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Untangling the Neoproterozoic-Early Paleozoic Tectonic Evolution of the Eastern Sierras Pampeanas Hidden in the Isotopical Record

Mónica G. López de Luchi, Carmen I. Martínez Dopico,
Klaus Wemmer, and Siegfried Siegesmund

Abstract

The Sierras Pampeanas of Central Argentina are an example of a continuous and fast overlap of episodes of high- to medium- and low-grade metamorphism, deformation, anatexis, magmatism and mineralization along the belts that bounded the margins of the South American cratons during the Neoproterozoic and early Paleozoic. A compilation and critical revision of the massive isotopic and geochemical data for the basement rocks of the Eastern Sierras Pampeanas is presented. The Eastern Sierras Pampeanas are defined by three main events: the Ediacaran to early Cambrian (580–530 Ma) Pampean, the late Cambrian–Ordovician (500–460 Ma) Famatinian and the Devonian-Carboniferous (400–350 Ma) Achalian orogenies. The mean average crustal residence age (Sm–Nd T_{DM}) varies between 1.8 and 1.7 Ga with $\varepsilon Nd_{(540)}$ (−6 to −8). Pampean and Famatinian granitoids exhibit a similar T_{DM} interval except for the Ordovician TTG suites of the Sierras de Córdoba (T_{DM} 1.3–1.0 Ga). Achalian magmatism exhibits more radiogenic $\varepsilon Nd_{(540)}$ values (0.5 to −4) and T_{DM} ages younger than 1.3 Ga. Two types of Pampean-related mafic rocks are recognized: one with a depleted mantle signature and another younger group with an enriched mantle signature, which is associated with the peak of metamorphism. Ordovician mafic-ultramafic rocks result from mixing/assimilation of depleted mantle and crustal magmas. Detrital zircon data for the metaclastic sequences indicates mainly Grenvillian and Brasiliano sources. The difference between the measured crystallization age for detrital zircon grains and the depositional age of the succession indicates that most of the Pampean basins are collisional, i.e. foreland basins except for Sierra Norte metaclastic host rocks that correspond to a convergent setting signature. The results for the post-Pampean Famatinian basins indicate mostly collisional convergent settings for the Ambato and La Cébila (type locality) metamorphic complexes and the Olta (northern sector) and Achavil formations. The Negro Peinado Formation is the only post-Pampean basin that corresponds to a collisional setting. Data from Green Quarry, Nogolí Metamorphic Complex, and Olta Formation in the central part of the Sierra de Chepes, Suri Formation and La Cébila Metamorphic Complex at Quebrada La Rioja yielded convergent margin settings. These exhibit the highest correlation among the post-Pampean basins showing a relatively large number of Middle Cambrian age detrital zircons apart from having Ordovician magmatic zircons in the detrital record. Sources were apparently more restricted than in the rest of the Famatinian post-Pampean basins. The two samples of post-Pampean basins that exhibit Río de la Plata age peaks (2.2–2.0 Ga; Paleoproterozoic) belong to the collisional convergent group. Therefore the exhumed Pampean rocks probably formed a drainage divide that blocked westward transport of the Río de la Plata-derived sediments. Intense erosion owing to an unstable tectonic scenario would have led to the progressive appearance of the Río de la Plata signature.

Keywords

Early Paleozoic orogenies • Eastern Sierras Pampeanas
Detrital zircon data • Sm–Nd systematic

M. G. López de Luchi (✉) · C. I. Martínez Dopico
Instituto de Geocronología y Geología Isotópica (INGEIS),
CONICET-UBA, Intendente Güiraldes 2160, Ciudad
Universitaria, C1428EGA Buenos Aires, Argentina
e-mail: deluchi@ingeis.uba.ar

C. I. Martínez Dopico
e-mail: carmenmd@ingeis.uba.ar

K. Wemmer · S. Siegesmund
Geoscience Centre of the University of Göttingen (GZG),
Goldschmidtstr. 3, 37077 Göttingen, Germany
e-mail: kwemmer@gwdg.de

S. Siegesmund
e-mail: ssieges@gwdg.de

S. Siegesmund et al. (eds.), *Geology of Southwest Gondwana*, Regional Geology Reviews,
https://doi.org/10.1007/978-3-319-68920-3_16

16.1 Introduction

The Sierras Pampeanas of Central Argentina (26°–33°S) are a series of N-S striking blocks that originated during the latest Neoproterozoic–Early Paleozoic as a result of the evolution of the tectonically active western margin of Gondwana, after the breakdown of Rodinia. The final uplift of the Sierras Pampeanas was closely related to the inception of the Miocene to Recent 27°–33° 30′S flat-slab segment of the Nazca plate in the Southern Andes (Ramos et al. 2002; Bense et al. 2013a). Previous exhumation processes causing the main uplift were pinpointed in Devonian, Carboniferous, Permian and late Cretaceous times (Steenken et al. 2010; Löbens et al. 2011, 2013a, 2013b, 2016; Bense et al. 2013b, 2017). The Sierras Pampeanas developed at a key locality for understanding the orogenic events along the Pacific margin of SW Gondwana, between the Paleoproterozoic Río de la Plata Craton against which they probably have a continental fault contact (Booker et al. 2004; Rapela et al. 2007; Peri et al. 2015) to the east and the Grenvillian (e.g., Kay et al. 1996; *c.* 1200–1000 Ma), Cuyania-Precordillera Terrane to the west. An overview of the geology of the Río de la Plata Craton is presented by Oyhantçabal et al. (2010), Rapela et al. (2011) and in Chap. 4. The final configuration of Gondwana was achieved by the end of the mid-Ediacaran time at the latest stages of the Pan-African/Brasiliano cycle. The main cratons involved were Amazonia–Arequipa–Río Apa, Kalahari, Río de la Plata, Congo, East Antarctica etc. Several collisional orogenic belts resulted, notably the East Africa–Antarctica, Brasiliano–Panafrican, Pampean–Saldania, and the Ross–Delamerian orogens (Casquet et al. 2012). The Brasiliano orogenic collage took place in four distinct phases: (a) Late Tonian (*c.* 800–740 Ma); (b) Late Cryogenian-Early Ediacaran (*c.* 660–610 Ma); (c) Early-Middle Ediacaran (*c.* 590–560 Ma); and (d) Late Cambrian (520–500 Ma; de Brito Neves et al. 2014). For the latest discussion see Oriolo et al. (2017).

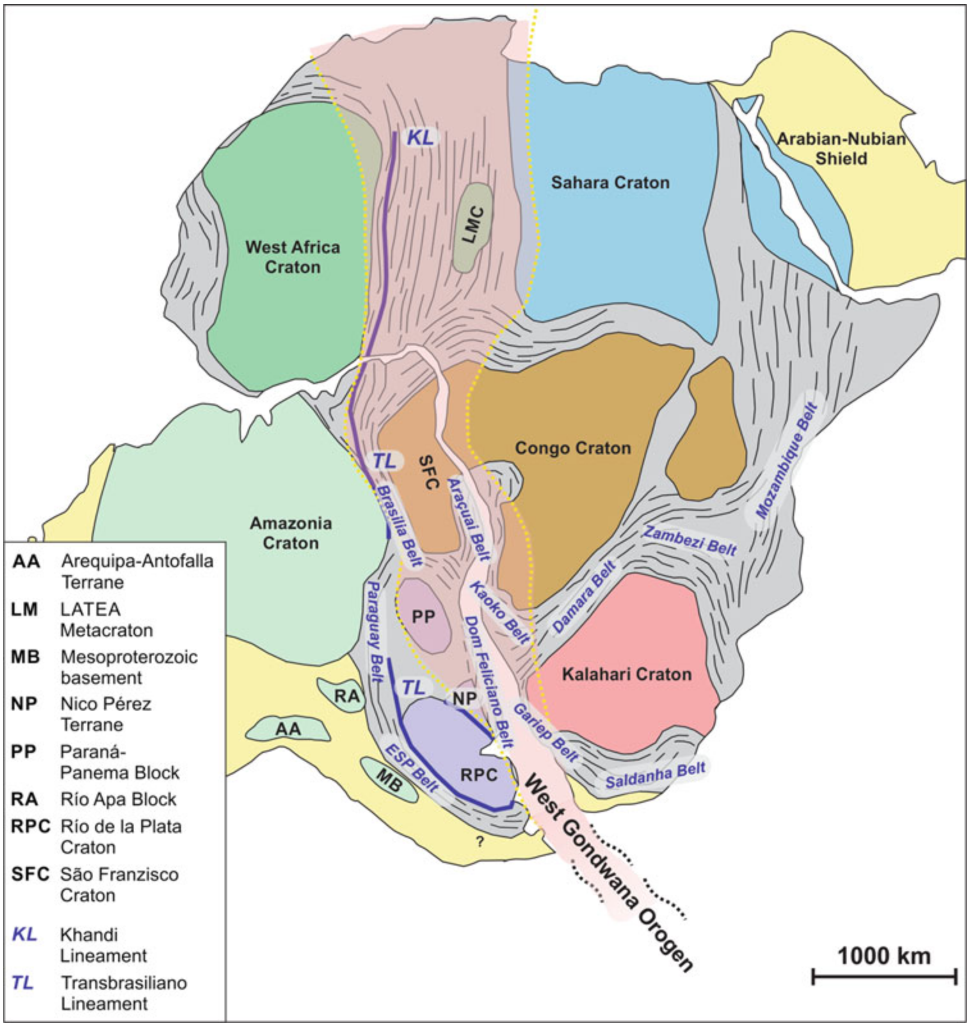
Caminos (1979) established a lithologically based separation of the Sierras Pampeanas into (1) the Eastern Sierras Pampeanas dominated by abundant granites, metasedimentary and metaigneous rocks; and (2) the Western Sierras Pampeanas characterized by abundant metabasic, ultrabasic and calcsilicate rocks. This original separation corresponds to different geological provinces. The Western Sierras Pampeanas (WSP) expose (1330–1030 Ma 'Grenville-age') Mesoproterozoic crystalline basement intruded by relatively scarce Ordovician granites of the Famatinian cycle (e.g., Pankhurst and Rapela 1998; Casquet et al. 2001, 2006, 2008; Sato et al. 2003; Vujovich et al. 2004; Rapela et al. 2010). The Eastern Sierras Pampeanas (ESP) constitute a polyphase deformed morphotectonic unit which is defined by three main events: the late Ediacaran to mid-Cambrian (560–510 Ma) Pampean, the late Cambrian–Ordovician (500–445 Ma) Famatinian, and the Devonian (400–360 Ma) Achalian orogenic cycles (Aceñolaza and Toselli 1981; Ramos et al. 1986; Aceñolaza et al. 1988; Sims et al. 1997, 1998; Rapela et al. 1998a, b, 2001, 2007; Stuart-Smith et al. 1999; Siegesmund et al. 2004, 2010; Steenken et al. 2004, 2006, 2011; López de Luchi et al. 2007; Drobe et al. 2009). These mountain-building processes are related to the accretion of different terranes integrated into the proto-Andean margin of Gondwana. Moreover, a distinct Carboniferous tectonic event in the Sierras Pampeanas, which post-dates the late Devonian collisional orogeny and pre-dates the Permian San Rafael orogenic phase, has been proposed by Dahlquist et al. (2015). This possible transtensional event played a major role during the development and evolution of the late Paleozoic Paganzo Basin as well as during the emplacement of alkaline magmatism in the retroarc (Alasino et al. 2012; Löbens et al. 2016).

The aim of this contribution is to present an overview of the main geological features of the Eastern Sierras Pampeanas, mainly focused on the area located between 28° and 38°S (Figs. 16.1 and 16.2). Furthermore, the investigation aims to set time constraints for the metamorphism and magmatism, to analyse petrogenetic events and track potential sources through the WR Sm–Nd isotopic system. A comprehensive statistical approach to detrital zircon ages of the metaclastic rocks and the WR Sm–Nd data provides a summary that may support a revised model for the Early Paleozoic tectonic evolution.

16.2 Geological Background of the Main Basement Units of the Eastern Sierras Pampeanas

The Eastern Sierras Pampeanas extend from southern Salta (26°S) to the San Luis and La Pampa provinces (38°S) (Fig. 16.2). The eastern sector of these basement units is mainly affected by the Pampean orogeny, which is characterized by late Neoproterozoic sedimentation and Ediacaran to Cambrian deformation, magmatism and medium- to high-grade metamorphism (e.g., Rapela et al. 2007; Siegesmund et al. 2010; Steenken et al. 2011). The western sector is dominated by the Famatinian orogeny, which is characterized by Late Cambrian to Early-Middle Ordovician marine and volcaniclastic successions and Early to Middle Ordovician I- and S-type intrusions, minor tonalite–trondhjemite–granodiorite suites in the foreland (e.g., Sims et al. 1998; Steenken et al. 2006). The Famatinian belt is characterized by low- to high-grade temperature, low- to intermediate-pressure metamorphism coeval with foliation development, folding and thrusting. The Achalian orogeny

Fig. 16.1 Cratonic blocks and Neoproterozoic belts of southwestern Gondwana. ESP = Eastern Sierras Pampeanas



overprints this basement (Sims et al. 1998; Siegesmund et al. 2010). The 393 and 360 Ma mid- to late-Devonian magmatism is developed mainly in the Sierra de Córdoba and San Luis, whereas Lower Carboniferous magmatism is more conspicuous towards the north in the Sierra de Velasco (Fig. 16.2). Most of these Devonian batholiths display elongated shapes and are composed of plutons that have discordant contacts with the country rocks and produce wide thermal aureoles. Post-Pampean cooling of the basement domains in the Cambrian to Early Ordovician is related to imbrication and uplift along different shear zones, while Middle to Late Silurian K/Ar biotite ages register different stages of the exhumation (Steenken et al. 2010).

The Late Ediacaran to Early Paleozoic low- to high-grade metasedimentary successions of the Eastern Sierras Pampeanas were considered to be an extension of the very low- to low-grade turbiditic metaclastic rocks of the Puncoviscana Formation that developed along the Cordillera Oriental (e.g., Schwartz and Gromet 2004; Steenken et al. 2004; Zimmermann 2005; Drobe et al. 2009, 2011). These protoliths were deposited in the Late Ediacaran (Omarini et al. 1999) to Early Cambrian and were folded in the Early Cambrian (537–523 Ma; Escayola et al. 2011 and references therein).

The Sierras de Córdoba and the Sierra Norte as well as its continuation in the Sierras de Ambargasta and Sumampa are the easternmost group of the Sierras Pampeanas. These Sierras are made up of Neoproterozoic to Paleozoic plutonic-metamorphic basement intruded by Paleozoic intermediate and felsic plutonic rocks. They consist of a series of submeridian mountain chains limited by west-vergent reverse thrust faults on its western side and separated by intermontane Mesozoic and Cenozoic sediments, which are partially covered by a localized series of trachiandesitic volcanites and pyroclastic deposits of the upper Tertiary age.

The Sierra Norte, Sierras de Ambargasta and Sumampa constitute a major block mostly made up of the Ediacaran to early Cambrian arc-related I-type calcalkaline Sierra Norte-Ambargasta Batholith (SNAB) (Lira et al. 1997; Siegesmund et al. 2010). This batholith comprises

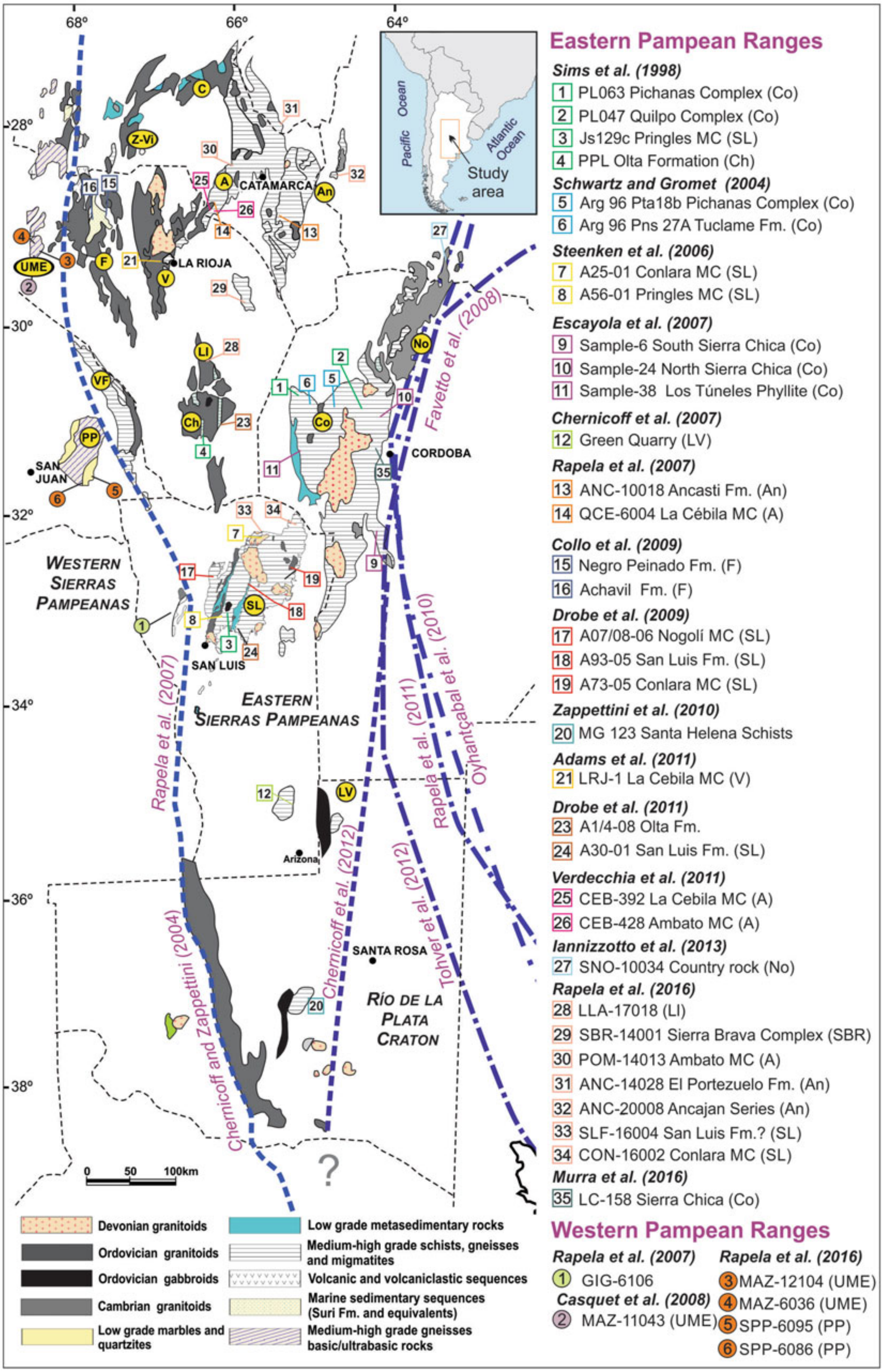


Fig. 16.2 Schematic map of the Early Paleozoic sequences of the Sierras Pampeanas between 24°–38° S with location of the published detrital zircon data. Limits of the Pampean and Famatinian orogens were taken from Chernicoff and Zappettini (2004) and Rapela et al. (2007). Limits of the Rio de la Plata craton were taken from Favetto et al. (2008), Oyhantçabal et al. (2010), Rapela et al. (2011), Chernicoff et al. (2012), and Tohver et al. (2012). Code for the names of the mountain chain A = Ambato; An = Ancasti; C = Capillitas; Co = Sierras de Córdoba; Ch = Chepes; F = Famatina; Fi = Fiambalá; LL = Llanos; No = Sierra Norte de Córdoba; PP = Pie de Palo; SL = Sierra de San Luis; LP = La Pampa hills; LV = Lonco Vaca; UME = Umango, Espinal, Maz; V = Velasco; VF = Valle Fértil; Z-Vi = Zapata-Vinquis

537 ± 4 Ma granitoid rocks affected by Pampean D_2 dextral shearing and mylonitization, and 530 ± 4 Ma granitoids emplaced after deformation had ceased (Iannizzotto et al. 2013). Late magmatic activity represented by a series of volcanic rocks associated with an extensional regime took place at 521 ± 2 Ma (Ramos et al. 2015; Oncan Rhyolite). The SNAB encloses large roof pendants of a locally contact overprinted late Neoproterozoic metamorphic complex (Siegesmund et al. 2010 and references therein). Ignimbritic rhyolites intercalated in conglomerates record synsedimentary igneous activity at 584 + 22/−14 Ma (conventional U–Pb zircon date in Llambías et al. 2003).

The Sierras de Córdoba constitute a unit of igneous and low- to high-grade metamorphic para- and orthogneisses, pure and impure marbles, amphibolites and ultramafic rocks, which underwent local partial melting processes that led to migmatitization (migmatite massifs are described in Gordillo 1979; Guereschi and Martino 2014). Several peraluminous and metaluminous granites and granodiorites of different ages intrude this basement (Rapela et al. 1998a, b; Sims et al. 1998; Escayola et al. 2007). The polymetamorphic complexes are organized in lithological and structural domains (Fig. 16.2): Sierra Chica, Sierra Grande and Sierra de Comechingones separated by ductile shear zones and mafic and ultramafic rocks (Siegesmund et al. 2010 and references therein). These domains are characterized by Ediacaran sedimentation and late Ediacaran to Cambrian deformation, magmatism and metamorphism with a more restricted Ordovician magmatic event (Rapela et al. 1998b; Siegesmund et al. 2010) and voluminous intrusions of Devonian granites (López de Luchi et al. 2007, 2017; Dahlquist et al. 2015). A NNW trending trending belt of granulite facies metamorphic rocks and related S-type granitoids (Fig. 16.2) can be split into the San Carlos Massif in the northwest and the Sierra de Comechingones in the southeast (Guereschi and Martino 2008, 2014 and references therein). Low-grade metamorphic rocks, the Los Túneles Phyllites are recognized in the western region of the Sierra Grande. Geochronological data supports the idea that the most important tectonothermal evolution of the crystalline rocks of Córdoba occurred in the lower Cambrian during the Pampean orogeny (*c.* 530–577 Ma; Rapela et al. 1998a, b; Sims et al. 1998; Siegesmund et al. 2010). In Table 16.1 a characterization of the metamorphic events is depicted. Peraluminous granitoids associated with migmatites were generated during a high-grade regional 525–519 Ma post M2 metamorphic event (Escayola et al. 2007; Siegesmund et al. 2010). Ordovician (480–490 Ma) granitoids are non-collisional, whereas the Devonian granitoids, e.g. Achala Batholith, Cerro Aspero pluton and Capilla del Monte pluton are alkali-calcic peraluminous granitoids.

Mafic and ultramafic rocks that were in part interpreted as ophiolites were separated into an eastern Sierra Chica belt and a western Sierra Grande belt (Mutti 1992; Escayola et al. 1996). The NNW-SSE trending eastern belt is composed of lherzolites with interlayered websterites and subordinate harzburgites, abundant pyroxenites and gabbros. Escayola et al. (1996) assigned the tectonic setting to an ensialic back-arc basin, whereas Ramos et al. (2000) interpreted the belt as ophiolite remnants of a back-arc zone. The ultramafic-mafic units of the western belt (Escayola et al. 2007), which extend for *c.* 300 km, are made up of harzburgites that were metasomatized by intrusive basaltic dykes. Escayola et al. (1996) interpreted the western belt as a suture zone and considered the ultramafic rocks as dismembered ophiolites with MORB affinities. Escayola et al. (2007) suggested a suprasubduction-zone ophiolitic complex formed during a back-arc stage. Sm–Nd dating of basalt and gabbro dykes, pyroxenites and impregnated peridotites of the western belt yielded an isochron age of 647 ± 77 Ma. The $\varepsilon\mathrm{Nd}_{(647)}$ value of 5.2 is consistent with an oceanic or back-arc origin for the ophiolite sequence (Escayola et al. 2007).

The Sierra de Comechingones encompasses the Calamuchita, Monte Guazú and Achiras complexes (Otamendi et al. 2004). The Calamuchita Complex, north of the Cerro Aspero, is essentially composed of a metasedimentary sequence synkinematically intruded at the time of the metamorphic peak by an OIB-type affinity basic unit (Tibaldi et al. 2008). The Monte Guazú and Achiras complexes appear at the southern end of the mountain range and are in tectonic contact with the Las Lajas shear belt (Stuart-Smith and Skirrow 1997; Otamendi et al. 2000). The Monte Guazú Complex, interpreted by Fagiano et al. (2008) as the southern culmination of the Calamuchita complex, is characterized by an Ordovician (?) calcalkaline igneous rock suite (Gromet et al. 2001; Drobe et al. 2011) intercalated with gneisses, schists, amphibolites and marbles, derived from protoliths similar to the Calamuchita Complex (Otamendi et al. 2000; Fagiano et al. 2008). The Achiras Complex is a monotonous sequence comprising discontinuous belts of greywacke-derived amphibolite-facies gneisses and schists that are interlayered with peraluminous leucogranites. The complex is intruded by Devonian peraluminous granites, like the Achiras Granite (384 ± 6 Ma, U–Pb zircon; Stuart-Smith et al. 1999). Fagiano et al. (2008) proposed an Ordovician age for the metamorphism of the Achiras Complex.

Gabbros with OIB signatures located in the Sierra de Comechingones, which result from the subduction of a mid-ocean ridge beneath the paleo-Pacific Gondwana margin, are related to the high-grade metamorphic overprint (Tibaldi et al. 2008 and references therein). Intrusive relationships suggest that emplacement varied from preceding (e.g., Cerro San Lorenzo; Chincarini et al. 1998) to coincident with peak metamorphism of the metasedimentary rocks (e.g., Suya Taco mafic complex; Otamendi et al. 2005; Tibaldi et al. 2008). Gromet et al. (2005) reported an U–Pb

Table 16.1 Mineral paragenesis, P–T constraints and ages of the Pampean metamorphic events

Basement unit	Paragénesis	P–T	Metamorphic age (Ma)	References
North Sierra de Córdoba	Cdr+Sill migmatites Cdr+Sill+Grt gneiss Bt-Grt gneiss	M2 P = 5.5–6.5 kbar T = 700 ºC M1	543 ± 4* 531 ± 10* 561 ± 10* 581 ± 16*	Martino and Guereschi (2005) Siegesmund et al. (2010) Sims et al. (1998)
Sierra Chica de Córdoba	Cdr migmatite Grt+Cdr+Sill Metabasites	M2 P = 6 kbar T = 820 ºC	522 ± 8**	Rapela et al. (1998a)
Santa Rosa Massif *Sierra de Córdoba*	Grt+Cdr migmatite	M2 = ca. 7.5 kbar T = 850–900 ºC	536 ± 11* 511 ± 6**	Siegesmund et al. (2010) Otamendi et al. (2006)
Tala Cruz Stromatite Cañada del Sauce Diatexite *South Sierra de Córdoba*	Bt–Sil–Grt–Kfs migmatite Bt–Grt–Crd–Kfs diatexite	M1 P = 8.5–9 kbar T = 810–840 °C P = 5.5–6 kbar T = 725–780 °C	553 ± 3* 577 ± 11*	Guereschi and Martino (2008) Siegesmund et al. (2010)
Monte Guazu Complex *Sierra de Comechingones*	Kfs+Sil+Grt Gneiss	P = 7 kbar T = −∼750 ºC		Fagiano (2007)
Conlara Metamorphic Complex *Sierra de San Luis*	Bt±Grt±Ms schist-gneiss	M1 P = 5–6 kbar T = 550−600 ºC	564 ± 21***	Siegesmund et al. (2010)
Ancasti Formation *Sierra de Ancasti*	Ms+Bt+And+Crd Schist		524 ± 28****	Knüver (1983)
Santa Helena schist *La Pampa province*	Bt+Grt Schist Sill+Grt gneiss	T > 700 ºC	Pre 530 Ma	Zappetini et al. (2010) Chernicoff et al. (2008)

*Zircon ages; **Monazite ages; ***garnet(PbSL); ****Rb–Sr WR isochron

age of 520 ± 2 Ma for igneous monazites in a migmatitic granitic leucosome developed in the aureole of a mafic dyke in the Suya Taco mafic complex. This age might be considered to be a cooling age from granulite facies metamorphism since an age of 536 ± 11 Ma was calculated for the Santa Rosa Grt-Crd granulite, which is the restite of the melting event that is associated with the metamorphic peak (Siegesmund et al. 2010).

The Sierra de San Luis (Figs. 16.1 and 16.2) records an Ediacaran to Devonian metamorphic and magmatic evolution and comprises three NNE-SSW striking basement domains of amphibolite- to granulite-facies complexes: the Nogolí, Pringles and Conlara metamorphic complexes (Sims et al. 1997). The domains are generally (Fig. 16.2) separated by an assemblage of metaquartz arenites and phyllites, e.g., the San Luis Formation (Prozzi and Ramos 1988; Wemmer et al. 2011) and are intruded by Ordovician and Devonian granitoids (Sims et al. 1997, 1998; López de Luchi et al. 2007). The metamorphic events and mineral parageneses are summarized in Table 16.1.

The Conlara Metamorphic Complex (CMC) is the easternmost of these complexes. The western margin of the CMC is affected by the Devonian Río Guzmán shear zone, whereas its eastern margin towards the Sierra de Comechingones (Sims et al. 1997) is controlled by the Guacha Corral shear zone. The metamorphic series of the CMC comprises NNE trending metagreywackes, and scarce metapelites and metaigneous components, which encompass basic sills and granitic rocks. These belong to two distinct groups: an earlier one represented by tonalitic to granodioritic orthogneisses, locally migmatites, and scarce amphibolites, and a later granitic one, which comprises Late Cambrian to Early Ordovician monzogranite to tonalitic plutons (i.e., La Tapera, El Peñón, El Salado plutons) and a group of large granitic pegmatites.

Within the CMC, ultramafic rocks have not been reported yet. However, scarce amphibolites are distributed in a N-S belt through Sierra del Morro, San Felipe and Villa de Praga. Some tungsten deposits are genetically related to them (de Brodtkorb et al. 2005).

The Pringles Metamorphic Complex (PMC) consists of paragneisses, mica schists, migmatites and amphibolites. Discontinuous lenses of mafic (mainly norites) to ultramafic units occur along a narrow NNE-SSW central belt concordant with the NNE trending S2 foliation. Granulite facies metamorphism, which would result from the emplacement of (ultra-)mafic intrusions, occurred during the beginning of the Famatinian cycle (Fig. 16.3; Hauzenberger et al. 2001; Steenken et al. 2005, 2006; Delpino et al. 2007). Granulite facies assemblages grade into amphibolite and greenschist facies assemblages (Hauzenberger et al. 2001; Delpino et al. 2007). Mafic to ultramafic 506–480 Ma Mid-Cambrian to earliest Ordovician complexes developed in the central part of the Sierra de San Luis (Sims et al. 1998; Hauzenberger et al. 2001; Steenken et al. 2008) and are represented by norites to gabbro-norites with minor ultramafic rocks. A back-arc or marginal basin setting was assigned to them. On both sides of the mafic to ultramafic bodies, a metamorphic gradient from granulite to greenschist facies is observed (Delpino et al. 2007), and therefore a contact metamorphic origin for the granulite facies paragenesis was suggested (Hauzenberger et al. 2001; Steenken et al. 2005, 2006).

The Nogolí Metamorphic Complex (NMC) is composed of paragneisses, mica schists, metaquartzites and migmatites, with minor amphibolite and small lenses of two mica- and garnet-leucogranites. González et al. (2009 and references therein) proposed that part of the amphibolites are metakomatiites and high-Fe tholeiite metabasalts, which are interlayered with marbles and banded iron formation. Prograde regional metamorphism accompanies deformational phases, from at least middle greenschist to high amphibolite facies conditions (Table 16.2).

The NNE trending low-grade phyllites and metaquartz-arenites of the San Luis Formation appear in two belts (Fig. 16.2). NNE trending large-scale tight folds with a moderate plunge is the main deformation style in the San Luis Formation, corresponding to the Famatinian event recorded in the PMC (Steenken et al. 2006, 2008). Isoclinally folded quartz layers within the phyllites (Fig. 16.2) suggest a preceding deformation phase. The transpressional Río Guzmán shear zone that accommodated the 'east-side-up' displacement of the CMC is developed along the eastern belt of this formation (Steenken et al. 2008).

López de Luchi et al. (2007) considered that the Late Cambrian– Early Ordovician granitoids (*c*. 500–470 Ma) of the Sierra de San Luis are synkinematic with compressive deformation related to the early stages of Famatinian convergence. The melting of crustal sources relates to a contemporaneous mafic magmatism as expressed in the mafic-ultramafic complexes of the Pringles Metamorphic Complex (Steenken et al. 2006, 2008 and references therein). Devonian magmatism is represented by voluminous alkali-calcic ellipsoidal granodiorite-monzogranite units, for example, Las Chacras-Potrerillos and Renca batholiths and La Totora, San José del Morro and El Hornito plutons associated with lamprophyres and monzonitic rocks, which suggest an enriched lithospheric mantle source (López de Luchi et al. 2017).

The Sierras de Chepes, Malanzán and Los Llanos (Fig. 16.2) are composed of Early Ordovician metaluminous, calcalkaline I-type granitoids and restricted peraluminous S-type monzogranites (Dahlquist and Galindo 2004). Small bodies of gabbro diorite or quartz diorite frequently mingled with the surrounding granodiorites. The most important granitoid unit is the 490 ± 5 Ma Chepes granodiorite (Pankhurst et al. 1998). The greenschist to amphibolite grade metasedimentary rocks are largely metapelites with intercalations of metarenites that occur as discontinuous

Table 16.2 Mineral paragenesis, P–T constraints and ages of the Ordovician metamorphic events

Basement unit	Paragenesis	P–T	Metamorphic age (Ma)	References
Pringles metamorphic complex	Grt–Sill Gneiss Pl–Kfs–Grt–Bt–Sil±Cd	P = 5.7–6.4 kbar T = 740– 790 °C	498 ± 10*	Steenken et al. (2006)
Nogolí metamorphic complex	Bt+Sill+Pl migmatite Bt–Pl–Grt gneiss	Not determined P = 5–8 kbar T = 660–791 °C	478 ± 4* 475–457*	Steenken et al. (2006)
Green quarry	Qz–Feld–Bt schist		Between 502–480	Chernicoff et al. (2007)
Olta Fm Sierra de Chepes	Low to medium grade schist	P = 3 kbar 400–700 °C	Pre 478 ± 9 (age anatetic Tuani granite)	Pankhurst et al. (2000)
La Cebila metamorphic complex Sierra de Velasco	Low to medium grade gneiss/migmatites	P = 5–7 kbar T = 600 °C	478 ± 4**	Larrovere et al. (2011)
El Portezuelo Igneous metamorphic complex	Bt+Pl±Grt metatexites Bt+Pl+Grt+Crd migmatites	P = 4.5–5.3 kbar T = 670–820 °C	477–470**	Larrovere et al. (2011)
Sierra Brava metamorphic complex	Bt+Pl+Grt gneiss/migmatite Bt+Pl schist Metabasite			Dahlquist et al. (2010)
Achavil Fm	Ill+Chl	P = 2.5–3 kbar T = 200–280	495–475 ± 15***	Collo et al. (2009)
Negro Peinado Fm	Ili+Chl±Bt	P = 4.5 kbar T = 290–400 °C	463 ± 14***	Collo et al. (2009)

*Zircon ages; **Monazite ages; ***garnet(PbSL); ****Rb–Sr WR isochron roof pendants (Pankhurst et al. 1998). The first metamorphic imprint is represented by fine-grained inclusions of biotite, magnetite, quartz and muscovite preserved in cordierite porphyroblasts that would correspond to a second low-pressure metamorphism, which was considered to be coeval with the granitoid emplacement (Pankhurst et al. 1998) grades from phyllite to an anatectic zone (Pankhurst et al. 1998). Metamorphic events and mineral parageneses are summarized in Table 16.2.

The Sierra de Velasco (Fig. 16.2) is dominated by granitoids (Grosse et al. 2011). Low-grade metamorphic rocks (Table 16.2) consisting of phyllites and mica schists correlate with the lower Ordovician metasedimentary La Cébila Formation (Verdecchia et al. 2007; Larrovere et al. 2012). These separate the Sierra de Velasco from the Sierra de Ambato (Fig. 16.2) and are only present as small outcrops along the eastern flank. The granitoid units (Báez et al. 2005; Toselli et al. 2005) comprise older weakly to strongly foliated Ordovician metagranitoids and younger Lower Carboniferous undeformed granites. These plutons share some features with the Devonian granitoids: nearly circular shapes, general lack of pervasive solid-state deformation, discordant relationships with the host rocks, shallow emplacement in the upper crust, syeno- to monzogranitic compositions, high SiO_2 and K_2O content, porphyritic textures and an abundance of pegmatites and aplites. They were possibly generated during a regional crustal heating event (e.g., 96.Miller and Söllner 2005) that occurred during the Late Devonian–Early Carboniferous.

The Sierra de Ancasti (Fig. 16.2) comprises three metamorphic domains:

- The Sierra Brava Complex in the eastern flank is composed of paragneisses and migmatites (Aceñolaza and Toselli 1981), as well as marbles, schists and amphibolites.
- The central sector is mostly formed by folded NNE-SSW trending banded schist of the Ancasti Formation (Aceñolaza and Toselli 1981; Willner 1983), which records a low-pressure Pampean metamorphism that was overprinted by a syndeformational medium grade event (Knüver 1983; Willner 1983; Gaido 2003).
- In the western flank, metasediments of the Ancasti Formation prograde into gneiss and migmatites of the El Portezuelo Formation (Willner 1983). Ordovician granites and tourmaline- and beryl-bearing pegmatites are emplaced in the Sierra Brava Complex and the Ancasti Formation (Dahlquist et al. 2012).

The metamorphic events and mineral parageneses are summarized in Table 16.2.

The Sierra de Ambato (Fig. 16.2) is dominated by widespread metasedimentary rock sequences metamorphosed under different metamorphic grades, such as granulites, migmatites, gneisses, schists, phyllites, shales, and less abundant marbles, amphibolites, and calcsilicate rocks (Caminos 1979; Toselli et al. 1996). The Ambato metamorphic complex mainly consists of high-grade metasedimentary rocks (migmatites and gneisses) and discordant granitic and pegmatitic bodies (Caminos 1979). Larrovere et al. (2009) obtained an Early to Mid-Ordovician metamorphic age on monazites from a migmatite in the central northern part of the Sierra de Ambato. At the southern tip of the Sierra de Ambato, next to Sierra de Velasco, the Ambato Metamorphic Complex overlies the low-grade successions of the La Cébila Metamorphic Complex across a west-directed Cenozoic reverse fault (Larrovere et al. 2011, 2012). Granitoid rocks intruded the metamorphic basement of both Sierra de Ancasti and Ambato during two major events: the Lower–Middle Ordovician (Famatinian) and Late Devonian–Early Carboniferous (Achalian). The former and more important is characterized by the Capillitas Batholith, an Ordovician peraluminous S-type Famatinian granitoid (Toselli et al. 1996; Pankhurst et al. 2000; Rossi et al. 2005). On the other hand, the Devonian–Carboniferous magmatism in the area is restricted to discordant bodies.

The Famatina Ranges are characterized by widespread Cambro-Ordovician volcanosedimentary successions and Ordovician plutonism. These low-grade rocks accumulated adjacent to a subduction-related magmatic arc (Astini 2003) and are separated into three units: the Negro Peinado, Achavil and La Aguadita Formations (Astini et al. 2005; Collo and Astini 2008). The dominance of Cambrian detrital ages in the Negro Peinado Formation suggests derivation principally from the eastern Pampean Belt, whereas the dominance of late Neoproterozoic ages in the Achavil Formation suggests that input from the Pampean Belt was overwhelmed by older sources (Collo et al. 2009). Detrital zircon ages from the La Aguadita Formation, a synorogenic Ordovician clastic wedge developed along a retroarc foreland basin, indicate that it must be younger than 452 ± 6 Ma (Astini 2003; Astini et al. 2005). The Achavil Formation is unconformably covered by the Late Cambrian–Early Ordovician Volcancito Formation. Calcalkaline metaluminous epizonal granitoids are recognized in the central area of the Famatina Ranges (Dahlquist et al. 2008 and references therein). Contact metamorphism is observed in the Negro Peinado/La Aguadita Formations (Rossi et al. 2005).

The Neoproterozoic–Cambrian basement of south-central La Pampa province is represented by metapelites and metapsammites, and minor quartzites and marbles. These rocks formed in a platform environment on the Río de la

Plata Craton, called the Las Piedras Formation. On the Pampia Terrane, minor outcrops are grouped in the Santa Helena Schist and the Green Quarry Schist. The first one represents a metamorphic unit, whose protolith was deposited after *c*. 556 Ma and prior to *c*. 530 Ma in a foreland basin possibly equivalent to the late Puncoviscana basin (Zappetini et al. 2010). Cambrian metasedimentary rocks are grouped in the Green Quarry Schist, whose protolith was deposited after *c*. 515–500 Ma and prior to *c*. 465 Ma (Chernicoff et al. 2007) in a post-collisional basin (mostly sourced from the Pampean orogen), slightly post-dating Middle- to Late Cambrian sedimentation that occurred in more northern latitudes

The 528 Ma Carancho Igneous Complex (Chernicoff et al. 2012) comprises calcalkaline metadiorites and metagranites, as well as tholeiitic metapyroxenites, and is considered as representing remains of the suture between the Río de la Plata Craton and the Pampia Terrane. Both the Santa Helena Schist and the Green Quarry Schist are the host rocks of Ordovician metaigneous rocks.

16.3 Time Constraints on the Pampean Metamorphism

The maximum depositional age for the sedimentation of the protolith that later constituted the Pampean metamorphic basement is provided by the 560–555 Ma detrital zircon age peak of the Puncoviscana Formation (Adams et al. 2008; Escayola et al. 2011) and equivalent rocks from various sectors (Sierras de Córdoba, the Conlara Metamorphic Complex and the Sierra de Ancasti, Sierra Brava) of the Eastern Sierras Pampeanas and further south in La Pampa province (the Santa Helena Schist) (e.g., Schwartz and Gromet 2004; Steenken et al. 2004, 2006, Escayola et al. 2007; Rapela et al. 2007, 2016; Drobe et al. 2009, 2011; Siegesmund et al. 2010; Zappetini et al. 2010; Ramos et al. 2015).

In the Sierra de Córdoba, complex metamorphic textures and parageneses (Guereschi and Martino 2003, 2008, 2014) suggest the existence of at least two metamorphic regional events (here named M1 and M2). The zircon ages of older rims reported by Sims et al. (1998) indicate a stage of zircon growth between 580 and 560 Ma. In the Sierra de Comechingones the first metamorphic peak produced a high-grade paragenesis (M1) and the first anatexis generating the stromatic migmatites, named the Tala Cruz type (Martino et al. 1994; Guereschi and Martino 2002, 2003, 2008). These rocks yield a concordant SHRIMP U/Pb zircon age of 553.5 ± 3.2 Ma (2σ, $n = 5$; Siegesmund et al. 2010). This age corresponds closely to the *c*. 560 Ma peaks on the detrital zircon patterns of the metaclastic Pampean rocks. In this connection, rocks that were considered as resulting from anatexis of the abovementioned stomatites, the Cañada del Sauce diatexite (M4 after Guereschi and Martino 2008), yielded a population of an igneous zircon concordant cluster at 581 ± 15 Ma (2σ, $n = 3$) (Siegesmund et al. 2010). The calcalkaline signature of this rock led Guereschi and Baldo (1993) to consider it to be an indication of an arc-related magmatism.

The M2 event (810–840 °C and 8.5–9 kbar) was determined in the Sierra Grande, where the metamorphism pre-dating the formation of cordierite was dated by the Sil + Kfs Las Palmas Gneiss, which lacks cordierite and yielded a zircon age of 543.1 ± 3.6 Ma (Siegesmund et al. 2010). Together with datings on the cooling history recorded by a 534 ± 13 Ma K/Ar hornblende age (Steenken et al. 2010), a metamorphic history pre-dating the onset of a lower pressure event is reinforced. Cordierite formation and diatexites resulting from a second melting of the metatexites at 725–780 °C and 5.5–6 kbar occurred at *c*. 530 Ma or even slightly later (Rapela et al. 1998b). Martino (2003) and Otamendi et al. (2005) related the main migmatization event and cordierite formation to the decompressional history of the M2 metamorphic stage, which lasted approximately 10 Ma (Camacho and Ireland 1997; Lyons et al. 1997; Fantini et al. 1998; Gromet and Simpson 1999; Rapela et al. 1998a, b; Sims et al. 1998). However, Ramos et al. (2015) proposed that granulite formation, migmatites and peraluminous granites at *c*. 518 Ma are related to an important heat advection in the crust, which controlled anatexis and cordierite formation in localized areas. Acidic volcanic rocks and peraluminous granitic intrusions that are widespread within the northern portion of the San Carlos massif as well as in the Sierra Norte yielded ages between 530 and 520 Ma (Rapela et al. 1998a, b; Escayola et al. 2007; Siegesmund et al. 2010; Ramos et al. 2015). The early Cambrian M2 event in the Sierra de Comechingones is recorded by the 536 ± 11 Ma Santa Rosa Grt-Crd granulite (Siegesmund et al. 2010). Monazite grains from this granulite yield a mean $^{206}\mathrm{Pb}/^{232}\mathrm{Th}$ age of 507.4 ± 6.1 Ma, supporting PbSL dating of one titanite separate (A20-04), yielding 506 ± 26 Ma. Additional constraints on the younger limit of the metamorphism are provided by PbSL titanite ages obtained from calcsilicate intercalations in the San Carlos Massif, which yielded an age of 505.7 ± 7.3 Ma (Siegesmund et al. 2010). This new Pb/Pb titanite age supports the earlier U–Pb data on titanite at 509 ± 2 Ma (Fantini et al. 1998). This closely follows a U/Pb age of 515 ± 2 Ma for prograde metamorphic monazite in the Guamanes shear zone in the east of the massif (Gromet and Simpson 1999). K/Ar and Ar/Ar muscovite ages starting at 502 Ma were also reported (Krol and Simpson 1999; Steenken et al. 2010). Most of these micas are large muscovite booklets from pegmatites and are considered to have a closure temperature significantly higher than 410 °C (Willigers et al. 2001), as pointed out by Steenken et al. (2008). These ages would mark the end of the last metamorphic episode in the ESP. The mid to low temperature ages that cluster around Mid-Cambrian times denote a short-lived cooling, and hence an accelerated exhumation of the Pampean orogen at around 505 Ma and a fast formation of the Famatinian arc (Ducea et al. 2017). This correlates with the first activation of the different compressional shear zones along the western margin of the basement complex, the Guacha Corral and Los Túneles shear zones (Steenken et al. 2010), and with the initiation of compression along the Gondwana margin (Steenken et al. 2006 and references therein) and, albeit more speculative, with the Irúyica deformation phase (Astini et al. 2008).

Therefore, portions of the basement of the Sierra de Córdoba might be older than the ubiquitous metamorphic equivalents of the Puncoviscana Formation. Geochemical results may indicate a calcalkaline granodioritic composition for the Cañada del Sauce diatexite (Guereschi and Baldo 1993), relating this intrusion to an active continental margin. The first metamorphic (M1) overprint took place at *c*. 560 Ma, whereas the more extended prograde event (M2) is constrained between 540 and 533 Ma.

Post-liminary fusion (migmatization), granulite formation (i.e., M3, M4 Steenken et al. 2011) and derived peraluminous magmatism are younger than 530 Ma and probably unrelated to the decompressional evolution of the main clockwise metamorphic overprint. Table 16.1 gives a summary of the Pampean metamorphism.

16.4 Time Constraints for the Famatinian Metamorphism

Famatinian metamorphism is mostly related to the closure of ensialic back-arc or inter-arc basins developed east of or inside the main Famatina magmatic arc (Balhburg 1991; Steenken et al. 2006; Dahlquist et al. 2008; Collo et al. 2009 and references therein).

Steenken et al. (2004) were the first to recognize at least two different sedimentary protoliths (P) in the Eastern Sierras Pampeanas: (P1) a *c*. 560 Ma source and (P2) a younger post-Pampean. Steenken et al. (2006) suggested resumption of subduction and sedimentation in the west of the Pampean terrane in the Late Cambrian and the subsequent amalgamation of the Cuyania and/or Precordillera Terrane during the Famatinian orogenic cycle (Ramos et al. 1986; Sims et al. 1998). Drobe et al. (2009) indicate that extensional tectonics after the Pampean metamorphism controlled the development of the back-arc basins in which the protoliths of the Nogolí and Pringles metamorphic complexes were deposited. Age constraints between 500 and 480 Ma for the (ultra)mafic rocks (Sims et al. 1998; Hauzenberger et al. 2001; Steenken et al. 2008) of the Pringles Metamorphic Complex argue for a thermal input and LP/HT metamorphism in a thinned crust. On both sides of the mafic to ultramafic bodies a metamorphic gradient from granulite to greenschist facies is observed (Delpino et al. 2007), and therefore a contact metamorphic origin for the granulite facies paragenesis was suggested (Hauzenberger et al. 2001; Steenken et al. 2005, 2006). Ducea et al. (2010) proposed that high-grade Famatinian metamorphism and migmatization were synchronous with 485–465 Ma magmatic emplacement. Inherited ages in some of the plutonic rocks as well as detrital zircons in the metasedimentary framework suggest that the Famatinian arc was emplaced into a thick miogeoclinal cover to the thinned margin of the proto-South American continent. Therefore it could be proposed that the Famatinian orogen consists of two stages of compression: one related to the onset of subduction at around 500 Ma and another stage related to docking of the Precordilleran terrane at 480 Ma (Ramos et al. 1998) and the cessation of the Famatinian magmatism.

Table 16.2 gives a summary of the P–T constraints and ages of the Famatinian metamorphism.

16.5 Magmatism in the Eastern Sierras Pampeanas

Rapela et al. (1990) separated granitoids of the Sierras Pampenas into G1, G2 and G3, which roughly correspond in time to the Pampean (Ediacaran to Cambrian), Famatinian (Ordivician) and Achalian (Devonian to Carboniferous) orogenies. Pampean magmatism is arc-related with minor post-collisional peraluminous intrusions and minor acid volcanic rocks. Famatinian magmatism is also arc-related. I-type granitoids occur in the west and grade into S-type towards the continent. Devonian magmatism would be either arc-related with a Devonian subduction along the western margin of the Precordillera Terrane or the result of post-Ordovician uplift.

16.5.1 Ediacaran to Cambrian Granitoids

Calcalkaline magmatic rocks of Ediacaran and Cambrian age form the batholith of Sierra Norte and extend discontinuously southwards along the Sierra Chica (Rapela et al. 1998b) and in Sierra de Guasayan (Dahlquist et al. 2016; Fig. 16.2). These rocks consist of calcalkaline metaluminous to weakly peraluminous granodiorites and monzogranites that intruded a metasedimentary host. Aplites, porphyries and pegmatites intrude all the sequences (Rapela et al. 1991). Late-stage rhyodacites and miarolitic monzogranites are also present (Lira et al. 1997). Geochemistry suggests (Lira et al. 1997) subduction related magmatism and convergent margin tectonics along the western margin of Gondwana in the late Neoproterozoic. Geochronological studies from the Sierra Norte granitoids (Schwartz et al. 2008) indicate a period of emplacement bracketed between 550 and 530 Ma. Cambrian granitoids (the G1b suite of Rapela et al. 1998a, b) located in the Sierra Grande and Sierra de Comechingones are peraluminous anatectic melts of the metasedimentary country rocks (Rapela et al. 1998a, b). The ages of these groups of plutons is bracketed between 523 ± 2 Ma for El Pilón (Rapela et al. 1998a, b, 2001) and 529 ± 3.4 Ma for the Juan XXIII pluton (Escayola et al. 2007). The 521.1 ± 1.6 Ma Oncan rhyolite cuts the older granites of the Sierra Norte and is considered to be related to the Cambrian peraluminous granites (Iannizzotto et al. 2013; Ramos et al. 2015).

Cambrian magmatism is represented further south in the La Pampa province by the 528 Ma El Carancho Igneous Complex (Chernicoff et al. 2012), which comprises calcalkaline metadiorites and metagranites, as well as tholeiitic metapyroxenites, and is considered as representing remains of the suture between the Río de la Plata Craton and the Pampia Terrane. Alternatively, this complex might represent the mantle-derived rocks related to the slab break-off that post-dated the M2 Pampean metamorphic event.

16.5.2 Ordovician Magmatism

The Famatinian magmatic belt was short-lived (*c.* 20 Ma) and without a significant asthenospheric contribution (Pankhurst et al. 1998; Dahlquist et al. 2008 and references therein). Three distinct 484–463 Ma Famatinian granite types were identified (Pankhurst et al. 2000): a dominant I-type, a small-scale S-type and a tonalite-trondhjemite-granodiorite. I-type magmatism is represented by tonalites, granodiorites, minor monzogranites and gabbros that can be traced from the northwest in Catamarca and La Rioja down to the Sierras de Valle Fértil-La Huerta in the San Juan province and the Sierra de San Luis. The S-type granitoids are developed in Sierra de Velasco, Sierra de Chepes and in a sector of the Sierra de San Luis (Steenken et al. 2006, 2008; Grosse et al. 2011). The TTG group (i.e., El Hongo, Calmayo, La Fronda, Guiraldes, La Playa and Paso del Carmen tonalite to granodiorite plutons) is emplaced in the Sierra Grande except for the Calmayo group (500–480 Ma, U–Pb zircon; Rapela et al. 1998a, b), which intrudes the Sierra Chica Complex.

López de Luchi et al. (2007) considered that the Late Cambrian-Ordovician granitoids (*c.* 500–470 Ma) of the Sierra de San Luis (Fig. 16.2) are synkinematic with compressive deformation related to Famatinian convergence. They proposed a separation of these granitoids into an Ordovician tonalite suite (OTS; metaluminous to mildly peraluminous calcic tonalite-granodiorites) and an Ordovician granodiorite-granite suite (OGGS; peraluminous calcic to calcalkaline granodiorite-monzogranites).

Lower Ordovician metaluminous calcalkaline granitoids, which constitute the principal lithologies of the Sierras de Chepes, Malanzán and Los Llanos (Fig. 16.2), frequently exhibit mingling relationships with gabbro diorite or quartz diorite. The most important granitoid unit is the 490 ± 5 Ma (U/Pb zircon age) Chepes Granodiorite (Pankhurst et al. 1998).

Weakly to strongly foliated (481 ± 3 Ma, Pankhurst et al. 2000; 481 ± 2 Ma, Rapela et al. 2001) peraluminous metagranitoids are the dominant lithology of the Sierra de Velasco (Báez et al. 2005; Toselli et al. 2005). Subordinate varieties include strongly peraluminous porphyritic biotite-cordierite metamonzogranites and moderately peraluminous coarse- to medium-grained biotite metagranodiorites and metatonalites. In the southern part of the Sierra de Velasco, the main lithologies are metaluminous to weakly peraluminous biotite-hornblende metagranodiorites and metatonalites (Bellos 2005).

In the Sierra de Ancasti, Ordovician granitoid magmatism is represented by calcalkaline metaluminous plutons, minor garnet-bearing two-mica granite stocks (Reissinger 1983; Rapela et al. 2005) and tourmaline- and beryl-bearing pegmatites that are emplaced in the Sierra Brava Complex and in the Ancasti Formation (Dahlquist et al. 2010, 2012). The metaluminous Las Cañadas Complex was emplaced close to 468 Ma (Dahlquist et al. 2012).

Calcalkaline metaluminous epizonal granitoids are recognized in the central area of the Famatina Ranges (Dahlquist et al. 2008 and references therein). Rapela et al. (1999) calculated a U–Pb SHRIMP zircon age of 484 ± 5 Ma for the Cerro Ñuñorco biotite granite. Rubiolo et al. (2002) determined a U–Pb age of 485 ± 7 for the Narváez Granite. Dahlquist et al. (2008) calculated U–Pb zircon SHRIMP ages of 481 ± 4 Ma for a tonalite of the weakly peraluminous Cerro Toro Complex and of 463 ± 4 Ma for a monzogranite of the Ñuñorco Complex. Rhyolites interbedded with sediments of a volcanic sedimentary sequence yielded 477 ± 4 Ma.

Further south, in the southernmost tip of the Sierras Pampeanas, both the Santa Helena Schist and the Green Quarry Schist are found intruded by Ordovician metaigneous rocks: the 475.7 ± 2.3 Ma (U–Pb zircon SHRIMP age) Paso del Bote metaquartz-diorites and associated rocks

(Chernicoff et al. 2010) and the *c*. 450 Ma (U–Pb zircon SHRIMP age) back-arc-related Valle Daza metagabbros (Chernicoff et al. 2009). Both Famatinian rock assemblages define narrow, roughly N-S trending belts.

16.5.3 Devonian to Lower Carboniferous Magmatism

Paleozoic magmatism in the Sierras Pampeanas ended with the intrusion of a suite of Middle Devonian to Lower Carboniferous batholiths (López de Luchi 1996; Stuart-Smith et al. 1999; Pinotti et al. 2002; Siegesmund et al. 2004). They were considered to be either post-Famatinian or post-orogenic (Rapela et al. 1998a, b) or the Achalian granite group. In the Sierra de Córdoba, the Devonian Achala Batholith is a peraluminous, alkali-calcic granitoid with sharp and discordant contacts, which would represent variable proportions of a juvenile mantle component and crustal melts formed by dehydration melting of biotite-bearing gneisses (Rapela et al. 2008). The Cerro Aspero Batholith in Sierra de Comechingones is made up of biotite-granite and leucogranites (Pinotti et al. 2002). In the Sierra de San Luis, voluminous Devonian batholiths (Las Chacras-Potrerillos, Renca, La Totora and El Hornito) are made up of an I-type hybrid monzonite-granite suite with metaluminous alkali-calcic (393–385 Ma) monzonite-quartz monzonite-granodiorite ± monzogranite and peraluminous alkali-calcic monzogranites (Fig. 16.3a, b; López de Luchi et al. 2007, 2017). Undeformed granites of Lower Carboniferous age (Dahlquist et al. 2008) are located in the north (Báez et al. 2005) and central (Grosse et al. 2008) parts of the Sierra de Velasco. They have some features in common with the Devonian granitoids, in particular, nearly circular shapes, general lack of pervasive solid-state deformation, discordant relationships with the host, shallow emplacement, syeno- to monzogranitic compositions, porphyritic textures and an abundance of pegmatites and aplites. Miller and Söllner (2005) proposed that the Devonian to Early Carboniferous magmatism is related to post-orogenic regional crustal heating, whereas Grosse et al. (2008) proposed that the general age decrease in the Achalian magmatism from south to north could have been related to a progressive delamination of the crust coupled with upper mantle upwelling from south to north.

16.5.4 Neoproterozoic-Ordovician (Ultra-)Mafic Rocks

Mafic to ultramafic units of the Sierras de Córdoba which trend NNW to NE and dip between 50°–70° to the east were separated into an eastern and a western belt (Escayola et al. 1996). The NNW-SSE trending eastern belt is composed of lherzolite with interlayered websterite and subordinate harzburgite, abundant pyroxenite and gabbros. Escayola et al. (1996) assigned the tectonic setting to an ensialic back-arc basin, whereas Ramos et al. (2000) interpreted the belt as ophiolite remnants of a back-arc zone. The mafic to ultramafic units of the western belt, which extend along *c*. 300 km, are made up of harzburgites that were metasomatized by intrusive basaltic dykes (Escayola et al. 2007). Escayola et al. (1996) interpreted the western belt as a suture zone and considered the ultramafic rocks as dismembered ophiolites with MORB affinities, whereas Escayola et al. (2007) suggested a suprasubduction zone ophiolitic complex formed at a back-arc stage. Sm–Nd dating of basalt and gabbro dykes, pyroxenites and impregnated peridotites of the western belt yielded an isochron age of 647 ± 77 Ma. The $\varepsilon\mathrm{Nd}_{(647)}$ value of 5.2 is consistent with an oceanic or back-arc origin for the ophiolite sequence (Escayola et al. 2007).

Gabbros with OIB signatures located in the Sierra de Comechingones, which result from the subduction of a mid-ocean ridge beneath the paleo-Pacific Gondwana margin, are related to the high-grade metamorphism of the Sierras de Córdoba (Tibaldi et al. 2008).

In the Cambrian Pampean Conlara Metamorphic Complex, ultramafic rocks have not been reported yet. Scarce amphibolites are distributed in a N-S belt through Sierra del Morro, San Felipe and Villa de Praga. Tholeiitic metapyroxenites of the 528 Ma Carancho Igneous Complex (Chernicoff et al. 2012) were interpreted as slivers of arc and back-arc rocks related to the inferred Late Early Cambrian suture between the Río de la Plata Craton and the Pampia Terrane.

Mafic to ultramafic Ordovician complexes are developed in the central part of the Sierra de San Luis, whereas gabbros are recognized as small plutons associated with the calcalkaline granitoids of the Sierra de Chepes, in the Famatina Complex in the Sierra de Famatina and in La Pampa province as scattered outcrops (Fig. 16.2). In the Pringles Metamorphic Complex, mafic to ultramafic bodies for which back-arc or marginal basin settings were proposed (Sims et al. 1998; Steenken et al. 2008, 2011) are represented by norites to gabbro-norites with minor ultramafic rocks. They have ages between 506 and 480 Ma. Amphibole-bearing gabbros associated with the calcalkaline complexes of the Sierra de Chepes, e.g. the Tama gabbro exhibit mingling relationships with the host granodiorite (Pankhurst et al. 1998).

Combined geological, geochronological, geochemical and geophysical studies have led to the identification of a large (*c*. 300 km long, *c*. 5 km wide) N-S trending belt of metagabbros in the province of La Pampa. This belt, though only poorly exposed in the localities of Valle Daza and Sierra de Lonco Vaca, stands out in the geophysical data (aeromagnetics and gravity) (Chenicoff et al. 2010). The main rock type is metagabbro with relict magmatic nucleii where layering is preserved. A counterclockwise P–T evolution affected these rocks, while during the Middle Ordovician the protolith reached an initial granulite facies metamorphism (M1), evolving to amphibolite facies (M2).

16.6 Isotopic Constraints for the Neoproterozoic-Early Paleozoic Geodynamic Evolution of the Gondwana Margin

16.6.1 Sm–Nd Fingerprints for the Metamorphic Rocks

Steenken et al. (2011) suggested that the T_{DM1} = 1.8–1.7 Ga and $\varepsilon Nd_{(540)} = -6$ to -8 can be considered as the mean average crustal composition of the Eastern Sierras Pampeanas (Fig. 16.3). Younger T_{DM1} ages and more radiogenic εNd values in metaclastic rocks could reflect the addition of juvenile material, a different source or a petrogenetic process. Although detrital zircon age patterns for the Pampean metaclastic rocks exhibit prominent Mesoproterozoic and Brasiliano peaks (Fig. 16.4), those for the post-Pampean metamorphic rocks independent of their metamorphic grade show Pampean and a less pronounced Grenvillian peak, suggesting a limited Grenvillian input. The average interval of crustal residence ages is similar. Since the Sm–Nd model ages (Fig. 16.3) are older than these peaks, the magmatic-metamorphic events sampled by the zircon cores/rims mainly represent recycled crustal sources.

Metamorphic rocks of the same grade from different parts of the Eastern Sierras Pampeanas show similar T_{DM1} ages (Fig. 16.3a), and crustal $f_{Sm/Nd}$ values but less radiogenic $\varepsilon Nd_{(540)}$ for those with older metamophic ages, which implies a higher degree of recycling between the Pampean and the Famatinian metaclastic rocks. Therefore increasing the amount of zircons from Grenville sources introduces a relatively more positive epsilon (Steenken et al. 2011). The La Cébila Metamorphic Complex at Quebrada La Rioja is not part of this trend, since T_{DM1} values vary between 1.3 and 1.5 Ga, and $\varepsilon Nd_{(t)}$ is less negative (Larrovere et al. 2011). Interestingly, the 526 Ma youngest age peak (Fig. 16.4) calculated for samples from this area is defined by only 20 zircons.

Increasing metamorphic grade in rocks with similar sources and detrital ages as in the Sierras de Córdoba is associated with a younging of the T_{DM} ages and less negative $\varepsilon Nd_{(540)}$ values. Pampean high-grade metaclastic rocks of the Sierra Grande de Córdoba and Sierra de Comechingones (Fig. 16.3) show consistently younger T_{DM} ages of 1.6–1.5 Ga and $\varepsilon Nd_{(540)}$ values between −4 and −3, whereas the $f_{Sm/Nd}$ value is above the average of −0.4 for the continental crust. Although SHRIMP zircon data for the granulite of the Sierra de Comechingones is scarce, the persistence of the Mesoproterozoic peaks (Fig. 16.4) in the Pampean units (Siegesmund et al. 2010) suggests that younging of the T_{DM1} ages results from a petrogenetic process. Otamendi et al. (2006) proposed that the granulites formed by a short-lived event of local thermal input, which enhanced the possibility that melting was rapid enough to trigger isotopic disequilibrium. The Ordovician high-grade rocks of the Pringles Metamorphic Complex exhibit T_{DM1} ages of 1.7–1.6 Ga (Fig. 16.3a) and a mean $\varepsilon Nd_{(540)}$ value of −5.5 (Fig. 16.3a). To summarize, increasing metamorphic grade in rocks with similar sources and ages as in the Sierras de Córdoba is associated with a younging of the T_{DM} ages and less negative $\varepsilon Nd_{(540)}$ values. In metaclastic rocks of different metamorphic age, the Grenville source introduces a relatively less negative epsilon. The change in source between the Pampean and the Famatinian metaclastic rocks is related to a higher degree of recycling as evidenced by the less radiogenic epsilon.

16.6.2 Sm–Nd Fingerprints for the Magmatic Rocks

16.6.2.1 Mafic-Ultramafic Complexes

Sm/Nd isotopes and Nd model ages of whole rock samples give the crustal residence age, this T_{DM} age corresponds to the actual mantle separation age of the crustal material (Patchett 1992). However, in the case of igneous rocks directly derived from the mantle, as can be expected for the (ultra-)mafic rocks, the T_{DM} model age could be close to the crystallization age if the rocks were stabilized over a period of up to 100 Ma (Patchett 1992) and thus provide the age of formation of the rock. Based on the analysis of the Sm–Nd results from various authors, Steenken et al. (2011) indicate two types of Pampean-related mafic rock in which no interaction with the continental crust in the sense of mixing or assimilation process is recognized. One group shows a depleted mantle-like signature and LREE depleted sources, which could indicate a stage of ocean crust formation, and the other, younger group has a signature of enrichment (cf. Siegesmund et al. 2010).

Rocks of the Eastern belt located in the Sierra Chica de Córdoba (Fig. 16.2) can be separated into two groups: (1) mafic to ultramafic rocks with a depleted mantle signature with $\varepsilon Nd_{(today)}$ values around 11, $\varepsilon Nd_{(540)}$ values of 6–8.5 and positive $f_{Sm/Nd}$; and (2) another more variable group located in the northern sector of the Sierras de Córdoba in which amphibolites are characterized by negative $\varepsilon Nd_{(today)}$ values,

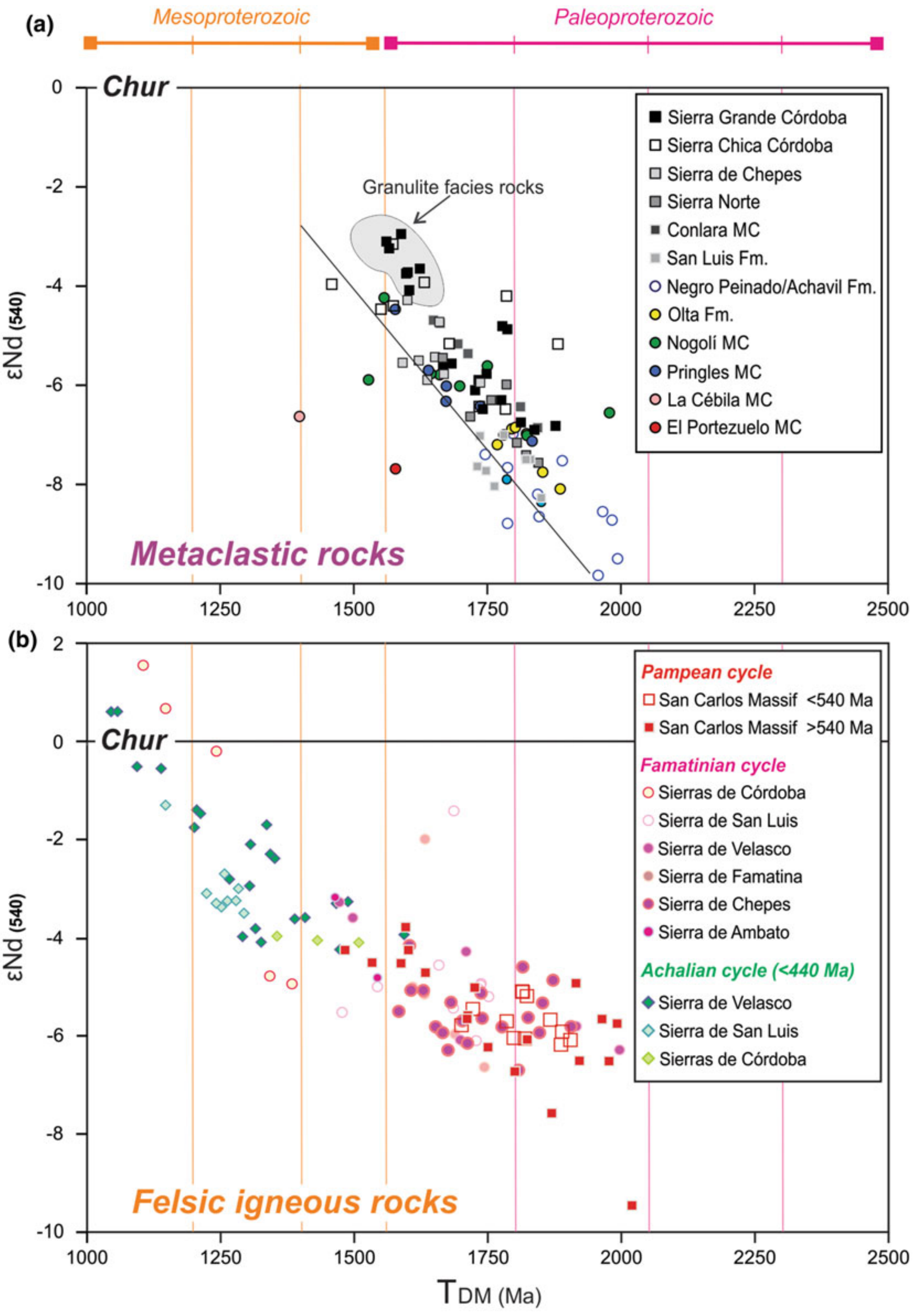


Fig. 16.3 **a** $\varepsilon Nd_{(t)}$ versus T_{DM} diagrams showing the Eastern Sierras Pampeanas clastic and felsic igneous rocks. Most of the metamorphic rocks are restricted to crustal values of fSm/Nd Pampean high grade rocks metaclastic rocks show more radiogenic compositions. Felsic igneous rocks exhibit more ample the data and the overlapping of the Ordovician and Pampean granitoids. Pampean rocks older than 540 Ma separate into two groups. The group with $\varepsilon Nd_{(t)} > -5$ corresponds to the rocks that show $T_{DM} < 1.6$ Ga. Rocks with less radiogenic $\varepsilon Nd_{(t)}$ correspond to older T_{DM} which suggest recycling nevertheless $\varepsilon Nd_{(t)}$ is more radiogenic in comparison with metaclastic rocks of similar T_{DM}. Devonian granitoids exhibit the more radiogenic signature. **b** $\varepsilon Nd_{(t)}$ versus fSm/Nd showing the Eastern Sierras Pampeanas mafic rocks in relation to the depleted mantle (DM) and the enriched reservoir represented by the oldest recorded crust i.e. T_{DM} between 1.7 and 1.8 Ga. Any mixture between these two reservoirs should lie along the mixing line (crust incorporation-thick line). Field limited for dashed line corresponds to the dominant values for the Eastern Sierras Pampeanas continental crust. See text for further explanation

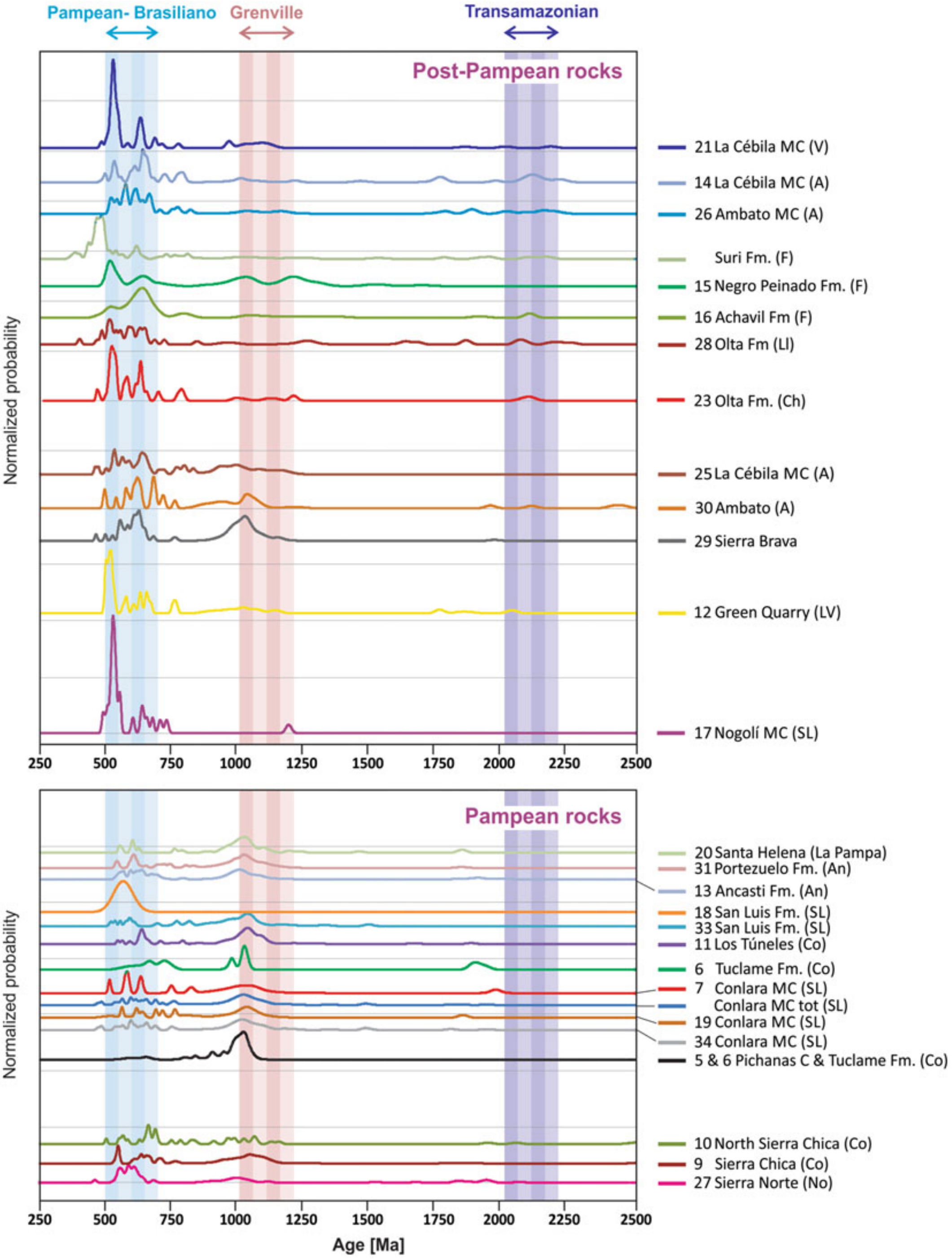


Fig. 16.4 Comparison of the detrital zircon provenance of the metaclastic rocks of the Pampean and post-Pampean basins using normalized zircon plots (Gehrels 2014). Data are display ordered from east (bottom side of the each figure) to west. Note that although the dominant peaks are of Brasiliano and Grenvillian ages, the relative proportion of each zircon population changes. See text for comments. Sample codes as in Fig. 16.2

positive $\varepsilon Nd_{(540)}$ values and negative $f_{Sm/Nd}$. T_{DM2} ages vary between 570 and 560 Ma. In the western belt in the Sierra de Comechingones, two groups can be recognized (Fig. 16.2). One includes serpentinites and metabasaltic dykes in which $\varepsilon Nd_{(today)}$ values can be higher or lower than the depleted mantle. The other group is made up of the gabbros related to the granulite facies metamporphism, such as Intihuasi, Suya Taco and Sol de Mayo in which the $\varepsilon Nd_{(today)}$ values are significantly lower than that of the contemporary depleted mantle. If the serpentinites and peridotites are discarded owing to the complex processes involved in their genesis, the remaining primitive rocks yield a T_{DM2} age range of 700–600 Ma. Therefore formation of oceanic crust in the Sierra Chica could have been active up to 560 Ma, whereas in the Sierra de Comechingones formation of oceanic crust was active up to 700–600 Ma (Steenken et al. 2011).

Ultramafic rocks of the Pringles Metamorphic Complex exhibit positive $\varepsilon Nd_{(480)}$ with T_{DM2} ages between 1100 and 800 Ma, whereas the norites with a negative $\varepsilon Nd_{(480)}$ plot on a mixing line (Fig. 16.3b) connecting the DM signature at 600–500 Ma with the average values for the metaclastic rocks of the complex. Amphibolites and komatiitic basalts of the Nogolí Metamorphic Complex, which show negative $\varepsilon Nd_{(480)}$ and $f_{Sm/Nd}$ values, and T_{DM2} ages between 1500 and 1300 Ma, suggest a process of mixing with a crustal component. A T_{DM2} age of 1100 Ma is calculated for the mafic rocks with positive $\varepsilon Nd_{(480)}$. The only available data for the gabbros of Sierra de Famatina and Sierra de Chepes exhibit negative $\varepsilon Nd_{(480)}$ and T_{DM2} ages between 1600 and 1500 Ma. The $\varepsilon Nd_{(480)}$ values for the Valle Daza gabbro are positive and show T_{DM2} ages of 800 Ma. Therefore the geodynamic scenario for the Ordovician mafic rocks could imply thicker continental crust than for the emplacement of the Pampean mafic rocks or, alternatively, a fast extensional process (extensional collapse) could prevent modification of the mafic melts in their ascent to the emplacement level.

The less evolved Devonian magmatic rock of the Las Chacras batholith indicates a T_{DM} age younger than any granitoid of the Sierra de San Luis (Fig. 16.3b) coupled with an εNd value of around zero.

16.6.2.2 Felsic Rocks

The overall high abundance of felsic intrusive rocks and the paucity of intermediate compositions suggest that crystal fractionation of mafic parental magmas was not an important process. Partial melting of crustal sources during basalt underplating or during crustal delamination is the dominant process at least during the Pampean and Famatinian cycles. Pre-540 Ma (Figs. 16.2 and 16.3), granitoids of the Sierra de Córdoba make up two clusters: one that would be entirely crustal, since they have T_{DM} ages older than 1.75 Ga and $\varepsilon Nd_{(540)}$ values of less than −5, and another cluster which corresponds to samples from Sierra Chica de Córdoba in which $\varepsilon Nd_{(540)}$ values are more radiogenic and T_{DM} ages are younger. In the first cluster, melting of a large amount of old sedimentary protoliths can be suggested, whereas in the second, since the host is less radiogenic and older, melting of a variably rejuvenated crust could be assumed. In this area, tonalites and granodiorites carry dioritic enclaves, suggesting a mixture of a mafic precursor with metasedimentary sources (Siegesmund et al. 2010; Iannizzotto et al. 2013).

In spite of the episode of mafic magma input to the crust as shown by the *c.* 540 Ma gabbros of the Sierra de Comechingones, no evidence of this juvenile input is seen in the post-540 Ma Pampean felsic rocks, which would represent partial melts derived from the older than 1.75 Ga basement (Fig. 16.3). Therefore it could be proposed that Cambrian OIB mafic magma should have just controlled the melting process through delamination or underplating.

Although juvenile input is suggested by the mafic rocks in the Pringles Metamorphic Complex or the gabbros of the Sierra de Chepes, widespread Famatinian magmatism with the exception of the minor TTG suites reworked the old lithospheric sources (Pankhurst et al. 1998). Mostly evolved granitoids dominate the rock spectrum in the Famatinian subduction stage. This fact suggests melting of mostly crustal sources as expected in regions of thick continental crust.

Sources for TTG suites of the Sierra de Córdoba are younger. Data from the Guiraldes and La Fronda throndhjemite together with a leucogranite from the northern sector of Sierra Chica de Córdoba cluster at a T_{DM} age of *c.* 1.3 Ga (see Fig. 16.3), which is accompanied by an $\varepsilon Nd_{(540)}$ value of −5 and a $f_{Sm/Nd}$ value of −0.55. One throndhjemite of the San Carlos Massif and the La Playa granodiorite shows the youngest T_{DM} age and a positive or slightly negative value of $\varepsilon Nd_{(470)}$, which suggests a juvenile input (Steenken et al. 2011).

16.7 Detrital Zircon Constraints on Protoliths of the Metaclastic Sequences

All 28 detrital zircon samples included in this study were compiled from the literature (see captions for references). Data was filtered by accepting values of less than 10% discordant for ages younger than 800 Ma. All the provenance and maximum depositional age analyses are based on published data that was obtained by different sampling strategies and numbers of points per sample. Differences in the method of age peak calculations, number of data, errors and strategy of measurement have a strong influence on the results, and thus interpretations must be careful and combine several analytical approaches.

To simplify the process of statistical provenance analysis, samples were separated into those belonging to Pampean and post-Pampean basins following first the idea proposed by Steenken et al. (2004) in which the post-Pampean basins have detrital zircons younger than 530 Ma. Post-Pampean basins were further separated on the basis of a westwards rejuvenation of the protoliths of the Famatinian back-arc metamorphic complexes owing to the development of less stable basins associated with the emplacement to the west of the Ordovician orogenic front on the Gondwana margin (Larroverre et al. 2012).

Samples were analysed using normalized plots, the K–S test, the age pick 2010 test and by plotting the distribution of the difference between the measured crystallization ages (CA) of individual zircon grains present in the sediment and the depositional age (DA) of the (meta)sediment (Cawood et al. 2012)

Normalized probability plots are formed by calculating a normal distribution for each age from the data, summing the probability distribution of all accepted analyses and dividing the area under the curve by the number of analyses. Peak heights are a function of the number of grains at a particular age and the precision associated with that analysis. On normalized probability diagrams, significant age populations are only defined as clusters with three or more overlapping analyses (Gehrels et al. 2006).

The K–S test (Kolmogorov-Smirnov test) is a statistical approach that compares the maximum probability difference between two cumulative distribution function (CDF) representations of the age spectra. The test is a measure of the percentage of statistically indistinguishable spectra in a given grouping. Dependence on the CDF, which sums probabilities with increasing age, results in heightened sensitivity to the relative abundance of age peaks rather than their presence or absence. This characteristic caused samples with fewer grains or skewed peak abundances to be rejected (Laskowski et al. 2013).

The age pick test was performed to identify the statistically significant peaks in a set of detrital zircon ages. The youngest peak is considered to be a maximum depositional age.

Cumulative proportions were calculated using as the maximum depositional age the youngest calculated peak and as the minimum depositional age the fossil record or the age of the oldest intrusion.

On the normalized plots, samples are ordered from east to west based on the growth of successive Pampean and Famatinian belts towards the west from the margin of the Río de la Plata Craton. As previously mentioned, in various provenance studies the two main provenance populations are from Grenvillian and Brasiliano sources (e.g., Schwartz and Gromet 2004; Steenken et al. 2006; Escayola et al. 2007; Rapela et al. 2007; Drobe et al. 2009, 2011; Ramos et al. 2010).

Samples from the easternmost sector of the Eastern Sierras Pampeanas (Fig. 16.2) exhibit a predominance of Brasiliano sources while towards the west, Grenvillian sources exhibit a more prominent normalized plot peak, as for example point 6 in Fig. 16.4. This observation was first mentioned by Escayola et al. (2007), who considered a Grenvillian source towards the west. This feature of increasing the presence of Grenvillian sources in the detrital record is also observed in the rest of the Pampean basins, except for one sample from the San Luis Formation, A93-05 (Drobe et al. 2009), which has a small amount of data.

The available data for the retroarc post-Pampean Famatinian basins from San Luis, e.g. the Nogolí Metamorphic Complex and the fine-grained schist of the Green Quarry exhibit a well-defined Brasiliano-Pampeano peak and a less pronounced Grenvillian one. In the remaining three samples, data from the Sierra Brava Complex and one sample from the Ambato Complex are more similar to the Pampean basins, whereas the Ambato Formation sample has a predominance of Brasiliano-Pampeano peak. All these detrital patterns show peaks at *c*. 500 Ma or older than 480 Ma.

The Famatinian arc basins are characterized by the Brasiliano-Pampean peaks and the peaks between 520 and 480 Ma or younger in the case of the Suri Fm. Verdecchia et al. (2011) mentioned that the presence of a Paleoproterozoic age peak between 2.2 and 2.0 Ga, which is an indication of a Río de la Plata Craton source, indicates that sedimentation post-dates the juxtaposition with the craton which would have occurred after the main Pampean tectonothermal event (530–520 Ma). This peak is exhibited by the Olta, La Cébila and Achavil Formations (Fig. 16.4). Nevertheless, all these detrital records exhibit peaks in the normalized plots between 500 and 480 Ma, a feature that is preserved even if the Río de la Plata Craton imprint is absent from the rest of the post-Pampean basins.

An alternative explanation is that the exhumed Pampean rocks formed a drainage divide that blocked westward transport of the Río de la Plata-derived sediments. Intense erosion owing to an unstable tectonic scenario led to the progressive appearance of the Río de la Plata signature.

Application of the K–S test to the Pampean and post-Pampean basins indicates a high degree of correlation, which would suggest common/recycled sources. Detrital zircon data from Pampean basins exhibits the highest internal correlation. This may suggest a relatively stable availability of sources, whereas post-Pampean basins show, on average, lower values, which could be an indication of greater instability or barrier erosion (Figs. 16.4 and 16.5; Table 16.3)

Age peak analysis results can be seen in Table 16.4, together with the total number of zircons and the number of zircons that define the peak. Only samples with more than 25 zircon data were used. Based on the calculated peak ages, sample POM14013 from the Ambato Metamorphic-Igneous Complex and SBR14001 from the Sierra Brava

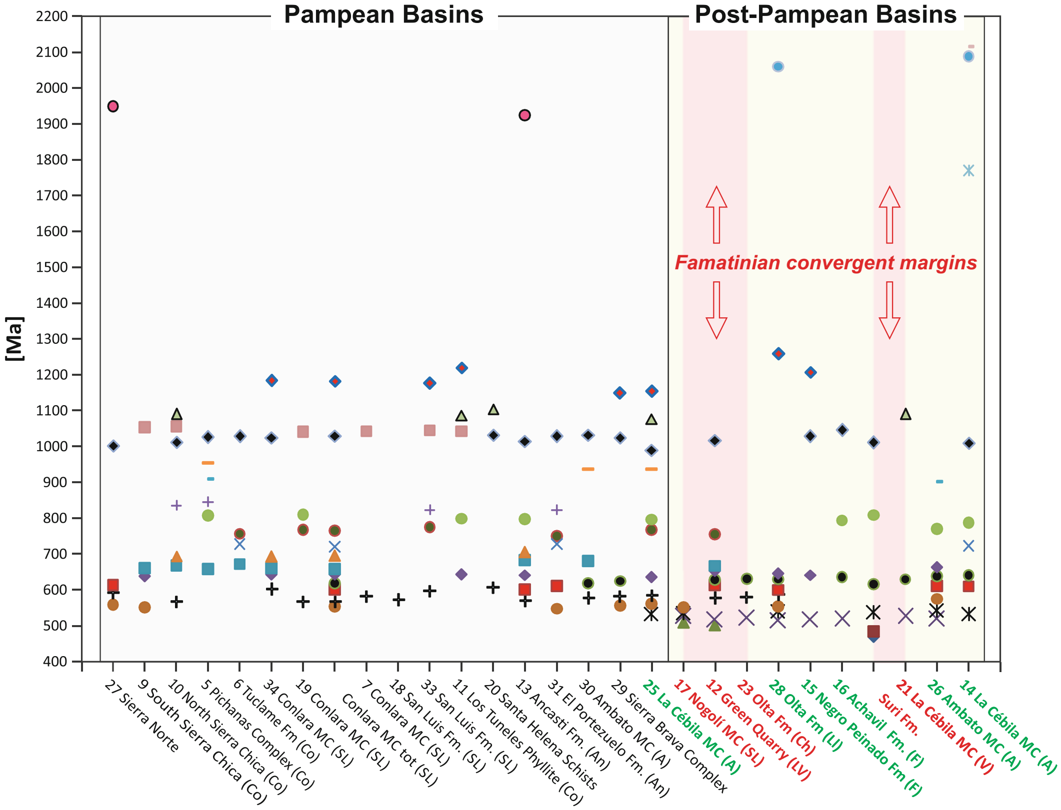


Fig. 16.5 Age picks of the detrital zircons of the metaclastic rocks derived from the Pampean and post Pampean basins. Dominant peaks are of Brasiliano and Grenvillian ages, Pampean ages appear in the post Pampean basins. Note that although several Post Pampean basins have a record of Ordovician sedimentation based not only on the youngest zircon data but also in the fossil record, the only Famatinian Ordovician peak corresponds to the Suri Formation (sample FAM 7082; Rapela et al. 2007). Sample codes as in Fig. 16.2

Metamorphic Complex are akin to the Pampean basins, since they lack younger peaks despite having younger detrital ages, as can be seen in the normalized plots (Fig. 16.4). All the samples exhibit Brasiliano peaks, which correspond to the mean peaks as defined by de Brito Neves et al. (2014; Early Cryogenian to Middle Ediacaran ages for the Pampean basins plus the Late Ediacaran–Middle Cambrian peak for the post-Pampean basins) and Grenvillian peaks. The Río de la Plata peaks are only present in one of the samples of the Olta Formation and in the sample of the La Cébila Metamorphic Complex. This corresponds to the one that Verdecchia et al. (2011) interpreted as an indication of its sedimentation post-dating the juxtaposition of the Río de la Plata Craton. Figure 16.5 is a graphical representation of the age peaks in which the importance of Late Cambrian to Early Ordovician peaks towards the west stand out.

Detrital zircon spectra reflect the tectonic setting of the basin in which they are deposited (Cawood et al. 2012). Convergent plate margins are characterized by a large proportion of zircon ages close to the depositional age of the sediment, whereas sediments in collisional, extensional and intracratonic settings contain greater proportions of older ages that reflect the history of the underlying basement. The difference between the measured crystallization ages (CA) of individual zircon grains present in the sediment and the depositional age (DA) of the sediment provides an indication of the type of margin. The youngest peak calculated from the Age Pick calculus routine (Table 16.4; Fig. 16.5) values was used as the CA, whereas the DA was obtained from the age of cover sequences, plutons intruding the units or the fossil record (Fig. 16.6).

The analysis of the cumulative distribution function (Table 16.4; Fig. 16.6) of highly concordant detrital zircon ages suggests, from east to west, that:

- ‘Pampean’ basins (sensu Steenken et al. 2011) are collisional, they are foreland basins except for Sierra Norte metaclastic host rocks that yield a convergent setting, and Sierra Brava and one of the samples from the Conlara Metamorphic Complex that would correspond to a collisional-convergent margin.
- Results from the post-Pampean Famatinian basins (Fig. 16.6b) indicate mostly collisional-convergent basins for the Ambato Metamorphic Complex, La

Table 16.3 K–S statistical correlation analysis of detrital zircon ages of the Pampean and Post Pampean metaclastic rocks

	Post-Pampean											Pampean											
Sample	Ch 23	Ch 28	F 16	F 15	A 26	A 25	A 14	GQ 12	V 21	SL 8	SL 17	Co 11	SL 33	Co 10	Co 9	No 27	SL 34+19	SBR. 29	An 13	An 31	SL 7	AIMC 30	SHS 20
23	■	0,052	0,025	0,000	0,013	0,001	0,000	0,442	0,985	0,250		0,000	0,000	0,000	0,000	0,001	0,000	0,000	0,000	0,000	0,000	0,000	0,000
28	0,052	■	0,279	0,207	0,569	0,105	0,081	0,220	0,047			0,000	0,001	0,019	0,019	0,163	0,001	0,013	0,051	0,001	0,001	0,175	0,000
16	0,025	0,279	■	0,003	0,428	0,069	0,032	0,131	0,004			0,000	0,000	0,000	0,001	0,070	0,000	0,003	0,002	0,000	0,000	0,053	0,000
15	0,000	0,207	0,003	■	0,110	0,178	0,009	0,098	0,004			0,101	0,193	0,450	0,448	0,357	0,202	0,049	0,220	0,121	0,193	0,496	0,097
26	0,013	0,569	0,428	0,110	■	0,476	0,291	0,019	0,000			0,000	0,001	0,009	0,053	0,948	0,001	0,079	0,098	0,000	0,001	0,347	0,000
25	0,001	0,105	0,069	0,178	0,476	■	0,032	0,088	0,002			0,002	0,019	0,064	0,141	0,805	0,016	0,417	0,215	0,022	0,019	0,879	0,001
14	0,000	0,081	0,032	0,009	0,291	0,031	■	0,019	0,000			0,013	0,015	0,058	0,153	0,186	0,026	0,002	0,060	0,005	0,015	0,176	0,005
12	0,442	0,220	0,131	0,098	0,033	0,088	0,019	■	0,782	0,083		0,000	0,000	0,005	0,012	0,000	0,002	0,010	0,003	0,000	0,000	0,032	0,000
21	0,985	0,047	0,004	0,004	0,000	0,002	0,000	0,782	■	0,177	0,065	0,000	0,000	0,000	0,000	0,000	0,000	0,000	0,000	0,000	0,000	0,000	0,000
8	0,250							0,083	0,177	■	0,313												
17									0,065	0,313	■												
11	0,000	0,000	0,000	0,101	0,000	0,002	0,013	0,000	0,000			■	0,909	0,418	0,926	0,001	0,922	0,016	0,245	0,800	0,909	0,018	0,998
33	0,000	0,001	0,000	0,193	0,001	0,019	0,015	0,000	0,000			0,909	■	0,484	0,761	0,007	0,894	0,057	0,160	0,987	1,000	0,034	0,843
10	0,000	0,019	0,000	0,450	0,009	0,064	0,058	0,005	0,000			0,418	0,484	■	0,508	0,089	0,366	0,301	0,393	0,389	0,484	0,140	0,508
9	0,000	0,019	0,001	0,448	0,053	0,141	0,153	0,012	0,000			0,926	0,761	0,508	■	0,045	1,000	0,271	0,893	0,983	0,761	0,360	0,509
27	0,001	0,163	0,070	0,357	0,948	0,805	0,186	0,000	0,000			0,001	0,007	0,089	0,045	■	0,007	0,237	0,108	0,000	0,007	0,555	0,004
34+19	0,000	0,001	0,000	0,202	0,001	0,016	0,026	0,001	0,000			0,922	0,894	0,366	1,000	0,007	■	0,085	0,626	0,961	0,894	0,111	0,622
29	0,000	0,013	0,003	0,049	0,079	0,417	0,002	0,010	0,000			0,016	0,057	0,301	0,271	0,237	0,085	■	0,691	0,065	0,057	0,805	0,042
13	0,000	0,051	0,002	0,220	0,098	0,215	0,060	0,003	0,000			0,245	0,160	0,393	0,893	0,108	0,626	0,691	■	0,310	0,160	0,575	0,077
31	0,000	0,001	0,000	0,121	0,000	0,022	0,005	0,000	0,000			0,800	0,987	0,389	0,983	0,000	0,961	0,065	0,310	■	0,987	0,087	0,764
7	0,000	0,001	0,000	0,193	0,001	0,019	0,015	0,000	0,000			0,909	1,000	0,484	0,761	0,007	0,894	0,057	0,160	0,987	■	0,034	0,843
30	0,000	0,175	0,053	0,496	0,347	0,879	0,176	0,032	0,000			0,018	0,034	0,140	0,360	0,555	0,111	0,805	0,575	0,087	0,034	■	0,004
20	0,000	0,000	0,000	0,097	0,000	0,001	0,005	0,000	0,000			0,998	0,843	0,508	0,509	0,004	0,622	0,042	0,077	0,764	0,843	0,004	■

P values > 0.05 indicate that two detrital zircon populations are statistically indistinguishable. Blue color indicate the highest degree of correlation from 0.8 to 1.0, yellow from 0.7 to 0.5

Note that the Pampean basins exhibit the highest degree of correlation. Samples with less than 25 zircon age data were excluded. Sample codes as in Fig. 16.2

Cébila Metamorphic Complex at the type locality, the sample from the northern sector of the Olta Formation (LLA-17018) and Achavil Formation in the Sierra de Famatina.

- Data from Green Quarry, Nogolí Metamorphic Complex, Olta Formation in the central part of the Sierra de Chepes, Suri Formation and La Cébila Metamorphic Complex at Quebrada La Rioja (LRJ1-1) yielded convergent margins. Negro Peinado Formation is the only post-Pampean basin sample that yielded a collisional setting. This could be an artefact of sampling, data acquisition routine or even misrepresentation. The two samples that exhibit Río de la Plata age peaks (2.2–2.0 Ga; Paleoproterozoic), LLA-17018 and CEB-428, belong to the collisional-convergent group. Samples that yielded convergent margin patterns in the cumulative distribution function plots exhibit the highest correlation among the post-Pampean basins (Table 16.3). They show a relatively large number of Middle Cambrian age zircons that define the corresponding peak, apart from having Ordovician magmatic zircons in the detrital record. It seems as if sources were more restricted than in the rest of the Famatinian to post-Pampean basins.

16.8 Orogenic Events of the Eastern Sierras Pampeanas

16.8.1 Pampean Orogeny

The western margin of South America and west Gondwana between 22° and 38°S (present coordinates) underwent an almost continuous convergent history extending from the Neoproterozoic–Early Cambrian to the Devonian through collisional/accretionary mechanisms, recycling processes and limited juvenile additions related to the Pampean, Famatinian and Achalian orogenies (Ramos 1988; Sims et al. 1998; Steenken et al. 2011). Final Gondwana amalgamation corresponds to the Cambrian closure of the Clymene Ocean that separated Amazonia from proto-Gondwana (Trindade et al. 2006). A large orogenic belt encompassing

Table 16.4 Main statistical peaks calculated with Age Pick 2010 (Gehrels 2014) for the Pampean and Post Pampean detrital zircon data. Colors of the columns correspond to the colors of the lines in Fig. 6.5

		PAMPEAN METACLASTIC ROCKS																										POST-PAMPEAN METACLASTIC ROCKS																					
	Code in Fig. 16.2	No		Co		Co		Co		Co		SL		Co		SL		An		An		A		SBR		LaPampa		Ch		Ch		F		Suri Fm		V		GQ		SL		A		A		A		F	
		27		10		9		5		6		7		11		33		31		13		30		29		20		23		28		16		FAM7082		21		GQ		17		25		26		14		15	
	NZ	55		41		38		51		16		66		58		42		69		67		42		67		69		50		59		46		59		52		33		26		60		64		62		49	
		peak	#	peak	#	peak	#	peak	#	peak	#	peak	#	peak	#	peak	#	peak	#	peak	#	peak	#	peak	#	peak	#	peak	#	peak	#	peak	#	peak	#	peak	#	peak	#	peak	#	peak	#	peak	#	peak	#	peak	#
Paleozoic	Ordovician																															**468**	**15**														
																																483	17														
	Cambrian																																			**501**	4	**508**	3								
																										520	**13**	**513**	6	**519**	6			**526**	19	515	6	526	9			**519**	5			**516**	10
																												538	4			536	3					534	7	**531**	5	541	5	**532**	6		
Neoproterozoic	Ediacaran	558	8			551	4					553	4					547	3					555	7					553	3									551	3								
				568	3							568	5							570	5	576	3																			561	4	574	8				
		591	8									601	5			598	3							582	5			580	6	586	4							576	3			585	5						
	Cryogenian	612	8									618	4					611	9	600	5	618	5			607	3			599	5			616	4			613	3					611	9	609	6		
						638	3					642	5	642	7					641	7			624	10			630	6	629	5	636	19			629	8	628	4			635	8	637	5	640	10	641	6
				668	5	660	3	657	3	671	3	658	6									680	5							645	4							652	5										
																																						666	5					663	7				
				694	4							695	4					704	3	682	4																												
										727	3	720	3					728	3																														
	Tonian							807	3			765	6	798	4	774	3	749	6	797	3																	754	6			767	4			722	3		
				835	3			845	3							823	4	823	5													793	3	808	4							796	5	770	4	787	4		
								932	20													936	5																			937	10	902	3				
Mesoprot-	Stenian			1012	3																													1010	3			1016	4			990	13			1009	5	1028	10
		1002	9	1055	5	1053	12	1025	32	1028	4	1029	37	1042	25	1044	18	1028	34	1014	29	1032	8	1024	26	1032	16					1045	3			1090	5					1077	7						
				1091	6									1087	17											1104	6			1259	5											1154	6					1207	10
												1182	7	1220	4	1176	3							1148	5																					1770	4		
Paleoprot	Orosirian-Rhyacian	1948	3																	1923	3									2059	3															2088	4		
																																												2116	5		
																																												2214	3		

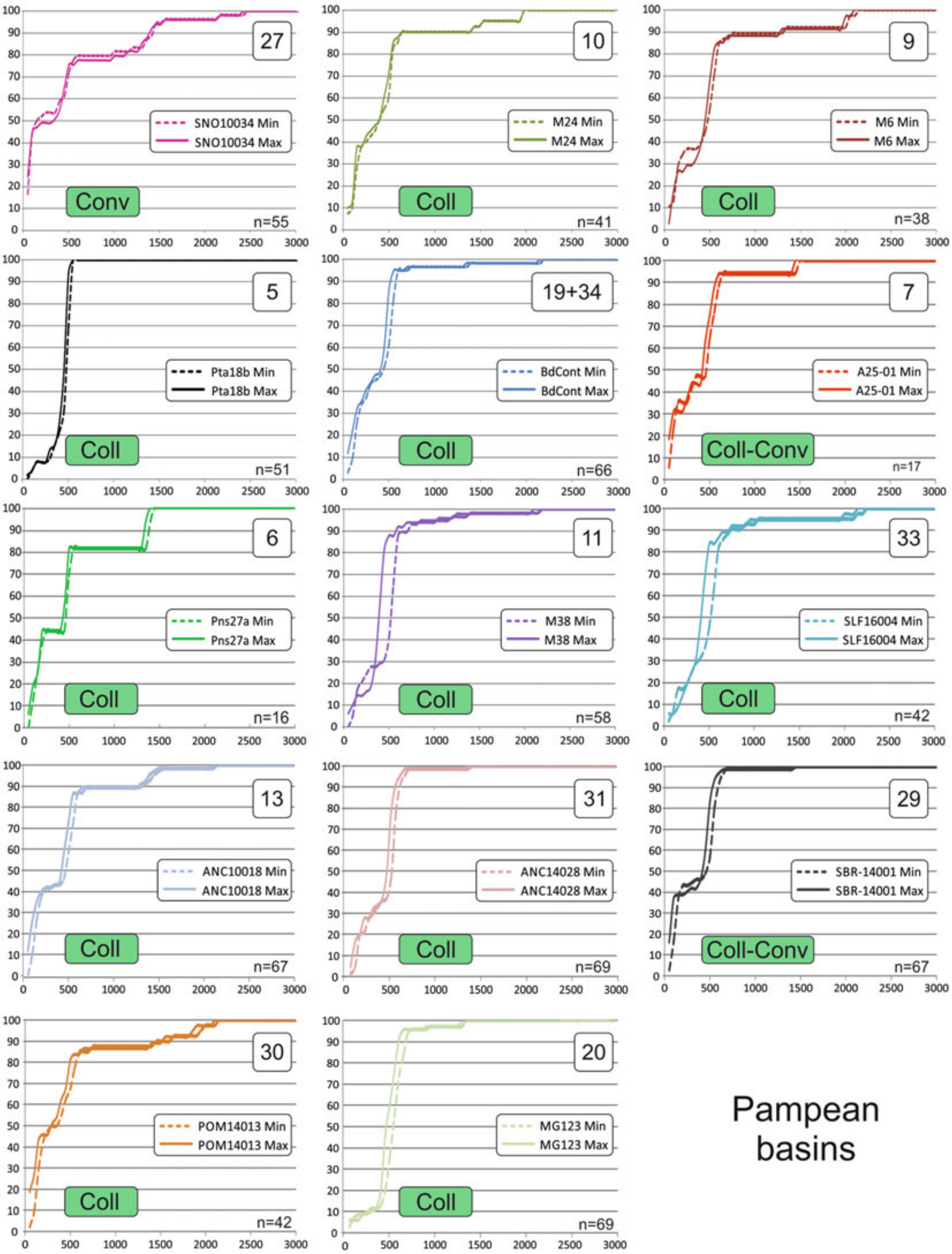


Fig. 16.6 Cumulative proportion curves (Cawood et al. 2012) for the **a** Pampean and **b** Post Pampean basins. Variation of the difference between the measured crystallization age for the youngest detrital zircon grain and the depositional age of the succession lead to the identification of the tectonic regime of the depositional system. *Coll* collisional setting, *Conv* convergent setting, *Coll-Conv* transitional between collisional and convergent. *N* number of zircon data. *Min* minimun depositional age, *Max* maximum depositional age

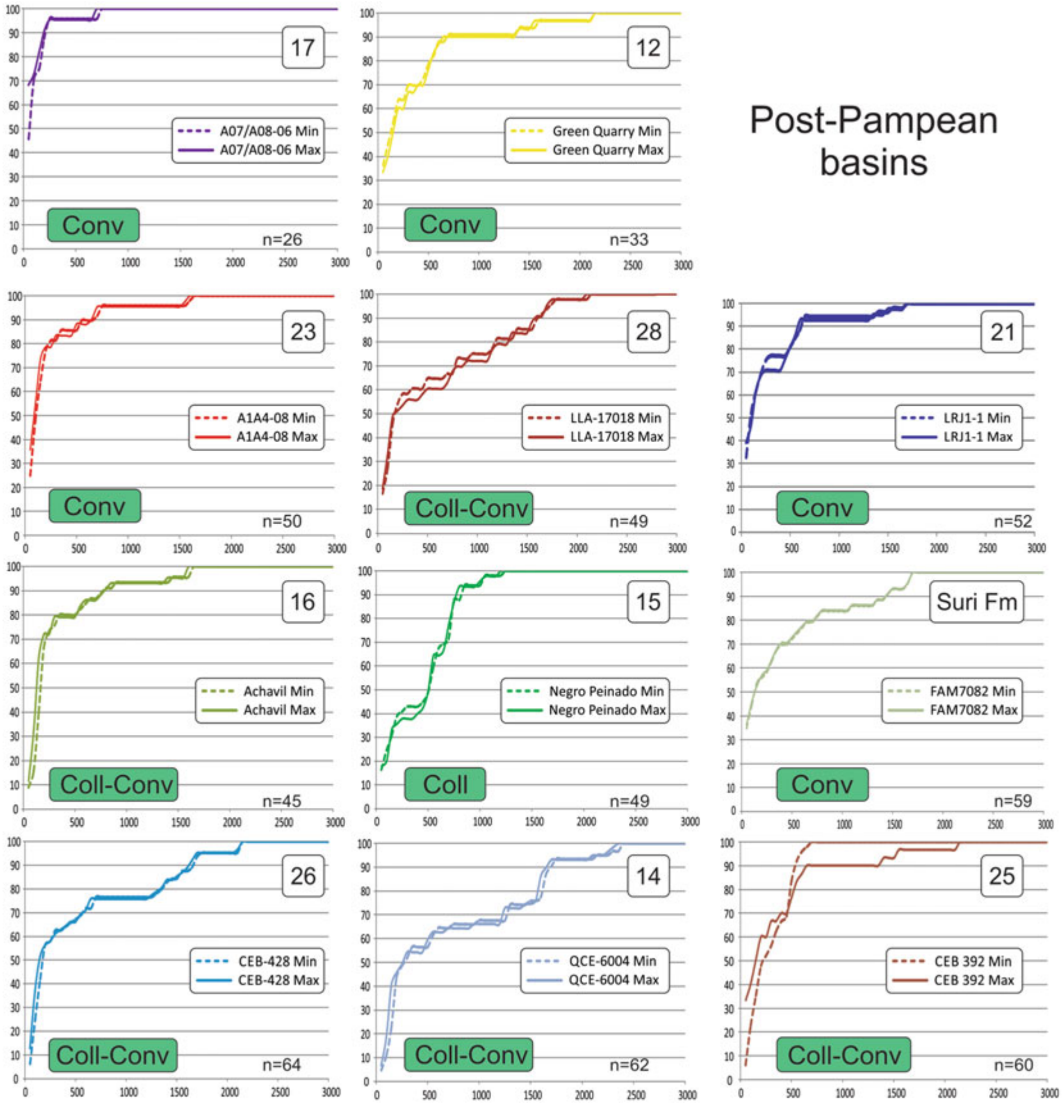


Fig. 16.6 (continued)

the Araguaia, Paraguay and Pampean Belts would record the closure of this ocean, which occurred between 530 and 520 Ma (Trindade et al. 2006).

Driving mechanisms for the Early Cambrian Pampean Orogeny involve competing scenarios of accretion of an allochthonous/para-autochthonous terrane or major block to the Río de la Plata Craton or subduction-related orogeny without any terrane addition (Piñán-Llamas and Simpson 2006; Schwartz et al. 2008) (Fig. 16.7a–d). These different hypotheses are based on deformational history, polymetamorphic evolution, detrital zircon patterns and/or magmatic signatures.

The terrane collision accretion models (Fig. 16.7a–c) involve the hypothesis that the substratum of the metamorphic basement of the Eastern Sierras Pampeanas is the Pampia Terrane (Ramos 1988), the Arequipa-Antofalla Terrane (Escayola et al. 2011) or that its basement is limited to the west by the MARA block (Rapela et al. 2016 and references therein). The Pampia Terrane was considered to be allochthonous and colliding with the Río de la Plata Craton in the Late Proterozoic (Ramos 1988; Kraemer et al. 1995), or para-autochthonous with an Early Cambrian collision with the Río de la Plata Craton (e.g., Rapela et al. 1998a, b). Ramos et al. (2010) proposed that it was originally attached to the Amazon craton since the Mesoproterozoic, jointly colliding with the Paranapanema block during the Neoproterozoic, and finally colliding with the Río de la Plata Craton during Early Cambrian. On the other hand, Chernicoff et al. (2012) considered that the Pampia Terrane derived from a Mesoproterozoic belt that detached from the Río de la Plata Craton to collide back at around 530 Ma. The collision of an island arc or a minor

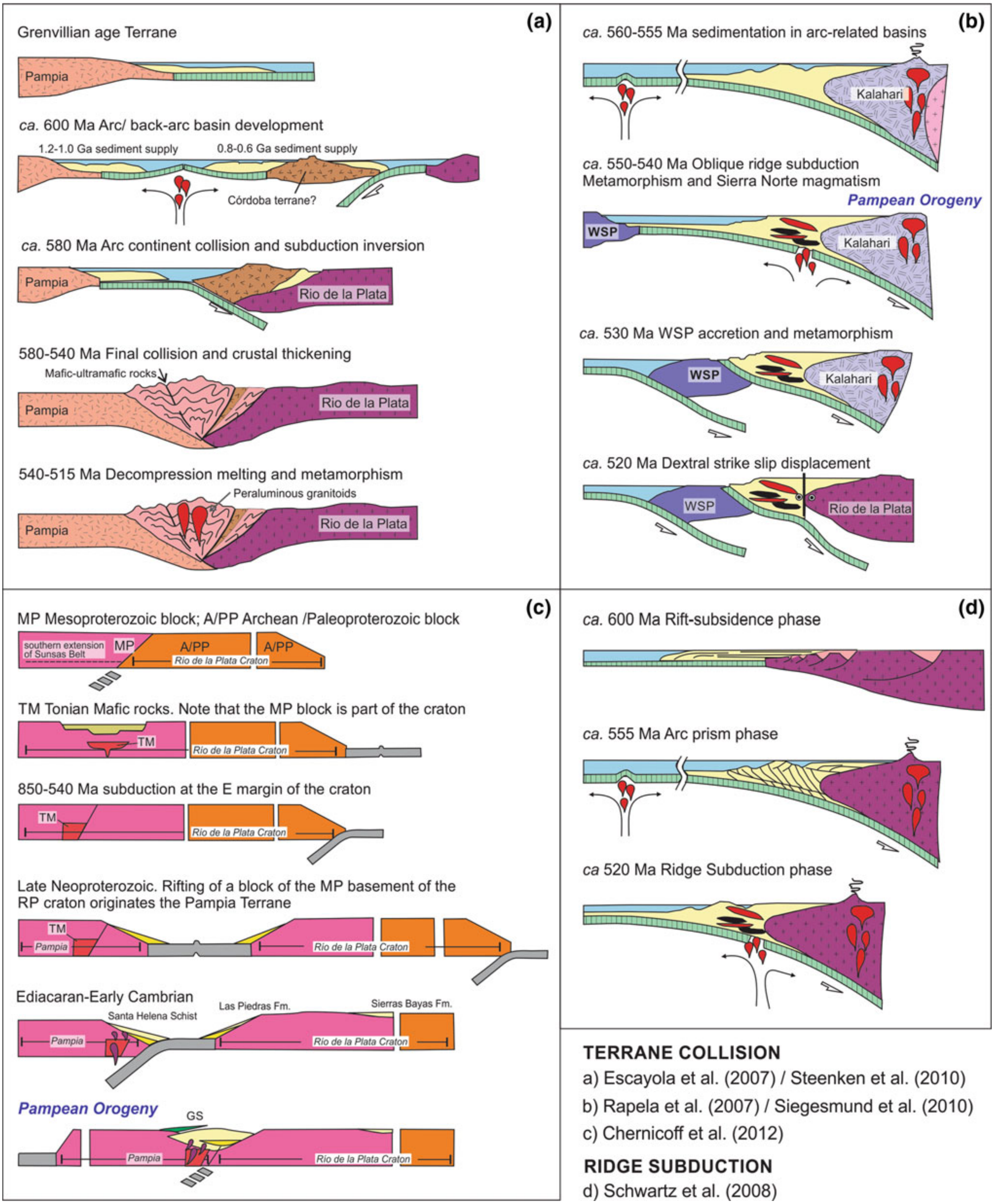


Fig. 16.7 Different models for the Pampean orogen. **a**, **b** and **c** are terrane accretion models whereas **d** correspond to the ridge subduction model. See text for comments

Mesoproterozoic block (Steenken et al. 2011) terrane against the Río de la Plata Craton would have preceded the collision of the Pampia Terrane (Ramos et al. 2015).

Rapela et al. (2007, 2016) proposed the collision of a ridge with the Kalahari Craton, and the subsequent displacement of the Pampean Belt along a transform fault was related to the collision/accretion with the MARA main terrane that includes the Western Sierras Pampeanas Terrane. This process ended with the juxtaposition with the Río de la Plata Craton.

Some variations on the terrane collisional model hypotheses correspond to Siegesmund et al. (2010) and Steenken et al. (2011), who combine the proximity to the Kalahari craton, the development of a magmatic arc at 580 Ma, with the collision of a Mesoproterozoic block. Chernicoff et al. (2012) proposed that the sedimentation and collisional episodes were developed on the Pampia Terrane facing the Río de la Plata Craton.

Ramos (1988) and later Kraemer et al. (1995) and Escayola et al. (2007) proposed that the evolution of the Eastern Sierras Pampeanas involved orthogonal subduction and terrane collision with the Río de la Plata Craton (Fig. 16.7a). An initial stage involved the Neoproterozoic accretion of the Córdoba Terrane, an island-arc terrane, against the Río de la Plata Craton following west-dipping (present coordinates) subduction (Escayola et al. 2007). Steenken et al. (2011) proposed a Mesoproterozoic terrane as responsible for this accretion that would have closed an ocean as indicated by ophiolite remnants of the Sierras de Córdoba. This early stage is supported by the M1-D1 metamorphic ages of 560 Ma and arc magmatism dated at 580 Ma at the south of the Sierras Córdoba (Sims et al. 1998; Siegesmund et al. 2010; Guereschi and Martino 2014 and references therein). The accretion of the Pampia Terrane (a Grenville age basement terrane or a Mesoproterozoic basement) is initiated by the subduction of a ridge to the already amalgamated Córdoba/Mesoproterozoic basement terrane in the Neoproterozoic-earliest Cambrian (Escayola et al. 2007; Ramos et al. 2010; Steenken et al. 2011). This collision controlled the M2 metamorphic event in the Sierras de Córdoba and produced the Pampean orogen with subsequent decompression melting and intrusion of peraluminous granitoids.

Based on a geochronological study of the Río de la Plata Craton, Rapela et al. (2007) proposed that these supracrustal sequences or protoliths of the metasedimentary rocks of the Eastern Sierras Pampeanas could initially have been deposited as large submarine fans at the southern tip of the Río de la Plata Craton and the Kalahari Craton, and fed by a magmatic arc located at the present African side. They concludes that the closing of the Clymene Ocean (Trindade et al. 2006), which might have separated the Río de la Plata Craton from the Grenvillian terranes such as Amazonia, Arequipa-Antofalla and Western Sierras Pampeanas, would have led to a right-lateral accretion between the MARA/Western Sierras Pampeanas Craton and the Pampean Belt at *c.* 540–520 Ma. After the collision, a continued right-lateral movement would have displaced the Pampean mobile belt to its present position alongside the Río de la Plata Craton through the dextral Trans-Brasiliano shear zone.

Schwartz et al. (2008) (Fig. 16.7d) proposed a long-lived early subduction between 555 and 525 Ma and a calcalkaline magmatic arc, which could partially provide the sediments to build a long accretionary prism. This process would have ended with the subduction of a seismic ridge beneath the accretionary prism developed in the Río de la Plata Craton, excluding a continental collision (Schwartz and Gromet 2004).

As noted previously by Escayola et al. (2007) and Ramos et al. (2015), the models, apart from those put forward by Rapela et al. (2007) and Drobe et al. (2011), have two statements in common: (1) the subduction of oceanic lithosphere was east dipping beneath the Río de la Plata Craton for the main metamorphic event; and (2) the supracrustal sequences of the Eastern Pampean Ranges represent passive/active margin deposits along the continental platform of that craton or facing the craton on the Pampia Terrane.

Detrital zircon ages for the Pampean basins show important Brasiliano and Grenvillian age peaks: Mesoproterozoic ages are typical of the (1250–950 Ma) Sunsás–Aguapeí province, located in southern Brazil (Teixeira et al. 2010). The lack of additional isotopic data, as for example Hf data, hampers our ability to determine whether the Grenvillian ages are derived from the Namaqualand Belt or the Sunsás-Aguapeí. The lack of a Río de la Plata signature in the Pampean basins could be derived from the existence of orogenic barriers, for instance, related to the Córdoba or island arc *c.* 580–560 Ma accretion that controlled the exhumation of Brasiliano rocks. After the final collision of the Pampia/Arequipa-Antofalla terrane at *c.* 530 Ma, cannibalization of the older granites together with recycling of the metasediments fed the post-Pampean foreland basins. Figure 16.7 illustrates the different models for the Pampean orogen.

16.8.2 Famatinian Orogeny

The Famatinian orogen consists of a continuous upper-plate continental arc, associated with some contribution from juvenile mantle material and ensialic basins (Dahlquist et al. 2008; Collo et al. 2009), which would stretch between southern Peru and northern Patagonia (Casquet et al. 2006; Chew et al. 2007; Martínez Dopico et al. 2011). The

Famatina Belt between 28° and 38°S has been interpreted as part of the accretionary margin built along Western Gondwana during the early Paleozoic after the Pampean Orogeny. The Famatinian arc started at *c*. 495 Ma when subduction was established along the outboard boundary of the Pampean Orogen. The final stages at *c*. 465 Ma (Ramos 1988; Thomas and Astini 2003) related to the collision of the Metamorphic Complex are akin to the Pampean basins allochthonous Laurentia-derived Precordillera/Cuyania Terrane during the Ordovician Ocloyic Orogeny (see Dahlquist et al. 2008 and references therein). An alternative para-autochthonous model proposed that the Cuyania Terrane (sedimentary sequence of the Precordillera of Argentina plus Grenville basement of the Western Sierras Pampeanas) migrated along a transform fault, from a position on the southern margin of West Gondwana (present coordinates) in the mid-Ordovician to its modern position outboard of the Famatina magmatic belt in Devonian time (Finney 2007 and references therein).

Dahlquist et al. (2008) proposed a model for the development of a relatively short-lived (481–463 Ma) Famatina Complex (Famatina magmatic arc and related ensialic basins) along the border of the Pampean basement, related to an ongoing subduction along the margin of the Western Sierras Pampeanas Terrane. Magmatism and closure of the ensialic basin was related to a compressive event. Steenken et al. (2006, 2008, 2011) proposed that the interval between the end of the Pampean orogen and the initiation of the Famatinian subduction encompasses the uplift and erosion of the Pampean-related rocks that fed Middle Cambrian foreland basins, which developed along the margin of the Pampean terrane and the initiation of a new subduction along the western margin of the Pampean Terrane. The first deformation and metamorphism in these basins were related to the onset of the Famatina subduction at around 500 Ma. Subsequently an extensional back-arc regime was established. The initiation of granulite facies metamorphism due to the transient anomaly associated with the 506–480 Ma mafic and ultramafic intrusions is contemporaneous or closely followed by the emplacement of crustal-derived granitoids, such as the northern stock of the 491 Ma Paso del Rey granite, parallel to the S1 foliation of the country rock and the 480–460 Ma OTS tonalite-granodiorite in the low-grade phyllites. The inferred post-D1 emplacement of those granitoids indicates that D1 and M1 pre-date the generally assumed mid-Ordovician accretion of the Precordillera/Cuyania Terrane. If Steenken et al. (2006) are right, the mafic to ultramafic rocks of the PMC emplaced in a back-arc basin at the lower levels of a continental crust had undergone D1 and amphibolite facies M1. Higher-grade metamorphism M2 and crustal melting closely followed this event. This magmatic arc would have developed in a crust that was already at amphibolite facies conditions owing to D1 that would be related to the Iruyic deformation. Collo et al. (2009) suggested a foreland tectonic context for the quartz-rich Mesón Group (NW Argentina), which unconformably covers the folded Puncoviscana Formation, and proposed that the initiation of the Famatinian orogenic cycle could correspond to the Iruyic unconformity in NW Argentina. Extensional basins post-dating the initiation of the Famatinan arc were developed along the entire margin of the Pampean Terrane (Collo et al. 2009). Ordovician volcanic rocks and volcaniclastic successions of the northern and central sector of the Famatina Complex (Dahlquist et al. 2008 and references therein) form an essentially bimodal association of basalt and subalkaline rhyolite interpreted as ensialic back-arc or inter-arc basins (e.g., Astini 2003 and references therein).

16.8.3 Achalian Orogeny

Achalian orogeny or the Achalian cycle (Sims et al. 1997) resulted from the collision of the allochthonous Chilenia terrane with Gondwana (Ramos et al. 1986; Ramos 1988; Sims et al. 1997, 1998; Quenardelle and Ramos 1999). This event roughly corresponds to the late to post-Famatinian events of the Famatinian cycle of Dalla Salda et al. (1998). The Achalian cycle is a period of heterogeneous deformation along crustal-scale fault lines that may have resulted from the resumption of the convergence on the western margin of Gondwana (Siegesmund et al. 2004). Achalian granitoids intruded the older basement and are especially widespread in the Sierra de San Luis and the Sierras de Córdoba (Sims et al. 1997; Siegesmund et al. 2004; López de Luchi et al. 2007; Rapela et al. 2007). The voluminous Achalian granite magmatism has raised much speculation regarding the underlying geodynamic causes (López de Luchi et al. 2007). It was first discussed as post-tectonic in relation to the Famatinian cycle (Llambías et al. 1998) or as resulting from slab break-off during the late stages of the subduction that ended with the collision of Chilenia (López de Luchi 1996; López de Luchi et al. 2007).

16.9 Concluding Remarks and Critical Topics for a Renewed Proposal for the Early Paleozoic Tectonic Evolution of the Eastern Sierras Pampeanas

A geodynamic model for the early Paleozoic orogenic evolution of the proto-Andean margin of Gondwana must combine the reliable geochronological evidence with petrological observations, structural features and physical parameters of the involved lithospheric segments (Fig. 16.8).

In evaluating alternatives for a tectonic model, critical data are as follows:

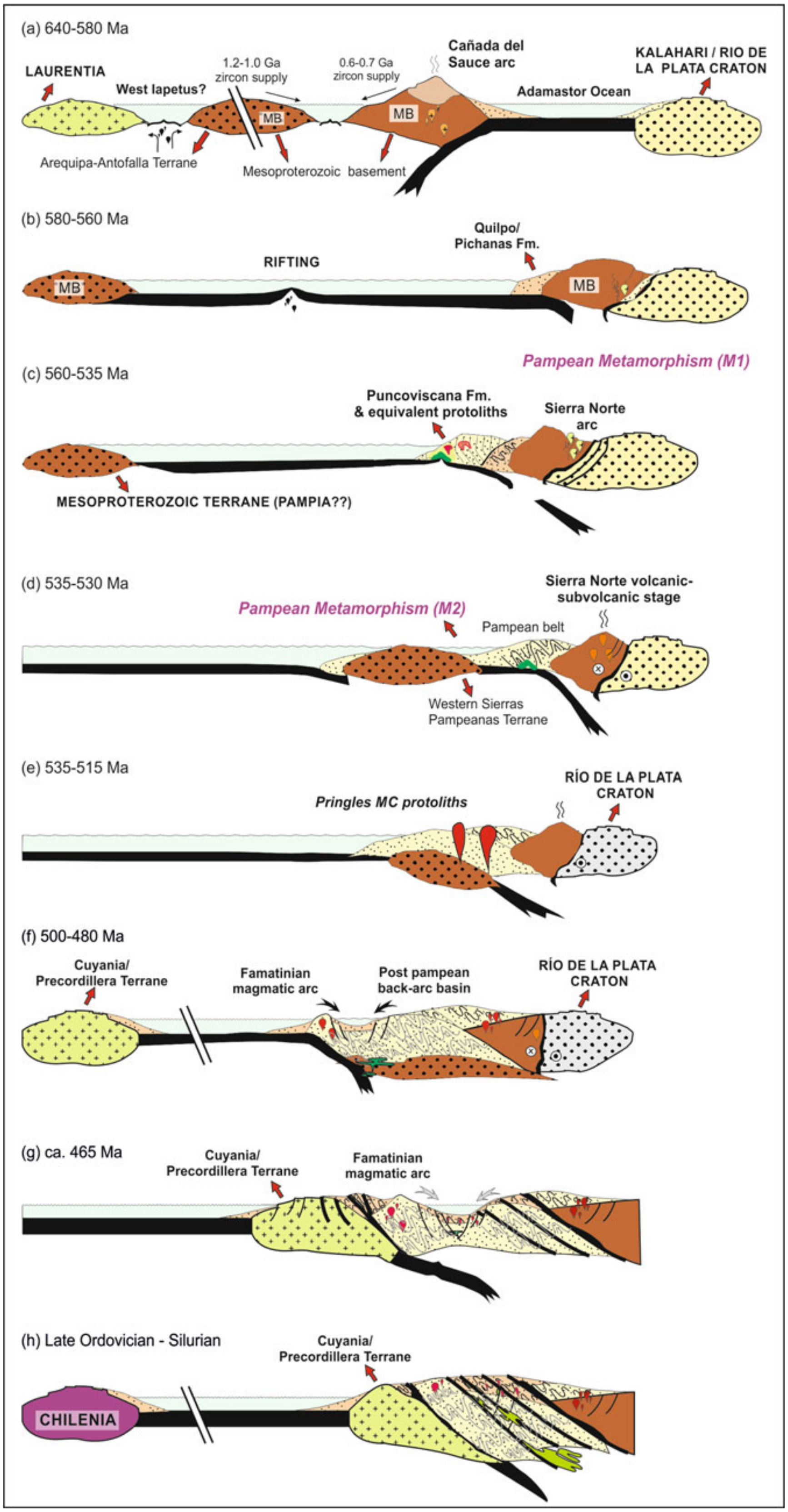

(a) 640-580 Ma
LAURENTIA
West Iapetus?
1.2-1.0 Ga zircon supply
0.6-0.7 Ga zircon supply
Cañada del Sauce arc
KALAHARI / RIO DE LA PLATA CRATON
Adamastor Ocean
MB
MB
Arequipa-Antofalla Terrane
Mesoproterozoic basement
(b) 580-560 Ma
RIFTING
Quilpo/ Pichanas Fm.
MB
MB
Pampean Metamorphism (M1)
(c) 560-535 Ma
Puncoviscana Fm. & equivalent protoliths
Sierra Norte arc
MESOPROTEROZOIC TERRANE (PAMPIA??)
(d) 535-530 Ma
Pampean Metamorphism (M2)
Sierra Norte volcanic-subvolcanic stage
Pampean belt
Western Sierras Pampeanas Terrane
(e) 535-515 Ma
Pringles MC protoliths
RÍO DE LA PLATA CRATON
(f) 500-480 Ma
Cuyania/ Precordillera Terrane
Famatinian magmatic arc
Post pampean back-arc basin
RÍO DE LA PLATA CRATON
(g) ca. 465 Ma
Cuyania/ Precordillera Terrane
Famatinian magmatic arc
(h) Late Ordovician - Silurian
Cuyania/ Precordillera Terrane
CHILENIA

Fig. 16.8 Schematic evolution of the Eastern Sierras Pampeanas between 30 and 32° SL from the Neoproterozoic to the Silurian. **a** Back-arc extension according to Escayola et al. (2007) and Steenken et al. (2011). Note that arc magmatism is developed in an inferred Mesoproterozoic (Arequipa derived?) block. Rapela et al. (2007) proposed that the location of this margin was close to the Kalahari craton. **b** Initial development of accretionary prism along an active margin that started after the closure of the Adamastor or Clymene ocean (Frimmel and Fölling 2004). Ages to constrain a *ca* 560 Ma metamorphic overprint are taken from Sims et al. (1998) and Siegesmund et al. (2010). **c** Late Neoproterozoic to Early Cambrian development of the Sierra Norte arc magmatism post dating M1 and concomitant deposition of accretionary prism and foreland basin sediments. **d** Metamorphism and deformation of the Puncoviscana Formation equivalents and cessation of arc magmatism is ascribed to the accretion of a slice of Mesoproterozoic basement that may correspond to the Pampia Terrane. **e** Post collisional collapse as indicated by Ramos et al. (2015). **f** Initiation of the subduction that migrated to the western border of the Pampean belt. Incipient development of extensional basins. Age constraints for the initiation of sedimentation are interpreted from the detrital data of the Post Pampean basins. **g** Accretion of the Precordillera Terrane. Age constraints for the Pringles Metamorphic Complex according to Steenken et al. (2006). Synorogenic deposition of Negro Peinado and Achavil formations according to Collo et al. (2009). **h** Post-Famatinian margin of Gondwana. **i** Convergence of Chilenia based on Ramos (1988)

- High-grade metamorphism either Pampean or Famatinian was ascribed to the thermal input by mafic intrusions. Most of the physical processes capable of raising temperatures in the upper mantle involve upwelling of hot asthenospheric material following attenuation or removal of the mantle lithosphere via convective thinning, delamination detachment via ‘slab break-off’, or extension either preceding or post-dating shortening (Bodorkos et al. 1999 and references therein). Attenuation of the mantle lithosphere is often accompanied by decompression melting and the emplacement of mantle-derived mafic magmas at the base of the crust as an underplate (Bodorkos et al. 1999 and references therein).
- In the Pampean orogen, high-grade granulite metamorphism post-dates a collisional stage and would result from the collapse of the subducting slab and delamination of the mantle lithosphere. This combination of processes resulted in the upwelling of hot asthenospheric material (the OIB signature of Tibaldi et al. 2008) beneath the crust of the Sierras de Córdoba, triggering plutonism and metamorphism in the middle crust. Uplift, which would be the isostatic response to asthenospheric upwelling and the initiation of the mantle-related temperature anomaly, would enhance melting of the middle crust and would control the *c*. 530–520 Ma crustal magmatism.
- In the Famatinian orogen the close temporal and spatial relationships between mid-crustal felsic plutonism, mafic magmatism and zones of high-grade metamorphism are controlled by active subduction. High-grade metamorphism in the Pringles Metamorphic Complex could result from a mantle-related transient thermal anomaly initiated at around 500 Ma as the primary heat source for high-T, relatively low-P metamorphism in an extensional setting. Subsequent compression during active convergence may explain the extensive crustal and mafic derived melts. At higher crustal levels, as exposed in the Sierra de Chepes, Ordovician metamorphism is driven by magma advection because it is typically localized and characterized by high lateral metamorphic field gradients in the vicinity of intrusions.
- Grenvillian detrital ages in Pampean metaclastic rocks are independent of the present location of the sampled rock. The average crustal residence time is *c*. 1.8 Ga, significantly older than the dominant detrital ages, a fact which implies recycled sources. These Grenvillian detrital ages could be ascribed to the approaching Mesoproterozoic terranes or, as proposed by Rapela et al. (2016 and references therein), either to the Mesoproterozoic rocks of the Natal–Namaqua Belt, southern Kalahari Craton or to the Brasiliano–Panafrican granites of southern Africa and SE Brazil and Uruguay (Schwartz and Gromet 2004; Rapela et al. 2007). Detrital zircons in the Rocha Group of SE Uruguay and the Oranjemund Group of the Gariep Belt of southwest Africa show similar bimodal patterns with a 1.0 Ga component (Basei et al. 2005), suggesting extensive offshore source rocks of these ages. In addition, Peri et al. (2013) showed that the tectonic limit between the Río de la Plata Craton and the basement of the Eastern Sierras Pampeanas dips to the east, which makes it difficult to explain a transcurrent regional displacement from the Kalahari to the Río de la Plata Craton.
- Concerning the lack of ‘Río de la Plata’ 2.2–2.0 Ga sources in Sierras Pampeanas, the model proposed initially by Ramos (1988) involving a ‘Córdoba terrane’ implies that its accretion could lead to exhumation of the associated metamorphic rocks, which would constitute the barrier for the detrital input of the Río de la Plata Craton, an idea put forward by Ramos et al. (2014).
- Ocean crust formation at 640 Ma at the present site of the Sierra de Comechingones is coherent with the existence of the Adamastor (or Clymene) ocean, which may have survived up to 560 Ma based on the data of the Sierra Chica mafic rocks with a respective DM signature. The closure of the Adamastor Ocean would be reflected in the older metamorphic age of 553 Ma recorded by the Tala Cruz metatexite (Siegesmund et al. 2010). Guereschi and

Martino (2008) calculated P–T constraints of 810–840 °C and 8.5–9 kbar for the metamorphism of the Tala Cruz stromatites.

- The *c*. 580 Ma metamorphic/igneous ages found in the easternmost extension of the Eastern Sierras Pampeanas determined by Siegesmund et al. (2010) for the Cañada del Sauce diatexite, and previously mentioned by Sims et al. (1998), would support the existence of a Mid-Ediacaran magmatic arc. Interestingly, the so-called Ancajan Series of Rapela et al. (2016 and references therein) could correspond to relicts of the host rocks of the 580 Ma magmatic rocks.
- The penetrative deformational phase at around 530 Ma is recognized in the Sierra Norte and in the Sierra Chica orthogneiss, and would correspond to the collision of the Pampia Terrane.
- The emplacement of OIB-type magmas at *c*. 535 Ma is closely associated in space with the generation of granulite facies rocks, where subsequent anatexis was associated with a decompressional path and massive anatectic crustal derived granitoids (younger than 530 Ma).
- Intrusive ages in the Sierra Norte between 555 and 535 Ma post-date the 560 Ma M 1 metamorphism and pre-date the M2 metamorphic peak and deformation history at *c*. 530 Ma (Rapela et al. 1998a) that represented the final closure of the Puncoviscan-like basins. Two stages during the magmatism in the Sierra Norte are further indicated by the intrusion of undeformed granite. Later exhumation is indicated by the *c*. 521 Ma rhyolites.
- The interval of Famatinian magmatism predates 460 Ma, which was considered to be the age of accretion of Cuyania. Therefore magmatism is related to subduction and no important collision-related Ordovician magmatism is present in the Eastern Sierras Pampeanas.
- The degree of interaction between the crust and the mafic magmas indicates a thicker crust and a protracted emplacement process in the Ordovician, which involves I-and S-type coeval magmatism.
- Detrital zircon spectra indicate a minor contribution of Grenvillian sources to the post-Pampean basins, which suggest either a barrier or the separation from these sources as a result of the relative displacement.
- Achalian Cycle magmatism is unrelated to Famatinian history and involves a transient heat anomaly that controlled extensive melting of an enriched mantle source, and probably a segment of a crust different from the Neoproterozoic to Ordovician continental crust of the Eastern Sierras Pampeanas.

Acknowledgements We gratefully acknowledge the support we have received from the German Science Foundation, which has funded our research projects in Argentina over the years (Grants Si 438/16, Si 438/17 and Si 438/24, Si 438/31), as well from the DAAD-ANTORCHAS programme. M. López de Luchi also thanks the DAAD for its support through grants A/03/39422 and A/ 07/10368. The authors thank Gabriel Giordanengo and F. Wilski for the final polishing of all figures. We are thankful to A. Steenken, R.D. Martino, C. Costa, M. Drobe, S. Löbens, F.A. Bense and numerous students from Göttingen for fruitful fieldwork and discussion.

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