
New SHRIMP U-Pb data from the Famatina Complex: constraining Early–Mid Ordovician Famatinian magmatism in the Sierras Pampeanas, Argentina

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ABSTRACT

New SHRIMP U-Pb zircon ages are reported for igneous and sedimentary rocks of the Famatina Complex, constraining the age of the magmatism and the ensialic basins. Together with whole-rock and isotope geochemistry for the igneous rocks from the complex, these ages indicate that the voluminous parental magmas of metaluminous composition were derived by partial melting of an older lithosphere without significant asthenospheric contribution. This magmatism was initiated in the Early Ordovician (481 Ma). During the Mid-Late Ordovician, the magmatism ceased (463 Ma), resulting in a short-lived (no more than ~20 Ma) and relatively narrow (~100–150 km) magmatic belt, in contrast to the long-lived cordilleran magmatism of the Andes. The exhumation rate of the Famatina Complex was considerably high and the erosional stripping and deposition of Ordovician sediments occurred soon after of the emplacement of the igneous source rocks during the Early to mid-Ordovician. During the upper Mid Ordovician the clastic contribution was mainly derived from plutonic rocks. Magmatism was completely extinguished in the Mid Ordovician and the sedimentary basins closed in the early Late Ordovician.

KEYWORDS | Famatina Complex. SHRIMP U-Pb data. Magmatism. Ensialic Basins.

INTRODUCTION

Located in NW Argentina, the Sierras Pampeanas are a series of mountainous ranges, comprising crystalline basement intruded by diverse Palaeozoic igneous rocks, which were elevated during Miocene to recent compressional (Andean) tectonics (e.g., Jordan and Allmendinger, 1986) (Fig. 1). The igneous rocks were generated in three main orogenic events: (a) Pampean (latest Neoproterozoic–mid Cambrian), (b) Famatinian (early–mid Ordovician), and (c) Achalian (late Devonian–early Carboniferous) (Aceñolaza et al., 1996; Sims et al., 1998; Pankhurst et al., 1998; Rapela et al., 1998a, b; Miller and Söllner, 2005; Büttner et al., 2005; Dahlquist et al., 2005a, b, 2006).

Extensive literature on the Famatinian magmatism is reported by Aceñolaza et al. (1996 and references therein), Sims et al. (1998), Rapela et al. (1998b, 1999a, 2001), Rapela (2000), Pankhurst et al. (1998, 2000), Dahlquist and Galindo (2004), Miller and Söllner (2005), Dahlquist et al. (2005a, b). Age constraints for the plutonic rocks of

the Famatina Complex are loosely defined, with only three published SHRIMP U–Pb ages so far (Pankhurst et al., 2000; Dahlquist et al., 2007). These SHRIMP U–Pb zircon ages are older than K–Ar and Rb–Sr isotopic ages less than 460 Ma previously obtained for Famatinian granites (Linares and González, 1990; Aceñolaza et al., 1996; Pankhurst et al., 1998; Saavedra et al., 1998). These younger ages can be explained as the result of an incomplete resetting of the Rb–Sr system during a regional ductile deformation event at 459–452 Ma indicated by, Ar/Ar dating of micas stables in the mylonitic fabric that overprint the Ordovician granitoids of the Sierra de Chepes (Sims et al., 1998; Fig. 1).

In this article, we present new ages using SHRIMP U–Pb zircon dating of igneous and sedimentary rocks from the Famatina Complex. We combine these with previous U–Pb zircon dating, whole-rock geochemistry and isotopic data of igneous rock outcrops in the Famatinian magmatic belt of the Sierras Pampeanas as a contribution to the understanding of Famatinian magmatism in the Sierras Pampeanas.

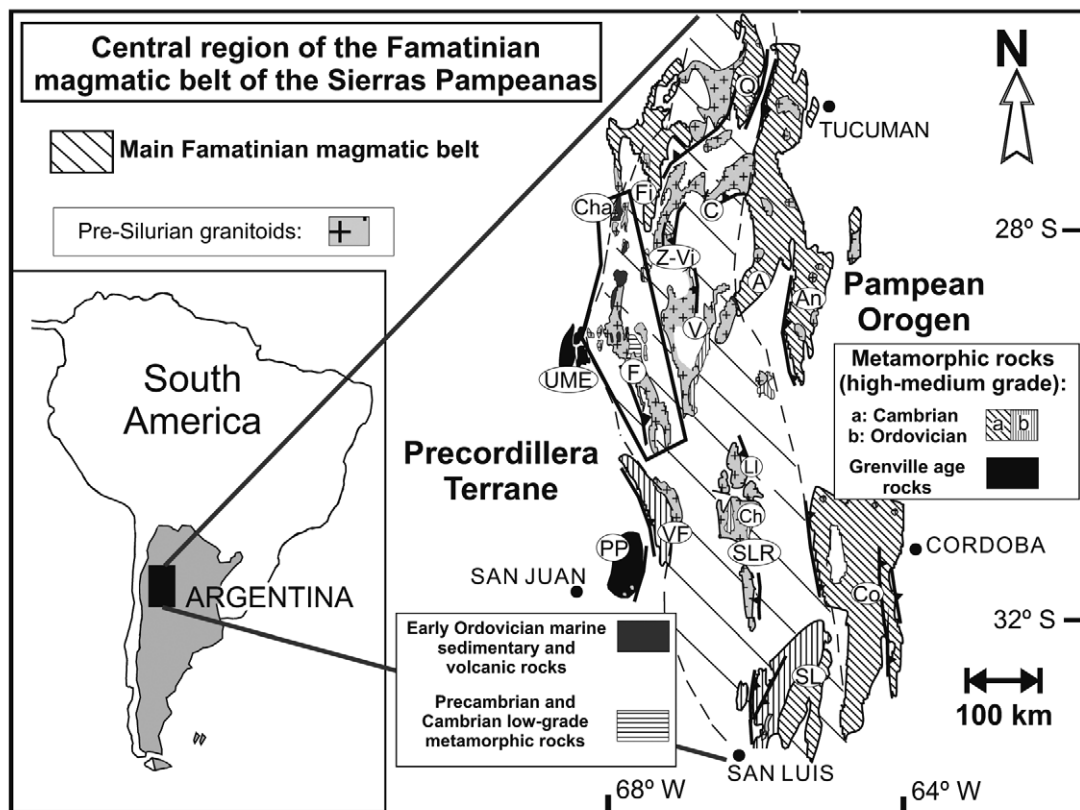


FIGURE 1 Generalized sketch map of the pre-Silurian geology of the Sierras Pampeanas, the Early Palaeozoic Famatinian belt developed on the Gondwana margin represented by the Pampean (Cambrian) and Grenville-age (~1000 Ma) basement. Letters identify the main mountain ranges; Q: Quilmes; Fi: Fiambalá; C: Capillitas; Z-Vi: Zapata y Vinquis; An: Ancasti; A: Ambato; V: Velasco; F: Famatina (Cha = Chaschuil); UME: Umango, Maz, Filo del Espinal; Co: Córdoba; SLR: Sur de La Rioja (Ch = Chepes, LI = Llanos); VF: Valle Fértil; PP: Pie de Palo; SL: San Luis. The framed area represents the studied zone in the Famatina Complex as well as other areas mentioned in the text (e.g., Chaschuil).

GEOLOGICAL SETTING

The Famatinian arc was developed on continental crust as argued by Pankhurst et al. (1998, 2000) and Dahlquist and Galindo (2004) and accepted widely (e.g., Miller and Söllner 2005). However, the evolution and geodynamic setting of the Sierras Pampeanas during the Mid-Late Ordovician is still the subject of controversy and debate: (i) one hypothesis invokes the collision of the Precordillera Terrane (the sedimentary sequence of the Precordillera of Argentina), rifting from Ouachita embayment of Laurentia in the Early Cambrian, drifting across the Iapetus ocean as a micro-continent, and docking with the proto-Andean margin of Gondwana in the Mid to Late Ordovician (Thomas and Astini, 2003 and references therein), whereas (ii) another hypothesis invokes a parautochthonous model, where 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).

The granitoids of the Famatina ranges were emplaced in Ordovician times and were part of the Famatinian magmatic arc, which occurred along the proto-Andean margin of Gondwana. This study focuses on igneous rocks and a local marine sedimentary outcrop in the west-central area of the Famatina ranges, located in the central part of the Sierras Pampeanas (Fig. 1). Geographically, the Famatina ranges lie between the Eastern Sierras Pampeanas and the exotic terrane of the Precordillera of Mendoza, San Juan, and La Rioja to the west (Figs. 1 and 2). The Famatina ranges are characterized by widespread Ordovician plutonism and local volcanic and sedimentary outcrops that are absent in the Eastern Sierras Pampeanas (for reviews see Aceñolaza et al., 1996; Saavedra et al., 1998; Miller and Söllner, 2005). These sedimentary and volcanic rocks, although of limited extent, are critically important to the understanding of the overall Early to Mid Ordovician geodynamic evolution of western Gondwana (Astini et al., 2007). Recently Miller and Söllner (2005) have suggested changing the term 'Sistema de Famatina', used by Argentinian geologists (Rapela et al., 1999a; Aceñolaza et al., 1996) for the geological entity of the Sierra de Famatina and the immediately surrounding ranges, composed mainly of granitic rocks, to 'Famatina Complex' in order to avoid confusion with the general stratigraphical meaning of 'System'. This denomination is used in this work.

MAGMATISM AND ENSIALIC BACK-ARC BASINS IN THE CENTRAL REGION OF THE SIERRAS PAMPEANAS: LITHOLOGY AND SPATIAL DISTRIBUTION

Within the Sierras Pampeanas (NW Argentina), only the Famatina Complex preserves Ordovician igneous rocks together with coetaneous marine sedimentary rocks, which are thus critically important to the understanding of the overall Early and Mid Ordovician geodynamic evolution of western Gondwana (Astini et al., 2007).

The lithology and spatial distribution of the granitoids and volcanic rocks of the Famatinian magmatic arc is summarized below, together with a brief description of the ensialic back-arc basins of the Famatina Complex.

Plutonism in the central Sierras Pampeanas

In the Famatinian magmatic belt of the Sierras Pampeanas, Pankhurst et al. (2000) have identified three distinct granite-types in the Famatinian orogen belt of the Sierras Pampeanas (Fig. 1): dominant I-type, small-scale S-type, and tonalite-trondhjemite-granodiorite (TTG, constrained to the Sierras de Córdoba, Fig. 1), which can be distinguished petrologically, geochemically and spatially, although all were essentially contemporaneous within the 484–466 Ma interval (the new ages reported in this work constrain the interval time between 484 to 463 Ma). Detailed petrological and geochemical studies of these rocks are given by Aceñolaza et al. (1996), Saavedra et al. (1998), Pankhurst et al. (1998, 2000), Dahlquist (2001a, b, 2002), Dahlquist and Galindo (2004), Miller and Söllner (2005), Dahlquist et al. (2005a, b), Dahlquist et al. (2007).

The widespread Famatinian magmatism yielded large I-type suites (most tonalites, granodiorites and minor monzogranites, gabbros) with $\epsilon_{\text{Nd}_t} = -5$ to -6 , although rare gabbros reached $\epsilon_{\text{Nd}_t} = -2.4$, and subordinate isolated Ordovician plutons of Na-rich granites located in the Pampean belt foreland (TTG suites, $\epsilon_{\text{Nd}_t} = +1.6$ to -0.2) (Rapela et al., 1998; Pankhurst et al., 1998, 2000; Rapela, 2000; Dahlquist and Galindo 2004; Miller and Söllner, 2005; Rapela et al., 2008). Most of the granitic rocks show T_{DM} ages between 1.7 and 1.5 Ga and Nd isotopic signatures, indicating derivation from a whole Palaeoproterozoic lithospheric section that included lower and upper crust sources as well as the sub-lithospheric mantle (Pankhurst et al., 1998, 2000; Dahlquist and Galindo, 2004). Thus, the Famatinian magmatic arc (with the exception of the minor TTG suites) reworked old lithospheric sources, with very little addition of juvenile material. Two dominant granitic lithology of calc-alkaline and metaluminous compositions are recognized in the central area of the Famatina Complex (Saavedra et al., 1992,

1998; Toselli et al., 1996; Dahlquist et al., 2007): (i) Cerro Toro granitic complex, with their composition ranging from gabbro to monzogranite and (ii) Ñuñorco granitic complex, with a narrow compositional range, from felsic granodiorite to monzogranite. Small-scale S-type plutons in the Los Llanos and Chepes Ranges ($\epsilon_{\text{Nd}t} = -6$ to -6.7) and the Famatina Complex ($\epsilon_{\text{Nd}t} = -5$ to -5.9 , Dahlquist et al., 2007) occur in the roof zones of I-type granitoids or in the highest-grade migmatites, and are associated with the peak of the high-temperature, low-to-intermediate pressure (M_2) metamorphic event (Dahlquist et al., 2005a; Dahlquist and Alasino, 2005; Dahlquist et al., 2007).

Early- and Mid-Ordovician ensialic back-arc basins and volcanism in the Sierra de Famatina

Ordovician volcanic rocks and volcanoclastic successions crop out in the northern and central sector of the Famatina Complex (Rapela et al., 1992; Mannheim and Miller, 1996; Clemens and Miller, 1996; Saavedra et al., 1998; Astini and Dávila, 2002) as shown in Fig. 2. The petrology and geochemistry of the volcanic rocks show that they form an essentially bimodal association of basalt and subalkaline rhyolite, with less common intermediate compositions (Mannheim and Miller, 1996; Fanning et al., 2004; Miller and Söllner, 2005). Saavedra et al. (1998) indicated that the mafic volcanic rocks in the Famatina Complex have an intraplate alkaline basalt signature, with a slight tendency to plate-margin subalkaline basalt, suggesting a dominant extensional setting. The isotopic and geochemical data for the volcanic rocks (e.g., Mannheim and Miller, 1996; Saavedra et al., 1998; Fanning et al., 2004; this work) suggest that the parental magma was mainly derived by melting of old (~1.43–1.66 Ga) meta-igneous rocks in the middle to lower crust such as the plutonic magmas. However, Fanning et al. (2004), based on an initial Nd isotope composition, suggest a rather less “evolved” source, or one with a lower contribution of older continental crustal material, than for the majority of Famatinian granites (Pankhurst et al., 1998, 2000).

Most of the Ordovician sedimentary outcrops are disconnected as a result of late Palaeozoic and Tertiary (Andean) tectonics (Astini, 2003). Thus, modern thrust faults caused the emplacement of the Ordovician sedi-

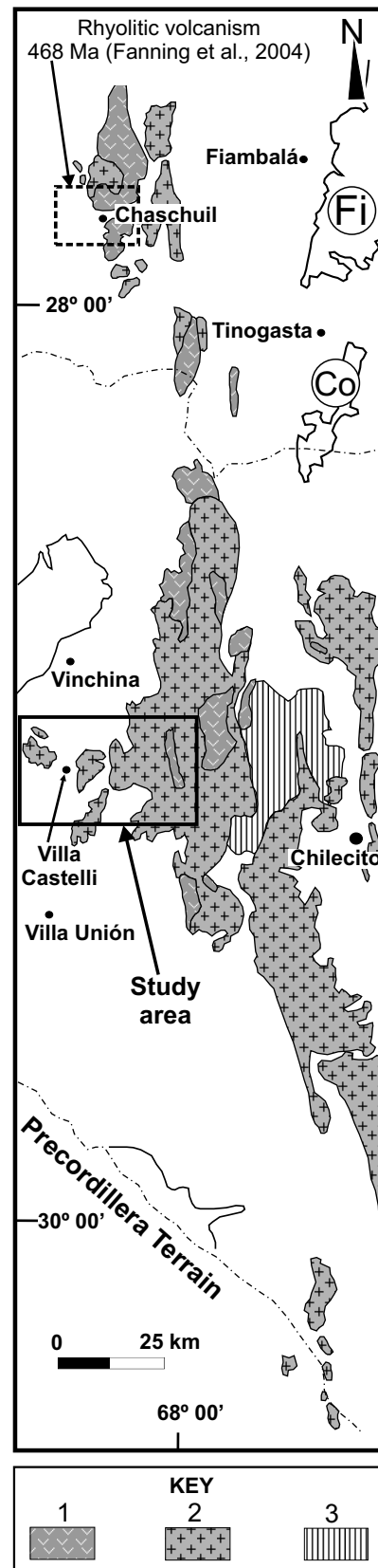


FIGURE 2 | Study area in the Famatina Complex, modified from Astini et al. (2007) and Aceñolaza et al. (1996). The Ordovician volcanic record from Fanning et al. (2004) and the Ordovician ensialic basins in the Famatina Complex from Astini and Dávila (2004), Astini et al. (2007), and this study. Abbreviation: Fi: Fiambalá; Co: Copacabana: 1: Ordovician marine sediments interbedded with volcanoclastic deposits and volcanic rocks; 2: Ordovician granitoids; 3: Early-Middle Cambrian low-grade metamorphic rocks.

mentary sequences within the outcrops of granitic rocks (e.g., Ramos et al., 2002). A composite section shows fairly continuous deposition through the Early and early Mid Ordovician (Astini, 2003; Astini et al., 2007). Some of the main formations have been grouped into higher-class units, the Famatinian Group (Lower Ordovician) and the Cerro Morado Group (Middle Ordovician) being the most commonly recognized (Astini and Dávila, 2002; Astini, 2003; Astini et al., 2007). A distinctive feature of the Famatinian and Cerro Morado groups is the inclusion of epiclastic deposits interbedded with volcanoclastic and volcanic rocks (e.g., volcanic flow, distinctive spherulitic rhyolites), together with intervals of fossiliferous siltstones with shallow marine fauna such as brachiopods, trilobites and conodonts (e.g., Albanesi and Vaccari, 1994; Mángano and Buatois, 1996; Toselli et al., 1996; Benedetto, 2003; Astini, 2003). Shell concentrations (mostly brachiopods) are observed in the Cerro Morado Group interpreted as a storm-influenced shelf, affected by contemporaneous volcanism (Astini, 2003). Thus, these characteristics, indicate strong volcano-sedimentary interactions within a shallow-marine volcanic arc setting, and are interpreted as ensialic back-arc or inter-arc basins (e.g., Astini and Dávila, 2002; Astini, 2003). The upper limit of the Cerro Morado Group is tentatively assigned to the Middle Ordovician (Llanvirn or younger), although the precise age remains unknown (Astini and Dávila, 2002; Astini, 2003).

SAMPLING AND METHODS

Petrographic investigations were conducted on 131 samples collected from the rock outcrops in the Fig. 3. Whole-rock major and trace elements were determined for 15 representative samples of the granitic rocks using ICP and ICP-MS (following the procedure 4-lithoresearch code at Activation Laboratories, Ontario, Canada). Rb-Sr and Sm-Nd determinations of three samples were carried out at the Geochronology and Isotope Geochemistry Centre of the Complutense University (Madrid, Spain). Isotopic analyses were made on an automated multicollector VG[®] SECTOR 54 mass spectrometer. Errors are quoted throughout as two standard deviations from measured or calculated values. Analytical uncertainties are estimated to be 0.01% for $^{87}\text{Sr}/^{86}\text{Sr}$, 0.006% for $^{143}\text{Nd}/^{144}\text{Nd}$, 1% for $^{87}\text{Rb}/^{86}\text{Sr}$, and 0.1% $^{147}\text{Sm}/^{144}\text{Nd}$. Replicate analyses of the NBS-987 Sr-isotope standard yielded an average $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.710247 ± 0.00003 ($n=524$). Fifty six analyses of La Jolla Nd-standard over year gave a mean $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.511846 ± 0.00003 . Zircons were separated from the three igneous samples and the metasandstone at NERC Isotope Geosciences Laboratory, U.K. (zircon was present in all samples and abundant in the tonalite). The zircon crystals are fairly uniform in

appearance: clear, colourless, 100–300 μm in length with length/breadth ratios of 2–4 (occasional long needles), euhedral with well-developed prismatic terminations, and with relatively few inclusions. Those from monzogranite FAM-7083 are heavily cracked, probably due to damage caused by their higher U-contents. All show concentric oscillatory zoning.

THE FAMATINA COMPLEX

Petrological and geochemical characteristics

The west-central area of the Famatina Complex (Fig. 2), the specific subject of this paper, is formed by widespread plutonic rocks and restricted volcanic and sedimentary outcrops. A geological map showing the localities of the samples used in our study is presented in Fig. 3. Petrological, geochemical and isotopic information of the dated samples is summarized below.

Igneous rocks

Within the part of the Famatina Complex considered here, two main and distinctive granitic complexes can be distinguished (Table 1): (a) the Cerro Toro Complex (CTC) or Cerro Toro granite (nomenclature of Toselli et al., 1996; Saavedra et al., 1996; Rapela et al., 1999a), and (b) the Ñuñorco Complex (ÑuC) or Ñuñorco granite (nomenclature of Toselli et al., 1996; Saavedra et al., 1996; Rapela et al., 1999a). The CTC was emplaced at greater depth than the ÑuC, and they are in tectonic contact (Toselli et al., 1996).

Detailed petrological and geochemical information for the granitoids of the Famatina Complex are given by Saavedra et al. (1992, 1998), Aceñolaza et al. (1996 and references therein), Dahlquist and Alasino (2005), and Alasino (2007). We summarize here the essential compositional features of these two complexes. The CTC is dominated by hornblende- and biotite-bearing tonalites and granodiorites, with less abundant monzogranites, leucogranites and hornblende-bearing gabbros and diorites; the tonalites and granodiorites are equigranular and medium-grained (1.5 cm), with the typical presence of mafic microgranular enclaves. The magmatic mineral assemblage is \pm microcline (Mc), plagioclase (Pl), biotite (Bt) and hornblende (Hbl) and the accessory minerals are epidote (Ep), apatite (Ap), zircon (Zrn), monazite (Mon), \pm allanite (Aln), and \pm titanite (Ttn) (abbreviations from Kretz, 1983) and oxides (mainly hematite after magnetite).

The ÑuC is dominated by monzogranite and felsic granodiorite, where intermediate or mafic rocks not being

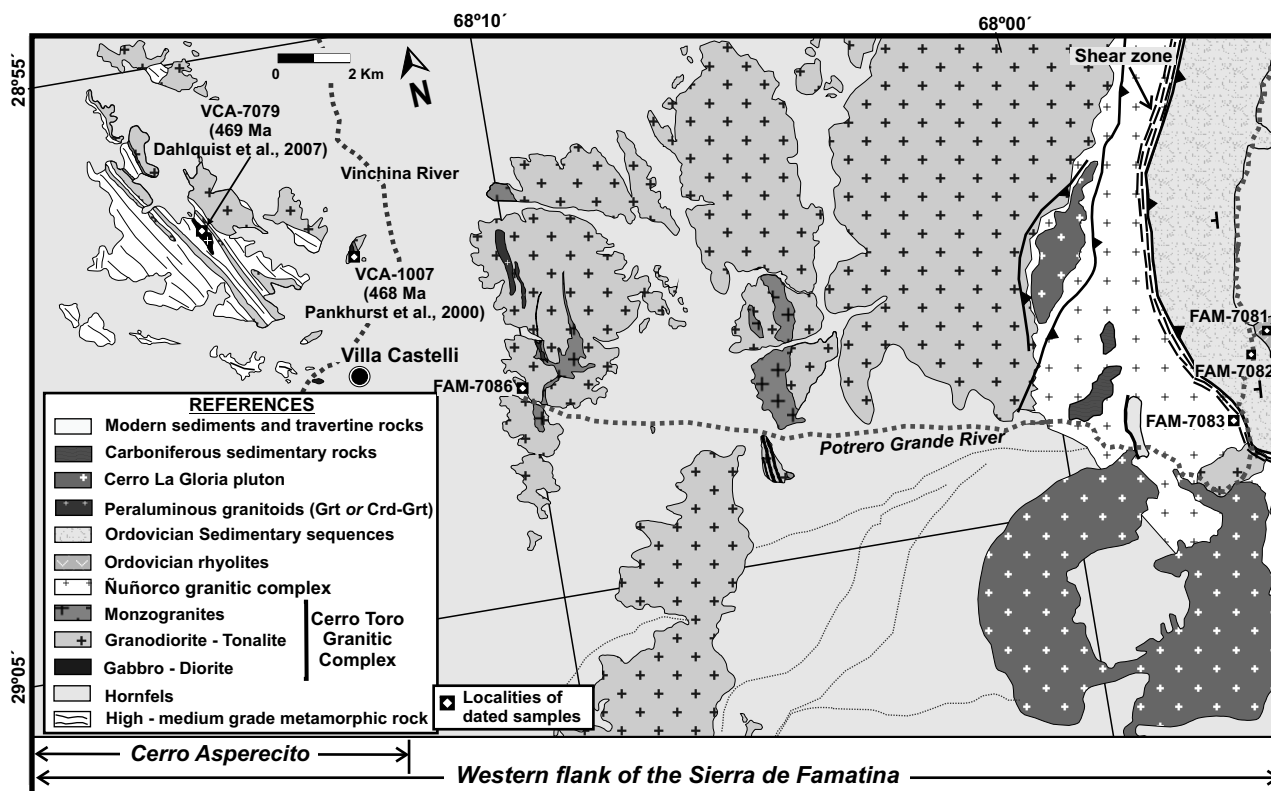


FIGURE 3 | Geological map of west-central Famatina Complex showing distribution of major magmatic units and the Ordovician sedimentary marine sequence interbedded with volcanoclastic and volcanic rocks. Previous and dated samples in this study are located in the geological map.

present. These rocks are coarse-grained (2 cm) and characterized by large alkali-feldspar and quartz crystals and the scarcity or absence of mafic microgranular enclaves. The typical mineral paragenesis is Pl, Mc, Qtz and Bt, with Zrn-oxides-Aln-Ep-Ap as accessory minerals. Major element geochemistry (Table 1) indicates that the CTC is metaluminous to slightly peraluminous (alumina saturation index using average values: ASI = 0.75 for gabbros, 1.03 for tonalites-granodiorites, and 1.14 for monzogranites, with agpaite indices of 0.12 for gabbros, 0.45 for tonalites-granodiorites, and 0.76 for monzogranites), with a wide compositional range between 45.77% and 73.81% SiO₂. Typical contents of K₂O of 0.17– 5.91% and a Peacock index of 62.3 indicate a calcic association, as reported for other Famatinian granitic suites (e.g., Pankhurst et al., 1998). According to Brown (1982) this Peacock index indicates that these granitic rocks were formed in an immature continental magmatic arc or, in other words, the continental magmatic arc was short-lived. The ÑuñC is essentially slightly peraluminous (ASI average = 1.03 and agpaite index = 0.76), with a restricted compositional range between 73.21 and 75.58% SiO₂, and with a typical content of K₂O between 3.02% and 3.90% (Table 1).

The samples analysed in this work (Table 1 and Figs. 2 and 3) are (i) FAM-7086, a tonalite from the Cerro Toro Complex, (ii) FAM-7083, a monzogranite from the Ñuñorco Complex, and (iii) FAM-7081, a K-rich (K₂O = 9.1%) rhyolite from an outcrop in the Potrero Grande canyon on the western flank of the Sierra de Famatina. The rhyolite is part of a sedimentary sequence described in the next section (metasedimentary rocks).

ASI values are 1.16 for the tonalite, 1.04 for the monzogranite and 1.02 for the rhyolite, i.e., they are metaluminous to slightly peraluminous igneous rocks. The tonalite has a SiO₂ content of 62.7%, with Pl(48%)-Qtz(27%)-Bt(20%)±Mc±Hbl and Zrn-Aln-Ep-Monazite-oxides as accessory minerals. Its REE pattern is characterized by relatively high REE_{total} (419 ppm), with [La/Yb]_N = 3.00, and negative Eu anomalies (Eu/Eu* = 0.36), suggesting plagioclase fractionation and strong control by accessory minerals. The monzogranite has SiO₂ = 75.8%, with Pl(35%)-Mc(21%)-Qtz(39%)-Bt(3%) and Zrn-oxides-Aln-Ep-Ap as accessory minerals, with a protomylonitic texture. Its REE pattern is characterized by relatively high LREE_{total} = 200, with [La/Yb]_N = 9.60, and a negative

TABLE 1 | Representative chemical analysis of the Early to mid-Ordovician granitoids and dated samples in the Famatina Complex.

Igneous rocks	I-types granitoids						Felsic Volcanic Rocks	
	Hbl-rich Gb	Tn & Gd	ASP-119	FAM-7086	Mzg	FAM-7083	Chaschuil Rhyolite	FAM-7081
Samples	CTC (n = 2)	CTC (n = 9)	CTC Mzq	CTC Tn	ÑuC (n = 3)	ÑuC Mzq	rhyolite (n = 2)	K ₂ O-rich rhyolite
<i>wt %</i>								
SiO ₂	46.33	63.39	73.81	62.73	74.66	75.58	76.47 ± 0.54	74.73
TiO ₂	0.53	0.65	0.14	17.16	0.23	0.16	0.14 ± 0.01	0.79
Al ₂ O ₃	17.02	16.02	13.96	6.48	12.78	12.31	12.29 ± 0.20	12.42
Fe ₂ O ₃ ¹	12.64	6.46	1.6	0.15	2.18	2.05	0.55 ± 0.15	0.81
FeO	nd	nd	nd	nd	nd	nd	1.48 ± 0.02	nd
MnO	0.17	0.14	0.04	2.22	0.06	0.06	0.05 ± 0.01	0.01
MgO	9.55	2.22	0.32	3.23	0.44	0.19	0.38 ± 0.18	0.13
CaO	11.47	4.62	0.92	3.61	1.48	1.04	0.38 ± 0.14	0.14
Na ₂ O	1.06	2.91	2.91	2.76	3.66	3.48	4.36 ± 0.12	1.25
K ₂ O	0.24	2.24	5.36	0.64	3.46	3.9	3.35 ± 0.47	9.11
P ₂ O ₅	0.04	0.17	0.21	0.05	0.05	0.03	0.03 ± 0.01	0.02
LOI	1.28	0.99	0.65	1.03	0.56	0.43	nd	0.35
H ₂ O+	nd	nd	nd	nd	nd	nd	0.87 ± 0.12	nd
H ₂ O-	nd	nd	nd	nd	nd	nd	0.19 ± 0.04	nd
Total	100.33	99.81	99.92	100.06	99.56	99.23	100.54	99.76
<i>ppm</i>								
Cs	0.65	4.67	8.68	10.1	2.37	1.6	0.30 ± 0.00	0.8
Rb	7.5	104	243	176	112	100	81.00 ± 3.00	203
Sr	140	183	54	122	94	72	45.00 ± 1.00	48
Ba	35	391	226	152	619	896	613.00 ± 38.00	1,960
La	2.94	33.71	26.45	73.9	38.17	46.7	35.20 ± 2.10	16
Ce	7.71	74.5	56.25	171	83.07	106	72.40 ± 2.60	41.1
Pr	0.9	7.59	5.99	15.8	8.18	9.78	8.79 ± 0.36	4.45
Nd	4.21	29.31	22.35	59.4	29.83	36.6	34.05 ± 0.95	18.5
Sm	1.11	6.53	5.52	14.2	6.19	7.65	7.36 ± 0.09	5.72
Eu	0.41	1.3	0.67	1.75	1.07	1.27	1.29 ± 0.09	0.794
Gd	1.22	6.11	4.85	15.3	5.04	5.68	7.66 ± 0.20	6.79
Tb	0.21	1.1	0.9	3.14	0.89	0.87	1.43 ± 0.05	1.29
Dy	1.37	6.9	5	21.9	5.35	5.65	8.67 ± 0.48	8.13
Ho	0.29	1.45	0.9	4.9	1.1	1.14	1.95 ± 0.11	1.68
Er	0.84	4.48	2.54	15.6	3.4	3.32	6.04 ± 0.36	5
Tm	0.13	0.72	0.36	2.73	0.55	0.52	0.93 ± 0.05	0.79
Yb	0.79	4.44	2.07	16.5	3.49	3.25	5.81 ± 0.31	4.69
Lu	0.12	0.67	0.3	2.6	0.52	0.5	0.86 ± 0.04	0.735
U	0.14	1.61	2.23	4.37	2.15	2.27	3.34 ± 0.50	3
Th	0.61	11.31	12.32	37	15.6	21.7	15.50 ± 0.80	19.1
Y	7.45	40.06	24.26	145	30.53	32.1	52.50 ± 2.90	45.3
Nb	1.75	12.25	16.87	29	10.43	8	9.10 ± 1.00	12.5
Zr	16	178	87	210	153	179	201.00 ± 4.00	127
Hf	0.55	4.71	3.17	5.3	4.43	4.8	6.10 ± 0.20	4.5
Ta	0.06	0.83	1.56	2.28	1.86	0.5	0.80 ± 0.10	1.19
Ga	13.5	13.67	17	23	14.33	13	15.50 ± 1.50	13
Ge	1.65	1.7	2.1	2	1.57	1.5	nd	1.5
ASI	0.75	1.03	1.14	1.16	1.03	1.04	1.07	1.02
AI	0.12	0.45	0.76	0.52	0.76	0.81	0.88	0.96
TAS*	Gb	Tn	Mzg	Tn	Mzg	Mzg	Rhy	Rhy

All major element oxides were analysed by ICP and trace element were analysed by ICP-MS in ACTLABS Canada. Total iron as Fe₂O₃; major element oxides in wt %, trace element in ppm. Abbreviations: Gb: Gabbro; Tn: Tonalite; Gd: Granodiorite; Mzg: Monzogranite; Rhy: Rhyolite; LOI: loss on ignition; n: samples number; CTC: Cerro Toro Complex; ÑuC: Ñuñorco Complex; nd: not determined; ASI: aluminium saturation index; AI: agpaicit index; *TAS: Total alkalis-silica classification following Middlemost (1994) for plutonic rocks and Le Maitre et al. (1989) for volcanic rocks. Data for Chaschuil rhyolites from Fanning et al. (2004).

Eu anomaly (Eu/Eu* = 0.59). The rhyolite (Table 1) has SiO₂ = 74.7%, with abundant Qtz and Kfs-Pl as phenocrysts and Zrn-oxides as accessory minerals. It presents a distinctive spherulitic texture. Chl is a typi-

cal secondary mineral. It has distinctive high K₂O (9.11%) and low Na₂O contents, that together with the low initial ⁸⁷Sr/⁸⁶Sr ratio (value reported in the section U-Pb SHRIMP geochronology and isotopic data), sug-

gest post-crystallization modification of its original alkali content and the Rb–Sr system. Cenozoic hydrothermal alteration has produced travertine rocks near the studied outcrops (Fig. 3). The REE pattern of the rhyolite is characterized by relatively low $REE_{total} = 116$, with $[La/Yb]_N = 2.30$, and a negative Eu anomaly ($Eu/Eu^* = 0.39$). The REE pattern is very similar to those reported by Fanning et al. (2004) for Famatinian rhyolites in the Chaschuil region, ~100 km to the north (Table 1 and Fig. 2).

Metasedimentary rocks

Detrital zircon was obtained from FAM-7082 greenish sandstone included in a sedimentary sequence that crops out in the Potrero Grande canyon (Figs. 2 and 3). In this area, the sampled sedimentary sequence is formed by epiclastic deposits interbedded with volcanoclastic and volcanic rocks. A shallow marine fauna is present in fossiliferous intervals (e.g. brachiopods and trilobites assigned to the Llanvirn by Lavandaio, 1973). The abundance and size diversity of the fossil remains suggest that the brachiopods may have been extinguished by a catastrophic event, such as a volcanic explosion. The studied volcano-sedimentary sequence terminates in the east with spherulitic rhyolitic rocks. Regrettably, the modern sedimentary cover does not allow us to estimate the upward continuation of the sedimentary sequence (Fig. 3).

In agreement with the lithological features and the stratigraphical criteria of Astini and Dávila (2002) and Astini (2003) this sedimentary sequence is tentatively correlated with the *Cerro Morado Group*.

The analyzed sample (FAM-7082) is formed of abundant angular to subangular clastic quartz (~0.1 mm, occasionally 0.41 mm) and clastic muscovite, plagioclase, oxides, and zircon grains in a clay matrix. The matrix is cut by parallel fractures that are filled with iron oxides (probable hematite).

U–Pb SHRIMP GEOCHRONOLOGY AND ISOTOPIC DATA

Igneous rocks

U–Pb isotopic analysis was performed using SHRIMP II at The Australian National University, Canberra (as in Williams, 1998). Results are given in Table 2 and plotted in Tera–Wasserburg diagrams together with cathodo-luminescence images in Fig. 4. The majority of analyses for each sample, uncorrected for common Pb, nevertheless concentrate around Concordia. The weighted mean ^{238}U – ^{206}Pb crystallization ages for the samples are 481 ± 4 Ma (tonalite), 463 ± 4 Ma (monzogranite), and 477 ± 4 Ma (rhyolite).

Initial $^{87}Sr/^{86}Sr$ and ϵNd_t values calculated for the times of emplacement (Table 3) are respectively, 0.7077 and -5.9 for the tonalite, and 0.7066 and -3.0 for the monzogranite, although the Sr compositions may have been partially modified during the shearing event that affected this rock. Multistage T_{DM} model ages are 1.66 Ga (tonalite), 1.43 Ga (monzogranite) and 1.44 Ga (rhyolite). These data suggest that the parental magma was probably derived by partial melting of old lithosphere, with little or no asthenospheric contribution (Fig. 5) in strongly contrast with the typical Andean-model for the generation of magmas, where an important asthenospheric component is invoked (Parada et al., 1999). The anomalously low initial $^{87}Sr/^{86}Sr$ ratio for the rhyolite (0.7000) suggests post-crystallization modification of its Rb–Sr system, and is not viable as a precise initial $^{87}Sr/^{86}Sr$ ratio. However, its ϵNd_t value of -2.9 at 477 Ma (Table 3) suggests a rather less “evolved” source, or one with a lower contribution of older lithosphere con-

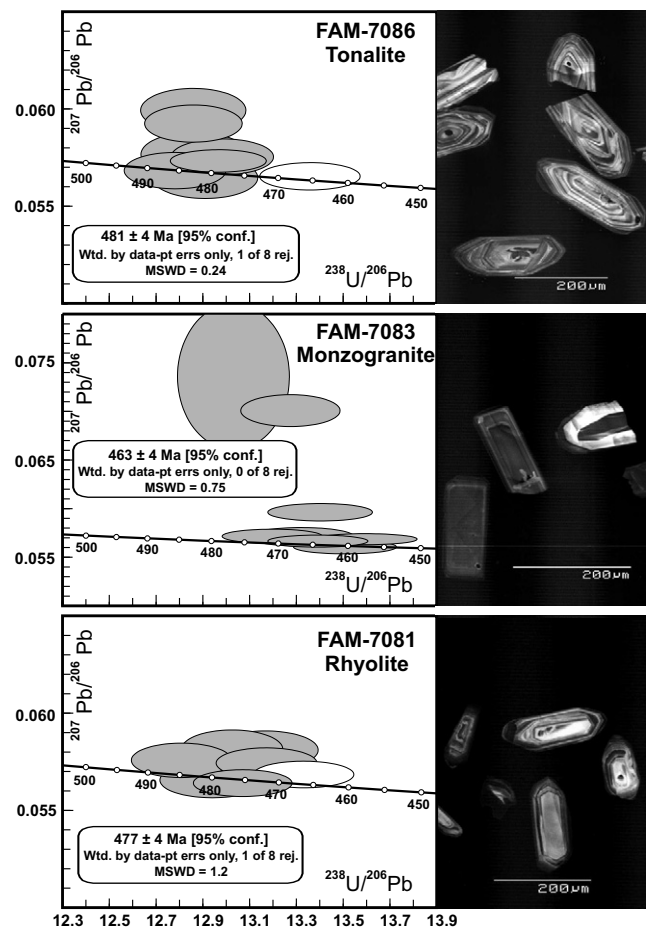


FIGURE 4 | Tera–Wasserburg diagrams for U–Pb SHRIMP zircon analyses of the studied samples (data uncorrected for common-Pb). Ellipses represent 68.7% confidence limits for each measurement, open ellipses are for points ignored as having been affected by Pb-loss. Age estimates are given for each sample (weighted mean ^{238}U – ^{206}Pb ages).

TABLE 2 | Summary of SHRIMP U-Pb zircon results.

Grain. spot	U (ppm)	Th (ppm)	Th/U	²⁰⁶ Pb* (ppm)	²⁰⁴ Pb/ ²⁰⁶ Pb	f ₂₀₆ %	²⁰⁶ Pb/ ²³⁸ U	Total			Radiogenic		Age (Ma)	
								±	²⁰⁷ Pb/ ²⁰⁶ Pb	±	²⁰⁶ Pb/ ²³⁸ U	±	²⁰⁶ Pb/ ²³⁸ U	±
FAM-7086 Tonalite														
1.1	176	183	1.04	11.8	0.000167	0.11	12.86	0.152	0.0576	0.0008	0.0777	0.0009	482.2	5.6
2.1	456	52	0.11	29.3	0.000001	0.02	13.356	0.143	0.0565	0.0005	0.0749	0.0008	465.4	4.9
3.1	416	208	0.5	27.8	0.000134	0.31	12.855	0.138	0.0592	0.0006	0.0776	0.0008	481.5	5.1
4.1	248	199	0.8	16.7	0.000042	<0.01	12.776	0.144	0.0568	0.0006	0.0783	0.0009	485.8	5.4
5.1	661	134	0.2	43.8	0.00005	0.08	12.963	0.136	0.0573	0.0004	0.0771	0.0008	478.7	4.9
6.1	194	168	0.87	13	0.000281	0.39	12.855	0.15	0.0599	0.0008	0.0775	0.0009	481.1	5.5
7.1	259	214	0.83	17.1	0.000116	0.11	12.98	0.145	0.0575	0.0006	0.077	0.0009	478.0	5.2
8.1	196	146	0.74	13	0.000213	<0.01	12.906	0.149	0.0564	0.0007	0.0775	0.0009	481.3	5.5
FAM-7083 Monzogranite														
1.1	241	211	0.88	15.5	0.000114	0.09	13.318	0.15	0.0571	0.0006	0.075	0.0009	466.3	5.2
2.1	403	266	0.66	25.7	0.000098	<0.01	13.506	0.145	0.056	0.0005	0.0741	0.0008	460.5	4.9
3.1	411	193	0.47	26.8	0.000077	0.08	13.193	0.142	0.0571	0.0005	0.0757	0.0008	470.6	5.0
4.1	280	126	0.45	18.5	0.001292	2.12	13.028	0.159	0.0735	0.005	0.0751	0.0011	467.0	6.4
5.1	288	163	0.57	18.4	0.000157	0.41	13.399	0.148	0.0596	0.0006	0.0743	0.0008	462.2	5.0
6.1	504	280	0.55	32.3	0.000113	0.04	13.39	0.142	0.0566	0.0004	0.0747	0.0008	464.1	4.8
7.1	595	126	0.21	38.5	0.000837	1.71	13.271	0.141	0.07	0.0011	0.0741	0.0008	460.6	4.9
8.1	587	203	0.35	37.1	0.000027	0.09	13.597	0.144	0.0568	0.0004	0.0735	0.0008	457.1	4.8
FAM-7081 Rhyolite														
1.1	293	171	0.59	19.3	0.00013	0.19	13.021	0.143	0.0581	0.0007	0.0767	0.0009	476.1	5.1
2.1	254	169	0.66	16.9	0.000001	<0.01	12.937	0.148	0.0565	0.0006	0.0773	0.0009	480.1	5.4
3.1	305	146	0.48	20.4	0.000016	0.09	12.803	0.142	0.0576	0.0006	0.078	0.0009	484.4	5.3
4.1	427	50	0.12	27.5	0.000068	0.06	13.329	0.143	0.0568	0.0005	0.075	0.0008	466.1	4.9
5.1	327	185	0.57	21.4	0.000093	0.11	13.17	0.143	0.0574	0.0005	0.0758	0.0008	471.3	5.0
6.1	214	164	0.76	14	0.000135	0.2	13.165	0.15	0.0581	0.0006	0.0758	0.0009	471.1	5.2
7.1	417	301	0.72	27.4	0.000029	<0.01	13.067	0.14	0.0564	0.0005	0.077	0.001	475.0	5.0
8.1	355	220	0.62	23.8	0.000028	0.01	12.816	0.139	0.0569	0.0005	0.078	0.001	484.0	5.0

Notes: 1. Uncertainties given at the 1s level. 2. Error in FC1 reference zircon calibration was 0.16% for the analytical session (not included in above errors but required when comparing data from different mounts). 3. f₂₀₆ % denotes the percentage of ²⁰⁶Pb that is common Pb. 4. Correction for common Pb made using the measured ²³⁸U/²⁰⁶Pb and ²⁰⁷Pb/²⁰⁶Pb ratios following Tera and Wasserburg (1972) as outlined in Williams (1998).

TABLE 3 | Rb-Sr and Sm-Nd data for Early to Mid Ordovician dated plutonic and volcanic rocks in the Famatina Complex.

	SiO ₂ (wt %)	Age (Ma)	Rb	Sr	⁸⁶ Rb/ ⁸⁷ Sr	(⁸⁷ Sr/ ⁸⁶ Sr) _{initial}	(⁸⁷ Sr/ ⁸⁶ Sr) _t	εSr(t)			
FAM-7086	62.73	481	176	122	41.856	0.736436	0.707749	54.21			
FAM-7083	75.58	463	100	72	40.284	0.733173	0.706601	37.59			
FAM-7081	74.73	477	203	48	123.271	0.783788	0.700008	-55.82			
	SiO ₂ (wt %)	Age (Ma)	Sm	Nd	¹⁴⁷ Sm/ ¹⁴⁴ Nd	(¹⁴³ Nd/ ¹⁴⁴ Nd) _{today}	(¹⁴³ Nd/ ¹⁴⁴ Nd) _t	εNd(t)	T _{DM} [*] (Ga)		
FAM-7086	62.73	481	14.2	59.4	0.1445	0.512172	0.511717	-5.9	1.66		
FAM-7083	75.58	463	7.65	36.6	0.1264	0.51227	0.511887	-3.0	1.43		
FAM-7081	74.73	477	5.72	18.5	0.1869	0.512457	0.511873	-2.9	1.44		
CHA-3008	75.93	470	7.96	35.3	0.1364	0.512403	511983	-1.0	1.28		

The decay constants used in the calculations are the values $\lambda^{87}\text{Rb} = 1.42 \times 10^{-11}$ and $\lambda^{147}\text{Sm} = 6.54 \times 10^{-12}$ year⁻¹ recommended by the IUGS Subcommission for Geochronology (Steiger and Jaeger, 1977). Epsilon-Sr (εSr) values were calculated relative to a uniform reservoir present day: (⁸⁶Rb/⁸⁷Sr)_{today}UR = 0.0827; (⁸⁷Sr/⁸⁶Sr)_{today}UR = 0.7045. Epsilon-Nd (εNd) values were calculated relative to a chondrite reservoir with present-day parameters: (¹⁴³Nd/¹⁴⁴Nd)_{today}CHUR. CHA-3008 = Chaschuil rhyolite from Fanning et al. (2004). References for samples in Table 1.

mental material than for the majority of metaluminous Famatinian granites, as was noted by Fanning et al. (2004) for the Chaschuil rhyolites (Table 3 and Fig. 2). Famatinian granites have typical εNd_t ranging between -4.2 and -6.3

(average of 22 samples = -5.4; using data from Pankhurst et al., 1998; Rapela, 2000; Dahlquist and Galindo, 2004). The average initial ⁸⁷Sr/⁸⁶Sr ratios and εNd₄₆₈ values of four samples from the western flank of Famatina Complex

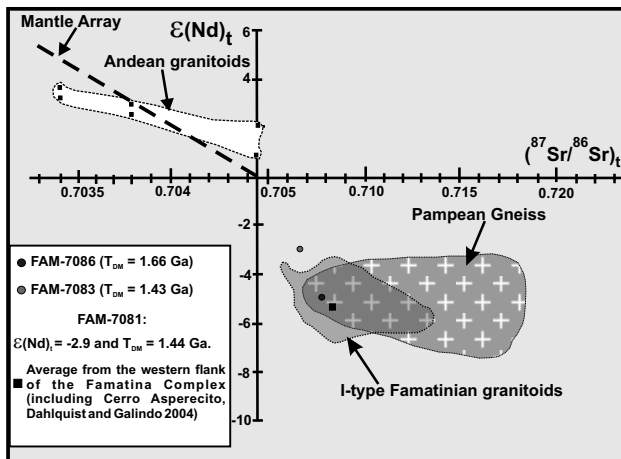


FIGURE 5 | $\epsilon(Nd)_t$ vs. initial $^{87}Sr/^{86}Sr$ diagram for Andean (black squares) and Famatinian granitoids. Data from Pankhurst et al. (1998), Pankhurst et al. (2000), Dahlquist and Galindo (2004), and dated igneous rocks in this study. FAM-7081 rhyolite has an anomalously low initial $^{87}Sr/^{86}Sr$ of 0.7000 which suggests post-crystallization modification of its Rb-Sr system.

(including Cerro Aspercito, Fig. 3) are 0.70836 and -5.3 (Dahlquist and Galindo, 2004; Fig. 5).

Metasedimentary rocks

U–Pb SHRIMP data for the sandstone FAM-7082 and its full provenance pattern have been published elsewhere in a regional context (Rapela et al., 2007), but a detail of the Phanerozoic portion shown in Fig. 6 shows that the youngest detrital zircon ages are Ordovician, with a major peak at 486 ± 5 Ma and possibly a younger one at 463 ± 7 Ma (see Fig. 6 caption). All the grains that gave Ordovician ages were small ($<100 \mu m$) and mostly well rounded by erosion, although there were a few with fractured ends. There were very few elongated grains with eroded prismatic terminations, and no signs of axial gas tubes that would have indicated a volcanic origin (e.g., from contemporaneous volcanism). The data and the nature of the grains are consistent with derivation by erosion of the plutonic rocks cropping out in the Famatina Complex, and should reflect the crystallization age of the source rocks.

DISCUSSION: AN AGE CONSTRAINT FOR FAMATINIAN MAGMATISM IN THE SIERRAS PAMPEANAS

Although initial K–Ar and Rb–Sr ages from the Famatina Complex (for review see Aceñolaza et al., 1996 and references therein as well as Saavedra et al., 1998) initially suggested that magmatism in this mountain range had a protracted development (Ordovician to Devonian), all SHRIMP U–Pb zircon ages for Famatinian granitoids (e.g., Sims et al., 1998; Stuart-Smith et al., 1999; Rapela et al., 1999b; Pankhurst et al., 2000), together with those

reported here indicate that magmatism was essentially constrained to the Early–Mid Ordovician and was coeval with development of the Famatinian ensialic basins (Astini et al., 2007; Figs. 1 and 2).

The new SHRIMP ages reported in this work for the central–west region of the Famatina Complex support two major and distinctive tectonothermal phases as originally postulated by Rapela et al. (2001) for the Famatinian magmatic belt. The main intrusive period of Famatinian granitoids of the Sierras Pampeanas began in Early Ordovician times at ~ 484 Ma: only two ages of >490 Ma have been reported (Sims et al., 1998 and Pankhurst et al., 1998 in the

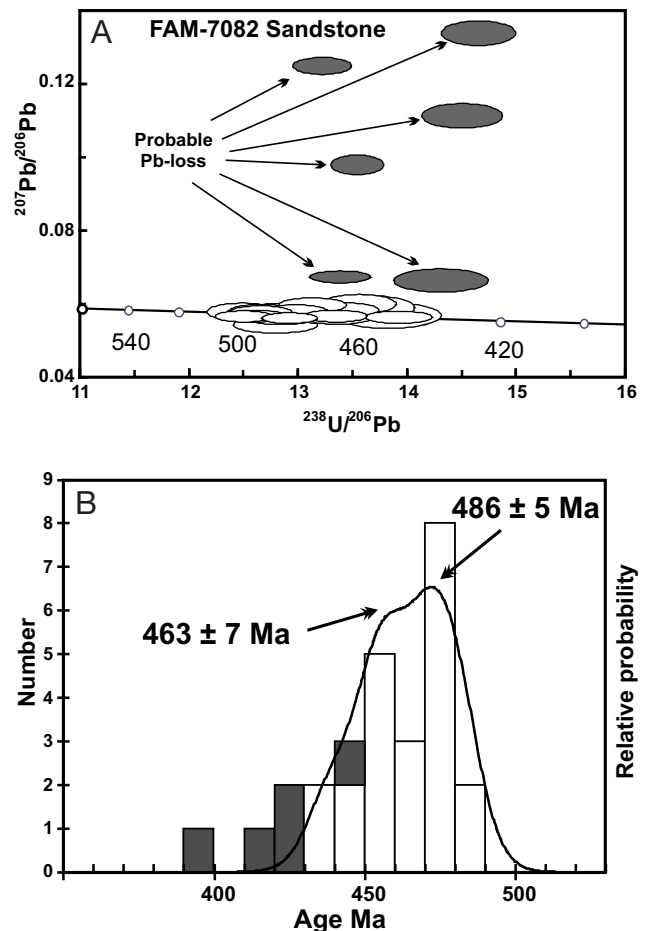


FIGURE 6 | U–Pb detrital zircon ages for FAM-7082 sandstone. A) Tera–Wasserburg plot for grains <500 Ma (see Rapela et al. 2007 for consideration of the older provenance of this sample). Error ellipses are shown at 86.3% confidence limits; individual ages are generally $\pm \sim 10$ Ma. Those shown in grey plot well above the Concordia, almost certainly indicating Pb-loss in their history, and these data are ignored. The remaining data yield ellipses that fall on the Concordia within error. A few scatter at the younger end of ~ 450 – 460 Ma and may also have suffered the effects of some Pb-loss, but the majorities are densely packed in the interval 460 – 490 Ma. B) The relative probability curve indicates a major peak at 486 ± 5 Ma with a younger shoulder at 463 ± 7 Ma (ages estimated by the Sambridge and Compston, 1994, unmixing algorithm). These fall within the known age range of Famatinian magmatism.

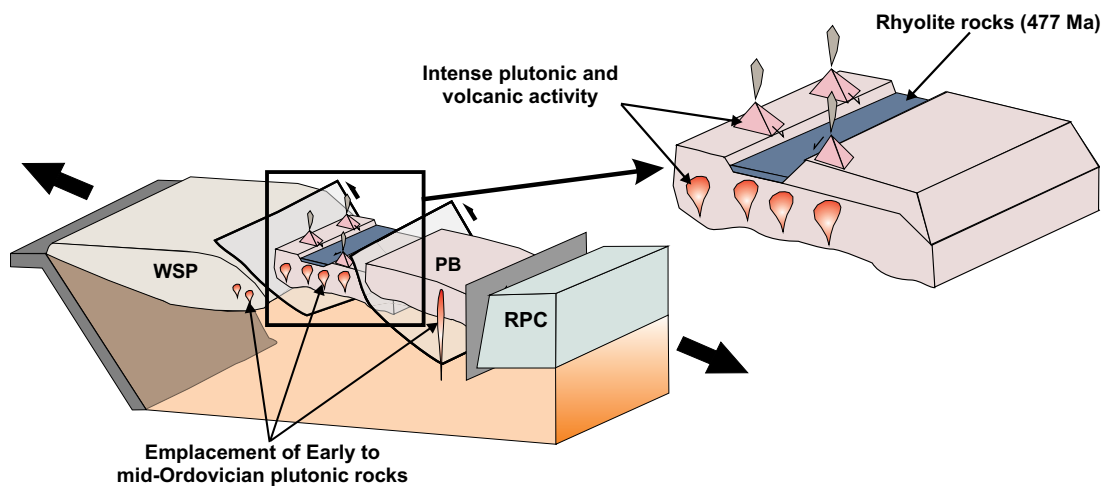
Sierras del Sur de La Rioja), but such older SHRIMP ages may have been biased by unresolved inheritance (Pankhurst et al., 2000, p. 157). Famatinian granitoids ceased at ~ 463 Ma (Pankhurst et al., 2000; Dahlquist et al., 2007; and data here presented), with largely contemporaneous and cogenetic arc volcanism (Fanning et al., 2004; and data here presented). The magmatic arc was thus short-lived (~ 20 Ma) and without a significant asthenospheric contribution (note the total absence of asthenospheric signature in Fig. 5 as well as the previous isotopic data reported by Pankhurst et al., 1998, 2000; Rapela,

2000; Dahlquist and Galindo, 2004; Dahlquist et al., 2007), in contrast to the long-lived (~200 Ma) cordilleran magmatism of the Andes with direct asthenospheric participation (e.g., Parada et al., 1999).

Stratigraphical data suggest that a dominant extensional event in this region could have resulted in arc splitting and the formation of crust slivers and isolated ensialic basins during the Early to Mid-Ordovician time (Astini and Dávila, 2004; Astini et al., 2007). In agreement with stratigraphical

Tectonic evolution of the Famatina Complex

a) ~481 to 477 Ma: Development of magmatism and ensialic basins



b) ≤463 Ma: Extinction of magmatism and subsequent closure of ensialic basins

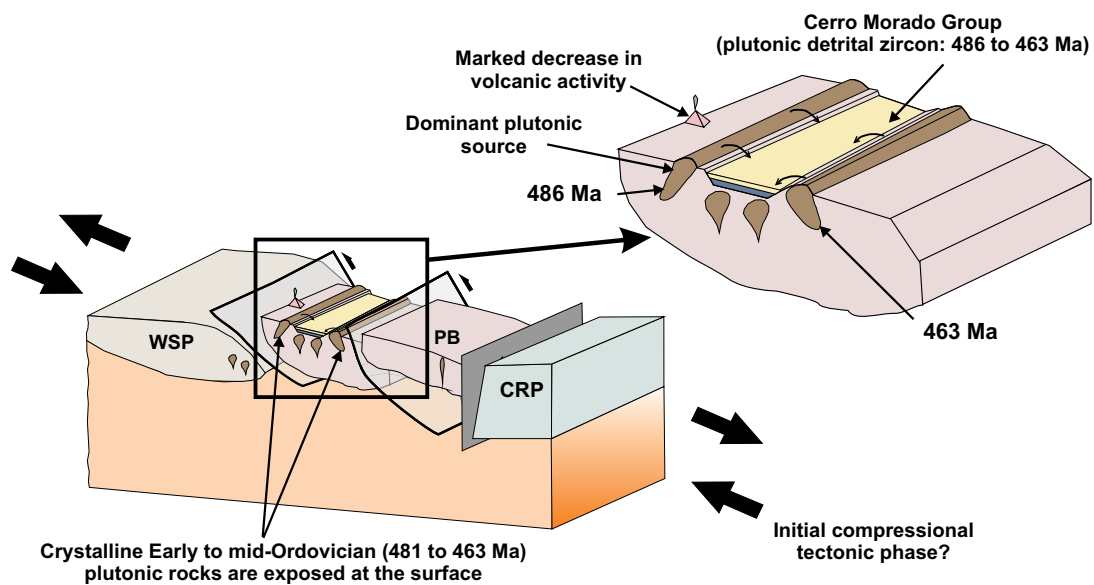


FIGURE 7 | Sketch showing the evolution of the Famatina Complex in the context of the western Gondwana margin. Abbreviations: RPC: Río de la Plata Craton; WSP: Western Sierras Pampeanas; PB: Pampean basement. The location of the basement blocks RPC, WSP, and PB during the Ordovician time is based on the data and the geotectonic proposals reported by Galindo et al. (2004) and Rapela et al. (2007).

data reported by Astini and Dávila (2002) and Astini (2003), the youngest provenance age peak of 463 Ma of detrital zircon can be tentatively associated with the *Cerro Morado Group* indicating the presence of the ensialic basin during the Mid Ordovician (Darriwilian according to Gradstein and Ogg, 2004; and the International Commission on Stratigraphy, 2006). The data and the nature of the detrital zircon grains recorded in the Ordovician basins are consistent with derivation by erosion of the plutonic rocks cropping out in the Famatina Complex, thus implying that the exhumation occurred short after granitoid intrusion and with a relatively high rates of exhumation and unroofing of the magmatic arc. A similar high exhumation rate was previously reported by Astini et al. (2003) who suggested derivation and deposition of the Ordovician sediments soon after Early to mid-Ordovician emplacement of the igneous source rocks. The presence of the volcanoclastic record in the sampled sedimentary sequence is a direct evidence of volcanic activity, although the detrital zircon record indicates a dominant plutonic source. The most direct interpretation suggests a marked decrease in volcanic activity at this time that is not seen in the subsequent sediments of the La Aguadita Formation (Astini et al. 2003; Astini and Dávila, 2004).

Remarkably, the Early Ordovician age obtained for the rhyolite FAM-7081 (477 ± 4 Ma) is coincident with that of the lower part of the *Suri Formation*, dated as Early-Mid Arenig (or Late Tremadocian) by its graptolite record (Toro and Brusca 1997). Thus, if the rhyolite is assumed to be at the base of the sedimentary sequence, this basin must have had a life of ~ 14 Ma (477–463 Ma), recording a depositional gap during the Early Ordovician. Subsequently, Famatinian magmatism closed in the upper Mid Ordovician (~ 463 Ma) and the already-emplaced granites underwent a compressional tectonic phase, resulting in the formation of well-known late Famatinian mylonitic belts (e.g., Sims et al., 1998; Pankhurst et al., 1998; Rapela et al., 2001; Astini and Dávila, 2004) and retrograde metamorphism of Late Ordovician age (e.g., Sims et al., 1998; Grissom et al., 1998; Büttner et al., 2005). Perhaps, the high exhumation rate was related with the initiation of compressional tectonics phase that culminated with the closure of the basins in the early Late Ordovician (Astini and Dávila, 2004; Astini et al., 2007).

In summary, the magmatic arc was short-lived (~ 20 Ma), with relatively high rates of exhumation, and was extinguished in the upper Mid Ordovician, coeval with the closure of the ensialic basins during the Late Ordovician (see fig. 4 of Astini and Dávila, 2004; Astini et al., 2007).

CONCLUSIONS

All the available data lead to the conclusions summarized below and to propose a sketch representing the evo-

lution of the Famatina Complex in the context of the western Gondwana margin (Fig. 7). Our data have revealed that: (i) the ages of the analysed granitoids and rhyolite from the west-central area of the Famatina Complex are similar to those obtained in other parts of the Famatinian magmatic belt of the Sierras Pampeanas; (ii) all available SHRIMP U-Pb zircon data indicate that Famatinian magmatism was brief ($\sim 484 \sim 463$ Ma), in remarkable contrast to the long-lived cordilleran magmatism of the Andes; (iii) the Early to mid-Ordovician development of ensialic marine basins was synchronous with the emplacement of conspicuous lithosphere-derived magmatism in the central region of the Famatinian orogenic belts of the Sierras Pampeanas, which strongly contrasts with the Andean-type model for the production of magmas; (iv) the magmatism was derived largely by melting of older lithosphere, although the volcanic rocks have a rather less “evolved” source, or one with a lower contribution of older lithosphere continental material, than the majority of Famatinian granitoids; and (v) the exhumation rate of the Famatina Complex was considerable and the derivation and deposition of Ordovician sediment occurred soon after of the emplacement of the igneous source rocks during the Early to mid-Ordovician. During the upper Mid Ordovician the clastic contribution was mainly derived from plutonic rocks, whereas the clastic contribution from the volcanic activity was of minor relevance for the ensialic basins. During the Mid Ordovician the magmatism was completely extinguished and the sedimentary basins closed in the early Late Ordovician.

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