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ORIGINAL PAPER

SHRIMP U–Pb geochronological constraints on the timing of the intra-Alcudian (Cadomian) angular unconformity in the Central Iberian Zone (Iberian Massif, Spain)

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Abstract New SHRIMP U–Pb ages of detrital zircons from Neoproterozoic low-grade metasandstones of the Schist–Graywacke Complex (Central Iberian Zone, Iberian Massif) sampled just below and above the intra-Alcudian unconformity at two selected locations contribute to reconstruct the geodynamic evolution of Iberia during the Cadomian orogeny in the north Gondwana margin. The youngest zircons (i.e., maximum depositional age) in the Lower Alcudian are c. 580–576 Ma, while those in the Upper Alcudian are c. 555–552 Ma. The obtained remarkable time gap of about 21 Ma supports the existence of a tectonic event in between. This event resulted in moderate folding (without related foliation/metamorphism) that verticalized the Lower Alcudian previous to the deposition of the Upper Alcudian. Additional evidence of late Cadomian tectonothermal events elsewhere in Iberia also fit in the interval c. 560–550 Ma. Combined with other geological data, the most probable maximum depositional ages are c. 580–560 Ma for the Lower Alcudian (previous to the late Cadomian folding event) and c. 550–540 Ma for the Upper Alcudian (previous to the deposition of the overlying Pusian Group and Lower Cambrian sandstones and limestones). A comparison of the new zircon age spectra with possible source areas verifies recent studies that point to the Cadomian foreland in the north Gondwana continent affected by the Pan-African orogeny: the West African Craton and/or the Saharan Metacraton. Furthermore, ongoing Cadomian arc-related magmatism in Iberia (c. 605–545 Ma) could have contributed as a local zircon source. The end of the Cadomian activity is marked by a transient stage (ephemeral Lower Cambrian platform) which preceded widespread Cambro–Ordovician rifting of north Gondwana.

Keywords Detrital zircon · Alcudian unconformity · Central Iberian Zone · SHRIMP U–Pb geochronology

Introduction

Knowledge of the timing of tectonic deformation events is a key point to understand the ongoing orogeny. The dating of tectonometamorphic events and related igneous rocks from exhumed internal domains of recent to ancient orogens is the best procedure to establish the chronology of the different phases of the orogeny. Additional information can be indirectly extracted from the study of synorogenic sediments deposited in basins formed adjacent or in the interior of the orogenic mountain belts. Geochemical and detrital provenance studies together with relative dating of these sedimentary successions also give a first approximation to the tectonic history of an adjacent orogen. The progressive deformation of these basins may give way to discrete mass slides resulting in local strata truncations. Moreover, the recognition of angular unconformities at the basin scale is one of the most important chronological indicators, i.e., older strata deformed/tilted and truncated by overlying younger ones. Such unconformities imply a sedimentary time gap that includes a tectonic deformation event

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and basin emersion/erosion, followed by renewed subsidence, since it was first shown by James Hutton in 1787. Since then, the relative dating of such time gaps has been quite common in unconformities involving fossiliferous Phanerozoic sedimentary successions. On the contrary, the absolute dating is a more troublesome task. In some cases, the relative dating can be improved by radiometric ages of interbedded volcanic layers close to an unconformity (e.g., Gutiérrez-Alonso et al. 2007). More rarely, datable minerals incorporated in the unconformable surface constrain the age of uplift/emersion (e.g., volcanic zircons in bauxite layers, Wang et al. 2010). An alternative approach to constrain the sedimentary time gap in azoic successions, e.g., Precambrian, is by obtaining the absolute ages of detrital mineral populations of sediments from beds located just above and below the unconformable surface. This is best done by U–Pb precise dating of zircon grains. The obtained age patterns provide some information about the probable provenance sources (yet local or far field), and the youngest ages indicate the maximum depositional ages of the sediments.

In this paper, we present the SHRIMP U–Pb dating of detrital zircons from two successions separated by an angular unconformity (the 'intra-Alcudian' unconformity). This unconformity separates Neoproterozoic–Lower Cambrian sedimentary successions included in the Schist–Graywacke Complex (SGC) of the Central Iberian Zone (CIZ) (Iberian Massif, European Variscan belt) (Rodríguez Alonso et al. 2004a). These successions are poorly dated since their fossil content is reduced to a few remains of vague Ediacaran to Cambrian microfauna and ichnofossils. Our results contribute to the reconstruction of the geodynamic evolution of Iberia during the Cadomian (Neoproteozoic) orogeny that took place along the northern margin of the supercontinent Gondwana (Murphy and Nance 1991).

Geological setting

The CIZ is the innermost zone of the Iberian Massif, one of the most complete exposures of the European Variscan belt (Fig. 1b). It is mainly composed of Late Proterozoic to Paleozoic metasediments, felsic intrusive and extrusive pre-Variscan igneous rocks, and syn- or late kinematic Variscan granitoids (Martínez Catalán et al. 2004). The zone is divided into two domains, the small Ollo de Sapo Domain (OSD) to the north and the large SGC Domain (SGCD) to the south (Fig. 1b) (Martínez Catalán et al. 2004). The first is characterized by the presence of Cambro–Ordovician metavolcanic and metagranitic rocks (the Ollo de Sapo Formation). The second is characterized by the occurrence of the SGC (Carrington da Costa 1950).

The SGC is a thick (8–11 km), monotonous succession of alternating thin pelitic and psammitic beds with minor intercalations of other lithologies and is considered to be Ediacaran to Cambrian in age (Rodríguez Alonso et al. 2004a and references therein). It is exposed on ~70,000 km^2 in the central part of the Iberian Massif in the cores of large Variscan antiforms in central Portugal and west-central Spain. The SGC has been divided into a number of lithostratigraphic units depending on the studied sector and authors, but, broadly speaking, two main units, separated by an unconformity (see next section), can be recognized (Rodríguez Alonso et al. 2004a).

The *Lower Unit* of the SGC is a monotonous succession of shales and sandstones with minor intercalations of conglomerates, chaotic beds and few volcaniclastic horizons which appear in the upper part of the series. In Spain, it corresponds to the Domo Extremeño Group (western SGCD), which is correlated with the Lower Alcudian (eastern SGCD). In Portugal, the Lower Unit is considered to be represented by the Beiras Group. Its thickness is unknown because the lower boundary is not exposed. The succession has been interpreted as turbidites deposited in submarine fans, slopes and channels and has been considered to be Ediacaran in age based on rare microfossils, such as metazoans (Cloudina Carinata from Upper Ediacaran, Cortijo et al. 2010), ichnofossils (Vidal et al. 1994a; Jensen et al. 2007) and acritarchs (Vidal et al. 1994b).

The *Upper Unit* of the SGC is mainly composed of pelites with occasional beds of black lutites, conglomerates and sandstones, together with limestones and discontinuous olistostromic beds and intercalations of phosphates and volcanic and volcaniclastic rocks. Due to its more lithological variety, the Upper Unit has been divided into a collage of lithostratigraphic elements. Despite oversimplification, from bottom to top, these are the Upper Alcudian (interbedded shales and sandstones with intercalations of conglomerates and limestones) and the Pusian Group (shales with discontinuous megabreccias, olistostromes and chaotic deposits toward the base). The chaotic deposits are carbonatic, siliciclastic and volcanoclastic in composition and lay on an erosive surface that may rest in places directly onto the Lower Unit. The Upper Unit has a spatial and temporal evolution that is more complex than that of the Lower Unit and reflects deposition in turbiditic channel, mixed platform, siliciclastic platform and continental environments. The Upper Unit has been considered to include the Proterozoic–Cambrian transition, reaching the Lower Cambrian, based on the presence of occasional fossils such as acritarchs (Synsphaeridium and Micrhystridium, Díez Balda and Fournier Vinas 1981; Vidal et al. 1994b), shelly metazoan (Cloudina, Vidal et al. 1994b), small shelly fossils (Anabarella, Vidal et al. 1999), trilobites (Rebelo and Romano 1986) and trace fossils (García Hidalgo 1993).

Geochemical studies of the minor volcanic and volcaniclastic rocks in both units of the SGC suggest magmatic

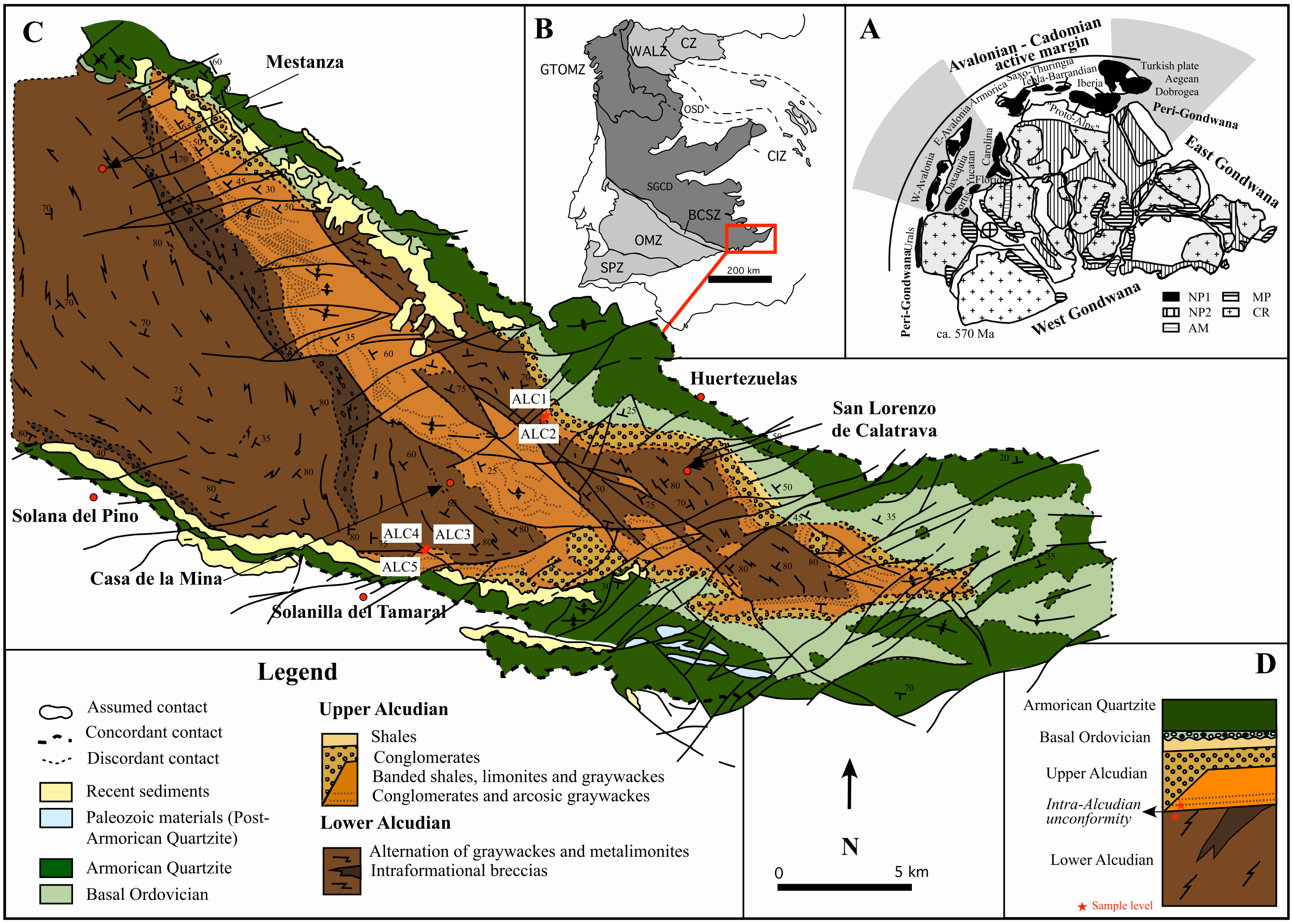

Fig. 1 Location of the samples: **a** schematic paleogeographical map of the Gondwana supercontinent at c. 570–560 Ma with possible paleoposition of Iberia. Modified from Linnemann et al. (2004) and references therein. Capital letters:*NP1* Neoproterozoic mobile belts of per-Gondwana (Cadomian and related events), *NP2* Neoproterozoic mobile belts of Gondwana (Pan-African and related events), *AM* 1.1–1.3 Ga Megashear event in Amazonia, *MP* Mesoproterozoic mobile belts (Grenville and related events), *CR* Cratons (Archean-Paleoproterozoic); **b** simplified map of the Iberian Massif modified from Pérez-Estaún et al. (2004). Capital letters: *CZ* Cantabrian Zone, *WALZ* West Asturian-Leonese Zone, *CIZ* Central Iberian Zone, *OSD* Ollo de Sapo Domain, *SGCD* Schist–Graywacke Complex Domain, *GTOMZ* Galicia Tras-os-Montes Zone, *BCSZ* Badajoz-Córdoba Shear Zone, *OMZ* Ossa-Morena Zone, *SPZ* South Portuguese Zone; **c** simplified geological map of the eastern part of Alcudian anticline (Pieren 2000); **d** schematic pre-Ordovician stratigraphy of the Schist–Graywacke Complex Domain in the Alcudian region

evolution in the same tectonic setting (Rodríguez Alonso et al. 2004b). They record two kinds of sources including (1) detrital material derived from recycled orogens and (2) a coeval Cadomian juvenile contribution that governs their isotopic signature (Rodríguez Alonso et al. 2004b). The results are indicative of an active geodynamic setting (the Cadomian orogeny at the northern margin of Gondwana with an associated magmatic arc and related basins) as a direct contributor to the synsedimentary and minor magmatic content of the Neoproterozoic–lowermost Cambrian successions in the CIZ (Rodríguez Alonso et al. 2004b). The chemical and isotopic data of the siliciclastic sediments of the SGC have led to controversial interpretations: as formed in synorogenic arc-related Cadomian basins (e.g., Nägler et al. 1995), or as deposited in a post-Cadomian, immature passive margin setting (Valladares et al. 2002 and references therein).

The precise dating and correlation of the SGC with other series is difficult due to the deformation (pre-Variscan discordances and Variscan folds and faults) and the scarcity of index fossils or reference strata. The microfossils and trace fossils are rare and suggest a Late Ediacaran age for the Lower Unit. In the Upper Unit, the age interval of these fossils includes the Precambrian–Cambrian limit (542 ± 1 Ma; Gradstein et al. 2004) and reaches the lowermost Cambrian. Besides paleontological studies, several U–Pb studies have been performed in the SGC over last decade. Ugidos et al. (2003) dated two sandstones using isotope dilution thermal ionization mass spectrometry (ID-TIMS) on multiple zircon grains. The U–Pb data were very discordant, indicating a complex geological history that prevented a straightforward geochronological interpretation. Despite the discordant data, these authors proposed a Gondwana (West African Craton) provenance for the sedimentary rocks based on their petrological, geochemical and isotopic similarities with sediments from other areas of the CIZ and Iberia. Gutiérrez-Alonso et al. (2003) dated a scant zircon population from three graywackes of the Upper Unit of the SGC, using laser ablation (LA-ICPMS). The few concordant ages reported provide no tighter constraint on the depositional age, since their youngest zircons grains yielded a Cryogenian age. These authors suggested that the high proportion of Late Mesoproterozoic zircons were derived from the Amazonian Craton. However, more recent age compilations suggest that the Central Iberia zircon inheritance may derive, besides local sources, from the Pan-African orogen of north Gondwana, particularly the Saharan Metacraton in west Egypt (Bea et al. 2010). Pereira et al. (2012a) dated two graywackes from the Beiras Group (considered to be the correlative of the Lower Unit of the SGC in Portugal) using LA-ICPMS. The U–Pb data obtained yielded the youngest zircon populations at 578 ± 5 and 560 ± 6 Ma and the youngest zircon ages at 562 ± 6 and 550 ± 4 Ma, thus suggesting a Late Ediacaran maximum depositional age (c. 578–560 Ma). The age spectra also revealed a source area during the Neoproterozoic associated with a long-lived system of magmatism (c. 850–545 Ma) that developed along or in the vicinity of the northern Gondwana margin (Pan-African suture and Cadomian arc). Talavera et al. (2012) dated eight samples from different successions ascribed to the SGC from the central and western parts of the SGCD using SHRIMP. The youngest detrital zircon populations of samples ascribed to the Lower Unit yielded mean ages of 578 ± 9 Ma (seven zircons grains), 578 ± 7 Ma (eight grains), 577 ± 7 Ma (seven grains) and 546 ± 13 Ma (thirteen grains); the youngest individual zircon grains of samples of the Lower Unit yielded ages of 559 ± 10, 550 ± 6, 542 ± 9 and 537 ± 10 Ma. Taken together, these data gave a Late Ediacaran–Lower Cambrian maximum depositional age for the Lower Unit. As for the sample ascribed to the Upper Unit (sample from the Pusian Group), the youngest zircon population yielded a mean age of 545 ± 9 Ma (five grains) and the youngest zircons yielded ages of 536 ± 13 and 533 ± 17 Ma, suggesting a Cambrian age at least for the younger parts of the Upper Unit. Moreover, two samples attributed to the SGC unexpectedly yielded Cambro–Ordovician zircon populations. The authors suggested that they had to belong to younger than Lower Ordovician successions that, up to that moment, had not been differentiated from those of the SGC. The age spectra of the SGC pointed to the Saharan Metacraton as a contributing source for these SGC sediments plus Cadomian granitoids that could have been a local Neoproterozoic source. Fernández-Suárez et al. (2013) dated a sandstone of the Lower Unit from west Spain using LA-ICPMS; the youngest zircon population yielded a mean age of 555 ± 14 Ma (four zircon grains). All the aforementioned studies report U–Pb detrital zircon data from the west and central parts of the SGCD, where distinguishing between the Lower Unit and Upper Unit is not obvious. Our study, however, reports data from the eastern SGCD, where the distinction between both units, clearly unconformable, is mapped (see below).

The boundaries of the SGC are not exposed, but possible lateral correlatives are found to the north and south. To the north, below the Ollo de Sapo Formation, there is a locally metamorphic series of schists and migmatitic gneisses (González Lodeiro et al. 2004). In north Iberia (West Asturian-Leonese and Cantabrian Zones, Fig. 1b), Neoproterozoic successions of slates with sandstones and volcanoclastic intercalations (Villalba, Tineo and Narcea successions) lie unconformably below Cambrian sedimentary formations (Pérez-Estaún 2004; Marcos et al. 2004). These Neoproterozoic successions contain some Cryogenian (early Cadomian) detrital zircons, as young as c. 640 Ma (Fernández-Suárez et al. 2000; Gutiérrez-Alonso et al. 2003), and are intruded by c. 605–580 Ma (Cadomian) granitoids (Fernández-Suárez et al. 1998) and also contain c. 560 Ma rhyolitic layers (Gutiérrez-Alonso et al. 2004). The abundance of Ediacaran zircons (c. 620–550 Ma) in the discordant Cambrian sediments points to the age of such arc-related (late Cadomian) basins in north Iberia (Fernández-Suárez et al. 2000). More recently, Fernández-Suárez et al. (2013) subdivided the Narcea succession into two units. The lower one was intruded by c. 590–580 Ma granitoids and contained a youngest zircon population at 599 ± 3 Ma. The upper one had the youngest zircon population at 553 ± 4 Ma. They might correlate with the Lower and Upper Units of the SGC, respectively. To the south, in the southernmost part of the CIZ and in the Ossa-Morena Zone (OZM), a thick (>3 km) Late Neoproterozoic succession called the Serie Negra underlies the Lower Cambrian sedimentary formations (Azor et al. 2004). The Serie Negra consists of graphite-rich schists, slates and metagreywackes with black quartzites, amphibolites and minor marble intercalations (Azor et al. 2004). Widespread, though not very thick, calc-alkaline magmatic rocks are located in the uppermost part of the Serie Negra and in the overlying volcanoclastic Malcocinado Formation, as well as a number of Ediacaran granitoids. These granitoids are usually interpreted as arc-related magmatic products of Cadomian subduction (Sánchez Carretero et al. 1989, 1990; Pin et al. 2002; Bandrés et al. 2004; Simancas et al. 2004). The stratigraphic relationships between the Serie Negra/Malcocinado Formation and the SGC remain unresolved. It has been argued that the Serie Negra could extend northwards beneath the SGC (e.g., Martínez Poyatos et al. 2001), but a lateral change has also been proposed (e.g., Vidal et al. 1994a). The youngest U–Pb detrital zircon ages of metasedimentary rocks from the Serie Negra range between 565 and 545 Ma (Schäfer et al. 1993; Fernández-Suárez et al. 2002; Linnemann et al. 2008; Pereira et al. 2012b), thus overlapping with those from the SGC. According to the interpretation of Pereira et al. (2012a), the Serie Negra would be coeval to the Beiras Group (considered to represent the Lower Unit of the SGC in Portugal) and to the Villalba and Lower Narcea successions (north Iberia), all of them being somehow affected by a late Cadomian tectonothermal overprint (see discussion).

Conformably overlying the Upper Unit of the SGC (and the Malcocinado Formation to the south) there are, at some locations, fossiliferous formations (from bottom to top: sandstones, limestones and slates), well dated to the lower/middle stages of the Lower Cambrian (c. 530–520 Ma; Liñán et al. 2002). Like their correlatives from Morocco to central Europe, these deposits are interpreted as having formed on a shallow-marine platform during a short period of tectonic quiescence after the latest Cadomian activity and prior to widespread Cambro–Ordovician rifting (Simancas et al. 2004). Finally, the Lower Ordovician Armorican Quartzite Formation unconformably overlies the Ediacaran to Cambrian successions. This sedimentary unit is composed of white orthoquartzitic layers containing Cruziana and was deposited along the peri-Gondwana continental margin during a transgressive event. It is attributed to the Floian stage of the Lower Ordovician on the basis of its fossil and ichnofossil content (Moreno et al. 1976; Pickerill et al. 1984).

The intra-Alcudian unconformity

The main purpose of this work is to constrain the age of the angular unconformity that separates the Lower and Upper Alcudian successions of the SGC. Since the unconformity was first described by Bouyx (1970), some features, such as existence, regional extent and origin, have been a matter of debate (see reviews by Palero 1993; Vidal et al. 1994b; Valladares et al. 2002; Rodríguez Alonso et al. 2004b). The interest in the differentiation of the Lower and Upper Units in the SGC arose in the 1980s due to the existence of intercalations of phosphates, both in nodules and layers, at different levels of the Upper Alcudian succession and in the overlying olistostromic megabreccias and Pusian black shales, but absent in the Lower Alcudian succession.

In the central/eastern part of the SGCD, Bouyx (1970) described an outcrop where the Lower Alcudian beds appears in the core of an anticline surrounded by the Upper Alcudian beds, showing the former vertical attitude disrupted by gentle dips of the latter. Besides the angular relationships, Crespo and Rey (1971) described localities where the lower strata appeared refolded with upright fold hinges, in contrast with the subhorizontal upper strata.

Later regional stratigraphic studies did not recognize the existence of the unconformity, considering the Alcudian series as deposited during a unique sedimentary cycle (Tamain 1972; Parga and Vegas 1974; Capote et al. 1977; Herranz et al. 1977; Moreno 1977a, b; Vegas et al. 1977; Roiz 1979; Roiz and Vegas 1980). Moreover, some of them suggested the possibility of intraformational stratigraphic irregularities formed by gravitational mass slides. These studies, based on stratigraphic criteria, considered the bottom of the Upper Alcudian located at the base of a common conglomeratic graywacke bed. However, in most cases, the Upper Alcudian starts with a massive dark quartzite or with shaly–sandy fine-banded facies, and the first conglomerate lenses are located tens of meters up (Palero 1993).

Ortega and González Lodeiro (1986) described not only the angular/truncated geometrical relationships of the strata, but also structural elements. Above the unconformity, the Upper Alcudian rocks show low-dipping stratification and low-plunging intersection lineation (intersection between the stratification and the Variscan foliation). Below the unconformity, the Lower Alcudian rocks systematically show very high dip and plunge (close to vertical) of the stratification and the intersection lineation, respectively, in addition to folds that are oblique to the foliation and fold interference patterns. The data led to propose the existence of a tectonic phase of moderate folding (without related foliation/metamorphism) that verticalized the Lower Alcudian previous to the deposition of the Upper Alcudian. Subsequent systematic cartographic works (which included structural data) showed the existence of tectonic deformation previous to the Upper Alcudian at different locations of the southeastern part of the SGCD (Amor and Ortega 1987; Ortega and Sánchez 1987; García Sansegundo et al. 1987). Later studies confirmed the existence of the intra-Alcudian unconformity at different locations not only in the central, eastern and southern parts of the SGCD (San José 1984; Pieren et al. 1987; Nozal et al. 1988a, b; García Hidalgo 1988; Palero 1993) but also in the north (Robles and Álvarez Nava 1988; Martín Herrero 1989). Some of them also indicated the existence of a paleoalteration band that obliterates the internal features of the beds just below the unconformable surface.

More recent stratigraphical studies of the SGC successions, based on their ichnofossil record and sedimentary geochemistry, argue that a proposed intra-Alcudian unconformity is neither regional nor does it mark a significant hiatus (Vidal et al. 1994b; Valladares et al. 2002 and references therein). They suggest a virtually flysch continuous succession in which marine slope channels could have been cut by major slump masses, thus resulting in the local development of 'pseudo-unconformable' truncations between the Lower and Upper Alcudian. Regarding that interpretation, the basal part of the Pusian Group (accumulations of discontinuous megabreccias, olistostromes and chaotic deposits of carbonate, terrigenous and volcaniclastic composition) is commonly interpreted as a continuous succession that could have been separated into units by major slump masses, channels and aprons as a result of tectonic instability, marking the collapse of the adjacent mixed terrigenous-carbonate platform. That could explain the local development of intraformational unconformities and also the local emersion of some areas as a consequence of a major fall in the sea level (Valladares et al. 2002). It is noteworthy that these chaotic deposits wedge out laterally at regional scale, in some places resting directly on Lower Alcudian deposits. That feature may have led the above authors to mistake between an intraformational (Upper Unit) erosional surface and the older intra-Alcudian tectonic unconformity. Key studies to clarify this issue need to constrain the time gap in sedimentation involved in those cases showing evidences of tectonic deformation, a time gap that paleontology has not elucidated yet.

The last detailed structural work of Palero (1993) at two new localities in the eastern Alcudian anticline (southernmost SGCD; Fig. 1c) has been the reference to our geochronological study of the intra-Alcudian unconformity. Palero (1993) described at various sectors the orientation of the Upper Alcudian rocks (stratification, foliation, intersection lineation) as coherent with the Variscan folding. However, orientation data from the Lower Alcudian rocks (stratification with high dip and variable strike, intersection lineation with high plunge) systematically call for a previous (intra-Alcudian) folding phase. The first selected locality is northwest of San Lorenzo de Calatrava village (Fig. 1c) (Gargantilla streamlet, near Casa del Huerto). In this locality, the Lower Alcudian strata (decimeter-scale alternances of slates and quartz wackes) are oriented N40E/90. Toward the NE, these strata are cut and covered by the Upper Alcudian strata (centimeter- to decimeter-scale alternances of slates and quartz wackes) oriented N120/50N (Fig. 2a, b). The existence of an irregular band of Lower Alcudian rocks is noteworthy that is several meters wide and is located just below the unconformity that originates a topographic crest (Fig. 2a). This could have been the result of lithification (i.e., subsoil cementation) below the erosive surface formed, as a hardground, before the Upper Alcudian deposition started. As for the foliation, it runs N120E/70N, being the intersection lineation subvertical in the Lower Alcudian and subhorizontal in the Upper Alcudian.

The second selected locality is northeast of Solanilla del Tamaral village (Fig. 1c) (Fresnedas river, near Casas del Chorrillo). In this locality, the Lower Alcudian strata are oriented N20E/70W, and they are cut and overlain by N60/30 N oriented Upper Alcudian strata. In between, the unconformity (located at the top of a cliff formed in harder Lower Alcudian rocks) features an open upright antiform (Fig. 2c). As in the case of the first locality, the harder rocks forming the cliff might have also been the result of more lithification in a hardground formed just below the unconformable surface. The foliation, oriented N120/90, is the axial planar of the anticline. The intersection lineation is subvertical in the Lower Alcudian and low plunging in the Upper Alcudian (Fig. 2d).

Samples and analytical methods

We took five samples from the sandstones which overlie and underlie the unconformity surface at the two described localities (Fig. 2): samples ALC-1 and ALC-2 are from northwest of San Lorenzo de Calatrava, and samples ALC-3, ALC-4 and ALC-5 are from northeast of Solanilla del Tamaral. Two of them (ALC-1 and ALC-3) belong to the Lower Alcudian unit, and the other three (ALC-2, ALC-4 and ALC-5) belong to the Upper Alcudian unit (Table 1 of electronic supplementary material).

Overall, samples are poorly sorted wackes with ~25 % of matrix and variably foliated (Fig. 3). The sandy fraction is fine-grained, and it is composed of variably rounded grains of predominant quartz, less feldspar and minor muscovite, chlorite and Fe oxides. The matrix is composed of sericite, chlorite, Fe oxides and silt-sized grains of quartz and feldspar. Besides quartz and feldspar, sample ALC-5 also contains abundant clasts of chlorite displaying internal folds (Fig. 3f), which is an evidence of deformation prior to the Upper Alcudian deposition. From a sedimentary point of view, our samples are relatively homogeneous in terms of clast composition and texture, suggesting a common or similar source area and transport process for both the Lower and Upper Alcudian successions.

Zircon grains were separated at the University of Granada using conventional magnetic and heavy liquid techniques. They were mounted in 35-mm-diameter epoxy disks together with the U-concentration standard SL13 (238 ppm U) and the U–Pb standard TEMORA-2 (416.8 Ma; Black et al. 2003) and polished. The growth textures of the sectioned grains were studied by cathodoluminescence (CL) imaging using a scanning electron microscope at the Scientific Instrumentation Center of the University of Granada (SIC-UGR), and then selected areas on selected crystals were analyzed for U–Th–Pb using the SHRIMP IIe ion microprobe at Geoscience Australia (GA) and the SHRIMP IIe/mc at the SHRIMP Ion Microprobe Laboratory (IBERSIMS) of the University of Granada. Samples ALC-1, ALC-2, ALC-4 and ALC-5 were analyzed

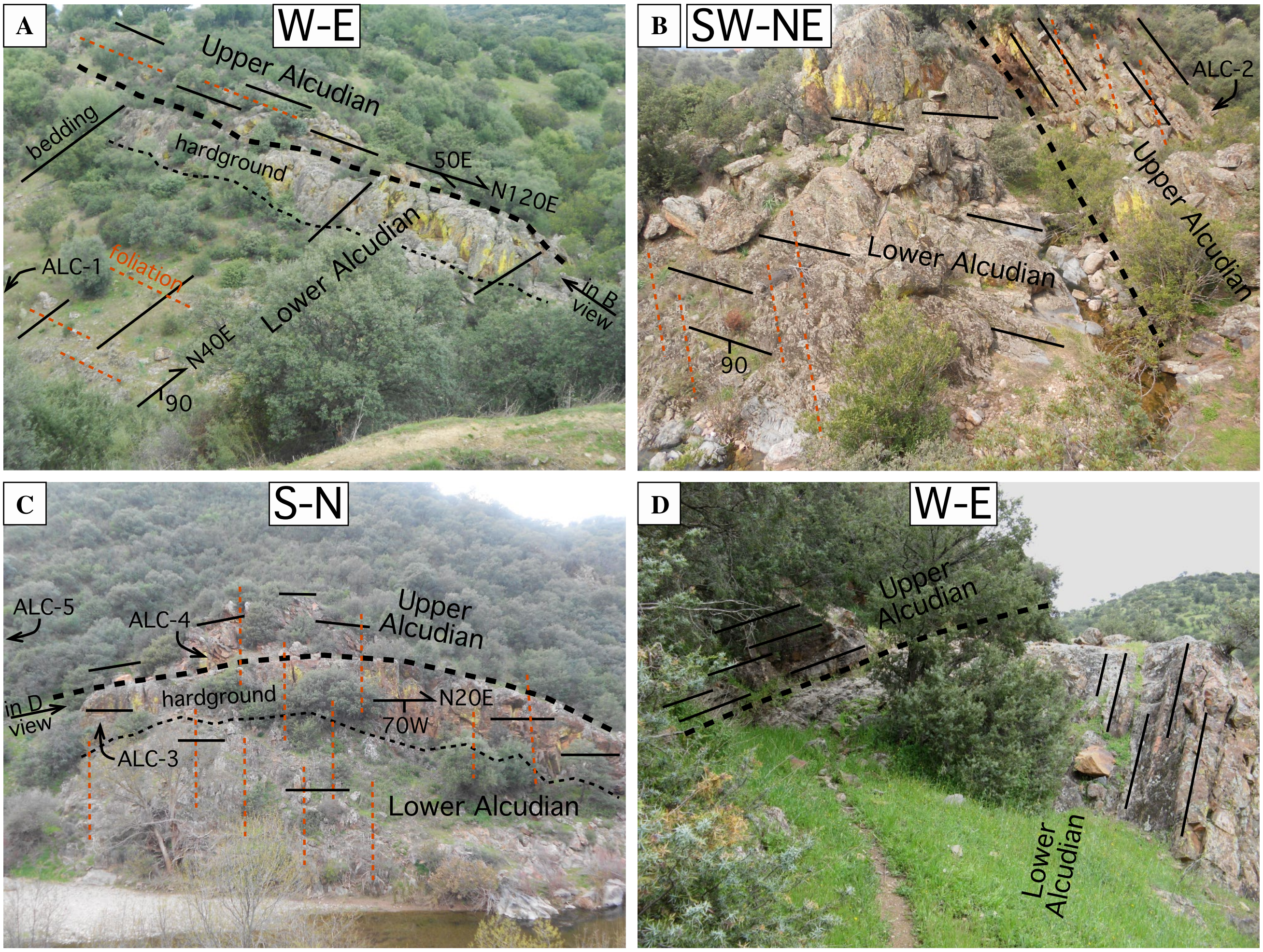

Fig. 2 Field relationships for the intra-Alcudian unconformity at the two selected locations (Palero 1993): San Lorenzo de Calatrava (**a**, **b**) and Solanilla del Tamaral (**c**, **d**)

at GA, and sample ALC-3 was analyzed at IBERSIMS. To minimize the possibility of Pb loss, we handpicked clear, transparent, bipyramidal and free-inclusion zircon grains and selected relatively bright zircons (that is, relatively low U zircons) from the CL imaging for the SHRIMP dating. Grains with fractures were also avoided when possible.

Ion microprobe analytical methods broadly follow those described by Williams and Claesson (1987). The analyses were made using a 10-kV, ~2 nA negative O_2 primary ion beam focused to 20 µm diameter. Positive secondary ions were accelerated to 10 kV, mass analyzed at high resolution (R5000) and counted using a single ETP electron multiplier by stepping the analyzer magnet. Each analysis consisted of five scans through the Zr, Pb, Th and U species of interest across the mass range from $^{196}Zr_2O$ to ^{254}UO. ^{204}Pb was also monitored and was reasonably small and had a negligible influence on the interpreted age in most cases. When necessary, common lead was corrected using the "207-correction" which is calculated by projecting the uncorrected analyses onto concordia from the assumed common $^{207}Pb/^{206}Pb$ present-day composition and from the measured ^{204}Pb using the $^{204}Pb/^{207}Pb$ ratios provided by the Stacey and Kramers (1975) model at the calculated age. All ages are calculated using the decay constant recommendations of Steiger and Jäger (1977). The data were processed using the SHRIMPTOOLS software developed by F. Bea (downloadable from http://www.ugr.es/~fbea/fbea/Software.html) using the STATA™ programming language. The data tables of the five samples from Lower and Upper Alcudian are listed in Tables 2–6 with 1σ precision (electronic supplementary material).

Zircon U–Pb results

The SHRIMP zircon U–Pb data of these samples were plotted on Tera–Wasserburg and Wetherill diagrams (Fig. 4). Analyses for which the ratio between the $^{206}Pb/^{238}U$ and $^{207}Pb/^{235}U$

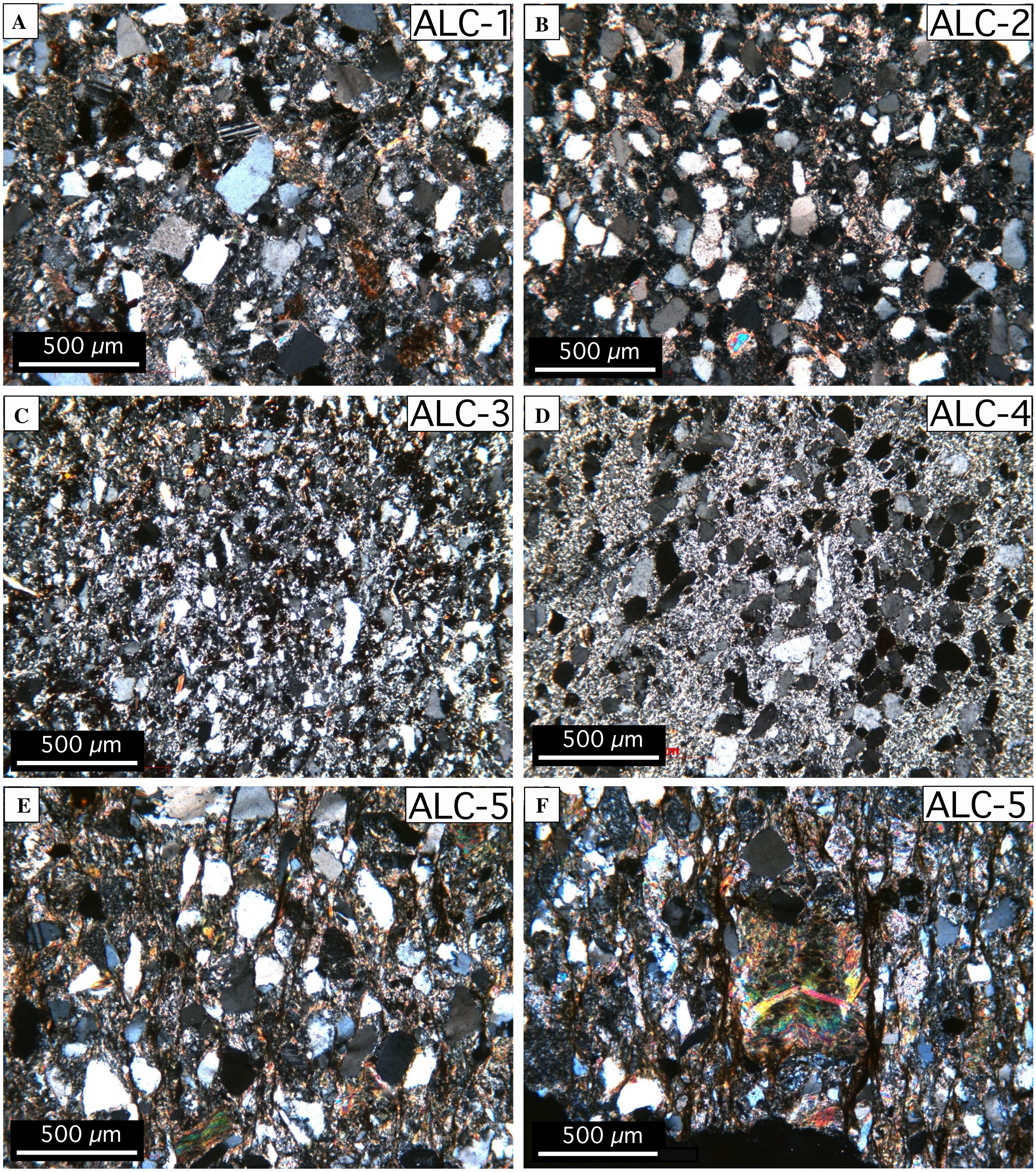

Fig. 3 Photomicrographs of studied samples, all in crossed nicols. Sections are cut normal to bedding and foliation (both visible in outcrop), and the latter is shown *vertical*

ages was outside the range 0.95–1.05 were rejected as excessively discordant and are not listed in the data Tables 2–6 (electronic supplementary material) or plotted in Fig. 4. The zircon populations were divided into different age groups, and the mean ages of these populations were calculated using the ^{207}Pb corrected ^{206}Pb/^{238}U ages for ages <1000 Ma and the ^{207}Pb/^{206}Pb ages for ages >1000 Ma. The U–Pb zircon data obtained from each sample are as follows:

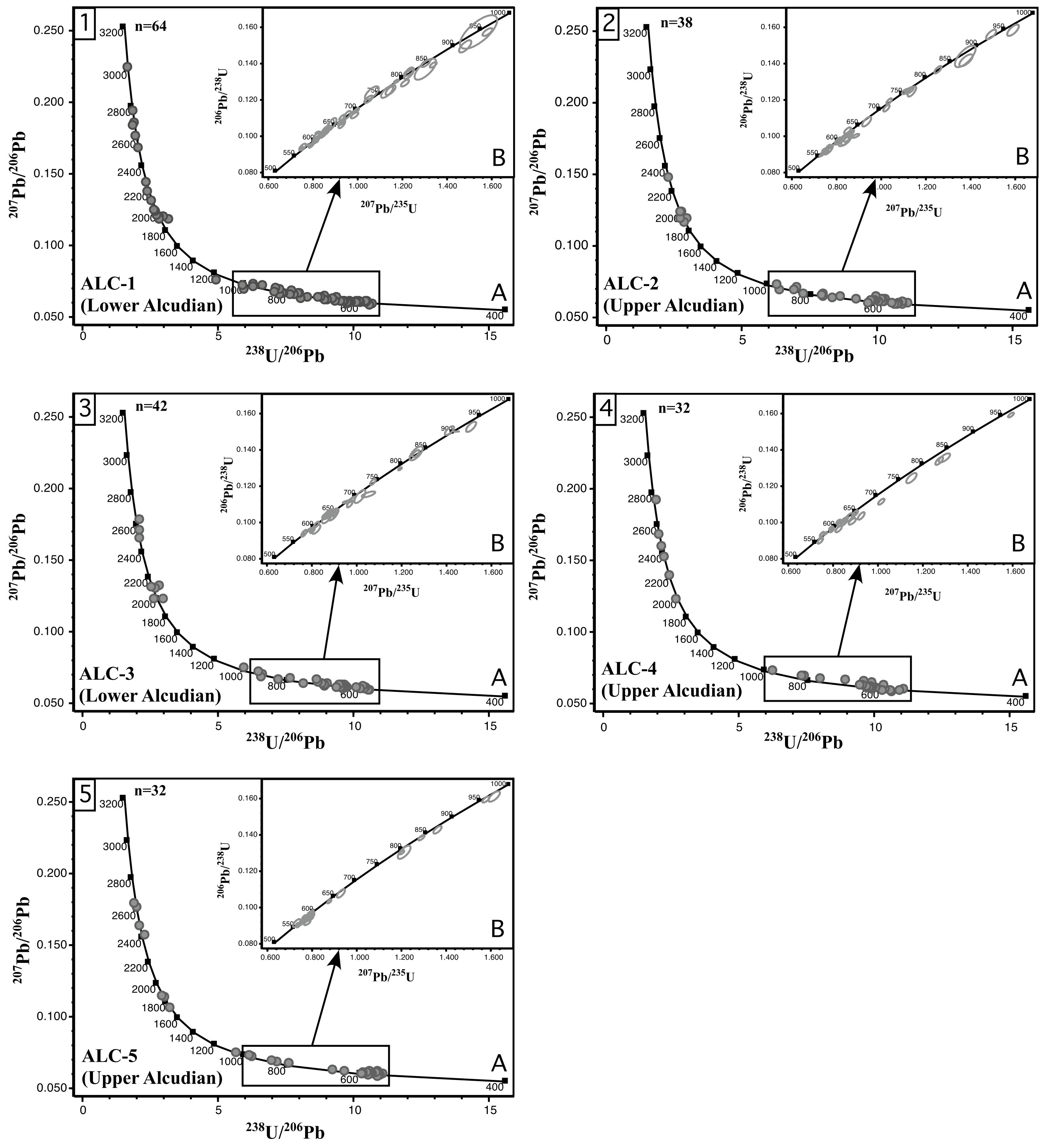

Fig. 4 U–Pb data plotted in Tera–Wasserburg (*a*) and Wetherill (*b*) concordia diagrams of 4.1-ALC-1 and 4.2-ALC-2 from San Lorenzo de Calatrava; 4.3-ALC-3, 4.4-ALC-4 and 4.5-ALC-5 from Solanilla del Tamaral

San Lorenzo de Calatrava locality

The two samples of this locality (ALC-1 and ALC-2) are rich in zircon grains. They are slightly to very rounded indicating that they have been affected by a process of erosion and transport. The CL imaging mainly shows magmatic oscillatory zoning in grains with no inherited cores and a few zircons with cores overgrown by thin rims which are mostly in sample ALC-1 (Fig. 5).

Fig. 5 Cathodoluminescence imaging of zircons of ALC-1 and ALC-2 from San Lorenzo de Calatrava; ALC-3, ALC-4 and ALC-5 from Solanilla del Tamaral

In the arkosic wacke sample (ALC-1), fifty-nine of 64 analyses were concordant with ages between 576 ± 3 and 3011 ± 9 Ma (Table 2 and Fig. 4a). About 77 % of those ages clustered around six populations; the most abundant had a mean age of 623 ± 7 Ma ($n = 16$) and the other five populations had mean ages of 582 ± 4 ($n = 5$), 674 ± 12

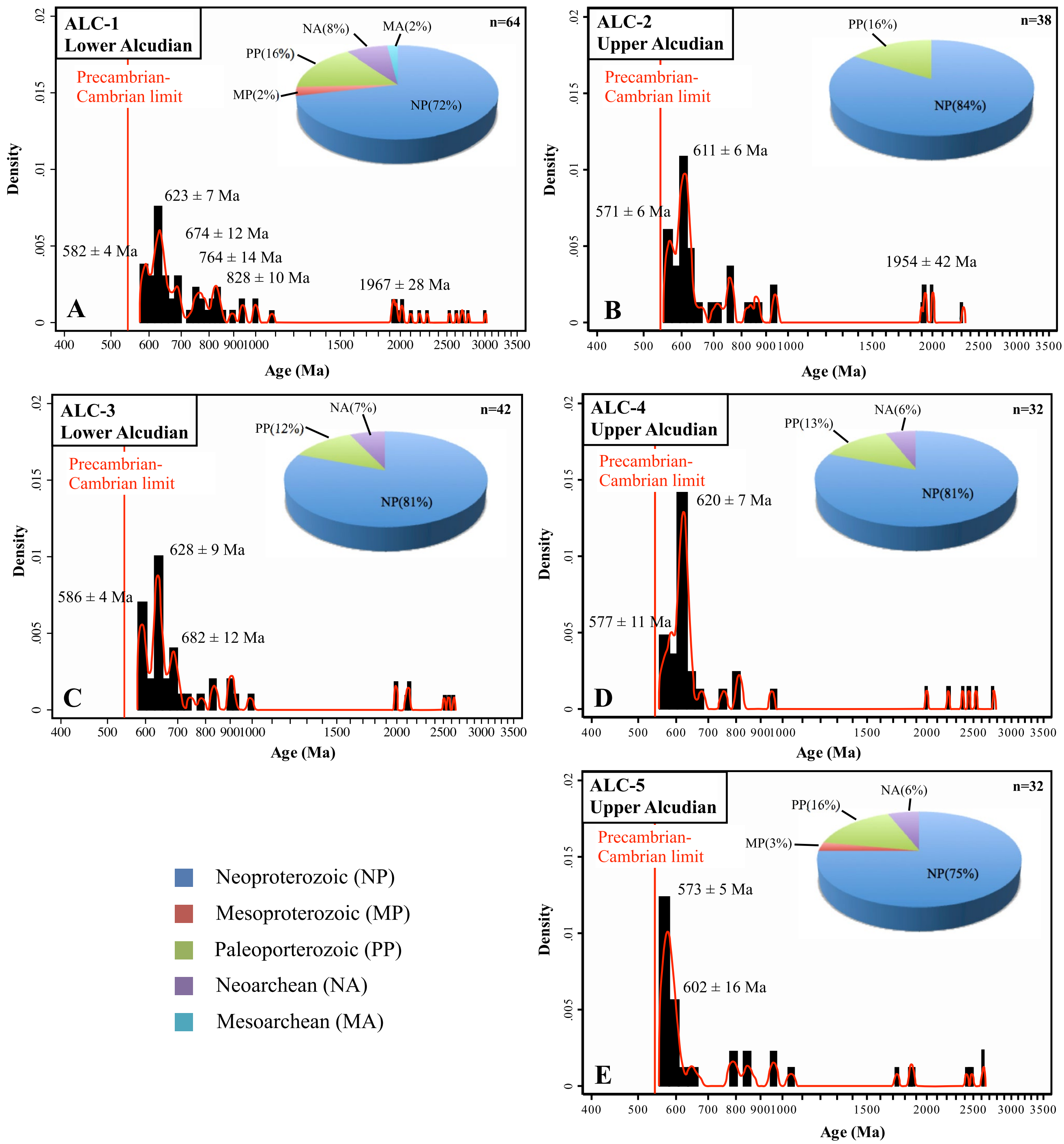

Fig. 6 U–Pb zircon age pattern for **a** ALC-1 and **b** ALC-2 from San Lorenzo de Calatrava; **c** ALC-3; **d** ALC-4 and **e** ALC-5 from Solanilla del Tamaral. Capital letters: *NP* Neoproterozoic, *MP* Mesoproterozoic, *PP* Paleoproterozoic, *NA* Neoarchean, *MA* Mesoarchean

($n = 8$), 764 ± 14 ($n = 7$), 828 ± 10 ($n = 6$) and 1967 ± 28 ($n = 7$) Ma (Fig. 6a). The remaining 23 % of the concordant ages were Tonian (c. 895–942 Ma), Mesoproterozoic (c. 1081 Ma), Rhyacian (c. 2.10–2.29 Ga), Neoarchean (c. 2.53–2.77 Ga) and Mesoarchean (c. 3.01 Ga) in age.

Thirty-eight zircons from the arkosic wacke (ALC-2) were analyzed, and all provided concordant dates ranging from 552 ± 2 to 2317 ± 18 Ma (Table 3 and Fig. 4b). Those ages mainly gathered in three populations, two Ediacaran (571 ± 6, $n = 6$; and 611 ± 6 Ma, $n = 14$) and one Orosirian (1954 ± 42 Ma, $n = 5$) (Fig. 6b). Besides those, there are also other ages that were Lower Ediacaran (c. 552 Ma), Cryogenian (c. 660–826 Ma), Tonian (c. 852–945 Ma) and Siderian (c. 2.3 Ga).

Solanilla del Tamaral locality

Zircons are quite abundant in samples ALC-4 and ALC-5, whereas sample ALC-3 is poor in zircon and only a few dozens were found. As in San Lorenzo de Calatrava locality, grains are also slightly to very rounded. According to the CL imaging, most zircons display continuous oscillatory zoning and a few zircons have partially dissolved cores surrounded by younger overgrowths (Fig. 5).

In the arkosic wacke sample (ALC-3), forty-two of 46 zircons yielded concordant Ediacaran to Neoarchean dates (Table 4 and Fig. 4c). Twenty-six of those ages clustered around three populations (Fig. 6c) with mean ages of 586 ± 4 ($n = 5$), 628 ± 9 ($n = 14$) and 682 ± 12 ($n = 7$) Ma. There are also older grains which are Cryogenian (c. 745–834 Ma), Tonian (c. 901–992 Ma), Orosirian (c. 2.0 Ga), Rhyacian (c. 2.09–2.12 Ga) and Neoarchean (c. 2.51–2.64 Ga) (Fig. 6c).

The thirty-two zircon grains analyzed from the quartz wacke sample (ALC-4) were all concordant with ages between 555 ± 6 and 2756 ± 3 Ma (Table 5 and Fig. 4d). The Neoproterozoic ages mostly clustered in two Ediacaran populations with mean ages of 577 ± 11 ($n = 4$) and 620 ± 7 ($n = 16$) Ma (Fig. 6d). There are also six Neoproterozoic ages: one Lower Ediacaran (c. 555 Ma), four Cryogenian (c. 678–817 Ma) and one Tonian (c. 951 Ma). Apart from those, there are also six other ages: one Orosirian (c. 2.0 Ga), one Rhyacian (c. 2.21 Ga), two Siderian (c. 2.37–2.45 Ga) and two Neoarchean (c. 2.53–2.76 Ga) in age.

Thirty-five analyses were made on 35 zircons from this wacke (ALC-5), 32 of which were concordant with ages from 555 ± 4 to 2641 ± 10 Ma (Table 6 and Fig. 4e). Two Ediacaran populations were distinguished in the zircon age pattern with mean ages of 573 ± 5 ($n = 10$) and 602 ± 16 Ma ($n = 5$) (Fig. 6e). There are also two younger Ediacaran ages at 555 ± 4 and 560 ± 3 Ma. Besides those, there are also other ages which are Cryogenian (c. 659–836 Ma), Tonian (c. 859–962 Ma), Mesoproterozoic (c. 1043 Ma), Statherian (c. 1.73 Ga), Orosirian (c. 1.85–1.87 Ga), Siderian (c. 2.42–2.49 Ga) and Neoarchean (c. 2.61–2.64 Ga).

Discussion

Age of the intra-Alcudian stratigraphic gap and the Cadomian deformations in Iberia

Our detrital zircon SHRIMP U–Pb data give information on the time gap in sedimentation in relation to the intra-Alcudian unconformity in the southern part of the SGCD, southern CIZ. Five sedimentary rocks from the Lower and Upper Alcudian successions were sampled just below and above the unconformity at two localities. The youngest detrital ages may provide the maximum depositional ages of both lithostratigraphic successions although the possibility of Pb loss cannot be completely excluded. In that scenario, the maximum depositional age may not be defined by the youngest zircon grains but by the youngest zircon populations. However, we have tried to minimize this possibility following the methodology described in "Samples and analytical methods" section.

The youngest zircon populations of the two samples (ALC-1 and ALC-3) from the Lower Alcudian yielded mean ages of 582 ± 4 Ma (five grains, sample ALC-1) and 586 ± 4 Ma (five grains, ALC-3). As for the three Upper Alcudian samples (ALC-2, ALC-4 and ALC-5), the youngest zircon populations are 571 ± 6 Ma (six grains, ALC-2), 577 ± 11 Ma (four grains, ALC-4) and 573 ± 5 Ma (ten grains, ALC-5) (Tables 2–6 and Fig. 6). A closer view is brought by the youngest zircon grain of each sample. The youngest zircon grains of the samples from the Lower Alcudian yielded concordant ^{207}Pb-corrected $^{206}Pb/^{238}U$ ages of 576 ± 3 Ma (ALC-1) and 580 ± 7 Ma (ALC-3) (Tables 2 and 4), and the ones of the samples from the Upper Alcudian gave concordant ^{207}Pb-corrected $^{206}Pb/^{238}U$ ages of 552 ± 2 Ma (ALC-2), 555 ± 6 Ma (ALC-4) and 555 ± 4 Ma (ALC-5) (Tables 3, 5 and 6). Thus, a maximum depositional age of c. 580–576 Ma for the Lower Alcudian overlap with the range 580–560 Ma obtained in most samples from previous studies in other parts of the Lower Unit of the SGCD (Pereira et al. 2012a; Talavera et al. 2012). As for the Upper Alcudian, a depositional age younger than c. 555–552 Ma agrees with the Lower Cambrian paleontological evidence in the Alcudia area (García Hidalgo 1993). The youngest zircon grains (maximum depositional ages) of our study give, on average, a time gap of c. 21 Ma (from c. 580–576 to c. 555–552 Ma) for the intra-Alcudian unconformity in the Alcudia area. This time gap has to be considered as a maximum interval, and it could be shorter whether a new magmatic event had provided zircons during this interval. Thus, the hiatus in sedimentation (and therefore the time window for the unconformity) may be smaller than the gap in the detrital zircon record.

It is noteworthy that some of the previously reported youngest detrital zircon ages of the Lower Unit (Pereira et al. 2012a; Talavera et al. 2012; Fernández-Suárez et al. 2013) overlap with our proposal of the intra-Alcudian time gap, and they even overlap with the estimated age for the Upper Unit (Upper Alcudian in our study). This discrepancy may be attributed either to the fact that the previous reported data came from the northwestern sectors of the SGCD or to the Pb loss. In the northwestern sectors, the unconformity between Lower Unit and Upper Unit has been poorly described, and thus, the differentiation of both units, when lithologies are similar, is not as clear as in the case of the southeastern sectors of the domain (e.g., Alcudia area). As it was pointed out by Talavera et al. (2012), a stratigraphic revision of the SGCD is needed in order to clarify the correlation of the multiple-defined units and subunits (which can be diachronous indeed) across the large SGCD.

We argue below that the maximum depositional ages might not differ substantially to their respective depositional ages. First, the SGC successions incorporated not only old inherited material but also igneous juvenile material (Rodríguez Alonso et al. 2004b). This juvenile material could be derived from a number of late Cadomian c. 600–540 Ma old granitoids, and volcanics, that intruded the Ediacaran successions in Iberia (see "Comparison with other samples of the SGC and zircon provenance" section). Second, the regional geology constrains the depositional age of the Alcudian successions: (1) the deposition of the Upper Alcudian and Pusian successions must be older than the overlying fossiliferous sandstones and limestones, dated as the lower/middle stages of the Lower Cambrian (c. 530–520 Ma; Liñán et al. 2002), and (2) the Lower Alcudian was affected by a late Cadomian tectonic event (see "The intra-Alcudian unconformity" section) as well as other lateral correlative successions in surrounding areas. Evidence of late Cadomian deformations is found to the north and south of the CIZ.

To the north, Díaz García (2006) described, in the Neoproterozoic Narcea Slates (West Asturian-Leonese and Cantabrian Zones, Fig. 1b), a Cadomian deformation event (c. 560–540 Ma) that formed large asymmetric NW-vergent folds, with related axial planar cleavage, and was developed in the lower greenschist facies. These folds are truncated by the Lower Cambrian angular unconformity. Lower Cambrian conglomerates contain pebbles with predepositional microstructures that are widespread in the underlying Neoproterozoic rocks. Gutiérrez-Alonso et al. (2005) obtained $^{40}Ar/^{39}Ar$ age data of single detrital muscovite grains from Ediacaran and Cambrian sedimentary rocks of northwest Iberia, which were mostly derived from proximal sources and record low-temperature processes. A sample from the Ediacaran Narcea Slates yielded an age interval of c. 785–590 Ma, interpreted as derived from the Cadomian/Avalonian/Pan-African arc. A sample from the discordant Lower Cambrian sandstones, although displayed a more complex age spectrum with Neoproterozoic (c. 550–640 Ma), Mesoproterozoic–Neoproterozoic boundary (c. 920–1060 Ma), Paleoproterozoic (c. 1580–1780 Ma) ages, yielded a youngest age of 550 ± 3 Ma (excluding two Eocene ages interpreted as reset by fluid circulation during the Eocene uplift of the Cantabrian Mountains in north Iberia). This age perfectly matches the late Cadomian low-temperature folding event described by Díaz García (2006).

To the south, the imprint of the Cadomian orogeny below the Lower Cambrian sediments in SW Iberia (OMZ and southernmost CIZ) has been a controversial issue over the last few decades (Ábalos et al. 1991; Quesada 1991; Azor et al. 1993, 1994; Eguíluz et al. 2000; Expósito 2000; Simancas et al. 2001, 2004). The Cadomian imprint mostly consists of widespread, though not very voluminous, calc-alkaline, Late Ediacaran arc-related magmatism (Sánchez Carretero et al. 1989, 1990; Pin et al. 2002; Bandrés et al. 2004; Simancas et al. 2004). Locally, the Serie Negra in the southernmost CIZ shows a main foliation and a low- to medium-grade metamorphism which must have developed before the Variscan orogeny (Capdevila et al. 1971). This metamorphism has been dated by $^{40}Ar/^{39}Ar$ in two localities. In Peraleda del Zaucejo, a muscovite concentrate from a graphite biotite schist yielded a mean age of 550 ± 10 Ma (Blatrix and Burg 1981). In Oliva de Mérida, Dallmeyer and Quesada (1992) obtained an interval age c. 560–550 Ma on two hornblende and one muscovite concentrates from foliated amphibolites and graphite schist. It is noteworthy that some of the reported youngest detrital zircon ages of the Serie Negra in the OMZ (c. 545 Ma; Fernández-Suárez et al. 2002; Linnemann et al. 2008; Pereira et al. 2012b) are slightly younger than the mentioned metamorphic ages. This difference may be explained by the existence of diachronous sedimentary and/or metamorphic processes across these large areas, the presence of inherited grains in the Ar/Ar multigrain samples or minor Pb loss on the youngest U–Pb ages.

In summary, all the evidences of Cadomian tectono-metamorphic imprint in Iberia point to the interval c. 560–550 Ma, affecting correlated Neoproterozoic successions (Narcea in north Iberia, Lower Alcudian/Beiras Group in the CIZ, Serie Negra in the southernmost CIZ). This time interval overlap with (1) the time gap in sedimentation that we have shown in the intra-Alcudian unconformity (which includes a folding event), and (2) the young c. 555–552 Ma detrital zircons that is present in the Upper Alcudian but is absent in the Lower Alcudian. Furthermore, this Cadomian orogenic imprint has been recognized in other massifs of the European Variscan belt (such as the Armorican Massif, Graindor 1957). In the Saxo-Thuringian Zone (Central Europe), a Cadomian magmatic arc also developed during c. 750–570 Ma (Linnemann et al. 2007, 2014). After cessation, it was followed by the closure of the back-arc basin (Cadomian unconformity) between c. 570 and 542 Ma. This late Cadomian evolution is quite similar (in terms of unconformities and zircon ages) to the one shown for Central Iberia in this paper. They point out the involvement of large Ediacaran arc-related volcano-sedimentary complexes in the late Cadomian deformation along the active northern border of Gondwana.

Characterization of the Lower and Upper Alcudian zircon spectra

The combination of the new SHRIMP U–Pb data reported in this paper defines the zircon age pattern of the Lower and Upper Alcudian in the southeastern SGCD (Fig. 7). On the one hand, the zircon age pattern of the two samples from the Lower Alcudian (ALC-1 and ALC-3) is mainly composed of Neoproterozoic (76 %) and Paleoproterozoic (14 %) ages, with minor Neoarchean (8 %), Mesoproterozoic (1 %) and Mesoarchean (1 %) ages (Fig. 7a). Moreover, these zircon ages cluster in six populations: two Ediacaran (c. 580–590 and c. 620–630 Ma), three Cryogenian (c.670–680, c. 760–770 and c. 830–840 Ma) and one Orosirian (c. 1.97–1.98 Ga) (Fig. 7a).

On the other hand, the zircon age spectrum of the three samples from the Upper Alcudian (ALC-2, ALC-4, ALC-5) mostly comprises Neoproterozoic (80 %) and Paleoproterozoic (15 %) ages, with minor Neoarchean (4 %) and Mesoproterozoic (1 %) ages (Fig. 7b). The zircon ages gather in six populations: three Ediacaran (c. 550–560, c. 570–580 and c. 610–620 Ma), two Cryogenian (c. 760–770 and c. 830–840) and one Tonian (c. 940–950 Ma) in age (Fig. 7b).

Despite the similar zircon patterns of both Lower and Upper Alcudian, composed of Neo- and Paleoproterozoic ages and minor Neoarchean and Mesoproterozoic, some differences can be distinguished. The proportion of Archean zircons is higher in the samples from the Lower Alcudian (9 %) than in the ones from the Upper Alcudian (4 %). The samples from the Lower Alcudian also have Mesoarchean ages, absent in the ones from the Upper Alcudian. A Late Cryogenian population (c. 670–680 Ma) is only found in the samples from the Lower Alcudian, and a Tonian population (c. 940–950 Ma) is only described in the Upper Alcudian. These differences may reflect inheritance from slightly different far-field eroding basements, or from an eroding local source (evolving Cadomian arc). At this respect, the youngest Late Ediacaran population (c. 550–560 Ma) is only found in the samples from the Upper Alcudian.

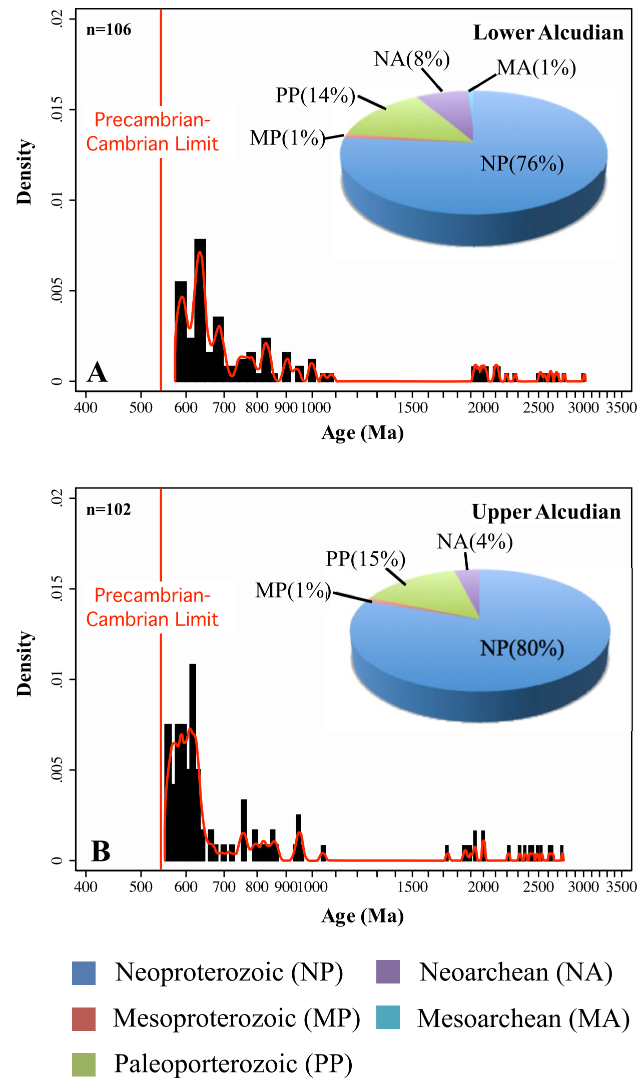

Fig. 7 U–Pb zircon age pattern for **a** Lower Alcudian (ALC-1 and ALC-3) and **b** Upper Alcudian (ALC-2, ALC-4 and ALC-5). Capital letters: *NP* Neoproterozoic, *MP* Mesoproterozoic, *PP* Paleoproterozoic, *NA* Neoarchean, *MA* Mesoarchean

Comparison with other samples of the SGC and zircon provenance

Samples from north and west part of the SGC have been recently dated by Pereira et al. (2012a) (Beiras Group) and Talavera et al. (2012) (Lower and Upper Units). The samples from both studies showed similar zircon patterns to the ones of the Lower and Upper Alcudian (hereafter called Alcudian) (Fig. 8). The zircon pattern of the Beiras Group is mainly composed of Neoproterozoic (75 %) and Paleoproterozoic (12 %) ages and minor Mesoproterozoic (6 %), Neoarchean (5 %), Mesoarchean (0.5 %) and Paleoarchean (2 %) ages (Fig. 8a), and the zircon pattern of the Lower and Upper Units mainly comprises Neoproterozoic (73 %) and Paleoproterozoic (15 %) ages and minor Mesoproterozoic (3 %), Neoarchean (7 %), Mesoarchean (2 %) and Cambrian (1 %) ages (Fig. 8b). However, very slight differences can be distinguished. First, samples from Beiras Group have Paleoarchean zircons, absent in the Alcudian and Lower and Upper Units. Second, Cryogenian zircons are more abundant in samples from the Beiras Group. These differences may point to slightly different sources of the sediments.

As a whole, the Alcudian metasedimentary rocks, as the Lower and Upper Units, and Beiras Group metasedimentary rocks, are characterized by a high proportion of

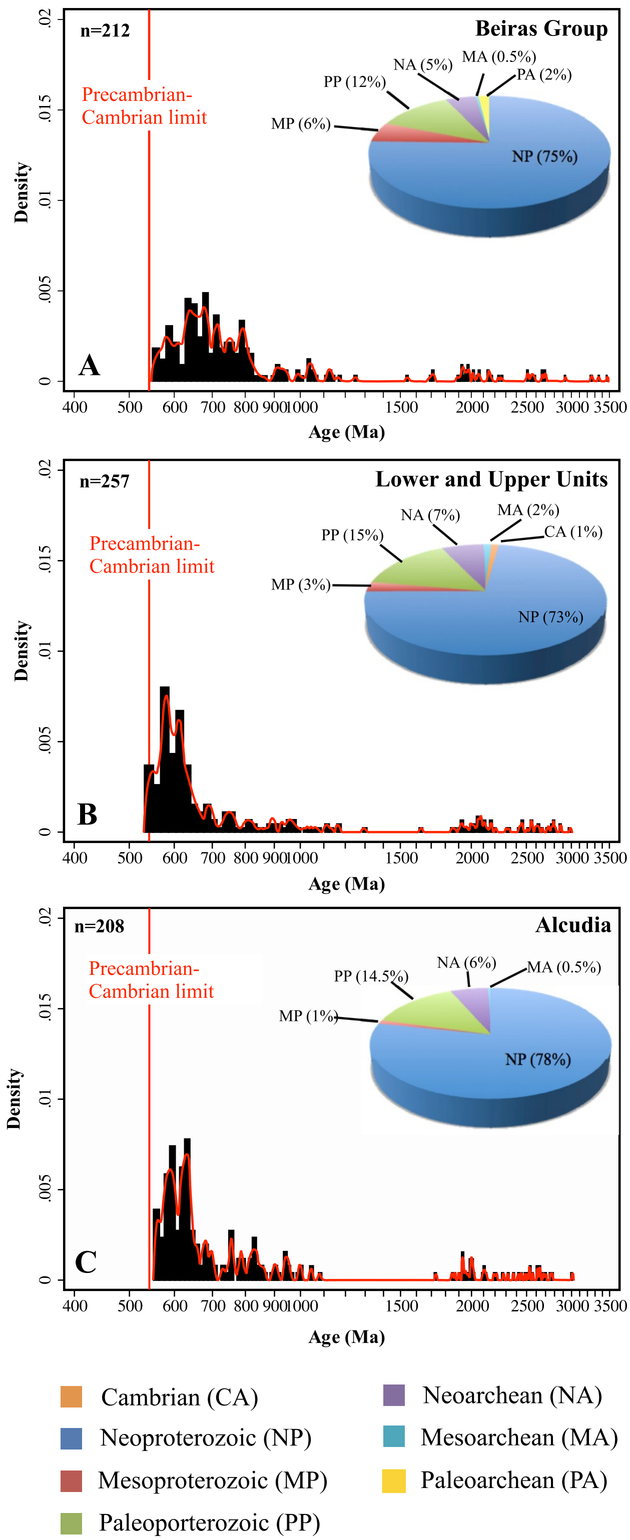

Fig. 8 U–Pb zircon age pattern for **a** Beiras Group (Pereira et al. 2012a); **b** Lower and Upper Units and (Talavera et al. 2012); and **c** Alcudia (this study). Capital letters: *CA* Cambrian, *NP* Neoproterozoic, *MP* Mesoproterozoic, *PP* Paleoproterozoic, *NA* Neoarchean, *MA* Mesoarchean, *PA* Paleoarchean

Neoproterozoic zircon that mainly clusters in three Ediacaran (c. 550–560, c. 570–580 and c. 610–620 Ma) and one Cryogenian populations (c. 670–680 Ma) and coincide with four magmatic events occurred at the northern margin of Gondwana during the Pan-African orogeny (Thomas et al. 2002; Gasquet et al. 2005) (Fig. 8c). According to Pereira et al. (2012a) and Talavera et al. (2012), the potential source of these Neoproteozoic ages are located throughout the northern margin of Africa, in the West African Craton (Anti-Atlas, Morocco) and in the Saharan Metacraton (Algeria, western Egypt and Sudan). Local potential sources of Ediacaran zircons are also found in a number of Cadomian granitoids (c. 605–545 Ma) in the CIZ (Valle de la Serena: Ordóñez Casado 1998; La Almohalla: Bea et al. 2003; Mérida: Bandrés et al. 2004; Aljucén: Talavera et al. 2008; Vila Nova: Reis et al. 2010), in the OMZ (Mosquil: Ochsner 1993; Ahillones: Ordóñez Casado 1998; Lora del Río: Ordóñez Casado 1998) and in north Iberia (Pola de Allande: Fernández-Suárez et al. 1998). Two other minor Cryogenian populations of the Alcudian samples (c. 760–770 and c. 830–840 Ma) are worthy of consideration and also coincide with tectonomagmatic events on the northern margin of Gondwana (Samson et al. 2004; D'Lemos et al. 2006; El Hadi et al. 2010) (Fig. 8c).

Paleoproterozoic, Neo- and Mesoarchean zircons are less abundant in the Alcudian sediments (Fig. 8c), as in the Lower and Upper Units and Beiras Group. Paleoproterozoic ages are related to the Eburnian orogeny, igneous rocks of which are located in two areas of north Africa: in the West African Craton (western and central Anti-Atlas, Morocco: Malek et al. 1998; Walsh et al. 2002; Thomas et al. 2002; Barbey et al. 2004; Gasquet et al. 2004) and in the Saharan Metacraton (Western Egypt: Bea pers. com.). Neo- and Mesoarchean zircons might also originate from the same places, that is, the Saharan Metacraton (western Egypt: Bea et al. 2011) and the West African Craton (Reguibat Shield: Potrel et al. 1998; Key et al. 2008; Man Shield: Barth et al. 2002).

It is noteworthy that the Alcudian metasedimentary rocks, as the Lower and Upper Units, and Beiras Group sediments, have a low percentage of Mesoproterozoic zircons with ages from 1.0 to 1.1 Ga that enables to discriminate the probable source of these sediments (Fig. 8c). According to Fernández-Suárez et al. (2000), Murphy et al. (2008) and Abati et al. (2010), the presence of Mesoproterozoic zircon was a distinctive feature of sediments derived from the Amazonian Craton due to the absence of those zircons in the West African Craton. However, new U–Pb dating of Pan-African igneous rocks from Algeria and west Egypt in the Saharan Metacraton has shown a small proportion of Mesoproterozoic inherited zircons with ages between 1.0 and 1.35 Ga (Henry et al. 2009; Bea et al. 2010; Fezaa et al. 2010). This age range matches the one of the Mesoproterozoic inherited zircons of the Alcudian sediments and suggests that the Saharan Metacraton could have been the source of these sediments. The hypothesis, that links the Neoproterozoic sediments of the SGC with the Saharan Metacraton, has been recently made by Talavera et al. (2012) and is in agreement with the previous U–Pb dating in sedimentary rocks from the NW Iberian and Bohemian massifs (Díez Fernández et al. 2010; Drost et al. 2011).

Conclusions

The youngest concordant U–Pb ages of detrital zircons from sandstones sampled just below and above the intra-Alcudian angular unconformity in the SGC (CIZ) yielded c. 580–576 Ma and c. 555–552 Ma, respectively. These maximum depositional ages support the existence of a relevant sedimentary time gap of about 21 Ma between the Lower and Upper Alcudian successions which is difficult to explain in terms of a unique and continuous sedimentary cycle for the complex. Concerning the angular relationships, the time gap supports the existence of a tectonic folding event instead of synsedimentary slump-related strata truncations.

The time gap of the intra-Alcudian unconformity is consistent with the regional geology. The combination of the geochronological and structural data suggests that the most probable maximum depositional ages are c. 580–560 Ma for the Lower Alcudian metasedimentary rocks (previous to the late Cadomian folding event) and c. 550–540 Ma for the Upper Alcudian metasedimentary rocks.

Sedimentation took place in basins contemporaneous to Ediacaran arc magmatism (the Cadomian arc of north Gondwana), but was interrupted due to a late Cadomian orogenic tectonothermal event (c. 560–550 Ma) that differently imprinted various parts of Iberia. As the CIZ concerns, the intra-Alcudian unconformity has been widely recognized (and now constrained in age) in the southeastern part. More stratigraphic, structural and geochronological studies are needed to clarify its extent to the rest of the zone.

The zircon age patterns suggest, in agreement with other recent data, a sedimentary provenance from the northern margin of Gondwana affected by the Pan-African orogeny: the West African Craton and/or the Saharan Metacraton. Cadomian young magmatism (granitoids and synsedimentary volcanics) could have also contributed as a local source. In this respect, it is noteworthy the existence of a c. 555–552 Ma old zircon population in the Upper Alcudian (but absent in the Lower Alcudian) which match the late Cadomian orogenic activity in Iberia.

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