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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  $\sim$ 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 1. Introduction

The Planchón-Peteroa Volcanic Complex -PPVC- (35.223° S, 70.568° W; 2 see location in Figure 1) is located in the Andes at the international border 3 between Argentina and Chile. The PPVC is composed of three main volcanic 4 edifices, i.e., the Azufre, the Planchón, and the Peteroa, out of which the 5 latter is the current active volcano. The PPVC presents overlapped calderas 6 originated from the destruction of several volcanic structures during past explosive events (Tormey, 1989). Through analyses of its historical activity 8 and products, this volcanic system is ranked as the most hazardous volcano 9 in Argentina (Elissondo and Farías, 2016) and the eighth most risky volcano 10

in Chile (Technical sheet, Observatorio Volcanológico de los Andes del Sur,
OVDAS-SERNAGEOMIN, Chile).

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

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

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

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

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

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

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

#### 83 2. Data

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

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

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

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

#### **3.** Application and results

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

$$\sum_{sources} \left\{ \left[ T(\mathbf{x}_{\mathbf{A}}, -t) * s_i(-t) * T(\mathbf{x}_{\mathbf{A}}, t) * s_i(t) \right] \otimes \left[ s(-t) * s(t) \right]_i \right\}$$
$$\approx -R(\mathbf{x}_{\mathbf{A}}, -t) + \delta(t) - R(\mathbf{x}_{\mathbf{A}}, t) \quad , \quad (1)$$

which states that the reflection response  $R(\mathbf{x}_{\mathbf{A}}, t)$  can be retrieved at 111 the station A located (at  $\mathbf{x}_{\mathbf{A}}$ ) at the surface through the autocorrelation 112 of a recorded transmitted wavefield  $T(\mathbf{x}_{\mathbf{A}}, t)$ . The operator \* indicates 113 convolution,  $\otimes$  means deconvolution, and  $\delta$  is the Dirac's delta. The fac-114 tor  $[s(-t) * s(t)]_i$  corresponds to the autocorrelated source time function 115 (ASTF), which allows the deconvolution of each source time function  $s_i(t)$ . 116 Even though Equation 1 requires sources over the whole stationary phase 117 area (i.e., the Fresnel Zone), seismic events present a non-uniform spatial 118 distribution. Therefore, performing a selection of the seismic sources to be 119 used is essential for a proper application of SIbyA. In order the transmission 120 response of the propagated seismic energy to be accurately estimated by 121 the vertical component of the records, we select only seismic events with 122 P-wave seismic energy arriving (sub) vertically to the station at the surface. 123 The retrieved reflection response (from here on,  $R_v(\mathbf{x}_{\mathbf{A}}, t)$ ) is related to a 124 seismic source co-located to the station at the surface, radiating P-wave 125 energy vertically downwards. 126

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

<sup>138</sup> in the subsurface below the station.

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

#### 147 3.1. Pre-processing

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

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

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

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

Provided the origin time of the selected events, obtained the accurate arrival times, and examined the (sub)vertical incidence of the P-wave energy and high signal-to-noise ratio of the records, we extract the verticalcomponent records of the selected events at each of the used stations.

## 187 3.2. Processing

The vertical-component records of seismic events with P-wave energy arriving vertically at a station represent an accurate estimate of the P-wave transmission response of such propagated wavefield (provided the discontinuities are not excessively inclined; Nishitsuji et al. (2016)).

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

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

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

<sup>217</sup> according to their vertical flux of seismic energy.

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

Figure 6 presents the autocorrelation of the time windows for the station PV01 and the frequency range  $[0.3 \ 0.8]$  Hz, in which each trace is deconvolved by its previously estimated ASTF. This figure shows the dominance of the main lobe in the autocorrelated deconvolved traces. These features close to 0 s are mainly non-physical amplitudes remaining from the deconvolution. Therefore, these amplitudes are removed through windowing.

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

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

#### 243 4. Interpretation and discussion

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

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

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

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

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

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

of the asthenospheric low-velocity zone ( $\sim 340$  km). A small difference in 296 depth is observed for the intra-lithospheric discontinuity and the top of the 297 subducting slab; even though greater depths are observed in the results of 298 the OVDAS stations, these differences would not be significant based on 299 the vertical resolution of the results. A greater difference is observed for the 300 bottom of the subducting slab, i.e.,  $\sim 15$  km greater for the OVDAS stations. 301 Although a dominant positive arrival is expected at the depth of the 302 Moho, a dominant negative amplitude is retrieved in the results for most 303 of the stations. Based on the retrieved waveforms, we interpret the pres-304 ence of a complex area at  $\sim$ 40-55 km depth, causing a perturbation of the 305 amplitudes retrieved for these depths, in particular for those related to the 306 Moho. 307

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

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

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

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

Based on the average depth of the reflectors interpreted in the seismic results, the available scientific information about the subsurface in the PPVC, the proposed structure of the crust for the Central Andes (Farías et al., 2010; Giambiagi et al., 2012), and the physics of magma storage in the crust Jackson et al. (2018), we propose a model for the distribution of magma reservoirs in depth in relation to the main subsurface structures in the crust (see Figure 12).

Through comparison of the average depth of the interpreted reflectors below the stations and the proposed structure of the crust (Farías et al., 2010; Giambiagi et al., 2012), we associate the interpreted reflectors at  $\sim$ 12,  $\sim$ 18, and  $\sim$ 32 km depth as the intra-crustal discontinuity (rigid-ductile discontinuity in the upper crust), the discontinuity between the upper and lower crust, and the rigid-ductile discontinuity in the lower crust, respectively (see Figure 12).

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



Even though no amplitude information is available for depths lower than

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

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

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

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

#### 400 5. Conclusions

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

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

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

Based on the results for the two higher-frequency ranges ([0.8 2.1] Hz and [1 3.2] Hz) and previous geological, geochemical, and geophysical information, we propose a model which describes the structure of the crust and the subsurface regions storaging magma bodies down to the Moho. Three regions of sufficiently great volume of magma emplaced at  $\sim 4$  km,  $\sim 28$  km, and  $\sim 35$  km depth, respectively are indicated.

The present work provides valuable information about the subsurface conditions of an active volcanic system -the CVPP. We expect the obtained knowledge to be employed in future research aiming to better understand 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  $30^{\circ}$  and magnitudes Mw > 5. Circles indicate events with epicentral distances greater than  $30^{\circ}$  and less than  $120^{\circ}$ , and magnitudes Mw > 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.,  $[0.3 \ 2.1]$  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 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  $\delta t_i$ . Operator \*\* means correlation. Dashed lines indicate equal time values.



(a) AD2 [0.3 0.8] 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).

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



Figure 10: Interpretation of the results at the stations of the PV array for the three frequency ranges: (a) [0.3 0.8] Hz, (b) [0.8 2.1] Hz, y (c) [1 3.2] Hz. Filled rectangle areas show the local maximum amplitudes, i.e., the interpreted subsurface discontinuities below each station. Rectangles with dashed line borders indicate a higher uncertainty at the identification of a discontinuity. Discontinuities interpreted only in (c) are marked with a small circle in the left bottom corner of each rectangle. Figure 10c also shows the interpreted discontinuities at depths close to those interpreted in (b).  $L_m$  represent the minimum depth level for the Moho (interp**33**ed in (a)). Arrows indicate zones of likely emplacement of magma.

(a)



Figure 11: Interpretation of the results at the stations of the OVDAS array for the three frequency ranges: (a) [0.3 0.8] Hz, (b) [0.8 2.1] Hz, y (c) [1 3.2] Hz. Filled rectangle areas show the local maximum amplitudes, i.e., the interpreted subsurface discontinuities below each station. Rectangles with dashed line borders indicate a higher uncertainty at the identification of a discontinuity. Discontinuities interpreted only in (c) are marked with a small circle in the left bottom corner of each rectangle. Figure 11c also shows the interpreted discontinuities at depths close  $\mathfrak{H}4$  those interpreted in (b).  $L_m$  represent the minimum depth level for the Moho (interpreted in (a)). Arrows indicate zones of likely emplacement of magma. The dashed arrow represents an uncertainty of interpretation higher than in Figure 10.

(a)



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.