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Geochronology and geochemistry of the Parashi granitoid, NE Colombia: Tectonic implication of short-lived Early Eocene plutonism along the SE Caribbean margin





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ABSTRACT

The Parashi granitoid of northeasternmost Colombia intrudes the Upper Cretaceous to Lower Paleocene accretionary complex formed by the collision of the Caribbean arc and the continental margin of South America. This granitoid presently separated of the continental margin includes a major quartzdiorite body with andesite to dacite dikes and mafic enclaves. Zircon U-Pb LA-MC-ICP-MS and K-Ar geochronology on the quartzdiorite and the dikes suggest that crystallization extended from ca. 47 to 51 Ma. Major and trace elements are characterized by a medium-K, immature continental arc signature and high Al₂O₃, Na₂O and Ba-Sr contents. Initial ⁸⁷Sr/⁸⁶Sr isotopic values range between 0.7050 and 0.7054, with 143 Nd/ 144 Nd = 0.51235–0.51253, ϵ Nd and ϵ Hf values from -0.81 to -4.40 and -4.4 and -5.2. Major and trace element ratios and isotopic modeling suggest that sedimentary and/or quartzofeldspathic crustal sources were mixed with a mafic melt input. The petrotectonic and geological constraints derived from this granitoid suggest that Parashi plutonism records an immature, oblique subduction-zone setting in which the presence of a high-temperature mantle realm and strong plate coupling associated to upper crust subduction caused the partial fusion of a previously tectonically underplated mafic crust and associated metasediments exposed in the continental margin. The limited temporal expression of this magmatism and the transition to a regional magmatic hiatus are related to a subsequent change to strongly and slow oblique tectonics in the Caribbean-South America plate interactions and the underflow of a relatively thick slab of Caribbean oceanic crust.

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1. Introduction

Plate tectonic reconstructions have suggested that after the Late Cretaceous to Paleocene collision of the Caribbean arc and oceanic plateau with northwestern South America (Pindell, 1993; Kerr et al., 1997; Spikings et al., 2001; Pindell et al., 2005; Luzieux et al., 2006; Vallejo et al., 2006; Weber et al., 2009, 2010a; Cardona et al., 2010; 2011a; van der Lelij et al., 2010; Villagómez et al., 2011), plate convergence in northwestern South America shifted to an oblique and transcurrent system (Pindell et al., 1998; Müeller et al., 1999; Montes et al., 2005; Pindell and Keenan, 2009), and the

* Corresponding author. E-mail address: agcardonamo@unal.edu.co (A. Cardona). convergence between North and South America (Pindell et al., 1998, 2005; Müeller et al., 1999) forced the Caribbean plate to be subducted under the South American continent.

Despite the major effects that such tectonic changes must leave within the upper continental plate, the geological record associated to this transition is not fully constrained (Montes et al., 2010; Bayona et al., 2011, 2012; Cardona et al., 2011a,b; Villagómez et al., 2011).

It's well known that the spatio-temporal and compositional features of magmatic rocks, when integrated with other geological information are sensible recorders of past tectonic setting and provide a robust limit to the timing of successive tectonic events (Pearce et al., 1984, 1996; Barbarin, 1999).

We present field and petrographic relationships together with U–Pb LA-MC-ICP-MS zircon and K–Ar geochronology, whole-rock

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geochemistry, Sr-Nd-Hf isotopes, from the Parashi Stock in the northeasternmost Caribbean region of Colombia (Fig. 1a, b). The temporal and petrotectonic character of this plutonism isolated from the Andean mountain as a consequence of subquent dispersion tectonics, record the early evolution of a new subduction system formed on the northwestern continental margin of South America due to oblique convergence of the Caribbean oceanic plate. The relatively short duration of the Eocene magmatism is also a major consequences of the changing nature of the convergence relation in the continental margin, that shift towards a more oblique subduction that limit magma generation in the arc system, and is responsible for block translation and deformation along the margin (Montes et al., 2005, 2010; Bayona et al., 2012).

2. Geological setting

In northern Colombia the continuous physiography of the Andean chain changes to a series of discontinuous uplifted massifs surrounded by thick Cenozoic basins that extend into the Caribbean Sea (Fig. 1a). Ranges or Serranías, in the northernmost Guajira Peninsula consist of isolated massifs surrounded by broader plains (MacDonald, 1964; Lockwood, 1965; Alvarez, 1967).

Palinspastic restoration suggests that these Serranías remained attached to the northernmost extension of the pre-Andes uplift that included the southwestern Santa Marta Massif and the Central Cordillera of the Andes at least until the Palaeogene when the front of the Caribbean plate collided with the continental margin (Alvarez, 1967; Montes et al., 2005, 2010; Pindell et al., 2005). The post-Eocene dextral strike-slip dominated movement of the Caribbean plate and the northern displacement of the northern Andean block caused the formation of several pull-apart basins in the northern South America plate, and caused the displacement and isolation of different crustal domains in northern Colombia (Pindell, 1993; Macellari, 1995; Trenkamp et al., 2002; Montes et al., 2010; Cardona et al., 2011a,b).

Geologically, the Guajira Serranías are composed of three lithostratigraphic belts that can be correlated with the adjacent Santa Marta region and the Colombian Andean Cordilleras (Alvarez, 1967, Fig. 1b). From southeast to northwest they include: (1) A weakly deformed Mesozoic volcano-sedimentary units correlatable with coeval units exposed in the southeastern Santa Marta massif foothills and the Perija range (MacDonald, 1964; Villamil, 1999). (2) A composite Grenvillian and Late Palaeozoic to Triassic metamorphic domain (Alvarez, 1967; Cordani et al., 2005; Cardona-Molina et al., 2006; Weber et al., 2010) which is intruded by Upper Jurassic plutons similar to the parautochthonous basement of the Andes (Aspden et al., 1987; Cardona-Molina et al., 2006), (3) Upper greenschist facies metamorphosed Cretaceous volcanosedimentary units with intercalated mafic-ultramafic plutonic rocks (Alvarez, 1967; Weber et al., 2007, 2010), together with suprasubduction-zone mafic and ultramafic rocks that crop out as an isolated remnant in the coastal region, and represent a remnant of the colliding Caribbean intra-oceanic arc (Alvarez, 1967; Weber et al., 2009). This belt represents a collage of mixed intra-oceanic arc and passive margin sediments, metamorphosed due to the collision of the Caribbean plate with the South American continent after ca. 76 Ma and before the intrusion of the Eocene igneous rocks (Weber et al., 2009, 2010).

A quartzdiorite stock (Parashi Stock) and a series of andesite to dacite dikes, which are the object of this contribution, intrude the Upper Cretaceous to Palaeocene collision-related greenschist facies rocks of the northern Guajira region (Lockwood, 1965; Weber et al., 2010). Previous K–Ar hornblende ages presented by Lockwood (1965) indicate that this magmatism is Early Eocene in age. However due to the nature of this geochronological method, which may record cooling after magmatic crystallization or may not allow to discriminate between Ar in excess or the presence of scarce radiogenic Ar – therefore this granitoid deserve a more appropriate geochronological constraints such as the new U–Pb age single crystal ages which are presented here.

Continental plutonic rocks of similar age have also been found in the adjacent Santa Marta Massif and the Central Cordillera of Colombia and record a margin scale convergent margin for the Eocene (Tschanz et al., 1974; Aspden et al., 1987; Cardona et al., 2011a; Bayona et al., 2012).

2.1. Geology and petrography of the Parashi Stock

The Parashi Stock (Lockwood, 1965) is exposed in the Jarara Serranía of the Guajira Peninsula (Fig. 2). The plutonic body covers an



Fig. 1. A. Digital elevation model of northern Colombia, Caribbean region. SM = Santa Marta Massif, GS = Guajira Serranías. B. Geological map of the Guajira Peninsula, northern Colombia, including the Parashi granitoid (modified from Lockwood (1965) and Gómez et al. (2007)).



Fig. 2. Geological map of the Parashi Stock and its host rocks modified from Lockwood (1965) and Gómez et al. (2007).

area of \sim 56 km² in and elongated SW–NE direction and comprises quartzdiorite to granodiorite facies. Our petrographic observations indicates that this pluton is medium grain equigranular, composed of quartz (10-30%), zoned plagioclase (30-45%), biotite (5-20%) and hornblende (20-65%). Accessory minerals include apatite, zircon and titanite. Plagioclase is usually oscillatoy zoned, commonly with discontinuous patterns, whereas some biotite crystals are rimmed by amphibole. Porphyritic dikes intrude the quartzdiorite as well as the greenschist Upper Cretaceous host rocks, following a predominant N-NE trend (Fig. 2). Compositionally, these dikes are andesites and dacites. Rounded to weakly angular mafic enclaves (<30 cm in size) are widespread within the quartzdiorite. The enclaves are characterized by a porphyritic texture and are composed of quartz, plagioclase, hornblende and biotite. Poikilitic hornblende with inclusions of plagioclase, biotite, apatite and zircon is very common. They varied in shape from rounded to irregular with diffuse contacts. Feldspar crystal growth along the contact between the host rocks and the enclave suggesting that both were in magmatic stages, whereas in some cases rim of finner grain felsic material surround the enclaves which reflect rapid cooling.

Contacts with the Cretaceous metamorphic rocks of the Etpana Formation (Fig. 2) are intrusive, as seen from cross-cutting relationships as well as from a well-defined ~50 m thermal aureole and numerous veins in the host rocks (Lockwood, 1965; Martínez and Zuluaga, 2010). The formation of quartz-biotite-oligoclase after sericite-chlorite-quartz and albite, as well as the presence of andalusite in the thermal aureole, suggests that the thermal metamorphism may have attained the amphibolite horn-fels facies (Lockwood, 1965; Martínez and Zuluaga, 2010; Weber et al., 2010).

The pluton and the dikes are unconformably overlain by Oligo-Miocene conglomerates and carbonates (Lockwood, 1965; Zapata et al., 2010), which together with U–Th/He apatite and zircon thermochronological constraints suggests that this granitoid had been exhumed and exposed at the surface by the Early-Middle Oligocene (Cardona et al., 2011b).

3. Analytical methods

3.1U/Pb LA-MC-ICP-MS

U/Pb analyses were carried at the Arizona LASERCHRON laboratory following the procedures described by Gehrels et al. (2008). Results are included in Table 1. Zircon crystals were analyzed in

Table 1					
U-Pb zircon	geochronological	data from	the	Parashi	Stock.

Analysis	U (ppm)	²⁰⁶ Pb ²⁰⁴ Pb	U/Th	²⁰⁶ Pb* ²⁰⁷ Pb*	± (%)	²⁰⁷ Pb* ²³⁵ U*	± (%)	²⁰⁶ Pb* ²³⁸ U	± (%)	Error corr.	²⁰⁶ Pb* ²³⁸ U*	\pm (Ma)	²⁰⁷ Pb* ²³⁵ U	\pm (Ma)	²⁰⁶ Pb* ²⁰⁷ Pb*	\pm (Ma)	Selected age (Ma)	\pm (Ma)
CM-3-7-1	108	416	2.3	11.7554	24.1	0.1039	24.1	0.0089	1.5	0.06	56.8	0.8	100.4	23.0	1317.1	473.2	56.8	0.8
CM-3-7-2	145	650	1.8	18.4403	29.9	0.0587	30.2	0.0078	3.9	0.13	50.4	1.9	57.9	17.0	380.5	686.3	50.4	1.9
CM-3-7-3	520	21314	1.9	4.4450	1.4	19.0875	2.1	0.6154	1.6	0.75	3091.4	39.5	3046.2	20.7	3016.6	22.8	3016.6	22.8
CM-3-7-4	144	1030	2.2	14.4927	16.5	0.0760	16.6	0.0080	1.6	0.10	51.3	0.8	74.4	11.9	898.7	342.6	51.3	0.8
CM-3-7-5	91	872	2.5	11.9382	30.9	0.0963	31.5	0.0083	6.2	0.20	53.5	3.3	93.4	28.1	1287.1	617.0	53.5	3.3
CM-3-7-6	120	437	2.1	17.7552	13.0	0.0613	13.8	0.0079	4.9	0.35	50.7	2.5	60.4	8.1	465.0	288.0	50.7	2.5
CM-3-7-7	93	322	2.0	23.9107	25.1	0.0456	25.4	0.0079	3.3	0.13	50.7	1.7	45.2	11.2	-237.3	643.0	50.7	1.7
CM-3-7-8	56	416	1.6	24.0757	79.1	0.0435	79.3	0.0076	5.0	0.06	48.8	2.4	43.2	33.6	-254.6	2426.0	48.8	2.4
CM-3-7-9	163	336	2.1	24.8731	19.1	0.0439	19.3	0.0079	2.5	0.13	50.9	1.3	43.7	8.2	-337.8	495.4	50.9	1.3
CM-3-7-11	495	641	2.1	11.4381	21.2	0.1154	21.4	0.0096	3.0	0.14	61.4	1.8	110.9	22.4	1369.9	411.9	61.4	1.8
CM-3-7-11A	116	535	1.6	18.3379	31.9	0.0632	32.1	0.0084	3.2	0.10	53.9	1.7	62.2	19.4	393.1	733.6	53.9	1.7
CM-3-7-12	174	1138	2.2	17.9897	15.3	0.0616	15.6	0.0080	2.8	0.18	51.6	1.5	60.7	9.2	435.9	342.5	51.6	1.5
CM-3-7-13	76	326	2.7	24.9379	50.0	0.0476	50.2	0.0086	5.0	0.10	55.2	2.8	47.2	23.1	-344.5	1363.6	55.2	2.8
CM-3-7-14	87	549	1.8	21.9396	82.1	0.0515	82.2	0.0082	2.0	0.02	52.6	1.0	51.0	40.9	-24.5	2485.7	52.6	1.0
CM-3-7-15	191	1320	1.7	26.9449	51.2	0.0390	51.7	0.0076	7.3	0.14	49.0	3.6	38.9	19.7	-548.3	1460.3	49.0	3.6
CM-3-7-16	80	549	2.3	18.6389	37.3	0.0554	38.5	0.0075	9.8	0.25	48.1	4.7	54.8	20.6	356.4	869.2	48.1	4.7
CM-3-7-17	96	482	2.4	27.8690	222.0	0.0401	222.1	0.0081	7.0	0.03	52.0	3.6	39.9	87.1	-639.9	0.0	52.0	3.6
CM-3-7-18	90	459	1.7	20.0405	37.3	0.0522	37.4	0.0076	3.5	0.09	48.7	1.7	51.7	18.9	190.3	894.9	48.7	1.7
CM-3-7-19	108	791	2.1	24.4527	26.0	0.0430	26.2	0.0076	3.6	0.14	49.0	1.8	42.7	11.0	-294.1	671.9	49.0	1.8
CM-3-7-20	131	1036	1.6	29.8843	29.5	0.0355	29.5	0.0077	1.9	0.06	49.4	0.9	35.4	10.3	-835.3	857.1	49.4	0.9
CM-3-7-21	70	349	2.1	55.3561	69.7	0.0193	70.0	0.0078	7.0	0.10	49.8	3.5	19.4	13.5	-3109.8	294.3	49.8	3.5
CM-3-7-22	167	935	1.5	26.4997	39.7	0.0410	39.8	0.0079	2.6	0.07	50.6	1.3	40.8	15.9	-503.7	1093.8	50.6	1.3
CM-3-7-23	101	263	1.9	31.9176	33.2	0.0326	33.8	0.0076	6.6	0.20	48.5	3.2	32.6	10.9	-1027.5	1010.7	48.5	3.2
CM-3-7-24	98	539	2.4	23.6527	55.7	0.0431	56.5	0.0074	9.6	0.17	47.5	4.5	42.9	23.7	-209.9	1506.7	47.5	4.5
CM-3-7-25	105	734	2.0	33.8257	47.1	0.0321	47.7	0.0079	7.6	0.16	50.5	3.8	32.0	15.0	-1204.2	1530.3	50.5	3.8
CM-3-7-25A	118	680	2.0	21.8294	34.6	0.0472	34.9	0.0075	4.4	0.13	48.0	2.1	46.8	16.0	-12.4	859.0	48.0	2.1
CM-3-7-26	150	452	2.0	23.2599	23.0	0.0479	23.1	0.0081	2.4	0.10	51.9	1.2	47.5	10.7	-168.1	579.4	51.9	1.2
CM-3-7-27	110	524	2.5	22.0255	41.6	0.0504	41.7	0.0081	3.2	0.08	51.7	1.7	49.9	20.3	-34.1	1048.9	51.7	1.7
CM-3-7-28	112	280	2.0	30.3131	18.4	0.0434	18.5	0.0095	1.9	0.10	61.2	1.1	43.1	7.8	-876.2	535.6	61.2	1.1
CM-3-7-29	299	1788	1.6	19.8899	13.5	0.0612	13.8	0.0088	3.3	0.23	56.7	1.8	60.3	8.1	207.8	313.2	56.7	1.8
CM-3-7-30	118	1009	2.0	31.7086	96.7	0.0336	96.9	0.0077	5.6	0.06	49.6	2.8	33.5	31.9	-1008.0	1875.4	49.6	2.8
CM-3-7-30A	91	449	2.1	36.0544	57.9	0.0287	58.1	0.0075	3.8	0.07	48.2	1.8	28.7	16.5	-1406.9	2028.2	48.2	1.8
CM-3-7-30AA	92	468	2.0	18.3579	26.7	0.0595	27.2	0.0079	5.3	0.20	50.9	2.7	58.7	15.5	390.6	608.6	50.9	2.7
CM-3-7-30AAA	158	326	2.2	16.0734	18.4	0.0662	18.6	0.0077	2.1	0.12	49.6	1.1	65.1	11.7	681.5	396.9	49.6	1.1
CM-3-7-31	175	1008	2.0	26.5105	18.8	0.0411	19.0	0.0079	2.5	0.13	50.8	1.2	40.9	7.6	-504.8	505.7	50.8	1.2
CM-3-2-1	249	1287	2.62	29.739	24.4	0.0362	24.602	0.0078	3	0.13	50.191	1.62	36.14556	8.74	-821.4	703.3	50.191	1.62
CM-3-2-2	124	946.5	2.52	22.363	34.1	0.0452	34.097	0.0073	1	0.04	47.113	0.57	44.91196	15	-71.09	854.1	47.113	0.57
CM-3-2-11	74.7	7616	5.7	14.228	2.71	1.399	3.1005	0.1444	2	0.48	869.3	12.2	888.4967	18.4	936.58	55.66	936.58	55.7
CM-3-2-10	92.4	11280	4.05	14.581	1.34	1.4151	2.3834	0.1496	2	0.83	898.98	16.5	895.2994	14.2	886.23	27.73	886.23	27.7
CM-3-2-12	233	1088	2.16	28.27	20	0.0358	20.043	0.0073	2	0.08	47.116	0.77	35.69271	7.03	-679.2	556.6	47.116	0.77
CM-3-2-13	322	6204	3.77	20.255	1.9	0.2397	3.2543	0.0352	3	0.81	223.11	5.79	218.1911	6.39	165.42	44.46	223.11	5.79
CM-3-2-14	126	435	2.13	22.538	38.5	0.0435	39.156	0.0071	7	0.18	45.627	3.29	43.18973	16.6	-90.14	975.3	45.627	3.29
CM-3-2-15	167	859.5	1.8	35.436	33.2	0.0285	33.453	0.0073	4	0.12	47.084	1.83	28.55684	9.42	-1351	1090	47.084	1.83
CM-3-2-16	161	631.5	1.95	39.275	37.8	0.029	38.405	0.0083	7	0.18	52.99	3.56	29.00366	11	-1695	1354	52.99	3.56
CM-3-2-17	248	1169	1.59	30.923	25.8	0.0323	25.968	0.0073	3	0.12	46.57	1.49	32.30544	8.26	-934.1	762.7	46.57	1.49
CM-3-2-18	219	1052	1.52	31.487	28.6	0.0318	28.965	0.0073	4	0.15	46.657	2.06	31.795	9.07	-987.2	859.9	46.657	2.06
CM-3-2-19	160	921	1.46	40.064	41.5	0.0256	41.542	0.0074	3	0.06	47.787	1.26	25.67339	10.5	-1764	1520	47.787	1.26
CM-3-2-20	160	835.5	1.66	31.868	25.9	0.0313	26.245	0.0072	4	0.16	46.525	2	31.33219	8.1	-1023	781.6	46.525	2
CM-3-2-21	211	1008	1.69	28.593	19	0.0392	19.043	0.0081	1	0.06	52.203	0.58	39.05058	7.3	-710.7	532.9	52.203	0.58
CM-3-2-22	287	991.5	1.32	19.814	32.3	0.0516	32.485	0.0074	4	0.11	47.648	1.75	51.11154	16.2	216.67	765	47.648	1.75
CM-3-2-23	103	558	2.02	35.683	54.7	0.0282	55.057	0.0073	7	0.12	46.861	3.04	28.22927	15.3	-1373	1881	46.861	3.04

CM-3-2-24	214	1133	1.46	27.079	17.4	0.0391	17.704	0.0077	3	0.17	49.311	1.5	38.94379	6.76	-561.6	473.1	49.311	1.5
CM-3-2-27	133	528	1.8	22.478	32.6	0.0438	32.727	0.0071	3	0.08	45.89	1.21	43.54958	14	-83.54	817.7	45.89	1.21
CM-3-2-28	145	708	1.53	32.353	25.2	0.0316	25.301	0.0074	3	0.11	47.619	1.27	31.58726	7.87	-1068	766.7	47.619	1.27
CM-3-11-1	946	6864	3.2	20.1565	3.3	0.1566	3.7	0.0229	1.7	0.45	145.9	2.4	147.7	5.1	176.8	76.6	145.9	2.4
CM-3-11-3	410	2698	2.5	20.2551	6.1	0.1240	6.6	0.0182	2.6	0.39	116.3	2.9	118.7	7.4	165.4	142.5	116.3	2.9
CM-3-11-4	721	4878	3.2	19.5735	4.2	0.1242	5.0	0.0176	2.8	0.56	112.6	3.1	118.8	5.6	244.8	95.8	112.6	3.1
CM-3-11-5	405	2415	2.0	18.9337	6.0	0.1852	6.0	0.0254	1.0	0.17	161.9	1.6	172.5	9.6	320.9	135.4	161.9	1.6
CM-3-11-6	617	2391	3.2	20.6321	5.1	0.0626	5.2	0.0094	1.0	0.19	60.1	0.6	61.6	3.1	122.2	120.1	60.1	0.6
CM-3-11-7	1012	2155	2.7	20.6458	32.2	0.0515	32.2	0.0077	1.0	0.03	49.6	0.5	51.0	16.0	120.6	777.3	49.6	0.5
CM-3-11-8	373	3045	0.8	19.5780	3.7	0.2266	4.1	0.0322	1.6	0.39	204.1	3.2	207.4	7.6	244.3	86.4	204.1	3.2
CM-3-11-9	194	1981	1.0	19.3600	5.3	0.2831	5.4	0.0397	1.1	0.20	251.3	2.6	253.1	12.0	270.1	120.8	251.3	2.6
CM-3-11-10	116	374	2.4	21.1275	19.5	0.0530	20.0	0.0081	4.4	0.22	52.2	2.3	52.5	10.2	66.0	467.6	52.2	2.3
CM-3-11-10A	151	505	2.0	21.2172	14.4	0.0517	14.9	0.0080	3.7	0.25	51.0	1.9	51.1	7.4	55.9	346.0	51.0	1.9
CM-3-11-11	1155	3781	3.1	20.0873	4.2	0.0617	8.6	0.0090	7.5	0.87	57.7	4.3	60.8	5.1	184.8	97.3	57.7	4.3
CM-3-11-13	1256	969	1.8	21.0558	6.2	0.0492	6.3	0.0075	1.0	0.16	48.2	0.5	48.7	3.0	74.1	147.5	48.2	0.5
CM-3-11-14	375	3032	2.7	14.3056	2.7	0.3569	8.3	0.0370	7.9	0.95	234.4	18.1	309.9	22.2	925.5	55.3	234.4	18.1
CM-3-11-15	797	1898	1.7	20.2650	11.7	0.0592	11.9	0.0087	1.8	0.15	55.9	1.0	58.4	6.7	164.3	274.8	55.9	1.0
CM-3-11-15A	664	7399	5.1	15.1015	2.0	0.1849	4.3	0.0202	3.8	0.89	129.2	4.9	172.2	6.9	813.3	41.9	129.2	4.9
CM-3-11-15AA	140	8030	3.4	13.7118	2.0	1.2547	3.1	0.1248	2.4	0.76	758.0	16.9	825.5	17.5	1012.0	40.5	758.0	16.9
CM-3-11-16	659	1376	2.7	16.7565	6.0	0.1141	6.7	0.0139	2.9	0.43	88.8	2.5	109.7	7.0	591.9	131.1	88.8	2.5
CM-3-11-17	75	4647	3.5	14.1887	3.4	1.5327	3.8	0.1577	1.6	0.43	944.1	14.3	943.6	23.4	942.3	70.5	944.1	14.3
CM-3-11-18	1383	689	6.6	19.6310	13.5	0.0557	13.5	0.0079	1.3	0.10	51.0	0.7	55.1	7.3	238.1	311.9	51.0	0.7
CM-3-11-19	2814	16468	17.8	20.7970	4.2	0.1204	7.3	0.0182	6.0	0.82	116.0	6.9	115.4	8.0	103.4	100.1	116.0	6.9
CM-3-11-19A	1961	15887	9.1	20.6852	6.1	0.1648	6.3	0.0247	1.4	0.22	157.4	2.1	154.9	9.0	116.1	144.3	157.4	2.1
CM-3-11-20	1373	1812	2.2	21.1815	5.7	0.0494	5.8	0.0076	1.0	0.17	48.7	0.5	49.0	2.8	59.9	137.1	48.7	0.5
CM-3-11-20A	827	1600	3.9	21.1622	8.0	0.0494	8.1	0.0076	1.0	0.12	48.7	0.5	49.0	3.9	62.1	191.3	48.7	0.5
CM-3-11-20AA	645	1268	4.1	21.0658	7.4	0.0491	7.7	0.0075	2.3	0.29	48.2	1.1	48.7	3.7	72.9	175.6	48.2	1.1
CM-3-11-20AAA	683	1166	3.3	20.6725	11.7	0.0507	11.7	0.0076	1.0	0.09	48.8	0.5	50.2	5.7	117.6	275.5	48.8	0.5
CM-3-11-21	88	332	2.9	32.1194	54.9	0.0321	55.2	0.0075	4.8	0.09	48.1	2.3	32.1	17.4	-1046.4	1758.4	48.1	2.3
CM-3-11-22	112	319	2.7	28.0850	36.7	0.0369	37.6	0.0075	8.0	0.21	48.3	3.8	36.8	13.6	-661.1	1040.7	48.3	3.8
CM-3-11-23	128	354	2.2	20.5975	45.9	0.0501	46.0	0.0075	2.7	0.06	48.1	1.3	49.7	22.3	126.1	1136.7	48.1	1.3
CM-3-11-25	131	6966	2.9	13.5399	1.4	1.8381	1.7	0.1805	1.0	0.58	1069.7	9.9	1059.2	11.3	1037.5	28.1	1037.5	28.1

Table 2
K-Ar amphibole and biotite from the Parashi Stock.

Sample	North	West	Mineral	Rock	K%	K error (%)	Ar ⁴⁰ Rad ccSTP/g (*10 ⁻⁶)	Ar ⁴⁰ Atm (%)	Age (Ma)
CM-5-7	12° 13′ 43″	71° 41′ 23.9″	Biotite	Quartdiorite	72,101	0.5	12.8	37.89	45.5
CM-4-16	12° 13′ 46″	71° 41′ 20.9″	Amphibole	Quartdiorite	0.5869	1,113	1.1	68.56	47.7

polished epoxy grain mounts with a Micromass IsoProbe Multicollector mass spectrometer (MC-ICP-MS) equipped with nine Faraday collectors, an axial Daly collector and four ion-counting channels. The IsoProbe is equipped with an ArF Excimer laser ablation system, which has an emission wavelength of 193 nm. The collector configuration allows measurement of ²⁰⁴Pb in the ioncounting channel while ²⁰⁶Pb, ²⁰⁷Pb, ²⁰⁸Pb, ²³²Th and ²³⁸U are simultaneously measured with Faraday detectors. All analyses were conducted in static mode with a laser beam diameter of 35–50 µm, operated with an output energy of ~32 mJ (at 23 kV) and a pulse rate of 9 Hz. Each analysis consisted of one 20-s integration on

Table 3	
Geochemical and Sr-Nd isotope d	lata from the Parashi Stock.

peaks with no laser firing and 20 one-second integrations on peaks with the laser firing. Hg contribution to the ²⁰⁴Pb mass position was removed by subtracting on-peak background values. Inter-element fractionation was monitored by analysing an in-house zircon standard, which has a concordant TIMS age of 564 ± 4 Ma (2σ) (Gehrels, unpublished data). This standard was analyzed once for every five unknowns in detrital grains. Uranium and Th concentrations were monitored by analysing a standard (NIST 610 Glass) with ~500 ppm Th and U. The lead isotopic ratios were corrected for common Pb, using the measured ²⁰⁴Pb, assuming an initial Pb composition according to Stacey and Kramers (1975) and respective

	CM3-11p	CM3-7p	CM4-16	CM5-19	CM5-21	CM5-22g	CM5-22x	CM4-14
	Porphyritic dike	Porphyritic dike	Quartzdiorite stock	Quartzdiorite stock	Quartzdiorite stock	Quartzdiorite stock	Mafic enclave	Mafic enclave
SiO ₂	67.23	67.45	58.89	62.46	62.13	64.11	54.73	55.06
TiO ₂	0.25	0.29	0.66	0.57	0.59	0.51	0.85	0.83
Al_2O_3	15.97	16.8	16.71	17.21	17.36	16.84	17.66	16.01
Fe ₂ O ₃	2.36	2.78	6.26	5.15	5.19	4.72	8.27	8.11
MnO	0.05	0.05	0.14	0.12	0.12	0.1	0.23	0.19
MgO	1.46	1.32	3.06	1.95	2.09	1.74	3.73	5.34
CaO	2.91	4.05	5.5	5.48	5.71	5.11	7.5	7.79
Na ₂ O	4.84	4.5	3.35	3.84	3.76	3.81	4.21	3.6
K ₂ O	2.25	1.91	2.49	1.96	1.85	2.03	1.57	0.93
P ₂ O ₅	0.1	0.1	0.22	0.24	0.23	0.2	0.25	0.17
LÕI	2.2	0.4	2.5	0.7	0.8	0.5	0.7	1.8
К	18678	15855	20670	16271	15357	16852	13033	7720
Ba	1365.9	1157.3	1046.1	1173.3	1128.3	1179.7	820.4	538.4
Rb	55.7	48.4	50.3	49.1	49.4	47.7	25.8	18.8
Sr	809.4	752.7	525.1	624.9	634.2	579.3	490	443.3
Cs	1.3	0.9	0.4	1.2	1.1	1.3	0.7	0.4
Ga	19.4	19.9	17.7	18.9	19.6	18.6	19.2	17.9
TI	0	01	0	02	01	02	01	0
Ta	03	0.4	04	0.5	0.81	0.5	0.7	02
Nb	52	59	5.5	7.6	14	5.5	9.8	43
Hf	3	31	3.6	43	4	37	3.9	26
7r	86.6	109.1	97.6	132.3	132.1	116.9	114	72.9
Ti	1499	1739	3957	3417	3537	3057	5096	4976
v	11	17.55	21.7	21.5	22.2	15 5	377	21.5
Th	28	36	5.2	52	ΔΔ ΔΔ	4.8	56	46
II	1.0	11	1.4	17	1.6	1.5	3	1.0
Ni	12504	1196 5	508.9	1069	104	1.5	1090 3	524.5
Co	5.8	100.5	117	78	103.2	68	11.2	10.7
Sc	5	5	11.7	0	105.2	8	11.2 22	15.2 27
V	16	16	120	9	96	0 77	1/1	27
v Cu	40	40	117	51	11	55	141	241
Dh	10.8	9	2.4	0.9	1.4	0.0	11.2	20.2
PD	2.4	0.5	2.4	0.0	1	0.9	1.1	2.0
La	11	26.2	20.5	20.0	22 47 4	21.2 41.6	61.8	246
De De	23.2	4.09	42.7	54.1	47.4 5.40	41.0	7 95	4 10
PI Nd	2.05	4.06	5 20.2	0.07	2.49	4.01	7.00	4.19
INU Smr	10.5	10.0	20.2	25	25.2	10.0	55.7	17.7
5111	2.1	3.1	4.3	4.2	4.5	3.4	0.8	3.8 1.20
EU	0.61	0.85	1.10	1.28	1.23	0.96	2.02	1.28
Ga	1.85	2.18	3.4	3.65	3./5	2.72	6.19	3.7
ID	0.3	0.36	0.63	0.59	0.66	0.47	1.04	0.61
Dy	1.69	2.23	3.86	3.52	3.85	2.57	6.14	3.79
Ho	0.31	0.4	0.7	0.68	0.71	0.52	1.23	0.72
Er	0.97	1.13	2.07	2.11	2.15	1.59	3.//	2.19
Im	0.18	0.18	0.34	0.34	0.33	0.23	0.57	0.32
Yb	1.04	1.09	2.05	2.24	2.01	1.78	3.82	2
°'Sr/°°Sr	0.70546	nd	0.70538	0.70543	nd	0.70523	nd	0.70505
¹⁴³ Nd/ ¹⁴⁴ Nd	0.51235	nd	0.51244	0.51237	nd	0.51241	nd	0.51253
North	12° 13′ 49″	12° 13′ 49″	12° 13' 46"	12° 13′ 43″	12° 13′ 43″	12° 13′ 43″	12° 13′ 43″	12° 13′ 46″
West	71° 40′ 13.2″	71° 40′ 13.2″	71° 41' 20.9"	71° 41′ 23.9″	71° 41′ 23.9″	71° 41′ 23.9″	71° 41′ 23.9″	71° 41′ 20.9″

 Table 4

 Hf isotope results from a quartzdiorite samples from the Parashi Stock.

Sample	¹⁷⁶ Hf/ ¹⁷⁷ Hf	2SE	¹⁷⁶ Lu/ ¹⁷⁷ Hf	2SE	¹⁷⁶ Yb/ ¹⁷⁷ Hf	2SE	Age (Ma)	$^{176}\text{Hf}/^{177}\text{Hf}_{(i)}$	$\epsilon H f_{(0)}$	2SE	$\epsilon H f_{(i)}$	2SE
CP7_1	0.282656	0.000026	0.00137	0.00008	0.03330	0.00170	50	0.282655	-4.6	0.9	-3.5	0.9
CP7_2	0.282638	0.000026	0.00088	0.00006	0.02070	0.00170	50	0.282637	-5.2	0.9	-4.1	0.9
CP7_3	0.282661	0.000022	0.00060	0.00001	0.01450	0.00032	50	0.282660	-4.4	0.8	-3.3	0.8
CP7_4	0.282649	0.000023	0.00081	0.00002	0.01960	0.00052	50	0.282648	-4.8	0.8	-3.7	0.8
N12° 13.5	51 W71° 40.760)										

uncertainties of 1.0, 0.3 and 2.0% for $^{206}\text{Pb}/^{204}\text{Pb},\,^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}.$

The uncertainty on the age is determined as the quadratic sum of the weighted mean error plus the total systematic error for the set of analyses. The systematic error, which includes contributions from the standard calibration, age of the calibration standard, composition of common Pb, and U decay constants, is generally $\sim 1-2\%$ (2-sigma). For these samples the systematic errors range from ~ 1.0 to 1.4% for 206 Pb/ 238 U and from ~ 0.8 to 1.1% for 206 Pb/ 207 Pb.

3.1.1. K-Ar geochronology

Biotite and amphibole were analyzed by the K–Ar method at the Geochronological Research Center **of** São Paulo **University** (CPGeo-USP). Two aliquots from the same sample were separated for the K and Ar analysis. Potassium analyses of each pulverized sample were carried out in duplicate, coupled to an ultra-vacuum system. A spike of ³⁸Ar was added and the gas was purified in titanium and copper ovens. Final argon determinations were carried out in a Reynolds-type gas spectrometer. Analytical precision for K is 5% whereas for Ar it is around 0.5%. Decay constants for calculation are after Steiger and Jager (1977). Results are presented in Table 2.

3.2. Geochemistry

Bulk whole-rock chemical analyses of eight samples were performed by inductively coupled plasma-mass spectrometry (ICP-MS) at Acme Analytical Laboratories Ltd. in Vancouver, Canada. A 0.2 g aliquot was placed into a graphite crucible and mixed with 1.5 g of LiBO₂ flux. The crucibles were placed in an oven and heated to 1050 °C for 15 min. The molten sample was dissolved in 5% HNO₃. Calibration standards and reagent blanks were added to the sample sequence. Sample solutions were aspirated into an ICP emission spectrograph (Jarrel Ash Atom Comb 975) for determining major oxides and certain trace elements (Ba, Nb, Ni, Sr, Sc, Y and Zr), and aspirated into an ICP-MS (Perkins–Elmer Elan 6000) for determining trace elements, including rare earth elements. Results are presented in Table 3. The interpretation and analysis of the geochemical database was done using the GCDkit software (Janoušek et al., 2006).

3.2.1. Nd-Sr isotopes

Whole-rock samples were analysed by the Sm–Nd and Rb–Sr methods at the Centre for Geochronological Research of the University of São Paulo (CPGeo-USP). For the Sm–Nd method the



Fig. 3. Concordia and weight average U-Pb ages from the Parashi Stock. a. and b. Quartzdiorite samples and c. Porphyritic dike.

analytical procedures followed Sato et al. (1995). ¹⁴³Nd/¹⁴⁴Nd were obtained using a multi-collector mass spectrometer, with an analytical precision of 0.0014% (2 σ). Experimental error for the ¹⁴⁷Sm/¹⁴⁴Nd ratios is on the order of 0.1%. La Jolla and BCR-1 standards yielded ¹⁴³Nd/¹⁴⁴Nd = 0.511849 ± 0.000025 (1 σ) and 0.512662 ± 0.000027 (1 σ) respectively during the period in which the analyses were performed. eNd values were calculated following De Paolo (1988), and the constants used include ¹⁴³Nd/¹⁴⁴Nd (CHUR) = 0.512638 and ¹⁴⁷Sm/¹⁴⁴Nd (CHUR)₀ = 0.1967. Rb–Sr analyses followed procedures presented by Tassinari et al. (1996). Rb and Sr values were obtained by X-ray fluorescence, and ⁸⁷Sr/⁸⁶Sr ratios were undertaken using a VG-sector mass spectrometer and corrected for isotopic fractionation during thermal ionization using a ⁸⁷Sr/⁸⁶Sr value of 01194. Analytical results are presented in Table 3.

3.3. Hf isotopes in zircon

Hf isotope analyses in zircons were conducted at the Geoanalytical Lab of the Washington State University, using a ThermoFinnigan NeptuneTM MC-ICP-MS equipped with 9 F collectors interfaced with a New WaveTM 213 nm UP Nd-YAG laser. The laser was operated at a pulse of rate 10 Hz and a fluence of 10–12 J/cm2. Laser spot size was 40 μ m. Carrier gas consisted of purified He plus small quantities of N₂ to minimize oxide formation and increase Hf sensitivity. The total Hf signal achieved was between 2 and 6 V. The data were acquired in static mode with 60 one second integrations. Details of analytical procedures and data treatment are after (Vervoort et al., 2004; Dufrane et al., 2007). For the Hf-depleted mantle model ages (Hf_{TDM}) we used ¹⁷⁶Hf/¹⁷⁷Hf and ¹⁷⁶Lu/¹⁷⁷Hf for the individual zircon samples to determine their initial ¹⁷⁶Hf/¹⁷⁷Hf ratios at their crystallization ages. Projection back from zircon crystallization was calculated using a present value of 0.0093 for ¹⁷⁶Lu/¹⁷⁷Hf crust (Amelin et al., 2000; Vervoort and Patchett, 1996). The depleted mantle Hf evolution curve was calculated from present-day depleted mantle values of ¹⁷⁶Hf/¹⁷⁷Hf DM(0) = 0.283225 and ¹⁷⁶Lu/¹⁷⁷Hf DM(0) = 0.038512 (Vervoort and Blichert-Toft, 1999). Analytical results are presented in Table 4.

4. Results

4.1. U-Pb zircon crystallization and K-Ar cooling ages

Two quartzdiorite samples and one porphyritic dike that intrude the quartzdiorite and the host Cretaceous metamorphic rocks were selected for zircon U–Pb geochronological analyses (Fig. 2). Results of the analyses are presented in Table 1.

Tips from 29 zircons from a quartzdiorite (sample CM-3-7) yielded an average age of 50.7 \pm 1.0 Ma (Fig. 3a). Spots on 14 zircons tips from the other quartzdiorite sample (CM-3-2) yielded a mean age of 47.3 \pm 0.9 Ma (Fig. 3b). Three zircon cores showed



Fig. 4. a. A/CNK versus A/NK (Shand, 1943). b. AFM tholeiitic versus calc-alkaline discrimination diagram (Irvine and Baragar, 1971). c. SiO₂ versus K₂O discrimination of magmatic compositional suites (Peccerillo and Taylor, 1976). Green triangles, quartzdiorite; blue circles, porphyritic dikes; black crosses, enclaves.

older Triassic (223 Ma) and Grenvillian ages (936 Ma and 882 Ma). Thirteen zircons from a porphyry andesite (CM-3-11) yielded a concordia age of 49.0 ± 1.0 Ma (Fig. 3c). Older cores (n = 14) include 110-160 Ma, 200-250 Ma, and older Grenvillian zircon ages. These Eocene ages represent the magmatic crystallization ages of the granitoid and porphyritic dikes, and suggest that the dikes are temporally related. The variable inherited zircon ages are similar to detrital zircon ages from the host metasediments and suggest that they could be a magmatic source (Weber et al., 2010).

Two K–Ar ages were obtained for a biotite and an amphibole from quartzdiorite samples CM-5-7-1 and CM-4-16. Analytical and age results are presented in Table 2. Results show that the amphibole age (47.7 ± 5.5 Ma) is older but within error of the biotite age

(45.1 \pm 2.5 Ma). These ages are similar to the amphibole K–Ar age obtained by Lockwood (1965) of 48 \pm 4 Ma that almost overlap also the U–Pb crystallization ages suggest fast cooling after magmatic emplacement.

4.2. Geochemistry

Table 3 presents major and trace elements from eight samples: four of the quartzdiorite, two of the porphyritic dikes, and two of the associated mafic enclaves. Together they indicate a compositional trend from quartzdiorite to diorite, with SiO₂ values ranging from 54.7% to 67.4%. Al₂O₃ varies from 15.97% to 17.66%, and K₂O/Na₂O ratios are low (0.25–0.74). MgO values range from 1.3 to 3.1%



Fig. 5. Harker type bi-variant diagrams from selected elements. Open triangles (quartzdiorite), open circles (porphyritic dikes), black cross (enclaves).

in the quartzdiorite and porphyritic rocks to 3.7–5.3% within the mafic enclaves, whereas the Mg# also varies from 29–40 to 33–42.

All rocks are metaluminous and subalkaline in character (Fig. 4a). Their AFM and SiO₂ versus K₂O trends are characteristic of a medium-K calc-alkaline series (Fig. 4b, c, Irvine and Baragar, 1971). On selected Harker's variation diagrams the samples follow a near-linear trend of decreasing TiO₂, Fe₂O₃, MnO, MgO and CaO with increasing SiO₂ with the dykes and enclaves recording the higher values (Fig. 5).

Chondrite-normalized REE patterns (Fig. 6a) show a well-defined LREE enrichment with (La_N/Yb)_N values of 5.18–11.58 and a weakly inclined HREE trend (Gd_N/Yb_N = 1.21–1.59). The enclaves are less enriched in the LREE than are the other rocks. All samples show a weakly negative to almost absent Eu anomaly with (Eu/Eu)_N varying from 0.92 to 1.00 in the quartzdiorite to 0.95–1.01 in the porphyritic rocks and 1.05–0.96 in the enclaves.

Primitive mantle-normalized multi-element diagrams for the Parashi Stock rocks are characterized by Large Ion Lithophile Elements (LILE) enrichment (Fig. 6b) with notably high Ba and Sr values (538–1366 ppm and 443–809, respectively). Relative depletion in the High Field Strength Elements (HFSE) together with a well-defined Nb and Ti anomaly are also characteristic (Fig. 6b).

4.3. Sr-Nd isotopes

Nd–Sr isotopic data for the analyzed samples are presented in Table 4 and Fig. 7. Initial 87 Sr/ 86 Sr values for the granitoid, the enclaves and the porphyritic rocks ranges between 0.7050 and 0.7054. Initial 143 Nd/ 144 Nd ratios calculated for a crystallization age of ca.

50 Ma are between 0.51235 and 0.51253, and ϵ Nd values vary from -0.81 to -4.40. Nd T_{Dm} model ages range between 0.9 and 1.1 Ga. In general, isotopic values from the enclaves are relatively more juvenile than those from the quartzdiorites and the porphyritic dikes (Fig. 7).

4.4. Hf isotopes

Four zircon crystals were analyzed from a quartdiorite of the Parashi stock (CP7) (results in Table 3). Initial ¹⁷⁶Hf/¹⁷⁷Hf ratios in the quartz-diorite range from 0.282637 to 0.282660, whereas initial ε Hf values are between -4.4 and -5.2. These results are significantly below the CHUR thus suggesting a major radiogenic input (Fig. 8) that can be related to older crust input to the magma.

5. Discussion

5.1. Petrotectonic constraints

The ubiquitous presence of hornblende and apatite together with the metaluminous character (A/(CNK) values < 1.0) are characteristic of an I-type related granitoid.

Fractional crystallization processes within the main quartzdiorite body can be evaluated using Harker type bi-variate plots. TiO₂, Fe₂O₃, MgO, MnO and CaO contents decrease with increasing SiO₂ concentrations (Fig. 5) and can be related to the crystallization of mafic phases, such as amphibole and biotite. However, the weak and inconsistent variations in trace elements, such as Rb, Sr, La, Th and Zr, and the lack of correlation between K and either Rb or Ba



Fig. 6. a. Chondrite normalized REE patterns. b. Primitive mantle normalized multi-element diagram (Sun and McDonough, 1989), Green triangles, quartzdiorite; blue circles, porphyritic dikes; black crosses, enclaves.



Fig. 7. a. Nd vs. ⁸⁷Sr/⁸⁶Sr for the Eocene Parashi Stock: open triangles = quartzdiorite; open circles = porphyritic dikes; black crosses = enclaves. Other fields depicted are from MORB (White and Hofman, 1982) and Caribbean PIA (Jolly et al., 2006), Atlantic sediments (White et al., 1985), Guajira region (Cabo de la Vela) mafic rocks (Weber et al., 2009) and Caribbean plateau (compilation in Thompson et al., 2004). **b** Single binary magma mixing model using two end-members (review in Faure and Mensing, 2005). Field from the Cabo de la Vela gabbroic and basaltic crust is after Weber et al. (2009). Field from the Jurassic is from Ordoñez-Carmona (2001) and Cardona-Molina et al. (2006).



Fig. 8. Hf isotopes in zircon from the analyzed quartzdiorite sample.

(Fig. 9), which are expected to be concentrated in the late magmatic stages as well as the lack of correlation between LREE enrichment and SiO_2 content in the three different rock groups (quartzdiorite, porphyry and diorite) suggest that magma mixing was common during the magmatic evolution (Wilson, 1989; Rollinson, 1993).

The presence of mafic microgranular enclaves hosted by more felsic rocks, biotite crystals rimmed with hornblende, and plagioclase showing discontinuous oscillatory zoning patterns are also features that have been related to compositional and/or temperature modifications due to magma mixing processes (Barbarin and Didier, 1992). Variations in εNd values can be also related to magma mixing from different sources, whereas the weak ⁸⁷Sr/⁸⁶Sr isotope variation reflect the faster Sr isotopic homogenization during mixing (Lesher, 1990).

The variations in isotope ratios, together with the limited correlation between SiO_2 and $({}^{87}Sr/{}^{86}Sr)$ and the MgO vs Nb/U and Ce/Pb relations (Fig. 10) suggest that magmatic assimilation of the host rocks was not important (Faure, 2001; Liu et al., 2009).

Within the major element R1-R2 tectonic discrimination diagram the Parashi Stock samples share similarities with precollisional granitoids formed in convergent margins (Fig. 11a). Whereas trace element relationships, such as the Rb versus Y + Nbgranitoid tectonic discrimination diagram, together with the presence of LILE enrichment when compare with the HFSE and the well-defined Nb and Ti anomalies, also suggest that this plutonism has a subduction related affinity (Figs. 5b and 11b). Similar results are seen within the Th—Yb and Ta/Yb geochemical indices for felsic to intermediate rocks also were this plutonic rock are similar to arcs formed within a continental margin (Fig. 11c).

Brown et al. (1984) have shown that the Nb versus Rb/Zr relationship trace the degree of arc maturity trough time. The analyzed samples are within the primitive continental arc field (Fig. 11d), falling outside the field of intra-oceanic granitoids such as those found in the Caribbean (White et al., 1999; Rojas-Agramonte et al., 2004; Proenza et al., 2006; Marchesi et al., 2007; Wörner et al., 2009) and therefore indirectly confirming that this granitoid is not part of the Caribbean oceanic arcs that accreted against the continental margin during the Late Cretaceous (Weber et al., 2009; Cardona et al., 2011a).

The Parashi Stock presents high Ba and Sr contents (Ba > 500 and Sr > 300), a weakly negative Eu anomaly, the Nb depletion, low Y values in all but one sample (11–22.2), high Sr/Y (12–74), high Al₂O₃ contents (15.9–17.6%) and low K₂O/Na₂O ratios (0.25–0.74) which are characteristics similar to those reported in rocks formed by partial melting of mafic rocks such as adakites or high Ba–Sr granitoids (Martin, 1999; Tarney and Jones, 1994; Quian et al., 2003). In any case the high Ba–Sr signature could be related to an input from subducted sediments (Rüpke et al., 2002; Kilian and Behrmann, 2003).

The relatively higher HREE and Y concentrations, the concaveup REE pattern, low Eu anomaly (Fig. 6a), high Al₂O₃ and Na₂O are more compatible with relatively lower melting in the amphibole than the garnet stability field (Wilson, 1989; Martin, 1999).

As already inferred from the geochemical data and is well known from arc systems, granitoid rocks are commonly formed by the melting of multiple sources. For granitoid rocks in the northern margin of South America these sources ranges from radiogenic South American crust and/or their derived sediments and underplated sediments, accreted Cretaceous mafic oceanic terranes and the mantle wedge.

Comparison with basement rocks in the adjacent South American margin shows that the Sr–Nd isotopic values from the Precambrian to Palaeozoic rocks are more radiogenic in character than the Parashi rocks. Their ε Nd values calculated for 50 Ma range between -6.9 and -16.5, Nd model ages from 1.6 to 2.1 Ga and



Fig. 9. Bivariate plots showing correlations among SiO₂ and selected trace elements and trace element ratios. Green triangles, quartzdiorite; Blue circles, porphyritic dikes; black crosses, enclaves.

⁸⁷Sr/⁸⁶Sr ratios range between >0.7748 (Fig. 7a, Restrepo-Pace et al., 1997; Cordani et al., 2005; Cardona-Molina et al., 2006). Similarly, Eocene to recent sediments from the Barbados and Demarara plain in the Atlantic Ocean and the Orinoco River that may reseamble the composition of sediments that will be subduct to the trench show high εNd values ranging from -11.2 to -14.9 and less variable but radiogenic initial ⁸⁷Sr/⁸⁶Sr ratios between 0.7121 and 0.7219 (White et al., 1985; Goldstein et al., 1997).

The Caribbean plateau, which is representative of the oceanic crust in the Caribbean region, has 87 Sr/ 86 Sr values that vary between radiogenic and nonradiogenic (0.7025–0.7055) and positive ϵ Nd values of 6.5–11 (Kerr et al., 1997; White et al., 1999; Thompson

et al., 2004). Isotopic characteristics of the Cabo de la Vela maficultramafic complex which may be part of the substrate of the Parashi Granitoid and the northern Lesser Antilles (White et al., 1985, Weber et al., 2009), which lack a radiogenic Sr component (ca. 7.03) and also have ε Nd values relatively more radiogenic (4.1– 7.5).

We have modeled single binary magma mixing using two endmembers following the equation presented in Faure and Mensing (2005). Selected end-members for calculations were Atlantic-type sediments for the crustal end-member (White et al., 1985), together with depleted MORB mantle, Caribbean arc or plateau crust (White et al., 1999; Thompson et al., 2004; Weber et al., 2009).



Fig. 10. Selected bivariate plots to test assimilation processes (Faure, 2001; Liu et al., 2009). Green triangles, quartzdiorite; Blue circles, porphyritic dikes; black crosses, enclaves.

Results from the model show a concave pattern characteristic of a two component mixing (Hildreth and Moorbath, 1988; Millar et al., 2001) including a mafic crust mixed with a radiogenic continental basement or a (meta) sedimentary source (Fig. 7b). These sources were melt at relatively shallow conditions as suggested by their HREE patterns that record the presence of amphibole instead of garnet as a residual phase (Peacock et al., 1994; Rapp and Watson, 1995; Kay et al., 2005).

Models that account for melting of mafic crust in subduction settings include slab melting in anomalously hot geothermal conditions (subduction of <5 Ma oceanic plate, subduction initiation and ridge subduction), melting of previously underplated mafic arc

crust or fore-arc subduction erosion transfer to the mantle wedge and flat slab subduction (Martin, 1999; Gutscher et al., 2000; Castillo, 2006; Kay et al., 2005). Within the southern Caribbean margin and the associated Parashi granitoid magmatism the age of the plate, the existence of an underplated mafic arc crust or ridge subduction conditions do not fit with geodynamic conditions during the Eocene (Pindell and Keenan, 2009) including the relatively young nature of the arc. Adakites associated to flat slab subduction are predicted to be formed far (by ca. 300 Km) from the trench (Gutscher et al., 2000), a situation that do not fit the position of the Parashi granitoid and the possibility that the flat slab setting was formed after the early Eocene (Bayona et al., 2012).



Fig. 11. Geochemical tectonic discrimination diagrams. **a.** Multication R1–R2 (Batchelor and Bowden, 1985). **b.** Granitoid discrimination diagram after Pearce et al. (1984). **c.** Intermediate rocks tectonic discrimination diagram (Gortonand and Schandl, 2000), **d.** Nb versus Nb/Zr arc maturity discrimination diagram (Brown et al., 1984). **e.** Low and high ba discrimination diagram (after Tarney and Jones, 1994). Green triangles, quartzdiorite; Blue circles, porphyritic dikes; black crosses, enclaves.



Fig. 12. Tectonic model in two stages. a. Maastrichtian to Palaeocene oblique arc-continent collision, followed by subduction initiation. b. Intrusion of the Eocene Parashi Stock (modified from Pindell et al., 2005; Giunta et al., 2006; Pindell and Keenan, 2009).

We therefore suggest that the mixed mafic and upper plate sourced Early Eocene magmatism that formed the Parashi granitoids was formed during the young stages of the subduction zone were mantle upwelling together with strong plate coupling within the fore-arc promote extensive subduction erosion of the upper plate, causing the melting of the older and previously tectonically underplated arc-continent crust.

5.2. Eocene tectonics

Geophysical and paleogeographical models and Miocene magmatism have argued for Cenozoic subduction of the Caribbean plate under the South American continental margin (Kellog and Bonini, 1982; Toto and Kellogg, 1992; van der Hilst and Mann, 1994; Taboada et al., 2000; Cortés and Angelier, 2005; Pindell et al., 2005; Miller et al., 2009; Mantilla et al., 2013), how those this long term subduction initiate and how is recorded on the upper continental plate along the Cenozoic is not fully resolved (Kellog and Bonini, 1982; Spikings et al., 2001; Vallejo et al., 2006; Bayona et al., 2011; Cardona et al., 2011a,b; Mantilla et al., 2013; Bissig et al., 2013).

Late Cretaceous to Eocene tectonics of northwestern South America was controlled by an oblique interaction of the front of the allochthonous Caribbean oceanic plate with the continental margin (Burke, 1988; Pindell, 1993; Pindell et al., 2005; Spikings et al., 2001, 2005; Vallejo et al., 2006). This culminated in the Late Maastrichtian to Palaeocene collision of island arc fragments and oceanic plateau remnants on the Colombian margin (Fig. 12a, Kerr et al., 1997; Spikings et al., 2001; Pindell et al., 2005; Vallejo et al., 2006; Cardona et al., 2010; Weber et al., 2010; van der Lelij et al., 2010; Villagómez et al., 2011; Wright and Wyld, 2011). Palinspatic restoration of northeastern Colombia, including the Guajira region, evidences for arc-continent collision is recorded by the presence of accreted 90–76 Ma intraoceanic arc fragments and a series of middle to high pressure metamorphic complexes of Late Cretaceous to Palaeocene age (Weber et al., 2007, 2010; Cardona et al., 2010). Following this event, plate tectonic reconstructions have suggest the existence of major changes along the plate boundaries between the Caribbean and North American plates: (1) During the Early Eocene the northwestern front of the Caribbean plate collided with the North American Bahamas platform, and (2) North and South America changed their convergence towards a more orthogonal relation (Pindell et al., 1998; Müeller et al., 1999). This situation made the Caribbean plate to approach in an eastsoutheast direction against South America (Müeller et al., 1999), which probably caused the stress necessary to allow to appropriately establish a new subduction zone along the South American continental margin (Gurnis et al., 2004).

We suggest that the new subduction zone and its associated arc magmatism represented by the Parashi Stock and other correlative plutons in the southeastern Santa Marta region and the Central Cordillera (Tschanz et al., 1974; Cardona et al., 2009; 2011a; Bayona et al., 2012) was active by ca. 65–60 Ma (Figs. 12b and 13). The installation of this arc seems to be diachronous, starting by 65–60 Ma in the Central Cordillera and become younger to the north in Santa Marta (58 Ma) the Guajira region at 50 Ma (Bayona et al., 2012).

5.3. Late Eocene to Oligocene tectonics

Eocene plutonism in the Colombian Andes is exposed the Central Cordillera, Sierra Nevada Santa Marta and the described Parashi Stock within the Guajira Peninsula extends until ca. 47 Ma. After this magmatic event during the late Eocene-Oligocene a regional magmatic hiatus was took place in northwestern Colombia (Bayona et al., 2012). This hiatus was associated to margin segmentation with associated block rotation, basin opening and deformation in other segments of the margin (Macellari, 1995; Montes et al., 2005, 2010; Beardsley and Ave Lallemant, 2007).

During this time interval the Parashi Granitoid document a Late Eocene-Early Oligocene cooling event (Cardona et al., 2011b) followed by submergence and basin formation including carbonate deposition (Zapata et al., 2010).



Fig. 13. Paleogeographic reconstruction of the northern Andes and southern Caribbean at near 50 Ma, based on the palinspastic reconstruction of (Montes et al., 2012a; Pindell and Kennan, 2009), modified for Maracaibo from (Escalona and Mann, 2010); Central America (Meschede and Frisch, 1998); Leeward Antilles (Muessig, 1984; Priem et al., 1986; Stearns et al., 1982). References for plutonic bodies: 1: Mande batholith (Montes et al., 2012b; Villagómez et al., 2011); 2: Mamoni-Cerro Azul intrusive suite (Montes et al., 2012b; Wegner et al., 2011); 3: Azuero intrusives (Montes et al., 2012a; Wegner et al., 2011); 4: Santa Barbara batholith (Ordoñez-Carmona et al., 2011); 5: Manizales stock (Bayona et al., 2012); 7: Sonson batholith (Ordoñez-Carmona et al., 2011); 8: Antioquia batholith (Bayona et al., 2012); 9: Santa Marta batholith (Cardona et al., 2012); 10: Parashi stock (Cardona et al., 2009). References for stratigraphic sections: 11: Gatuncillo Formation (Woodring, 1964); 12: Tonosi Formation (Guidice and Recchi, 1969; Herrera et al., 2012; Kolarsky et al., 2007); 16: Soebi Blanco Formation (Borrero et al., 2012); 14: Lower Amaga Formation (Silva Tamayo et al., 2008); 15: Toluviejo Formation (Priem et al., 2012); 10: Marta Stock (2010); 19: Misoa Formation (Priem et al., 2012); 10: Marta Jabaco Formation (Rollins, 1965); 18: Tabaco Formation (Bayona et al., 2011); 20: Mirador Formation (Mora et al., 2010); 21: La Paz Formation (Nie et al., 2010); 22: Bogota Formation (Moron et al., 2013); 23: Almacigos Member of the San Juan de Rio Seco Formation (Gomez et al., 2003); 24: Palermo Fm. of the Gualanday Group (Anderson, 1972; Caicedo and Roncancio, 1994); 25: Pepino Formation (Borrero et al., 2012).

Plate tectonic reconstructions have suggested that the convergence rates on the South American margin were below 3.7 mm/y between 55.5 Ma and 38.4 Ma (Müeller et al., 1999; Pindell and Keenan, 2009) and convergence relations between South American and the Caribbean apparently changes towards a more oblique strike-slip system. We therefore suggest that the oblique and relatively slow plate convergence caused the magmatic hiatus within the continental margin and succession of transpression and transtension causing exhumation, and disruption of the Guajira Region and must of the South American margin from a coherent Early Paleogene orogen (Fig. 13).

6. Conclusions

The origin of the Parashi Stock includes the mixing of (meta) sedimentary component with mafic component. The spatio-

temporal and regional constraints suggest that this Early Eocene magmatism is related to the formation of a new subduction setting in the northwestern margin of South America, characterized by high convergence obliquity. Strong heat advection linked to this tectonic configuration facilitated the melting of the former tectonically underplated mafic crust with associated sediments. This configuration evolved to an even stronger oblique relationship that facilitated block rotation and pull-apart basin formation on the southern Caribbean margin.

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