Palaeozoic to Early Jurassic history of the northwestern corner of Gondwana, and implications for the evolution of the Iapetus, Rheic and Pacific Oceans

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Abstract

The Palaeozoic to Mesozoic igneous and metamorphic basement rocks exposed in the Mérida Andes of Venezuela and the Santander Massif of Colombia are generally considered to define allochthonous terranes that accreted to the margin of Gondwana during the Ordovician and the Carboniferous. However, terrane sutures have not been identified and there are no published isotopic data that support the existence of separate crustal domains. A general paucity of geochronological data led to published tectonic reconstructions for the evolution of the northwestern corner of Gondwana that do not account for the magmatic and metamorphic histories of the basement rocks of the Mérida Andes and the Santander Massif. We present new zircon U-Pb (ICP-MS) data from 52 igneous and metamorphic rocks, which we combine with whole rock geochemical and Pb isotopic data to constrain the tectonic history of the Precambrian to Mesozoic basement of the Mérida Andes and the Santander Massif. These data show that the basement rocks of these massifs are autochthonous to Gondwana and share a similar tectono-magmatic history with the Gondwanan margin of Peru, Chile and Argentina, which evolved during the subduction of oceanic lithosphere of the Iapetus Ocean. The oldest Palaeozoic arc magmatism is recorded at ~500 Ma, and was followed shortly by Barrovian metamorphism. Peak metamorphic conditions at upper amphibolite facies are recorded by anatexis at ~477 Ma and the intrusion of synkinematic granitoids until ~472 Ma. Subsequent retrogression resulted from localised back-arc or intra-arc extension at ~453 Ma, when volcanic tuffs and interfingering sedimentary rocks were deposited over the amphibolite facies basement. Continental arc magmatism dwindled after ~430 Ma and terminated at ~415 Ma, coevally with most of the western margin of Gondwana. After Pangaea amalgamation in the Late Carboniferous to Early Permian, a magmatic arc developed on its western margin at ~294 Ma as a result of subduction of oceanic crust of the palaeo-Pacific ocean. Intermittent arc magmatism recorded between ~294 and ~225 Ma was followed by the onset of the Andean

subduction cycle at ~213 Ma, in an extensional regime. Extension was accompanied by slab roll-back which led to the migration of the arc axis into the Central Cordillera of Colombia in the Early Jurassic.

1. Introduction

The Andes of Venezuela and Colombia host exposures of Precambrian to Phanerozoic orogenic belts (Fig. 1), which record the tectonic history of northwestern Gondwana. We combine a large quantity of new data with pre-existing information to reconstruct the Palaeozoic history of northwest Gondwana. Most attempts to reconstruct the pre-Mesozoic tectonic history have focused on crustal fragments that separated during the disassembly of Pangaea, and now form large parts of Central America and Mexico. Recent studies of basement rocks in Mexico (e.g. Keppie et al., 2008, 2012; Vega-Granillo et al., 2009a and references therein) and Central America (e.g. Ratschbacher et al., 2009; Estrada-Carmona et al., 2012 and references therein) have contributed large amounts of new data and palaeogeographic models. These models assume that the basement rocks of Mexico were juxtaposed against northwestern Gondwana in the Early Palaeozoic, which is supported by sparse Pb isotopic data from the Garzón Massif and the Sierra Nevada de Santa Marta of Colombia (Fig. 1; e.g. Yañez et al., 1991; Ruiz et al., 1999) and faunal similarities between Early Palaeozoic sedimentary rocks exposed in Mexico and the Mérida Andes of Venezuela (e.g. Stewart et al., 1999). However, most models of the Palaeozoic evolution of the basement rocks exposed in the Northern Andes imply that they are allochthonous to northwestern Gondwana, and accreted during the Ordovician and the Carboniferous (e.g. Bellizzia and Pimentel, 1994; Aleman and Ramos, 2000; Ramos, 2009; Ramos, 2010). Therefore, it is unclear whether the basement of the Northern Andes or Central America can be used to constrain the Palaeozoic tectonic history of northwestern Gondwana, and for which time periods.

We present new in-situ zircon U-Pb laser-ablation inductively coupled plasma-mass spectrometry (LA-ICP-MS) ages, whole rock geochemistry and Pb radiogenic isotope data from the basement of the Mérida Andes of Venezuela and the Santander Massif of Colombia. These data are used to (i) test the hypothesis that there are allochthonous terranes in the Northern Andes, which accreted in the Ordovician and the Carboniferous, (ii) determine the tectonic origin of basement rocks exposed in the Northern Andes, and use these to propose plate reconstructions for the evolution of the southern Iapetus and Rheic oceans, and (iii) investigate the possible connections between the basement sequences in northwestern South America and Mexico.

2. Geological Framework

The present-day configuration of northwest South America is a consequence of interactions between the Caribbean, South American and Pacific Plates, which have formed the South Caribbean Plate Boundary Zone (Fig. 1). This broad, diffuse plate boundary hosts discrete tectonic blocks that have undergone alternating periods of dextral transtension and transpression since the Late Cretaceous (e.g. Colletta et al., 1997; van der Lelij et al., 2010; Bermúdez et al., 2011), generating regions of high topography isolated by rapidly subsiding basins. A prominent structural entity of the Southern Caribbean Plate Boundary Zone is the Maracaibo Triangular Block, which incorporates the Mérida Andes, the Santander Massif, the Sierra Nevada de Santa Marta and the basement of Lake Maracaibo (Fig. 1).

2.1. The Mérida Andes

2.1.1. Igneous and Metamorphic basement

The metamorphic basement of the Mérida Andes consists of Precambrian to Palaeozoic schists and gneisses, which are intruded by numerous granitoids (Fig. 2; Gonzales de Juana et al., 1980). These rocks are exposed within a transpressive orogen whose architecture is controlled by the active SW-NE trending Boconó Fault (Colletta et al., 1997).

Metasedimentary schists, paragneisses and orthogneisses of the Iglesias Complex form the oldest rocks exposed in the Mérida Andes (Gonzales de Juana et al., 1980). The Iglesias Complex is subdivided by metamorphic facies into the greenschist facies Tostós and Bella Vista suites and the amphibolite facies Sierra Nevada Suite (Fig. 2; Odremán and Useche, 1997). Mapping has shown that the transition between the Tostós and Sierra Nevada suites is mainly conformable and follows metamorphic isograds (e.g. Case et al., 1990), which has led to the suggestion that the basement rocks in the Mérida Andes were metamorphosed simultaneously (Maréchal, 1983).

The Iglesias Complex is considered to be Proterozoic by most authors (e.g. Odremán and Useche, 1997; Aleman and Ramos, 2000), although there is a paucity of constraining data. Previous geochronological studies of the Iglesias Complex utilized the K/Ar, Rb/Sr and multigrain U-Pb methods (Fig. 3). Interpretations of those dates are complicated by inheritance of ancient radiogenic components and daughter isotope loss due to thermal overprinting or chemical alteration. The only Precambrian U-Pb dates are upper intercepts of strongly discordant multigrain fractions of the Upper Caparo, Valera and Pedraza granitoids, which intrude the Iglesias Complex and yield ages of 615 ± 40 , 593 ± 16 and 585 ± 25 Ma, respectively (Fig. 3; Burkley, 1976).

Most previous U-Pb geochronology focused on undeformed igneous rocks, and the only reported U-Pb data from the high grade Sierra Nevada Suite yielded an Ordovician age (Burkley, 1976). Rb/Sr dates acquired from metasedimentary units suggest the rocks were metamorphosed in the Early and Late Palaeozoic (Fig. 3; e.g. Burkley, 1976; Kovach et al., 1977), and most K/Ar dates from the Iglesias Complex are Mesozoic.

Early Palaeozoic regional metamorphism of the Iglesias complex is inferred from the structural concordance of the metamorphic Sierra Nevada Suite with magmatic bodies that underwent syn-intrusive deformation (Schubert, 1969; Burkley, 1976; Kovach et al., 1977; Gonzales de Juana et al., 1980; Maréchal, 1983). The amphibolite facies metasedimentary rocks of the Sierra Nevada Suite record Barrovian metamorphism with clockwise pressure-temperature paths (Grauch, 1972). Granitic gneisses from the Sierra Nevada Suite near La Grita (Fig. 2) yield a 6 point Rb/Sr whole rock isochron age of 440 ± 40 Ma (Kovach et al., 1977), which was interpreted to estimate the timing of amphibolite facies metamorphism.

The Tostós Suite mainly consists of greenschist facies phyllites, schists, quartzites and slates, and may represent a lower grade equivalent of the Sierra Nevada Suite (Shagam and Grauch, 2006). It is intruded by the Pueblo Hondo granodiorite (08VDL04; Fig. 2) and several smaller bodies.

The Bella Vista Suite is a greenschist facies metasedimentary unit consisting of schists, slates and phyllites, and forms the basement of the Caparo Block (Fig. 2). No geochronological data have been obtained from this unit, but its deposition is loosely constrained to the Precambrian by the intrusive contact of several granitoids which yield strongly discordant Late Neoproterozoic to Palaeozoic U-Pb ages (615 ± 40 to 350 ± 50 Ma; Fig. 3). The Bella Vista Suite is inferred to be unconformably overlain by Sandbian (Late Ordovician) sedimentary rocks of the Caparo Fm. (Gonzales de Juana et al., 1980), although field relationships between these units have not been observed.

2.1.2. Palaeozoic to Jurassic sedimentary rocks

The oldest, non-metamorphosed sedimentary rocks crop out in the Caparo Block (Fig. 2). The Caparo Fm. consists of micaceous fine to medium grained sandstones and siltstones, which yield fauna suggesting a Sandbian age (Odremán and Useche, 1997, Gutiérrez-Marco et al., 2011) of ~ 461 -456 Ma (Gradstein et al., 2012). Early Palaeozoic granitoids that are intrusive into the Caparo Fm. generated chistolite bearing contact aureoles (Gonzales de Juana et al., 1980) suggesting

shallow intrusion at less than ~3.5 kbar. A conglomerate containing clasts of basement rocks separates this unit from the overlying El Horno Fm., which consists of siltstones and lithic sandstones yielding brachiopods of Early Llandovery – Early Wenlock age (Gonzales de Juana et al., 1980; 444-426 Ma). Basal conglomerates of the El Horno Fm. unconformably overlie the Islitas tonalite (08VDL55, Fig. 2, Gonzales de Juana et al., 1980), which constrains exhumation of this intrusion to the Ordovician.

The Mucuchachi Formation (Fig. 2) unconformably overlies amphibolite facies metamorphic basement and mainly consists of green to black carbonaceous slates, shales and phyllites, which locally yield Pennsylvanian fauna (Odremán and Useche, 1997), although most sections have an undifferentiated Palaeozoic age (Gonzales de Juana et al., 1980). Shales from the Mucuchachi Fm. are intercalated with a 2000m thick pile of volcanic tuffs associated with a volcanogenic massive sulphide deposit (Carlson, 1977; Staargaard and Carlson, 2000) at its westernmost exposure near Bailadores (08VDL15; Fig. 2). Low grade metasedimentary rocks of the Mucuchachi Fm. yield an imprecise Rb/Sr whole rock isochron age of 293 ± 30 Ma (Burkley, 1976), which was interpreted to constrain the timing of metamorphism. The Mucuchachi Fm. is unconformably overlain by the middle Pennsylvanian Sabaneta Formation, which crops out near the city of Mérida and in the Caparo Block (Fig. 2). This sedimentary unit consists of fluvial and floodplain deposits that grade up into the Middle Permian marine sedimentary rocks of the Palmarito Formation (Laya and Tucker, 2012). The Palmarito Fm. may record a marine transgression in a foreland basin related to Permian orogenesis (Laya and Tucker, 2012). A sedimentary hiatus spans the Guadalupian (~260 Ma) to the latest Triassic-earliest Jurassic, when an extensional regime led to the formation of grabens that host volcanic rocks and red siliclastic strata of the La Quinta Formation (Maze, 1984; Ostos et al., 2005; Barrett et al., 2008).

Several low grade metasedimentary units of unknown age unconformably overlie the high grade metamorphic basement complex. These include the Los Torres, Cerro Azul and El Aguila Suites, and are considered to be Palaeozoic (e.g. Schubert, 1969; Kovisars, 1971).

2.2. The Santander Massif

2.2.1. Igneous-Metamorphic basement

The Santander Massif is the northern continuation of the Eastern Cordillera of Colombia and forms the southwestern margin of the Maracaibo Triangular Block (Fig. 1). Its eastern boundary is defined by late Cainozoic thrusts that separate it from the Mérida Andes (Macellari, 1984). Northwards

from Bucaramanga, its western margin is sharply defined by the Santa Marta – Bucaramanga Fault (SMBF) along a NNW trend, although similar basement units can be found on both sides of the same fault where it meets the Floresta Massif (Fig. 1).

The oldest unit exposed in the Santander Massif is the Bucaramanga Gneiss (Fig. 4), which consists of high grade pelitic to silicic paragneisses with lesser amphibolites, marbles and migmatites (Ward et al., 1974). Metamorphic conditions of the Bucaramanga Gneiss reach medium pressure – high temperature amphibolite facies conditions of 5.5 to 7.2 kbar and 660 to 750°C in the central Santander Massif, with clockwise PT paths (García et al., 2005; Castellanos et al., 2008; Urueña-Suárez and Zuluaga, 2011). Many geochronological data have been collected from the basement of the Santander Massif (Fig. 5) and several analyses of the Bucaramanga Gneiss yield Precambrian ages. A hornblende gneiss to the west of Ocaña (Fig. 4) yields a hornblende K/Ar date of 945 ± 40 Ma (Ward et al., 1974). Chloritized hornblendes from the same area yield strongly disturbed $^{40}\text{Ar}/^{39}\text{Ar}$ spectra with individual steps spanning from ~200 to ~900 Ma (Restrepo-Pace et al., 1997). Zircons from a stromatic migmatite in the Bucaramanga Gneiss yield discordant, multigrain U-Pb data that define a discordia between 981 ± 85 Ma and 224 ± 76 Ma (Restrepo-Pace and Cediell, 2010), suggesting that protoliths containing Proterozoic zircons experienced Pb loss or metamorphic overgrowth in the Triassic. Single grain zircon U-Pb SHRIMP analyses yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates that range between 860 and 1200 Ma, which may reflect several magmatic and metamorphic episodes (Cordani et al., 2005).

The Silgará Formation mainly consists of phyllites, slates, quartzites and metasiltstones (Ward et al., 1974). It is inferred to overlie the Bucaramanga gneiss, and is overlain by mainly non-metamorphic Middle Devonian sedimentary rocks of the Floresta Fm. in several locations (Ward et al., 1974; Boinet et al., 1986), which suggests that metamorphism occurred prior to the Middle Devonian. Contacts with the Bucaramanga Gneiss are difficult to identify and frequently follow isograds (Ward et al., 1974). Metamorphic assemblages from the Silgará Formation to the east of San Gil (Fig. 4) reveal a wide range of metamorphic conditions and a clockwise P-T path from 495-518°C at 4.4-5.5 kbar to peak pressures of 6.6-7.5 kbar at 590-612°C, and peak temperatures of 660-700°C at 5.5-7.2 kbar (Ríos et al., 2003). Near Berlin (Fig. 4), the Silgara Formation reaches similar metamorphic conditions to the Bucaramanga Gneiss with peak temperatures of 630-704°C at pressures of 5-9.5 kbar (García et al., 2005; Castellanos et al., 2008).

The Silgará Formation and the Bucaramanga Gneiss are both intruded by the strongly foliated quartzo-feldspathic Berlin Orthogneiss (Fig. 4). Previous studies of the microtexture and the structural and isograd concordance of this unit with its host metasedimentary rocks suggest that the emplacement of these bodies was coeval with metamorphism (Ward et al., 1974), which may have occurred at 477 ± 16 Ma (multigrain U-Pb; Restrepo-Pace and Cediel, 2010). A U-Pb LA-ICP-MS age of 477 ± 2 Ma was obtained from calc-alkaline metadiorites that intrude the Bucaramanga Gneiss and share a sub-parallel foliation with their host rock (Mantilla et al., 2012). The deformed rocks are intruded by undeformed Early Palaeozoic granitoids and numerous large Mesozoic batholiths of the Santander Plutonic Group (Fig. 4), which yield late Triassic to early Jurassic ages and have continental arc geochemistry (Fig. 5; Goldsmith et al., 1971; Dörr et al., 1995; Mantilla et al., 2013).

2.2.2. Palaeozoic to Jurassic sedimentary rocks

There are no recognised Early Palaeozoic sedimentary rocks in the Santander Massif and the metamorphic basement is unconformably overlain by Late Palaeozoic units. The Middle Devonian Floresta Fm. comprises marine sedimentary rocks that unconformably overlie the Silgará Fm in several locations (Ward et al., 1974; Boinet et al., 1986). These rocks, together with the overlying Late Carboniferous Diamante Fm., were metamorphosed under low grade conditions, but their isograd patterns do not coincide with those of the basement and metamorphism appears to be related to a different event that predates the deposition of the younger Tiburrón, Bocas, Jordan and Girón Formations (Ward et al., 1974). Within these overlying sedimentary rocks only the Bocas Fm. yields fossils, which include *Glossopteris* sp. and *Classopollis* sp., yielding a Jurassic age (Mojica et al., 1996 and references therein). However, the proposed Latest Permian extinction of *Glossopteris* (McLoughlin et al., 1997) questions the accuracy of this age determination. The Girón Fm. comprises a series of red beds that contain clasts of pink granitoids from the Triassic-Jurassic Santander Plutonic Group, and is unconformably overlain by Lower Cretaceous sedimentary rocks (Ward et al., 1974). Hence, its age may span the Jurassic.

3. Previous models for the tectonic evolution of northwestern Gondwana

Existing models place the basement rocks of the Mérida Andes and the Santander Massif into different terranes, namely the Mérida (Bellizzia and Pimentel, 1994; Aleman and Ramos, 2000) and Chibcha terranes (e.g. Ramos, 2009). These suspect terranes are inferred to be allochthonous because the Northern Andes host Precambrian to Early Palaeozoic metamorphic basement that lies outboard of Early Palaeozoic pericratonic basins (e.g. Llanos Basin, Fig. 1; Sempere, 1995). Models

proposing an allochthonous Mérida Terrane argue that it formed on the northern margin of the Rheic Ocean and accreted to the autochthonous Caparo Block within Gondwana (Fig. 2) during the Carboniferous (Bellizzia and Pimentel, 1994; Aleman and Ramos, 2000). These models utilised ambiguous Rb/Sr model dates (Fig. 3) to suggest that a Carboniferous arc formed on the Mérida Terrane as it converged with Gondwana. The inferred absence of metamorphism in Early Palaeozoic rocks to the south of the Caparo-Soledad Fault (Fig. 2) was used to suggest that it was the Rheic Ocean suture between the Mérida Terrane and the autochthonous Caparo Block of Gondwana. However, (i) the Caparo-Soledad Fault is an active reverse fault with dextral strike slip components and exposes deeper levels of the crust to the northwest, in its hanging wall (Audemard and Audemard, 2002), (ii) Early Palaeozoic metamorphic basement may extend to the Apure Fault, which underlies the the Llanos Basin (Fig 1; Feo-Codecido et al., 1984), (iii) stratigraphic differences across the Caparo-Soledad Fault include an absence of documented Triassic granitoids in its footwall (Figs. 2 and 3), challenging the hypothesis that this structural divide is a Carboniferous feature, (iv) the Caparo-Soledad fault exposes no high pressure metamorphic assemblage or any remnants of entrained oceanic crust, (v) the Cerro Azul Suite of Early Palaeozoic or older age appears to crop out on both sides of the fault near its northeastern termination (Schubert, 1969), and (vi) the model is constructed on the assumption that all pre-Cretaceous K/Ar and Rb/Sr whole rock and mineral dates, and strongly discordant multigrain U-Pb data (Fig. 3), are accurate crystallisation ages and are evidence for a Late Palaeozoic arc (Bellizzia and Pimentel, 1994; Aleman and Ramos, 2000). The published geochronological data can not be unambiguously interpreted as crystallisation ages given the evidence for several phases of deformation and metamorphism (Kovach et al., 1977; Maréchal, 1983). We will refer to the area north of the Caparo-Soledad fault as the Mérida block, rather than the Mérida Terrane because there does not seem to be any clear evidence for oceanic separation or suturing between the Mérida and Caparo blocks, or any basis for considering it to be allochthonous to South America.

The Chibcha Terrane (e.g. Toussaint and Restrepo, 1988; Bellizzia and Pimentel, 1994) forms the basement of the Eastern Cordillera of Colombia and includes the Santander Massif, the Santa Marta Massif, the Périja Mountains, the Serranía de San Lucas and the Quetame Massif (Fig. 1). Existing models propose that the Chibcha Terrane formed part of Laurentia and collided with the passive margin of Gondwana in the Late Ordovician, forming a suture in the eastern Quetame Massif (Fig. 1; e.g. Ramos, 2009). Orogenesis during collision drove widespread deformation in the Quetame and Santander massifs (Forero Suarez, 1990; Ramos, 2009; Ramos, 2010; Restrepo-Pace and Cediél, 2010). However, (i) a Laurentian provenance is considered based upon similarities in

Devonian fauna between Laurentia and the Northern Andes (Forero Suarez, 1990) instead of pre-Ordovician faunal similarities, which are required by the model of Ramos (2009; 2010), (ii) rocks from the eastern Quetame Massif interpreted as Ordovician ophiolitic remnants (Ramos, 2009; 2010) were either originally reported as submarine intracontinental volcanic rocks of unknown age (Ariari basalts; Cáceres et al., 2003) or are Silurian or younger gabbros and diorites that were emplaced into the Silurian Quetame Group (Forero Suarez, 1990; Ordóñez-Carmona et al., 2006), and are not ophiolites (e.g. Steinmann et al., 2003), and (iii) the Early Ordovician “graptolite facies” rocks interpreted as the autochthonous Gondwanan pre-collision passive margin by Ramos (2009), traverse across the proposed suture zone from southwest of the San Lucas Range (Fig. 1), which forms part of the “Chibcha Terrane” (Ordóñez-Carmona et al., 2006), to the eastern Llanos Basin (Fig. 1), which overlies cratonic basement (Mojica and Villarroel, 1990). The presence of Early Palaeozoic sedimentary basins inboard of Palaeozoic orogenic belts along the Gondwanan margin (e.g. Sempere, 1995) may be a result of a rise in global sea level by almost 200m between the Early Cambrian and Late Ordovician (Haq and Schutter, 2008).

The Late Palaeozoic history of the Santander Massif is largely unconstrained by geochronological data (Fig. 5). Low grade metamorphism of the Late Carboniferous Diamante Fm. is considered to be coeval with a regional Late Permian metamorphic event within the Mérida Andes (Ward, 1974), which was inferred from Late Palaeozoic Rb/Sr and K/Ar dates. Regional models suggest that Latest Permian deformation recorded in the Sierra Nevada de Santa Marta and the Central Cordillera of Colombia (Fig. 1) may be related either to terrane accretion or increased plate coupling in a magmatic arc setting (e.g. Vinasco et al., 2006; Cardona et al., 2010). This period of compression was followed by Early to Middle Triassic extension recorded in the Central Cordillera of Colombia and in Ecuador by the formation of sedimentary basins and the coeval intrusion of T-MORB amphibolites (e.g. Villagómez et al., 2011; Cochrane et al., 2014; Spikings et al., 2014). The effect of Early to Middle Triassic extension on the rocks of the Mérida Andes and the Santander Massif, where there are no known Triassic sedimentary rocks, is not clear.

The Mesozoic batholiths of the Santander Plutonic Group (Fig. 4) yield zircon U-Pb ages ranging from ~210 Ma to ~196 Ma and calc-alkaline, continental arc geochemistry (Fig. 5; Dörr et al., 1995; Mantilla et al., 2013), although all of these data were obtained from a relatively small area near our sample 10VDL52 (Fig. 4) in the central part of the Santander Massif. These granitoids are interpreted to have formed as a result of subduction of oceanic crust beneath a continental arc, and

are older than arc related granitoids in the Central Cordillera (<189 Ma; Bustamente et al., 2011; Cochrane et al., 2014; Spikings et al., 2014).

4. Analytical Methods

4.1. Zircon U-Pb Geochronology

Weathered portions of samples were removed with a saw, and the remaining sample material was crushed with a jaw crusher. The crushed rock fraction was split and a representative fraction was separated for whole rock and isotope geochemistry. Zircons were extracted from the remainder of the crushed rock using conventional heavy liquid and magnetic mineral separation techniques. Hand picked, inclusion free, euhedral zircons were mounted in epoxy discs and polished to expose inner portions. The polished disks were coated in carbon and the internal structure of the zircons was imaged with cathodoluminescence on a CamScan MV2300 scanning electron microscope. Zircons were analysed for U-Pb geochronology on a Thermo Scientific Element XR ICP-MS at the Institute of Mineralogy and Geochemistry, University of Lausanne, Switzerland. Selected 35 μm spots were ablated from zircons using a NewWave UP-193 ArF excimer ablation system at 5 Hz for 30 seconds, with a beam intensity of 2.2-2.6 J/cm². Ablated material was transported to the instrument using purified helium as a carrier gas. Primary standardization was made using GEMOC GJ-1 (CA-ID-TIMS ²⁰⁶Pb/²³⁸U age of 600.5 $\pm$ 0.4 Ma; Schaltegger et al., unpublished). The secondary standards Harvard 91500 (1065 $\pm$ 0.4 Ma ; Wiedenbeck et al., 1995) and Plešovice zircon (337.13 $\pm$ 0.37 Ma; Slama et al., 2008) were utilised for calibration and routine monitoring of the consistency of measured U-Pb dates. Ages and uncertainties were calculated using LAMTRACE (Jackson, 2008) using the ratio of the mean intensities (Ulianov et al., 2012). Signals for every analysis were routinely monitored for isotopic fractionation, common Pb and Hg-induced isobaric interference by inspection of time resolved intensity ratios and masses 204 and 202, respectively. Analyses with detectable Hg or common Pb were discarded. One secondary standard was measured for each three sample zircons, and measured average ages and their respective standard deviations were 1057.1 $\pm$ 11.5 Ma for Harvard 91500 (251 analyses) and 337.1 $\pm$ 3.7 Ma for Plešovice zircon (63 analyses). Statistical analysis of filtered data was performed using ISOPLOT 3.75 (Ludwig, 2003). For magmatic and metamorphic samples (Table 1), concordia plots were constructed and all discordant points were discarded unless they defined a discordia between two populations. Weighted mean ²⁰⁶Pb/²³⁸U ages with a mean square of weighted deviates (MSWD) larger than 2.5 indicate excess scatter that may reflect either minor inheritance or Pb loss – our data was acquired without zircon chemical abrasion (e.g. von Quadt et al., 2014) and do not permit distinction between these possibilities. For detrital sample 08VDL05, random zircons were analysed and all

points that were >5% discordant were discarded. Dates included in the interpretation of data acquired from detrital zircons are $^{206}\text{Pb}/^{238}\text{U}$ dates for analyses <900 Ma, and $^{207}\text{Pb}/^{206}\text{Pb}$ dates for analyses >900 Ma. All ages are reported with an analytical uncertainty of $\pm 2\sigma$.

4.2. Geochemistry

Geochemical and Pb isotopic analyses were performed on representative whole rock powder fractions prepared using an agate mill. Major and trace elements were measured using a Philips PW2400 X-Ray Fluorescence (XRF) spectrometer. Additional trace elements and rare earth element abundances were measured at the University of Lausanne using a Perkin Elmer ELAN 6100 DRC quadrupole LA-ICP-MS on the fused glass disks used for XRF, with NIST SRM610 as the external standard. The GeoLas 200 M 193 nm excimer laser was fired at 10Hz with a spot size of 120 μm and ablation energy of 140 mJ. Sr concentration previously determined by XRF was utilised as the internal standard. When Sr concentration was lower than 100 ppm, the CaO concentration determined by XRF was used. Concentrations were calculated offline using LAMTRACE (Jackson, 2008), and the reported values are the averages of 3 spots per sample. Uncertainties between the results of repeated points of the same sample are estimated to be <10% for REE and <5% for the other trace elements.

4.3. Pb isotope geochemistry

Pb isotopic compositions were determined from whole rock powders of 36 igneous and metamorphic rocks. Approximately 100 mg of whole rock powder was dissolved in 4 ml of concentrated HF and 1 ml of 15M HNO_3 in closed Teflon vials at 140 °C for seven days. The samples were dried down and re-dissolved in 3 ml of 15M HNO_3 before being dried down again. Pb chemical separation was achieved following the methods outlined in Pin et al. (1997) and Chiaradia et al. (2011). Pb isotopic measurements were made using a Thermo Finnigan Triton Thermal Ionization Mass Spectrometer (TIMS) or a Nu Instruments Neptune MC-ICP-MS at the University of Geneva, Switzerland. Typical analytical uncertainties were 0.0075% on all measured ratios.

5. Results

5.1. Petrology and geochronology

The rock type nomenclature of igneous rocks (Table 1) is based on the geochemical classification of De La Roche et al. (1980). K/Ar dates or Rb/Sr model ages have previously been obtained from a majority of the rock units that we analysed (Figs. 3 and 5), and these are briefly referred to below. A summary of all new zircon U-Pb data is presented in Table 1.

5.1.1. Caparo Block

U-Pb zircon ages were obtained from four granitoids of the Caparo Block (Fig. 2), which exhibit subsolidus deformation microfabrics of quartz with undulose extinction and variable grain size reduction. The oldest $^{206}\text{Pb}/^{238}\text{U}$ age of 499.4 ± 2.7 Ma was obtained from the Cerro Azul granodiorite (08VDL35, Table 1; Fig. 2), which was sampled from a large river boulder within an alluvial fan that is fed by a stream which only drains the Cerro Azul pluton and phyllites of the Cerro Azul Formation. The coarse, non-foliated Tapo granite is located in the central part of the Caparo Block and yields a $^{206}\text{Pb}/^{238}\text{U}$ age of 492.3 ± 5.3 Ma (08VDL57; Fig. 2; Table 1), providing a minimum Cambrian depositional age for the metasedimentary Bella Vista Suite, which it intrudes. The fine grained, two-mica El Cambur granite crops out 5 km upstream from granite 08VDL47 and gave a $^{206}\text{Pb}/^{238}\text{U}$ age of 492.3 ± 8.4 Ma (08VDL56; Table 1). Finally, the youngest granitoid is the medium-grained Las Islitas hornblende-biotite tonalite, which yielded a $^{206}\text{Pb}/^{238}\text{U}$ age of 473.1 ± 3.5 Ma (08VDL55; Table 1; Fig. 2). The tonalite is unconformably overlain by Early Llandovery (~444 Ma: Gradstein et al., 2012) marine sedimentary rocks of the El Horno Formation (Burkley, 1976; Gonzales de Juana et al., 1980), indicating that the tonalite was unroofed by ~444 Ma.

5.1.2 Mérida Block

5.1.2.1. Metamorphic basement: Sierra Nevada Suite

A paragneiss from the Sierra Nevada Suite (08VDL05; Fig. 2) hosts idiomorphic biotite and minor sillimanite indicating that it was metamorphosed at amphibolite facies conditions below the granitic solidus. Zircons lack metamorphic rims and yielded 109 concordant U-Pb age analyses, with peak $^{206}\text{Pb}/^{238}\text{U}$ age probabilities at ~525 Ma, 700-550 Ma, ~800 Ma, ~1000 Ma, and ~1100-1300 Ma (Fig. 6). Minor peaks occur at ~1350-1500 Ma and ~2000 Ma. The youngest detrital zircon yields an age of 505.0 ± 9.4 Ma, and the youngest cluster of 5 statistically equivalent analyses yields a weighted mean age of 518.7 ± 2.5 Ma (MSWD = 0.83), which is interpreted to indicate the maximum depositional age for the sedimentary protolith.

Orthogneiss 08VDL11 resembles biotite microgranite with up to ~10mm elongated aggregates of fibrolitic sillimanite, and small grains of anhedral orthopyroxene. Grain boundaries are lobate, suggesting that this unit was partially melted and recrystallized. Muscovite replaces sillimanite (Fig. 7A) and muscovite and biotite form symplectites with myrmekite (Fig. 7B). These microtextures are typical of post-anatexis retrograde reactions (e.g. Peterman and Grove, 2010). An older population of four concordant analyses yields a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 1008.6 ± 6.7 Ma

(Fig. 8), which may represent the age of the magmatic protolith. A younger population of three concordant analyses yields a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 454 ± 10 Ma (Table 1), which may reflect the timing of anatexis. Six analyses yield ages which define a chord between the two populations on a U-Pb Concordia diagram (Fig. 8).

The La Raya granitic gneiss (08VDL27) is an equigranular, muscovite – garnet leucogranite which yields a zircon $^{206}\text{Pb}/^{238}\text{U}$ age of 488.5 ± 6.5 Ma (Fig. 2; Table 1). Muscovite is aligned and quartz grains have undulose extinction but grain boundaries do not exhibit any strain or recrystallisation. These features define a fabric which is indicative of syn-magmatic deformation (Passchier and Trouw, 2005). The La Raya gneiss and the adjacent Mitisus striped gneiss (08VDL26; Fig. 2) are considered to be parts of the Sierra Nevada Suite (Schubert, 1969). The microtexture of Mitisus gneiss 08VDL26 features aligned quartz ribbons in addition to macroscopic quartz and feldspar augen porphyroclasts with small biotite crystals in strain shadows (Fig. 7C), which may have formed during high grade sub-solidus shearing (Passchier and Trouw, 2005). Analysed zircons from the Mitisus gneiss yield a range of concordant, non-equivalent ages, which may be a result of Pb loss during deformation. The oldest group of equivalent, concordant analyses yield a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 488.2 ± 4.3 Ma ($n=6$; MSWD=2.0), which is statistically equivalent to the age of the adjacent La Raya granitic gneiss (488.5 ± 6.5 Ma), suggesting that the Mitisus striped gneiss may be a sheared member of the latter. The foliation planes of the Mitisus striped gneiss are roughly parallel to the Boconó Fault system (Schubert, 1969, Audemard and Audemard, 2002) and dip steeply to the SE, with a NE-SW strike direction. The timing of high-grade deformation of this unit is not constrained by our data.

Strongly foliated augen gneisses of the Sierra Nevada Suite near Puente Real (08VDL36; Fig. 2) yield a weighted mean zircon $^{206}\text{Pb}/^{238}\text{U}$ age of 472.3 ± 3.4 Ma (Table 1). Burkley (1976) utilised structural observations to conclude that the Puente Real gneiss was emplaced during metamorphism. Our sample has a metamorphic overprint defined by quartz recrystallisation and the growth of epidote, which may reflect post-magmatic deformation under greenschist conditions.

A strongly foliated gneiss from the Sierra Nevada Suite was collected at La Playita (08VDL22; Fig. 2), and yields a zircon $^{206}\text{Pb}/^{238}\text{U}$ age of 449.7 ± 4.4 Ma (Table 1). Larger biotite and feldspar porphyroclasts define an augen texture, whereas smaller biotite and muscovite crystals are undeformed and aligned. Quartz crystals between the mica layers are preferably elongated and

partially recrystallized, suggesting that much of the deformation in our sample arose from post magmatic low to medium grade shearing (Passchier and Trouw, 2005).

5.1.2.2 Granitoids intruding the Iglesias Complex

Zircon U-Pb ages are reported from eleven granitoids which intrude the Iglesias Complex metamorphic basement. These units are generally unfoliated and yield weighted mean Early Palaeozoic zircon $^{206}\text{Pb}/^{238}\text{U}$ ages that range between 471.8 ± 4.0 Ma and 414.5 ± 3.9 Ma (Table 1).

A rhyolitic tuff (08VDL15) from the footwall of the Bailadores volcanogenic massive sulphide (VMS) deposit (Carlson, 1977; Staargaard and Carlson, 2000) yields a weighted mean zircon $^{206}\text{Pb}/^{238}\text{U}$ age of 452.6 ± 2.7 Ma. The 1-2 km thick tuffs unconformably overlie the Sierra Nevada Suite and are interfingered with low grade metasedimentary rocks of the Mucuchachi Formation at their top (Carlson, 1977). VMS mineralisation is of the bimodal-siliciclastic or Kuroko type (Staargaard and Carlson, 2000), suggesting that the tuffs were deposited in a basin over the metamorphic basement.

The Pueblo Hondo biotite granite (08VDL04) has a surface area of ~ 200 km² (Fig. 2), intrudes the Tostos Suite, and yields a weighted mean zircon $^{206}\text{Pb}/^{238}\text{U}$ age of 434.2 ± 4.4 Ma. The Soledad granodiorite (08VDL32) gave the youngest Palaeozoic zircon mean $^{206}\text{Pb}/^{238}\text{U}$ age in the Mérida Andes of 414.5 ± 3.9 Ma. The emplacement of this unit generated an andalusite hornfels contact aureole in the Cerro Azul Suite (Gonzales de Juana et al., 1980), indicating an Early Devonian minimum age for the host rock. The Cerro Azul Suite crops out on both sides of the Caparo-Soledad fault (Schubert, 1969), and its Early Palaeozoic or older age suggests that basement blocks on either side of that fault were continuous in the Early Palaeozoic, challenging the hypothesis that the Caparo Soledad Fault is a Carboniferous suture (Bellizzia and Pimentel, 1994; Aleman and Ramos, 2000).

Four granitoids and one dacite collected from the Mérida Andes yield Triassic weighted mean zircon $^{206}\text{Pb}/^{238}\text{U}$ ages, which range between 243.5 ± 3.4 Ma and 202.0 ± 1.8 Ma (Fig. 2, Table 1).

The basal interval of the syn-rift La Quinta Formation (e.g. Maze, 1984) is composed of ~ 150 m thick biotite bearing rhyodacite tuffs (08VDL18; Fig. 2), which yield a zircon U-Pb age of 202.0 ± 1.8 Ma (Table 1). The tuffs unconformably overlie Palaeozoic low grade metasedimentary rocks of the Mucuchachi Fm. The La Quinta Formation includes all non-metamorphic, red

continental siliciclastic rocks in the Mérida Andes, and published palaeontological data suggest that the top of the formation may be as young as Cretaceous (Maze, 1984). Our U-Pb age of 202.0 ± 1.8 Ma is a maximum age for the base of the La Quinta Fm in the Merida Andes.

5.1.3. Granitoids from isolated outcrops in Venezuela

Three magmatic intrusions were collected from isolated outcrops located around the northern shore of Lake Maracaibo (Fig. 1). The Amparo hornblende-biotite diorite (08VDL64), located in the centre of the Paraguaná peninsula yields a weighted mean zircon U-Pb age of 272.2 ± 2.6 Ma (Fig. 1). The pluton is unconformably overlain by Upper Jurassic sedimentary rocks (Odremán and Useche, 1997).

Toas Island (Fig. 1) is composed of an outcrop of granitic basement along the Oca Fault, which forms the northern boundary of the Maracaibo Block. The Toas Island granodiorite (08VDL59) gave a zircon U-Pb age of 248.9 ± 1.7 Ma and is locally overlain by red sandstones of the Mesozoic La Quinta Formation.

A weathered pink equigranular biotite granite was collected from the foothills of the Serranía del Perijá (Fig.1; 08VDL60), which yielded a zircon U-Pb age of 225.1 ± 1.5 Ma. This granite is unconformably overlain by Jurassic rocks of the La Quinta Formation (e.g. Maze, 1984; Shagam et al., 1984).

5.1.4. Santander Massif

5.1.4.1. Metamorphic Basement

The Bucaramanga gneiss (Fig. 4), located in the central Santander Massif, hosts subparallel stromatic leucosomes which locally form isoclinal s-folds. An undeformed decimetric dike of the Mesozoic Santander Plutonic Group crosscuts foliation of the outcrop. Leucosome 10VDL50 is composed mainly of quartz, feldspar, plagioclase and minor biotite. Many zircons from this sample are rounded, suggesting that they were inherited from a sedimentary protolith. Rims from 9 euhedral zircons yield a $^{206}\text{Pb}/^{238}\text{U}$ age of 477.0 ± 5.3 Ma, with a large scatter (MSWD = 7.5; Table 1) that may be a result of accidental analysis of small zircon cores or partial Pb loss, possibly during intrusion of the crosscutting dike.

A second sample of Bucaramanga Gneiss (10VDL23) forms alternating 5 cm thick bands. The dark bands are mainly composed of hornblende and feature plagioclase poikiloblasts incorporating

hornblende and chlorite, which alternate with leucocratic bands composed mainly of plagioclase, minor hornblende and fresh biotite. Zircon forms inclusions in all major mineral phases in both bands, and all 18 analysed oscillatory zoned zircons define a single population with a weighted mean age of 461.0 ± 2.1 Ma. These relationships suggest that this rock had a mafic to intermediate igneous protolith which underwent post-magmatic deformation at lower amphibolite facies. Our zircon U-Pb age most likely records the time of crystallisation of the igneous protolith.

A foliated granite of the Berlin Orthogneiss (10VDL44; Fig. 4) hosts thin shear zones with <http://www.geosociety.org/pubs/editorsCall.htm> muscovite, biotite and epidote that define a strong foliation, although it is unclear from the microtexture of our sample whether or not deformation was syn- or post-magmatic. Oscillatory zoned zircon rims yield a weighted mean zircon $^{206}\text{Pb}/^{238}\text{U}$ age of 473.6 ± 2.5 Ma.

A sample of granodiorite orthogneiss from the Páramo de Berlin (10VDL51) contains biotite and minor fibrolitic sillimanite which define a strong foliation around aggregates of felsic minerals. Unaligned but slightly undulose quartz forms highly lobate grain boundaries and fills holes in aligned biotite, which suggests that melt was present after grain alignment. Migmatites are absent from the outcrop. Seventeen zircons yield concordant dates that range between 483.3 ± 5.7 Ma and 462.7 ± 3.5 Ma, and define a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 472.5 ± 3.4 Ma. The high age dispersion (MSWD = 11.5, Table 1) suggests that zircons from this sample may either record both crystallisation and subsequent anatexis or include some small xenocrystic cores which were not detectable by cathodoluminescence image analysis. Alternatively, zircons may have undergone minor Pb loss as a result of reheating during the intrusion of the Páramo Rico pluton at ~205 Ma (Dörr et al., 1995), as suggested by early Jurassic K/Ar (biotite) dates obtained from biotite gneiss in this area (Goldsmith et al., 1971).

A tonalitic hornblende-biotite orthogneiss (10VDL37) forms a sill that intrudes the Silgará Formation and yields a zircon $^{206}\text{Pb}/^{238}\text{U}$ age of 451.5 ± 1.3 Ma. The host rocks at this location (Fig. 4; Table X) have a lower metamorphic grade (Ward et al., 1974). Foliation in this sample is less regular and defined by slightly folded biotite, which is wrapped around dislocation-rich plagioclase porphyroclasts, indicating that deformation in this rock may have occurred below the solidus. This sill is crosscut by deformed migmatitic leucosomes (10VDL39) and undeformed dikes (Fig. 9). Leucosome 10VDL39 yields a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 208.8 ± 1.2 Ma.

An undeformed pegmatite (10VDL43) crosscuts the foliation of the host orthogneiss and contains zircons which yield two distinct U-Pb age populations with weighted mean ages of 460.3 ± 3.5 Ma (6 analyses; Table 1) and 209.2 ± 3.4 Ma (5 analyses, Table 1). Additional analyses from 6 zircons define a discordia between these two age populations. We interpret the older population as the age of the orthogneiss host rock, and the younger population as the crystallisation age of the pegmatite.

Fine grained, massive hornblende-biotite gneiss from the north-northwest of Pamplona (10VDL49; Fig. 4) contains oscillatory zoned zircons that yield a weighted mean zircon $^{206}\text{Pb}/^{238}\text{U}$ age of 479.8 ± 3.1 Ma. The gneiss is mapped as the Silgará Formation and has a dark matrix with aligned hornblende and biotite, and fine leucocratic bands giving it a strong foliation. Plagioclase in this rock is mostly replaced by prehnite and epidote, revealing a low grade metamorphic overprint.

5.1.4.2. Palaeozoic plutons

Three unfoliated plutonic rocks which crop out in the Santander Massif yield Early Palaeozoic weighted mean zircon $^{206}\text{Pb}/^{238}\text{U}$ ages of 483.7 ± 5.9 (10VDL47), 443.4 ± 3.2 Ma (10VDL55) and 439.2 ± 4.7 Ma (10VDL46). The oldest of these, gabbrodiorite 10VDL47, was collected from a 3m^3 boulder which lay below a ridge, and a local source was inferred although its field relationships are unclear. The two younger granitoids crosscut foliation of the Silgará Formation and thus postdate regional metamorphism.

A dike of grey vesicular porphyritic rhyolite crosscuts the Bocas Formation to the west of the Santa Marta - Bucaramanga Fault (Fig. 4; 10VDL58). The strained nature of some quartz grains compared with the generally round shape of vesicles suggests that this rock was contaminated by older material, which is reflected in the zircon U-Pb ages. The youngest coeval zircons ($n=6$) yield a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 250.7 ± 4.3 Ma, and 10 inherited grains have ages that range between 449.9 ± 9.7 Ma and 1316.9 ± 25.7 Ma. If the youngest coeval zircon population crystallised from the rhyolitic melt, then the Bocas Fm. is Permian or older, although a range of ages from Carboniferous to Jurassic were previously assigned to this unit (Ward et al., 1974; Mojica et al., 1996).

5.1.4.3. Mesozoic igneous rocks

Intermediate to felsic granitoids of the Triassic – Jurassic Santander Plutonic Group (Fig. 4) are exposed over a surface area of more than 4000 km^2 (Fig. 4) and yield 11 weighted mean $^{206}\text{Pb}/^{238}\text{U}$ zircon ages ranging between 202.2 ± 1.0 and 195.8 ± 1.5 Ma (Table 1). Several of these rocks have

porphyritic or granophyric textures suggesting that they cooled rapidly at shallow levels in the crust. Granodiorite 10VDL61 contains numerous 1-20 cm sized mafic enclaves composed of biotite pseudomorphed by epidote, with corundum, white mica and sericite. Euhedral zircons from this sample yield two populations (Table 1), which are (i) a younger population of 10 grains with a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 200.0 ± 1.5 Ma, which we interpret to be the crystallisation age of the granodiorite, and (ii) an older population of 6 grains that yield a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 1018.3 ± 8.9 Ma (Fig. 10), which may be related to the numerous mafic enclaves. The mineral assemblage of these enclaves suggests either a pelitic or silica undersaturated igneous protolith, although we favour a metaigneous protolith because of the euhedral shape and age distribution of the zircons in two distinct, clearly defined populations.

5.2. Geochemistry

The granitoids and felsic volcanic rocks exposed in the Santander Massif and the Mérida Andes have a wide range of SiO_2 content ranging from 54% to 75% and generally plot along a high-K calc-alkaline trend (Fig. 11A). Most of the Phanerozoic igneous rocks are mildly to strongly peraluminous although rocks in the Santander Massif are generally less peraluminous than coeval igneous rocks in the Mérida Andes (Fig. 11b). The large range in Shand's Index (A/CNK ; Maniar and Piccoli, 1989) suggests that some of the studied granitoids may be derived from the anatexis of pelitic rocks, while others may be derived from igneous sources. Large variations in Na_2O compared to K_2O (Fig. 11c) suggest that both S- and I-Type granitoids (Chappell and White, 1974) were coeval. For example, monzodiorite 08VDL47 and granodiorite 08VDL45 are coeval within uncertainty (Table 1) and were collected from the same batholith, although the former is mildly peraluminous ($\text{A}/\text{CNK}=1.03$, appendix) and plots in the I-Type field (Fig. 11c), whereas the latter 08VDL45 is strongly peraluminous ($\text{A}/\text{CNK}=1.20$, appendix) and plots in the S-Type field (Fig. 11c). Early Palaeozoic igneous rocks may be a result of mixing of variable proportions of melts derived from different sources (e.g. Patiño-Douce, 1999; Maas et al., 1997).

Late Triassic to Jurassic granitoids from the Santander Massif are mildly peraluminous and have K_2O and Na_2O compositions typical of I-Type granites (Fig. 11c), corroborating the presence of hornblende and titanite. In contrast, late Triassic granitoids from Venezuela are more peraluminous with muscovite and biotite as coexisting phases, and contain K_2O and Na_2O compositions typical of both I- and S-Type granites.

The magmatic rocks yield trace element compositions that are very similar to average upper continental crust (Taylor and McLennan, 1985), with the exception of Nb and Ta concentrations, which are depleted in rocks of all age groups (Fig. 12). The low Nb and Y contents of most rocks (Fig. 11d) are characteristic of rocks that formed in convergent margins (Pearce, 1996). REE abundances are generally similar to upper continental crust, with variable positive or negative europium anomalies (Fig. 12).

5.3. Pb isotope geochemistry

Thirty-six intrusive and volcanic rocks from the studied regions yield present day Pb compositions with more radiogenic $^{207}\text{Pb}/^{204}\text{Pb}$ compared to $^{206}\text{Pb}/^{204}\text{Pb}$ ratios relative to the Stacy and Kramers (1975) terrestrial Pb evolution curve (Fig. 13, Table 2). Samples collected from the Mérida Block, the Caparo Block and the Santander Massif yield indistinguishable Pb isotopic trends (Fig. 13). Rocks from Toas Island, the Perijá Foothills and the Paraguana Peninsula yield slightly more juvenile Pb compositions, although they lie on the same general trend. The crustal evolution trend of all analysed samples is similar to that of rocks exposed in the Guyana Shield in southern Venezuela, the Garzón Massif in Colombia and the Amazonian Craton in western Brazil (Fig. 13). In contrast, Laurentian cratonic rocks or the Laurentia – derived Cuyania Terrane (e.g. Astini et al., 1995 and references therein) yield very different trends, which lie below the terrestrial Pb evolution curve (Fig. 13).

6. Interpretation

6.1. Periods of magmatic activity in Venezuela and the Santander Massif

6.1.1 Late Mesoproterozoic

An orthogneiss (08VDL11) and a granodiorite (10VDL61) of the Sierra Nevada Suite and the Santander Plutonic Group, respectively, yield two populations of zircons. The older, concordant analyses define weighted mean ages of 1008.6 ± 6.7 Ma (Fig. 8) and 1018.3 ± 8.9 Ma (Fig. 10). Orthogneiss 08VDL11 underwent anatexis at ~ 454 Ma, whereas granodiorite 10VDL61 may have crystallised at ~ 200 Ma while incorporating Proterozoic material. These older populations provide indirect evidence for ~ 1 Ga basement in the Mérida Andes and support previous hypotheses that ~ 1 Ga basement is present in the central Santander Massif (Fig. 1; e.g., Restrepo-Pace et al., 1997), although it was largely reworked by Phanerozoic magmatism and metamorphism. Reliable U-Pb data and Sm-Nd isochrons reveal the presence of ~ 1 Ga basement in the northern Santander Massif, the Sierra Nevada de Santa Marta, the Garzón Massif and beneath the Llanos Basin (Fig. 1, e.g.

Restrepo-Pace et al., 1997; Cordani et al., 2005; Ibanez-Mejia et al., 2011), and rocks with similar ages can be found along extensive tracts of the Andean margin (e.g. the Arequipa Terrane; Ramos, 2010 and references therein).

6.1.2 Late Neoproterozoic

Late Neoproterozoic rocks were not sampled in this study, although previous authors (Aleman and Ramos, 2000; Chew et al., 2008) infer that they exist within the Mérida Andes based on a Rb/Sr model age and U-Pb data (Fig. 3; Compañía Shell de Venezuela and Creole Petroleum Corporation, 1964; Burkley, 1976). However, (i) all of the Late Neoproterozoic U-Pb dates obtained from the Mérida Andes are strongly discordant (Burkley, 1976) and their accuracy is questionable, and (ii) no supporting data for the Rb/Sr model age of ~660 Ma (Compañía Shell de Venezuela and Creole Petroleum Corporation, 1964) has been published and its age may be anomalously old because of incorporated excess ^{87}Sr during crystallisation. The only precise Late Neoproterozoic U-Pb data from Venezuela or Colombia are a zircon U-Pb age of $577.8 \pm 6.3/-9$ Ma (LA-ICP-MS; Arango Mejía et al., 2012) obtained from a nepheline syenite with within-plate geochemistry, located in an inlier in the Llanos Basin (Fig. 1). There is stronger evidence for Neoproterozoic basement in Perú (~752 Ma - 691 Ma; Chew et al., 2008; Mišković et al., 2009) than in the Andes of Colombia and Venezuela.

6.1.3. Early Palaeozoic

The Mérida Andes and the Santander Massif host Early Palaeozoic magmatic rocks which yield $^{206}\text{Pb}/^{238}\text{U}$ ages that range between $499.4 \pm 2.7 - 414.5 \pm 3.9$ Ma and $483.7 \pm 5.9 - 439.2 \pm 4.7$ Ma, respectively (Fig. 14). In Venezuela, Early Palaeozoic igneous rocks are also found in the Caribbean Mountains (Sisson et al., 2005), the El Baúl Massif (Viscarret et al., 2009) and in the Llanos Basin (Feo-Codecido et al., 1984) north of the Apure Fault (Fig. 1). Early Palaeozoic magmatism in Colombia is also recorded in the Quetame and Floresta massifs of the Eastern Cordillera (Fig. 1), where granitoids yield zircon U-Pb ages that range between 520 and 420 Ma (Horton et al., 2010), and it occurs in the Central Cordillera (La Miel Gneiss; e.g. Villagómez et al., 2011).

Published in-situ U-Pb zircon ages from all basement regions in the Northern Andes (Fig. 1) suggest that a regional magmatic hiatus lasted from ~415 Ma until ~294 Ma. This observation is supported by a paucity of Devonian to Carboniferous detrital zircons in the Eastern Cordillera of Colombia (Horton et al., 2010). We do not consider the previous two Devonian U-Pb dates from the

Caparo Block (Fig. 3; Burkley, 1976) to be accurate because (i) a lower intercept date of 350 ± 50 Ma was obtained from a mix of rounded zircons and unidentified material (not zircon; Burkley, 1976) extracted from a metasedimentary rock, and its significance is unclear, and (ii) the 390 ± 30 Ma date is a $^{207}\text{Pb}/^{206}\text{Pb}$ intercept from discordant data resulting from large amounts of incorporated common Pb (Burkley, 1976).

The only Permian granitoid analysed during this study is the El Amparo Diorite, which crops out in the Paraguana Peninsula (08VDL64; 272.2 ± 2.6 Ma; Fig. 1). This intrusion is located in the Bonaire Block, which has accommodated dextral displacement between the Caribbean and the South American plates since the Palaeocene (e.g. van der Lelij et al., 2010). Several Permian granitoids crop out in the Sierra Nevada de Santa Marta (Fig. 1; Cardona et al., 2010), the Central Cordillera of Colombia (Cochrane et al., 2014) and the El Baúl Massif (Fig. 1; Viscarret et al., 2009), which are interpreted to be continental arc rocks.

6.1.4. Mesozoic

Early to Middle Triassic U-Pb ages were obtained from small granitic bodies in the Perijá foothills, Toas Island, Mérida Andes and a rhyolitic dike crosscutting the Bocas Fm. in the Santander Massif (Fig. 14; Table 1). Permian to Middle Triassic igneous and metamorphic rocks are widespread in the Central Cordillera of Colombia and the Sierra Nevada de Santa Marta (e.g. Vinasco et al., 2006; Montes et al., 2010; Restrepo et al., 2011; Villagómez et al., 2011; Spikings et al., 2014). Late Triassic to Early Jurassic magmatism may have been relatively continuous in the Mérida Andes and Santander Massif, spanning the interval 213.2 ± 1.7 Ma to 195.8 ± 1.5 Ma (Fig. 14). There are no known coeval magmatic rocks in the rest of the Northern Andes, and the earliest Jurassic magmatism in the Central Cordillera (Fig. 1) commenced at ~ 189 Ma (Bustamante et al., 2010).

6.2. Conditions and timing of metamorphism

6.2.1. Mérida Block

The Sierra Nevada Suite underwent amphibolite facies Barrovian metamorphism characterised by clockwise PT paths (e.g., Grauch, 1972). The youngest detrital zircon within amphibolitic metasedimentary rocks of the sampled Sierra Nevada Suite yields a $^{206}\text{Pb}/^{238}\text{U}$ age of 505.0 ± 9.4 Ma, representing a maximum stratigraphic age. Structural mapping suggests that some granitoids were emplaced prior to or during metamorphism (e.g. Schubert, 1969; Kovach et al., 1977), and existing maps include these foliated granitoids within the Sierra Nevada Suite metamorphic basement complex. Four foliated magmatic rocks from the basement complex yield ages ranging between

488.5±6.5 Ma and 449±4.4 Ma. The youngest of these samples exhibits sub-solidus, mylonitic deformation and is located close to the Boconó Fault, suggesting that its deformation may be unrelated to a regional metamorphic event.

The foliation of the Sierra Nevada Suite is crosscut by numerous granitoids with ages that range between 471.8±4.0 Ma (08VDL21) and 447.9±1.8 Ma (08VDL45), suggesting that metamorphism occurred in the interval ~505 – 472 Ma. A biotite-muscovite-sillimanite orthogneiss (08VDL11; Fig. 2) sampled from the same fault block studied by Grauch (1972) exhibits a microtexture that is typical of retrogression from partial melting, and we interpret the younger population (454±10 Ma; Fig. 8; Table 1) to reflect the timing of anatexis under upper amphibolite facies conditions during retrogression. Many Early Palaeozoic granitoids have epidote and chlorite replacing amphibole and biotite, suggesting low grade alteration.

6.2.2. Caparo Block

The greenschist grade Bella Vista phyllites of the Caparo Block were intruded by non-deformed to weakly foliated granitoids between ~492 Ma and ~473 Ma (Fig. 14), and are unconformably overlain by the strongly folded but non-metamorphosed Late Ordovician Caparo Formation (Odremán and Useche, 1997). This suggests that the Caparo Block was exhumed before the Late Ordovician. Field relationships are not sufficiently clear to determine the temporal relationship between the small intrusions in the Caparo Block and the metamorphism of the host rock, but the Late Ordovician Caparo Fm. (Gonzales de Juana et al., 1980) provides a minimum age for the greenschist facies metamorphism of the Bella Vista Suite.

6.2.3. Santander Massif

Recent studies describe medium pressure, Barrovian metamorphism of the Bucaramanga Gneiss and the Silgará Formation in the central Santander Massif (García et al., 2005; Castellanos et al., 2008; Urueña-Suárez and Zuluaga, 2011). In this locality, isograd boundaries have a north-south orientation and metamorphic grades of the Silgará Fm. increase westwards towards the Bucaramanga Gneiss, near samples 10VDL50 and 10VDL51 (Fig. 4). A leucosome within the Bucaramanga Gneiss (10VDL50) was sampled in the highest grade zone in the central Santander Massif, and yields a weighted mean age of 477.0±5.3 Ma. Migmatization of the Bucaramanga gneiss may have occurred near peak conditions, and this Early Ordovician age is our best estimate for the timing of peak metamorphism in this region.

The regional metamorphic foliation of the Bucaramanga Gneiss and Silgará Formation is parallel with the pre-full crystallisation foliation of the intruding orthogneiss units (Ward et al., 1974). Recent work suggests that the orthogneiss in the central Santander Massif was deformed as it intruded during metamorphic retrogression (García et al., 2005; Castellanos et al., 2008), which is consistent with our petrographic observations. A sample of sillimanite orthogneiss from the central Santander Massif yields an Early Ordovician age (472.5 ± 3.5 Ma; 10VDL50), coeval with intrusion of orthogneiss 10VDL44 at 473.5 ± 2.5 Ma in the eastern Santander Massif near Pamplona (Table 1, Fig. 4). These relationships suggest that the basement rocks of the central Santander Massif were on the retrograde path above the sillimanite isograd at ~ 472 Ma. The foliation of the orthogneiss and the Silgará Formation is crosscut by the undeformed Pamplona Granite (10VDL46, Ward et al., 1974), suggesting that regional metamorphism had terminated by 439.2 ± 4.7 Ma. Consequently, Barrovian metamorphism may have been coeval in the Mérida Andes and the Santander Massif.

Ordovician U-Pb ages of 461.0 ± 2.1 Ma (10VDL23) and 451.5 Ma (10VDL37) have been obtained from the Bucaramanga Gneiss and orthogneiss. These rocks show evidence for post-magmatic deformation at upper greenschist to lower amphibolite facies conditions, and our age data from cross-cutting rocks indicate that they were deformed before ~ 209 Ma (Fig. 9). Clockwise PT paths indicating similar conditions to the central Santander Massif have been obtained from rocks of the Silgará Formation in the southwestern Santander Massif (Ríos et al., 2003). Our age data only constrain metamorphism in the southwestern Santander Massif to have occurred prior to $\sim 199.1 \pm 1.3$ Ma (granodiorite 10VDL22; Fig. 4).

6.3. The Northern Andes: autochthonous blocks or allochthonous terranes?

Several lines of evidence strongly support an autochthonous origin for the basement sequences of the Merida Andes and the Santander Massif. First, the basement rocks from the Caparo Block, Mérida Andes and the Santander Massif collectively yield Pb isotopic ratios which plot in a narrow field (Fig. 13), suggesting that they originated from crustal sources with a similar magmatic and metamorphic history. Their isotopic composition is compatible with crustal sources within the Amazonian Craton and the Guyana Shield, but it is distinct from Laurentian Grenvillian-aged basement. MP-HT metamorphism occurred between ~ 505 Ma and ~ 472 Ma in both the Santander Massif and the Mérida Andes (Fig. 14), and was coeval with unconformities between metamorphic basement and Late Ordovician sedimentary rocks in the Caparo Block, which suggests that all three areas were on the same margin during the Early Palaeozoic. The lower degree of metamorphism

within the exposed basement rocks of the Caparo Block may be related to differential exhumation along the northwest dipping Caparo-Soledad reverse fault (Fig. 2; e.g. Colletta et al., 1997).

Second, Late Neoproterozoic ages were obtained from more than a third of the detrital zircons in Sierra Nevada gneiss 08VDL05 (Fig. 6). Late Neoproterozoic felsic magmatism is scarce in Laurentia and Baltica (e.g. Nance et al., 2008 and references therein), suggesting that detritus in the Sierra Nevada Fm. was derived from Gondwanan (e.g. Chew et al., 2008) or peri-Gondwanan (e.g. Nance et al., 2008) sources in the Late Cambrian prior to metamorphism in the Early Ordovician.

Unfortunately, there are no interpretable palaeomagnetic data from Palaeozoic rocks of the Northern Andes. Palaeomagnetic data from Jurassic and Cretaceous strata suggest that the rocks of the Mérida Andes lack latitudinal displacement relative to the Guyana Shield (Castillo et al., 1991; Bayona et al., 2010), in contrast to the Sierra Nevada de Santa Marta and the Floresta Massif (Fig. 1) which may have moved northwards by $\sim 10^\circ$ since the Jurassic (Bayona et al., 2010). Miocene transpression led to the northeastern extrusion of the Maracaibo Block by approximately ~ 290 km (Dewey and Pindell, 1985).

Insufficient data are available from Palaeozoic units in the pericratonic basins to establish reliable correlations with the metamorphic rocks of the Mérida Andes and the Santander Massif. The oldest known sedimentary sequences in the pericratonic Venezuelan Llanos Basin are Early Cambrian graben filling sedimentary rocks in the Espino Graben (Fig. 1; e.g. Feo-Codecido et al., 1984), which are unconformably overlain by Permian to Cretaceous sequences. New seismic data hints at an Ordovician fold and thrust belt in the Colombian Llanos Basin (e.g. Moreno-Lopez, 2012; Suárez Díaz and Solano, 2012), arising from early Palaeozoic compression which may have been contemporaneous with Barrovian metamorphism in the Mérida Andes and the Santander Massif. This suggests that the Llanos Basin, the Mérida Andes and the Santander Massif were simultaneously subjected to similar tectonic regimes.

Geochronological and isotopic analyses of basement blocks in the Santander Massif and the Mérida Andes are consistent with them being derived from Gondwanan sources, and these blocks have experienced very similar tectono-metamorphic histories (Fig. 14) since at least ~ 500 Ma. Multiple accretionary episodes of allochthonous terranes are not required to explain any existing data obtained from rocks exposed in the Mérida Andes and the Santander Massif, and we recommend that the terms “Chibcha Terrane” and “Mérida Terrane” should no longer be used. This conclusion

is similar to that reached by Spikings et al. (2014) for the Tahami and Chaucha Terranes of Colombia and Ecuador, who suggest that these are autochthonous or paraautochthonous sequences.

7. Tectonic setting of the basement of the Santander Massif and the Merida Andes

7.1 Precambrian

Our data demonstrate the presence of Precambrian protoliths with ages of ~1018.3 Ma to ~1008.6 Ma (08VDL11; Fig. 8) although these ages alone can not be used to determine their tectonic setting. Anhedral orthopyroxene in orthogneiss 08VDL11 may indicate granulite facies conditions for the protolith (Spear et al., 1999). Rocks of similar age in the Garzon Massif, the Sierra Nevada de Santa Marta and the Serranía de San Lucas (Fig. 1) are frequently associated with granulite facies metamorphism, and are interpreted to have formed during the assembly of Rodinia (e.g. Ordoñez-Carmona et al., 2006; Ibanez-Mejia et al., 2011; Cuadros et al., 2014, and references therein). Rocks with crystallisation ages of ~1 Ga are exposed along large tracts of the western margin of South America (Sunsás belt; e.g. Fuck et al., 2008; Ramos, 2010), and have originated within a common tectonic setting

7.2 Early Palaeozoic

Upper amphibolite facies Barrovian metamorphism (Kovisars, 1971; García et al., 2005; Castellanos et al., 2008) affected Cambrian (e.g. 08VDL05) and Precambrian sedimentary and igneous rocks in the Mérida Andes and Santander Massif. Geochronological analyses of migmatites and granitoids with syn-magmatic deformation suggest that peak metamorphic conditions were reached between ~489 and ~472 Ma. An anatectic orthogneiss may date a point on the retrograde path at 454 ± 10 Ma (08VDL11), and isotropic granitoids with ages between ~472 and ~448 Ma crosscut the metamorphic foliation. The amphibolitic rocks are unconformably overlain at ~453 Ma by low grade, felsic metavolcanic rocks of the Mucuchachi Formation near Bailadores (08VDL15), which are interfingered with massive-sulphide mineralized metasedimentary rocks (e.g., Staargaard and Carlson, 2000). Volcanogenic massive sulphide mineralisation commonly forms in extensional settings related to magmatic arcs (e.g. Schandl and Gorton, 2002; Shanks and Thurston, 2012).

These relationships require (i) tectonic burial of Cambrian and Precambrian metasedimentary rocks to 5-8 kbar between ~500 and ~472 Ma, implying that a period of crustal thickening and orogenesis occurred (Fig.15A), followed by (ii) exhumation during extension between ~472 and ~453 Ma (Fig. 15B), ending with basins forming over the amphibolite facies metamorphic rocks in the Bailadores

region and in the Caparo Block (Caparo Fm.). Sufficient heating of the crust to produce granitic melts lasted from ~500 Ma until ~415 Ma.

The crustal blocks that host the Andes have undergone alternating periods of compression and extension since at least the Mesozoic over broad regions of the western margin of South America (Ramos, 2009; Spikings et al., 2014). Orogenic phases along active margins are generally considered to be a result of compression of crust that was previously thinned during trench roll-back (e.g. Dalziel, 1986; Collins, 2002; Collins and Richards, 2008; Kemp et al., 2009), and are caused by enhanced compressive coupling between subducting oceanic crust and the overriding continental crust (e.g. Gutscher, 2002; Ramos, 2009; Spikings and Simpson, 2014). Early Ordovician metamorphism is widespread along the western Gondwanan margin and is documented in the Marañon Complex of Peru (~476 Ma; e.g. Chew et al., 2007; Mišković et al., 2009), the Famatinian Orogeny of Chile and Argentina (477-470 Ma; e.g. Büttner et al., 2005; Larrovere et al., 2011), and may have formed part of an Early Palaeozoic orogen that extended for more than 15000km (Cawood, 2005). Parts of the Early Palaeozoic orogeny of western Gondwana may have been influenced by the docking of terranes (e.g. Cuyania and Paracas terranes; Rapela et al., 1998; Ramos, 2009), although several studies have shown that the extent of the Famatinian orogenic belt is far wider than the size of the Cuyania Terrane. Alternatively, Early Ordovician orogenesis in many parts of western Gondwana may be related to alternating cycles of slab advance and retreat, inducing compression and extension in the upper plate (e.g. Lucassen and Franz, 2005; Miller and Söllner, 2005; Larrovere et al., 2011). There is no evidence to support terrane collision outboard of the basement blocks of the Mérida Andes and the Santander Massif in the Ordovician, and we hypothesize that Barrovian metamorphism and subsequent extension in this region was not driven by the collision of allochthonous blocks.

Post-orogenic extension exhumed the amphibolite facies basement and opened a back-arc or intra arc basin by ~453 Ma (Fig. 15B). This basin is preserved in Venezuela as the lower part of the Mucuchachi Fm. (Fig. 14), which unconformably overlies the Sierra Nevada Suite (e.g. Staargaard and Carlson, 2000). The Caparo Fm. in the Caparo Block may have formed in a similar overall setting (e.g. Compañía Shell de Venezuela and Creole Petroleum Corporation, 1964), although it lacks volcanic rocks. The general paucity of Early Palaeozoic metasedimentary rocks (Fig. 14) suggests that this basin may have been short lived, and basin closure (Fig. 15C) may have been synchronous with the development of an unconformity spanning the interval ~453 Ma- ~445 Ma in the Caparo Block (Caparo Fm. - El Horno Fm; Fig. 14). Magmatism lasted until ~415 Ma although

there is insufficient data to reliably document periods of extension and compression between ~445 Ma and ~415 Ma. Melting of the continental crust between ~499 and ~415 Ma to produce felsic magmatic rocks may have been facilitated by heat transfer from underplated mafic melts (Huppert and Sparks, 1988; Cochrane et al., 2014), and an active continental margin is a plausible setting for this process over a duration of ~85 Ma.

7.3. Permian

The only Late Palaeozoic U-Pb data presently known in Venezuela are four U-Pb ages obtained from rhyolites and granites in the El Baúl Massif (Fig. 1), which range from 294.1 ± 3.1 Ma to 283.3 ± 2.5 Ma (Viscarret et al., 2009), and one age of 272.2 ± 2.6 Ma from the Paraguana Peninsula (this study). The El Baúl Massif (Fig. 1) may have been closer to the continental margin before northeastward movement of the Maracaibo Triangular Block during the Cainozoic (Pindell and Dewey, 1982), and it may have been transported eastwards along the southern Caribbean Plate boundary in the Cainozoic (e.g. van der Lelij et al., 2010). These translations would place the rocks of the El Baúl massif in a similar palinspastic position as the Permian rocks of the Sierra Nevada de Santa Marta and the Paraguana Peninsula. However, there are still insufficient data from the El Baúl massif and its surrounding region to further constrain its geological history.

Permian magmatism along the northwestern margin of Gondwana may have been coeval with sedimentation of the Diamante, Tiburón and Bocas formations in the Santander Massif, and the Sabaneta and Palmarito formations in the Mérida Andes (Fig. 14). Carboniferous to Permian sedimentary successions in the Mérida Andes have been interpreted to form part of a foreland basin (Laya and Tucker, 2012), which may have formed during the final stages of the Alleghanian Orogeny, culminating with the amalgamation of Pangaea by the Early Permian (Hatcher, 2002).

Cardona et al. (2010) suggest that the Permian magmatic rocks in the Sierra Nevada de Santa Marta (Fig. 1) were emplaced in a magmatic arc. The basement of the Paraguana Peninsula (Fig. 1; Diorite 08VDL64) may have been separated from the Sierra Nevada de Santa Marta during Cainozoic dextral displacement between the Caribbean and the South American plates (van der Lelij et al., 2010). Therefore, Permian felsic magmatism in Venezuela may also be related to subduction of oceanic crust beneath western Pangaea (Cardona et al., 2010; Laya and Tucker, 2012).

7.4. Mesozoic

There was sparse felsic magmatism between ~250 and ~225 Ma in the Santander Massif and the Mérida Andes, and the geochemical data are not diagnostic of any particular tectonic setting (Fig. 11). A maximum stratigraphic age of ~202 Ma for the La Quinta formation (Mérida Andes; 08VDL18) and a minimum stratigraphic age of ~250 Ma for the Bocas Fm. (Santander Massif; 10VDL58) suggest that Triassic basins may be absent from both of these regions (Fig. 14). In contrast, Early Triassic basins are documented in the Central Cordillera of Colombia and in Ecuador, which were coeval with numerous mafic intrusions, crustal anatexis and foliated S-type granitoids (e.g., Jaillard et al., 1995; Villagómez et al., 2011; Cochrane et al., 2014; Spikings et al., 2014). These formed during extension and rifting along the Pacific margin during the disassembly of Pangaea between 240 Ma and 216 Ma (Cochrane et al., 2014; Spikings et al., 2014). We propose that magmatism between ~250 and ~196 Ma in the Mérida Andes and the Santander Massif formed in a continental arc or in a back arc. Early Triassic back-arc extension along the western margin of Pangaea did not affect the rocks of the Mérida Andes and the Santander Massif, perhaps because they were too distant from the locus of extension. The magmatic rocks with ages spanning the interval ~250 – 225 Ma from the Serranía del Perijá and Toas Island have similar geochemistry to the latest Triassic to Earliest Jurassic granitoids in the Santander Massif, and we propose that they may also be related to subduction of oceanic crust of the Pacific Ocean.

Magmatism at ~213 - ~211 Ma produced one large peraluminous pluton in the Mérida Andes (Fig. 2), which was followed by the intrusion of pegmatite dikes and migmatites at ~209 Ma in the Santander Massif (Fig. 14). Subsequently, the locus of magmatism migrated westward to produce voluminous shallow batholiths in the Santander Massif. Extension after ~202 Ma is documented in the Mérida Andes and the Santander Massif by the syn-rift La Quinta, Jordán and Girón Formations (Fig. 14; Maze, 1984; Kammer and Sánchez, 2006).

Magmatism and extension between ~213 and ~196 Ma in the Northern Andes was coeval with graben development and red bed deposition in many parts of the Andean range (e.g. Ramos, 2009). The oldest reliable date for Jurassic arc magmatism in the Central Cordillera of Colombia (Fig. 1) is ~189 Ma (zircon U-Pb; Bustamante et al., 2010), and we propose that magmatism spanning the interval ~213 - ~196 Ma in the Mérida Andes and the Santander Massif was related to a continental active margin. The arc axis may have migrated westward between ~205 Ma and ~189 Ma during extension, as a result of slab roll-back (Spikings et al., 2014).

The timing of rifting which led to complete separation of North and South America and the formation of the oceanic crust in the Caribbean Sea may not be constrained by our data, although previous work suggests that it may have occurred in several stages from the Triassic to Middle Jurassic. (e.g. Bartok, 1993; Beutel, 2009). Although the sedimentary features and spatial distribution of the La Quinta Fm. are consistent with formation in grabens (Maze, 1984; Gonzales de Juana et al., 1980; Bartok, 1993; Odremán and Useche, 1997), these may record several reactivation episodes throughout the Jurassic (Maze, 1984). Rhyodacite 08VDL18 (~202 Ma; Table 1) located at the base of the La Quinta Fm. provides evidence for the earliest Mesozoic rifting in the Mérida Andes. This is indistinguishable from the timing of the Central Atlantic Magmatic Province at ~201 Ma (e.g. Marzoli et al., 1999), which formed during the initial stages of rifting of the Central Atlantic Ocean. However, the Central Atlantic Magmatic Province mainly consists of tholeiitic basalts, and no felsic calc alkaline rocks have been identified (e.g. Menzies et al., 2002 and references therein). Consequently, we consider the La Quinta rhyodacite to be related to an active margin in an extensional regime rather than a western remnant of the Central Atlantic Magmatic Province.

8. Tectonic model

Early Palaeozoic rocks in the Northern Andes record continental arc magmatism between ~499 and ~415 Ma, orogenesis between ~500 and ~473 Ma (Fig. 15a, Fig. 16), followed by intra-arc or back-arc extension and basin formation at ~453 Ma (Fig. 15b). The arc rocks are coeval with the Famatinian arc (e.g. Bahlburg et al., 2009 and references therein) and arc-related intrusions in the Eastern Cordillera of Peru, and probably represent the northern extent of the same arc-subduction system (Fig. 16).

Existing plate tectonic models for northwestern Gondwana show the rift to drift transition which opened the Rheic Ocean occurred between ~505 and ~480 Ma, and this separated several peri-Gondwanan terranes from Gondwana (Avalonia, Carolinia and Ganderia; e.g. Murphy et al., 2004; Linnemann et al., 2008; Nance et al., 2010; van Staal et al., 2012 and references therein). However, autochthonous, Early Palaeozoic rocks of the Northern Andes record orogenesis between ~505 and ~472 Ma, which is not consistent with a rift to passive margin transition. Therefore, it appears unlikely that the rifted margin of the Rheic Ocean was outboard of the Mérida Andes and the Santander Massif between ~500 and ~472 Ma (Fig. 16a). The Palaeozoic Espino Graben (Feo-Codecido et al., 1984) lies to the East of the Mérida Andes (Fig. 1) where Precambrian and Palaeozoic rocks are truncated by the Caribbean Plate and the Atlantic Ocean. We interpret this

graben to represent the westernmost extension of the rift which resulted in the opening of the Rheic Ocean between ~500 and ~480 Ma (Fig. 16a).

Several crustal fragments may have been adjacent to northwestern Gondwana during the Palaeozoic, including rocks presently found in Guatemala (Maya Block; e.g. Ratschbacher et al., 2009; Estrada-Carmona et al., 2012; Solari et al., 2013 and references therein), Belize (Maya Block; e.g. Martens et al., 2010) and Mexico (Acatlán Complex; e.g. Keppie, 2004; Keppie et al., 2008; Vega-Granillo et al., 2007, 2009). No consensus exists on the Early Palaeozoic tectonic setting of any of these terranes, which are considered to record either (i) Ordovician to Silurian rifting and passive margin development on the southern margin of the Rheic Ocean (e.g. Estrada-Carmona et al., 2012; Keppie et al., 2008, 2012b), or (ii) Ordovician orogenesis and Ordovician to Silurian continental arc magmatism (e.g. Ratschbacher et al., 2009; Vega-Granillo et al., 2009; Solari et al., 2013). These strongly contrasting interpretations have been used to suggest that either the Rheic Ocean opened along the western margin of Gondwana, with its southern extent in front of present day Peru (e.g. Pollock et al., 2009; van Staal et al., 2012), or opened >2700 km further along-strike to the northeast, possibly outboard of Eastern Venezuela (Fig. 16a; this study). This controversy is a problem for any Ordovician tectonic reconstruction of Gondwana and Avalonia, and consequently for Silurian tectonic reconstructions of Baltica and eastern Laurentia.

The new evidence presented in this paper contrasts with some interpretations of the geology of southern Mexico (Acatlán Complex; e.g., Keppie et al., 2008), which may have been a part of northwestern Gondwana from ~1 Ga until its separation at ~220 Ma (e.g. Noble et al., 1997; Keppie and Ortega-Gutierrez, 1999; Ruiz et al., 1999; Keppie, 2004; Cochrane et al., 2014; Spikings et al., 2014). The Acatlán Complex hosts continental bimodal magmatic rocks spanning the interval ~470 - ~440 Ma (e.g. Keppie et al., 2012b and references therein), which are interpreted to record a rift to passive margin transition on the southern margin of the Rheic Ocean between ~500 and ~440 Ma, (Keppie et al., 2008). However, the hypothesis of passive margin development during this period is challenged by evidence of high pressure, low temperature metamorphism in the Acatlán Complex at ~478 Ma, which is diagnostic of active margins but not of passive margin development (e.g. Vega-Granillo et al., 2009 and references therein). Furthermore, although the bimodal magmatic assemblage suggests extension, this may be related to a continental back-arc environment (e.g. Shinjo and Kato, 2000; Nivia et al., 2006; Helbig et al., 2012b). Finally, crystallisation of MORB-like tholeiitic rocks in the Acatlán Complex in the interval ~470 – ~440 Ma (e.g. Keppie et al.,

2012b and references therein) post-dates the drift stage of the Rheic Ocean (505-480 Ma; e.g. Linnemann et al., 2008; Nance et al., 2010; van Staal et al., 2012) by at least ~10 – 40 Ma.

We hypothesize that high pressure metamorphism at ~478 Ma documented in the Acatlán Complex (e.g. Vega-Granillo et al., 2009) and bimodal magmatism between ~470±10 Ma and 442±1 Ma (e.g. Keppie et al., 2012b and references therein) may be related to (i) compression at ~478 Ma followed by (ii) periods of extension between ~470 and ~440 Ma in an active margin along western Gondwana. This hypothesis is consistent with the calc-alkaline geochemistry and arc signature of granitoids in the Acatlán Complex which intruded between ~488 Ma and ~442 Ma (e.g. Miller et al., 2007; Keppie et al., 2010, 2012a) and their association with tholeiitic basalts which formed between ~470 Ma and ~440 Ma (Keppie et al., 2012b).

The Acatlán Complex and other fragments of Central America were separated from Gondwana during the disassembly of Pangaea (e.g. Keppie and Ortega-Gutierrez, 1999) during the Triassic (e.g. Cochrane et al., 2014; Spikings et al., 2014). Recent suggestions for its position prior to Pangaea breakup include a position outboard of the Mérida Andes and Santander Massif (see Helbig et al., 2012a). Arc magmatism occurred in the intervals ~500 - ~415 Ma and ~250 - ~195 Ma in autochthonous rocks of the Mérida Andes and the Santander Massif, and the presence of large portions of the basement of southern Mexico outboard of the pre-Cainozoic basement of Colombia and Venezuela (Fig. 1; e.g. Keppie, 2004) implies an arc to (Pacific) trench distance in excess of 1000 km (Fig. 17). We suggest that the Acatlán Complex and other Central American terranes may have been along-strike either southward (Fig. 17) or northwestward from the basement rocks of the Santander Massif and the Mérida Andes during the Palaeozoic (e.g. Helbig et al., 2012a; Keppie and Keppie, 2012). A potential position for the Acatlán Complex prior to its separation from Gondwana may be outboard of the Triassic magmatic belt in Ecuador and southern-central Colombia (Fig. 16a-e, Fig. 17), where (i) Early Palaeozoic igneous or metamorphic basement is not reported, (ii) the Precambrian basement may be much closer to the pre-Cainozoic margin of South America than in Colombia or Venezuela (Fig. 1), and (iii) Triassic basement rocks record dextral shearing during removal of a conjugate margin at ~240-216 Ma (e.g. Noble et al., 1997, Cochrane et al., 2014; Spikings et al., 2014). This hypothesis is supported by data suggesting that the Acatlán Complex was on the western margin of Pangaea at ~250 Ma, prior to Pangaea breakup (e.g. Helbig et al., 2012a, b; Kirsch et al., 2012).

There may have been a lack of magmatism in northwest South America between ~415 Ma and ~300 Ma, as suggested by our data (Table 1) and by detrital zircon U-Pb data (Horton et al., 2010) within the study areas. Similarly, Devonian igneous and high grade metamorphic rocks are not documented over an along-strike distance of > 5000 km on the western margin of Gondwana (Bahlburg et al., 2009, Reitsma, 2012), with the exception of the Achala Batholith in Argentina (379 Ma - 369 Ma; e.g. Dahlquist et al., 2012). The low volume of Devonian magmatism has led to suggestions that the margin became passive at the end of the Silurian (Fig. 16C; e.g. Bahlburg et al., 2009). It is not clear how a convergent margin could become passive along most of the western margin of Gondwana. Subduction termination by large scale terrane accretion or removal of the arc by faulting are not documented. Fauna and flora in autochthonous Devonian rocks in the Santander Massif, the Sierra de Perijá and Bolivia show affinities to both Gondwana and Laurentia (e.g. Berry et al., 2000; Janvier and Villarreal, 2000; Troth et al., 2011), which supports proximity between Laurentia and Gondwana in the Devonian (Fig. 16d; e.g. McKerrow et al., 2000; Torsvik and Cocks, 2004). However, there is no evidence for deformation resulting from collision between Gondwana and Laurentia in the Devonian.

Known Devonian tectonic rearrangements which affected Gondwana and Laurentia include, (i) closure of the northern Iapetus Ocean, resulting in the formation of Laurussia by ~430-420 Ma (Fig. 16c; e.g. Cocks and Torsvik, 2002), followed by (ii) the onset of subduction of Rheic Ocean crust beneath the Appalachian margin at ~410 Ma (Fig. 16d; e.g. Nance et al., 2010), and (iii) changes in directions of rotation of Gondwana from clockwise at ~420 Ma to counter clockwise at ~400 Ma (Torsvik et al., 2012). We hypothesize that these changes resulted in either highly oblique, amagmatic subduction, or cessation of subduction beneath the western margin of South America in the interval ~415 - ~350 Ma.

Late Palaeozoic arc magmatism started at ~335 Ma in the Eastern Cordillera of Peru (Mišković et al., 2009), at ~293 Ma in the Sierra Nevada de Santa Marta (Fig. 16e; Cardona et al., 2010), and perhaps in the El Baúl Massif at ~294 Ma (Viscarret et al., 2009). Minor arc magmatism is recorded between ~250 and 225 Ma in Toas Island, the Perijá Andes (Fig. 1), the Santander Massif and in the Mérida Andes, where it was followed by more voluminous intrusions at ~213 Ma. Slab roll back occurred during arc magmatism between ~202 Ma and ~196 Ma (Fig. 17), was coeval with the formation of back-arc basins, and was accompanied by the migration of the arc axis to the west into the Central Cordillera of Colombia at ~189 Ma (Spikings et al., 2014).

9. Conclusions

- 1) There is no evidence for the existence of allochthonous terranes in the Merida Andes or the Santander Massif, and we suggest that the Merida Terrane and the Chibcha Terrane, as they were originally defined, do not exist.
- 2) The Santander Massif and the Mérida Andes host the remnants of an Early Palaeozoic active margin which lasted from $> \sim 500$ Ma to $< \sim 415$ Ma
- 3) The onset of active margin magmatism was followed shortly by Barrovian metamorphism related to compressional orogenesis associated with trench advance. Migmatites and syn-deformational granitoids record peak to retrograde conditions between ~ 477 and ~ 454 Ma.
- 4) Trench roll back resulted in the formation of a back-arc or intra-arc basin over the metamorphic basement at ~ 453 Ma. The duration of extension is uncertain but the highly discontinuous sedimentary record suggests that the depocenters closed rapidly, or are not preserved.
- 5) Cessation of arc magmatism at ~ 415 Ma along most of the western Gondwanan margin may have resulted from amagmatic subduction or cessation of subduction during changes in rotation direction of Gondwana (Fig. 16).
- 6) Peri-Gondwanan terranes involved in the Caledonian and Appalachian orogenies (e.g. Ganderia, Carolinia and Avalonia, Fig. 16a) may have originated from the region to the east of the Palaeozoic orogenic belt exposed in the Mérida Andes, rather than originating outboard of this region. These relationships constrain the Early Palaeozoic western limit of the Rheic Ocean, and the extent of preserved autochthonous Early Palaeozoic continental arc rocks within Gondwana.
- 7) There are no reliable published constraints on late Palaeozoic deformations in the Merida Andes or the Santander Massif during the amalgamation of Pangaea, perhaps because these regions were too far from the collision zone (Fig. 17). However, it has recently been proposed that Permian sedimentary sequences in the Merida Andes were deposited in a foreland basin (e.g. Laya and Tucker, 2012).
- 8) Late Palaeozoic to Early Jurassic arc magmatism resulted from subduction of the Pacific Ocean beneath western Pangaea (Fig. 17). Voluminous igneous activity between ~ 205 Ma and ~ 196 Ma in an extensional regime marks the beginning of the Andean subduction cycle in the Northern Andes.

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Supplementary items:

Figure 1: U-Pb Concordia diagrams for rocks from Venezuela, Figure 2: Concordia diagrams for rocks from the Santander Massif, Table 3: geochemical data, Table 4: U-Pb data and Table 5: detrital zircon data available from the online version of the journal at: [yyyyyyyyy](#)

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Figure Captions

Figure 1. Digital relief map (Amante and Eakins, 2008) showing the tectonic provinces of northern South America modified from Feo-Codecido et al. (1984) and Ostos et al. (2005). The locations of geological maps in Figures 2 and 4 are highlighted in dashed red boxes. BB: Bonaire Block, CC: Central Cordillera, CM: Caribbean Mountains, EB: El Baúl Massif, EC: Eastern Cordillera, FM: Floresta Massif, GM: Garzón Massif, LB: Llanos Basin, MA: Mérida Andes, MTB: Maracaibo Triangular Block, PP: Paraguana Peninsular, QM: Quetame Massif, SL: Serranía de San Lucas, SMBF: Santa Marta-Bucaramanga Fault, SM: Santander Massif, SNSM: Sierra Nevada de Santa Marta, SP: Sierra del Perijá, TI: Toas Island.

Figure 2: Geological map of the Mérida Andes, Venezuela, simplified and modified from Hackley et al. (2005). Locations of all analysed samples are shown. Sample codes in parentheses (XX) correspond to full sample codes 08VDLXX. All quoted ages are new weighted mean zircon $^{206}\text{Pb}/^{238}\text{U}$ dates obtained during this study with uncertainties at the 2σ confidence level. Location of the main faults and the Caparo Block are shown.

Figure 3: Compilation of previously obtained geochronological data from the Mérida Andes (Bass and Shagam, 1959; Compañía Shell de Venezuela and Creole Petroleum Corporation, 1964; Martin-Bellizzia et al., 1968; Schubert, 1969; Burkley, 1976; Kovach et al., 1977; Schubert et al., 1979; Etchart and Cordani, 1981; Cordani et al., 1985). Colour coding is according to analytical method utilised, and box size is a function of the quoted uncertainty (confidence levels are unspecified). Initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios used for Rb/Sr model ages = 0.705.

Figure 4: Geological map of the Santander Massif, Colombia, simplified and modified from Gómez et al. (2007). Sample codes in parentheses (XX) correspond to full sample codes 10VDLXX. All quoted ages are new weighted mean zircon $^{206}\text{Pb}/^{238}\text{U}$ dates obtained during this study with uncertainties at the 2σ confidence level. The Bocas Fm. (Permian?) is included with Jurassic sedimentary rocks as available base maps do not differentiate them.

Figure 5: Compilation of previously obtained geochronological data from the Santander Massif (Goldsmith et al., 1971; Boinet et al., 1985; Cordani et al., 2005; Dörr et al., 1995; Restrepo-Pace, 1995, 1997; Restrepo-Pace and Cediél, 2010; Mantilla et al., 2009, 2012, 2013). $^{40}\text{Ar}/^{39}\text{Ar}$ dates include integrated dates from discordant spectra. Initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios used for Rb/Sr model ages = 0.705. Note the change in scale of the time axis at 300 Ma.

Figure 6: Probability density plot for U-Pb dates from detrital zircon sample 08VDL05. The dates used are $^{206}\text{Pb}/^{238}\text{U}$ dates for spots younger than 0.9 Ga and $^{207}\text{Pb}/^{206}\text{Pb}$ dates for spots older than 0.9 Ga.

Figure 7: Photomicrographs of microtextures discussed in the text. Bt: biotite, chl: chlorite, fsp: feldspar, hbl: hornblende, ms: muscovite, myr: myrmekite, pl: plagioclase, sil: sillimanite, sph: sphene. A) Crossed polarized light photomicrograph of sample 08VDL11 (Micarache anatectic orthogneiss, Mérida Andes) showing randomly oriented muscovite replacing sillimanite. In the top left, muscovite forms symplectites with feldspar and quartz. B) Another photomicrograph of 08VDL11 showing the symplectite of muscovite, biotite and myrmekite. C) Crossed polarised light photomicrograph of Mitisus striped gneiss sample 08VDL26, Mérida Andes, showing typical elongated quartz crystals around feldspar augen with biotite growing in strain shadows. D) Plane polarized light photomicrograph of the Bucaramanga gneiss sample 10VDL23, Santander Massif, showing the poikiloblastic structure and its relationship to chlorite and hornblende.

Figure 8: Concordia plot of data obtained from the Micarache orthogneiss, 08VDL11. Error ellipses are $\pm 2\sigma$. Clustered blue ellipses are interpreted to reflect the timing of protolith crystallisation and anatexis, and

orange ellipses are considered to be mixed domains between the two populations. Quoted ages are weighted mean $^{206}\text{Pb}/^{238}\text{U}$ ages. N: number of individual zircon ages included in weighted mean calculation. MSWD: Mean Squared Weighted Deviates.

Figure 9: Photograph showing crosscutting relationships between different rock units in the Santander Massif. Orthogneiss (A) is probably related to Ordovician sample 10VDL37, and forms schollen in the migmatite structure (B). The fabric is crosscut by an undeformed granitic dike (C), which may relate to Early Jurassic granitoid 10VDL05 (Corcova granitodiorite, Table 1). The migmatites in this picture (10VDL39) yield a weighted mean zircon $^{206}\text{Pb}/^{238}\text{U}$ age of 208.8 ± 1.2 Ma. The hammer in the picture is for scale.

Figure 10: Concordia plots of zircon U-Pb data obtained from the Rio Surata granodiorite (10VDL61). Error ellipses are $\pm 2\sigma$. Blue ellipses were used to calculate weighted mean $^{206}\text{Pb}/^{238}\text{U}$ ages, and the orange ellipse was discarded from calculations as it may reflect Pb loss. The older population is considered to be derived from one of the numerous enclaves in this rock. N: number of individual zircon ages included in weighted mean calculation. MSWD: Mean Squared Weighted Deviates.

Figure 11: Geochemical data plots for major and trace element data obtained during this study. Colour coding of symbols is identical in all graphs, and reflects sample region and age as shown in the legend. (A) K_2O vs SiO_2 plot with fields from Peccerillo and Taylor (1976). (B) Shand's index plot with fields from Maniar and Piccoli (1989). (C) Na_2O vs K_2O I-Type / S-Type discriminant after Chappell and White (1974). (D) Tectonic discrimination plot for granitoids with fields from Pearce (1996); ORG: ocean ridge granite, PC: post-collisional granite, syn-COLG: syn-collisional granite, VAG: volcanic arc granite, WPG: within plate granite. Fields for data from Permian to Triassic (orange outlines) and Jurassic (blue outlines) granitoids from Colombia and Ecuador are compiled from Litherland et al. (1994), Vinasco et al. (2006) and Cardona et al. (2010, 2014).

Figure 12: Trace element and rare earth element plots for unaltered samples from the Merida Andes and the Santander Massif. All plots are normalized to upper continental crust (Taylor and McLennan, 1985).

Figure 13: Isotope correlation diagram displaying present-day uranogenic Pb isotopic compositions. The field for age-corrected initial ratios is shaded in grey. See Figure 11 for colour coding of samples. Pb isotopic fields from major basement units in Colombia (Ruiz et al., 1999) and western Gondwana and Laurentia (Tosdal, 1996 and references therein) are shown for comparison. The thick black curve shows the terrestrial evolution trend (Stacey and Kramers, 1975), and numbers adjacent to tick marks represent Ga before present.

Figure 14 : Synoptic stratigraphic columns for the main basement areas studied in this paper, based on our new data. Pink boxes outline the main periods of magmatic activity. Dated magmatic members of sedimentary formations are also highlighted in pink. Question marks denote an uncertain extent. The only age constraint on the Cerro Azul Suite is that it is Silurian or older (see text for details), and the Bella Vista Suite and Silgará Formation predate the end of the Cambrian. Zigzag lines indicate unconformities discussed in the text. The Pre-Devonian stratigraphy of the Sierra del Perijá and the field relationships of Toas Island and Paraguaná Peninsula are essentially unknown and have not been included here.

Figure 15: Schematic cross-sections for the Early Palaeozoic evolution of the northern Andes based on this study, illustrating the changing geodynamic regime. See text for discussion. AF: Apure fault; CB: Caparo Block, MA: Mérida Andes; SM: Santander Massif. Cardinal directions are in present day coordinates. The Apure Fault is the apparent limit of Early Palaeozoic deformation in Venezuela (see also Feo-Codecido et al. (1984) and its location is marked on Fig. 1.

Figure 16: Proposed plate reconstruction and palaeogeography for the Palaeozoic margin of Gondwana, modified and simplified from Van Staal and Hatcher (2010) and Cocks and Torsvik (2006) with focus on the Gondwanan margin (Famatinian Arc). The region studied in this paper is highlighted with a red box. The structural features of Laurentia, Baltica, Siberia, and peri-Gondwanan terranes are beyond the scope of this paper and are simplified or omitted. The peri-Gondwanan position of the Carolina Terrane may have been similar to Ganderia (van Staal and Hatcher, 2010) and is not shown. Rotation directions for Gondwana at 420

Ma and 400 Ma (Torsvik et al., 2012) are indicated by black arrows. AC: our preferred position for the Acatlán Complex. For contrasting views see Keppie et al. (2012) and Vega-Granillo et al. (2009), and references therein. CT: Cuyania Terrane (e.g. Astini et al., 1995).

Figure 17: Proposed reconstruction for the evolution of the continental margin between the end of the Permian (a) and the Early Jurassic (b), modified after Weber et al., 2007; Cochrane et al., 2014 and Spikings et al., 2014. Pink shaded areas highlight the approximate extent of active margin magmatism during each period. The Ouachita-Marathon suture is the main collision zone between Gondwana and Laurentia (e.g. Ross, 1986). The Acatlán Complex is included in the Central American block.

Table 1: Summary of geochronological data collected during this study. Full sample codes and GPS Coordinates are shown. Weighted mean zircon $^{206}\text{Pb}/^{238}\text{U}$ ages were calculated using concordant analyses only. Th/U ratios are mean ratios of analyses included in the weighted mean age calculation. The coordinates of sample 08VDL60 are estimated from map coordinates as a result of a GPS malfunction during sampling. See supplementary files for full dataset.

Table 2: Whole rock common Pb isotopic compositions of analysed igneous and metamorphic rocks from the Mérida Andes and the Santander Massif. Typical analytical uncertainties are 0.0075% on all measured ratios. Initial compositions were calculated using the measured zircon U-Pb age of the rock.

Figure 1

Early Palaeozoic Mérida Andes

Early Palaeozoic Santander Massif

Permian - Middle Triassic

Late Triassic - Jurassic

A) ~500-472 Ma

B) ~472-452 Ma

C) ~450-440 Ma

a: 490 Ma

b: 450 Ma

c: 420 Ma

d: 400 Ma

e: 280 Ma

Table 1 - GPS coordinates and summary of geochronological data

Sample Data			GPS Coordinates			²⁰⁶ Pb/ ²³⁸ U (zircon) data			
Sample	Location	Unit name	Latitude N	Longitude W	Alt (m)	Age ±2σ (Ma)	MSWD	spots	Th/U
08VDL03	Mérida	Verdalito granodiorite	8°28'49"	71°34'13"	363	449.3 ± 2.5	2.0	16	0.34
08VDL04	Mérida	Pueblo Hondo granite	8°17'53"	71°53'58"	2010	434.2 ± 4.4	1.0	12	1.02
08VDL05	Mérida	Sierra Nevada paragneiss	8°29'59"	71°12'04"	2247	<505 Ma; detrital; see figure 6			
08VDL11	Mérida	Micarache Orthogneiss	8°41'41"	70°53'31"	3500	454.0 ± 10.0	2.5	3	0.97
		08VDL11, protolith				1008.6 ± 6.7	0.4	4	0.34
08VDL14	Mérida	El Carmen granodiorite	8°45'07"	70°57'33"	3195	211.2 ± 1.2	2.3	14	0.40
08VDL15	Mérida	Bailadores alkali rhyolite	8°11'51"	71°50'09"	2058	452.6 ± 2.7	2.7	12	0.56
08VDL18	Mérida	La Quinta rhyodacite	8°08'56"	72°00'55"	1138	202.0 ± 1.6	2.0	14	0.78
08VDL21	Mérida	La Grita Diorite	8°08'28"	71°56'15"	2240	471.8 ± 4.0	2.0	13	0.91
08VDL22	Mérida	La Playita orthogneiss	8°11'38"	71°53'02"	2489	449.7 ± 4.4	9.3	12	0.48
08VDL23	Mérida	Mucubaji granite	8°48'36"	70°48'47"	3465	237.1 ± 1.4	3.8	12	0.12
08VDL26	Mérida	Mitisus striped gneiss	8°52'31"	70°37'48"	1669	488.2 ± 4.3	2.0	6	0.36
08VDL27	Mérida	La Raya granitic gneiss	8°51'34"	70°34'51"	1427	488.5 ± 6.5	2.0	11	0.38
08VDL32	Mérida	La Soledad granodiorite	8°49'48"	70°31'47"	961	414.5 ± 3.9	5.6	14	0.36
08VDL35	Caparo	Cerro Azul granodiorite	8°53'14"	70°23'34"	848	499.4 ± 2.7	1.7	12	0.81
08VDL36	Mérida	Puente Real augen gneiss	8°27'53"	71°26'50"	572	472.3 ± 3.4	4.1	14	0.59
08VDL38	Mérida	Macanas alkali granite	8°20'19"	71°37'21"	1497	454.8 ± 4.2	1.5	11	0.60
08VDL40	Mérida	La Culata granodiorite	8°48'27"	71°00'04"	4566	213.2 ± 1.7	3.0	15	0.32
08VDL41	Mérida	Chachopo granite	8°55'55"	70°47'31"	3119	454.5 ± 3.1	1.0	12	0.29
08VDL44	Mérida	Miraflores granodiorite	9°10'42"	70°38'08"	867	466.8 ± 2.5	1.4	13	0.14
08VDL45	Mérida	Valera granodiorite	9°11'11"	70°37'51"	838	447.9 ± 1.8	1.7	14	0.14
08VDL46	Mérida	La Quebrada granite	9°10'25"	70°32'24"	2358	243.5 ± 3.4	2.9	5	0.70
08VDL47	Mérida	Carmania monzodiorite	9°15'59"	70°38'18"	789	450.4 ± 2.5	1.5	12	1.12
08VDL49	Mérida	La Puerta granodiorite	9°07'11"	70°42'39"	1920	453.1 ± 1.8	2.1	14	0.16
08VDL50	Mérida	Timotes granite	9°00'52"	70°44'25"	1931	445.2 ± 3.9	8.2	15	0.10
08VDL55	Caparo	Islitas tonalite	7°49'01"	71°21'01"	340	473.1 ± 3.5	2.0	16	0.60
08VDL56	Caparo	El Cambur granite	7°52'29"	71°16'03"	303	492.3 ± 8.4	2.0	8	0.71
08VDL57	Caparo	El Tapo granite	7°51'00"	71°13'41"	259	492.3 ± 5.3	4.8	8	0.52
08VDL59	Toas Isl.	Toas Island granodiorite	10°57'23"	71°37'45"	1	248.9 ± 1.7	1.1	17	0.66
08VDL60	Perijá	Riecito del Palmar granite	10°37'56"	72°21'26"	175	225.1 ± 1.5	1.1	16	0.90
08VDL64	Paraguaná	El Amparo diorite	11°55'54"	69°57'36"	156	272.2 ± 2.6	2.3	13	0.61
10VDL05	Santander	Corcova granodiorite	7°06'03"	73°00'36"	2793	198.3 ± 1.8	2.2	7	0.77
10VDL22	Santander	Pescadero granodiorite	6°49'48"	72°59'27"	636	199.1 ± 1.3	2.5	15	0.88
10VDL23	Santander	Bucaramanga gneiss	6°56'35"	72°58'00"	2055	461.0 ± 2.1	2.6	18	0.49
10VDL28	Santander	Onzaga granodiorite	6°22'31"	72°49'06"	1876	200.4 ± 0.7	1.3	16	0.87
10VDL31	Santander	Onzaga granodiorite	6°24'28"	72°49'08"	1800	201.0 ± 0.9	2.9	17	0.79
10VDL32	Santander	Mogotes granodiorite	6°25'22"	72°49'29"	1782	198.0 ± 0.8	1.3	17	0.81
10VDL35	Santander	Rio Surata Diorite	7°10'22"	73°05'08"	840	201.1 ± 1.4	1.9	12	1.17
10VDL37	Santander	Berlin orthogneiss	7°11'24"	72°58'41"	1712	451.5 ± 1.3	1.0	13	0.38
10VDL39	Santander	Pegmatite from orthogneiss	7°10'46"	72°59'48"	1513	208.8 ± 1.2	2.2	12	0.03
10VDL43	Santander	Migmatite from orthogneiss	7°14'18"	72°49'30"	3379	209.2 ± 3.4	6.0	5	0.02
		10VDL43, Host rock				460.3 ± 3.5	1.8	6	0.53
10VDL44	Santander	Berlin orthogneiss	7°20'40"	72°42'59"	3023	473.5 ± 2.5	4.6	14	0.29
10VDL46	Santander	Pamplona granite	7°16'29"	72°39'55"	2458	439.2 ± 4.7	8.0	9	0.46
10VDL47	Santander	gabbrodiorite	7°28'23"	72°41'44"	2615	483.7 ± 5.9	0.7	11	0.75
10VDL49	Santander	Silgará Fm. Gneiss	7°29'20"	72°42'16"	2360	479.8 ± 3.1	1.7	17	0.20
10VDL50	Santander	Migmatite from paragneiss	7°15'22"	72°53'45"	3710	477.0 ± 5.3	7.5	9	0.82
10VDL51	Santander	Berlin orthogneiss	7°15'17"	72°53'48"	3707	472.5 ± 3.4	11.5	17	0.30
10VDL52	Santander	Paramo Rico tonalite	7°13'54"	72°53'54"	3731	199.8 ± 1.2	1.2	18	0.97
10VDL54	Santander	Ocana alkali granite	8°09'45"	73°17'59"	1268	195.8 ± 1.5	1.2	14	0.80
10VDL55	Santander	Tarra granodiorite	8°03'10"	73°04'31"	1506	443.4 ± 3.2	3.8	12	0.45
10VDL56	Santander	Aguablanca granite	8°07'08"	72°56'51"	1636	202.2 ± 1.0	2.3	16	0.79
10VDL58	Santander	Bocas Fm. Rhyolite	7°26'27"	73°13'24"	569	250.7 ± 4.3	1.9	6	0.89
10VDL59	Santander	Rio Negro tonalite	7°17'13"	73°08'46"	710	196.0 ± 1.1	1.5	14	1.29
10VDL61	Santander	Rio Surata granodiorite	7°09'59"	73°05'17"	827	200.0 ± 1.5	1.2	10	1.13
		10VDL61, enclaves				1018.3 ± 8.9	2.7	6	0.31

Table 2. Pb isotopic data for basement rocks in the Mérida Andes, western Venezuela and the Santander Massif

Sample Data			Concentration			Present day isotopic ratios			Initial isotopic ratios			
Sample	Location	Age $\pm 2\sigma$ (Ma)	Pb (ppm)	Th (ppm)	U (ppm)	$^{206}\text{Pb}/^{204}\text{Pb}$	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{204}\text{Pb}$	$^{206}\text{Pb}/^{204}\text{Pb}_i$	$^{207}\text{Pb}/^{204}\text{Pb}_i$	$^{208}\text{Pb}/^{204}\text{Pb}_i$	
08VDL03	Mérida	449.3 $\pm$ 2.5	14.48	13.21	3.86	19.574	15.729	39.413	18.314	15.659	38.026	
08VDL04	Mérida	434.2 $\pm$ 4.4	16.91	5.61	0.59	18.628	15.645	38.566	18.475	15.636	38.091	
08VDL11	Mérida	454.0 $\pm$ 10.0	23.72	8.34	1.62	19.152	15.706	39.215	18.402	15.652	38.009	
08VDL14	Mérida	211.2 $\pm$ 1.2	17.48	10.26	3.17	19.021	15.691	39.104	18.630	15.671	38.691	
08VDL15	Mérida	452.6 $\pm$ 2.7	8.32	19.34	4.11	20.040	15.732	40.569	17.636	15.597	36.930	
08VDL21	Mérida	471.8 $\pm$ 4.0	16.91	11.58	2.32	18.737	15.643	38.793	18.069	15.605	37.722	
08VDL22	Mérida	449.7 $\pm$ 4.4	17.57	13.57	2.26	19.161	15.716	39.041	18.559	15.683	37.878	
08VDL27	Mérida	488.5 $\pm$ 6.5	24.44	23.43	9.57	20.944	15.854	39.712	18.877	15.736	38.088	
08VDL32	Mérida	414.5 $\pm$ 3.9	24.46	19.93	9.76	19.505	15.707	39.116	17.781	15.612	37.981	
08VDL38	Mérida	454.8 $\pm$ 4.2	17.58	18.28	3.94	19.221	15.702	39.074	18.159	15.642	37.488	
08VDL40	Mérida	213.2 $\pm$ 1.7	15.30	8.09	1.54	18.938	15.671	39.005	18.720	15.660	38.631	
08VDL41	Mérida	454.5 $\pm$ 3.1	17.87	11.41	8.25	20.322	15.777	39.024	18.103	15.653	38.037	
08VDL44	Mérida	466.8 $\pm$ 2.5	15.60	15.26	3.67	19.545	15.702	39.704	18.387	15.636	38.154	
08VDL45	Mérida	447.9 $\pm$ 1.8	25.23	13.46	2.71	18.803	15.683	38.902	18.306	15.655	38.108	
08VDL47	Mérida	450.4 $\pm$ 2.5	14.02	15.03	2.73	19.235	15.686	39.376	18.319	15.635	37.753	
08VDL49	Mérida	453.1 $\pm$ 1.8	17.96	10.86	3.17	19.851	15.732	38.789	19.014	15.685	37.867	
08VDL50	Mérida	445.2 $\pm$ 3.9	22.92	9.30	6.97	19.746	15.709	38.673	18.335	15.630	38.067	
08VDL35	Caparo	499.4 $\pm$ 2.7	115.74	7.01	1.75	18.595	15.712	38.736	18.517	15.708	38.636	
08VDL55	Caparo	473.1 $\pm$ 3.5	11.07	8.57	2.46	19.628	15.673	39.057	18.527	15.611	37.824	
08VDL56	Caparo	492.3 $\pm$ 8.4	59.63	13.47	1.90	18.797	15.705	38.932	18.634	15.695	38.562	
08VDL57	Caparo	492.3 $\pm$ 5.3	14.93	18.36	4.72	19.779	15.763	39.985	18.123	15.669	37.916	
08VDL59	Toas Island	248.9 $\pm$ 1.7	42.84	6.85	1.23	18.450	15.611	38.013	18.378	15.607	37.884	
08VDL60	Perijá	225.1 $\pm$ 1.5	6.67	1.81	0.68	18.408	15.604	37.937	18.180	15.593	37.739	
08VDL64	Paraguaná	272.2 $\pm$ 2.6	8.97	5.53	1.66	18.617	15.637	38.443	18.262	15.618	37.893	
10VDL05	Santander	198.3 $\pm$ 1.8	15.86	22.66	3.12	19.272	15.706	39.543	18.871	15.686	38.593	
10VDL22	Santander	199.1 $\pm$ 1.3	12.01	19.66	2.65	19.256	15.700	39.520	18.805	15.678	38.428	
10VDL23	Santander	461.0 $\pm$ 2.1	11.99	6.80	2.36	19.381	15.676	39.020	18.435	15.623	38.143	
10VDL37	Santander	451.5 $\pm$ 1.3	7.73	4.08	1.17	19.310	15.693	38.891	18.599	15.653	38.093	
10VDL39	Santander	208.8 $\pm$ 1.2	49.35	0.34	1.15	19.074	15.721	39.034	19.024	15.718	39.029	
10VDL43	Santander	209.2 $\pm$ 3.4	12.56	5.24	1.08	19.523	15.734	39.414	19.107	15.711	38.765	
10VDL44	Santander	473.5 $\pm$ 2.5	27.50	18.33	2.64	19.232	15.743	39.631	18.754	15.716	38.565	
10VDL47	Santander	483.7 $\pm$ 5.9	5.06	6.37	1.24	19.642	15.778	39.976	18.382	15.707	37.896	
10VDL50	Santander	477.0 $\pm$ 5.3	22.48	29.88	3.93	19.639	15.738	40.245	18.760	15.689	38.096	
10VDL51	Santander	472.5 $\pm$ 3.4	8.37	17.53	3.02	20.170	15.787	40.671	18.329	15.683	37.237	
10VDL55	Santander	443.4 $\pm$ 3.2	15.19	10.91	2.39	19.010	15.679	38.885	18.288	15.638	37.824	
10VDL61	Santander	200.0 $\pm$ 1.5	11.93	10.88	1.51	19.337	15.701	39.504	19.076	15.688	38.892	

a: 490 Ma

b: 450 Ma

c: 420 Ma

d: 400 Ma