U-Pb ages of detrital zircons from the Basal allochthonous units of NW Iberia: Provenance and paleoposition on the northern margin of Gondwana during the Neoproterozoic and Paleozoic

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ABSTRACT

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LA-ICP-MS U-Pb ages of detrital zircons from eight siliciclastic samples from the Basal units of the Variscan allochthonous complexes of NW Iberia are used to establish the maximum depositional age and provenance of two tectonically-stacked metasedimentary sequences deposited on the outermost margin of Gondwana, and subsequently involved in the Rheic Ocean suture. The maximum depositional ages for the two sequences is latest Neoproterozoic and latest Cambrian, respectively. The age spectra are also used to discuss the paleoposition of the NW Iberian basement on the continental margin of Gondwana prior to the opening of the Rheic Ocean, which is tentatively placed in northern Africa, between the West African and Saharan cratons. Based on similarities and differences with age data from the NW Iberian autochthon and other allochthonous terranes involved in the Rheic suture, the relative proportions of Mesoproterozoic zircons in both assemblages are proposed as markers of proximity to the eastern part of the West African craton during late Neoproterozoic and late Cambrian. The geodynamic processes that took place along this part of Gondwana during the late Neoproterozoic, late Cambrian and Early Ordovician are discussed in the light of the LA-ICP-MS results, as well as the sedimentological record, magmatic evolution and plate tectonic setting of NW Iberia. These processes are linked to late Neoproterozoic and Cambro-Ordovician subduction events beneath the northern Gondwanan margin.

1. Geological setting

The allochthonous complexes of NW Iberia are a stack of nappes thrust over the sequences of the Iberian autochthon (Fig. 1a), which formed part of the northern margin of Gondwana for the entire Paleozoic (Martínez Catalán et al., 2007, 2009).

Three groups of allochthonous units have been distinguished (Fig. 1b and c). The Upper units, on top of the nappe pile, are pieces of a Cambro-Ordovician ensialic island arc (Abati et al., 1999, 2003; Andonaegui et al., 2002; Santos et al., 2002) inferred to have been detached from northern Gondwana by the roll-back of the subducting slab of the Iapetus-Tornquist ocean, to leave a new oceanic realm, the Rheic Ocean, in its wake (Stampfl i and Borel, 2002; Stampfl i et al., 2002; Winchester et al., 2002; von Raumer et al., 2003; Abati et al., 2010).

Relicts of oceanic floor form the middle allochthonous units, known as the Ophiolitic units (Díaz García et al., 1999; Pin et al., 2002, 2006; Arenas et al., 2007; Sánchez Martínez et al., 2007) and possibly remnants of older oceanic domains (Sánchez Martínez et al., 2006; Sánchez Martínez, 2009).

The Basal units, at the base of the nappe stack, represent distal parts of the Gondwana continental margin. They preserve calc-alkaline igneous rocks roughly coeval with the Cambro-Ordovician suite of the Upper units (Abati et al., 2010), and experienced extension and rift-related magmatism during the Ordovician (Floor, 1966; Ribeiro and Floor, 1987; Pin et al., 1992), while the Rheic oceanic lithosphere was being created. Subsequently, the Basal units were subducted beneath Laurussia at the onset of Variscan collision (Arenas et al., 1995, 1997; Santos Zalduegui et al., 1995; Rodríguez et al., 2003; Abati et al., 2010), and exhumed by crustal-scale thrusting accompanied by recumbent folding and tectonic denudation during the Variscan Orogeny (Martínez Catalán et al., 1996, 1997; Díez Fernández and Martínez Catalán, 2009).

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An imbricate thrust sheet, composed of siliciclastic sedimentary and volcanic rocks known as the Parautochthon, Lower allochthon or Schistose domain, separates the allochthonous complexes from the autochthon. The latter consists of a thick metasedimentary sequence which, as with the Lower allochthon and the Basal units, was deposited along the northern margin of Gondwana. The autochthonous sedimentary sequences of NW Iberia record the late Neoproterozoic and early
Paleozoic evolution of the northern Gondwana margin, including the Avalonian–Cadomian active margin in the Neoproterozoic (Rodríguez Alonso et al., 2004), the development of a passive margin during the Cambrian, the Cambro-Ordovician opening of the Rheic Ocean, and the return to passive-margin conditions until the onset of the Variscan collision in late Devonian times (Martínez Catalán et al., 2007, 2009; von Raumer and Stampfl, 2008).

2. The stratigraphic succession of the Basal units

The Basal allochthonous units comprise two tectonically juxtaposed metasedimentary sequences (Fig. 1c), the Upper sequence representing a paleogeographic domain distinct from that of the Lower sequence. Both sequences were involved in the initial subduction that preceded Variscan collision, but the Lower sequence developed eclogite facies
metamorphism (Gil Ibarguchi and Ortega Gironés, 1985) whereas the Upper sequence reached only blueschist facies conditions (López Carmona et al., 2010). The upper limit of the Upper sequence is a partly reworked thrust, and the original thrust contact between both sequences was also reworked during the Variscan orogenic collapse (Gómez Barreiro et al., in review). However, differences between the two sequences occur not only in their metamorphic evolution but also in their lithostratigraphy.

2.1. The Lower sequence

The strong heterogeneity of deformation in this sequence during the subduction and exhumation processes allows the original sedimentary stratigraphy to be characterized in the less deformed domains. These are preserved as meso- to macro-scale boudins that frequently reach map-scale size. The Lower sedimentary sequence consists of a thick, monotonous pile of immature sandstones (greywackes) alternating with minor layers or lenses of pelites, graphitic schist, calc-silicate layers and quartzites (Fig. 2a). The sandstones preserve Bouma sequences, crossed bedding, erosive contacts and normal graded bedding. They are clast-supported sedimentary rocks containing angular fragments of feldspars (the most abundant clasts), quartz and detrital micas in a clay-rich matrix with carbonaceous material. Pelitic horizons are common among the sandstones, and carbonaceous matter within them may become so abundant as to form graphitic horizons.

The sandstones form sequences up to 50 m thick. The normal thickness of the layers ranges from less than 1 dm to more than 1 m and often occur as a monotonous alternation of thin-bedded sandstones. The thicker the layers, the more massive the internal structure. No conglomeratic levels have been found, but it is possible to find grains bigger than 2 mm in the less deformed domains.

The quartzites (Portela Quartzites of Marquinez Garda, 1984) occur relatively high in the sequence, and according to the restoration of Variscan structures made by Martínez Catalan et al. (1996), they lie in a paleogeographic position relatively close to the Gondwanan mainland. In the moderately to highly deformed and metamorphosed areas, the sequence consists of albite-bearing schists and paragneisses alternating with centimetre-thick lenses of mica schists, graphite-bearing schists, and calc-silicate lenses, in which the sedimentary layering can still be seen. Intruding the sediments are mafic dykes (alkali basalts; Marquinez Garda, 1984; Rodríguez, 2005), calc-alkaline granitoids (tonalites and granodiorites), and rift-related granitoids (alkaline to peralkaline and alkali-feldspar granites; Floor, 1966; Ribeiro and Floor, 1987; Pin et al., 1992), all transformed during the Variscan Orogeny into eclogites, amphibolites and orthogneisses (Rodriguez, 2005). The age of the oldest calc-alkaline orthogneisses (495–500 Ma; Abati et al., 2010) establishes a minimum depositional age for the Lower sequence.

The original thickness of the Lower sequence cannot be calculated because of the ductile deformation accompanying subduction, thrusting, recumbent folding and tectonic denudation. A minimum present thickness of 4 km can be estimated, but this value probably represents less than half of the original thickness.

2.2. The Upper sequence

No significant deformation partitioning is visible during the subduction-exhumation process in the Upper sequence and all rocks have been strongly deformed. The following description is based on the characteristics of the metamorphic rocks.

The Upper sequence consists of a thick, monotonous pile of mica schists alternating with minor lenses of amphibolites, graphitic schists, metacherts, calc-silicate lenses, greywackes and quartzites (Fig. 2b). A
3. Sample description

Eight samples of the most representative lithologies were collected (Fig. 1b), five in the Lower sequence (BA-4, BA-5, BA-6, BA-7 and BA-8) and three in the Upper sequence (BA-1, BA-9 and BA-12). Their position in the stratigraphic column is shown in Fig. 2.

Samples BA-4 to BA-8 are representative of the Lower sequence. BA-4, 5 and 6 are greywackes from the lower part of the sequence, BA-5 representing a more deformed and metamorphosed greywacke than the massive BA-4 and BA-6. Sample BA-7 corresponds to the quartzites in the upper part of the sequence, and BA-8 represents the more pelitic facies in the Lower sequence.

BA-4 is a mildly deformed metagreywacke sampled from a massive layer 40 cm thick, in which irregular clasts of feldspar, quartz and mica are surrounded by a partly recrystallized pelitic matrix. Small albite porphyroblasts include tiny crystals of garnet, white mica and chlorite, although their metamorphic growth does not disturb the massive internal structure of the sandstone. The detrital grains show undulose extinction. Their relationship to the Variscan deformation cannot be clearly established but assumed.

Sample BA-5 is an albite-bearing paragneiss with a crenulation cleavage folding a previous amphibolite facies fabric that represents the main foliation. The latter is related to exhumation during Variscan collision following early Variscan subduction (Martínez Catalán et al., 1996; Llana-Fuñe, 2001; Rodríguez, 2005). The paragneiss is composed of the stable assemblage quartz + white mica + biotite + albite ± garnet ± ilmenite. Albite porphyroblasts include an internal schistosity defined by garnet + phengite + garnet ± rutile ± epidote, which represents a high-pressure, medium temperature assemblage that records the initial subduction event (Gil Ibaruguchi and Ortega Girónés, 1985). The main foliation is bent by microfolds with vertical axial planes and gently plunging axes associated with the development of a late crenulation cleavage defined by the reorientation of the plagioclase, biotite, quartz, and white mica of the main foliation, and the growth of new biotite, white mica, and quartz defining a tectonic banding.

BA-6 is a massive metagreywacke transformed into an albite-bearing paragneiss with a poorly developed planar foliation defined by quartz, biotite, and white mica. The main foliation is oblique to the compositional layering, which can be recognized at the layer boundary where the sample was collected. The albite porphyroblasts also include the same high-pressure assemblage described in BA-5.

BA-7 is a fine-grained quartzite with small amounts of white mica. The main foliation is formed by the statistical orientation of quartz grains, micas and minor quantities of opaque minerals. Quartz grains show internal deformation features, but the white mica does not define a true tectonic banding. The sample was collected from a homogeneous layer surrounded by mica-rich quartzites.

BA-8 is an albite-bearing mica schist with a foliation typical of the chlorite-biotite zone, composed of chlorite + quartz + albite + white mica ± biotite ± opaque minerals. The albite porphyroblasts include an internal fabric with folded patterns strongly oblique to the external foliation. The latter is formed by phengite, chlorite, quartz, rutile and rare garnet, and is comparable to the high-pressure internal foliation described in other samples (see BA-5).

The samples from the Upper sequence are pelitic rocks (BA-1 and BA-12) that occasionally include more arenaceous horizons (BA-9). The three samples are garnet-bearing mica schists with a main foliation whose stable mineral assemblage varies. For BA-1, the foliation is defined by garnet + white mica + biotite + albite + quartz ± ilmenite, whereas in BA-9 and BA-12, it consists of garnet + quartz + white mica + albite + chloritoid + epidote ± rutile. In all of the samples the albite porphyroblasts contain an internal fabric defined by garnet + phengite + rutile + quartz ± glaucophane, which is also preserved as helicitic or straight inclusions in porphyroblastic garnet (BA-9 and BA-12) and chloritoid (BA-9). C' extensional shear bands composed of quartz ± white mica ± chlorite crosscut and retrograde the main foliation in all samples. The main foliation in each sample represents a different stage of exhumation from blueschist facies (preserved in the internal fabric) to greenschist facies conditions (extensional shear.
4. Analytical methodology: U–Pb zircon dating

Zircon was recovered by the usual procedure of separation of heavy minerals: crushing, sieving and concentration of the heavy fraction using a Wilfley table, followed by magnetic and density separation. The final mineral fractions were hand-picked under a binocular microscope, mounted in Epoxy resin, and polished to an equatorial section of the grains.

All samples contain subidiomorphic zircons with rounded rims and variable shape and size. Their colour varies from clear and colourless to pinkish and almost opaque. The poorest sorting and less rounded shapes are in the greywackes, whereas grains from the mica schists and quartzites are more homogeneous, of smaller size (smaller fractionation and instrumental mass discrimination were corrected by normalization to the reference zircon GJ-1 (Jackson et al., 2004)). Prior to this normalization, the drift in inter-elemental fractionation (Pb/U) during 30 s sample ablation was corrected for the individual analysis.

The correction was done by applying a linear regression through all measured ratios, excluding the outliers (± 2 standard deviation; 2SD), and using the intercept with the y-axis as the initial ratio. The total offset of the measured drift-corrected $^{206}\text{Pb}/^{238}\text{U}$ ratio from the “true” IDTIMS value of the analyzed GJ-1 grain was typically around 3–9%. Reported uncertainties (2σ) of $^{206}\text{Pb}/^{238}\text{U}$ were propagated by the quadratic addition of the external reproducibility (2SD) obtained from the standard zircon GJ-1 ($n = 12$; 2 SD = 3–13%) during the analytical sequence (55 unknowns plus 12 GJ-1) and the within-run precision of each analysis (2 SE; standard error).

The external reproducibility of the $^{207}\text{Pb}/^{206}\text{Pb}$ (GJ-1) was about 0.9% (2 SD). However, as $^{207}\text{Pb}/^{206}\text{Pb}$ uncertainty during LA-SF-ICP-MS analysis is directly dependent on $^{206}\text{Pb}$ signal strength, uncertainties were propagated following Gerdes and Zeh (2009). The $^{235}\text{U}$ was calculated from the $^{238}\text{U}$ divided by 137.88 and the $^{207}\text{Pb}/^{206}\text{Pb}$ uncertainty by the quadratic addition of the $^{206}\text{Pb}/^{238}\text{U}$ and the $^{207}\text{Pb}/^{206}\text{Pb}$ uncertainty.

5. Age spectra

Only concordant or nearly concordant (<10% discordant) data were considered for interpretation of detrital zircon age. Fig. 3 includes all the U–Pb concordia diagrams showing the results of LA-ICP-MS dating with 2σ errors for the ellipses. The enlarged regions show the Paleozoic and Neoproterozoic zircons. Ratios and ages (in bold) for the selected analyses are given in Tables 1–8 (supplementary item in data repository). Ages younger than 1 Ga are reported based on $^{206}\text{Pb}/^{212}\text{Pb}$ ratios corrected for common Pb. Older ages are reported based on their $^{206}\text{Pb}$-corrected $^{206}\text{Pb}/^{206}\text{Pb}$ isotopic ratio. Figs. 4 and 5 include all the probability and frequency diagrams made for each sample. As the samples belonging to each sedimentary sequence do not show significant differences, they have been integrated in single diagrams plotted for each representative lithology (plots with a light-grey background).

5.1. The greywackes of the Lower sequence

Samples BA-4, BA-5, BA-6 and BA-8 represent the greywackes of the Lower sequence and have been plotted in Fig. 4a, b, c, and d. The main group of zircons has Neoproterozoic $^{206}\text{Pb}/^{238}\text{U}$ ages between 540 and 750 Ma, with the maximum density around 650 Ma. The second group includes Paleoproterozoic $^{206}\text{Pb}/^{238}\text{U}$ ages ranging between 1850 and 2250 Ma. Three relative maxima can be identified in this group. The first is around 1950 Ma, the second at 2050 Ma, and the third around 2150–2200 Ma. A widespread population of analyses with Archean ages represents the third group. It ranges between 2500 and 3500 Ma with two relative maxima at 2700 and 2850 Ma. A few clusters of Mesoproterozoic zircons also occur. Their ages form part of a continuous interval from 750 Ma (Neoproterozoic) to 1250 Ma, with several analyses around 1500 and 1600 Ma, but with a maximum abundance located between 800 and 1100 Ma.

The youngest zircon found yielded an age of 537 ± 17 Ma (98% concordance) and the youngest population age is 566 Ma old. The oldest zircon yielded an age of 3522 ± 11 Ma (101% concordance).

5.2. The quartzites of the Lower sequence

Sample BA-7 represents the quartzites of the Lower sequence and has been plotted in Fig. 4e. The main group is represented by 70 analyses with Neoproterozoic $^{206}\text{Pb}/^{238}\text{U}$ ages ranging between 551 and 993 Ma, with the maximum density around 650 Ma. The second group includes 58 analyses with Paleoproterozoic $^{206}\text{Pb}/^{238}\text{U}$ ages ranging between 1650 and 2450 Ma. Three main sub-maxima can be identified in this group. The first is around 1930 Ma, the second...
Fig. 3. U-Pb concordia diagrams showing the results of LA-ICP-MS dating of detrital zircons for all the samples. Error ellipses represent 2σ uncertainties. Phanerozoic and Neoproterozoic ages have been enlarged for clarity.
5.3. The sediments of the Upper sequence

Probability and frequency diagrams of samples BA-1, BA-9 and BA-12 appear in Fig. 5a, b, and c, and all ages have been plotted together in Fig. 5d. The main group of zircons is represented by the analyses with Neoproterozoic $^{206}\text{Pb}/^{238}\text{U}$ ages between 500 and 750 Ma, with the maximum density around 650 Ma. The second group includes analyses with Paleoproterozoic $^{207}\text{Pb}/^{206}\text{Pb}$ ages between 1850 and 2200 Ma. Three different relative maxima can be identified in this group: 1950 Ma, 2050 Ma, and 2150-2200 Ma. A widespread population of analyses with Archean ages ranging between 2500 and 3500 Ma represents the third group. A few clusters of Neo- to Mesoproterozoic zircons occur that complete a continuous interval from 800 Ma (Neoproterozoic) to 1250 Ma. The maximum input ranges between 900 and 1050 Ma. A few analyses around 1400 and 1600 Ma also occur.

The youngest zircon has an age of 497 ± 22 Ma (93% concordance) and the youngest population age is 512 Ma. The oldest zircon is 3537 ± 14 Ma old (102% concordance).

6. Data integration and geological implications

6.1. Ages and geodynamic setting

The zircon age populations of the Basal allochthonous units studied here can be easily correlated with those of the autochthonous sequences of NW Iberia published by Fernández-Suárez et al. (2002a,b), Gutiérrez-Alonso et al. (2003), and Martínez Catalán et al. (2004b, 2008), which suggests that the Basal units are only moderately allochthonous, representing far-travelled domains but not exotic terranes. Our data support the idea that the Basal allochthonous units remained attached to the northern margin of Gondwana during the opening of the Rheic Ocean. This would explain why their subduction that occurred in the Late Devonian was the first Variscan deformation to affect this part of the outermost margin of Gondwana.

Our study also places tight constraints on the age of sedimentation. Given the statistical uncertainty of a single analysis, the youngest zircon population would represent a good approximation of the maximum age of sedimentation. Since the Lower sequence is intruded by late Cambrian calc-alkaline granitoids (ca. 495 Ma; Abati et al., 2010), and its youngest detrital zircon population is 566 Ma for the greywackes and 557 Ma for the quartzites, deposition is bracketed between the Neoproterozoic and the middle Cambrian. The Lower sequence represents deposition in a tectonically active setting, so the youngest population likely represents a good approximation of the age of sedimentation. Given the high content of Paleoproterozoic and Archean ages in all the samples from the Lower sequence (Fig. 4e and f), we suggest that the quartzites and metagreywackes were derived from a more cratonic source than the sediments of the Upper sequence. This input likely reflects not only the cratonic influence of a passive margin, but also the denudation of an arc built upon an old continental crust.

650 Ma is a common maximum for zircon age populations in sediments related to the Avalonian-Cadomian active margin along northern Gondwana. This maximum occurs in sediments that range in age from 545 to 570 Ma (Linnemann et al., 2004), which is consistent with the age proposed for the greywackes. Accordingly, we propose a late Neoproterozoic age for the sediments of the Lower sequence. They were probably laid down during the late pulses of the Avalonian-Cadomian arc-system, either in a back arc or retroarc setting (Murphy and Nance, 1991; Fernández-Suárez et al., 2000; Murphy et al., 2002; Pereira et al., 2006; Linnemann et al., 2007, 2008). In this scenario, the quartzites could represent the passive margin on the Gondwanan side of the arc (Fig. 6a).

The Upper sequence represents deposition that either pre-dates or is caused by the opening of the Rheic Ocean, as constrained by ca. 497 Ma granitoids intruded into the back-arc related ophiolite of Vila de Cruces (Arenas et al., 2007). The age of the youngest population (512 Ma) and the youngest zircons (around 500 Ma) in the Upper sequence coincide with one of the two main magmatic episodes registered in the Basal and Upper allochthonous units. The abundance of middle to late Cambrian ages in this sequence (Fig. 5d) suggests a direct linkage with the arc-related event, rather than the rift-related event dated as Early Ordovician (Rodríguez et al., 2007).

There is also a significant population of Avalonian–Cadomian zircons in the Upper sequence, which is the main population in the late Neoproterozoic greywackes of the Lower sequence. Erosion of the Cadomian arc-system would supply the same age population in both the Ediacaran and early Paleozoic. However, recycling of the Cadomian basement can also be envisaged, since Cambro-Ordovician magmatism is widespread in the Iberian autochthon and the Lower sequence of the Basal allochthonous units (Valverde-Vaquero and Dunning, 2000; Beá et al., 2006; Montero et al., 2007; Díez Montes et al., 2010).

The volcanic input in the Upper sequence can be placed into an extensional context, in which N-MORB basalts would have been incorporated during back-arc spreading (Fig. 6b). This interpretation is consistent with the sedimentological record, since the volcanic rocks were emplaced at the time of basin development and widening. This setting is considered to represent the first stage in the evolution of the Rheic Ocean by Sánchez Martínez (2009).

At the same time, flyschoid deposits with a large input of Cambrian zircons were accumulating close to the drifting arc (Fernández-Suárez et al., 2003; Fuenlabrada et al., 2010). These sediments are preserved in the Upper allochthonous units of the European Variscides, and may have been located along the more active side of the back-arc basin, attached to the arc-system. We consider them to be the proximal facies of the Cambrian arc, whereas the Upper sequence of the Basal allochthonous units represents contemporaneous distal deposits along the Gondwanan side of the basin during the middle to late Cambrian (Fig. 5b). Even if they are related to an active Cambrian setting, the flyschoid deposits of the Upper allochthonous units still include a remarkable number of cratonic Paleoproterozoic and Archean ages, as well as the Avalonian–Cadomian signature, which points to the ensialic character of the arc.

6.2. Paleogeographic constraints

No significant differences in cratonic input have been found in the sequences of the two tectonic units. They show the same dominant age populations and even the same minor peaks in the Paleoproterozoic input, although the latter are more pronounced in the older sediments of the Lower sequence. Furthermore, no differences appear to exist between the age populations of the two units and those of the NW Iberian autochthon. However, the relative proportions of early Neoproterozoic and late Mesoproterozoic zircons (corresponding to the 750–1150 Ma interval) are different. Whereas in the Basal units they represent 2–8% of the concordant analyses (Figs. 4 and 5), in late Ordovician, early Silurian and early Devonian quartzites of the autochthon, they amount to 24–38% (Martínez Catalán et al., 2004b). Similar abundances also occur in early Ordovician quartzites and greywackes, early Cambrian quartzites and late Neoproterozoic sandstones (Fernández-Suárez et al., 2000, 2002a; Gutiérrez-Alonso et al., 2003).
Neoproterozoic, Paleoproterozoic and Archean ages in both the Basal allochthonous units and the NW Iberian autochthon strongly suggest a West African craton provenance for most of the zircons (Nance and Murphy, 1994). But the middle Neoproterozoic to Mesoproterozoic signature (750-1150 Ma) is absent in the West African craton (Rocci et al., 1991; Boher et al., 1992; Potel et al., 1998; Hirdes and Davis, 2002; Guye et al., 2007). The two closer source areas identified to date for the Mesoproterozoic zircons are the couple formed by the Saharan craton and the Arabian–Nubian shield (Loizenbauer et al., 2001; Abdelsalam et al., 2002; Avigad et al., 2003, 2007; de Wit et al., 2005; Stern, 2008; Be’eri-Shlevin et al., 2009) in the eastern branch of the West African craton, and the Amazonian craton to the west (Bermasconi, 1987; Santos et al., 2000; Cordani and Teixeira, 2007).

The early Neoproterozoic (Tonian) and late Mesoproterozoic (Stenian; Walker and Geissman, 2009) record in the Basal units and the autochthon shows a continuous presence of zircon ages in the interval 750-1250 Ma (Figs. 3-5). The largest population occurs between 800 and 1100 Ma (Figs. 4e, f, and 5d), followed by a subordinate population between 1500 and 1650 Ma. Fig. 7 shows a simplified map of Gondwana at ca. 570 Ma with the main cratonic areas (and their characteristic isotopic ages) that may have acted as sources of detrital zircons for the northern Gondwana margin, according to Linnemann et al. (2007), and the main active zones during the late Neoproterozoic which could control the main detritus influx, including the Avalonian–Cadomian belt (Nance and Murphy, 1994) and the Trans-Brasiliano-Hoggar megasuture (Cordani and Teixeira, 2007).

Since NW Iberia cannot be placed far from either the West African craton or a Mesoproterozoic source, two possibilities arise for its paleoposition during the late Neoproterozoic to late Cambrian. The entire record of age populations roughly fits a paleoposition either between the eastern West African craton and the Saharan craton/Arabian–Nubian shield, or between West African craton and Amazonia (Friedl et al., 2000; Fernández-Suárez et al., 2002b; see also Mellet et al., 2010). Several arguments point to the first option, as discussed below.

Ages ranging 750-900 Ma, which represent most of the early Neoproterozoic zircons in our samples, are absent in the Amazonian craton, but exist along its eastern rim (Santos et al., 2000; Cordani and Teixeira, 2007), and also in the western rim of the Saharan craton (Abdelsalam et al., 2002 and references therein). These ages can be used as a guide to place the NW Iberian autochthon, because during that time interval, widespread magmatic and tectonic activities occurred in Amazonia (Pimentel et al., 1997; Cordani and Teixeira, 2007, and references therein) and Africa (Caby, 2003, and references therein), linked to the evolution of the Pampéan–Goiás–Pharusian oceanic lithosphere (Cordani et al., 2003; Kröner and Cordani, 2003). This activity included extensive calc-alkaline magmatism corresponding to tectonomorphic episodes of the Pan-African/Brasiliano orogenic cycle. The 750-900 Ma ages provide tight constrains on the paleoposition of sediments deposited in the northern margin of Gondwana, since the peri-Gondwana and Pan-African/Brasiliano provenances strongly suggest a position between the West African and Saharan cratons (Fig. 7).

An overview of the timing and building processes in the Amazonian craton compiled by Cordani and Teixeira (2007) and Cordani et al. (2009) provides the age spectra that can be expected from its erosion and/or recycling. It includes, in the Rondônia–San Ignacio Province, an important source of zircons ranging 1300-1500 Ma, either derived from magmatic or metamorphic events. This interval is represented by extremely scarce zircon populations in our samples, which is difficult to explain if an Amazonian provenance is considered (Keppie et al., 2003) for the paleoposition of NW Iberia during the late Neoproterozoic, as suggested by Fernández-Suárez et al. (2002b) and Nance et al. (2008). Alternatively, potential African sources for the 1300-1500 Ma populations also exist in the Saharan craton (Abdelsalam et al., 2002).

Geochronology of late Neoproterozoic siliciclastic rocks in an area of the NW Iberian autochthon suggests a homogeneous and recycled source, and favours a West Africa craton provenance (Ugidos et al., 2003a). Sm–Nd data from these sediments suggest a contribution of juvenile material, much younger than 1 Ga, and probably derived from pan-African orogens (650–700–900 Ma; Ugidos et al., 2003b).

Detrital micas from Cambrian sediments in the NW Iberian autochthon include the following populations: 550–640 Ma, c. 920–1060 Ma and 1580–1780 Ma (Gutiérrez-Alonso et al., 2005). The 550–640 Ma interval fits quite well in the Avalonian–Cadomian setting, whereas the two others fit both the African (Saharan) and Amazonian provenance. The presence of Mesoproterozoic detrital micas indicates the existence of rocks formed at that age that were eroded during the Cambrian. These ages, together with detrital zircon ages and geochemical data, have been interpreted as an evidence of a Mesoproterozoic basement occurring in the core of the Iberian-armorican arc (Gutiérrez-Alonso et al., 2005; Murphety et al., 2008). However, such interpretation is not supported by the inherited zircon ages of widespread Cambro-Ordovician granitoids and felsic volcanics intruded and erupted in the Iberian autochthon, Lower Allochthon and Basal units, where Mesoproterozoic ages are statistically meaningless. In fact, the zircon ages rather suggest a magmatic source mostly consisting of Neoproterozoic rocks (Valverde-Vaquero et al., 2005; Bea et al., 2007; Montero et al., 2007, 2009; Castiñeiras et al., 2008a,b; Díez Montes et al., 2010; Abati et al., 2010).

No late Neoproterozoic sutures or dextral mega-shear zones affecting the rim of northern Gondwana have been identified so far either in Iberia or the European Variscides. The models approximating Amazonia-derived terranes to northern African domains by along-strike movements rely only in the interpretation of Mesoproterozoic populations as derived from the Amazonian craton.

In short, we believe that the Mesoproterozoic ages interpreted as of Amazonian derivation in previous works are best explained by northern African sources in the light of new data in both South America and Africa (Santos et al., 2000; Loizenbauer et al., 2001; Abdelsalam et al., 2002; Avigad et al., 2003, 2007; de Wit et al., 2005; Cordani and Teixeira, 2007; Cordani et al., 2009; Emnén and Liégeois, 2008; Be’eri-Shlevin et al., 2009). The new sources support the building of the Cadomian belts by recycling of Paleoproterozoic basement, and provide a new role for the pan-African mobile zones involved in Gondwana assembly. For instance, the Trans-Brasiliano-Hoggar megasuture occurring along the eastern border of the West African and Amazonian cratons and the western border of the Saharan craton (Caby, 2003; Caby and Moïné, 2003; Emnén and Liégeois, 2008; Santos et al., 2008) could be the source of Mesoproterozoic zircons ranging 900–530 Ma.

Assuming an African provenance, the relative proportions of Mesoproterozoic zircons in the two sequences analyzed, compared to those of the NW Iberian autochthon suggest that the Basal allochthonous units occupied a more western position, closer to the West African craton. Conversely, the Upper allochthonous units of NW Iberia are characterized by the absence of zircon age populations between the Ediacaran and the Paleoproterozoic (Fernández-Suárez et al., 2003), which suggests a more western derivation for this exotic terrane, the zircon ages record of which appears to have been derived almost exclusively from the West African craton (Fig. 7).

Fig. 4. Frequency (bars) and probability density distribution (curves) of ages, and relative abundance of significant age populations of detrital zircon grains of metagreywackes and quartzites from the lower sequence. Plot (f) with a light-grey background, represents a cumulative diagram made by merging the data from all the greywackes of the lower sequence. The quartiles (e) have also been shaded. Only concordant or subconcordant analyses (< 10% discordant) have been used for interpretation. n: number of analyses with < 10% discordance/total number of analyzed grains; conc.: concordance.
Fig. 5. Frequency (bars) and probability density distribution (curves) of ages, and relative abundance of significant age populations of detrital zircon grains of metapelites from the Upper sequence. Plot (d), with a light-grey background, represents a cumulative diagram made by merging the data from all the metapelites of the Upper sequence. Only concordant or subconcordant analyses (<10% discordant) have been used for interpretation. n: number of analyses with < 10% discordance/total number of analyzed grains; cone.: concordance.
Late Neoproterozoic
Cadomian-Avalonian arc dismantlement
Passive margin?

Middle-Late Cambrian
Future opening of the Rheic Ocean

7. Conclusions
Two different metasedimentary sequences have been identified in the Basal Allochthonous units of NW Iberia. The sequences occur in thrust sheets whose boundaries were reworked during late Variscan extensional collapse, and have been mapped in the allochthonous complexes of Ordones and Malpica-Tui. The Lower sequence is characterized by turbiditic facies, thick greywacke horizons, and relative proximity to the source area, probably the Neoproterozoic Avalonian-Cadomian arc. The Upper sequence is essentially pelitic with alternations of amphibolites with N-MORB affinity, and represents a more distal paleoenvironment in the back-arc basin of a Cambro-Ordovician volcanic arc.

Detrital zircons from five samples of the Lower sequence and three samples of the Upper sequence have yielded late Ediacaran and late Cambrian maximum depositional ages, respectively. The zircon age populations show no significant differences in the cratonic input of both sequences, although the cratonic influence is more pronounced in the Lower sequence. The age spectra are similar to published detrital zircon ages from the NW Iberian autochthon. This fact establishes the Basal allochthonous units as the outermost recognizable pieces of Gondwanan continental crust, and supports the idea that the early Paleozoic northern Gondwanan platform was extremely wide.

The zircon age populations of the Basal units have been used to locate the paleoposition of NW Iberia along the northern margin of Gondwana.
A position to the north of the West African craton is the most reasonable option, because (i) the sources of Neoproterozoic ages are no longer exclusive of the Amazonian craton according to recent data, (ii) the tracers and implications derived for the Amazonian basement in NW Iberia are less clear than those for African, and (iii) the 750-900 Ma age population gives better constrains on tectonomagmatic processes affecting its eastern rim and the western rim of the Saharan craton than a South American source for the early Neoproterozoic zircons. This correlation does not require large-scale along-strike movements during the late Neoproterozoic to late Cambrian at the periphery of northern Gondwana, and gives a still poorly understood role to other pan-African sutures in the European Variscan terranes. A widely accepted dextral component of convergence between Laurussia and Gondwana during the closure of the Rheic Ocean may have joined realms that were once in lateral continuity and showed small differences in their provenance along the northern margin of Gondwana.

The sedimentological features of the two sequences described, their age, and the timing of magmatic events, have been integrated into a geodynamic model linked to the long-lived history of Neoproterozoic and Cambro-Ordovician subduction beneath northern Gondwana. The model involves an Avalonian–Cadomian active margin during the late Ediacaran, and a new arc that developed during the middle to late Cambrian with a back-arc basin between it and the newly formed passive margin of Gondwana. Sea floor spreading in the back arc gave birth to the Rheic Ocean around the Cambrian–Ordovician boundary.

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Appendix A. Supplementary data

