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article in press

Geological nature of the North Schofield Dam, a major hydrogeological barrier in the central O‘ahu groundwater flow system (Hawai‘i, USA)

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30 Abstract

The Schofield groundwater body is one of the most important freshwater reservoirs on the island of O‘ahu and has enigmatically high hydraulic head as compared to the adjacent basal aquifers. To better understand the geologic and fluid-flow conditions across this aquifer's northern boundary, the so-called North Schofield Dam, we collected self-potential and seismic ambient noise data, revealing two distinct
35 geophysical domains separated by a sharp N25° boundary. We interpret that the northwestern domain contains low-permeability buried valley-fill material, while the southeastern zone is dominated by more permeable lava flows from the Wai‘anae and Ko‘olau volcanoes. This explains the abnormally high groundwater levels observed in the Schofield groundwater unit. The buried paleo-valley geometry also provides new constraints on volcanic stratigraphy, indicating a minimum of ~100 m of Ko‘olau lavas
40 overlying the eroded Wai‘anae edifice in this area. Our study elucidates the orientation, underground

geometry, and geological nature of the North Schofield Dam, highlighting the critical role of geological complexity in volcanic environments in influencing groundwater storage and flow.

1. Introduction

Water resources research has become increasingly important due to more frequent extreme and unpredictable climatic conditions. In such conditions, population growth and land transformation exert increasing pressure on water resources, making water research a priority for environmental management. This situation is particularly critical on oceanic islands, where access to fresh water is limited, and accessing external resources is both expensive and logistically difficult.

The indigenous people of Hawai'i learned where to find freshwater resources in a challenging volcanic environment. They learned where the sources are and which paths they go through, from the mountain to the ocean, learning how to use this resource sustainably. This knowledge forms the foundation of the 'Ike Wai project ['Ike Wai Project 2024] of which the present study is an integral component. Recent scientific studies have employed modern geophysical methods to validate what indigenous knowledge had previously discerned. For instance, using high-resolution marine CSEM (Controlled Source Electro-Magnetics) imaging, Attias et al. [2020] successfully imaged transport of freshwater from coastal to offshore areas through the layered structure of Hualālai Volcano, in the western part of the Big Island of Hawai'i. Alternating permeable and impervious layers, typical of volcanic environments, are crucial in guiding freshwater to offshore underwater reservoirs.

One consequence of freshwater discharge at the coast of Hawai'i and the strong permeability contrasts of volcanic deposits is the formation of anchialine pools (or anchialine ponds) along the Hawaiian coastlines [Holthuis 1973; Maciolek & Brock 1974]. These pools provide a habitat for endemic species like the 'ōpae 'ula (*Halocaridina rubra*; Wong 1975; Brock 1985; Brock & Kam 1997). Any changes in the dynamics of the hydrogeological system of the islands can have a disruptive effect on these ecosystems. For instance, variations in the annual rainfall volume, rising ocean levels, or human activities like pumping, pollution, construction, and land use conversion can all disrupt groundwater-dependent ecosystems.

Here we present a geophysical study conducted in the central part of the island of O'ahu (Figure 1), focusing on one of the hydrogeological barriers that significantly affects the groundwater flow system: the so-called North Schofield Dam (shown in Figure 2). The objective of our study is to determine the geological nature, geometry, and hydrogeological role of the North Schofield Dam by integrating self-potential mapping and

ambient noise surface wave tomography. We directed our efforts toward this structure rather than the South Schofield Dam due to its accessibility and relative lack of anthropogenic electromagnetic signals. We used Seismic Ambient Noise Surface Wave Tomography (ANSWT) combined with the Self Potential (SP) method for a joint interpretation. While the latter method has proven effective in numerous hydrogeological studies [e.g., Aubert and Atangana, 1996; Jouniaux et al., 2009; Revil et al., 2023], application of ANSWT and the joint use of both techniques is fairly novel [e.g., Lecocq et al. 2017; Clements & Denolle 2018; Garambois et al. 2019] and had been applied only once in Hawai'i [Grobbe et al. 2021]. With 40560 m of linear profiles at a 20 m measurement spacing covering a 12 km² study area for SP, and ANSWT coverage across 7.15 km², our study is the first to present such an extensive and detailed dataset of this type in Hawai'i. We used a cluster analysis on the SP/elevation data to evaluate whether it exhibits a natural statistical separation consistent with the contrast observed in the SP map. The SP method primarily enables 2D mapping, providing insights into groundwater circulation contrasts in the unsaturated zone and offering indirect information about structural features that influence groundwater flow. The ANSWT method provides structural and geological information through three-dimensional imagery based on seismic shear wave velocity analysis. We analyzed the record of the vadose zone thickness variations from nearby water wells [U.S. Geological Survey 2024] to complement our study.

2. Geological context

Located in the Central Pacific, O'ahu is the third largest and most populated Hawaiian island and home of the State capital, Honolulu. The subaerial part of the island results from volcanic activity, caldera collapse, large-scale landsliding/flank collapse, and erosion of two shield volcanoes: Wai'anae and Ko'olau. Ka'ena, a third shield volcano predating the others, was initially proposed by Moore et al. [1989] and later confirmed by Sinton et al. [2014]. Geological investigation and K/Ar dating indicate that Ka'ena was emergent at least by 3.5 Ma, reaching a maximum height of ~4000 m above the ocean floor and ~1000 m above sea level. The subaerial activity of Wai'anae Volcano lasted from about 3.93 ± 0.08 to about 2.9 Ma [Presley et al. 1997; Guillou et al. 2000], while the subaerial activity of Ko'olau Volcano lasted from about 3.2 to 2.1 Ma [McDougall 1964; Ozawa et al. 2005]. All three volcanoes have undergone massive collapses into the ocean toward the North (Ka'ena), the West (Wai'anae), and the East (Ko'olau) of the island and, today, only the Wai'anae and Ko'olau formations constitute the subaerial part of the island (Moore et al., 1989; Presley et al., 1997).

Wai'anae and Ko'olau are characterized by three rift zones radiating from a central caldera [e.g., Stearns & Vaksvik, 1935; Takasaki & Mink, 1985]. Both volcanoes have two main rift zones and a third more diffuse one [Oki 1998]. Rift zones are defined by dense swarms of dikes that intruded vertically or nearly vertically, constituting at least 10% of the total rock volume [Takasaki and Mink 1985]. At Ko'olau, the main NW

and SE rift zones extend along the island's eastern coast, whereas the third SW rift zone, the Ka'au Rift Zone, is defined by a broad zone of dispersed dikes East of Honolulu (Figure 2). Wai'anae's NW and S rift zones are well-defined, but its proposed NE rift zone is more diffuse and its extent remains uncertain due to burial beneath younger volcanic and non-volcanic deposits [e.g. Sinton and Rowland 2022]. Lava flows from Ko'olau Volcano accumulated against the eastern flank of Wai'anae, and shaped the central saddle area between the two edifices. In the saddle area, intense tropical weathering and local topographic conditions have produced thick sequences of laterite or saprolite [Miller 1987]. Our study area, in the Schofield Plateau, is underlain by the Wai'anae and Ko'olau volcanic sequences, separated by a regional unconformity formed by erosion along the Wai'anae- Ko'olau contact [Nichols et al., 1996]. Regional geological mapping shows that younger Ko'olau shield lavas onlap and partly bury the eroded eastern flank of Wai'anae Volcano across the Schofield Plateau [e.g. Sherrod et al., 2007, 2021]. This superposition is well established at the island scale, but existing mapping and well data in central O'ahu do not provide direct constraints on the depth of the Ko'olau-Wai'anae overlap in our study area. In particular, the nearby Heleman Deep Monitor Well [Board of Water Supply & Commission on Water Resource Management, 2002] penetrates more than 300 m of basalt, clinker, and saprolite without recording any lithological change that would distinguish Ko'olau flows from older Wai'anae units. Other wells in the Schofield region either lack detailed lithological descriptions or do not reach sufficient depths to encounter the volcanic contact. Stearns & Vaksvik (1935) mapped an erosional unconformity between the Wai'anae and Ko'olau volcanic formations in the Schofield area but no depth or thickness constraints were identified. As a result, while the geological superposition is firmly established regionally, its precise subsurface geometry beneath the Schofield Plateau remains poorly constrained. Our study area lies within this altered and structurally complex transitional zone.

3. O'ahu's Schofield hydrogeological unit

In this section, we summarize the hydrogeological framework of central O'ahu, emphasizing the characteristics and uncertainties of the Schofield high-level groundwater body, one of O'ahu's most important natural groundwater storage areas.

O'ahu's hydrogeological system is fundamentally controlled by its geological structure as the eroded remnant of the Wai'anae and Ko'olau Volcanoes. Its main aquifers thus consist largely of highly permeable stacks of thin lava flows. The anisotropic nature of these lava flows, with horizontal hydraulic conductivity significantly greater than vertical, explains preferential horizontal movement parallel to bedding planes [Paillet & Hess, 1995]. Other geological structures compartmentalize the groundwater system by acting as barriers to groundwater flow. Hunt [1996] shows that gently dipping lava flows and sedimentary rocks, intersected at high angles by eruptive feeder dikes in rift-zones and weathered valley-fill sediments,

combine to host the regional aquifer system. In Hawai‘i, groundwater impounded by low-permeability dikes within otherwise permeable lava-flow sequences is commonly referred to as dike-impounded groundwater.

These structures disrupt lateral groundwater flow, producing abrupt discontinuities in hydraulic head delineating distinct groundwater areas and defining the main aquifers (Figure 2 modified from Oki 1998).

The central O‘ahu groundwater flow system is bounded by the Ko‘olau Range to the east, the Ka‘au rift zone to the southeast, the sedimentary deposits and rocks forming the coastal caprock to the south and north, and the crest of the Wai‘anae Range to the west (see Figure 8 in Oki 1998). The system is further subdivided by additional hydrogeological barriers, including valley fills, which have been shown to influence hydrogeological processes across the island [e.g. Grobbe et al. 2021; Izuka et al. 2021].

The Schofield area is located between the two major dike-impounded systems on O‘ahu, the Ko‘olau and Wai‘anae water bodies. These systems host the island’s highest groundwater heads and therefore exert a first-order control on regional groundwater gradients (Dale & Takasaki, 1976). The Ko‘olau dike-impounded water body, which underlies some of O‘ahu’s highest rainfall zones, receives much more recharge than the Wai‘anae system. A persistent head difference of approximately 185 m exists between the Ko‘olau dike-impounded water body and the Schofield high-level water body. Long-term tunnel discharge data indicate that this head difference is remarkably stable, suggesting that the Ko‘olau high-level system provides a steady, sustained inflow into the Schofield groundwater body [Dale & Takasaki, 1976].

The Schofield groundwater body is considered one of the main natural freshwater reservoirs in central O‘ahu. It maintains an anomalously high hydraulic head, sharply contrasting with the adjacent basal aquifers (H. Stearns et al., 1940; Dale & Takasaki, 1976) despite the lack of exposed surface structures to explain such strong groundwater impoundment. While its eastern and western hydrogeological boundaries coincide with the Ko‘olau and Wai‘anae volcanic rift zones, the northern and southern boundaries remain debated. Even their geographic positions are uncertain, as reflected by variations among published studies [Oki 1998]. Initially referred to as groundwater dams by Dale and Takasaki [1976] these features have since been described as Schofield groundwater barriers [e.g. Hunt 1996].

Between these barriers, the groundwater level is nearly constant at about 85 m a.s.l., but declines steeply across the northern and southern boundaries towards the <10 m heads of the adjacent Waialua and Pearl Harbor basal aquifers [Dale and Takasaki, 1976]. Although high-level groundwater bodies occur elsewhere on volcanic islands, the Schofield system is unusual because it does not match any of the classical hydrogeological settings typically observed in Hawaiian and ocean-island aquifers.

First, it cannot be interpreted as a basal aquifer. The term “basal groundwater” has traditionally been used in Hawai‘i to refer to freshwater lenses floating on denser underlying saltwater, typically characterized by relatively low water-table elevations above sea level [Hunt, 1996; Izuka et al., 2016]. The hydraulic head

of the Schofield groundwater body (~85 m a.s.l.) is far too high for a Ghyben–Herzberg freshwater lens, and the system shows no hydraulic connection to the ocean [e.g. Gingerich & Oki, 2000].

Second, it does not correspond to a perched aquifer, which by definition is separated from underlying groundwater by unsaturated rock (Izuka et al., 2016). Indeed, wells drilled beneath Schofield to depths below sea level remained saturated, confirming that the Schofield groundwater body is not perched (H. Stearns et al., 1940; Oki, 1998; Izuka et al., 2016)

Finally, the Schofield groundwater body does not show the compartmentalized hydraulic behavior characteristic of typical dike-impounded groundwater. Groundwater occurs mainly in permeable Ko‘olau lavas, and the water level is relatively flat and continuous across much of the Schofield Plateau, with no dike system clearly identified that could account for the observed impoundment (Hunt, 1996). The result is an aquifer whose hydraulic behavior and anomalously elevated but relatively uniform head, do not conform to any of the expected configurations. Thus Schofield continues to be described as a groundwater enigma in recent studies (Izuka et al., 2016).

Dale and Takasaki [1976] used a hydrologic budget to show that the Schofield water body receives an average recharge of 481 thousand m³/day. At that time, approximately 28 thousand m³/day were extracted through pumping, meaning that about 453 thousand m³/day were transferred to neighboring basal-water bodies. Of this surplus, more than 80% flowed southward toward the Pearl Harbor basal-water body, while less than 20% flowed northward toward the Waialua basal-water body. O‘ahu’s primary groundwater supply thus depends strongly on the Schofield hydrogeological unit.

Together, the Ko‘olau and Wai‘anae high-level systems, the Schofield reservoir, and these groundwater dams form an interconnected volcanic aquifer network in which high-head dike-impounded systems feed the Schofield plateau aquifer, which in turn spills water, either through, below, or over the dams into the basal systems that supply most of O‘ahu’s potable water. Understanding how volcanic structures and associated permeability contrasts control groundwater transfer and storage in such complex aquifer systems is important for understanding groundwater resources on O‘ahu and can provide transferable knowledge applicable to other volcanic islands.

4. Methods: data acquisition and processing

Hawaiian volcanic aquifers present significant imaging challenges for many standard geophysical techniques due to high seismic velocities, strong attenuation, fracturing, cultural noise, and high electrical contact resistance. In this context, SP and ANSWT represent two robust and complementary methods. SP is sensitive to preferential groundwater flow and near-surface permeability contrasts [e.g. Revil et al., 2023b], whereas ANSWT provides 3D shear-wave velocity information that reflects lithology and degree

of alteration at depth [Brenquier et al., 2007]. Their successful application in a comparable volcanic watershed on the Kaiwi Coast [Grobbe et al., 2021] demonstrated that combining SP and ANSWT helps overcome the limitations of individual methods and offers a more complete view of subsurface hydrogeological structure. For these reasons, we selected this integrated approach for the North Schofield Dam investigation.

4.1. Self-Potential (SP)

From October 28 to November 21, 2019, we acquired a dataset of 2029 SP measurements on agricultural land of North O‘ahu, on the Ko‘olau Volcano units, with a 20 m spacing between measurements along the profiles (Figure 1B). The SP profiles were distributed as evenly as possible, avoiding a major ditch, adjusting the repartition as required by the continued agricultural field operations, and connecting all the measurements to obtain a completely interconnected network. We chose a 20 m-spacing between consecutive measurements to detect short-wavelength variations related to small-scale features such as faults and abrupt lithological changes, while ensuring a reasonable acquisition time.

We acquired and processed the SP data following the procedures described in detail in Barde-Cabusson et al. [2021]. We measured the SP values between two Cu-CuSO₄ non-polarizable electrodes made of a polyvinyl chloride (PVC) tube with a porous ceramic tip. The electrodes were connected to a commercial high-impedance voltmeter ($>10\text{M}\Omega$) through a 300 m-long wire. Applying the fixed reference method, one reference electrode was positioned at the beginning of a profile while a roving electrode was moved systematically every 20 m along predefined profiles. Upon reaching the end of the 300-meter wire, the roving electrode became the new reference electrode. Simultaneously, the prior reference electrode was removed from its original position and repurposed as the succeeding roving electrode for the subsequent 300 meters. This cycle continued iteratively until completion of the acquisition campaign.

We first measured the resistance between the reference and roving electrodes at each station. This step ensures the quality of the electric contact before proceeding to the actual SP measurement (in mV). Typically, the measured resistance was $< 5\text{ k}\Omega$, indicating excellent electrical contact. We recorded each SP measurement after it had stabilized, usually within a few seconds. If the value remained unstable at the initial location, we repeated the measurement within a 10 cm radius at a different spot. If the measurement still failed to stabilize after a few attempts, we calculated the mean value of five measurements taken within a 50 cm radius. Measurements were excluded when their instability exceeded the range of variation that could reasonably be expected from the local spatial trend observed at neighboring stations. The data processing involved applying a reference correction and a closure (or loop, or drift) correction [e.g., Barde-Cabusson et al. 2021]. We made these corrections to establish a unique reference for the entire dataset and

account for temporal variations accumulated during data acquisition, including those associated with changes in environmental conditions (e.g., rain, temperature variations during the day, etc.).

The results presented herein are relative to an arbitrary reference point selected at a marked location in Figure 1B (denoted by the yellow star). Consequently, our discussion will revolve not around the absolute values of the SP but rather on the observed spatial variations.

4.2. Cluster analysis

As a complementary approach, we performed a basic clustering analysis of our SP dataset. To our knowledge, this is the first time that this has been tested on SP data. We used version 3.35 of Orange, an open-source data visualization and analysis toolbox in Python [Demšar et al. 2013]. This tool provides a visual programming interface and supports various clustering algorithms. We used k-means clustering, which is justified based on several factors. First, the structure of our data shows distinct patterns suitable for partitioning into groups, which aligns well with the strengths of k-means clustering. It also allows validation of the number of clusters with silhouette scores, determining the optimal clustering solution. This partitioning method divides a dataset into k distinct, non-overlapping clusters.

We defined a non-fixed number of clusters between 2 and 8 and checked the Silhouette Scores table displayed by the toolbox, which helped to determine the best number of clusters in our dataset without forcing it. A high silhouette score (> 0.5) suggests that the clusters are well-defined and distinct from each other. Additionally, the Interactive k-means option allows visualizing each cluster and its centroids and refining the result progressively. No manual re-assignment of data points was performed; the refinement refers only to the visualization of the iterative centroid updates provided by Orange.

A centroid serves as a representative point or center of a cluster identified by the algorithm. In the case of two clusters, the first centroid is initially placed randomly and assigned to a group of data points, while the second centroid is positioned among the remaining data points. Next, the centroids are iteratively adjusted based on the best distribution of data points assigned to each cluster. The position of a centroid corresponds to the average of all data points within its cluster along each attribute in the dataset. In our case, the attributes are the SP and the elevation values. Elevation was included because, in volcanic hydrogeological settings, SP commonly varies with elevation as a result of groundwater circulation. Spatial changes in this SP/elevation relationship can therefore indicate changes in water-table geometry or subsurface hydrogeological properties (Aubert & Atangana, 1996; Lénat, 2007). Other studies have explored the application of k-means clustering analysis to distinguish lithological units, for instance within the Quaternary aquifer system around Debrecen in Hungary [Mohammed et al. 2024] or in the Alps, to detect serpentinites embedded in the sedimentary cover [Revil et al., 2024]. Here, k-means clustering is used as an exploratory tool to test whether the SP dataset exhibits a natural statistical partition consistent with the

spatial contrast observed in the SP data. We note that k-means minimizes within-cluster variance using Euclidean distance, assumes convex cluster geometry, and can vary slightly depending on the initial centroid placement. In our case, SP (mV) and elevation (m) have comparable numerical scales, so no additional normalization was required. Given that the dataset consists of 2-D Euclidean points, k-means provides a simple and transparent method to evaluate whether two dominant SP domains emerge from the data without imposing additional parameters. The resulting clusters are examined in attribute space (SP vs elevation). This analysis is therefore exploratory and confirmatory rather than interpretative of the lithology.

4.3. Ambient Noise Surface Wave Tomography (ANSWT)

ANSWT relies on the fact that different frequencies of a horizontally propagating surface wave penetrate to different depths into the ground. Depending on the lithology, the velocity at which the waves travel is different. Seismic dispersion analysis measures these velocity changes across frequencies, providing insights into the subsurface structure and composition. This method helps identify variations in rock types and their conditions, such as porosity, which is crucial for understanding groundwater flow and storage.

At the end of 2019, we deployed a total of 53 MagseisFairfield seismic stations, including 28 1-component stations (measuring the vertical component particle velocity) and 25 3-component stations (measuring two horizontal components and the vertical component of particle velocity), on an area slightly smaller than the SP map (Figure 1). We recorded ambient seismic noise for 20 consecutive days, from November 13 to December 2. For this study, we only used the vertical component measurements, which correspond to the vertical component particle velocity movement of the ground resulting from seismic wave propagation. The processing consisted of computing cross-correlations between all the possible pairs of stations to retrieve approximations of the Rayleigh wave Green's functions between them. We used the open-source software MSNoise [Lecocq et al. 2014] to process the noise data and compute cross-correlations between the various station pairs. We obtained a total of 1378 vertical-to-vertical correlations. Following Mordret et al. [2015] and Grobde et al. [2021], we picked the Rayleigh wave fundamental mode group velocity dispersion curves using an automatic picker based on Frequency-Time analysis of the signal. This picker has a graphical interface allowing the operator to validate, correct, or reject the picked dispersion curve. This procedure resulted in 1030 dispersion curves spanning 0.4 to 5 seconds of period. We used a two-step surface-wave inversion scheme, where we first regionalize the dispersion curves at each period onto a Cartesian grid covering the area of interest. Then, for each pixel of this grid, we inverted the resulting dispersion curve into a 1D shear-wave velocity (V_s) profile with depth. Gathering all 1D profiles produces a pseudo-3D V_s volume model of the subsurface. In detail, the area is divided into square pixels of $0.001^\circ \times 0.001^\circ$ ($\sim 100\text{m} \times 100\text{m}$) between -158.0849° and -158.0313° in longitude and between 21.5123° and 21.5642° in latitude. The rest of the two-step inversion procedure is similar to the one used by Grobde et al. [2021],

using straight-ray linear inversion for the regionalization and the Neighborhood Algorithm with cubic splines on top of a power-law backbone to invert and describe the Vs profiles with depth. The main differences are the size of the Gaussian smoothing window during the regionalization of the dispersion curves, which we set to be 150 m wide, and the maximum inversion depth, which here is 2 km below the surface, though the better-resolved part of the model is the first 500 m below the surface.

5. Results

The SP and ANSWT datasets provide complementary information. The SP signal reflects near-surface electrokinetic processes and lateral variations in hydraulic conductivity and vadose-zone flow, allowing us to identify hydrological contrasts and potential flow boundaries. The ANSWT shear-wave velocity model images the 3D distribution of seismic velocities, which relate to lithology, degree of alteration, and the presence of major geological structures at depth. We therefore present the results of each method separately before integrating them into a joint geological and hydrogeological interpretation.

5.1. Self-potential

In our study area, the SP values range from -90 mV to +130 mV relative to a reference point established at an arbitrary location in the study area (see Figure 1 and 3). The SP map displays some local anomalies that are limited in their spatial extension and amplitude. For instance, Waialua pump 26 and the Helemano well exhibit a low SP anomaly that influences an area approximately 100 m in diameter around the wells (Figure 3A). Here, an "anomaly" refers to a localized deviation from the surrounding background SP level, characterized by a distinct amplitude contrast and limited spatial extent. In this context, the background SP level refers to the smoothly varying local SP trend observed elsewhere in the map.

The dominant feature observed in the SP map is a distinct contrast between a NW zone and a SE zone. We determine a demarcation line running NE-SW, approximately delineated by the sharpest spatial variation in the SP data (red dashed line in Figure 3A).

We divided the dataset into two distinct series of data points based on the proposed NE-SW separation to validate the NW and SE zones. The assignment of each SP measurement is based on its spatial position relative to the NE-SW demarcation line on the SP map (Figure 3A) and does not result from any statistical optimization or breakpoint detection. The resulting SP/elevation graph (Figure 3B) displays the complete dataset with different colors assigned to the two series. We excluded data collected in the anomalous area around the office and storage infrastructures near the Helemano well (Figures 1B and 3A), as these points,

mostly above +60 mV, clearly reflect anthropogenic interference. Considering only the remaining SP data, the SP/elevation diagram, although forming a continuous point cloud, reveals two coherent and distinct trends. These trends are effectively highlighted by plotting regression lines for the two data series and determining the SP/elevation gradient (C_e) graphically for each series [e.g., Jackson & Kauahikaua 1987; Lénat 2007]. The two SP–elevation regressions therefore characterize these spatially defined zones. The NW zone contains 1122 measurements with values ranging from 155 m to 272 m and -94 mV to +13 mV, while the SE zone contains 875 measurements ranging from 259 m to 307 m and -58 mV to +56 mV. The SE zone displays a steeper regression line (C_e of +0.83 mV/m) compared to the NW zone (C_e of +0.29 mV/m).

To assess the validity of sorting the SP data into two groups, we conducted a basic clustering analysis of our dataset (Figure 4), using version 3.35 of Orange [Demšar et al. 2013]. The highest silhouette score (0.519) was obtained for $k=2$, indicating that a two-cluster solution best captures the statistical structure of our dataset. Through multiple iterations, the Interactive k-means option allowed us to visualize each cluster and its centroids (Figure 4A.). This analysis confirmed two clusters of data corresponding roughly to the NW and SE zones of our study area. The SP/elevation gradients (C_e) values obtained using this tool show slight variations from those determined through the manual sorting of the data (Figure 4B.). The C_e values obtained from the clustering analysis (+0.67 mV/m in the SE zone and +0.16 mV/m in the NW zone) are consistent with those derived from the manual sorting, yielding similar absolute values and, importantly, preserving the same strong contrast between the two domains. The SE gradient remains roughly four times larger than the NW gradient, which confirms the clear difference in SP–elevation behavior between the zones. The difference between the results of the two methods is primarily attributable to an inherent characteristic of computing tools like Orange, which prohibits cluster overlapping. Both methods possess distinct advantages and drawbacks. Automated tools are much faster, but generate approximations within transition zones among clusters, especially in datasets comprising continuous point clouds. Conversely, the manual method, while time-consuming due to the necessity of manual data sorting, can offer more precise results. Indeed, in this case, use of the SP map and contour lines as visual indicators enhanced the precision of manually sorting data between the NW and SE zones. On the other hand, human judgment may occasionally introduce biased interpretations. Hence, comparing both methods on our dataset has been useful to validate the result.

5.2. 3D Shear-wave velocity model

Horizontal cross sections extracted from the final 3D shear wave velocity model reveal generally increasing velocity values with depth (Figure 5). Velocities range from approximately 800 m/s at the land surface to

1800 m/s in some deeper parts of the model. To approximately 100 m a.s.l., lateral velocity variations are relatively limited compared with the more pronounced spatial contrasts observed at greater depth. From 0 m a.s.l. downward, the seismic model exhibits a clear contrast between the NW and the SE halves of the study area, characterized by a strong velocity gradient from lower seismic velocities in the NW to higher velocities in the SE. The division between these two areas lies near the center of the study area and roughly coincides with the NE–SW boundary identified in the SP map. In the deepest part of the model, this NW–SE contrast fades, with more homogeneous velocities between -300 and -400 m a.s.l. (Figure 5).

In shield volcanoes, the large-scale volcanic architecture is dominated by gently inclined lava-flow sequences. Horizontal cross-sections may therefore intersect geological layers and potentially lead to misinterpretation of lateral velocity contrasts. Vertical cross-sections (Figure 6) provide a complementary visualization of the velocity contrasts at depth and across the identified NE–SW limit.

Figure 6 displays five vertical profiles whose traces are shown on the 100 m a.s.l. horizontal slice. The shallow lower-velocity layer (< 1400 m/s) is visible and continuous throughout the entire area, up to a depth of 200 m under the topographic surface - i.e., to elevations between 0 m and 200 m a.s.l., with some variations due to topography. At greater depths, we observe a contrast between the NW and the SE areas covered by the seismic tomography, aligning with the NE–SW boundary in the SP map. The transition from the shallow low-velocity layer to higher velocities in the deeper units is sharper in the SE half of the study area. This is well-illustrated by cross-section DD' located in the SE zone, close to the NE–SW delimitation. Conversely, lower velocities extend to deeper levels in the NW half of the study area. The deepest layers display progressively higher velocities, although several low-velocity columns (Figure 6, profiles AA'', CC'', and EE'') are detected down to the maximum depth of investigation.

6. Discussion

Our study area is located at the North Schofield dam, a transition zone between abnormally high groundwater levels in the island's central region, specifically the Schofield groundwater area, and "normal" groundwater levels in the northern region (Figure 2). Both the seismic tomography and self-potential data detect this transition. Although historical water-level data exist for the broader Schofield area, the spatial distribution of wells in our study area is sparse. Dale & Takasaki (1976) identified the high-level Schofield groundwater body but wells are widely separated and provide only limited spatial constraints on the position and geometry of the North Schofield Dam. Geophysical data are therefore essential to complement the available water-level information. One of the main results of our study, is the interpretation of the location, orientation, and subsurface geometry of the North Schofield Dam. However, further discussion and interpretation are required to determine its geological nature.

O‘ahu, as previously described, originated from the coalescence of two primary shield volcanoes, Ko‘olau and Wai‘anae. The storage and transfer conditions of fresh groundwater in O‘ahu are highly dependent on local geological factors associated with these volcanic edifices, such as the presence of dikes, valley fills, and caprock [Hunt 1996]. Important high-level groundwater bodies on volcanic islands typically include dike-impounded aquifers and perched aquifers [Takasaki & Mink 1985]. The Schofield high-level groundwater body is still considered a groundwater enigma [Izuka et al., 2016] because it does not fit into these expected hydrogeological settings.

According to authors like Hunt [1996], dike-impoundment is not the primary hypothesis for explaining the Schofield water body. However, given our study area's geographic position between the Ko‘olau and Wai‘anae shield volcanoes, and coincidence with the assumed location of the Wai‘anae North-East Rift Zone, the dike-impounded aquifer hypothesis should be considered in light of our new geophysical dataset. Volcanic dikes form as magma intrudes within the Earth’s crust, solidifying into vertical or near-vertical structures as they cool. This process, and associated hydrothermal alteration, results in massive units with low effective porosity. The intrusion process itself involves compression and, thus, a decrease in the porosity of the surrounding rocks [Heap et al. 2022]. Consequently, dikes impede horizontal groundwater flow, reducing the overall permeability and storage capacity of the aquifers they intersect [Izuka et al. 2018]. However, in some cases, widely spaced dikes intruded in permeable volcanic units can impound groundwater and, by raising water levels, increase storage [Takasaki et al. 1969; Takasaki & Mink 1985; Hunt 1996; Houben et al. 2018; Marazuela et al. 2023]. To explain the Schofield groundwater enigma, Oki [1998] proposed either stray dikes striking transversely to the trend of the main Ko‘olau and Wai‘anae rift zones, or former ridge spurs of the Wai‘anae Range buried under younger Ko‘olau lava flows. In the latter case, the north Schofield dam could be attributed to the existence of an ancient valley originating from the Wai‘anae volcano. The valley would have developed and accumulated sediment predominantly before the coalescence of the Wai‘anae and Ko‘olau volcanoes, and was subsequently buried by younger Ko‘olau lava flows. Its sedimentary fill and associated weathered materials could therefore constitute a low-hydraulic-conductivity unit.

6.1. Multiple sources to the NW-SE SP zonation

SP mapping allows identifying a zonation between a NW and a SE zone, consistent with the clustering analysis, which identified the same two-cluster statistical separation visible in the SP map. Due to the multiple sources contributing to SP signals [e.g., Revil et al. 2023], the NW-SE zonation can be interpreted in various ways: as a sharp variation in aquifer depth, hydrothermal upwelling in delimited areas, heterogeneity of resistivity in the porous medium, and/or spatial variations of meteoric recharge. In this

section, we evaluate these possible interpretations in light of the combined geophysical and geological evidence.

We examined spatial differences in vadose-zone thickness to assess whether they could contribute to the observed NW–SE SP contrast. Three water wells in our study area had head measurements recorded at the time of drilling (Table 1, and Figure 3 for localization). These three wells indicate nearly constant water-table elevations across the area, ranging from 213.3 m to 232.3 m. We need to ensure (1) that the correlation of data from the three wells drilled in different years is meaningful and (2) that we can extrapolate this information to the time when our SP data was acquired in 2019. Unfortunately, no water-level monitoring data is available for these three wells since their drilling. Direct water-level measurements in those wells were not possible during the 2019 survey because these boreholes were not accessible at the time of acquisition. We therefore relied on long-term records from the nearest USGS monitoring well within the Schofield hydrogeological unit, approximately 4 km south of our study area (Figure 1 and 2 for localization; 3-2901-02 Schofield MW W330-B, well ID: 212927158014801; U.S. Geological Survey 2024). This allowed us to constrain vadose-zone thickness during our measurements and to assess the potential influence of groundwater-level variations on the SP data.

The USGS record of the depth to water level below the land surface for this well is available from January 16, 1966, to present, with approximately one measurement per month and some extrapolatable gaps (Figure 7). This time span encompasses the drilling of the other three wells and our study period. This allows us to make meaningful comparisons among the three wells considered in this discussion. The record indicates a maximum amplitude of 4.67 m for the water-level variation in the Schofield well over its entire recording period until the date of access on January 10, 2024. Consequently, we can reasonably assume that the thickness of the vadose zone has remained relatively constant over time in our study area. Using the steepest SP/elevation gradient ($C_e = 0.83$ mV/m) from our study area (Figure 3), the maximum 4.67 m change in water-table elevation would produce less than $|3.9 \text{ mV}|$ of SP variation. The maximum vadose-zone thickness difference documented between wells on opposite sides of the NE–SW transition is 10.4 m (Table 1). Assuming the same maximum C_e value, such a difference would result in only $|\sim 9 \text{ mV}|$ of potential SP variation, an order of magnitude less than the observed contrast between the NW and SE domains. The combined effect of both independent factors, i.e. the measured variations of the vadose zone thickness in time and space, is thus negligible compared to the SP difference across the NE-SW transition. Consequently, it appears that the contrast in SP values between the NW and SE zones is not related to differences in the thickness of the vadose zone.

Table 1. Groundwater data from the U.S. Geological Survey [2024] for wells located in and near our study area. Water-level data are from the time of drilling. 'Schofield MW' refers to 'Schofield Monitor Well'.

well ID	Name	Year drilled	Head (m, amsl)	Elevation (m, amsl)	X (UTM)	Y (UTM)	vadose zone thickness (m)
3-2901-02	Schofield MW W330-B	1966	87.3	261	600762.5	2376430.8	173.7
3-3102-001	Helemano	1972	80.8	294.1	599214.2	2380667.7	213.3
3-3203-002	Waialua Pump 26	1975	50.3	282.6	598582	2382435	232.3
3-3204-001	Kaheaka Obs.	1994	3.7	225.6	595437.4	2382717.8	221.9
3-3405-005	Helemano Deep Monitor	2002	5	123.8	594837.6	2385179.9	118.8

In the absence of significant signs of hydrothermal activity in this part of the island [Ito et al. 2017], we can also likely dismiss the hypothesis of hydrothermal upwelling.

When considering the potential correlation between meteoric water infiltration and SP zonation, we note that in the NW zone, there is more evidence of surface drainage, notably around a deep ditch that aligns well with our most significant SP minima. However, the enhanced surface drainage in the NW is localized, and the exact process responsible for developing a more extensive stream network, potentially linked to greater groundwater infiltration, remains unclear.

One possibility is valley retrogression, a process during which ditches deepen and extend upstream, gradually exposing deeper layers after eroding the superficial cover, here clay-rich saprolites. Conversely, the surface water network could be the result of infiltration and erosion initiated in more permeable rock beneath the saprolite layer, mostly in the NW compared to the SE. Both hypotheses involve relatively superficial processes occurring in the vadose zone, which might contribute to the observed SP signal.

Another potential factor contributing to the SP contrast between the NW and SE could be heterogeneity of resistivity in the underlying terrains, as observed in various volcanic settings [Nishida & Tomiya 1987; Matsushima et al. 1990; Lénat 2007]. Volcanic edifices are typically heterogeneous, with older, more altered zones having lower resistivity. Volcano-tectonic faults may separate formations with different lithology or degrees of alteration. Consequently, the resistivity of the medium may vary from one area to another, resulting in different C_e values. In our study area, lateral variation is evident in the SP signal and C_e values from the NW and SE zones (Figure 3), suggesting a potential lithological contrast. C_e values can

505 vary depending on the medium's composition, porosity, and moisture content. However, a lower C_e (as observed in the NW zone) generally indicates higher electrical resistivity and lower hydraulic conductivity. While the SP method is primarily sensitive to processes in the vadose zone, changes in resistivity beneath the water table can also affect the SP signal, which is thus compatible with the hypothesis of lower-permeability geological formations in the NW.

510 **6.2. ANSWT as a key to the interpretation**

The ANSWT results provide crucial information for refining interpretations. Our seismic tomograms can be interpreted and extrapolated to depth in terms of lithology based on field observations, literature information, and a previous ANSWT study conducted on O'ahu [Grobbe et al. 2021].

515 We examine and interpret the ANSWT tomograms for the NW and SE zones separately, starting with the SE, where the tomograms show mostly flat layering. Based on field observations, we can associate the superficial mostly flat, low-velocity layer with soil and saprolites. O'ahu's saprolites are formed through the chemical weathering of volcanic deposits, which are primarily basalt [Sherrod et al. 2007]. This process
520 can substantially alter the hydrological characteristics of the original material, leading to decreased hydraulic conductivity as secondary clay minerals fill in pore spaces. Seismic velocities for saprolite are usually <1200 m/s [Flinchum et al. 2022], which corresponds to approximately the first 100-150 m of our ANSWT profiles (Figure 6). Below that depth, the velocity increases progressively to ~1400-1500 m/s at 200 m below the surface. From 200 m to the maximum depth of investigation, values range from 1500 to
525 1800 m/s, corresponding to values encountered for the Ko'olau basalts in Makapu'u [Grobbe et al. 2021]. Drill cores collected 10 km south of our study area demonstrate the presence of saprolite to depths of at least 30.5 m [Miller 1987]. At Schofield Barracks, about 4.5 km south of our study area, Nelson and McBride [2019] collected horizontal/vertical spectral ratio data with a 3-component seismometer and compared their results with well data. They determined that the boundary between saprolites and bedrock
530 lies 32 m to 42 m below the land surface, which, in our velocity model, approximately corresponds to velocities <900 m/s. We previously established that the water table across most of study area is located between depths of 213.3 m and 232.3 m below the land surface (Table 1). Thus we infer that the low-velocity layers beneath the saprolite-bedrock boundary, from approximately 40 m to 200 m below the land surface, are unsaturated basalts. At greater depths, the same basaltic bedrock is saturated with groundwater.
535 Heap [2019] shows that S-wave velocity decreases with porosity but depends mostly on the aspect ratio of the pores. However, if you replace the air in the pores with water, the decrease in S-wave velocity is very small compared to the effects of porosity, alteration, or pore-shape. In the tomography, we are therefore sensitive to the combination of all these factors but cannot directly determine which one is the most

influential. The increase in S-wave velocity with depth is likely more due to lithology, potentially associated with a decrease in porosity with depth or a change in the aspect ratio of the pores, rather than pore-water saturation. Below the low-velocity upper layer (<900 m/s), the increasing velocity values likely result from a decreasing degree of alteration with depth. Weathering of basalt, associated with changes including increased porosity and mineral alteration, generally reduces seismic velocities, and highly weathered horizons have been associated with lower shear-wave velocities (Von Voigtlander et al., 2018). A velocity threshold of approximately 1500 m/s can indicate a demarcation point between more altered and less altered basalt. We interpret the gradient in velocity between 900 m/s and 1500 m/s as indicative of an alteration continuum.

We expect the same superficial lithology in the NW zone, reflecting low-velocity soil and saprolites. However, there the low-velocity zone extends deeper, resembling a valley delineated by the 1400-1500 m/s isolines (Figure 6). The low velocities extending to depth argue against the hypothesis of a dense dike network as the barrier to groundwater flow, instead suggesting the possible presence of altered or detritic material similar to the valley-fill material found near the Kaiwi coast [Grobbe et al. 2021]. The valley-like structure widens from SW to NE along the EE", CC", and AA" profiles. We interpret that, as hypothesized by Oki [1998], the NW zone is intersected from SW to NE by an ancient valley associated with the Wai'anae volcano, subsequently filled with detrital material and later covered by volcanic deposits from Ko'olau. Deep paleo-valleys on Hawaiian shield volcanoes are common features of the post-shield erosional phase, where intense fluvial incision can form depressions several hundred meters deep (H. T. Stearns & Vaksvik, 1935; Hunt, 1996). Valleys with erosional relief exceeding 300 m are observed today on the flanks of the Wai'anae volcano, indicating that the depth of the structure imaged here (based on section EE") is consistent with the known geomorphological evolution of the edifice. Additional context comes from classic geomorphic interpretations of the Schofield region. Stearns & Vaksvik (1935) noted that the large valleys on the eastern side of Waianae are difficult to reconcile with modern rainfall patterns and therefore likely represent erosional features formed mainly prior to the construction of the Ko'olau edifice. Before Ko'olau grew to sufficient height to intercept the trade winds, the northeast flank of Wai'anae would have been the windward, high-rainfall side of the island, favoring the development of deep fluvial valleys. Our geophysical evidence is fully consistent with this interpretation: the valley-fill unit inferred here is best understood as an ancestral, predominantly pre-Ko'olau valley that was later buried by younger Ko'olau lava flows. Such an origin also implies that the fill has had several million years to accumulate, compact, and undergo chemical weathering, leading to the low-permeability, low-velocity material required to create the modern North Schofield hydrogeological barrier.

Underneath that paleo- riverbeds, infiltration could have significantly increased the vertical extent of weathering, potentially explaining the low-velocity vertical columns now visible below. Infiltration would also contribute to lower hydraulic conductivity at depth. Alternatively, and considering the ~1.5 km width of the structure, the imaging may indicate the presence of two merged paleo-valley branches, a pattern that is also commonly observed in modern topography. The paleo-valley direction is consistent with radial erosion of the Wai‘anae shield, with valley axes pointing away from the eruptive center.

We cannot rule out that stray dikes or dike swarms from the Wai‘anae NE rift zone may contribute as well, either at depths beyond the coverage of our study or in neighboring areas.

Figure 8 summarizes our interpretations regarding geological settings and fluid flow patterns. Profile EE” is taken as an example to provide an interpretative cross-section. The values in the SP plot (Figure 8A) have been extracted along the EE” line, revealing a sharp decrease of the SP signal from the SE to NW, which we interpret as reflecting the presence of valley fill and the associated water circulation disruption. However, at the northwestern end of the profile, the SP values do not increase as anticipated upon crossing the paleo-valley border. This discrepancy may be attributed to the water table returning to its normal deeper level on that side of the paleovalley. This is consistent with the head values observed at the Helemano Deep Monitor well, located northwest of our study area in the Waialua groundwater area (Figure 1, 2 and Table 1). The valley-fill materials likely have substantially lower hydraulic conductivity than the surrounding lava-flow aquifer. The Ko‘olau lavas may have filled part of the valley's upper section and may also be more altered within the paleovalley, as suggested by the downward extension of the lower seismic velocities, potentially reducing their hydraulic conductivity (Figure 8C).

These coupled processes have created an obstacle that dampens groundwater flow and plays a major role in the existence of the Schofield groundwater body. By reducing hydraulic conductivity, the valley fill impedes groundwater flow, and causes hydraulic heads to rise until groundwater reaches the more permeable Ko‘olau lava flows overlying the paleo-valley. This explains the abnormally-high water levels observed in the wells of this hydrogeological zone. To the NW, beyond the paleo-valley, water levels decrease toward the lower heads observed in the Waialua groundwater area, where groundwater flows through more permeable units from surficial to deeper levels.

We know of no available well log in the Schofield region, including that of the deep Helemano Monitor Well (State Well 3405-05), which provides lithological markers that constrain the thickness of Ko‘olau lavas above Wai‘anae lavas at depth [State of Hawai‘i, DLNR & Commission on Water Resource Management, 2002]. Early geological mapping indicates an erosional unconformity between the Wai‘anae and Ko‘olau volcanic formations in the Schofield area [H. T. Stearns & Vaksvik, 1935] but no depth or

thickness information was reported. Thus, existing well data neither resolve the volcanic contact nor provide thickness constraints. On another hand, the paleo-valley carved into the Wai‘anae flank and interpreted from our ANSWT model (Figure 8B), must have formed during or after the waning of Wai‘anae volcanism and was later buried by younger Ko‘olau basaltic flows. Thus, our study suggests a minimum thickness of 100 m of Ko‘olau lavas overlying the edge of the paleo-valley, i.e. the eroded Wai‘anae surface, in this sector of the Schofield Plateau.

7. Conclusions

Our study improves the characterization of the Central O‘ahu groundwater flow system and its underlying geological controls. Integrating SP data, ambient noise seismic tomography, geological observations, and hydrogeological interpretation has allowed us to identify major features governing groundwater flow in the Schofield region.

Our investigation has revealed a clear zonation of the self-potential signal across the North Schofield Dam, matched at depth by the distribution of seismic velocities. Together, these results delineate the orientation and geometry of one of the most significant hydrogeological boundaries on the island. The nearly constant thickness of the vadose zone across the study area cannot account for the observed NW-SE SP contrast, while the lower C_e value in the NW zone, is consistent with spatial variations in the groundwater flow regime, which may include changes in flow direction and/or velocity. Additionally, the lower C_e value, suggesting higher electrical resistivity, can be explained by lower effective porosity of dry clay-rich alluvium or detrital material. This inference, and the valley-shaped low-velocity structure in the NW zone, combine to support the interpretation of an ancient valley carved into the Wai‘anae volcano and later covered by Ko‘olau volcanic activity. We interpret the abnormal groundwater behaviour in Central O‘ahu to the presence of low hydraulic conductivity valley fill material. Identifying and characterizing this paleo-valley provides essential context for understanding the region’s groundwater dynamics.

The paleo-valley geometry imaged in our ANSWT model also offers new constraints on the volcanic stratigraphy beneath the Schofield Plateau. While existing well logs do not distinguish Ko‘olau from Wai‘anae units, and therefore cannot quantify the thickness of the Ko‘olau cover, our geophysical interpretation indicates a minimum of ~100 m of Ko‘olau lavas overlying the eroded Wai‘anae flank in this area.

Our findings also contribute to the understanding of how groundwater is transferred from inland recharge areas toward the coastal aquifers of O‘ahu. Our interpretation of the North Schofield Dam as a deeply buried, low-permeability paleo-valley, strongly supports its role as a hydrogeological barrier influencing

regional water partitioning. This structure may influence the relative proportions of freshwater that flows northward toward the Waialua basal aquifer and southward toward the Pearl Harbor basal aquifer. These insights are essential for refining island-scale groundwater models, improving water-resource management, and understanding volcanic hydrogeological systems.

This study paves the way for future hydrogeophysical investigations on O‘ahu. While we believe we have identified the nature of the North Schofield Dam, further work is required to determine whether this structure is unique or if multiple transverse paleo-valleys exist and may also explain the South Schofield Dam. Numerical simulations of groundwater flow, incorporating the constraints developed here, will be an important next step for informing freshwater-management strategies on O‘ahu. Because such hydrogeological barriers are common on volcanic islands, the conceptual model presented here may also prove valuable for other tropical volcanic settings.

Author contributions

S. Barde-Cabusson: Conceptualization, Methodology, Validation, Formal analysis, Investigation, Data curation, Writing – original draft, Writing – review & editing, Visualization. **A. Mordret:** Conceptualization, Methodology, Software, Formal analysis, Investigation, Data curation, Writing – original draft, Writing – review & editing, Visualization. **N. Grobbe:** Conceptualization, Methodology, Formal analysis, Investigation, Writing – review & editing. **D. Dores:** Conceptualization, Investigation, Writing – review & editing. **N. Lautze:** Conceptualization, Writing – review & editing. **J. Sinton:** Conceptualization, Writing – review & editing. **E. Wallin:** Conceptualization, Writing – review & editing.

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Data Availability

The data supporting the conclusions of this study are publicly available on HydroShare; The SP data at Barde-Cabusson, S., N. Grobbe (2024). Dole Plantation Self Potential map, HydroShare, <https://doi.org/10.4211/hs.31b908b569b64beabfd9e7762c060110>; the ANSWT data at Barde-Cabusson, S., N. Grobbe (2024). Continuous Ambient Noise Seismic Data at Central O‘ahu, Hawai‘i, HydroShare, <https://doi.org/10.4211/hs.a0fc9c33ed9c430e9b5299d38369c3d0>.

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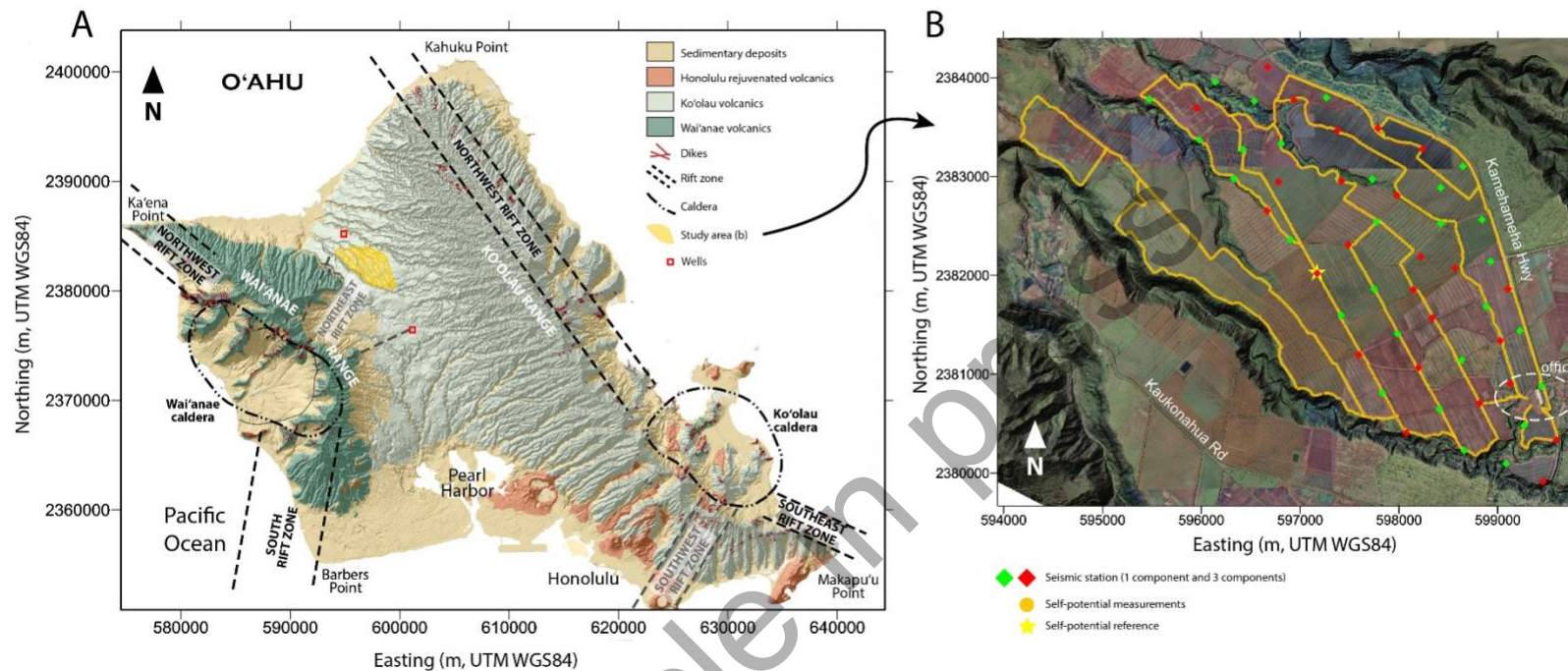


Figure 1: Study site. A: Simplified geological map (adapted from Sherrod et al., 2007) overlain on a Digital Elevation Model (from the Hawai'i Statewide GIS Program, <https://geoportal.hawaii.gov/>). Structural data are from Hunt (1996) and Sinton and Rowland (2022). Two wells of regional importance, Helemano Deep Monitor Well (northern red square) and Schofield MW W330-B (southern red square), are shown here. Three additional wells inside the study area are visible in Figure 3; B: Localization of the geophysical datapoints: orange dots correspond to the locations of the SP measurements; The seismic array of 53 sensors used in this study is shown with the 25 3-component sensors in red and the 28 1-component sensors in green. The office and storage facilities of the agricultural facilities are highlighted as "office".

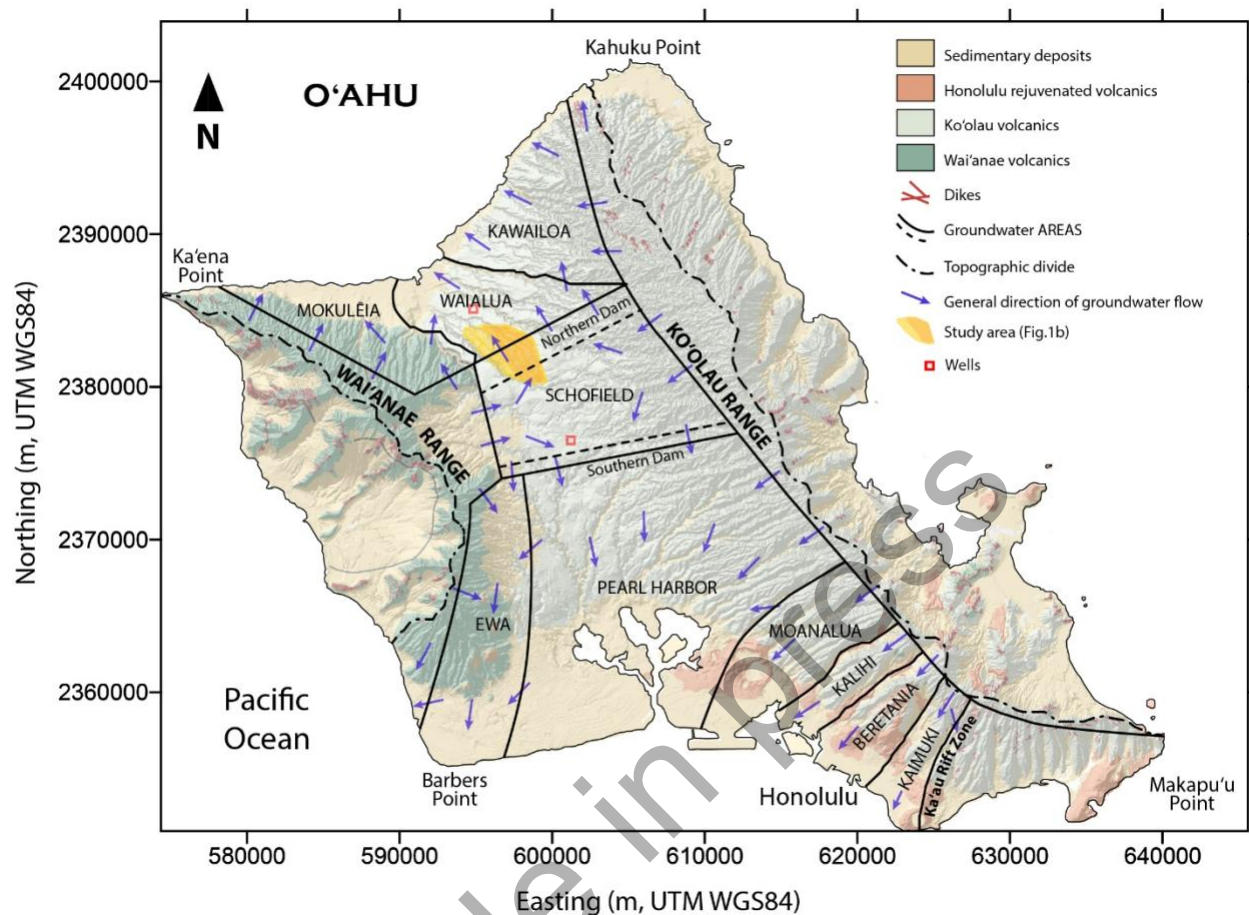


Figure 2: Groundwater areas within the central O'ahu groundwater flow system. Groundwater information (based on Oki, 1998) overlain on the simplified geological map (adapted from Sherrod et al., 2007). The SP survey area is shown in light orange, with the overlapping ANSWT survey area highlighted in darker orange. Two wells of regional importance, Helemano Deep Monitor Well (northern red square) and Schofield MW W330-B (southern red square), are shown here. Three additional wells inside the study area are visible in Figure 3.

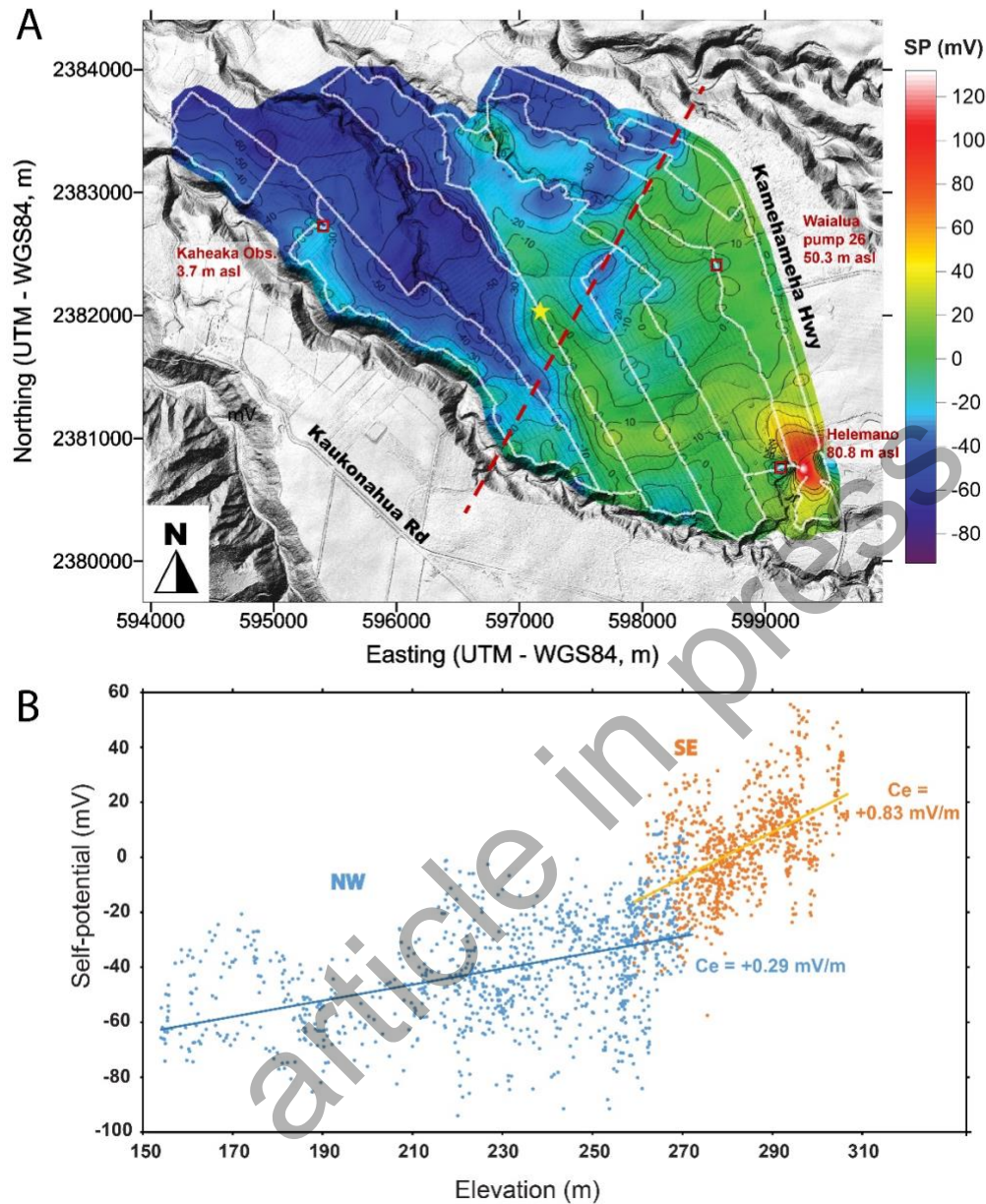


Figure 3: Self-Potential (SP) results. A: SP map of the study area overlain on a 1 m Digital Elevation Model (DEM) from Hawaii Statewide GIS Program, <https://geoportal.hawaii.gov/>. The color bar indicates the SP values relative to an arbitrary reference (yellow star). White dots represent SP measurements; the red dashed line highlights the NE-SW demarcation line delineating the sharpest spatial variation in the SP data; Three water wells are indicated with red squares. B: SP/elevation plot of the dataset, except for the data acquired near the office area (red tones at the SE tip of the map). The SP/elevation gradients (Ce) of the NW and SE SP-zones are indicated. The two zones are those separated by the red dashed line in A.

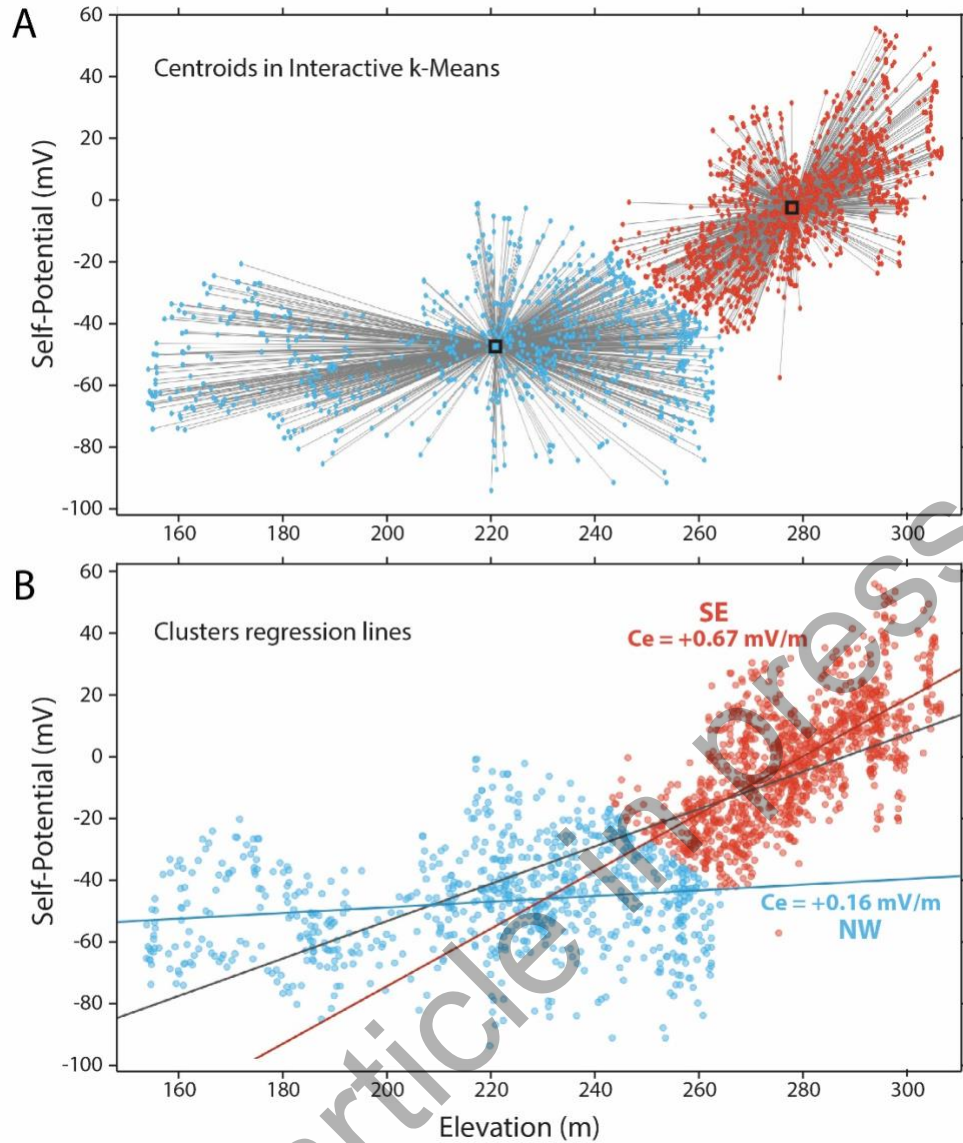


Figure 4: k-means clustering analysis of our SP dataset. We used the 3.35 version of Orange, an open-source data visualization and analysis toolbox in Python (Demšar et al., 2013). A: Interactive k-means option to visualize the two clusters and their centroids and to refine the result; B: SP/elevation gradients (Ce) values obtained for the two clusters. The black line is the regression line for the complete dataset.

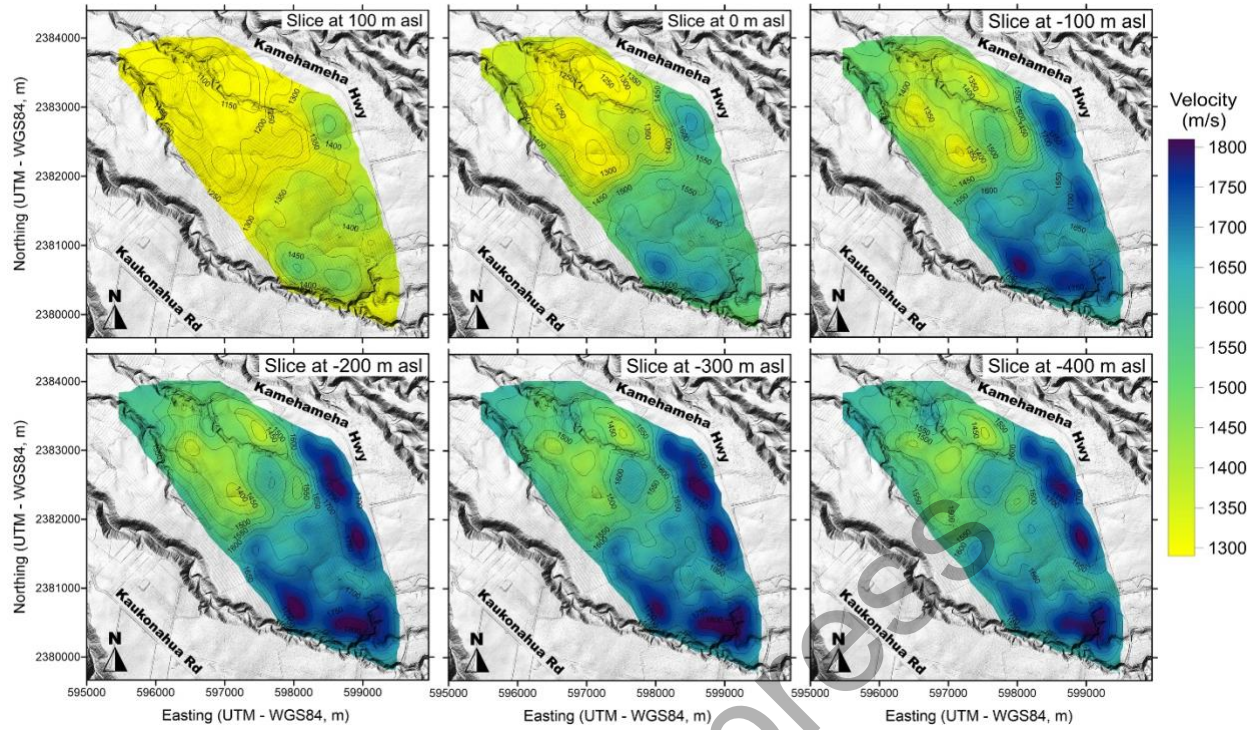


Figure 5: Seismic Ambient Noise Surface Wave Tomography (ANSWT) results. Horizontal slices in the final 3D shear-wave velocity model at 100, 0, -100, -200, -300, and -400 m above sea level. The surface wave tomography results are draped on a 1 m-resolution DEM (Hawaii Statewide GIS Program, <https://geoportal.hawaii.gov/>) of the studied area. Yellow represents slower velocities and the contour lines are displayed on each slice.

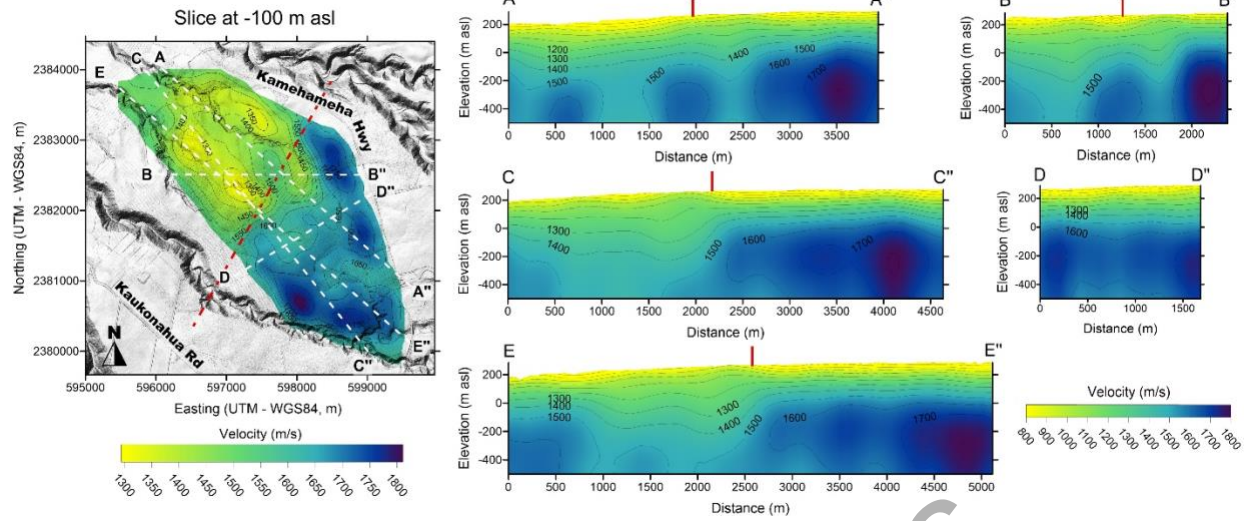
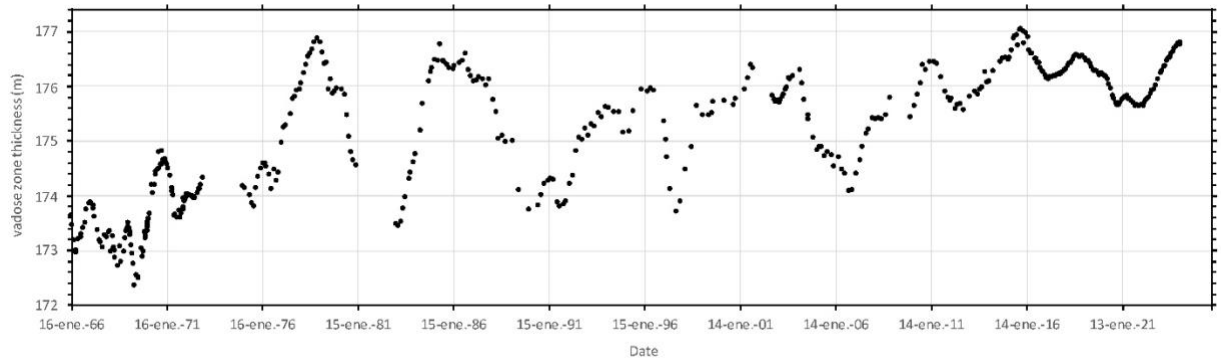


Figure 6: Vertical cross sections through the 3D shear-wave velocity model. The traces of the sections are shown in the left panel, which is a horizontal slice of the shear wave velocity model at -100 above sea level. The red dashed line highlights the NE-SW demarcation delineating the strongest spatial variation in the SP data (see Figure 3) and red markers on the vertical sections indicate where this demarcation intersects.



900 Figure 7: Record of vadose zone thickness variations near our study site, 1966-2024. Data recorded at a
 well located ~4 km south of our study area (3-2901-02 Schofield MW W330-B, well ID:
 212927158014801; see Figure 1 and 2 for localization); data available from the U.S. Geological Survey
 (2024). Record from January 16, 1966 to the January 10, 2024, with approximately one measurement per
 month (https://waterdata.usgs.gov/monitoring-location/USGS-
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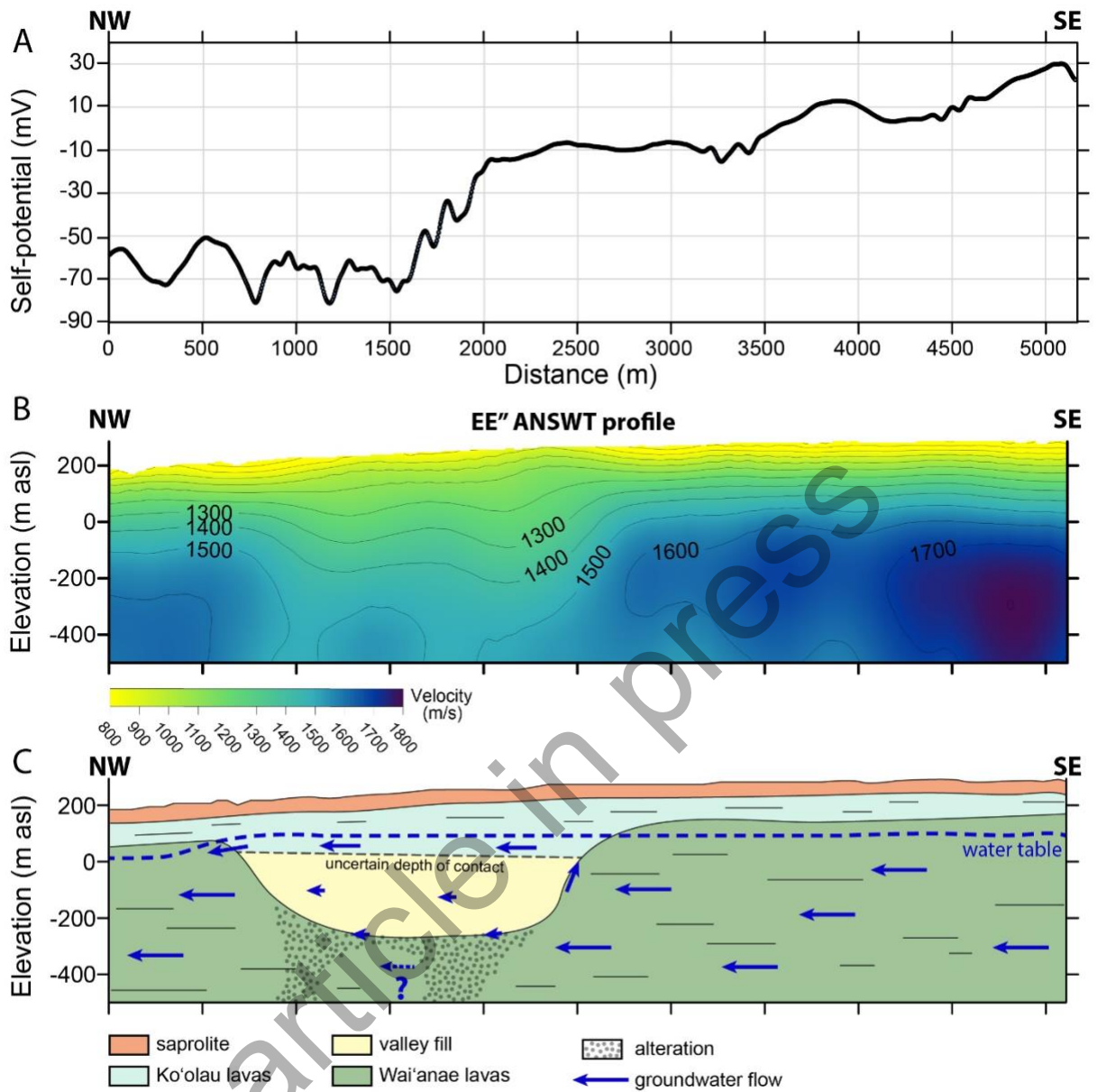


Figure 8: Interpretative synthesis. A: SP plot extracted along the EE'' ANSWT profile in B (see location in
 910 Figure 6); C: geological and hydrogeological interpretation, with water table depth inferred from wells
 listed in Table 1.