

Chapter 4.1

EXTENSION AND GENERAL EVOLUTION OF THE RIO DE LA PLATA CRATON *

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4.1.1. INTRODUCTION

4.1.1.1. *Origin, extension and boundaries of the Río de la Plata cratonic region*

Following the fragmentation of the Rodinia supercontinent, Archaean to Mesoproterozoic cratonic blocks were amalgamated to constitute the Gondwana continent during Neoproterozoic-Cambrian times (Hoffman, 1999; Brito Neves et al., 1999; Fuck et al., 2008). One of these continental blocks of western Gondwana is the Río de la Plata Craton (Almeida *et al.*, 1973; 1976; 2000; Cordani et al., 2000), which comprises part of southern Brazil, Paraguay, Uruguay and the central-eastern sector of Argentina (Fig. 4.1.1). However, the main exposures of the Río de la Plata Craton (RPC), are located within its eastern border and include the Asunción arch, the Taquarembó Block, and the Nico Pérez, Piedra Alta and Tandilia terranes (Fig. 4.1.2). The Nico Pérez Terrane is the oldest and includes Paleoarchaean granite-greenstone sequences (Hartmann et al., 2001).

To the north, the RPC (Fig. 4.1.2) is bounded by the southern Amazonian craton (Bolivia-Brazil), whereas to the east it is bounded by the allochthonous Cuchilla Dionisio suspect terrane (Uruguay). The RPC is limited towards west by the Pampia terrane of Argentina (Ramos, 1988; Rapela *et al.*, 2007). According to geochronological studies of basement rocks (Cingolani and Varela, 1973; Rapela *et al.*, 2003) and based on at least two boreholes in the Claromecó basin (Lesta and Sylwan, 2005), it was deduced that the RPC was not part of the basement of the Sierra de la Ventana fold and

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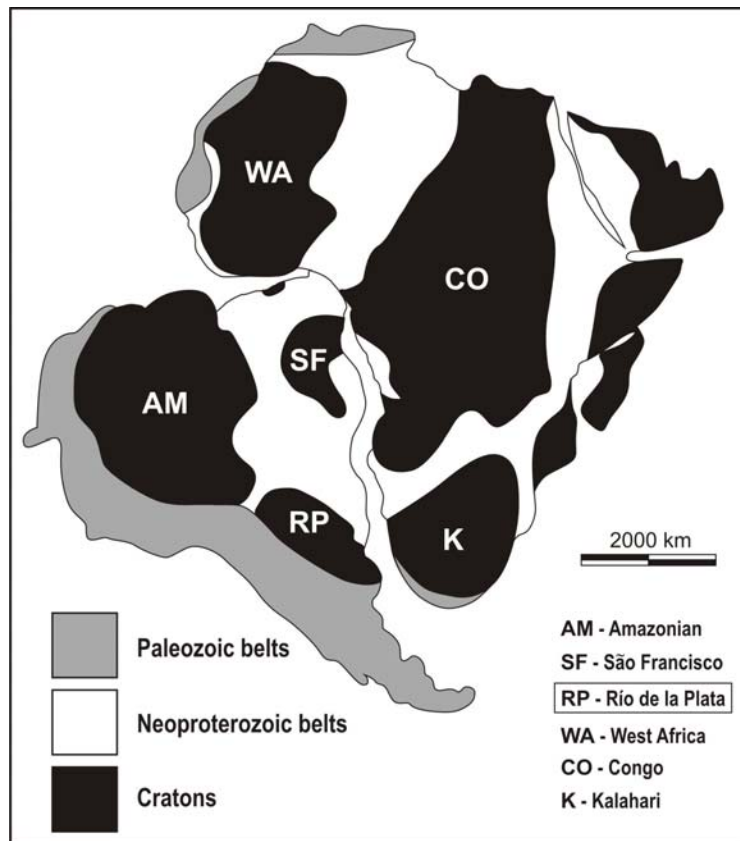


Fig. 4.1.1: Pre-drift configuration of Western Gondwana showing the location of the Río de la Plata Craton.

thrust belt. It is important to note that low-grade metapelites occurring in the Punta Mogotes borehole have been correlated to similar rocks (Rocha Formation) of the Cuchilla Dionisio terrane in Uruguay (Cingolani and Bonhomme, 1982; Ramos, 1988), showing similar detrital zircon ages (Rapela et al., 2008).

The southwestern, Pacific margin of Gondwana records a protracted terrane accretion history during Neoproterozoic-Paleozoic times simplified as “Terra Australis Orogen” by Cawood (2005).

4.1.1.2. Geophysical data

Extensive areas of the RPC are covered by Early Paleozoic to Cenozoic sedimentary basins, therefore knowledge regarding its subsurface continuation is based on geophysical research and borehole data (Rapela *et al.*, 2007 and references therein). Geophysical studies (Booker et al., 2004), indicate that the boundary between the RPC and the Early Paleozoic Pampia terrane could be a complex faulted margin (Ramos et al., 2002; von Gosen and Prozzi, 2005) as a continuation of the Transbrasiliano shear

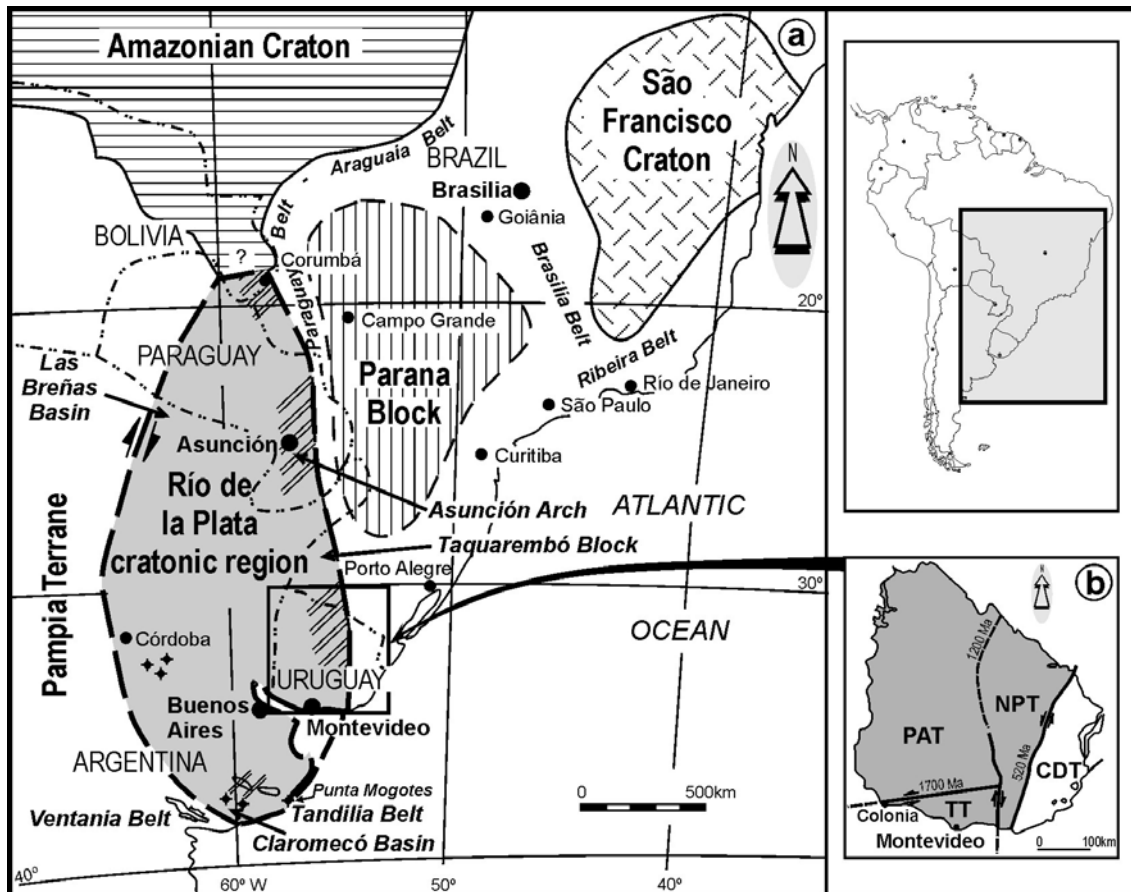


Fig. 4.1.2: (a) Southeastern South American sector showing the inferred limits of the Río de la Plata cratonic region (modified from Gaucher *et al.*, 2003). (b) Detail of the Uruguayan basement, comprising the Piedra Alta (PAT), Tandilia (TT) and Nico Pérez (NPT) terranes as part of Río de la Plata cratonic region, and the exotic Cuchilla Dionisio Terrane (CDT).

zone (Cordani *et al.*, 2001), that was probably active as early as the late Neoproterozoic. The recognition that longitudinal, dextral, ductile shear zones were associated with magmatism suggests transpressional movements.

Late Proterozoic to Early Paleozoic paleomagnetic poles were obtained from the Sierra de Ánimas rift-related magmatic complex (520 Ma), exposed in the Piriápolis area (Uruguay; Sánchez Bettucci and Rapalini, 2002) and at the Sierra de los Barrientos area of Tandilia, Argentina (Rapalini, 2006).

4.1.1.3. Neoproterozoic-Lower Paleozoic sedimentary cover

The north-south extension of the RPC is based on correlation of similar remnants of Neoproterozoic-Lower Paleozoic sedimentary covers. The main units are: Corumbá Group (Brazil), Itapucumí Group (Paraguay), Las Ventanas Formation and Arroyo del Soldado Group (Uruguay) and Sierras Bayas Group-Cerro Negro Formation (Argentina;

Gaucher et al., 2003; 2005a; 2008a; Blanco & Gaucher, 2005; Poiré and Spalletti, 2005). It is noteworthy that westwards of the mentioned platform, the Neoproterozoic (Brasiliano cycle) belts have not been neither directly observed nor detected using geophysical methods within the inferred Argentinian limits of the RPC.

4.1.1.4. Palaeocontinent concept

The ‘Río de la Plata palaeocontinent’ concept geographically coincides with the Río de la Plata craton as originally proposed by Almeida et al. (1976). On the other hand, Dalziel (1992) used the ‘Plata Block’ term for his paleogeographic reconstructions. Other authors continued naming craton all the regions with rock ages more than 2000 ± 100 Ma that represent a foreland with respect to the Brasiliano mobile belt (500-700 Ma; Dom Feliciano Belt *sensu* Fragoso-César, 1987; Basei et al., 2006; Oyhançabal et al., 2006, 2007). New geological, paleontological, geochemical and geochronological evidence suggest that the term ‘palaeocontinent’ is the best to be used in order to analyze all the geological information without any preconception, such as a ‘stable craton with a mobile belt’.

4.1.2. MAIN OUTCROP AREAS

4.1.2.1 Southern Brazil

As shown in Fig. 4.1.2, the southernmost portion of the Brazilian shield (Hartmann et al., 2002a; Gastal et al., 2005) comprises three main geotectonic domains: (1) the Río de la Plata craton represented by the Taquarembó block; (2) the Dom Feliciano terrane or collisional belt (550-600 Ma); and (3) the São Gabriel arc (700-900 Ma). The Late Neoproterozoic magmatic events were contemporaneous with the beginning of a large foreland system or extensional event: the Camaquã basin active between 605 and 470 Ma (Babinski et al., 1996; Paim et al., 2000). The São Gabriel arc includes metamorphosed volcanosedimentary sequences (Vacacaí Group) intruded by low-K calc-alkaline metagranitoids (Cambai Complex). All these rocks formed during the 700-900 Ma São Gabriel Orogeny (Gastal et al., 2005; Saalman et al., 2006; see Chapter 4.6), and are largely characterised by a depleted juvenile signature. The relationship between the RPC and the Camaquã basin, a thick Neoproterozoic to Ordovician volcanosedimentary

succession (Paim *et al.*, 2000; Gastal and Lafon, 2001), will be discussed in detail below (see Chapter 4.6).

In southwestern Brazil (Mato Grosso do Sul state), remnants of a carbonate-siliciclastic platform cover bearing Ediacaran fossils are included in the Corumbá Group (Boggiani *et al.*, 1993; Fig. 4.1.2). All lines of evidence presented by Gaucher *et al.* (2003) indicate that this Group is coeval with the Arroyo del Soldado Group (Uruguay), showing similar depositional environments and eustatic sea-level changes as a part of the same shelf. These units were deposited in a passive, Atlantic-type continental margin (Boggiani *et al.*, 1993; Boggiani, 1998; Gaucher *et al.*, 2003). The platform deepened to the east, with granitic-metamorphic source areas located to the west. The age of deformational events affecting the Corumbá Group are constrained around 540-490 Ma (see Chapter 2).

4.1.2.2. *Paraguay*

At the Asunción-Arch (Fig. 4.1.2), a high-grade Paleoproterozoic basement complex is composed of orthogneisses, banded granulites, amphibolites and quartzites, which record reworking events at 620-500 Ma (Cordani *et al.*, 2001). Ramos (1988) suggested the possibility that the Río de la Plata cratonic region could extend to the Asunción region. Predominantly carbonatic rocks of the Itapucumí Group were assigned to the Ediacaran on the basis of the occurrence of *Cloudina* (Boggiani and Gaucher, 2004), and correlated in part with the Corumbá Group (Boggiani, 1998). Due to the collision between the RPC and the ensialic Paraná-Paranapanema continental block located to the east, the Paraguay-Araguaia belt (ancient “Brazilides Ocean”) was formed and the Neoproterozoic-Cambrian sedimentary basin was closed and deformed.

4.1.2.3. *Uruguay*

The main outcrops of the Río de la Plata cratonic region are located in Uruguay. It was considered tectonically homogeneous (Dalla Salda *et al.*, 1988; Bossi & Navarro, 1991 and references therein), but detailed studies show that it is composed of three tectono-stratigraphic terranes named Piedra Alta, Tandilia and Nico Perez, separated by first-order shear zones (Bossi and Campal, 1992; Bossi *et al.*, 1993c; Bossi *et al.*, 2005). As

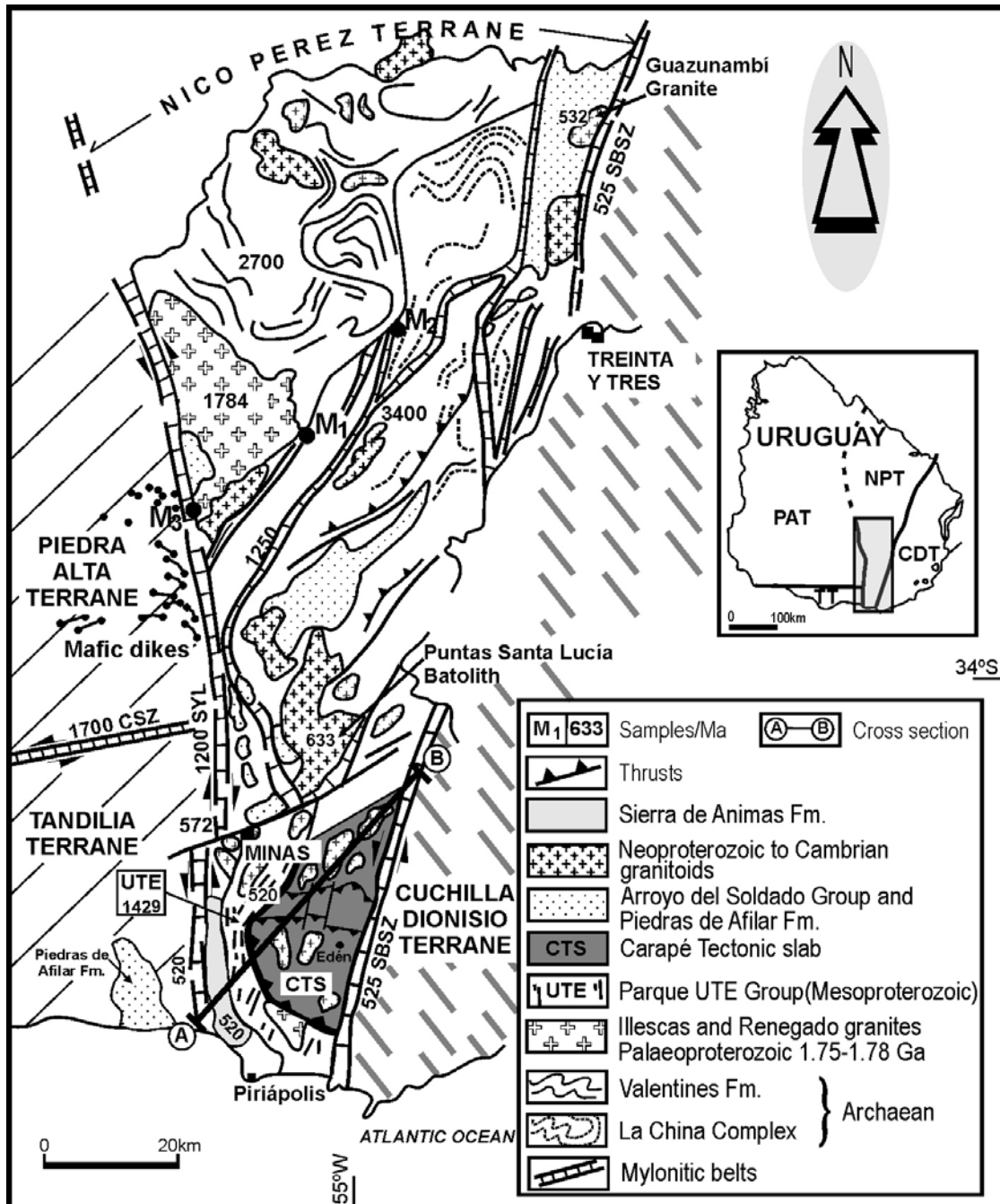


Fig. 4.1.3: Main geological units and boundaries of the Nico Pérez Terrane and neighbouring areas modified after Bossi and Gaucher (2004). CSZ: Colonia shear zone; SBSZ: Sierra Ballena shear zone; SYL: Sarandí del Yí lineament. Section AB: see Fig. 4.1.4.

shown in Fig. 4.1.2b, the Piedra Alta terrane (PAT) is separated from the Tandilia terrane (TT) by the Colonia Shear Zone of ca.1700 Ma. Neoproterozoic, siliciclastic-carbonate sedimentary cover remnants (Piedras de Afilar Formation), unconformable overlying the Palaeoproterozoic basement, are preserved in the southern portion of this terrane. Pamoukaghlian et al. (2006) assigned this unit to the Neoproterozoic based on

carbon isotope chemostratigraphy (see Chapters 4.2-4.4). Dolerites and the La Tuna Granite, probably of Lower Cambrian age, intrude the Piedras de Afilas Formation.

The suture between the Nico Pérez terrane (NPT) and the Piedra Alta and Tandilia terranes is the Sarandí del Yí lineament (SYL), recording dextral shear at 1200-1250 Ma. The age of the Sarandí del Yí Lineament of ca. 1200 Ma was obtained by K-Ar on synkinematic muscovites from pegmatites crystallised along thrust planes (1253 ± 32 Ma; Cingolani, in Bossi et al., 1998) and by ^{40}Ar - ^{39}Ar on mafic dikes deformed by the shear zone, yielding ages between 1370 and 1170 Ma (Teixeira et al., 1999; see Chapter 4.6). The Nico Pérez terrane is mainly composed of a Palaeoarchean basement (granite-greenstone belt after Hartmann et al., 2001), Paleoproterozoic orthogneisses and supracrustal volcano-sedimentary sequences (Bossi et al., 1965; Bossi and Navarro, 1991; Hartmann et al., 2001), intruded by the 1784 ± 3 Ma Illescas rapakivi granite (Fig. 4.1.3). A 'Grenvillian' metamorphic complex (Campal et al., 1995) occurs in the NPT. An important thrust separates the Grenvillian rocks from the Carapé tectonic slab (Bossi et al., 2007), which has been dated K-Ar on synkinematic muscovite at 572 ± 11 Ma (Cingolani, in Bossi and Campal, 1992). The eastern side of the RPC is limited by the sinistral Sierra Ballena megashear of ca. 530 Ma (SBSZ), separating the Nico Pérez Terrane from the allochthonous Cuchilla Dionisio Terrane (CDT, Fig. 4.1.3). The suture between cratonic areas to the west and remnants of a magmatic arc to the east (CDT) are characterised by the absence of a continuous Neoproterozoic, Brasiliano-aged schist belt. These facts could indicate that the amalgamation of the CDT to NPT was due to lateral rather than orthogonal accretion (Basei et al., 2000; Bossi and Gaucher, 2004; Basei et al., 2005).

The interpretation that the NPT is a para-autochthonous terrane displaced to the north during Cambrian times was suggested (Gaucher, 2000; Bossi and Gaucher, 2004), related to a sinistral reactivation of the Sarandí del Yí Lineament. Main evidence for the north displacement of the NPT is the presence of Archean detrital zircons in Neoproterozoic sediments of the Tandilia System (Gaucher et al., 2008b), presently 600 km to the south.

In this review of the Uruguayan part of the RPC we will discuss the following events, which still remain controversial or poorly studied: **granitogenesis between 630 and 530 Ma**; the compressive event represented by the Carapé tectonic slab; the

significance of shear zones at c. 525 Ma and the Sierra de Ánimas alkaline magmatism. Finally, the 'Lavalleja Group' will be discussed, because there are conflicting Neoproterozoic and Mesoproterozoic ages for the sedimentation and/or metamorphism of this unit.

4.1.2.3.1. Neoproterozoic-Cambrian granitic magmatism in the RPC

The NPT records magmatic activity in the late Neoproterozoic and Cambrian that can be grouped in three main events at 660-630 Ma, 590-580 Ma and 540-530 Ma. The geochemical signature, origin and evolution of the small granitic intrusions in NPT are not well known. Umpierre and Halpern (1971), Soliani (1986), Preciozzi et al. (1993) and Hartmann et al. (2002a) suggested, based in isotopic data, a relationship with a magmatic arc during the Brasiliano cycle, with west-directed subduction and a long distance from the trench. Metamorphic rocks of the Brasiliano cycle are lacking within the terrane, but a conspicuous late Ediacaran to basal Cambrian, folded sedimentary cover (Arroyo del Soldado Group) is preserved, which bears abundant fossils (Gaucher, 2000). It is clear that the Sierra Ballena Shear Zone affected this sedimentary cover (Gaucher et al., 1998). Furthermore, the Arroyo del Soldado Group overlies with erosional unconformity the Puntas de Santa Lucía Granite (633±12 Ma, U-Pb SHRIMP: Hartmann et al., 2002a) and the Mangacha Granite of 583±7 Ma (U-Pb SIMS, Gaucher et al., 2008b). On the other hand, the Arroyo del Soldado Group is intruded by the Guazunambí Granite which yielded a Rb-Sr isochronic age of 532±11 Ma (Kawashita et al., 1999; Fig. 4.1.3). K-Ar ages ranging from 532±16 to 492±14 Ma for pelites of the Arroyo del Soldado Group constrain the age of anchimetamorphism and cooling of the unit (Cingolani et al., 1990b; Gaucher, 2000). Thus, all lines of evidence suggest that between 580 and 540 Ma stable conditions prevailed at the eastern margin of the RPC. Gaucher et al. (2008a) argue that magmatism pre-dating the Arroyo del Soldado Group (580 Ma or older) is associated to extensional tectonics (see also Chapter 4.6). Magmatism post-dating the Arroyo del Soldado Group is clearly linked to tangential continental collision (Bossi and Gaucher, 2004). For a different interpretation of these events see Oyhantçabal et al. (2007). One apparent contradiction arises from K-Ar ages of 572±11 Ma on synkinematic muscovites (Cingolani, in Bossi and Campal, 1992) associated to the emplacement of the Carapé tectonic slab. If this single age is taken at face value, it would indicate an important compressive event coeval with the transition

from rifting to drifting in the Arroyo del Soldado basin, which is clearly not compatible (see below).

In the northern part of the Tandilia Terrane sensu Bossi et al. (2005; Fig. 4.1.2) granitic rocks assigned to the La Paz Granite crop out 20 km to the north of Montevideo (34°50' S-56°10'W). It is a coarse-grained, homogeneous, pink granite with scarce ferromagnesian minerals (Cardelino and Ferrando, 1969). A Rb-Sr age for this massif of 565±15 Ma was reported by Umpierre and Halpern (1971). This posttectonic magmatic event intruded a Paleoproterozoic cratonic area of 2200 to 2000 Ma in age. The granite is typically medium-grained (0.5-1.5 mm) and is made up of 70% perthitic-microcline, 29% quartz (mainly idiomorphic) and 1% hornblende and biotite. Chemical analyses show SiO₂: 72%, K₂O: 5.6%, Na₂O: 3.1% and CaO; 1%. Oyhantçabal et al. (2006) have recently demonstrated the existence of an important sinistral shear zone striking N60E, associated with muscovite-rich peraluminous granites dated at 1980±60 Ma by K-Ar. This lineament may have provided a weakness zone for the intrusion of the La Paz Granite in the middle of the cratonic area. Precise U-Pb ages of the La Paz Granite will help to assign this intrusion to one of the different magmatic events mentioned above.

4.1.2.3.2. *Carapé tectonic slab*

A medium-grade metamorphic complex was recognised in the southeastern part of the NPT. This complex is composed of BIFs, marbles, amphibolites, gneisses and micaschists. Bossi (1983) erected the Carapé Group to include these lithologies. On the other hand, Sánchez Bettucci (1998) proposed another denomination for the same unit, the Zanja del Tigre Formation. Bossi and Navarro (2001) reviewed the nomenclature and proposed to maintain the Carapé Group as priority name. It is noteworthy that the U-Pb data on detrital zircons obtained by Basei et al. (2008) and chemostratigraphic data reported by Chiglino et al. (2008) suggest that the Zanja del Tigre Formation was deposited between 1800 and 1400 Ma.

The Carapé Complex as described by Sánchez Bettucci (1998) and Sánchez Bettucci et al. (2003) is composed of pre-, syn- and post-tectonic granites and anorogenic granitoids associated to the Zanja del Tigre Formation. Mallmann et al. (2007) concluded that the Carapé Complex was formed at 1.75 Ga, mainly by partial melting of Archaean crust (T_{DM} model ages between 2.8 and 3.0 Ga with strong

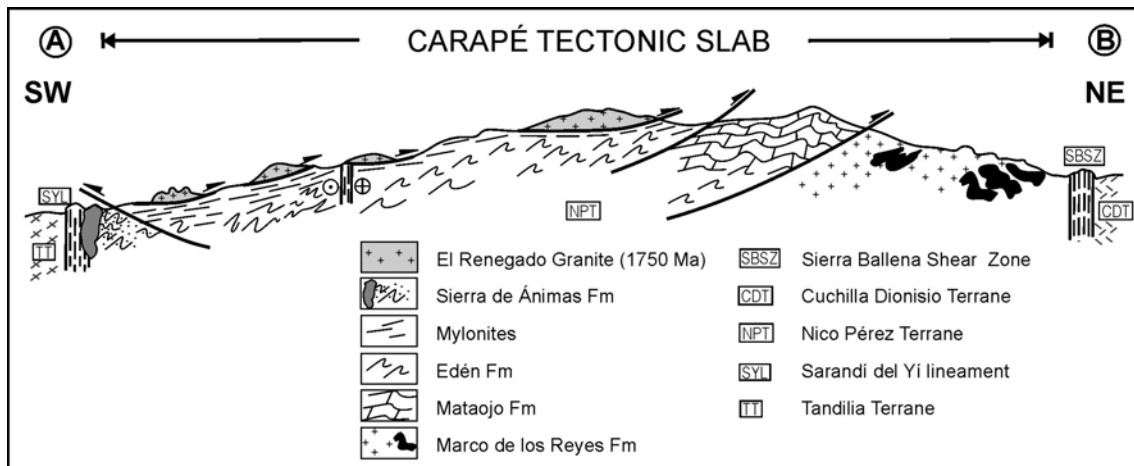


Fig. 4.1.4: Schematic cross section (AB) of the Carapé tectonic slab (Nico Pérez Terrane, Uruguay). Location of the section: see Fig. 4.1.3.

negative $\varepsilon_{Nd}(t)$ values). All mentioned data allow us to confirm that, for the most part, the complex is not of Neoproterozoic age.

Detailed geological mapping of the south-eastern NPT (Bossi et al., 2007) allowed to separate a number of fault-bounded lithologic blocks, grouped into the Carapé tectonic slab, which represents a displaced block of the NPT. The Carapé tectonic slab (Fig. 4.1.3) is made up of the following fault-bounded units (Fig. 4.1.4):

- (a) *Marco de los Reyes Formation*, medium-grade metamorphic rocks that include BIFs, micaschists, gneisses and pure limestones;
- (b) *Mataojo Formation*, recrystallised dolostones, micaschists and gneisses;
- (c) *Eden Formation*, amphibolites and paragneisses;
- (d) *El Renegado Granite*, that represents a nappe thrust onto the above units (Fig. 4.1.4). The crystallisation of this granite has been dated by U-Pb (SHRIMP and TIMS) at 1754 ± 7 Ma and 1735 ± 32 Ma (Mallmann et al., 2007 and Sánchez Bettucci et al., 2004 respectively).

Intrusive magmatic rocks include: (a) pre-tectonic granitic sills within the Marco de los Reyes Formation, typically boudinaged; and (b) voluminous, later post-tectonic granite intrusions into the Marco de los Reyes Formation. The age of the post-tectonic granitoids is constrained between 498 and 572 Ma, but there is a lack of U-Pb geochronological data of the pre- and syn-tectonic granitoids.

The tectonic evolution of the Carapé slab is complex, but the following events were recognised (Bossi et al., 2007; Fig. 4.1.4):

- (1) Evidence of north-vergent overthrusting and sinistral transcurrence, along the Puntas del Pan de Azúcar lineament (Oyhantçabal et al., 2001).
- (2) Crystallisation of muscovite-bearing pegmatites into the thrust plane, that yielded K-Ar ages of 572 ± 11 Ma (Cingolani, in Bossi and Campal, 1992).
- (3) Drag folding of previous structures with a radius of 10 km and sinistral sense was recognised near Eden town.
- (4) Emplacement of the El Renegado Nappe on top of the other units (Fig. 4.1.4).

Thus, as proposed by Bossi et al. (2007), the Carapé tectonic slab comprises various and complex geological events which demand further study.

4.1.2.3.3. Shear zones

The Nico Pérez Terrane is cross-cut by several mylonite bands (Fig. 4.1.3) of different ages, as a consequence of conspicuous tectonic events in the Archaean, Mesoproterozoic, Neoproterozoic and Cambrian. The eastern boundary of the Nico Pérez terrane is the Sierra Ballena Shear Zone (SBSZ, Oyhantçabal, 2005), of possible Cambrian age (Bossi and Gaucher, 2004; Rapalini and Sánchez Bettucci, 2008). This shear zone has not been precisely dated, but it obviously affected the Guazuambí Granite with a Rb-Sr age of 532 ± 11 Ma (Kawashita et al., 1999; Gaucher, 2000), and did not deform the Sierra de Ánimas magmatic unit of 520 ± 5 Ma (Rb-Sr, Bossi et al., 1993b). Umpierre and Halpern (1971) dated mylonitic rocks east of the Nico Pérez town (sample M1 in Fig. 4.1.3), at the cerro La Palma on road number 19, 22 km to the east of the Valentines locality (sample M2) and in the SYL on road number 7 (sample M3) as shown in Fig. 4.1.3. All those rocks yielded Rb-Sr average ages of 535 ± 20 Ma and were originally described as ‘migmatites or embrechitic’ rocks. These ages can now be easily explained, because they are coincident with the age of the Sierra Ballena shear zone, showing that the Sarandí del Yí lineament was sinistrally reactivated during the tangential collision of the Cuchilla Dionisio Terrane (Gaucher et al., 2008b; Fig. 4.1.5).

4.1.2.3.4. Sierra de Ánimas Formation

Cambrian magmatic activity ended in Uruguay with extensional magmatism of the Sierra de Animas Formation. It represents a north-trending extensional basin, 50 km long and 10 km wide, where mesosilicic rocks including syenites and trachytes but also

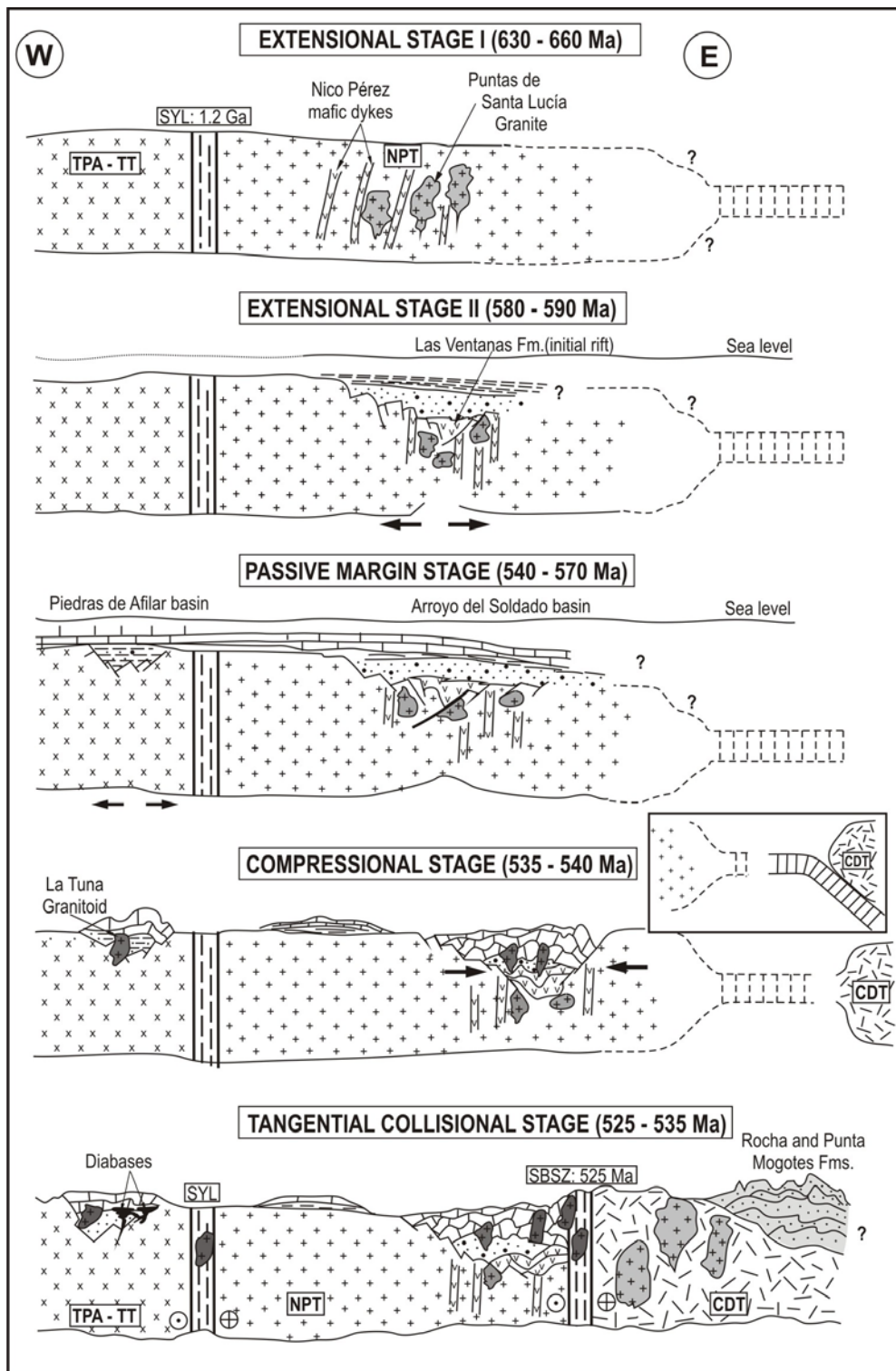


Fig. 4.1.5: Interpretative geotectonic evolution in the Uruguayan sector of the Río de la Plata cratonic region beginning with two extensional stages (660-630 and 590-580 Ma) and a passive margin stage (570-540 Ma) where the Piedra Alta and Tandilia terranes are already amalgamated with the Nico Pérez terrane along the Sarandí del Yí lineament (SYL). The compressional stage (535-540 Ma) shows intrusions and deformation of the sedimentary cover, in a transpressive regime with subduction towards de south-southwest beneath the Cuchilla Dionisio Terrane (CDT). After that the tangential collisional stage occurred (525-535 Ma) and a sinistral shear zone was developed (SBSZ) at ca. 530 Ma. The eastern part of the Nico Pérez terrane may have been displaced to the north, as well as hypothetical metamorphic rocks of the Brasiliano cycle. Granitoid intrusions occur, which are mainly associated to shear zones.

trachybasalts occur; rhyolites developed as a later rock facies. Bossi (1966), Preciozzi et al. (1985), Bossi et al. (1993b) and Sánchez Bettucci (1998) described the petrography and the magmatic structure. Umpierre and Halpern (1971), Bossi et al. (1993b), Sánchez Bettucci and Linares (1996) and Oyhantçabal et al. (2007) published several isotopic ages. A whole rock Rb-Sr isochronic age on five different lithologies yielded 520 ± 5 Ma (Bossi et al., 1993b), which is here considered the best available age estimate.

Some authors mentioned that the Valle Chico Massif, located to the northeast of Mariscal city, is a Cambrian event due to the lithological similarity with the Sierra de Ánimas Formation (Bossi and Navarro, 1991; Bossi et al., 1998), but the age is poorly constrained. Muzio and Artur (2000) reported an U-Pb zircon age of 129 ± 5 Ma, derived from the lower intercept in the discordia diagramme. This discordant age could be alternatively interpreted as due to Pb loss caused by voluminous Mesozoic magmatism affecting the massif.

4.1.2.3.5. Comments about the Lavalleya Group

Caorsi and Goñi (1958) were the first to use the name “Lavalleya Series” for the volcano-sedimentary succession, but they also included other, older supracrustal rocks. Bossi et al. (1965), Preciozzi et al. (1985), and Bossi and Navarro (1991) suggested restricting the name to the Neoproterozoic supracrustal rocks (Lavalleya Group). According to Gaucher et al. (2004c), the Lavalleya Group (*s.s.*) only occurs between Minas and Pan de Azúcar cities (Fig. 4.1.3). This lithostratigraphic unit has been used as a “waste bin” and is the main cause of the geological controversies surrounding the Neoproterozoic evolution of Uruguay.

The original definition included a low-grade metamorphic succession with northeast regional strike and exposed between Pan de Azúcar and Parao (Bossi et al., 1965). In the prevailing ‘geosynclinal theory’ of the time it represented the western border of the orogenic cycle with high-grade metamorphic and granitoid rocks within the central part. The pioneer geochronological work carried out by Hart (1966) and Umpierre and Halpern (1971) identified a 500-700 Ma mobile belt (Ferrando and Fernández 1971) and a cratonic region characterised by rocks of ca. 2000 Ma. Almeida et al. (1976), assigned these two terrains according to their ages to the Brasiliano and Transamazonian Cycles, respectively. The definition of the Nico Pérez Terrane (Bossi and Campal, 1992) opened a new chapter on the study of the Precambrian rocks in

Uruguay: the Lavalleya Group was restricted to its type area, and the folded but unmetamorphosed sedimentary facies were included in the Arroyo del Soldado Group (Gaucher et al., 1996). Part of what was previously assigned to the Lavalleya Group because of its greenschist-facies metamorphism yielded Archean ages (Hartmann et al., 2001). Moreover, U-Pb zircon ages of 1492 and 1429 Ma were obtained for metarhyolites and metagabbros in the type area of the Lavalleya Group (Oyhantçabal et al., 2005). This reinforces previous Pb-Pb ages of 1300 to 1500 Ma on galenas (Garau, in Bossi and Ferrando, 2001) and K-Ar ages of 1208±10 Ma in metabasalts (Gómez Rifas, 1995), showing that part of the former Lavalleya Group is Mesoproterozoic in age. Finally Mallmann et al. (2007), report U-Pb ages of 590 Ma for metavolcanics, indicating a very complex scenario. Two distinct signatures have been recognised, one similar to mid-ocean ridge basalts and another one with characteristics of ocean island basalts (Mallmann et al., 2007).

Therefore, it is clear that the Lavalleya Group includes a number of unrelated units of different ages and geotectonic settings, thus becoming obsolete, a view already expressed by Bossi et al. (2002) and Gaucher et al. (2004c). The name, however, has become established in the literature and authors continue to use it. To overcome this problem, the type outcrop areas between Minas and Pan de Azúcar with reliable Mesoproterozoic ages were assigned by Bossi et al. (2008) to a new stratigraphic unit, the Parque UTE Group.

4.1.2.4. Argentina

The southernmost outcrops of the Río de la Plata cratonic region are exposed in the Tandilia System (Fig. 4.1.7). It is a 350 km long, northwest-trending orographic belt, located in the central part of the Buenos Aires province. It comprises an igneous-metamorphic Paleoproterozoic basement named Buenos Aires Complex ('Tandiliano cycle'), which is covered by the Neoproterozoic Sierras Bayas Group and Cerro Negro Formation. An extensive blanket of Ordovician to Silurian quartz-arenites (Balcarce Formation) partially cover the older units, reflecting an independent basin evolution (Teruggi et al., 1989; Ramos, 1999; Cingolani and Dalla Salda, 2000 and references therein).

Ages of basement granites, gneisses and migmatites were determined by K-Ar,

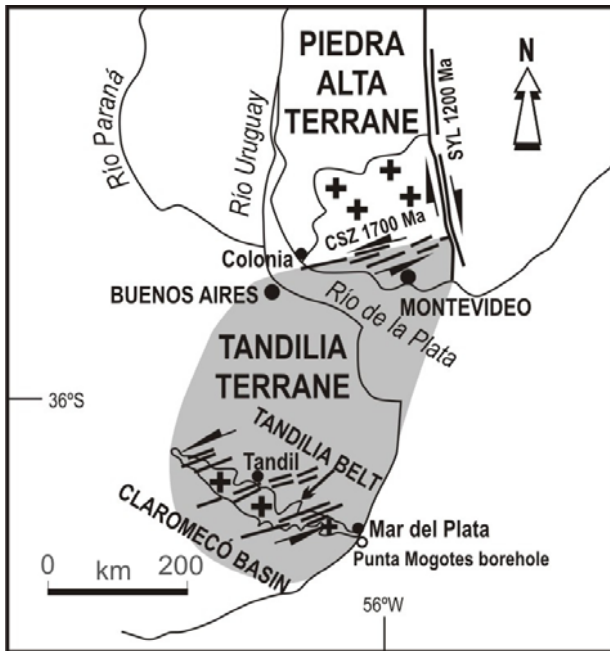


Fig. 4.1.6: Inferred extension of the Tandilia terrane after Bossi et al. (2005), which is exposed both in southern Uruguay and in the Buenos Aires Province (Argentina). CSZ: Colonia Shear Zone.

Rb-Sr, Sm-Nd and U-Pb SHRIMP zircon dating from several localities in the Tandilia System. Ages range mainly between 2.2 and 2.08 Ga (Cingolani and Dalla Salda, 2000; Pankhurst et al., 2003; Hartmann et al., 2002b, Cingolani et al., 2000, 2005 and references therein). Calc-alkaline dykes are coeval with the youngest granites and are considered to represent transtensional stages. Sm-Nd T_{DM} model ages range between 2.7 and 2.4 Ga. According to Hartmann et al. (2002b), the main tectonic activity has been linked to the 2.25-2.12 Ga accretionary Encantadas Orogeny and to the 2.1-2.08 collisional overprint, the Camboriú Orogeny.

It is important to mention that Bossi et al. (2005) postulated a correlation between the Buenos Aires Complex in Tandilia and the southwestern part of the Precambrian basement in Uruguay, including the Montevideo Formation (Pando Belt) and several Palaeoproterozoic granitoids. Both compose the Tandilia Terrane (Fig. 4.1.6), which is characterised by Palaeoproterozoic igneous and metamorphic rocks with similar Nd and U-Pb isotopic data that were not reworked during the Brasiliano orogenic cycle. The accretion of the Tandilia Terrane to the Piedra Alta Terrane along the Colonia Shear Zone took place between 1.79 and 1.59 Ga, as demonstrated by cross-cutting relationships of the shear zone with two mafic dyke swarms (Bossi et al., 2005). In the Neoproterozoic, the main geological events were the deposition of a passive margin sedimentary succession and brittle tectonism (Fig. 4.1.8).

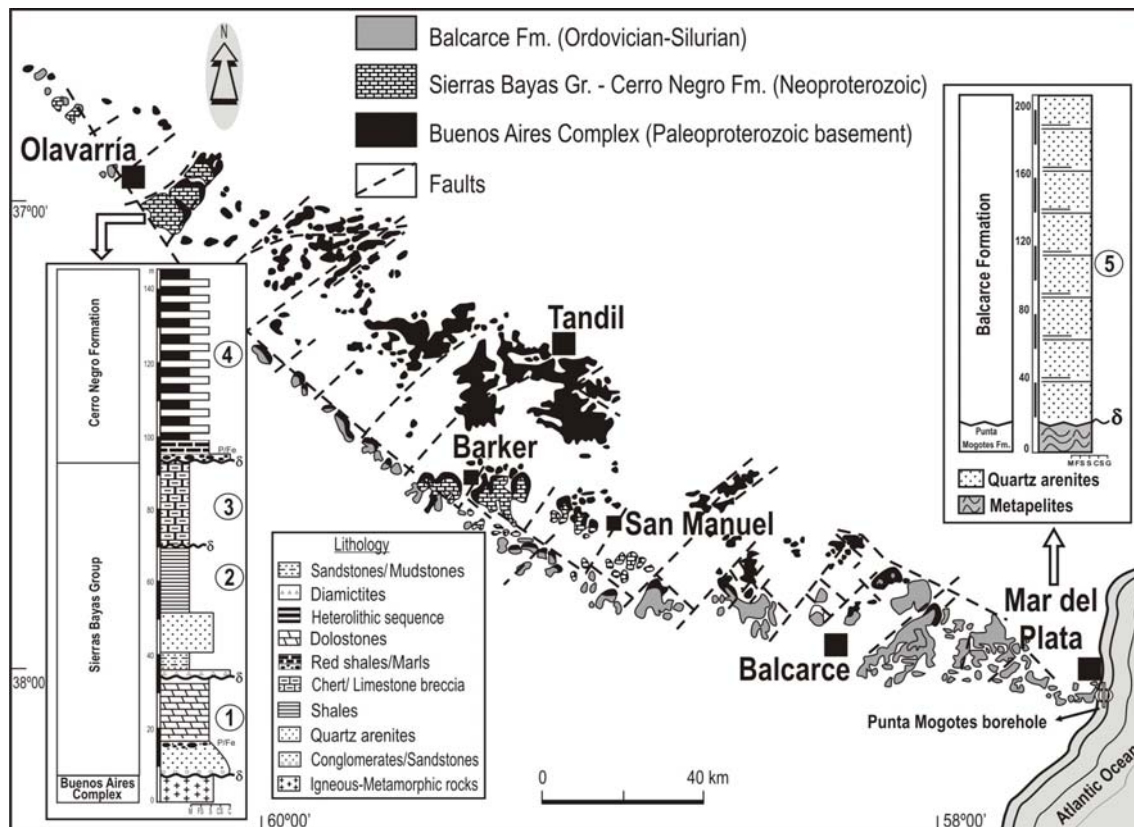


Fig. 4.1.7: Geological map of the Tandilia System showing the main outcrop areas of the Neoproterozoic-Lower Paleozoic successions. On the left, the stratigraphic column of the Sierras Bayas Group-Cerro Negro Formation in the Olavarría region is shown (Poiré et al., 2005). The stratigraphic column on the right shows the Punta Mogotes Formation overlain by the Balcarce Formation at the Punta Mogotes borehole.

4.1.2.4.1 Neoproterozoic sedimentary succession

The Neoproterozoic marine deposits, formerly known as 'La Tinta Group' and composed by the Sierras Bayas and Balcarce formations, were correlated to the Nama Group in southwest Africa (Dalla Salda, 1982), recording the assembly of southwestern Gondwana during the Neoproterozoic. Aceñolaza (1978) mentioned this correlation and described the common Neoproterozoic-Lower Paleozoic sequences as the "Nama-La Tinta" basin.

In the currently accepted stratigraphic scheme (Dalla Salda and Iñiguez, 1979; Iñiguez et al., 1989 and references therein; Poiré et al., 2005; Gomez Peral et al., 2007), the **Sierras Bayas Group** is a Neoproterozoic sedimentary cover 175 m in thickness comprising –from base to top– the Villa Mónica, Cerro Largo, Olavarría and Loma Negra formations (see Chapter 4.2). The Villa Mónica Formation was defined as the "Lower Quartzites" and "Dolostones", and bears phosphate levels and a rich stromatolite assemblage (Poiré, 1993) suggesting a Riphean age. Rb-Sr geochronology of illitic pelites within the dolostones yielded an age of 793 ± 32 Ma (Cingolani and

Bonhomme, 1982). Detrital zircon geochronology (Rapela et al., 2007; Gaucher et al., 2008b), show a unimodal population of Palaeoproterozoic age. This is a clear indication that the sandstones were mostly derived from the underlying Buenos Aires Complex, which has a restricted age range. The Cerro Largo Formation begins with a shale-sandstone association (Poiré, 1993), unconformable on top of the Villa Mónica Formation. It is followed by quartz-arenites which bear simple trace fossils (see Chapter 4.3). According to geochronological studies (Gaucher et al., 2007, 2008b), the sandstones are characterised by a dominant Palaeoproterozoic detrital zircon population, but also important Archean to lowermost Paleoproterozoic and Mesoproterozoic ages. The presence of Archean detrital zircons in sandstones of the Sierras Bayas Group suggests that the Nico Pérez Terrane (Uruguay), as a source of detritus, was closer to Tandilia (Argentina) than it is today (Gaucher et al., 2008b). The absence of Neoproterozoic detrital zircons confirms the deposition in a stable continental margin. Pelites compose the overlying Olavarría Formation. Rb-Sr ages on illitic shales point to a Neoproterozoic age (Bonhomme and Cingolani, 1980). Up section, the 45 m thick Loma Negra Limestone occurs (Borrello, 1966), and has been mined for a century for the cement industry. It was preserved in the Sierras Bayas and Barker areas (Fig. 4.1.7). In Barker, karstic breccias developed on top of the Loma Negra Formation, pointing to a last regression. The occurrence of *Cloudina*, low diversity acritarchs including *Leiosphaeridia* and chemostratigraphic data (Kawashita et al., 1999b; Gaucher et al., 2005a; Gómez Peral et al., 2007) all suggest a late Ediacaran age for the upper Sierras Bayas Group.

In the Olavarría-Sierras Bayas region, Iñiguez and Zalba (1974) erected the **Cerro Negro Formation** as a 150 m thick unit made up of illite-chlorite rich, green and reddish pelites. This unit unconformably overlies the Loma Negra Formation. At the base of the Cerro Negro Formation, Leanza and Hugo (1987) described a phosphate layer (“phosphate member”) that suggests the infilling of a flat basin on a paleosurface of the Loma Negra Formation during a marine regression. Radiometric data suggest a Neoproterozoic age, as well as acritarch biostratigraphy (Cingolani et al., 1991, 1999; Gaucher et al., 2005a). Detrital zircon U-Pb ages show the same polymodal pattern reported for the Cerro Largo Formation (C. Cingolani, unpublished data).

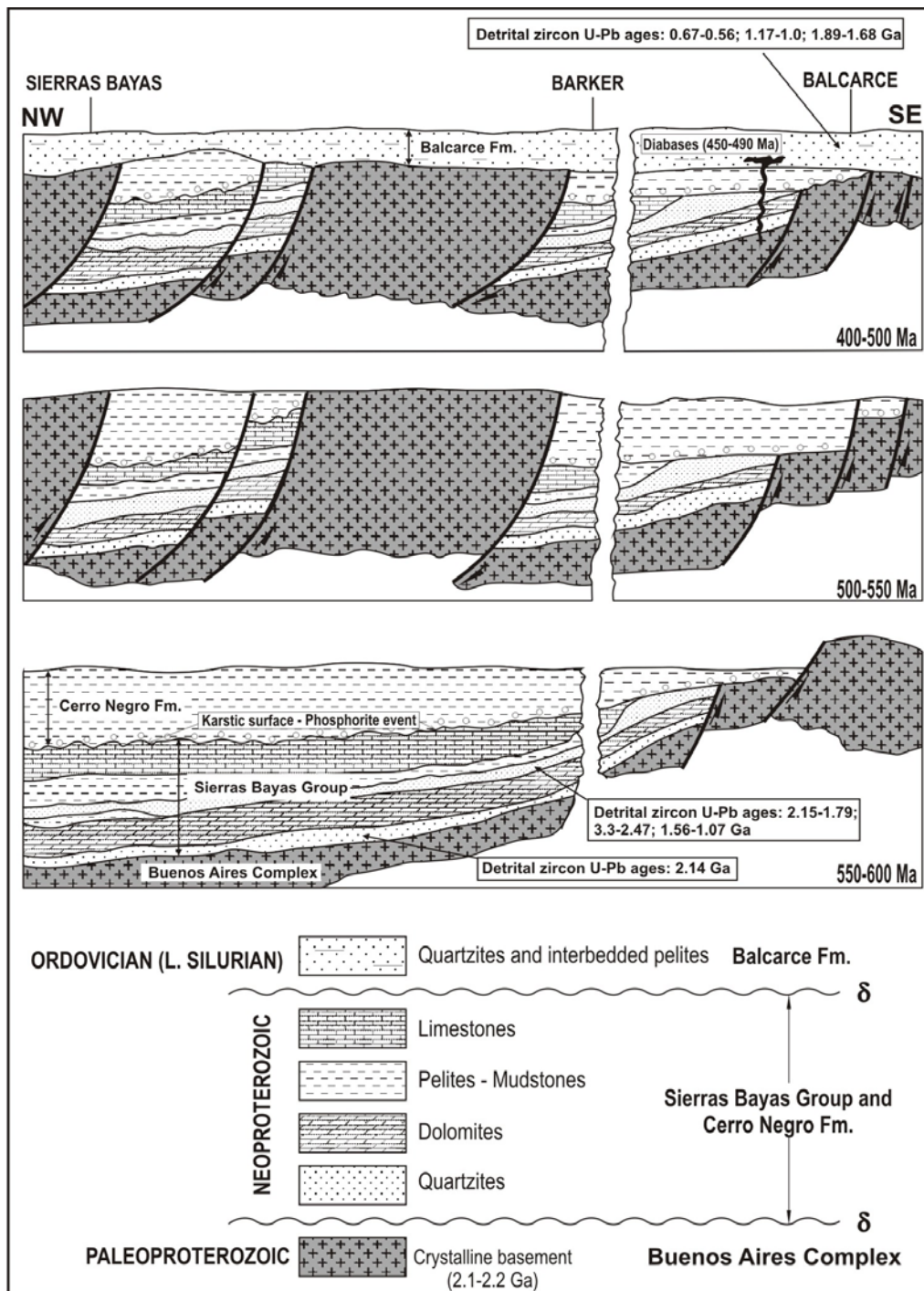


Fig. 4.1.8: Tectono-sedimentary model of the Tandilia System in the Neoproterozoic-Lower Paleozoic, modified after Iñiguez et al. (1989). See text for explanation.

4.1.2.4.2. Lower Palaeozoic sedimentary cycle

The last sedimentary cycle, represented by the **Balcarce Formation**, crops out along the southern edge of Tandilia from the Olavarría area to Mar del Plata at the Atlantic coast (Fig. 4.1.7) The thickest sections are exposed between Balcarce and Mar del Plata. The Balcarce Formation unconformably overlies all the Precambrian units. Its average

thickness is 75 to 90 m, reaching 450 m in the Punta Mogotes borehole near Mar del Plata. The unit is composed of quartzites, fine-grained quartzitic conglomerates and shales. The base of the sequence displays pyroclastic rocks that crop out near the Cerro del Corral area, which were affected by hydrothermal solutions. The Balcarce Formation bears abundant trace fossils that were assigned to the “*Cruziana* facies” by Borrello (1966) and Aceñolaza (1978). Subalkaline diabase sills intrude kaolinitic shales in Los Barrientos area (Rapela et al., 1974). More recent trace fossil studies (Seilacher et al., 2002) describe new *Cruziana* species, suggesting an Upper Ordovician to Lower Silurian age. Detrital zircon U-Pb ages were obtained by Rapela et al. (2007), showing a distinctively different age pattern compared with that of the Sierras Bayas Group. Neoproterozoic, Mesoproterozoic and Late Paleoproterozoic grains dominate the zircon population. The youngest zircon age of 480 Ma provides a maximum age constraint for the Balcarce Formation.

4.1.2.4.3 Deformation and tectonic evolution

The Tandilia Neoproterozoic-Lower Paleozoic sedimentary rocks were not deformed and remained flat-lying (Iñiguez et al., 1989). The Brasiliano cycle only affected the crystalline basement and its sedimentary cover by faulting with vertical and horizontal components in response to southwest-directed stress (Fig. 4.1.8). Due to vertical tectonic movements, the Palaeoproterozoic basement was uplifted but the depressed blocks preserved most of the sedimentary cover (Sierras Bayas, Barker areas).

It is important to note that the southern extension of the Neoproterozoic-Lower Paleozoic sedimentary cover of Tandilia was recorded in the Claromecó foreland (or foredeep) basin on several seismic profiles and boreholes (Lesta and Sylwan, 2005; Fig. 4.1.2). Finally, Zalba et al. (2007) confirmed that after the uplift and deformational phase that occurred during the Middle Permian, the Tandilia System has remained as a positive area.

4.1.3 CONCLUSIONS

1. The main outcrops of the Río de la Plata cratonic region are located in Uruguay, where three tectono-stratigraphic terranes (Piedra Alta, Tandilia and Nico Pérez) are separated by first order shear zones. The Nico Pérez terrane records several Neoproterozoic-Cambrian tectono-magmatic events. The most relevant are:

granitogenesis at 660-630, 590-580 and 540-530 Ma; the compressive event responsible for the emplacement of the Carapé tectonic slab (572 ± 11 Ma, K-Ar); tangential collision at ca. 530 Ma represented by the Sierra Ballena Shear Zone, and the Sierra de Ánimas alkaline magmatism (520 ± 5 Ma). The Nico Pérez terrane is a para-autochthonous, north-displaced terrane, related to the sinistral reactivation of the Sarandi del Yí Lineament during the Cambrian.

2. In the Tandilia region (Buenos Aires Province, Argentina) the igneous-metamorphic Paleoproterozoic complex that characterises the southern part of the RPC was not reworked during the Brasiliano orogenic cycle. Only passive margin deposition and subsequent brittle tectonics were recorded during Neoproterozoic-Lower Paleozoic times. Several sedimentary cycles separated by regionally extensive unconformities resulted from sea-level changes during the evolution of the basin (Sierras Bayas Group-Cerro Negro Formation).

3. The high-grade Paleoproterozoic basement complex (Asunción Arch) in Paraguay, record reworking at 620-500 Ma. Ediacaran carbonates of the Itapucumí Group represent a cratonic cover correlative in part to the Corumbá Group.

6. In the RPC, it is evident that the final amalgamation of Gondwana occurs during Cambrian times (ca. 530 Ma), when the tangential collision of the Cuchilla Dionisio Terrane with the RPC generated an important megashear zone (Sierra Ballena).

7. The geotectonic evolution of the RPC during the Neoproterozoic and Cambrian is still under discussion. The rift-drift transition is clearly shown by the remnants of an extensive Neoproterozoic-Lower Paleozoic platform succession. However, the abundant granitic intrusions between 630 and 530 Ma in the Nico Pérez terrane, coeval with the mentioned sedimentary cover, could be explained by rift related granitoids or transpressive regime with subduction towards the north-northwest during the Brasiliano cycle, or a combination of both models. Subsequently, the sinistral tangential collision with the Cuchilla Dionisio Terrane may have displaced part of the Nico Pérez terrane as well as the metamorphic rocks formed during closure of the Brazilides Ocean (see Chapter 4.6).

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LITHOSTRATIGRAPHY*

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4.2.1 INTRODUCTION

Two main areas in the Río de la Plata Craton (RPC) show important exposures of Neoproterozoic sedimentary successions, comprising both siliciclastic and carbonate rocks: the Nico Pérez Terrane in Uruguay and Tandilia in Argentina (Figs. 4.1.3, 4.1.7). The aim of this chapter is to describe the lithostratigraphy of the Neoproterozoic and early Palaeozoic successions occurring there.

4.2.2 TANDILIA SYSTEM, ARGENTINA

The Tandilia System (Nágera, 1940) is an orographic belt located in the Buenos Aires Province, between latitudes 36° 30' - 38° 10' South and longitudes 57° 30' - 61° West (Fig. 4.1.7). Its maximum length is 350 km in the NW-SE direction. The hills are composed of an igneous metamorphic basement and a Neoproterozoic and early Palaeozoic sedimentary cover, which displays horizontal to subhorizontal bedding (<12°).

Whereas the Neoproterozoic sedimentary succession crops out in the north-western part, around Olavarría and Barker-San Manuel region (Fig. 4.1.7), Lower Palaeozoic rocks occur in the western part of the range and –especially- towards the southeast (Balcarce-Mar del Plata area, Fig. 4.1.7). These deposits overlie a Palaeoproterozoic crystalline basement (Buenos Aires Complex, Marchese and Di Paola, 1975; Pankhurst et al., 2003), which comprises granitoids, migmatites, ectinites, mylonites, amphibolites and basic dykes.

* Poiré, D., Gaucher, C., 2009. Lithostratigraphy. Neoproterozoic-Cambrian evolution of the Río de la Plata Palaeocontinent. In: Gaucher, C., Sial, A.N., Halverson, G.P., Frimmel, H.E. (Eds.): Neoproterozoic-Cambrian tectonics, global change and evolution: a focus on southwestern Gondwana. *Developments in Precambrian Geology*, Vol. 16, Elsevier, pp.

Many authors have contributed to our knowledge of the Tandilia sedimentary succession (see Poiré and Spalletti, 2005 and references therein). The most widely accepted stratigraphic scheme for the Neoproterozoic units was originally proposed by Dalla Salda and Iñiguez (1979) and subsequently modified by Poiré (1987, 1993) and Iñiguez et al. (1989).

From a lithostratigraphic standpoint (Fig. 4.2.1) the Neoproterozoic sedimentary successions comprise the Villa Mónica Formation (Poiré, 1993), Cerro Largo Formation (Poiré, 1993), Las Aguilas Formation (Zalba, 1978), Olavarría Formation (Andreis et al., 1996) and Loma Negra Formation (Borrello, 1966), all of these grouped into the Sierras Bayas Group (Dalla Salda and Iñiguez, 1979; Poiré, 1993). The Cerro Negro Formation (Iñiguez and Zalba, 1974) overlies the Sierras Bayas Group with erosional unconformity and is of latest Neoproterozoic age. Finally, an early Palaeozoic succession is represented by the Balcarce Formation (Dalla Salda and Iñiguez, 1979).

Between the crystalline basement and the sedimentary cover, arkosic and quartz-kaolinitic saprolites represent palaeoweathering surfaces (Zalba et al., 1992). Diamictites of the Sierra del Volcán Formation (Spalletti and del Valle, 1984) occur between the crystalline basement and the Balcarce Formation. Its age is still controversial and will be discussed below. Another unit of restricted areal extent is the sub-surface Punta Mogotes Formation, which was only recorded in the homonymous borehole at 400 m depth in the Mar del Plata area (Fig. 4.1.7).

From a sequence-stratigraphic point of view, Iñiguez et al. (1989) distinguished three Neoproterozoic sequences, one Ediacaran-Cambrian sequence and a final Ordovician sequence, to which Poiré and Spalletti (2005) added one more Neoproterozoic sequence (Diamante sequence). Most of the four sedimentary sequences of the Sierras Bayas Group and Cerro Negro Formation are bounded by regional unconformities (Poiré, 1987, 1993, 2003, 2004; Leanza and Hugo, 1987; Barrio et al., 1991; Gómez Peral, 2008). Thus, six depositional sequences are currently recognised in Tandilia: the Neoproterozoic Tofolletti, Malegni, Diamante and Villa Fortabat sequences, the La Providencia sequence (latest Neoproterozoic-Cambrian) and the Ordovician-Silurian Batán sequence (Spalletti et al., 1996; Andreis and Zalba, 1998; Poiré et al., 2003a;

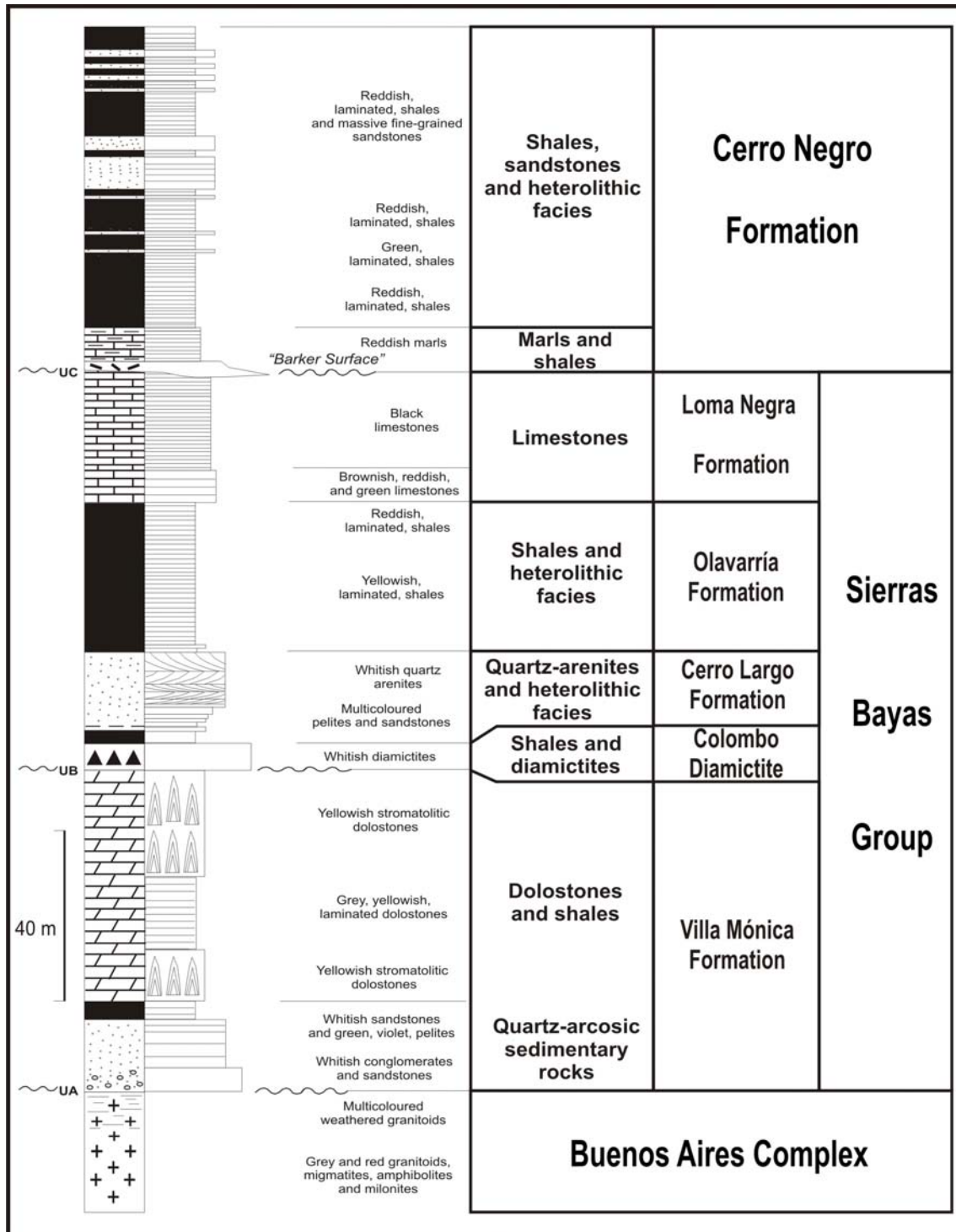


Fig. 4.2.1: Stratigraphic column of the Sierras Bayas Group and Cerro Negro Formation in the western part of Tandilia (Olavarría region).

Poiré and Spalletti, 2005). The confirmation of the stratigraphic position of the Las Aguilas Formation below the limestones of the Loma Negra Formation (Poiré and Spalletti, 2005), as first suggested by Leveratto and Marchese (1983), helps to clarify and simplify the stratigraphy of the sedimentary cover in Tandilia.

4.2.2.1. *Sierras Bayas Group*

The Neoproterozoic sedimentary cover of Tandilia in the Sierras Bayas range (close to Olavarría, Fig. 4.1.7) comprises a 185 m thick succession (Sierras Bayas Group, Fig. 4.2.1) composed of three depositional sequences separated by regional unconformities (Poiré, 1987; 1993) and overlain by the latest Ediacaran Cerro Negro Formation.

4.2.2.1.1. *Weathered crystalline basement*

The oldest regional unconformity (UA, Fig. 4.2.1) is developed between the crystalline basement and the sedimentary cover. An arkosic saprolite occurs there, representing a palaeoweathering surface (Poiré, 1987; Zalba et al., 1992). This altered level is 4 m thick and well exposed in the Villa Mónica Quarry (Olavarría area, Fig. 4.1.7). It is composed of yellow, red, light green, and grey, “pseudostratified”, unconsolidated saprolite. Mineralogically, it is made up of quartz, microcline, plagioclase, biotite, muscovite and abundant Fe-oxides and illite, thus reflecting basement rock mineralogy.

4.2.2.1.2. *Villa Mónica Formation (Tofoletti Sequence)*

The Villa Mónica Formation, the oldest depositional sequence (Tofoletti sequence, 52-70 m thick), comprises two sedimentary facies associations: a siliciclastic member at the base and a carbonate member at the top (Fig. 4.2.1). Whereas the basal member is composed of shallow-marine siliciclastic rocks (conglomerates, quartz-arenites, arkoses, diamictites and shales), the upper member is characterised by shallow marine stromatolitic dolostones and shales. The sandstone unit is interpreted as marine sediments, which covered the weathered and peneplainised RPC during a transgressive event. The contact with the underlying palaeosols (saprolite) is well exposed in quarries in the Olavarría area (Poiré, 1993). Gómez Peral et al. (2005) and Poiré et al (2005b) reported phosphates and hydrothermal pyrophyllite in shales of the upper part of the siliciclastic member.

The overlying dolostone member includes very conspicuous stromatolites representing a stromatolitic platform, and associated clastic, non-stromatolitic carbonates, probably deposited in an intertidal environment. Detailed studies have allowed to identify the following groups and forms (Poiré, 1987, 1989, 1990): *Colonella* fm., *Conophyton* ?*ressotti*, *Conophyton* fm., *Cryptozoon* fm., *Gongylina* fm., *Gymnosolem* fm., *Inzeria*

fm., *Jacutophyton* fm., *Jurusonia nisvensis*, *Katavia* fm., *Kotuikania* fm., *Kussiella* fm., *Minjaria* fm., *Parmites* fm., *Parmites cf. cocrescens* and *Stratifera* fm (see Chapter 4.3). Total organic carbon (TOC) content of dolostone samples from the Villa Mónica Formation average 0.8 % (Gómez Peral et al., 2007). On the basis of stromatolite biostratigraphy, Poiré (1987) assigned dolostones of the Villa Mónica Formation to the Riphean (ca. 800-900 Ma). A Rb-Sr age from interbedded shales of 793 ± 32 Ma was reported by Cingolani and Bonhomme (1988), but its significance remains controversial. Carbon isotope data were presented by Gómez Peral et al. (2007), and will be discussed in Chapter 4.4.

An unconformity (UB, Fig. 4.2.1) separates the Tofoletti sequence (Villa Mónica Formation) from the overlying Malegni sequence (Cerro Largo Formation). The top of the dolostones is irregular, as clearly shown by its variable thickness (36 to 52 m). Above the unconformity UB, predominantly diamictitic deposits occur, which were recently named as Colombo Diamictites (Poiré and Gaucher, 2007, see below).

Diagenesis of the dolostones of the Villa Mónica Formation can be described in terms of two dolomitisation processes that affected these rocks, which probably were developed from low-magnesium calcite precursor (Gómez Peral, 2008). Three different cement generations were recognised: Stage 1, crystallisation of a dolosparitic cement occasionally showing crystals with iron-rich nuclei; Stage 2, quartz cement precipitation in form of euhedral crystals; and Stage 3, high-Mg-calcite cement made up of anhedral to euhedral macrosparitic crystals.

4.2.2.1.3. *Colombo Diamictite*

The Colombo Diamictite (8 m; Fig. 4.2.1) is composed of whitish, massive mudstones and claystones bearing lonestones up to 2.5 m in diameter and showing convolute bedding (Fig. 4.5.1F). Chert breccias and fine orthoconglomerate also occur. This sedimentary package has been considered as the base of the Cerro Largo Formation (Poiré, 1993) but its possible separation as a new unit developed between the Villa Mónica and Cerro Largo formations is being considered. Evidences of syn-sedimentary

tectonism and its stratigraphic position suggest correlation of the Colombo diamictite with the rift-related Las Ventanas Formation in Uruguay (see Chapter 4.5).

4.2.2.1.4. *Cerro Largo Formation (Malegni Sequence)*

The Cerro Largo Formation represents the second depositional sequence (Malegni sequence, 40 m thick). As discussed above, it is not clear whether the Colombo Diamictite belongs to this sequence or represents a separate unit. Finely bedded, varicoloured, glauconitic sandstones, heterolithic facies and cross-bedded quartz-arenites characterise the Cerro Largo Formation. Outcrops of the white, well-sorted quartz-arenites of the Cerro Largo Formation typically occur at the core of the main hills (up to 300m high) of the Olavarría region. This coarsening- and shallowing-upward sequence represents the transition from subtidal nearshore to shoreface environments, suggesting progradational processes.

Trace fossils have been reported from sandstones of the Cerro Largo Formation (*Palaeophycus* isp. and *Didymaulichnus* isp.) but they are dubious and were re-interpreted as desiccation cracks associated to biomats (Porada and Bougouri, 2008; see Chapter 4.3).

The contact between the Cerro Largo and Olavarría formations (Malegni and Diamante sequences) in the Olavarría region is transitional, but in the Barker area there are breccias at the base of the younger sequence (Las Aguilas Formation).

4.2.2.1.5. *Olavarría and Las Aguilas formations (Diamante Sequence)*

In the Olavarría region (Fig. 4.1.7) quartz-arenites of the Cerro Largo Formation pass transitionally into siltstones and claystones of the Olavarría Formation (Andreis et al., 1996; Fig. 4.2.1). This contact is well exposed at the Volquetes roadcut and in the Cementos Avellaneda quarry.

The Olavarría Formation (maximum thickness: 37 m) consists of two sections, from base to top: yellowish heterolithic facies (27 m thick) and reddish claystones (8 m thick). Paleoenvironmental interpretations suggest shallow marine deposits in a transgressive systems tract (TST).

At the Cuchilla de las Aguilas (Barker area, Fig. 4.1.7), three lithofacies occur in the Las Aguilas Formation, from base to top (Zalba et al. 1988, Andreis, 2003): silicified calcareous breccias with silicified oolites and peloids; reddish to withish claystones; and a coarsening-upward, heterolithic (sandstone-shale) sequence. Sedimentary structures in the latter facies include plane-parallel lamination, ripples, hummocky cross bedding and synaeresis and desiccation cracks.

In the Diamante quarry (Barker area), the middle part of the unit comprises red claystones with high iron content (32 to 70% Fe₂O₃) up to 9 m in thickness (Alló, 2001), which could be correlated with other late Neoproterozoic iron deposits, such as the Jacadigo Group in SW Brazil (Mato Grosso do Sul state), or BIF in the lower Arroyo del Soldado Group (Gaucher, 2000; Gaucher et al., 2003, 2004b; Fig. 4.2.5). No glacial deposits are associated to the ironstones of the Las Aguilas Formation.

4.2.2.1.6. *Loma Negra Formation (Villa Fortabat sequence)*

The Loma Negra Formation, the youngest depositional sequence of the Sierras Bayas Group (Villa Fortabat sequence, 40 m thick), is composed almost exclusively of red and black micritic limestones, deposited by suspension fall-out in open marine ramp and lagoonal environments.

This sequence can be divided from base to top in two members based on lithology: reddish micritic limestones (8m thick) followed by black micritic limestones (up to 32 m thick). In these limestones, the following diagenetic processes were recognised by Gómez Peral et al. (2007): *Stage 1*: Recrystallisation of the calcitic precursor to non-planar micrite (1-10 µm), microsparite (10-25 µm) or sparite, preserving the existing microstructure (1-25 µm). *Stage 2*: Chemical dissolution generates porosity in form of veins and voids. *Stage 3*: Precipitation of calcite cements filling porosity. *Stage 4*: irregular stylolites due to pressure–solution. *Stage 5*: Silicification along veins and as partial replacement of carbonate crystals.

On top of the Loma Negra Formation a regional unconformity occurs (Barrio et al., 1991), which has been recently named “Barker Surface” (Poiré and Gaucher, 2007; Fig. 4.2.1; Chapter 4.5). This surface has been related to a sea-level drop that exposed the Loma Negra shelf. Meteoric dissolution of carbonates led to an expressive karstic surface, in which residual clays and brecciated chert accumulated.

Gaucher et al. (2005a) reported for the first time the occurrence of *Cloudina* cf. *C. riemkeae* Germs (1972) in the reddish micritic limestones (see Chapter 4.3). *Helminthopsis* isp. and probable medusa resting traces have been found in the reddish limestones (Poiré et al, 2003a).

No glacial sediments occur at the base of limestones of the Loma Negra Formation, and there is no sign of an erosional unconformity either (Poiré, 2003). Therefore, carbonates of the Loma Negra Formation do not represent “cap carbonates” (Hoffman et al., 1998). Cozzi et al. (2003) discussed the Snowball Earth hypothesis as exemplified by the Ediacaran Shuram Formation of Oman. The Sierras Bayas Group and its correlation with the Arroyo del Soldado Group support a “slushball” Earth scenario, at least for Ediacaran glacial events (Poiré, 2003; Gaucher et al., 2004b; see Chapter 4.5).

4.2.2.2 *Cerro Negro Formation (La Providencia sequence)*

The Cerro Negro Formation (La Providencia Depositional Sequence) unconformably overlies the Loma Negra Formation, filling palaeovalleys and a karst relief. It is a 100-400 m thick unit characterised by reddish, greenish or brown-olive claystones and heterolithic facies (fine-grained sandstone–claystone interbeds), mainly deposited in upper to lower intertidal conditions.

The lower part of the Cerro Negro Formation consists of reddish residual clays, micritic limestones and marls. Chert breccias and phosphatic rocks occur as karst infill (Leanza and Hugo, 1987; Barrio et al., 1991). The upper contact with the Balcarce Formation is not exposed.

The Cerro Negro Formation yielded abundant acritarchs mainly assigned to *Leiosphaeridia*, which are consistent with a latest Ediacaran age for the unit (Cingolani et al., 1991; Gaucher et al., 2005a, see Chapters 4.3 and 9.1).

4.2.2.3 *Balcarce Formation (early Palaeozoic)*

The Balcarce Formation has been studied in detail by del Valle (1987a). It overlies the crystalline basement or the older sedimentary units (Cerro Negro Formation, Sierra del Volcán Diamictite and Punta Mogotes Formation).

The Balcarce Formation (100-400 m thick) is composed of white quartz-arenites and granule-conglomerates with subordinate levels of kaolinitic mudstones and quartz-pebble conglomerates. The geometry of the sandstone beds is sheet-like; most sedimentary bodies are bound by convex-upward surfaces, although some wide channel-like features are also present. Planar and tangential cross-bedding is the dominant structure within sandstone bedsets, and large-scale sigmoidal bodies are common in most sections. Sheet-like and lenticular, sandstone-mudstone interbeds are commonly intercalated with sandstone beds. Trace fossils are abundant at the top bedding surface of sandstones in sandstone-mudstone intercalations. The quarries in the Batán and Chapadmalal area (Fig. 4.1.7) expose the stratigraphic architecture of the Balcarce Formation. Based on their contrasting geometry, two main groups can be defined in this siliciclastic succession: one group is characterised by a subhorizontal stacking pattern (aggradational geometry) and the other shows very well developed depositional clinoforms (progradational geometry).

Tidal processes are inferred for the cross-bedded sandstone facies (bars) and heterolithic facies. Large- to medium-scale, laterally persistent bodies of cross-bedded sandstones, exhibit rhythmic lateral variations in the thickness of foresets and in clay content due to spring and neap tide alternation. Clay drapes covering foresets and other depositional surfaces, herringbone cross-bedding, opposite palaeocurrent trends in successive sedimentary bodies and the occurrence of reactivation surfaces also suggest tidal deposition. The migration and accretion of bidimensional sand bars seem to be controlled by highly asymmetrical time-velocity tidal currents. Subordinated, high-energy storm episodes are suggested by hummocky cross-bedded sandstones, sheet conglomerates armouring previous tidal sand bodies, and heavy mineral concentrations in the wavy sandstone laminae of heterolithic facies.

An epicontinental, shallow marine open shelf is inferred for the Balcarce Formation in Tandilia. Most sedimentary facies were developed in the nearshore and inner shelf environments of a tide-dominated and storm influenced platform. The internal anatomy

of the clinoforms of the Balcarce Formation clearly shows that controls on sedimentation failed to remain uniform as progradation occurred. The interdigitation of interbar heterolithic deposits and subtidal sand bar deposits within this falling stage systems tract suggests that the shallow marine system was affected by higher order oscillations in relative sea level.

The progradation to the south of the clinoforms reported by Poiré et al (2003a) confirms that the margin of the Balcarce basin was located to the north of the Tandilia region (Teruggi, 1964; Dalla Salda and Iñiguez, 1979).

The Balcarce Formation contains abundant and diverse trace fossils (Poiré et al., 1984; del Valle, 1987b; Poiré and del Valle, 1996; Spalletti and Poiré, 2000; Seilacher et al., 2002; see Chapter 4.3). The age of the Balcarce Formation has been considered as early Palaeozoic. The unit lacks body fossils that could allow a more precise age assignment. Trace fossils broadly constrain the age between Cambrian and Silurian, but Seilacher et al. (2002) favor a Lower Silurian age based on the occurrence of *Cruziana ancora*. More recently, Rapela et al. (2007) reported U-Pb detrital zircon ages as young as 475-480 Ma (Early Ordovician) from the Balcarce Formation, suggesting a Late Ordovician to Lower Silurian age for the unit.

4.2.2.4 Sierra del Volcán Diamictite

The Sierra del Volcán Diamictite (Spalletti and del Valle, 1984) is a thin sedimentary unit (4 m) cropping out between the crystalline basement and quartz-arenites of the Balcarce Formation at the homonymous hill. This sequence is composed of two diamictite facies (Spalletti and del Valle, 1984): (a) basal, brownish, coarse- to fine-grained sandstone, sandy shales and yellowish shales with scattered, often faceted dropstones of variable size. Dropstones are polyhedral, often in vertical position, affecting both ripple bedding and lamination structures. (b) Grey and whitish, muddy or sandy, massive diamictite (Fig. 4.5.1G) bearing orthoquartzite and granitoid boulders up to 30 cm in diameter. They grade up-section into poorly-stratified, finer diamictites with local syndepositional corrugations.

On the basis of grain-size studies and field occurrence, Spalletti and del Valle (1984) suggested that the basal sandstones were deposited in a marine, littoral environment,

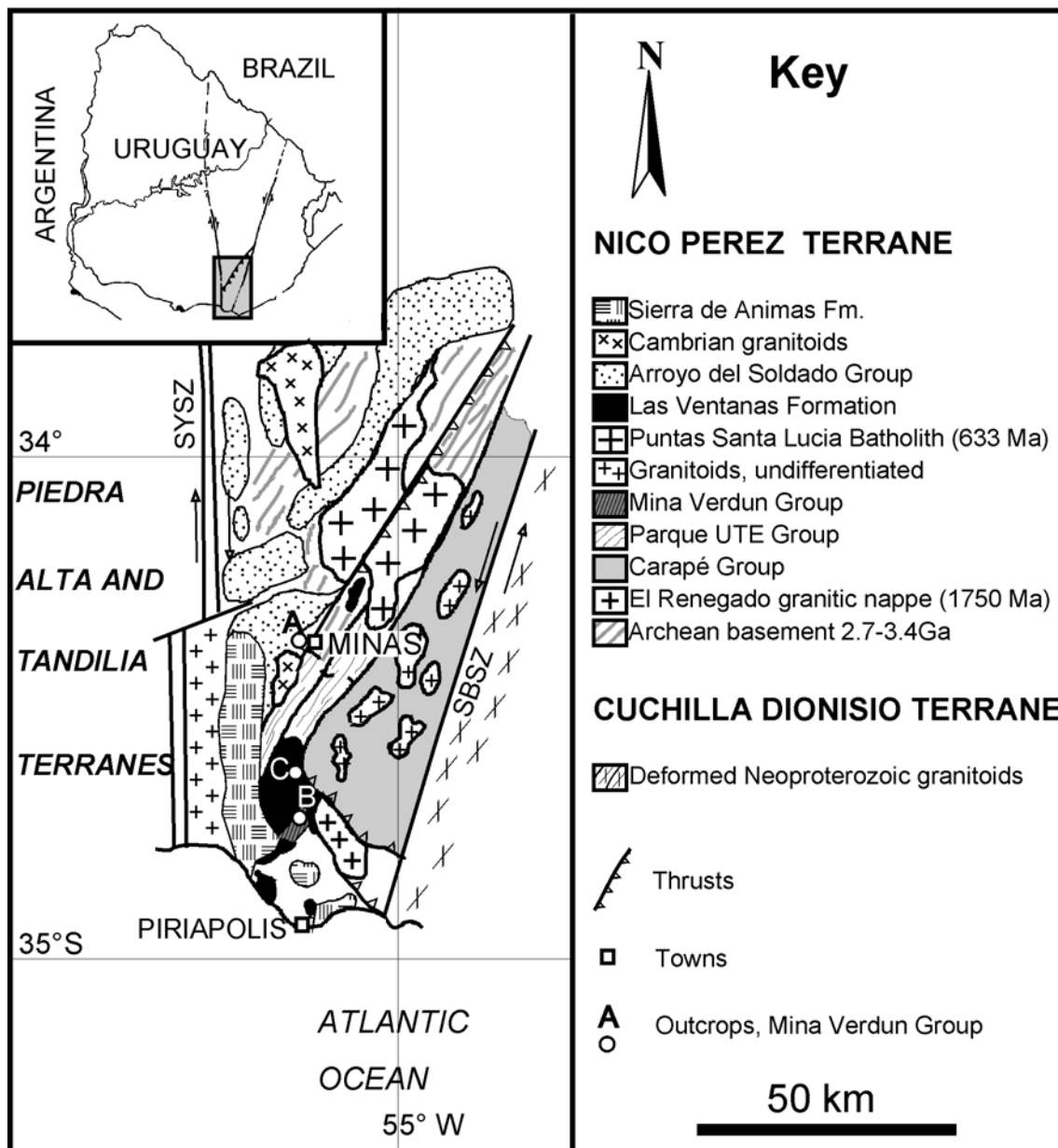


Fig. 4.2.2: Geological map of the southern Nico Pérez Terrane, showing outcrops of the Mina Verdún Group. SYSZ: Sarandí del Yi Shear Zone, SBSZ: Sierra Ballena Shear Zone. A: Mina Verdún Quarry; B: BURGUEÑO Quarry; C: Paso del Molino section. Modified from Gaucher et al. (2008a).

between the beach and the nearshore-offshore transition zone. The diamictites were inferred to be glaciomarine deposits developed near a peneplanised continent in the shallow platform of an epeiric sea.

Based on its stratigraphic position, the Sierra del Volcán Diamictite was originally considered to represent an Ediacaran glacial event (Spalletti and del Valle, 1984; Pazos et al. 2008). However, Poiré and Spalletti (2008) suggested a possible Ashgillian (Hirnantian) age for these glacial deposits on the basis of Lower Ordovician U-Pb ages

of detrital zircons reported by van Staden et al. (in review) from the Sierra del Volcán Diamictite. Therefore, this glacial unit should be referred to the Hirnantian, which has a very important record in southern South America and South Africa.

The geochronological information provided by detrital zircons from the Sierra del Volcán Diamictite and the Balcarce Formation suggests that the kaolinite-rich Balcarce Formation could represent the onset of Silurian greenhouse conditions. The Lower Silurian age for the Balcarce Formation suggested by Seilacher et al. (2002) on the basis of trace fossils is consistent with this scenario.

4.2.3 NICO PEREZ TERRANE, URUGUAY

The Nico Pérez Terrane of Uruguay represents the oldest crustal block belonging to the Río de la Plata Craton. Its pre-Neoproterozoic basement is composed mainly by Archean and Palaeoproterozoic units (Bossi and Ferrando, 2001; Hartmann et al., 2001). Mesoproterozoic volcano-sedimentary successions occur in the southern part of the terrane (see Chapter 4.6).

Thick Neoproterozoic sedimentary successions include the Mina Verdún Group (Poiré et al., 2003b, 2005a), the Las Ventanas Formation (Midot, 1984) and the Arroyo del Soldado Group (Gaucher et al., 1996; Gaucher, 2000).

The controversy regarding the age of the former Lavalleja or Fuente del Puma Group has been dealt with in the previous chapter. It suffices here to say that its Neoproterozoic age has not been thoroughly demonstrated yet, and an important portion is actually Mesoproterozoic in age (Parque UTE Group: Bossi et al., 2008). All Neoproterozoic detrital zircon U-Pb SHRIMP ages reported by Basei et al. (2008) from this unit are moderately to highly discordant, except for only one grain that yielded 715 ± 26 Ma. On the other hand, the same sandstones are intruded by gabbros that yielded an U-Pb zircon age of 1492 ± 4 Ma (Oyhantçabal et al., 2005). Thus, it is still uncertain if there are any Neoproterozoic rocks in the unit formerly known as Lavalleja Group.

4.2.3.1 *Mina Verdun Group*

The Mina Verdún Group (MVG) was erected by Poiré et al. (2003b, 2005a) to include a sedimentary succession exposed in the homonymous mine near Minas (Fig. 4.2.2). The age of the unit is loosely constrained between late Mesoproterozoic and early Neoproterozoic (Tonian) by carbon isotope chemostratigraphy and field relationships (Gaucher et al., 2006, 2008a; see 4.4).

The original definition included four formations (Poiré et al., 2003b, 2005a), from base to top: (a) Don Mario Formation (black shales); (b) La Toma Formation (greenish marls); (c) El Calabozo Formation (limestones); and (d) Gibraltar Formation (massive dolostones, dolomitic marls and subordinate red limestones). It is worth noting that at the stratotype base and top are covered. Subsequently, Gaucher et al. (2004a) assigned carbonates exposed in the Burgueño Quarry, 40 km to the south of Minas (Fig. 4.2.2) to the Mina Verdún Group. There, acid volcanic and pyroclastic rocks are concordantly folded with the rest of the succession, and were grouped into the Cerro de las Víboras and Nueva Carrara formations, respectively (Gaucher et al., 2007a). Thus the currently accepted lithostratigraphy of the MVG comprises six formations, which will be described below in ascending order.

Cerro de las Víboras Formation. Its type area is located in the homonymous hill to the NW of Burgueño Quarry (Gaucher et al., 2007a; Fig. 4.2.2). The Cerro de las Víboras Formation, 80 m in thickness, represents the base of the MVG there; its lower contact is not exposed. The unit is made up of foliated, greenish-gray metarhyolites with β quartz phenocrysts and sericitic matrix. Lapilli-tuff intercalations showing black fiammes are common. Rhyolites are very resistant to weathering and erosion, resulting in the highest and steepest hills of the area. Up section, acid pyroclastic rocks are interbedded with pelites and sandstones of the lower Don Mario Formation, showing that the contact between both units is transitional (Fig. 4.2.3).

Don Mario Formation (Poiré et al., 2003b). It is composed of massive and laminated, organic-rich, black shales 40 m in thickness. At the base, arkosic sandstone and pyroclastic intercalations occur (Gaucher et al., 2007a). XRD analyses of shales show quartz, calcite, clay minerals (illite, chlorite, illite-smectite and chlorite-smectite interlayer) and feldspars. Their geochemistry suggests a high degree of weathering and

MINA VERDUN GROUP

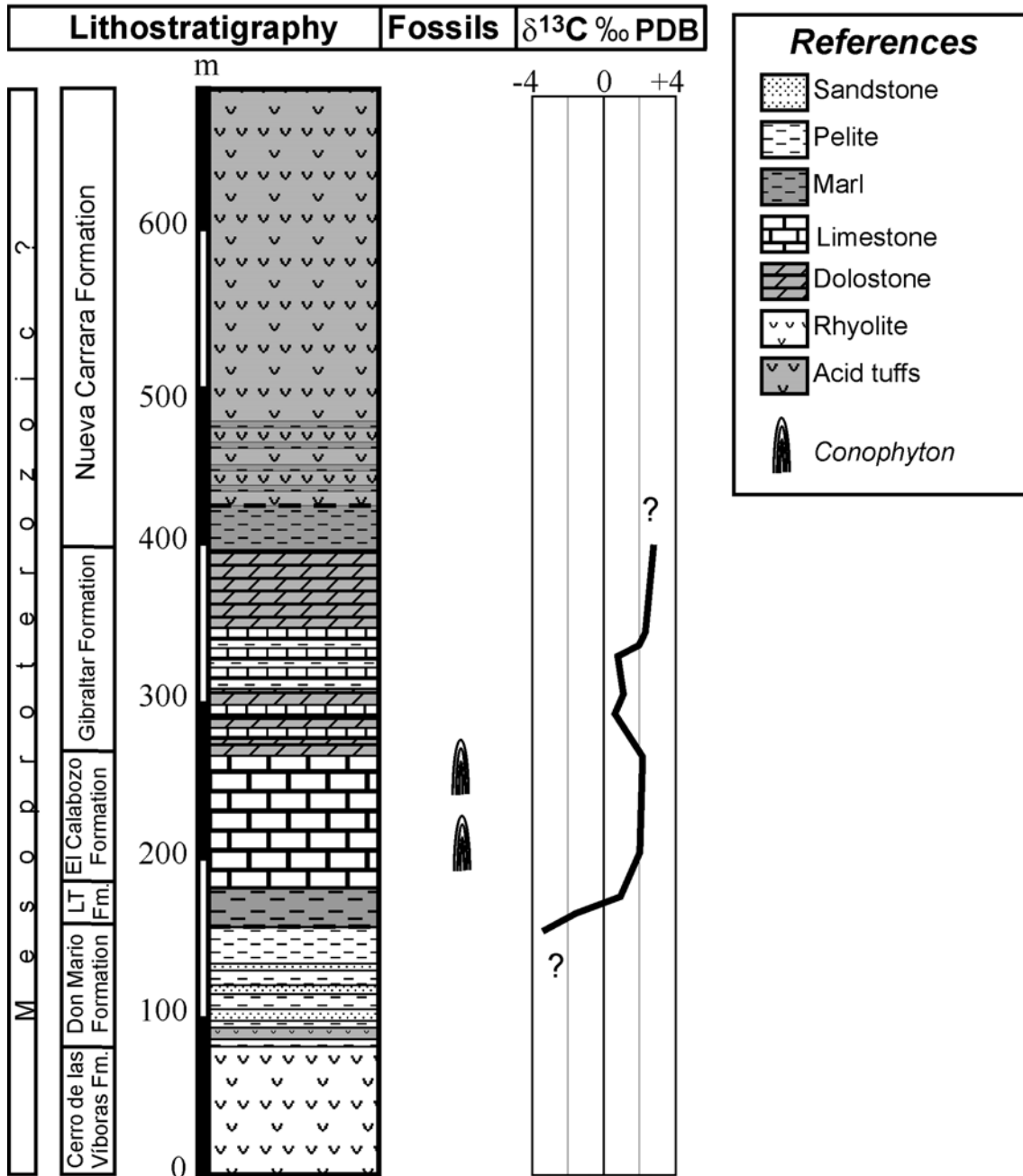


Fig. 4.2.3: Simplified stratigraphic column of the Mina Verdún Group, after Gaucher et al. (2007a). LTFm: La Toma Formation.

reworking in the source area, which was made up of differentiated upper continental crust (Poiré et al., 2005a). Anoxic conditions prevailed in the basin during deposition of the Don Mario Formation, leading to high TOC values and the occurrence of disseminated sulphides.

La Toma Formation. Dark green marls up to 15 m in thickness concordantly overly black shales of the Don Mario Formation (Poiré et al., 2003b; Fig. 4.2.3), representing the onset of carbonate deposition. Marls are composed of up to 45 % SiO₂ of silica and 35 % CaCO₃ (Poiré et al., 2005a). Carbonate minerals are represented by minute grains of both calcite and dolomite, the former being more abundant. Clay minerals are composed of illite and scarce chlorite. The unit is concordantly overlain by massive and stromatolitic limestones of the El Calabozo Formation.

El Calabozo Formation. This is probably the most characteristic and economically important unit of the Mina Verdún Group. It comprises massive, laminated, and stromatolitic limestones typically showing coniform stromatolites (*Conophyton*), up to 170 m in thickness (Poiré et al., 2003b; Fig. 4.2.3). Limestones are very pure, with up to 93 % CaCO₃ at its stratotype or even 95% at Burgueño Quarry (Fig. 4.2.2), making them an important raw material for the cement and lime industries. In thin section they show a homogeneous mosaic of micritic and subsparitic calcite with stylolites and corrugated microbial lamination. Subordinate quartz, feldspars and illite have been observed by XRD. Collapse breccias probably resulting from karstic phenomena also occur. Columnar stromatolites assigned to *Conophyton* fm. are by far the most abundant group. Scarce but well-preserved, dendroid, branching and digitate stromatolites were also observed (see Chapter 4.3). In the Burgueño Quarry, a pervasive decoloration of limestones has been explained by contact metamorphism of nearby granites, causing the reaction of organic carbon with carbonates and affecting its carbon isotopic composition (Gaucher et al., 2006, 2007a; see Chapter 4.4).

Gibraltar Formation (Poiré et al., 2003b). An erosional unconformity separates the El Calabozo Formation from the overlying Gibraltar Formation (Fig. 4.2.3), which has been interpreted as a karstic palaeorelief by Poiré et al. (2005a). This contact is also deformed due to boudinage of the more competent dolostones, as shown by Gaucher et al. (2007a). A 4 m-thick diagenetic breccia is often at the base of the Gibraltar Formation. The unit, 60 m in thickness, comprises the following lithologies from base to top: (a) black marls and shales with pink limestone intercalations, (b) pink and red limestones, often dolomitic, and (c) massive, mainly light yellow or reddish dolostones, showing abundant carbonate veins. Dolostones are the most abundant and characteristic

rocks of the Gibraltar Formation. They are in turn overlain by green marls at Burgueño Quarry, where the top of the unit is well exposed (Gaucher et al., 2007a).

The occurrence of stromatolites in the Gibraltar Formation was recently recognised in two drill-cores at its stratotype (Poiré et al., 2006). The strong dolomitisation of the Gibraltar Formation and its poor exposure do not allow a detailed palaeoenvironmental interpretation (Poiré et al., 2005a). However, the preservation of the stromatolites in cores is surprisingly good and can be used to determine environmental conditions. *Gymnosolenid* and few *Conophyton* stromatolites have been so far identified.

Nueva Carrara Formation. The Gibraltar Formation transitionally passes into greenish tuffs of the Nueva Carrara Formation (Gaucher et al., 2007a; Fig. 4.2.3). Green, chloritic marls occur at the base, and are often interbedded with green, acid lapilli-tuffs. Up section, 300 m lapilli-tuffs with subordinate metamarl intercalations occur. Tuffs show β quartz and alkali feldspar phenocrysts and a matrix composed of chlorite and quartz. Pumice and fiammes made up of chlorite, muscovite and quartz occur, as well as accessory lithic and crystal fragments of dacite, microcline and perthitic feldspars.

In all outcrop areas, the Mina Verdún Group is overlain with angular and erosional unconformity by polymictic orthoconglomerates and red diamictites of the Las Ventanas Formation (Blanco and Gaucher, 2005; Gaucher et al., 2008a), of early Ediacaran age. Thus, the MVG may be considerably older than early Ediacaran, probably of Mesoproterozoic age (Gaucher et al., 2006, 2008a; see Chapter 4.4). Both the MVG and Las Ventanas Formation are intruded by a subvolcanic, longitudinal (E-W) mafic dyke swarm that yielded a K-Ar whole rock datum of 485 ± 13 Ma considered as a minimum age (Gonzalez et al., 2004). These dykes may belong to the 581 ± 13 Ma Nico Pérez mafic dyke swarm (see Chapter 4.6).

4.2.3.2 *Las Ventanas Formation*

The Las Ventanas Formation, erected and first mapped by Midot (1984), is composed of a thick, fining- and thinning-upward volcanosedimentary sequence covering an area of ca. 120 km² of the southern Nico Perez Terrane (Blanco and Gaucher, 2005; Fig. 4.2.2).

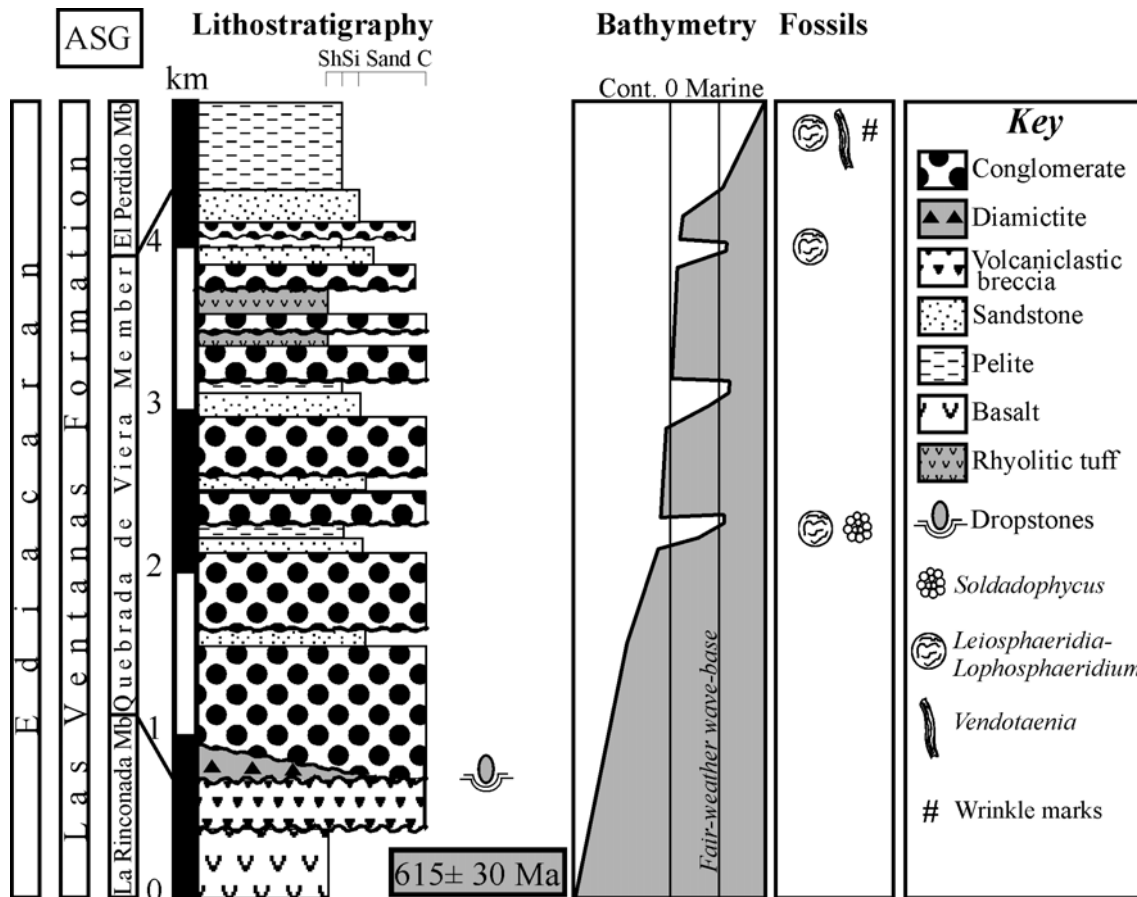


Fig. 4.2.4: Synthetic stratigraphic column of the Las Ventanas Formation, modified from Gaucher et al. (2008a). ASG: Arroyo del Soldado Group.

Polymictic orthoconglomerates predominate, passing into sandstones and siltstones up-section. The structure in its type area corresponds to a broad syncline (Cerro Las Ventanas Syncline) with an axis oriented S20W.

Masquelin and Sánchez Bettucci (1993) first correlated outcrops of Las Ventanas Formation with the predominantly conglomeratic Playa Hermosa Formation, which they considered Neoproterozoic or Cambrian in age. This correlation has been confirmed by Blanco and Gaucher (2005), which regarded the Playa Hermosa Formation as a deep-water equivalent of the Las Ventanas Formation. A detailed stratigraphic and sedimentologic analysis of the Playa Hermosa Formation is due to Pazos et al. (2003, 2008).

The Las Ventanas Formation is subdivided into the following members, from base to top (Blanco and Gaucher, 2005; Fig. 4.2.4):

- (a) La Rinconada Member, made up of basic volcanics, mainly basalts (44-52% SiO₂), volcanoclastic breccias and subordinate basic tuffs, reaching 500 m in thickness. Basalts are often vesicular and represent stacked flows. Volcanoclastic breccias (locally named “Darwin breccia”) are more common in the upper part of the unit.
- (b) The Quebrada de Viera Member, up to 3.800 m in thickness, overlies the Rinconada Member with erosional unconformity. It is composed of polymictic, mostly clast-supported orthoconglomerates, grading to conglomerate-sandstone intercalations up section. Glacigenic, reddish and greenish diamictites occur only at the base of the unit (Fig. 4.2.4), filling a karst surface developed on top of the Mina Verdún Group (Gaucher et al., 2008a; see Chapter 4.5). Lithoclast composition of orthoconglomerates shows provenance from a bimodal volcanic and granitic source area. Up to 37 % basalt clasts in certain conglomerate levels suggest little chemical weathering and/or a steep palaeorelief. Cross bedding and normal grading are common sedimentary features of the conglomerates. Sandstones are texturally and mineralogically immature arkoses, often showing hummocky cross-stratification at the top of this member (Blanco and Gaucher, 2005). Whereas basalt clasts are more common at the base, rhyolitic clasts and intercalated rhyolitic tuffs occur at the top.
- (c) El Perdido Member concordantly overlies the Quebrada de Viera Member and is characterised by laminated siltstones and shales up to 600 m in thickness (Fig. 4.2.4), showing reddish colours at the base and green and grey colours at the top (Blanco and Gaucher, 2005). Abundant pyrite at the top of this member indicates dysoxic or anoxic conditions.

Conglomerates of the Las Ventanas Formation represent alluvial fan and fan delta deposits (Pecoits et al., 2004; Blanco and Gaucher, 2005), on the basis of (Einsele, 2000): (1) predominance of unchannelised, conglomerate deposits; (2) occurrence of both stream deposits (predominant) and sedimentary gravity flows; (3) great thickness of conglomerates; and (4) intercalation of conglomerates with marine siltstones and sandstones at the top of the Quebrada de Viera Member.

Thus, the Las Ventanas Formation represents a deepening-upward sequence (Fig. 4.2.4), recording evolution from an alluvial fan-dominated environment to shallow marine conditions with occasional storms. Sedimentary structures, thickness of conglomerate deposits, abundance of unstable lithoclasts and immaturity of sandstones point to a steep palaeorelief. Bimodal volcanism evolves from mainly basic (basalts) at the base to

mainly acid (rhyolitic) at the top. An extensional geotectonic setting, possibly a rift, was postulated by Blanco and Gaucher (2005) for the Las Ventanas Formation (see Chapter 4.6). In favour of this hypothesis are the bimodal, synsedimentary volcanism, strong palaeorelief, great thickness of alluvial fan conglomerates and the evolution from continental to open marine environments.

Acritarchs and other organic-walled microfossils occurring in palynological macerations of siltstones of Las Ventanas Formation (Quebrada de Viera and El Perdido members,) were described and illustrated by Blanco and Gaucher (2005) and Gaucher et al. (2008a, see Chapter 4.3).

As noted already by Blanco and Gaucher (2005), the lithostratigraphic column presented by Pecoits (2003), although correct in its lower part, duplicates the stratigraphy of the Las Ventanas Formation by the failure to recognise the structure, which is an open syncline. Grading in the lower half of the column presented by Pecoits (2003) and Pecoits et al. (2008) is normal, and turns to reverse in the upper half. Likewise, clast composition is symmetrical, the upper part being a specular image of the lower. This is easily explained by tectonic duplication, as mentioned by Blanco and Gaucher (2005).

The recent proposal of the “Maldonado Group”, made up of the Las Ventanas and San Carlos formations (Pecoits et al., 2008), also deserves consideration here. The Las Ventanas and San Carlos formations crop out in different tectonostratigraphic terranes (Nico Perez Terrane and Cuchilla Dionisio Terrane, respectively) and are separated by the crustal-scale Sierra Ballena Shear Zone, comprising up to 10 km-thick ultramylonites. Both units were never observed in contact. U-Pb ages of detrital zircons, Sm-Nd isotopic signature of granitoids and its general geologic evolution show that the Cuchilla Dionisio Terrane (=Punta del Este Terrane) is an exotic block of African affinity (Bossi and Gaucher, 2004; Basei et al., 2005, 2008; Gaucher et al., 2008b; see Chapter 4.6). According to the International Stratigraphic Guide, a group is “a sequence of two or more **contiguous or associated** formations with significant and diagnostic lithologic properties in common” (Salvador, 1994: p. 35). None of these criteria apply to the San Carlos and Las Ventanas formations, therefore the Maldonado Group is not a valid lithostratigraphic unit.

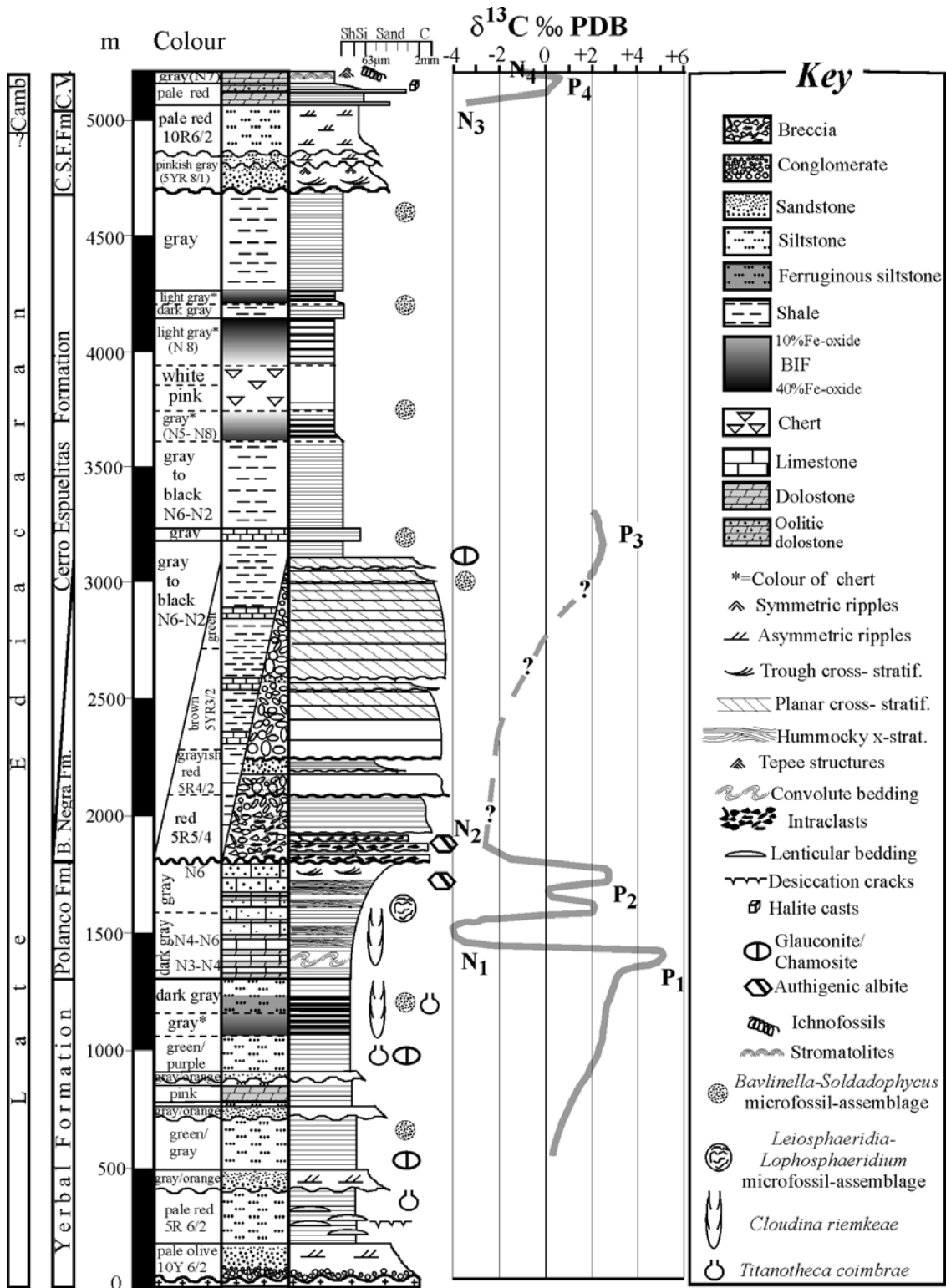


Fig. 4.2.5: Simplified stratigraphic column of the Arroyo del Soldado Group, modified from Gaucher (2000). CSF Fm: Cerros San Francisco Formation. CV: Cerro Victoria Formation. A more detailed discussion of the bio- and chemostratigraphic data is given in Chapters 4.3 and 4.4, respectively.

4.2.3.3. Arroyo del Soldado Group

The Arroyo del Soldado Group (ASG) is a key unit for unravelling the Ediacaran to Cambrian evolution of southwestern Gondwana because of its position at the boundary between the African and South American cratons and also due to its fossiliferous nature. It was originally described from restricted areas in the northwestern and central Nico Pérez Terrane (Gaucher et al. 1996). The ASG occurs over an area in excess of 30.000 km² (Figs. 4.1.3 and 4.2.2), with a maximum NS dimension of 320 km and EW dimension of 130 km. Outcrops of the ASG cover an area of ca. 3000 km² (Gaucher, 2000). The main geological features (lithostratigraphy, sedimentology, deformation) were described in detail by Gaucher (2000), therefore only the lithostratigraphic framework and recently acquired data will be discussed here.

4.2.3.3.1. Lithostratigraphy and sedimentary environments

The ASG is a platform succession unconformably overlying a mainly Archean to Neoproterozoic basement in the Nico Pérez Terrane, representing the drift phase of the oceanic basin opened during rifting represented by the Las Ventanas Formation (Blanco and Gaucher, 2005; Gaucher et al., 2008a). The group is characterised by an alternation of predominantly siliciclastic (Yerbal, Barriga Negra, Cerro Espuelitas and Cerros San Francisco formations) and thick carbonate units (Polanco and Cerro Victoria formations), which extend laterally over hundreds of kilometers with little facies variability. These formations, in ascending order, are characterised as follows (Gaucher, 2000; Fig. 4.2.5):

- (a) The *Yerbal Formation* is a fining-upward, mainly siliciclastic unit more than 1500 m thick. Sandstones occur at the base, siltstones dominate up section, and BIF, dolostone and chert intercalations occur at the top (Gaucher et al., 2004b).
- (a) The *Polanco Formation* is composed of ~900 m of bluish-gray, pure limestones, limestone-dolostone rhythmites and, more rarely, pure dolostones (Gaucher, 2000; Gaucher et al., 2004b). It is the most geographically extensive unit of the ASG.
- (b) The *Barriga Negra Formation* is composed of 1500 m of conglomerates and breccias. Whereas carbonate breccias directly derived from the underlying Polanco Formation characterise the base, an increasing proportion of extra-basinal (basement) clasts occur up section (Gaucher, 2000). At the top, chloritic and glauconitic sandstones and conglomerates are interbedded with siltstones containing marine acritarchs (Gaucher, 2000; Fig. 4.2.5). Thus, depositional environments of

the Barriga Negra Formation are transitional at the base, continental in the middle and again marine at the top, recording a regression and shelf exposure followed by renewed flooding. The unit wedges out toward the east, showing that the deeper areas of the shelf were not exposed during the regression.

- (c) The *Cerro Espuelitas Formation* is 1200 m thick and composed of intercalated dark shales and chemical sediments, such as oxide-facies BIF and chert (Gaucher, 2000).
- (d) The ~300 m-thick *Cerros San Francisco Formation* is composed of mature quartz arenites and subarkoses (Montaña and Sprechmann, 1993; Gaucher, 2000). It rests with erosional unconformity on shales of the Cerro Espuelitas Formation or directly on the pre-Ediacaran basement.
- (e) The *Cerro Victoria Formation* is 400 m thick and comprises stromatolitic dolostones with trace fossils indicative of Cambrian age (Sprechmann et al., 2004; Gaucher et al., 2007b).

Volcanic, volcanoclastic and pyroclastic rocks do not occur in the ASG. Carbonates were deposited on pericontinental, storm-dominated, carbonate ramps. Siliciclastic units were deposited in a basin with a gentle palaeoslope, resulting in sandstones being texturally and mineralogically mature quartz arenites or subarkoses (Gaucher, 2000). Palaeocurrent analyses, thicknesses and facies distribution suggest that the basin opened to the east and south. Together, these features indicate a stable, Atlantic-type continental margin as the geotectonic setting of the Arroyo del Soldado basin (Gaucher, 2000), which has been recently confirmed by detrital zircon U-Pb ages of Arroyo del Soldado sandstones (Gaucher et al., 2008b; Blanco et al., in press; see Chapter 4.6). This platform was probably the depositional setting of the correlative Sierras Bayas Group of Argentina (Gaucher et al., 2005a) and the Corumbá Group of Brazil (Gaucher et al., 2003; Fig. 4.1.2).

In a recent paper, Pecoits et al. (2008) confirm the stratigraphic scheme put forward by Gaucher (2000), but correlate the Cerros San Francisco Formation with the Piedras de Afilas Formation, thereby synonymizing both units as Piedras de Afilas Formation. However, a number of reasons militate against this correlation (Gaucher et al., 2008b), which will be discussed in section 4.2.4. Correlations of the ASG with other units of the Río de la Plata Craton will be dealt with in Chapters 4.3-4.4.

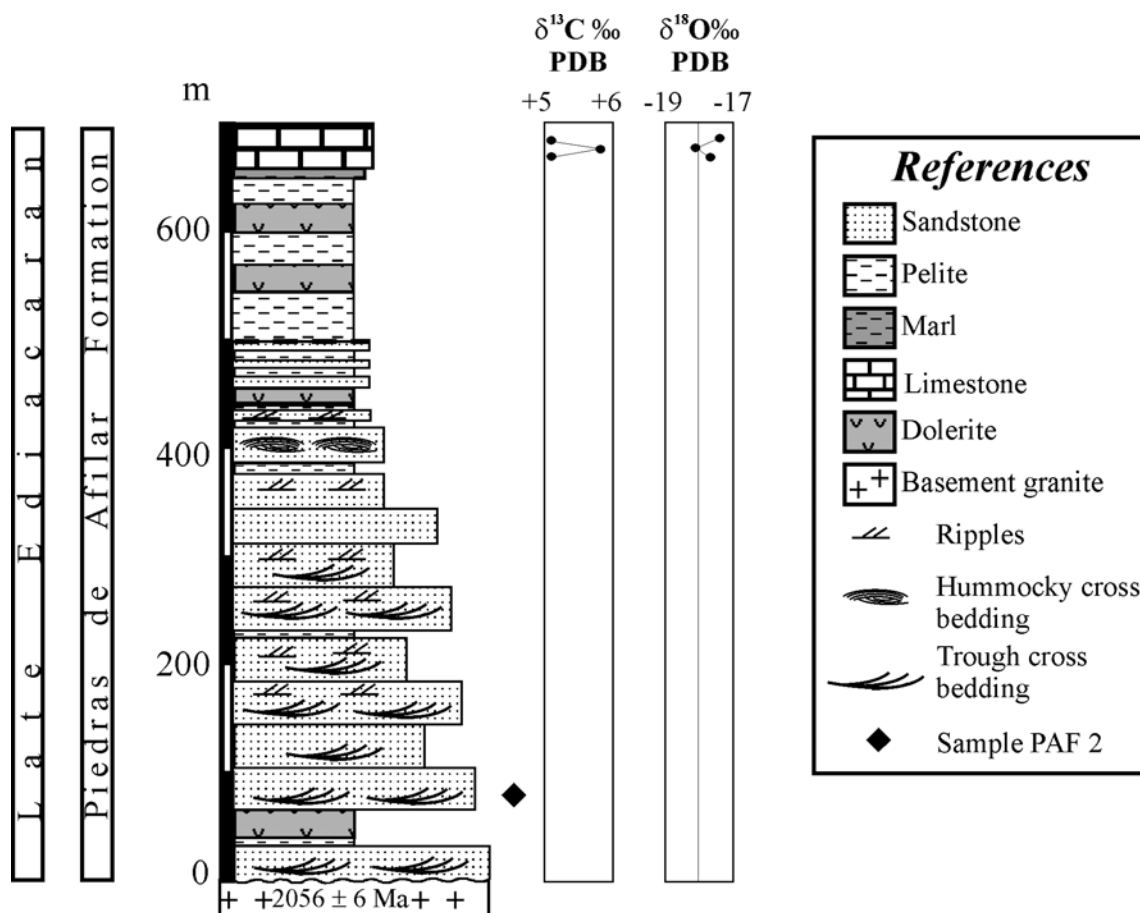


Fig. 4.2.6: Simplified stratigraphic column of the Piedras de Afilar Formation, based on Pamoukaghlián et al. (2006), Coronel et al. (1982), Gaucher et al. (2008b) and Pecoits et al. (2008). Basement ages according to Santos et al. (2003). U-Pb detrital zircon ages for sample PAF 2 were reported by Gaucher et al. (2008b, see Chapter 4.6).

4.2.3.3.2. Age constraints

The age of the ASG is radiochronologically constrained by: (a) a maximum U-Pb SHRIMP age of $633 \pm 12 \text{ Ma}$ for the Puntas del Santa Lucía pluton (Hartmann et al., 2002a), which the ASG unconformably overlies; (b) a better maximum age constraint is provided by an U-Pb SIMS zircon age of $583 \pm 7 \text{ Ma}$ for the Arroyo Mangacha Granite, which is overlain by sandstones of the Cerros San Francisco Formation (Gaucher et al., 2008b); (c) a detrital zircon U-Pb LA-ICP MS age of $566 \pm 8 \text{ Ma}$ for sandstones of the Barriga Negra Formation (Blanco et al., in press); and (d) a minimum Rb-Sr isochron age of $532 \pm 11 \text{ Ma}$ for the Guazunambí Granite (Kawashita et al., 1999a), which intrudes the ASG. In addition, K-Ar ages ranging between 532 ± 16 and $492 \pm 14 \text{ Ma}$ have been reported for the recrystallisation of pelites within the group (Cingolani et al., 1990; Gaucher, 2000). Thus, deposition of the ASG is bracketed between 583 and 532 Ma

(late Ediacaran-Lower Cambrian), with the Barriga Negra and younger formations being deposited after 566 ± 8 Ma.

Biostratigraphic data further allow to assign the lower and middle ASG to the late Ediacaran, and the upper ASG to the lowermost Cambrian. Most significant is the occurrence in the lower-middle ASG of an assemblage of shelly fossils including *Cloudina riemkeae* Germs (1972) and *Titanotheca coimbrae* Gaucher and Sprechmann (1999), as well as a low-diversity acritarch assemblage dominated by *Bavlinella faveolata*, *Soldadophycus* and *Leiosphaeridia* (Gaucher et al., 1996, 1998, 2004c; Gaucher, 2000; Fig. 4.2.5, see Chapter 4.3). Carbon and strontium isotope chemostratigraphy supports a late Ediacaran age for the lower-middle ASG (Gaucher et al., 2004b, 2004d, 2007b). Substantial evidence supports a Lower Cambrian correlation for the Cerro Victoria Formation. Large ichnofossils representing complex infaunal burrow systems, classified by Sprechmann et al. (2004) as *Thalassinoides* isp., are common in the formation. Carbon-isotopic curves for the Cerro Victoria Formation (Gaucher et al., 2007b) closely resemble Lower Cambrian (lower Nemakyt-Daldyn or Fortunian) $\delta^{13}\text{C}$ global curves between 542 and 535 Ma (Derry et al., 1994; Montañez et al., 2000; Maloof et al., 2005), characterised by low-amplitude, high-frequency secular variations (see Chapter 4.4).

4.2.4 PIEDRAS DE AFILAR FORMATION

The Piedras de Afilar Formation crops out in a restricted area of the recently recognised Tandilia Terrane (Bossi et al., 2005; Figs. 4.1.3, 4.1.6) in southern Uruguay (Jones, 1956; Bossi et al., 1998; Spoturno et al., 2005). It is more than 700 m thick and made up of a basal sandstone succession that passes upwards into shales and then carbonates at the top (Fig. 4.2.6). The sedimentary succession, as described by Coronel et al (1982), is composed of, in ascending order: 350 m of predominantly quartzitic sandstones and shale interbeds with cross stratification, ripple marks and flute casts common in the sandstones; 350 m of laminated siltstones; and fine-grained, intensively folded limestones. Dolerite sills intrude and significantly alter the stratigraphic succession at many levels (Fig. 4.2.6). According to Pecoits et al. (2008), the Piedras de Afilar Formation comprises, at its stratotype, 430 m coarse to very fine, cross-bedded

sandstones, followed by 150 m of laminated siltstones and capped by 25 m of greenish and grey limestones (Fig. 4.2.6).

The basement of the Piedras de Afilas Formation is made up of amphibolite-facies metasedimentary rocks of the Montevideo Formation (Pando Belt), as well as the Soca rapakivi granite (Oyhantçabal et al., 1998). The intrusion yielded an U-Pb SHRIMP zircon age of 2056 ± 6 Ma (Santos et al., 2003). The Piedras de Afilas Formation rests with erosional and angular unconformity on these basement units, thus representing an unmetamorphosed sedimentary cover of the Río de la Plata Craton.

4.2.4.1. Age and correlations

A maximum age constraint is provided by the youngest detrital zircon U-Pb ages of 1008 ± 13 Ma and 1006 ± 31 reported by Gaucher et al. (2008b) for the Piedras de Afilas Formation. Pamoukaghlián et al. (2006) reported $\delta^{13}\text{C}$ values between +5.05 and +5.80‰ PDB for carbonates in the upper Piedras de Afilas Formation (Fig. 4.2.6). $\delta^{13}\text{C}$ values between +5 and +6 ‰ PDB are more often encountered in Neoproterozoic carbonate successions, especially in the Cryogenian and Ediacaran (Halverson et al., 2005; see Chapter 10). Cambrian carbonates, on the other hand, are characterised by $\delta^{13}\text{C}$ values between -4 and +2.5‰ PDB (Montañez et al., 2000). Therefore, the data reported by Pamoukaghlián et al. (2006) suggest a Cryogenian or Ediacaran age for the Piedras de Afilas Formation.

Pecoits et al. (2008) suggest correlation of the Piedras de Afilas and Cerros San Francisco formations on lithostratigraphic grounds alone. Four independent lines of evidence militate against this correlation: (a) $\delta^{13}\text{C}$ values of carbonates of the upper Piedras de Afilas Formation (Pamoukaghlián et al., 2006) are much higher than $\delta^{13}\text{C}$ of carbonates overlying the Cerros San Francisco Formation (Cerro Victoria Formation), which range between -3 and +0.6 ‰ PDB (Gaucher et al., 2007b; Fig. 4.2.5); (b) stromatolitic carbonates of the Cerro Victoria Formation are distinctively different from marls and clastic carbonates of the upper Piedras de Afilas Formation; (c) source areas of both units were very different; the Piedras de Afilas Formation being derived mainly from Meso- and Palaeoproterozoic sources, and the Cerros San Francisco Formation from Archean and Palaeoproterozoic rocks (Gaucher et al., 2008b; Blanco et al., 2007,

in press; see Chapter 4.6); and (d) the poorly-preserved acritarchs reported by Pecoits et al. (2008) for the Piedras de Afilar Formation actually favor correlation with the Yerbal Formation and not with the Cerros San Francisco Formation, as pointed out by Pamoukaghlian et al. (2006).

The available data point to a correlation of the Piedras de Afilar Formation with the Yerbal Formation (ASG) and the Cerro Largo-Olavarría formations (Sierras Bayas Group), because: (a) they are all overlain by carbonates characterised by positive $\delta^{13}\text{C}$ values up to 5.5‰ PDB (Gaucher et al., 2004b; Pamoukaghlián et al., 2006; Gómez Peral et al., 2007; see 4.4); (b) they represent deepening-upward sequences, with quartz-arenites at the base and pelites at the top; (c) in all cases provenance is dominated by Palaeoproterozoic detritus and a significant (but variable) proportion of Mesoproterozoic grains (Gaucher et al., 2008b; Blanco et al., in press; see Chapter 4.6); and (d) the depositional environment of the three units was an Atlantic-type shelf deepening to the east and south. According to this correlation, the Piedras de Afilar Formation represents a shallower lateral equivalent of the Yerbal Formation.

Regardless the preferred correlation of the Piedras de Afilar Formation, the latter should be maintained as a separate unit. The formation has its own, distinctive lithologic characteristics, its outcrops are separated from the possible correlates by the crustal-scale Sarandí del Yí shear zone, and the name is well established in the literature since Jones (1956). More specifically, the International Stratigraphic Guide states that "...in case of subtle lateral transition, where boundaries have to be arbitrary, the use of one name in a broad sense may be preferable to the use of two or more names" (Salvador, 1994: p. 40). Considering that important facies changes (not "a subtle transition") and a large shear zone occur between the Piedras de Afilar Formation (not an "arbitrary boundary") and its possible correlates, it becomes clear that the name should be maintained.

If the correlation of the unit with the Yerbal Formation is confirmed, as seems the case, the Piedras de Afilar Formation may be included in the Arroyo del Soldado Group as a lateral equivalent of the Yerbal Formation in the Tandilia Terrane. This is at odds with the proposal of Pecoits et al. (2008), which suggested to apply the name "Piedras de Afilar Formation" also to the outcrops assigned to the Cerros San Francisco Formation in the Nico Pérez Terrane.

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BIOSTRATIGRAPHY*

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4.3.1. INTRODUCTION

Neoproterozoic stromatolites were first recognised in the Río de la Plata Craton (RPC) by González Bonorino (1954) in the Sierras Bayas Group (Argentina). The first body fossils were also described from this unit by Fairchild (1978) and Pothe de Baldis et al. (1983), consisting of organic-walled microfossils (acritarchs). In Uruguay, Montaña and Sprechmann (1993) were the first to recognise Cambrian stromatolites in the Cerro Victoria Formation. The first report of Neoproterozoic body fossils (acritarchs) in Uruguay is due to Gaucher and Schipilov (1994) and Gaucher et al. (1996). Since these first reports, around a dozen papers appeared that deal with Neoproterozoic and Cambrian fossils in the RPC. The sedimentary successions deposited on and around the RPC record the existence of a diversified biota in Ediacaran times, including cyanobacteria, acritarchs, shelly fossils, stromatolites, trace fossils and –possibly- rare members of the Ediacara fauna.

4.3.2. TONIAN AND CRYOGENIAN FOSSIL RECORD

Due to the fact that rifting of the Río de la Plata Craton from Rodinia begun around 650 Ma (see Chapter 4.6), the early Neoproterozoic sedimentary record is very poor and controversial. Two units of potentially lower Neoproterozoic age are fossiliferous, namely the Mina Verdún Group in Uruguay and the Villa Mónica Formation (Sierras Bayas Group) in Argentina.

* Gaucher, C., Poiré, D., 2009. Biostratigraphy. Neoproterozoic-Cambrian evolution of the Río de la Plata Palaeocontinent. In: Gaucher, C., Sial, A.N., Halverson, G.P., Frimmel, H.E. (Eds.): Neoproterozoic-Cambrian tectonics, global change and evolution: a focus on southwestern Gondwana. *Developments in Precambrian Geology*, Vol. 16, Elsevier, pp.

The Mina Verdún Group is a marine succession occurring in the Nico Pérez Terrane (Figs. 4.2.2-4.2.3). Limestones of the Calabozo Formation contain an almost monospecific assemblage of stromatolites, dominated by *Conophyton* (Poiré et al., 2003b, 2005a; Gaucher et al., 2004a, 2006, 2007a). They consist of unbranching, subcylindrical columnar stromatolites with typical conical laminae (Figs. 4.3.1A-B). Their profile is angulate to geniculate, and their cross section circular to oblong (Poiré et al., 2005a). Column height reaches 70 cm, and diameter 25 cm. Rare occurrences of digitate columnar stromatolites were reported from the Mina Verdún Group by Poiré et al. (2005a, 2006).

A minimum age of the Mina Verdún Group is provided by the unconformably overlying Las Ventanas Formation, of early Ediacaran age (Blanco and Gaucher, 2005; Gaucher et al., 2008a; see below). Carbon isotopic data presented by Gaucher et al. (2006, 2007a) show low-amplitude $\delta^{13}\text{C}$ oscillations and a plateau around 2 ‰ PDB, thus consistent with a late Mesoproterozoic to Tonian age for the Mina Verdún Group (Kah et al., 1999; see Chapter 4.4). Other *Conophyton*-dominated stromatolite assemblages include: (1) the Itaiacoca Group (Sallun Filho and Fairchild, 2005) of SE-Brazil (Ribeira Belt), with a minimum U-Pb TIMS age of 934 ± 36 Ma (Siga Jr. et al., 2006); (2) the Paranoá Group of central Brazil (Brasília Belt), possibly deposited between 1200 and 900 Ma (Sallun Filho & Fairchild 2005), and showing a C-isotope curve similar to the Mina Verdún Group (Santos et al., 2000), and (3) the Uluksan and Nunatsiaq groups of the basal Bylot Supergroup (Canada), dated by Pb-Pb method at 1200 Ma (Sherman et al. 2002). According to Walter (1994), the period between 1350 and 1050 Ma marks the highest abundance and diversity of coniform stromatolites. Thus, it is still uncertain if the Mina Verdún Group was deposited in the late Mesoproterozoic or early Neoproterozoic. U-Pb dating of interbedded rhyolites and tuffs described by Gaucher et al. (2007a) might help clear this issue.

The Villa Mónica Formation (lower Sierras Bayas Group) hosts a diverse stromatolite assemblage described in detail by Poiré (1987). Stromatolites were grouped into 16 morphotypes by Poiré (1987, 1993), namely (Figs. 4.3.1C-F): *Colonella* fm., *Conophyton resotti*, *Conophyton* fm., *Cryptozoon* fm., *Gongylina* fm., *Gymnosolen* fm., *Inzeria* fm., *Jacutophyton* fm., *Jurusonia nisvensis*, *Katavia* fm., *Kotuikania* fm.,

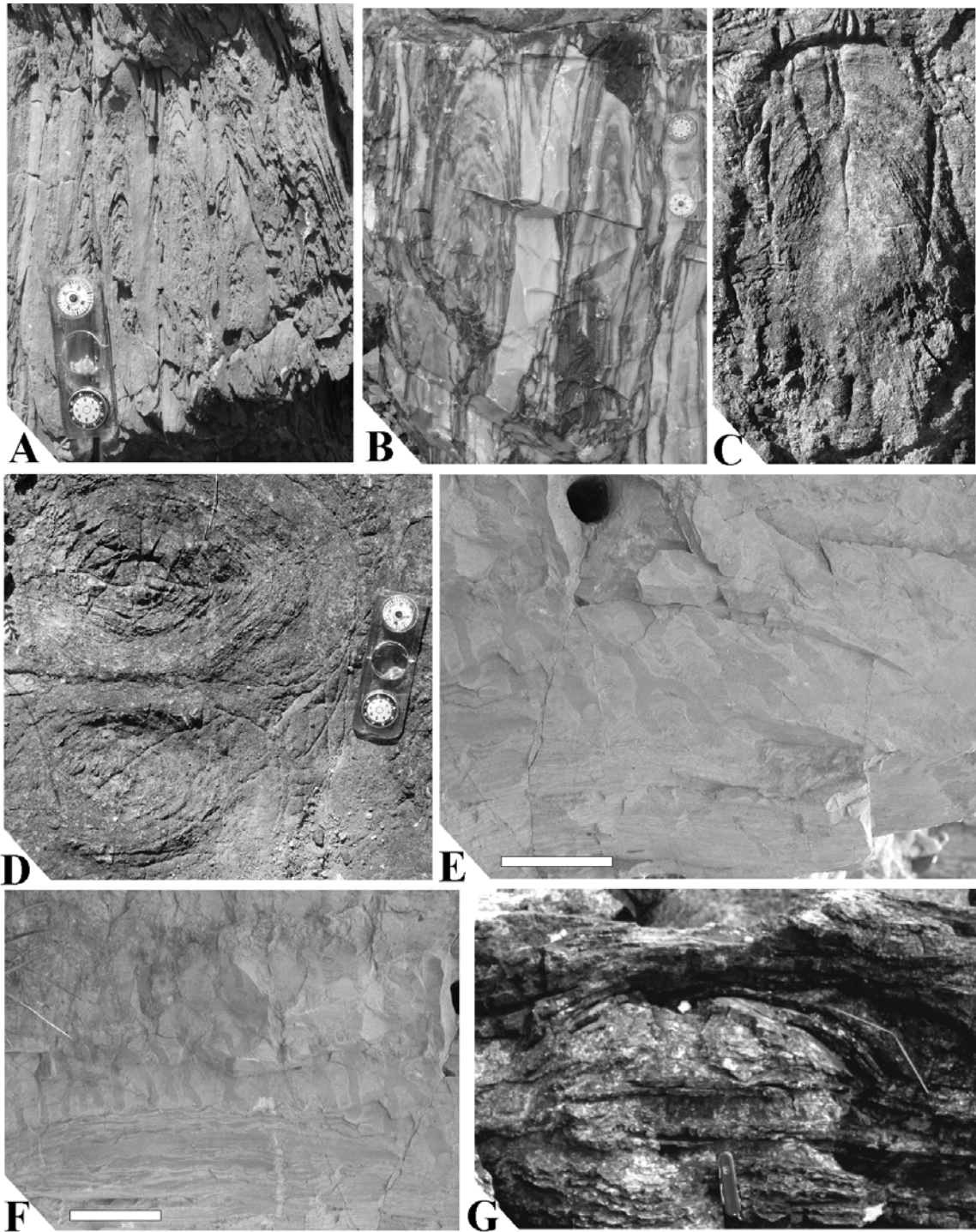


Fig. 4.3.1. Stromatolites of the Mina Verdún (A-B), Sierras Bayas (C-F) and Arroyo del Soldado (G) groups. (A) *Conophyton* from the Mina Verdún Group at its stratotype. (B) Larger, adjacent *Conophyton* columns from the Mina Verdún Group at Burgueño Quarry. (C-D) *Conophyton* from the Villa Mónica Formation at Piedra Amarilla Quarry, in longitudinal (C) and cross section (D). Height of column in C: 21 cm. (E-F) Gymnosolenids developed on top of stratiform stromatolites, Villa Mónica Formation, same locality as C-D. (G) Large, domical stromatolite of the Cerro Victoria Formation at its stratotype, from Gaucher (2000). Scale is 8 cm long in A, B, D, 10 cm in E-F and 9 cm in G.

Kussiella fm., *Minjaria* fm., *Parmites* fm., *Parmites* cf. *cocrescens* and *Stratifera* fm. (Poiré, 1987, 1993).

Acritarchs assigned to the species *Leiosphaeridia minutissima*, *Leiosphaeridia tenuissima* and *Synsphaeridium* sp. were described from shales of the Villa Mónica Formation by Gaucher et al. (2005a; Fig 4.3.2I). The assemblage is characterised by low-diversity, dominance of leiosphaerids and large size of *L. tenuissima*, reaching 450 µm (Fig. 4.3.2I).

Whereas stromatolite biostratigraphy (Poiré, 1987) and Rb-Sr ages of pelites between 793 ± 32 Ma and 805 ± 28 Ma (Bonhomme and Cingolani, 1980) point to a Tonian or early Cryogenian age, the acritarch assemblage described by Gaucher et al. (2005a) favors an early Ediacaran age.

4.3.3. EDIACARAN FOSSIL RECORD

The Ediacaran is well represented in the RPC by a number of sedimentary successions, namely the Las Ventanas Formation, Arroyo del Soldado Group and Sierras Bayas Group-Cerro Negro Formation (see Chapter 4.2). The fossil content of each of these units will be discussed below.

4.3.3.1 Las Ventanas Formation

This unit represents a taphrogenic succession associated to rifting of the RPC from Rodinia (see Chapter 4.6). Sedimentary environments range from continental alluvial fans at the base, to fan deltas in the middle part and open marine shales at the top. Acritarchs were described by Blanco and Gaucher (2004, 2005) and Gaucher et al. (2008a) from the upper Las Ventanas Formation (Quebrada de Viera and El Perdido members). The assemblage is leiosphaerid-dominated and includes the following species (Figs 4.3.2F-H): *Leiosphaeridia tenuissima*, *Leiosphaeridia minutissima*, *Lophosphaeridium* sp., *Soldadophycus bossii*, *Soldadophycus major* and *Soldadophycus* sp. The vendotaenid *Vendotaenia antiqua* also occurs (Gaucher et al., 2008a). The microflora is characterised by relatively low diversity, dominance of *Leiosphaeridia tenuissima*, absence of acanthomorphic acritarchs, and large size of sphaeromorphs, reaching 400 µm in diameter. A number of species are shared with acritarch assemblages preserved in the overlying Arroyo del Soldado Group. Differences between

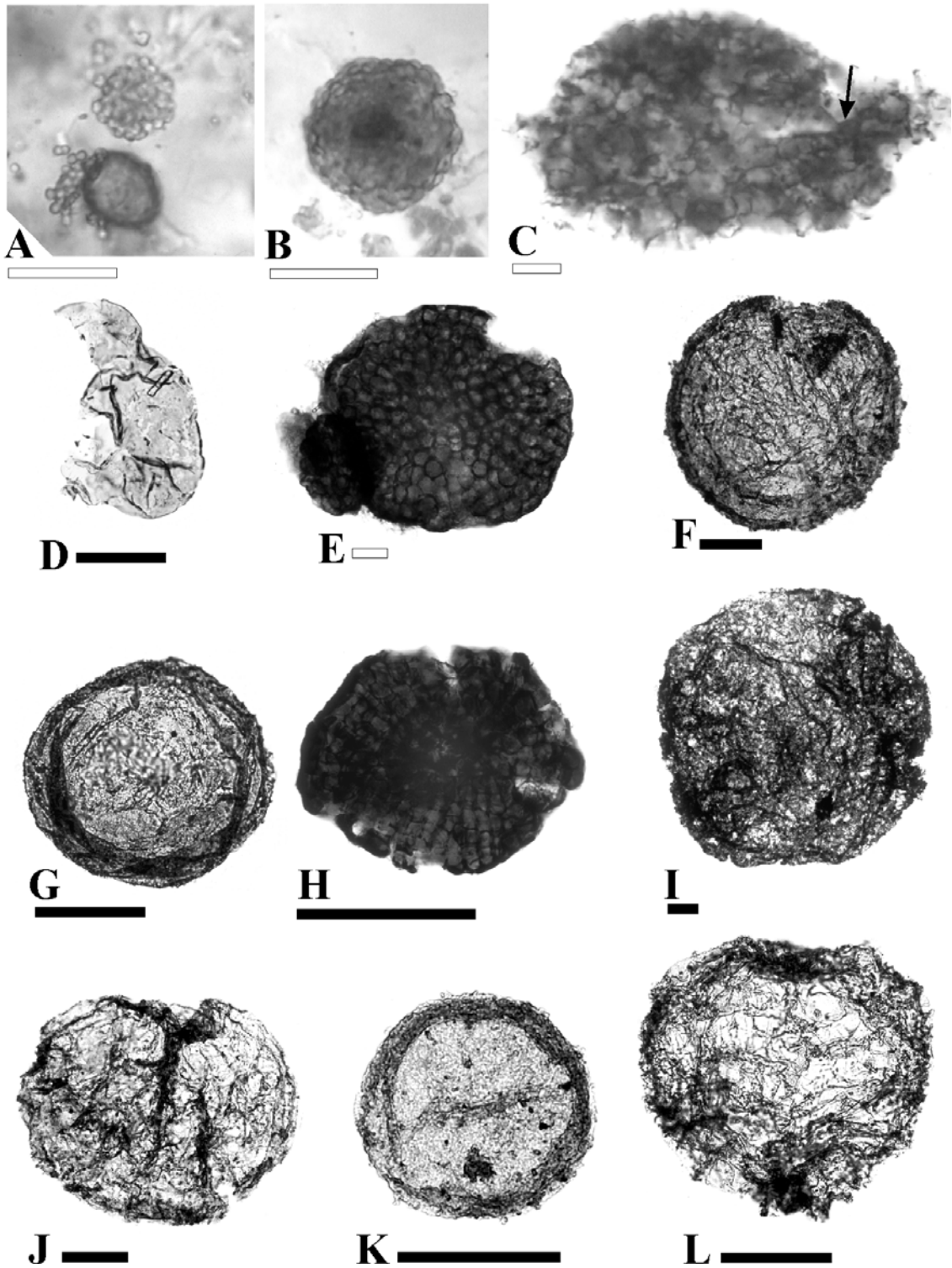


Fig. 4.3.2. Organic-walled microfossils of the Arroyo del Soldado (A-E) and Sierras Bayas (I-K) groups and Las Ventanas (F-H) and Cerro Negro (L) formations. Specimens isolated by means of acid maceration otherwise indicated. (A-B) *Bavlinella faveolata* in thin sections from the uppermost Barriga Negra (A) and Yermal (B) formations. Note in A the sheathed specimen releasing baeocytes (from Gaucher, 2000). (C) *Soldadophycus* cf. *S. major* Gaucher (2000) in a thin section of chert from the uppermost Yermal Formation, La Salvaje Anticline. Note transition from sphaeroidal into filamentous cells (arrowed). (D) *Leiosphaeridia tenuissima*, Polanco Formation (from Gaucher, 2000). (E) *Soldadophycus bossii*, saucer-shaped colony with “daughter” colony, Yermal Formation (from Gaucher et al., 2004c). (F) *Leiosphaeridia tenuissima*, Las Ventanas Formation (from Gaucher et al., 2008a). (G) *Lophosphaeridium* sp. showing prominent concentric folds and small verrucae, Las Ventanas Formation

(from Gaucher et al., 2008a). (H) *Soldadophycus* sp. saucer-shaped colony with radial rows of cells (Las Ventanas Formation; Gaucher et al., 2008a). (I) Large *Leiosphaeridia tenuissima*, Villa Mónica Formation (from Gaucher et al., 2005a). (J-K) *Leiosphaeridia tenuissima*, Cerro Largo Formation (from Gaucher et al., 2005a). (L) *Leiosphaeridia tenuissima*, Cerro Negro Formation (Gaucher et al., 2005a). Note tenuous, strongly plicated vesicle wall (same as in F and J). Hollow scale bars represent 10 μm , solid scale bars 50 μm .

both assemblages include the occurrence of abundant *Bavlinella faveolata* and smaller size of leiosphaerids in the Arroyo del Soldado Group (Fig. 4.3.6). The assemblage occurring in the Las Ventanas Formation is assigned to the early Ediacaran Leiosphere Palynoflora (Grey, 2005; EELP: see Chapter 9.1), which possibly spans the interval between the base of the Ediacaran (end of the Marinoan Glaciation) and the termination of the Gaskiers Glaciation (582 Ma) or the Acraman impact event (Grey, 2005). An early Ediacaran age is also supported by a K-Ar age of 615 ± 30 Ma for basalts (Sánchez Bettucci and Linares, 1996) of the lower Las Ventanas Formation (La Rinconada Member) and an Ar-Ar age of 579 ± 1.5 Ma for the Pan de Azúcar Pluton, which intrudes the Las Ventanas Formation (Oyhantçabal et al., 2007).

4.3.3.2 Lower and middle Arroyo del Soldado Group

The fossil record of this unit is the richest and –thus far– most intensively studied in the whole RPC. Whereas the Yermal, Polanco, Cerro Espuelitas and Barriga Negra formations have been assigned to the Ediacaran, the upper Arroyo del Soldado Group was deposited in the Lower Cambrian (Gaucher, 2000; Gaucher et al., 2004b, 2007b). The Ediacaran fossil groups occurring in the unit are acritarchs and shelly fossils, which are briefly described below.

4.3.3.2.1. Acritarchs

Organic-walled microfossils (acritarchs, filamentous microfossils, colonial sphaeroids) were the first fossils reported from the Arroyo del Soldado Group (Gaucher and Schipilov, 1994; Gaucher et al., 1996, 1998). Two different assemblages have been recognised by Gaucher (2000), namely:

- (a) A *Bavlinella-Soldadophycus* assemblage, strongly dominated by *Bavlinella faveolata*, *Soldadophycus bossii* and/or *Soldadophycus major* occurs in the Yermal and Cerro Espuelitas formations (Figs. 4.3.2A-C, E). It represents a low-

diversity, high-abundance assemblage including sphaeromorphic acritarchs, colonial sphaeroids and filamentous microfossils (Gaucher, 2000; Gaucher et al., 2004c). Sphaeromorphs are characteristically small (<100 µm).

- (b) A *Leiosphaeridia-Lophosphaeridium* assemblage is restricted to the Polanco Formation, showing somewhat higher diversity. Typical species are *Leiosphaeridia minutissima*, *Leiosphaeridia tenuissima* (Fig. 4.3.2D), *Lophosphaeridium* spp., colonial sphaeroids (*Coniunctiophycus*, *Glenobotrydion*, *Myxococcoides*) and large, branched filaments (Gaucher, 2000). In contrast to the previous assemblage, microfossils of the *Leiosphaeridia-Lophosphaeridium* assemblage are sparse.

Gaucher (2000) interpreted the different assemblages as reflecting different palaeoenvironmental conditions. The *Bavlinella-Soldadophycus* assemblage represents an eutrophic community, similar to other *Bavlinella*-dominated assemblages worldwide (Mansuy and Vidal, 1983; Knoll and Swett, 1987; Palacios, 1989; Vidal and Nystuen, 1990; Gaucher et al., 2003, 2005b, Gaucher and Germs, 2006, 2007). On the other hand, the *Leiosphaeridia-Lophosphaeridium* assemblage probably represents a plankton community adapted to lower nutrient levels (oligotrophic).

Biostratigraphically, both communities are comparable to other late Ediacaran acritarch assemblages, characterised by long-ranging species, comparatively small sphaeromorphs and low diversity (Vidal and Moczydlowska, 1997; Knoll, 2000). The fact that four species are shared with the Las Ventanas Formation (*L. tenuissima*, *L. minutissima*, *S. bossi* and *S. major*: Fig. 4.3.6), suggest that there is no large hiatus between both units (Gaucher et al., 2008a). However, important differences were noted by Gaucher et al. (2008a): (a) occurrence of *Bavlinella faveolata* as a dominant species in the Arroyo del Soldado Group, unknown in the Las Ventanas Formation; (b) occurrence of large leiosphaerids up to 400 µm in diameter in the latter unit; and (c) dominance of *Leiosphaeridia* in Las Ventanas Formation, unlike the Arroyo del Soldado Group.

Decreasing maximum leiosphaerid diameter characterise the transition from the Las Ventanas Formation into the overlying Arroyo del Soldado Group: from 400 µm in the Las Ventanas Formation (Fig. 4.3.2F) to 125 µm in the Polanco Formation of the

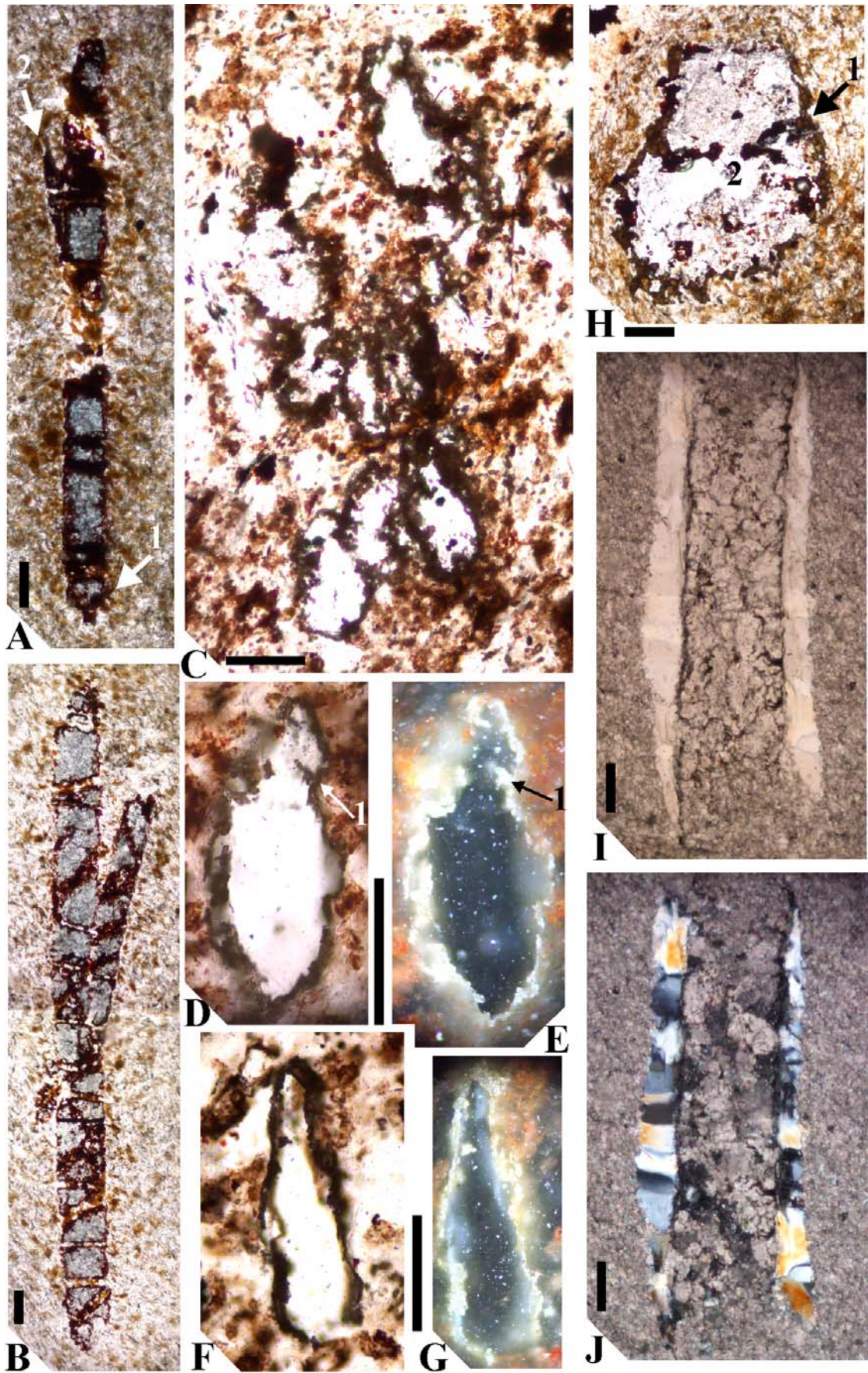


Fig. 4.3.3. *Waltheria*, *Titanotheca* and *Cloudina* in thin sections from the Arroyo del Soldado Group. (A) Holotype of *Waltheria marburgensis* Gaucher and Sprechmann (1999), Yermal Formation. Note septate

shell, spherical chamber (arrowed, 1) and small branch (arrowed, 2). (B) *Waltheria marburgensis* Gaucher and Sprechmann (1999), same unit and locality as previous. (C) *Titanotheca coimbrae* Gaucher and Sprechmann (1999), group of attached, gregarious specimens. Yermal Formation, 5 km to the N of Minas. (D-E) *Titanotheca coimbrae*, solitary specimen with curved neck and septum (arrowed, 1) in transmitted (D) and reflected (E) light; same locality as previous. Note highly reflective rutile grains that compose the shell. (F-G) *Titanotheca coimbrae*, flask-shaped specimen, in transmitted and reflected light, same locality as previous. (H) *Titanotheca coimbrae* figured by Gaucher and Sprechmann (1999), bilocular specimen with a septum (1) and a pore connecting both chambers (2), Yermal Formation. (I-J) Fragment of *Cloudina riemkeae* Germs (1972b) from the Polanco Formation. Replacement of original shell minerals by quartz is clearly visible with cross nicols (J). All scale bars represent 0.1 mm.

Arroyo del Soldado Group (Fig. 4.3.2D). This trend has been reported in other shallow-water, marine Ediacaran successions elsewhere (Butterfield, 1997; Gaucher et al., 2005a). Leiospherids occurring in early Ediacaran successions of Australia (Ediacaran Leiosphere Palynoflora) reach 240 μm in diameter (Grey, 2005).

According to the biostratigraphic schemes of Vidal and Moczydlowska (1997), acritarchs of the Arroyo del Soldado Group match the Kotlin-Rovno assemblage, of late Ediacaran age (Fig. 4.3.6). This is in accordance with the above mentioned assignment of the leiosphaerid assemblage of the Las Ventanas Formation to the older Early Ediacaran Leiosphere Palynoflora (Grey, 2005; see Chapter 9.1).

4.3.3.2.2. *Shelly fossils*

A surprisingly diverse assemblage of skeletal fossils has been described by Gaucher and Sprechmann (1999) and Gaucher (2000) from the lower Arroyo del Soldado Group, which is the most diverse so far reported from the Ediacaran worldwide. It comprises calcified tubes, agglutinated shells, one species with a possibly phosphatic skeleton, and other forms with yet undetermined shell mineralogy.

Cloudina riemkeae Germs (1972b) occurs mainly in the upper Yermal Formation (Gaucher and Sprechmann, 1999; Gaucher, 2000). The fossils, which are hematized and rather poorly preserved (Fig. 4.3.4D), occur in banded siltstones in life position (Gaucher and Sprechmann, 1999). This is shown by: (1) the high angle between the shell axis and bedding plane, (2) lack of size sorting and preferred orientation, and (3) absence of “event beds” with higher shell concentration. The opposite is true for *Cloudina riemkeae* occurring in the overlying Polanco Formation. We report here the occurrence of fragmented *Cloudina* shells in storm deposits of unit B of the Polanco Formation at the Calera de Recalde Syncline (Gaucher et al., 2004b; Figs. 4.3.3I-J,

4.3.4A,C). Although a few specimens preserve their original calcitic shell, most of the specimens are replaced by quartz (Fig. 4.3.3J). Fragments are small (diameter: 0.16-0.54 mm, mean=0.30 mm, N=14), and show a narrow size-range due to sorting by transport processes. Hummocky cross-bedding in the host beds shows that the fossils represent an event-accumulation (tempestite), typical for other *Cloudina* occurrences elsewhere (e.g. Grant, 1990; Gaucher et al., 2003). Typical cone-in-cone structures have been observed in these samples, confirming the assignment to *Cloudina riemkeae* (Figs. 4.3.4A-B). The genus *Cloudina* is a globally-distributed index fossil of the late Ediacaran (Grant, 1990; Knoll, 2000; Grotzinger et al., 2000; Gaucher et al., 2003; Amthor et al., 2003).

Titanotheca coimbrae Gaucher and Sprechmann (1999) is common in the Yermal Formation, especially in its upper half, and co-occurs with *Cloudina*. The fossils are sphaeroidal, vase- or flask-shaped and have one or two chambers separated by a perforated septum (Figs. 4.3.3D-H). The shell is made up of tiny, agglutinated rutile grains (Gaucher and Sprechmann, 1999; Gaucher, 2000). The same type of test composition is known from modern foraminiferans of the genus *Bathysiphon* (Lowenstam and Weiner, 1989). On this basis, and considering the anatomy of *Titanotheca*, Gaucher and Sprechmann (1999) assigned the species to the Foraminiferida (Suborder Textulariina), representing the oldest record of this important protists worldwide. Reproduction was probably by budding (Gaucher and Sprechmann, 1999). Maximum diameter of test varies between 0.1 and 1.5 mm, and length between 0.1 and 5 mm. The gregarious nature of this species is demonstrated by samples of the upper Yermal Formation, where a number of attached specimens have been observed (Fig. 4.3.3C). After the first description of the species from the Arroyo del Soldado Group, other occurrences have been described from a number of Ediacaran successions of Brazil (Gaucher et al., 2003, Teixeira and Gaucher, 2004) and Namibia (Gaucher et al., 2005b). Thus, *Titanotheca* is another potential index fossil of the late Ediacaran, not least because of its unusually high preservation potential.

Waltheria marburgensis Gaucher and Sprechmann (1999; Figs. 4.3.3A-B) occurs exclusively in green and banded siltstones of the upper Yermal Formation, co-occurring with both *Cloudina* and *Titanotheca*. Fossils are hematized, like co-occurring *Cloudina*, but up to 5 wt % P in the shell (corresponding to 25-30 % fluor-apatite) could be

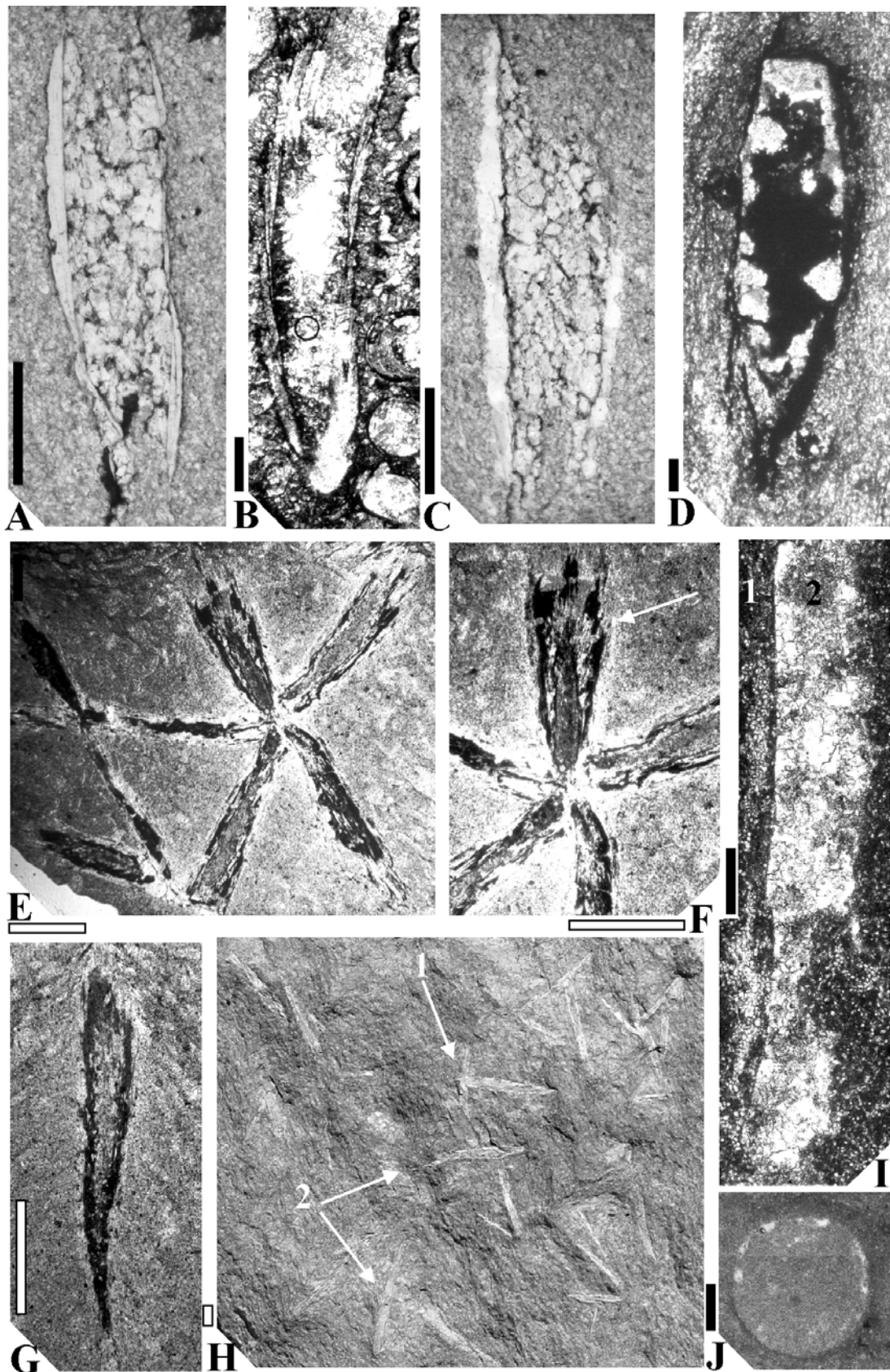


Fig. 4.3.4. *Cloudina* from the Arroyo del Soldado, Nama and Sierras Bayas groups and tubular-conical fossils of the Yermal Formation (thin sections). (A, C) *Cloudina riemkeae* Germs (1972b) in limestones of the Polanco Formation (Unit B) with silicified shells and sparitic infill. (B) *Cloudina riemkeae* Germs (1972b) from the type area in the Nama Group. Note cone in cone structure and same anatomy as specimen figured in A. (D) *Cloudina riemkeae* in siltstones of the upper Yermal Formation, showing

hematised shell and infill composed of quartz and hematite (Gaucher et al., 2003). (E-F) Star-shaped aggregate of conical fossils in a bedding-parallel section of siltstone from the middle Yerbal Formation, Quebrada de los Cuervos Member (Gaucher, 2002). Note deeply nested cones (arrowed). (G) Solitary and sinuous cone from the same unit. Note sediment infill. (H) Conical-tubular fossils as observed on bedding planes of siltstones from the Quebrada de los Cuervos Member. Note (1) branching at 90° and (2) solitary, straight or sinuous tubes. (I) *Cloudina riemkeae* Germs (1972b) in limestones of the Loma Negra Formation (longitudinal section). Note (1) recrystallised outer tube, and (2) inner cavity filled with sparitic calcite (from Gaucher et al., 2005a). (J) *Cloudina riemkeae* from the same locality as previous, cross section (from Gaucher et al., 2005a). Solid scale bars represent 0.25 mm, hollow scale bars 2.5 mm.

detected by means of EDS-microprobe analyses (Gaucher and Sprechmann, 1999). This could represent a relict of the original shell composition, because phosphate is almost absent (<0.12 wt%) from the host sediment. The shell is cylindrical, septate, commonly shows branching, and has a multi-layered, complex structure (Figs. 4.3.3A-B; Gaucher and Sprechmann, 1999). Transitions between single tubes, tube bundles and discs have been reported (Gaucher and Sprechmann, 1999; Gaucher, 2000). Diameter varies between 10 µm and 1 mm, and length between 0.8 and 10 mm. Fossils usually occur in large numbers, both in life position (perpendicular to bedding) and reworked by currents. The palaeobiological affinity of *Waltheria* remains speculative.

Two other species of skeletal fossils from the Yerbal Formation were formally described by Gaucher and Sprechmann (1999) and Gaucher (2000), which are less abundant and show a restricted stratigraphic range, namely *Soldadotubulus siderophoba* and *Palaeodiscus mendezalzolai*. Both species occur as steinkerns (internal moulds), thus the original mineralogy and structure of the shell is unknown.

The oldest skeletal fossils occurring in the Arroyo del Soldado Group, however, were first illustrated by Gaucher (2002) and have not been formally described yet. These fossils, collectively known as the “Quebrada de los Cuervos fauna”, occur in bluish-gray siltstones interbedded with graded, fine sandstone strata of the middle-lower Yerbal Formation. Fossils are conical (Fig. 4.3.4G), either solitary or arranged in “X” or star-shaped clusters (Figs. 4.3.4E-F), straight to sinuous, and occur in large numbers on bedding planes (Fig. 4.3.4H). Cone-in-cone structure is typical for the species, but differs from the structure of *Cloudina* shells in that the cones share the same apex. Diameter ranges between 0.4 and 3 mm, and length reaches 30 mm. Shell walls are mostly replaced by chamosite, as indicated by EDS microprobe analyses and petrography. The inner cavities are filled with sediment and/or chamosite. A common

feature is the growth of sericite crystals perpendicular to the shell in the inner cavities. The described preservation is almost identical to that of fossils of the Burgess Shale Lagerstätte (Petrovich, 2001 and references therein), currently explained by replacive precipitation of chamosite under suboxic diagenetic conditions and high Fe/C ratios (Petrovich, 2001). The original mineralogy of the shell is thus unknown, but the existence of a mineralised exoskeleton is indicated by strong orientation by currents, occurrence of a few specimens perpendicular to bedding (life position?) and shell fragments.

4.3.3.3. *Sierras Bayas Group and Cerro Negro Formation*

The fossil record of the Sierras Bayas Group and Cerro Negro Formation includes acritarchs, shelly fossils, trace fossils and discoidal fossils of unknown affinity. Fossils of the lowermost Villa Mónica Formation (stromatolites, acritarchs), of still uncertain age, were already discussed above.

4.3.3.3.1. *Acritarchs*

Pothe de Baldis et al. (1983) reported *Chuarina olavarriensis*, *Leiosphaeridia* sp. and *Paleorivularia ontarica* from the upper Cerro Largo Formation (now Olavarría Formation, see Chapter 4.2), which were reassigned by Gaucher et al. (2005a) as *Chuarina circularis* Walcott and *Leiosphaeridia minutissima* (Naumova) Jankauskas et al. (1989; Figs. 4.3.2J-K). In the case of *Paleorivularia*, Gaucher et al. (2005a) considered them as dubiofossils (pyrite framboids?). Pothe de Baldis et al. (1983) also described colonial acritarchs and assigned them to *Leiosphaeridia* sp., but they are better classified as *Synsphaeridium* Eisenack (Gaucher et al., 2005a).

A detailed palynological study of the Cerro Negro Formation is due to Cingolani et al. (1991), who described *Leiosphaeridia* sp., *Trachysphaeridium* sp., *Synsphaeridium* sp. and “type A“ granular sphaeromorphs from the mentioned unit. Gaucher et al. (2005a) apply the taxonomic schemes of Jankauskas et al. (1989) and Butterfield et al. (1994) to the sphaeromorphs, reassigning the material to *Leiosphaeridia jacutica* Timofeev, *Lophosphaeridium* and *Synsphaeridium*. As for *Trachysphaeridium* sp., fossils assigned by Cingolani et al. (1991) to this taxon range between 60 and 80 µm in diameter, but do not show the spongy or alveolate surface texture typical of the genus (Vidal, 1976).

Thus, Gaucher et al. (2005a) conclude that they may represent strongly corroded *Leiosphaeridia minutissima* and *L. tenuissima* (Fig. 4.3.2L).

Palynological macerations of shales of the Cerro Largo and Olavarría formations carried out by Gaucher et al. (2005a) yielded *Leiosphaeridia tenuissima* as dominant species (Figs. 4.3.2J-K), and a single vesicle assigned to *Leiosphaeridia minutissima*, which add to the species described by Pothe de Baldis et al. (1983) for the same units. Gaucher et al. (2005a) also report acritarchs assigned to *Leiosphaeridia tenuissima* from shales of the Cerro Negro Formation (Fig. 4.3.2L).

Considering all the available data, it is clear that acritarchs of the Sierras Bayas Group and Cerro Negro Formation are all leiosphere-dominated, low-diversity assemblages. Size of the vesicles, though, varies considerably in the different units. A tendency towards smaller sphaeromorphs from base to top is observed: from 450 µm for the Villa Mónica Formation to 380 µm for the Cerro Largo Formation and finally 200 µm for the Cerro Negro Formation (Gaucher et al., 2005a). This trend has been reported for other shallow-water, marine Ediacaran successions elsewhere (Butterfield, 1997; Gaucher et al., 2008a). On the basis of low diversity and dominance of sphaeromorphs, the acritarch assemblage of the Sierras Bayas Group-Cerro Negro Formation can be assigned either to the Ediacaran Leiosphere Palynoflora of Grey (2005) or the younger Kotlin-Rovno assemblage of Vidal & Moczydlowska (1997). Other biostratigraphic and chemostratigraphic data, however, confidently place the Cerro Largo, Olavarría and Loma Negra formations in the late Ediacaran (see below), allowing to assign the corresponding acritarch assemblages to the Kotlin-Rovno assemblage, or Late Ediacaran Leiosphere Palynoflora (LELP: see Chapter 9.1). As already discussed above, available data are ambiguous regarding the age assignment and correlation of the Villa Mónica Formation. The apparent similarity of leiosphere-dominated assemblages of the Villa Mónica Formation and the rest of the Sierras Bayas Group raises the possibility of the Villa Mónica Formation being also Ediacaran in age. No Cambrian acritarchs were found in the Cerro Negro Formation, which is here assigned to the terminal Ediacaran.

4.3.3.3.2. *Shelly fossils*

Gaucher et al. (2005a) report for the first time the occurrence of *Cloudina* cf. *C. riemkeae* Germs (1972b) in micritic limestones of the Loma Negra Formation (Figs. 4.3.4I-J). These fossils show circular cross sections (Fig. 4.3.4J), poorly-preserved

calcite walls (Fig. 4.3.4I) and diameters ranging between 0.15 and 0.80 mm (mean=0.42 mm, N=12). Maximum length of the tubes reaches 2.5 mm. According to Gaucher et al. (2005a), poor preservation of *Cloudina* is due to a certain amount of recrystallisation and partial dissolution of the shells. As already mentioned, the genus *Cloudina* is a globally-distributed index fossil of the late Ediacaran (Grant, 1990; Knoll, 2000; Grotzinger et al., 2000; Gaucher et al., 2003; Amthor et al., 2003).

4.3.3.3.3. Trace fossils

Trace fossils are scarce and show low ichnodiversity. *Palaeophycus* isp. and *Didymaulichnus* isp. have been described from the Cerro Largo Formation (Poiré et al., 1984). Porada & Bouougri (2008) re-interpreted the trace fossils of Cerro Largo Formation as biomat desiccation cracks and other biomat structures (Fig. 4.3.5D). *Helminthopsis* isp. and probable medusa resting traces have been found in the Loma Negra Formation. *Skolithos* isp. was reported from the lower part of the Cerro Negro Formation (Poiré et al., 1984), which may indicate a Cambrian age for part of this unit but contradicts other biostratigraphic data (see above). However, *Skolithos* isp. has been reported from the late Ediacaran Kuibis Subgroup of the Nama Group (Germs, 1972c). The early Palaeozoic Balcarce Formation bears abundant and diverse trace fossils (Borrello, 1966; Aceñolaza, 1978; Seilacher et al., 2002; Poiré et al., 2003a), most of which can be attributed to the activities of trilobites.

4.3.4 CAMBRIAN FOSSIL RECORD

Cambrian fossils are scarce in the eastern RPC, and mostly restricted to the sub-trilobitic Lower Cambrian, the reason being that tangential collision with the Cuchilla Dionisio-Pelotas Terrane and basin closure took place around 530 Ma (Bossi and Gaucher, 2004).

4.3.4.1. Upper Arroyo del Soldado Group

The Cerro Victoria Formation is characterised by shallow-water dolostones of Lower Cambrian age (Sprechmann et al., 2004; Gaucher et al. 2007b). Stromatolites and large trace fossils occur in this unit.

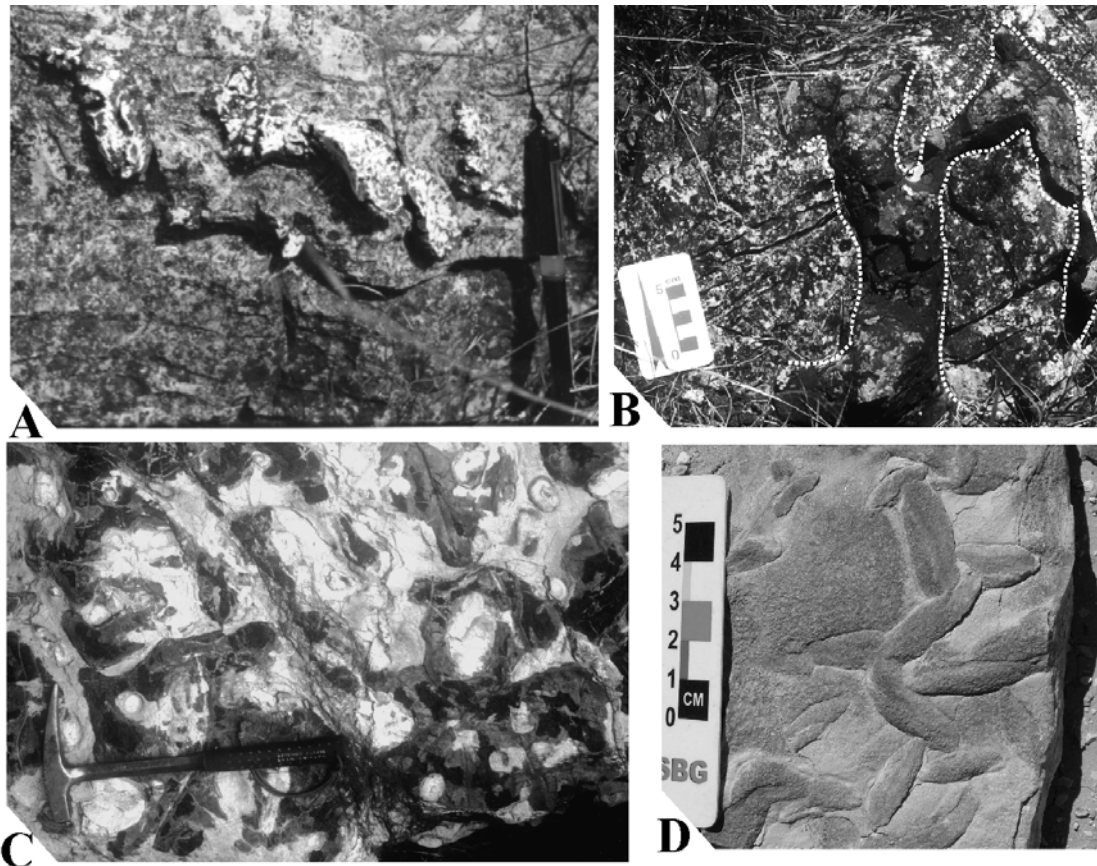


Fig. 4.3.5. Trace fossils and biomat structures. (A) Helicoidal burrow (*Gyrolithes*), Cerro Victoria Formation at its stratotype. (B) *Thalassinoides*, Cerro Victoria Formation at its stratotype. (C) Large *Thalassinoides* burrows (plan view, dark material) from the Late Cambrian upper La Flecha Formation (Argentina, coordinates: 29°29'59''S, 68°40'05''W), showing secondary diagenetic overgrowth and silicification (see Sial et al., 2008). (D) Biomat desiccation structures (*sensu* Porada and Bouougri, 2008), Cerro Largo Formation.

4.3.4.1.1. *Stromatolites*

Montaña and Sprechmann (1993) reported the occurrence of stromatolites in the Cerro Victoria Formation, uppermost Arroyo del Soldado Group (Gaucher, 2000). Sprechmann et al. (2004) described in detail the low-diversity stromatolite community of the Cerro Victoria Formation, consisting of non-branching morphogroups. The following morphotypes were recognised (Sprechmann et al., 2004; Gaucher et al., 2007b): (1) planar stromatolites (cryptalgal laminites); (2) domal stromatolites assigned by Sprechmann et al. (2004) to the LLH-C, LLH-S groups of Logan et al. (1964), up to 1.6 m in diameter (Fig. 4.3.1G); (3) short columns of the SH-V and SH-C groups of Logan et al. (1964); (4) microdigitate stromatolites, and (5) rare oncoids. Stromatolites form sheet-like bioherms up to 4 m in height and elongated biostromes of low relief (Gaucher et al., 2007b). A number of different coated grains (oooids, pisoids) occur in

the Cerro Victoria Formation (Montaña and Sprechmann, 1993; Sprechmann et al., 2004; Gaucher et al., 2007b).

The low-diversity stromatolite community of the Cerro Victoria Formation, consisting of non-branching morphogroups is consistent with a Lower Cambrian age (Walter, 1994). Thickness of carbonate deposits (400 m), occurrence of abundant ooids and associated evaporites suggest the return of tropical conditions to the RPC in the Lower Cambrian.

4.3.4.1.2. Trace fossils

No trace fossils occur in the lower and middle Arroyo del Soldado Group. Montaña and Sprechmann (1993) described large burrow systems from the Cerro Victoria Formation. Sprechmann et al. (2004) assign these ichnofossils to the ichnogenera *Thalassinoides*, *Gyrolithes* (Fig. 4.3.5A) and *Palaeophycus*. *Thalassinoides*, by far the most common (Fig. 4.3.5B), comprises cylindrical burrows, forming 3-dimensional branching systems consisting of bedding-parallel networks or boxworks presenting dichotomic branching and connected to the surface by subvertical to vertical shafts (Sprechmann et al., 2004). All trace fossils show siliceous diagenetic overgrowths (flint).

The trace fossil assemblage of the Cerro Victoria Formation is biostratigraphically relevant, strongly suggesting a Lower Cambrian or younger age (Brasier et al., 1994), in accordance with chemostratigraphic data (Gaucher et al., 2007b). *Thalassinoides* showing virtually the same preservation and size as in the Cerro Victoria Formation occurs in the upper La Flecha and lower La Silla formations of the Precordillera Terrane (Argentina) at Cerro la Silla and Quebrada la Angostura (Fig. 4.3.5C), which has been assigned to the Upper Cambrian (Peralta, 2000; Sial et al., 2008). The palaeoecological analogy between both occurrences is further indicated by stromatolitic and thrombolitic carbonates in the La Flecha and La Silla formations, similar to that described above for the Cerro Victoria Formation (Sial et al., 2008).

4.3.5. BIOSTRATIGRAPHIC CORRELATIONS AND DISCUSSION

The poor to nonexistent Tonian and Cryogenian sedimentary record in the RPC restricts biostratigraphic inferences to the Ediacaran and basal Cambrian. However, stromatolitic assemblages of the Mina Verdún Group and Villa Mónica Formation are more easily explained if these successions are older than the Ediacaran, and for the former unit,

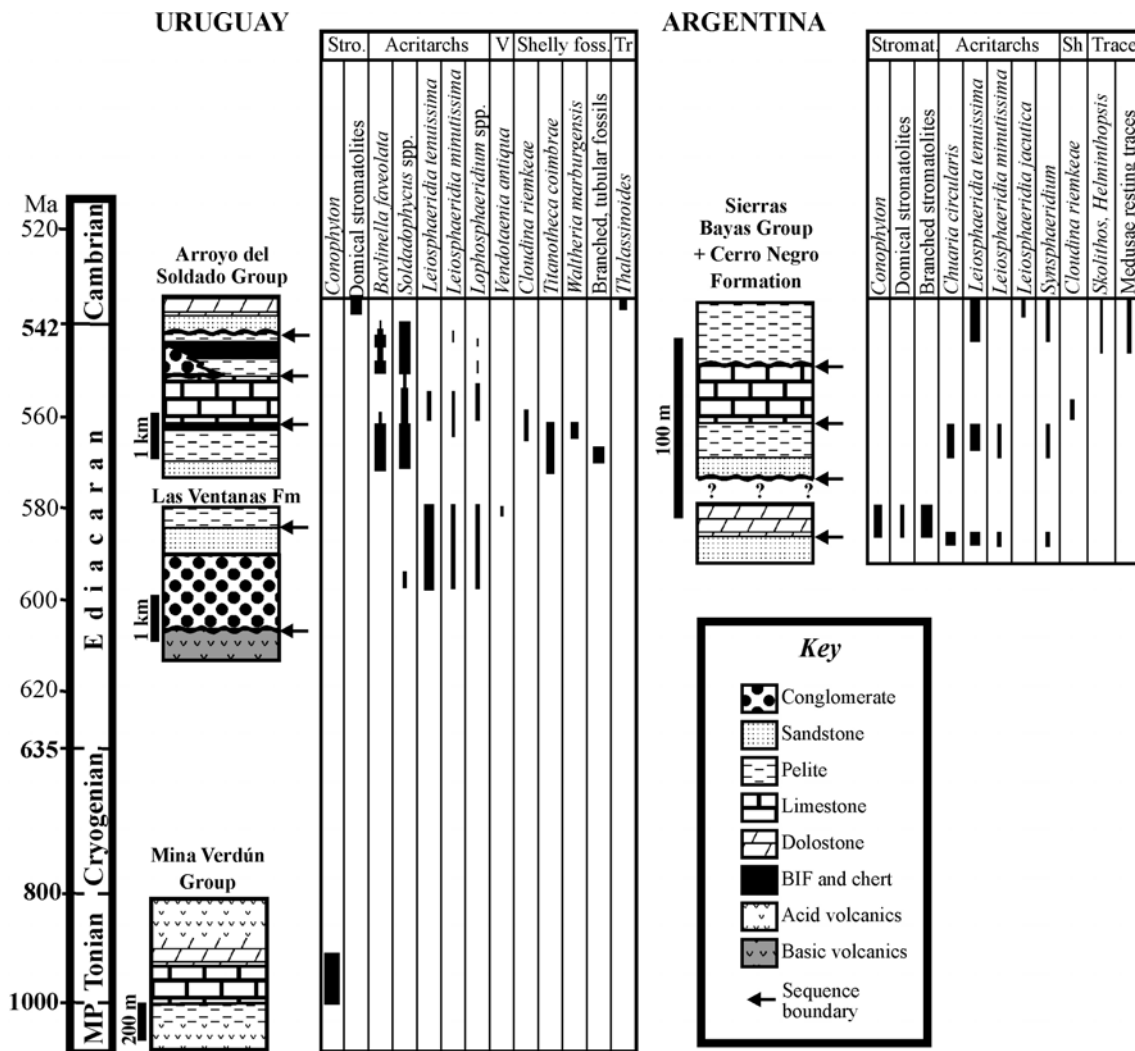


Fig. 4.3.6. Stratigraphic ranges of biostratigraphically most significant taxa in Neoproterozoic-Lower Cambrian successions in the RPC. Data from Gaucher and Sprechmann (1999), Gaucher (2000), Gaucher et al. (2005a, 2008a), Poiré (1987), Poiré et al. (2005) and this work.

maybe even Mesoproterozoic in age (Fig. 4.3.6). Stromatolite biostratigraphy is not without problems, though, because stromatolite morphology is strongly dependent on palaeoenvironmental conditions (Logan et al., 1964).

Ediacaran acritarchs in the RPC are without exception low-diversity, sphaeromorph-dominated assemblages. Diverse acanthomorphic acritarchs reported from the Ediacaran of Australia, China and Siberia (e.g. Grey, 2005; Zhou et al., 2007) do not occur in the RPC. According to Zhou et al. (2007) the complex acanthomorphs were extant between 635 and 580 Ma, suggesting that their absence from the RPC is due to the fact that most sedimentary successions there were deposited after 580 Ma. According to Gaucher et al. (2008b), the Mangacha Granite, overlain by sandstones of the Arroyo del Soldado Group with erosional unconformity, yielded an U-Pb SIMS zircon age of 583 ± 7 Ma.

Blanco et al. (in press) report an age of 566 ± 8 Ma for the youngest detrital zircon of the Barriga Negra Formation, also showing that the base of the Arroyo del Soldado Group is probably no older than 570 Ma. On the other hand, it is difficult to explain why the acritarch assemblage of the lower Ediacaran Las Ventanas Formation is also leiosphere-dominated (Fig. 4.3.6; see Chapter 9.1). The short-lived acanthomorph assemblage (<15 Myr: Grey et al., 2003) may have been either overlooked or not preserved in the RPC.

The litho- and chemostratigraphic correlation between the Sierras Bayas and Arroyo del Soldado groups (see above) is biostratigraphically supported by (Gaucher et al., 2005a): (1) *Cloudina* occurring in the Loma Negra, lower Polanco and upper Yerbál formations (Figs. 4.3.3-4); and (2) similar, sphaeromorph-dominated acritarch assemblages. The differences between the *Bavlinella*-dominated assemblages of the Arroyo del Soldado Group and the *Leiosphaeridia*-dominated assemblages of the middle-upper Sierras Bayas Group are probably due to the latter unit being more proximal, as suggested by Butterfield (1997) for other Neoproterozoic successions elsewhere.

Ediacaran shelly fossils of the Arroyo del Soldado and Sierras Bayas groups (Figs. 4.3.3-4) show considerable diversity and potential as index fossils for subdividing the late Ediacaran. Ongoing research aimed at constraining their precise ranges will contribute toward this goal.

Soft-bodied fossils of the Ediacara Fauna are unknown in the RPC and also in the rest of South America, with the possible exception of the Corumbá Group (see Chapter 2). Discoidal fossils 2 to 5 cm in diameter have been recently found in the Cerro Negro Formation (Cingolani and Gaucher, in preparation), but their taxonomic assignment is still under study.

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CHEMOSTRATIGRAPHY*

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4.4.1. INTRODUCTION

Isotopic chemostratigraphy was first applied to Neoproterozoic successions in the RPC by Boggiani (1998) and Kawashita et al. (1999a,b), and a compilation of the data was published by Gaucher (2000). A relatively large database is now available for the Mina Verdún, Arroyo del Soldado and Sierras Bayas groups (Gaucher et al., 2003; 2004b,d; 2006, 2007a,b; Gómez Peral et al., 2007), as well as preliminary data for the upper Piedras de Afilar Formation (Pamoukaghlián et al., 2006).

4.4.2. CARBON ISOTOPE CHEMOSTRATIGRAPHY

4.4.2.1. Mina Verdún Group

A total of eighty-eight carbon- and oxygen-isotope analyses on carbonates of the Mina Verdún Group were presented by Gaucher et al. (2006, 2007a). Three sections were studied, namely: (1) the stratotype of the group at Mina Verdún, (2) Burgueño Quarry, and (3) Paso del Molino (Fig. 4.2.2). $\delta^{13}\text{C}$ vs. $\delta^{18}\text{O}$ crossplots display strong covariation of both parameters and more negative values for the latter two sections, showing that C isotope composition has been affected by post-depositional processes (Fig. 4.4.1). However, corresponding Mn/Sr values are 0.3 for limestones at the Paso

* Gaucher, C., Sial, A.N., Poiré, D., Gómez-Peral, L., Ferreira, V.P., Pimentel, M.M., 2009. Chemostratigraphy. Neoproterozoic-Cambrian evolution of the Río de la Plata Palaeocontinent. In: Gaucher, C., Sial, A.N., Halverson, G.P., Frimmel, H.E. (Eds.): Neoproterozoic-Cambrian tectonics, global change and evolution: a focus on southwestern Gondwana. *Developments in Precambrian Geology*, Vol. 16, Elsevier, pp.

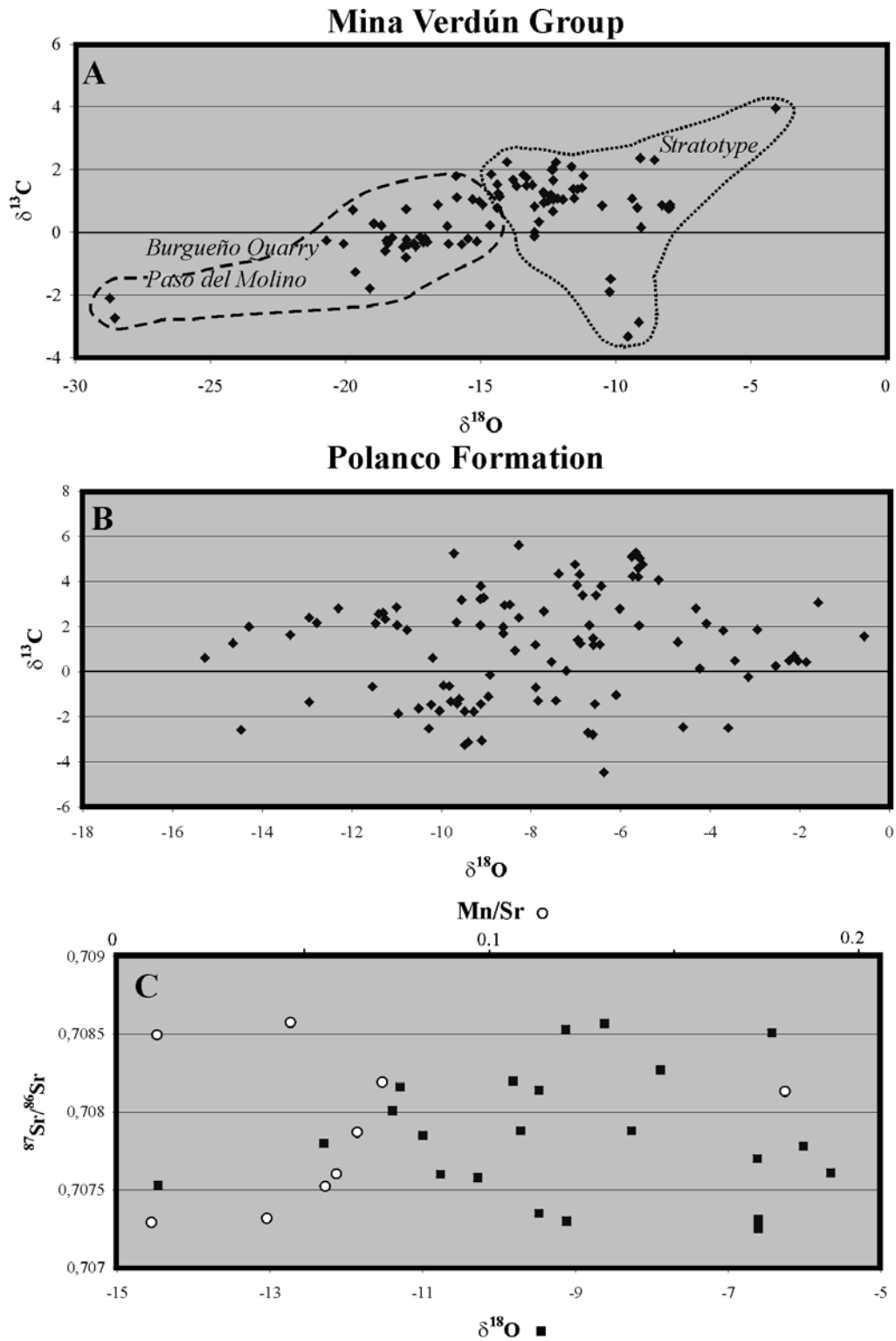


Fig. 4.4.1. (A) $\delta^{13}\text{C}$ vs. $\delta^{18}\text{O}$ crossplot for 88 samples of the Mina Verdún Group. Note altered nature (linear array) of samples from the Burgueño Quarry and Paso del Molino sections, contrasting with the primary isotopic signature at the stratotype. (B) $\delta^{13}\text{C}$ vs. $\delta^{18}\text{O}$ crossplot of more than 100 samples from the Polanco Formation throughout the basin. Note lack of correlation between both parameters. (C) $^{87}\text{Sr}/^{86}\text{Sr}$ vs. $\delta^{18}\text{O}$ (black squares) and $^{87}\text{Sr}/^{86}\text{Sr}$ vs. Mn/Sr (white circles) for limestones of the Polanco Formation (22 samples). Lack of co-variance strongly suggest that $^{87}\text{Sr}/^{86}\text{Sr}$ values reflect seawater composition.

del Molino section, suggesting that the isotopic ratios are near-primary (Jacobsen and Kaufman, 1999). The strong decoloration of limestones at Burgueño Quarry and Paso del Molino, compared to the same levels at the stratotype, shows that organic carbon was oxidated, leading to production of CO₂ enriched in ¹²C. This was later incorporated into the carbonates during early diagenesis, resulting in lower δ¹³C ratios (Murata et al. 1969; De Giovanni et al., 1974). These reactions do not necessarily involve Sr or Mn remobilisation, thus explaining the low Mn/Sr ratios.

At the stratotype in Mina Verdún, no co-variation of δ¹³C vs. δ¹⁸O is observed (Fig. 4.4.1), and a very consistent trend from negative to positive values is recorded (Gaucher et al., 2006). δ¹³C is higher by 3 ‰ PDB than at corresponding stratigraphic levels in the other sections, and organic matter is well preserved. Therefore, we conclude that carbon isotopic ratios at Mina Verdún reflect coeval seawater composition. The δ¹³C curve thus obtained (Fig. 4.4.2) shows a negative excursion at the base (La Toma Formation) to -3.3 ‰ PDB, and steadily rising to values around +1.5 ‰ PDB up section. For 240 m of section (El Calabozo and Gibraltar formations) only positive δ¹³C values occur, varying between +0.15 and +4.0 ‰ PDB, but mostly around a plateau centered at +2 ‰ PDB (Gaucher et al., 2006). This plateau is the most prominent chemostratigraphic feature of the Mina Verdún Group (Fig. 4.4.2).

The δ¹³C curve obtained has the following salient features: (a) moderate amplitude of δ¹³C excursions of up to 5.5 ‰ (transition between La Toma and El Calabozo formations); (b) δ¹³C values ranging between -3.3 and +2.4 ‰ PDB, with only one analysis yielding +4.0 ‰ PDB, and (c) occurrence of a plateau around +2 ‰ PDB for most of the section. These features strongly resemble the global C-isotopic curve the late Mesoproterozoic to early Neoproterozoic (Tonian), between 1300-800 Ma (Kah et al., 1999; Bartley et al., 2007). Early Mesoproterozoic (>1300 Ma) carbonates show isotopic invariance around 0 ‰ PDB, and late Neoproterozoic (Cryogenian-Ediacaran) successions are characterised by rapid, large-amplitude isotopic excursions and heavy δ¹³C values between +4 and +12 ‰ PDB (Kah et al., 1999; Halverson et al., 2005; Bartley et al., 2007). A similar age around the Meso-Neoproterozoic boundary is inferred by Santos et al. (2000) for the Paranoá Group of central Brazil, which shows similar δ¹³C values and a stromatolite assemblage also dominated by *Conophyton*.

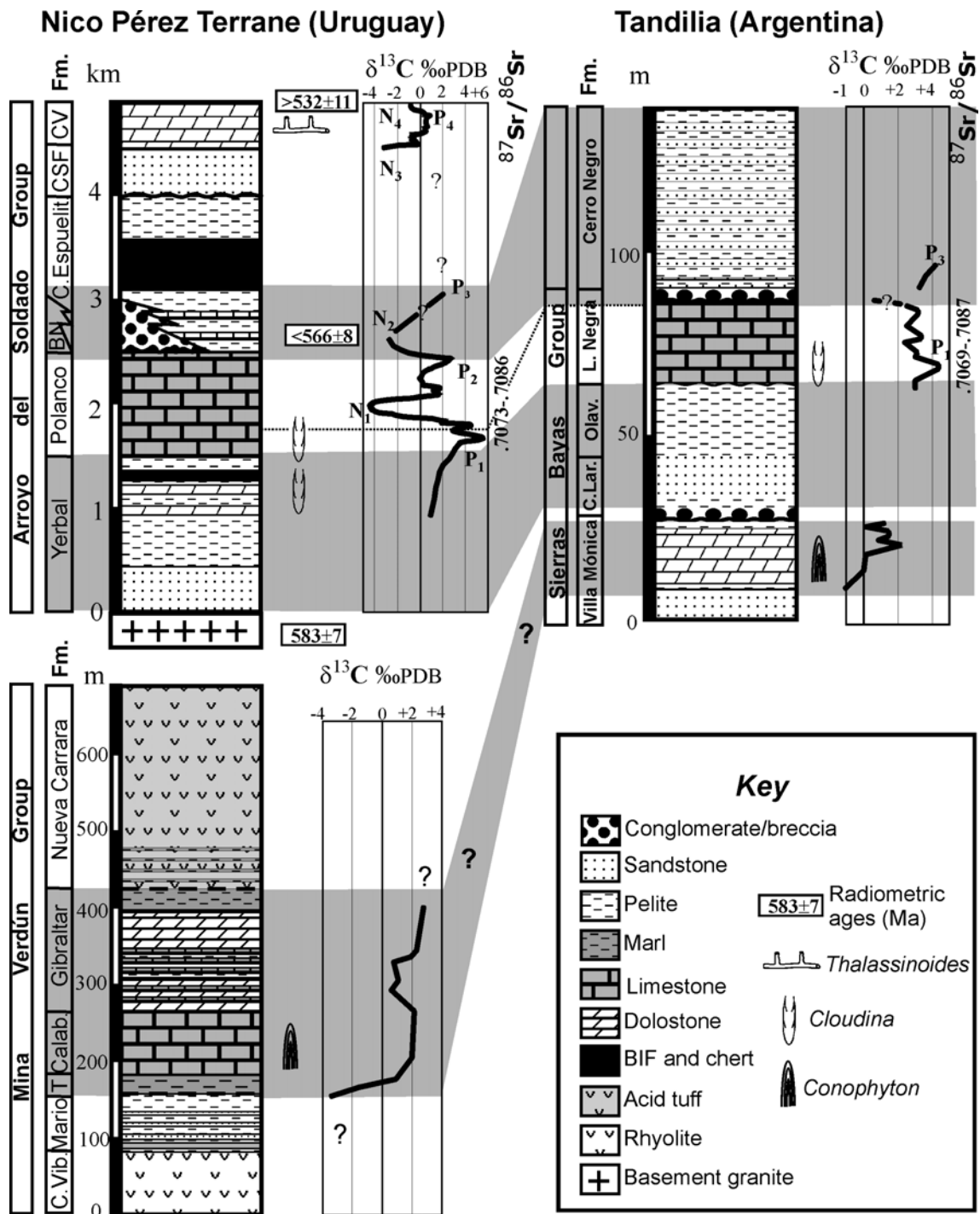


Fig. 4.4.2. Available chemostratigraphic data for the Arroyo del Soldado, Mina Verdún and Sierras Bayas groups and inferred correlations. Arroyo del Soldado Group: BN: Barriga Negra Formation; C.Espuelit.: Cerro Espuelitas Formation; CSF: Cerros San Francisco Formation; CV: Cerro Victoria Formation. Radiometric ages shown are 583 ± 7 Ma for the Mangacha Granite (Gaucher et al., 2008b); 566 ± 8 Ma for the youngest detrital zircon in the Barriga Negra Formation (Blanco et al., in press) and 532 ± 11 Ma for the Guazuñambí Granite, intrusive in the ASG (Kawashita et al., 1999a). Sierras Bayas Group: C.Lar: Cerro Largo Formation, Olav.: Olavarría Formation, L. Negra: Loma Negra Formation. Chemostratigraphic data from Gómez Peral et al. (2007) and Kawashita et al. (1999b). Mina Verdún Group (Gaucher et al., 2007a): C.Vib.: Cerro de las Víboras Formation; Mario: Don Mario Formation; T: La Toma Formation; Calab.: El Calabozo Formation. Modified from Gaucher et al. (2005a).

4.4.2.2. Arroyo del Soldado Group

The thick and pure carbonate deposits of the Polanco and Cerro Victoria formations of the Arroyo del Soldado Group (ASG) provide an ideal record of the isotopic fluctuations of coeval seawater. Thus far, the database comprises 175 $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ analyses on carbonates, and a dozen sections have been studied (Boggiani, 1998; Kawashita et al., 1999a; Gaucher, 2000; Gaucher et al., 2003, 2004b, 2004d, 2007b). Preliminary $\delta^{13}\text{C}$ analyses on organic matter from the ASG were presented by Velásquez et al. (2007, 2008).

Whereas moderate to large-amplitude $\delta^{13}\text{C}$ oscillations (up to 10 ‰ PDB) occur in the lower and middle ASG, low-amplitude variations (2 ‰ PDB) characterise the upper ASG (Gaucher et al., 2007b; Fig. 4.4.2). Four positive and negative $\delta^{13}\text{C}$ excursions occur in the ASG, which are described below (Figs. 4.4.2-4.4.3). Basinal sections of the ASG (Polanco Formation) record more negative values than proximal sections (Gaucher et al., 2004b), suggesting a strong C-isotopic gradient in the Ediacaran consistent with a stratified ocean (Calver, 2000; Frimmel, 2004; Shen et al., 2005).

4.4.2.2.1. Upper Yerbal-Lower Polanco P_1 positive excursion

This is a long-lived feature, encompassing more than 1.000 m of section (Gaucher et al., 2004b; Figs. 4.4.2-3). A steady rise from +1 to +5.6 ‰ PDB is observed, which is concomitant to sea-level rise, shift from siliciclastic to carbonate deposits and climate warming, as also indicated by changes in clay-mineral assemblages (Gaucher, 2000; Pamoukaghlian et al., 2004). The transition to negative values up section is rather abrupt and occurs in less than 70 m (Fig. 4.4.3), in which $\delta^{13}\text{C}$ values drop from +5 to -4.5 ‰ PDB.

4.4.2.2.2. Middle Polanco N_1 negative excursion

This prominent feature comprises 200 m of carbonates of Unit B of the Polanco Formation (Gaucher et al., 2004b; Fig. 4.4.3). The most negative values of this excursion were given as -3.3 ‰ PDB by Gaucher et al. (2004b; Fig. 4.4.3), but recently acquired data show even more depleted carbonates with $\delta^{13}\text{C}$ of -4.5 ‰ PDB. Thus, both because of the magnitude of the $\delta^{13}\text{C}$ drop and its significant stratigraphic expression, this negative excursion is the most important in the ASG. Sea-level drop,

Calera de Recalde Syncline

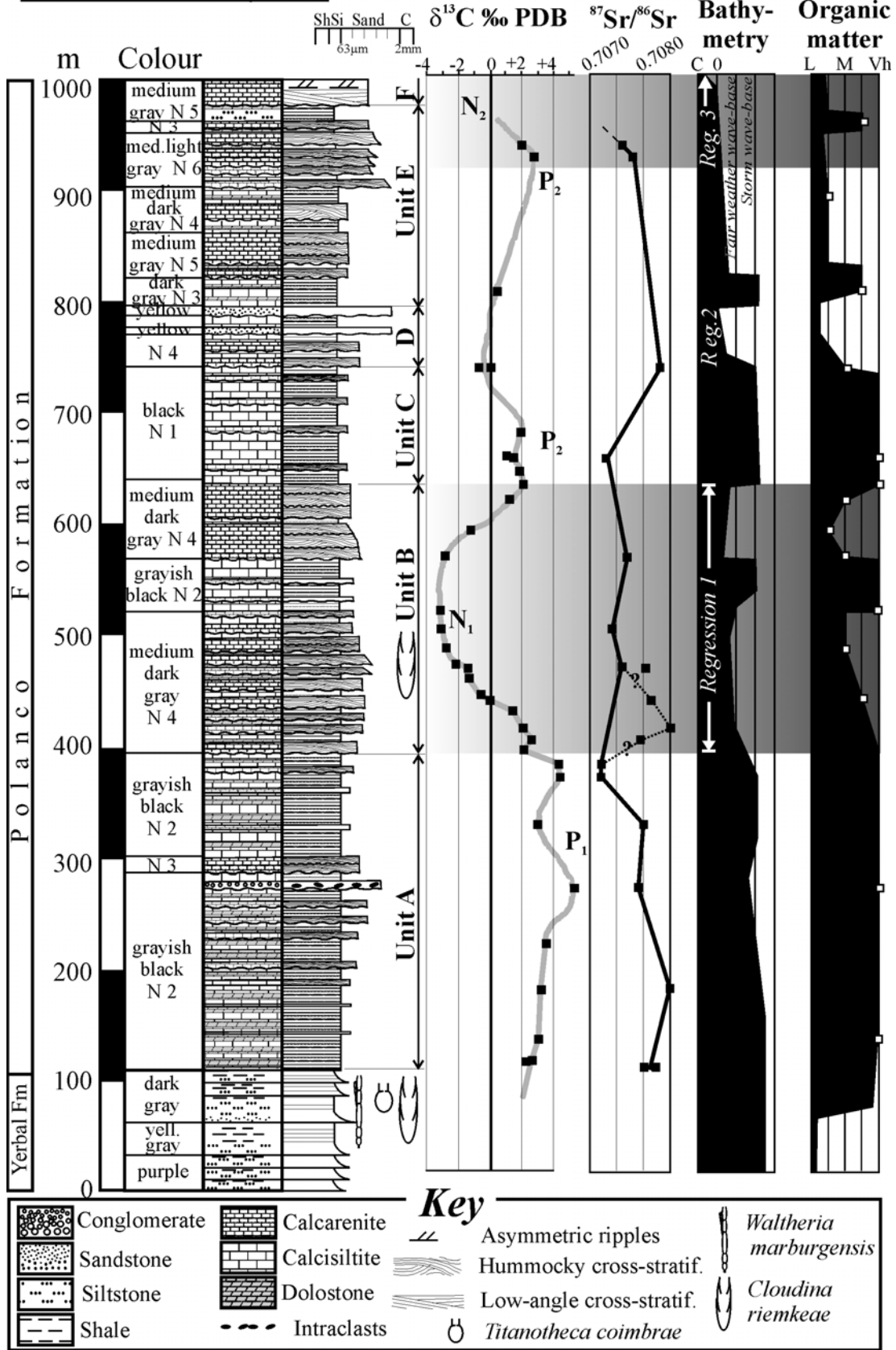


Fig. 4.4.3. C and Sr chemostratigraphy and lithostratigraphy of the Polanco Formation at Calera de Recalde Syncline, modified from Gaucher et al. (2004b, d). Inferred bathymetry (C: continental, 0: swash zone) and semi-quantitative organic matter content (L: low, M: medium, Vh: very high) of carbonates is shown.

reduction of organic matter content (TOC) of carbonates, a drop in $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (see below; Fig. 4.4.3) and diminished acritarch diversity are associated to excursion N_1 , suggesting that it represents a non-snowball glacial event (Gaucher et al., 2004b, 2004d).

4.4.2.2.3. Upper Polanco P_2 positive excursion

Positive $\delta^{13}\text{C}$ values of up to +3.3 ‰ PDB return in units C to E of the Polanco Formation, comprising more than 300 m of section (Figs. 4.4.2-3). Gaucher et al. (2004b) proposed that a small negative excursion may be recorded in unit D, but later work showed that $\delta^{13}\text{C}$ values never reach negative values but remain around 0 to +0.5 ‰ PDB. Therefore, this positive excursion has a central indentation (Fig. 4.4.3), as also observed for feature N_1 . Sea-level rise, higher TOC and higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios characterise carbonates of P_2 (Fig. 4.4.3).

4.4.2.2.4. Terminal Polanco-Barriga Negra negative excursion N_2

A significant negative excursion marks the transition between carbonates of the Polanco Formation and carbonate breccias and conglomerates of the Barriga Negra Formation (Gaucher, 2000; Gaucher et al., 2004b). $\delta^{13}\text{C}$ values drop to -1.8 ‰ PDB (Fig. 4.4.2), and in the shallower areas shelf exposure and erosion characterise this interval, the Barriga Negra Formation being directly derived from underlying sedimentary rocks. A significant, regional palaeokarst occurs on top of the Polanco Formation, known as the Barker Surface (Barrio et al., 1991, see Chapters 4.2 and 4.5), as a result of possibly glacioeustatic sea-level drop. In deeper settings, however, the shelf was not exposed and shales of the Cerro Espuelitas Formation directly overly the Polanco Formation (Gaucher, 2000). There, carbonates interbedded at the base of the Cerro Espuelitas Formation also show negative $\delta^{13}\text{C}$ values, thus recording N_2 feature in a deeper setting (Gaucher et al., 2004b).

4.4.2.2.5. Cerro Espuelitas positive excursion P_3 and basal Cerro Victoria negative excursion N_3

Carbonates are very rare between the lower Cerro Espuelitas/lower Barriga Negra Formation and the lower Cerro Victoria Formation. One carbonate bed of the middle Cerro Espuelitas Formation at its stratotype yielded $\delta^{13}\text{C}$ of +2.4 ‰ PDB (Fig. 4.4.2).

Microfossils of the Cerro Espuelitas Formation allow to place it in the Ediacaran (Gaucher, 2000), thus showing that the negative excursion at its base is older than the basal Cambrian negative excursion (Derry et al., 1994; Amthor et al., 2003; Maloof et al., 2005). The latter is likely represented in the ASG by the negative excursion at the base of the Cerro Victoria Formation, as shown by trace fossils occurring in that unit (Sprechmann et al., 2004; Gaucher et al., 2007b; Fig. 4.4.2). Therefore, negative excursions N₂ and N₃ are of different age, the positive, uppermost Ediacaran $\delta^{13}\text{C}$ excursion in between (P₃) being well represented in the Nama Group (Grotzinger et al., 1995), in the Corumbá Group (Misi et al., 2007) and elsewhere.

4.4.2.2.6. Cerro Victoria positive (P₄) and negative (N₄) excursions

After ca. 130 m of carbonates yielding negative $\delta^{13}\text{C}$ values, slightly positive values around +0.3‰ PDB are recorded at the transition between units B and C of the Cerro Victoria Formation (Gaucher et al., 2007b). $\delta^{13}\text{C}$ in positive excursion P₄ reaches +0.64‰ PDB and shows fairly constant values for ca. 180 m of mainly stromatolitic carbonates (Fig. 4.4.2). Up section (lower unit D), very low-amplitude, high-frequency oscillations around a mean value of 0 ‰ PDB are recorded in *Thalassinoides*-bearing dolostones (Sprechmann et al., 2004; Gaucher et al., 2007b). Finally, increasingly negative values of up to -1.1‰ PDB (N₄ excursion) characterise uppermost carbonates of the Cerro Victoria Formation (Fig. 4.4.2). The described pattern is identical to the Lower Cambrian $\delta^{13}\text{C}$ global curve between 542 and 535 Ma (lower Nemakyt-Daldyn) presented by Maloof et al. (2005). Multiple, successive negative excursions in this interval (Maloof et al., 2005), however, prevent an unambiguous assignment of negative excursions N₃ and N₄.

4.4.2.3 Sierras Bayas Group

Gómez Peral et al. (2007) report $\delta^{13}\text{C}$ analyses for carbonates and organic matter of the Villa Mónica, Loma Negra and lower Cerro Negro formations. The Sierras Bayas Group is a shallow-water succession characterised by multiple erosional unconformities which, in practical terms, means that the $\delta^{13}\text{C}$ record is fragmentary and mostly reflects periods of high sea level.

Stromatolitic dolostones of the Villa Mónica Formation record a positive $\delta^{13}\text{C}$ excursion between -1.4 and $+2.2$ ‰ PDB (Gómez Peral et al., 2007), being truncated at the top by an erosional unconformity (Figs. 4.2.1, 4.4.2). Consistently positive $\delta^{13}\text{C}$ values up to $+4.5$ ‰ PDB characterise carbonates of the overlying Loma Negra Formation, which are truncated by the Barker palaeokarst (see Chapters 4.2 and 4.5; Fig. 4.4.2). $\delta^{13}\text{C}$ of kerogens from the same unit vary between -28 and -27 ‰ PDB, $\Delta\delta^{13}\text{C}$ ranging between 30.5 and 32 ‰ PDB (Gómez Peral et al., 2007). Marls of the basal Cerro Negro Formation resting atop the Barker surface also show rising, positive $\delta^{13}\text{C}_{\text{carbonate}}$ values between $+3.5$ and $+4.3$ ‰ PDB (Fig. 4.4.2), and corresponding $\Delta\delta^{13}\text{C}$ values around 32 ‰ PDB. However, the fact that an important erosional surface separates the Loma Negra and Cerro Negro formations shows that the mentioned positive $\delta^{13}\text{C}_{\text{carbonate}}$ values in fact represent two different excursions (correlated here to P_1 and P_3 ; Fig. 4.4.2) and not just one.

4.4.2.4 Piedras de Afilar Formation

Limestones occur at the top of the Piedras de Afilar Formation, and yielded $\delta^{13}\text{C}$ between $+5.05$ and $+5.80$ ‰ PDB, with little variation in 10 m of section (Pamoukaghlian et al., 2006). Rather negative $\delta^{18}\text{O}$ values between -18.1 and -17.5 ‰ are best explained by the thermal effects of nearby dolerites, which probably did not alter the carbon isotopic composition.

4.4.3 STRONTIUM ISOTOPE CHEMOSTRATIGRAPHY

Strontium isotope data are available for carbonates of the Arroyo del Soldado Group (Kawashita et al., 1999a; Gaucher et al., 2004d, 2007b) and Sierras Bayas Group (Kawashita et al., 1999b). In this work, we report $^{87}\text{Sr}/^{86}\text{Sr}$ analyses of limestones from the Mina Verdún Group.

4.4.3.1. Mina Verdún Group

Ten analyses of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were carried out on limestones of the El Calabozo and lower Gibraltar formations at their stratotype. Six analyses from stromatolitic, pure limestones of the El Calabozo Formation yielded values between 0.7081 and 0.7189 .

Four analyses of limestones from the lower Gibraltar Formation at the same section yielded very radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ ratios mostly between 0.7153 and 0.7174. All values are too high, and obviously reflect either post-depositional resetting and/or input of radiogenic strontium from silicates. Only the least altered samples from the El Calabozo Formation, yielding ratios of 0.7081 and 0.7084, could be regarded as near-seawater values. However, according to Shields (2007) and Halverson et al. (2007), seawater $^{87}\text{Sr}/^{86}\text{Sr}$ ratios ranged between 0.7050 and 0.7065 in the upper Mesoproterozoic-Tonian (1300-800 Ma). Therefore, the currently available $^{87}\text{Sr}/^{86}\text{Sr}$ data for the Mina Verdún Group cannot be used to constrain its depositional age.

4.4.3.2. Arroyo del Soldado Group

A total of 47 $^{87}\text{Sr}/^{86}\text{Sr}$ analyses are known for the ASG, 33 being from limestones of the Polanco Formation (Gaucher et al., 2004d, 2006, this work) and 14 from dolostones of the Cerro Victoria Formation (Gaucher et al., 2007b). The behaviour of the Sr isotopic system in these two units differ markedly from each other. Whereas dolostones of the Cerro Victoria Formation yield too radiogenic values (>0.7106 : Gaucher et al., 2007b), limestones of the Polanco Formation yield mostly near-primary $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (Fig. 4.4.1C). This is readily explained by the lower Sr concentrations that dolomite allows in its lattice, which renders it more susceptible to post-depositional alteration (Gaucher et al., 2007d). On the other hand, limestones of the Polanco Formation are characterised by high Sr concentrations of up to 2250 ppm (mean=1025 ppm, standard deviation=470 ppm, N=88), and Rb concentrations typically less than 10 ppm, denoting a negligible proportion of radiogenic Sr from silicates. $^{87}\text{Sr}/^{86}\text{Sr}$ and $\delta^{13}\text{C}$ crossplots versus common alteration proxies (Fig. 4.4.1C) show the unaltered nature of these isotopic signatures. Considering this, only the results from the Polanco Formation will be discussed, which are the ones useful for strontium chemostratigraphy.

All samples from the Polanco Formation show $^{87}\text{Sr}/^{86}\text{Sr}$ values within 0.7073 and 0.7086, the only exceptions being samples near the contact of the intrusive Minas Granite, with values up to 0.7156. Even there, one sample retained a near-primary value of 0.7077 (Gaucher et al., 2006). The variation of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in limestones of the Polanco Formation is not random, but shows secular variations of an amplitude of 0.0010 (Gaucher et al., 2004d; Fig. 4.4.3). The lowermost Polanco Formation is

characterised by declining $^{87}\text{Sr}/^{86}\text{Sr}$ ratios between 0.7085 and 0.7080 (Fig. 4.4.3), dropping to a nadir of 0.7073 at peak $\delta^{13}\text{C}$ values of the uppermost unit A (Gaucher et al., 2004d). Throughout units B and C, $^{87}\text{Sr}/^{86}\text{Sr}$ values remain low (0.7073 to 0.7077), only increasing in unit D to a maximum of 0.7083. One possible exception could be the apparent surge of $^{87}\text{Sr}/^{86}\text{Sr}$ values from 0.7073 to 0.7086 at the base of unit B at Calera de Recalde Syncline (Fig. 4.4.3), but this needs confirmation from other sections. A final drop to 0.7076 is observed at the top of the Polanco Formation (unit E; Fig. 4.4.3). The consistency of the observed pattern, along with the geochemical indications of a primary isotopic signature (Fig. 4.4.1), suggest that the observed $^{87}\text{Sr}/^{86}\text{Sr}$ oscillations reflect secular variations of seawater composition. This has profound implications for both Neoproterozoic Sr-isotope chemostratigraphy and palaeoclimate models (see below).

4.4.3.3. *Sierras Bayas Group*

Gómez Peral et al. (2007) show, on the basis of different geochemical and isotopic proxies, that limestones of the Loma Negra Formation retain primary C and Sr isotopic signatures. Kawashita et al. (1999b) reports $^{87}\text{Sr}/^{86}\text{Sr}$ values between 0.7069 and 0.7087 for the Loma Negra Formation, with the least altered values between 0.7069 and 0.7075. As noted by Gaucher et al. (2005a), the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the Loma Negra Formation are almost identical to the ones reported for the Polanco Formation of the ASG. Both units were previously correlated on the basis of litho- and biostratigraphy (Gaucher et al., 2005a, see Chapters 4.2-4.3; Fig. 4.4.2).

No $^{87}\text{Sr}/^{86}\text{Sr}$ analyses are available for dolostones of the Villa Mónica Formation, which usually contain less than 45 ppm Sr (Gómez Peral et al., 2007).

4.4.4. SYNTHESIS

The age of the Mina Verdún Group is still poorly constrained, but the $\delta^{13}\text{C}$ plateau around 2 ‰ PDB observed in the El Calabozo and Gibraltar formations is typical of late Mesoproterozoic to Tonian carbonates (1300-800 Ma: Bartley et al., 2007). U-Pb dating of interbedded acid volcanics described by Gaucher et al. (2007a) will help improve the still poorly-constrained pre-Cryogenian global $\delta^{13}\text{C}$ curve.

Poiré and Gaucher (2007) pointed out that dolostones of the Villa Mónica Formation show a $\delta^{13}\text{C}$ curve similar to carbonates of the Mina Verdún Group, both units being pre-Ediacaran in age and containing *Conophyton* stromatolites (Fig. 4.4.2). However, similar, leiosphere-dominated acritarch assemblages in the Villa Mónica Formation and overlying units of the Sierras Bayas Group militate against this interpretation (Gaucher et al., 2005a; see Chapter 4.3). Thus, the issue of the correlation of the Villa Mónica Formation remains controversial.

The Arroyo del Soldado Group contains without doubt the most detailed late Ediacaran chemostratigraphic record in whole southwestern Gondwana between ca. 570 and 550 Ma. Positive excursion P_1 is recorded in the Yermal and Polanco Formations in the ASG, and in the Loma Negra Formation in the Sierras Bayas Group (Fig. 4.4.2). In both units, *Cloudina riemkeae* and low-diversity acritarch assemblages occur (Gaucher et al., 2005a; see Chapter 4.3). The impressive negative excursion N_1 that follows may be correlated to the Shuram–Wonoka anomaly (Halverson et al., 2005), which ended at 551 ± 1 Ma and reached its nadir before 555 ± 6 Ma in south China (Condon et al., 2005; Zhang et al., 2005). $\delta^{13}\text{C}$ values in the Shuram–Wonoka anomaly are as low as -10 ‰ PDB or less (Halverson et al., 2005), but in the Polanco Formation $\delta^{13}\text{C}$ does not drop beyond -4.5 ‰ PDB. This, however, can be due to ocean stratification or other local effects (Calver, 2000; Frimmel, 2004; Shen et al., 2005). The sedimentary record of the Polanco Formation favors a Phanerozoic-type glaciation as the cause of the negative anomaly (Gaucher et al., 2004d; Gaucher, 2007, see Chapter 4.5), as also observed for other Neoproterozoic negative anomalies. Most interestingly, $^{87}\text{Sr}/^{86}\text{Sr}$ reaches minimum values before the onset of N_1 , and stays low (0.7075) well into the following positive excursion (P_2 ; Figs. 4.4.2,3). The decrease of $^{87}\text{Sr}/^{86}\text{Sr}$ values during a glacial event could be due to: (a) ice cover of continents, diminishing the input of ^{87}Sr to the oceans (Jacobsen & Kaufmann, 1999); (b) pre-glacial, enhanced hydrothermal activity in mid-ocean ridges leading to plankton blooms, CO_2 sequestration and glaciation (Gaucher, 2007); and (c) a combination of both. Considering that $^{87}\text{Sr}/^{86}\text{Sr}$ ratios begin to fall well before the onset of negative anomaly N_1 (Fig. 4.4.3), enhanced hydrothermal activity linked to rifting events seems a more attractive hypothesis.

The same sequence of events is observed during the terminal Polanco-Barriga Negra negative excursion N_2 , which predates platform exposure both in the Arroyo del

Soldado and Sierras Bayas Group (Fig. 4.4.2), leading to the regionally important Barker palaeokarst (Barrio et al., 1991; Poiré and Gaucher, 2007). A tentative correlation of this surface with the Vingerbreek Member in the Nama Group, constrained between 549 ± 1 Ma and 545 ± 1 Ma (Grotzinger et al., 1995; Saylor et al., 1998), showing platform reworking, a negative $\delta^{13}\text{C}$ excursion to -2 ‰ PDB and evidences of a glacial event (Germs, 1995) is considered plausible (see Chapter 4.5).

Carbonates of the Cerro Victoria Formation, are confidently assigned to the Lower Cambrian on the basis of trace fossils and carbon isotopes (Gaucher et al., 2007b and references therein). The $\delta^{13}\text{C}$ curve reported for this unit is identical to the interval between 542 and 535 Ma (lower Nemakiy-Daldyn) in the high-resolution global $\delta^{13}\text{C}$ curve presented by Maloof et al. (2005). This is firm evidence that coeval Cambrian carbonates in the Cuyania-Precordillera Terrane in Argentina have potential counterparts in core areas of Gondwana (Finney, 2007), apart from Laurentia. The tectonic evolution of the area may explain the absence of carbonates –or other sedimentary rocks- younger than 535 Ma in the RPC (see Chapter 4.6).

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PALAEOCLIMATIC EVENTS*

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4.5.1. PALAEOCLIMATIC EVOLUTION OF THE RÍO DE LA PLATA PALAEOCONTINENT IN THE NEOPROTEROZOIC-CAMBRIAN

Neoproterozoic palaeoclimate in the Río de la Plata Craton (RPC) can be understood in terms of a warm, possibly tropical, “background” climate punctuated by several cold or even glacial events. Evidences of a dominantly tropical palaeoclimate are:

- (1) Thick carbonate deposits blanketed the marginal areas of the RPC in the Neoproterozoic, beginning in the late Mesoproterozoic-Tonian (Mina Verdún Group) and especially in the late Ediacaran-Early Cambrian (Sierras Bayas and Arroyo del Soldado groups).
- (2) Clay minerals derived from the granitic-metamorphic basement were predominantly kaolinitic during certain periods (Poiré, 1987; Gaucher, 2000; Pamoukaghlián et al., 2004).
- (3) Preliminary palaeomagnetic data suggest a tropical latitude for the late Ediacaran and Early Cambrian (Sánchez Bettucci and Rapalini, 2002)

On the other hand, clear evidences of ice-rafting and glacially-influenced deposits were reported from the Playa Hermosa and Las Ventanas formations (Pazos et al., 2003, 2008; Gaucher et al., 2008a), which could be lateral equivalents (Blanco and Gaucher, 2005). These evidences will be discussed in detail below.

* Gaucher, C., Poiré, D., 2009. Palaeoclimatic events. Neoproterozoic-Cambrian evolution of the Río de la Plata Palaeocontinent. In: Gaucher, C., Sial, A.N., Halverson, G.P., Frimmel, H.E. (Eds.): Neoproterozoic-Cambrian tectonics, global change and evolution: a focus on southwestern Gondwana. Developments in Precambrian Geology, Vol. 16, Elsevier, pp.

4.5.2. DIAMICTITES OF THE LAS VENTANAS AND PLAYA HERMOSA FORMATIONS

4.5.2.1 *Las Ventanas Formation*

Conglomerates of the Las Ventanas Formation (Quebrada de Viera Member) are mostly orthoconglomerates without signs of glacial influence (Blanco and Gaucher, 2005).

A diamictite level occurs at the Mina Verdún quarry, which is overlain with erosional unconformity by typical orthoconglomerates of the Quebrada de Viera Member (Gaucher et al., 2008a). Basalts and breccias of the La Rinconada Member are not present there, and the Quebrada de Viera Member overlies carbonates of the Mina Verdún Group (Poiré et al., 2003b, 2005a; Gaucher et al., 2008a, see Chapter 4.2) with erosional and angular unconformity. Similar stratigraphic relationships were observed also at Burgueño Quarry (Gaucher et al., 2007a, 2008a). A prominent palaeokarst is developed at both localities on top of carbonates of the Mina Verdún Group and filled with diamictites. These are made up of angular to subrounded clasts up to 25 cm in diameter (Figs. 4.5.1A-C), composed of underlying carbonates, granitoids, rhyolite, metasilstone, sandstone, quartzite and epidote-rich metavolcanics (Gaucher et al., 2008a). Clasts are embedded in a reddish or greenish, ferruginous mudstone matrix mainly composed of clay and silt. The matrix averages 35 to 40 % by volume, but transitions to clast-free mudstones are common. Diamictites are massive and chaotic or slightly graded. Lonestones made up of limestone, metasilstone and granite occur in laminated, red siltstones (Figs. 4.5.1A-C). In a few cases the long axis of the clasts is perpendicular to bedding plane (Fig. 4.5.1C), allowing its interpretation as a “bullet-clast” dropstone. Deformation of underlying laminae, and onlap of overlying strata on the top of the dropstones confirms this interpretation. All petrographic features of the described diamictites, along with the occurrence of dropstones, suggest a glacial origin (Gaucher et al., 2008a). However, no cap carbonate, like those often seen on top of other Neoproterozoic glacial units (e.g. Kennedy, 1996; Hoffman and Schrag, 2002), occurs on top of the diamictites. This may be due to truncation during the events responsible for the overlying, clast-supported conglomerates or, alternatively, the absence of a cap carbonate may represent a primary feature of the glacial event in Uruguay, as observed in the type region for the 583 Ma Gaskiers glaciation (MacGabhann, 2005).

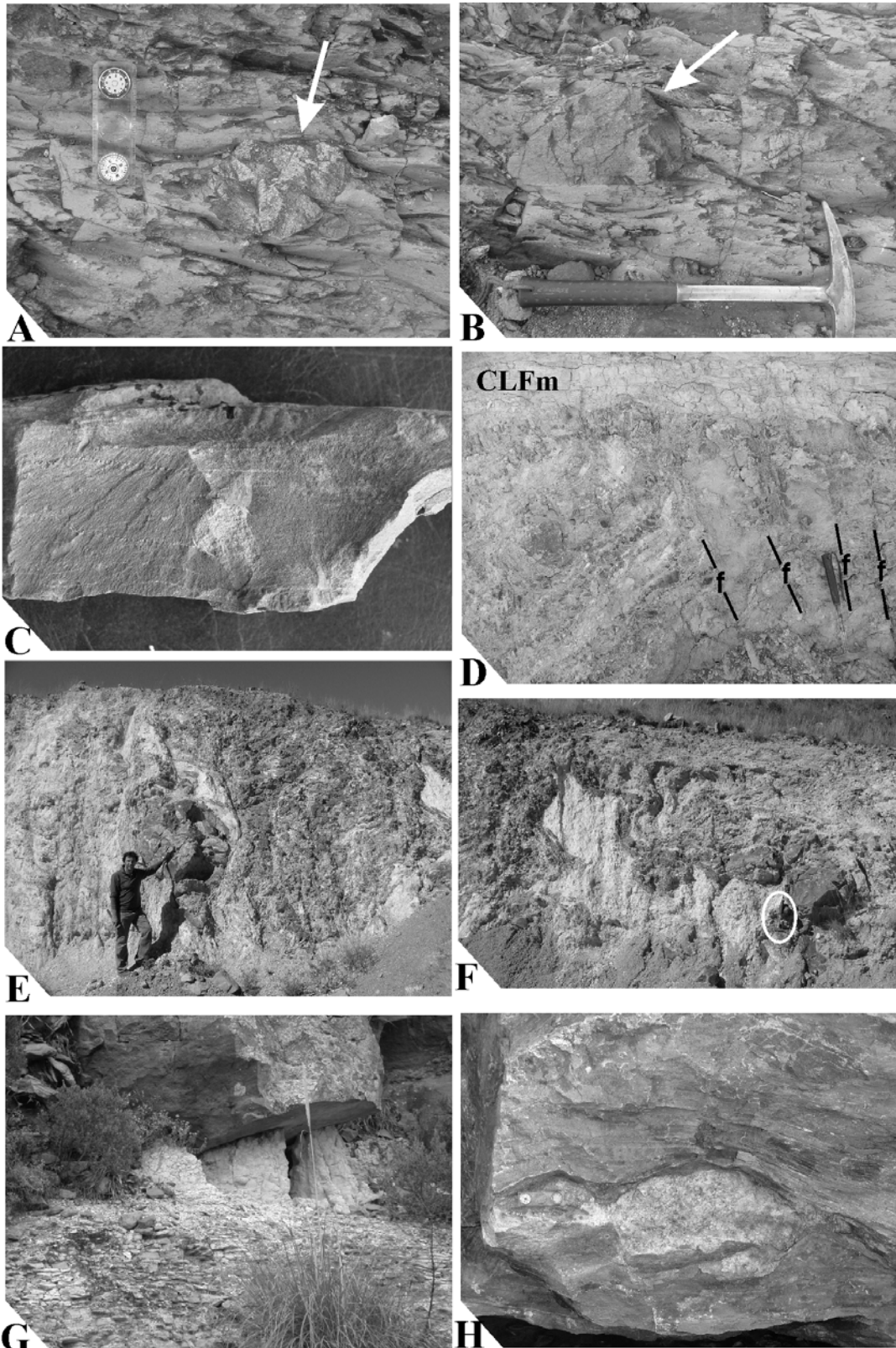


Fig. 4.5.1. Neoproterozoic diamictites in the RPC. (A-B) Out-sized clasts in rain-out diamictite of the Las Ventanas Formation in the vicinity of Burgueño Quarry. Clasts (arrowed) are composed of granite (A) and metapelite (B). Scale in A is 8 cm long. (C) Granitic bullet-clast (long axis is 2 cm long) in red siltstones, lower Las Ventanas Formation at Mina Verdún (from Gaucher et al., 2008a). (D) Normal faults (f) cross-cutting diamictites of the Colombo Member at Piedra Amarilla Quarry. Note truncation of faults by overlying sandstones of the Cerro Largo Formation (CLFm). (E) Large boulder (2 m in diameter) in the Colombo Member, Piedra Amarilla Quarry. (F) Convolute bedding (slump), Colombo Member at

Piedra Amarilla Quarry. Hammer is indicated for scale. (G) Sierra del Volcán diamictites overlain by sandstones of the Balcarce Formation. Height of the outcrop is 1.5 m. (H) Boudinaged granite in limestones of the Carapé Group at Floridán Quarry. Scale (center left) is 8 cm long.

The main age constraints for the diamictite level of the lower Las Ventanas Formation are:

- (1) Tholeiitic andesites tectonically intercalated in the Fuente del Puma Group and dated U-Pb SHRIMP at 590 ± 2 Ma by Mallmann et al. (2007, see Chapter 4.6) probably correlate with basalts of the La Rinconada Member of the Las Ventanas Formation (615 ± 30 Ma, K-Ar). If this is accepted, diamictites of the lower Las Ventanas Formation are younger than 590 Ma.
- (2) Acritarchs of the middle and upper Las Ventanas Formation have been assigned by Gaucher et al. (2008a) to the Ediacaran Leiosphere Palynoflora (Grey et al., 2003; Grey, 2005), loosely constrained between the base of the Ediacaran Period and the Acraman impact (ca. 570 Ma; see Chapter 9.1).

Therefore, the mentioned data strongly suggest that the Las Ventanas diamictites were deposited during the Gaskiers glaciation (582 ± 0.4 Ma: Bowring, in MacGabhann, 2005; Bowring et al., 2003), as suggested by Blanco and Gaucher (2005).

4.5.2.2 Playa Hermosa Formation

The Playa Hermosa Formation (Masquelín & Sánchez Bettucci, 1993) crops out in a narrow strip along the coast near Piriápolis, a section described in detail by Pazos et al. (2003). The unit is made up of diamictites, breccias, conglomerates, sandstones and rhythmites, showing evidence of slope instability (i.e. slumping) and resedimentation processes (Pazos et al., 2003). Sandstone layers bear all the hallmarks of turbidites, and also indicate a relatively steep palaeoslope. Lonestones up to 1 m in length were interpreted as dropstones by Pazos et al. (2003, 2008), although sliding processes may have also played an important role in a tectonically unstable, steep-sloped basin.

Thus, the Playa Hermosa Formation differs markedly from the Las Ventanas Formation, because:

- (1) it represents deep-water, slope sediments;
- (2) resedimentation and turbidity currents were the most important depositional processes;

(3) lithologically, conglomerates of the Playa Hermosa Formation contain up to 50% intraclasts and 30% quartzites (Blanco and Gaucher, 2005), in contrast to the granite-dominated conglomerates of the Las Ventanas Formation.

The contact between both units is tectonic, but it is reasonable to postulate that the Playa Hermosa Formation is the deep-water lateral equivalent of the shallower Las Ventanas Formation (Blanco and Gaucher, 2005).

No direct age constraints are available for the Playa Hermosa Formation. However, both the correlation with the Las Ventanas Formation (Blanco and Gaucher, 2005) and the soft-sediment interaction with basic lavas of the La Rinconada Member suggest an age around 590-580 Ma (Pazos et al., 2008). Therefore, the unit may record the Gaskiers glacial event in a more distal, slope environment.

4.5.3. DIAMICTITES IN TANDILIA

4.5.3.1. *Sierra del Volcán diamictites*

In the southeastern part of Tandilia, a thin (<8 m) diamictite unit known as Sierra del Volcán Formation occurs (Spalletti and del Valle, 1984). It rests with erosional unconformity on the Palaeoproterozoic Buenos Aires Complex, and according to Pazos et al. (2008) is overlain with angular unconformity by the Middle Ordovician to Lower Silurian Balcarce Formation (Seilacher et al., 2002; Rapela et al., 2007; Fig. 4.5.1G). However, Poiré and Spalletti (2008) suggested a possible Ashgillian (Hirnantian) age for these glacial deposits on the basis of Lower Ordovician U-Pb ages of detrital zircons reported by Van Staden et al. (in review) from the Sierra del Volcán Diamictite. The apparent angular unconformity between the Sierra del Volcán and Balcarce formations (Fig. 4.5.1G) is probably due to their different rheology. Therefore, this glacial unit should be referred to the Hirnantian, which has a very important record in southwestern Gondwana. The geochronological information provided by detrital zircons from the Sierra del Volcán Diamictite and the Balcarce Formation suggests that the kaolinite-rich Balcarce Formation could represent the onset of Silurian greenhouse conditions.

4.5.3.2. *Colombo diamictites*

In the Sierras Bayas Group, a diamictite horizon (Colombo Member) overlies a karstic surface on top of dolostones of the Villa Mónica Formation, and is in turn overlain with erosional unconformity by glauconitic sandstones of the lower Cerro Largo Formation. It is made up of sandstone, chert and breccia clasts up to 2.5 m in diameter, embedded in a greenish pelitic matrix (Fig. 4.5.1E). Thickness of the Colombo Member never exceeds 10 m. So far, no evidences of a glacial origin have been reported for this unit. Furthermore, a number of features suggest that gravity flows in a tectonically-unstable environment may be responsible for deposition of this diamictite level, namely:

- (a) Large-scale slumps occur, as evidenced by large-scale convolute bedding (Fig. 4.5.1F).
- (b) Normal faults disturb bedding, but are truncated by the overlying sandstones of the Cerro Largo Formation (Fig. 4.5.1D), showing that they represent syn-sedimentary extensional faulting during deposition of the Colombo Member.
- (c) Clast-supported conglomerates and bedded diamictites occur, which are not of glacial origin.

The stratigraphic position of these mass-flow deposits is analogous to the Las Ventanas and Playa Hermosa formations in Uruguay. The geotectonic setting, like in the latter unit, was probably the border of a rift basin, characterised by a steep palaeorelief and tectonic instability (Masquelín and Sánchez Bettucci, 1993; Pazos et al., 2003; Blanco and Gaucher, 2005; Gaucher et al., 2008a). However, in the case of the Colombo Member, evidence of glacial processes is lacking. It is worth noting that the bulk of conglomerates of the Las Ventanas Formation is not of glacial origin, but were deposited in alluvial fans (Blanco and Gaucher, 2005; Gaucher et al., 2008a).

4.5.4 OTHER UNITS

Pazos et al. (2008) report the occurrence of limestones in laminated limestones of the Zanja del Tigre Formation (Sánchez Bettucci and Ramos, 1999), now assigned to the Marco de los Reyes Formation of the Carapé Group (Bossi et al., 2007; see Chapter 4.1). Pazos et al. (2008) interpret these limestones, composed of granite, quartzite and gabbro (Fig. 4.5.1H), as dropstones. Apart from the great uncertainties regarding the age assignment of the Carapé Group (Bossi et al., 2007; Basei et al., 2008), there are serious

objections to the glacial interpretation of Pazos et al. (2008). The occurrence of boudinaged granites (Fig. 4.5.1G) is a characteristic feature of limestones of the Carapé Group (Bossi and Navarro, 1991; Rossini, 2002; Bossi et al., 2007). Disrupted quartz-arenite layers also occur, resulting in boudinaged quartzite blocks within the less competent, sheared limestone. Deformation of the limestones is ductile and very significant (Rossini, 2002; Bossi et al., 2007), as shown by:

- (a) vicinity (ca. 10 km) to the continental-scale Sierra Ballena Shear Zone (Fig. 4.1.4),
- (b) preservation of limestones as thin basement inliers between large granite intrusions (Fig. 4.1.4),
- (c) tight anisopachous folding with subvertical axial planes, and
- (d) obliteration of all primary sedimentary structures.

That the limestones are not of glacial origin is shown by granite “clasts” often connected to each other (Fig. 4.5.1H), typical of rheologically more resistant, boudinaged layers within a plastic matrix. Finally, $\delta^{13}\text{C}$ of the Marco de los Reyes limestones is predominantly positive (Chiglino et al., 2008), unlike glacially-related units (e.g. Halverson et al., 2005).

4.5.5. INDIRECT EVIDENCE OF LATE EDICARAN GLACIATION

4.5.5.1 Middle Polanco Formation

Gaucher et al. (2004b,d) showed that an important $\delta^{13}\text{C}$ negative excursion encompassing 200m of carbonates occurs in the middle Polanco Formation (Unit B: see Chapter 4.4). The following changes in measured parameters occur concomitantly to a drop in $\delta^{13}\text{C}$ (Fig. 4.4.3):

- (a) Remarkable sea-level drop of > 50 m, as shown by sedimentary structures.
- (b) Positive to negative $\delta^{13}\text{C}$ excursion, between +5.5 and -4.5 ‰ PDB
- (c) Low $^{87}\text{Sr}/^{86}\text{Sr}$ values between 0.7073 and 0.7077 throughout Unit Polanco B, as compared to $^{87}\text{Sr}/^{86}\text{Sr}$ ratios for underlying carbonates of 0.7080-0.7085.

- (d) Lower organic matter contents.
- (e) Lower diversity of organic-walled microfossils.

The most plausible explanation for these combined phenomena is a severe but non-global glacial event. Glaciation would have caused lowering of sea-level, decrease in plankton bioproductivity, reduction of the amount of organic matter exported to the sediments, environmental stress leading to lower acritarch diversity, a negative $\delta^{13}\text{C}$ excursion and lowering of $^{87}\text{Sr}/^{86}\text{Sr}$ values due to diminished chemical weathering and continental runoff (see Chapter 4.4 for a full discussion). Absence of glacial deposits indicates that the glacial event did not reach the Polanco shelf, which probably was at tropical latitudes. According to palaeomagnetic data reported by Sánchez Bettucci and Rapalini (2002), the Río de la Plata Craton was located at 17.5° latitude by 550 Ma.

The middle Polanco glacial event may have been a significant perturbation of the ocean-climate system. As shown in Chapter 4.4, the middle Polanco negative anomaly probably correlates with the Shuram-Wonoka anomaly (Halverson et al., 2005), which ended at 551 ± 1 Ma and reached its nadir before 555 ± 6 Ma in south China (Condon et al., 2005; Zhang et al., 2005). Both the occurrence of *Cloudina riemkeae* in Unit B of the Polanco Formation (see Chapter 4.3), and an U-Pb detrital zircon age of 566 ± 8 Ma in the overlying Barriga Negra Formation (Blanco et al., in press) are consistent with this interpretation. A glacially-related interpretation of the Shuram-Wonoka anomaly in Oman has been put forward by Cozzi et al. (2002), and is in agreement with our results.

4.5.5.2 *Barker surface*

The uppermost Polanco Formation-basal Barriga Negra Formation are also characterised by a negative $\delta^{13}\text{C}$ excursion, lowering of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (Fig. 4.4.3), changes in acritarch diversity, and –most significantly- drop in sea-level leading to platform exposure and erosion (Gaucher, 2000; Gaucher et al., 2004b,d; Figs. 4.5.2C-D). The same event is recorded in the Sierras Bayas Group in the uppermost Loma Negra Formation, which was karstified and filled with karstic breccias as a result of this regression (Barrio et al., 1991; Figs. 4.5.2A-B). The karstic surface has been termed “Barker Surface” by Poiré and Gaucher (2007). Age constraints for this event are

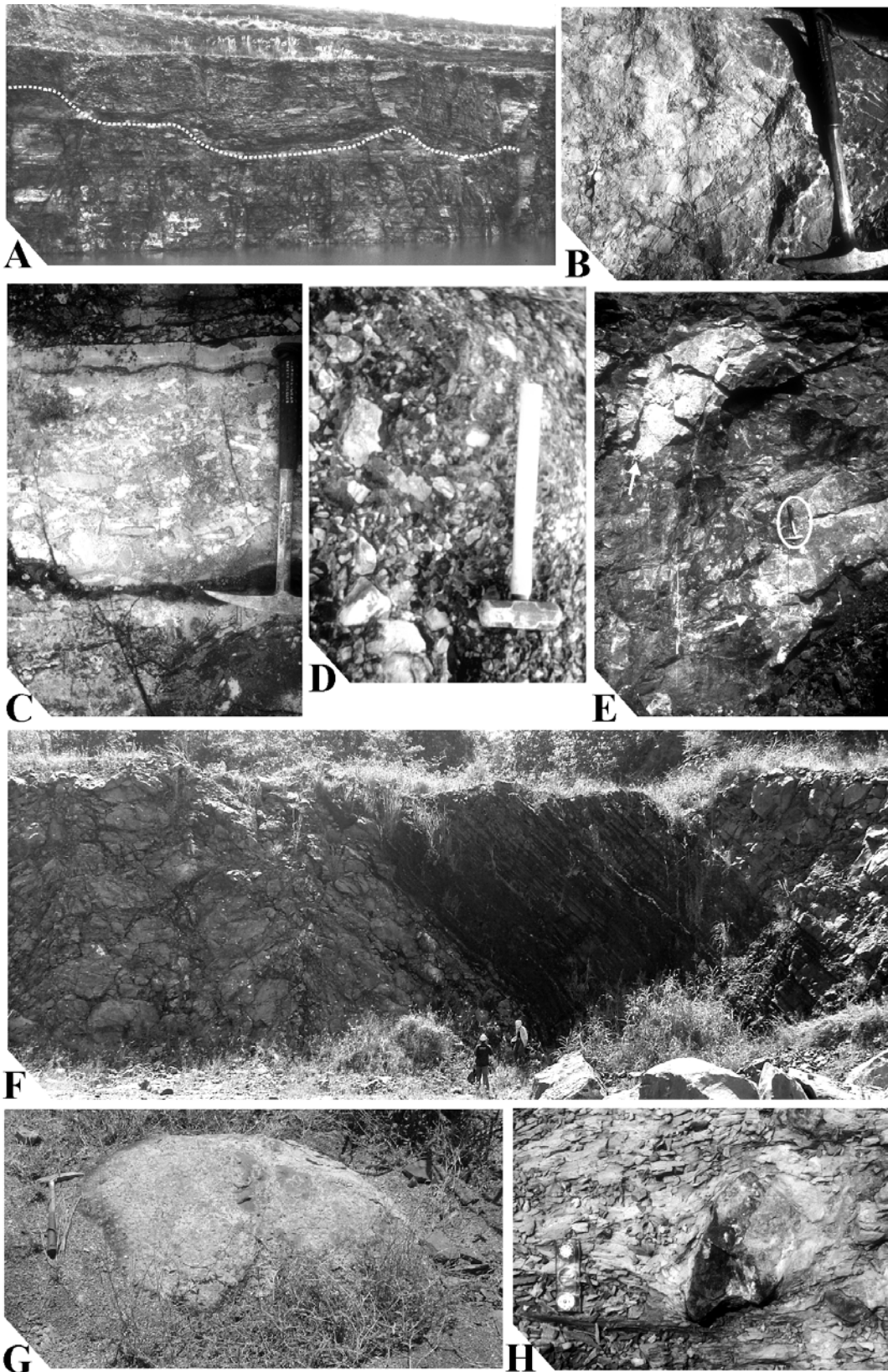


Fig. 4.5.2. Barker surface and possible correlates in the Paraguay Belt. (A) Palaeokarst on top of the Loma Negra Formation (dotted) and covered by pelites and sandstones of the Cerro Negro Formation, Loma Negra Quarry near Olavarría. (B) Limestone breccias filling the palaeokarst at Polvorín Quarry (Olavarría area). (C) Graded carbonate breccias at the Polanco-Barriga Negra transition (from Gaucher, 2000). (D) Limestone breccia, lower Barriga Negra Formation. (E) Metric clasts (arrowed) in breccias of the lower Tamengo Formation, Laginha Quarry (see F). Hammer is shown as scale (circle). (F) Overview of breccias in the Lower Tamengo Formation, separated by fine-grained, dark limestones and marls

yielding negative $\delta^{13}\text{C}$ values. (G) Laginha diamictite: outsized clast 2 m in diameter, made up of oncolitic limestone, Laginha Quarry. (H) Laginha diamictite: outsized clast in a siltstone matrix, showing deformation features characteristic of dropstones (Laginha Quarry).

provided by (a) *Cloudina riemkeae* occurring in the underlying Polanco and Loma Negra formations; and (b) an U-Pb detrital zircon age of 566 ± 8 Ma for sandstones of the lower Barriga Negra Formation (Blanco et al., in press). Therefore, the Barker Surface event is latest Ediacaran in age, allowing a preliminary correlation with the Vingerbreek Member in the Nama Group, constrained between 549 ± 1 Ma and 545 ± 1 Ma (Grotzinger et al., 1995; Saylor et al., 1998). There, platform reworking, a negative $\delta^{13}\text{C}$ excursion to -2 ‰ PDB, striated pavements and other evidences of a glacial event have been described (Germs, 1995 and references therein, see Chapter 5.4).

4.5.6. RECORD OF CAMBRIAN GLOBAL WARMING

During deposition of the Cerro Victoria Formation (ASG) tropical climate prevailed once again in the RPC, as demonstrated by thick carbonate deposits (400 m), abundance of ooids and evaporitic minerals (Montaña and Sprechmann, 1993; Sprechmann et al., 2004; Gaucher et al., 2007b). Common occurrence of calcareous storm deposits confirms this interpretation (Gaucher, 2000). As noted by Gaucher et al. (2007b), climate stabilisation is indicated by the decreasing amplitude of $\delta^{13}\text{C}$ excursions in the ASG, being only 2 ‰ PDB in the Cerro Victoria Formation (see Chapter 4.4).

Lower Cambrian carbonate ramps are scarce in southwestern Gondwana. However, a number of carbonate successions broadly correlative with the Cerro Victoria Formation have been described, suggesting tropical conditions at least in part of Gondwana. These are:

- (a) the thick carbonate deposits of the Precordillera terrane, and especially La Laja Formation, which according to Sial et al. (2003) is of Lower Cambrian age (Chapter 11.1). Affinity of the Precordillera terrane is currently the subject of intense debate, but data presented by Finney (2007) seem to favour a Gondwanic affinity;
- (b) the Nelson Limestone Formation, exposed in the Pensacola Mountains (Transantarctic Province) of Antarctica, comprises oolitic, pisolitic and

archaeocyathan limestones of Lower to Middle Cambrian age (Laird, 1995). Rhyolites occurring in the conformably overlying Gambacorta Formation yielded an Upper Cambrian age, confirming the biostratigraphic data (Schmidt, in Laird, 1995); and

- (c) the Tucavaca Group of Bolivia comprises stromatolitic limestones conformably overlain by siliciclastics bearing *Treptichnus (Phycodes) pedum*, suggesting a Lower Cambrian age for the unit (Trompette, 1994).

Carbonate deposition was probably diminished due to high erosion rates in the mountain ranges generated during the Pan African-Brasiliano system of orogens (Trompette, 1994). Therefore, most Cambrian sedimentary successions in western Gondwana are siliciclastic and carbonate ramps were restricted to areas with low clastic input. Most interesting, Early Cambrian successions comprising carbonate ramps in western Gondwana occur in blocks located at tropical paleolatitudes (Antarctica, Río de la Plata Craton and Argentine Precordillera), according to the palaeogeographic reconstruction of Finney et al. (2003).

4.5.7. DISCUSSION AND SUMMARY

It becomes clear that in the late Neoproterozoic the RPC was predominantly under tropical climate (Fig. 4.5.3). If carbonate deposits of the Mina Verdún Group turn out to be Tonian in age (see Chapters 4.2-4.4), tropical conditions may have lasted for most of the Neoproterozoic. However, due to the gap in the sedimentary record, which encompasses at least the whole Cryogenian (Fig. 4.5.3), no data exists on the effects of the most severe Neoproterozoic glacial events on the palaeocontinent. This might explain the “paradox” put forward by Eerola (2002) regarding the absence of Neoproterozoic glacial deposits in the RPC and neighbouring areas. On the other hand, diamictites of the Las Ventanas and Playa Hermosa formations show that the RPC was briefly covered by ice during the Gaskiers glaciation. The limited extent of glacially-related units suggests that this event was short-lived in the RPC, as also shown by U-Pb dating of ash beds at its type locality (Bowring et al., 2003; MacGabhann, 2005).

One of the most interesting features of the Neoproterozoic paleoclimatic record in the RPC is the strong evidence supporting post-Gaskiers glaciation (Fig. 4.5.3). The glacial

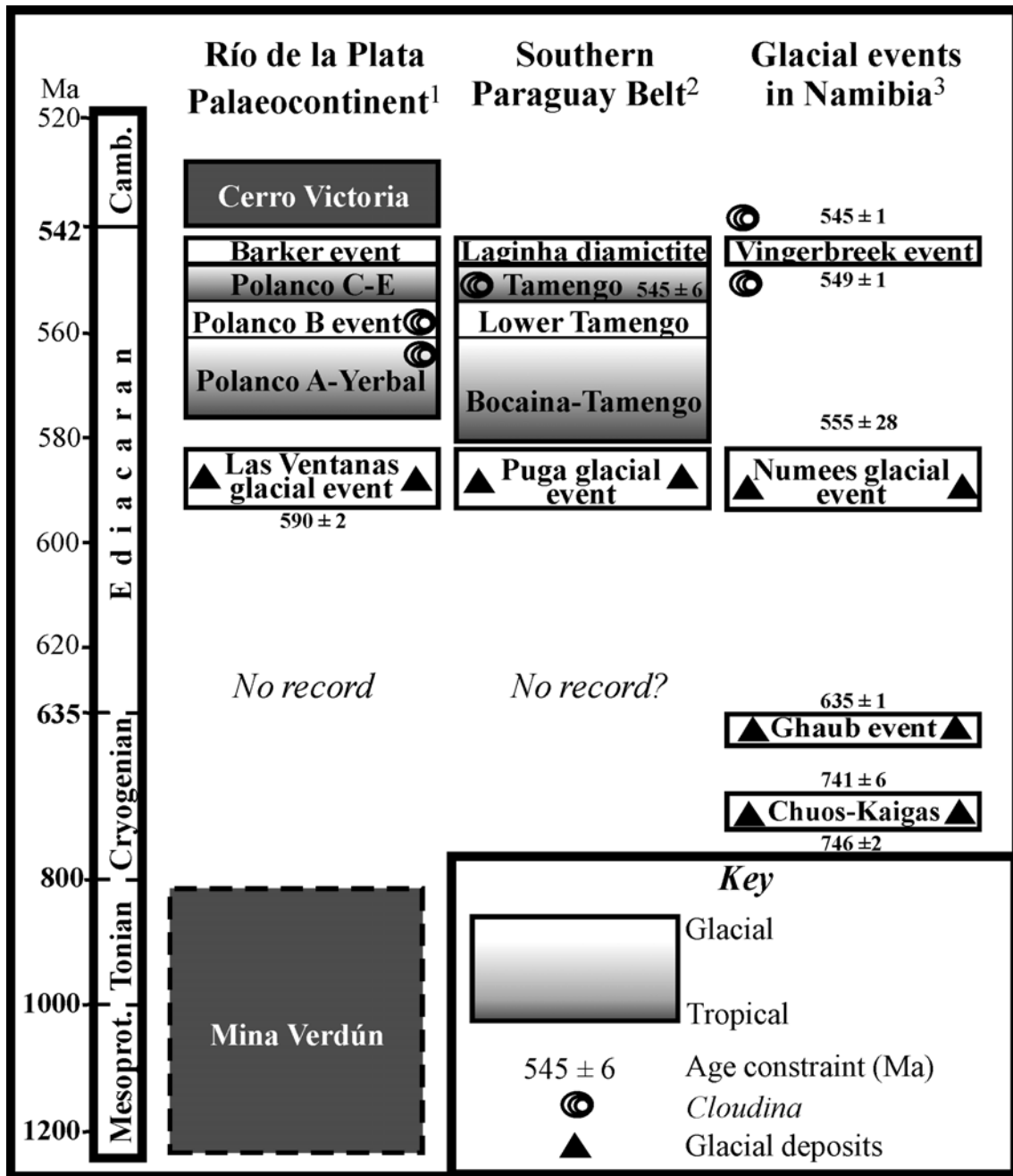


Fig. 4.5.3. Summary of Neoproterozoic-Cambrian palaeoclimatic events in the RPC, and comparison with events recorded in the southern Paraguay Belt and Namibia (Gariiep and Damara belts). Note changes in the time scale (not linear). Absolute ages according to: (1) Mallmann et al. (2007); Barker event: Blanco et al. (in press; 566 ± 8 Ma) ; (2) Babinski et al. (2007); (3) from base to top: Hoffman et al. (1996), Frimmel et al. (1996), Hoffmann et al. (2004), Fölling et al. (2000) and Grotzinger et al. (1995).

events represented by Unit B of the Polanco Formation and the Barker Surface event are significantly younger (> 20 My) than the age currently accepted for the Gaskiers event, which ended at 582 ± 0.4 Ma (Bowring, in MacGabhann, 2005). Despite the fact that ice did not cover the RPC, compelling evidence supporting Phanerozoic-type glacial events includes: sea-level drop, shelf exposure and erosion, negative $\delta^{13}\text{C}$ excursions, diminished bioproductivity, impact on acritarch diversity and lowering of $^{87}\text{Sr}/^{86}\text{Sr}$

seawater ratios. Whereas the Polanco B glacial event is a likely correlative to the global Shuram-Wonoka anomaly (ca. 560-551 Ma), the Barker Surface is correlated to the Vingerbreek glacial event (Germs, 1995 and references therein), constrained between 549 ± 1 Ma and 545 ± 1 Ma (Fig. 4.5.3). These glacial events may have had a profound impact on the oxygenation of the atmosphere, evolution of eukaryotes and metazoans in particular (Fike et al., 2006).

On a regional scale, the glacial record of the RPC bears close resemblance to that encountered in the southern Paraguay Belt (SW-Brazil, see Chapter 2; Fig. 4.5.3), which was probably also deposited on the Río de la Plata Palaeocontinent (Gaucher et al., 2003, 2008b; see Chapter 8). Glacial deposits of the Las Ventanas/Playa Hermosa Formation probably correlate to glacial diamictites of the Puga Formation in the southern Paraguay Belt (Boggiani and Alvarenga, 2004). The Polanco B glacial event may be represented by breccias and calcarenites of the lower Tamengo Formation (Corumbá Group, Boggiani, 1998; Gaucher et al., 2003; Figs. 4.5.2E-F), which show a well-developed, negative $\delta^{13}\text{C}$ excursion to -3 ‰ PDB (Misi et al., 2006; Velásquez et al., 2008). A recent U-Pb SHRIMP zircon age of 543 ± 2 Ma for an ash bed at this level (Babinski et al., 2008) is considered too young for the following reasons (Fig. 4.5.3): (a) 100 m of *Cloudina*-bearing limestones and marls occur above that level, which should be accommodated in the remaining 1 My of the Neoproterozoic (Amthor et al., 2003), implying unreasonably high sedimentation rates for the upper Tamengo Formation; (b) the dated ash bed occurs within a negative excursion in the Laginha Quarry, on top of which a $\delta^{13}\text{C}$ plateau around $+3$ ‰ occurs (Babinski et al., 2008). The opposite is observed in Namibia and Oman, where ash beds dated at 543 Ma occur within the top of this plateau (Grotzinger et al., 1995; Amthor et al., 2003). A correlation of the negative excursion at the base of the Tamengo Formation with the Shuram-Wonoka anomaly is thus preferred here.

A diamictite horizon (Laginha diamictite) occurs atop karstified limestones of the Tamengo Formation (Boggiani et al., 2004), at the same stratigraphic level of the Barker Surface (Figs. 4.5.2G-H). *Cloudina* occurring in underlying limestones of the Tamengo Formation (Gaucher et al., 2003) confirms this correlation. A preliminary U-Pb SHRIMP zircon age of 545 ± 6 Ma (Babinski et al., 2006) for ash beds interbedded in the same unit is also consistent with this interpretation.

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PALAEOGEOGRAPHY*

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4.6.1. RIO DE LA PLATA CRATON: A PART OF RODINIA

The apparent absence of Mesoproterozoic rocks in the Río de la Plata Craton (RPC) was interpreted in the past as evidence of this block being unrelated to Rodinia (e.g. Cordani et al., 2003). Other authors, however, pointed out that geological similarities between the Amazonia and Río de la Plata paleocontinents suggested that they were linked in the Mesoproterozoic, and were part of Rodinia (e.g. Meert and Torsvik, 2003).

In Uruguay, the occurrence of Mesoproterozoic tectonomagmatic events in the RPC has been advocated by different authors in the last 15 years. While mapping in detail the Piedra Alta mafic dyke swarm, dated at 1790 ± 5 Ma (U-Pb on beddeleyite: Halls et al., 2001; Ar-Ar: Teixeira et al., 1999), Bossi and Campal (1992) recognised and described a dextral megashear zone (Sarandí del Yí Shear Zone, SYSZ) with mylonites up to 8 km thick (Fig. 4.6.1). The curvature of the eastern part of the dyke swarm is consistent with dextral shear, unlike all “Brasiliano-Pan African” structures in the area. The age of the SYSZ is probably late Mesoproterozoic as shown by post-emplacement thermal overprinting of the Piedra Alta mafic dike swarm (Bossi et al., 1993a) between 1370 and 1170 Ma, reflected in initial release portions of ^{40}Ar - ^{39}Ar spectra and Rb-Sr mineral isochrons (Teixeira et al., 1999). Bossi and Navarro (2001) report an ^{40}Ar - ^{39}Ar age of 1240 ± 5 Ma for the same event (Fig. 4.6.1). Cingolani (in Bossi et al., 1998) reported K-Ar ages of 1253 ± 32 Ma for synkinematic muscovites that crystallised along southwest-vergent thrust planes in the Nico Pérez Terrane (Fig. 4.6.1), thus consistent with the above-mentioned shear sense and timing of this tectonic event. The SYSZ is a

* Gaucher, C., Bossi, J., Blanco, G., 2009. Palaeogeography. Neoproterozoic-Cambrian evolution of the Río de la Plata Palaeocontinent. In: Gaucher, C., Sial, A.N., Halverson, G.P., Frimmel, H.E. (Eds.): Neoproterozoic-Cambrian tectonics, global change and evolution: a focus on southwestern Gondwana. Developments in Precambrian Geology, Vol. 16, Elsevier, pp.

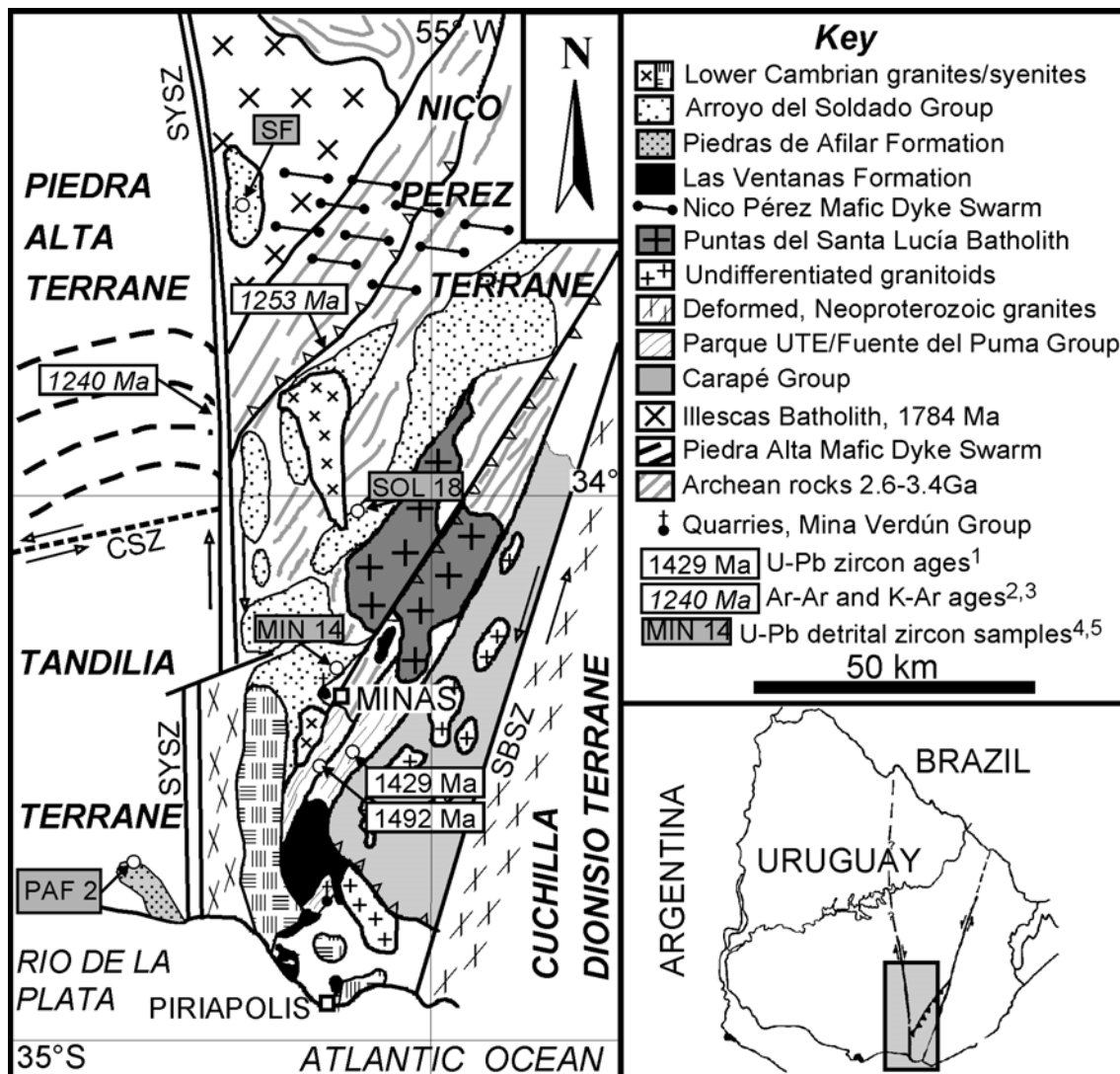


Fig. 4.6.1. Geological sketch map of the southern Nico Pérez Terrane and surrounding areas. Sources of Mesoproterozoic radiometric ages: 1: Oyhantçabal et al. (2005); 2: Bossi and Navarro (2001), Teixeira et al. (1999); 3: Cingolani in Bossi et al. (1998); 4: Gaucher et al. (2008b), 5: Blanco et al. (in press), only sample SF. SYSZ: Sarandí del Yí Shear Zone (shear sense indicated for the main Mesoproterozoic event), SBSZ: Sierra Ballena Shear Zone; CSZ: Colonia Shear Zone. Modified from Gaucher et al. (2008b).

continental-scale megashear (Unrug, 1996), which suggests significant reworking of the RPC during the Grenvillian Orogeny.

Bossi et al. (1998) assigned a number of volcanosedimentary successions in the southern Nico Pérez Terrane (Fuente del Puma and Carapé groups) to the Mesoproterozoic. This age assignment is confirmed by: (a) U-Pb zircon ages of 1429 ± 21 Ma for metarhyolites of the upper Fuente del Puma Group, and 1492 ± 4 Ma for metagabbros intruding the succession elsewhere (Oyhantçabal et al., 2005; Fig. 4.6.1); (b) one K-Ar age of 1208 ± 10 Ma for metabasalts of the Fuente del Puma Group (Gómez Rifas, 1995), (c) Pb-Pb measurements of galenas from the same unit yield ages between 1500 and 1200 Ma (Garau, in Bossi and Navarro, 2001); (d) carbon isotope

curves of carbonates previously assigned to the Fuente del Puma Group (now Parque UTE Group) and Carapé Group (sensu Bossi et al., 2007) are consistent with a Mesoproterozoic depositional age (Chiglino et al., 2008; Bossi et al., 2008); and (e) U-Pb detrital zircon ages of sandstones of the Carapé Group are consistent with a Mesoproterozoic age for part of this unit (Basei et al., 2008). It is worth noting that an extensional tectonic event is represented by the 1588 ± 11 Ma mafic dike swarm of Tandilia (U-Pb on baddeleyite: Teixeira et al., 2002).

Independent evidence of one or more Mesoproterozoic tectonic events in the RPC is provided by U-Pb ages of detrital zircons from Neoproterozoic sandstone units deposited on the RPC in Uruguay and Argentina (Figs. 4.6.1, 4.6.3). The Piedras de Afilar Formation in Uruguay shows a dominant Mesoproterozoic zircon population (62% of total) with four distinct peaks at 1.49, 1.35, 1.24 and 1.01 Ga (Gaucher et al., 2007c, 2008b; Fig. 4.6.3). In the Cerro Largo Formation (Sierras Bayas Group), Mesoproterozoic zircons represent 15% of the population and define distinct peaks at 1.55, 1.23 and 1.05 Ga (Fig. 4.6.3), roughly coincident with the Piedras de Afilar data. Rapela et al. (2007) report detrital zircon ages for one sample of the Villa Mónica Formation in the Barker area, with a subordinate Mesoproterozoic peak centered at 1135 Ma. Finally, Blanco et al. (2007, in press) report that 17% of detrital zircons from the Yermal Formation (Arroyo del Soldado Group) yielded U-Pb ages between 1063 and 1009 Ma (Fig. 4.6.3). This shows that extensive areas of Mesoproterozoic rocks were exposed in the RPC and acted as the source of detritus in the late Neoproterozoic. Gaucher et al. (2007c, 2008b) showed that the abundance of Mesoproterozoic detrital zircons in Neoproterozoic sandstones increases toward the west of the RPC. A proto-Andean, Mesoproterozoic belt fringing the western RPC has been suggested as the source of the Mesoproterozoic zircons (Gaucher et al., 2007c, 2008b), which may represent the southward extension of Mesoproterozoic units of the Amazonian Craton (Sunsás Orogenic Cycle, Fig. 4.6.4).

It is significant that a number of Mesoproterozoic units and blocks fringe the Río de la Plata and Amazonian cratons to the west. Chew et al. (2007) suggest that the Arequipa-Antofalla Block accreted to Amazonia during the Grenvillian-Sunsás Orogeny. In the Sierra de Pie de Palo of the Cuyania Terrane (western Argentina), a Grenvillian (1.1-1.0 Ga) metamorphic basement known as Pie de Palo Complex is exposed. Recent data on

the timing of tectonic events in the Pie de Palo range by Mulcahy et al. (2007) support the idea of this unit being a para-autochthonous block of Gondwanan affinity.

Thus, the available evidence strongly suggests that: (a) the RPC was part of the tectonic collage ultimately leading to the accretion of Rodinia, and (b) the Amazonian Craton and the RPC were probably attached in the late Mesoproterozoic (Gaucher et al., 2003), representing the core of proto-South America. They were fringed to the west by a Grenvillian (Sunsás), proto-Andean orogenic belt. Remnants of Mesoproterozoic volcanosedimentary successions occur also at the eastern boundary of the RPC (Parque UTE and Carapé groups: Bossi et al., 2007, 2008).

4.6.2. LATEST CRYOGENIAN-EARLY EDIACARAN RIFTING

Evidence of Neoproterozoic rifting is compelling in the Nico Pérez Terrane in Uruguay. Mafic dyke swarms, anorogenic granitoids and rift-related sedimentary successions occur there, and are broadly constrained between 635 and 580 Ma. This is significantly younger than Neoproterozoic rifting in the Kalahari and Congo cratons (see Chapters 5.2 and 8), but older than the opening of the Iapetus Ocean between Laurentia and Gondwana.

4.6.2.1. Nico Pérez mafic dyke swarm

A mafic dyke swarm occurs in the central Nico Pérez Terrane (Fig. 4.6.1). Dykes strike roughly EW and range in composition from basalts to tholeiitic andesites. They intrude a mainly Archean basement, but also late Palaeoproterozoic rapakivi granites and Mesoproterozoic peraluminous granites (Fig. 4.6.1). A detailed geochemical study has been presented by Mazzucchelli et al. (1995) and Rivalenti et al. (1995), and a recent compilation of all available data is due to Bossi and Schipilov (2007). The age of the dykes is constrained by: (a) a K-Ar age of 581 ± 13 Ma on biotite at the contact between the dykes and the country rocks, and (b) a poor Rb-Sr isochron of 665 ± 203 Ma (Rivalenti et al., 1995).

In either case, the dykes record rifting in the RPC around 600 Ma, thus immediately prior to deposition of shelf sediments of the Arroyo del Soldado and upper Sierras Bayas groups.

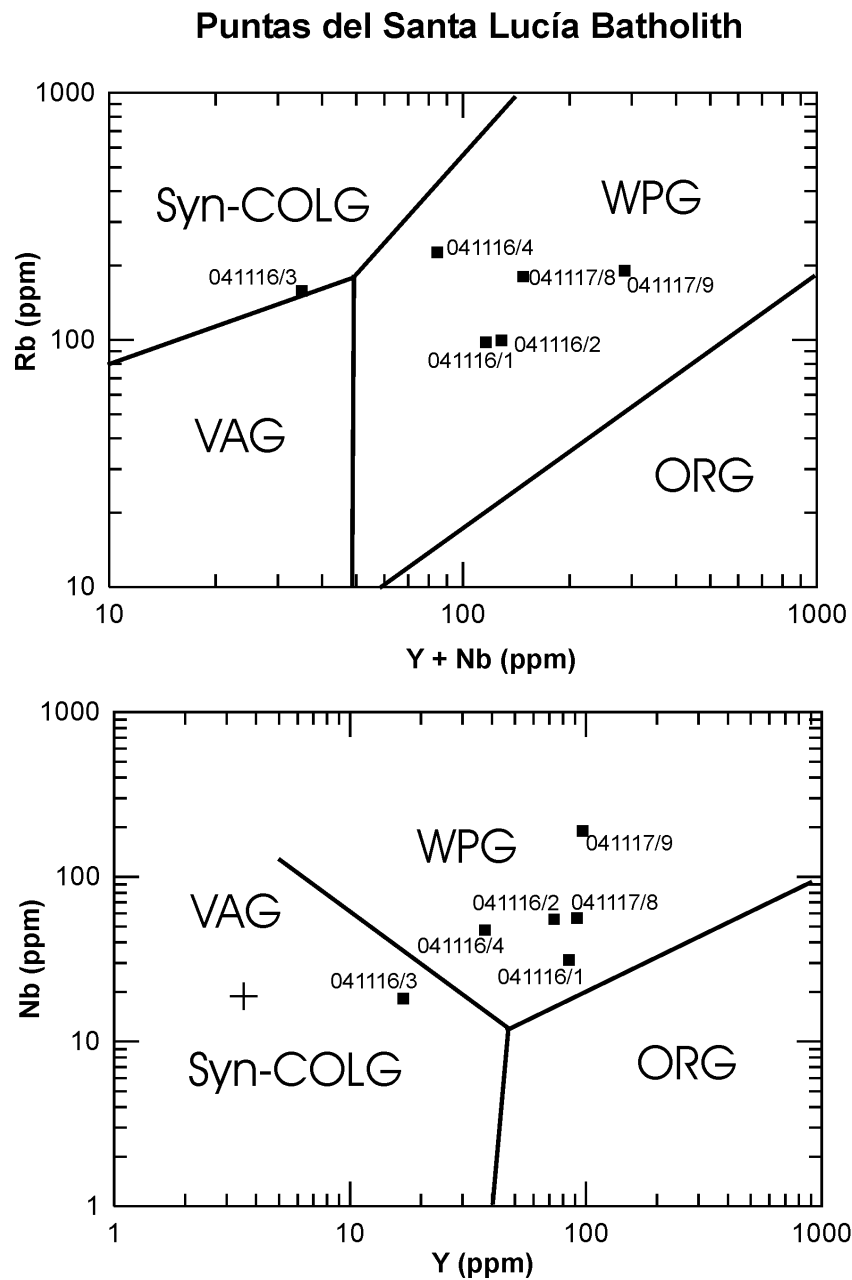


Fig. 4.6.2. Composition of six granitoid samples (Table 1) from the Puntas del Santa Lucía Batholith plotted in the diagrams of Pearce et al. (1984). WPG: Within-plate granites; syn-COLG: Syn-collisional granites; VAG: Volcanic arc granites; ORG: Ocean ridge granites. Note clustering of samples in the within-plate field.

4.6.2.2. Early Ediacaran within-plate basalts

Porphyritic metabasalts occur in the Fuente del Puma Group sensu Bossi et al. (1998), for which Mallmann et al. (2007) report an U-Pb SHRIMP zircon crystallisation age of 590 ± 2 Ma. This age is within error of the above mentioned K-Ar age of the Nico Pérez Dyke Swarm. The geochemical signature of the dated basalts falls between the within-

Sample	041116/1	041116/2	041116/3	041116/4	041117/8	041117/9
SiO₂	67,13	71,31	73,52	78,55	77,92	78,50
Al₂O₃	12,61	12,09	12,08	10,80	12,64	11,54
MgO	0,10	0,06	0,82	0,20	0,02	0,04
TiO₂	0,23	0,25	0,58	0,09	0,09	0,07
Fe₂O₃	4,90	4,60	3,70	1,28	1,56	1,04
CaO	1,57	1,11	1,55	0,53	0,04	0,41
Na₂O	3,79	3,57	2,99	2,96	4,12	3,53
K₂O	5,18	5,12	4,08	4,28	4,73	4,29
MnO	0,10	0,08	0,04	0,01	0,06	0,00
P₂O₅	0,06	0,02	0,21	0,03	0,01	0,01
<i>Sum %</i>	95,68	98,21	99,57	98,72	101,19	99,42
Ba	447,50	105,60	932,20	273,70	558,30	93,30
Nb	31,30	55,20	18,20	47,40	56,20	189,70
Pb	11,70	15,70	19,20	23,50	8,60	24,60
Rb	97,70	99,40	157,60	226,00	180,40	190,40
Sr	64,00	19,20	245,50	68,80	202,40	17,80
Th	1,70	28,00	13,80	36,90	19,00	26,50
Y	84,80	73,30	16,80	37,30	91,80	97,00
Zr	651,00	489,90	320,10	123,60	242,80	212,50
Ce	60,30	417,60	88,30	98,40	143,80	65,50
Co	-0,90	-0,60	3,20	-0,10	-2,40	0,50
Cr	17,50	17,50	18,60	15,80	15,70	10,70
Cu	3,20	2,70	3,60	2,10	1,60	1,40
Ga	25,20	31,30	21,10	25,60	19,40	31,90
Hf	17,90	16,30	9,90	4,50	11,50	10,40
La	158,10	599,60	39,20	48,40	104,20	20,20
S	-352,90	-1671,80	42,10	22,00	-127,20	112,60
Zn	95,60	155,00	74,20	54,10	69,90	21,20
Ni	6,20	7,10	7,00	6,90	8,60	9,30
V	8,50	7,60	32,90	5,50	3,50	5,70

Table 4.6.1: X-ray fluorescence analyses of samples from the Puntas del Santa Lucía Batholith plotted in Fig. 4.6.2. Oxide concentration in weight %, trace elements in ppm.

plate and E-MORB fields (Mallmann et al., 2007), thus consistent with a rifting event at 590 Ma. It is worth noting that the Fuente del Puma Group, or the roughly equivalent Lavalleja Metamorphic Complex, contains according to Mallmann et al. (2007) at least two different groups of basaltic rocks, as deduced from their different geochemistry, Sm-Nd T_{dm} model ages and Sr isotopic characteristics. As noted in Chapter 4.1, the unit undoubtedly includes a Mesoproterozoic volcanosedimentary succession (Parque UTE Group: Bossi et al., 2008), tectonically intercalated with the Neoproterozoic basalts.

4.6.2.3. *A-Type granitic magmatism*

The Puntas del Santa Lucía Batholith is a relatively large intrusion (Fig. 4.6.1) including granodiorites, granites, syenites and rhyolites (A. Toselli, pers. comm., 2007). Hartmann et al. (2002a) report an U-Pb SHRIMP zircon age of 633 ± 10 Ma for this batholith. Samples of the different facies of the intrusion plot in the within-plate (anorogenic) field in the diagrams of Pearce et al. (1984; Fig. 4.6.2; Table 4.6.1), probably representing the earliest record of extensional Neoproterozoic magmatism in the RPC (Blanco and Gaucher, 2005).

4.6.2.4. *Las Ventanas Formation: early Ediacaran rift deposits*

An extensional geotectonic setting was first suggested by Bossi and Navarro (1991) and Masquelín and Sánchez Bettucci (1993) for the possibly correlative Las Ventanas and Playa Hermosa formations in Uruguay (see Chapter 4.2). The latter authors mention evidences of syndepositional extensional tectonics. Pazos et al. (2003) also postulate the interaction of extensional faulting and glacial processes for the Playa Hermosa Formation.

A rift setting was postulated by Blanco and Gaucher (2005) for Las Ventanas Formation. In favour of this hypothesis are:

- (a) the bimodal, synsedimentary volcanism, with basalts at the base (La Rinconada Member) and rhyolites at the top (Quebrada de Viera Member) of the unit (Fig. 4.2.4),
- (b) strong palaeorelief, as shown by high-energy, immature conglomerate deposits,
- (c) great thickness of alluvial fan conglomerates (Fig. 4.2.4), and
- (d) the progressive evolution from continental to open marine environments, suggesting a taphrogenic basin setting (Miall, 2000).

The onset of this rifting event is poorly constrained by one K-Ar age of 615 ± 30 Ma for basalts of the La Rinconada Member (Sánchez Bettucci and Linares, 1996) and datings of a number of magmatic units probably associated with this extensional event. The basalts tectonically intercalated in the Fuente del Puma Group and dated U-Pb SHRIMP at 590 ± 2 Ma by Mallmann et al. (2007, see above) probably represent the same magmatic event. If this is accepted, the age of the basal Las Ventanas Formation is ca.

590 Ma, in accordance with biostratigraphic data (Gaucher et al., 2008a; see Chapter 4.2) and field relationships with other volcanic units (Pazos et al., 2008). On the other hand, mafic dykes crosscut conglomerates of the Las Ventanas Formation at the Mina Verdún quarry, nearby Minas (Blanco and Gaucher, 2005; Fig. 4.6.1). These dykes yielded a minimum K-Ar age of 485 ± 13 Ma (González et al., 2004), and could be part of the Nico Pérez Dyke Swarm (581 ± 13 Ma, see above). Thus, a high sedimentation rate could be responsible for the accumulation of more than 3.5 km of conglomerates in ca. 10 Myr, a typical feature of rift basins (Miall, 2000).

The Las Ventanas rifting was clearly influenced by pre-existing lineaments, notably the Sarandí del Yí Shear Zone. This fact explains the roughly NS trend of the inferred rift axis (Fig. 4.6.1), and the occurrence of a thicker basalt pile near the shear zone (Blanco and Gaucher, 2005).

In the Tandilia area only little evidences of rifting were preserved. A thin (<10 m) diamictite level (“Colombo Diamictite”) between the Villa Mónica and Cerro Largo formations shows evidence of slumping, gravity flows and syn-depositional, extensional tectonics (Figs. 4.2.1, 4.5.1D-F). The stratigraphic position of these mass-flow deposits, suggest correlation with the Las Ventanas Formation in Uruguay.

Lower Ediacaran rifting in the RPC heralded the opening of the Brazilides Ocean and the development of the Sierras Bayas-Arroyo del Soldado-Corumbá shelf (Gaucher et al., 2003, 2005a), representing the drift phase of the platform (Fig. 4.6.4).

4.6.2.5. Camaquã Basin and São Gabriel Arc: part of the Río de la Plata palaeocontinent ?

The only area in Brazil currently accepted as part of the RPC is the Taquarembó Block, close to the Uruguayan border (Fig. 4.1.2). The roughly NW-trending Ibaré Shear Zone in southernmost Brazil could represent the boundary of the RPC there. To the north of this shear zone, a juvenile magmatic arc (São Gabriel Orogen) occurs, represented by the Vacacaí and Cambaí groups. A juvenile Neoproterozoic source for the Yermal Formation is suggested in NE-Uruguay (Isla Cristalina de Rivera) by T_{DM} model ages of 1.2 Ga and ϵ_{Nd} between -0.6 and -1.1, which are in contrast with T_{DM}

ages of around 2.1 Ga and ϵ_{Nd} as low as -17 for outcrops of the same units in the south (Blanco et al., 2007; in press). Blanco et al. (in press) suggested that the Vacacaí and Cambaí groups acted as main sources, both units showing positive ϵ_{Nd} and T_{DM} model ages between 0.9 and 1.2 Ga (Gastal et al., 2005). Paleocurrent directions in the Yerbal Formation, from the northeast (Gaucher, 2000), are in agreement with this interpretation. Thus, according to these data, the São Gabriel Arc was accreted to the RPC before deposition of the Yerbal Formation, and according to a number of authors at ca. 700 Ma (Babinski et al., 1996; Saalman et al., 2005).

The Camaquã Basin records episodic sedimentation between ca. 630 Ma and the Ordovician and is developed mainly to the NE of the Ibaré Shear Zone (Paim et al., 2000; Borba et al., 2008). It represents a series of different, stacked basins that share the same depositional locus (Paim et al., 2000). Andesitic and rhyolitic volcanism has been dated U-Pb SHRIMP at 630±3 Ma (pyroclastic clasts, Maricá Formation: Borba et al., 2008), 592±5 Ma (andesites of the Hilário Formation: Remus et al., in Paim et al., 2000), and 573±18 Ma or 549±5 Ma for rhyolites of the Acampamento Velho Formation (Paim and Fonseca, 2004 and Sommer et al., in Borba et al., 2008). Granitic magmatism in the area is constrained by U-Pb SHRIMP zircon ages between 592 and 541 Ma (Paim et al., 2000 and references therein). Whereas the 630 and 592 volcanic units match well the extensional magmatism recorded in the Nico Pérez Terrane (see above), voluminous granitic magmatism and bimodal volcanics between 580 and 541 Ma in the Camaquã Basin contrast with tectonic quiescence in the RPC. This fundamental difference militates against the Camaquã Basin being associated to the RPC between 580 and 541 Ma. One possibility would be that the block containing the basin rifted from the RPC between 630 and 592, and evolved separately in the late Ediacaran. During the Lower Cambrian Búzios Orogeny, the basin was probably once again accreted against the RPC. However, this is at odds with the results of Blanco et al. (2007) showing that the juvenile São Gabriel Arc probably acted as a source of the Yerbal Formation in northern Uruguay. It is worth noting that the nature of the contact between the São Gabriel Orogen and its sedimentary cover may be tectonic, as suggested by the latter not being sourced in the São Gabriel juvenile rocks (Borba et al., 2008 and references therein). This might explain the conflicting results regarding the relationship between the Camaquã Basin and the RPC.

4.6.3. EDIACARAN DRIFT PHASE AND PASSIVE MARGIN

Extensive deposits of the Arroyo del Soldado and Sierras Bayas groups and correlative units (see Chapters 4.1-4.2) represent the drift phase of the platform. This passive margin probably reached as far north as Corumbá (Gaucher et al., 2003; Fig. 4.6.4), where the Corumbá Group records essentially the same evolution.

An Atlantic-type continental shelf was interpreted as the geotectonic setting for deposition of the mentioned units (Poiré, 1987; Gaucher, 2000; Gaucher et al., 2003, 2005a; Bossi and Gaucher, 2004), on the bases of: (a) sandstone petrography, which are predominantly mature quartz-arenites and subarkoses; (b) lateral persistence of lithofacies over hundreds of kilometers and markedly tabular geometry; (c) absence of synsedimentary volcanism, (d) occurrence of thick (up to 1000 m) carbonates and chemical deposits (BIF, chert), denoting low palaeorelief and little siliciclastic input, and (e) palaeocurrent pattern.

More recently, detrital zircon ages and Sm-Nd model ages of sediments confirmed the geotectonic setting postulated for the Arroyo del Soldado and Sierras Bayas groups and Piedras de Afilas Formation. Sm-Nd analyses of sandstones and shales from the Arroyo del Soldado Group yield T_{DM} model ages between 1.6 and 2.9 Ga and ϵ_{Nd} values of approximately -10 (Blanco et al., 2007; Mallmann et al., 2007), showing provenance from old continental crust. U-Pb LA-ICP MS ages of detrital zircons from the mentioned units show (Figs. 4.6.3) that the main source areas were Archean, Palaeoproterozoic and Mesoproterozoic rocks (Gaucher et al., 2008b; Blanco et al., 2007, in press). Neoproterozoic zircons are rare (Fig. 4.6.3): one grain from a sandstone sample of the Yermal Formation yielded an age of 664 ± 14 Ma (0.8% of concordant grains), another grain from the Cerros San Francisco Formation was dated at 605 ± 53 Ma (0.8% of concordant grains) and 28% of zircons from a sample of the Barriga Negra Formation define an age population of 631 ± 12 to 566 ± 8 Ma. These ages match the rift-related magmatic units mentioned above, such as the Puntas del Santa Lucía Batholith. The higher abundance of Neoproterozoic zircons in the Barriga Negra Formation can be explained by deeper erosion of basement units compared to the other studied formations, which allowed exhumation of deep-seated anorogenic intrusions. A source area dominated by Archean and Palaeoproterozoic rocks is corroborated by T_{DM} model ages of the Barriga Negra Formation between 1.6 and 2.7 Ga (Blanco et al., in press).

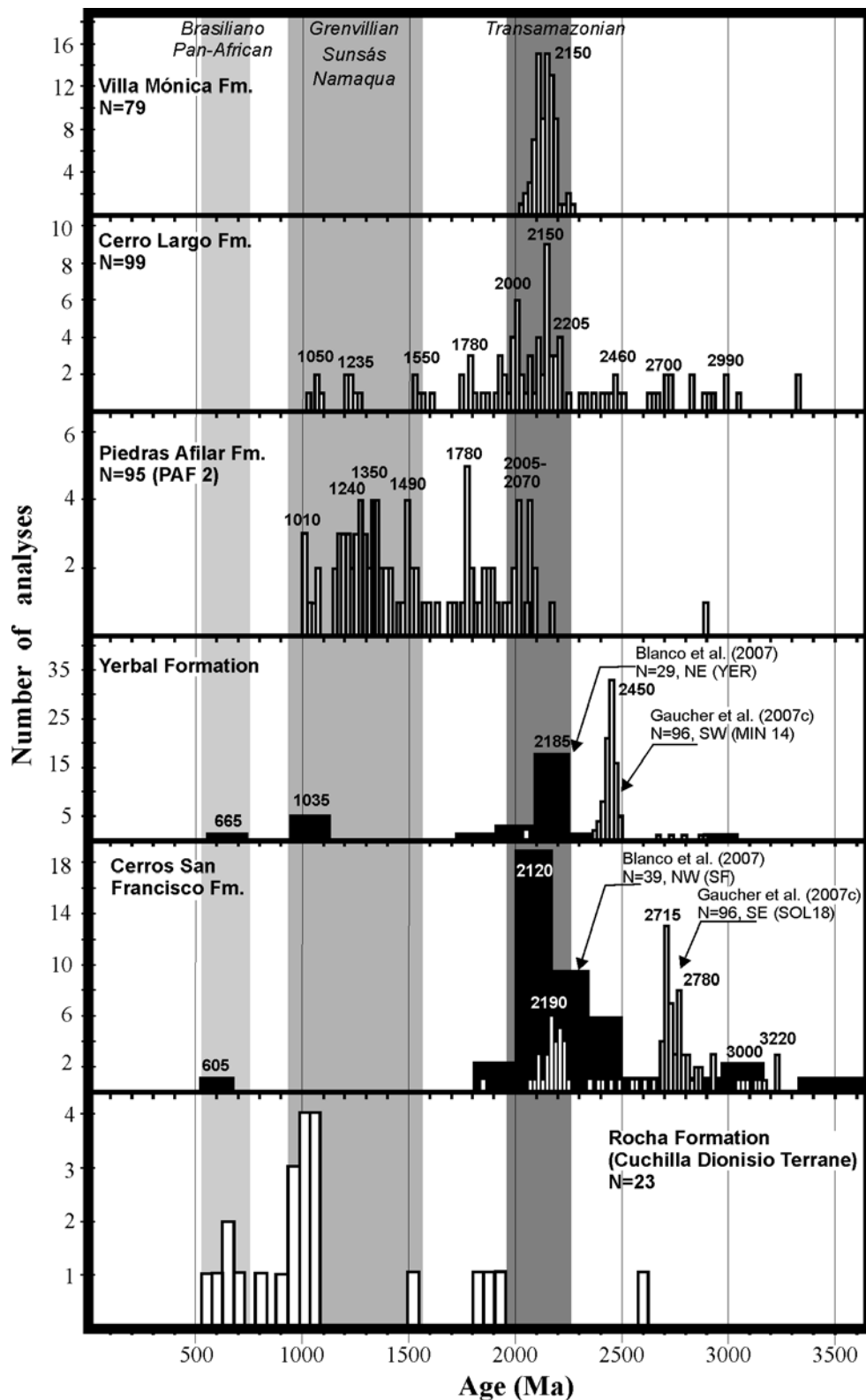


Fig. 4.6.3. Detrital zircon age spectra for Neoproterozoic-Lower Cambrian sandstones of the RPC and Cuchilla Dionsio Terrane. Location of samples in Uruguay: see Fig. 4.6.1. Sample from the Rocha Formation from Basei et al. (2005), all other samples from Gaucher et al. (2008b) otherwise indicated (Blanco et al., in press). Age of the main peaks (in Ma) and range of important orogenic events is indicated. Note (a) distinctively different spectra for the Rocha Formation compared to the samples from the RPC; (b) paucity of Neoproterozoic zircons in the RPC samples; and (c) significant Mesoproterozoic populations.

Due to the fact that conglomerates of the Barriga Negra Formation wedge out toward the basin and are absent in deep-water sections, Gaucher (2000) suggested that a large-amplitude sea-level drop was responsible for deposition of the unit. Shallow areas of the shelf were exposed and reworked, allowing rivers to cut deeply through the stratigraphy of the Arroyo del Soldado Group and its basement. This explanation is consistent with similar conglomerate levels occurring at the same stratigraphic position in the Sierras Bayas, Corumbá, Nama and Cango Caves Group (see Chapter 4.5).

Thus, available data are consistent with provenance of the Arroyo del Soldado and Sierras Bayas groups from the peneplainised RPC, and effectively rule out an input from a Neoproterozoic orogenic belt (Fig. 4.6.4), as in the case of a foreland basin (e.g. Basei et al., 2000; Pecoits et al., 2008). The most important controls on deposition were climate and associated sea-level fluctuations.

4.6.4. EVOLUTION OF THE WESTERN BOUNDARY OF THE RÍO DE LA PLATA PALAEOCONTINENT

The western boundary of the RPC has remained controversial due to the extensive Phanerozoic cover of the Paraná Basin. Almeida (2004) placed the western boundary at the Eastern Sierras Pampeanas, which has been confirmed by Rapela et al. (2007) by U-Pb SHRIMP dating of basement samples from boreholes drilled through the Paraná Basin in Argentina. All studies regarding the western boundary of the RPC have systematically considered Palaeoproterozoic (Transamazonian) ages between 2.2 and 2.0 Ga as typical of the craton. However, as demonstrated by detrital zircon dating of Neoproterozoic sedimentary covers of the RPC (Gaucher et al., 2008b), Mesoproterozoic belts are also an integral part of the craton (Figs. 4.6.3-4). This means that the RPC may extend farther to the west than currently accepted, probably including a Grenvillian belt surrounding the Palaeoproterozoic-Archean core (see above).

The main events linked to the evolution of the western margin of the RPC have been outlined by Rapela et al. (1998; Fig. 4.6.5). The Puncoviscana Formation, a sedimentary succession of latest Ediacaran to Lower Cambrian age (Aceñolaza, 2003, 2004 and references therein, see Chapter 6), fringes the RPC to the west (Fig. 4.6.4). This unit was deposited in a narrow ocean between the RPC and the Pampia Terrane (see Chapter

6), that probably opened as a result of late Ediacaran rifting between the RPC and Laurentia (Grunow et al., 1996). A prominent angular unconformity (Tilcaric unconformity) between the Puncoviscana Formation and the Middle Cambrian Mesón Group shows that closure of the Puncoviscana Basin took place in the Lower Cambrian (Aceñolaza, 2003). This short-lived orogenic event, known as the Pampean Orogeny, is also characterised by calc-alkaline granite intrusions between 530 ± 4 and 523 ± 2 Ma (U-Pb on zircon: Rapela et al., 1998). Mulcahy et al. (2007) suggest that following the accretion of the Pampia Terrane to the RPC at 515 ± 2 Ma (Ar-Ar on mylonites from a thrust plane), eastward subduction stepped westward, establishing a new convergent margin along the Famatina arc, active until the Middle Ordovician. It is worth noting that while in the eastern RPC extensional tectonics led to the intrusion of a syenitic anorogenic complex, at the western boundary of the craton granulite-facies metamorphism was underway (Rapela et al., 1998).

The paucity or absence of Palaeoproterozoic detrital zircons in late Ediacaran-Cambrian metasediments from the Eastern Sierras Pampeanas has been suggested as proof of these rocks being sourced in a pre-Pampean, Neoproterozoic arc and not in the RPC (Escayola et al., 2007). However, as has been recently shown by detrital zircon data from autochthonous Neoproterozoic covers in the eastern RPC (Blanco et al., 2007; Gaucher et al., 2008b), Palaeoproterozoic detritus becomes progressively less important toward the west of the RPC, and Mesoproterozoic zircons dominate at the western boundary (Schwartz and Gromet, 2004). Therefore, the logical explanation is that the metasedimentary rocks of the Eastern Sierras Pampeanas were sourced mainly in the proto-Andean, Grenvillian belt fringing the RPC to the west (Fig. 4.6.4).

Detrital zircon ages of Ordovician-Silurian sandstones from the Balcarce Formation show Cambrian to Neoproterozoic, Mesoproterozoic and late Palaeoproterozoic peaks (Rapela et al., 2007). In this case, the paucity of Transamazonian ages is easily explained by the docking in the Lower Cambrian of the Cuchilla Dionsio-Pelotas Terrane, which provided Cambrian-Neoproterozoic and late Mesoproterozoic detritus typical of that block (Basei et al., 2005). This is confirmed by west-directed palaeocurrents in the Balcarce Formation (Poiré, 1987; Rapela et al., 2007). The Cuchilla Dionsio Terrane was probably an important mountain range undergoing active erosion in Ordovician times, thus overshadowing the sediment input of the low-relief, Transamazonian units of the RPC.

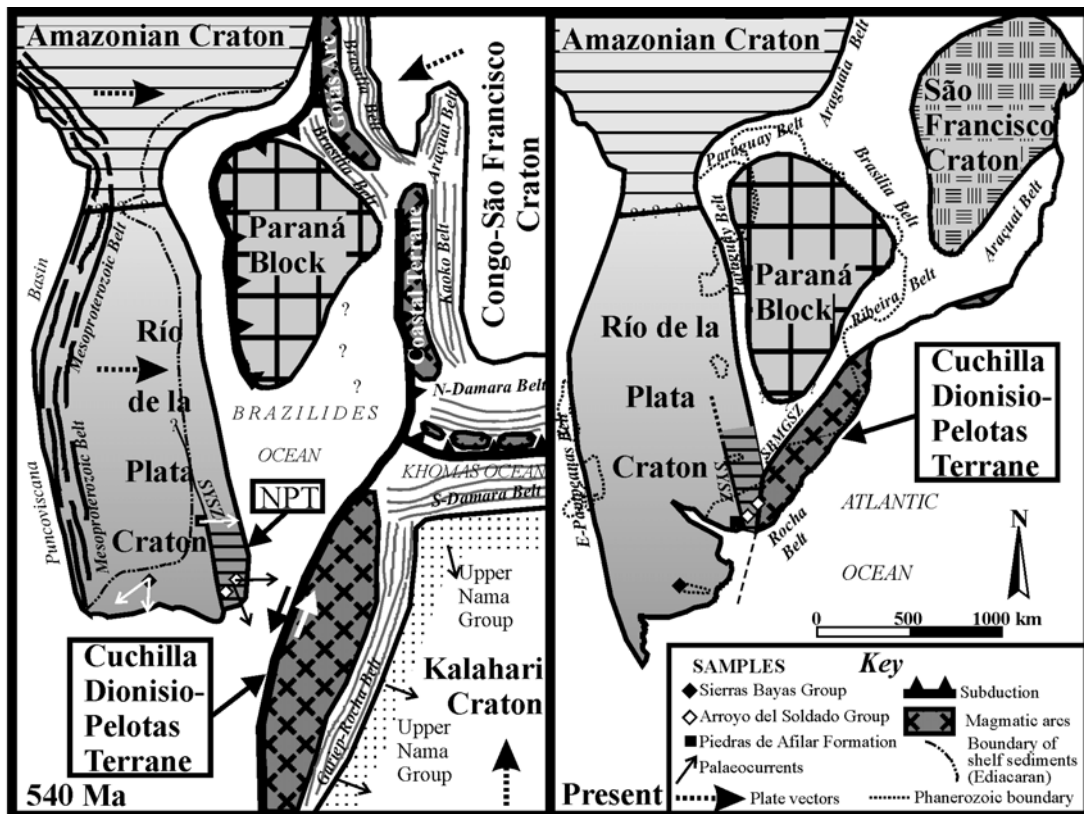


Fig. 4.6.4. Palaeogeographic reconstruction of the RPC in the earliest Cambrian and present configuration of cratons and belts. Note location of detrital zircon samples of Neoproterozoic-Lower Cambrian sedimentary successions discussed in the text, showing corresponding palaeocurrent pattern. Modified from Gaucher et al. (2008b) and sources therein.

4.6.5. LOWER CAMBRIAN BUZIOS OROGENY

Bossi and Gaucher (2004) proposed that the Cuchilla Dionisio-Pelotas Terrane is an allochthonous block juxtaposed to the RPC by means of tangential tectonics (see Chapter 8). Although it represents the roots of a magmatic arc active between 650-550 Ma, as already envisaged by Ramos (1988), its present position is out of context and does not reflect the origin of the magmatic arc. Detrital zircon ages of the Ediacaran Rocha Formation (Fig. 4.6.3), sourced in the Cuchilla Dionisio Terrane, show prominent peaks at 1.0 Ga and between 764 to 596 Ma (31% of all zircons; Basei et al., 2005), typical ages of the basement and granitic intrusions of the terrane. This is in sharp contrast to the above mentioned detrital zircon ages for coeval units deposited in the RPC, showing that the RPC and Cuchilla Dionisio Terrane were separated by a large (oceanic) basin in the Ediacaran. Given that the youngest units of the Arroyo del Soldado Group were deposited between 542 and 535 Ma (Sprechmann et al., 2004; Gaucher et al., 2007b; see Chapter 4.4), the tangential collision took place no earlier

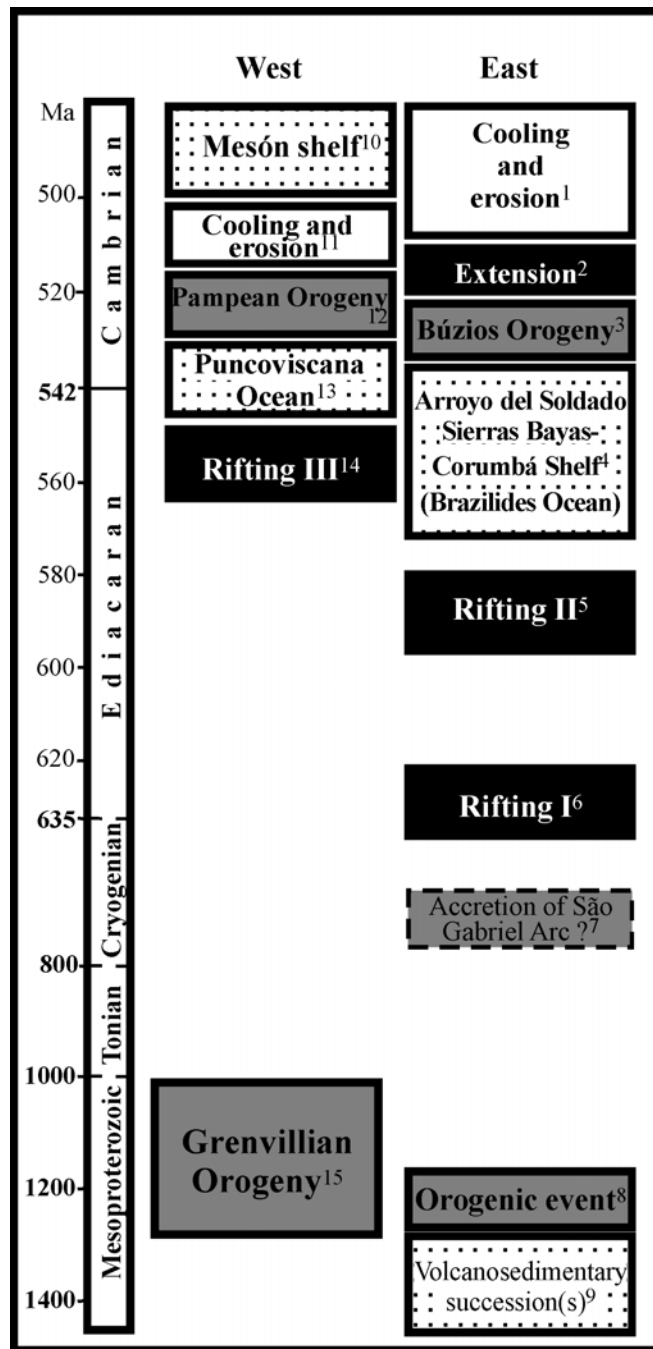


Fig. 4.6.5. Palaeogeographic evolution of the RPC in time, between the Mesoproterozoic and Cambrian. Sources of data: 1: Cingolani et al. (1990, in Gaucher, 2000); 2: Bossi et al. (1993b); 3: Bossi and Gaucher (2004) and references therein; 4: Gaucher (2000); Gaucher et al. (2003, 2004b, 2008b); 5: Blanco and Gaucher (2005), Mallmann et al. (2007); 6: Hartmann et al. (2002a); 7: Blanco et al. (in press), Babinski et al. (1996); 8: Bossi et al. (1998); Teixeira et al. (1999); Bossi and Navarro (2001); 9: Oyhantçabal et al. (2005); Gaucher et al. (2006); 10: Aceñolaza (2003); 11: Mulcahy et al. (2007); 12: Rapela et al. (1998); Mulcahy et al. (2007); 13: Aceñolaza (2003, 2004) and references therein; 14: Grunow et al. (1996); 15: Gaucher et al. (2008b, and references therein).

than the Lower Cambrian, possibly around 530 Ma (Fig. 4.6.4). This is in accordance with Rb-Sr and U-Pb zircon ages of associated granites of 532 ± 11 Ma (Kawashita et al., 1999a) and 530 ± 14 Ma (Oyhantçabal et al., 2005), K-Ar ages of recrystallised shales between 532 ± 16 and 492 ± 14 Ma (Cingolani et al., 1990 in Gaucher, 2000), and

remagnetisation of the Arroyo del Soldado Group around 525 Ma (Rapalini and Sánchez Bettucci, 2008). Typical Brasiliano compressional tectonics, such as in the Brasília and Araçuaí belts, are distinctively older than 530 Ma. Therefore, we assign the tangential collision of the RPC with the Cuchilla Dionisio Terrane to the Búzios Orogeny (Schmitt et al., 2004, 2008), which represents the last event leading to Gondwana amalgamation. The Búzios Orogeny is the result of the collision of the block conformed by the Cuchilla Dionisio-Pelotas Terrane (*Arachania*, see Chapter 8) and the Kalahari Craton with the Río de la Plata, Amazonian, Paranapanema and Congo-Sao Francisco cratons. A reconstruction of this complex tectonic event was published by Gaucher et al. (2008b), and is shown in Fig. 4.6.4.

4.6.6. CAMBRIAN ANOROGENIC MAGMATISM

The onset of post-orogenic relaxation and extensional tectonics is marked by the intrusion of the Sierra de Animas Formation in Uruguay. This unit comprises syenites, trachytes, rhyolites and basalts, yielding a Rb-Sr isochronic age of 520 ± 5 Ma (Bossi et al., 1993a). The Sierra de Animas Formation shows a subalkaline to alkaline affinity (Bossi et al., 1993a) and is largely undeformed. It is worth noting that syenites were emplaced along the Sarandí del Yí Shear Zone. The undeformed nature of the Sierra de Animas Formation shows that compressive tectonics likely finished before 520 Ma. The reactivation of the Sarandí del Yí lineament is constrained by Rb-Sr ages of 530 ± 10 Ma on intensely deformed granites emplaced in the shear zone (Umpierre and Halpern, 1971; see Chapter 4.1), and by the age of the upper Arroyo del Soldado Group (542-535 Ma), which is affected by the fault (Gaucher et al., 2007b). A northward displacement of several hundreds of kilometers (Fig. 4.6.4) is indicated for the Nico Pérez Terrane by detrital zircon age spectra of several Ediacaran units in the RPC (Gaucher et al., 2008b).

4.6.7. PALAEOMAGNETIC CONSTRAINTS

A recent compilation of Neoproterozoic palaeomagnetic data by Trinidade and Macouin (2007) lists only two palaeopoles for the RPC that fulfill the necessary quality criteria and represent primary magnetisation: one from the Sierras Bayas Group (“La Tinta Formation” pole: Valencio et al., 1980) and one from the Campo Alegre Group. The latter unit is not associated to the RPC but to the Luis Alves block, a microplate

possibly accreted in the Cambrian (D'Agrella-Filho and Pacca, 1988; see Chapter 7.2). The age of the palaeopole from the Sierras Bayas Group is poorly constrained, given the uncertainties regarding the age of the Villa Mónica Formation. Furthermore, in a recent, comprehensive work, Rapalini and Sánchez Bettucci (2008) demonstrate the non-primary nature of the “La Tinta” palaeopole. According to these authors, a number of Neoproterozoic units, including the Arroyo del Soldado Group, are remagnetised, and cannot be used for palaeogeographic reconstruction. Other published palaeopoles lack the necessary tests to ascertain their primary nature or their age is poorly constrained (Sánchez Bettucci and Rappalini, 2002; Rapalini and Sánchez Bettucci, 2008). Therefore, the paucity of palaeomagnetic data for the RPC does not allow yet to carry out palaeogeographic reconstructions independently from the geological data.

4.6.8. CONCLUSIONS

Several lines of evidence suggest that the RPC was part of Rodinia, and that a proto-Andean Grenvillian belt is an integral part of the craton (Figs. 4.6.4-4.6.5). The RPC rifted off Rodinia between 635 and 580 Ma. Evidences of rifting include mafic dyke swarms, anorogenic granites, within-plate basalts and rift sedimentary deposits. Coeval volcanism in the Camaquã Basin in southern Brazil suggest that this block may have rifted off the RPC, evolving separately until its accretion in the Lower Cambrian, although other data suggest that the São Gabriel juvenile arc was accreted to the RPC at ca. 700 Ma.

Following rifting, an extensive passive margin was established on the eastern margin of the RPC, including deposits of the Sierras Bayas, Arroyo del Soldado and Corumbá groups. Shelf deposits were deformed and weakly metamorphosed due to tangential collision of the Cuchilla Dionisio-Pelotas Terrane (Arachania) with the RPC. This collision, assigned here to the Búzios Orogeny, likely took place around 530 Ma, representing the last event leading to final accretion of Gondwana. Anorogenic, syenitic magmatism dated at 520 Ma marks the end of compressional tectonics in the eastern RPC and the onset of post-orogenic relaxation. Roughly at the same time (530-515 Ma) in the western RPC, the Pampean Orogeny led to deformation and metamorphism (up to granulite facies) of latest Ediacaran-Lower Cambrian siliciclastic successions (Puncoviscana Formation and equivalents; Fig. 4.6.5). Once Pampia was accreted to the

RPC, eastward subduction likely migrated westwards and continued until the Middle Ordovician.

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