99.1	309.6	3.9	311.0	8.9	9	0.03	40771.1 [*]	267.2	311.4	3.1	0.006161
	BW-132	hbl, 125–250 μm [60]	10	3.1	97.7	321.9	3.3	323.6	4.1	11	5.6	453.2	124.5	314.6	9.4	0.004425
	BW-134	hbl, 125–250 μm [60]	8	1.0	100.0	319.1	3.0	318.8	3.6	8	0.9	271.7	29.5	319.5	3.0	0.004425
5	BW-134	hbl, 125–250 μm [60]	8	11.8	99.1	315.6	3.5	315.4	3.6	12	8.0	267.3	199.8	316.4	7.5	0.004425
	BW-118	hbl, 125–250 μm [60]	9	1.8	98.4	313.7	3.7	316.3	7.2	10	3.3	358.6	75.9	308.8	9.6	0.004425
	BW-120	bt, 250–500 μm [2]	5	0.3	92.5	311.8	3.0	310.8	3.7	6	1.7	159.2	45.7	312.4	3.1	0.006097
	BW-120	Kfs, 250–500 μm [1]	9	45.8	82.8	296.4	3.0	296.3	3.2	10	14.5	878.3	1816.2	291.5	20.2	0.006136
6	BW-112	hbl, 125–250 μm [60]	9	5.4	98.8	320.3	3.2	318.9	3.5	10	18.2	119.3	190.4	322.6	8.3	0.004425
	BW-116	bt, 250–500 μm [3]	4	0.7	69.6	309.7	4.3	303.6	11.4	8	1.7	138.3	93.1	311.2	4.6	0.006161
	BW-116	Kfs, 250–500 μm [1]	7	20.3	87.7	305.3	3.8	306.9	3.3	8	2.2	2424.7	2390.5	296.9	13.0	0.006136
	BW-44	bt, in matrix														0.005364
7	BW-44	bt, incl. in Kfs														0.005364
	BW-44	Kfs, 250–500 μm [1]	11	24.1	84.3	305.1	3.6	307.2	3.2	12	14.0	518.8	358.9	302.8	11.2	0.005364
	BW-46	bt, 250–500 μm [2]	8	0.8	100.0	313.2	3.2	313.0	5.3	8	0.7	172.6	92.8	314.5	3.3	0.006097
	BW-46	bt, 250–500 μm [2]	12	1.7	100.0	311.5	3.7	312.1	8.7	12	1.6	108.2	119.7	314.4	3.5	0.006161
	BW-46	Kfs, 250–500 μm [1]	9	2.8	96.5	319.7	3.1	318.4	3.4	12	0.2	20010.0 [*]	269.5	70.1	39	0.006136
	BW-46	Kfs, 250–500 μm [1]	3	2.3	84.0	317.8	3.1	318.0	3.3	9	5.7	3047.0 [*]	297.6			0.006136
	BW-90	bt, in matrix														0.005154
	BW-90	bt, incl. in grt														0.005154

^a Mineral abbreviations according to Kretz (1982); incl., inclusion; [1], number of grains^b n number of heating steps included in the plateau age calculation^c Percentage of ^{39}Ar included in the plateau age calculation^d PL Plateau^e Quoted errors include the uncertainty on the irradiation parameter J , the uncertainty on the mass spectrometric analyses of the Ar intensities (including errors on the blanks), the mass interferences of radiogenic Ar with Ar derived from K, Ca and Cl and the error on the weighted mean of the individual steps (Dalrymple and Lanphere 1971)^f TF Total fusion^g n Number of single grains analysed^h Mean age calculated from the total fusion ages of single grainsⁱ Standard deviation of the mean age^j Apparent age of the final heating step (no plateau age could be calculated)^k Asterisk indicates error too large to be adequately displayed

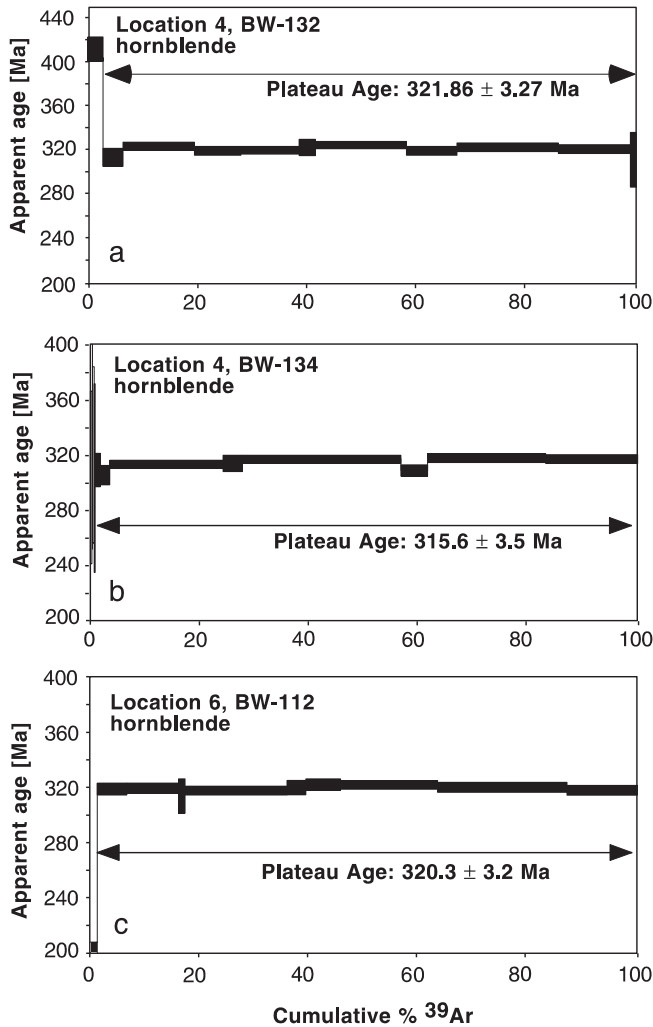


Fig. 5a–c Results of incremental heating experiments of hornblende separates from amphibolite samples BW-132 (a), BW-134 (b) and BW-112 (c). Errors are given at the 2σ level. For further explanation see section on $^{40}\text{Ar}/^{39}\text{Ar}$ dating

Spot fusion ages of K-feldspar from sample sections could not be obtained. It was not possible to fuse the spots by hitting them with the focused laser beam for several seconds. As longer laser treatment would have thermally activated other parts of the section, the spot fusion experiments on K-feldspar were not continued. K-feldspar was broken from the sections and fragments of grains were then analysed by incremental heating experiments. Incremental heating experiments were performed on single K-feldspar fragments from locations 1, 5, 6 and 7 (Table 4). In general, the age spectra obtained from K-feldspar are not as even as those of biotite and hornblende (Fig. 7). Plateau ages that include more than 80% of the released Ar and that coincide with the integrated total fusion ages could be calculated for samples BW-44, BW-46 and BW-116. For a K-feldspar of sample BW-120, a plateau age including only 60% of the released Ar, but coinciding with the integrated total fusion age, could be calculated. The two K-feldspars of sample BW-67 are characterised by a

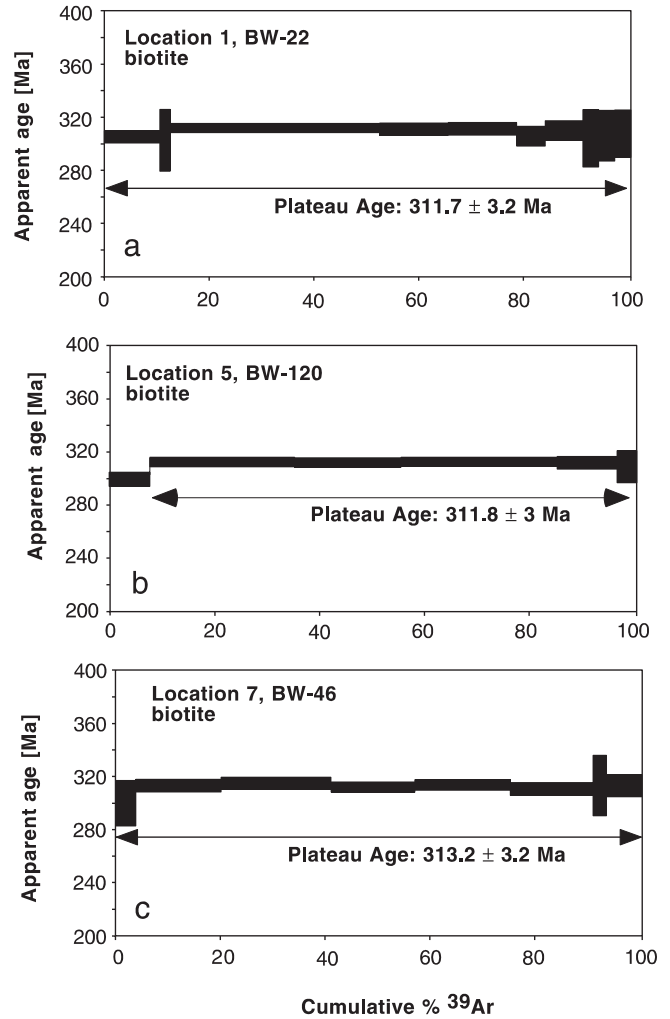


Fig. 6a–c Results of incremental heating experiments of biotite from migmatite samples BW-22 (a), BW-120 (b), and BW-46 (c). Errors are given at the 2σ level. For further explanation see section on $^{40}\text{Ar}/^{39}\text{Ar}$ dating

gentle staircase-like increase in apparent ages and by the lack of plateaus. The integrated ages are in good agreement with the total fusion ages for each sample, whereas the apparent ages of the final degassing step are slightly higher (Table 4). The K-feldspar plateau ages (ages of the final degassing step for sample BW-67) form two distinct groups: K-feldspars from locations 2, 5 and 6 have ages of 299.3 ± 2.9 , 300.0 ± 2.9 , 296.4 ± 3.0 and 305.3 ± 3.8 Ma, the latter being identical to an age of 305.1 ± 3.6 Ma of one K-feldspar from locality 7. Two other K-feldspars from location 7 have considerably higher ages of 319.7 ± 3.1 and 317.8 ± 3.1 Ma.

Significance of mineral ages and link to the P–T path

U–Pb zircon ages

Microstructural data and modes show the leucosomes of migmatites from the Bayerische Wald to have either been

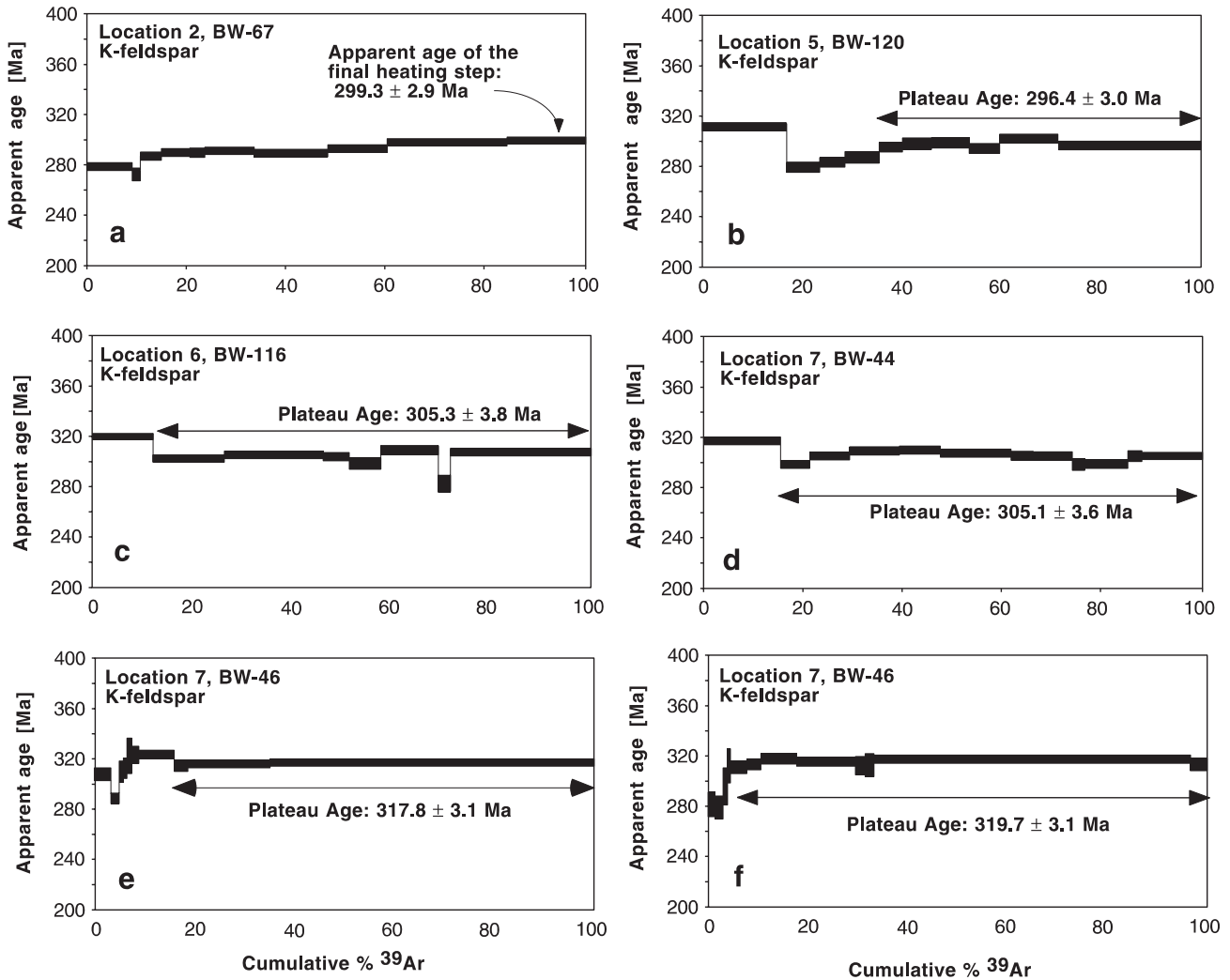


Fig. 7a–f Results of incremental heating experiments of K-feldspar from migmatite samples BW-67 (a), BW-120 (b), BW-116 (c), BW-44 (d) and BW-46 (e–f). Errors are given at the 2σ level. For further explanation see section on $^{40}\text{Ar}/^{39}\text{Ar}$ dating

pure melt (Qtz–Kfs leucosomes) or melt and residual crystals (Plg–Qtz–Kfs leucosomes, Berger and Kalt 1999). Therefore, the analysed zircon tips from the leucosomes must have coexisted with melt at some stage. The euhedral shapes of the zircons strongly suggest that the tips grew during HT–LP metamorphism by precipitation from a melt. This interpretation is supported by the fact that most crustal lithologies contain a sufficient amount of zirconium (Zr) and light rare earth elements (LREE) to saturate a melt in these components and allow precipitation of accessory phases such as zircon and monazite (see Watt and Harley 1993 for a discussion). The case may be different in small-volume disequilibrium melts (Sawyer 1991), but estimates of former melt volume fractions in migmatites of the Bayerische Wald yield values of 20–40% (Berger and Kalt 1999). The core-rim structures of zircon grains visible under the binocular microscope probably represent residual zircons that were entrained in the partial melt formed during HT–LP

metamorphism, partly dissolved and later overgrown during zircon precipitation. Hence, the concordant U–Pb ages of the zircon tips must be crystallisation ages as opposed to resetting ages. The ages do not record cooling. This interpretation is generally accepted for zircons, based on low diffusivities for U, Pb and Th (Cherniak et al. 1997; Lee et al. 1997) and very high closure temperatures for Pb in zircon (Lee et al. 1997), implying that volume diffusion is not the dominant process for Pb loss in zircon (Mezger and Krogstad 1997).

Zircon incorporates large amounts of incompatible elements (Zr, Hf, Y, U, LREE, e.g. Hanchar and Miller 1993). In the metapelitic migmatites of the Bayerische Wald, these incompatible elements became available as soon as in situ melting started during prograde stage B dehydration reactions through the dissolution of existing zircon (Watson 1996). The numerical simulations of Watson (1996) show that diffusion-controlled zircon dissolution in melts of granitic composition at increasing temperatures is a continuous process whereby dissolution rates increase exponentially with temperature in the range 650–850 °C. Diffusion-controlled zircon growth in granitic melts is also favoured by increasing tempera-

tures (Watson 1996). Therefore, the zircon tips used for dating here should reflect growth beginning at peak metamorphic conditions, estimated to be about 850 °C (Kalt et al. 1999), down to temperatures at which all melt crystallised. This was at the end of stage B and during stage C (Fig. 1). Experiments in metapelitic systems indicate 800 °C as approximate minimum temperature for biotite dehydration melting (e.g. Le Breton and Thompson 1988; Vielzeuf and Montel 1994) and thermobarometry on subsolidus stage D phase assemblages yields 770–846 °C as a minimum temperature for melt crystallisation in migmatites of the Bayerische Wald (Kalt et al. 1999). Therefore, the U–Pb ages of zircon tips date a minimum temperature of approximately 800 °C. This temperature is still below common estimates of Pb closure temperature and thus supports the interpretation as crystallisation ages (Lee et al. 1997).

U–Pb monazite ages

In general, the interpretation of U–Pb ages is not as straightforward for monazites as it is for zircons because monazite closure temperatures for U, Th and Pb are still a matter of debate. Comparatively high temperatures of ≥ 750 °C are suggested by experimental diffusion data (Smith and Giletti 1997) and crystal-chemical considerations (Dahl 1997). Slightly lower minimum estimates of 680–720 °C emerge from the presence of inherited monazites in leucogranites (Copeland et al. 1988; Kingsbury et al. 1993; Harrison et al. 1995), from partial resetting of monazites at amphibolite-facies conditions (Parrish et al. 1990) and from electron microprobe analyses of zoned monazites (Suzuki et al. 1994). However, slow diffusion velocities, the preservation of radiogenic lead in inherited monazite grains from magmatic and metamorphic rocks and postmagmatic monazite recrystallisation (Hawkins and Bowring 1997) argue that recrystallisation rather than volume diffusion controls the U–Pb systematics of monazite. This view is also supported by the presence of very sharp boundaries in line scans of Th, U and Pb across granulite-facies monazites (Zhu and O’Nions 1999). Peak metamorphic temperatures in migmatites of the Bayerische Wald are above some of the assumed U–Pb closure temperatures for monazite. However, a number of observations suggests that the obtained U–Pb monazite ages reflect crystallisation and inheritance rather than monazite resetting and/or cooling through a closure temperature.

The euhedral shape of the monazites from leucosomes suggest that at least their rims record growth in the presence of a melt and not resetting of older crystals. The presence of excess ^{206}Pb (see section on U–Pb dating) in monazites from leucosomes and mesosomes is the strongest argument for the monazite ages reflecting crystallisation. If resetting of older monazites by diffusion had occurred during HT–LP metamorphism, this disequilibrium effect would not have been preserved. Both mesosome samples (BW-70 and BW-90) contain

simple, euhedral to subhedral inherited monazite grains with distinctly higher ages (Figs 2b, 3b). In mesosome BW-70, the inherited monazite was included in early grown garnet that shielded it from recrystallising or dissolving in a partial melt during HT–LP metamorphism. This effect was described by DeWolf et al. (1993) for monazites from the high-grade core of the Wind River Range, Wyoming. In mesosome BW-90, the monazite carrier phase is not known. The rest of the monazites from the mesosome of sample BW-70 cluster on concordia due to the absence of inherited monazite domains, while those from the mesosome of sample BW-90 show a scatter that can be best explained by inheritance of older concordant monazites.

The U–Pb systematics of monazites from migmatite BW-116 (leucosome and mesosome could not be separated) most clearly reveal the effects of inheritance (Fig. 4). Three concordant grains with ages between 323 and 337 Ma as well as one slightly discordant grain demonstrate that monazite crystallised at 322.9 ± 1.8 Ma during HT–LP metamorphism may contain varying proportions of older monazite with either undisturbed or partially reset U–Pb systems. Monazites of sample BW-120 reveal a simpler picture (Fig. 4a) and are best interpreted as reflecting crystallisation without inheritance. Future study on monazite composition may help to distinguish inherited and newly crystallised parts.

In summary, the monazite U–Pb data reveal crystallisation at the same time as the precipitation of zircon from partial melts and must hence record virtually the same minimum temperature of approximately 800 °C. Monazites in all samples except BW-120 show inheritance of older monazite to varying degrees. The fact that these older monazites survived metamorphic conditions and partial melting at temperatures of approximately 850 °C places new constraints on the U–Pb closure temperature of monazite. There is a slight trend towards higher U–Pb ages of zircon and monazite from the northwestern area around Cham. Considering the fact that the monazites from all samples are effected by variable degrees of inheritance of older grains, it is possible that the slightly higher U–Pb monazite ages from the northwestern area reflect inheritance. However, zircon tips that should be free of inherited components are also older in the northwestern area, suggesting that the age difference is geologically meaningful. The very young U–Pb age of 317 Ma for one monazite grain of sample BW-70 cannot be reasonably explained at the moment. Perhaps it is due to local late-stage hydrothermal features, although a U–Pb isotope study on hydrothermally treated monazites suggests that monazite becomes discordant due to lead loss under these conditions (Teufel and Heinrich 1997).

U–Pb titanite ages

Several U–Pb dating studies document that titanite may remain closed to Pb diffusion at elevated temperatures,

may be reset during later metamorphism or may recrystallise even at low temperatures (Corfu and Grunsky 1985; Corfu et al. 1994; Corfu 1996). Other studies suggest that while at temperatures above 700 °C titanite is affected by Pb diffusion, recrystallisation is the dominant process below approximately 700 °C (Scott and St-Onge 1995; Verts et al. 1996). Calculations of U–Pb closure temperatures from magmatic titanite with inherited Pb components yield minimum estimates of 712 °C (Zhang and Schärer 1996). These estimates coincide quite well with calculations from diffusion experiments (Cherniak 1993). As titanite in the amphibolites of the Bayerische Wald is not only included in amphibole but also in plagioclase that was formed during earlier granulite-facies metamorphism, it most likely crystallised at temperatures (850 °C) considerably exceeding any estimates of closure temperature. Thus, the U–Pb ages of titanites most likely do not record crystallisation but cooling to temperatures of approximately 700 °C.

$^{40}\text{Ar}/^{39}\text{Ar}$ ages

Diffusion of Ar in amphibole, biotite and K-feldspar is comparatively fast, even at fairly low temperatures (e.g. Baldwin et al. 1990; Lovera et al. 1991; Dahl 1996 and references therein). As in many metamorphic terranes the attained peak temperatures are considerably higher than the closure temperatures calculated from the Ar diffusion velocities, K–Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages of amphibole, biotite and K-feldspar can be interpreted as cooling ages, provided that thermally induced volume diffusion according to Fick's Law is the dominating process and that the minerals did not form below their closure temperatures. There are also other mechanisms of Ar loss or exchange such as fluid- or deformation-assisted recrystallisation (e.g. Wijbrans and McDougall 1986; Villa 1998; Parsons et al. 1999), but these mainly apply to Ar-bearing minerals that were substantially affected by hydrothermal overprint (metamorphism) and/or deformation after their crystallisation. These complex histories usually result in disturbed Ar spectra. In the Bayerische Wald, migmatites and amphibolites experienced one single HT-metamorphic event, peak metamorphic temperatures approximated 850 °C, the investigated rocks and minerals were not subject to late-stage fluid- or deformation-assisted recrystallisation (see sections on geological setting and on characteristics of the phases), and apart from two feldspars, all analysed Ar-bearing minerals yielded good plateaus (Figs. 5–7). Therefore, it seems reasonable to interpret the $^{40}\text{Ar}/^{39}\text{Ar}$ ages as recording cooling below closure temperatures.

For amphiboles it has been discussed whether the Ar retentivity and thus the closure temperatures depend on composition. Whereas several geochronological studies indicated the possibility of a compositional dependence (e.g. O'Nions et al. 1969; Berry and McDougall 1986; Onstott and Peacock 1987), experimental determination

of Ar diffusion velocities in metamorphic hornblende showed no correlation between X_{Mg} and activation energy (Baldwin et al. 1990). On the basis of the ionic porosity model of Fortier and Giletti (1989), Dahl (1996) argued that there should be differences in closure temperature due to compositional effects, but that these compositional effects would probably become lost within analytical uncertainty once the cooling rate exceeded 10 °C/my. The amphiboles examined here by incremental heating experiments are magnesio-hornblende (BW-112), edenitic hornblende (BW-134) and actinolitic hornblende (BW-118, BW-132). The calculations of Dahl (1996) suggest that pure edenite should have higher closure temperatures (600–520 °C) than pure ferro-actinolite (440–470 °C) at cooling rates of 10–100 °C/my, which are realistic for the Bayerische Wald (see section on cooling rates). This compositional effect cannot be seen in the $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende total fusion ages obtained here (Tables 2 and 4), because the investigated amphiboles have intermediate, but not endmember compositions. Magnesio-hornblende and edenitic hornblende (320.3 ± 3.2 , 319.1 ± 3.0 , 315.6 ± 3.5 Ma) do not yield significantly higher ages than actinolitic hornblende (321.9 ± 3.3 , 313.7 ± 3.7 Ma). Therefore, combining the hornblende compositions of the samples with the ionic porosity model of Dahl (1996), it can be assumed that the hornblende ages obtained record cooling of the amphibolites below approximately 500–570 °C.

As for amphibole, it has been discussed whether the closure temperatures of biotite may vary with composition. Whereas Harrison et al. (1985) found a negative correlation between X_{Mg} and activation energy for the diffusion of $^{40}\text{Ar}^*$ in their experiments, hydrothermal degassing experiments (Grove and Harrison 1996) yielded nearly identical results for Fe-rich biotite ($X_{\text{Mg}} = 0.26$) and biotite of intermediate composition ($X_{\text{Mg}} = 0.47$). Instead it was suggested that high halogen contents may increase Ar retentivity. Applying single-domain volume diffusion models for argon transport in biotite of intermediate composition and assuming geologically reasonable cooling rates yields closure temperatures between 310 and 345 °C (Harrison et al. 1985; Grove and Harrison 1996 and references therein).

The halogen contents of the biotite grains investigated in this study are not known. The X_{Mg} values range from 0.32 (BW-116) to 0.70 (BW-132). No systematic relationship between X_{Mg} value and age can be recognised (Tables 2 and 4). Moreover, all biotite ages coincide within error limits. Therefore, the obtained Ar–Ar ages are interpreted to record cooling of the samples below approximately 310–345 °C.

The Ar diffusion systematics of K-feldspar are generally more complicated than those of micas and hornblende, and many $^{40}\text{Ar}/^{39}\text{Ar}$ spectra obtained by incremental heating are rather complex. While earlier studies suggested that many K-feldspars may not be reliable geochronometers due to the continuous loss of $^{40}\text{Ar}^*$ at low temperatures and the incorporation of

excess argon (e.g. Faure 1977), increasing knowledge on the transport behaviour of argon in feldspars has considerably increased the ability to extract geologically meaningful ages (see Harrison and McDougall 1982; Lovera et al. 1989; Foster et al. 1990 and references therein). Further studies indicated that the apparently anomalous $^{40}\text{Ar}/^{39}\text{Ar}$ spectra of K-feldspar reflect the presence of diffusion domains of variable length scales (e.g. Zeitler 1987; Lovera et al. 1989; Foster et al. 1990; Fitz Gerald and Harrison 1993; Lovera et al. 1993) and that this behaviour can be modelled by assuming a discrete distribution of non-interacting domains of variable size (Lovera et al. 1989). The resulting closure temperatures for K-feldspar span up a range between 150 and 300 °C. Recent studies of alkali feldspar microtextures (Parsons et al. 1999) provide a critical view of the multiple diffusion domain theory. They suggest that deuterically altered zones formed by fluid–feldspar reactions at temperatures below 500 °C provide fast pathways for Ar that will be metastable in step-heating experiments. However, the study being performed on very slowly cooling, strongly zoned crystals that reacted with fluid, its implications are not necessarily applicable to other rocks.

Most of the above studies and considerations apply to $^{40}\text{Ar}/^{39}\text{Ar}$ spectra of K-feldspar that do not show plateaus but saddle- or U-shaped spectra, gradients or spectra with two plateau segments (Foster et al. 1990). Most of the K-feldspars analysed in this study show broad plateaus with the exception of the first few steps of incremental heating that mainly have lower, but sometimes also higher apparent ages (Fig. 7). The fact that the domains with anomalous behaviour are degassed during the low-temperature steps suggests that these domains are either located near crystal surfaces, cleavage planes or cracks and may correspond to altered material. Considering a closure temperature between 150 and 300 °C for K-feldspar, the plateau ages of 296.4 ± 3.0 Ma (BW-120), 305.3 ± 3.8 Ma (BW-116) and 305.1 ± 3.6 Ma (BW-44) fit very well to the cooling ages obtained from hornblende and biotite (see above and Table 4). K-feldspars from sample BW-67 do not yield plateaus, but show increasing apparent ages from approximately 280 to 300 Ma, the final degassing steps yielding ages (299.3 ± 2.9 , 300.0 ± 2.9 Ma, Table 4) in agreement with the plateau ages of samples BW-116, BW-120 and BW-44. The rimward decrease in age may be explained by loss of $^{40}\text{Ar}^*$ or by progressive closure of the Ar system down to very low temperatures during slow cooling.

In contrast, the two K-feldspar ages of 319.7 ± 3.1 and 317.8 ± 3.1 Ma (BW-46) from location 7 are significantly higher than the biotite ages from the same location. Moreover, another K-feldspar grain from the same location yields a plateau age of 305.1 ± 3.6 Ma (BW-44). Therefore, the age difference between the grains cannot be due to different cooling rates at the various locations but must reflect either loss of $^{40}\text{Ar}^*$ from the apparently younger grains, or incorporation of

excess argon by the apparently older grains, or different Ar retentivities of the K-feldspars. Loss of $^{40}\text{Ar}^*$ is not apparent from the spectrum of the feldspar from BW-44 and, moreover, its plateau age agrees with those from samples BW-116 and BW-120 (see above). All hornblende and biotite data and the other K-feldspar data represent an internally consistent set, suggesting that incorporation of excess argon is not a problem in these rocks.

Different Ar retentivities are the best explanation for the high ages. They could result from different structural states of K-feldspars. However, as all the K-feldspars are chemically and structurally very similar, we expect the diffusion parameters to be in the same range. Therefore, it is suggested that the effective diffusion radii of the dominant subgrains in the two K-feldspars of sample BW-46 are substantially larger than those of the other K-feldspars. Another possible reason for the elevated Ar–Ar ages of the two K-feldspars is that they contain numerous inclusions of plagioclase. Whereas K-feldspar most likely crystallised from melt produced during dehydration melting reactions, plagioclase could partly represent an older stage of metamorphism as indicated by residual plagioclase crystals in some leucosomes (Berger and Kalt 1999). This hypothesis is supported by partly low K/Ca ratios (9–87) obtained during the incremental heating experiments on feldspars of sample BW-46 compared to significantly higher K/Ca ratios in all other analysed K-feldspars.

In conclusion, most of the analysed K-feldspar grains yield plateau ages or apparent ages of the final steps between 296 and 305 Ma. These ages fit well with those obtained from biotite and hornblende at the same or adjacent localities and probably record cooling below 150–300 °C. Staircase patterns may record closure of the Ar system down to lower temperatures, and older plateau ages of 318–320 Ma most likely reflect diffusion domains with higher closure temperatures.

Cooling rates, duration of metamorphism and possible heat sources

As discussed in the previous sections, U–Pb zircon and monazite ages are interpreted to reflect crystallisation from a melt at approximately 800 °C. For the north-western area, mean leucosome and mesosome monazite ages of 325.3 and 326.1 Ma are considered to date this temperature, although uncertainty remains about inheritance problems. U–Pb titanite ages are interpreted as cooling below 700 °C and $^{40}\text{Ar}/^{39}\text{Ar}$ ages of hornblende and biotite are considered to reflect cooling below 500–570 and 310–345 °C respectively. Most K-feldspars probably record cooling below 150–300 °C. Two feldspar grains from location 7 (BW-46) most likely have higher closure temperatures and are excluded in the following. Taking the mean ages for the minerals and average cooling ages, cooling paths can be constructed for the southeastern area around Bodenmais (locations

1–5) and for the northeastern area around Cham (locations 6–7; Fig. 8).

The two cooling paths are quite similar, although the closure temperatures of the K-feldspars and hence the low-temperature part of the paths are not well-constrained. Both paths are characterised by rapid cooling in the high-temperature range and a non-linear decrease in the cooling rate towards lower temperatures. Considering an age difference of approximately 1 million years between zircon and monazite ages and titanite ages from the southeastern area, cooling rates must have been in the order of 100 °C/my. Further cooling down to the hornblende closure temperature is characterised by a rate of approximately 58 °C/my. For the northwestern area, no titanite ages are available. An average cooling rate of about 46 °C/Ma can be calculated for the temperature span between 800 °C and hornblende closure temperature. The samples further cooled down to the biotite closure temperature at rates of approximately 16 °C/my (southeastern area) and 11 °C/my (northwestern area). Considering the average K-feldspar age of 300 Ma (excluding sample BW-46) to be geologically meaningful, the further cooling rate is 1–17 °C/my. In summary, cooling from 800 to 150–300 °C lasted approximately 20–30 my, provided that exhumation was a continuous process.

Ideally, in order to constrain possible external heat sources and mechanisms of heat input, not only the cooling rates but also the timing of the peak and prograde parts of the P–T path should be known. The only mineral that potentially records all three parts of the P–T path is garnet. Biotite included in garnets does not yield ages significantly older than those of biotite in the matrix. The monazite that is preserved in a garnet core records older ages than those in the matrix. However,

this places no time constraints on prograde or peak metamorphism as the age is not known precisely enough and as it may record either metamorphism of the host rock or detrital inheritance or a combination of the two. More data are needed to test whether the obtained age is representative and reproducible.

The zoning patterns of some very large garnets (5 mm diameter), particularly their bell-shaped Mn profiles, and their inclusion relations indicate that growth of these garnets started at prograde conditions (Kalt et al. 1999, their Fig. 6a). Chakraborty and Ganguly (1991) have examined the extent of relaxation of initial garnet zoning patterns in relation to grain-size, time and diffusion coefficients. Their model calculations for Mn (their Fig. 13) predict that an initially growth-zoned garnet of 5 mm diameter would have been homogenised with respect to Mn by diffusion if temperatures of 850 °C (the estimated peak temperature for migmatites of the Bayerische Wald) prevailed longer than approximately 2.5 million years. As there are many uncertainties in this calculation (the initial zoning pattern is not known, processes other than diffusion may have contributed to the zoning pattern, the fixed edge composition assumed in the calculations may not be valid), the result may just be taken to indicate the order of magnitude for the maximum duration of peak metamorphic conditions. A short heating period fits well with the survival of older monazite grains and with the core-rim structures of zircon grains that also suggest inheritance. No time information can be gained on the prograde part of the P–T path.

As reasoned above, the T–t path of the Bayerische Wald is characterised by a short thermal peak and exponentially decreasing cooling rates. As stated in the introduction, peak temperatures of 850 °C at 0.5–0.7 GPa can only be achieved through external heat sources. Also, exponentially decreasing cooling rates with time are diagnostic of transient heat sources. The rapid heating indicated by the short thermal peak suggests an advective rather than a conductive mechanism of heat transport into the crust. The high peak temperatures attained exclude the abundant granite plutons as heat sources and require the presence of hot, mafic or ultramafic material near to the crustal level represented by the presently exposed gneisses and migmatites of the Bayerische Wald. Indeed, we propose that granite emplacement was the result of massive crustal heating. Therefore, in principle, the thermal evolution of the Bayerische Wald may be modelled by assuming as a heat source a shallow-level asthenosphere in the geodynamic context of lithospheric delamination subsequent to crustal thickening (see Introduction). However, the decompressional features expected in metamorphic rocks that experienced the extension accompanying lithospheric delamination cannot be observed in the migmatites of the Bayerische Wald. These features are observed in regions where lithospheric delamination is independently constrained by geophysical data (Alboran Sea and Tibet).

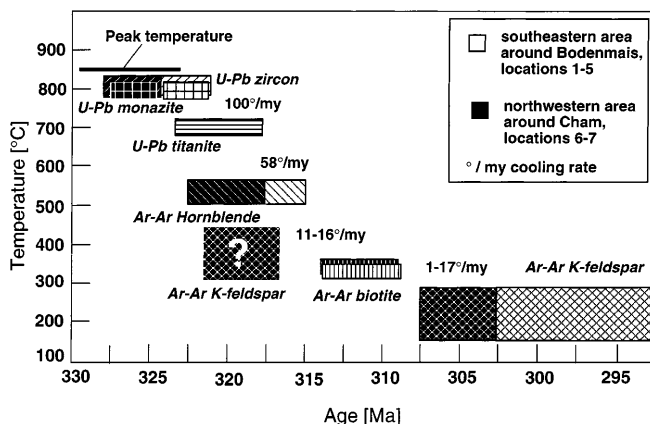


Fig. 8 Temperature–time path derived from the U–Pb and $^{40}\text{Ar}/^{39}\text{Ar}$ dating results. The zoning patterns of large garnets give an upper limit for the duration of peak temperatures (≤ 850 °C) of approximately 2.5 Ma. Note that although no errors can be calculated for the published closure temperatures and the calculated cooling rates, the uncertainty on the given values and ranges is large. For further explanation see section on cooling rates, duration of metamorphism and possible heat sources

Also, the cooling rates obtained for the Bayerische Wald differ from those observed with late-orogenic extensional settings. At high temperatures, the cooling rates of the Bayerische Wald are high (58–100 °C/my), but still lower than the rapid cooling rates (>100–350 °C/my) obtained for example in the Nanga Parbat region (Winslow and Zeitler 1996) or in the Alboran Sea (Soto and Platt 1999 and references therein). At low temperatures, the cooling rates of the Bayerische Wald decrease and are much lower than in the Nanga Parbat or the Alboran Sea. In these settings, rapid denudation in addition to asthenospheric heating is required and cooling rates do not decrease with time which points to exhumation-controlled cooling. The lack of decompressional features and the decrease in cooling rates observed for the Bayerische Wald points to only moderate exhumation rates and to cooling rates controlled by cooling of a transient heat source following magmatic emplacement, as observed with contact metamorphism. In summary, the boundary conditions for the Bayerische Wald established here and by Kalt et al. (1999) suggest that the thermal evolution of a part of the internal Variscan belt may be different from the settings described above. However, modelling the thermal evolution of the Bayerische Wald is beyond the scope of this paper and will be the issue of a separate study.

Summary

A T–t path has been constructed for migmatites of the Bayerische Wald based on the petrological data presented by Kalt et al. (1999), on U–Pb ages of zircon, monazite and titanite and on $^{40}\text{Ar}/^{39}\text{Ar}$ ages of hornblende, biotite and K-feldspar. Additionally, crude estimates of the duration of peak metamorphism were derived from garnet zoning patterns. The obtained T–t path is characterised by a short thermal peak (<2.5 Ma) at 850 °C/0.5–0.7 GPa between 326 and 323 Ma in the southwestern part of the investigated area and between 329 and 326 Ma in the northwestern area. Whether this age difference is geologically meaningful is rendered uncertain by the inheritance features of the monazites and the duration of the heating event. The fact that inherited monazite is preserved at temperatures of 850 °C places constraints on the U–Pb closure temperature. Interpretation of monazite ages may be complicated under such conditions. The further T–t path of the migmatites is virtually identical in both investigated areas of the Bayerische Wald. It is characterised by a decrease in cooling rates from 100 °C/my between 322 and 321 Ma to 11–16 °C/my between 318 and 311 Ma. The further cooling path is not well constrained, but the $^{40}\text{Ar}/^{39}\text{Ar}$ ages of K-feldspar suggest even slower cooling. The high peak temperatures (850 °C) at comparatively shallow crustal levels along with the short thermal peak are compatible with external heating by hot mafic or ultramafic material. The obtained moderate cooling rates and the decline of cooling rates with time are not

easily reconciled with a model of asthenospheric heating, rapid uplift, denudation and extension due to lithospheric delamination as described for example from the Alboran sea. Instead, cooling after underplating of mafic magmas seems to be a more viable mechanism.

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