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Palaeogeography and crustal evolution of the Ossa–Morena Zone, southwest Iberia, and the North Gondwana margin during the Cambro-Ordovician: a review of isotopic evidence

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ABSTRACT

Cambro-Ordovician palaeogeography and fragmentation of the North Gondwana margin is still not very well understood. Here we address this question using isotopic data to consider the crustal evolution and palaeogeographic position of the, North Gondwana, Iberian Massif Ossa–Morena Zone (OMZ). The OMZ preserves a complex tectonomagmatic history: late Neoproterozoic Cadomian orogenesis (ca. 650–550 Ma); Cambro-Ordovician rifting (ca. 540–450 Ma); and Variscan orogenesis (ca. 390–305 Ma). We place this evolution in the context of recent North Gondwana Cambro-Ordovician palaeogeographic reconstructions that suggest more easterly positions, adjacent to the Sahara Metacraton, for other Iberian Massif zones. To do this we compiled an extensive new database of published late Proterozoic–Palaeozoic Nd model ages and detrital and magmatic zircon age data for (i) the Iberian Massif and (ii) North Gondwana Anti-Atlas West African Craton, Tuareg Shield, and Sahara Metacraton. The Nd model ages of OMZ Cambro-Ordovician crustal-derived magmatism and Ediacaran-Ordovician sedimentary rocks range from ca. 1.9 to 1.6 Ga, with a mode ca. 1.7 Ga. They show the greatest affinity with the Tuareg Shield, with limited contribution of more juvenile material from the Anti-Atlas West African Craton. This association is supported by detrital zircons that have Archaean, Palaeoproterozic, and Neoproterozoic radiometric ages similar to the aforementioned Iberian Massif zones. However, an OMZ Mesoproterozoic gap, with no ca. 1.0 Ga cluster, is different from other zones but, once more, similar to the westerly Tuareg Shield distribution. This places the OMZ in a more easterly position than previously thought but still further west than other Iberian zones. It has been proposed that in the Cambro-Ordovician the North Gondwana margin rifted as the Rheic Ocean opened diachronously from west to east. Thus, the more extensive rift-related magmatism in the westerly OMZ than in other, more easterly, Iberian Massif zones fits our new proposed palaeogeographic reconstruction.

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1. Introduction

The Iberian Massif (Pérez-Estaún and Bea 2004 and references therein) is the best preserved segment of the European Variscides, the orogenic belt that resulted from collision of Gondwana and Laurussia during the late Palaeozoic (Bard *et al*. 1973, 1980; Burg *et al*. 1981; Matte 1986). The Iberian Massif consists of several zones with a roughly symmetric disposition and distinct stratigraphic, structural, magmatic, and metamorphic characteristics. There are broadly from north to south: Cantabrian; West Asturian-Leonese; Galicia-Tras-os-Montes; Central Iberian; Ossa–Morena Zone (OMZ); and South Portuguese (Figures 1 and 2; Lotze 1945; Julivert *et al*. 1972; Ribeiro *et al*. 1990; Martínez-Catalán *et al*. 1999).

New palaeogeographic models have recently been published for several of the Iberian zones during the Ediacaran–lower Palaeozoic. Bea *et al*. (2010) proposed a new position for the Central Iberian Zone, north of the Sahara Metacraton (Figure 1), further east than previously positioned (Eguiluz *et al*. 2000; Fernández-Suárez *et al*. 2000; Gutiérrez-Alonso *et al*. 2003). Fernández-Suárez *et al*. (2014) located the Cantabrian Zone next to the Sahara Metacraton and Arabian Nubian Shield and suggested that it represents a passive margin fragment of the northern fringe of Gondwana (cf. Pastor-Galán *et al*. 2013a) (Figure 1). Shaw *et al*. (2014) proposed a comparable position for the West Asturian-Leonese Zone (Figure 1). Likewise, Díez-Fernández *et al*. (2010) placed the Galicia Tras-os-Montes parautochthonous units close to the Sahara Metacraton, whereas allochthonous units are related to the Anti-Atlas (Albert *et al*. 2015) or Avalonia (Henderson *et al*. 2016) (Figure 1).

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Supplemental data for this article can be accessed here.

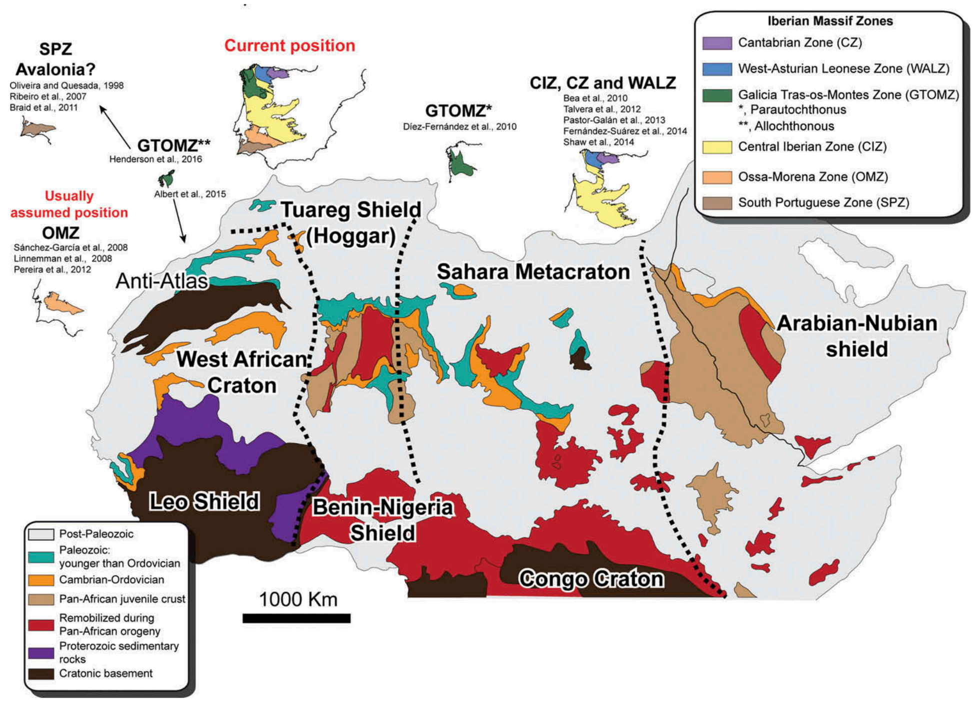

Figure 1. The Iberian Massif zones and their Cambro-Ordovician palaeographic positions with respect to the north African areas. The cratonic region ages are taken from Avigad *et al*. (2003). Note that the Central Iberian Zone (CIZ), Cantabrian Zone (CZ) and West-Asturian Leonese Zone (WALZ) are close to the Sahara Metacraton and Arabian Nubian Shield (Bea *et al*. 2010; Talavera *et al*. 2013; Pastor-Galán *et al*. 2013a; Fernández-Suárez *et al*. 2014; Shaw *et al*. 2014). The parautochthonous units from the Galicia-Tras-os-Montes Zone (GTOMZ) are located between the Sahara Metacraton and Tuareg Shield (Díez-Fernández *et al*. 2010) whereas the allochthonous units are associated with the Anti-Atlas (Albert *et al*. 2015) or Avalonia (Henderson *et al*. 2016). The South Portuguese Zone (SPZ) has been considered part of the Avalonia microplate (e.g. Ribeiro *et al*. 2007; Braid *et al*. 2011). The Ossa–Morena Zone (OMZ) has traditionally been located close to the west Anti-Atlas West African Craton (Nance and Murphy 1994; Fernández-Suárez *et al*. 2002; Linnemann *et al*. 2004, 2008; Pereira *et al*. 2008, 2011, 2012c).

During the Cambro-Ordovician, the palaeogeographic position of the OMZ and South Portuguese Zone, by contrast, are not so well constrained. The South Portuguese Zone has been associated with the Avalonia microplate (e.g. Ribeiro *et al*. 2007; Braid *et al*. 2011) (Figure 1). On the other hand, the OMZ has been considered part of the Armonica microplate (e.g. Matte 2001) or a continental block that represents the most northern margin of Gondwana (e.g. Robardet and Gutiérrez-Marco 2004). In the latter model the OMZ is positioned close to the Anti-Atlas West African Craton (Figure 1; e.g. Nance and Murphy 1994; Linnemann *et al*. 2008; Pereira *et al*. 2011).

The objective of this study is to determine the Cambro-Ordovician palaeogeographic position of the OMZ prior to its amalgamation with other Iberian Massif zones during the Variscan Orogeny. In particular we aim to understand how the OMZ fits into new models of: (i) early Palaeozoic palaeogeographic reconstructions that place other Iberian zones further east than previously thought (Figure 1); and (ii) Rheic Ocean opening (Linnemann *et al*. 2004, 2008; Nance *et al*. 2010, 2012).

To do this we compiled an extensive database (see supplementary material I and II) of published geochemical and geochronological data. The former includes major and trace elements and Nd isotopic data. The latter comprises Nd model ages, detrital zircon ages magmatic and pre-magmatic zircon, and whole-rock ages. The data set includes Ediacaran to Ordovician sedimentary rocks and Cambro-Ordovician igneous rocks from: the OMZ, the other Iberian Massif zones; and three main North Gondwana, African, areas. These areas, from which the OMZ may have detached, are the Anti-Atlas West African Craton, the Tuareg Shield and the Sahara Metacraton. The data have permitted us to develop new models for the palaeogeographic position of the OMZ during the Cambro-Ordovician.

2. Regional setting

The OMZ preserves evidence of complex evolution during the late Neoproterozoic Cadomian Orogeny, ca. 650–550 Ma, and subsequent Cambro-Ordovician, ca. 510–480 Ma, extension (e.g. Quesada *et al*. 1991; Eguiluz *et al*. 2000; Expósito *et al*. 2003; Sánchez-García *et al*. 2003, 2010; Silva and Pereira 2004; Extebarria *et al*. 2006; Pereira *et al*. 2006; Chichorro *et al*. 2008; Montero

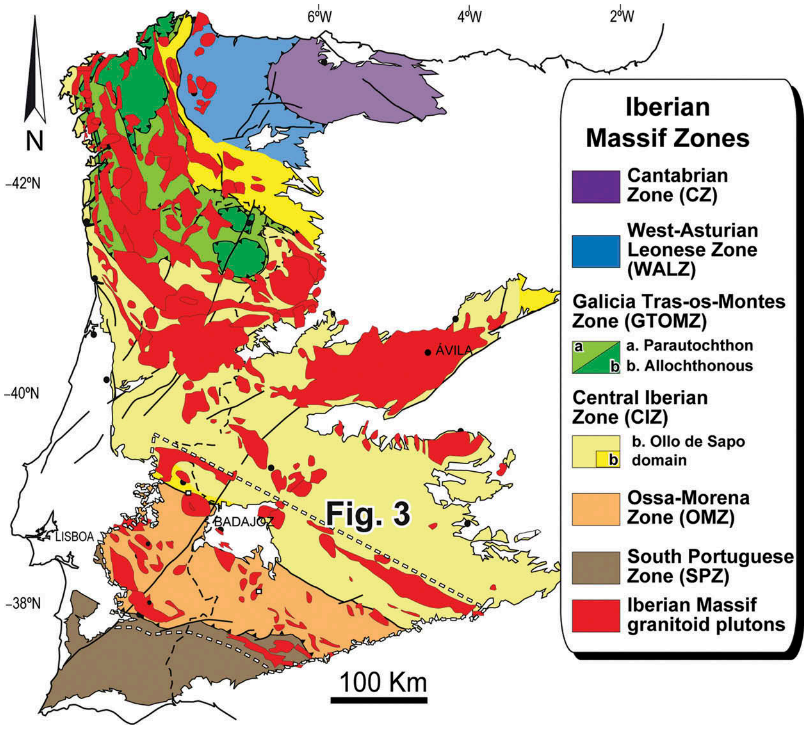

Figure 2. Zones of the Iberian Massif indicating the distribution of granitoids, modified from Bea *et al*. (2006) and Martínez-Catalán (2011).

et al. 2009a). Furthermore, during the Carboniferous Variscan Orogeny the region was deformed, as reflected in two complex contacts that register convergence at its northern and southern margins (Bard 1977; Bard and Moine 1979; Quesada 1991; Matte 2001) (Figures 2 and 3).

The contact of the OMZ with the Central Iberian Zone, to the north, is marked by the Badajoz–Córdoba shear zone.

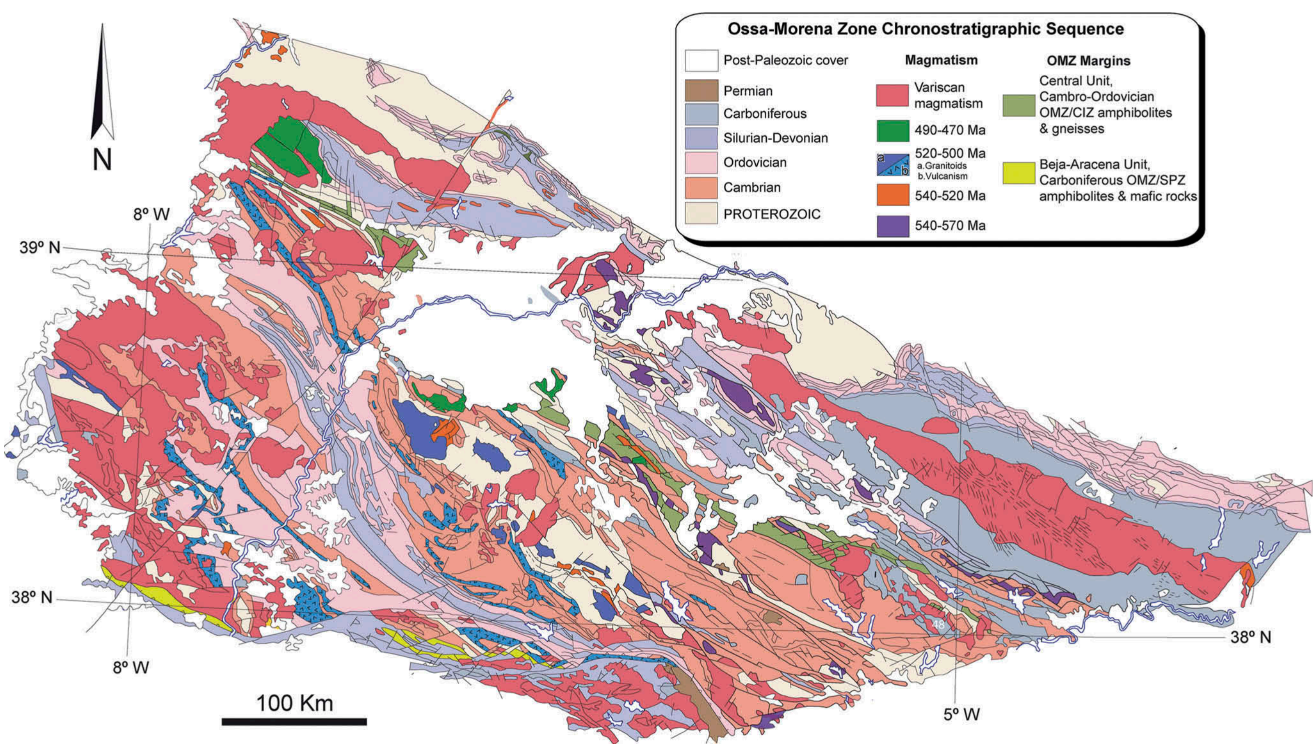

Figure 3. Chronostratigraphic sequence of the Ossa–Morena Zone also showing the different magmatic stages and the Ossa–Morena Zone boundaries with the Central Iberian Zone and the South Portuguese Zone, modified from Gabaldón (2001).

Some authors consider that this shear zone was originally active during the Cadomian Orogeny and was then subsequently reactivated during the Variscan (Ábalos *et al*. 1991; Quesada 1991; Ábalos and Díaz-Cusi 1995; Eguiluz *et al*. 1995, 2000). However, other authors consider it to be a major intra-continental, solely Variscan, shear zone (e.g. Burg *et al*. 1981; Azor *et al*. 1994; Pereira *et al*. 2007, 2009; 2010a, 2010b, 2012a). Identification of mid-ocean rigde basalt (MORB)-like amphibolites, that represent metamorphosed early Palaeozoic proto-oceanic crust, led various authors (Gómez-Pugnaire *et al*. 2003; Simancas *et al*. 2005) to consider the shear zone as a 'mini' (cf. Dewey 1977) Variscan suture (Figures 2 and 3).

The contact of the OMZ with the South Portuguese Zone, to the south, is evidenced by the Beja–Acebuches amphibolites and the South Iberia shear zone (Fonseca and Ribeiro 1993; Quesada *et al*. 1994; Araujo *et al*. 2005; Azor *et al*. 2008; Braid *et al*. 2010) (Figures 2 and 3). The amphibolites have been interpreted as Rheic Ocean remnants (e.g. Quesada *et al*. 1994; Castro *et al*. 1996a, 1996b). However, more recently, this idea has been refuted. Azor *et al*. (2008) dated the mafic protolith of the Beja–Acebuches amphibolites at ca. 340–332 Ma, U–Pb SHRIMP, remarkably close to the accepted exhumation age, ca. 330 Ma amphibole, Ar–Ar (Dallmeyer *et al*. 1993; Castro *et al*. 1999). Furthermore, MORB-like metabasalts also crop out in a postulated subduction-related accretionary prism, the Pulo do Lobo unit (e.g. Silva *et al*. 1990; Braid *et al*. 2010). The MORB-like protoliths are the same age as the Beja–Acebuches amphibolites ca. 341–333 Ma (Dahn *et al*. 2014; Pérez-Cáceres *et al*. 2015). This age range, for possible Rheic Ocean remnants, is significantly younger than the date of the initial collision related to final closure of this ocean during the latest Devonian–earliest Carboniferous, ca. 370–355 Ma (cf. Nance *et al*. 2010, 2012; Braid *et al*. 2011; Pereira *et al*. 2012b). The MORB-like magmatism preserved in the OMZ southern contact was, rather, coeval with post-collisional, ca. 350–330 Ma, ultramafic-mafic to intermediate-felsic magmatism in the OMZ (Cambeses *et al*. 2015 and references therein).

Between the northern and southern margins the complex evolution of the OMZ is preserved in structures related to various transpressional and transtensional events from the late Neoproterozoic, Cadomian Orogeny, to the early Carboniferous, Variscan Orogeny (e.g. Abalos *et al*. 1991; Simancas *et al*. 2001; Pereira *et al*. 2003; Silva and Pereira 2004). Pre-Variscan structures were overprinted by the Variscan deformation producing complex structural relationships characteristic of the OMZ (e.g. Apalategui *et al*. 1990; Azor *et al*. 2004). The Variscan structures preserve evidence of an initial collisional event between ca. 390–345 Ma (Simancas *et al*. 2001, 2003; Expósito *et al*. 2002; Braid *et al*. 2011; Pereira *et al*. 2012b). This was followed, *c*.345–330 Ma, by extension/transtension (Apraiz and Eguiluz 2002; Simancas *et al*. 2003; Pereira *et al*. 2007, 2009; Rosas *et al*. 2008) and a subsequent, *c*.330–305 Ma, second collisional event (Simancas *et al*. 2003, 2006; Azor *et al*. 2008).

These events are also recorded in the metamorphism that affected the OMZ, this was generally low-grade although some areas record high grade conditions:

(i) Pre-Variscan ca. 532–500 Ma (Schäfer 1990; Oschner 1993; Ordóñez-Casado 1998; Montero *et al*. 1999, 2000) HT-LP conditions are preserved in the Valuengo and Monesterio complexes (e.g. Apraiz and Eguiluz 1996; Expósito *et al*. 2003; Simancas *et al*. 2004). This high-grade metamorphism has been related to a Cambro-Ordovician rifting context associated with Rheic Ocean opening (e.g. Sánchez-García *et al*. 2003, 2010; Nance *et al*. 2010, 2012).
(ii) Variscan ca. 390–370 Ma (Araujo *et al*. 2005; Moita *et al*. 2005) LT/HT-HP collisional-related metamorphism has been identified in the Badajoz–Cordoba shear zone: the Safira-Viana do Alentejo eclogites; Moura Phyllonitic Complex; and the Cubito-Moura unit (e.g. Fonseca *et al*. 1999; López Sánchez-Vizcaíno *et al*. 2003; Araujo *et al*. 2005; Booth-Rea *et al*. 2006; Ribeiro *et al*. 2007; Rubio-Pascual *et al*. 2013).
(iii) Variscan decompression and shearing metamorphism began later, at ca. 340 Ma, e.g. Campo Maior–Arronches–Crato region (Pereira *et al*. 2010a, 2010b). Whereas younger Variscan ca. 340–323 Ma (e.g. Ordóñez-Casado 1998; Castro *et al*. 1999; Pereira *et al*. 2009) extension/transtension-related HT-LP metamorphism is preserved in the Évora–Aracena–Lora del Rio metamorphic belt (e.g. Chichorro *et al*. 2008; Pereira *et al*. 2009).

The sedimentary, stratigraphic, and igneous, tectonomagmatic, records of crustal evolution of the OMZ are discussed below.

3. Late Neoproterozoic to late Palaeozoic crustal evolution of the Ossa–Morena Zone: stratigraphic record

The OMZ stratigraphic succession includes Ediacaran to early Carboniferous rocks (Figures 3 and 4).

3.1. Precambrian to Cambrian

The OMZ basement is formed of the Ediacaran Serie Negra, which comprises black shales, quartzites, metagreywackes and intercalations of black cherts and

Age	Tectonic context	Representative unit	Lithologies
Upper Carboniferous			
Lower Carboniferous	Syn-orogenic deposits	Santos de Maimona Fm	Metagreywackes, breccias, shales and volcanites
Devonian	Shallow-water series	Verdugo Fm	Shales and greywackes
Silurian	Passive margin deposits	Xistos Raiados Fm	
Ordovician		Barrancos Fm	Shales, sandstones and conglomerates
Upper Cambrian	Rift-drift transition	Volcano-sedimentary succession	Shales and metagreywackes
Middle Cambrian	Main rift-related event		Bimodal volcanism and siliciclastic deposits
Lower-Middle Cambrian		Upper detrital Fm Jerez volcano-detrital Fm	Turbidities, siliciclastic deposits and bimodal volcanism
Lower Cambrian		Carbonate Fm Alconeras Fm	Limestones
	Early rift-related event	Lower detrital Fm Torreárboles Fm	Shales, sandstones, conglomerates and breccias
Ediacaran-Lower Cambrian	Arc-related sucessions	Malococinado Fm	Volcanoclastic sucessions, phyllites, sandstones and conglomerates
Ediacaran	Back-arc basin fill deposits	Serie Negra Tentudía Fm	Black shales, quartzites and metagreywackes amphibolites
	Passive margin sediments	Montemolín Fm	

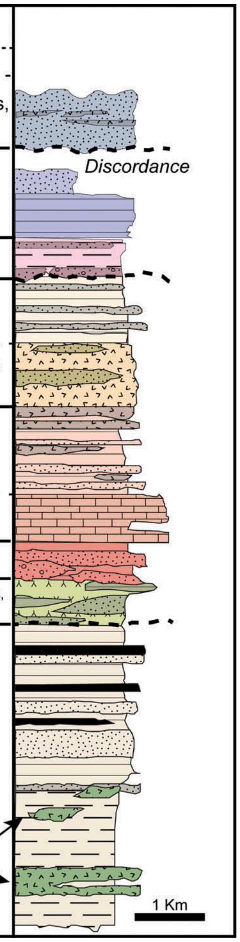

Figure 4. Summary of the Ossa–Morena Zone stratigraphic sequence (modified from Expósito 2000; Robardet and Gutiérrez-Marco 2004; Sánchez-García *et al*. 2010) showing the age, tectonic context, representative units and the lithologies in the area. Dashed lines mark unconformities.

tholeiitic basalts (Figures 3 and 4) (e.g. Alía 1963; Carvalhosa 1965; Gonçalves 1971; Chacón *et al*. 1984; Gonçalves and Oliveira 1986; Oliveira *et al*. 1991; Schäfer *et al*. 1993; López-Guijarro 2006; Sánchez-García *et al*. 2016). The Serie Negra is composed of the Montemolín Formation at the base, formed of passive margin sediments, and the Tentudía Formation at the top, consisting of back-arc basin fill deposits (Eguiluz 1988; Quesada *et al*. 1990a; Bandrés *et al*. 2002). Radiometric U–Pb ages on detrital zircons indicate a maximum sedimentation age of ca. 590 Ma for the Montemolín Formation (Ordóñez-Casado 1998) and ca. 565–541 Ma for the Tentudía Formation (Schäfer *et al*. 1993; Linnemann *et al*. 2008).

The Serie Negra is overlain by the Malcocinado Formation, which is composed of volcanoclastic materials, metagreywackes, phyllites, sandstones and conglomerates, the last with Serie Negra pebbles (Figure 4) (e.g. Fricke 1941; Sánchez-Carretero *et al*. 1989, 1990; Quesada *et al*. 1990a; Pereira and Silva 2002; Perejón *et al*. 2004; Pereira *et al*. 2006). The Malcocinado Formation is locally intruded by late Ediacaran to early Cambrian diorites to granites (e.g. Sanchez-Carretero *et al*. 1989, 1990). The youngest detrital zircon from the Malcocinado Formation indicates a maximum depositional age of ca. 522 Ma (Ordóñez-Casado 1998).

3.2. Cambrian to Ordovician

The Precambrian to Cambrian transition is characterized by an unconformable contact associated with the aforementioned rifting (Figures 3 and 4) (e.g. Liñán 1978, 1984; Liñán and Quesada 1990; Liñan and Gámez-Vitaned, 1993; Expósit-o *et al*. 2003; Sánchez-García *et al*. 2003, 2010). The Cambrian rift-related succession consists of four main components.

The first, the lower detrital formation, is early Cambrian (Liñán and Quesada 1990; Liñán *et al*. 2002; Pereira *et al*. 2011) and is formed of fluvial to shallow marine shelf deposits related to an early transgression, e.g. the Torrearboles Formation (Figure 4, cf.

Liñán 1984). The lower detrital formation is overlain by the second component, an early Cambrian shallow marine carbonate formation, e.g. the Alconera Formation (Figure 4) (Liñán 1978; Liñán and Perejón 1981). This is overlain by a third early to middle Cambrian upper detrital formation that consists of turbiditic and shelf siliciclastic sediments deposited during a collapse-related extensional process, e.g. Jerez volcano-detrital Formation (Figure 4) (Liñán and Quesada 1990; Sánchez-García *et al*. 2003). The uppermost, fourth, part of the formation is a volcano-sedimentary succession dated as middle Cambrian (Figure 4) (Liñán and Quesada 1990; Pereira *et al*. 2006; López-Guijarro *et al*. 2008; Sánchez-García *et al*. 2008). In some units a late Cambrian age was attributed to this succession, e.g. Playon Beds (Palacios 1993). However, this age is poorly constrained, and a middle Cambrian age is generally more accepted for the succession (Figure 4; e.g. Liñán and Quesada 1990; Sánchez-García *et al*. 2003). In fact, a gap in sedimentation took place in the late Cambrian transition as the result of a period in which the whole OMZ was uplifted and eroded related to emplacement of extension-related magmatic bodies (Quesada 1991; Sánchez-García *et al*. 2003, 2010; Quesada *et al*. 2006). Early Cambrian sedimentary sandstone from the lower detrital formation, close to the boundary between the OMZ and the CIZ, has a maximum sedimentation age of ca. 567 Ma for the Oguela Detritic-Carbonate Complex (Linnemann *et al*. 2008). Although Pereira *et al*. (2011) recorded younger ages ca. 536–532 Ma for the Freixo–Segóvia volcanic–sedimentary complex, sandstone, and Ouguela Tectonic Unit, quartzite, from the same region.

Early Ordovician deposits, with a basal erosive discontinuity marked by sandstones and conglomerates, crop out above the Cambrian sedimentary rocks (Figures 3 and 4; e.g. Oliveira 1983; Oliveira *et al*. 1992; Piçarra 1997; 2000; Giese *et al*. 1994a). Higher in the series the Ordovician sedimentary rocks are dominated by shales with intercalations of sandstones and greywackes e.g. Barrancos Formation. (Figures 3 and 4) (e.g. Gutiérrez-Marco *et al*. 1984; Oliveira *et al*. 1992; Giese *et al*. 1994a; Gutiérrez-Marco and Robardet 2004). These sediments are considered to be passive margin deposits (Figure 4) (Robardet 1981; Gutiérrez-Marco *et al*. 1984, 1990, 2002). At the northern contact of the OMZ, a siliciclastic deposit with a Central Iberian Zone palaeogeographic affinity, the Ordovician Armonican Quartzite Formation has a maximum sedimentation age of ca. 522 Ma, (Linnemann *et al*. 2008).

3.3. Silurian to carboniferous

Overlying the Ordovician sedimentary rocks is a monotonous series of Silurian, Landovery to Ludlow, black shales and black cherts (Robardet and Gutiérrez-Marco 1990a, 2004; Gutiérrez-Marco *et al*. 1998). This is conformably covered by late Silurian, Ludlow–Pridoli, to Early Devonian, Lochkovian, black shales with intercalation of lutites and fine sandstones, e.g. Xistos Raiados and Verdugo formations (Figure 4; e.g. Schneider 1951; Jaeger and Robardet 1979; Perdigão *et al*. 1982; Racheboeuf and Robardet 1986; Piçarra 1998; Robardet *et al*. 1998). An ensuing gap in sedimentation occurred in the Middle–Late Devonian, Eifelian–Famennian (Figure 4, e.g. Racheboeuf and Robardet 1986; Robardet and Gutiérrez-Marco 1990a; 1990b; 2004). An unconformity separates the Siluro-Devonian sedimentary rocks from a synorogenic succession of early Carboniferous greywackes, shales, and volcano-sedimentary sequences (Wagner 1978; Wagner *et al*. 1983; Quesada *et al*. 1990b; Giese *et al*. 1994b; Pereira *et al*. 2012b). Early Carboniferous turbidites, Tournaisian–Visean, in the OMZ, Cabrela Formation, have a detrital zircon maximum sedimentation age of ca. 352 Ma (Pereira *et al*. 2012b).

4. Late Neoproterozoic to late Palaeozoic crustal evolution of the Ossa–Morena Zone: tectonomagmatic record

The age of the OMZ igneous rocks ranges from Precambrian to Permian, thus recording the transition from the Cadomian to Variscan orogenies. The OMZ magmatism includes the following:

(i) Neoproterozoic–Cambrian (ca. 590–540 Ma) Cadomian collisional magmatism (Figure 3) (e.g. Oschner 1993; Linnemann *et al*. 2008; Henriques *et al*. 2015; Sánchez-Lorda *et al*. 2016).
(ii) Early–middle Cambrian (ca. 540–500 Ma) continental extension-related magmatism (Figure 3) (e.g. Expósito 2000; Sánchez-García *et al*. 2003, 2010; Simancas *et al*. 2004), which progressed during the Cambro-Ordovician (ca. 490–470 Ma) to incipient ocean basin magmatism (Figure 3) (e.g. Quesada 1991; Sánchez-García *et al*. 2003, 2016; Chichorro *et al*. 2008).
(iii) Carboniferous (ca. 350–330 Ma) Variscan collision-related magmatism (Figure 3) (e.g. Dallmeyer *et al*. 1995; Ordóñez-Casado 1998;

Montero *et al*. 2000; Gladney *et al*. 2014; Cambeses *et al*. 2015; Pereira *et al*. 2015).

The Neoproterozoic–Cambrian, ca. 590–540 Ma, magmatism mainly comprises subduction-related diorites to granites and basalts, the latter are preserved as metabasite amphibolites (Bellon *et al*. 1979; Schäfer 1990; Oschner 1993; Henriques *et al*. 2015). This magmatism shows south to north N-MORB to calc-alkaline magmatic polarity in the OMZ which led Sánchez-Lorda *et al*. (2013a, 2013b) to postulate Gondwana-ward subduction at that time.

During the early Cambrian magmatism changed from collision- to extension-related (Sánchez-García *et al*. 2013). The last Cadomian arc manifestation is represented by the I-type Culebrin tonalite 532 ± 4 Ma (Figures 5 and 6; Montero *et al*. 2000), which is roughly coeval with peraluminous granitoids produced during initiation of extension, the so-called 'early rift-related event' (e.g. Galindo 1989; Ochsner 1993; Galindo and Casquet 2004; Sánchez-García *et al*. 2010; 2013). This magmatism was followed by A_2-type granites (Figures 5 and 6), such as the Calera de León granite 524 ± 4 Ma (Montero *et al*. 2000), associated with the initial extension in the region (Sánchez-García *et al*. 2003; 2013).

The ca. 520–500 Ma 'main rift-related event' (Sánchez-García *et al*. 2003, 2016) produced abundant plutonic and volcanic rocks. The earliest magmatism associated with this phase is E-MORB-like mafic lavas and tuffs with ages of ca. 517–512 Ma (Figure 7; Sánchez-García *et al*. 2008; 2010). This was followed by widespread OIB-like, transitional alkaline to tholeiitic, volcanic, and plutonic mafic magmatism that occurred from ca. 512 to 505 Ma (Figure 7; Galindo *et al*. 1990, Sánchez-García *et al*. 2010). This mafic magmatism has εNd_t from 5.6 to 1.5 (Figure 8(b); Sánchez-García *et al*. 2010; Sarrionandia *et al*. 2012). It apparently resulted from asthenospheric upwelling and proto-ocean basin development with variable amounts of crustal contamination (Ordóñez-Casado 1998; Chichorro *et al*. 2008). The mafic magmatism was coeval with emplacement of crustally-derived anatectic peraluminous intermediate to felsic granitoids such as the Monesterio granodiorite (510 ± 4 Ma, Montero *et al*. 1999); mantle-derived extensional peralkaline A-type complexes such as the bimodal Barcarrota plutonic complex (505 ± 5 Ma Rb–Sr WR, Galindo *et al*. 1990; 501 + 5-2 Ma U–Pb TIMS; 501 + 5–2 Ma U–Pb TIMS, Oschner 1993); and the felsic Castillo granite (502 ± 8 Ma KOBER method, Montero *et al*. 1999) (Figures 5 and 6). The A-type magmatism also reflects the presence of a crustal component (Figures 5 and 6) e.g. the Évora orthogneiss (517 ± 15 and 505 ± 5 Ma, U–Pb SHRIMP, Chichorro *et al*. 2008).

From ca. 490 to 470 Ma, the Cambro-Ordovician rifting produced T- and N-MORB-like mafic rocks (Figure 7), with εNd_t from 10.9 to 1.4 with a main cluster at 6–7.5 (Figure 8(c); Ordóñez-Casado 1998; Gomez-Pugnaire *et al*. 2003; Sola, 2007; Chichorro *et al*. 2008). These rocks are notably more primitive than the ca. 520–500 Ma mafic magmatism (Figure 8(b)) (Sánchez-García *et al*. 2010; Sarrionandia *et al*. 2012). Felsic magmatism during the ca. 490–470 Ma interval was also extension-related: peralkaline anorogenic and peraluminous anatectic (Figures 5 and 6) (Sánchez-Carretero *et al*. 1999; Solá *et al*. 2008; Díez-Fernández *et al*. 2014). The former includes peralkaline and alkaline A-type orthogneisses close to the northern boundary of the OMZ (ca. 490–470 Ma, LA-ICP-MS, Díez-Fernández *et al*. 2014). The most important manifestation of peraluminous crustal-derived magmatism is the volcanic and subvolcanic Urra Formation (488 ± 5 Ma U–Pb SHRIMP Solá *et al*. 2008). Mixing of mantle- and crust-derived magmas occurred in this period (Figures 5 and 6). The Carrascal granite (486 ± 7 Ma, U–Pb LA-ICP-MS, Solá 2007) and the Ribera del Fresno and Las Minillas orthogneisses (473 + 2–3 Ma, U–Pb TIMS, Schäfer unpublished data in Oschner 1993), for example, are interpreted as crustally contaminated alkaline magmas (Figure 6(e)) (Oschner 1993; Solá 2007).

At some point after ca. 450 Ma, ocean opening was arrested in the OMZ prior to broad ocean basin development. Then, before ca. 390 Ma, regional rift-related extension changed to collision (Simancas *et al*. 2001; Braid *et al*. 2011; Pereira *et al*. 2012b). Carboniferous Variscan collision-related magmatism comprises ultrabasic to acid, metaluminous alkaline to calc-alkaline plutons and peraluminous dykes (e.g. Montero *et al*. 2000; Casquet and Galindo 2004; Moita *et al*. 2009; Pereira *et al*. 2009; Gladney *et al*. 2014; Cambeses *et al*. 2015, 2015).

5. Isotopic correlation of the Ossa–Morena Zone Cambro-Ordovician palaeogeographic position

With the aim of determining the Cambro-Ordovician palaeogeographic position of the OMZ we combine data from various geochronological data sets (see supplementary material I and II):

(i) The Nd model ages of: OMZ Cambro-Ordovician igneous rocks; OMZ Ediacaran to Ordovician sedimentary rock; and, also, Ediacaran to Cambrian sedimentary rocks from the other Iberian zones (Figure 8).
(ii) The Nd model ages of potential sources for the OMZ Ediacaran to Ordovician sediments, specifically Pan-African granitoids from the three North Gondwana, African, areas: the Anti-Atlas West

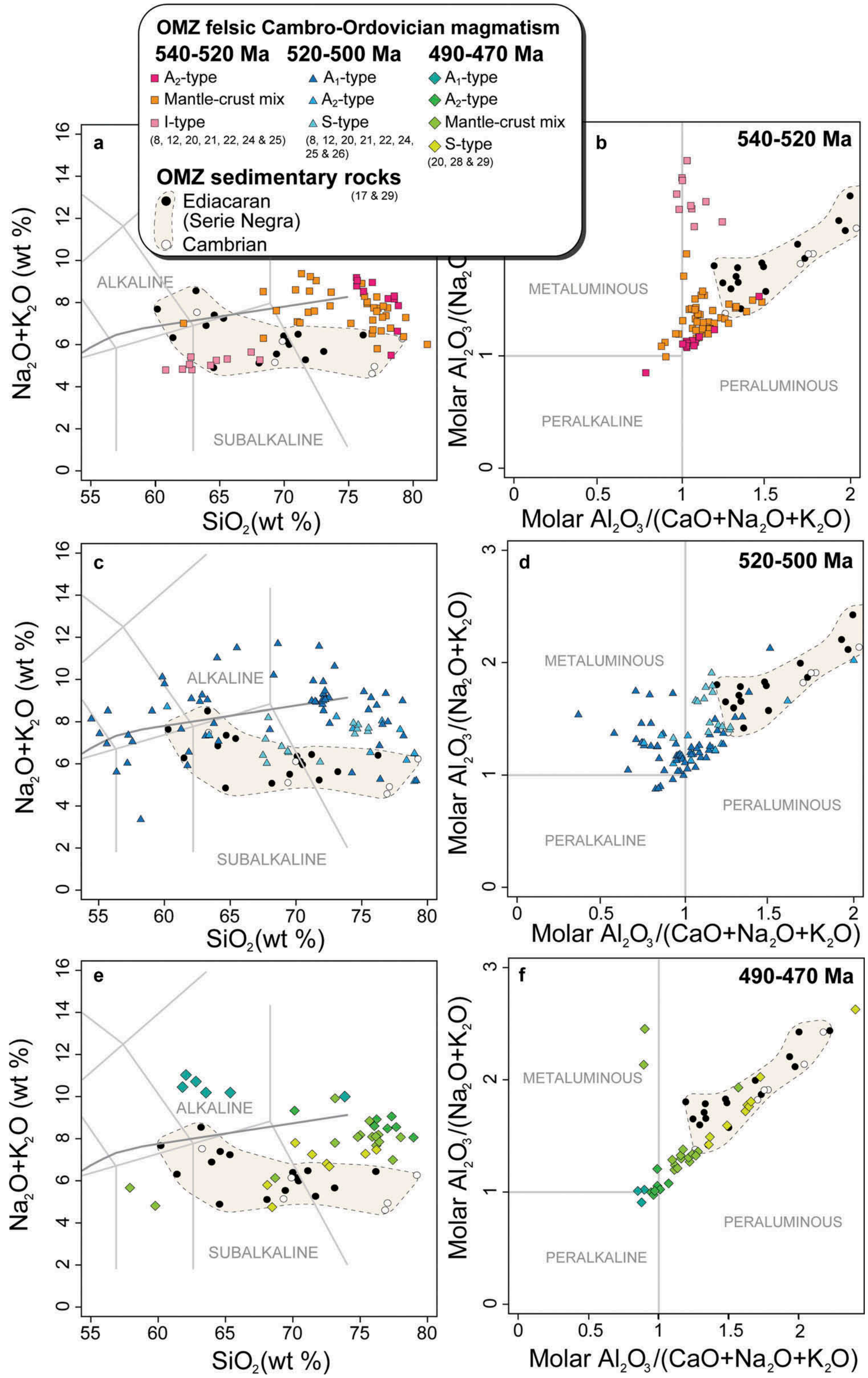

Figure 5. Whole-rock major element composition of the ca. 540–520 Ma, ca. 520–500 Ma, and ca. 490–470 Ma Ossa–Morena Zone (OMZ) igneous rocks and OMZ sedimentary rocks from Serie Negra and Cambrian units. (a, c, e) TAS plots, note the discrimination of alkaline and sub-alkaline compositions (fields from Le Maitre 1989). (b, d, f) Molar ($Al_2O_3/(Na_2O+K_2O)$) vs. molar ($Al_2O_3/(CaO+Na_2O+K_2O)$) plots show the compositional variation of metaluminous, peralkaline and peraluminous samples (fields from Shand 1947). Numbers in brackets correspond to the data sources in supplementary material I.

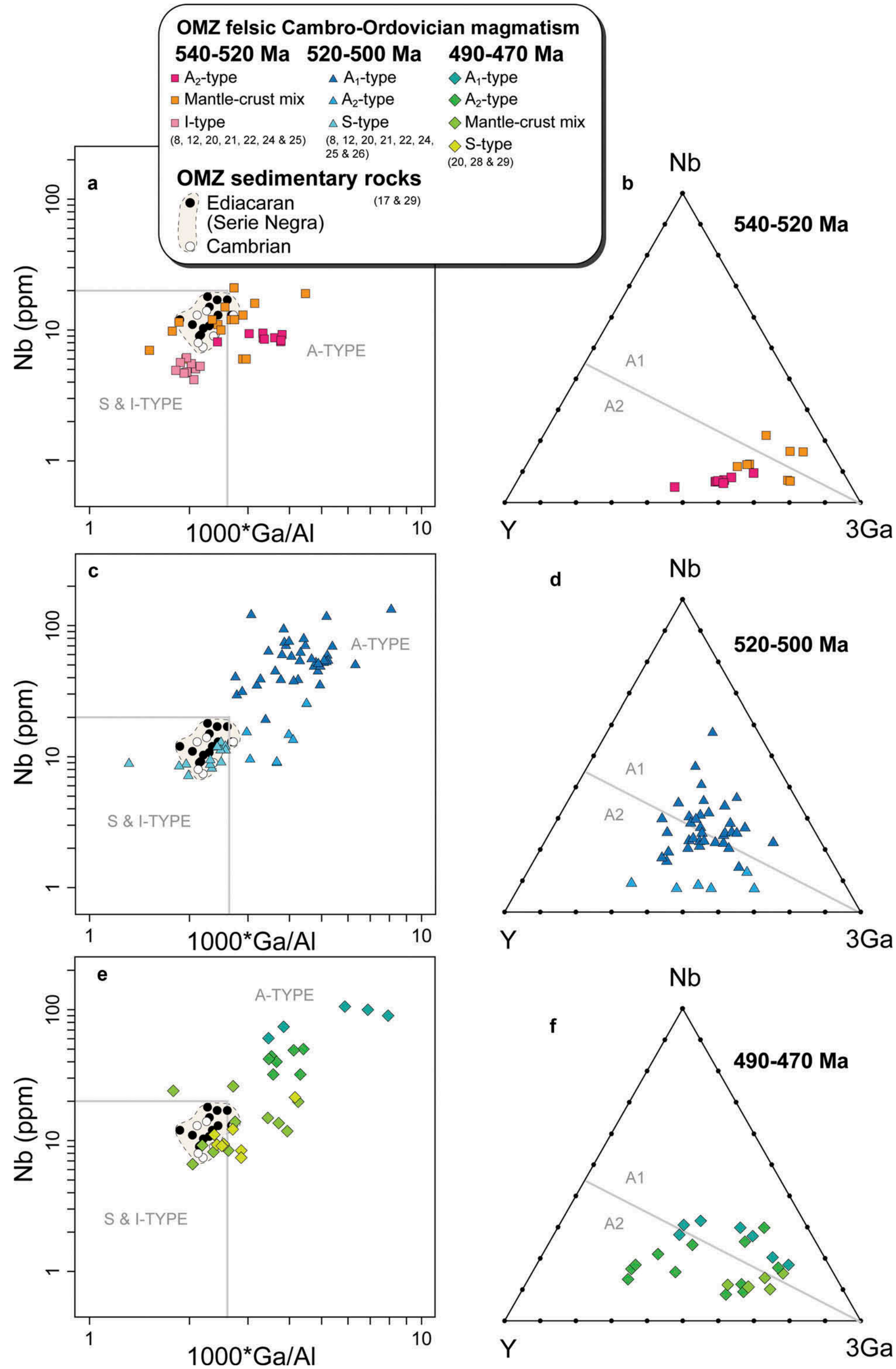

Figure 6. Granitoid discrimination diagrams. (a, c, e) The Nb vs. 1000*Ga/Al (Whalen *et al*. 1987) and (b, d, f) the Nb-Y-3 Ga (Eby 1992) show the ca. 540–520 Ma, ca. 520–500 Ma, and ca. 490–470 Ma Ossa–Morena Zone igneous rocks and sedimentary rocks from Serie Negra and Cambrian units. Numbers in brackets correspond to the data sources in supplementary material I.

African Craton; the Tuareg Shield; and the Sahara Metacraton (Figure 8).

(iii) Single and population detrital zircon ages of Ediacaran to Ordovician sedimentary rocks from the OMZ, other Iberian zones (Figure 9) and the three aforementioned North Gondwana, African, areas (Figures 10 and 11).

(iv) Single and population pre-magmatic and magmatic zircon and whole-rock ages of Ediacaran to Ordovician igneous rocks from the OMZ,

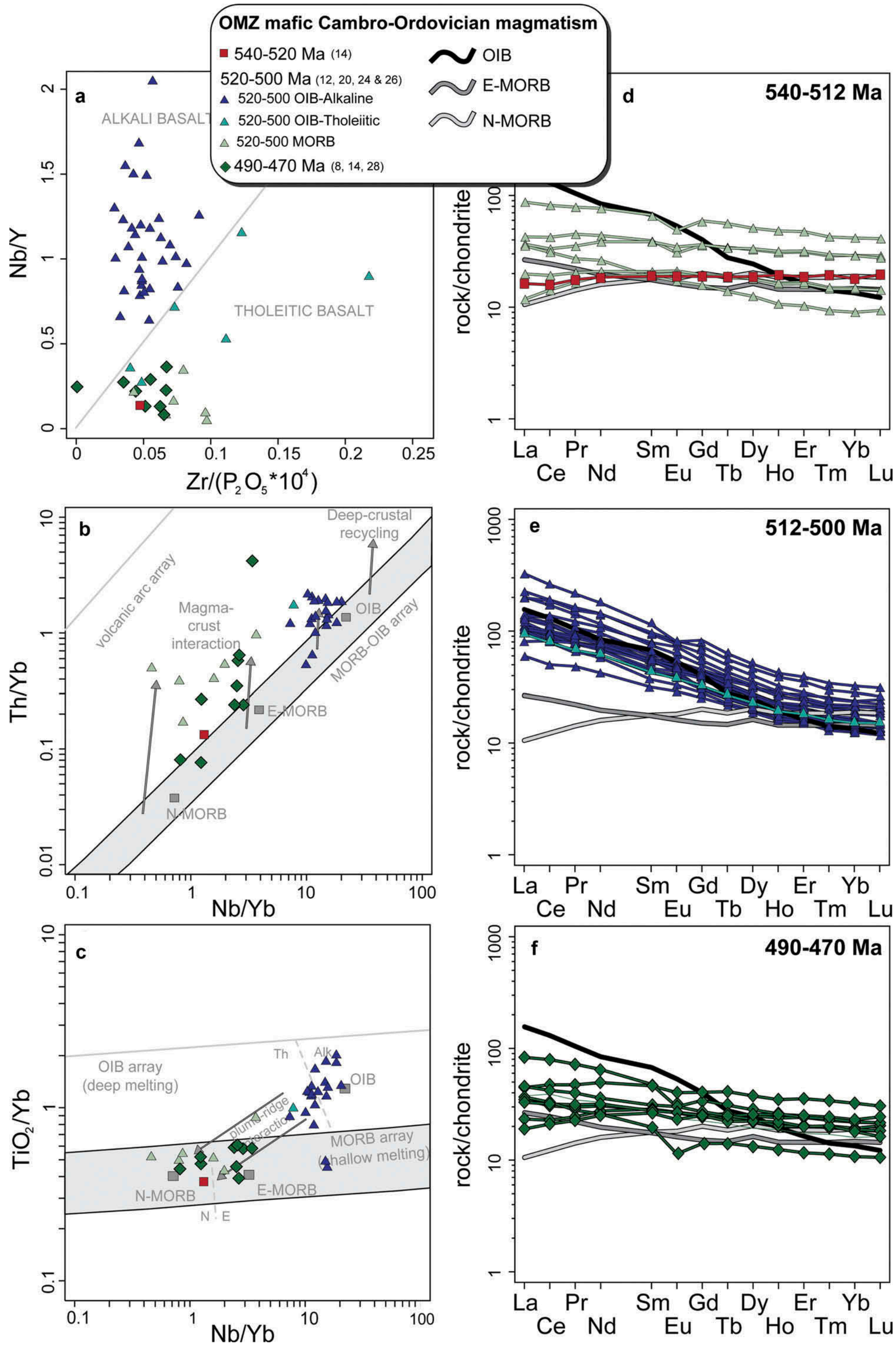

Figure 7. Whole-rock minor and trace element diagrams for the Ossa–Morena Zone (OMZ) mafic igneous rocks ca. 540–512 Ma, ca. 512–500 Ma, and ca. 490–470 Ma. (a) Alkaline and tholeiitic compositions (after Floyd and Winchester 1975), (b, c) Mantle source affinity of each group in Pearce (2008) plots. Chondrite-nomalized REE plots of (d) ca. 540–512 Ma E-MORB-like rocks, (e) ca. 512–500 Ma alkaline/tholeiitic OIB rocks, and (f) ca. 490–470 Ma T- and N-MORB rocks. Normalization values after McDonough and Sun (1995), the reference mantle source patterns from Sun and McDonough (1989) are shown in(d)–(f). Numbers in brackets correspond to the data sources in supplementary material I.

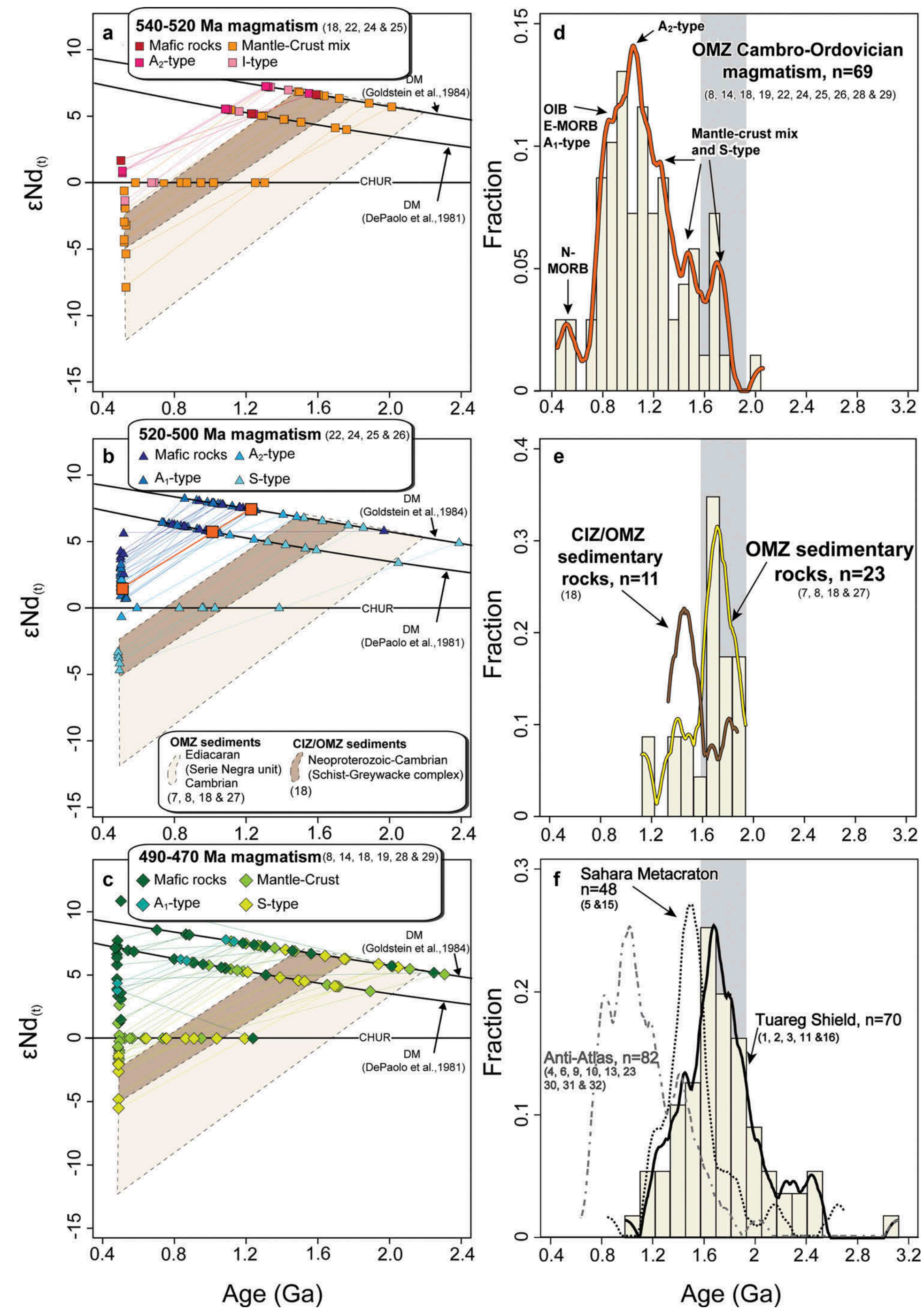

Figure 8. (a)–(c) Nd isotope compositions of ca. 540–520 Ma, ca. 520–500 Ma, and ca. 490–470 Ma Ossa–Morena Zone (OMZ) Cambro-Ordovician igneous rocks. In addition we include Nd isotopic data for OMZ basement metasedimentary rocks from the Ediacaran Serie Negra and Cambro-Ordovician units as well as OMZ-Central Iberian Zone (CIZ) boundary metasediments from the Ediacaran Schist-Greywacke Complex and Cambrian units. For all data, the Nd model ages were calculated based on the method of DePaolo (1981) (T_{DM}) and Goldstein *et al.* (1984) (T_{CR}). This summary of Nd model (T_{DM}) age distributions includes: (d) Histogram of Cambro-Ordovician OMZ igneous rocks, note the variation in age according to mantle or crustal character; (e) Histogram of the OMZ metasedimentary rocks, note the main peak at ca. 1.7 Ga similar to crust-derived or -contaminated igneous rocks in (d), the OMZ/CIZ boundary metasediments are shown for comparison; and (f) Histogram of the Tuareg Shield Pan-African granitoids, a potential source of OMZ sedimentary rocks (solid line and thick grey vertical band in (d)–(f)) and the other north African regions: The Anti-Atlas West African Craton (dashed line); and the Sahara Metacraton (dotted line). Notably the Tuareg Shield also has a main peak at ca. 1.7 Ga as detected in the OMZ basement in (e). Numbers in brackets correspond to the data sources in supplementary material I.

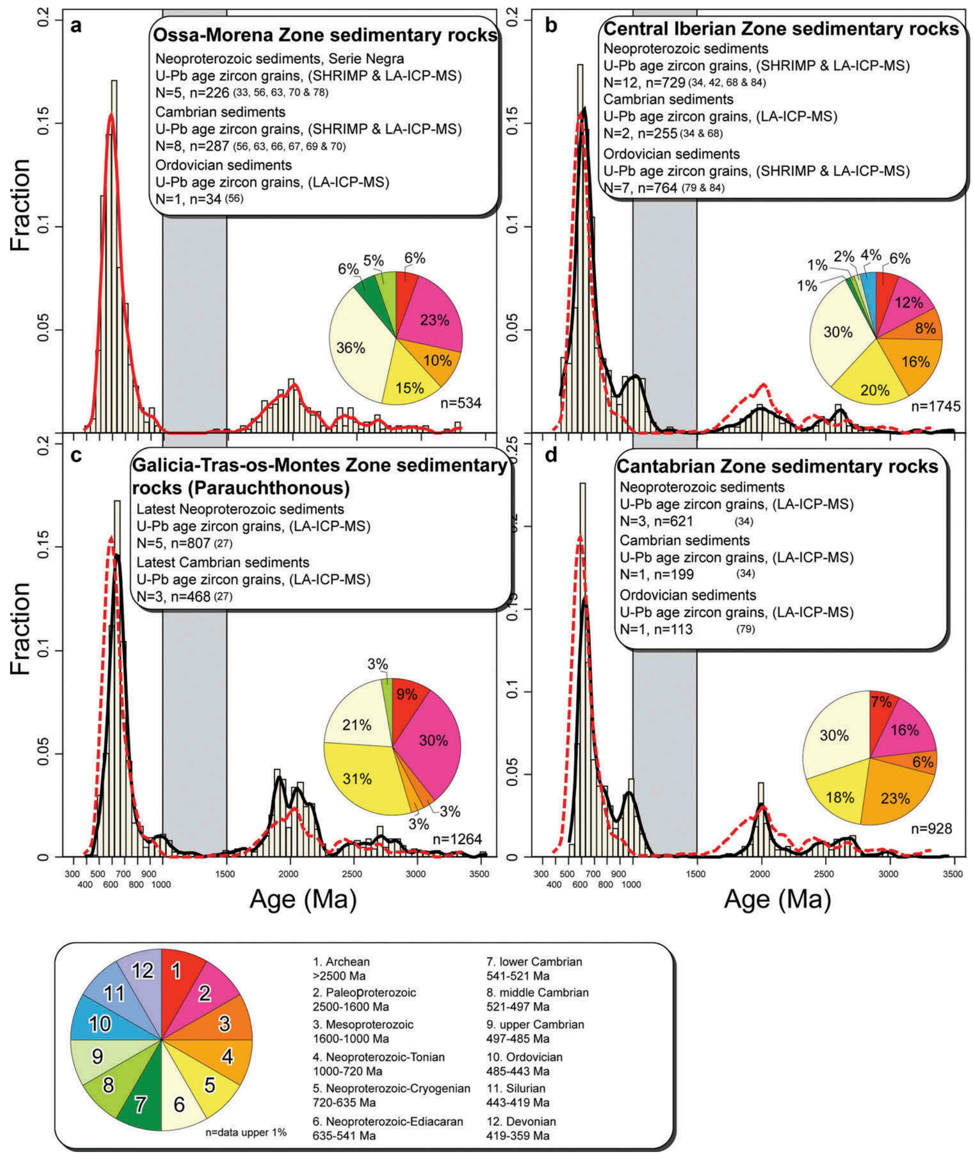

Figure 9. Frequency and density distribution of U–Pb SHRIMP and LA-ICP-MS age data for detrital zircons from: (a) Ossa–Morena Zone (OMZ) Neoproterozoic and Cambro-Ordovician sedimentary rocks. (b) Central Iberian Zone Neoproterozoic and Cambro-Ordovician sedimentary rocks. (c) Galicia-Tras-os-Montes Zone parautochthonous units latest Ediacaran to latest Cambrian sedimentary rocks. (d) Cantabrian Zone Neoproterozoic and Cambro-Ordovician rocks. To facilitate comparison all plots include the kernel density distribution of the OMZ ages (thick red/grey line), the Mezoproterozic age range (thick grey vertical band) and pie diagram. Note that Mesoproterozoic zircons are abundant in the Central Iberian Zone and Cantabrian Zone, less abundant in the parautochthonous units from Galicia-Tras-os-Montes Zone and absent in the OMZ. Numbers in brackets correspond to the data sources in supplementary material II N corresponds to individual samples and n with zircon ages.

other Iberian Massif zones (Figure 12) and the three North Gondwana, African, areas (Figure 13).

5.1. Nd model ages

5.1.1. OMZ Cambro-Ordovician igneous rocks

The Nd model ages of the OMZ Cambro-Ordovician igneous rocks range from ca. 2.0–0.5 Ga, with main clusters at ca. 1.7 Ga, ca. 1.5 Ga, ca. 1.3 Ga, ca. 1.1 Ga and ca. 0.5 Ga (Figure 8(a–d)).

Mafic rocks may be divided into two main age groups; on the one hand the rocks with the youngest Nd model ages, ca. 0.6–0.5 Ga (Figure 8(d)), close to the ca. 480 Ma crystallization age, are those with primitive tholeiitic N-MORB compositions (Figure 8 (c)). By contrast, igneous rocks with E-MORB and OIB compositions, with an apparently variable crustal input, have older Nd model ages in the range ca. 1.2–0.8 Ga (Figure 8(d)).

Contemporaneous felsic igneous rocks fall into three age groups. The first group are A-type peralkaline–

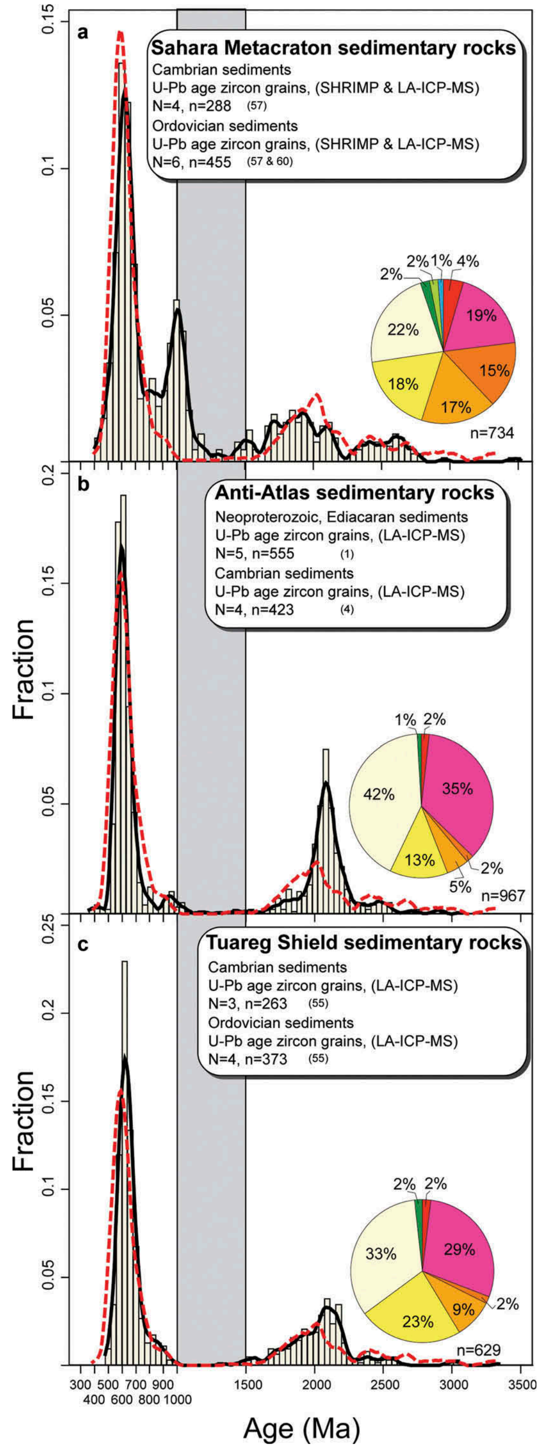

Figure 10. Frequency and density distribution of U–Pb SHRIMP and LA-ICP-MS age data for detrital zircons from; (a) Anti-Atlas West African Craton, Neoproterozoic to Cambrian sedimentary rocks. (b) Sahara Metacraton Cambro-Ordovician sedimentary rocks and (c) Tuareg Shield Cambro-Ordovician rocks. To facilitate comparison, all plots include the kernel density distribution of the Ossa–Morena Zone (OMZ) ages (thick red/grey line). Mezoproterozic age range (thick grey vertical band) and pie diagram as in Figure 9. Note the coincidence of the OMZ and Tuareg Shield patterns, in both Mezoproterozic ages are absent. Numbers in brackets correspond to the data sources in supplementary material II. N corresponds to individual samples and n with zircon ages.

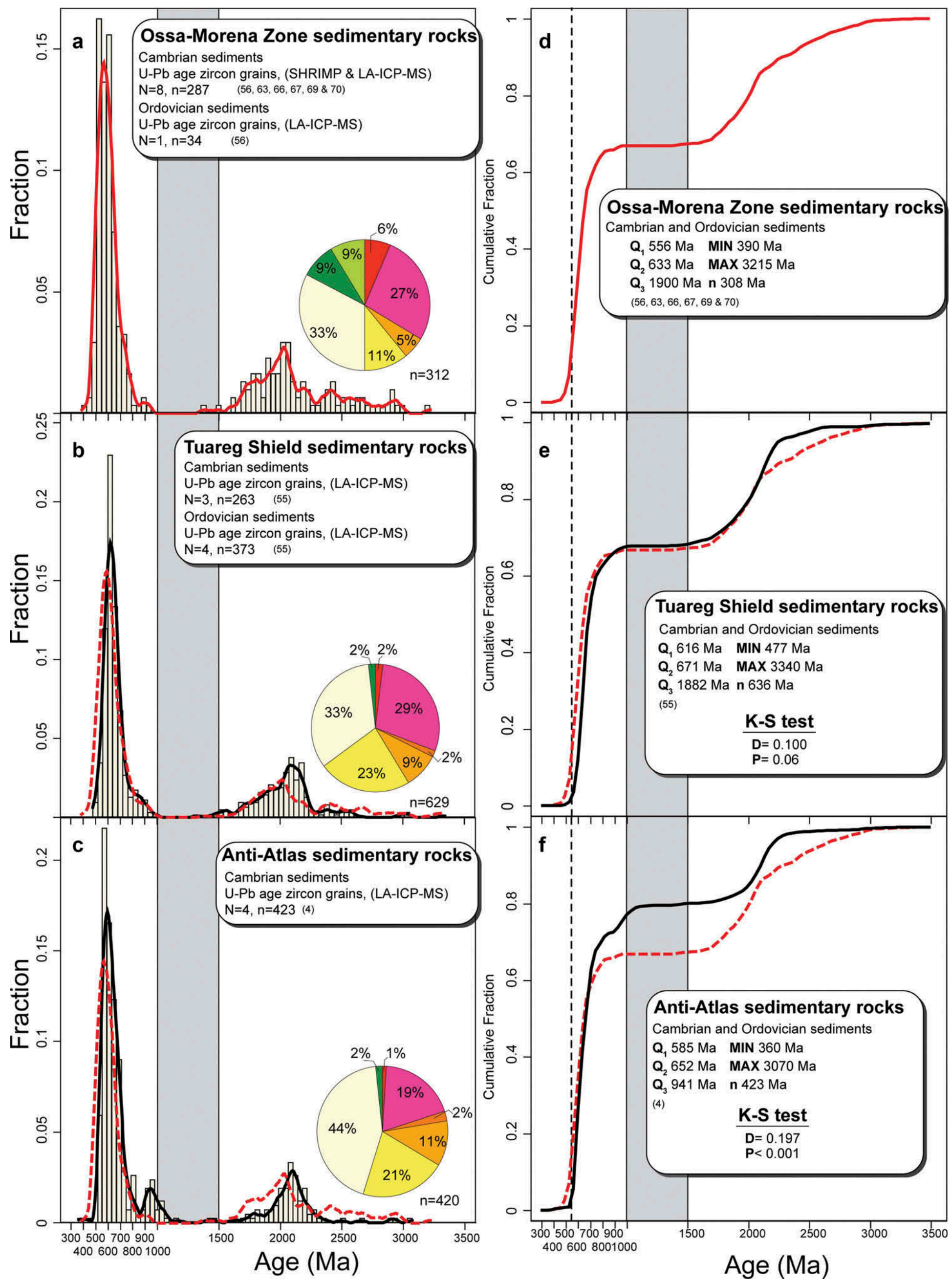

Figure 11. Frequency and density distribution of U–Pb SHRIMP and LA-ICP-MS age data for detrital zircons from: (a) Ossa–Morena Zone (OMZ) Cambro-Ordovician sedimentary rocks. (b) Tuareg Shield Cambro-Ordovician sedimentary rocks. (c) Anti-Atlas West African Craton, Cambrian sedimentary rocks. Cumulative fraction diagrams from the same rocks of: (d) OMZ, (e) Tuareg Shield, and (f) Anti-Atlas West African Craton. To facilitate comparison, all plots include the kernel density distribution of the OMZ ages in Figure (b) to (c) and cumulative fraction in Figure (e) and (f) (thick red/grey line), the Mesoproterozic age range (thick grey vertical band) and pie diagram as in Figure 9. Statistical parameters for zircon population and the Kolmogorov–Smirnov test (K-S) results for the OMZ and the Tuareg Shield and Anti-Atlas sedimentary rocks in Figure d, e and f. Note the coincidence of the OMZ and Tuareg Shield patterns. Numbers in brackets correspond to the data sources in supplementary material II. N corresponds to individual samples and n with zircon ages.

alkaline rocks with a Nd model age of ca. 0.8 Ga (Figure 8(d)), e.g. the Almendral granite and Aceuchal–Almendradejo orthogneiss (Oschner 1993; Casquet and Galindo 2004). The second group of rocks have an older ca. 1.0 Ga Nd model age (Figure 8(d)). This age, it seems, is mixed, indicated by a transitional composition from A-type *ss.* to OMZ basement (Figure 8), e.g. the Castillo granite (Oschner 1993; Salman 2002) and the Loma del Aire unit (Sánchez-García *et al.* 2016). The last group contains coeval crust-derived and strongly crustally

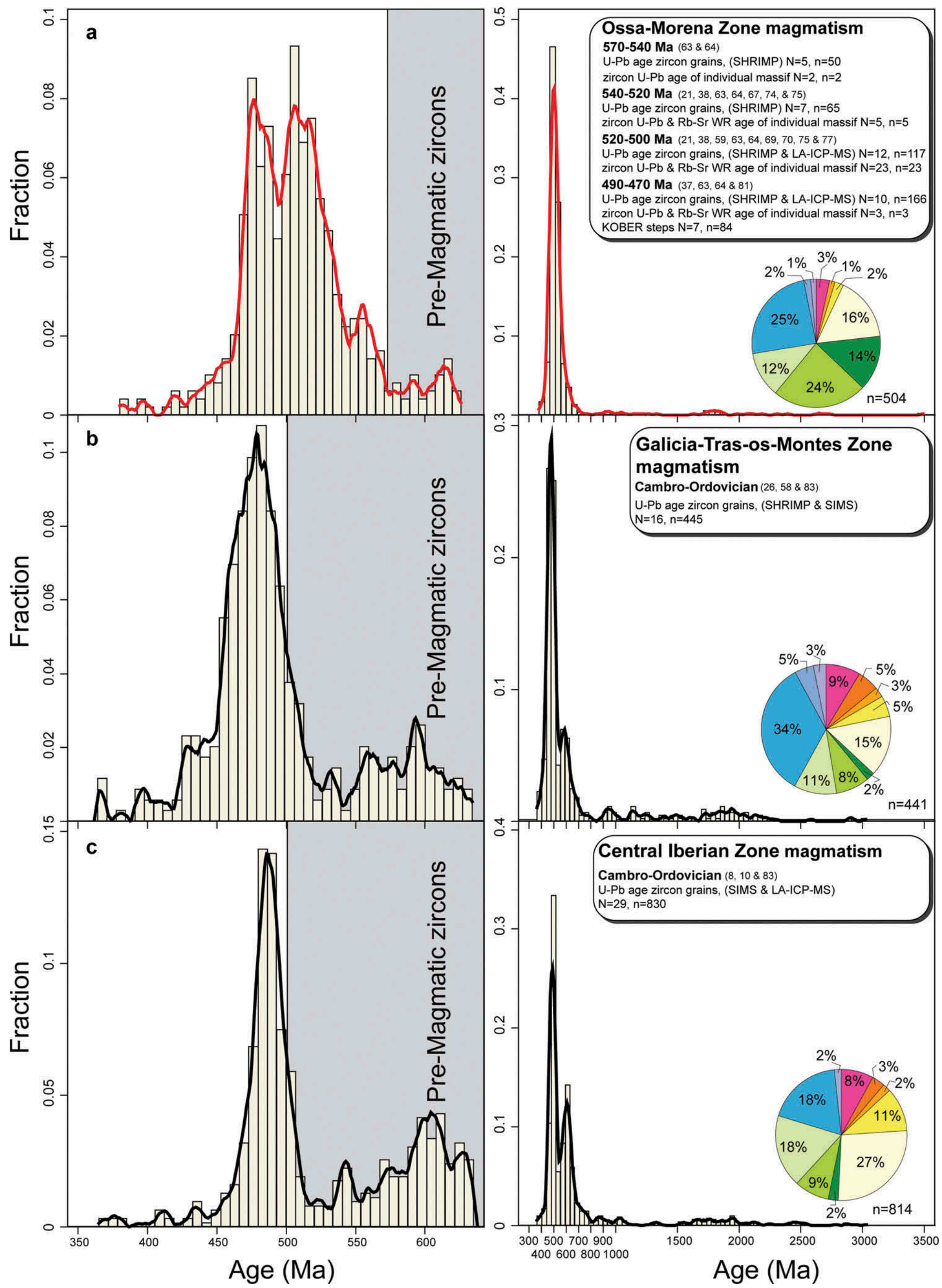

Figure 12. Frequency and density distribution of U–Pb ion microprobe, LA-ICP-MS and whole-rock ages of igneous rocks from: (a) Ossa–Morena Zone Cambro-Ordovician rocks, note the three main magmatic peaks at ca. 540 Ma, ca. 512 Ma, and ca. 480 Ma. Ages older than ca. 570 Ma correspond to pre-magmatic zircons. (b) Galicia-Tras-os-Montes Zone parautochthonous units with a Cambro-Ordovician magmatic event with an age range of ca. 490–470 Ma centred at ca. 480 Ma. Note the abundant pre-magmatic zircons at ca. 615 Ma. (c) Central Iberian Zone Cambro-Ordovician igneous rocks, note the main ca. 480 Ma peak and the extraordinary abundance of pre-magmatic, here ages older than ca. 490 Ma are pre-magmatic, centred at ca. 615 Ma. Pie diagram age sections as in Figure 9. Numbers in brackets correspond to the data sources in supplementary material II. N corresponds to individual samples and n with whole-rock and zircon ages.

contaminated OMZ magmatism with older Nd model ages still, of ca. 2.0–1.3 Ga, with clusters at ca. 1.7 Ga, ca. 1.5 Ga, and ca. 1.3 Ga (Figure 8(d)), e.g. Monesterio anatectic dome (Salman 2002) and Urra volcano–sedimentary formation (Solá *et al*. 2008).

5.1.2. OMZ Ediacaran–Ordovician sedimentary rocks

The OMZ Ediacaran to Ordovician sedimentary rocks have an Nd model age range of ca. 1.9–1.1 Ga (Figure 8(e)) (Schäfer *et al*. 1993; Mullane 1998; cited in Fernández-Suarez *et al*. 2002; López-Guijarro *et al*. 2008).

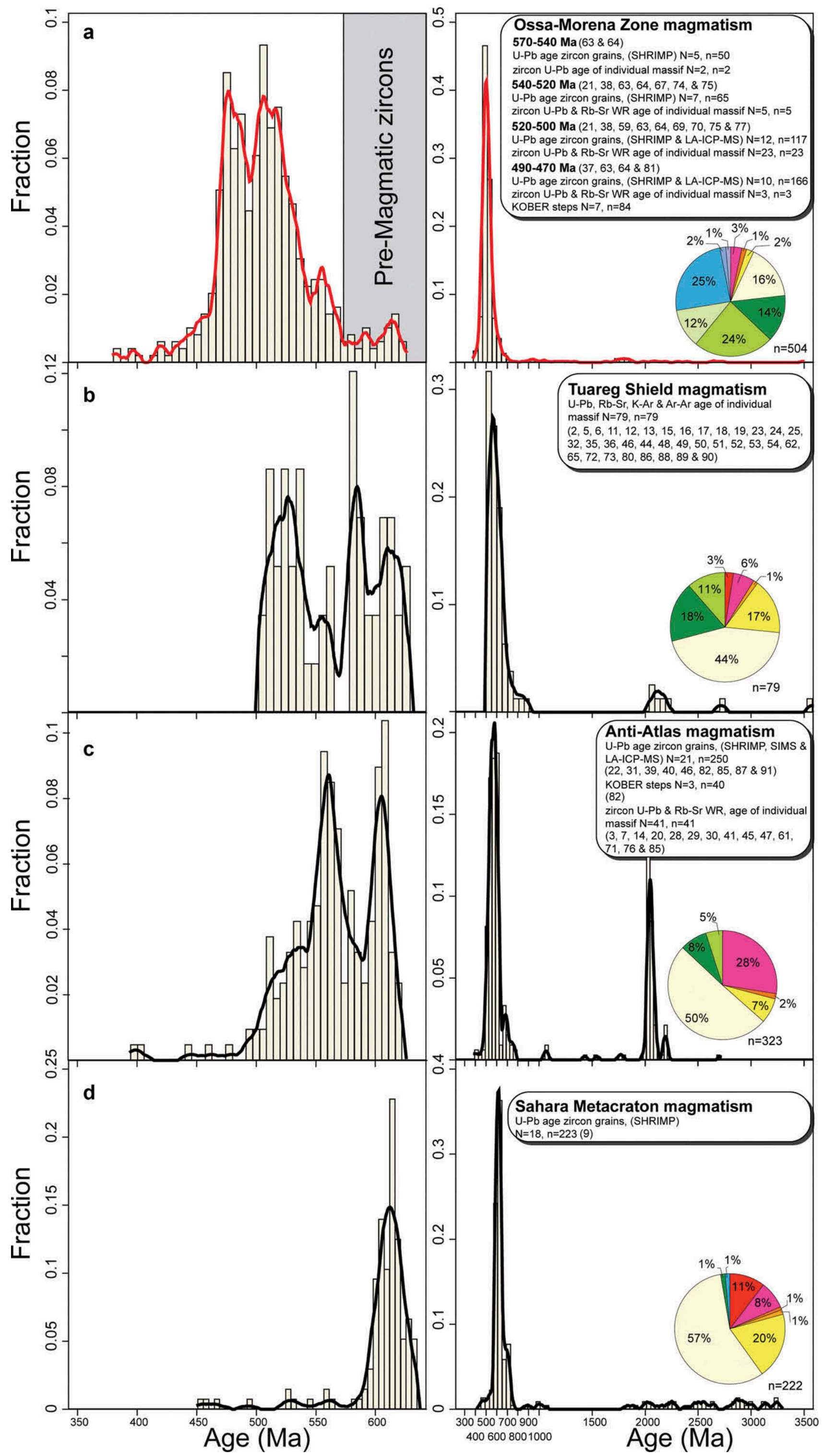

Figure 13. Frequency and density distribution of U–Pb ion microprobe, LA-ICP-MS and whole-rock ages of igneous rocks from: (a) Ossa–Morena Zone Cambro-Ordovician rocks, note the three main magmatic peaks at ca. 540 Ma, ca. 512 Ma, and ca. 480 Ma. Ages older than ca. 570 Ma correspond to pre-magmatic zircons. change to: (b)Tuareg Shield granitoids whole-rock ages with a main peak at ca. 525 Ma, here all ages are magmatic. (c) Anti-Atlas West African Craton granitoids with a main cluster at ca. 570 Ma and ca. 600 Ma, note the asymmetric distribution and that the ca. 480 Ma magmatic event is scarcely represented. (d) Sahara Metacraton granitoids, migmatites, and orthogneisses ages, note the peak at ca. 615 Ma. Pie diagrams age periods as in Figure 9. Numbers in brackets correspond to the data sources in suplementary material II. N corresponds to individual samples and n with whole-rock and zircon ages.

The Ediacaran Serie Negra has an older Nd model age range of ca. 1.9–1.7 Ga, with a mode of ca. 1.7 Ga (Schäfer *et al*. 1993; Casquet *et al*. 2001; Chichorro *et al*. 2008; López-Guijarro *et al*. 2008). Passive margin and back-arc basin contexts, both of which are consistent with an input of continental crust, have been suggested (Eguiluz 1988; Quesada 1991) for the formation of these sedimentary rocks with an old crustal signature (López-Guijarro *et al*. 2008). The Nd model age range from the base to the top of the OMZ Ediacaran–early Cambrian Malcocinado Formation is ca. 1.6–1.1 Ga (López-Guijarro *et al*. 2008). The younger Nd model ages of the Cambrian sedimentary rocks are a consequence of progressive increase in input of juvenile crust or mantle material associated with arc-related magmatism followed by rifting (López-Guijarro *et al*. 2008). Ordovician–Early Devonian OMZ sedimentary rocks, by contrast, have a Nd model age range of ca. 1.8–1.6 Ga, related to continental crust input in a passive margin context (López-Guijarro *et al*. 2008).

As noted above, some of the OMZ Cambro-Ordovician mafic and felsic igneous rocks have an Nd model age that corresponds to a mixture between old material derived from the OMZ basement and young mantle-derived magmatism, giving a composite main Nd model age cluster at ca. 1.0 Ga (Figure 8(d)). Therefore, the true Nd model age of the OMZ sediment end-member should be older than ca. 1.0 Ga. So, of the OMZ Ediacaran to Ordovician sedimentary rocks, the best estimate Nd model age for the OMZ sediment source may be the Serie Negra or the Ordovician sedimentary rocks which have a range of ca. 1.9–1.6 Ga and a mode at ca. 1.7 Ga (Figure 8(e)). Significantly, the Cambro-Ordovician crust-derived magmatism has a similar Nd model age range ca. 1.9–1.5 Ga with mode of ca. 1.7 Ga (Figure 8(d)).

5.1.3. The other Iberian Massif zones Ediacaran–Ordovician rocks

Neoproterozoic to Ordovician Central Iberian Zone and Cantabrian Zone sedimentary rocks have Nd model ages in the range ca. 2.0–1.3 Ga similar to comparable age rocks in the OMZ, but with a younger cluster at ca. 1.5 Ga (Nägler *et al*. 1995; Fernández-Suarez *et al*. 1998; Gutiérrez-Alonso *et al*. 2003; Bea *et al*. 2010; Pastor-Galán *et al*. 2013b; Villaseca *et al*. 2014; Rubio-Ordóñez *et al*. 2015; Fuenlabrada *et al*. 2016; Ugidos *et al*. 2016). Furthermore, the Cambro-Ordovician crust-derived, Central Iberian Zone Ollo de Sapo orthogneisses also have a Nd model age mode of ca. 1.5 Ga (Bea *et al*. 2007, 2010; Montero *et al*. 2007; Talavera *et al*. 2013). Even the southerly Central Iberian Zone Neoproterozoic Greywacke Schist Complex and Cambro-Ordovician sedimentary rocks just to the north of the boundary with the OMZ have an Nd model age mode of ca. 1.5 Ga, within a range ca. 1.9–1.3 Ga (Figure 8(e)) (López-Guijarro *et al*. 2008).

5.1.4. North Gondwana, African, basement granitoids

Bea *et al*. (2010) summarized published Nd model ages of Pan-African granitoids from the three main north African Gondwana areas: the Sahara Metacraton, the Anti-Atlas West African Craton, and the Tuareg Shield. These authors considered the Pan-African granitoids as a potential source for the Central Iberian Zone Ediacaran to Cambrian sedimentary rocks. Similarly, we use the Nd model age data collated in the present work to identify potential North Gondwana African sources of the OMZ Ediacaran to Ordovician sedimentary rocks (supplementary material I).

The Sahara Metacraton Pan-African granitoids have Nd model ages with a mode of ca. 1.5 Ga in a range of ca. 2.3–1.1 Ga and a minor older cluster at ca. 2.7 Ga (Figure 8(f)) (Harms *et al*. 1990; Bea *et al*. 2010, 2011). The Anti-Atlas West African Craton Pan-African granitoids have an asymmetric Nd model age distribution with a continuous range from 1.8–0.8 Ga with a cluster at ca. 1.0 Ga and small populations at ca. 2.2, ca. 2.7 and ca. 3.0 Ga (Figure 8(f)) (e.g. Blanc *et al*. 1992; Gasquet *et al*. 2005; Tahiri *et al*. 2010). The Tuareg Shield Pan-African granitoids have Nd model ages from ca. 2.5–1.0 Ga with a mode of ca. 1.7 Ga and a minor cluster at ca. 3.1 Ga (Figure 8(f)) (e.g. Liégios *et al*. 1994; Ferré *et al*. 1996; Abdallah *et al*. 2007).

5.1.5. Comparison of the OMZ Nd model ages with other Iberian Massif zones and the North Gondwana, African, potential sources

It is worth mentioning that like Bea *et al*. (2010), we rule out the Arabian Nubian Shield as a possible OMZ sediment source from the outset because its Nd model age is too young, ca. 0.74 Ga (Stern 2002; Moreno *et al*. 2014). In addition, we also rule out the regions of the Sahara Metacraton, West African Craton and Tuareg Shield with crustal signatures older than ca. 2.5 Ga, because they are scarce and older than the OMZ main cluster ca. 1.7 Ga (Peucat *et al*. 1996, 2003, 2005; Barth *et al*. 2002; Bea *et al*. 2014).

The Sahara Metacraton Pan-African granitoids Nd model ages match the northwest Iberia Ediacaran to Ordovician sedimentary rocks with a range of ca. 2.3–1.0 Ga and a mode of ca. 1.5 Ga (e.g. Fernández-Suárez *et al*. 2014) and those of the Central Iberian Zone Cambro-Ordovician crust-derived Ollo de Sapo orthogneisses (Bea *et al*. 2010; Talavera *et al*. 2013).

Accordingly, a link between the Central Iberian Zone and the Sahara Metacraton was proposed (Bea *et al.* 2010; Talavera *et al.* 2013). The Central Iberian Zone Ediacaran to Ordovician sedimentary Nd model age main cluster is younger, ca. 1.5 Ga, than the same age OMZ sedimentary rocks Nd model age distribution, which is centred at ca. 1.7 Ga (Figure 8(e)). This leads us to exclude the Sahara Metacraton as the source of the OMZ sedimentary rocks (Figure 8(e,f)).

In comparison with the Anti-Atlas West African Craton Pan-African granitoids, the OMZ Ediacaran to Ordovician sedimentary rocks have generally older Nd model ages (Figure 8(e,f)). The former have more abundant relatively young ages centred at ca. 1.0 Ga (Figure 8(f)). Moreover, ca. 1.9–1.6 Ga Nd model ages, which are common in the OMZ, are scarce to non-existent in the Anti-Atlas West African Craton granitoids (Figure 8(f)). However, the OMZ Cambrian sedimentary rocks do include some younger Nd model ages of ca. 1.6–1.1 Ga (López-Guijarro *et al.* 2008). These younger ages may be explained by an input of juvenile material from: (i) mixed mantle-crustal source consistent with the OMZ Cambrian rifting or (ii) input of juvenile crust from Cadomian arc sediments. The Anti-Atlas West African Craton Pan-African granitoids are apparently not the main source of the OMZ Ediacaran and Ordovician sedimentary rocks old-crust signature. Nevertheless, we consider that an input from these granitoids may explain the juvenile-crust component character of the OMZ Cambrian sedimentary rocks.

The Pan-African granitoids from the Tuareg Shield show a wide range of Nd model ages, ca. 2.5–1.0 Ga, centred at ca. 1.7 Ga (Figure 8(f)). These granitoids match the main OMZ sedimentary rocks Nd model age distribution. In addition, like the OMZ sedimentary rocks, the Tuareg Shield Pan-African granitoids also have a population of younger Nd model ages. Considered together these characteristics make the Tuareg Shield granitoids the best fit as a source for the OMZ late Ediacaran to early Cambrian sedimentary rocks (Figure 8(e,f)).

This, thus, implies that the OMZ was geographically close to the Tuareg Shield but could also have received a sediment contribution from the Anti-Atlas West African Craton placing it to the north-northwest of the former and northeast of the latter (present-day coordinates).

5.2. Single and population zircon and whole-rock ages

To clarify further the OMZ position during the Cambro-Ordovician, we undertook a study of OMZ detrital zircons from sedimentary rocks and pre-magmatic and magmatic zircons from igneous rocks and compared them with data from the autochthonous Central Iberian Zone and Cantabrian Zone; the parautochthonous Galicia-Tras-os-Montes Zone and North Gondwana (supplementary material II).

5.2.1. Sedimentary rocks detrital zircons

5.2.1.1. The Ossa–Morena Zone. The U–Pb ion microprobe and LA-ICP-MS age determinations on detrital zircons from the OMZ Neoproterozoic to Ordovician sedimentary rocks reveal an abundance of Neoproterozoic ages (Figure 9(a)). Some 51% of the data are Cryogenian to Ediacaran, ca. 720–541 Ma. These data have a Pan-African peak, so typical of North Gondwana, at ca. 600 Ma (Figure 9(a)). Notably, older Neoproterozoic Tonian ages, ca. 1.0–0.72 Ga, represent 10% of data (Figure 9(a)). Palaeoproterozic, ca. 2.5–1.6 Ga, and Archaean, >2.5 Ga, ages are also abundant, 29% of the data, with main clusters at ca. 2.9, ca. 2.4, and ca. 1.6 Ga (Figure 9(a)). Remarkably, Mesoproterozoic, ca. 1.0–1.6 Ga ages are lacking in the OMZ sedimentary rocks (Figure 9(a)).

5.2.1.2. The other Iberian Massif zones. In the Central Iberian Zone Neoproterozoic to Ordovician sedimentary rocks, Neoproterozoic Cryogenian to Ediacaran ages also abundant, 60% of the data, with a peak at ca. 615 Ma (Figure 9(b)). Tonian, 16%, and Mesoproterozoic, 8%, ages are also relatively common yielding a peak at ca. 1.0 Ga (Figure 9(b)). Palaeoproterozoic and Archaean ages comprise 18% of data, but the main peaks are tighter than in the OMZ, ca. 2.9–2.5 Ga and ca. 2.3–1.9 Ga (Figure 9(b)).

In the Cantabrian Zone Neoproterozoic to Ordovician sedimentary rocks, Cryogenian to Ediacaran Neoprotezoic ages are again abundant, 58% of the data, with a peak at ca. 620 Ma (Figure 9(d)). As in the Central Iberian Zone, the Cantabrian Zone contains both Tonian and Mesoproterozic zircon populations, 29% of the data, with a well-defined peak at ca. 1.0 Ga (Figure 9(d)). Palaeoproterozoic and Archaean zircons comprise 23% of the data, with peaks at ca. 2.7–2.4 Ga and ca. 2.0 Ga (Figure 9(d)).

In the Galicia-Tras-os-Montes Zone, the ages of zircons from the parautochthonous units sedimentary rocks were considered. These rocks were deposited from the latest Neoproterozoic to the latest Cambrian (Díez-Fernández *et al.* 2010). Their zircon populations include Neoproterozoic Cryogenian to Ediacaran ages, 52% of the data, with a main cluster at ca. 650 Ma (Figure 9(c)). Tonian and Mesoproterozoic zircons are also present, 6% of the data, but are less abundant than in the Central Iberian Zone or the Cantabrian Zone (Figure 9(c)). In addition, Palaeoproterozoic zircons, 30% of the data, are numerous with main populations at ca. 2.3–2.1 Ga and ca. 1.9 Ga (Figure 9(c)). Archaean zircons comprise 9% of the data with a range of ca. 3.5–2.5 Ga and a peak at ca. 2.7 Ga (Figure 9(c)).

5.2.1.3. The North Gondwana, African, areas. The Sahara Metacraton, and the Arabian Nubian Shield, Cambrian to Ordovician sedimentary rocks have abundant Neoproterozoic zircons (Kolodner *et al*. 2006; Meinhold *et al*. 2011; Morton *et al*. 2013; compiled in Meinhold *et al*. 2013). Neoproterozoic Cryogenian and Ediacaran ages comprise 40% of the data, with a main peak at ca. 615 Ma (Figure 10(a)). The Sahara Metacraton sedimentary rocks also have a significant population of Mesoproterozoic and Tonian ages, 32% of the data, with a main peak at ca. 1.0 Ga (Figure 10(a)). They have abundant Palaeoproterozoic and Archaean ages, some 23% of the data, with small peaks at ca. 2.8–2.5 Ga and ca. 2.2–1.7 Ga (Figure 10(a)).

The sediment detrital zircons age pattern of the Neoproterozoic to Cambrian rocks from the Anti-Atlas West African Craton, reveals even more abundant Neoprotezoic Cryogenian and Ediacaran ages, 55% of the data, with the typical Pan-African North Gondwana peak at ca. 600 Ma (Figure 10(b)). Tonian, 5%, and Mesoproterozic, 2%, zircons, although uncommon are present (Figure 10(b)). Palaeoproterozic ages are abundant, 35% of the data, with a main peak centred at ca. 2.2 Ga and a minor peak ca. 2.5 Ga (Figure 10(b)) whereas the Archaean zircon population is scarce, only 2% (Figure 10(b)).

The Tuareg Shield Cambro-Ordovician sedimentary rocks present a bimodal detrital zircon age distribution. Neoproterozoic Cryogenian and Ediacaran ages comprise 66% of the data with a peak at ca. 615 Ma (Figure 10(c)). Tonian ages, 9% of the data, are present but these sedimentary rocks also have a Mesoproterozic gap: only 2% of data have an age of ca. 1.5 Ga, and there is no detectable ca. 1.0 Ga peak (Figure 10(c)). Palaeoproterozic and Archaean ages are abundant, 31% of the data, with a main peak at ca. 2.2–1.5 Ga (Figure 10(c)).

5.2.1.4. Comparison of the OMZ detrital zircon ages with other Iberian Massif zones and the North Gondwana, African, potential sources. Comparison of detrital zircon populations in Neoproterozoic to Ordovician sedimentary rocks from the Iberian Massif zones with the north African areas (Figures 9 and 10) revealed a northeast Gondwana (current coordinates) source for the zircon populations. The Central Iberian Zone, Cantabrian Zone and West-Asturian Leonese Zone have been linked to the Sahara Metacraton (Bea *et al*. 2010; Talavera *et al*. 2012; Henderson *et al*. 2016) and the Arabian Nubian Shield (Fernández-Suárez *et al*. 2014; Shaw *et al*. 2014).

Díez-Fernández *et al*. (2010), on the other hand, suggested an intermediate position between the Sahara Metacraton and Anti-Atlas West African Craton, for the Galicia-Tras-os-Montes Zone parautochthonous units. Their new position was based on the abundance of Cryogenian, Tonian and Mesoproterozoic ages, ca. 1.2–0.75 Ga, which are less abundant in the Anti-Atlas West African Craton, but well represented in the Sahara Metacraton (Figures 10(a,b)).

The main differences between the OMZ and the Central Iberian Zone, Cantabrian Zone, and Galicia-Tras-os-Montes Zone parautochthonous units are that the OMZ sedimentary rocks do not contain Mesoproterozoic zircons, ca. 1.0 Ga, or an abundance of Cryogenian and Tonian ages (Figure 9(a–d)). Thus the Sahara Metacraton is once more ruled out as a possible palaeogeographic source for the OMZ (Figure 10(a)).

The two other potential North Gondwana sources for the OMZ detrital zircons are the Anti-Atlas West African Craton and the Tuareg Shield (Figure 10(b,c)). The detrital zircon age distributions of the Anti-Atlas West African Craton and Tuareg Shield sedimentary rocks are both similar to those of the OMZ, for example the distribution of Neoproterozoic, Palaeoproterozoic, and Archaean ages (Figure 10). Also, Mesoproterozoic ages, ca. 1.0 Ga, are absent in the three areas (Figure 10(b,c)). However, there are some differences between the Anti-Atlas West African Craton and the OMZ to 'zircon age distributions': Cryogenian and Tonian ages are more abundant in the OMZ than the Anti-Atlas West African Craton, in particular for comparable Ediacaran age sedimentary rocks (Figure 10(b,c)). Also, the Palaeoproterozic zircon distribution in the Anti-Atlas West African Craton is centred at ca. 2.2 Ga with few representative ages in the range ca. 1.9–1.5 Ga which are common in the OMZ and also, notably, in the Tuareg Shield (Figure 10(b,c)). Significantly, Linnemann *et al*. (2011) noted that the ca. 1.9–1.5 Ga zircon population is absent in the Anti-Atlas West African Craton but is typical in rocks from the western Hoggar of the Tuareg Shield (cf. Drost *et al*. 2011) (Figure 10(b,c)).

To look in more detail at the similarities and differences in the distribution of the detrital zircon ages, we focussed on the zircon populations in just the Cambro-Ordovician sedimentary rocks from the OMZ, Tuareg Shield, and Anti-Atlas West African Craton (Figure 11). The zircon age distribution in the OMZ Cambro-Ordovician sedimentary rocks is quite similar to the whole detrital zircon population that also includes the Ediacaran sedimentary rocks (Figures 9(a) and 11(a)). This is not the case for the Anti-Atlas West African Craton Ediacaran and Cambro-Ordovician sedimentary rocks zircon populations (Figures 10(b) and 11(c)). The OMZ detrital zircon age distribution matches the Tuareg Shield ages better than those of the Anti-Atlas West African Craton (Figure 11(b)). The differences are particularly appreciable in the Anti-Atlas West African Craton Neoproterozoic, late Cryogenian to early Mesoproterozic ages, ca. 1.2–0.95 Ga, which are scarce to absent in both the OMZ and the Tuareg Shield (Figure 11(b,c)). The Palaeoproterozic zircons are more abundant in the OMZ and the Tuareg Shield, ca. 28% of the data, than in the Anti-Atlas West African Craton, 19% of the data. This difference in Palaeoproterozic variation is also marked by the aforementioned gap in the range ca. 1.9–1.5 Ga in the Anti-Atlas West African Craton compared with the other regions (e.g. Abati *et al*. 2010a; Linnemann *et al*. 2011; Avigad *et al*. 2012) (Figure 11(a–c)). Another important difference is in the Archaean zircons, which make up 6% of the OMZ data and are present, although scarce, in the Tuareg Shield, 2%, but virtually absent, ca. 1%, in the Anti-Atlas West African Craton Cambro-Ordovician sedimentary rocks (Figure 11(a–c)).

The differences between the zircon populations in the three zones are clear in cumulative fraction plots (Figure 11(d–f)). We used the two-sample Kolmogorov–Smirnov (K-S) test to make a quantitative comparison of the zircon populations (cf. Berry *et al*. 2001). This test compares the maximum difference of the cumulative fraction function for two different distributions, i.e. here, the detrital zircon age distributions. When the maximum difference between two populations (D) is significant, the null hypothesis is rejected, thus, the two distributions come from different populations. The D value of two distribution depends on the number of observations, $n > 20$, as well as on the similar abundance of observations in the compared distributions (Guynn and Gehrels 2010). Two populations can be considered to be from the same source if the probability value (p) corresponds to a confidence level of 95% ($p \geq 0.05$) (Guynn and Gehrels 2010). Shaw *et al*. (2014) performed a K-S test on the Ordovician Central Iberian Zone, Cantabrian Zone and West-Asturian Leonese Zone Amorican quartzite. The results showed that potential source areas are the Sahara Metacraton and Arabian Nubian Shield.

In the present work we carried out this test to assess the similarities and differences between the OMZ, Anti-Atlas West African Craton and Tuareg Shield. We took into account the analytical error of each measurement (cf. Guynn and Gehrels 2010). Another factor that had to be considered was the relative abundance of Cambrian zircons in the OMZ compared with similar age rocks from the Tuareg Shield and Anti-Atlas West African Craton. So the K-S test was only performed on Precambrian >541 Ma detrital zircons from the Cambro-Ordovician sedimentary rocks from the three regions (Figure 11(d–f)).

The K-S test revealed a great affinity between the Precambrian detrital zircon age distribution in the OMZ and Tuareg Shield Cambro-Ordovician sedimentary rocks (D 0.100, p 0.06), and so supports the connection between these two regions observed in the age histograms and density distributions (Figure 11(e)). The Anti-Atlas West African Craton Cambrian sedimentary rocks detrital zircons, by contrast, have a cumulative distribution different from the OMZ and Tuareg Shield and the K-S test rejects the possibility that this population had the same source as those regions (D 0.197, $p < 0.001$) (Figure 11(f)).

5.2.2. Igneous rocks pre-magmatic and magmatic zircons

To complement the Nd model age and detrital zircon data, we also consider the pre-magmatic and magmatic zircon age data from the Cambro-Ordovician extension-related igneous rocks from the OMZ, Central Iberian Zone and Galicia-Tras-os-Montes Zone parautochthonous units. The compiled geochronological data also includes Rb–Sr WR and U–Pb population zircon ages. These data are compared with similar data from the three north African areas (supplementary material II).

5.2.2.1. The Ossa–Morena Zone. Cadomian-arc and rift-related magmatic events were recorded by the OMZ zircons with peaks at ca. 550 Ma, ca. 520 Ma and ca. 480 Ma, 75% of the data (Figure 12(a)). The OMZ pre-magmatic ages include Neoproterozoic, Tonian, Cryogenian, and mostly Ediacaran centred at ca. 615 Ma 19% of the data, as well as less common Palaeoproterozoic ages, 3%, conspicuously once more Mesoproterozoic ages are absent (Figure 12(a)).

5.2.2.2. The other Iberian Massif zones. In the Galicia-Tras-os-Montes Zone parautochthonous units, Cambro-Ordovician magmatism comprises varied calc-alkaline, peraluminous, alkaline and peralkaline meta-granites and metavolcanic rocks (e.g. Arenas 1984; Montero *et al*. 1998). The magmatism occurred from ca. 500 to 470 Ma (Figure 12(b)). There was a change in the composition of the magmatism over time: (i) Calc-alkaline ca. 500–490 Ma (Rodríguez *et al*. 2007; Abati *et al*. 2010b) and (ii) Peraluminous felsic ca. 498–462 Ma

(Talavera *et al*. 2013) coeval with alkaline to peralkaline felsic ca. 475–470 Ma (Montero *et al*. 2009b; Díez-Fernandez *et al*. 2012). The latter event is comparable with the Central Iberian Zone that had a magmatic peak at ca. 480 Ma, 34% of the data (Figure 12(b)) (Montero *et al*. 2009b; Abati *et al*. 2010b; Díez-Fernandez *et al*. 2012; Talavera *et al*. 2013; Gutierrez-Alonso *et al*. 2016). The Galician peraluminous metagranites and metavolcanic rocks also have abundant pre-magmatic zircons (Figure 12(b)). These include: lower to upper Cambrian ages, 21% of the data (Figure 13(b)); Neoproterozoic Tonian, Cryogenian and Ediacaran ages, 23% of data centred at ca. 615 Ma; Mesoproterozoic, Stenian ages 5% of the data, in the range ca. 1.2–1.0 Ga (Figure 12 (b)); and Palaeoproterozoic ages of ca. 2.3–1.6 Ga, 9%, are also present (Figure 12(b)).

In the Central Iberian Zone, Cambro-Ordovician magmatism is only preserved in the peraluminous metavolcanic rocks and metagranites of the Ollo de Sapo Formation (Parga-Pondal *et al*. 1964; Montero *et al*. 2009a; Díez-Montes *et al*. 2010). The magmatism occurred from ca. 496 to 474 Ma with the peak of activity, marked by new zircon formation, at ca. 480 Ma, 18% of the data (Figure 12(c)) (Bea *et al*. 2007; Montero *et al*. 2007; Talavera *et al*. 2013). Pre-magmatic ages are extraordinarily abundant in these rocks. They include lower to upper Cambrian ages, 29% of the data (Figure 13(c)); Neoproterozoic ages, again Tonian, Cryogenian and Ediacaran centred at ca. 615 Ma that comprise 40% of the data (Figure 12(c)); Mesoproterozoic ages are also recognizable, 3% of the data, centred at ca. 1.1 Ga; as well as Palaeoproterozoic ages, 8% of the data (Figure 12(c)).

5.2.2.3. The North Gondwana, African, areas. The Sahara Metacraton granitoids do not record Cambro-Ordovician magmatism but have Neoproterozoic magmatic zircon ages, 78% of data centred at ca. 615 Ma (Figure 13(c)). Zircons with these ages are also found, pre-magmatic, in the OMZ, Central Iberian Zone and Galicia-Tras-os-Montes Zone parautochthonous units. Furthermore, the Sahara Metacraton granitoids also contain abundant Palaeoproterozoic and Archaean ages, 19% of data (Figure 13).

The Neoproterozoic to lower Cambrian magmatic events recorded in the Anti-Atlas West African Craton have an asymmetric distribution, with main Ediacaran peaks at ca. 600 Ma and ca. 560 Ma, 58% of the data (Figure 13(c)). A few middle Cambrian magmatic ages, 5% of data, are also detected (Figure 13(c)). This region also records abundant Palaeoproterozoic magmatism at ca. 2.0 Ga, 28% of the data, that is not present in the OMZ (cf. Barbey *et al*. 2004; Peucat *et al*. 2005) (Figure 13(c)).

Regarding the Tuareg Shield, the first thing to mention is that, unfortunately, there are few published single zircon ages. As mentioned above like in the other regions, the compiled geochronological data also includes Rb–Sr WR and U–Pb population zircon ages (Figure 13(b)).

The youngest igneous event in the Tuareg Shield was a major latest Neoproterozoic–Cambrian intrusion forming the alkaline–peralkaline Iforas province, ca. 556 Ma, the Tisselliline pluton, ca. 555 Ma, and the Taourirt magmatic province at ca. 525 Ma (Figure 13(b)). Azzouni-Sekkal *et al*. (2003) interpreted these as post-collisional alkaline magmatism related to the transtensional mega-shear zones that provoked asthenosphere upwelling and melting. Bertrand *et al*. (1986) dated Ediacaran, ca. 615 Ma, granitoids in the north of the region, contemporaneous magmatism was also dated in Central Hoggar (Liégeois *et al*. 2003; Abdallah *et al*. 2007) and in the Eastern Nigeria province at ca. 650–570 Ma (e.g. Ferré *et al*. 1996; Ekwueme and Kröner 1998). Other regional Neoproterozoic, mainly Cryogenian, magmatism includes granitoids in the Nigerian Air region, ca. 750–650 Ma (Liégeois *et al*. 1994) (Figure 13(b)). Somewhat older Neoproterozoic Tonian ca. 870–800 Ma, granitoids crop out in western-central Hoggar as do Palaeoproterozoic Eburnean, ca. 2.5 and ca. 2.0 Ga granitoids (Figure 12(b)) (e.g. Caby 2003). In addition, Palaeoproterozoic zircons, ca. 1.9–1.8 Ga, were described in the western Hoggar high-temperature metamorphic rocks (Peucat *et al*. 1996, 2003; Bruguier *et al*. 2008). Archaean zircons, ca. 3.5 Ga, are present as pre-magmatic ages in Pan-African granites from the Hoggar (Abdallah *et al*. 2007) and some West Nigeria outcrops (Kröner *et al*. 2001).

5.2.2.4. Comparison of the OMZ igneous pre-magmatic and magmatic zircon ages with other Iberian Massif zones and the North Gondwana, African, areas potential sources. The pattern of Central Iberian Zone and Galicia-Tras-os-Montes Zone parautochthonous units magmatic ages is quite different from the OMZ. The first two only show a magmatic peak at ca. 480 Ma whereas the OMZ, in addition to a ca. 480 Ma peak, also has older peaks at ca. 520 Ma and ca. 550 Ma (Figure 12). The difference in the age of the magmatic events led Montero *et al*. (2009a) to propose a different, albeit unspecified, palaeogeographic position for the OMZ relative to the Central Iberian Zone. This suggestion is also supported by the pre-magmatic age data, as stated previously the Central Iberian Zone and Galicia-Tras-os-Montes Zone both have

Mesoproterozoic ages that are very sparse in the OMZ (Figure 12).

Bea *et al*. (2010) compared pre-magmatic and magmatic zircon age data from the Central Iberian Zone orthogneisses and north African granitoids in their study of the early Palaeozoic palaeographic position of the Central Iberian Zone. Based on this and the aforementioned Nd model ages they concluded that, at that time, the Central Iberian Zone was located further east than had previously been thought, next to the Sahara Metacraton. A similar scenario was proposed for the Galicia-Tras-os-Montes Zone parautochthonous units based on the compositional characteristics of the peraluminous magmatism (Talavera *et al*. 2013).

Considering other regions, the Anti-Atlas West African Craton magmatic events are not a good fit with the timing of the OMZ events (Figure 13). Cambrian events preserved in the former are older than in the OMZ or any of the other Iberian Massif zones (Figure 13). Furthermore, the Ordovician ca. 480 Ma magmatic event identified in the OMZ, Central Iberian Zone, and Galicia-Tras-os-Montes Zone parautochthonous units is hardly represented in the Anti-Atlas West African Craton magmatic rocks (Figure 13 (c)). In fact, the Anti-Atlas West African Craton and OMZ rift-related magmatism was, seemingly, diachronous. It began in the latest Ediacaran in the former (Thomas *et al*. 2002, 2004; Walsh *et al*. 2012) but later, in the Cambrian 'early and main rift-related events', in the latter (Sánchez-García *et al*. 2008, 2010).

Revealingly perhaps, the youngest Tuareg Shield Taourit magmatic province alkaline event (Azzouni-Sekkal *et al*. 2003) was contemporaneous with the OMZ 'early rift-related event' at ca. 540–520 Ma (Sánchez-García *et al*. 2008, 2010). Consistent with this, the OMZ magmatism has an alkaline composition with a clear crustal contribution that, we have suggested, resulted in a mixed Nd model age range of ca. 1.7–1.0 Ga (Figure 8(a)). This is in agreement with the mixed origin suggested by Azzouni-Sekkal *et al*. (2003) for the Tuareg Shield alkaline magmatism. The recognition of a comparable coeval tectonomagmatic event in both the OMZ and the Tuareg Shield provides further support for the suggestion that the regions were proximal during the Cambro-Ordovician.

6. The palaeogeographic position of the Ossa–Morena Zone during the Cambro-Ordovician

Stampfli *et al*. (2013) and Torsvik and Cocks (2013) considered an early Palaeozoic rifting event in the northern part of Gondwana in the transition between the Cadomian and Variscan orogenies. Associated magmatism has been suggested to be a consequence of west to east (present-day coordinates) Rheic ocean opening and associated separation of a ribbon continent from the northern margin of Gondwana (Figure 14 (a)) (Murphy *et al*. 2006; Nance *et al*. 2010, 2012).

The question is then, whether the new palaeogeographic positions recently suggested for the Iberian Massif zones, as detailed above, and the interpretation of their Cambro-Ordovician crustal evolution, in particular magmatism, are consistent with this scenario. We consider this and also the Cambro-Ordovician stratigraphic similarities and differences between the OMZ and the other Iberian Massif zones and the north African areas. Geochronological data combined with magmatic and stratigraphic interpretations permits us to propose a tectonomagmatic scenario for the OMZ during the Cambro-Ordovician.

6.1. The previously accepted palaeogeographic position of the Ossa–Morena Zone

The OMZ has been linked with the western region of the Anti-Atlas West African Craton during the Cambro-Ordovician (Murphy and Nance 1989; Nance and Murphy 1996). This scenario was proposed based on palaeontological data that indicated the OMZ and other regions of Cadomia had fauna typical of the periphery of Gondwana (Robardet *et al*. 1994; Robardet and Gutiérrez-Marco 2004). Palaeomagnetic data, albeit inconclusive, was also interpreted to suggest a north Africa position for the OMZ and other Cadomian regions at that time (e.g. Torsvik *et al*. 1992; Stampfli *et al*. 2002). Other data that led to the suggestion that the OMZ was positioned close to the Anti-Atlas West African Craton, was the age distribution of the detrital zircon population in the sedimentary rocks (Nance and Murphy 1994; Fernández-Suárez *et al*. 2002; Linnemann *et al*. 2008; Pereira *et al*. 2011, 2012c). As noted above, this connection was based on the abundance of OMZ Neoproterozoic, Palaeoproterozoic, and Archaean ages and the lack of Mesoproterozoic ages, ca. 1.0 Ga in the two regions (Figure (10)).

6.2. Towards a new palaeogeographic position for the Ossa–Morena Zone

Álvaro *et al*. (2014) reviewed the relationship between the western sector of the Anti-Atlas West African Craton and the OMZ. They noted that there was a difference in the age and timing of several important lithologies. First, in the western Anti-Atlas West African Craton Cryogenian and Ediacaran sedimentary sequences are well developed, whereas in the OMZ only Ediacaran,

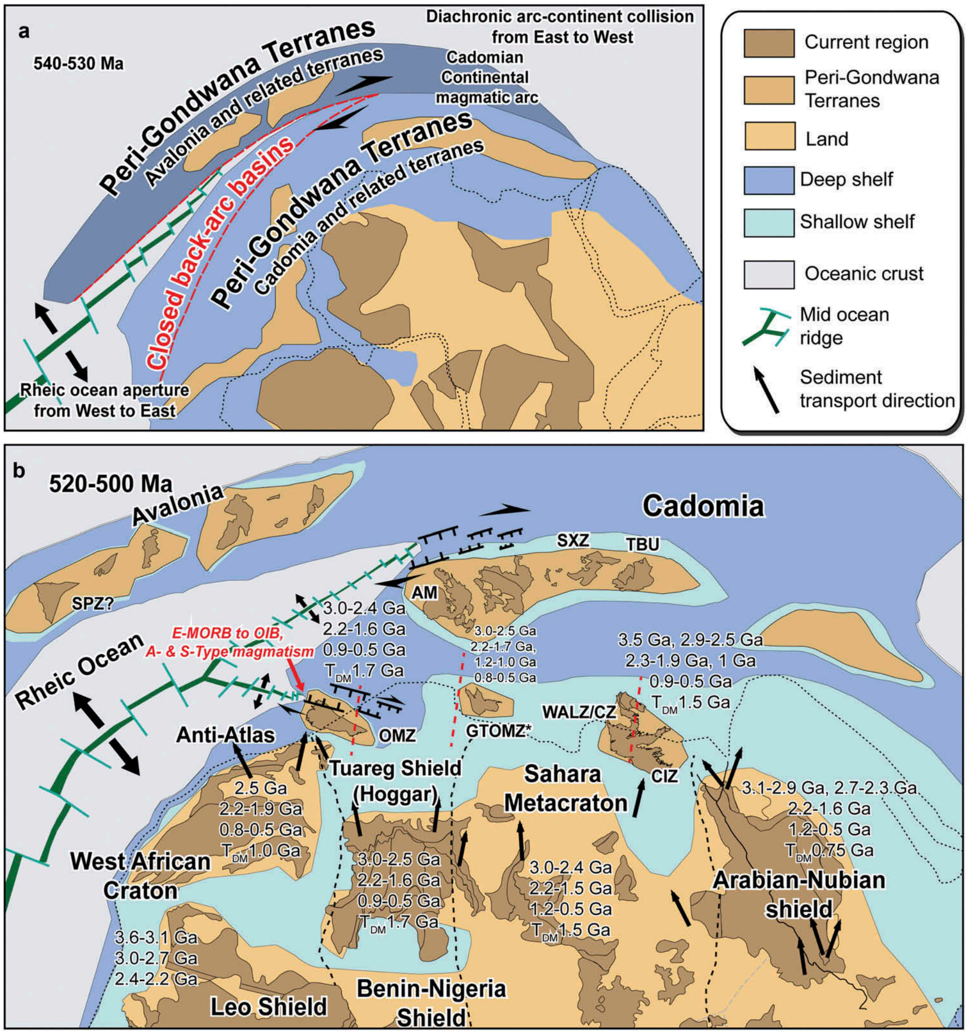

Figure 14. Palaeogeographic reconstructions of northern Gondwana: (a) Transition from continental arc to west-to-east Rheic Ocean opening that resulted in ribbon continent separation from northern Gondwana during the Precambrian–Cambrian transition (based on Linnemann *et al*. 2008; Torsvik and Cocks 2013). (b) Middle to late Cambrian rift progression provoked the separation of peri-Gondwana terranes Avalonia, South Portuguese Zone-like (SPZ?) then Cadomia (AM: Armonican Massif; SXZ: Saxo-Thuringian Zone; TBU: Tepla-Barrandian Unit) (after Linnemann *et al*. 2008). This ca. 520–500 Ma period, contemporaneous with the 'main rift-related event' in the Ossa–Morena Zone (OMZ), is associated with initiation of a 'Gondwana-ward' branch of the Rheic Ocean to the west (palaeogeographic position) of the OMZ close to Tuareg Shield. Our new more easterly palaeoposition for the OMZ is adjacent to that recently defined for the Galicia-Tras-os-Montes Zone parautochthonous units (GTOMZ*), West-Asturian Leonese Zone (WALZ), Cantabrian Zone (CZ) and Central Iberian Zone (CIZ) close to the Sahara Metacraton. (c) Ordovician, ca. 490–470 Ma, rifting evolution with passive margin formation that resulted in generation of oceanic crust in the west in the OMZ, whereas in the Galicia-Tras-os-Montes Zone and Central Iberian Zone magmatism was incipient rift propagation related. For clarity, we include the main detrital zircon ages of the OMZ, Central Iberian Zone and the three north African areas as well as a summary of magmatism during Rheic Ocean branch opening from west to east. In addition, shallow and deep ocean and land morphology are shown, from Torsvik and Cocks (2013) as is the bathimetric gradient established for the OMZ and Central Iberian Zone after Robardet and Gutiérrez-Marco (2004). Thick dashed lines in (b) and (c) correspond to Figure 15 cross-sections. Cratonic regions age ranges are taken from the compilation in the present work and from compilations by Pereira *et al*. (2008) and Drost *et al*. (2011). Sediment transport directions are from Avigad *et al*. (2003) and Shaw *et al*. (2014).

Serie Negra, sedimentary rocks crop out. Second, the two regions have different zircon ages in arc-related syn-orogenic basins, specifically the Ediacaran sedimentary rocks in both areas have different maximum sedimentation ages: ca. 630–610 Ma in the Anti-Atlas West African Craton and ca. 590–545 Ma in the OMZ. Third, they noted a variation in the timing of key tectonomagmatic events, such as calc-alkaline arc-related magmatism, in the Anti-Atlas West African Craton, ca. 615–579 Ma, and the OMZ, ca. 582–535 Ma (Figure 13 (a,c)). Also, as noted above, the post-collisional magmatism in the Anti-Atlas West African Craton occurred at

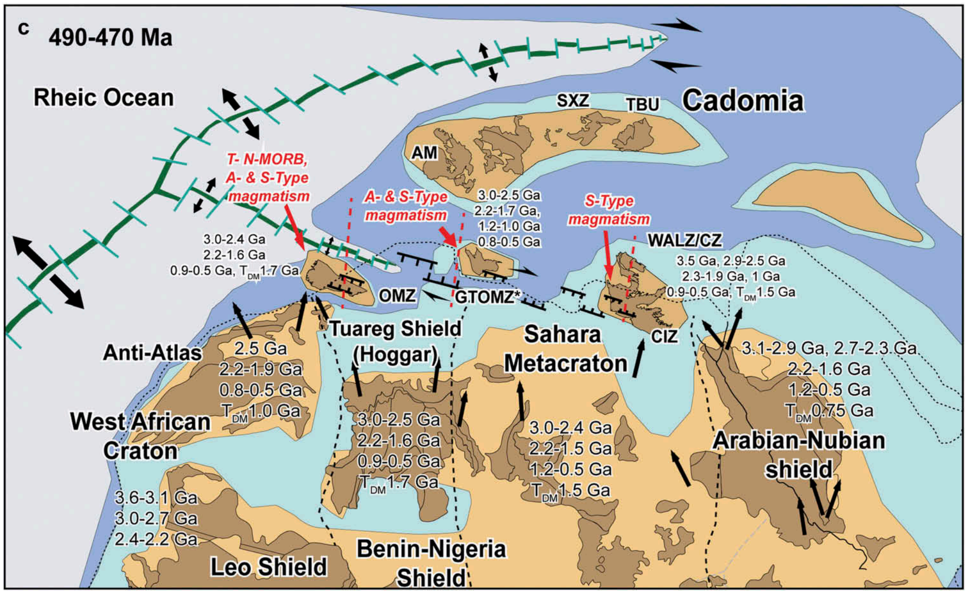

Figure 14. (Continuted).

ca. 577–552 Ma, whereas in the OMZ comparable magmatism, i.e. the ‘early rift-related event’, happened later at ca. 530–527 Ma (Figure 13(a,c)). Notably, Bea *et al.* (2015) dated the first evidence of peralkaline magmatism generated in an intraplate rifting environment in the West African Craton at ca. 527–517 Ma. This magmatism is compositionally comparable to the peralkaline magmatism associated with rifting progression in the OMZ at ca. 490–470 Ma. Rift propagation in both areas marks a southwest to northeast trend, present-day coordinates.

So, the Cambro-Ordovician progression of events in the Anti-Atlas West African Craton and the OMZ was apparently diachronous from the southwest to the northeast. This leads us to propose that if the OMZ was close to the Anti-Atlas West African Craton then it would have been in a more northeasterly rather than westerly position. In agreement with this, the data presented here link the OMZ and the Tuareg Shield. All available geochronological data, be it Nd model ages, dates of detrital zircons in sedimentary rocks, or pre-magmatic and magmatic ages, is consistent with the OMZ having a clear input from, and being adjacent to, the Tuareg Shield during the early Palaeozoic (Figure 14(b)).

Furthermore, similarities can be established between the sedimentary sequences in the OMZ and Tuareg Shield Tassili Ouan Ahaggar basin. In both cases the sedimentation record reflects an important change from continental deposits to marine deposition with a clear input of Pan-African rocks (Beuf *et al.* 1971; Liñán and Quesada 1990; Ghienne *et al.* 2007a, 2007b). Moreover, Cambrian deposits unconformably overlie Precambrian basement. The early Cambrian series in the OMZ and the Tuareg Shield are fluvial to shallow marine deposits that have been related to a Rheic Ocean rifting context (cf. Sánchez-García *et al.* 2010; Linnemann *et al.* 2011). Also, early Ordovician deposits, in both the OMZ and Tuareg Shield Tassili Ouan Ahaggar basin, are detrital to open marine sediments although these are somewhat deeper in the former (Beuf *et al.* 1971; Giese *et al.* 1994a). Similarly, late Ordovician to Silurian deposits in both the Tuareg Shield Tassili Ouan Ahaggar basin and the OMZ are characterized by monotonous black marine shelf shales (Beuf *et al.* 1971; Gutiérrez-Marco *et al.* 1998), which subsequently progressed to more shallow-marine Early Devonian terrigenous series (Beuf *et al.* 1971; Oliveira *et al.* 1991).

7. Cambro-Ordovician rifting of the North Gondwana margin related to west to east opening of the Rheic Ocean

7.1. Stratigraphic and palaeontological data

The indication of a somewhat deeper marine sedimentation in the OMZ than in the Tuareg Shield is consistent with a more northerly marginal palaeoposition relative to the northern margin of Gondwana. It also fits with a Cambro-Ordovician northeasterly sediment transport direction (Beuf *et al.* 1971). So, the overall correlation between the two regions outlined above and established here link them to Cambro-Ordovician rifting and Rheic Ocean opening off North Gondwana (Murphy *et al.* 2006; Nance *et al.* 2010).

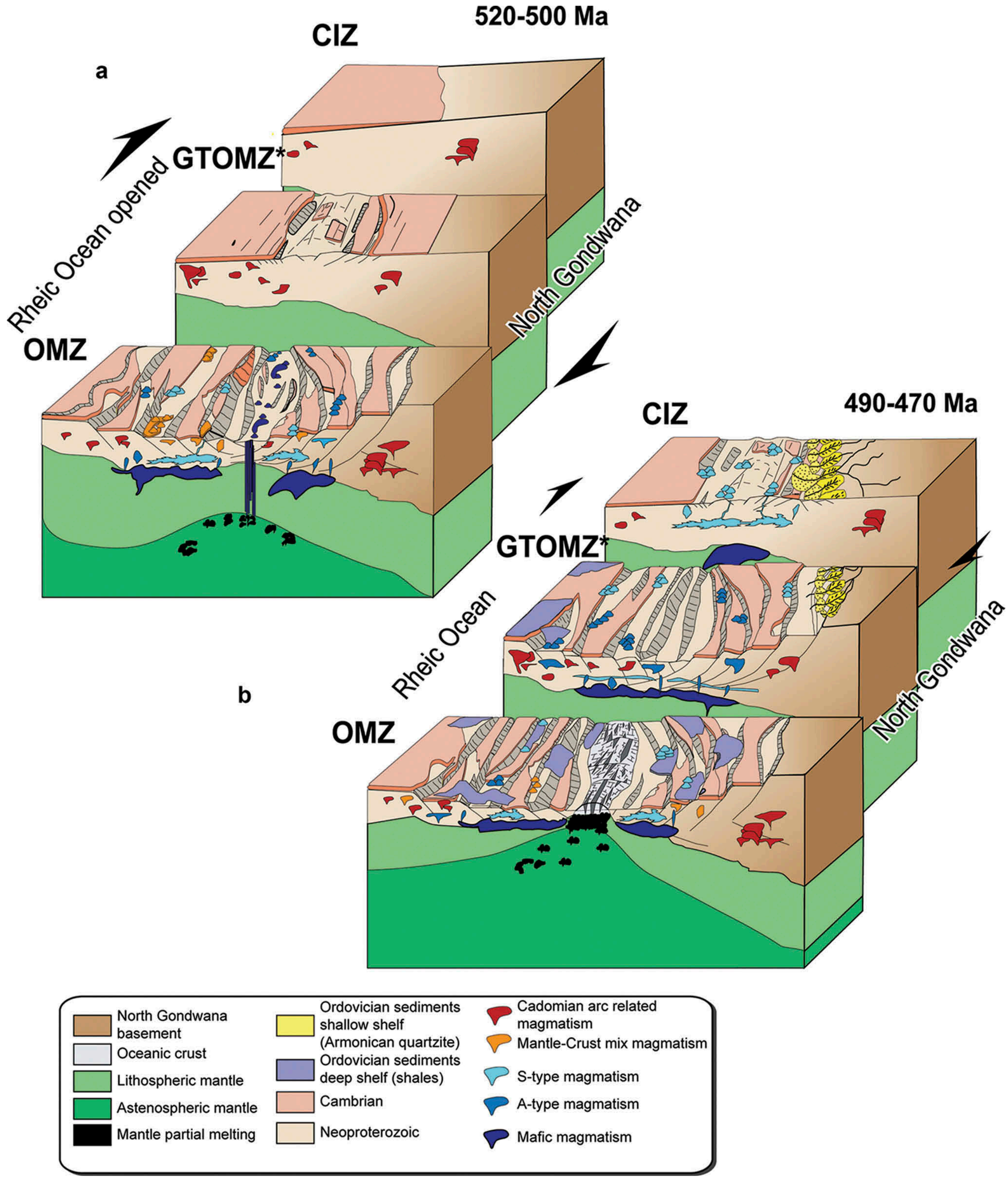

Figure 15. Model for the formation of the 'Gondwana-ward' Rheic Ocean branch that propagated from west to east during: (a) Cambrian ca. 520–500 Ma, 'main rift-related event' period during which there was: generation of A-type, S-type, mantle-crust mixed and E-MORB to OIB mafic magmatism in the Ossa–Morena Zone (OMZ); and, collapse-related extension that affected Cambrian sedimentary successions. Notably this magmatic event is absent in the Galicia-Tras-os-Montes Zone parautochthonous units (GTOMZ*) and the Central Iberian Zone (CIZ) that were further east close to the Sahara Metacraton. (b) Cambro-Ordovician, ca. 490–470 Ma, oceanic crust generation in the OMZ resulted from rift progression, with generation of A- and S-type magmatism further east in the GTOMZ*, whereas in the CIZ more limited rifting generated S-type magmatism during the same period. Note that during the Ordovician a deeper marine context is suggested for the OMZ and shallow deposition conditions were defined for the CIZ from Armorican quartzite sedimentation (Robardet and Gutiérrez-Marco 2004.

Analogous, albeit diachronous, rifting is also recorded elsewhere along the North Gondwana margin. The stratigraphic sequences of the Anti-Atlas West African Craton and part of the European Variscan Massif preserve evidence of an early Cambrian rifting stage that evolved to a passive margin context during the Ordovician (Drost *et al*. 2011; Álvaro *et al*. 2014). In addition, Cambro-Ordovician rifting, admittedly not as extensively developed as in the OMZ, has also been described in other Iberian Massif zones, for example in the Central Iberian Zone and Cantabrian Zone (cf. Robardet and Gutiérrez-Marco 2004; Pastor-Galán *et al*. 2013a). Early Cambrian Central Iberian Zone shallow marine deposits (Liñán *et al*. 2002) are coeval with the OMZ Cambrian marine carbonate unit and upper detrital formation (Liñán and Quesada 1990). However, these rocks have been interpreted differently in the two regions. In the Central Iberian Zone calm shallow-marine deposition is suggested to have occurred prior to Cambro-Ordovician rifting (Simancas *et al*. 2004) whereas in the OMZ coeval sedimentation apparently resulted from turbiditic and shelf siliciclastic deposition during a collapse-related extensional process (Figure 14 (b,a)) (cf. Sánchez-García *et al*. 2003).

Drawing on palaeontological data, Robardet and Gutiérrez-Marco (1990a, 1990b, 2004) concluded there was no doubt that both the OMZ and Central Iberian Zone have North Gondwana type sediments and faunas. However, whereas the Cambrian sedimentation was similar in the two regions, the lower Ordovician–Lower Devonian successions differ appreciably. The stratigraphic and palaeontological indications suggest that the OMZ was situated in a deep ocean that shallowed eastward to the Galicia-Tras-os-Montes Zone and became shallower still further east in the West-Asturian Leonese Zone, Cantabrian Zone, and Central Iberian Zone (Figures 14(c) and 15(b)).

Arguments for the eastward shallowing also include the lower Ordovician, ca. 477–465 Ma, Armorican Quartzite (e.g. Gutiérrez-Marco *et al*. 2002). This facies is interpreted as an off-shore Gondwana passive margin unit deposited on a wide stable shallow marine platform. The quartzite is well-exposed in the Central Iberian Zone and Cantabrian Zone. It is not present, however, in the OMZ (Figures 14(c) and 15(b)) (Robardet and Gutiérrez-Marco 1990a, 1990b, 2004) because there the succession comprises deeper water shaley and silty deposits characteristic of a more distal deeper marine environment (Figures 14(c) and 15(b)). Accordingly, a bathimetric gradient has been established from deeper, OMZ, to shallow, Central Iberian Zone (Figures 14(b–c) and 15(a,b)) (Hammann and Henry 1978; Gutiérrez-Marco *et al*. 1998). This earlier initiation and more extensive progression of rifting is, once more, consistent with the west-to-east opening of the Rheic Ocean (cf. Nance *et al*. 2012).

Linking back to geochronological constraints, Shaw *et al*. (2014) studied U–Pb detrital zircon age data from the Armorican Quartzite to determine the provenance variability of the Cantabrian Zone, West-Asturian Leonese Zone, and Central Iberian Zone along the early Ordovician North Gondwana margin. They concluded that their data were consistent with the proposed easterly off-shore Sahara Metacraton and Arabian Nubian Shield location of the zones proposed by Bea *et al*. (2010), Pastor-Galán *et al*. (2013a), Talavera *et al*. (2013), and Fernández-Suárez *et al*. (2014).

7.2. *Magmatic data*

It is worth underlining that the timing and composition of early Palaeozoic magmatism in the Iberian Massif zones detailed above are equally consistent with the west-to-east opening of the Rheic Ocean and also with the new palaeogeographic positions. As can be deduced from the range of magmatic ages presented, Cambro-Ordovician magmatism began earlier and was more protracted in the more westerly OMZ than the more easterly, Cambro-Ordovician coordinates, Central Iberian Zone (Figure 15(a)). Magmatism was also more prolonged in the OMZ than in the Galicia-Tras-os-Montes Zone parautochthonous units, which have also recently been situated in an eastern position off the Sahara Metacraton (Figures 14(b) and 15(a)).

The OMZ Cambro-Ordovician plutonic rocks, ca. 520–470 Ma, which include peralkaline anorogenic magmatism, are related to progression of extension that started in the early Cambrian (Figure 15). This extensional event is characterized by bimodal magmatism comprising mafic and felsic rocks best exposed in the west of the OMZ, and by the generation of alkaline magmas, (Figure 15(a,b)).

The Central Iberian Zone magmatism during the same period is restricted to ca. 480 Ma calc-alkaline to peraluminous S-type rocks preserved as metagranite and metavolcanic orthogneisses that contain abundant pre-magmatic zircons. Bea *et al*. (2007) interpreted the numerous inherited zircons to reflect fast crustal melt magma generation and emplacement related to intrusion of mantle-derived mafic magmas at the base of the crust in an extensional environment (Figure 15(b)) (cf. Díez-Montes *et al*. 2010; Talavera *et al*. 2013). This led Bea *et al*. (2007) to conclude that their model was consistent with the hypothesis of separation and dispersal of ribbon-continent fragments from the northern margin of Gondwana. Accordingly, Rubio-Ordóñez *et al*.

(2012) recently identified an early Ordovician tonalitic-granodioritic belt in the Central Iberian Zone.

Similar to the Central Iberian Zone, the Galicia-Tras-os-Montes Zone parautochthonous units contain ca. 480 Ma S-type and A_1-type granites, no coeval mafic rocks crop out indicating that rifting did not progress as much as in the OMZ (Figure 15(b)). The weaker Ordovician magmatic expression in the Galicia-Tras-os-Montes Zone and Central Iberian Zone father east, Cambro-Ordovician coordinates, than the OMZ, is consistent with production related to an eastward propagating rift.

The pattern of the OMZ Cambro-Ordovician magmatism is consistent, temporally and compositionally, with the diachronous development of a narrow ocean basin related to a rifted volcanic margin, comparable say to the Ethiopian rift Red Sea type model (Pearce 2008). There, the progression was from E-MORB tholeiitic to OIB-like alkaline and finally T- and N-MORB with a declining crustal input over a period of some 30 million years. Initial felsic magmatism, comparable to the OMZ 'early rift-related event' at ca. 540–520 Ma, changed to predominantly intermediate and then bimodal basic-felsic as rifting proceeded, equivalent to the OMZ 'main rift-related event' at ca. 520–500 Ma. Finally, the Red Sea context evolved from continental to oceanic crust generation (Wolfenden *et al*. 2005), as preserved in the OMZ in the ca. 490–470 Ma T- and N-MORB mafic and alkaline and peralkaline felsic magmatism. We suggest that the OMZ preserves a Gondwana-ward southerly branch of the Rheic Ocean that opened diachronously from west to east rather than the main ocean (cf. Linnemann *et al*. 2008) (Figures 14 and 15). Further east, in the Galicia-Tras-os-Montes Zone, coeval extension is reflected by Ordovician mantle-derived A_1-type granites and subsequent crustal melt S-type granitoids (Figures 14 and 15). Further eastward still in the Central Iberian Zone, where rifting did not progress so extensively, no mantle-derived magmatic expression is described other than provocation of a crustal partial melting event that produced the S-type Ollo de Sapo orthogneisses with abundant pre-magmatic zircons (Figures 14 and 15).

The sedimentological and magmatic data lead us to suggest that the OMZ was located to the west of the other Iberian Massif zones during the Cambro-Ordovician. Nevertheless, stratigraphic data, as detailed above, and the apparent sedimentary contributions from both the Tuareg Shield and the Anti-Atlas West African Craton, place it further east than its previously accepted palaeogeographic position off the western Anti-Atlas West African Craton (Murphy and Nance 1991; Nance and Murphy 1994; Sánchez-García *et al*. 2008; Pereira *et al*. 2012c) (Figure 14(b)). Recent Cambro-Ordovician palaeogeographic reconstructions of North Gondwana establish a more easterly palaeogeographic position for Cadomia (Stampfli *et al*. 2002; Simancas *et al*. 2009; Torsvik and Cocks 2013). This is consistent with our new scenario that suggests a more easterly northeast Anti-Atlas West African Craton to northwest Tuareg Shield, Cambro-Ordovician palaeogeographic position of the OMZ.

8. Conclusions

(i) Isotopic data, including Nd model ages, dates of detrital zircons in Ediacaran to Ordovician sedimentary rocks and pre-magmatic and magmatic zircons in Cambro-Ordovician igneous rocks permit the reinterpretation of the palaeogeographic position of the OMZ during the Cambro-Ordovician. At that time the OMZ was situated adjacent to the Tuareg Shield and to the northeast, present-day coordinates, of the Anti-Atlas West African Craton. This new position is consistent with recent Cambro-Ordovician palaeogeographic reconstructions of North Gondwana that also place other Iberian zones further east, present-day coordinates, than previously thought.

(ii) Along the North Gondwana margin, a rifting context was recorded during the Cambro-Ordovician associated with Rheic Ocean opening diachronously from west to east. The magmatic expression of this extension is weaker in the more easterly Iberian Massif zones than the westerly location of the OMZ proposed in the present work. These differences are reflected in: the more prolonged mafic and felsic magmatism in the OMZ (ca. 540–470 Ma); the temporally more restricted peralkaline-alkaline and peraluminous magmatism in the parautochthonous units from the Galicia-Tras-os-Montes Zone; and, calc-alkaline to peraluminous magmatism in the Central Iberian Zone (ca. 480 Ma).

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