Distinct sources for syntectonic Variscan granitoids: Insights from the Aguiar da Beira region, Central Portugal

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A B S T R A C T

The Variscan syntectonic granitoid plutons from the Aguiar da Beira region (central Portugal) were emplaced into metasediments of Late Proterozoic–Early Cambrian age during the last Variscan ductile tectonic event (D3), which is related to dextral and sinistral shearing. The older intrusion, with a U–Pb ID–TIMS zircon age of 2118 ± 2.0 Ma, consists of porphyritic biotite granodiorite–granite with transitional I–S type geochemical signature, relatively low ⁸⁷Sr/⁸⁶Sr322 ratios (0.7070–0.7074), εNd317 values of –3.9 to –4.6 and whole-rock and zircon δ¹⁸O values of 10.6‰ and 8.0‰, respectively. By contrast, the younger intrusion is an S-type muscovite–biotite leucogranite, emplaced at 317.0 ± 1.1 Ma, showing more radiogenic ⁸⁷Sr/⁸⁶Sr317 = 0.7104–0.7146, lower εNd317 values of –7.7 to –8.7 and higher δ¹⁸O-wr = 11.3‰ and δ¹⁸O-zr = 9.5‰. The combined isotopic and geochemical evidence supports a lower crustal origin for the biotite granodiorite–granite, involving the anatexis of lower crustal metagneous protoliths, and possible hybridization with mantle-derived magmas. A shallower origin, at mid crustal levels, from pure crustal derivation, through moderate degrees of partial melting of Proterozoic–Cambrian metasediments is proposed instead for the muscovite–biotite leucogranite.

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1. Introduction

The Iberian Variscan belt, also known as Iberian Massif, was formed in Carboniferous times after the complete subduction of the Palaeozoic Rheic Ocean, during amalgamation of Pangaea by the complex and poly-cyclic collision of Laurussia with Gondwana. Continental collision began around 365 Ma (Dallmeyer et al., 1997), with crustal thickening (D1) and associated prograde metamorphism of Barrovian type. Continued shortening is thought to have been followed by the exhumation of the thickened continental crust (D2) (e.g. Arenas and Catalán, 2003; Escuder Viruete et al., 1994; Martínez Catalán et al., 2008; Valle Aguado et al., 2005), and final deformation (D3) is considered to be related to crustal-scale strike-slip ductile shear zones that produced large-wavelength upright folds (Martínez Catalán et al., 2009). Migmatization and low-pressure/high-temperature (LP–HP) metamorphism started during D3, continued throughout D3 and culminated with the emplacement of abundant syntectonic granitoids. Immediately after D3, large volumes of late- to post-tectonic granite magmas intruded the Variscan continental crust, strongly overprinting its previous tectono-metamorphic record.

To other authors the tectono-metamorphic evolution of the Iberian Variscan belt is better explained as the result of two compressive deformation phases (D1 and D2), during which the crust reached its maximum thickness, followed by a late, post-thickening, ductile tectonic event (D3) marking the end of the continental collision in Iberia (Dias and Ribeiro, 1995; Noronha et al., 1981; Ribeiro et al., 1990). However, the P–T conditions attained during D2, inferred from the mineral assemblages related to the Sr fabrics, are considered as a strong evidence for the extensional character of this event (Valle Aguado et al., 2005) in opposition to the compressive regime interpretation. Nevertheless, irrespective of the model accepted, D3 is considered in both interpretations to be a post-thickening strike-slip phase responsible for the final structural arrangement of the Iberian Massif.

Regional constraints show that Variscan granite plutonism in Iberia post-dates syn-collisional deformation (D1 + D2), being predominantly correlated with intracontinental shearing (D3). The diversity of granite types observed in this segment of the Variscan belt represents the sum of a series of processes, ranging from partial melting of distinct source rock materials, mixing and/or mingling of crustal- and mantle-derived magmas, fractional crystallization, crustal contamination and more complex models involving concurrent assimilation and fractional
crystallization (AFC) (e.g. Azevedo et al., 2005; Beetsma, 1995; Castro et al., 1999; Dias et al., 2002; Fernández-Suárez et al., 2011; Neiva and Gomes, 2001; Villaseca et al., 1999). In order to assess the factors controlling the evolution of each specific granitoid suite, it is crucial to understand the information stored in rock structures and to investigate the main vectors of chemical and isotopic change involved in their petrogenesis.

In this study petrographic, geochemical, Sr–Nd–Os isotopic data and high-precision U–Pb ID–TIMS zircon and monazite ages are used to discuss the origin of two particular syntectonic intrusive suites from the central sector of the Central-Iberian Zone. The aim was to contribute to a better understanding of the origin, genesis and evolution of these magmas during the Variscan orogeny and constrain the nature of their potential sources, as well as the timing and duration of magmatic and deformation events in the studied area.

2. Geological background

Based on tectonostratigraphic criteria, the Iberian Variscan belt is divided into, from north to south, the Cantabrian, West Asturian-Leonese, Galicia-Trás-os-Montes, Central-Iberian, Ossa-Morena and South Portuguese zones (Fig. 1). In its internal domains, such as in the Central-Iberian Zone, where the study area is located, the stratigraphic record includes a Proterozoic Gondwanan basement of Cadomian (West African) affinity overlain by Early Palaeozoic, mostly siliciclastic sequences, deposited on the passive margin of northern Gondwana (e.g. Fernández-Suárez et al., 2002, 2011; Gutiérrez-Alonso et al., 2011; Martínez Catalán et al., 2007, 2009; Murphy et al., 2008; Robardet, 2002, 2003).

The Aguiar da Beira region is located within the axial zone of the Iberian Massif (Fig. 1), in central northern Portugal. The region is mainly composed of Variscan granitoids emplaced into Neoproterozoic–Early Palaeozoic sedimentary sequences, which were variably affected by regional metamorphism and deformation during the Variscan orogeny (Fig. 2). The oldest rocks exposed in this area belong to the so called “Schist–Greywacke Complex” (SGC), a thick turbidite-like formation of Neoproterozoic–Early Cambrian age consisting of metapelites and metagreywackes interlayered with relatively thin calc-silicate and metaconglomerate horizons (Oliveira et al., 1992; Rodríguez Alonso et al., 2004; Sousa, 1984). The SGC is unconformably overlain by an Ordovician clastic succession of stable marine platform sediments, spatially confined to the core of the NW–SE trending Porto–Sátão syncline that crosses the southwestern corner of the studied area (Fig. 1). A narrow deposit of Late Carboniferous molasse occurs in contact with the Ordovician rocks (Fig. 1).

The overall structure of the region can be ascribed to the combined effects of two main Variscan deformation events (D1 + D3). The first, D1, affects all the pre-Carboniferous sequences and resulted in the generation of NW–SE striking subvertical folds with a penetrative axial plane schistosity (S1), whilst D3 is related to several crustal-scale transcurrent shear zones that have accommodated part of the shortening related to the final stages of the continental collision (Valle Aguado et al., 2005). One of these major shear zones, the NW–SE striking sinistral Douro–Beira shear zone, coincident with the Porto–Sátão syncline (Fig. 1), led to the development of a subvertical mylonitic S3 fabric recorded in both the pre-Carboniferous and the Carboniferous metasedimentary units (Valle Aguado et al., 2005).

Regional metamorphism in this area corresponds to a low-grade metamorphic epizonal domain in which metamorphic recrystallization...
occurred under greenschist facies conditions not exceeding the biotite zone. The emplacement of voluminous amounts of granitic magmas during and slightly after D3 generated low-P–high-T contact aureoles.

Variscan intrusions, spanning the entire spectrum of granitoid types found in the NW Iberian Variscan belt, dominate the Aguiar da Beira region, covering about 75% of the total exposure (Fig. 2). The intrusive sequence consists of an earlier syntectonic (syn-D3) association, which is described in detail below, and a volumetrically dominant late- to post-tectonic association formed by biotite monzogranites, biotite–muscovite granites and minor bodies of muscovite leucogranites.

3. Syntectonic association — field occurrence

The syn-D3 association is composed of two intrusive suites with specific compositional, petrographic and geochemical characteristics: (a) an early biotite granodiorite–granite and (b) a slightly younger, highly peraluminous leucogranite. Zircon and/or monazite geochronology yield
U–Pb ages of 321.8 ± 2.0 Ma for the biotite granodiorite–granite and 317.0 ± 1.1 Ma for the highly peraluminous muscovite–biotite leucogranite (this study).

3.1. Sym-D3 biotite granodiorite–granite

The medium- to coarse-grained porphyritic biotite granodiorite–granite forms an irregularly shaped intrusion of about 2 × 4 km and two smaller exposures in the northern part of the area (Fig. 2). All these occurrences are surrounded by the sym-D3 muscovite–biotite leucogranite and seem to represent remnants of an earlier intrusive complex. The biotite granodiorite–granite (Fig. 3A) locally shows a N130° trending magmatic flow foliation, due to the alignment of K-feldspar megacrysts. The NW–SE foliation is concordant with D3 regional structures and might be an evidence for syntectonic emplacement. Rounded and ellipsoidal mafic microgranular enclaves with sizes ranging from a few centimeters to several decimeters are common. As a result of an extensive interaction with the host granitoids, the original mineralogical, textural and geochemical characteristics of the mafic microgranular enclaves are no longer preserved. Some light colored enclaves of fine-grained granite with cognate appearance have also been observed.

3.2. Sym-D3 highly peraluminous muscovite–biotite leucogranite

Unlike the biotite granodiorite–granite, the peraluminous leucogranite crops out as a large NW–SE to E–W trending intrusion with long and short axes of about 21 and 8 km, respectively (Fig. 2). Field relationships indicate emplacement during D3, after the biotite granodiorite–granite previously described. The contacts between the two granitoids are sharp (Fig. 3B). A younger, circular, late- to post-tectonic granitoid pluton separates the leucogranite intrusion into two portions (Fig. 2). In the western part, the massif consists of a heterogeneously deformed medium-grained muscovite–biotite granite, showing a gneissic foliation (Fig. 3C, D) defined by the sub-parallel alignment of biotite and muscovite. This foliation is steeply inclined and runs parallel to the Douro–Beira D3, transcurrent shear zone suggesting that the emplacement of this granite was controlled by shearing. In the eastern part, the granite is finer-grained and shows no evidence of solid state deformation. The absence of evidence for deformation in the whole eastern part of the muscovite–biotite leucogranite and in the biotite granodiorite–granite suggests that the Douro–Beira shear zone did not affect this sector. Mafic microgranular igneous enclaves are absent throughout the entire massif. By contrast, centimeter-sized xenoliths of metamorphic country-rocks are abundant.

4. Analytical procedures

Representative rock samples from the two intrusions were collected, crushed and milled to fine powders in an agate mortar for whole-rock geochemistry and Rb–Sr, Sm–Nd and 618O isotope analyses. Major- and trace-element data (including REE) were obtained by ICP-AES and ICP-MS, respectively, at the Activation Laboratories, Ltd. (Canada). Precision is better than 4% for major elements and 10% for trace elements. The detection limits are 0.001 wt.% for MnO and TiO2, 0.01 wt.% for the remaining oxides, and lower than 1 ppm for most trace elements, except V and Pb (5 ppm), Cr and Ni (20 ppm), Cu (10 ppm) and Zn (30 ppm). FeO was determined by titration with a standardized potassium permanganate solution, at the Earth Sciences Department of the University of Coimbra, Portugal (precision ±1%).

Whole-rock powders for strontium and neodymium isotope analyses were dissolved in a HF–HNO3 mixture in pressurized Parr vessels for 3 days, then dried and taken up in HCl for chemical separation. Cation exchange resin (Biorad AG50W) in quartz glass columns was used to separate Sr and the REE fraction. Neodymium was separated from the other REE in quartz glass columns filled with LN-resin from Eichrom. 87Sr/86Sr and 143Nd/144Nd isotope ratios were measured on a VG Sector 54 mass spectrometer operating in dynamic mode, at the Laboratory of Isotope Geology of the University of Aveiro, Portugal. By maintaining a 1.5–2 V 88Sr beam for 50–100 cycles, an internal precision of 20 ppm (1-sigma) on 87Sr/86Sr was consistently achieved. Fractionation was corrected with an exponential law relative to 86Sr/88Sr = 0.1194. During the measuring campaign, the 87Sr/86Sr ratio of SRM-987 was 0.710249(6) (6 ppm, 1-sigma, n = 61; literature value = 0.710248). Neodymium isotope ratios were measured as metal ions in a Ta–Re–Ta triple filament assembly in a three sequence dynamic routine using seven Faraday cups. By maintaining a 1 V 146Nd beam for 50–100 cycles, an internal precision of 20 ppm (1-sigma) was obtained. The 143Nd/144Nd ratios were corrected for mass fractionation using the 146Nd/144Nd ratio of 0.7219. Repeated analysis of the JNd-1 standard gave an average value of 0.5121037(41) for 16 measurements. Analytical blanks for Sr and Nd are lower than 250 pg. The Rb, Sr, Sm and Nd concentrations used for calculation of 87Rb/86Sr and 143Sm/144Nd isotope ratios were obtained either by isotopic dilution or by ICP-MS at the Activation Laboratories, Canada. The initial s Nd values were calculated using 143Sm/144Nd = 0.1967 and 147Nd/144Nd = 0.512638 for the Chondritic Uniform Reservoir (CHUR) following Jacobsen and Wasserburg (1980). A linear model with parameters 143Sm/144Nd = 0.2136 and 147Nd/144Nd = 0.513151 was used for the Depleted Mantle reservoir (DM) according to Goldstein and Jacobsen (1988).

Oxygen-isotope analyses of whole-rock samples were performed by gas mass spectrometry according to the methodology of Clayton and Mayeda (1963), at the Department of Earth Sciences of the University of Western Ontario, Canada. Repeated analysis of biotite NBS-30 and quartz NBS-28 standards gave values of 4.99% and 9.72‰, respectively.

Mineral compositions were determined with a Cameca Camebax SX-100 electron microprobe at the Scientific-Technological Centre of the University of Oviedo, Spain, operating in wavelength dispersive mode with an acceleration voltage of 15 kV and a beam current of 15 nA. Analytical precision was better than 2%.

Zircons and monazites, used for isotope dilution TIMS U–Pb analyses, were separated from ~420 μm size fractions by standard mineral separation techniques using a Frantz® magnetic separator and heavy liquids. The crystals were subsequently handpicked under a binocular microscope and mechanically abraded to remove external disturbed domains (Davies et al., 1982; Krogh, 1982). On the basis of optical microscopy the best quality zircon grains (free of visible cracks, overgrowths or inclusions) were cleaned with HNO3, acetic and ultra-pure H2O. A Fig. 4. Aluminium and magnesium compositions of biotite (atoms per 11 O) from the Aguiar da Beira syntectonic granitoids. Fields from different magmatic series taken from Nachit et al. (1985).
mixed $^{202}\text{Pb} - ^{205}\text{Pb}$ U tracer solution was added prior to dissolution. Chemical separation and purification of uranium and lead were performed in small Teflon® columns, according to the procedure described by Krogh (1973) and modified by Corfu (2004). Uranium and lead isotopic compositions were measured on a Finnigan MAT 262 multicollector mass spectrometer at the Department of Geosciences of the University of Oslo, Norway. The samples, as well as a mixed NBS 982Pb + US00 standard, were loaded on single rhenium filaments using a silica-gel activator and H$_3$PO$_4$. Total procedure blanks were measured for each element in the standard relative to its known concentration, allows the determination of the calibration factors that were applied to all sample data.

5. Petrography and mineral chemistry

5.1. Syn-D$_3$ biotite granodiorite–granite

The syn-D$_3$ medium- to coarse-grained porphyritic biotite granodiorite–granite contains quartz (22–25%), plagioclase (27–44%), microcline (12–36%), biotite (12–20%), apatite, zircon, monazite and ilmenite. Quartz is characteristically interstitial and shows undulational extinction. Plagioclase (An$_{39}$–An$_{59}$) occurs in the matrix of the rock as weakly zoned subhedral crystals containing small inclusions of biotite, quartz and apatite. Strontium contents in plagioclase are high (608–1168 ppm; see Electronic Appendix, Table 2A) and define irregular zoning patterns, suggesting disequilibrium crystalization conditions. K-feldspar is slightly perthitic microcline and constitutes large phenocrysts, up to 7 × 3 cm, with compositions ranging from Or$_{50}$ to Or$_{90}$, enclosing quartz, plagioclase and biotite. Microcline is also present in the groundmass as small interstitial crystals (Or$_{80}$–Or$_{95}$). Mafic microgranular equigranular granite composed of quartz (32–37%), perthitic...
microcline (26–36%), plagioclase (16–23%), muscovite (4–11%), biotite (0–6%), and minor amounts of tourmaline, zircon, apatite, monazite and ilmenite (very scarce). In the more deformed samples, a gneissic foliation defined by the alignment of muscovite and biotite flakes can be observed. Quartz is mostly anhedral and shows undulose extinction. K-feldspar is perthitic microcline (Or87–Or92) and occurs as anhedral interstitial crystals, frequently enclosing small inclusions of plagioclase, biotite and muscovite. In the samples affected by regional deformation, the primary K-feldspar crystals are often recrystallized into a mosaic of smaller grains of K-feldspar, albite plagioclase and quartz (Fig. 5). Plagioclase (An0–An2) is essentially subhedral, weakly zoned and may show bent twin lamellae and recrystallization borders in the more deformed samples. The Sr and Ba contents in plagioclase are uniformly low (Sr = 4–36 ppm; Ba = 0–2 ppm), suggesting near equilibrium crystallization conditions. Biotite has a foxy red-brown color and is often intergrown with muscovite. In intensively deformed domains, the micas are concentrated into long bands and the larger crystals may exhibit sigmoidal shapes (Fig. 5). Biotite compositions (Al0.87–Al0.95) and ilmenite (very scarce). In the more deformed samples, a gneissic foliation de

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granodiorite–granite and the muscovite–biotite leucogranite suggests that the emplacement of the latter may have triggered monazite recrystallization.

6.2. Muscovite–biotite leucogranite (sample GN-3)

Sample GN-3 from the muscovite–biotite leucogranite yielded a much lower amount of zircon crystals than the sample from the biotite granodiorite–granite. Zircons extracted from this sample are transparent, pale yellow to colorless and practically devoid of inclusions. Small, rounded euhedral crystals and short prisms are dominant, but some long prismatic crystals were also identified. Given the scarcity of long prismatic zircons, the fractions selected for analysis include tips and long prismatic grains (two fractions) and fragments of a large abraded crystal (~350 × 630 μm) (two fractions). The four zircon fractions yielded a spread of concordant ages between 314 Ma and 317 Ma (Fig. 6B), reflecting partial Pb loss. The oldest concordant zircon age (317.2 ± 1.1 Ma) gives the best approximation to the crystallization age of the muscovite leucogranite (Fig. 6B), reflecting partial Pb loss. The oldest concordant zircon age (317.2 ± 1.1 Ma) gives the best approximation to the crystallization age of the intrusion. Regression passing through the four zircon fractions results in an upper intercept age of 315.0 ± 6.1 Ma (Fig. 6B) providing further support for the age obtained.

The only monazite fraction analyzed from this granite consists of small rounded and ovoid crystals and yielded an almost concordant U–Pb age of 300.8 ± 0.7 Ma, which is significantly younger than the age of coexisting zircon. As mentioned before, the U–Pb monazite ages can be partially reset during late thermal events. In the Aguiar da Beira region, the effects of a younger overprint associated with the emplacement of large volumes of late-to post-tectonic granitoids, with the biotite leucogranite in the Aguiar da Beira region, the effects of a younger overprint associated with the emplacement of large volumes of late-to post-tectonic granitoids, with the biotite leucogranite yielded a spread of concordant ages between 314 Ma and 317 Ma (Fig. 6B), reflecting partial Pb loss. The oldest concordant zircon age (317.2 ± 1.1 Ma) gives the best approximation to the crystallization age of the intrusion. Regression passing through the four zircon fractions results in an upper intercept age of 315.0 ± 6.1 Ma (Fig. 6B) providing further support for the age obtained.

7. Whole-rock geochemistry

Representative whole-rock chemical analyses for the two syntectonic granitoid intrusions from the Aguiar da Beira region are given in Table 2. The biotite granodiorite–granite is slightly peraluminous (A/CNK = 1.01–1.02) with very homogeneous SiO₂ contents (65.59–67.61 wt.%). Its composition plots essentially within the field of the monzogranites in the R₁–R₂ diagram (Fig. 7), defining an “alumino-cafemic” association of calc-alkaline type in the A–B diagram (Fig. 8A), and plots in the mafic granitic field in the Fe number vs. % SiO₂ diagram (Fig. 8B). Amongst the trace elements, Sr abundances are as great as 735 to 772 ppm, Ba reaches values of 1397–1691 ppm, Zr = 202–223 ppm (Fig. 9), and XREE ranges from 474 to 507 ppm. Chondrite normalized REE patterns for the biotite granodiorite–granite samples are characterized by high LREE/HREE ratios (LάN/LuN = 55–58) and small Eu negative anomalies (Eu/Eu* = 0.71–0.80) (Fig. 10). In the multi-element variation diagrams normalized to primordial mantle there are strong negative anomalies for Nb, Ta, Sr, P and Ti and a positive anomaly for Th (Fig. 11A). The significant Ba and Sr enrichment, A/CNK values < 1.1, highly fractionated REE patterns, and marked LILE/HFSE enrichment are typical of high-K calc-alkaline I-type magmas (Roberts and Clemens, 1993; Wilson, 1989).

By contrast, the Aguiar da Beira muscovite–biotite leucogranite is strongly peraluminous (A/CNK = 1.18 and 1.36) with high contents of silica (71.91–73.90 wt.%), P₂O₅ (0.39–0.48 wt.%) and high Rb/Sr ratios (5–17), coupled with low CaO (0.29–0.63 wt.%), MgO (0.14–0.41 wt.%), Ba (81–295 ppm), Sr (31–77 ppm) and ΣREE (41–158 ppm) abundances. These rocks are classified as syenogranites in the R₁–R₂ diagram (Fig. 7) and plot in the leucogranitoid field of the peraluminous domain in the A–B classification diagram (Fig. 8A). The diagrams by Frost et al. (2001) also evidence the lower contents of MgO and CaO that characterize this leucogranite (Fig. 8B₁ and B₂). Despite their relatively narrow compositional range, the samples of this intrusion define coherent evolutionary curvilinear trends in most major- and trace-element variation diagrams, characterized by an increase in SiO₂, F, Rb, Cs and Ta and a decrease in TiO₂, MgO, CaO, K₂O, V, Zr, Sr, Ba, Hf and Th with decreasing total FeO (Fig. 9, Table 2), suggesting fractionation of feldspar, biotite and accessory minerals. Their REE chondrite normalized patterns are sub-parallel, characterized by an increase in Eu negative anomalies in Nb, Sr and Ti and positive anomalies for Nb and Pb (Fig. 11B). Based on its major and trace element compositions, the two-mica leucogranite bears a strong resemblance to other S-type granitoids from the Iberian Variscan belt.

The temperature at which zircon starts to crystallize depends on the amount of zirconium present in the melt and this relation can be used as a geothermometer (Watson and Harrison, 1983). Zircon saturation temperatures (T₂₀) calculated from bulk rock compositions are 796 °C for the biotite granodiorite–granite and 731 °C for the muscovite–biotite leucogranite. Due to possible occurrence of zircons with inheritance in both intrusions, particularly in the muscovite–biotite leucogranite, the obtained T₂₀ are probably overestimated (Miller et al., 2003). However,
the higher T estimates for the biotite granodiorite–granite are broadly consistent with the petrogenetic models discussed in a later section.

8. Sr–Nd and δ¹⁸O (whole-rock and zircon) isotopic data

The Rb–Sr, Sm–Nd and δ¹⁸O isotopic data for the samples of the Aguiar da Beira syntectonic intrusions are listed in Table 3. As shown in the ⁸⁷Sr/⁸⁶Sr–εNd diagram (Fig. 12), the biotite granodiorite–granite has relatively low ⁸⁷Sr/⁸⁶Sr (0.7070–0.7074) and εNd varying between −3.9 and −4.6, while the muscovite–biotite leucogranite shows distinctly more radiogenic Sr isotopic signatures (⁸⁷Sr/⁸⁶Sr = 0.7104–0.7146) and lower εNd values (−7.7 to −8.7). One sample from the muscovite–biotite leucogranite (GN-6) has a much lower ⁸⁷Sr/⁸⁶Sr of 0.70495, probably reflecting a disturbance in the Rb–Sr isotopic system. Its high Rb/Sr ratio of 57 (vs. 12–31 in the others) suggests a recent introduction of Rb or Sr loss. The Nd model ages (TDM) (DePaolo, 1981, 1988) range between 1.0 and 1.1 Ga in the biotite granodiorite–granite and between 1.6 and 2.1 Ga in the muscovite–biotite leucogranite.

Whole-rock oxygen isotopic compositions are high and match the trend observed in Nd and Sr, with higher values of δ¹⁸O-wr = 11.33‰ and δ¹⁸O-zr = 9.5 ± 0.2‰ in the leucogranites than in the granodiorite which is characterized by lower δ¹⁸O-wr = 10.6‰ and δ¹⁸O-zr = 8.0 ± 0.2‰.

9. Discussion

9.1. Time constraints

The U–Pb zircon crystallization ages of the Aguiar da Beira syntectonic granitoids provide time constraints for the duration of magmatic and tectonic events in the area and give new insights...
into the tectonothermal evolution of this sector of the Iberian Variscan belt. The Variscan plutonic activity started with the intrusion of the biotite granodiorite, at c. 322 Ma, and was shortly followed by the emplacement of the muscovite leucogranite, at c. 317 Ma. As both granitoids were intruded synkinematically during D2, these ages reveal that the last Variscan ductile deformation event spanned a minimum of about 5 Ma. They also yield an upper age limit for the movement of the sinistral Douro–Beira shear zone which has controlled the emplacement of the muscovite–biotite leucogranite (c. 317 Ma). The geochronological data obtained in this study partially overlap the U–Pb ages determined for other syntectonic granitoids from the NW Iberian Variscan belt (Dias et al., 1998, 2002; Fernández-Suárez et al., 2000; Teixeira et al., 2012; Valle Aguado et al., 2005) and are consistent with the structural and metamorphic evolution proposed for this sector of the CIZ (e.g. Valle Aguado et al., 2005).

9.2. Magma sources

The compositional and isotopic differences between the syntectonic biotite granodiorite–granite and the muscovite–biotite leucogranite from Aguiar da Beira strongly suggest that these magmas were either produced by partial melting of distinct source rocks, or that different petrogenetic processes were involved in their origin (partial melting and/or magma mixing).

9.2.1. Biotite granodiorite–granite

The major and trace element characteristics of the biotite granodiorite–granite are typical of metaluminous to slightly peraluminous high-K calc-alkaline I-type magmas with A/CAK values < 1.1, significant Ba and Sr enrichment, high K/Rb ratios, fractionated REE patterns, marked LILE/HFSE enrichment, and negative anomalies of Nb–Ta, Ti, Y and HREE. This favors an origin by partial melting of feldspar-rich metaigneous crustal protoliths (high Ba and Sr), leaving a residue depleted in feldspar and biotite and enriched in mineral phases capable of retaining Nb, Ta, Ti, Y and HREE. This favors an origin by partial melting of feldspar-rich metaigneous crustal protoliths (high Ba and Sr), leaving a residue depleted in feldspar and biotite and enriched in mineral phases capable of retaining Nb, Ta, Ti, Y and HREE.

The lower crustal felsic xenoliths plotted in Fig. 12 have been interpreted as the residues left after the extraction of granitic melts from lower crustal sources (Villaseca et al., 1999), and they suggest that the composition of the lower crust underneath Iberia, although not accurately known, might be more felsic and LILE and REE enriched than model compositions proposed for other areas (McLennan and Taylor, 1996; Villaseca et al., 1999; Wedepohl, 1995). The hypothetical derivation of the biotite granodiorite–granite from a protolith similar to that which produced the residual felsic xenoliths, is not only supported by their matching Sr and Nd isotopic signatures (Fig. 12), but also by the trace element composition (Fig. 11A). The compositions of a lower crustal xenolith and of an I-type granite from the Spanish Central System (Fig. 12) supporting an origin through partial melting of similar sources.

Whole-rock δ18O (δ18O wr) and zircon δ18O (δ18O-zr) values of 10.6‰ and 8.0‰, respectively, are common in felsic meta-igneous granulites (Hoefs, 2004).

The Lower crustal felsic xenoliths plotted in Fig. 12 have been interpreted as the residues left after the extraction of granitic melts from lower crustal sources (Villaseca et al., 1999), and they suggest that the composition of the lower crust underneath Iberia, although not accurately known, might be more felsic and LILE and REE enriched than model compositions proposed for other areas (McLennan and Taylor, 1996; Villaseca et al., 1999; Wedepohl, 1995).
Much of the chemical and isotope composition of this intrusion can alternatively be explained by simple mixing between metaluminous basic magmas of mantle origin and crust-derived peraluminous partial melts and/or AFC-type processes. The widespread occurrence of mafic microgranular enclaves would be consistent with hybridization processes between coeval mafic and felsic magmas (e.g. Barbarin, 2005; Barbarin and Didier, 1992; Vernon, 1984, 1990, 1991). In the Iberian Variscan belt, mafic igneous rocks and I-type or transitional S-I type granitoids with low degrees of crustal contamination yield TDM values ranging between ca. 0.95 and 1.2 Ga (Beetsma, 1995; Castro et al., 2003; Dias et al., 1998; Fernández-Suárez et al., 2011). These low TDM ages may imply some input of juvenile mantle magmas into the crust and, in this case, the TDM ages would correspond to mixed model ages between the juvenile source and an older crustal contaminant (Beetsma, 1995; Dias et al., 1998). However, these values can also result from a maetagegne basement extracted from a subcontinental enriched lithospheric mantle between ca. 0.9 and 1.1 Ga (Fernández-Suárez et al., 2011; Murphy et al., 2008).

Fig. 9. Selected major (wt.%) and trace (ppm) element variation diagrams for the Aguiar da Beira syntectonic granitoids. Symbols as in Fig. 4.

Fig. 10. Chondrite-normalized REE patterns for the samples of the two syntectonic granitoids from Aguiar da Beira. Normalizing values from Haskin et al. (1968).

Fig. 11. Trace-element composition of the Aguiar da Beira syntectonic granitoids (normalizing values from Sun and McDonough, 1989). A — biotite granodiorite–granite samples, Spanish Central System I-type granite composition and lower crustal maetagegne granulitic xenolith sample (Villaseca et al., 1999); B — muscovite–biotite leucogranite samples and metagreywacke sample from the Schist–Greywacke Complex (Teixeira, 2008).
Thus, the possible lineage of the biotite granodiorite–granite suite remains controversial, as the available data do not allow effective discrimination between the two hypotheses. At the present level of exposure, there is no field evidence either for feldspar-rich protoliths from the deeper crust, or for the occurrence of mafic rocks, and the nature of the underlying lower crust is also unknown.

9.2.2. Muscovite–biotite leucogranite

Both the elemental and isotopic compositions of the muscovite–biotite leucogranite (A/CNK values > 1.1; high SiO₂ contents, low CaO, MgO, Ba, Sr, 87Sr/86Sr > 0.710, Ndᵣ = −7.7 to −8.7, δ¹⁸O-WR = 11.33‰, δ¹⁸O-Zr = 9.5 ± 0.2‰) support a major involvement of metasedimentary sources in the genesis of this magma. Trace element primordial mantle-normalized patterns for the samples from this granite broadly overlap with those of low-grade metapelites and metagreywackes of the SGC, but are comparatively depleted in Sr, HREE and HFSE and enriched in Rb, LREE and P (Fig. 11). The observed HREE and HFSE depletions are consistent with the presence of biotite and minor amounts of garnet in the residual mineral assemblage (Table 4), which would retain the HFSE and the HREE, respectively. The more elevated Rb, LREE and P contents suggest that feldspar + muscovite + biotite + apatite + monazite were consumed during partial melting reactions. The relatively flat HREE patterns of this granite (Gdₙ₀/Vₙₐ = 3.4 to 8.6; Fig. 10) indicate a minor role for garnet in the source and, therefore, partial melting at mid crustal levels. This feature is common to many strongly peraluminous granites from the Iberian Variscan belt (e.g. Azevedo et al., 2005; Villaseca et al., 1998, 2008).

Given the high mobility of Rb and Sr during post-magmatic processes, the Sr isotope ratios can be ambiguous. Correlations with potential protoliths must therefore rely mainly on the Nd isotope compositions. The basement in the Aguiar da Beira region is largely composed of Neoproterozoic–Early Cambrian metasediments of the SGC, which includes pelitic sediments (mica-rich, feldspar-poor) and metagreywackes (more enriched in feldspar and quartz). The εNd values for SGC materials, calculated for an age of 320 Ma, range between −8.6 and −11.1 (Teixeira, 2008; Teixeira et al., 2012), providing suitable source lithologies for the leucogranite suite (Fig. 12). The participation of this metasedimentary crust in the genesis of the muscovite–biotite leucogranite can account for both the Nd compositions (εNd317 = −7.7 to −8.7) and the relatively high oxygen isotopic compositions (δ¹⁸O-WR = 11.33‰, δ¹⁸O-Zr = 9.5 ± 0.2‰).

Fractional crystallization of a mineral association composed by K-feldspar + plagioclase + quartz + biotite + apatite + monazite + zircon + ilmenite is likely to have played an important role in the evolution of this magma, as suggested by the differentiation trends in variation diagrams, REE patterns and mass balance calculations (discussed below).

10. Petrogenetic modeling of the syntectonic muscovite–biotite leucogranite

Simple mass balance calculations for major elements were used to test if the least evolved sample of the muscovite–biotite leucogranite (GN-5) could have been produced by partial melting of a metagreywacke (R14) from the SGC (Teixeira, 2008) composed of quartz, plagioclase, K-feldspar, biotite, muscovite, sillimanite and ilmenite. Attempts to use metapelites from the same metasedimentary sequence as potential protoliths have failed because the composition of the samples did not fit the model and are not presented here.

The model was calculated by adjusting the relative proportions of minerals entering the melt to the composition of the expected magma (GN-5). The quality of the model was assessed through the sum of the squares of the differences between the compositions of the calculated melt and the expected liquid for each element (ΣK²). For ΣK² values less than 1.2, the model was considered acceptable (Wyers and Barton, 1986). The results obtained (Table 4) show that it is possible to replicate
the major element composition of sample GN-5 by the anatexis of SGC metagreywacke, leaving a residuum composed by biotite (42%), quartz (21%), sillimanite (20%), plagioclase (6%), garnet (3%) and K-feldspar (1%), for a degree of melting (F) of 36% (ΣR² = 0.23).

Melting reactions are likely to be similar to the following (Vielzeuf and Montel, 1994; Vielzeuf and Schmidt, 2001): Ms + Ab + Qtz → Bt + Al + Kfs + melt, and Bt + Al + Pl + Qtz → Grt + Kfs + Melt, where mineral abbreviations are as recommended by Kretz (1983).

Comparison with the major element compositions of glasses obtained from partial melting experiments (e.g. Castro et al., 1999; Montel and Vielzeuf, 1997; Patiño-Douce and Johnston, 1991; Thompson, 1982) reveals, however, that the least evolved samples of muscovite–biotite leucogranite plot close to the boundary between the fields of the melts produced from metagreywacke and muscovite-rich sources in the Al₂O₃ / (MgO + FeOt) vs CaO / (MgO + FeOt) diagram (Fig. 13A) and within the domains defined by the same melt types in the Na₂O + K₂O + Fe₂O₃ vs CaO / (MgO + Fe₂O₃) diagram (Fig. 13B). This suggests that both sources could have been involved in the generation of the muscovite–biotite

Fig. 12. $^{87}$Sr/$^{86}$Sr and εNd initial isotopic compositions of samples from the Aguiar da Beira syntectonic granitoids. The plotted fields are recalculated to 320 Ma and were taken from: southern sector of Schist–Greywacke Complex and Ordovician and Silurian metasediments (Beetsma, 1995), northern sector of Schist–Greywacke Complex (Teixeira, 2008), lower crustal pelitic and metagranitic xenoliths (Villaescu et al., 1999).


<table>
<thead>
<tr>
<th>Whole rock wt.%</th>
<th>Protolith composition — metagreywacke (RI4 — Teixeira, 2008)</th>
<th>Produced melt</th>
<th>Residue</th>
<th>Calculated composition</th>
</tr>
</thead>
<tbody>
<tr>
<td>Minerals</td>
<td>Sample GN5 Calculated composition</td>
<td>Calculated</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bt</td>
<td>34.31 63.39 36.89 100.00 0.00 48.29 72.72 72.81 51.32</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>IIm</td>
<td>21.12 19.31 42.00 0.01 0.00 0.00 0.00 0.00 10.70</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SiO₂</td>
<td>36.39 19.03 43.00 0.00 0.00 0.00 0.00 0.00 6.69</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>TiO₂</td>
<td>3.16 8.05 0.00 0.00 0.00 0.00 0.00 0.00 1.26</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>22.06 20.05 0.00 0.00 0.00 10.70 0.00 0.00 10.00</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MnO</td>
<td>0.46 0.04 0.00 0.00 0.00 0.00 0.00 0.00 3.15</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MgO</td>
<td>0.00 0.00 0.00 0.00 0.00 0.00 0.00 0.00 3.87</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.00 0.00 0.00 0.00 0.00 0.00 0.00 0.00 5.88</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>K₂O</td>
<td>0.00 0.00 0.00 0.00 0.00 0.00 0.00 0.00 0.49</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.00 0.00 0.00 0.00 0.00 0.00 0.00 0.00 0.38</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total</td>
<td>100.00 100.00 100.00 100.00 100.00 98.20 100.00 100.00 96.98</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 4 Muscovite–biotite leucogranite partial melting major element modeling results.

Bt — biotite; Qtz — quartz; Ilm — sillimanite; Pl — plagioclase; Cd — cordierite; Grt — garnet; Fk — K-feldspar; Ap — apatite; Ilm — ilmenite.
leucogranite. In fact, the dehydration of metapelites, that are also part of the SGC, and mica-enriched, must have played an important role in providing some of the water required to melt the metagreywackes.

The suggested and modeled derivation from mid-crustal metasedimentary protoliths for the muscovite–biotite leucogranite from Aguiar da Beira is in agreement with other studies of similar syntectonic intrusions (e.g. Azevedo et al., 2005; Beetsma, 1995; Neiva and Gomes, 2001; Teixeira, 2008; Valle Agudo et al., 2005; Villaseca et al., 2008). Furthermore, it evidences that the petrogenetic processes and/or the protoliths involved in magma genesis changed from the production of syntectonic to the post-tectonic Iberian Variscan leucogranitic magmas, as for the later leucogranites, given their less radiogenic (Sr–Nd–O) signatures, other origins have been proposed, such as a derivation from metagneous lower crustal sources (Villaseca et al., 2012) or the addition of mantle magmas to crustal-derived melts (e.g. Castro et al., 1999; Dias et al., 2002; Silva et al., 2000).

The fractional crystallization hypothesis has also been tested using the general least-squares mixing equation devised by Le Maitre (1981). Examples of the output are given in Table 5. In the calculation of this model, one of the most primitive samples was chosen to represent the composition of the presumed parental magma and the sample with the highest SiO₂ content (GN-6) to simulate the fractionated melt. Major-element based least-squares modeling reveals that the geochemical variation observed within the muscovite–biotite leucogranite can be accounted for by up to 34% fractionation of a mineral assemblage consisting of K-feldspar (43%), plagioclase (23%), quartz (21%), biotite (12%), apatite (2%) and ilmenite (1%). Statistical fit is excellent ($\Sigma R^2 = 0.06$).

On the basis of the major element model results and partition coefficients from the literature (Table 5), further calculations were performed using the Rayleigh's fractionation equation (Arth, 1976) to predict the behavior of some trace elements (Ba, Rb, Sr) during fractional crystallization. Cumulate compositions and $F$ values estimated from major element modeling were used to derive the Ba–Rb–Sr models and the results demonstrate that crystal fractionation has probably played a significant role in the evolution of the Aguiar da Beira peraluminous leucogranite suite.

### Table 5

<table>
<thead>
<tr>
<th>Parental magma composition</th>
<th>Residual melt</th>
<th>Cumulate</th>
</tr>
</thead>
<tbody>
<tr>
<td>Whole rock wt.%</td>
<td>Minerals</td>
<td>GN-6 Calculated composition</td>
</tr>
<tr>
<td>SiO₂</td>
<td>72.82</td>
<td>35.15</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.26</td>
<td>2.60</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>15.19</td>
<td>18.72</td>
</tr>
<tr>
<td>FeO</td>
<td>1.26</td>
<td>22.79</td>
</tr>
<tr>
<td>MnO</td>
<td>0.02</td>
<td>0.50</td>
</tr>
<tr>
<td>MgO</td>
<td>0.31</td>
<td>4.55</td>
</tr>
<tr>
<td>CaO</td>
<td>0.63</td>
<td>0.00</td>
</tr>
<tr>
<td>Na₂O</td>
<td>3.31</td>
<td>0.09</td>
</tr>
<tr>
<td>K₂O</td>
<td>5.78</td>
<td>9.23</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.41</td>
<td>0.00</td>
</tr>
<tr>
<td>Total</td>
<td>100.00</td>
<td>93.63</td>
</tr>
<tr>
<td>% crystallization</td>
<td>34%</td>
<td>Cumulate mineral composition</td>
</tr>
<tr>
<td>$\Sigma R^2$</td>
<td>0.055</td>
<td>$0.43F_{K} + 0.23Pl + 0.21Qtz + 0.12Bt + 0.02Ap + 0.01Ilm$</td>
</tr>
</tbody>
</table>

### Trace element modeling

Rayleigh's equation — $CL / CL_0 = F^{n-1}$

<table>
<thead>
<tr>
<th>$CL_0$</th>
<th>$F = 0.66$</th>
<th>KDS used</th>
</tr>
</thead>
<tbody>
<tr>
<td>$CL_0$</td>
<td>$GN_3$</td>
<td>$GN_6$</td>
</tr>
<tr>
<td>Sr</td>
<td>56</td>
<td>31</td>
</tr>
<tr>
<td>Ba</td>
<td>191</td>
<td>81</td>
</tr>
<tr>
<td>Rb</td>
<td>475</td>
<td>538</td>
</tr>
</tbody>
</table>

$CL_0$, trace element composition in the residual melt; $CL_0$, trace element composition in the parental magma; $F$, residual melt fraction; $D$, global partition coefficient. Sr, Ba and Rb contents in ppm. $Bt$, biotite; $Qtz$, quartz; $Pl$, plagioclase; $Fk$, K-feldspar; $Ap$, apatite; $Ilm$, ilmenite.

11. Concluding remarks

Given the apparent synchronous emplacement of both Aguiar da Beira syntectonic granitoids (322–317 Ma) and the intracratonal shearing that characterized the D2 Variscan deformation event, it may be presumed that the steep D3 shear zones could have acted as structural conduits for the ascent of these magmas and that their emplacement occurred in an essentially transcurrent regime related to strike-slip, sinistral and dextral subvertical shear zones (D3). While the overall chemical and isotopic signature of the muscovite–biotite leucogranite can well be explained by partial melting of an old supracrustal metasedimentary source, with high $87Sr/86Sr$ and low $\varepsilon Ndt$, the genesis of the biotite granodiorite–granite suite appears to require a less radiogenic metagneous crustal source or mixing of lower crustal and mantle derived magmas. Decompression melting of mid-upper crust metasedimentary rocks, triggered by the D2 extensional event related to crustal gravitational collapse, could account for the origin of the leucogranite suite, whereas mantle melting and concomitant underplating of the lower crust is one of the possible mechanisms to explain the generation of the biotite granodiorite–granite suite.

### Acknowledgments

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Appendix 1. Zircon oxygen isotopic individual analysis of each of the samples analyzed from the biotite granodiorite–granite and the muscovite–biotite leucogranite

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Background corrected</th>
<th>δ18O V-SMOW</th>
<th>±1 σ (abs)</th>
<th>±2 σ (abs)</th>
</tr>
</thead>
<tbody>
<tr>
<td>01 – lp tip</td>
<td>0.00201777</td>
<td>9.37</td>
<td>±0.31</td>
<td>±0.62</td>
</tr>
<tr>
<td>01b – lp tip</td>
<td>0.00201743</td>
<td>9.00</td>
<td>±0.33</td>
<td>±0.66</td>
</tr>
<tr>
<td>02a – lp tip</td>
<td>0.00201830</td>
<td>8.34</td>
<td>±0.34</td>
<td>±0.68</td>
</tr>
<tr>
<td>02b – lp edge of tip</td>
<td>0.00201457</td>
<td>6.47</td>
<td>±0.31</td>
<td>±0.62</td>
</tr>
<tr>
<td>03 – lp center</td>
<td>0.00201774</td>
<td>8.07</td>
<td>±0.29</td>
<td>±0.58</td>
</tr>
<tr>
<td>03a – lp tip</td>
<td>0.00201852</td>
<td>8.46</td>
<td>±0.29</td>
<td>±0.58</td>
</tr>
<tr>
<td>03b – lp center</td>
<td>0.00201757</td>
<td>7.99</td>
<td>±0.30</td>
<td>±0.60</td>
</tr>
<tr>
<td>03c – lp tip</td>
<td>0.00201805</td>
<td>8.23</td>
<td>±0.29</td>
<td>±0.58</td>
</tr>
<tr>
<td>04 – lp tip</td>
<td>0.00201716</td>
<td>7.78</td>
<td>±0.33</td>
<td>±0.66</td>
</tr>
<tr>
<td>04b – lp tip</td>
<td>0.00201652</td>
<td>7.46</td>
<td>±0.30</td>
<td>±0.60</td>
</tr>
<tr>
<td>05a – lp edge of tip</td>
<td>0.00201519</td>
<td>6.80</td>
<td>±0.29</td>
<td>±0.58</td>
</tr>
<tr>
<td>05b – lp tip</td>
<td>0.00201752</td>
<td>7.98</td>
<td>±0.32</td>
<td>±0.64</td>
</tr>
<tr>
<td>06b – lp center, alt</td>
<td>0.00201739</td>
<td>7.90</td>
<td>±0.34</td>
<td>±0.68</td>
</tr>
<tr>
<td>07a – lp tip</td>
<td>0.00201740</td>
<td>5.05</td>
<td>±0.39</td>
<td>±0.78</td>
</tr>
<tr>
<td>Average of 12 spots (without 2b, 5a)</td>
<td></td>
<td>8.05</td>
<td>±0.31 (2σ)</td>
<td>±0.62 (2σ)</td>
</tr>
</tbody>
</table>

Muscovite–biotite leucogranite — sample CN3

<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Background corrected</th>
<th>δ18O V-SMOW</th>
<th>±1 σ (abs)</th>
<th>±2 σ (abs)</th>
</tr>
</thead>
<tbody>
<tr>
<td>01 – isolated tip</td>
<td>0.00202040</td>
<td>9.39</td>
<td>±0.32</td>
<td>±0.64</td>
</tr>
<tr>
<td>03 – lp tip</td>
<td>0.00202046</td>
<td>9.42</td>
<td>±0.33</td>
<td>±0.66</td>
</tr>
<tr>
<td>04 – lp tip</td>
<td>0.00202017</td>
<td>9.37</td>
<td>±0.36</td>
<td>±0.72</td>
</tr>
<tr>
<td>05a – lp tip</td>
<td>0.00202109</td>
<td>9.75</td>
<td>±0.29</td>
<td>±0.58</td>
</tr>
<tr>
<td>07a – lp center</td>
<td>0.00202065</td>
<td>9.53</td>
<td>±0.33</td>
<td>±0.66</td>
</tr>
<tr>
<td>07b – lp tip</td>
<td>0.00202083</td>
<td>9.61</td>
<td>±0.31</td>
<td>±0.62</td>
</tr>
<tr>
<td>08a – lp tip</td>
<td>0.00202091</td>
<td>9.65</td>
<td>±0.32</td>
<td>±0.64</td>
</tr>
<tr>
<td>08b – lp tip</td>
<td>0.00201991</td>
<td>9.23</td>
<td>±0.31</td>
<td>±0.62</td>
</tr>
<tr>
<td>Average of 8 spots</td>
<td></td>
<td>9.48</td>
<td>±0.19 (2σ)</td>
<td>±0.38 (2σ)</td>
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</table>

References


