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Transcurrent displacement of the Cadomian magmatic arc

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Abstract

The Ossa-Morena Zone in Iberia, and equivalent terranes in northern France and Central Europe, are thought to be paleogeographically linked to the West African Craton at Ediacaran to early Paleozoic times. Evidence is mainly based on metasedimentary rock detrital zircon age spectra, characterized by two dominant populations of Paleoproterozoic and Cryogenian-Ediacaran ages, with a systematic lack of late Stenian – early Tonian zircon grains. We report here U-Pb-Hf results on detrital zircon grains from six Ordovician-Devonian metasedimentary rocks from the Ossa-Morena Zone. In addition to the Cryogenian-Ediacaran and Paleoproterozoic populations already recognized in previous studies on uppermost Ediacaran and lower Cambrian rocks, the samples ubiquitously show a late Stenian – early Tonian population centered at ≈ 1 Ga and representing ≈ 20 % of the concordant dates. The εHf *versus* age plot of the studied samples mainly depicts two vertical arrays corresponding to the Cryogenian-Ediacaran and Stenian-Tonian detrital zircon populations, which spread from εHf values of ≈ 10 down to ≈ -20. The detrital zircon age distribution and Hf isotope signature of the studied rocks point to the Sahara metacraton as the most plausible sediment source. The eastward translation of the continental ribbon represented by the Ossa-Morena Zone and equivalent domains of the Variscides from an original position close to the West African Craton to an Ordovician-Devonian location close to the Sahara metacraton probably occurred at latest Ediacaran – earliest Cambrian times in a dextral strike-slip tectonic setting that post-dated Pan-African collision and Cadomian subduction.

Key-words: Detrital zircon grains, U-Pb geochronology, Hf isotope signature, Ossa-Morena Zone, Sahara metacraton, West African Craton.

1. Introduction

Paleogeographic and paleotectonic reconstruction of pre-Mesozoic orogens is subjected to a number of uncertainties regarding the number of intervening plates/microplates, as well as the extent and age of subducted oceanic domains in-between the amalgamated continental pieces. A first step to addressing this classic issue is producing good quality geological maps at the scale of the orogen, where the location and lateral continuity of ophiolitic and/or high-pressure belts must be carefully established. Obviously, different transects of the same orogen often offer very different perspectives on the number of putative oceanic and continental domains implied, so correlation at the scale of the whole orogen becomes controversial. The combination of field geology with other geological, paleontological, geophysical and geochronological data can sometimes shed some light on the orogenic picture, though consensus is rarely reached. Among these new data, the systematic U-Pb dating of detrital zircon populations has become a powerful tool to elucidate the derivation of the different continental terranes welded during collisional and pre-collisional orogenic evolution (e.g., Miller et al., 2006; Wang et al., 2007; Gehrels, 2012 and 2014).

The Variscides represent a paradigmatic example of an orogen where uncertainty remains regarding the number and extent of oceanic domains subducted at Late Paleozoic time, before assembly of the supercontinent Pangea. Interpretations of the paleogeographic/paleotectonic evolution vary from those considering only a single and wide oceanic realm (Rheic Ocean) between two main continental plates (Laurussia and Gondwana; e.g., Robardet, 2002; Kroner and Romer, 2013; Romer and Kroner, 2019; Stephan et al., 2019a and b), to the existence of a number of different oceanic and continental domains between the two main continental blocks (e.g., Matte, 2001; Stampfli et al., 2013; Franke, 2000; Franke et al., 2017 and 2019). The variety of models results from overemphasis on one kind of data over others. Proposals relying only on faunal differences (e.g., Robardet, 2002; Robardet and Gutiérrez-Marco, 2004), geochemical data (e.g., Romer and Kroner, 2019), or detrital zircon data (Stephan et al., 2019a and b) favour a single and wide oceanic domain. In contrast, those models based on a multi-method approach support that at least two oceanic domains of varying importance were involved in Variscan orogenesis (e.g., Franke, 2000; Matte, 2001; Simancas et al., 2002 and 2009).

Independently of the number of oceanic domains involved, all available paleogeographic reconstructions consider that the Variscides resulted from the amalgamation of a number of Gondwana-derived terranes, whose primary position at the periphery of the supercontinent is well established in some cases, but unclear in others. Studies on detrital zircon populations from Ediacaran to Paleozoic rocks have provided key paleogeographic evidence (e.g., Linnemann et al., 2008; Pereira et al., 2011 and 2012a, Pérez-Cáceres et al., 2017; Collet et al., 2020). Avalonian terranes are characterized by detrital zircon spectra with a distinctive multi-peak distribution between 1 and 1.8 Ga, as well as a zircon-forming event at Late Ordovician-Silurian time (Hamilton and Murphy, 2004; Braid et al., 2012; Pérez-Cáceres et al., 2017; Pereira et al., 2017; Herbosch et al., 2020; Sorger et al., 2020). These terranes were attached to the western segment of Northern Gondwana at Ediacaran time, but during Rheic Ocean in the Early Paleozoic, they drifted from Gondwana and collided with Laurentia during the Silurian (e.g., Nance et al., 2002). The remaining Gondwana-derived continental pieces involved in the Variscan orogenesis would have never drifted far from Gondwana, having evolved either as ribbon continental pieces separated by minor oceanic realms from the mainland, or as aborted rifts and passive margins attached to it. These peri-Gondwanan terranes have detrital zircon spectra dominated by a prominent Ediacaran peak attributed to subduction-related Cadomian/Avalonian arc magmatism developed all along the northern Gondwana margin (e.g., Gutiérrez-Alonso et al., 2003; Linnemann et al., 2014). Minor peaks at 2 Ga are always present, while a 1 Ga population is only present in some zones (Central Iberian Zone, West Asturian-Leonese Zone, and Cantabrian Zone in Iberia; e.g., Gutiérrez-Alonso et al., 2015) and has been attributed to eastern Arabian-Nubian shield or Sahara metacraton zircon sources (Bea et al., 2010; Díez-Fernández et al., 2010; Meinhold et al., 2013). The absence of a 1 Ga detrital zircon population of in the Ossa-Morena Zone (OMZ) and equivalent terranes in northern Brittany and Central Europe (Saxo-Thuringian Zone) has served to locate the OMZ close to the West African Craton (WAC) in Ediacaran to early Paleozoic times, west of the other Variscan zones (Fernández-Suárez et al., 2002; Linnemann et al., 2008; Pereira et al., 2011 and 2012a). Nevertheless, there is no consensus on this paleogeographic attribution. Cambeses et al. (2017) used the Sm-Nd isotope signature of Ediacaran to Cambro-Ordovician rocks to propose that the OMZ was located close to the Tuareg Shield in Cambrian-Ordovician times, in a more easterly position than previous studies have claimed. In contrast, López-Guijarro et al. (2008), Fuenlabrada et al. (2020) and Rojo-Pérez et al. (2021) also utilised Sm-Nd data to propose a western OMZ paleoposition, close to the WAC.

This paper aims to elucidate the paleogeographic and paleotectonic meaning of the OMZ, based on the previous geological knowledge, new U-Pb ages and the new Hf isotope signatures of detrital zircon grains from its Ordovician-Devonian sequence. In particular, our study seeks evidence of a 1Ga zircon population in Paleozoic OMZ samples in an effort to evaluate the Paleozoic location of this piece of the Iberian Massif, and its counterparts in other regions of the Variscides.

2. Geological setting

The Variscan Orogen extends from Central and Western Europe to Northwestern Africa, with the northern foreland (Laurussia) cropping out in England, Belgium and Germany. The southern one (Gondwana) is mostly obscured due to Alpine reworking, except in Morocco where it is exposed in the Anti-Atlas ranges (Fig. 1A). The Iberian Massif (Central-Western Spain and Portugal) constitutes the largest outcrop of the Variscan orogen, having been traditionally divided into a number of zones with different stratigraphic, tectonometamorphic, and magmatic features (e.g., Julivert et al., 1974; Simancas, 2019 and references therein). This classic division (Fig. 1B) comprises the Cantabrian (CZ), Western Asturian-Leonese (WALZ), Central Iberian (CIZ), Galicia - Tras-Os-Montes (GTMZ), Ossa-Morena (OMZ) and South Portuguese (SPZ) Zones. The correlation of these zones with other European Variscan massifs through the Cantabrian arc usually considers the CIZ, WALZ and CZ to represent the Gondwanan margin, while the SPZ represents the Avalonian foreland with counterparts in Southern England, Belgium and Germany (Rheno-Hercynian Zone) (e.g., Matte, 1991; Franke, 2000; Simancas et al., 2005; Murphy et al., 2016). The OMZ extends through the Cantabrian arc in northernmost Brittany, Belgium, Germany and Czech Republic, where it is referred to as the Saxo-Thuringian Zone.

The OMZ is characterized by particular stratigraphic and magmatic features, which have served, despite controversies, to elucidate its paleogeographic and paleotectonic evolution. In the stratigraphic record, both Ediacaran and Lower Paleozoic sequences in the OMZ are distinctive with respect to other Variscan Zones in the Iberian Massif. The OMZ is also characterized by an abundance of upper Ediacaran and lower-middle Cambrian to Ordovician volcanic and plutonic rocks (e.g., Simancas et al., 2004; Sarrionandia et al., 2020).

On a broad scale, the OMZ can be considered as a continental segment deformed between two Variscan suture-type contacts, namely the contacts with the CIZ (Burg et al., 1981; Azor et al., 1994 and 2019; Simancas et al., 2001; Gómez-Pugnaire et al., 2003; López Sánchez-Vizcaíno et al., 2003) and SPZ (Bard, 1977; Crespo-Blanc and Orozco, 1991; Fonseca and Ribeiro, 1993; Quesada et al., 1994; Castro et al., 1996; Azor et al., 2008 and 2019; Pérez-Cáceres et al., 2015 and 2020). These two tectonic boundaries attest to oceanic and/or continental subduction and subsequent collision at Devonian-Carboniferous times, which deformed the whole OMZ and SPZ interiors (e.g., Simancas et al., 2013). The OMZ/SPZ boundary is unanimously considered as the Rheic Ocean suture, though the contact is marked by a Carboniferous MORB-featured amphibolitic unit that postdates ocean consumption (Azor et al., 2008) and makes the suture appear as cryptic (Pérez-Cáceres et al., 2015). On the contrary, the OMZ/CIZ boundary, namely Badajoz-Córdoba Shear Zone (Figs. 1B and 2), has been the subject of an extensive debate regarding its tectonic significance. Some authors have considered this boundary as a Variscan intracontinental shear zone and, hence, have claimed paleogeographic continuity of the CIZ and OMZ at Early Paleozoic times along the Gondwana margin (Abalos et al., 1991; Quesada, 1991; Robardet, 2002; Ribeiro et al., 2007). Others have interpreted this contact as representing a second-order Variscan suture, where the ophiolite-bearing allochthonous units of NW Iberia would be rooted (Matte, 1991; Simancas et al., 2002). The available petrological, geochronological and geochemical data have shown that the entire metamorphic evolution, including an HP event, is Variscan in age (Ordóñez-Casado, 1998; Pereira et al., 2010a and 2012b; Abati et al., 2018). Furthermore, some of the MORB-featured mafic rocks have Cambrian to Early Ordovician ages (Ordóñez-Casado, 1998; Gómez-Pugnaire et al., 2003), thus confirming that the CIZ/OMZ boundary records the closure of a minor oceanic realm between Gondwana and a ribbon continental segment (Azor et al., 2019).

2.1. Stratigraphic framework

The OMZ shows an almost continuous stratigraphic record from late Ediacaran to late Carboniferous times (Figs. 2 and 3), with several unconformities of diverse importance and meaning (e.g., Quesada et al., 1990; Azor et al., 2004; Robardet and Gutiérrez-Marco, 2004). The whole sequence is dominated by siliciclastic rocks, with some carbonate-dominated periods (Early Cambrian and Mississippian); volcanic-plutonic rocks with variable geochemical signatures are common at latest Ediacaran, Cambrian, early Ordovician, and Mississippian times (e.g., Sánchez Carretero et al., 1990; Ordóñez Casado, 1998; Galindo and Casquet, 2004). From a tectonic point of view, the pre-orogenic sequence is mainly affected by km-scale SW-vergent recumbent folds, with an associated axial-plane foliation; a second folding event gave way to NW-SE striking upright folds with an associated crenulation cleavage (e.g., Expósito, 2000; Azor et al., 2019). This second event is the only ductile deformation observed in the syn-orogenic sequence. Left-lateral strike-slip faults occurred during a late Variscan deformational event (e.g., Pérez-Cáceres et al., 2016).

The oldest outcropping rocks belong to the so-called Serie Negra Group (Black Series), which consists of slates, schists and greywackes, with amphibolite and black quartzite intercalations (e.g., Quesada et al., 1990) (Fig. 3). Based on the age of the youngest detrital zircon population in samples from the Serie Negra Group (Schäfer et al., 1993; Fernández-Suárez et al., 2002; Linnemann et al., 2008; Pereira et al., 2011 and 2012a), a late Ediacaran maximum depositional age was established. This age is in agreement with the stratigraphic position of the Serie Negra Group (underlying paleontologically-dated Cambrian rocks), as well as with some radiometric ages obtained on the amphibolites (Sánchez-Lorda et al., 2016) and cross-cut relationships with Early Cambrian granitic rocks (Pereira et al., 2011; Sanchéz-García et al., 2014). A bimodal volcano-sedimentary sequence with abundant associated plutonic rocks (Malcocinado Formation) unconformably overlies the Serie Negra Group. Numerous radiometric ages on the volcanic and plutonic rocks associated with this formation yielded latest Ediacaran to earliest Cambrian ages (Oschner, 1993; Ordóñez-Casado, 1998; Salman, 2004; Sarrionandia et al., 2020). The Malcocinado Formation is traditionally interpreted as representing a supra-subduction zone / Cadomian volcanic arc developed all along the northern Gondwanan continental margin (Sánchez Carretero et al., 1990; Pin et al., 2002; Simancas et al., 2004).

The Cambrian sequence unconformably overlies the Serie Negra Group and the Malcocinado Formation (Fig. 3), attesting to the onset of a new Wilson cycle that gave way to the opening of the Rheic ocean. The lowermost Cambrian formation (Torreárboles Formation) is made up of arkosic sandstones with conglomerate and slate intercalations (Liñán, 1978). Onto the Torreárboles formation, a lower Cambrian succession of limestones with slate intercalations crops out (Pedroche formation; Liñán, 1974). A lower-middle Cambrian slate-sandstone dominated sequence overlies the Pedroche Formation. Finally, a thick sequence of middle-upper Cambrian slate and fine-grained sandstones with abundant felsic/mafic volcanic intercalations (Sánchez-García, 2001) witnesses the rifting stage that preceded Rheic ocean opening (Sánchez-García et al., 2003, 2010 and 2019; Chichorro et al., 2008; Palacios et al., 2021).

The entire Ordovician-Devonian succession is interpreted as a passive-margin sequence deposited on the Gonwana margin coevally with Rheic –and other minor– ocean expansion (e.g., Robardet and Gutiérrez-Marco, 1990). In the localities of Cerrón del Hornillo and El Valle, the Ordovician (upper Floian-Hirnantian) consists of a ≈500 m-thick sequence dominated by slates, siltstones and sandstones, with a 15-20 m-thick limestone intercalation of Katian age (e.g., Robardet, 1976; Robardet et al., 1998; Robardet and Gutiérrez-Marco, 2004) (Fig. 3); in other outcrops, the Ordovician sequence is entirely siliciclastic and reaches thicknesses of up to 1000 m. The Silurian comprises a 100-150 m-thick condensed sequence dominated by graptolite-bearing black slates, with some black chert, sandstone and limestone intercalations. Among these intercalations, the most important is the so-called Scyphocrinites Limestone (lower Pridoli), a 10-15 m-thick alternating black limestone and slate series that divides the Silurian sequence into a 120 m-thick lower part (lower graptolitic slates) and a 20 m-thick upper part (upper graptolitic slates). However, in some of the transects the Silurian does not intercalate any cabonatic level. The Devonian succession starts with black slates of Lochkovian age in stratigraphic continuity with Silurian slates. The remaining Lower Devonian pre-orogenic sequence is dominated by green to brown shales with siltstone intercalations and has been dated as Praguian-Emsian according to the abundant fossil content (e.g., Robardet et al., 1998; Robardet and Gutiérrez-Marco, 2004).

The syn-orogenic deposits of the OMZ start with marine greywackes and conglomerates concentrated in the so-called Terena Flysch syncline. Actually, the Terena Flysch comprises two different successions (Expósito, 2000): the Lower Devonian Terena sediments (Piçarra 1996; Piçarra et al. 1998) are the first record of syn-orogenic deposits; the upper Terena sediments, which unconformably lie over the lower Terena or directly overlie the pre-orogenic succession, are of Famennian to Visean age (Boogaard and Vazquez, 1981; Giese et al., 1994). In other outcrops there are only lower Carboniferous (Tournaisian-Visean) conglomerates, shales, greywackes and limestones, with volcanic intercalations (e.g., Azor et al., 2004).

2.2. Previous detrital zircon geochronological data

A number of studies have reported detrital zircon ages of late Ediacaran (Serie Negra and Malcocinado formations) and Cambrian rocks of the OMZ (Fernández-Suárez et al., 2002; Linnemann et al., 2008; Pereira et al., 2011 and 2012a). Despite the limited number of analyzed grains, these studies showed a dominant peak of Ediacaran-Cryogenian age (550-750 Ma), attributed to the Pan-African orogeny and/or the Cadomian arc magmatism, and minor peaks centered at 2 Ga, which are interpreted as material derived from the West African Craton (WAC). Furthermore, the absence of a late Stenian – early Tonian (≈ 1 Ga) detrital zircon population has been used to reinforce the western paleoposition of the OMZ, close to the WAC along the northern Gondwanan margin.

The detrital zircon ages of the Carboniferous syn-orogenic deposits have also been investigated (Pereira et al., 2012c and 2020; Dinis et al., 2018). The results show the presence of Early Carboniferous, Late Devonian, Ordovician, Cambrian, Neoproterozoic and Paleoproterozoic detrital zircon populations at variable percentages, which are compatible with the amalgamation and recycling of variably sourced grains during the Variscan collision.

3. Samples and methods

This study focuses on the OMZ Ordovician-Devonian sequence at two localities, namely the Cerrón del Hornillo and El Valle synclines (**Error! Reference source not found.** 2 and 3), where both sedimentary facies and faunistically-based stratigraphic ages are very well known (Robardet, 1976; Robardet et al., 1998; Robardet and Gutiérrez-Marco, 2004). The whole Ordovician-Devonian succession was sampled, though the Silurian rocks are monotonous black shales and ampelites with very minor fine-grained silt intercalations that did not yield detrital zircon grains. In contrast, Ordovician and Devonian rocks are mostly sandy levels with abundant detrital zircon grains. We analyzed 4 samples from the Cerrón del Hornillo syncline (CH2, CH3, and CH4 of Middle Ordovician age, and CH6 of Early Devonian age) and 2 samples from the El Valle syncline (EV1 and EV2 of Late Ordovician age; Figs. 2 and 3; Table 1).

Under the microscope, all of the studied samples are non-foliated quartz-rich sandstones. Quartz represents up to 90 % of the rock content, with opaque minerals being also present in all of the samples, often at interstitial positions together with phyllosilicate minerals. Quartz grains generally have angular shapes and variable diameters from 100 to 400 microns, with mean values of ≈ 200-250 microns. Feldspar, zircon, rutile and tourmaline have been recognized as accessory minerals.

For each sample site, 4-5 kg of rock were collected and processed at the laboratories of the University of Granada (Spain). A statistically significative amount of detrital zircon was separated from each sample using mechanical smashing in a jaw-crusher, sorting by sieving, manual panning, and, finally, handpicking. Cathodoluminescence (CL) images (Fig. 4) were taken on a Carl Zeiss SIGMA HD VP Field Emission SEM at the School of GeoSciences, University of Edinburgh (Scotland, United Kingdom), and a Mira3 FESEM instrument at the John de Laeter Centre (JdLC), Curtin University (Perth, Australia) in order to recognize zoning and alteration textures within the zircon grains. U-Th-Pb analyses were conducted using a Laser Ablation Inductively Coupled Plasma Mass Spectrometry (LA-ICPMS) at the GeoHistory Facility, JdLC. In order to have a representative number of analyses, some additional U-Th-Pb analyses were performed on small zircon grains of sample EV2 using a Secondary Ion Mass Spectrometry (SIMS) at the NERC Ion Microprobe Facility of the University of Edinburgh (UK) (Table 1). A detailed description of the analytical methods can be found in Appendix A.

Raw data were analyzed using Isoplot (Ludwig, 2003 and 2009) and SQUID II (for SHRIMP analyses). All the dates with a discordance higher than 10% were discarded. $^{206}Pb/^{238}U$ dates were used for zircon grains younger than 1.5 Ga and $^{207}Pb/^{206}Pb$ dates for >1.5 Ga zircon grains because of the significant increase in error of $^{206}Pb/^{238}U$ ratios for > 1.5 Ga zircon grains. The mean square weighted deviation (MSWD) of the youngest detrital zircon population was calculated using IsoplotR online (Vermeesch, 2018). Finally, Kernel Density Estimators (KDE) and histograms were calculated using the software DensityPlotter 8.4 (Vermeesch, 2012) and applying a KDE bandwidth and histogram bin of 20 Ma. Errors are expressed at 1σ level.

Hf isotope analysis was untertaken in the GeoHistory Facility, JdLC. The analyses were performed on previously dated zircon grains using a Resonetics resolution M-50A excimer laser, coupled to a Nu Plasma II multicollector inductively coupled plasma mass spectrometer. The analyzed isotopes were: ^{180}Hf, ^{179}Hf, ^{178}Hf, ^{177}Hf, ^{176}Hf, ^{175}Lu, ^{174}Hf, ^{173}Yb, ^{172}Yb, and ^{171}Yb. The ^{176}Lu decay constant of Scherer et al. (2001) and the Chrondritic Uniform Reservoir (CHUR) values of Blichert-Toft and Albarède (1997) were applied in the calculation of εHf values. Errors are expressed at 2σ level. A full description of this method can be found in Appendix A.

4. Results

4.1. U-Pb geochronology of detrital zircon grains

Sample CH2 (Middle Ordovician; Fig. 3): We carried out 150 analyses on 147 zircon grains, having obtained 138 concordant dates (Fig. 5A; Appendix B). Most ages define a dominant Cryogenian-Ediacaran population peaked at ≈ 630 Ma (c. 554-703 Ma; 51 analyses; 37 % of the data); two second-order peaks within this population occur at ≈ 590 and 680 Ma. The youngest detrital zircon population in this sample, based on 7 Ediacaran dates, has a mean $^{206}Pb/^{238}U$ age of 592.6 ± 1.7 Ma (MSWD = 1.24). A late Stenian – early Tonian population peaked at ≈ 1 Ga accounts for 27 dates and represents 19.6 % of the data, while second-order maxima appear at ≈ 780 and 885 Ma. Furthermore, third-order peaks appear at ≈ 1070, 1880 (c. 1955-1808 Ma; 8 analyses; 5.8 % of the data) and 2025 Ma (c. 2170-1986 Ma; 10 analyses; 7.2 % of the data); the two Paleoproterozoic peaks can be interpreted as featuring an Orosirian population that would include 18 dates (13 % of the data). Finally, a few scattered data yielded Neo-Archean dates (2.5-2.8 Ga).

Sample CH3 (Middle Ordovician; Fig. 3): 150 analyses were performed on 145 zircon grains, yielding 137 concordant dates (Fig. 5B; Appendix B). The majority of data can be grouped into a Cryogenian-Ediacaran population with two maxima at ≈ 600 and 635 Ma (c. 721-541 Ma; 64 analyses; 46.7 % of the data; Table 2). The $^{206}Pb/^{238}U$ mean age of the youngest detrital zircon population in this sample is 592.8 ± 1 Ma (MSWD = 1.07), being defined by 12 Ediacaran dates. A late Stenian – early Tonian population is also present, peaked at 992 Ma (c. 1085-881 Ma; 24 analyses; 17.5 % of the data; Fig. 5B). A minor Paleoproterozoic (Orosirian) population centered at c. 2 Ga is also present (16 dates; 11.6 % of the data), with peaks at 1874 (c. 1911-1834 Ma), 1970 (c. 1986-1957 Ma) and 2063 Ma (c. 2079-2045 Ma). Finally, a very minor Neo-Archean-Siderian population can be defined (11 dates; 8 % of the data), with peaks at ≈ 2455 (c. 2497-2428 Ma) and 2650 Ma (c. 2693-2601 Ma).

Sample CH4 (Middle Ordovician; Fig. 3): We carried out 150 analyses on 150 zircon grains, yielding 139 concordant dates (Appendix B). A prominent Cryogenian-Ediacaran population peaked at ≈ 595 and 645 Ma dominates this sample (c. 734-565 Ma; mean age 628.8±0.6 Ma; 53 dates; 38.1 % of the data; Fig. 5C and Table 2); a second-order peak centered at ≈ 710 Ma can also be included within this population. The youngest detrital zircon population of this sample is late Ediacaran in age (585.6±1.7 Ma; 6 dates; MSWD = 1.03). The late Stenian – early Tonian population defines a narrow band centered at ≈ 1 Ga (c. 1071-913 Ma; mean age 988.0±1.3 Ma; 30 dates; 21.6 % of the data). A third population of Paleoproterozoic (Rhyacian-Orosirian) age is also observed in this sample (21 dates; 15.2 % of the data), with maxima at ≈ 1865 (c. 1892-1848 Ma), 1970 (c. 1981-1965 Ma) and 2080 Ma (2154-2021 Ma). The older dates define a very minor Neo-Archean population centered at ≈ 2630 Ma (c. 2678-2563 Ma; 8 dates; 5.8 % of the data). There are also a few scattered analyses at ≈ 1225, 1425 and 2340 Ma, and 4 grains with early Cambrian dates (c. 545-520 Ma; Table 2).

Sample CH6 (Lower Devonian; Fig. 3) yielded 137 concordant dates obtained from 150 analyses on 135 zircon grains (Appendix B). The main population of this sample is Cryogenian-Ediacaran in age and peaks at ≈ 640 Ma, (c. 715-546 Ma; mean age 639.1±0.7 Ma; 54 analyses; 39.4%; Fig. 5D and Table 2); a third-order late Tonian peak centered at ≈ 775 Ma (c. 813-755 Ma; 7 dates; 5.1 % of the data) appears very close to the Cryogenian-Ediacaran population. The youngest detrital population of this sample, based on 7 analyses, is Ediacaran in age (590.2±1.7 Ma; MSWD = 1.26). A second noticeable population is of late Stenian – early Tonian age, with a main peak at ≈ 1 Ga (c. 1091-886 Ma; 37 dates; 27 % of the data). A Paleoproterozoic (Rhyacian-Orosirian) population is also present in this sample, with four separate peaks at ≈ 1890, 1970, 2055 and 2135 Ma (26 dates; 11.6 % of the data). Finally, a Neo-Archean population is defined by a single peak centered at ≈ 2615 Ma (5 dates; 3.6 % of the data).

Sample EV1 (Upper Ordovician; Fig. 3): We obtained 133 concordant dates from 148 analyses on 131 zircon grains (Appendix B). Most dates form a Cryogenian-Ediacaran population (c. 770-578 Ma; mean age 648.1±0.5 Ma; 67 dates, 50.4% of the data; Fig. 5E and Table 2) peaking at ≈ 650 Ma, with two second-order maxima at ≈ 600 and 690 Ma. A late Stenian –early Tonian population (c. 1141-803 Ma; 30 dates; 23.6 % of the data) peaks at ≈ 1070 Ma, with second-order maxima at ≈ 985 and 825 Ma. Paleoproterozoic (Rhyacian-Orosirian) and Neo-Archean minor populations are also present with ages ranging from c. 2138 to 1963 Ma (9 dates; 6.8 % of the data) and from 2719 to 2506 Ma (13 dates; 9.8 % of the data), respectively. Finally, a few dates define an early Cambrian peak centered at ≈ 500 Ma (5 dates with a mean age of 498.7±1.5 Ma; 3.8 % of the data). Paradoxically, the youngest detrital zircon population of this sample is Ediacaran (591.4 ± 1.8 Ma; 4 dates; MSWD = 1.17).

Sample EV2 (Upper Ordovician; Fig. 3): We obtained 114 concordant dates out of 122 analyses (80 by LA-ICPMS and 42 by SIMS; Table 1) of 122 zircon grains (Appendix B). The dominant population is Cryogenian-Ediacaran in age (c.714-546 Ma; mean age 632.1±0.5 Ma; 46 dates; 40.4% of the data), with a main peak at ≈ 630 Ma and minor peaks at ≈ 580 and 670 Ma (Fig. 5F; a third-order latemost Tonian peak centered at ≈ 735 Ma (10 dates; 8.8 % of the data) appears very close to the Cryogenian-Ediacaran population and separated from an older Stenian-Tonian population. This latter population is centered at ≈ 1 Ga (c. 1081-979 Ma; mean age 1023.2±1.2 Ma; 22 dates; 19.3%of the data) and shows a second-order peak at ≈ 910 Ma (c. 958-862 Ma; 10 dates; 8.8% of the data). A third and less relevant population is of Paleoproterozoic (Rhyacian-Orosirian) age, characterized by two maxima at ≈ 1900 and 2000 Ma (19 dates; 16.7 % of the data). Finally, a few scattered dates group at ≈ 2650 Ma and can be considered to represent a very scarce Neo-Archean population. Two dates of ≈ 500 Ma could be taken as evidence of the presence of Cambrian detrital zircon grains. The youngest detrital population of this sample, based on 4 dates, is late Ediacaran (582.4±1.9 Ma; MSWD = 0.79).

To sum up, all of the studied samples show a quite similar detrital zircon record characterized by (i) a dominant Ediacaran-Cryogenian population with maxima at c. 595-650 Ma, (ii) a noteworthy late Stenian – early Tonian population with an almost invariable peak at ≈ 1 Ga, (iii) a subordinate Paleoproterozoic population usually depicting several peaks between c. 1.8 and 2.1 Ga, and (iv) very scarce Neo-Archean-Siderian scattered dates that range from c. 2.4 to 2.7 Ga.

4.2. Hf isotope signature

In all of the studied samples, Lu–Hf isotopes were analyzed on the same zones of the zircon grains where concordant U–Pb zircon ages were obtained.

Sample CH2 (Middle Ordovician; Fig. 3): One hundred and thirty-eight Lu-Hf isotope measurements were carried out in this sample, yielding $^{176}Hf/^{177}Hf$ ratios of 0.280865–0.28271 and $^{176}Hf/^{177}Hf$ initial ratios of 0.280689–0.282677 (Appendix C). Seventy-six per cent of the CHUR-normalized (εHf) values are negative, ranging from -0.1 to -27.3, while supra-CHUR values vary from 0 to 10.9. Cryogenian-Ediacaran zircon grains show a vertical arrangement that extends from the Depleted Mantle (DM) line at εHf 10.9 down to εHf -27.3 (Fig. 6A; Table 2); negative εHf values clearly dominate over positive values, with most εHf clustered between 3.4 and -20. The late Stenian – early Tonian population depicts εHf values between 10.5 and -22.2 (Fig. 6A; Table 2); the zircon grains peaked at ≈1 Ga define an almost continuous and vertical trend with a range of εHf from 6.3 to -18.8; as for the two minor Tonian peaks, the younger (≈ 780 Ma) shows dominant positive εHf values between 7 and 3, while the older peak (≈ 885 Ma) plots following a vertical arrangement with rather dispersedly distributed values from 10.9 to -18.8. The Orosirian zircon population dominantly shows negative εHf values in the range -0.2 to 18.5. Finally, the scarce and scattered Neo-Archean zircon grains have εHf values that vary between 4.3 and -10.0.

Sample CH3 (Middle Ordovician; Fig. 3): One hundred and eight Lu-Hf isotope measurements were carried out in this sample, yielding $^{176}Hf/^{177}Hf$ ratios of 0.280902–0.282667 and $^{176}Hf/^{177}Hf$ initial ratios of 0.280859–0.282645 (Appendix C). Seventy-two per cent of the CHUR-normalized (εHf) values are negative, ranging from -0.1 to -27.7, while supra-CHUR values vary from 0 to 11.7. The Cryogenian-Ediacaran zircon grains are organized in a vertical array in the εHf *versus* age plot that covers an εHf interval of 9.7 to -27.7, with most data concentrated between εHf 5 and -15 (Fig. 6B). The late Stenian – early Tonian population also displays a vertical arrangement that extends from εHf 11.7 down to -23.8; the younger and older subpopulations (centered at ≈ 875 and 1070 Ma, respectively) dominantly show supra-CHUR εHf values, while ≈1 Ga zircon grains define a sub-CHUR vertical array (εHf = 0 to -16.9). The Paleoproterozoic (Orosirian) and Neo-Archean-Siderian zircon grains mostly have sub-CHUR εHf values, ranging from 0 to -20.7 (Fig. 6B).

Sample CH4 (Middle Ordovician; Fig. 3): One hundred and thirty-nine Lu-Hf isotope analyses were carried out in this sample, yielding $^{176}Hf/^{177}Hf$ ratios of 0.280927–0.28262 and $^{176}Hf/^{177}Hf$ initial ratios of 0.280826–0.282595 (Appendix C). Seventy-one per cent of the CHUR-normalized (εHf) values are negative, ranging from -0.1 to -24.2, while supra-CHUR values vary from 0 to 11.7. The Cryogenian-Ediacaran detrital zircon population depicts a vertical array in the εHf *versus* age diagram (Fig. 6C); more precisely, peak ages (≈ 630 Ma) cluster between εHf 5.7 and -21.2, while older ages (≈ 715 Ma) mostly show supra-CHUR values (3.2-8.7). The late Stenian – early Tonian population also defines a vertical trend in the εHf *versus* age diagram, with most of the values clustering between εHf 4.4 and -13 (Fig. 6C). Paleoproterozoic zircon grains are distributed between εHf 4.1 and -13.7, with older ages rather clustered at supra-CHUR values and younger ages progressively plotted at sub-CHUR values. Finally, the scarce Neo-Archean zircon grains show a subvertical spread, covering an εHf interval from 2.3 to -10.2.

Sample CH6 (Lower Devonian; Fig. 3): One hundred and thirty-six Lu-Hf isotope analyses were carried out in this sample, yielding $^{176}Hf/^{177}Hf$ ratios of 0.280984–0.282692 and $^{176}Hf/^{177}Hf$ initial ratios of 0.280917–0.282658 (Appendix C). Seventy-nine per cent of the CHUR-normalized (εHf) values are negative, ranging from -0.2 to -24.5, while positive values vary from 0 to 10.3. The dominant Cryogenian-Ediacaran population defines a heterogeneously distributed vertical array, extending from εHf 10.3 down to -24.5, with most values clustered between εHf 2.3 and -12.4 (Fig. 6D); a minor late Tonian population shows a similar vertical trend (εHf from 5.1 to -13). The late Stenian – early Tonian zircon grains also depict a remarkable vertical array with εHf values extending from 7 to -18.6 (Fig. 6D). The Paleoproterozoic zircon grains are distributed from εHf 4.1 down to -14.4, with older ages (peaked at ≈ 2135 Ma) defining a vertical trend (εHf from 4.1 to -3.6) and younger ones mostly clustered at sub-CHUR εHf values (0 to -14.4). Finally, the scarce Neo-Archean zircon grains arrange vertically in the sub-CHUR field (0.1 to -7; Fig. 6D).

Sample EV1 (Upper Ordovician; Fig. 3): One hundred and thirty Lu-Hf isotope analyses yielded ^{176}Hf/^{177}Hf ratios of 0.280618–0.282637 and ^{176}Hf/^{177}Hf initial ratios of 0.280525–0.282629 (Appendix C). Seventy-eight per cent of the CHUR-normalized (εHf) values are negative, ranging from -0.1 to -32.1, while positive values vary from 0.3 to 9.3. The Cryogenian-Ediacaran population is organized in a vertical array that extends from εHf values of 9.3 down to -24.3, with most data concentrated in the sub-CHUR field between -3.3 and -12.5 (Fig. 6E); late Tonian zircon grains also distribute vertically, but supra-CHUR εHf values dominate over sub-CHUR ones. The scarce Cambrian zircon grains mostly show εHf values close to CHUR (0.7 to -3.2), except for one of the measurements (-32.1). The late Stenian – early Tonian population defines an apparent vertical array with εHf values varying from 4.8 to -23.7 (Fig. 6E). The subordinate Paleoproterozoic and Neo-Archean populations dominantly show negative εHf values, though a few supra-CHUR values are also present.

Sample EV2 (Upper Ordovician; Fig. 3): The small size of a good number of the dated zircon grains did not allow Hf isotope analysis. Thus, only 59 Lu-Hf isotope analyses were conducted, yielding ^{176}Hf/^{177}Hf ratios of 0.281068–0.282675 and ^{176}Hf/^{177}Hf initial ratios of 0.281013–0.282666 (Appendix C). Sixty-four per cent of the CHUR-normalized (εHf) values are negative (-0.4 to -32.7) and the remaining positive (0.1 to 10.4). The analysed Cryogenian-Ediacaran and late Stenian – early Tonian zircon grains depict vertical trends in the εHf *versus* age diagram, extending from 10.4 to -32.7 and 9.1 to -16.0 εHf values, respectively (Fig.6F). In-between the two dominant detrital zircon populations, late Tonian detrital zircon grains cluster in the supra- (εHf 3.5 to 7.3) and sub-CHUR (εHf -3.7 to -6.2) fields. Finally, the scarce Paleoproterozoic and Neo-Archean zircon grains appear scattered between εHf values of 2.8 and -9.4.

To sum up, all of the samples studied are characterized in the εHf *vs* age diagram by two well-defined vertical arrays corresponding to the Cryogenian-Ediacaran and late Stenian – early Tonian detrital zircon populations, which extend from supra-CHUR values of ≈ 10 down to sub-CHUR values of ≈ -20 (Fig. 6). Paleoproterozoic and Neo-Archean detrital zircon populations are also mostly arranged in vertical trends, with supra-CHUR εHf values clustered at the ≈ 0 to 5 interval and sub-CHUR values ranging from 0 to -10.

5. Discussion

5.1. Stability of detrital zircon sources at Ordovician-Devonian time

All of the samples studied show very similar detrital zircon age spectra, with comparable percentages of both dominant and subordinate populations (Fig. 5). The Hf isotope signature is also uniform (Fig. 6), independent of the stratigraphic age of the sample. The simplest way of interpreting these results is by invoking stable detrital zircon sources at Ordovician-Devonian time, which, in turn, can be related to the tectonic setting: a passive-margin sequence is the obvious tectonic setting for the OMZ Ordovician-Devonian rocks.

The youngest detrital zircon populations can serve to constrain maximum depositional ages (MDA) of the studied sedimentary rocks. True depositional ages (TDA), usually based on paleontological dating of the sedimentary rocks and/or direct geochronological dating of volcanic intercalations, can be slightly delayed –or even be geologically coeval– with respect to MDA in some cases, but they can also be separated by hundreds of millions of years in others (e.g., Sharman and Malkoswki, 2020). In our case, the scarcity of Cambrian - Lower Ordovician detrital zircon grains in the OMZ Ordovician-Devonian means that there is a long time lapse between MDA and TDA, in agreement with the inferred passive-margin setting of the OMZ throughout the Ordovician - Devonian period. The most plausible explanation for the scarcity of Cambrian - Lower Ordovician detrital zircon grains is that the present-day exposed igneous rocks of Cambrian – Early Ordovician age did not reach the surface until the collisional and/or post-collisional stages of the Variscan orogeny, i.e., they would not have been exposed at surface until Carboniferous time.

5.2. Potential sources of detrital zircon grains in the Ordovician-Devonian sedimentary sequence of the Ossa-Morena Zone

The most important –and unexpected- finding of our study is the recognition of an important late Stenian – early Tonian detrital zircon population peaked at ≈ 1 Ga, with percentages of ≈ 20 % in all of the studied samples. The presence of Tonian detrital zircon grains in the OMZ was previously reported by Pereira et al. (2014), who noted a progressive increase from the middle-late Cambrian (Ossa and Fatuquedo Formations) to the Ordovician-Silurian (Colorada Formation). The appearance of a noticeable Stenian – early Tonian detrital zircon population in Ordovician-Devonian OMZ rocks has important paleogeographic and tectonic consequences, which will be discussed separately in sub-section 5.3.

5.2.1. Neo-Archean and Paleoproterozoic detrital zircon populations

These populations are always present in the studied samples and depict three-order peaks in the OMZ Ordovician-Devonian rocks (Fig. 5). These populations are traditionally thought to have been derived from the WAC, where igneous and metamorphic rocks of 1.8-2.1 and 2.5-2.75 Ga crop out extensively (Abati et al., 2010 and 2012 and references therein; Bea et al., 2020). However, other cratons and metacratons cropping out in northern Africa (Arabian-Nubian shield, Sahara metacraton, Tuareg shield) also contain rocks of these two age intervals (see for instance Pereira et al., 2008, Drost et al., 2011, and Cambeses et al., 2017, for compilations), and could also be invoked as primary sources of both Paleoproterozoic and Neo-Archean detrital zircon grains. Therefore, the pre-Mesoproterozoic detrital zircon populations in the Ordovician-Devonian OMZ rocks cannot be used in isolation to unequivocally constrain their primary source.

5.2.2. Cryogenian-Ediacaran detrital zircon population

This population is dominant in all of the studied samples, representing ≈ 40-50 % of the concordant analyses (Figs. 5 and 7). It is also prevalent in upper Ediacaran- lower Cambrian samples of the OMZ (Fernández-Suárez et al., 2002; Linnemann et al., 2008; Pereira et al., 2011 and 2012a), and some Carboniferous samples (Dinis et al., 2018; Pereira et al., 2012c and 2020). Beyond the OMZ, Cryogenian-Ediacaran detrital zircon grains are the most conspicuous population in the CIZ, WALZ, CZ and GT MZ (and counterparts through the Cantabrian arc (e.g., Martínez Catalán et al., 2020), as well as in the Moroccan mesetas (Accotto et al., 2019 and 2021; Fig. 7). The source of this population is unanimously attributed to a prominent Cadomian/Avalonian magmatic arc that would have extended all along the northern Gondwanan continental margin at Cryogenian-Ediacaran time (e.g., Murphy et al., 2018, and references therein). This arc was apparently long-lasting (≈ 550 to 700 Ma) and laterally continuous along thousands of km, developed over different segments of older continental crust from the WAC to the Arabian Nubian shield (e.g., Fig. 3 of Murphy et al., 2018). The εHf isotope signature of the Cryogenian-Ediacaran detrital zircon grains can serve to discriminate between zircon grains sourced from different segments of the Cadomian/Avalonian magmatic arc. In this regard, the OMZ Cryogenian-Ediacaran εHf values compare quite well with those from the WAC and the Sahara metacraton (Fig. 8; Henderson et al., 2016 and references therein). In contrast, the Cryogenian-Ediacaran εHf signature from the Arabian-Nubian shield does not show the strong vertical distribution on an age *versus* εHf plot that was observed in the OMZ, being dominated by supra-CHUR values (Henderson et al., 2016). Therefore, the εHf data suggest that the OMZ detrital zircon of Ediacaran-Cryogenian age might have been derived from either the WAC or the Sahara metacraton segments of the Cadomian/Avalonian magmatic arc. A more careful analysis of the εHf data shows that the older ages of the Cryogenian-Ediacaran population mainly plot in the supra-CHUR field in the case of the OMZ and the Sahara metacraton, while the West African Craton values do not (Fig. 8). The evidence is meager, but it might favor a dominant derivation of Cryogenian-Ediacaran detrital zircon grains in the OMZ Ordovician-Devonian rocks from a source located in the Sahara metacraton segment of the long-lived Cadomian/ Avalonian magmatic arc.

5.2.3. Cambrian – Early Ordovician detrital zircon grains

The very scarce Cambrian – Early Ordovician detrital zircon grains found in the studied samples were probably derived from local igneous sources (e.g., Táliga-Barcarrota and Salvatierra de los Barros plutons; Fig. 2; see Simancas et al., 2004, for a compilation), formed during the rifting episode that preceded the passive margin stage. The εHf isotopic signature of one of the Cambrian zircon grains is compatible with a juvenile mantle-derived origin, while the remainder show sub-CHUR εHf values indicative of recycling of older crustal rocks (Fig. 6). Therefore, the εHf signature supports the provenance of these zircons from igneous rocks formed mostly by the partial melting of the crust underlying the OMZ at that time.

5.3. Late Stenian – early Tonian detrital zircon population: new insights into the paleogeography and tectonics of the OMZ during late Neoproterozoic – early Cambrian times

The late Ediacaran – early Cambrian rocks of the OMZ include a dominant Cryogenian-Ediacaran detrital zircon population and a subordinate Paleoproterozoic population, with a systematic late Stenian – early Tonian (≈1 Ga) gap (Fernández-Suárez et al., 2002; Linnemann et al., 2008; Pereira et al., 2011 and 2012a). However, the new U-Pb-Hf data reported here show the presence of a significant (≈ 20 %) late Stenian – early Tonian (≈1 Ga) detrital zircon population in Ordovician-Devonian OMZ rocks. To explore the paleogeographic and tectonic consequences of this finding, the context provided by additional geological data is essential.

The discussion that follows addresses two main issues. The first is establishing the primary source of the late Stenian – early Tonian detrital zircon grains, bearing in mind that despite the potential existence of intermediate sediment repositories (ISR) (Pereira and Gama, 2021), a direct connection with a primary source seems justified by the high percentage of this population in the studied Ordovician-Devonian OMZ samples. In this respect, the middle Cambrian siliciclastic rocks from the Anti-Atlas contain a meager 1 Ga detrital zircon population, which cannot be invoked as an ISR for the Stenian – early Tonian detrital zircon grains found in the Ordovician-Devonian OMZ rocks (Fig. 7), since sediment recycling could maintain the detrital zircon content in the primary/intermediate source (Pereira and Gama, 2021) but not selectively increase the content of one of the populations. The second issue to be addressed is why late Stenian – early Tonian detrital zircon grains are not found in the late Ediacaran – early Cambrian rocks of the OMZ, while they constitute a significant population in the Ordovician-Devonian rocks.

5.3.1. Primary late Stenian – early Tonian *detrital zircon sources*

Two potential late Stenian – early Tonian primary sources exist, separated by the Paleoproterozoic WAC: i) to the west the Grenville terranes (Henderson et al., 2016); and ii) to the east the Sahara metacraton / Arabian-Nubian shield (Bea et al., 2010, Cambeses et al., 2017). The fact that Mesoproterozoic detrital zircon ages in

Grenvillian terranes show a multi peak distribution between 1 and 1.6 Ga (e.g., Henderson et al., 2016; Pérez-Cáceres et al., 2017) can serve to cast doubt on a Grenvillian provenance of late Stenian – early Tonian zircon grains in the Ordovician-Devonian OMZ rocks, which invariably show a single peak population centered at 1 Ga (Figs. 5 and 7). On this ground, the northern Africa cratonic areas seem to be more plausible sources. Furthermore, the εHf isotopic signature of the late Stenian – early Tonian detrital zircon population in the OMZ Ordovician-Devonian rocks shows a remarkably similar distribution than those of the Sahara metacraton and the Arabian-Nubian shield (Iizuka et al., 2013; Henderson et al., 2016; Fig. 8).

5.3.2. Right-lateral translation of the Ossa-Morena Zone at late Ediacaran – early Cambrian times

In addition to the arguments given above, the geological context provides support for the origin of late Stenian – early Tonian OMZ zircon grains being the Sahara metacraton. The most plausible key mechanisms to explain the 1 Ga detrital zircon provenance are putative late Ediacaran – early Cambrian right-lateral displacements of the OMZ, modifying its position with respect to the CIZ and the Precambrian cratons along the northern Gondwanan margin (Fig. 9). However, a first consideration to be made is whether the 1 Ga detrital zircon grains were supplied by a relatively close source or by a distant one. In this respect, the OMZ might have been located close to the WAC all along the Ediacaran-Devonian timespan, with the 1 Ga detrital zircon grains coming from a distant Sahara metacraton source. This possibility is undermined by the systematic presence and relative abundance of Stenian – early Tonian zircon grains with respect to Paleoproterozoic ones in the Ordovician-Devonian OMZ rocks (Fig. 5). Distant but intermittent NE African detrital zircon sources have been invoked to explain the presence of a 1 Ga population in Ordovician-Devonian rocks from the Eastern Moroccan Meseta (Accotto et al., 2019). Nevertheless, in the Eastern Moroccan Meseta samples the 1 Ga detrital zircon population is neither systematic nor so abundant as in the similar age OMZ samples (Fig. 7). On this ground, we can plausibly discard a long-lasting and distant NE African provenance for the 1 Ga detrital zircon population in the Ordovician-Devonian OMZ rocks. Hence, in the following paragraphs we will assume a relatively close Sahara metacraton source for the Stenian – early Tonian detrital zircon population in the studied samples.

The proposed tectonic evolution is based on a critical review of three main issues: i) the boundary between the OMZ and the CIZ (Figs. 1 and 2); ii) the correlation between the OMZ and the STZ in Central Europe (Figs. 1A and 9A); and iii) the kinematics of the Cadomian late Ediacaran subduction (Fig. 9B and C).

i) According to the Variscan evolution, the boundary between the OMZ and the CIZ is the Badajoz-Córdoba shear zone (BCSZ; Fig. 2), with convergent-transcurrent left-lateral kinematics and Devonian-to-earliest Carboniferous activity (Burg et al., 1981; Azor et al., 1994 and 2019; Simancas et al., 2001). The BCSZ is a first-order tectonic structure developed over an early Paleozoic rifting band, which eventually gave way to MORB-type mafic rocks (e.g., Gómez-Pugnaire et al., 2003). These mafic rocks underwent eclogite facies metamorphism during early Variscan tectonic evolution

(López Sánchez-Vizcaíno et al., 2003). Despite this dominant view, some authors have located the OMZ/CIZ boundary north of the BCSZ, along the late Carboniferous Pedroches Batholith (Fig. 2). Actually, the Ediacaran formations on both sides of this batholith show significant differences (San José et al., 2004): southwards, the Serie Negra formation contains Paleoproterozoic detrital zircon grains but not late Stenian – early Tonian grains (Fernández-Suárez et al., 2002; Pereira et al., 2011 and 2012a; Cambeses et al., 2017), while northwards the Lower Alcudian formation yielded a late Stenian – early Tonian detrital zircon population (Talavera et al., 2012; Fernández-Suárez et al., 2014). Therefore, the existence of a hidden tectonic line underneath the Pedroches Batholith has been proposed (e.g., Pérez-Cáceres et al., 2017; Fig. 9A). This concealed tectonic boundary would have been active at latest Ediacaran - earliest Cambrian time, since the Cambrian formations are similar on both sides. Consequently, this tectonic line should not be used to differentiate two Variscan zones of the Iberian massif, i.e. the Variscan zonation should be based on the Paleozoic evolution. Nevertheless, the hidden tectonic line seems to have played the main role in the lateral translation of the OMZ to a location where late Stenian – early Tonian detrital zircon grains were available (see below).

ii) The along-strike orogenic correlation between the OMZ in Iberia and the STZ in Central Europe is implicit in most of Variscan reconstructions (e.g., Matte, 2001; Simancas et al., 2005; Franke, 2006; Martínez Catalán et al., 2020). Thus, the outcrops of the OMZ and the STZ would form part of a ribbon-shaped continental terrane (Fig. 9A), which, based on the nature of its basement, would be linked to the Paleoproterozoic WAC. Direct evidence of the Paleoproterozoic age of the OMZ/STZ basement has been found in the granulites of the Galicia Bank (Guerrot et al., 1989; Gardien et al., 2000) and in Proterozoic outcrops of the Cherbourg-Trégor region in NW France (Calvez and Vidal, 1978; Samson and D'Lemos, 1998; Inglis et al., 2004). Furthermore, the presence of Paleoproterozoic detrital zircon grains but not late Stenian – early Tonian grains characterizes the Ediacaran rocks of both OMZ and STZ (Linnemann et al., 2014; Pérez-Cáceres et al., 2017). Finally, Sm-Nd isotope geochemistry on the Ediacaran rocks of the OMZ and STZ supports an affinity between these rocks and the Paleoproterozoic WAC (Fuenlabrada et al., 2020; Rojo-Pérez et al., 2021).

iii) The broad-scale plate tectonic displacement related to the Cadomian orogeny is very difficult to constrain, since almost no direct data on the subduction kinematics are available. Nevertheless, important -though indirect- pieces of evidence can be inferred from the strip geometry of the OMZ/STZ terrane, which connects westwards with the Paleoproterozoic WAC (Fig. 9A), depicting a tail-shaped terrane. We suggest that the OMZ/STZ terrane was primarily a continental ribbon formed by dextral strike-slip displacement of a piece of the northern border of the Paleoproterozoic WAC. Regarding direct Cadomian kinematic data, Linnemann et al. (2008) proposed a tectonic frame characterized by oblique left-lateral Ediacaran subduction, which would have turned into right-lateral displacements during latest Ediacaran - early Cambrian times (Fig. 9B). Furthermore, considering the Pan-African suture cropping out in the Anti-Atlas (El Hadi et al., 2010), a broad-scale kinematic scenario can be proposed by combining the convergence vector of the Pan-African collision with the oblique left-lateral Cadomian subduction on the northern border of the WAC (Fig. 9B; Linnemann et al., 2008). Thus, the blockade of the Pan-African convergence at around 580 Ma (El Hadi et al., 2010) would result in dextral kinematics, which, in turn, would have displaced eastwards a strip from the northern border of WAC, namely the OMZ/STZ terrane with Paleoproterozoic basement and Ediacaran magmatic arc rocks (Fig. 9B; Pérez-Cáceres et al., 2017). Indeed, present-day oblique subduction in many areas worldwide has usually given way to strain partitioning with decoupling of segments of the overriding magmatic arc along transcurrent faults subparallel to the trench (e.g., Sumatra, Fitch, 1972; Kurile arc, Kimura 1986; Burma, Maung, 1987). Alternatively, drawing on the recent evolution of western North America (e.g., Atwater, 1970), other authors have proposed that the subduction of a ridge triggered the shift from convergence to lateral displacement at the end of the Cadomian orogenesis (e.g., Linnemann et al., 2008; Sánchez-García et al., 2019). Independently of the reason behind dextral displacements, a subsequent kinematic change would have occurred at early-middle Cambrian time, when a rifting environment existed at the northern Gondwanan margin, lasting until the Early Ordovician and marking the onset of the Variscan cycle (e.g., Sánchez-García et al., 2019; Simancas, 2019).

In summary, the above-presented data lead us to propose that the OMZ-STZ domain is part of an originally continuous and thin continental terrane with Paleoproterozoic basement and located close to the WAC. This terrane would have been displaced eastwards from the northern part of the WAC to a position close to the Sahara metacraton. In SW Iberia, the transcurrent fault or shear zone permitting dextral displacement would be currently concealed underneath the late Carboniferous Pedroches Batholith (Figs. 2 and 9A), which bounds two contrasting types of Ediacaran rocks. The transcurrent tectonics must have occurred at latest Ediacaran - earliest Cambrian time.

As a corollary, we can now answer the two main issues raised at the beginning of this section regarding the presence of late Stenian – early Tonian c detrital zircons in Ordovician-Devonian OMZ rocks. The primary source of the ≈1 Ga detrital zircon population was probably the Sahara metacraton. Before early Cambrian time, the OMZ-STZ domain was located at the border of the WAC and received detrital zircon grains from the Paleoproterozoic Eburnian crust and the Neoproterozoic magmatic arc, developed at the northern Gondwanan margin. Once the dextral displacement that placed the OMZ-STZ terrane close to the Sahara metacraton occurred, Mesoproterozic detrital zircon grains coming from this source started to appear in the Paleozoic sedimentary rocks of the OMZ. The neighboring CIZ would have been located at Ordovician-Devonian time close to the Sahara metacraton too, but westwards with respect to the OMZ (Fig. 9B), in order to account for the important left-lateral displacement related to the whole Variscan collisional evolution of SW Iberia (e.g., Pérez-Cáceres et al., 2016).

A latest Ediacaran - earliest Cambrian age of the OMZ-STZ dextral displacement fits with late Stenian – early Tonian detrital zircon grains occurring in the Cambrian rocks of this terrane. Nevertheless, this detrital zircon population was not detected in the two lower Cambrian samples analyzed until now (Linnemann et al., 2008; Pereira et al., 2012a). This is an open issue that needs more data to be properly explored, but considering putative changes in the sedimentary paleogeography due to the Cambro-Ordovician rifting seems to be an appropriate approach.

6. Concluding remarks

i) Ordovician-Devonian sedimentary rocks from the OMZ show a similar detrital zircon content characterized by a dominant Cryogenian-Ediacaran population with a maximum at 595-650 Ma, a noteworthy late Stenian – early Tonian population with an almost invariable peak at ≈ 1 Ga, and a subordinate Paleoproterozoic population usually depicting several peaks between 1.8 and 2.1 Ga.

ii) The Hf isotope signature of the studied samples features two well-defined vertical arrays with age corresponding to the Cryogenian-Ediacaran and Stenian-Tonian detrital zircon populations, which extend from supra-CHUR εHf values of ≈ 10 down to sub-CHUR values of ≈ -20. Neo-Archean and Paleoproterozoic detrital zircon grains have a vertical spread, with supra-CHUR εHf values clustered at the ≈ 0 to 5 interval and sub-CHUR values ranging from 0 to -10.

iii) The similarity of both detrital zircon spectra and Hf isotope signature in all of the studied samples fits with the passive margin setting that other geological data (particularly stratigraphy) also suggest for the OMZ at Ordovician–Devonian times.

iv) The OMZ and equivalent zones in Central Europe were located close to the WAC along the northern Gondwanan margin at Ediacaran times, according to the presence of ≈ 2 Ga orthogneiss remnants and the detrital zircon record of Ediacaran rocks, characterized by a dominant Cryogenian-Ediacaran population, a subordinate Paleoproterozoic one, and a systematic lack of late Stenian – early Tonian zircon grains.

v) The unexpected presence of a ≈ 1 Ga detrital zircon population in Ordovician-Devonian OMZ sedimentary rocks requires relocating this continental piece at that time in a position close to the Sahara metacraton, which represents the most probable source of late Stenian – early Tonian detrital zircon grains.

vi) The translation of the OMZ and equivalent zones eastwards from an original position close to the WAC to an Ordovician-Devonian location close to the Sahara metacraton probably occurred at latest Ediacaran – earliest Cambrian time in a dextral strike-slip tectonic scenario that postdated Pan-African collision and Cadomian subduction.

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Figure captions

Figure 1: A) Plate reconstruction at latest Paleozoic time to show the distribution of Variscan sutures, the deformation fronts, and the location of the Iberian Massif and

other European Variscan domains. **B)** Main Variscan zones in the Iberian Massif, with location of the map shown in Figure 2 (red polygon).

Figure 2: Schematic lithological map of the Ossa-Morena Zone and the southernmost Central Iberian Zone. The two blue stars locate the areas sampled in this study (EV: El Valle; CH: Cerrón del Hornillo). Two of the numerous Cambrian-Ordovician intrusive bodies have been marked (T-B: Táliga-Barcarrota plutons; S: Salvatierra de los Barros pluton). The boundary between the Ossa-Morena and the Central Iberian zones (Badajoz-Córdoba Shear Zone) has also been labelled (BCSZ).

Figure 3: General stratigraphic column of the Ossa-Morena Zone (left), and detailed Ordovician-Devonian sequence adapted from Robardet et al. (1998) (right) with location of the studied samples (CH1, CH2, CH4, CH6, EV1 and EV2).

Figure 4: Cathodoluminescence images of detrital zircon grains from the studied samples with the U-Pb analytical spots marked by circles and the obtained ages and ε_{Hf} values. High-U sectors of the crystals correspond to darker zones in the images.

Figure 5: A-F) U-Pb detrital zircon age distribution in the studied Ordovician-Devonian rocks of the Ossa-Morena Zone. The results are presented as Kernel Density Estimates (KDE, black lines) and histograms (grey bars) on frequency *versus* age plots, as well as on pie-charts. Colors in the frequency *versus* age plots and pie-charts correspond to the different periods as specified in the legend.

Figure 6: A-F) εHf values *versus* U-Pb ages for the studied samples. CHUR: Chondritic Uniform Reservoir (Bouvier et al., 2008); NC: new crust (Dhuime et al., 2011). Colors in the plots correspond to the different periods as specified in Figure 5.

Figure 7: Comparison among KDE of detrital zircon spectra for different ages and from different areas of the Variscides and surrounding regions. The following references have been used as primary sources of U-Pb detrital zircon ages: Abati et al. (2010); Accotto et al. (2019, 2021 and in press); Avigad et al. (2012); Díez Fernández et al. (2010); Fernández-Suárez et al. (2002 and 2014); Gutiérrez-Alonso et al. (2003); Linnemann et al. (2008 and 2011); Meinhold et al. (2011); Ordóñez-Casado (1998); Pereira et al. (2010a and b, 2011, 2012a and d); Schäfer (1990); Shaw et al. (2014); Talavera et al. (2012).

Figure 8: Comparison of εHf results in the studied samples **(A)** with those of the West African Craton **(B)**, Sahara metacraton **(C)** and Arabian-Nubian shield **(D)**, redrawn from Henderson et al. (2016). The following references have been used as primary sources of εHf versus age data: Abati et al. (2012); Ali et al. (2013); Avigad et al. (2012); Be'eri-Shlevin et al. (2014); Gärtner et al. (2014); Iizuka et al. (2013); Linnemann et al. (2014); Meinhold et al. (2014); Morag et al. (2011 and 2012); Robinson et al. (2014). CHUR: Chondritic Uniform Reservoir (Bouvier et al., 2008); NC: new crust (Dhuime et al., 2011). Colors in the plots correspond to the different periods as specified in Figure 5.

Figure 9: A) Broad-scale reconstruction of the Variscan belt at latest Carboniferous time, emphasizing the connection of the Ossa-Morena / Saxo-Thuringian ribbon-shaped terrane with the West African Craton. The evidence supporting this connection is discussed in the text. **B)** Tentative kinematic scenario for the Cadomian orogenesis depicting the change from variably oblique subduction throughout the northern Gondwanan margin at Ediacaran time to a single dextral subduction at the vanishing stages of the Pan-African convergence (latest Ediacaran – earliest Cambrian). Once Pan-African subduction was blocked, strain partitioning would have given way to transcurrent tectonics, translating eastwards a strip of the Cadomian/Avalonian magmatic arc. The translated ribbon-shaped arc fragment represents the crustal basement of the Ossa-Morena and Saxo-Thuringian Variscan zones. The dextral kinematics would have relocated the Ossa-Morena/Saxo-thuringian zones close to the Sahara metacraton, which, in turn, would have sourced the detrital zircon grains analyzed in the Ordovician-Devonian rocks. $V_{NA/CS}$: assumed motion of the Cadomian subducted plate with respect to the North Africa complex basement (Tuareg shield, Sahara metacraton and Arabian-Nubian shield); $V_{WA/NA}$: assumed motion of the North Africa complex basement with respect to the West African Craton; $V_{WA/CS}$: assumed motion of the Cadomian subducted plate with respect to the West African Craton. See text for further explanations and alternative proposals.

Table Captions

Table 1: Location of the samples and number and type of analyses carried out; (*) Total number of U-Pb analyses performed and concordant values (in bold).

Table 2: Summary of U-Pb-Hf results.