The amphibolites from the Ossa–Morena / Central Iberian Variscan suture (Southwestern Iberian Massif): geochemistry and tectonic interpretation

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Abstract

Basic rocks included in the Ossa–Morena/Central Iberian Variscan suture of SW Iberia have been studied in order to decipher the pre-orogenic evolution along this major boundary. These rocks appear as garnet-free amphibolites and garnet-bearing amphibolites. Previous geochronological data on these basic rocks suggest the existence of two different protolith ages: Upper Precambrian and Lower Palaeozoic (Ordovician). Upper Precambrian amphibolites form small-size (up to several metres thick), lense-, dike- or elongated-shaped bodies, while Lower Palaeozoic ones form elongated bodies with thicknesses up to several hectometres.

The chemical compositions of the amphibolites enable us to separate cumulate rocks from those representing more basaltic magmas. As for the originally basaltic rocks, their geochemical characteristics indicate a shallow melting with different proportions of at least three end-members, namely (i) a MORB-like asthenospheric source, (ii) a more enriched (plume-derived or subcontinental lithosphere) mantle source, and (iii) a continental crust component. Moreover, Upper Precambrian and Lower Palaeozoic amphibolites have different geochemical affinities. In the former group, we envisage an advanced intra-continental rifting situation or an intra-continental back-arc setting related to the Cadomian orogeny, while for the latter, we propose an environment of oceanic crust formation predating Variscan subduction and collision. The recognition of these remains of oceanic-affinity basic rocks along the OMZ/CIZ boundary allows us to propose that some sort of ocean may well have existed between the two zones during Early Palaeozoic times, and that this ocean was presumably closed at a later stage, probably during the Devonian, by oceanic subduction under the CIZ crust.

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1. Introduction

Metamorphosed basic and ultrabasic rocks are of substantial importance in establishing the tectonic
evolution of mountain belts. These rocks usually mark those places where subduction of oceanic lithosphere took place before the collision of two continents. Four different aspects need to be addressed in order to understand the significance of basic rocks in the orogens: (i) ages of protolith and metamorphism, (ii) geochemistry, (iii) metamorphic evolution, and (iv) structural evolution. In particular, the geochemical features of these rocks, together with the protolith age, can be used to characterize the pre-collisional history of the orogens in terms of the tectonic scenario where the basic rocks were formed, e.g. large or narrow ocean, back-arc basin, continental rift, subduction zone, etc. In this respect, the recognition of remains of ophiolitic sequences is of particular interest. The study of the metamorphic and structural evolution of basic rocks provides information on possible subduction events responsible for the closure of an intervening ocean, as well as on the subsequent exhumation of these rocks after the collision.

In the case of Palaeozoic and older orogens, the use of basic rocks in the reconstruction of pre-collisional tectonic settings is much more difficult than in the case of more recent ones. The difficulties in question arise from a number of uncertainties concerning the age of the protoliths along with complicated metamorphic and structural evolutions —sometimes polyorogenic—that may obliterate the primary geochemical signatures of the rocks. Despite these limitations, careful and systematic geochemical studies on basic rocks, together with other geological and geophysical data, can be regarded as the basic tools to produce well-supported and coherent plate tectonic reconstructions.

Basic rocks in the Central Unit, one of the sutures of the SW Iberian Massif, are relatively well known, especially from the standpoint of their metamorphic and structural evolution. However, their geochemical features have not received due attention in the literature so far. This paper aims to study the magmatic significance of these basic rocks, which were strongly deformed and metamorphosed under high-pressure conditions (Abalos et al., 1991; López Sánchez-Vizcaíno et al., 2003) during the Variscan orogeny (Azor et al., 1993, 1995; Ordóñez Casado, 1998). First, we will outline the geochemical features of these rocks, and then will go on to address their likely significance and tectonic implications in the context of their location along one of the main boundaries of the Variscan belt in the Iberian Peninsula.

2. Geological setting

The southwestern Iberian Massif comprises three zones separated by major tectonic boundaries (Fig. 1a): the South Portuguese Zone (SPZ), the Ossa–Morena Zone (OMZ) and the Central Iberian Zone (CIZ). The SPZ/OMZ contact has been considered to be a Variscan suture according to the existence of basic igneous rocks (amphibolites and basalts) with oceanic affinity (Bard, 1977; Dupuy et al., 1979; Munhá, 1983; Munhá et al., 1986). However, the lack of high-pressure metamorphic assemblages within the basic rocks (Bard, 1977) may indicate a very limited subduction along this contact.

In terms of its metamorphic and tectonic evolution, this contact has been considered to be one of the sutures of the Variscan orogen in the Iberian Peninsula (Matte, 1986; Azor et al., 1994; Simancas et al., 2001). The boundary itself is marked by a complex unit, namely, the Central Unit (Azor et al., 1994), made up of orthogneisses, amphibolites and metasediments. Moreover, Ordovician–Devonian stratigraphic and palaeontological differences at both sides of this boundary (Robardet, 1976; Robardet and Gutiérrez Marco, 1990) suggest the possible existence of an ocean between the OMZ and CIZ in Lower Palaeozoic times. Nevertheless, oceanic subduction prior to collision along this suture has not yet been proved. In this regard, the existence of sheet-shaped strongly deformed bodies of basic rocks included in this contact may well prove useful to confirm or discard the hypothesis of an oceanic subduction along this suture. Thus, the geochemical characterisation of these basic rocks, which is the target of our study, can contribute to the deciphering of the pre-collisional Early Palaeozoic evolution of the OMZ/CIZ suture.

The amphibolites crop out as metre- to hectometre-thick bodies parallel to the foliation and intercalated within orthogneisses. The rocks are garnet-free amphibolites and garnet-bearing amphibolites (see López Sánchez-Vizcaíno et al., 2003, for a detailed petrographic description). Both types of rocks under-
Fig. 1. (a) Schematic geological map of the Iberian Massif with the main zones and tectonic boundaries of the Variscan belt. (b) Very simplified geological map of the boundary between the Ossa–Morena and the Central Iberian Zones; asterisks depict the localities where the amphibolites were sampled. (c) Geological cross-section across the Ossa–Morena/ Central Iberian contact, showing the lithological succession and the structural position of the Central Unit underneath the Central Iberian Zone; location in (b).
Table 1
Major (wt.%) and trace element (ppm) contents of Upper Precambrian amphibolites (Group 1)

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$X_{Si}$ 0.40 0.43 0.44 0.49 0.51 0.45 0.51 0.46 0.46 0.50 0.51 0.55 0.54 0.57 0.29 0.49 0.44

Li 3.5 4.3 16.6 22.9 46.8 14.4 17.8 7.6 18.2 12.3 3.1 9.9 7.6 5.4 10.3 27.8 11.3
Rb 10.6 18.5 11.2 20.4 41.1 15.3 17.7 8.6 16.6 57.2 5.9 20.0 10.8 6.4 11.7 41.6 13.6
Cs 0.2 0.2 0.6 1.8 3.6 0.6 2.0 0.3 0.8 1.4 0.4 0.8 1.0 1.4 0.2 0.5 0.4
Be 1.7 0.4 0.5 1.1 1.5 1.1 1.5 0.7 1.2 1.3 1.2 0.9 0.3 3.1 1.4 1.8
Sr 232.8 175.6 564.1 256.5 274.0 255.9 432.3 207.5 487.2 321.1 371.9 317.5 328.6 332.1 183.5 230.1 360.9 212.8
Ba 151.1 159.3 124.5 201.1 289.1 146.5 160.5 117.5 110.6 332.6 106.5 112.6 112.2 41.9 172.1 538.7 1777.6
Sc 45.8 32.1 34.9 39.2 40.7 39.7 40.6 47.2 36.1 31.3 43.7 40.0 34.6 45.0 35.5 42.5 39.9
V 385.8 437.3 401.7 281.8 277.1 324.7 330.2 346.5 384.0 235.2 325.2 254.8 246.6 461.1 234.4 333.0 307.5
Cr 52.2 19.8 3.7 65.0 65.4 53.5 69.9 104.8 8.1 68.2 135.7 103.3 126.9 256.4 0.0 104.9 50.3
Co 48.2 72.1 60.4 40.6 39.2 39.2 32.1 49.8 47.9 31.0 42.2 46.3 38.9 52.6 31.3 25.5 39.0
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Cu 47.2 84.6 58.0 20.1 20.4 50.1 13.5 48.1 59.2 83.4 31.2 30.9 21.8 41.4 31.8 14.2 40.5
Zn 124.5 90.7 107.3 109.6 105.3 110.6 83.8 111.3 93.7 74.2 95.9 94.1 95.3 61.4 163.9 258.9 143.2
Ga 22.6 16.6 23.6 21.7 21.7 19.7 21.5 21.8 22.4 18.3 21.6 20.4 18.9 16.5 27.2 20.6 22.7
Y 60.7 34.1 24.9 38.6 39.5 51.2 47.0 45.1 23.3 33.0 42.4 35.4 35.6 9.0 116.0 49.6 46.5
| Element | Nb  | Ta  | Zr  | Hf  | Mo  | Sn  | Ti  | Pb  | U   | Th  | La  | Ce  | Pr  | Nd  | Sm  | Eu  | Gd  | Tb  | Dy  | Ho  | Er  | Tm  | Yb  | Lu  | Ne  | Di  | Hy  | Ol  |
|---------|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|-----|
|        | 7.2 | 0.6 | 246.4 | 7.5 | 1.5 | 3.1 | 0.1 | 4.4 | 0.5 | 1.8 | 17.1 | 42.6 | 6.6 | 31.4 | 2.8 | 10.0 | 1.7 | 10.8 | 2.4 | 6.2 | 1.0 | 6.0 | 0.9 | 4.8 | 21.9 | 0.0 | 16.0 |
|        | 4.6 | 0.5 | 49.4 | 2.2 | 1.9 | 1.5 | 0.1 | 3.7 | 0.3 | 1.0 | 11.9 | 29.5 | 4.5 | 23.2 | 6.4 | 7.2 | 1.1 | 6.3 | 1.3 | 3.1 | 0.4 | 2.6 | 0.4 | 2.3 | 21.1 | 12.1 | 18.2 |
|        | 3.0 | 0.3 | 56.4 | 4.1 | 0.1 | 2.7 | 0.1 | 7.1 | 0.2 | 0.7 | 9.1  | 18.9 | 3.2 | 19.5 | 4.6 | 4.7 | 0.7 | 4.6 | 1.0 | 2.5 | 0.6 | 2.4 | 0.6 | 3.6 | 10.4 | 13.1 | 23.5 |
|        | 3.6 | 0.6 | 175.8 | 5.5 | 1.5 | 1.3 | 0.1 | 4.2 | 0.3 | 1.1 | 16.5 | 24.8 | 4.0 | 19.3 | 5.8 | 6.5 | 1.1 | 7.0 | 1.5 | 4.0 | 0.6 | 3.8 | 0.6 | 3.0 | 10.4 | 13.1 | 23.5 |
|        | 3.5 | 0.8 | 317.2 | 10.2 | 2.1 | 1.9 | 0.1 | 2.6 | 0.3 | 1.5 | 15.3 | 43.5 | 4.1 | 29.6 | 8.1 | 9.2 | 1.5 | 7.1 | 1.5 | 3.9 | 0.6 | 4.6 | 0.7 | 6.4 | 19.7 | 21.1 | 17.1 |
|        | 10.8 | 0.8 | 310.7 | 8.7 | 1.2 | 2.6 | 0.1 | 2.5 | 0.3 | 2.1 | 18.8 | 38.0 | 5.8 | 25.5 | 6.8 | 8.0 | 1.3 | 9.3 | 2.0 | 2.5 | 0.6 | 5.0 | 0.8 | 3.6 | 10.4 | 13.1 | 23.5 |
|        | 9.0 | 0.8 | 252.5 | 7.6 | 3.5 | 7.3 | 0.1 | 4.5 | 0.4 | 1.0 | 17.7 | 18.5 | 5.8 | 27.0 | 7.2 | 7.8 | 1.0 | 8.3 | 1.8 | 2.3 | 0.6 | 5.6 | 0.7 | 4.0 | 19.7 | 21.1 | 17.1 |
|        | 8.8 | 0.8 | 113.2 | 6.5 | 0.6 | 6.3 | 0.1 | 2.5 | 0.3 | 1.2 | 22.4 | 28.8 | 3.1 | 18.2 | 6.4 | 7.4 | 0.9 | 8.2 | 1.7 | 2.5 | 0.5 | 5.6 | 0.7 | 4.3 | 16.3 | 19.7 | 17.1 |
|        | 3.1 | 0.3 | 127.7 | 3.7 | 0.6 | 6.0 | 0.1 | 2.0 | 0.3 | 1.1 | 17.3 | 27.0 | 1.0 | 15.9 | 5.1 | 5.9 | 1.0 | 7.2 | 1.8 | 2.0 | 0.6 | 5.4 | 0.7 | 4.4 | 16.3 | 19.7 | 17.1 |
|        | 4.2 | 0.4 | 167.6 | 6.6 | 0.3 | 3.5 | 0.1 | 2.7 | 0.2 | 1.1 | 16.7 | 28.7 | 1.0 | 18.8 | 5.1 | 5.1 | 1.0 | 7.0 | 1.8 | 2.0 | 0.6 | 5.4 | 0.7 | 4.3 | 16.3 | 19.7 | 17.1 |
|        | 4.6 | 0.4 | 153.5 | 6.3 | 0.3 | 3.0 | 0.1 | 3.1 | 0.2 | 1.3 | 17.2 | 18.5 | 1.2 | 22.4 | 5.1 | 5.1 | 1.0 | 5.7 | 1.8 | 2.0 | 0.6 | 5.3 | 0.7 | 4.4 | 16.3 | 19.7 | 17.1 |
|        | 2.8 | 0.3 | 157.4 | 6.3 | 0.3 | 3.0 | 0.1 | 3.1 | 0.2 | 1.3 | 18.5 | 28.7 | 1.2 | 17.3 | 5.1 | 5.1 | 1.0 | 5.7 | 1.8 | 2.0 | 0.6 | 5.3 | 0.7 | 4.4 | 16.3 | 19.7 | 17.1 |
|        | 3.7 | 0.3 | 14.1 | 7.4 | 0.3 | 3.0 | 0.1 | 3.1 | 0.2 | 1.3 | 16.7 | 27.0 | 1.2 | 15.9 | 5.1 | 5.1 | 1.0 | 5.7 | 1.8 | 2.0 | 0.6 | 5.3 | 0.7 | 4.4 | 16.3 | 19.7 | 17.1 |
|        | 0.4 | 0.1 | 543.0 | 7.4 | 0.3 | 3.0 | 0.1 | 3.1 | 0.2 | 1.3 | 22.4 | 28.8 | 1.2 | 15.9 | 5.1 | 5.1 | 1.0 | 5.7 | 1.8 | 2.0 | 0.6 | 5.3 | 0.7 | 4.4 | 16.3 | 19.7 | 17.1 |
|        | 21.1 | 0.8 | 212.2 | 4.0 | 0.3 | 3.0 | 0.1 | 3.1 | 0.2 | 1.3 | 22.4 | 28.8 | 1.2 | 15.9 | 5.1 | 5.1 | 1.0 | 5.7 | 1.8 | 2.0 | 0.6 | 5.3 | 0.7 | 4.4 | 16.3 | 19.7 | 17.1 |
|        | 9.1 | 0.7 | 232.2 | 0.7 | 0.1 | 3.0 | 0.1 | 3.1 | 0.2 | 1.3 | 22.4 | 28.8 | 1.2 | 15.9 | 5.1 | 5.1 | 1.0 | 5.7 | 1.8 | 2.0 | 0.6 | 5.3 | 0.7 | 4.4 | 16.3 | 19.7 | 17.1 |

\(X_{\text{Mg}}=\frac{\text{Mg}}{\text{Mg} + \text{Fe}^{2+}}\) assuming Fe\(_2\text{O}_3/\text{FeO} = 0.15.

Ne, Di, Hy, Ol = CIPW-normative nepheline, diopside, hypersthene and olivine.
## Table 2

Major (wt.%) and trace element (ppm) contents of Precambrian cumulates (Group 2) and Lower Palaeozoic amphibolites (Group 3)

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<td>X₅Mg</td>
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<td>0.69</td>
<td>0.67</td>
<td>0.48</td>
<td>0.63</td>
<td>0.63</td>
<td>0.55</td>
<td>0.48</td>
<td>0.44</td>
<td>0.41</td>
</tr>
</tbody>
</table>

Li 16.6 42.8 11.1 18.7 9.5 15.9 30.0 6.4 15.2 9.0
Rb 0.6 1.3 0.4 6.0 0.5 0.7 3.6 0.1 0.4 0.4
Be 0.6 0.3 0.4 0.6 0.7 0.6 1.0 0.8 0.9 0.8
Sr 104.8 119.3 80.6 88.1 58.4 156.5 321.8 117.6 305.3 220.7
Ba 46.0 44.8 39.3 44.8 28.3 46.7 43.4 46.3 35.4 40.1
V 577.6 149.9 182.7 497.1 143.7 365.3 262.6 366.9 240.0 291.3
Cr 222.0 369.1 422.4 18.7 146.6 376.2 342.3 201.2 325.1 179.3
Co 48.1 51.5 46.2 38.3 42.1 49.9 51.6 50.9 50.3
Ni 79.7 112.0 97.7 58.9 58.0 83.3 141.8 59.4 115.2 81.0
Cu 32.4 142.5 34.2 15.7 34.3 7.4 50.5 67.2 85.5 77.5
Zn 78.2 41.8 36.9 84.2 56.3 163.6 157.3 122.7 110.0 139.9
Ga 17.6 16.0 16.2 16.4 16.2 18.7 17.3 19.3 18.6
Y 12.6 9.0 8.1 12.0 10.9 13.1 28.5 36.6 28.9 30.5
Nb 1.1 0.3 0.4 5.3 1.5 1.0 3.9 3.0 10.5 6.9
Ta 0.2 0.1 0.2 0.6 0.2 0.1 0.3 0.3 0.8 0.6
Zr 22.5 12.9 13.0 56.2 30.4 22.1 57.6 106.0 137.5 115.0
Hf 1.0 0.7 0.7 2.6 1.4 1.1 3.0 5.6 7.4 5.2
Mo 0.6 0.3 0.6 1.1 1.3 0.5 0.8 2.2 0.7 0.8
Sn 0.7 0.7 0.3 1.5 3.6 1.2 4.8 1.2 4.1 1.4
Ti 0.1 0.2 0.1 0.1 0.1 0.3 0.1 0.1 0.1 0.1
Pb 4.5 2.6 3.2 2.8 2.5 2.8 9.4 5.4 6.1 19.3
U 0.2 0.0 0.1 0.2 0.2 0.6 0.2 0.2 0.5 0.3
Th 0.5 0.1 0.1 0.3 0.6 0.3 0.4 0.3 1.3 0.7
La 3.8 1.0 1.4 3.0 3.3 1.9 4.0 4.5 10.8 7.8
Ce 6.4 2.6 3.2 6.5 7.4 4.8 10.1 12.9 25.8 19.5
Pr 1.2 0.5 0.5 1.0 1.0 0.8 1.7 2.1 3.7 2.9
Nd 5.2 2.5 2.3 4.9 4.6 4.0 8.8 11.5 16.4 14.5
Sm 1.5 0.9 0.8 1.5 1.4 1.4 2.9 3.9 4.6 4.1
Eu 0.7 0.6 0.5 0.8 0.6 0.7 1.1 1.3 1.7 1.5
Gd 1.9 1.3 1.2 1.8 1.6 1.9 3.9 5.0 5.4 5.0
Tb 0.3 0.2 0.2 0.3 0.3 0.3 0.7 0.9 0.9 0.8
Dy 2.2 1.6 1.3 2.1 1.9 2.3 4.8 6.3 5.7 5.5
Ho 0.5 0.4 0.3 0.5 0.4 0.5 1.1 1.4 1.1 1.2
Er 1.3 1.0 0.8 1.3 1.2 1.4 3.1 4.0 3.2 3.2
Tm 0.2 0.2 0.1 0.2 0.2 0.2 0.5 0.6 0.5 0.5
went a common metamorphic evolution, characterized by an initial eclogite-facies event with peak pressures above 1.9 GPa and temperatures around 750 °C followed by a strong decompression accompanied by cooling to amphibolite facies \((T = 700 \pm 50 \, ^\circ C, \, P = 1.4 \pm 0.2 \, GPa)\) and greenschist facies (López Sánchez-Vizcaíno et al., 2003). Estimates of the ages of the protoliths of these amphibolites were obtained by the Sensitive High Resolution Ion Microprobe (SHRIMP) U–Pb method on zircons (Ordoñez Casado, 1998).

Two groups with different ages and geochemical features will be considered here for geochemical description: (i) Lower Palaeozoic amphibolites and (ii) Upper Precambrian amphibolites. The Lower Palaeozoic ages obtained by Ordoñez Casado (1998) are 483 ± 13 Ma for the Pocico outcrop (Fig. 1b), 490 ± 17 Ma for a sample taken in Portugal near Arronches, and 525 ± 13 Ma for a sample from Arroyo Argallón (Fig. 1b). The Upper Precambrian ages provided by this author are 577 ± 26 Ma for a sample from La Cardenchosa, 596 ± 14 Ma for a sample from the Cerro de las Mesas, and 566 ± 9 Ma for another sample from La Cardenchosa (Fig. 1b).

We must emphasize that these two groups of amphibolites appear in the field with very different dimensions. Upper Precambrian amphibolites crop out as small-size bodies (some centimetres to several metres thick) parallel to the main foliation. By contrast, Lower Palaeozoic amphibolites at Pocico form an 800-m-thick strongly deformed, elongated body with lateral dimensions of several kilometres, lacking any intercalation of orthogneissic rocks. This means that Upper Precambrian amphibolites are likely to represent former dikes or sills intruded in the gneissic lithologies, while Lower Palaeozoic amphibolites may well represent former gabbro stocks emplaced in the middle or lower crust.

3. Geochemistry of amphibolites

In this work we study amphibolites from seven different outcrops (Fig. 1b) belonging to the lower part of the sequence of the Central Unit. Five of these outcrops occur in the southeastern part of the unit (Fig. 1b), near La Cardenchosa (samples BCA, CA and LC), Arroyo Argallón (samples AA) and Cerro de las Mesas (samples CM). The other two outcrops appear more than 50 km to the northwest at Cortijo de Peñarresbalá (samples CPR) and Pocico (samples PC).

Major-element determinations for the bulk rocks were performed by X-ray fluorescence (XRF) (Philips PW1404), calibrated against international standards of appropriate composition, after fusion with lithium tetraborate. Precision was typically better than ±1.5% for a concentration of 10 wt.%. Zirconium was determined by XRF on pressed pellets, with a precision better than ±4% at a 100 ppm level. Trace-element determinations were done by ICP-mass spectrometry (ICP-MS) after HNO3 + HF digestion of 100 mg sample powder in a Teflon-lined vessel at 180 °C and 14 bar during 30 min, evaporation to dryness, and subsequent dissolution in 100 ml of 4 vol.% HNO3. Instrument measurements were carried out in triplicate with a PE SCIEX ELAN-5000 spectrometer using Rh as internal standard. Precision was better than ±2% and ±5% relative for concentrations of 50 and 5 ppm, respectively.

\[X_{Fe} = \frac{[Mg/(Mg + Fe^{2+})]}{Fe_{2}O_{3}/FeO = 0.15.}\]

Ne, Di, Hy, Ol = CIPW-normative nepheline, diopside, hypersthene and olivine.

Table 2 (continued)

<table>
<thead>
<tr>
<th>Sample Group</th>
<th>AA-275</th>
<th>AA-368</th>
<th>BCA-6</th>
<th>CM-6</th>
<th>LC-2</th>
<th>LC-3</th>
<th>AA-217</th>
<th>PC-1</th>
<th>PC-4</th>
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</tr>
</thead>
<tbody>
<tr>
<td>Yb</td>
<td>1.2</td>
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<td>0.7</td>
<td>1.3</td>
<td>1.2</td>
<td>1.4</td>
<td>3.0</td>
<td>3.7</td>
<td>3.1</td>
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<td>Lu</td>
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<td>0.1</td>
<td>0.1</td>
<td>0.2</td>
<td>0.2</td>
<td>0.2</td>
<td>0.5</td>
<td>0.6</td>
<td>0.5</td>
<td>0.4</td>
</tr>
<tr>
<td>Ne</td>
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<td>0.0</td>
<td>12.1</td>
<td>0.0</td>
<td>0.0</td>
<td>0.9</td>
<td>0.0</td>
<td>3.5</td>
<td>7.1</td>
</tr>
<tr>
<td>Di</td>
<td>17.0</td>
<td>17.3</td>
<td>16.4</td>
<td>21.9</td>
<td>13.1</td>
<td>21.9</td>
<td>27.2</td>
<td>23.3</td>
<td>20.2</td>
<td>39.8</td>
</tr>
<tr>
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<td>13.3</td>
<td>25.3</td>
<td>0.0</td>
<td>3.9</td>
<td>11.8</td>
<td>0.0</td>
<td>12.0</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>Ol</td>
<td>10.4</td>
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<td>1.1</td>
<td>12.7</td>
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<td>11.8</td>
<td>17.0</td>
<td>11.8</td>
<td>16.5</td>
<td>8.5</td>
</tr>
</tbody>
</table>

\[\frac{87}{86}Sr/\frac{143}{144}Nd\] analyses were performed using a Finnigan MAT 262 RPQ spectrometer after separation by ion-exchange resins. External precision (2σ) measured in 10 replicates of the standard WS-E (Govindaraju et al., 1994) was approximately ±0.003% relative for \[\frac{87}{86}Sr\] and ±0.0026% relative for \[\frac{143}{144}Nd\]. \[\frac{87}{86}Sr\] and \[\frac{147}{144}Nd\]
proportions were obtained by ICP-MS, with an external precision \((2\sigma)\) better than \(\pm 1.2\%\) and \(\pm 0.8\%\) relative, respectively (Montero and Bea, 1998).

### 3.1. Geochemical classification

According to previous geochronological data (Ordóñez Casado, 1998), two groups of amphibolites can be differentiated: (i) Lower Palaeozoic amphibolites and (ii) Upper Precambrian amphibolites. Tables 1 and 2 show the chemical composition (major and trace elements) of these amphibolites. For classification and plotting purposes, major element data have been recalculated on an anhydrous basis, with FeO/Fe₂O₃ = 0.15.

On the basis of their SiO₂ and alkaline contents, and their CIPW norms, most of the protoliths of these metabasites may be classified as basalts and trachybasalts (Fig. 2a), with subalkaline to moderately alkaline affinity (with up to 5% normative nepheline). However, special care should be taken with this classification because, as will be discussed below, in many of the rocks analysed, substantial evidence can be found that these can be cumulate rocks. In addition, since the rocks have undergone at least one metamorphic event, it can be expected that they should have suffered significant element mobility, especially involving the alkali and LIL elements. Therefore, the alkaline character of some of the igneous protoliths should also be regarded with caution. A better classification can be attempted drawing on those elements less sensible to the metamorphic mobility, such as the high field strength elements (HFSE). Accordingly, the low Nb/Y ratios (< 0.7) can be concluded to reveal a clear tholeiitic affinity in the samples studied (Floyd and Winchester, 1975; Fig. 2b).

### 3.2. Major elements

Most of the samples are characterized by moderate to high Al₂O₃ (12.56–19.59 wt.%), and CaO (7.43–13.09 wt.%) contents, and variable TiO₂ (0.39–3.34 wt.%) and P₂O₅ contents (0.01–0.25 wt.%). On a Mg-number \([i.e. X_{Mg} = \text{molar MgO/(MgO + FeOt)}]\) vs. TiO₂ plot, there are three groups of samples that can be distinguished (Fig. 2c): (1) a group with high Mg-number and low to high TiO₂, interpreted as cumulates (see below), (2) high Ti metabasaltic samples (TiO₂>2%), and (3) low Ti metabasaltic samples (TiO₂ < 2%). In general, a neat correlation can be established between TiO₂ and P₂O₅ contents. Two samples show very high
P₂O₅ (1.96 wt.%) and TiO₂ (6.54 wt.%) contents, which can be explained in terms of an extreme accumulation of apatite and ilmenite (and/or hornblende), respectively. In fact, abnormal abundance of apatite and a TiO₂-rich phase (rutile) is petrographically observed in some of these amphibolites. The Na₂O and K₂O contents are variable and range between 0.92–4.78 and 0.01–1.96 wt.%, respectively. The MgO and Fe₂O₃t contents range from 4.77 to 9.36, and from 7.75 to 15.75 wt.%, respectively. There is not any clear regular variation of major elements with any common fractionation index, although a rough increase in Fe₂O₃, Na₂O, TiO₂ and P₂O₅, and a decrease in Al₂O₃ and CaO contents with increasing Zr or decreasing XMg may be observed (Fig. 3), with the exception of the Pocico samples (Lower Palaeozoic amphibolites) and other samples interpreted here as cumulates. A better positive correlation can be observed between Zr and TiO₂ and P₂O₅, which indicates an effective ilmenite and/or Ti-rich amphibole fractionation except

Fig. 3. Selected major and trace element bivariate plots for the studied rocks using Zr as differentiation index. Symbols indicate samples from different outcrops. Light shaded area marks the position of Group 2 rocks, interpreted as cumulates (see text). Dark grey line highlights the trend of Group 3 (Lower Palaeozoic) amphibolites.
for cumulates samples. These observations highlight the importance of fractional crystallization in the petrogenesis of the rocks studied, but also imply the existence of several magmatic lineages, deriving from different parental magmas, and probably originated in different scenarios (see below).

3.3. Trace and rare earth elements

There cannot be said to exist a good correlation between most minor elements and Zr (or $X_{Mg}$). As a matter of fact, this is to be expected if one takes into account the wide extent of ages of the different Upper Precambrian amphibolites. The best linear correlation is between REE, Ta (Nb) and Y with respect to Zr (Fig. 3). According to the chondrite-normalized REE plots shown in Fig. 4, three groups of rocks can be distinguished: one corresponds to Lower Palaeozoic amphibolites and the other two (liquids and cumulates, see below) to Upper Precambrian amphibolites.

Some of the Upper Precambrian amphibolites (Group 1 in the following) represent magmatic liquids and display roughly parallel, slightly LREE-enriched patterns [(La/Yb)$_N$ = 1.35–3.07], with the REE global content ranging between 20 and 70 chondrite, and commonly a relative depletion in La and Ce (Fig. 4a and b) that may be explained by the extraction of LREE-rich minerals such as apatite. There is not any apparent Eu anomaly. This fact indicates that the generation of this group cannot simply be explained in terms of different degrees of fractional crystallization. The uniform increase in global REE contents may
be the result of shallow magmatic differentiation. However, there cannot be established any clear correlation between total REE content and $X_{\text{Mg}}$, neither in the same outcrop nor in all the sampled outcrops. In fact, the lowest REE contents are not always represented by samples with the lowest $X_{\text{Mg}}$ (CA-2, Fig. 4b) or, alternatively, almost identical REE contents are displayed by samples with different $X_{\text{Mg}}$ (BCA samples, Fig. 4b). This, together with the slightly sigmoidal LREE patterns, can be taken to indicate that there is not a simple fractionation relationship between the different samples. Therefore, the samples concerned do not represent a real liquid-line of descent (see Fig. 3), probably reflecting that most of them must have undergone different degrees of crystal accumulation. In addition, most of the samples are high Ti mafic rocks, but low Ti ones are also present, thus implying the existence of several lines of migmatic evolution from different magmatic parental batches.

The remaining Upper Precambrian amphibolites samples (Group 2 in the following) display the lowest total REE contents (6–11 chondritic values, Fig. 4c), almost flat to slightly LREE-depleted patterns ($\text{La/Yb}}_N=0.730–1.823$) and a distinct positive Eu anomaly ($\text{Eu/Eu}^*_N=1.20–1.64$). These patterns may be produced by the accumulation of a mixture of plagioclase ($\text{La} > \text{Yb}$), pyroxene ($\text{La} < \text{Yb}$) and minor olivine ($\text{La} = \text{Yb}$), in liquids slightly LREE richer than E-MORB. The high Mg-number and $\text{SiO}_2$ content, the compatible element contents, such as Cr and Ni, and the lower contents in all the incompatible elements (see Fig. 3 and Table 2) are consistent with a cumulate assemblage of pyroxene, plagioclase, and minor olivine.

Lower Palaeozoic amphibolites (Group 3 hereafter) were sampled in the Pocico outcrop. Other samples from the Arroyo Argallón outcrop (AA-217) show similar geochemical characteristics and have been tentatively considered as belonging to the same group. These rocks are mostly low Ti ($\text{TiO}_2 < 2\%$) mafic rocks. Two samples are slightly LREE-depleted ($\text{[La/Yb}}_N=0.82–0.90$), and have 18–20 chondritic total REE contents. HREE patterns are flat, and cut some of the patterns of Group 1 amphibolites (Fig. 4d). A minor negative Eu anomaly can be observed in sample PC-1. These features are similar to transitional MORB patterns (Saunders, 1984). The other PC samples display a similar geochemical signature, despite their LREE-enriched patterns ($\text{[(La/Yb}}_N=1.77–2.35$; Fig. 4c). These differences in ($\text{La/Yb}}_N$, however, can easily be explained by a variable degree of melting of the same mantle source.

MORB-normalized diagrams of trace element abundances are displayed in Fig. 5. The three groups defined on the basis of REE data show also distinctive trace element patterns.

Samples from Group 1 display gradually increasing contents from La to Rb, with values always exceeding MORB abundances (Fig. 5a and b). LIL elements (Li, Rb, Ba, Cs, Th, U and Pb) are variably enriched, and there is a minor but significant negative anomaly in Nb. Compatible element contents are lower than typical MORB values. Sample AA-364 (Fig. 5a) shows a strong positive P anomaly and high Ti content. This amphibolite is very rich in apatite and, to a lesser extent, ilmenite, which were probably accumulated together with plagioclase, according to its high $\text{Al}_2\text{O}_3$ and $\text{CaO}$ and low $\text{MgO}$, $\text{Cr}$ and $\text{Ni}$ contents (Table 1).

The patterns of Group 2 samples are more variable than the other amphibolites but commonly display more moderate enrichments in alkaline metals and Ba. U and Th are relatively depleted with respect to other LIL elements (Fig. 5c and d). Incompatible HFS and other immobile elements are strongly depleted. Compatible element contents (Cr, Ni and Sc) are variable, some of these samples showing higher values than in Group 1 samples. The Sr contents are similar to Group 1 amphibolites, while P outlines a clearly marked negative peak. There is also a remarkable negative Nb anomaly. The differences in chemical composition among samples of this group can be explained in terms of variable plagioclase, olivine, pyroxene, and magnetite modal contents in the original protolith.

The patterns of Group 3 samples are more variable than the other amphibolites but commonly display more moderate enrichments in alkaline metals and Ba. U and Th are relatively depleted with respect to other LIL elements (Fig. 5c and d). Incompatible HFS and other immobile elements are strongly depleted. Compatible element contents (Cr, Ni and Sc) are variable, some of these samples showing higher values than in Group 1 samples. The Sr contents are similar to Group 1 amphibolites, while P outlines a clearly marked negative peak. There is also a remarkable negative Nb anomaly. The differences in chemical composition among samples of this group can be explained in terms of variable plagioclase, olivine, pyroxene, and magnetite modal contents in the original protolith.

The samples of Lower Palaeozoic amphibolites (Group 3) show patterns rather similar to those of Group 1 (Fig. 5c), with a somewhat lower enrichment in LILE and flat patterns for the immobile elements. There is also a very small negative Nb anomaly in some of the samples. Compatible elements are similar to N-MORB, and higher than in Group 1.

In short, the trace element contents may be explained in terms of the different modal proportion of cumulate phases (probably plagioclase, pyroxene, and minor olivine, apatite and ilmenite) in the original rocks. Thus, Groups 1 and 3 probably correspond to liquid-rich, porphyritic rocks with widely variable, but
low, phenocryst contents, while Group 2 samples are clearly cumulate rocks of the phases in question with minor trapped liquid. The Ba, Sr, Rb and other LIL elements are always relatively enriched and their abundances do not define any regular trend with common differentiation indexes (either using Zr or Mg-number), thus indicating the post-magmatic mobility of these elements.

3.4. Sr–Nd isotope geochemistry

Sr and Nd isotopes have been measured on samples from the three groups of rocks previously defined. Table 3 shows the analytical results. Geochronological and Nd isotope data from samples of the same outcrops have been reported by Ordoñez Casado (1998). The ages proposed by this author have been used in Table 3 to compute initial isotopic ratios and epsilon values.

Fig. 6 shows the variation in Sr and Nd initial values. εNd varies within the range 2.12–7.53, corresponding to a slightly to moderately depleted source. Sample CPR-12 from Group 1 has an initial εNd of 6.31. Sample PC-1 from Group 3 amphibolites shows the most depleted character (εNd = 7.53), a value close to the Depleted Mantle (DM) component for the considered age. The εNd values for Group 2 range from 2.12 to 5.64. The widespread Nd initial ratios indicate the participation of a more enriched component in their genesis. All the samples have relatively high Sr initial ratios and plot slightly to the right of the mantle array, recalculated for the range of ages considered in this work (483–595 Ma). Two samples show very high εSr values, and plot in the....

Fig. 5. MORB-normalized plot for the samples under study (normalizing values modified after Pearce (1982)). (a, b) Plots for the Group 1 samples. (c) Plot for the Group 2 samples, interpreted as cumulates, and for the Group 3 samples (Lower Palaeozoic amphibolites). (d) Comparison of the three groups.
upper right quadrant of the $^{143}\text{Nd}/^{144}\text{Nd}–^{87}\text{Sr}/^{86}\text{Sr}$ diagram. These high $\varepsilon$Sr values may be due either to crustal contamination with a low-Nd high-Sr contaminant, or most probably to post-magmatic alteration. A likely contamination prevents the use of their Sr isotope composition as a mantle source petrogenetic indicator. However, their Nd initial ratios cannot be taken to be seriously affected by this alteration and, what is more, they plot within the range depicted by the rest of samples.

3.5. Petrogenesis

In the previous sections, we have shown that the compositional features of some of the Upper Precambrian amphibolites indicate a cumulative origin for these rocks. For this reason, in what follows we will concentrate on the origin of the rest of Upper Precambrian amphibolites (Group 1) and all of the Lower Palaeozoic ones (Group 3), which are more representative of the original basaltic melts.

The $(\text{La}/\text{Yb})_N$ values for both groups of liquids are low (Upper Precambrian samples = 1.28–3.07, Lower Palaeozoic samples = 0.82–2.35) and show a narrow variation. This is consistent with a generation by partial melting of a mantle source in which garnet does not remain as a residual phase. Therefore, the magma generation depth should have been at < 60–80 km, within the spinel peridotite field (Watson and McKenzie, 1991). This is also consistent with the narrow ranges in Sc/V and Ti/Y (see Tables 1 and 2).

Fractional crystallization is unable to produce significant variations in incompatible element ratios. Instead, such variations can be produced in response to different degrees of partial melting from a homogeneous source. Accordingly, the positive slope in Fig. 7a displayed by Group 3 samples can be attributed to variable degrees of partial melting from a common source, with higher melting degree in the lower end of this array, where the most depleted samples plot (PC-1 and AA-217). For Group 1 samples, the overall tendency is horizontal, which can be attributed to the predominance of low pressure frac-
tional crystallization. Nevertheless, there is a slight spread of Ce/Yb ratio, which could reflect minor differences in the degree of melting or source heterogeneity. The Cr vs. Y plot in Fig. 7b (Pearce, 1982; Pearce and Parkinson, 1993) also shows that the variations among Group 1 samples are better explained in terms of variable degrees of fractional crystallization, while Group 3 samples may have been produced by a slightly higher degree of melting (likely at shallower depths). Both groups plot in the overlapping field of MORB and/or within-plate (WP) basalts proposed by Pearce (1982), and outside the volcanic arc basalt (VAB) field.

The variations observed in Fig. 8a for Group 1 amphibolites can also be better explained in terms of an evolution dominated by low pressure fractional crystallization from MORB-like asthenospheric melts, comparable to Group 3 amphibolites, the most depleted samples. These variations in the Group 1 rocks are also compatible with addition of crustal material to MORB-like, tholeiitic basalt. Those crustal materials which are likely to supply this component could be similar to some of the nearby Upper Precambrian orthogneisses, which appear intercalated with the amphibolites in the lithostratigraphic sequence (Azor et al., 1995). The addition of this Zr-rich material could be taken to be responsible for

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**Fig. 7.** (a) Ce vs. Ce/Yb diagram for Groups 1 and 3; see text for further explanation. (b) Y vs. Cr diagram (Pearce, 1982). Most samples plot in the MORB and within-plate basalt (WPB) fields, with the exception of three very differentiated samples which are outside the volcanic arc basalt (VAB) field.

**Fig. 8.** (a) Zr vs. Ce/Zr diagram for the samples studied. (b) Sm vs. Ce/Yb diagram. Arrows show the trends of Groups 1 and 3. Composition of E-MORB, N-MORB, PM, OIB and Continental Crust according to Sun and McDonough (1989) and Taylor and MacLennan (1985). The effects of fractional crystallization and zircon addition are also shown.
the rather negative slope of the observed trend. The same effect is also observed in Fig. 8b.

Additional information regarding the mantle source of the basaltic magmas can be obtained from the diagrams in Fig. 9. In these diagrams, the observed trends can be explained by considering at least three components:

1. A depleted, N-MORB-like asthenospheric source is predominant in Group 3. Many geochemical features of these rocks match those of average N-MORB magmas, such as their high initial Nd isotope ratio ($\varepsilon_{\text{Nd}(483)} = 7.53$). In Fig. 10 we have represented the immobile element composition normalized to the fertile MORB mantle (FMM) of Pearce and Parkinson (1993). The plot shows a negative slope with very highly incompatible (VHI) elements $>$ highly incompatible (HI) elements $>$ moderately incompatible (MI) elements. These patterns are usually produced by a moderate melting degree from an undepleted FMM source (Pearce and Parkinson, 1993).

2. A more enriched source is required to explain the shift towards higher Ce/Yb, Ce/Nb, Ti/Y, Zr/Y and Ta/Yb, and to lower Zr/Nb and Y/Nb ratios (Figs. 7–9). The influence of a more enriched component is more evident in the Group 1 rocks than in those of Group 3. This enriched source could be a (plume-related?) OIB, an E-MORB component present in the asthenosphere source or, alternatively, a subcontinental lithospheric mantle (SCLM). The isotopic variation towards lower $\varepsilon_{\text{Nd}}$ values in the analysed samples is also consistent with a mixing of melts from depleted and enriched reservoirs. This fact is more evident again in the Group 1 amphibolites.

3. A limited amount of crustal material is required to produce the variation towards higher Th, Zr, Zr/Y, Th/Nb and Th/Yb, and lower Ce/Zr values observed in some samples (Figs. 8 and 9). The nature of this crustal material is unknown, but we have argued previously that it could be similar to some of the nearby Upper Precambrian orthogneisses. In Fig. 9c, we have represented in a Ta/Yb vs. Th/Yb plot the average upper continental crust value from Taylor and MacLennan (1985). Interestingly, most of the Groups 1 and 3 samples plot on the mantle array, between the compositions from Sun and McDonough (1989) and Taylor and MacLennan (1985).
tions of typical N- and E-MORB, while some Group 1 samples show an increase in the Th/Yb ratio, a feature commonly attributed to the introduction of a “subduction” component. The same variation could alternatively be obtained by mixing asthenospheric basalts with acidic rocks (arrow “a” in Fig. 9c) or with average upper crust (arrow “b” in Fig. 9c). The slightly negative Nb anomaly could also be produced in the same way (Wilson, 1989). The enrichment in LIL elements displayed in the MORB-normalized diagrams (Fig. 5) could also be attributed to a “subduction component” or to mixing with crustal melts. However, this should be considered with caution, because the original LILE contents could have been largely modified by metamorphism, due to the mobility of these elements.

To conclude, the different geochemical features indicate that Lower Palaeozoic amphibolites (Group 3) have numerous features similar to present-day MORB magmas and derive from a more depleted source than Upper Precambrian amphibolites (Group 1). Furthermore, some of the Group 1 amphibolites are more affected by crustal contamination than Group 3 ones.

4. Discussion

Despite the limited outcrop extent of amphibolites in the Central Unit, their origin and tectonic signifi-
cance are key aspects to fully understand the significance of the OMZ/CIZ boundary. According to previous geochronological data (Ordóñez Casado, 1998), two groups of amphibolites can be differentiated: (i) Upper Precambrian amphibolites and (ii) Lower Palaeozoic amphibolites. Both types of amphibolites underwent a common tectono-metamorphic evolution during the Variscan orogeny. Subduction and continental collision were followed by fast exhumation owing to extensional processes (Azor et al., 1994; López Sánchez-Vizcaíno et al., 2003).

Our geochemical data have revealed that the two groups of amphibolites also have different origins. As for Upper Precambrian amphibolites (Group 1), we envisage an advanced intra-continental rifting situation or an intra-continental back-arc setting related to the Cadomian orogeny, while for Lower Palaeozoic ones (Group 3), we propose an environment of oceanic crust formation related to the Variscan orogeny. Ordóñez Casado (1998) proposed a similar scenario for both groups of amphibolites drawing on the 143Nd/144Nd isotopic values.

Basalts ranging from depleted to slightly enriched, such as those of Group 1, can be generated in an intra-continental rifting stage (including intra-continental back-arc settings) (cf. Holm, 1985). In modern analogues, this setting usually includes alkaline as well as tholeiitic associations, depending on the degree of extension in the rifting process and the source composition. In areas of advanced degree of extension, asthenospheric melting would produce depleted tholeiitic suites, similar to those from Group 3. However, the passage of basaltic magmas through thinned subcontinental lithosphere and continental crust is usually reflected in the composition of the final magmas, which then incorporates an enriched and/or crustal component. As described above, these enriched and crustal components, superimposed on an overall MORB chemistry, are clearly present in the trace element and isotopic composition of Upper Precambrian amphibolites. In this respect, these amphibolites are regionally related to felsic rocks of alkaline affinity located near to La Cardenchosa, although their age is poorly constrained [690 ± 134 Ma, Rb–Sr/whole rock method; Azor et al. (1995)]. It is clear, however, that magmas generated in this setting incorporate components from far too many sources, which cannot be easily discriminated by their geochemical signature.
The geochronological data available (cf. Ordoñez Casado, 1998) record long-lasting, likely pulsating magmatic activity in the Upper Precambrian. Thus, the overlapping of temporally different magmatic events further contribute to the complexity of the chemical features, and explain the scatter observed in the geochemical plots.

Lower Palaeozoic amphibolites (Group 3) have compositions similar to N- or E-MORB recent basalts. Accordingly, the trace element and isotopic data point to a moderate melting of depleted asthenosphere as the main source of this group of samples. The transition from positive-slope, depleted REE patterns (e.g. sample PC-1, Table 2) to negative, slightly LREE-enriched patterns (Fig. 5), may be effectively related to variable melting degrees from a homogeneous source, or by involving an enriched mantle component, which can be either a plume component present in the asthenosphere (OIB or E-MORB), or subcontinental lithospheric mantle (SCLM). The trace element and isotopic data available are unfortunately not conclusive enough so as to distinguish between these possibilities. In addition, both the REE pattern and the depleted isotopic signature of sample PC-1 indicate that the amount of crustal addition, if present at all, should be very low. A number of geodynamic scenarios are coherent with the features of the Early Palaeozoic basic magmas: a mid-ocean ridge, a very extended ensialic rift, and a back-arc rift. Nevertheless, other geological data (mainly stratigraphical) point to a mid-ocean ridge as the most likely situation. In this respect, the Lower Palaeozoic succession of the OMZ denotes the beginning of a rifting process in Cambrian times with rift-related sediments and widespread basaltic volcanism in Middle Cambrian (cf. Liñán and Quesada, 1990; Giese and Bührn, 1993). In this context of progression of the rift during Ordovician times, the formation of an oceanic crust of this age seems to be quite plausible (cf. Pin, 1990; Crowley et al., 2000; Floyd et al., 2000), although a very extended continental rift should not be completely ruled out. Moreover, the relatively large dimensions of the Pocico basic body and the presence of dismembered ultramafic bodies in nearby areas (Apaleategui et al., 1988) are also in agreement with this hypothesis. In this respect, the presence of a magnetic anomaly just to the NE of the OMZ/CIZ boundary in the Hornachos area (Fig. 1b) has been related to a 3- to 6-km-thick body of basic and/or ultramafic rock located at depth (Aller, 1996). This anomaly may well indicate the existence at depth of huge volumes of basic and ultramafic rocks similar to those cropping out in the Central Unit. Their present position to the NE of the OMZ/CIZ has been related to subduction of the OMZ crust underneath the CIZ one (Azor et al., 1994; Aller, 1996; Simancas et al., 2001).

The recognition of remains of oceanic-affinity basic rocks along the OMZ/CIZ boundary changes the interpretation of this major contact of the Variscan belt in SW Iberia. This contact marks a change in vergence where high-pressure metamorphism and intense ductile deformations took place during the Variscan orogeny. On these grounds, Simancas et al. (2001) propose that this boundary is a suture contact where significant intra-continental subduction of OMZ lower crust under CIZ crust took place in Devonian–Early Carboniferous times, thus accommodating large amounts of shortening in the OMZ upper crust. Drawing on the new geochemical data, we can propose in this respect that some sort of incipient ocean between the OMZ and CIZ blocks may probably have existed during Early Palaeozoic times and that the ocean in question was closed later, probably during the Devonian, by oceanic subduction under the CIZ crust.

The chronology favoured here, with ocean generation in Ordovician times and ocean subduction in Devonian times, fits in neatly with the chronology recently proposed by Ordoñez Casado et al. (2001) for the Cabo Ortegal complex in NW Iberia. This complex, as the Central Unit, includes oceanic-affinity rocks having undergone a high-pressure/high-temperature metamorphism during the Variscan orogeny. In addition to this similar chronology, it must be stressed that the Central Unit (Badajoz–Córdoba Shear Zone in previous works) has been considered by some authors as the root zone of the allochthonous complexes with oceanic rocks outcropping in NW Iberia (e.g. Matte, 1986). However, the different subduction polarities in NW and SW Iberia have led Simancas et al. (2002) to propose that both domains were connected in pre-orogenic times by a transfer fault coincident with the present position of the Porto–Tomar Fault. This hypothesis is reinforced now with the new data provided here, which are consistent with the existence of an oceanic domain between OMZ and CIZ in Lower Palaeozoic times.
The situation described, with basic-dominated magmatism associated to rifted continental margins, eventually leading to ocean production, has also been recognized in other sectors of the Variscides, such as the N Bohemian Massif, the Massif Central, the Maures Massif, and NW Iberia (cf. Piboule and Briand, 1985; Pin and Marini, 1993; Briand et al., 1995, 2002; Santos Zalduegui et al., 1996; Crowley et al., 2000; Floyd et al., 2000; Kryza and Pin, 2002), being commonly interpreted as a consequence of an episode of considerable rifting and fragmentation of the northern Gondwana margins in Early Palaeozoic times. In these regions, the metabasites observed in different crustal blocks include variably depleted and enriched tholeiitic basalts as well as alkali basalts. These rocks are often associated with felsic rocks, resulting in bimodal suites in some of the aforementioned Variscan massifs. The geochemical variations of the metabasites have been attributed to a number of factors, such as variable participation of a sediment-contaminated lithospheric mantle, a possible contribution of a plume component, crustal contamination, and different degrees of lithospheric thinning and ocean-floor production.

5. Conclusion

Amphibolites included in the OMZ/CIZ contact appear as small-size dike or lens-shaped bodies of Upper Precambrian age and an elongated hectometre-thick body of Lower Palaeozoic age. According to their geochemical features, Upper Precambrian amphibolites seem to be generated in an advanced intra-continental rift or an intra-continental back-arc setting related to the Cadomian orogeny. Lower Palaeozoic amphibolites were formed, more likely than not, in an oceanic scenario which predated Variscan subduction and collision. A situation of widely extended continent rift cannot, however, be completely discarded for these amphibolites. In Devonian times, the oceanic domain was consumed by subduction underneath the CIZ, as attested by eclogite facies assemblages preserved in the amphibolites. Finally, exhumation of the high-pressure rocks must have taken place in the Late Devonian–Early Carboniferous by means of an intense ductile shearing. The Lower Palaeozoic oceanic domain, represented by some of the amphibolites studied, was probably connected with the one represented by the oceanic-affinity rocks included in the allochthonous complexes of NW Iberia.

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