

Zircon ages of the metavolcanic rocks and metagranites of the Ollo de Sapo Domain in central Spain: implications for the Neoproterozoic to Early Palaeozoic evolution of Iberia

P. MONTERO*, F. BEA*§, F. GONZÁLEZ-LODEIRO†,
C. TALAVERA* & M. J. WHITEHOUSE‡

*Department of Mineralogy and Petrology, Campus Fuentenueva, Univ. Granada, 18002 Granada, Spain

†Department of Geodynamics, Campus Fuentenueva, Univ. Granada, 18002 Granada, Spain

‡Swedish Museum of Natural History, Box 50007, SE-104 05 Stockholm, Sweden

(Received 8 October 2006; accepted 21 February 2007)

Abstract – Dating the pre-Middle Ordovician metavolcanic rocks and metagranites of the Ollo de Sapo Domain has, historically, been difficult because of the small compositional variation, the effects of the Variscan orogeny and, as revealed in this paper, the unusually high fraction of inherited zircon components. The first reliable zircon data (U–Pb ion microprobe and Pb–Pb stepwise evaporation) indicate that the Ollo de Sapo volcanism spanned 495 ± 5 Ma to 483 ± 3 Ma, and was followed by the intrusion of high-level granites from 483 ± 3 Ma to 474 ± 4 Ma. In both metavolcanic rocks and metagranites, no less than 70–80 % of zircon grains are either totally Precambrian or contain a Precambrian core overgrown by a Cambro-Ordovician rim. About 80–90 % of inherited zircons are Early Ediacaran (602–614 Ma) and derived from calc-alkaline intermediate to felsic igneous rocks generated at the end of the Pan-African arc–continent collision. In the Villadepera region, located to the west, both the metagranites and metavolcanic rocks also contain Meso-Archaean zircons (3.0–3.2 Ga) which ultimately originated from the West African Craton. In the Híndelencina region, located to the east, both the metagranites and metavolcanic rocks lack Meso-Archaean zircons, but they have two different inherited zircon populations, one Cryogenian (650–700 Ma) and the other Tonian (850–900 Ma), which suggest older-than-Ediacaran additional island-arc components. The different proportion of source components and the marked variation of the $^{87}\text{Sr}/^{86}\text{Sr}_{\text{init}}$ suggest, at least tentatively, that the across-arc polarity of the remnants of the Pan-African arc of Iberia trended east–west (with respect to the current coordinates) during Cambro-Ordovician times, and that the passive margin was situated to the west.

Keywords: magmatism, Ordovician, zircon, dating, U/Pb, Pb/Pb.

1. Introduction

Growing geochronological evidence indicates that the Cambro-Ordovician magmatism of South and Central Europe was more voluminous than previously believed (e.g. Delaperrière & Respaut, 1995; Schaltegger & Gebauer, 1999; Valverde & Dunning, 2000; Barbey, Cheilletz & Laumonier, 2001; Deloule *et al.* 2002; Schaltegger, Abrecht & Corfu, 2003; Friedl *et al.* 2004; Mazur, Turniak & Bocker, 2004; Roger *et al.* 2004; Helbing & Tiepolo, 2005). Igneous rocks of this age are well represented in the Central Iberian Zone where pre-Middle Ordovician felsic vulcanites and granitoids, strongly deformed and variably metamorphosed during Variscan times, crop out following three NW–SE roughly parallel curved lineaments (Fig. 1): the first, and most voluminous, comprises the metavolcanic rocks and metagranites of the Ollo de Sapo Formation (Parga-Pondal, Matte & Capdevila, 1964), located near the boundary with the Western-Astur Leonian Zone; the second comprises the metagranites of the northern

part of the Schist-Greywacke Complex (Valverde & Dunning, 2000; Bea, Montero & Zinger, 2003; Zeck *et al.* 2004); and the third, and least voluminous, includes the metavolcanic rocks and the metagranites of the Urro Formation (Solá *et al.* 2006), at the boundary with the Ossa-Morena Zone.

The Ollo de Sapo (Toad's Eye) Formation crops out in the core of a Variscan anticlinorium extending for about 600 km from the west Cantabric Coast to central Spain (Fig. 1a, b) (see a recent overview in Díaz Montes *et al.* 2004). The metavolcanic rocks, which originally consisted of dacitic to rhyolitic ignimbrites and tuffs (Navidad, Peinado & Casillas, 1992), are currently represented by augen-gneisses with large, and locally rapakivi, megacrysts of K-feldspar, within a fine- to medium-grained and strongly foliated felsic peraluminous groundmass. The metagranites, which originally consisted of small high-level granites, are currently represented by coarse-grained augen-gneisses with intrusive contacts and recognizable aplopegmatitic dykes and enclaves (González-Lodeiro, 1981a; Iglesias & Ribeiro, 1981).

§Author for correspondence: fbea@ugr.es

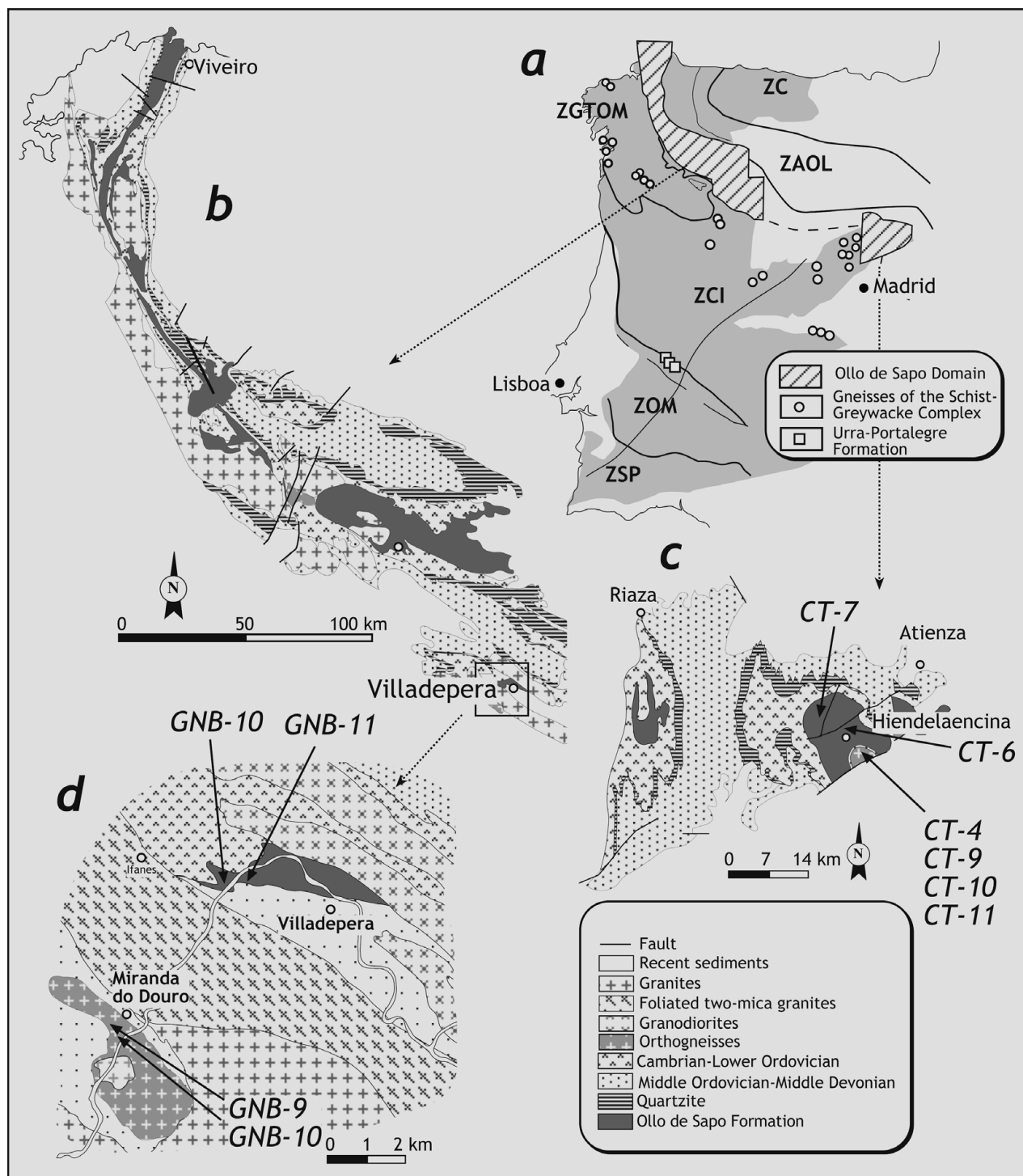


Figure 1. Geological scheme of the Ollo de Sapo Domain and maps of the studied regions. In the large-scale map (a), the zones of the Iberian massif (in grey), according to Farias *et al.* (1987), are ZC – Cantabrian Zone; ZAOL – Western-Astur Leonian Zone; ZGTOM – Galicia Tras-os-Montes Zone; ZCI – Central Iberian Zone; ZOM – Ossa Morena Zone; ZSP – South Portuguese Zone. (b) and (c) represent the Ollo de Sapo Domain, and (d) is a detail of the Villadepera region. Alphanumeric labels indicate sample localities; see Table 1.

The age of the Ollo de Sapo metavolcanic rocks is still poorly known, with estimates ranging from Ediacaran to Ordovician. Stratigraphically, they are located below siliciclastic series of probable Arenig age, but the age of these sediments has never been precisely determined. Radiometrically, no attempt to date them by Rb–

Sr, Sm–Nd or conventional U–Pb on zircon concentrates has succeeded so far. Only the El Cardoso gneiss in the eastern Guadarrama mountains (to the south of Riaza, Fig. 1c), which although it does not belong to the Ollo de Sapo Formation bears many similarities to it, has been dated at 480 ± 2 Ma (U–Pb on zircons:

Valverde & Dunning, 2000). Equally problematic is the age of the spatially associated metagranites, in which U–Pb dating of zircon concentrates has characteristically yielded highly discordant and very imprecise ages, from 618 Ma to 465 Ma (Lancelot, Allegret & Iglesias, 1985; Wildberg, Bischoff & Baumann, 1989).

Recently, Bea *et al.* (2006) revisited the Miranda do Douro metagranite using cathodoluminescence imaging and single zircon U–Pb microanalysis. These authors obtained a concordant crystallization age of 483 ± 3 Ma and concluded that the upper intercept at 618 ± 9 Ma found by Lancelot, Allegret & Iglesias (1985) reflects that about 70–80 % of zircon grains have pre-magmatic cores, most of them with ages clustering around 605 Ma. Such unusually high zircon inheritance seems to be a feature of the magmatism in the Ollo de Sapo Domain (see Section 4) and, undoubtedly, was one of the main reasons for the historical difficulties in dating this unit using zircon concentrates. On the other hand, when suitable microanalytical techniques are used, zircon inheritance can be a useful window to look at the history of the magmatic sources, especially in those terranes such as central Iberia, where later orogenies have obliterated all direct evidence.

This paper is a U–Pb ion microprobe, U–Pb laser ablation ICPMS, and Pb–Pb stepwise evaporation study aimed at determining the crystallization age and the protolith history of the Ollo de Sapo metavolcanic rocks and related metagranites in central Spain (regions of Hiendelaencina and Villadepera, Fig. 1c, d). We have three main objectives: (1) the Ollo de Sapo metavolcanic rocks at Hiendelaencina, (2) the spatially-associated Antoñita metagranite and (3) the Ollo de Sapo metavolcanic rocks at Villadepera, near the already dated Miranda do Douro metagranite. In Hiendelaencina, where the Ollo de Sapo volcanites are thicker, we studied three samples with different stratigraphic positions, trying to determine the time span of the volcanism. One sample consisted of zircon included in K-feldspar megacrysts, which were also dated with a Rb–Sr internal isochron. The new radiometric data combined with element geochemistry and Sr and Nd isotopes enabled us to identify different components in the Pan-African rocks from which the Cambro-Ordovician magmas were derived. Lastly, the spatial distribution of these components permitted us to discuss the across-arc polarity of the Pan-African magmatism in central Iberia.

2. Geological setting and petrography

2.a. Hiendelaencina region

In the Hiendelaencina region, the Ollo de Sapo Formation crops out in the core of a late Variscan antiform (Fig. 1). It is located below a siliciclastic series of Ordovician to Early Devonian age (Schäfer, 1969; Bultynck & Soers, 1971) and over a sequence of

metapelites and metapsammities with some levels of marbles and amphibolites of probable Early Cambrian age (González-Lodeiro, 1981a). Emplaced within this underlying metasedimentary sequence there is a body of metagranites customarily named the Antoñita gneiss (Schäfer, 1969).

The Ollo de Sapo Formation consists of fine-grained and coarse-grained augen-gneisses, mostly derived from felsic volcanic rocks, interbedded with micaceous schists, sandstones and quartzites, with a total thickness of about 2 km. The coarse-grained augen-gneisses crop out preferentially in the lower part of the series. They contain huge (> 10 cm) K-feldspar megacrysts, locally with rapakivi structures, euhedral oligoclase phenocrysts (up to 3 cm), and rounded and frequently embayed phenocrysts of quartz (up to 1.5 cm) which, when the metamorphic grade is low, have a noticeable blue colour due to inclusions of sagenitic rutile. The phenocrysts are surrounded by a fine-grained groundmass of quartz, K-feldspar, muscovite, biotite and occasional albite. The accessory assemblage consists of apatite, zircon, Fe–Ti oxides, monazite, rare xenotime and irregularly distributed Fe–Cu sulphides.

The fine-grained augen-gneisses are concentrated in the upper part of the series. They consist of polycrystalline (up to 2 cm) augen of K-feldspar or, rarely, plagioclase and rounded (< 2 cm) bluish quartz phenocrysts within a groundmass identical to the augen-gneisses. Where the deformation and the metamorphism are less intense, the gneisses present recognizable volcanic structures, ignimbritic in the coarse-grained, and tuffaceous in the fine-grained varieties (Navidad, Peinado & Casillas, 1992).

The associated Antoñita metagranite is a sill-like intrusive body with sharp contacts and a thickness of 300–400 m (González-Lodeiro, 1981b). It is composed of coarse-grained gneisses with abundant aplopegmatitic dykes, which are also gneissic, and conspicuous xenoliths of metasediments. The major minerals are large crystals of K-feldspar (up to 5–7 cm) within a coarse-grained groundmass formed of quartz, oligoclase, K-feldspar, biotite, muscovite and occasional tourmaline and garnet. Accessory minerals consist of apatite, Fe–Ti oxides, zircon, monazite and rare xenotime and huttonite. The Antoñita gneiss was intensely deformed by a Variscan subhorizontal shear zone resulting in a plano-linear fabric.

According to Wildberg, Bischoff & Baumann (1989), this area has undergone two metamorphic phases, one of intermediate pressure that occurred at 370–380 Ma, and another of low pressure that occurred between 300 and 280 Ma with peak conditions 4.5 to 6 kbar and 550 °C.

2.b. Villadepera region

In the Villadepera region (Fig. 1d) the Ollo de Sapo Formation crops out in the northern flank of the

Miranda do Douro antiform below a sequence of metapelites and quartzites of Ordovician age, and over the pelites and greywackes of the Cambro-Ediacaran Schist-Greywacke Complex (Iglesias & Ribeiro, 1981). It mostly consists of fine-grained metavolcanic augen-gneisses almost identical to those of Hiendelaencina. In Villadepera, however, the metavolcanic rocks are considerably thinner (300–400 m) and locally appear slightly migmatized. They contain small (< 3–4 cm) polycrystalline augen of K-feldspar and phenocrysts (< 1.5 cm) of quartz or, rarely, plagioclase, within a fine-grained groundmass which, under the microscope, appears composed of alternating thin micaceous and thick quartz-feldspathic bands. The former contain similar proportions of biotite and muscovite with a few small garnet crystals. The latter consist of a granoblastic aggregate of quartz, oligoclase, minor K-feldspar and a few large crystals of biotite. The accessory assemblage consists of abundant apatite, zircon, ilmenite, monazite and a few irregularly distributed Fe–Cu sulphides.

The associated Miranda do Douro metagranite is located in the central part of the Miranda do Douro antiform (Iglesias & Ribeiro, 1981), where it intrudes the metasediments of the Schist-Greywacke Complex that underlie the Ollo de Sapo metavolcanic rocks. The metagranite is a mesocratic augen-gneiss which locally is slightly migmatized. Under the microscope it has a foliated, almost hypidiomorphic granular texture with no cataclasis and few strained grains. The augen are formed either by single crystals of K-feldspar or by syneusis of plagioclase. The groundmass is coarse-grained and consists of quartz, zoned plagioclase crystals, abundant biotite and K-feldspar. As accessories it contains apatite, zircon, monazite, rare xenotime, ilmenite and occasional Fe sulphides.

3. Samples and methods

Finding samples of the Ollo de Sapo metavolcanic rocks and associated metagranites suitable for chemical and isotopic studies is not an easy task, owing to the intense weathering that often affects them. For this work we were able to collect ten fresh representative samples: two from the metavolcanic rocks at Villadepera, two from the neighbouring Miranda do Douro metagranite, two from the metavolcanic rocks at Hiendelaencina, and four from the Antoñita metagranite. All were analysed for major and trace elements, and Sr and Nd isotopes. The results and the UTM coordinates are shown in Table 1.

Zircon was extracted from six samples: one Antoñita metagranite and five metavolcanic rocks. These are the four mentioned previously, plus a hand-picked concentrate of K-feldspar megacrysts collected at Hiendelaencina. Several fresh K-feldspar megacrysts were sliced into 5 mm thick slabs that were then powdered and analysed for Rb and Sr. One of them

yielded enough Rb/Sr variation to make an internal Rb–Sr isochron. Zircon concentrates were obtained using conventional magnetic and heavy-liquid techniques. Once mounted and polished, zircon grains were studied by cathodoluminescence imaging and analysed for U–Th–Pb using the Cameca IMS1270 ion microprobe of the Nordsim facility in Stockholm and the LA-ICPMS of the University of Granada, and for Pb–Pb using the stepwise evaporation method at the TIMS facility of the University of Granada.

Ion microprobe analytical methods broadly follow those described by Whitehouse, Kamber & Moorbath (1999, and references therein). U/Pb and Th/Pb ratios were calibrated using the Geostandards 91500 reference zircon (1065 Ma: Wiedenbeck *et al.* 1995) and include a propagated error component from replicate analyses of 91500 during the analytical session. Errors on $^{207}\text{Pb}/^{206}\text{Pb}$ ratios are either the observed analytical uncertainty or the counting statistics error, whichever is highest. Common Pb corrections assume that most contaminant Pb is present on the surface of the analysed grains, introduced from the sample preparation process, and has a composition that can be approximated, using the Stacey & Kramers (1975) model, for the present day. Table 1ESM (electronic supplementary material, available from the authors, or online at <http://earthref.org/cgi-bin/erda.cgi?n=722>) presents the ‘207-corrected’ ages which are calculated by projecting the uncorrected analysis onto concordia from the assumed common $^{207}\text{Pb}/^{206}\text{Pb}$ composition. In most cases, however, the amount of common Pb, revealed by monitoring ^{204}Pb , is relatively small and has little influence on the calculated age. All ages are calculated using the decay constant recommendations of Steiger & Jäger (1977).

LA-ICPMS analyses of U, Th and Pb isotopes were carried out with a Nd-YAG 213 µm Mercantek laser and a torch-shielded quadrupolar Agilent 7500 ICP-MS spectrometer. The laser beam was set at a diameter of 60 µm, with a repetition rate of 10 Hz and an output energy of 1 mJ per pulse. The ablation time was 60 s and the spot was pre-ablated during 45 s with a laser output energy of 0.3 mJ per pulse. The ablation was done in a He atmosphere. ^{91}Zr was used as an internal standard. The external standard was the NIST-610 glass, which contains 439.9 ppm Zr, 417.7 Hf, 409 ppm Pb, 457.1 ppm U and 450.6 ppm Th (Pearce *et al.* 1997). The following isotope ratios, determined by TIMS at the University of Granada, were also used: $^{204}\text{Pb}/^{206}\text{Pb} = 0.0611$, $^{207}\text{Pb}/^{206}\text{Pb} = 0.9127$, $^{208}\text{Pb}/^{206}\text{Pb} = 2.1898$, $^{206}\text{Pb}/^{238}\text{U} = 0.2501$, $^{208}\text{Pb}/^{232}\text{Th} = 0.5402$. LA-ICPMS U–Pb ages are in good agreement with ion-microprobe data but show more dispersion and tend to be more discordant. The precision (1σ) estimated on ten replicates of the NIST-610 analysed in the same run was better than 2.5 % for element ratios and 1.3 % for isotope ratios. The U–Pb age of each zircon population was estimated by averaging the 207-corrected age of all ion microprobe

Table 1. Location (UTM, zones 29T and 30T), major elements (wt %), trace elements (ppm) and Sr and Nd isotope composition of the studied samples

	CT-6	CT-7	CT-4	CT-10	CT-11	CT-9	GNB-9	GNB-8	GNB-11	GNB-10
X	0495168	0494044	0501166	0507849	0508099	0507849	739245	739245	727601	727601
Y	4554194	5440728	4545047	4552296	4552140	4552296	4604942	4604942	4596900	4596900
SiO ₂	68.06	71.92	76	71.46	72.37	70.25	68.02	70.67	64.62	65.54
TiO ₂	0.50	0.40	0.09	0.32	0.35	0.43	0.68	0.54	0.62	0.67
Al ₂ O ₃	15.76	14.81	12.88	14.45	13.92	15.07	15.34	14.44	16.07	15.79
FeO ^{tot}	3.32	2.59	1.14	2.10	2.36	2.61	3.73	3.45	3.83	4.25
MgO	1.49	1.18	0.12	0.59	0.77	0.89	1.44	0.92	2.23	2.62
MnO	0.04	0.02	0.01	0.03	0.03	0.04	0.05	0.04	0.06	0.06
CaO	1.38	1.03	0.41	0.80	1.08	0.91	1.28	1.08	2.93	2.88
Na ₂ O	2.73	2.85	2.29	2.74	3.14	2.55	2.17	3.35	3.21	3.10
K ₂ O	4.18	3.14	5.83	5.27	4.28	5.39	4.39	4.08	4.03	3.36
P ₂ O ₅	0.18	0.16	0.15	0.17	0.17	0.20	0.25	0.19	0.35	0.35
L.O.I.	1.43	1.18	0.53	1.11	0.85	0.97	1.82	0.82	1.60	0.97
Total	99.07	99.28	99.45	99.04	99.32	99.31	99.17	99.58	99.55	99.59
Li	42	26	15	21	51	35	122	263	102	159
Rb	159	124	213	190	187	213	173	203	227	215
Cs	4.86	2.78	5.72	3.97	9.23	6.77	20.5	50.7	14.5	15.9
Be	2.22	2.71	0.92	2.06	2.36	2.26	2.91	2.30	5.91	5.01
Sr	146	176	28.2	102	89	128	133	131	335	313
Ba	712	863	59	615	459	724	757	753	904	799
Sc	7.1	6.4	1.8	4.2	4.1	6.0	9.4	8.8	9.9	10
V	61	49	9	31	32	43	63	53	66	66
Cr	57	48	21	32	35	46	47	42	83	107
Co	128	105	35	26	34	38	60	73	66	73
Ni	16	11	< 1	7	8	10	22	16	34	41
Cu	7	11	3	11	2	6	15	17	15	18
Zn	42	40	27	44	41	50	75	78	77	84
Ga	20	18.1	15.1	17.2	17.5	20.9	22	18.4	24.3	23.7
Y	28.2	19	13.5	25	16.1	18.1	21.6	17.8	24	22.1
Nb	10.5	8.4	6.7	7.9	8	10.6	11.3	10	11.8	12.9
Ta	1.1	0.9	0.9	0.9	0.6	1.1	1.2	1	1.2	1.4
Zr	162	136	68	136	151	194	217	182	232	247
Hf	5.2	3.6	2.1	4.3	4.9	6.0	5.0	5.0	5.7	5.9
Mo	0.56	0.4	0.11	0.32	0.36	0.37	0.63	0.22	0.78	0.73
Sn	4.07	2.91	7.49	6.2	7.14	8.43	4.63	3.23	7.45	6.89
Tl	0.92	0.62	1.14	1.09	1.03	1.16	0.97	1.16	1.47	1.54
Pb	22.1	21.7	25.1	23.9	21.9	23.3	18.8	48.3	40.8	32.0
U	3.42	3.22	2.22	3.44	3.21	3.96	3.12	2.73	8.29	8.07
Th	11.1	10.3	6.86	9.82	10.1	12.1	12.5	12.1	33.5	35.8
La	26.8	31.3	7.96	22.7	22.2	27.5	35.1	30.4	83.2	82.9
Ce	58.6	50.5	18.5	48.3	46.7	59.2	74.3	63.6	160	161
Pr	7.04	7.84	2.26	5.86	5.79	7.04	8.99	7.73	18.4	18.4
Nd	28.6	27.9	8.52	20.9	21.9	27.8	34.8	30.3	63.7	66.1
Sm	6.17	5.86	2.29	4.73	4.92	6.13	7.53	6.46	10.3	10.3
Eu	1.22	1.46	0.18	0.94	0.89	0.99	1.16	1.03	1.52	1.5
Gd	5.07	5.26	2.32	4.81	4.34	4.97	6.02	4.93	6.80	6.44
Tb	0.86	0.75	0.41	0.75	0.67	0.75	0.83	0.73	0.89	0.87
Dy	5.05	3.83	2.49	4.54	3.68	4.03	4.24	3.70	4.34	4.09
Ho	1.01	0.71	0.51	0.91	0.64	0.72	0.76	0.64	0.78	0.75
Er	2.55	1.76	1.39	2.3	1.50	1.61	1.76	1.49	2.09	1.88
Tm	0.37	0.23	0.21	0.34	0.19	0.22	0.25	0.21	0.28	0.27
Yb	2.14	1.44	1.34	2.01	1.18	1.3	1.52	1.15	1.74	1.47
Lu	0.33	0.22	0.21	0.3	0.17	0.2	0.21	0.17	0.24	0.22
Age (Ma)	495	495	475	475	475	475	483	483	483	483
⁸⁷ Rb/ ⁸⁶ Sr	3.164	2.047	22.089	5.45	6.147	4.839	3.782	4.508	1.963	1.993
⁸⁷ Sr/ ⁸⁶ Sr	0.73271	0.72429	0.86098	0.74794	0.75328	0.74329	0.73302	0.73857	0.71793	0.71892
⁸⁷ Sr/ ⁸⁶ Sr _i	0.71039	0.70985	0.71150	0.71105	0.71168	0.71054	0.70700	0.70755	0.70442	0.70521
εSr _i	92	84	107	101	110	94	43	51	7	18
¹⁴⁷ Sm/ ¹⁴³ Nd	0.1305	0.1268	0.1631	0.1368	0.1355	0.1335	0.1308	0.1291	0.0979	0.0946
¹⁴³ Nd/ ¹⁴³ Nd	0.512219	0.512214	0.512407	0.512244	0.512307	0.512268	0.512174	0.512199	0.512232	0.512136
¹⁴³ Nd/ ¹⁴³ Nd _i	0.511804	0.511811	0.511889	0.511809	0.511876	0.511844	0.511758	0.511789	0.511921	0.511835
εNd _i	-4.07	-3.94	-2.43	-3.98	-2.67	-3.3	-4.97	-4.38	-1.8	-3.47
Nd _(CHUR) MA (Ma)	964	925	1047	1002	825	892	1073	989	627	750
Nd _(DM) MA (Ma)	1708	1645	2239	1798	1648	1679	1797	1716	1213	1302

CT-6 and CT-7 are metavolcanic rocks of Hiendelaencina; CT-4, CT-10, CT-11 and CT-9 are from the Antoñita gneiss of Hiendelaencina; GNB-9 and GNB-8 are metavolcanic rocks of Villadepera; GNB-11 and GNB-10 are from the Miranda do Douro gneiss. See location in Figure 1.

and LA-ICPMS determinations, and the errors are reported at 95 % confidence level.

Single-zircon evaporation analyses (Kober, 1987) were performed using a SEM-RPQ multicollector Finnigan Mat 262 thermal ionization mass spectrometer with a double filament ion-source arrangement, operated with the RunIt262 Spectromat software.

The zircon grain was mounted on a canoe-shaped Re evaporation filament and heated until the lead beam was intense enough (~ 200–400 ²⁰⁶Pb ions per second). The lead was collected on the ionization filament for 20–30 minutes, and was then analysed in five blocks with seven scans per block. Once the analysis was finished, a new step was started by heating the zircon on

the evaporation filament at a higher temperature than in the previous step (usually increasing the current by 50–100 mA) and analysing, as before, the Pb deposited anew on the ionization filament. The procedure was repeated until all the lead was exhausted from the zircon. The number of steps depended on the size and lead content of each zircon. Data acquisition was performed in dynamic mode (peak hopping), using a secondary electron multiplier as detector with the 206–204–206–207–208 mass sequence. The mass-ratio 204/206 was monitored to detect and, if necessary, correct for common lead. Factors for common lead correction were calculated by iteration from the $^{204}\text{Pb}/^{206}\text{Pb}$ and $^{204}\text{Pb}/^{207}\text{Pb}$ ratios provided by the Stacey & Kramers (1975) model at the calculated age, until convergence to a constant value. For 207/206, mass fractionation in the detector was corrected by multiplying the measure value by $\sqrt{(207/206)}$. Standard errors for each step were calculated according to the formula: $\text{SE} = 2\sigma/\sqrt{n}$. The age is calculated for the 95 % confidence level of the mean of all steps.

Samples for Sr and Nd isotope analysis (0.1000 g) were digested with $\text{HNO}_3 + \text{HF}$ in a Teflon[®]-lined vessel. The elements were separated with ion-exchange resins, and the Sr and Nd isotope ratios were determined by thermal ionization mass spectrometry (TIMS) with a Finnigan Mat 262 at the University of Granada. All reagents were ultra-clean. Normalization values were $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ and $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$. Blanks were 0.6 and 0.09 ng for Sr and Nd. The external precision (2σ), estimated by analysing ten replicates of the standard WS-E (Govindaraju *et al.* 1994), was better than $\pm 0.003\%$ for $^{87}\text{Sr}/^{86}\text{Sr}$ and $\pm 0.0015\%$ for $^{143}\text{Nd}/^{144}\text{Nd}$. $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ were directly determined by ICP-MS at Granada following the method developed by Montero & Bea (1998), with a precision better than $\pm 1.2\%$ and $\pm 0.9\%$ (2σ), respectively.

Major elements and Zr were determined at the University of Granada by X-ray fluorescence after fusion with lithium tetraborate. Typical precision was better than $\pm 1.5\%$ for an analyte concentration of 10 wt %, and $\pm 5\%$ for 100 ppm Zr. Trace elements, except Zr, were determined by ICP-mass spectrometry (ICP-MS) after $\text{HNO}_3 + \text{HF}$ digestion in a Teflon[®]-lined vessel, evaporation to dryness, and subsequent dissolution in 100 ml of 4 vol. % HNO_3 . Precision was better than $\pm 5\%$ for analyte concentrations of 10 ppm.

4. Dating results

4.a. The age of the Ollo de Sapo metavolcanic rocks at Villadepera

U–Pb ion-microprobe and LA-ICPMS data from the Ollo de Sapo metavolcanic rocks at Villadepera (table 1ESM, available from the authors, or online at <http://earthref.org/cgi-bin/erda.cgi?n=722>) are, in

general, highly concordant. The two techniques yield similar results (Fig. 2). The most abundant concordant population is Ordovician and yielded 483 ± 4 Ma. The second most abundant concordant population is Ediacaran and yielded 614 ± 12 Ma. A third, nearly concordant, Orosirian population yielded an upper intercept at 1968 ± 35 Ma. Lastly, there are a few Meso-Archaean grains defining a discordia line with an upper intercept at 2997 ± 37 Ma.

Remarkably, the metavolcanic rocks have almost exactly the same age distribution as the neighbouring Miranda do Douro metagranite, which has a crystallization age of 483 ± 3 Ma, an inherited concordant zircon population at 605 ± 13 Ma, and a few discordant Meso-Archaean grains (Bea *et al.* 2006). The only difference is that the Orosirian population of the metavolcanic rocks has not yet been found in the metagranite.

4.b. The age of the Ollo de Sapo metavolcanic rocks at Hiendelaencina

4.b.1. The lower metavolcanic rocks

The U–Pb data of the two samples from the lower metavolcanic gneisses, one augen and the other fine-grained, are indistinguishable, so they will be discussed together. The most abundant population is Late Cambrian and yielded a concordant age of 495 ± 5 Ma (Fig. 3), confirmed by a Pb–Pb plateau at 494 ± 4 Ma (see data in table 3ESM, available from the authors, or online at <http://earthref.org/cgi-bin/erda.cgi?n=724>). The second and almost equally abundant population is Ediacaran and yielded 603 ± 8 Ma. Additionally, there are three other inherited zircon populations. Two are concordant or nearly concordant, one Cryogenian with 650–700 Ma and another Tonian with 850–900 Ma. The last population comprises some discordant Orosirian grains with an upper intercept at 2016 ± 49 Ma.

4.b.2. The upper metavolcanic rocks

Slabs sliced from one fresh K-feldspar megacryst had sufficient Rb/Sr variation to obtain a Rb–Sr isochron. This yielded 490 ± 65 Ma with $^{87}\text{Sr}/^{86}\text{Sr}_{\text{init}} = 0.7123 \pm 0.0017$ and $\text{MSWD} = 0.92$ (Fig. 4; data in table 2ESM, available from the authors, or online at <http://earthref.org/cgi-bin/erda.cgi?n=723>). Although the confidence interval is large, because the $^{87}\text{Rb}/^{86}\text{Sr}$ coordinate only ranges from 1.67 to 2.21, the excellent fit permits us, in principle, to consider the isochron as a good estimate of the crystallization age of the K-feldspar megacryst.

U–Pb data from zircon extracted from the megacrysts tend to be slightly more discordant than in the lower metavolcanic rocks (Fig. 5). The most abundant concordant population yielded 485 ± 6 Ma, in good agreement with the Rb–Sr age. The second most

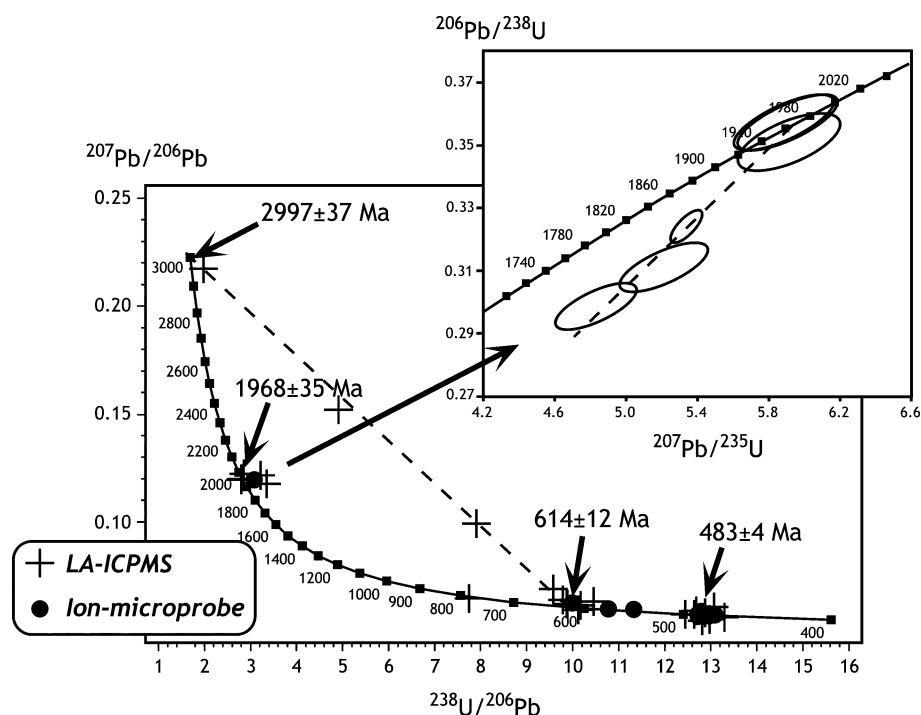


Figure 2. Tera-Wasserburg concordia plot of zircons from the Ollo de Sapo metavolcanic rocks at Villadepera. The most abundant concordant population yielded an age of 483 ± 4 Ma, identical to the crystallization age of the neighbouring Miranda de Douro metagranite (Bea *et al.* 2006). The Ediacaran population at 614 ± 12 Ma, interpreted as the magma source's age, and the Meso-Archaeon discordant population are also found in the metagranite, but the Orosirian nearly concordant population at *c.* 2.0 Ga, plotted in a conventional concordia in the inset, is exclusive to the Ollo de Sapo metavolcanic rocks. See text for more details.

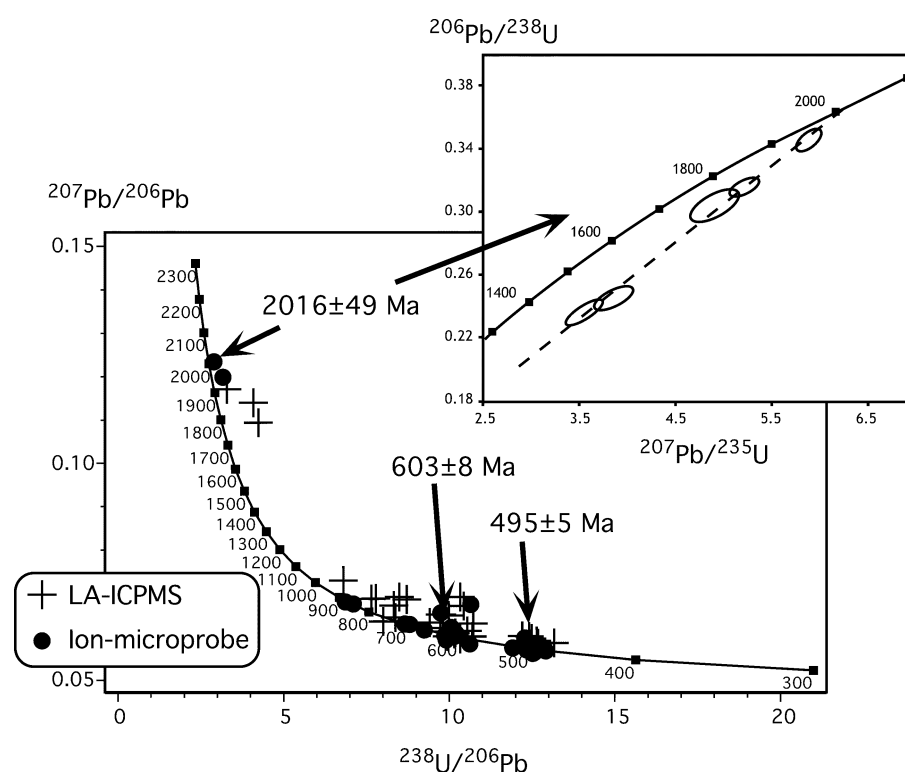


Figure 3. Tera-Wasserburg concordia plot for the lower metavolcanic rocks of Ollo de Sapo at Hiendelaencina. The most abundant concordant population yielded an age of 495 ± 5 Ma. The Ediacaran population yielded 603 ± 8 Ma. Note two small concordant populations at 650–700 Ma (Cryogenian) and 850–900 Ma (Tonian), as well as the discordant Orosirian population, plotted in a conventional concordia in the inset, also found in Villadepera.

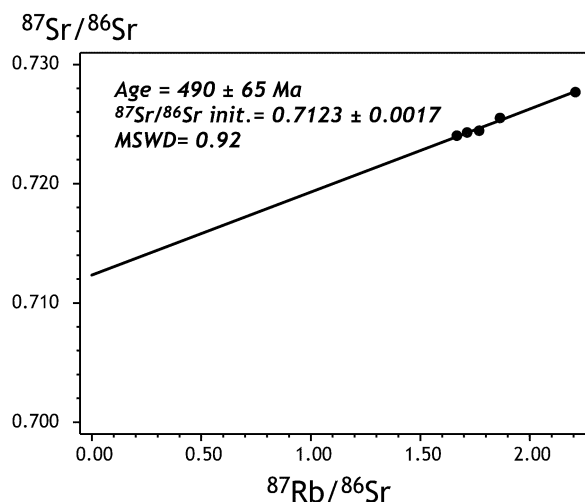


Figure 4. Rb–Sr internal isochron of a sliced K-feldspar megacryst from the upper metavolcanic rocks of the Ollo de Sapo at Hiendelaencina. The error is very high owing to the low dispersion of the $^{87}\text{Rb}/^{86}\text{Sr}$ coordinate, but the MSWD is so low that it fits York's model 1 (York, 1969). The age is almost identical to the zircon age and indicates the magmatic origin of the K-feldspar megacrysts, something that has been occasionally questioned.

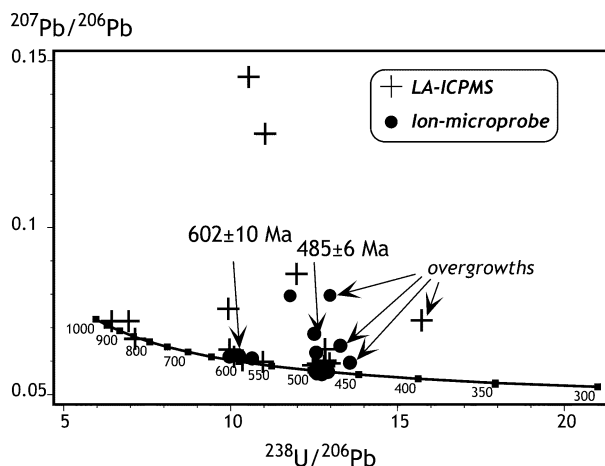


Figure 5. Tera–Wasserburg concordia plot for zircons separated from the K-feldspar megacrysts of upper metavolcanic rocks of the Ollo de Sapo at Hiendelaencina (see Rb–Sr isochron in Fig. 4). The most abundant concordant population yielded an age of 485 ± 6 Ma, and the Ediacaran population yielded 602 ± 10 Ma, very similar to the metavolcanic rocks at Villadepera. Here, however, the oldest concordant zircons are Tonian. Note how the younger rims are highly discordant and have 207-corrected ages younger than the Rb–Sr age.

important population, also concordant, yielded 602 ± 10 Ma, and a few concordant or nearly concordant grains yielded ages between 850 and 900 Ma. There are also texturally discontinuous rims with highly discordant and unreasonably young 207-corrected U–Pb ages (see Section 6.a).

Ten zircon grains were also studied with the Pb–Pb stepwise evaporation method (Kober, 1987), capable of yielding accurate crystallization ages for zircons with complex discordant patterns (Karabinos, 1997). Nine-

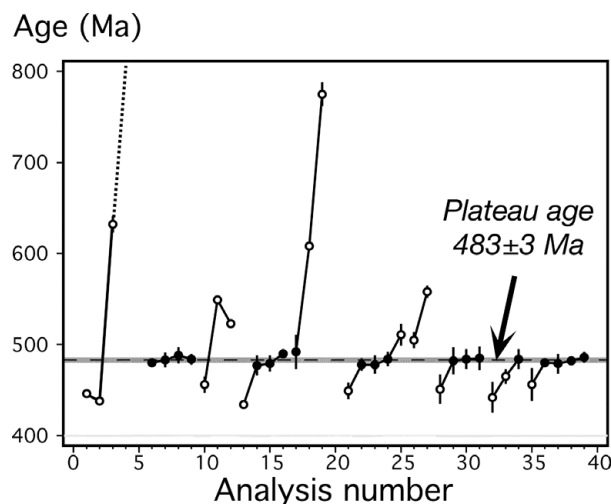


Figure 6. $^{207}\text{Pb}/^{206}\text{Pb}$ age plot of ten zircon grains separated from the K-feldspar megacrysts from the upper metavolcanic rocks of the Hiendelaencina Ollo de Sapo. The plateau, calculated from the analyses shown as solid dots, yields 483 ± 3 Ma. This confirms the Rb–Sr and U–Pb ages and confirms that the upper metavolcanic rocks are about 12 million years younger than the lower metavolcanic rocks. As in the Tera–Wasserburg concordia (Fig. 5), most grains show younger rims which cannot be correlated to any major geological event.

teen out of 39 measurements yielded a plateau age of 483 ± 3 Ma (Fig. 6, table 3ESM, available from the authors, or online at <http://earthref.org/cgi-bin/erda.cgi?n=724>). Since this value agrees (within error) with the Rb–Sr and U–Pb estimates, we regard it as the best estimate of the crystallization age. Seven grains yielded younger overgrowths with a minimum age of about 445 Ma, younger than the Rb–Sr age. Four grains also had older cores, in one case with a minimum age close to 2000 Ma.

4.c. The Antoñita metagranite

Despite the intense Variscan deformation that affected them, the four whole-rock samples of the Antoñita metagranite fit well into a Rb–Sr isochron (Fig. 7, data in table 2ESM, available from the authors, or online at <http://earthref.org/cgi-bin/erda.cgi?n=723>) that yielded 477 ± 30 Ma, with $^{87}\text{Sr}/^{86}\text{Sr}_{\text{init.}} = 0.7107 \pm 0.0027$ and a MSWD = 1.03. The ion-microprobe and the LA-ICPMS U–Pb zircon data, on the other hand, are too discordant for any reliable age estimation. This, however, was possible using the stepwise Pb–Pb evaporation technique. Ten grains yielded a plateau at 474 ± 4 Ma (Fig. 8, data in table 3ESM, available from the authors, or online at <http://earthref.org/cgi-bin/erda.cgi?n=724>); since this value matches the result of the Rb–Sr isochron, we consider it to represent the best estimate of the crystallization age. Six grains also have younger overgrowths with ages down to 400 Ma, a value that roughly agrees with the c. 380 Ma U–Pb lower intercept previously found by Wildberg, Bischoff & Baumann (1989). Most zircon

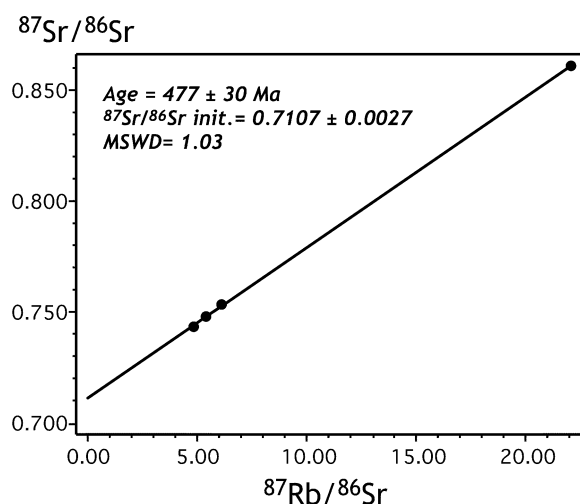
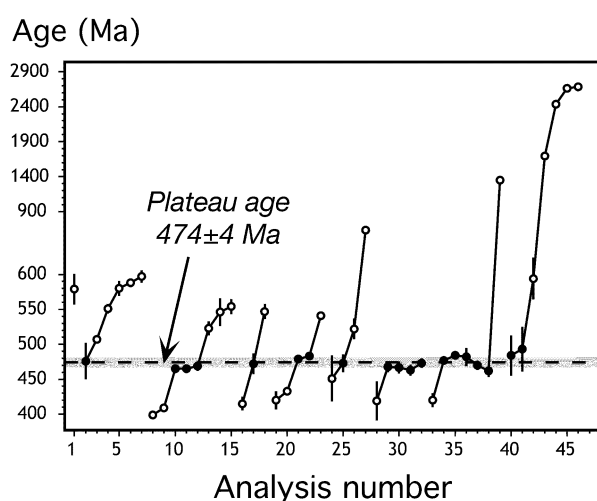


Figure 7. Rb–Sr isochron of the Antoñita metagranite.


 Figure 8. $^{207}\text{Pb}/^{206}\text{Pb}$ age plot of ten grains from the Antoñita metagranite. The plateau, calculated from the analyses shown as solid dots, yields 474 ± 4 Ma. This confirms the Rb–Sr ages and is, therefore, considered as the best estimate of the crystallization age. Note the complex pattern with younger rims shown by most grains. One core shows a plateau at c. 2.68 Ga.

grains also have older cores, the precise age of which cannot be estimated because of the lack of plateaus, except in one case that yielded a Neo-Archaeon age of 2.68 Ga.

5. Chemical and isotopic composition

To elaborate a detailed geochemical study of the Ollo de Sapo metavolcanic rocks and related metagranites is beyond the scope of this paper, basically because the number of samples suitable for chemical and isotopic studies that we were able to collect was severely limited by the scarcity of fresh-rock outcrops. What follows, therefore, is just an outline intended to ascertain whether volcanic and intrusive rocks from a specific locality could be cogenetic, and to recognize regional differences between Villadepera and Hiendelaencina.

The overall chemistry of all these rocks is similar to felsic peraluminous igneous rocks with $\text{K}_2\text{O} > \text{Na}_2\text{O}$ and mol. $\text{Fe}/\text{Fe}+\text{Mg} \approx 0.49\text{--}0.60$ (Table 1). The metagranites are granodioritic to granitic with the aluminium saturation index (ASI) $\approx 1.07\text{--}1.29$, and the metavolcanic rocks are rhyodacitic to dacitic with ASI $\approx 1.21\text{--}1.49$. Metavolcanic rocks and metagranites from the same region have similar Nb/Zr, Ni/MgO, Sc/V (Fig. 9b) and $^{87}\text{Sr}/^{86}\text{Sr}_i$ (Fig. 10), thus indicating they derived from similar sources. Notably, the largest differences in chemical composition occur in Villadepera, where the metagranite is richer in LREE, Zr, Th, U, Cr, Ni, Sr and Be, but poorer in Cs and Li than the metavolcanic rocks, despite both being coeval (Fig. 9a).

The two rock types, on the other hand, show marked regional variations. In Villadepera both metavolcanic rocks and metagranites are richer in LREE (Fig. 9a), HFSE and LIL elements than in Hiendelaencina, and define distinct positive trends between Ni and MgO, and Sc and V (Fig. 9b). Especially important are the differences in Sr isotopes (Fig. 10): whereas at Villadepera $^{87}\text{Sr}/^{86}\text{Sr}_i$ is always lower than 0.7075, in Hiendelaencina this ratio is always higher than 0.7098. Nd isotopes do not show meaningful regional variations, but there are differences between the metavolcanic rocks and the metagranites, especially in Villadepera (Fig. 10). The $\text{Nd}_{(\text{CHUR})}$ model age for the Miranda do Douro metagranite, which has the most primitive isotope and chemical composition, is 0.69 Ga, slightly older than the zircon age of the source, whereas for the Villadepera metavolcanic rocks the $\text{Nd}_{(\text{CHUR})}$ model age is 1.0 Ga. In Hiendelaencina, both rock types yielded the same value of 0.94 Ga.

6. Discussion

6.a. The significance of younger rims

Texturally discontinuous rims with U–Pb and Pb–Pb ages younger than host Rb–Sr ages occur in the Hiendelaencina upper metavolcanic rocks and the Antoñita metagranite. In the former, Bea, Montero & Ortega (2006) have suggested that the rims may represent solid-state overgrowths related to the exsolution of Zr during the subsolidus recrystallization of Zr-bearing feldspars (see fig. 18 of these authors). In the Antoñita metagranite, the younger rims could be related either to this phenomenon or to processes of zircon dissolution–precipitation induced by fluids during intense Variscan shearing deformation. In any case, the age of the younger rims was imprecisely determined and is therefore of no use for petrogenetic interpretations.

6.b. The age of crystallization

The Ollo de Sapo metavolcanic rocks in Villadepera, where they are 300–400 m thick, yielded a

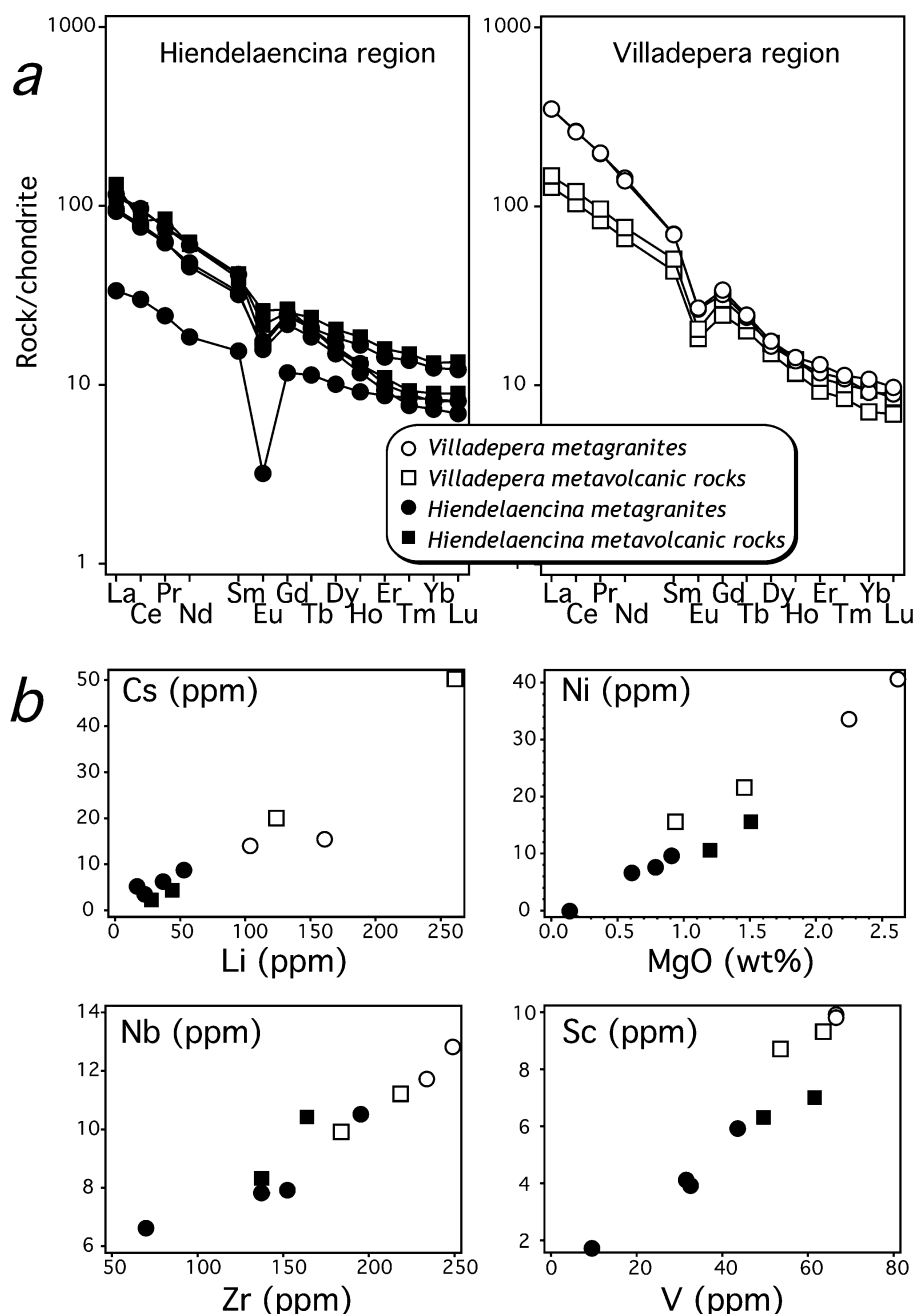


Figure 9. (a) Chondrite-normalized REE patterns of the Ollo de Sapo metavolcanic rocks and related metagranites. Note the enrichment in LREE in the Villadepera metavolcanic rocks and the Miranda do Douro metagranite. (b) Selected binary element plots. Note how the metavolcanic rocks and the metagranites of each zone plot on the same line, thus suggesting they are genetically related.

crystallization age of 483 ± 4 Ma. In Hiendelaencina, where they are more than 2000 m thick, the upper metavolcanic rocks yielded a crystallization age of 483 ± 3 Ma and the lower metavolcanic rocks yielded a crystallization age of 495 ± 5 Ma. These data indicate that the volcanic event that produced the Ollo de Sapo finished at about the same time in these two regions, over a distance of more than 300 km (Fig. 1), and that in Hiendelaencina the volcanism lasted for about 12 million years. In the Villadepera area, the Miranda do Douro metagranite (483 ± 3 Ma; Bea *et al.* 2006) is coeval with the neighbouring metavolcanic

rocks, but in the Hiendelaencina area the Antoñita metagranite (474 ± 4 Ma) is about 10 Ma younger than the youngest Ollo de Sapo metavolcanic rocks. The intrusion of granites, therefore, started at the end of the volcanism and continued for at least another 10 million years, probably longer if we also consider the Ordovician orthogneisses of the Schist-Greywacke Domain (Viallette *et al.* 1987; Valverde & Dunning, 2000; unpublished data of the present authors) (Fig. 1). This same pattern is found in the southern boundary of the Central Iberian Zone where the volcano-sedimentary Urra Formation, equivalent

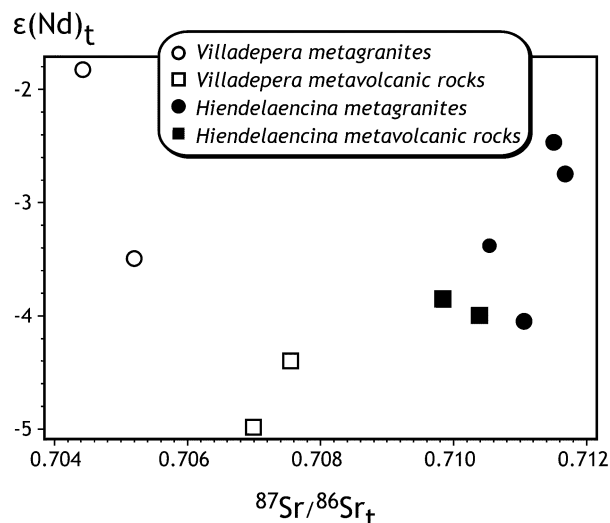


Figure 10. $\epsilon(\text{Nd})_t$ v. $^{87}\text{Sr}/^{86}\text{Sr}_t$ for the Ollo de Sapo metavolcanic rocks and related metagranites. Note the marked difference in $^{87}\text{Sr}/^{86}\text{Sr}_t$ between the two regions. This coincides with the presence of older crustal components in the protolith of the Hiendelaencina rocks, as revealed by the ages of inherited zircons, and might indicate the across-arc magmatic polarity of the Pan-African magmas; see text for discussion.

to the Ollo de Sapo but much smaller, has yielded crystallization ages of 495 ± 7 Ma and 488 ± 5 Ma (Solá *et al.* 2006), whereas the spatially associated Carrascal granites and diorites yielded *c.* 484–470 Ma (Solá *et al.* 2005). These data reveal important Cambro-Ordovician magmatic activity in the two boundaries of the Central Iberian Zone, especially in the north, which caused felsic peraluminous volcanism from *c.* 495 Ma to 483 Ma and high-level granite intrusions from 483 Ma to 465 Ma.

Remarkably, this magmatism occurred with no evident connection to any major tectonic or metamorphic event. The volcanism coincides with the Toledanic unconformity at the base of the Ordovician (Gutiérrez Marco *et al.* 2002), but the intrusion of high-level granites was contemporaneous with the sedimentation of the Armorican quartzite in an almost stable platform.

6.c. The history of the magmatic sources

One of the most conspicuous aspects of the Ollo de Sapo metavolcanic rocks and related metagranites is the abundance of inherited zircon, much higher than is usual even in high-inheritance ‘cold’ granites (Miller, McDowell & Mapes, 2003). In both metavolcanic rocks and metagranites, no less than 70–80 % of zircon grains consist of a Precambrian core, most often Early Ediacaran, overgrown by a Cambro-Ordovician rim. Such a textural relationship indicates that the older zircon components of the Ollo de Sapo were not due to a hypothetical sedimentary contribution to the volcanic deposits, but a primary feature of the Cambro-Ordovician magmas also found in other sectors of the West-European Variscan belt (e.g. Deloule *et al.* 2002;

Laumonier *et al.* 2004; Helbing & Tiepolo, 2005). The fact that zircons separated from rapakivi K-feldspar megacrysts (of indisputable magmatic origin) have the same inherited populations as zircons separated from the whole rock adds corroborative evidence to this interpretation. Although discussing the reasons for this unusually high zircon inheritance is far beyond the scope of this article, it is worth mentioning here that they point to extremely fast processes of melting, melt segregation and emplacement, thus not giving enough time for zircon dissolution. This aspect is currently being studied.

Notably, about 80–95 % of all inherited zircons yielded virtually the same age in all the studied rocks, from 602 ± 10 Ma to 614 ± 12 Ma. This reveals that the source of the Cambro-Ordovician magmas ultimately derived from Early Ediacaran igneous rocks, of which unfortunately no remnant has been preserved. In the Villadepera region, the Miranda do Douro metagranite rarely contains older-than-Ediacaran zircons except for a few Meso-Archaean grains. The spatially related metavolcanic rocks, on the other hand, contain two older populations. One, highly discordant, is also Meso-Archaean, indistinguishable from the equivalent population of the metagranite. The other, nearly concordant, is Orosirian with an upper intercept at 1.97 Ga. The latter seems exclusive to the metavolcanic rocks, thus suggesting that their source contained a component not found in the metagranite, despite the same age and spatial proximity. The few chemical and isotopic data available (Figs 9, 10) are consistent with this interpretation. The two sources might well have consisted of variable proportions of two components: juvenile island-arc igneous rocks (or immature sediments derived from them) with 600–615 Ma zircon, and old crustal rocks with 1.97 Ga zircon. Whereas the metagranite source was almost exclusively composed of the former, as suggested by its lower $^{87}\text{Sr}/^{86}\text{Sr}_{\text{init}}$ and the similarity between its U–Pb zircon age (0.61 Ga) and the $\text{Nd}_{\text{(CHUR)}}$ model age (0.69 Ga), the metavolcanic rocks source also had an appreciable fraction of the latter. The Archaean component detected by zircon ages probably just consisted of exotic detrital zircon grains, since it apparently caused no impact on the Sr and Nd isotope composition of the products.

The pre-Ediacaran zircons in the Hiendelaencina metavolcanic rocks show marked differences from Villadepera. There is a similar Orosirian population but it lacks the Meso-Archaean population and has two new concordant populations, one Cryogenian (650–700 Ma) and the other Tonian (850–900 Ma). The age of the pre-Ordovician zircon populations of the Antoñita metagranite, on the other hand, was not precisely established, so that we cannot use this criterion for detecting whether it derived from the same magmatic source as the neighbouring metavolcanic rocks. Nonetheless, the chemical and isotopic

composition of the two rock formations are notably similar (Fig. 9), thus indicating that they derived from similar sources and these probably had more sedimentary components (higher $^{87}\text{Sr}/^{86}\text{Sr}$) than in Villadepera.

As mentioned previously, no remnant of Early Ediacaran or older magmatic rocks has been identified in Iberia. Fortunately, we can get an insight into the Pan-African magmatism of Iberia by looking at the magmatic sequence of the Moroccan Anti-Atlas (Gasquet *et al.* 2005; Tahiri *et al.* 2005), because this region was close to Iberia during the Pan-African orogeny (e.g. Ennih & Liégeois, 2001, 2003) and was barely affected by the Variscan orogeny.

According to Gasquet *et al.* (2005), in North Africa the oldest magmatic events occurred at 2050 Ma and 1760 Ma and correspond to pre-Pan-African Eburnian basement. The Pan-African began with tholeiitic magmatism related to ocean opening at about 870–840 Ma (see also Reischmann *et al.* 1992) and probably lasted until about 740 Ma. This was followed by subduction-related calc-alkaline magmatism between 740 Ma and 690 Ma. The production of calc-alkaline magmas, however, reached a maximum during the arc–continent collision and ocean closure. This process was episodic, with the first and less voluminous episode between 690 Ma and 660 Ma, and the second, and more voluminous, episode between 620 Ma and 600 Ma. From this point onwards, the tectonic regime changed from orogenic to anorogenic, which produced abundant high-K calc-alkaline magmatism between 594 Ma and 545 Ma and then some minor alkaline volcanism at 530 Ma.

Placing the ages of the zircons inherited by the Iberian Cambro-Ordovician magmas into this framework gives the following results. The Ediacaran population, which is ubiquitous and by far the most abundant, almost perfectly matches the 600–620 Ma episode that occurred at the end of the arc–continent collision. The Hiendelaencina Cryogenian population matches the 660–690 Ma episode that occurred at the beginning of the arc–continent collision and thus it represents older island-arc components. The Tonian population from the same area, on the other hand, matches the opening of the Pan-African palaeo-ocean, so that it might represent early-formed oceanic crust obducted over the arc during the final stages of the ocean closure. As indicated before, the widely distributed Orosirian population may represent the contribution of the pre-Pan-African basement to the magma source.

The uneven distribution of the various components in the Cambro-Ordovician rocks from the two Iberian regions studied suggests, at least tentatively, an orientation of the across-arc polarity of the Pan-African sources. In Villadepera, to the west, the presence of Meso-Archaeozoic zircons indicates proximity to the West African Craton, and the presence of just one Pan-African zircon population suggests that the fertile Pan-African arc component might have derived exclusively

from the last magmas produced during the orogenic stage. This might indicate that this region was close to the passive margin. In Hiendelaencina, to the east, both the lack of Meso-Archaeozoic zircons and the presence of Cryogenian and Tonian zircons suggest that the Pan-African source was far from the West African Craton and contained older island-arc components. This suggests that this region was located well within the colliding arc. It seems, therefore, that the across-arc polarity of the Pan-African arc of Iberia had a marked east–west (with respect to the current coordinates) factor during Cambro-Ordovician times, and that the passive margin was situated to the west.

7. Conclusions

The most important conclusions of this work can be summarized as follows:

The volcanism that produced the Ollo de Sapo metavolcanic rocks spanned from about 495 Ma to 483 Ma. It was followed by the intrusion of high-level granites, now preserved as gneissic metagranites, from 483 Ma to 474 Ma (or 465 Ma if the Ordovician metagranites of the Schist-Greywacke domain are also considered). The magmatic event that generated these rocks was apparently unrelated to any major metamorphic or tectonic event. It might represent the initial stages of the rifting process that caused peralkaline magmas of the same age in the Ossa Morena Zone to the south.

The Cambro-Ordovician igneous rocks of Central Iberia contain an unusually high fraction of inherited zircon; no less than 70 % to 80 %, in some cases nearer 100 %, of zircon grains are either totally Precambrian or contain a Precambrian core overgrown by a Cambro-Ordovician rim. In all the studied rocks, about 80 % of inherited zircons are Early Ediacaran, with ages from 602 ± 10 Ma to 614 ± 12 Ma. These zircons come from the dominant component in the source of the Cambro-Ordovician magmas, which probably consisted of calc-alkaline intermediate to felsic igneous rocks (or young, immature sediments derived from them) generated at the end of the Pan-African arc–continent collision and ocean closure.

In the Villadepera region, both the Miranda do Douro metagranite and the Ollo de Sapo metavolcanic rocks contain Meso-Archaeozoic zircons with ages of 3.0–3.2 Ga. These zircons were probably added as exotic detrital grains ultimately derived from the West-African Craton. Their presence does not imply the participation of fragments of an Archaeozoic crust in the source, as indicated by the low $^{87}\text{Sr}/^{86}\text{Sr}_{\text{init}}$ of the metagranite and the good correspondence between the $\text{Nd}_{\text{(CHUR)}}$ model age of this and the U–Pb age of the Ediacaran zircons. The Ollo de Sapo metavolcanic rocks also contain an Orosirian zircon population derived from an old crustal component. In the Hiendelaencina region, by contrast, the Ollo de Sapo metavolcanic

rocks and the Antoñita metagranite have similar Sr and Nd isotope composition, in both cases with $^{87}\text{Sr}/^{86}\text{Sr}_{\text{init}}$ significantly higher than in Villadepera. The presence of two new inherited zircon populations, one Cryogenian (650–700 Ma) and the other Tonian (850–900 Ma), indicates that, besides the Ediacaran component, the Hiendelaencina sources also had older island-arc components.

The different proportion of source components in the two studied regions, and the marked variation of the $^{87}\text{Sr}/^{86}\text{Sr}_{\text{init}}$, suggest that the across-arc polarity of the Pan-African arc of Iberia trended east–west (with respect to the current coordinates) during Cambro-Ordovician times and that the passive margin was situated to the west.

Acknowledgements. We are indebted to B. Murphy and one anonymous referee whose suggestions greatly contributed to improving the original manuscript, and to J. H. Scarrow for her valuable comments and assistance with the English. We are also grateful to J. Holland for editorial assistance. This work was financially supported by the Spanish grant CGL2005–05863/BTE, the SYNTHESIS grant SE-TAF-528 and the NATO grant EST-CLG-978997. This is Nordsim publication no. 175.

References

- BARBEY, P., CHEILLETZ, A. & LAUMONIER, B. 2001. The Canigou orthogneisses (Eastern Pyrenees, France, Spain): an early Ordovician rapakivi granite laccolith and its contact aureole. *Comptes rendus Geoscience* **332**, 129–36.
- BEA, F., MONTERO, P. & ORTEGA, M. 2006. A LA-ICPMS evaluation of Zr reservoirs in common crustal rocks: implications for Zr and Hf geochemistry, and zircon-forming processes. *Canadian Mineralogist* **44**, 693–714.
- BEA, F., MONTERO, P., TALAVERA, C. & ZINGER, T. 2006. A revised Ordovician age for the oldest magmatism of Central Iberia: U–Pb ion microprobe and LA-ICPMS dating of the Miranda do Douro orthogneiss. *Geologica Acta* **4**, 395–401.
- BEA, F., MONTERO, P. & ZINGER, T. 2003. The Nature and Origin of the Granite Source Layer of Central Iberia: Evidence from Trace Element, Sr and Nd Isotopes, and Zircon Age Patterns. *Journal of Geology* **111**, 579–95.
- BULTYNCK, P. & SOERS, E. 1971. Le Silurien supérieur et le Dévonien inférieur de la Sierra de Guadarrama (Espagne Central). *Bulletin Institute Royal Sciences Naturelles de Belgique* **47**, 1–22.
- DELAPERRIÈRE, E. & RESPAUT, J. P. 1995. Un âge ordovicien de l'orthogneiss de la Preste par la méthode d'évaporation directe du plomb sur monozircon remet en question l'existence d'un socle précambrien dans le Massif du Canigou (Pyénées Orientales, France). *Comptes rendus Geoscience* **320**, 1179–85.
- DELOULE, E., ALEXANDROV, P., CHEILLETZ, A., LAUMONIER, B. & BARBEY, P. 2002. In-situ U–Pb zircon ages for Early Ordovician magmatism in the eastern Pyrenees, France: the Canigou orthogneisses. *International Journal of Earth Sciences* **91**, 398–405.
- DÍAZ MONTES, A., NAVIDAD, M., GONZÁLEZ-LODEIRO, F. & MARTÍNEZ CATALÁN, J. R. 2004. El Ollo de Sapo. In *Geología de España* (ed. J. A. Vera), pp. 69–72. Madrid: SGE-IGME.
- ENNIH, N. & LIÉGEOIS, J. P. 2001. The Moroccan Anti-Atlas: the West Africa craton passive margin with limited Pan-African activity. Implications for the northern limit of the craton. *Precambrian Research* **112**, 289–302.
- ENNIH, N. & LIÉGEOIS, J. P. 2003. The Moroccan Anti-Atlas: the West Africa craton passive margin with limited Pan-African activity. Implications for the northern limit of the craton: reply to comments by E. H. Bouougri. *Precambrian Research* **120**, 185–9.
- FARIAS, P., GALLASTEGUI, G., GONZÁLEZ-LODEIRO, F., MARQUÍNEZ, J., MARTÍN-PARRA, L. M., MARTÍNEZ CATALÁN, J. R., DE PABLO MACIÁ, J. G. & RODRÍGUEZ FERNÁNDEZ, L. R. 1987. Aportaciones al conocimiento de la litoestratigrafía y estructura de Galicia Central. *Memorias Museu Laboratorio Mineralogia Geologia Faculdade Ciências Universidade do Porto* **1**, 411–31.
- FRIEDL, G., FINGER, F., PAQUETTE, J. L., VON QUADT, A., MCNAUGHTON, N. J. & FLETCHER, I. R. 2004. Pre-Variscan geological events in the Austrian part of the Bohemian Massif deduced from U–Pb zircon ages. *International Journal of Earth Sciences* **93**, 802–23.
- GASQUET, D., LEVRESSE, G., CHEILLETZ, A., AZIZI-SAMIR, M. R. & MOUTTAQI, A. 2005. Contribution to a geodynamic reconstruction of the Anti-Atlas Morocco) during Pan-African times with the emphasis on inversion tectonics and metallogenic activity at the Precambrian–Cambrian transition. *Precambrian Research* **140**, 157–82.
- GONZÁLEZ-LODEIRO, F. 1981a. Posición de las series infraordovícicas en el extremo oriental del Sistema Central y su correlación. *Cuadernos Laboratorio Xeoloxico de Laxe* **2**, 125–34.
- GONZÁLEZ-LODEIRO, F. 1981b. La estructura del anticlinorio del “Ollo de Sapo” en la región de Hiendelaencina (extremo oriental del Sistema Central Español). *Cuadernos Geología Ibérica* **7**, 535–45.
- GOVINDARAJU, K., POTTS, P. J., WEBB, P. C. & WATSON, J. S. 1994. 1994 Report on Whin Sill Dolerite WS-E from England and Pitscurrie Microgabbro PM-S from Scotland: assessment by one hundred and four international laboratories. *Geostandards Newsletter* **18**, 211–300.
- GUTIÉRREZ MARCO, J. C., ROBARDET, M., RÁBANO, I., SARMIENTO, G. N., SAN JOSÉ LANCHÁ, M. A., HERRANZ, P. & PIEREN PIDAL, A. P. 2002. Ordovician. In *The Geology of Spain* (eds W. Gibbons & T. Moreno), pp. 31–49. London: Geological Society.
- HELBING, H. & TIEPOLO, M. 2005. Age determination of Ordovician magmatism in NE Sardinia and its bearing on Variscan basement evolution. *Journal of the Geological Society, London* **162**, 689–700.
- IGLESIAS PONCE DE LEÓN, M. & RIBEIRO, A. 1981. Position stratigraphique de la formation Ollo de Sapo dans la région de Zamora (Espagne)–Miranda do Douro (Portugal). *Comunicações Serviço Geológico de Portugal* **67**, 141–6.
- KARABINOS, P. 1997. An evaluation of the single-grain zircon evaporation method in highly discordant samples. *Geochimica et Cosmochimica Acta* **61**, 2467–74.
- KOBER, B. 1987. Single-zircon evaporation combined with Pb+ emitter-bedding for $^{207}\text{Pb}/^{206}\text{Pb}$ age investigations using thermal ion mass spectrometry and implications to

- zirconology. *Contributions to Mineralogy and Petrology* **96**, 63–71.
- LANCELOT, J. R., ALLEGRET, A. & IGLESIAS PONCE DE LEÓN, M. 1985. Outline of Upper Precambrian and Lower Palaeozoic evolution of the Iberian Peninsula according to U–Pb dating of zircons. *Earth and Planetary Science Letters* **74**, 325–37.
- LAUMONIER, B., AUTRAN, A., BARBEY, P., CHEILLETZ, A., BAUDIN, T., COCHERIE, A. & GUERROT, C. 2004. On the non-existence of a Cadomian basement in southern France (Pyrenees, Montagne Noire): implications for the significance of the pre-Variscan (pre-Upper Ordovician) series. *Bulletin Société Géologique de France* **175**, 643–55.
- MAZUR, S., TURNIAK, K. & BROCKER, M. 2004. Neoproterozoic and Cambro-Ordovician magmatism in the Variscan Kłodzko Metamorphic Complex (West Sudetes, Poland): new insights from U/Pb zircon dating. *International Journal of Earth Sciences* **93**, 758–72.
- MILLER, C. F., MCDOWELL, S. M. & MAPES, R. W. 2003. Hot and Cold Granites? Implications of Zircon Saturation Temperatures and Preservation of Inheritance. *Geology* **31**, 529–32.
- MONTERO, P. & BEA, F. 1998. Accurate determination of $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{143}\text{Sm}/^{144}\text{Nd}$ ratios by inductively-coupled-plasma mass spectrometry in isotope geoscience: an alternative to isotope dilution analysis. *Analytica Chimica Acta* **358**, 227–33.
- NAVIDAD, M., PEINADO, M. & CASILLAS, R. 1992. El magmatismo pre-Hercínico del Centro Peninsular (Sistema Central Español). In *Palaeozoic Inferior de Iberoamérica* (eds J. C. Gutiérrez-Marco, J. Saavedra & I. Rábano), pp. 485–94. Badajoz: Univ. Extremadura.
- PARGA-PONDAL, I., MATTE, P. & CAPDEVILA, R. 1964. Introduction a la geologie de “l’Ollo de Sapo”, formation porphyroide antésilurienne du nord ouest de l’Espagne. *Notas y Comunicaciones del Instituto Geológico y Minero de España* **76**, 119–53.
- PEARCE, N. J. G., PERKINS, W. T., WESTGATE, J. A., GORTON, M. P., JACKSON, S. E., NEAL, C. R. & CHENERY, S. P. 1997. A Compilation of New and Published Major and Trace Element Data for NIST SRM 610 and NIST SRM 612 Glass Reference Materials. *Geostandards Newsletter* **21**, 115–44.
- REISCHMANN, T., BACHTADSE, V., KRÖNER, A. & LAYER, P. 1992. Geochronology and paleomagnetism of a late Proterozoic island arc terrane from the Red Sea Hills, northeast Sudan. *Earth and Planetary Science Letters* **114**, 1–15.
- ROGER, F., RESPAUT, J. P., BRUNEL, M., MATTE, P. & PAQUETTE, J. L. 2004. Première datation U–Pb des orthogneiss illés de la zone axiale de la Montagne noire (Sud du Massif central): nouveaux témoins du magmatisme ordovicien dans la chaîne Varisque. *Comptes rendus Geoscience* **336**, 19–28.
- SCHÄFER, G. 1969. Geologie und Petrographie im östlichen Kastilischen Hauptscheidegebirge (Sierra de Guadarrama, Spanien). *Münstersche Forschungen zur Geologie und Paläontologie* **10**, 1–207.
- SCHALTEGGER, U. & GEBAUER, D. 1999. Pre-Alpine geochronology of the Central, Western and Southern Alps. *Schweizerische Mineralogische und Petrographische Mitteilungen* **79**, 79–87.
- SCHALTEGGER, U., ABRECHT, J. & CORFU, F. 2003. The Ordovician orogeny in the Alpine basement: constraints from geochronology and geochemistry in the Aar Massif (Central Alps). *Schweizerische Mineralogische und Petrographische Mitteilungen* **83**, 183–95.
- SOLÁ, A. R., MONTERO, P., RIBEIRO, M. L., NEIVA, A. M. R., ZINGER, T. & BEA, F. 2005. Pb/Pb zircon age of the Carrascal Massif, central Portugal. *Geochimica et Cosmochimica Acta* **69** (10, Supplement 1, Goldschmidt Conference Abstracts), A856.
- SOLÁ, A. R., PEREIRA, M. F., RIBEIRO, M. L., NEIVA, A. M. R., WILLIAMS, I. S., MONTERO, P., BEA, F. & ZINGER, T. 2006. The Urro Formation: Age and Precambrian inherited record. *Resumos do VII Congresso Nacional de Geologia de Portugal, Universidade de Evora* **1**, 29–32.
- STACEY, J. S. & KRAMERS, J. D. 1975. Approximation of terrestrial lead isotope evolution by a two-stage model. *Earth and Planetary Science Letters* **26**, 207–21.
- STEIGER, R. H. & JÄGER, E. 1977. Subcommission on Geochronology. Convention on the use of decay constants in geo- and cosmochronology. *Earth and Planetary Science Letters* **36**, 359–62.
- TAHIRI, A., SIMANCAS, J. F., EL HADI, H., GONZÁLEZ-LODEIRO, F., AZOR, A. & MARTÍNEZ POYATOS, D. 2005. A comparison of the Neoproterozoic/Lower Palaeozoic lithostratigraphy of Morocco and southwestern Iberia. *Geodynamic Interpretations. Geogaceta* **38**, 203–6.
- VALVERDE VAQUERO, P. & DUNNING, G. R. 2000. New U–Pb ages for Early Ordovician magmatism in Central Spain. *Journal of the Geological Society, London* **157**, 15–26.
- VIALETTE, Y., CASQUET, C., FÜSTER, J. M., IBARROLA, E., NAVIDAD, M., PEINADO, M. & VILLASECA, C. 1987. Geochronological study of orthogneisses from the Sierra de Guadarrama (Spanish Central System). *Neues Jahrbuch für Mineralogie, Monatshefte* **10**, 465–79.
- WHITEHOUSE, M. J., KAMBER, B. S. & MOORBATH, S. 1999. Age significance of U–Th–Pb zircon data from early Archaean rocks of west Greenland – a reassessment based on combined ion-microprobe and imaging studies. *Chemical Geology* **160**, 201–24.
- WIEDENBECK, M., ALLÉ, P., CORFU, F., GRIFFIN, W. L., MEIER, M., OBERLI, F., VON QUADT, A., RODDICK, J. C. & SPIEGEL, W. 1995. Three natural zircon standards for U–Th–Pb Lu–Hf trace element and REE analysis. *Geostandards Newsletter* **19**, 1–23.
- WILDBERG, H. G., BISCHOFF, L. & BAUMANN, A. 1989. U–Pb ages of zircons from meta-igneous and meta-sedimentary rocks of the Sierra de Guadarrama: Implications for the Central Iberia crustal evolution. *Contributions to Mineralogy and Petrology* **103**, 253–62.
- YORK, D. 1969. Least squares fitting of a straight line with correlated errors. *Earth and Planetary Science Letters* **5**, 320–4.
- ZECK, H. P., WINGATE, M. T. D., POOLEY, G. D. & UGIDOS, J. 2004. A Sequence of Pan-African and Hercynian Events Recorded in Zircons from an Orthogneiss from the Hercynian Belt of Western Central Iberia – an Ion Microprobe U–Pb Study. *Journal of Petrology* **45**, 1613–29.