New ideas on the Proterozoic-Early Palaeozoic evolution of NW Iberia: insights from U–Pb detrital zircon ages

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Abstract

U–Pb ages were obtained on single detrital zircon grains separated from six samples of Neoproterozoic and Lower Palaeozoic sedimentary and volcanosedimentary rocks from NW Iberia using the laser ablation microprobe-inductively coupled plasma mass spectrometry (LAM-ICP-MS) method. Precambrian greywackes yielded abundant zircons with Neoproterozoic (800 – 640 Ma) and Mesoproterozoic (0.9 – 1.2 Ga) ages, and a smaller proportion of Palaeoproterozoic (1.8 – 2 Ga) and Archaean zircons. Palaeozoic samples (Lower Cambrian and Ordovician) yielded abundant zircons with younger Neoproterozoic (ca. 550 and 620 Ma) and Mesoproterozoic (0.9 – 1.2 Ga) ages. Palaeoproterozoic (1.8 – 2 Ga) and Archaean zircons were also found. This data set, used in conjuction with previous paleogeographic and isotopic studies sheds new light on the Precambrian-early Palaeozoic evolution of NW Iberia and is consistent with the following sequence of events: (1) Early Cadomian-Avalonian subduction and arc construction (ca. 800 – 640 Ma). This magmatic episode created the main arc edifice (Avalonia); (2) full development of a back arc basin upon which the Neoproterozoic sediments were deposited (ca. 640 – 600 Ma). The combined U–Pb ages of detrital zircons and Nd isotopic features of these sedimenary rocks suggest that they were mostly shed from the main magmatic arc. On the basis of the presence of Grenvillian age detrital zircons with short waterborne transport before incorporation in the sediment, we propose that the basin was possibly located in a peri-Amazonian realm close to West Avalonian terranes. These basins were developed upon a cratonic basement that possibly involved both Grenvillian (ca. 0.9 – 1.2 Ga) and Transamazonian (ca. 1.9 – 2.1) igneous rocks. The reported zircon ages suggest a long-lived subduction, starting at ca. 800 Ma and terminated by ca. 580 – 570 Ma with no geological record of a final collision event; (3) the continuation of extension gave rise to the undocking of Avalonia from the back-arc. Detrital zircon ages in Lowermost Cambrian strata suggest that the main arc edifice had drifted away by ca. 550 – 540 Ma and was no longer shedding detritus into the back-arc basin. (4) During the Lower Ordovician, further extension of an already thinned crust gave rise to the Lower Ordovician 'Ollo de Sapo' magmatic event (ca. 480 Ma). Coeval volcanism in neighbouring areas displaying within-plate geochemical signatures is consistent with an extensional setting for the generation of the Lower Ordovician igneous and sedimentary rocks. Detrital zircon ages and Nd

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isotopic features of the Ordovician greywackes reflect both an increase in the contribution from older crustal components and the addition of newly accreted crust. A progressively thinning crust is a likely scenario that would explain the simultaneous exhumation of lower crustal (Grenvillian+ Transamazonian/Icartian) material and the generation of coeval magmatism. This latter scenario is consistent with models proposed for other circum-North Atlantic Avalonian-Cadomian terranes where repeated episodes of melting occurred in response to subduction and subsequent rifting events.

Keywords: Iberia; Proterozoic; Early Palaeozoic; U–Pb; Laser ablation; Detrital zircons

1. Introduction

Knowledge of the nature, age and distribution of basement terranes in orogenic zones provides constraints on the evolution of igneous and sedimentary rocks formed during subsequent orogenic cycles. In areas where cratonic basement is not exposed, constraints on the ages and nature of older basement can be acquired by: presence of inherited zircons in magmatic rocks; detrital zircons in sedimentary rocks; and by neodymium isotopic studies. The significance and interpretation of these types of data are not always straightforward. For example, neodymium model ages/isotopic compositions of granitoids and detrital sedimentary rocks may not be interpretable without knowledge of the ages of crustal reservoirs that constituted sources for those granitoid melts or sedimentary detritus. Interpretation of detrital zircon ages can be hampered by uncertainties about transport distances (cf. Rainbird et al., 1997). Nonetheless, these types of studies provide data that, when used in conjunction with other geological criteria, are important in constraining the tectonic and paleogeographic evolution of ancient orogens.

The Variscan belt of Western Europe is a well known example of an orogenic belt lacking exposures of cratonic igneous or metamorphic basement. With the exception of the Palaeoproterozoic gneisses exposed in the Channel Islands (Calvez and Vidal, 1978; Samson and D'Lemos, 1998), little is known of the western European pre-Neoproterozoic basement. In the northwestern Iberian Massif (NW Iberia), thick Neoproterozoic successions of pelites and greywackes are the oldest exposed rocks and there is great uncertainty about the nature and age of the basement upon which they were deposited. NW Iberia is considered to have been part of the Neoproterozoic assemblage terranes (e.g. Nance and Murphy, 1996) that comprise the Cadomian-Avalonian magmatic arc and associated basins that developed on an active Gondwanan margin. Studies devoted to different aspects of the geological history of these terranes have provided indirect evidence for the existence of an older pre-Neoproterozoic basement at depth (e.g. Liew and Hofmann, 1988; Nägler et al., 1995; Nance and Murphy, 1996; Keppie et al., 1998). Direct evidence is, however, rather scarce (e.g. data compilation in Nance and Murphy, 1996; van Staal et al., 1996; Keppie et al., 1998; Samson and D'Lemos, 1998; Sheridan et al., 1999). In NW Iberia, there is some U–Pb zircon evidence for Archaean (ca. 2.7 Ga) and Palaeoproterozoic (ca. 1.9 Ga) crustal components (Guerrot et al., 1989; Gebauer, 1993). Furthermore, the neodymium isotopic composition of Neoproterozoic to Carboniferous sedimentary rocks and Cadomian and Variscan granitoids requires at least some component derived from Mesoproterozoic or older rocks (e.g. Beetsma, 1995; Nägler et al., 1995; Moreno-Ventas et al., 1995; Fernández-Suárez et al., 1998).

In order to place constraints on the Proterozoic-Early Palaeozoic evolution of NW Iberia and to better integrate this area in paleogeographic reconstructions of circum-Atlantic Gondwanan terranes, we have undertaken a study of the U–Pb ages of detrital zircons obtained from Neoproterozoic, Cambrian and Ordovician sedimentary and volcanosedimentary rocks. These data were obtained by laser ablation microprobe-inductively coupled plasma-mass spectrometry (LAM-ICP-MS). The U–Pb ages reported here have significant implications that may inspire further work and discussion on the role of Iberia in

Avalonian-Cadomian and subsequent early Palaeozoic events.

2. Geological and tectonic framework

The autochthon of the Iberian Variscan belt (IVB) in NW Iberia is a well known example of a Palaeozoic collisional orogen (ca. 365 – 280 Ma) and has been extensively studied (Pérez-Estaún et al., 1991; Martínez Catalán et al., 1996, 1997; Dallmeyer et al., 1997; and references therein). The NW Iberian Variscan belt (Fig. 1) has been divided into four main paleogeographic zones each consisting of sedimentary and igneous rocks of Neoproterozoic and Palaeozoic age. The differ- ent zones have a common Variscan tectonic evo- lution with varying styles of deformation, and increasing metamorphic grade and abundance of granitoids from the foreland (Cantabrian Zone) towards the suture. Prior to the Variscan evolu- tion, there is a record of late Cambrian and early Ordovician magmatic and tectonic events. Ob- scured by these Palaeozoic events, the Precam- brian evolution of NW Iberia is still poorly constrained.

An important feature of the autochthonous units of the IVB is the existence of large exposures of Neoproterozoic pelite-greywacke sedimentary sequences (Fig. 1). These rocks are exposed at antiformal culminations in the Narcea Antiform (Cantabrian Zone– West Asturian Leonese Zone limit), the Lugo Dome (West Asturian Leonese Zone) and in the Central Iberian Zone (Fig. 1). The Neoproterozoic sedimentary rocks are interpreted to have been deposited in an extensional basin generated during the Cadomian orogeny (Vidal et al., 1994). The basement upon which these rocks were deposited is not exposed anywhere in the autochthonous zone of the IVB.

In the Narcea Antiform (Fig. 1) two Neoproterozoic series of clastic sedimentary rocks (lithic arenites, greywackes and shales) have been distinguished: a lower series with abundant volcanics and intruded by scarce granitoids (Tineo Series, Gutiérrez-Alonso and Fernández-Suárez, 1996); and an upper series dominated by pelitic rocks with rare volcanics. In the Lugo dome (Fig. 1) the Lower and Upper Villalba Series have the same characteristics as their counterparts in the Narcea Antiform, except for a higher metamorphic grade (low- to medium-grade) (Capdevila, 1969; Martínez Catalán et al., 1996). The ca. 580 – 600 Ma granitoids (Fig. 1) (Fernández-Suárez et al., 1998) intruding the Tineo Series provide a lower age limit for the deposition of these rocks, for which there is otherwise poor age control. Equivalent rocks in the Central Iberian Zone are also intruded by ca. 580 – 600 Ma granitoids (Leterrier and Noronha, 1998; Valle et al., 1999).

The uppermost Proterozoic-Lower Cambrian (Crimes et al., 1977) Cándana Formation unconformably overlays the Neoproterozoic rocks (Fig. 1) and is composed of quartzites, quartz-arenites, greywackes, arkoses and shales, interpreted to have been deposited in a braided-plain delta environment. Paleocurrents indicate that the provenance of the sediments is from the East– North East (van den Bosch, 1969; Crimes et al., 1977). In the Cantabrian and West Asturian Leonese zones, the Cándana Formation is overlain by a continuous Cambro-Ordovician passive continental margin sedimentary sequence. In the easternmost Central Iberian Zone, the Ordovician ‘Ollo de Sapo complex’, which includes both volcanosedimentary and magmatic rocks occurs in place of the passive continental margin sequence (Fig. 1). U– Pb geochronological data indicate Lower Ordovician ages for the magmatic rocks in the OS (Gebauer, 1993; Valverde-Vaquero and Dunning, 1997). The structure and origin of the OS is currently under debate and several hypothesis have been proposed (Capdevila, 1969; Martínez Catalán et al., 1997; Bastida et al., 1984; Ortega et al., 1996; Valverde-Vaquero and Dunning, 1997, 1999; Fernández-Suárez et al., 1998, 1999).

3. Samples and zircon morphology

Sample locations are shown in Fig. 1 and their stratigraphic position is schematically shown in the simplified columns of Fig. 2. The three Neoproterozoic rocks sampled for this study are

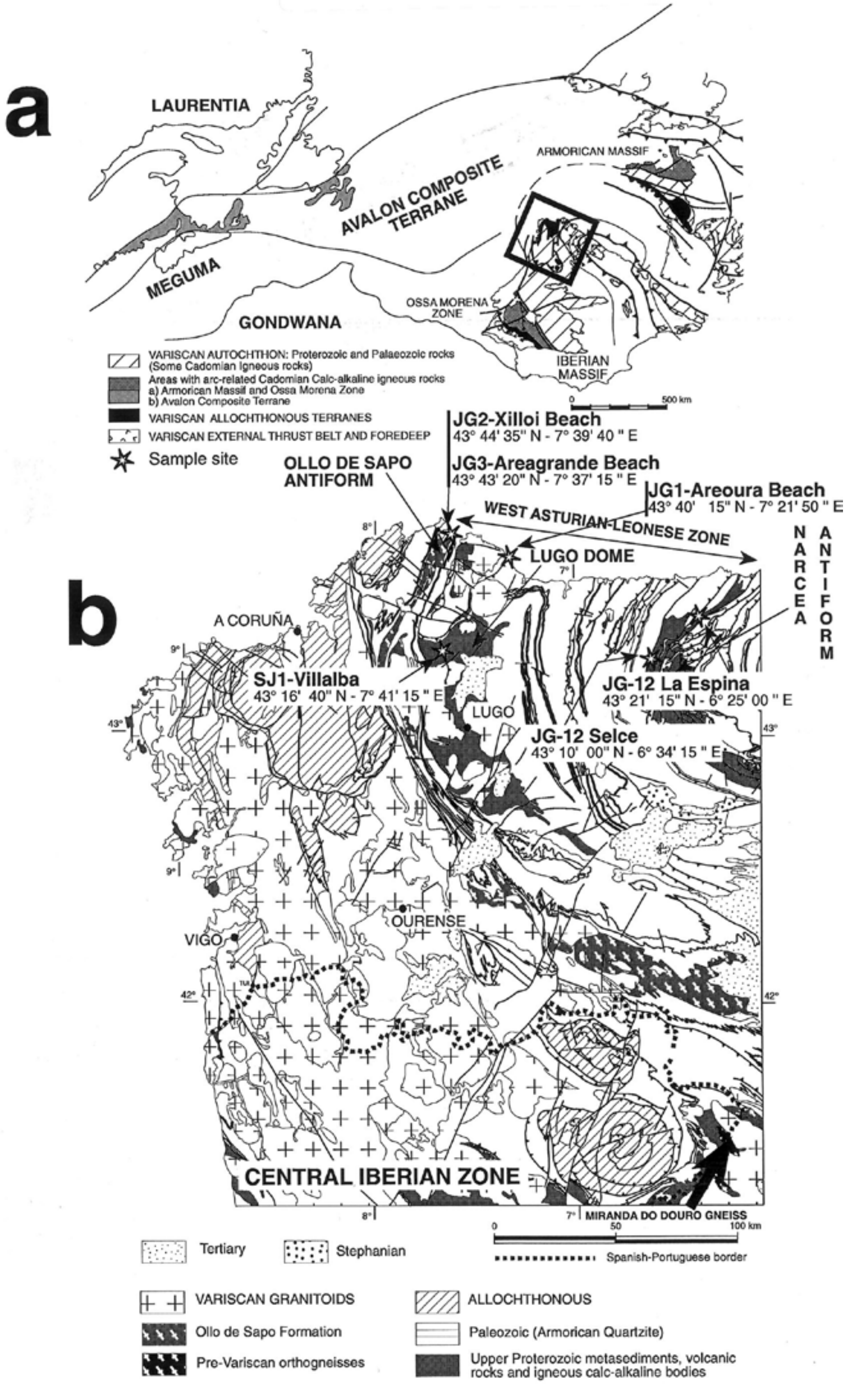

Fig. 1. (a) Reconstruction of Pangea around Iberia depicting the Ibero-Armorican Arc and the situation of the Avalon and Meguma terranes (after Martínez Catalán et al., 1997). (b) Geological map of part of the NW Iberian Variscan belt showing the main units, tectonic features and granitoids. Location of samples collected for this study is also indicated.

metamorphosed (low-grade) medium-grained feldspathic greywackes from the Tineo Series (JG-12), Lower Villalba Series (JG-1) and Upper Villalba Series (SJ-1). Zircons separated from these samples are very similar in morphology, with a predominance of subhedral (facetted to subfacetted) long prismatic, tabular or stubby prisms (Fig. 3). This suggests that sediments are proximal to their source and that these zircons are unlikely to have been involved in more than one sedimentary cycle.

The Cambrian sample (JG-16) is a medium grained sub-arkose collected at the lower part of the Cándana Formation, approximately 10 m above the unconformity with the Tineo-Narcea Series. Zircons separated from this sample (Fig. 3) are yellow to light pink subhedral prismatic crystals also suggesting a short sedimentary transport. A small proportion of zircons were transparent sub-rounded grains with a size of less than ca. 30 microns which renders them unsuitable for LAM-ICP-MS dating (see below).

The two Ordovician samples are coarse- (JG2), and medium- (JG3) grained feldspathic greywackes. These rocks have been interpreted as immature sediments containing significant amounts of re-deposited volcanic component (Ortega et al., 1996 and references therein). Mineralogical and textural features such as poor sorting, high amount of matrix, occurrence of euhedral-subhedral quartz and feldspars, embayed quartz and zoned plagioclase suggest that: (i) these rocks were derived mainly from erosion of igneous sources (Capdevila, 1969; Ortega et al., 1996); and (ii) the sedimentary transport distance was short. Most zircons separated from the greywackes (and all the grains analysed) were euhedral to subhedral long to short prismatic or tabular suggesting a short transport and the improbability of their having undergone more than one sedimentary cycle.

Based on the immature (proximal) nature of the sedimentary rocks and on the morphological features of the zircons, we consider that the studied samples are unlikely to contain recycled zircons of exotic provenance (i.e. very long sedimentary transport, cf. Rainbird et al., 1997). It is also important to note that our results show no apparent correlation between the morphology of zircon grains and their U–Pb age.

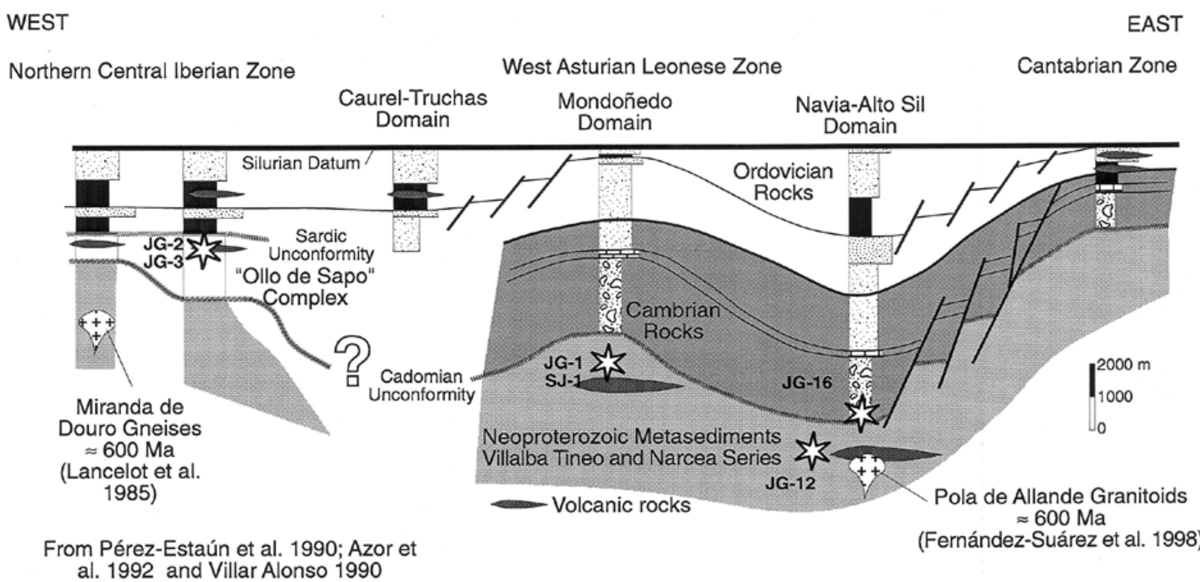

Fig. 2. Simplified stratigraphic column showing sample location in relation to the stratigraphic record.

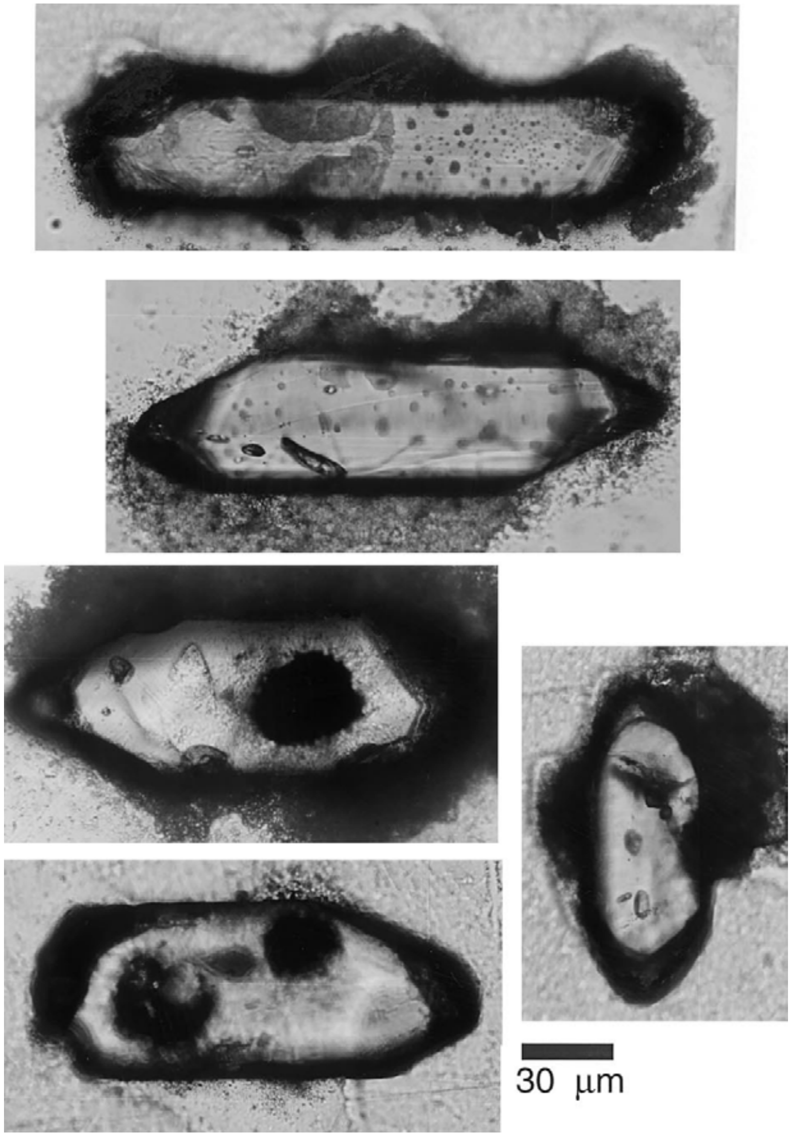

Fig. 3. Selected photographs of zircons from the studied samples.

4. Sample preparation and analytical method

Zircons were separated using conventional separation techniques (Wilfley table, heavy liquids and magnetic separation). Hand picked zircon grains were mounted in epoxy blocks, then polished to obtain an even surface, and cleaned in an acid bath prior to analysis. The LAM ICP-MS instrumentation, analytical conditions and data reduction software, were developed at Memorial University of Newfoundland and are partially described in Jackson et al. (1997). The instrumentation consists of an in-house built laser ablation microprobe incorporating a frequency quadrupled Nd:YAG laser (,= 266 nm in the UV). Ablated material is analysed by an enhanced sensitivity ICP-MS (FISONS VG PQII+'S'). This instrumentation has low ppb detection limits for U and

Pb at 30 – 50 micron sampling resolution. Ablation of the zircons is accompanied by some elemental fractionation so to minimize this we: (a) use a defocused laser beam to reduce thermal edge effects; (b) direct a jet of cooling Ar at the ablation point; and (c) use identical ablation conditions on the sample and standard (including the interval integrated for each analysis). Calibration is made using an in-house concordant zircon standard (Oslo Rift 02123 @ 295 ± 1 Ma by thermal ionization mass spectrometry). Three measurements on the zircon standard and one on NBS 612 glass are made at the beginning and end of each run of 20 measurements. High instrumental Hg background prohibits accurate measurement of ^{204}Pb, so no common Pb correction is made. Pb background is minimised by careful attention to system cleanliness, and monitored throughout a run. Data are acquired in peak jumping mode using time-resolved software. Each analysis requires about 60 s of background, followed by 10 – 40 s of ablation depending on the sample size. The time-resolved signal data is reduced off-line, where it is possible to see the change in isotopic ratio profiles to detect zoning, and/or heterogeneities. Appropriate signal intervals are then integrated accordingly. ^{235}U counts are calculated from ^{238}U. The strategies described allow the ratios of $^{206}Pb/^{238}U$, $^{207}Pb/^{235}U$, and $^{208}Pb/^{232}Th$ to be measured with precision (1 r.s.d.) of about ± 5% and $^{207}Pb/^{206}Pb$ to ± 1%. Two sigma uncertainties on the ages determined using the individual ratios are given in Table 1. Accuracy using these procedures is estimated to be ± 10–20 Ma based on comparing TIMS analyses of concordant zircons ranging from ca. 290 – 2700 Ma, with those obtained by LAM-ICP-MS on zircons from the same sample (sometimes the same grain) (Jackson and Dunning, unpublished).

5. Data presentation and U–Pb results

LAM-ICP-MS is a relatively new technique and there is little background concerning the interpretation, discussion and presentation of analytical data, in comparison with the large amount of studies based on ion microprobe (SHRIMP, SIMS) U– Pb single zircon dating. In the latter case, results of U– Pb analyses are currently reported in different ways such as Tera-Wasserburg ($^{207}Pb/^{206}Pb$ vs $^{238}U/^{206}Pb$) concordia diagrams, conventional ($^{206}Pb/^{238}U$ vs $^{207}Pb/^{235}U$) concordia plots, age histograms and cumulative probability plots. A common feature to these techniques, in contrast to single grain TIMS U– Pb analyses, are the data reduction strategies used (cf. Hirata and Nesbitt, 1995) and the assumptions made for isotopic ratio corrections. The above, combined with a lower capacity for isotope mass separation and measurement, results in less precise data and sometimes yields discordant data points that are not interpretable in a straightforward manner. In the present study, which reports a set of 83 zircon U– Th– Pb LAM-ICP-MS analyses with varying degrees of concordancy and precision, our major aim has been to identify the modes in the age spectra. We have been able to use the geological constraints on the data as a means for testing the meaningfulness of actual ages outside of the purely analytical interpretation. In addition, data sets were acquired on different selections of zircons in a number of samples, over a period of time, by three different analysts. These data sets all yielded consistent results. These analytical results are discussed in the following sections.

The analytical data containing isotope ratios, ages and errors are reported in Table 1. We report $Pb/^{238}U/^{207}Pb/^{235}U$ concordant analyses on conventional concordia plots (Fig. 4). These plots were made using an unpublished program from the Royal Ontario Museum. In addition, a frequency histogram of $^{207}Pb/^{206}Pb$ ages (Fig. 5) is presented, that includes data from concordant and non-concordant analyses. We now briefly comment upon the criteria followed to select the best age estimate when all analyses are not concordant within error limits. For younger zircons in the age range from about 500 to 800 Ma, the $^{206}Pb/^{238}U$ ages are preferred because $^{207}Pb/^{206}Pb$ ages are compromised by low ^{207}Pb count rates. In these younger zircons, $^{208}Pb/^{232}Th$ ages (Table 1) are often consistent with the $^{206}Pb/^{238}U$ age. In discordant zircons older than ca. 1.5 Ga, we use $^{207}Pb/^{206}Pb$ ages because they have comparatively

Table 1
LAM-ICP-MS U–Pb results

Analysis	$^{207}Pb/^{206}Pb$	$^{208}Pb/^{232}Th$	$^{207}Pb/^{235}U$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$		$^{206}Pb/^{238}U$		$^{207}Pb/^{206}Pb$		$^{208}Pb/^{232}Th$	
	ratio	ratio	ratio	ratio	Age (Ma)	2	Age (Ma)	2	Age (Ma)	2	Age (Ma)	2
Sample JG1												
Z1	0.06422	0.04784	1.1732	0.1320	788	42	802	18	748	106	945	93
Z2	0.06666	0.06150	1.1587	0.1261	781	84	765	12	810	210	807	150
Z3	0.06616	0.04297	1.1005	0.1206	754	36	734	15	810	120	820	35
Z4	0.06187	0.03790	1.0523	0.1234	730	50	750	24	670	144	752	90
Z5	0.07133	0.04990	1.6029	0.1630	971	96	973	29	966	248	984	71
Z6	0.07472	0.05445	1.7730	0.1721	1036	67	1024	24	1060	108	1072	71
Z7	0.06781	0.05985	1.4735	0.1576	920	55	943	38	870	90	1175	55
Z8	0.19366	0.11707	9.2877	0.3478	2367	81	1924	98	2772	152	2238	200
Z9	0.07124	0.04059	1.3420	0.1366	864	49	826	44	964	170	904	83
Z10	0.06973	0.05943	1.7695	0.1841	1034	81	1089	61	920	218	1168	78
Z11	0.07608	0.04474	1.5921	0.1518	967	55	911	42	1096	198	885	117
Z12	0.06392	0.02684	0.8283	0.0940	613	50	579	52	738	134	685	62
Z13	0.07005	0.04998	1.7446	0.1806	1025	44	1070	48	928	90	986	53
Z14	0.06762	0.04838	1.4920	0.1600	927	50	957	55	856	126	955	40
Z15	0.06310	0.03606	1.1159	0.1283	761	36	778	20	710	160	716	14
Z16	0.06765	0.04712	1.4536	0.1558	911	47	934	57	856	112	931	59
Z17	0.06453	0.03107	1.0206	0.1147	714	59	700	63	758	154	618	53
Z18	0.10063	0.07156	2.6214	0.1889	1307	29	1116	7	1634	64	1397	49
Sample SJ1												
Z1	0.07324	0.05062	1.6371	0.1621	985	30	968	24	1020	84	998	22
Z2	0.07135	0.03451	1.1868	0.1206	794	67	734	26	966	292	686	37
Z3	0.06853	0.02844	1.0489	0.1110	728	47	679	13	884	178	667	32
Z4	0.08298	0.06352	2.1493	0.1879	1165	22	1110	13	1268	70	1245	55
Z5	0.07521	0.04585	1.5077	0.1454	934	21	875	16	1074	72	926	18
Z6	0.06233	0.03150	0.8950	0.1041	649	39	639	19	684	176	627	33
Z7	0.18392	0.13036	10.7787	0.4250	2504	26	2283	61	2688	50	2477	44
Z8	0.18117	0.11694	9.3416	0.3740	2372	33	2048	62	2662	46	2235	51
Z9	0.16074	0.09445	6.7988	0.3066	2086	34	1724	71	2462	40	1824	137
Z10	0.12238	0.08180	4.5926	0.2720	1748	37	1551	29	1990	94	1589	30
Z11	0.07437	0.03050	0.9333	0.0910	669	44	561	36	1050	146	607	55
Z12	0.06495	0.02268	0.6997	0.0781	539	26	485	9	772	140	753	36
Z13	0.06378	0.02448	0.6776	0.0770	525	21	478	8	734	110	789	27
Z14	0.08931	0.04848	1.8428	0.1496	1061	60	899	31	1410	140	957	63
Z15	0.06759	0.03317	1.1645	0.1249	784	26	759	12	854	92	660	40
Z16	0.07406	0.05227	1.7962	0.1758	1044	59	1044	77	1042	82	1030	127
Z17	0.13953	0.09612	3.0919	0.1607	1431	86	960	46	2220	190	1855	243
Z18	0.07752	0.06718	2.2928	0.2145	1210	45	1253	57	1134	82	1314	72
Z19	0.06715	0.03946	1.1057	0.1194	756	40	727	36	842	122	782	61

Table 1 (*Continued*)

Analysis	$^{207}Pb/^{206}Pb$	$^{208}Pb/^{232}Th$	$^{207}Pb/^{235}U$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$		$^{206}Pb/^{238}U$		$^{207}Pb/^{206}Pb$		$^{208}Pb/^{232}Th$	
	ratio	ratio	ratio	ratio	Age (Ma)	2	Age (Ma)	2	Age (Ma)	2	Age (Ma)	2
Sample JG12												
Z1	0.06086	0.03637	0.9072	0.1081	656	55	662	32	634	146	722	49
Z2	0.06326	0.03552	0.9887	0.1133	698	19	692	17	716	112	705	14
Z3	0.07485	0.04627	1.6027	0.1553	971	56	931	22	1064	62	994	68
Z4	0.07859	0.05579	1.6059	0.1482	972	97	899	80	1160	192	1097	74
Z5	0.06443	0.04122	1.0360	0.1166	722	61	711	27	754	236	816	121
Z6	0.11505	0.08049	4.7499	0.2994	1776	37	1688	35	1880	90	1690	22
Z7	0.06463	0.03994	1.1580	0.1300	758	91	788	14	650	290	781	33
Z8	0.11257	0.12303	5.1013	0.3287	1836	65	1832	55	1840	146	1945	111
Z9	0.07376	0.05312	1.8595	0.1828	1067	83	1082	32	1034	230	1046	88
Z10	0.09780	0.07265	2.5323	0.1878	1281	87	1109	52	1582	198	1481	119
Z11	0.07196	0.04886	1.6706	0.1684	997	90	1003	74	984	152	964	145
Z12	0.06624	0.04231	1.2514	0.1370	824	35	828	23	814	86	838	39
Z13	0.08740	0.07322	2.5964	0.2154	1300	73	1260	56	1358	224	1228	120
Sample JG16												
Z1	0.05531	0.02567	0.6758	0.0886	524	60	547	17	484	228	512	32
Z2	0.07170	0.02782	0.9208	0.0931	663	96	574	29	976	206	555	74
Z3	0.06329	0.03176	0.8854	0.1014	644	73	623	22	718	270	632	63
Z4	0.16592	0.10206	7.9378	0.3470	2224	101	1920	86	2516	208	1964	211
Z5	0.15332	0.10491	7.8107	0.3695	2209	26	2027	22	2382	48	2017	51
Z6	0.06604	0.02785	0.8266	0.09098	612	68	560	28	806	294	573	45
Z7	0.07302	0.04098	1.3646	0.1355	874	62	819	28	1014	100	972	83
Z8	0.17240	0.10083	9.3368	0.3928	2372	46	2136	52	2580	60	1942	54
Z9	0.06020	0.02688	0.7513	0.0905	569	34	559	16	610	186	536	31
Z10	0.06545	0.02891	0.8248	0.0985	611	20	600	8	620	94	576	54
Z11	0.06764	0.03715	1.1561	0.1240	780	31	753	21	856	114	737	41
Z12	0.07296	0.04341	1.4440	0.1435	907	81	865	30	1012	258	959	38
Z13	0.07613	0.03344	1.2212	0.1163	810	52	709	23	1098	174	995	34
Z14	0.06097	0.02640	0.8266	0.0983	612	43	605	15	638	206	567	34
Z15	0.06445	0.02111	0.7679	0.0864	579	19	534	16	756	106	522	18
Sample JG2												
Z1	0.11890	0.10200	5.0798	0.3537	1946	19	1952	24	1938	32	1963	30
Z2	0.07611	0.05145	1.9224	0.1832	1089	20	1084	14	1098	60	1014	92
Z3	0.07478	0.05106	1.9251	0.1867	1090	16	1104	15	1062	54	1017	35
Z4	0.06047	0.02678	0.7694	0.0923	579	29	569	18	620	120	534	32
Z5	0.05605	0.01992	0.5507	0.0713	445	29	444	16	454	140	400	30
Z6	0.05418	0.01870	0.5315	0.0711	433	28	443	8	380	190	374	30
Z7	0.20557	0.010283	12.1646	0.4292	2617	18	2302	26	2840	24	1978	32
Z8	0.11769	0.07568	4.8652	0.2998	1796	24	1690	23	1920	52	1475	39

Table 1 (*Continued*)

Analysis	$^{207}Pb/^{206}Pb$	$^{208}Pb/^{232}Th$	$^{207}Pb/^{235}U$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$		$^{206}Pb/^{238}U$		$^{207}Pb/^{206}Pb$		$^{208}Pb/^{232}Th$	
	ratio	ratio	ratio	ratio	Age (Ma)	2	Age (Ma)	2	Age (Ma)	2	Age (Ma)	2
Z9	0.07862	0.06650	1.5280	0.1410	942	48	850	47	1162	114	1301	107
Sample JG3												
Z1	0.12189	0.08705	5.9961	0.3568	1975	35	1967	37	1984	68	1687	67
Z2	0.12349	0.09507	5.7901	0.3353	1933	37	1864	31	2006	80	1836	58
Z3	0.07760	0.05865	2.1556	0.2015	1167	19	1183	12	1136	58	1152	48
Z4	0.06133	0.02819	0.8174	0.0967	607	21	595	9	650	98	562	16
Z5	0.058782	0.02916	0.7992	0.1002	596	12	616	7	522	56	581	11
Z6	0.05925	0.03633	0.7886	0.0965	590	36	594	30	576	100	546	40
Z7	0.05440	0.02123	0.5546	0.0739	448	16	460	6	386	92	425	11
Z8	0.11200	0.10426	4.8299	0.31270	1790	32	1754	34	1832	54	2005	41
Z9	0.07669	0.06984	2.3241	0.2198	1220	24	1281	20	1112	64	1265	45

smaller uncertainties owing to higher amounts of ^{207}Pb. In the case of zircons of intermediate age (1–1.5 Ga) $^{207}Pb/^{206}Pb$ age is preferred when it is consistent with the $^{208}Pb/^{238}Th$ age. The $^{208}Pb/^{238}U$ age is otherwise preferred. Similar weighting of young ages to $^{206}Pb/^{238}U$ and older ages to

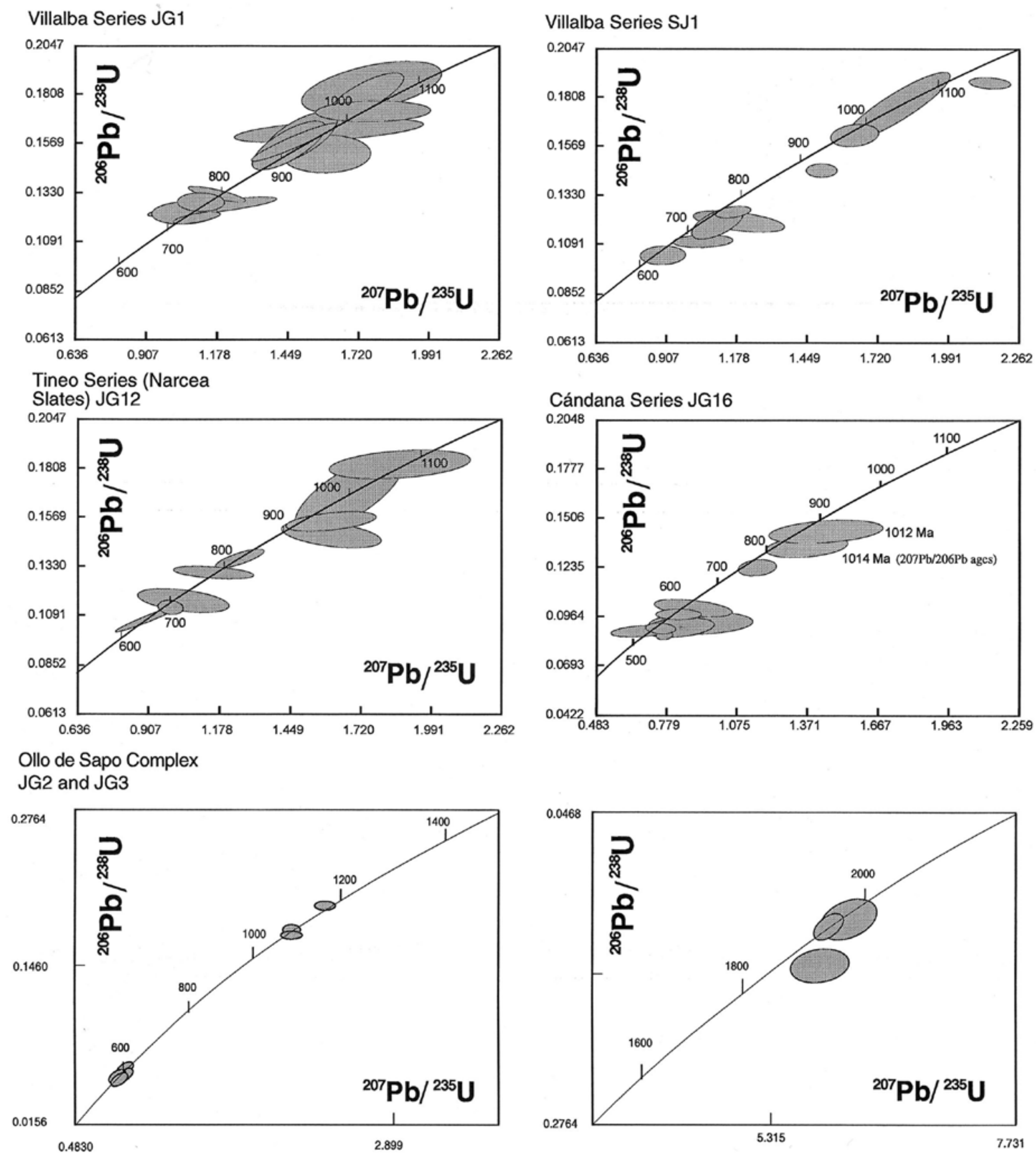

Fig. 4. Concordia plots of U–Pb analytical data. Ellipses represent 2 uncertainties.

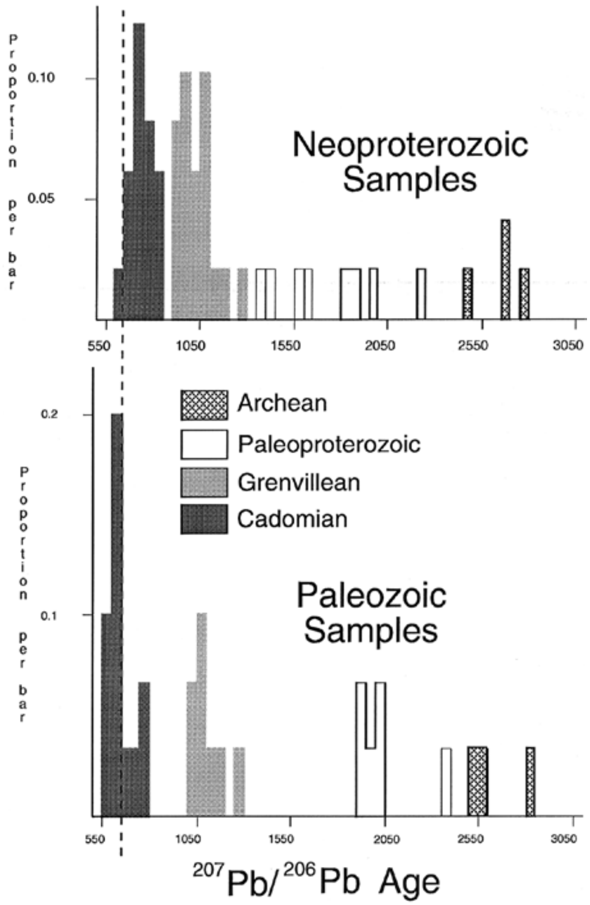

Fig. 5. Histograms of $^{207}Pb/^{206}Pb$ ages. (a) Neoproterozoic samples; (b) lower Palaeozoic samples.

$^{207}Pb/^{206}Pb$ is used in ion probe work (cf. Ireland et al., 1998).

5.1. Proterozoic metagreywackes (JG1, JG12 and SJ1) (Fig. 4a, b, c)

Zircons in sample JG-1 (Lower Villalba Series) yielded a number of sub-concordant analyses with Neoproterozoic $^{206}Pb/^{238}U$ ages ranging from ca. 800 to ca. 700 Ma and Middle Proterozoic $^{206}Pb/^{238}U$ ages ranging from 911 to 1089 Ma. Zircons Z8 and Z18 are discordant and have Achaean (ca. 2.77 Ga) and Palaeoproterozoic (ca. 1.63 Ga) $^{207}Pb/^{206}Pb$ ages. Z12 has a somewhat ambiguous age around 600 Ma, while Z9 is best described by an age of ca. 800−850 Ma.

Five zircons from sample SJ-1 (Upper Villalba Series) yielded Late Proterozoic subconcordant ages (Table 1, Fig. 4b) with $^{206}Pb/^{238}U$ ages ranging from 639 to 759 Ma. Z12 and Z13 are discordant and yielded Neoproterozoic $^{207}Pb/^{206}Pb$ ages of 772 and 734 Ma, respectively. Note that Z12 is a rim to Z19. Zircons Z1, Z16 and Z18 are subconcordant and yielded Middle Proterozoic $^{206}Pb/^{238}U$ ages of 968, 1044 and 1253 Ma, respectively. Zircons Z4, Z5 and Z11 are discordant and yielded Mesoproterozoic $^{207}Pb/^{206}Pb$ ages of 1268, 1074 and 1050 Ma, respectively. Zircons Z10 and Z17 are discordant and yielded Palaeoproterozoic $^{207}Pb/^{206}Pb$ ages of 1990 and 2220 Ma, respectively. Discordant analysis Z14 yielded a late Mesoproterozoic $^{207}Pb/^{206}Pb$ age (1410 Ma). Finally, discordant analyses Z7, Z8 and Z9 yielded Archaean $^{207}Pb/^{206}Pb$ ages between 2462 and 2688 Ma.

Five zircons from sample JG-12 (Tineo Series) yielded Late Proterozoic ages (Fig. 4c, Table 1), with $^{206}Pb/^{238}U$ ages ranging from 662 to 828 Ma. Three zircons yielded concordant Mesoproterozoic ages with $^{206}Pb/^{238}U$ ages of 1082, 1003 and 1260 Ma, respectively. Z3 and Z4 yielded subconcordant data points with $^{206}Pb/^{238}U$ ages of 931 and 900 Ma and $^{207}Pb/^{206}Pb$ ages of 1064 and 1164 Ma, respectively. In this case, the $^{207}Pb/^{206}Pb$ ages are preferred as they are more consistent with the Th/Pb ages (Table 1). Zircon Z8 yielded a concordant Palaeoproterozoic age of ca. 1.83 Ga ($^{206}Pb/^{238}U$ age) (not shown in Fig. 4). Zircons Z6 and Z10 are discordant and yielded Palaeoproterozoic $^{207}Pb/^{206}Pb$ ages of ca. 1.88 and 1.58 Ga, respectively.

5.2. Cambrian sandstone (Cándana Formation, JG16)

Eight zircons separated from the Lower Cambrian arkose yielded Late Proterozoic ages younger than those of zircons found in the Neoproterozoic greywackes, the $^{206}Pb/^{238}U$ ages ranging from ca. 540 to ca. 623 Ma. Sub-concordant (Z7, Z12) and discordant (Z13) zircons yielded $^{207}Pb/^{206}Pb$ Mesoproterozoic ages of 1014, 1012 and 1098 Ma, respectively. Although the uncertainties associated to the $^{207}Pb/^{206}Pb$ ages are quite large, these ages are in this case preferred to the U–Pb ages since they are more consistent with the Th/Pb ages (Table 1). In addition, three discordant zircons (Z4, Z5 and Z8) have $^{207}Pb/^{206}Pb$ ages of ca. 2.4 – 2.6 Ga (Archaean-Early Proterozoic boundary).

5.3. Ordovician volcanosedimentary metagreywackes (Ollo de Sapo complex, JG2 and JG3)

Zircons from these samples yielded mostly concordant U–Pb ages that fall into four age groupings: Ordovician (ca. 440 – 460 Ma), Late Neoproterozoic (ca. 570–620), Mesoproterozoic (ca. 1–1.2 Ga) and Palaeoproterozoic (ca. 1.9–2 Ga). The youngest U–Pb age grouping (ca. 440–460, Table 1) is considered to represent the age of the volcanic component of the OS rocks. This age is younger than ages reported for other magmatic units of this complex in areas situated further to the S-SE along the OS antiform (ca. 465–490 Ma, Lancelot et al., 1985; Gebauer, 1993; Valverde-Vaquero and Dunning, 1997). One zircon from sample JG2 (Z4) and three from sample JG-3 (Z4, Z5, Z6) yielded Late Proterozoic concordant ages (Fig. 4e). The $^{206}Pb/^{238}U$ ages range from ca. 569 to ca. 616 Ma. The third group, of which three concordant ages are reported in Fig. 4e, represents the recycling of a mid-Proterozoic crustal component of ca. 1–1.2 Ga. Two zircons from sample JG-2 (Z2 and Z3) yielded overlapping concordant ages of ca. 1080 – 1105 Ma and a zircon from sample JG3 (Z3) yielded an older concordant age of ca. 1183 Ma. Zircons Z9 (JG2) and Z9 (JG3) yielded discordant data points with $^{207}Pb/^{206}Pb$ ages of 1162 and 1112 Ma, respectively. The fourth group of recycled zircons is represented by two concordant and partially overlapping U–Pb ages of ca. 1952 Ma (Z1, sample JG2) and ca. 1967 Ma (Z1, sample JG3). A second zircon from JG3 (Z2), although not concordant, yielded a $^{207}Pb/^{206}Pb$ age of 2 Ga, within error limits of the concordant zircon from the same sample. Discordant analyses Z8 (JG2) and Z8 (JG3) yielded $^{207}Pb/^{206}Pb$ ages of 1920 and 1832 Ma, respectively. Finally, a discordant zircon from sample JG-2 (Z6) has a $^{207}Pb/^{206}Pb$ age of ca. 2.8 Ga, the oldest Archaean age found in this study.

6. Discussion

Before discussing the main implications of the ages reported in this study, we must consider how representative the samples analysed are. It is difficult to ascertain how many individual analyses might be required; for instance Ireland et al. (1998) use 50 zircons per sample. However, we consider that given: (a) the geological consistency of the data; (b) our own tests, wherein we analysed additional grains (eight to ten) on separate occasions but found no new age groups; and (c) the implications of actual ages and apparent age-group distribution; the results of this study provide sufficient grounds for a preliminary hypothesis on their geological implications.

6.1. Comparison of zircon ages with tectonothermal and crustal growth events in Gondwana

The U–Pb ages of detrital zircons in Neoproterozoic and early Palaeozoic rocks from NW Iberia reveal a varied provenance including newly accreted crust and old cratonic crustal sources. The youngest zircon age population found in the Neoproterozoic greywackes (ca. 640–800 Ma) corresponds to the period of early arc magmatism in the peripheral orogens around Gondwana, in which subduction related magmatic activity began at ca. 800 Ma (e.g. Murphy and Nance, 1991). Although the most complete record of early arc activity is preserved in the Arabian shield (e.g. Kröner et al., 1988), there is also a record of this early stage in the Avalonian-Cadomian belt (O'Brien et al., 1996; Egal et al., 1996). The data reported in this study suggest that the Neoproterozoic sedimentary rocks in NW Iberia contain a significant detrital component generated by erosion of igneous rocks formed in the early stages of the construction of the Cadomian-Avalonian magmatic arc. The data also bracket the deposition age of the Neoproterozoic sedimentary sequences between ca. 600 Ma (age of granitoids intruded into the Tineo Series, Fernández-Suárez et al., 1998) and ca. 640 Ma (age of the youngest detrital zircons found in these Formations). This age interval matches the period of full development of back-arc basins in the Avalon Composite Terrane (ca. 630 – 610 Ma) proposed by Keppie et al. (1998).

Cambrian and Ordovician samples in contrast (Figs. 4 and 5 and Table 1) yielded very few zircons in the 640 – 800 age group and instead contain abundant younger zircons (ca. 550 – 620 Ma). This age interval corresponds to the crystallisation ages of Cadomian granitoids intruded into the Neoproterozoic sequences in NW Iberia (Lancelot et al., 1985; Fernández-Suárez et al., 1998; Leterrier and Noronha, 1998; Valle et al., 1999). The data indicate that by the Early Cambrian, the arc source furnishing the ca. 640 – 800 Ma zircons had either been eroded/submerged or had drifted away from the platform upon which the Palaeozoic rocks were deposited, or a combination of both.

The ca. 0.9 – 1.2 Ga old zircons from Neoproterozoic and Palaeozoic samples are the first evidence for a Grenvillian age crustal component in NW Iberia, which has important implications for its Neoproterozoic position (see below). The data indicate that Mesoproterozoic zircons found in Neoproterozoic-Lower Palaeozoic sedimentary rocks from NW Iberia have ages that match the span of the Grenville event in Laurentia or the Sunsas/Rondonian event in the Amazon craton (e.g., Nance and Murphy, 1996; Keppie et al., 1998 and references therein). Similar ages have been reported from the Moldanubian zone of the Bohemian Massif and from the French Massif Central (Gebauer et al., 1989). The provenance area of the Grenvillian zircons could be sought either in exposed basement rocks of the Avalonian/Cadomian arc (van Staal et al., 1996; Sheridan et al., 1999) or in a cratonic source located in the margin of Gondwana during the Neoproterozoic/Early Palaeozoic. The proximal character of an igneous Grenvilli an age source is supported by the morphological features of zircons, indicating a short waterborne transport history, and the improbability of their being involved in more than one sedimentary cycle. It is also noteworthy that the Neoproterozoic sedimentary rocks seem to contain rare zircons of pre-Mesoproterozoic age relative to zircons of Grenvillian and Early Cadomian age (Table 1, Fig. 5). This may suggest that the arc contributed more detritus to the basin where Neoproterozoic rocks were deposited than the continental cratonic source. In addition, Nd isotope signatures of Cadomian granitoids in NW Iberia intruded into the Neoproterozoic sedimentary rocks at ca. 600 Ma (Fernández-Suárez et al., 1998; Fig. 4) suggest the possible existence of a basement component with a broadly Grenvillian mantle extraction age, a feature frequently found in other Cadomian/Avalonian circum-Atlantic terranes (e.g. Murphy et al., 1996).

Zircons with pre-Mesoproterozoic ages are present in all the sedimentary rocks studied. However, only three concordant analyses were obtained in zircons from samples JG-12, JG-2 and JG-3 (Table 1). These data are consistent with the existence of a Palaeoproterozoic (ca. 1.8 – 2.0 Ga) crustal component in the basement of NW Iberia (cf. Gebauer, 1993). This age corresponds to the Icartian-Eburnian-Transamazonian events recorded in different terranes of Africa and South America (e.g. Keppie et al., 1998 and references therein). In western Europe, the ca. 1.8 – 2.1 Ga basement is represented by the Icartian granitic orthogneisses (2061 ± 2 Ma) exposed in the Channel Islands (North Armorican Massif of Cadomia) (Samson and D’Lemos, 1998). Similar ages (1.8 – 1.9 Ga) were reported from acid granulite boulders collected in the submarine environment of the northern coast of Spain (Guerrot et al., 1989). Gebauer et al. (1989) reported similar ages (ca. 1.85 Ga) for detrital zircons in high-grade metasedimentary rocks from the Montaigne Noire of the French Massif Central. The presence of ca. 1.8 – 2 Ga zircons in sedimentary rocks from NW Iberia is in agreement with its peri-Gondwanan location during Neoproterozoic-early Palaeozoic times, but can be interpreted in different ways: (i) the zircons are derived from igneous sources like those represented by the Icart gneisses; (ii) they represent inherited xenocrysts in Grenvillian or Early Cadomian igneous rocks; (iii) they are recycled from a post-Icartian sedimentary formation that is no longer exposed. The third possiblity is not favoured on the basis of zircon morphology and the features of the host sedimentary rocks (see above). The second possibility also seems improbable given the concordant nature of the analyses. If these zircons were hosted by rocks that suffered high-grade metamorphism and anatexis, they are likely to have suffered Pb loss. Therefore we suggest that they are likely derived from Icartian-Eburnean-Transmazonian igneous sources. The reported Archaean $^{207}Pb/^{206}Pb$ ages (2.5 – 2.8 Ga) match the age span of the Central Amazonian tectonothermal events in the Amazon craton (Teixeira et al., 1989).

6.2. Insights from comparison of U–Pb and Sm–Nd data

Geochemical and Sm– Nd data are available for Neoproterozoic and Palaeozoic metasedimentary rocks from NW Iberia (Beetsma, 1995; Ortega, 1995; Nägler et al., 1995). These data are summarised in Fig. 6 and briefly discussed below in the light of the ages of detrital zircons obtained in this study. The major and trace element signatures of Neoproterozoic rocks in NW Iberia are compatible with an origin in an active margin-continental arc setting (Beetsma, 1995; Nägler et al., 1995). This is consistent with relatively high ε_{Nd} values (> -5) at the time of deposition and low T_{DM} values (1.3 – 1.6 Ga, Fig. 6), that indicate an important contribution of juvenile crust in the Neoproterozoic sedimentary rocks. This is consistent with the ages of detrital zircons reported in this study, which confirm a dominant input of Early Cadomian/Avalonian (ca. 800 – 640 Ma) crust in the Neoproterozoic sedimentary rocks and also a significant input from a likely igneous Grenvillian source. As illustrated in Fig. 6, the Nd isotopic composition of Neoproterozoic sedimentary rocks is compatible with a dominant input from Cadomian and Grenvillian juvenile crust, plus a minor component of older more enriched sources (Palaeoproterozoic, Archaean).

In contrast, Cambrian sedimentary rocks (even the lowermost Cambrian) have significantly lower ε_{Nd} (T) values (< -6) and higher T_{DM} values

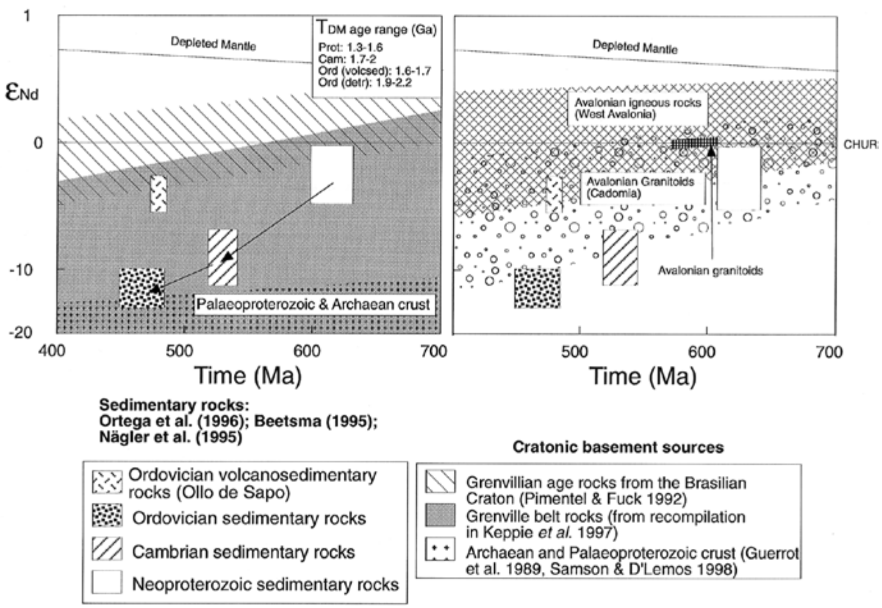

Fig. 6. Plot of ε_{Nd} (T) versus Age of deposition of Neoproterozoic and early Palaeozoic sedimentary rocks from NW Iberia (sources of analytical data: Beetsma, 1995; Nägler et al., 1995; Ortega, 1995). T_{DM} values shown as inset have been recalculated using the equation of Liew and Hofmann (1988). For comparison Nd envelopes of Grenvillian, Palaeoproterozoic, and Archaean crust are shown in addition to envelopes for Avalonian granitoids from Iberia (Fernández-Suárez et al., 1998), West Avalonia (Nance and Murphy, 1996; Keppie et al., 1997) and Cadomia (from compilation in Nance and Murphy, 1996).

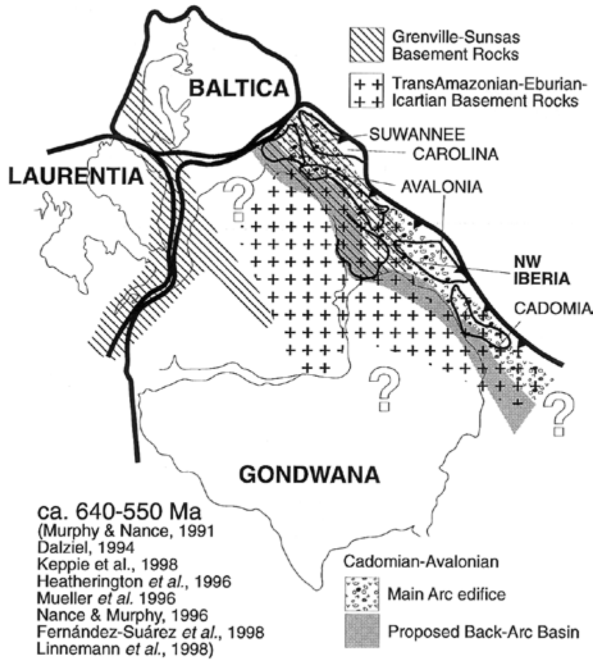

Fig. 7. Proposed paleogeographic reconstruction of Neoproterozoic peri-Gondwanan terranes (based on Nance and Murphy, 1996; Keppie et al., 1998). NW Iberia is considered to have occupied a peri-Amazonian location (see text for discussion) and was possibly situated further towards the continent with respect to West Avalonia.

(1.7 – 2 Ga, Fig. 6). This change in the Nd isotopic signature is an indication that the contribution of juvenile crust dwindled significantly in a relatively short time span. In the light of zircon ages, this is reflected in the almost complete lack of zircons in the age range 800 – 640 Ma in the Lower Cambrian sub-arkose (JG-16). Zircons with Cadomian/Avalonian ages in this sample fall in the age range 570 – 615 Ma, which matches the age range of the scarce Late Cadomian granitoids intruded into the Neoproterozoic sedimentary rocks, and have been interpreted to be generated through lower crustal melting (Fernández-Suárez et al., 1998).

Detrital Ordovician sedimentary rocks (Fig. 6) have even more negative s_{Nd} (T) values and higher average T_{DM} values than the Cambrian sedimentary rocks. This is consistent with a progressively waning contribution of Cadomian/Avalonian newly accreted crust. However, in Ordovician volcano-sedimentary greywackes (JG2, JG3) there is an increase of s_{Nd} (T) values and a lowering of T_{DM} values in relation to an addition of juvenile crust related to the 'Ollo de Sapo' magmatic event (ca. 480 – 460 Ma) (see also above). The U– Pb zircon ages in samples JG-2 and JG-3 are consistent with Nd isotopic signatures reported in similar rocks, which could result from a mixture of Palaeoproterozoic (ca. 2 Ga), Mesoproterozoic (ca. 1 Ga) Late-Cadomian (ca. 550 – 620) and Ordovician (Ollo de Sapo) juvenile components.

6.3. Paleogeographic implications

A significant implication of the ages obtained in this study concerns the location of NW Iberia in pre-Palaeozoic paleogeographic reconstructions. It is currently accepted that Avalonian/Cadomian circum-Atlantic terranes are dismembered fragments of an accretionary orogenic belt developed on an active margin of Gondwana (e.g. Nance and Murphy, 1996). This peripheral orogen was adjacent to cratonic provinces recording different events of continental assembly, and therefore the Cadomian-Avalonian terranes were most likely developed on different cratonic basements. On the basis of available geological, geochronologic and Nd isotopic signatures of these terranes it has been considered so far that West Avalonia, Carolina and Florida were situated in the periphery of the Amazonian craton during the Neoproterozoic, whereas Cadomia, Iberia and portions of East Avalonia are thought to have laid adjacent to (and evolved upon) the West African cratonic crust at that time (e.g. Kröner et al., 1988; Guerrot et al., 1989; Dalziel, 1992; Nance and Murphy, 1996; Keppie et al., 1998).

The presence of seemingly proximal Grenville-Sunsas/Rondonian age zircons in Neoproterozoic sedimentary rocks from NW Iberia argues against a peri-West African location during the Neoproterozoic, unless a yet unidentified Grenvillian event is also present within the West African craton. With the evidence available to date, the data reported in this study suggest a peri-Amazonian location of Iberia closer to West Avalonia, Florida and Carolina during the Neoproterozoic, as tentatively depicted in Fig. 7. In addition, the few discordant zircons with $^{207}Pb/^{206}Pb$ ages around 1.4 – 1.6 Ga (Table 1, Fig. 5) could have their provenance in the Rio Negro orogen of the Amazon craton (cf. Keppie et al., 1998), an orogenic event that is not recorded in the West African craton. Furthermore, the fact that Neoproterozoic NW Iberia mainly records back-arc basin sedimentation and relatively scarce magmatic activity (cf. Fernández-Suárez et al., 1998) suggests a continentwards location such as that depicted in Fig. 7. The hypothesis of a peri-Amazonian location of Iberia in Neoproterozoic times is strengthened by its position after undoing the Alpine and late/post Variscan rotations, which suggests a more north westerly paleo-location of this area with respect to Cadomia and East Avalonia.

Available Sm– Nd data on ca. 600 My old granitoids from the Narcea Antiform (Fig. 6) (ε_{Nd}� 0, T_{DM}� 1.2 Ga, Fernández-Suárez et al., 1998) are compatible with Sm– Nd signatures of Avalonian granitoids from terranes considered to have evolved upon Amazonian cratonic basement (cf. Nance and Murphy, 1996). The Nd isotopic data from those Avalonian/Cadomian granitoids of NW Iberia, taken together with detrital zircon ages (this study) suggest that NW Iberia might have evolved upon a basement containing a significant component of Grenvillian age crust and occupied a location close to South America during the Neoproterozoic where it would have had access to detrital zircons from the Amazon craton.

The Palaeoproterozoic (Eburnian-Icartian-Transamazonian) ages obtained (ca. 1.8 – 2 Ga) are consistent with the peri-Gondwanic nature of Iberia and the other terranes mentioned above since ca. 1.9 – 2.1 Ga events are recorded in both the West African craton (Eburnian event) and the Amazon craton (Trans-Amazonian event). The data also confirm the involvement of Archaean crustal components in the Neoproterozoic and subsequent events in NW Iberia (cf. Guerrot et al., 1989; Gebauer, 1993). The apparent absence of zircons with $^{207}Pb/^{206}Pb$ ages older than ca. 2.8 Ga could also be an indication that NW Iberia did not occupy a peri-west African location during the Neoproterozoic, as the Archaean crust in the West African craton records Archaean tectonothermal events older than ca. 2.7 Ga (Rocci et al., 1991).

6.4. Tentative scenario of Proterozoic-Early Palaeozoic in NW Iberia

Fig. 8 depicts the tentative scenario proposed for the Neoproterozoic-early Palaeozoic evolution of NW Iberia. The data reported here and available U– Pb and Nd isotope data from other areas point to the existence of a complex cratonic basement in the European Variscides including Archaean, Palaeoproterozoic and Mesoproterozoic components (Liew and Hofmann, 1988; Guerrot et al., 1989; Pin and Duthou, 1990; Turpin et al., 1990; Downes et al., 1990; Gebauer, 1993; Beetsma, 1995; Nägler et al., 1995; Samson and D'Lemos, 1998; Zeck and Whitehouse, 1999). However, actual contributions of the proposed crustal components to the basement of pre-Neoproterozoic Iberia and other areas of the European Variscides are poorly constrained. Our contention, based on the abundance of Grenvillian age zircons and model ages of Cadomian granitoids in NW Iberia is that the pre-Neoproterozoic basement might have had a significant Grenvillian crustal component.

The pre-Neoproterozoic basement (Fig. 8a) is taken as a starting point for the proposed scenario of Neoproterozoic-early Palaeozoic evolution of NW Iberia. Reconstructions of Gondwana during Late Proterozoic times show a peripheral arc edifice related to an oblique subduction event on its margin (Fig. 8b). The magmatic activity along this arc (Fig. 8b, c) is thought to have reached its climax between ca. 640 and 580 Ma (Murphy and Nance, 1991; Nance et al., 1991; Nance and Murphy, 1996; Keppie et al., 1998) although magmatic activity related to arc construction is considered to have started some 100 – 150 Ma before (Nance, 1990; Murphy and Nance, 1991; Egal et al., 1996; O'Brien et al., 1996). Back arc basins were developed during this subduction event (Nance et al., 1991; Murphy and Nance, 1991; Nance and Murphy, 1996; Fernández-Suárez et al., 1998) and the Neoproterozoic sediments are thought to have been deposited in these basins and to derive mainly from the arc (Murphy and Nance, 1989, 1991).

Fig. 8c depicts the continuation of subduction and arc-related magmatism down to ca. 560 Ma.

The back arc basin shown in this figure is considered to represent the locus of deposition for the Neoproterozoic sedimentary rocks of NW Iberia (Fernández-Suárez et al., 1998). It is proposed that these rocks contain detritus from both the main arc edifice and the margin of Gondwana. The presence of detrital components from the main arc edifice is represented by the ca. 640 – 800 Ma zircon population found in the Neoproterozoic rocks. This is in agreement with Nd isotopic signatures that indicate a significant component of newly accreted crust in these sedimentary rocks (see Section 6.2). The Grenvillian age zircons (0.9–1.27 Ga) may have reached this basin(s)

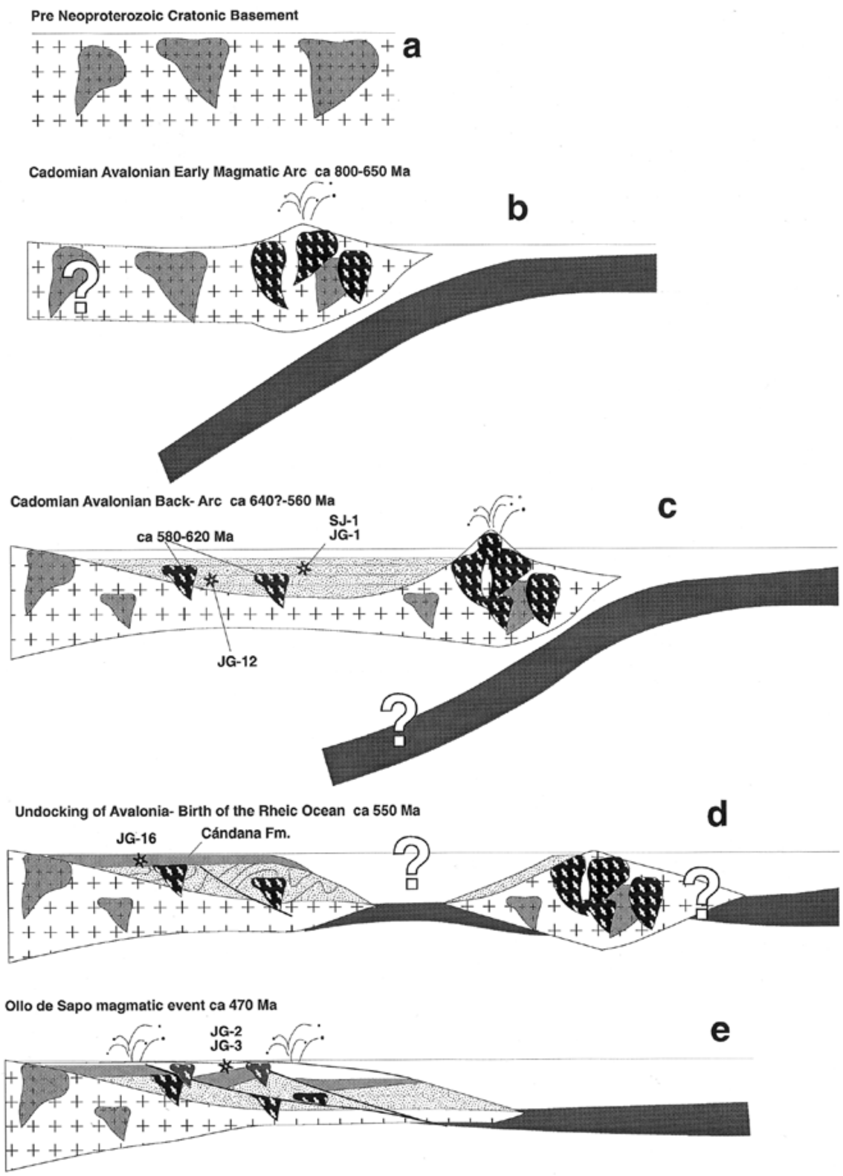

Fig. 8. Proposed schematic scenario for the Proterozoic-early Palaeozoic geological evolution of NW Iberia. See full discussion in text.

either from exposed fragments of the arc basement or from the continentwards cratonic source (see Section 6.3 above). The data suggest that the most likely cratonic source is the Sunsas/Rondonian crust found in the Amazonian craton of Gondwana. Older zircons (Lower Proterozoic and Archaean) would also have their source in their counterpart cratonic portions of Amazonia. The latter is in agreement with geochemical data suggesting the involvement of a stable cratonic source in the genesis of the Neoproterozoic sedimentary rocks from the Spanish Central Iberian Zone (Ugidos et al., 1997).

In NW Iberia, as in other areas of the Avalonian/Cadomian belt (cf. Murphy and Nance, 1991) geological evidence (e.g. lack of widespread deformation or significant crustal thickening) suggests that the termination of the subduction event was not due to final continent– continent collision. We concur with the views of Murphy and Nance (1989) who proposed that the subduction was replaced by transform/extensional activity related to the breakup of the late Precambrian supercontinent. Based upon the data from NW Iberia, our hypothesis is that following the arrest of subduction or a change in subduction regime (from ridge push to slab pull?), further extension of the basin resulted in the undocking of Avalonia (main arc edifice) from Gondwana, the opening of the Rheic ocean and eventually, the genesis of the Iberian-Armorican and Meguma platforms on both sides of the ocean (Fig. 8d) (see also Fortey et al., 1988; Soper, 1988). This event is reflected in the scarcity of ca. 640 – 800 Ma detrital zircons in the Lower Cambrian Cándana Formation (JG-16), that lies unconformably over the Neoproterozoic sequence. The Cándana Formation does contain abundant late Cadomian zircons (550 – 620 Ma) likely recycled through exhumation and erosion of the granitoids intruded into the Neoproterozoic sediments of NW Iberia (Fig. 8c, d). The Grenvillian age zircons and those with pre-Mesoproterozoic ages could have their source either in the cratonic areas of Gondwana or in the Neoproterozoic sedimentary rocks, which already contain these age populations. However, the morphological features of zircons suggest that the latter case is unlikely. Available Nd isotopic data (Fig. 6) suggest that the actual contribution of juvenile crust is not very significant in Cambrian rocks, in spite of the seeming abundance of ca. 550 – 620 Ma zircons (Fig. 4d).

In the Ordovician, it is proposed that further extension of an already thinned crust (Fig. 8e) gave rise to the magmatic Ollo de Sapo event (ca. 460 – 490 Ma) in Iberia, likely to have been triggered by mantle upwelling and upper mantle-lower crustal melting in a progressively thinning lithosphere (Fernández-Suárez et al., 1998). The genesis of these rocks in an extensional intracontinental setting is consistent with the within-plate trace element signatures of coeval alkalic basaltic rocks (Loeschke and Zeidler, 1982; Valverde-Vaquero, 1992) and the presence of Early Ordovician peralkaline granitoids in the margin of Gondwana (Van Calsteren et al., 1979). In the Ollo de Sapo greywackes, the juvenile volcanic component is reflected both directly by the 440 – 460 Ma zircon population and higher s_{Nd} values and younger T_{DM} ages with respect to Ordovician detrital (no volcanic components) sediments. As stated earlier, these samples contain Cadomian, Grenvillian and Palaeoproterozoic zircons from a proximal Gondwanan source. In NW Iberia, the oldest T_{DM} ages and maximum spread of T_{DM} values in sedimentary rocks correspond to an age of ca. 440 – 470 Ma (i.e. overlapping and following the Ollo de Sapo event). The combined T_{DM} and s_{Nd} spread (Beetsma, 1995; Nägler et al., 1995) reflect both an increase in the contribution from older crustal components and the addition of newly accreted crust. We suggest that this increase in addition of older material could be related to the genesis of the OS rocks in a crust undergoing progressive extension (see above). A progressively thinning crust is a likely scenario that would explain the simultaneous exhumation of lower crustal material and the generation of coeval magmatism. This latter scenario is consistent with models proposed for other circum-North Atlantic Avalonian/Cadomian terranes where repeated episodes of melting occurred in response to subduction and subsequent rifting events (Murphy et al., 1996; Mancuso et al., 1996). This scenario is also in agreement with current models for the subsequent evolution of Iberia from Silurian to

Late Carboniferous times (e.g. Martínez Catalán et al., 1997).

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