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Closure of the Clymene Ocean and formation of West Gondwana in the Cambrian: Evidence from the Sierras Australes of the southernmost Rio de la Plata craton, Argentina

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ABSTRACT

The formation of Gondwana took place across a series of Brasiliano–Pan African suture zones that record late Neoproterozoic to earliest Paleozoic collisions between Precambrian cratons. In South America, an internal suture zone marks the disappearance of the Clymene Ocean that separated the Amazon craton from the São Francisco and Rio de la Plata cratons. New geochronological data from the southern end of this massive collision zone in the Sierras Australes of central-eastern Argentina document Paleoproterozoic crust and suggest an Ediacaran age for the oldest sedimentary rocks. These two observations extend the known limit of the Rio de la Plata craton at least 200 km SW of previous estimates. New data also confirm the occurrence of late Ediacaran to late Cambrian magmatism in the Sierras Australes. The age of these hypabyssal to volcanic rocks corresponds to igneous events in the Pampean belt along the western margin of the Rio de la Plata craton, although the shallow level magma emplacement in the Sierra da Ventana study area contrasts with the deeply exhumed rocks of the Pampean orogeny type locality. These new age data are compared with a broad compilation of geochronological age Clymene collision belts to the north, the Paraguai and Araguaia belts. The close overlap of the timing of orogenesis indicates the age of Clymene ocean closure in its northern reaches. In the south, the Pampean belt was unconfined, allowing continued tectonic activity and crustal accretion throughout the Paleozoic.

1. Introduction

The dispersion of Rodinia and the parallel assembly of Gondwana dominate the rapidly evolving paleogeography of the late Neoproterozoic and early Phanerozoic. In South America, the record of collisions that created the Gondwana supercontinent is contained in Brasiliano mobile belts broadly recognized as late Neoproterozoic to Cambrian in age. Geological and isotopic evidence for Neoproterozoic juvenile material in Brasiliano belts (Pimentel and Fuck, 1992; Babinski et al., 1996; Pedrosa-Soares et al., 1998; Murphy, and Nance, 2003; Hartmann et al., 2011) supports the contention that oceanic plates were subducted along these belts. More recently, the evaluation of geological and paleomagnetic evidence (Cordani et al., 2003; Tohver et al., 2006; Spagnuolo et al., 2011) from individual cratons of South America suggests a division between the western cratons (Amazon *cum* West Africa) and the central Gondwanan cratons (São Francisco *cum* Congo, and Rio de la Plata). The collision between these different elements of west Gondwana would form the Clymene suture zone, which bisects the continent in a predominantly N–S direction.

At present, only loose constraints are established from the sparse paleomagnetic data from the individual cratons in the time leading up to the formation of Gondwana. Therefore, the paleogeography and chronology of the construction of this supercontinent must derive from the geological history of individual mobile belts. In this contribution, we report new U–Pb and $^{40}Ar/^{39}Ar$ geochronological data from the southernmost portion of the Clymene Ocean, the Sierra Australes belt found along the southern margin of the Rio de la Plata craton of east-central Argentina. These new data are integrated with a regional compilation of igneous and metamorphic age constraints from the Pampean, Paraguai, and Araguaia belts, demonstrating that these belts were active in the late Ediacaran to late Cambrian along the entire ca. 3000 km length of the Clymene suture zone (Fig. 1).

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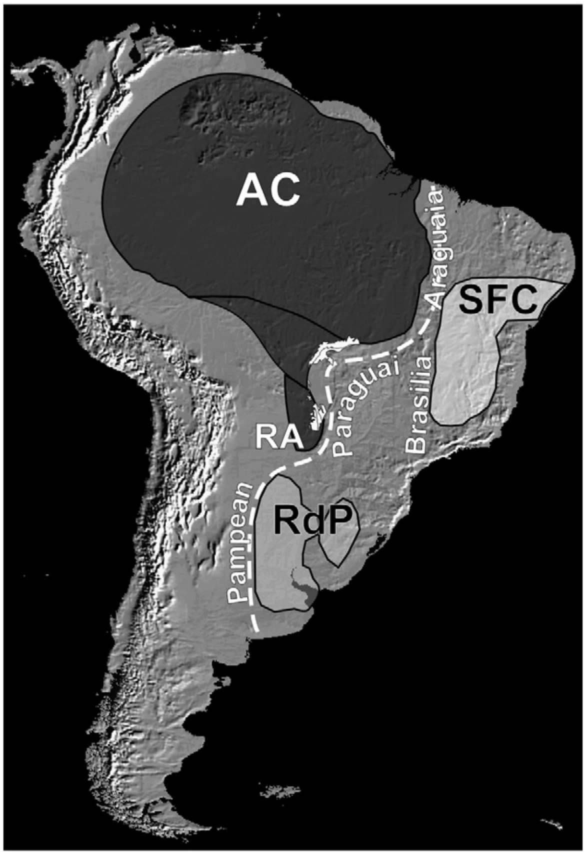

Fig. 1. General location figure depicting the principal cratons of South America, with the correlation of the Clymene belts along the dashed white line.

2. Regional geology

The Rio de la Plata craton is the southernmost Precambrian unit of South America and comprises predominantly juvenile, Paleoproterozoic with small volumes of Archean rock (Hartmann et al., 2002; Pankhurst et al., 2003; Rapela et al., 2007). The chief exposures of the Rio de la Plata basement are found in Uruguay, with limited exposures near Tandilia, in the Buenos Aires province of Argentina, and in southern Brazil. The northern boundary of the craton is not exposed, lying beneath the Paleozoic–Mesozoic sedimentary rocks and Cretaceous flood basalts of the Paraná basin. Ramos et al. (2010) infer that the craton's northwestern limit lies to the south and west of the Rio Apa craton and Paranapanema blocks (Mantovani and Brito Neves, 2005). The westernmost extent of the craton proposed by Ramos (1988) was recently established by U–Pb dating of drill core samples of basement just east of the Sierras de Cordoba in Argentina (Rapela et al., 2007). The eastern boundary of the Rio de la Plata basement in Uruguay is formed by the 530 Ma Sierra Ballena shear zone, which separates autochthonous basement rocks from the Cuchilla de Dionisio terrane, which extends into southern Brazil as part of the Dom Feliciano orogenic belt (Basei et al., 2005). In southern Brazil, the stable portion of the craton is overlain by mixed volcaniclastic and fluvio-lacustrine sedimentary rocks that range in age from Ediacaran to early Cambrian (Almeida et al., 2010; Janikian et al., 2008, this issue). In Uruguay, the late Ediacaran was marked by deposition of carbonate and siliciclastic sedimentary rocks of the Arroyo Soldado Group (Gaucher et al., 2004), with glacial sediments in these rocks considered to reflect the Gaskiers event at ca. 580 Ma (Pazos et al., 2008). Further south in Argentina, the Rio de la Plata craton is overlain by Ediacaran siliciclastic succession in the Tandilia region, traditionally considered to be the southernmost extent of the Rio de la Plata craton (Hartmann et al., 2001; Pankhurst et al., 2003; Rapela et al., 2007). An alternative view was advanced by Ramos (1988), who suggested that the boundary lay at the Colorado River, the putative boundary with Patagonia.

Approximately 200 km to the WSW of the Tandilia region lies the Sierras Australes, where late Paleozoic deformation has given rise to the prominent topography of the Sierra de Ventana region (Fig. 2). The level of exhumation of the Sierras Australes increases to the SW, ranging from undeformed foreland basin deposits in the NE, into a prominent fold-thrust belt in the central domain (Cobbold et al., 1991; von Gosen et al., 1991). The deepest crustal levels are found in more subdued topography of the SW portion of the belt, where the felsic basement rocks and the oldest sedimentary rocks are observed in isolated outcrops. Deformation and exhumation control the geographic distribution of the three major groups of Paleozoic strata found in the region. These packages were divided by Harrington (1947), from bottom to top: the quartz arenites and conglomerates of the Curamalal Group; the siltstones and sandstones of the Ventana Group; and the glacio-marine tillites overlain by organic-rich shales and immature greywackes of the Pillahuinco Group.

Evidence for the age of the Curamalal Group is ambiguous, but has been considered to be Silurian to Ordovician on the basis of trace fossil occurrences (Harrington, 1947). However, regional maps produced by Kilmurray (1975) subdivided the older Curamalal Group into the Los Chilenos Formation, an older group of quartzose conglomerates rocks at the base of the Curamalal Group, as discussed by Llambías and Prozzi (1975). Firmer age constraints for the other two sedimentary packages are provided by Malvino-kaffric fauna that indicate an early Devonian age for the Ventana Group. The *Glossopteris* and *Eurydesma* flora in the upper Pillahuinco Group are Permo-Carboniferous in age, overlying glacial diamictites of the Sauce Grande Formation (Frakes and Crowell, 1969). Further age constraints for the age of the upper Pillahuinco Group are provided by syn-folding magnetizations (Tomezzoli and Vilas, 1999), interpreted as being early-middle Permian according to the Apparent Polar Wander Path of Tommezoli (2008).

The deformed rhyolites and granites that constitute the regional basement of the Sierras Australes form the basis for this study (Fig. 2). The contact between the basement and the overlying sedimentary packages is generally unconformable (Harrington, 1947), with the exception of an intrusive contact reported for aplitic dykes of the Los Chilenos granite in the Cerro Colorado locality (Massabie, et al., 1999). Late Paleozoic deformation has affected the basement rocks, which are marked by brittle–ductile deformation in protomylonitic shear zones (Buggisch, 1987; von Gosen et al., 1990). This deformation has resulted in some isotopic resetting in these rocks, a fact that plagued early geochronological studies of these basement rocks using Rb–Sr and K–Ar methodologies (summarized by Cingolani and Dalla Salda, 2000). More recent work has proceeded by SHRIMP analysis of zircon, coupled with whole rock Sm–Nd data (Rapela et al., 2003). Whole rock geochemistry work was carried out by Grecco et al. (1997).

3. Methods

The U–Pb analyses were conducted using the WA Consortium SHRIMP II ion microprobe housed at Curtin University of Technology. Isotopic ratios were monitored with reference to Sri Lankan gem zircon standard BR266, with an age of 559.1 Ma, and Pb/U ratios in the samples were corrected using the ln(Pb/U)/ln(UO/U) relationship as measured on zircons standard. All uncertainties in the calculated intercept or Concordia ages are reported at 95% confidence limits, unless stated otherwise. The results of the spot analyses are presented in a supplementary appendix.

The $^{40}Ar/^{39}Ar$ analyses were performed at the Western Australian Argon Isotope Facility at Curtin University, operated by a consortium consisting of Curtin University and the University of Western Australia. Samples were loaded into one large well of one 1.9 cm diameter and 0.3 cm depth aluminum disc. This sample well was

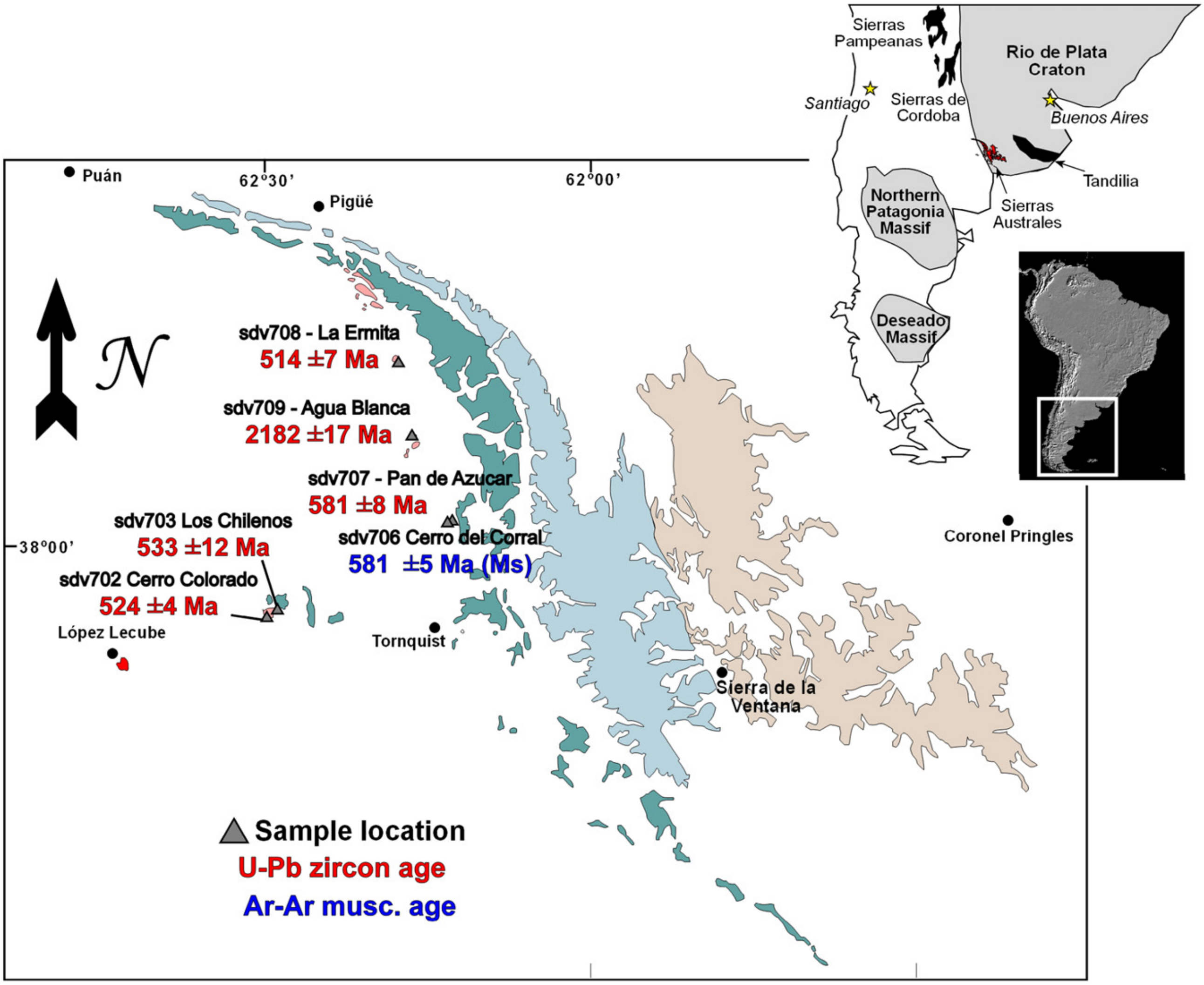

Fig. 2. Location map of the Sierra Australes with geochronological information and sample locations shown. Outcrops of basement are restricted to small occurrences in the SW quadrant of the sigmoidal belt.

surrounded by smaller wells that included Fish Canyon sanidine (FCs) used as a neutron flux monitor for which an age of 28.03 ± 0.08 Ma was adopted (Jourdan and Renne, 2007). The discs were Cd-shielded (to minimize undesirable nuclear interference reactions) and irradiated for 25 h in the Hamilton McMaster University nuclear reactor (Canada) in position 5C. The mean J-values computed from standard grains is 0.0087750 ± 0.0000430 determined as the average and standard deviation of J-values of the small wells for each irradiation disc. Mass discrimination was monitored using an automated air pipette and applied correction factors for interfering isotopes were $(^{39}Ar/^{37}Ar)_{Ca} = 7.30 \times 10^{-4}$ (± 11%), $(^{36}Ar/^{37}Ar)_{Ca} = 2.82 \times 10^{-4}$ (± 1%) and $(^{40}Ar/^{39}Ar)_{K} = 6.76 \times 10^{-4}$ (± 32%). The ages are calculated using the new K decay constant reported by Renne et al. (2010).

The samples were step-heated using a 110 W Spectron Laser Systems, with a continuous Nd–YAG (IR; 1064 nm) laser rastered over the sample during 1 mn to ensure a homogenously distributed temperature. The gas was purified in a stainless steel extraction line using three SAES AP10 getters and a liquid nitrogen condensation trap. Ar isotopes were measured in static mode using a MAP 215-50 mass spectrometer (resolution of ~600; sensitivity of 2×10^{-14} mol/V) with a Balzers SEV 217 electron multiplier mostly using 9 to 10 cycles of peak-hopping. Ar isotopic data corrected for blank, mass discrimination and radioactive decay are given in a supplementary appendix.

4. Sample description and geochronology

The Cerro Colorado granite (SdV702) is a high silica, slightly peraluminous magmatic body that outcrops in the eponymous quarry (Grecco et al., 1997). The fluorite-bearing granite exhibits a large number of low-grade mylonitic shear zones dipping to the SW, with S–C fabrics and asymmetric porphyroclasts indicate thrusting to the NE. A weak to well-developed fabric in the granite is defined by foliated zones of sericite and polygonal, recrystallised quartz interspersed with porphyroclasts of flame-textured K-feldspar mesoperthites. Muscovite is intergrown with biotite, and late chlorite is observed. The overall grain size is bimodal, ranging from 1 to 10 mm, with fine-grained zones serving as the locus of deformation-induced recrystallization, the intensity of which is correlated with distance from these shear zones.

The U–Pb SHRIMP analysis reveals a homogeneous, concordant population of zircons with no evidence of inheritance of an older population. The concordia age of 524 ± 4 Ma was from SHRIMP analyses of 16/17 zircon grains, marginally younger than the intercept age of 531 ± 4 Ma (95% confidence) reported by Rapela et al. (2003) (Fig. 3). We note that the new, younger age assignment is a marginal improvement on the previous age assignment, being based on larger number of analyses, with a smaller overall correction for common Pb measured by ^{204}Pb, versus the excess ^{207}Pb used by Rapela and colleagues.

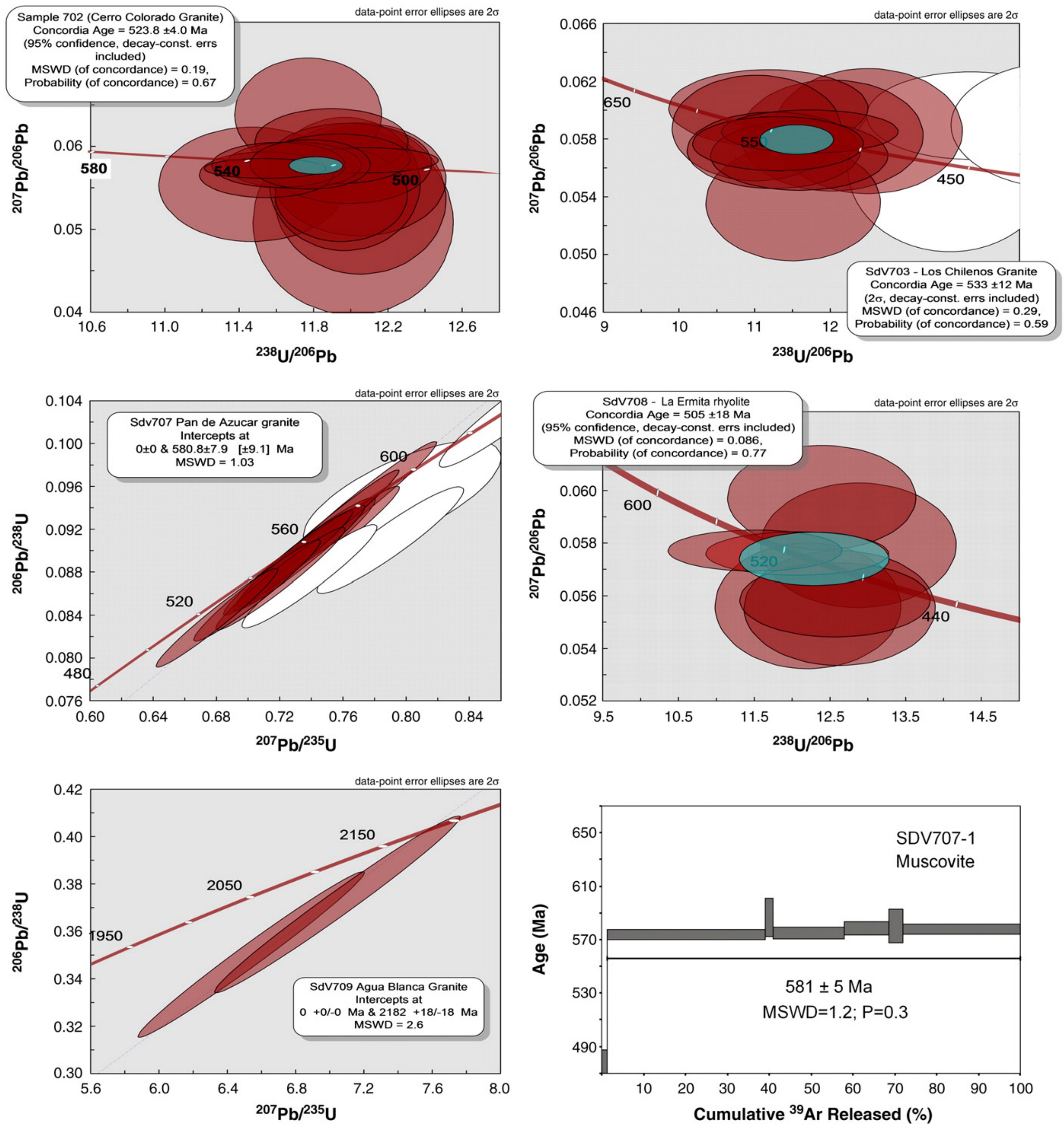

Fig. 3. U–Pb Concordia plots for each of the analysed samples and $^{40}Ar/^{39}Ar$ data for sample SdV 706 showing plateau age for muscovite.

Sample SdV703 was taken from the Los Chilenos Granite, an aplitic, K-spar rich-body collected from outcrops ca. 300 m outside the Cerro Colorado quarry. The fine-grained texture of these granites relative to the Cerro Colorado granite serves as a basis for the distinction, although similar aplitic textures were also observed within the Cerro Colorado granite by von Gosen et al. (1991). The distinction of two granite bodies is supported by the intrusive contact described by Massabie et al. (1999), including chilled margins at the contact with the arenose conglomerates of the basal Curamalal Group (Mascota Formation) and also with the Cerro Corral granite. The presence of an older (meta)sedimentary country rock supports the stratigraphic columns proposed by Kilmurray (1975). The Rb–Sr data reported by Massabie et al. (1999) for this fine-grained granite are perturbed by the altered mineral assemblage observed in thin section (sericitized K-feldspar, quartz, and plagioclase, ± muscovite and opaques). Zircon from sample 703 yielded an age of 533 ± 12 Ma, with the poor age precision reflecting the effects of minor Pb loss (Fig. 3). As a result, we cannot statistically distinguish the age of the Los Chilenos granite from that of the Cerro Colorado granite (sample SdV702), so the two bodies may represent a single magmatic event, with the finer-grained texture of Los Chilenos facies possibly reflecting proximity to an intrusive contact with the country rock.

Sample SdV706 and SdV707 are the from the Pan de Azucar/Cerro do Corral locality, where basement is exposed in the core of an asymmetric anticlinal structure that verges to the NE. Sample SdV707 was taken approximately 100 m from the Cerro del Corral sample of Rapela et al. (2003). Greenschist facies deformation has affected the overlying conglomeratic sedimentary rocks (La Lola Formation of the lower Curamalal Group) with the lower limb having a vertical to slightly overturned attitude marked by high strain fabrics. Lower greenschist facies deformation of the basement rock resulted in the development of an anastamosing protomylonitic foliation. In thin section, this foliation is formed by oriented masses of white mica and chlorite with pull-apart clasts of K-feldspar and tails of recrystallised, polygonal quartz. Euhedral porphyroclasts of muscovite are also observed with deformation microstructures (undulose extinction, kink-bands) that indicate overprinting during the late Paleozoic orogeny. Petrographic observations of sample SdV706 reveal massive, igneous textures that have been extensively metasomatized: heavily sericitized K-feldspars and fracture-filling carbonate cements. Both samples are muscovite bearing, suggesting that the samples are the product of the same episode of peraluminous magmatism with variable overprinting by the late Paleozoic deformation.

Zircon analysed by SHRIMP from sample SdV707 yields a concordant intercept age of 581 ± 8 Ma (Fig. 3), based on a coherent population of the youngest grains. A small number of >600 Ma grains are observed in the sample, suggestive of an inherited population. The Ediacaran emplacement age of this body is substantiated by the $^{40}Ar/^{39}Ar$ age of muscovite age from sample SdV706, which yields a six-step plateau age of age of 576 ± 5 Ma (MSWD = 1.2; P = 0.3) including 99% of ^{39}Ar released (Fig. 3).

Sample SdV708 is a rhyolitic tuff with pumice fragments and flow-banding exposed near the La Ermita convent. A strongly developed, lower greenschist fabric verges to the NE, concordant to the regional fabric imparted by the late Paleozoic deformation. In thin section, the sample is characterized by a fine-grained, sub-millimeter framework of polygonal quartz that is overgrown by radiating arrays of chlorite interspersed with zeolite facies minerals. SHRIMP analysis of zircon from the La Ermita rhyolite yields a Concordia age of 505 ± 18 Ma (Fig. 3), in good agreement with the more precise 509 ± 5 Ma age reported by Rapela et al. (2003).

Sample SdV709 is from a fluorite- and cassiterite-bearing granite exposed in a quarry on the Agua Blanca plantation. This silica-rich monzogranite (Rapela et al., 2003) is massive in texture, with no obvious foliation apart from steeply-dipping, cataclastic to mylonitic shear zones that verge to the NE. Both muscovite and biotite are present in the coarse-grained samples (5–15 mm), which are unfoliated apart from the shear zones. Previous attempts to obtain a U–Pb age from this body (Rapela et al., 2003) were foiled by a lack of zircon (Zr <100 ppm, Grecco et al., 1997). The two zircons we recovered from heavy mineral separates of the Agua Blanca granite yield an upper intercept age of 2200 Ma, with a small degree of discordance in both grains. We cannot exclude the possibility of inheritance for these grains, other than noting that both grains are euhedral with little indication of detrital transport.

5. Geochronology results

The geochronological results fall into three general age groups: Paleoproterozoic, Ediacaran, and Cambrian. The oldest ages are from two zircons recovered from the Agua Blanca granite. Assuming that the two zircons in the Agua Blanca granite are not inherited, the Paleoproterozoic age represents the first documented Rio de la Plata basement material in the Sierra de Ventana region and extends the occurrence of basement from the Tandilia region, 200 km to the NW.

The Ediacaran age determined for the Pan de Azucar granites is confirmed by both U–Pb crystallization age of zircon and the $^{40}Ar/^{39}Ar$ cooling age of muscovite. The preservation of the ca. 581 Ma muscovite age indicates that the regional temperatures during the late Paleozoic Gondwanide orogeny were below 400 °C. More significantly for the Neoproterozoic history of the region, the coeval ages from zircon and muscovite suggest a fast cooling rate for this body, consistent with a shallow intrusion.

The early Cambrian age (533 ± 12 Ma) for the Los Chilenos granite overlaps with age of the Cerro Colorado granite (523 ± 4 Ma), at the 2σ level, so both granites may have formed during the same magmatic event. The shallow intrusive level for these granites is supported by the presence of a ~2.0 Ga xenocrystic zircon, likely a Rio de la Plata-derived detrital grain from country rock that was incorporated into the granite body during intrusion. If so, the arenose conglomerates that are typical of the lower Curamalal Group are older than ca. 533 Ma, and are possibly correlated with the Ediacaran strata deposited on the Rio de la Plata craton in the Tandilia region. The peralkaline rhyolites of the La Ermita Formation represent the last phase of magmatism in the region. These high-silica, spherulitic rhyolites erupted at ca. 508 Ma, with a crustal source suggested from the peraluminous chemistry of these rocks. The absence of mafic rocks associated with these rhyolites contradicts the extensional, continental rifting environment proposed by Rapela et al. (2003). Also, the preservation of these subaerially erupted bodies indicates minimal little exhumation of the Sierra de Ventana crust took place prior to the late Paleozoic deformation, so the late Cambrian ushered in a period of crustal stability for the Sierras Australes until the late Paleozoic orogeny.

In summary, the southern margin of the Rio de la Plata craton may be extended by at least 200 km from the Tandilia region based on two new observations, the presence of Paleoproterozoic crust (the 2.2 Ga Agua Blanca granite), and the presence of Ediacaran to earliest Cambrian sedimentary rocks that are intruded by ca. 533 Ma Los Chilenos Granite. The corresponding geology of the Tandilia region is also marked by the presence of a 2.2 Ga basement complex overlain by Ediacaran-aged sediments. The presence of Ediacaran- to Cambrian-aged magmatism distinguishes the geology of the Sierra da Ventana region from that of Tandilia, and suggests a position closer to the boundary of the Rio de la Plata craton. Intriguingly, the Cambrian igneous events in the Sierra da Ventana region are similar in age to the Pampean orogeny along the western boundary of the Rio de la Plata craton, and demonstrate a common geological history.

6. Discussion

The cluster of Ediacaran- to Cambrian-aged orogenies in Africa and South America are historically termed Pan-African or Brasiliano. The geodynamic significance of these belts relative to the supercontinent cycle lies in their record of collisions between different cratons. The separate drift histories of these cratons end with the orogenies that mark the coalescence of India, Australia, Antarctica and the east African cratons into East Gondwana, (Meert, 2003; Collins and Pisarevsky, 2005), the grouping of the Congo, Kalahari, and Rio de la Plata cratons into central Gondwanan (Gray et al., 2009), and the addition of Amazon *cum* West Africa to form Western Gondwana (Trindade et al., 2003; Tohver et al., 2010).

The central theme of this contribution and special issue is the assembly of western Gondwana. In the discussion that follows, we summarize the salient features of the belts that faced the Clymene Ocean. The shared history of these belts, the Araguaia, Paraguai, and the Pampean belts, marks the final stages in the closure of the Clymene Ocean, forming a large collisional belt separating the Amazon craton from the other large Precambrian cratons of South America, the São Francisco and Rio de la Plata. To examine this hypothesis, we review the basic geology and existing geochronology for the individual belts that comprise this greater suture zone.

6.1. Geology and age of Brasiliano and Pampean orogenies

The component belts of the Clymene suture zone are from north to south: 1) the N–S trending Araguaia belt on the eastern margin of the

Amazon craton; 2) the Paraguai belt on the southeastern margin of the Amazon craton, marked by two 90° changes in strike from E–W at 10–14°S latitude to NNW–SSE along its southern extension that defines the eastern limit the Rio Apa basement province, and 3) the Pampean belt found chiefly in Argentina, where it marks the western margin of the Rio de la Plata craton. In this review, we will focus chiefly on the Araguaia, Paraguai, and Pampean belts along the margin of the Amazon and Rio de la Plata cratons. The coeval Brasilia belt along the western margin of the São Francisco craton is not included in this analysis, having been reviewed by Pimentel et al. (2000); and Valeriano et al. (2004). Also, the Brasilia belt marks the ca. 600 Ma closing of the Goianides oceanic basin (Brito Neves et al., 1999) between the São Francisco craton and independent microcontinents (Goiás massif) and island arcs (e.g., Arenopolis, Maria Rosa arcs; Pimentel et al., 2000) . Isotopic and geochronological analysis of the detrital sediments in this belt (Pimentel et al., 2001; Rodrigues et al., 2010) indicate derivation from the São Francisco craton with an increasing contribution from juvenile, Neoproterozoic arc material until ca. 600 Ma, suggesting that the Brasilia belt and the Goianides ocean basin were distanced from the Amazon craton by a wide Clymene Ocean.

6.2. Araguaia belt

The N–S trending Araguaia belt forms the eastern boundary of the Amazon craton, stretching from the mouth of the Amazon River at the equator to approximately 11°S. Historically, the remoteness of the Araguaia belt has been a significant obstacle to the understanding of its geology. The modern outlines of Araguaia belt geology were established by Almeida and Hasui (1984) with an English language summary by Herz et al. (1989), and a more recent overview by Alvarenga et al. (2000).

Most workers divide the belt into an inner zone to the east, where gneiss domes sheathed by supracrustal rocks were metamorphosed at upper amphibolite–facies conditions, and an outer zone to the west comprised of low grade to unmetamorphosed sedimentary rocks atop the Paleoproterozoic to Archean basement rocks of the Amazon craton itself (Table 1). The lowest stratigraphic units of inner zone metasediments are known as the Baixa Araguaia Supergroup, subdivided into the basal Estrondo Group, which grades upwards from arenitic quartzites and micaceous quartzites (Morro do Campo Formation) into biotite and muscovite schists (Xambioá Formation) interbedded with amphibolites and dolomites. Detrital zircons from the Xambioá Formation yield chiefly late Archean ages of 2.6–2.8 Ga (Moura et al., 2008).

Overlying the Estrondo Group is the Tocantins Group, which displays the same decrease in metamorphic from east to west. In the east, greenschist facies conditions are commonly preserved but the equivalent rocks in the west appear to be nearly unmetamorphosed. The Tocantins Groups consists of the basal Couto Magalhães Formation (mudstones and sandstones) that grades upward into the phyllites, quartzites, and marbles of the Pequizeiro Formation. Detrital zircons from quartzites of the Tocantins Group range in age from Archean to dominantly late Mesoproterozoic (Moura et al., 2008). Serpentinized ultramafic bodies are commonly associated with the Tocantins Group, and are considered to be either intrusive into the older metasediments (Hasui et al., 1984), or tectonically-imbricated bodies within metasedimentary units as part of a dismembered ophiolitic complex (Moura et al., 2008; Paixão et al., 2008). Dating of zircon from associated gabbroic bodies yield a $^{207}Pb/^{206}Pb$ age of 817 ± 5 Ma (Gorayeb et al., 2004 in Moura et al., 2008), similar to a 757 ± 49 Ma Sm–Nd whole rock isochron from mafic dykes of the Quatipuru complex (Paixão et al., 2008). Peak metamorphic anatectic melts in the high grade interior of the belt were dated by Pb–Pb zircon evaporation, yielding a range of ages from 540 Ma to 2200 Ma, with the younger part of this range a maximum constraint on the timing of melting (Teixeira et al., 2002).

Table 1
Proposed lithostratigraphic correlations for the Araguaia Belt according to various authors. A major discrepancy regards the intrusive versus fault contact proposed for the Quatipuru mafic rocks.

Araguaia Belt						
Hasui et al (1984)				Moura et al. (2008)		
EoPaleozoic	Granite intrusions, basic dykes			Cambrian		Syntectonic Granites (Ramal do Lontra)
Neoproterozoic	Basic dykes			Neoproterozoic		
Neoproterozoic	Baixo Araguaia Supergroup	Grupo Tocantins	Pequizeiro Fm.	Neoproterozoic	Baixo Araguaia Supergroup	Tocantins Group
Neoproterozoic	Baixo Araguaia Supergroup	Grupo Tocantins	Ultramafic & mafic intrusives	Neoproterozoic	Baixo Araguaia Supergroup	
Mesoproterozoic	Baixo Araguaia Supergroup	Grupo Tocantins	Couto Magalhães Fm.	Neoproterozoic	Baixo Araguaia Supergroup	Mafic-ultramafic Quatipuru ophiolite complex
Mesoproterozoic	Baixo Araguaia Supergroup	Grupo Estrondo.	Canto da Vazante Fm.	Mesoproterozoic	Baixo Araguaia Supergroup	Mafic-ultramafic Quatipuru ophiolite complex
Mesoproterozoic	Baixo Araguaia Supergroup	Grupo Estrondo.	Xambioá Fm.	Mesoproterozoic	Baixo Araguaia Supergroup	Estrondo Group
Mesoproterozoic	Baixo Araguaia Supergroup	Grupo Estrondo.	Morro do Campo Fm.	Mesoproterozoic	Baixo Araguaia Supergroup	Estrondo Group
Archean-Paleoproterozoic	Colmeia Complex (Gneiss domes in interior belt)			Paleoproterozoic		Colmeia Complex
Archean-Paleoproterozoic	Xingu Complex (Undeformed Amazon basement)			Paleoproterozoic		Colmeia Complex

Brasiliano-aged deformation and metamorphism in the Araguaia belt are linked (Fig. 4), with migmatites and banded gneisses at the core of structural domes (e.g., Colmeia complex). Pelitic rocks preserve Barrovian-type metamorphism that ranges in grade from chlorite-in reactions to the K-feldspar isograd associated with partial melts. The general N–S trend of metamorphic isograds in map view is interpreted as reflecting nappe structures (Abreu et al., 1994), suggested by Campanha and Brito Neves (2004) to be the result of orthogonal convergence across the Araguaia belt. More recent studies have identified sinistral strike-slip shear zones active during a more advanced stage of convergence (Fonseca et al., 2004). Furthermore, the identification of extensional shear zones (Alvarenga et al., 2000; Fonseca et al., 2004) along the borders of gneissic domes structures in the interior of the belt (e.g. Xambioá dome) are suggestive of late-orogenic extensional collapse. This deformation sequence and dating of late-stage granites suggests peak metamorphic conditions

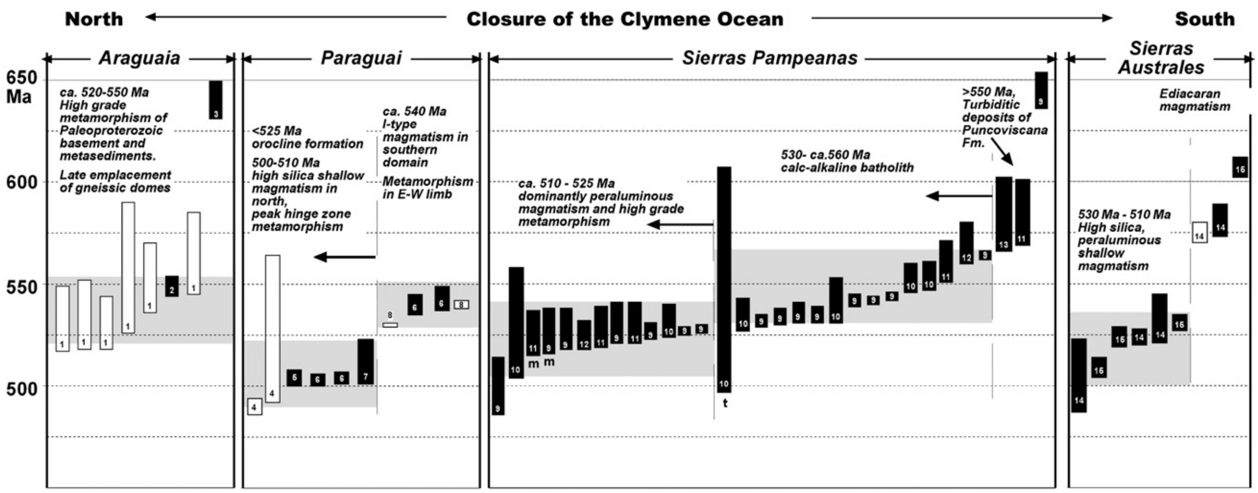

Fig. 4. Geochronological data from the various belts of the Clymene suture zone, plotted at the 2σ level of uncertainty. White symbols are for $^{40}Ar/^{39}Ar$ data and black symbols are for U–Pb data, chiefly zircon except where noted as m—monazite or t—titanite. Numbers refer to the data source as follows: 1 Amaral (1974), 2 Alves (2006), 3 Moura and Gaudette (1993), 4 Tohver et al. (2010), 5 Pinho et al. (2001), 6 Manzano (2009), 7 Tohver (unpublished data), 8—Geraldes et al. (2008), 9—Rapela et al. (1998), 10—Schwartz et al. (2008), 11—Sims et al. (1998), 12—Casquet et al. (2008), 13—Llambías et al. (2003), 14—this study, 15—Rapela et al. (2003).

at ca. 550 Ma (maximum age?) followed by exhumation and cooling of the orogen at ca. 500–530 Ma.

6.3. Paraguai belt

The Paraguai belt marks the SE perimeter of the Amazon craton (Almeida and Hasui, 1984), and the eastern margin of the Rio Apa basement province in its southernmost extension into Paraguay. The Paraguai belt is notable for the two 90° inflections in strike, first, at the poorly exposed juncture with the N–S Araguaia belt where trends switch to an E–W orientation, and a second inflection in the well exposed, western hinge zone where average strike varies smoothly from E–W to NNE–SSW. The southern portion of the belt is dominantly NNW–SSE trending, extending for ca.1000 km. Deformation within the Paraguai fold-thrust belt is thin-skinned, signifying that the regional basement is generally not observed. The youngest Amazon basement rocks are ca. 950 Ma dykes nearly 300 km west of the belt (Elming et al., 2009) overlain by ca. 1150 Ma sedimentary rocks of the Aguapeí belt (D'Agrella-Filho et al., 2008). To the east, basement rocks are putative microcontinents within the Brasilia belt, such as the Goias massif (Pimentel et al., 2000) The presence of the Parapanema block is inferred from geophysical sounding of the crust (Mantovani and Brito Neves, 2005) below sedimentary rocks and flood basalts of the Paraná Basin.

Like the Araguaia belt, the Paraguai belt is divided into a metamorphosed interior and a lower grade exterior belt (Table 2). Generally, the overall metamorphic grade Paraguai belt is lower than the Araguaia belt, not exceeding greenschist facies in its interior (Alvarenga and Trompette, 1993). The geology of the Paraguai belt is dominated by stratified rocks, with an abrupt geomorphologic contrast that distinguishes the peneplaned interior of the belt, where only the metamorphosed Cuiabá Group is exposed, from the ridge-and-valley topography of the lower grade, exterior belt where the Alto Paraguay Group is found. The Cuiabá Group comprise a 4–6 km thick sequence of greywackes, pelites, polymict conglomerates, with subordinate quartzites, graphitic schists, and marbles, and localized occurrences of thin iron formations, imbricated by intense folding and faulting (Barros et al., 1982). The Amazon provenance of the Cuiabá Group is established by the abundance of Paleoproterozoic Nd model ages (Dantas et al., 2009).

The exterior, fold-thrust belt exposes a 2–3 km thick sedimentary sequence of carbonates and siliciclastic rocks known as the Alto Paraguay Group in the northern domain near the central hinge zone. A similar, but younger package of sedimentary rocks in the NNW-trending portion of the belt south of the central hinge zone is known as the Corumbá Group. The stratigraphy of the Alto Paraguai Group itself is well established, with glacial diamictites of the Puga Formation overlain by cap carbonates of the Araras Formation (or Group, Nogueira et al., 2007). Carbon isotope values suggest that these rocks were deposited in the wake of the Marinoan glaciation (~630 Ma), a supposition supported by whole rock Pb–Pb isochron (Babinski et al., 2006). Clastic sedimentary rocks (Raizama, Sepotuba, and Diamantino Formations) overlie the carbonate rocks. This contact has been interpreted as unconformable, with Alto Paraguay Group representing molasse deposited during basin inversion (Almeida and Hasui, 1984; Nogueira et al., 2007; Bandeira et al., submitted for publication).

The stratigraphy of the Corumbá Group to the south has also been extensively studied, but there is persistent uncertainty regarding the stratigraphic position of the diamictites and manganiferous banded iron formations of the Urucum Formation (sensu Almeida, 1954) that outcrops along the tops of fault-bounded blocks uplifted ca.750 m above the Corumbá Group. Recently, Boggiani et al. (2010) proposed that the Cuiabá Group represents the more deformed and metamorphosed distal facies of the Corumbá Group, but the others have suggested an unconformable relationship (e.g., Almeida, 1965).The Corumbá group comprises siliciclastic rocks (Cerradinho Formation) overlain by dolostones (Bocaina Formation) and organic-rich carbonates (Tamengo Formation). These carbonates preserve meter to decametre scale stromatolites and classic Ediacaran fauna such as *Cloudina* and *Corumbella werneri* (Boggiani et al., 2010). Recent dating of ash layers interbedded with the bituminous limestones of the upper Tamengo Formation demonstrates that these rocks straddle the Ediacaran–Cambrian boundary (Babinski et al., 2008).

Only a handful of radiogenic isotope ages directly constrain the age of Paraguai belt deformation and metamorphism (Fig. 4). Regional metamorphism of the Cuiabá Group ~200 km east of the central hinge zone was dated by $^{40}Ar/^{39}Ar$ analysis of three biotites, yielding ages of approximately 530–540 Ma (Geraldes et al., 2008). Peak metamorphism within the hinge zone of the interior belt was determined from $^{40}Ar/^{39}Ar$ dating of different size fractions of illite from strongly-cleaved phyllites, yielding a tightly clustered range of ages from 495 Ma to 500 Ma (Tohver et al., 2010). Shallow level granite intrusions in the interior portion of the belt also provide minimum age constraints on the age of deformation, given the presence of cordierite-bearing hornfels fabrics in country rock around the São Vicente granite (Almeida, 1954) near the central hinge zone, and cordierite-bearing pelitic xenoliths found in the Araguainha granite near the northern hinge zone (Lana et al., 2007).

Table 2
Proposed lithostratigraphic correlations from various authors for the northern (top) and southern domains (bottom) of the Paraguai Belt. For both domains, the nature of the contact between the Cuiabá Group and the Puga Fm. is poorly established.

Paraguai Belt Northern Domain

Almeida (1965b)

Ordovician	Alto Paraguai Gp.	Diamantino Fm.
Ordovician	Alto Paraguai Gp.	Sepotuba Fm.
Ordovician	Alto Paraguai Gp.	Raizama Fm.
Eocambrian - Upper Cambrian	Araras Gp.	
Eocambrian - Upper Cambrian	Araras Gp.	
Eocambrian - Upper Cambrian	Jangada Gp.	Marzagao Fm.
Eocambrian - Upper Cambrian	Jangada Gp.	Bauxi Fm.
Eocambrian - Upper Cambrian	Jangada Gp.	Engenho Fm.
Eocambrian - Upper Cambrian	Jangada Gp.	Acorizal Fm.
Pre €	Cuiabá Series	

Barros et al. (1982)

Precambrian	Alto Paraguai Gp.	Diamantino Fm.
Precambrian	Alto Paraguai Gp.	Sepotuba Fm.
Precambrian	Alto Paraguai Gp.	Raizama Fm.
Precambrian	Alto Paraguai Gp.	Araras Fm.
Precambrian	Alto Paraguai Gp.	Moenda Fm.
Precambrian	Alto Paraguai Gp.	Bauxi Fm.
Precambrian	Cuiabá Series	

Almeida and Hasui (1984)

Cambrian	Alto Paraguai Gp.	Diamantino Fm.
Cambrian	Alto Paraguai Gp.	Raizama Fm.
Precambrian	Araras Corumba Gp.	Tamengo Fm.
Precambrian	Araras Corumba Gp.	Bocaina Fm.
Precambrian	Araras Corumba Gp.	Cerradinho Fm.
Precambrian		Puga Fm.
Precambrian		Bauxi Fm.
Precambrian	Boqui - Jacadigo Gp.	Urucum Fm.
Precambrian	Boqui - Jacadigo Gp.	Santa Cruz Fm.
Precambrian	Cuiabá Series	

Alvarenga and Trompette (1993)

		"External"	"Internal"
Neoproterozoic	"Detrital Units"	Diamantino Fm.	
Neoproterozoic	"Detrital Units"	Raizama Fm.	
Neoproterozoic	"Carbonate Units"	Araras Fm.	Guia Fm.
Neoproterozoic	Glacio-marine	Puga Fm.	Bauxi Fm.
Neoproterozoic	Lower – turbiditic unit	Cuiabá Gp.	

Nogueira and Riccomini (2006)

Cambrian	Alto Paraguai Gp.	Diamantino Fm.
Cambrian	Alto Paraguai Gp.	Sepotuba Fm.
Ediacaran	Alto Paraguai Gp.	Raizama Fm.
Ediacaran	Araras Gp.	Nobres Fm.
Ediacaran	Araras Gp.	Serra do Quilombo Fm
Ediacaran	Araras Gp.	Guia Fm.
Ediacaran	Araras Gp.	Mirassol D'Oeste
630 Ma	Puga Fm.	

Paraguai Belt Southern Domain

Almeida (1945)

Serra da Bodoquena	Tamengo Gp.
Serra da Bodoquena	Bocaina Gp.
Jacadigo Gp.	Santa Cruz Gp.*
Jacadigo Gp.	Urucum Gp.*

*Subdivided by Dorr (1945) into three formations: Urucum, Corrego das Pedras, and Band'Alta Fms.

Almeida (1965a)

Eocambrian	Corumba Gp	Guaicurus Fm.
Eocambrian	Corumba Gp	Tamengo Fm.
Eocambrian	Corumba Gp	Bocaina Fm.
Eocambrian	Corumba Gp	Cerradinho Fm.
Proterozoic		Puga Fm.
Proterozoic	Cuiabá Series	
Proterozoic	Cadieus Fm	

Del'Arco et al (1982)

		Serra da Bodoquena	Corumbá
Upper Precambrian	Corumba Gp - Alto Paraguai Gp.	Tamengo Fm.	
Upper Precambrian	Corumba Gp - Alto Paraguai Gp.	Bocaina Fm.	
Upper Precambrian	Corumba Gp - Alto Paraguai Gp.	Cerradinho Fm.	
Upper Precambrian	Corumba Gp - Alto Paraguai Gp.	Puga Fm - Moenda Fm.	Jacadigo Gp.: Santa Cruz Fm.
Upper Precambrian	Cuiabá Series		Jacadigo Gp.: Urucum Fm.
Upper Precambrian			Cuiabá Series

Boggiani et al. (2010)

	Cratonic	Proximal	Distal
Cambrain	Guaicurus Fm.	Guaicurus Fm.	Guaicurus Fm.
Cambrain	Tamengo Fm.	Bocaina Fm.	Bocaina Fm.
Ediacaran	Morraria do Sul stromatolites		
Ediacaran	Cerradinho Fm.	Cerradinho Fm.	Cerradinho Fm.
Ediacaran	Cadieus Fm.	BIFs, ferriferous diamictite	Cuiabá Group
Ediacaran		Puga Fm.	

These granites were emplaced at shallow conditions in a *tardi-* to *post-*tectonic phase, with U–Pb SHRIMP ages of zircon from the Araguainha granite yield a crystallization age of 510 ± 12 Ma, with rapid cooling through monazite (Tc ~ 700 °C) and feldspar (Tc ~ 300 °C) closure temperatures occurring at ca. 490 Ma and ca. 430 Ma, respectively (Tohver et al., in review). The nearby Lajinha granite was dated by conventional U–Pb zircon analysis at 505.4 ± 4.1 Ma (Manzano, 2009), and zircon from the São Vicente granite was dated by LA-ICP-MS to 518 ± 4 Ma (McGee et al., submitted for publication). A recent summary of the Paraguai belt magmatism by Manzano (2009) suggests that magmatic rocks in the southern, NNW-trending portion of the belt are older and more primitive. The U/Pb age of zircon from these bodies constrains the timing of crystallization: Taboco (ca. 540 Ma), Rio Negro, Coxim (540–546 Ma), and Sonora granitoids (Manzano et al., 2008).

Paleomagnetic study provides independent age constraints on regional deformation. A pervasive remagnetization in the Araras Group carbonates (Trindade et al., 2003) overlaps the direction established by the 525 Ma Itabaiana dykes reference pole (Trindade et al., 2006), in agreement with the ca. 528 Ma established by $^{40}Ar/^{39}Ar$ encapsulation dating of authigenic clays linked to the remagnetization (Tohver et al., 2010). Ongoing deformation in the Paraguai belt resulted in the 90°, clockwise rotation of the E–W limb of the belt after ca. 528 Ma (Tohver et al., 2010), a rotation that was probably accommodated by strike-slip motion along E–W striking duplex stuctures (Martinelli, 1998). Taken together, these observations suggest that early thrusting in the Parguaia belt at ca. 528 Ma was followed by E–W strike-slip motion, so convergence in the Paraguai belt took place in the Cambrian and was largely E–W directed.

6.4. Sierras Pampeanas

The type locality of the Pampean orogeny lies in the ca. 250 km-wide Eastern Sierras Pampeanas between 26°S and 34°S, where high-relief uplift of the basement occurs above a zone of flat-slab subduction of the Nazca plate (e.g. Isacks, 1988). The eastern limit of the effects of the Pampean orogeny defines the margin of the Rio de la Plata craton, which extends in the subsurface to just east of the Sierras de Cordoba (Ramos, 1988; Rapela et al., 2007). To the west of the Sierras Pampeanas lies the Famatinian belt, a broad zone of rocks reworked or formed during the Cambro-Ordovician orogeny (Ramos, 1988; Kraemer et al., 1995; Rapela et al., 1998; Sato et al., 2003; Vaughan and Pankhurst, 2008). The northern limit of the Pampean belt is not well defined; according to Ramos et al. (2010) the belt terminates in either the NW–SE trending Tucavaca aulacogen of southern Bolivia outboard of the Sunsas Province (Litherland et al., 1989; Teixeira et al., 2009), or, according to Brito Neves et al. (1999), the trace of the northern Pampean belt is continuous with the Paraguai belt (Brito Neves et al., 1999). The southern extent of the Pampean belt is exposed in isolated outcrops, and is inferred from aeromagnetic data to follow a N–S trend parallel to the Rio de la Plata cratonic margin (Chernicoff and Zappettini, 2004; Chernicoff et al., 2010, this issue).

The oldest rocks in the belt belong to the Puncoviscana Formation (Table 3), a thick Ediacaran- to early Cambrian-aged sequence of turbidites that extend to southern Bolivia (Aceñolaza and Aceñolaza, 2007). The Puncoviscana Formation is overlain in these northern reaches by the littoral facies sandstones of the Meson Group (Aceñolaza, 2003), the late Cambrian age of which brackets the timing of Pampean orogenesis. Paleocurrent directions within the Puncoviscana indicate deposition from an eastern source region, but, curiously, the detrital zircon populations from the Puncoviscana Formation are dominated by Mesoproterozoic "Grenville-aged" and late Neoproterozoic (600–700 Ma) ages (Sims et al., 1998; Schwartz and Gromet, 2004; Adams et al., 2008). The lack of detrital input from the Rio de la Plata craton itself suggests a subdued topography at this time and that the principal sediment sources were located elsewhere, such as the Sunsas province to the N (Schwartz and Gromet, 2004). A more recent detrital zircon study by Escayola et al. (2007) posits a western source region, as yet unidentified Grenvillian basement within the Pampian terrane. Another possibility is that the sediment source lies even further east, in the Kalahari craton, an idea explored by Spagnuolo et al. (this volume). According to these workers, the inferred Kalahari provenance of detrital material is linked to paleomagnetic observations of systematic clockwise rotations, both caused by the strike-slip migration of the Pampia terrane from the Panthalassan margin of the Kalahari craton to the Rio de Plata craton in the late Cambrian.

Geochronological constraints for the high temperature history of the Pampean orogeny are provided by U/Pb and Ar/Ar geochronological constraints from individual localities (Fig. 4) within the Sierras Pampeanas: the Sierra do Córdoba (Rapela et al., 1998; Sims et al., 1998; Escayola et al., 2007); the Sierra Norte do Córdoba (Schwartz et al., 2008), and the Sierra de San Luís (Sims et al., 1998). Generally, speaking the Pampean belt is marked by an early phase of calc-alkaline volcanism at 555–525 Ma (Rapela et al., 1998; Escayola et al., 2007). Peak metamorphism attains the granulite to amphibolite facies, with thermobarometric work in the aluminous greywackes (Puncoviscana equivalent?) of the Calamuchita metamorphic complex indicating peak temperatures of 750–850 °C at 700–800 MPa (Otamendi et al., 1999). Isothermal decompression in hot, orogenically-thickened crust is inferred from garnet-consuming, plagioclase-producing reactions associated with widespread emplacement of cordierite-bearing, peraluminous magmas at 525–515 Ma (Rapela et al., 1998, 2007; Schwartz et al., 2008). Otamendi et al. (2004) interpret the rapid onset of this decompression as tectonically-driven, rather than the result of slower isostatic reequilibration via erosional unloading. Different tectonic mechanisms for this rapid exhumation have been proposed; collisional docking of the Pampia terrane (e.g. Ramos et al., 2010), or subduction-related magmatism followed by ridge subduction (Schwartz et al., 2008).

Table 3
Proposed lithostratigraphic correlation for the Pampean belt by Aceñolaza (2003).

		Pampean Belt
		Aceñolaza (2003)
Paleozoic	Ordovician	Santa Victoria Formation
	Cambrain	Meson Group
		Puncoviscana Formation
Neoproterozoic	Ediacaran	
	Cryogenian	

7. Conclusions

Our new geochronological data provides evidence for two critical observations regarding the Rio de la Plata craton and its record of the assembly of western Gondwana (Fig. 5). First, the southern limit of the Rio de la Plata craton extends ca. 200 km farther to the south than previously supposed. This observation is based on the presence of Paleoproterozoic crust (Agua Blanca granite) and the early-Cambrian–Ediacaran(?) sedimentary rocks that are intruded by the ca. 533 Ma granites in the Cerro Colorado region. Secondly, the Ediacaran through late Cambrian magmatism along the southern margin of the Rio de la Plata craton overlaps the age of the Pampean orogeny. Unlike the Pampean belt type locality, the magmatic rocks of the Sierras Australes are volcanic or hypabyssal, and there is no high-grade metamorphic overprint in this region. Thus, the Pampean-aged magmatic products in the Sierras Australes reflect volcanism along the southern margin of the Rio de la Plata craton inboard of the active deformation and crustal reworking in the Pampean belt itself.

Our review of the tectonic histories of the three Clymene belts indicates some common features, which reflect a shared geodynamic

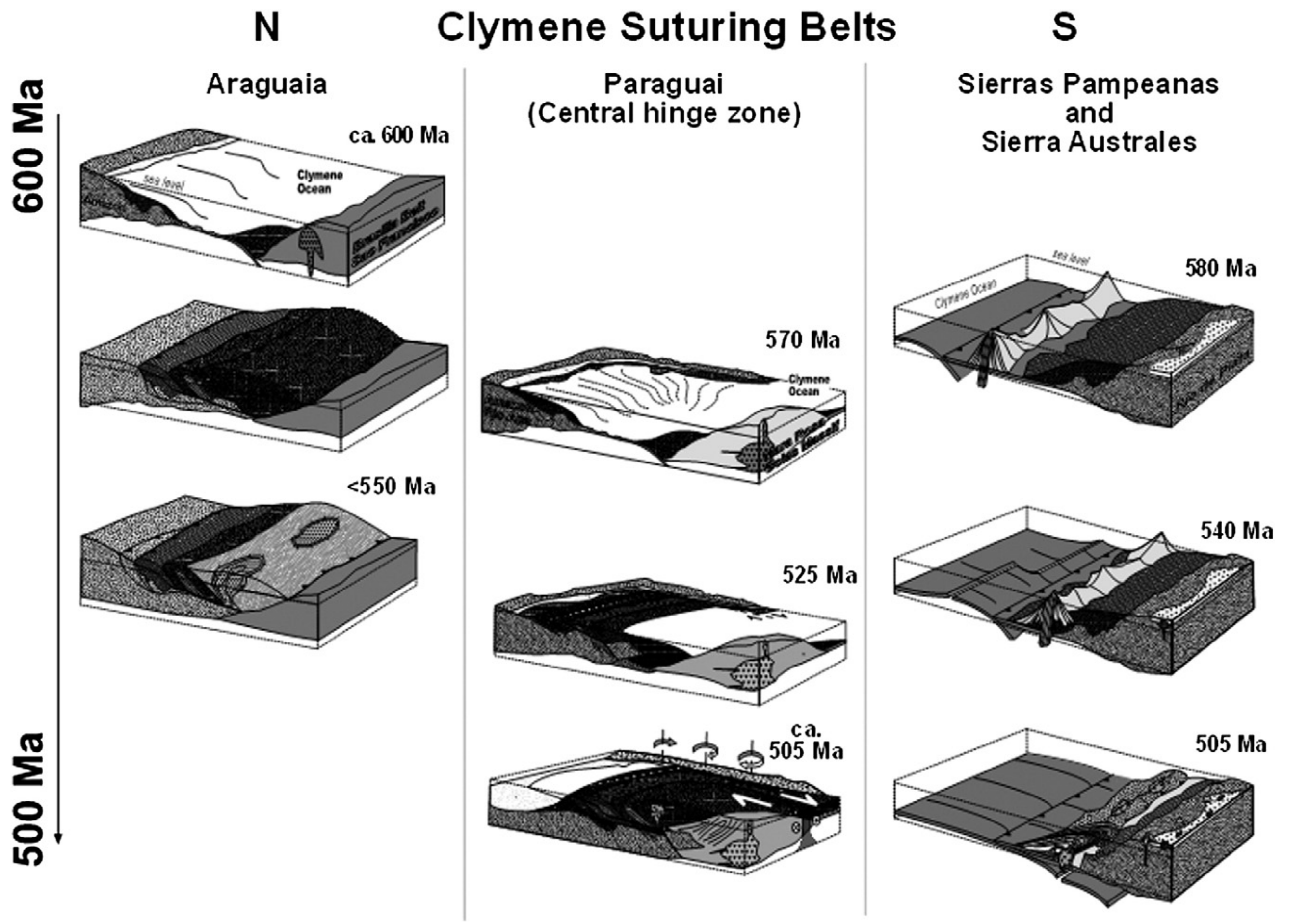

Fig. 5. Diagram illustrating the events and relative age of major tectonic episodes across the Clymene suture zone reconstructed from survey of regional literature. The timing of events in the Araguaia belt is the least certain, and U–Pb data probably represent maximum ages. The structural rotations in the Paraguai belt are known to have occurred after ca. 528 Ma, based on paleomagnetic studies. Note that the Goianides Ocean would be located between the Goiás massif-Mara Rosa arc and the São Francisco craton (not pictured here). For the Pampean belt, the Puncoviscana Formation is depicted as accumulating on a passive western margin of the Rio de Plata craton, although alternative models are discussed in the text. The link between ridge subduction and peraluminous magmatism in the late Cambrian of the Pampean belt follows the model of Schwartz et al. (2008). Note the lack of a colliding block with Pampean belt, which is left unconfined and available for continuing reactivation throughout the Paleozoic, contrary to the Brasiliano belts to the north.

setting. The early history of sedimentation generally reflects mature, craton-derived material such as pelites, or carbonates in more quiescent settings (Fig. 5). The Araguaia and Paraguai Belts receive most of their sedimentary input from the Amazon craton (Moura et al., 2008; Dantas et al., 2009), whereas the source of detrital sediments in the Pampean belt is still uncertain. The transition from passive accumulation of craton-derived material in Ediacaran times to magmatism, metamorphism, and deformation took place in the period from the 550 to 500 Ma, as indicated by our synthesis of geochronological data from all three belts (Fig. 4). In all three localities, early, I-type magmatism in the 530–560 Ma period was followed by more differentiated, high silica magmatic or peraluminous products in the 500–520 Ma period (Fig. 5).

Some general differences between the three belts also emerge from this overview. For example, the metamorphic record of crustal thickening is pronounced in the Araguaia and Pampean belts, where U–Pb whole zircon age data may partly reflect the effects of this high grade metamorphism. The presence of reactivated basement domes in the Araguaia belt suggests a phase of post-orogenic collapse, mirrored by the large-scale anatectic melting and S-type magmatism in the Pampean belt (Fig. 5). In contrast, the peak greenschist metamorphism observed in the Paraguai belt suggests that crustal thickening was subdued in this portion of the belt, possibly as a consequence of the irregular shape of cratonic margin before the formation of the Paraguai belt (Tohver et al., 2010). In the Sierra da Ventana region of the greater Pampean belt, regional metamorphism in Ediacaran to Cambrian times is absent, demonstrating a more stable position away from the cratonic margin.

The most significant difference that emerges from comparison of these three belts lies in their subsequent geological histories. Throughout the early Paleozoic, the Pampean belt remained unconfined along its western margin, with tectonic activity continuing throughout the Cambro-Ordovician Famatinian orogen. The continuous westward migration of the locus of Paleozoic tectonism suggests that the growth of the South American protocontinent along its SW margin resulted from accretionary processes along an active margin, curtailed only by the docking of the Precordillera terrane in the early-middle Paleozoic (Thomas and Astini, 1996). Thus, the Pampean record of magmatism coupled with deformation appears to be the first step in a long-lived process that contrasts with the quiescence of the confined Clymene belts to the north, where the bookend cratons shielded the belts from reactivation.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at doi:10.1016/j.gr.2011.04.001.

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