

GEOLOGICAL, GEOPHYSICAL AND GEOCHEMICAL STUDIES ON THE HYDROTHERMAL SYSTEM OF MENENGAI GEOTHERMAL FIELD, KENYA

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GEOLOGICAL, GEOPHYSICAL AND GEOCHEMICAL STUDIES ON THE HYDROTHERMAL SYSTEM OF MENENGAI GEOTHERMAL FIELD, KENYA

by

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Requirement for the Degree of

Doctor of Engineering
(Earth Resources Engineering)

Supervised by

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TABLE OF CONTENTS

LIST OF FIGURES	i
LIST OF TABLES	viii
ACKNOWLEDGMENTS	ix
DEDICATION.....	xi
ABSTRACT.....	xii
1. GENERAL INTRODUCTION	1-1
1.1. Background of the study	1-1
1.2. Geothermal status in Kenya	1-3
1.3. Previous studies.....	1-4
1.4. Purpose and plan of study	1-6
1.5. Structure of the dissertation.....	1-8
References	1-10
2. FIELD CHARACTERISTICS OF MENENGAI GEOTHERMAL AREA	2-1
2.1. Introduction	2-1
2.2. Geology and structural setting.....	2-1
2.2.1. Regional setting	2-1
2.2.2. Local geology.....	2-3
2.2.3. Lineament studies	2-7
2.3. Hydrological setting	2-10
2.3.1. Regional Relief and Rainfall Distribution	2-10
2.3.2. Regional hydrology.....	2-12
2.3.3. Hydrological models.....	2-13
2.4. Conclusion.....	2-17
References	2-18
3. GRAVITY DATA ANALYSIS AND INTERPRETATION.....	3-1
3.1 Introduction	3-1

3.2	Gravity method: theory and application	3-1
3.3	Gravity survey	3-3
3.4	Gravity data reduction and Bouguer density estimation	3-4
3.4.1	Gravity data correction	3-4
a).	Latitude correction (normal gravity)	3-5
b).	Free-air correction	3-5
c).	Bouguer correction	3-6
d).	Terrain correction	3-6
3.4.2	Bouguer density estimation	3-8
a).	Simple G-H method.....	3-8
b).	F-H method.....	3-8
c).	Extended F-H.....	3-9
d).	Akaike's Bayesian Information Criterion (ABIC).....	3-10
e).	Comparison with the Variance of the Upward-Continuation (CVUR).....	3-11
3.5	3-D gravity inversion	3-12
3.6	Results and Discussion.....	3-16
3.6.1	Bouguer density estimation	3-16
3.6.2	Bouguer anomaly.....	3-21
3.6.3	3-D gravity modeling.....	3-23
3.7	Conclusion.....	3-32
References		3-33
4. RESISTIVITY STUDIES USING MAGNETOTELLURIC METHOD		4-1
4.1.	Introduction	4-1
4.2.	The magnetotelluric method.....	4-1
4.3.	Magnetotelluric in geothermal exploration.....	4-4
4.4.	Magnetotelluric data acquisition, processing and analysis	4-8
4.4.1.	Data acquisition	4-8
4.4.2.	Distortion and static shift in MT methods	4-9
4.4.3.	Dimensionality analysis.....	4-12
a).	Swift-skew and ellipticity.....	4-13
b).	Phase tensor	4-15
4.5.	1-D and 2-D Occam's inversion of MT data.....	4-19
4.6.	3-D ModEM inversion of MT data	4-20
4.7.	Results and Discussion.....	4-22
4.7.1.	MT inversion results	4-24
4.8.	Conclusion.....	4-31

References	4-32
-------------------------	-------------

5. ISOTOPE AND WATER CHEMISTRY STUDIES	5-36
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5.1. Introduction	5-36
5.1.1. Isotope geochemistry	5-36
5.1.2. Reactive transport modeling	5-38
5.2. Geochemistry in geothermal exploration	5-40
5.3. Physical and chemical characteristics of Menengai geothermal field.....	5-40
5.4. Results and discussion.....	5-43
5.4.1. Major chemical elements studies	5-44
5.4.2. Isotope Studies	5-49
5.5. Reactive geochemical modeling.....	5-53
5.5.1. Physical and chemical characteristics of MW-01	5-53
5.5.2. Mineral and geochemical data	5-55
5.5.3. Equilibrium and kinetic considerations	5-61
5.5.4. Model setup.....	5-64
5.5.5. Results and discussion	5-66
5.6. Conclusion.....	5-73

References	5-74
-------------------------	-------------

6. COMBINED INTERPRETATION AND DISCUSSION	6-85
--	-------------

3.8 Introduction	6-85
3.9 Menengai caldera	6-85
3.10 The Olrongai area.....	6-93
3.11 Conclusion.....	6-99

References	6-100
-------------------------	--------------

7. GENERAL CONCLUSIONS	7-1
-------------------------------------	------------

7.1. Introduction	7-1
7.2. Field characteristics.....	7-1
7.3. Gravity data analysis and interpretation.....	7-2
7.4. Resistivities studies using the MT method.....	7-3
7.5. Isotope and water chemistry studies.....	7-3
7.6. Combined interpretation and discussion	7-4

LIST OF FIGURES

Figure 1.1: Map of Kenya showing the location of geothermal sites (in red) and the rift structures (black lines). Also indicated are water bodies (shown in light blue polygons) and major cities.	1-2
Figure 1.2: Schematic diagram that shows the stages and targets of the present studies. The properties of wells are represented by x , P , T , h and M which represents chemical composition, Pressure, Temperature, enthalpy, and mass flow, respectively.....	1-8
Figure 2.1: Location map of the study area. The high altitude (>1500 m above sea level) area shows part of the Kenya dome.	2-3
Figure 2.2: Descriptive map of the study area showing the elevation (m), faults, fracture and location of fumaroles.	2-5
Figure 2.3: Surface geology map of the study area, modified from previous work done by McCall (1967), Jennings (1971) and Walsh (1969).....	2-6
Figure 2.4: Lineament density map of the study area per km^2 . The Rose diagrams show the azimuths direction of the local distribution of lineaments, with radial coordinate indicating the number of lineament.	2-9
Figure 2.5: Annual average rainfall and relief map of Kenya.	2-11
Figure 2.6: Main geological structures and local hydrology. The study area is shown in grey, and the green line is considered as an area of interest from a hydrological perspective. Major faults, perennial rivers and ephemeral rivers are indicated as broken red lines, blue lines and broken blue lines, respectively.....	2-12
Figure 2.7: Hydrogeological controls of the Menengai geothermal area (adapted from Mungania & Lagat, 2004). Boreholes are represented by circles.	2-14
Figure 2.8: Hydrogeological model of the study area indicating possible fluid recharge sources of the geothermal system.....	2-16
Figure 3.1: Geometry representation of the points K and Q and distance r between them; notations of equation [3.3].....	3-2

Figure 3.2: Reference model from the gridded density contrast of the down hole density estimated from the initial 10 geothermal wells drilled in Menengai.....	3-14
Figure 3.3: A 3-D Gravity inversion grid. The cell size is 620 m × 650 m × 50 m with the height component set to increase by a factor of 1.08 for each successively deeper layer. The colored map represents complete Bouguer anomaly (CBA) with values ranging from -189.4 mGal to -134.7 mGal.	3-15
Figure 3.4: Bouguer density by G-H method.....	3-18
Figure 3.5: Bouguer density by the Simple F-H method.....	3-19
Figure 3.6: (a) Bouguer density by Extended F-H method, (b) Extended F-H density frequency distribution.	3-20
Figure 3.7: The Bouguer density by Comparison with Variance of the Upward-Continuation (CVUR).....	3-21
Figure 3.8: Bouguer anomaly map calculated for a Bouguer density of $2.23 \times 10^3 \text{ kg m}^{-3}$	3-23
Figure 3.9: A planar view of density distribution at 1000 m asl elevation.....	3-25
Figure 3.10: A planar view of density distribution at sea level.	3-26
Figure 3.11: A planar view of density distribution at 1000 m b.s.l elevation.....	3-27
Figure 3.12: A planar profile and isosurface ($2.45 \times 10^3 \text{ kg m}^{-3}$) inferred from the density model of Menengai area.....	3-28
Figure 3.13: A display of geological faults as seen from the surface and fumaroles location superimposed against a deep planar layer at sea level and an isosurface of $2.45 \times 10^3 \text{ kg m}^{-3}$ to highlight the dense body of Olrongai hill.....	3-29
Figure 3.14: A display of geological faults/caldera rim ring-fracture as seen from the surface and fumaroles location superimposed against a deep planar layer at sea level and an isosurface of $2.45 \times 10^3 \text{ kg m}^{-3}$ to highlight the dense body of Menengai caldera. The inferred faults are presented as broken lines.....	3-30
Figure 4.1: Resistivity and conductivity values of various rock-forming materials (from González-Álvarez et al., 2014; adapted from Palacky, 1988).....	4-6
Figure 4.2: A conceptual model of a generic geothermal system with a 250 to 300°C geothermal reservoir with isotherms, alteration zones, and structures.	4-7
Figure 4.3: Schematic representation of the field set-up for a typical 5-channel MT sounding (adapted from Phoenix Geophysics Ltd., 2010). E_x and E_y dipoles represent the non-	

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-
- polarizing electrodes for measuring the electric field components. The magnetic induction coils indicated by H_x and H_y measure the horizontal components of the magnetic field, and H_z shows its vertical component. The directions are labeled as x, y, and z, whereas the E and H are the electric field and the magnetic field, respectively. Hence, the acquisition unit records the time-series data of all five electromagnetic field components (i.e., E_x , E_y , H_x , H_y , and H_z). 4-8
- Figure 4.4. Typical magnetotelluric data collected in the study area displayed as apparent resistivity and phase curves. The curves on the left side show the raw data collected while those on the right side display the results after the removal of static shift and distortion. Period acts as a proxy for depth, with short periods sensing shallow structures and long periods sensing deeper structures. 4-11
- Figure 4.5: Plots of all MT sounding curves of TE and TM modes from the 212 stations in Menengai field. The red dots indicate the TE mode and the blue dots indicate TM mode. The consistent shape of all the curves is displayed. 4-12
- Figure 4.6: Swift-skew and ellipticity are plotted for selected MT stations. 4-14
- Figure 4.7: Graphical representation of the phase tensor as an ellipse. The magnitudes of the minimum and maximum principal axes are given by Φ_{min} and Φ_{max} , respectively. If the phase tensor is non-symmetric, a third coordinate invariant is required to describe the tensor, which is the skew angle, β . The angle $\alpha - \beta$ gives the orientation of the major axis of the ellipse and defines the link between the tensor and the observational reference frame (x_1 and x_2); adapted from Caldwell et al. (2004). 4-16
- Figure 4.8: Phase tensor maps for periods of (a) 0.0033 s, (b) 0.01 s, (c) 1 s, (d) 10 s, (e) 100 s, and (f) 300 s. 4-18
- Figure 4.9: Example of data misfit between observed and modeled station responses of the Z_{xy} (*red dots*) and Z_{yx} (*blue*) components for some selected stations. The dots and the lines represent the measured values and modeled response, respectively. 4-23
- Figure 4.10: Menengai caldera map showing 3-D MT inversion profiles (in blue) along E-W directions (P1-P4) and S-N direction (P5-P8), and 2-D MT inversion profiles (in red) P9-P12. Four MT stations selected for 1-D Occam's inversion are encircled in green. The orientation of 2-D profiles was chosen to ensure it crosses near as much data as

possible. The black and orange dots show the MT stations and the caldera rim, respectively.	4-24
Figure 4.11: MT profiles along W-E directions (P1-P4).	4-25
Figure 4.12: MT profiles along S-N directions (P5-P8).	4-26
Figure 4.13: Horizontal slices at different heights above sea level through the 3-D resistivity model derived from inversion of the caldera MT dataset. The black and orange dots show the MT stations and the caldera rim, respectively.	4-27
Figure 4.14: Horizontal slices at different heights above sea level through the 3-D resistivity model derived from inversion of Olrongai area MT dataset. The orange dots and brown lines denote fumaroles and faults, respectively.	4-28
Figure 4.15: 2-D resistivity models for profiles P9, P10, P11 and P12 of Menengai caldera as shown in Figure 4.10.	4-29
Figure 4.16: 1-D resistivity models of the caldera area from MT stations 052, 006, 007 and 164 align from north to south (see Figure 4.10), respectively, and situated right above the low resistive body (CC2).	4-30
Figure 5.1: Governing equations for a single phase coupled reactive transport system (Steefel et al., 2005).	5-38
Figure 5.2: Cl-SO ₄ -HCO ₃ ternary diagram (after Giggenbach, 1988) showing the distribution of major anions in the geothermal fluid of Menengai wells. The water type is predominantly bicarbonate (HCO ₃ ⁻). Various colored symbols are used to distinguish different wells.	5-41
Figure 5.3: Scatter plots of SiO ₂ versus CO ₂ (mg/kg) of Menengai well discharge waters collected during the testing period. Similarly colored symbols represent different samples from the same well while the orange lines indicate hypothetical mixing boundaries.	5-45
Figure 5.4: Scatter plots of Na versus SiO ₂ (mg/kg) of Menengai well discharge waters collected during the testing period. Similarly colored symbols represent different samples from the same well while the orange lines indicate hypothetical mixing boundaries.	5-46
Figure 5.5: Scatter plots of CO ₂ versus Na (mg/kg) of Menengai well discharge waters collected during the testing period. Similarly colored symbols represent different samples from the same well while the orange lines represents hypothetical mixing boundaries..	5-47

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-
- Figure 5.6: Scatter plots of CO₂ versus Cl (mg/kg) of Menengai well discharge waters collected during the testing period. Similarly colored symbols represent different samples from the same well while the orange lines indicate hypothetical mixing boundaries. 5-48
- Figure 5.7: Scatter plots of Na versus Cl (mg/kg) of Menengai well discharge waters collected during the testing period. The symbols represent different wells while the orange lines indicate hypothetical mixing boundaries. 5-49
- Figure 5.8: δD (²H) and $\delta^{18}O$ of some of the Menengai well fluid (red dots); cold borehole (blue dots) and warm borehole water (blue dot enclosed in an orange circle), and; rift valley lakes (stars). Also presented are the Global Meteoric Water Line - GMWL (Craig, 1961), Kenya Rift Meteoric Water Line - KRMWL (Allen & Darling, 1987), Central Kenya Meteoric Water Line – CKMWL and Continental African Rain Line –CARL, defined by Ármannsson (1994). The boreholes are named after their well number with prefix ‘M’ for Menengai, and suffixes ‘ws’ and ‘wb’ are abbreviations for the names of the sampling points, i.e., for Webre steam-water separator and weir-box, respectively. 5-50
- Figure 5.9: (a) Various processes which shift the δD and $\delta^{18}O$ values from the MWL. Evaporation shifts both δD and $\delta^{18}O$ values. The δD is displaced as a result of an isotopic exchange with volcanic CO₂ and limestone while $\delta^{18}O$ is shifted due to exchange with H₂S and silicate hydration. (b) δD and $\delta^{18}O$ plot for continental precipitation (MWL = Meteoric Water Line; local corresponds to the Mediterranean precipitation) with examples of various mixing lines. The MWL of Pleistocene palaeowater may be apart from the MWL (adapted from Geyh et al., 2008)..... 5-51
- Figure 5.10: R/Ra and ³He/⁴He values versus ⁴He/²⁰Ne ratio diagram of sampled gas data from Menengai wells (blue dots), a borehole (blue dot encircled with a black circle) and a fumarole (blue dot encircled with an orange circle) samples. The theoretical lines represent binary mixing trends for mantle-derived, crustal and atmospheric helium. The end member values for ⁴He/²⁰Ne ratios are 1000 for crust and mantle and 0.318 for air. The end-members for the ³He/⁴He is 0.01 Ra for the crust, 1 Ra for air and 8 Ra for MORB. The ratio (R) of ³He to ⁴He is often used to represent ³He content and R usually is given as a multiple of the present atmospheric ratio (Ra)..... 5-52
-
-

Figure 5.11: Map of Menengai caldera showing the location of geothermal well MW-01 and groundwater borehole BH01. The location of fumaroles is represented by asterisks symbols.	5-53
Figure 5.12: Downhole temperature and pressure profiles measurement of MW-01 measurement obtained after the 44 days heating-up after the first work-over operation. Also indicated are the inferred aquifer depths and boiling point curve with depth.	5-55
Figure 5.13: Schematic diagram describing the processes occurring in the depressurization zone around wells (Arnórsson et al., 2010). Where $h^{d,t}$ is discharge enthalpy (kJ/kg); P^d is Sampling pressure (bar abs.); T^f is aquifer temperature (°C), and; $X^{f,v}$ is Initial aquifer fluid vapor fraction. $m_i^{f,v}$, $m_i^{f,l}$, $m_i^{d,v}$ and $m_i^{d,l}$ are concentration of component i in the initial aquifer steam, the initial aquifer water, the discharged steam, and the discharged water, respectively. $M^{e,v}$, $M^{e,l}$ and Q^e are mass flow of steam, mass flow of water and mass flow enthalpy in the intermediate zone, respectively. $M^{f,t}$ and $M^{d,t}$ are the total mass flow at initial aquifer and discharge conditions, respectively.....	5-56
Figure 5.14: Downhole secondary minerals distribution of MW-01 recorded during the drilling operation.....	5-57
Figure 5.15: Total alkali versus silica (TAS) diagram of Menengai volcanics. The red dots represent samples that were collected on 2017 survey, while the green dots are reported in Macdonald et al. (2011).	5-59
Figure 5.16: The model setup for the 1-D reactive transport simulation.....	5-65
Figure 5.17: Comparison between the results of calcite deposition, Ca^{2+} concentrations, and diopside dissolution after 1000 years of simulation, and observed field data.	5-67
Figure 5.18: Comparison between the results of illite after 1000 years of simulation and observed field data.....	5-68
Figure 5.19: Comparison between the results of zeolites after 1000 years of simulation and observed field data.	5-69
Figure 5.20: Comparison between the results of quartz after 1000 years of simulation and observed field data.....	5-71
Figure 5.21: Comparison between the results of chalcedony after 1000 years of simulation and observed field data.	5-71

Figure 5.22: Comparison between the results of epidote after 1000 years of simulation and observed field data.	5-72
Figure 5.23: Change of porosity after 1000 years of simulation.	5-73
Figure 6.1: Menengai caldera map showing the location of the 3-D MT inversion profiles (blue line) along West-East and South-North directions. The cross sections are shown in Figure 6.2 and Figure 6.3. A density isosurface of $2.55 \times 10^3 \text{ kg m}^{-3}$ extracted from 3-D gravity model is also plotted. Temperature (ranging from 20 – 57 °C) of water drawn from boreholes spread around the study area is indicated by colored dots. The fumaroles are presented as red stars.....	6-87
Figure 6.2: South-North cross-section of 3-D resistivity model and a density isosurface of $2.55 \times 10^3 \text{ kg m}^{-3}$ (shown as a mesh) extracted from the 3-D gravity model.	6-88
Figure 6.3: West-East cross-section of the 3-D resistivity model and a density isosurface of $2.55 \times 10^3 \text{ kg m}^{-3}$ (shown as a mesh) extracted from the 3-D gravity model.	6-89
Figure 6.4: 3-D perspective view of the resistive isosurface (blue blocks (CR2), $>90 \Omega.m$) and 3D geometry of the dense body isosurface (transparent mesh, $>2.55 \times 10^3 \text{ kg m}^{-3}$) of caldera area. Planar view is a horizontal slice of density anomaly at 3000 m bsl. .	6-90
Figure 6.5: A perspective model of the 3-D gravity and magnetotelluric inversions of the Menengai caldera area. The block represents magnetotelluric results whereas the transparent mesh indicates the dense body isosurface ($2.55 \times 10^3 \text{ kg m}^{-3}$).....	6-91
Figure 6.6: A joint model 3-D gravity and magnetotelluric inversions, and $^3\text{He}/^4\text{He}$ isotopes contour map (referenced at an elevation of -500 m a.s.l) of the Menengai caldera area. The block represents magnetotelluric result ($<10 \Omega.m$) whereas the transparent mesh indicates the dense body isosurface ($2.55 \times 10^3 \text{ kg m}^{-3}$).	6-92
Figure 6.7: Cross sections of 3-D resistivity model (b and c) along profiles ‘west-east’ and ‘south-north’ indicated on (a) in the Olrongai area. On each of the cross-section, a density isosurface of $2.5 \times 10^3 \text{ kg m}^{-3}$ extracted from the 3-D gravity model is also plotted (also plotted on (a)). Circles in Figure 6.7a show temperature (ranging from 20 – 57 °C) of water drawn from boreholes spread around the study area. The orange line, the red stars (shown as black squares in b and c) and brown lines in Figure 6.7a are the Menengai caldera rim, fumarole locations, and geological faults, respectively.	6-94

Figure 6.8: 3-D perspective view of the resistive isosurface (blue blocks, $>90 \Omega.m$) and 3D geometry of the dense body isosurface (transparent mesh, $>2.5 \times 10^3 \text{ kg m}^{-3}$) of Olongai. Planar view is a horizontal slice of density anomaly at 3000 m bsl.	6-96
Figure 6.9: A perspective model of 3D gravity and magnetotelluric inversions in the Olongai area. The block model represents magnetotelluric results whereas the transparent mesh indicates the dense body isosurface ($2.5 \times 10^3 \text{ kg m}^{-3}$).	6-98
Figure 6.10: A conceptual model of the Menengai geothermal area.	6-99

LIST OF TABLES

Table 3.1: Different rock types and their depths of occurrence from the initial 10 geothermal wells drilled in Menengai geothermal field (adapted from GDC, 2018).	3-13
Table 5.1: Mineral composition and their chemical formula of phenocrysts	5-59
Table 5.2: Minerals and their volume proportion obtained from the CIPW normative scheme. The surface area and reaction mechanisms are also shown.	5-60
Table 5.3: Initial rock mineral composition and potential alteration minerals along with parameters for calculating kinetic rates for mineral dissolution and precipitation. The kinetic parameters used are from Palandri & Kharaka (2004).	5-63

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*So whether you eat or drink or whatever you
do, do it all to the Glory of God.*

1 Cor. 10:31

ABSTRACT

Menengai volcano is one of the late Quaternary caldera volcanoes formed on a massive shield in the inner-trough of the Kenya rift valley associated with a high thermal gradient as a result of shallow magmatic intrusions. Menengai geothermal area lies within the central part of this rift section, which is characterized by a complex geological set up associated with regional uplift referred to as the Kenya dome, with Menengai caldera being central for its geothermal resource and conspicuously the major geological feature in the area. The tectonic setting of Menengai is two-fold and consists of a much older Olrongai system of NNW trending faults and the NNE trending Solai system which formed later and constitutes a thermally active zone. The geothermal activities is manifested by the occurrence of fumaroles, steaming-gas boreholes, hot-warm water in boreholes and altered grounds. While drilling of geothermal wells in this field proved the existence of exploitable steam, the complexity of its geological setup has led to various technical setback where promising well targets turned out to be unproductive. These challenges were attributed partly to lack of enough knowledge of factors controlling the geothermal system. There is a need, therefore, of more detailed study to delineate the structures that are deemed to constraint the resource. In this study, investigations were conducted by use of 166 magnetotelluric (MT) soundings and 1823 gravity stations. The purpose of these geophysical data was to facilitate imaging of the subsurface geothermal structure with a view of confirming and delineating the extent of the geothermal resource. The distribution of the acquired MT data was concentrated around two areas; the caldera and the Olrongai areas, which are prominent for their thermal manifestations. It was necessary to subdivide the data into these two parts in order to obtain a better inversion result and this also helped ease the computation. Both analyses of the MT data revealed that the dimensionality of the subsurface shows a complex scenario where the 1-D and 2-D dimensionality effects are prominent at shallow depth whereas 3-D characteristics are detected at deeper levels. These insights from dimensionality analysis informed the need for a 3-D inversion to image the complex subsurface structures at a deeper depth. Moreover, to support the 3-D results, 1-D and 2-D inversion were carried out. The three models, 1-D, 2-D and 3-D, were able to reveal similar subsurface resistivity structure in the area of study. From the gravity data, an

improved Bouguer anomaly map of the area was generated using a Bouguer density of $2.23 \times 10^3 \text{ kg m}^{-3}$, which was obtained from the gravity dataset by using different optimization methods. The constructed map presents a dominant high Bouguer anomaly underlying Menengai caldera and Olrongai hill, associated with the presence of trachytic rocks and occurrence of fumarolic activities. A 3-D gravity inversion for predicting the geothermal heat source architecture within the Menengai caldera and Olrongai hill area is also presented. The resultant model provides an in-depth insight into the subsurface structure from potential field perspective by constraining deep bodies in 3-D space, which is significant for geological modeling in such a complex geothermal system. The model is then interpreted together with surface manifestations, geology, and geological structures. A lineament map was developed and used to enrich the geological structures map. The gravity model shows a close relationship between faults and inferred geometry of subsurface volcanic complexes displayed as discrete dike-like bodies in the near-surface environment and believed to form an integral part of the volcanic plumbing system. These structures act as conduits of magma to shallow levels, supplying the heat to the geothermal reservoir. A joint analysis and interpretation of 3-D density and 3-D resistivity inversion models of Menengai geothermal area was then conducted. The resulting gravity model reveals the presence of a structurally-controlled high-density body within the area, while the resistivity model shows the three main resistivity structures of a high-enthalpy geothermal system. These resistivity structures included a near-surface resistive layer associated with unaltered volcanic ($> 80 \text{ } \Omega\cdot\text{m}$) that is underlain by pockets of the conductive region related to the presence of clay minerals ($< 10 \text{ } \Omega\cdot\text{m}$). Finally, a resistive layer, which appears to be intersected intermittently by deeper dike-like conductive bodies in the caldera area while for Olrongai area the resistive layer increases gradually with depth. Integration of gravity data and other geophysical data show a good relationship between density and resistivity where regions with relatively low densities corresponded well with high resistivity area. For instance, a high-resistive body at Olrongai appears to mantle a high-density structure showing that the two have different formation properties. The presence of fumarolic activities and low resistivity in the region is attributed to the rise of volcanic fluids beneath the summit of the dense body. This decrease in resistivity may have resulted from the effect of acidic waters formed by the interaction of local groundwater with rising steam and gases of volcanic origin causing intense hydrothermal alteration of the volcanic rocks at the upflow zone and fades laterally away at the outflow. The anomalous $^3\text{He}/^4\text{He}$ isotopes from the geothermal

CHAPTER ONE

GENERAL INTRODUCTION

1.1. Background of the study

Kenya lies on the equator and overlies the East African Rift covering an area of 581,309 km² with diverse and expansive terrain that extends roughly from Lake Victoria to Lake Turkana and further south-east to the Indian Ocean. Uganda borders it to the west, South Sudan to the north-west, Ethiopia to the north, Somalia to the north-east and Tanzania to the south and south-west (Figure 1.1).

High-enthalpy geothermal fields in Kenya is confined to the Kenya Rift which forms part of the East Africa Rift, a narrow zone that is a developing divergent tectonic plate boundary, where the African Plate is in the process of splitting into two tectonic plates, the Somali Plate and the Nubian Plate, at a rate of 2–7 mm annually ([Fernandes et al., 2004](#)). At least 14 high-enthalpy prospects are already identified (see Figure 1.1), and three of which (Olkaria, Eburru, and Menengai) proven through deep drilling for geothermal production. These prospects associate with rifting-related volcanic activity in the rift floor. The geothermal areas with documented thermal manifestations outside the rift are Mwananyamala and Homa Hills, which are characterized by low- to medium-enthalpy resources are associated with the existence of carbonatite volcanic centers ([e.g., Tole, 1988, 1996](#)).

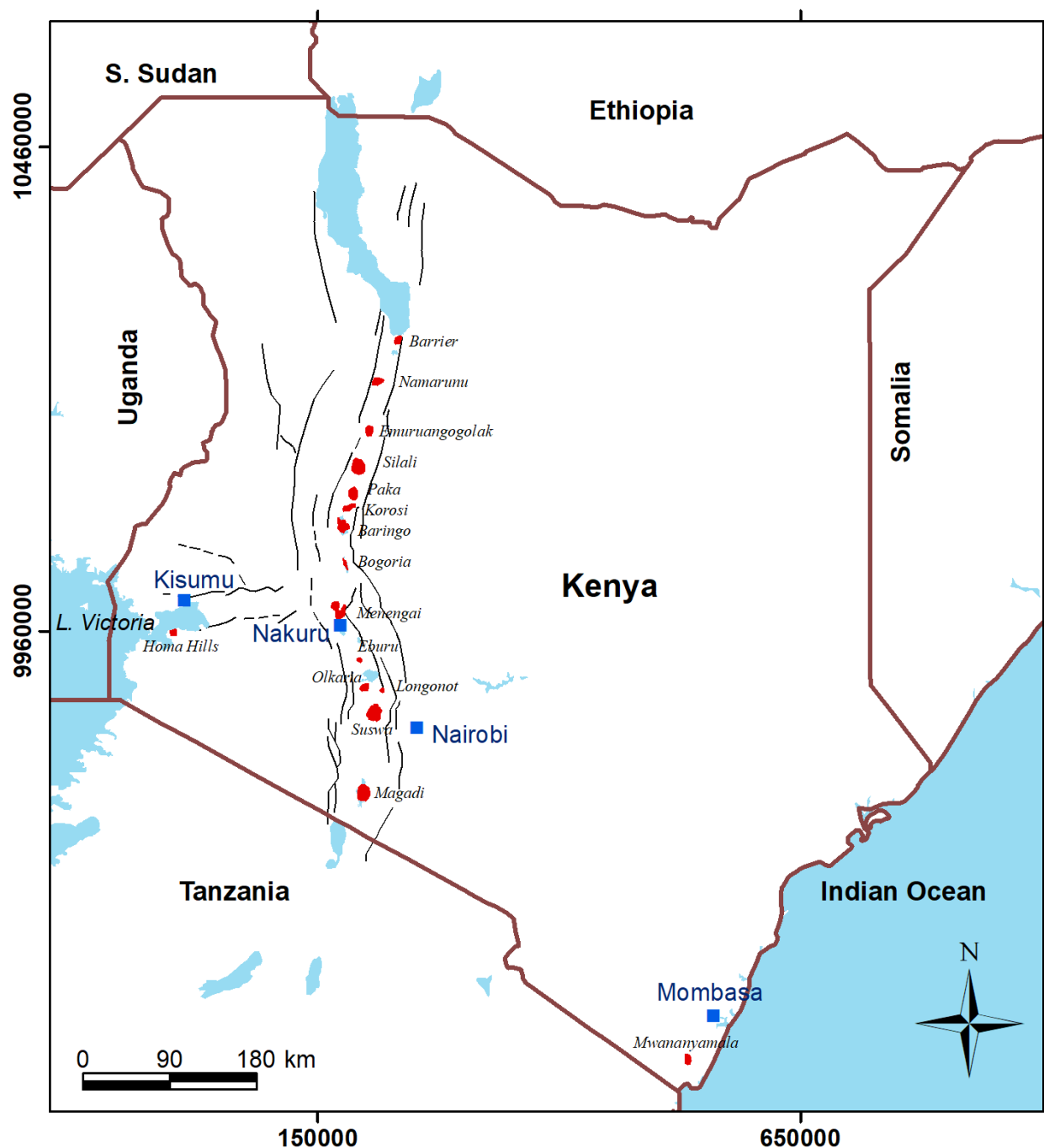


Figure 1.1: Map of Kenya showing the location of geothermal sites (in red) and the rift structures (black lines). Also indicated are water bodies (light blue polygons) and major cities.

The Menengai geothermal field is located in the central sector of the Kenya Rift Valley and is one of the high-temperature geothermal systems. There are two main geothermal sites within the

Menengai area; the Menengai caldera (hereafter simply referred to as ‘caldera’) and Olrongai. Deep drilling of geothermal wells in this field started in February 2011 and proved the existence of exploitable steam within the caldera. Presently, plans are underway to have the first 105 MWe of electricity connected to the national grid with production drilling ongoing. However, the geology is quite complex and geothermal development in the field has faced a technical setback where sited wells have shown unpredictable discharge behavior including varied forms of chemical scaling, cold fluid incursion, inconsistent discharge output, and high gas/steam ratio. These challenges have so far complicated planning for the development and management of the field. They are loosely attributed to intersected multiple aquifers and complex geology mainly caused by many eruption episodes. Initial estimates suggested that the field hosts ~ 1600 MWe worth of steam and this might not be readily feasible due to impending challenges already experienced in the field.

1.2. Geothermal status in Kenya

Kenya is endowed with quite enormous geothermal resources owing to its geographical position about the Kenya Rift system. Geothermal energy has relatively low electricity production costs making it attractive to investors. The current total geothermal installed capacity amounts to nearly 650 MW providing to almost 50% of total power generation in the country ([MoE Kenya, 2018](#)). Exploration for geothermal resources in Kenya shows that the Quaternary volcanic complexes of the Kenya Rift Valley offer the most promising prospects for geothermal exploration due to their geodynamic setting and high temperatures indicative of magmatic activity close to the surface producing a substantial geothermal anomaly. These thermally advantaged zones associate with local ‘highs’ of the hot asthenosphere roof, from where the preferential magmatic mass transfer towards the surface occurs ([Geotermica Italiana, 1987](#)). Such encouraging indications of the exploitable geothermal resources has so far intensified exploration studies on these areas.

Exploration for geothermal energy in Kenya started in the 1950’s with surface exploration that led to the drilling of geothermal wells at Olkaria, Eburru and Menengai geothermal fields. Currently, over 200 wells have been drilled in the Greater Olkaria geothermal field while six wells drilled in Eburru, and at least fifty at Menengai ([GDC, 2018](#); [Omenda & Simiyu, 2015](#)). Geothermal development is currently being fast-tracked in Kenya with drilling ongoing in Menengai, and

Olkaria geothermal fields with the country's geothermal potential estimate exceed 7,000 MWe (Simiyu, 2010).

The location of high enthalpy prospects is mainly within the Kenya Rift Valley where they associate closely with Quaternary volcanoes. Olkaria geothermal field is so far the leading producer with a current installed capacity of 630 MWe from five power plants owned by Kenya Electricity Generating Company (KenGen) and Orpower4. About 10 MW_t is being utilized to heat greenhouses and fumigate soils at the Oserian flower farm. The Oserian flower farm also has 4 MWe installed for own use. Power generation at the Eburru geothermal field stands at 2.5 MWe from a pilot plant. Production drilling for the additional 560 MWe power plants to be developed under the Public Private Partnership (PPP) arrangement between KenGen and private sector is ongoing (Omenda & Simiyu, 2015).

The Geothermal Development Company (GDC) is currently undertaking production drilling at the Menengai geothermal field, and the construction for the first 3×35 MWe power plants is underway (GDC, 2018). Over fifty deep wells of depths varying between 2,100 m and 3,200 m are now complete since exploration drilling started in 2011. Steam production from the wells varies from almost a megawatt up to ~30 MWe per well, and currently, over 130 MWe of steam equivalent is on the wellhead ready for full steam production. Reservoir temperatures of up to 400°C at 2,000 m have been encountered in at least four of these wells, making it the hottest proven geothermal system in Kenya. Detailed exploration is continuing in Suswa, Longonot, Baringo, Korosi, Paka, and Silali geothermal prospects and exploration drilling are expected to commence soon in Baringo–Silali geothermal areas.

1.3. Previous studies

A number of geoscientific studies have been so far carried out on the Kenya Rift region and a few of them on geophysics with focus on seismic, resistivity and gravimetric studies both in a regional scale (e.g. Bullard, 1936; Fairhead, 1976; Fairhead & Reeves, 1977; Girdler, 1983; Girdler & Sowerbutts, 1970; Searle, 1970; Simiyu & Keller, 1998, 2001; Sippel et al., 2017; Smith & Andrew, 1962) and more specific to geothermal resource of the study area (e.g., *Geotermica Italiana*, 1987;

Lagat et al., 2010; Mungania & Lagat, 2004; Simiyu, 2013; Wamalwa et al., 2013). Exceptional attention was drawn to an East Africa's salient oval-shaped negative Bouguer anomaly that covers the whole of east and west branches of the rift whose minimum is centred over the East Africa rift system (EARS) near the equator (within the study area) and a major axis that strikes NNE (e.g., Girdler, 1975, 1983; Girdler & Sowerbutts, 1970; Swain, 1992). The negative anomaly zone suggests an isostatic disequilibrium within the region (McCall, 1967), considered to be produced by the low density of the surface volcanics (Fairhead, 1976) or perceived asthenospheric intrusions (Chorowicz, 2005).

Geothermal activity in the study area is manifested by the occurrence of fumaroles, steaming and gas-discharging boreholes, hot-to-warm water boreholes, and altered grounds. These manifestations relate closely with the occurrence of the younger trachytic outcrops implying that this young formation confines thermal anomalies in the area. Gas geothermometers analysis suggest reservoir temperature of more than 250°C suggesting a possible geothermal resource underneath both Menengai caldera and Olrongai area (Lagat et al., 2010).

Geotermica Italiana (1987) conducted a reconnaissance survey in the study area with particular attention focused upon geological and geochemical features relevant to the identification of thermally anomalous zones and those hydrogeological and structural elements required for the existence of the high-enthalpy geothermal system. The geochemical survey revealed the existence of steam leakages and soil gas anomalies, which was evidence for the existence of high-temperature geothermal fluids at depth, within the study area. These studies gave promising results confirming the existence of exploitable geothermal resource, and this later laid the foundation for further exploration studies.

Mungania & Lagat (2004) later carried out a detailed geoscientific survey which included geophysical, geochemical, heat loss measurement and geological studies and came up with the initial conceptual model of the field. Their results echoed and updated the work done by Geotermica Italiana (1987) stating that there exists a shallow active magma based on the surface manifestation of fumarolic activity and the morphological buildup of lava pile dome inside Menengai caldera. They also noted that local volcano-tectonic axis caused by extensional tectonics allows upflow and

controls the lateral flow of geothermal fluids within the reservoir. Geophysical studies revealed a dense body that lies within the depth of 10 km below the surface and also resistivity structure typical of a geothermal system. Temperature estimation using the gas geothermometers, indicated that the reservoir temperatures beneath Menengai caldera are $>250^{\circ}\text{C}$ taking both H_2 and H_2S gases into account. The area gave anomalously high ^{222}Rn counts for both the soil gas and in the fumarole steam, which would infer good permeability and a possible high-temperature geothermal reservoir. This high ^{222}Rn anomaly coincides with high CO_2 values in this area. The $^{222}\text{Rn} / \text{CO}_2$ ratios were also high implying that the CO_2 emanating from this zone could be of magmatic origin. Surface heat loss studies conducted in the Menengai prospect indicated that a natural heat lost in the area is over 3536 MWt. Out of this amount, the caldera floor releases about 2690 MWt while the fumaroles lose totals to about 2440 MWt by convection. This substantial heat loss could be an indicator of a massive heat source underneath the prospect.

Lagat et al. (2010) conducted further infill studies and updated previous model inferring the depth and location of heat sources, delineating the reservoir and recharge zones, as well as the reservoir capping system. They recommended drilling of at least four deep directional exploratory wells in the caldera, most of which later proved productive, as was initially proposed by Mungania & Lagat (2004). They also proposed exploration wells for the Olrongai area.

Simiyu (2013) mapped the spatial seismic intensity, hypocenter distribution, event magnitude, depth of clusters with time and shear wave attenuating structures, and observed that a large magma body underlies the study area and extending southwards to Menengai caldera. Shear wave attenuation and the brittle-ductile transition depth suggests that the molten material is at a very shallow depth of about 6 to 8 km, at which any new injection into the magma chamber from the mantle will lead to volcanic eruptions. Velocity ratio analysis shows that this is a quite resilient heat source, which is unlikely to be cooling.

1.4. Purpose and plan of study

The purpose of this work is to describe the characteristics of the Menengai geothermal system intended to improve on well siting strategies which will lead to better targeting of producing zones.

The study attempts to establish essential sources of fluids feeding the geothermal system, and also, looks at their movement and interaction with the lithology. It also attempts to delineate the heat sources and reservoirs of Menengai geothermal system.

This objective was achieved by applying an integrated approach involving geophysical, geological, hydrological and geochemical techniques as well as reservoir properties (Figure 1.2). The geology and structures were reviewed, and the lineaments were extracted from satellite images to corroborate the understanding of local structures. Subsurface density distribution study using gravity data was carried out to help infer the extent and geometry of heat sources. Resistivity study was used to help refine the geometry of heat source and also define the extent of the geothermal reservoir. Recharge mechanisms were inferred based on stable isotope geochemistry and piezometric information while magmatic contribution was assessed using Helium isotope ratios. Finally, reactive geochemical modeling was then conducted to determine the geochemical processes occurring in the reservoir.

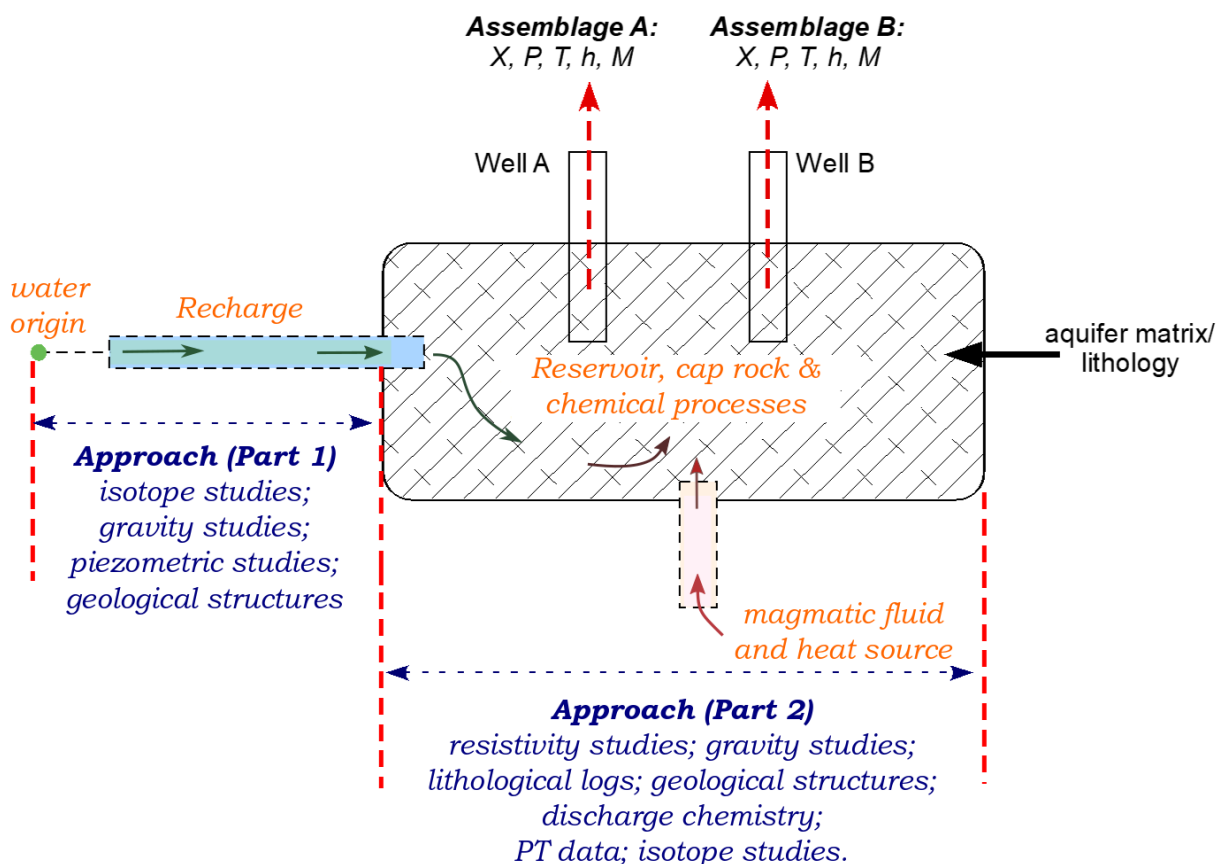


Figure 1.2: Schematic diagram that shows the stages and targets of the present studies. The properties of wells are represented by x, P, T, h and M which represents chemical composition, Pressure, Temperature, enthalpy, and mass flow, respectively.

1.5. Structure of the dissertation

This dissertation consists of 7 chapters, which describe the research carried out on the integration of geology, geochemical, geophysical and hydrological studies for the assessment of mechanism controlling the occurrence and extent of geothermal resource of the Menengai geothermal area.

Chapter 1: This chapter describes the general background of the study as well as the general information about the geothermal status in Kenya as well as documented studies that were conducted by preceding researchers. The history and current status of geothermal development in Kenya, with a particular focus on the Menengai geothermal area, is given. Likewise, a description of the purpose, the plan and the structure of the thesis are also made.

Chapter 2: This chapter elaborates on the various characteristics that define the Menengai geothermal area. The chapter starts with a detailed review of the geology and structural setting of the area from both local and regional viewpoints. Then a description of the methodology used in the study of lineaments from Landsat images is presented. The description includes the extraction of lineaments from the satellite image, their analysis, and interpretation, and presentation of the results using a density map and rose diagrams. The last section reviews the hydrological constraints, both local and regional, and suggests possible sources of the fluid recharging the geothermal system.

Chapter 3: This chapter describes the gravity study. A comprehensive review of the gravity method both theory and applications is given then a description of gravity data and how the survey was conducted in the field. Besides, some corrections were done in the field, and others applied later to the data are also discussed. Bouguer anomaly map is presented and the methods used to estimate the Bouguer density are discussed. The 3-D inversion technique used is described, and the resulting model is presented and discussed in the last part of this chapter.

Chapter 4: This chapter gives subsurface resistivity structure of the study area using the MT method. It gives a review of the MT method, its application in geothermal exploration and how measurements were done in the field. The data acquired suffered from distortion, and static shift and corrective measures were applied. Dimensionality analysis was carried out to determine the nature of the subsurface structures. MT inversion modeling (1-D, 2-D, and 3-D) was conducted, and the inversion results were interpreted and discussed in the last part of this chapter.

Chapter 5: This chapter discusses the various geochemical methods that were considered and deemed essential for the present study. It begins by describing the theoretical background that forms the basis of this study, which includes isotope geochemistry and reactive transport modeling. A brief review is given on the geochemistry in geothermal exploration as well as the physico-chemical features of the geothermal system. The results of major chemical elements and isotope analysis are presented and discussed. Finally, the output of the reactive transport simulation is interpreted and discussed.

Chapter 6: This chapter brings together the results from previous chapters and presents an integrated interpretation. The discussion in this chapter is divided into two parts; the Menengai caldera and Olrongai area.

Chapter 7: This chapter presents the conclusions of the study, and this includes a summary of the conclusions made in preceding chapters.

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CHAPTER TWO

FIELD CHARACTERISTICS OF MENENGAI GEOTHERMAL AREA

2.1. Introduction

The regional geology and structural setting of the study area have attracted much researches, and some review papers have been written. This chapter describes the geology and structural makeup of the Menengai area both regional and local. It also looks at the hydrological setting of the area and proposes possible sources of fluid recharging Menengai geothermal system based on known hydrological constraints.

2.2. Geology and structural setting

The general geological background of the study area has been incorporated both in the quantitative and qualitative interpretation phase of the geochemical and geophysical results in the present study. A clear understanding of the regional geology of the central Kenyan rift and the adjacent areas, therefore, is required. To define the spatial extent of the major rock units, four geological maps with a scale of 1:125,000 from the geological reports of Nakuru and Thomson's Falls – Lake Hannington areas ([McCall, 1967](#)), Kabarnet – Eldama Ravine area ([Walsh, 1969](#)), and Molo area ([Jennings, 1971](#)) were used to construct a geology map of the study area.

2.2.1. Regional setting

The East African Rift System (EARS) has long been recognized as a typical illustration of a continental rifting ([Chorowicz, 2005](#)), which forms part of the largest Tertiary–Quaternary rift system that extends from Syria, Jordan, and Israel in the north to Mozambique in the south. The

Kenyan portion is characterized by extension tectonism where the E-W tensional forces resulted in block faulting, which includes leant blocks, evidently seen in both the floor and rift scarps. The rift trough is truncated by numerous normal faults, which represent persistent and wide-ranging tectonism under the rift floor. The geometry and sizes of well preserved erosional surfaces of Miocene as well of Late Pliocene ages give evidence for three central phases of epeirogeny and a crustal flexure that consist of episodic up-arching of central Kenya and local down-warping of the coastal zone (e.g., Baker & Wohlenberg, 1971; Saggerson & Baker, 1965). Like the rest of the EARS, it has undergone a very complex geological evolution and tectonic movements.

The Menengai geothermal field is located in the central segment of the Kenya rift and is hosted partly within a major Quaternary volcano with Menengai caldera being central for its geothermal resource and conspicuously the major geological feature in the area (see Figure 2.1). This area is characterized by a complex geological setting associated with uplifted region the so-called Kenya dome, a geographical upwelling province of anomalously high elevations situated at the confluence of structures that resulted from major tectonic events and is covered for the most part by Cenozoic volcanic rocks (e.g., Corti, 2011; Davis & Slack, 2002; Leat, 1991; Savage & Long, 1985). This elliptical-shaped hoisted Kenya dome is a local culmination on the eastern rim of the East African plateau (e.g., Baker & Wohlenberg, 1971) conspicuously marked by three major central volcanoes; Mounts Kenya and Elgon in Kenya, and Mount Kilimanjaro in the Tanzania side.

Volcanisms in the Kenya rift occurred nearly relentlessly from Early Miocene to Holocene times, with succession in the central part of the rift consisting of Pliocene and Pleistocene trachytes, with subordinate basalts, overlying Miocene phonolites which are in turn resting on the Precambrian metamorphic basement (e.g., Leat, 1984; Williams, 1978). The domal uplift in the central rift during the Late Miocene led to massive fissure phonolitic eruptions and emplacement (Baker et al., 1971), locally referred to as Rumuruti phonolite (McCall, 1967). Pliocene phonolitic trachytes outcrop in the Olrongoi area, whereas younger trachytic flows cover the caldera floor, and both are associated with the occurrence of fumarolic activities. However, a more substantial part of the study area is covered by volcanic soils and sediments.

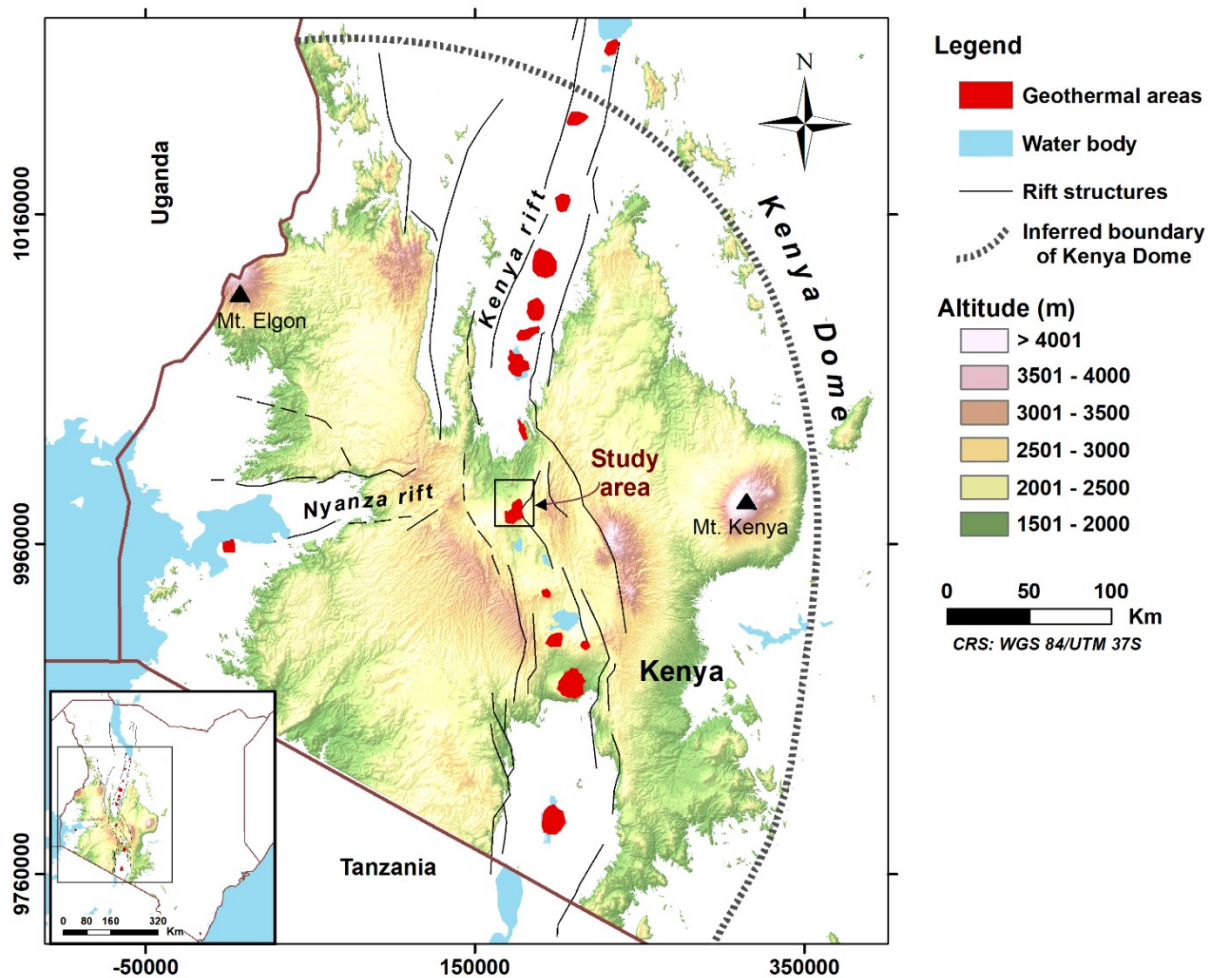


Figure 2.1: Location map of the study area. The high altitude (>1500 m above sea level) area shows part of the Kenya dome.

2.2.2. Local geology

The caldera volcanoes of the Kenyan rift are all to some extent unique, despite their close geographic proximities. Their variation is evidenced by their structural setting, the geological makeup, the interplay of their petrogenetic processes, and the mechanism of caldera collapse. All these factors encompass many of the features of peralkaline silicic systems globally (Macdonald, 2012). At Menengai, two episodes of caldera formation of Krakatoan-type were accompanied by an eruption of a major ignimbrite and associated pumice and ash falls (Leat, 1984).

Menengai volcano has been active between about 0.4 and 0.3 Ma B.P (Gislason, 1989). According to Leat et al. (1984), the formation of shield volcano began about 200,000 years ago and then was

followed by the eruption of two large ash-flow tuffs, each preceded by significant pumice falls. More than 70 post-caldera lava flows cover the caldera floor, the youngest of which may be only a few hundred years old. The caldera of Menengai volcano lies immediately north of Lake Nakuru, but ignimbrites and air-fall tuffs from the volcano cover some 1350 km².

The volcano is thought to be underlain by a voluminous magma chamber, with a 12 x 8 km caldera which has almost vertical, embayed walls up to 300 m high (Leat, 1984), the rocks either composed of peralkaline or metaluminous trachyte, or pantellerite (McCall, 1967). Pantellerite is an iron-rich, peralkaline, silica-oversaturated volcanic rock that occurs most frequently as the felsic-end member of some bimodal suites in continental rifts and oceanic island settings (White et al., 2009). Pantellerite originates probably from trachyte through crystal fractionation of alkali feldspar-rich assemblage at a reasonably low pressures and oxygen fugacities at or below the fayalite–magnetite–quartz (FMQ) buffer, although the origin of trachyte remains unclear (Avanzinelli et al., 2004; Civetta et al., 1998; White et al., 2005, 2009).

The Menengai caldera is distinguished by the existence of a graben confined within a fault segment, and its partial collapse features suggest that portions of adjoining caldera wall also represent surface expressions of an embayed ring-fracture (see Figure 2.2). Such faults are not common in the Kenya rift valley, which is a tectonic province typically described by numerous parallel normal faults (e.g., Fairhead et al., 1972; Leat, 1984; McCall, 1967). Leat (1984) observed that the ring-fault architecture must have been controlled by a local crustal anomaly at Menengai and is well explained by causing collapse into a magma chamber, and the voluminous ash-flow tuff as a single flow unit suggests that the magma must have been stored, prior to eruption, in a vast, shallow reservoir.

The occurrence of fumarolic activities is associated with the presence of fault structures that provide a means of access for hydrothermal fluids to reach the surface. In the same way, diffuse soil degassing study conducted shows that Carbon Dioxide (CO₂) and Radon 220 emanating from faults in Menengai were measured to be anomalously higher than background values, which was taken to insinuate existence of deep-seated crustal faults that tap and channel magmatic CO₂ to the

surface where it is released through the soil (Kanda, 2010b). These studies are part of more considerable attempts to understand the regional lithospheric structure of the EARS.

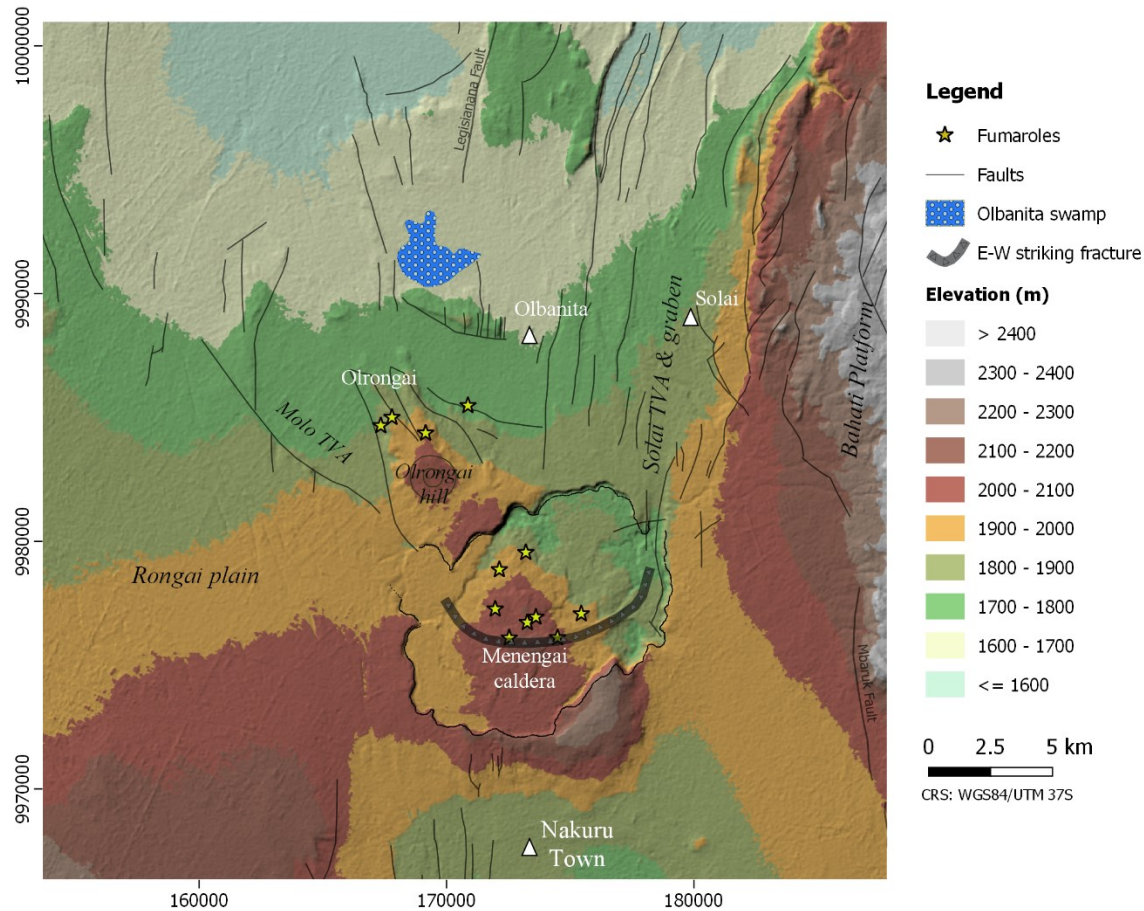


Figure 2.2: Descriptive map of the study area showing the elevation (m), faults, fracture and location of fumaroles.

The local surface geology, shown in Figure 2.3, is composed mainly of unconsolidated low-density volcanic soil and sediments that cover a total surface area of 876 km² (72% of the total area), and tuffs & ignimbrites covering 160 km² (13% of the total area). The higher density formations consist of phonolites, trachytes, and basalts with a combined surficial area of 15%.

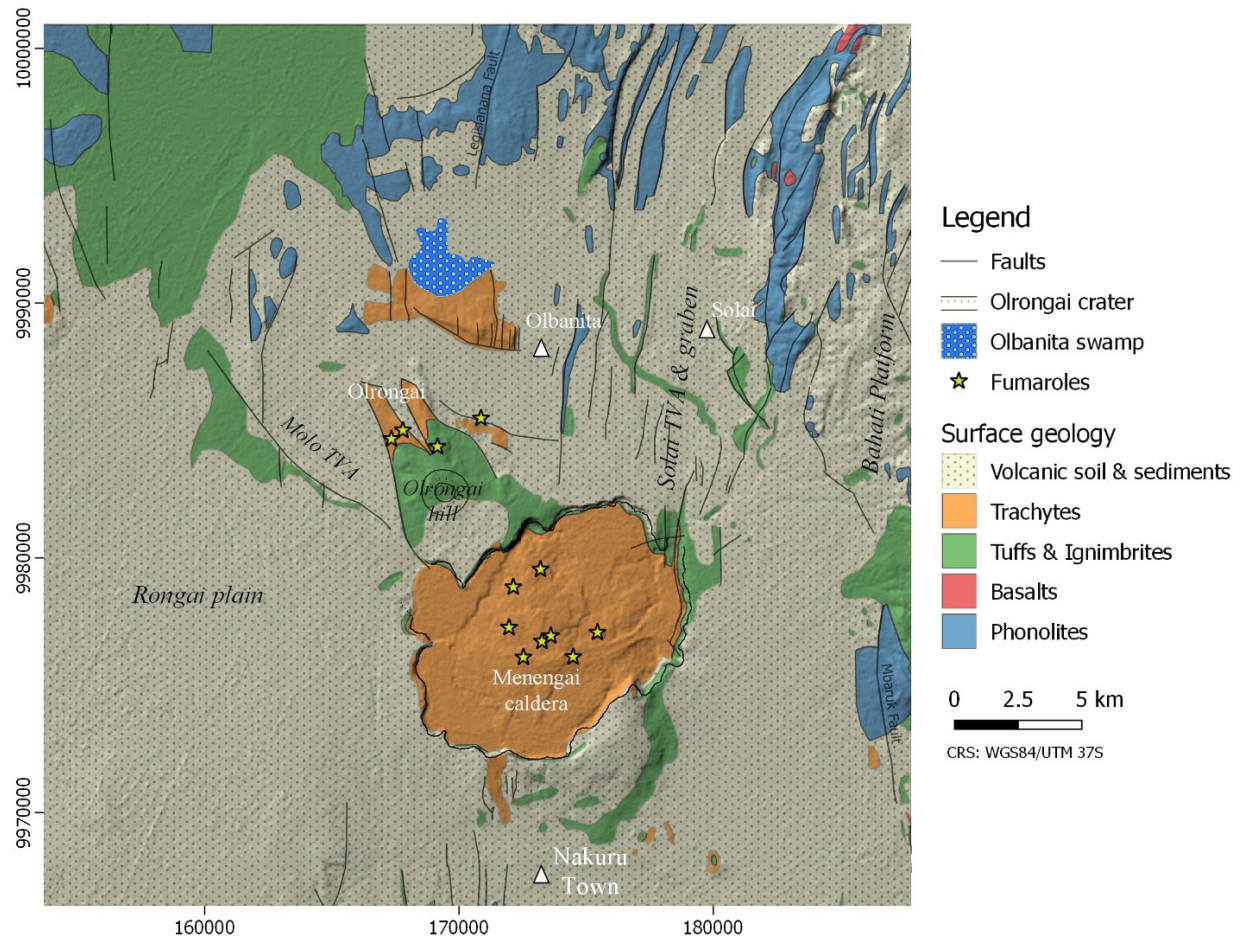


Figure 2.3: Surface geology map of the study area, modified from previous work done by [McCall \(1967\)](#), [Jennings \(1971\)](#) and [Walsh \(1969\)](#).

Recent lithostratigraphical studies penetrated by about 50 geothermal wells within the caldera with varying depths of up to 3 km have shown that the rock formation is predominantly composed of trachyte, tuff, and pyroclastics. However, erratic lenses of syenitic intrusives and basalt have also been encountered in some of the wells. These syenite intrusives were intercepted intermittently in some wells, suggesting that magma pulses from the main magma chamber were injected into the overlying formations in the form of dikes and other intrusives. At least four of these wells that recorded bottom-hole temperature close to 400°C at depths of about 2.1 km, and presence of fresh glass encountered by some wells indicate a possible encounter with magmatic intrusive. This inference is supported by drilling challenges that were experienced at these depths during offloading, the torque, weight on bit, pressure and temperature increased substantially.

Menengai Geothermal Prospect itself is located within an area characterized by an intricate tectonic activity linked with the rift triple junction. The triple junction is a zone at which the failed rift arm of the Nyanza rift joins the central Kenyan rift. Two rift floor volcano-tectonic axes that are important in controlling the geothermal system in the study area are the Molo and the Solai TVA. The main trend of the tectonic structures in the study area is the same as that of the rest of Kenyan Rift, which is mainly of N-S, NE-SW, and NW-SW, predominantly lying to the north of the caldera (see Figure 2.3).

2.2.3. Lineament studies

A quantitative study of lineaments was carried out to determine the length and directional patterns of lineament sets about geological structures. In this study, a band combination of cloud-free Landsat 7 Enhanced Thematic Mapper Plus (ETM+) of path/row 169/60 acquired on May 10, 2016, was integrated with the geological map and analyzed alongside a shaded relief map created from Digital Elevation Model (DEM). The DEM was extracted from Shuttle Radar Topography Mission (SRTM) data retrieved from the online Data Pool, courtesy of the NASA Land Processes Distributed Active Archive Center (LP DAAC), USGS/Earth Resources Observation and Science (EROS) Center. The preprocessing of Landsat image included conversion from a calibrated digital numbers (DNs) to physical units, such as sensor radiance and surface reflectance (SR), using Landsat calibration and FLAASH (Fast Line-of-Sight Atmospheric Analysis of Spectral Hypercubes) algorithm of ENVI software (e.g., Gawad et al., 2016; Matthew et al., 2002).

The edge detection algorithm of Canny (1986) was used to retrieve the lineaments. This algorithm is famously known for its suitability based on the three criteria of proper localization, good detection and single response to an edge (e.g., Ding & Goshtasby, 2001; Heath et al., 1996). The Canny algorithm uses a filter based on the first derivative of a Gaussian for the reason that it is prone to noise present on the unprocessed image data (Marghany & Hashim, 2010). The operation of retrieving lineaments was carried out using arguably the most widely used software for the automatic lineament extraction, the LINE module of the PCI Geomatica (PCI Geomatics, 2015). The linear density of lineaments was then calculated as their total length per unit of the area using Line Density tool built in ESRI ArcGIS © software which computes the density (in units of length per unit of area) of linear features in the neighborhood of each output raster cell.

Spatial properties of lineaments related to geological displacement have played an essential role in tectonic studies for the delineation of structural units as well as in identification of geological boundaries (e.g., Abebe et al., 1993; Chandrasiri Ekneligoda & Henkel, 2010; Hashim et al., 2013; Kageyama et al., 2000; Marghany & Hashim, 2010; Mountrakis & Luo, 2011; Noltimier et al., 1998; Saadi et al., 2011; Salati et al., 2011). Analysis of these structures (fractures, faults, dikes, and lithological contacts), from remotely sensed Satellite data is widely used as the basis of information for geologists to map lineaments at local and regional scales and provides useful information for energy and mineral exploration.

Lineaments in the present work are analyzed using density maps and rose diagrams, which are some of the conventional methods generally used (e.g., Hung et al., 2005; Karnieli et al., 1996; Zakir et al., 1999). In this study area, a total of 8,838 lineaments (with length ranging from 60 m to 24,338 m) were extracted with lengths, comprising of both natural features owing to geology-related structures and artificial lineaments that arise due to human-made features such as roads, farm boundaries, among others. Lineaments considered as products of human activities were excluded by use of DEM map with multiple sun azimuth directions in addition to field observations.

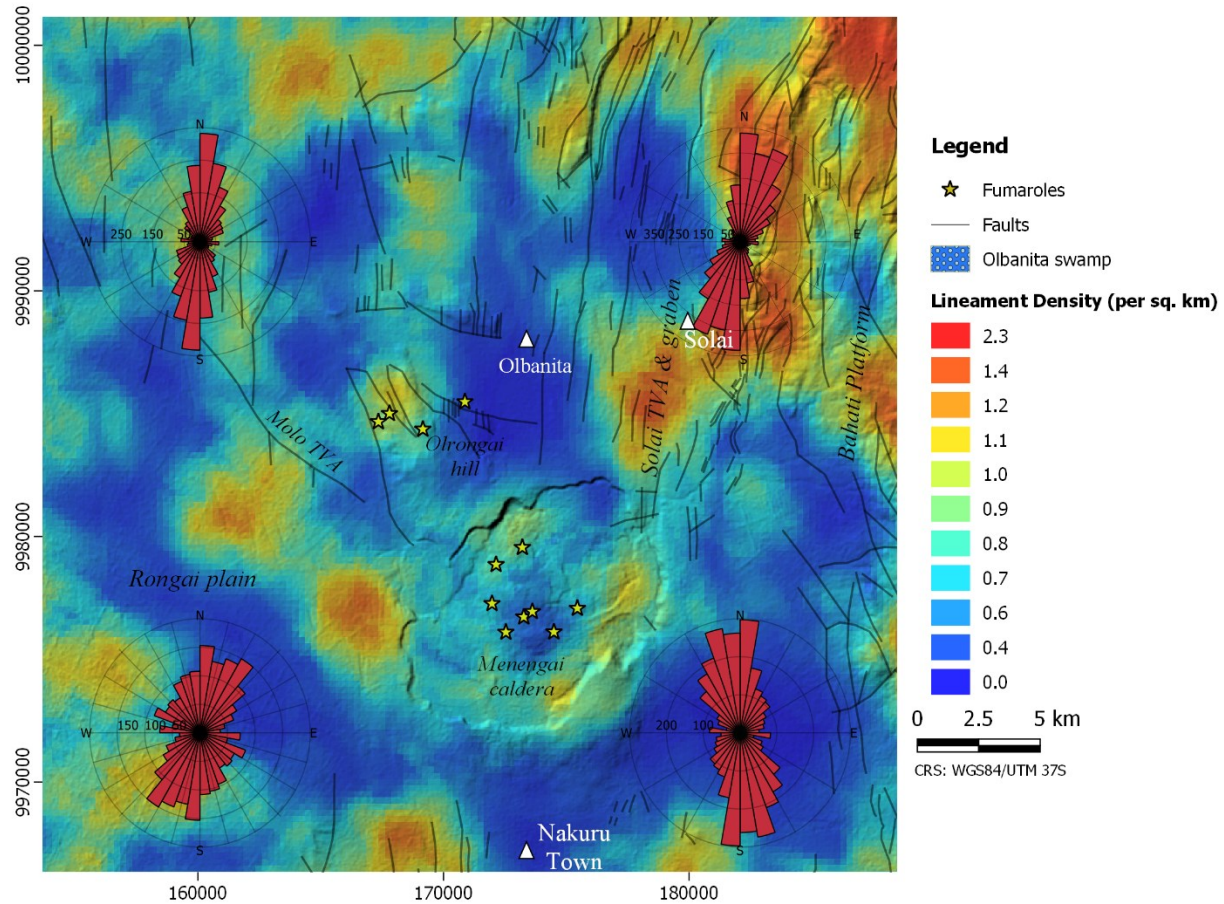


Figure 2.4: Lineament density map of the study area per km². The Rose diagrams show the azimuths direction of the local distribution of lineaments, with radial coordinate indicating the number of lineament.

Lineament density was calculated based on the number of lineaments per unit area, with the unit area taken as a square kilometer and the resulting map was then superimposed over the DEM map for easy interpretation. Figure 2.4 shows a quantitative study of lineaments conducted to determine the length and directional patterns of lineament sets. In the present study, the density was used to find the link between the distribution of existing geological faults and the concentration of lineaments within the study area. Density distribution in the study of lineaments has been widely used as a tool in the delineation of fracture zones based on lineament indices extracted from satellite imagery (e.g., Hung et al., 2005), and it reveals the distribution of lineaments per unit area. The geological map was used to corroborate the density concentrations and known lithological units in the study area.

The lineament density map (Figure 2.4) shows that the high-density values are concentrated mostly in areas occupied by phonolites and trachytes and more so to the NE area where these rocks are reasonably well exposed, while most of the faults are located adjacent to zones with average to high-density values (see Figure 2.3). On the contrary, the results obtained from terrains masked by volcanic soil and sediments, as one might have expected, show low to moderate density values.

Analysis of the orientation of structures completed using rose diagrams helps in ascertaining the most common directions of the lineaments, which then can be compared with directions of existing faults. The study area was subdivided into four equal zones (namely NE, SE, SW, and NW) for the sake of interpretation, based on the general orientations of mapped faults. With an angular spacing of 10°, rose diagrams shown in Figure 2.4 indicate that the most prevailing directions of lineaments extracted from the satellite image are N 0° - N 30°, N 330° - N 10°, N 0° - N 40° and N 350° - N 20°, in the NE, SE, SW and NW zones, respectively. This angular tendency corresponds pretty much to that of the faults, and despite lack of faults in the SW zone, the general trend from lineaments matched with the regional trend of faults, and this underscores the importance of lineaments in the mapping of geological structures in inaccessible areas.

2.3. Hydrological setting

Hydrological constraints of Menengai geothermal system have been examined using some set of data. The average annual rainfall data is from 2225 rain gauge stations fairly distributed within the country. Digital Elevation Map (DEM) for relief studies was extracted using ASTER data. Structural maps were studied using geological maps ([McCall, 1967](#)) and lineaments processed from Landsat imageries.

2.3.1. Regional Relief and Rainfall Distribution

Kenya lies astride the equator confined by latitudes and longitudes from 5.5°N to 5°S and from 34°E to 42°E, and therefore enjoys a tropical climate. It is temperate inland, hot and humid at the coast, and very dry in the north and northeast parts of the country. The seasonal movement of the Inter-Tropical Convergence Zone (ITCZ) and the coinciding precipitation pattern strongly influence the regional climate ([Nicholson, 1996](#)). Rainfall associated with the transition of the

ITCZ follows the highest sun in March and September with a lag of about 3 to 4 weeks. Therefore, the field of study receives most of its precipitation during the “long rains” in April to May and the “short rains” in November. A third lesser precipitation maximum occurs in September when the Congo air masses transport moisture from the Congo basin to East Africa (Vincent et al., 1979). The amount of precipitation is strongly linked to topography. Thus, the highest rates occur in the high altitude parts of the basin in the west and east, whereas the lower areas are relatively dry.

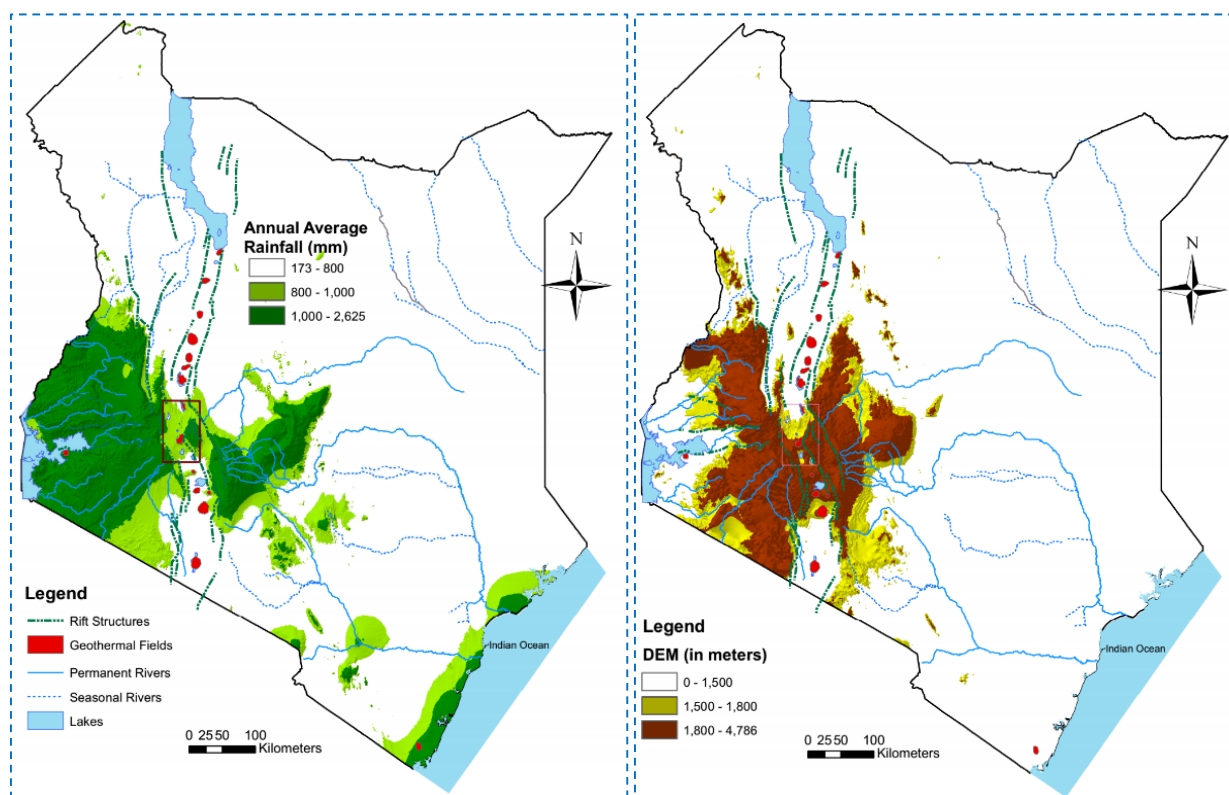


Figure 2.5: Annual average rainfall and relief map of Kenya.

The long-term hydrologic budget of the area is primarily controlled by the spatial and temporal distribution of rainfall, superimposed on tectonically-driven influences on the drainage network and basin geometry (Strecker et al., 1990). Kenya’s complex topography, e.g. large lakes, Great Rift Valley and snow-capped mountains, dramatically influences circulation and alters the climate of various parts of the country. As shown in Figure 2.5, Kenya’s rainfall distribution is mostly highest in areas of higher relief associated with the formation of the Great Rift Valley, mainly the upwelling of the Kenyan Dome (see Figure 2.1). The other significant aspect in controlling rainfall

distribution is the effect of two large water bodies; Indian Ocean and L. Victoria, which lies to the east and west of the country, respectively.

2.3.2. Regional hydrology

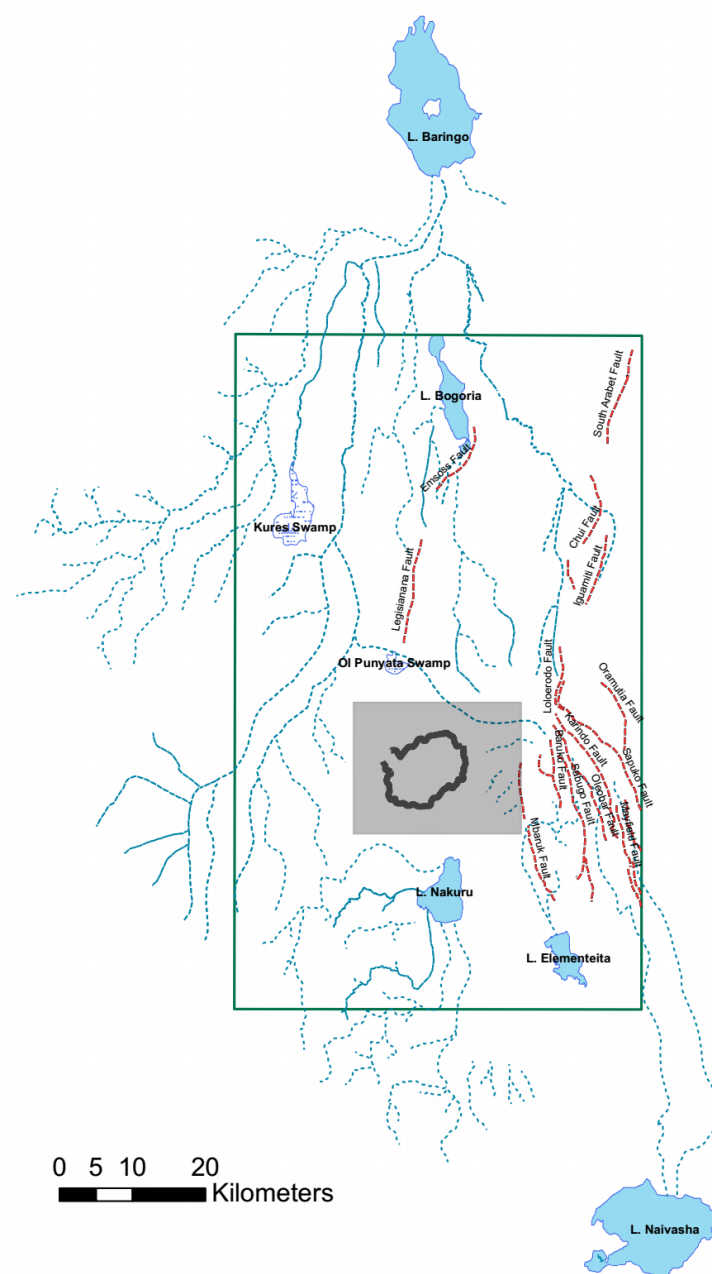


Figure 2.6: Main geological structures and local hydrology. The study area is shown in grey, and the green line is considered as an area of interest from a hydrological perspective. Major faults, perennial rivers and ephemeral rivers are indicated as broken red lines, blue lines and broken blue lines, respectively.

Most of the lakes in Kenya are located within the Kenya rift valley and derive most of the water from springs, perennial rivers, and ephemeral streams during rainy seasons (see Figure 2.6). Lake Naivasha and Lake Baringo are the only freshwater lakes which suggest the presence of underground drainage since there is no known surface outflow.

2.3.3. Hydrological models

This section examines six possible sources of fluid recharging Menengai geothermal system by discussing their hydrological constraints observed in the preceding sections. These concepts are graphically demonstrated in Figure 2.8

Concept 1: Shallow groundwater and runoff

Menengai area lies on the rift floor that gently slopes northwards. To the east, the scarps bound the eastern inner-rift trough and are relatively higher and wetter grounds. The prevalence of rainfall particularly towards the eastern flanks of the rift scarp poses a higher likelihood of recharge source for the Menengai geothermal system. To the west, the rift floor slopes gently into a series of ridges connecting hills. The surface drainage system is therefore mostly internal from the east and the western scarps (Kapija, 2008).

The drainage on the rift floor is mainly from Menengai Caldera northwards except the drainage from the southern slopes of Menengai Caldera into Lake Nakuru. The caldera springs/streams, which occur at the NE foot of the caldera confirm that the southern part of Solai tectonic axis is essential in restraining groundwater movement. The occurrence of cold springs in the northern part of L. Nakuru tends to support a possible near-surface flow of water that recharges a deeper reservoir upon intercepting the caldera rim. The concept is supported by the presence of perennial caldera streams and some highly productive boreholes drilled at the northwesterly edge of the caldera (see Figure 2.7).

Concept 2: Deep westerly groundwater flow

The deep-seated faults of the rainfall-prone eastern scarp act as conduits of water that most likely finds its way to the caldera, ultimately recharging the geothermal reservoir. The presence of N-S

truncated by NE-NW and NW-SW trending faults/fractures system provides underground channels from the wetter eastern area. This has resulted in the seepage water to the subsurface environment, and as a result the steady flows in some streams have been disrupted at some points.

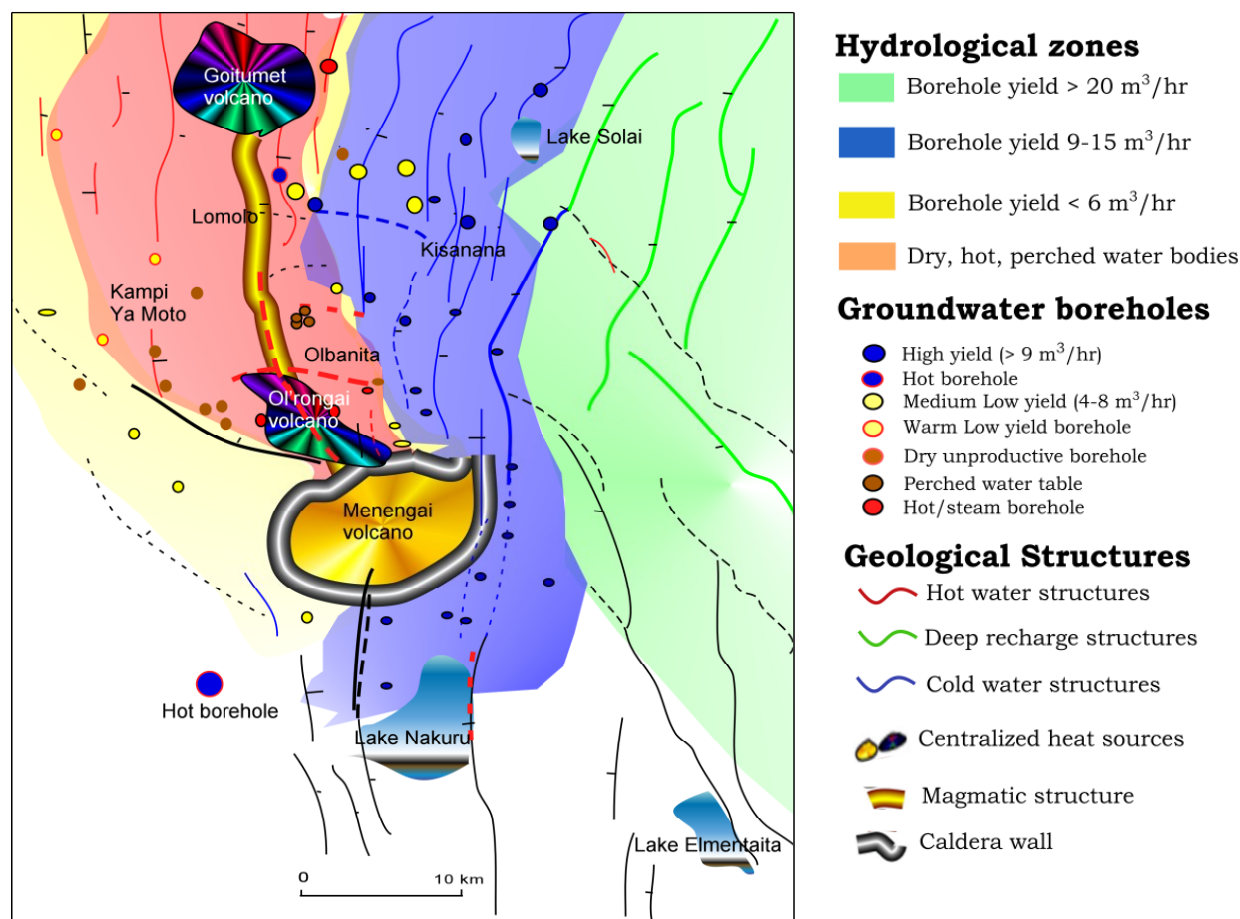


Figure 2.7: Hydrogeological controls of the Menengai geothermal area (adapted from [Mungania & Lagat, 2004](#)). Boreholes are represented by circles.

About five water boreholes drilled to an average depth of 220 m supply water to the project, with an average productivity of the boreholes of $12 \text{ m}^3/\text{hr}$, which is currently adequate for all drilling operations. These high productive boreholes are typical of most of the other boreholes drilled to the east of the caldera relative to the boreholes to the west, which are characterized with less productive and deeper aquifers ([Kanda, 2010a](#)). These boreholes insinuate a potential deep-seated circulation of groundwater feeding the geothermal reservoir.

Concept 3: Near-surface flow from L. Nakuru

At an elevation of about 1,760 m above sea level (asl), Lake Nakuru is one of the highest lakes in the Kenya dome of the central rift valley, and its high elevation has hydrological effects on regional water flow pattern. Unlike other low-lying Kenya rift lakes that have a plentiful water supply through a series of hot and freshwater springs, there is nominal underground inflow into Lake Nakuru through the axial fault system (Muhandiki & Kulecho, 2006). The lake's hydrology is reliant on on catchment supply from rivers, demanding catchment reliability in order to sustain the lake water level. Previous geochemical and isotopic studies suggest that Menengai reservoir waters (situated at elevation of < 1,100 m asl) seem to result from the mixing of local shallow groundwater with waters of similar composition with that of Lake Nakuru (Kanda & Suwai, 2013; Sekento, 2012).

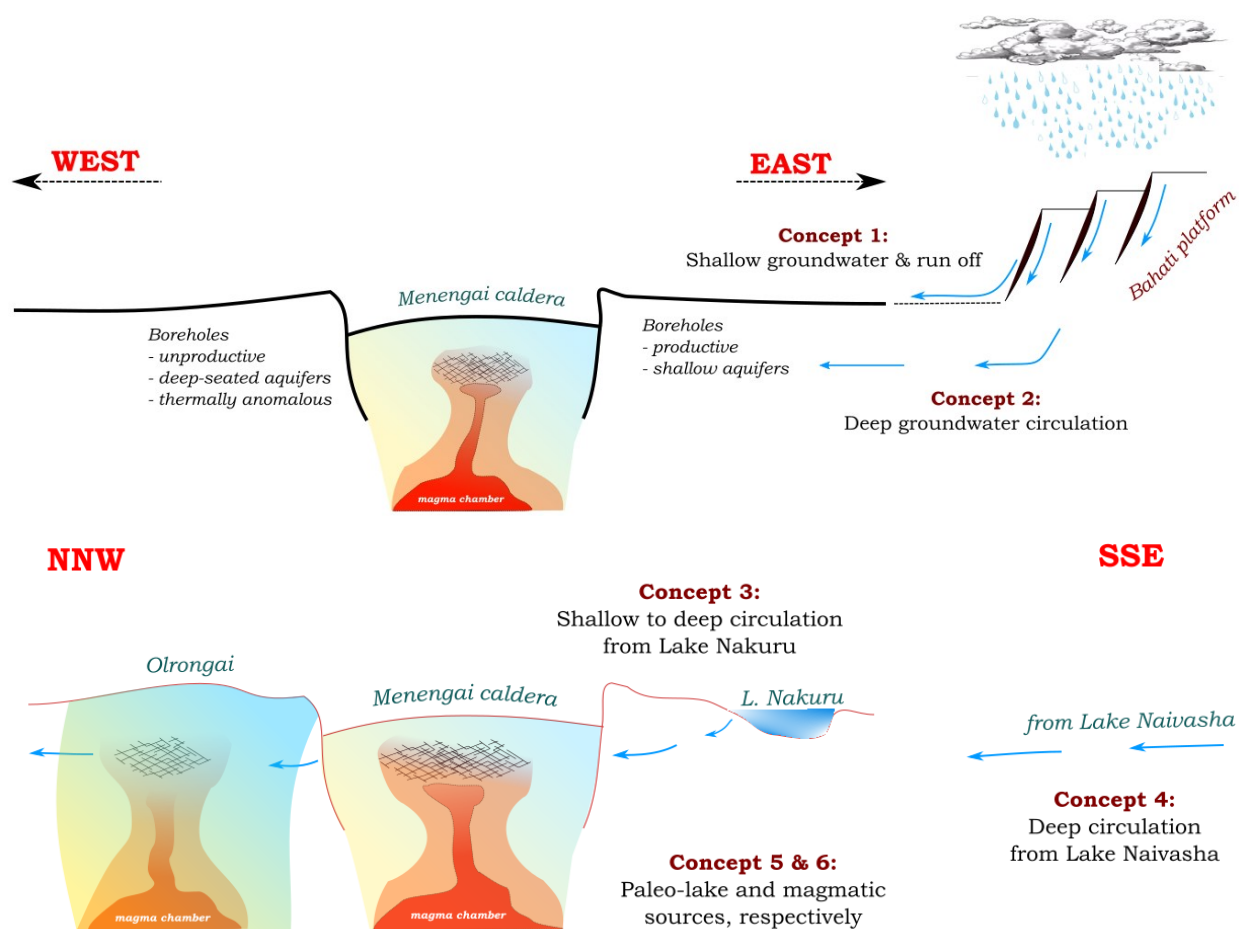


Figure 2.8: Hydrogeological model of the study area indicating possible fluid recharge sources of the geothermal system.

Concept 4: Deep northerly flow from L. Naivasha

Lake Naivasha sits on a higher altitude (1,884 m) compared to the rest of rift valley lakes. It is a freshwater lake despite having no surface outflow suggesting an underground outflow. Structural setting supports a northward flow of water and possibly recharging Menengai geothermal field, which also has been discussed by some previous authors (e.g., [Becht et al., 2006](#)).

Concept 5: Presence of paleolake

The caldera floor is covered by rock formations and cinder cone structures of the post-caldera stage of Menengai volcanic evolution where $\sim 23 \text{ km}^3$ of post-caldera volcanics erupted. The other common lithological unit is the lake sediments in the northeast portion of the caldera. The intra-caldera lake sediments prove the existence of a lake which flooded the NE part of the caldera, through which a prominent phreatomagmatic deposit, the Ruplax Tuff, was erupted ([Leat et al., 1984](#); [Mibei, 2012](#)). This formation is an evidence of a paleolake presumed to be from 8,000-10,000 yrs. BP as fronted by [Leat \(1984\)](#) and could have formed minor components of the recharge to the Menengai geothermal system. The lake was formed as a result of the subterranean outflow of Lake Nakuru through structures and faults related to the Solai system. The diatomaceous deposits associated with this paleolake are overlain by post-caldera volcanics ([Mibei, 2012](#); [Omondi, 2011](#)). The likely remnant of paleowaters overlaying the basement rock, trapped within the caldera formation could be a possible source of geothermal reservoir water.

Concept 6: Magmatic Contributions

The relative abundance of gases in geothermal fluids will depend both on the fluid type (dilute/saline) and temperature ([Arnorsson, 1986](#); [Kharaka, 1994](#)). These gases can originate in the crust, mantle, and atmosphere. In magmatic systems, the source of the reactive components is believed to be solidifying magma bodies. The proportions of these gases may differ from system to system or even in different parts of the same system. As [Kipng'ok et al. \(2014\)](#) describes, CO_2 is the most predominant gas in the discharging fluid, which is characteristics of most geothermal fields along the rift valley.

2.4. Conclusion

The Menengai geothermal area is located in the central segment of the Kenya rift and lies within a complex geological and structural setting, which associates with the formation of the Kenya dome. The Menengai caldera is the most noticeable feature and also a significant source of the geothermal resource. The local surface geology is composed mostly of unconsolidated low-density volcanic soil and sediments. Tuffs and ignimbrites also cover a considerable portion of the surface somehow associated with limited outcrops of phonolites, trachytes, and basalts.

The tectonic activity in the study area is associated with the rift triple junction where the failed rift arm of the Nyanza rift joins the central Kenya rift. Two rift floor volcano-tectonic axes that are important in controlling the geothermal system in the study area are the Molo and the Solai TVA. The primary trend of the tectonic structures in the study area is N-S, NE-SW, and NW-SW and similar to the rest of Kenya Rift. High-density values in the analysis of lineaments are concentrated mostly in areas covered by phonolites and trachytes and more so to the NE area where the rocks are reasonably well exposed. Most of the faults are located adjacent to zones with average to high-density values. However, regions masked by volcanic soil and sediments show low to moderate density values. The orientation of these lineaments corresponds pretty much to the faults, and their general trend matches the regional tendencies of faults. This underscores the importance of lineaments in the mapping of geological structures in inaccessible areas.

The concepts of the reservoir recharge review possible feeder sources of fluids that are thought to contribute to the hydrology of the Menengai geothermal system. At least six possible fluid recharge sources exist in the study area. These sources include; shallow groundwater and runoff from the east, deep westerly groundwater flow, near-surface flow from L. Nakuru, deep northerly flow from L. Naivasha, the contribution of paleolake, and magmatic contributions.

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CHAPTER THREE

GRAVITY DATA ANALYSIS AND INTERPRETATION

3.1 Introduction

A local change in a gravity field is interpreted as arising from density variations in the basic model. The genesis of these variations could be structure related such as fault zones, crack density, rock porosity or perhaps associated with a temperature field. Gravity is commonly employed geophysical tool for prospecting unknown geothermal reservoirs and for monitoring reservoir changes during exploitation. During the exploration stage of geothermal development, gravity results are beneficial in delineating geological structures. In this chapter, the gravity method is utilized in understanding the subsurface structures that are controlling the placement of heat sources and thermal fluid flow in the case of the Menengai geothermal field.

3.2 Gravity method: theory and application

The Newton's law of gravitation states that the force (N) between two point masses m_1 , m_2 (kg) is equal to Gm_1m_2/r^2 , where r (m) is the distance between the masses and $G = 6.67 \times 10^{-11} m^3 kg^{-1} s^{-2}$ (e.g., Hinze et al., 2013). The force is one of the attraction, which acts along the line joining m_1 and m_2 within a gravitation field. The force acting on the point mass m_1 in the gravitation field of m_2 is also given by ' $Force(N) = Mass(kg) \times Acceleration (m/s^2)$ '. A body will experience an acceleration if placed in the vicinity of a point mass of another body's gravitational field, and thus the total force on the mass due to the body in any given direction can be obtained by applying Newton's law to infinitesimal, point mass components of the body and integrating over the entire volume ([Parasnis, 1986](#)).

The gravitation potential (Jkg^{-1}) due to a point mass m at a distance r from itself is given by

$$V = \frac{Gm}{r} \quad [3.1]$$

The acceleration a of any point mass towards m is given by differentiating equation [3.1] with respect to r .

$$a = -\frac{Gm}{r^2} \quad [3.2]$$

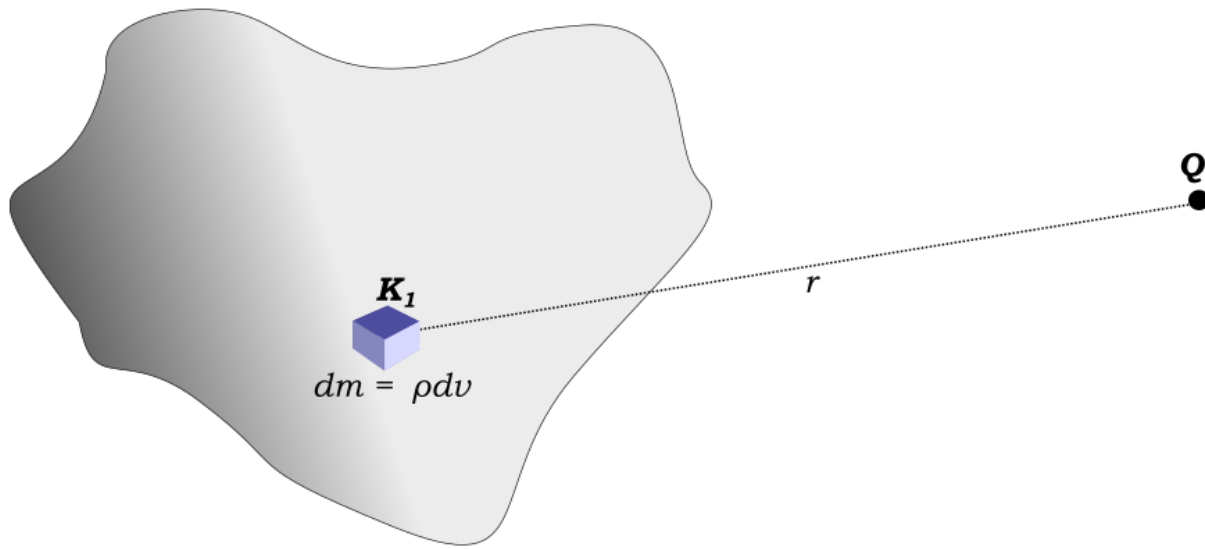


Figure 3.1: Geometry representation of the points K and Q and distance r between them; notations of equation [3.3].

In general, the acceleration of a point mass in a particular direction is given by the derivative of V in that direction. For an arbitrarily shaped body at a point K , the gravitational potential anywhere in the body is

$$V = \int \frac{G\rho dv}{r} \quad [3.3]$$

Where ρdv is the mass of an infinitesimal volume element dv positioned at point Q of the body, ρ is the density at Q , r is the distance KQ and the integral is taken over the entire volume of the body (Figure 3.1).

Gravity is a geophysical method which has been extensively used to examine density inhomogeneities of deeper subsurface bodies and has successfully been used in exploring geological structures (e.g., [Nishijima & Naritomi, 2017](#)), groundwater (e.g., [Carmichael & Henry, 1977](#); [Chen et al., 2014](#); [Feng et al., 2013](#); [Tiwari et al., 2009](#)), magma chambers (e.g., [Battaglia et al., 1999](#); [Fedi et al., 2018](#); [MacQueen et al., 2016](#)), and geothermal resource (e.g., [Guglielmetti et al., 2013](#); [Hunt, 1970](#); [Mariita, 2012](#); [Tontini et al., 2015](#)). The interpretation of lateral structures suffers from an inherent non-uniqueness. That is, an infinite number of mass distributions may give rise to the same potential on the surface ([Phillips & Lambeck, 1980](#)). Nonetheless, a good model should be more sensible and comparable to other available geophysical and geological data.

3.3 Gravity survey

The gravity dataset used in this study covers an area of approximately 880 km² and a total of 1823 stations. The data consist of those we acquired during a field survey conducted in 2016-2017, combined with previous data by [Mungania & Lagat \(2004\)](#), [Geotermica Italiana \(1987\)](#) and Bureau Gravimétrique International – BGI ([Drewes et al., 2016](#)). Measurement stations focused on areas with unreliable and insufficient data, as well as in rather flat and easily accessible terrain settings. Areas situated within protected zones, densely populated urban settlements, heavily forested regions, private firms, and rugged terrains were not accessible.

During the survey, measurements were carried out using a Scintrex CG-5 which has a reading resolution of 0.001 mGal, and repeatability accuracy of 0.005 mGal ([Scintrex, 2012](#)). The stations were positioned by using a Differential Global Positioning System (DGPS), whose

vertical uncertainty is lower than 0.03 m. This form of GPS uses a static base station as the reference position to adjust real-time GPS signals in order to eliminate pseudo-range errors. DGPS data provided accurate relative positions for all the gravity stations measured during the survey.

Usually, the readings of gravimeters drift with time due to elastic change in the springs causing an apparent change in the gravity at any given station. These fluctuations may vary from as low as 0.01 mGal to about 1 mGal per hour (Hinze et al., 2013; Parasnis, 1986). Measurements were done at the beginning of the survey of any given day at a designated station (reference station), repeated at the close of the daily survey. The differences obtained were plotted against the time between the two readings at a station in order to correct the drift problem. The corrections were then obtained from a ‘drift curve’ since most gravimeters drift linearly with time.

3.4 Gravity data reduction and Bouguer density estimation

3.4.1 Gravity data correction

Gravity measurements are usually not made on the Earth’s ellipsoid in actual practice. Hence, appropriate corrections must be applied to the measured gravity to reduce it to sea-level equivalent. These corrections allow the determination of an anomaly by subtracting the normal gravity value from the actual measured value. Several corrections (latitude correction, free-air correction, Bouguer correction and terrain correction), have had to be done to the observed gravity to obtain the Bouguer anomaly.

Bouguer anomaly (BA, *mGal*) can be loosely defined as the measured gravity corrected for the known or modeled gravity effects at a planetary scale (Karcot et al., 2017), and written in the form of a simplified equation (e.g., Yamamoto, 1999):

$$BA = g - \gamma + \beta h - 2\pi G\rho h + \rho T \quad [3.4]$$

where g ($mGal$) is the observed absolute gravity, γ ($mGal$) is the normal gravity value, β ($mGal/m$) is the elevation-dependent free-air term, and T is the terrain correction per unit density. G ($m^3.kg^{-1}.s^{-2}$), ρ ($kg m^{-3}$) and $h(m)$, are the universal gravitational constant, density and elevation terms, respectively.

a). Latitude correction (normal gravity)

Latitude correction γ , deals with the shape of the earth where the radius of the earth's poles is smaller than the radius along earth's equator. However, there is an opposing centrifugal acceleration at the equator and more mass under the equator. The net effect is that gravity at the poles is greater than the equator.

The gravity effect of a reference ellipsoid is represented by normal gravity as a geometry, which can be defined to represent the shape of the Earth. The computation of the gravity effect is straightforward since the ellipsoid can be defined mathematically. The Somigliana–Pizetti formula (e.g., Götze, 2011; Hofmann-Wellenhof & Moritz, 2006) expresses a closed-form expression for calculating the normal gravity on the surface of a geocentric reference ellipsoid, and is expressed by

$$\gamma = \gamma_e \frac{1 + k \sin^2 \phi}{\sqrt{1 - e^2 k \sin^2 \phi}} \quad [3.5]$$

where k is the normal gravity constant (0.00193185265241), γ_e is the normal gravitational acceleration at the equator ($9.7803253359 \text{ m s}^{-2}$), ϕ is the geodetic latitude, and e^2 is the square of the first numerical eccentricity of the ellipsoid ($6.69437999014 \times 10^{-2}$). The values k , γ_e and e^2 are WGS 84 reference ellipsoid derived constants (NIMA, 1984).

b). Free-air correction

The free-air correction β , accounts for gravity disparities caused by elevation changes in the measurements point, p , which is almost never located precisely on the surface of the reference ellipsoid. This implies that the model value at p above the reference ellipsoid will be less than γ , and this is accounted free-air correction, which uses the rate of change in gravity above the Earth

surface to correct for the drop in gravity with elevation. The value of the free-air correction was traditionally approximated as 0.3086 mGal/m, the gradient above a spherical Earth, and can be expressed as

$$\beta = \frac{\partial g}{\partial h} h \quad [3.6]$$

c). Bouguer correction

The Bouguer correction, BGC , is a first-order reduction to account for the excess mass underlying measurement point located at elevations higher than the reference ellipsoid, and also for a mass deficiency at measurement point positioned below the reference ellipsoid. The Bouguer correction is computed by representing the body of rock between the observation point and surface of the reference ellipsoid as a slab that extends horizontally to infinity. The most sophisticated illustrations of the rock mass below the measurement station take accounts for the curvature of the Earth's surface, and the common representation is the spherical cap, δ_{gB}^{cap} , the gravity effect given by [LaFehr \(1991a\)](#):

$$\delta_{gB}^{cap} = 2\pi G\rho[\mu t - \lambda(R_E + t)] \quad [3.7]$$

where ρ is the assumed density of the slab, t is the thickness of the cap which is generally equivalent to h , and G is the universal gravitational constant, $6.67 \times 10^{-11} \text{ m}^3 \cdot \text{kg}^{-1} \cdot \text{s}^{-2}$. R_E is the mean radius of the reference ellipsoid, μ and λ are dimensionless coefficients connected to R_E , and whose angle is subtended by the spherical cap at the center of the Earth. [Bullard \(1936\)](#) originally selected a spherical cap that extends to a distance of 166.735 km from the observation point in order to minimize the difference between the infinite slab and the gravity effect of the spherical cap for a range of elevations ([LaFehr, 1991b](#)).

d). Terrain correction

The terrain correction, T , accounts for disparities in the observed gravitational acceleration caused by the differences in topography of areas surrounding the measurement point. Terrain correction is positive irrespective of whether the local topography is mountains or a valley due

to the assumptions made during the Bouguer correction. In general, computation of terrain corrections is divided into two segments which independently consider the near and far topography (Hinze et al., 2013). A number of methods used in terrain correction have been proposed and reviewed in previous works (e.g., Banerjee & Das Gupta, 1977; Hammer, 1939; Nagy, 1966; Nowell, 1999; Yamamoto, 2002).

In this study, a code written by Nishijima (2009) is used to compute spherical terrain corrections for gravity data using ASTER GDEM v2 Worldwide Elevation Data (1-arc-second resolution) obtained from the NASA Land Processes Distributed Active Archive Center (LP DAAC), USGS/Earth Resources Observation and Science (EROS) Center. This method is based on a standard right rectangular prism model in a distant zone of a given gravity station.

The vertical elements of the gravitational attraction at the origin due to a right rectangular prism bounded by planes $x = x_1$, $x = x_2$ ($x_2 > x_1$); $y = y_1$, $y = y_2$ ($y_2 > y_1$); and $z = h > 0$; is given by

$$g_p = G\rho\{F(x_2, y_2, h) - F(x_1, y_2, h) - F(x_2, y_1, h) + F(x_1, y_1, h)\} \quad [3.8]$$

where G is the universal gravitational constant and ρ is the density of the prism (Banerjee & Das Gupta, 1977), and the function F is given by

$$\begin{aligned} F(x, y, h) = & x \ln \left(\frac{y + \sqrt{x^2 + y^2}}{y + \sqrt{x^2 + y^2 + h^2}} \right) \\ & + y \ln \left(\frac{x + \sqrt{x^2 + y^2}}{x + \sqrt{x^2 + y^2 + h^2}} \right) \\ & + h \arctan \left(\frac{xy}{h\sqrt{x^2 + y^2 + h^2}} \right) \end{aligned} \quad [3.9]$$

3.4.2 Bouguer density estimation

To evaluate the Bouguer anomaly, a priori knowledge about gravity reductions is necessary. Performing a complete Bouguer reduction demands proper selection of the reduction density. The commonly accepted density value of $2.67 \times 10^3 \text{ kg m}^{-3}$ is a reasonable compromise for a specific group of geological rocks (Bowie, 1912), which assumes that the mean density of continental surface rocks is crystalline and of granitic in composition (Hinze, 2003). A quantitative study of gravity anomalies demands an accurate Bouguer reduction density, and considering the geology of Menengai area that predominantly consists of tuffs and sediments, this standard value may not give the optimum value required. Estimation of Bouguer density from the gravity data is deemed the best way to calculate the optimum Bouguer density (Murata, 1993).

a). Simple G-H method

The Simple G-H method generally does not take into account terrain corrections. Rearranging equation [3.4], we obtain

$$BA = (g_{obs} - \gamma + \rho T) + (\beta - 2\pi G\rho)h \quad [3.10]$$

Substituting the first term of the right-hand side with Q and the coefficient of h in the second term with α of equation [3.10], then the equation becomes

$$BA = Q + \alpha h \quad [3.11]$$

The optimum density can then be obtained from the slope of the plot of $Q (= g - \gamma)$ against h .

b). F-H method

The F-H technique was initially proposed by Parasnis (1952, 1986). In this method equation [3.4] is rearranged as

$$B = (g_{obs} - \gamma + \beta h) - (2\pi Gh + T)\rho \quad [3.12]$$

From equation [3.12], the second term is plotted against the coefficient of ρ , and the gradient of the straight line is then determined by least squares technique and the value is taken as the optimum reduction density. The gradient is equivalent to assuming Bouguer anomaly to be a random error. This method minimizes the summation of the difference between each Bouguer anomaly value and the representative $\bar{B}$ of the smoothed profile of Bouguer anomaly, which is averaged over the area of interest. This equation [3.12] can be modified and reduced to

$$B = F - H\rho \quad [3.13]$$

where F is a free-air anomaly and H a topography term.

c). Extended F-H

This technique, which considers a smooth Bouguer anomaly profile, was proposed by [Fukao et al. \(1981\)](#) obtained by dividing the area under consideration into a series of grids of equal size. The method is based on F-H method in which the estimation of Bouguer anomalies is made by a piecewise step function in a 2-D space, taking into account the correlation between the long-wavelength elements of Bouguer anomalies and the fact that topography tends to be supported by dynamic force or in isostatic equilibrium within the earth in a larger area. Rewriting equation [3.4] in a 2-D space, we obtain

$$B_{ij} = g_{ij} - \gamma_{ij} + \beta h_{ij} - 2\pi G\rho h_{ij} + \rho T_{ij} \quad [3.14]$$

$$= (g_{ij} - \gamma_{ij} + \beta h_{ij}) - (2\pi Gh_{ij} - T_{ij})\rho \quad [3.15]$$

$$= F_{ij} - H_{ij}\rho \quad [3.16]$$

moreover, if we assume the number of grids to be M and the number of stations to be N_j within the J th grid, the minimization condition can be obtained as follows

$$S = \sum_{j=1}^M \sum_{i=1}^{N_j} \{F_{ij} - H_{ij}\rho - \bar{B}_j\}^2 \Rightarrow \text{minimum} \left(\bar{B}_j = \frac{\sum_{k=1}^{N_j} B_{jk}}{N_j} \right) \quad [3.17]$$

$$\frac{\partial S}{\partial \bar{B}_j} = \frac{\partial S}{\partial \rho} = 0 \Rightarrow \text{solution} \quad [3.18]$$

The above minimization condition yields the estimated density, ρ_{EF}

$$\rho_{EF} = \frac{\sum_{j=1}^M \sum_{i=1}^{N_j} (H_{ij} - \bar{H}_j)(F_{ij} - \bar{F}_j)}{\sum_{j=1}^M \sum_{i=1}^{N_j} (H_{ij} - \bar{H}_j)^2} \quad [3.19]$$

In general, the resulting estimated Bouguer density depends on the size of the grid.

d). Akaike's Bayesian Information Criterion (ABIC)

Based on the assumption that the Bouguer anomaly varies smoothly with topography, [Murata \(1990, 1993\)](#) proposed the ABIC method, approximating the Bouguer anomaly by a smooth function. In his method, the optimum terrain density is obtained by minimizing the function

$$\sum_{i=1}^N \{F_i - \rho H_i - f(x_i, y_i | \mathbf{s})\}^2 + \sum_{k=1}^2 \omega_k \int_v \int_v \|\nabla^k f\|^2 dx dy \quad [3.20]$$

$$\Rightarrow \text{minimum}$$

where f is the cubic B spline function that is fitted to the Observed Bouguer anomalies, $F_i - \rho H_i$, $\nabla^k f$ is the k -th differentiation of the function f , $\mathbf{s}$ is a vector of spline parameters, and ω_k is the trade-off parameters. The roughness of the first-order gradient of f and second-order curvature of f , of the Bouguer anomaly surface are controlled, by the trade-off parameters ω_1 and ω_2 , respectively, and a fitting choice of these parameters is made by minimizing ABIC ([Akaike, 1998](#)). Equation [3.20] above can be reduced in form of a matrix as

$$\|F - \rho H - Es\|^2 + \|D_\omega s\|^2 \Rightarrow \text{minimum} \quad [3.21]$$

where F and H are vectors that represents free-air anomalies and terrain-Bouguer corrections per unit density, respectively. The term s is the M parameters of the function f whereas D_ω and E are an $N \times M$ matrices containing ω_1 and ω_2 parameters and spline values at the stations, in that order.

e). Comparison with the Variance of the Upward-Continuation (CVUR)

The CVUR method was proposed by [Komazawa \(1995\)](#). The method assumes that the surface layer produces shorter wavelength elements of Bouguer anomalies, and in the case where the density corresponds with the surface layer density, the first layer will then appear transparent and the short wavelength elements caused by the surface topography will have to be attenuated. The more natural way is to compare the variances obtained from band-pass or high-pass filtered residuals and to determine the appropriate density as the one giving the minimum variance. These variances are taken to be a measure of roughness or randomness resulting from the skin effects. The method is applied to the meshed dataset, and the grid at a point $p(i, j)$ are interpolated from the Bouguer anomalies. Assuming the correction is made with an estimated density ρ followed by an upward-continuation to a height h , the Bouguer anomaly at point p can be defined as $g_p(\rho, h)$, and the variance $a_p^M(\rho; h_1 h_2)$ is then defined by the differences between the two upward-continued anomalies and given by

$$a_p^M(\rho; h_1 h_2) = \frac{\sum_q^M W_q \{g_{p-q}(\rho, h_1) - g_{p-q}(\rho, h_2) - (g_p^M)\}^2}{\sum_q^M W_q} \quad [3.22]$$

$$= \sum_q^M W_q \{\nabla g_{p-q}(\rho; h_1 h_2) - (g_p^M)\}^2 / \sum_q^M W_q \quad [3.23]$$

where M is the number of range of sampling, q is the convolution parameter of grid point p , h_1 and h_2 are continuation heights, $\sum_q^M$ denotes the summation with respect to q within a circular region ($|q| \leq M$). The term g_p^M is the weighted mean value and W_q is the weight adopted for smoothing.

3.5 3-D gravity inversion

The implementation of 3-D modeling of gravity has become an essential tool in the exploration of geological resources (Li & Oldenburg, 1998). However, the inversion of the field gravity data is one of the most ambiguous problems in exploration geophysics studies (Oldenburg, 1974). Inversion of gravity data is usually fraught with problems due to data noise and inhomogeneity of geological bodies even with a reasonably accurate gravity measurement and data reduction. Therefore, carrying out gravity inversion of field data and generating geologically realistic results may not be easy because the inverse problem in potential field modeling suffers from ambiguity and non-uniqueness of its solutions.

To remedy the non-uniqueness problem and improve the inversion performance, one may opt for regularization or instead addition of *a priori* knowledge as constraints to reduce the ambiguity of the inverse problem. The application of *a priori* data, such as, drilling or geological information to constrain gravity inversions has been increasing in the last two decades (e.g., Dahlin et al., 2002; Fullagar et al., 2008; Li & Oldenburg, 1996, 2000; Represas et al., 2012).

In this study, *a priori* information from borehole geology from the initial ten geothermal wells drilled in Menengai geothermal field has been integrated to produce a reasonable source model approximation (Table 3.1). These wells are located within the caldera and reveal the geological formation intercepted down to a minimum depth of 2077 m and a maximum of 3189 m. The lithostratigraphy is predominantly composed of trachytes, tuffs, and syenitic intrusives. Trachyte is the most dominant rock-type. Tuffs occur as thin intercalations between the thick trachyte lavas, and in some wells, syenitic intrusives occur intermittently. The lithology of these wells generally shows very close similarity, but the lack of a clear marker horizon makes it difficult to correlate the various stratigraphic units (GDC, 2018).

Table 3.1: Different rock types and their depths of occurrence from the initial 10 geothermal wells drilled in Menengai geothermal field (adapted from [GDC, 2018](#)).

Well	Total depth (m)	Depths (m) of occurrence		
		Tuff	Trachyte	Syenite
M01	2195	-	1050	1900
M02	3189	520, 2000	1100, 2350	2700
M03	2196	650	1350	2100
M04	2106	800	1300	2080
M05A	2085	-	1100	2085
M06	2171	680	1350	2150
M07	2136	580	850, 1600	220
M08	2345	1090	2000	-
M09	2077	940	1200	-
M10A	2150	1620	1000, 2100	1945

The actual measurements of rock densities from these wells (or of surface geology within the study area) were not available, and as a result, representative densities were approximated. The densities for volcanic sediments ($1.85 \times 10^3 \text{ kg m}^{-3}$), tuffs ($2.11 \times 10^3 \text{ kg m}^{-3}$), trachyte ($2.63 \times 10^3 \text{ kg m}^{-3}$), phonolite ($2.53 \times 10^3 \text{ kg m}^{-3}$), basalt ($2.82 \times 10^3 \text{ kg m}^{-3}$) and syenite ($2.75 \times 10^3 \text{ kg m}^{-3}$) rock formations were estimated from the literature ([Daly, 1935](#); [Jacoby & Smilde, 2009](#); [Schön, 2015](#); [Tenzer et al., 2011](#)).

These approximations gave a density contrast of $0.97 \times 10^3 \text{ kg m}^{-3}$ from an average of $2.32 \times 10^3 \text{ kg m}^{-3}$, which is somehow higher than the computed density of $2.23 \times 10^3 \text{ kg m}^{-3}$ that was used to reduce the gravity data. The discrepancy is largely attributed to the dominance of the low-density sediments and tuffs, which tends to lower the calculated Bouguer density. Despite this, the Bouguer density value was removed for each of the rock types, which corresponds to the removal of a uniform background from the observed data. As a result, a density contrast of between $-0.38 \times 10^3 \text{ kg m}^{-3}$ and $0.59 \times 10^3 \text{ kg m}^{-3}$ was obtained. The resulting density contrast was then empirically gridded to a distance of 50 m using Kriging method to obtain a reference model (Figure 3.2).

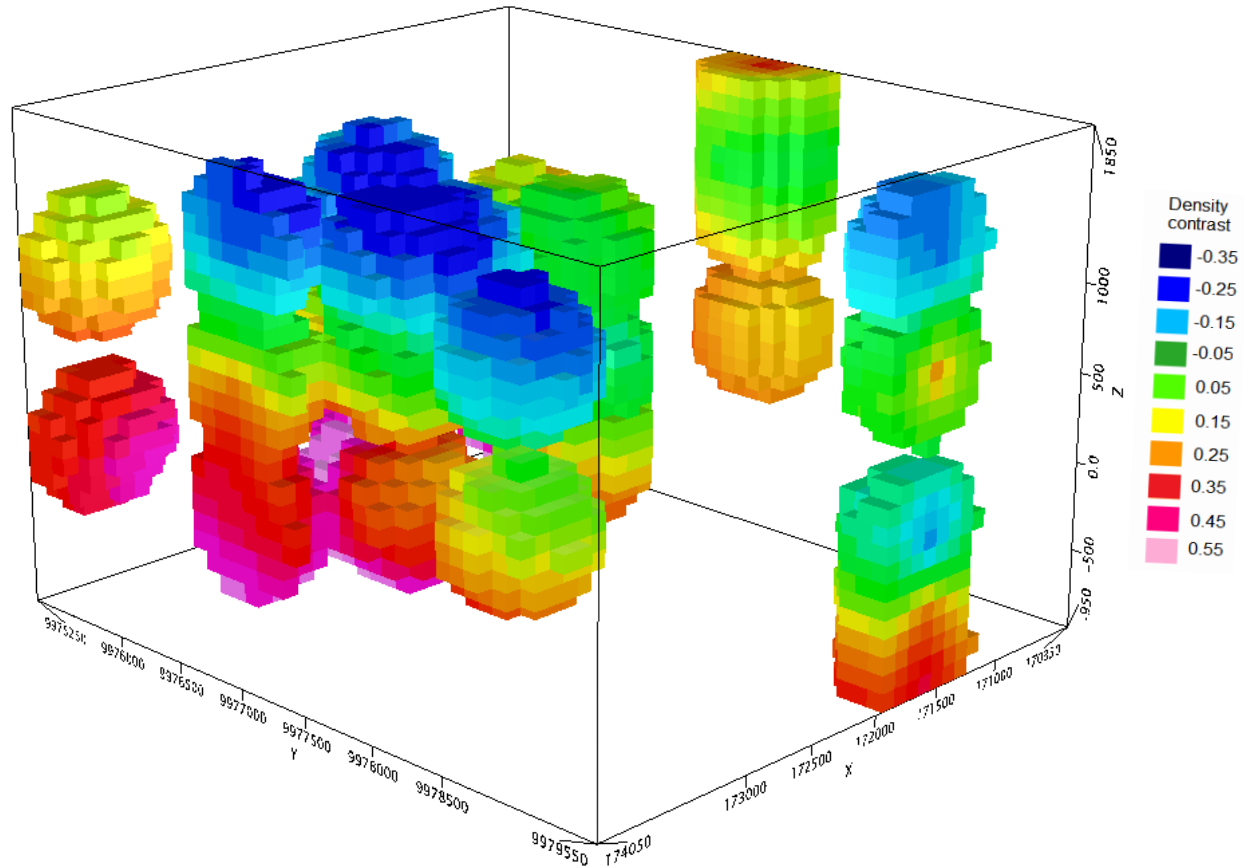


Figure 3.2: Reference model from the gridded density contrast of the down hole density estimated from the initial 10 geothermal wells drilled in Menengai.

To determine the density distribution of the subsurface structure, we carried out a 3-D inversion of the Bouguer gravity anomaly using the VOXI Earth Modelling tool, a voxel-based tool that is built within the Geosoft platform. This platform is a cloud-based computing service, and this makes it possible to calculate a large-scale 3-D model at high speed without depending on the local computer specification. It applies a Cartesian Cut Cell method developed by [Ingram et al. \(2003\)](#), which allows accurate geometric representation of geological surfaces ([Ellis & MacLeod, 2013](#)). The solution search method uses the iterative re-weighting least squares (IRLS) algorithm initially developed by [Lawson \(1961\)](#). The IRLS approach solves certain optimization problems using weighted least squares calculations, iteratively. In order to obtain a stable solution, the objective function is regularized by the misfit between observed and calculated values. The reduced chi-squared misfit ([Taylor, 1997](#)) is then used to measure the degree which indicates that the predicted response agrees with the observed data.

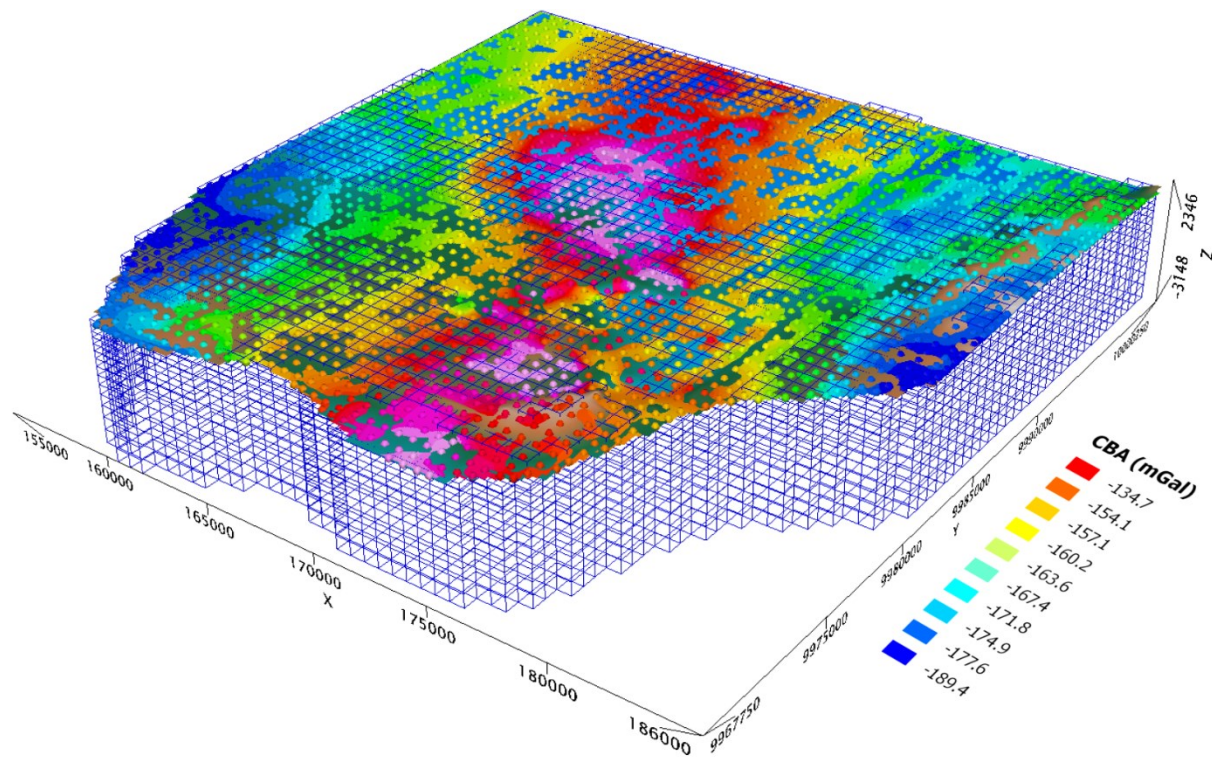


Figure 3.3: A 3-D Gravity inversion grid. The cell size is $620 \text{ m} \times 650 \text{ m} \times 50 \text{ m}$ with the height component set to increase by a factor of 1.08 for each successively deeper layer. The colored map represents complete Bouguer anomaly (CBA) with values ranging from -189.4 mGal to -134.7 mGal .

The subsurface of the study area was discretized into $58 \times 60 \times 49$ rectangular cells in the x, y and z-direction, respectively (Figure 3.3). The size of a rectangular cell was set to 620 m and 650 m along respective x and y directions. The height of the cells increases from a top layer of 50 m by a factor of 1.08 for each successive deeper layer. The inversion process involves varying the trade-off regularization parameter in an attempt to fit the data, and the inversion improves the model iteratively until the predicted data satisfy the residual chi-squared criterion to obtain the optimal model.

The reference model generated from the density contrasts was used to constrain the inversion. To ensure that the results are within an acceptable range, lower-bound and upper-bound

constraints of $-0.38 \times 10^3 \text{ kg m}^{-3}$ (volcanic-sediment formations) and $0.59 \times 10^3 \text{ kg m}^{-3}$ (basaltic formations), respectively, were set. These values are based on the density variation derived from the estimated densities of local geology. Also, an Iterative Reweighting Inversion Focusing (IRIF) constraint was applied to enhance the model and produce a more geologically plausible inversion result (Ellis, 2012). IRIF uses the output of the inversion as a weighting function in successive iterations and this results in a more focused model. The weighting function makes sharpening of contrasts possible by enhancing boundaries around distinctive anomalous features (MacLeod & Ellis, 2015). In the present case, we applied two passes of IRIF to emphasize both positive and negative ends of density variation.

The inversion will improve the model iteratively until the predicted data satisfy an assigned tolerance criterion. In this study, a maximum absolute error was initially set at 0.05 mGal but then adjusted until an acceptable fit tolerance of 0.02 mGal was achieved. The simulation ran for 21 iterations and resulted in a 3-D model with density contrasts that range from $-0.35 \times 10^3 \text{ kg m}^{-3}$ to $0.58 \times 10^3 \text{ kg m}^{-3}$ and lie within a geologically reasonable density variation for the study area. The Bouguer density was then added to each cell to obtain the density distribution. Finally, the results were gridded to a higher resolution of $120 \times 120 \times 120$ rectangular cells using an isotropic inverse distance weighting interpolator (Shepard, 1968) to enhance the presentation of the model recovered from the inversion.

3.6 Results and Discussion

3.6.1 Bouguer density estimation

According to Hay et al. (1995), the rift infill consists of a lower layer, up to 500 m thick, of volcanic rocks and sediments with a density of $2.45 \times 10^3 \text{ kg m}^{-3}$ expressive of predominantly down-faulted Plateau phonolites. These phonolites are in turn overlain by later volcanic rocks and sediments with a density of about $2.2 \times 10^3 \text{ kg m}^{-3}$, ranging up to 2500 m thick, representing a significant proportion of low-density tuffs and alluvium. Mungania & Lagat (2004) prepared plots for density values ranging between $1.9 - 2.5 \times 10^3 \text{ kg m}^{-3}$ using Nettleton's method (Nettleton, 1939, 1976). They found that to get the best fit for the region of the rift shoulders a

Bouguer density of $2.5 \times 10^3 \text{ kg m}^{-3}$ had to be used, while for the rift floor, a density of $2.0 \times 10^3 \text{ kg m}^{-3}$ was the most appropriate.

[Simiyu & Keller \(2001\)](#) assessed density measurement of more than 2000 samples from drill holes to a depths of about 3000 m in the geothermal prospects. Shallow drill holes outside the rift reveal densities of less than $2.4 \times 10^3 \text{ kg m}^{-3}$ up to the depths of 2 km (i.e. at sea level), rising to more than $2.5 \times 10^3 \text{ kg m}^{-3}$ as the depth increases. The mean drill core density value is between 2.5 and $2.6 \times 10^3 \text{ kg m}^{-3}$, and the in-situ density measurements carried out for geothermal exploration in the volcanic terrains show basement density values that range from $2.58 - 2.88 \times 10^3 \text{ kg m}^{-3}$.

The evaluation of Bouguer anomaly value requires *a priori* information about the gravity reduction density, and in this study, some methods were selected and used to ensure a better approximation of Bouguer density. The G-H method ([Rikitake et al., 1965](#)) equation [3.11], F-H method ([Parasnis, 1952, 1986](#)) equation [3.13], Extended F-H ([Fukao et al., 1981](#)) equation [3.19], the Akaike's Bayesian Information Criterion (ABIC) method ([Murata, 1990, 1993](#)) equation [3.21], and Comparison with Variance of the Upward-Continuation (CVUR) method ([Komazawa, 1995](#)) equation [3.23], gave comparable Bouguer densities of $2.10 \times 10^3 \text{ kg m}^{-3}$ (see Figure 3.4), $2.22 \times 10^3 \text{ kg m}^{-3}$ (see Figure 3.5), $2.33 \times 10^3 \text{ kg m}^{-3}$ (see Figure 3.6), $2.24 \times 10^3 \text{ kg m}^{-3}$, and $2.25 \times 10^3 \text{ kg m}^{-3}$ (Figure 3.7), respectively. An average density of $2.23 \times 10^3 \text{ kg m}^{-3}$ was selected and deemed enough to produce a smooth anomaly. This average value lies within the range of density values proposed in some previous studies and owes to the large proportion of tuffs and sediments estimated by [Hay et al. \(1995\)](#) to be as thick as 2.5 km and a mean density of $2.20 \times 10^3 \text{ kg m}^{-3}$.

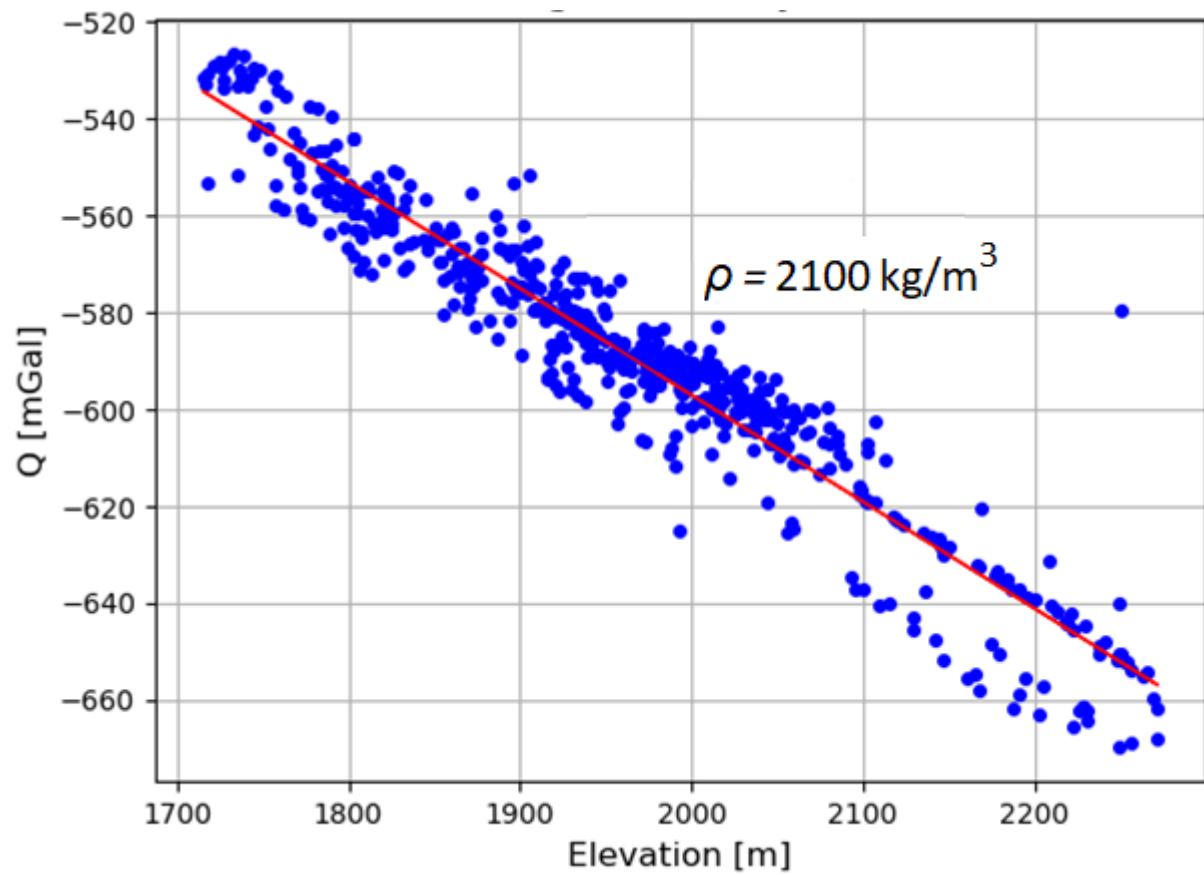


Figure 3.4: Bouguer density by G-H method

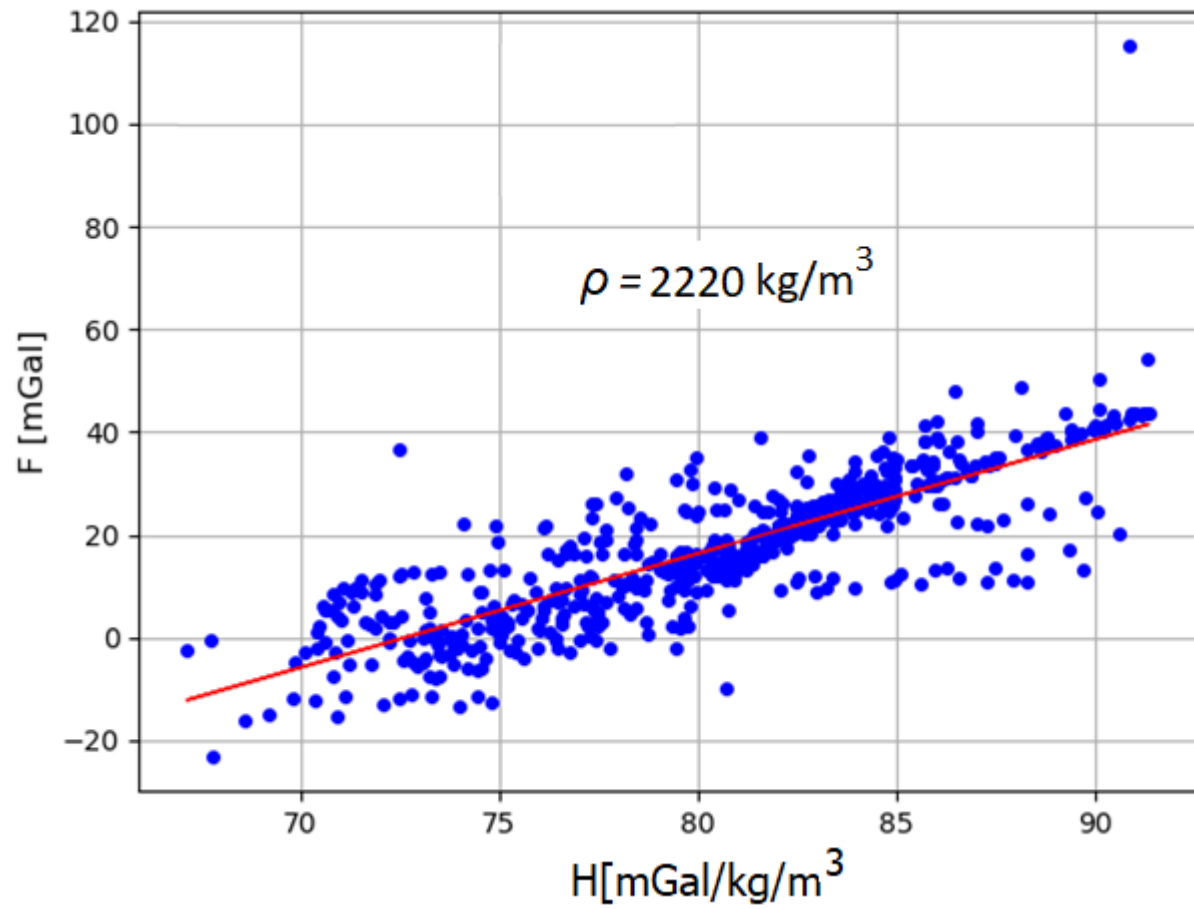


Figure 3.5: Bouguer density by the Simple F-H method

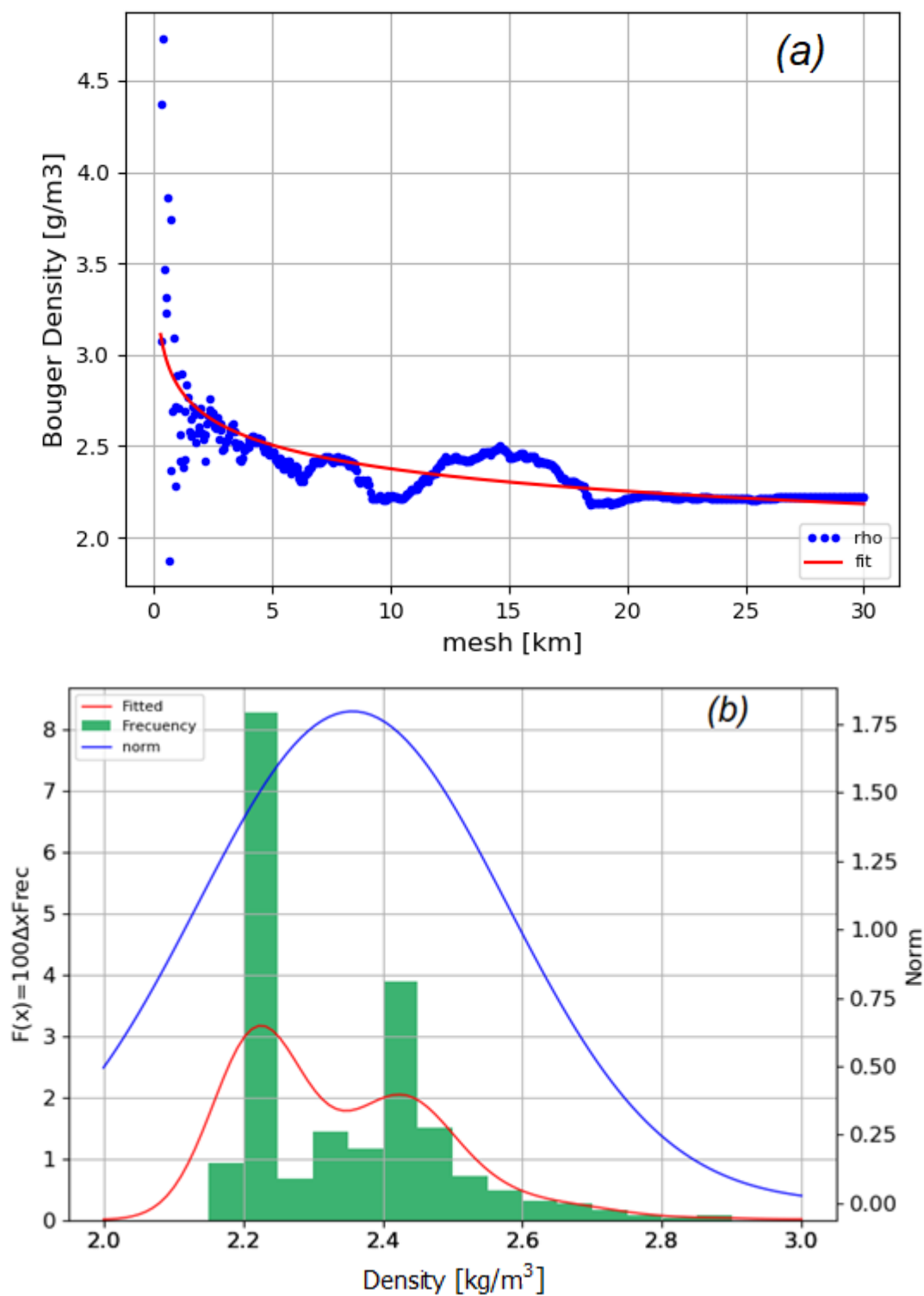


Figure 3.6: (a) Bouguer density by Extended F-H method, (b) Extended F-H density frequency distribution.

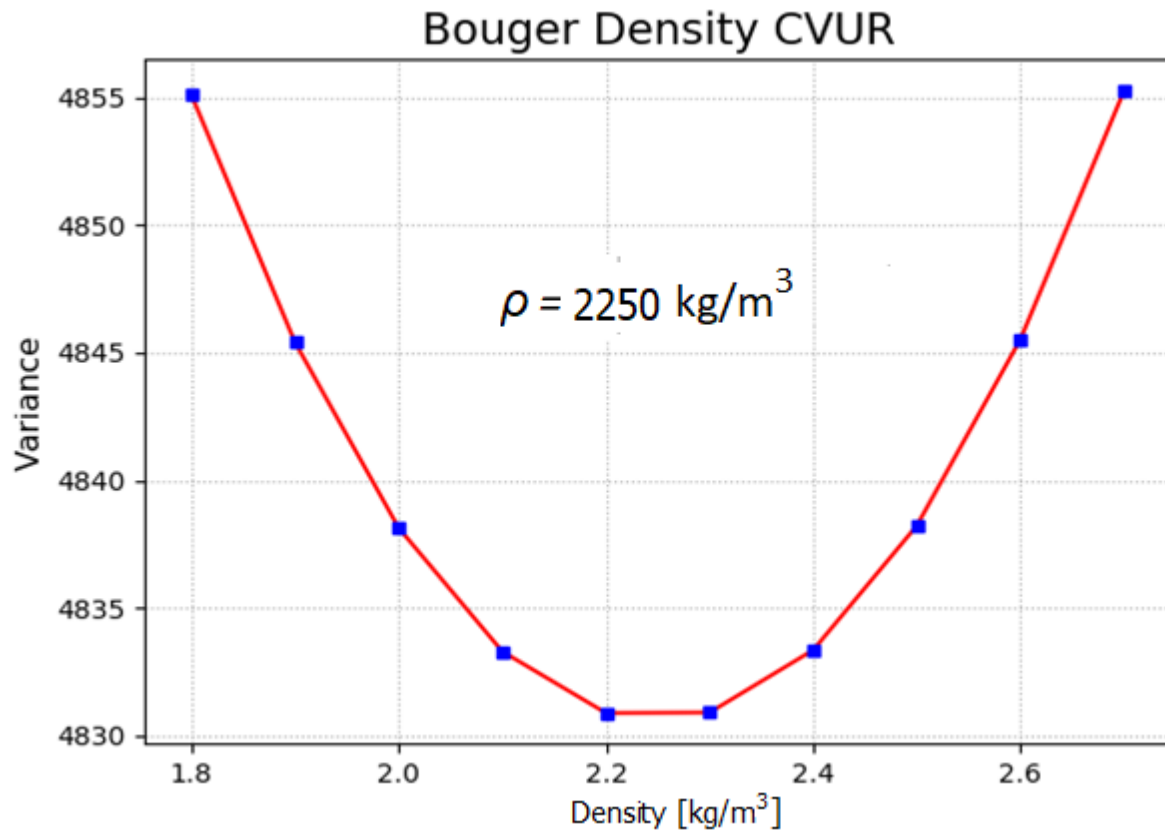


Figure 3.7: Bouguer density by Comparison with Variance of the Upward-Continuation (CVUR).

3.6.2 Bouguer anomaly

Interpretation of Bouguer anomaly was limited to the distribution of survey stations shown in Figure 3.8. Areas with no stations were removed and structures located in areas with sparse stations interpreted with caution, nevertheless, the areas considered relevant to this study appear to be well covered during the survey. The complete Bouguer anomaly was interpolated to a 250-m grid with the minimum curvature method with tension (Smith & Wessel, 1990), using Surface module implemented in the Generic Mapping Tools system (Wessel et al., 2013).

The minimum curvature method is widely used but old principle for constructing smooth surface estimates from irregularly spaced data (Briggs, 1974). The estimated surface of minimum curvature relates to the minimum of the Laplacian differential equation or, in an alternative formulation, satisfies the biharmonic differential equation. Gridding with tension is an improvement to the minimum curvature methods as it adds a degree of freedom and relaxes the

minimum-curvature constraint by introducing some tension into the surface. This technique tries to force a thin elastic plate to go through randomly-spaced (x,y,z) dataset and generates a binary grid file of gridded values $z(x, y)$ by solving:

$$(1 - T) \times L(L(z)) + T \times L(z) = 0 \quad [3.24]$$

where T is a tension factor between 0 and 1, L indicates the Laplacian operator, and z is small displacements a thin elastic plate of constant flexural rigidity.

The Surface module achieves this by finding a function that satisfies the following constraints (Smith & Wessel, 1990):

$$\begin{aligned} Z(x_k, y_k) &= z_k, & \text{for all data } (x_k, y_k, z_k), k = 1, n \\ (1 - t)\nabla^4 Z - t\nabla^2 Z &= 0 & \text{elsewhere} \end{aligned} \quad [3.25]$$

where t is the tension, $0 \leq t \leq 1$.

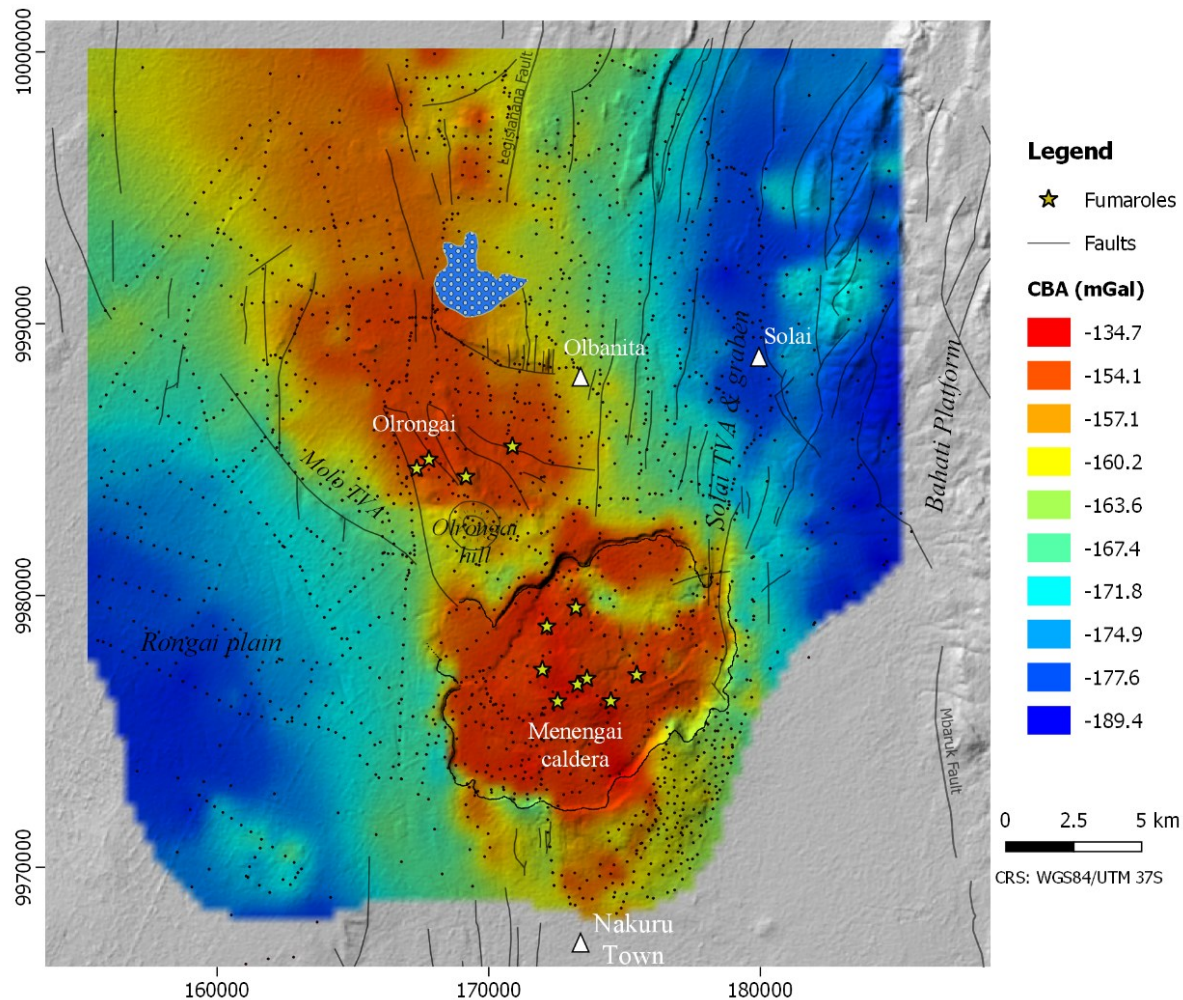


Figure 3.8: Bouguer anomaly map calculated for a Bouguer density of $2.23 \times 10^3 \text{ kg m}^{-3}$.

The Bouguer anomaly observed has high values trending NNW-SSE that changes from -189 mGal to -134 mGal, with a dominant two-fold anomaly located to the north-northwest of the caldera and within the caldera (see Figure 3.8). The low Bouguer anomaly is associated with Solai graben and Rongai plain located to the north-east and south-west areas, respectively. As seen from local geology, these low anomaly zones are covered by volcanic soil and sediments. The high Bouguer anomaly appears to be confined within the caldera and to the immediate north-west of the caldera (Olrongai hill) and is associated with the occurrence of trachytic rocks as well as geothermal surface manifestations as evidenced by fumarolic activities. The caldera rim and the extensive faulting seem to control the extent of the hydrothermal system.

3.6.3 3-D gravity modeling

The planar slices at elevations of 1000 m asl (Figure 3.9), sea level (

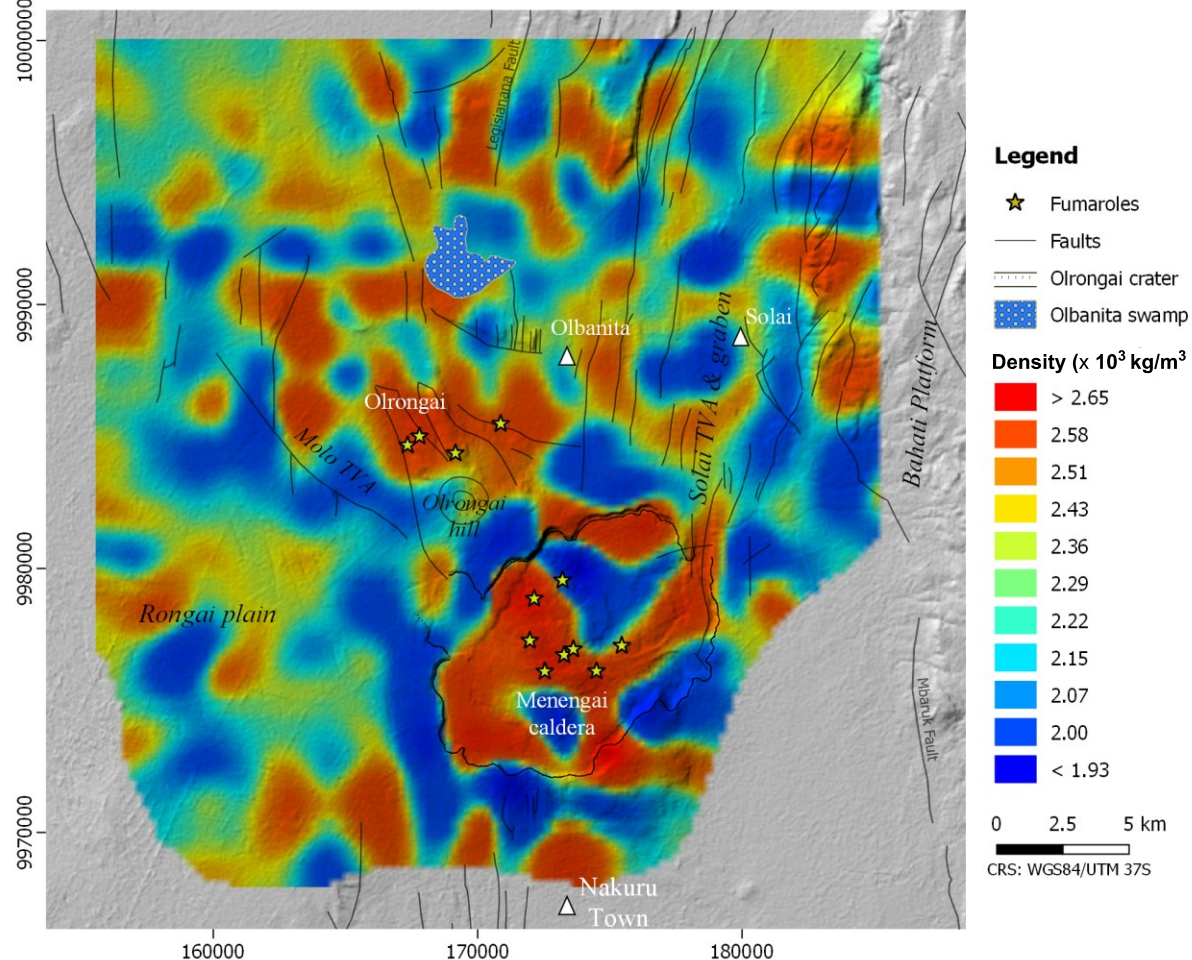


Figure 3.10) and 1000 m bsl (Figure 3.11) were extracted from the inversion model, then overlain on a shaded relief map and plotted alongside faults. These anomalous areas are viewed in shallower levels as discrete bodies but converge with depth to form two main bodies; lying within the caldera and the Olrongai area. The flow of magma through and along the faults in the shallow parts of the system seems to have caused the differences of density distributions with depth. These two main bodies coincide with the presence of trachytes while the dense bodies north of Olbanita and also north and east of Solai seem to be associated with the presence of phonolites.

A striking relationship is also observed between the occurrence of the high-density anomalous structure and the existence of geological faults. The two appear to coexist, and this association becomes even more distinct with increasing depth, implying that these structures acted as

conduits during the ascent and emplacement of at least the phonolites and trachytes. The high-density anomalous structure beneath the caldera appears to be confined mainly within the caldera walls while NNE-SSE and N-S trending structures enclose the anomaly beneath Olrongai. This relationship underscores the significance of these structures (i.e., ring structures and faults) in guiding and restraining the up flow of magma and also the hydrothermal system.

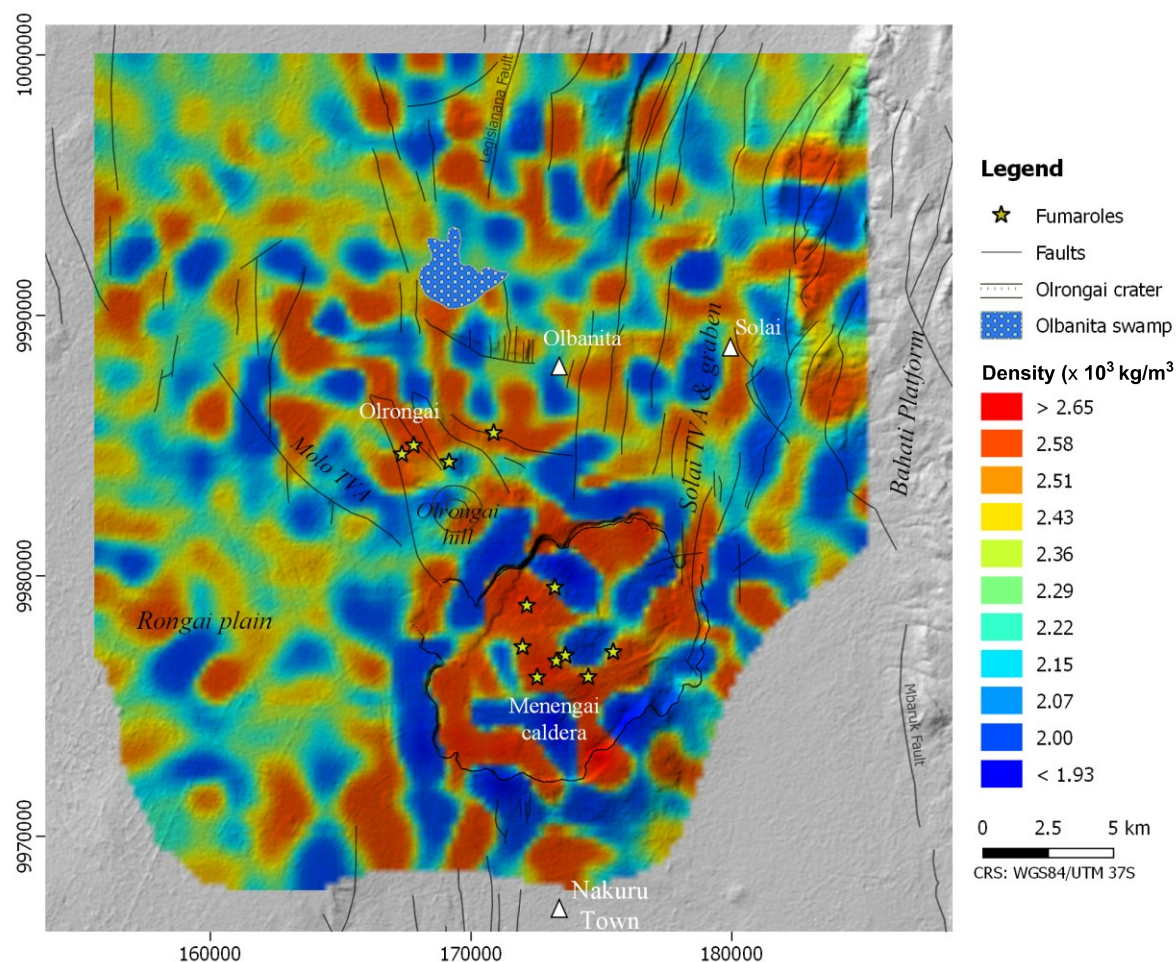


Figure 3.9: A planar view of density distribution at 1000 m asl elevation.

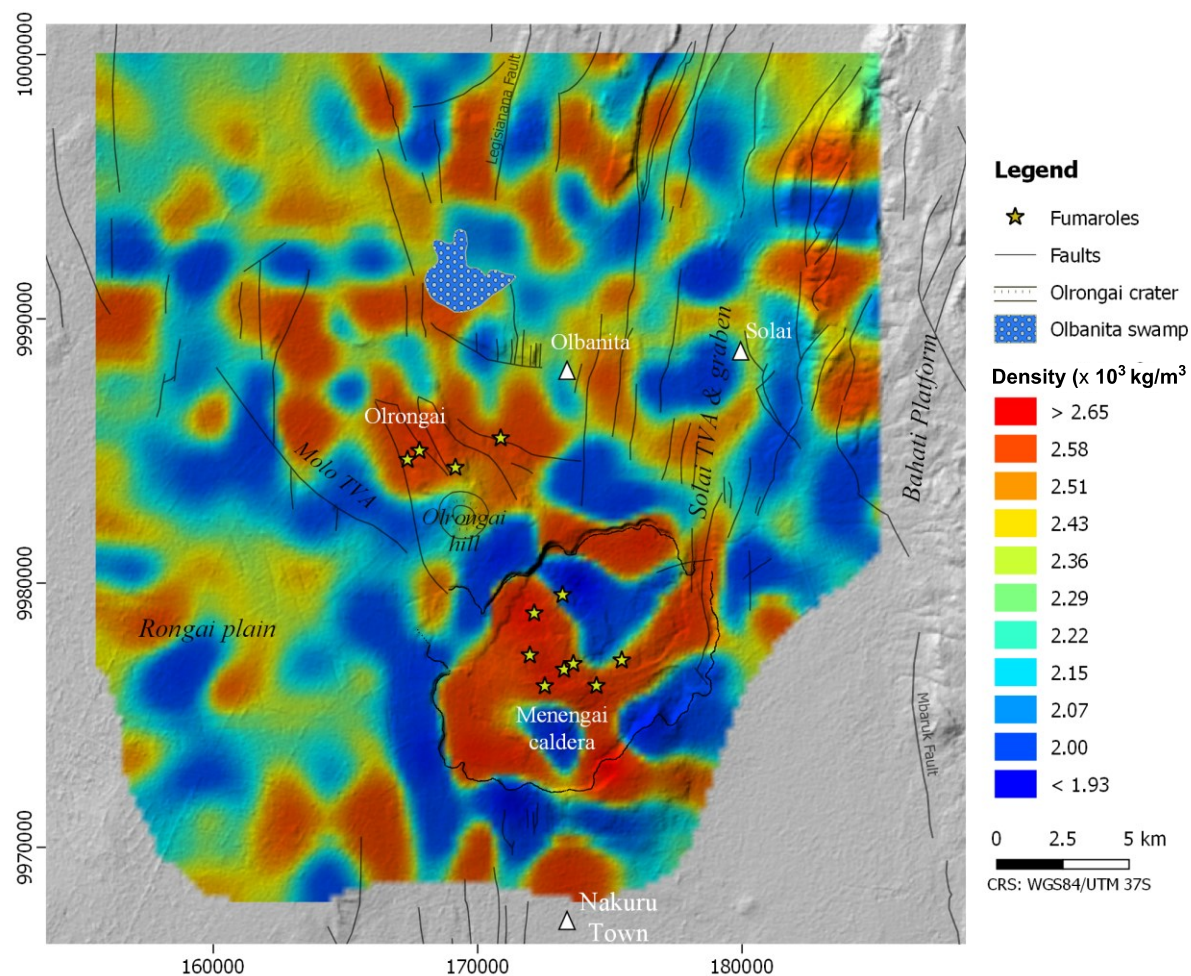


Figure 3.10: A planar view of density distribution at sea level.

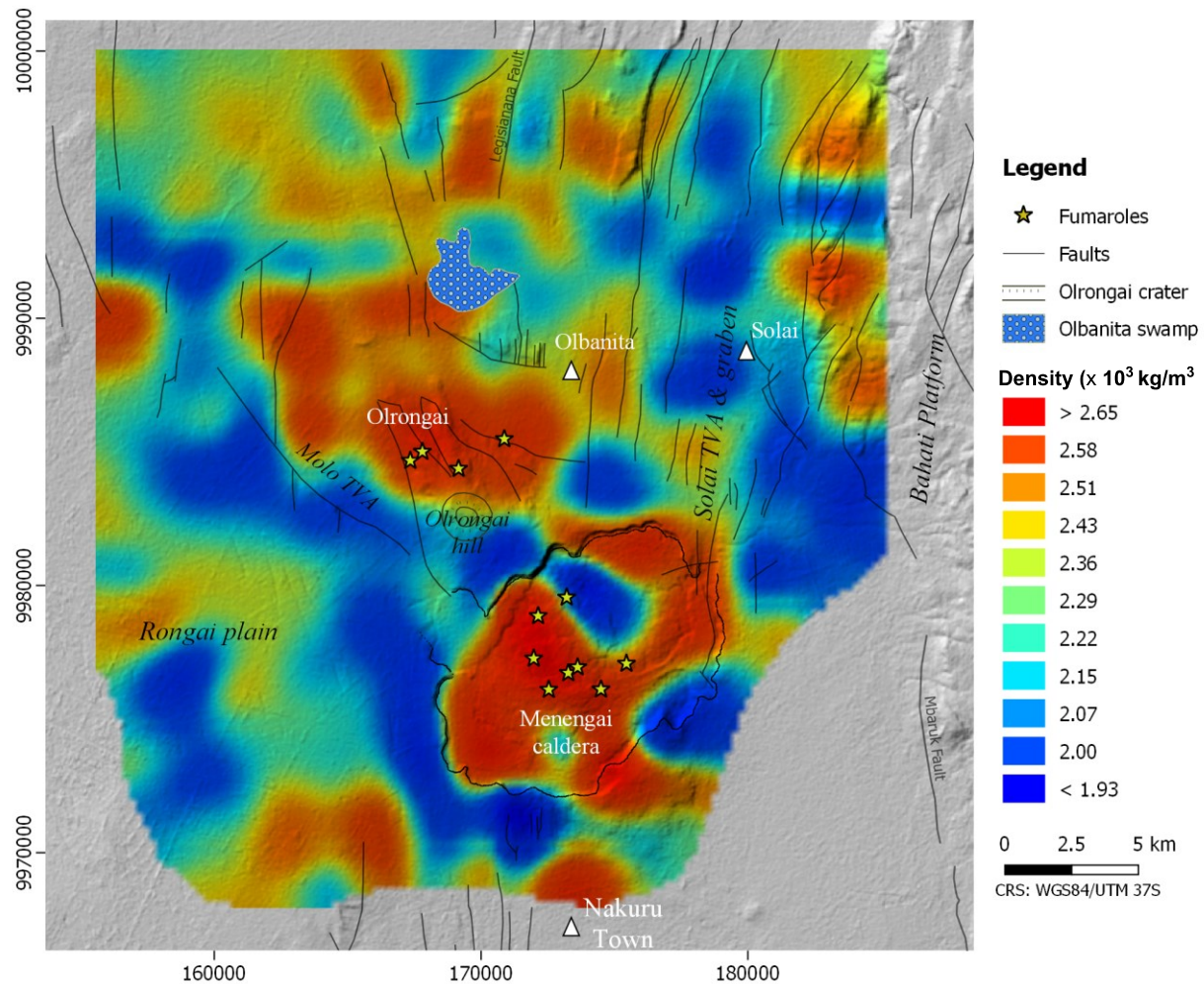


Figure 3.11: A planar view of density distribution at 1000 m b.s.l elevation.

The isosurface of $2.45 \times 10^3 \text{ kg m}^{-3}$ in Figure 3.12 infers the geometry and positions of shallow-crustal high-density volumes located beneath the caldera and Olrongai hill. The isosurface displays the geometry of the high-density structure of the study area, and similar isosurface was also used in Figure 3.13 and Figure 3.14. The two masses appear as discrete bodies, but they probably may be connected at depth. These two high-density features underlie Menengai caldera and Olrongai areas that have been explored for geothermal power generation. Menengai caldera and Olrongai are known by their surface manifestation of the high-temperature geothermal system, and the presence of dense bodies suggest the heat source of this area.

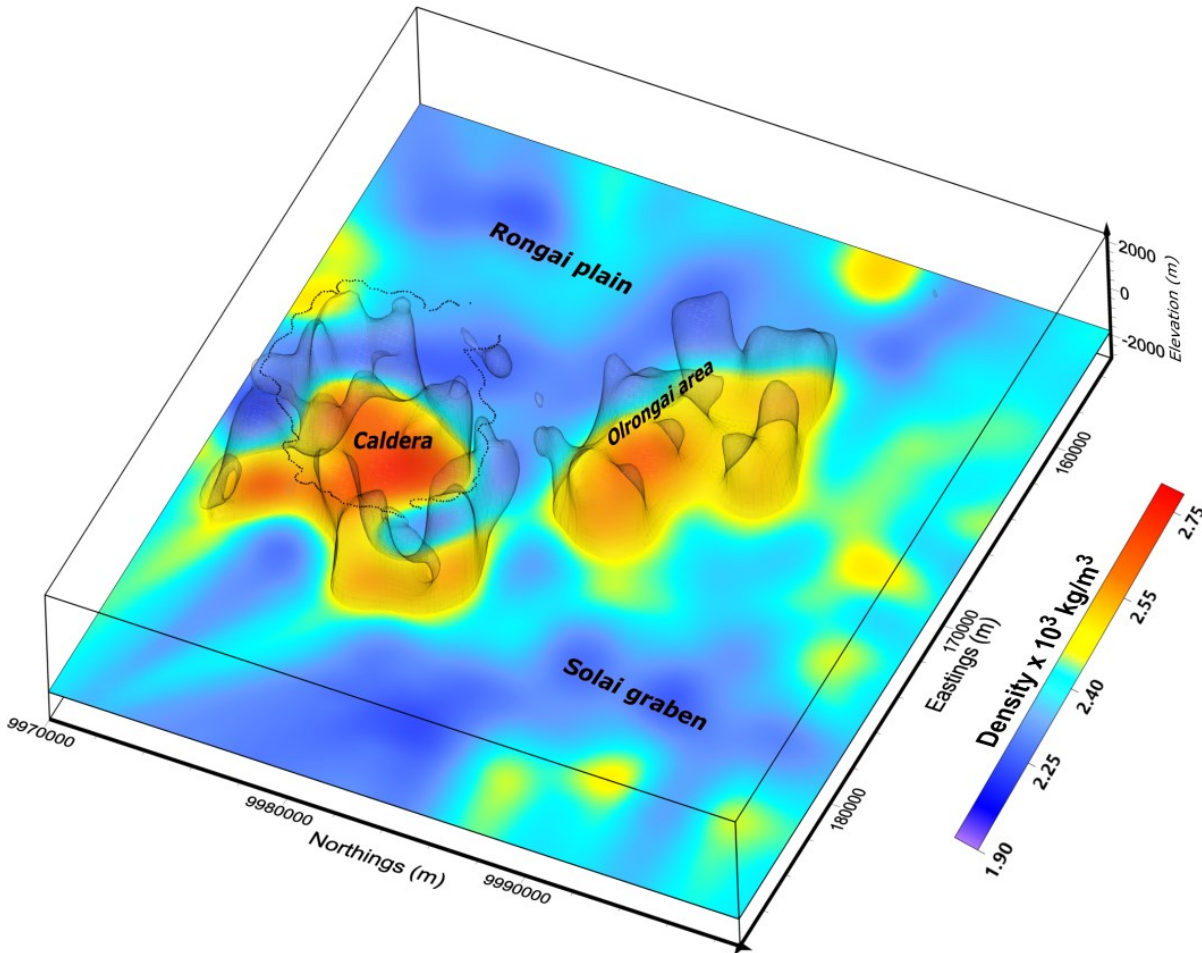


Figure 3.12: A planar profile and isosurface ($2.45 \times 10^3 \text{ kg m}^{-3}$) inferred from the density model.

The orientation of tectonic faults shown in Figure 3.13 appears to delimit the lateral extent and intensity in distributions of the dense body beneath Olrongai hill. The relationship is still apparently visible despite the depth difference between surface structures and the dense body. These Upper Pleistocene faults (McCall, 1967) are relatively younger and comprise of new faults and renewals of old fractures. The presence of fumaroles at boiling temperatures (Mungania & Lagat, 2004), located adjacent to faults suggests that the geothermal reservoir is positioned at a relatively shallow depth and likely beneath the fumaroles.

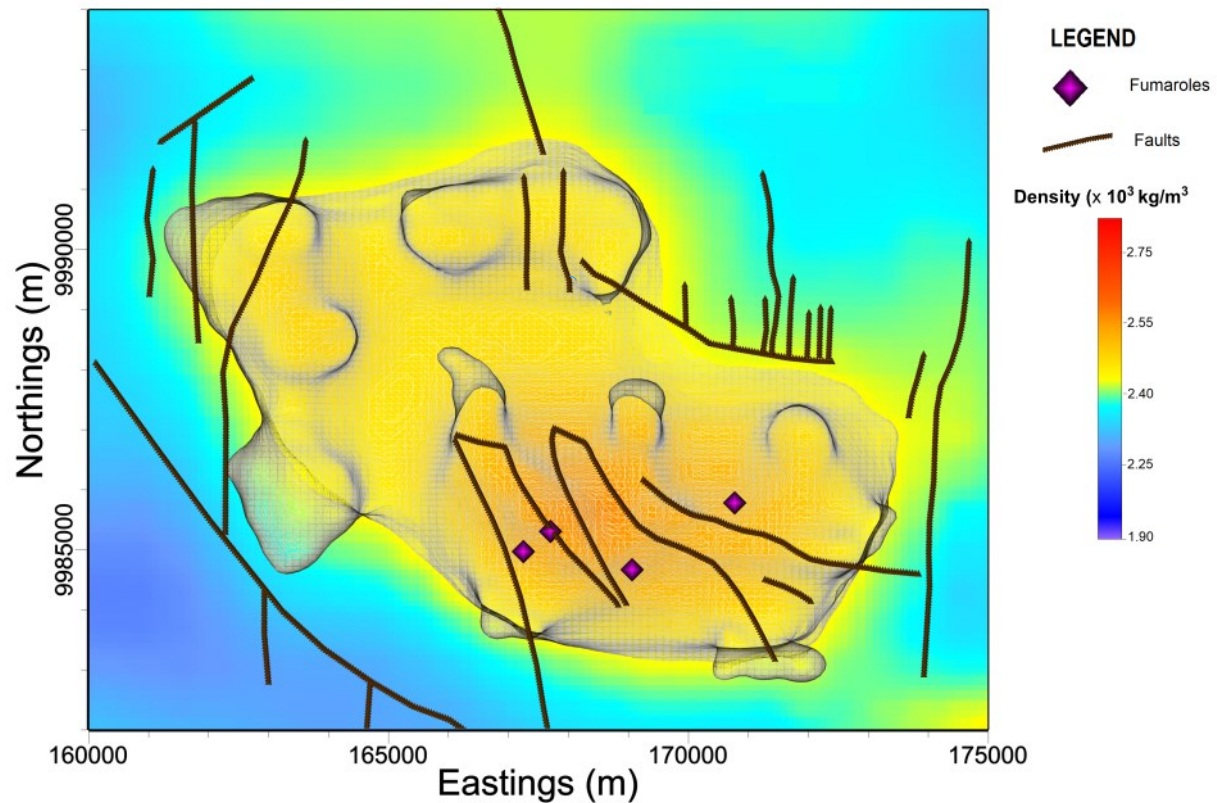


Figure 3.13: A display of geological faults as seen from the surface and fumaroles location superimposed against a deep planar layer at sea level and an isosurface of $2.45 \times 10^3 \text{ kg m}^{-3}$ to highlight the dense body of Olrongai hill.

Unlike Olrongai area, the caldera in Menengai has no clear subsurface structures other than the E-W arcuate structure, which appears to have aided the ascend of younger trachytic magma to the surface (Figure 2.2). The pattern of this structure coincides with a high-density anomaly shown in Figure 3.14. The eastern end of this structural feature seems to be connected with a series of NNE trending normal faults. These faults consequentially render the caldera wall incomplete as it cuts through the northeastern rim by north-south trending segment graben (McCall, 1967), marking the southern end of the Solai graben. Some inferred faults are presented in Figure 3.14 based on the shape of the dense body, the caldera rim, and the pattern displayed by the locations of fumaroles.

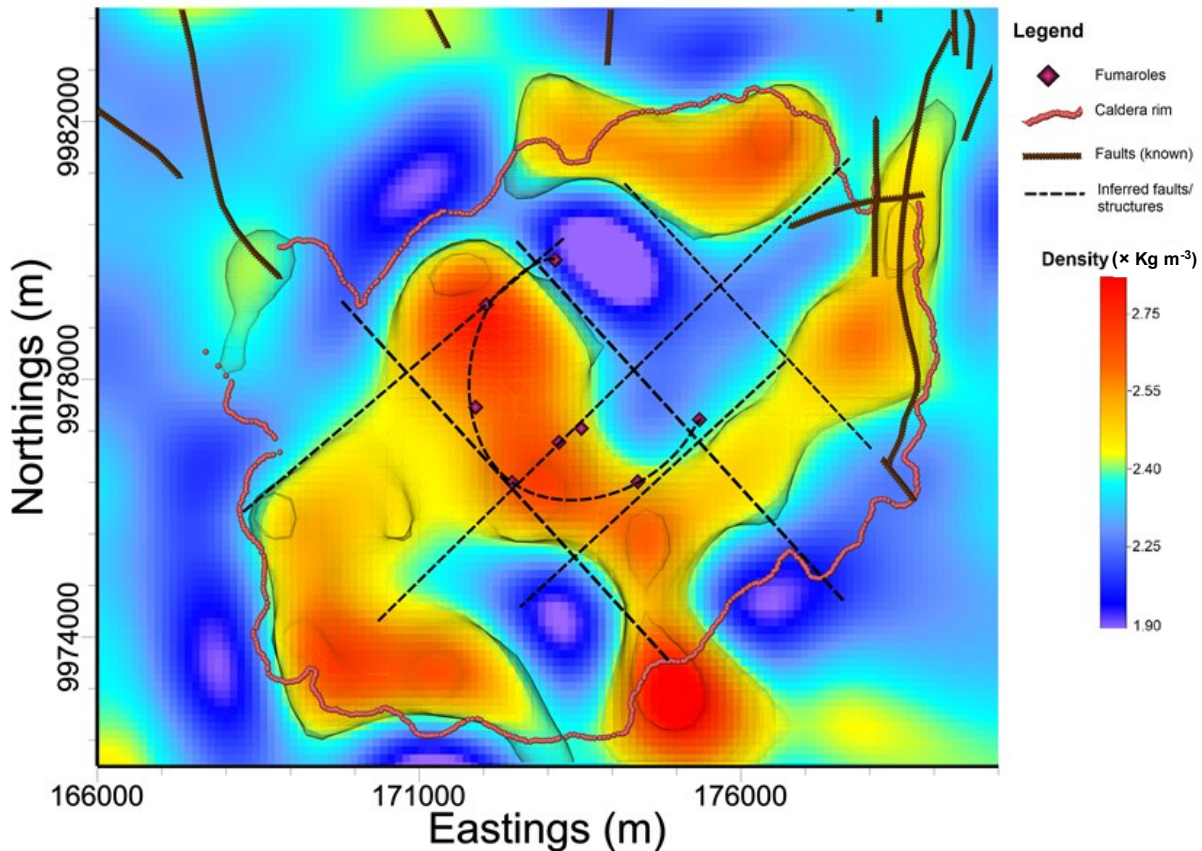


Figure 3.14: A display of geological faults/caldera rim ring-fracture as seen from the surface and fumaroles location superimposed against a deep planar layer at sea level and an isosurface of $2.45 \times 10^3 \text{ kg m}^{-3}$ to highlight the dense body of Menengai caldera. The inferred faults are presented as broken lines.

The location of fumarolic activities in most cases describes the zones of high thermal anomalies generally located close to or on the fault structures, which offer a pathway for fluids to rise to the surface showing zones of hydrothermal upwelling. A magma emplaced at shallow levels beneath the caldera may cause localized but significant deformation that might have led to the growth of the dome-shaped forced structure and possible fracturing. The loop-shaped pattern portrayed by the locations of fumaroles inside the Menengai caldera in Figure 3.14 seems to suggest the presence of an arcuate fracture system, which can be interpreted as surface-expression of a ring-fault created during the post-caldera collapse activity. Such arcuate fracture systems normally relate with the local intensification of the tension resulting from inflation of underlying magma bodies commonly seen or inferred at volcanoes (e.g., Anderson, 1936).

A closer look at the caldera wall, the distribution of fumaroles and the shape of the dense body appears to propose an alternative view of NE and NW striking network of faults (Figure 3.14), which would support a possible extension of post-caldera faults across the caldera superficially obscured by younger deposits. The fumaroles appear to be located right on top of the dense trachytic formation but on the shallower levels of ~1000 m asl (see Figure 3.9), they appear close to the edges. This is probably because the hydrothermal system occurs at lithological contact zone where the permeable host-formation allows circulation of water that interacts and is heated by the intrusives embayed within the dense formation. The generation of a network of faults or a ring fracture and subsequent uprising of fumarolic fluids along these openings could hypothesize a possible coupling of the hydrothermal-magmatic system, considering the indication of magma pockets or dikes interpreted from the borehole data to be as shallow as 2 km (i.e. at sea level).

Although the results obtained from the inversion model are encouraging, there are some limitations. First, a significant part of the caldera floor is known by its rugged terrain that is mostly inaccessible ([Mungania & Lagat, 2004](#)). Therefore, the gravity data coverage is somewhat sparse (Figure 3.8), and may not be conclusively used to infer small changes in density or explicitly define lithological contacts. Hence, this study could not image these pockets of magma since they are confined within the trachytic and syenitic formations.

Secondly, other than the E-W arcuate structure that is readily observable, the presence of faults inside the caldera floor is poorly known although some significant attempts have been made to uncover them (e.g., [Riedl et al., 2015](#); [Strecker & Melnick, 2013](#)). The caldera floor is covered with more than 70 post-caldera lava flows, pumice deposits and strombolian cinder cones ([Macdonald et al., 2011](#)). This makes it difficult to establish any existence of geological faults that would be useful in adding value to the model. Nonetheless, the imaging of the dense body beneath the E-W arcuate structure confirms the validity of the 3-D inversion model.

Lastly, we constrained the model in this study by including density estimates of lithologies interpreted from drill hole cuttings. However, intermittent losses of circulation affected the recovery process and led to incomplete lithological logs ([GDC, 2018](#)). Consequently, the depths

affected by these losses were avoided and this narrowed the scope of available geology. To improve the robustness and reduce any uncertainty on the inversion model, a rigorous characterization of the rock unit densities from measurements of downhole and surface samples collected from the study area is necessary

3.7 Conclusion

The main aim focus of conducting gravity study was to assess the relationship between geological structures and a geothermal heat source in the Menengai geothermal area. An improved Bouguer anomaly map of the Menengai area using updated gravity dataset presented here reveals a dominant two-fold high Bouguer anomaly underlying the caldera and Olrongai areas, and are associated with the occurrence of young trachytic rocks and active fumarolic activities. Away from this northerly trend, low-density sectors extend over the volcanic-sediment terrains of Rongai plain and Solai graben.

A 3-D gravity inversion for predicting the geothermal heat source architecture within the Menengai caldera and Olrongai is also presented. The model provides an in-depth insight into the subsurface structure from potential field perspective by constraining deep bodies in 3-D space. This is significantly important for geological modeling in such a complex geothermal system. The host-rock boundary is highlighted by strong density gradient that suggests a precipitous contact between two lithologies delineated by tectonic faults, more so evident in Olrongai.

The complex nature of post-caldera emplacement of trachytic formation and ash-flow tuffs is related to many eruptions. The products of these eruptions cover the caldera floors, hence make gravity techniques one of the best method for inferring buried faults. The inversion model reveals information about the undulation of the dense bodies beneath Olrongai hill and Menengai caldera. The location of the two main high-density structures in the shallower portions of the model corresponds with eruption centers, manifested by the existence of younger volcanic rocks and fumarolic activities, and its geomorphology at depth is interpreted to embay the heat source crucial for the geothermal system.

These results show the significance of the gravity inversion method to provide information of subsurface structure beneath the Menengai geothermal field. The subsurface model obtained adds to the existing volcanological and geological information, giving valuable details about heat source geometry and geological structures that control its lateral extent. This model can be used to delineate target areas for more detailed geothermal investigations.

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CHAPTER FOUR

RESISTIVITY STUDIES USING MAGNETOTELLURIC METHOD

4.1. Introduction

The MT method is a geophysical technique that utilizes temporal fluxes of the naturally occurring electromagnetic (EM) fields of the Earth to resolve spatial variations in the electrical resistivity of the beneath the Earth's surface. This technique works because measurements taken at the surface for both the amplitude and phase fields are reliant on the electrical resistivity of the subsurface. The magnetotelluric method (MT) method is useful in the exploration of the geothermal resource since electrical resistivity is sensitive to the presence of thermal fluids, which is the goal of this chapter. In this chapter, 1-D, 2-D and 3-D inversions of MT data are used to explain the resistivity study of Menengai geothermal field. A review of the MT method in geothermal exploration, and description of MT data acquisition, removal of distortion and static shifts, as well as dimensionality analysis, were done. It also presents the models derived from the 1-D, 2-D and 3-D inversions inversion, deductions of the resistivity structures beneath the caldera and Olrongai area, and description of the fitting of the measured MT data to the model response.

4.2. The magnetotelluric method

The magnetotelluric (MT) exploration technique was initially proposed by [Tikhonov \(1950\)](#) but stayed almost unknown until later when [Cagniard \(1953\)](#) suggested and developed the theory behind the MT method. The technique involves recording electric and magnetic field variations in the Earth caused by naturally occurring, time-varying geomagnetic field. MT is a

passive electromagnetic method capable of determining the electrical resistivity response in the subsurface on depth scales extending from a few tens of meters down to the mantle transition zone, and therefore it is possible to map the subsurface resistivity structure up to several tens of kilometers. Various authors have extensively reviewed the theory and applications of MT (e.g., [Berdichevsky & Dmitriev, 2008](#); [Chave & Jones, 2012](#); [Simpson & Bahr, 2005](#)).

The distance over which EM fields diffuse depends on the frequency of the variations and the resistivity of the Earth, which allows MT to yield both lateral and vertical constraints. These geomagnetic variations range between periods of 10^{-3} s and 10^5 s (or frequencies from 10^3 Hz to 10^{-5} Hz) depending on their origin. The passive source EM signals used in the MT method arise from two distinct processes, each has a distinctive frequency content ([Heise et al., 2006](#)). Signals of short periods below 1 s are mainly due to meteorological processes such as lightning activity at equatorial latitudes. The EM signal discharged by lightning, known as ‘sferics’, may saturate amplifiers from the vastly unsettled equatorial regions, which travels around the Earth within a resistive waveguide bounded by the conductive ionospheric plasma region and the relatively conductive surface of the Earth ([Simpson & Bahr, 2005](#)).

EM signal at longer periods greater than 1 s result from solar-driven ionospheric electric currents produced by the interaction between the ionosphere and Earth’s magnetosphere. This solar wind is a stream of plasma, that continually radiates electrons and protons from the Sun. The solar wind varies in both intensity and velocity, and this can alter the Earth’s magnetosphere, producing constant magnetic field variations which are measurable at the surface of the Earth. The Earth itself has a strong internal magnetic field thought to be produced by electrical currents in the conductive part of the core driven by a dynamo process ([Lowrie, 2007](#)). However, the internal magnetic field has variations on the timescales of years, and so is of limited use for geophysical exploration techniques, and thus the external magnetic field is responsible for the primary source of EM signals for MT exploration in geothermal systems.

In MT surveys, two components of the electric field (E_x and E_y) and three components of the magnetic field (H_x , H_y and H_z) are measured simultaneously on the earth's surface in MT surveying. The apparent resistivity of the subsurface is determined from the relationship of these

components. By assuming that the EM field contains a vertically propagating plane wave, the electromagnetic diffusion equations describe electromagnetic induction in the subsurface as follows

$$\nabla^2 \mathbf{E} = \mu\sigma \frac{\partial \mathbf{E}}{\partial t} = i\omega\mu\sigma \mathbf{E} \quad [4.1]$$

$$\nabla^2 \mathbf{H} = \mu\sigma \frac{\partial \mathbf{H}}{\partial t} = i\omega\mu\sigma \mathbf{H} \quad [4.2]$$

where $\mathbf{E}$ is an electric field in $\frac{V}{m}$, $\mathbf{H}$ is magnetic field intensity in $\frac{A}{m}$, ω is the angular frequency, μ is the magnetic permeability of free space, and σ is conductivity of the Earth. The orthogonal elements of the horizontal electric ($E_{x(\omega)}, E_{y(\omega)}$) and magnetic ($H_{x(\omega)}, H_{y(\omega)}$) fields are related via a complex electromagnetic impedance tensor, $Z_{xy(\omega)}$, and can be written as

$$Z_{xy} = \frac{E_{x(\omega)}}{H_{y(\omega)}} \quad [4.3]$$

The impedance (has units of ohms, Ω) is a complex quantity that has both imaginary and real components, which depends on the angular frequency of the EM signal. For a 1-D Earth, wherein conductivity varies only with depth, the diagonal elements of the impedance tensor, Z_{xx} and Z_{yy} (which couple parallel electric and magnetic field components) are zero, whilst the off-diagonal components (which couple orthogonal electric and magnetic field components) are equal in magnitude, but have opposite signs, i.e.

$$\left. \begin{array}{l} Z_{xx} = Z_{yy} = 0 \\ Z_{xy} = -Z_{yx} \end{array} \right\} \text{1-D} \quad [4.4]$$

For a 2-D Earth, in which conductivity varies along one horizontal direction as well as with depth, Z_{xx} and Z_{yy} are equal in magnitude but have opposite sign, while Z_{xy} and Z_{yx} differ, i.e.

$$\left. \begin{array}{l} Z_{xx} = -Z_{yy} \\ Z_{xy} \neq -Z_{yx} \end{array} \right\} \text{2-D} \quad [4.5]$$

For a 2-D Earth with the x- or y-direction aligned along electromagnetic strike, Z_{xx} and Z_{yy} are again zero. Mathematically, a 1-D anisotropic Earth is equivalent to a 2-D Earth. Similarly, the individual components of Z of 3-D Earth can be defined in tensor notation as,

$$\begin{pmatrix} E_x \\ E_y \end{pmatrix} = \begin{pmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{pmatrix} \begin{pmatrix} H_x \\ H_y \end{pmatrix} \quad [4.6]$$

where Z_{xx} , Z_{xy} , Z_{yx} and Z_{yy} are the components of the impedance tensor. Z is a complex second-rank frequency dependent matrix and contains the information about the subsurface conductivity structure. The apparent resistivity, ρ_{ij} , and phase, Φ_{ij} , are calculated from impedance components, Z_{ij} , written as

$$\rho_{ij} = \frac{1}{\omega\mu_0} |Z_{ij}|^2, \quad [4.7]$$

$$\Phi_{ij} = \tan^{-1} \left(\frac{\text{Im}(Z_{ij})}{\text{Re}(Z_{ij})} \right), \quad [4.8]$$

where $\text{Im}(Z_{ij})$ and $\text{Re}(Z_{ij})$ are imaginary and real part of Z_{ij} , respectively.

4.3. Magnetotelluric in geothermal exploration

Geophysical methods can be used to provide constraints on the physical properties of the subsurface structure related to geology. Its application to geothermal exploration contributes to a better understanding of the location and geometry of both reservoir and heat source in geothermal systems. These systems are generally characterized by faulted and fractured zones filled with geothermal fluids. The geothermal fluids tend to have high concentrations of dissolved salts, and whose conductivity intensifies with temperature (Ucok et al., 1980), and therefore these salts provide conducting electrolytes in the rock matrix. The highly mineralized waters of high-

enthalpy geothermal resources give attractive targets for electromagnetic methods since they produce substantial variations in subsurface electrical resistivity (Spichak & Manzella, 2009). Such variation renders resistivity an essential physical parameter to image the subsurface structure of a geothermal system and determine its characteristics.

The MT method can map deep subsurface conductive bodies and has proved to be valuable in the exploration of high enthalpy geothermal systems in the Kenya rift (e.g., Lichoro, 2010; Lichoro et al., 2017; Wamalwa et al., 2013). Temperatures of high-enthalpy (magmatic or volcanic-associated) hydrothermal systems of volcanic derivation are generally more than 150-200°C (e.g., Muñoz, 2014). Such high-enthalpy systems are found commonly in tectonically active regions such as the rift zones and young volcanic regions where the high thermodynamic energy favorably generates the temperatures.

Figure 4.1 shows the resistivity and conductivity values of various rock-forming materials. They were calculated in the laboratory, from the resistivity values obtained in the field, and therefore inferences about the lithological properties of the geothermal system can be made. Factors such as fractures, porosity, and fluid saturation are known to influence conductivity measurements, and thus a wide range of values for the same substance exists in the literature (González-Álvarez et al., 2014).

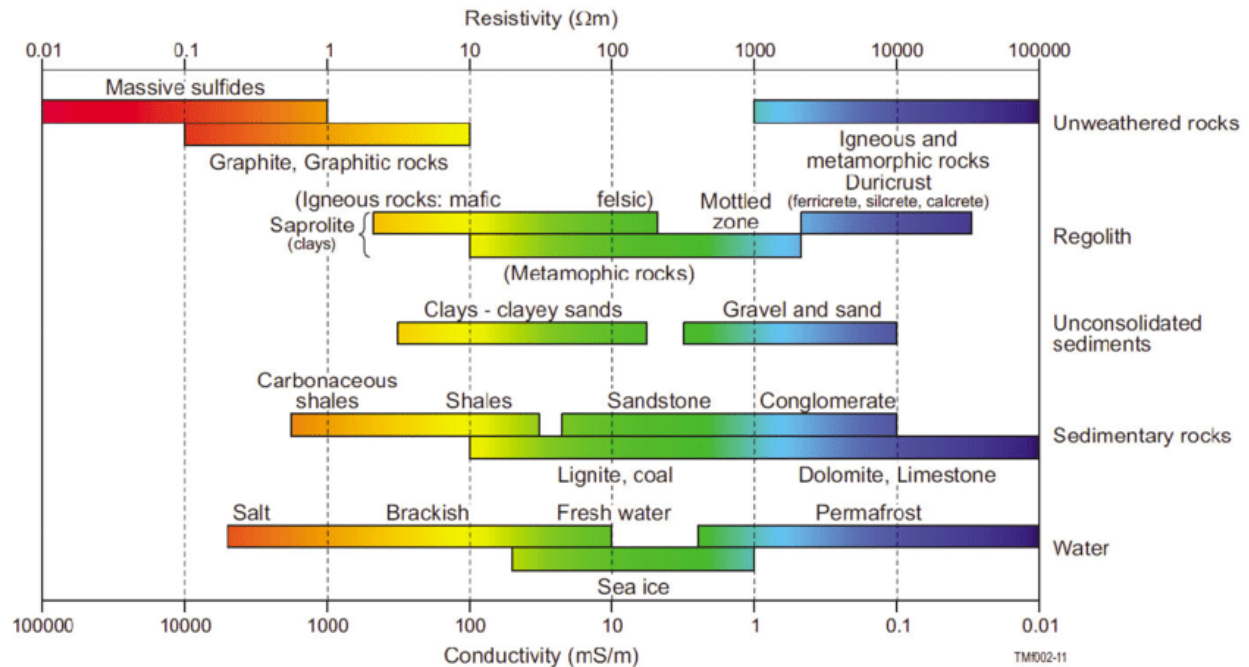


Figure 4.1: Resistivity and conductivity values of various rock-forming materials (from [González-Álvarez et al., 2014](#); adapted from [Palacky, 1988](#)).

Some review studies have been extensively done on the use of electromagnetic methods in geothermal (e.g., [Meju, 2002](#); [Muñoz, 2014](#); [Pellerin et al., 1996](#); [Spichak & Manzella, 2009](#); [Ussher et al., 2000](#)). The geology of geothermal systems is more often complex due to hydrothermal fluid circulation and alteration. Resistivity surveys have the potential to isolate active from relict areas of a geothermal system ([Ussher et al., 2000](#)), as it helps to delineate the geothermal reservoirs and the conductive clay cap.

Figure 4.2 below shows a typical structure of a high-enthalpy geothermal system. It consists of a colder upper zone characterized by smectite alteration minerals, which is relative of low resistivity and forms at a temperature of 70 °C and above ([Anderson et al., 2000](#)). Smectite is more often interlayered with illite, a less conductive clay and whose percentage rises with increasing temperature forming about 70% of the mixture at 180 °C. Beyond this temperature, the smectite decreases while the illite content increases steadily replacing the remaining smectite. Pure illite usually appears above 220 °C at the propylitic alteration zone dominated by other high-temperature alteration minerals such as epidote and chlorite.

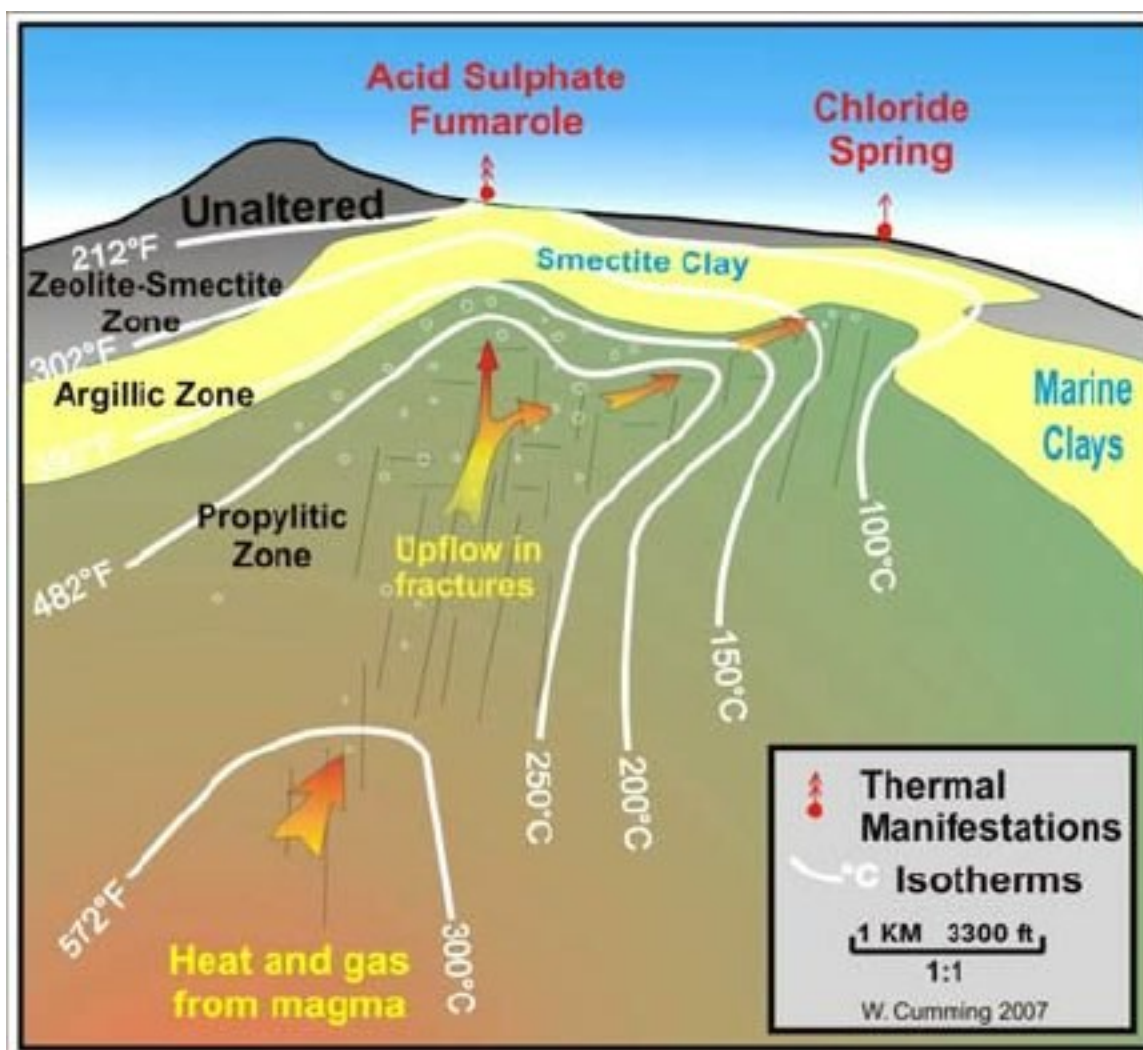


Figure 4.2: A conceptual model of a generic geothermal system with a 250 to 300°C geothermal reservoir with isotherms, alteration zones, and structures (Cumming & Mackie, 2010).

The resistivity of smectite zone is generally between 1 – 10 $\Omega\cdot\text{m}$ and is determined by intensity and type of alterations, temperature and the extent of water saturation. Resistivity increases at a higher temperature and as such depends mostly on temperature, salinity, and porosity. The geothermal structure in a flat terrain will form the smectite layer of low resistivity right above the high-temperature geothermal reservoir, and thus shallow resistivity anomalies can reliably indicate the general location of the geothermal resource. However, in steeper terrain, the overall structure of the geothermal system is more complicated due to the existence of a substantial hydrological gradient below the surface (Anderson et al., 2000).

4.4. Magnetotelluric data acquisition, processing and analysis

4.4.1. Data acquisition

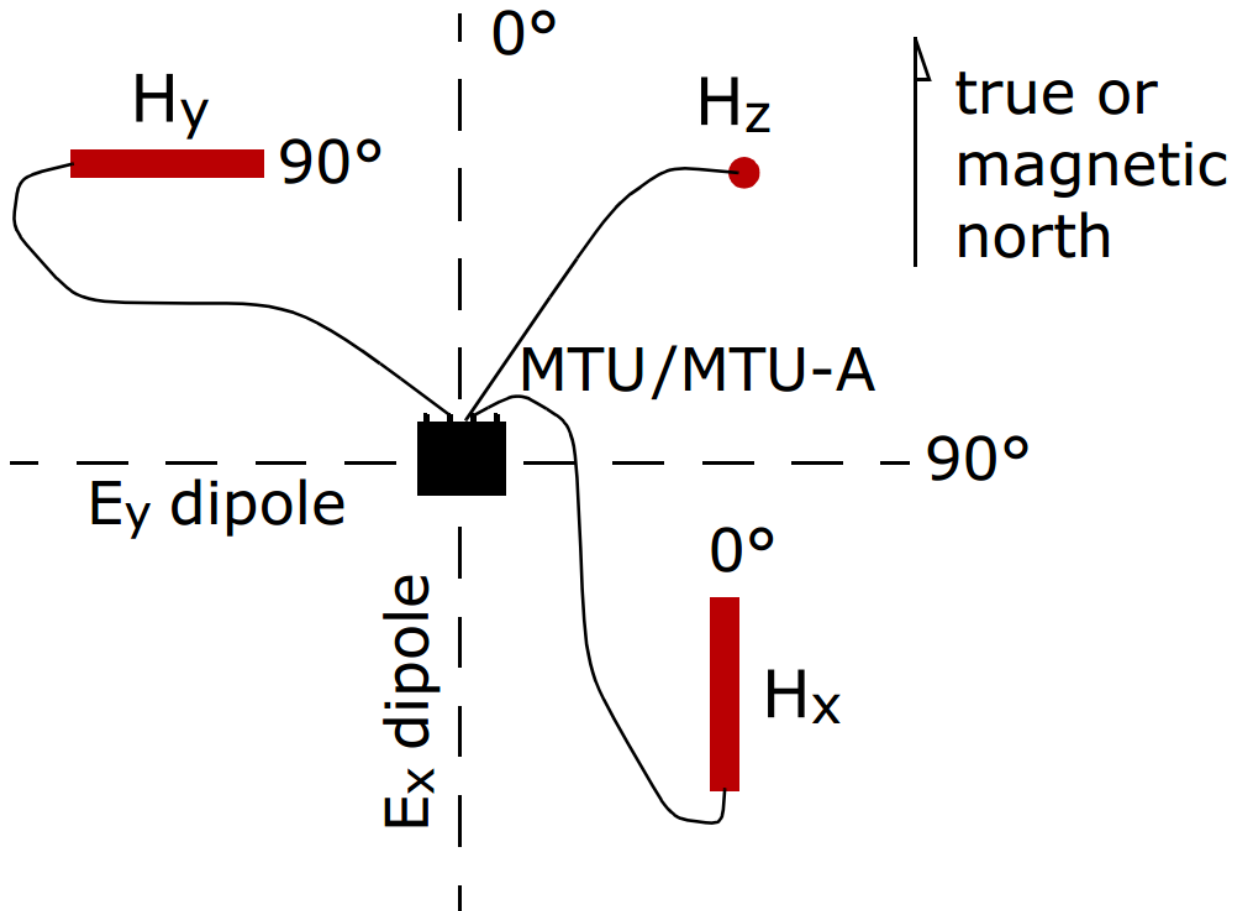


Figure 4.3: Schematic representation of the field set-up for a typical 5-channel MT sounding (adapted from [Phoenix Geophysics Ltd., 2010](#)). E_x and E_y dipoles represent the non-polarizing electrodes for measuring the electric field components. The magnetic induction coils indicated by H_x and H_y measure the horizontal components of the magnetic field, and H_z shows its vertical component. The directions are labeled as x, y, and z, whereas the E and H are the electric field and the magnetic field, respectively. Hence, the acquisition unit records the time-series data of all five electromagnetic field components (i.e., E_x , E_y , H_x , H_y , and H_z).

A total of 212 MT soundings used in this study were collected and recorded by the Geothermal Development Company (GDC) and described in [Lagat et al. \(2010\)](#). The MT team acquired between two to six stations concurrently where each of the stations is independent of each other. At each of this MT station, five measurements (channels) were recorded; the magnetic field in two horizontal directions, one vertical direction, in addition to the electric field in two horizontal directions as displayed in Figure 4.3. The two horizontal measurements are always set perpendicular to each other. The vertical electric field usually is not measured, because it is presumed to be either zero or negligibly insignificant.

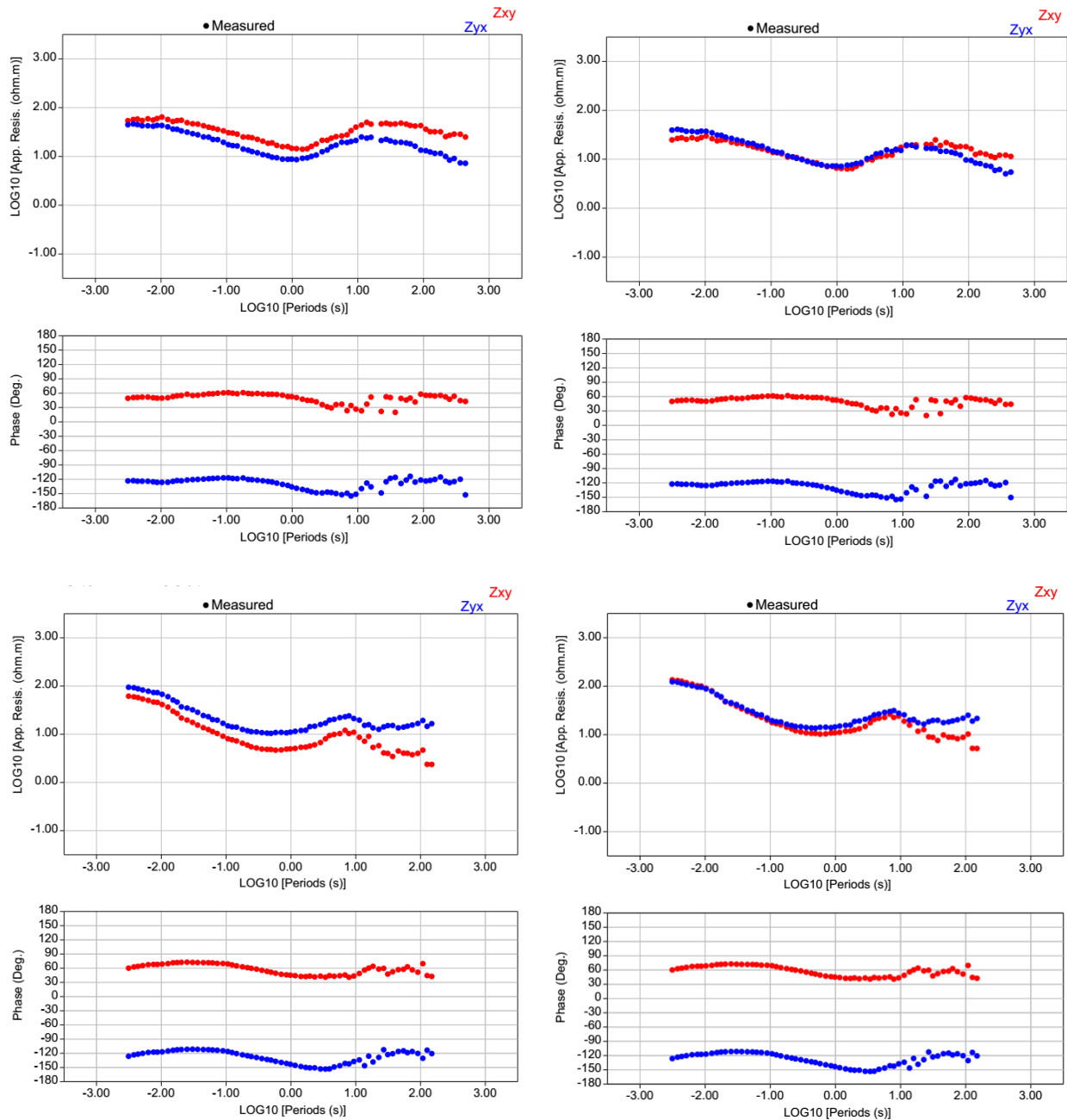
4.4.2. Distortion and static shift in MT methods

The objective of MT data analysis is to develop a suitable dataset for defining the resistivity model of the study area from observed MT data. Near-surface inhomogeneities due to galvanic and inductive effects can cause distortions of the measured electromagnetic fields, which may mask the large-scale structure usually targeted during MT surveys ([Bibby et al., 2005](#); [Chave & Jones, 2012](#); [Jiracek, 1990](#)). Deep-seated resistivity variation produced by changes in lithology and geological structures can as well cause distortion and dimensional effects in the magnetotelluric data.

Static shift effect is a small-scale distortion that causes parallel shifts in the apparent resistivity curve of magnetotelluric sounding and is a common problem in geothermal exploration ([Árnason, 2015](#); [Sternberg et al., 1988](#)). Apparent resistivity curves shift by a constant, real scaling factor (thus the term ‘static’), and therefore the shape of the curve remains unchanged on a log-log plot of apparent resistivity versus frequency since the distortion is frequency independent. The curve can then be translated vertically because the static shift coefficient scales the apparent resistivities (in the y-axis). However, static shift affects only the apparent resistivity curve but not the phase response. It is easy to identify the presence of a static shift in measured data in which apparent resistivities shifts relative to one other, on the other hand, the impedance phase remains unchanged.

Most of the data acquired suffered from some degree of distortions and contamination from noise particularly on the longer periods. It was necessary to discard data that were strongly affected by noise. The distortion removal technique was then applied to the remaining data following the

method by Bibby et al. (2005). Removal of distortion does not necessarily remove static shift, and thus, a spatial median filter with a diameter of 3 km was also applied to remove static shifts, which significantly minimized the static shift problem to an acceptable level on our view. Figure 4.4 shows the apparent resistivity and phase curves before and after removal of both static shift and distortion. Figure 4.5 displays the apparent resistivities curves for all the stations, which are characterized by different starting values but similar structure.



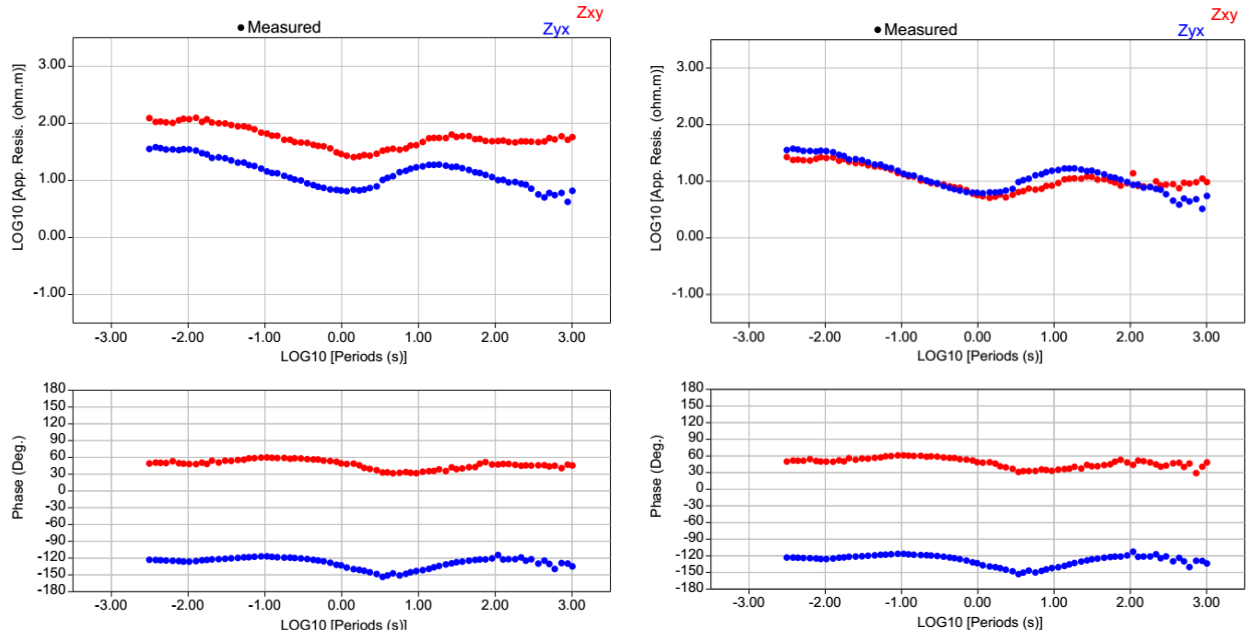


Figure 4.4. Typical magnetotelluric data collected in the study area displayed as apparent resistivity and phase curves. The curves on the left side show the raw data collected while those on the right side display the results after the removal of static shift and distortion. Period acts as a proxy for depth, with short periods sensing shallow structures and long periods sensing deeper structures.

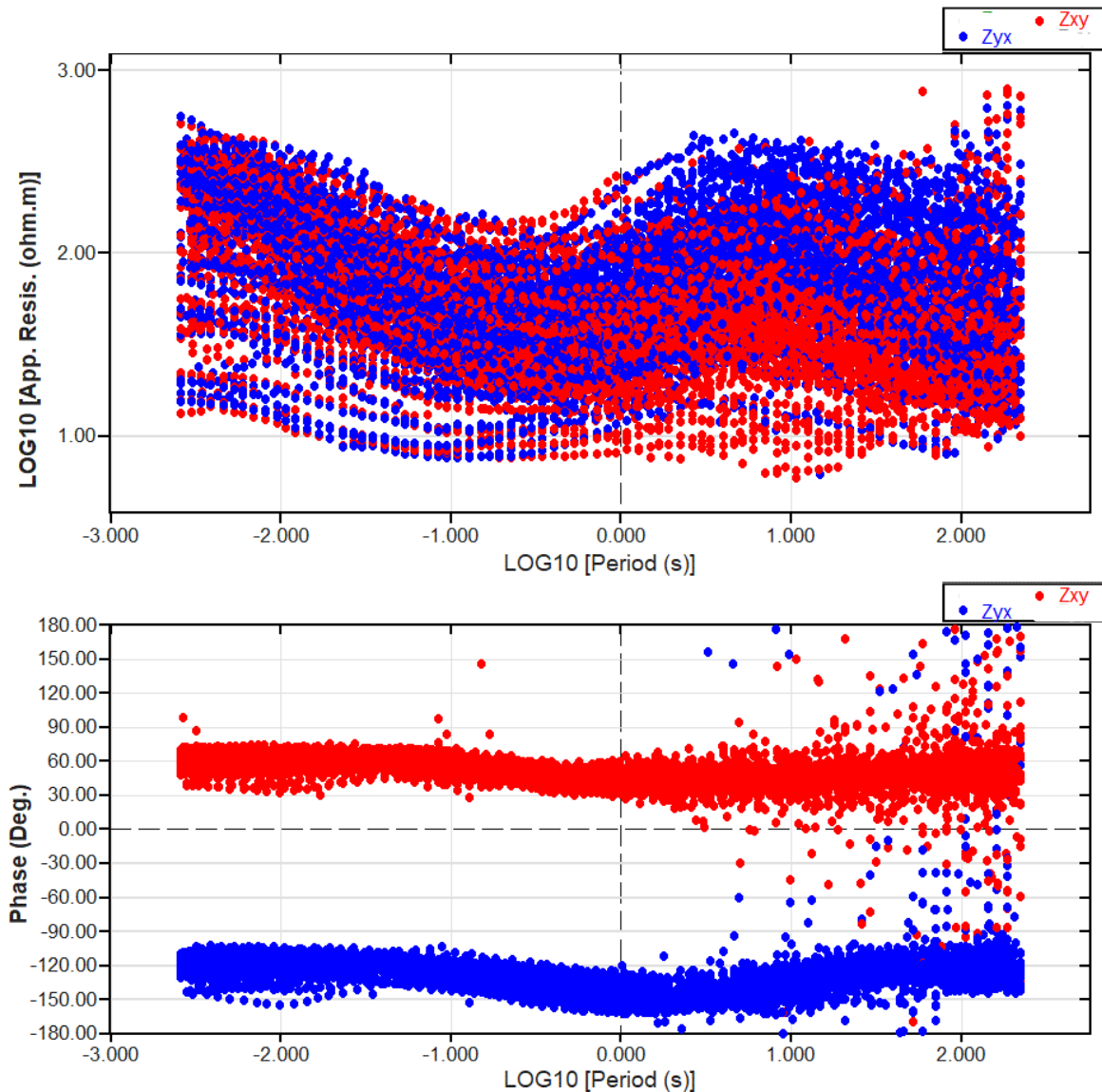


Figure 4.5: Plots of all MT sounding curves of TE and TM modes from the 212 stations in Menengai field. The red dots indicate the TE mode and the blue dots indicate TM mode. The consistent shape of all the curves is displayed. However, the starting resistivity values are different for all stations indicating an inhomogeneous system.

4.4.3. Dimensionality analysis

The results of a chosen modeling approach are strongly influenced by whether the magnetotelluric data has 1-D, 2-D or 3-D characteristics. This implies that the dimensionality of the geological structures in the subsurface should limit the dimensionality of the inversion useful to the

magnetotelluric data (e.g., Bibby et al., 2005; Cumming & Mackie, 2010; Ledo, 2005). Therefore, it is necessary to consider the dimensionality of the MT responses and determine the most appropriate modeling approach.

a). Swift-skew and ellipticity

Swift-skew (Swift, 1967) and ellipticity can be used to show whether response at an MT station is 1-D, 2-D or 3-D. Swift-skew is the ratio of the magnitude of the second invariant to that of the third invariant and defined as

$$Swift - skew = \frac{|Z_{xx} + Z_{yy}|}{|Z_{xy} - Z_{yx}|} \quad [4.9]$$

where Z_{xx} , Z_{xy} , Z_{yx} and Z_{yy} are the impedance tensor elements. This quantity does not depend on the rotation angle of the data or the coordinate system used, and is therefore rotationally invariant (Vozoff, 1991). Impedance ellipticity on the other hand is a quantity that can vary with the direction of the setup and is defined by the ratio of the impedance ellipse divided by the major axis. Ellipticity can be expressed as

$$Ellipticity(\theta) = \frac{|Z_{xx}(\theta) - Z_{yy}(\theta)|}{|Z_{xy}(\theta) + Z_{yx}(\theta)|} \quad [4.10]$$

The zero values indicate perfect 1-D and 2-D structures and values over 0.3 are indicative of 3-D inductive effects (e.g., Bahr, 1991; Lezaeta, 2002), and in some cases, values less than 0.3 are commonly assumed that the data represent the responses of a 2-D structure (e.g., Garcia et al., 1999). This method should, however, be used with caution as it is sensitive to noise (Chave & Jones, 2012; Lahti et al., 2005).

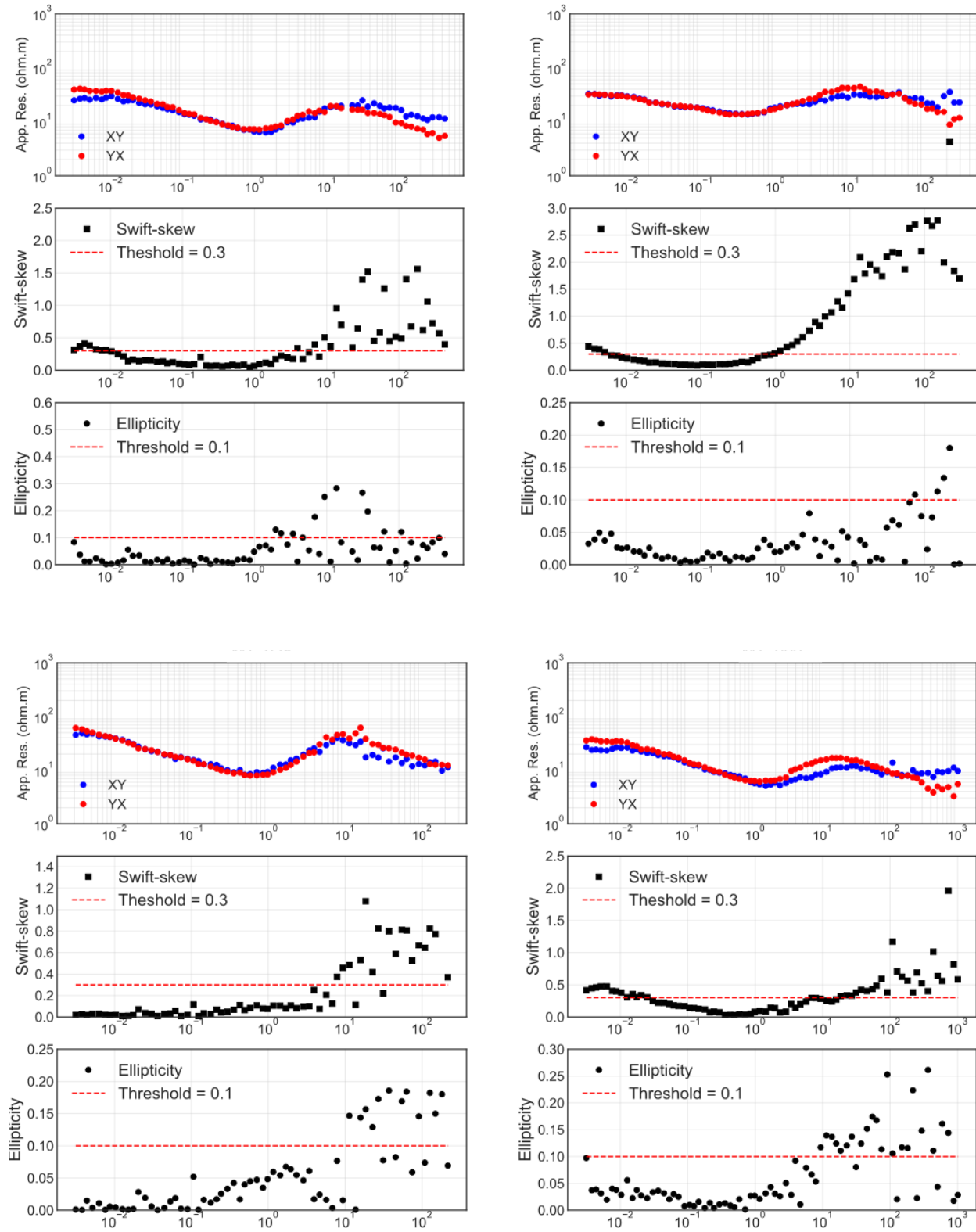


Figure 4.6: Swift-skew and ellipticity are plotted for selected MT stations.

In general, Swift-skew and ellipticity of most of the stations are below the threshold (Figure 4.6). They approach zero at lower periods (< 10 s), and this implies that at shallow depth a 1-D and 2-D dimensionality characteristics are exhibited. On the longer periods (>10 s), high Swift-skew and ellipticity feature observed indicate 3-D structures at deeper depths.

b). Phase tensor

Dimensionality analysis of the impedance tensor is necessary to determine if a 1-D, 2-D, or 3-D approach to data analysis is required. If the data are found to be 1-D, then resistivity only varies with depth. If the data characterize a 2-D condition, then the resistivity varies both with depth and horizontally, across the geoelectrical strike direction. For the 2-D case, a directionality analysis is requisite to determine the geoelectrical strike direction, and there is no defined strike direction when the data are 3-D. For an entirely 2-D condition, the impedance tensor will have diagonal elements equal to zero when the impedance tensor rotates to the geoelectrical strike direction. However, when the data are noisy or not precisely 2-D, which is usually the case, then the diagonal elements may be non-zero, but small and negligible (Jones & Groom, 1993). However, rotating the impedance tensor to a strike angle minimizes the diagonal elements and maximizes the off-diagonal elements, and therefore the Earth can be estimated as a 2-D structure.

The MT phase tensor is a property of the MT impedance that is unaffected by a common form of distortion due to unresolvable local structures and therefore can give a good indication of the dimensionality of a regional conductivity structure. The phase tensor approach (Booker, 2014; Caldwell et al., 2004) is defined by

$$\Phi = X^{-1}Y = \begin{bmatrix} \Phi_{xx} & \Phi_{xy} \\ \Phi_{yx} & \Phi_{yy} \end{bmatrix} \quad [4.11]$$

where X and Y are the real and imaginary parts of Z, the impedance tensor. The phase tensor can be plotted either in map or pseudo-section view as an ellipse defined by the maximum phase (Φ_{max}) and the minimum phase (Φ_{min}), which correspond to the major and minor axes of the ellipse with their directions indicating the two orthogonal electrical principal axes or two possible strike directions, and the skew angle (β). In a 1-D condition, the maximum and minimum phases

are equal and the ellipse is a circle. The phase tensor is symmetric (i.e., the skew angle is equal to zero) for a 1-D and 2-D resistivity structures and is defined by non-zero in the case of a 3-D resistivity structure, and represents the rotation of the major axis of the phase tensor ellipse (Caldwell et al., 2004). Figure 4.7 shows the phase tensor represented as an ellipse.

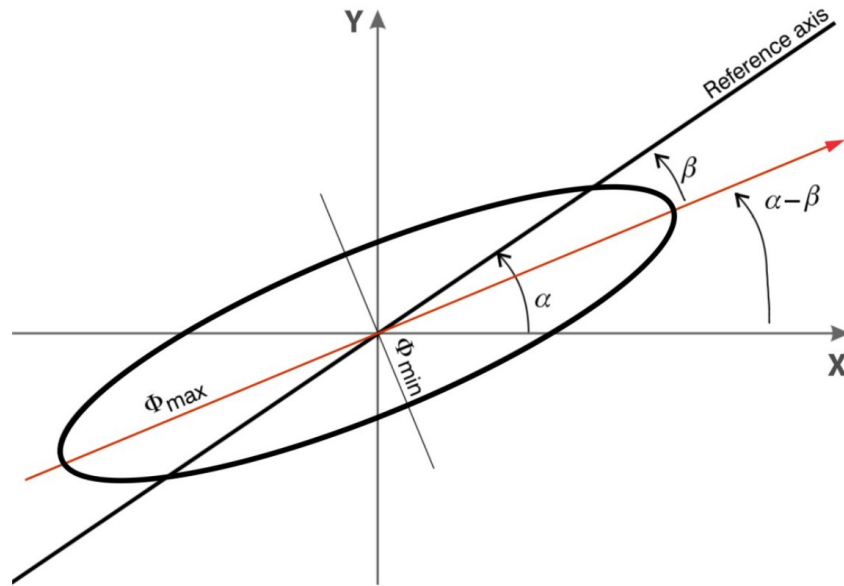
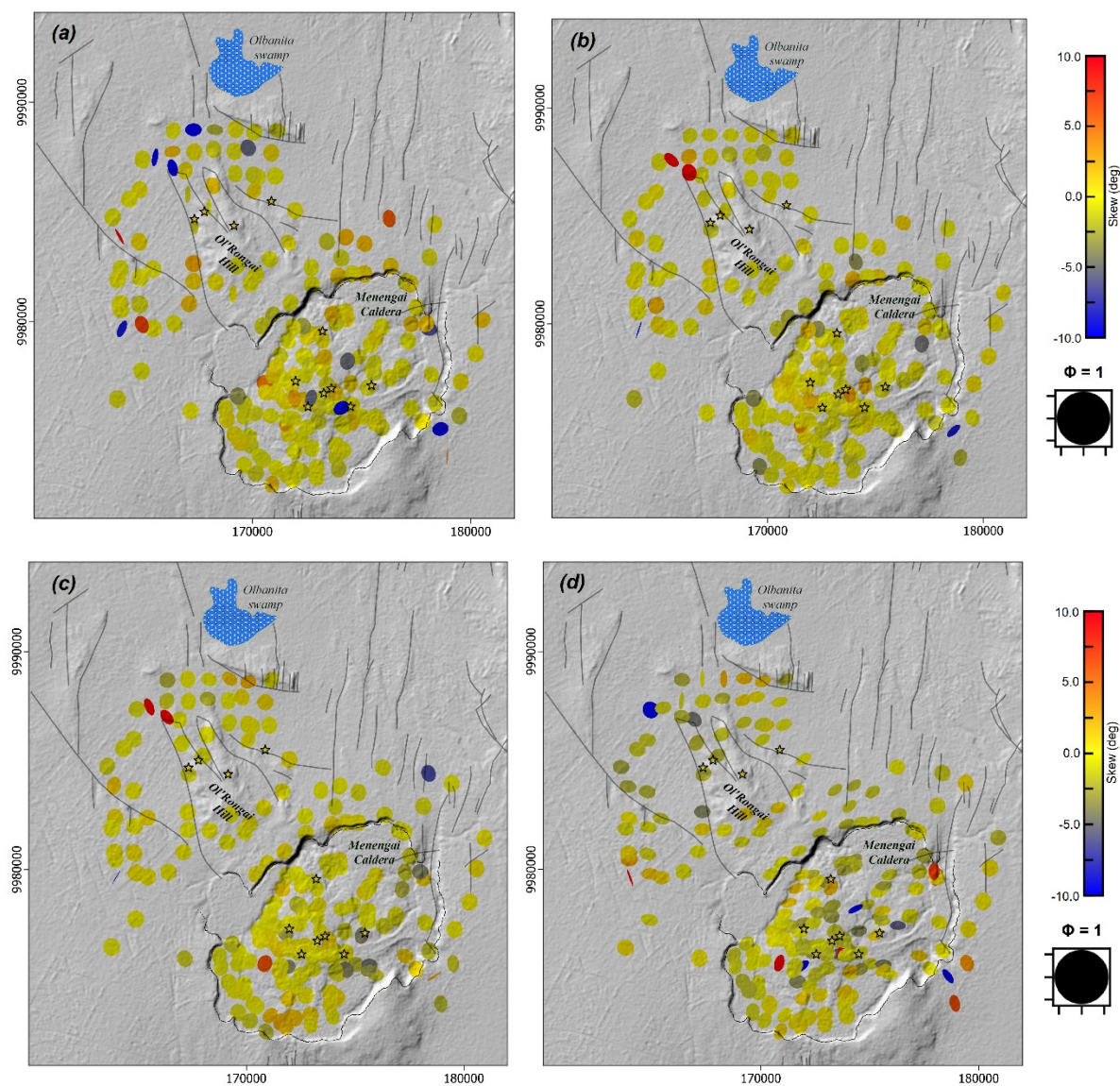


Figure 4.7: Graphical representation of the phase tensor as an ellipse. The magnitudes of the minimum and maximum principal axes are given by Φ_{min} and Φ_{max} , respectively. If the phase tensor is non-symmetric, a third coordinate invariant is required to describe the tensor, which is the skew angle, β . The angle $\alpha - \beta$ gives the orientation of the major axis of the ellipse and defines the link between the tensor and the observational reference frame (x_1 and x_2); adapted from Caldwell et al. (2004).



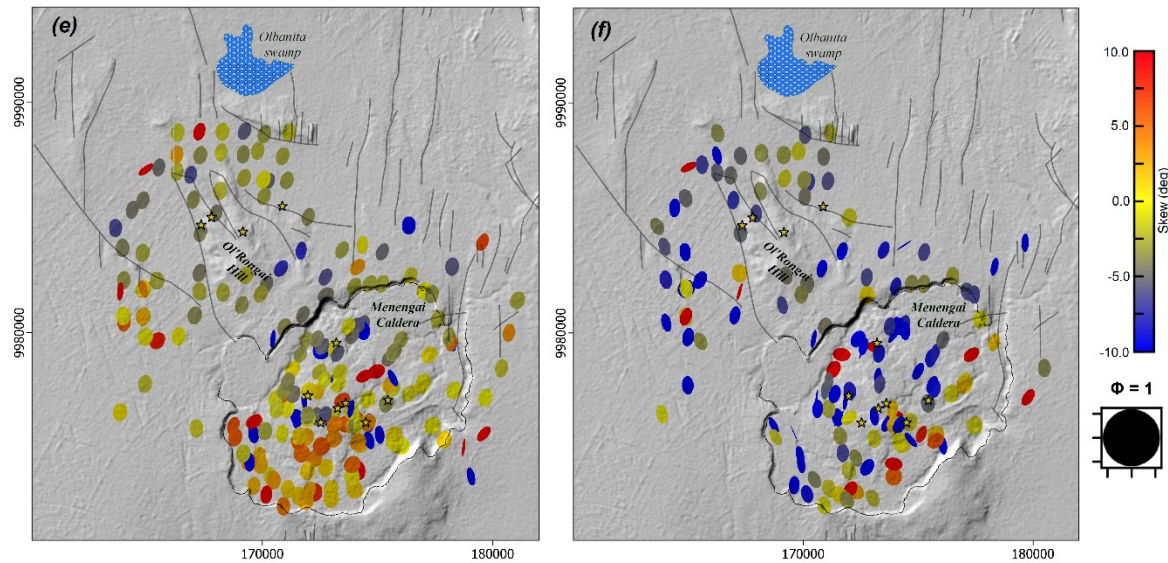


Figure 4.8: Phase tensor maps for periods of (a) 0.0033 s, (b) 0.01 s, (c) 1 s, (d) 10 s, (e) 100 s, and (f) 300 s.

Figure 4.8 displays the phase tensor maps for periods of 0.0033 s, 0.01 s, 1 s, 10 s, 100 s, and 300 s. At short periods, most of the stations are circular and showed relatively small skew angles ($-3^\circ < \beta < 3^\circ$), exhibiting a 1-D characteristics as shown in Figure 4.8a-b, and yet some few stations indicate a 2-D (elliptical with $\beta < 3^\circ$) and 3-D (elliptical with $\beta > 3^\circ$) dimensionalities. However, at longer periods shown in Figure 4.8c-d, most of the stations begin to exhibit a more complex aspect that points to a 3-D dimensionality effect. This suggest that at these periods the presence of a 1-D resistivity structure is no longer supported, and the effects of 2-D characteristics are more localized and therefore sporadically distributed. In general, the phase tensor analysis shows a complex scenario where the 1-D and 2-D dimensionality effects are prominent at shallow depth whereas 3-D characteristics are detected at deeper levels. These insights from dimensionality analysis informed the need for both 2-D and 3-D inversions to adequately image the complex subsurface structures Menengai area.

The plot strike angles are estimated from the invariants of the impedance tensor defined by [Weaver et al. \(2000\)](#). It can be seen that the scattering of these directions is quite spread at the shortest period (see Figure 4.8a), and the most dominant direction cannot be established (although it shows a mean of N15°E). This is probably due to the 1-D dimensionality of the shallow volcanic sediment

layer in the Olrongai area. Conversely, from periods of greater than 0.01 s, the strike directions become more focused, that is to ~N105°E, ~N40°E and ~N10°E, for periods 0.01 s, 10 s, and 300 s, respectively (see Figure 4.8b, d and f). The inconsistency of these strike directions, despite showing a less scatter behavior, relates to the complexity of local structural setting. Each of these directions appears to parallel with individual directions of geological structures closely.

4.5. 1-D and 2-D Occam's inversion of MT data

The 1D inversion method for the MT data was processed using Occam's inversion algorithm. The 1D inversion of electromagnetic sounding data using Occam-type inversion proposed initially by [Constable et al. \(1987\)](#) and shortly after [deGroot-Hedlin & Constable \(1990\)](#) introduced the 2-D Occam's inversion. Since then Occam's inversion has been widely used to interpret MT data and other electromagnetic data. This inversion scheme is an approach to the regularization of nonlinear inversion problems, for the most part, those relating electromagnetic forward solutions known to be virtually nonlinear.

Regularization refers to the use of a penalty on the model, and the penalty is then minimized subject to the constraint that the model predicts the observed data. This inversion scheme yields a smooth model that fits an observed dataset within specific tolerances. By applying smoothness constraint, an inverted model may not produce the best fit to the data, but instead, it remains acceptable. The true least-squares solution in electrical methods produces physically unrealistic models with conductivity contrasts, and therefore a smoothing penalty is necessary ([deGroot-Hedlin & Constable, 1993](#)).

In general, the 1-D MT forward problem that relates the sought after model parameters to the observed data is expressed as:

$$d = F[m] \quad [4.12]$$

where $\mathbf{d}$ is the data matrix, $\mathbf{m}$ is the matrix of model parameters and $\mathbf{F}$ is the forward functional. The estimated model parameters at the $k+1$ th iteration can be obtained by solving the following equation:

$$m_{k+1}(\mu) = [\mu \partial^T \partial + (\mathbf{WJ}_k)^T \mathbf{WJ}_k]^{-1} (\mathbf{WJ}_k)^T \mathbf{Wd}_k \quad [4.13]$$

where μ is Lagrange multiplier, $\mathbf{W}$ is the weighted diagonal matrix $M \times M$, and $\mathbf{J}$ is the Jacobian matrix.

In 2-D Occam's inversion, equation [4.13] above can be expanded as follows

$$m_{k+1} = [\mu(\partial_y^T \partial_y + \partial_z^T \partial_z) + (\mathbf{WJ}_k)^T \mathbf{WJ}_k]^{-1} (\mathbf{WJ}_k)^T \mathbf{W} \quad [4.14]$$

Where ∂_y is roughness matrix for a lateral description of different model parameter whereas ∂_z is the roughness matrix for a vertical description of the model parameter.

4.6. 3-D ModEM inversion of MT data

Application and interpretation of 1-D and 2-D electromagnetic data often result in inaccurate conclusions regarding the boundaries of the geothermal reservoir particularly in deeper parts hence the 3-D model becomes more suitable (Spichak & Manzella, 2009). The need for 3-D resistivity modeling makes the removal of 3-D topographic effects possible through tensor stripping since the geometry can be obtained. It is also useful for the evaluation of regional 3-D effects, adequate in testing the applicability of distortion removal schemes and proper handling of 3-D near-surface effects (Jiracek, 1990).

In this study, we inverted the MT data to image the 3-D resistivity structure of the study area using ModEM inversion code (Egbert & Kelbert, 2012). This modular system includes a variant of the master-worker parallelization over forward problems after the scheme of Meqbel et al. (2009), for

frequency-domain EM geophysical data. The 3-D forward scheme is constructed on finite differences while the inverse scheme uses the nonlinear conjugate gradient method. The model regularization is applied in a flexible manner with the ability to select different sets of smoothing and trade-off parameters. The implemented 3-D inversion in ModEM is based on the minimization of the penalty functional,

$$U(\mathbf{m}, \mathbf{d}) = (\mathbf{d} - F(\mathbf{m}))^T \mathbf{C}_d^{-1} (\mathbf{d} - F(\mathbf{m})) + \lambda (\mathbf{m} - \mathbf{m}_0)^T \mathbf{C}_m^{-1} (\mathbf{m} - \mathbf{m}_0) \quad [4.15]$$

where the first term in the right side of the equation describes the data misfit between the measured ($\mathbf{d}$) and forward model response, $F(\mathbf{m})$, whereas the second term presents the model update between the prior model ($\mathbf{m}_0$) and the estimated model ($\mathbf{m}$). $\mathbf{C}_m$, $\mathbf{C}_d$, and λ are the model covariance matrix, the data covariance matrix, and the regularization parameter, respectively. Minimization of the penalty functional in ModEM is achieved using the nonlinear conjugate gradient algorithm (NLCG), which introduces the line search process that has a small memory usage by avoiding explicit computing of sizeable Hessian matrix and the storage of Jacobian matrix. However, the 3-D forward component is based on the finite difference method.

The MT data used formed two clusters, one centered at the caldera and the other within Olrongai area. The simulation was carried out focusing on the two areas separately to save on computation time and required resources. A dense network of 118 MT sites both on and around Menengai caldera occupying an area of 11×11 km was chosen for the first 3-D inversion. The model constructed for the inversion process was composed of 55 × 55 × 48 cells in the x, y and z directions, respectively, with a total of 145,200 cells, where the x and y cell dimensions are taken to be 200 × 200 m. The other dataset was composed of 48 MT sites acquired from Olrongai section of the study area. These data spanned an area of 8.8 × 9.9 km. The model constructed for this area had also consisted of the same number of cells (55 × 55 × 48 cells in the x, y and z directions, respectively), adding up to 145,200 cells. The cell dimensions, however, was taken as 160 × 180 m to accommodate the scope occupied by the MT sites.

In both models, a total of 48 layers were used in the vertical grid with the topmost-layer thickness of 10 m and set to increase subsequently by a factor of 1.2. An additional 10 padding cells with an increasing cell-size factor of 1.3 were affixed to the mesh borders in both ways of the x and y directions for an accurate model boundary conditions, and this enlarged the total number of cells to 270,000 (dimension of 33.16×33.16 km and 29.84×29.84 km for the caldera and Olrongai area, respectively). The full impedance tensor was inverted in the frequency range of 320 Hz to 0.00034 Hz for a total of 80 frequencies, and the starting model for both was a homogenous, isotropic half space of 50 Ω .m. The initial RMS at the start of the inversion of caldera portion was 23.3, decreasing to 3.69 after 106 iterations, and the Olrongai sector had a starting RMS of 58.4, reducing to 4.78 after 88 iterations. The data misfit of the final models was considered satisfactory (see Figure 4.9 for some of the results).

4.7. Results and Discussion

The curves in Figure 4.9 show a good fitting of data between the measured and predicted. All the stations display a similar resistivity trend with high value (unaltered zone) followed by a low value (clay zone) then the high value (propylitic zone) and finally a low value interpreted as the heat source.

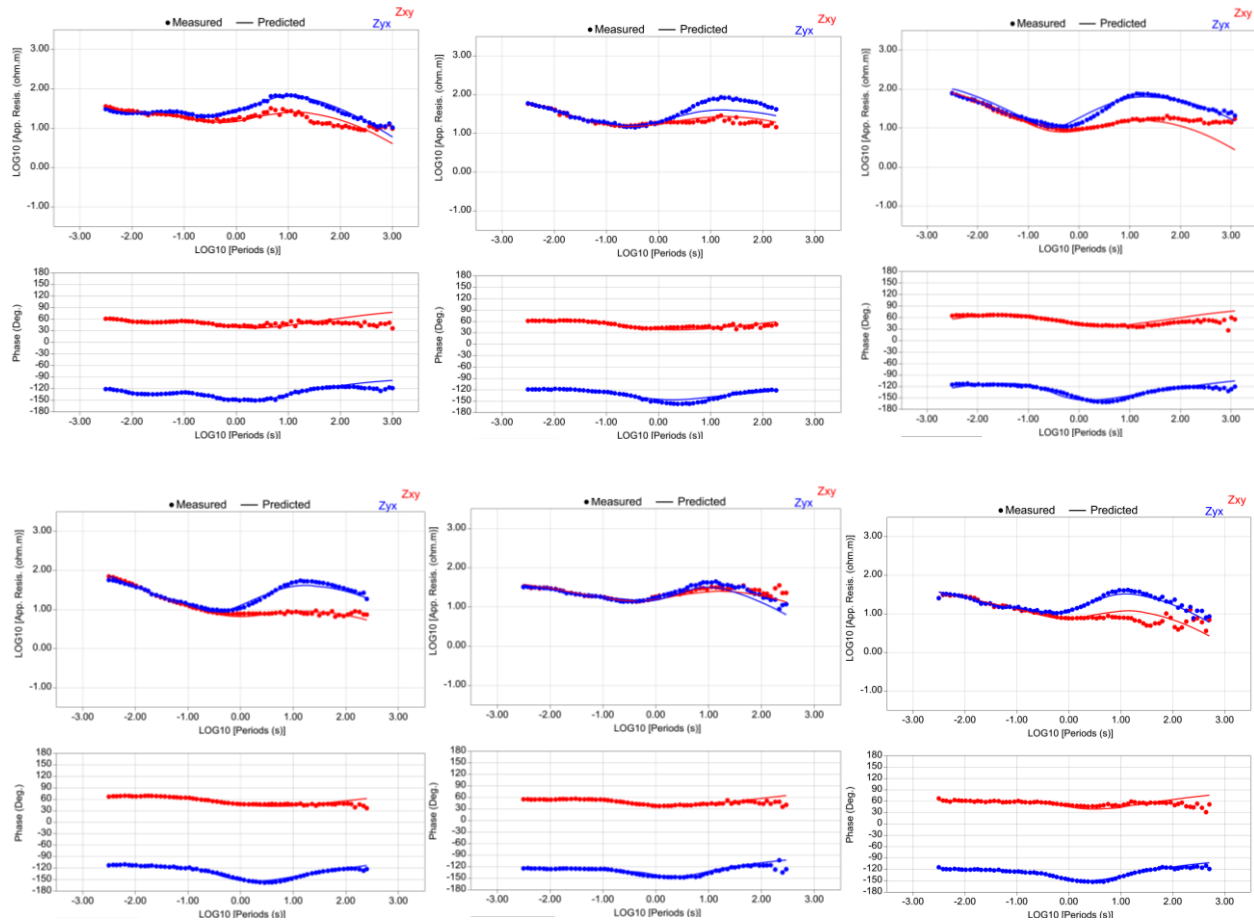


Figure 4.9: Example of data misfit between observed and modeled station responses of the Z_{xy} (red dots) and Z_{yx} (blue) components for some selected stations. The dots and the lines represent the measured values and modeled response, respectively.

Resistivity distribution maps based on both 3-D ModEM, and 1-D and 2-D Occam's inversions were prepared. Maps were generated and displayed as contour and color filled maps of varying values related to a given parameter at a project station. Gridding was achieved using spatial interpolation technique Inverse Distance to a Power, a weighted average method which assumes that contiguous sites are more alike than those located farther from each other. Inverse Distance is a simple method that estimates the variable of interest by allocating more weight to closer sites. Despite their simplicity, the Inverse Distance estimator proves to be quite sensitive to the type of data characteristics (for instance displaying skewed or unskewed distribution, isotopic or anisotropic occurrences, among others), to the number of adjacent sites used in the approximation, and to the exponent of distance used in weighting (Weber & Englund, 1992, 1994).

4.7.1. MT inversion results

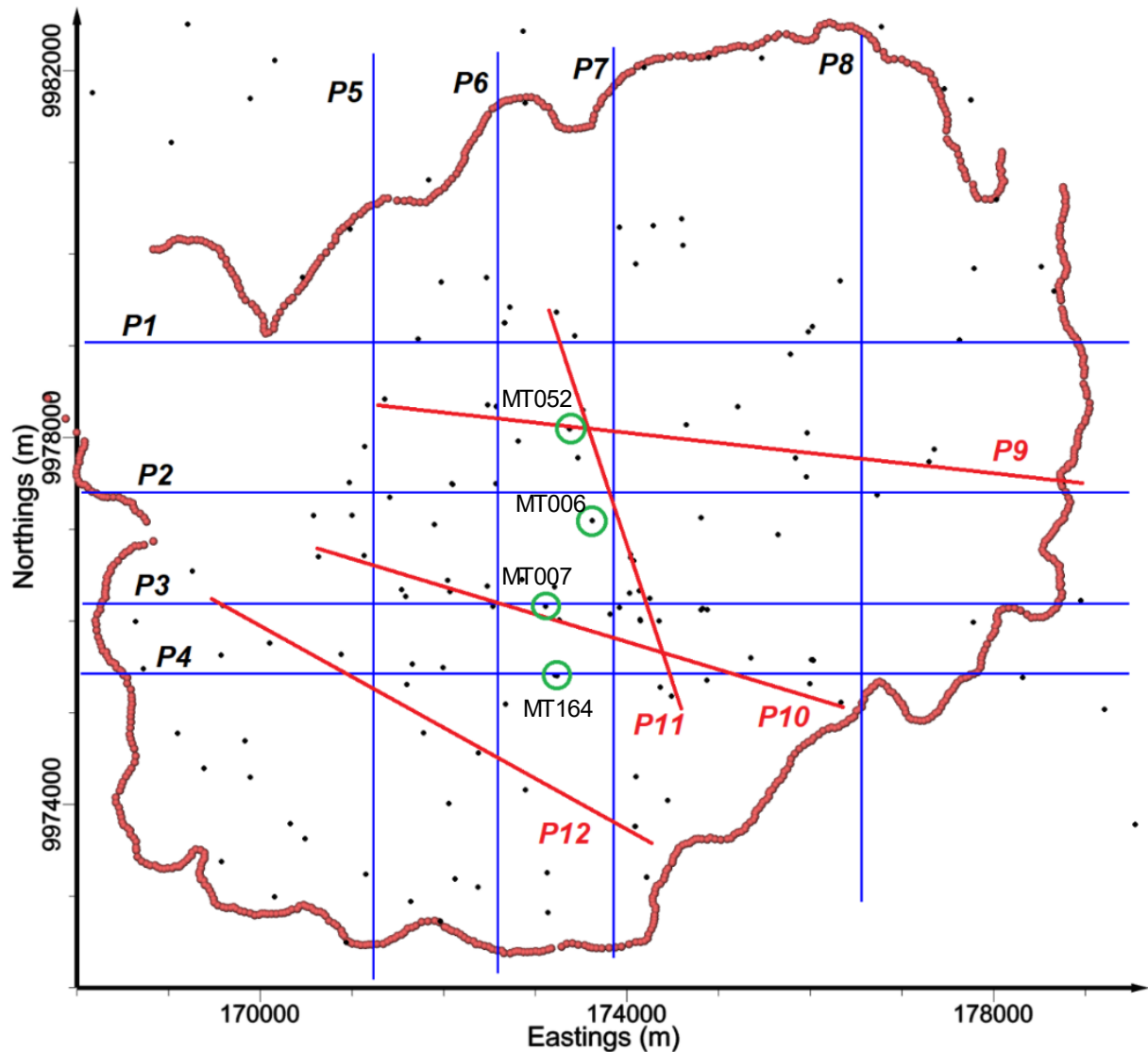


Figure 4.10: Menengai caldera map showing 3-D MT inversion profiles (in blue) along E-W directions (P1-P4) and S-N direction (P5-P8), and 2-D MT inversion profiles (in red) P9-P12. Four MT stations selected for 1-D Occam's inversion are encircled in green. The orientation of 2-D profiles was chosen to ensure it crosses near as much data as possible. The black and orange dots show the MT stations and the caldera rim, respectively.

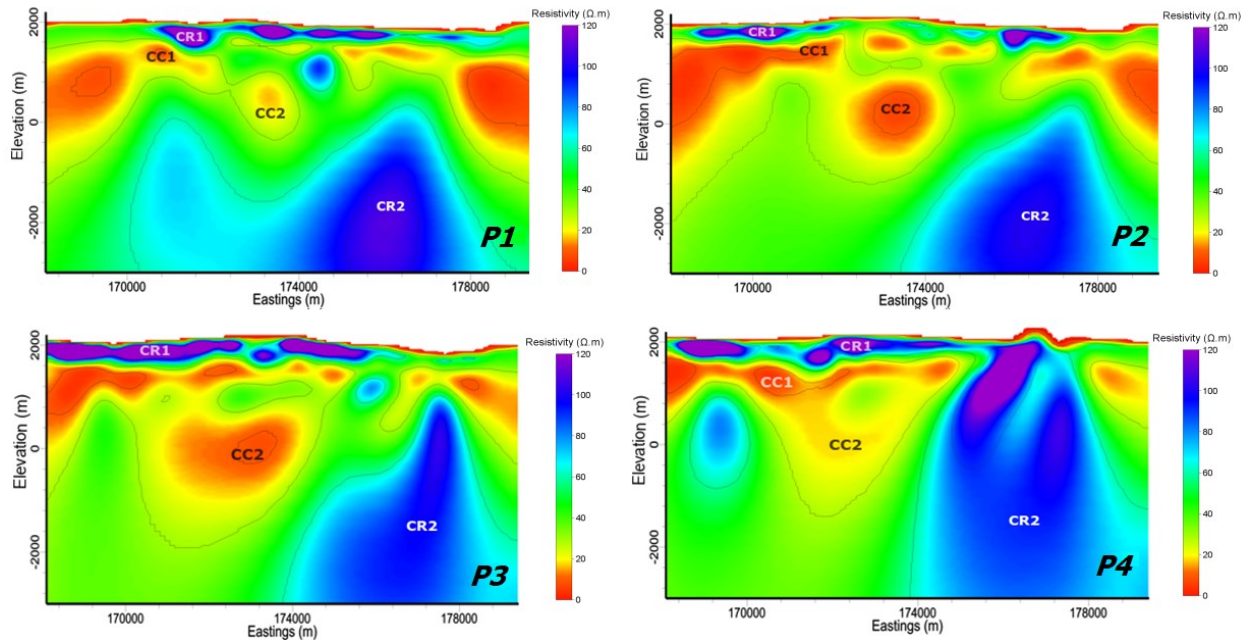


Figure 4.11: MT profiles along W-E directions (P1-P4).

The profiles in Figure 4.11, from the trends in Figure 4.10, aligns from north-south in the order of P1, P2, P3, and P4, and since they lie parallel to each other, they tend to illustrate related resistivity discontinuities. From these profiles, some important resistivity structures are identifiable. At the shallower level, an extensive resistive layer CR1 ($>100 \Omega.m$) is seen and interpreted as unaltered formation. This resistive layer is evident in profiles P1, P3, and P4 but less prominent in profile P2. Profile P2 cuts across the highest elevation within the caldera floor. The prominence of CR1 in some parts of the caldera and not the summit may have been caused by a continued erosion that resulted in the removal of the resistive layer from the caldera summit or by a preferential flow and deposition of resistive lava to the lower part of the caldera floor. The rise of hot magmatic fluid might have also led to a relatively higher rate of hydrothermal alteration of the resistive layer.

Another conspicuous feature is the conductive body CC1 ($<30 \Omega.m$) recognizable in all the four profiles shown as pockets of different shapes and sizes and interpreted as a clay cap. The irregular placed clay cap is expected of such system which has undergone several episodes of volcanic activity, up to 70 different eruptions (Leat, 1984), that led to the formation of an inhomogeneous system. The resistive body CR2 ($>100 \Omega.m$) seen on the eastern section of these profiles can be construed as an impervious body that does not allow water to permeate.

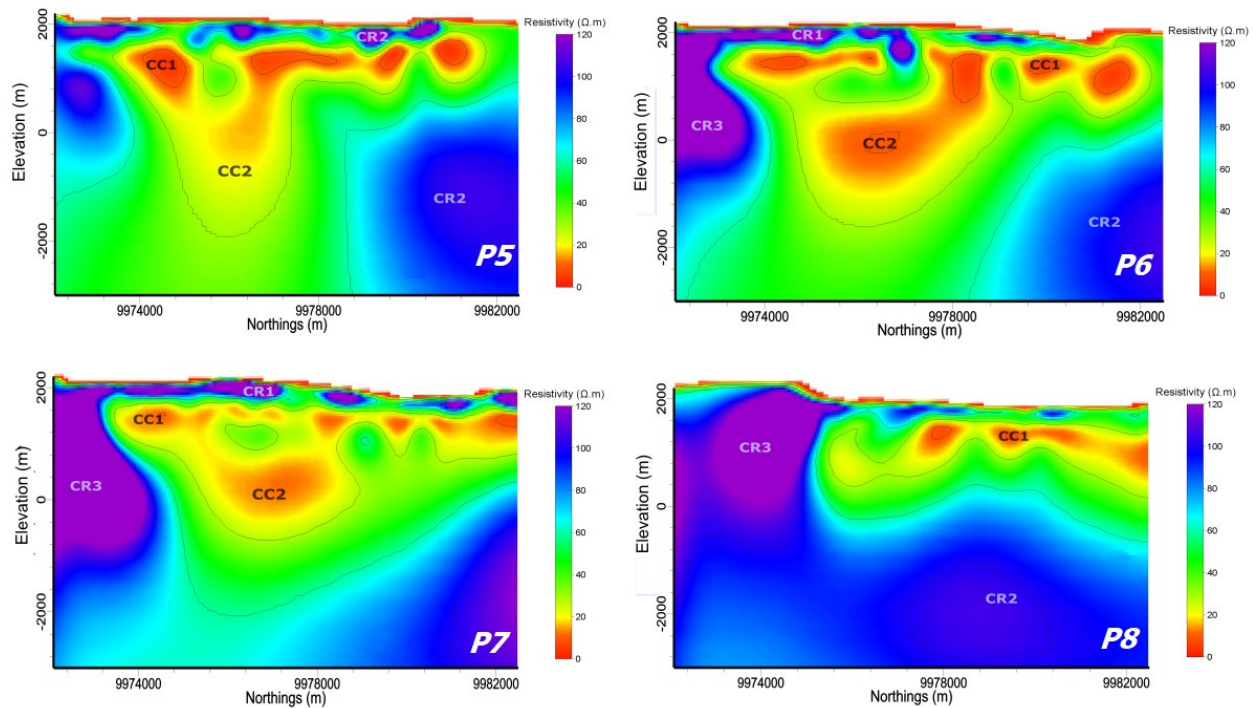


Figure 4.12: MT profiles along S-N directions (P5-P8).

The S-N trending profiles (Figure 4.12) shown in Figure 4.10 are arranged from west to east and follow the order of P5, P6, P7, and P8. Similarly, these profiles identify some prominent resistivity structures. The high resistive body on the southern segment of these profiles, CR3, observed particularly on P6, P7, and P8, coincide with the caldera wall illustrated by the topography change. This section of the caldera wall forms part of the pre-caldera collapse shield volcano most prominent in the southern flanks of the caldera (Leat, 1984). The other resistive body (CR2) interpreted as an impervious formation is observed to the east of this profile and underlies the conductive layer CC1. CC1 forms a consistent layer (< 1000 m a.s.l) across the caldera floor and appears to connect to a much deeper conductive body (CC2). As observed in Figure 4.11, the distribution of the near-surface high resistive formation (CR1) within the caldera is to a certain extent erratic.

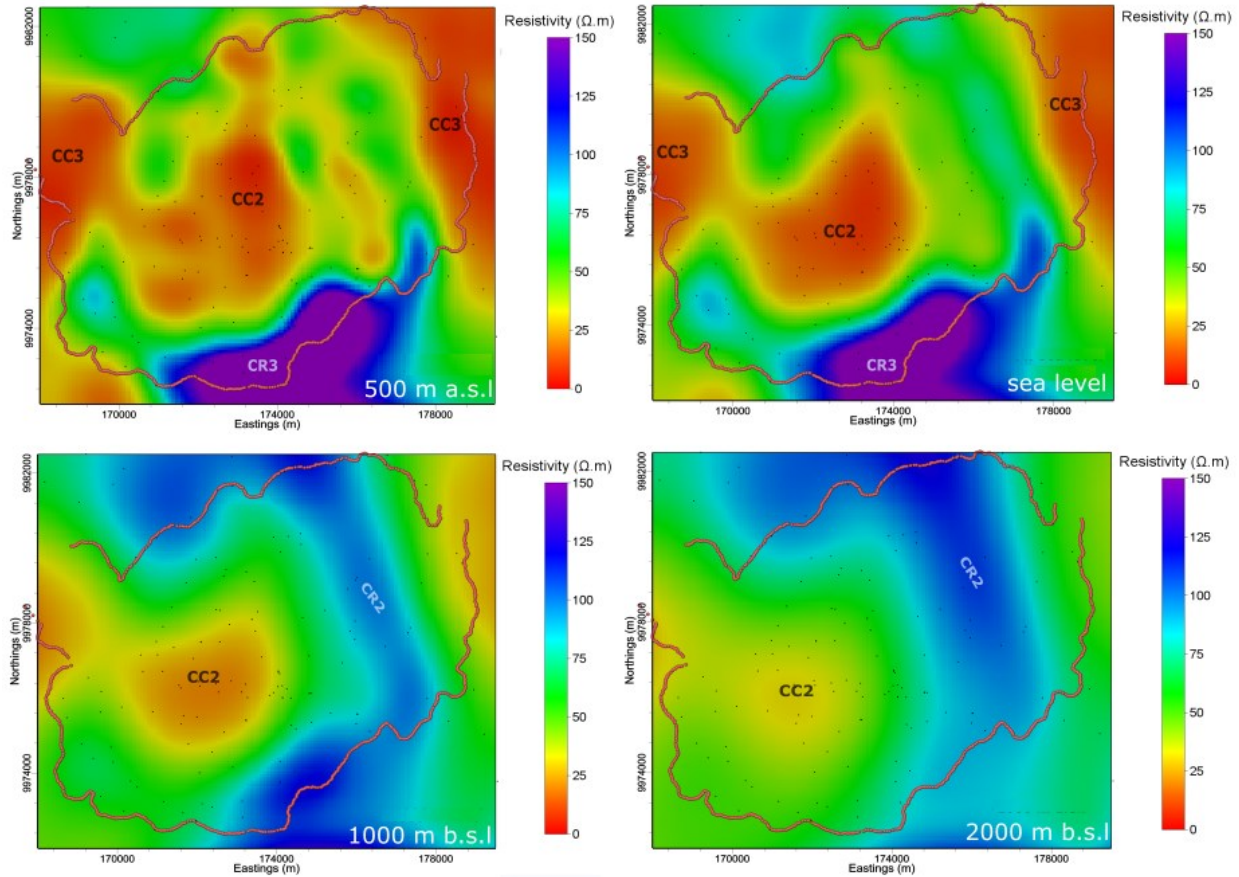


Figure 4.13: Horizontal slices at different heights above sea level through the 3-D resistivity model derived from inversion of the caldera MT dataset. The black and orange dots show the MT stations and the caldera rim, respectively.

Figure 4.13 shows horizontal slices of Menengai caldera obtained from the 3-D resistivity model at 500 m b.s.l, sea level, 1000 m b.s.l and 2000 m b.s.l. Within the caldera, there is a consistently massive conductive body (CC2) seen in all the profiles which is interpreted as shallow partial melt that provides the heat for the geothermal system. The conductive body is surrounded by a loop-shaped resistive zone (CR2), which crosses the caldera in the eastern segment and appears to separate the conductive anomaly from another conductive zone (CC3). CC3 extends outside the eastern rim of the caldera and is more conspicuous on elevations above sea level. However, only a few data points support the CC3 anomaly (both the eastern and western flanks of the caldera), and may not be useful in making important inferences.

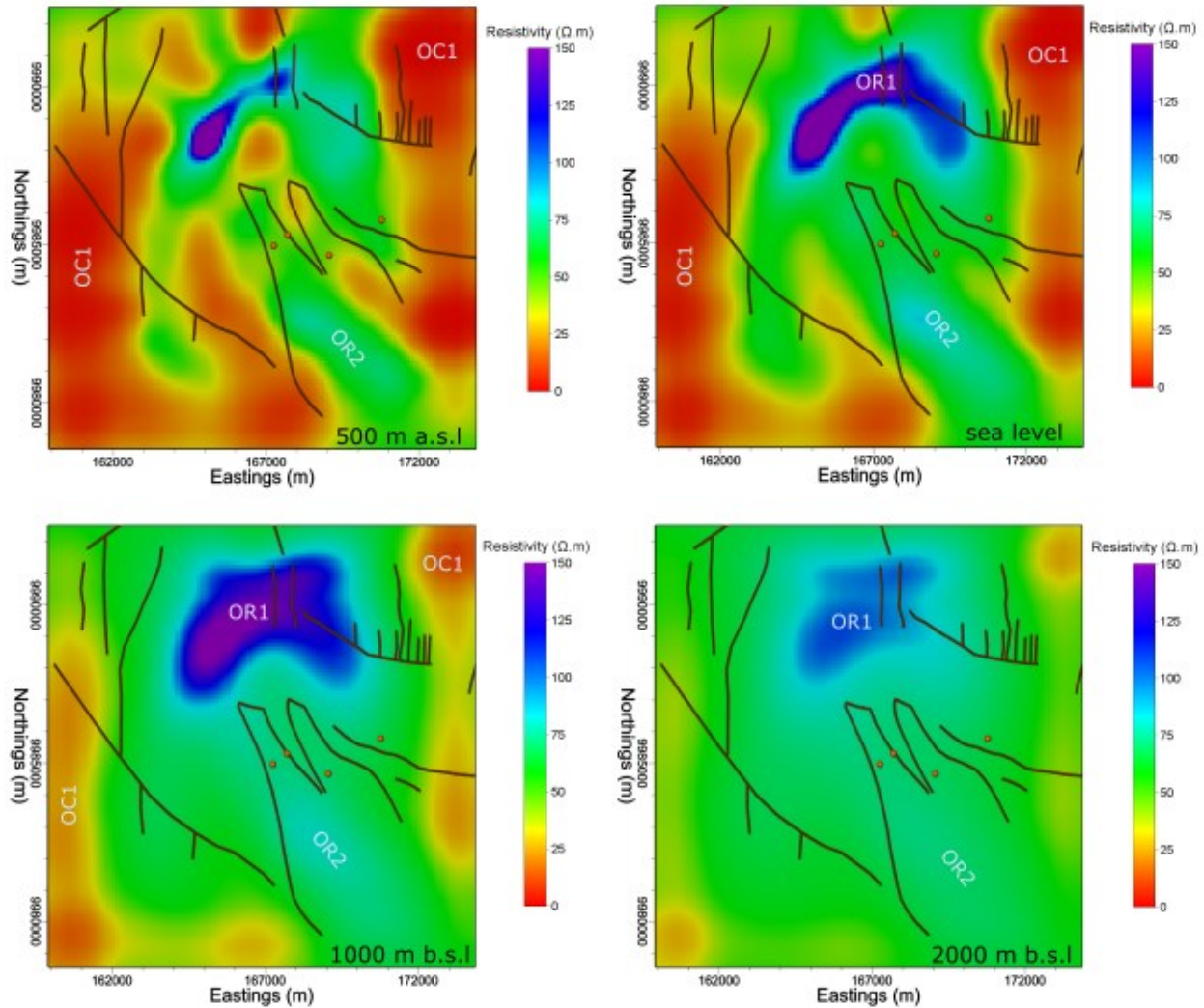


Figure 4.14: Horizontal slices at different heights above sea level through the 3-D resistivity model derived from inversion of Olrongai area MT dataset. The orange dots and brown lines denote fumaroles and faults, respectively.

Figure 4.14 shows four map view slices of Olrongai area at elevations that spans between 500 m a.s.l and 2000 m b.s.l. At higher elevations (the profile at sea level and 500 m a.s.l), a low resistive body (CR1, with $<30 \Omega.m$) situated to the east and west of the start area is quite dominant. A curved body of high resistivity (OR1, with $>120 \Omega.m$) lies towards the northern segment of the study region becoming more pronounced at sea level and elevation of 1000 m b.s.l. Similar structure (OR2) forms in the south-east from heights as high as from 500 m a.s.l., but fades away

at 2000 m b.s.l. At these deeper depths, the dominance of the low resistivity segments becomes more peripheral and increases away from the faulted zones.

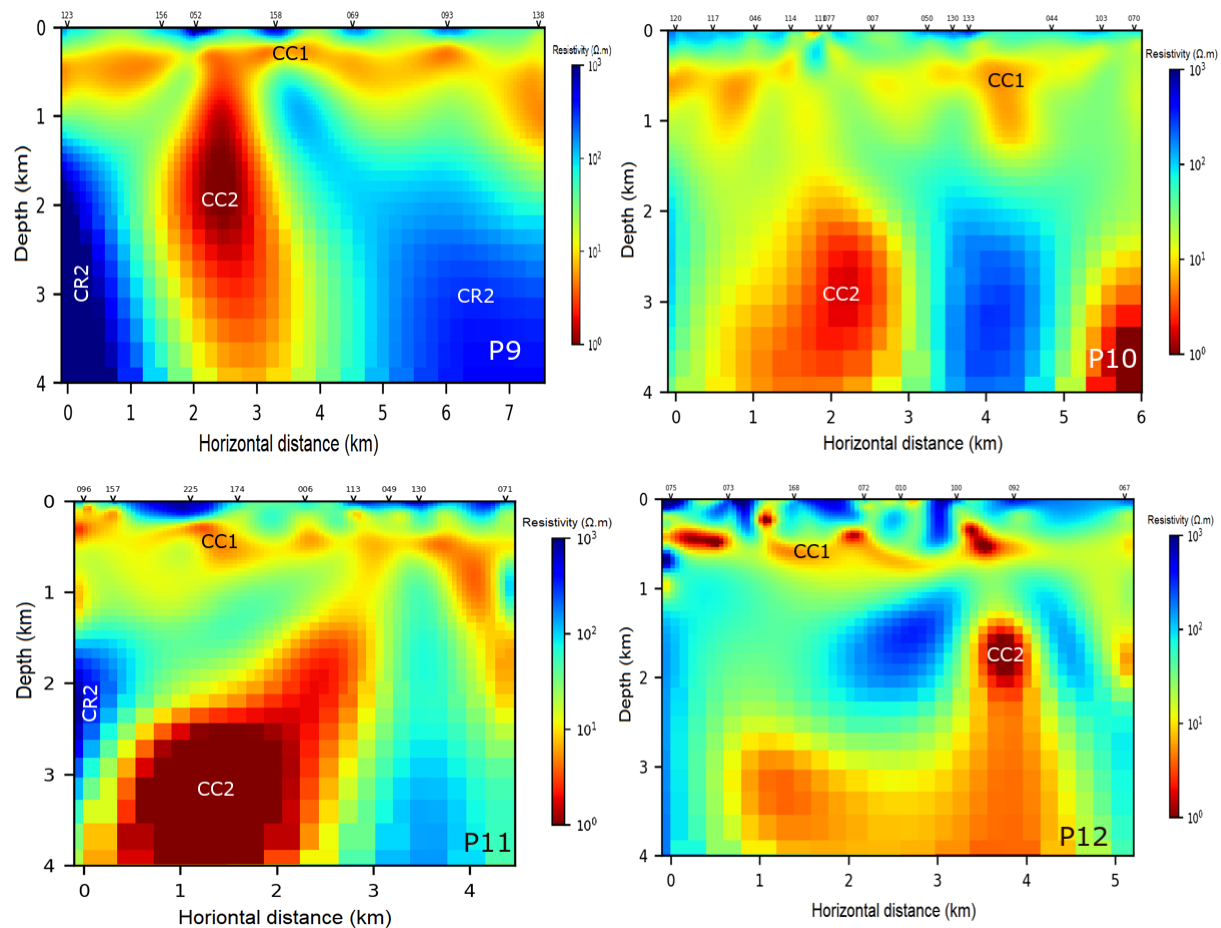


Figure 4.15: 2-D resistivity models for profiles P9, P10, P11 and P12 of Menengai caldera as shown in Figure 4.10.

A 1-D and 2-D profiles of Menengai caldera were plotted and analyzed to confirm the resistivity structures observed in the 3-D models. Figure 4.15 shows the 2-D resistivity models with the reference point being at the surface with no consideration of changes in topography. The Occam's 2-D inversion was used to obtain these models derived along Profiles P9, P10, P11, and P12 of Figure 4.10. All these profiles trend almost WNW-ESE direction apart from P11, which trend in the NNW-SSE direction. The orientation of 2-D profiles was chosen to ensure they constrain the major resistivity structures detected in the 3-D models and also target areas with closely spaced

MT stations for better inversion results. These profiles clearly show the main resistivity features importantly illustrated by the dike-shaped conductive structure CC2 and the resistive body CR2 features. These 2-D inversion models complement the results and interpretation made from the 3-D models.

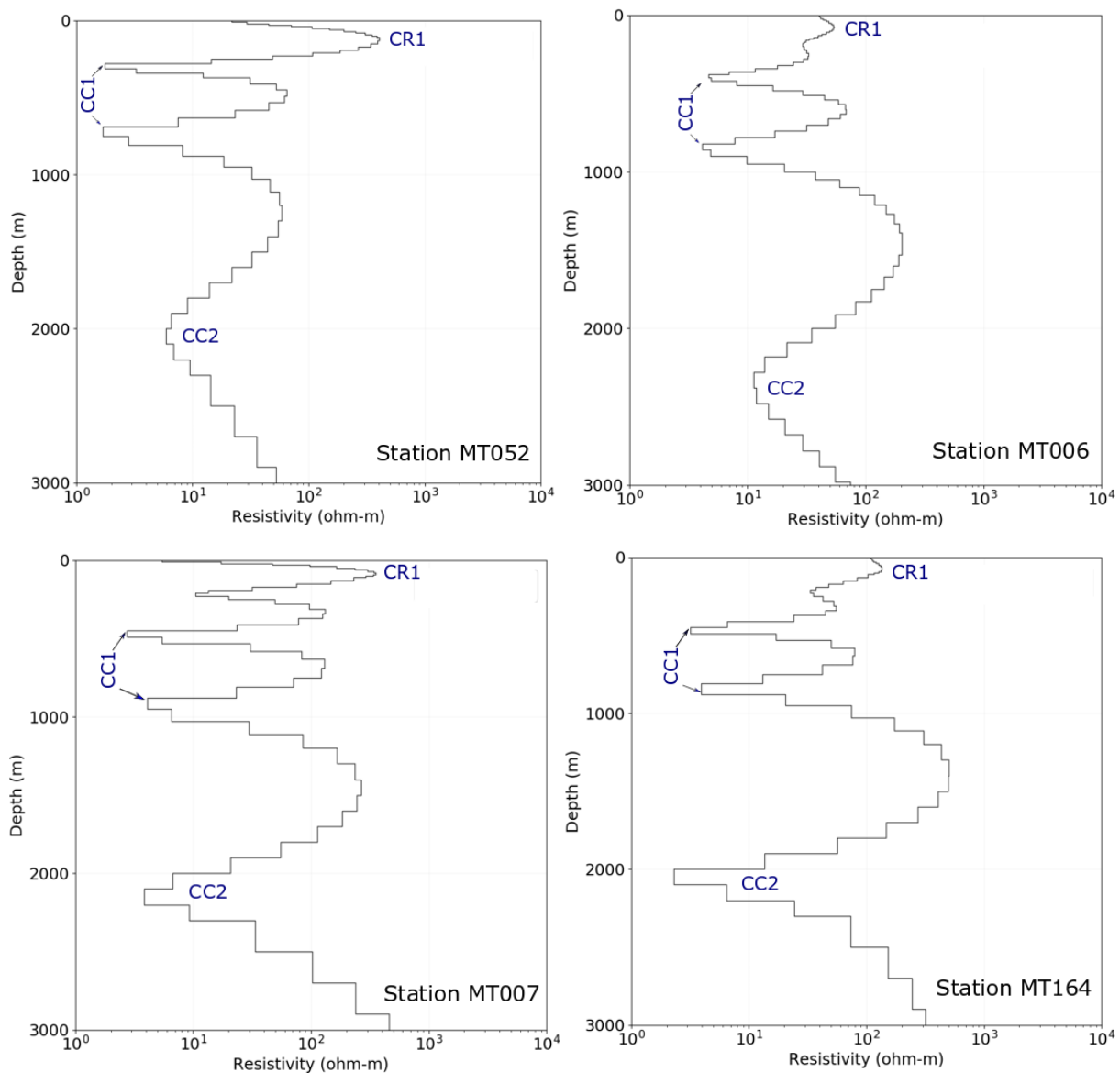


Figure 4.16: 1-D resistivity models of the caldera area from MT stations 052, 006, 007 and 164 align from north to south (see Figure 4.10), respectively, and situated right above the low resistive body (CC2).

The four stations in Figure 4.10, selected for the 1-D resistivity inversion are located vertically above the deep conductive anomaly and were meant to reaffirm the anomalies observed in 3-D (Figure 4.11-Figure 4.13) and corroborated by 2-D models (Figure 4.15). These profiles show similar resistivity trend with a consistent conductive body of less than 10 $\Omega\cdot\text{m}$ at the depths of about 1500-2400 m below which the resistivity increases gradually. The near-surface conductive body (CC1) generally occurs in depths of < 1000 m, as two distinct layers; at average depths of 400 m and 900 m for the lower the upper and lower CC1, respectively.

4.8. Conclusion

This chapter presents the 1-D, 2-D, and 3-D resistivity models of the MT data. Two 3-D resistivity models were constructed separately for the caldera and Olrongai areas. The caldera field was also handled independently for the 1-D and 2-D Occam's inversion. The use of 1-D and 2-D models were to help constraint the distribution of resistivity features within the study area. The three models were able to highlight the dominant geophysical structures which appear to control hydrothermal processes within the geothermal system. The 1-D resistivity model (Figure 4.16), 2-D resistivity model (Figure 4.15) and 3-D resistivity model for Menengai caldera (Figure 4.11-Figure 4.13) show similar major resistivity features.

The 3-D resistivity models of Menengai and Olrongai area show three structures: (1) a shallow surface inhomogeneous resistive layer ($>100 \Omega\cdot\text{m}$); (2) a near-surface conductive layer ($<10 \Omega\cdot\text{m}$), and; (3) a deeper resistive body ($30-70 \Omega\cdot\text{m}$) interpreted as the geothermal reservoir. The near-surface conductor is confined above 1 km above sea level symbolized by low resistivity ($< 10 \Omega\cdot\text{m}$) and interpreted as the cap rock. A hydrothermal alteration likely causes it in a shallow aquifer within the trachytic rocks intercalated by tuffaceous formation. Additionally, two distinct resistivity features are also observed; a dike-like conductive body that cuts through the deeper resistive layer at the heart of caldera, and a loop-shaped high resistive body that is observed in both the caldera and Olrongai areas. The near-surface conductive layer causes the upper-minimum in the apparent resistivity curves while the deeper conductive body, interpreted as a shallow cooling pocket of partial melts is responsible for the lower-minimum of the apparent resistivity curves in Figure 4.9.

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CHAPTER FIVE

ISOTOPE AND WATER CHEMISTRY STUDIES

5.1. Introduction

The primary goal of geochemical exploration in geothermal is to obtain the subsurface composition of the fluids in the system and to use the composition to get evidences on temperature, origin, and flow direction that will help locate the geothermal reservoir. Equilibrium speciation is achieved using speciation programs and simulation of processes such as boiling and cooling to get more information to predict potential deposition and corrosion. Environmental effects can be predicted, and the general information used to contribute to the model of the geothermal system. This chapter presents the chemical analysis and interpretation of major elements, stable and helium isotopes stable as well as the construction and analysis of 1-D reactive transport model.

5.1.1. Isotope geochemistry

The application of isotope techniques in hydrological studies of geothermal systems spans more than five decades, and as a result, many significant contributions have been made worldwide to the development of geothermal energy. The practical applications of isotope methods in the exploration of geothermal resources are dated back to 1970s, and widespread applications reported in the late 1980s and early 1990s ([Pang, 2005](#)). Isotope techniques are virtually indispensable tools in geothermal exploration, and since isotope ratios are sensitive not only to temperature but also to natural physical and chemical processes such as interaction of water and rock, fluid mixing, and steam separation, they can be resourcefully used to trace the origin of geothermal fluids and also processes undergone by the fluids at subsurface. Isotope techniques have also been useful in the estimation of reservoir temperature of geothermal reservoirs and the subsurface residence time of

waters, and also they play a vital role in search of origins of geothermal fluids (e.g., Bergfeld et al., 2001).

Some review articles and books have been written extensively regarding the theory and applications of isotope techniques (Giggenbach, 1992; Hoefs, 2018; Rasskazov & Brandt, 2010). Presentation of isotopic values are usually in the delta notation and fractionations or separation factors in the alpha notation:

$$\delta_A = \left(\frac{R_A}{R_{std}} - 1 \right) \cdot 10^3 \text{ and } \alpha_{A-B} = \left(\frac{R_A}{R_{std}} \right) \quad [5.1]$$

where δ reported in *per mil* (‰), A and B represent the different phases and *std* represents standard. R is the ratio of different isotopes for a specific element and α is the fractionation between phase A and B .

Helium element has two isotopes derived from different origins. ^3He , which is referred as primordial isotope, was produced during the formation of the Earth during its initial accretion and in the subsequent long-term acquisition of late veneer material, and is therefore enriched in the mantle relative to the atmosphere in the order between 8 and 50 relative to ^4He (Graham, 2002; Stuart et al., 2003). However, ^3He is no longer generated in any significant quantities by radiogenic decay and is thus no longer added to the Earth system at any significant rate.

On the other hand, ^4He is produced from the alpha decay of U and Th, and accumulates over time which gets concentrated in the Earth's crust. The absolute abundance of helium in rocks is hard to understand as helium is so mobile and therefore, usually the $^3\text{He}/^4\text{He}$ ratio is used as a proxy for ^3He content. The ration (R) is expressed generally as a multiple of the present-day atmospheric $^3\text{He}/^4\text{He}$ ratio, R_a , which is 1.39×10^{-6} (Craig & Lupton, 1976; Mamyrin, 1970). Generally, high R_a of the volcanic system is usually assumed to be derived from the mantle and taken as evidence for the existence of a mantle reservoir that has mainly remained less degassed since the Earth's accretion. Ocean island basalts (OIB) from the deep, undegassed mantle reservoir are characterized by a high $^3\text{He}/^4\text{He}$ ratio that rises to 50 R_a , which is higher than the mid-ocean ridge basalts

(MORB) that are released in the upper asthenospheric mantle (Hilton et al., 2011; Kurz & Jenkins, 1981).

5.1.2. Reactive transport modeling

Numerical simulation is a powerful tool for understanding and predicting the complex interplay of physical and chemical processes in geologic media. Injection of meteoric water and magmatic components into a geothermal reservoir in most cases disturbs local equilibrium conditions when it flows through it. The extent of which chemical reactions occur depends on the magnitude of the flow and transport processes within this system. Therefore, there is the need to couple the static reservoir system to the flow and transport process. Coupled models for mass transport and chemical reactions in subsurface systems simulate how the geochemical reactions evolve in space and time.

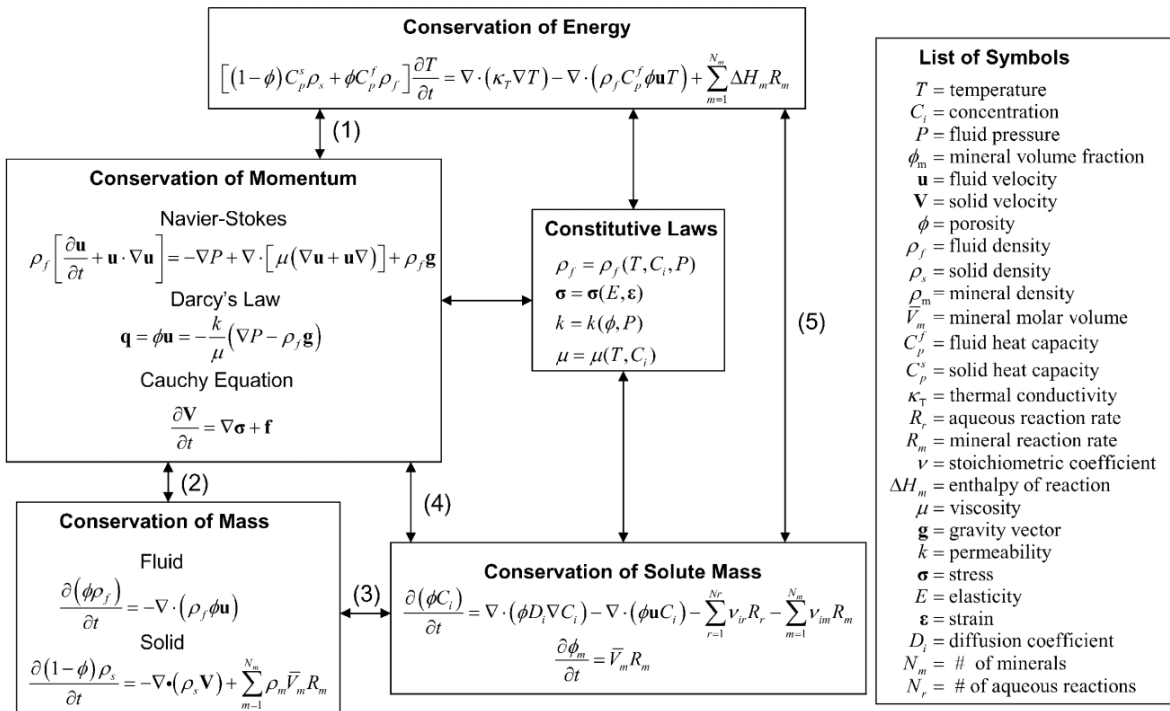


Figure 5.1: Governing equations for a single phase coupled reactive transport system (Steefel et al., 2005).

Theoretical formulations of the governing equations and the multi-process coupling in a reactive transport modeling have been extensively reviewed (e.g., [Lichtner, 1996](#); [Steeffel & Lasaga, 1994](#); [Yeh & Tripathi, 1989](#)). The main components of reactive transport computations are mass flow equations of fluids, transport equations of dissolved chemical species and equations describing homogenous and inhomogeneous chemical reactions in the coupling of various processes that involve two sets of equations which include the partial differential equations that define flow, transport and reactions kinetics and nonlinear algebraic equations that describe chemical reactions. Figure 5.1 gives the governing equations for a single phase system ([Steeffel et al., 2005](#)).

Reactive transport modeling is an essential tool for the analysis of complex physical and chemical coupled processes in the subsurface environment. Recent decades have seen the growing number of developments of reactive transport models and their applications to industrial and natural processes in both spatial and temporal scales processes involving water-rock interactions, such as the formation of ore deposits, transport and geological sequestration of carbon dioxide (CO₂) and geothermal resource development.

Earliest multi-component reactive transport models that could couple transport and biogeochemical processes are dated back to the mid-1980s with primary focus on developing theoretical and numerical techniques necessary for reactive transport simulation spanning a range of environments ([Steeffel et al., 2005](#)). Some of these studies included the basic theory of continuum model for a reactive transport ([Lichtner, 1985](#)), theoretical and numerical basis on how to treat reactive contaminant transport ([Yeh & Tripathi, 1989](#)), and reactive flow and transport model for non-isothermal, kinetically controlled water-rock interaction and fracture sealing in hydrothermal systems ([Steeffel & Lasaga, 1994](#)).

Continuum multicomponent reactive technique coupled for mass and fluid transport in porous media was developed in 1980's and have been widely applied in the past decades to predict and quantify chemical processes and reaction in groundwater flow, geological carbon sequestration, geothermal systems and other subsurface processes ([Bethke, 2008](#); [Li et al., 2017](#); [Lichtner, 1985](#); [Schwartz & Smith, 1988](#)). A straightforward approach for flow in a fractured media is the useful continuum model where individual fractures are considered to be related to that of individual pores

in porous media. In each of the computational grid blocks, the fractures and matrix are assumed to have the same pressure, temperature and chemical compositions. Since this approximation relies on the relative time scales for global flow and local equilibrium conditions, the method is more suitable for the case that requires a long-term, steady-state solution on considerable time and spatial scales of concern (Wu, 1999).

5.2. Geochemistry in geothermal exploration

Comprehensive reviews on the geochemical features of geothermal fluids exist in the literature (e.g., Arnórsson, 2000; Arnórsson et al., 2007; D'Amore, 1985; Ellis, 1967; Giggenbach, 1980, 1981; Henley et al., 1984; Henley & Ellis, 1983). Geochemical studies are particularly important in the assessment of geothermal resource in the initial stages of exploration. Chemical geothermometers may provide the only evidence on possible reservoir temperatures that precedes drilling. Similarly, geochemical studies may be used to delineate the spatial extent of the geothermal system before carrying out geophysical surveys. Even after the resistivity surveys, geochemical data can still offer crucial supporting evidences to claims of reservoir extent established from the results of the resistivity surveys.

5.3. Physical and chemical characteristics of Menengai geothermal field

About 50 deep geothermal wells with depths varying between 2,100 and 3,200 drilled within Menengai caldera proved the existence of the exploitable geothermal resource (GDC, 2018). The temperature and pressure recorded at the bottom-hole of these wells show temperatures more than 300 °C with some close to 400 °C indicating a typically high-enthalpy geothermal system. These wells tap mainly two aquifers where the shallow one occurs around 800 – 1200 m depth characterized by low-enthalpy fluids. The deeper aquifer occurs between the depths of 1600 m and below 2000 m dominated by high-enthalpy fluids. Consequently, the enthalpy of individual wells varies extensively from water enthalpy of less than 1000 kJ/kg to a dry steam enthalpy of about 2700 kJ/kg. Both single and two-phase fluids discharge from these wells with the steam-water ratio varying considerably.

Fluid samples collected during flow tests show that the chemistry of discharged fluid extensively exhibits erratic chemical behavior from well to well, which points to the inhomogeneous nature of the aquifers feeding the wells. The total dissolved solutes (TDS) ranges from 1,700 to 7,000 mg/kg whereas the pH values lie between 8.5 and 9.5. The fluid discharge is of sodium bicarbonate (NaHCO_3) facies (Figure 5.2), which forms most of the dissolved content, although significant chloride (from 370 to 1,050 mg/kg) and silica (250 mg/kg to 750 mg/kg) concentrations are also noted. The gas concentration in the steam phase range between 1 and 12 by weight percent and dominated by CO_2 (> 90 %). The latter being the major component of the non-condensable gas (NCG).

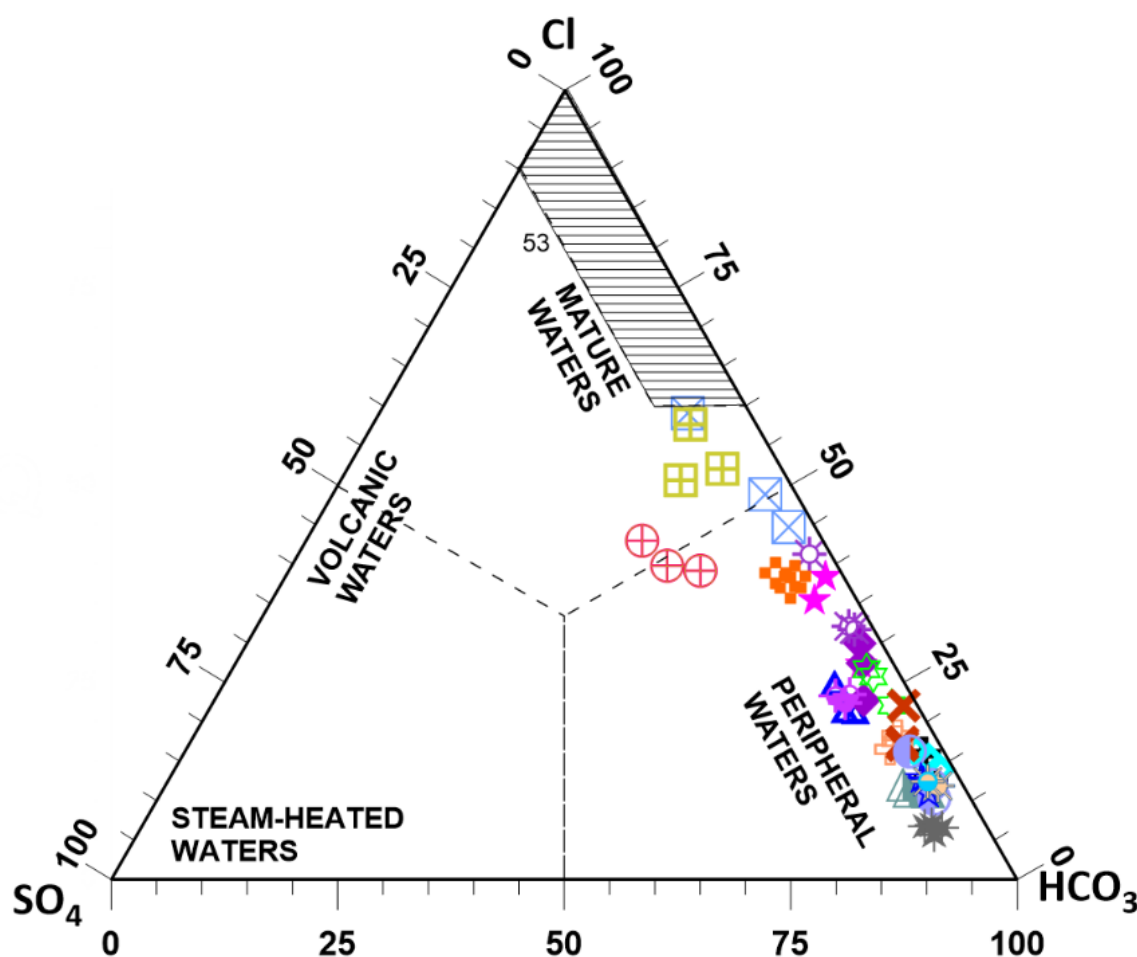
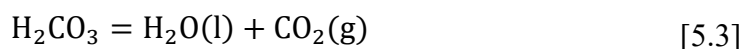
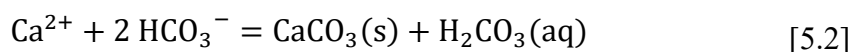


Figure 5.2: $\text{Cl-SO}_4\text{-HCO}_3$ ternary diagram (after Giggenbach, 1988) showing the distribution of major anions in the geothermal fluid of Menengai wells. The water type is predominantly bicarbonate (HCO_3^-). Various colored symbols are used to distinguish different wells.

Scaling has been reported in some of the discharging wells in both single and two-phase steam wells and was predicted by saturation indices and confirmed by a decrease in production because scaling has the potential obstruct upflow of fluid from the reservoir. Downhole samples retrieved from some of the wells further confirmed the existence of scale deposits in the system. Calcite scaling consequently removes calcium from the system, and this might have led to the low calcium concentration (between 0.07 and 1.2 mg/kg) reported in the discharged water.

The scaling characteristics of any given geothermal water are determined mainly by the composition of the geothermal fluids and physical conditions of the well. Since the composition of the fluids within Menengai varies significantly, the same thermal water can be exceedingly aggressive at one well and may show a different trend towards scaling at another, owing to changes in its physical and chemical parameters. Various authors have documented information on the solubility of calcite in water at elevated pressure and temperature conditions underpinning the processes of calcite deposition (e.g., Ellis, 1963; Ellis & Golding, 1963; Segnit et al., 1962). Calcite deposition has been known to increase with an increase in partial pressure of CO₂ and with a decrease in temperature. In a calcite saturated geothermal reservoir with the isothermal two-phase condition, a considerable drop in pressure would likely lead to calcite deposition. Simplified equilibrium reactions of scaling from a simplified thermal-water model system (CaCO₃ – CO₂ – H₂O) is described as (Stáhl et al., 2000)



Boiling in the well leads to the transfer of gases to the steam phase, and subsequently to a sharp reduction in CO₂ partial pressure. The decrease in pressure forces a shift of reaction in equation [5.2] to the right by the loss of CO₂ from the liquid into the gas phase upon flashing causing calcite formation. At reservoir conditions, a constant concentration of dissolved species of reactants, Ca²⁺ and HCO₃[−], and product, H₂O & CO₂ (i.e. H₂CO₃ or dissolved CO₂) reveals the state of old

equilibrium. The rapid removal of dissolved CO_2 , as H_2CO_3 from the fluids during boiling stresses the old equilibrium leading to the establishment of a new state of equilibrium. This degassing of CO_2 increases pH, hence, leading to an increase in carbonate ion concentrations. Increase in carbonate ions is thought to be responsible for making initially saturated geothermal water supersaturated with calcite (Arnórsson, 1989). Boiling has been suggested to be a tipping point of already supersaturated species, i.e., excess Ca^{2+} or HCO_3^- , at reservoir conditions (Sanchez et al., 2005).

CO_2 is practically always the most abundant fumarolic and well discharge gas (on a steam-free basis) by a considerable margin in the Kenyan rift system with concentrations often more than 90%. Similar phenomena have been observed in the producing wells of Menengai geothermal field from the existing data, and also reported in preceding studies (e.g., Auko et al., 2015; Kipngok et al., 2015; Mibei, 2012; Sekento, 2012). In most of the places where the geothermal activity is studied, juvenile CO_2 is thought to be the primary heating agent. The sheer amount of CO_2 outgassing has led authors to suggest that oxidized carbon must be derived predominantly from a mantle reservoir (e.g., Bailey, 1978; Darling et al., 1995; Lee et al., 2016). Releases of mantle-derived CO_2 are thought to be sourced mainly at volcanic centers. About 90% of total natural CO_2 emitted from the earth is released through active degassing of volcanic plumes or passive, diffuse emissions around volcanic edifices, mid-ocean ridges, and volcanic lakes (Burton et al., 2013; Lee et al., 2016; Robertson et al., 2016). The flexing and fissuring caused by horizontal movement of the anisotropic continental plate reactivates preexisting zones of weakness that channel the release of CO_2 -rich gas from the deeper mantle.

5.4. Results and discussion

A set of geochemical data was acquired from GDC and comprised of data that acquired since the first well was discharge-tested in 2011. A further collection of water and gas samples were also taken from the wells and analyzed for helium ^3He and ^4He as well as for δD and $\delta^{18}\text{O}$ isotopes.

5.4.1. Major chemical elements studies

Geothermal fluids contain a wide-ranging variety and concentration of dissolved components. The nature and content of dissolved chemical species in geothermal fluids are functions of local geology and temperature. Lower temperature systems in most cases have smaller quantities of solutes as compared to systems of higher temperature. This section discusses the relationship of some of the major ions in the Menengai geothermal field which include Na, dissolved CO₂, SiO₂, and Cl.

Figure 5.3 shows the scatter plot of SiO₂ versus CO₂. The concentration of SiO₂ appears to decrease with an increase in the concentration of CO₂, and this suggests that the two constituents do not have the same source. The wells with the highest amount of CO₂ correspond to a relatively low SiO₂, and vice versa. An imaginary line has been drawn to demonstrate the mixing boundary. In general, CO₂ is most prevalent in areas of volcanic activity where mixing of metamorphic or mantle-derived, magmatic CO₂ with groundwater occurs. CO₂-waters may also form by condensation of CO₂-containing vapor in perched aquifers in volcanic geothermal systems, or by mixing of percolating CO₂-rich condensate with the deep-seated parental geothermal fluid. CO₂-waters may as well form by mixing of high-enthalpy fluid that is yet to undergo fluid phase separation with cold groundwater.

SiO₂, on the other hand, is mainly affected by changes in temperature. High temperature leads to the dissolution of SiO₂ and hence its content in solution increases accordingly. The plot suggests the existence of an end-member fluid that has undergone extensive water-rock interactions leading to high SiO₂ and another end-member fluid that is high in CO₂ but is relatively younger in the system.

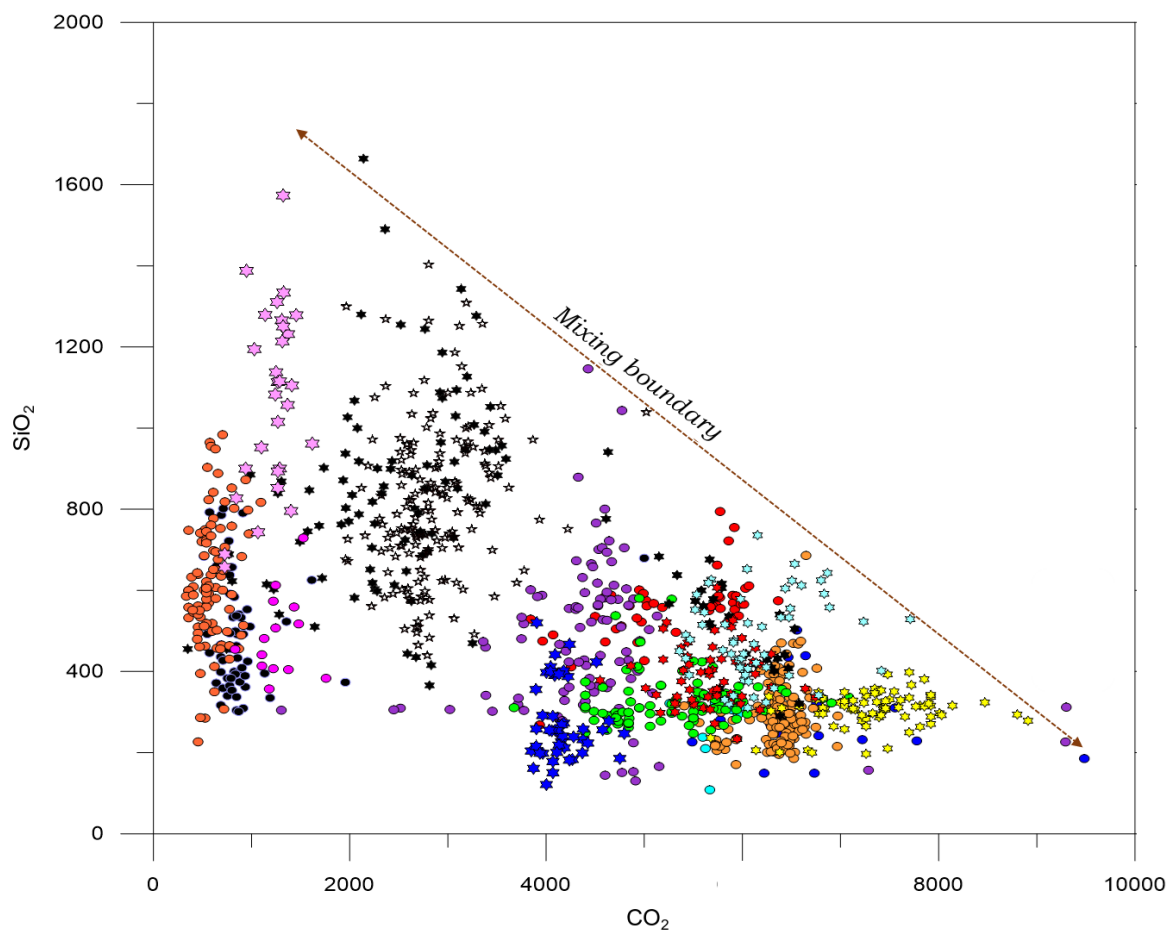


Figure 5.3: Scatter plots of SiO_2 versus CO_2 (mg/kg) of Menengai well discharge waters collected during the testing period. Similarly colored symbols represent different samples from the same well while the orange lines indicate hypothetical mixing boundaries.

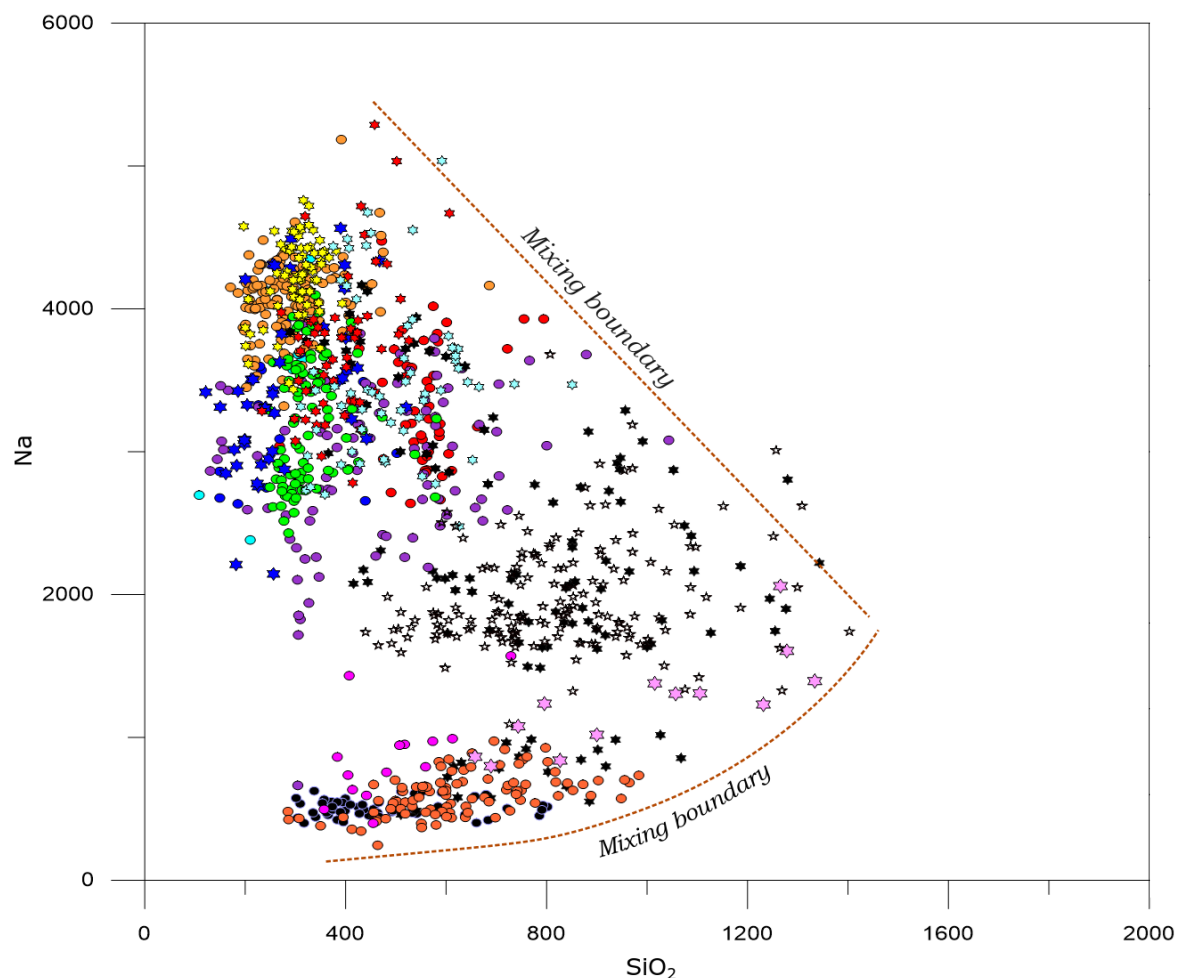


Figure 5.4: Scatter plots of Na versus SiO_2 (mg/kg) of Menengai well discharge waters collected during the testing period. Similarly colored symbols represent different samples from the same well while the orange lines indicate hypothetical mixing boundaries.

Figure 5.4 gives a comparable trend where Na relates similarly to CO_2 with regard to SiO_2 although the lower boundary of Na rises gently with a corresponding increase in SiO_2 concentrations. The relationship implies that both Na and CO_2 chemical constituents originate either from the same source or from a similar geological setting, and the linear correlation of the two elements (CO_2 versus Na) displayed in Figure 5.5 further confirmed this relationship. The linear mixing trend is a characteristic observed in all the well samples. The relationship of Cl vs. CO_2 (Figure 5.6) and also Cl versus Na (Figure 5.7) demonstrates an intersecting mixing trend that indicates the Cl concentrations are the highest where CO_2 and Na concentrations are in their transitional concentration.

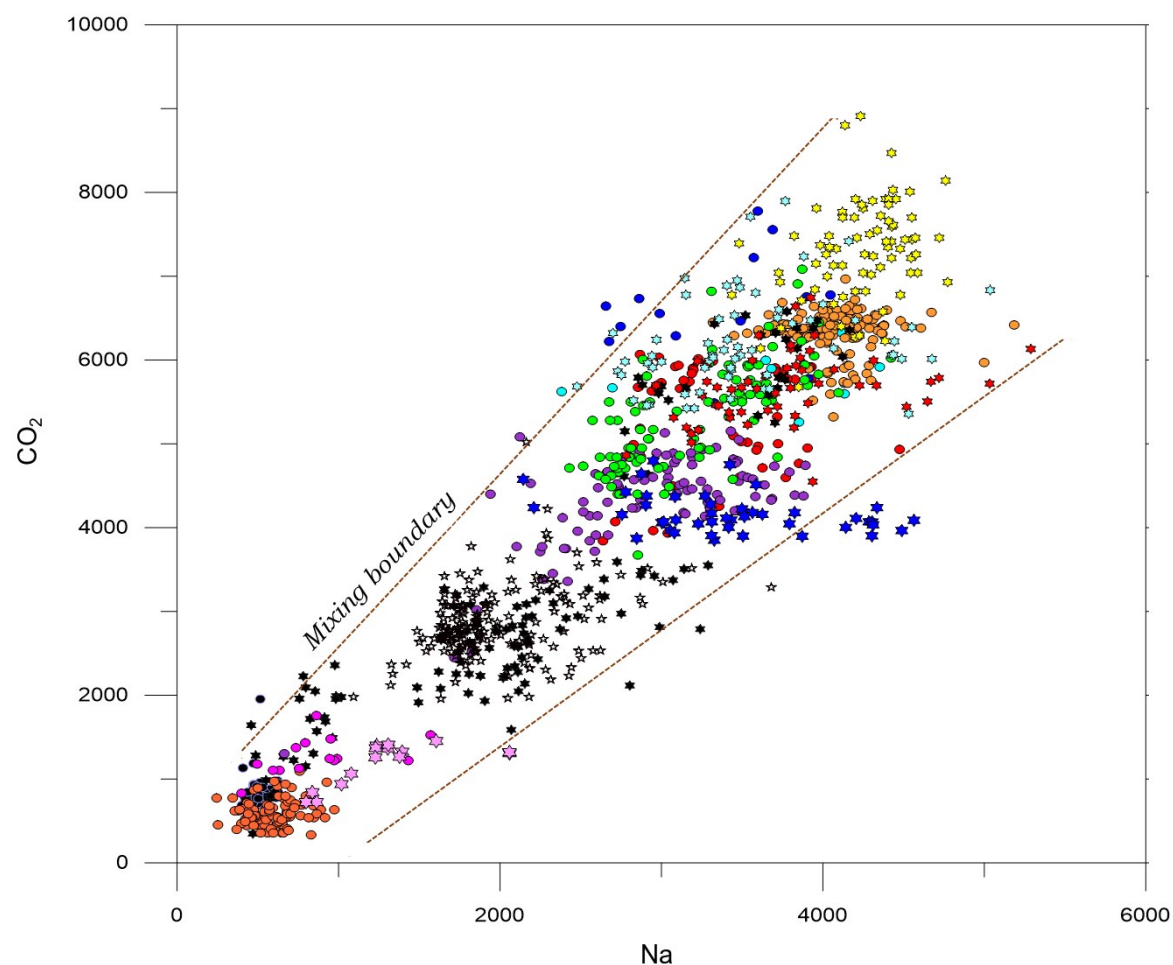


Figure 5.5: Scatter plots of CO₂ versus Na (mg/kg) of Menengai well discharge waters collected during the testing period. Similarly colored symbols represent different samples from the same well while the orange lines indicate hypothetical mixing boundaries.

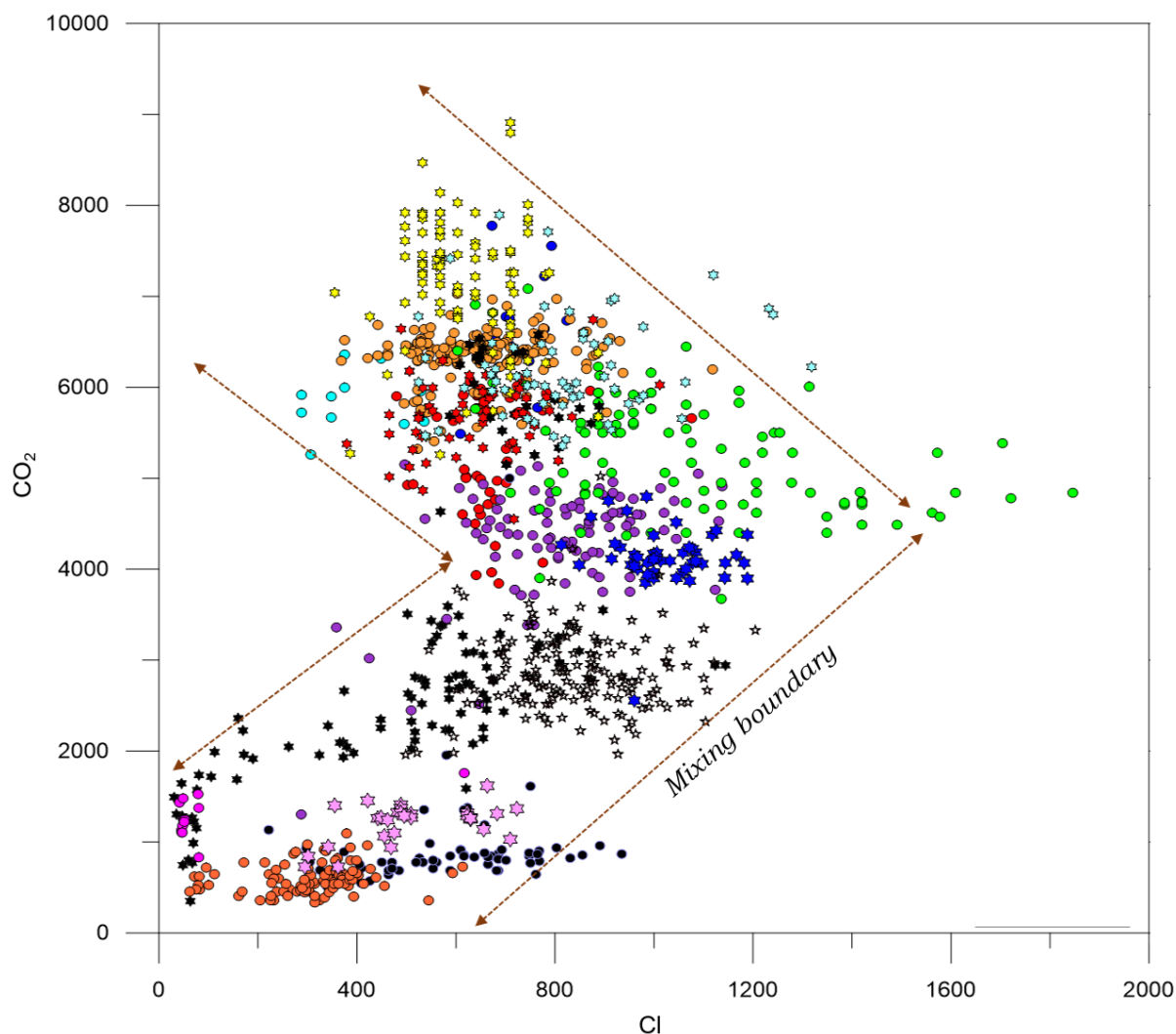


Figure 5.6: Scatter plots of CO_2 versus Cl (mg/kg) of Menengai well discharge waters collected during the testing period. Similarly colored symbols represent different samples from the same well while the orange lines indicate hypothetical mixing boundaries.

Cl concentrations is highest (> 1600 mg/kg) at moderate CO_2 (about 5000 mg/kg) and Na (3000 mg/kg) concentrations in Figure 5.6 and Figure 5.7, respectively, but decreases away from this peak values. From this, it appears that the Cl -rich fluid which in most geothermal represents mature waters is diluted progressively with two sources of fluid; Na - CO_2 -poor and Na - CO_2 -rich waters.

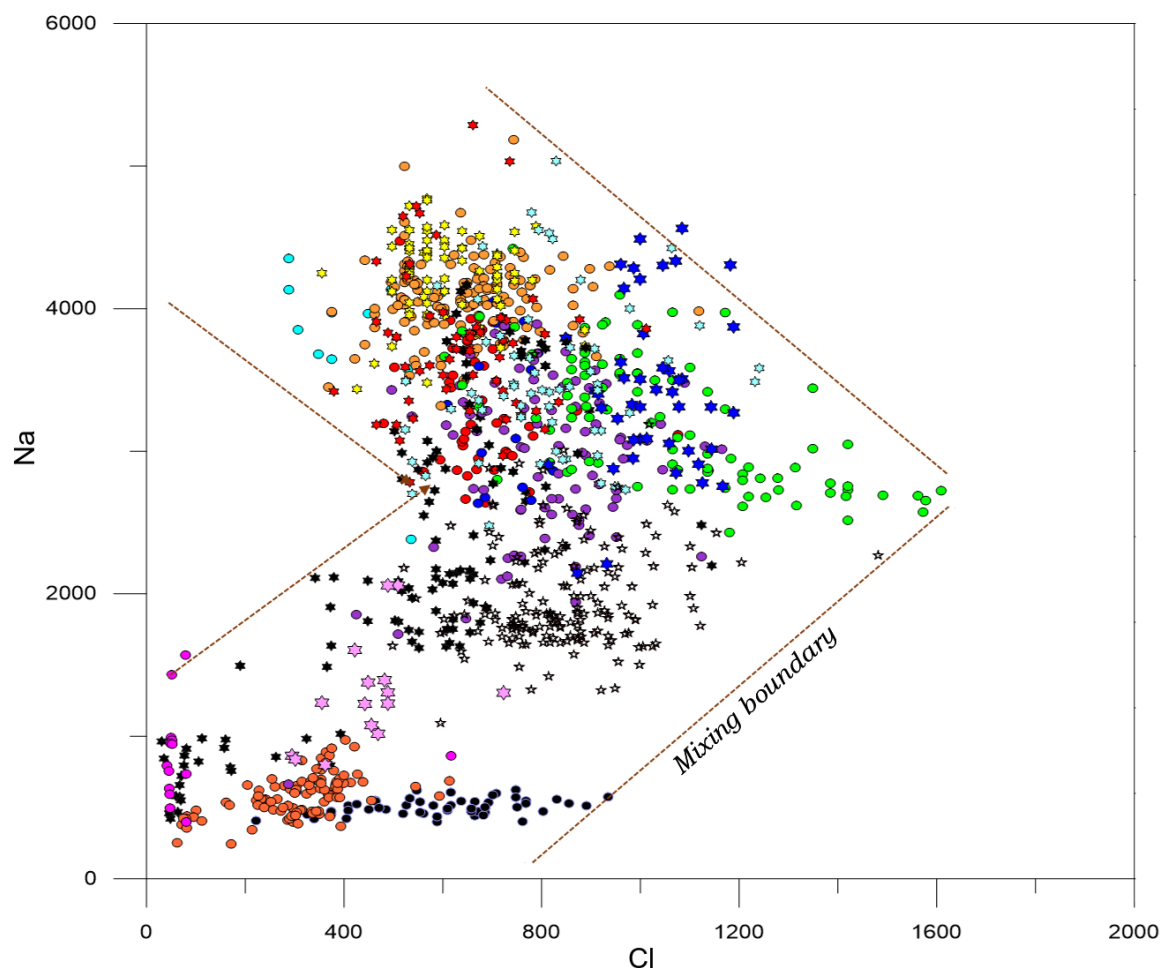


Figure 5.7: Scatter plots of Na versus Cl (mg/kg) of Menengai well discharge waters collected during the testing period. The symbols represent different wells while the orange lines indicate hypothetical mixing boundaries.

5.4.2. Isotope Studies

Generally, water isotopes are used in geothermal studies to identify the origin of geothermal fluids and to define the physicochemical characteristics of the geothermal reservoir for optimal exploitation. The variation in $^{18}\text{O}/^{16}\text{O}$ and D/H ratios about the Standard Mean Ocean Water (SMOW) are usually used to report the isotope composition of hydrogen and oxygen. The ocean is the primary water reservoir and quite homogeneous, and therefore it is preferably used as a reference standard for the δ scale of both hydrogen and oxygen isotopes in water samples (Craig, 1961). The isotopic composition can be expressed as per mil difference relative to SMOW. The δ - (per mil) value is given by:

$$\delta = \left[\frac{R_{\text{sample}}}{R_{\text{SMOW}}} - 1 \right] \times 1000 \quad [5.4]$$

where R is the atomic ratio D/H and $^{18}\text{O}/^{16}\text{O}$, e.g., for hydrogen:

$$\delta D = \left[\frac{D/H_{\text{sample}} - D/H_{\text{SMOW}}}{D/H_{\text{SMOW}}} \right] \times 1000 \quad [5.5]$$

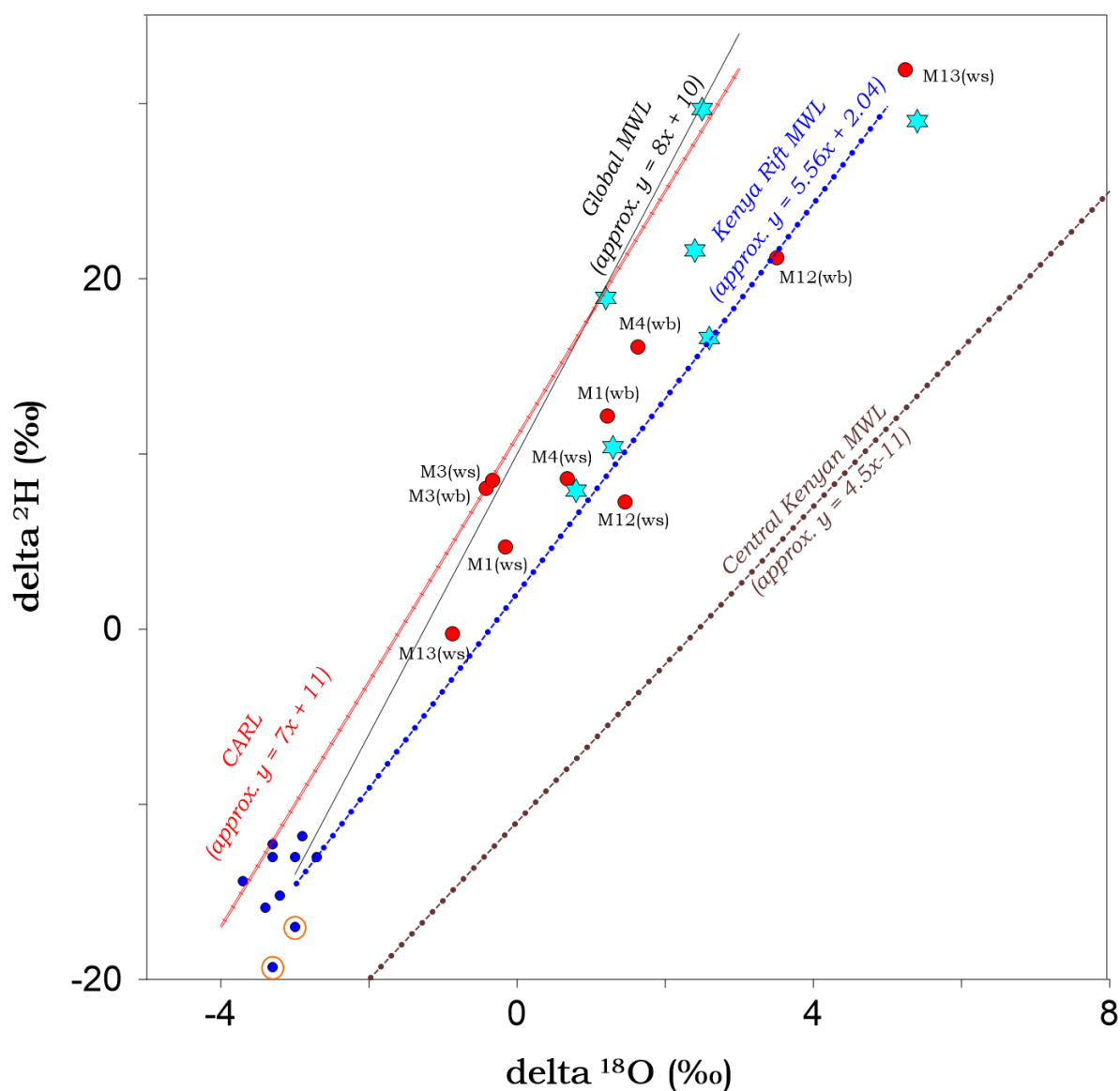


Figure 5.8: δD (^2H) and $\delta^{18}\text{O}$ of some of the Menengai well fluid (red dots); cold borehole (blue dots) and warm borehole water (blue dot enclosed in an orange circle), and; rift valley lakes (stars).

Also presented are the Global Meteoric Water Line - GMWL (Craig, 1961), Kenya Rift Meteoric Water Line - KRMWL (Allen & Darling, 1987), Central Kenya Meteoric Water Line – CKMWL and Continental African Rain Line –CARL, defined by Ármannsson (1994). The boreholes are named after their well number with prefix ‘M’ for Menengai, and suffixes ‘ws’ and ‘wb’ are abbreviations for the names of the sampling points, i.e., for Webre steam-water separator and weir-box, respectively.

Figure 5.8 displays the δD and $\delta^{18}O$ analyzed from various water sources within and adjacent to the study area. Most cold groundwater resources are of meteoric origin (meteoric groundwater). There is a strong relationship between the δD , and $\delta^{18}O$ values of groundwater reflective of the CARL, GMWL and KRMWL plots. The δD and $\delta^{18}O$ values of all borehole samples appear to be depleted relative to those of the geothermal well and the Lake Nakuru samples typical of a water body which has not undergone evaporation or intense water-rock interaction. The samples from the boreholes discharging warm water, however, appear to be more depleted of δD relative to $\delta^{18}O$ compared to other borehole samples. These thermal water boreholes seem to tap from much older water bodies (referred to as palaeowater in Figure 5.9) and probably controlled by its MWL which shifted away from the main MWL.

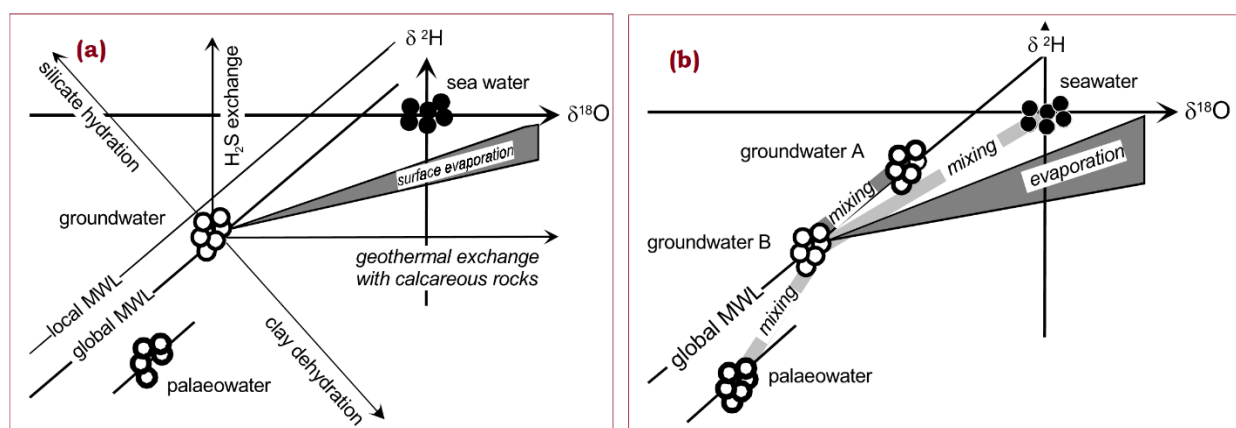


Figure 5.9: (a) Various processes which shift the δD and $\delta^{18}O$ values from the MWL. Evaporation shifts both δD and $\delta^{18}O$ values. The δD is displaced as a result of an isotopic exchange with volcanic CO₂ and limestone while $\delta^{18}O$ is shifted due to exchange with H₂S and silicate hydration. (b) δD and $\delta^{18}O$ plot for continental precipitation (MWL = Meteoric Water Line; local corresponds

to the Mediterranean precipitation) with examples of various mixing lines. The MWL of Pleistocene palaeowater may be apart from the MWL (adapted from Geyh et al., 2008).

Most of the lake water is aligned close to the KRMWL. Well M3 lies adjacent to M4 and derives its discharge fluid from a liquid-dominated system with a high liquid-steam ratio, presently utilized as a source of hot water for various direct use pilot projects (GDC, 2018). Although, the well plots along CARL, it may be interpreted locally as being influenced by hydration of silicate minerals and thus δD is shifted relative to $\delta^{18}O$ according to Figure 5.9a.

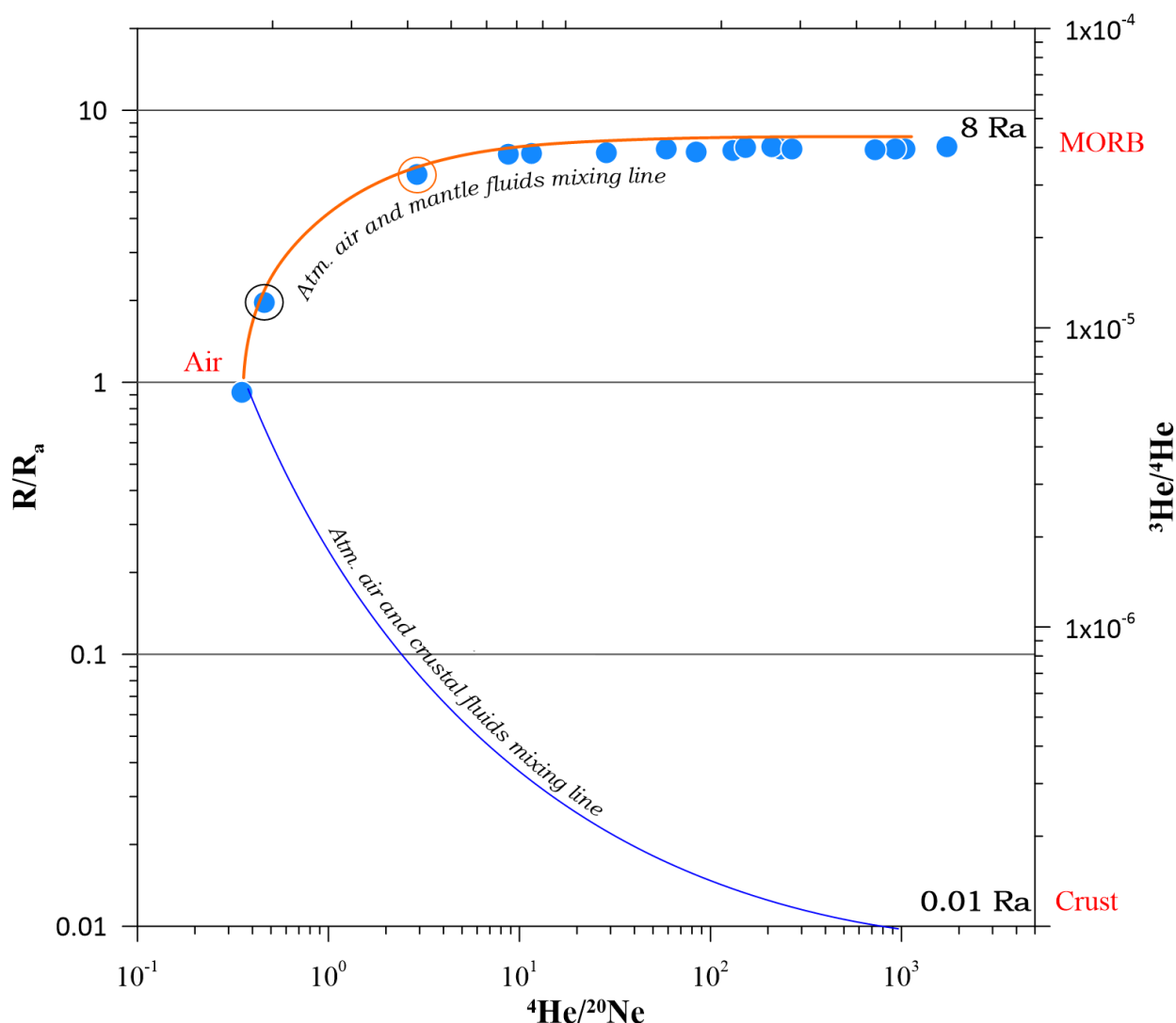


Figure 5.10: R/R_a and $^3\text{He}/^4\text{He}$ values versus $^4\text{He}/^{20}\text{Ne}$ ratio diagram of sampled gas data from Menengai wells (blue dots), a borehole (blue dot encircled with a black circle) and a fumarole (blue dot encircled with an orange circle) samples. The theoretical lines represent binary mixing

trends for mantle-derived, crustal and atmospheric helium. The end member values for $^4\text{He}/^{20}\text{Ne}$ ratios are 1000 for crust and mantle and 0.318 for air. The end-members for the $^3\text{He}/^4\text{He}$ is 0.01 Ra for the crust, 1 Ra for air and 8 Ra for MORB. The ratio (R) of ^3He to ^4He is often used to represent ^3He content and R usually is given as a multiple of the present atmospheric ratio (Ra).

The primary source of ^3He is the mantle while Neon (Ne) is a species primarily of pure meteoric (atmospheric) origin (e.g., Craig & Lupton, 1976; Farley & Poreda, 1993; Giggenbach & Poreda, 1993). In this regard, Ne contents may be used to assess the amount of atmospheric or meteoric influences on hydrothermal systems relative to their helium isotope ratios. Figure 5.10 shows that the values show that the He contents of this area are almost purely of upper mantle origin and aligns along the theoretical atmospheric-mantle mixing line.

5.5. Reactive geochemical modeling

5.5.1. Physical and chemical characteristics of MW-01

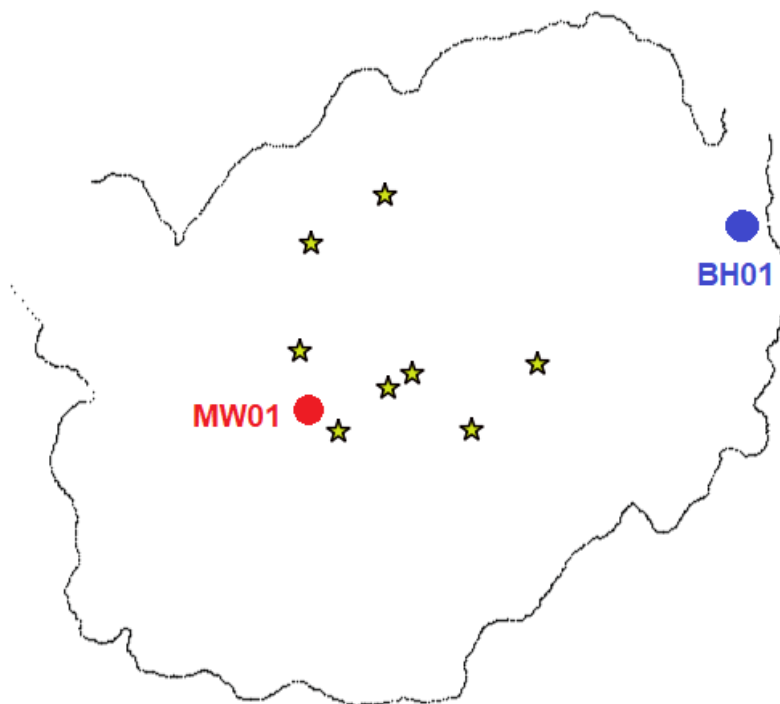


Figure 5.11: Map of Menengai caldera showing the location of geothermal well MW-01 and groundwater borehole BH01. Location of fumaroles represented by asterisks symbols.

Menengai Well 1 (MW-01) was selected for this study because of its much data acquired over the years, and more extensive studies carried out relative to the rest of the wells. MW-01 was the first exploration well to be drilled in Menengai geothermal field to determine the presence and nature of the geothermal reservoir (Figure 5.11). The well was drilled to a vertical depth of 2,200 m, with the production casing shoe set at 831 m depth. During drilling, sampling and analysis of rock cuttings at regular depths of 2 m was carried out.

After completion of the drilling process, the well took 25 days to recover fully and was then allowed to discharge. The discharge test was occasionally carried out from 25th May 2011 until May 2012. Work-over operations were carried out in July 2012 to remove a scale that formed a few months after the initial well discharge. The well was then discharged again in September 2012, with intermittent shut-ins. Collection of gas and water samples for chemical analysis was done during the discharge period. A second work-over operation was carried out from 25th July to 7th August 2013. The well was again discharged on 5th September 2013 after about a month of heat up to test calcite inhibitor. Flow and chemical analysis was carried out in all the discharge periods to determine the flow and chemical characteristics of the well as well as the efficacy of geothermal scale inhibitor.

Figure 5.12 presents measured temperature and pressure profiles of MW-01 taken during the warm-up and flowing conditions. The profile immediately after drilling indicates feed zones at around 1000 m, 1300 m, 1700 m and below 2000 m. The injection profile after well work-over shows feed zones around 1000 m and another at around 1850 m. During warm-up, feed zones below the casing depth, around 1700 m and at the bottom of the well became evident.

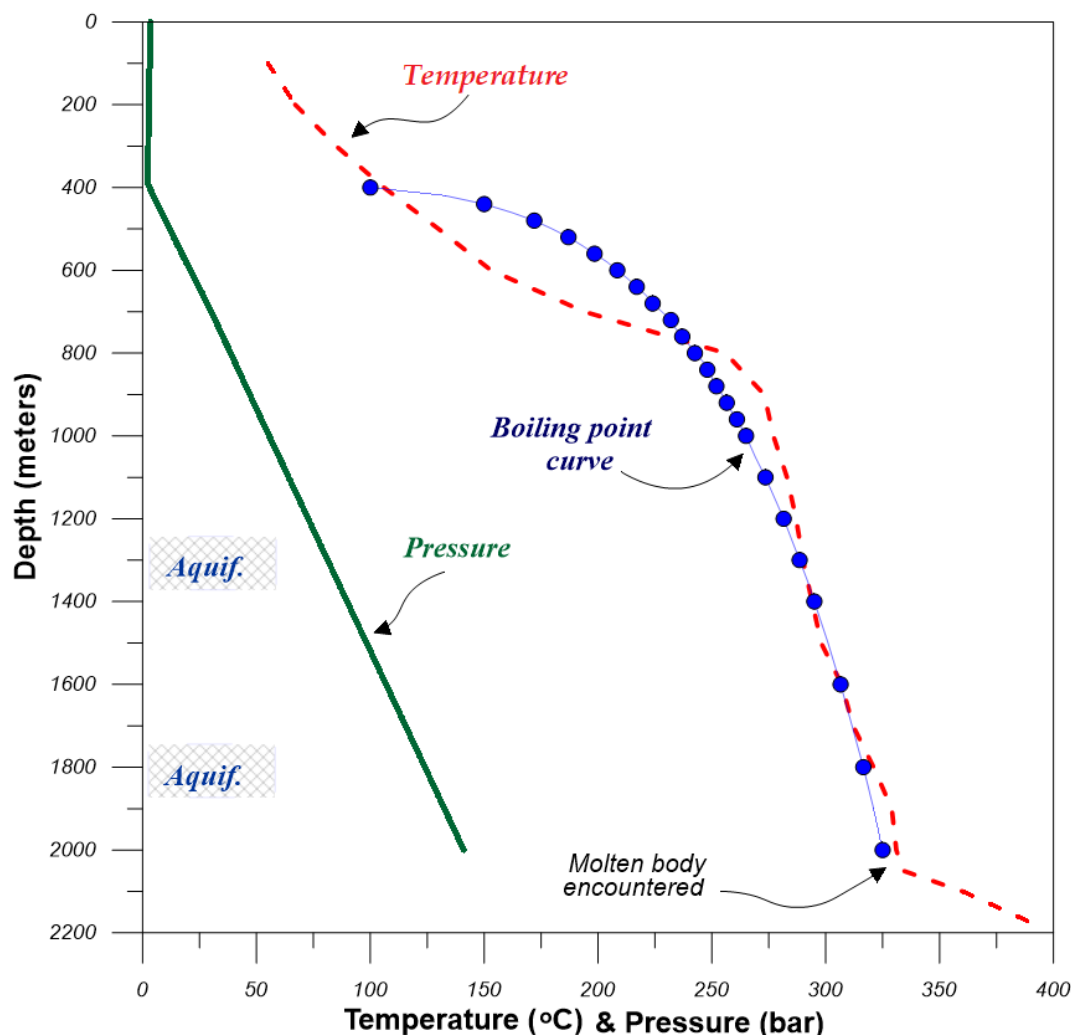


Figure 5.12: Downhole temperature and pressure profiles measurement of MW-01 measurement obtained after the 44 days heating-up after the first work-over operation. Also indicated are the inferred aquifer depths and boiling point curve with depth.

5.5.2. Mineral and geochemical data

The calculation of the aquifer fluid composition is based on a model discussed by (Arnórsson et al., 2007, 2010) and was conducted with the help of the WATCH chemical speciation program (Arnórsson et al., 1982; Bjarnason, 2010). The calculations were based on the fact that the excess enthalpy from the field resulted from phase segregation, which occurred at 180°C (10 bars) and the fluid represented by liquid water only (Karingithi et al., 2010). The model of phase segregation is displayed in Figure 5.13 and is discussed in Arnórsson et al. (2010) and brings into attention two versions of the model. The first version was adopted in this study and is based on selecting a

phase segregation temperature, assuming that the vapor fraction in the initial aquifer fluid is zero. The aquifer temperature is selected based on quartz, and Na/K geothermometers results and the model of choice utilizes the vapor fraction in the initial aquifer fluid, i.e., the fluid beyond the depressurization zone around wells.

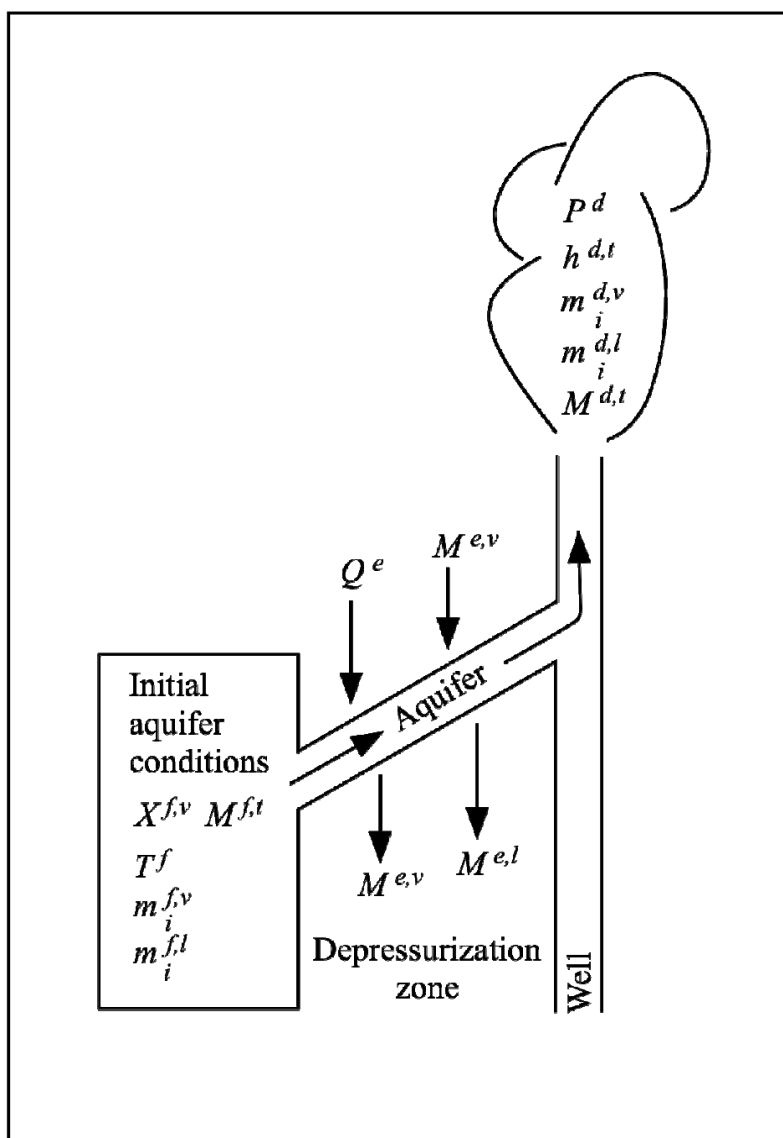


Figure 5.13: Schematic diagram describing the processes occurring in the depressurization zone around wells (Arnórsson et al., 2010). Where $h^{d,t}$ is discharge enthalpy (kJ/kg); P^d is Sampling pressure (bar abs.); T^f is aquifer temperature ($^{\circ}\text{C}$), and; $X^{f,v}$ is Initial aquifer fluid vapor fraction. $m_i^{f,v}$, $m_i^{f,l}$, $m_i^{d,v}$ and $m_i^{d,l}$ are concentration of component i in the initial aquifer steam, the initial aquifer water, the discharged steam, and the discharged water, respectively. $M^{e,v}$, $M^{e,l}$ and Q^e are

mass flow of steam, mass flow of water and mass flow enthalpy in the intermediate zone, respectively. $M^{f,t}$ and $M^{d,t}$ are the total mass flow at initial aquifer and discharge conditions, respectively.

The lithology of wells drilled in Menengai is discussed in previous studies (e.g., Lagat et al., 2010; Mbia, 2014; Mibei, 2012; Mungania & Lagat, 2004; Omondi, 2011). Description of the lithological formations observed in MW-01 aided by petrographic thin sections analysis reveals that well MW-01 penetrates through a thick fine to medium-grained post-caldera lavas of pantelleritic trachytes with minor intercalations of tuffs at some depths. The logs illustrated strong silica saturated hypocrySTALLINE and fine to medium grained plagioclase phyric trachytes. It also has clinopyroxene, aegirine-augite-hedenbergite, and amphibolic arfvedsonite-riebeckite microphenocrysts within microlitic alkali feldspar-rich glassy matrix, alongside disseminated flecks of iron oxides.

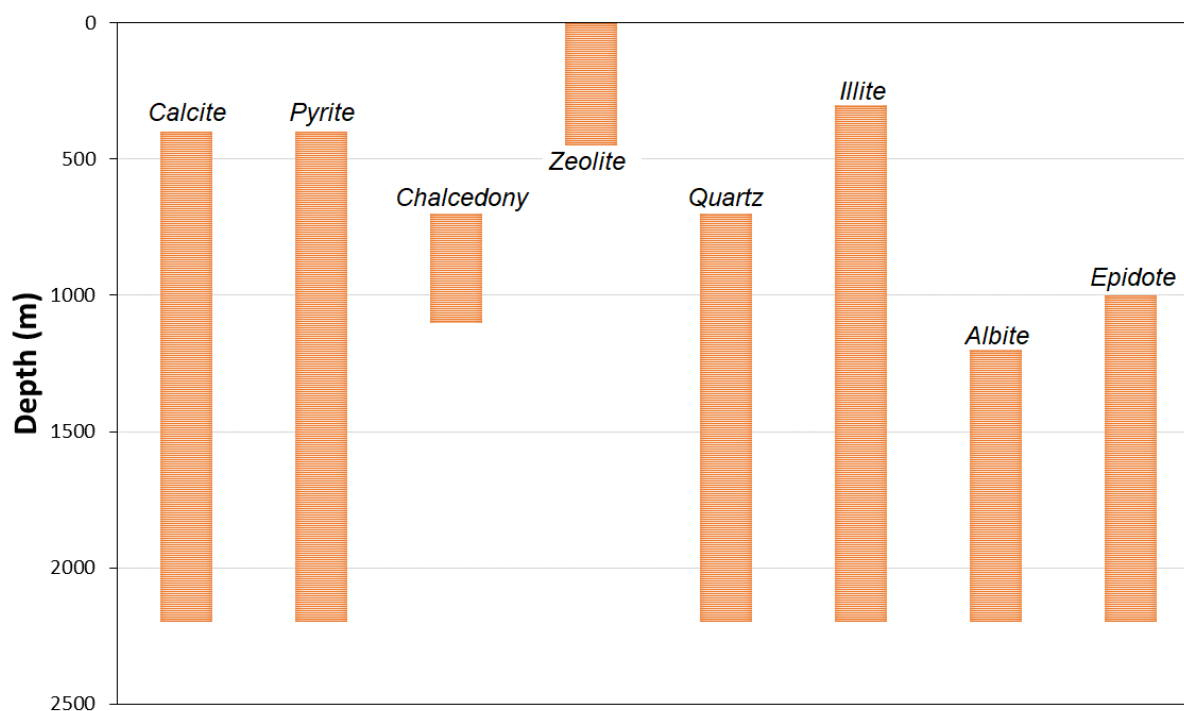


Figure 5.14: Downhole secondary minerals distribution of MW-01 recorded during the drilling operation.

Hydrothermal alteration in well MW-01 is mostly dependent on the chemical compositions of the rocks, temperature, permeability and hydrothermal fluid composition. Figure 5.14 shows distributions of secondary minerals encountered during analysis of the cuttings. These minerals in MW-01 included chalcedony, zeolites, quartz, calcite, illite, albite, epidote and pyrite. Precipitation of these minerals is not continuous as shown but are irregularly distributed owing mainly to lack of samples retrieved during drilling resulting from loss of fluid circulation.

The bulk of Menengai volcano eruptives is phenocryst-poor, with modal phenocrysts usually ranging only between 0 and 6%, with exceptions to the last erupted portions of the ash-flow tuffs that have modal abundances of up to 35% ([Macdonald et al., 2011](#)). Therefore, observations by petrographic microscope might not adequately give representative assemblage of the contained mineralogy. Table 5.1 presents the mineral composition of the phenocryst. For a better bulk representative of rock chemical composition, an idealized estimate of mineralogy was achieved through the use of CIPW norm of the crystallized melt. In this regard, some rock samples were taken from the outcrops within and outside the caldera and analyzed for their chemical composition. Figure 5.15 shows the total alkali versus silica (TAS) diagram of the Menengai volcanics.

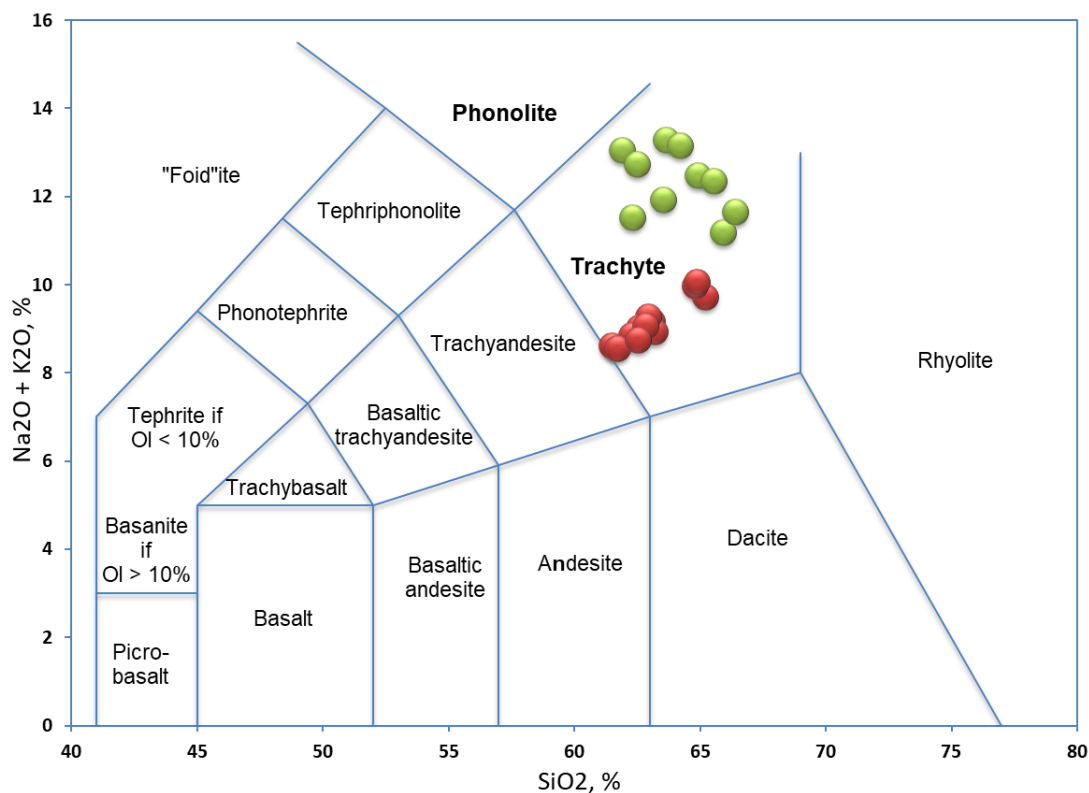


Figure 5.15: Total alkali versus silica (TAS) diagram of Menengai volcanics. The red dots represent samples that were collected on 2017 survey, while the green dots are reported in [Macdonald et al. \(2011\)](#).

Table 5.1: Mineral composition and their chemical formula of phenocrysts

Phenocryst	Chemical formula	Volume (%)
Sanidine	$K(AlSi_3O_8)$	40
Augite	$(Ca, Na)(Mg, Fe, Al, Ti)(Si, Al)_2O_6$	15
Hypersthene	$(Mg, Fe^{+2})_2Si_2O_6$	10
Magnetite	Fe_3O_4	7
Riebeckite	$Na_2Mg_3Fe_3 + 2Si_8O_{22}(OH)_2$	8
Apatite	$Ca_5(PO_4)_3(OH, F, Cl)$	5
Others		15

The CIPW normative ([Cross et al., 1902](#)) is a useful algorithm proposed at the beginning of the twentieth century designed for estimation of standard mineral assemblages for igneous rocks. The

algorithm takes the chemical composition of the minerals and computes the corresponding anhydrous assemblage that could crystallize from such a composition at low pressures. In general, CIPW norms provide the foundation for the comparison of aphanitic igneous rocks with their phaneritic equivalents. However, the Fe-oxidation ratios ($\text{FeO}/\text{Fe}_2\text{O}_3$) of igneous requires a standardized way fixing it because it may erroneously modify the amount and nature of the normative minerals of the rocks (Middlemost, 1989).

To reduce this problem, Verma et al. (2002) documented three different methods for adjusting the Fe-oxidation ratio, which include a proposed actual analysis of iron composition (Le Bas et al., 1986), solving the equation of Le Maitre (1976), and use of the ratio involving the bulk chemical composition of the rocks proposed by Middlemost (1989). For this study, the CIPW norm was calculated assuming a $\text{Fe}_2\text{O}_3/(\text{Fe}_2\text{O}_3+\text{FeO})$ ratio of 0.5 for trachytic rocks proposed by Middlemost (1989), a method also preferred by previous authors (e.g., Pruseth, 2009; Verma et al., 2002). Table 5.2 presents the resulting mineral composition reported in volume percent. The dominant minerals include feldspars (Albite and orthoclase) which contributes 57 % and quartz which constitutes about 18 % of the total bulk composition. Diopside (7.8%) and hypersthene (12.9%) also form a considerable bulk of the assemblage. Apatite, halite, fluorite, ilmenite, magnetite and riebeckite also occur but as minor constituents.

Table 5.2: Minerals and their volume proportion obtained from the CIPW normative scheme. The surface area and reaction mechanisms used in the simulation are also shown.

Mineral	Chemical formula	Solid volume (%)	Surface area (cm^2/g)	Reactions mechanism
Quartz	SiO_2	18.41	10	kinetic
Diopside	$\text{CaMgSi}_2\text{O}_6$	7.83	10	kinetic
Hypersthene	$(\text{Mg,Fe})\text{SiO}_3$	12.91	10	kinetic
Albite	$\text{NaAlSi}_3\text{O}_8$	28.98	10	kinetic
Orthoclase	KAlSi_3O_8	28.45	10	Kinetic
Apatite	$\text{Ca}_5(\text{PO}_4)_3(\text{OH, F, Cl})$	0.25	Not considered	Missing data
Halite	NaCl	0.46	Not considered	Inadequate data

Fluorite	CaF_2	0.72	10	<i>kinetic</i>
Ilmenite	$FeTiO_2$	0.74	10	<i>kinetic</i>
Magnetite	$Fe^{2+} Fe^{3+} O_4$	0.21	10	<i>kinetic</i>
Riebeckite	$Na_2(Fe^{2+} Fe^{3+})$ $Si_3O_8(OH)_2$	0.24	Not considered	<i>Missing data</i>

5.5.3. Equilibrium and kinetic considerations

Kinetic and thermodynamic usually control chemical reactions of a natural system, where the former is confined typically to fast reactions. Concentrations of species involved in fast reactions such as aqueous complexation relate through the laws of mass action (equilibrium constants).

All minerals, except calcite, were set to react under kinetic constraints. Calcite was assumed to be at equilibrium because its reaction rate is rapid compared to silicate minerals. A general form of the rate law used is based on the Henry Eyring's 1935-formulation of transition state theory (Laidler & King, 1983). The rate law requires understanding, for each mineral phase of interest, the thermodynamic solubility constant, the reactive surface area, and the additional rate law parameters for individual reaction mechanisms. The rate law is a product of the rate constant and reactive surface area (Lasaga et al., 1994; Steefel & Lasaga, 1994):

$$r = kA \left[1 - \left(\frac{Q}{K} \right)^\theta \right]^\eta \quad [5.6]$$

where r is the kinetic rate of dissolution (when r results in positive values) or precipitation (when r results in negative values); k is the temperature dependent rate constant; A is the reactive surface area; K is the equilibrium constant mineral-water reaction; Q is the reaction quotient. The parameters θ and η are empirical power-law positive-number exponents, whose values are set to unity in most cases.

For many minerals, the full rate equation includes a term for those in pure H₂O (neutral pH mechanism), and those catalyzed by H⁺ (acid mechanism) and OH⁻ (base mechanism), and the

constant kinetic k , and is often written as the sum of these three mechanisms (Palandri & Kharaka, 2004):

$$\begin{aligned}
 \text{Rate} = & k_{25}^{nu} \exp \left[\frac{-E_A^{nu}}{R} \left(\frac{1}{T} - \frac{1}{298.15} \right) \right] \\
 & + k_{25}^H \exp \left[\frac{-E_A^H}{R} \left(\frac{1}{T} - \frac{1}{298.15} \right) \right] a_{H^+}^n \\
 & + k_{25}^{OH} \exp \left[\frac{-E_A^{OH}}{R} \left(\frac{1}{T} - \frac{1}{298.15} \right) \right] a_{OH^-}^n
 \end{aligned} \quad [5.7]$$

where the superscripts nu , H , OH denote neutral, acid and base mechanisms, respectively; k_{25} is the rate constant at 25°C; E_a is the activation energy; R is the gas constant; T is the absolute temperature; a is the activity of the aqueous species; and n is a power term (constant). The rate constant k_{25} can be dependent on other species, for instance, $O_2(aq)$ controls pyrite in regards the neutral mechanism, whereas the acid mechanism has two species involved: H^+ and Fe^{3+} (Palandri & Kharaka, 2004; Xu et al., 2012). Table 5.3 gives the parameters used for kinetic rate formulations for the different minerals.

The subject of surface area in kinetics is less studied, and as a result, its understanding is marred with uncertainties. Consequently, there is no entirely accepted method for approximating average mineral reactive surface areas. Reactive surface area, A_r , for reactive transport models is often estimated as the geometric surface area (GSA) assuming an ideal geometry of one or more particle sizes (cubic or spherical) and a smooth or rough surface modified with a surface roughness factor (Steeffel & Lasaga, 1994; Steefel & Lichtner, 1998; Xu and Pruess, 2004). GSA can then be normalized to the mass of mineral to obtain specific surface area (SSA).

Table 5.3: Initial rock mineral composition and potential alteration minerals along with parameters for calculating kinetic rates for mineral dissolution and precipitation. The kinetic parameters used are from Palandri & Kharaka (2004).

Mineral	Vol% of solid	$A \left(\frac{cm^2}{g} \right)$	Parameters for kinetic rate law					
			Neutral mechanism		Acid mechanism		Base mechanism	
			$k_{25} \left(\frac{mol}{m^2 \times s} \right)$	$E_a \left(\frac{kJ}{mol} \right)$	$k_{25} \left(\frac{mol}{m^2 \times s} \right)$	$E_a \left(\frac{kJ}{mol} \right)$	$k_{25} \left(\frac{mol}{m^2 \times s} \right)$	$E_a \left(\frac{kJ}{mol} \right)$
Primary								
Quartz (alpha)	18.41	10	1.023×10^{-14}	87.7				
Diopside	7.83	10	7.762×10^{-13}	80	9.549×10^{-10}	80 $n(H^+) = 0.6$		
Enstatite	12.91	10	1.905×10^{-12}	40.6	4.365×10^{-7}	96.1 $n(H^+) = 0.71$		
Albite	28.98	10	9.12×10^{-13}	69.8	1.3489×10^{-10}	65 $n(H^+) = 0.457$	1.047×10^{-17}	71.0 $n(H^+) = -0.572$
K-feldspar	28.45	10	3.89×10^{-13}	38	8.709×10^{-11}	51 $n(H^+) = 0.51$	6.309×10^{-22}	94.1 $n(H^+) = -0.823$
Fluorite	0.72	10	1.62×10^{-14}	73	7.2×10^{-8}	73 $n(H^+) = 1$		
Ilmenite	0.74	10	6.918×10^{-12}	37.9	4.47×10^{-9}	37.9 $n(H^+) = 1$		
Magnetite	0.21	10	1.659×10^{-11}	18.6	2.57×10^{-9}	18.6 $n(H^+) = 0.279$		
Riebeckite	0.24	10	6.918×10^{-13}	47.2	1.995×10^{-8}	56.6 $n(H^+) = 0.7$		
Secondary								
Calcite		n/a						
Pyrite		10	4.00×10^{-11}	56.9	3.02×10^{-8}	56.9 $n(H^+) = -0.5$ $n(Fe^{3+}) = 0.5$	2.818×10^{-5}	56.9 $n(O_2) = 0.5$
Chalcedony		10	1.25×10^{-14}	87.5				
Mordenite		10	2.50×10^{-13}	58				
Laumontite		10	5.01187×10^{-13}	58				
Riebeckite		10	6.309×10^{-13}	47.2	1.995×10^{-8}	56.6 $n(H^+) = 0.7$		
Quartz		10	1.023×10^{-14}	87.7				
SiO ₂ (amorphous)		10	5.888×10^{-13}	75.4				
Illite (clay)		151.6	1.659×10^{-13}	35	1.0471×10^{-11}	23.6 $n(H^+) = 0.34$	3.02×10^{-17}	58.9 $n(H^+) = -0.4$
Albite low		10	9.12×10^{-13}	69.8	1.3489×10^{-10}	65 $n(H^+) = 0.457$	1.047×10^{-17}	71.0 $n(H^+) = -0.572$
Epidote		10	1.023×10^{-12}	70.47	2.7118×10^{-11}	71.1 $n(H^+) = 0.338$	4.677×10^{-18}	79.1 $n(H^+) = -0.556$

Approximation of SSA can also be made through laboratory measurement using, for instance, the Brunauer-Emmett-Teller (BET) gas adsorption technique (Brunauer et al., 1938). The reactive surface area also depends on the fraction of the SSA that is not clogged by surface layers or other

mineral grains that causes pore surface area exclusion (Beckingham et al., 2016; Bourg et al., 2015). The accessibility of these sites determines whether or not these parts of the SSA can take part in the reaction. These surface areas are related to each other by a scaling factor applied to account for reduced surface reactivity, which is usually assigned an arbitrary value in practice or adjusted to fit measured reaction rates (e.g. Bourg et al., 2015; Liu et al., 2011; Xu et al., 2007; Beni et al., 2014). However, mineral reactive surface areas in the subsurface are mostly unknown.

In the present study, reactive surface areas are based on the work of Sonnenthal et al. (2005), who estimated using the geometric area of a cubic array of truncated spheres to make up the rock framework constituting the rock framework. A surface roughness of 10 was used based from some previous studies (e.g., White & Peterson, 1990; Zerai et al., 2006; Zhang et al., 2009; Xu et al., 2007) and taken as the ratio of the true (BET) surface area to the corresponding geometric surface area. However, illite was assigned a much larger surface area owing to its small particle size. The surface area for illite is calculated by supposing the grains have a tabular morphology.

5.5.4. Model setup

In this study, a one-dimensional reactive transport model was constructed assuming that the fluid flow is dominantly vertical. In this case, the fluid flows upwards from a depth of 2100 m below the surface through the porous section, intersecting relatively homogenous rock properties, infinite-acting peralkaline trachytic formation. The model was aligned to flow in a possible path for fumaroles discharge illustrated by the conceptual model shown in Figure 5.16. The model consists of 84 discrete grid blocks each having a thickness of 25 m and a linear configuration, same thickness, and identical geometry. Geothermal fluid (with dissolved CO₂) is injected in the bottom grid block, and the top grid block set at a constant boundary condition. The fluid flow together with geochemical transport simulation was set to run for 1000 years, presumed to be adequate timescale to model the development of static state condition in the hydrothermal system.

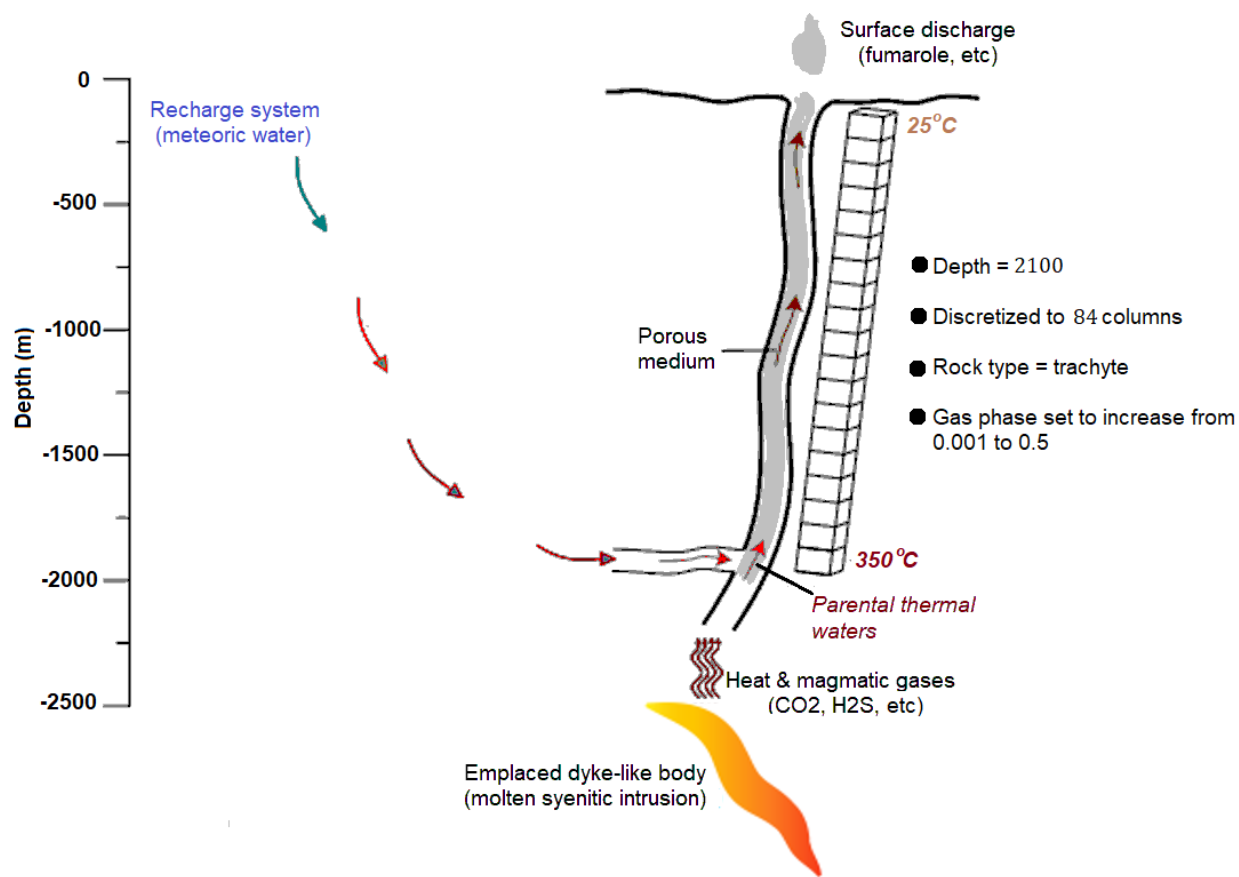


Figure 5.16: The model setup for the 1-D reactive transport simulation.

The temperature and pressure of the fluid range between 350°C – 25°C and 155 bars – 1 bar, at the base and top of the model, respectively, and are based field measurements (Figure 5.12). The hot fluid flow flashes on its way up, through the liquid-saturated formation, initially at a temperature significantly below saturation. The model assumes that the fumaroles produce fluid from the formation at or only just below saturation temperature, so the fluid flashes because of the pressure reduction. The pressure reduction works its way linearly upwards from the base, and as it does so the saturation temperature in the fluid falls. Well MW-01 discharge properties show that the mass fluid flows of steam and water phases are 89.8 t/hr and 100 t/hr, respectively. Subsequently, the fluid was set to flash progressively until a steam-water ratio of 0.5:1 was achieved assuming the prevalence of similar conditions in the geothermal system.

The simulation was performed using the parallelized computer code TOUGHREACT that employs sequential (or operator-splitting formulation) iterative approach (SIA) in solving solute transport

and chemical reaction processes (Xu et al., 2011). This simulation code for chemically reactive non-isothermal flows of multiphase fluids in both porous and fractured media introduces reactive chemistry into the existing framework of multiphase fluid and heat flow simulator TOUGH2 V2 (Pruess, 2004).

5.5.5. Results and discussion

Alteration minerals in geothermal systems provide useful information about the physical and chemical conditions of past and present hydrothermal activity. Clay minerals, in particular, have been studied as markers of circulation zones (e.g., Beaufort et al., 1992; Flexser, 1991; Marks et al., 2010; Mas et al., 2006; Reyes, 1990). The type of clay minerals that can form is dependent on the temperature of the system, time, fluid/rock ratio as well as their chemistries.

Geothermal reservoir waters are known to be saturated with calcite although some waters with a temperature higher than 300°C may be undersaturated (e.g., Arnórsson, 1989). During boiling the CO₂ partial pressure decreases drastically and that may then lead to a supersaturation concerning calcite and subsequent precipitation. The extent of precipitation is controlled mostly by the salinity of the geothermal water and the initial temperature in the system where precipitation increases with a decrease in temperature and increase in salinity.

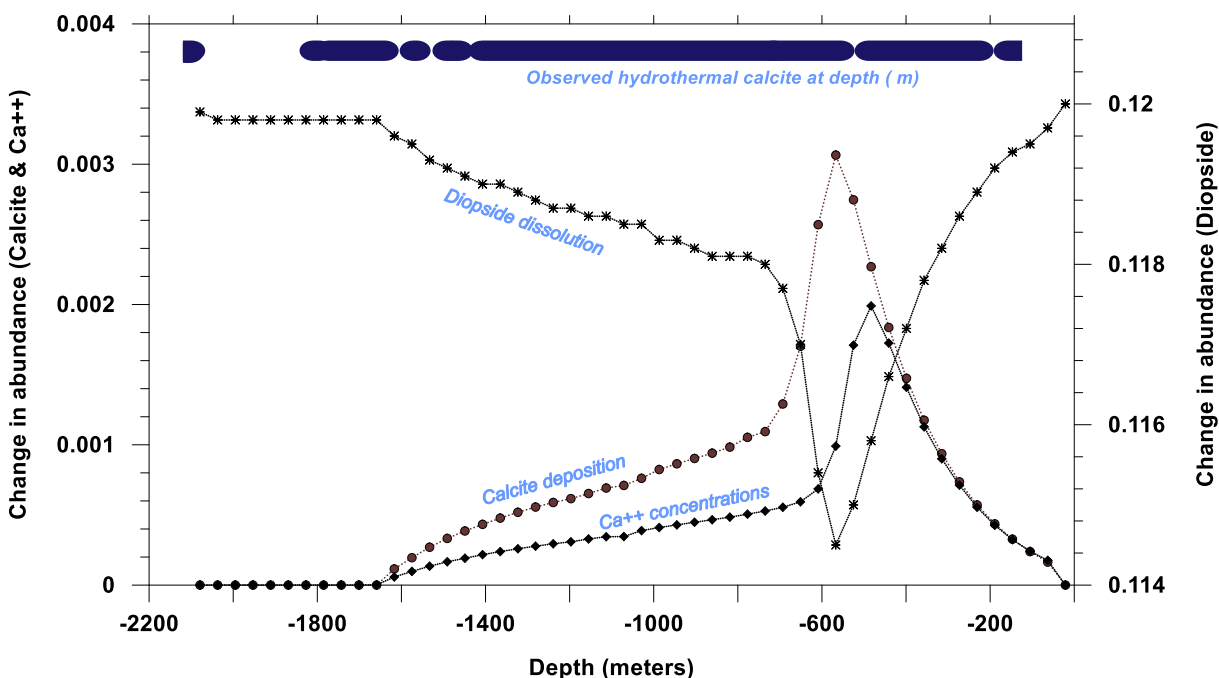
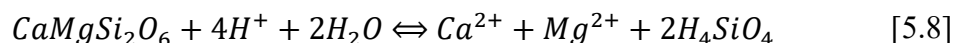


Figure 5.17: Comparison between the results of calcite deposition, Ca^{2+} concentrations, and diopside dissolution after 1000 years of simulation, and observed field data.

Figure 5.17 shows the results of calcite precipitation, Ca^{2+} concentrations, and diopside dissolution after 1000 years of simulation plotted alongside field data on the occurrence of calcite deposition at depth. Deposition occurs, although in small quantities, throughout the formation increasing steadily from the depth of 1700 m to 700 m and peaks sharply at about 600 m before it declines gradually. Deposition of calcite appears to coincide with the dissolution of diopside and the concentration of dissolved calcium. The same trend of calcite forming in the system from the simulation results is also seen in the observed field data at least from a qualitative sense. These results may suggest that the dissolution of diopside yields Ca^{2+} ions, which are requisite for the formation of calcite, although other external sources are possible such as those added by water recharging the system. The presence of Ca^{2+} ions may also be derived from the dissolution of apatite and fluorite minerals, and however, their concentrations in the system from Figure 5.17 are relatively small.

The formation of hydrothermal calcite relates to the movement of CO₂ in a geothermal system and is controlled by physicochemical processes which include dilution, boiling, and condensation. The dissolution of diopside can be described by



The deposition of calcite in hydrothermal systems is promoted by the dissolution of silicate rocks that release divalent metal cations such as Ca²⁺ to the fluid phase which can react with dissolved CO₂ and precipitate. The rate-limiting step mineralization of CO₂ is commonly thought to be the release of divalent cations into the solution through the dissolution of silicate rocks (Oelkers et al., 2008). The results of previous experimental investigations suggest that the direct effects of dissolved CO₂ (up to 1 bar P_{CO2}), HCO₃⁻ and CO₃²⁻ ions are mostly negligible, and pH is the primary controlling parameter of silicate dissolution (Marini, 2007), where dissolution rate decreases with increase in pH particularly with Ca-Mg-silicate minerals (Knauss et al., 1993).

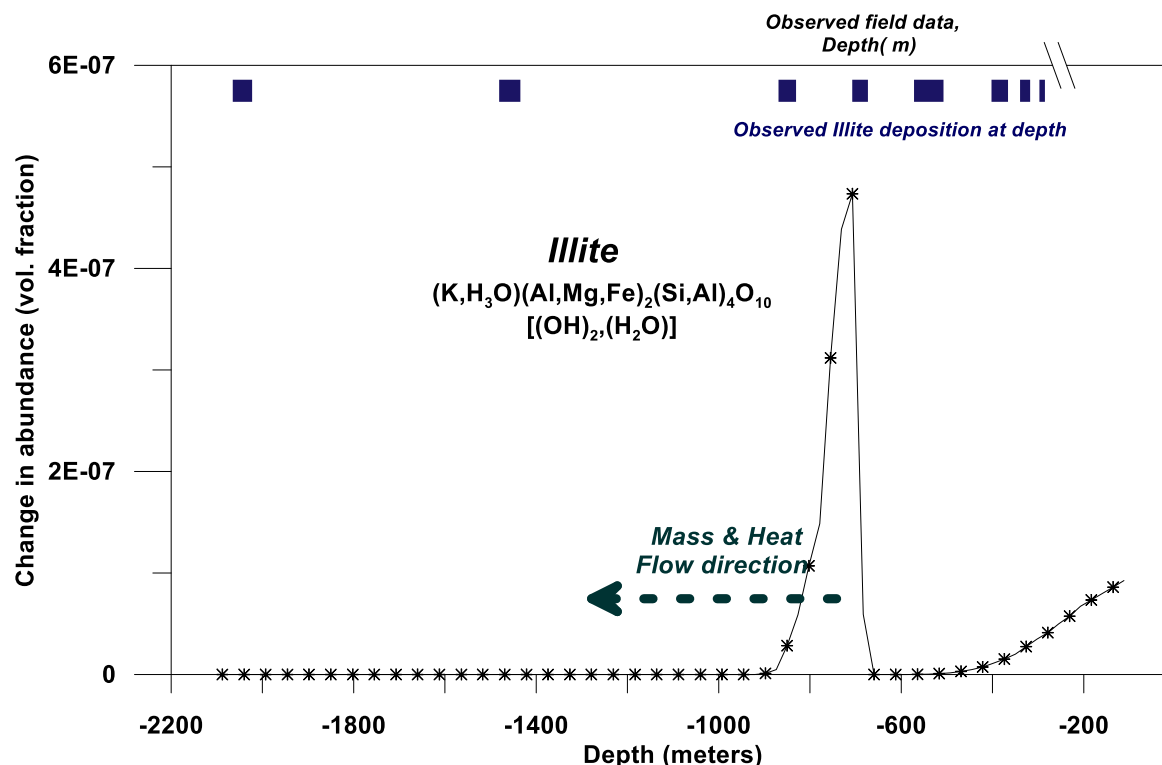


Figure 5.18: Comparison between the results of illite after 1000 years of simulation and observed field data.

Illite minerals occur commonly as a detrital form and also as product of alterations in hydrothermal systems. Because of its abundance in many geothermal systems, illites secondary minerals from hydrothermal alteration is used importantly as a marker of propylitic and argillic alteration zones (Srodon & Eberl, 1984). The occurrence of illite minerals represents the highest-temperature alteration conditions and also solutions with a higher potassium content. However, higher temperatures and potassium activity produce the potassic alteration phases characterized by potassium feldspar crystallization (Meunier & Velde, 2004). Figure 5.18 shows the distribution of illite formation in the simulation compared against data observed in the field. From the results, the formation of illite appears to form mainly in the upper section of the hydrothermal system (< 900 m), which corresponds to the depths where illite deposits occurred in the field. Calcite formation in the well, however, occur also at the depths of 1450 m and 2050 m, which is not accounted for by the simulation results.

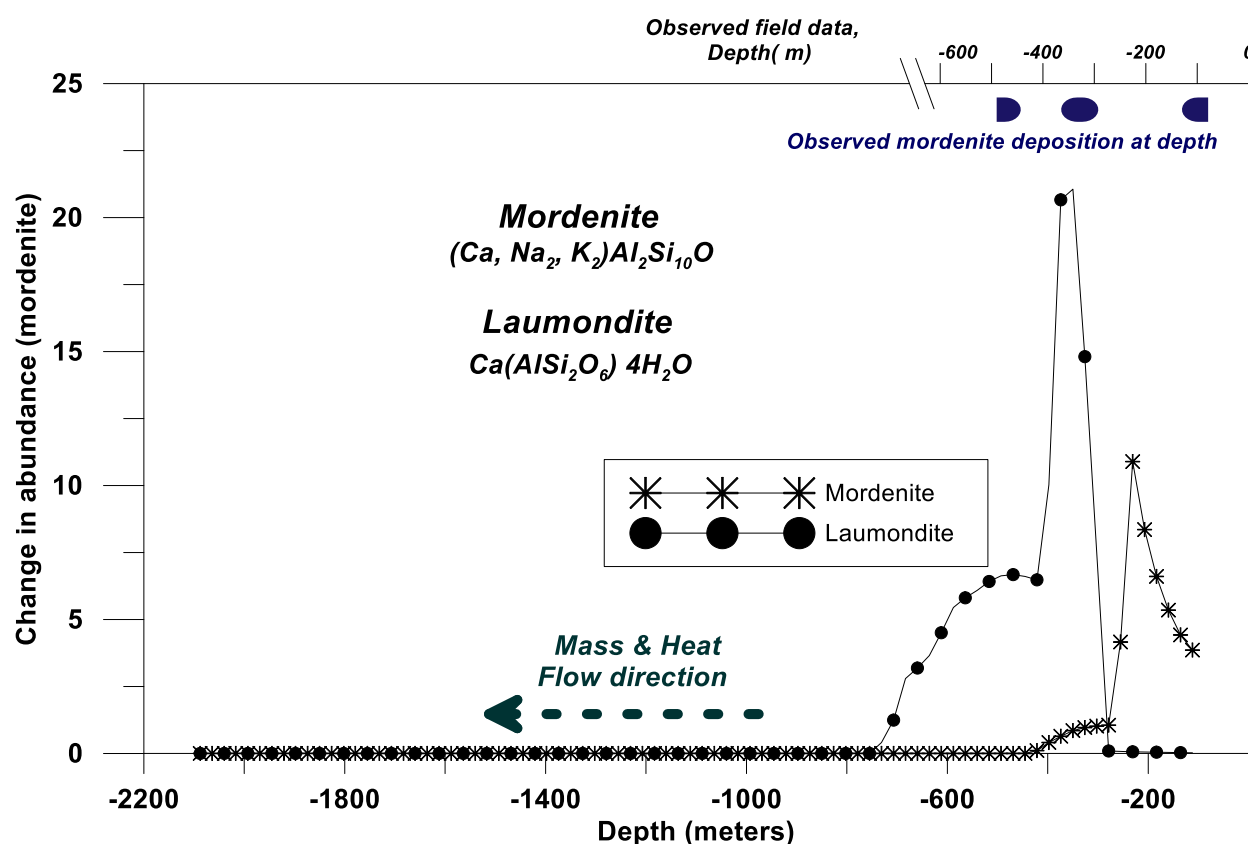


Figure 5.19: Comparison between the results of zeolites after 1000 years of simulation and observed field data.

Zeolites are a frequent constituent of volcanoclastic rocks which have gone through hydrothermal system forming mostly by the alteration of the original glass. Generally, the composition of the parent rock and also the conditions necessary for zeolitization such as pressure, temperature, and hydrology controls the occurrence of zeolite species in a given geological environment (e.g., Hall, 1998). In an open hydrological condition, mordenite is one of the commonly present minerals when the parental glass is acid-intermediate in composition (Hall & Moss, 1997). Mordenite (high-silica zeolite mineral) produced by the circulation of hydrothermal fluids within a volcanoclastic sequence occurs commonly within active geothermal systems as one of the earliest formed zeolites and (e.g., Kitsopoulos, 1997). Hydrothermal formation of laumontite is mostly known as the filling of fractures, vesicles and veins in igneous rocks setting (e.g., Hall & Moss, 1997).

Figure 5.19 shows a plot of both mordenite and laumontite from the simulation results. The occurrence of the mineral is reported only as zeolite from the field data, and for the sake of this study, the two minerals were chosen to represent zeolite formations. The formation of laumontite occurs at a higher temperature than mordenite, although both forms within the zeolite zone shown by the field data. The modeled results suggest the possible presence of more than one zeolite species forming in the system.

In geothermal systems, there are many polymorphs of silica minerals that can be observed such as quartz, chalcedony, among others. Figure 5.20 and Figure 5.21 show the relationship between the simulations and the field observations of quartz and chalcedony minerals, respectively. The quartz precipitation increases with depth and appears to match field data. On the other hand, the simulation fails to reproduce the occurrence chalcedony in the field.

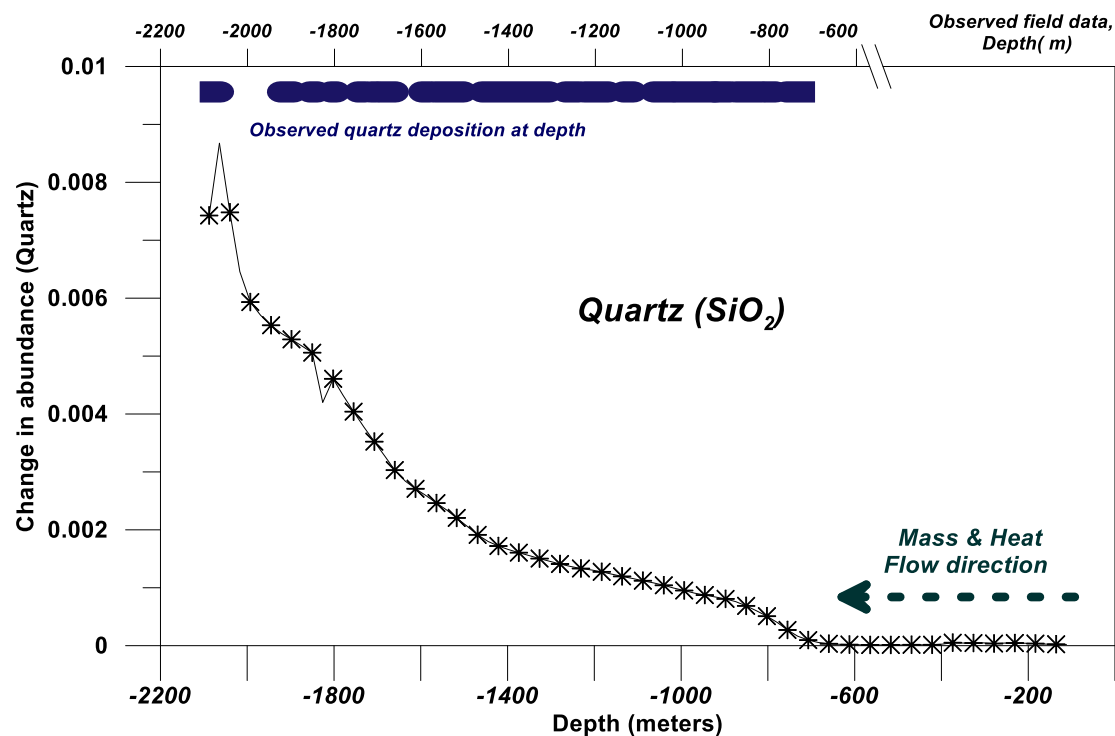


Figure 5.20: Comparison between the results of quartz after 1000 years of simulation and observed field data.

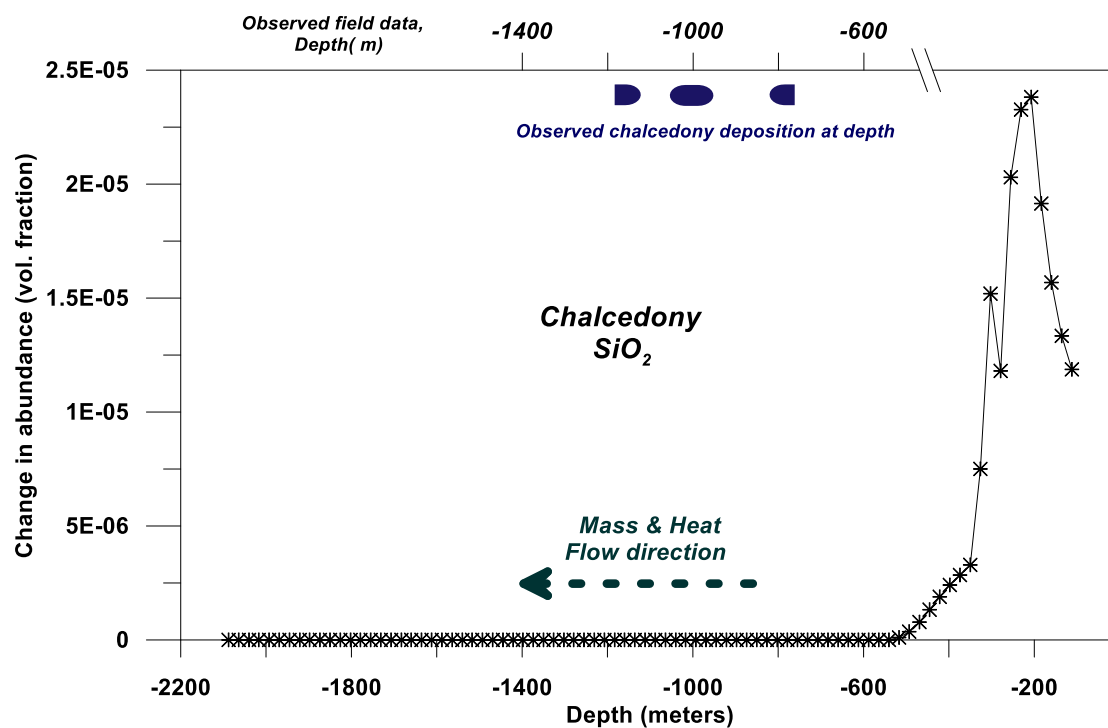


Figure 5.21: Comparison between the results of chalcedony after 1000 years of simulation and observed field data.

The solubility of silica minerals in most hydrothermal and geothermal systems is generally low (Fournier, 1985). The concentration of SiO_2 in high enthalpy geothermal systems (200–350°C) is about 300–700 mg/kg and is governed by the solubility of quartz (Fournier & Rowe, 1966). Silica exhibits prograde solubility attribute where it becomes more soluble with an increase in fluid temperature. During cooling the saturation is reached with respect to amorphous silica at temperatures of $\sim 100 - 200^\circ\text{C}$. Deposition of amorphous silica is quite fast and may result in potential silica scaling when the temperature decreases below the saturation temperature (Arnórsson & Gudmundsson, 2003).

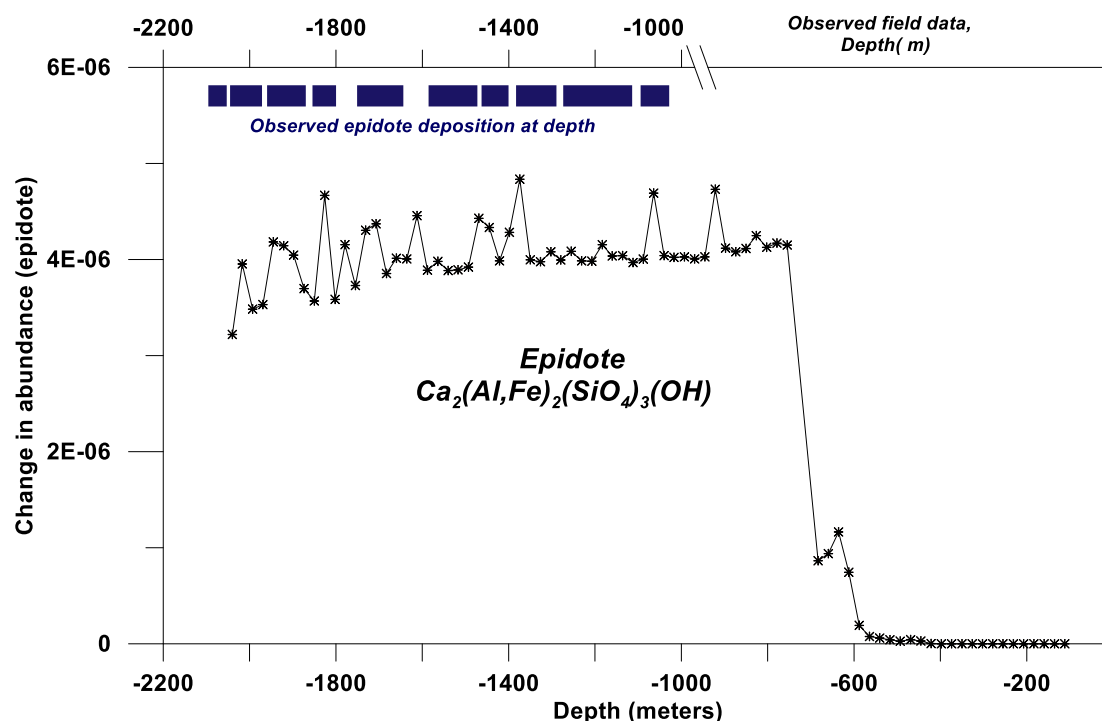


Figure 5.22: Comparison between the results of epidote after 1000 years of simulation and observed field data.

Epidote is one of the most common alteration minerals that occur in high geothermal systems of temperature $> 220^\circ\text{C}$ (Reyes, 2000). As shown in Figure 5.22, the occurrence of epidote matches field observations. Epidote is one of the sinks of Ca^{2+} in a hydrothermal environment, but there seems to be no clear relationship between the dissolution of diopside and the formation of epidote.

This points out to alternative sources of Ca^{2+} such as the water recharging the system or other dissolution of other silicate minerals not considered in the present simulation.

Figure 5.23 shows the net changes in porosity due to dissolution and precipitation processes for the 1000 years period used in the simulation. Below the depth of ~ 800 m, there is a gradual increase in porosity to the well bottom, which indicates the extent to which minerals were dissolved and carried alongside the rising thermal fluids. Above this depth, porosity decreases most probably due to precipitation of fine-grained clay minerals whose tangential packing arrangement reduces the porosity in a significant amount.

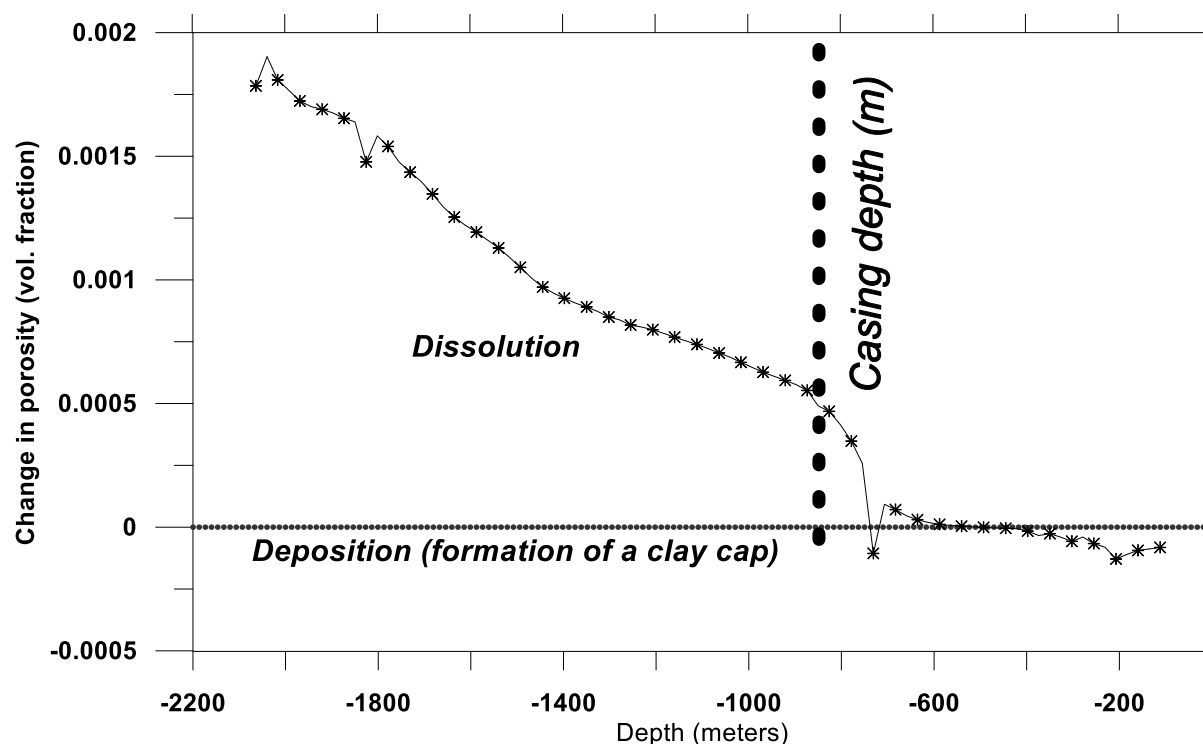


Figure 5.23: Change of porosity after 1000 years of simulation and the casing depth of MW-01.

5.6. Conclusion

The magma reservoir acts as a stable and protracted heat source for the thermal fluid circulation, and indeed the occurrence of prolonged hydrothermal activity seems to reflect the presence of a sizeable magma body at depth. Geochemical fluid can give a vital indicator of the kind of

hydrothermal activities that takes place, and this can help in to locate target areas for geothermal exploitation.

The analysis of major chemical elements from the samples obtained from discharging wells suggests that there are at least two sources of water within the geothermal system. These sources include a Na-dissolved CO₂ end-member on one end and SiO₂ end-member on the other end. An alternative interpretation is the existence of Cl-rich water considered to be mature thermal water is diluted progressively by these two end-members.

The δD and $\delta^{18}O$ stable isotopes show that the Menengai geothermal waters are influenced mainly by waters of meteoric origin. The δD and $\delta^{18}O$ values of boreholes samples appear to be depleted relative to the geothermal well and lake samples. However, samples from boreholes discharging warm water is more depleted of δD relative to $\delta^{18}O$ compared to other borehole samples, and seemingly they tap from much older water bodies. The concentration of helium in the fluids is sourced almost entirely from the upper mantle, and this implies that there exists a shallow magmatic body that releases some of its constituents to the geothermal system.

The reactive transport simulation results show that the secondary minerals seem to reproduce the observed field data reasonably well. This phenomenon is seen with epidote, quartz, calcite, mordenite, laumontite, chalcedony, and illite minerals. Calcite scaling in the system relies primarily on the presence of Ca²⁺ ions in the formation, and the dissolution of diopside minerals controls their concentrations, although other sources are possible.

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CHAPTER SIX

COMBINED INTERPRETATION AND DISCUSSION

6.1 Introduction

The Menengai caldera and the Olrongai areas are manifested by signatures of high-temperature geothermal resource that include relatively young volcanism represented by numerous recent eruptions within and outside the caldera. The occurrence of active fumaroles (with temperatures between ~62 and ~88°C), steaming ground and warm water boreholes (e.g., [Mungania & Lagat, 2004](#)), implies that the area overlays an active hydrothermal system. The occurrence of such hydrothermal systems inside the caldera has already been confirmed by drilling (e.g., [Omenda et al., 2014](#)).

The Bouguer anomaly reveals high-density values trending NNW-SSE with a dominant two-fold anomaly located to the north-northwest of the caldera and within the caldera. The low Bouguer anomalies are associated with Solai graben and Rongai plain located to the north-east and south-west areas, respectively. The high Bouguer anomalies appear to be confined within the caldera and the Olrongai area. This chapter gives a combined interpretation of the results obtained from this study focusing on the caldera and Olrongai areas.

6.2 Menengai caldera

The nature of post-caldera emplacement of trachytic formation and ash-flow tuffs is quite complex and is related to many eruptions. The products of these eruptions cover the caldera floors and make it difficult to delineate important geological structures essential for geothermal exploitation

without the help of geophysical methods. The Menengai caldera is formed on a large exposed joint at the tailcrack culmination of the northerly striking dextral normal oblique slip-fault region (Chorowicz, 2005), which seems to be connected with a series of NNE trending normal faults. These faults cut across the caldera wall, and they perhaps contributed to the existence of the caldera.

This large open crack curls across the caldera from east to west and appears to have acted as a conduit of eruptives to the surface as supported by an underlying dense body. The faults consequentially render the caldera wall incomplete as they cut through the northeastern rim by north-south trending segment graben (McCall, 1967), marking the southern end of Solai graben. This graben runs along a swarm of very young faults that cut the rift valley floor from Menengai northwards. Kinematic assessment inside the caldera relates these faults with strike-slip and oblique-slip transfer faults, commonly seen across the central Kenyan rift (Riedl et al., 2015). The activation of these regional faults allowed trachytic magma to ascend to shallower levels, which then flows laterally being accommodated by permeable fault zones.

The idea of interpreting the two geophysical methods together was to give a better understanding of the structures controlling the geothermal system in the study area. The two methods were based on density and resistivity distribution below the earth surface. To achieve this, the resistivity profiles were chosen to intersect the high-density body in order to support the obtained results hence to give more insights into major subsurface structures governing the geothermal system (Figure 6.1). Since the two resistivity profiles intersect each other, they appear to image the same geophysical structures.

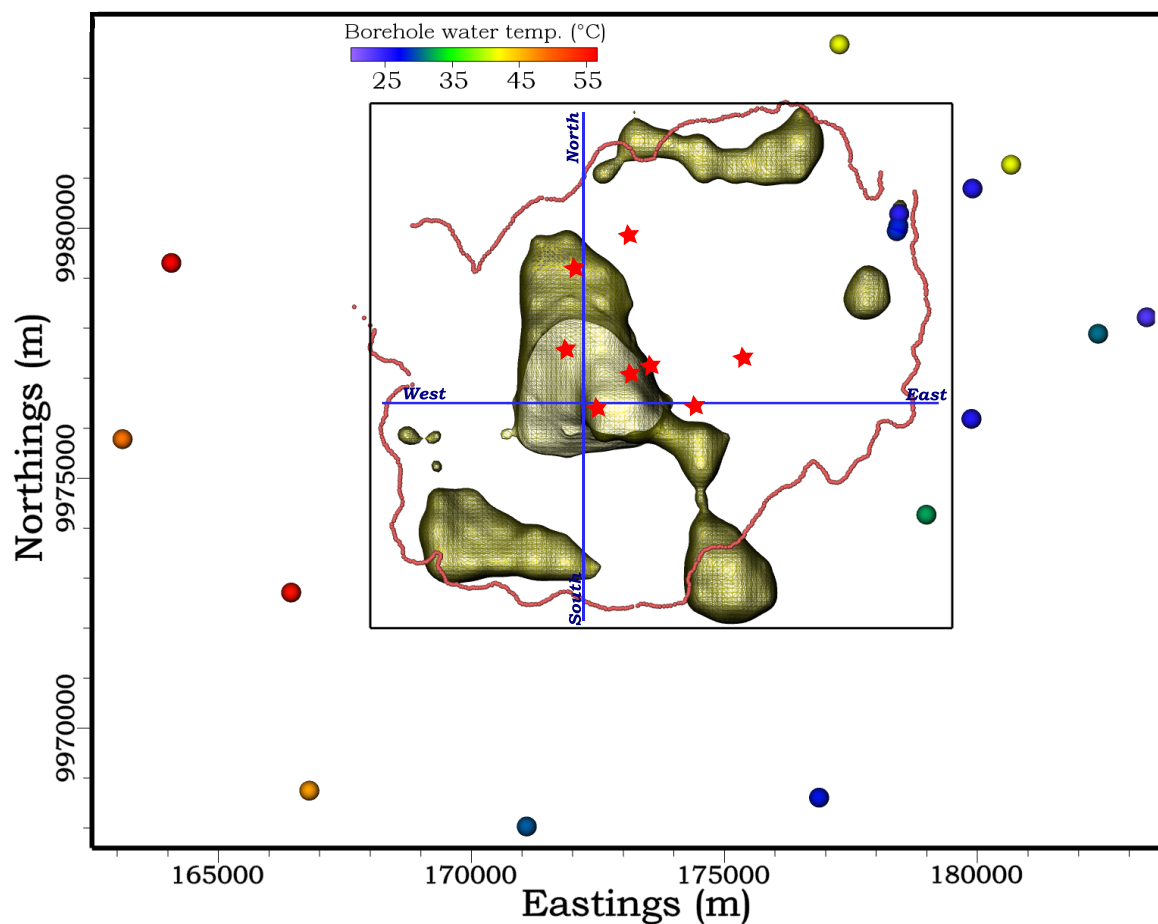


Figure 6.1: Menengai caldera map showing the location of the 3-D MT inversion profiles (blue line) along West-East and South-North directions. The cross sections are shown in Figure 6.2 and Figure 6.3. A density isosurface of $2.55 \times 10^3 \text{ kg m}^{-3}$ extracted from 3-D gravity model is also plotted. Temperature (ranging from 20 – 57 °C) of water drawn from boreholes spread around the study area is indicated by colored dots. The fumaroles are presented as red stars.

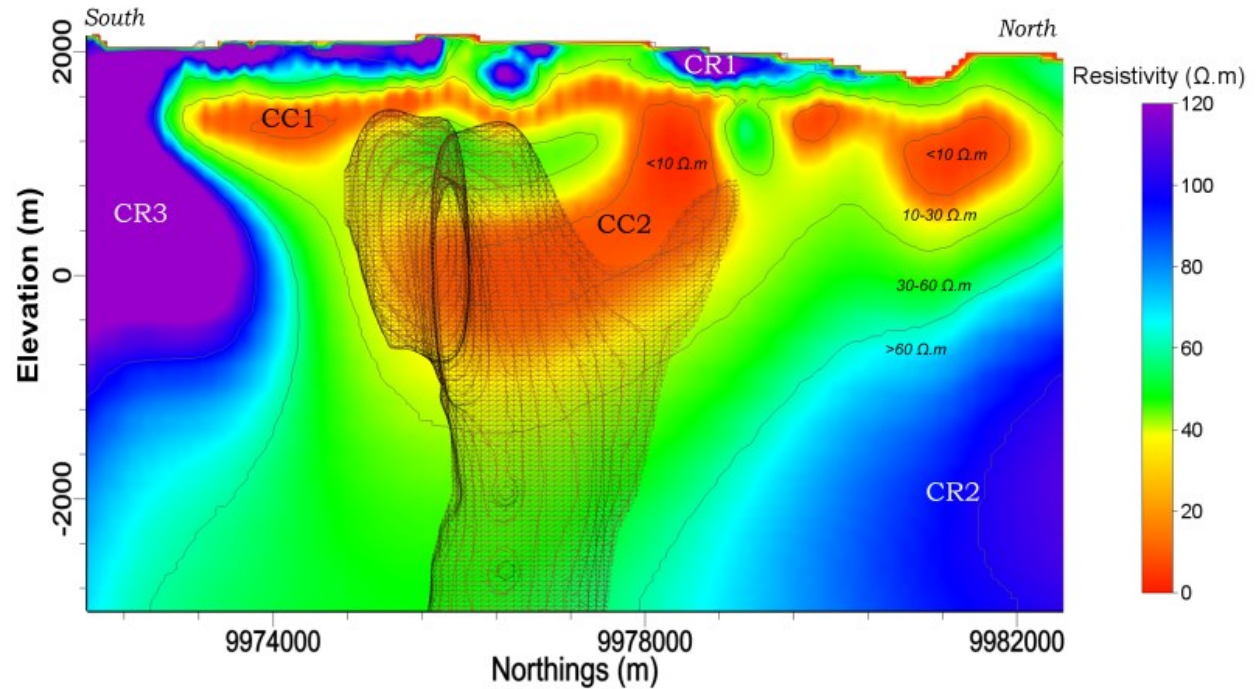


Figure 6.2: South-North cross-section of 3-D resistivity model and a density isosurface of $2.55 \times 10^3 \text{ kg m}^{-3}$ (shown as a mesh) extracted from the 3-D gravity model.

Figure 6.2 shows the south-north resistivity section that was obtained from the 3-D MT inversion and density isosurface of $2.55 \times 10^3 \text{ kg m}^{-3}$ taken from the 3-D gravity model. The conspicuous highly resistive body (CR3) to the south coincides with the caldera wall while the other resistive body (CR2) is observed on the northern part of the caldera, which can be interpreted as a cold formation. The other obvious feature is the conductive layer (CC1) which appears to connect a relatively deeper conductor (CC2). This CC2 appears to lie on top of the dense body from the gravity model suggesting a close relationship between the bodies.

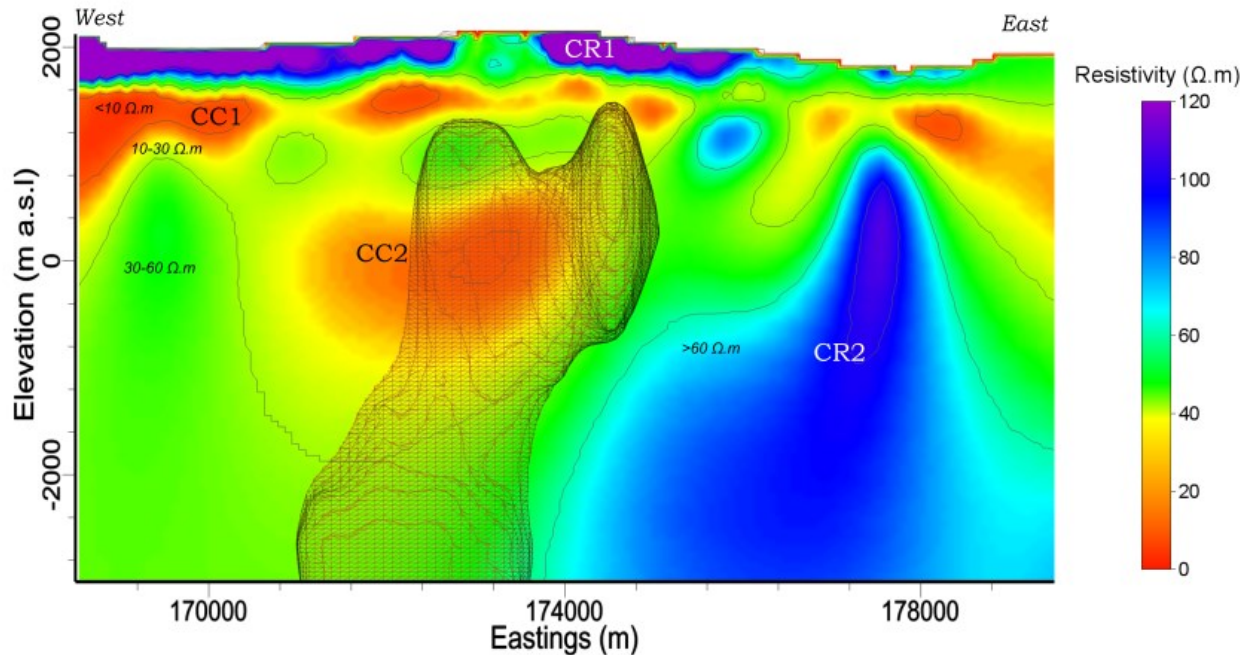


Figure 6.3: West-East cross-section of the 3-D resistivity model and a density isosurface of $2.55 \times 10^3 \text{ kg m}^{-3}$ (shown as a mesh) extracted from the 3-D gravity model.

The cross-section in Figure 6.3 is aligned in a west-east direction and displays noticeable resistivity structures. One of these structures is a resistive layer (CR1, $>120 \text{ } \Omega\cdot\text{m}$) that can be interpreted as unaltered formation near the ground surface, which spreads across, particularly to the west of the caldera floor. A deeply seated resistive body (CR2, $>100 \text{ } \Omega\cdot\text{m}$) is found on the eastern part which perhaps can be regarded a cold formation. Conductors CC1 and CC2 are imaged the same way as in Figure 6.2.

The high resistive body CR2 in Figure 6.4 appears to surround the high-density anomaly, and this suggests the occurrences of the two geophysical structures are interrelated. This dense structure is located within the conductive zone, away from the resistive structure and this might be caused by the deep circulation of chemically altered hydrothermal fluid. Such deep fluid movement is likely to occur with the help of geological structures that can allow free fluid motion, and this can be corroborated by the occurrence of fumarolic activities confined within this region. However, the dense structure and the resistive body (CR3) appear to coexist on the southern segment of the caldera wall that forms the remaining part of the Menengai shield volcano (Leat, 1991).

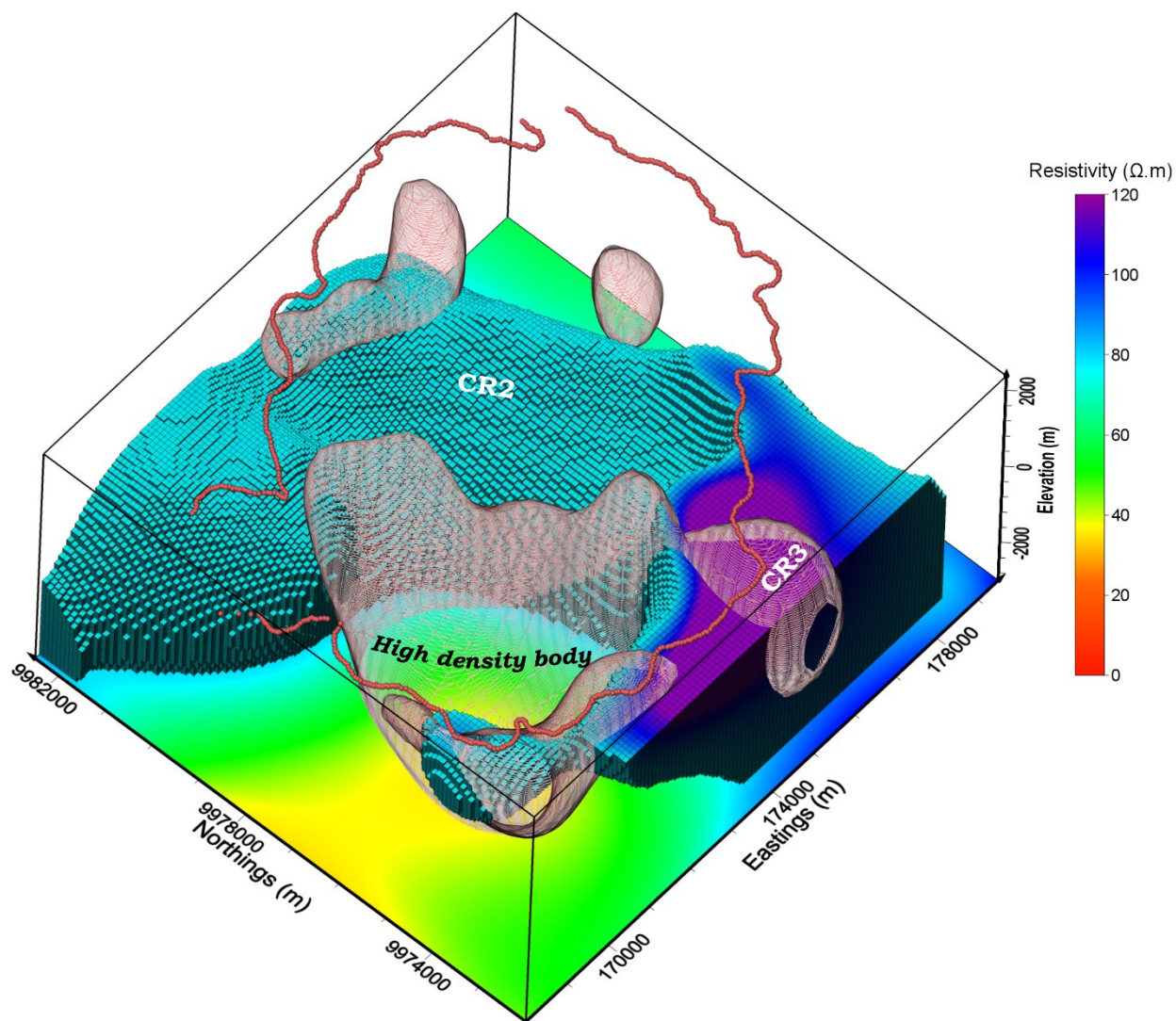


Figure 6.4: 3-D perspective view of the resistive isosurface (blue blocks (CR2), >90 Ω.m) and 3D geometry of the dense body isosurface (transparent mesh, >2.55 × 10³ kg m⁻³) of caldera area. The planar view is a horizontal slice of density anomaly at 3000 m bsl.

Figure 6.5 presents a perspective view of 3-D gravity and resistivity models of the caldera. The main features identified include the near surface conductor overlaying the dense body and stretching to the west interpreted as the upflow and outflow zones, respectively. In Figure 6.1, the distribution of boreholes with measured water temperature ranging from 20-57 °C is presented. Those groundwater boreholes with higher temperatures are located to the west and south-west of the caldera and perhaps fed by thermal fluid outflowing from the Menengai geothermal system.

The outflow might then be responsible for the formation of the alteration zone to the west of the caldera as shown in Figure 6.3 and Figure 6.5.

The deeper conductive zone positioned together with the dense body is likely to be the heat source. Most of the production wells are located above or close to this deeper conductor, and the hottest wells have intercepted temperatures close to 400 °C (GDC, 2018). The proposed recharge zone lies within a resistive region which might be caused by cold water incursion. The location of this zone also is in line with the assumption that the general flow of fluid from this region is northwards as supported by piezometric and structural studies (Mungania & Lagat, 2004).

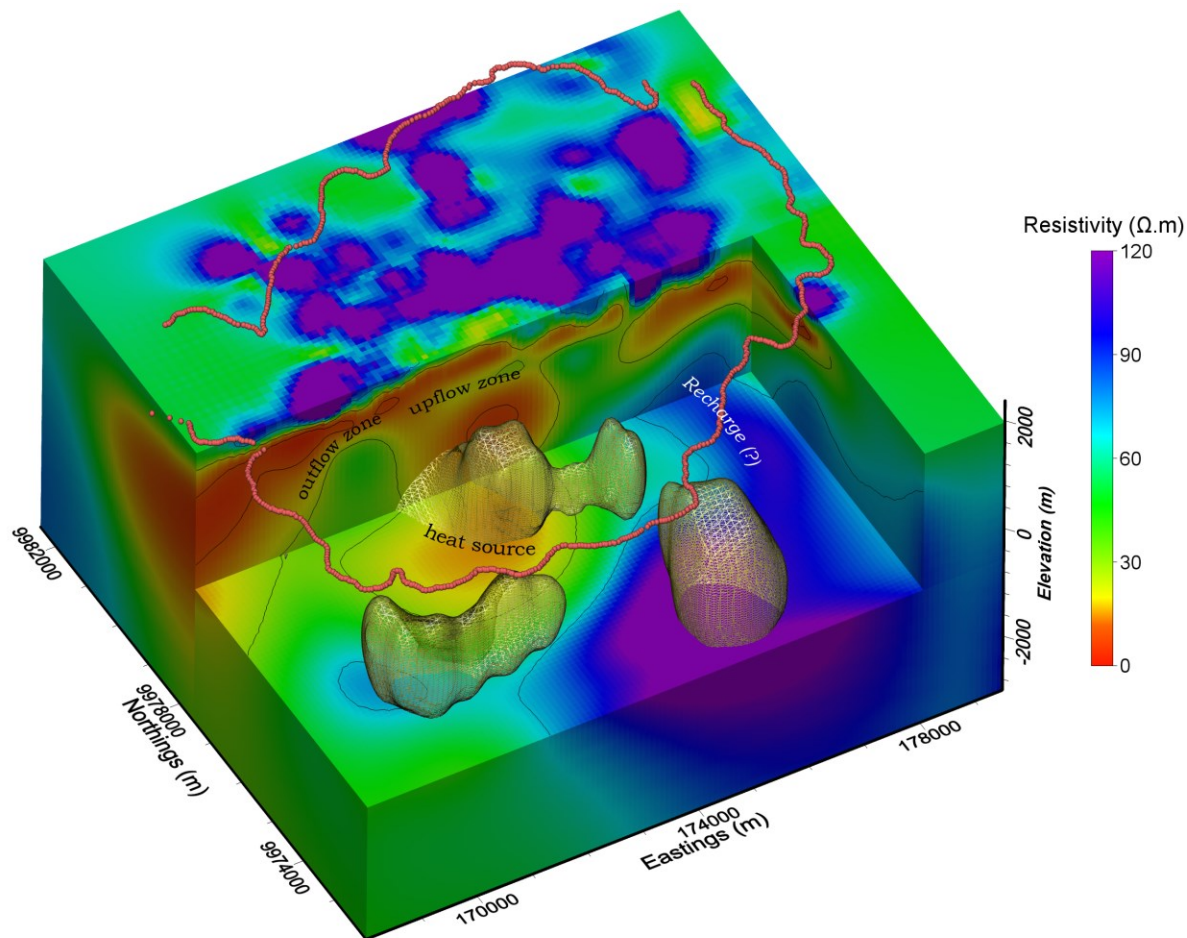


Figure 6.5: A perspective model of the 3-D gravity and magnetotelluric inversions of the Menengai caldera area. The block represents magnetotelluric results whereas the transparent mesh indicates the dense body isosurface ($2.55 \times 10^3 \text{ kg m}^{-3}$).

Figure 6.6 shows that the high $^3\text{He}/^4\text{He}$ coincides with the highly conductive body beneath Menengai caldera. The anomalous $^3\text{He}/^4\text{He}$ isotopes and the conductive layer appear to sit on top of the dense body. This implies that the highly conductive zone is a cooling body of magma or a partial melt that was emplaced as a dike and is connected to a deeper magma chamber. The magma chamber taps from the mantle and hence the high $^3\text{He}/^4\text{He}$ anomaly appears.

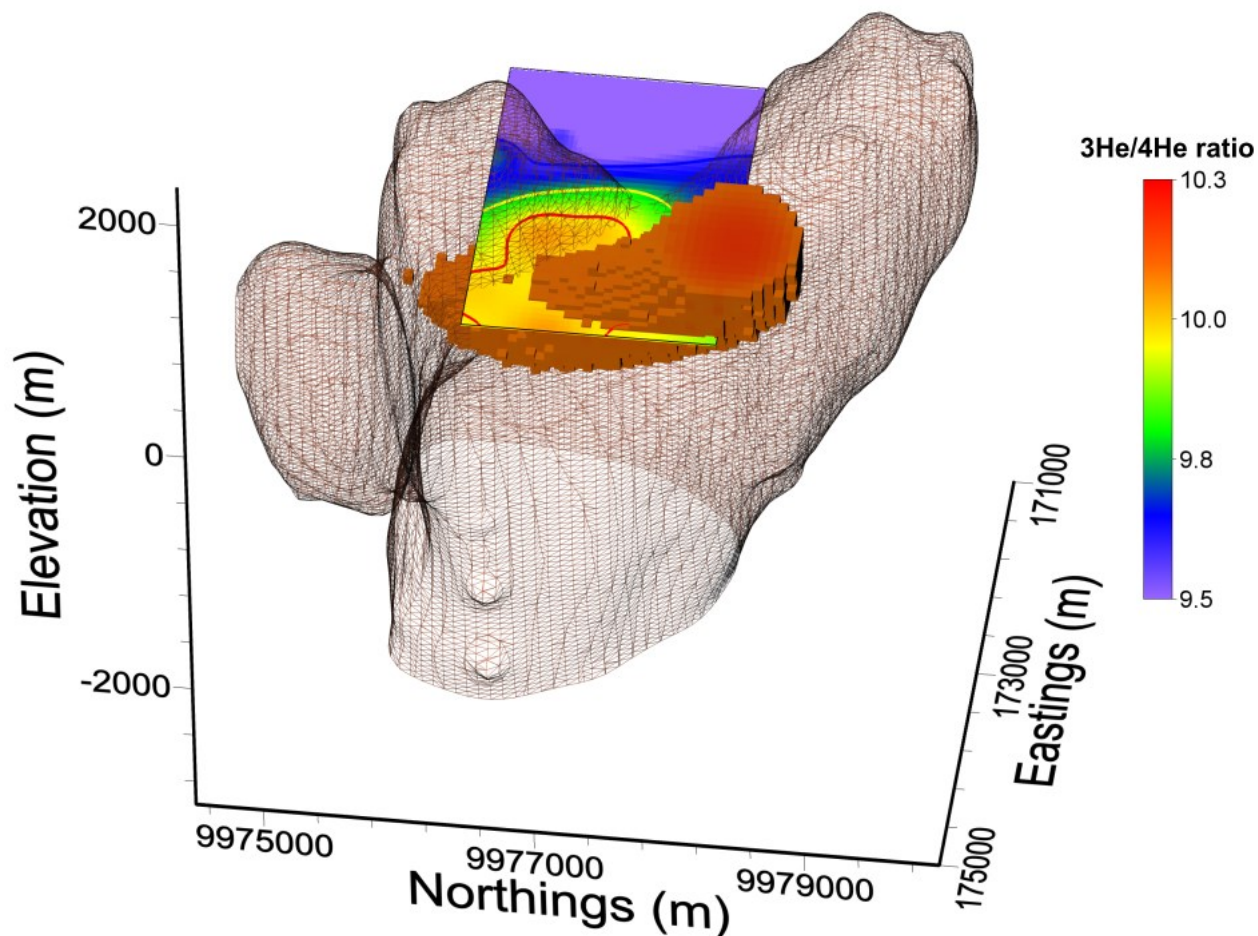
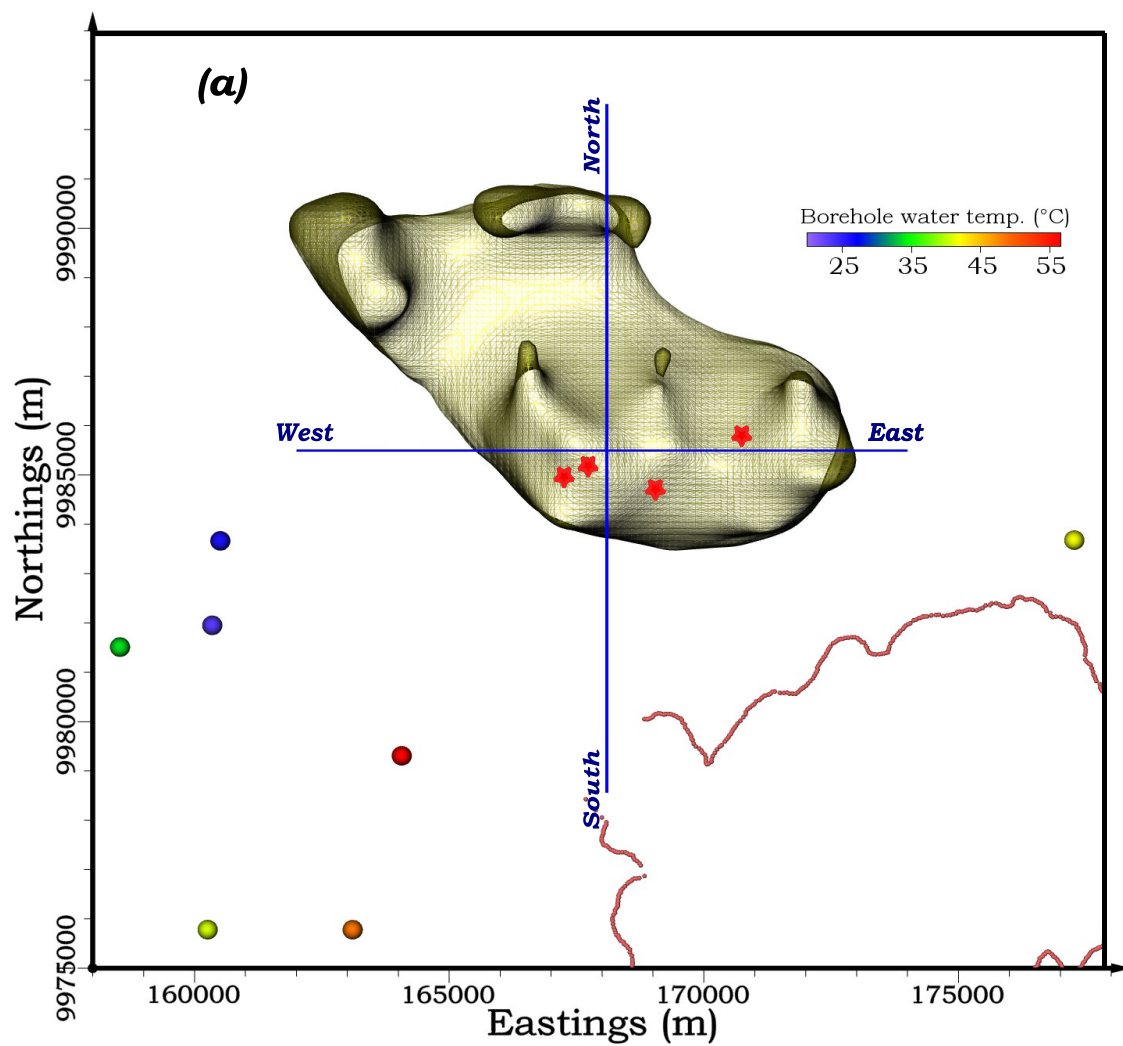


Figure 6.6: A joint model 3-D gravity and magnetotelluric inversions, and $^3\text{He}/^4\text{He}$ isotopes contour map (referenced at an elevation of -500 m a.s.l) of the Menengai caldera area. The block represents magnetotelluric result ($<10 \Omega.m$) whereas the transparent mesh indicates the dense body isosurface ($2.55 \times 10^3 \text{ kg m}^{-3}$).

6.3 The Olrongai area



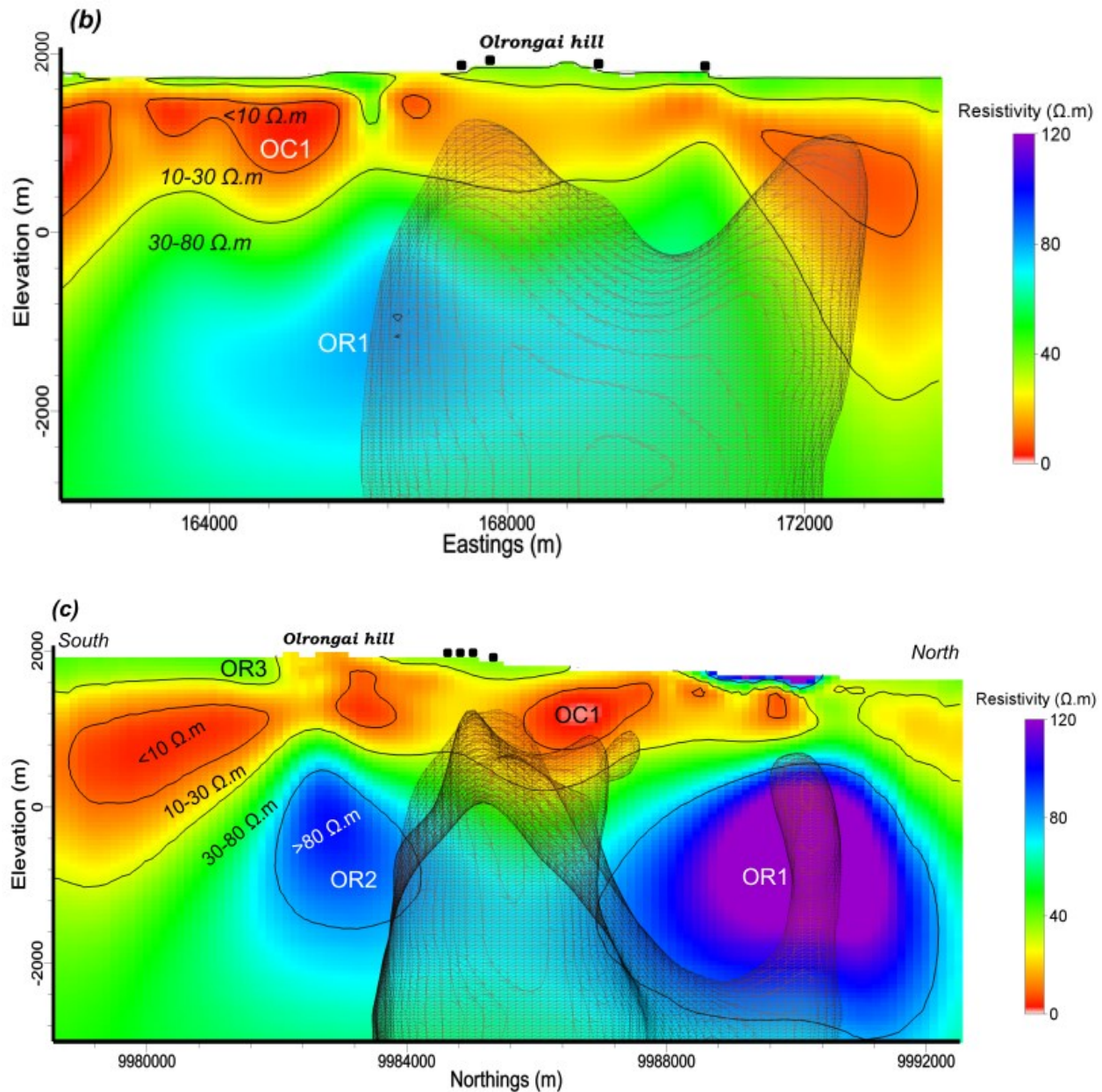


Figure 6.7: Cross sections of 3-D resistivity model (b and c) along profiles ‘west-east’ and ‘south-north’ indicated on (a) in the Olrongai area. On each of the cross-section, a density isosurface of $2.5 \times 10^3 \text{ kg m}^{-3}$ extracted from the 3-D gravity model is also plotted (also plotted on (a)). Circles in Figure 6.7a show temperature (ranging from 20 – 57 °C) of water drawn from boreholes spread around the study area. The orange line, the red stars (shown as black squares in b and c) and brown lines in Figure 6.7a are the Menengai caldera rim, fumarole locations, and geological faults, respectively.

Figure 6.7b-c show the west-east and south-north resistivity cross-section extracted from the inversion model and density isosurface of $2.5 \times 10^3 \text{ kg m}^{-3}$ extracted from the density model. The resistivity sections describe the main features of the Olrongai subsurface structures. Generally, the shallow depths are characterized by a resistive layer (OR3) of more than $60 \text{ } \Omega\cdot\text{m}$ followed by pockets of conductive zones (OC1) with a resistivity value below $10 \text{ } \Omega\cdot\text{m}$ sandwiched within transitional areas, whose resistivity range from $10 - 30 \text{ } \Omega\cdot\text{m}$. Beneath this layer is a region with resistive values greater than $30 \text{ } \Omega\cdot\text{m}$ (OR1). In this region, resistivity increases gradually up to $>120 \text{ } \Omega\cdot\text{m}$ away from the dense body, and particularly to the north of the dense body as shown in Figure 6.7c. The upper boundary of OR1 zone ($> 30 \text{ } \Omega\cdot\text{m}$) varies across the study area and ranges between elevations of 2000 m a.s.l (right above OR1) and sea level in the east and west (Figure 6.7b), and south (Figure 6.7c) directions.

The low resistivities observed in the near-surface section is caused inferably by hydrothermal alteration of the young rocks of Athinai Trachyte (and probably also Olrongai Ignimbrites) formations, due to a higher temperature and acidic fluid emanating from the magma body. The resistivity of this layer, commonly known as the smectite zone (mainly in the range of $1-10 \text{ } \Omega\cdot\text{m}$), is usually determined by the type and intensity of alteration process, transformed by the amount of water saturation and temperature (e.g., Anderson et al., 2000; Ussher et al., 2000). At depths greater than $\sim 1000 \text{ m}$, the resistivity values increase most probably due to changes in the clay-type ensuing from hydrothermal alteration process. The alteration products at higher temperatures are less conductive and are typically illitic and chloritic clays replacing the smectite, the latter is unstable at higher temperatures. Therefore, the hotter part of the geothermal system beneath the highly conductive clay cap, interpreted as the heart of the reservoir, is characterized by relatively higher resistivities. However, away from the active part of the system, resistivity decreases as the depth increases. The relatively low resistive region to the west (see Figure 6.7b, Figure 6.8 and Figure 6.9) can be interpreted as the recharge zone for the geothermal system.

Figure 6.7a shows distributions of boreholes with measured water temperature ranging from $20 - 57 \text{ } ^\circ\text{C}$. Those with higher temperatures are located in the southern part of the area and beyond. Based on the resistivity results of a south-north cross-section of the area shown in Figure 6.7c and

the north-south trending structures, these boreholes appear to be tapping from the outflow zone of the Olrongai geothermal system.

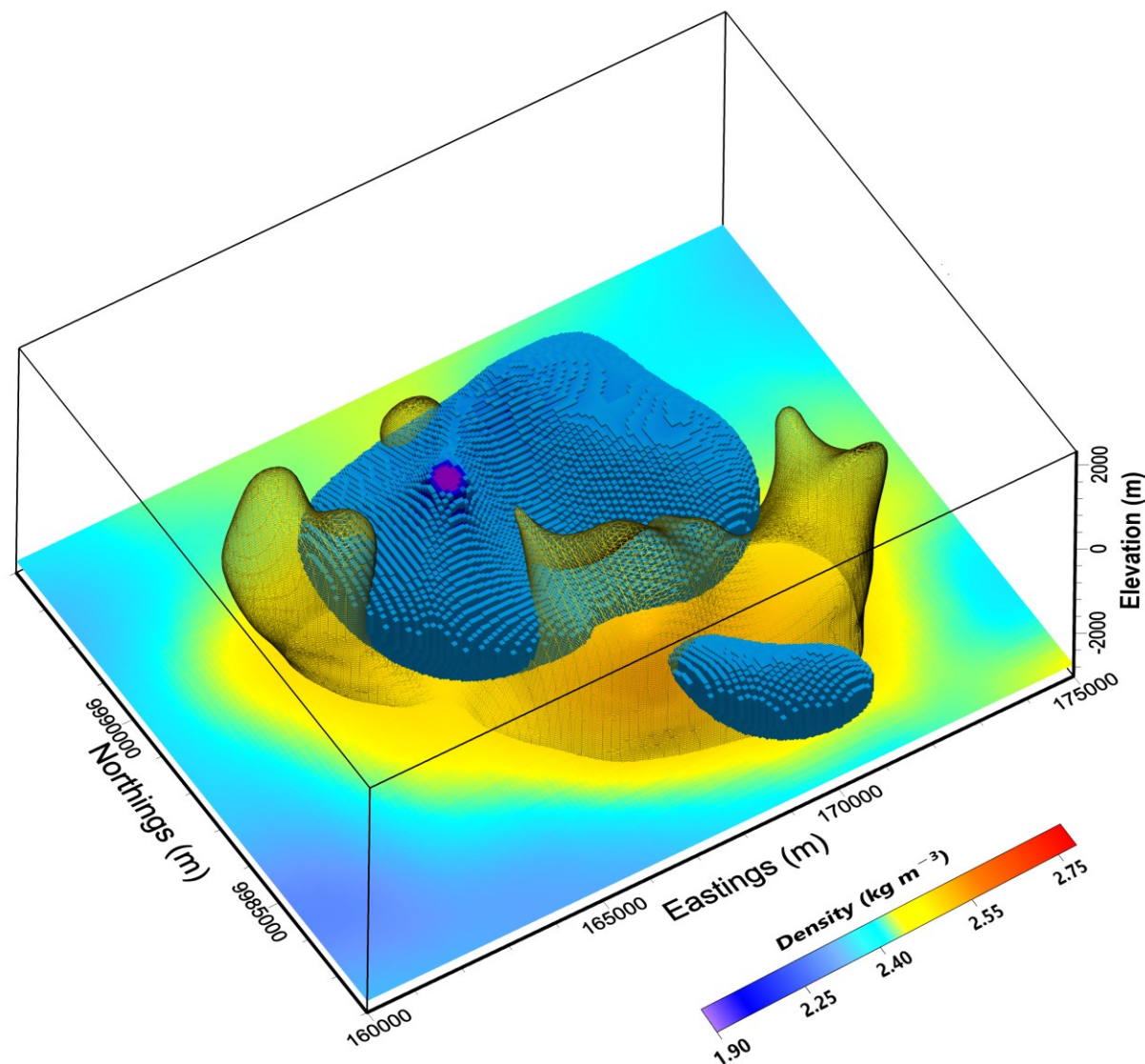


Figure 6.8: 3-D perspective view of the resistive isosurface (blue blocks, $>90 \Omega.m$) and 3D geometry of the dense body isosurface (transparent mesh, $>2.5 \times 10^3 \text{ kg m}^{-3}$) of Olrongai area. The planar view is a horizontal slice of density anomaly 3000 m bsl.

One of the most attractive features of the gravity and MT inversion models is the relationship in the geometry of the dense body relative to the resistive body. Figure 6.8 displays the 3-D view of the high resistive isosurface ($>90 \Omega.m$) and the high density ($>2.5 \times 10^3 \text{ kg m}^{-3}$) bodies from the inversion models. The dense body appears to mantle the resistive part showing that the two have different formation properties. There is evidence of volcanic fluids rising beneath the summit of

the dense body as interpreted by the presence of fumarolic activities and low resistivity in the region. This decrease in resistivity may have resulted from the effect of acidic waters formed by the interaction of local groundwater with rising steam and gases of volcanic origin causing intense hydrothermal alteration of the volcanic rocks at the upflow zone and fades laterally away at the outflow. However, there is no indication that the volcanic-hydrothermal system extends laterally much to the north of the study area. Figure 6.9 shows a clay layer that deepens away from the dense body and suggesting that the fluid flow away from this central part. The proposed recharge zone is based on the assumption that the general flow of fluid from this region is northwards as supported by piezometric and structural studies ([Mungania & Lagat, 2004](#)).

Mungania & Lagat (2004) carried out fumarole flow rates, and heat loss measurements in this area from shallow temperature gradient. The temperatures were obtained on the surface and at the depths of 25 cm, 50 cm, and 100 cm in the holes they drilled. The natural heat flow results showed that areas with intense thermal activity near the summit of Olrongai hill coincided with the locations of fumaroles. Beneath this thermally anomalous region is the heart of the dense body and extends in east and west directions, which can be interpreted as the Athinai Trachytes ([Leat, 1991](#); [Macdonald et al., 2011](#)).

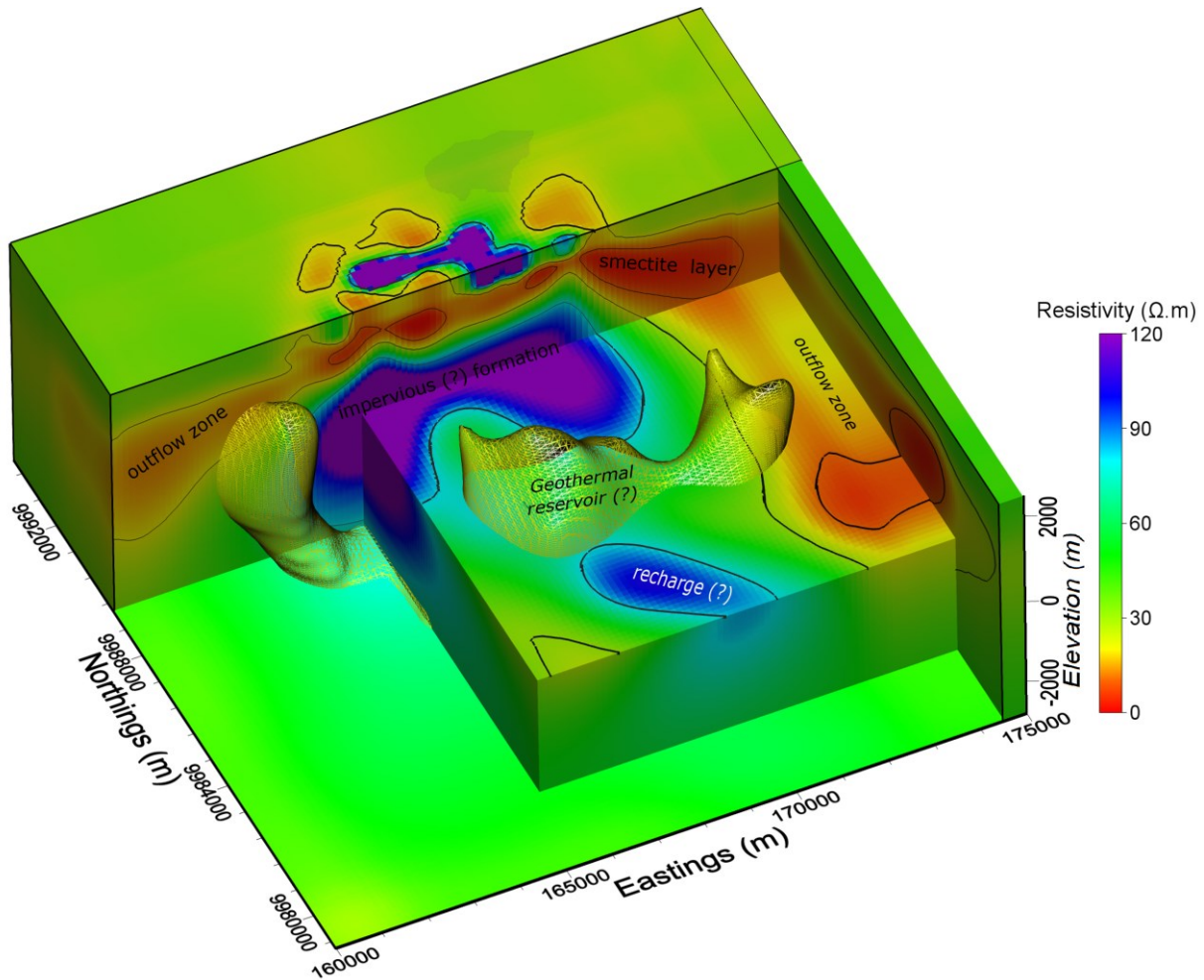


Figure 6.9: A perspective model of 3D gravity and magnetotelluric inversions in the Olrongai area. The block model represents magnetotelluric results whereas the transparent mesh indicates the dense body isosurface ($2.5 \times 10^3 \text{ kg m}^{-3}$).

The Olrongai area is dominated by dry, perched-aquifer and thermally anomalous boreholes, where the productive ones are very shallow with low-yield ($< 6 \text{ m}^3/\text{hr}$) that get depleted fast since the deeper formations are impervious (Mungania & Lagat, 2004). The aquifers of the low-yield boreholes are located at the contacts between tuff and ignimbrite beds. The perched ones are adjacent to the Olbanita swamp (Figure 2.3), which lies right above the northern end of the highly resistive body (Figure 6.6c & Figure 6.9). This swamp is fed by a seasonal river and owes its existence to impervious ignimbrite beds that have been affected by hydrothermal alteration. The ignimbrite is densely welded, with rheomorphic flow structures and consists of trachytic pumices

and lithic clasts mostly trachyte or trachyphonolite lava (Leat, 1991). Ignimbrite usually is characterized by low density ($1.21\text{--}1.93 \times 10^3 \text{ kg m}^{-3}$) and high porosity (18–51%) (e.g., Moon, 1993) and from the model we infer that the high resistive and relatively low-density body beneath the Olrongai area is predominantly composed of an impervious Olrongai Ignimbrite formation with a conservative computed volume of $\sim 63 \text{ km}^3$. Nonetheless, Leat (1991) reported the presence of much younger air-fall pumice and tuffs that were emplaced alongside Athinai Trachytes. Alternatively, this resistive body could as well be interpreted as cold formation. In this case, the water from the swamp Olbanita might have seeped down through the existing faults possibly recharging the geothermal system.

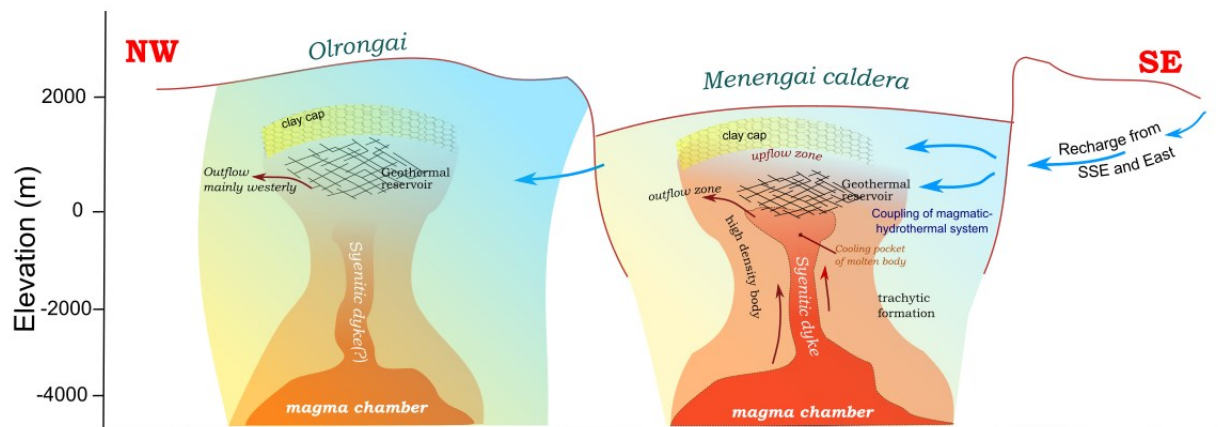


Figure 6.10: A conceptual model of the Menengai geothermal area.

Figure 6.10 shows a generalized conceptual model of the Menengai geothermal area. From the model, the water feeding the geothermal reservoir enters from SSE and East of the study area. It then flows in the NNE direction along the eastern rim of the caldera and the Olrongai tectonic structures recharging both the systems.

6.4 Conclusion

This chapter presents the integrated interpretation of various methods used in this study to characterize the subsurface structures of the Menengai geothermal area. From the inversion results

of gravity and MT data, the relationship in the geometry of the dense body could be identified relative to the resistive body. The high-density body of Olrongai appears to mantle the resistive part while the resistive body inside Menengai caldera seems to surround a high density structure. This relationship shows that the two geophysical structures have different properties. The resistive body is interpreted as cold or relatively impervious structures. A cold formation/ structure would suggest a possible recharge to the geothermal reservoir.

The presence of fumarolic activities and low resistivity in the region is attributed to the rise of volcanic fluids beneath the summit of the dense body where the heat source is believed to be hosted. This decrease in resistivity may have resulted from the effect of acidic waters formed by the interaction of local groundwater with rising steam and gases of volcanic origin causing intense hydrothermal alteration of the volcanic rocks at the upflow zone and fades laterally away at the outflow.

The gases appear to emanate from the mantle as confirmed by the presence of anomalously high $^3\text{He}/^4\text{He}$ from well discharge chemistry. The anomalous $^3\text{He}/^4\text{He}$ isotopes from the geothermal wells located on top of the dense body imply that the highly conductive zone is likely to be a cooling body of magma or a partial melt emplaced as a dike and is connected to a deeper magma chamber.

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CHAPTER SEVEN

GENERAL CONCLUSIONS

7.1. Introduction

This study attempts to establish essential sources of fluids feeding Menengai geothermal system, and also, it looks at their movement and interaction with the lithology. The objective was achieved by applying an integrated approach involving geophysical, geological and geochemical techniques as well as reservoir properties. The geology and structures were reviewed, and the lineaments were extracted from satellite images to corroborate the understanding of local structures. Subsurface density distribution study using gravity data was carried out to help infer the extent and geometry of heat sources. Resistivity study was used to help refine the geometry of heat source and also define the extent of the geothermal reservoir. Recharge mechanisms were inferred based on stable isotope geochemistry and piezometric information while magmatic contribution was assessed using Helium isotope ratios. Finally, reactive geochemical modeling was then conducted to determine the geochemical processes occurring in the reservoir.

7.2. Field characteristics

The Menengai geothermal area is located in the central segment of the Kenya rift and lies within a complex geological and structural setting, which associates with the formation of the Kenya dome. The Menengai caldera is the most noticeable feature and also a significant source of the geothermal resource. The local surface geology is composed mostly of unconsolidated low-density volcanic soil and sediments. Tuffs and ignimbrites also cover a considerable portion of the surface somehow associated with limited outcrops of phonolites, trachytes, and basalts. Two rift floor

volcano-tectonic axes that are important in controlling the geothermal system in the study area are the Molo and the Solai TVA. The primary trend of the tectonic structures in the study area is N-S, NE-SW, and NW-SW and similar to the rest of Kenya Rift.

High-density values in the analysis of lineaments are concentrated mostly in areas covered by phonolites and trachytes and more so to the NE area where the rocks are reasonably well exposed. Most of the faults are located adjacent to the zones with average to high-density values. The orientation of these lineaments corresponds pretty much to the faults, and their general trend matches the regional tendencies of faults. Six recharge sources models are reviewed thought to contribute to the hydrology of the Menengai geothermal system. These sources include; shallow groundwater and runoff from the east, deep westerly groundwater flow, near-surface flow from L. Nakuru, deep northerly flow from L. Naivasha, the contribution of paleolake, and magmatic contributions.

7.3. Gravity data analysis and interpretation

An improved Bouguer anomaly map of the Menengai area using updated gravity dataset presented here reveals a dominant two-fold high Bouguer anomaly underlying the caldera and Olrongai areas, and is associated with the occurrence of young trachytic rocks and active fumarolic activities. A 3-D gravity inversion for predicting the geothermal heat source architecture within the Menengai caldera and Olrongai is also presented. The model provides an in-depth insight into the subsurface structure from potential field perspective by constraining deep bodies in 3-D space.

The host-rock boundary is highlighted by a strong density gradient that suggests a precipitous contact between two lithologies delineated by tectonic faults, more so evident in Olrongai. The inversion model reveals information about the undulation of the dense bodies beneath Olrongai hill and Menengai caldera. The location of the two main high-density structures in the shallower portions of the model corresponds with eruption centers, manifested by the existence of younger volcanic rocks and fumarolic activities, and its geomorphology at depth is interpreted to embay the heat source crucial for the geothermal system.

7.4. Resistivity studies using the MT method

Two 3-D resistivity models were constructed separately for the caldera and Olrongai areas. The caldera field was also handled independently for the 1-D and 2-D Occam's inversion. The use of 1-D and 2-D models were to help constraint the distribution of resistivity features within the study area. The three models were able to highlight the dominant geophysical structures which appear to control hydrothermal processes within the geothermal system.

The 3-D resistivity models of Menengai and Olrongai area shows three structures: (1) a shallow surface inhomogeneous resistive layer ($>100 \Omega.m$); (2) a near-surface conductive layer ($<10 \Omega.m$, and; (3) a deeper resistive body ($30-70 \Omega.m$). The near-surface conductor is confined above the elevation of 1000 m asl indicated by low resistivity ($< 10 \Omega.m$). Additionally, two distinct resistivity features are also observed; a dike-like conductive body that cuts through the deeper resistive layer at the heart of caldera, and a loop-shaped high resistive body that is observed in both the caldera and Olrongai areas. The near-surface conductive layer causes the upper-minimum in the apparent resistivity curves while the deeper conductive body is interpreted as a shallow cooling pocket of partial melt.

7.5. Isotope and water chemistry studies

The analysis of major chemical elements from the samples obtained from discharging wells suggests that there are at least two sources of water within the geothermal system. These sources include a Na-dissolved CO_2 end-member on one end and SiO_2 end-member on the other end. An alternative interpretation is the existence of Cl-rich water considered to be mature thermal water, which is diluted progressively by these two end-members.

The δD and $\delta^{18}O$ stable isotopes show that the Menengai geothermal waters are influenced mainly by waters of meteoric origin. The δD and $\delta^{18}O$ values of the borehole samples appear to be depleted relative to the geothermal well and lake samples. However, samples from boreholes discharging warm water is more depleted of δD relative to $\delta^{18}O$ compared to other borehole samples, and seemingly they tap from much older water bodies. The concentration of helium in the fluids is

sourced almost entirely from the upper mantle, and this implies that there exists a shallow magmatic body that releases some of its constituents to the geothermal system.

The reactive transport simulation results show that the secondary minerals seem to reproduce the observed field data reasonably well. This phenomenon is seen with epidote, quartz, calcite, mordenite, laumontite, chalcedony, and illite. Calcite scaling in the system relies primarily on the presence of Ca^{2+} ions in the formation, and the dissolution of diopside minerals controls their concentrations, although other sources are possible.

7.6. Combined interpretation and discussion

From the inversion results of gravity and MT data, a clear relationship in the geometry of the dense body relative to the resistive body exists. The high-density body of Olorongai appears to mantle the resistive part while the resistive body inside Menengai caldera seems to surround a high density structure. This suggests that the two geophysical structures have different properties. The presence of fumarolic activities and low resistivity in the region is attributed to the rise of volcanic fluids beneath the summit of the dense body. The anomalous $^3\text{He}/^4\text{He}$ isotopes from the geothermal wells located on top of the dense body imply that the highly conductive zone is likely to be a cooling body of magma or a partial melt emplaced as a dike and is connected to a deeper magma chamber.