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1 **Structure of the crust and the lithosphere beneath the southern Puna plateau from**
2 **teleseismic receiver functions**

3

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17

18 **ABSTRACT**

19 We present a teleseismic P and S receiver function study using data from a temporary passive-
20 source seismic array in the southernmost Puna Plateau and adjacent regions. The P receiver
21 function images show the distribution of crustal thickness and Vp/Vs ratio for this area. Over much
22 of the southern Puna plateau, the crustal thickness is 50-55 km, whereas to the west a thicker
23 crust (~60 to 75 km) is observed beneath much of the Andean volcanic arc region. From the Puna

24 southward, there is little obvious change in the crustal thickness across the border of the plateau
25 (south of 28°S). The crust is seen to progressively thin towards the east in the Pampean Ranges
26 where it is 35-40 km thick. The southern Puna plateau is characterized overall by a low crustal
27 Vp/Vs ratio (less than 1.70), implying a felsic crustal composition. An anomalously high Vp/Vs ratio
28 of 1.87 is observed beneath the Cerro Galan volcanic center, in the region where a prominent
29 crustal low-velocity zone identified below ~10 km depth probably extends into the lower crust. The
30 crustal thickness determined under the Cerro Galan area (59 km) is close to that of the rest of the
31 southern Puna. The prominent high Vp/Vs ratio and low velocity zone beneath the Galan region
32 implies the presence of a zone of partial melt or a magma chamber is consistent with hypothesis
33 calling for lithospheric delamination beneath the Galan caldera. A widespread crustal low-velocity
34 layer observed beneath much of the Southern Puna, correlates well with crustal low-velocity
35 anomalies observed by teleseismic tomography. The lithosphere-asthenosphere boundary beneath
36 the array can be clearly observed by both P and S receiver functions at depths of 70-90 km in
37 agreement with previous studies suggesting a thin lithosphere beneath the high elevated plateau.
38 The mantle transition zone discontinuities appear at expected depths.

39 *Keywords: Central Andes, Puna plateau, teleseismic receiver functions, crustal thickness,*
40 *lithosphere-asthenosphere boundary.*

41

42 **1. Introduction**

43

44 The Altiplano-Puna region (marked as Puna in Fig. 1) in the Central Andes is a high elevated
45 plateau that was formed in a subduction regime, which is flanked to the north and south by
46 segments of flat subduction. The inherited plateau geological history, uplift, shortening and
47 magmatism have been studied over the last decades (e.g., Cahill and Isacks, 1992; Allmendinger
48 et al., 1997; Oncken et al. 2003; Kay and Coira, 2009 and references therein) and distinct
49 differences have been recognized between the Altiplano to the north and the Puna to the south. In

50 this paper we focus on the crustal and lithospheric structure beneath the southern end of the Puna
51 plateau, until recently a geophysically poorly studied area. In a new study, Bianchi et al. (2013)
52 gave an account of available geophysical information on the southern Puna Plateau based on a
53 number of studies, which began with the pioneering Magnetotelluric experiments of Febrer et al.
54 (1982). To the south of 30°S, a number of seismic studies provide clues to the Moho, the
55 lithosphere and the seismicity in the area located above the Chilean flat-slab region (e.g. Fromm et
56 al., 2004; Alvarado et al., 2005 and 2007; Gilbert et al., 2007; Gans et al., 2011; Porter et. al.,
57 2012, Ward et al., 2013).

58

59 The main features distinguishing the southern Puna plateau from the northern Puna and the
60 Altiplano are: (a) a thinner continental lithosphere (Whitman et al. 1992; Heit et al., 2007), (b) the
61 subducting Nazca plate begins to shallow to the south (e.g., Cahill and Isacks 1992; Mulcahy et al.,
62 2013) and (c) a distinctive sedimentary, magmatic and structural history (e.g. Coira et al. 1993; Kay
63 et al., 1994, 1999; Allmendinger et al., 1997; Kay and Coira, 2009). Tectonically, the southern Puna
64 has experienced less shortening than the Altiplano and the deformation is mainly concentrated in
65 the eastern border of the plateau in the Eastern Cordillera and the Pampean ranges (e.g., Kley and
66 Monaldi, 1998; Kley et al., 1999). The large ignimbritic deposits in the area of the Cerro Galan
67 caldera are suggested to be a consequence of piecemeal delamination of lithospheric material in
68 the southern Puna (Kay and Kay, 1993; Kay et al. 1994, 2011).

69

70 Previous geophysical studies revealed low velocities in the crust and asthenosphere beneath the
71 southern Puna plateau, suggesting higher temperatures than under the northern Puna and the
72 Altiplano (e.g. Isacks, 1988; Whitman et al., 1992, 1996; Heit, 2005; Heit et al., 2007, 2008;
73 Woelbern et al., 2009; Bianchi et al. 2013). The maximum crustal thickness along a profile at
74 25.5°S latitude obtained from a receiver function image, was nearly 60 km and the tomographic
75 images show low-velocity anomalies in the crust that lessen into the mantle, consistent with

76 asthenospheric material just below the Moho (Heit, 2005; Woelbern et al., 2009).

77

78 Here we apply the receiver function method to the teleseismic data recorded by the southern Puna
79 passive source seismic array, consisting of 74 stations (Fig. 1). In the analysis, we examine crustal
80 thicknesses and intracrustal layers with P receiver functions (PRF) and the lithosphere-
81 asthenosphere boundary with S- receiver functions (SRF).

82

83 **2. Data and methodology**

84

85 The seismic stations of the southern Puna array (Fig. 1) were distributed in two orthogonal profiles
86 with a center at approximately 26.5°S and 67.5°W, which were surrounded by a sparser 2D array
87 covering an area of approximately 300x300 km. For details concerning the station spacing and
88 instrumentation see Bianchi et al. (2013). The data are archived at IRIS and GEOFON data
89 centers.

90

91 We performed the teleseismic PRF receiver function analyses using events from the PDE catalog
92 for which teleseismic P and PP phases were recorded at epicentral distances between 30° and
93 180° and for which magnitudes (mb) were greater than 5.5. For the SRF receiver function
94 (including S and SKS phases), the epicentral distances were between 60° and 115° and
95 magnitudes were greater than 5.5. The events used are shown in Fig. 2. The PRF computation
96 was performed following the approach described by Yuan et al. (1997), whereas the SRF
97 computation was performed using the approach by Kumar et al. (2006) and Yuan et al. (2006).

98

99 Seismograms with high signal/noise ratio have been visually inspected and manually selected for

100 calculation of the P and S receiver functions. They were rotated, deconvolved and move out
101 corrected for a constant reference slowness of 6.4 s/deg. The PRF were then divided into boxes
102 according to piercing point locations at 60 km depth and stacked inside the boxed regions shown in
103 Figures 3 and 4 to enhance the signal-to-noise ratios of single traces (Dueker and Sheehan, 1997).
104 We maximized the signal-to-noise ratios and homogenized the resolution based on the distribution
105 of the stations in the array. We also enhanced coherent signals, like the Moho phase, at the same
106 time that we suppressed random noise effects by stacking single traces from different events
107 around stations in 0.5 x 0.5 degrees boxes. Along the north-south and east-west profiles where the
108 station spacing is about 10 km (Fig. 3), the boxes overlap by 0.1 degrees. For the stations
109 surrounding the two orthogonal profiles, similar sized boxes (i.e. 0.5 x 0.5 degree) do not overlap
110 due to the larger station spacing (i.e. 50 km to 100 km; Fig. 4). In most cases, the Moho converted
111 phase can be clearly recognized providing reliable estimates of the Moho depth.

112

113 Using the information provided by different crustal reverberations also known as Moho multiples,
114 we estimated the Moho depth and the crustal Vp/Vs ratio following the H-K stacking approach (Zhu
115 & Kanamori, 2000). The H-K stacks are done by boxes along the profiles north-south and west-
116 east and by station piercing points elsewhere in the array since there are no overlaps where the
117 station spacing is big. Some examples are presented in Fig. 5. Within each group a grid search is
118 performed for estimates of Moho depth and crustal Vp/Vs ratio as indicated by the position of the
119 maximum. The resulting Vp/Vs map shown in Fig. 6 illustrates the Vp/Vs variations in the region.
120 Using the Moho depth values obtained by the H-K stacking method, we obtained a Moho depth
121 map for the region (Fig. 7).

122

123 **3. Results**

124 **3.1 The Moho from P receiver functions**

125

126 We present the PRF results in order to discuss the time arrivals of the Moho phase across the
127 study area. As the station spacing is very close along the north-south and west-east profiles, we
128 choose to stack the PRF along these lines to get an idea of the Moho time arrival variations across
129 and along the strike of the subduction. In Fig. 3, we present the stacked traces per box for each
130 profile that represents the Moho arrival times beneath the boxes along each profile. Along the
131 west-east profile (26.8° S) the Moho phase (marked as M) is not clear at Box 1 but is well
132 recognized in Box 2 at 7.9 s. If we consider this to be the Moho beneath the volcanic arc then we
133 could also pick the small positive phase at 8.2 s to be the correct Moho phase in Box 1. The
134 situation becomes more complicated for Box 3 at the position of the volcanic arc ($\sim 68^{\circ}$ W) where
135 we detect the most prominent phase at 4.4 s not marked by the blue Moho line. This is an
136 unexpected arrival time for the Moho-like phase as the traces from nearby boxes show clear later
137 arrivals of the Moho, suggesting the phase at 4.4 s might be generated by a local feature. We
138 suspect this phase to be an intra-crustal phase in an area where the Moho is invisible for receiver
139 functions as we discuss below. In Box 4, the Moho phase is detected at 5.9 s while it arrives later
140 at 6.5 s in Box 5 and then towards the east it arrives earlier where it can be seen at 6.3 s (Box 6),
141 6.1 s (Box 7), 5.7 s (Box 8) and finally 5.2 s in Box 9. The distribution of negative phases in the
142 crust is well represented west of 66° W between Boxes 1-7 where they are found to arrive earlier
143 beneath the volcanic arc at 2.5 s and at 3.2 s in the center of the plateau. The slab can be traced
144 as a positive phase on the western side at 10.8 s (Box 1-2) and 11.6 s (Box 3). Eastward from Box
145 4, the slab should be around 30 s for Box 9 but it is for us impossible to determine its arrival time
146 with our data. The earthquakes plotted in this figure (300 km swath from the line of every profile i.e
147 150 km to the north and 150 km to the south of the line) are located by Mulcahy et al. (2013) using
148 the same network and have been used to improve the trace of the slab originally defined by Cahill
149 and Isacks et al. (1992). We choose to plot a wide band of earthquakes to avoid assumptions of
150 the position of the slab due to the scarcity of the slab seismicity. By plotting more events from the
151 slab we expect the slab feature to become more stable (for more details about the slab geometry

152 see Mulcahy et al., 2013).

153

154 Along the South-North profile (67.8°W-67.2°W), it is possible to identify the Moho phases marked
155 as M in Fig. 3. The Moho phase is identified at 6.7 s in Box 1 and arrives earlier at Box 2 (5.7 s).
156 The Moho phase is then interpreted to arrive later in Box 3 (6.4 s) and it arrives progressively
157 earlier towards the center of the profile at Boxes 4 and 5 (both at 5.9 s). At the northernmost end of
158 the profile, the Moho phase arrival is identified at 7.0 s. In the crust, strong positive phases can be
159 observed close to 0s (followed by a negative phase) in Boxes 1 to 3. The continuity of these intra-
160 crustal phases is then interrupted to the south where a break could be related to the tectonic
161 border between the Puna to the north and the Pampean Block to the south in the region of the
162 Fiambala basin as discussed later. To the north of Box 3, the presence of negative phases without
163 any clear correlated positive phase close to 0 s is consistent with the distribution of low velocity
164 anomalies described in the crust by Bianchi et al. (2013) using a tomographic approach. The
165 oceanic slab is not recognized along the profile in the area known as the Antofalla intermediate
166 depth seismic gap in the subducted plate where sparsely distributed earthquakes (Cahill and
167 Isacks, 1992; Mulcahy et al., 2013) suggest a slab position at around 170 km depth in the north
168 and 130 km depth in the south where the slab begins to shallow (110 km) into the flatslab region at
169 29°S. The approximate position of the slab is marked based on the data of Mulcahy et al. (2013).

170

171 The results for the stations away from the W-E and S-N profiles are presented in Fig. 4, where the
172 Moho phases are picked in different sections from North (Profile 1) to South (Profile 6). As in Fig. 3
173 the blue line marks the position of the Moho phase. As with most stations under the Puna in Fig. 3,
174 the Moho phases arrives earlier than the Moho converted phases to the west beneath the volcanic
175 arc. Beneath the Cerro Galan station we obtained a late arrival of the Moho phase at 8.7 s (Fig. 4,
176 Profile 2 at 67 °W). The Moho conversion times are observed earlier towards the eastern border of
177 the plateau beneath the Sierras Pampeanas. The slab seismicity from Mulcahy et al. (2013) helps

178 to infer the changing position of the subducted oceanic crust along the different profiles.

179

180 We use the Moho and its multiples to estimate the depth of the Moho and the crustal Vp/Vs ratio
181 (Zhu & Kanamori, 2000). In Fig. 5, some examples of the H-K stacks of receiver functions are
182 displayed. The boxes presented in Fig. 5 belong to profile that crosses the Puna Plateau from west
183 to east with a high number of stations with overlapping piercing points as can be seen in Fig. 3.
184 Within each group, a grid search is performed for estimates of Moho depth and crustal Vp/Vs ratio
185 which is indicated by the position of the maximum amplitude. In each example we also show the
186 stacked receiver functions moveout corrected for the Ps and the multiple phases (PpPs and PpSs),
187 respectively. A good practice is that the phases are enhanced after the correct moveout correction
188 and agree with the arrival times predicted by the final model. For some boxes the H-K stack does
189 not give a reliable measurement, which is mostly due to the Moho multiples being too weak. In this
190 case, we estimate the Moho depth by the arrival time of the Moho Ps phase using an average
191 Vp/Vs ratio for the entire study area (1.74). The overall Vp/Vs map displayed in Fig. 6 shows a
192 Vp/Vs ratio smaller than 1.7 beneath most of the southern Puna, whereas the region around Galan
193 has a high value (1.87).

194

195 Fig. 7 displays the resultant Moho depth map where it is possible to observe a deep Moho in the
196 northwest between 69°W and 68°W where the volcanic arc has a crustal thickness of 70 km. The
197 Moho beneath the Puna plateau is seen to be relatively flat with depths varying between 50 km
198 and 60 km. The Moho is shallower outside of the plateau where it can be observed between 35 km
199 and 45 km in the Sierras Pampeanas (Pampean Ranges) to the south and east.

200

201 **3.2 The crust beneath the Cerro Galan caldera**

202

203 At the Cerro Galan station we obtained a late arrival of the Moho phase of 8.7 s (Fig. 4, Profile 2 at
204 67 °W). This is an unexpected value (corresponding to a depth of 75 km using a V_p/V_s ratio of
205 1.74), because the Moho is relatively flat at a depth of ~55 km beneath all nearby stations.

206

207 In Figure 8, data from the station located inside the Galan caldera is presented for both P and S
208 receiver function results. Due to bigger station separation little overlapping between data from
209 neighboring stations is observed. In order to evaluate the results for the anomalously late Moho
210 arrival beneath the Galan caldera, we present the individual PRF in Fig. 8a and compare them with
211 the SRF (Fig. 8b). The Moho can be clearly identified in the PRF stack at 8.7 s, as well as in the
212 individual traces. Although the quality and number of traces of the SRF do not compare to those of
213 the PRF, the Moho phase clearly shows up in the SRF stack at 8.0 s, close to the Moho arrival in
214 the PRF.

215

216 A strong negative phase can be identified in both datasets, arriving at 2.5 s in the PRF and at 3.5 s
217 in the SRF. This is an indication of a pronounced low-velocity zone (LVZ) in the crust, that might
218 account for the late arrivals of the Moho phases. Although the positions of the Moho and LVZ
219 phases are similar in both methods, there are small differences in the arrival times, which can be
220 explained by different piercing point locations of the PRF and SRF and lateral variations in the
221 Moho depth and crustal heterogeneity, as well as different frequency contents of the P-to-S and S-
222 to-P converted waves.

223

224 We performed waveform modeling of the P receiver functions for the Cerro Galan station (Fig. 9)
225 by a linear inversion approach (Kind et al., 1995). It is well known that receiver function inversion is
226 highly non-unique (Ammon, 1990). The final model obtained by a linear inversion tends to follow
227 the start model, therefore, diverse nonlinear inversion approaches have been proposed (e.g.,

228 Sambridge, 1999). However, none of them are satisfied, due to the fact that the high-frequency
229 receiver functions are sensitive to the seismic discontinuities, but do not constrain the absolute
230 crustal velocity. Our receiver function modeling for the Cerro Galan station serves as a model
231 check, rather than a unique inversion. We used fine layers with a layer thickness of 1 km and a
232 strong smoothness control to avoid unrealistic velocity oscillation between neighboring layers. The
233 waveform fitting is mainly applied for the crustal velocities, whereas a large damping factor is given
234 to the uppermost mantle. A perfect waveform fitting could thus be hampered by the smoothness
235 regularization. A strong crustal low velocity zone is needed in the S wave velocity model in order to
236 fit the strong negative phase detected at ~2.5 s in the receiver function. In the model in Fig. 9, the
237 Moho required to match the stacked P receiver function and the synthetic curve needs to be
238 located at 55-60 km depth. By performing the H-K analysis, we obtain a Moho depth of 59 km and
239 a crustal V_p/V_s ratio of 1.87 beneath the station. This implies that the Cerro Galan region has a
240 crustal thickness in agreement with the thicknesses obtained across the southern Puna plateau
241 and overlies a zone of melts in the crust that is seen in our data as a prominent low-velocity zone.

242

243 In the model in Fig. 9, the low-velocity zone begins at a depth of ~10 km with a sharp velocity
244 reduction and a further velocity reduction occurs at a depth of ~20 km. Compared to the global
245 reference IASP91 model the shear wave velocity remains low in almost the entire crust below 10
246 km depth with an elevated V_p/V_s ratio. The stepwise velocity reduction and the gradual nature of
247 the Moho can explain the time difference of the PRF and SRF observations of the LVZ and Moho
248 phases.

249

250 **3.3 PRF and SRF migrated sections: the Moho and the LAB**

251

252 We present the migrated sections incorporating all available data in Fig. 10. In addition to the Moho

253 interface identified by the PRF, we look at the LAB discontinuity in order to answer the question,
254 whether the Puna plateau has a thin lithosphere as suggested by the delamination hypothesis for
255 this region (Kay and Kay, 1993; Kay et al., 1994). To better constrain the depth of the LAB, we also
256 use the SRF method as it is better suited to investigate the upper mantle as has been successfully
257 done in the Central Andes (Heit et al., 2007 and 2008) and elsewhere (e.g., Li et al., 2004; Kumar
258 et al., 2005a, 2005b; Sodoudi et al., 2006; Angus et al., 2006) as it is free of multiples. We used a
259 single scatterer approach (Kind, et al., 2002) for the PRF and SRF migration. P-to-S and S-to-P
260 converted amplitudes are back-projected to the space along the corresponding raypaths within a
261 width of a Fresnel's zone. A 10x10 km smoothing window is applied over the migration sections.
262 Ray-tracing is based on a 1D velocity model. Slightly dipping interfaces, such as the subducted
263 slab in the present study with a dip angle $<15^\circ$, can thus be properly reconstructed. For larger dips
264 ($>30^\circ$) a correction is necessary (Schneider et al., 2013).

265

266 The migrated PRF and SRF sections along the south-north and west-east profiles in Fig. 10 show
267 clear images of the Moho and the LAB. The profiles incorporate data available in a 200 km wide
268 swath on both margins of the profile and down to a depth of 200 km. The inclusion of data in the
269 marginal swaths of the profile helps to increase the robustness of the results as the use of more
270 events facilitates comparisons between the migrated sections and stacked traces in Figs. 3 and 4.

271

272 The west-east migrated sections in Fig. 10 clearly show the Moho topography (marked as M) in
273 both the PRF and SRF profiles. The Moho is at depths near 45 km in the east beneath the
274 Pampean Range, near 75 km in the west beneath the volcanic arc and near 50-60 km beneath the
275 Puna plateau. Although in detail, the Moho appears to be at slightly different depths in the PRF and
276 the SRF profiles, the depth trend is similar in both profiles and is in agreement with the result of the
277 H-K analysis (Fig. 7). The negative phases (blue) below the Moho represent the lithosphere-
278 asthenosphere boundary (marked LAB) which can be identified at similar depths between 70 km

279 and 80 km in the PRF and SRF sections. The trace of the slab is recognized in the PRF section
280 where the positive phase in red is coincident with the local seismicity detected by our network
281 (Mulcahy et al., 2013). The dispersion of the earthquake hypocenters on the profile is in part due to
282 the north to south change in depth to the Wadati-Benioff zone across the array. A cluster of
283 negative amplitudes at depth of ~150 km below the earthquakes in the PRF profile, which are also
284 observed in the SRF section, may represent the base of the oceanic lithosphere. The positive slab
285 converter is not as clear in the SRF profile. The thickness of the subducted Nazca lithosphere can
286 be estimated to be about 50 km.

287

288 The PRF migrated section for the north-south profile clearly shows the Moho depth variations. The
289 Moho is seen at 55 km in the north and deepens around 27°S at the southern border of the Puna
290 plateau. In contrast, the SRF migrated section shows the position of the Moho phases to be
291 deeper in the north (60 km) and to shallow to 55 km at the southern end of the Puna plateau at
292 27°S. This discrepancy in Moho topography in the north-south PRF and SRF profiles could be due
293 to the strong low-velocity anomalies observed in the crust. A negative phase above the Moho in the
294 middle crust can be traced along the PRF profile north of 28°S. An intra-crustal negative phase is
295 even more clearly observed on the SRF profile north of 27°S where its presence coincides with the
296 Puna plateau and the presence of the Sierras Pampeanas block south of 27°S. This low velocity
297 zone as explained above (see the Cerro Galan section) might produce the observed effect on the
298 Moho, making it apparently to appear at greater depth. The PRF profile shows a clear negative
299 phase at greater depth beneath the Moho that we interpret as the LAB. On the SRF profile, the
300 LAB is only observed at 27°S (80 km) probably due to the low quality of the SRF. The down-going
301 slab is not clearly observed in either the PRF or the SRF sections, although its position can be
302 inferred when correlated with the sparse seismicity (see Mulcahy et al. for more details). The slab
303 is difficult to image in this area where there is a pause in seismicity in the downgoing slab. This is
304 known as the Antofalla seismicity gap and our results are suggesting and supporting the idea of a

305 hot mantle and crust beneath the southern Puna. This could be related with recent lithospheric
306 delamination beneath the region.

307

308 **4. Discussion**

309

310 The small differences in depth determinations of the Moho and LAB between the results from the
311 PRF and SRF analysis can be explained partly by different locations of P-to-S and S-to-P piercing
312 points and partly by different frequency contents of the PRF and SRF. Specifically, the P-to-S
313 conversions are closer to the stations whereas the S-to-P conversions sample areas farther away
314 from the stations. The PRF has higher frequency content, therefore, is sensitive to sharper velocity
315 contrast, whereas the SRF is more sensitive to gradual interfaces. The PRF could be affected by
316 numerous crustal multiples that can mask the position of primary converted phases. The LAB is
317 thus best resolved by the SRF as the profiles are free of disturbing multiples. The SRF is, however,
318 noisier than the PRF, because the number of teleseismic events useful for calculation of the SRF is
319 limited for South America (Kind et al., 2012). For this reason, the use of SKS arrivals is imperative
320 as demonstrated by Kind et al. (2012).

321

322 The data provided by all stations represent well constrained results for Moho depths that
323 correspond to the different morpho-structural units and major volcanic centers in the southern
324 Puna plateau (Fig. 7). The Moho is deepest beneath the volcanic arc on the west where it reaches
325 a depth of 75 km in the northwest of the array. Such a thick crust can be explained by large
326 amounts of magmatic addition in combination with ductile thickening of the lower crust in response
327 to crustal shortening (e.g. Isacks, 1988; Allmendinger et al., 1997; Oncken et al., 2006) with the
328 resultant isostatic response leading to high topography. The thick crust is observed northwest of
329 the Antofalla volcano in an area where some active volcanoes are located (i.e. Lastarria, see Fig.

1). The position of this thickened crust beneath the volcanic arc speaks for magmatic addition at the base of the crust. The Lastarria and nearby volcanoes are located at the western end of the Archibarca lineament (see Fig. 1) which runs through the Antofalla and Galan volcanoes in the northern edge of the southern Puna plateau. The Archibarca lineament is part of number of northwest-southeast trending fault systems considered to be zones of lithospheric weakness (e.g. Bianchi et al., 2013). The center of the Puna plateau shows a relatively thin crust with a Moho depth generally between 50 and 55 km. This is very thin compared to the Moho depth obtained beneath the volcanic arc. On average the crust beneath the southern Puna is 5-10 km thinner than that beneath the northern Puna and 20 km thinner than beneath the Altiplano (Yuan et al., 2002, Woelbern et al., 2009). Towards the eastern border of the Puna plateau (at 67°W) the crust is slightly thicker than the central part of the plateau what could be indicating that the thickening is related to shortening of the crust at the border between the Puna and the Pampean Ranges.

A receiver function Moho under the volcanic arc in the W-E profile in Box 3 on Figure 3 appears to be invisible in the same locality that a strong P-to-S converted phase in box 3 provides evidence for a strong interface at a depth near 35 km. A Moho at ~35 km in this region is clearly inconsistent with the Moho depths in the migrated PRF section (Fig. 10) for the region, which show a crustal thickness near 70 km. An explanation for the lack of a clear Moho phase at the base of the crust and a strong intracrustal discontinuity at this depth in the box 3 profile is, however, consistent with eclogitization of the lower crust. Such a proposal for detecting the presence of eclogite in the lower crust by RFs was made by Gilbert et al. (2006) in the high cordillera region in the flat-slab region just to the south. In detail, they argued that an intracontinental discontinuity marking the top of eclogitized lower crust would create a larger impedance contrast than the boundary between the base of eclogitized lower crust and the mantle. This would lead to a situation in which the Moho interface would be nearly invisible to RFs due to the similarities of velocities of the mantle and the eclogitic lower crust. Such an explanation for the crust of the cordillera in the flatslab region was

356 noted to be consistent with previous inferences for the presence of an eclogitized crust based on
357 geochemical data from Neogene lavas (e.g., Kay et al., 1987; Mpodozis and Kay, 2002). A similar
358 argument can be made to explain the seismic observations in box 3 under the volcanic arc on
359 Figure 3, which is the region where the crust might be expected to be the thickest. As in the flatslab
360 segment, eclogitization of the crust can be inferred to be related to magmatic addition in
361 conjunction with ductile shortening of the lower crust in response to crustal shortening. In parallel
362 with the flat-slab region, Kay and Coira (2009) have proposed that the peak of crustal thickening in
363 this region could have accompanied the Miocene shallowing of the subducting Nazca plate under
364 the southern Puna, which was followed by resteepeening.

365

366 Recently, Carbonell et al. (2013) suggested that a geophysically invisible Moho might be an
367 indication of a thick crust-mantle transition zone where the impedance changes very slowly
368 between values for the crust to values of the mantle. They claim that this situation could be directly
369 related to a fast tectonic crustal thickening due to a continental orogenesis which can be the case
370 in the Central Andes.

371

372 The southern Puna plateau is characterized by low V_p/V_s ratios (less than 1.7), similar to the
373 Bolivian Altiplano to the north where Zandt et al. (1994) reported a V_p/V_s ratio of 1.6. Beck and
374 Zandt (2002) also obtained similar results for the Altiplano plateau (1.7) and suggested a felsic
375 crust is responsible for this value.

376

377 A low velocity crust with a high V_p/V_s ratio of 1.87 is observed in the crust beneath the Cerro
378 Galan caldera (Fig. 6), consistent with a zone of partial melt under the region. This interpretation is
379 consistent with the extremely high seismic attenuation observed beneath Cerro Galan by Liang et
380 al., (2013) using the same dataset. Surface wave tomography studies from the same array further

381 reveal pronounced low velocities in the crust in this area (Calixto et al. 2013). All of these
382 observations imply that a zone of partial melt or a magma chamber exists in the crust beneath
383 Cerro Galan as could result from the aftermath of lithospheric delamination. In accord with Kay et
384 al. (1994, 2011), Bianchi et al. (2013) suggested that a delamination event in this area could have
385 been triggered by pre-existing dense and thick crust. The region is also located in an area where a
386 prominent Ordovician arc occurred (e.g., Rapela et al., 1992; Coira et al., 1999), which could have
387 played a role in subsequent delamination. Thus, the thin mantle lithosphere with a depth of the
388 LAB at 70-80 km (Fig. 10) could be interpreted as supporting a delamination model in this area.

389

390 Another region of high Vp/Vs is also observed south of the Puna plateau in the Pampean Ranges
391 (see Fig. 6). No active volcanic edifice occurs in this region with the nearest center being the late
392 Miocene Farallon Negro volcano to the west (see Fig. 1). The cause of the high Vp/Vs anomaly is
393 still unclear.

394

395 In the upper-mantle, the slab can be traced down to 170 km depth as is better seen on the
396 migrated section in Fig. 11 where all available data is included (i.e. 300 km wide swath from 25°S
397 to 28°S). Thus, Fig. 11 shows the complete geodynamic picture of the study area including the low
398 velocity zone in the crust, the Moho, the LAB, the subducting slab and the upper-mantle
399 discontinuities. In this profile, the mantle transition zone (MTZ) appears coincident with global
400 averages elsewhere. In detail, the 410 km discontinuity is clear from 71°W to 66.5°W and then
401 becomes less clear near the impingement of the slab and the MTZ in the east near 64°W. The 660
402 km is rather masked by noise but is partly identifiable in the west where it is at the expected global
403 average depth. Finally, Fig. 11 is essential to demonstrate that the anomalous mantle transition
404 zone suggested by Woelbern et al. (2009) to the north of the present array can not be confirmed in
405 this study.

406

407 5. Conclusions

408

409 The P and S receiver functions presented here provide for the first time an image of the crust and
410 upper mantle in the Southern Puna plateau. The crust is the thickest beneath the volcanic arc with
411 a maximum depth of 75 km. In the Southern Puna plateau the crustal thickness is 50-55 km. The
412 thinnest crust occurs to the east of the plateau in the Sierras Pampeanas where the thickness is
413 35-40 km.

414

415 The crust of the plateau has a low V_p/V_s ratio (< 1.7), implying a felsic crustal composition. The RF
416 results revealed a widespread crustal low velocity zone in the Southern Puna plateau, consistent
417 with the crustal low velocity body observed by the teleseismic P wave tomography that occurs
418 across the central part of the southern Puna plateau and is related to various volcanic edifices
419 (Bianchi et al., 2013). Such a widespread crustal low velocity zone in the southern Puna is also in
420 accord with surface wave (Calixto et al., 2013) and body wave attenuation (Liang et al., in press)
421 tomographic images using teleseismic data and the earthquakes recorded in the study by Mulcahy
422 et al. (2013).

423

424 A low velocity crust with a high V_p/V_s ratio (1.87) observed beneath Cerro Galan provides
425 evidence for a partial melt zone or magma chamber in the underlying crust that could have
426 developed in the aftermath of lithospheric delamination. This hypothesis is in accord with the
427 proposal of Kay et al. (1994, 2011) that the giant eruptions of the Cerro Galan ignimbrite complex
428 were a consequence of delamination of the underlying crust and lithosphere, which triggered
429 decompression melting in the underlying mantle wedge leading to extensive crustal melting.

430

431 The subducted Nazca plate has been clearly observed down to depths of 170 km in a region of

432 transition between normal to flat subduction. The mantle transition zone discontinuities at 410 and
433 660 km appear at typical depths.

434

435 The receiver function images presented here are in accord with the continental lithospheric
436 delamination events in this region suggested by previous authors (e.g., Kay et al., 1994; Bianchi et
437 al., 2013). In detail, the PRF and SRF images reveal a relatively thin crust beneath the Puna
438 plateau (50-60 km) compared to the Altiplano and the presence of a shallow LAB (70-90 km).
439 Crustal thickening and eclogitization of the lower crust in response to crustal shortening and
440 magmatic addition along with resteepeening of a formerly shallower subduction zone (e.g., Kay and
441 Coira, 2009) are consistent with the triggering of delamination events that left behind the thinned
442 crust and shallow LAB now observed in the southern Puna

443

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451

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647 **Figure Captions**

648

649 Figure 1: Map showing the distribution of stations (blue inverted triangles) and the tectonic units
650 (grey lines) in the southern Puna. Red triangles denote volcanoes. The red dashed line is the
651 Chile/Argentina border. Volcanic centers inside the array are marked with black circles; At:
652 Antofalla; G: Cerro Galan Caldera; Pn: Peinado; Cb: Cerro Blanco Caldera; Os: Ojos del Salado;
653 Fn: Farallon Negro. The blue dashed line is the trace of the Archibarca lineament that runs from
654 the northwest at the position of the Lastarria volcano (marked as L) to the southeast in the
655 southern Puna plateau.

656

657 Figure 2: Map of teleseismic earthquakes used in this study. Blue circles denote epicenters in a
658 distance range of 30° - 180° used for P and PP phases. Red circles denote earthquakes at
659 epicentral distances between 60° and 115° used for S and SKS phases. Black concentric circles
660 marked as reference for epicentral distances every 30° .

661

662 Figure 3: Upper panels: Map showing the stacking boxes (black squares with bold squares in the
663 center of each box) and piercing points (violet circles) for two orthogonal EW (left) and NS (right)
664 profiles with the highest station density. The piercing points are calculated at a depth of 60 km,
665 roughly representing the depth of the Moho. Receiver functions are stacked for piercing points
666 falling into each box with a size of 0.5° overlapping by 0.1° . Lower panels: Stacked receiver
667 functions along the two profiles. Positive amplitudes are shaded in red, negatives in black.
668 Topography is sampled every 1 km across the red line on the maps. Black dots on the profiles are
669 hypocenters from Mulcahy et al. (2013). The Moho phases are marked by the blue line and
670 indicated as M on the profiles.

671

672 Figure 4: Map of piercing points (upper panel) at a depth of 60 km grouped by boxes within which
673 receiver functions are stacked and displayed along 6 profiles (marked as red lines on the upper
674 panel map). Receiver functions shown in Fig. 3 are excluded here. Size of the boxes is the same
675 as in Fig. 3. Black dots on the profiles denote hypocenters from Mulcahy et al. (2013). The Moho
676 phases are marked by the blue line and indicated as M on the profiles. The black dashed line
677 represents the approximate trace of the slab after Mulcahy et al. (2013).

678

679 Figure 5: Sample H-K stacks of receiver functions for three groups indicated in Fig. 3 (WE profile
680 Boxes 2, 4, 9) and also shown in Fig. 6. Within each group a grid search is performed for estimates
681 of Moho depth and crustal Vp/Vs ratio, indicated by the position of the maximum. The three traces
682 on the right side of each panel are stacked receiver functions after moveout correction for the Ps,
683 PpPs and PpSs phases, respectively. Red bars denote the predicted times of the three phases by
684 the model obtained for each group.

685

686 Figure 6: Map of crustal Vp/Vs ratio determined by the H-K analysis (Zhu & Kanamori, 2000). The
687 Ps Moho conversion and the crustal multiples (PpPs and PpSs) are used to determine the crustal
688 thickness and Vp/Vs ratio. Black lines mark the tectonic units as shown in Fig. 1. The southern
689 Puna is characterized by a low crustal Vp/Vs ratio (<1.70), indicating a felsic crustal composition.
690 The Vp/Vs ratio beneath Cerro Galan is anomalously high (1.87), implying presence of melts. The
691 position of Galan is included as a reference.

692

693 Figure 7: Map of crustal thickness in the study area. Small circles represent the center of the boxes
694 used to obtain the Moho depth. Most of the values are obtained by the H-K analysis (indicated by
695 bold squares). For some of the boxes where the H-K analysis failed, an average Vp/Vs ratio of
696 1.74 is used to estimate the Moho depth (indicated by light circles). Black lines mark the tectonic

697 units as in shown Fig. 1. The crust is thicker (>60 km) in the west and northwest beneath the
698 volcanic arc. The Moho in the Puna region is relatively flat with depths varying between 50-60 km.
699 To the southeast, the Sierras Pampeanas have a crustal thickness between 35-45 km. The position
700 of Galan is included as a reference.

701

702 Figure 8: P (a) and S (b) receiver functions for the station located at the Cerro Galan caldera.
703 Individual receiver functions are sorted by back azimuth and displayed equally spaced. Sample
704 back azimuth every 5 traces are indicated on the left side of each panel. Summation traces are
705 displayed in the top panels. Positive amplitude are shaded in black, negative in gray. Two coherent
706 conversion phases, the positive Moho conversion and a negative crustal conversion (LVZ), are
707 indicated on the top of the summation traces. Both negative LVZ phases can be clearly observed
708 by the P and the S receiver functions.

709

710 Figure 9: Waveform modeling of P receiver functions for the Cerro Galan station. (a) S wave
711 velocity model obtained by waveform fitting. A prominent crustal low velocity zone is needed to fit
712 the strong negative phase at ~ 2.5 s in the receiver function shown in (b). In the model, the Moho is
713 a gradient zone from 55 to 60 km depth. The dashed line is the global reference IASP91 model. (b)
714 Stacked P receiver function (gray curve) and the synthetic one (black curve) corresponding to the
715 model shown in (a). (c) H-K analysis showing a Moho depth of 59 km and a crustal V_p/V_s ratio of
716 1.87 beneath the station.

717

718 Figure 10: Migrated sections performed using receiver functions with piercing points falling within a
719 100 km wide band along the profiles defined in Fig. 3. The upper two panels are west-east (left)
720 and north-south (right) sections of P receiver functions. The lower panels are the corresponding S
721 receiver function sections. Positive amplitudes are coded in red, while negative are coded in blue.

722 Along all profiles the positive Moho phase (50-75 km) and the negative LAB phase (70-90 km) can
723 be observed. Earthquakes from Mulcahy et al. (2013) within a same swath width along the profiles.
724 The approximate position of the slab is marked by the dashed line.

725

726 Figure 11: Migrated upper mantle section along 27° S using data from all stations. Red and blue
727 colors denote positive and negative amplitudes, respectively. The Moho in the Puna plateau
728 (dotted line), as well as the LVZ, the LAB, the oceanic slab and the mantle transition zone
729 discontinuities (dashed lines) can be well recognized in the figure. The CVZ (volcanic arc) and the
730 Galan volcano are marked as a reference on the profile. On the right, the summation trace (sum)
731 highlights the phases seen on the profile where it is possible to recognize that both 410 and 660
732 arrive at expected global averaged depths.

***Highlights (for review)**

Highlights

Thin lithosphere due to lithospheric delamination beneath the Puna plateau

Low-velocity zone and high V_p/V_s ratio beneath the Cerro Galan volcano

Intra-crustal low velocity zones along the entire southern Puna

These figures represent our attempt to demonstrate that the features we observe remain stable after changing the size of the boxes and the overlapping. The original submitted figure is in the middle (0.5x0.5).

Changing the latitude and longitude size for the West-East profile

Changing the longitude side for the South-North profile

Changing the latitude and longitude size for the South-North profile

Figure1

[Click Here to download Figure-06.pdf](#)

Figure2

[Click here to download Figure: figure-02.pdf](#)

Figure3-70°
[Click here to download Figure: figure-03.pdf](#)

Figure4

[Click here to download Figure: figure-04.pdf](#)

Figure7
[Click here to download Figure: figure-07.pdf](#)

Figure8
[Click here to download Figure: figure-08.pdf](#)

Figure 9
[Click here to download Figure: figure-09.pdf](#)

Figure10

[Click here to download Figure: figure-10.pdf](#)

PRF migrated sections

SRF migrated sections

Figure11

[Click here to download Figure: figure-11.pdf](#)