

Updated 3D resistivity model of the Tres Vírgenes geothermal field (Mexico) from magnetotelluric data: Implications for geothermal exploration and development

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ABSTRACT

A case study is presented involving the 3D inversion of a legacy magnetotelluric (MT) dataset from the Tres Vírgenes Volcanic Complex (TVVC), Baja California Sur, Mexico, which is a significant geothermal energy resource with an installed capacity of 10 MWe. The MT dataset consists of 76 stations acquired between 1992 and 1999 (bandwidth: 0.003 – 100 Hz), and was previously analyzed using 1D, 2D, and 3D approaches, revealing a layered resistivity structure. However, significant uncertainty surrounds the capability of this dataset to resolve deep resistivity structures associated with the magmatic system. This study applies a 3D inversion approach that incorporates topography and bathymetry and inversion tests to assess the sensitivity of the model features to the data. Well-logs and laboratory measurements of resistivity were used to constrain the interpretation. The integrated methodology enabled a more accurate delineation of the resistivity structure of the TVVC, mapping an extensive conductor primarily associated with a brine-saturated smectite clay cap (10 – 60 % clay fraction, 150 – 200 °C). Internal variability within the conductor suggests the presence of a small sill with andesitic mush (30 – 85 % melt fraction) north of El Azufre Volcano, possibly associated with sulphate-type hydrothermal activity along the El Azufre Fault Zone. Moreover, a zone of lower resistivity to the north of the main conductor suggests a possible intrusion of seawater that remains isolated from the main hydrothermal systems in TVVC. Sensitivity tests indicate that the dataset lacks the resolution required to image deep bodies beneath the main conductor; consequently, deep conductors modelled in previous studies are likely inversion artifacts, and are unrelated to the deep magmatic system. These findings have the potential to improve the delineation of the geothermal system boundaries, and provide insights for future geothermal and magnetotelluric exploration.

1. Introduction

Geothermal energy is a component of the global energy portfolio that represents a renewable source of baseload power, with comparatively low greenhouse gas emissions, and considerable potential for expansion (Tester et al., 2007; Fridleifsson et al., 2008; USDoE, 2019). The heat from the interior of the Earth can be used in a wide range of industrial and commercial applications, including electricity generation, desalination, and district heating. An advantage of geothermal energy systems is their suitability for cascade utilization, where geothermal fluids are used in a series of sequential processes operating at progressively lower

temperatures, improving the efficiency and return on investment of these systems (Clauser, 2005; Dickson and Fanelli, 2005; Glassley, 2015; Huenges, 2010; Stober and Bucher, 2021). In geothermal systems, fluids circulate through hot rock at depths ranging from hundreds to thousands of meters, where they absorb heat that is transported to the surface for use. The residual fluids are typically reinjected into the ground, forming a semi-closed loop.

Geothermal exploration employs a wide array of geological, geochemical, and geophysical methods to identify potential targets where heat can be economically extracted using currently available technology (IGA, 2014). Geophysical methods, such as gravity,

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magnetics, seismology, or electromagnetics, play an important role in exploration, by imaging spatial variations in the physical properties of the Earth (e.g., density, magnetic susceptibility, seismic velocity, or electrical resistivity) (Georgsson, 2009). Electromagnetic (EM) methods, which include magnetotellurics, allow the identification of electrical resistivity variations. Conductive anomalies are often associated with geothermal features, such as hydrothermal alteration, fluids, or magma (Unsworth and Rondenay, 2012; Börner et al., 2015).

The magnetotelluric (MT) method (Cagniard, 1953) is widely applied in geothermal exploration due to its ability to resolve resistivity features from the surface to depths greater than 10 km, making it particularly useful in volcanic studies (Comeau et al., 2016; Cordell et al., 2018; Hanneson and Unsworth, 2023). MT images the subsurface resistivity using natural EM fields generated by solar and atmospheric processes. These signals attenuate as they travel downward into the subsurface, with the penetration depth increasing as frequency decreases. The attenuating fields convey information about the resistivity structure with which they interact. The relationship between orthogonal electric and magnetic field components (in the frequency domain) is described by the impedance tensor, typically expressed in terms of apparent resistivity and phase (Simpson and Bahr, 2005; Bedrosian, 2007; Chave and Jones, 2012). In complex volcanic settings, where resistivity structures are often non-layered, 3D inversion techniques are required to obtain a reliable resistivity model from MT data (Gailler et al., 2018; Gresse et al., 2021; Tanaka et al., 2021). However, topography and proximal conductive bodies, such as seawater, can distort the

EM fields, affecting depth of investigation and model resolution (Parkinson and Jones, 1979; Jiracek, 1990; Brasse and Eydum, 2008; Bertrand et al., 2009). Accounting for these factors improves the reliability of the model and enhances interpretation.

The Tres Vírgenes Geothermal Field (TVGF) (Fig. 1) provides base-load electric power to remote, off-grid communities in Baja California Sur, Mexico (Quijano-León and Gutiérrez-Negrín, 2003). In response to recent growth in population and tourism, energy demand in this arid region has increased substantially, especially for domestic use and desalination (Alcocer and Hiriart, 2008; Bermúdez-Contreras et al., 2008; Gutiérrez and Espíndola, 2010; Pelliccioli, 2020). The TVGF could play an important role in meeting these demands; however, this would require a sustainable expansion of its current installed capacity (10 MWe). Consequently, detailed geothermal exploration is necessary to gain a better understanding of the resource base in the region and guide future development decisions.

This paper describes a study from the TVGF, and illustrates how the MT method can contribute to geothermal exploration. It demonstrates the value of full 3D inversion that includes both topography and bathymetry. It also illustrates how resolution tests can be used to determine the sensitivity of resistivity model features to the measured MT data. Compared to previous EM studies (Romo-Jones et al., 2000; Romo-Jones et al., 2005; Flores, 2003; Antayhua-Vera, 2017), our methodology offers a more refined and detailed resistivity model. The results obtained in this study were interpreted in an integrated approach that also included geological and fluid geochemistry data. The findings

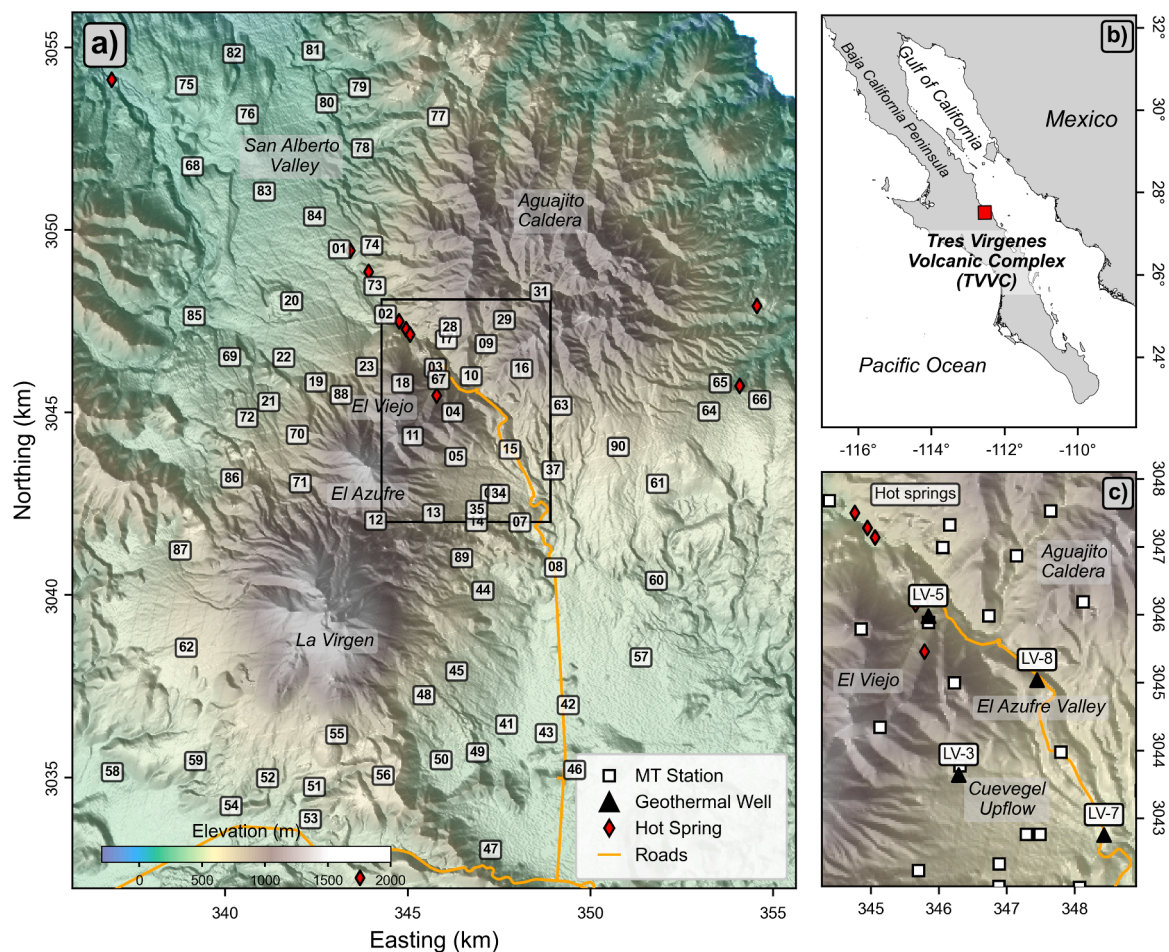


Fig. 1. Location and features of the Tres Virgenes Volcanic Complex (TVVC). a) Model domain displaying the magnetotelluric (MT) dataset locations. The map highlights some key geographic features referenced within this work, including the Aguajito Caldera, El Azufre, El Viejo and La Virgen Volcanoes; b) Regional map showing the location of the TVVC within the Baja California Peninsula, Mexico (coordinates in WGS84). c) Detailed view of the Tres Virgenes Geothermal Field, indicating the position of the wells and hot springs referred to throughout this study. Coordinates used in this study are in UTM Zone 12N unless otherwise specified.

from this study offer insights into the boundaries and possible active processes within the geothermal system, and has the potential to guide future geothermal development in the region.

2. Tres vírgenes volcanic complex (TVVC)

2.1. Geologic and tectonic setting

The TVVC is located in the Baja California Peninsula (BC), and its stratigraphic and structural evolution is closely related to (1) the interaction between the Pacific and North America plates, and (2) the opening of the Gulf of California. Subduction of the Pacific Plate beneath western Mexico ceased around 12.5 Ma, when the Gulf of California began its formation with N-S to NW-SE oblique-strike slip movements (Stock and Hodges, 1989; Lonsdale, 1991). Subsequently, the deformation was controlled by the opening of the Gulf of California (Conly et al., 2005), which was characterized by a NW-oriented compressive stress field sub-parallel to the main rift margin faults, accompanied by E-W and NE-SW extensional deformation (Bonini et al., 2019). During the Pliocene, the extensive deformation originated NW-SE, N-S, and NE-SW striking normal faults (Coletta and Angelier, 1981; Stock and Hodges, 1989; Garduño-Monroy et al., 1993).

At 12.4 Ma, the Pacific-North America plate boundary changed from an E-W convergent margin to a N-S to NW-SE oblique-strike-slip regime

(Stock and Hodges, 1989; Lonsdale, 1991). At approximately 6 Ma, the Pacific-North America plate boundary was transferred to the Gulf of California (Oskin et al., 2001). In the TVVC area, this tectonic activity is expressed by regional NW-SE fault systems, such as the Mezquital, Bonfil, and Campamento, which are predominantly of normal type with a right-lateral component (e.g., Garduño-Monroy et al., 1993; Macías-Vázquez and Jiménez-Salgado, 2012). These faults are located northwest and southwest of the TVVC. La Reforma Fault is a right-lateral strike-slip structure with a minor normal component. El Aguajito Caldera is bounded to the west and east by the El Aguajito and El Alamo faults, respectively. N-S to NE-SW faults, including the El Cimarrón Fault (CF), traverse the Aguajito Caldera and further continue southwestwards through the TVVC.

The Mezquital Fault delineates the western boundary of the San Francisco range (SFR), while the Campamento Fault defines the eastern limit of the Santa Lucia Range (SLR). The La Reforma Fault is approximately tangential to the La Reforma caldera (LRC), and together with the El Azufre Fault (EAFZ), is connected locally by the Las Víboras Fault, forming a continuous NW-SE fault system. To the northwest of La Virgen Volcano (LVV), the La Virgen Fault (LVF) is interpreted as the north-western extension of the El Campamento Fault (ECF). Local fault mapping by the Comisión Federal de Electricidad identified additional normal faults (e.g. EAFZ, El Partido (EPO), El Viejo 1 (EV1), El Viejo 2 (EV2), and El Volcán (EVF) Faults), that affect the El Viejo (EVV) and El

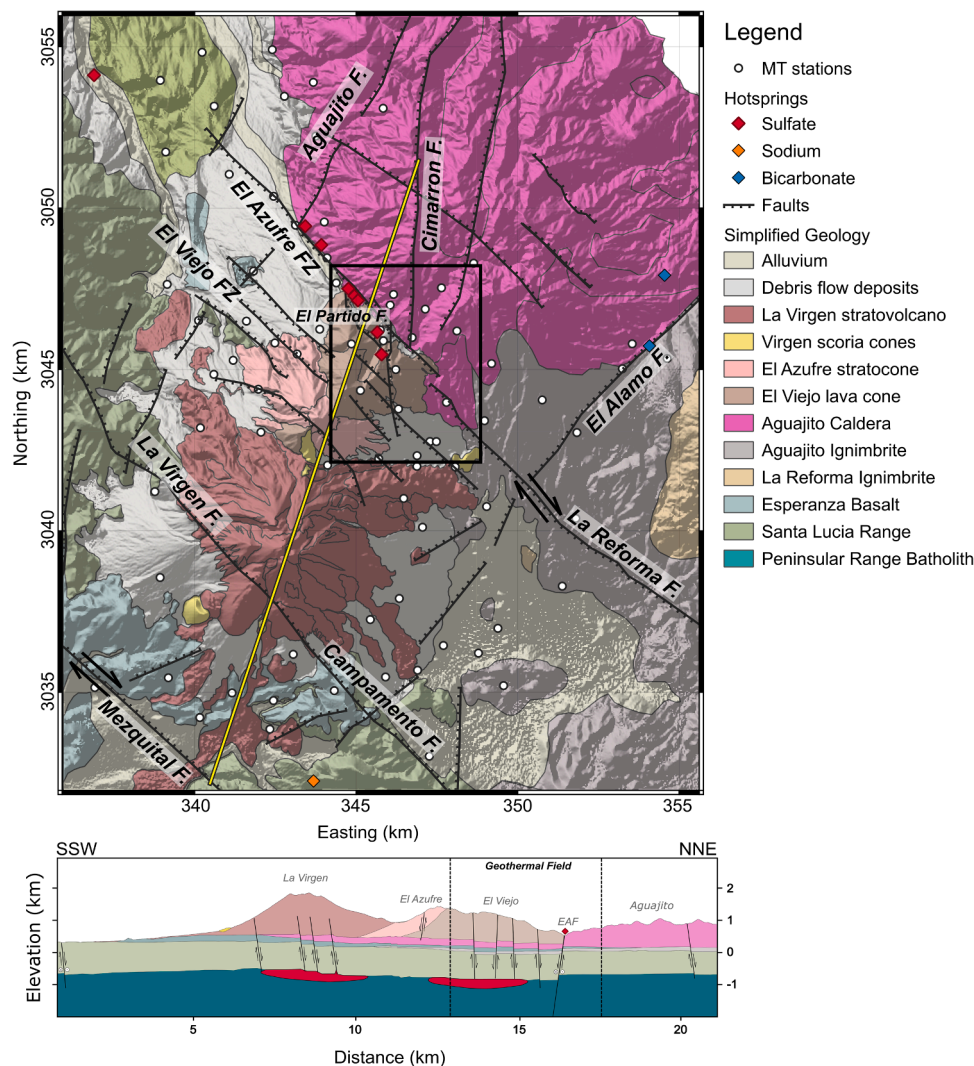


Fig. 2. Simplified geology of the Tres Virgenes Volcanic Complex. Map modified after Avellán et al. (2018), Antayhua-Vera (2017), and Servicio Geológico Mexicano (1997). Cross-section modified after Avellán et al. (2018). Hot spring location and categorization from Portugal et al. (2000).

Azufre (EAV) volcanoes. The main geologic units and faults of the TVVC are illustrated in the simplified geologic map shown in Fig. 2, adapted from Avellán et al. (2018).

The basement in the study area consists of the Peninsular Ranges Batholith (PRB), which intruded northern BC during the Cretaceous (Wetmore et al., 2003) and is inferred to extend beneath southern BC (Langenheim and Jachens, 2003; García-Abdeslem, 2014). The crystalline basement is overlain by Cenozoic post batholithic subduction-related volcanic rocks, emplaced between 24 and 11 Ma (Gastil et al., 1979; Sawlan and Smith, 1984; Sawlan, 1991). These units, including the Miocene Comondú Formation, are composed primarily of homogeneous calc-alkaline volcanics. Following the cessation of subduction, volcanism persisted giving rise to a variety of volcanic rocks. Calc-alkaline volcanism continued into the Pliocene and Quaternary during the syn-rift stage of the Gulf of California, leading to eruptions at the Reforma and Aguajito Calderas (Demant, 1981, 1984; Demant and Ortlieb, 1981; Garduño-Monroy et al., 1993; Osorio-Ocampo et al., 2019; Schmitt et al., 2006, 2010), and at the TVVC (Gastil et al., 1975; 1979; Sawlan, 1981; Capra et al., 1998; Schmitt et al., 2007; Avellán et al., 2018).

The stratigraphic sequence in TVVC consists of the PRB, which is covered by the Comondú Formation. In turn, this unit is overlain by an ignimbrite originating from the Aguajito Caldera. These units constitute the TVVC basement. The PRB was intersected at a depth of 1,129 m below the surface in well LV-2 (Garduño-Monroy et al., 1993). Outcrops of the Comondú Formation are present at 1) the San Francisco Range (SFR), NW of the volcanic complex, and 2) to the SW of the Santa Lucia range. At LV-2, the Santa Lucia Formation directly overlies the PRB. Tholeiitic basalts of the La Esperanza Formation cover the Santa Lucia Formation. This post-subduction volcanic sequence comprises the substratum upon which the TVVC was emplaced. The TVVC consists, from NE to SW, of the El Viejo, El Azufre, and La Virgen volcanoes (Fig. 1).

2.2. Previous studies of the TVVC and its geothermal system

In 1982, Comisión Federal de Electricidad (CFE) initiated an exploration program to establish the feasibility of exploitation of the TVVC geothermal resources. The initial studies included geology, geophysics, geochemistry, and hydrology. Most of these initial studies consist of unpublished internal reports. Based on these pioneering studies, the first steam-producing well was drilled in 1988. The early research has been published as follows. López (1998) summarized the geology of the Las Tres Vírgenes geothermal zone; mapping of local and regional fault systems was reported by Rocha and Romero (2009) and Gómez and Rocha (2009). García-Estrada and González-López (1998) summarized the gravity and magnetic studies. An open report of the DC geoelectric study was first published in 1998 (Bisurra-Pimentel, 1998).

After this first development phase, collaborative studies between CFE and scientific institutions and universities enabled a better understanding of the TVVC and its geothermal system in particular. Scientific research groups, driven by academic interest in the origin and evolution of the TVVC, have also made significant contributions.

The chemical composition of fluids from geothermal surface manifestations, as well as groundwater from surrounding areas of the TVVC, helped establish hydrogeological models for both the groundwater and the geothermal system (Portugal et al., 2000; González-Partida et al., 2005; Verma et al., 2006; Barragan et al., 2009; Viggiano-Guerra et al., 2009; Estrada and Sandoval, 2011; Benton et al., 2011; Prol-Ledesma et al., 2016). Temperature distribution in the geothermal system was determined by means of geothermometers, fluid inclusion studies, and analysis of hydrothermal alteration of clay minerals in the stratigraphic columns cut by the wells.

The seismicity of the geothermal field has been monitored and analyzed (Lermo et al., 2009; Lorenzo-Pulido and Soto-Peredo, 2013; Soto-Peredo, 2014; Antayhua-Vera et al., 2015). Based on this information, the distribution of b-values has been established

(Antayhua-Vera, 2017; Antayhua-Vera et al., 2015, 2022), with anomalously high values reported along the El Azufre and El Volcan Faults, which suggests active fracturing and fluid overpressure. Seismic anisotropy has been reported (Chacón-Hernández et al., 2021), and subsurface volumes with a high fracture density has been mapped beneath El Viejo and El Azufre volcanoes (Chacón-Hernández et al., 2024). A seismic tomography study has also been conducted (Vilchis-García et al., 2019), which identified an upflow zone beneath the El Azufre Fault.

Recently, detailed geological mapping of the TVVC and its geochronological evolution has been published (Avellán et al., 2018; Sosa-Ceballos et al., 2019). Detailed geophysical studies have been conducted between the El Azufre and Reforma faults (Sena-Lozoya et al., 2020; Casallas-Moreno et al., 2021). The depth to the Curie isotherm has also been established (Campos-Enríquez et al., 2019).

2.2.1. Electromagnetic and electrical studies at TVVC

The measurement of electrical resistivity in geothermal exploration in active volcanic settings can help identify and characterize different components of the hydrothermal and magmatic systems, such as clay caps, geothermal reservoirs, partial melt, and other heat sources. Multiple electrical and electromagnetic studies have been conducted in the TVVC region to image the subsurface resistivity structure, which include transient electromagnetics (TEM), vertical electric sounding (VES), and magnetotellurics (MT).

Palma-Guzmán (1998) applied a 1D approach to the same MT dataset used in this study. The location of this dataset is illustrated in Fig. 1. The area of interest was a graben structure bounded by the El Azufre and El Partido Faults, as well as the intersection of the Cimarrón Fault and other NW-SE trending faults.

Subsequently, Romo-Jones et al. (2000) combined 2D MT inversion, TDEM, and VES data with coda wave analysis, and identified multiple conductive zones ($< 5 \Omega\text{m}$) in the depth range 250 – 2500 m. These zones suggested that there was high geothermal energy development potential along the El Azufre Valley, at the boundary between TVVC and Aguajito Caldera (Fig. 1c). Some of these conductive zones were associated with high seismic attenuation, and were interpreted as zones of intensely fractured rocks containing high-temperature fluids associated with the El Azufre Fault.

Flores (2003) applied 1D TEM and VES to map the resistivity structure of the top 650 m, identifying a persistent lateral conductor within the region of geothermal interest. This conductor was interpreted as a regional aquifer. Lateral discontinuities and thickness variations within this conductor were interpreted as a convective system, referred to as the Cuevegel hydrothermal upflow zone (Fig. 1c). The location of this feature may correlate with the upflow zone identified by Vilchis-García et al. (2019).

Romo-Jones et al. (2005) implemented a novel approach that employed series-parallel (S-P) response functions as a substitute for the impedance tensor. Through 2D inversion of S-P responses, these authors imaged a continuous lateral shallow conductor, as well as some conspicuous deep vertical conductors below 12 km depth. The lateral conductor was interpreted as being constrained to andesitic and pyroclastic deposits from the Santa Lucia Formation and the Comondú Group, whereas the deep vertical anomalies were labelled as possible methodological artifacts.

Antayhua-Vera (2017) undertook additional analysis of all of the previously collected MT data. Analysis showed that the MT data required a 2D/3D resistivity structure. Antayhua-Vera (2017) implemented the first 3D inversion of the TVVC MT data using the WSINV3DMT algorithm (Siripunvaraporn et al., 2005, 2006). A rectangular grid was used, but the topography and bathymetry were not included in the analysis. This study identified a horizontal conductor ($< 10 \Omega\text{m}$) in the depth range 50 to 1,500 m depth, embedded in more resistive media. Two similar deep conductive zones at 2.5–5.5 km and > 6.5 km depth were observed in some of the models. The lateral

conductor was interpreted as highly fractured, highly altered rocks from the Santa Rosalía, Santa Lucía and Comondú Formations. This conductor was interpreted as the cap seal, with the reservoir located in the underlying resistive zone linked to the granodioritic basement. The geometry of the deep conductors was suggestive of magmatic activity. Resolution tests showed that the model was well resolved to a depth of 5 km, and conductors at greater depth were not required by the data.

2.3. Need for additional analysis of the MT data

In this study, the previous analysis of the MT data was extended by implementing a full 3D inversion using the ModEM algorithm (Kelbert et al., 2014). Topography and bathymetry were included to obtain a more realistic model of the geothermal system than those obtained from previous inversions. A sensitivity analysis is an essential part of an inversion study and was undertaken to investigate the nature of the deep conductors, such as those imaged by Romo-Jones et al. (2005) and Antayhua-Vera (2017), which may be either numerical artifacts or model features required by the measured MT data.

3. 3D inversion of the Tres Virgenes MT data set

3.1. Tres Virgenes MT dataset

The dataset used in this study was acquired with funding from the Comisión Federal de Electricidad of Mexico during multiple campaigns from 1992-1999 (Antayhua-Vera, 2017; Romo-Jones et al., 1994; Romo-Jones et al., 2005). This legacy collection consists of processed MT full tensor measurements from 79 stations distributed around the vicinity of the Tres Virgenes Volcanic Complex and Aguajito Caldera, included in a rectangular region of 20 km x 24 km, with an average site spacing of 1200 m (Fig. 1). The data span the frequency range 0.003 to 100 Hz, and were recorded using an EMI, Inc. system (Palma-Guzmán, 1998). Remote referencing (Gamble et al., 1978) was applied during

robust processing of the time series data (Palma-Guzmán, 1998).

Based on sounding curve patterns, two distinct types of responses can be recognized in the MT data. Fig. 3 illustrates these types of MT response in plots of apparent resistivity and phase as functions of frequency, where the x- and y-axes are aligned with the northing and easting direction, respectively.

Most sites exhibit curves with a pattern of high-low-high apparent resistivity as frequency decreases (identified as type A curves in Fig. 3b), suggesting a three-layered structure consisting of a resistive-conductive-resistive configuration with increasing depth. The corresponding phase curves are consistent with apparent resistivity with phase $> 45^\circ$, when apparent resistivity decreases, and phase $< 45^\circ$ when apparent resistivity increases. The observation of phase values greater than 45° at the highest measured frequencies is common in this dataset, and suggests the presence of partially resolved near-surface resistivity structures. The presence of near-surface structures may also introduce static shifts or galvanic distortion effects in MT data. Modest static shifts were observed at a number of stations. Some variations in this pattern are observed based on the comparison of the values of the ρ_{xy} and ρ_{yx} components. At some sites, these values are approximately equal (e.g. TV06), while in other sites one of the components shows higher values than the other at lower frequencies (e.g. site TV20 where $\rho_{xy} > \rho_{yx}$, or site TV07 where $\rho_{xy} < \rho_{yx}$).

Another type of response is characterized by a pattern of low-high-(low) apparent resistivity (type B curves, Fig. 3c). This pattern suggests a shallow conductor underlain by more resistive media and a deeper conductor. Some variations in this pattern were also recognized based on the comparison of ρ_{xy} and ρ_{yx} values. For example, the response from station TV41 ($\rho_{xy} = \rho_{yx}$) suggests a laterally isotropic, two-layered structure. In contrast, the responses from stations TV17 ($\rho_{xy} > \rho_{yx}$) and TV60 ($\rho_{xy} < \rho_{yx}$) are indicative of lateral resistivity contrasts, possibly affected by static shift or galvanic distortion caused by topography or near-surface resistivity variations. Proximity to fault zones (e.g. El

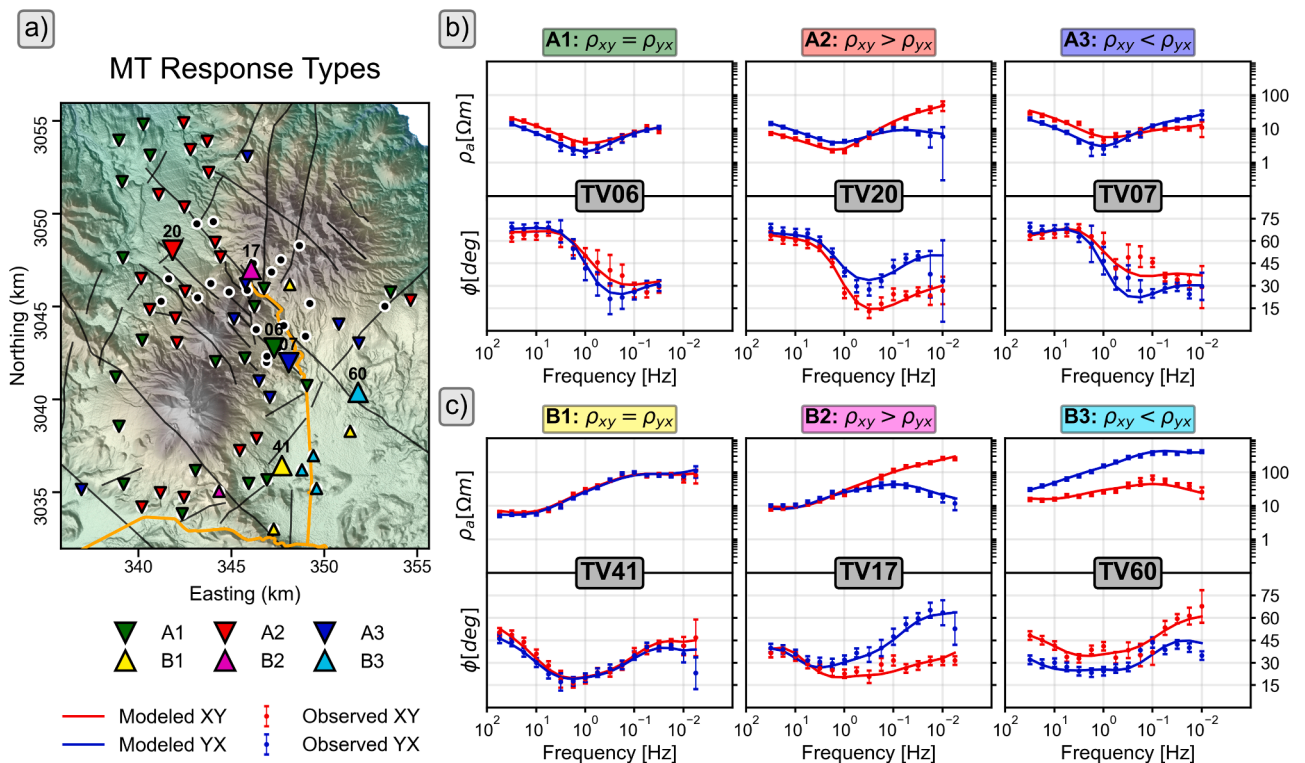


Fig. 3. Magnetotelluric response classification in TVVC. a) Map showing spatial distribution of MT response types. b) Examples of type A curves at sites TV06 (Type A1), TV20 (type A2), TV07 (type A3). c) Examples of type B curves at sites TV41 (Type B1), TV17 (type B2), TV60 (type B3). Modelled responses (solid lines) correspond to inversion TVV20. Reference frame alignment: X and Y axes correspond to the northing and easting directions, respectively.

Azufre, La Reforma, El Alamo) is often correlated with a split between the xy and yx response curves (Fig. 3a).

Fig. 4a,b,c shows the entire dataset as pseudo-sections of apparent resistivity, phase and tipper. The data are projected onto the N-S direction, and frequency represented on the vertical axes. The diagonal

components of apparent resistivity (ρ_{xx} , ρ_{yy}) are several orders of magnitude lower than the off-diagonal components (ρ_{xy} , ρ_{yx}), except at low frequencies. Both off-diagonal components have similar values, and exhibit a three-layered pattern with low resistivity typically concentrated in the 0.3-10 Hz band. Phase in the diagonal components (ϕ_{xx} ,

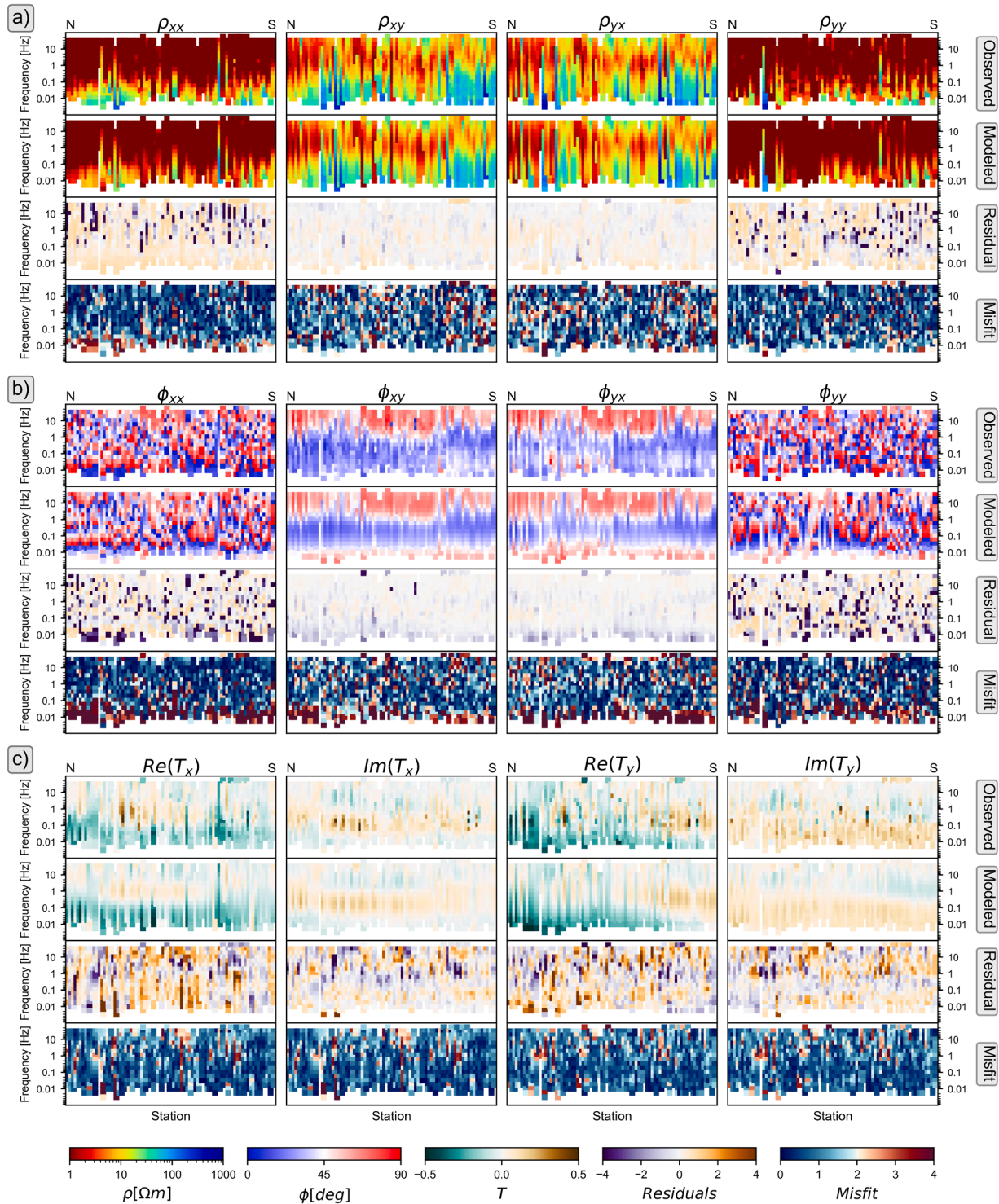


Fig. 4. Measured data and response of the preferred inversion model (TVV20) misfit. a) Pseudosections showing observed and modeled responses for the four components of apparent resistivity. b) Pseudosection comparison between observed and modeled (TVV20) responses for the four components of phase. c) Pseudosection comparison between observed and modeled responses for the tipper vector. d) Map of apparent resistivity of the impedance tensor determinant for selected frequencies (10, 1, 0.1, 0.01 Hz). e) Map of phase of the impedance tensor determinant for selected frequencies (10, 1, 0.1, 0.01 Hz). f) Map of induction vectors using the Parker convention at selected frequencies (10, 1, 0.1, 0.01 Hz). g) Map of normalized misfit from all response components at selected frequencies (10, 1, 0.1, 0.01 Hz). Same map scale and dimensions as in Fig. 1a (northing: 24km, easting: 19.8 km).

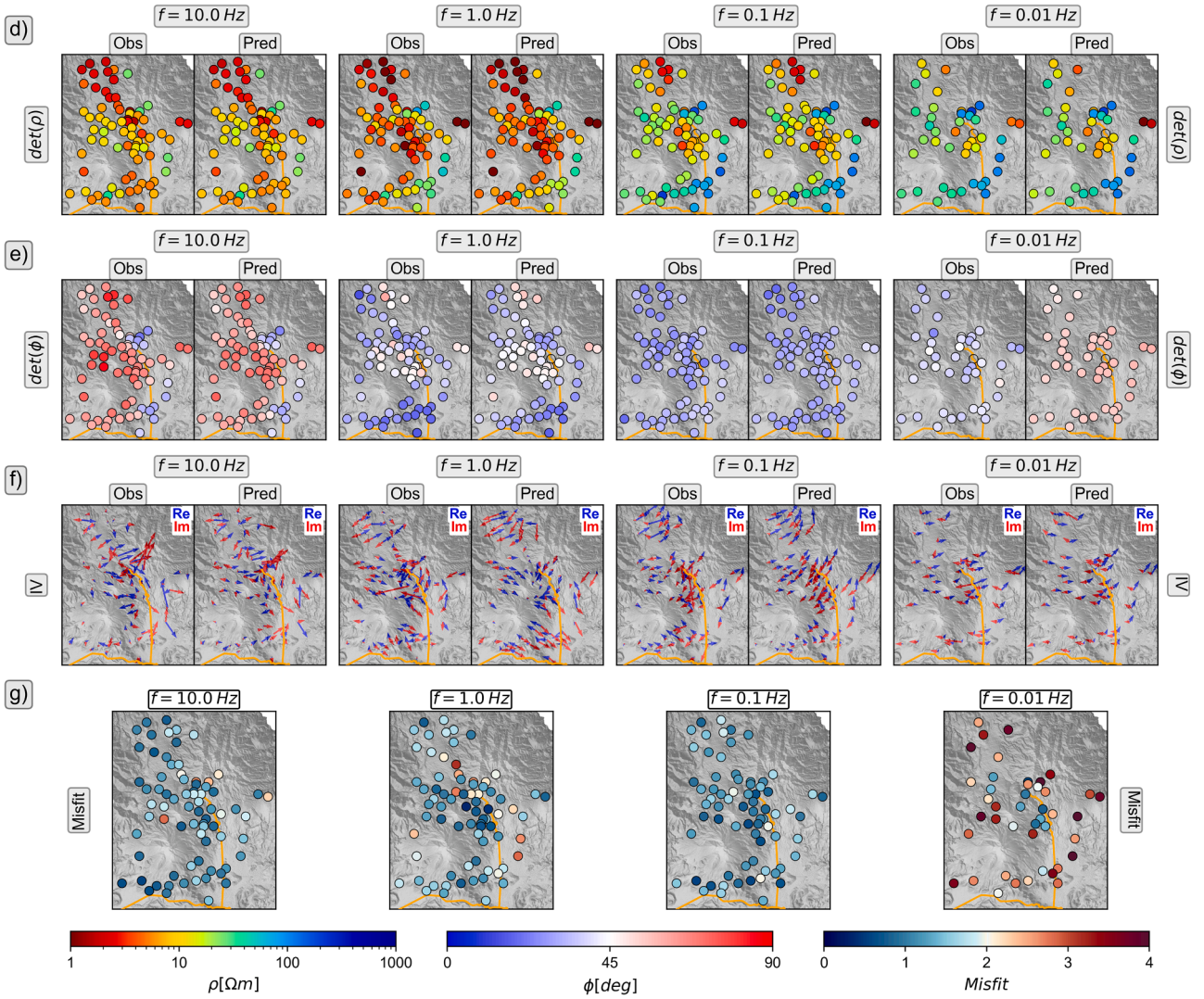


Fig. 4. (continued).

ϕ_{yy}) is randomly distributed across most of the spectrum, whereas the off-diagonal components (ϕ_{xy} , ϕ_{yx}) consistently show high values ($> 45^\circ$) at higher frequencies shifting to low values at low frequencies. The spatial distributions of the determinant apparent resistivity, determinant phase and induction vectors at selected frequencies are shown in Fig. 4d,e,f.

Phase tensor analysis (Caldwell et al., 2004; Booker, 2014) was applied to the dataset to investigate the dimensionality of the dataset. Fig. 5 shows a north-south pseudo-section illustrating variations with frequency, represented as ellipses colored according to the β -skew angle. The analysis suggests near-isotropic conditions at higher frequencies, where the ellipses tend to be nearly circular with low skew angles. At lower frequencies, the ellipses tend to be elongated, indicating a preferred strike direction. This direction shifts from NNE-SSW to NE-SW with decreasing frequency, possibly related to structural variations at depth. This observation aligns with a comparable reorientation of fast shear-wave polarization directions, from NW-SE to NE-SW at 4-6 km depth, obtained through shear-wave splitting analysis (Chacón-Hernández et al., 2024). The skew angle is generally low except for some localized subsets at higher frequencies, and a more widespread subset at lower frequencies. This configuration suggests an overall 2D resistivity structure near the surface shifting to a deep 3D structure in several parts of the study area, which requires a 3D inversion approach.

3.2. Inversion of MT data

Since the data have a significant 3D component, a 3D inversion approach using the *ModEM* algorithm was implemented (Egbert and Kelbert, 2012; Kelbert et al., 2014). This algorithm calculates the solution of the inverse problem minimizing a scalar penalty functional, Φ , which depends on the data d and model m vectors:

$$\Phi(d, m) = (d - f(m))^T C_d^{-1} (d - f(m)) + \lambda (m - m_0)^T C_m^{-1} (m - m_0)$$

$$\Phi(d, m) = \Phi_d + \lambda \Phi_m$$

This objective function consists of a data misfit term Φ_d that measures how the model fits the observed data, and a regularization term $\lambda \Phi_m$ that penalizes model roughness.

The misfit term contains the sum of the differences between the observed data (d) and the forward model response ($f(m)$), normalized by the error covariance matrix (C_d). For future reference (Section 3.2.2), it is important to mention that the magnitude of this term is influenced by the data errors, as well as the number of elements in the data vector (e.g. number of sites, number of transfer function polarizations, or number of evaluation frequencies used).

In contrast, the regularization term represents the sum of the differences between the modelled resistivity (m) and the starting (prior) model (m_0) normalized by the model covariance matrix (C_m), and scaled

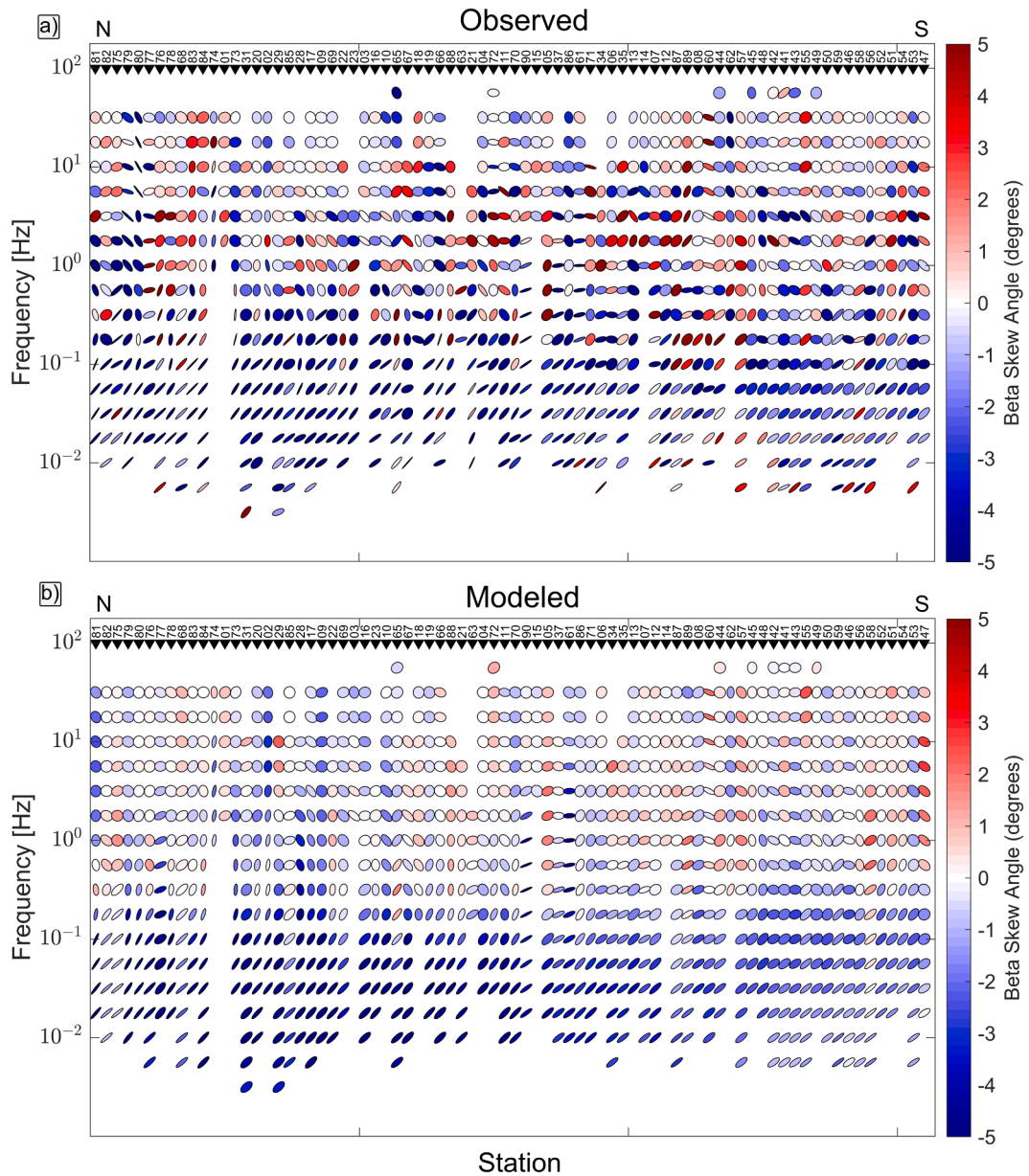


Fig. 5. Phase tensor pseudosection indicating isotropic conditions near the surface, and a prevalent electromagnetic strike direction below 1 Hz shifting from NNE-SSW to NE-SW. Low skew angle suggests a 2D structure for higher frequencies, with some 3D effects at lower frequencies. a) observed data; b) modeled data (model TVV20).

by a trade-off regularization factor λ . The model covariance matrix incorporates a user-defined array of directional smoothing factors that control the relative weighting of the directional model gradients. The starting model remains fixed, while the modelled resistivity is updated at each iteration. The magnitude of the regularization term is influenced by the number of parameters in the model vector (i.e. number of cells).

The minimization of the objective function is achieved iteratively using the non-linear conjugate gradients (NLGC) method (Rodi and Mackie, 2001). This process ensures rapid descent through the solution space, moving in a direction based on the objective function gradient $-\nabla\Phi(m)$, with the step length α obtained through line search. Traditionally, the Jacobian matrix ($J = \frac{\partial f(m)}{\partial m}$) is used for gradient calculation; however, the NLGC method applies a computationally efficient adjoint technique not requiring the full evaluation of the Jacobian (Egbert and Kelbert, 2012). The gradient and step length values are used to iteratively update the model by introducing a perturbation (i.e. $m_{new} = m - \alpha \nabla\Phi(m)$). The forward response of the updated model is then

calculated, and new gradients and step sizes are estimated to adjust the search direction. The process is repeated until convergence is achieved. For future discussion (Section 3.2.2), it should be noticed that the magnitude of the step size (obtained through a line search) and the function gradient are also influenced by the number of data points and model parameters.

To reduce the likelihood of the inversion converging to local minima, the regularization factor λ is progressively reduced when the change in RMS misfit between iterations is less than a specified misfit reduction threshold (i.e. 1E-3). Lower values of λ reduce the influence of the regularization term, favoring larger model perturbations. This allows the inversion to explore a broad set of the solution space with increasingly rough models (i.e. higher norms, more complex structure), and obtaining a closer fit to the observed MT data. The inversion process continues until the regularization factor reaches a user-defined termination threshold value (i.e. $\lambda = 1E-10$). This feature allows escaping local minima and fitting finer-scale structures by gradually reducing the

influence of the regularization term at later iterations.

3.2.1. Selected MT data

For the inversion process, 76 of the acquired 79 MT stations were selected containing full impedance tensor and tipper vector responses. The off-diagonal components of the impedance tensor were edited using the D+ algorithm (Parker, 1980; Beamish and Travassos, 1993; Parker and Booker, 1996). An error floor of 5 % of the magnitude of each component was used to prevent overfitting when calculating the D+ responses. The average RMS misfits for the Z_{xy} and Z_{yx} impedance components were 0.53 ± 0.22 and 0.40 ± 0.19 . A table listing the RMS misfit of the D+ models for the Z_{xy} and Z_{yx} components for all sites is included in the Supplemental Material (S1). The data were interpolated to obtain a common set of 18 frequencies at all MT stations, with four frequencies per decade, in a range of 0.003–100 Hz. This choice of frequencies was selected to balance spectral resolution and computational cost.

3.2.2. Grid design

A heuristic approach was adopted to define a model domain sufficiently large to model the attenuation of the source EM fields, while also being discretized into a computationally feasible grid with appropriate spatial resolution (Table 1).

An initial test was conducted, using a large grid with a maximum extent of three times the skin depth of the lowest frequency (0.003 Hz) with a constant resistivity of 100 Ωm (TVV01, Supplemental Material S2A). The model grid included topography and bathymetry data from GEBCO (2024). Standard fixed resistivity values were assigned to air and seawater cells, i.e. 10^{17} Ωm and 0.3 Ωm , respectively. The air cells were set to a value close to infinity to approximate an insulating boundary, as it has been used in several previous studies without causing instability or numerical artifacts (e.g. Comeau et al., 2016; Cordell et al., 2018; Hanneson and Unsworth, 2023). For the initial grid, a horizontal cell size of 300 m (one quarter of the typical station spacing) was sufficient to accommodate the dataset while avoiding placing multiple MT sites inside each cell. The cell heights in the vertical direction started at 25 m for cells within topography (air cells), and increased geometrically by a factor of 1.15 downwards. Horizontal padding cells were added at the margins, increasing in width by a geometric factor of 1.5. In total, the model consisted of 1,289,304 cells ($108 \times 94 \times 127$, x , y , z), covering a volume of 547 km $\times$ 543 km $\times$ 240 km. An initial regularization parameter value of 1 was used, as well as horizontal and vertical smoothing factors of 0.3 and 0.1, respectively. The initial test converged after 69 iterations, with the RMS misfit decreasing from 16.416 to 3.762. The solution was characterized by very faint anomalies in the resistivity model, showing minimal departure from the homogeneous 100 Ωm initial model (Supplemental Material S2A, Fig. S2). The response curves exhibited poor fits.

Two additional tests were conducted using a smaller number of cells in the resistivity model. Test TVV02 used a model domain with horizontal padding reduced to a value twice as large as the core area where the MT stations were located (Supplemental Material S2A). The grid was designed using similar parameters as in TVV01. This model consisted of $100 \times 86 \times 117$ cells, and a maximum extent of 126 km $\times$ 122 km $\times$ 117 km. The inversion converged after 124 iterations, reducing the RMS misfit from 16.433 to 3.332. The final resistivity model also exhibits faint anomalies; however, the model parameters show greater departure from the initial resistivity model than in TVV01. Similarly, the solution response curves exhibit poor fits.

To investigate this problem, test TVV03 used a further reduction in the extent of the padding area (i.e. equal to the data extent), as well as an increased air cell height of 35 m (Supplemental Material S2A). The resulting grid consisted of $92 \times 78 \times 83$ cells, and extended through a region of 100 km $\times$ 95 km $\times$ 83 km. The inversion converged after 179 iterations, significantly reducing the RMS misfit from 16.466 to 1.335. Compared to TVV01 and TVV02, this inversion had much greater

resistivity model variations and an improved data fit.

Fig. S1 in section S2A of the supplemental material presents the variation of convergence indicators throughout the inversion for the three inversions mentioned above. In all three cases, RMS misfit reached a floor before convergence, with the number of iterations inversely correlated with the number of model cells. This behavior can be explained by a faster reduction of the regularization factor λ , which is correlated with a rapid decrease in step size α of several orders of magnitude. Although smaller step sizes require more iterations for convergence, they are also associated with smaller perturbations. In turn, smaller perturbations are more likely to trigger a reduction of λ , thus requiring fewer iterations to meet the termination criteria. The inverse correlation between the number of cells and the step size can be attributed to the influence of model vector length (i.e. number of model parameters) in the NLCG algorithm.

Finely discretized models (i.e. longer model vector m) may result in larger contributions from the regularization term $\lambda\Phi_m$, compared to the data misfit term Φ_d . This results in the scalar objective function Φ being dominated by regularization, and the inversion algorithm favoring solutions that minimize model updates ($m - m_0$) instead of data misfit ($d - f(m)$). This type of model solution tends to remain close to the initial model, producing models characterized by faint resistivity anomalies and a minimal reduction of the RMS misfit. Overly constrained inversions may lead to smaller step sizes α , which can potentially cause convergence to local minima close to the initial model, i.e., with faint anomalies. Larger step sizes permit greater mobility to explore the solution space, resulting in higher model norms and a greater reduction of the objective function. This explains the minimal deviation from the starting resistivity and the higher data misfits observed in larger models, such as TVV01 and TVV02. Therefore, grid design must ensure an adequate balance between the number of model parameters (grid size), the number of data points, and the regularization scaling factor λ .

As a rule of thumb, the grid size must be sufficiently large to model attenuation of the electromagnetic fields in the subsurface, while the number of cells should not be significantly larger than the data vector to prevent unwanted effects caused by regularization dominance during inversion. In our preferred inversion, the model-to-data ratio was ~ 46.5 . The reduced model vector consists of 595,608 parameters ($92 \times 78 \times 83$ cells), whereas the data vector included 12,804 values, corresponding to 1,067 tipper and impedance data points (defined at 76 sites and 18 frequencies), 6 polarizations, and 2 complex components per polarization. This model-to-data ratio is comparable to values calculated from similar case studies, such as Cordell et al. (2018), where it was ~ 44 . In contrast, regularization effects were encountered for higher ratios of ~ 101 (TVV01) and ~ 79 (TVV02).

The air resistivity was set to 10^{17} Ωm to ensure that air cells behave as insulators. Sensitivity tests using alternative values (e.g. 10^8 Ωm , 10^{10} Ωm) produced negligible differences in the forward responses, indicating that the selected value does not have a critical impact on the inversion results.

Based on these considerations, the grid design from model TVV03 was selected for further analysis (Table 1).

3.2.3. Inversion parameter selection

A similar heuristic approach was applied to investigate appropriate values for the remaining inversion parameters, including the regularization factor λ , directional smoothing, and starting model resistivity ρ_0 . A summary of these inversion tests is presented in Table 1, as well as in the Supplemental Material S2.

Inversions with larger regularization factor λ values typically require a greater number of decimations and iterations to reach the termination threshold (i.e., $\lambda = 10^{-10}$). Greater values of λ also increase the influence of the regularization term in the objective function, and are associated with smaller step sizes and model norms. This behavior was observed when testing three different orders of magnitude for the initial

Table 1
Summary of inversion tests for TVVC dataset.

Inversion	Variable parameter	Iterations	Initial RMS misfit	Final RMS misfit	Error floor	λ_{start}	λ_{end}	ρ_{initial}	Horizontal smoothing	Vertical smoothing	Node count (nx, ny, nz)	Padding length (km)	Model dimensions (km)
TVV01	Extended model domain	69	16.416	3.762	5 %	1	1.2E-10	100	0.3	0.1	108×94×127	262×262	548×543×127
TVV02	Intermediate model domain	124	16.434	3.332	5 %	1	1.2E-10	100	0.3	0.1	100×86×117	51×51	126×122×117
TVV03	Reduced model domain	179	16.466	1.335	5 %	1	1.2E-10	100	0.3	0.1	92×78×83	37.8×37.8	100×95×83
TVV04	$\lambda = 1$	179	16.466	1.335	5 %	1	1.2E-10	100	0.3	0.1	92×78×83	37.8×37.8	100×95×83
TVV05	$\lambda = 10$	218	16.466	1.376	5 %	10	2.3E-9	100	0.3	0.1	92×78×83	37.8×37.8	100×95×83
TVV06	$\lambda = 100$	256	16.466	1.464	5 %	100	1.8E-10	100	0.3	0.1	92×78×83	37.8×37.8	100×95×83
TVV07	Horizontal smoothing: 0.1	128	16.469	2.184	5 %	1	1.2E-10	100	0.1	0.1	92×78×83	37.8×37.8	100×95×83
TVV08	Horizontal smoothing: 0.3	179	16.466	1.335	5 %	1	1.2E-10	100	0.3	0.1	92×78×83	37.8×37.8	100×95×83
TVV09	Horizontal smoothing: 0.5	110	16.469	1.468	5 %	1	1.2E-10	100	0.5	0.1	92×78×83	37.8×37.8	100×95×83
TVV10	Vertical smoothing: 0.1	179	16.466	1.335	5 %	1	1.2E-10	100	0.3	0.1	92×78×83	37.8×37.8	100×95×83
TVV11	Vertical smoothing: 0.3	128	16.469	1.636	5 %	1	1.2E-10	100	0.3	0.3	92×78×83	37.8×37.8	100×95×83
TVV12	Vertical smoothing: 0.5	104	16.469	1.436	5 %	1	1.2E-10	100	0.3	0.5	92×78×83	37.8×37.8	100×95×83
TVV13	Initial $\rho = 1 \text{ } \Omega\text{m}$	188	5.657	1.178	5 %	1	1.2E-10	1	0.3	0.1	92×78×83	37.8×37.8	100×95×83
TVV14	Initial $\rho = 10 \text{ } \Omega\text{m}$	113	4.681	0.981	5 %	1	1.2E-10	10	0.3	0.1	92×78×83	37.8×37.8	100×95×83
TVV15	Initial $\rho = 100 \text{ } \Omega\text{m}$	179	16.466	1.335	5 %	1	1.2E-10	100	0.3	0.1	92×78×83	37.8×37.8	100×95×83
TVV16	Initial $\rho = 250 \text{ } \Omega\text{m}$	197	28.773	1.366	5 %	1	7.5E-9	250	0.3	0.1	92×78×83	37.8×37.8	100×95×83
TVV17	Initial $\rho = 500 \text{ } \Omega\text{m}$	186	42.518	1.665	5 %	1	1.2E-10	500	0.3	0.1	92×78×83	37.8×37.8	100×95×83
TVV18	Initial $\rho = 750 \text{ } \Omega\text{m}$	183	52.496	1.901	5 %	1	1.2E-10	750	0.3	0.1	92×78×83	37.8×37.8	100×95×83
TVV19	Initial $\rho = 1,000 \text{ } \Omega\text{m}$	221	61.445	1.966	5 %	1	2.3E-10	1,000	0.3	0.1	92×78×83	37.8×37.8	100×95×83
TVV20	Selected parameters	113	4.681	0.981	5 %	1	1.2E-10	10	0.3	0.1	92×78×83	37.8×37.8	100×95×83

regularization parameter, i.e., $\lambda = 1$ (TVV04), $\lambda = 10$ (TVV05), and $\lambda = 100$ (TVV06) (Supplemental Material S2B). The value $\lambda = 1$ was selected, as it was associated with larger step size values, fewer iterations, and larger misfit reduction. These characteristics improved the efficiency of the inversion process.

Directional smoothing factors of 0.1, 0.3, and 0.5 were investigated for horizontal smoothing (HS, TVV07–TVV09) and vertical smoothing (VS, TVV10–TVV12). Horizontal smoothing of 0.3 was correlated with sufficiently large step size values and more iterations, enabling the greatest reduction in RMS misfit and objective function values (Supplemental Material S2C). In comparison, a vertical smoothing (VS) factor of 0.1 produced the greatest reduction in misfit and objective function (Supplemental Material S2D). These values resolved well-defined, strong, and laterally continuous anomalies, preventing the introduction of spurious features such as two-layered stacked conductors (e.g. HS 0.5, VS 0.1, Supplemental Material S2C).

3.2.3.1. Initial resistivity model. The starting model determines the initial RMS misfit and influences the step size and number of iterations required for convergence. The resistivity of the starting model, ρ_0 , was tested using values of 1 Ωm (TVV13), 10 Ωm (TVV14), 100 Ωm (TVV15), 250 Ωm (TVV16), 500 Ωm (TVV17), 750 Ωm (TVV18) and 1000 Ωm (TVV19). These tests revealed both commonalities and important variations. As shown in Fig. 6, a distinct conductive layer is consistently resolved in all cases. The conductance of this feature is relatively stable; however, its thickness, depth, and resistivity are sensitive to the starting model. More specifically, the thickness and depth of the conductive layer increase with higher ρ_0 , while its resistivity decreases.

It is well established that the resistivity of the starting model can have a major impact on the final inversion model (Robertson et al., 2020; Sharma and Verma, 2015). This is illustrated in Fig. 6 and Figs S9 and S10 in Supplemental Material S2E. This non-uniqueness can be addressed by comparing the resistivity model with well log data, if available. At Tres Vírgenes, well logs from LV-3 and LV-11 are publicly accessible, including temperature, resistivity, and mineral alteration records (González-Partida et al., 2005; Gutiérrez-Negrín et al., 2001; Santoyo et al., 2000) (Fig. 7). These logs suggest the presence of two

distinct low-resistivity zones. The first zone, observed in hydrothermal alteration records from LV-11, shows alteration intensity between 10–60 % containing smectite, occurring between 290 masl and 64 mbsl (Fig. 7c). The second zone, reported as an interval of “low to very low resistivity” in qualitative resistivity logs at LV-11, corresponds to depths of 908–1021 mbsl (Fig. 7a; Gutiérrez-Negrín et al., 2001). Additionally, well LV-11 exhibits two permeable zones at depths of 410–480 mbsl and 900–1110 mbsl, which partially overlap the production zone from well LV-3, located between 481–931 mbsl (Fig. 7a,c,d).

It is important to mention that well logs provide a detailed sampling of the local structures surrounding boreholes, whereas MT data senses a larger volume of the subsurface. Consequently, integrating well log data does not provide exact constraints for the broader region probed by the MT data. Nevertheless, these observations strongly suggest that inversions starting with a half-space resistivity of 10 Ωm yield a low-resistivity layer in a similar depth range as the smectite-rich alteration zone in LV-11.

Furthermore, a starting model resistivity of 10 Ωm produced the lowest misfit values (both initial and final) and the fastest convergence (i.e. misfit reduced from 4.681 to 0.981 after 113 iterations). These findings suggest that $\rho_0 = 10 \Omega\text{m}$ is the closest half-space approximation to the actual resistivity values in the study area compared to the other tested initial resistivities, resulting in a solution with minimal misfit. In addition, this model also resulted in faster convergence and objective function minimization (Supplemental Material S2E).

3.3. Preferred 3-D inversion model

We finally selected TVV14 as the preferred inversion model and also renamed it TVV20 (Table 1). It was obtained from a homogeneous half-space (10 Ωm) with an initial regularization parameter $\lambda = 1$ and smoothing factors of 0.3 (horizontal) and 0.1 (vertical). After 113 iterations, it achieved an RMS misfit of 0.981. Model TVV20 provides a close fit to the data and exhibits smoother, more laterally continuous resistivity features that align with borehole log observations. The smoothing factors also help mitigate artifacts such as spurious two-layered conductors. The preferred model TVV20 is shown in Fig. 8.

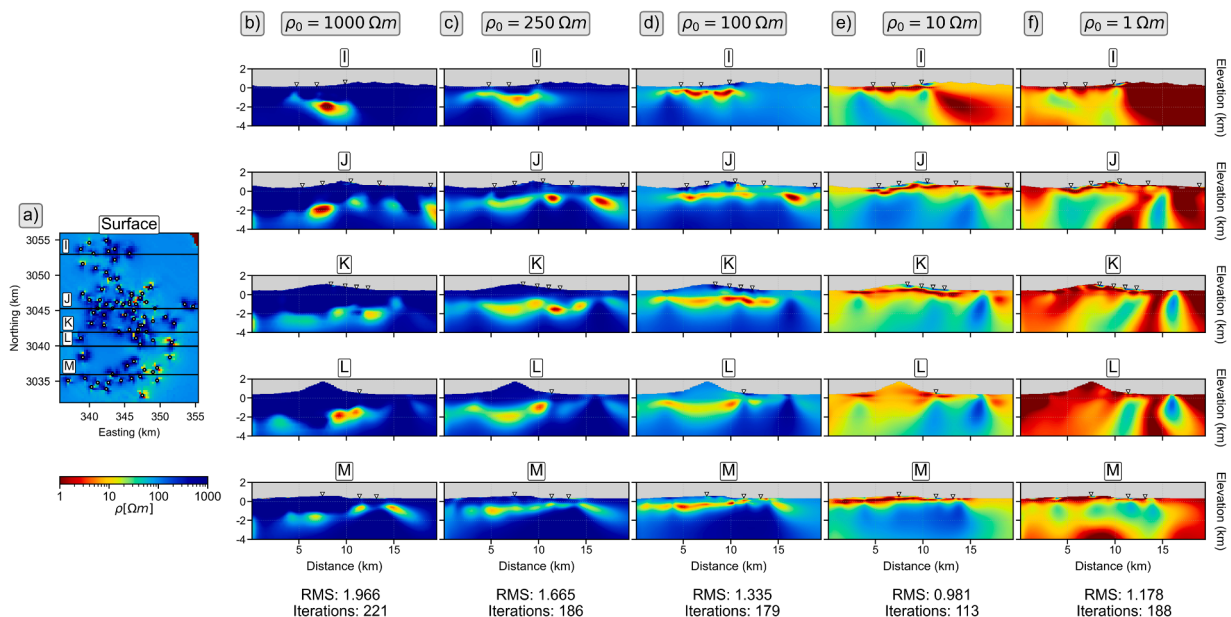


Fig. 6. Comparison of inversion tests for different half-space starting models. Each column represents an initial resistivity value, while each row represents a longitudinal cross-section as illustrated in panel a). Columns: b) 1000 Ωm , c) 250 Ωm , d) 100 Ωm , e) 10 Ωm , f) 1 Ωm . In all cases, the overall misfit values were relatively low, suggesting that the dataset supports multiple families of models. In all cases, a layered structure was consistently resolved at relatively stable conductance. Thickness and modelled resistivity are highly sensitive to the initial model, requiring ancillary information to select the optimal starting resistivity value.

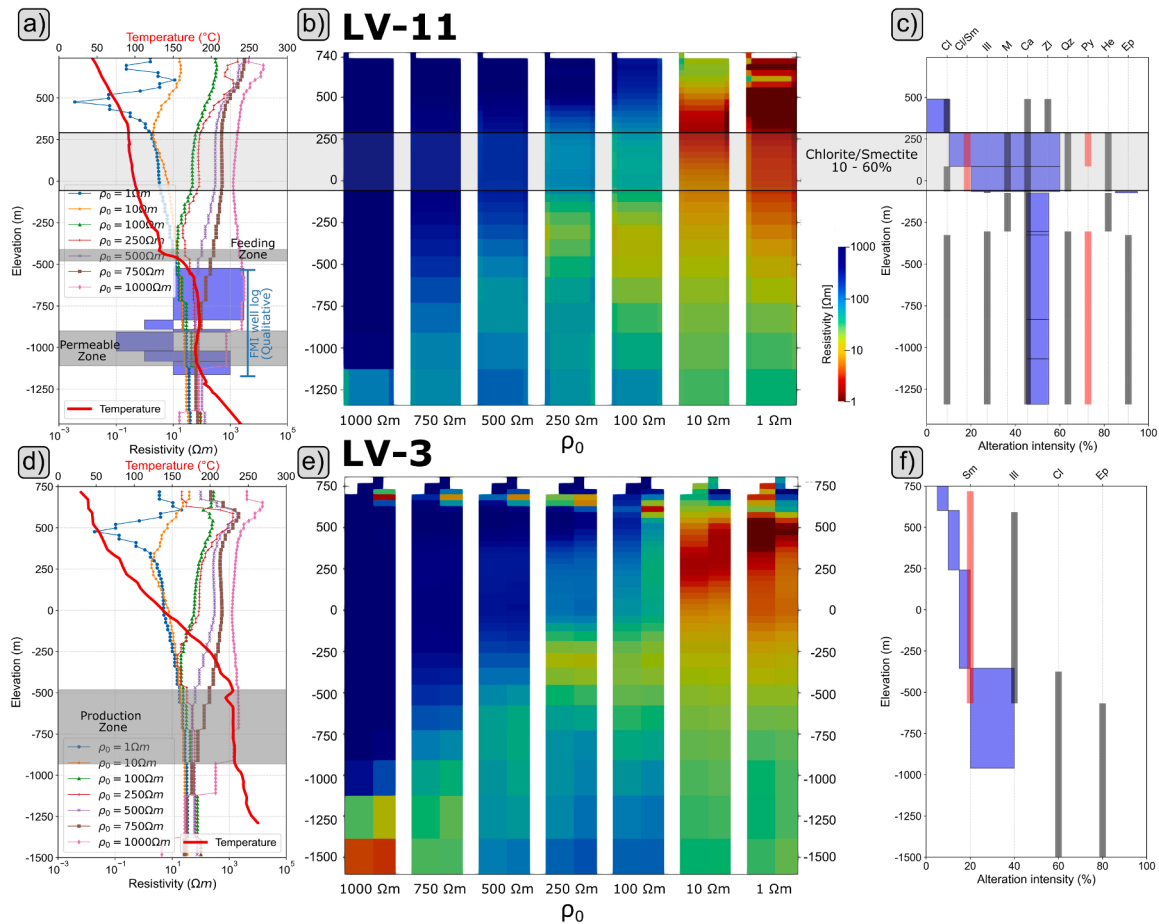


Fig. 7. Schematic comparison of well log data for LV-11 and LV-3 and modelled resistivity obtained from different starting models (i.e. 1, 10, 100, 250, 500, 750, 1000 Ωm). a) temperature (red line) and qualitative resistivity (blue rectangles) well logs, with highlighted permeable zones (gray boxes) (Gutiérrez-Negrín et al., 2001). Modelled resistivity profile curves (this work) at LV-11 shown for comparison. b) resistivity models (this work) from different starting models sampled using a cylinder of 400m diameter centered at the wellhead of LV-11. c) hydrothermal alteration mineral records at well LV-11 (Gutiérrez-Negrín et al., 2001). d) temperature well log (red) with highlighted production zone (gray box) (Santoyo et al., 2000) and modelled resistivity profiles (this work) for LV-11. e) resistivity models (this work) from different starting models sampled using a cylinder of 400m diameter centered at the wellhead of LV-3. f) hydrothermal alteration mineral zones at well LV-3 interpreted from González-Partida et al., 2005).

The fit of the impedance response of the preferred model is illustrated in Fig. 4, both as pseudo-sections and as maps for selected frequencies. This comparison reveals a smoothing effect in the modelled responses compared to the observed data, as presented in the first and second rows from top to bottom in Fig. 4a-c. Apparent resistivity (Fig. 4a), phase (Fig. 4b), and tipper (Fig. 4c) values show a close agreement with the observed data at most sites. The normalized misfits (or residuals) for each impedance tensor component (third and fourth rows from top to bottom in Fig. 4a-c) indicate acceptable fits for the diagonal components (Z_{xx} , Z_{yy}) across the entire frequency range. The off-diagonal components (Z_{xy} , Z_{yx}) show higher misfits at some stations in the southern part of the study area.

3.3.1. Characterization of the resistivity structure of inversion TVV20

The preferred resistivity model (TVV20) is characterized by a layered resistivity structure, as shown in Fig. 8. It consists of a prominent regional conductive layer (C2) with resistive layers above and below (R1, R2), and adjacent to a local shallow conductor (C1). The regional conductor exhibits significant horizontal variation, displaying lower resistivity values towards the northeast (C2a) and center (C2b) of the study area. The main conductor is also affected by a spheroidal-shaped anomaly that interrupts its lateral continuity near the El Azufre Fault (C2c). The overall layered configuration is consistent with the MT responses described in section 3.1, where the prevalent pattern exhibits a

three-layer structure indicated by high-low-high and low-high-low trends in the apparent resistivity and phase curves, respectively, which in turn are locally influenced by galvanic effects from shallow structures.

The main resistivity features identified in the TVVC include:

- Resistor R1. Consists of scattered and discontinuous near-surface regions showing resistivity values in range 20 – 1000 Ωm .
- Conductor C1. Includes a near-surface continuous thin layer, adjacent to the El Azufre Valley, and associated with hydrothermal surface manifestations (white diamonds in Fig. 8). It forms a distinguishable feature above 600 masl, where it appears as a two-level structure partially overlying the regional conductor C2. It exhibits resistivity values between 0.1 and 5 Ωm .
- Conductor C2. This is the regional conductor, forming a continuous layer extending throughout the entire study area. Local variations caused by resolution artifacts and/or internal structure heterogeneity are observed. Typical thickness ranges between 250–500 m, with highly variable depth ranging from a few hundred meters to sub-aerial exposure. The top of this layer mirrors topography, especially beneath the El Viejo and El Azufre volcanoes. This layer has a resistivity in the range 0.1 – 5 Ωm .
- Conductor C2a. The northern part of the regional conductor exhibits lower resistivity values, and forms a strong conductor between 100

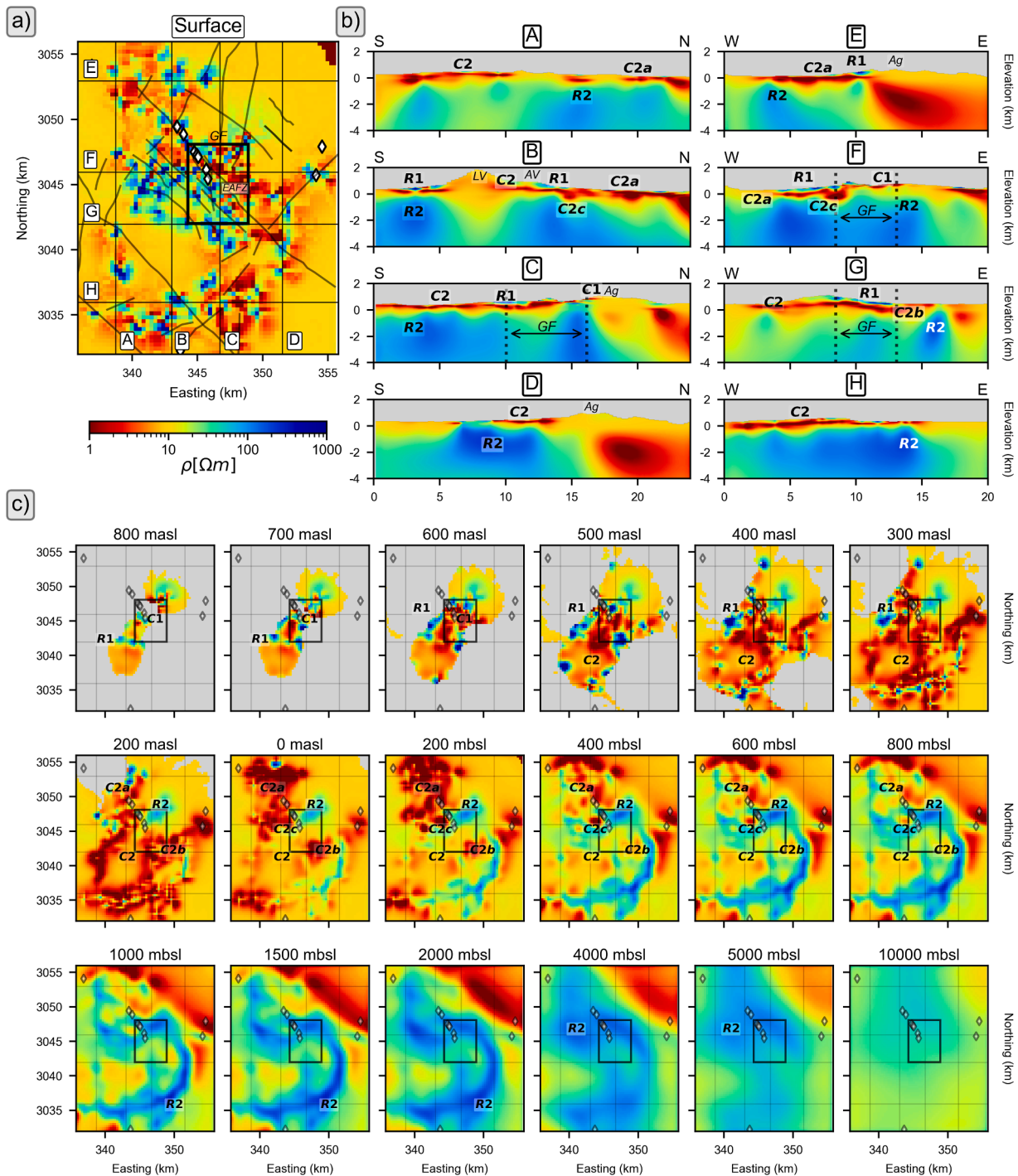


Fig. 8. Preferred model (TVV20) that minimizes RMS misfit (i.e. 0.981 after 113 iterations starting from a 10 Ωm half-space). a) Surface resistivity map showing the location of cross-sections A-H and MT stations. b) Vertical cross-sections oriented N-S and W-E. c) Horizontal cross sections at selected depths. Abbreviations: La Virgen Volcano (LV), El Azufre Volcano (AV), Aguajito Caldera (Ag), Geothermal Field (GF). Hot spring locations: white diamonds. Animated view is included in the supplemental materials.

masl and 100 mbsl. The northern edge of this feature is in proximity to the Gulf of California.

- Conductor C2b. The central part of the regional conductor forms a distinguishable feature between 200 masl and 300 mbsl. This part of the conductor extends across the geothermal field (Fig. 8b, section G).
- Conductor C2c. A small strong conductor is resolved as a spheroidal shaped anomaly straddling the El Azufre Fault. This region occurs near the boundary between C2a and C2b.

- Resistor R2. The deep portion of the model consists of a resistive region between 20 and 300 Ωm that extends to the resolution limits of the model. This resistor forms a curved ridge separating conductive regions in the southeast of the study area between 0 and 750 mbsl. Notably, no deep conductors are observed within the basal resistor beneath the regional conductor.

The layered resistivity structure was observed in a broad range of inversions, indicating that it was required to fit the measured MT data.

Near-surface resistors (R1) are required to fit the high frequency responses and are part of a three-layer structure: R1 (resistor), C2 (regional conductor), and R2 (deep basement resistor). This structure explains the V-shaped apparent resistivity trends shown in Fig. 3. Local deviations, such as shallow conductors C1, introduce site-specific complexity. Moreover, a small portion of the model domain, especially in peripheral regions distant from the MT station coverage, shows minimal to no deviation from the initial resistivity model (10 Ω m). These regions of the model consistently showed low responsiveness throughout the inversion process, indicating parts of the model domain that are poorly constrained by the observed data. Similarly, strong edge effects are seen as a deep strong conductor in the northeast corner of the study area between 0 and 1,250 mbsl. These artifacts are only present in models that use starting resistivities of 10 and 1 Ω m (Fig. 6e,f).

The resistivity model presented in this study has several features that are similar to those identified in earlier works, while also introducing new features not previously reported. For instance, the main regional conductor C2 was previously reported by Romo-Jones et al. (2000, 2005) and Antayhua-Vera (2017). However, the inclusion of topography and the 3D inversion of full impedance tensor and tipper vector data allow a closer fit to the observed MT responses. This revealed the presence of a local conductive structure C1 that is spatially associated with hydrothermal activity along the El Azufre Fault (Figs. 2, 8a). Similarly, the top of the main conductor C2 can be more accurately resolved and delineated. Furthermore, no evidence was found for deep conductive structures, such as the vertical features between 3.5 and 12 km depth reported by Romo-Jones et al. (2005), or the deep conductors between 2 and 7 km depth described by Antayhua-Vera (2017). The shallow conductive layer identified by Flores (2003) as the Cuevel hydrothermal upflow zone is comparable to the C2b conductor observed in Fig. 8c. A comparison of the 2D model from Romo-Jones et al. (2005) and the TVV20 3D model, extending to 20 km b.s.l., is included in the Supplemental Material S4.

3.4. Sensitivity analysis

The absence of deep conductors in model TVV20 (Fig. 8) marks a significant departure from previous inversions of this MT data set (Romo-Jones et al., 2005; Antayhua-Vera, 2017), and may have significant implications for the geothermal and volcanic systems in the TVVC. A sensitivity analysis was therefore conducted to assess whether these differences are required by the MT data. This assessment is a fundamental component of magnetotelluric analysis, as it allows the identification of inversion artifacts. EM methods have a limited capacity to resolve resistive structures beneath conductive layers. This limitation arises from the physics of EM signal propagation, where strong attenuation within conductors hinders the detection of deeper conductive features (Telford et al., 1990).

Given this limitation, a series of forward modeling tests was performed to evaluate the actual resolution capabilities of the dataset. Four tests were designed to compare the preferred model (TVV20) with synthetic responses generated from simplified resistivity models that approximate the expected configuration in the TVVC. The complexity of the simulated structures increased progressively with each test:

- TVV01-S: A three-layered model representing the main regional conductor C2.
- TVV02-S: A four-layered model including a local shallow conductor C1.
- TVV03-S: A five-layered model containing an intermediate resistivity volume in the geothermal field, simulating the reservoir beneath the conductive cap seal.
- TVV04-S: An eight-layered model with deep, strong conductors, simulating the configuration of the magmatic system proposed by Avellán et al. (2018).

The forward response was calculated at the same locations and frequencies as in the observed MT data, incorporating 5 % random noise. This response was subsequently inverted using an initial homogeneous model at 10 Ω m, using similar inversion parameters to TVV20 (Table 1). The objective of these tests is not to reconstruct the true model in full detail, but rather to assess the ability of the dataset to resolve specific features. These models have been deliberately simplified to represent first-order resistivity structures.

3.4.1. TVV01-S: Simplified three-layered model

The first synthetic test evaluates the simplest resistivity structure capable of replicating the MT response of the preferred model. Based on the types of MT response curves described in Section 3.1, and the modelled resistivity structure (TVV20), a three-layered synthetic model incorporating topography was constructed (Fig. 9a,b). This model consists of:

- L1: A top resistive layer (100 Ω m) extending from 200 masl to the surface, representing R1.
- L2: An intermediate conductive layer (5 Ω m), extending from 50 mbsl to 200 masl, corresponding to C2.
- L3: A deep resistive layer (250 Ω m) below 50 mbsl, representing R2.

The RMS misfit was 5.966 for the starting model and was reduced to 0.775 after 107 iterations. The inversion results (Fig. 9c) confirm that the basic three-layer geometry of the true model is resolved. The geometry of the conductive layer is more accurately delineated in regions with denser site coverage, while areas with sparse station distribution exhibit greater distortion. Modelled resistivity values show spatial heterogeneity, correlating with station distribution. L1 resistivity tends to be overestimated, while L2 and L3 are generally underestimated. Moreover, the top layer L1 is only partially resolved beneath the MT sites, and its thickness tends to be underestimated.

The limits of the resolution capabilities of the dataset are also evident. In the northeast corner of the study area, a strong conductive region appears as a resolution artifact (Section I, Fig. 9). Similarly, topographic features above MT stations exhibit low sensitivity in the inversion (e.g. La Virgen Volcano, in section L, Fig. 9). Differences in station coverage produce lateral variability in all layers.

3.4.2. TVV02-S: Simplified layered model with thin shallow conductors

The second synthetic test simulated the presence of a shallow conductor C1, introducing a rectangular conductor L4 to the TVV01-S model (Section S3B in Supplemental Material):

- L4: A near-surface rectangular body representing the geometry of C1. It has a resistivity of 5 Ω m, with dimensions of 7000 m x 7000 m x 250 m, and with a main axis oriented 60° to the NW.

The inversion converged after 143 iterations, reducing RMS misfit from 6.840 to 0.796 (see Fig. S12 in Supplemental Material S3B). The inversion results indicate that L2 is well resolved, especially in regions with the densest MT station coverage. The shallow conductor L4 was distinctively resolved with MT data, but it introduced local perturbations in the lateral continuity of L2.

Compared to the TVV20 inversion, the continuous conductive layer L2 acts as a low-resolution approximation of the modelled anomaly C2. The shallow conductors are not necessarily resolved as discrete structures, but rather manifest as discontinuities in depth, thickness, or resistivity within the main conductor.

3.4.3. TVV03-S: Simplified layered model with permeable reservoir

The next synthetic inversion builds upon TVV02-S, incorporating a region with medium resistivity (L5) beneath the main conductor. This configuration simulates the presence of a geothermal reservoir with intermediate resistivity beneath a conductive clay cap (Pellerin et al.,

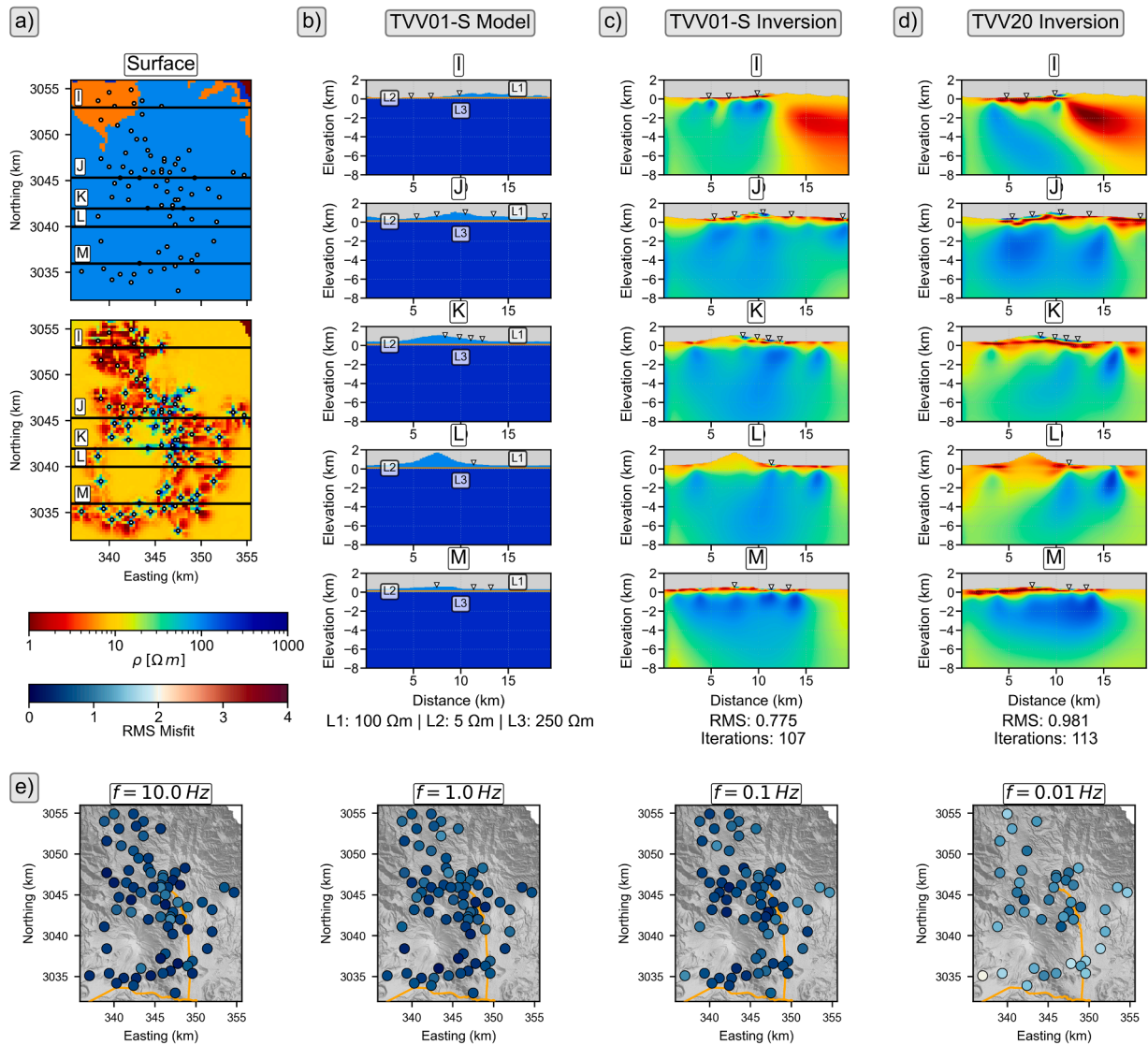


Fig. 9. Simplified three-layered synthetic model (TVV01-S). a) Surface view of the true (top) and inverted (bottom) models, showing the I-M cross-sections location; b) synthetic model dissected at sections I-M; c) results from the inversion of the forward responses from the synthetic model; d) preferred TVV20 model for reference; e) normalized misfit at selected frequencies.

1996), and it maintains a distance from the cap seal to the production zone as encountered in well LV-3 (Fig. 7):

- **L5:** A rectangular volume (30 Ωm) measuring 3300 $\times$ 3300 $\times$ 670 m, and aligned with the N-S axis. It is located in the center of the study area, within the geothermal field.

RMS misfit was reduced from 6.840 to 0.822 after 145 iterations. The inversion results are presented in Supplemental Material S3C.

Subtle differences between models TVV02-S and TVV03-S were observed in the vicinity of L5. However, it remains uncertain whether these variations result from shading effects induced by L4 or partial resolution of L5 (Section J, in Supplemental Material S3C).

3.4.4. TVV04-S: Simplified layered model with magmatic conductors (following Avellán et al., 2018)

The fourth synthetic inversion evaluated the sensitivity of the dataset to deep conductive features associated with the TVVC magmatic system. The geometry of these conductors was based on the magmatic configuration described by Avellán et al. (2018). These authors used petrological and magnetic data to infer the presence of two small, shallow

magma reservoirs (500 – 1000 mbsl) and a larger, deep reservoir at depths of 5000 and 7000 mbsl. These features were incorporated into the TVV03-S synthetic model (Fig. 10) as follows:

- **L6:** An oblate conductive lens with a resistivity of 1 Ωm , oriented along the N-S axis, measuring 3300 m $\times$ 2600 m, embedded within the main anomaly L5 and the deep resistive layer L3, located beneath El Azufre and El Viejo Volcanoes (Figs 2 and 10b Section J).
- **L7:** A conductive lens similar to L6, measuring 3600 m $\times$ 4500 m with a resistivity of 1 Ωm . It is located beneath La Virgen Volcano (Figs 2 and 10b Sections K-M).
- **L8:** A conductive spheroid with a resistivity of 1 Ωm . It is oriented along a major axis trending N60E, its dimensions are 12 km $\times$ 8.5 km $\times$ 8 km, and is embedded within the main resistor L3.

A resistivity of 1 Ωm was assigned to these conductors to simulate zones of partial melting. The TVV03-S model described above was used as the base configuration, with the resistivity of the shallow conductor L5 modified to 5 Ωm . This adjustment was introduced to investigate the attenuation effects induced by stronger conductors.

The inversion results, presented in Fig. 10c, reveal no clear anomaly

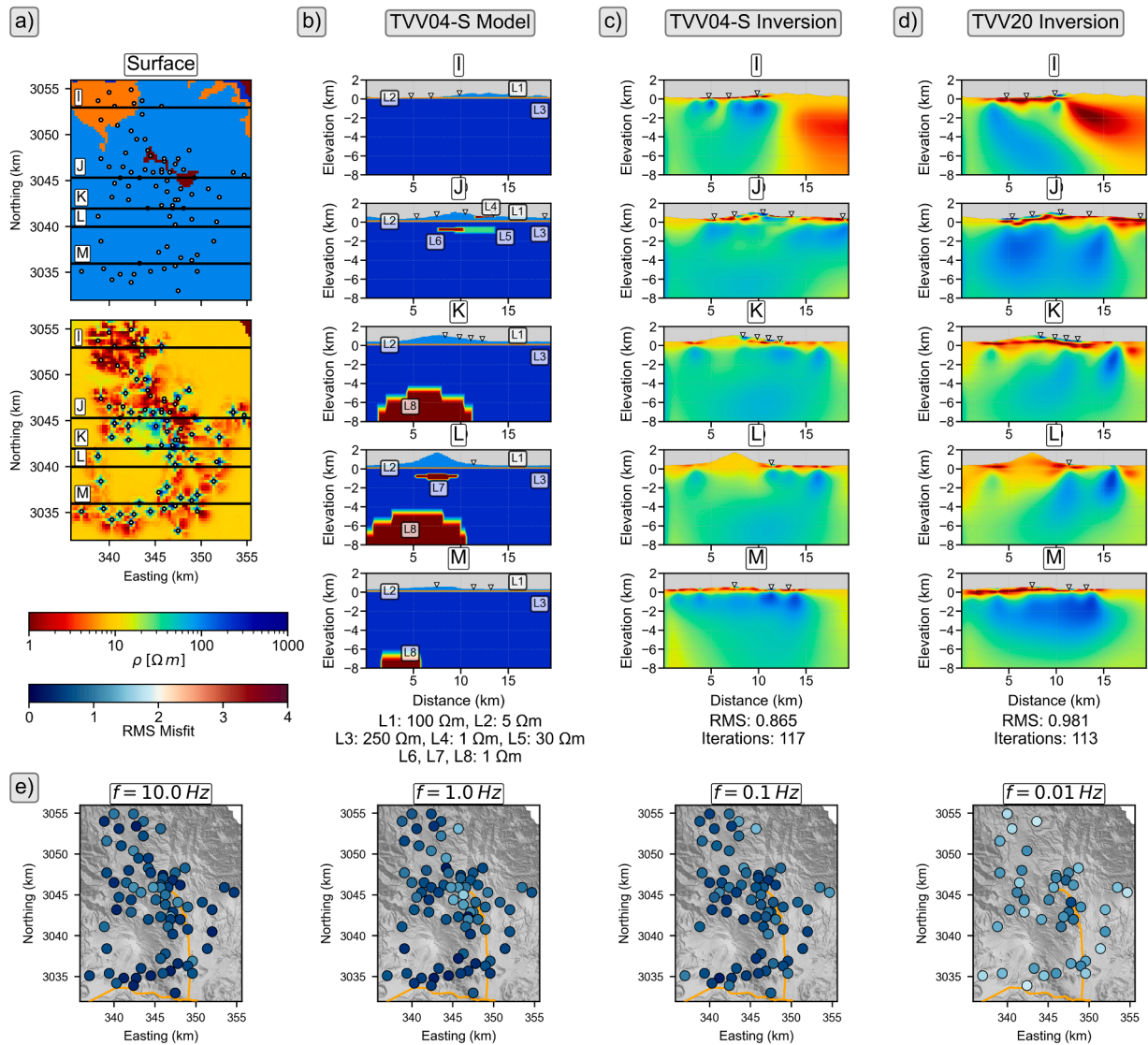


Fig. 10. Synthetic model (TVV04-S) with deep magmatic structure as depicted in [Avellán et al. \(2018\)](#). a) Surface view of the true (top) and inverted (bottom) models, showing the I-M cross-sections location; b) synthetic model dissected at sections I-M; c) results from the inversion of the forward responses from the synthetic model; d) preferred TVV20 model for reference; e) normalized misfit at selected frequencies.

associated with the deep conductor *L8*. However, a faint discontinuity in the basal resistor *L3* is observed, but it remains almost indistinguishable from resolution artifacts. This suggests that the dataset exhibits low sensitivity to large conductive bodies beneath the main conductive layer, requiring lower frequencies than currently available for distinct resolution.

In contrast, the flat top of *L2* is well-resolved except for a noticeable gap beneath *L4*. The oblate conductors introduce distortions to the lateral continuity of the main conductor, making it appear deeper and more conductive. Compared to the TVV20 model, the lateral heterogeneities within the main conductor *C2* can be attributed to the presence of embedded stronger conductors.

These four synthetic inversions suggest that the minimum structural complexity required to explain the TVVC resistivity model includes four layers: 1) a basal resistor *R2*, 2) a regional conductive layer *C2*, 3) a shallow resistor *R1*, and 4) a thin, near-surface conductor *C1*. Additionally, the synthetic tests indicate that deep conductors, associated with the magmatic system of TVVC, cannot be distinctively resolved with the current dataset. Moreover, strong, smaller conductors adjacent to the regional conductor induce localized distortions, altering the lateral continuity, conductivity and interpreted depth of the main

conductor.

4. Resistivity model interpretation

In active volcanic settings, the resistivity structure is influenced by multiple factors commonly associated with hydrothermal and magmatic processes. To support the interpretation of anomalies observed in inversion models, laboratory studies of rock resistivity are required. In this section, we examine the potential contributions of the three most common conductive phases found in volcanic environments, with reference to conditions present in the TVVC.

4.1. Possible cause of low resistivity in TVVC

Electrical resistivity is a highly variable petrophysical property, particularly in geothermal and volcanic environments, where it can be influenced by factors that include temperature, mineralogy, permeability, pore fluid, and melt fraction ([Ucock et al., 1980](#); [Caldwell et al., 1986](#); [Lašovičková, 1991](#); [Tyburczy, 2007](#); [Unsworth and Rondenay, 2012](#); [Pommier, 2014](#)). These parameters can have complex interrelationships that require prior knowledge of the geothermal system,

together with laboratory measurements to accurately interpret the resistivity values derived from MT data.

To support the model interpretation, a set of empirical constraints from experimental studies was established. This analysis examines the relationship between bulk resistivity and variable parameters, including brine-saturated porosity, clay fraction, and melt fraction. Given the inherent uncertainty in these parameters, a Monte Carlo approach was applied to quantify their effects on modelled resistivity.

Crystalline rocks, with minimal fracturing and hydrothermal alteration, generally exhibit high resistivity values, often exceeding 1,000 Ωm . In contrast, a number of materials can produce low resistivity, often orders of magnitude lower. These materials are systematically described in the following subsections. A common scenario in Earth systems is that the conducting material occupies a relatively small fraction of the rock volume. This means that a calculation is needed to compute the overall (bulk) resistivity that results from the high resistivity host rock and one or more low resistivity phases. A number of empirical relations, known as mixing laws, have been proposed. In this study, we apply a generalized form of Archie's law for multiple conducting phases (Archie, 1942; Glover, 2010) to calculate the ranges of conductive phase fractions that explain the modelled bulk resistivity under different conditions.

4.1.1. Low resistivity due to brines in unaltered rock

In the first scenario, low resistivity is caused by conduction through a brine contained in fractured rock. To analyze this system, two steps are needed: 1) calculate the resistivity of the brine, and 2) determine the bulk resistivity when the brine is present in a rock with a specified porosity and fluid distribution. For the first step, fluid resistivity (ρ_w in Ωm) is controlled by salinity (g/L) and temperature ($^{\circ}\text{C}$) (Sen and Goode, 1992) (Fig. 11a). For the second step, bulk resistivity is calculated using Archie's law (Archie, 1942). This mixing law relates bulk resistivity (ρ), fluid resistivity (ρ_w), porosity (ϕ), and cementation factor (m):

$$\rho = \rho_w \phi^{-m}$$

The cementation factor describes the interconnectivity of the pore space, with higher values indicating a lower degree of fluid interconnection. As a result, this factor is directly related to permeability. Fig. 11c illustrates the relationship between porosity and bulk resistivity for unaltered rocks saturated with brine for different cementation factors, and at different temperatures. The bulk resistivity for rock saturated with seawater (salinity 34.4 g/L) at selected cementation factors and temperatures (red lines) is shown for reference. The Fig. indicates a rapid decrease in resistivity with increasing saturated porosity, with faster decay observed in higher permeability media.

To explore the range of possible temperature, salinity, cementation and porosity values that could explain the resistivity of conductor C2 in TVVC, a stochastic estimation approach was employed, treating these parameters as random variables. The temperature was modelled using a triangular probability distribution, with values ranging from 24 $^{\circ}\text{C}$ (ambient T) to 300 $^{\circ}\text{C}$ (maximum reservoir T) (Portugal et al., 2000; Prol-Ledesma et al., 2016). Salinity values from fluid inclusions measured by Verma et al. (2006), ranging from 7 to 192 g/L, were used as bounds for uniformly distributed random realizations (for comparison, the average salinity in the Gulf of California is 34.4 g/L (Reagan et al., 2024)). This wide range of salinities also includes the values of borehole samples measured by Birkle et al. (2016) (9.6–15.3 g/L). The hypersaline values have been reported in the southern part of the Tres Virgenes geothermal field, which has been interpreted as dissolution of local evaporitic lenses (Hernández-Hernández et al., 2019). The cementation factor and porosity values were uniformly distributed in the ranges 1–2 and 0–1, respectively. The random variables used in this calculation are listed in Table 2.

These parameters were used in a Monte Carlo simulation with 10 million realizations. The porosity ranges corresponding to observed bulk resistivity, given the conditions of temperature, salinity, and

cementation expected in TVVC, are presented in Fig. 11c. A list of porosity ranges estimated for selected resistivity values is presented in Table 3.

4.1.2. Partial melt

In the second scenario, low resistivity is attributed to the presence of partial melt. Conduction is primarily electrolytic, controlled by the melt fraction and its degree of interconnectivity, with some conduction also occurring through the rock matrix. The resistivity of the andesitic melts is primarily controlled by temperature, pressure, water content, and the presence of Na^+ ions as a charge carrier (Guo et al., 2017). To estimate andesitic melt resistivity, the experimental model of Guo et al. (2017) was used (Fig. 11b). The resistivity of the solid rock grains was estimated using the model from Hashim et al. (2013).

Resistivity decreases with increasing temperature, due to enhanced mobility of the charge carriers. Similarly, water in the melt dissociates into H^+ and OH^- ions, further reducing resistivity. By selecting representative values for pressure, temperature, and water content, the resistivity of the melt and rock was estimated.

Since Archie's law accounts for only one conducting phase, a generalized form of the law was applied for n conducting phases (Glover, 2010). This form accounts for contributions from both melt and rock matrix:

$$\rho = \left(\sum_i^n \rho_i^{-1} \phi_i^m \right)^{-1}$$

where, ρ_i , m_i , and ϕ_i represent the resistivity, cementation factor, and relative fraction of phase i , respectively. Cementation exponents were calculated iteratively using the formula:

$$m_j = \log(\phi_j^{m_j}) / \log(\phi_j)$$

and meeting the conditions presented in Glover (2010).

These values were used in the generalized Archie's equation to determine the relationship between melt fraction and bulk resistivity in a partially molten zone. To account for variations in relevant parameters in TVVC, water content, temperature, and pressure were treated as random variables. Symmetric triangular distributions were used to represent their expected ranges based on empirical data. Water content and temperature values from Guo et al. (2017) were used, and ranged from 0.01–5.9 wt % of H_2O and 1164–1573 K. Pressure estimates were derived from the lithostatic pressure of a 1 km thick overburden of variable density (i.e. 23–28 MPa). The melt fraction ranges associated with different values of bulk resistivity are presented in Fig. 11d and Table 3.

Fig. 11d illustrates the relationship between bulk resistivity and melt fraction for different cementation factors. In the preferred model (TVV20), most conductive features are laterally continuous and relatively shallow, making a magmatic origin unlikely. However, conductor C2c exhibits a spheroidal shaped geometry with bulk resistivity in the range 1–10 Ωm . The resistivity range encountered in C2c corresponds to melt fractions in the range 33–85 % (Table 3), depending on melt interconnectivity. Since these values are physically plausible, the presence of a localized andesitic mush lens remains a viable interpretation for C2c.

4.1.3. Hydrothermally altered non-saturated rock

The third scenario explores the contribution of clay minerals to the bulk resistivity of non-saturated rock. In geothermal systems, water-rock interaction transforms volcanic glass and plagioclase into smectite, illite, and chlorite (Steiner, 1968). When clay minerals are present, surface conduction occurs. These clay minerals enhance electrical conductivity through a double layer of ions on the crystal surfaces. The magnitude of this process is controlled by the cation exchange capacity (CEC) of the minerals. Consequently, smectite (CEC range: 80–150 meq/100g, where

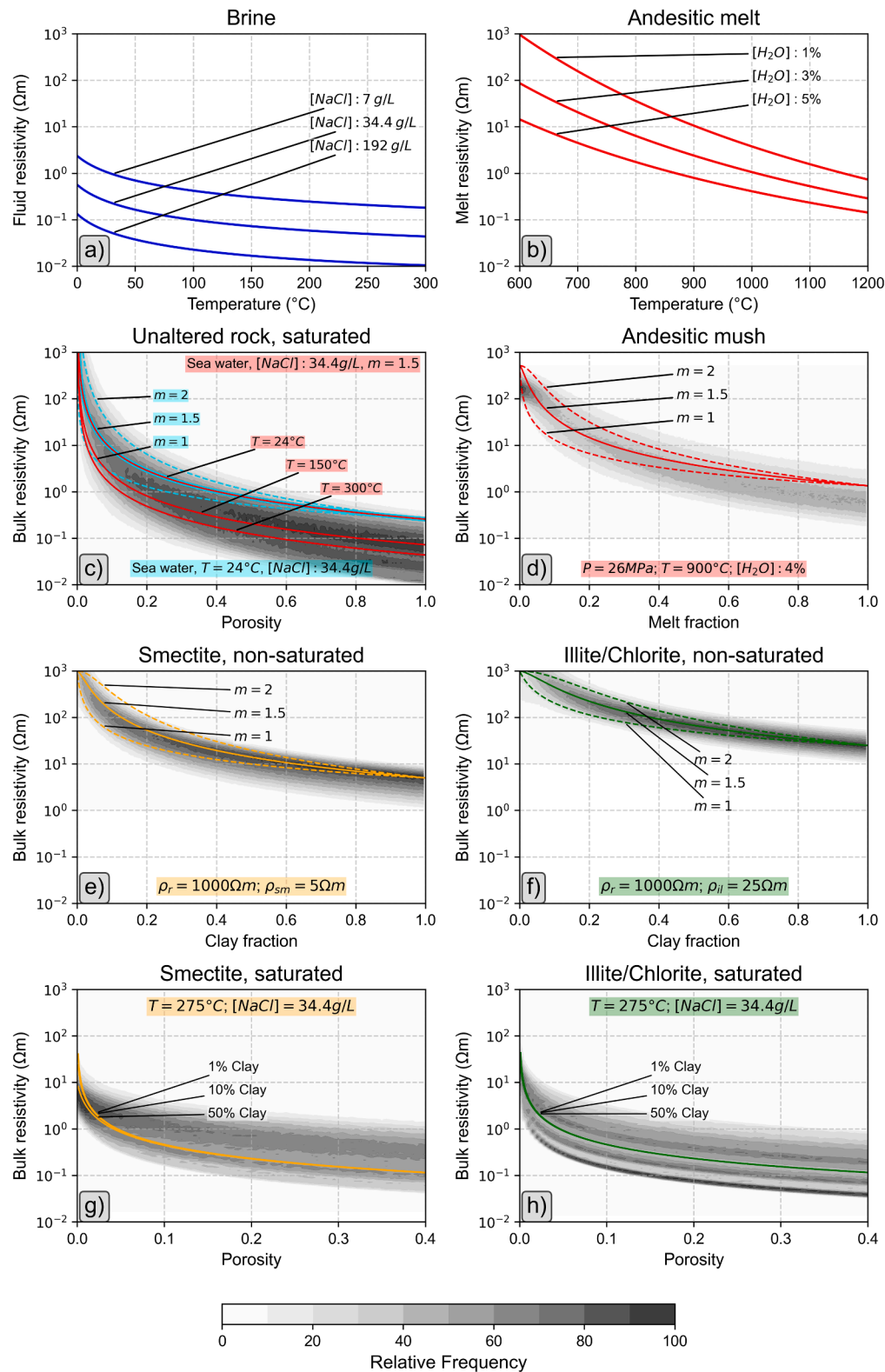


Fig. 11. Estimation of the proportion of conductive phase fraction required to explain modelled bulk resistivity in TVVC under different scenarios. a) relationship between temperature and fluid resistivity at different salinities, b) resistivity of andesitic melt as a function of temperature for different water content. Six different scenarios are considered where low resistivity is caused by: c) brines in fractured unaltered rock (sea water, with salinity of 34.4 g/L for reference); d) partial melt of andesitic composition (melt at 26 MPa, 900°C, and 4 % $[H_2O]$ for reference); e) non-saturated rock with smectite alteration; f) non-saturated illite or chlorite alteration; g) brine-saturated rock, smectite-altered; h) brine-saturated rock, illite- or chlorite-altered. Shaded areas indicate the relative frequency of the bulk resistivity, estimated using a stochastic approach using parameter variability ranges expected in TVVC. Colored lines indicate the bulk resistivity obtained using fixed parameters (displayed for reference).

Table 2

Distribution of random variables used in the stochastic analysis of porosity/clay fraction/melt fraction for the different scenarios considered.

Case	Variable	Distribution	Minimum	Mode	Maximum
Unaltered, brine-saturated rock	Fluid temperature (°C)	Triangular	24	270	300
	Salinity (g/L)	Uniform	7		192
Hydrothermal alteration, non-saturated rock	Clay fraction	Uniform	0		1
	Resistivity: smectite (Ωm)	Triangular	1	5	10
	Resistivity: illite/chlorite (Ωm)	Triangular	10	25	60
Hydrothermal alteration, brine-saturated rock	Fluid temperature: smectite (°C)	Uniform	40		150
	Fluid temperature: illite/chlorite (°C)	Uniform	150		200
	Melt temperature (°C)	Triangular	900	1100	1300
Andesitic mush	Lithostatic pressure (MPa)	Triangular	23	25.5	28
	Water content (%)	Triangular	1	2	3
	Resistivity: rock (Ωm)	Triangular	100	1000	10000
All	Porosity/clay fraction/melt fraction	Uniform	0		1
	Cementation	Triangular	1.0	1.5	2.0

Table 3

Stochastic ranges of porosity, clay fraction, or melt fraction (as percentages) required to explain selected bulk resistivity values for different conductivity mechanism scenarios. The third, fourth, and fifth columns represent the bounds and most likely values of the 95 % confidence interval for the stochastic model.

Porosity, clay and melt fractions required to explain modelled bulk resistivity for selected values (95 % confidence interval)				
Case	Bulk resistivity (Ωm)	Lower limit (%)	Most likely (%)	Upper limit (%)
Unaltered, brine-saturated rock (Porosity) Fig. 11c	300	0.30	3.31	7.23
	100	0.70	5.25	11.25
	50	0.94	7.25	13.29
	20	1.73	9.12	19.61
	10	2.26	13.81	26.23
	5	3.50	16.60	32.82
	1	8.32	34.68	67.89
Smectite alteration, non-saturated rock (Clay fraction) Fig. 11e	300	0.66	6.64	12.21
	100	3.77	13.90	22.10
	50	8.22	22.77	33.75
	20	16.75	40.45	59.98
	10	26.67	66.30	88.37
	5	41.56	92.11	99.15
	1	Out of range	Out of range	Out of range
Illite/chlorite alteration, non-saturated rock (Clay fraction) Fig. 11f	300	2.53	18.47	29.91
	100	19.15	39.50	62.00
	10	Out of range	Out of range	Out of range
	1	Out of range	Out of range	Out of range
	1	Out of range	Out of range	Out of range
Smectite alteration, brine-saturated rock (Porosity) Fig. 11g	300	Out of range	Out of range	Out of range
	100	Out of range	Out of range	Out of range
	50	0.01	0.33	1.22
	20	0.01	0.53	2.68
	10	0.03	1.52	5.53
	5	0.12	2.66	12.12
	1	3.20	25.42	76.19
Illite/chlorite alteration, brine-saturated rock (Porosity) Fig. 11h	300	Out of range	Out of range	Out of range
	100	0.00	0.06	0.23
	50	0.01	0.17	0.52
	20	0.04	0.40	1.42
	10	0.15	1.11	3.11
	5	0.35	2.20	6.91
	1	2.00	13.05	36.34
Andesitic mush (Melt fraction) Fig. 11d	300	0.13	3.77	13.85
	100	0.36	9.09	23.48
	50	3.12	13.55	34.61
	20	6.78	23.72	60.08
	10	10.81	32.66	71.31
	5	17.10	41.51	89.24
	1	40.55	85.22	98.27

meq = milliequivalents) exhibits a much lower resistivity, i.e. 6 to 10 times lower than illite and chlorite (CEC range: 10-40 meq/100g) (Ussher et al., 2000).

The contribution of clay minerals to bulk resistivity has been modelled by many authors (Waxman and Smits, 1968; Clavier et al., 1984; Caldwell et al., 1986). These models empirically compensate for surface conduction instead of accounting for conductivity mixing laws. Since the generalized multi-phase Archie's equation from Glover (2010) circumvents this limitation, it was also used for this setting. Here, the clay fraction and the matrix of unaltered rock correspond to two phases contributing to bulk resistivity.

Based on borehole log data from wells LV-3, LV-5, LV-7 and LV-8, González-Partida et al. (2005) presented cross sections including a hydrothermal alteration characterization for Tres Vírgenes, based on the occurrence of (a) smectite, (b) smectite and illite, (c) smectite, illite and incipient chlorite, (d) prevalent chlorite, illite and smectite. In these wells, the degree of hydrothermal alteration increases with depth, ranging from 0 to 40 % clay fraction. A diagram summarizing the details presented in that study that are relevant to this analysis, and comparing them with modelled resistivity, is illustrated in Fig. 12. In addition, Gutiérrez and Espíndola (2010) report the occurrence of chlorite/smectite 10-60 % hydrothermal alteration between 290 masl and 64 mbsl at well LV-11, as previously seen in Fig. 7c.

Two cases were considered for analysis: (1) smectite, and (2) a combination of illite and chlorite. This was designed to identify the end members of a mixture continuum. Illite and chlorite were lumped together since they exhibit similar CEC and resistivity ranges. Figs 11e and 11f show the relationship between bulk resistivity and clay fraction for pure smectite and pure illite.

The stochastic estimation of this relationship for pure clay minerals treated clay and rock resistivity as random variables. These parameters were sampled from a triangular distribution defined by specified ranges (Table 2), i.e. 1-10 Ωm (smectite), 10-60 Ωm (illite/chlorite), and 100-10,000 Ωm (fresh rock) (Pellerin et al., 1996). The cementation factor for the clay (m_c) and rock (m_r) fractions is uniformly distributed, ranging between 1-3 for m_c . The values of m_r were calculated to comply with the condition $\sum_i^n \phi_i^{m_i} = 1$ (Glover, 2010). The clay fraction (ϕ_c) is uniformly distributed between 0 and 1, while the remaining part was assigned to the rock fraction ($\phi_r = 1 - \phi_c$).

In the preferred model (TVV20), the conductive layers C1 and C2 are associated with bulk resistivity ranging between 1 and 10 Ωm . Fig. 11e,f illustrates the relationship between clay fraction and bulk resistivity for the parameter ranges expected in TVVC. Table 3 lists the 95 % confidence intervals for selected resistivity values. In the conductors C1 and C2, the observed resistivity can be explained by rocks with smectite alteration >66 %; however, no amount of pure illite (or chlorite) can account for this range of resistivity.

4.1.4. Hydrothermally altered rock, brine-saturated

The fourth scenario examines hydrothermally altered rock saturated

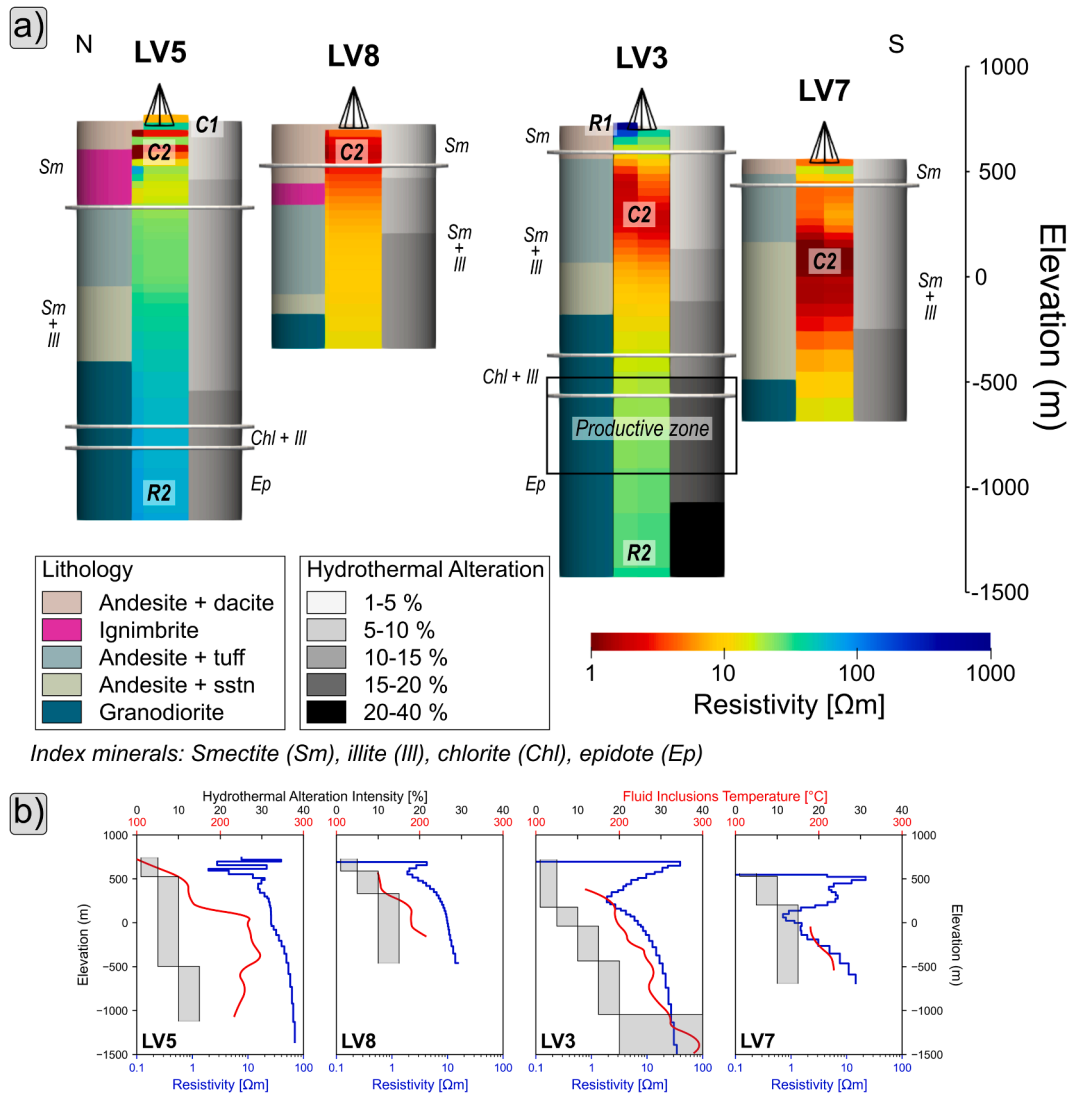


Fig. 12. a). Comparison of the resistivity model and the observed lithology and hydrothermal alteration intensity at wells LV7, LV3, LV8, and LV5, with data from González-Partida et al. (2005). Resistivity is visualized as a cylindrical sample of the model centered at the wellhead. Lithology and hydrothermal alteration intensity are represented as cylindrical columns to the left and right of the resistivity cylinder, respectively. The white rings surrounding the columns represent the boundaries of index mineral zones: Sm (smectite), Ill (illite), Chl (chlorite), Ep (epidote). Black rectangle around LV3 represents the productive zone, where convective heat transfer is dominant. **b)** Comparison of modelled vertical resistivity (TVV20) profiles at selected wells (blue), fluid inclusion temperature (red) and hydrothermal alteration intensity (gray), with data digitized from González-Partida et al. (2005).

with brine. In this context, conduction is predominantly electrolytic, with minor contributions from surface conduction on clay crystals and almost negligible bulk conduction through the rock matrix. The analysis uses the generalized Archie's law (Glover, 2010) to account for three phases: brine, clay, and the rock matrix. Fluid resistivity was calculated using the Sen and Goode (1992) model, as in Section 4.1.1, with temperature constrained by the thermal stability of the respective clay species (i.e. 40–150 °C for smectite and 150–230 °C for illite/chlorite). The cementation exponents were calculated following the method proposed by Glover (2010).

Fig. 11g–h illustrates the relationship between porosity and bulk resistivity for smectite and illite-altered rocks with clay fractions of 1 %, 10 % and 50 %. This example shows that clay fraction provides only a minor contribution to bulk resistivity when highly saline fluids are present. Moreover, the distinction between the different clay species is noticeable only when porosity is below 5 %, where surface conduction is not negligible compared to its electrolytic counterpart.

Fig. 11g–h also presents the stochastic estimation of these ranges, accounting for the variability of clay fractions, temperature, and salinity

expected in TVVC. The estimated porosity ranges required to explain selected resistivity values are listed in Table 3. The presence of brine and clay requires lower porosities (1.5–25.4 %) to explain bulk resistivity between 1–10 Ωm compared to unaltered rock (13.8–34.7 %) (Fig. 11c,g, h, Table 3). Consequently, the presence of a wet clay cap seal constitutes a more likely explanation for the conductors C1 and C2.

4.2. Implications for understanding of the Tres Virgenes geothermal system

Throughout the inversion process, a layered resistivity structure consistently appeared in the models. However, the depth and resistivity values of these layers were found to be highly sensitive to the resistivity of the starting model. The high degree of uncertainty in the modelled depth and resistivity requires the use of ancillary information to constrain the interpretation.

Fig. 12 presents a comparative diagram illustrating the correlation between resistivity, lithology, and hydrothermal alteration from bore-hole records in TVVC. Fig. 12b presents a comparison of the modelled

resistivity (TVV20) with lithology and hydrothermal alteration observed in wells LV5, LV8, LV3, and LV7, as reported by [González-Partida et al. \(2005\)](#). This comparison strongly suggests that low resistivity is more strongly correlated with hydrothermal alteration than with lithology. Fig. 12a also shows that the regional conductor C2 occurs in regions containing smectite, with varying intensities of hydrothermal alteration, and across different rock types. The cross sections in [González-Partida et al. \(2005\)](#) also include an upflow zone interpreted from isotherms constructed using fluid inclusion data.

Fig. 13 compares the resistivity model with published cross sections.

Section T correlates resistivity from model TVV20 with the isotherms and hydrothermal alteration zones reported in [González-Partida et al. \(2005\)](#). In the same Fig., sections N and O contrast the model TVV20 with the results from [Romo-Jones et al. \(2000\)](#). Similar comparisons are included for the models from [Antayhua-Vera \(2017\)](#) (sections P and Q), and from [Flores \(2003\)](#) (sections R and S). Fig. 14 offers additional context, representing the resistivity (TVV20) next to geological models adapted from [Avellán et al. \(2018\)](#) (sections U, W), and from [Gómez and Rocha \(2009\)](#) in [Antayhua-Vera \(2017\)](#).

Fig. 15 presents a panoramic 3D view of the resistivity model,

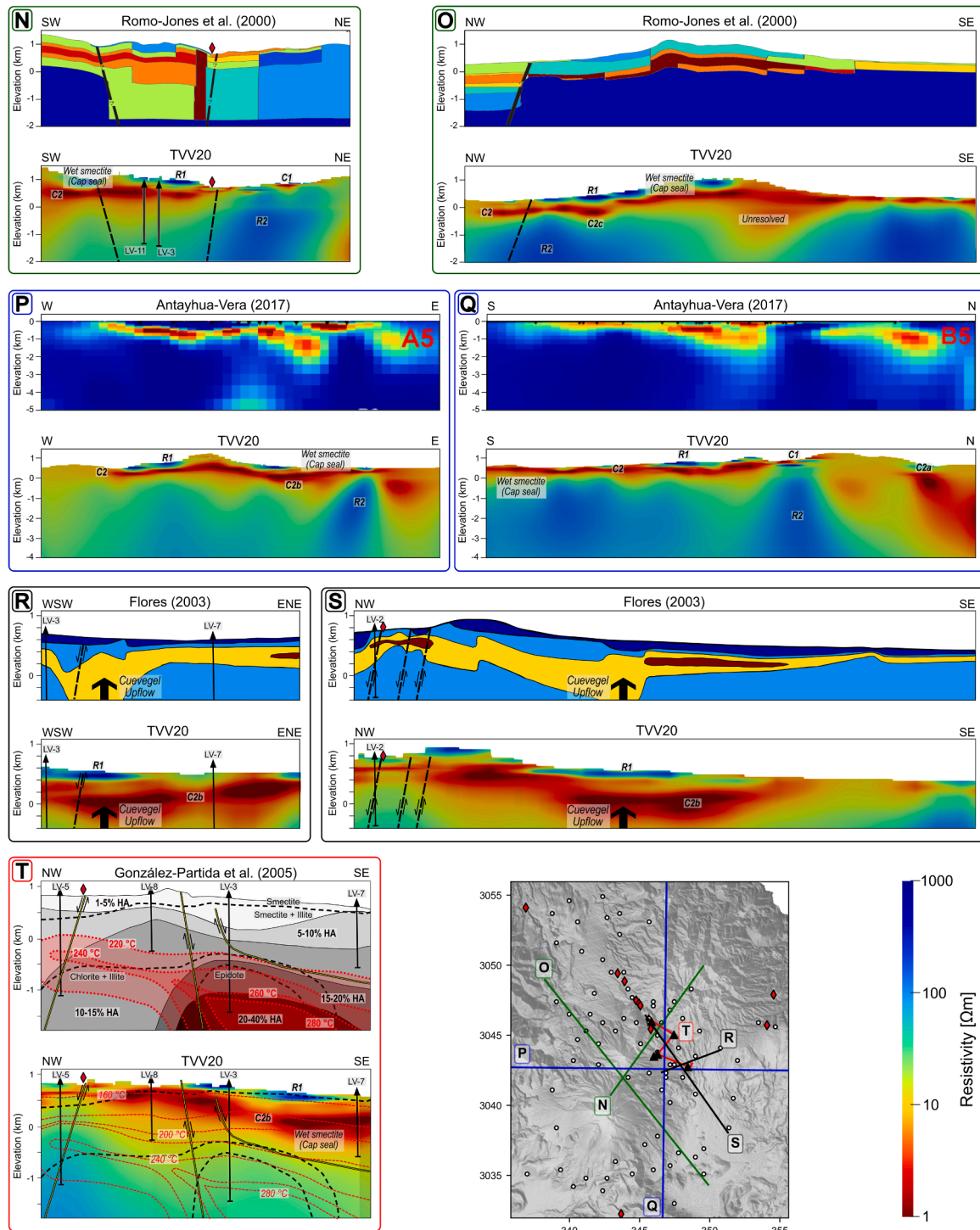


Fig. 13. Comparison of previous EM studies with preferred model TVV20. Sections N, O: [Romo-Jones et al. \(2000\)](#); sections P, Q: [Antayhua-Vera \(2017\)](#); sections R, S: [Flores \(2003\)](#); section T: [González-Partida \(2005\)](#). Location of cross sections is illustrated on the map in the lower right corner.

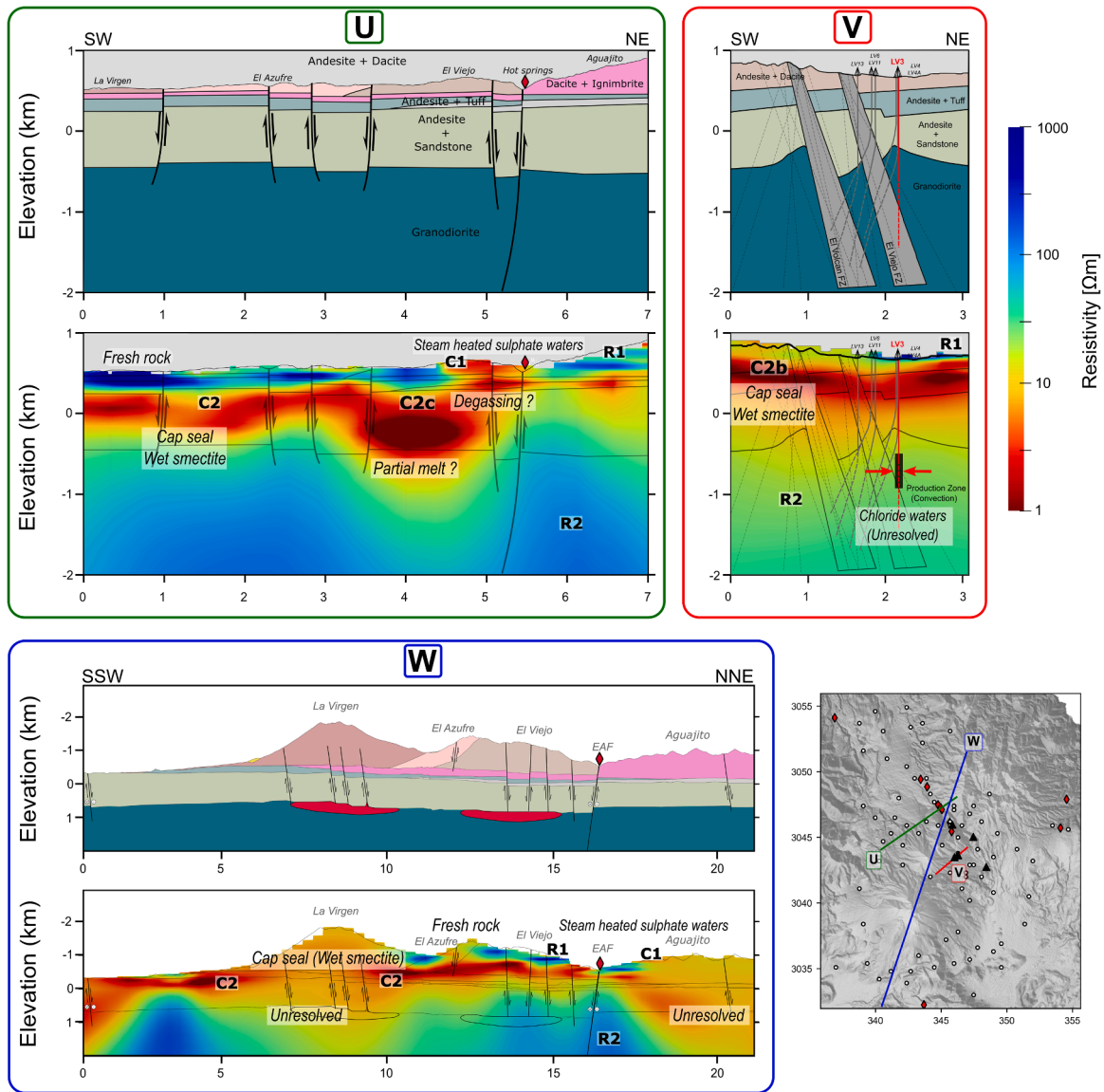


Fig. 14. Comparison of geological cross-sections with the preferred model TVV20. Section U is normal to the El Azufre Fault, and includes anomaly C2c, which could be associated with a lens of andesitic mush (30–85 % partial melt), which could be associated with sulphate steam-heated hydrothermal surface manifestations. Section V was adapted from Gómez and Rocha (2009) in Antayhua-Vera (2017), showing a detail of the geothermal field, and the anomaly C2b as a wet smectite cap seal. Section W was adapted from Avellán et al. (2018), and it illustrates that the conductive structure beneath C2 was not resolved with the dataset used in this study. Location of cross sections is illustrated on the map in the lower right corner.

including a volumetric representation of the resistivity anomalies described in Section 3.3.1. This perspective focuses on the geothermal field and highlights how hot springs, production, and reinjection wells are spatially related to the conductive structure at TVGF.

This information was used to assist in the interpretation of the conductive structure in the preferred model TVV20.

4.2.1. R1 (Volcanic sequence)

The near-surface resistor (R1) exhibits resistivity values in the range 20–1000 Ωm . This unit is associated with lavas, pyroclastic deposits and sedimentary units with varying degrees of hydrothermal alteration, and is primarily found in the edifices of the El Viejo, El Azufre and La Virgen volcanoes (sections N and W of Figs 13 and 14, respectively). While fluids may have been present in the porous units, primary permeability has largely been replaced by calcareous cementation and illite alteration. However, some fluids may still persist in areas with secondary permeability, near the damage zones of active faults (Guerrero et al., 2023).

4.2.2. C1 (El Azufre Fault Zone)

Conductor C1 is interpreted as a shallow aquifer system associated with El Azufre Fault Zone (EAFZ). Multiple hydrothermal surface manifestations with sulfate-type waters occur along this fault, particularly west of its intersection with the Cimarrón Fault (Portugal et al., 2000; Pelliccioli, 2020). The distribution of this aquifer is lithologically controlled, as evidenced by varying degrees of smectite alteration and the presence of veins of quartz, calcite, and plagioclase reported in outcrops of the Aguajito Ignimbrite (Pelliccioli, 2020). The presence of veins and smectite indicates self-sealing processes that confine the aquifer. The fluid composition suggests a steam-heated reservoir containing fossil meteoric waters mixed with magmatic H_2S , exhibiting near-neutral pH, which is indicative of intense water-rock interaction (Portugal et al., 2000).

Effective porosity estimates range from 0.4 % to 7 % (Cadenas and de la Torre, 1998; Portugal et al., 2000; Pelliccioli, 2020). According to the stochastic estimation presented in section 4.1 and Fig. 11g, this porosity range is consistent with bulk resistivity values between 3 and 20 Ωm in a

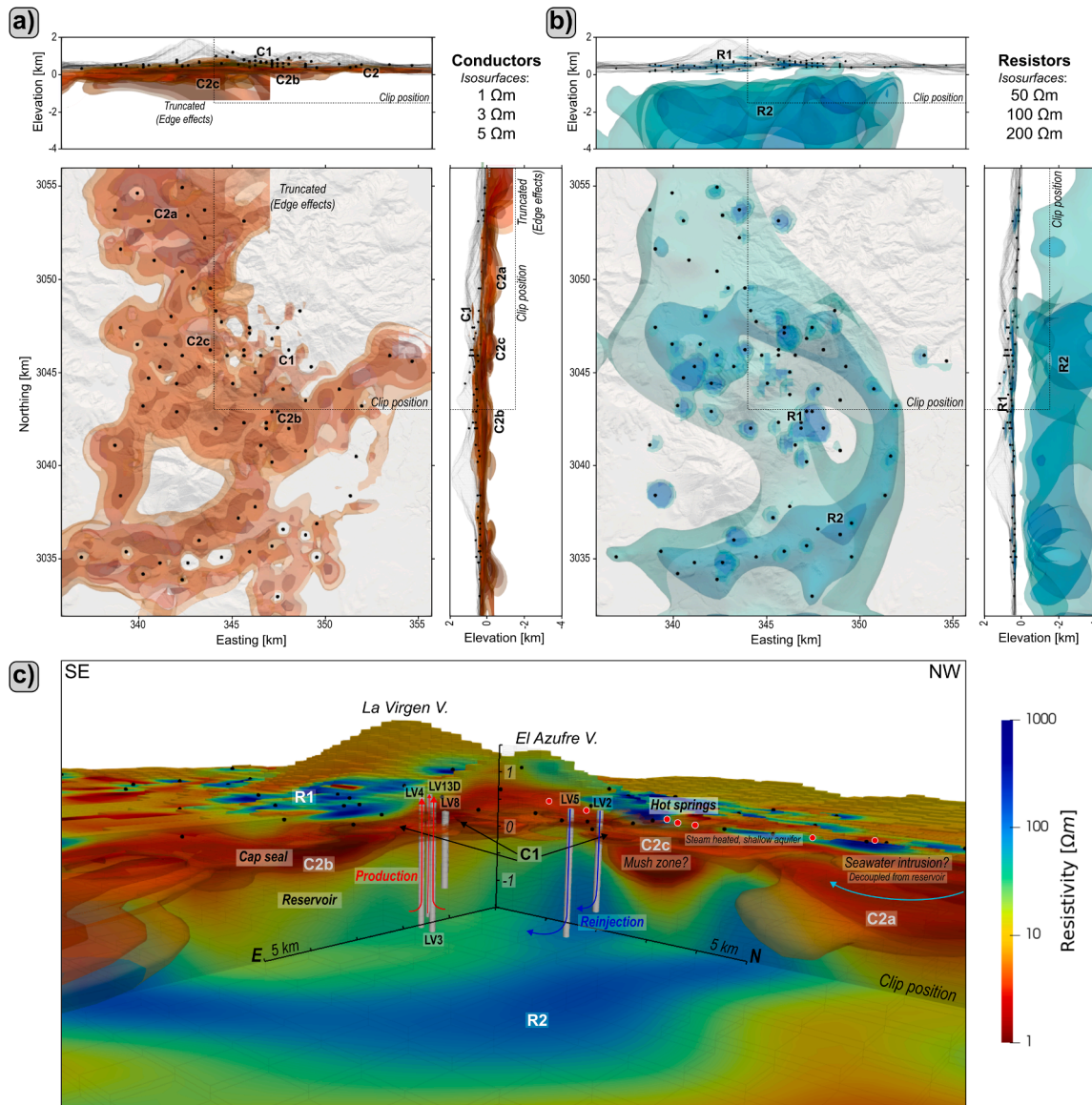


Fig. 15. Overview of the updated resistivity model for TVVC. a) Orthographic 3D representation of conductive anomalies, showcasing isosurfaces at 1, 3, and 5 Ωm . Edge effects observed in Fig. 8 have been truncated. b) Orthographic 3D representation of resistive anomalies, including isosurfaces at 50, 100, and 200 Ωm . c) Panoramic view of the resistivity model near the geothermal field; the clipped portion of the model was extracted along the dotted line in panels a) and b); the isosurfaces for the conductive structures (1, 3, 5 Ωm) are shown; the geothermal well locations are represented as gray cylinders. No vertical exaggeration. Animated view is included in the supplemental materials.

brine-saturated medium with smectite alteration (Supplemental materials S5E). This range overlaps with the resistivity range observed in C1 (0.1 – 5 Ωm).

4.2.3. C2 (Main regional conductor)

A regional conductive layer extends across the well-resolved portion of the model domain. The depth of this feature is highly dependent on the starting model resistivity (Section 3.2.3.1 and Figs. 6, 7). The preferred model ($\rho_0 = 10 \Omega\text{m}$) was selected to align the C2 conductor with a 10–60 % chlorite/smectite hydrothermal alteration zone in well LV11 (Santoyo et al., 2000). The preferred model also gives good agreement with conductive layers identified in previous studies, including Romo-Jones et al. (2000) (sections N and O, Fig. 13), Antayhua-Vera (2017) (sections P and Q, Fig. 18), and Flores (2003) (sections R and S, Fig. 13).

According to the stochastic estimation from section 4.1, a 10–60 % fraction of smectite in non-saturated rock is associated with bulk

resistivity between 15–150 Ωm . This range is higher than the inverted resistivity within C2 (i.e., 0.1 – 5 Ωm). In contrast, brine-saturated media with smectite alteration require porosity ranging from 1.8 to 92 %. For hot (150 – 200 °C), hypersaline (190 g/L HCl) brines, resistivity values tend to be closer to the lower end of this range.

Consequently, the conductor C2 is interpreted as a wet smectite cap seal with a 10–60 % clay fraction, where self-sealing processes occur. Intense self-sealing processes are evidenced by: a) calcite and illite cementation observed throughout the sedimentary units of the Comondú Group (Viggiano-Guerra et al., 2009), b) a massive layer of calcite and quartz reported at the unconformity surface between the Miocene deposits and the Peninsular Ranges Batholith (González-Partida et al., 2005), c) the presence of veins in surface outcrops (Pellicoli, 2020), and d) scaling processes affecting geothermal wells (Barragán et al., 2009, Viggiano-Guerra et al., 2009). Furthermore, the main regional conductor coincides with a region with V_p/V_s values below 1.65, interpreted as a liquid- and steam-saturated reservoir

(Vilchis-García et al., 2019).

The cap seal delineates the upper limit of the deep hydrothermal system. In section T of Fig. 13, C2 dips southward, following a distinct isotherm upflow pattern (González-Partida et al., 2005). This trend extends from wells LV3 and LV7 and protrudes northwards in the direction of well LV5, suggesting a deep convective system confined beneath C2.

The deepening of C2 is categorized in section 3.3.1 as the sub-conductor C2b. This deeper region aligns with the Cuevegel Upflow Zone (Flores, 2003), as illustrated in sections R and S of Fig. 13. Flores (2003) interpreted the regional conductor as an aquifer system and identified a discontinuity at its base, suggesting a deep upflow zone. Although the geometry of the regional conductor in TVV20 is similar, C2b is not interpreted as an upflow zone in this study. The deep hydrothermal system occurs in a depth range between 0 and 2,000 mbsl (Section T, Fig. 13), well below the regional conductor. The synthetic tests TVV03-S and TVV04-S (Supplemental material S3C and S3D) indicate that the MT dataset is not sensitive to conductive features beneath the regional conductor. A similar pattern is illustrated in cross sections V in Fig. 14, where the productive zone of well LV3 is not associated with any resistivity anomalies. The absence of a resistivity footprint for the hydrothermal system is likely due to shading effects from the overlying conductor, rather than the reservoir being associated with intermediate resistivity values (20–50 Ωm).

Anomaly C2b is located in proximity to the El Azufre Fault Zone and the El Viejo and El Azufre volcanoes. This area is characterized by high seismicity (Antayhua-Vera et al., 2015), elevated *b*-values (Antayhua-Vera et al., 2022), and intense fracturing interpreted from shear-wave splitting (Chacón-Hernández et al., 2024). These features strongly suggest enhanced secondary permeability and fluid circulation associated with the fault damage zone. Although the upflow zone interpreted from seismic tomography (Vilchis-García et al., 2019) coincides with the Cuevegel upflow zone described by Flores (2003), the observed high conductivity is attributed to the cap seal, rather than to fluid circulation processes occurring underneath.

Another anomalous region within C2 was identified in the northern part of the study area as sub-conductor C2a, which exhibits lower resistivity values. Given its proximity to the Gulf of California, this anomaly suggests a possible seawater intrusion in contact with the smectite layer. However, there is no conclusive evidence of a seawater component neither in the shallow nor the deep hydrothermal systems in TVVC (Portugal et al., 2000; Hernández-Hernández et al., 2019). Therefore, this hypothetical seawater intrusion needs to be restricted to the northern block of EAFZ to the west of Aguajito Caldera, with a permeability barrier preventing mixing with the geothermal reservoir.

Another region of interest within C2 is anomaly C2c, located north of El Azufre Volcano. Section U of Fig. 14 dissects this feature, showing its relationship to the geological structure. C2c corresponds to a spheroidal-shaped feature located between the Santa Lucia Fm. and the Peninsular Ranges Batholith, within the damage zone of the EAFZ. The top of this anomaly is deeper than the rest of C2, aligning with a graben structure parallel to El Azufre Valley. This anomaly is adjacent to multiple hot springs with sulfate-type waters and conductor C1. The low resistivity encountered at this location (1–10 Ωm) is consistent with local pockets of andesitic mush, where melt fraction ranges between 30 % and 85 %. Therefore, the favored interpretation for this anomaly is a shallow sill with remnant andesitic mush, where SO_2 and H_2S from the degassing magma ascend through the permeable damage zone of the EAFZ, and mix with the shallow aquifer system associated with C1. Interaction with calcareous levels neutralizes the otherwise acidic steam-heated waters (Portugal et al., 2000). An alternative interpretation for C2c is that this region corresponds to a local zone of increased conductance, with greater thickness but similar resistivity as the rest of C2. In this scenario, the anomaly would indicate a thicker clay cap.

4.2.5. R2 (Basement)

The deepest part of the model is dominated by the resistive unit R2

(20–300 Ωm), interpreted as granodiorite from the Cretaceous Peninsular Range Batholith (PRB). Thermochronological data indicate that the PRB underwent significant exhumation (> 10 km) following its emplacement (Kimbrough et al., 2001). Exhumation is often associated with exfoliation and thermal fracturing processes, which likely enhanced permeability at the pluton margins. Additionally, hydrothermal alteration surrounding magmatic zones enhances the porosity and permeability of the host rock, facilitating the formation of a reservoir. This reservoir contains chloride waters of magmatic origin, which mixes with paleometeoric waters confined in pre-existing reservoirs.

Laboratory measurements indicate an effective porosity of 1.1–5 % and permeability in the range 0.01 to 0.9 mD (Portugal et al., 2000). No conductive units were identified within R2, which is the result of resolution limitations of the MT dataset rather than the complete absence of deep conductive bodies. The resistor R2 hosts microseismic activity concentrated along the volcanic axis, recorded between 2003 and 2013 (Antayhua-Vera et al., 2015; Antayhua-Vera, 2017). This seismically active region contains clusters with high *b*-values, suggesting recent magmatic emplacement (Antayhua-Vera et al., 2022). Given the limitations of the current dataset, additional MT data with high-quality, low-frequency data is required to resolve the deep magmatic system.

5. Conclusions

The systematic application of 3D inversion MT techniques to a legacy MT dataset from TVVC has produced a refined resistivity model and improved on previous analyses. This case study implemented a 3D inversion using the ModEM algorithm (Kelbert et al., 2014), which included topography and bathymetry to enhance resistivity model accuracy. The resolution capacity of the MT dataset in a 3D resistivity model was investigated, including topography effects and synthetic inversion tests, offering valuable insights into the resolution limits of the dataset beneath the regional conductor. These insights, together with a stochastic estimation of bulk resistivity for different conductive phases, facilitated the interpretation of resistivity anomalies.

Through the 3D inversion of the legacy dataset, a layered structure was resolved, consisting of the following units:

- Near-surface resistor (R1) with high resistivity due to the presence of fresh, unaltered volcanic rocks and pyroclastic deposits.
- Local conductor (C1) located adjacent to the EAFZ and several sulfate-type hydrothermal surface manifestations. The low resistivity is due to a shallow aquifer system with steam-heated sulfate waters of near-neutral pH and smectite alteration.
- Regional conductor (C2) which is the dominant feature in the TVVC resistivity model. The low resistivity is due to a brine-saturated, smectite-rich clay cap overlying the geothermal reservoir. This layer exhibits internal heterogeneity characterized by:
- C2a: An anomaly with lower resistivity in the northern part of the study area. Given its geometry and proximity to the Gulf of California, the resistivity of this feature may be due to the presence of a seawater intrusion in contact with the smectite layer. These fluids are likely isolated from the TVVC hydrothermal systems.
- C2b: A deeper region of C2, located in the central part of the study area. The resistivity of this zone is interpreted as being due to a wet smectite cap seal confining a deep convective system that remains unresolved due to limitations in data sensitivity.
- C2c: A distinctive spheroidal feature located north of El Azufre Volcano. The resistivity of the region could result from a zone of aqueous fluids with higher porosity than in C2, or more plausibly, the presence of andesitic partial melt. The existing data cannot distinguish between these possibilities.
- Basal resistor (R2): Located beneath the regional conductor and extending to the resolution limits of the model. The absence of detected conductive features within this unit results from resolution limitations inherent to the dataset rather than the actual absence of

causative bodies. Small conductive features could be present, but are not imaged by the data.

Previous geophysical studies indicate the presence of a deep magmatic system beneath TVVC (Vilchis-García et al., 2019; Avellán et al., 2018; Antayhua-Vera et al., 2015; Antayhua-Vera, 2017; Guerrero et al., 2021; Antayhua-Vera et al., 2022). However, bandwidth restrictions of the dataset prevented imaging the deep magmatic system. Sensitivity tests revealed that the dataset (0.01–100 Hz bandwidth) was unable to resolve large conductive features beneath the main conductor. This limitation is attributed to strong signal attenuation at multiple horizontal conductors (C1, C2), and the reduction of depth of investigation caused by topographic effects. Coastal effects were found to be negligible, as signal attenuation within local topography and subsurface conductors prevents EM energy from reaching the seawater cells. Considering these factors, future data acquisition should focus on expanding the current bandwidth to include frequencies below 0.01 Hz, in order to improve sensitivity to deeper conductors associated with the deep magmatic system.

Nevertheless, the refined resistivity model validates previous studies based on EM methods (Romo-Jones et al., 2000; Flores, 2003; Romo-Jones et al., 2005; Antayhua-Vera, 2017). The revisited model improved the delineation of the cap seal of the geothermal system, and expanded the modelled area in a 3D space. The final resistivity model is available in NetCDF and VTK file formats in the supplemental materials, along with 3D animations illustrating the main model features.

The findings of this study have direct implications for geothermal development in the TVVC. The main conductor (C2), representing a well-defined cap seal, indicates the upper limit of the geothermal system. The presence of remnant andesitic-dacitic mush lenses is supported by the resistivity model, and is likely associated with surface hydrothermal activity along the EAFZ. Near-surface local conductors are associated with shallow aquifers, supporting the presence of two hydrogeologic systems within the geothermal field (Portugal et al., 2000).

A graphical summary of the updated resistivity model is presented as a 3D map in Fig. 15.

Declaration of Generative AI and AI-Assisted Technologies in the Writing Process

During the preparation of this work, the authors used GPT-4o in order to improve the readability and language of the manuscript, and to expedite the generation of Python scripts used in the preparation of many of the Figs and animations included in this work. After using this tool, the authors reviewed and edited the content as needed and take full responsibility for the content of the published article.

Elsevier Policy for the use of generative AI on AI-assisted technologies in writing.

Author contributions (CRediT taxonomy):

- Javier González-García: Conceptualization; Methodology; Investigation; Software; Funding acquisition; Writing – original draft.
- Martyn Unsworth: Supervision; Writing – review & editing.
- Oscar Campos-Enríquez: Investigation; Writing – review & editing.
- Yanet Antayhua-Vera: Writing – review & editing (minor revisions)

CRediT authorship contribution statement

Javier González-García: Writing – review & editing, Writing – original draft, Visualization, Methodology, Funding acquisition, Formal analysis, Data curation, Conceptualization. **Martyn Unsworth:** Writing – review & editing, Writing – original draft, Supervision, Resources, Project administration, Funding acquisition. **Oscar Campos-Enríquez:** Writing – review & editing, Writing – original draft, Validation. **Yanet Antayhua-Vera:** Writing – review & editing.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Supplementary materials

Supplementary material associated with this article can be found, in the online version, at doi:10.1016/j.geothermics.2025.103506.

Data availability

The authors do not have permission to share data.

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