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# Exploring the shallow structure of the San Ramón thrust fault in Santiago, Chile ( $\sim 33.5^\circ$ S), using active seismic and electric methods

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## Abstract

The crustal-scale west-vergent San Ramón thrust fault system at the foot of the main Andean Cordillera in central Chile is a geologically active structure with Quaternary manifestations of complex surface rupture along fault segments in the eastern border of Santiago city. From the comparison of geophysical and geological observations, we assessed the subsurface structure pattern affecting sedimentary cover and rock-substratum topography across fault scarps, which is critical for evaluating structural modeling and associated seismic hazard along this kind of faults. We performed seismic profiles with an average length of 250 m, using an array of twenty-four geophones (GEODE), and 25 shots per profile, supporting high-resolution seismic tomography for interpreting impedance changes associated to deformed sedimentary cover. The recorded traveltimes and reflections were jointly inverted by using a 2-D tomographic approach, which resulted in variations across the scarp axis in both velocities and reflections interpreted as the sedimentary cover-rock substratum topography. Seismic anisotropy observed from tomographic profiles is consistent with sediment deformation triggered by west-vergent thrust tectonics along the fault. Electrical soundings crossing two fault scarps supported subsurface resistivity tomographic profiles, which revealed systematic differences between lower resistivity values in the hanging wall with respect to the footwall of the geological structure, clearly limited by well-defined east-dipping resistivity boundaries. The latter can be interpreted in terms of structurally driven fluid content-change between the hanging wall and the footwall of a permeability boundary associated with the San Ramón fault. The overall results are consistent with a west-vergent thrust structure dipping  $\sim 55^\circ$  E at subsurface levels in piedmont sediments, with local complexities being probably associated to fault surface rupture propagation, fault-splay and fault segment transfer zones.

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## 1 Introduction

The San Ramón Fault (SRF) is a kilometric crustal-scale west-vergent thrust fault system, located east of the highly populated city of Santiago, at the piedmont of the main Andean Cordillera of central Chile (Fig. 1). The SRF is a N–S fault situated along the eastern border of Santiago central valley, formed by N–S to NNW-striking fault segments evidenced by conspicuous Quaternary fault scarps disrupting alluvial sediments of the piedmont units (Armijo et al., 2010). SRF zone has been associated with compressive tectonics that raised the main Andean Cordillera with respect to the central valley of Santiago, likely since the Miocene in central Chile (Armijo et al., 2010; Farías et al., 2010; Rauld, 2011).

Recent tectonic activity has been evidenced through the analysis of 5 m-high fault scarps affecting Late Quaternary alluvial units (Fig. 1), according to structural-geomorphological work (Armijo et al., 2010; Rauld, 2011) and ongoing paleoseismological studies from trenches (Vargas and Rebolledo, 2012). Alluvial piedmont sediments in this area, located at around 800–900 m.a.s.l., correspond mostly to coarse massive and poorly sorted debris flow deposits and localized relatively well-sorted lenses of channel-flow deposits in proximal areas of alluvial-fan geomorphologic units. These units are disposed at the foot of the mountain front, which constitutes the western limit of the main Andean Cordillera composed mostly by Cenozoic volcanic rocks, reaching up to 3240 m.a.s.l. at the San Ramón hill.

The objective of the study herein is to assess the subsurface surroundings of the SRF, from the shallow velocity ( $V_p$ ) structure using 2-D joint refraction and reflection seismic tomography (Korenaga et al., 2000). This method allows inverting directly the seismic velocities and thicknesses of a single layer when refractions and reflections are available. We record refractions through the shallow sedimentary layer and reflections from the top of the volcanic rock substratum assigned to Abanico Formation (Cenozoic). Thus, velocities and substratum topography are simultaneously derived. Along with the active seismic experiment, electrical resistivity measurements along two pro-

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files crossing fault scarps of the SRF provide us information on subsurface electrical properties at both sides of the fault zone. Determination of seismic velocities and electrical resistivity of the sedimentary layers above the substratum together with their behavior across the fault zone, allow improving the understanding about the geometry and Quaternary tectonics associated to the SRF in the eastern border of the highly populated Santiago city.

## 2 Geophysical measurements

### 2.1 Seismic experiment description

The seismic experiment was carried out in September–October 2011 and comprises four seismic lines. The purpose of the experiment was to detect small-scale structures with wavelengths smaller than 500 m, bellow one of the youngest fault scarps recognized along the SRF (Armijo et al., 2010). Thereby we employed a large number of shots with geophone spacing ranging from 5 to 10 m.

Figure 2 shows the location of the four seismic profiles (P1 to P4), which were designed so as to study the 5 m high fault scarp. The objective is to detect possible velocity changes from the west to the east of the scarp, which runs approximately N–S (Figs. 1 and 2). P1 and P4 cross the scarp axis, while P2 and P3 run parallel to the scarp axis, eastward and westward of it respectively (see Fig. 2b). Table 1 summarizes the experiment details. It is worth noting that eastward of the study area there are urban roads and electric power towers that limit the prolongation of the seismic lines further east (see Fig. 2a).

The seismic equipment employed comprises a 24-channel GEODE recorder, a seismic cable with a 10 m-separation among channels and standard vertical geophones with a natural frequency of 14 Hz. The topography used for our seismic modeling was provided by differential GPS data (Armijo et al., 2010; Rauld, 2011). As a seismic source, we used a 12-pounds hammer hitting an iron sphere, which allowed us to

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record refractions imaging up to 50 m in depth and reflections mapping the top of the substratum at depths of 80–100 m.

In order to generate a detailed tomography across the scarp, the central part of profiles P1 and P4 were measured using a geophone spacing of 5 m. On the other hand, since the north-south topographic gradients are small, profiles P2 and P3 were registered using a nominal geophone spacing of 10 m. These considerations allow recording of reflections from the top of the substratum. As a result of the dense vegetation in the studied zone, it was necessary to slightly change the direction of the seismic lines, which did not affect our 2-D modeling at the spatial-scale analysis.

## 2.2 *P* wave velocity modeling

### 2.2.1 Forward modeling

In order to derive an initial model for the 2-D tomographic inversion, we first modeled the 1-D velocity structure for individual shots in each seismic profile. The 1-D velocity model consists of a combination of homogeneous and constant-velocity gradient layers in which thicknesses and top and bottom velocities can be interactively modified to adjust the traveltimes of different seismic arrivals seen in the data (Vera et al., 1990). Figure 3 shows examples of the seismic records associated with individual shots, including the predicted traveltimes from the 1-D model. 1-D models of each single shot registered in profiles P1 and P4 does not show significant variation concerning the velocity structure with depth across the lines. Hence for these two profiles we use the 1-D model of the shot showing the best signal (Fig. 3a and f) as a reference to build the initial model of the 2-D tomographic inversion. On the other hand, the 1-D models calculated using the shots located near the ends of the lines P2 and P3 show variations in the velocity structure with depth in the north-south direction (see Fig. 3b and c for the profile P2, and Fig. 3d and e for the line P3). Therefore we use two 1-D models in each profile, as a reference for generating initial models for the 2-D inversion.

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## 2.2.2 2-D traveltimes inversion

We obtained the  $P$  wave velocity-depth structure based on the joint refraction and reflection traveltimes inversion method of Korenaga et al. (2000). This method allows joint inversion of seismic refraction and reflection traveltimes data for a 2-D velocity field. Traveltimes and raypaths are calculated using a hybrid ray-tracing scheme based on the graph method and the local ray-bending refinement (van Avendonk et al., 1998). Smoothing constraints using predefined correlation lengths and optimized damping constraints for the model parameters are employed to regularize an iterative linearized inversion (Korenaga et al., 2000).

The 2-D initial models are based on the 1-D forward modeling previously calculated (see Sect. 2.2.1). A Digital Elevation Model was obtained from topographic data resulting from D-GPS measurements taken in the area. The top of the substratum and sedimentary velocities are jointly inverted using refractions through the sediment layer ( $P_s$ ) and reflections from the top of the substratum (PbP). The horizontal grid spacing of the model used for the velocity inversion is 1 m, whereas the vertical grid spacing varied from 0.2 m at the top of the model to 0.5 m at its bottom. Depth nodes defining the reflector for the joint inversion (top of the substratum) are spaced 5 m apart. We used horizontal correlation lengths (average-smoothness window) ranging from 0.5 m at the top to 2 m at the bottom of the model, and vertical correlation lengths varying from 0.2 m at its top to 1 m at its bottom. Different tests showed that varying the values of correlation lengths by 50 % does not significantly affect the results. As a result of the trade-off between correlation lengths and smoothing weights, we tried to apply shorter correlation lengths and larger smoothing weights to reduce memory requirements (Korenaga et al., 2000; Contreras-Reyes et al., 2008). Depth and velocity nodes are equally weighted in the refraction and reflection traveltimes inversions. Table 2 summarizes the details regarding the final velocity model inverted.

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## 2.3 Electrical resistivity measurements

DC geoelectrics is one of the earliest geophysical exploration techniques, going back to the first decades of the last century. It basically consists in feeding a (direct) current into the ground and measuring the resulting voltage drop. The vertical electrical sounding (VES) perhaps is the most commonly used strategy in the application of DC electrical methods.

The VES is based on the concept that the electrical structure of the earth can be described by a one-dimensional resistivity function,  $\rho = \rho(z)$ , with the resistivity of a rock varying only with depth within the earth. Assuming that the lateral change in resistivity of the earth described by  $\rho = \rho(x, y)$  is so slow at any one location, the profile  $\rho(z)$  can be determined independently of the lateral variations,  $\rho(x, y)$ . Then, the lateral changes can be determined by a series of vertical electric soundings made along a profile-or over an area-and interpreted as a set of slowly changing  $\rho(z)$  functions. For more details on VES, see Zhdanov (2009).

In practice a VES consists of the injection of a direct current through two electrodes connected to the ground and of the measurement of the corresponding electrical potential through two other electrodes, following a symmetrical setting (e.g. Wenner, Schlumberger, etc). The concept of sounding relates to the increment of the current electrode spreading, which allows the inference of the electrical properties of the deeper layers of the studied zone. The use of Schlumberger settings for determining the subsurface resistivities is widespread in geophysical applications, mainly because of its simplicity and speed of survey.

During our work a 2-D electrical resistivity imaging was performed along with the normal VES measurements. Electric Resistivity Tomography (ERT) is a proven imaging technique and both its theory and application are well documented in geophysical research literature (e.g. Griffiths and Barker, 1993). It has been proven that ERT is a useful tool in mineral exploration, as well as in hydrogeology or exploration of sed-

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imentary basins (e.g. van Schoor and Duvenhage, 2000; van Schoor, 2002; Colella et al., 2004).

## 2.4 Data acquisition and processing for ERT measurements

The multi-electrode resistivity technique uses multi-core cables (ABEM SAS 1000 unit) with as many conductors (between 81 and 121 in the study herein) as electrodes are plugged into the ground at fixed spacing. The two-location survey was carried out using a symmetric Wenner–Schlumberger configuration. The unit electrode spacing was 7 m along the two profiles (see Table 3). The location of both profiles can be seen in Figs. 2 and 4.

To calculate the resistivity of the material, the electrical potential difference created by an electrical current passing through the ground via the conductors is measured. The different combination of current and potential pairs of electrodes results in the mixed sounding and profiling section with the corresponding maximum depth of investigation, which in turn depends on the length of the cable and the type of configuration used for the survey. The changes of resistivity related to depth allow the construction of a 2-D section of the subsurface resistivity values.

These data were inverted using the RES2DINV inversion routine, which uses a non-linear smoothness-constrained least-squares technique to calculate the resistivity of the model blocks (de Groot-Hedlin and Constable, 1990; Sasaki, 1992; Loke et al., 2003), and a finite-element approach to generate apparent resistivity values. An optimization process tries to reduce iteratively the difference between the calculated and measured apparent resistivity values. The percent error, considering  $N$  points of comparison for each data set, was calculated as a data fit indicator. For a comparison of measured and calculated apparent resistivity pseudosections along profiles L1 and L2, see Appendix (Figs. A3 and A4).

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### 3 Results

#### 3.1 Seismic measurements

The 1-D velocity-depth models (Fig. 3) provide key information to define the general structure of the sedimentary layer. These models show typical sedimentary velocities with moderate compaction ( $V_p < 2 \text{ km s}^{-1}$ ). A shallow layer with  $V_p \sim 0.7 \text{ km s}^{-1}$  is observed, probably made of soft soil and poorly compacted sediments. Deeper, the 1-D models show an increase of velocity indicating the increase of sediment compaction. A strong velocity discontinuity detected in the 1-D velocity models was interpreted as a reflector associated with the sedimentary cover-volcanic rock substratum interface. The modeled reflector becomes systematically shallower northward (perpendicular to the direction of the topographic gradient in the region) as it is observed in Fig. 3. This geometry change in the substratum-reflector topography is clearer in the 2-D tomographic models (Figs. 5–8), but it is remarkable that the simple 1-D modeling provides information about its shape's differences.

The 2-D velocity-depth models present more detailed images, including topographic effects and velocity heterogeneities (Figs. 5–8). The latter was possible due to the large number of shots and the narrow spacing between geophones in accordance with the spatial-scale of the analysis (1–100 m). The main observations from the 2-D tomographic profiles are:

*P1 and P4 (Figs. 5 and 6):* both profiles cross the fault scarp in an approximate E–W direction. In profile P1 (Fig. 5), located in the southern part of the study area, the velocity-depth model shows a decrease in the vertical gradient of velocity marked by the iso-velocity contour of  $1.5 \text{ km s}^{-1}$  (segmented red line in Fig. 5a). This change is interpreted as the transition from a poor compacted shallower sedimentary unit to a more competent layer at depth. The location of the SRF from paleoseismological studies in trenches (Vargas and Rebolledo, 2012), is indicated by a continuous black line. In the hanging wall of the SRF fault scarp the interpreted shallower sedimentary unit shows a homogenous thickness and internal velocity gradient, while in the foot-

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5 wall this unit becomes thicker and its internal vertical gradient of velocity decreases. A localized deepening of the iso-contours of velocities at  $x \sim 40$  m in the footwall of the fault scarp suggests a complexity in the subsurface soil structure that could be associated with another fault. The deepest reflector (blue line) interpreted as the change between sedimentary-cover and substratum-rocks, shows a similar inclination with respect to the topography. The eastward shallowing of this interpreted substratum could be related to the presence of the SRF, but the distribution of the reflected rays does not allow to properly define the shape of it with a high enough resolution to support this hypothesis (Fig. 5b).

10 Profile P4 is parallel to P1 and it is located in the northern part of the study area. Seismic velocities along P4 (Fig. 6a) are similar to those observed along P1. The iso-velocity contour of  $1.5 \text{ km s}^{-1}$  becomes remarkably deeper westward. A localized deepening of the iso-contours of velocities at  $x \sim 40$  m suggests, as in profile P1, local complexity that could be associated with another thrusting. The sediment–substratum interface was detected only by reflections in the central portion of the profile (Fig. 6b), but its trend indicates a relatively constant depth of  $\sim 70$  m below the topography. The smooth shape of the substratum, well defined by reflections, suggests that this is not cut by the SRF in this short segment of the profile P4. For more details on the estimation of the substratum depth, see Appendix (Fig. A2).

20 *P2 and P3 (Figs. 7 and 8)*: both profiles are located parallel to the fault scarp. Profile P2 (Fig. 7) is located eastward from the scarp, in the hanging wall of the observed fault at surface and it strikes approximately in a N–S direction. The velocity-depth model is well explained by a combination of flat layers (1-D) with a slight increase in velocities around  $x = 40$  m (Fig. 7a). The main feature of this model is the clear reduction in depth of the reflector around a distance of 80 m along the profile. This observation is well constrained by the distribution of reflected rays which support a smooth deepening southward of the substratum.

25 Profile P3 (Fig. 8) is located westward from the scarp, in the footwall of the observed fault at surface and like P2 it is well represented by a simple combination of flat layers

that present a local increase in velocities around 80 m distance along the profile. The slight variations seen along profiles P2 and P3 can be explained by differences in the amount and compaction of sedimentary flows sourced mainly from the east. Similar to P2, the velocity depth model along P3 shows a southward deepening of the substratum with a 20 m variation in depth between the southern and northern segments of the profile.

### 3.2 Resistivity measurements

ERT experiments resulted in two profiles, one crossing the Late Quaternary fault scarp of the SRF where seismic profiles were realized (L1, Quebrada Macul area, Fig. 2), and one crossing a more evolved fault scarp which probably represents a longer history of the SRF during the Quaternary (L2, La Reina area, Fig. 4). The inversion of ERT data along the L2 profile resulted in a model with resistivity values between 1 and 600  $\Omega\text{m}$  (see Fig. 9), and a percent error of 4.4 % between measured and modeled data (see Fig. A3 of the Appendix). The resistivity distribution in this profile shows clearly two different areas. The western part, from 0 to 280 m along horizontal distance presents much higher resistivity values (100–600  $\Omega\text{m}$ ) compared with the eastern part of the profile (280–530 m), where the observed resistivity values were as low as 5–50  $\Omega\text{m}$ . The resistivity contrast at 280 m appears clearly in the ERT section. This change can be observed at depth in this profile, defining a sharp lateral resistivity contrast-zone dipping 60–54° E, which can be interpreted like structural control on electric properties associated to the SRF.

Two VES using Schlumberger array (200 m maximum spread) were carried out at the eastern and western part of the L2 profile (Fig. 4), to check the strong resistivity discontinuity observed there. The results confirmed the observations from profile L2, with resistivity values between 100 and 200  $\Omega\text{m}$  at the western part of the profile and of 10–20  $\Omega\text{m}$  at its eastern part (see Appendix, Fig. A5).

The inversion of ERT data along the L1 profile resulted in a model with a percent error of 2.8 % (see Figs. 10, and A4 of the Appendix). The resistivity distribution along

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5 indicating a relative higher compaction to the east of the structure (Fig. 2). This E–W change in vertical  $V_p$  gradients in the western side with respect to the eastern side of the fault scarp, can be interpreted like the result of thicker poorly compacted alluvial sediments in the footwall with respect to the hangingwall of the fault, and could be associated with the development of a syntectonic basin resulted from Late Quaternary kinematics along the SRF. Thus, the overall observations are consistent with sedimentary units deformed by an inverse west-vergent fault at subsurface levels. In addition, the substratum geometry interpreted from the seismic experiments suggests that at deep levels ( $\sim 60$ – $80$  m), the San Ramon fault presents north-south variations, probably associated with fault transfer zones, and that this structure could strike NNW, as indicated by the abrupt change in the depth of the substratum observed in profiles P2 and P3, from which a  $\sim N30^\circ W$ -striking structure may be deduced. In addition, these variations in substratum depth along profiles P2 and P3 could be also associated with erosional surfaces produced by ancient streams of from the main quebradas in the area.

15 The electrical resistivity values obtained from two profiles crossing the San Ramon fault zone vary from 10 to 800  $\Omega m$ . The L2 profile shows a strong lateral variation, with values of 10–20  $\Omega m$  at the eastern side of the fault zone, interpreted close to the base of the fault escarpment, and of 100–200  $\Omega m$  at its western side. The effect of porosity and fluid presence in the electrical resistivity of rocks has been widely observed, particularly in sedimentary environments and also in fault zones (e.g. Caputo et al., 2007; Colella et al., 2004; Unsworth et al., 2000), confirming the strong correlation between fluid presence in highly porous media and high (electrolytic) conductivity. As water content is one of the parameters that strongly control the resistivity both in sedimentary layers and in fractured substratum rocks, this difference could be related to diminished permeability at the fault core zone, acting as a barrier for fluids possibly due to localized slip along the fault planes at shallower levels (Caine et al., 1996), as well as to a lithological change with shallower substratum in the hanging wall with respect to the footwall of the fault. This could cause a zone of low resistivity by trapping



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groundwater uphill of the fault zone keeping the sedimentary alluvial layers dryer in the footwall due to a sharp drop of the water-table, which results in higher resistivity values immediately westward of the geologic structure. The influence of the SRF in the water control and transport down to the valley can be also inferred from the location of the Baños de Apoquindo spring (see Fig. 1), very close to the fault scarp in that area.

A similar behavior can be interpreted from the L1 profile, where lower resistivity values at depth tend to remain at the eastern side of the profile, while at the western side there is scarcely any conductivity increment with depth, which can be interpreted like structural control on the groundwater system resulting in systematically higher water-tables in the hangingwall with respect to the footwall of the structures. Besides, four prominent features are observed along profile L1. At the easternmost part of this profile, a sharp lateral resistivity change can be interpreted as a structure dividing zones with high (downhill) and low (uphill) resistivity values below 30 m depth, similarly to the structure interpreted from the L2 profile; in both cases those structures are located at the base of the main topographic fault escarpments associated to the San Ramón fault (“A” in Figs. 9 and 10). The westernmost structures observed in profile L1 (“C” and “D” in Fig. 10), are characterized by relative increased conductivity, which probably reflects localized increased permeability resulting from recent surface ruptures along faults in the area, in accordance with complexities that characterizes permeability changes in fault zones (Caine et al., 1996). One of those structures (“C”, Fig. 10), corresponds to an observed fault affecting recent sediments, according to ongoing paleoseismological studies (Vargas and Rebolledo, 2012), showing a remarkable association with shallow seismic results as well (see Fig. 11). If the reflection located around 70 m below de surface topography here, is effectively associated to the volcanic rock substratum-sedimentary cover transition (P4, Fig. 10), it is possible to interpret that the low resistivity unit characterizes relatively humid sediments confined bellow a dry surface sediment layer of about 50 m thickness at the location of the young fault scarp, and above a highly fractured substratum characterized by higher Vp and lower electrical resistivity (Fig. 10).

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The comparison between the paleoseismological trench with the seismic and resistivity tomographic profiles (Figs. 10 and 11), reveals a change in the dip of the fault between the most surface levels and deeper zones. The fault was observed dipping 17–20° E at surface (Fig. 11), but it quickly evolves to a fault dipping around 55° at depths greater than 10–20 m, according our interpretation from the seismic and resistivity profiles (Figs. 10 and 11), and in agreement with the dominant dip of the SRF proposed by Armijo et al. (2010). Similarly, other inferred structures (“D” and “B”) shown in L1 (QM-resistivity profile, Fig. 10), could be associated with local complexities like fault splay and fault segment transfer zones (Figs. 1 and 2).

## 5 Conclusions

We interpreted seismic and resistivity profiles across one of the youngest and evolved Quaternary escarpments of the San Ramón Fault, which evidenced clear localized Vp and resistivity changes at subsurface associated with main and secondary structures. Seismic and electric models are consistent with sedimentary units deformed by reverse west vergent localized faults, dipping around 55° E at depth, cutting the piedmont sedimentary cover and substratum-rocks, and possibly controlling the position of the water-table in the eastern border of the Santiago valley. The geophysical observations suggest also local complexities that can be associated with fault splay or fault segment transfer zones.

Given the acute urbanization in the eastern border of the city of Santiago, future measurements should focus on a better coverage along the fault zone through geophysical techniques, which together with detailed paleoseismological and geomorphological results will improve critic knowledge concerning the geometry, Quaternary kinematics and potential seismic hazard related to the San Ramón Fault. This information will be extremely valuable in order to assess a better territorial planning in the still growing city of Santiago.

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**Table 1.** Specifications of the seismic experiment.

Profile	Orientation	Length [m]	Number of shots	Geophone spacing [m]
<b>P1</b>	WE	269	27	5 and 10
<b>P2</b>	SN	388	17	10
<b>P3</b>	SN	373	17	10
<b>P4</b>	WE	271	17	5

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**Table 2.** Summary of travel time picks and details of the average final velocity-depth model shown in Figs. 5–8. The nomenclature is as follows: Ps (refractions), PbP (reflections from the top of the substratum),  $\Delta T$  (average travel time uncertainty), and TRMS (root mean square travel-time misfit).

Profile	Ps ( $\Delta T$ ) [ms]	PbP ( $\Delta T$ ) [ms]	Ps (TRMS) [ms]	Ps + PbP (TRMS) [ms]
<b>P1</b>	5	10	3.42	5.15
<b>P2</b>	5	10	4.81	4.68
<b>P3</b>	5	10	5.57	5.49
<b>P4</b>	5	10	4.31	4.80

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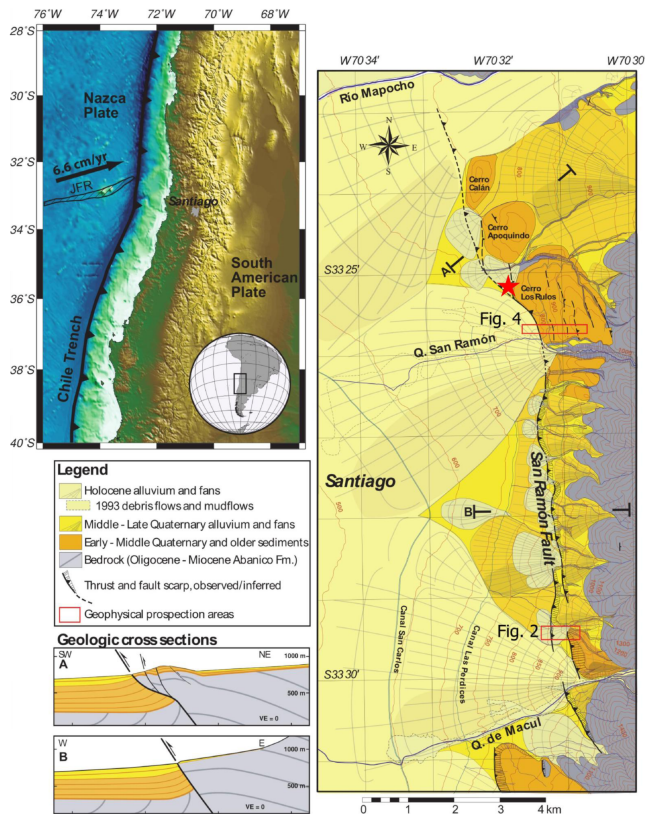
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**Table 3.** Specifications of the ERT measurements.

Profile	Orientation	Length [m]	Electrode spacing [m]
<b>L1</b>	WE	840	7
<b>L2</b>	WE	560	7



**Fig. 1.** Geological chart and cross sections of the study zone showing the San Ramón fault system along the eastern border of Santiago city, and location of the geophysical measurements (modified from Armijo et al., 2010). Location of geophysical measurements is marked by red rectangles, enclosing the areas shown in Fig. 2. (Quebrada de Macul area) and Fig. 4 (La Reina area). A red star marks the position of the Baños de Apoquindo spring.

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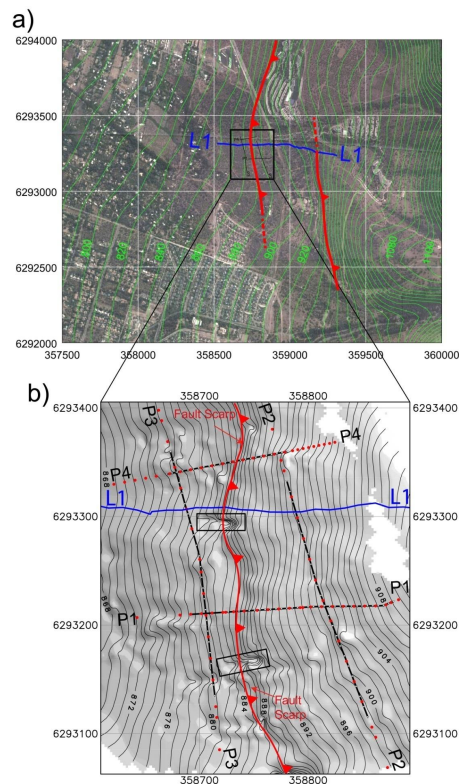
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**Fig. 2.** Location of seismic (P1–P4) and electric (L1) profiles at the Quebrada de Macul area. Green lines in (a) are contour lines of the topography based on SRTM data. Black lines in (b) are topographic contour lines based on D-GPS data, red dots mark shot locations and white triangles along black lines represent geophone positions. Black boxes mark the position of trenches used for ongoing paleoseismological studies in the area (Vargas and Rebolledo, 2012). Red line represents the approximate location of the SRF at surface according Armijo et al. (2010). Coordinate system UTM zone 19S, WGS84.

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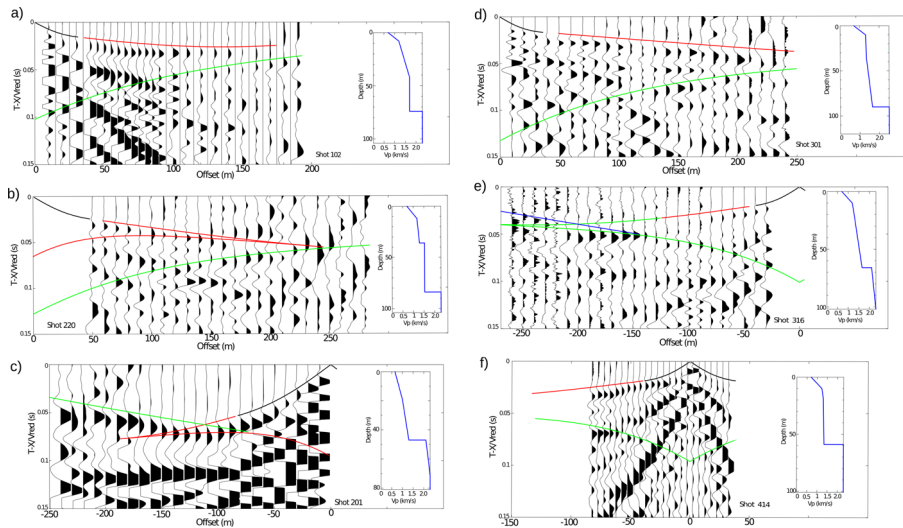
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**Fig. 3.** (a) Shot at  $-10$  m, profile P1. (b) Shot at  $-50$  m, profile P2. (c) Shot at  $240$  m, profile P2. The reduction velocity is  $1.5 \text{ km s}^{-1}$ , and the horizontal coordinate is shot-geophone distance (offset). (d) Shot at  $-10$  m, profile P3. (e) Shot at  $260$  m, profile P3. The reduction velocity is  $1.5 \text{ km s}^{-1}$ , and the horizontal coordinate is shot-geophone distance (offset).

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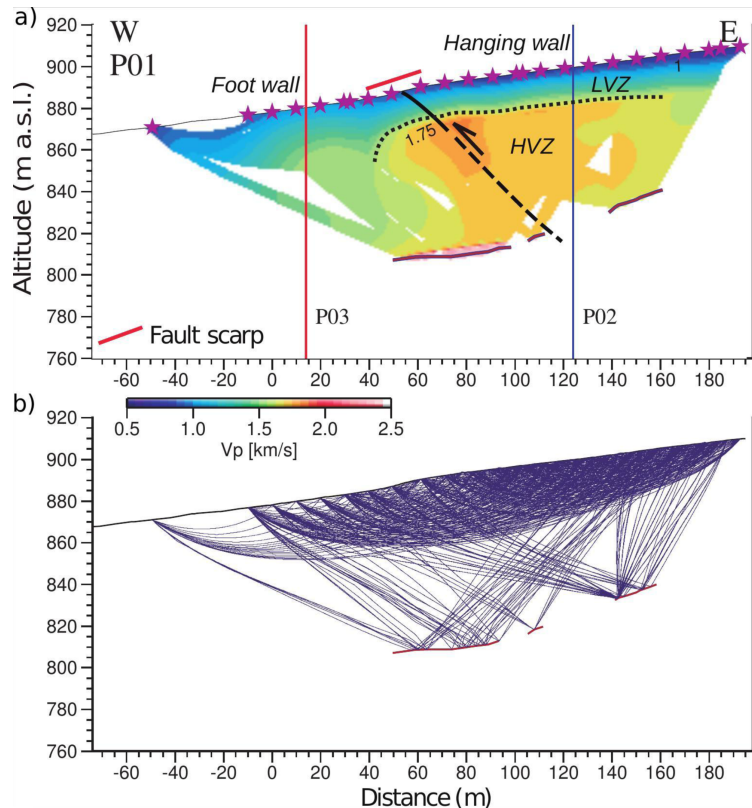




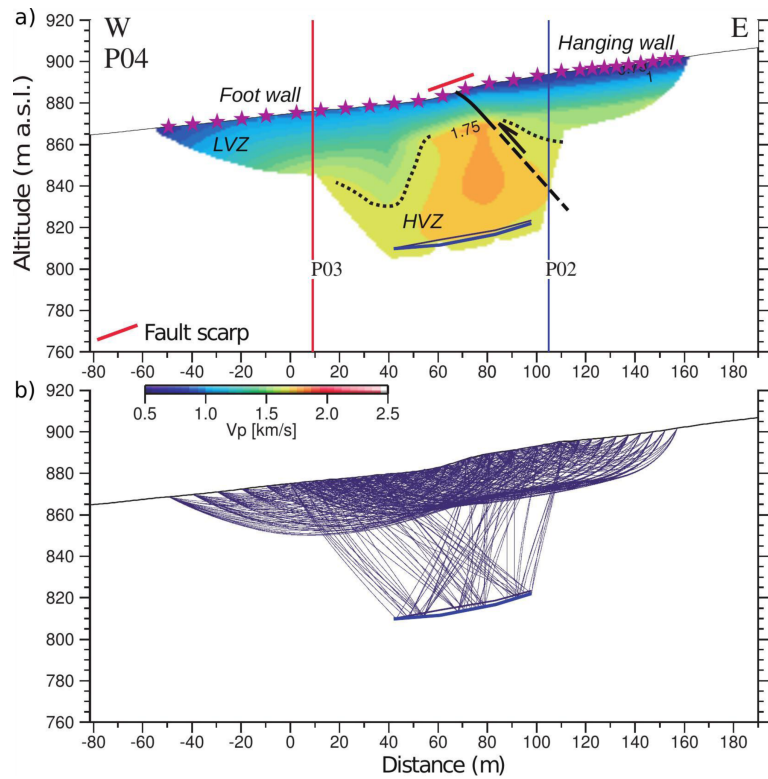


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**Fig. 5.** Results of seismic profile P1. **(a)** Final velocity model including the inverted reflector (blue line). Stars represent shot positions. Zero distance represents position of the first geophone along the profile. Black line represent the SRF system derived from paleoseismological studies (Vargas and Rebolledo, 2012), while dashed black lines are inferred structures related to the SRF. Dashed red line represents shallow sedimentary units. The intersections between P1 with profiles P3 and P2 are also indicated (see Fig. 2). **(b)** Associated ray-paths.



**Fig. 6.** Results of seismic profile P4. **(a)** Final velocity model including the inverted reflector (blue line). Stars represent shot positions. Zero distance represents position of the first geophone along the profile. Black line represent the SRF derived from paleoseismological studies (Vargas and Rebolledo, 2012), while dashed black lines are inferred structures related to the SRF. Dashed red line represents shallow sedimentary units. The intersections between P4 with profiles P3 and P2 are also indicated (see Fig. 2). **(b)** Associated ray-paths.

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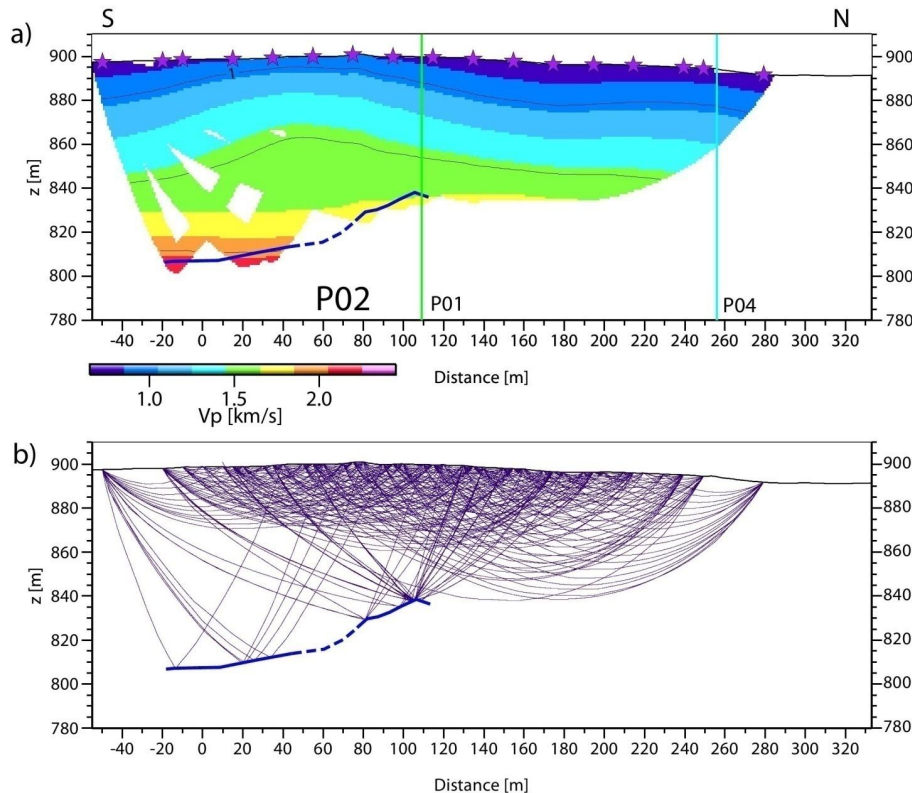
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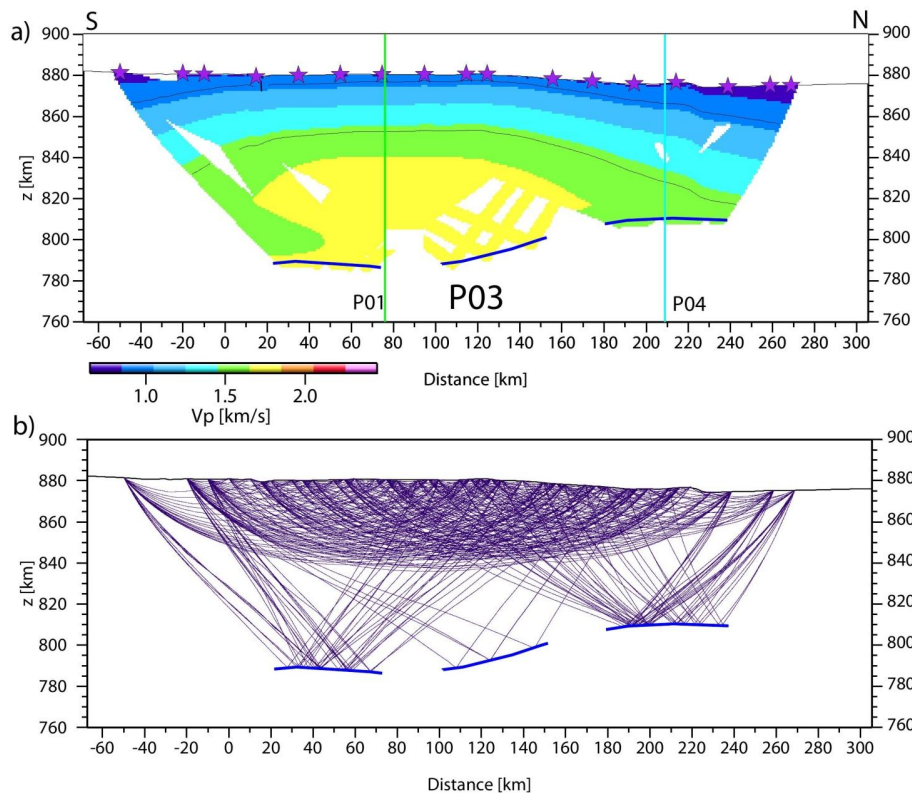
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**Fig. 7.** Results of seismic profile P2. **(a)** Final velocity model including the inverted reflector (blue line). The segmented blue line indicates the regions where the reflector is poorly constrained. Stars represent shot positions. Zero distance represents position of the first geophone along the profile. The intersections between P2 with profiles P1 and P4 are also indicated (see Fig. 2). **(b)** Associated ray-paths.

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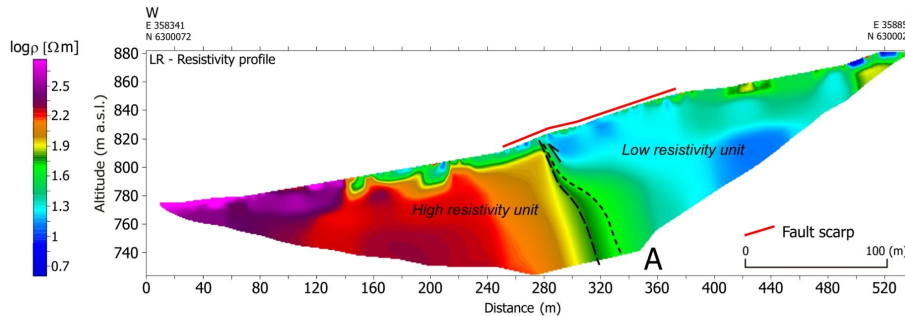
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**Fig. 8.** Results of seismic profile P3. **(a)** Final velocity model including the inverted reflector (blue line). Stars represent shot positions. Zero distance represents position of the first geophone along the profile. The intersections between P3 with profiles P1 and P4 are also indicated (see Fig. 2). **(b)** Associated ray-paths.

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**Fig. 9.** Inversion result along the La Reina (L2) profile, including main structures interpreted in the text. Resistivity values show a strong contrast between the eastern (conductive) and western (resistive) part of the profile.

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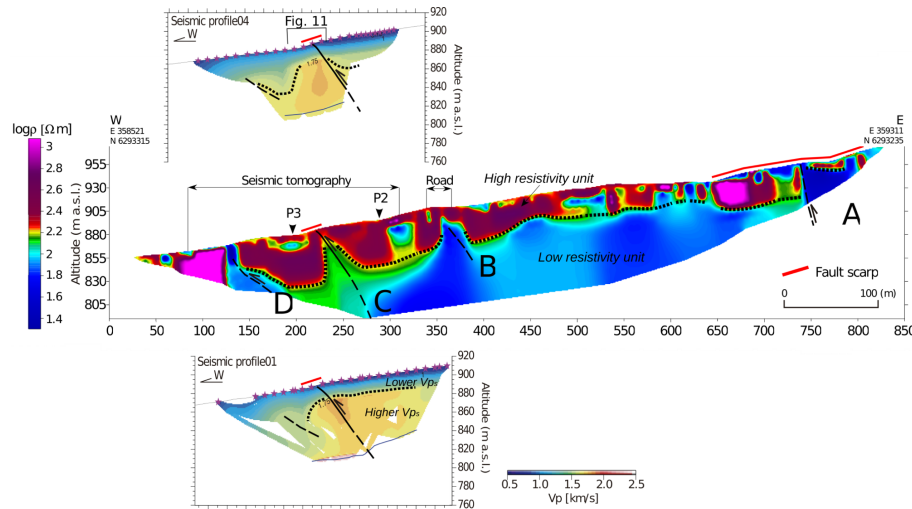
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**Fig. 10.** Inversion result close to along the Quebrada de Macul (L1) profile, including main structures interpreted in the text. Lower resistivity values at depth tend to remain at the eastern side of the profile. Note that the structure marked as A could be the same structure observed in both zones (see Fig. 9). Results from seismic profiles P01 and P04 have been added for comparison. See zoomed view of P04 together with a corresponding outcrop from ongoing paleoseismological studies in Fig. 11.

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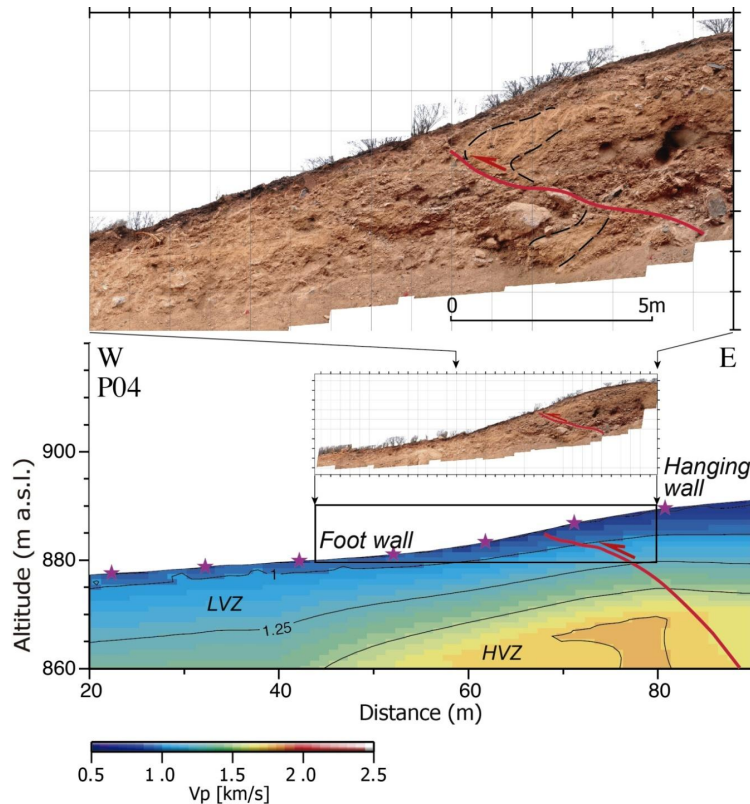
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**Fig. 11.** Surface observation of a fault affecting recent sediments close to seismic profile P04, from ongoing paleoseismological studies (Vargas and Rebolledo, 2012), together with a zoomed view of the results obtained along this profile.

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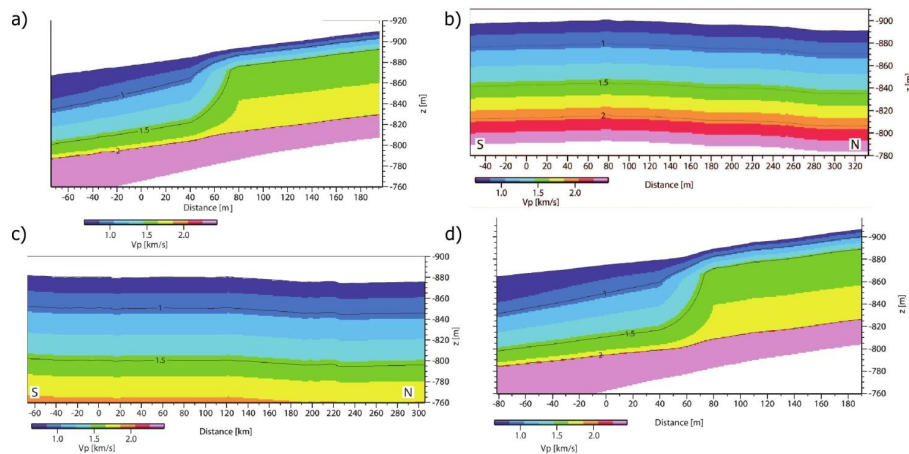
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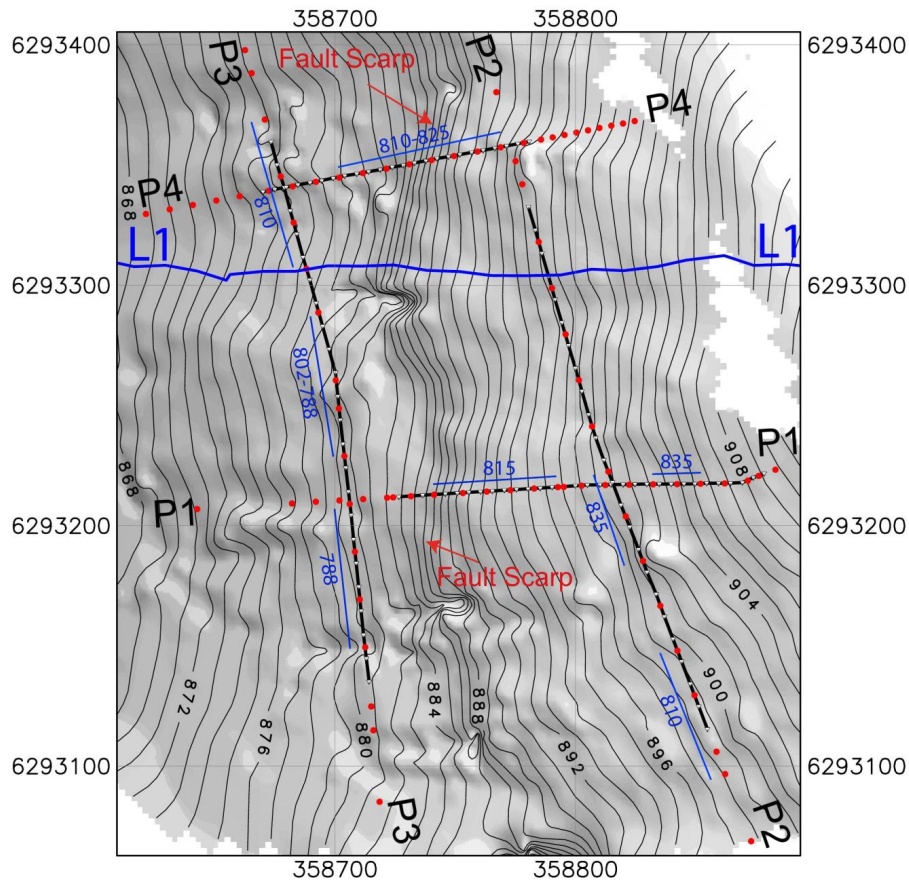
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**Fig. A1.** Initial models. Initial velocity models for profiles (a) P1, (b) P2, (c) P3 and (d) P4.



**Fig. A2.** Estimated substratum depth. Blue lines along seismic profiles P1–P4 mark the substratum depth, based on the 2-D seismic velocities models.

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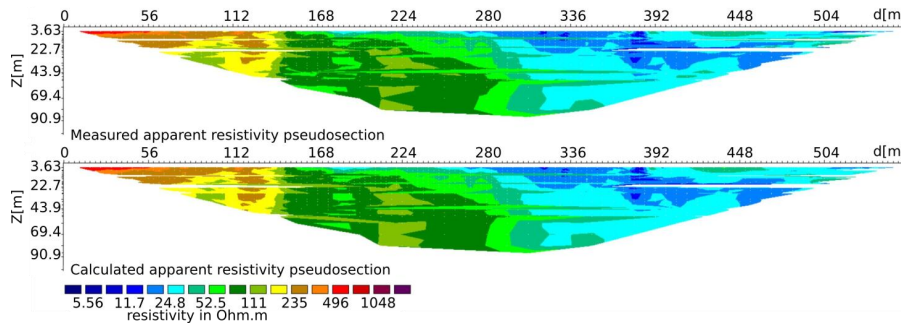
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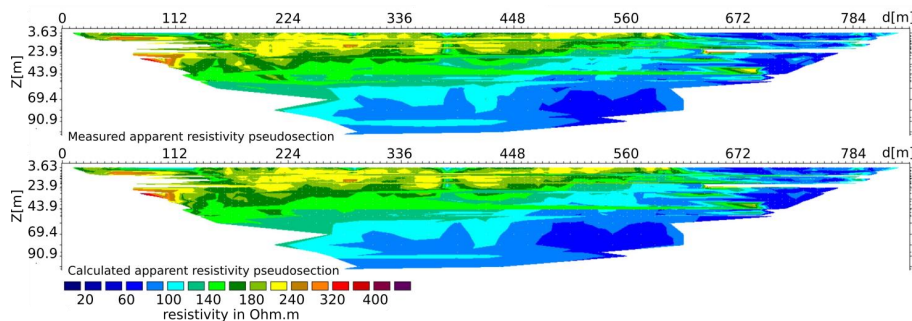
**Fig. A3.** Apparent resistivity pseudosections. Measured (up) and obtained from inversion (down) apparent resistivity pseudosections along L2 profile. Two VES were carried out using a Schlumberger configuration approximately 100 m at the eastern and western sides of the San Ramón fault scarp observed in that area.

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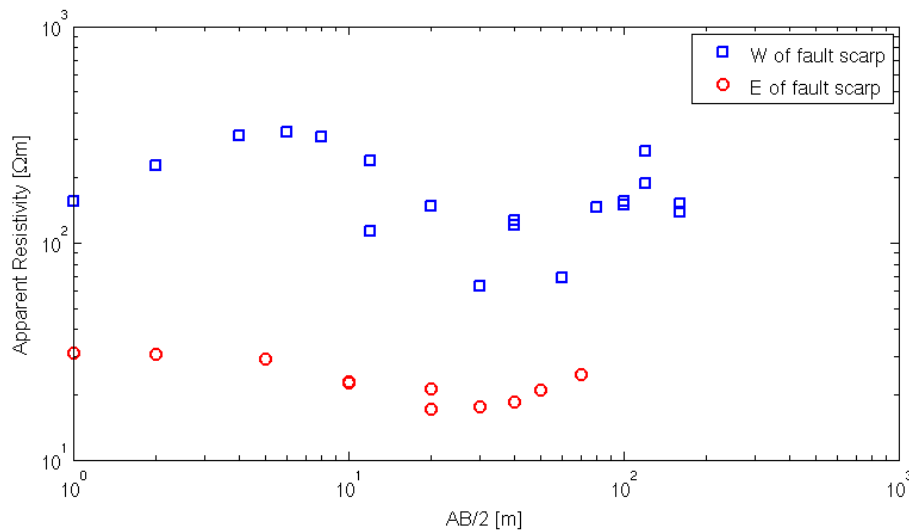


**Fig. A4.** Apparent resistivity pseudosections. Measured (up) and obtained from inversion (down) apparent resistivity pseudosections along L1 profile.

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**Fig. A5.** VES close to L2 profile. The results of both VES are shown in this figure as blue squares for apparent resistivity at the western side of the fault scarp (SEV1 in Fig. 4) and red circles at the eastern side of the fault scarp (SEV2 in Fig. 4).

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