Subsurface hydrothermal alteration mapping in the Reykjanes Geothermal area using a combined geoelectrical approach.

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Abstract

Most high-temperature geothermal areas have a similar resistivity signature, reflecting the alteration state of the system, as is the case for the Reykjanes high temperature system. A geothermal system has a high resistivity core, overlain by a low resistivity cap, at Reykjanes this cap reaches the surface. Hence, the study of the shallow subsurface can provide insights into the state of the system and deeper processes. Traditionally, geothermal systems are studied using electromagnetic methods, which have a large penetration depth but a low resolution, this is sufficient to characterize the system, but not to capture variations, unless they are significantly large. In this study, we explore the potential of the combined use of three geo-electric methods: electrical resistivity tomography (ERT), induced polarization (IP), and self-potential (SP), to characterize the shallow (<50 m) subsurface at Reykjanes and interpret it in a dynamic context. The observed resistivity signature reflects the typical resistivity distribution known at the site. The addition of SP allows the identification of active geothermal processes, which are highly variable and localized. The IP signal revealed a shallow (<20m) sealing structure, prohibiting fluid and gas migration, causing the absence of hydrothermal surface expressions. Such a seal can be potentially hazardous due to over-pressurization and could not be identified from resistivity imaging alone. Here we demonstrate that shallow structures can act as a proxy for deep processes, furthermore, we show that the combination of the tree methods is invaluable in studying these complex systems and recommend this for future studies.

Keywords: Alteration mapping; Shallow; Geo-electric; Reykjanes.

1 Introduction

Most volcanoes on Earth host a volcanic hydrothermal system (VHS), where complex interactions occur between the solid, liquid, and gas phases [Caudron et al. 2018a]. These VHS are often related to phreatic or gas eruptions, the processes occurring within VHS leading up to a gas eruption remain poorly known. [e.g. Caudron et al. 2018b; 2019]. Seismic methods have been used in the past for the study of VHS, here we explore the potential of a different group of geophysical methods, geo-electrics, to study VHS as they are sensitive to saturation, temperature, salinity, and porosity, which are some of the key parameters likely to change in an active VHS. More specifically, in this work, we explore the potential of geo-electric methods to delineate spatio-temporal variations of these parameters. Methods like self-potential (SP), electrical resistivity tomography (ERT), and

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induced polarization (IP) depend on the electrical properties of the subsurface, either as a passive signature (SP) or as a response to an applied electrical field (ERT/IP). They are sensitive to the presence of groundwater or groundwater flow, porosity, temperature, salinity, clay content, and mineral precipitations [Binley and Slater 2020], making them especially suited for studying hydrothermal areas and their related alteration. The advantage of geophysical methods is that they provide quasi-continuous temporal or spatial data, and they are non- to minimally invasive [Binley and Slater 2020]. The downside is that they are influenced by a large number of physical parameters which makes them difficult to use independently, especially in volcanic areas where many processes occur simultaneously. This can be overcome by combining different (geo-electric) methods [Ghorbani et al. 2018].

Our study area, the Reykjanes system, is not currently volcanically active but hosts a hydrothermal system expressed at the surface as boiling mud pools, steam vents, and fumaroles. Characterization of the geothermal system has been done using electromagnetic methods [Karlsdóttir et al. 2020], and attempts have been made to monitor the impact of the exploitation of the field [Darnet et al. 2019], but did not yield conclusive results. Monitoring using other resistivity methods such as ERT or IP has not been explored on the field. EM methods have been favored over them due to their larger depth penetration, but have insufficient resolution to detect changes in the system unless changes occur over a significantly large volume [Darnet et al. 2019]. The top of the geothermal system cuts the surface at Reykjanes, shallow observations hence provide direct insights in the state of the geothermal system with the advantage of having a high resolution.

In this contribution, we present a characterization of the shallow geothermal field where pressurization can occur through sealing [e.g. Christenson et al. 2010; Ardid et al. 2021], using SP, ERT and IP. To our knowledge, this is the first time the three are combined to characterize a hydrothermal system. Characterization of the shallow subsurface in Reykjanes enables us to identify areas of activity that are not visible from the surface and hence establish the potential of monitoring deep activity from shallow measurements. Although we use preliminary monitoring of the site to identify potential active areas, long-term monitoring will be presented in future contributions.

2 Geology and Geothermal setting

Due to its geodynamic context, Iceland is mostly made up of basalts; the neovolcanic zone (FFigure 1) marks the youngest rocks and active volcanism on Iceland, dividing older rocks east and west from the active spreading center [Sigmundsson et al. 2020]. The spreading rate at the mid Atlantic ridge (MAR) has been estimated between 18 and 20 mm/y using ocean floor reversal patterns [Demets et al. 1994] and GPS measurements [LaFemina et al. 2005]. The island is characterized by a complex plate deformation zone caused by the interaction between the plate boundary and hotspot, the MAR splits into a series of sub-areal and submarine rift- and transform segments [Einarsson 2008]. Sæmundsson et al. [2020] defines the four main tectonic structures as spreading zones, fracture zones, trans-tensional (oblique) zones, and flank zones. Most volcanism occurs in spreading zones but is also present in the oblique zones, as is the case on the Reykjanes peninsula. Iceland presently hosts three active rift segments: the Reykjanes-Langjökull Rift Zone, the Eastern- and Northern Rift Zones [Khodayar et al. 2018].

The Reykjanes Ridge (RR) segment of the MAR continues onshore on the Reykjanes Peninsula (RP). Both on- and offshore segments are arranged in a sub-parallel, staggered manner, but are distinguished by a different strike direction [Clifton and Kattenhorn 2006; Pałgan et al. 2017]. These sub-parallel fractures are characteristic of volcanically active areas [Thordarson and Höskuldsson 2008] and form the main structural feature within the RP [Khodayar et al. 2018]. The RP rift splits into narrower, parallel fissure swarms, each hosting a volcanic system. Historically, five systems are defined (from west to east): Reykjanes, Eldvörp-Svartsengi, Krýsuvík, Brennsteinsfjöll, and Hengill [Clifton and Kattenhorn 2006]. Recently, a sixth volcanic system, Fagradalsfjall, has been defined, situated between Eldvörp-Svartsengi and Krýsuvík [Sæmundsson et al. 2020]. The Reykjanes Peninsula has become volcanically active again after 800 years of rest [Sæmundsson et al. 2020] marked by periodic eruptions of Fagradalsfjall in 2021, 2022 and 2023 [Einarsson et al. 2023; Krmíček et al. 2023] and Svartsengi in 2023, 2024 and 2025 [IMO 2024; Troll et al. 2024]. Our study area is located on the Reykjanes high temperature field (Figure 1), 25 km west of Fagradalsfjall and 15 km west of Svartsengi, and has not had volcanic activity since the onset of the new episode.

Fissure swarms within the neo-volcanic zone are host to central volcanoes and associated high-temperature geothermal systems, the heat source can be a magmatic intrusion or a cooling magmatic body from an old eruptive episode [Eysteinsson et al. 1994]. Most high-temperature geothermal systems have a similar resistivity signature [Eysteinsson et al. 1994; Flóvenz et al. 2012]. Close to the surface, basalts are cold and unaltered with high resistivity (1000 Ωm), which is several orders of magnitude higher in dry conditions. A low resistivity cap (1-10 Ωm) can be found underneath, coinciding with the smectite-zeolite alteration zone. Smectite has a high cation-exchange capacity (CEC), causing the low resistivity [Lévy et al. 2018], surface conductivity is the main conductive mechanism. The cap covers a high resistivity core (30-100 Ωm), coinciding with a change in alteration minerals to chloride-epidote, these mixed-layer clays have a lower CEC [Deer 1978], here, the conductivity is controlled by surface and electrolytic conductivity. Alteration sequences correspond to specific temperature intervals, the smectite-zeolite zone occurs in the temperature range of 110-220°C, the chlorite-epidote zone at temperatures exceeding 240°C. Consequentially, temperature can be interpreted from the resistivity structure in basaltic high temperature systems. Historically, DC resistivity methods like ERT were deployed to map the geothermal structure [Arnason et al. 2000]. Currently, the joint use of transient electromagnetics (TEM) and magneto-tellurics (MT) is preferred due to its higher depth penetration [Árnason 2015] and has been successfully deployed to characterize high temperature geothermal systems in Iceland, including Krafla, Hengill [Gasperikova et al. 2011; Rosenkjaer et al. 2015], Krýsuvík [Hersir et al. 2020], Reykjanes, Eldvörp, and Svartsengi [Karlsdóttir et al. 2020].

The Reykjanes geothermal area differs from the classic resistivity signature due to its high salinity nature, where fluid chemistry dominates. Eysteinsson et al. [1994] reported no observable resistivity layering in brine systems, however, more recent studies found a similar resistivity structure albeit with lower values, in the range of 1-3 Ωm for the low resistivity cap and 10-30 Ωm for the high resistivity core [Friðleifsson et al. 2003; Karlsdóttir et al. 2020]. The cap reaches the surface along an area of approximately 1 km^2, with intense hydrothermal surface manifestations such as steam vents, mudpools, and fumaroles [Björnsson et al. 1970], furthermore, there is history of episodic hydrothermal explosions in this area [Friðleifsson et al. 2003]. A large-scale 3D resistivity model was published by Karlsdóttir et al. [2020] based on the joint inversion of MT and TEM data, they found that the cap is comprised of connected low resistivity anomalies rather than a continuous zone. Furthermore, the model reveals an area of lower resistivity within the high-resistivity core, which is suggested to be a zone of higher permeability and temperature, facilitating the circulation of hydrothermal fluids. A low resistivity zone is commonly present at depth, commonly found around 3-5 km, originating from the heat source, however, this feature can not be resolved from the resistivity distribution, potentially due to different tectonic settings [Sigurdsson 2010; Karlsdóttir et al. 2020]. Characterization surveys were done mainly for the utilization of the natural resources of the geothermal field by field operator HS Orka [Friðleifsson et al. 2020], furthermore, the field has been studied extensively as a potential site for the Icelandic Deep Drilling Project (IDDP) [Friðleifsson et al. 2003]. Resistivity methods have also been deployed to asses the effect of the exploitation of the geothermal field. Controlled-Source EM (CSEM) measurements were repeated prior to and after the installation of the IDDP2 well [Darnet et al. 2019; Bretaudeau et al. 2021], joint inversion with MT data revealed no significant difference, Darnet et al. [2019] suggest that only changes over a significant reservoir volume can result in detectable signal, furthermore, inadequate repeatability between the measurements introduces error in the timelapse analysis. Magnusdottir and Jonsson [2020] propose a new method to monitor the state of the geothermal field by utilizing the borehole casings as long electrodes in a DC resistivity survey. Measuring the resistance difference between different combinations of boreholes allows for determining fracture connectivity and productivity of the field.

The Reykjanes geothermal field is unique in Iceland because the fluid recharge of the hydrothermal system is 100% seawater, other systems in the proximity of the ocean always have at least a small fraction of freshwater recharge, like the Svartsengi system [Kadko et al. 2007; Friðleifsson et al. 2020]. Silica, calcium and potassium levels are increased, which can be caused by the dissolution of the basaltic host rock. Sulphur and magnesium are depleted, likely due to the deposition of secondary minerals [Arnórsson 1995]. Because of its fully seawater-recharged nature, the Reykjanes system is considered an on-land black smoker analog, though under low hydrostatic pressure, and has been studied extensively in this context [Marks et al. 2010; Fowler et al. 2015; Friðleifsson et al. 2020]. The host rocks are young, highly permeable basalts [Björnsson et al. 1970], hydrothermally altered due to the presence of the high-temperature field. Geological logs show a progressive hydrothermal alteration with increasing depth, described in detail by Marks et al. [2010]. Mineral assemblages suggest the alteration sequence has a composition transitional between a Mid Ocean Ridge basalt and an Ocean Island Basalt [Marks et al. 2010]. Due to its high permeability and the heavily fractured nature of the basalt, rainwater infiltrates in the shallow subsurface forming a thin freshwater lens on top of the chemically altered seawater. In the Reykjanes area, the freshwater lens has a thickness of approximately 30 meters, increasing landward, reaching 60 meters at Svartsengi [Sigurðsson 1986].

3 Geophysical Methods

3.1 Electrical Resistivity Tomography and Induced Polarization

Electrical resistivity is a measure of the ability of a medium to allow the flow of electrical current, which is inversely related to the electrical conductivity. It can provide valuable information about the hydrological and geological properties of the subsurface [Loke et al. 2013]. Resistivity is sensitive to, amongst others, porosity, salinity, saturation, temperature, and the presence of (semi) conducting minerals and clays. It has been applied on volcanoes to characterize their structure and associated hydrothermal system and to image the presence of hot conductive fluids [Rosas-Carbajal et al. 2016; Troiano et al. 2019]. While

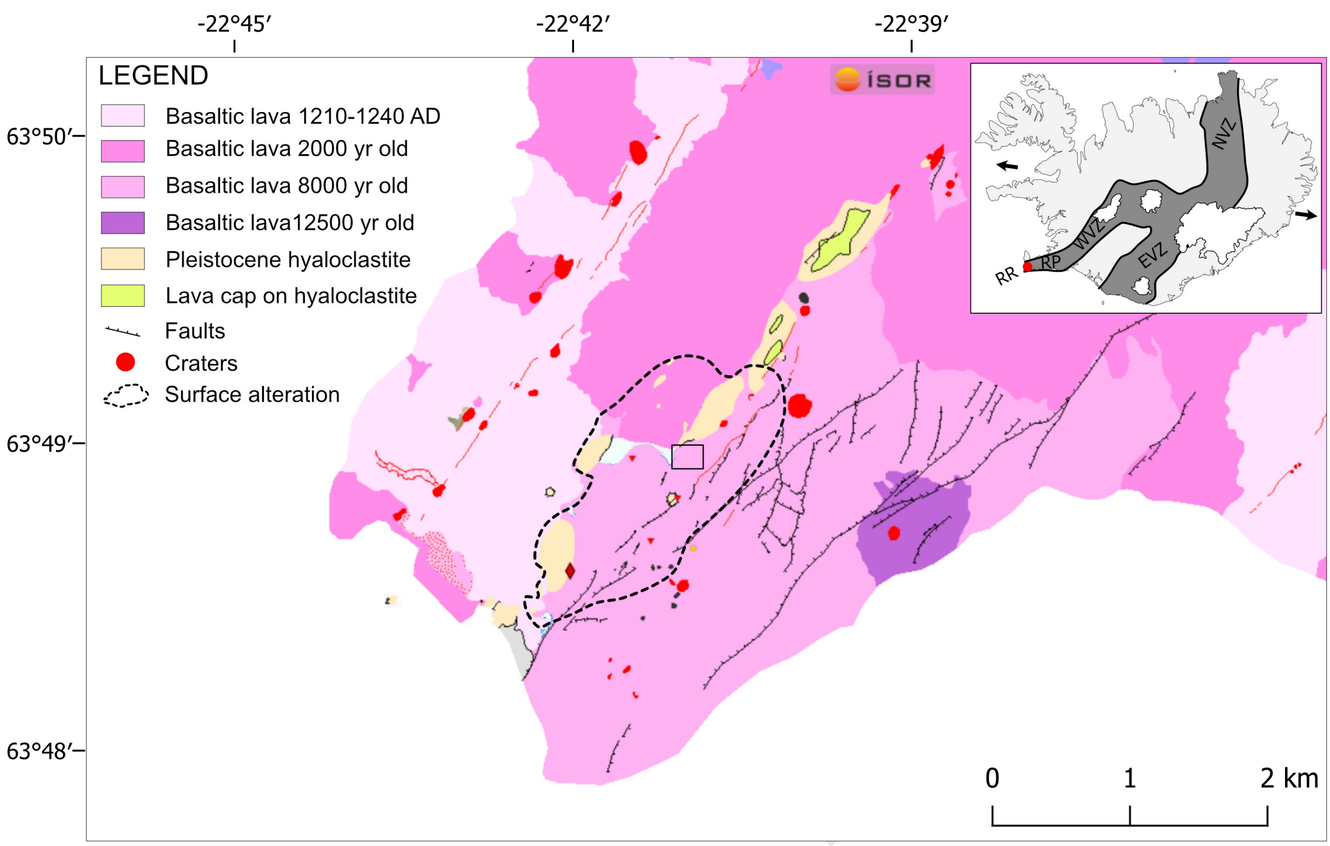

Figure 1: Geology of the Reykjanes Peninsula, adapted from Sigurdsson [2010]. The study area is marked by the black square in the center of the alteration zone. The neovolcanic zone is indicated on the map of Iceland with its main tectonic elements: Reykjanes Ridge (RR), Reykjanes Peninsula (RP), and the Western, Eastern, and Northern Volcanic zones (WVZ, EVZ, NVZ).

resistivity provides information about how well a medium can conduct current, chargeablility represents the ability to store electrical charges. Charges are retained as a result of polarization at the interface between grain and electrolyte in the so-called electrical double layer (EDL). In case of metallic minerals, the polarization of charges also takes place within the matrix [Binley and Slater 2020], hence the method is commonly known as Induced Polarization (IP). IP is sensitive to the presence of sulfides, iron-oxides [Pelton et al. 1978; Gurin et al. 2015] and clay minerals like smectite [Weller et al. 2013] which are common alteration minerals in magmatic-hydrothermal systems [Meunier 2005], making the method especially well suited for alteration mapping [Revil et al. 2002; Ghorbani et al. 2018].

An ERT measurement involves the injection of an electrical current between two electrodes (A and B) and the subsequent measurement between two different electrodes (M and N). A section is computed through the inversion of data collected with different combinations of electrodes (quadrupoles). Quadrupoles can be arranged in different formations, influencing the depth of investigation, horizontal and vertical resolutions [Dahlin and Zhou 2004]. IP can be measured in the time domain (TDIP) or frequency domain (FDIP), in this study, the former is used. TDIP measures the voltage after the current shutoff, ERT and IP measurements are done subsequently, resistivity is measured before voltage shutoff, IP windows are measured at set intervals following a delay. Such a delay is commonly used to reduce the contamination of the data due to parasitic electromagnetic fields, like electromagnetic or inductive coupling [Dahlin et al. 2002; Flores-Orozco et al. 2018].

3.2 Self Potential

The self-potential (SP) method is a passive technique that studies the electrical fields in the Earth by measuring the electrical potential, and is mainly influenced by fluid flow, oxidation-reduction processes, temperature, and concentration gradients [Revil et al. 2023]. SP is well established in the study of volcanic hydrothermal systems because of its sensitivity to fluid flow [Revil and Jardani 2013]. SP signals are generally positive in the flow direction, hence uprising fluids generate a positive anomaly [Revil et al. 2002], as a result, a typical w-shaped anomaly is often present over the summit of a volcano, constraining the hydrothermal system [Revil et al. 2008; Aizawa et al. 2009; Revil and Jardani 2013]. Other volcanological applications include constraining structural features like fracture systems [Di Maio et al. 1998] and edifice stability [Chaput et al. 2019]. As the SP method is passive, it can be limited by low signals or the discrimination of different processes underlying the recorded data.

To avoid polarization effects at the electrode surface, SP is measured at the ground surface, between two non-polarizable electrodes consisting of a metal rod submersed in a saturated solution of the corresponding salt, housed in a porous container to facilitate electrical contact [Lowrie 2007]. An SP profile consists of a base station, which is the reference point and is artificially set a 0 mV. Measurements are performed between the fixed electrode at the base station and a moving electrode. In this study, the moving electrode is measured three consecutive times at each station to assess the measurement error. When the connecting cable reaches its full range, a new base station is set to continue profiling [Barde-Cabusson et al. 2021; Revil et al. 2023]. For generating large maps, profiles are measured in closed loops, meaning a return to the base station at the end of the profile to check for the existence of a closure error. Since all measurements are relative to an artificially set base, the potential itself becomes meaningless hence, interpretation should be strictly qualitative [Revil et al. 2008].

4 Data and Processing

4.1 Field Survey

On the field, profiles were measured in a cross-hatched pattern (Figure 2). The profiles are 355 m long, consisting of 72 electrodes with a 5 m spacing, and are measured using the multiple gradient (MG) protocol, known for its good compromise between resolution and good signal-to-noise ratio [Dahlin and Zhou 2006]. Stacking was limited to two stacks to reduce the polarization at the electrodes. Measurements are done using the *Syscal Pro* from IRIS instruments, injecting for 1s using a 50% duty cycle. IP windows are measured in 20 semi-logarithmically ordered time windows, after a 40 ms delay. Stainless steel electrodes are mounted in the surface, and where necessary, clay was used to ensure good electrical contact. Processing and inversion of the ERT and IP data are discussed in Section 4.2.

Since ERT is sensitive to temperature, eight TOMST temperature and soil moisture sensors [Wild et al. 2019] were installed along profile 1, two *extreme* sensors, which can accurately measure temperatures up to 80°C, and six *long* sensors, which can be buried up to 40 cm in the soil. The temperature is measured at three locations, in the air, at the soil-air interface, and in the soil [Wild et al. 2019].

The SP survey was conducted at the same electrode locations as the ERT lines but not simultaneously, to avoid polarization effects. Measurements were done using the total-field method because it has a lower cumulative error compared to the gradient

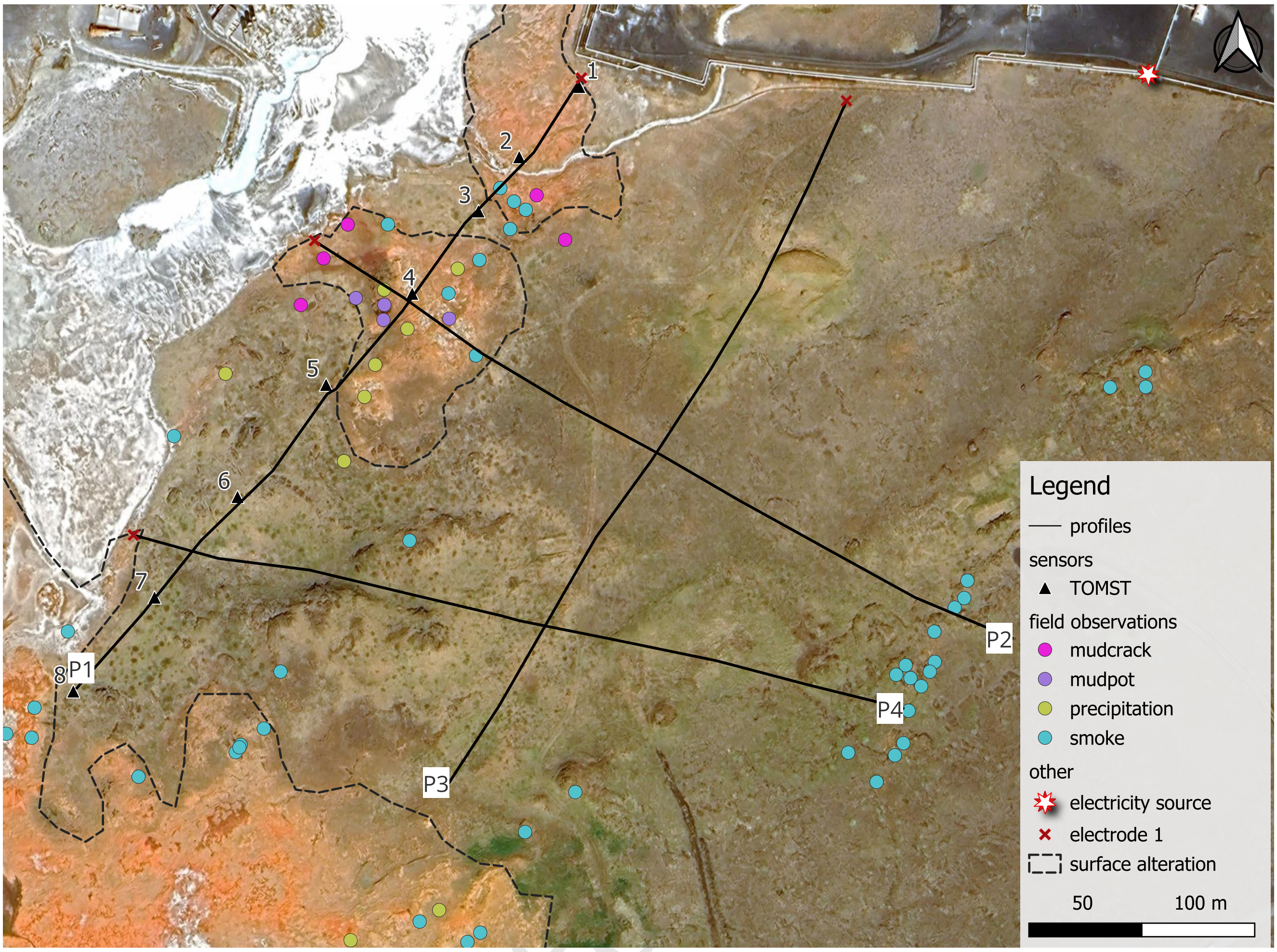

Figure 2: Localization of the ERT profiles (P1-4) and surface observations in the Reykjanes Geothermal Field. The profiles are measured from North to South and West to East, the first electrode is indicated for each profile. The surface alteration is delineated using Google imagery combined with field observations.

method and provides more flexibility during measurements, making it better suited for difficult terrains [Lowrie 2007]. To ensure good electrical contact, the electrodes were mounted in shallow holes (5-10 cm) filled with a bentonite-salt slurry [Robert et al. 2011]. On the field, the connecting wire had a length of only 30 m, requiring frequent relocation of the base station. Processing involves a correction at each base station to ensure continuity of the signal and a correction for the drift, which is estimated as a linear interpolation from the multiple base measurements [Revil et al. 2023]. Since the profiles are arranged in a cross-hatched pattern, three points of intersection were used to perform an additional in-between profile drift correction, using the profile with the most stable signal as a baseline. The closing error at the last intersection point is used to assess the level of error on the relative SP signal.

4.2 Signal Processing and Inversion

4.2.1 *Error considerations and Reciprocal measurements*

Misinterpretation of ERT images can be prevented with accurate quantification of the measurement error or noise [Labrecque et al. 1996]. The defined noise level can have a big influence on the images as an overestimation can result in over-smoothing of the resistivity whilst an underestimation can introduce artifacts in the inversion [Hermans et al. 2012]. A common way to estimate the noise level in ERT is by measuring the reciprocal profile. Relying on the principle of reciprocity, when the current and potential dipoles are interchanged, the resulting response should be identical, given a linear [Parasnis 1988], but not necessarily homogeneous [Labrecque et al. 1996] behavior of the subsurface.

The reciprocal error is defined as $e_{N/R} = R_N - R_R$ with R_N and R_R representing the normal and reciprocal resistance, respectively. Slater et al. [2000] propose an error model where the reciprocal error increases linearly with the average resistance:

$$|e_{N/R}| = a + bR \tag{1}$$

where a represents the absolute error, in Ω, and b is a relative increase with the resistance. The parameters can be determined by the envelope function as proposed by Slater et al. [2000]. This approach overestimates the error in all data points, which will result in an over-smoothed conservative image. Alternative approaches derive a less conservative error model by balancing over- and underestimation of the error, allowing the recovery of more detail in the resulting image [e.g. Koestel et al. 2008]. Here we use a linear regression model based on a least-square fit [Van Riet et al. 2022].

4.2.2 *Data Processing*

The ERT data processing is limited to removing negative values and bad datapoints based on the reciprocal data described above. The reciprocal approach can also be used for the TDIP data but is not sufficient to identify outliers and readings with poor S/N ratios. The TDIP method measures the voltage drop as a decay curve, the apparent chargeability is calculated as the area under the decay curve [Binley and Slater 2020]. Hence all curves that show erratic or non-decaying behaviour are rejected. Additionally, curves that asymptotically decay towards a non-zero value are rejected as well [Evrard et al. 2018; Thibaut et al. 2021]. This decay curve analysis is common practice in processing TDIP signals [Evrard et al. 2018] and relies on the fitting of a decreasing exponential or power function to the data [Flores-Orozco et al. 2018]. An rms-error is calculated between the measured and fitted data, all curves not meeting a predefined threshold value are rejected. Here we use an exponential function with a threshold of 60%.

4.2.3 *Inversion*

The inverse electrical problem is ill-posed, meaning that there is no unique model to explain the data. Adding constraints to the inversion algorithm can contribute to finding a physically plausible model [Tikhonov and Viak 1977]. This is done by minimizing the objective function of a regularized problem:

$$\psi(m) = \psi_d(m) + \lambda\psi_m(m) \tag{2}$$

with $\psi_d(m)$ the data misfit, $\psi_m(m)$ the model misfit and m the inverted parameter. The regularization parameter λ quantifies the compromise between the two terms. The data misfit term contains a measure of the error, which is commonly quantified by the data noise level, as discussed in the previous section. The minimization of the objective function is done by iteratively using a smoothness-constrained inversion algorithm (CRTomo) [Kemna 2000]. The inversion result can be further improved by including some a priori information [Oldenburg and Li 1994; Caterina et al. 2014] and is added here under the form of a reference model. In practice, a background resistivity value can be defined for each cell in the model, accompanied by a closeness factor, weighing the importance of each cell [Dumont et al. 2016]. Note that the choice of background resistivity and the closeness factor can introduce some subjectivity in the inversion [Hermans et al. 2014]. Here we use a homogeneous background resistivity of 1 Ω with a weighing factor of 0.01 for each cell. The low resistivity was chosen to reflect the nature of the field site where the combination of high temperature and high salinity will result in a very conductive subsurface at depth larger than 10 m. The data misfit can be estimated from the noise level meaning that the data are fitted within their noise level [Kemna 2000].

The CRTomo code solves for the distribution of the natural logarithm of the complex resistivity $m = \rho^*$, calculations expressed in terms of the amplitude of the resistivity and the phase shift ϕ. The measured TDIP data consists of chargeability measurements that need to be converted to the phase shift in the frequency domain in order to execute inversion [Kemna 2000; Thibaut et al. 2021]. After inversion, the phase shift is back-transformed to chargeability. Because the chargeability is closely related to the resistivity, the latter can have a significant impact on the measured chargeability. Hence to isolate the polarization effect, the chargeability can be normalized by the resistivity $MN = M/\rho$ [e.g Lesmes and Frye 2001; Benoit et al. 2019]. TDIP datasets are often heavily filtered and can result in a significant loss of data points, this is much less so in resistivity data. Therefore, the resistivity datasets are first inverted independently, i.e., without including the phase shift, followed by an inversion of the complex resistivity for the IP dataset.

Assessing the quality of the inverted image is of vital importance for the interpretation of the data [Caterina et al. 2013]. Here we apply the cumulative sensitivity matrix [Kemna 2000], which is representative of how well the data is resolved in the inversion model [Paepen et al. 2022].

5 Results and Discussion

5.1 Data Quality and Error Quantification

The measured resistances are distributed between 0.0001 and 3 Ω (Figure 3), the distribution of profile 1 is skewed to the left with a significant peak between 0.0001 and 0.001 Ω, profiles 2 and 4 are skewed to the right. The average resistance of profile 1, 0.13 Ω, is significantly lower than the other profiles, approximating 0.80 Ω. This is reflected in the reciprocal data filtering, the applied method removes points where the two measurements deviate 10% from each other (Table 1) and will have a larger effect on smaller values because of the presence of some systematic noise, (Figure 4). On profile 1, 73.96% of the data is retained after filtering, on the other profiles significantly more data is retained (Table 1). A global error model was determined by reconstructing a trendline on the remaining points. The analysis was done on all profiles simultaneously to deduce a single linear error model with parameters $a = -0.0002$ Ω and $b = 0.38\%$. The a factor originates from the regression and is unrealistic, given the underlying numerical error of CRTomo [Flores-Orozco et al. 2012], a value of 0.001 Ω was chosen instead.

Decay curve analysis was done to process the IP data involving curve-fitting to filter out most of the erratic curves as shown for profile 1 (Figure 5). Using a stricter threshold, for example, 0.75 instead of the used 0.60 would improve the outcome. However, profile 1 already has a low acceptance rate, where only 29.85% of the data were retained after the analysis, increasing

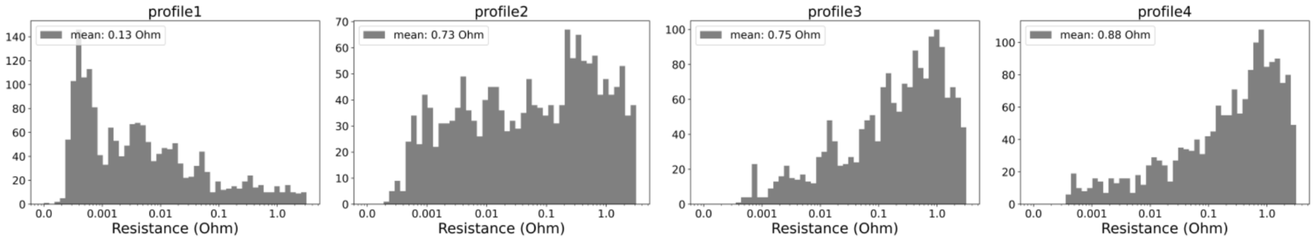

Figure 3: Resistance distribution in the four profiles, the resistance is plotted logarithmically.

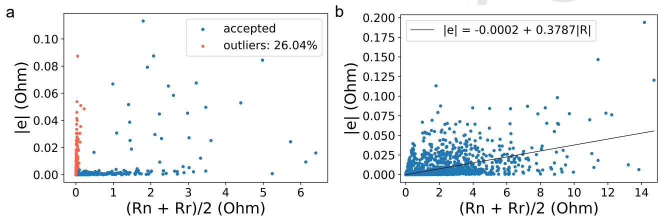

Figure 4: a: example of the data filtering for profile 1, outliers are determined based on the normal/reciprocal discrepancy, with the average resistance plotted against the residual error $|e|$. The orange dots represent the points filtered out using the method described by Slater et al. [2000]. b: global error model reconstruction (black line) fitted on the accepted points of the four profiles combined.

Table 1: Percentage of accepted points after the filtering steps for the resistance and chargeability data, i.e. the proportion of data used for inversion.

Profile	Resistance (%)	Chargeability (%)
1	73.96	28.69
2	89.78	72.10
3	96.85	84.91
4	97.23	87.15

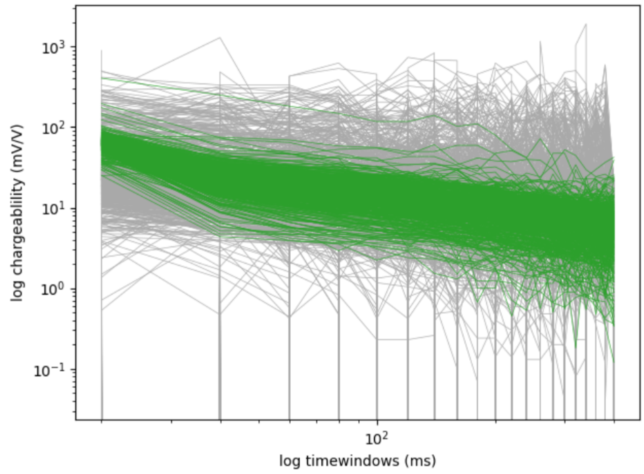

Figure 5: Example of the decay curve analysis for profile 1, a curve is accepted (green lines) when it deviates no more than 60% from a fitted exponential function. Note that the vertical lines represent negative values as they cannot be shown logarithmically.

the threshold would result in an even sparser dataset for inversion. Profiles 2, 3, and 4 have a substantially higher acceptance rate (Table 1). The areas where more points are rejected overlap with the low resistance (Figure 6), which can be attributed to the presence of high-salinity groundwater at depth [Binley and Slater 2020] which reduces the signal-to-noise ratio [Cong-Thi et al. 2024].

5.2 Inversion Results

The inversion results are reliable in the top layer of the model as can be seen in Figure 7, the sensitivity shows the data is well resolved in the first 20 to 30 meters. The inversion results are shown with the self-potential data (Figure 8). All four profiles show a zone of high resistivity (50-600 Ωm) at the surface of maximum 20 and minimum 10 m thickness. The high resistivity zone is bound at the bottom by low resistivity (1-10 Ωm). The low resistivity also reaches the surface in some profiles, on profile 1 between 0-180m and 310-400m, on profile 2 between 0-90 m and on profile 4 between 0-5m. The depth where the groundwater becomes saline according to Sigurðsson [1986] coincides with the model sensitivity. The high conductivity leads to a strong signal decrease at depth, which impacts the reliability of the inversion. The areas where the low resistivity reaches the surface coincide largely with the observation of surface alteration but not completely.

Due to the highly conductive nature of the subsurface, the majority of the deep data points (<30m) were removed from the IP dataset as their IP signal is below the noise level. The remaining dataset is sparse and is dominated by the points with a higher resistivity for which the voltages, and thus the signal-to-noise ratio, are larger. After inversion, the chargeability is larger in the top 20 m for the four profiles. The normalized chageability is also shown in order to isolate the polarizable areas. On profile 1 the normalized chargeability is increased between 0-150m and 290-330m, with values from 2-10 mS/m. Profile 2 also has an area of increased normalized chargeability, between 0-100 m with values between 2-4 mS/m, slightly lower than the anomaly on profile 1. Profiles 3 and 4 have no observable anomalies in normalized chargeability.

Given the nature of the SP method, the absolute value of the measurements should be disregarded and only trends and relative variations should be interpreted. The first part of profile 1 (0-150m) is erratic, with many fluctuations in the signal,

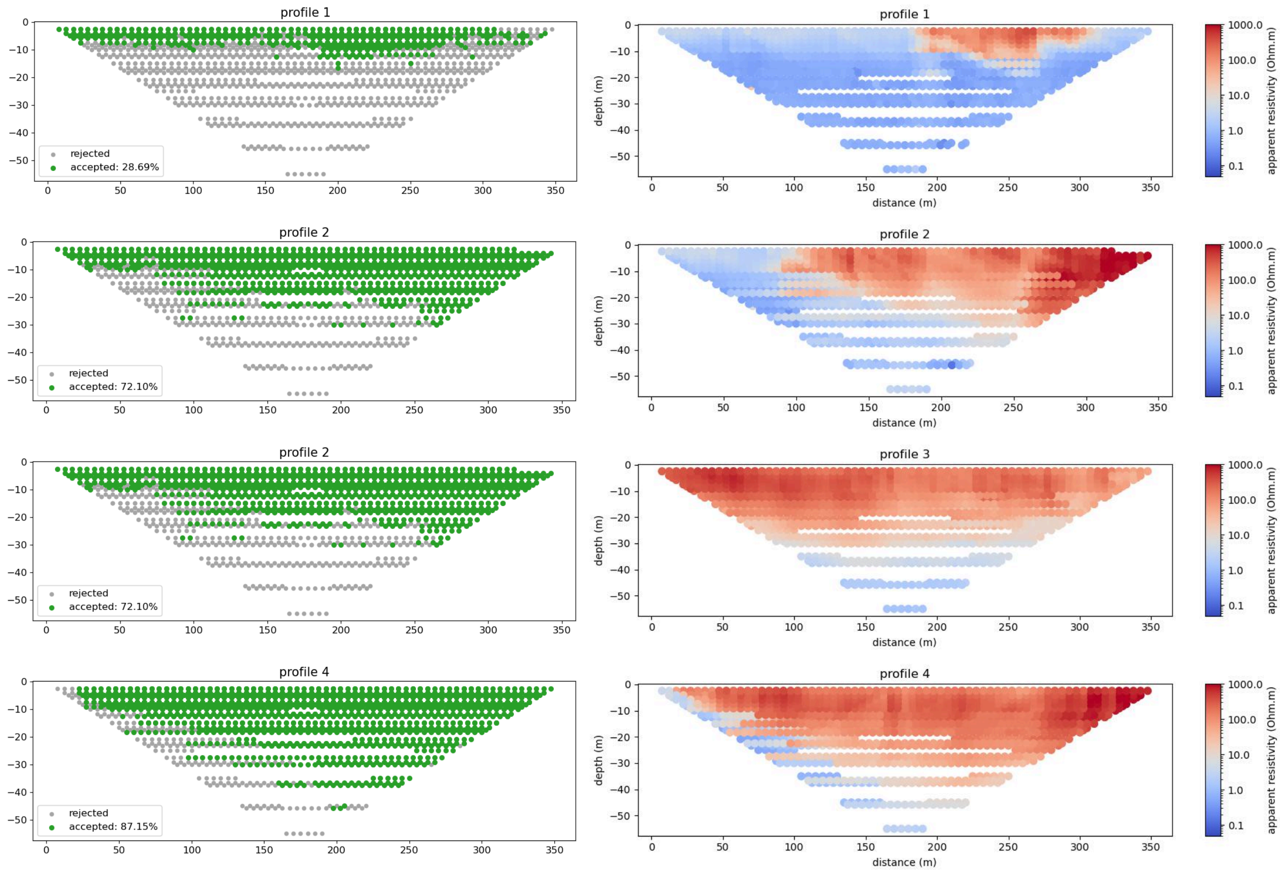

Figure 6: Acceptance rate of the decay curve analysis plotted as a pseudosection, plotted alongside the apparent resistivity, areas where points are rejected coincide with low Resistance.

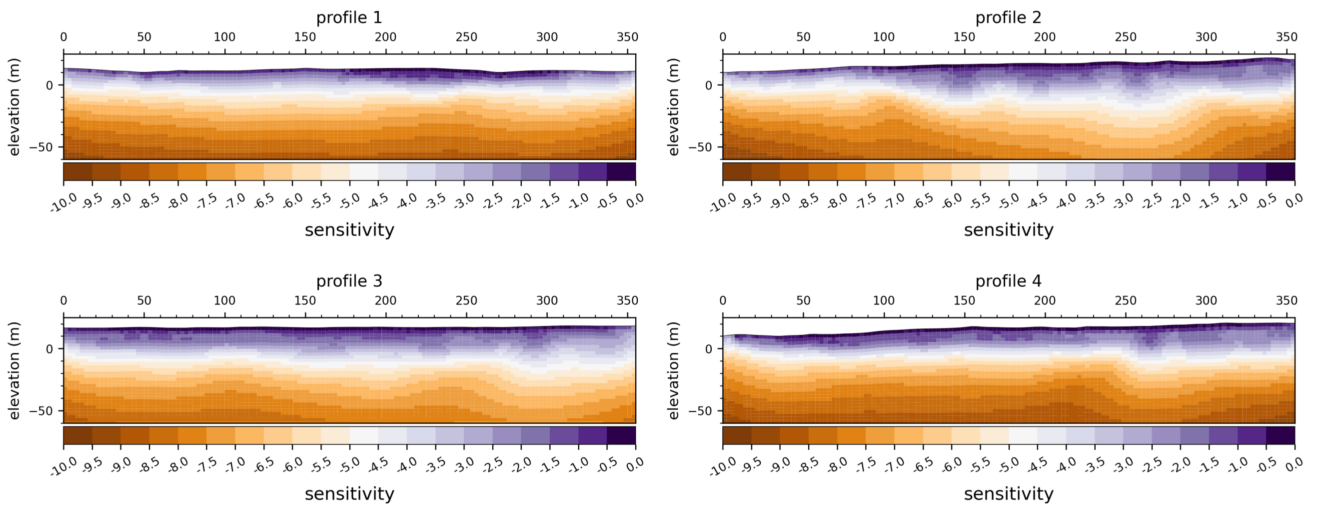

Figure 7: The sensitivity of the four profiles based on the resistivity inversion, a high value reflects that the data is well resolved in the inversion model.

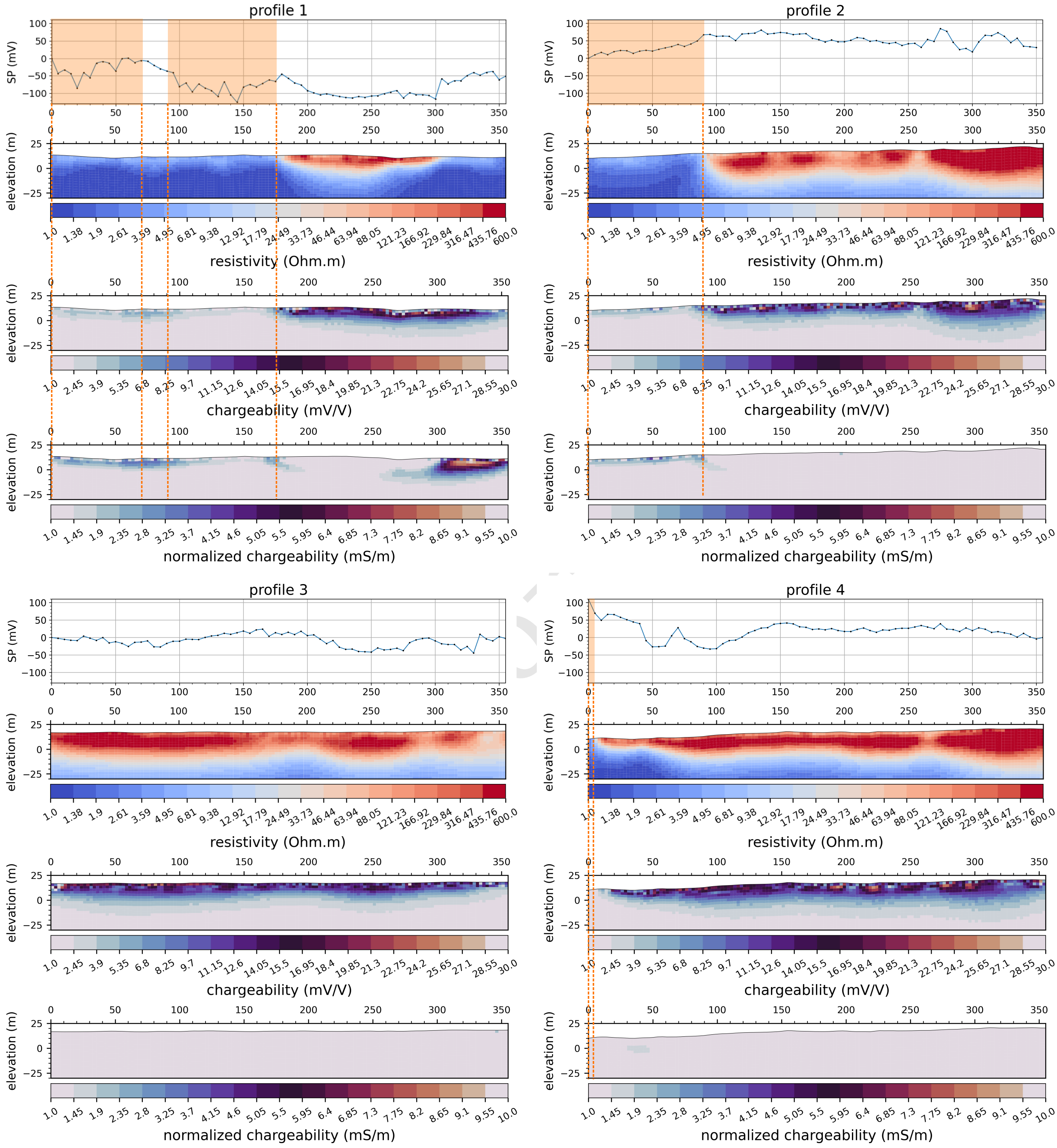

Figure 8: The results from the field campaign, for each profile, there is the corrected SP signal in mV, the resistivity in Ωm, the chargeability in mV/V and the normalized chargeability in mS/S. The orange-shaded areas and extended dashed red lines indicate where surface alteration was observed. Profiles 1 and 3 are measured north to south, and profiles 2 and 4 east to west.

Table 2: Temperature measurements on 30/10/2022 at 8:00, the sensor locations are indicated on Figure 2. The temperature gradient is calculated between the soil and surface sensors, note that the surface sensor of the *XTreme* type is just under the soil surface, for the *long* type at 12.5 cm depth, hence the term surface temperature does not apply but should be seen as shallow soil temperature.

Sensor id	Sensor Type	Soil Temperature (°C)	Gradient (°C/cm)	Atm. Temperature (°C)
1	*Long*	46.50	1.53	6.63
2	*XTreme*	55.75	5.58	8.13
3	*Long*	25.25	0.64	7.06
4	*XTreme*	64.00	6.56	8.25
5	*Long*	34.13	0.72	6.90
6	*Long*	23.62	0.42	5.90
7	*Long*	19.25	0.27	5.87
8	*Long*	17.25	0.47	6.00

there is a general trend of a plateau until 60m followed by a decrease until 150m. After that, the signal increases and follows a smooth concave trend between 180 and 280m, coinciding with the high resistivity area. The last part is characterized by an increase towards a plateau. Profile 2 has an increasing signal between 0-100m, reaching a relatively stable plateau until 260m. The last part has two big peaks, from 260-280m and from 300-340m. Profile 3 is relatively smooth with the absence of strong fluctuations, between 80 and 250 m there is a subtle concave feature. A stronger concave feature is present between 280 and 320 m. Profile 4 starts with a generally decreasing signal between 0 and 60 m, followed by a strong peak between 60 and 90m. The remainder of the profile is relatively stable with a few smaller peaks.

Table 2 shows the temperature gradients in the soil calculated from the TOMST sensors. The gradient is calculated between the soil and surface sensor, and given as an increase in °C per cm. The *XTreme* sensors are installed on the areas with the most intense surface activity, and have a very steep temperature gradient of more than 5°C/cm. The remaining sensors have a lower gradient, between 0.27 and 0.72 °C/cm with the exception of sensor 1 with an intermediate gradient of 1.53°C/cm. Table 2 also shows that where the soil temperature is high, the atmospheric temperature is elevated as well, indeed the atmospheric temperature is around 6°C for most of the sensors but around 8°C where the soil temperature is high. Hence heat radiates from the surface and can still influence temperature at 15cm above surface level.

5.3 Discussion

5.3.1 *Resistivity Zoning*

The resistivity distribution is typical of that of a seawater brine high-temperature geothermal system [Eysteinsson et al. 1994]. A high resistivity zone (100-600 Ωm) is overlying a low resistivity cap (1-5 Ωm). Karlsdóttir et al. [2020] reported that the low resistivity cap consists of smaller, intercalated lenses rather than a continuous layer. This is reflected by the shallow (< 30m) low resistivity; here, the cap reaches the surface, reflected by intense surface alteration, present on profiles 1 and 2. The high resistivity zone is described in literature as cold, unaltered basalt [Eysteinsson et al. 1994], heterogeneity present in the ERT images points to zones within this stable block that have been subject to more progressive alteration. Resistivity alone only provides information on the historically most progressive state of alteration [Karlsdóttir et al. 2020]. The addition of SP allows identification of whether these conduits are hydrothermally active or have been abandoned, as a positive SP anomaly corresponds to upflow of hydrothermal fluids [Grobbe and Barde-Cabusson 2019]. On profiles 2 and 4, at 260m, a positive SP anomaly corresponds to a lower resistivity (120 Ωm), meaning the hydrothermal fluids percolate towards the surface. Similar resistivity features in the absence of the SP anomaly indicate zones of past activity. Notably, the identified active fluid pathways do not always correspond to surface manifestations like steam vents. The low resistivity cap is also prone to heterogeneity in the SP signal, with strongly varying signals. Water and gas likely follow complex flow paths linked to open and sealed fractures, which can potentially explain the strongly varying SP signal but cannot be confirmed with the available data. The low resistivity at depth (> 30 m) is a consequence of the thermal brine and corresponds to the depth where the saltwater interface has been identified. Sigurðsson [1986] reports a sharp interface, but a gradual transition is observed in the ERT which can be explained by smoothing effects of the regularization.

The variation in temperature gradients and the presence of the high gradient in sensors 2 and 4 are not uncommon in volcanic hydrothermal systems [e.g. Chiodini et al. 2005]. Chiodini et al. [2005] measured shallow (up to 40cm) vertical soil temperature profiles in the Solfatara crater and classified the resulting curves into four categories based on the gradient and shape of the curve. These differences are mainly caused by the initial gas concentration and the flux. In the Reykjanes field, degassing is primarily controlled by local tectonics [Fridriksson et al. 2006]. The published CO_2 flux map by Fridriksson et al. [2006] shows a highly localized character of the anomalies, unfortunately, the map does not overlap our study area but we can assume a similar situation. This confirms that geothermal processes are more intense in the areas with the surface manifestation of the alteration, as both are significantly higher compared to the sensors installed in the unaltered areas.

5.3.2 *Sealing Structure*

In the traditional resistivity signature, the low resistivity cap consists of polarizable clay minerals like smectite, corresponding to an anomaly in normalized chargeability. In areas with intense surface manifestations, the normalized chargeability anomaly is weak (2-3 mS/m), while in the area with a strong signal (8-10 mS/m) these hydrothermal manifestations are absent (profile 1, 310-355 m). Due to the substantial NM difference, we suspect the precipitation of secondary (polarizable) minerals such as iron oxides or sulphides, yielding a stronger signal. These minerals form a shallow sealing structure, restricting the migration of hydrothermal fluids towards the surface and hence prohibiting further alteration. Temperature sensors 7 and 8 do not exhibit elevated temperature, confirming the absence of hydrothermal fluids directly beneath the surface. However, the SP signal shows some erratic anomalies, which can be related to the presence of the (semi-) conductive minerals, fluid flow under the seal or a combination of the two. The presence of a sealing structure has major implications on the stability of the field, overpressure can build up under a seal, resulting in hydrothermal explosions [Browne and Lawless 2001], there is a history of explosive hot-spring activity on the field [Friðleifsson et al. 2003], but it is not well documented.

5.3.3 *Potential and limitations*

The addition of the SP measurements to the ERT allowed us to determine that active hydrothermal processes are present in the shallow part. Profile 1 was measured again two, three, and four months later, Figure 9 shows the resistivity change with the first day. On 26/10/2022, strong variations are present, in the low resistivity areas a significant decrease in resistivity is present which can be related to an increase in temperature or saturation. The months after that, then intensity of the

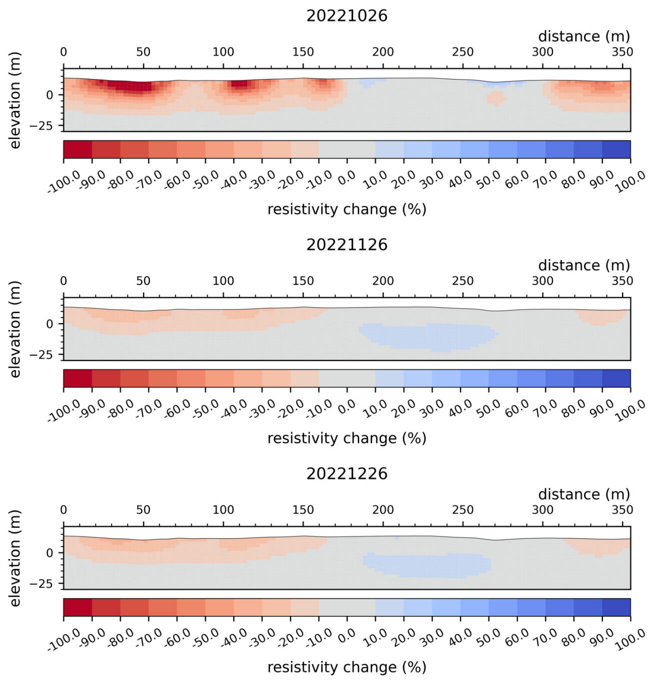

Figure 9: The resistivity change of profile 1 between the measurement shown in Figure 8 and the same profile measured again on 26/10/2022, 26/11/2022 and 26/12/2022. The changes are given in a relative change of resistivity, a positive number reflects an increase and vice versa.

signal descreased but variations remain present. Further investigations of the monitoring signal are out of the scope of this contribution. However, the presence of said variations confirms that geothermal processes are indeed active in these areas, as could be interpreted from the static survey, also where no visible surface activity is present.

The above paragraphs illustrate that by combining ERT, IP, and SP, we are able to identify areas of alteration and hydrothermal activity, even where surface manifestations are not present, and hypothesize the presence of a sealing layer. According to the literature, IP can distinguish between different alteration minerals [e.g. Lévy et al. 2018; Revil et al. 2022]. Textural information can only be quantified through analysis of the frequency-dependence of the polarization response, for instance, through spectral-induced polarization (SIP) measurements, and is thus not possible with TDIP. However, if those data were available it is not an assurance that it could be successfully applied on the field given the complex nature of volcanic areas [Lévy et al. 2019] and low registered signal-to-noise ratios. A mineralogical analysis is needed to further analyze alteration processes. The alteration is only identified until 25 to 30m depth, but this does not exclude that it is not present in deeper parts, due to the saline nature, the electrolytic conductivity dominates the ERT signal. Indeed Revil et al. [2022] stated that for salinities above 10 S/m the bulk conductivity dominates the surface conductivity hence chargeability measurements are not reliable which is coherent with our observations. The salinity at the site is that of seawater so we cannot confirm nor deny the presence of (alteration) clay below 30m for this site. Karlsdóttir et al. [2020] reported heterogeneity in the upper layer, our study sheds more light on the type of heterogeneities and closes the gap. The results from this study are summarized in a conceptual model of the shallow part of the Reykjanes geothermal field (Figure 10) where we show the highly localized nature of geothermal alteration on the field and indicate that shallow and surface observations can potentially be a proxy for deeper processes

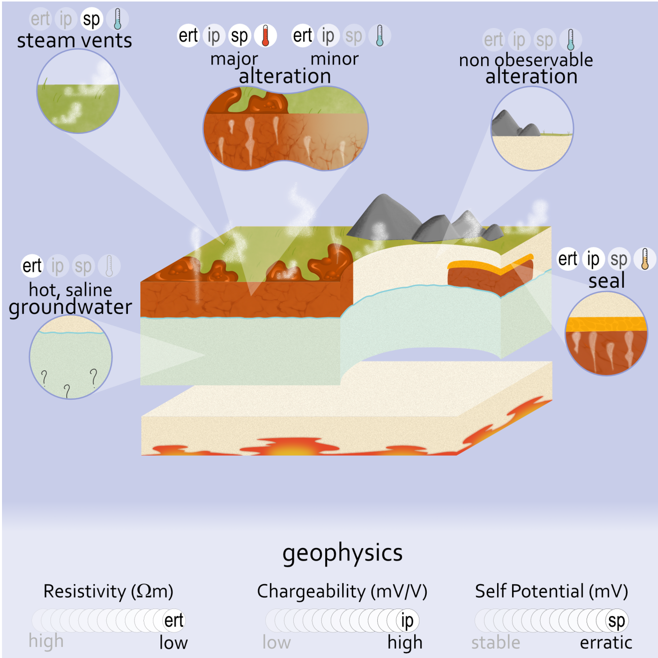

Figure 10: Conceptual model showing the main findings of this research, the different features are indicated alongside the geophysical methods used to identify said features. The brightness of the icons represents the strength of the signal of the parameter. Note that there was no temperature data available in the 'hot saline groundwater' area since our sensors are mounted in the shallow soil.

6 Conclusion

In this study, we used geo-electric methods to characterize the shallow electrical signature of the Reykjanes geothermal field and show that study of the near surface can give insight in the state of the geothermal system.

The resistivity structure is that typical of a high temperature geothermal field, and concurs with previous studies where it was found that the capping structure of the geothermal field is highly heterogeneous [Karlsdóttir et al. 2020]. Zones of intense activity are characterized by a low resistivity and a weak, but present signal in the normalized chargeability, due to the presence of alteration clays. The IP data reveals a previously unknown shallow sealing structure, which has the same resistivity signature as the alteration clays of the geothermal cap but a significantly stronger polarization response. This cap prohibits the migration of geothermal fluids towards the surface, resulting in an absence of surface manifestations. Such a seal can be potentially hazardous due to overpressurization resulting in explosions. The resistivity and chargeability provide information about the geology and maximum progression of geothermal alteration, but on their own carry no information regarding the state of the system. Therefor, we can not conclude whether hydrothermal processes are still ongoing. The addition of the spontaneous potential overcomes this, as it is strongly related to fluid flow mechanisms. Positive SP anomalies are present throughout the area, showing that there is hydrothermal activity in the shallow subsurface but also that it is not limited to the areas with intense alteration. This means that the field is currently still undergoing progressive alteration.

Previous attempts to use resistivity methods to monitor the field were unsuccessful [Darnet et al. 2019] and mainly related to the resolution as they target a large depth (>3km), compromising on the resolution. Our study has a very high resolution at shallow depth, enabling us to capture changes in the system on a short timescale. Given the nature of the site, with the cap of the geothermal system reaching the surface, changes in the shallow cap can be directly related to deeper processes.

At Reykjanes, we show that the combination of these three geophysical methods is essential to provide valuable insight into the shallow structures of a geothermal system, which can act as a proxy for deeper processes, with the advantage that they are cost-effective and relatively easy to deploy. We recommend the joint interpretation for complex environments such as volcanoes and geothermal systems.

Author Contributions

LV: survey design, data processing and writing the draft; EV: ERT and SP data processing; WD: IP data processing; AFO: IP data processing and revision of the article; OF: mapping of the site; KJ and BB: coordination and survey design; CC and TH: survey design, revision and project funding. The field campaign was conducted by LV, EV, WD, OF, TH and CC

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Data Availability

All new data from the ERupT project have been deposited in the Zenodo repository and are freely accessible at the following link: 10.5281/zenodo.14251329.

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