Review of the Cambrian Pampean orogeny of Argentina; a displaced orogen formerly attached to the Saldania Belt of South Africa?

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Abstract

The Pampean orogeny of northern Argentina resulted from Early Cambrian oblique collision of the Paleoproterozoic MARA block, formerly attached to Laurentia, with the Gondwanan Kalahari and Rio de la Plata cratons. The orogen is partially preserved because it is bounded by the younger Córdoba Fault on the east and by the Los Túneles Ordovician shear zone on the west. In this review we correlate the Pampean Belt with the Saldania orogenic belt of South Africa and argue that both formed at an active continental margin fed with sediments coming mainly from the erosion of the Brasiliano–Pan-African and East African–Antarctica orogens between ca. 570 and 537 Ma (Puncoviscana Formation) and between 557 and 552 Ma (Malmesbury Group) respectively. Magmatic arcs (I-type and S-type granitoids) formed at the margin between ca. 552 and 530 Ma. Further right-lateral oblique collision of MARA between ca. 530 and 520 Ma produced a westward verging thickened belt. This involved an upper plate with high P/T metamorphism and a lower plate with high-grade intermediate to high P/T metamorphism probably resulting from crustal delamination or root foundering. The Neoproterozoic to Early Cambrian sedimentary cover of MARA that was part of the lower plate is only recognized in the high-grade domain along with a dismembered mafic–ultramafic ophiolite probably obducted in the early stages of collision. Uplift was fast in the upper plate and slower in the lower plate. Eventually the Saldania and Pampean belts detached from each other along the right-lateral Córdoba Fault, juxtaposing the Rio de la Plata craton against the internal high-grade zone of the

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Pampean belt.

1. Introduction

The Pampean Belt of central and North Western Argentina is part of a long chain of Cambrian orogenic belts that formed in southwestern Gondwana. They include the Araguaia belt of Brazil, the Paraguay belt of southwestern Brazil and eastern Bolivia (Fig. 1), and the Saldania Belt of South Africa. The belt extends further east into the Transantarctic Mountains of Antarctica and the Delamerian belt of southern Australia. These belts concluded the amalgamation of Gondwana in the Early Cambrian.

The Pampean orogeny was first recognized in the Sierras Subandinas and the Sierras Orientales of North Western Argentina (Aceñolaza and Toselli, 1973, 1976). The orogeny was inferred from the Tilcarian unconformity between the allegedly Late Precambrian to Early Cambrian, strongly folded, low-grade metasedimentary Puncoviscana Formation and the overlaying post-orogenic middle to Late Cambrian Mesón Group. This constrained its timing to Early-to-Middle Cambrian (e.g., Aceñolaza and Toselli, 1981). Further work, including precise U-Pb geochronology (for a review see Baldo et al., 2014), has confirmed that the orogeny can also be recognized in the easternmost Sierras Pampeanas, where there are metasedimentary rocks equivalent to the Puncoviscana Formation and regional metamorphism and magmatism has been dated between ca 545 and 520 Ma, i.e., Early Cambrian (Rapela et al., 1998; Otamendi et al., 2004; Schwartz and Gromet, 2004; Escayola et al., 2007; Ianizzotto et al., 2013; Murra et al., 2016). Cambrian magmatism has also been reported from Patagonia (Hervé et al., 2010; Pankhurst et al. 2014).

The Pampean orogeny involved strong folding accompanied by penetrative foliation, shearing and low to high-grade regional metamorphism under high P/T to intermediate and low P/T conditions. The significance of this orogeny has long been a matter of much debate but there is agreement now that it involved the closure of an ocean to the west and concluded with the collision of continental blocks (e.g., Ramos et al., 1988; Rapela et al., 1998). The main current models represent two alternative tectonic interpretations: 1) orthogonal collision involving subduction beneath the Rio de la Plata craton in its present relative position (Ramos et al. 2015, Fig. 10 A; and references therein) or 2) transpressional orogeny that juxtaposed the orogenic belt and the Rio de la Plata craton by right-lateral displacement at the end of or after the orogeny

(Rapela et al., 2007, 2016; Casquet et al., 2012).

In this contribution we present a model of the Pampean orogeny in the Sierras Pampeanas and North Western Argentina particularly focused on the structural, metamorphic and magmatic evolution. Correlation of the Pampean belt with the Saldania Belt of southern Africa is as part of the hypothesis of significant right-lateral translation (e.g., Rapela et al., 2007). The tectonic model proposed here accounts for the similarities and explains the final displacement of the Pampean section of the orogenic belt relative to the Saldanian orogen.

2. Definition and Boundaries

The Pampean belt crops out in the westernmost Sierras Pampeanas (Sierras de Córdoba and Sierra Norte) and in the Cordillera Oriental of North Western Argentina. The Sierras Pampeanas constitute a morphotectonic region of elongated outcrops (sierras) of pre-Andean basement resulting from Cenozoic reverse faulting of the Andean foreland. The width of this belt is low (ca. 100 kilometres at most) because the original belt is actually truncated on both sides (Fig. 2).

On the eastern side is the Córdoba Fault, an important geological and geophysical discontinuity that separates the Sierras Pampeanas from the Río de la Plata craton (Favetto et al., 2008, Ramé & Miró, 2011; Peri et al., 2013). The latter is a Paleoproterozoic block that shows no evidence of reworking by the Pampean orogeny (Rapela et al., 2007): it reached its present position after right-lateral displacement during and immediately after the Pampean orogeny in the Early–Middle Cambrian (Rapela et al., 2007; Siegesmund et al., 2010; Drobe et al., 2009; Spagnuolo et al., 2012). The Córdoba Fault has been correlated with the transcontinental Transbrasiliano Lineament of Schobbenhaus Filho (1975) and Cordani et al. (2003) (Rapela et al., 2007; Ramé & Miró, 2011) (Figs. 1 & 2). On its western side the Pampean orogen is juxtaposed against the Ordovician Famatinian orogenic belt across the anastomosed Los Tuneles–Guacha Corral ductile westward thrust (Figs. 2 & 3), which superposed the internal high-grade zone of the Pampean orogen over rocks of the Conlara Complex of the eastern Sierras de San Luis, probably at 440–430 Ma (Martino, 2003; Whitmeyer & Simpson, 2003: Steenken et al., 2010). The Conlara Complex underwent medium-grade regional metamorphism in the Early to Middle Ordovician (Steenken et al., 2006). The western boundary of the Pampean orogen can be traced northward into the Sierras

Subandinas of North Western Argentina (Fig. 2).

3. Metamorphic and Tectono-stratigraphic Domains.

A primary internal division of the Pampean orogen can be made on the basis of metamorphic grade. Most of the orogen consists of low-grade rocks (Fig. 2) exposed in a large region embracing the Sierras Norte de Córdoba and NW Argentina together with minor outcrops in the Sierra de Guasayán, Sierras de Córdoba and probably in the Sierra de Ancasti (Ancasti Formation) and the eastern Sierra de San Luis (Conlara Complex). In the latter cases no evidence of Pampean metamorphism has been preserved because the rocks were overprinted by medium- to high-grade metamorphism during the Famatinian orogeny. The second metamorphic domain corresponds for the most part to the Sierras de Córdoba (Fig. 2), where Pampean metamorphism attained high-grade conditions and migmatites are widespread. The boundary between these two domains remains to be precisely defined. Metamorphic continuity is everywhere disrupted by younger shear zones or faults such as the Carapé fault in northernmost Sierra Chica (Fig. 2) which underwent significant displacement in the Ordovician (Martino, 2003). However the absence of a medium-grade domain could mean that it was thinned out late during the Pampean orogeny and that the overall picture was that of a mantled gneiss dome (e.g., An Yin, 2004) before reworking along younger faults or shear zones.

The Pampean orogeny has long been considered as a case of continental collisional, mainly on the evidence of metamorphic *P-T* conditions and magmatism (Rapela et al., 1998; Ramos et al., 2010). Recent work on U-Pb dating of detrital zircon and Sr isotope blind dating of marbles strengthens this interpretation. In fact here werer two contrasting sedimentary successions (see below) involved in the orogenic belt that were apparently sourced from opposite continents (Casquet et al., 2012; Rapela et al., 2016; Murra, 2016). This in turn implies that two main tectono-stratigraphic domains exist in the Pampean orogen representing upper and lower plates. The boundary between them apparently coincides with a dismembered mafic-ultramafic complex whose outcrops are scattered across the Sierra Grande and Sierra Chica de Córdoba. The complex consists of meta-peridotite, meta-pyroxenite, meta-gabbro, massive chromitite, and minor leucocratic rocks that have been interpreted as an ophiolite (upper mantle and oceanic crust), i.e., relics of a suture (e.g., Ramos et al., 2000; Escayola et al., 2007; Proenza et al., 2008; Martino et al., 2010, among others).

The two tectono-stratigraphic domains would thus represent the two continental margins that collided to produce the Pampean orogeny and the dismembered mafic-ultramafic complex would be the relic of the intervening Neoproterozoic to Early Cambrian oceanic lithosphere. The ocean correlates with the southern extension (present coordinates) of the Clymene Ocean that existed between Amazonia and other Gondwanan blocks (Trindade et al., 2006). Continental collision brought to an end the amalgamation of Gondwana (Rapela et al., 2016; Murra et al., 2016).

The boundaries between the metamorphic domains and the Pampean suture are not coincident. The upper plate is for the most part represented by the low-grade domain although it was also imbricated in the high-grade domain. The lower plate however is only preserved in the high-grade domain. Figure 3 shows a schematic cross-section showing the hypothetical relationships between upper and lower plates. Colliding blocks in Figure 3 are MARA, i.e., a Paleoproterozoic block named after three of its alleged outcrops: Sierra de **M**AZ, **A**requipa (Peru) and **R**io **A**pa (southern Brazil) according to Casquet et al. (2012) (see section 6.1) and the Kalahari - Rio de la Plata cratons.

4. The Low-Grade Domain

4.1 The Punscoviscana Formation

The Puncoviscana Formation (Turner, 1960) is a thick, mainly siliciclastic, partly turbiditic succession with minor limestone and volcanic beds (Ježek, 1990; Omarini et al., 1999; Zimmermann, 2005; Aceñolaza and Aceñolaza, 2007) that is widespread in NW Argentina. Its age and the tectonic setting of sedimentation have been controversial. The term Puncoviscana Formation in the literature embraces rocks stratigraphically older than the unconformably overlying Middle to late Cambrian Meson Group (e.g., Omarini et al., 1999; Adams et al., 2008, 2011, and references therein). Originally described as the “basal Precambrian shield” it was first recognized as of Late Neoproterozoic to Early Cambrian age on the basis of trace fossils (Aceñolaza and Durand, 1973; Aceñolaza and Toselli, 1976). Correlation with high-grade metamorphic rocks in the eastern Sierras Pampeanas was first established by Rapela et al. (1998) and confirmed by subsequent isotope studies (Schwartz and Gromet 2004; Rapela et al. 2007, 2016; Murra et al., 2011; Escayola, 2007, among others). This thus presents the main evidence for the Early Cambrian Pampean orogeny. The Puncoviscana Formation characteristically contains an almost bimodal detrital zircon population with major peaks at 1100–960 Ma and 680–570 Ma and a few grains of 1.7–2.0 Ga and ca. 2.6 Ga; it lacks grains derived from the nearby Rio de la Plata craton (2.02–2.26 Ga) (Schwartz and Gromet, 2004; Escayola et al., 2007; Rapela et al., 2007). Sedimentation took place between ca. 570 Ma and ca. 537 Ma (Rapela et al., 2007, 2016; Escayola et al., 2011; Aparicio González et al., 2014). Contrasting sedimentary settings have been proposed, such as a passive continental margin (Do Campo & Ribeiro Guevara, 2005; Piñan-Llamas & Simpson, 2006) and a shallow extensional aulacogen (Aceñolaza and Toselli, 2009). A forearc basin on an active continental margin resulting from oblique closure of the Clymene Ocean was suggested by Rapela et al. (2007). The basin was probably not adjacent to the Rio de la Plata craton until both became juxtaposed through right-lateral displacement along the Cordoba Fault (Schwartz and Gromet, 2004; Rapela et al., 2007; Verdecchia et al., 2011; Casquet et al., 2012). This kinematic model was accepted by Drobe et al. (2009), Siegesmund et al. (2010) and Llamas & Escamilla (2013) and right-lateral displacement was found to be compatible with paleomagnetic evidence (Spagnuolo et al., 2012). If *P* values of 8 - 9 kbar at 240º - 300º C attained during metamorphism (Do Campo et al., 2013) are confirmed by future work, the Puncoviscana Formation could have originated as an accretionary prism coeval with oblique eastward subduction of the Clymene Ocean.

Rocks with similar detrital zircon U-Pb ages are also recognized further west in the Sierras de Ancasti (as the Ancasti Formation) and San Luis (Conlara Complex) (Rapela et al., 2007, 2016; Steenken et al., 2006), but in both places Famatinian deformation and metamorphism has overprinted the earlier structures of the Puncoviscana Formation, masking the evidence for a pre-Famatinian event (Steenken et al., 2006). Verdecchia et al. (2012) showed that medium-grade metamorphism of the Ancasti Formation resulted from a single event of Famatinian age, so that this and the Conlara Complex were probably originally akin to the low-grade Puncoviscana Formation elsewhere.

A large outcrop of phyllite equivalent to the Puncoviscana Formation is found near Los Tuneles (the Los Túneles Phyllites of Rapela et al., 1998; Escayola et al., 2007) (Figs. 2 & 3) (Baldo et al., 1996; Rapela et al., 1998; Escayola et al., 2007), where it is separated by faults and shear zones from the high-grade domain (Martino et al., 2003).

One sample from the Los Túneles Phyllites (TLT-2069) has been analysed for U-Pb SHRIMP zircon chronology and $\delta^{18}O$ and Hf isotope measurements on dated zircon (analytical methods as in Rapela et al., 2016) for comparison with North Western Argentina and the Saldania Belt (see Discussion section). The results from TLT-2069 are shown in Tables 1 and 2, and detrital zircon ages represented in Fig. 4. This age spectrum shows two well defined peaks in the ranges 562–690 Ma and 953–1100 Ma typical of the Puncoviscana Formation (Rapela et al., 2016). There are a few single grains ages of 777, 890, 1240 1880 and 2505 Ma. The $\delta^{18}O$ values range from +5.74‰ to +10.71‰. The εHf_t values are positive (+0.7 to +12.0) and Hf model ages (single-stage) are in the range 710 –1440 Ma.

4.2 The magmatic arc

The mostly low-grade upper plate domain comprises low-grade clastic metasedimentary rocks equivalent to the monotonous Late Neoproterozoic to Early Cambrian Puncoviscana Formation. It hosts a Cordilleran calc-alkaline magmatic arc of Early Cambrian age – the Sierra Norte–Ambargasta batholith, the Tastil and other plutons in NW Argentina. The arc rocks are I-type metaluminous to slightly peraluminous, ranging from diorite to leucogranite, dacite porphyry and even volcanic rocks (rhyolite) and tuffs. U-Pb zircon ages of plutons, dykes and tuffs that can be confidently attributed to the Pampean orogeny range from 541 ± 4 to 523 ± 2 Ma (Lyons et al., 1997; Rapela et al., 1998; Stuart-Smith et al., 1999; Leal et al., 2003; Hauser et al., 2011; Escayola et al., 2007, 2011; Tibaldi et al., 2008; Aparicio Gonzalez et al., 2011; Hong et al., 2010; Siegesmund et al., 2010; Iannizzotto et al., 2013; von Gosen et al., 2014; Dahlquist et al., 2016) (Table 3). The metaluminous Sierra Norte–Ambargasta batholith was emplaced within a very short period of time: the main intrusive pulse took place between 537 ± 4 and 528 ± 2 Ma (based on 11 results) with most ages in the range 530 ± 4 Ma (Rapela et al., 1998; Leal et al., 2003; Siegesmund et al., 2010; Iannizzotto et al., 2013; von Gosen et al., 2014; Table 3) and thus constitutes a magmatic flare-up. This pulse was largely coeval with regional compressional strain (Iannizzotto et al., 2013; Von Gosen et al., 2014). Volcanic rocks of 531 ± 4 Ma (Agua del Rio dacite porphyry) (Table 3) and 531 ± 3 Ma (Rodeito rhyolite-dacite; age recalculated from $^{206}Pb/^{238}U$ ages after Von Gosen et al., 2014) attest to their contemporaneity with plutonism.

A group of rhyolites and granites with ages between 519 ± 4 and 512 ± 4 Ma (Table 3) is probably related to late orogenic uplift (Ramos et al., 2015). However, younger ages should be viewed with care because of Pb-loss since the Pampean realm (both upper and lower plates) underwent significant reheating during the Famatinian orogeny (490–440 Ma) (Rapela et al., 1998). In any case, ages younger than ca. 520 Ma quoted by other authors (see Table 3) are taken here as resulting from post-Pampean orogeny processes.

Hf isotope composition of zircon (Table 3) is so far only available for the Tastil pluton, a dacitic porphyry in NW Argentina (Hauser et al., 2011) and the small granite body of Guasayán pluton (533 ± 4 Ma, Dahlquist et al., 2016) (Fig. 2). The εHf_t values range from +1.1 to -6.9 and the Hf T_{DM} model ages are in the range 1.3 to 1.7 Ga. Nd-isotope compositions are available for the Sierra Grande–Ambargasta batholith and for NW Argentina (Hauser et al., 2011; Iannizzotto et al., 2013) (Table 3): εNd_t values range between ca. -2 and -10 and the Nd T_{DM} model ages 1.4 to 2.0 Ga. The Nd model ages of the Puncoviscana Formation represent the crustal age of the upper plate, ranging from ca. 1.7 to 2.0 Ga (most are ca. 1.7 Ga) (Bock et al., 2000; Lucassen et al., 2000; Rapela et al., 1998). This range of Nd model ages is compatible with that found for the Pampean magmatic arc (1.5 – 2.0 Ga), perhaps with the additional effect of a minor juvenile component.

4.3 Structural features and metamorphic conditions in the upper plate

Strain in the upper plate was distributed between sub-domains with upright to west-vergent folds with axial-planar foliation and dextral NE–SW trending mylonite corridors such as the large Sauce Punco shear zone in Sierra Norte (Martino, 1999; von Gosen and Prozzi, 2010). Magma emplacement was largely focused along shear zones and plutons are both earlier and later with respect to shearing (e.g., Iannizzotto et al., 2013). This domain continues in North Western Argentina where metasedimentary rocks of the Puncoviscana Formation are abundant. Metamorphism here was very low-grade to low-grade and was coeval with folding and shearing (von Gosen & Prozzi, 2010). *P-T* conditions for $M1_{up}$ (up indicating upper plate) were recently rated at 8 – 9 kbar and 240 – 300ºC, i.e., a high *P/T* type of metamorphism (Do Campo et al., 2013); this was followed by isothermal decompression through 275 – 350 ºC and 0.7 – 3.0 kbar ($M2_{up}$) (Table 4 ; Fig. 5). Rapid uplift and erosion resulted in volcanism with ages indistinguishable within error from those of the underlying plutonic rocks (see below).

In the northern Sierras Pampeanas a discordant fanning foliation ($S1_{up}$) is associated with ubiquitous metric/decametric-scale chevron folds that have subvertical axial planes. These structures resulted from a single major deformation episode (Piñán-Llamas & Simpson, 2006). Tectonic foliation overprints pressure-solution cleavage and banding is interpreted as a compaction-related primary foliation (Piñán-Llamas & Simpson, 2009).

Outcrops of low-grade Puncoviscana Formation are also found to the west of the high-grade domain as isolated lenses separated from the latter domain by shear zones and faults. Such is the case of the Los Túneles Phyllites in the westernmost Sierra Grande de Córdoba (Martino et al., 2003; Escayola et al., 2007). The phyllites underwent Pampean metamorphism at 525 ± 18 Ma (whole rock Rb-Sr isochron, MSWD = 25; Rapela et al., 1998).

5. The High-Grade Domain

5.1 The Sierras de Córdoba Metasedimentary Series

This domain consists for the most part of high-grade metasedimentary migmatitic gneisses of the Puncoviscana Formation and migmatite gneisses, marbles and calc-silicate rocks that were recently included within the Sierras de Cordoba Metasedimentary Series by Murra et al. (2016). Late Ediacaran to Early Cambrian ages were indirectly determined for the latter from the Sr-isotope composition of chemically screened samples of almost pure calcite marble and were further constrained by C- and O-isotope data and U-Pb SHRIMP detrital zircon ages of an interbedded paragneiss (Murra et al., 2016). The marbles show two groups of $^{87}Sr/^{86}Sr$ ratios (ca. 0.7075 and 0.7085), inferred as corresponding to Early Ediacaran (620 to 635 Ma) and Late Ediacaran to Early Cambrian, respectively. Interbedded migmatitic gneisses have detrital zircon age patterns that show a group of ages between ca. 700 and 1650 Ma; there is a notable peak at ca.1190 Ma (range1100–1250 Ma), and an older population with ages of ca. 1950 – 2060 Ma (in contrast, the Puncoviscana Formation lacks ages between ca.1.2 and 1.65 Ga but has a characteristic Neoproterozoic peak between 570 and 680 Ma that is missing here). The Sierras de Cordoba Metasedimentary Series pattern instead resembles those of rocks from the Western Sierras Pampeanas, such as the Difunta Correa Sedimentary Sequence and the Ancaján Series, which are also interbedded with Ediacaran marbles (Ramacciotti et al., 2015; Rapela et al., 2016). This correlation implies that all belong to an originally extensive sedimentary cover to Mesoproterozoic (Grenvillian *s.l.*) basement (Murra et al., 2016). The source of these metasedimentary series has been ascribed to the Mesoproterozoic (and Paleoproterozoic) basement of the Western Sierras Pampeanas and further west (Laurentia?) (Ramacciotti et al., 2015; Rapela et al., 2016). In contrast, the Late Ediacaran to Early Cambrian Puncoviscana Formation of NW Argentina, northern Sierra Chica and Sierra Norte, is thought to have had sedimentary input from Gondwana continental sources (Murra et al., 2016).

5.2 The ophiolite

The mafic–ultramafic complex consists of meta-peridotite, meta-pyroxenite meta-gabbro, massive chromitite, and minor leucocratic rocks that were collectively interpreted as an ophiolite complex (upper mantle and oceanic crust) and hence a relict suture (e.g., Ramos et al., 2000, Escayola et al., 2007, Proenza et al., 2008, Martino et al., 2010, among other). The ophiolite probably formed in a supra-subduction setting on the basis of basalt chemistry that ranges from N-MORB to OIB; it has yielded a Sm-Nd age of 647 ± 77 Ma (Escayola et al., 2007). Ophiolite remnants are found in the inner part of the Pampean orogenic wedge now exposed in the Sierras de Córdoba. They underwent Pampean high-grade metamorphism and deformation (e.g., Martino et al., 2010; Tibaldi et al., 2008) that overprinted the obduction-related structures that preceded continental collision. Remarkably the ophiolite outcrops are often spatially associated with marbles, calc-silicate rocks and gneisses of the Sierras de Córdoba Metasedimentary Series (Fig. 1, Kraemer et al., 1995; Martino, 2003), strengthening the idea that the ophiolite was obducted onto the carbonate platform at the western margin of the Clymene Ocean in a similar manner to the Oman ophiolite (Escayola et al., 2007). The ophiolite was then involved in the Pampean orogenic wedge and imbricated with slivers of rocks of the upper plate (Puncoviscana Formation).

N-MORB type tholeiitic amphibolites are also found as disrupted bodies closely associated with the Sierras de Cordoba Metasedimentary Series (Rapela et al., 1998). The age and significance of these rocks, whether related to the ophiolite or to early processes along the active margin remain unknown.

5.3 Structural Geology

The large scale structural geology of the high-grade Pampean domain is poorly known. Early work by Martino et al. (1995) and Baldo et al. (1996) recognized a regional shallow-dipping axial planar foliation (S2) related to west-verging isoclinal folds (D2) and coeval with high-grade intermediate *P/T* metamorphism (M1) which peaked at 530 –520 Ma (Rapela et al., 1998; Murra et al., 2016). After allowing for block tilting during Cenozoic compression, the regional foliation is almost flat-lying. An older foliation is locally preserved in rafts in migmatitic granitoids and is interpreted as an S1 foliation re-folded and transposed by S2. Continued deformation led to shearing along discrete zones (D3 according to Martino et al. (2010). D2 folding and D3 shearing was accompanied by crustal thickening, with recorded P values of up to 7 – 9 kbar under upper amphibolite to granulite facies conditions (see below; Rapela et al., 1998; Otamendi et al., 1999; Martino et al., 2010). Mafic and ultramafic bodies are often aligned with bands of marble, calc-silicate hornfelses and gneisses of the Sierra de Córdoba Metasedimentary Series, and with the S2 foliation. They were formerly considered to define two regional strips (e.g., Kramer et al., 1995) but this interpretation has been recently challenged, i.e., mafic–ultramafic bodies are repeated by folding and shearing and there is no regular regional pattern (Fig. 3) (Martino et al., 2010). These bodies underwent M2 metamorphism (see below) (Rapela et al., 1998; Tibaldi et al., 2008; Anzil et al., 2012). The high-grade domain further underwent uplift at ca. 525–520 Ma, accompanied by strongly peraluminous magmatism such as the El Pilón cordieritite (see below). The latter complex resulted from magma displacement from ca. 6 to 3.7 kbar at 523 ± 2 Ma (Rapela et al., 1998, 2002) favoured by regional decompression, probably during mantled-gneiss dome formation. Structures related to uplift of the Pampean orogen are as yet poorly known: one example is the eastern side of the large Guacha Corral shear zone in the Santa Rosa area, where dextral movement combined with extension in narrow mylonitic belts was recorded by Martino et al. (1994; Fig.5).

Foliation S2 in the lower plate may be correlated with the S1 foliation dominant in the low-grade domain of the upper plate. However the thermal peak may have been attained later in the high-grade domain (527 ± 3 Ma) than in the low-grade upper plate domain (530 ± 4 Ma; minimum age).

5.4 Regional metamorphism and anatectic magmatism

Metamorphism in the Pampean orogen high-grade domain has been the subject of work by many authors (Table 4). Most have focused on metapelitic gneisses and migmatites and a few on mafic granulites (metabasites) (Table 4). Migmatites are represented by metatexites and diatexites mostly consisting of garnet–biotite–plagioclase–quartz–K feldspar ± sillimanite ± cordierite (see Guereschi and Martino, 2014). The metabasites consist of plagioclase–orthopyroxene–biotite–quartz–garnet–amphibole (e.g., Rapela et al., 1998; Otamendi et al., 2005). There is some evidence of an older $M1_{lp}$ metamorphic event (lp = lower plate) in the form of aligned inclusions in garnet, but its *P*-*T* conditions are unknown. $M2_{lp}$ corresponds to the high-grade event under granulite-facies conditions. Garnet ± cordierite migmatites from central to northern and eastern Sierras de Córdoba yield *P*-*T* estimates (conventional thermobarometry) of 750–850 ºC for peak *T* and 7–8 kbar for the maximum recorded *P* (Fig. 5) (Baldo et al., 1996; Rapela et al., 1998; Otamendi et al., 1999, 2005). However N–S longitudinal *P*-*T* gradients probably existed.

The ages obtained for the peak of metamorphism are apparently younger that those in the low-grade domain, between 530 and 520 Ma (Lyons et al., 1997; Rapela et al., 1998 among others). A recent U-Pb SHRIMP zircon age of 527 ± 3 Ma was reported by Murra et al. (2016). The high *P* values for the high-grade granulite-facies domain suggest that underthrusting played a role and that an orogenic root probably formed within a few million years.

Metamorphism-related S-type peraluminous magmatism (Table 3) is represented by several granitic complexes such as El Pilón (ASI = 1.08–1.40) dated at 523 ± 2 Ma (conventional U-Pb, Rapela et al., 1998, 2000), the San Carlos migmatitic massif (529 ± 3 Ma; Escayola et al., 2007), Suya Taco (520 ± 3 Ma, U-Pb on monazite, Tibaldi et al., 2008). These ages suggest that peraluminous magmatism took place after the I-type magmatism and was related to the high-grade $M2_{lp}$ event. Nd-isotope data from the El Pilón and San Carlos complexes yielded consistent values of εNd_t of ca. -5.7 and Nd model ages (T_{DM}) of 1.6–1.7 Ga, compatible with derivation by melting of fertile supracrustal rocks.

The conditions of retrograde metamorphism are poorly constrained (Table 4; Fig. 5). After intrusion at a depth of ca. 6 kbar (and ca. 780ºC), the El Pilón granitic complex re-equilibrated with the host migmatites at T = 555 ± 50ºC and P = 3.3 ± 0.3 kbar (Rapela et al., 2002). These values imply a gross uplift rate of ca. 2.4 mm/a accompanied by cooling of 280 – 180ºC subject to analytical errors. In Fig. 5, an estimate uplift *P-T* path has been drawn based on data from several sources.

Evidence of thermal events older than $M2_{lp}$ is contentious. Metamorphism related to early ridge subduction was invoked by Simpson et al. (2003), Gromet et al. (2005) and Guereschi & Martino (2008) to explain the high temperatures attained in the high-grade domain but this mechanism is not supported by the P values of up to 8 kbar referred to above and chronological constraints – in fact the peak of metamorphism (M2) is younger than the magmatic arc in the Pampean belt. Siegesmund et al. (2010) showed that some high-grade gneisses and diatexites from the Eastern Sierras Pampeanas contain zircon grains with low U/Th overgrowths interpreted as of metamorphic origin and dated at ca. 550–540 Ma. If confirmed by further research this cryptic metamorphism could correspond to the early (phase I) S-type granitic magmatism in the Saldania Belt (see below).

6. Discussion

6.1 The geodynamic framework

Current models on the evolution of the Pampean belt fall into one of two types. According to Escayola et al. (2007) and Ramos et al. (2014) an ocean existed on the eastern side of a Grenvillian terrane called Pampia before 0.7 Ga. The Puncoviscana Formation began to form on this margin, receiving zircon from the west with ages between 1.0 and 1.2 Ga. Subsequent intra-oceanic west-directed subduction started at 700–600 Ma, with development of a magmatic island-arc and a back-arc region between this and Pampia. The arc supplied zircon of 0.6–0.7 Ga to the adjacent continental margin thus explaining, according to these authors, the typical Puncoviscana detrital zircon pattern. A consequence of this model is that the Puncoviscana Formation formed in part after 1.1 Ga and before 700 Ma, and in part during and after the island arc activity when zircon from Pampia and the arc mingled in the sediment. However no sedimentary rocks have ever been found from the Puncoviscana Formation, i.e. with the characteristic detrital zircon age peak between 950 and 1100 Ma (Rapela et al., 2016), as old as Tonian (> ca. 720 Ma), nor have rocks been found with a single peak of 0.6–

0.7 Ga, as might be expected close to the island arc source area. All the evidence from detrital zircon ages and paleontology (see above) suggest that the Puncoviscana Formation is younger than 570 Ma and that the sources were in the east. Moreover most zircon from the Puncoviscana Formation between 560 and 700 Ma have negative εHf_t values suggesting a continental provenance. The next step in this type of model is obduction of the island arc over the Rio de la Plata craton at 600–580 Ma, accompanied by a flip of the subduction zone from west-dipping to east-dipping, for which there is no direct evidence. The Rio de la Plata craton was not affected by the Pampean orogeny. Collision between Pampia and the Rio de la Plata craton took place between 580–540 Ma and was followed by decompression melting and metamorphism between 540 and 515 Ma. However the thickening-related M2 metamorphism in the high-grade zone that yielded P values of up to ca. 8 kbar took between 530 and 525 Ma (see above). Furthermore no account is taken in this model of sedimentary rocks such as the late Neoproterozoic Sierras de Córdoba Metasedimentary Series of age (Murra et al., 2016), for which Grenville-age (0.95–1.45 Ga) zircon was sourced from the Western Sierras Pampeanas basement and yet further west from Laurentia (Rapela et al., 2016).

In another view the collisional Pampean orogeny resulted from the closure of the intervening Clymene Ocean hypothesized on the basis of paleomagnetic evidence as located between Amazonia and Laurentia on one side and West Gondwana cratons such as Rio de la Plata and Kalahari on the other (Trindade et al., 2006; Rapela et al., 2007). Colliding blocks were considered either para-autochthonous to the Gondwana margin (Rapela et al., 1998; Ramos el al., 1988, 2010) or allochthonous (Rapela et al., 2007). The latter case invokes a Paleoproterozoic block formerly attached to Laurentia and Amazonia during Mesoproterozoic (Grenvillian *s.l.*) continental collisions. This block was named MARA after three of its alleged outcrops: Sierra de **M**AZ in the Western Sierras Pampeanas, **A**requipa (Peru) and **R**io **A**pa (southern Brazil) (Casquet et al., 2012). This block along with Amazonia rifted away from Laurentia during the Early Cambrian opening of the Iapetus Ocean (Dalziel, 1997; Casquet et al., 2012; Rapela et al., 2016). At the same time oblique right-lateral subduction of the Clymene Ocean started under the West Gondwana cratons, ending with right-lateral collision of MARA+Amazonia and consequent development of the Pampean, Paraguay and Araguaia collisional belts (Fig. 1) (Casquet et al., 2012; Rapela et al., 2016). In this model the mainly turbiditic sediments of the Puncoviscana Formation were derived from Gondwana sources in the east (in the present-day sense) while the Sierras de

Córdoba Metasedimentary Series was sourced from the west, i.e. from the MARA block and from Laurentia prior to break-up. Evidence for this interpretation is consistent with the U-Pb detrital zircon evidence (see above) and the Early Ediacaran and Late Ediacaran to Early Cambrian age of marbles (Murra et al., 2016). In this model the Pampean orogen was displaced right-laterally during subduction and collision and attained its present position relative to the Rio de la Plata craton after orogenic uplift.

6.2 Comparison of the Pampean Belt with the Saldanian Belt of South Africa

The Saldania Belt of South Africa consists of scattered outcrops and inliers of Ediacaran to Early Cambrian low-grade metasedimentary rocks and Cambrian granitoids (the Cape Granite Suite): these are unconformably overlain by Gondwanan Permo-Triassic sedimentary rocks of the Cape Fold Belt. The Saldanian orogeny took place in the Early Cambrian (pre-520 Ma) (Rozendaal et al., 1999; Curtis, 2001; Chemale et al., 2011). The metasedimentary rocks are partly a para-autochthonous cover to the Late Mesoproterozoic (Grenvillian) Natal–Namaqua basement in the northeast (i.e., the Boland Group), but most have been assigned to the Malmesbury Group (the allochthonous Malmesbury Terrane) which has unknown relationships to the basement (Rozendaal et al., 1999; Frimmel et al., 2013).

The Malmesbury Group turbidites (and the Kangoo Caves Group in eastern inliers, Naidoo et al., 2013) bear a strong resemblance to the Puncoviscana Formation of Argentina. In particular the detrital U-Pb zircon age pattern shows many similarities. Armstrong et al. (1998) found in one turbidite a bimodal distribution of ages peaking at 900–1050 Ma and 575–700 Ma, with a youngest zircon of 560 Ma. Frimmel et al. (2013) analyzed several rocks (one argillite and three greywackes) from the Malmesbury Group. The youngest concordant zircon had ages of 550–600 Ma in three cases, and sedimentation age was constrained to between 557 and 552 Ma (Late Ediacaran). Most grains are Neoproterozoic with many minor peaks. One main group of ages can be recognized as ca. 580–700 Ma and a second group as 700–960 Ma. Grenville-age detrital grains constitute a smaller group between 960 and 1100 Ma. Very few grains of ca. 1.2, 1.3, 2.1 and 2.0 Ga and few Archean grains of ca. 2.7 Ga were also found (Fig. 4). As noted above, the U-Pb zircon age pattern of the Puncoviscana Formation is bimodal with major peaks at 1100–960 Ma and 680–570 Ma and few grains of 1.7-2.0 Ga and ca. 2.6 Ga (Rapela et al., 2007, 2016; Adams et al., 2008;

Hauser et al., 2011). In Fig. 4, we compare the zircon age pattern from the Tygerberg Formation of the Malmesbury Group (HFS08-06; Frimmel et al., 2013) with the Puncoviscana sample TLT-2069. Both age patterns have the same two main groups – one between 560 and 690 Ma and a Grenvillian group between 960 and 1100 Ma. A small group of grains between 700 and 900 Ma in the Malmesbury sample is poorly represented in the Puncoviscana sample (two grains only of 777 and 890 Ma), but is more significant in other Puncoviscana samples (e.g. sample RCX-1; Adams et al., 2008). The few Mesoproterozoic ages (1.2 Ga in TLT-2069 and 1.3–1.35 Ga in the Tygerberg Formation) are not coincident, although the 1.2 Ga peak is recognized in other samples from the Malmesbury Group. Significantly, the Puncoviscana Formation and the Malmesbury Group share one important feature feature, i.e., the absence of zircon with the Rio de la Plata craton ages between 2.02 and 2.26 Ga (Rapela et al., 2007).

The comparison can be extended to the Hf composition of detrital zircon (Fig. 6). We have chosen the time interval 570–680 Ma, corresponding to the Brasiliano–Pan-African orogeny, for this comparison, using data from North Western Argentina (Hauser et al., 2011; Augustsson et al., 2016) and sample TLT-2069 (Table 2) along with data from the Saldania Belt (Frimmel et al., 2013). The εHf_t values of the latter (+4.7 to -19.2) are for the most part negative and coincident with those of the Puncoviscana Formation (+ 5.8 to -18.4), suggesting Brasiliano–Pan African sources for both (Frimmel et al., 2013). However sample TLT-2069 yields mainly positive εHf_t values between + 0.7 and +12.0 and the source of these zircon grains must be different, perhaps in the Neoproterozoic East African–Antarctic orogen (EAAO) as formerly suggested by Rapela et al. (2007, 2016). In Fig. 6, zircon within the chosen age range from the Mecuburí Formation of NE Mozambique (Thomas et al., 2010) yielded mostly positive εHf_t values in the range +9.9 to -3.8 similar to those of sample TLT-2069. The EAAO was in fact a major source of molasse sediments to the southern Kalahari continental margin in Late Neoproterozoic–Early Paleozoic times (Jacobs & Thomas, 2004).

We conclude that the Puncoviscana Formation embraces Ediacaran turbidite sediments derived from different continental sources with a detrital zircon age pattern resembling that of the Malmesbury Group of the Saldania Belt. This pattern is in fact typical of southern Gondwanan sources in general (Kristoffersen et al., 2016). One source probably was in the Brasiliano–Pan-African orogenic belt that resulted from the closure of the Adamastor Ocean (Frimmel et al., 2013; Rapela et al., 2011) and references therein); another source probably was in the EAAO (Rapela et al., 2007, 2016).

Moreover, tectonic structure of the Pampean Puncoviscana Formation and the Saldanian Malmesbury Group are very similar. Both are simple consisting of essentially uprigh folds with axial planar foliation (Piñan-Llamas & Simpson, 2006; Rozeendal et al. 1994; Buggisch et al., 2010; Rowe et al., 2010).

6.3 The Cape Granite Suite and the Pampean magmatic arc

The Pan-African Cape Granite Suite (CGS) is predominantly composed of 552 to 533 Ma S- and I-type granitoids that formed during the Saldanian orogeny (Table 3). Scheepers (1995, and references therein) and da Silva et al. (2000) divided the orogenic magmatism into two main episodes: phase I, S-type (ASI = 0.98 to 1.66) synorogenic granites, and phase II, late-orogenic calc-alkaline I-type (ASI = 0.86 to 1.08) granites. Phase I is bracketed between 552 ± 4 Ma and 533 ± 2 Ma (with many ages close to 540 Ma), and Phase 2 is dated at 536 ± 5 Ma (da Silva et al., 2000; Scheepers and Armstrong, 2002; Chemale et al., 2011; Villaros et al., 2012). Minor post-orogenic S-type granites and volcanics and A-type plutonic bodies intruded granites of both phases as well as the Malmesbury Group metasediments between 527 ± 8 Ma and 510 ± 4 Ma (Scheepers and Armstrong, 2002; Chemale et al., 2011). The post-orogenic S-type granites (ca. 527 Ma) may have been emplaced in a different geodynamic setting to that invoked for Phase I. The youngest peraluminous magmatism was the extrusion of S-type ignimbrites at 516± 3 Ma (Scheepers and Poujol 2002). A-type post-orogenic granitoids and the late ignimbrites will not be dealt with further here. Phase I S-type granites occur inward, in the Tygerberg terrane and are usually deformed; Phase II I-type granites occur outward, i.e., towards the craton, and are generally undeformed (Chemale et al., 2011).

Nd- and Hf isotope data from the Cape Granite Suite have been reported by Chemale et al. (2011) (Table 3 & 5). The εNd_t values of S-type granitoids (at ca. 550Ma) range from -3.2 to -4.9 and the Nd model ages (T_{DM}) from 1.5 to 1.9 Ga. Late peraluminous granitoids yield εNd_t values of -5.9 and T_{DM} = 1.7 Ga. The εNd_t values of high K-calc-alkaline I-type granitoids range from -1.4 to -3.9 and the T_{DM} ages from 1.0 to 2.0 Ga . The εNd_t values of the I-type granitoids are in general higher than for the S-type granites suggesting a larger juvenile component in magma evolution (Chemale et al., 2011). Hf isotope compositions of zircon have been reported from S-type granites of the Cape Granite Suite by Villaros et al. (2012) and Farina et al. (2014) (Table 3). Magmatic zircon and magmatic overgrowths show a restricted range of εHf_t between -8.6 and +1.5 (Villaros et al., 2012), interpreted as indicating that the granitic magma resulted from the anatexis of Malmesbury Group metasedimentary rocks, thus confirming the S-type signature of these granites.

A comparison between the I- and S-type Cape Granite suite and the Pampean magmatic arc rocks is shown in Table 5. I-type magmatism was roughly coeval in the two belts with most ages mainly between 535 and 525 Ma and post-orogenic peraluminous S-type granites of ca 527–524 Ma in the Cape Granite suite (Chemale et al., 2011) are coeval with those of the high-grade domain in the Sierras de Córdoba (e.g., ca. 523 Ma; Rapela et al., 1998). Nd- and Hf isotope data of I-type granitoids are slightly more juvenile in the Cape Granite Suite than in the Sierras Pampeanas and NW Argentina.

Phase I S-type granites between ca. 552 and 533 Ma (most values ca. 540Ma) represent an earlier event in the Saldania Belt. Anatexis of Malmesbury Group sediments apparently involved fast heating (ca. 30º/Ma) to ca. 850ºC because of the short time span between the youngest detrital zircon and the age of magmatism (Farina et al., 2014). This older event apparently preceded the formation of the I-type Cordilleran magmatic arc and has not been recognized in Argentina; although Siegesmund et al. (2010) interpreted that some high-grade gneisses and diatexites in the Eastern Sierras Pampeanas record a metamorphism at ca. 550-540 Ma. The significance of an early, high-T metamorphism and related S-type magmatism (in the Saldania Belt, at least) remains unknown. In this regard, both the hypothesis of a ridge-subduction stage proposed by Gromet & Simpson (2003) for the Pampean orogeny and of an extension of the accretionary prism previous to continental collision remain potential options.

7. A paleogeographic and dynamic model

The Pampean orogeny took place between ca. 545 (and may be as old as 550Ma) and 520 Ma and involved early subduction of the Clymene Ocean, development of a magmatic arc and final continental collision with preservation of a suture recognized in the Sierras de Córdoba. The similarities shown here between the sedimentary rocks of the Malmesbury Terrane (Frimmel et al., 2013) and those of the Puncoviscana Formation strengthens the hypothesis already suggested by others that both formations were laid down in the same sedimentary basin probably along the southern margin of the Kalahari craton (Fig. 7). The U-Pb detrital zircon age patterns are quite similar and typical of Gondwana provenance. Grenville-age zircon of ca. 1.0 Ga can be sourced in the Natal–Namaqua belt of the southern Kalahari. Moreover, Hf isotope composition of the Cryogenian to Ediacaran zircon is compatible with sediment sources in the slightly older (650–570 Ma) Brasiliano–Pan-African orogen and EAAO, as proposed by Rapela et al. (2016). In the Saldania Belt the time of sedimentation is bracketed between 557 and 552 Ma in the Malmesbury terrane and between 609 and 532 Ma in the outermost autochthonous Boland Zone (Frimmel et al., 2013). In the first case the lower age corresponds to the older Cape Granite suite plutons that intruded the Malmesbury Group. In NW Argentina deposition of the Puncoviscana Formation can less precisely be bracketed between 570 and ca. 537 Ma (Escayola et al., 2011; Casquet et al., 2012). In consequence both the Puncoviscana and the Malmesbury Group sediments were laid down between 570 Ma (probably after 555 Ma) and ca. 537 Ma. However while in the Saldania Belt sedimentation of the Malmesbury Group preceded the Cape Granite suite, i.e., it was older than 552 Ma, in the Puncoviscana case the youngest sediments were deposited at the same time as I-type arc magmatism in the upper plate (Escayola et al., 2011).

Remarkably both sedimentary series lack zircon with Rio de la Plata craton ages between 2.02 and 2.26 Ga (Rapela et al., 2007), implying that the craton was probably not adjacent to the sedimentary basin between 570 and 537 Ma. Since Pampean folding and I-type magmatism in the upper plate were synchronous with right-lateral shearing (Iannizzotto et al., 2013; Van Gosen & Prozzi, 2010), this could mean that the Rio de la Plata craton only reached its present relative position after the main tectonothermal event (530-520 Ma) (Verdecchia et al., 2011). Displacement was focused along the Córdoba Fault, a crust-scale strike-slip and geophysical discontinuity correlated with the Transbrasiliano Lineament (Rapela et al., 2007; Ramé & Miró, 2011) (Figs. 1 & 2) and the displacement juxtaposed the high-grade zone of the Pampean orogen with the Rio de la Plata craton. The latter did not undergo Pampean metamorphism thus implying that the lateral displacement was large and final docking was younger than exhumation of the high-grade domain at ca. 520 Ma.

8. Conclusions

Combining the evidence given above we hypothesize the following evolution for the Pampean/Saldanian orogeny summarized in Fig. 7:

a) The Punscoviscana Formation and the Malmesbury group were deposited on a continental margin to the south of the Kalahari craton between 570 and 537 Ma and 570 and 552 Ma respectively. Both were probably separated from the autochthonous Boland terrane through the Piketberg–Wellington Fault (Frimmel et al., 2013). Sediments came from sources in the Brasiliano–Pan-African orogenic belt and the EAAO.
b) Magmatism started at ca. 552 Ma with intrusion of the Cape Granite Suite S-type granitoids. A cordilleran I-type magmatic arc formed afterwards, between 537 and 528 Ma (most ages are ca. 530Ma), coeval with Puncoviscana sedimentation. Magmatism evolved from S-type to I-type and then back to S-type again over time. The later S-type magmatism (between 530 and 520 Ma) is only recognized in the Sierras Pampeanas and NW Argentina. Fast uplift took place by the end of the I-type magmatism producing dacitic volcanism on top of almost coeval, eroded plutonic rocks of the magmatic arc.
c) Subduction was oblique with strain distributed in the upper plate. Right-lateral shear-zones controlled magma emplacement. Folding in the upper plate during the subduction stage produced upright folds. Piñán-Llamas and Simpson (2006) invoke a tectonic model that involves the buttressing of scraped-off Puncoviscana Formation over the subducting slab. This model is compatible with our interpretation of the Puncoviscana Formation as formed in a forearc setting along the eastern margin of the Clymene Ocean.
d) Metamorphism during the subduction stage was of high *P/T* type as recognized in the Puncoviscana Formation as it evolved from passive margin sediment into a forearc accretionary prism. Age of this metamorphism remains unknown.
e) Continental collision started at ca. 530 Ma resulting in juxtaposition of the Puncoviscana Formation/Malmesbury Group upper plate against the MARA

continental block. The latter consisted of a Grenvillian basement with Laurentian affinities and a sedimentary cover of Ediacaran to Early Cambrian age (Murra et al., 2016).

f) The intermediate Clymene Ocean was consumed and relics of it were preserved defining a paleo-suture in the Sierras de Córdoba. The age of the oceanic crust was estimated as 647± 77 Ma by Escayola et al. (2007), which is compatible with a recent estimate of 620 - 635 Ma for the Early Ediacaran marbles of the sedimentary cover to the MARA block (Murra et al., 2016). The oceanic crust was probably overthrust (obducted) onto the platform of MARA before collision in a manner similar to obduction of the Oman ophiolite (Escayola et al., 2011).

g) Collision led to strong deformation and metamorphism between 530 and 520 Ma, i.e., younger than the I-type magmatic peak. Sedimentary rocks of the MARA platform and the Puncoviscana Formation, respectively on opposite sides of the suture, were folded and dragged down to as deep as 30 km (8 kbar) at temperatures of up to ca. 800ºC. This high-T domain is not found in the Saldania Belt because it was transferred to the Pampean belt in the Sierras Pampeanas. We further infer that this domain underwent uplift and detachment with respect to the low-grade domain by 525–520 Ma, along with strongly peraluminous S-type magmatism. The high-grade domain probably became a mantled gneiss-dome. The origin of heat remains elusive: because igneous rocks of that age are minor we suggest that crustal delamination or crustal foundering played a role.

h) Juxtaposition of the Rio de la Plata craton with the Pampean belt across the right-lateral Córdoba Fault took place after the high-grade domain was exhumed, i.e., after 520 Ma. The fault was slightly oblique with respect to axis of the Pampean orogen and probably played a major role in separating the Pampean orogen from the Saldania Belt. The timing of docking probably in the late Early to Late Cambrian remains to be precised.

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Figure Captions

Fig. 1. Schematic geological map of South America showing Paleoproterozoic to Archean cratons, Mesoproterozoic mobile belts, and the Neoproterozoic-to-early Cambrian orogens.

Fig. 2. Schematic geological map of the Sierras Pampeanas showing the Pampean belt (ca. 545–520 Ma) and the Ordovician accretionary-type Famatinian belt (490–440Ma). The inferred Pampean suture is indicated. Ruled decoration shows Pampean orogen reworked by the Famatinian orogeny. NWA is North Western Argentina. CP: Carapé Fault (Martino, 2003).

Fig. 3. Schematic cross-section of the Pampean orogen sandwiched between the Paleoproterozoic Rio de la Plata craton (RPC) and the Ordovician Famatinian orogen.

Fig. 4. Probability plot with histograms and TW plots of samples from the Puncoviscana Formation-equivalent Pocho Phyllites (TLT-2069) and the Malmesbury Group (HFS08-06; Frimmel et al., 2013) from the Saldanian orogen in South Africa.

Fig. 5. Plot of P-T conditions of Pampean metamorphism recovered from conventional thermobarometry by different authors (see footnote for references). M2 and M3 events in migmatites, gneisses and granulites from (a) Do Campo et al. (2013), (b, c) Rapela et al. (1998), (d, e) Rapela et al.(2002), (f, g, h) Otamendi et al.(1999), (i) Otamendi et al. (2005), (j) (Martino et al. (2010). M2 conditions are shown as white boxes and circles and black line, whereas grey boxes and black circles represent M_3 conditions. The broken lines join thermobarometric calculations made on the same metamorphic rocks. Where uncertainties are available they are indicated. The thick dashed curves embrace the probable clockwise metamorphic P-T path for both the low-grade high-P domain and the high-grade domain.

Fig. 6. Plot of εHft values *vs*. age of detrital zircon between 570 and 700 Ma from the Puncoviscana Formation of NW Argentina (Hauser et al., 2011; Augustsson et al., 2016), sample TLT-2069 from this work (Table 2), the Saldanian belt (Frimmel et al., 2013) and the Mecuburí Group of the East Africa-Antarctica Orogen (Thomas et al., 2010).

Fig. 7. Paleogeographic and dynamic model for the origin and evolution of the Pampean–Saldanian orogeny. Modified after Rapela et al. (2007). The orogeny resulted from oblique subduction of the Clymene Ocean beneath a continental margin with right-laterally displacement relative to the Kalahari and Rio de la Plata cratons. A) The margin was fed with turbidite sediments (Puncoviscana Formation and Malmesbury Group) derived from the erosion of the Neoproterozoic large East African-Antarctica Orogen in the east and the Brasiliano–Pan-African orogens in the west. Displacement was focused along an earlier continent-scale fault (probably a transform fault). B) The margin became active and magmatic arcs developed (S-type and I-type) between 552 and 530 Ma. Sedimentation of the Puncoviscana Formation continued in the forearc as an accretionary prism. C) Closure of the Clymene Ocean at ca. 530 Ma brought arc magmatism to an end and resulted in continental collision between MARA (that had formerly rifted away from Laurentia) and the active margin between ca. 530 and 520 Ma. High P/T metamorphism is recorded from the upper plate while intermediate to low P/T metamorphism took place in the lower plate. An obducted ophiolite in the Sierras Pampeanas is evidence for the continental suture. Collision took place along with continuous right-lateral displacement of the closed margin. The resulting transpressional orogen was westward vergent (westward and upright folds in the Pampean belt; upright to weakly westward folds in the Saldanian belt). D) The Pampean orogen records uplift between 525 and 520 Ma. Renewed right-lateral movement along the Córdoba Fault eventually juxtaposed the Rio de la Plata craton against the internal part of the Pampean orogen. This fault strikes at a low angle relative to the orogenic grain, suggesting that it played a major role in the detachment of the Pampean belt from the Saldanian belt.

Grain. spot	U (ppm)	Th (ppm)	Th/U	^{204}Pb/ ^{206}Pb	f_{206} %	Total ratios ^{238}U/ ^{206}Pb	±	^{206}Pb/ ^{207}Pb	±	Radiogenic Ratios ^{206}Pb/ ^{238}U	±	^{207}Pb/ ^{235}U	±	^{207}Pb/ ^{206}Pb	±	Ages (in Ma) ^{206}Pb/ ^{238}U	±	^{207}Pb/ ^{235}U	±	^{207}Pb/ ^{206}Pb	±	% Disc
1.1	217	107	0.49	#######	0.19	9.867	0.147	0.0620	0.0007	0.1012	0.0015					621	9					
2.1	436	95	0.22	#######	0.14	10.074	0.132	0.0616	0.0005	0.0991	0.0013					609	8					
3.1	215	115	0.54	#######	0.21	9.021	0.154	0.0639	0.0009	0.1106	0.0019					676	11					
3.2	272	186	0.68	-	0.07	10.065	0.133	0.0610	0.0010	0.0993	0.0013					610	8					
4.1	533	48	0.09	#######	0.00	5.950	0.074	0.0747	0.0005	0.1681	0.0021	1.730	0.026	0.0747	#####	1001	12	1020	10	1059	14	5
5.1	389	109	0.28	#######	0.18	5.770	0.081	0.0755	0.0010	0.1730	0.0024	1.766	0.041	0.0740	#####	1029	13	1033	15	1042	35	1
6.1	406	203	0.50	#######	0.33	5.721	0.070	0.0748	0.0006	0.1742	0.0021	1.730	0.031	0.0720	#####	1035	12	1020	12	987	24	-5
7.1	267	9	0.03	#######	0.07	9.211	0.142	0.0627	0.0010	0.1085	0.0017					664	10					
8.1	304	<1	<0.01	#######	0.09	10.473	0.142	0.0604	0.0014	0.0954	0.0013					587	8					
9.1	92	77	0.84	-	<0.01	2.186	0.038	0.1646	0.0013	0.4576	0.0080	10.393	0.209	0.1647	#####	2429	36	2470	19	2505	13	3
10.1	67	36	0.53	#######	0.42	10.933	0.333	0.0631	0.0019	0.0911	0.0028					562	16					
11.1	304	100	0.33	-	0.01	9.891	0.153	0.0608	0.0009	0.1011	0.0016					621	9					
12.1	485	57	0.12	#######	0.19	6.636	0.085	0.0703	0.0007	0.1504	0.0019	1.425	0.026	0.0687	#####	903	11	899	11	890	23	-1
13.1	501	108	0.22	#######	0.18	5.977	0.072	0.0743	0.0006	0.1670	0.0020	1.676	0.027	0.0728	#####	996	11	1000	10	1008	19	1
14.1	653	87	0.13	#######	0.07	6.289	0.073	0.0765	0.0005	0.1589	0.0018	1.662	0.024	0.0759	#####	951	10	994	9	1092	14	13
15.1	137	39	0.28	-	<0.01	5.547	0.091	0.0744	0.0010	0.1804	0.0030	1.861	0.042	0.0748	#####	1069	16	1067	15	1064	28	0
16.1	516	253	0.49	#######	0.03	9.723	0.120	0.0613	0.0005	0.1028	0.0013					631	7					
17.1	230	59	0.26	#######	0.10	6.237	0.097	0.0761	0.0010	0.1602	0.0025	1.661	0.038	0.0752	#####	958	14	994	15	1074	30	11
18.1	939	460	0.49	#######	0.32	9.166	0.319	0.0770	0.0005	0.1088	0.0038					666	22					
19.1	537	198	0.37	#######	3.11	17.094	0.800	0.0791	0.0015	0.0567	0.0027					355	16					
20.1	290	89	0.31	#######	0.03	3.165	0.055	0.1153	0.0011	0.3159	0.0055	5.010	0.105	0.1150	#####	1770	27	1821	18	1880	17	6
21.1	488	36	0.07	#######	0.04	5.520	0.072	0.0742	0.0007	0.1811	0.0024	1.845	0.031	0.0739	#####	1073	13	1062	11	1039	18	-3
21.2	275	49	0.18	#######	0.14	5.498	0.070	0.0741	0.0010	0.1816	0.0023	1.826	0.038	0.0729	#####	1076	13	1055	14	1011	31	-6
22.1	562	8	0.01	#######	0.05	7.293	0.095	0.0655	0.0005	0.1371	0.0018	1.230	0.019	0.0651	#####	828	10	814	9	777	15	-7
23.1	153	88	0.58	-	<0.01	5.458	0.085	0.0752	0.0007	0.1835	0.0029	1.929	0.045	0.0763	#####	1086	16	1091	16	1102	31	1
24.1	370	50	0.13	#######	0.00	10.040	0.140	0.0605	0.0005	0.0996	0.0014					612	8					
25.1	103	50	0.49	#######	0.45	5.659	0.085	0.0764	0.0012	0.1759	0.0027	1.762	0.060	0.0726	#####	1045	15	1031	22	1004	59	-4
26.1	69	24	0.34	-	0.52	9.407	0.374	0.0658	0.0019	0.1058	0.0042					648	25					
27.1	396	309	0.78	#######	2.67	14.800	0.520	0.0768	0.0017	0.0658	0.0023					411	14					
28.1	185	63	0.34	#######	0.01	5.352	0.082	0.0739	0.0009	0.1868	0.0028	1.901	0.040	0.0738	#####	1104	15	1081	14	1036	26	-7
29.1	674	401	0.59	#######	0.16	8.849	0.098	0.0641	0.0008	0.1128	0.0013					689	7					
30.1	708	81	0.11	#######	<0.01	9.036	0.108	0.0605	0.0005	0.1110	0.0013					678	8					
31.1	456	89	0.19	#######	0.06	9.547	0.139	0.0618	0.0010	0.1047	0.0015					642	9					
32.1	563	418	0.74	#######	0.03	9.812	0.127	0.0615	0.0005	0.1019	0.0013					625	8					
33.1	136	72	0.53	#######	0.39	5.582	0.102	0.0779	0.0008	0.1785	0.0033	1.836	0.046	0.0746	#####	1059	18	1059	17	1059	31	0
34.1	280	174	0.62	#######	0.11	6.046	0.236	0.0756	0.0012	0.1652	0.0064	1.702	0.075	0.0747	#####	986	36	1010	28	1061	32	7
35.1	344	128	0.37	#######	<0.01	9.326	0.120	0.0601	0.0012	0.1075	0.0014					658	8					
36.1	64	24	0.37	#######	0.28	6.035	0.109	0.0755	0.0015	0.1653	0.0030	1.666	0.054	0.0731	#####	986	17	996	21	1018	51	3
37.1	430	138	0.32	-	<0.01	10.162	0.126	0.0604	0.0006	0.0984	0.0012					605	7					
38.1	128	36	0.28	-	<0.01	5.950	0.097	0.0732	0.0010	0.1684	0.0027	1.738	0.041	0.0749	#####	1003	15	1023	15	1065	31	6
39.1	328	34	0.10	#######	0.04	6.148	0.089	0.0736	0.0006	0.1626	0.0024	1.642	0.029	0.0733	#####	971	13	987	11	1021	17	5
40.1	498	192	0.39	#######	0.09	5.841	0.080	0.0744	0.0006	0.1710	0.0023	1.736	0.030	0.0736	#####	1018	13	1022	11	1031	18	1
41.1	255	130	0.51	#######	0.36	6.420	0.101	0.0738	0.0007	0.1552	0.0024	1.516	0.034	0.0709	#####	930	14	937	14	953	28	2
42.1	13	3	0.19	#######	0.40	9.879	0.346	0.0640	0.0031	0.1008	0.0036					619	21					
43.1	314	99	0.31	#######	0.13	4.792	0.073	0.0829	0.0009	0.2084	0.0032	2.350	0.054	0.0818	#####	1221	17	1228	16	1240	30	2

Notes :

1. Uncertainties given at the one σ level.
2. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
3. For areas >800 Ma, correction for common Pb made using the measured ^{204}Pb/^{206}Pb ratio.
4. For areas <800 Ma, correction for common Pb made using the measured ^{238}U/^{206}Pb and ^{207}Pb/^{206}Pb ratios following Tera and Wasserburg (1972) as outlined in Compston *et al* . (1992).
5. For % Conc., 100% denotes a concordant analysis.

spot number	spot age Ma	±	$\delta^{18}O$‰	± 1σ	$^{176}Hf/^{177}Hf$	± 2σ	$^{176}Lu/^{177}Hf$	± 2σ	εHf(t)	± 2σ	T_{DM} Ga
1	621	9	7.1537	0.181	0.282626	0.000025	0.001139	0.000051	7.77	0.89	0.99
2	609	8	8.4290	0.180	0.282428	0.000013	0.000563	0.000002	0.74	0.45	1.43
3	676	11	7.2280	0.178	0.282438	0.000015	0.000376	0.000003	2.68	0.52	1.36
5	1029	13	6.6960	0.181	0.282343	0.000017	0.000809	0.000032	6.87	0.59	1.37
6	1035	12	6.8572	0.179	0.282393	0.000027	0.000912	0.000026	8.72	0.96	1.26
7	664	10	8.7694	0.177	0.282406	0.000015	0.000329	0.000012	1.27	0.53	1.44
13	996	11	9.5896	0.179	0.282522	0.000047	0.001510	0.000065	12.00	1.66	1.02
15	1069	16	5.7397	0.177	0.282415	0.000015	0.000853	0.000008	10.29	0.52	1.18
21	1073	13	10.3911	0.179	0.282440	0.000037	0.001433	0.000025	10.85	1.31	1.15
24	612	8	10.7095	0.185	0.282761	0.000035	0.001937	0.000085	12.04	1.26	0.71
28	1104	15	6.0949	0.179	0.282308	0.000021	0.000484	0.000011	7.55	0.73	1.39
30	678	8	8.5188	0.179	0.282426	0.000018	0.000695	0.000034	2.12	0.62	1.40
32	625	8	6.5692	0.178	0.282474	0.000026	0.000724	0.000020	2.66	0.94	1.32
35	658	8	7.4773	0.184	0.282589	0.000028	0.001082	0.000057	7.30	0.98	1.05

Table 2. O- and Hf-isotope composition of zircon spots from sample TLT-2069

Age (Ma)	method	Lithology	Geological Unit	εHf	εNd	Hf T_{DM}	Nd T_{DM}	References
			Eastern Cordillera (NOA)					
536 ± 5	1	Rhyolitic tuff	Puncoviscana Formation					Escayola et al. 2011
523 ± 5	1	Granodiorite	Cañani batholith					"
526 ± 13	1	Dacite	"					Hongn et al. 2010
534 ± 7	2	Gray granodiorite	"	+1.1 to -6.9	-5.1/-9.8	1.45	1.57/2.0	Hauser et al. 2011
541 ± 4	2	Red granitic facies	"		-5.0			"
523 ± 5	2	Porphyritic dacite	"	+0.4 to -3.8	-4.4 /-4.7	1.32/1.54	1.50	"
533 ± 2	2	Granite porphyry	Granite dykes					"
			Sierras Pampeanas of Córdoba and Guasayán					
537 ± 4	1	Granodiorite (Hbl+Bt)	Sierra Norte-Ambargasta batholith		-5.8		1.69	Iannizzotto et al. 2013
537 #	3	I-type granitoids	"		-1.8/-5.4		1.39/1.66	"
530 ± 4	1	Granite	"		-5.5		1.67	"
535 ± 5	1	Metarhyolite	"					Von Gosen et al. 2014
534 ± 5	1	Granite porphyry	"					"
533 ± 4	1	Granite mylonite	"					"
531 ± 4	1	Dacitic phorphyry	"					"
530 ± 4	1	Granite	"					"
523 ± 5	1	Rhyolite to dacite	"					"
521 ± 4	1	Granite	"					"
519 ± 4	1	Rhyolite to dacite	"					"
533 ± 12	1	Porphyritic tonalite gneiss	"					Siegesmund et al. 2010
533 ± 2	1	Metaluminus O-gneiss	"		-5.8		1.7	Rapela et al. 1998
529 ± 2		Hb-Bt-Granodiorite			-4.3		1.6	"
532 ± 2	1	Dacite ??	"					Leal et al. 2003
512 ± 4	1	Dacite ????	"					"
515 ± 4	1	Granite	"					Stuart-Smidth et al. 1999
540 #	3	I-type granitoids	"		-7.9		2.09	Escayola et al 2007
528 ± 2	1	Granodiorite	Sierra Norte-Ambargasta batholith? (Ascochinga unit)		-5.0		1.62	Rapela et al. 1998
527	1	Peraluminus granite	Pichanas					Lyons et al.997
ca. 548	1	S-type porphyritic granite	El Pilón granite complex					Stuart-Smidth et al. 1999
523 ± 2	1	S-type granite	"		-5.6		1.69	Rapela et al. 1998
520 ± 3	1	Anatectic granite (U-Pb in Mo)	Suya Taco igneous complex					Tibaldi et al. 2008
529 ± 3.4	1	S-type granite	Juan XXIII pluton		-5.7		1.6/1.7	Escayola et al. 2007; Rapela et al. 1998
533 ± 4	2	Porphyritic granite	Guasayan pluton	-0.12 to -4.76		1.45 to 1.74		Dahlquist et al. 2016
			Saldania belt. Cape Granite Suite					

510-523	1	A-type syenogranite	Darling batholith		+5.1		0.67	Chemale et al. 2011
524.2 ± 8.1	2	Syenite (A-type granite)	"		-3.66		0.76	"
547 ± 6	1	S-type granite	"		-3.5		1.56	da Silva et al. 2000
527.5 ± 8.2	2	Granite	George pluton		-5.8		1.71	Chemale et al. 2011
538.2 ± 1.9	2	Granodiorite	Peninsula batholith	-10.7/-2.3		1.39/1.71		Villaros et al. 2012
532.7 ± 1.9	2	Granite	"					"
536.2 ± 2.4	2	S-type microgranodioritic enclave	"	-6.3/ +0.7		1.24/1.60		"
538.3 ± 1.5	2	Granodiorite	"	-1.5/+2.1		1.19/1.52		"
537.8 ± 1.6	2	S-type granite	"	-7.6/+1.2		1.10/2.16		Farina et al. 2014
536 ± 5	1	I-type granitoids	Robertson pluton		-3.1		1.63	da Silva et al. 2000
552 ± 4	1	S-type granite	Saldanha batholith					Scheepers and Armstrong 2002
540 ± 4	1	S-type granite	"					"
539 ± 4	1	S-type granite	"					"
515.5 ± 3	2	S-type igninmbrite	Postberg ignimbrites					Scheepers and Pujol 2002

Saldania belt. Cape Granite Suite (cont.)

Estimated age					εNd		Nd T_{DM}	
550-530	3	Granites	Maalgaten granite		-4.47		1.88	Chemale et al. 2011
550-531	3	"	Olifantskop granite		-3.29		1.72	"
550-532	3	"	Darling batholith		-4.25		1.54	"
550-533	3	"	Woodville granite		-4.94		1.60	"
550-534	3	"	Rooiklip granite		-5.85		1.71	"
540	3	"	Riviera pluton		-2.1		1.01	"
540	3	"	Haelkraal granite		-2.78		1.99	"
540	3	"	Paarl pluton		-1.87		123	"
540	3	"	Paarl pluton		-1.92		1.89	"
540	3	"	Greyton granite		-3.63		1.49	"
540	3	"	Robertson pluton		-3.08		1.41	"
540	3	"	Schapenberg granite		-1.44		1.32	"
540	3	"	Swellendam granite		-3.89		1.45	"
540	3	"	Worcester mylonite		-1.78		1.21	"
540	3	"	Cape Columbine granite		-2.56		1.39	"

Table 3. Compilation of ages of Pampean-Saldanian igneous rocks and of isotope data (Nd and Hf in zircons)

Region and lithology	Metamorphic event	Mineral assemblage	*P-T* conditions	References
Low- to mid-temperature domain				
Puncoviscana Formation NW Argentina	M_1 (HP/LT)	Wm+Chl+Qz	***b* parameter** of 9.035-9.055 Å (intermediate-high pressure) **CIS** 0.23-0.36 ºΔ2θ (anquizone-epizone) 240-300 ºC, 8-9 kbar **(1)**	Do Campo y Nieto (2003) y Do Campo et al (2013)
	M_2 (LP/LT)	Wm+Chl+Qz	275-350 ºC, 0.7-3 kbar **(1)**	
High-temperature domain (Sierras de Córdoba)				
Cordierite diatexite (El Pilón)	M_2 (MP/HT)	Grt (core)+Pl (core)+Crd1 (matrix)+Sil+Qz	780 ºC, 5.9 kbar **(1)**	Rapela et al. (2002)
Restite in monzogranite (el Pilón)	M_3 (LP/MT-HT)	Crd+Bt+Ms+Sil+Qz+Pl±K fs	550 ±50 ºC, 3.3 ± 0.6 kbar **(1)**	
Garnet-cordierite diatexite. Central Sierra Chica	M_2 (MP/HT)	Grt+Sil+Crd+Qz+Bt+Kfs	820±25 ºC, 5.7±0.4 kbar **(1)**	Baldo et al. (1996) and Rapela et al (1998)
	M_3 (LP/MT-HT)	Sp+Sil+Crd+Kfs	715±15 ºC, 4±0.5 kbar **(1)**	
Garnet-biotite gneis. Central Sierra Chica	M_2 (MP/HT)	Grt+Kfs+Qz+Pl+Sil+Bt and relictic Ky	820±60 ºC, 6.3±1 kbar **(1)**	
Banded garnet gneisses. Southern Sierra Chica	M_2 (MP/HT)	Grt (core)+Bt+Pl+Sil+Qz	715-814 ºC, 7.3-8.6 kbar **(2)**	Martino et al. (2010)
	M_3 (LP/MT-HT)	Grt (rim)+Bt+Pl+Sil+Qz	598-710 ºC, 5.2-7.2 kbar **(2)**	
Garnet-biotite gneisses. Northern Sierra de Comechingones	M_2 (MP/HT)	Grt+Bt+Qz+Pl+Rt	760±30 ºC, 3±0.5 kbar **(3)**	Otamendi et al. (1999)
	M_3 (LP/MT-HT)		600 ºC, 5.8 kbar **(3)**	
Garnet±cordierite migmatites. Northern Sierra de Comechingones	M_2 (MP/HT)	Grt+Bt+Qz+Pl+Sil+Rt+Ilm	800-900 ºC, 7-8.3 kbar **(3)**	
	M_3 (LP/MT-HT)	Grt+Bt+Qz+Pl+Sil	700-750 ºC, 6.5-6.9 kbar **(3)**	

Garnet-orthopyroxene granulite. Northern Sierra de Comechingones	M_2 (MP/HT)	Grt+Opx+Pl+Qz	850±50 ºC, 7.1-8.5 kbar **(3)**	Otamendi et al. (2005)
Garnet-cordierite granulite. Northern Sierra de Comechingones	M_2 (MP/HT)	Grt+Crd+Pl+Sil+Qz	790 ºC, 8±0.5 kbar **(3)**	Otamendi et al. (1999)

Table 4. Summary of representatively thermobarometry data from low- to mid-temperature and high-temperature domains. (1) TWQ method (Berman, 1991). (2) Thermometer GB (garnet-biotite; Holdaway et al., 1997) and barometer GASP (garnet, sillimanite, quartz and plagioclase; Koziol, 1989). (3) Conventional thermometry (multi-equilibrium).

	S. Pampeanas + NOA		Saldanian Belt	
parameter	S-type	I-type	S-type	I-type
Age (Ma)	529 ± 3 to 520 ± 3	541 ± 4 to 523 ± 5	552 ± 4 to 533 ± 2	536 ± 5
	ca. 523	ca. 530	ca. 540	
			527 ± 8 (late)	
A/CNK	1.1 - 1.4	0.95 - 1.03	1.0 - 1.7	0.9 - 1.1
eNd	-5 / -6	-4 / -10	-3 / -5	-1.4 / -3.9
TDM (Ga)	1.6 - 1.7	1.5 - 2.0	1.5 - 1.9	1.0 - 2.0
eHf		+1.1 / -6.9	-11 / -0.2	
TDM (Ga)		1.3 - 1.7	1.1 - 2.0	

Table 5. Summary of age and gechemical characteristics of igneous rocks from the Pampean and the Saldanian belts

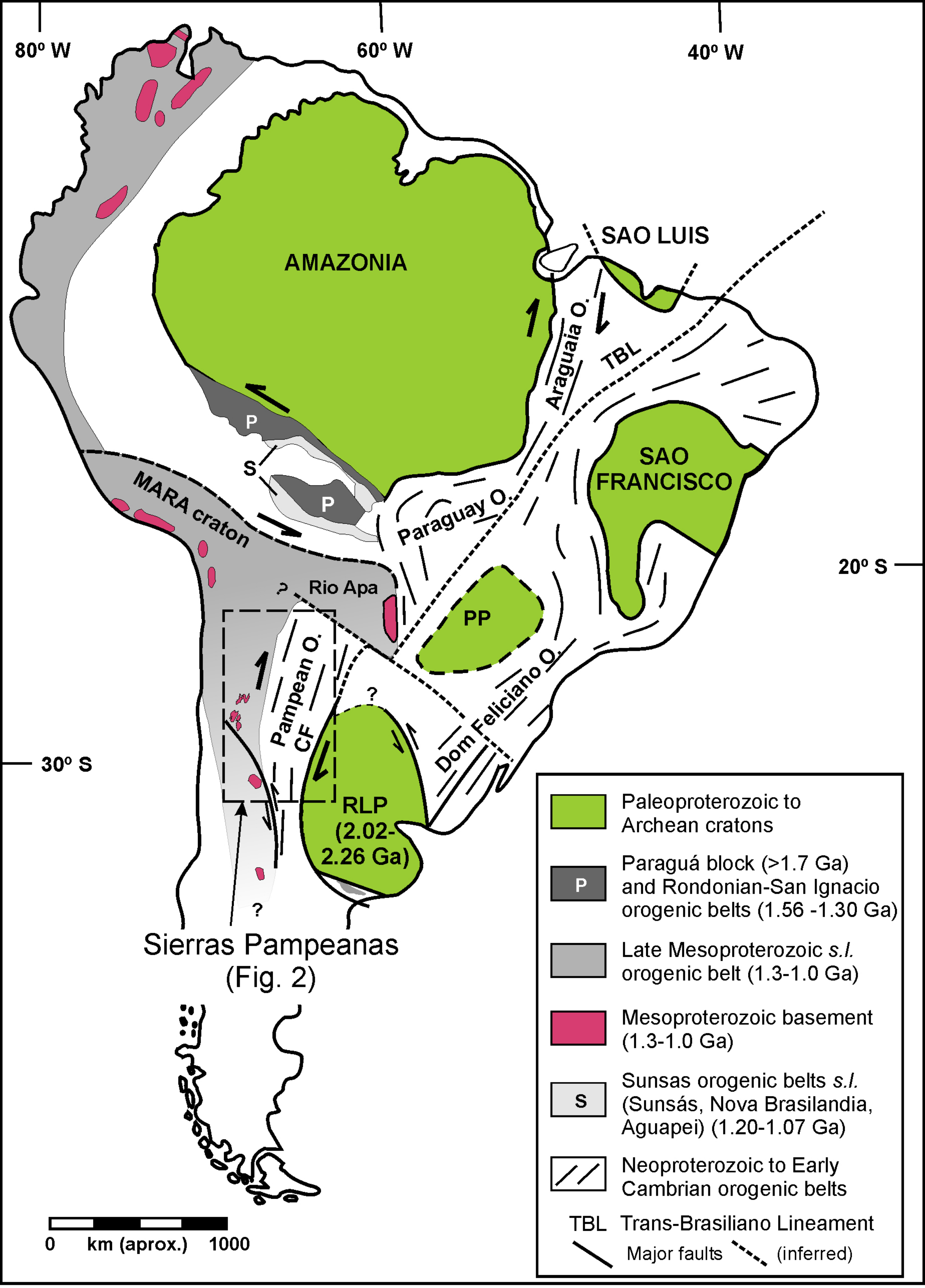

80º W
60º W
40º W
20º S
30º S
AMAZONIA
SAO LUIS
Araguaia O.
TBL
SAO FRANCISCO
P
S
P
MARA craton
Paraguay O.
Rio Apa
PP
Pampean O.
CF
Dom Feliciano O.
RLP (2.02-2.26 Ga)
?
Sierras Pampeanas (Fig. 2)
0 km (aprox.) 1000
Paleoproterozoic to Archean cratons
Paraguá block (>1.7 Ga) and Rondonian-San Ignacio orogenic belts (1.56 -1.30 Ga)
Late Mesoproterozoic s.l. orogenic belt (1.3-1.0 Ga)
Mesoproterozoic basement (1.3-1.0 Ga)
Sunsas orogenic belts s.l. (Sunsás, Nova Brasilandia, Aguapei) (1.20-1.07 Ga)
Neoproterozoic to Early Cambrian orogenic belts
TBL Trans-Brasiliano Lineament
Major faults
(inferred)

Figure 1

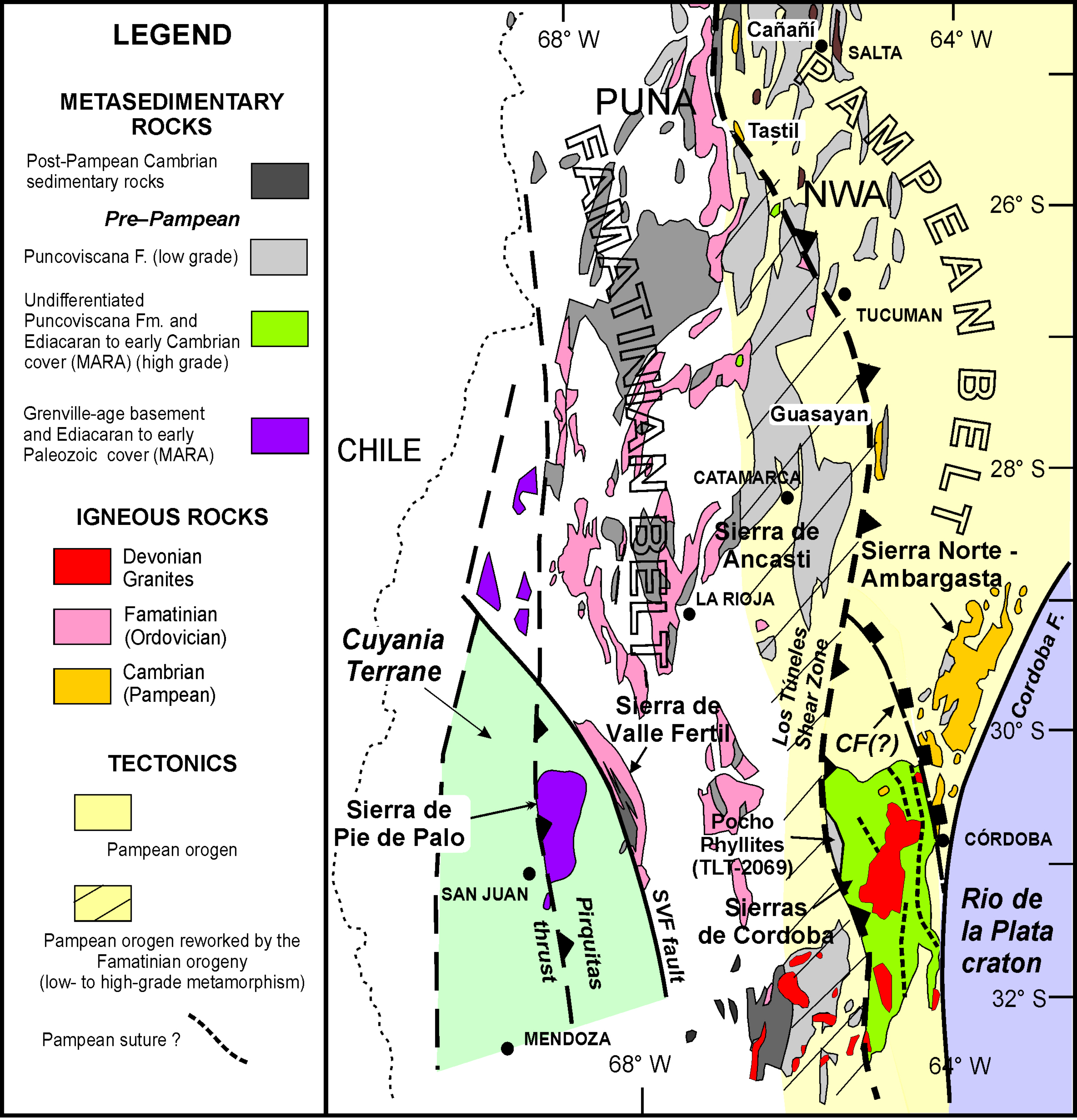
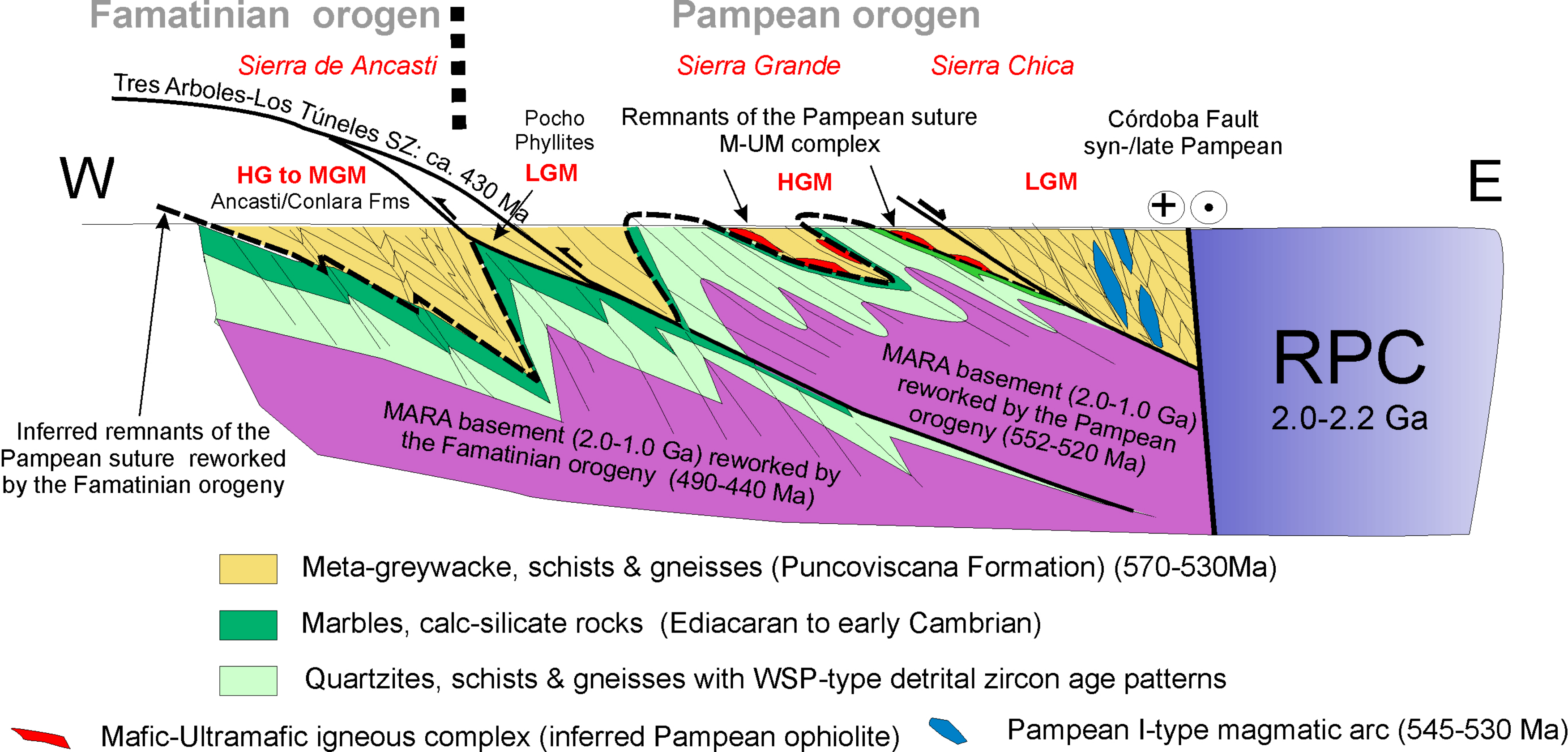
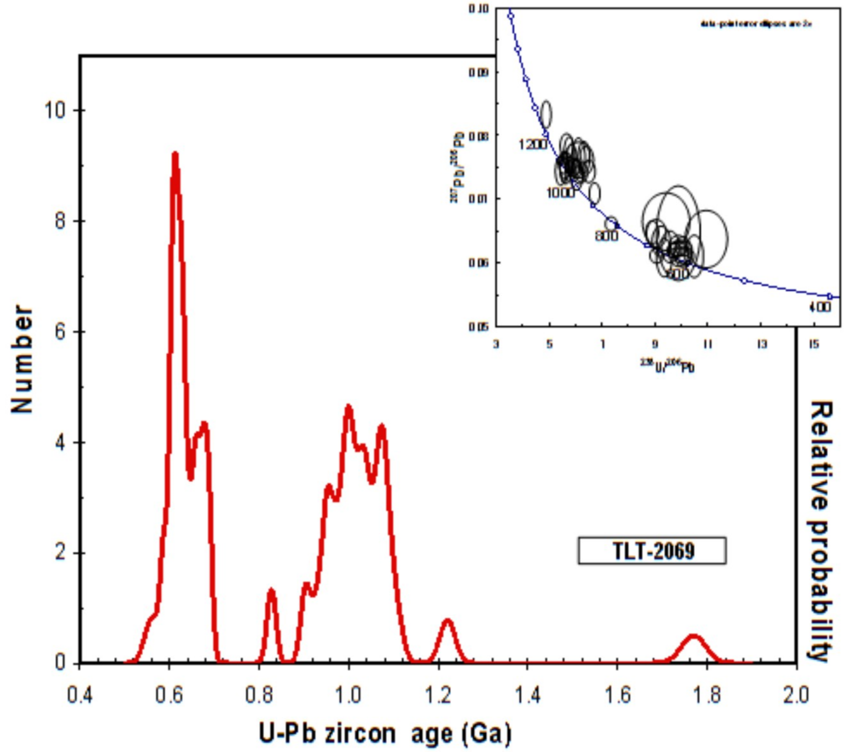
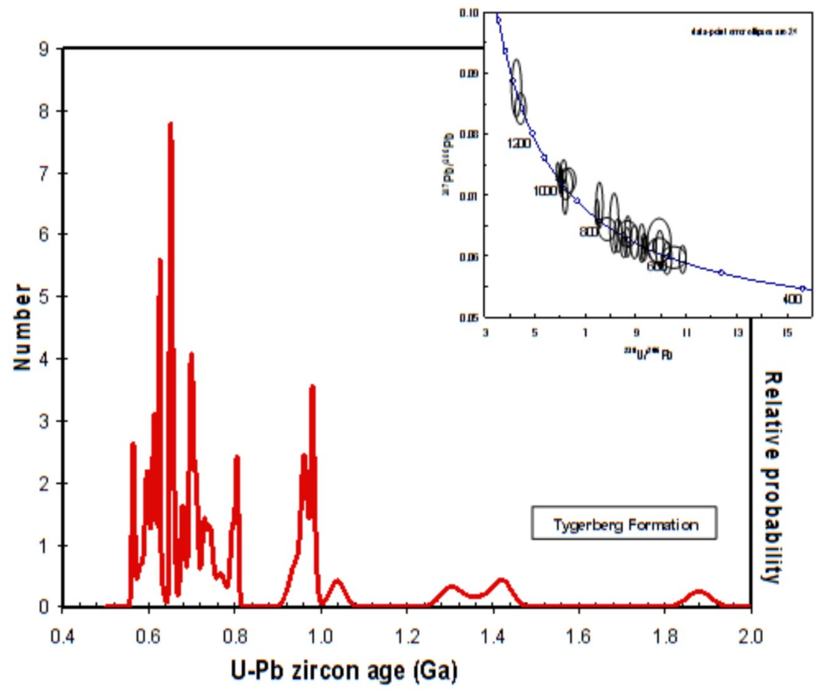
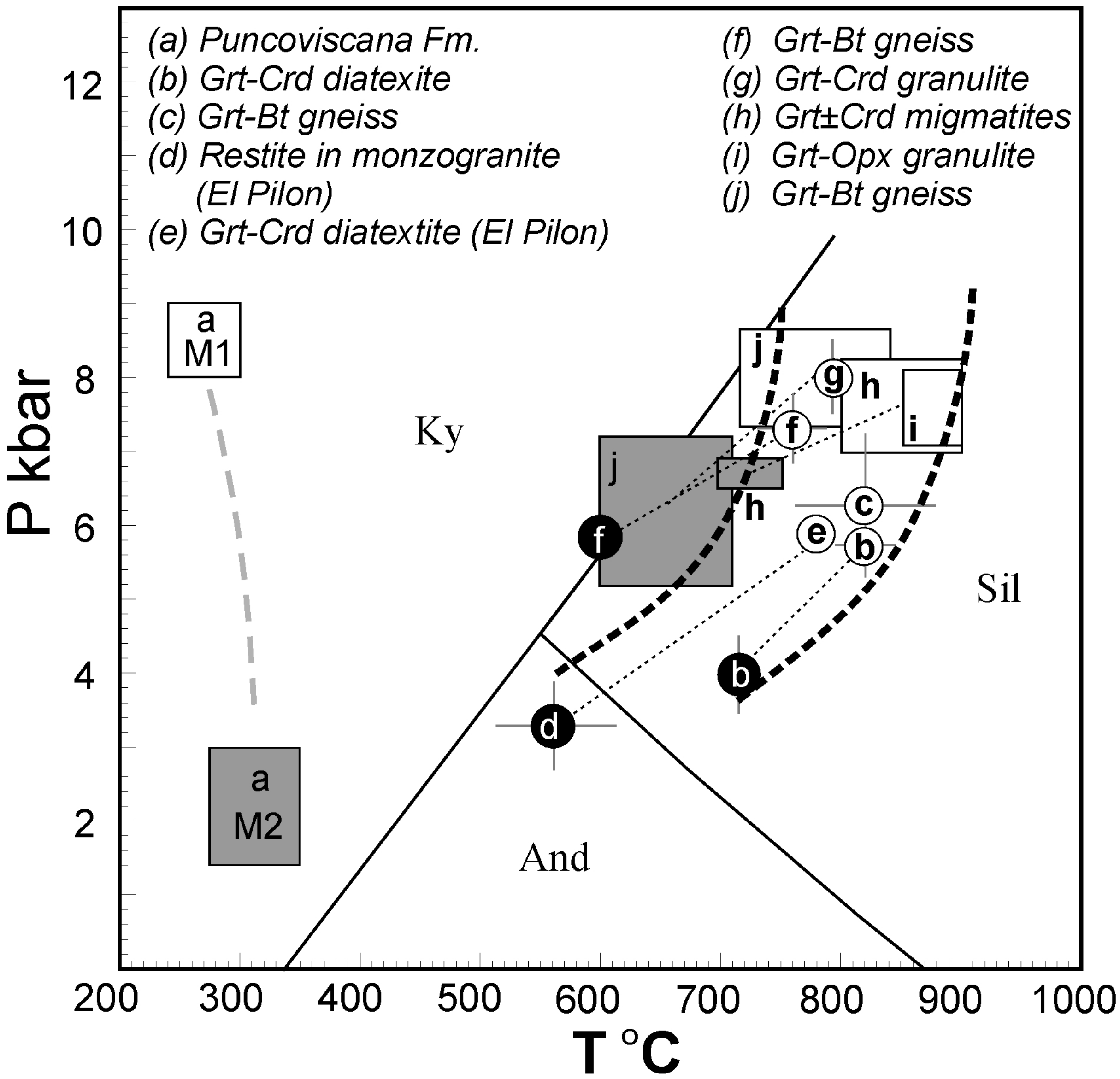

LEGEND
METASEDIMENTARY ROCKS
Post-Pampean Cambrian sedimentary rocks
Pre–Pampean
Puncoviscana F. (low grade)
Undifferentiated Puncoviscana Fm. and Ediacaran to early Cambrian cover (MARA) (high grade)
Grenville-age basement and Ediacaran to early Paleozoic cover (MARA)
IGNEOUS ROCKS
Devonian Granites
Famatinian (Ordovician)
Cambrian (Pampean)
TECTONICS
Pampean orogen
Pampean orogen reworked by the Famatinian orogeny (low- to high-grade metamorphism)
Pampean suture ?
68° W
64° W
26° S
28° S
30° S
32° S
Cañañí
SALTA
PUNA
Tastil
NWA
PAMPEAN BELT
FAMATINIAN BELT
TUCUMAN
Guasayan
CHILE
CATAMARCA
Sierra de Ancasti
Sierra Norte - Ambargasta
LA RIOJA
Cuyania Terrane
Los Túneles Shear Zone
Cordoba F.
Sierra de Valle Fertil
CF(?)
Sierra de Pie de Palo
Pocho Phyllites (TLT-2069)
CÓRDOBA
SAN JUAN
Pirquitas thrust
SVF fault
Sierras de Cordoba
Rio de la Plata craton
MENDOZA

Fig. 2

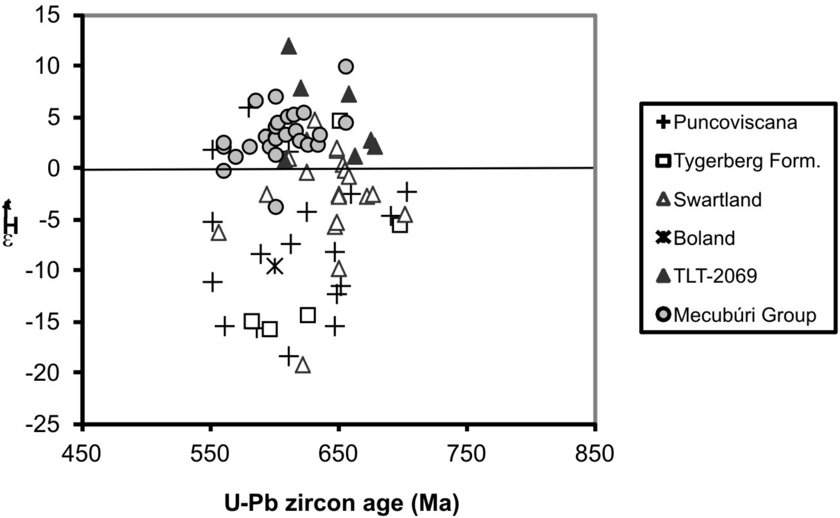

Famatinian orogen
Pampean orogen
Sierra de Ancasti
Sierra Grande
Sierra Chica
Tres Arboles-Los Túneles SZ: ca. 430 Ma
Pocho
Phyllites
Remnants of the Pampean suture
M-UM complex
Córdoba Fault
syn-/late Pampean
W
E
HG to MGM
Ancasti/Conlara Fms
LGM
HGM
LGM
RPC
2.0-2.2 Ga
MARA basement (2.0-1.0 Ga) reworked by the Pampean orogeny (552-520 Ma)
MARA basement (2.0-1.0 Ga) reworked by the Famatinian orogeny (490-440 Ma)
Inferred remnants of the Pampean suture reworked by the Famatinian orogeny
Meta-greywacke, schists & gneisses (Puncoviscana Formation) (570-530Ma)
Marbles, calc-silicate rocks (Ediacaran to early Cambrian)
Quartzites, schists & gneisses with WSP-type detrital zircon age patterns
Mafic-Ultramafic igneous complex (inferred Pampean ophiolite)
Pampean I-type magmatic arc (545-530 Ma)


Number
U-Pb zircon age (Ga)
Relative probability
TLT-2069
data-point error ellipses are 2σ
$^{207}Pb/^{206}Pb$
$^{238}U/^{206}Pb$
1200
1000
800
600
400

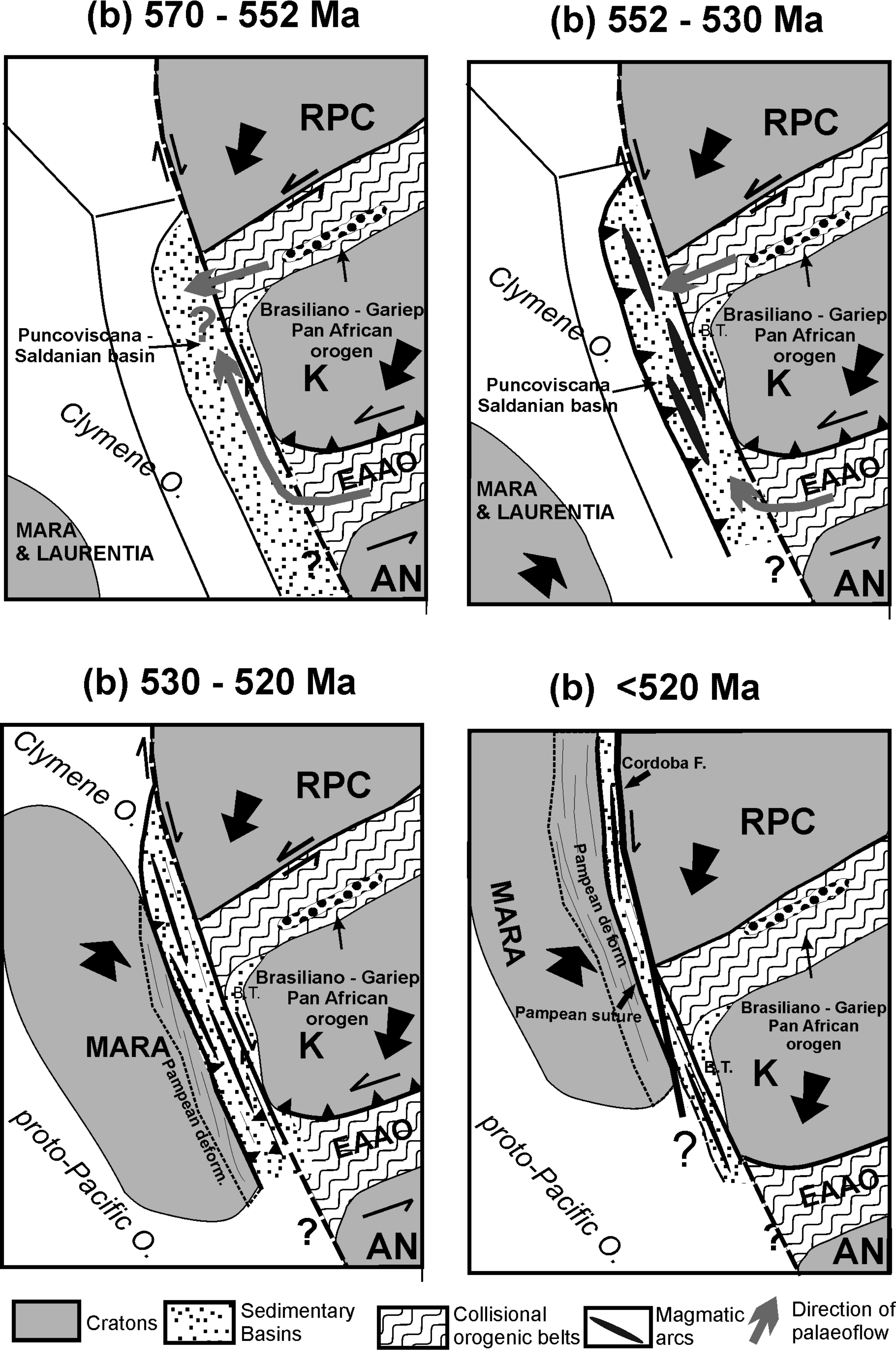

Number
U-Pb zircon age (Ga)
Relative probability
Tygerberg Formation
data-point error ellipses are 2σ
$^{207}Pb/^{206}Pb$
$^{238}U/^{206}Pb$
1200
1000
800
600
400



(a) Puncoviscana Fm.
(b) Grt-Crd diatexite
(c) Grt-Bt gneiss
(d) Restite in monzogranite
(El Pilon)
(e) Grt-Crd diatextite (El Pilon)
(f) Grt-Bt gneiss
(g) Grt-Crd granulite
(h) Grt±Crd migmatites
(i) Grt-Opx granulite
(j) Grt-Bt gneiss
a
M1
a
M2
Ky
Sil
And
j
h
i
g
f
c
e
b
d
P kbar
T °C
2
4
6
8
10
12
200
300
400
500
600
700
800
900
1000



15
10
5
0
-5
-10
-15
-20
-25
εHf t
450
550
650
750
850
U-Pb zircon age (Ma)
Puncoviscana
Tygerberg Form.
Swartland
Boland
TLT-2069
Mecubúri Group



(b) 570 - 552 Ma
RPC
Brasiliano - Gariep
Pan African
orogen
K
Puncoviscana -
Saldanian basin
Clymene O.
EAAO
MARA
& LAURENTIA
AN
(b) 552 - 530 Ma
RPC
Clymene O.
Brasiliano - Gariep
Pan African
orogen
B.T.
K
Puncoviscana
Saldanian basin
EAAO
MARA
& LAURENTIA
AN
(b) 530 - 520 Ma
Clymene O.
RPC
Brasiliano - Gariep
Pan African
orogen
B.T.
K
MARA
Pampean deform.
EAAO
proto-Pacific O.
AN
(b) <520 Ma
Cordoba F.
RPC
MARA
Pampean deform
Pampean suture
Brasiliano - Gariep
Pan African
orogen
B.T.
K
EAAO
proto-Pacific O.
AN
Cratons
Sedimentary
Basins
Collisional
orogenic belts
Magmatic
arcs
Direction of
palaeoflow

Figure 7