High-resolution images above the Pampean flat slab of Argentina (31-32°S) from local receiver functions: Implications on regional tectonics

Jean-Baptiste Ammirati ^{a,*}, Sofía Pérez Luján ^{a,1}, Patricia Alvarado ^a, Susan Beck ^b, Sebastián Rocher ^c, George Zandt ^b

*a) Centro de Investigaciones de la Geósfera y Biosfera (CIGEOBIO-CONICET),
Universidad Nacional de San Juan, Argentina*

b) Department of Geosciences, University of Arizona, USA

*c) Centro Regional de Investigaciones y Transferencia Tecnológica de La Rioja
(CRILAR-CONICET), Argentina*

** Corresponding author, contact mail: jb.ammirati@unsj-cuim.edu.ar*

*¹ Now at Instituto de Estudios Andinos Don Pablo Groeber (IDEAN-CONICET),
Universidad de Buenos Aires, Argentina*

Abstract

In the flat slab region of the South Central Andes (~31-32°S), geological observations suggest that the regional crustal structure is inherited from the accretion of different terranes during the Ordovician. These structures were later reactivated, first in extension

20 during the Triassic and later in compression during the Andean uplift since the
21 Miocene. Seismological observations confirmed that those fault structures extend to
22 depth with décollement levels that accommodate current crustal shortening in the
23 region. In order to get better insight on the regional tectonics we computed higher
24 frequency receiver functions from local slab seismicity of intermediate ~100 km depth.
25 Using a common conversion point (CCP) stacking method we obtained cross sections
26 showing high vertical resolution crustal structure at the transition between the
27 Precordillera and the Frontal Cordillera. In addition we performed a joint inversion of
28 our high frequency receiver functions with surface wave data from ambient noise
29 tomography allowing us to constrain absolute seismic wave velocities. Our higher
30 resolution images reveal more structural details down to a depth of 80 km and laterally
31 over the flat slab in good agreement with previous studies. Our results help to better
32 identify very shallow discontinuities in seismic velocities. Recent petrological analyses
33 combined with our high-resolution RF structure correlates with a crustal mafic
34 composition and partial eclogitization in the lower crust. We observe a shift in the
35 crustal structure between the Precordillera (east) and the Frontal Cordillera (west).
36 Regional seismicity and previously determined focal mechanisms superimposed over
37 our images indicate this shifting is a thrust structure extending down to a depth of 40
38 km. Our results combined with petrological analyses suggest the presence of a master
39 décollement between the Cuyania (Western Precordillera) and Chilenia (Frontal
40 Cordillera) terranes that probably accommodates the crustal deformation in the
41 Pampean flat slab region since the Late Ordovician.

43 **Key words:** Terrane, Andean retroarc, Subduction, Body waves, Surface waves,
44 Eclogite, South America

45

46 **1. Introduction and geological setting**

47

48 Between 30° and 33°S, the tectonics of the South American western margin is
49 characterized by a series of foreland uplifts related to the flat subduction of the Nazca
50 plate under the South American plate (Ramos 1988). Well recorded in the South Central
51 Andes segment (30-33°S), the structural, magmatic and sedimentary response to the
52 Cenozoic slab flattening has been intensely studied over the years and, combined with
53 seismological observations, has defined the Pampean (Chilean) flat slab region in time
54 and space (e.g. Stauder 1973; Allmendinger et al. 1990; Kay & Abbruzzi 1998; Ramos
55 et al. 2002; Anderson et al. 2007).

56 The flat slab region is composed of three main morpho-structural units (Fig. 1): i) The
57 eastern Sierras Pampeanas (in the Pampia terrane), ii) The Precordillera and the western
58 Sierras Pampeanas (Cuyania terrane) and iii) Cordillera (Chilenia terrane), which
59 includes the Frontal Cordillera (east) and the Principal Cordillera (west). Regional
60 tectonics involves the Famatinian (Early Ordovician-Early Carboniferous), Gondwanic
61 (Early Carboniferous-Early Cretaceous) and Andean (Early Cretaceous-present)
62 orogenic cycles (Ramos 1988).

63 The western Sierras Pampeanas are characterized by tilted blocks of Mesoproterozoic
64 ultramafic and mafic rocks, mainly metagabbros and garnet-amphibolites well exposed
65 in the Cuyania terrane basement (Kay et al. 1996). The Precordillera is mainly
66 composed of Paleozoic marine carbonate and siliciclastic platform sequences of
67 Cambrian to Devonian ages (Baldi et al. 1982). Western Precordillera rocks consist of
68 slope and deep marine deposits affected by tight folding and low-grade metamorphism
69 (Pérez Luján et al. 2015). The presence of an Ordovician mafic-ultramafic belt that
70 outcrops in the Central and Western Precordillera is generally considered the exposed
71 boundary between the Cuyania and Chilenia terranes (Kay et al. 1984; Davis et al.
72 1999; Pérez Luján et al. 2015).

73 The Uspallata-Calingasta valley, a large tectonic depression, separates the Frontal
74 Cordillera from the Precordillera (Fig. 2). Most of the current Frontal Cordillera formed
75 during the Gondwanic orogeny (Neopaleozoic to Triassic times). During this period, the
76 Frontal Cordillera recorded significant extensional tectonics characterized by the
77 Permian-Triassic large plutonic-volcanic Choiyoi complex (Heredia et al. 2002; Rocher
78 et al. 2015). More to the west, the Principal Cordillera is composed of Mesozoic
79 continental and marine deposits exhibiting shallow thin- and thick-skin deformation (La
80 Ramada fold and thrust belt; Cristallini & Ramos 2000).

81 Recent focal mechanism solutions (Alvarado et al. 2010; Ammirati et al. 2015) show
82 that the Cuyania-Chilenia boundary is affected by moderate magnitude earthquakes (4.2
83 $< M_w < 5.4$) with reverse focal mechanisms, down to ~ 30 km depth. In this work, we
84 present new higher-resolution crustal velocity structure images for the same Andean
85 region above the Pampean flat slab. We compare our improved crustal velocity structure

determined from local slab earthquake receiver functions and ambient noise with petrological analyses. In particular, we investigate the transition zone between the Cuyania and Chilena terranes in order to propose a seismotectonic model that could explain the current configuration of this boundary in terms of rock composition and tectonic evolution.

2. Previous works

For this study we mainly used IRIS-PASSCAL data from the Sierras Pampeanas Experiment using a Multicomponent Broadband Array (SIEMBRA). This experiment consists in 43 broadband stations deployed in the Pampean flat slab region in December 2007 and operated for two years. The CHile-ARgentina Geophysical Experiment (CHARGE) deployed in the region in 2001 also allowed us to compute some good quality local RFs beneath the Sierra de Pie de Palo (station JUAN: 31.60°S; 68.23°W) in the western Sierras Pampeanas (Fig. 1). Another seismic experiment deployed in 2008-2010 in the eastern Sierras Pampeanas (the ESP experiment) was not directly used for the computation of RFs in this work but contributed to determine phase velocity dispersion from ambient noise tomography (Porter et al. 2012), which we jointly inverted with our local RFs to obtain absolute shear-wave velocities (Fig. 1 & S1).

The aforementioned experiments deployed in the Pampean flat slab region contributed to better understand the flat subduction geometry and the associated tectonic response of the overriding plate. For example, Anderson et al. (2007) analyzed the slab seismicity recorded during the CHARGE experiment to propose new slab top contours refining

108 those determined by Cahill & Isacks (1992) and Pardo et al. (2002) and showing nicely
109 how the flat portion of the subducting slab coincides with the projection of the path of
110 the Juan Fernández Ridge (JFR) under the South American plate (Fig. 1).

111 Using the same data, Wagner et al. (2005) and Porter et al. (2012) have performed body
112 wave and ambient noise tomography respectively, in the region and proposed
113 interpretations about how the flat portion of the subducting slab is probably still
114 hydrated and contrasts with the overlying drier and cooler mantle wedge. Teleseismic
115 RF and converted-wave analysis (Regnier et al. 1994; Gilbert et al. 2006; Perarnau et al.
116 2010; Gans et al. 2011; Ammirati et al. 2013, 2015) show an abnormally thick crust for
117 the western Cuyania terrane (~65 km). Some of these studies have observed high shear-
118 wave velocity for depths greater than 40 km likely associated with partial eclogitization
119 in the lower crust. Most of these RF studies also reveal the presence of mid-crustal
120 discontinuities associated with regional seismicity and décollements accommodating
121 global shortening at crustal levels.

122 Although the crustal structure of the Sierras Pampeanas and the Precordillera has been
123 extensively studied, the western part of the Pampean flat slab region, especially the
124 transition zone between Cuyania and Chilenia and the Frontal Cordillera remain poorly
125 interpreted in terms of seismic velocity structure.

126 Some studies have focused on regional rock composition and seismic characteristics in
127 terms of wave velocities. For example, Castro de Machuca et al. (2012) have
128 determined the main mineral assemblages for the most representative rocks from the
129 Sierra de Pie de Palo complex (WSP) compared with a database of typical subduction

130 rocks and estimated their physical and seismic properties. Assuming pressure-
131 temperature (P-T) peak metamorphic conditions, the results were then compared with a
132 regional velocity model obtained by Perarnau et al. (2010) to estimate rock facies as a
133 function of depth. They concluded that the lower crust deeper than 28 km depth had
134 high S-wave velocities and low V_p/V_s representative of crustal densification and mafic
135 rocks in the upper amphibolite to granulite/eclogite facies. Exploring P- and S-wave
136 crustal velocities, Venerdini et al. (2016) have confirmed these observations.

137 Similar studies were performed to the west in the Central and Western Precordillera by
138 Pérez Luján et al. (2015). Shear-wave velocity estimations obtained by them from
139 representative rock samples collected in the Sierra de la Invernada, Sierra del Tigre and
140 Rodeo areas (Fig. 2), can be projected to depths greater than 30 km strengthening the
141 hypothesis of a mafic-ultramafic composition for the middle and lower crust beneath the
142 Central and Western Precordillera. Their results support the idea of a deformed mafic-
143 ultramafic belt at the transition zone between the Precordillera and the Frontal
144 Cordillera extending in depth to levels deeper than 50 km. Boedo et al. (2013) show
145 how the same exposed belt extends in the south to about 33°S where the Precordillera
146 ends.

147 In this study we combine high-frequency RF, ambient noise dispersion data and
148 petrological analyses to better constrain the Cuyania-Chilenia terrane boundary.

149 **3. Data and Methods**

150 **3.1. Local receiver functions**

151

152 RF analysis uses broadband waveforms recorded by a 3-components seismometer to
153 identify P-to-S conversions from discontinuities beneath the station. The waveforms
154 coming from distant earthquakes will cross discontinuities in seismic velocities at a
155 steep angle and generate P- to S-wave conversions that reverberate beneath the receiver.
156 Hence, a receiver function is a time series of Gaussian pulses where each spike marks a
157 phase arrival (direct wave and/or associated multiples) directly related to the Earth
158 structure (Langston, 1979).

159 In this study, we compute receiver functions from local intermediate-depth seismicity.
160 Local earthquakes are characterized by higher frequency content than that one from
161 teleseismic earthquakes; this allows better vertical resolution since high frequency
162 waveforms are sensitive to smaller velocity variations (Calkins et al. 2006; Perarnau et
163 al. 2012). Local seismograms are also characterized by a higher signal-to-noise ratio
164 allowing us to work with lower magnitude earthquakes.

165 The dataset used in this work corresponds to the slab seismicity (183 events) recorded
166 by the SIEMBRA array between January 2008 and November 2009 (Fig. 3a). For each
167 station, we select earthquakes with a local magnitude M_L greater than 3.0 and an
168 epicentral distance less than 0.7 degree (~78 km) from the receiver. The epicentral
169 distance has been calculated considering an average focal depth of 100 km and a
170 maximum incidence angle of 35° allowing the assumption of plane waves and ensuring
171 traces that exhibit no interferences with the possible generation of surface waves (SW).
172 The average number of events per station is 54 with a maximum of 135 at station Río
173 Saso (SASO) in the Central Precordillera and a minimum of 1 at station Los Colorados
174 (COLO) in the Sierras Pampeanas. In general, we expect stations located above the flat

175 portion of the subducting slab to record a higher number of events since many studies
176 concluded that most of the seismicity occurs in the flat portion of the slab at around 100
177 km depth (Anderson et al. 2007).

178 Before computing the local RFs, we removed the mean and rotated the north-south and
179 east-west components into radial and tangential components respectively, along the
180 great circle path. Resulting traces were high-pass filtered (corner frequency of 0.5 Hz)
181 to avoid long-period trends. The difference is small comparing RFs obtained from
182 filtered and unfiltered traces since the local waveforms have a higher frequency content
183 (> 1 Hz) while long-period trends are more common in the long-period signals (< 0.5
184 Hz). However, we generally observe better quality RFs for filtered traces. To compute
185 local RFs, we use an iterative pulse stripping deconvolution method (Ligorria &
186 Ammon 1999) with 600 iterations maximum or an improvement of the fit between two
187 iterations lower than 0.01%. We use a Gaussian width of 5 in the deconvolution which
188 corresponds to a low-pass filter of corner frequency 2.4 Hz. A higher Gaussian width,
189 such as 10 (5 Hz) resulted in excessively noisy RFs. The Gaussian filter controls the
190 wavelength size, which is directly related to the maximum vertical resolution
191 recoverable. In this work, the RF vertical resolution is ~ 0.5 km.

192 The high frequency content of the waveforms combined with a high Gaussian width
193 value used to compute local RFs allows us to image more details at crustal levels in
194 comparison with teleseismic RFs (Fig. 4a). However, this implies more primary
195 conversions and thus, more multiples further back in time, making the RF noisy.
196 Moreover, the deconvolution is much more unstable when using higher frequency
197 waveforms. A challenging aspect consists of differentiating complex traces containing

198 bumps related to the structure from unusable noisy traces. For each station, RFs are thus
199 cross-correlated with each other in order to identify traces exhibiting the same pattern
200 (Fig. 4b). Selected traces were then visually inspected to eventually refine the RF
201 selection. By applying this method, the amount of bad quality RFs is significantly
202 reduced so we are confident that the remaining traces mainly contain information about
203 the structure. Nevertheless, some of the stations lack good quality data. This is the case
204 for stations GUAR and REGU located quite far west and close to our epicentral distance
205 limit criteria. Stations SASO and ZOND also present noisy traces likely due to site
206 effects.

207 The top of the slab has been imaged in previous studies lying around 95-105 km depth
208 (Gans et al. 2011; Ammirati et al. 2013, 2015). It appears as a strong negative arrival
209 that marks a decrease in seismic velocities due to the presence of hydrated oceanic
210 crustal material. Theoretically, high resolution RFs from the slab would be able to
211 image the top of the slab. However, because the slab top signal is expected 12-13 s after
212 first P-wave arrival, this arrival would be merged with the crustal multiples and thus
213 hard to identify. For this reason we only interpret crustal RF arrivals (between 0 and 80
214 km depths).

215 Figure S2 shows the average radial and tangential RFs obtained after move out
216 correction for each station used in this work. Theoretically, for wave propagation in a
217 homogenous and isotropic media, we do not expect transverse P-wave signal. The low
218 P-wave amplitudes observed for transverse RFs, in contrast to the corresponding high
219 radial RF amplitudes, support our assumption of homogeneity and isotropy in the
220 sampling region beneath each receiver. Only radial RFs were subsequently stacked.

221 Once computed and selected, the local RFs are stacked using the Common Conversion
222 Point (CCP) stacking method (details of this methodology can be seen in Dueker &
223 Sheehan 1997). Briefly, stacking receiver functions from multiple stations reduces the
224 noise and enhances the conversions and their lateral extent.

225 Traces are first converted from time to depth using the following travel time equation
226 (Gurrola et al. 1994):

$$\Delta T_{Ps}(z, p, V_s, V_p) = \int_0^z \left(\sqrt{V_s^{-2} - p^2} - \sqrt{V_p^{-2} - p^2} \right) dz \quad (1)$$

227 A discretized version of Eq. 1 calculates the travel time delay (ΔT_{Ps}) for each depth
228 increment ($dz_i = 1$ km in our case). Then, RFs are resampled with a point at every time
229 increment to the corresponding depth (z). The process automatically corrects the traces
230 from moveout since the ray parameter (p) of each RF is taken into account. Finally,
231 traces with common conversion points are binned and stacked. In this study, a total of
232 171 local RFs were stacked. The average number of local RFs is 11 per seismic station
233 and the minimum hit per bin is 5. The common conversion depth has been set to 60 km
234 which corresponds to the average Moho depth in the region of study (Gans et al. 2011,
235 Ammirati et al. 2015). The stacking bins have a radius of 20 km and are spaced by 10
236 km; hence two bins contain overlapping data, which provide lateral smoothing.

237 P- and S-wave velocities in Eq. 1 (V_p and V_s , respectively) come from a calibrated
238 velocity model obtained by Ammirati et al. (2015). This model contains crustal details
239 and has been tested for regional seismic source characterization. The uncertainties on
240 the Ammirati et al.'s (2015) crustal seismic velocities (0.3 and 0.2 km/s for P- and S-

241 wave velocities, respectively) cause inaccuracies up to 1.1 s in travel time estimations
242 when applying Eq. 1. Considering an average crustal shear-wave velocity of 3.5 km/s,
243 we estimate the uncertainty on crustal thickness to be ~4 km.

244 Three CCP cross sections are shown in Figs. 3b and 5. Since wave velocity is expected
245 to increase with increasing depth within the upper plate we chose not to interpret
246 negative arrivals. Those can be either noise-induced or PsPs+PpSs phases (negative
247 amplitude reverberations).

248 **3.2. Joint inversion**

249

250 We use the ANT phase velocity dispersion data obtained by Porter et al. (2012) to do a
251 joint inversion with our RF dataset. ANT is more sensitive to crustal shear-wave
252 velocities defining phase velocities for shorter periods (< 10 s) and makes it appropriate
253 to combine with high frequency RFs in order to get more detailed absolute velocity
254 structures. Porter et al. (2012) phase velocities are defined for periods ranging from 8 s
255 to 40 s.

256 For each station, depending on its location, we select the ANT dispersion curve
257 corresponding to the closest grid point on the Porter et al. (2012) model (Fig. S1). Then
258 the RFs are individually inverted with the dispersion data resulting in one velocity
259 model per RF. Those individual velocity models are then averaged to produce the final
260 station S-wave velocity model (Fig. S3). Since all receiver functions do not fit the data
261 in the same way, the averaged model is weighted with the RF corresponding variance
262 reductions.

263 We performed the joint inversion using the joint96 code (Julià et al. 2000; Herrmann &
264 Ammon 2002). The code will iteratively invert the datasets looking for the best velocity
265 model that minimizes both RF and SW residuals. A damping value of 0.5 was chosen as
266 the best compromise between data fit and unrealistic jumps in the resulting model. The
267 p-factor, that controls the influence of each dataset in the joint inversion, has been set to
268 0.3 which means that the RF dataset accounts for 70 % while the ANT dispersion
269 accounts for 30 % in the inversion. We observed that keeping the p-factor at low values
270 helps to preserve a good level of details in the resulting model although averaging the
271 models will also have a smoothing effect. Figure S3 shows the resulting models with 1
272 standard deviation confidence interval.

273 The initial velocity model consists of a series of 2-km-thick layers down to 150 km
274 depth characterized by constant seismic wave velocities and density with $V_p = 7$ km/s,
275 $V_p/V_s = 1.75$ and $\rho = 3$ g/cm³.

276 Between the stations, the shear-wave velocities have been extrapolated to produce the
277 image shown in Figure 6a. The RF positive arrivals interpreted as discontinuities in
278 seismic wave velocities (see next section) have been reported in Figure 6b in order to
279 compare the results from our CCP stacking and from the joint inversion.

280 **4. Results and interpretations**

281 **4.1. The Precordillera and western Sierras Pampeanas**

282

283 In the eastern part of the Pampean flat slab region, beneath the Sierra de Pie de Palo
284 (WSP), we observe a relatively simple shallow structure associated with a relatively
285 high shear-wave velocity (Fig. 5b & 6). This confirms that the structure of the WSP
286 mainly involves cored-based deformation of the Cuyania terrane as illustrated by the
287 rapid uplift of the Sierra Pie de Palo with 2 décollement levels around 20 and 30 km
288 depths (Ramos et al. 2002). Those observations are consistent with previous RF results
289 by Calkins et al. (2006) and Perarnau et al. (2010) and P- and S-wave velocity
290 estimations by Venerdini et al. (2016). The S-wave velocity obtained in this sector is
291 consistent with a Cuyania basement mostly characterized by metamorphic rocks in the
292 upper greenschist to amphibolite facies and upper amphibolite to granulite/eclogite
293 facies for rocks located at depths greater than 30 km (Castro de Machuca et al. 2012).

294 To the west, beneath the Eastern Precordillera several discontinuities are observable at
295 very shallow depths (< 10 km) associated with low S-wave velocities ($V_s < 3.2$ km/s)
296 (Fig. 6), generally associated with sedimentary rocks (Brocher 2005). In our study
297 region, the Eastern Precordillera outcrops are mainly composed of Cambrian limestones
298 (Marquesado Group) compatible with such low S-wave velocities. These limestones
299 mainly expose west-verging faulting extending beneath the Tulum Valley to the east
300 (Fig. 2), which is filled with Tertiary deposits (Baldis et al. 1982). According to our
301 results, we interpret the Tulum basin deposits to be ~ 5 km thick (corresponding to the
302 $V_s < 3.0$ km layer) and the base of the Ordovician stratification at ~ 10 km depth,
303 corresponding to a décollement level that separates the sedimentary strata from the
304 Cuyania basement. At deeper levels beneath the Eastern Precordillera we observe the

305 lateral continuation of the 32 km discontinuity that deepens to ~40 km depth toward the
306 northwest as observable along the cross section E-F (Fig. 5c).

307 The 18-20 km discontinuity observable in each cross section (Fig 5), beneath the
308 Central and Western Precordillera, has been interpreted in previous studies as the main
309 thrust between the thin-skinned deformed Precordillera and the Cuyania basement. Our
310 stacked RFs allow us to better constrain the lateral extent of this structure both in the E-
311 W and SE-NW directions. The relatively low S-wave velocity (~3.4 km/s) mainly
312 characterizing sedimentary rocks (Brocher 2005) observed in Figure 6 down to ~20 km
313 depth confirms a major lithology change at this depth.

314 Interestingly, we note that the regional seismicity (Linkimer 2011; Marot et al. 2013;
315 Venerdini et al. 2016) affects the crust mostly between ~20 and ~40 km corresponding
316 to the two aforementioned discontinuities (Fig. 5 & 6c). Shear-wave velocities observed
317 for these levels are comparable to those observed for the Sierra de Pie de Palo ($3.4 < V_s$
318 < 3.6 km/s) which is consistent with a similar type of metamorphic rocks forming the
319 Cuyania basement. The 40 km discontinuity coincides with a sharp decrease in the
320 Cuyania basement seismicity, which has been observed and interpreted as the
321 brittle/ductile transition for the Cuyania terrane (Regnier et al. 1994; Ammirati et al.
322 2013).

323 All cross sections show major arrivals beneath the WSP (~54 km), the Eastern
324 Precordillera (53-55 km) and Central and Western Precordillera (~65 km). In good
325 agreement with previous studies, we interpret these arrivals as the Moho discontinuity.
326 Figure 5a exhibits quite an important jump in Moho depth (~65 km beneath the

327 Precordillera and ~52 beneath the Eastern Precordillera). An interesting observation is
328 the superposition of two Moho arrivals beneath the Central Precordillera (well visible in
329 Fig. 5b). This double Moho configuration coincides with zones of particularly poor
330 Moho signal observed in previous teleseismic RF studies (Gilbert et al. 2006; Ammirati
331 et al. 2013) which had then been interpreted as partial eclogitization of the lower crust.
332 Our high resolution local RFs allow us to better resolve the Moho signal beneath the
333 Central Precordillera, that would be in fact formed by two smaller arrivals at (60-65 km
334 and 50-55 km). Although this double Moho could be a result of combining a complex
335 3D geometry for the Moho and projection effects when realizing the cross section, Gans
336 et al. (2011) and Ammirati et al. (2015) showed that in the flat slab region, the Moho is
337 clearly and rather abruptly deepening to the west. Cross section C-D is quite
338 perpendicular to the N-S structures that would affect the Moho geometry. Projection
339 effects are thus unlikely.

340 We suggest that the double Moho signal could delimitate a zone where most of the
341 Cuyania crustal material is turned to eclogites. Our absolute S-wave velocities of 4.2 -
342 4.3 km/s obtained for this zone are consistent with eclogites S-wave velocities
343 (Christensen & Mooney 1995). Previous observations from Gilbert et al. (2006), Porter
344 et al. (2012) and Ammirati et al. (2015) at lithospheric scale also suggest that the flat
345 slab is releasing water that could contribute to the lower crust eclogitization by
346 migration from the slab to the base of the crust.

347 Our shear-wave velocities for depths greater than 40 km (> 3.8 km/s) combined with
348 petrological analyses (Pérez Luján et al. 2015) show that the Cuyania terrane mainly
349 consists of mafic rocks (such as leuco-gabbros, gabbros, olivine-enriched gabbros and

350 wherlites) (Fig. 7). The main mineral composition mostly consists in Ca-rich
351 plagioclase (labradorite and bitownite) and clinopyroxene (diopside and augite).
352 Primary mineral assemblage exhibits low pervasive alteration where plagioclase is
353 variably altered to epidote, calcite, sericite and chlorite. Clinopyroxene is mostly fresh
354 and sometimes presents rims of pseudomorphic tremolite and chlorite, which implies
355 low-grade metamorphism.

356 Previous studies using microscopic Fourier transform infrared spectroscopy have
357 revealed that clinopyroxene and plagioclase (nominally anhydrous minerals) commonly
358 contain a significant amount of water as OH (Yang et al. 2008). The breakdown of
359 plagioclase in gabbroic lithologies, favored by the presence of water, would drive to
360 eclogitization (Hacker, 1996). In the Cuyania terrane, we expect partial eclogitization
361 occurring deeper than 40 km with an increasing fraction of eclogite with increasing
362 depth.

363 **4.2. The Frontal Cordillera crust**

364

365 Although some features of the Cordilleran crustal structure appear similar to the ones
366 beneath the Precordillera, their interpretation diverges due to significant differences in
367 terrane composition and geological history. Also, due to the restriction on the epicentral
368 distance, stations in the Frontal Cordillera (GUAR, REGU and CASP) produce fewer
369 RFs, which imply that the stacking may have not reduced the noise significantly.
370 Although a poorer quality for the images obtained in this region, some features appear
371 well identifiable and quite consistent with geological observations.

372 Cross sections in Figure 5 at 69.4°W, show shallow arrivals at ~5 km depth that could
373 be related to the bottom of the thick Tertiary/Quaternary deposits filling the Uspallata-
374 Calingasta valley. This 5-km depth conversion seems to extend and deepen to the west
375 and merge with another arrival at 10 km depth (Fig 5a & c). Both arrivals could be
376 generated by the presence of the Frontal Cordillera thrust belt décollement (Heredia et
377 al. 2002). Geological studies in the region indicate that the basement consists of a thick
378 (~2.5 km) rhyolitic sequence of lavas, ignimbrites and volcanoclastic sediments
379 characterizing the Permian-Triassic Choiyoi Group, intruded by Triassic granitoids
380 (Cristallini & Ramos 1999; Gambiagi & Ramos 2002; Rocher et al. 2015) (Fig. 2). We
381 observe a major arrival at a depth of approximately 15 km (Fig. 5a & c) that seems to
382 reach 20 km depth more in the south (Fig. 5b). Since there is no geological evidence for
383 a drastic change in lithology at this depth and considering that the Chilenia basement is
384 mainly composed of quartz enriched material (Cristallini & Ramos 1999), we interpret
385 this 15-20 km arrival as the brittle-ductile transition generally expected around such
386 depths in quartz-enriched formations (granite and granodiorite).

387 In terms of absolute velocities (Fig. 6) we observe that lower shear-wave velocities at
388 both shallow and deeper levels generally characterize the Frontal and Principal
389 Cordillera crust. Between 0 and 10 km depths low shear-wave velocities (< 3.4 km/s)
390 characterize the intensely deformed Mesozoic sediments. We note that some shallow
391 seismicity (Fig. 6c) in this area seems to coincide with the discontinuity at ~10 km and
392 would thus mark the main thrust décollement separating the deformed Mesozoic cover
393 from the underlying basement. Relatively low S-wave velocities are observed in the
394 Chilenia basement (~3.5 km/s) down to 45-50 km depths, compatible with quartz-

395 enriched material (Christensen & Mooney 1995). A very clear arrival between 67 and
396 69 km on all studied cross sections (Fig. 5) corresponds to the Moho. These Moho
397 depths are in good agreement with Gans et al. (2011) results although our higher
398 resolution images allow for a sharper imaged Moho conversion.

399 At depth, the increase in crustal pressure and temperature (200 MPa - 550°C at the base
400 of the crust, according to Marot et al. 2014) will eventually allow the transition to felsic
401 granulite and/or eclogites. We believe that this transition occurs at 50-55 km since all
402 cross sections show arrivals at this depth. We also observe a significant increase in
403 shear-wave velocities (from 3.6 to > 3.8 km/s) compatible with an eclogite facies
404 beneath the Principal and Frontal Cordillera.

405 Interestingly, comparing our results with recorded seismicity shows a cluster of seismic
406 activity around 70.2°W (Fig. 6c). We believe that this seismicity is related to the
407 presence of the La Ramada thrust belt affecting the Principal Cordillera with a
408 discontinuity at ~10 km separating the Mesozoic strata from the basement and a
409 décollement level at ~20 km depth that coincides with the brittle-ductile transition, in
410 good agreement with Cristallini & Ramos (2000).

411 **4.3. The Cuyania-Chilenia terrane boundary**

412

413 In the south central Andes, the evolutionary history for the western margin of Gondwana
414 is still widely debated and mostly depends on the presence (or absence) of a Devonian
415 magmatic arc between the Chilenia and Cuyania terranes. At 31-32°S, there is no
416 evidence for such magmatism. Moreover, Ordovician basalts in the Calingasta and

417 Rodeo areas (Fig. 2) present geochemical characteristics of early stages of extending
418 oceanic ridge magmatism (Kay et al. 1984), which suggests that the Chilenia and
419 Cuyania terranes were probably part of the same block that spread apart during the
420 Ordovician. The extension probably resulted in subsidence and areas of low elevation
421 that were filled with the fine sedimentary material interlayered with the mafic-
422 ultramafic outcrops observable in the current Western and Central Precordillera (Fig. 7).
423 Extension in this region, however, was probably not significant enough to generate
424 oceanic crust. During the Late Ordovician, when the eastern margin of the Chilenia-
425 Cuyania block docked to the western margin of Gondwana, the extension stopped and
426 the ocean recently formed between the Chilenia and Cuyania terranes would have
427 closed without generating arc type magmatism. The low metamorphic grade and the
428 double vergence observed by Pérez Luján et al. (2015) in the Sierra de la Invernada and
429 Rodeo area (Fig. 2) supports this hypothesis of a narrow ocean closed by the tectonic
430 inversion of previous extensional structures at this latitude. We believe that our RF
431 results presented in the previous section allowed us to identify one of these structures
432 (Fig. 5 & 6). This would be a major Ordovician west dipping master fault affecting the
433 whole crust and then inverted in reverse faulting from late Ordovician to Middle
434 Devonian, driving to the formation of a proto-Precordillera (current Western and
435 Central Precordillera). Under compression, this structure would lead the Cuyania
436 basement under the Chilenia basement in a continental collision.

437 Later, the Chilenia terrane recorded intense orogenic magmatism as suggested by the
438 presence of the Permian-Triassic Choyoi formation (Fig. 8a). The Choyoi is almost
439 absent in the Precordillera, east of the Uspallata-Calingasta valley. The presence of a

major west dipping structure between the Frontal Cordillera and the Precordillera is consistent with this observation. The Choiyoi Group has been defined as a volcanic complex developed in extensional conditions whose emplacement and deposition occurred in a series of half-grabens controlled by normal faults with downthrown western blocks (Heredia et al. 2002). In the Frontal Cordillera, the Choiyoi Group is more than 2.5 km thick and associated with large granitic plutons (Heredia et al. 2002; Rocher et al. 2015). In contrast, the Precordillera, Permian-Triassic magmatic rocks are very scarce and mostly restricted to the southern Precordillera of Mendoza at ~33°S (Strazzere et al. 2006). The structure recognized in this work could have played an important role during the Permian-Triassic, accommodating extensional tectonics between Cuyania and Chilenia (Fig. 8a). In this scenario the magmatic activity and extensional deformation mainly affected the hanging-wall, which in this case corresponds to the Frontal Cordillera (Chilenia), leaving the foot-wall (Cuyania) less affected by the Choiyoi magmatism (e.g. Ebinger & Casey 2001). Also, the Choiyoi magmatism probably changed the composition of the Chilenia basement by intrusion of felsic material (Rocher et al. 2015) and more importantly, probably contributed to the creation of new, granitic continental crust.

During the Miocene, the Andean compression inverted previous structures both inherited from the Ordovician and the Permian-Triassic. The important shallow seismic activity recorded in the region (Fig. 6c) demonstrates that this global shortening is still active. We suggest that our master structure has been reactivated in reverse faulting and accommodates crustal deformation down to ~40 km depth by thrusting the Chilenia terrane over the Cuyania terrane. Reverse focal mechanisms for crustal intermediate

463 magnitude earthquakes (Alvarado et al. 2010; Ammirati et al. 2015) support this
464 hypothesis. Moreover we observe a cluster of seismicity around (69.3°W) presenting a
465 west-dipping trend at the transition zone between the Cuyania and Chilenia terranes
466 where our master structure is expected.

467 Although not very well constrained, some features suggested by our seismic results
468 allow us to propose a hypothesis that supports shortening mechanisms in the lower
469 crust. The possibility exists that the described west-dipping Ordovician structure would
470 affect the whole crust by simple shear thrusting. However in this case, Moho
471 deformation should be observed beneath the Frontal and Principal Cordillera as
472 observed in other orogens (e.g. Giese et al. 1999; Zhang et al. 2014). Also, because our
473 structure would drive Cuyanian material to the west, under the Chilenia terrane, it
474 would be difficult to explain the abnormally thick crust observed beneath the Central
475 Precordillera. According to Allmendinger & Judge (2014), this phenomenon could
476 rather be explained by an eastward material transference possibly driven by subduction
477 erosion. The presence of a double Moho beneath the Precordillera (Fig. 5) would
478 provide more constraints about such a process. To accommodate material transference
479 from Chilenia under the Cuyania terrane, we need an east-dipping shearing structure
480 that would connect the bottom of our aforementioned Ordovician structure to the double
481 Moho inferred from our local RF analysis (Fig. 5 & 6). Our hypothesis thus involves an
482 eclogite wedge migrating eastward from the Chilenia lower crust underneath the
483 Cuyania terrane accommodating shortening in the lower crust. Under compression, as
484 the Cuyania terrane remains rigid down to ~40 km depth, it would act like a “planer”,
485 cutting through the much softer Chilenia terrane. Below 40 km, both Cuyania and

486 Chilenia lower crust are likely turned into eclogites that would be driven to the east by
487 subduction erosion contributing to the crustal thickening observed beneath the Central
488 Precordillera (Fig. 8b).

489 **5. Conclusions**

490

491 In this study, we computed high-resolution local receiver functions from slab seismicity
492 in the Pampean flat slab region of Argentina. The RFs have been stacked to produce 2D
493 cross sections imaging the velocity structure and their lateral extent across the region of
494 study. Then, we jointly inverted phase velocity dispersion with our local RFs to
495 constrain high-resolution absolute shear-wave velocity structure and combined them
496 with our stacked RFs images. Our improved level of crustal details allowed us to better
497 constrain regional tectonic features such as décollement levels accommodating regional
498 crustal shortening as well as small structures related to very shallow changes in
499 lithology such as sedimentary basins (Tulum Valley, Calingasta Valley).

500 We suggest the presence of a major structure that separates the Cuyania and Chilenia
501 terranes. We geometrically constrained this structure using the crustal seismicity
502 recorded by several seismic experiments deployed in the region and focal mechanisms
503 from regional moment tensor inversion. This structure is well constrained down to ~40
504 km depth, dipping to the west and exhibiting reverse motion. Petrological analyses
505 show that the Precordillera recorded evidence of extension during the Late Ordovician
506 with spreading ridge magmatism. From Late Ordovician to Middle Devonian, this
507 narrow ocean was closed, probably by the tectonic inversion of one major extensional

508 structure leading to the collision of the Chilenia and Cuyania terranes. This structure
509 would correspond to the one observed in our stacked RFs images. The collision between
510 the two terranes likely led to crustal thickening. The Permian-Triassic extension in the
511 region considerably changed the geological configuration. At this time, Chilenia
512 recorded intense magmatism that contributed to its expansion and a major change of its
513 rock composition probably due to the creation of new granitic crust. This major
514 structure apparently played an important role in controlling the regional deformation
515 during the Permian-Triassic. To explain crustal shortening accommodation in the lower
516 crust, we propose the hypothesis of an eclogite wedge migrating from Chilenia beneath
517 Cuyania contributing to the complex Moho geometry and the abnormally high crustal
518 thickness generally observed beneath the Precordillera and western Sierras Pampeanas.

519 **6. Acknowledgments**

520

521 This work has been supported by the Ministerio de Ciencia, Tecnología e Innovación
522 Productiva of Argentina (PICT2011-160). We thank Ryan Porter for providing with the
523 ANT dispersion data used in this work. We are grateful to the Incorporated Research
524 Institutions for Seismology (IRIS) for the easy download of the data presented in this
525 study. The SIEMBRA project was supported by the National Science Foundation of the
526 USA (NSF, EAR 0510966 and 0907880). We also thank Suzanne M. Kay, Brian Horton
527 and Chelsea Mackaman-Lofland for constructive discussions about the geochemistry of
528 the Western Precordillera and suggestions of very useful references. Most of the figures
529 presented in this manuscript were made using the Generic Mapping Tools (GMT). The

530 joint inversion of RF and Surface wave dispersion has been performed using the
531 Computer Programs in Seismology (CPS) package, version 3.30 and the Seismic
532 Analysis Code (SAC). We particularly thank the editor, Peter Shearer as well as two
533 anonymous reviewers for their very helpful comments.

534 **7. References**

535

536 Allmendinger, R.W., Figueroa, D., Snyder, D., Beer, J., Mpodozis, C., Isacks, B.L., 1990.
537 Foreland shortening and crustal balancing in the Andes at 30°S latitude. *Tectonics*,
538 9, 789-809

539 Alvarado, P., Sánchez, G., Saez, M., Castro de Machuca, B., 2010, Nuevas evidencias de
540 la actividad sísmica del terreno Cuyania en la región de subducción de placa
541 horizontal de Argentina: *Rev. Mex. Cienc. Geol.*, 27(2), 278-291

542 Ammirati, J.-B., Alvarado, P., Perarnau, M., Saez, M., Monsalvo, G., 2013. Crustal
543 structure of the Central Precordillera of San Juan, Argentina (31°S) using
544 teleseismic receiver functions, *J. South Am. Earth Sci.*, 46, 100-109,
545 doi:10.1016/j.jsames.2013.05.007

546 Ammirati, J.-B., Alvarado, P., Beck, S., 2015. A lithospheric velocity model for the flat
547 slab region of Argentina from joint inversion of Rayleigh wave phase velocity
548 dispersion and teleseismic receiver functions, *Geophys. J. Int.*, 202, 224–241, doi:
549 10.1093/gji/ggv140

550 Anderson, M., Alvarado, P., Zandt, P., Beck, S., 2007. Geometry and brittle
 551 deformation of the subducting Nazca Plate, Central Chile and Argentina,
 552 Geophys. J. Int., 171, 419-434, doi: 10.1111/j.1365-246X.2007.03483.x

553 Baldis, B., Beresi, M., Bordonaro, O., Vaca, A., 1982. Síntesis evolutiva de la Precordillera
 554 Argentina, 5º Congres. Latinoam. Geol., Actas 4, 399-445, Buenos Aires,
 555 Argentina.

556 Boedo, F.L., Vujovich, G.I., Kay, S.M., Ariza, J.P., Pérez Luján S.B., 2013. The E-MORB
 557 like geochemical features of the Early Paleozoic mafic-ultramafic belt of the
 558 Cuyania terrane, Western Argentina, J. South Am. Earth Sci., 48, 73-84.

559 Brocher, T.M., 2005. Empirical relations between elastic wavespeeds and density in the
 560 Earth's crust, Bull. Seism. Soc. Am., 95, 2081-2092, doi: 10.1785/0120050077

561 Cahill, T., Isacks, B.L., 1992. Seismicity and the shape of the subducted Nazca plate, J.
 562 Geophys. Res., 97, 17503-17529, doi: 10.1029/92JB00493

563 Calkins, J., Zandt, G., Gilbert, H., Beck, S., 2006. Crustal images from San Juan,
 564 Argentina, obtained using high frequency local event receiver functions, Geophys.
 565 Res. Lett., 33, 1-4, doi: 10.1029/2005GL025516

566 Castro de Machuca, B., Perarnau, M., Alvarado, P., López, G., Saez, M., 2012. A
 567 seismological and petrological crustal model for the southwest of the Sierra de Pie
 568 de Palo, Province of San Juan, Rev. Asoc. Geol. Argent., 69, 179-186.

569 Christensen, N.I., Mooney W.D., 1995. Seismic velocity structure and composition of the
 570 continental crust: A global view, J. Geophys. Res., 100, 9761-9788, doi:
 571 10.1029/95JB00259

572 Cristallini, E.O., Ramos, V.A., 2000. Thick-skinned and thin-skinned thrusting in the La
 573 Ramada fold and thrust belt: crustal evolution of the high Andes of San Juan,
 574 Argentina (32°S), Tectonophysics, 317, 205-235, doi:10.1016/S0040-
 575 1951(99)00276-0

576 Davis, J.S., Roeske, S.M., McClelland, W.C., Snee, L.W., 1999. Closing the ocean
 577 between the Precordillera terrane and Chilenia: Early Devonian ophiolite
 578 emplacement and deformation in the southwest Precordillera, in Ramos V.A.
 579 and Keppie, J.D., eds., Laurentia Gondwana connections before Pangea: Geol.
 580 Soc. Am. Special paper 336, 115-138, Boulder, Colorado, USA.

581 Dueker, K.G., Sheehan A.F., 1997. Mantle discontinuity structure from midpoint stacks
 582 of converted P to S waves across the Yellowstone hotspot track, J. Geoph. Res.,
 583 102, 8313-8327

584 Ebinger, C.J., Casey, M., 2001. Continental breakup in magmatic provinces: An
 585 Ethiopian example, Geology 29(6), 527-530, doi:
 586 10.1130/00917613(2001)029<0527:CBIMPA>2.0.CO;2

587 Gans, C.R., Beck, S.L., Zandt, G., Gilbert, H., Alvarado, P., Anderson, M., Linkimer, L.,
 588 2011. Continental and oceanic crustal structure of the Pampean flat slab region,
 589 western Argentina, using receiver function analysis: new high-resolution results,
 590 Geophys. J. Int., 186, 45-58, doi: 10.1111/j.1365-246X.2011.05023.x

591 Giambiagi, L.B., Ramos, V.A., 2002. Structural evolution of the Andes in a transitional
 592 zone between flat and normal subduction (33°30'-33°45'S), Argentina and Chile,
 593 J. South Am. Earth Sci., 15, 101-116

594 Giese, P., Scheuber, E., Schilling, F., Schmidz, M., Wigger, P., 1999. Crustal thickening
 595 processes in the Central Andes and the different natures of the Moho-
 596 discontinuity, J. South Am. Earth Sci., 12, 201-220, doi:10.1016/S0895-
 597 9811(99)00014-0

598 Gilbert, H., Beck, S., Zandt, G., 2006. Lithospheric and upper mantle structure of central
 599 Chile and Argentina, Geophys. J. Int., 165, 383-398, doi: 10.1111/j.1365-
 600 246X.2006.02867.x.

601 Gurrola, H., Minster, J.B., Owens, T., 1994. The use of velocity spectrum for stacking
 602 receiver functions and imaging upper mantle discontinuities, Geoph. J. Int., 117,
 603 427-440

604 Hacker, B.R., 1996. Eclogite formation and the Rheology, Buoyancy, Seismicity, and H₂O
 605 content of Oceanic Crust. In: Geophysical Monograph, vol. 96, American
 606 Geophysical Union, pp. 337-346, Washington D. C., USA.
 607 <http://dx.doi.org/10.1029/GM096p0337>

608 Heredia, N., Rodríguez Fernández, L.R., Gallastegui, G., Busquets, P., Colombo, F.
 609 2002. Geological setting of the Argentine Frontal Cordillera in the flat-slab
 610 segment (30°00' - 31°30'S latitude), J. South Am. Earth Sci. 15, 79–99

611 Herrmann, R. B., Ammon, C.J., 2002. Computer programs in seismology – 3.30: surface
 612 waves, receiver functions and crustal structure,
 613 www.eas.slu.edu/People/RBHerrmann/CPS330.html

614 Julià, J., Ammon, C.J., Herrmann, R.B., Correig, A.M., 2000. Joint inversion of receiver
 615 function and surface wave dispersion observations, Geophys. J. Int., 143, 99-112

616 Kay, S.M., Ramos, V.A., Kay, R.W., 1984. Elementos mayoritarios y trazas de las
 617 vulcanitas ordovícicas de la Precordillera Occidental: Basaltos de rift oceánico
 618 temprano (?) próximos al margen continental, IX Congreso Geológico Argentino,
 619 S.C. de Bariloche, Rio Negro, Argentina, Actas II, 48-65

620 Kay, S., Abbruzzi, J., 1996. Magmatic evidence for Neogene lithospheric evolution of the
 621 Central Andean flat-slab between 30 and 32°S, Tectonophysics, 259, 15-28,
 622 doi:10.1016/0040-1951(96)00032-7

623 Kay, S.M., Orrell, S. Abbruzzi, J.M., 1996. Zircon and whole rock Ns-Pb isotopic evidence
 624 for a Greenville age and Laurentia origin for the basement of the Precordilleran
 625 Terrane in Argentina, Geol. Soc. Am. South-Central Sec. Abs. Prog., 28(1), 21-22

626 Langston, C.A., 1979. Structure under Mount Rainier, Washington, inferred from
 627 teleseismic body waves, J. Geophys. Res., 84, 4749-4762

628 Ligorria. J. and Ammon, C.J., 1999. Iterative deconvolution and receiver function
 629 estimation, Bull. Seism. Soc. Am., 89, 1395-1400

630 Marot, M., Monfret, T., Pardo, M., Ranalli, G., et Nolet, G., 2013. A double seismic
 631 zone in the subducting Juan Fernández Ridge of the Nazca Plate (32°S), central
 632 Chile, J. Geophys. Res., doi: 10.1002/jgrb.50240

633 Marot, M., Monfret, T., Gerbault, M., Nolet, G., Ranalli, G., Pardo, M., 2014. Flat
 634 versus normal subduction zones: a comparison based on 3-D regional traveltimes
 635 tomography and petrological modelling of central Chile and western Argentina
 636 (29°-35°S), Geophys. J. Int., 199, 1633-1654, doi: 10.1093/gji/ggu355

637 Mulcahy, P., Chen, C., Kay, S.M., Brown, L.D., Isacks, B.L., Sandvol, E., Heit, B., Yuan,
 638 X., Coira, B.L., 2014. Central Andean mantle and crustal seismicity beneath the
 639 Southern Puna plateau and the northern margin of the Chilean-Pampean flat slab,
 640 Tectonics, 33 (8), 1636-1658, doi:10.1002/2013TC003393

641 Perarnau, M., Alvarado, P., Saez, M., 2010. Estimación de la estructura cortical de
 642 velocidades sísmicas en el suroeste de la Sierra de Pie de Palo, Provincia de San
 643 Juan, Rev. Asoc. Geol. Argent., 64, 473-480

644 Perarnau, M., Gilbert, H., Alvarado, P., Martino, R., Anderson, M., 2012. Crustal
 645 structure of the Eastern Sierras Pampeanas of Argentina using high frequency
 646 local receiver functions, Tectonophysics, 580, 208-217,
 647 doi:10.1016/j.tecto.2012.09.021

648 Pérez Luján, S.B., Ammirati, J.-B., Alvarado, P., Vujovich, G.I., 2015. Constraining a
 649 mafic thick crust model in the Andean Precordillera of the Pampean flat slab
 650 subduction region, J. South Am. Earth Sci., 64, 325-338, doi:
 651 10.1016/j.jsames.2015.09.005

652 Porter, R., Gilbert, H., Zandt, G., Beck, S., Warren, L., Calkins, J., Alvarado, P.,
 653 Anderson, M., 2012. Shear wave velocities in the Pampean flat-slab region from
 654 Rayleigh wave tomography: implications for slab and upper mantle dehydration,
 655 J. Geophys. Res., 117, doi:10.1029/2012JB009350

656 Ragona, D., Anselmi, G., González, P., Vujovich, G., 1995. Mapa geológico de la
 657 provincia de San Juan, República Argentina, Secretaría de Minería, Dirección
 658 Nacional del Servicio Geológico, Buenos Aires, Argentina

659 Ramos, V.A., Jordan, T.E., Allmendinger, R.W., Kay, S.M., Cortes, J.M., Palma, M.A.,
 660 1984. Chilenia: Un terreno alóctono en la evolución paleozoica de los Andes
 661 Centrales, IX Congreso Geológico Argentino, S.C. de Bariloche, Rio Negro,
 662 Argentina, Actas II, 84-106

663 Ramos, V.A., Cegarra, M., Cristallini, E., 1996. Cenozoic tectonics of the hHigh Andes
 664 of west-central Argentina (30-36°S latitude), Tectonophysics, 119, 185-200

665 Ramos, V.A., Cristallini, E.O., Pérez, D.J., 2002. The Pampean flat-slab of the central
 666 Andes, J. South Am. Earth Sci. 15, 59-78

667 Regnier, M., Chiu, J.-M., Smalley, R., Isacks, B.L., Araujo, M., 1994. Crustal thickness
 668 variation in the Andean foreland, Argentina, from converted waves, Bull. Seism. Soc.
 669 Am., 84(4), 1097-1111

670 Rocher, S., Vallecillo, G.M., Castro de Machuca, B., Alasino, P.H., 2015. El Grupo Choiyoi
 671 (Pérmico temprano-medio) en la Cordillera Frontal de Calingasta, San Juan,
 672 Argentina: Volcanismo de arco asociado a extensión, Rev. Mex. Cienc. Geol., 32(3),
 673 415-432

674 Stauder, W., 1973. Mechanism and spatial distribution of Chilean earthquakes with
 675 relation to subduction of the oceanic plate, J. Geophys. Res., 78, 5033-5061, doi:
 676 10.1029/JB078i023p05033

677 Strazzere, L., Gregori, D.A., Dristas, J.A., 2006. Genetic evolution of Permo-Triassic
 678 volcanoclastic sequences at Uspallata, Mendoza Precordillera, Argentina,
 679 Gondwana Research 9, 485–499, doi:10.1016/j.gr.2005.12.003

680 Venerdini, A., Sánchez, G., Alvarado, P., Bilbao, I., Ammirati, J.-B. 2016, Nuevas
681 determinaciones de velocidades de ondas P y ondas S para la corteza sísmica del
682 terreno Cuyania en el retroarco andino, *Rev. Mex. Cienc. Geol.*, 33(1), 59-71

683 Wagner, L.S., Beck, S., Zandt, G., 2005. Upper mantle structure in the south central
684 Chilean subduction zone (30° to 36°S), *J. Geophys. Res.*, 110, doi:
685 10.1029/2004JB003238

686 Yang, X., Deloule, E., Xia, Q., Fan, Q., Feng, M., 2008. Water contrast between
687 Precambrian and Phanerozoic continental lower crust in eastern China. *J. Geophys.*
688 *Res.* 113, B08207

689 Yáñez, G., Cembrano, J., Pardo, M., Ranero, C., Selles, D., 2002. The Challenger-Juan
690 Fernández – Maipo major tectonic transition of the Nazca-Andean subduction
691 system at 33-34°S: geodynamic evidence and implications, *J. South Am. Earth Sci.*,
692 15, 23-38

693 Zangh, Z., Wang, Y., Houseman, G.A., Xu, T., Wu, Z., Yuan, X., Chen, Y., Tian, X., Bai,
694 Z., Teng, J., 2014. The Moho beneath western Tibet: Shear zones and eclogitization
695 in the lower crust. *Earth and Planet. Sci. Lett.*, 408, 370-377,
696 doi:10.1016/j.epsl.2014.10.022

697

702 **Figure 1:** Location map centered on the Pampean flat slab region of Argentina. The thick
703 black lines mark terrane boundaries (Ramos et al., 2002). Slab top contours are shown by
704 thin black lines (Anderson et al. 2007) and thin black dashed lines (Mulcahy et al. 2014).
705 The white diamonds mark the location of the SIEMBRA seismic stations. The upside-
706 down green triangles mark the location of the CHARGE stations and the magenta triangles
707 mark the location of the ESP stations. The red line shows the path of the JFR under the
708 South American plate (Yáñez et al. 2002). SPP stands for Sierra de Pie de Palo in the
709 western Sierras Pampeanas. The square shows the emplacement of Figure 2.

710

712 **Figure 2:** Main geological formations characterizing the Cuyania-Chilenia boundary
713 simplified from Ragona et al. (1995). The white diamonds mark the location of the
714 SIEMBRA stations used to compute the local RFs presented in this work. The thick grey
715 dashed line marks the approximate boundary between the Cuyania terrane to the east and
716 the Chilenia terrane to the west. This boundary is generally associated with the presence of
717 a belt of mafic-ultramafic rocks that outcrops in the Western and Central PreCORDILLERA. The
718 thin grey dashed line delimitates the Frontal Cordillera from the Principal Cordillera.
719 Letters A and D refer to the pictures shown in Figure 7a & d.

720

721

722

723

724 **Figure 3: A)** NEIC-USGS slab seismicity ($M_L > 3.0$) recorded by the SIEMBRA array
725 (white diamonds). The black lines represent the slab contours (Anderson et al. 2007). Note
726 the regional seismicity is concentrated in the flat portion of the slab. **B)** The colored circles
727 represent the piercing points corresponding to local RFs computed in this work. Red, blue
728 and green correspond to the A-B, C-D and E-F cross sections, respectively. The asterisks
729 mark bin centers for the common conversion point (CCP) stacks. The thick black lines
730 represent the top of the slab contours (Anderson et al 2007). The black lines indicate the
731 cross sections shown in Figure 5.

732

733

734

735

736

737 **Figure 4: A)** Comparison between local (left, 26 traces) and teleseismic (right, 24 traces)
738 receiver function recorded at station DOCA (Central Precordillera). The traces were
739 stacked after moveout correction and time to depth conversion. Local data have been
740 corrected from moveout using the 1D regional velocity model obtained by Ammirati et al.
741 (2015). Teleseismic data are from Ammirati et al. (2013). **B)** Cross correlation matrix.
742 Traces are correlated with one another and return a correlation coefficient value. Pairs of
743 traces with good correlation (correlation coefficient > 0.8) are selected. The ensemble of
744 good pairs allows us to identify a waving pattern related to crustal velocity structures and
745 thus differentiate usable traces from noisy and/or poor quality RFs.

746

747

748

749

750

751 **Figure 5:** Cross sections showing our stacked RFs. See Figure 3 for location of the cross
752 sections. The CCP stacked receiver functions have been projected along the corresponding
753 lines. The time to depth conversion has been performed using the calibrated regional
754 velocity model obtained in Ammirati et al. (2015). Positive amplitudes are in red, negative
755 amplitudes are in blue. Red triangles are the SIEMBRA stations for which we computed
756 the local RFs. FC: Frontal Cordillera; WP: Western Precordillera; CP: Central
757 Precordillera; EP: Eastern Precordillera; WSP: Western Sierras Pampeanas. Note the
758 structure change between the Cuyania and Chilenia terranes indicated by the dark grey
759 line. Our interpretations for the Moho signal correspond to the grey dashed lines. The black
760 dashed line marks a possible shear zone accommodating crustal shortening between the
761 Cuyania and Chilenia terranes in the lower crust.

762

763

764

765

766 **Figure 6: A)** Joint inversion model for the C-D cross section (see Fig. 3b for location).
767 Low velocities appear in red tones, high velocities appear in blue tones. Contour spacing is
768 0.1 km/s (see main text for details of the joint inversion process). **B)** Same as (A) overlaid
769 with the CCP stack discontinuities shown in Fig. 5b. Note that only positive arrivals
770 interpreted as discontinuities (in the main text) have been reported. **C)** Same as (B)
771 overlaid by regional crustal seismicity (white dots) recorded by the OVA99, CHARAME,
772 CHARGE and SIEMBRA experiments (Linkimer 2011, Marot et al. 2013). Focal
773 mechanisms are from Ammirati et al. (2015). Clusters of seismic activity related with
774 regional tectonics as discussed in the main text, have been circled. Red triangles represent
775 the seismic stations used for both CCP stacking and joint inversion.

776

777

Figure 7: A series of pictures from the Central Precordillera illustrating the mafic-ultramafic belt at the transition zone between the Chilenia and Cuyania terranes (Fig. 2).

A) Thick diorite and wehrlite sills interlayered in metagraywacke layers. **B)** Diorite sample in hand. Primary minerals (mainly hornblende and plagioclase) can be easily recognized due to their coarse granular texture. **C)** Partially altered diorite sample under cross-polarized light showing the hypidiomorphic texture. Note the presence of primary (amphibole, plagioclase, clinopyroxene) and secondary minerals (such as chlorite) minerals. **D)** Gabbro sill in tectonic contact with metapelites layers. **E)** Massive gabbro with medium granular texture. **F)** Same sample as (E) in thin section under cross-polarized light. Note the presence of well-preserved primary minerals such as clinopyroxene and

790 plagioclase showing ophitic and subophitic texture. **G)** Coarse granular wehrlite rock
791 sample (Same as in A). **H)** Wehrlite sample in thin section under cross-polarized light
792 showing alotriomorphic texture. Olivine is mostly altered to serpentine. Pl: plagioclase;
793 Kfs: K-feldspar, Amp: amphibole, Ol: olivine; Cpx: clinopyroxene; Chl: chlorite; Srp:
794 serpentine.

795

796

797

798

799 **Figure 8: A)** Western margin of Gondwana during the Permian-Triassic in the region of
800 study. Choiyoi volcanism probably changed the composition of the Chilenia terrane and
801 contributed to its expansion. Extensional stresses due to the break-up of Gondwana led to
802 normal faulting. The master structure suggested by the geophysical analyses in this work is
803 a good candidate to accommodate the extension between Cuyania and Chilenia also
804 controlling the deposition of the Choiyoi Group on the Chilenia side. **B)** Present day
805 western margin of South America at ~31-32°S. As a result of 20 Ma of Andean
806 compression, previous structures may have been reactivated as reverse faults.
807 Simultaneously, the region recorded slab flattening coeval with the uplift of the Eastern
808 Precordillera and Sierras Pampeanas (more to the east). PC: Principal Cordillera, FC:
809 Frontal Cordillera, CV: Calingasta valley, WP: Western Precordillera, CP: Central
810 Precordillera, EP: Eastern Precordillera. Circled letters refer to the lithology described in
811 Figure 7.

812