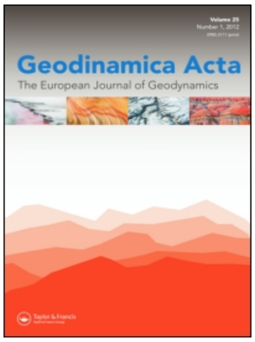

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The Beja Layered Gabbroic Sequence (Ossa-Morena Zone, Southern Portugal): geochronology and geodynamic implications

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Abstract

The Beja Igneous Complex (BIC) is a major geological feature in SW Iberian Variscides, consisting of three main units developed during different stages of the oblique collision between the Ossa Morena Zone (OMZ) upper plate and South Portuguese Zone (SPZ) lower plate, namely: (1) *ca*. 355 Ma to *ca*. 345 Ma Layered Gabbroic Sequence formed in early stages of collision magmatism; (2) the *ca*. 335-330 Ma to *ca*. 320 Ma Cuba-Alvito (gabbro-diorite) Complex formed throughout the late-collision magmatic event; and (3) the Baleizão Porphyry Complex corresponding to the period of post-collision magmatism, *ca*. 300 Ma. The new SHRIMP U-Pb age of 342±9 Ma reported here for amphibole-bearing pegmatite dykes cutting the layered gabbros is interpreted as dating the development of late fluid-rich melts in the Layered Gabbroic Sequence, synchronous with Fe-Cu-Co sulphide deposition. The close agreement between this data and available amphibole $^{40}Ar/^{39}Ar$ ages of BIC, Beja-Acebuches Ophiolite and other geological units of the OMZ southern border, may be taken as evidence for a moderate to rapid regional crustal uplift episode at *ca*. 340±5 Ma; this data, coupled with structural constraints, also allow to estimate the age for the transition between the D_{2a} – D_{2b} deformation phases of Variscan continental collision.

A complex wedge system within the SW Iberian Variscides developed during this collision, involving the OMZ upper plate to the north and the SPZ passive margin in the lower plate. The Évora-Beja-Aracena Domain, located in the upper plate above the N-dipping subduction zone, is re-interpreted as a retro-wedge domain that was kinematically coupled to the SPZ pro-wedge and subduction system. Retro-wedge growth is linked to upper plate uplift (early collision) and a late-orogenic wedge thickening. The early stages of magmatism in the retro-wedge are related to asthenospheric mantle upwelling induced by the slab break-off. Regional LP-HT metamorphism and subsequent magmatic events in the retro-wedge domain were caused by long term high heat flow sustained by (1) mafic magma underplating, (2) stacking of high-heat producing upper-crustal lithologies, and reinforced (3) by (moderate to) rapid crustal uplifting. Mass advection and orogenic architecture were strongly affected by asymmetric removal towards the lower-part foreland basin and by transient mechanical properties of the wedge system associated with the anomalous thermal regime.

Keywords: Beja Igneous Complex; U-Pb geochronology; Palaeozoic magmatic activity in SW Iberian Variscides

1. Introduction

The Ossa-Morena Zone (OMZ) is an inner tectono-stratigraphic unit of the SW Iberian Variscides, exposed between the Central-Iberian Zone (CIZ) to the north and the South Portuguese Zone (SPZ) to the south [1-4]. Different belts in the OMZ can be recognised on the basis of palaeo-geographic, magmatic, metamorphic and structural criteria. Nonetheless,

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questions remain concerning the geodynamic evolution of the OMZ on Cadomian and Variscan times. The OMZ southern border corresponds to a distinct, curved and wedge shaped belt known as the Évora-Beja-Aracena Domain [4, 5] or as the Southern Crystalline Complexes [6] (Fig. 1). The transition towards SPZ is marked by deformed and metamorphosed Exotic Terranes, namely the Beja-Acebuches Ophiolite Complex (BAOC) and the Pulo do Lobo Terrane (PLT) [7-9].

The Beja Igneous Complex (BIC) dominated by large scale Palaeozoic igneous activity is a major geological feature of the Portuguese segment of the OMZ southern border in SW Iberian Variscides. The formation of this large Complex remains, however, somewhat controversial. In earlier conceptions, the onset of BIC development is envisaged as a result of mantle-derived magma rise triggered during active subduction of oceanic lithosphere, in view of the OMZ-SPZ collision in pre-Famennian times [8, 10-12]. More recently, Pin *et al*. [13] suggested that the emplacement of the BIC gabbroic unit took place in an extensional (probably, transtensional) tectonic setting at *ca*. 350 Ma (Late Tournaisian) as indicated by U-Pb ID-TIMS zircon ages. Both perspectives have significant consequences on the interpretation of the geodynamic evolution experienced by the SW Variscides, thus demanding new geochronological data and integration of multidisciplinary information concerning the OMZ-SPZ boundary. In this context, the discussion concerning the mechanisms that can provide and sustain the heat sources required both for syn-convergent metamorphism/magmatism and post-convergent magmatism, emerge as a fundamental issue.

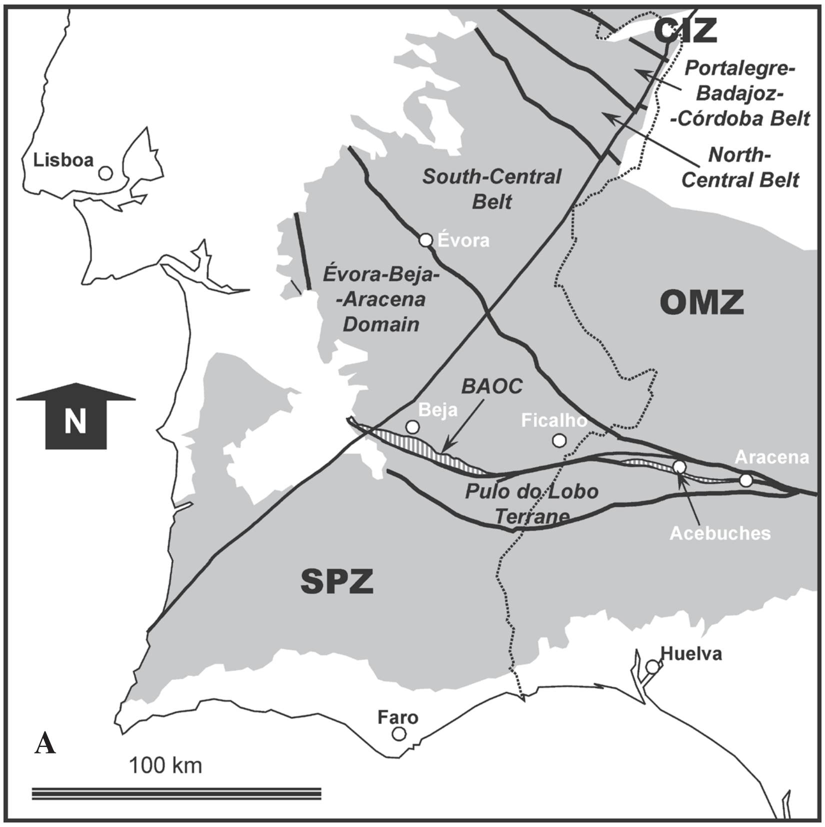


Fig. 1: A: Schematic representation of the SW Iberian Variscides tectono-stratigraphic units. **B**: Simplified geological map of the OMZ southern border in Portugal (adapted from the Geological Map of Portugal, 1:500000 scale, 1992 – J.T. Oliveira and E. Pereira, compilers); note the location of BIC. The location of Beja and Évora cities is also given as B and E, respectively.

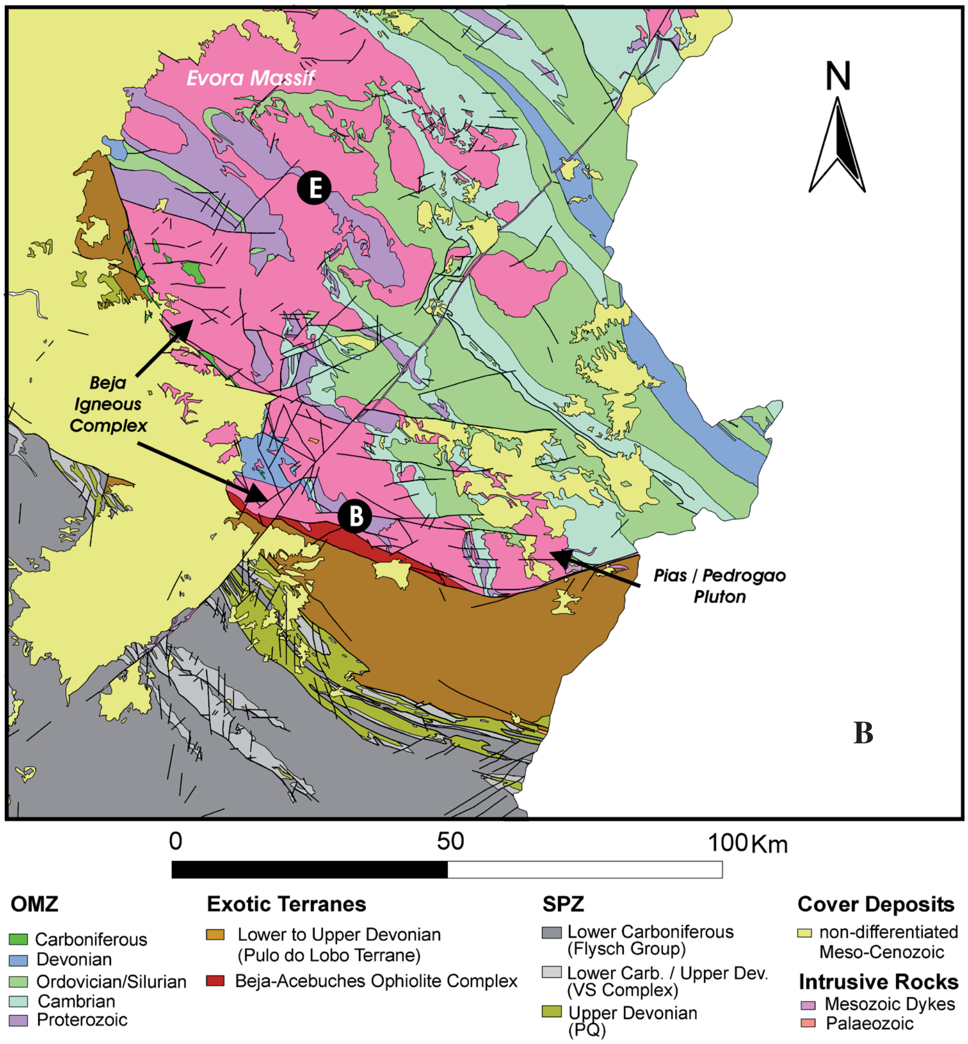
In this paper, a new U-Pb SHRIMP date for a BIC rock of the Évora-Beja-Aracena Domain is reported and compared with recent dating presented by different authors for other sectors of this igneous complex. Regional geochronological constraints are also evaluated, integrating published data for the Variscan igneous activity (and metamorphism) recorded either in the OMZ southern border or in the adjoining Exotic Terranes and SPZ northern border. On this basis, an interpretative general geodynamic evolution is reconstructed, allowing to get an insight on the processes that may account for the thermal regime (and high-T metamorphism) experienced by the Évora-Beja-Aracena Domain from early to late stages of the OMZ-SPZ collision an its relation with widespread magmatism.

2. Geodynamic framework of the SW Iberian Variscides: setting of the Évora-Beja-Aracena Domain

In order to better constrain the geodynamic framework of the SW Iberian Variscides it is fundamental to build a broader regional perspective on the subject, incorporating the available data for the adjoining Exotic Terranes (Beja-Acebuches Ophiolite Complex – BAOC, and Pulo do Lobo Terrane) and the SPZ northern border.

The BAOC is a tectonically dismembered ophiolite sequence within the South Iberia Variscan Suture showing transitional characteristics between MORB and island arc basalts compatible with a back arc basin tectonic setting [4, 7-9, 12, 14-20]. BAOC was emplaced by antithetic obduction towards North, as evidenced by early-developed anisotropic fabrics indicating northwards shearing [12, 15, 17] preserved in the lower and intermediate sections comprising meta-peridotites and meta-gabbroic rocks (well represented in the Portuguese part of the belt). The upper sections of BAOC consist essentially of amphibolites derived from tholeiitic basaltic lavas. Both the northern and southern limits of BAOC are tectonic, and correspond to (re-activated) reverse-sinistral WNW-ESE shear zones with strong dip towards SSW. Geochronologic data for the Portuguese part of BAOC are limited to $^{40}Ar/^{39}Ar$ measurements on amphibole concentrates carried out by Dallmeyer *et al*. [21]. The reported dates (Table 1) are consistent with other dates obtained for the Oceanic Domain of the Aracena Metamorphic Belt [22], and have been interpreted as the age of a regional cooling to under ≈ 500ºC at *ca*. 345 Ma.

The Pulo do Lobo Terrane (PLT) comprises distinct meta-sedimentary successions that locally include bimodal meta-



volcanics (with dominantly N-MORB type mafic rocks) that display strong multiphase deformation [12]. Three main stratigraphic Formations can be grouped according to the available palynological data, ranging from Lower-Middle to Late Devonian (Givetian-Frasnian) [10, 14, 15, 23-27]. Metamorphic conditions are typically of lower greenschist facies, although rocks belonging to the uppermost Formation reveal very low grade metamorphism [28]. The PLT displays strong multiphase deformation that can be ascribed to three main Variscan deformation phases with folding/axial plane cleavage development, plus important thrust systems with prevailing southwards shear sense [15, 24, 29-31]. It should be noted, however, that structures formed in the northern sectors of the Terrane are consistent with a dominant sinistral strike-slip tectonic regime while those developed in the southern sectors reflect a chief dextral strike-slip tectonic regimes. In fact, the contact of the Pulo do Lobo Terrane with the SPZ northern border is characterized by dextral (to reverse-dextral) shearing structures (from Paymogo to Campofrio areas), which seem to be related to the late syntectonic emplacement of the Gil Márquez granodiorite dated 328±2 Ma [30, 32, 33 – Table 1]. According to Onézime *et al.* [31], this dextral (to reverse-dextral) shearing typifies the last deformation episode recorded in both the Pulo do Lobo Terrane and the SPZ northern border, and may be related to dextral transtension associated with the emplacement of late-stage pulses of the (tonalitic-granitic) Sierra Norte Batholith.

The SPZ northern border corresponds mostly to the Iberian Pyrite Belt (IPB) classically described as a succession of three major stratigraphic Formations dated from Devonian to Early Carboniferous (Dinantian) ages and affected by low-grade metamorphism. The Phyllite-Quartzite Formation (Fammenian- Upper Devonian) [34-37] is the older meta-sedimentary sequence, whose base has never been observed. Lying upon the Phyllite-Quartzite Formation is the Volcano-sedimentary Formation, composed of a bimodal volcanic suite (despite the large predominance of felsic units, hosting world-class VMS deposits) inter-fingered with diverse meta-sedimentary rocks [38-46]. Bio-stratigraphic data provided early Fammenian to Late Visean ages (Upper Devonian to Lower Carboniferous) for this Formation [47, 48], indicating as well a very narrow period (Strunian) for the development of VMS ores [49]. Geochronological data obtained in the last decade for different igneous rocks of this Formation, prove to be either internally consistent or compatible with the available litho- and bio-stratigraphic informations, bracketing the more significant magmatic/hydrothermal activity between *ca.* 355 Ma and *ca.* 345 Ma [50-56 – Table 1]. The late stratigraphic succession in the IPB consists of the flysch sequence (starting with the Mértola Formation) dated as Visean age [38, 48]. To the northeast, the early magmatic rocks related to Sierra Norte Batholith intrude the IPB, being coeval with the igneous activity in the IPB [50, 55-57]. The major structural feature of the IPB consists of a south-verging thrust-fold belt complex that caused significant crustal thickening and stacking. This structural configuration, compatible with a thin-skinned geometry, was developed in the course of two main deformation phases, presumably on top of a major cryptic *décollement* correlated with the lower-upper Devonian transition within the Phyllite-Quartzite Formation [29, 31, 58-62]. Data gathered so far support the conclusion that the structural evolution experienced by the IPB is controlled by the OMZ-SPZ oblique collision.

3. The Évora-Beja-Aracena Domain and present geochronological constraints

The internal litho-stratigraphic arrangement of the Évora-Beja-Aracena Domain reflects overprinting of effects caused by a complex geodynamic evolution, such as: 1) the profusion of diversified Variscan igneous intrusions suites, particularly well represented in its NW part; 2) the widespread development of Variscan thrust zones, leading to extensive tectonic imbrication and strong strain partitioning; and 3) the extensive formation of late, sub-vertical shear zones, which often result in strong dismembering of the pre-existent geological framework. On the basis of the available data [63-67], the evolution of the Portuguese part of the Évora-Beja-Aracena Domain can be briefly outlined as follows.

Upper Proterozoic sequences corresponding to the "Série Negra" (black-schists meta-cherts, mica-schists, amphibolites and felsic gneisses [*e.g.* 64]) are mostly preserved in the NW of the Évora-Beja-Aracena Domain and display deformation and LP metamorphism. Metasedimentary and metavolcanic sequences formed during the pre-orogenic Cambrian-Ordovician period often record a complex internal architecture as a consequence of thrusting and emplacement of eclogite/blueschist tectonic slices. Thick metasedimentary sequences (with minor metavolcanics) tentatively ascribed to Silurian that can be found at SE has been interpreted as tectonic *mélanges* [69, 70]. Finally, the uppermost meta-sedimentary sequences (Devonian to Carboniferous) correspond to synorogenic deposits of predominant flyshoid nature and are usually preserved in small windows within igneous complexes and syntectonic basins [*e.g* 64].

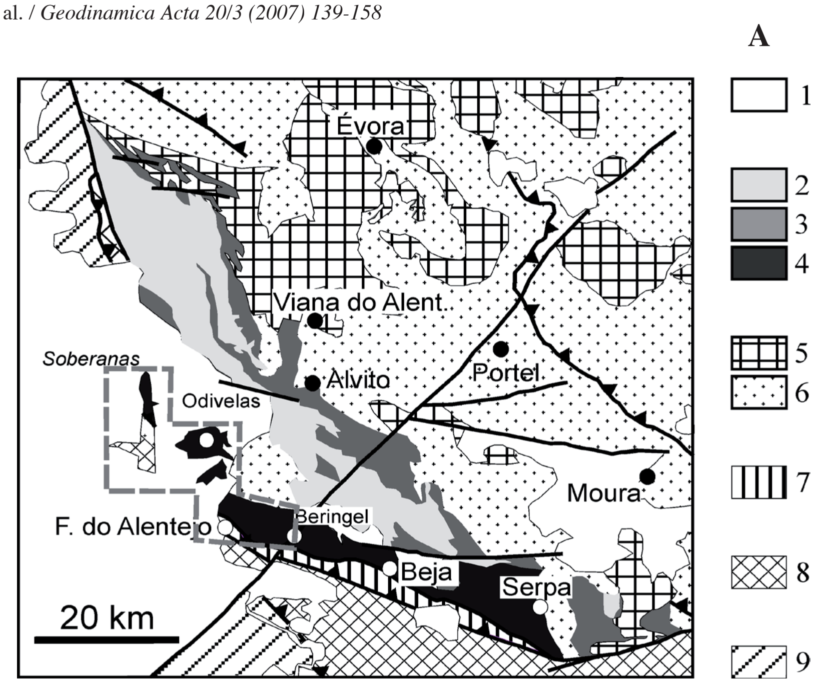
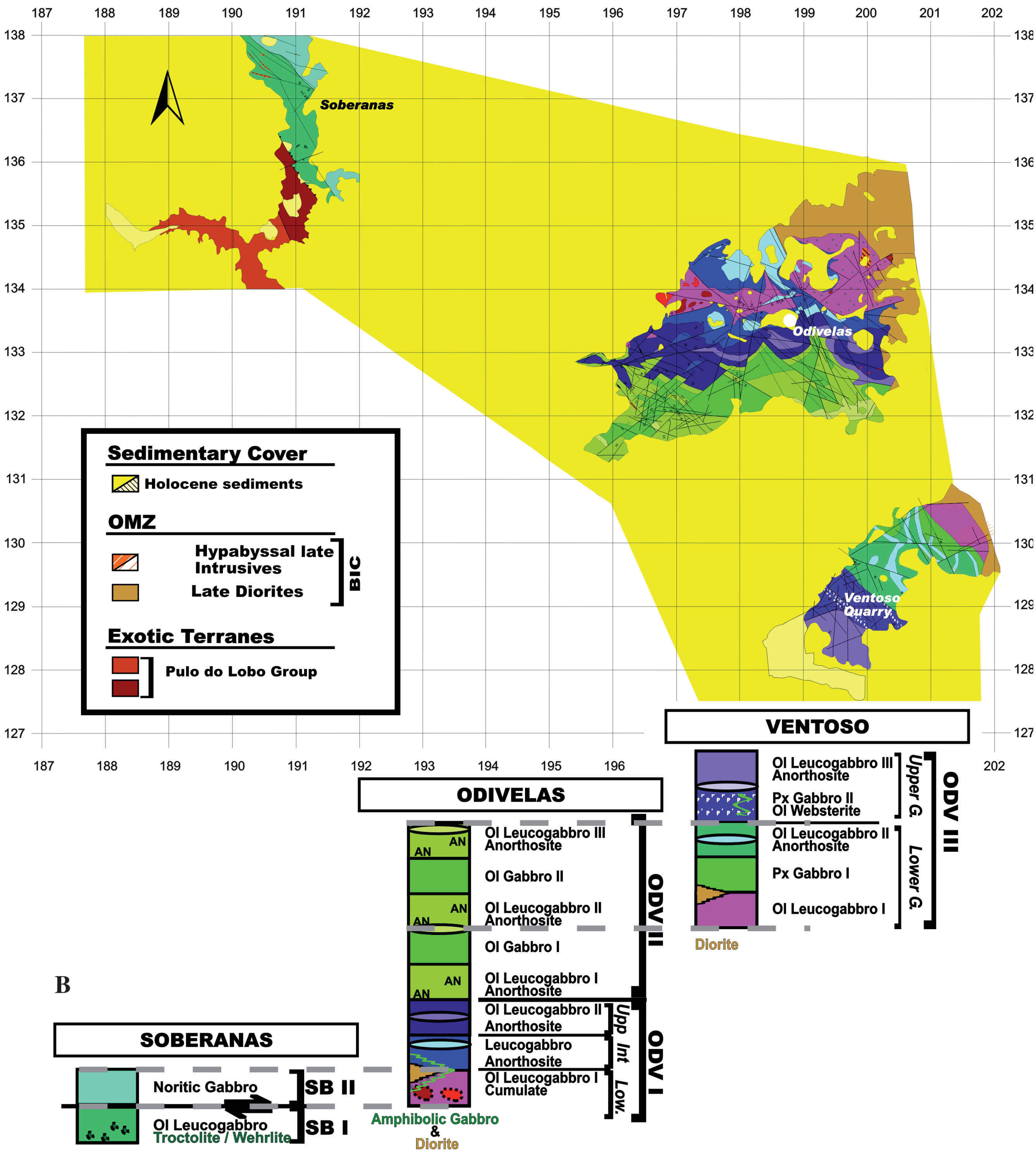


Fig. 2: A: Geological map of BIC illustrating the spatial distribution of its major units; adapted from Andrade (1983), considering field and petrographic data acquired in the last decade [93-98]: **1**, Cenozoic sedimentary cover; **2**, Baleizão Porphyry Complex; **3**, Cuba-Alvito Complex; **4**, Layered Gabbroic Sequence; **5**, Undifferentiated Variscan granitoids; **6**, Undifferentiated meta-sedimentary, meta-volcanic sequences and ultramafic rocks; **7**, Beja-Acebuches Ophiolite Complex; **8**, Meta-sediments and meta-volcanics of Pulo do Lobo Group; **9**, Meta-sedimentary and meta-volcanic sequences of SPZ. The dashed polygon corresponds to the LGS western compartment, presently mapped in detail.
B: Extract of the LGS western compartment geological map, covering the area of interest for this work.

In the Évora-Beja-Aracena Domain, Variscan deformation is multiphase and its effects are observable at all scales [12, 17, 66, 67, 69]. Structures formed during the first phase of deformation (D_1), usually striking WNW-ESE with a predominant shear sense towards N, are associated with the HP-LT event responsible by the development of eclogite/blueschist rocks (preserved at Safira, Viana do Alentejo – Alvito, Vidigueira – Vila de Frades) and the obduction / emplacement of BAOC shortly after [68, 69, 74-76]. For the older eclogite facies metamorphic event (dated at 371±17 Ma; garnet-whole-rock Sm-Nd) there is a clockwise P-T path peaking at 16-18 kbar / 600º-650ºC, subsequently followed by near isothermal decompression to 12-10 kbar before final cooling (T<550ºC) to greenschist facies conditions [68]. Most $^{40}Ar/^{39}Ar$ ages of amphibolitized eclogites range from 371 ± 11 Ma to 360 ± 4 Ma [68] and indicate that eclogite facies metamorphism in the OMZ was shortly followed by exhumation during (D_1) nappe emplacement. A considerably younger $^{40}Ar/^{39}Ar$ eclogite amphibole date (316 ± 6Ma) [68] is interpreted here as resulting from amphibole recrystallization (below 500-550ºC) due to thermal doming related to the late-stage widespread intrusion of granodiorite bodies in the OMZ Évora-Beja Domain [67, 73].

The multitude of intrusive igneous bodies that follow the early HP-LT event in the Évora-Beja-Aracena Domain represent a magmatic arc, supporting a northward polarity of Variscan subduction [*e.g.*, 12]. Fairly accurate tectono-magmatic associations can be made with the well-established major deformation regional events if the structural/field features associated to these magmatic suites are taken in account together with the geochronological constraints for their emplacement/crystallization/cooling.

Although D_2 is diachronic from SW to NE, several field-based studies support its subdivision into D_{2a} and D_{2b}, based on the distinct structural arrays ascribable to this deformation [17, 66, 72,]. Calc-alkaline to shoshonitic gabbro-dioritic intrusions in the NE region of the OMZ (Vale de Maceira – Veiros) are contoured by the (W verging NW-SE) D_{2a} regional foliation and have been dated at 364±12 Ma (mean weight average of 369±17 Ma and 358±18 Ma dates obtained for Vale de Maceira feldspar-whole rock Rb-Sr and amphibole K/Ar, respectively [71, Table 1]), representing mantle derived orogenic magmatism triggered by subduction, coeval with HP (eclogite facies) metamorphism (*ca.* 370 Ma; see above). The late 315±3 Ma biotite-whole rock Rb-Sr date for Vale de Maceira intrusion [71 – Table 1] represents biotite isotope resetting during regional greenschist/lower-amphibolite facies metamorphism (300-350ºC < T < 500 – 550ºC) associated with D_{2b} tectonothermal event.



Although D_2 metamorphism usually grades from greenschist to lower-amphibolite facies conditions, local anatexis is documented by migmatites and related granitoid rocks in the Montemor-o-Novo – Évora region [66, 72]. K-Ar and Ar/Ar amphibole/biotite dating of syn-D_{2b} Évora Massif Alvito diorites (318±11 Ma; Table 1, [67]) and Hospitais (Montemor-o-Novo) tonalitic (323±5 Ma to 317±8 Ma; Table 1, [73]) bodies are consistent with the younger ages obtained for Vale de Maceira and the Alvito-Viana eclogites (see above), providing further support for $D2_b$ tectonothermal event at *ca.* 320 Ma.

Structures developed throughout the third deformation phase (D_3) display a preferred NW-SE trend and preserve kinematical criteria indicative of prominent transport towards SW [17, 66]. Crystallization/cooling dates obtained for the post-collisional granitic bodies intruding the core of D_3 struc-

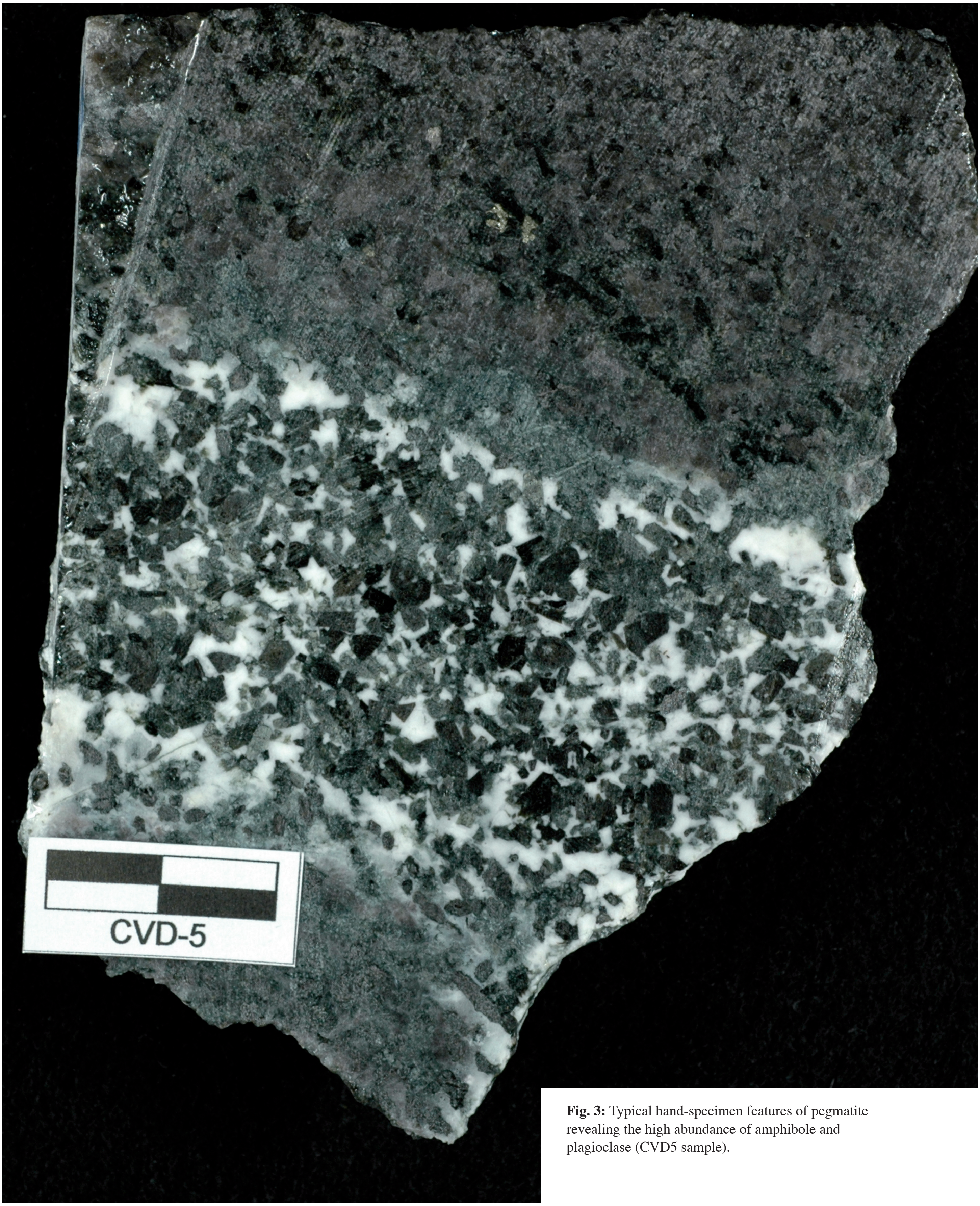


Fig. 3: Typical hand-specimen features of pegmatite revealing the high abundance of amphibole and plagioclase (CVD5 sample).

tures give ages of 308±4 Ma [77] and 306±8 Ma [71], whereas biotite K/Ar resetting related with D_3 deformation ranges from 307±8 Ma [73] to 298±3 Ma (Table 1). Thus, the final stages of OMZ-SPZ oblique collision may be bracketed between 312 Ma and 295 Ma, i.e. *ca*. 300 Ma.

Additional geochronological constraints for the geodynamic evolution of the Évora-Beja-Aracena Domain obtained in the Spanish counterparts are also summarized in Table 1. At the Aracena sector the LP-HT metamorphic have lasted for a long period during Early Carboniferous (343±1 Ma to 328±1 Ma, see refs. in [78]), postdating the earlier HP-LT metamorphic event documented by the eclogite rocks by *ca*. 10-40 million years. In the Aracena region, syn- to late-tectonic noritic bodies (high-Mg andesites with boninitic affinity) and post-tectonic intrusions of gabbros, diorites and meta-aluminous granites emplaced into the Continental Domain of the Aracena Metamorphic Belt, display an age interval consistent with that found in the Portuguese part of the Évora-Beja-Aracena Domain ([79-83], Table 1). Evidence for orogenic magmatism triggered by subduction and for late-collision processes can also be found to the North, particularly in the SE part of the South-Central Domain of the OMZ (Fig. 1). This is interpreted as due to the wedge-shape geometry shown by the SE-closure of the Évora-Beja-Aracena Domain in Spain. Crystallization and cooling ages for these igneous bodies of tonalitic to granitic nature were obtained with different methods (Table 1, [84-88]) ranging from 354±17 Ma (granodiorites from St. Olalla pluton) to *ca*. 279 ± 10 Ma (late collisional granite from S. Guillermo intrusive).

4. Beja Igneous Complex; main lithologic features and geochronology

The BIC extends for *ca*. 100 km along the southern edge of the Évora-Beja-Aracena Domain (Fig. 2) and comprises three main units [11, 89-96]: 1) the Beja Layered Gabbroic Sequence, composed of a wide range of gabbroic rocks rimmed by heterogeneous diorites; 2) the Cuba-Alvito Complex, mostly consisting of gabbro to (grano-)diorite rocks; and 3) the Baleizão Porphyry Complex, consisting of different granitoid rocks. As a whole, the BIC has tholeiitic (high-Al) to calc-alkaline geochemical affinities, the latter being more evident in the evolved/crustal contaminated units of Cuba-Alvito and Baleizão [91, 93-98].

Layered Gabbroic Sequence (LGS) can be divided in two major compartments, separated by the Messejana strike-slip fault zone. The western compartment, from W of Torrão to

Table 1: Summary of the most relevant geochronologic data for igneous and metamorphic rocks belonging to the OMZ southern border, Exotic Terranes and SPZ northern border

Age (Ma)	Method	Mineral / WR	Rock type	Study area	Reference
A) South Portuguese Zone Northern Border					
347±25	Rb-Sr		Massive ore	Neves Corvo	[54]
350±1	U-Pb	Zircon	Rhyolite	Riotinto	[55]
354±1	U-Pb	Zircon	Dacitic tuff	Las Cruces	[55]
353±2	U-Pb	Zircon	Green tuff	Aljustrel	[55]
356±1	U-Pb	Zircon	Rhyolitic tuff	Lagoa Salgada	[55]
346±26	Re-Os	Sulphide	Massive ore	Riotinto	[51]
353±44	Re-Os	Sulphide	Massive ore	Tharsis	[51]
348±2	U-Pb	Zircon	Rhyolite	Zufre	[56]
353±2	U-Pb	Zircon	Rhyolite	Nerva	[56]
351±8	Re-Os	Sulphide	Massive ore	Aznalcóllar	[53]
346±5	U-Pb	Zircon	Dacitic tuff	Los Frailles	[52]
347±2	U-Pb	Zircon	Ignimbrites	IPB	[115]
355±5	U-Pb	Zircon	Ignimbrites	IPB	[115]
368±26	$^{206}Pb/^{204}Pb$	Sulphide	Massive ore	IPB	[116]
336±98	Rb-Sr	Whole-rock?	Gabbros	N Sevilla	[117]
300±6	Rb-Sr	Whole-rock	Granite	El Berrocal	[50]
B) Exotic Terranes					
300±10	Rb-Sr	Whole-rock	Granodiorite	Gil Márquez	[33]
328±2	U-Pb	Zircon	Granodiorite	Gil Márquez	[32]
330±3	$^{40}Ar/^{39}Ar$	Biotite	Granodiorite	Gil Márquez	Ref in [31]
347±3	$^{40}Ar/^{39}Ar$	Amphibole	Amphibolite	BAOC	[21]
343±1	$^{40}Ar/^{39}Ar$	Amphibole	Amphibolite	BAOC	[21]
328±4	Rb-Sr	Whole-rock	Metanorite	Los Molares	[22]
340±23	Sm-Nd	Whole-rock	Metanorite	Los Molares	[22]
331±27	Rb-Sr	Whole-rock	Migmatite	Los Romeros	[22]
351±58	Rb-Sr	Whole-rock	Migmatite	Cortegana	[22]
328±4	Rb-Sr	Whole-rock	Metanorite	Los Molares	[118]
327(+51/-12)	U-Pb	Zircon	Tonalite	Aroches	[119]

▶

C) Ossa-Morena Zone Southern Border					
337±2	$^{40}Ar/^{39}Ar$	Amphibole	Amphibolite	S. Brissos	[21]
341±1	$^{40}Ar/^{39}Ar$	Amphibole	Amphibolite	Ventosa	[21]
338±1	$^{40}Ar/^{39}Ar$	Amphibole	Gabbro	Odivelas	[21]
340±1	$^{40}Ar/^{39}Ar$	Amphibole	Amphibolic Gabbro	Odivelas	[21]
350±4	U-Pb	Zircon	Gabbro	Serpa	[13]
352±4	U-Pb	Zircon	Gabbro	Torrão	[13]
331±7	K-Ar	Amphibole	Diorite	Alvito	[99]
371±17	Sm-Nd	Garnet-WR	Gt- Eclogite	Safira	[68]
371±11	$^{40}Ar/^{39}Ar$	Amphibole	Amph-Eclogite	Alvito	[68]
360±4	$^{40}Ar/^{39}Ar$	Amphibole	Amph-Eclogite	Alvito	[68]
316±6	$^{40}Ar/^{39}Ar$	Amphibole	Amph-Eclogite	Alvito	[68]
325±10	K-Ar	Amphibole	Gabbro-diorite	Viana-Alvito	[67]
318±11	K-Ar	Amphibole	Diorite	Viana-Alvito	[67]
324(+8/-5)	K-Ar	Muscovite	Porphyry	Alcaçovas	[100]
323±5	$^{40}Ar/^{39}Ar$	Amphibole	Tonalite	Montemor-o-Novo	[73]
317±8	K-Ar	Amphibole	Tonalite	Montemor-o-Novo	[73]
307±8	K-Ar	Biotite	Tonalite	Montemor-o-Novo	[73]
321±8	$^{40}Ar/^{39}Ar$	Biotite	Tonalite	Montemor-o-Novo	[73]
358±18	K-Ar	Amphibole	Gabbro	Vale Maceira	[71]
369±17	Rb-Sr	Feldspar-WR	Gabbro	Vale Maceira	[71]
315±3	Rb-Sr	Biotite-WR	Gabbro	Vale Maceira	[71]
306±8	Rb-Sr	Feldspar-WR	Granite	Sesmarias	[71]
298±3	Rb-Sr	Biotite-WR	Granite	Sesmarias	[71]
308±4	K-Ar	Muscovite	Granite	Pias-Pedrogão	[77]
348±4	Pb Evap	Zircon	Granite	Teuler	[118]
332±3	Pb Evap	Zircon	Tonalite	Sta Olalla	[118]
342±4	Pb Evap	Zircon	Mafic rock	Valuengo	[118]
340±7	Pb Evap	Zircon	Granite	Brovales	[118]
354±17	Rb-Sr	Whole-rock	Granodiorite	Sta Olalla	[86]
338±15	U-Pb	Allanite	Diorite	Burguillos	[86]
330±9	Rb-Sr	Whole-rock	Leucogranite	Burguillos	[120]
335±5	$^{40}Ar/^{39}Ar$	Amphibole	Diorite	Burguillos	[85]
328±10	K-Ar	Biotita	Granodiorite	Burguillos	[84]
279±10	K-Ar	Muscovite	Granite	S Guillermo	[84]
305±10	K-Ar	Muscovite	Granite	Brovales	[84]

Beringel, is much wider and comprises different successions of layered gabbroic rocks, the lower suites showing the most primitive Sr-Nd isotopic signature [13]. In the eastern compartment, from Beringel to Serpa, magmatic layering is quite difficult to observe at meso- to macro-scales, and evidence of crustal assimilation is much more common, as documented by Sr-Nd isotopic data [13] and a higher abundance of amphibole-rich diorites [e.g. 91]. $^{40}Ar/^{39}Ar$ (amphibole) and U-Pb (ID TIMS - zircon) dating on rocks from both compartments of this sequence is reported by [21] and [13], respectively (Table 1). U-Pb zircon ages (350±4 Ma and 352±4 Ma for samples collected E of Serpa and W of Torrão, respectively) date the emplacement and first crystallization step of the earlier gabbroic series. $^{40}Ar/^{39}Ar$ dating of LGS amphiboles (340±1 Ma and 343±1 Ma in Odivelas, E of Torrão) should represent the cooling age of the pluton below 550-500ºC [21], being consistent with those obtained for different amphibolite rocks belonging either to the OMZ southern border autochthonous formations (337±2 Ma and 341±1 Ma, [21]) and to the BAOC suite (347±3 and 343±1 Ma, [21]), thus providing evidence for a significant regional thermal event at about 345-335 Ma.

Recent data obtained for the westernmost portion of the LGS western compartment reveal a succession of distinct gabbroic rocks presently grouped in five series, supporting a multiphase history characterized by several magma replenishments [93, 94]. These rocks display a well-developed NW-SE layering, commonly associated with a magmatic lamination and gently dipping (< 35º) to the SW. Magmatic fabrics in LGS western compartment, as those in the Serpa region, [20] are geometrically consistent with D_{2b} regional foliation. From NE to SW, the five series are labelled Soberanas I and II, and Odivelas I, II and III (Fig. 2B). Soberanas I comprises primitive troctolite and wehrlite cumulates and is in tectonic contact with the more evolved leuconorite and leucogabbro of Soberanas II. The latter series records the early stages of the oxide-forming event responsible for the development of Fe-Ti-V mineralizations hosted by olivine leucogabbros and cumulates in the lower section of Odivelas I [94-96]. Odivelas II is a rhythmic succession of olivine gabbro and olivine leucogabbro that records a new magma input [93, 95] and ranges upwards to the Odivelas III series of (leuco-)gabbro, pyroxenite and pyroxenic gabbro, which contains disseminated Fe-Ni-Cu-Co sulphide blebs. Anastomosing veins of pyrrhotite

+ chalcopyrite ± pyrite are related to late metasomatic haloes and pegmatitic dykes that cut through the gabbroic rocks in the upper section of Odivelas III [93, 97, 98]. Zircons selected for U-Pb SHRIMP analysis reported below (see 3.1) were extracted from these pegmatites.

According with Andrade [91] the Cuba-Alvito gabbro-dioritic Complex may result from magma mixing and crustal contamination at the margin of the intrusion. Whole-rock and mineral (amphibole, biotite) K-Ar geochronology of this Complex in the Viana-Alvito region [67, 99] indicates ages from 331±7 to 325±10 Ma (Table 1). The Baleizão Porphyry Complex consists of a late epizonal-intrusive unit composed of a great diversity of granitoid rocks forming fairly small bodies of variable geometry that intersect the two aforementioned BIC units. Cartographic criteria and field relationships suggest that the Baleizão Porphyry Complex has a multistage development related to a long-lived magma system, probably beginning at Late Visean times, as indicated by K-Ar muscovite ages of 324±8 Ma ([100], Table 1).

4.1. SHRIMP U/Pb Zircon Geochronology

4.1.1. Petrographic features of pegmatite dykes and host rocks

Pegmatite dykes used for zircon dating are well exposed in the upper section of Odivelas III at the Ventoso quarry (N of Ferreira do Alentejo); they intersect mainly pyroxenic (porphyroid) gabbro layers, forming several sets of swarms (Fig. 3). Within a few millimeters from the contact with the pegmatites, the gabbros show metassomatic alteration of olivine (Fo_{74}) and orthopyroxene ($X_{Mg} = 0.78$), to serpentine ± magnetite ± actinolite + chlorite and actinolite + chlorite ± talc assemblages. Intercummulus clinopyroxene ($En_{44}Wo_{44}$) is rimmed by Mg-hornblende compositionally similar to the hornblendes in the pegmatites. Plagioclase displays strong compositional gradients with An_{62} fresh cores bordered by symplectic intergrowths of high Ca-plagioclase (An_{88}) ± quartz ± chlorite and phlogopite. Such textural and compositional features reflect introduction of K and preferential dissolution of Na in the original plagioclase grains; whereas the anorthitic component is readily re-precipitated as high-Ca plagioclase. The vermicular quartz, chlorite and phlogopite by-products of this process are incorporated in the symplectites, whereas the excess Na^+ is incorporated by the albite of the pegmatites. Single pyroxene thermobarometry [101] on selected (finer-grained, homogeneous) gabbroic rocks indicates primary crystallization conditions of BIC at 1154 ± 37 ºC and 4 ± 1.5 kb, whereas experimental data [102] suggests that the symplectitic formation processes occurred at *ca.* 700ºC. All these features support a late–magmatic origin for the pegmatite dykes, constraining their emplacement during the late-stages of magmatic evolution within the gabbroic system, coeval with the metasomatic haloes and Fe-Cu-Co sulphide deposition. The sulphide deposition probably resulted from admixing of sulphide-rich residual gabbroic melts with fluid-rich magmas similar to those that formed the late pegmatites [93, 97].

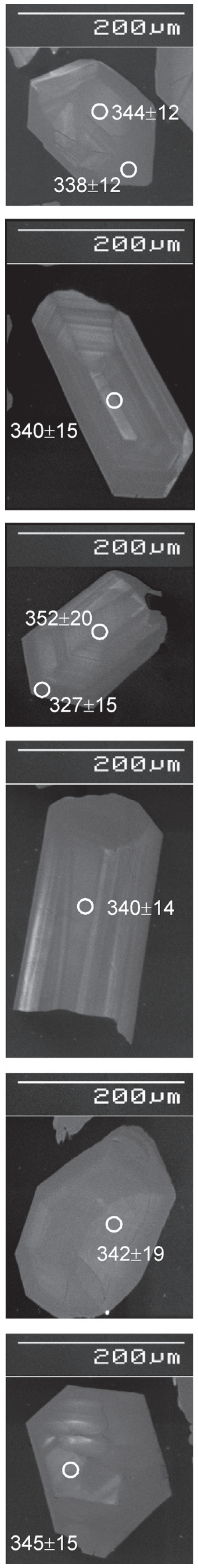


Fig. 4: CL images of the examined zircon grains from CVD5 sample.

Detailed petrographic / mineralogical analyses were performed on the pegmatite sample (CVD5) from which zircons were extracted. The pegmatite is essentially composed of an interlocking framework of euhedral Mg-horneblende (0.61 ≤ Al^t ≤ 1.19; 0.70 ≤ #Mg ≤ 0.78) and subhedral albite (An_{4-6}). Amphibole-plagioclase geothermometry [103] indicates equilibration temperatures from 680 ºC to 500ºC. Amphibole often includes corroded grains of titanite (Al : Fe ≈ 1.0) and, occasionally, can be seen nucleating on clinopyroxene cores of iron-rich augitic composition ($Fs_{32}En_{40}Wo_{28}$). Prehnite and calcite, both fill small late-stage vugs and micro-fractures. Microscopic observation suggests that zircons are mostly sited within the early formed amphibole, accompanying titanite.

4.1.2. Sample preparation and analytical techniques

Approximately 10 kg of pegmatite were collected at fresh outcrops (sample CVD5) and crushed with a Retsch jaw-crusher. Rock fragments were sieved and ground in a roller-crusher in order to maximise the amount of the size-fraction $106\mu m < x < 150\ \mu m$. This granulometric fraction was subsequently washed in compressed air stirred water-columns and oven dried. Sodium heteropolytungstate (LST Fastfloat) and di-iodomethane, coupled with isodynamic magnetic separation, were used to obtain a zircon-rich concentrate that was then purified by hand picking through a binocular lens.

Zircon U/Pb isotopic data were obtained from the Australian National University SHRIMP I instrument, using a ~30 μm diameter spot. Calibration methods and analytical procedures are those described in [104, 105]. $^{206}Pb/^{238}U$ ratios have an error component (typically) 1.5% to

Table 2: Summary of ion microprobe U-Pb results for zircon grains from sample CVD5, Ventoso quarry (N of Ferreira do Alentejo) – Layered Gabbroic Sequence

Grain Spot	Grain Type	U ppm	Th ppm	Th/U	Pb* ppm	$^{204}Pb/^{206}Pb$	$^{207}Pb/^{206}Pb$	±	$^{206}Pb/^{238}U$	±	$^{206}Pb/^{238}U$ Age (Ma)	±
1.1	c osc bip	1284	1084	0.84	80	0.000001	0.0536	0.0008	0.0548	0.0020	344	12
1.2	e osc bip	2632	3210	1.22	175	0.000025	0.0536	0.0006	0.0538	0.0019	338	12
2.1	c osc p	874	639	0.73	52	0.000011	0.0548	0.0008	0.0542	0.0024	340	15
3.1	c osc eq	1032	754	0.73	64	0.000001	0.0535	0.0013	0.0561	0.0034	352	20
3.2	e osc eq	1829	1255	0.69	104	0.000056	0.0519	0.0010	0.0520	0.0024	327	15
4.1	c osc p f	996	715	0.72	60	0.000001	0.0543	0.0007	0.0542	0.0023	340	14
5.1	c anh f	2427	1920	0.79	150	0.000097	0.0543	0.0009	0.0555	0.0029	348	18
6.1	c osc eq	1210	698	0.58	70	0.000040	0.0516	0.0015	0.0545	0.0030	342	19
7.1	c osc anh	452	208	0.46	26	0.000033	0.0547	0.0015	0.0545	0.0033	342	20
8.1	m anh	442	213	0.48	26	0.000571	0.0479	0.0046	0.0570	0.0031	357	19
9.1	c osc bip	814	406	0.50	47	0.000095	0.0525	0.0009	0.0550	0.0025	345	15

p = prismatic; **eq** = equant; **bip** = bipyramidal; **c** = core; **m** = middle; **osc** = fine-scale zoning; **anh** = anhedral; **f** = fragment
Uncertainties given at the one sigma level. Correction for common Pb made on the basis of extrapolation to concordia along a mixing line with common Pb, following Tera & Wasserburg (1972).

2.0%, from calibration measurements using the standard zircons. This error was added into each analysis and not to the final calculated data. U abundance was calibrated against 238 ppm U (< ±10 %) fragments of the single crystal SL-13 standard and Pb/U was calibrated against the multi-crystal standard AS57 (1100 Ma, [106]). All errors take into account non-linear fluctuations in ion counting rates beyond that expected from counting statistics (e.g. [104]). Age calculations were performed using the Ludwig (1998) ISOPLOT/Ex program.

4.1.3. Results

SHRIMP data are summarized in Table 2 and displayed on the $^{207}Pb/^{206}Pb$ – $^{238}U/^{206}Pb$ Tera-Wasserburg plot of Figure 5. The studied pegmatite sample (CVD5) yielded 250 - 300 μm long, mostly euhedral, prismatic to stubby zircons. In cathodoluminescence (CL) images, these display predominantly fine-scale oscillatory-zoned patterns typical of magmatic zircons. The studied zircons also display diffuse, CL-bright internal or thin-rimmed domains (Fig. 4) similar to those reported by Pidgeon *et al.* [107]. Ion-microprobe analysis produced relatively high Th/U ratios (0.46 to 1.22), and variable U (442 to 2632 ppm) and Th (208 to 3210 ppm) contents, both increasing from core to rim zones of the analysed zircons. The data suggest a roughly inverse correlation between U, Th contents and $^{238}U/^{206}Pb$ ages with the higher U analysed border domains giving younger ages (Fig. 4). However, all analysis sites yielded dates indistinguishable from each other, with a weighted mean $^{238}U/^{206}Pb$ date of 342±9 Ma (95% confidence level, MSWD = 0.25). The 342±9 Ma age is marginally younger than the ID-TIMS U-Pb ages of 350 ± 4 and 352 ± 4 Ma reported by Pin *et al.* [13] for the gabbroic facies hosting the pegmatites. Given the observed spread in ages related to textural and compositional features in the zircons, 342±9 Ma might be a slight underestimate (but still within error of) of the true age of the zircons and hence the pegmatite formation.

5. Discussion: geodynamic consequences

The new U-Pb SHRIMP data provide important constraints on the late-stage evolution of the Layered Gabbroic Sequence of BIC, and the average 342±9 Ma age should date the development of water-rich differentiated/contaminated magmas represented by the (amphibole-) pegmatite intrusions into the main gabbroic series. In this context, the close agreement with the available $^{40}Ar/^{39}Ar$ ages for rock-forming amphiboles belonging to BIC, BAOC (see section 3) and other geological formations of the OMZ southern border may be taken as evidence for a moderate to rapid regional crustal uplift episode at *ca.* 340±5 Ma, *i.e.* approximately 10 Ma after the emplacement of the earlier gabbroic series of BIC. This episode of uplift, causing significant modifications in the crustal heat flow and influencing decisively the mechanical properties of crustal rocks in the Évora-Beja-Aracena Domain, may also represent the ultimate cause for differences in strain accommodation during the second phase of Variscan deformation (D_2). Therefore, the 340±5 Ma age corresponds to a good estimation for the D_{2a} – D_{2b} transition.

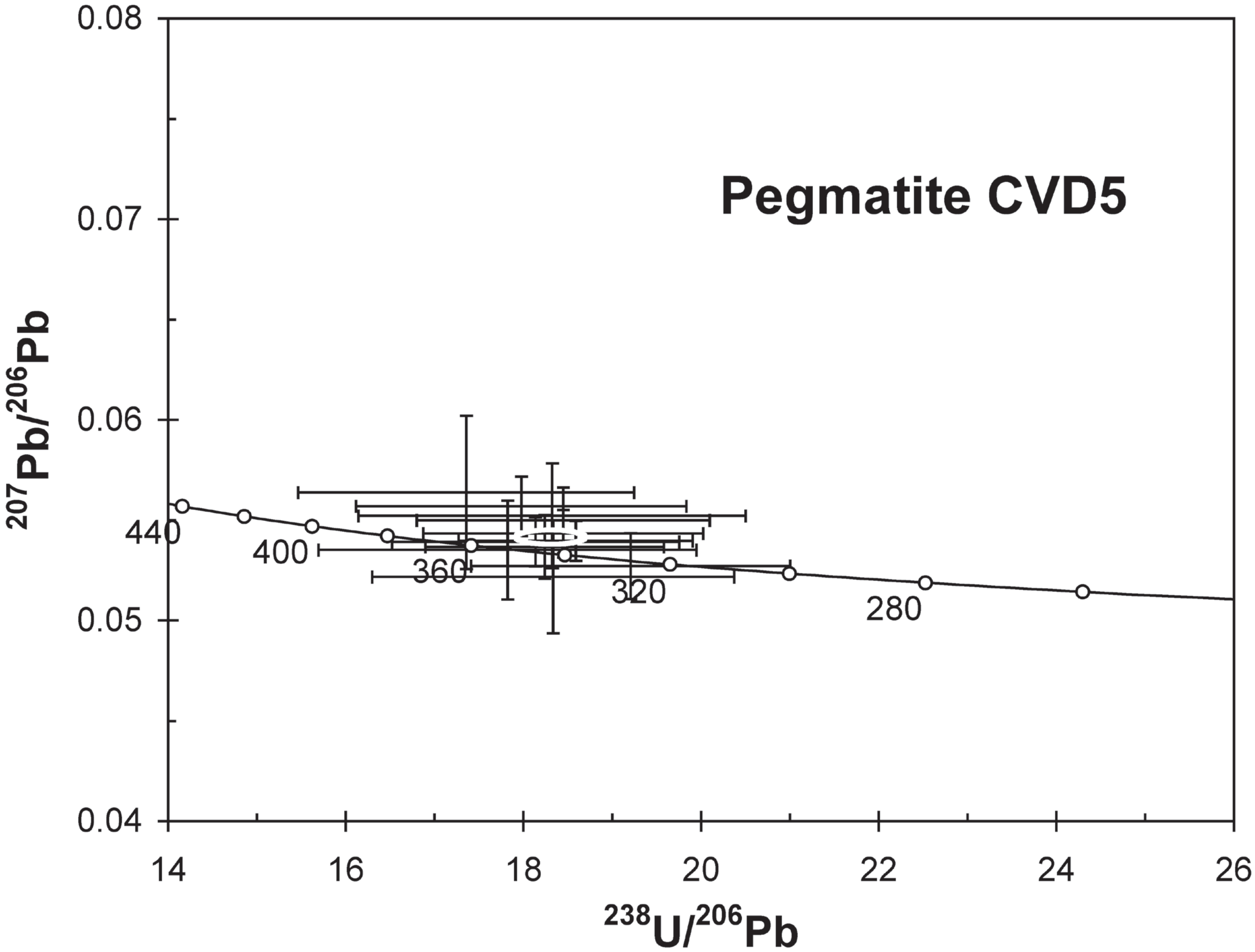


Fig. 5: Tera-Wasserburg plot ($^{207}Pb/^{206}Pb$ versus $^{238}U/^{206}Pb$) for zircons from the pegmatite of the Ventosa quarry (sample CVD5).

5.1. Late- and post-collision magmatism

The SW Iberian Variscides evolution can be reassessed using the geochronological data for the BIC Layered Gabbroic Sequence within the general data set available for the OMZ-SPZ boundary, briefly reviewed in sections 3 and 4. The schematic chrono-stratigraphic chart presented in Fig. 6 intends to illustrate the main features emerging when plotting the data set roughly fom SW to NE and making also use of the existing bio-stratigraphic and litho-stratigraphic constraints for key-formations in the Pulo do Lobo Terrane and IPB. The most striking feature is the relatively narrow interval of *ca*. 355 Ma to *ca*. 345 Ma for igneous activity on both sides of the OMZ-SPZ boundary (Fig. 6), clearly postdating the earliest record of 364±12 Ma orogenic magmatism triggered by subduction and the eclogite/ophiolite nappe emplacement (at *ca*. 370 - 360 Ma) in OMZ (see section 3). A broadly similar geological record occurs in the Spanish counterpart of the OMZ-SPZ boundary. Any differences can be explained by geometry of the plate boundary [61]. From this, it seems plausible to conclude that the time span ranging from *ca*. 355 to *ca*. 345 Ma should correspond to the early stages of the OMZ-SPZ oblique collision, dating also the main events ascribable to D_{2a} (presumably starting at *ca*. 360 Ma). The development of the Layered Gabbroic Sequence of BIC should have occurred during this period.

In the OMZ southern border and adjoining Pulo do Lobo Terrane (Gil Márquez granodiorite) late-collisional magmatism started at 328±2 Ma [32]. The petrologic features clearly indicate that late-collision igneous rocks in these geological contexts resulted from different and more evolved magmas with magma mixing and strong to very strong crustal contamination. This geochronology continuum strongly suggests that the high thermal crustal regime should have persisted for a relatively long period of time, being to a great extent synchronous of the LP-HT Variscan metamorphic event recorded in the Évora-Beja-Aracena Domain (see section 3). In BIC, this magmatic episode is responsible for the development of the large gabbro-diorite envelop known as the Cuba-Alvito Complex, besides some older but minor parts of the Baleizão Porphyry Complex. Following this reasoning, the granitoids dated of *ca*. 320 Ma, like those included in the Évora Massif

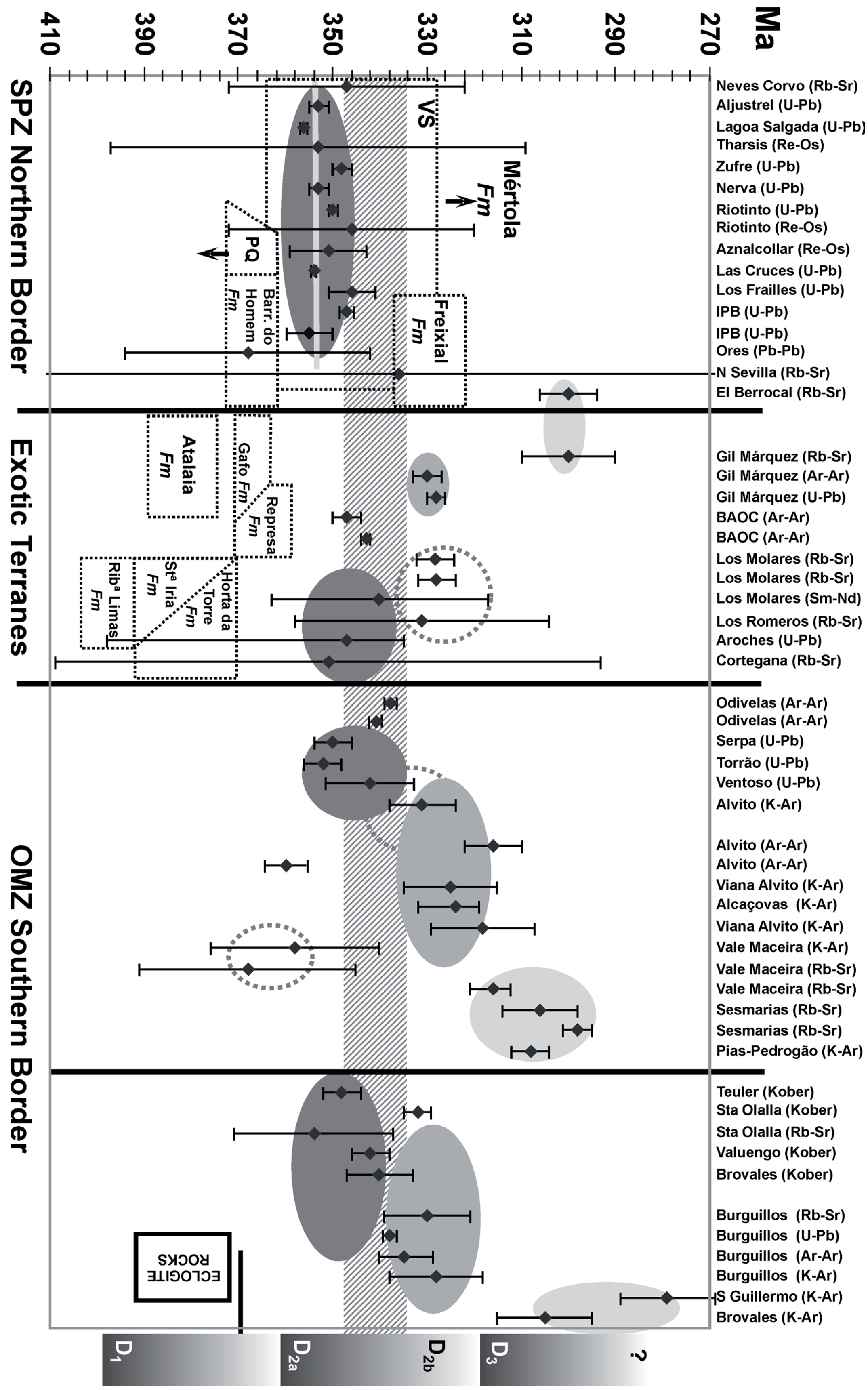


Fig. 6: Schematic chrono-stratigraphic chart for OMZ southern border, Exotic Terranes and SPZ northern border, using the available bio-stratigraphic and litho-stratigraphic data (for the Pulo do Lobo Terrane and IPB) and the most pertinent geochronology data for igneous rocks (as compiled on Table 1)

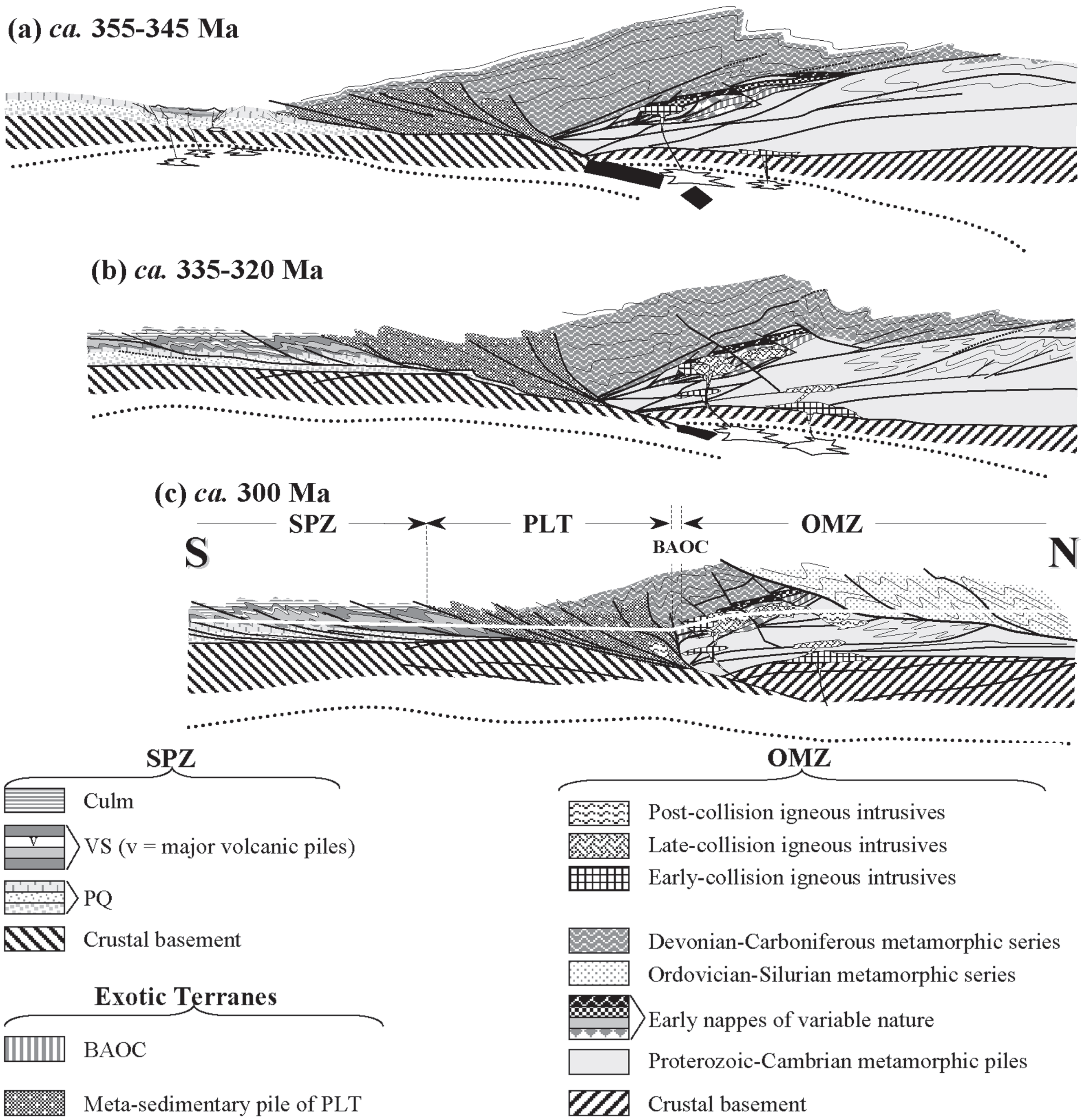


Fig. 7: Simplified and interpretative section across the SPZ-OMZ boundary, laterally projecting the major geological formations and not considering the displacements caused by late shear zones and strike-slip fault zones; post-collision granitoids intruding the Pulo do Lobo Terrane do not outcrop but their possible existence is documented by geophysical data (Almeida *et al.*, 2005): **(a)** early stages of continental oblique collision at *ca*. 355-345 Ma; **(b)** intermediate stages of collision, after the major uplift episode; **(c)** late stages of continental oblique collision at *ca*. 300 Ma; present day surface is roughly represented by the thick white line .

(*e.g*. Montemor-o-Novo), should represent the final stages of this late orogenic high-T regime, being emplaced and crystallized after the aforementioned (moderate to) rapid crustal uplift episode placed at *ca*. 340±5 Ma. Accordingly, the 320 Ma age should roughly represent the late stages of $D_{2b,}$ being consistent with the generalized isotopic resetting in the OMZ southern border.

The second magmatic event is recorded by the crystallization of late, essentially undeformed granite bodies during the final stages of D_3 in OMZ, thus including the main epizonal intrusive units of the Baleizão Porphyry Complex that are part of BIC. This post-collision igneous activity, dated *ca*. 300 Ma and locally inducing isotopic resetting of biotite represents, therefore, the late stages of the OMZ-SPZ oblique collision.

5.2. An inferred geodynamic evolution

Figure 7 presents a reconstruction of the geodynamic evolution of the SW Iberian Variscides. This provides a mean to visualize the processes that, from the early to the late stages of the OMZ-SPZ oblique collision, may have favoured the heat supply necessary for the widespread magmatism and the HT metamorphism in those geotectonic units and in the Exotic Terranes placed between them. The crustal architecture and, consequently, the structural and metamorphic features across the plate boundary and in the Exotic Terranes trapped there are quite different. The Pulo do Lobo Terrane is commonly interpreted as an oceanic accretionary wedge developed from Lower-Middle Devonian [9, 12]. The thrust belt progression in the IPB proceeds by continuous foreland accretion until, at least, Upper Visean (the age of the older formations in the flysch sequence) above a major cryptic *décollement* correlated with the early-late Devonian transition [58, 59]. A stage of major internal reactivation and thickening in large parts of the accretionary wedge appears to have occurred at around 350-340 Ma, possibly resulting from changes in the foreland mechanics as suggested by the results of numerical modeling of similar geodynamic frameworks [109-110]. In the IPB, the extensive imbricate thrust belt with predominantly localised deformation along thrusts and significant total shortening controls the progress of sedimentary processes, behaving as a typical orogenic pro-wedge system. Therefore, the sedimentary/structural evolution recorded by the Pulo do Lobo Terrane and the IPB, together with their typical low-grade metamorphic zoning (decreasing towards SW, [28, 41-43, 46]) can be directly related to the advance of the upper-plate (OMZ) deformation. All these data strongly suggest the formation of a kinematically coupled system including a pro-wedge (IPB), an axial zone (Pulo do Lobo Terrane) and a retro-wedge (Évora-Beja-Aracena Domain) operating during the Carboniferous sinistral transpressive collision.

As proposed for other Variscan sectors, namely for the Saxo-Thuringian Zone in the Central European Variscides [108], the evolution of the Évora-Beja-Aracena retro-wedge should be intimately related to two main collision episodes: (1) beginning of continental subduction/collision and subsequent lithospheric delamination (thermal erosion) induced long-term magma underplating as well as the uplift of parts of the upper plate; and (2) mechanically controlled changes during the foreland accretion in the IPB, leading to a significant increase of the angle (wedge taper) between the topographic slope and the basal detachment developed on top of the OMZ crustal basement (presumably of granulitic nature, [111, 112]). This would explain the formation of a narrow internal thrust stack (that can be interpreted as a retro-shear belt) in the retro-wedge domain during the early stages of the OMZ-SPZ oblique collision (*ca*. 355-345 Ma), followed by the development of a subsequent fold belt during D2b. Consequently, the fact that most of the crustal shortening operated in the Évora-Beja-Aracena Domain is localised in the internal thrust stack would not be coincidental; subsequent (heterogeneous) crustal shortening was accommodated in the much wider external belt. In these circumstances, the earlier internal thrusting seems to be strongly restrained by the proximity to the crustal domain that accommodated the uplift of the forearc/arc crust during the magma underplating. Conversely, the younger external belt mostly involves the outer meta-sedimentary sequences and, apparently, does not have any major mechanical discontinuity; note that this regime of strain accommodation was also favoured by the fact that late deformation progression has occurred under relatively high averaged temperature conditions. Finally, it should be emphasized that the wedge taper increase in retro-wedge systems is often taken as an indication of relative weakening of basal detachments (with respect to their hanging walls) through time ([108] and references therein); this feature is probably related to radiogenic heating during crustal thickening, as suggested by the increasing abundance of crustal post-collision melts (see below).

5.3. Thermal regimes

An important feature of the Évora-Beja-Aracena Domain is the persistent high thermal gradient, which should have influenced the deformation style and the large amount of upper-plate shortening during the OMZ-SPZ oblique collision. This feature is compatible with the presence of a thermally weakened retro-wedge system, whose internal part went through a complex P-T path during the pre-collision evolution from *ca*. 400-390 Ma to *ca*. 370 Ma, involving the opening and closure of a back-arc basin, BAOC emplacement via antithetic obduction, and development of HP-LT metamorphic events [68, 69, 113]. Therefore, significant variation in time of the subduction rate and plane dip should have occurred along the OMZ-SPZ boundary, preceding the collision stages and determining the evolution of a thermal regime that will affect mainly the Évora-Beja-Aracena Domain. Initially, a steep (N-directed) subduction zone might have promoted the rollback of the lower plate hinge, thus favouring the back-arc extension; in this situation, the hinge retreat rate is expected to be higher than the convergence rate. Subsequent gradual decrease of the subduction angle and relative increase of the convergence rate, should have led to compression of the back-arc domain and, further, to the arc system overthrusting, decisively contributing to the Évora-Beja-Aracena Domain thickening [69, 70]. Later on, during the dominantly transpressive collision stages, as the continental crust (SPZ) began to subduct and the OMZ sourthern border was compressed and (re-)stacked, a different thermal regime developed.

In the OMZ, the early orogenic magmatism triggered by subduction is dated at 364±12 Ma (see section 3). However, in the Évora-Beja-Aracena retro-wedge all the processes appear to have proceeded at high temperature conditions from ca. 355 Ma onwards, allowing the progression of LP-HT metamorphism and multistage magmatism. A similar record for the igneous activity is found in the Pulo do Lobo Terrane. In the IPB, significant magmatism occurred between ca. 355 Ma and ca. 345 Ma. The origin of this late orogenic (early to late collision stages) magmatism and HT regime is controversial, although apparently characteristic of many segments of the European Variscides [e.g., 114].

From the scenario depicted in Fig. 7a, it might be expected that the localised (but strong) SPZ thinning, favoured by subduction blocking and subsequent slab break-off, caused the underplating of basaltic magmas due to asthenospheric upwelling. The underplating of these magmas at, and within, the base of the SPZ crust and the combination of conductive (hot mantle) and advective (basalt magma) heat transport would promote the igneous activity typical of the IPB. Concurrently, the slab break-off at the subduction zone allows the local incursion of the asthenospheric mantle; subsequent lithospheric thermal erosion may have led to detachment of the dense, relatively cold, lithospheric mantle root [114], and perhaps even the mafic lower crust of the Évora-Beja-Aracena Domain. The resulting transient transtensional tectonic regime coupled with the juxtaposition of hot mantle and subducted rock, and sudden added buoyancy (as the dense down-dragging force of the orogenic root was removed), favoured melting of metasomatised (low melting-T) components of the mantle. These processes might have caused the early stages of magmatism in the retro-wedge system, primarily traced by the rise of primitive melts like those found in the earlier series of the Layered Gabbroic Sequence of BIC.

The stacking of high heat-producing upper-crustal lithologies over a period of ca. 20 Ma (from ca. 360 Ma to 340 Ma, during D2a) would provide the (radiogenic) heat needed to sustain the LP-HT metamorphism and, alongside with residual thermal effects related to the early magmatic event, the initial steps of the late-collision magmatism, involving mixing of mantle-derived and crust-derived melts. Later on, in the course of the moderate to rapid crustal uplift episode (at ca. 340±5 Ma), the rising of the brittle-ductile transition would result in a significant reduction of crustal strength, since the mechanisms responsible for the generated heat anomalies occurred much more faster than dissipation of heat by conduction and fluid convection did. Consequently, an increase of crustal contamination of mafic magma chambers and the formation of anatectic magmas would be gradually favoured. The emplacement of the remaining late-collision granitoids and post-collision granites (immediately above that mechanical transition) would occur, therefore, at progressively shallower depths.

6. Conclusions

Previous work [*e.g.*, 11, 91] shows that the Beja Igneous Complex is a major geological feature of the Évora-Beja-Aracena belt, being also an outstanding mark of the Palaeozoic igneous activity in SW Iberian Variscides. It consists of three main units that were developed during different stages of the OMZ-SPZ oblique collision, namely: (1) the Layered Gabbroic Sequence in the course of the early stages of collision magmatism, from *ca*. 355 Ma to *ca*. 345 Ma; (2) the Cuba-Alvito (gabbro-diorite) Complex throughout the late-collision magmatic events, from *ca*. 335-330 Ma to *ca*. 320 Ma ; and (3) the Baleizão Porphyry Complex mostly during the post-collision magmatic events at *ca*. 300 Ma.

The new SHRIMP U-Pb age of 342±9 Ma reported here for amphibole-bearing pegmatite dykes cutting the layered gabbros is interpreted as dating the development of late-stage fluid-rich melts in the Layered Gabbroic Sequence, synchronous with Fe-Cu-Co sulphide deposition. This result suggests a hydrous contamination of the late diferentiated melts in BIC, coeval with amphibolite-greenschist facies retrogradation of BAOC and other OMZ southern border geological formations, at *ca*. 340±5 Ma, immediately prior or during rapid regional crustal uplift. It also gives an estimation for the transition between the D2a – D2b phases of Variscan deformation (Fig. 6). Within this context, the new geochronological results, together with the already published data also supports a sustained, long-term regional high heat flow (≈ 30-40 Ma) within the southern OMZ border.

The complex wedge system within the SW Iberian Variscides developed during Carboniferous collision of two continental fragments, the OMZ upper plate to the north and the SPZ passive margin in the lower plate. The Évora-Beja-Aracena Domain, located in the upper plate above the N-dipping subduction zone, is re-interpreted as a retro-wedge domain that was kinematically coupled to the SPZ pro-wedge and subduction system. The two major steps in retro-wedge growth are linked to (1) the onset of collision, causing the upper plate uplift and (2) a widespread late-orogenic stage of wedge thickening. The early stages of magmatism in the retro-wedge are related to the first step of its growth, as a result of the local incursion of asthenospheric mantle melts allowed by slab break-off; at the same time, localized (but strong) thinning of the lower plate favoured the underplating of basaltic magmas due to asthenospheric upwelling. The heat needed to sustain the regional LP-HT metamorphism and the subsequent magmatic events in the retro-wedge domain was provided through the stacking of radioactive upper-crustal lithologies, reinforced and sustained for a longer period due to the anomalous heat flow related to the (moderate to) rapid crustal uplift episode. The retro-wedge accumulated mostly diffuse shortening versus the shortening by imbrication in the lower plate. Mass advection and orogenic architecture were strongly affected by asymmetric removal towards the lower-part foreland and by transient mechanical properties of the wedge system.

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