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U-Pb zircon constraints on the age and provenance of the Rocas Verdes basin fill, Tierra del Fuego, Argentina

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[1] The Late Jurassic to Early Cretaceous Rocas Verdes basin constitutes one of the most poorly understood components of the southernmost Andes. As a result, accurate reconstructions and interpretations of deformation associated with the Andean orogeny and the kinematics of Scotia arc development also remain poorly constrained. In this data brief, we report U-Pb zircon ages from sandstones of the Rocas Verdes basin fill and from a crosscutting pluton in the southernmost Andes of Argentine Tierra del Fuego. Detrital samples contain predominant Early to early Middle Cretaceous (circa 130–105 Ma) U-Pb zircon age populations, with very small or single-grain middle Mesozoic and Proterozoic subpopulations. A very small subpopulation of Late Cretaceous ages in one sample raises the unlikely possibility that parts of the Rocas Verdes basin are younger than perceived. A sample from a crosscutting syenitic pegmatite yields a crystallization age of 74.7 +2.2/−2.0 Ma. The data presented herein encourage further geochronologic evaluation of the Rocas Verdes basin in order to better constrain the depositional ages and provenance of its contents.

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1. Introduction

[2] The Rocas Verdes basin of southern Argentina and Chile (Figure 1) is composed of a thick succession of siliciclastic detritus that accumulated on submarine mafic and transitional crust near the Pacific margin of southern South America during the Late Jurassic and Early Middle Cretaceous [*Katz*, 1972; *Dalziel et al.*, 1974]. The basin formed in an extensional setting between the nascent Patagonian magmatic arc to its west [*Hervé et al.*, 2007] and the Patagonian craton. The Rocas Verdes basin occupied an important period in southern South America's evolution, recording the transition between Jurassic extension associated with the breakup of Gondwana and formation of the Chon Aike silicic large igneous province, and Late Cretaceous-Oligocene contraction associated with the Magallanes foreland basin [*Wilson*, 1991; *Pankhurst et al.*, 2000; *Fildani et al.*, 2003; *Fildani and Hessler*, 2005; *Calderón et al.*, 2007; *Romans et al.*, 2009]. Quantification of the deformation that occurred during the Late Cretaceous-Oligocene Andean orogeny requires a detailed understanding of the Rocas Verdes basin as it is variably interpreted to have accommodated small amounts [e.g., *Wilson*, 1991; *Ghiglione and Cristallini*, 2007] or as much as 430 km [*Kraemer*, 2003] of upper crustal shortening, the resolution of which has significant implications for the development of the Scotia arc and Drake Passage [*Kraemer*, 2003; *Ghiglione and Cristallini*, 2007; *Ghiglione et al.*, 2008; *Barbeau et al.*, 2009; *Gombosi et al.*, 2009].

[3] Despite its importance for regional and global geologic problems, an understanding of the Rocas Verdes basin is hampered by poor chronostratigraphic control caused by the limited diversity of lithologies in its basin fill [e.g., *Winn*, 1978; *Olivero and Martinioni*, 2001; *Olivero and Malumián*, 2008], coupled with its internal deformation [*Halpern and Rex*, 1972; *Dalziel et al.*, 1974; *Bruhn*, 1979]. Moreover, most recent comprehensive studies of the basin have focused on outcrops in Chilean Patagonia [*Wilson*, 1991; *Fildani and Hessler*, 2005; *Calderón et al.*, 2007], although the basin continues more than 500 km south and east along the spine of the Patagonian and Fuegian Andes along which the continent's architecture and kinematic history varies considerably. Toward improving our understanding of the Rocas Verdes basin, we present U-Pb detrital zircon geochronology data collected from its sedimentary basin fill along the northern margin of the Beagle Channel in Argentine Tierra del Fuego. We also present the U-Pb zircon crystallization age of a crosscutting pluton.

2. Geologic Background

[4] The Rocas Verdes basin is named for a long, discontinuous belt of mafic rocks that occurs inboard of the Patagonian batholith, near the crest of the southern Andes (Figure 1) [*Katz*, 1972; *Dalziel et al.*, 1974]. Composed of pillow basalts, dikes and layered gabbro [*Winn*, 1978; *Saunders et al.*, 1979], the Rocas Verdes have been interpreted as the upper parts of an ophiolite [*Dalziel*, 1981; *Allen*, 1983; *Stern and de Wit*, 2003] that floored large parts of a marine basin [*Fildani and Hessler*, 2005; *Calderón et al.*, 2007]. These mafic rocks have a composition similar to mid-ocean ridge basalt [*Alabaster and Storey*, 1990; *Stern et al.*, 1992] and were obducted onto the South American margin during Cretaceous and younger inversion [*Dalziel*, 1986; *Dalziel and Brown*, 1989; *Wilson*, 1991; *Kraemer*, 2003; *Fildani and Hessler*, 2005]. Filling the basin floored by the Rocas Verdes is a thick Upper Jurassic to Middle Cretaceous succession of predominantly fine-grained volcaniclastic and terrigenous detritus known collectively as the Yahgán, Beauvoir, Río Jackson, Río García, Vicuña, La Paciencia, and Zapata Formations [*Wilson*, 1991; *Alvarez-Marrón et al.*, 1993; *Olivero and Malumián*, 2008, and references therein]. Whereas the northern part of the Rocas Verdes basin fill (e.g., Zapata Formation) contains very little medium- and coarse-grained material [*Fildani and Hessler*, 2005], the southern basin fill contains both coarse- and fine-grained lithofacies [*Suárez and Pettigrew*, 1976; *Olivero and Malumián*, 2008]. In the southern part of the basin, these formations are locally cut by multiple Middle Jurassic-Neogene plutonic suites [*Halpern and Rex*, 1972; *Halpern*, 1973; *Hervé et al.*, 1984; *Acevedo et al.*, 2002; *Peroni et al.*, 2009; *González Guillot et al.*, 2009] whose geochemistry indicate formation by subduction magmatism [*Cerredo et al.*,

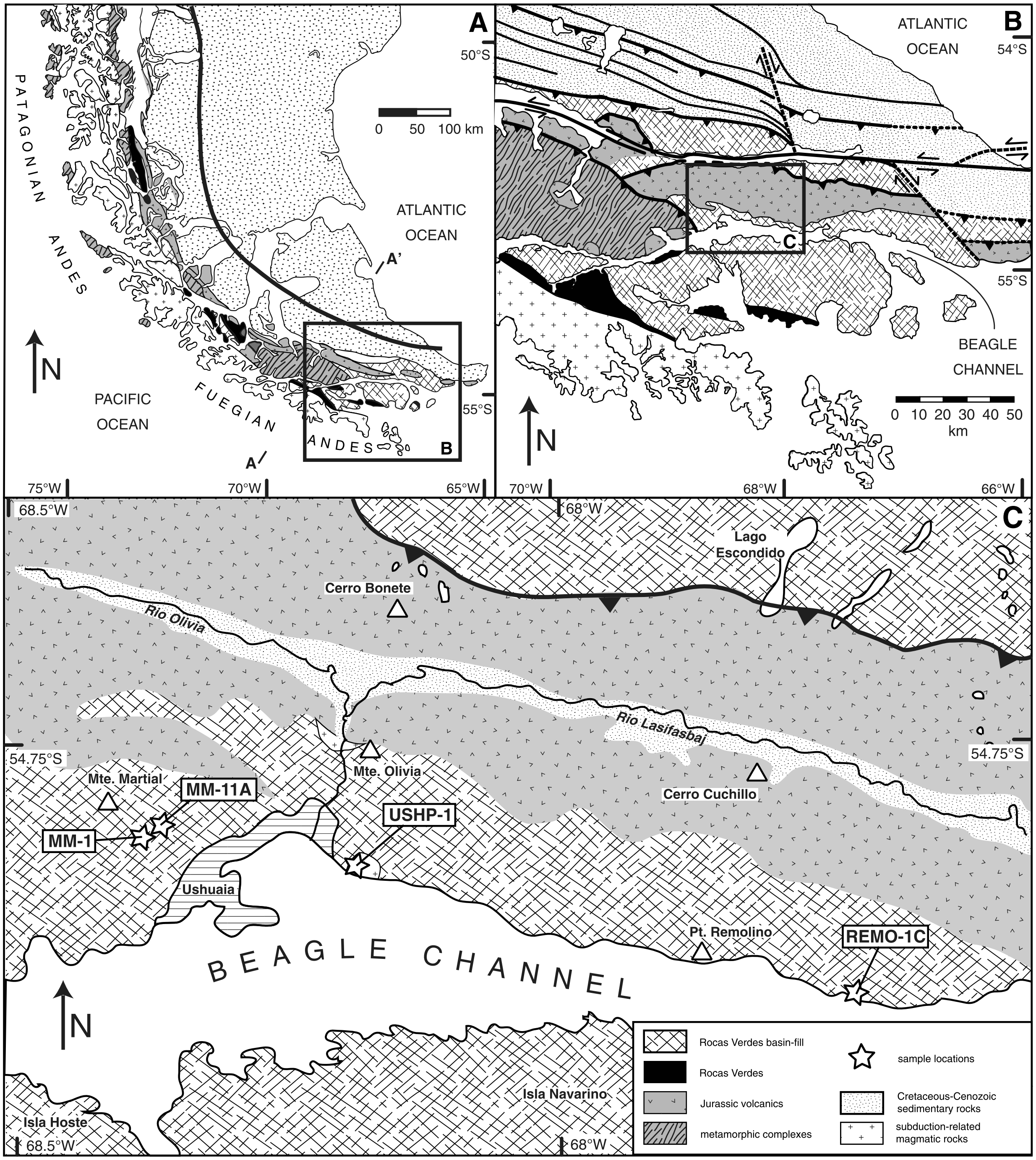

Figure 1. Geologic maps of study area depicting locations of samples reported in this study. Geology was derived from a range of sources [*Wilson*, 1991; *Fildani and Hessler*, 2005; *Olivero and Malumián*, 2008; *Barbeau et al.*, 2009] and field work conducted by the authors.

2007; *González Guillot et al.*, 2009] but occur significantly inboard of the Patagonian magmatic arc.

3. Samples

3.1. Rocas Verdes Basin Fill Detrital Samples

[5] Samples MM-1, MM-11A and REMO-1C come from the Yahgán Formation, whose protolith is composed of a succession of alternating mudstones and sandstones with subordinate chert and conglomerate [*Winn*, 1978; *Olivero and Martinioni*, 2001] that have been metamorphosed to greenschist facies [*Cunningham*, 1994], foliated, folded and/or tilted. Geometric relationships and facies analysis suggest the Yahgán Formation constitutes volcaniclastic sediments deposited in a submarine fan setting [*Kranck*, 1932; *Suárez and Pettigrew*, 1976; *Olivero and Malumián*, 2008]. The formation is well exposed along the eastern part of the northern margin of Beagle Channel on Isla Grande de Tierra del Fuego, interpreted correlatives of which occur along the southern margin of the Beagle Channel within the Chilean archipelago [*Winn*, 1978; *Suárez et al.*, 1985]. Although body fossil contents are sparse in the Yahgán Formation, late Albian (circa 105–100 Ma) inoceramids occur east of the study area [*Olivero and Martinioni*, 1996]. Localities south of the Beagle Channel contain Aptian-Albian (circa 125–100 Ma) bivalves and corals [*Dott et al.*, 1977], and Tithonian-"Neocomian" (circa 150–130 Ma) ammonites and belemnites [*Suárez et al.*, 1985]. In the broadly equivalent Zapata Formation of the Patagonian Andes, inoceramid, belemnite and ammonite paleofauna are late Tithonian to Albian (circa 150–100 Ma) in age [*Katz*, 1963; *Stewart et al.*, 1971; *Fuenzalida and Covacevich*, 1988] (also B. Aguirre-Urreta (personal communication with F. Hervé, 2002), reported by *Fildani and Hessler* [2005]). Detrital zircons from the Zapata Formation are no younger than circa 132 Ma, supporting a Hauterivian or younger depositional age [*Calderón et al.*, 2007]. Detrital zircons from the lowermost part of the overlying and conformable Punta Barrosa Formation indicate a depositional age of circa 92 Ma [*Fildani et al.*, 2003].

[6] Samples MM-1 and MM-11A were collected from thinly and tabularly bedded fine- to medium-grained quartzites from Monte Martial in westernmost Argentine Tierra del Fuego. Sample MM-1 (54°47.296′S, 068°23.867′W) was collected from a 20 cm thick bed of medium-grained quartzite within a thick succession of interbedded, ~10 cm thick, black metapelites and subordinate low-grade, fine-grained quartzites. The sampled quartzite fines upward and has a sharp basal contact with an underlying metapelite, suggesting that its protolith was deposited as a low-density turbidity current, consistent with interpretations of similar lithofacies in other parts of the basin [*Winn*, 1978; *Wilson*, 1991]. Quartzites in the sampled interval contained small quartz veinlets oriented orthogonal to bedding that do not penetrate into the underlying or overlying metapelites. Sample MM-11A (54°47.427′S, 068°22.272′W) was collected from a 15 cm thick, light gray lithic quartzite interbedded with 5 cm thick, black metapelites. The sampled quartzite fines upward and contains evidence of structureless and relict horizontal stratification characteristic of the T_a and T_b facies, respectively, of Bouma turbidites [*Bouma*, 1962]. Steeply dipping penetrative foliation that is oblique to bedding occurs in the metapelite facies and is oriented roughly parallel to the axial planes of small-scale, slightly recumbent folds that deform the section. Sample REMO-1C (54°52.829′S, 067°44.450′W) comes from a 80 cm thick, well-sorted, fining upward, medium- to coarse-grained light green quartzite within a succession of tabularly bedded, coarse-grained granular quartzites and subordinate interbedded mudstones east of Punto Remolino along the northern shore of the Beagle Channel.

3.2. Crosscutting Granitoid Sample

[7] Sample USHP-1 comes from a syenitic pegmatite of the Ushuaia pluton on the northern shore of the Beagle Channel, approximately nine km east of Ushuaia, Argentina. The Ushuaia pluton is one of several small igneous bodies associated with subduction magmatism that occur >50 km inboard of the Patagonian batholith [*Olivero and Malumián*, 2008; *González Guillot et al.*, 2009]. Although largely mafic-ultramafic in composition, comagmatic enclaves and veins of felsic pegmatites are known locally along the southwestern margin of the pluton within a mixed syenite-hornblende igneous facies. Al-in-amphibole barometry suggest emplacement at 6–8 kbar, whereas hornblende alkali exchange thermometry indicates formation at ~950°C [*Acevedo et al.*, 2002]. Whole-rock $^{40}K/^{40}Ar$ isotope analysis from two samples of the mafic facies suggest crystallization at 113 ± 5 Ma and 100 ± 6 Ma [*Acevedo et al.*, 2002], but widespread hydrothermal alteration in the Ushuaia pluton and the possibility of excess Ar incorporation limits the reliability of these ages. Related plutons have returned similar but occasion-

ally younger results including $^{40}K/^{40}Ar$ ages of 88 ± 3 Ma [*Acevedo et al.*, 2002], 86.9 ± 1.8 Ma [*Elsztein*, 2004], 77 ± 3 Ma [*Ramos et al.*, 1986]; and a $^{87}Rb/^{87}Sr$ age of 115 ± 3 Ma [*González Guillot et al.*, 2009]. Sample USHP-1 (54°49.084′S, 068°11.242′W) was collected from an epidotized leucosome within a dominant mafic assemblage of the Ushuaia pluton. Zircon and apatite (U-Th-Sm)/He low-temperature (~180°C, ~50°C) thermochronologies have yielded cooling ages of 46.4 ± 4.2 Ma and 14.6 ± 0.8 Ma, respectively [*Gombosi et al.*, 2009].

4. Methods

[8] For each of the detrital samples MM-1, MM-11A and REMO-1C, approximately 5–10 kg of quartzite was collected from outcrops with minimal evidence of possible contaminants. For igneous sample USHP-1, ~5 kg of material was collected from a small, pegmatitic felsic leucosome with significant evidence of hydrothermal alteration. Zircon separates were acquired using standard disaggregation, density and magnetic separation techniques following *Barbeau et al.* [2009]. U-Pb zircon geochronology was conducted by laser ablation multicollector inductively coupled plasma–mass spectrometry (LA-MC-ICPMS) following the techniques described by *Gehrels et al.* [2006], using a 193 nm ArF laser with spot diameters of 20–50 μm depending on grain size. Interpreted ages are based on $^{206}Pb/^{238}U$ for <1.0 Ga grains and on $^{206}Pb/^{207}Pb$ for >1.0 Ga grains. Following LA-MC-ICPMS analysis, selected zircons were imaged with a photomultiplier cathodoluminescence (CL) detector attached to a JEOL JXA8600 Microprobe using a 15 kV accelerating voltage and ~25 nA current.

5. Results

[9] Table 1 contains all data obtained from the zircons analyzed in this study. CL scanning electron micrographs occur in Figure 2. Figures 3 and 4 depict concordia and histogram probability plots of U-Pb age distributions.

[10] All three detrital quartzite samples collected from the Rocas Verdes basin fill contain predominant Early to early Middle Cretaceous zircon U-Pb age populations. Sample MM-1 contains circa 110–130 Ma grains, constituting 85% (61) of the sample's analyzed zircons. Sample MM-11A contains circa 105–122 Ma grains, constituting 85% (23) of the sample's analyzed zircons. Sample REMO-1C contains circa 107–123 Ma grains, constituting 100% (103) of the sample's analyzed zircons. The remaining grains from sample MM-1 constitute a five-grain subpopulation between circa 75–83 Ma (77.7 ± 3.0 Ma mean age, MSWD = 1.7, probability = 0.15), and isolated grains with ages of 91.7 ± 2.4 Ma, 99.9 ± 5.1 Ma, 615.1 ± 57.6 Ma, 1025.3 ± 30.7 Ma and 2115.0 ± 262.7 Ma. The remaining four grains from sample MM-11A have isolated ages of 135.7 ± 2.4 Ma, 147.9 ± 2.0 Ma, 617.8 ± 12.9 Ma, and 873.2 ± 23.8 Ma.

[11] Igneous pluton sample USHP-1 contains a broad peak of 71–91 Ma zircons, constituting 98% (45) of the sample's analyzed grains. The sole remaining grain from sample USHP-1 has an age of 1492.1 ± 88.1 Ma. Integration of laboratory analytical errors with the TuffZirc age extraction algorithm [*Ludwig and Mundil*, 2002] indicates a crystallization age of 74.7 +2.2/−2.0 Ma (Table 1). The single Proterozoic grain indicates incorporation of inherited zircons from underlying country rock, which may also explain the older (>77 Ma) Late Cretaceous grains excluded by the TuffZirc algorithm.

6. Discussion

6.1. Depositional Age and Provenance of the Yahgán Formation

[12] The predominant age peaks of all three Yahgán Formation samples are broadly coincident with the biostratigraphic age assignments for the Aptian-Albian (circa 125–100 Ma) components of the Rocas Verdes basin fill [*Dott et al.*, 1977; *Olivero and Martinioni*, 1996] and strongly suggest that the analyzed strata are distinct from (i.e., younger than) those bearing Tithonian-"Neocomian" (circa 150–130 Ma) fossils [*Katz*, 1963; *Suárez et al.*, 1985]. In samples MM-11A and REMO-1C, these predominant age peaks constitute the youngest populations in each sample, suggesting an Albian maximum depositional age that is consistent with existing biostratigraphic age assignments for broadly equivalent strata. In contrast, the small but significant population of 75–83 Ma zircons (mean age of 77.7 ± 3.0 Ma) in sample MM-1 are drastically younger than any other zircons reported from the Rocas Verdes basin fill or immediately superjacent or subjacent units [*Fildani et al.*, 2003; *Calderón et al.*, 2007; this study]. In light of the quartz veinlets recognized in the sampled stratigraphy and existing biostratigraphic constraints on interpreted stratigraphic equivalents, parsimony suggests this subpopulation

Table 1 (Sample). U/Pb Zircon Geochronology of the Rocas Verdes Basin Fill and Crosscutting Plutons[a] [The full Table 1 is available in the HTML version of this article]

			Isotope Ratios								Apparent Ages (Ma)							
	U (ppm)	U/Th	$^{206}Pb/^{204}Pb$	$^{206}Pb^*/^{207}Pb^*$	± (%)	$^{207}Pb^*/^{235}U^*$	± (%)	$^{206}Pb^*/^{238}U$	± (%)	Error Correlation Coefficient	$^{206}Pb^*/^{238}U^*$	± (Ma)	$^{207}Pb^*/^{235}U$	± (Ma)	$^{206}Pb^*/^{207}Pb^*$	± (Ma)	Preferred Age (Ma)	± (Ma)
REMO-1C, Yahgan Formation, Punta Remolino, 54°52.829′S, 067°44.450′W																		
REMO1C-01	193	2.0	862	22.3862	8.2	0.1061	8.3	0.0172	1.1	0.14	110.1	1.2	102.4	8.1	-	-	110.1	1.2
REMO1C-02	293	1.3	1660	21.5521	5.3	0.1126	5.4	0.0176	1.0	0.19	112.5	1.1	108.4	5.5	-	-	112.5	1.1
REMO1C-03	204	1.9	1478	13.1996	8.1	0.1774	8.2	0.0170	1.4	0.17	108.6	1.5	165.8	12.6	-	-	108.6	1.5
REMO1C-04	254	2.9	1786	13.9261	6.8	0.1674	7.1	0.0169	2.2	0.31	108.1	2.3	157.1	10.3	-	-	108.1	2.3
REMO1C-05	195	1.9	990	14.8327	8.4	0.1590	8.5	0.0171	1.2	0.14	109.3	1.3	149.8	11.8	-	-	109.3	1.3
REMO1C-06	610	3.0	3110	14.6145	3.4	0.1677	3.6	0.0178	1.2	0.32	113.6	1.3	157.4	5.2	-	-	113.6	1.3
REMO1C-07	106	2.6	800	16.5162	25.4	0.1477	25.4	0.0177	1.0	0.04	113.1	1.1	139.9	33.2	-	-	113.1	1.1
REMO1C-08	357	1.7	1824	12.6087	5.6	0.1899	5.9	0.0174	1.8	0.31	111.0	2.0	176.6	9.5	-	-	111.0	2.0
REMO1C-09	338	1.7	3384	13.6931	12.9	0.1780	12.9	0.0177	1.0	0.08	113.0	1.1	166.3	19.8	-	-	113.0	1.1
REMO1C-10	230	1.7	1790	13.2900	10.3	0.1803	10.3	0.0174	1.0	0.10	111.1	1.1	168.3	16.0	-	-	111.1	1.1
REMO1C-11	405	1.8	2656	12.7406	5.3	0.1840	5.7	0.0170	2.0	0.36	108.7	2.2	171.5	9.0	-	-	108.7	2.2
REMO1C-12	496	0.9	1922	12.4271	4.5	0.2068	4.8	0.0186	1.7	0.36	119.0	2.1	190.9	8.4	-	-	119.0	2.1
REMO1C-13	155	1.4	1708	15.1983	18.5	0.1580	18.6	0.0174	1.0	0.05	111.3	1.1	149.0	25.7	-	-	111.3	1.1
REMO1C-14	108	1.3	752	15.0521	17.0	0.1635	17.1	0.0179	2.3	0.13	114.1	2.6	153.8	24.5	-	-	114.1	2.6
REMO1C-15	440	1.4	3468	21.4614	5.3	0.1148	5.4	0.0179	1.0	0.19	114.2	1.1	110.3	5.6	-	-	114.2	1.1
REMO1C-16	368	1.2	2914	21.6898	7.7	0.1063	7.9	0.0167	1.9	0.24	106.9	2.0	102.6	7.7	-	-	106.9	2.0
REMO1C-17	111	1.7	872	22.1778	12.5	0.1069	12.6	0.0172	1.0	0.08	109.9	1.1	103.1	12.3	-	-	109.9	1.1
REMO1C-18	329	3.1	2642	19.6049	9.9	0.1299	10.0	0.0185	1.0	0.10	117.9	1.2	124.0	11.6	-	-	117.9	1.2
REMO1C-19	564	1.7	5422	20.8510	4.0	0.1170	4.1	0.0177	1.2	0.28	113.0	1.3	112.3	4.4	-	-	113.0	1.3
REMO1C-20	403	1.8	4052	21.0608	2.7	0.1158	2.9	0.0177	1.0	0.34	113.0	1.1	111.2	3.1	-	-	113.0	1.1
REMO1C-21	60	1.8	436	20.4253	23.6	0.1182	23.6	0.0175	1.7	0.07	111.9	1.9	113.4	25.4	-	-	111.9	1.9
REMO1C-22	116	1.6	956	25.8437	21.6	0.0950	21.6	0.0178	1.0	0.05	113.8	1.1	92.2	19.0	-	-	113.8	1.1
REMO1C-23	192	1.4	1762	23.2067	11.9	0.1069	12.2	0.0180	2.5	0.21	115.0	2.9	103.2	11.9	-	-	115.0	2.9
REMO1C-24	148	1.6	1360	21.9163	9.4	0.1122	9.4	0.0178	1.0	0.11	113.9	1.1	108.0	9.7	-	-	113.9	1.1
REMO1C-25	144	1.8	1696	21.3639	18.8	0.1172	18.8	0.0182	1.0	0.05	116.0	1.2	112.6	20.0	-	-	116.0	1.2
REMO1C-26	166	2.1	1766	22.9891	10.2	0.1066	10.4	0.0178	1.6	0.15	113.6	1.8	102.9	10.1	-	-	113.6	1.8
REMO1C-28	581	1.5	4092	21.3416	5.8	0.1141	6.4	0.0177	2.7	0.42	112.9	3.0	109.7	6.6	-	-	112.9	3.0
REMO1C-29	121	1.7	1114	22.7480	10.5	0.1081	10.6	0.0178	1.5	0.14	114.0	1.7	104.2	10.5	-	-	114.0	1.7
REMO1C-31	308	1.3	2602	20.7390	4.8	0.1189	5.0	0.0179	1.0	0.20	114.2	1.1	114.0	5.3	-	-	114.2	1.1
REMO1C-32	232	1.4	2096	21.3193	5.7	0.1148	6.0	0.0178	1.8	0.30	113.4	2.0	110.4	6.3	-	-	113.4	2.0
REMO1C-33	302	2.0	3868	21.6567	6.3	0.1151	6.4	0.0181	1.1	0.17	115.5	1.2	110.7	6.7	-	-	115.5	1.2
REMO1C-34	451	2.6	6114	21.1441	3.4	0.1240	3.9	0.0190	1.9	0.48	121.4	2.2	118.7	4.3	-	-	121.4	2.2

[a] USHP-1 crystallization age per TuffZirc age extraction algorithm [*Ludwig and Mundil*, 2002] and analytical errors: 74.7 +2.2/−2.0 Ma (96.4% conf., from coherent group of 18 grains). All uncertainties are reported at the 1-sigma level, and include only measurement errors. Systematic errors would increase age uncertainties by 1–2%. U concentration and U/Th are calibrated relative to a Sri Lanka standard zircon, and are accurate to ~20%. Common Pb correction is from ^{204}Pb, with composition interpreted from *Stacey and Kramers* [1975] and uncertainties of 1.0 for $^{206}Pb/^{204}Pb$, 0.3 for $^{207}Pb/^{204}Pb$, and 2.0 for $^{208}Pb/^{204}Pb$. U/Pb and $^{206}Pb/^{207}Pb$ fractionation is calibrated relative to fragments of a large Sri Lanka zircon of 564 ± 4 Ma (2-sigma). U decay constants and compositions as follows: $^{238}U = 9.8485 \times 10^{-10}$, $^{235}U = 1.55125 \times 10^{-10}$, $^{238}U/^{235}U = 137.88$.

10.1029/2009GC002749

Figure 2. Cathodoluminescence scanning electron micrographs of randomly selected grains from igneous sample USHP-1. Boxed numbers indicate analysis number depicted in Table 1. Ages are from $^{206}Pb/^{238}U$ and are reported in Ma at the 1σ level.

of detrital zircons was derived from hydrothermal or magmatic contamination. However, the striking absence of known Late Cretaceous zircon-bearing intrusions within 10 km of Monte Martial and in the immediate subsurface raises the possibility that parts of the Yahgán Formation may be considerably younger (Campanian) than perceived. If so, inversion of the Rocas Verdes basin by Andean orogenesis would have occurred more than 20 Myr later than previously thought [e.g., *Fildani and Hessler*, 2005]. We consider this unlikely, but worthy of additional consideration and sample analysis.

[13] Our detrital zircon results are in accord with the interpretation of the Rocas Verdes basin as receiving detritus dominantly from the adjacent Patagonian arc [*Winn*, 1978; *Fildani et al.*, 2003; *Calderón et al.*, 2007]. However, we note that despite indistinguishable differences in lithofacies, comparable stratigraphic elevations and close proximity, the samples MM-1 and MM-11A have distinct detrital zircon age populations that fail the Kolmogorov-Smirnov comparison test (p = 0.002), indicating a high likelihood of having been derived from distinct sediment sources within the arc.

6.2. Age of the Ushuaia Pluton and Associated Rocks

[14] Cathodoluminescence imaging of Late Cretaceous USHP-1 zircons (Figure 2) reveals widespread oscillatory zoning and an absence of metamorphic rims, indicating formation from a magma with negligible secondary growth. Thus, our reported zircon mean age (74.7 +2.2/−2.0 Ma; Figure 4) likely records emplacement and crystallization of the Ushuaia pluton. These results call into question the previously reported 113 ± 5 Ma and 100 ± 6 Ma ages [*Acevedo et al.*, 2002] derived from the less reliable whole-rock K-Ar isotope system. As a result, parts or all of the retroarc "Shoshonitic Rock Complex" intrusives to which the Ushuaia pluton is interpreted to belong [*González Guillot et al.*, 2009] may be considerably younger than currently perceived, and/or they might be better equated with the Beagle Plutonic Suite, separate plutons of which have yielded U-Pb zircon ages of 90 ± 2 and 69 ± 1 Ma [*Kohn et al.*, 1995, and references therein].

[15] The USHP-1 ages also constrain the depositional and deformational age of the crosscut Yahgán

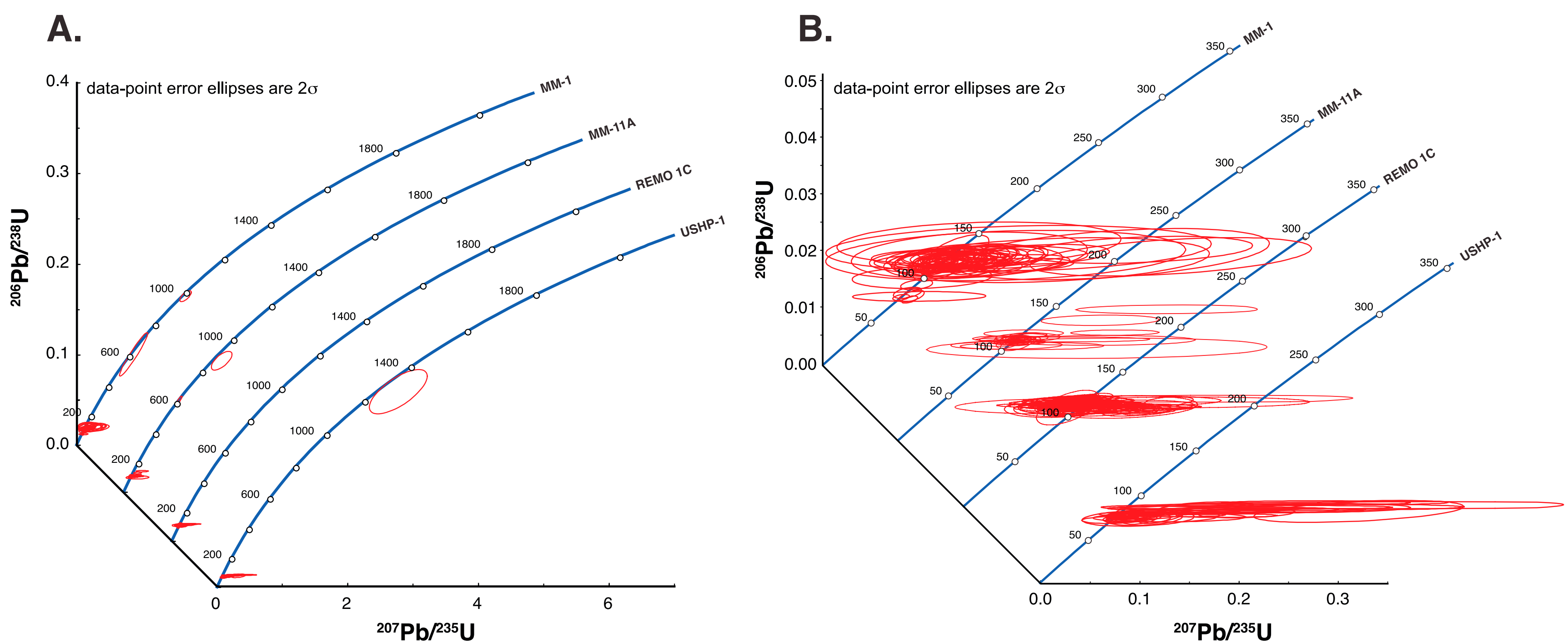

Figure 3. U-Pb concordia plots for samples analyzed in this study.

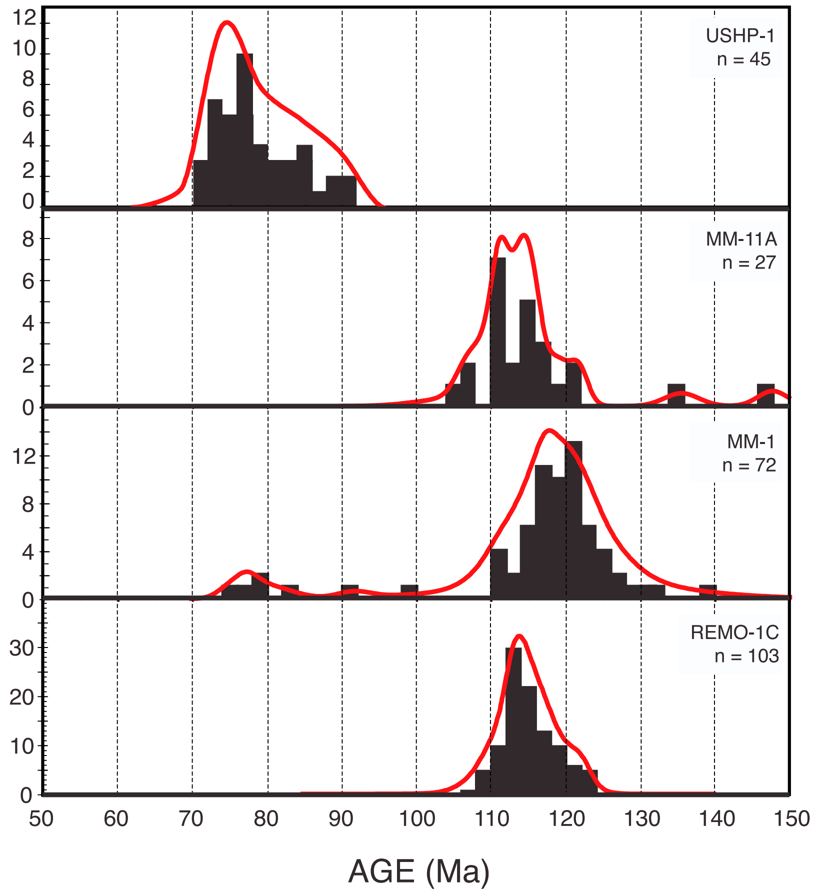

Figure 4. Histogram (2 Myr bins) and probability plots for U-Pb zircon data reported in this study.

Formation to the Campanian or older. The 74.7 +2.2/−2.0 Ma age we determined for the Ushuaia pluton indicates that the Rocas Verdes basin fill in the immediate study area is no younger than circa 73 Ma, in line with all available depositional age constraints. If the Late Cretaceous (77.7 ± 3.0 Ma) population of zircons in detrital sample MM-1 are not contaminants, deposition of the studied Yahgán Formation could be constrained to between circa 73 and 81 Ma. However, as stated above, the vast majority of existing chronostratigraphic data strongly indicate that the Rocas Verdes basin fill is most likely no younger than late Albian. Hence until further geochronology of the southern Rocas Verdes basin can evaluate the surprisingly young result obtained from detrital sample MM-1, we favor the older age interpretation of the Rocas Verdes basin fill.

7. Summary

[16] In addition to allowing more detailed stratigraphic division of the Rocas Verdes basin fill, our new U-Pb zircon ages suggest that comprehensive analysis of various detrital zircon age spectra of the Rocas Verdes strata may enable improved structural tie points and cutoffs required for accurate estimation of intrabasinal shortening that is essential to constraining the tectonic history of southern South America and the Scotia arc. Further, these and future data should provide a valuable tool for comparison of the southern basin preserved in the Fuegian Andes with the well-studied Rocas Verdes units of Chilean Patagonia. The variations in sediment composition, together with the new ages of plutonism will allow for a reevaluation of the source region evolution during the development of the Rocas Verdes basin in detail and the Scotia arc in general.

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