Journal of South American Earth Sciences 71 (2016) 108–130

Contents lists available at ScienceDirect

Journal of South American Earth Sciences

journal homepage: www.elsevier.com/locate/jsames

U–Pb detrital zircon ages from some Neoproterozoic successions of Uruguay: Provenance, stratigraphy and tectonic evolution

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ARTICLE INFO

Article history:
Received 7 March 2016
Received in revised form
21 June 2016
Accepted 1 July 2016
Available online 12 July 2016

Keywords:
Neoproterozoic
Provenance
Geochronology
Paleogeography
Uruguay

ABSTRACT

The Neoproterozoic volcano-sedimentary successions of Uruguay have been the subject of several sedimentologic, chrono-stratigraphic and tectonic interpretation studies. Recent studies have shown, however, that the stratigraphy, age and tectonic evolution of these units remain uncertain. Here we use new U–Pb detrital zircon ages, combined with previously published geochronologic and stratigraphic data in order to provide more precise temporal constraints on their depositional age and to establish a more solid framework for the stratigraphic and tectonic evolution of these units. The sequence of events begins with a period of tectonic quiescence and deposition of extensive mixed siliciclastic-carbonate sedimentary successions. This is followed by the development of small fault-bounded siliciclastic and volcaniclastic basins and the emplacement of voluminous granites associated with episodic terrane accretion. According to our model, the Arroyo del Soldado Group and the Piedras de Afilar Formation were deposited sometime between ~1000 and 650 Ma, and represent passive continental margin deposits of the Nico Pérez and Piedra Alta terranes, respectively. In contrast, the Ediacaran San Carlos (<552 ± 3 Ma) and Barriga Negra (<581 ± 6 Ma) formations, and the Maldonado Group (<580–566 Ma) were deposited in tectonically active basins developed on the Nico Pérez and Cuchilla Dionisio terranes, and the herein defined Edén Terrane. The Edén and the Nico Pérez terranes likely accreted at ~650–620 Ma (Edén Accretionary Event), followed by their accretion to the Piedra Alta Terrane at ~620–600 Ma (Piedra Alta Accretionary Event), and culminating with the accretion of the Cuchilla Dionisio Terrane at ~600–560 Ma (Cuchilla Dionisio Accretionary Event). Although existing models consider all the Ediacaran granites as a result of a single orogenic event, recently published age constraints point to the existence of at least two distinct stages of granite generation, which are spatially and temporally associated with the Edén and Cuchilla Dionisio accretionary events.

1. Introduction

The analysis of the U–Pb age spectra on detrital zircons is a useful tool for determining the provenance, maximum depositional age and stratigraphic correlations of clastic sedimentary successions. This is particularly the case in the Precambrian rock record where biostratigraphical markers are lacking and whose original stratigraphic relationships have been obliterated by deformation and metamorphism. Despite the alleged importance of the Neoproterozoic successions of Uruguay, which have been suggested to record global climatic, biogeochemical and biotic events, a recent review by Aubet et al. (2014) concluded that the stratigraphy and age of these units remain ambiguous. Consequently, existing basin models and their tectonic evolution remain poorly resolved.

With the exception of the easternmost Cuchilla Dionisio Terrane (Fig. 1), the early Neoproterozoic (1000–700 Ma) rock record of Uruguay is characterized by the absence of evidence indicating significant tectono-magmatic activity (Aubet et al., 2014). In

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http://dx.doi.org/10.1016/j.jsames.2016.07.003
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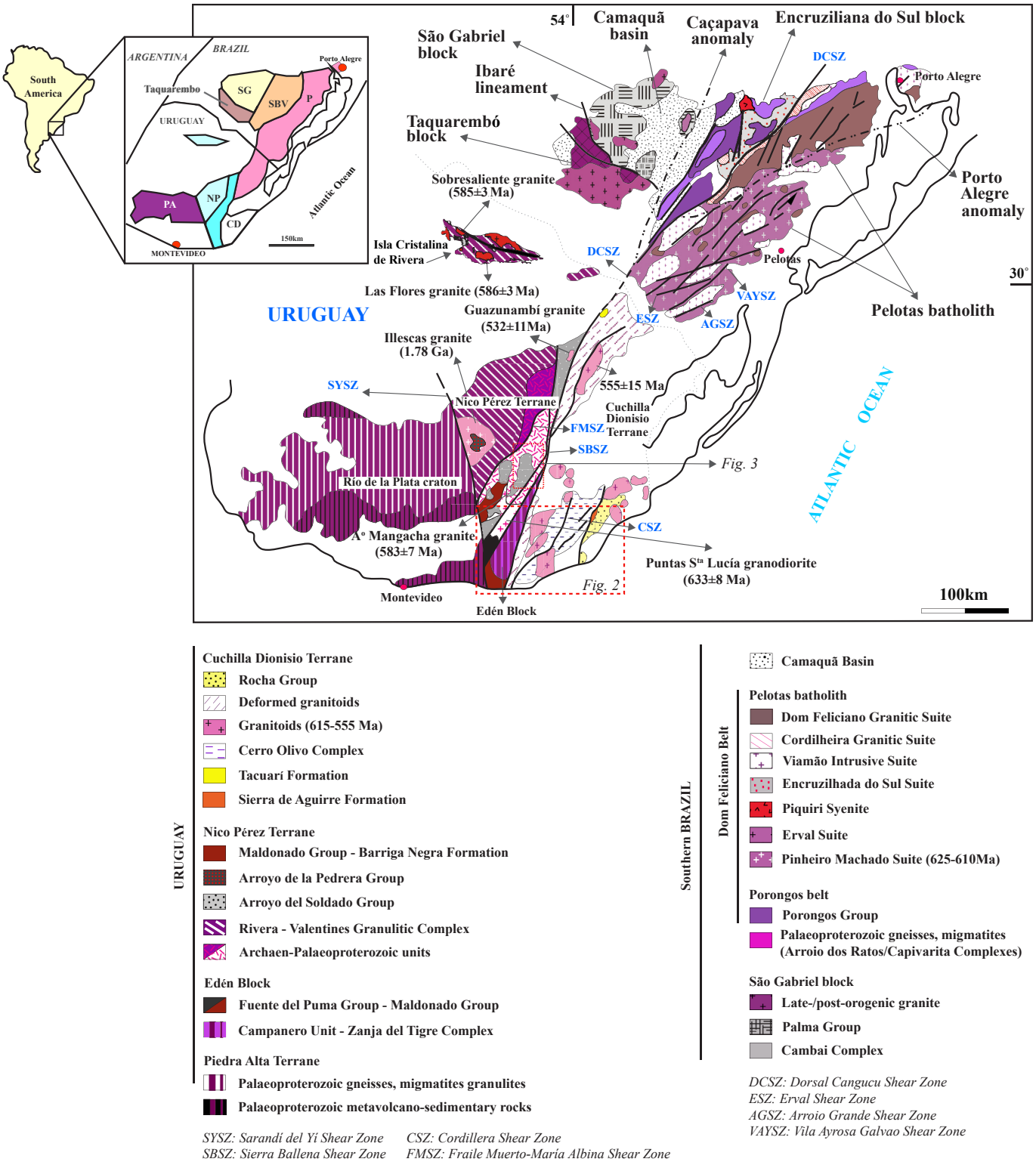

Fig. 1. Simplified geologic map of Uruguay and southern Brazil (modified from Saalmann et al., 2011 and references therein). Inset: PA-Piedra Alta, NP-Nico Pérez, and CD-Cuchilla Dionisio terranes of Uruguay; SG-São Gabriel, SBV-Santana da Boa Vista, P-Pelotas and Taquarembó blocks of southern Brazil.

contrast, part of the late Neoproterozoic (560–650 Ma) record is punctuated by extensive magmatism accompanied by major changes in the regional tectonic regime (Oyhantcabal et al., 2007), i.e., accretion and transport of allochtonous terranes, the so-called Brasilian/Pan-African Orogenic Cycle. This differential tectono-magmatic activity was expressed by the deposition of volcano-sedimentary successions showing a distinct stratigraphic architecture. The early Neoproterozoic record, represented by the Arroyo del Soldado Group, comprises epicontinental deposits that covered the passive margin of the Nico Pérez Terrane, while late

Neoproterozoic successions, which includes the Playa Hermosa, Las Ventanas, San Carlos, Barriga Negra, Fuente del Puma, Sierra de Aguirre, Tacuarí and Rocha formations, points to an active tectono-magmatic settings for their deposition in both, the Nico Pérez and the Cuchilla Dionisio terranes (Fig. 1). However, for most of these units the depositional age is still controversial due to the lack of radiometric ages. Additionally, the lack of a solid chronostratigraphic framework has prevented the development of a coherent tectonic model during terrane accretion.

The primary objective of this study is to constrain the maximum depositional age and provenance of Neoproterozoic volcano-sedimentary successions of Uruguay. In this regard, detrital zircon geochronology from these units is a crucial step for testing previous stratigraphic models and for determining depositional basin histories. Furthermore, by comparing the detrital zircon age provenance patterns in samples investigated in this study with published geochronological studies on basement units, we aim to establish the relationship between these Neoproterozoic successions and the surrounding tectono-stratigraphic terranes.

2. Geological setting

The crystalline basement of Uruguay comprises three major terranes separated by large-scale shear zones: (1) the Paleoproterozoic Piedra Alta Terrane, located to the west of the Sarandí del Yí Shear Zone, (2) the Archean to Paleoproterozoic Nico Pérez Terrane, between the Sarandí del Yí and the Sierra Ballena Shear Zones, and (3) the Neoproterozoic Cuchilla Dionisio Terrane, to the east of the Sierra Ballena Shear Zone (Fig. 1). The Piedra Alta and Nico Pérez terranes have been traditionally considered as forming part of the Río de la Plata Craton. However, Oyhantçabal et al. (2011) recently proposed that the Nico Pérez Terrane should be considered an allochtonous block, and thus the Río de la Plata Craton in Uruguay is restricted to the Piedra Alta Terrane. In contrast to the Piedra Alta Terrane with ages between 2.0 and 2.2 Ga, the Nico Pérez and the Cuchilla Dionisio terranes have been strongly reworked during the Brasilian Cycle forming the Neoproterozoic Dom Feliciano Belt (Oyhantçabal et al., 2009), thought to have occurred between 650 and 550 Ma (Cordani et al., 2000). According to Basei et al. (2000), this orogenic belt is composed of three distinct domains: the eastern granitoid belt, the central supracrustal schist belt and the western foreland basin domain, wherein the latter is less affected by deformation and metamorphism than the adjacent schist belt. Following this model, it has recently been proposed that the eastern boundary of the Nico Pérez Terrane should be placed at the Fraile Muerto-María Albina Shear Zone and not at the Sierra Ballena Shear Zone (Fig. 1) (Preciozzi et al., 1979; Basei et al., 2000; Oyhantçabal et al., 2011). Therefore, the basement block (Campanero Unit and Zanja del Tigre Complex (Sánchez-Bettucci and Ramos, 1999; Sánchez-Bettucci et al., 2003)) located to the west of the Sierra Ballena Shear Zone is considered as a basement inlier of the Dom Feliciano Belt and not part of the Nico Pérez Terrane (Oyhantçabal et al., 2011) (Fig. 1).

The units focused on in this study have previously been included in both, the schist belt and the foreland belt (including syn-to post-collisional strike slip settings) and thought to have been deposited between ~1000 and 450 Ma (see Aubet et al., 2014 for a recent review). For instance, the Las Ventanas Formation (Fig. 2) was long considered to be an alluvial-fan system operating during the Ordovician (Midot, 1984; Pazos et al., 2003). Alternatively, Pecoits (2003) suggested an Ediacaran age (570–590 Ma) and redefined the unit as a volcano-sedimentary succession deposited in a fan-delta sedimentary system formed in a strike-slip basin. Similarly, the Playa Hermosa Formation (Fig. 2) was deposited in a tectonically active basin with locally fast subsidence rates and correlated to the Varanger (Ediacaran) glacial event by Pazos et al. (2003). Although the less studied San Carlos Formation (Fig. 2) was thought to represent a meandering fluvial depositional system of Ordovician age (Sánchez-Bettucci, 1998), a more recent study suggested that the unit possesses similar sedimentological characteristics to those shown by the Las Ventanas Formation, and is likely similar in age (Pecoits et al., 2008). The Playa Hermosa and Las Ventanas formations were grouped into the Maldonado Group (Pecoits et al., 2004), however, their relationship with the San Carlos Formation (located to the east of the Sierra Ballena Shear Zone) remains unknown. The age and depositional environment of the Arroyo del Soldado Group are also controversial. The group has been assigned to the upper Ediacaran (<583 Ma) and to the Cryogenian-Tonian (~700–1000 Ma) deposited in conflicting tectonic settings, either in a passive margin or in a foreland basin (Aubet et al., 2014 and references therein). Furthermore, some units that were originally included in this group were later shown to be independent successions. For example, the Barriga Negra Formation (Fig. 3) has been considered as a late-to post-orogenic (Midot, 1984; Fambrini et al., 2005), or alternatively as a pre-orogenic unit when included within the Arroyo del Soldado Group (Gaucher et al., 2008).

The depositional ages of other units are better constrained, including the Rocha, Sierra de Aguirre and Tacuarí formations (Figs. 1 and 2). According to Basei et al. (2005), sedimentation of the Rocha Formation took place sometime between ~600 and 550 Ma, constrained by the youngest detrital zircon grains and the onset of post-tectonic alkaline magmatism, respectively. Based on their similar lithologies, comparable structural setting and the detrital zircon age patterns, Basei et al. (2005) further suggested that the Oranjemund Group (Namibia) and the Rocha Formation are stratigraphic equivalents, and probably represent sediment fills of the same basin. The volcaniclastic Sierra de Aguirre Formation was deposited in a shallow marine and deltaic settings and is thought to have accumulated in a pull-apart basin (Campal and Schipilov, 2005). A dacite from this unit yielded a SHRIMP U–Pb concordia age of 571 ± 8 Ma (Hartmann et al., 2002). The glacially influenced Tacuarí Formation is the best-constrained unit. U–Pb zircon ages (LA-MC-ICPMS and SHRIMP) obtained from a cross-cutting granitic dyke places a 585 ± 3 Ma minimum depositional age for this unit. A maximum depositional age for the Tacuarí Formation is 600 ± 9 Ma, based on the youngest detrital zircon age cluster recorded (Pecoits et al., 2012, 2014).

3. Methodology

Zircon was recovered from six sandstone samples using standard mineral separation techniques. Approximately 2–3 kg of rock from each sample was pulverized to a fine (80 mesh) powder using a jaw crusher and a rotary disc mill. Zircon concentrates were obtained by a combination of hydro-gravimetric (Wilfley table), magnetic (Frantz isodynamic separator) and density (methylene iodide) separation techniques followed by handpicking individual crystals under a binocular microscope. Zircon grains were secured in epoxy mounts, polished to expose the interior of the crystals, and imaged using a custom modified Zeiss Axioskop 40 cathodoluminescence microscope. The U–Pb analyses were performed with a Nu Plasma I multi-collector ICP-MS equipped with 12 F detectors and 3 ion counter detectors. The zircons were ablated with a 213 nm New Wave laser (4 Hz; 1.13 J/cm^3) and typical analysis spots were 40 μm in diameter. Lead isotope fractionation was monitored by simultaneously aspirating a thallium-doped solution. Three zircon standards were analyzed with each sample, a Paleoproterozoic zircon (LH94-15; 1830 Ma) was used to monitor U/Pb fractionation and a Neoproterozoic (GJ-32–1; 605 Ma) and

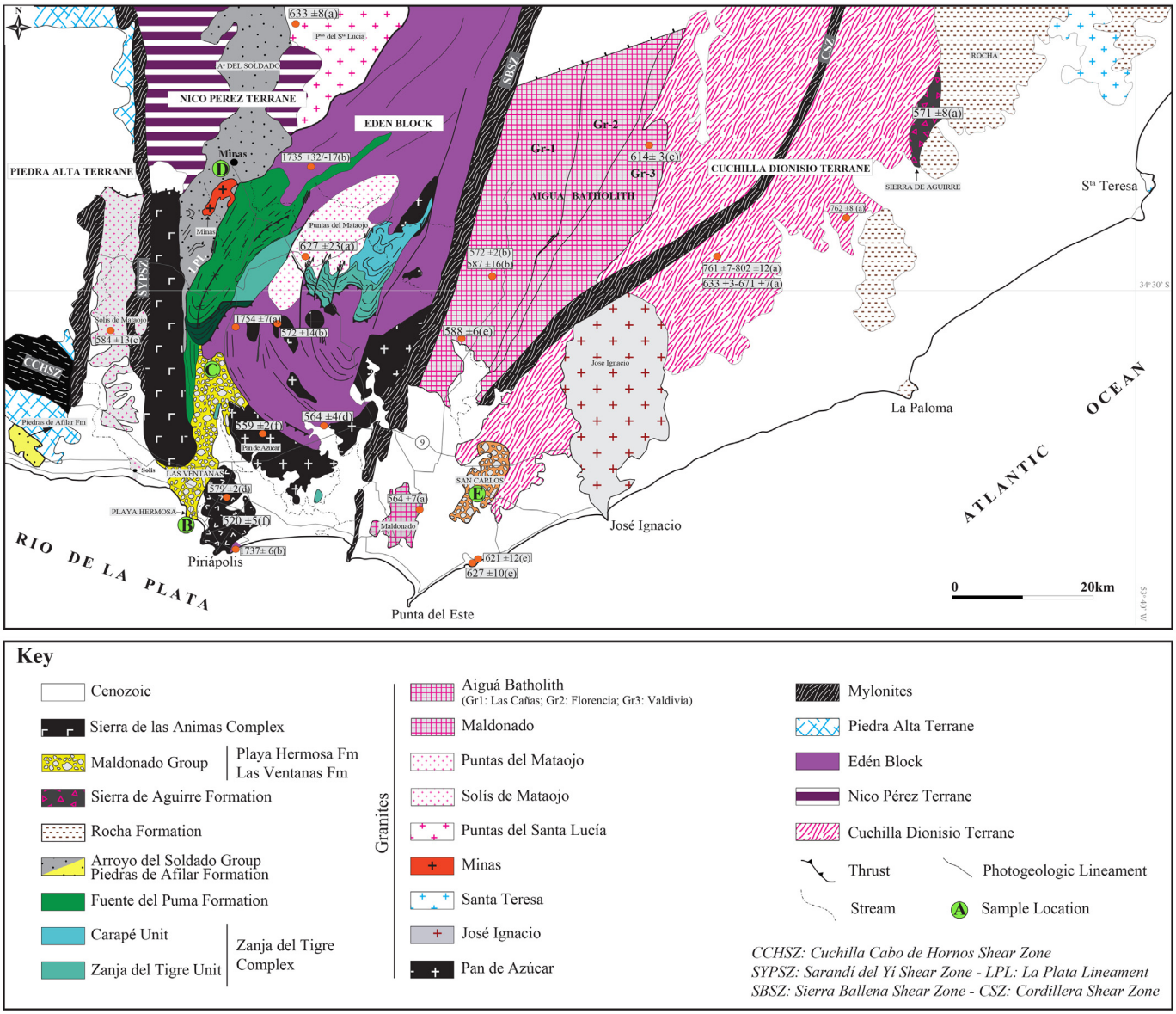

Fig. 2. Geologic map of southeastern Uruguay showing published U–Pb geochronological data and sample locations. Modified from Oyhantçabal (2005), Oyhantçabal et al. (2009) and Pecoits et al. (2011). Radiometric ages: (a) U–Pb SHRIMP zircon, (b) U–Pb TIMS zircon, (c) Pb–Pb titanite, (d) Ar–Ar hornblende, (e) K–Ar muscovite, (f) Rb–Sr whole rock.

Mesoarchean (OG-1; 3465 Ma) zircons were analyzed as blind standards to evaluate the accuracy of the method, a summary of the blind standard results for each session is presented in the supplementary material. In general the results for the blind standards are in agreement with or slightly younger than their recommended ages. The exception is GJ-32–1 analyzed during session 4 where the $^{206}Pb/^{238}U$ date is 10% lower than the recommended age. Common lead corrections were not applied to the data. Long-term monitoring of zircon standards with a range of ages using this protocol indicates that uncorrected $^{206}Pb/^{238}U$ dates are more accurate for grains younger than 800 Ma, for all other analyses $^{207}Pb/^{206}Pb$ dates are used to compile probability density plots. Details of the U–Pb LA-MC-ICPMS technique used at the University of Alberta are outlined in Simonetti et al. (2005, 2006). Data reduction was conducted with an in-house program and data presentation (concordia diagrams, probability density plots) was constructed using Isoplot/Ex (Ludwig, 2008). In some samples the identification of age populations was evaluated using a mixture modelling treatment (Sambridge and Compston, 1994). Ages were calculated using the uranium decay constants and $^{238}U/^{235}U$ value recommended by Jaffey et al. (1971). The analytical data referred to in this paper are available as supplementary data (Tables S1–S7).

4. Results

4.1. U–Pb detrital zircon geochronology

A total of 662 zircons were analyzed in six samples from five different lithostratigraphic units, namely: Yerbal (Arroyo del Soldado Group), Playa Hermosa, Las Ventanas (two samples), San Carlos and Barriga Negra formations. Results are displayed in $^{206}Pb/^{238}U$ and $^{207}Pb/^{206}Pb$ age histograms and probability density plots calculated using the Isoplot 3.7 software by Ludwig (2008). Sample locations (coordinates) and their stratigraphic positions are provided in supplementary material (Table S8) and shown in Figs. 2–4.

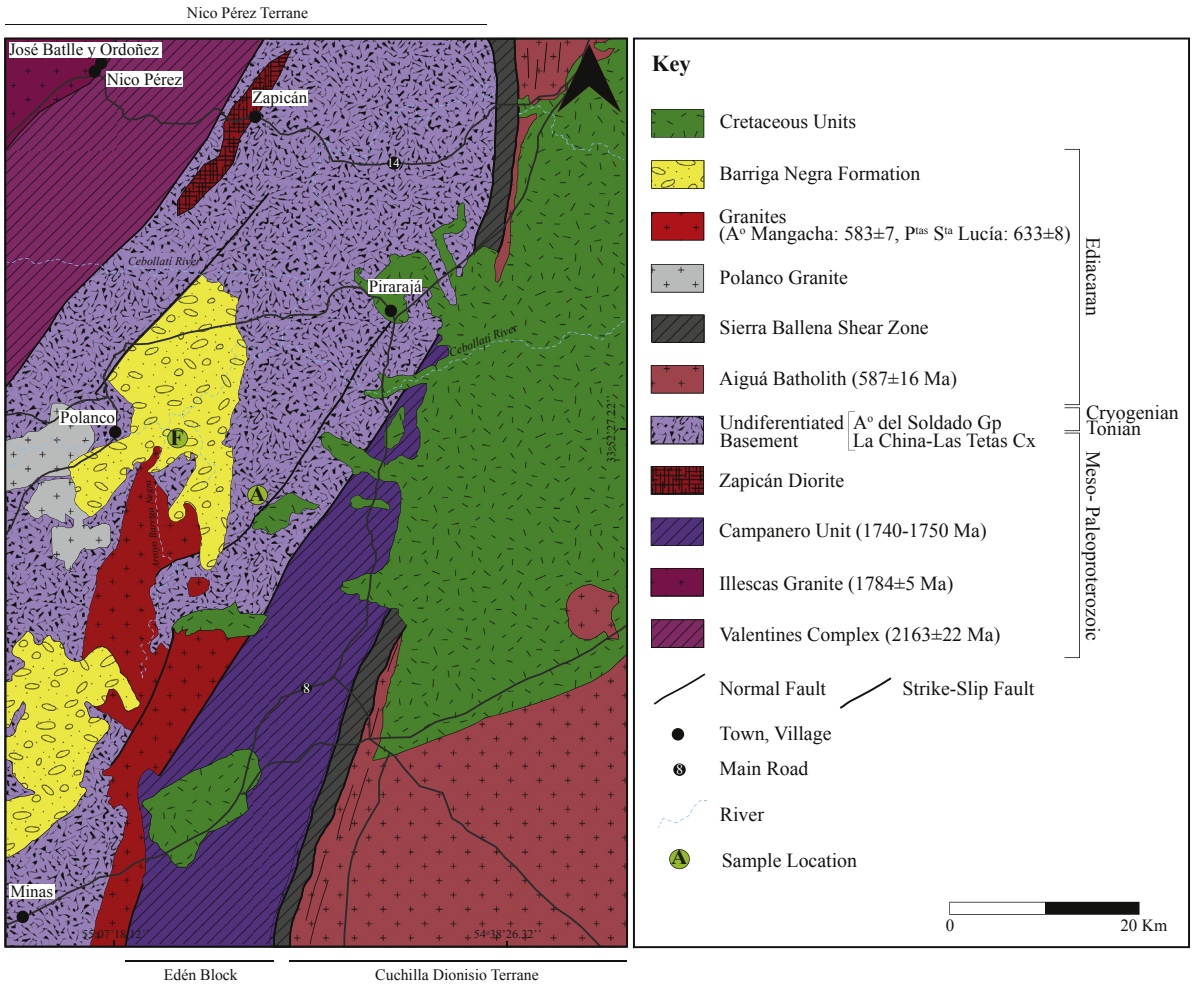

Fig. 3. Simplified geologic map of the type area of the Barriga Negra Formation. Modified from Fambrini et al. (2005) and references therein.

4.1.1. Yerbal Formation

A total of 105 zircon ages were obtained from a coarse sandstone interbedded with dark siltstones and shales of the uppermost Yerbal Formation (sample 101115/1). At this location (Arroyo Tapes Grande area), the unit is overlain by a thick package of limestones and dolostones of the Polanco Limestones Formation and dark shales of the Cerro Espuelitas Formation (Fig. 4A). The probability density plot for this sample displays two distinct age clusters, a Paleoproterozoic population at 2212 ± 4 Ma (n = 24) and a larger Mesoarchean population (n = 57) with two nodes computed from the unmixing treatment at 2915.8 ± 2.0 Ma (34%) and 3038.1 ± 1.3 Ma (66%). The youngest detrital zircon (grain 80) has a $^{207}Pb/^{206}Pb$ date of 1794 Ma and provides a maximum age for deposition of the Yerbal Formation (Fig. 5A).

4.1.2. Playa Hermosa formation

A total of 107 zircon grains were analyzed from a pebbly sandstone of the Playa Hermosa Formation (sample 101026/2). The sample was collected from the upper part of the stratotype of the unit, in the Hermosa Beach near Piriápolis city, where it is interbedded with siltstones and conglomerates (Fig. 4B). This sandstone contains two distinct zircon populations, an Ediacaran cluster at 563 ± 13 Ma (n = 6), providing a maximum age for deposition of this unit, and a dominant Paleoproterozoic cluster with 90 analyses recording dates between 2000 and 2320 Ma. A mixture modelling treatment of the Paleoproterozoic cluster identifies two similar abundance populations at 2073.2 ± 3.1 and 2128.7 ± 2.7 Ma. Fig. 5B displays the zircon distribution of the Playa Hermosa Formation where there is a clear dominance of Paleoproterozoic zircons. Only one zircon (#12) has a near-concordant date that plots between these two peaks at ~1730 Ma.

4.1.3. Las Ventanas formation

Two pebbly sandstone units were analyzed from the Las Ventanas Formation. At its type area (sample 101012/2; Fig. 4C), detrital zircon spot analyses (n = 111) display a prominent peak at 590 ± 5 Ma (n = 42) and two less pronounced peaks at 2127.9 ± 2.1 and 2649.4 ± 2.8 Ma based on an unmixing treatment of the ages between 2.0 and 2.7 Ga (Fig. 5C). The second sample (n = 115), taken from west of Minas city (sample 101026/1; Fig. 4D), exhibits two main populations at 548 ± 19 and 2083 ± 11 Ma, and contains a broad range of zircon ages between ~1050 and 1,950, and between ~2250 and 2650 Ma (Fig. 5D).

4.1.4. San Carlos formation

For the San Carlos Formation (sample San Carlos), 111 zircon dates were obtained from a pebbly sandstone at its type area, wherein sandstone layers are interbedded with grey plane parallel stratified siltstones in the upper part of the unit (Fig. 4E). Three age clusters are observed at 552 ± 3, 647 ± 10 and 764 ± 12 Ma, along with Meso- and Paleoproterozoic zircon grains with dates between ~1100 and 2000 Ma (Fig. 5E, F).

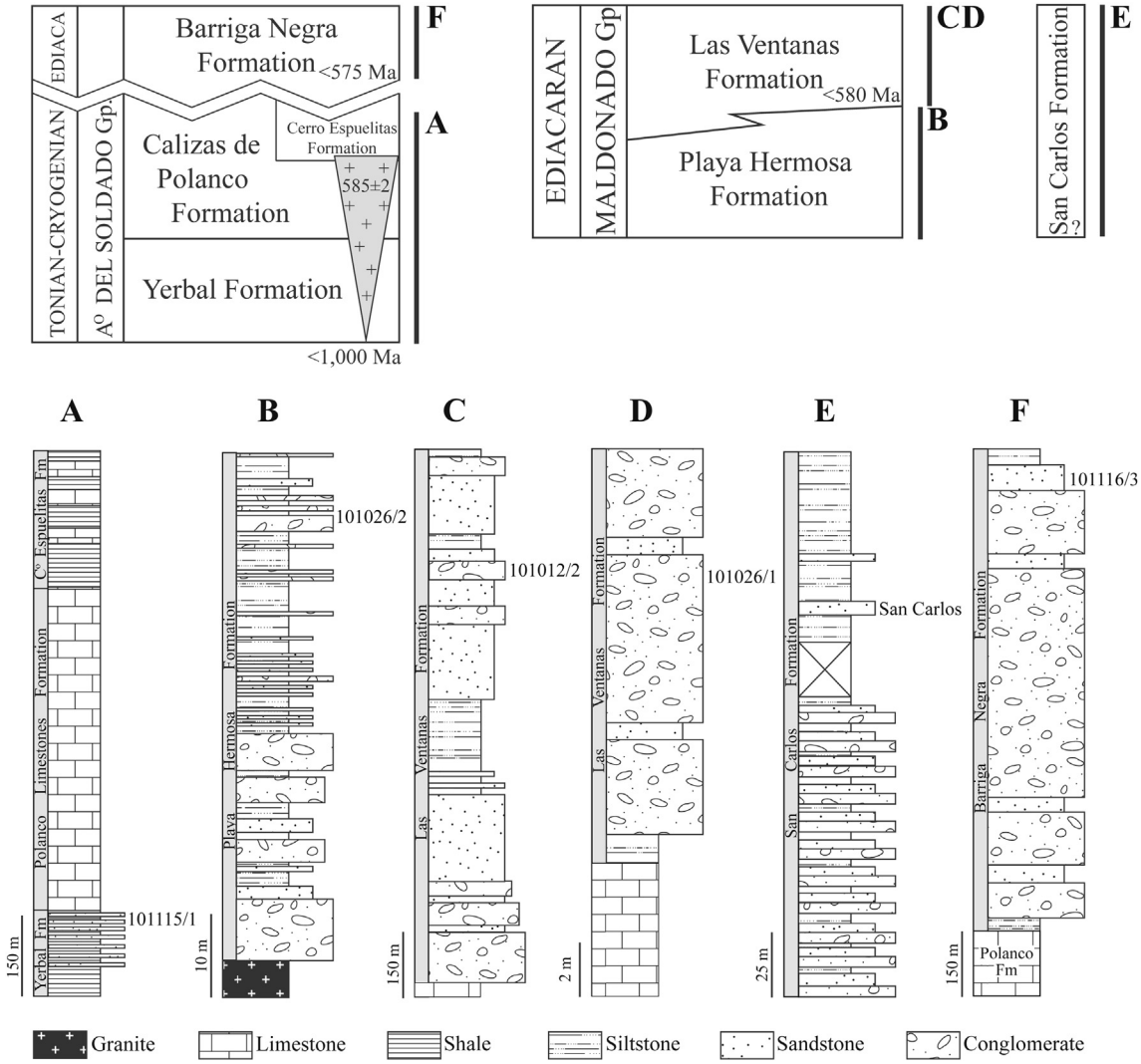

Fig. 4. Lithostratigraphic units sampled in this study (modified from Aubet et al., 2014) and stratigraphic profiles at sample locations (palinspastic restoration was not considered). **(A)** Yerbal Formation (sample 101115/1). **(B)** Playa Hermosa Formation (sample 101026/2). **(C)** Las Ventanas Formation (sample 101012/2). **(D)** Las Ventanas Formation (sample 101012/1). **(E)** San Carlos Formation (sample San Carlos). **(F)** Barriga Negra Formation (sample 101116/3). See Figs. 2 and 3 for sample locations.

4.1.5. *Barriga Negra formation*

The Barriga Negra sample (101116/3) was collected from a sandstone layer overlying coarse conglomerates of the type area of the unit (Francisco Vidal Farm; Figs. 3 and 4F). Detrital zircons (n = 115) from the Barriga Negra Formation display two dominant peaks at 581 ± 6 and 2028 ± 15 Ma. Archean zircons are also abundant and range from ~2600 to 3400 Ma with a noticeable cluster at 2669 ± 15 Ma. A few scattered zircons occur between ~1000 and 1800 Ma (Fig. 5).

5. Discussion

5.1. *Possible source areas*

5.1.1. *Yerbal formation*

The sample from the Yerbal Formation is dominated by Paleoproterozoic and Archean detrital zircon grains (Fig. 5A). Possible sources for the 2212 ± 4 Ma zircons include Paleoproterozoic units of the Piedra Alta and Nico Pérez terranes (Figs. 1 and 2). For example, similar U–Pb ages were obtained from orthogneisses of the San José (~2200 Ma) and Montevideo belts (~2160 Ma) (Table 1). Similarly, this zircon population mirrors inherited (~2220 Ma) and magmatic (~2160–2114 Ma) ages of the Valentines and Rivera Granulitic complexes of the Nico Pérez Terrane. Archean ages (~2700–3400 Ma) are restricted to the La China and Las Tetas complexes of the Nico Pérez Terrane (Fig. 1). Although statistically insignificant, the only zircon at ~1794 Ma was most likely derived from Nico Pérez Terrane (Illescas Granite) (Table 1). Other potential sources, such as metatuffs of the Paso Severino Belt in the Piedra Alta Terrane or the Campanero Unit, are unlikely given the younger ages shown (1730–1760 Ma). Furthermore, the latter unit is also associated with younger Mesoproterozoic volcano-sedimentary successions of the Zanja del Tigre Complex (1435–1490 Ma) and the absence of zircons with this age in the Yerbal Formation suggests that these Paleo- and Mesoproterozoic units (herein included within the Edén Terrane; see below) were not part of the source area.

Two samples were previously analyzed from the Yerbal Formation. One sample (92 zircon grains) yielded an essentially unimodal concentration centered at 2450 Ma, two grains showed an age of 2045 Ma and five grains were dated between 2660 and 2895 Ma (Gaucher et al., 2008; their sample 'MIN 14'). The second

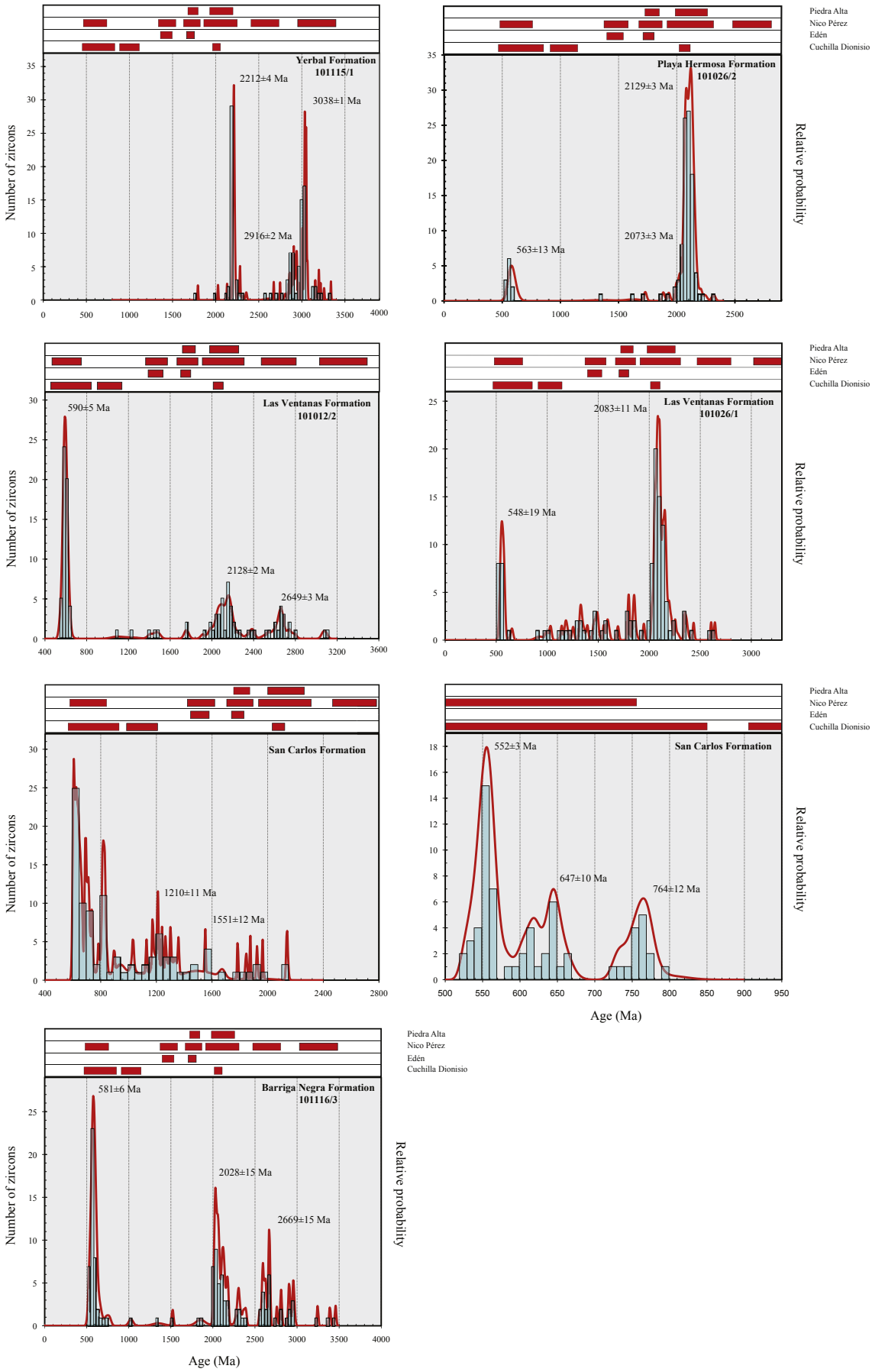

Fig. 5. Combined relative probability density and histogram plots. **(A)** Yerbal Formation (sample 101115/1, n = 105). **(B)** Playa Hermosa Formation (sample 101026/2, n = 107). **(C)** Las Ventanas Formation (sample 101012/2, n = 111). **(D)** Las Ventanas Formation (sample 101012/1, n = 115). **(E–F)** San Carlos Formation (sample San Carlos, n = 111). In (F) results are displayed in $^{206}Pb/^{238}U$ ages. **(G)** Barriga Negra Formation (sample 101116/3, n = 115). Representative detrital zircon grains from each sample are shown in Supplementary Material (Fig. S1).

sample (29 zircon grains) showed a predominance of Paleoproterozoic zircons (76%), with ages ranging from 1900 to 2220 Ma and a smaller late Mesoproterozoic cluster between 1010 and 1065 Ma (17%). The oldest zircon displays a Mesoarchean age (3030 Ma), while the youngest and the only Neoproterozoic aged zircon was dated at 664 ± 14 Ma (Blanco et al., 2009; their sample 'Yerbal Fm.'). As discussed by Aubet et al. (2014), the interpretation of the Neoproterozoic zircon in this sample is problematic because this single grain is statistically insignificant and it has not been recorded in any other sample from the Yerbal Formation, and thus it should be taken with caution. Here, we further support this suggestion based on the following arguments: (1) the new dated sample from the Yerbal Formation does not show any Neoproterozoic zircon; in fact, none of the three samples dated from that unit have zircon grains younger than 1036 Ma; (2) none of the Neoproterozoic volcano-sedimentary units focused on in this study (located to the west of the Sierra Ballena Shear Zone) show detrital zircons of that age (all of them show a clear gap between ~650 and 1000 Ma); and (3) as pointed out by Aubet et al. (2014), the same gap exists in basement rocks, which would explain the lack of detrital zircons of that age in the Neoproterozoic cover. Hence, the 664 ± 14 Ma-age zircon reported by Blanco et al. (2009) is not considered valid herein.

When the three Yerbal samples are analyzed in detail, some differences arise. The two northernmost samples ('Yerbal Fm.' and 101115/1) display a marked peak at ~2200 Ma (76% and 37%, respectively), while the southernmost sample ('MIN 14') shows no zircons at that age but it displays a significant cluster at 2450 Ma (92%). Similarly, the former two samples show relatively abundant Archean zircons whereas the latter only shows four Archean grains. Another striking difference is the presence of Mesoproterozoic (~1036 Ma) zircons in the northernmost sample ('Yerbal Fm.'), which are completely absent in the other two samples. Taken together, these results suggest that the Yerbal Formation was mainly sourced from the Nico Pérez Terrane. Even though the Piedra Alta Terrane could source 2200 Ma zircons, their presence in the Yerbal samples is accompanied by Archean zircon grains, which only occur in the Nico Pérez Terrane and are not accompanied by younger (~1750–2100 Ma) grains also common in the Piedra Alta Terrane (Table 1). Furthermore, the southernmost Yerbal sample ('MIN 14'), which is closest to the Piedra Alta Terrane at its present position, shows no zircons from this terrane. Strikingly, this sample displays a Paleoproterozoic cluster at 2450 Ma. There is no record of rocks with this age in the Nico Pérez Terrane and the only relatively nearby rocks of similar age (2480–2440 Ma) are found in the São Gabriel Block, in southern Brazil (Fig. 1) (Hartmann et al., 2004). However, this sample is a very coarse-grained subarkose characterized by abundant relatively fresh feldspars, which would indicate a proximal source area (Gaucher et al., 2008), and discount recycled-secondary sources. Hence, we predict that future studies will identify basement rocks of this age in the Nico Pérez Terrane that are hitherto unknown in Uruguay.

Similarly, the ~1036 Ma peak recorded in the Yerbal Formation does not correspond to any known basement unit of the Nico Pérez Terrane. These zircons could derive either from (1) the Cuchilla Dioniso Terrane, (2) Mesoproterozoic rocks from elsewhere, (3) basement rocks not yet identified in the Nico Pérez Terrane or, alternatively, (4) the source rocks were simply eroded away. The Cuchilla Dionisio and Nico Pérez terranes were suggested to have amalgamated in the mid-Ediacaran (Pecoits et al., 2004, 2008; Oyhantçabal et al., 2009), and their proximity during the deposition of the Tonian-Cryogenian Yerbal Formation (see section 5.2 below) is unknown. Although similar ages (~1000 Ma) were obtained in the Cuchilla Dioniso Terrane (Preciozzi et al., 1999b), they correspond to inherited zircon grains (Oyhantçabal et al., 2009) and new geochronological studies on this terrane indicated magmatism at ~800–760 Ma, followed by high-grade metamorphism at ~670–630 Ma (Hartmann et al., 2002; Oyhantçabal et al., 2009; Lenz et al., 2011; Masquelin et al., 2012; Basei et al., 2011). Therefore, it is very unlikely that those Mesoproterozoic zircons were sourced from this terrane without showing any evidence of the more typical Neoproterozoic ages (Table 1 and Fig. 2). The same holds true if an African source is considered. This is further supported by paleocurrent directions measured in the Yerbal Formation indicating source areas located to the northwest. Hence, possible source areas were located to the west of the Sierra Ballena Shear Zone and not to the east. In this regard, alternative sources for these Mesoproterozoic zircons are represented by the 'Grenvillian-type age' Sunsas orogenic belt at the southwestern corner of the Amazon Craton (Cordani et al., 2010), and the proto-Andean belt in the western Río de la Plata Craton (Rapela et al., 2010; Ramos, 2010). In either case, the Mesoproterozoic zircons found in the Yerbal Formation (Blanco et al., 2009 their Fig. 7) are less rounded than other Paleoproterozoic-Archean zircons sourced from local areas (Zimmermann, 2011). This last point is difficult to reconcile with the more than 1000 km that separates these areas from the Arroyo del Soldado outcrops at their present position. Therefore, excluding the above mentioned hypotheses to explain the presence of the ~1036 Ma zircons in the Yerbal Formation we are left with three alternatives: (1) the Nico Pérez Terrane was closer to the above mentioned (Sunsas and proto-Andean belts) or other Grenville-age source areas during the deposition of the Yerbal Formation (Arroyo del Soldado Group); (2) ~1036 Ma rocks have not yet been identified in the Nico Pérez Terrane; or (3) they have been eroded away.

5.1.2. Playa Hermosa formation

The dominance of Paleoproterozoic dates with two age nodes at 2073.2 ± 3.1 and 2128.7 ± 2.7 Ma in the detrital zircon age spectra of the Playa Hermosa Formation is readily explained when considering Paleoproterozoic units of the Piedra Alta and Nico Pérez terranes as possible source areas. For example, the oldest ages overlap with U–Pb ages obtained from granites and gneisses of the Montevideo Belt and metavolcanics of the Paso Severino Belt, whereas the youngest peak is similar within error to the age of late tectonic granites, such as the Isla Mala Granitic Suite of the Piedra Alta Terrane (Table 1). Similarly, these two zircon populations mirror the magmatic (~2160–2114 Ma) and metamorphic (~2060–2090 Ma) ages of the Valentines and Rivera Granulitic complexes of the Nico Pérez Terrane (Fig. 1). The only zircon at ~1730 Ma could also derive from any of these two terranes and even from the Edén Terrane (Campanero Unit: 1730–1760 Ma). Considering the lack of the typical Archean ages found in the Nico Pérez Terrane and present in all the Neoproterozoic successions located to the west of the Sierra Ballena Shear Zone, we suggest that the Piedra Alta was the most important source area, if not the only one, for the Playa Hermosa Formation. This is further supported by paleocurrent and paleoslope directions measured in this unit indicating source areas located to the southwest (Pazos et al., 2003; Pecoits et al., 2008). A maximum depositional age for the Playa Hermosa Formation is constrained by the youngest zircon population at 563 ± 13 Ma. This age coincides with that obtained from basal volcaniclastic deposits in the Las Ventanas Formation at 573 ± 11 Ma (Table 1). Intense magmatism represented by volcanic, sub-volcanic and intrusive units has been reported in the area, but no U–Pb ages are available. The most reliable age is that obtained from the Pan de Azúcar Pluton at 579 ± 2 Ma ($^{40}Ar/^{39}Ar$ in hornblende; Oyhantçabal et al., 2007), which are similar to both the volcaniclastic deposits and the youngest zircon population found in the Playa Hermosa Formation.

Recently, Rapalini et al. (2015) reported detrital zircon ages (41

Table 1
Compilation of published U–Pb zircon (unless otherwise indicated) geochronology data (see also Figs. 1 and 2).

Terrane	Unit	Rock Type	Age	Method	Reference
Piedra Alta	San José	Amphibolic gneiss	2202 ± 8	TIMS	Peel and Preciozzi (2006)
	Montevideo Granite (Pajas Blancas)	Porphyritic granite	2158 + 24/−23	TIMS	Peel and Preciozzi (2006)
	Montevideo Gneiss	Orthogneiss	2165 ± 38	SHRIMP	Cordani in Santos et al. (2003)
	Paso Severino	Metadacite	2146 ± 8	SHRIMP	Santos et al. (2003)
	Isla Mala Pluton	Granodiorite	2122 ± 4	SHRIMP	Santos et al. (2003)
	Isla Mala Pluton	Monzogranite	2111 ± 9	SHRIMP	Santos et al. (2003)
	Isla Mala Pluton	Granodiorite	2088 ± 12 (MSWD = 36)	TIMS	Preciozzi et al. (1999a)
	Isla Mala Pluton	Granodiorite	2086 ± 11 (MSWD = 100)	TIMS	Peel and Preciozzi (2006)
	Rospide Gabbro	Gabbro	2086 ± 7	SHRIMP	Hartmann et al. (2000)
	Soca Granite	Monzogranite (charnockite)	2078 ± 8	TIMS	Peel and Preciozzi (2006)
	Rospide Gabbro	Gabbro	2076 ± 6	SHRIMP	Hartmann et al. (2008)
	Isla Mala Pluton	Granodiorite	2074 ± 6	SHRIMP	Hartmann et al. (2000)
	Isla Mala Pluton	Monzogranite	2065 ± 9	SHRIMP	Hartmann et al. (2000)
	Soca Granite	Monzogranite (charnockite)	2056 ± 6	SHRIMP	Santos et al. (2003)
	Soca Granite	Monzogranite (charnockite)	2054 ± 11	SHRIMP	Bossi et al. (2001)
	Cufré Granite	Granite	2053 ± 14	TIMS	Peel and Preciozzi (2006)
	Granodiorite Florida	Porphiritic Granodiorite	2011 + 15/−5	TIMS	Peel and Preciozzi (2006)
	Boca del Rosario Migmatite	Melanosome	2007 ± 14	TIMS	Peel and Preciozzi (2006)
	Florida Mafic Dyke Swarm	Dolerite	1790 ± 5	TIMS (Baddeleyite)	Halls et al. (2001)
	Paso Severino	Felsic metatuffs	1753 + 6/−4	TIMS	Peel and Preciozzi (2006)
	La Paz	Granite	587 ± 8	LA-ICP-MS	Cingolani et al. (2012)
Sarandí del Yí Shear Zone	Solís de Mataojo Granite	Tonalite	584 ± 13	Pb–Pb (Sphene)	Oyhantçabal et al. (2007)
Nico Pérez	La China Complex	Metatonalite orthogneiss	3404 ± 8	SHRIMP	Hartmann et al. (2001)
	La China Complex	Metatonalite orthogneiss	3340 ± 19	SHRIMP	Hartmann et al. (2001)
	La China Complex	Metatonalite orthogneiss	3281 ± 17	SHRIMP	Hartmann et al. (2001)
	La China Complex	Metatonalite orthogneiss	3101 ± 4	SHRIMP	Hartmann et al. (2001)
	La China Complex	Metatonalite orthogneiss	3096 ± 45	TIMS (upper intercept)	Gaucher et al. (2011)
	La China Complex	Metatonalite orthogneiss	2721 ± 7	SHRIMP	Hartmann et al. (2001)
	Las Tetas Complex	Muscovite quartzite	<3145 ± 4	SHRIMP	Hartmann et al. (2001)
	Las Tetas Complex	Metaconglomerate	<2762 ± 8	SHRIMP	Hartmann et al. (2001)
	Valentines Complex	Tonalitic grannulite	2619 ± 8	SHRIMP	Santos et al. (2003)
	Valentines Complex	Tonalitic grannulite	2600	SHRIMP	Hartmann et al. (2001)
	Valentines Complex	Tonalitic grannulite	2535 ± 12	SHRIMP	Santos et al. (2003)
	Valentines Complex	Tonalitic grannulite	2224 ± 4	SHRIMP	Hartmann et al. (2001)
	Valentines Complex	Granitic granulite	2224 ± 4	SHRIMP	Santos et al. (2003)
	Rivera Granulitic Complex	Granulitic gneiss	2172 ± 8	SHRIMP	Oyhantçabal et al. (2012)
	Rivera Granulitic Complex	Trondhjemitic gneiss	2163 ± 8	SHRIMP	Santos et al. (2003)
	Valentines Complex	Tonalitic grannulite	2163 ± 22	SHRIMP	Santos et al. (2003)
	Las Flores Granite	Granite	2161 ± 14	SHRIMP	Oyhantçabal et al. (2012)
	Rivera Granulitic Complex	Granulitic gneiss	2147 ± 9	SHRIMP	Oyhantçabal et al. (2012)
	Rivera Granulitic Complex	Trondhjemitic gneiss	2140 ± 6	SHRIMP	Santos et al. (2003)
	Rivera Granulitic Complex	Granulitic gneiss	2114 ± 3	SHRIMP	Oyhantçabal et al. (2012)
	Rivera Granulitic Complex	Granulitic gneiss	2093 ± 36	SHRIMP	Oyhantçabal et al. (2012)
	Las Flores Granite	Granite	2083 ± 14	SHRIMP	Oyhantçabal et al. (2012)
	Rivera Granulitic Complex	Trondhjemitic gneiss	2077 ± 6	SHRIMP	Santos et al. (2003)
	Valentines Complex	Tonalitic grannulite	2058 ± 3	SHRIMP	Santos et al. (2003)
	Rivera Granulitic Complex	Granulitic gneiss	1975 ± 5	EMP (Monazite)	Oyhantçabal et al. (2012)
	Illescas Granite	Rapakivi granite	1784 ± 5	TIMS	Heaman in Campal and Schipilov (1995)
	Puntas del Santa Lucía Pluton	Granite	633 ± 8	SHRIMP	Hartmann et al. (2002)
	Sobresaliente Granite	Granite	585 ± 3	SHRIMP	Oyhantçabal et al. (2012)
	Las Flores Granite	Granite	586 ± 3	SHRIMP	Oyhantçabal et al. (2012)
	Arroyo Mangacha Granite	Granite	583 ± 7	SIMS	Gaucher et al. (2008)
Edén Block	Zanja del Tigre Complex	Metasandstone	1800–3400	SHRIMP	Oyhantçabal et al. (2005)
	Zanja del Tigre Complex	Sandstone	1950–3550	SHRIMP	Mallmann et al. (2007)
	Campanero Unit	Orthogneiss	1754 ± 7	SHRIMP	Mallmann et al. (2007)
	Campanero Unit	Orthogneiss	1737 + 6/−8	TIMS	Oyhantçabal et al. (2005)
	Campanero Unit	Orthogneiss	1735 + 32/−17	TIMS	Sánchez-Bettucci et al. (2003)
	Zanja del Tigre Complex	Metagabbro	1492 ± 4	TIMS	Oyhantçabal et al. (2005)
	Zanja del Tigre Complex	Volcaniclastics	1433 ± 6	TIMS	Gaucher et al. (2011)
	Zanja del Tigre Complex	Volcaniclastics	1429 ± 21	TIMS	Oyhantçabal et al. (2005)
	Fuente del Puma Formation (Lavalleja)	Matabasalt	667 ± 4	TIMS	Sánchez-Bettucci et al. (2003)
	Puntas del Mataojo Granite	Granite	627 ± 23	SHRIMP	Oyhantçabal et al. (2009)
	Fuente del Puma Formation (ex-Lavalleja)	Metasandstone	700–3400	SHRIMP	Oyhantçabal et al. (2005)
	Las Ventanas Formation	Gabbro	590 ± 2	SHRIMP	Mallmann et al. (2007)
	Sierra de Animas Complex	Syenite	587 ± 7	SHRIMP (upper intercept)	Rapalini et al. (2015)
	Sierra de Animas Complex	Rhyolite	582 ± 3	SHRIMP	Rapalini et al. (2015)

Table 1 (*continued*)

Terrane	Unit	Rock Type	Age	Method	Reference
	Sierra de Animas Complex	Microsyenitic Dyke	575 ± 8	SHRIMP	Rapalini et al. (2015)
	Las Ventanas Formation	Volcaniclastic	573 ± 11	SHRIMP	Oyhantçabal et al. (2009)
	Carapé Complex	Granite	572 ± 14 (MSWD = 93)	TIMS	Sánchez-Bettucci et al. (2003)
	Unamed Alkaline Granite	Granite	530 ± 14	TIMS	Oyhantçabal et al. (2005)
Cuchilla Dionisio	Cerro Olivo Complex	Orthogneiss	2058 ± 10	SHRIMP	Hartmann et al. (2002)
	Cerro Olivo Complex	Orthogneiss	1073 ± 28	SHRIMP	Basei et al. (2011)
	Cerro Olivo Complex	Orthogneiss	1081 ± 58	TIMS	Basei et al. (2011)
	Cerro Olivo Complex	Orthogneiss	1008 ± 60	TIMS	Preciozzi et al. (1999b)
	Cerro Olivo Complex	Orthogneiss	1006 ± 37	TIMS	Preciozzi et al. (1999b)
	Cerro Olivo Complex	Orthogneiss	1005 ± 20	TIMS	Basei et al. (2011)
	Cerro Olivo Complex	Orthogneiss	1001 ± 17	TIMS	Basei et al. (2011)
	Cerro Olivo Complex	Orthogneiss	ca. 990–1000	SHRIMP (xenocryst)	Masquelin et al. (2012)
	Cerro Olivo Complex	Orthogneiss	984 ± 64	TIMS	Preciozzi et al. (1999b)
	Cerro Olivo Complex	Orthogneiss	961 ± 3	TIMS	Basei et al. (2011)
	Cerro Olivo Complex	Orthogneiss	802 ± 12	SHRIMP	Lenz et al. (2011)
	Cerro Olivo Complex	Orthogneiss	796 ± 8	SHRIMP	Lenz et al. (2011)
	Cerro Olivo Complex	Orthogneiss	797 ± 8	SHRIMP	Lenz et al. (2011)
	Cerro Olivo Complex	Orthogneiss	795 ± 8	SHRIMP	Lenz et al. (2011)
	Cerro Olivo Complex	Orthogneiss	794 ± 8	SHRIMP	Lenz et al. (2011)
	Cerro Olivo Complex	Orthogneiss	793 ± 4	SHRIMP	Lenz et al. (2011)
	Cerro Olivo Complex	Orthogneiss	788 ± 6	SHRIMP	Lenz et al. (2011)
	Cerro Olivo Complex	Orthogneiss	786 ± 9	SHRIMP	Lenz et al. (2011)
	Cerro Olivo Complex	Orthogneiss	780 ± 5	SHRIMP	Lenz et al. (2011)
	Cerro Olivo Complex	Orthogneiss	778 ± 7	SHRIMP	Masquelin et al. (2012)
	Cerro Olivo Complex	Orthogneiss	776 ± 12	SHRIMP	Oyhantçabal et al. (2009)
	Cerro Olivo Complex	Orthogneiss	771 ± 6	SHRIMP	Lenz et al. (2011)
	Cerro Olivo Complex	Orthogneiss	767 ± 9	SHRIMP	Lenz et al. (2011)
	Cerro Olivo Complex	Orthogneiss	762 ± 8	SHRIMP	Hartmann et al. (2002)
	Cerro Olivo Complex	Orthogneiss	761 ± 7	SHRIMP	Basei et al. (2011)
	Cerro Olivo Complex	Orthogneiss	753 ± 14	SHRIMP	Basei et al. (2011)
	Cerro Olivo Complex	Orthogneiss	743 ± 7	SHIRMP	Hartmann et al. (2002)
	Cerro Olivo Complex	Migmatite	704 ± 17	TIMS	Preciozzi et al. (2001)
	Cerro Olivo Complex	Orthogneiss	671 ± 7	SHRIMP	Masquelin et al. (2012)
	Cerro Olivo Complex	Migmatite	655 ± 6	TIMS	Basei et al. (2011)
	Cerro Olivo Complex	Paragneiss	645 ± 4	TIMS (Monazite)	Basei et al. (2011)
	Cerro Olivo Complex	Orthogneiss	641 ± 17	SHIRMP	Oyhantçabal et al. (2009)
	Cerro Olivo Complex	Orthogneiss and Pegmatitic Veins	637 ± 11	SHIRMP	Masquelin et al. (2012)
	Cerro Olivo Complex	Augen Gneiss	636 ± 19	TIMS	Basei et al. (2011)
	Cerro Olivo Complex	Paragneiss	633 ± 3	TIMS	Basei et al. (2011)
	Rocha Group	Sandstone	629 ± 17–2611 ± 7	SHRIMP	Basei et al. (2005)
	Cerro Olivo Complex	Migmatite	626 ± 25	TIMS (Zircon+Monazite)	Preciozzi et al. (2001)
	Cerro Olivo Complex	Paragneiss	616 ± 19	TIMS	Basei et al. (2011)
	Valdivia Granite	Granite	614 ± 3	TIMS (Titanite)	Oyhantçabal et al. (2007)
	Aiguá Granite	Granite	587 ± 16	TIMS	Basei et al. (2000)
	Tacuarí Granite	Granite	585 ± 3	LA-ICP-MS/TIMS/ SHRIMP	Pecoits et al. (2012)
	Florencia Granite	Granite	572 ± 2	TIMS	Basei et al. (2000)
	Sierra de Aguirre Formation	Dacite	571 ± 8	SHRIMP	Hartmann et al. (2002)
	Maldonado Granite	Granite	564 ± 7	SHRIMP	Oyhantçabal et al. (2009)
	Cerro Olivo Complex	Migmatite (leucosome)	546 ± 69	TIMS	Preciozzi et al. (2001)
	Cerro Olivo Complex	Migmatite (leucosome)	520–540	TIMS	Preciozzi et al. (1999b)
	El Pintor Granite	Monzogranite	525 ± 9	TIMS (lower intercept)	Basei et al. (2011)
	Cerro Olivo Complex	Migmatite (leucosome)	510 ± 130	TIMS (lower intercept)	Preciozzi et al. (1999b)

zircon grains) from a sample collected at the basal levels of this formation. These grains define a bimodal population with ~60% of zircons yielding ages between 2050 and 2200 Ma while the remaining 40% defines a mean of ~615 + 6/−10 Ma (youngest analytical result: 594 ± 16 Ma; Rapalini et al., 2015). Furthermore, the same authors presented three U/P (SHRIMP) ages from syentic and rhyolitic rocks of the Sierra de Animas Complex, which show a grand mean weighted age of 578 ± 4.3 Ma. Of particular interest is the microsyentic dyke intruding the upper levels of the Playa Hermosa Formation (their sample SdAn1), which yields a "concordia" age of 574.5 ± 8.1 Ma (calculated from six grains). Based on these results, Rapalini et al. (2015) suggested a deposition age sometime between 594 and 578 ± 4 Ma for the lower Playa Hermosa Formation.

The combination of these data with our results suggests that the unit was likely deposited between ~576 and 566.4 Ma, the approximate oldest age of our youngest detrital zircon peak and the youngest age of the intruding microsyenitic dyke, respectively.

5.1.3. Las Ventanas formation

At its type area, the zircon age spectra of the Las Ventanas Formation show three main age clusters: (1) 2649 ± 3; (2) 2128 ± 2;

and (3) 590 ± 5 Ma. The oldest zircon population (2649 ± 3 Ma) suggests that basement of the Nico Pérez Terrane contributed to this unit because no igneous Archean rocks have been found elsewhere in the basement of Uruguay. For example, similar ages have been obtained in La China, Las Tetas and Valentines complexes of the Nico Pérez Terrane (Table 1). The second cluster (2128 ± 2 Ma) could be sourced from the Piedra Alta or Nico Pérez terranes. Both terranes have units with ages well within this range (Table 1). The few zircon grains dated at ~3,100, 2,400, and between 1750 and 1100 Ma, can also be explained by inputs from the basement of the Nico Pérez and Edén terranes. Based on the available U–Pb data, source units for these zircons include, La China (2720–3400 Ma) and Las Tetas (2762–3262 Ma) complexes in the Nico Pérez Terrane, and the Campanero Unit (1737–1754 Ma) and Zanja del Tigre Complex (1429–1492 Ma) of the Edén Terrane (Table 1). The Piedra Alta Terrane could only source 1750 Ma zircons because similar ages were obtained in felsic metatuffs of the Paso Severio Belt (Table 1). Hence, the unit seems to have been mainly sourced from the Nico Pérez Terrane and Edén terranes. The large Ediacaran cluster at 590 ± 5 Ma defines the maximum depositional age for the Las Ventanas Formation. Similar to the Playa Hermosa Formation, this peak is most likely related to the Ediacaran magmatism described in the area, and it fits with the age reported for basal volcaniclastic rocks of the Las Ventanas Formation at 573 ± 11 Ma (Table 1).

The sample taken from the conglomerates near Minas city shows two prominent peaks at 548 ± 19 Ma, and between 2083 ± 11 and 2150 Ma. The latter range is virtually the same to that shown by the Playa Hermosa Formation, and as such, it can largely be explained by contributions from the Piedra Alta Terrane. However, this sample also shows a wide range of zircon grains with ages between ~1050 and 2650 Ma. Considering that the Piedra Alta Terrane does not record magmatic activity, with the exception of the La Paz Neoproterozoic granite and some Paleoproterozoic metatuffs (~1753 Ma) after the intrusion of the Florida dyke swarm at 1790 ± 5 Ma (Halls et al., 2001), it can be argued that both the Piedra Alta and the Nico Pérez terranes were source areas. The same is true for the Archean zircons. As stated above, there is no record of ~1050 Ma zircons in basement rocks of the Nico Pérez Terrane. However, the detrital zircon spectra of the Arroyo del Soldado Group also shows zircons of this age, which indicate that the ~1050 Ma zircons recorded in the Las Ventanas Formation were likely sourced by the Arroyo del Soldado Group. If so, these zircons would represent second-cycle grains. On the other hand, with the exception of the Illescas Rapakivi granite (1784 ± 5 Ma; Fig. 1), zircons between ~1200 and 1800 Ma are not recorded in the basement of the Nico Pérez Terrane or in the Arroyo del Soldado Group implying that rocks from the Edén Terrane (1400–1800 Ma) also sourced the Las Ventanas Formation at this location. The youngest zircon cluster at 548 ± 19 Ma constrains the maximum age of deposition for this conglomerate, which overlaps with that of the Playa Hermosa Formation and is slightly younger than that of the Las Ventanas Formation at its type area.

In summary, the results for the Playa Hermosa and Las Ventanas formations (Maldonado Group) show that, although spatially variable, the Piedra Alta, Nico Pérez and Edén terranes were important source areas. Both samples from the Las Ventanas Formation show a wide range of zircons that could only be sourced from the Nico Pérez and Edén terranes. Although the Playa Hermosa Formation also shows Paleoproterozoic zircons that could be sourced from the same areas, the lack of Archean, Meso- and upper Paleoproterozoic ages (except for one single grain at ~1730 Ma), also typical in the Nico Pérez Terrane, seems to indicate that the Piedra Alta Terrane was the ultimate origin of the above mentioned Paleoproterozoic cluster in the Playa Hermosa Formation. Therefore, we suggest that the main source area of the Playa Hermosa Formation was the Piedra Alta Terrane, although minor inputs from the Nico Pérez cannot be excluded. The sedimentological features and internal architecture of the Maldonado Group, together with the variations observed in the zircon age distribution in all the three samples, most likely reflects tectonic adjustments during basin(s) infill, and thus, changes in the contribution of the source areas. At first glance, when combining the youngest detrital zircon populations of the Las Ventanas and Playa Hermosa formations at their respective type areas, we can conclude that the maximum age of the deposition is ~580 Ma. However, if we also consider the data from conglomerates (Las Ventanas Formation) near Minas city, the maximum age can be further constrained to ~570 Ma. Both ages are within error of that obtained in basal volcaniclastic rocks of the Las Ventanas Formation at its type area dated by U–Pb SHRIMP to 573 ± 11 Ma (Oyhantçabal et al., 2007) and could simply reflect different stratigraphic positions or source areas. The intrusive mycrosientic dyke dated by Rapalini et al. (2015) at 574.5 ± 8.1 Ma provides the best approximation to the minimum age of deposition for the entire group. Hence, we suggest that the deposition of the group occurred between ~580 and 566 Ma, as indicated by the maximum age for the Playa Hermosa strata and the approximate oldest age possible for the basal volcanoclastics of the Las Ventanas Formation, and the youngest age possible for the intrusive mycrosyenite, respectively.

5.1.4. San Carlos formation

The age pattern for the San Carlos Formation shows a strong contribution from a Neoproterozoic basement that can be correlated with the gneissic-migmatitic and intrusive rocks of the Cuchilla Dioniso Terrane (Table 1). Although the Cuchilla Dionisio Terrane and the Nico Pérez-Piedra Alta terranes were amalgamated in the mid-Ediacaran and before the deposition of the San Carlos Formation (Pecoits et al., 2004, 2008; Oyhantçabal et al., 2009), the absence of typical ages recorded in the Piedra Alta and Nico Pérez terranes strongly suggests that these blocks were not source areas for the sediments of the San Carlos Formation. However, the occurrence of Meso- and Paleoproterozoic ages account for some contribution from the Edén Terrane, which typically shows the same ages (Table 1). A maximum depositional age for the unit is constrained by the youngest zircon cluster at 552 ± 3 Ma. Showing similar features to the Playa Hermosa and Las Ventanas formations, the San Carlos Formation was informally included in the Maldonado Group until new data became available (Pecoits et al., 2004). In this regard, two scenarios were proposed by the same authors: (1) the San Carlos Formation, along with the Las Ventanas and Playa Hermosa formations, were deposited in the same basin, which subsequently was dismantled by the Sierra Ballena Shear Zone (Fig. 2), or (2) the Maldonado Group *s.s.* and the San Carlos Formation, although most likely contemporaneous, were deposited in different basins. Based on the new data presented here, we suggest that the San Carlos Formation was deposited in a different basin whose main source was its own basement. Taking into account the youngest zircon cluster observed in the San Carlos Formation (552 ± 3 Ma), the initiation of deposition occurred later than that in the Playa Hermosa and Las Ventanas formations, which is constrained between ~580 and 566 Ma (see section 5.1.3. above).

5.1.5. Barriga Negra formation

The Barriga Negra Formation shows a zircon age spectra typical of that found in the basement of the Nico Pérez Terrane, ranging from ~1750 to 3400 Ma (Table 1). Similar to samples from the Las Ventanas Formation, there are some few grains (three) between ~1000 and 1500 Ma. The two zircons with ages at ~1300–1500 Ma could be sourced from the Edén Terrane, while that at ~1000 Ma probably corresponds to reworking of the underlying Arroyo del

Soldado Group outcropping in the area. A Cuchilla Dionisio source for this zircon is discounted based on the lack of the more widespread Neoproterozoic ages found in this terrane (Table 1 and Fig. 2). The youngest zircon population constrains the maximum depositional age of the Barriga Negra Formation to ~581 ± 6 Ma, which fits well within the chronostratigraphic framework established for the Las Ventanas and Playa Hermosa formations. As for the Maldonado Group, the San Carlos Formation and other Ediacaran units (e.g., Sierra de Aguirre Formation), the ubiquitous occurrence of Ediacaran zircons clearly suggests the presence of active magmatic activity at that time and the contemporaneous development of extensional basins, some of them accompanied with important volcanic activity in both the Nico Pérez and Cuchilla Dionisio terranes (see section 5.2 below).

Recent work by Aubet et al. (2014) showed compelling evidence demonstrating the existence of an angular unconformity between the slightly deformed Barriga Negra Formation (above) and the strongly deformed Arroyo del Soldado Group (below), and thus both units are demonstratively unrelated (see also Fragoso-Cesar et al., 1987; Fambrini et al., 2005). Therefore, the age presented here constraining the maximum depositional age for the Barriga Negra Formation (581 ± 6 Ma), also helps to constrain the minimum age for the Arroyo del Soldado Group. A similar maximum age constraint of ~589 Ma for the Barriga Negra Formation was presented by Blanco et al. (2009). Although the data presented by the authors have been strongly questioned (see for example Sánchez-Bettucci et al., 2010; Zimmermann, 2011; Aubet et al., 2014) some general comparison can be made. Besides the youngest cluster (~589 Ma), Blanco et al. (2009) reported a couple of important peaks at 1795–1,723, 631–566, and 2890–3155 Ma with only two grains around 2157–2284 Ma. As described above, our zircon age spectra is also characterized by Paleoproterozoic and Archean ages but it is noteworthy the absence of the 1795–1723 Ma cluster reported by Blanco et al. (2009). The sample dated by these authors comes from the middle-lower part of the Barriga Negra Formation, while our sample was taken from the upper most part of the unit. This difference, together with the sedimentological features of the succession, indicates active tectonism and changing source areas during deposition. In this scenario, the lower section of the Barriga Negra Formation would be dominated by Paleoproterozoic zircons sourced from the Edén Terrane (~1740–1750 Ma, Campanero Unit), located to the southeast (Fig. 3). Paleocurrent measurements obtained in the succession further support this source area (Fambrini et al., 2005). In contrast, the upper half of the unit is dominated by Ediacaran zircons (~581 Ma) and early Paleoproterozoic and Archean zircons from local granites and basement rocks of the Nico Pérez Terrane, respectively (Fig. 3).

5.2. Chronostratigraphy and geological setting

The evolution of the Neoproterozoic Dom Feliciano Belt in Uruguay has been subject of many geotectonic interpretations. Proposals have included subduction models towards the east and the west accompanied by the development of one or more volcanic arcs. For example, Fragoso-Cesar (1980) proposed subduction to the west, which would have produced a magmatic arc represented by the granite domain (Aiguá Batholith) of the Cuchilla Dionisio Terrane (Fig. 1). Based on this model, it was later suggested a back arc basin represented by the Neoproterozoic supracrustal succession (schist domain) was located to the west of the magmatic arc (Fernades et al., 1992; Sánchez-Bettucci and Ramos, 1999). In contrast, other models suggested the existence of a passive margin represented by the Arroyo del Soldado Group, which evolved from rift deposits of the Maldonado Group (e.g., Gaucher et al., 2008; Blanco et al., 2009). In an attempt to reconcile this inconsistency, Basei et al. (2000) suggested that the strata assigned to a passive margin setting (Arroyo del Soldado Group) were deposited instead in a foreland basin along with the volcano-sedimentary Maldonado Group, and that the rest of the surpacrustal rocks belonging to the schist domain (e.g., Fuente del Puma Formation) were produced in a different context to that of the magmatic arc (see also Pecoits et al., 2008). Hence, Basei et al. (2000, 2005, 2008) proposed eastward subduction toward the Kalahari Craton building up the granite domain, and that this granite belt (i.e., magmatic arc) and the schist domain were juxtaposed later during a late Brasilian Event. Oyhantçabal et al. (2007) proposed a west-directed subduction and suggested that slab beak-off was the most likely mechanism associated with the generation of the granitic magmas shortly after the collision, and thus the granites (Aiguá Batholith) were not part of the magmatic arc.

In a recent study, Aubet et al. (2014) concluded that the stratigraphic model suggesting the evolution from a rift basin (Maldonado Group) to a passive margin setting (Arroyo del Soldado Group) is not supported by stratigraphic relations. Indeed, when detrital zircon age spectra of the Playa Hermosa and Las Ventanas formations (Maldonado Group) are compared to those of the Yerbal Formation (Arroyo del Soldado Group), it is clear that differences in their tectonic settings exist. The Maldonado Group record important Ediacaran zircon clusters, whilst the Yerbal Formation does not record ages younger than ~1036 Ma (Gaucher et al., 2008; Blanco et al., 2009; this study). This is not a trivial aspect given the voluminous record of Ediacaran magmatism in the region, particularly in the Nico Pérez Terrane, which is the basement of the Arroyo del Soldado Group. Furthermore, this magmatism is accompanied by marked strike-slip tectonics, which also contradicts the idea of an Atlantic-type continental shelf during the Ediacaran-lowermost Cambrian, as proposed for the Arroyo del Soldado Group (Gaucher et al., 2008; Blanco et al., 2009). Therefore, our data indicates that the Maldonado Group is younger than the Arroyo del Soldado Group. The same conclusion was reached by Aubet et al. (2014), where the authors presented an extensive review of the stratigraphy, radiometric ages and existing models for the origin of those basins.

As shown in the present study, this tectono-magmatic activity is clearly recorded in the Maldonado Group, and in the San Carlos and Barriga Negra formations as evidenced by the presence of Ediacaran zircon ages in all these units - these are totally absent in the Arroyo del Soldado Group. With the exception of the San Carlos Formation, which is developed on the Cuchilla Dionisio Terrane, detrital zircons between ~1000 and 650 Ma are lacking in all the units studied here. This would suggest that magmatism was absent during that time in the Nico Pérez Terrane. Significantly, zircons younger than 1036 Ma are also absent in the Yerbal Formation of the Arroyo del Soldado Group. This is further supported by the lack of magmatic ages at that time in Piedra Alta and Nico Pérez terranes (Aubet et al., 2014), which explains their absence in the Barriga Negra Formation, Maldonado and Arroyo del Soldado groups. When U–Pb ages of the basement are considered (Table 1), it appears as though the last magmatic events (prior to deposition of the Arroyo del Soldado Group) in the Nico Pérez and Edén terranes occurred at ~1785 and ~1430 Ma, respectively (Fig. 6). This is further supported by considering all the available U–Pb, Rb–Sr and K–Ar ages of the basement, which shows that the same gap exists Aubet et al. (2014; their Fig. 5). Similarly, the first magmatic event after deposition of the Arroyo del Soldado Group is represented by the Puntas del Santa Lucía Batholith at ~633 Ma (Table 1). Hence, we suggest that the mixed carbonate-siliciclastic Arroyo del Soldado Group was deposited in a shallow epicontinental sea that transgressed the Nico Pérez Terrane sometime between ~1000 and 650 Ma (Fig. 6).

A radically different tectonic setting has been proposed for the

Playa Hermosa, Las Ventanas, San Carlos and Barriga Negra formations. Masquelin and Sánchez-Bettucci (1993) suggested that Las Ventanas and Playa Hermosa formations were deposited in a pull-apart basin. Later, Pecoits et al. (2008) proposed a strike-slip system for the Maldonado Group and the San Carlos Formation (see also Sánchez-Bettucci et al., 2009). A similar tectonic regime was also indicated for the Sierra de Aguirre (Campal and Schipilov, 2005) and Barriga Negra formations (Fambrini et al., 2005). A comparable scenario has been proposed for the Ediacaran-Early Cambrian (605–530 Ma) basins (e.g., Camaquã, Castro and Itajaí basins) in southern Brazil (Almeida et al., 2010). However, those authors suggested an extensional origin for this basin system and concluded that the strike-slip deformation occurred after basin formation, in the Early Cambrian. In Uruguay, the Ediacaran basins are spatially related to major strike-slip faults and seem to be directly related to the tangential collisional event. Furthermore, typical structures found in strike-slip basins, such as post-diagenetic inverse faulting in the Playa Hermosa Formation (Masquelin and Sánchez-Bettucci, 1993), apparent migration of the depocentre in the Las Ventanas Formation (Pecoits et al., 2011), and the rhombohedral shape of the Sierra de Aguirre Basin (Campal and Schipilov, 2005) have all been suggested. Most importantly, and as shown by radiometric ages, the synchronicity between deposition and regional shearing, at least for the best-constrained basins, seems clear. Therefore, we suggest that the deposition of the Maldonado Group (566–580 Ma), Sierra de Aguirre Formation

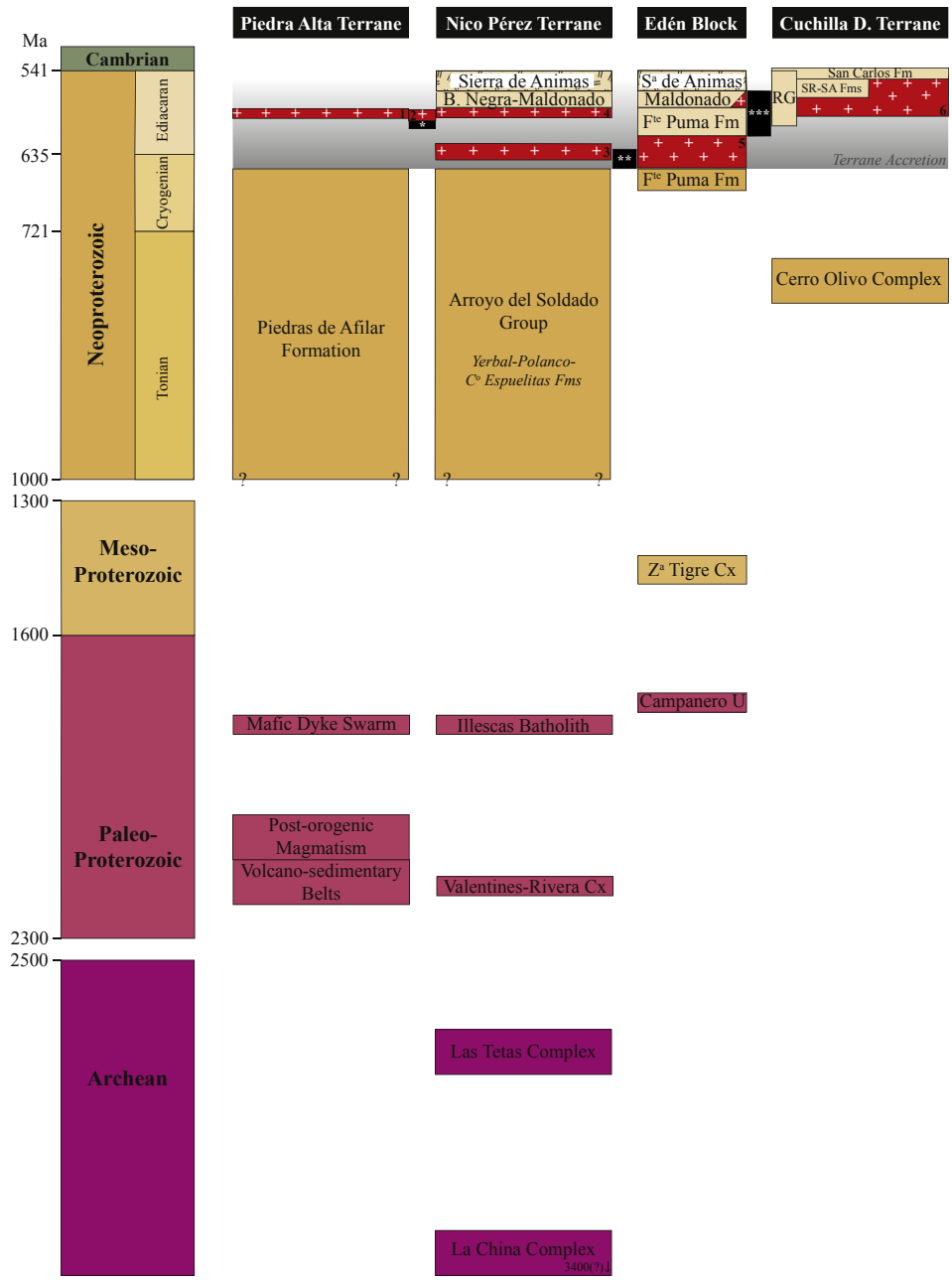

Fig. 6. Stratigraphic columns for tectono-stratigraphic terranes of Uruguay. Time scale from Gradstein et al. (2012). Granites: (1) La Paz; (2) Solís de Mataojo; (3) Puntas del Santa Lucía; (4) Sobresaliente, Mangacha, Las Flores; (5) Mataojo; (6) Tacuarí, Aiguá, Florencia, Maldonado. FP: Fuente del Puma Formation; RG: Rocha Group; SR-SA: Sierra de Ríos and Sierra de Aguirre formations. Accretionary events: (*) Piedra Alta; (**) Edén; (***) Cuchilla Dionisio. Note that the Piedras de Afilar and Arroyo del Soldado boxes do not represent the time elapsed during sedimentation but the period of time within which both units were deposited.

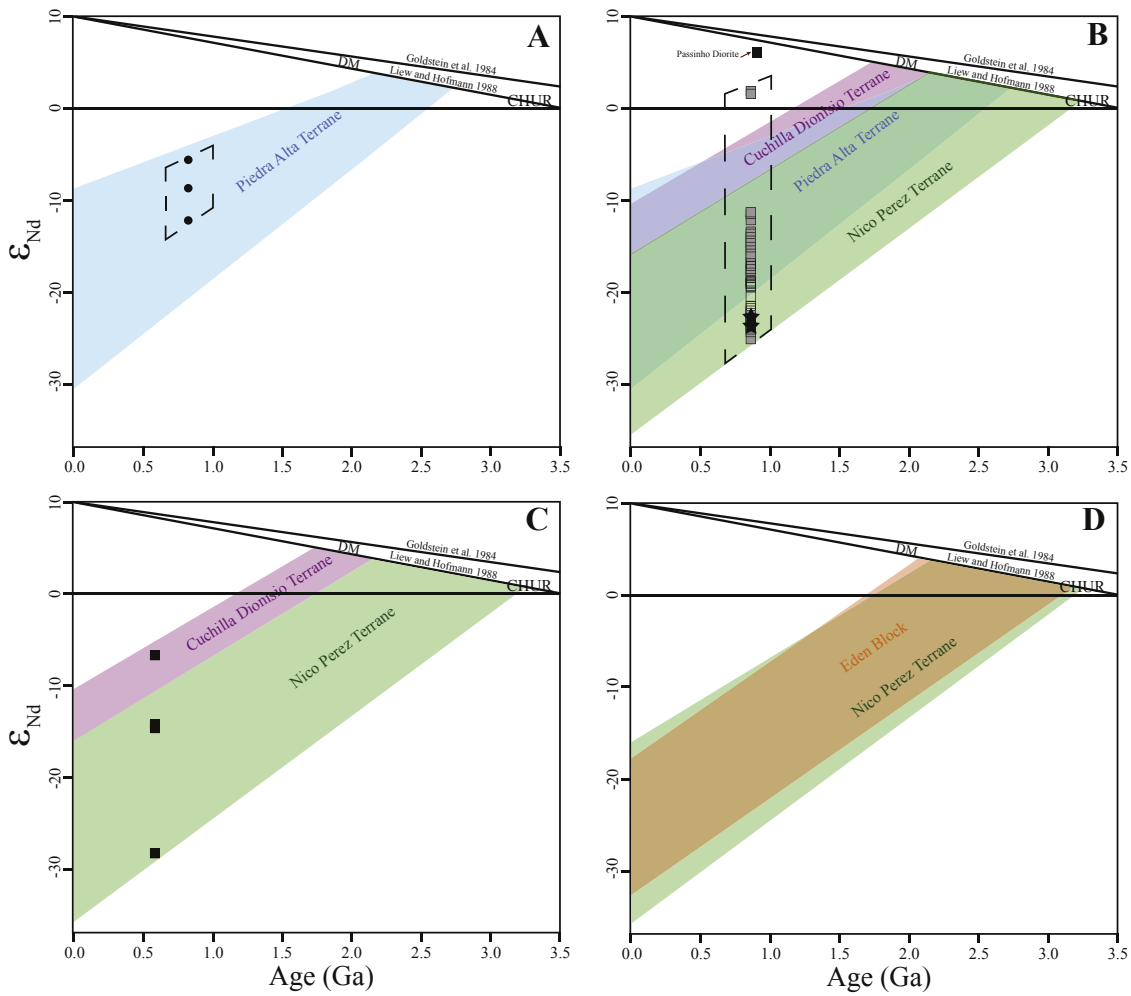

Fig. 7. Diagrams of εNd(t) vs. time (Ma). Compilation of Sm–Nd isotope compositions for **(A)** Piedras de Afilar Formation (n = 3); **(B)** Yerbal Formation (grey squares, n = 76) and Polanco Limestones (empty squares; n = 22) formations (Arroyo del Soldado Group), and Cerros San Francisco Formation (black stars, n = 2; Arroyo de la Pedrera Group); **(C)** Barriga Negra Formation (n = 4). Composition fields of metasedimentary and metaigneous rocks from Piedra Alta, Nico Pérez and Cuchilla Dionisio terranes are shown for reference. **(D)** Composition fields from Edén and Nico Pérez terranes. Data in A and B were plotted arbitrarily at 850 Ma but would plot somewhere within the dash-outlined areas between ~650 and 1000 Ma. Source of data: Bossi et al., 1993; Mazzuchelli et al., 1995; Rivalenti et al., 1995; Cingolani et al., 2002; Pankhurst et al., 2003; Gastal et al., 2005; Preciozzi and Peel, 2005; Peel and Preciozzi, 2006; Mallmann et al., 2007; Blanco et al., 2009; Oyhantçabal et al., 2011; Frei et al., 2011, 2013. DM: depleted mantle evolution lines after Goldstein et al. (1984) and Liew and Hofmann (1988). CHUR: chondrite uniform reservoir after DePaolo and Wasserburg (1976).

(571 ± 8 Ma), and Barriga Negra Formation (<581 ± 6 Ma) overlaps with the synkinematic granites, ranging from 584 ± 13 (Solís de Mataojo Granite) to 564 ± 7 Ma (Maldonado Granite), which support that these faults were active during basin infill.

One crucial aspect recognized by Aubet et al. (2014) is the difficulty in distinguishing between volcanic rocks previously assigned to the older volcanic pulse of the Sierra de Animas Complex, which were thought to be interdigitated with the Maldonado Group, and those traditionally assigned to Neoproterozoic Fuente del Puma Formation (ex Lavalleja Group) in the type area of the Las Ventanas Formation (Fig. 2). Unfortunately, the geochronological data from the Fuente del Puma Formation is scarce. Basei et al. (2008) reported U–Pb SHRIMP detrital zircon ages from one quartz-sericite schist of this unit (Fuente del Puma Formation) showing seven Archean zircons (2600–3400 Ma), six between (1780–2400 Ma), and five Neoproterozoic ages between 600 and 1060 Ma, with three values close to the youngest age (data not presented). However, Oyhantçabal et al. (2005) pointed out that zircon ages younger than 650 Ma are highly discordant, and thus only one single zircon dated at 715 ± 26 Ma would constrain the maximum depositional age of the unit. Sánchez-Bettucci et al. (2003) also reported some preliminary conventional U–Pb ages on zircon (four analyses) and one rutile from a metabasalt of this unit. According to the same authors (data not presented), it yielded a possible crystallization age of 667 ± 4 Ma and a metamorphic age of 624 ± 14 Ma. Hence, by combining the magmatic, metamorphic and detrital zircon ages reported for the Fuente del Puma Formation, magmatism and deposition seems to have occurred roughly between 660 and 690 Ma. Although discordant, the data presented by Basei et al. (2008) showing detrital zircon ages between 650 and 600 Ma could imply the presence of lithodemic units with different ages within the formation (Fig. 6). This would be in agreement with the apparent transition between volcanic rocks of the Fuente del Puma and Las Ventanas formations (Aubet et al., 2014; see below) and the postulated proximity of the Nico Pérez Terrane; the latter being the only potential source of the Archean detrital zircons present in the sample.

At its type area, the Las Ventanas Formation overlies the Mesoproterozoic Zanja del Tigre Complex with a marked angular unconformity, whereas its relation with the underlying volcanic and sedimentary rocks traditionally assigned to the Fuente del Puma Formation is less clear (Fig. 2). In the first instance, this basement-

cover relationship is clearly reflected in the detrital zircon spectra of the Las Ventanas Formation, which contains Mesoproterozoic zircon grains of the Zanja del Tigre Complex (see above). However, this is not the case for the Fuente del Puma Formation; i.e., being the Fuente del Puma Formation the basement of the Las Ventanas Formation, this is not reflected in the detrital age spectra of the latter. In this sense, there is a clear absence of magmatism between 1000 and 590 ± 5 Ma in the Las Ventanas Formation at its type area where the relationships with its basement (Fuente del Puma Formation and Zanja del Tigre Complex) is observed. Interestingly, Oyhantçabal et al. (2007) obtained an U–Pb SHRIMP age of 573 ± 11 Ma for volcaniclastic rocks of the Fuente del Puma Formation immediately underlying the Las Ventanas Formation. This age is similar to that reported by Mallmann et al. (2007) at 590 ± 2 Ma in gabbros also correlated by the same authors to the Fuente del Puma Formation. These ages are recorded in the detrital zircon spectra of the Las Ventanas Formation and indicate that the magmatism assigned to the Fuente del Puma Formation took place in two stages, approximately at 667, and from 590 to 560 Ma. Therefore, it can be concluded that while the first stage corresponds to the Fuente del Puma Formation in the strict sense, which did not constitute a source area for the Las Ventanas Formation, the younger magmatism is coeval with the deposition of the Las Ventanas Formation and would rather be part of this unit (this package corresponds to the Fuente del Puma box immediately underlying the Maldonado Group in Fig. 6).

Although further geochronologic studies are necessary to better constrain the Fuente del Puma Formation (*sensu stricto*; i.e., older box in Fig. 6), these ranges agree with the magmatic and detrital zircon data presented by Sánchez-Bettucci et al. (2003), Mallmann et al. (2007), Oyhantçabal et al. (2007) and Basei et al. (2008), and with the detrital zircon spectra reported in this study for the Las Ventanas Formation. We suggest that the volcanic rocks of the Las Ventanas Formation, previously assigned to the Fuente del Puma Formation, might represent the base of the rift basin wherein the Las Ventanas Formation was later deposited, accompanied and capped by volcanism of the Sierra de Animas Complex. A similar scenario was described for contemporaneous Ediacaran basins in southern Brazil where age constraints demonstrated that the Brazilian basins were formed between 605 and 530 Ma, and they record two major periods of basin formation and volcanic activity from 605 to 570 Ma and from 550 to 535 Ma (Almeida et al., 2010). The first period is characterized by thick volcano-sedimentary successions related to basic and intermediate volcanic rocks, with minor acid volcanics, and the second period is characterized by thick siliciclastic successions and discrete events of acid volcanism (Almeida et al., 2010). If the correlation suggested here between volcanic rocks once considered part of the Fuente del Puma Formation and basal volcanics of the Las Ventanas Formation is confirmed, an extensional origin for this basin, as suggested for the Ediacaran-Early Cambrian in southern Brazil (Almeida et al., 2010), is the most likely scenario.

5.3. Accretion of the Piedra Alta and Nico Pérez terranes

One important question regarding the configuration of the Río de la Plata Craton is, when did the Piedra Alta and Nico Pérez terranes accrete? This craton has been long considered as a single unit that was in place at least ~1200 Ma (Bossi and Ferrando, 2001). However, in a recent review, Oyhantçabal et al. (2011) did not include the Nico Pérez Terrane in the Río de la Plata Craton and pointed out that the accretion age is poorly constrained between approximately 1750 and 600 Ma. The last important movement of the Sarandí del Yí Shear Zone, which separates the Piedra Alta and Nico Pérez terranes, is geochronologically constrained (using U–Pb on sphene) by the age obtained for the emplacement of the syntectonic Solís de Mataojo Granitic Complex at 584 ± 13 Ma (Oyhantçabal et al., 2007, Figs. 1 and 2). The Paleo- or Mesoproterozoic age for accretion is based on the idea that the Piedra Alta mafic dyke swarm, dated at 1790 ± 5 Ma (U–Pb on baddeleyite; Halls et al., 2001), is affected by the Sarandí del Yí Shear Zone where the apparent curvature of the eastern end of the dyke swarm is consistent with a dextral shear sense (Bossi and Campal, 1992) and not with the sinistral sense recorded during the intrusion of the Ediacaran Solís de Mataojo Granitic Complex (Oyhantçabal et al., 2007). Low-grade hydrothermal overprint in two felsic veins and one associated dyke dated between 1370 and 1170 Ma (Ar–Ar and Rb–Sr; Teixeira et al., 1999) and one K–Ar age of 1253 ± 32 Ma (data not presented by the authors) for a synkinematic muscovite located on a thrust plane in the Nico Pérez Terrane have been interpreted to support the accretion of both terranes during the Mesoproterozoic (Bossi and Cingolani, 2009).

Extensive fieldwork has not confirmed the curvature of the Piedra Alta mafic dykes. The problem is probably rooted in the fact that most of these mafic dykes were identified in aerial photographs and only a small fraction could actually be recognized in the field. Hence, numerous dykes and the mentioned curvature simply correspond to photo-lineaments whose origins are not clear. The only unit that seems to have been dextrally deflected is the Cuchilla Cabo de Hornos Shear Zone (Spoturno et al., 2004, Fig. 2). This shear zone is composed of mylonites deriving from granitic and parametamorphic protoliths of the Piedra Alta Terrane, and although no ages are available, a Paleoproterozoic age for the mylonites was inferred (Spoturno et al., 2004); the age of deflection and its relationship with the Sarandí del Yí Shear Zone remains unknown. Furthermore, the use of few K–Ar and Rb–Sr ages on both sides of the Sarandí del Yí Shear Zone to place constraints on the age of accretion is very speculative. Teixeira et al. (2002) hypothesized that reactivation of the fault systems, which controlled the emplacement of the dyke swarm, could have been the probable mechanism through which low temperature hydrothermal fluids disturbed the K–Ar and Rb–Sr isotopic systems. Unfortunately, the nature and extension of this event remains unclear. Similarly, the K–Ar age reported from a pegmatite in a thrust plane in the Nico Pérez Terrane is located more than 10 km away from the Sarandí del Yí Shear Zone and not in the transcurrent shear itself; the relationship between these two structures also remains unknown (Oyhantçabal et al., 2011).

The analysis of detrital zircon spectra from Neoproterozoic sedimentary successions deposited on both sides of the Sarandí del Yí Shear Zone can shed some light on this issue. The Piedras de Afilar Formation is the only known sedimentary succession deposited on the Piedra Alta Terrane (Fig. 2). As explained below, the lack of Neoproterozoic detrital zircons, the absence of detrital zircons typically found in basement rocks of the neighboring Nico Pérez Terrane (located to the east), and the lack of evidence indicating a source from the western margin of the Río de la Plata Craton (proto-Andes) suggests that the Piedras de Afilar Formation was most likely deposited sometime between 1000 and 650 Ma, and that the Piedra Alta and the Nico Pérez terranes were not attached at that time. Previous proposals considered the Piedras de Afilar Formation as Ediacaran in age, while its Mesoproterozoic zircons (1000–1500 Ma) were likely sourced from the proto-Andean Mesoproterozoic Belt (Gaucher et al., 2008). However, the available evidence does not support this interpretation. If an Ediacaran age is considered, the absence of Neoproterozoic (pre-Ediacaran) zircons in the Piedras de Afilar Formation, which are extensively represented in the Nico Pérez Terrane, cannot be explained. Moreover, the extensive magmatic and tectonic activity recorded in the Ediacaran is not compatible with the stable tectonic

setting for the Piedras de Afilar Formation (Pecoits et al., 2008). While the maximum depositional age for the Piedras de Afilar Formation is constrained by the youngest detrital zircon cluster recorded at ~1000 Ma, its minimum depositional age is placed at 584 ± 13 Ma, which corresponds to the synkinematic magmatism recorded in the Sarandí del Yí Shear Zone. Additionally, the absence of Neoproterozoic (≤650 Ma) detrital zircons mirroring source areas from the once approaching Nico Pérez Terrane (i.e., the proximity of this terrane should have caused input of ≤650 Ma zircons), suggests a minimum depositional age of ~650 Ma.

The Piedras de Afilar Formation shows 99% of the zircons with ages between 1000–2070 Ma (Gaucher et al., 2008). Although one can argue that some of these zircons could be sourced from the Nico Pérez Terrane, the lack of typical Archean and Paleoproterozoic zircons in the Piedras de Afilar Formation, which are common in the Nico Pérez Terrane, argues against a source from this terrane. Of particular importance for our analysis are the zircon clusters between 1000 and 1900 Ma. Older populations (~2000 Ma) most likely derived from the local basement of the unit (Table 1). For most of these Mesoproterozoic zircons a proto-Andean source was proposed (Gaucher et al., 2008). In this regard, the basement of the Andes is formed by different blocks showing Grenville affinities, such as Chilenia, Cuyania, Antofalla, Arequipa, among others (Ramos, 2010). Particularly important for our discussion are those blocks that were left in Gondwana after the break up of Rodinia and could be part of the source areas of the Piedras de Afilar Formation. Such is the case of Pampia, which remained attached to Amazonia after the Rodinia amalgamation and never detached from it (Ramos et al., 2010). The Pampia terrane was part of the Rodinia supercontinent during the Mesoproterozoic with a Neoproterozoic ocean in its eastern side, which separated the Pampean terrane from the Río de la Plata Craton (Pampean, Brasilian or Clymene Ocean) (Rapela et al., 2007; Cordani et al., 2009, Fig. 8A). Escayola et al. (2007) pointed out that the western part of this terrane has a strong Grenville-age source while the eastern margin shows a Brasiliano source with ages between 700 and 600 Ma. They suggested a west-dipping subduction starting at ~700–650 Ma between the Pampia and Río de la Plata blocks, with the development of an island arc and associated back arc basin. This Neoproterozoic magmatic arc system was the source area of the Neoproterozoic zircons found in the pampean metasediments. Further to the north, correlative rocks of the Puncoviscana Formation (late Neoproterozoic-Early Cambrian) and the overlying Mesón Group (Late Cambrian) also show a predominance of Late Neoproterozoic zircons characteristic of the widespread Brasilian Cycle of south central and southeast Brazil (Adams et al., 2011). The presence of an ocean basin separating the Pampia and the Río de la Plata cratons and the absence of Neoproterozoic zircons (<1000 Ma) in the Piedras de Afilar Formation, therefore, argue against proto-Andes being its source area. Furthermore, nearly all the Neoproterozoic-Cambrian samples reported by Escayola et al. (2007) and Adams et al. (2011) show Mesoproterozoic detrital zircon spectra different from those displayed by the Piedras de Afilar Formation.

On the other hand, Paleoproterozoic zircons are almost absent from the provenance patterns of the above-mentioned Argentinian Neoproterozoic successions, suggesting that there was little contribution from the Río de la Plata Craton (Escayola et al., 2007; Adams et al., 2011). In turn, the Late Cambrian Mesón Group shows abundant Paleoproterozoic zircons (Adams et al., 2011). This clearly suggests that the Río de la Plata and Pampia were not attached in the Neoproterozoic (e.g., Rapela et al., 2007; Ramos et al., 2010), and thus a proto-Andean source for the Piedras de Afilar Formation during the Ediacaran finds no support. In this regard, it has been suggested that Amazonia, Pampia and Arequipa-Antofalla cratonic blocks were part of the Rodinia supercontinent (Brito Neves et al., 1999; Kröner and Cordani, 2003) whereas São Francisco, Paranapanema and Río de La Plata cratons were separated from those by the Transbrasilian lineament (Cordani et al., 2009). In this model, the Neoproterozoic-Early Cambrian Araguaia-Paraguai-Pampean belts were formed (along this lineament) and represent the main suture for closing the Clymene (Brasilian) Ocean (Rapela et al., 1998; Escayola et al., 2007; Ramos et al., 2010). Hence, the eastern margin of Pampia represents the southern extension of the Transbrasilian lineament, which juxtaposed the Río de La Plata Craton and Pampia (Ramos et al., 2010, Fig. 8). Differences exist in terms of the tectonic juxtaposition model. According to Rapela et al. (2007), the Río de la Plata craton was juxtaposed obliquely to the Pampean orogen, whereas Escayola et al. (2007) and Ramos et al. (2010) favored an orthogonal collision. In any case, the Río de la Plata Craton likely accreted by the Early Cambrian when it became the source for Paleoproterozoic detrital zircons in Mid-to Late Cambrian and Ordovician successions (Adams et al., 2011; Verdecchia et al., 2011). Later during the Late Cambrian-Early Ordovician, a significant displacement of the Pampia terrane with respect to the Río de la Plata (and other Gondwana cratons) through a dextral strike-slip shear positioned both cratons into their final position (Spagnuolo et al., 2012). Therefore, the detrital zircon ages gathered from the Piedras de Afilar Formation further support the allochtony of the Piedra Alta Terrane (Río de la Plata Craton after Oyhantçabal et al. (2011)) with respect to Pampia and the Nico Pérez terranes.

Further evidence indicating the exotic nature of the Piedra Alta Terrane with respect to the Nico Pérez Terrane comes from the detrital age spectra of the Arroyo del Soldado Group, located to the east of the Sarandí del Yí Shear Zone (Figs. 1 and 2). Considering the available U–Pb ages from the Nico Pérez basement, nearly all the detrital zircons from the Arroyo del Soldado Group can be readily explained as being sourced from it (see section 5.1). The only exception is the relatively small late Mesoproterozoic cluster (~1050 Ma). Rocks of this age are unknown in the Nico Pérez Terrane, and thus three possible explanations are put forward here to explain the presence of these zircons in the Arroyo del Soldado Group: (1) Mesoproterozoic (~1050 Ma) rocks have not yet been found in the Nico Pérez Terrane, (2) they were eroded away, or (3) these zircons were sourced from a now distant Mesoproterozoic belt. For the same reasons mentioned above for the Piedras de Afilar Formation, a Proto-Andean Mesoproterozoic source on the western side of the Río de la Plata Craton cannot explain the zircons found in the Arroyo del Soldado Group. A Piedras de Afilar source (reworking) can be also discounted for these Mesoproterozoic zircons because the Arroyo del Soldado Group does not record the same age spectrum shown by the Piedras de Afilar Formation. This implies that different areas sourced these units. It also supports the suggestion that the Nico Pérez and the Piedra Alta terranes were not accreted at the time of the Piedras de Afilar Formation and Arroyo del Soldado Group deposition.

Additional support for a Neoproterozoic accretion comes from Nd isotope data. In this regard, all the samples of the Piedras de Afilar Formation only plot within the range defined by basement rocks of the Piedra Alta Terrane (Fig. 7A). Conversely, the Yerbal and Polanco formations (Arroyo del Soldado Group), and the Cerros San Francisco Formation (Arroyo de la Pedrera Group) plot within the field of the Nico Pérez Terrane (Fig. 7B). While all the samples from the Arroyo de la Pedrera Group only plot within the field defined by this terrane some of the samples of the Yerbal and Polanco formations also overlap with that of the Piedra Alta Terrane. Nevertheless, the fact that not all of them do likely indicates that the source area for both of these units was the Nico Pérez Terrane. It is also noteworthy that the two northernmost samples of the Yerbal Formation show positive εNd(t) values and do not overlap with any

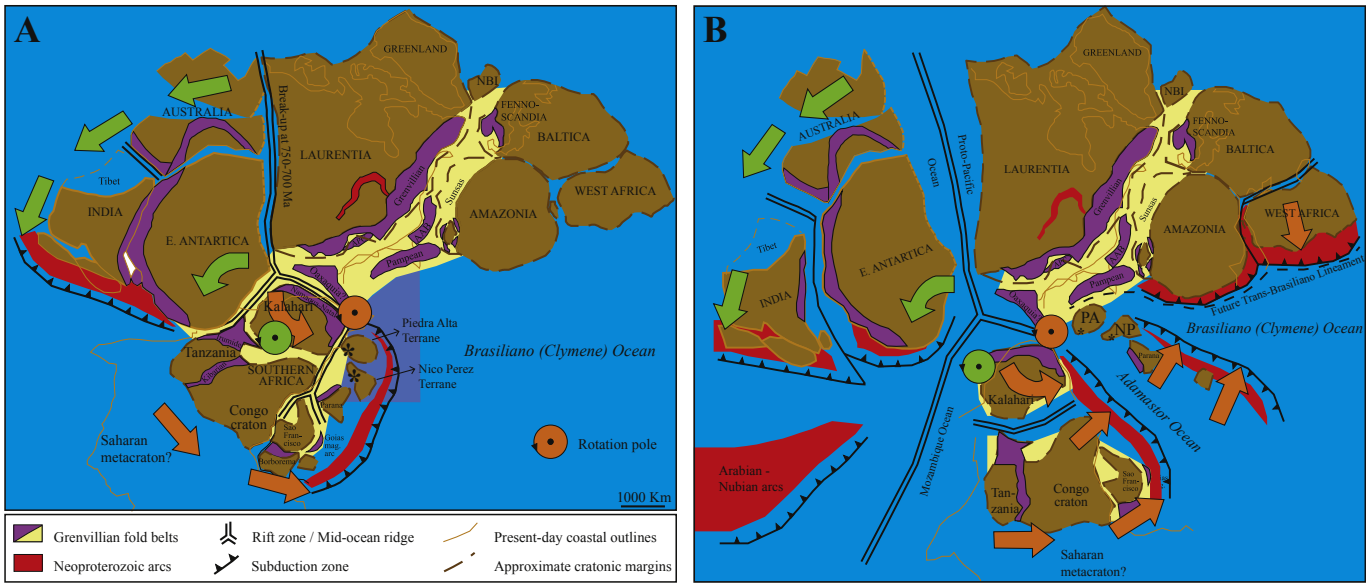

Fig. 8. **(A)** Reconstruction of Rodinia showing the initial break-up of its western part at ca. 850–750 Ma. Piedras de Afilar (Piedra Alta Terrane = Río de la Plata Craton) and Arroyo del Soldado (Nico Pérez Terrane) basins indicated by asterisks. The dark blue shaded area indicates possible location of Piedra Alta and Nico Pérez terranes in the Clymene Ocean (see text for explanation). **(B)** Shift from Rodinia to Gondwana ca. 750–650. Modified from Johansson (2014). AAB; Arequipa-Antofalla Basement. APC; Argentine Pre-Cordillera Basement. PA; Piedra Alta Terrane. NA; Nico Pérez Terrane. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

field defined by basement rocks of Uruguay (Fig. 7B). One possibility is that these rocks, attributed to the Yerbal Formation in the Isla Cristalina de Rivera (Fig. 1), do not belong to this unit. Another scenario is that the Yerbal Formation, at least it northernmost part, was sourced by rocks belonging to the São Gabriel Block in southern Brazil (Fig. 1). In contrast to the crustal reworking shown by other adjacent crustal blocks, including those located in Uruguay, Sm–Nd analyses of the São Gabriel block indicate a juvenile origin for this unit (Gastal et al., 2005; Saalmann et al., 2011). Particularly important for our analysis is the Passinho Diorite of the Cambaí Complex, with an age of ~880 Ma (U–Pb SHRIMP, Leite et al., 1998), located in the southern part of the São Gabriel Block and which shows positive εNd(t) values from +5.8 to +6.3 (Leite, 1997) (Fig. 7B). In contrast to the Arroyo del Soldado Group (650–1000 Ma), the Ediacaran Barriga Negra Formation (<581 ± 6 Ma) shows evidence of Nico Pérez and Cuchilla Dionisio sources (Fig. 7C), which further supports the idea that both terranes were accreted in the Ediacaran, but before the deposition of this unit.

As stated above, the only dextrally deflected unit in the Piedra Alta Terrane is the Cuchilla Cabo de Hornos Shear Zone (Spoturno et al., 2004). Interestingly, Oyhantcabal (2005) suggested that the curvature shown by the Campanero Unit (Sierra de Cabral fold), located to the east of the Sarandí del Yí Shear Zone, represents the counterpart of the abovementioned structure and that both high temperature mylonitic foliations were rotated as a consequence of the activity of the dextral Sarandí del Yí Shear Zone (Fig. 2). Unfortunately, no radiometric ages have been available to place constraints on the timing of this deflection, which would represent the accretion of the Piedra Alta and Edén terranes. Based on the ages obtained in this study, we suggest that this event took place between ~650 and 585 Ma, the minimum depositional age for the Arroyo del Soldado Group and the Piedras de Afilar Formation, and the sinistral shear reactivation of the Sarandí del Yí Shear Zone, respectively. As discussed in more detail in the following section, the accretion of the Edén Terrane, where the Sierra de Cabral fold occurs, took place ~650–620 Ma ago. Therefore, if the observation made by Oyhantcabal (2005) holds true, this would indicate that the accretion of the Piedra Alta and Nico Pérez terranes occurred afterwards, approximately between 620 and 600 Ma (see section 5.4 below).

5.4. Paleogeographic and paleotectonic implications

Paleogeographic reconstructions of the Río de la Plata Craton for the Meso- and Neoproterozoic differ widely. For example, Li et al. (2008) placed the Río de la Plata Craton at the heart of Rodinia adjacent to the southern Grenville margin of Laurentia between the Congo-São Francisco Craton and Laurentia, close to the Amazon and Kalahari Cratons (cf. Santos et al., 2008). In turn, Johansson (2014) positioned the Río de la Plata Craton at some distance away from Laurentia and Amazonia, and separated from the latter by a wide Brasilian (Clymene) ocean embayment; i.e., not forming part of Rodinia. Although Johansson (2014) located the Río de la Plata Craton adjacent to the present-day west side of the Kalahari Craton, the author pointed out that a position in between this craton and the Laurentian margin cannot be ruled out (Fig. 8A). The location of the Río de la Plata Craton between Laurentia and Congo cratons by Li et al. (2008) follows the idea of Fuck et al. (2008), who mentioned the occurrence of Mesoproterozoic units to the west (Pampia) and east (Cuchilla Dionisio Terrane) of the Río de la Plata Craton at its present position (Ramos and Vujovich, 1993; Preciozzi et al., 1999b; Basei et al., 2000). In the latter case, it is noteworthy that the basement rocks of the Cuchilla Dioniso Terrane were formerly correlated with the Namaqua Belt of South Africa (Preciozzi et al., 1999b; Basei et al., 2005, 2011). However, new studies have suggested a correlation with the Coastal Terrane of the Kaoko Belt located at the southwestern margin of the Congo Craton (Gross et al., 2009; Oyhantçabal et al., 2009; Konopásek et al., 2014). Thus, the idea proposed by Fuck et al. (2008), and followed by Li et al. (2008), at least for the eastern margin of the Río de la Plata Craton (the relation between its western margin and Laurentia is strongly debated), finds support in more recent regional studies. Nonetheless, as previously suggested (Pecoits et al., 2004, 2008;

Oyhantçabal et al., 2009), and further supported in this study, the Cuchilla Dioniso Terrane attached to the Nico Pérez Terrane in the mid-Ediacaran, and the latter was attached to the Piedra Alta slightly later. Thus, the correlation between the Cuchilla Dioniso and Coastal terranes should not be used to place precise paleogeographic constraints on the Río de la Plata Craton before the Ediacaran when these terranes were independent blocks (Fig. 8A). In turn, Johansson (2014) placed the Río de la Plata Craton attached to the Kalahari Craton through the Namaqua-Natal Belt based on the presence of Mesoproterozoic rocks of the Edén Terrane (Zanja del Tigre Complex). As for the Cuchilla Dioniso Terrane, the Edén Terrane is allochtonous with respect to the Río de la Plata Craton (i.e., the Piedra Alta and Nico Pérez terranes), and thus cannot be used either for paleogeographic reconstructions of the craton. Although both Li et al. (2008) and Johansson (2014) positioned the Río de la Plata Craton close to the Kalahari and Congo-São Francisco cratons, Johansson (2014) considered that all these blocks were not part of Rodinia (cf. Li et al., 2008).

Although it is still premature to draw a definite Paleogeographic model, the ages obtained in this and previous studies (Oyhantçabal et al., 2011, 2012 and references therein) accompanied with updated geological mapping (e.g., Spoturno et al., 2004, 2012) and a better understanding of the stratigraphy of the Neoproterozoic successions and their basement (Aubet et al., 2014) allow us to set a baseline for future research. As stated above, the Piedra Alta and the Nico Pérez terranes were independent blocks until ~620–600 Ma. The presence of clearly different Mesoproterozoic zircon age spectra in the Piedras de Afilar Formation and Arroyo del Soldado Group indicates that these units had different source areas and were deposited in different basins. We suggest that the position of the Piedra Alta and Nico Pérez terranes in Rodinia was somewhere in the Clymene (Brasilian) ocean between the Laurentian and Kalahari and Congo-São Francisco Cratons as hypothesized by Johansson (2014) (Fig. 8A). In this context, both sedimentary successions (Piedras de Afilar Formation and Arroyo del Soldado Group) were deposited in stable tectonic settings and sourced from the west-northwest (at present coordinates), and so any link with Mesoproterozoic orogenic systems should be located in that direction with respect to the Piedra Alta and Nico Pérez terranes. Likewise, any model should consider a stronger influence of Mesoproterozoic sources in the case of the Piedras de Afilar Formation (Piedra Alta Terrane) when compared with the Arroyo del Soldado Group (Nico Pérez Terrane).

The source of the Mesoproterozoic zircons in the Arroyo del Soldado Group (1050 Ma) and the Meso- and upper Paleoproterozoic zircons (1000–1900 Ma) in the Piedras de Afilar Formation remains unknown. Regardless, these units clearly had different source areas, which further supports that the Piedra Alta and Nico Pérez terranes were allochtonous with respect to each other at that time. Although both units show typical Grenville ages (900–1300 Ma; Rivers, 1997), the Piedras de Afilar Formation also display detrital zircon populations between 1350 and 1900 Ma that are characteristic of Grenville basement areas, such as the Laurentian craton interior in North America (e.g., Thomas et al., 2004) or its counterpart at the southwestern corner of the Amazon Craton in South America. The latter is represented by the Ventuari-Tapajós (2000–1800 Ma), Rio Negro-Juruena (1780–1550 Ma) and Rondonian-San Ignacio (1550–1300 Ma) tectonic (Cordani et al., 2010) or geochronological (Santos et al., 2008) provinces, whose orogenic evolution has been compared to that of the North American Grenvillian Province (Cordani et al., 2010). However, Laurentia along with some cratonic blocks from South America, including Amazonia, Pampia and Arequipa-Antofalla, were part of the Rodinia supercontinent, whereas São Francisco, Paranapanema and Río de La Plata cratons (along with Kalahari and Congo cratons from Africa) were separated from those by the Transbrasilian lineament (Fig. 8B) (Brito Neves et al., 1999; Kröner and Cordani, 2003; Cordani et al., 2009). If this holds true, the Piedras de Afilar (Piedra Alta Terrane) and the Arroyo del Soldado (Nico Pérez Terrane) were not sourced from Rodinia-Grenvillian Provinces. Based on available paleomagnetic data from Africa and South America, Tohver et al. (2006) placed the São Francisco-Congo, Kalahari and Río de la Plata cratons very close to each other and likely forming a single continental mass that agglutinated more or less contemporaneously to Rodinia. Therefore, some of the Mesoproterozoic mobile belts of central Brazil (e.g., Espinhaço) and central-southern Africa, collectively known as Kibaran (e.g., Kibaran, Irumide, Namaqua, Natal, Rehoboth), formed between 1400 and 1000 Ma (Cordani et al., 2010 and references therein), are favored here as the sources of the Uruguayan successions (Fig. 8A).

The global reconfiguration from Rodinia to Gondwana (~850–650 Ma) has been described as a counterclockwise rotation of western and southern Rodinia, relative to a fixed Laurentia, Amazonia, Baltica and West Africa cratons with the concomitant opening of the Proto-Pacific ocean (west of Laurentia) and closure of the Brasilian (Clymene) ocean (Johannsson, 2014) (Fig. 8). It was during this reconfiguration that the Arroyo del Soldado Group was likely deposited and subsequently the basin was closed. On a more regional scale, Saalmann et al. (2011) suggested that following the subduction of oceanic crust beneath the Río de la Plata Craton margin in its northeastern side the collision of a microcontinent (Encantadas) with the Río de la Plata Craton and the São Gabriel Block (Fig. 1) took place at ~730–690 Ma (São Gabriel Event) (cf. Silva et al., 2015). The presence of this microcontinent and the São Gabriel Event in Uruguay as responsible for the closure of the Arroyo del Soldado Basin, however, has not been previously discussed.

The Encantadas Block (in southern Brazil) comprises Paleoproterozoic (~2200 Ma) basement rocks of the Encantadas Complex, the Porongos schist belt and to east, separated by the transcurrent Dorsal de Canguçu Shear Zone, the Pelotas batholith (Fig. 1). In Uruguay, both units have been correlated with the Fuente del Puma Formation (ex-Lavalleja Group) and the Aiguá batholith, respectively. The Sierra Ballena Shear Zone is considered the southern extension of the Dorsal de Canguçu Shear Zone. In the first case, when comparing the Fuente del Puma Formation and the Porongos belt it is clear that more radiometric determinations, particularly in the Fuente del Puma Formation (see above), are necessary to draw definite conclusions. The ages reported for the Porongos belt range from ~790 to 570 Ma (Porcher et al., 1999; Saalmann et al., 2011; Pertille et al., 2015), however, recent studies suggest that the duration of the Porongos Group basin filling probably lasted from 650 to 570 Ma (Pertille et al., 2015). Based on previously reported data, this range is similar to that proposed here for the Fuente del Puma Formation (see Fig. 6). The Pelotas batholith comprises several suites of 635–590 Ma granites (Fig. 1) and shows T_{DM} model ages between 2000 and 1300 Ma (Saalmann et al., 2011 and references therein). Similarly, the Aiguá Batholith, situated to the east of the Sierra Ballena Shear Zone, shows an age between 560 and 615 Ma (Table 1) and a TDM model age between 1700 and 2200 Ma (Fig. 7B–C).

In contrast to the basement of these units in Brazil (Encantadas Complex), when the basement located on both sides of the Sierra Ballena Shear Zone in southern Uruguay is considered, some clear differences arise. The Edén Terrane, located to the west, comprises a Paleoproterozoic (~1750 Ma) gneissic basement (Campanero Unit) and Mesoproterozoic supracrustal rocks of the Zanja del Tigre Complex (~1430 Ma). In contrast, the Cerro Olivo Complex, situated to the east, comprises ortho- and para-derived metamorphic rocks with ages between 750 and 800 Ma for the magmatism and

650–600 Ma for the high-grade metamorphic event. Additionally, when TDM model ages from both sides of the Sierra Ballena Shear Zone are compared they also show differences. The Edén Terrane displays a TDM model age roughly between 2100 and 3300 Ma, which overlap with those of the Nico Pérez Terrane; see Fig. 7D), while the Cuchilla Dioniso Terrane shows younger ages between 1700 and 2200 Ma (Fig. 7C). Other differences include: (1) the absence of inherited ~1000 Ma zircons in ortho-derived rocks to the west of the Sierra Ballena Shear Zone, which are common in the Cerro Olivo Complex located to the east (Table 1); (2) only granites occurring in the Cuchilla Dionisio Terrane (e.g., Valdivia granite) show a source of Pb with ages of ~1000 Ma (Oyhantçabal et al., 2007); and (3) only granites found in the Edén Terrane (e.g., Puntas del Mataojo Granite) have inherited zircon cores displaying concordant Paleoproterozoic/Archean ages (see for example Oyhantçabal et al., 2009). Therefore, while favouring a possible affinity between the Edén Terrane and the Nico Pérez Terrane (see below), these features argue in favour of a different tectonic evolution for the Edén and Cuchilla Dioniso terranes. Hence, the boundary between these two domains (Sierra Ballena Shear Zone) would represent a suture zone.

The apparent similarities found in the basement on both sides of the Dorsal de Canguçu Shear Zone in southern Brazil led Saalmann et al. (2011) to propose a common tectonic evolution, thereby arguing against this shear zone being a suture zone. In this regard, the significance of the Dorsal de Canguçu and Sierra Ballena Shear Zones has been a matter of debate. While some thought it to represent a suture between two lithospheric plates (e.g., Basei et al., 2000), others suggested that it represents an intra-continental fault (e.g., Fernandes and Koester, 1999). In Uruguay, the Sierra Ballena Shear Zone was argued to represent an intra-continental shear zone based on the occurrence of calc-alkaline granitic intrusions on both sides with ages in the range of 650–600 Ma (Oyhantçabal et al., 2009). Under close scrutiny, however, the age and spatial distribution of these granites show differences. For example, the oldest U–Pb SHRIMP ages (~630 Ma; Table 1) are only found in the Nico Pérez Terrane and in the Edén Terrane. Available geochemical data for one of these granites (Puntas del Mataojo Granite) indicates a high-K_2O calc-alkaline signature compatible with a magmatic arc or a post-collisional tectonic setting (Sánchez-Bettucci et al., 2003; Oyhantçabal et al., 2009). Here, we suggest that these granites were most likely emplaced after the closure of the Arroyo del Soldado Basin and as a result of the collision between the Edén and the Nico Pérez terranes (Fig. 9). This event, herein named the "Edén Accretionary Event" might also be reflected in U–Pb ages (~630 Ma) of metamorphic rutile in the Fuente del Puma Formation (see above). Following this collision, the Cuchilla Dionisio Terrane accreted, via the Sierra Ballena Shear Zone, to the Nico Pérez and the Edén terranes producing a second magmatic event between 615 and 560 Ma (Fig. 9), which is manifested in all those blocks (Cuchilla Dionisio Accretionary Event; Table 1). Therefore, the Sierra Ballena Shear Zone would represent a suture whose main activity in Uruguay spans from 580 to 560 Ma, as indicated by the occurrence of the associated transpressional magmatism (see also Oyhantçabal et al., 2009). Older granites located to the west of this shear zone are related to an earlier accretionary event (Edén Event), located further to the west, and should not be considered in dating the evolution of the Sierra Ballena Shear Zone, which is rather associated to the accretion of the Cuchilla Dionisio Terrane.

The provenance of the Edén Terrane is uncertain, although it's basement and that of the Nico Pérez Terrane might have the same affinity. This observation is supported by the overlapping TDM model ages of Neoproterozoic granites, gneissic basement and supracrustal rocks from both domains (Fig. 7D). As stated above, the fact that the coeval Illescas Rapakivi Granite in the Nico Pérez Terrane corresponds to an anorogenic setting, which did not experience high-grade metamorphism, as did the arc-related protoliths of the Campanero Unit of the Edén Terrane, suggests that both units drifted away before 1780–1750 Ma. This is further supported by the lack of Mesoproterozoic (~1430 Ma) rocks in the Nico Pérez Terrane which can be used to argue that the Paleoproterozoic and Archean detrital zircons recorded in the Mesoproterozoic (~1430 Ma) Zanja del Tigre Complex were derived from a domain other than the Nico Pérez Terrane. As shown in this study, the Edén Terrane was not part of the basement of the Arroyo del Soldado Group, and therefore, we suggest that these two blocks were attached in the Neoproterozoic likely during the lower Ediacaran-upper Cryogenian (~650–620 Ma). Although fundamental questions still exist with regards to the nature and age of the Fuente del Puma Formation (*s.s*), this unit occurs within the Edén Terrane (e.g., Oyhantçabal et al., 2005). Thus, the boundary between this allochthonous block and the Nico Pérez Terrane is placed at the La Plata Lineament (Oyhantçabal, 2005, Figs. 2 and 3). As demonstrated by the occurrence of basement windows belonging to the Zanja del Tigre Complex (e.g., Burgueño and COMSA quarries) and orthogneisses of the Campanero Unit in Punta Rasa (Fig. 2), this block extends to the south under the Maldonado Group and Sierra de Animas Complex (Piriápolis area) up to the Atlantic coast. The boundaries of this block coincides with the once informally defined but later abandoned Serrana Block (Gaucher et al., 2004).

Sometime around 600–550 Ma, the closure of the Adamastor Ocean and collision with the Congo (Kalahari?) craton took place (Fig. 9). Eastward (Basei et al., 2008) and westward (Oyhantçabal et al., 2007) subduction models have been proposed. In either case, the collisional history in southeastern Uruguay was characterized by NE-SW plate convergence, evident as sinistral transcurrent shearing mainly through the Sierra Ballena Shear Zone. The absence of relicts preserving high-pressure paragenesis further supports an oblique collision. Additionally, syntectonic granite emplacement along the sinistral Sierra Ballena Shear Zone (and its equivalents in southern Brazil) became locally more important. In Uruguay, synkinematic granitoids include, (1) the Maldonado granite (564 ± 7 Ma) emplaced between the Sierra Ballena and Cordillera Shear Zones; (2) the Aiguá granite (587 ± 16 Ma), an elongated pluton in contact with mylonites of the Sierra Ballena Shear Zone to the west and the Florencia Granite to the east; (3) the Florencia granite (572 ± 2 Ma), an elongated pluton located between the Aiguá and Valdivia Granites; (4) the Arroyo de los Píriz Granite (588 ± 6 Ma), an elongate body emplaced directly in the eastern Sierra Ballena Shear Zone; and (5) the Solís de Mataojo granitic complex (584 ± 13 Ma), an elongated intrusion related to the southern extreme of the Sarandí del Yí Shear Zone (Oyhantçabal et al., 2005, 2007; 2009) (Fig. 2). Contrary to previous proposals, which considered this granitic magmatism to be the root of a continental arc (see Basei et al., 2008), Oyhantçabal et al. (2007) suggested that slab beak-off was the most likely mechanism associated with the generation of the granitic magmas shortly after the collision, and thus these granites were not part of the magmatic arc. Furthermore, Oyhantçabal et al. (2007) suggested that this magmatism shows an evolution beginning with highly-fractionated calc-alkaline granites (e.g., Solís the Mataojo Granitic Complex ~584) followed by volcanics of shoshonitic affinity and concluding with magmas with alkaline signature (Sierra de Animas Complex). The signature of this magmatism indicates a post-collision setting during the activity of the Sierra Ballena Shear Zone, which played a major role in the emplacement of these magmatic associations.

Brittle deformation was associated with the Sierra Ballena Shear Zone. Its conjugate shears or faults are related to anisotropy of the

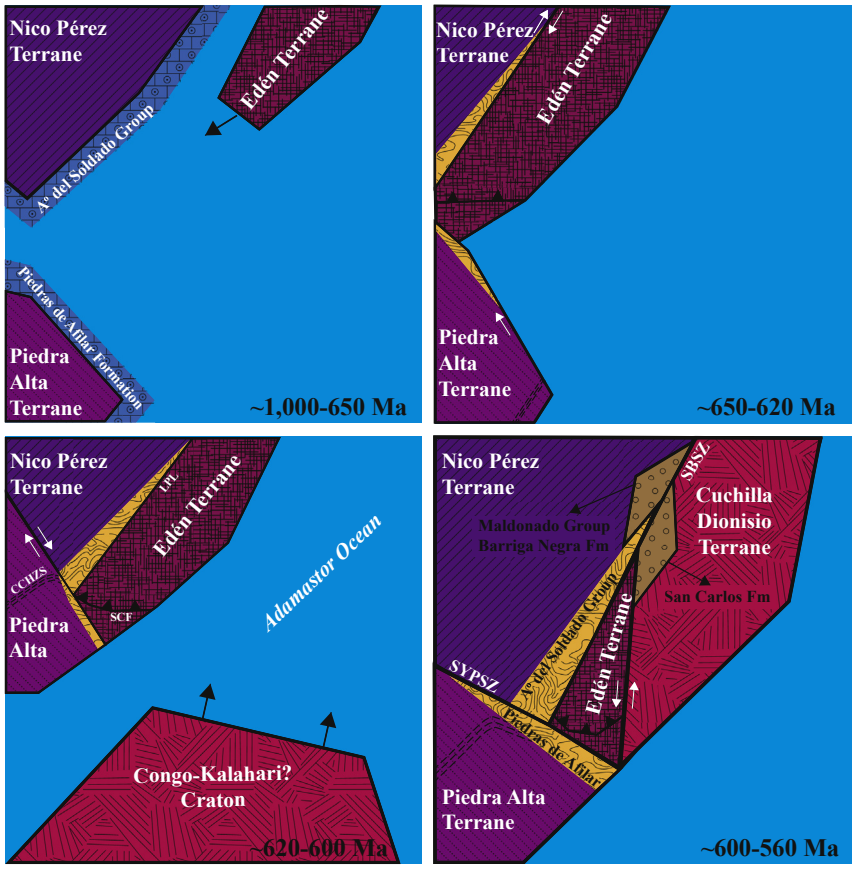

Fig. 9. Schematic reconstruction showing a possible scenario of episodic oblique terrane collisions during the Neoproterozoic (see text for details). CCHZS: Cuchilla Cabo de Hornos Shear Zone; SCF: Sierra de Cabral Fold; SYPSZ: LPL: La Plata Lineament; Sarandí del Yí-Piriápolis Shear Zone; SBSZ: Sierra Ballena Shear Zone.

local basement, and led to formation of fault bounded pull-apart basins in transtensional segments (Fig. 9). Prevailing fault orientations show NNE-SSW trends coinciding with the direction of main compressional deformation generating the NW-SE extension in these basins. Similar paleostress fields have been recognized in Ediacaran basins of Brazil (Almeida et al., 2010) suggested that the age constraints, lithological similarities, and structural aspects point to the correlation of all Ediacaran to Cambrian basins of southeastern South America in a common basin system, with recurrent events of subsidence, magmatic activity, and brittle deformation from 605 to 530 Ma. Once established, these pull-apart basins subsided very quickly and accumulate large thicknesses of alluvial, marine and volcanic deposits. Such is the case of the Maldonado Group (alluvial conglomerates, sandstones, sand-mud rhythmites, basic and acid volcanic and volcaniclastic rocks), San Carlos Formation (fluvial sandstones and conglomerates; siltstones), Cerros de Aguirre Formation (acidic volcanic and volcaniclastic rocks), Barriga Negra Formation (alluvial conglomerates, sandstones and mudstones), Sierra de Ríos Formation (acidic volcanic rocks), and the Tacuarí Formation (conglomerates, sandstones and sand-mud rhythmites).

The Rocha Formation, located in the extreme southeastern Uruguay, was also deposited in the Ediacaran (Basei et al., 2008) and similar to its own basement (Cerro Olivo Complex) shows an African affinity. In the latter case, magmatic ages at 800–750 Ma on zircon and high-temperature metamorphism dated at 650–600 Ma would support the correlation between the Cuchilla Dionisio Terrane and the Coastal Terrane of the Kaoko Belt (Oyhantçabal et al., 2009; Gross et al., 2009; Konopásek et al., 2014). The Rocha Formation has been correlated with successions in the Gariep belt in Southern Africa, based on detrital zircon record, suggesting that it was deposited in the back-arc to the Gariep magmatic arc (Basei et al., 2000, 2005; cf. Oyhantçabal et al., 2007).

6. Conclusions

The purpose of the present study was to constrain the maximum depositional age and provenance of Neoproterozoic volcano-sedimentary successions of Uruguay with the ultimate goal of better understanding the regional paleogeography, and the timing and dynamics of terrane accretion. The following conclusions can now be drawn:

(1) The mixed siliciclastic-carbonate Arroyo del Soldado Group was deposited during the Tonian-Cryogenian, sometime between ~1000 and 650 Ma, in a passive continental margin developed along the southeastern margin of the Nico Pérez Terrane. In contrast, the deposition of the siliciclastic and volcaniclastic Playa Hermosa, Las Ventanas, San Carlos and Barriga Negra formations occurred during the upper Ediacaran (<585 Ma) in small fault-bounded basins developed over the Nico Pérez, Edén and Cuchilla Dionisio terranes.

(2) Evidence of tectonic and magmatic activity during the Tonian-Cryogenian is absent in the Nico Pérez Terrane, which further supports the existence of a stable continental margin at that time. The Ediacaran, however, is characterized by extensive tectonic activity and voluminous magmatism in all

the domains. This tectonic and magmatic activity is largely associated with the accretion of continental blocks.

(3) Far from being a definite model, our results suggest that the present-day terrane configuration took place in a relatively short (~100 Ma long) period of time between ~650 and 560 Ma. In this regard, the Edén and Nico Pérez terranes likely accreted at ~650–620 Ma (Edén Accretionary Event), followed by their accretion to the Piedra Alta Terrane at ~620–600 Ma (Piedra Alta Accretionary Event), and culminating with the accretion of the Cuchilla Dionisio Terrane at ~600–560 Ma (Cuchilla Dionisio Accretionary Event).

(4) Based on the distinct stratigraphy, geology, U–Pb zircon ages and Nd isotopes, a new terrane, named Edén Terrane, is now defined. As for the above-mentioned terranes, the boundaries between this block and the adjacent terranes are highly tectonized, with large-scale ductile shear zones implying that the geodynamic history of the Ediacaran is dominated by oblique movements with episodes of emplacement of granite complexes during transpressional and posterior extensional events.

Acknowledgments

This study benefited from funding provided by the Natural Sciences and Engineering Research Council of Canada, the Comisión Sectorial de Investigación Científica, Conselho Nacional de Desenvolvimiento Científico e Tecnológico (CNPq) and Alberta Ingenuity Fund. EP would like to express his sincere thanks to Arnold for his constant inspiration. We are grateful for a thorough review by João Orestes Schneider Santos and an anonymous reviewer, which helped to improve this manuscript.

Appendix A. Supplementary data

Supplementary data related to this article can be found at http://dx.doi.org/10.1016/j.jsames.2016.07.003.

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