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Zircon thermometry and U–Pb ion-microprobe dating of the gabbros and associated migmatites of the Variscan Toledo Anatectic Complex, Central Iberia

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Abstract: In the Central Iberian Zone there are several large thermal domes in which small bodies of ultramafic, mafic and intermediate rocks appear intimately associated with crustal granites and migmatites. The closest spatial association between the ultramafic, mafic and intermediate rocks and migmatites is in the Toledo Anatectic Complex, where field relationships suggest that these rocks are coeval and have an age close to 340 Ma. This, and the recent discovery in the neighbouring Ossa Morena Zone of a large mid-crustal seismic reflector interpreted as a 335–350 Ma mafic sill, reinforce the hypothesis that heat for crustal melting was supplied from early Variscan mantle magmas emplaced in the middle crust. However, precise ion-microprobe U–Pb zircon dating and Ti-in-zircon thermometry in Toledo do not support this idea. Whereas the mean age of four mafic bodies is 307 ± 2 Ma, the migmatites are *c.* 25 Ma older. The migmatites hosting ultramafic, mafic and intermediate bodies have the same age and Ti-in-zircon temperatures as migmatites far from any mafic intrusion. These data reveal that ultramafic, mafic and intermediate magmas are late Variscan; they were emplaced in already cooling anatectic zones once the extensional collapse was initiated, and their thermal impact on the mid-crustal Variscan anatexis of Central Iberia was negligible.

The Iberian Massif (Fig. 1) is the largest and the best preserved segment of the Variscan belt of western Europe (Pérez-Estaún & Bea 2004). It consists of several zones with different stratigraphic, structural, magmatic and metamorphic characteristics. The largest is the Central Iberian Zone, which from *c.* 330 to 295 Ma was the locus of abundant crustal granite magmatism with little, if any, detectable mantle component (Bea 2004, and references therein). Roughly aligned along the axis of the Central Iberian Zone there are several thermal domes such as Toledo, Gredos and Sayago (also called Tormes Dome) (Fig. 1) in which small volumes of ultramafic, mafic and intermediate rocks appear intimately associated with large volumes of crustal granites and migmatites. Field relationships and Rb–Sr dating (e.g. Bea *et al.* 1999) suggest that the ultramafic, mafic and intermediate rocks predate or are coeval with the spatially associated granites and migmatites. This association is the main basis for the hypothesis, implicitly assumed by many workers although never demonstrated, that mantle-derived magmas have played an essential role in the high-*T* metamorphism and granite production in Central Iberia during the Variscan orogeny (e.g. Franco González & García De Figuerola 1986; Pinarelli & Rottura 1995; Castro *et al.* 2003; López-Moro & López-Plaza 2004).

The Ossa Morena Zone, to the south of Central Iberia (Fig. 1), also contains abundant Variscan granitoids although of a different nature. Their compositional spectrum is more varied and metaluminous varieties with detectable mantle components are common (Casquet & Galindo 2004). Here, the recent IBERSEIS deep-seismic reflection profile (Fig. 1) discovered a *c.* 200 km long, up to 1 km thick mid-crustal seismic reflector, the IBERSEIS Reflective Body, interpreted as a *c.* 335–350 Ma mafic–ultramafic sill that was probably the source of the *c.* 330–340 Ma ultramafic to intermediate rocks of that area (Simancas *et al.* 2003).

The finding of the IBERSEIS Reflective Body in Ossa Morena has reinforced the idea that in Central Iberia the heat for crustal melting and the material for the ultramafic, mafic and intermediate intrusions could also have been supplied from Early Variscan mafic magmas emplaced in the middle crust. Recent precise geochronological data, however, do not support this hypothesis: single-crystal zircon dating of similar rocks in Gredos (Fig. 1) has revealed that most ultramafic, mafic and intermediate bodies have ages between 310 and 313 Ma (Montero *et al.* 2004*a*), and are about 20 Ma younger than the anatexis peak and almost 40 Ma younger than the beginning of melting (Montero *et al.* 2004*b*), so they could hardly have provided the heat for anatexis. None the less, it could be argued that dating in Gredos is not conclusive because the studied ultramafic, mafic and intermediate bodies and migmatites are not directly in contact.

Undoubtedly, the best place in Central Iberia for checking the age relationships between mafic magmas and crustal melting is the Toledo Anatectic Complex, where the ultramafic, mafic and intermediate rocks (up to 20 wt% MgO) form small sills or irregular bodies emplaced into the migmatites. Although none of these rocks have been radiometrically dated, Barbero (1995) and Barbero & Villaseca (2004) suggested, on the basis of geological and petrographical evidence, that the ultramafic, mafic and intermediate rocks were emplaced just prior to the anatexis peak. Those workers assumed an age for both mafic intrusions and migmatites close to 340 Ma, identical to the Ossa Morena IBERSEIS Reflective Body. As this assumption is compatible with a cause-and-effect relationship between mafic magmatism and crustal melting, but contrary to what geochronology of

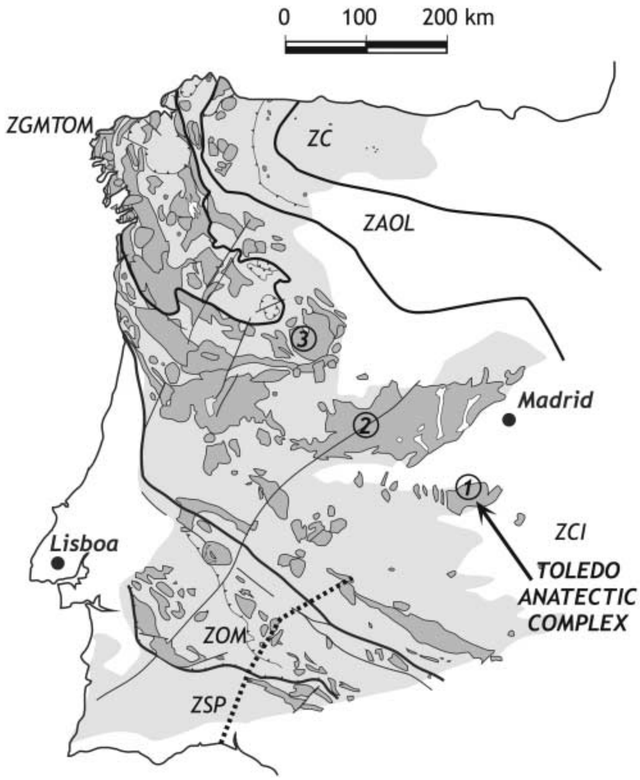

Fig. 1. Map of the Iberian Massif (light grey). Dark grey areas represent plutonic rocks, mostly granite–granodiorite. Bold continuous lines represent the boundaries between the palaeogeographical zones that make up the massif, according to Farias *et al*. (1987). ZC, Cantabrian Zone; ZAOL, Western Asturian Leonian Zone; ZGTOM, Galicia Medial Tras-os-Montes Zone; ZCI, Central Iberian Zone; ZOM, Ossa Morena Zone; ZSP, South Portuguese Zone. Numbers represent the three largest thermal domes of the Central Iberian Zone: 1, Toledo; 2, Gredos; 3, Sayago (also called Domo del Tormes). The dashed line represents the IBERSEIS seismic profile (Simancas *et al*. 2003).

practically identical rocks has shown in Gredos, we believed it urgent to check the two possibilities by radiometric methods.

This study aimed to obtain the precise zircon age of the ultramafic, mafic and intermediate rocks and migmatites of Toledo, to evaluate the thermal impact caused by ultramafic, mafic and intermediate intrusions, and to understand why age estimates based on Rb–Sr dating and field relationships are usually misleading. We studied zircon grains separated from the four main outcrops of mafic rocks and from two migmatites (one leucocratic diatexite collected at the contact with the gabbros (Fig. 2), and one mesocratic metatexite far from any mafic intrusion). Ti-in-zircon thermometry (Watson & Harrison 2005) and Zr/Hf systematics proved that zircon grains from the ultramafic, mafic and intermediate rocks are not xenocrystic, and U–Pb ion-microprobe dating showed they are considerably younger than the Variscan zircons of the migmatites, which, notably, were not rejuvenated near mafic intrusions. The nature of the accessories assemblage as well as Sr and Nd isotopes revealed that ultramafic, mafic and intermediate rocks, even the most magnesian, always represent a mixture between mantle and crust materials. This may cause $^{87}Sr/^{86}Sr$ v. $^{87}Rb/^{86}Sr$ regression lines with an excellent goodness of fit but wrong ages. Our results support the idea that the emplacement of mafic magmas

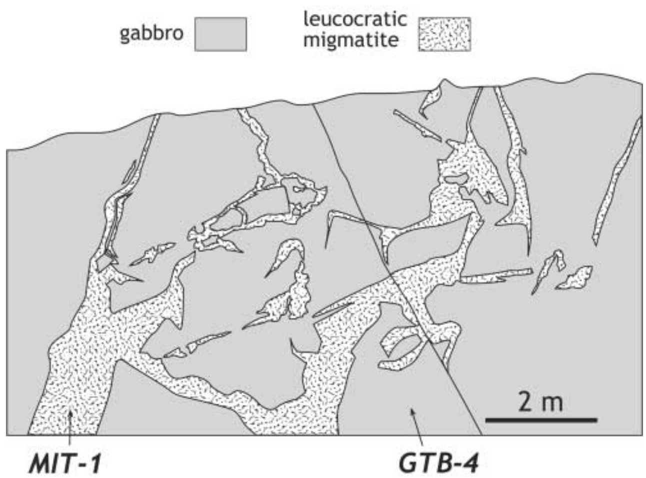

Fig. 2. Field relations between La Bastida gabbro sill and host migmatites (part of the Toledo Anatectic Complex; 1 in Fig. 1), drawn from a photograph. The gabbro locally shows fine-grained contact facies but, despite this, it is locally brecciated and invaded by pegmatitic leucosome-rich nebulites. The mean zircon U–Pb age of the gabbro is about 25 Ma younger than the migmatite. MIT-1 and GBT-4 indicate where these samples (see Table 1) were collected.

in the middle crust contributed little to the widespread Variscan anatexis in Central Iberia.

Geological setting

The Toledo Anatectic Complex (see Barbero & Villaseca 2004) is located in central Spain (Fig. 1). Its dimensions are about 90 km from east to west and up to 30 km from north to south. To the north, it is in contact with nonmetamorphic sediments as a result of a system of vertical faults. To the south, it is in contact with low-grade metasediments of early Palaeozoic–Neoproterozoic age and intrusive Variscan granites as a result of a listric late Variscan shear zone. Materials are mostly pelitic and psammitic migmatites equilibrated at peak conditions of 800 ± 50 °C and 4–6 kbar (Barbero 1995), and diverse peraluminous granitoids. Scattered in the complex there are small outcrops of ultramafic, mafic and intermediate rocks, which, according to Barbero & Villaseca (1989), may be grouped into two types: La Bastida-type and Toledo-type. The first is composed of appinitic high-magnesium gabbros to diorites; the second is composed of vaugneritic high-K amphibole–biotite gabbros to quartzdiorites (see also Bea 2004).

The La Bastida-type gabbros contain olivine (Fo_{73}), Ti-rich amphibole (kaersutite and Ti-pargasite with 4.2–3.8 wt% TiO_2, locally rimmed by cummingtonitic amphibole), Ti-rich phlogopite (5.7–7.1 wt% TiO_2), clinopyroxene, orthopyroxene (En_{75-77}) and plagioclase (An_{38-63}). Notably, anhedral phlogopite appears rimmed by Ti-rich amphibole or as tiny inclusions in amphibole and orthopyroxene, suggesting phlogopite resorption during magmatic crystallization. Subsolidus orthopyroxene coronas develop at the interface between olivine and plagioclase. The accessories are apatite, Fe–Ni sulphides, ilmenite, zircon accompanied by tiny (<50 µm) rounded grains of abundant monazite and rare $ThSiO_4$ (huttonite?) and uraninite, in most cases included in major minerals. Monazite, huttonite and uraninite with identical textural relationships are also found in the cortlanditic gabbros of Gredos, where they are considered xenocrystals as a result of crustal assimilation (Bea *et al*. 1999). As discussed below, this poses the problem of recognizing whether zircon currently found in gabbros is primocrystic or xenocrystic.

The Toledo-type gabbros contain neither olivine nor Ti-rich silicates, and orthopyroxene is scarce. Their Fe–Mg assemblage consists of minor augitic clinopyroxene and abundant pargasitic to edenitic amphibole and Mg-rich biotite (the latter being more dominant in the more silicic rocks) and occasional orthopyroxene. The plagioclase is An_{23-54}, and the most felsic facies contain quartz and, locally, K-feldspar. As accessories they have abundant apatite, ilmenite, zircon, and Fe–Ni, Fe–Cu and Pb sulphides; monazite and $ThSiO_4$ (huttonite?) are also common. Minor titanite and primary-looking epidote are also present in the more felsic (quartzdioritic) facies.

The migmatites show a large compositional variation from leucocratic to mesocratic varieties (e.g. Table 1). The leucocratic varieties are generally diatexitic, with a large leucosome to melanosome ratio. The leucosome has a hypidiomorphic granular texture, and is composed of quartz, K-feldspar, oligoclase, abundant prismatic cordierite, biotite and subordinate garnet, with apatite, zircon, monazite, huttonite and rare xenotime as accessories; myrmekites are common. Dispersed within the leucosome there are small streaks and rounded enclaves of melanosome composed of a granoblastic intergrowth of rhomboidal sillimanite, green spinel, ilmenite and minor sulphides. The enclaves of melanosome are frequently surrounded by large crystals of granoblastic cordierite. The mesocratic varieties are either diatexitic or metatexitic, with a leucosome similar to that described above, and a foliated mesosome composed of abundant biotite, cordierite and quartz, subordinate K-feldspar and plagioclase, and minor garnet.

Field relations between migmatites and gabbros are complex but, in general, they suggest that the two rocks are coeval. The gabbros are intrusive in the migmatites and locally show chilled margins but, at the same time, they may appear locally brecciated and invaded by migmatites (Fig. 2). It is worth mentioning that the composition of the migmatite ‘dykes’ that cut the gabbros usually changes with the width of the filled fracture: the widest ones consist of a normal migmatite, which gradually becomes more felsic and pegmatoid as the filled fracture narrows.

Samples and methods

Six samples, consisting of four mafic rocks (one from each of the four largest ultramafic, mafic and intermediate bodies: Toledo, Argés, La Bastida and Guajaraz) and two migmatites (MIT-1, a leucocratic nebulite just in contact with La Bastida gabbro (see Fig. 2), and MIT-2, a pelitic metatexitic migmatite far from any mafic body), were collected for zircon separation. The locations, major and some relevant trace elements, and Sr and Nd isotopes of these samples are given in Table 1. Two mafic bodies, Toledo and Argés, are vaugneritic, as reflected by the elevated K_2O and Zr contents of the samples. The other two, La Bastida and Guajaraz, are appinitic (Bea 2004).

Zircon was separated using conventional magnetic and heavy-liquid techniques. Once mounted and polished, zircon grains were studied by cathodoluminescence (CL) imaging and analysed for U–Th–Pb using a Cameca IMS1270 ion microprobe at the Nordsim facility in Stockholm, and for Ti and Zr/Hf using the laser-ablation inductively coupled plasma

Table 1. *Location (UTM, zone 30T), major and some trace element, and Sr and Nd isotope composition of studied samples*

Sample:	GBT-1	GBT-2	GBT-3	GBT-4	MIT-1	MIT-2
Lithology:	Gabbro	Tonalite	Gabbro	Gabbro	Migmatite	Migmatite
Mafic unit:	Toledo	Argés	Guajaraz	La Bastida	–	–
UTM coordinates						
Easting:	413022	405744	407963	410140	410140	411529
Northing:	4411969	4407888	4407299	4412553	4412553	4407933
Major elements (%)						
SiO_2	48.97	59.11	50.13	49.56	72.24	50.04
TiO_2	1.61	1.58	1.61	0.99	0.25	1.39
Al_2O_3	19.95	17.38	15.32	14.48	15.35	24.27
FeO tot.	8.18	6.16	9.32	9.09	1.12	10.35
MgO	4.99	3.08	10.28	13.97	0.29	5.93
MnO	0.11	0.08	0.15	0.15	0.01	0.18
CaO	8.13	4.92	6.95	6.98	0.86	0.98
Na_2O	2.76	1.78	1.4	1.9	2.45	0.98
K_2O	2.85	3.02	1.51	0.71	6.37	2.16
P_2O_5	0.54	0.61	0.52	0.14	0.21	0.08
LOI	1.17	1.14	0.98	0.38	0.55	1.67
Total	99.26	98.86	98.17	98.35	99.7	98.03
Trace elements and isotope ratios						
Zr (ppm)	355	368	275	37.3	94	206
Zr/Hf	37.1	43.2	37.1	38.1	28.5	36.9
Rb (ppm)	131	142.3	80	18.9	205	120
Sr (ppm)	1143	537.5	628	321.3	135	121
$^{87}Rb/^{86}Sr$	0.3304	0.766	0.368	0.17	4.402	2.871
$^{87}Sr/86Sr$	0.706683	0.71035	0.707103	0.705394	0.732721	0.727559
Nd (ppm)	63.28	69.49	54.77	12.94	20.37	21.78
Sm (ppm)	13.28	12.24	8.5	3.06	4.66	4.19
$^{147}Sm/^{144}Nd$	0.1269	0.1065	0.0939	0.143	0.1383	0.1163
$^{143}Nd/^{144}Nd$	0.512363	0.512198	0.512325	0.512496	0.512173	0.512158
Summary of zircon 207-corrected age (95% confidence interval)						
n determinations	16	4	12	13	10	4
Mean age (Ma)	308 ± 2	309 ± 4	311 ± 5	306 ± 2	332 ± 5	334 ± 14

Second row of $^{87}Sr/^{86}Sr$ and the $^{143}Nd/^{144}Nd$ values are measured values.

mass spectrometry (LA-ICP-MS) system of the University of Granada. Ion microprobe analytical methods broadly follow those described by Whitehouse *et al*. (1999, and references therein). U/Pb and Th/Pb ratios were calibrated using the Geostandards 91500 reference zircon (1065 Ma; Wiedenbeck *et al*. 1995) and include a propagated error component from replicate analyses of 91500 during the analytical session. Errors on $^{207}Pb/^{206}Pb$ ratios are either the observed analytical uncertainly or the counting statistics error, whichever is higher. Common Pb corrections assume that most contaminant Pb is present on the surface of the analysed grains, introduced from the sample preparation process, and has a composition that can be approximated using the Stacey & Kramers (1975) model for the present day. The '207-corrected' ages for grains with ages <500 Ma are available online at http://www.geolsoc.org.uk/SUP18242. A hard copy can be obtained from the Society Library. They were calculated by projecting the uncorrected analysis onto concordia from the assumed common $^{207}Pb/^{206}Pb$ composition. In most cases, however, the amount of common Pb, revealed by monitoring ^{204}Pb, is relatively small and has little influence on the interpreted age. All ages are calculated using the decay constant recommendations of Steiger & Jäger (1977).

LA-ICP-MS analyses of Ti, Zr, Hf, Th, U and Pb isotopes were carried out with a Nd-YAG 213 nm Mercantek laser and a torch-shielded quadrupole Agilent 7500 ICP-MS system. To avoid the isobaric interference caused by $^{96}Zr^{2+}$, Ti was determined on the isotope ^{49}Ti (5.5%) instead of the most abundant ^{48}Ti (73.8%). The laser beam was set at a diameter of 60 μm, with a repetition rate of 10 Hz and an output energy of 75%. The ablation time was 60 s and the spot was pre-ablated for 45 s with a laser output energy of 50%. The ablation was carried out in a He atmosphere. The internal standard was ^{91}Zr. The external standard was the NIST-610 glass with the element concentration values recommended by Pearce *et al*. (1997) and the isotope ratios determined by thermal ionization mass spectrometry (TIMS) at the University of Granada (Bea *et al*. 2006*b*). LA-ICP-MS U–Pb ages are in good agreement with ion-microprobe data but show more dispersion and tend to be more discordant. They were used only to determine whether the analysed spot in zircon from migmatites gave a Variscan age. The precision (1σ) estimated on 10 replicates of NIST-610 analysed in the same run was better than 2.5% for element ratios and about 0.3% for isotope ratios. The detection limit for Ti was about 0.4 ppm, and the precision (1σ) estimated on the standards NIST-610, NIST-612 and a homemade glass was 4%, 8% and 20% for Ti concentrations of 434 ppm, 48 ppm and 12 ppm, respectively. Ti-in-zircon temperatures were estimated with the formula of Watson & Harrison (2005): $T(\mathrm{K}) = 5080/(6.01 - \log_{10}(\mathrm{ppmTi}))$ assuming a TiO_2 activity is unity.

Samples for Sr and Nd isotope analysis (0.1000 g) were digested with HNO_3+HF in a Teflon-lined vessel at *c*. 180 °C and *c*. 1.38 GPa for 30 min, evaporated to dryness, dissolved in 6.5N HCl, separated with ion-exchange resins, and analysed by TIMS in a Finnigan Mat 262 at the University of Granada. All reagents were ultra clean. Normalization values were $^{86}Sr/^{88}Sr = 0.1194$ and $^{146}Nd/^{144}Nd = 0.7219$. Blanks were 0.6 and 0.09 ng for Sr and Nd, respectively. The external precision (2σ), estimated by analysing 10 replicates of the standard WS-E (Govindaraju *et al*. 1994), was better than 0.003% for $^{87}Sr/^{86}Sr$ and 0.0015% for $^{143}Nd/^{144}Nd$. $^{87}Rb/^{86}Sr$ and $^{147}Sm/^{144}Nd$ were directly determined by ICP-MS at Granada following the method developed by Montero & Bea (1998), with a precision better than 1.2% and 0.9% (2σ), respectively.

Major elements and Zr were determined at the University of Granada by X-ray fluorescence after fusion with lithium tetraborate. The precision (1σ) was better than 1.5% for an analyte concentration of 10 wt%, and ±5% for 100 ppm Zr.

Zircon morphology

Ultramafic–mafic–intermediate rocks

The four samples of ultramafic, mafic and intermediate rocks contained abundant zircon, but the size and morphology varied greatly between vaugneritic and appinitic types. The first type, from Toledo and Argés, contained the largest and most euhedral grains. Zircon grains from the Toledo gabbro are colourless, euhedral, commonly prismatic, up to 500 μm × 200 μm, with little zoning, no inclusions and very limpid, almost gem-quality (Fig. 3). They show a few embayments and small overgrowths.

Zircon from the Argés tonalite tends to form needle-like subhedral or anhedral crystals, up to 250 μm × 50 μm (Fig. 3). The zircon grains are colourless or, more frequently, pale yellow to brownish. The internal morphology is always complex, frequently with highly cathodoluminescent areas surrounding zones similar to zircon from the Toledo gabbro. Considering that this sample is a tonalite (see Table 1) the morphology of its zircon could have resulted from repeated resorption and growth of zircon grains, initially similar to Toledo, during protracted magmatic evolution.

Zircon morphology in the two appinitic gabbros is similar. It consists of anhedral, rarely subhedral, small (150 μm × 30 μm) yellow to brownish grains which, under cathodoluminescence, characteristically show irregularly distributed bright and black areas. Zircon grains from La Bastida gabbro (Fig. 3) frequently show one or more internal areas with fine oscillatory zoning and a highly cathodoluminescent irregular rim. Zircon grains from Guajaraz gabbro (Fig. 3) are more irregular. They commonly lack alternating zoning, except in a few grains (e.g. Fig. 3, grain c3) which contain a texturally discordant internal area with a tenuous alternating structure very similar to the zoning shown by the Variscan zircon of host migmatites. As described in the next section, these discordant zones yielded slightly abnormal older ages, which might indicate a xenocrystic origin.

Migmatites

Zircon grains from migmatites are either needle-like euhedral or subhedral prismatic crystals with Variscan ages (Fig. 3) or stubby prismatic to bipyramidal crystals usually with a Variscan rim and a pre-Variscan core. The morphology and internal structure is, in all cases, sharply different from that of gabbros. As mentioned above, only some zircon grains found in the Guajaraz appinitic gabbros show a few cores with morphology and age similar to those of the migmatites.

U–Pb dating

Ultramafic–mafic–intermediate rocks

Despite the large morphological diversity displayed by the zircon grains from the four ultramafic, mafic and intermediate rocks, they yield notably uniform concordant or nearly concordant U–Pb ages (Fig. 4) (The data are available online, see above.). Sixteen determinations on the Toledo gabbro yielded an average of 308.2 ± 1.4 Ma (confidence intervals are always reported at 95% confidence level). Four determinations on the Argés tonalite yielded 308.7 ± 4.4 Ma. Twelve determinations on the Guajaraz gabbro yielded 311.1 ± 5.2 Ma. Lastly, 13 determinations on the La Bastida gabbro yielded 305.6 ± 2 Ma. Considering the 45 U–Pb determinations together, the mean age of the Toledo mafic rocks is 308.3 ± 1.7 Ma. These data can be further refined if we consider that ages older than 315 Ma were found only in zircon cores from the Guajaraz gabbro (Fig. 3, grain c3), which may be xenocrystic and, therefore, yielded either migmatite or mixed migmatite–gabbro ages (Fig. 5). Excluding these values, the average of 40 determinations is 306.8 ± 1.2 Ma, in round numbers 307 ± 2 Ma, which seems to be the best estimate for the age of the Variscan mafic magmatism in the Toledo Anatectic Complex.

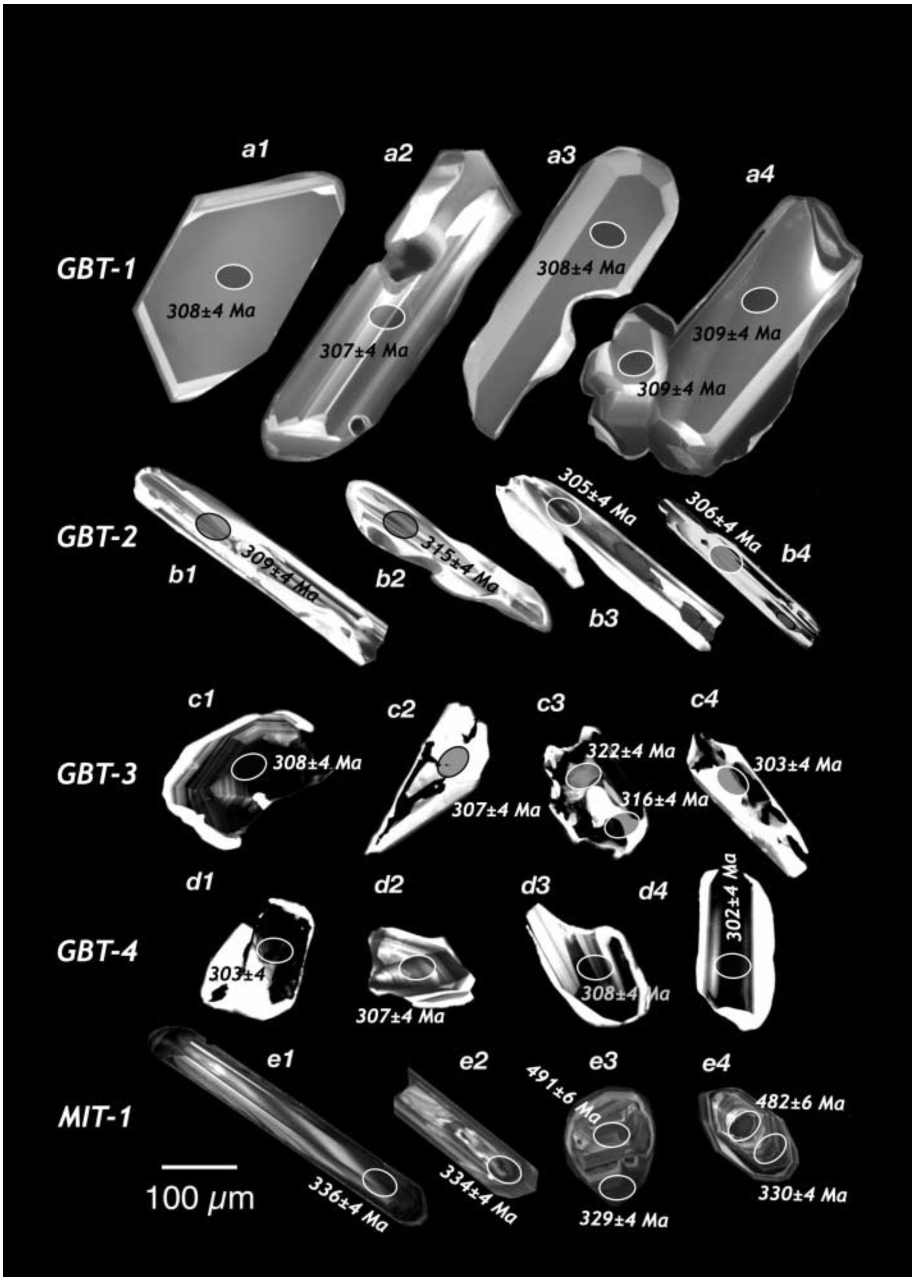

Fig. 3. Cathodoluminiscence images of zircons from gabbros and migmatites. The ellipses represent the spot analysed with the ion-microprobe and the neighbouring number the 207-corrected U–Pb age. The large variation in zircon morphology of gabbros should be noted. Zircon grains from the vaugneritic Toledo gabbro (GBT-1) are very limpid and uniform. Those from the vaugneritic Argés tonalite (GBT-2) show many texturally discordant zones, and the internal areas are somewhat similar to the zircon from the Toledo gabbro. Zircon grains from both appinitic gabbros, GBT-3 and GBT-4, are smaller, irregular, and characteristically contain alternating high- and low-cathodoluminescence areas. Despite this variety, they yielded a uniform age of 307 ± 2 Ma. Only in the Guajaraz gabbro, GBT-3, are these inherited cores, probably xenocrysts from the migmatites, which yielded ages that are either migmatite ages or, when the ion beam was larger than the core, mixed migmatite–gabbro ages.

Migmatites

In this study, we are primarily interested in the age of the migmatization, so that the pre-Variscan inherited ages found in zircon grains from the migmatites, mostly Early Ordovician to Cambrian, will be not discussed here. In the migmatite MIT-1, collected at the contact with La Bastida gabbro, we obtained 10 Variscan ages from 11 determinations, with an average of 331 ± 5 Ma (the data are available as a Supplementary Publication, see p. 4). In the pelitic migmatite MIT-2, far from any mafic body, we obtained four Variscan ages from seven determinations. The average is 334 ± 14 Ma. A *t*-test comparing the mean of the two migmatites shows that they are statistically indistinguishable. Considering all the Variscan data together, the age of the migmatization reflected by these samples is 332 ± 5 Ma, about 25 Ma older than the ultramafic, mafic and intermediate rocks (Figs 4 and 5).

Zircon thermometry and Zr/Hf variations

The Ti-in-zircon thermometer recently developed by Watson & Harrison (2005) seems highly promising for understanding the crystallization history of magmatic and high-grade metamorphic rocks, especially when it is combined with U–Pb and Zr/Hf studies (e.g. Lowery *et al*. 2006). Although the thermometer was originally calibrated for systems with TiO_2 activity of unity, Watson & Harrison, from the consideration that the same factors that led to zircon saturation also led to high TiO_2 activity, have stated that it will not underestimate the growth temperature of zircon by more than 50–60 °C in most rocks except peralkaline ones. It follows, therefore, that rocks with similar Ti-bearing mineral assemblages would give comparable results, and that the differences in temperature recorded by zircon grains from the same rock would accurately reflect its thermal history. Using this technique we studied zircon from three samples containing only

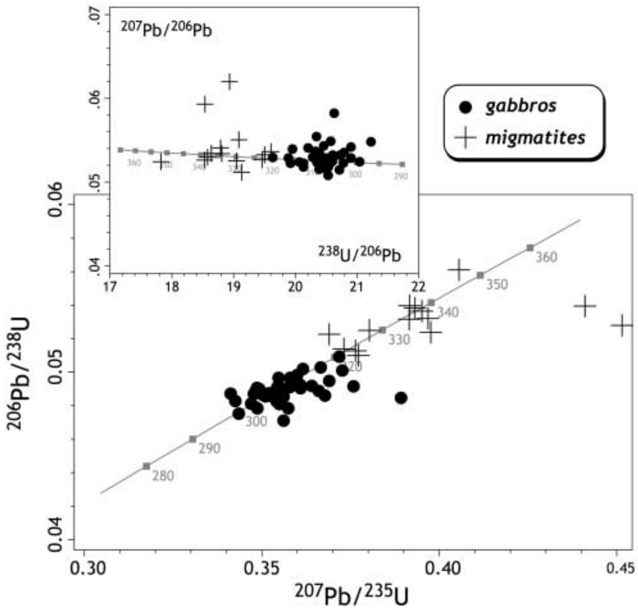

Fig. 4. Conventional and Tera–Wasserburg concordia plots of gabbros (●) and migmatites (+; only the Variscan ages). Data are common-lead uncorrected. Most points are concordant or nearly concordant; also, the difference between the two rock types should be noted.

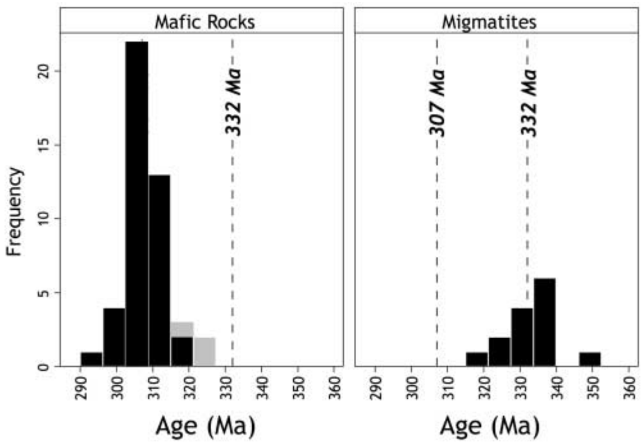

Fig. 5. Distribution of 207-corrected U–Pb zircon ages in gabbros ($n = 45$) and migmatites ($n = 14$). In gabbros, practically all values higher than 315 Ma appear in the Guajaraz gabbro (GBT-3), which contains inherited xenocrystic cores from the migmatites (represented as grey bars). Excluding these, the average of the remaining 40 determinations is 307 ± 2 Ma, which represents the best age estimate for the mafic magmatism of the Toledo Anatectic Complex. (Note how the distribution of Variscan ages in migmatites yields a maximum at *c*. 335 Ma and is asymmetrically tailed towards younger values.)

ilmenite and biotite (TiO_2 *c*. 2–3 wt%) as the only Ti-rich phases: the Toledo gabbro GBT-1, which also has the largest zircon crystals (Fig. 3), and the two migmatites MIT-1 and MIT-2, located, respectively, close to and far from mafic intrusions. Results (available as a Supplementary Publication, see p. 850) reveal the following.

Zircon from the gabbro shows a well-defined positive correlation between Ti-in-zircon temperatures and Zr/Hf (Fig. 6). As

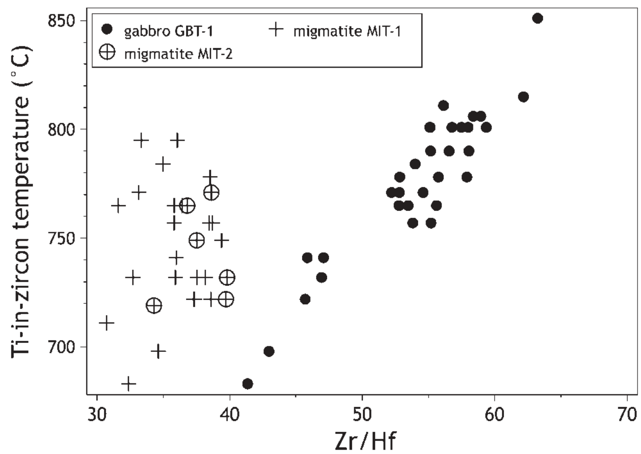

Fig. 6. Zr/Hf v. Ti-in-zircon temperature (Watson & Harrison 2005) plot for the gabbro GBT-1 and the two migmatites. The three rocks contain ilmenite and biotite (TiO_2 *c*. 2–3 wt%) as the only Ti-rich minerals and results are comparable. Remarkably, the highest temperature estimations for gabbros and migmatites agree well with the estimation using conventional thermometers (see text). The excellent positive correlation in the gabbro reveals that its zircon is magmatic. (Note how the two migmatites, one in contact with a mafic body and the other far from any intrusion, yield the same temperature, thus indicating that the thermal impact of mafic intrusions was minimal.)

discussed below, this is exactly what one would expect from sequential magmatic crystallization. Remarkably, the highest temperature recorded in zircon (851 °C) compares well with the temperature estimated for these rocks using the opx–cpx thermometer (*c*. 850 °C, Barbero & Villaseca 2004). Significantly, zircon from migmatites has much lower Zr/Hf than in the gabbros so that in Figure 6 they plot in different regions. The average temperature of the two migmatite samples is virtually the same, 746 °C in MIT-1 and 743 °C in MIT-2. It seems, therefore, that MIT-1 does not reflect any thermal impact related to the intrusion of La Bastida gabbro. Again, the highest Ti-in-zircon temperature recorded in the migmatites (791 °C) is very close to that yielded by the garnet–biotite thermometer (*c*. 800 °C, Barbero & Villaseca 2004).

Rb–Sr dating of ultramafic, mafic and intermediate rocks: misleading ages

It is worth considering why these mafic rocks, which are spatially related to migmatite, were classically thought to be coeval with or older than the associated granites and migmatites. This belief was founded not only on field relationships (see Fig. 2) discussed in detail below, but also on Rb–Sr dating. For example, the Prado de las Pozas body, in Gredos, initially dated by Rb–Sr at *c*. 440 Ma (Pereira *et al*. 1992), later showed a precise Pb–Pb single-zircon age of 312 ± 3 Ma (Montero *et al*. 2004*a*). The reasons for this effect are beautifully exemplified in rocks from the Toledo Anatectic Complex. A $^{87}Sr/^{86}Sr$ v. $^{87}Rb/^{86}Sr$ regression line for the four mafic samples studied in this work, each from a different massif, fits the isochron model 1 of York (1969), yielding an age of 584 ± 50 Ma with an MSWD of 1.89 (Fig. 7; data in Table 1). Were it not known that 40 nearly concordant U–Pb ion microprobe zircon data for the same samples yielded 307 ± 2 Ma, this regression line would surely lead anybody to assume a pre-Variscan age for these rocks. However, the reason

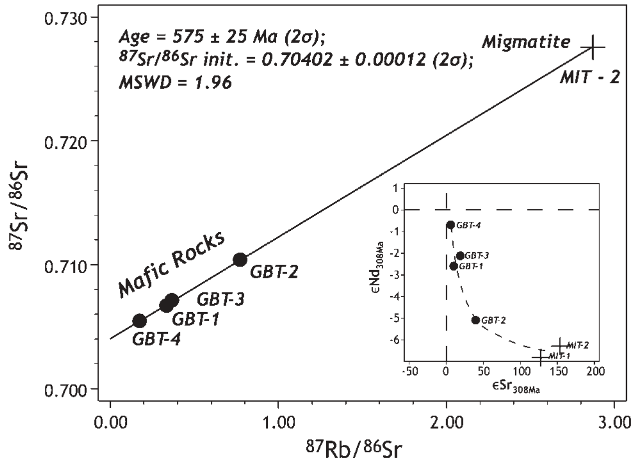

Fig. 7. Fictitious Rb–Sr isochron fitted by the four ultramafic–mafic–intermediate rocks and the regional migmatite MIT-2, which fits model 1 of York (1969). Excluding the migmatite, the four ultramafic–mafic–intermediate rocks yield almost the same result (see text). This is because the mafic magmas are hybrid, and probably have been so since their generation (Bea *et al*. 1999), so that the regression line represents in fact a mixing line between a mantle and a crustal end-member, as also reflected by Nd isotopes (see the inset). This effect is characteristically shown by the Variscan mafic rocks of Central Iberia (Montero *et al*. 2004*a*), which, therefore, cannot be dated accurately with the Rb–Sr method.

for this fictitious Rb–Sr isochron becomes clear when the sample of the metatexitic migmatite MIT-2 is also included in the figure, because it also plots exactly on the regression line of the gabbros. The five samples together, four gabbros and one migmatite, again fit York's model 1, yielding a fictitious isochron of 575 ± 25 Ma (2σ) with an MSWD of 1.96 (Fig. 7). Only from precise zircon U–Pb data can we realize that the Rb–Sr isochron age of gabbros is an artefact and the fitted line represents a mixing line between mafic or ultramafic magmas and migmatitic crustal materials. This idea is also supported by Nd isotopes (Fig. 7).

Discussion

Sr and Nd isotopes (Fig. 7) and the presence of crustal accessory phases, such as monazite, huttonite (or thorite) and uraninite, included in major minerals indicate that the gabbros were heavily contaminated with crustal materials. It is necessary, therefore, to ascertain whether zircon from the gabbros is magmatic or xenocrystic. The variations in Zr/Hf and Ti-in-zircon temperatures have been very useful for this purpose. The sharp difference in Zr/Hf between zircon from the gabbro (41–63) and the migmatites (31–40, Fig. 6) precludes the latter as a source of zircon in gabbros but does not, however, prove a magmatic origin; the best evidence of this comes from the higher than whole-rock Zr/Hf and the positive correlation between Zr/Hf and Ti-in-zircon temperatures shown by zircon from the gabbro. These features indicate that the system was depleted in Hf relative to Zr before zircon crystallization and that, once saturation was reached, zircon grew continuously from a progressively cooler, and increasingly depleted in Zr relative to Hf, environment. This scenario represents the crystallization of a mafic magma from which amphibole or clinopyroxene (up to 150 ppm Zr, Zr/Hf *c*. 20–27; Bea *et al*. 2006) was fractionated before it was saturated in zircon. As this is exactly what we infer to have happened in the Toledo gabbro, we must accept that the zircon found in it is magmatic.

Once this conclusion is reached for the gabbro GBT-1, the identical U–Pb age of zircon from the other ultramafic, mafic and intermediate rocks leads us to accept that the zircon in those rocks is also magmatic, and that the mean of 307 ± 2 Ma represents the real crystallization age of the mafic magmatism in the area. Not surprisingly, this value is similar to the age of the ultramafic, mafic and intermediate rocks of Gredos (310–312 Ma, Montero *et al*. 2004*a*) with which they share many petrographic and geological features. In the same way, the mean Variscan age of Toledo migmatites is 332 ± 5 Ma, exactly coincident with the peak of migmatization in the Peña Negra Complex of Gredos (Montero *et al*. 2004*b*). In both thermal domes, therefore, the ultramafic, mafic and intermediate rocks are 20–25 Ma younger than the peak of the migmatization.

It was discussed above why the Rb–Sr method applied to ultramafic, mafic and intermediate rocks leads to wrong, older, ages. To understand why field relationships between ultramafic, mafic and intermediate rocks and migmatites (Fig. 2) cause the impression that both groups are coeval, we first must bear in mind that the Variscan migmatization in Central Iberia was an extraordinarily long process: in Peña Negra it lasted about 55 Ma (Montero *et al*. 2004*b*). In Toledo our data point to similar figures: the growth of Variscan zircon began at *c*. 350 Ma, reached a maximum at *c*. 335 Ma, and then decreased gradually until 315 Ma (Fig. 5), with this time span representing a minimum estimate. The ultramafic, mafic and intermediate rocks, therefore, were intruded during the final stages of the evolution of the migmatite complexes. At that time the migmatites probably still contained a small fraction of residual melt, presumably silicic and water-rich, with a temperature close to that of the water-saturated haplogranitic system, i.e. 660–670 °C. In these conditions the migmatites were still capable of flowing plastically but could hardly grow new zircon (Watson 1996). The heat released by the crystallization of a small mafic intrusion would surely have locally augmented the melt fraction, and hence the plasticity, of the host migmatites so that, once crystallized, the dense, rigid and, in most cases, sill-like mafic body would be enclosed by a considerably more ductile and less dense host. Then, the mafic body would be easily brecciated, and the surrounding melt-enriched migmatite would easily migrate to fill the fractures, so producing a picture such as that shown in Figure 2. In this sense, therefore, migmatites and ultramafic, mafic and intermediate rocks are coeval despite the latter being *c*. 25 Ma younger than the anatexis peak. The chemical composition of the leucosome-enriched migmatite that fills the mafic breccia, characterized by notably lower than chondritic Zr/Hf, indicates that it contained an elevated percentage of a melt from which zircon had already fractionated (Bea *et al*. 2006; Lowery *et al*. 2006), and is consistent with the scenario postulated above of slowly cooling but not totally solidified migmatites.

If the heat released by the mafic intrusion was not great (i.e. if the intrusion was small) the effects of the thermal buffering at the solidus (Stuwe 1995) would keep the melt composition and temperature nearly constant regardless of the increase in the melt fraction. As a consequence, the capability for dissolving and precipitating new zircon would have remained very low. This may explain why the age and Ti-in-zircon temperatures of migmatites close to and far from the mafic intrusions are virtually the same. To put it simply, no new zircon was formed as a result of the intrusion of mafic bodies because they are too small. A rough estimation of the minimum size of ultramafic, mafic and intermediate intrusions required to cause a perceptible impact on zircon age of the aureole can be derived from the work of Montero *et al*. (2004*a*), who compared the zircon ages of the mafic bodies and host rocks in Gredos, and found a measurable rejuvenation only in the aureole of the Arenal body, which is by far the largest of the area, with an exposure of about 5 km^2.

The 20–25 Ma difference between the peak of migmatization and ultramafic, mafic and intermediate rocks revealed by zircon dating in the two already well-studied largest axial domes, Toledo and Gredos, discounts the possibility that heat transported by the mafic magmas could have been the main source for crustal metamorphism and anatexis in the Central Iberian Zone. The ultramafic, mafic and intermediate magmas appeared after the beginning of the extensional collapse of the thickened continental crust (Bea *et al*. 1999) and were emplaced into already cooling anatectic complexes, causing an insignificant thermal impact. On the other hand, mid-crustal anatexis in a thickened continental crust can be attributed to the abnormally elevated average heat production of regional migmatites in the axial thermal domes of Central Iberia (2.7–3.2 $\mu W\ m^{-3}\ s^{-1}$) if *c*. 30 Ma is allowed for thermal maturation (Bea *et al*. 2003).

Conclusions

The four studied ultramafic, mafic and intermediate bodies contain abundant magmatic zircon. The magmatic origin is proved by the excellent positive correlation between Zr/Hf and Ti-in-zircon temperatures shown by the Toledo gabbro. U—Pb ion microprobe dating yielded the following results: Toledo, 308.2 ± 1.4 Ma; Argés, 308.7 ± 4.4 Ma; Guajaraz, 311.1 ± 5.2 Ma, La Bastida, 305.6 ± 2 Ma. Excluding five zircon cores from the Guajaraz gabbro, which might be xenocrystic, the mean of 40 determinations is 306.8 ± 1.2 Ma, in round numbers 307 ± 2 Ma, which seems to be the best estimate for the age of the mafic magmatism in the Toledo Anatectic Complex.

The average Variscan age of the migmatites is 332 ± 5 Ma (i.e. about 25 Ma older). The migmatites have recorded no significant thermal impact related to the intrusion of mafic magmas. Those surrounding the ultramafic, mafic and intermediate bodies have the same age and average Ti-in-zircon temperatures as others located far from any mafic intrusion: 331 ± 5 Ma and *c*. 746 °C versus 334 ± 14 Ma and *c*. 743 °C.

The ultramafic, mafic and intermediate magmas, therefore, were emplaced into already cooling long-lived anatectic complexes. When they were intruded, the host migmatites still contained a small fraction of highly silicic and water-rich low-temperature melt with very little capacity for growing new zircon. The ultramafic, mafic and intermediate intrusions increased the melt fraction on a local scale, but the melt temperature and composition remained nearly the same. The ultramafic, mafic and intermediate bodies, especially those that are sill-like, were easily brecciated once crystallized, and the surrounding melt-enriched migmatite filled the fractures, so creating the wrong impression that mafic intrusions and the peak of regional migmatization were coeval.

This idea has been historically reinforced by Rb–Sr dating, which always yielded early Variscan, or even pre-Variscan, ages for the ultramafic, mafic and intermediate rocks. The reasons for this became evident in Toledo. A $^{87}Sr/^{86}Sr$ v. $^{87}Rb/^{86}Sr$ regression line for the four studied mafic samples yields a fictitious isochron at *c*. 584 Ma (MSWD = 1.9). Sr and Nd isotopes and, especially, the fact that migmatites plot on the same regression line reveal that it is not an isochron but a mixing line between mafic magmas and crustal materials.

Zircon dating has shown that in Central Iberia, in contrast to Ossa Morena, there is no early Variscan mafic magmatism. It also has narrowed the time gap between the appinites–vaugnerites and Early Permian camptonitic lamprophyres of the same region (Bea *et al*. 1999). Despite many differences, the latter also have an uncommon Ti-rich biopyribole mineralogy and are the only other mantle-derived rocks generated in Central Iberia during the late Variscan phase. It does not seem unreasonable, therefore, to venture a genetic link between them. The study of this possible connection may provide a new perspective for understanding the evolution of the subcontinental mantle during the evolution of intracrustal orogens.

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