



Petrogenesis of a late-Variscan rhyodacite at the Ossa Morena-Central Iberian zones boundary, Iberian Massif, Central Portugal: Evidence for the involvement of lithospheric mantle and meta-igneous lower crust

S.B.A. Henriques^{a,*}, A.M.R. Neiva^b, G.R. Dunning^c

^a Geobiotec, LNEG: Geology, Hydrogeology and Coastal Geology Department, Estrada da Portela, Bairro do Zambujal, Ap. 7586-Alfragide, 2610-999 Amadora, Portugal

^b Geobiotec, Department of Earth Sciences, University of Coimbra, 3030-790 Coimbra, Portugal

^c Earth Sciences Department, Memorial University, St. John's, NL A1B 3X5, Canada

ARTICLE INFO

Article history:

Received 18 February 2016

Received in revised form 8 June 2016

Accepted 11 June 2016

Editorial handling - A. Holzheid

Keywords:

Carboniferous

Igneous monazite

Inherited zircon

ID-TIMS U-Pb geochronology

Sr, Sm-Nd and $\delta^{18}\text{O}$ isotopes

ABSTRACT

A late-Variscan rhyodacite is exposed at the contact between the Ossa Morena Zone and the Central Iberian Zone of the Iberian Massif, Central Portugal. Dykes of rhyodacite intruded the Série Negra Unit and the Sardeal Complex that are part of the Cadomian basement. The igneous crystallization age of the rhyodacite (308 ± 1 Ma) was obtained on igneous monazite by the ID-TIMS U-Pb method. It is broadly coeval with the emplacement of late-Variscan granitoids during the last deformation phase of the Variscan Orogeny (ca. 304–314 Ma) and with the development of the large late-Variscan strike-slip shear zones (ca. 307 Ma). The rhyodacite samples are calc-alkaline, show identical composition and belong to the same magmatic sequence. The rhyodacite isotopic signatures (Sm-Nd and $\delta^{18}\text{O}$) are consistent with depleted-mantle juvenile sources and the contribution of the meta-igneous lower crust. The input of mantle juvenile sources is related to Variscan reactivation of lithospheric fractures. The inherited Neoproterozoic (ca. 619 Ma) and Mesoproterozoic (ca. 1054 Ma) zircon ages, are similar to those of the Central Iberian Zone. This suggests that lower crust of the Central Iberian Zone was involved in the magma generation of the rhyodacite. Coeval late-Variscan magmatic rocks display a larger contribution from ancient crustal components, which may be attributed to the smaller volume and faster cooling rate of the rhyodacite and consequent lower melting of the crust. Mixing of juvenile mantle-derived melts with melts from the lower continental crust was followed by fractional crystallization of garnet and amphibole that remained in the source. Fractional crystallization of plagioclase, biotite, quartz and zircon occurred in shallower magma chambers. Fractional crystallization of zircon was not significant.

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1. Introduction

The closure of the Rheic Ocean occurred in the Devonian-Carboniferous due to the continental collision between Laurussia, the peri-Gondwanan terranes and Gondwana that gave rise in Europe to the Variscan orogeny (e.g. Gutiérrez-Alonso et al., 2004; Matte, 2001; Murphy et al., 2014 and references therein). During the post collisional stage of the Variscan orogeny large volumes of granitoids were emplaced in the Central Iberian Zone and at the contact with the Ossa Morena Zone (e.g. Carracedo et al., 2009; Dias

et al., 2002; Ferreira et al., 1987). The late-Variscan granitoids (ca. 306–311 Ma; e.g. Dias et al., 2002) correspond to abundant S-type granites (e.g. Teixeira et al., 2011), rare I-type granites (e.g. Villaseca et al., 2009) and hybrid granites (e.g. Gomes and Neiva, 2005).

At the contact between the Ossa Morena Zone and the Central Iberian Zone, a late-Variscan rhyodacite crops out, near Abrantes, Central Portugal. This contact is a key locality for the understanding of the late-Variscan magmatism, because it represents an important Neoproterozoic suture reactivated during the Variscan orogeny (e.g. Eguiluz et al., 2000; Henriques et al., 2015; Ribeiro et al., 2009). In this paper we present the first study of the rhyodacite, which includes the description of field relationships, petrography, mineral chemistry, U-Pb geochronology, whole rock geochemistry and isotopic data (Sr, Sm-Nd, $\delta^{18}\text{O}$). The aim of this study is to characterize the magmatic sources of the rhyodacite as well as the main magmatic processes involved in its petrogenesis.

* Corresponding author.

E-mail addresses:

sbahgeo@gmail.com, susana.henriques@lneg.pt (S.B.A. Henriques), neiva@dct.uc.pt (A.M.R. Neiva), gunning@mun.ca (G.R. Dunning).

In order to contribute to a better understanding of the late-Variscan magmatism in the orogen a comparison with other correlated units is presented.

2. Geological setting

The study area is located in the Iberian Massif, at the contact between the Ossa Morena Zone and the Central Iberian Zone, within the Tomar-Badajoz-Córdoba Shear Zone, near Abrantes, Central Portugal (Fig. 1a). The Ossa Morena Zone and the Central Iberian Zone are tectonostratigraphic units of the Iberian Massif (e.g. Arenas et al., 1988; Julivert et al., 1972; Lotze, 1945; Ribeiro et al., 2010). The Ossa Morena Zone corresponds to a continental arc, adjacent to the West African craton (e.g. Nance et al., 2008) and accreted to the Iberian Autochthon during the Neoproterozoic–earliest Cambrian (e.g. Sánchez-García et al., 2008 and references therein) more precisely at ca. 540 Ma (Henriques et al., 2015). The Central Iberian Zone is a continental fragment of the Iberian Autochthon (e.g. Sánchez-García et al., 2008 and references therein). A complex tectonometamorphic evolution related to the Cadomian and Variscan orogenies has been documented by several authors for the Ossa Morena Zone (Eguíluz et al., 2000; Henriques et al., 2015; Quesada and Munhá, 1990; Ribeiro et al., 2009). The tectonostratigraphic sequence identified in the area comprises the following units from bottom to top: the Série Negra Unit, the Sardeal Complex and the Mouriscas Complex (Fig. 1b), (Henriques et al., 2015).

The Série Negra Unit is a volcano-sedimentary succession deposited in a back-arc basin of Ediacaran age (ca. 590–550 Ma; Linnemann et al., 2008), deformed and metamorphosed under variable *P-T* conditions (Alfá, 1963; Carvalhosa, 1965; Eguíluz et al., 2000). In the area, it consists of phyllite, quartz-phyllite, metagreywacke, schists (micaschist, quartz-mica schist, quartzofeldspathic schist) with interbedded black chert, metavolcaniclastic rocks, acid metavolcanic rocks, greenschist and amphibolite (Fig. 1b).

The Sardeal Complex comprises deformed metamorphic rocks mostly of igneous origin: quartzofeldspathic schist, felsic orthogneiss, migmatite, gneissic granite with amphibolite, amphibolitic gneiss and gabbro (Fig. 1b). It has an igneous age of ca. 692–548 Ma and a metamorphic age of ca. 540 Ma (Henriques et al., 2015).

The Mouriscas Complex corresponds to a deformed and metamorphosed predominantly mafic igneous complex. It consists of amphibolite, amphibolitic gneiss, garnet amphibolite, amphibolite dykes, protomylonite trondhjemite, metatexite and diatexite (Fig. 1b). Some of these rocks have igneous ages of ca. 544 Ma and 483–477 Ma (Henriques et al., 2015).

The geochronological and isotopic data allowed the establishment of a sequence of events: onset of island arc magmatism accompanied by deposition of fine-grained sediments in a Neoproterozoic basin; magmatism related to the Cadomian continental arc, with generation of subalkaline and peraluminous protoliths with the isotopic signature of old crustal sources at ca. 692 Ma and 569 Ma (Henriques et al., 2015); a final phase of the arc magmatism (ca. 548–544 Ma) that involved mainly partial melting of continental crust; continental arc accretion of the Ossa Morena Zone with the Iberian Autochthon passive margin (future Central Iberian Zone) at ca. 540 Ma; early Ordovician magmatism (ca. 483–477 Ma) related to the opening of Rheic Ocean and metamorphism at amphibolite facies conditions related to the Variscan orogeny (Henriques et al., 2015). These events have correlated units in other areas of the Cadomian orogen (Henriques et al., 2015).

The rhyodacites intruded the Série Negra Unit and the Sardeal Complex (Fig. 1b) and the contact with the host rocks is sharp

(Fig. 2a and b). These intrusions occur as dykes with a porphyritic texture (Fig. 2a–c). They are from a few meters to hundreds of meters long and from centimeters to several meters wide. The trend of the dykes is N-S, NNW-SSE, NW-SE and ENE-WSW (Zbyszewski et al., 1981).

3. Analytical methods

Most mineral analyses were performed on a CAMECA SX-100, at the Scientific-Technological Centre of the University of Oviedo, Spain. The operating conditions for silicates and oxides were 5 μm beam diameter, 15 kV accelerating voltage, and 15 nA beam current. For sulfides analytical conditions were 1 μm beam diameter, 20 kV accelerating voltage and 20 nA beam current. The peak counting times were 10 s for most elements except for V (20 s), Co and Ni (5 s). Some of the mineral analyses were carried out on a Jeol 8500-F at LNEG, Portugal, with identical operating conditions for silicates, oxides and sulfides. The peak counting times were 15 s. Precision is better than 2%.

ID-TIMS U-Pb geochronology was performed at the Earth Sciences Department of Memorial University of Newfoundland. The sample was crushed, pulverized and sieved to dimensions < 40 mesh. Heavy mineral fractions were separated using the Wilfley table, heavy liquids and the Frantz magnetic separator. The final fractions were observed under a binocular microscope and the best quality zircon and monazite grains were picked for analysis. Cathodo-luminescence images of zircon were obtained on a FEI Quanta 400 SEM. The selected zircon and monazite crystals were abraded using the Krogh (1982) technique. Mineral fractions were washed and weighted. Zircon fractions were mixed with a ^{205}Pb - ^{238}U spike (Parrish and Krogh, 1987) and dissolved with concentrated HF and 1 drop of 8 N HNO_3 in Krogh-type Teflon® dissolution bombs, that were kept in the oven at a temperature of 220 °C for 5 days. Monazite fractions were mixed with a ^{205}Pb - ^{235}U spike and dissolved with distilled 6.2 N HCl in Savillex® Teflon containers which were heated on a hot-plate for 5 days at 120 °C (Parrish, 1990). U and Pb were separated through ion exchange chemistry according to modified procedures for zircon chemistry after Krogh (1973), reducing the volume of the columns and the reagents by a ratio of 1/10, and HBr chemistry after Manhães et al. (1978) for monazite. The isotopic ratios were measured in a Finnigan MAT 262 thermal ionization mass spectrometer, and ages were calculated using the decay constants of Jaffey et al. (1977). The uncertainties on the isotopic ratios were calculated at 2σ using the same software and taking into account the uncertainties of the isotopic ratio measurements, the U and Pb fractionation, the laboratory procedure blanks (1–10 pg Pb) and the initial common Pb, which was corrected following the model of Stacey and Kramers (1975). The age of the rhyodacite is the weighted average $^{206}\text{Pb}/^{238}\text{U}$ age calculated with ISOPLOT (Ludwig, 2003) with uncertainties reported at the 95% confidence interval.

Whole rock major and trace elements analyses were obtained for 5 samples at Activation Laboratories–ACTLABS (Canada). Samples weighing more than 1 kg were crushed. The major elements abundances were determined by ICP-AES and the trace elements abundances, including the REE, were determined by ICP-MS (with the exception of Sc, V, Sr and Ba, which were determined by ICP-AES). The detection limits are 0.001 wt.% for MnO and TiO_2 and 0.01 wt.% for the remaining oxides and lower than 1 ppm for most trace elements, except for Sr (2 ppm), Ba (3 ppm), V (5 ppm), Cu (10 ppm), Cr and Ni (20 ppm) and Zn (30 ppm). The precision for major elements is better than 4% and better than 10% for trace elements. Whole rock FeO concentrations were determined by titration with a standardized potassium permanganate solution

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