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Seismic interferometry applied to regional and teleseismic events recorded at Planchón-Peteroa Volcanic Complex, Argentina-Chile

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Abstract

The Planchón-Peteroa Volcanic Complex (PPVC) is located in the Central Andes, Argentina-Chile. Even though this active volcanic system is considered one of the most dangerous volcano in the region, with more than twenty modest ($VEI < 4$) Holocene eruptions, knowledge of its subsurface structures, internal processes, dynamics, and their relation, is still limited.

Seismic interferometry (SI) is a high-resolution technique based on analyses of the interference of the propagated seismic energy at one or many stations. SI can be used to characterize the subsurface properties of a target area. In particular, previous SI studies performed in the area of the PPVC describe specific ranges of depth; therefore, more information is required for a thorough description of the subsurface features in the area and for a better understanding of the PPVC dynamics.

We apply SI based on autocorrelations of selected regional and teleseismic events to image the subsurface structures below stations located in

Argentina and Chile during 2012. The selection of the events is performed according to their location, magnitude, angle of incidence of P-wave seismic energy, and signal to noise ratio in the records. For each station, we extract time windows and we process them using three ranges of frequency, which are sensitive to different ranges of depths.

This work describes depths and zones previously not analyzed in the area. The results not only complement the available geological, geochemical, and geophysical information, but present new information for depths between 5 and ~ 400 km depth, increasing the general knowledge of the subsurface features in the PPVC. Finally, we also propose a model for the first 45 km of the subsurface (i.e., down to the Moho), which indicates the crustal structure and the likely distribution of magma bodies in depth.

Keywords:

Planchón-Peteroa Volcanic Complex, Seismic Interferometry, Regional and teleseismic events, Magma storage in depth

1. Introduction

The Planchón-Peteroa Volcanic Complex -PPVC- (35.223° S, 70.568° W; see location in [Figure 1](#)) is located in the Andes at the international border between Argentina and Chile. The PPVC is composed of three main volcanic edifices, i.e., the Azufre, the Planchón, and the Peteroa, out of which the latter is the current active volcano. The PPVC presents overlapped calderas originated from the destruction of several volcanic structures during past explosive events ([Tormey, 1989](#)). Through analyses of its historical activity and products, this volcanic system is ranked as the most hazardous volcano in Argentina ([Elissondo and Farías, 2016](#)) and the eighth most risky volcano

11 in Chile ([Technical sheet](#), Observatorio Volcanológico de los Andes del Sur,
12 OVDAS-SERNAGEOMIN, Chile).

13 The knowledge of the PPVC has been developed by the contribution
14 from several disciplines, i.e., geology ([Tormey, 1989](#); [Haller et al., 1994](#);
15 [Naranjo et al., 1999](#); [Tapia Silva, 2010](#); [Haller and Risso, 2011](#)), geochemistry
16 ([Benavente, 2010](#); [Tassi et al., 2016](#); [Benavente et al., 2016](#)), meteorology
17 ([Guzmán et al., 2013](#)), ash analysis ([Ramires et al., 2013](#)), seismology ([Casas](#)
18 [et al., 2014](#); [Manassero et al., 2014](#); [Olivera Craig, 2017](#); [Casas et al., 2018](#);
19 [Casas et al.](#)), gravimetry ([Tassara et al., 2006](#)), and risk analysis ([Haller](#)
20 [and Coscarella](#)). These studies contribute to the knowledge of the eruptive
21 history and the current subsurface conditions of this volcanic system. Nev-
22 ertheless, the dynamics the PPVC and their relation with the subsurface
23 structures are still poorly understood, increasing the local risk ([Elissondo](#)
24 [and Farías, 2016](#)).

25 A description of the subsurface structures (i.e., depth, associated dimen-
26 sions, density contrasts, etc.) is essential for developing accurate knowl-
27 edge of the dynamics of any volcanic system. In particular, knowledge of
28 subsurface discontinuities provides constraints for tomographic studies, for
29 magma-ascent modeling, among others, contributing to a better inference
30 of the subsurface conditions, and, therefore, leading to more reliable analy-
31 ses of likely future volcanic scenarios. Based on structural-geology analyses,
32 [Tapia Silva \(2010\)](#) describes the subsurface geological units located in the
33 very first 10 km of the subsurface in the area of the PPVC, and present
34 their distribution in depth. Even though no local studies have been applied
35 for describing the crustal structure in the PPVC, [Farías et al. \(2010\)](#) and
36 [Giambiagi et al. \(2012\)](#) provide a crustal structure as a function of depth
37 and the distance from the trench in the Central Andes. For the depth of

38 the subducting slab below the PPVC, they estimate four zones delimited in
39 depth at ~ 12 (the intracrustal discontinuity), ~ 27 , and 45 km depth -the
40 crust-mantle discontinuity (the Moho). The Moho is estimated at ~ 45 km
41 depth (Tassara et al., 2006); the intra-lithosphere discontinuity (top of litho-
42 spheric low-velocity zone), at ~ 70 km depth (Karato, 2012); and the top of
43 the subducting slab, at ~ 120 km depth (Tassara et al., 2006). Nevertheless,
44 more scientific evidence is required to increase the information about the
45 known subsurface structures, leading to a more accurate characterization of
46 their properties, as well as to describe the subsurface features previously not
47 analyzed. These goals motivate local studies, as the one presented in this
48 article.

49 Claerbout (1968) has constituted a frame over which the theory of seis-
50 mic interferometry developed. This passive seismic method -from here on,
51 Seismic Interferometry by Autocorrelations (SIbyA)- suggests that the au-
52 tocorrelation of a plane-wave transmission response propagating in a hori-
53 zontally layered medium, recorded at the surface, allows the retrieval of the
54 reflection response of a virtual source co-located to the recording station.
55 SIbyA has shown to be a robust method; it has been applied to different
56 type of seismic data, in several areas and at different scales. For example,
57 SIbyA was applied to global- and teleseismic phases to imaging the crustal
58 subsurface at regional scales (Ruigrok and Wapenaar, 2012; Nishitsuji et al.,
59 2016), to P-wave of microseismic events to imaging the shallow volcanic sub-
60 surface (Kim et al., 2017), and to ambient-noise seismic data at several scales
61 (Draganov et al., 2007; Gorbatov et al., 2013; Boullenger et al., 2014; Oren
62 and Nowack, 2017). The robustness of SIbyA has motivated its application
63 to local (Casas et al.), regional, and teleseismic seismic data recorded in the
64 area of the PPVC.

65 [Nishitsuji et al. \(2016\)](#) apply SIbyA to global seismic phases recorded
66 in the eastern flank of the Peteroa volcano during 2012. They confirm the
67 location of the Moho at ~ 45 km depth, and propose a deformation feature
68 in the subducting slab in the form of detachment, shearing, necking, or any
69 combination of them.

70 [Casas et al.](#) apply SIbyA to local seismic events to image the subsurface
71 below the stations located in the Argentine and Chilean sides of the PPVC
72 during 2012. They confirm the geological structure described for the first
73 4 km of the subsurface ([Tapia Silva, 2010](#)), provide information about re-
74 gions of higher heterogeneity caused by faulting and complex geochemical
75 processes, and support the presence of a magma body emplaced at ~ 4 km
76 depth (previously suggested by [Benavente \(2010\)](#)).

77 We apply SIbyA to regional and teleseismic events selected according to
78 their location, magnitude, angles of incidence of the P-wave seismic energy
79 at each station, and the signal to noise ratio in the records. The results
80 for three different frequency ranges allow the description of the subsurface
81 structures between ~ 5 and 400 km depth, and the inference of the crustal
82 structure and the location of magma bodies down to the Moho.

83 **2. Data**

84 The present application uses seismic data recorded by stations deployed
85 in Argentina and Chile during 2012 (see station distribution in [Figure 1](#)).

86 The temporal deployment of seismic instruments in an area of interest is
87 a widely used tool for reaching several goals, e.g., perform first analyses of
88 the propagated wavefield and the subsurface conditions, increase the num-
89 ber of the recording stations, extend the analyzed area, and improve the

90 accuracy of the results. The MalARRgue project (Ruigrok et al., 2012) was
 91 designed by institutions from The Netherlands (Delft University of Technol-
 92 ogy -TUDelft), Argentina (Comisión Nacional de Energía Atómica CNEA),
 93 and The United States (Boise State University -BSU). Its goal is imaging
 94 and monitoring the subsurface of the Malarge region (Mendoza, Argentina),
 95 an area of high scientific interest due to peculiar volcanic and tectonic pro-
 96 cesses (Stern, 2004). The MarlARRgue project consisted in a temporal
 97 deployment (from January 2012 to January 2013) of 38 stations, out of
 98 which six were deployed along the eastern flank of the PPVC (from here
 99 on, the PV array). The PV array was equipped with short-period (2 Hz)
 100 three-component (Sercel L-22) sensors.

101 Another source of data is provided by three broad-band stations of the
 102 Observatorio Volcanológico de los Andes del Sur (OVDAS-SERNAGEOMIN,
 103 Chile), which are located ~ 6 km northwards. These stations (from here on,
 104 OVDAS array) were active during 2012, through the same period as the PV
 105 array.

106 3. Application and results

107 SIbyA is described by the reciprocity theorem of correlation type (Wape-
 108 naar, 2003, 2004). Based on this theorem for transient sources (Wapenaar
 109 and Fokkema, 2006), and using autocorrelation in the time domain, we ob-
 110 tain:

$$\sum_{sources} \{ [T(\mathbf{x}_A, -t) * s_i(-t) * T(\mathbf{x}_A, t) * s_i(t)] \otimes [s(-t) * s(t)]_i \}$$

$$\approx -R(\mathbf{x}_A, -t) + \delta(t) - R(\mathbf{x}_A, t) \quad , \quad (1)$$

111 which states that the reflection response $R(\mathbf{x}_A, t)$ can be retrieved at
112 the station A located (at \mathbf{x}_A) at the surface through the autocorrelation
113 of a recorded transmitted wavefield $T(\mathbf{x}_A, t)$. The operator $*$ indicates
114 convolution, \otimes means deconvolution, and δ is the Dirac's delta. The fac-
115 tor $[s(-t) * s(t)]_i$ corresponds to the autocorrelated source time function
116 (ASTF), which allows the deconvolution of each source time function $s_i(t)$.

117 Even though [Equation 1](#) requires sources over the whole stationary phase
118 area (i.e., the Fresnel Zone), seismic events present a non-uniform spatial
119 distribution. Therefore, performing a selection of the seismic sources to be
120 used is essential for a proper application of SIbyA. In order the transmission
121 response of the propagated seismic energy to be accurately estimated by
122 the vertical component of the records, we select only seismic events with
123 P-wave seismic energy arriving (sub) vertically to the station at the surface.
124 The retrieved reflection response (from here on, $R_v(\mathbf{x}_A, t)$) is related to a
125 seismic source co-located to the station at the surface, radiating P-wave
126 energy vertically downwards.

127 A seismic source in the subsurface release energy that propagates to-
128 wards the surface in which it is reflected. This seismic energy is reflected,
129 refracted, and diffracted at the subsurface structures and heterogeneities (or
130 the surface), part of which arrives to the recording station at the surface.
131 Seismograms are then composed of direct waves followed by these reverber-
132 ated waves. SIbyA removes the times previous to the direct arrival, and
133 attenuates the incoherent noise, providing seismic evidence of the location
134 of the subsurface structures. [Figure 2](#) pictures the application of SIbyA in
135 an idealized horizontally layered 2-D medium, given a plane wavefield orig-
136 inated by a seismic source located exactly below the station. The obtained
137 reflection response can be used to know the depth of the reflectors located

138 in the subsurface below the station.

139 In the real Earth, nor the wave fronts are plane at local and regional
140 scales nor the subsurface is horizontally layered in volcanic zones. In highly
141 heterogeneous zones (as, for example, the area of the PPVC; [Manassero et al.](#)
142 [\(2014\)](#)), the location of a seismic source exactly below the station is not an
143 imperative condition for an accurate retrieval of the subsurface reflection
144 response $R_v(\mathbf{x}_A, t)$ ([Fan and Snieder, 2009](#)), i.e., the vertical component
145 of the records is still an accurate estimation of the transmission response.
146 Therefore, sources with small P-wave angles of incidence are selected.

147 *3.1. Pre-processing*

148 This section aims to get the input data and prepare it for the proper
149 application of the [Equation 1](#). Using the reference seismic catalogs (IRIS and
150 USGS), we select events occurred during the recording period (i.e., January
151 2012 until January 2013) and which are characterized by a sufficiently great
152 magnitude so that signal to noise ratio is high in the records of each station.
153 Due to likely variations of the local seismic wavefield in space and time, we
154 judge the signal to noise ratio of each event at each of the stations.

155 For the selection of seismic events, we use the software JWEED (Java
156 version of Windows Extracted from Event Data) developed by IRIS. Based
157 on restrictions in the origin time, the location, and the magnitude, we pre-
158 select events (see [Figure 3](#)). According to their epicentral distance, we clas-
159 sify them in two groups. One group is composed of events with epicentral
160 distances between 30° and 120° , and magnitudes greater than Mw. 6; each
161 event of this group guarantees a sufficiently small P-wave ray parameter
162 (< 0.08 s/km) so that seismic energy arrives (sub)vertically at the station,
163 i.e., with incident angles $< \sim 25^\circ$ ([Kennett et al., 1995](#)). The second group

164 is composed of events with epicentral distances lower than 30° and magni-
165 tudes greater than Mw. 5. These events present a wide range of possible
166 P-wave angles of incidence. Therefore, we perform an examination analysis
167 on this second group in order to select only those events with at least one
168 P-wave phase smaller than the adopted threshold (i.e., 0.08 s/km). The ray
169 parameters estimated by the regional velocity model ak135 (Kennett et al.,
170 1995) are appropriate for this analysis, as smaller angles of incidence of the
171 P-wave energy are expected in the real Earth, provided its higher hetero-
172 geneity (Fan and Snieder, 2009). Once the seismic events are selected, there
173 is no need to keep the distinction between the groups, i.e., they are equally
174 significant.

175 The origin time of the selected events is used to extract the seismic
176 waveforms from the records of the PV and OVDAS stations. A first estimate
177 of the P- and S-wave arrival times for each event is calculated using the
178 regional velocity model ak135. These times are then employed to manually
179 pick accurate P- and S-wave arrival times, as well as to get the frequency
180 range of a sufficiently high signal-to-noise ratio. We request a good ($>$
181 0.8) signal to noise ratio for the events to be processed, in order to avoid
182 non-interested high amplitudes.

183 Provided the origin time of the selected events, obtained the accurate
184 arrival times, and examined the (sub)vertical incidence of the P-wave en-
185 ergy and high signal-to-noise ratio of the records, we extract the vertical-
186 component records of the selected events at each of the used stations.

187 *3.2. Processing*

188 The vertical-component records of seismic events with P-wave energy
189 arriving vertically at a station represent an accurate estimate of the P-wave

190 transmission response of such propagated wavefield (provided the disconti-
191 nities are not excessively inclined; [Nishitsuji et al. \(2016\)](#)).

192 Out of the frequency range of processing previously selected for each
193 event according to its signal to noise ratio in the results, we use those fre-
194 quencies greater than 0.3 Hz, a threshold defined by the instrumental char-
195 acteristics of the PV-array stations ([Nishitsuji et al., 2014](#)). Furthermore,
196 we only use those frequencies shared through the events, i.e., [0.3 2.1] Hz.
197 In order to perform a better interpretation of the results through depth, we
198 segmented this frequency range in two sub-ranges, i.e., [0.3 0.8] Hz and [0.8
199 2.1] Hz. The separating frequency (0.8 Hz) is selected after a trial and error
200 approach, based on the observed coherency in the results for all the stations
201 in advanced stages of the processing.

202 In order to avoid the rise of non-physical arrivals caused by cross-terms
203 in the correlations, we extract the times between the first P-wave arrival
204 and the first S-phase arrival. As an example, [Figure 4](#) shows the processing
205 windows for the station PV04, for the complete range of frequencies (i.e.,
206 [0.8 2.1] Hz).

207 The higher value of the selected frequency range (i.e., 2.1 Hz) restricts
208 the resolution of the results for particular depths. Therefore, out of the
209 (previously tested) vertically arriving seismic events, we make a third group
210 composed of those with epicentral distances smaller than 20° . These events
211 are characterized by a sufficiently high signal-to-noise ratio up to 3.2 Hz.
212 As this group aims to provide information about shallower subsurface struc-
213 tures, we select a minimum frequency of 1 Hz. Therefore, we apply the
214 same processing workflow to the three selected frequency ranges, i.e., [0.3
215 0.8] Hz, [0.8 2.1] Hz, and [1 3.2] Hz. As the same importance is assigned to
216 the events of each of the three groups, we normalize the processing windows

217 according to their vertical flux of seismic energy.

218 As suggested by [Equation 1](#), we estimate and deconvolve the ASTF
219 from each of the autocorrelated time windows. The ASTF of each event is
220 estimated by the main lobe and the secondary monotonous-decreasing lobes,
221 as shown in [Figure 5](#) for the station AD2 and the frequency range [0.3 0.8]
222 Hz.

223 [Figure 6](#) presents the autocorrelation of the time windows for the station
224 PV01 and the frequency range [0.3 0.8] Hz, in which each trace is decon-
225 volved by its previously estimated ASTF. This figure shows the dominance
226 of the main lobe in the autocorrelated deconvolved traces. These features
227 close to 0 s are mainly non-physical amplitudes remaining from the decon-
228 volution. Therefore, these amplitudes are removed through windowing.

229 SIbyA is based on the autocorrelation of time windows extracted from
230 the records of selected seismic events. Note that the autocorrelation of
231 a extracted time window could arise non-physical arrivals at times equal
232 to the time interval between two P-wave arrivals, reducing the quality of
233 the results. However, these time intervals are a function of the epicentral
234 distance of the events. The seismic events used in this application present
235 a wide range of epicentral distances, so that the non-physical arrivals are
236 located at different times in the autocorrelations, leading to a destructive
237 interference of their energy during stacking (see [Figure 7](#)).

238 The last step in the application of [Equation 1](#) is stacking the resulting
239 seismic traces for each station, which enhances the energy from the sta-
240 tionary phase area. [Figure 8](#) shows the pre-stack panel (deconvolved and
241 windowed traces) and the stacked traces for stations AD2 and PV04, for the
242 three selected frequency ranges of processing.

243 4. Interpretation and discussion

244 Aiming to compare the seismic results with the known features of the
245 subsurface, we transform the time vector of the results to depth through
246 construction and utilization of a velocity model. This model is composed of
247 velocities provided by the regional model ak135 for depths greater than 60
248 km, and a modified version of the one obtained by [Bohm et al. \(2002\)](#) for
249 shallower depths (see used velocity model in [Figure 9](#)).

250 [Figure 10](#) and [Figure 11](#) show the stacked traces for the PV and OV-
251 DAS arrays, respectively, for each processing frequency range. These figures
252 also show the interpreted subsurface features for each of the stations. As a
253 complex impedance contrast through depth is expected for the area of the
254 PPVC, we only seek for the dominant amplitudes on the obtained reflection
255 responses, which are potentially related to the main subsurface discontinu-
256 ities. The lower frequency range (i.e., [0.3 0.8] Hz) leads to describe the
257 subsurface between ~ 40 and 400 km depth. The results for the other two
258 frequency ranges (i.e., [0.8 2.1] Hz, and [1 3.2] Hz) allow to interpret the
259 subsurface features for depths between 5 and ~ 45 km. The minimum depth
260 limit is set by the non-physical amplitudes removed from close to 0 s after
261 deconvolution. The maximum depth limit is set by the coherency in the
262 results for all the frequency ranges and all the used stations.

263 The interpretation of the results for the smallest frequency range ([0.3
264 0.8] Hz) is performed through contrast of the seismic results and the expected
265 location of the known subsurface features based on the geodynamic scenario
266 and the available geological information for the area of the PPVC ([Ferrán
267 and Martínez, 1962](#); [Tassara et al., 2006](#); [Benavente, 2010](#); [Tapia Silva, 2010](#);
268 [Karato, 2012](#)).

269 The results for the PV array (see [Figure 10a](#)) show six dominant ampli-
270 tudes (i.e., local maximum on the absolute values of the waveforms), which
271 we classify as potential subsurface discontinuities. The close location of the
272 identified features in the seismic results and the known subsurface features
273 lead to the interpretation of the Mohorovicic discontinuity at ~ 45 km depth,
274 the intra-lithospheric discontinuity at 65 km, the top of the subducting slab
275 between 110 and 120 km, the bottom of the subducting slab between 140
276 and ~ 160 km, the lithosphere-asthenosphere boundary between 230 and 255
277 km, and the top of the asthenospheric low-velocity zone between ~ 330 and
278 ~ 360 km depth.

279 The OVDAS array (see [Figure 11a](#)) is an array located ~ 6 km to the
280 north of the PV array, composed of half the stations of the latter, and
281 with greater longitudinal extension. The results for the OVDAS array al-
282 low to interpret the Mohorovicic discontinuity at ~ 45 km depth, the intra-
283 lithospheric discontinuity between 70 and 90 km, the top of the subducting
284 slab between 115 and 130 km, the bottom of the subducting slab between
285 ~ 165 and ~ 185 km, the lithosphere-asthenosphere boundary at ~ 250 km,
286 and the top of the asthenospheric low-velocity zone between ~ 310 and ~ 350
287 km depth.

288 Based on the seismic velocity values for the depths of interpretation and
289 the frequency range of processing, the resolution of the seismic results is 5
290 km ([Widess, 1973](#)). This value leads to interpret that the results for the
291 OVDAS array do not differ substantially from the results of the PV array,
292 what is expected provided the small geological variation in ~ 6 km along
293 the north-south direction for the used processing wavelengths. The best
294 correlation in depth is observed for the Mohorovicic discontinuity (43-48 km
295 depth), the lithosphere-asthenosphere boundary (~ 245 km), and the top

296 of the asthenospheric low-velocity zone (~ 340 km). A small difference in
297 depth is observed for the intra-lithospheric discontinuity and the top of the
298 subducting slab; even though greater depths are observed in the results of
299 the OVDAS stations, these differences would not be significant based on
300 the vertical resolution of the results. A greater difference is observed for the
301 bottom of the subducting slab, i.e., ~ 15 km greater for the OVDAS stations.

302 Although a dominant positive arrival is expected at the depth of the
303 Moho, a dominant negative amplitude is retrieved in the results for most
304 of the stations. Based on the retrieved waveforms, we interpret the pres-
305 ence of a complex area at ~ 40 - 55 km depth, causing a perturbation of the
306 amplitudes retrieved for these depths, in particular for those related to the
307 Moho.

308 Even though dipping structures in the subsurface restrict the reflection
309 energy arrived at the surface, we clearly recognize the depth of the top and
310 bottom of the subducting slab. Therefore, two hypotheses arise. One hy-
311 pothesis suggests a stair-like subduction, according to which the top and the
312 bottom of the oceanic slab present horizontal (or gently inclined) regions;
313 the different depths estimated in the results of the PV and the OVDAS ar-
314 rays for the bottom of the subducting slab could be caused by a local change
315 of the thickness of the subducting lithosphere. Nevertheless, this hypothesis
316 would not explain the lack of seismicity at the longitude of the stations and
317 depths of analysis (*US Geological Survey*; Nishitsuji et al. (2016)). A second
318 hypothesis (Nishitsuji et al., 2016) proposes a slab deformation in the form
319 of detachment, shearing, necking, or any combination. Then, a differential
320 deformation between the latitudes of the PV and OVDAS arrays would ex-
321 plain the estimated depths for the bottom of the subducting slab. Finally,
322 more information is required to elucidate the proper interpretation.

323 For the two higher ranges of frequencies (i.e., [0.8 2.1] Hz and [1 3.2] Hz)
324 (see [Figure 10b](#), [Figure 10c](#), [Figure 11b](#), and [Figure 11c](#)), the interpretation
325 is also based on the identification of the dominant amplitudes in the results,
326 and the depths for which the arrived reflected energy is particularly smaller,
327 a feature probably caused by the emplacement of a sufficiently great volume
328 of magma as to be manifested in the seismic results.

329 The results for the PV array and the frequency range [0.8 2.1] Hz (see
330 [Figure 10b](#)) indicate five clear dominant arrivals in most of the stations, out
331 of which four are between ~ 10 and ~ 30 km depth and another one at ~ 40 km
332 depth. Additionally, we identify an apparent lack of dominant amplitudes
333 for depths between ~ 30 and ~ 40 km (indicated with an arrow in [Figure 10b](#)).
334 The features identified for [0.8 2.1] Hz are supported by the results for the
335 frequency range [1 3.2] Hz ([Figure 10c](#)), which improve the depth of the
336 inferred subsurface discontinuities. In addition, these results manifest an
337 apparent low-amplitude region at ~ 25 km depth for the western stations of
338 the array. The results for this frequency range also show a dominant arrival
339 at ~ 6 km depth.

340 The results for the OVDAS stations agree with the interpretation per-
341 formed for the PV array, for the two analyzed frequency ranges. Therefore,
342 we identify local-maximum amplitudes, as well as apparent small-amplitude
343 zones, at roughly the same depths for the two arrays and for the two higher-
344 frequency ranges, even though the effect of attenuation increases for the
345 highest frequencies (around 3 Hz in this application) ([Schön, 2015](#)). Then,
346 these results allow the interpretation of the subsurface structures between 5
347 and ~ 45 km depth (the Moho).

348 Based on the average depth of the reflectors interpreted in the seismic re-
349 sults, the available scientific information about the subsurface in the PPVC,

350 the proposed structure of the crust for the Central Andes (Farías et al.,
351 2010; Giambiagi et al., 2012), and the physics of magma storage in the crust
352 Jackson et al. (2018), we propose a model for the distribution of magma
353 reservoirs in depth in relation to the main subsurface structures in the crust
354 (see Figure 12).

355 Through comparison of the average depth of the interpreted reflectors
356 below the stations and the proposed structure of the crust (Farías et al.,
357 2010; Giambiagi et al., 2012), we associate the interpreted reflectors at ~ 12 ,
358 ~ 18 , and ~ 32 km depth as the intra-crustal discontinuity (rigid-ductile dis-
359 continuity in the upper crust), the discontinuity between the upper and lower
360 crust, and the rigid-ductile discontinuity in the lower crust, respectively (see
361 Figure 12).

362 Jackson et al. (2018) models the formation, storage, and chemical differ-
363 entiation of magma in the Earth’s crust. According to the physics of magma
364 storage, the melt fraction is not homogeneously distributed through depth.
365 A great percentage of melt is located in the very upper part of a reservoir,
366 a low percentage is located through most of the reservoir, while a solid area
367 is present in the lower part. The seismic results are most probably evidence
368 of the solid lower section of the reservoir (Jackson et al., 2018). Therefore,
369 we interpret a region in depth as characterized by a magma emplacement
370 in case two conditions are satisfied: 1. the presence of an area of smaller
371 amplitudes in the seismic results, and 2. it is located above any of the inter-
372 preted subsurface reflectors. This circumstance is satisfied for two regions,
373 i.e., a shallower zone located above the rigid-ductile discontinuity in the
374 lower crust (i.e., ~ 32 km depth); and a deeper one at ~ 35 km depth, above
375 a reflector located at ~ 40 km.

376 Even though no amplitude information is available for depths lower than

377 5 km depth (which are removed after deconvolution), a subsurface model for
378 the area (Benavente, 2010) proposes a magma emplacement at ~ 4 km depth.
379 We identify a reflector at ~ 6 km depth, which motivates the incorporation
380 of such magma emplacement in our model.

381 Furthermore, two regions (indicated with a question mark in Figure 12)
382 satisfy only one of the imposed conditions, therefore, their interpretation as
383 regions of magma storage is subjected to extra information. These regions
384 are located above the reflectors interpreted at ~ 22 km depth and the Moho,
385 for which no apparent smaller amplitudes are observed, probably due to
386 its close location to another feature of the subsurface (upper-lower crust
387 discontinuity and the Moho, respectively), or the resolution of the seismic
388 results are not sufficiently great to recognize a region of limited vertical
389 extension of magma.

390 Our results support the information obtained for the subsurface in the
391 area (Yuan et al., 2006; Ward et al., 2013; González-Vidal et al., 2018)
392 which indicate (although with a limited resolution) low-velocity zones for
393 approximately the same range of depths. They are also consistent with
394 the conceptual model proposed for the area (Benavente, 2010) for depths
395 between 5 and 15 km depth, for which great volumes of magma storage are
396 not expected.

397 Finally, more research (e.g., local seismic velocity -or attenuation- tomog-
398 raphy studies) is required to accurately identify the location and dimensions
399 of the regions of magma emplacement.

400 **5. Conclusions**

401 Even though the Planchón-Peteroa Volcanic Complex (PPVC) is one of
402 the most hazardous volcanic systems in the Central Andes, knowledge of
403 its internal processes, structures, dynamics, and their relation are still not
404 satisfactorily understood.

405 We apply seismic interferometry by autocorrelations to regional and tele-
406 seismic data recorded by nine stations deployed in the area of the PPVC (six
407 in Argentina and three in Chile) during 2012. The events are selected accord-
408 ing their location, magnitude, angle of incidence of the P-wave energy, the
409 signal to noise ratio on the results, and the related useful frequency range.
410 In order to perform an appropriate description of the subsurface structures
411 below the stations, we use three frequency ranges ([0.3 0.8] Hz, [0.8 2.1] Hz,
412 and [1 3.2] Hz) which are sensitive to different range of frequencies.

413 The smallest frequency range ([0.3 0.8] Hz) is used to infer the tectonic
414 features, i.e., the Moho (at 43-48 km depth), the intra-lithospheric discon-
415 tinuity (~ 70 km), the top and bottom of the subducting slab (~ 120 and
416 ~ 150 -165 km), the lithosphere-asthenosphere boundary (~ 250 km), and the
417 top of the asthenospheric low-velocity zone (~ 340 km). The results support
418 the hypothesis of deformation in the form of detachment, searing, and/or
419 necking for the longitude of the used stations. Our results also suggest a
420 higher depth (~ 15 km) for the bottom of the subducting slab at the north
421 of the PPVC, likely caused by differential deformation along the latitude
422 direction.

423 Based on the results for the two higher-frequency ranges ([0.8 2.1] Hz
424 and [1 3.2] Hz) and previous geological, geochemical, and geophysical infor-
425 mation, we propose a model which describes the structure of the crust and

426 the subsurface regions storing magma bodies down to the Moho. Three
427 regions of sufficiently great volume of magma emplaced at ~ 4 km, ~ 28 km,
428 and ~ 35 km depth, respectively are indicated.

429 The present work provides valuable information about the subsurface
430 conditions of an active volcanic system -the CVPP. We expect the obtained
431 knowledge to be employed in future research aiming to better understand
432 the dynamics of the CVPP.

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574 **6. Figures**

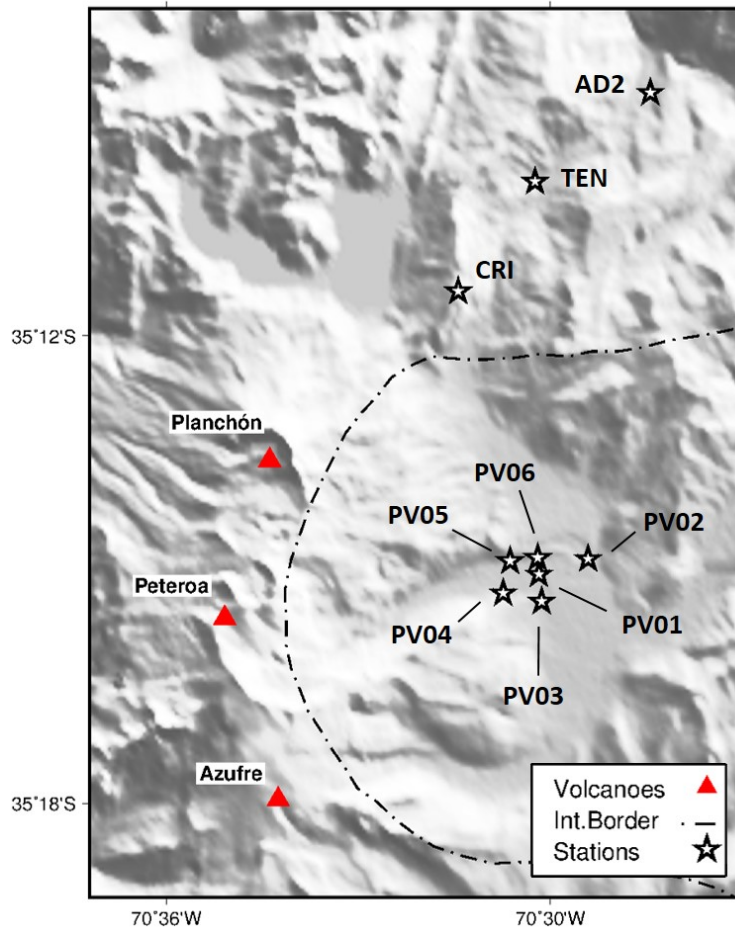


Figure 1: Distribution of the seismic stations used in the present application in relation to the main edifices of the Planchón-Peteroa Volcanic Complex (PPVC).

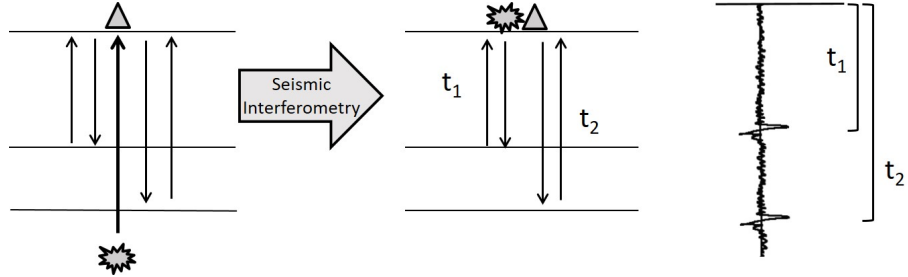


Figure 2: Seismic interferometry by autocorrelation applied to vertically arriving energy in a horizontally layered medium. t_j represents the two-way travel time between the station at the surface and the reflector j in the subsurface. The autocorrelation allows the retrieval of a seismogram composed of reflected energy released by a virtual source co-located at the position of the station.

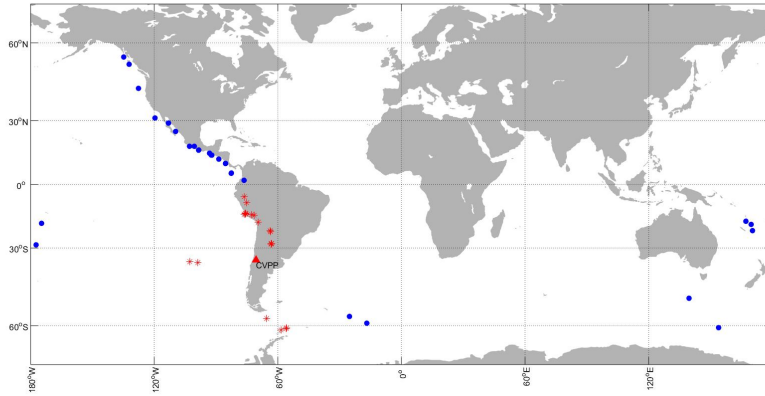


Figure 3: Location of seismic events pre-selected for the application of SIbyA in the area of the PPVC. A triangle indicates the location of the PPVC. Stars show the location of events with epicentral distances less than 30° and magnitudes $M_w > 5$. Circles indicate events with epicentral distances greater than 30° and less than 120° , and magnitudes $M_w > 6$.

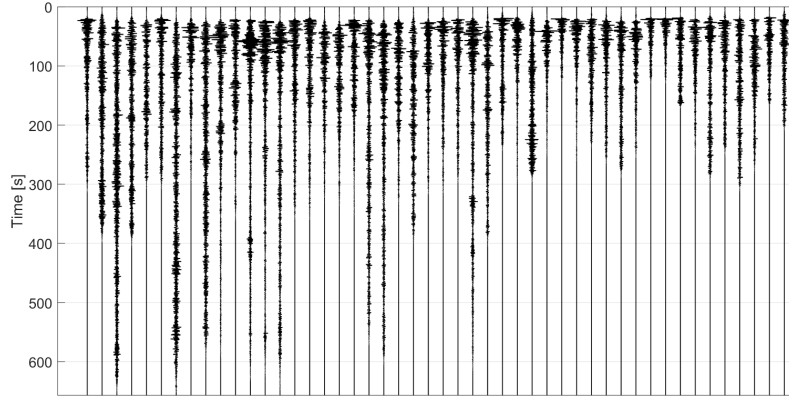


Figure 4: Processing time windows (P-wave codas) for each of the events selected for PV04 station in the complete range of frequencies, i.e., $[0.3 \text{ } 2.1]$ Hz. Each window is normalized according to its vertical energy flux. Vertical axis indicates propagation time. Each window is composed of a pre-event time (20 s) and the times between the first P- and S-wave arrival times.

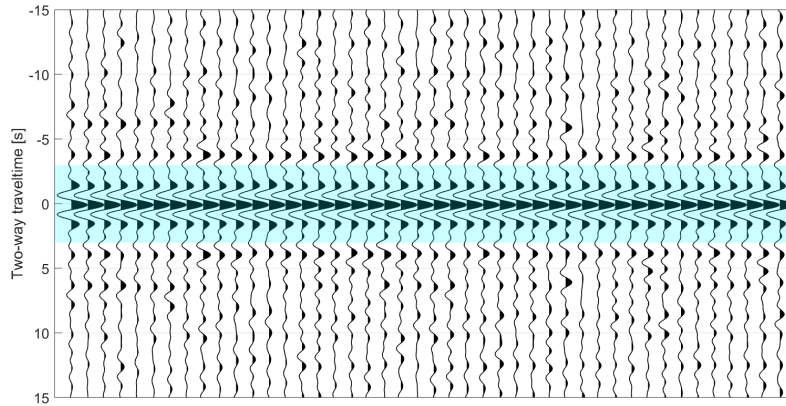


Figure 5: Autocorrelated source time functions (ASTFs) estimated for the station AD2 for the frequency range $[0.3 \text{ } 0.8]$ Hz. A shaded area shows the ASTFs in the autocorrelation panel (for graphical purposes, we only show the first 15 s).

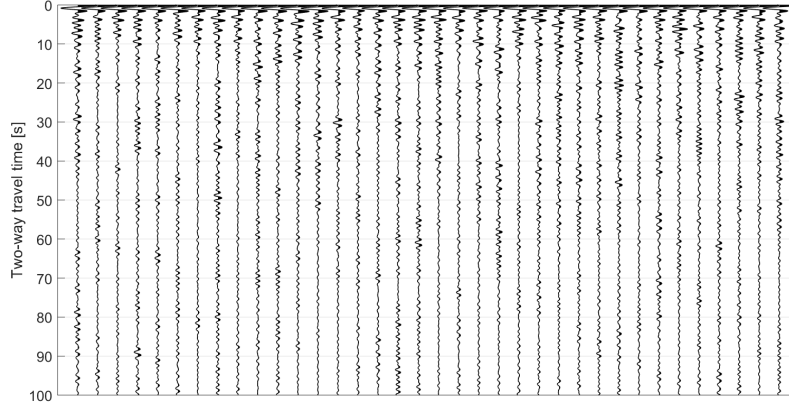


Figure 6: Autocorrelated time windows for the station PV01 in the frequency range [0.3 0.8] Hz. The vertical axis indicates two-way travel time. Each seismic trace is deconvolved by its previously estimated source time function.

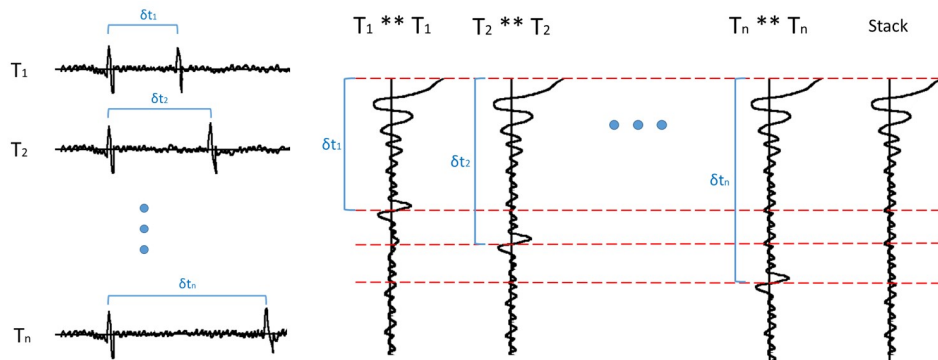
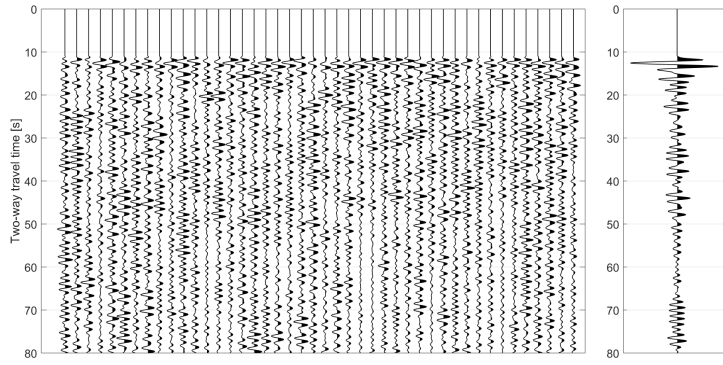
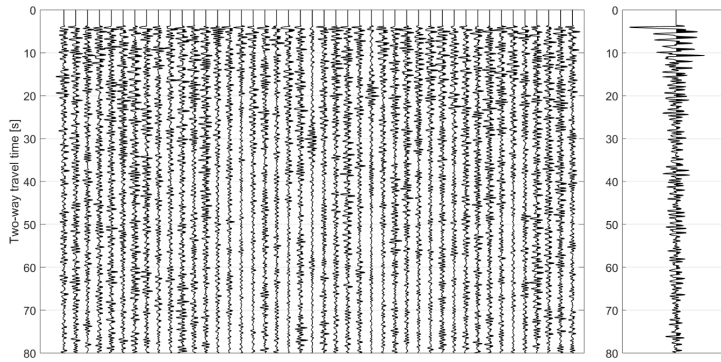


Figure 7: Cartoon illustrating the attenuation of non-physical arrivals originated in the correlation of a time window with several P-wave arrivals. Stacking seismic traces from events with different epicentral distances enhances features located in phase, so that non-physical arrivals due to several P-wave arrivals are attenuated. Without loss of generality, this figure shows the effect of stacking using time windows of events with different epicentral distances, each of them composed of two P-wave phases. T_i is the time window of the event i , which contains two P-wave arrivals separated in δt_i . Operator $**$ means correlation. Dashed lines indicate equal time values.

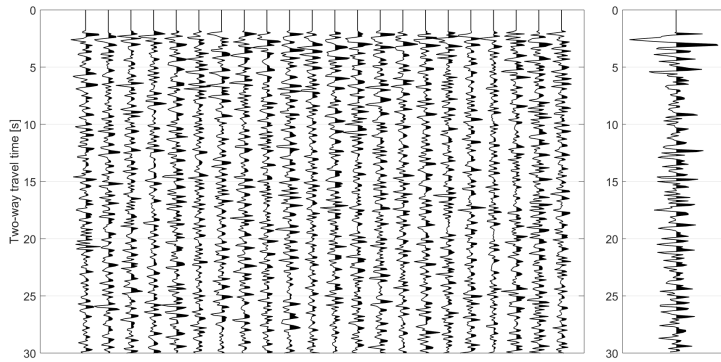
(a) AD2 [0.3 0.8] Hz



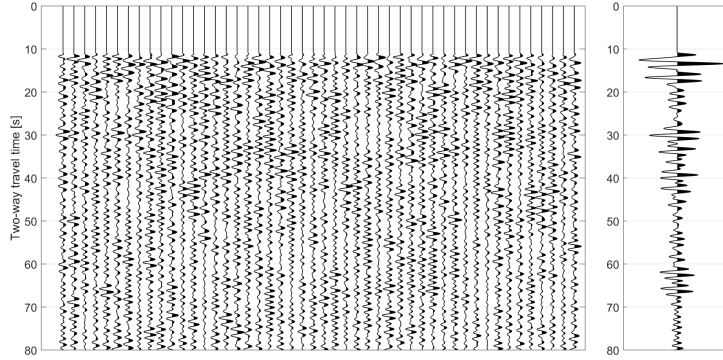
(b) AD2 [0.8 2.1] Hz



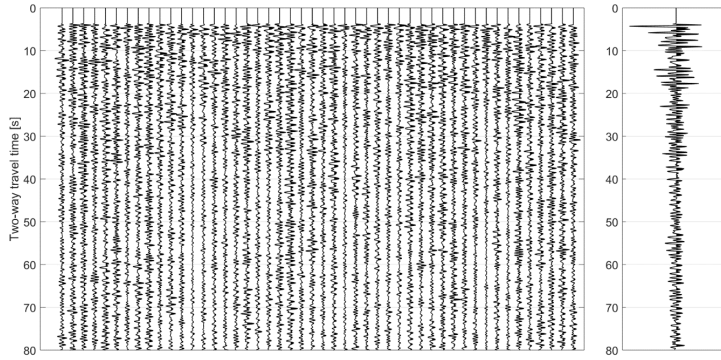
(c) AD2 [1 3.2] Hz



(d) PV04 [0.3 0.8] Hz



(e) PV04 [0.8 2.1] Hz



(f) PV04 [1 3.2] Hz

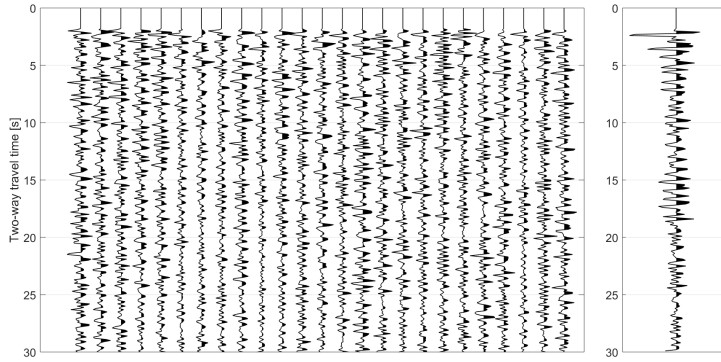


Figure 8: Pre-stacking panels and stacked seismic trace for the stations AD2 (a, b, c) and PV04 (d, e, f), for the frequency ranges [0.3 0.8] Hz (a, d), [0.8 2.1] Hz (b, e), and [1 3.2] Hz (c, f).

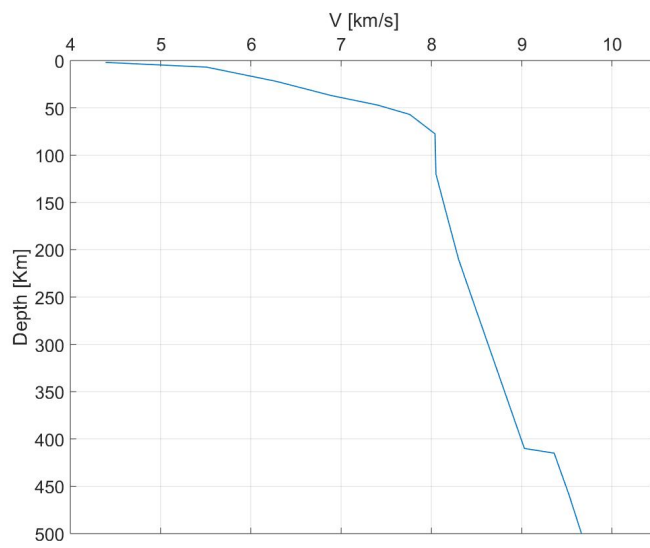


Figure 9: Velocity model used to perform the time-to-depth transformation of the seismic results.

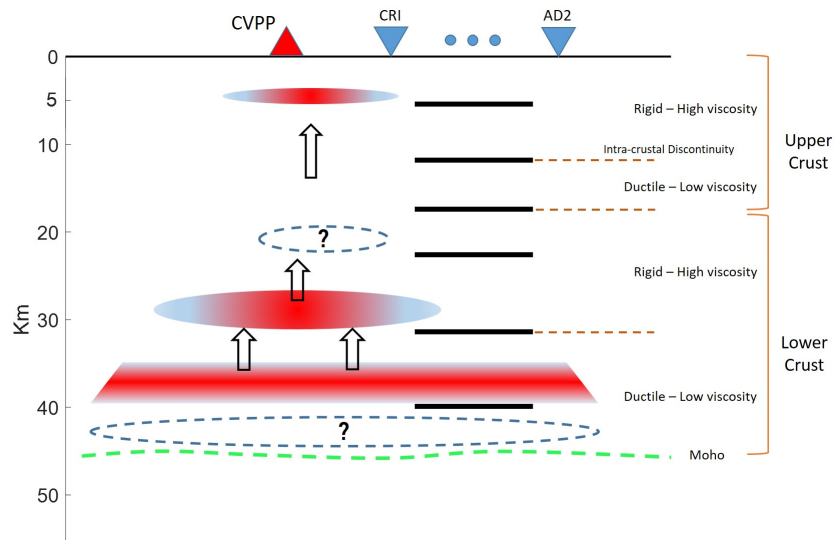


Figure 12: Proposed model of magma emplacement in relation to the structure of the crust down to the Moho in the area of the PPVC. Inverted triangles indicate the longitude of the stations. Thick horizontal lines below the stations show the average depth of the reflectors interpreted in the seismic results. Dashed lines are the interpreted discontinuities (based on [Farías et al. \(2010\)](#) and [Giambiagi et al. \(2012\)](#)) between the different regions of the crust. Arrows show the inferred direction of magma movement. Areas with a question mark inside indicate zones of higher ambiguity in the interpretation.