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DETRITAL ZIRCON GEOCHRONOLOGY OF JURASSIC SANDSTONES OF WESTERN CUBA (SAN CAYETANO FORMATION): IMPLICATIONS FOR THE JURASSIC PALEOGEOGRAPHY OF THE NW PROTO-CARIBBEAN

YAMIRKA ROJAS-AGRAMONTE*†, ALFRED KRÖNER*, JAMES PINDELL**, ANTONIO GARCÍA-CASCO***, DORA GARCÍA-DELGADO§, DUNYI LIU§§, and YUSHENG WANG§§

ABSTRACT. Clastic sediments of the early (?) to late Jurassic (Oxfordian) San Cayetano Formation of western Cuba are interpreted to reflect syn-rift sedimentation coeval with the breakup of Pangaea. This sedimentary unit is the oldest known in the Guaniguanico Mountains and Cuba. U-Pb SHRIMP dating of 19 detrital zircon grains from two samples of San Cayetano micaceous sandstone provided concordant ages ranging from ~398 to 2479 Ma. The oldest zircon population is of Paleoproterozoic age (~2479 – 1735 Ma), but most zircons have early Mesoproterozoic and Grenvillian ages (~1556 – 985 Ma), whereas still younger ages are Pan-African (561 Ma), Ordovician (451 Ma) and early Devonian (~398 Ma). We discuss the possible origin of these zircons and conclude that the most likely source terrain(s) are Precambrian and early Paleozoic massifs in northern South America (Colombia and/or Venezuela) and the Yucatán Peninsula in Mexico. This is compatible with paleogeographic reconstructions of the Caribbean that imply that sediments of the San Cayetano Formation were still part of the disintegrating supercontinent Pangea in pre mid-Oxfordian time.

INTRODUCTION

The importance of detrital zircon ages in sedimentary provenance studies is widely recognized. Detrital zircon ages provide important constraints on basin evolution, tectonic history and the paleoposition of crustal blocks as well as the erosional history of ancient orogenic systems (for example, Ross and Bowring, 1990; Gehrels and Dickinson, 1995; Gehrels, 2000; DeGraaff-Surpless and others, 2002, 2003; Hodges and others, 2005; Andersen, 2005). Provenance studies of zircons from the early to late Jurassic San Cayetano Formation in western Cuba (figs. 1 and 2) provide important constrains on the position of western Cuba during and after the breakup of Pangea, as well as providing a critical test for hypotheses concerning the formation's paleogeography.

The San Cayetano Formation forms part of the early(?) Jurassic to middle Eocene Guaniguanico Terrane (fig. 2B), which encompasses several tectonostratigraphic belts that record the transition from Pangea to Laurentian margin, Proto-Caribbean basin, and Tertiary foreland (Pszczółkowski, 1978, 1999; Pszczółkowski and Myczyñski, 2003; Iturralde-Vinent, 1994, 1998, 2006; Bralower and Iturralde-Vinent, 1997). It has been a matter of debate whether the Guaniguanico Terrane is *in-situ* (Alva-Valdivia, and others, 2001), or allochthonous (Rosencrantz and Pardo, 1993; Pindell, 1994; Iturralde-Vinent, 1994, 1998; Pszczółkowski, 1999; Kerr and others, 1999). However, most authors agree that the San Cayetano depocenter was somewhere in the southern

*Institut für Geowissenschaften, Universität Mainz, 55099 Mainz, Germany, and SHRIMP-Center, Institute of Geology, Chinese Academy of Geological Sciences, 26 Baiwanzhuang Road, Beijing, 100037, China

**Tectonic Analysis, Ltd., Cokes, Barn, West Burton, West Sussex RH20 1HD, United Kingdom

***Departamento de Mineralogía y Petrología and Instituto Andaluz de Ciencias de la Tierra, Universidad de Granada-CSIC, Avda. Fuentenueva sn, 18002, Granada, Spain

§Centro de Investigaciones del Petróleo, Washington 169, Habana 12000, Cuba

§§SHRIMP-Center, Institute of Geology, Chinese Academy of Geological Sciences, 26 Baiwanzhuang Road, Beijing 100037, China

†Corresponding author: rojas@uni-mainz.de

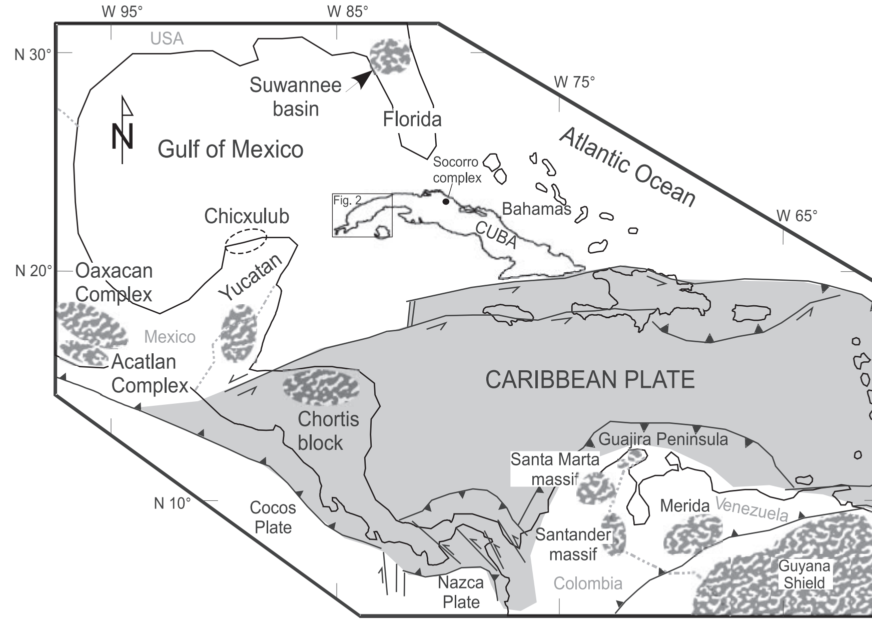

Fig. 1. Present-day map of the Caribbean region showing location of pre-Mesozoic rocks (filled elliptical areas) in Florida, central America and northern South America (modified after Hutson and others, 1998). Location of figure 2B is outlined.

Yucatan Peninsula and that the San Cayetano Formation records sedimentation in grabens and halfgrabens during the early breakup of Pangea and prior to formation of the Proto-Caribbean basin (Marton and Buffler, 1994, 1999; Lopez and others, 2003; Iturralde-Vinent, 2003).

Three regions have been considered as potential source areas for the siliciclastic rocks of the San Cayetano Formation (fig. 1): (1) southern North America (Florida), (2) central America (Maya Block, Yucatan Peninsula), and (3) northern South America (Colombia and Venezuela). For example, Anderson and Schmidt (1983), Ryabukhin and others (1984), Pszczółkowski (1987a) and Cobiella-Reguera (2000) suggested a South American source for the detritus, whereas Iturralde-Vinent (1994, 1996, 1998) and Pszczółkowski (1999) place the San Cayetano basin somewhere between the Yucatan block in Central America and northern South America, favoring input from both regions. Hutson and others (1998), on the other hand, suggested a more northerly-northwesterly source (North America and/or the Yucatan Peninsula).

Directions of sediment transport from the south or southwest, measured in sandstones of the San Cayetano Formation in the Sierra de los Órganos belt (fig. 2B; Haczewski, 1976), suggest a South American source for the detritus. The Jurassic paleogeographic reconstruction of Iturralde-Vinent (1994) and new stratigraphic data enabled Pszczółkowski (1999) to re-interpret the significance of original paleocurrent measurements from Haczewski (1976). Pszczółkowski (1999) argued that the main transport direction for the clastic material was probably from the southwest and not directly from the rift valley escarpments; and from the SE to NW in the southern Rosario Belt, west of La Palma town (fig. 2B). Based on detrital mica Ar-Ar ages, however, Hutson and others (1998) concluded that the source area was located in North America and/or the prolongations of North American Taconic and Acadian orogenic belts exposed on the Yucatan Peninsula.

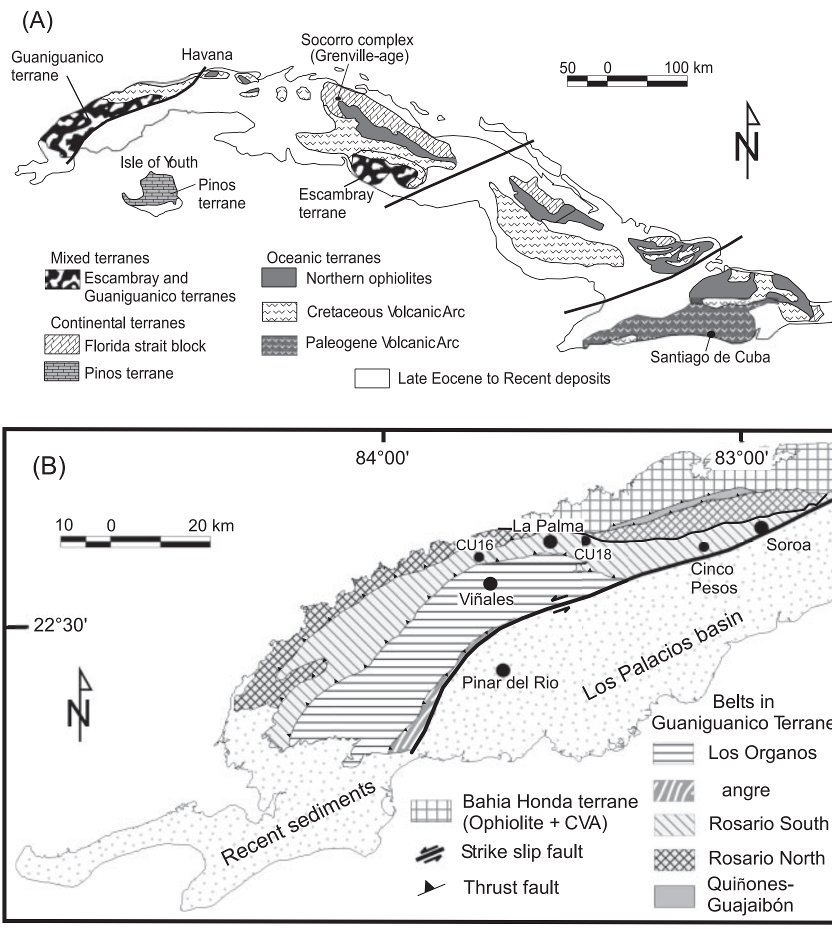

Fig. 2. (A) Schematic geologic map of Cuba, showing distribution of oceanic and continental terranes as well as mixed terranes (after Iturralde-Vinent, 1998). (B) Map of the study area in western Cuba showing belt subdivision within the Guaniguanico terrane (modified after Iturralde-Vinent, 1998 and Pszczółkowski, 1999). San Cayetano Formation sample localities CU 16 and CU 18 are also shown. CVA=Cretaceous Volcanic Arc.

Paleogeographic reconstructions for the Caribbean plate in the Jurassic show that separation of the Americas involved middle to late Jurassic rifting and the onset of seafloor-spreading in the Gulf of Mexico and the Proto-Caribbean to produce the American continental passive margin of the Caribbean region (Iturralde-Vinent, 1998; Pszczółkowski, 1999; Pindell and others, 2005 and references therein). According to Pindell and others (2000) the paleo-kinematics of the Proto-Caribbean region require westward-propagating early to middle Jurassic rifting, followed by late Jurassic seafloor-spreading. This is consistent with the presence of marine sediments in the upper part of the San Cayetano Formation, although overall transgression occurred during the Oxfordian (Jagua Formation; Pszczółkowski, 1978, 1987a, 1999).

Iturralde-Vinent (2003, 2006) proposed that a network of rift basins and valleys developed along the present continental margin of North America into the Gulf of Mexico, and along the Mexican terranes into northern South America, during the latest Triassic–Jurassic as a result of Pangea breakup. Rifting then continued along the suture between Gondwana and Laurentia to form precursors of the Caribbean basin and the Gulf of Mexico. Early to middle late Oxfordian fossils of marine and terrestrial origin in the Guaniguanico Terrane probably represent part of the emerging Florida-Yucatan ridge and the early Proto-Caribbean seaway. A thick Oxfordian basaltic unit in Guaniguanico corroborates the late Jurassic opening of the Caribbean as a seaway (Pszczółkowski, 1987a; Iturralde-Vinent, 1994, 2006).

There is still controversy regarding the provenance and original location of the San Cayetano basin which, as stated before, suggest three different possible scenarios (Central America, North, and South America). This has important consequences for the early evolution of the Caribbean plate in general and the palaeogeography of western Cuba in particular. We present SHRIMP U-Pb detrital zircon ages from micaceous sandstones of the San Cayetano Formation and discuss their implications for the Jurassic paleogeography in order to constrain the early stratigraphic and tectonic evolution of the Caribbean region.

GEOLOGIC SETTING

Regional Geology

Cuba, the largest of the Antilles islands (fig. 1), exposes Jurassic to middle Eocene sediments as well as minor Neoproterozoic rocks, which now form part of the Caribbean orogenic belt. This belt was produced by the convergence and collision of the Caribbean and North and South American plates during the Cretaceous and Tertiary (Iturralde-Vinent, 1994, 1996). In Cuba (fig. 2A), the rocks within the orogenic belt represent crustal fragments detached from the North American plate, the Caribbean crust and the Pacific realm (Iturralde-Vinent, 1998). Several terranes are recognized within the Cuban orogenic belt (Iturralde-Vinent, 1998; see also Iturralde-Vinent, 1994, 1996; fig. 2A): (1) continental terranes (Florida Strait block and Pinos Terrane), (2) mixed continental and oceanic terranes (Guaniguanico and Escambray Terranes), (3) oceanic rocks (the Northern ophiolite and Cretaceous and Paleogene island arcs), and (4) post-volcanic (Campanian-late Eocene) and foreland (Paleocene-late Eocene) basins.

Several authors have suggested that the Pinos, Guaniguanico and Escambray terranes were formed in the Caribbean borderland of the Maya Block (Iturralde-Vinent, 1996, 1998, 2006; Pindell and others, 2006; see fig. 1). Pszczółkowski (1999) thought it probable that the Mezosoic Guaniguanico successions were deposited some 100 km east and NE of the present Yucatan coast. The most widespread rocks in the Guaniguanico, Pinos and Escambray terranes are Mesozoic continental margin deposits although ophiolitic rocks and volcanic sequences have also been described from the Guaniguanico and Escambray terranes (Somin and Millán, 1972; Millán, 1996).

The tectonic models of Iturralde-Vinent (1996, 1998) and Pszczółkowski (1999) for western Cuba involve northeastward migration and early to middle Eocene collision of the arc with the passive margin of North America. In this model, northward thrusting led to imbrication of deep-water, slope, and platform deposits (including the San Cayetano Formation) and reversed the original paleogeographic position of the Guaniguanico units. As a result, the deep water and most distal deposits are the farthest travelled (Iturralde-Vinnent, 1994, 1996).

Belt Subdivision within the Guaniguanico Terrane

The Guaniguanico terrane, located in western Cuba (fig. 2B), is characterized by north-verging thrusts and partially superimposed belts of Jurassic-Cretaceous sedimentary sequences and syntectonic Paleogene foreland sediments (fig. 2B; Iturralde-Vinent, 1994, 1996; Bralower and Iturralde-Vinent, 1997; Pszczółkowski, 1999). Several authors have attempted to subdivide the sequences within this terrane into different belts according to lithostratigraphy and facies development (Pszczółkowski, 1985, 1994a, 1999; Iturralde-Vinent, 1994, 1996, 1998). In general, five tectono-stratigraphic belts are recognized (fig. 2B), namely the Cangre, Los Organos, Rosario South, Rosario North and Guajaibon-Quiñones belts, each represented by several smaller thrust sheets. These belts are abruptly truncated to the south by the Pinar fault zone and are tectonically overlain by allochthonous, north-verging ophiolites and Cretaceous volcanic arc units (fig. 2B; Pszczółkowski, 1978, 1999; Iturralde-Vinent, 1994, 1998).

The two dated samples come from the Rosario South belt, which occurs tectonically above the Los Organos belt. The Rosario South belt includes deep marine carbonate and clastic facies, ranging in age from late Jurassic to early Cretaceous, as well as more fragmented late Cretaceous carbonates (Pszczółkowski, 1994b). The youngest rocks within this belt are Paleocene carbonates, early Eocene flysch, and olistostromes deposits (Pszczółkowski, 1999). Thick basaltic pillow lavas intercalated with limestones and shales are also present within this belt (Pszczółkowski, 1999) and were interpreted by Iturralde-Vinent (1998) and Kerr and others (1999) as intracontinental rift basalts.

San Cayetano Formation

This formation was first defined by De Golyer (1918) and is the oldest stratigraphic unit known in the Guaniguanico Terrane and Cuba (fig. 2). It has a wide spatial distribution in the Cangre, Los Organos and Rosario Belts (fig. 2B). The formation is made up of 1000 to 3000(?) m of grayish, reddish and yellowish (although black when fresh) medium- to fine-grained micaceous sandstones, siltstones, mudstones, mica-rich gray shales, carbonaceous shales, and arkoses (fig. 3), with minor volcanic rocks reported from its base (Haczewski, 1976). Intercalations of limestone and conglomerate have also been reported with the limestones occurring mostly in the upper part (Pszczółkowski, 1999). The formation is locally cut by mafic dikes and sills. Abundant dark carbonaceous organic material occurs within the mudstones and shales. The San Cayetano Formation is conformably overlain in the Los Organos belt by the Jagua Formation and, in the Rosario belts, by the Francisco formation. All these units are unconformably covered by Neogene and Quaternary deposits along the southern boundary of the Guaniguanico Mountain, whereas late Eocene to Recent deposits occur in tectonic contact in the rest of the mountain (Pszczółkowski, 1999).

Detailed descriptions of the composition of the San Cayetano sandstone were provided by Pszczółkowski (1986), Cobiella and others (1997), and Hutson and others (1998). According to Pszczółkowski (1978, 1999) and Cobiella, (1997), the source area of the San Cayetano sediments probably consisted of metasedimentary, terrigenous and granitoid rocks that, together with prolonged transport, recycling and weathering, resulted in a high content of quartz. The age of the San Cayetano Formation has been reported as early? to late Jurassic (middle Oxfordian; Pszczółkowski, 1978, 1999). However, early Jurassic strata have never been identified with confidence (Pszczółkowski, 1978, 1999). Only the age of the upper part of the formation is well constrained by ammonites (Myczyñski and Pszczółkowski, 1976).

Hutson and others (1998) reported detrital mica Ar-Ar ages from the San Cayetano Formation ranging from 300 to 1950 Ma. They divided these ages into three groups, two of Precambrian age (1.95 – 1.625 Ga and 1.325 – 0.94 Ga) and one of dominantly Paleozoic age (0.47 – 0.308 Ga). These authors suggested that the first group could have been derived from North or South America, although they favored a North American source based on an age overlap with the North American Yavapai-Mazatzal, Granite-Rhyolite and Grenville crustal age provinces. They considered the Paleozoic age group to be related to the Taconic orogeny in eastern North America. Hutson and others (1998) consequently proposed that the San Cayetano basin formed

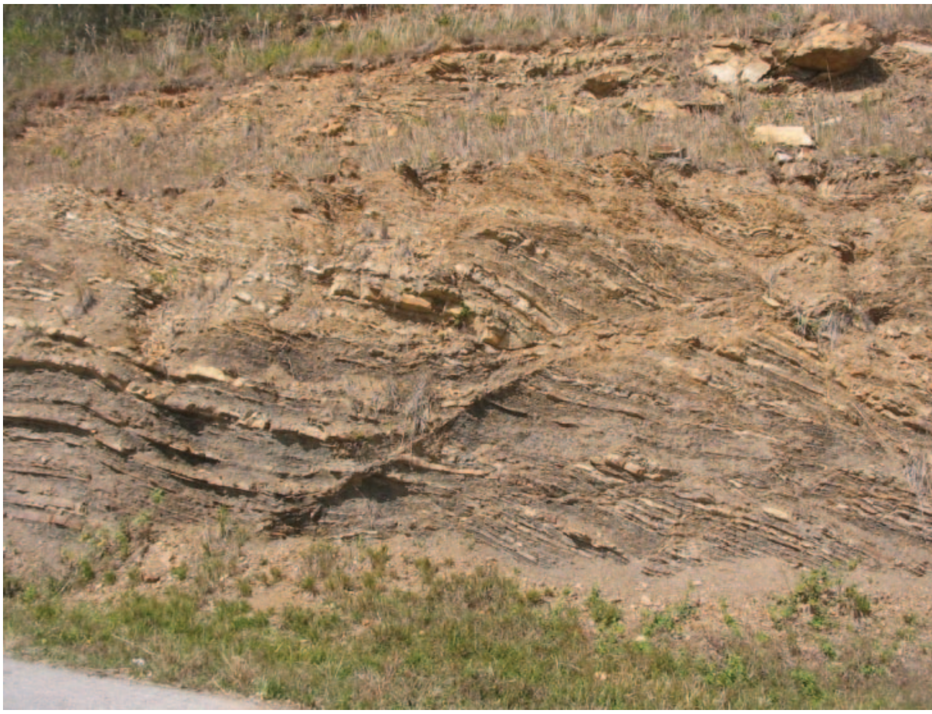

Fig. 3. San Cayetano Formation on road from Cinco Pesos to Soroa (fig. 2B for location). The exposure is characterized by fine interbedding of micaceous sandstone, siltstone and shales. Several thrust faults cut the sequence.

along the southeastern margin of the Yucatan Peninsula but was derived from North America and/or the prolongations of the North American Taconic and Acadian orogenic belts exposed on the Yucatan Peninsula.

Haczewski (1976) concluded that the uniformity in composition and thickness of the San Cayetano Formation over an outcrop length of ~100 km, suggested that these siliciclastic sediments were deposited with sheet-like geometry on a continental coastal plain in a rift setting near a stable cratonic margin. Sedimentary structures indicate that the rocks comprise intermingled terrestrial, alluvial, lagoonal and shallow marine beds deposited in a transitional rift-drift setting (Haczewski, 1976; Iturralde-Vinent, 1998).

SHRIMP DATING PROCEDURE

Zircons were handpicked and mounted in epoxy resin together with chips of the Perth Consortium standard CZ3. The mount was then polished, cleaned, and photographed in reflected and transmitted light and under cathodoluminescence (CL) to bring out the internal structures of the grains (fig. 4). CL images were obtained on a JEOL JXA-8900RL microprobe at the University of Mainz, and operating conditions were 15 kV accelerating voltage and 12 nA beam current.

Isotopic analyses were performed on the SHRIMP II ion microprobe in the SHRIMP Centre of the Chinese Academy of Geological Sciences in Beijing. Instrumental characteristics were outlined by De Laeter and Kennedy (1998). For data collection on detrital grains, four scans through the critical mass range were made, and counting times on individual masses were $^{196}Zr_2O$ = 2s, ^{204}Pb = 10s, background = 10s, ^{206}Pb =

20s, ^{207}Pb = 30s, ^{208}Pb = 10s, ^{238}U = 10s, ^{248}ThO = 5s, ^{254}UO = 2s. The analytical procedures are detailed in Claoué-Long and others (1995), Nelson (1997) and Williams (1998).

Precise dating of young zircons (<1000 Ma) by ion-microprobe is best achieved by using concordant ^{206}Pb/^{238}U-ages, whereas older zircons usually provide precise ^{207}Pb/^{206}Pb ages (see Black and others, 2003, for explanation). The reduced ^{206}Pb/^{238}U ratios were normalized to the Perth standard CZ3 (^{206}Pb/^{238}U = 0.09432, age: 564 Ma), and the error in the ratio ^{206}Pb/^{238}U during analysis of all standard zircons during this study was 1.35 percent. Analyses of samples and standards were alternated to allow assessment of Pb^+/U^+ discrimination. Primary beam intensity was about 3.5 nA, and a Köhler aperture of 100 μm diameter was used, giving a slightly elliptical spot size of about 25 μm. Peak resolution was about 5000, enabling clear separation of the ^{208}Pb-peak from the nearby HfO_2-peak. Sensitivity was around 22 to 24 cps/ppm/nA

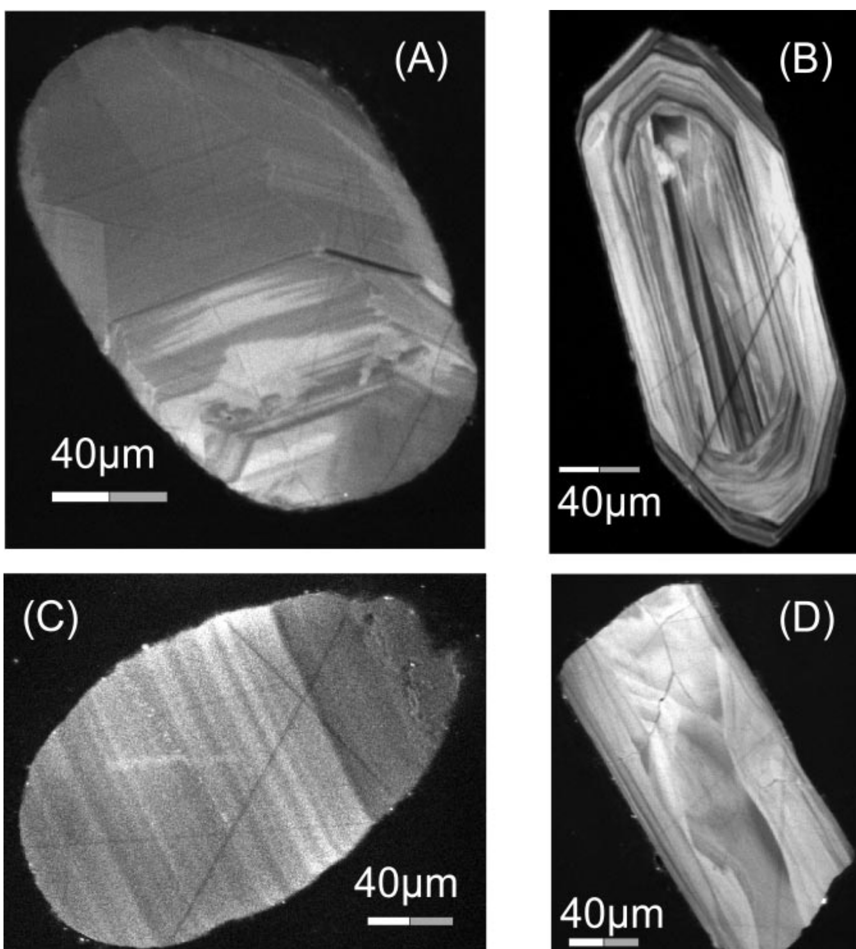

Fig. 4. Cathodoluminescence images of selected detrital zircon grains dated in this study. (A) Well rounded grain with broad magmatic zoning from sample CU 16; (B) near-idiomorphic grain with well developed oscillatory magmatic zoning from sample CU 16; (C) well-rounded grain with striped magmatic zoning from sample CU 18; (D) fragment of idiomorphic grain with diffuse zoning from sample CU 18.

Pb on standard CZ3. Raw data reduction followed the method described by Nelson (1997). Common Pb is considered to be surface-related (Kinny, 1986), and corrections have been applied using the ^{204}Pb-correction method and assuming the isotopic composition of Broken Hill lead (Cumming and Richards, 1975). The analytical data are presented in table 1. Errors given for individual analyses are based on counting statistics and are at the 1-sigma level and include the uncertainty of the standard added in quadrature. Stern (1997) provides a detailed account of the counting error assessment for SHRIMP analyses. Errors for pooled analyses are reported at the 2-sigma level.

ZIRCON AGES

Two samples of micaceous sandstone from the upper section of San Cayetano Formation were collected north of Viñales at the road intersection to La Palma (Cu16, N22° 41' 58.6"; W83° 44' 28.7") and along the road to La Palma (CU 18, N22° 45' 27.7"; W83° 28' 25.5") in the Rosario South belt. The lithology at both localities consists of well stratified yellowish fine to medium grained micaceous sandstone with intercalations of limestone. Detrital zircon ages were determined in order to constrain the provenance of this formation and place new constraints on the Mesozoic paleoposition of western Cuba.

The zircons from sample Cu 16 are relatively small (~50 – 150 μm) and predominantly short-prismatic although rare long-prismatic grains also occur (figs. 4A and 4B). Most grains are well rounded, indicating variable sedimentary transport, but angular fragments also occur, and these probably resulted from sample crushing. Cl images exhibit well preserved oscillatory magmatic zoning in many grains (fig. 4B), and some seem to have distinct cores surrounded by younger magmatic overgrowth. Variable U-content is indicated by low and high luminenscence.

Ten zircon grains were analyzed from sample Cu 16, and the results are listed in table 1 and are presented graphically in the Concordia diagram of figure 5A. All analyses are concordant, within error, and the ages range from 398 ± 7 to 2282 ± 10 Ma. Obviously, the number of analyses is too small to consider the age distribution as representative of the entire sample and therefore the absence of a particular age range may be a sampling bias, whereas the presence of any age is real data. It is therefore noticeable that there is only one Devonian grain, then a suite of Mesoproterozoic grains with ages between 985 and 1497 Ma, and one grain with a Paleoproterozoic age of 2282 Ma. The youngest grain shows least rounding, and this may signify relatively short sedimentary transport, whereas the oldest grain is the most rounded, which probably signifies derivation from a distant source or redeposition from a pre-San Cayetano clastic sediment such as the late Paleozoic sandstones of the Santa Rosa Formation (SE Mexico, Maya block) or equivalent units elsewhere in Florida (Suwannee basin), central America and northern South America (Weber and others, 2006). The remaining grains display variable degrees of rounding.

The zircons from sample CU 18 are noticeably larger (up to 200 μm) than those in CU 16 and are also better rounded, although near-euhedral and angular, broken grains also occur. Magmatic, oscillatory zoning is conspicuous in many grains, and some display the broad sector zoning frequently found in zircons of metamorphic origin (figs. 4C and 4D; Hoskin and Black, 2000; Corfu and others, 2003).

Nine grains were analyzed from sample CU 18, and the results are listed in table 1 and are plotted in the Concordia diagram of figure 5B. As in sample Cu 16, there is a Paleozoic group represented by two grains with nearly identical ages and a mean of 452 ± 2 Ma. These, and a late Neoproterozoic grain with an age of 561 ± 4 Ma, are least rounded and are interpreted to be derived from not-too-distant sources. The spread of ^{207}Pb/^{206}Pb ages from 1034 to 1735 Ma records input from Mesoproterozoic

TABLE 1

SHRIMP II analytical data for spot analyses of single zircons from clastic sediment samples CU16 and CU 18, San Cayetano Fm., western Cuba[1]

Samle No.	U ppm	Th ppm	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	$^{207}Pb/^{206}Pb$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{206}Pb/^{238}U$ age ± 1σ	$^{207}Pb/^{235}U$ age ± 1σ	$^{207}Pb/^{206}Pb$ age ± 1σ
Cu 16-1*	242	55	8738	0.0696 ± 16	0.0729 ± 8	0.1710 ± 41	1.7187 ± 472	1017 ± 23	1016 ± 17	1012 ± 22
Cu 16-2	220	104	9708	0.1426 ± 19	0.0728 ± 8	0.1707 ± 41	1.7137 ± 473	1016 ± 23	1014 ± 18	1008 ± 22
Cu 16-3	998	621	2520	0.1351 ± 17	0.0720 ± 7	0.1652 ± 40	1.6385 ± 445	985 ± 22	985 ± 17	985 ± 21
Cu 16-4	405	190	9018	0.1425 ± 13	0.0783 ± 6	0.1958 ± 47	2.1140 ± 549	1153 ± 25	1153 ± 18	1154 ± 14
Cu 16-5	128	47	11629	0.1033 ± 14	0.1445 ± 8	0.4239 ± 103	8.4468 ± 2162	2278 ± 46	2280 ± 24	2282 ± 10
Cu 16-6	443	859	1698	0.3850 ± 31	0.0816 ± 12	0.2136 ± 51	2.4032 ± 719	1248 ± 27	1244 ± 22	1236 ± 30
Cu 16-7	268	111	8849	0.1303 ± 17	0.0781 ± 8	0.1993 ± 48	2.1469 ± 580	1171 ± 26	1164 ± 19	1150 ± 19
Cu 16-8	500	200	3815	0.0903 ± 15	0.0935 ± 7	0.2596 ± 62	3.3448 ± 871	1488 ± 32	1492 ± 21	1497 ± 14
Cu 16-9	640	216	2930	0.0985 ± 33	0.0541 ± 3	0.0636 ± 11	0.4746 ± 152	398 ± 7	394 ± 10	375 ± 57
Cu 16-10	97	28	3246	0.0860 ± 43	0.0777 ± 9	0.1912 ± 33	2.0488 ± 652	1128 ± 18	1132 ± 21	1140 ± 49
Cu 18-1	406	153	599880	0.1128 ± 11	0.0726 ± 5	0.1741 ± 12	1.7433 ± 187	1034 ± 7	1025 ± 7	1004 ± 15
Cu 18-2	561	176	1440922	0.0899 ± 7	0.0732 ± 4	0.1841 ± 13	1.8591 ± 178	1089 ± 7	1067 ± 6	1021 ± 11
Cu 18-3	174	73	333333	0.1134 ± 10	0.1622 ± 8	0.4672 ± 36	10.4504 ± 995	2471 ± 16	2476 ± 9	2479 ± 8
Cu 18-4	91	82	25072	0.2908 ± 64	0.0561 ± 18	0.0681 ± 5	0.5263 ± 177	425 ± 3	429 ± 12	455 ± 71
Cu 18-5	308	97	806452	0.0903 ± 8	0.0956 ± 5	0.2707 ± 20	3.5683 ± 348	1544 ± 10	1543 ± 8	1540 ± 11
Cu 18-6	106	64	7061	0.1867 ± 63	0.0602 ± 25	0.0910 ± 7	0.7554 ± 327	561 ± 4	571 ± 19	611 ± 90
Cu 18-7	154	77	107492	0.1679 ± 33	0.0560 ± 11	0.0680 ± 5	0.525 ± 115	424 ± 3	428 ± 8	451 ± 44
Cu 18-8	79	80	744602	0.2825 ± 29	0.1054 ± 10	0.3089 ± 25	4.4901 ± 606	1735 ± 12	1729 ± 11	1722 ± 18
Cu 18-9	186	55	9460	0.0936 ± 16	0.0964 ± 8	0.2744 ± 50	3.6491 ± 767	1563 ± 25	1560 ± 17	1556 ± 16

[1]All errors are 1-sigma. *Cu 16-1 is spot 1 on grain 1, 16-2 is spot 1 on grain 2, *et cetera*

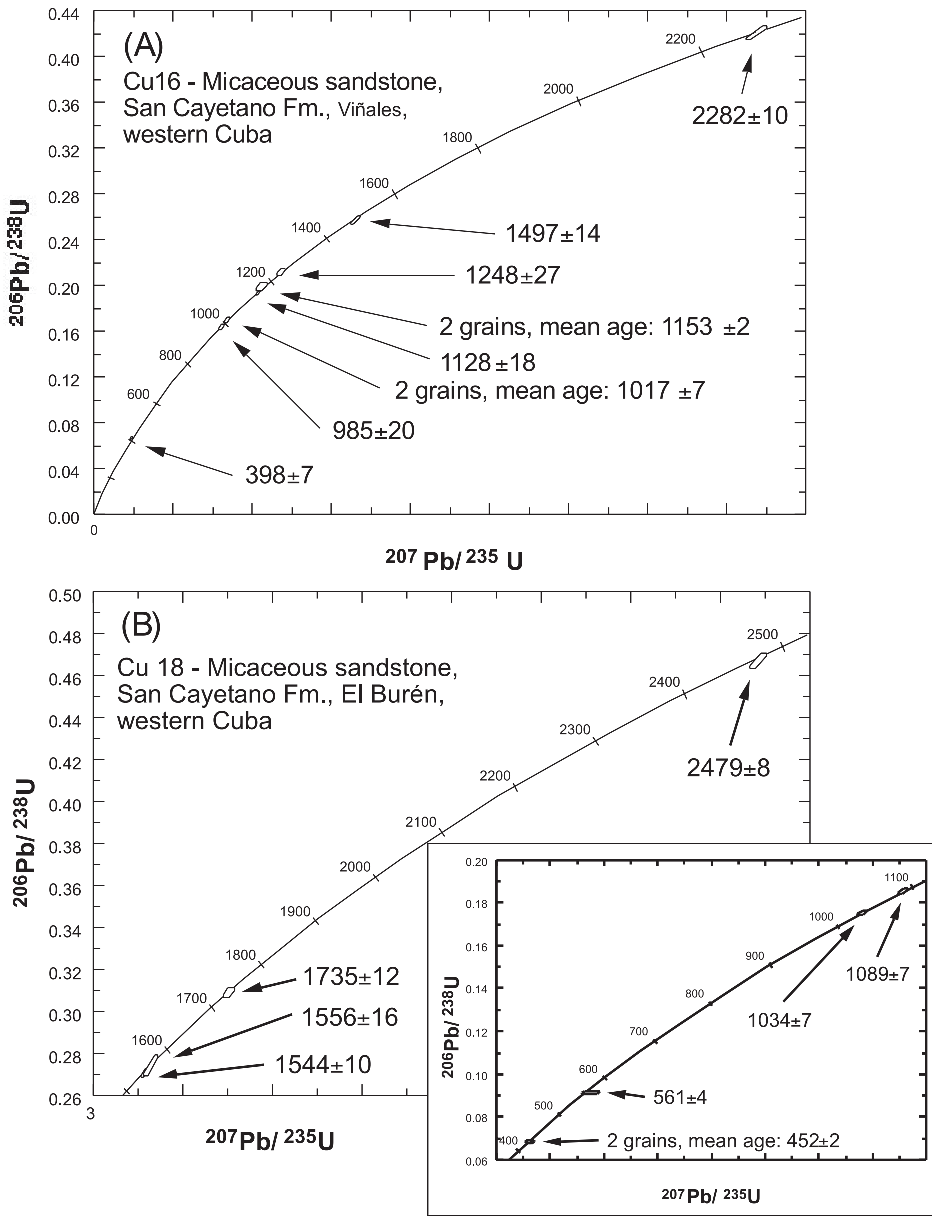

Fig. 5. Concordia diagrams showing SHRIMP analyses of zircons from micaceous sandstones CU 16 (A) and CU 18 (B) of the San Cayetano Formation. Data boxes for each analysis are defined by standard errors in $^{207}Pb/^{235}U$, $^{206}Pb/^{238}U$, and $^{207}Pb/^{206}Pb$. Errors of pooled mean ages are 2σ.

to late Paleoproterozoic terranes, whereas one early Paleoproterozoic grain with a $^{207}Pb/^{206}Pb$ age of 2479 ± 8 Ma is again the most rounded in the zircon population.

In combination, the 19 detrital zircons from sandstones of the San Cayetano Formation predominantly reflect sedimentary input from a Mesoproterozoic source. The following discussion focuses on the question of which source area(s) are most appropriate.

DISCUSSION

In view of the detrital zircon age spectrum (fig. 6) the source area(s) for the San Cayetano sediments must include rocks with Paleoproterozoic, Mesoproterozoic, Neoproterozoic, late Ordovician and early Devonian ages. In Central America and Mexico, the late Mesoproterozoic (Grenvillian) Oaxaca, Complex and neighboring Paleozoic Acatlán terrane may be considered as one possible source of detritus for the sediments of the San Cayetano Formation. The late Paleozoic Santa Rosa Formation, which occurs along the southern limit of the Maya block, north of the border between the North American and the Caribbean plates, could also be regarded as a possible source of detritus. Weber and others (2006) reported detrital Silurian, Pan-African/Brasiliano, Grenville, Mesoproterozoic, Paleoproterozoic and Archean ages from the Santa Rosa Formation. Muller and others (1994) and Heatherington and Mueller (1997) also reported single zircon detrital ages from Ordovician to Devonian sediments of the Suwannee basin in Florida. These authors identified two main populations, from 515 to 637 Ma and 1967 to 2282 Ma, but also found Archean (~2680 Ma) and Paleoproterozoic (2463 and 1750 Ma) grains. We do not entirely discard the possibility that North America could have provided detritus for the San Cayetano sediments but taking into account the detrital nature of these grains if they were eroded and transported again, most should be very well rounded and that is rarely the

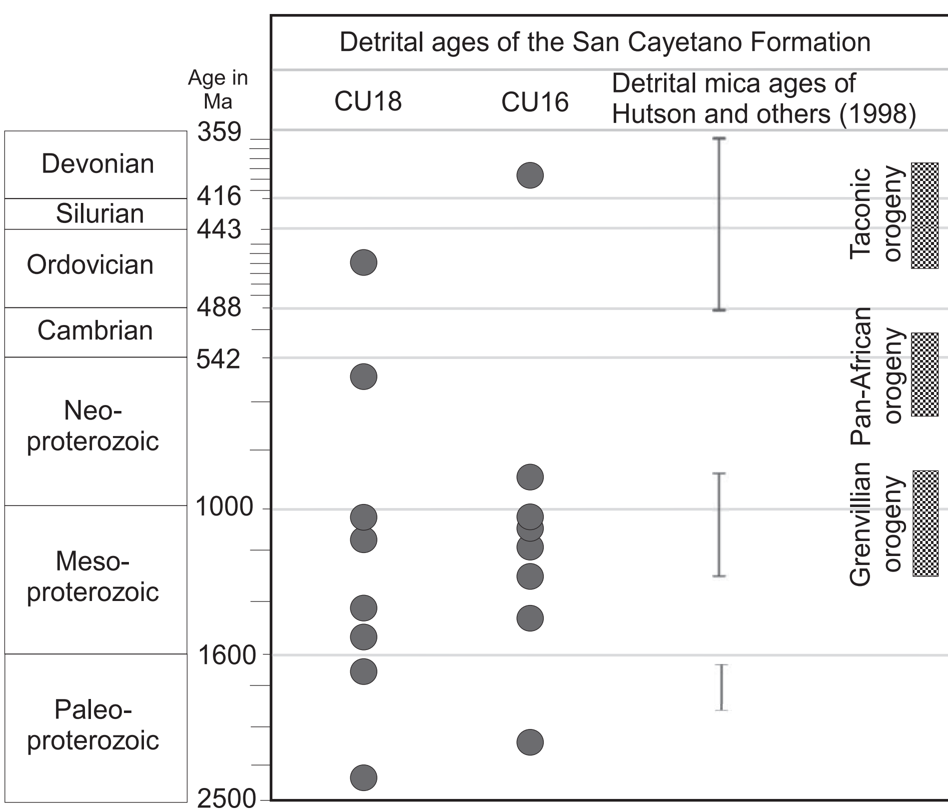

Fig. 6. Comparison of detrital mica ages of Hutson and others (1998) and zircon ages of this study (full circles) for micaceous sandstones of the San Cayetano Formation. The youngest zircon age of 308 Ma is not included in the figure. Timing of major orogenic cycles are shown on the right. Geologic time scale after Gradstein and others (2004).

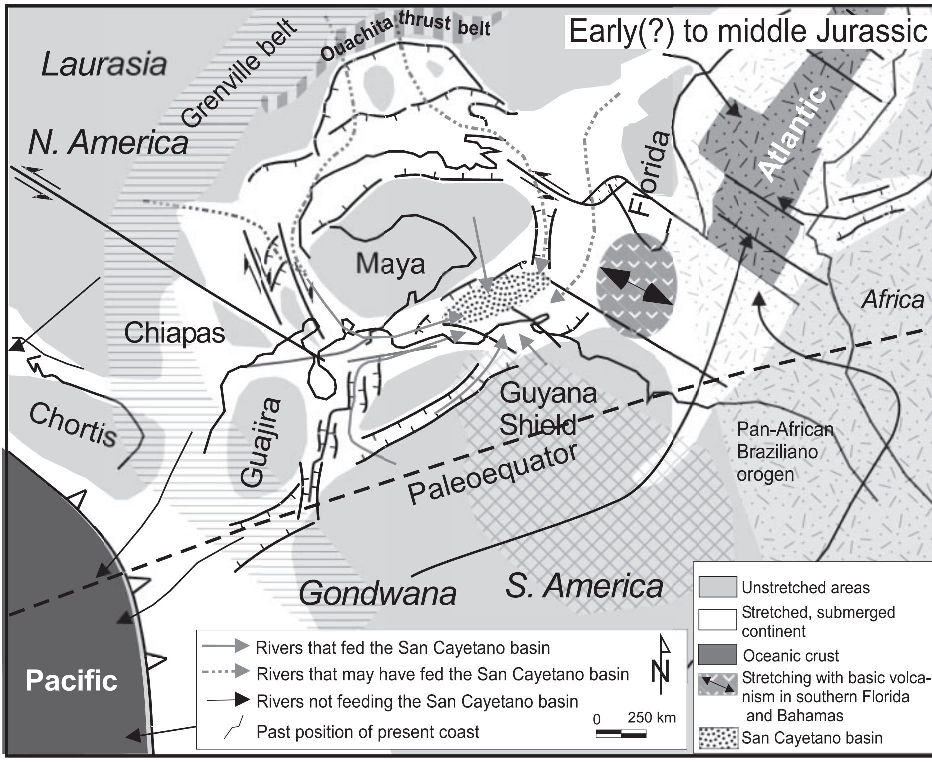

Fig. 7. Early (?) to middle Jurassic paleogeographic and tectonic reconstruction of western Pangea showing Precambrian and Paleozoic foldbelts and other units (modified after Pindell and others, 2000 and Pindell and Kennan, 2001). Arrows indicate probable sedimentary transport directions.

case here. Proterozoic and Paleozoic terranes are more abundant in northern South America and, as discussed below, provide a more likely location for the source rocks.

In the Mesozoic plate tectonic reconstructions of Pindell and others (2000, 2005) and Iturralde-Vinent (2003, 2006), the American plates began their separation in the latest Triassic (?), as early Jurassic rifts evolved into passive margins along northern South America and southern North America, and an east-dipping subduction zone defined the western limits of continental crust along the Cordilleran margin. In these models, the northern Venezuela-Trinidad margin rifted from eastern Yucatán, and the northwestern Colombia margin rifted from southern Mexico–eastern Chortis (fig. 1). Accordingly, northwestern South America must have occupied the present-day position of southern and central Mexico, and the San Cayetano clastic sediments may have been deposited in a depocenter that probably extended to the NW along the Yucatan margin. This region is now occupied by the oceanic Yucatan basin, which formed during the latest Cretaceous to early Tertiary (Rosencratz, 1996). This interpretation argues in favor of an allochtonous character for the Guaniguanico Terrane.

Constraints on Provenance and Correlations

In view of the reconstructions of Pindell and others (2000, 2005) and Iturralde-Vinent (2003), the most probable source regions for the Paleoproterozoic zircons (~1735, ~2288, ~2479 Ma) of the San Cayetano Formation are the central and western parts of northern South America (fig. 7). Paleoproterozoic zircon populations have been reported from the Guyana shield of the northern Amazonian craton in the Maroni-Itacaíunas Province (2.26 − 1.95 Ga; Tassinari and Macambira, 2004). Similar ages also occur in the Rio Negro and Juruena Provinces of the northern Amazon craton (1.8 – 1.55 Ga; Onstott and others, 1984; Teixeira and others, 1989; Tassinari, 1996) and the Imataca complex in the northern Guyana Craton (Onstott and others, 1984). The Guyana Shield, in the northwestern Amazon Craton, represents a large terrane that became tectonically consolidated during the Transamazonian orogenic cycle (2.26 – 1.95 Ga) and consists of extensive areas of Paleoproterozoic crust and two major Archean blocks, namely the Imataca Block in Venezuela and the Amapá Block in northern Brazil (da Rosa-Costa and others, 2006). However, Paleoproterozoic detrital populations (1.8 – 2.2 Ga) have also been reported from the late Paleozoic Upper Santa Rosa formation in SE Mexico and from similar rocks in the Suwannee basin in Florida (1.9 – 2.3 Ga; Mueller and others, 1994).

In the San Cayetano Formation, zircons of early Mesoproterozoic (~1497 Ma) and Grenvillian (~985 − 1248 Ma) age are the most abundant in both dated samples. They may have come from two different sources, namely the northern Colombian Andes (fig. 1; Guajira, Sierra Nevada de Santa Marta, Santander and Merida; Restrepo-Pace and others, 1997; Ordoñez, 1999; Cordani and others, 2005) and/or the Oaxacan complex of the Oaxaquia terrane in Mexico (fig. 1; Solari and others, 2003; Loewy and others, 2004). Cardona and others (2006) reported detrital zircons from a paragneiss sample in the Sevilla Complex (Sierra Nevada de Santa Marta) with ages varying from 500 to 1400 Ma, including two groups of Mesoproterozoic ages between 1120 and 1380 Ma and 920 to 1080 Ma. López and others (2001) recognized that Grenvillian inherited zircons found in cobbles and boulders in a Paleozoic conglomerate in north-central Mexico may have come from northern South America. Weber and others (2006) also reported detrital zircons of Mesoproterozoic (1.4 – 1.6) and Grenvillian (0.95 – 1.3 Ga) ages for the Upper Santa Rosa Formation, again suggesting derivation from Colombian or Venezuelan terranes in Northern South America or the Oxaquia terrane in Mexico.

The Mesoproterozoic Oaxacan complex constitutes the basement of Oaxaquia, a Precambrian-Paleozoic terrane, which underlies most of eastern Mexico and also appears to extend beneath the Chortis block of Honduras (Ortega-Gutiérrez and others, 1995; Keppie and Ortega- Gutiérrez, 1999). López and others (2001 and references therein) suggested that the Oaxacan complex was derived from Gondwana rather than representing an extension of the North American craton. Keppie and others (2001) proposed that Oaxaquia was a juvenile arc metamorphosed during a relatively late Grenvillian tectonothermal event at ~990 Ma and that it was located adjacent to northwestern Amazonia at ~1000 Ma.

Somin and Millán (1977) and Renne and others (1989) documented Grenvillian basement rocks (~904 Ma) from the Sierra Morena marbles in the small Socorro Complex of north-central Cuba (fig. 1) and proposed that these rocks formed part of a continuous band of Grenvillian basement extending from southwestern North America via Central America (Chortis block) to northwestern South America (fig. 7). In this context, Cardona and others (2006) also suggested that Proterozoic terranes in Mexico, the eastern Colombian Andes, and the southern Andes were all part of a long Mesoproterozoic Grenvillian collisional belt related to the formation of Rodinia. Other Grenvillian inherited zircon ages in Mexico come from the Ordovician Esperanza granitoids in the Acatlán complex (1165 ± 30 to 1043 ± 50 Ma, Talavera-Mendoza and others, 2005). Also, the metasedimentary rocks from the Piaxtla Suite in the Acatlán complex contain abundant Mesoproterozoic detrital zircons (1050 – 1250 Ma and few ages in the range of 900 – 992 and 1330 – 1662 Ma; Murphy and others, 2006). Meso- and early Neoproterozoic detrital ages (~1375 – 880 Ma) have also been reported from the Sierra Madre in Mexico (Nance and others, 2007).

Our single 561 Ma zircon grain falls within the age range of the Brasiliano/Pan-African orogenic cycle. Again northern South America (Santa Marta Massif in northern Colombia and Cordillera de Merida in northern Venezuela; López and others, 2001) could have contributed detritus to the San Cayetano basin. Burkley (ms, 1976) and Maurrasse (1990) also reported U–Pb zircon ages for Pan-African/Brasiliano magmatism in Venezuela, whereas Mueller and others (1994) reported detrital zircons from the Suwannee basin in Florida (North America) whose ages correspond to the Pan-African orogeny (550 – 650 Ma). Hutson and others (1998), however, noted a lack of Pan-African/Brasiliano ages in their study of detrital micas in the San Cayetano Formation, which could mean that such micas did not survive long sedimentary transport whereas the zircons did.

López and others (2001) also provided evidence for Pan-African/Brasiliano-age basement (580 Ma) in Mexico from cobbles and boulders in a Paleozoic conglomerate at Las Uvas Coahuila (north-central Mexico). Previous reports of Pan-African/Brasiliano ages in Mexico come from shocked zircon (545 ± 5 Ma) in an impact breccia of the Chicxulub crater in eastern Yucatan (Krogh and others, 1993). Therefore, we consider it likely that Mexico also provided some detritus to the San Cayetano basin. Nance and others (2007) reported Neoproterozoic to early Cambrian detrital zircon ages of ~650 to 525 Ma from the Sierra Madre terrane in Mexico and suggested provenance from the Maya terrane beneath the Yucatan Peninsula or the Brasiliano orogens of South America. The detrital ages of the Upper Santa Rosa Formation are dominated by Pan-African/Brasiliano ages (~630 Ma, 540 – 560 Ma, ~520 Ma; Weber and others, 2006), suggesting that it, too, could have provided detritus to the San Cayetano basin. Detrital Pan-African ages have also been reported from sandstone samples in Florida (515 – 637 Ma; Mueller and others, 1994). Weber and others (2006) propose a similar provenance (West Africa and northeastern South America) for the Florida sedimentary basement and the Upper Santa Rosa Formation.

The Paleozoic detrital mica ages reported by Hutson and others (1998) were linked by these authors to the Taconic orogeny in eastern North America since orogenic belts of this age were unknown in Florida and northernmost South America. However, Chew and others (2007) recently reported the presence of an early to middle Ordovician magmatic and metamorphic belt in the north-central Andes which continues from Patagonia to as far north as Colombia and Venezuela. They compared this belt in western Gondwana with the Taconic–Grampian belt along the eastern Laurentian margin. Our single late Ordovician zircon grain (~452 Ma) may therefore come from Venezuela (the Caparo arc of Bellizzia and Pimentel, 1994), or Colombia (Boinet and others, 1985) where Ordovician subduction-related magmatism has been discovered. Cardona and others (2006) reported a late Ordovician zircon core age (456 ± 10 Ma) from the Siapana granodiorite in the Guajira Peninsula. It remains a puzzling question why there are so few detrital zircons compared to Paleozoic detrital micas in sediments of the Cayetano Formation.

Late Ordovician (~470 – 440 Ma) ages have also been found in the megacrystics Esperanza granitoids of the Acatlán Complex of southern Mexico (Nance and others, 2006), and Ordovician and Silurian detrital zircon ages (~460 – 435 Ma) reported from the Sierra Madre terrane in central Mexico are interpreted to have been sourced in the Acatlán Complex (Nance and others, 2007). Our single early Devonian age (~398 Ma) is similar to detrital zircon ages reported for the Cosoltepec Formation (410 – 365 Ma) in the Acatlán Complex (Nance and others, 2006).

Finally, we have to admit that a North American source for at least some of the San Cayetano zircons cannot be entirely ruled out on the basis of the available age data alone. However, the paleogeographic reconstructions of Pszczółkowski (1999, his Figs. 15-17) and Pindell and others (2006, reproduced here as our fig. 7) show that the San Cayetano basin must have been situated much closer to northern South America and Yucatan than to Florida and thus make a North American source less likely.

CONCLUSIONS AND GEODYNAMIC IMPLICATIONS

It is still uncertain when, exactly, deposition of the San Cayetano Formation began. Most authors assume deposition began in the early Jurassic, but, as recently discussed by Iturralde-Vinent (2003, 2006; see also Pszczółkowski, 1978, 1999), the early Jurassic date is uncertain. This paleogeography allowed clastic material to be transported over considerable distances in river systems (or marine pathways if the stretched areas were already submerged). These discharged their detritus into the rift valleys, which gradually evolved into shallow marine basins during the Oxfordian and eventually into a passive margin (Pszczółkowski, 1999).

If this paleogeographic reconstruction is correct, it is likely that most of the far-traveled zircons in the San Cayetano sediments were derived from sources in northern Venezuela/Colombia and the Yucatán Peninsula. However, as seen from the reconstruction in figure 7, we do not completely exclude input from other areas such as North America. When the Proto-Caribbean seaway opened in the Late Jurassic (Oxfordian-Tithonian), the San Cayetano sediments, as part of the Guaniguanico Terrane and basin, were already separated from South America (Pszczółkowski, 1987; Iturralde-Vinent and MacPhee, 1999; Iturralde-Vinent, 2006 and references therein) and, therefore, could no longer receive ancient clastic material from northern South America.

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