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Recycling in the subduction factory: Archaean to Permian zircons in the oceanic Cretaceous Caribbean island-arc (Hispaniola).

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Abstract

Little mineralogical evidence is left of the recycling of continental and oceanic crust into the mantle at subduction zones. Zircon, because of its exceptional robustness, is probably the only surviving phase and the best mineral tracer of this global-scale process. This article combines new in-situ U-Pb dating and O and Hf isotope analyses on Cretaceous (co-magmatic) and pre-Cretaceous (inherited) zircons separated from Albian-Aptian arc-related igneous rocks from the Dominican Republic. The O and Hf systematics of Cretaceous zircons reflect derivation from predominantly juvenile sources and variable mixing with evolved melts, as expected for an oceanic island-arc. Inherited zircons yield U-Pb ages between 256 and 2923 Ma (n = 219). Most studied inherited zircons are Permian-Carboniferous, peaking at ca. 300 Ma, and formed in continental crust to judge from mineral inclusions (including quartz, orthoclase, muscovite and apatite), $\delta^{18}O$ (6.16 to 12.67 ‰) and εHf(t) (-11.9 to 1.12) values. Ordovician to Proterozoic zircons yield $\delta^{18}O$ between 4.93 and 9.21 ‰ and εHf(t) between -19.9 and 8.5 and hence crystallized in equilibrium with either mantle-derived (juvenile) magmas or magmas derived from melting of supracrustal rocks. For Archean zircons, $\delta^{18}O$ between 6.98 and 7.94 ‰ and εHf(t) between -18.5 and -1.6 might reflect contribution of supracrustal materials in the parent magmas. We suggest that arc magmas picked up the inherited zircon cargo from their mantle sources. Accounting paleogeographic and paleo-tectonic constraints, Mexican and Colombian terranes are identified as the potential primary (magmatic) and secondary (sedimentary) sources for the studied inherited zircons. We envisage that inherited zircons were transported downward to the mantle beneath the Caribbean plate as detrital grains on top of the subducting Proto-Caribbean and/or Pacific slabs.

Keywords: inherited zircon, crustal recycling, subduction zone, Caribbean island-arc, Dominican Republic

1. Introduction

Volcanic arcs are sites of intense magmatism and the factory where crust is primarily created (Stern, 2002; Davidson and Arculus, 2006; Jicha and Jagoutz, 2015). In such a factory, magma is generated in a mantle wedge bounded by the subducting slab and overlying crust, the latter either composed of oceanic or continental rocks (Leat and Larter, 2003). As such, subduction zones are, in turn, important loci for the recycling of oceanic and continental material back into the mantle. Much of the subducted continental crust is probably driven back to crustal levels (accretionary prism/orogen). However, the incoming of oceanic crust (igneous, commonly altered/hydrated and sedimentary cover) and tectonically eroded fragments of oceanic forearc provide exotic elements that might be transferred into the mantle at the relatively shallow depths of the mantle wedge

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through the asthenosphere and transition zone to the lower mantle (Scholl and von Huene, 2010, and references therein). Such contribution of crustal, geochemically evolved materials to the mantle modifies its composition, creating metasomatized elemental and isotopic mantle "reservoirs" identified in juvenile mantle-derived liquids (e.g., Nebel et al., 2011; Whattam and Stern, 2011; Pearce, 2014). However, little direct mineralogical evidence is observed in magmatic rocks influenced by this global-scale recycling process. Zircon, because of its exceptional physical and chemical robustness (Harley and Kelly, 2007; Bindeman and Melnik, 2016), is the most notable exception. Ample proof of this are the description of crustal-derived zircons in ophiolitic peridotites and chromitites (Yamamoto et al., 2013; McGowan et al., 2015; Robinson et al., 2015; Li et al., 2016; Rojas-Agramonte et al., 2016) and of inherited, much older than the carrier magmas, zircons in intra-oceanic arc basalts (Rojas-Agramonte et al., 2016).

During the course of radiometric dating of Cretaceous (Aptian-Albian) arc-related igneous rocks from central and eastern Dominican Republic, abundant inherited zircons with a wide spread of pre-Cretaceous ages (from Jurassic to Archean) were discovered. This finding aligns with a previous description of inherited zircons in Cuban ophiolitic and arc-related rocks of Cretaceous age by Rojas-Agramonte et al. (2016). The genesis of the igneous rocks hosting the xenocrystic zircons in the Dominican Republic, which belong to the Maimón and Los Ranchos formations, is framed in the early evolution of the Caribbean paleo-arc in an intra-oceanic scenario (Escuder-Viruete et al., 2006; Torró et al., 2016a, b, 2017a, b, c). As such, the occurrence in these rocks of inherited zircons, some of unequivocal crustal origin, raises important questions: were some segments of the Caribbean island arc built on continental substrates? Was a subcontinental lithospheric mantle of Gondwana affinity involved in the genesis of these arc magmas (cf. Kamenov et al., 2011)? Could the inherited zircons be effectively transported as detrital grains from neighboring continental masses to the nascent subduction zone (cf. Rojas-Agramonte et al., 2016)? In this study we present new in-situ SHRIMP U-Pb, oxygen and hafnium isotope data that together with the study of mineral inclusions in zircons are used to address the origin and potential provenance of the enigmatic pre-Cretaceous zircons.

2. Geological setting

2.1. General geodynamic framework

The modern Caribbean plate is suggested to be an allochthonous Farallon-derived plate fragment (Boschman et al., 2014; Pindell et al., 2012, and references therein). The progressive west to east insertion of the Caribbean plate from a Pacific into its current position (Fig 1a) was enabled by the Jurassic breakup of Pangea and the separation of the North America and Gondwana blocks. Amid

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these two drifting apart blocks, synchronous sea-floor spreading along the so-called Proto-Caribbean seaway developed an intervening oceanic basin, which was later overran by the Caribbean Plate. According to Pindell et al. (2012), the inception of the west-dipping subduction of the Proto-Caribbean lithosphere beneath the Farallon plate occurred along a sinistral inter-American transform fault at ca. 135 Ma (see also Cárdenas-Párraga et al., 2017). Such transform would have connected mature, east-dipping Cordilleran subduction zones along the western margins of North and South America. In the tectonic model of Pindell et al. (2012) hence, those HP/LT mélanges found along the Caribbean plate which are older than 135 Ma would have originated along the Benioff Zone of a long-lived east-dipping subduction system previously to the inception of the subduction of the Proto-Caribbean.

Arc magmatism in the Caribbean followed the onset of the subduction of the Proto-Caribbean in Hauterivian times (e.g., at ca. 135 Ma in central Cuba; Rojas-Agramonte et al., 2011; between 136-130 Ma in Jamaica, Hastie et al., 2009). Nevertheless, in the correlated Hispaniola (Haiti and the Dominican Republic) and eastern Cuba segments of the island-arc, arc magmatism would have apparently initiated at ca. 126-123 Ma (cf. Escuder-Viruete et al., 2014; Lázaro et al., 2016; Rojas-Agramonte et al., 2016; Torró et al. 2017a, b). Arc magmatism exposed along the Greater Antilles experienced a generalized cessation in middle Eocene (ca. 45 Ma; Lidiak and Anderson, 2015), by when the Antillean island-arc, leading the Caribbean plate in its northern edge, collided with the carbonate Bahamas platform, arresting the subduction (Mann et al., 1991). Early Cretaceous boninitic and island arc tholeiitic magmas broadly predated the calc-alkaline magmatic suite, which is more voluminous and emplaced mostly between the Late Cretaceous and Eocene times (ca. 94-45 Ma; Lidiak and Anderson, 2015). Nevertheless, a growing number of works infer various timings for the shift from island arc tholeiitic to calc-alkaline magmas from unit to unit, including, for example, Early Cretaceous calc-alkaline (Hastie et al., 2009; Torró et al., 2017c) and Late Cretaceous tholeiitic (Proenza et al., 2006; Hastie et al., 2009) rocks. This observation and the rather progressive chemostratigraphic changes observed in some Caribbean geologic units are used by some authors to disfavour an abrupt change in the affinity of the magmatism eventually caused by a polarity reversal in the Aptian-Albian (Marchesi et al., 2007; Hastie et al., 2009; Pindell et al., 2012).

2.2. The geology of central Dominican Republic

Primarily formed by island-arc growth, the island of Hispaniola is a tectonic collage of upper mantle rocks and oceanic crust units uplifted and accreted in the aftermath of the collision between the Caribbean island-arc and the Bahamas platform (Lewis and Draper, 1990; Mann et al., 1991).

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Oblique convergence between the Caribbean island-arc and the North American plate from middle Eocene to the present and associated left-lateral strike-slip tectonics developed parallel motion of tectonic blocks along the Septentrional and the Enriquillo-Plantain Garden fault zones (Fig. 1b) and the general alignment of the modern island with the Bahamas trench (Vila et al., 1987; Mann et al., 2002). Rock units in tectonic nappes comprising Hispaniola are representative of (1) ultramafic, mostly serpentinized peridotites of probable Late Jurassic or Early Cretaceous protolith age (Lewis et al., 2006), (2) volcano-plutonic assemblages representative of the Proto-Caribbean oceanic lithosphere (Escuder-Viruete et al., 2009), (3) basaltic volcanic rocks of the Caribbean Large Igneous Province (CLIP, which comprises the plume-related Caribbean-Colombian Oceanic Plateau-CCOP), of Jurassic to Late Cretaceous age (Kerr et al., 2003; Lidiak and Anderson, 2015; Escuder-Viruete et al., 2016), (4) igneous and sedimentary rocks associated to island arc(s) of Cretaceous age(Lidiak and Anderson, 2015, and references therein; Escuder-Viruete and Castillo-Carrión, 2016; Torró et al., 2016a, 2017a), and (5) high-pressure subduction complexes of Cretaceous to Oligocene age(Nagle, 1974; Krebs et al., 2008; Escuder-Viruete et al., 2011a, 2013a, b; Escuder-Viruete and Pérez-Estaún, 2013; Fernández et al., 2016), which in the northern edge of the island embrace allocthonous, North American derived terranes (Garcia-Casco et al., 2008; Escuder-Viruete et al., 2011b). These basement units are unconformably overlain by late Eocene to Holocene sedimentary rocks, which include siliciclastic and carbonate deposits. Areally restricted Quaternary alkaline volcanism is mapped across south-central Hispaniola, probably associated to the Enriquillo-Plantain Garden fault zone (Kamenov et al., 2011).

Most of the aforementioned oceanic-derived tectonostratigraphic units are represented in the Central Cordillera geological province of the Dominican Republic (and its extension to the Haitian Massif du Nord; *sensu* Lewis and Draper, 1990). They are tectonically bounded by left-lateral strike-slip, NNW-SSE to WNW-ESE trending faults, namely San Juan Restauración, Bonao-La Guácara and Hispaniola (Fig. 1c herein; Escuder-Viruete et al., 2008). In the eastern limit of the Central Cordillera, and cropping out along the Hispaniola fault zone, the serpentinized Loma Caribe peridotites occupy the core of the so-called Median belt (Lewis and Draper, 1990); their marked geochemical heterogeneity has been proposed to correlate with a supra-subduction zone position during the early evolution of the Caribbean island-arc (Marchesi et al., 2016). The peridotite body is flanked by metamorphic complexes (Fig. 1c). To the south-west, a variably metamorphosed belt conforming the Jarabacoa tectonic block (Escuder-Viruete et al., 2008) agglutinates (1) volcano-plutonic basic rocks representative of the Proto-Caribbean oceanic lithosphere (recorded in the Loma la Monja assemblage; Escuder-Viruete et al., 2009), (2) mafic igneous rocks and subordinate sediments deposited in a back-arc basin during the Early Cretaceous age(i.e., the Río Verde

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Formation; Escuder-Viruete et al., 2010), and (3) oceanic island arc basalts (OIB) and picrites likely associated to the CCOP of Early Cretaceous (i.e., the Duarte Complex) and Latest Cretaceous (non-metamorphic Siete Cabezas and Pelona-Pico Duarte formations) ages (Lewis et al., 2002; Escuder-Viruete et al., 2007, 2008, 2011c; Sandoval et al., 2015). To the north-east of the Hispaniola fault zone, a thin belt of non-metamorphic, undeformed arc-related volcanic and volcanosedimentary rocks of apparent Late Cretaceous age (i.e., the Peralvillo Sur Formation; Draper et al., 1996; Lewis et al., 2002) separates the peridotites from the metamorphic Maimón Formation (Fig. 1c). Further east, the Maimón Formation overthrusts the non-metamorphic Los Ranchos Formation, which in turn is unconformably overlain by the reefal limestones of the Hatillo Formation, of Albian to Cenomanian age (Kesler et al., 1991a; Myczynski and Iturralde-Vinent, 2005). Magmatic suites from both the Maimón and Los Ranchos formations are acknowledged to be representative of the most primitive island arc tholeiitic and boninitic magmatism of the Early Cretaceous Caribbean island-arc (Lidiak and Anderson, 2015). More detailed descriptions of both formations, sampled to develop the present study, are described below.

2.2.1. The Maimón Formation

The Maimón Formation, exposed in a 9 km wide and about 73 km long belt (Fig. 1c), is composed of bimodal mafic-felsic volcanic and volcaniclastic rocks and a thin belt of well-laminated rocks of sedimentary origin altogether deformed and metamorphosed to variable grade. Sedimentary rocks include fine-grained meta-tuffs, graphitic-shales, metacherts and marbles conformable with the volcanic sequence (Kesler et al., 1991b; Lewis et al., 2000). The rocks of the Maimón Formation are characterized by the development of syn-metamorphic ductile fabrics and structures, as observed particularly well in the Ozama shear zone, to the NW of the Fátima thrust fault (Fig. 2), in contrast to their much less intense development in the so-called El Altar zone (Draper et al., 1996). Rocks of the Ozama shear zone include greenschists and gneissic low-K meta-plagiorhyolites and are pervasively deformed and recrystallized, lacking magmatic remnants. Torró et al. (2016b) described metamorphic parageneses including chlorite, phengite, epidote, amphibole (actinolite ± winchite), albite and quartz, and constrained peak metamorphism conditions at the greenschist-blueschist facies transition (~ 8.2 kbar, 380 ºC). On the other hand, rocks from the El Altar zone include fine-grained plagioclase- and pyroxene-phyric, massive basalts and fine- to medium-grained quartz-phyric massive plagiorhyolites with limited metamorphic recrystallization to phengite and chlorite laths along with less abundant epidote and paragonite (Torró et al., 2016b).

Igneous protoliths of the Maimón Formation include mafic, intermediate and felsic rocks, with boninitic and island arc tholeiitic affinities (Lewis and Draper, 1990; Horan, 1995; Lewis et al.,

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2000, 2002; Torró et al., 2016a, b, 2017a). Taking into account lithogeochemical particularities, Torró et al. (2017a) subdivided the Maimón basaltic rocks into three subgroups: (1) lower LREE-depleted low-Ti island arc tholeiites similar to forearc basalts (cf. Reagan et al., 2010; Pearce, 2014), and overlying (2) boninites and (3) LREE-richer "normal" low-Ti island arc tholeiites. These authors also pointed to the hydrous melting of a spinel-facies depleted and/or very depleted MORB mantle types as the likely source(s) of the Maimón basalts. On the basis of chemostratigraphic relationships, Torró et al. (2017a) noted equivalence with the magmatic progression described for subduction-initiation ophiolites (cf. Whattam and Stern, 2011) and contextualized the Maimón Formation in the forearc of the nascent Caribbean island-arc, in a relatively hot subduction scenario.

The age of the Maimón Formation has been commonly assigned to the Early Cretaceous (Barremian-Early Aptian; Horan, 1995; Lidiak and Anderson, 2015; Torró et al., 2016b, 2017a) accounting their geochemical similitude to other arc units of this age described throughout the Caribbean; also, based on the proposed age for the emplacement of the first arc lavas in Hispaniola (Escuder-Viruete et al., 2014). Nevertheless, neither radiometric ages (e.g., U-Pb zircon, Ar-Ar amphibole) nor ages based on fossiliferous contents of the sediments have been reported to date. The fact that the Maimón Formation is completely fault bounded hinders its relative dating regarding stratigraphic relationships with surrounding units (Draper et al., 1996).

2.2.2. The Los Ranchos Formation

The Los Ranchos Formation crops out as a 100 km-long arched belt that extends from the Hatillo reservoir in the Central Cordillera, eastwards to the south shore of the Samaná Bay along the Eastern Cordillera (Fig. 1c herein; Bowin, 1966; Kesler et al., 1991a). With a stratigraphic thickness of more than 3 km, it is composed of bimodal volcanic, volcaniclastic and sedimentary rocks intruded by plutons and dykes of variable age and composition (Kesler et al., 1991a; Escuder-Viruete et al., 2006; Torró et al., 2017b, c). A lower basaltic sequence (i.e., the Cotuí member *sensu* Kesler et al., 1991a) is characterized by pillow lavas and minor lava flows with amygdaloidal and spherulitic textures and local interbedded fine volcaniclastic rocks. In contrast, the upper basaltic sequence (i.e., the Platanal member) is composed of massive flows of basaltic andesites with porphyritic textures. Volcaniclastic rocks in the Los Ranchos Formation (e.g., the Meladito and Zambrana fragmental members in the Pueblo Viejo district; Kesler et al., 1991a) comprise thick hyaloclastitic deposits (Torró et al., 2017c). The basaltic and volcaniclastic sequences were intruded by Cretaceous gabbros (e.g., the Doña Ruth gabbro; Torró et al., 2017c), tonalite plutons (e.g., Zambrana) and plagiogranite stocks (Kesler et al., 1991a, 2005a, b; Escuder-Viruete et al., 2006; Torró et al., 2017b, c), the latter locally associated with acid flows (i.e., the Quita Sueño member;

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Fig. 2). Volcanogenic and terrestrial sediments of the Pueblo Viejo member with Early Cretaceous plant fossils (Smiley, 2002) cap the Los Ranchos Formation as observed particularly well in the Pueblo Viejo mining area and in Loma Guaymarote, near Bayaguana (Kesler et al., 1991a; Torró et al., 2017b, c). In both districts, volcanic, volcaniclastic and sedimentary deposits were intruded by diorite and monzodiorite dykes and sills (Mueller et al., 2008); volcanic domes of intermediate and acid compositions also cut the sediments even if their carapaces of auto-breccia are locally interstratified with the upper portion of the epiclastic sediments (Nelson, 2000; Escuder-Viruete et al., 2006; Torró et al., 2017b).

Geochemically, the lower basalts of the Los Ranchos Formation are LREE-depleted low-Ti island arc tholeiites (Escuder-Viruete et al., 2006 and Torró et al., 2017b, c) though basalts with boninite-like compositions are locally present. On the other hand, the upper basalts and basaltic andesites have compositions similar to both normal and low-Ti island arc tholeiites with moderate to strong enrichment in LREE. Tonalites, plagiogranites and plagiorhyolites are of tholeiitic affinity (Escuder-Viruete et al., 2006; Torró et al., 2017b, c). In addition, some diorites have intermediate compositions between island arc tholeiitic and calc-alkaline affinities (Torró et al., 2017c), whereas monzodiorites and domes of intermediate composition crosscutting sediments of the Pueblo Viejo member are calc-alkaline. Based on chemostratigraphic reconstructions, Escuder-Viruete et al. (2006) and Torró et al. (2017c) proposed that Los Ranchos magmas record a magmatic evolution from subduction-initiation to true-subduction.

The age of the Los Ranchos Formation has been extensively addressed by means of paleontological and radiometric methods. Marine and terrestrial plant fossils from the Pueblo Viejo member indicate a Cretaceous middle Aptian to middle Albian age (Bowin, 1966; Smiley, 2002). U-Pb zircon dating constrains the age of the basaltic rocks between 122 and 112 Ma, and of the intrusion of the Doña Ruth gabbro between 118 and 116 Ma (Torró et al., 2017c). U-Pb zircon ages of plagiogranite stocks range between ca. 113 and 109 Ma (Kesler et al., 2005a; Torró et al., 2016a, 2017c), similar to U-Pb zircon ages for the Zambrana tonalite batholith, which are in the range between 113 and 110 Ma (Torró et al., 2017c); on the other hand, $^{40}Ar/^{39}Ar$ hornblende plateau ages for the same tonalite body were constrained at 109-106 Ma and interpreted as cooling ages by Escuder-Viruete et al. (2006). Based on U-Pb zircon ages, Mueller et al. (2008) determined the emplacement of a diorite body cutting sedimentary rocks of the Pueblo Viejo member at 109.6±0.6 Ma, and Torró et al. (2017c) indicated that transitional and calc-alkaline diorites and monzodiorites emplaced between 110 and 106 Ma; an emplacement at ca. 110-107 Ma of plagiorhyolite domes cutting that sedimentary member near Bayaguana has been reported by Torró et al. (2017b).

3. Samples and analytical methods

Zircon grains analyzed for this paper were extracted from rocks of the Maimón and Los Ranchos formations. Rock samples from the Maimón Formation (n = 2) are representative of meta-plagiorhyolitic rocks of unknown age. Samples from the Los Ranchos Formation (n = 14) were collected from the Pueblo Viejo district, in the Central Cordillera, and the Bayaguana district, in the Eastern Cordillera (Fig. 1c). Samples from the Pueblo Viejo district include tholeiitic basalt (n = 3), gabbro (n = 1), diorite (n = 2), tonalite (n = 1) and plagiogranite stock (n = 2), as well as calc-alkaline monzodiorite (n = 3) rocks (Fig. 2). Magmatic crystallization ages for rocks from the Pueblo Viejo district and individual U-Pb ages of the Cretaceous zircons they contain are reported in Torró et al. (2017c). For samples from the Bayaguana district, oxygen isotope data are provided for zircon grains hosted in felsic domes (n = 2). U-Pb ages of zircons from the samples from Bayaguana (including those of inherited zircons) were reported by Torró et al. (2017b). The samples are listed in Table 1, and the location of those from the Central Cordillera is shown in Fig. 2. The location of the studied samples from felsic domes in the Bayaguana district is shown in Fig. 2 in Torró et al. (2007b).

Rock samples were crushed and milled using a tungsten carbide mill, and the resulting material was sieved (grain fractions: 125, 100, 75 and 40 μm) by means of disposable sieves in order to avoid contamination. Zircon crystals were separated using panning in water. Non-magnetic concentrates, after eliminating the magnetic fractions with a Frantz® isodynamic LB-1separator, were processed by applying the hydroseparation technique (HS) at the HS-11 laboratory of the University of Barcelona to obtain high-density mineral concentrates. The resulting non-magnetic high-density concentrates went through acid digestion in open bombs with combinations of HF, HCl and HNO_3. Zircons were then handpicked under the binocular microscope and cast on “megamounts” (i.e., 35 mm epoxy discs fixed on the front of a mount holder so that no metallic parts or surface discontinuities faced the secondary ions extraction plate). Once mounted and polished, zircon grains were studied by optical and cathodoluminescence imaging prior to U–Th–Pb and oxygen isotope analyses.

Zircon grains were analyzed for U–Th–Pb and oxygen isotopes using a SHRIMP IIe/mc ion microprobe (IBERSIMS, University of Granada) following the methodology explained in APPENDIX 1. SHRIMP U–Th–Pb and SHRIMP oxygen results are presented in APPENDIX 2 and APPENDIX 3, respectively, and in Figs. 3 to 7. In addition, a second set of zircons separated from the sample APV-Diorite (Table 1) was dated by means of LA-ICP-MS at the Institute of Geology of the Czech Academy of Sciences, Prague (methodology explained in APPENDIX 1; results

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presented in APPENDIX 4). Concordia plots and calculated concordia and/or discordia and/or average ages for individual rock samples are gathered in APPENDIX 5. Hf-in-zircon isotope compositions were measured in situ, mostly on dated grains and on the same U-Pb spot, at the MC-LA-ICP-MS laboratory of the State Key Laboratory of Lithospheric Evolution, Institute of Geology and Geophysics, Chinese Academy of Sciences, Beijing; the analytical technique is described in APPENDIX 1 and the results are presented in APPENDIX 6. Cathodoluminescence (CL) images with spot SHRIMP U–Pb ages, δ^{18}O and εHf data are presented in APPENDIX 7. Mineral inclusions in zircon grains were studied by means of scanning electron microscopy and Raman spectroscopy (see APPENDIX 1 for details).

4. Results and interpretation

4.1. Cretaceous zircons

One single zircon grain, out of 69 grains analyzed for rocks from the Maimón Formation, yield a concordant ^{206}Pb/^{238}U Cretaceous age: 116.4 ± 1.7 Ma. This age hence represents the first absolute date constraining the minimum age of crystallization of igneous rocks from the Maimón Formation to the Early Cretaceous. On the other hand, it points out that the magmatism recorded in the Maimón and Los Ranchos formations were broadly coeval, even though these two units likely occupied different positions along the newly born Caribbean island-arc taking into account lithogeochemical and metamorphic characteristics (Torró et al., 2016b, 2017a).

The single Cretaceous zircon found in rocks from the Maimón Formation yields an δ^{18}O value of 5.01 ± 0.07 ‰. Cretaceous zircons from the Los Ranchos Formation yield δ^{18}O values (n = 58; APPENDIX 2) between 3.34 ± 0.07 and 9.10 ± 0.11 ‰ (n = 57; Fig. 7a), and εHf(t) values between -9.7 and +17.4. Within these ranges, zircons from basaltic rocks of tholeiitic affinity (comprising volcanic and intrusive lithotypes) yield δ^{18}O values between 4.28 ± 0.09 and 5.90 ± 0.06 ‰ and εHf(t) values between 6.3 and 17.4. The lightest oxygen isotopic values are registered in zircons from tonalite (δ^{18}O between 3.34 ± 0.07 and 5.34 ± 0.08 ‰) and plagiogranite stock (δ^{18}O between 3.98 ± 0.05 and 6.42 ± 0.07 ‰) rocks; the later yield εHf(t) values between 7.9 and 16.6. Zircons from calc-alkaline monzodiorites and rhyolitic domes yield the heaviest oxygen isotopic signatures (δ^{18}O between 4.22 ± 0.10 and 9.10 ± 0.11 ‰) and εHf(t) values between -9.7 to 16 (Fig. 7a).

A third of the studied Cretaceous zircons fall within the δ^{18}O 5.3 ± 0.3 ‰ range (accounting the associated analytical errors) suggesting hence crystallization from juvenile magmas (Valley et al., 1998). This constraint is in good agreement with the positive εHf(t) values in most of such zircons (cf. Belousova et al., 2006). On the other hand, higher δ^{18}O values (Fig. 7a) may reflect variable

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contribution of sedimentary components and/or of ocean crust altered at low temperature by surface waters into their parental magmas as postulated by Valley (2003) and Valley et al. (2005). The relatively wide variation in εHf(t) positive values (APPENDIX 7), from near depleted mantle (DM; Fig. 7b) values to 3.0, indicate considerable Hf isotopic heterogeneity in the zircon source probably involving mixing of old crust and mantle-derived material. The highest $\delta^{18}O$ and the lowest εHf(t) values were analyzed in zircons from rocks with calc-alkaline signatures (Fig. 7), suggesting magmas derived through melting of mantle with a component of old (or recycled) continental crust. These observations are in agreement with the conclusions reached by Torró et al. (2017a, 2017c) on the source of Maimón and Los Ranchos basalts during the subduction initiation stage of the Caribbean island-arc involving an increasing contamination of the mantle source by fluids/melts expelled from the downgoing slab as subduction progressed (see also McGowan et al., 2015).

Nearly a third of the Cretaceous zircons yield $\delta^{18}O$ values below the average values of juvenile (mantle-derived) zircons (Fig. 7a). Valley et al. (2005) postulated that such low $\delta^{18}O$ magmas may originate from melting of hydrothermally altered wall rocks, for example, in the volcanogenic massive sulfide (VMS) environment. Both the Maimón and the lower portion of the Los Ranchos formations are known for hosting VMS mineralization (Andreu et al., 2015; Torró et al., 2016a, 2017c), thus pointing to the widespread activity of hydrothermal cells during their deposition. Finally, it is noteworthy that the lowest $\delta^{18}O$ zircon values are recorded in Cretaceous zircons from acid intrusives, which according to Escuder-Viruete et al. (2006) and Torró et al. (2017a, c) might have formed after partial melting of the basaltic roots of the island-arc likely registering generalized hydrothermal alteration.

4.2. Pre-Cretaceous zircons

The presence of pre-Cretaceous (i.e., xenocrystic or *inherited*) zircons in Los Ranchos tholeiitic basalts and basaltic andesites (and their intrusive counterparts) is variable from sample to sample. Their very high proportion in meta-plagiorhyolite rocks from the Maimón Formation is outstanding. In stark contrast, no pre-Cretaceous zircons have been found in plagiogranite rocks from the Los Ranchos Formation, and only one pre-Cretaceous zircon was dated in samples from the Zambrana tonalite batholith. The concentration of inherited zircons is steadily high in calc-alkaline monzodiorite rocks (16 out of 17 zircon grains). In felsic domes from the Bayaguana area (Fig. 2), Torró et al. (2017b) found two pre-Cretaceous (out of 90) zircons. Despite these apparent variations in the proportion of inherited zircons between the two geological units and from lithotype to lithotype, zircon morphologies and U-Pb age populations are mostly repetitive in the whole suite of studied rocks (Table 1). As a rule, pre-Cretaceous zircon grains show a quite wide variety of sizes,

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shapes (Fig. 3) and mineral inclusions (Fig. 4), which are described in detail below. Cathodoluminescence images reveal that most of the studied zircons have complex internal structures including conspicuous oscillatory zoning and overgrowths pointing to several stages of growth; often, the very narrow character of the outer rims (< 10 μm) prevented their analysis.

A total of 203 U-Pb individual spot analyses yield concordant and near-concordant U-Pb ages range between 256.2 ± 8.9 ($^{206}Pb/^{238}U$) and 2937.5 ± 8.0 Ma ($^{207}Pb/^{206}Pb$; Figs. 5 and 6; APPENDICES 2 and 4). Indeed, a pronounced gap of ca. 130 My is observed between the age of the oldest Early Cretaceous zircon (ca. 135 Ma; Torró et al., 2017b, c; i.e., syn-magmatic crystallization of the hosts + antecrystic ± xenocrystic zircons) and the youngest pre-Cretaceous one (i.e., xenocrystic zircons). Most of the studied zircons are Permian-Carboniferous (n = 165, between 256.2 ± 8.9 and 361.0 ± 11.5 Ma, with a clear age peak at ca. 300 Ma; Fig. 6) and Proterozoic (n = 37, between 558.4 ± 6.9 and 2470.3 ± 9.8 Ma); two zircon grains yield Archaean ages. For the whole dataset, Th/U ratios span between 0.1 and 2.5 (av. 0.55), and no correlation is observed between these values and the respective U-Pb spot ages. The relatively high Th/U, mostly >> 0.1 in the studied zircons, is suggestive of a magmatic crystallization, in contrast with the low Th/U (< 0.1) of most metamorphic (or grown by solid state reactions) and hydrothermal (< 0.02) zircons (e.g., Schaltegger et al., 1999; Rubatto, 2002; Harley et al., 2007; Wang et al., 2011).

Permian-Carboniferous zircons are euhedral and have grain sizes between 40 and 200 μm across (Fig. 3, APPENDIX 7). They show the most marked disparity in morphologies, which range from stubby to very thin crystals and show aspect ratios (length/width) from ~ 1.5 to 6 (Fig. 3). Their $\delta^{18}O$ values range between 6.16 ± 0.10 and 12.67 ± 0.11 ‰, and are hence systematically higher than average values of zircons crystallized in mantle-like melts (Fig. 7a). Their εHf(t) values are mostly negative, between -11.9 and -0.9, with a single positive value of 1.2 (Fig. 7b). Therefore, both oxygen and hafnium isotope data suggest crystallization in equilibrium with an "evolved" magma involving variable amounts of crustal contributions (Valley et al., 2005; Belousova et al., 2006). Permian-Carboniferous zircons show abundant mineral inclusions of apatite, quartz and much less abundant potassium feldspar and muscovite as well as devitrified glass. Inclusions of individual crystals or composite assemblages of quartz, potassium feldspar and muscovite, which are consistent with crystallization from a granitic or syenitic liquid, are particularly abundant in zircons with relatively high $\delta^{18}O$ and low (negative) εH(t) (Fig. 4).

Earliest Ordovician to Proterozoic zircons also show variable sizes in the range of 50 to 200 μm across (Fig. 3). Their morphologies are rather stubby, with aspect ratios between ~ 1.5 and 3. They present blunt tips and, often, overall rounded shapes that contrast with the euhedral, sharp zircon

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crystals of Permian-Carboniferous age. Their $\delta^{18}O$ values range between 4.93 ± 0.09 and 9.21 ± 0.07 ‰, and their εH(t) values between -19.9 and 8.5 (Fig. 7), hence indicating crystallization in equilibrium with either mantle-derived (juvenile) magmas or magmas involving supracrustal rocks (Valley et al., 2005; Belousova et al., 2006). Mineral inclusions are of dominant apatite, and of less abundant quartz and muscovite as observed particularly well in zircons with high $\delta^{18}O$ and low εH(t) (Fig. 4).

Two zircons yield Archean ages. In one of them, an age of 2519.9 ± 18.4 Ma was determined in the core of the grain, which shows an outer overgrowth domain dated at 2369.5 ± 9.2 Ma. In the other, the core was dated at 2937.5 ± 8.0 Ma and one of the overgrowth domains, at 2856.7 ± 13.0 Ma (Fig. 3). The second of these grains presents a subhedral, equant morphology and markedly blunt tips. Zircon domains with Archean ages yield similar $\delta^{18}O$ values in the range 6.98 ± 0.06 to 7.94 ± 0.07 ‰ (i.e., they are higher than the average of juvenile zircons; Fig. 7a) and negative εHf(t) values between -18.5 and -1.6 (Fig. 7b). These data reflect a contribution of supracrustal materials in the parent magmas. It is important to highlight that our $\delta^{18}O$ value near 8 ‰ exceeds the 7.5 ‰ maximum of typical magmatic Archean zircons (Valley et al., 2005); nevertheless, Archean zircons with even higher $\delta^{18}O$ values have been recently reported by Ashwall et al. (2017). Mineral inclusions were not found in Archean zircons.

5. Discussion

The conspicuous occurrence of pre-Cretaceous inherited zircons in Aptian-Albian arc-related igneous rocks from Hispaniola demands assessment of their origin and provenance. Also of the unlikely survival (e.g., Watson and Harrison, 1983; Watson, 1996) of those xenocrystic zircons in high liquidus temperature, Zr-poor melt carriers as are island arc tholeiitic and boninitic magmas of the Maimón and Los Ranchos formations (Torró et al., 2017a, b, c). Answering this latter question is, however, beyond the scope of this study, in part because zircon grains were isolated for concentration purposes and any textural relationship with other rock-forming minerals is missing. Ultimately, what we regard as more significant is their general occurrence along the Caribbean volcanic arc and ophiolitic rocks (see Rojas-Agramonte et al., 2016).

5.1. On the origin of pre-Cretaceous zircon grains

A number of works have documented the occurrence and delved into the origin of exotic/inherited zircon grains whose U-Pb ages significantly predate the age of crystallization of their igneous hosts (e.g., Bea et al., 2001; Siebel et al., 2009; Stern et al., 2010; Yamamoto et al., 2013; Wang et al., 2014; McGowan et al., 2015; Li et al., 2016; Rojas-Agramonte et al., 2016; Ashwal et al., 2017).

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For a juvenile intra-oceanic environment, zircon inheritance is explained either by direct incorporation from the mantle source of the carrier magmas (cf. Rojas-Agramonte et al., 2016) or by the passage of such magmas through fragments of ancient continental crust during their rising (cf. Ashwal et al., 2017). Accounting the paleotectonic framework of the Early Cretaceous Caribbean island arc, both possibilities deserve attention.

The break-up of Pangea previous to the inception of the Proto-Caribbean subduction in Early Cretaceous time could have reasonably left a string of continental fragments along the newly created intervening oceanic domain between the Americas. These ancient continental strips might have acted as (micro-)continental basement(s) of proto- and first- island-arc lavas during the subduction initiation of the Caribbean arc. Nevertheless, on-land geologic and geophysical research has never raised evidence for a continental basement in Hispaniola or any other segment of the Greater and Lesser Antilles island arcs (Iturralde-Vinent, 1996; García-Lobón and Ayala, 2007; Neill et al., 2010; Smith et al., 2013). Instead, geological observations identify pre-arc basement in Hispaniola as of oceanic nature and likely representing fragments of the Proto-Caribbean oceanic crust (Escuder-Viruete et al., 2009). Analogous conclusions are put forth for pre-arc basements in Puerto Rico (Greater Antilles), Trinidad and La Désirade (Lesser Antilles) islands (Mattinson et al., 2008; Neill et al., 2010, 2014; Lidiak et al., 2011; Marchesi et al., 2011). On the other hand, a noticeable assimilation of continental crust materials during the petrogenesis of Early Cretaceous island arc tholeiitic and boninitic magmas has not been reported for the Caribbean island arc (Lidiak and Anderson, 2015). Even in Tertiary volcanic arc rocks of eastern Cuba emplaced after latest Cretaceous arc-continent collision (see Garcia-Casco et al., 2008), the influence of continental basement is lacking (Rojas-Agramonte et al., 2004).

Interestingly though, Kamenov et al. (2011) proposed the entrainment of fragments from an ancient subcontinental lithospheric mantle (SCLM) in the roots of the Caribbean island-arc after identifying a strong mantle component similar to the enriched mantle 1 (EM1) isotopic reservoir in Quaternary alkaline basalts in south-central Hispaniola. Accordingly, Kamenov et al. (2011) postulated that such basalts derived from melting of a SCLM of likely Gondwana affinity that, incidentally, had undergone melt extraction during a Grenvillian-age event. However, an EM1 isotopic signature has not been identified for Cretaceous arc magmas in the Greater Antilles (Lidiak and Anderson, 2015) and indeed the main mantle sources of basaltic rocks from the Maimón and Los Ranchos formations were a depleted and/or very depleted MORB mantle types (Escuder-Viruete et al., 2006; Horan, 1995; Torró et al., 2016, 2017a, b, c). This observation suggests that interactions with such an alleged Gondwanan SCLM were limited or nonexistent during Cretaceous arc magmatism in the region. Coinciding with Kamenov et al. (2011), we envisage that an extensive entrainment of a

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Gondwanan SCLM could become effective only during the west to east migration of the Caribbean island-arc and, in consequence, its geochemical imprint in, and the potential transference of inherited zircon to the studied early magmatic products of subduction initiation are unlikely.

Alternatively, and as proposed by Stern et al. (2010) and Rojas-Agramonte et al. (2016), Cretaceous arc-related carrier magmas could have loaded the inherited zircon cargo directly from their predominantly depleted MORB-type mantle sources. In the knowledge that most inherited zircons host mineral inclusions that recall crustal assemblages (Fig. 4), and that only a few of them yield $\delta^{18}O$ values comparable with those of juvenile zircons (Fig. 7a), such a hypothesis should account for the occurrence of zircons with crustal signatures in the asthenospheric mantle wedge below the newly born Caribbean island-arc. The occurrence of crustal, recycled zircons in the mantle is actually quite amply described from the benchmark work of Bea et al. (2001). These researchers determined a crustal origin for zircons found in dunites from Kytlym (Urals belt), a subduction-related dunite-clinopyroxenite-gabbro massif, and proposed that they were introduced in the mantle source via melting of subducted sediments. Likewise, Yamamoto et al. (2013) explain the occurrence of old zircons in Tibetan chromitites as a consequence of recycling of continental crust material in the upper mantle. The influence of subducted sediments in the petrogenesis of arc magmas has been identified in Cretaceous island-arc rocks of the Greater Antilles in general (Lidiak and Anderson, 2015), and in the Maimón and Los Ranchos formations in particular (Escuder-Viruete et al., 2006; Torró et al., 2016a, b, 2017b, c). Importantly, an incorporation of detrital zircons from subducted sediments to the mantle wedge would also account for the occurrence of pre-Cretaceous, inherited zircons with a similar age span (Fig. 6) and isotopic signatures in Cuban ophiolitic gabbros and volcanic arc rocks described in Rojas-Agramonte et al. (2016).

5.2. On the provenance of pre-Cretaceous zircon grains: primary and secondary suppliers

In order to trace back the primeval provenance of inherited zircons found in Maimón and Los Ranchos igneous rocks, geologic events and particular units representing potential primary suppliers have been scrutinized accounting: 1) the temporal coincidence with the ages of the studied inherited zircons, 2) a suitable paleo-geographic location previous to and during the subduction initiation at the Caribbean island arc, and 3) geologic/geodynamic environments compatible with a supracrustal, continental magmatism. Given the inception of the Caribbean island-arc in a near equatorial position of the paleo-Pacific margin along the western flank of the Americas, we will focus our attention on Mexican and northwestern Colombian terranes (Fig. 8).

Permian-Carboniferous zircons represent, by far, the most abundant population among the inherited zircons studied in this work. It is noteworthy that our peak age at 300 Ma (Fig. 6) is virtually the

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same as the peak registered in the global granitoid and detrital zircon record attributed to the assembly of the supercontinent Pangea (Condie and Aster, 2010; Condie et al., 2014). However, in the western flank of Pangea, it is the subduction of the paleo-Pacific plate during the Permian-Carboniferous what caused a major magmatic event in the form of a continental arc along the western margin of Gondwana, as it has been documented in Mexico (Centeno-García, 2005, 2017; Rosales-Lagarde et al., 2005; Nance et al., 2006; Ortega-Obregón et al., 2014) and Colombia (Cardona et al., 2010a; Leal-Mejía, 2011; Leal-Mejía et al., 2011). In Mexico, Permian-Carboniferous granitoids intruded between ca. 311 and 255 Ma (Keppie et al., 2004; Kirsch et al., 2012, 2013; Ortega-Obregón et al., 2014). In the northern Colombian Andes (El Bagre sector), Leal-Mejía (2011) and Leal-Mejía et al. (2011) have dated diorite and leucogranite intrusions at ca. 330-310 Ma. On the other hand, Permian arc magmatism is bracketed between ca. 288 and 265 Ma in the Santa Marta Region (Cardona et al., 2010a), and between ca. 275 and 260 Ma in Central Cordillera granitoids (Leal-Mejía et al., 2011).

The presence of zircons with Neoproterozoic, Pan-African/Braziliano ages in particular (Fig. 6) concurs with a Gondwana source (Nance and Murphy, 1996; Kröner and Stern, 2004; Adams et al., 2011). Neoproterozoic magmatic suites are described, even if poorly understood, in the crystalline basement of the Maya Terrane (sensu Centeno-Garcia, 2017; see also Barboza-Gudiño et al., 2010). From the geochronological study of granitoid cobbles and boulders in Paleozoic conglomerates from northeastern Mexico (state of Coahuila), Lopez et al. (2001) concluded that, in addition to dominant Grenvillian-age basement (U-Pb zircon ages at ca. 1232 to 1214 Ma, with inheritance ages of ca. 1850 Ma), Pan-African-age basement (U-Pb zircon ages at ca. 580 Ma) is present in that sector of Mexico. With regards to the Early Paleozoic magmatism in the region, Ordovician granitoids are dated in the Mexican Acatlán Complex between ca. 478-440 Ma (Talavera-Mendoza et al., 2005). In the Colombian Andes, Ordovician magmatism is recorded in the Cajamarca-Valdivia terrane, which is described as a parautochthonous island arc sutured to continental South America in Middle Ordovician to Silurian times (Cediel et al., 2003; Cediel., 2011).

Grenvillian-age crust (ca. 1300 to 900 Ma) is found in the Oaxaca complex (which contains the largest exposures of the Oaxaquia micro-continent, of Gondwanan affinity; Ortega-Gutiérrez et al., 1995; Centeno-García, 2005, 2017; Solari et al., 2014) and in northern Colombian inliers, including the Garzón, Santa Marta and Santander massifs, as well as the Guajira peninsula (Cordani et al., 2005; Cardona et al., 2010b). Grenvillian-age basement rocks (fragments from the rifted North American margin) were also reported in central Cuba (Socorro Complex; Renne et al., 1989). In general, there is consensus that such Mexican and Colombian units are remnants of a fragmented Grenvillian belt (Ruiz et al., 1999).

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It follows that both Mexican and northwestern Colombian terranes host igneous suites formed coevally to most of the inherited zircons reported in our study, and that might represent their primary, magmatic suppliers in which they crystallized. Unfortunately, the fact that Archean basement is not described neither in Mexico nor in northwestern Colombia makes it difficult, with the available data, to trace the provenance of our inherited Archean zircons. Nearby exposures of Archean igneous rocks restrict to the North American (north of the Ouachita orogenic belt) and Amazonian cratons. Multiple zircon inheritance during successive magmatic events operating in progressively younger substrates could explain the presence of the inherited Archean zircons, which, on the other hand, are conspicuously scarce in the studied rocks (Figs. 5, 6). Alternatively or complementarily, Precambrian and/or younger exogenous dynamics could have transported detrital Archean zircons to more proximal positions with regards to the area of influence delimited in our study. For example, detrital zircons as old as 3451 Ma have been described in Paleozoic sedimentary rocks of the Acatlán Complex in southern Mexico, and are interpreted to be derived from Archean Amazonian crust during the Mesoproterozoic or either from North American or Amazonia Archean crusts during the Permian-Carboniferous (see Fig. 7 of Talavera-Mendoza et al., 2005; see also Fig. 7 of Centeno-García, 2005).

Detrital zircons with Proterozoic and Phanerozoic ages of interest are also described in Paleozoic and Mesozoic sedimentary units of Mexico and northwestern Colombia. For example, Cardona et al. (2006) dated detrital zircon grains from a paragneiss of the Paleozoic Sevilla Complex, in the northern termination of the Colombian Andes, obtaining concordant ages between ca. 1400 and 500 Ma. In Mexico, old detrital zircons in sedimentary rocks from the Tiñu (Cambrian-Ordovician), Santiago (Mississippian) and Ixtaltepec (Pennsylvanian) formations of the Oaxaca Complex yield concordant ages mostly between ca. 1220 and 980 Ma (Gillis et al., 2005; Solari et al., 2014). Detrital zircons from Paleozoic sedimentary rocks of the Acatlán Complex yield ages in the range from ca. 300 to 2500 Ma with clear probability highs at Grenvillian and Pan-African ages (Talavera-Mendoza et al., 2005; Vega-Granillo et al., 2009; Kirsch et al., 2012).

Of special interest in our study are detrital zircons hosted by sedimentary rocks of the so-called Potosí fan described in Mexico, which deposited on the paleo-Pacific margin of Pangea in Late Triassic time (Fig. 8; Centeno-García, 2017). Such detrital zircons draw three main populations that matches ours at Grenvillian, Pan-African and more abundant Permian ages (Barboza-Gudiño et al., 2010). These authors concluded that sediments deposited in the Potosí fan were transported by a fluvial system (El Alamar) draining extensive areas of west equatorial Pangea (mostly Mexican terranes) including Grenvillian (Oaxaquia), Pan-African (Yucatan and Gondwana) and Paleozoic (Acatlán Complex and Permian-Triassic arc; see Fig. 14 of Barboza-Gudiño et al., 2010) sources.

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Detrital zircon transport in the Alamar-Potosí fluvial-fan delta and other sedimentary systems discharging in central paleo-Pacific would have been conductive to the incorporation of ancient detrital zircons in subduction dynamics operating in the area during the Triassic-Jurassic (Centeno-García, 2017 and references therein) and the Early Cretaceous (Pindell et al., 2012) as discussed below.

It is noteworthy also to recall here that most of our zircon grains with ages older than Carboniferous show differential rounded shapes relative to Permian-Carboniferous ones (Fig. 3; APPENDIX 7). The paleo-geographic upstream course of Mesozoic alluvial systems such as El Alamar would have favored the prominent erosion of proximal Permian-Carboniferous arc rocks (i.e., the likely source of our dominant zircon population; Fig. 6) rather than basement rocks containing older zircons. On the other hand, such alluvial systems could have incorporated older zircons from relative proximal ancient sedimentary rocks (e.g., Acatlán or Oaxaca complexes in Mexico; Sevilla Complex in Colombia) or from distal igneous sources both possibilities accounting for higher erosive modifications (i.e., roundness) of such zircons.

5.3. Deciphering the whole mechanism: conveyors and elevators

Lines of evidence discussed above suggest 1) that arc magmas in the Maimón and Los Ranchos formations picked up the inherited zircon cargo from their mantle sources, and 2) that potential primary (igneous) and secondary (sedimentary) suppliers of such zircons exist in continental Americas. Encompassing these two observations implies that the studied pre-Cretaceous zircons were transported downward from crustal levels to the asthenospheric mantle previously to their entrainment in Early Cretaceous arc magmas. Such transport can be contextualized in the process of transference and recycling of crustal materials into the mantle wedge above the Benioff zone of the subducting Proto-Caribbean slab from ca. 135 Ma (Rojas-Agramonte et al., 2011, 2016; Pindell et al., 2012) and/or any other previous subduction zone operating nearby the central segment of the western margin of Pangea from as early as Permian-Carboniferous (cf. Ortega-Obregón et al., 2014, and references therein).

The transference and recycling of sediments with detrital zircons of interest atop the subducting Proto-Caribbean oceanic crust to the asthenospheric mantle below the Greater Antilles paleo-arc is extensively discussed in Rojas-Agramonte et al. (2016; see their Fig. 1). We should note that basalts from the Maimón and lower Los Ranchos formations were emplaced in the forearc and axial arc regions of the Caribbean paleo-island arc during the subduction initiation stage (Lidiak and Anderson, 2015; Torró et al., 2017a). Therefore, such a hypothesis would imply that the studied detrital zircons were “available” in the western edge of the Proto-Caribbean slab, next to the inter-

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American transform fault (Pindell et al., 2012), by the time it sank into the asthenospheric mantle (i.e., previously to its true, steady-state subduction).

Alternatively, sediments of continental origin deposited along the Pangea-Pacific margin during Paleozoic and early Mesozoic could have been delivered to the mantle along successive subduction zones operating previously to the inception of the Proto-Caribbean subduction, as exemplified in Figure 8a-c. For example, the consumption of oceanic crust of the Arteaga basin and its sedimentary cover (Potosí fan) in Mexico has been documented to occur in the Late Triassic (Centeno-García, 2017, and references therein). Noteworthy, the tectonic models of Pindell and Kennan (2009) and Pindell et al. (2012) envisage that the sinistral component of the east-dipping subduction of the Farallón (Pacific) plate would have translated in a southward relative transport of geological materials along the subduction channel(s). Such process could hence lead to the contamination of the mantle below the Proto-Caribbean slab with North American-derived sediments containing the ancient zircons. During the embryonic stages of the Caribbean island arc (Fig. 8d), the sinking of the Proto-Caribbean slab into the asthenospheric mantle and its subsequent roll-back would have triggered the rising of deeper mantle materials contaminated with the detrital zircons of interest in the fore-arc of the arc system (McGowan et al., 2015; Griffin et al., 2016); also the transfer via return and slab-window flow of Proto-Caribbean (Atlantic) asthenospheric mantle to fill the newly created mantle wedge (Torró et al., 2017a; see also Pearce and Robinson, 2010 and Pearce et al., 2014). Further, Pindell et al. (2012) propose that during Neocomian times, any material entering the subduction channel in western Mexico was potentially dragged southeastward throughout the inter-American transform along which the subduction of the Proto-Caribbean took place (Fig. 8c); therefore, that material would have found itself in the hanging wall of the new west-dipping subduction zone right in the moment of the proto-Caribbean subduction initiation. Coevally to that process, the Maimón and Los Ranchos basaltic magmas would have picked up the pre-Cretaceous zircon cargo from their shallow asthenospheric mantle sources.

6. Conclusions and future developments

Our work reveals a conspicuous presence of ancient, Permian to Archean detrital zircon grains in Early Cretaceous oceanic arc-related rocks from central Hispaniola, and aligns with previous findings in analogous rocks and ophiolitic gabbros from Cuba. Oxygen and hafnium isotope relations measured on zircons and a detailed study of their mineral inclusions reveal a prevalent crustal origin for most of the detrital zircons, and potential primary (igneous) and secondary (sedimentary) zircon suppliers are identified in Mexican and northwestern Colombian terranes. Geologic evidence points to the transfer of detrital zircons to the asthenospheric mantle below the

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Caribbean paleo-arc along with sediments of continental origin previously to or during their entrainment in arc magmas from ca. 135 Ma. We suggest that the subducting Proto-Caribbean slab and/or subduction zones operating in the western margin of north-equatorial Pangea acted as the conveyor(s) that transported downward, and likely southward, the detrital zircon cargo previously and/or during the subduction initiation of the Caribbean island-arc.

Hopefully, the wealth of information on zircon inheritance in the Caribbean arc will keep growing and embrace other arc-related and ophiolitic rocks including also those of Late Cretaceous age, as well as plume-related rocks of the Caribbean large igneous province (CLIP). Such information should shed further light on the nature and origin of the mantle heterogeneities in the Caribbean realm and on the early tectonic relationships of tectonic blocks around the Proto-Caribbean and early Caribbean plates.

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Figure and table captions

Table 1. Sample locations, lithologies and geochemical affinities of rocks from the Maimón and Los Ranchos formations hosting zircon grains studied in this article.

Figure 1. (a) General tectonic map of the Caribbean region showing major faults (adapted from Lidiak and Anderson, 2015). (b) Main fault zones in the island of Hispaniola above an image from Google Earth©. Red box indicates the geographic location of the geological map shown in c. (c) Simplified geological map of the eastern segment of Central Cordillera and the Eastern Cordillera of the Dominican Republic (modified from Toloczyki and Ramirez, 1991 and Escuder-Viruete et al., 2010). S = Septentrional fault zone; H = Hispaniola fault zone; BG = Bonao-La Guácara fault zone; SJR = San Juan Restauración fault zone; EPG = Enriquillo-Plantain Garden fault zone.

Figure 2. Geologic map of the Cotuí-Maimón area in central Dominican Republic (see geographic location of the area in Fig. 1), showing the location of the studied rock samples. When available, the crystallization age of the igneous rock hosting the studied zircon grains is indicated ([1]age according to Torró et al., 2017c). Geological map adapted from Martín and Draper (1999), Nelson (2000) and Torró et al. (2017c).

Figure 3. Representative zircon cathodoluminescence images of inherited zircon grains from the Maimón and Los Ranchos formations. Yellow circles and text denote spot U-Pb ages, blue circles and text, $\delta^{18}O$ values, and green circles and text, εHf(t) values. A complete record of CL images with spot U-Pb, $\delta^{18}O$ and εHf(t) values is given in the supplementary material.

Figure 4. Representative Raman spectra of mineral inclusions within inherited zircons from the Maimón and Los Ranchos formations. The rock sample, U-Pb zircon age, the $\delta^{18}O$ and the εHf(t) of each hosting zircon is also given. Yellow filled circles indicate the location of laser analyses. Abbreviations: Ap = apatite; Ms = muscovite; Or = orthoclase; Qz = quartz; Zrn = zircon.

Figure 5. Wetherill concordia plot of all inherited zircons from the Maimón and Los Ranchos formations analyzed for this work (n = 198). The errors are at 95% confidence interval. Ages on the concordia curve are in Ma. Individual concordia plots for zircons (including discordant ones) from each rock sample are shown in the supplementary material.

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Figure 6. Probability density plot of spot SHRIMP U-Pb ages for all inherited zircons studied for this work (red line; n = 219). In addition, individual age probability curves for zircons from the Maimón (blue line; n = 69) and Los Ranchos (green line; n = 150) formations are shown. Note that zircons from both formations yield largely equivalent age distributions. U-Pb ages for inherited zircons found in Cuban arc-related and ophiolitic rocks of Cretaceous age are also shown for comparison (data from Rojas-Agramonte et al., 2016).

Figure 7. (a) Oxygen ($\delta^{18}O$) and (b) hafnium [εHf(t)] isotope composition vs. U-Pb age plots for magmatic and inherited zircons from the Maimón and Los Ranchos formations. The grey field in (a) indicate typical $\delta^{18}O$ compositions of mantle zircons according to Valley et al. (2005). The ages of magmatic (Cretaceous) zircons are from Torró et al. (2017a, b, c). DM = depleted mantle; CHUR = chondritic uniform reservoir.

Figure 8. Paleogeographic and tectonic reconstruction of the Mexico-Caribbean-northern South America region (after Pindell and Kennan, 2009, and Pindell et al., 2012). (a) During the Late Triassic (ca. 210 Ma), arc magmatism was inactive and the Paleo-Pacific margin of Pangea acted as a passive margin (Centeno-García, 2017). Extensive erosion of equatorial Pangea and deposition in the Paleo-Pacific margin resulted in the development of fan systems such as the Potosí fan delta. Detrital zircons in the Potosí fan were likely eroded from the exposed volcanic arc of Permian-Carboniferous age (Kirsch et al., 2012); Early Paleozoic, Pan-African and Grenvillian-age zircons were potentially eroded from the basement sources indicated with stars and explained in the main text. The location of the Alamar River and the Potosí fan system is based on Barboza-Gudiño et al. (2010). (b) During the latest Jurassic (ca. 148 Ma), rifting and ocean crust formation along the Proto-Caribbean seaway had separated southern Mexico from northwestern Colombia. Subduction of marine sedimentary covers (e.g., Potosí fan; Centeno-García, 2017) along the Cordilleran margin of Mexico, which had a strong sinistral component, resulted in the relative southward migration of that sediments into the position formerly occupied by Colombia. (c) Previously to the inception of the subduction of the Proto-Caribbean (ca. 141 Ma), a sinistral inter-American transform fault connected the west-directed subduction of the Paleo-Pacific beneath the Americas. Materials subducting along the Cordilleran margin of Mexico at that time were transported southward to the hanging wall of the transform, where the initial subduction of the Proto-Caribbean took place (Pindell et al., 2012). (d) During the early stages of evolution of the Caribbean island arc (ca. 125 Ma), proto- and first-arc magmas, of tholeiitic and boninitic affinities, recorded in the Maimón and Los Ranchos formations emplaced (Escuder-Viruete et al., 2006; Torró et al., 2016b, 2017b, c). These magmas transported upwards the inherited, pre-Cretaceous zircon cargo from the mantle wedge source. Such mantle source was contaminated by subducted sediments along the new

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Benioff zone, and potentially during previous subduction zones along the western margin of Pangea.

APPENDICES (Supplementary material)

APPENDIX 1. Analytical methods.

APPENDIX 2. U–Pb SHRIMP data for spot analyses of zircon grains from Early Cretaceous igneous arc-related Los Ranchos and Maimón formations, Dominican Republic. Details on the location and petrographic-lithogeochemical characterization of the hosting rock samples are given in Table 1.

APPENDIX 3. Oxygen SHRIMP data for spot analyses of zircons from the igneous arc-related, Early Cretaceous Los Ranchos and Maimón formations, Dominican Republic. Calibration data is also provided. Oxygen isotope ratios are expressed in the standard $\delta^{18}O$ notation, which represents deviation of the measured $^{18}O/^{16}O$ value from the Vienna standard mean ocean water (VSMOW) in parts per thousand.

APPENDIX 4. U-Pb LA-ICP-MS data for spot analyses of zircon grains from sample APV-Diorite from Early Cretaceous volcanic, arc-related Los Ranchos Formation, Dominican Republic.

APPENDIX 5. Wetherill concordia plot for inherited zircons studied from each rock sample.

APPENDIX 6. Hf isotopic compositions of zircons from igneous arc-related, Early Cretaceous Los Ranchos and Maimón formations, Dominican Republic. Analytical uncertainties are quoted at 2 sigma. The error in εHf (t) combines (in quadrature) the within-run uncertainties in interference-corrected $^{176}Hf/^{177}Hf$, measured $^{176}Lu/^{177}Hf$, crystallization age and the decay constant of ^{176}Lu. The initial Hf isotope ratios were calculated using the $^{206}Pb/^{238}U$ age for concordant zircons.

APPENDIX 7.Cathodoluminescence images with spot U-Pb ages and oxygen signatures of analyzed zircons from the igneous arc-related, Early Cretaceous Los Ranchos and Maimón formations, Dominican Republic.

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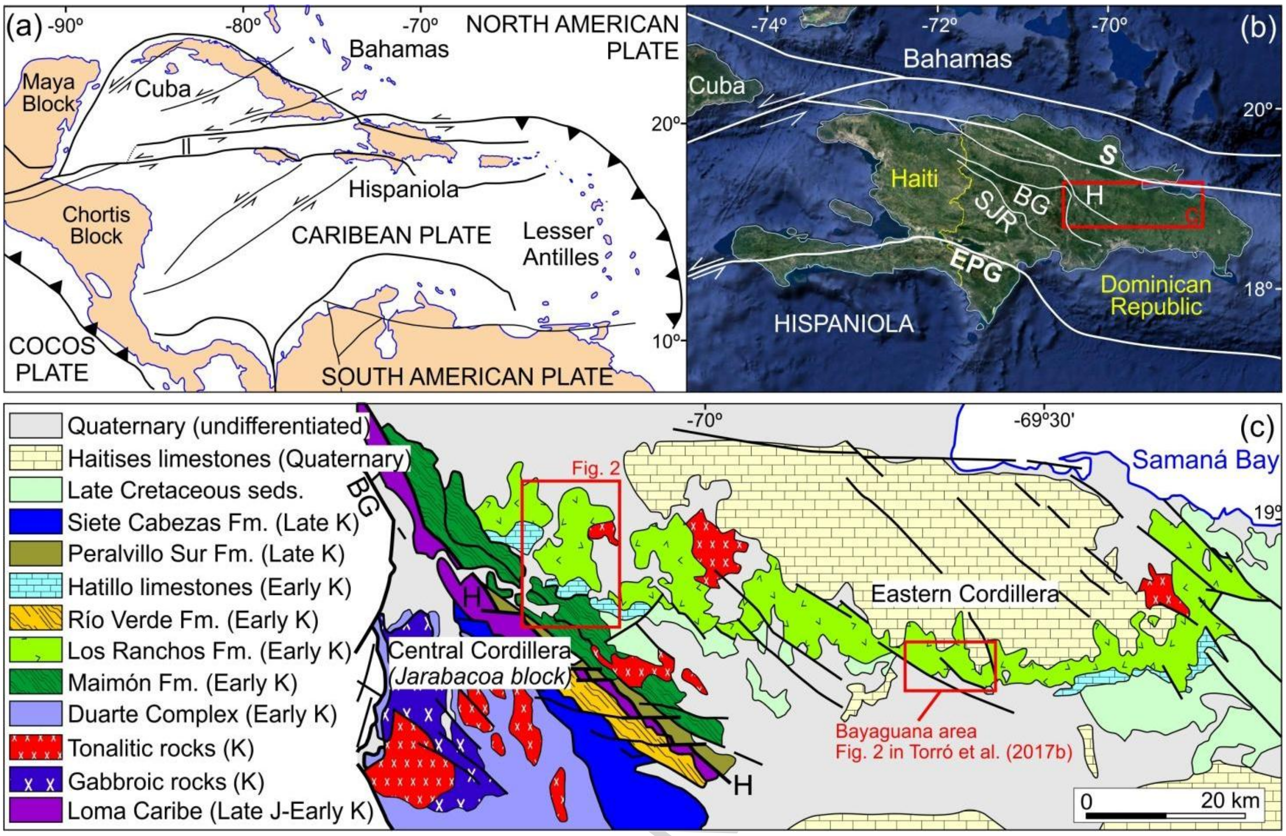
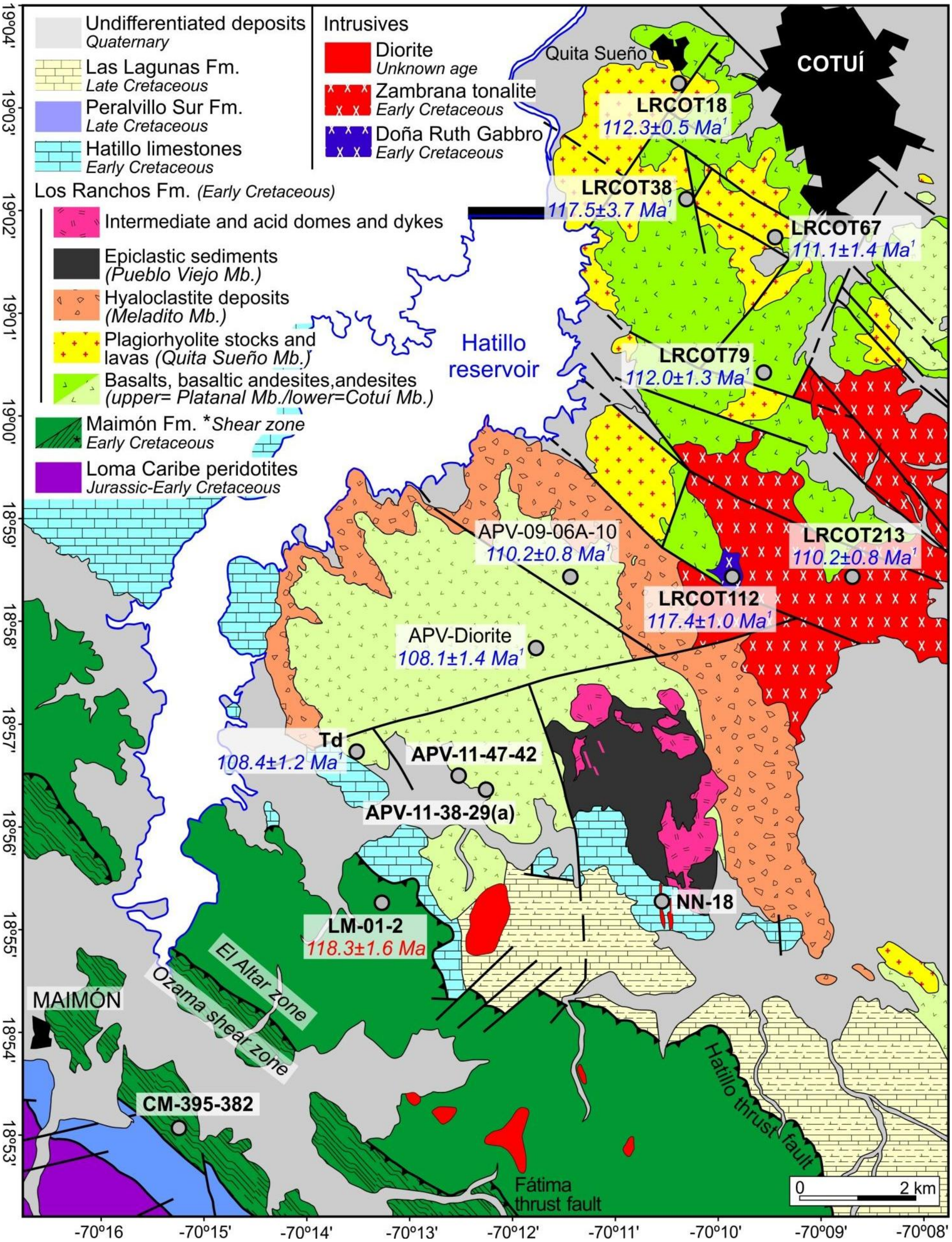

(a) -90º -80º -70º
NORTH AMERICAN PLATE
Bahamas
Maya Block
Cuba
20º
Hispaniola
Chortis Block
CARIBEAN PLATE
Lesser Antilles
10º
COCOS PLATE
SOUTH AMERICAN PLATE
-74º -72º -70º (b)
Bahamas
Cuba
20º
S
Haiti
H
BG
SJR
C
EPG
Dominican Republic
18º
HISPANIOLA
Quaternary (undifferentiated)
Haitises limestones (Quaternary)
Late Cretaceous seds.
Siete Cabezas Fm. (Late K)
Peralvillo Sur Fm. (Late K)
Hatillo limestones (Early K)
Río Verde Fm. (Early K)
Los Ranchos Fm. (Early K)
Maimón Fm. (Early K)
Duarte Complex (Early K)
Tonalitic rocks (K)
Gabbroic rocks (K)
Loma Caribe (Late J-Early K)
-70º -69º30' (c)
Samaná Bay
19º
Fig. 2
BG
H
Eastern Cordillera
Central Cordillera
(Jarabacoa block)
Bayaguana area
Fig. 2 in Torró et al. (2017b)
H
0 20 km

Fig. 1



Undifferentiated deposits
Quaternary
Las Lagunas Fm.
Late Cretaceous
Peralvillo Sur Fm.
Late Cretaceous
Hatillo limestones
Early Cretaceous
Los Ranchos Fm. (Early Cretaceous)
Intermediate and acid domes and dykes
Epiclastic sediments
(Pueblo Viejo Mb.)
Hyaloclastite deposits
(Meladito Mb.)
Plagiorhyolite stocks and lavas (Quita Sueño Mb.)
Basalts, basaltic andesites,andesites
(upper= Platanal Mb./lower=Cotuí Mb.)
Maimón Fm. *Shear zone
Early Cretaceous
Loma Caribe peridotites
Jurassic-Early Cretaceous
Intrusives
Diorite
Unknown age
Zambrana tonalite
Early Cretaceous
Doña Ruth Gabbro
Early Cretaceous
Quita Sueño
COTUÍ
LRCOT18
112.3±0.5 Ma1
LRCOT38
117.5±3.7 Ma1
LRCOT67
111.1±1.4 Ma1
LRCOT79
112.0±1.3 Ma1
Hatillo reservoir
APV-09-06A-10
110.2±0.8 Ma1
LRCOT213
110.2±0.8 Ma1
LRCOT112
117.4±1.0 Ma1
APV-Diorite
108.1±1.4 Ma1
Td
108.4±1.2 Ma1
APV-11-47-42
APV-11-38-29(a)
NN-18
LM-01-2
118.3±1.6 Ma
El Altar Zone
Ozama shear zone
MAIMÓN
CM-395-382
Hatillo thrust fault
Fátima thrust fault
0
2 km
19º04'
19º03'
19º02'
19º01'
19º00'
18º59'
18º58'
18º57'
18º56'
18º55'
18º54'
18º53'
-70º16
-70º15'
-70º14'
-70º13'
-70º12'
-70º11'
-70º10'
-70º09'
-70º08'

Fig. 2

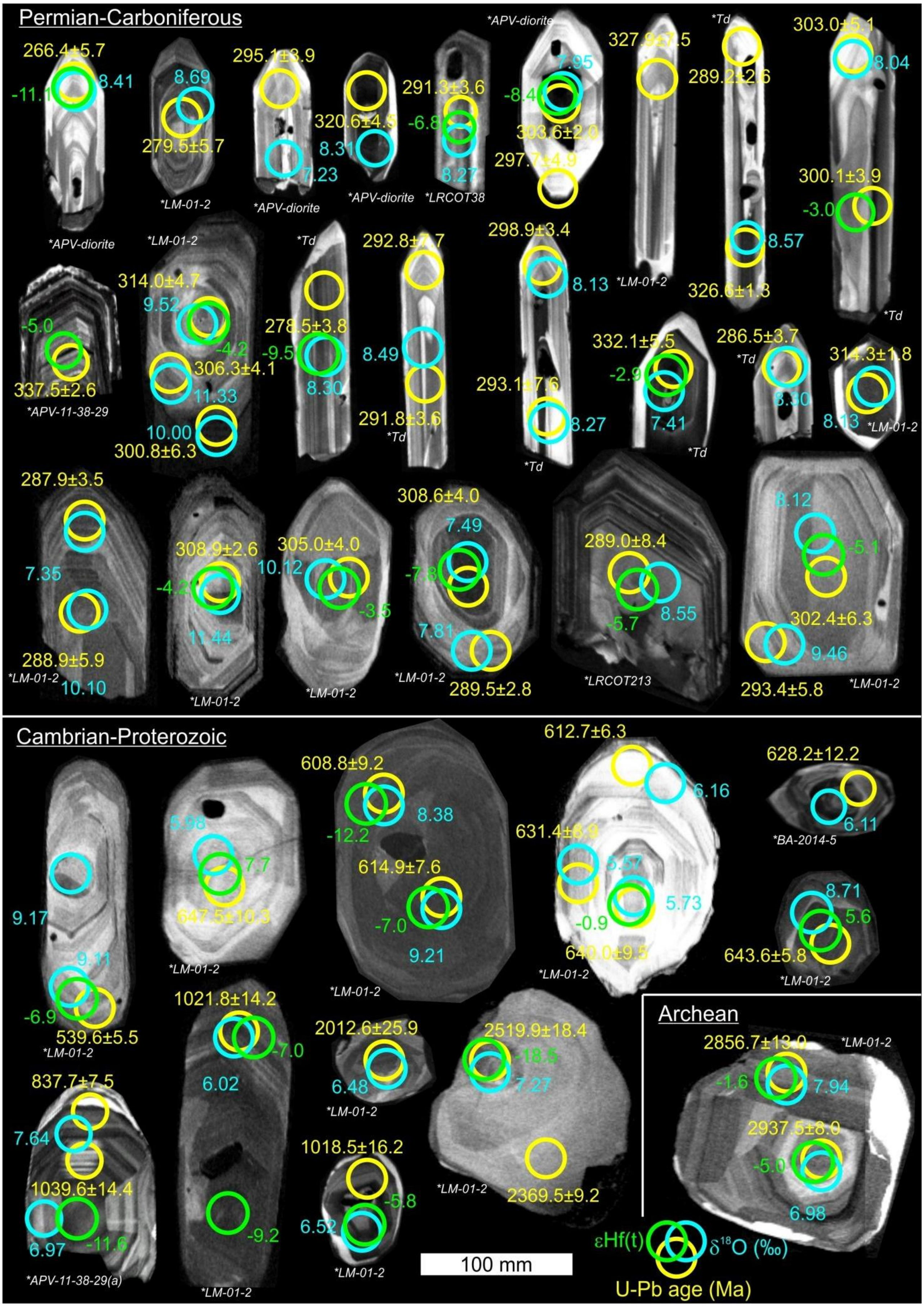

Permian-Carboniferous
266.4±5.7
8.41
8.69
279.5±5.7
*APV-diorite
*LM-01-2
295.1±3.9
7.23
*APV-diorite
320.6±4.5
8.31
*APV-diorite
291.3±3.6
-6.8
8.27
*LRCOT38
*APV-diorite
7.95
303.6±2.0
297.7±4.9
327.9±7.5
*Td
289.2±2.6
303.0±5.1
8.04
300.1±3.9
-3.0
8.57
326.6±1.3
*Td
*LM-01-2
-5.0
337.5±2.6
*APV-11-38-29
314.0±4.7
9.52
-4.2
306.3±4.1
11.33
10.00
300.8±6.3
*Td
278.5±3.8
-9.5
8.30
292.8±7.7
8.49
291.8±3.6
*Td
298.9±3.4
8.13
293.1±7.6
8.27
*Td
332.1±5.5
-2.9
7.41
*Td
286.5±3.7
*Td
314.3±1.8
8.13
*LM-01-2
287.9±3.5
7.35
288.9±5.9
*LM-01-2
10.10
308.9±2.6
11.44
*LM-01-2
305.0±4.0
10.12
-3.5
*LM-01-2
308.6±4.0
7.49
7.81
289.5±2.8
*LM-01-2
289.0±8.4
-5.7
8.55
*LRCOT213
8.12
-5.1
302.4±6.3
9.46
293.4±5.8
*LM-01-2
Cambrian-Proterozoic
9.17
-6.9
539.6±5.5
*LM-01-2
5.98
7.7
647.5±10.3
*LM-01-2
608.8±9.2
8.38
-12.2
614.9±7.6
-7.0
9.21
*LM-01-2
612.7±6.3
6.16
5.73
-0.9
*LM-01-2
628.2±12.2
6.11
*BA-2014-5
8.71
5.6
643.6±5.8
*LM-01-2
1021.8±14.2
-7.0
6.02
-9.2
*LM-01-2
837.7±7.5
7.64
1039.6±14.4
6.97
-11.6
*APV-11-38-29(a)
2012.6±25.9
6.48
*LM-01-2
1018.5±16.2
-5.8
6.52
*LM-01-2
2519.9±18.4
7.27
*LM-01-2
2369.5±9.2
Archean
*LM-01-2
7.94
-5.0
6.98
100 mm
εHf(t)
δ18O (‰)
U-Pb age (Ma)

Fig. 3

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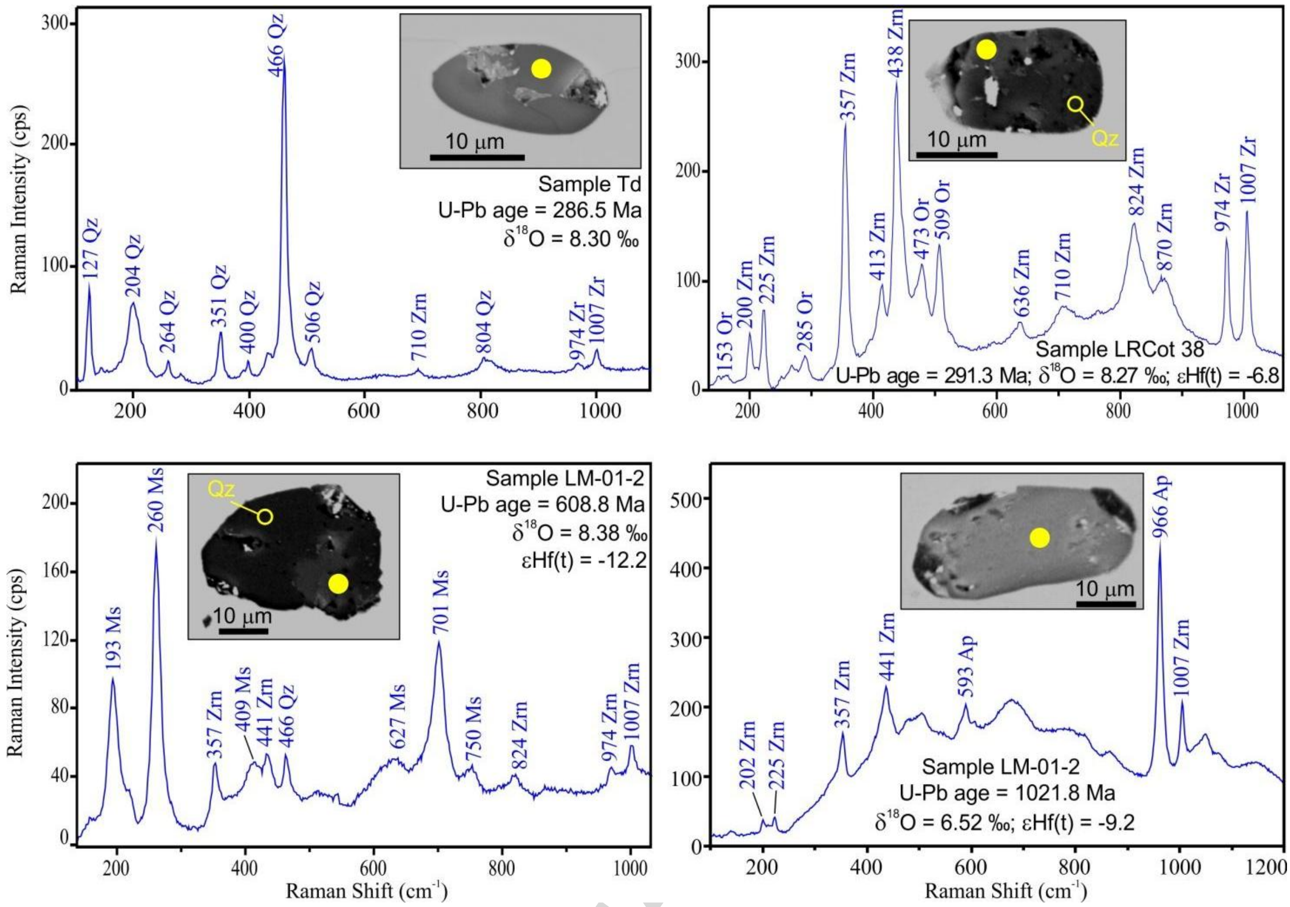

Fig. 4

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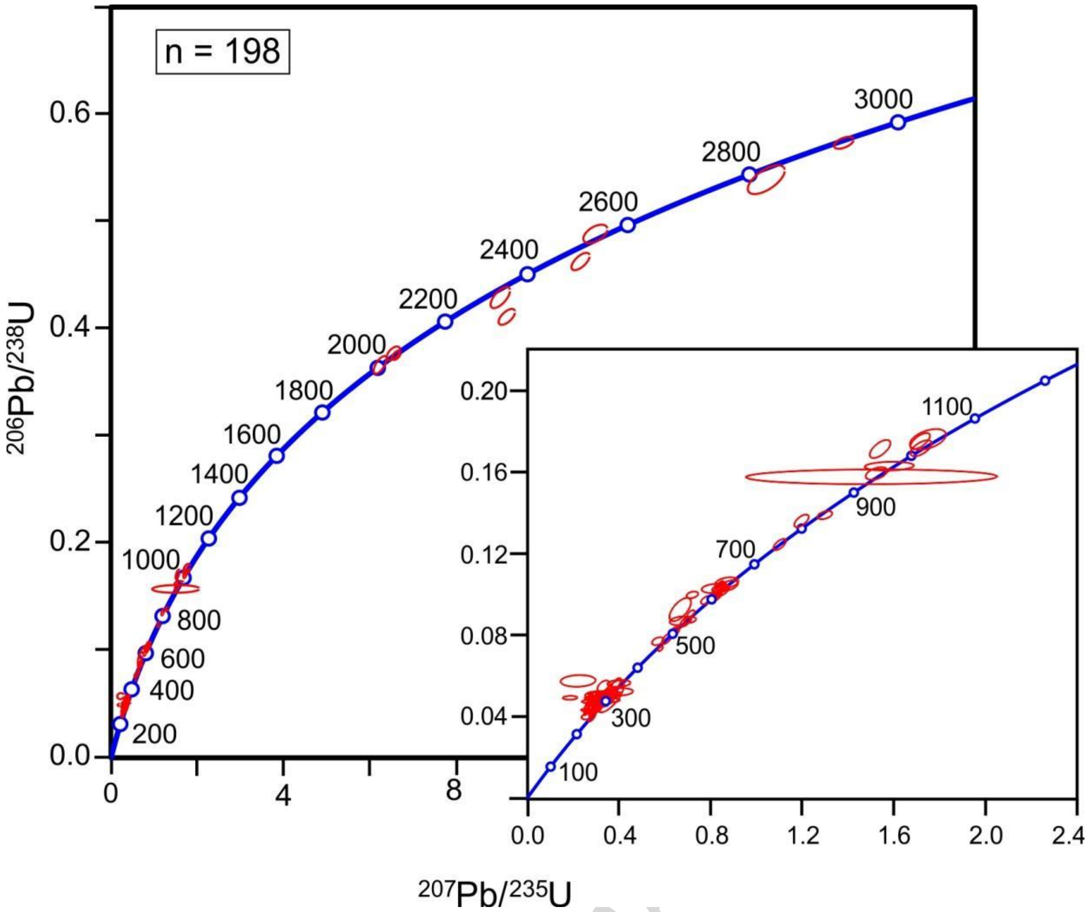

Fig. 5

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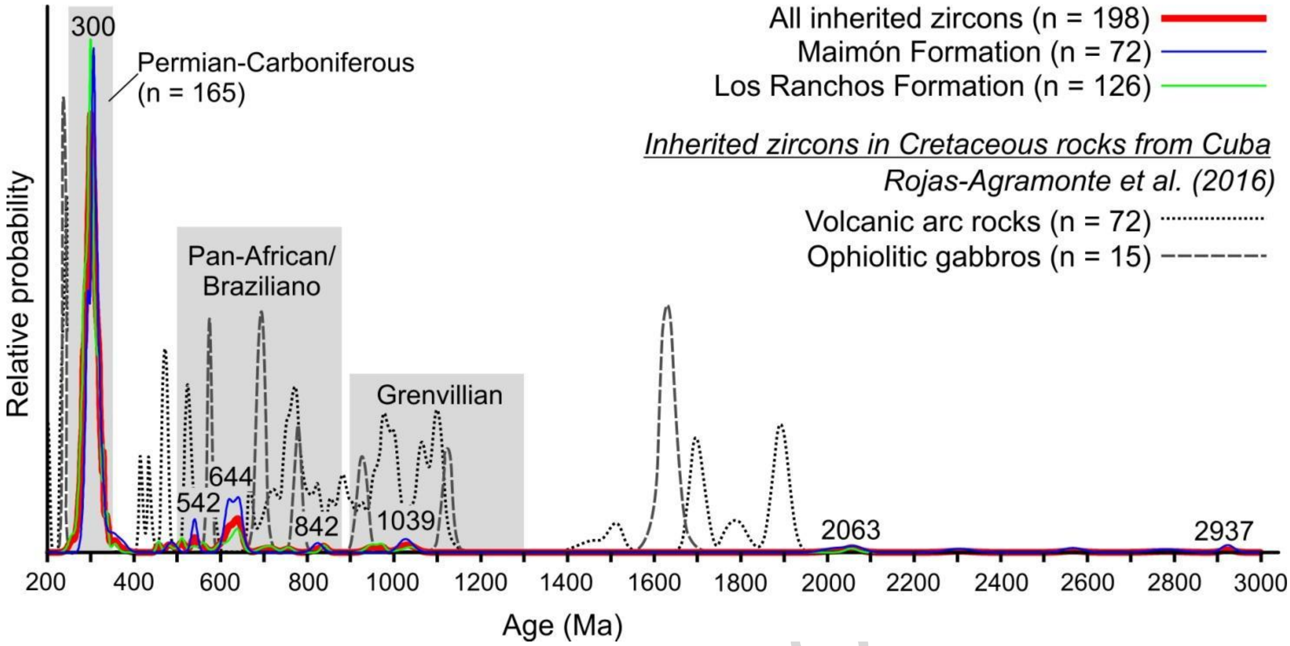

Fig. 6

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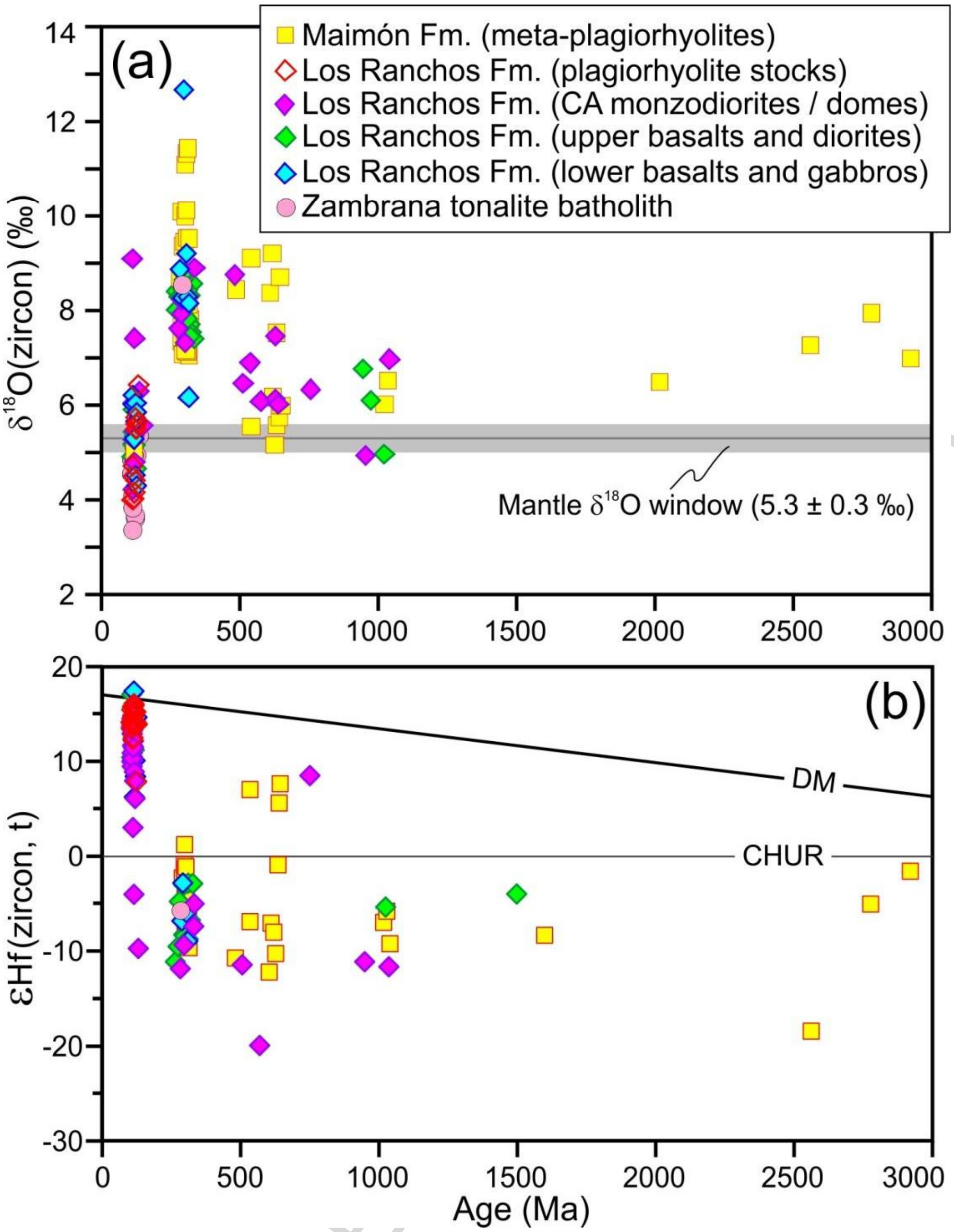
Maimón Fm. (meta-plagiorhyolites)
Los Ranchos Fm. (plagiorhyolite stocks)
Los Ranchos Fm. (CA monzodiorites / domes)
Los Ranchos Fm. (upper basalts and diorites)
Los Ranchos Fm. (lower basalts and gabbros)
Zambrana tonalite batholith
(a)
δ18O(zircon) (‰)
Mantle δ18O window (5.3 ± 0.3 ‰)
(b)
DM
CHUR
εHf(zircon, t)
Age (Ma)

Fig. 7

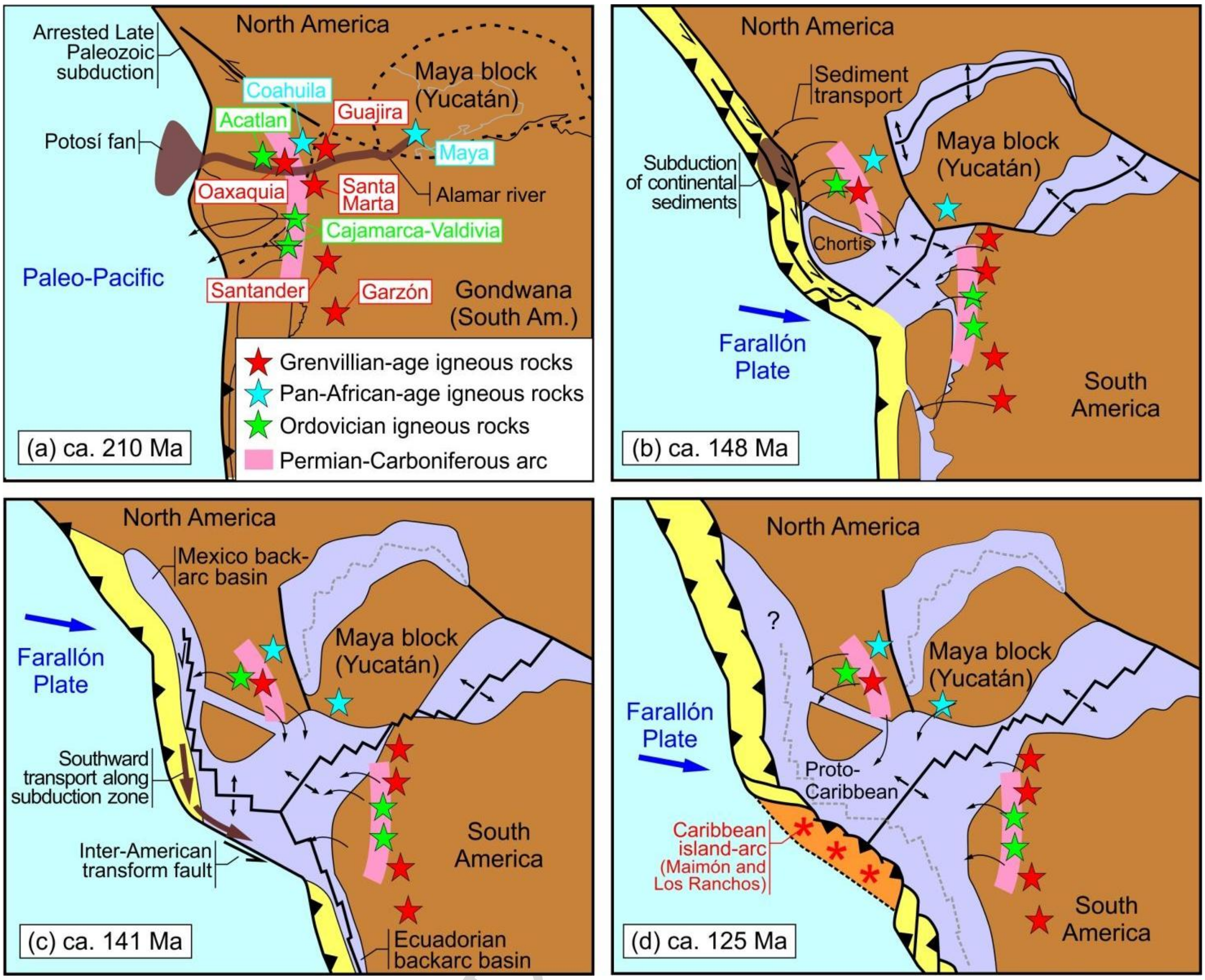

Fig. 8

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Table 1

Sample No.	Sample type	Coordinates[1]	DDH[2]/ depth (m)	Unit	Lithology	Geochemical affinity	Age (Ma; youngest zircon population)	Peak xenocrystic populations (Ma)
LM-01-2	Drill core	371888E 2092667N	LM-01-2/45.15	Maimón Fm.	Gnessic plagiorhyolite[a]	Tholeiitic, low-K[b]	118.3± 1.6	306, 341, 373; 487; 540; 617; 637; 823; 1030; 2038; 2304; 2566; 2780; 2923
CM-395-382	Drill core	369101E 2087513N	CM-395/382	Maimón Fm.	Gnessic plagiorhyolite[a]	-	?	224; 295
APV-09-06A-10	Drill core	374625E 2098672N	APV-09-06(A)/ 95.95	Los Ranchos Fm.	Semi-porphyritic basalt (spillite)	Low-Ti IAT[c]	110.2± 0.8[c]	300; 968
APV-11-38-29(a)	Drill core	373120E 2094990N	APV-11-38/223	Los Ranchos Fm.	Monzodiorite	Calc-alkaline[c]	110-107 (?[c])	302; 337; 541; 756; 839; 953; 1040
APV-11-47-42	Drill core	372687E 2095183N	APV-11-47/229	Los Ranchos Fm.	Monzodiorite	Calc-alkaline[c]	110-107 (?[c])	288
APV-Diorite	Outcrop	373992E 2098057N	-	Los Ranchos Fm.	Diorite	Low-Ti IAT[c]	108.1± 1.4[c]	252; 304; 637; 712
Td	Outcrop	370998E 2095450N	-	Los Ranchos Fm.	Diorite	Low-Ti IAT[c]	108.4± 1.2[c]	300, 457
NN-18	Outcrop	376460 E 2093090N		Los Ranchos Fm.	Monzodiorite	Calc-alkaline[c]	?	272; 289; 480; 508; 559; 629
LRCOT-18	Outcrop	376511E 2107120N	-	Los Ranchos Fm.	Plagiorhyolite	Tholeiitic, low-K[c]	112.3± 0.5[c]	-
LRCOT-38	Outcrop	376735E 2105151N	-	Los Ranchos Fm.	Pillowed basalts	LREE-depleted IAT[c]	117.5± 3.7[c]	296
LRCOT-67	Outcrop	378220E 2104372N	-	Los Ranchos Fm.	Plagiorhyolite	Tholeiitic, low-K[c]	111.1± 1.4[c]	-
LRCOT-79	Outcrop	377741E 2101983N	-	Los Ranchos Fm.	Amygdaloidal-spherulitic basalt	LREE-depleted low-Ti IAT[c]	112.0± 1.3[c]	-
LRCOT-112	Outcrop	377337E 2098534N	-	Los Ranchos Fm.	Plagioclase-amphibole gabbro	LREE-depleted low-Ti IAT[c]	117.4± 1.0[c]	283; 314; 598; 2055
LRCOT-	Outcrop	37932	-	Los	Tonalite	Tholeiitic,	110.2±	290

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213	p	3E 2098792N		Ranchos Fm.		low-K[c]	0.8[c]	
BA-2014-5	Outcrop	445431E 2079606N	-	Los Ranchos Fm.	Rhyodacite dome (breccia)	-	106.5± 2.6[c]	628
BA-2014-7	Outcrop	444902E 2080825N	-	Los Ranchos Fm.	Rhyodacite dome (breccia)	-	109.1± 1.7[c]	277

[1]For drill core samples, coordinates refer to the location of collars. UTM Zone 19 (NAD27 for US).
[2]DDH refers to the identification of the Diamanond Drill Holes from which the samples were collected.
[a]Torró et al. (2016b); [b]Torró et al. (2017a); [c]Torró et al. (submitted)

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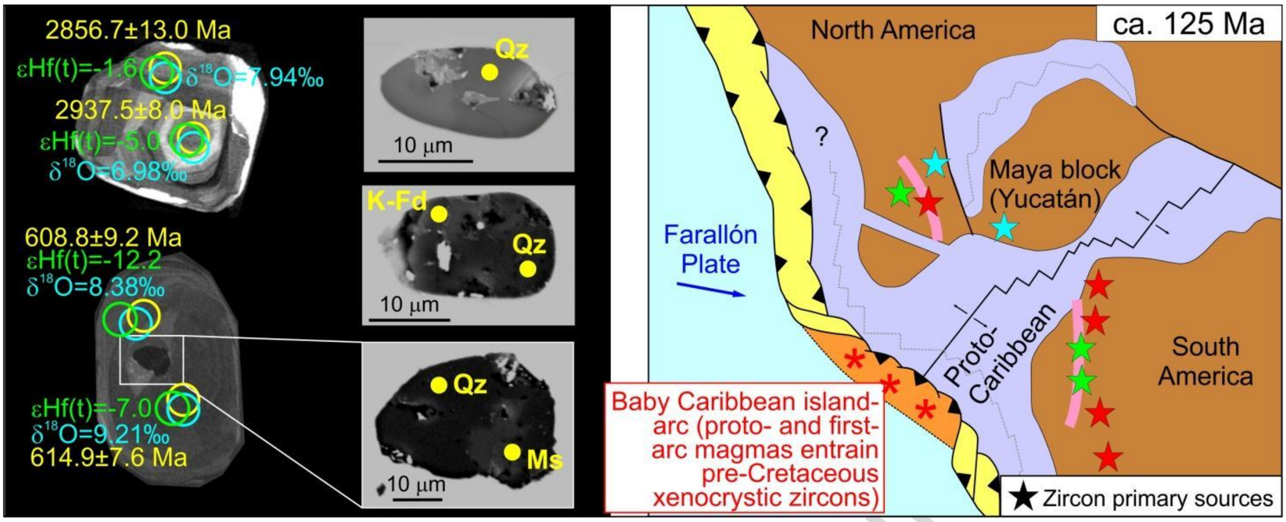

Graphical abstract

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HIGHLIGHTS

- Archean to Permian zircons in Cretaceous intra-oceanic arc rocks from the Caribbean
- Inherited zircons of crustal origin; potential sources in Mexico and Colombia
- Arc magmas picked up the ancient zircons from their sources in the mantle wedge

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