

ADDIS ABABA UNIVERSITY
SCHOOL OF GRADUATE STUDIES
COLLEGE OF NATURAL AND COMPUTATIONAL SCIENCES
SCHOOL OF EARTH SCIENCES



**3-D INVERSION OF GRAVITY AND MAGNETIC DATA TO STUDY THE
CRUSTAL STRUCTURE AND DEPTH TO THE MOHO OF THE
CENTRAL MAIN ETHIOPIAN RIFT**

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This is to certify that the thesis prepared by **Denberu Besfat**, entitled: **“3-D Inversion of Gravity and Magnetic Data to Study the Crustal Structure and Depth to the Moho of the Central Main Ethiopian Rift”** and submitted in partial fulfillment of the requirements for the degree of Master of Science in Applied Geophysics complies with the regulations of the university and meets the accepted standards with respect to originality and quality.

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DECLARATION

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ABSTRACT

3-D inversion of gravity and magnetic data is a powerful method for predicting geology beneath the surface through quantitative interpretation of the geophysical problem at hand. 3-D inversion of these two potential field methods increases the amount of information that can be extracted from the observed gravity and magnetic data.

The ultimate goal of this MSc thesis is to implement 3-D inversions of airborne gravity and ground magnetic data. Accordingly, 3-D inversions of airborne gravity data collected with an 8 miles' line interval for geodetic purposes (2006-2007) and magnetic data collected by the researcher along with existing magnetic data collected by previous MSc students of AAU have been used to study the crustal structure and determine the depth to Moho of the central part of the Main Ethiopian Rift (CMER) and adjacent plateaus.

The analysis of the isostatic gravity anomaly and the 3-D gravity inversion had helped to compute and compile the depth to Moho map of the study area (CMER). Consequently, the crust beneath the rift floor was found to be thin and shallow (36-38km) while the crust beneath the South Western (SW) and South Eastern (SE) plateau regions is found to be thick and deep (38-44 km).

Moreover, the 3-D gravity inversion and the analysis of isostatic anomalies had revealed that the crust beneath the rift floor is associated with positive density contrasts caused by magmatic intrusions intruding the crust over this region while the crust beneath the SW and SE plateau regions is associated with negative density contrasts indicating the crust beneath the plateau regions is embedded in to a high density mantle material and hence producing a negative density contrasts.

The 3-D gravity and magnetic inversion in this MSc thesis work has finally produced 3-D subsurface density and susceptibility models of the study area (CMER).

The principal inherent non-uniqueness problems faced with the inversion of gravity and magnetic data that exist based upon a static potential field were attempted to be constrained through incorporating as little information as obtained from previous studies and through by carefully designing a global objective function.

This MSc thesis illustrates how the University of British Columbia Geophysical Inversion Facility (UBC–GIF) 3-D gravity and magnetic inversion codes can be applied to study crustal structure by establishing the necessary results.

Henceforth, the 3-D inversion executed in this MSc thesis is the UBC-GIF's inversion methodology which is 3-D regularized pixel-based algorithm proposed by LI and OLDENBURG (1996, 1998) where right rectangular prisms are used to define unknown density and susceptibility model parameters and the solution to the inverse problem is obtained by posing the inverse problem as a mathematical optimization process.

Key words: 3-D gravity inversion, 3-D magnetic inversion, inversion codes, Central main Ethiopian Rift (CMER), density model, susceptibility model, Moho depth, Crustal thickness, isostatic anomaly, magmatic intrusions, Ethiopia.

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LIST OF ACRONYMS

EARS	East African Rift System
ERS	Ethiopian Rift System
MER	Main Ethiopian Rift
NMER	Northern Main Ethiopian Rift
CMER	Central Main Ethiopian Rift
SMER	Southern Main Ethiopian Rift
3-D	Three-dimensional
2-D	Two-dimensional
UTM	Universal Transverse Mercator
M.S.L	Mean Sea Level
MSc	Master of Science
SW	South Western
SDFZ	Silti Debrezeit Fault Zone
EAGLE	Ethiopia Afar Geoscientific Lithospheric Experiment
BF	Border Fault
FDMS	Fentale–Dofen Magmatic Segment
BMS	Boset Magmatic Segment,
GTMS	Gedemssa–Tullu Moye Magmatic Segment
DEM	Digital Elevation Model
GPS	Global Positioning System
GIS	Geographical Information System
NW-SW	North West South East
EW	East West
NE	North East
NNE–SSW	North North East - South South West
WFB	Wonji Fault Belt

LIST OF ACRONYMS

AAE	Addis Ababa Embayment
BT	Boru Toru structural high
MK	Midre Kebd structural high
YTVL	Yerer–Tullu Wellel volcano-tectonic lineament.
Ma	Million annum
GSE	Geological Survey of Ethiopia
EIGS	Ethiopian Institute of Geological Surveys
UNDP	United Nations Development Programme
c.g.s	centimeter gram second system
m.k.s	meter kilogram second system
Gal	Galileo
mGal	miliGal
g.u	gravity unit
IGSN	International Gravity Standardization Network
nT	Nano Tesla
IGRF	International Geomagnetic Reference Field
SRTM	Shuttle Radar Topography Mission
RAPS	Radially Average power Spectrum
THDR	Total Horizontal Derivative
TDR	Tilt Derivative
SED	Source Edge Detection
FFT	Fast Fourier Transform

LIST OF ACRONYMS

AS	Analytic Signal
MS	Magmatic segment
MABF	Munesa Asela Border Fault

CHAPTER ONE

INTRODUCTION

1.1 General Background

The Main Ethiopian Rift (MER) that forms one segment of the East African Rift System (EARS) is considered as magma related rift that possesses all the different stages of rift development starting from rift initiation to break up and incipient oceanic spreading (EBINGER, 2005). Both from geophysical and geological perspectives, the MER is considered as an interesting region to investigate the volcano-tectonic deformation of the entire crust and even the lithosphere with an insight to oceanic spreading development (e.g., CORTI, 2009). The existence of several previous geological and geophysical data in the MER gives us a general perspective of the volcano-tectonic activity and crustal deformation processes taking place in the region.

However, 3-D gravity and magnetic inversion investigations that could provide an enormous amount of information about the crustal structure through the distribution of density and magnetic susceptibility properties in three spatial dimensions have not yet been previously conducted in the CMER with in an adequate number.

The present MSc thesis research work, which is going to be conducted in the Central Main Ethiopian Rift (CMER) is aimed at 3-D inversion and analysis of Gravity and Magnetic data to study the crustal structure and determine the depth to the Moho of the CMER and adjacent plateaus. The ultimate goal of this research is to perform 3-D gravity and magnetic data inversions for recovering 3-D density and magnetic susceptibility models of any possible magnitude and geometric complexity with in the study area.

The nature and spatial variation of subsurface tectono-magmatic deformations in extensional rift zones like that of the MER can be better understood from 3-D inversions of gravity and magnetic data.

The inversion method presented in this study uses the 3-D regularized pixel-based algorithm proposed by LI and OLDENBURG (1996, 1998) where right rectangular prisms are used to define unknown density and magnetic susceptibility model space parameters.

1.2 Description of the Study Area

1.2.1 Location, Areal Extent and Accessibility of the Study Area

The study area that comprises of the Central Main Ethiopian Rift (CMER) is about 240 km south of Addis Ababa (Fig 1.1). Geographically, it is bounded between $7^{\circ} - 8^{\circ} 30'00''$ N latitude and $38^{\circ} 00' - 39^{\circ} 15'00''$ E longitude or with UTM coordinates of 389812m -530000m easting and 770000m-940000m Northing.

The study area can be accessed through the Addis Ababa – Mojo- Alemtena – Butajira- Hawassa Shashemene asphalt road. The total area of the study area is estimated to be 23,205 km². Hawassa, Shashemene, Butajira, and Alaba Kulito are among the major towns located in the study area.

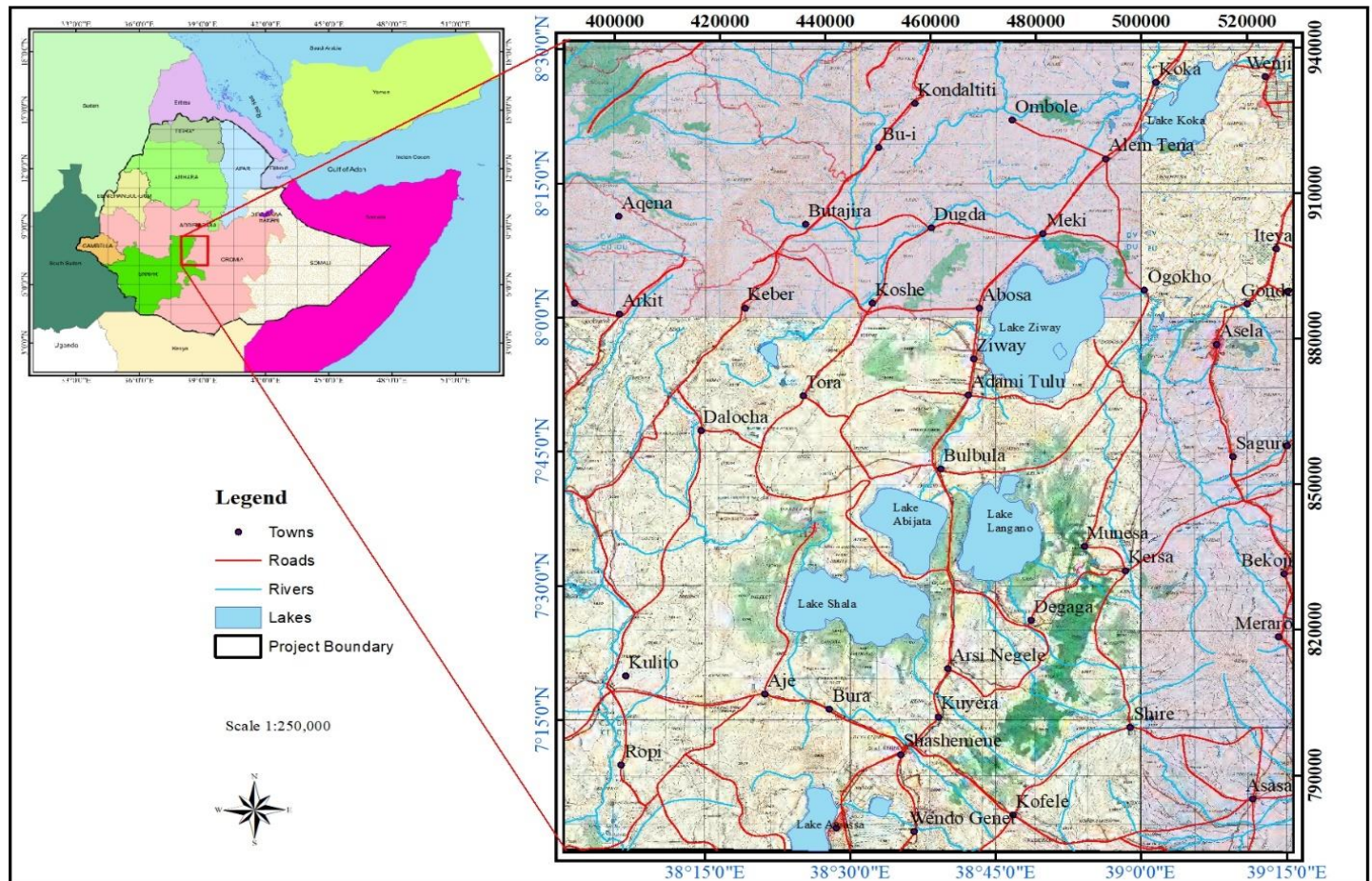


Figure 1.1 Location map of the study area including the major towns, lakes and road networks

1.2.2 Physiography of the Study Area

The study area (CMER) is bounded by highly elevated Arsi Bale mountain ranges (including Chilalo, Kaka, etc.) to the east and the Guraghe mountain ranges to the northwest. The elevation in the study area ranges from 1522 m in the central part to 4187 m above mean sea level (m.s.l) on the adjacent plateaus. The study area comprises most of the rift valley lakes including Lake Koka, Lake Hawassa, Lake Langano, Lake Abiyata, Lake Shalla and Lake Ziway.

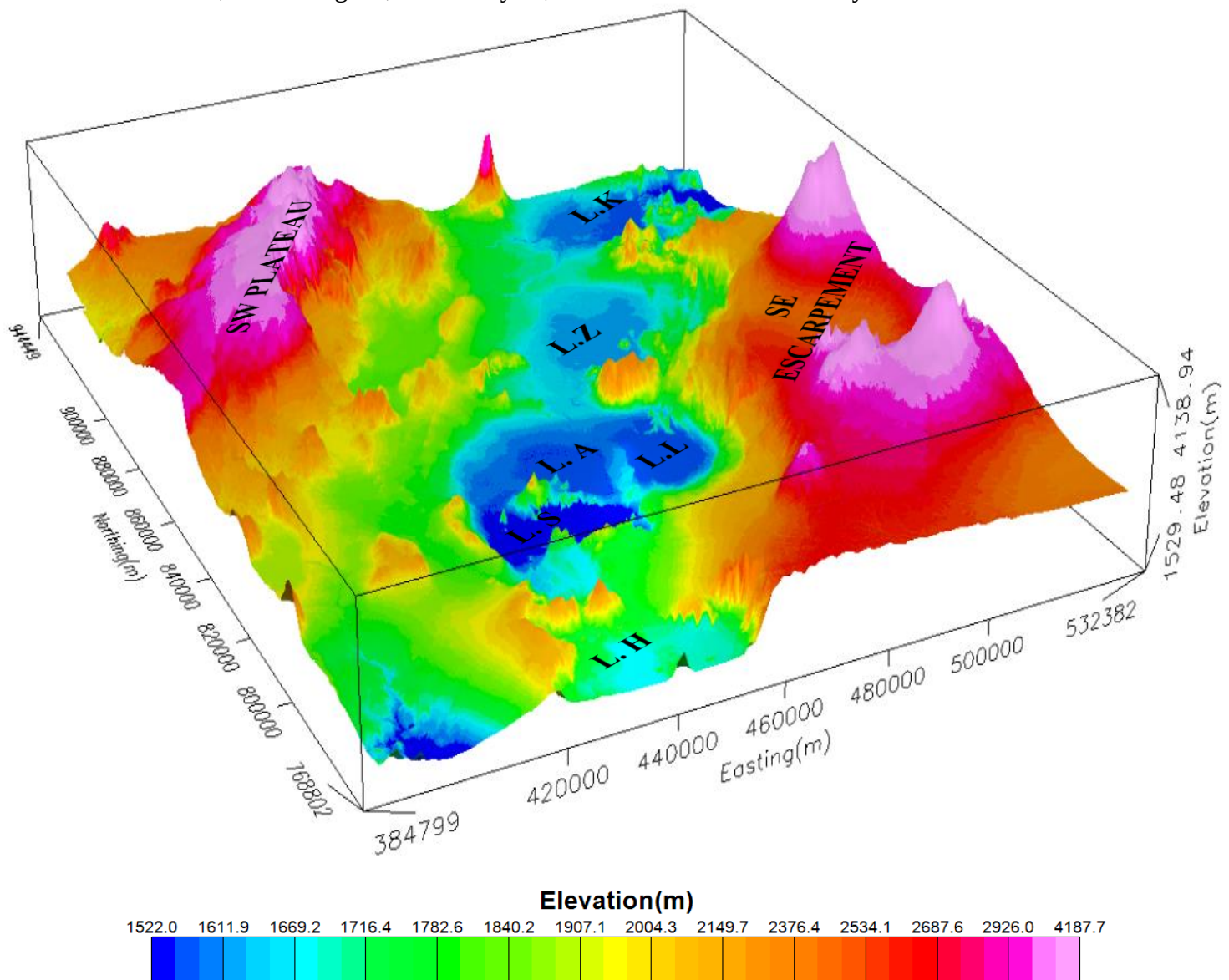


Figure 1.2. 3-D Terrain map of the CMER and adjacent plateaus. (Labels: L.H – Lake Hawassa, L.S – Lake Shalla, L.A – Lake Abiyata, L.L – Lake Langano, L.Z – Lake Ziway, L.K – Lake Koka).

1.3 Statement of the Problem

A number of detailed geophysical, geological, and geochemical investigations have been carried out in the Ethiopian Rift system in general and the MER (NMER, CMER, and SMER) in particular to study the crustal structure, nature, and evolution of the rift system.

However, studies involving 3-D inversions of gravity and magnetic data conducted in the central Main Ethiopian Rift for the study of crustal structure, nature, and evolution of the rift are not adequate (or totally missing).

In an attempt to contribute a solution to these existing problems, the present MSc thesis research work utilized 3-D Inverse modelling algorithms to determine the crustal structure, nature and thickness of the CMER and come up with 3-D density and magnetic susceptibility models of the CMER.

The 3-D gravity and magnetic inversion investigation performed for the first time in this MSc thesis research work is thought to add additional knowledge about the crustal structure and the general rift architecture of the CMER. Moreover, the results obtained here could be used as a reference for future investigations of the remaining segments of the MER (NMER, SMER).

1.4 Research Questions

The research questions that necessitated identifying the proposed research problem are:

1. What do the 3-D gravity and magnetic inversion results computed for the CMER and the adjacent plateaus look like in indicating the crustal structure and the thickness of the CMER?
2. What do the 3-D distribution of the subsurface density and magnetic susceptibility models of the CMER look like in identifying the subsurface anomaly sources?

In order to answer the above-formulated research questions, the general and specific objectives of this MSc thesis research work are designed as follows.

1.5 Objectives

1.5.1 General Objective

The general objective of this MSc thesis research work is to carry out 3-D inversion of airborne gravity and ground magnetic data and recover 3-D density and susceptibility models of shallow and deep geologic sources (any possible magnitude and geometric complexity) in the CMER.

1.5.2 Specific Objectives

The specific objectives of this MSc research are:

- ❖ Implement 3-D inversion of Gravity and Magnetic Data sets and produce 3-D density and magnetic susceptibility model for the CMER.
- ❖ To determine the depth to the Moho of the CMER and adjacent plateaus using gravity data
- ❖ To compile complete Bouguer gravity anomaly map, isostatic gravity anomaly map, depth to the Moho map and total magnetic anomaly map including their different derivative maps for the CMER.
- ❖ Identify the prominent geologic and tectonic features of the CMER and the adjacent plateaus by correlating the inversion results with the previous works in the study area.

1.6 Expected Outcome of the Research Work

The possible outcomes of the research work can be summarized as follows;

- ❖ 3-D density models of the CMER
- ❖ 3-D susceptibility models of the CMER
- ❖ Moho depth map of the CMER
- ❖ complete Bouguer gravity anomaly, isostatic regional anomaly, isostatic residual anomaly and total magnetic anomaly map including their different derivative maps for the CMER
- ❖ crustal structure and thickness of the central Main Ethiopian Rift and adjacent plateaus
- ❖ Identify areas associated with positive and negative Bouguer anomaly and its relationship with geologic and tectonic features.

1.7 Significance/Contribution of the Study

This MSc thesis research work, which could be considered as the first of its kind with the methodology it has applied in the CMER, is anticipated to contribute to the scientific community for the better understanding of the crustal structure, Moho depth and thickness of the CMER in particular and the MER in general.

The major contributions of this MSc thesis work could be summarized as,

- ❖ Provide 3-D density and susceptibility models derived from the 3-D inversions of gravity and magnetic data in the CMER, which is thought to be significant for any future Geoscientific work.
- ❖ Provide 3-D density and susceptibility models of the CMER, which can be used for the exploration (implications) of natural resource and geohazard mitigation works.
- ❖ Provide a more reliable and recent gravity and magnetic maps compiled using previously acquired and recent data sets.
- ❖ Provide a recent depth to the Moho depth, thickness and crustal structure of the CMER and adjacent plateaus.

1.8 Previous Geophysical Reviews

1.8.1 Geophysical Reviews

The Ethiopian Rift System (ERS) has long been recognized as one of the best places to investigate continental rifting, and, in particular, rift processes leading to seafloor spreading (MAKRIS & GINZBURG 1987; EBINGER & CASEY 2001; BASTOW & KEIR 2011).

Several geophysical, geological, and geochemical investigations have been carried out in the Ethiopian Rift system in broad and in the MER (NMER, CMER, SMER), in particular to determine the crustal structure, nature, and evolution of the rift system.

Attempts have been made to review previous geophysical investigations conducted in the Ethiopian rift system that are assumed relevant to this MSc work. Though, 3-D gravity and magnetic inversion investigations conducted in the ERS generally is few in number and below is the summary of the considered works.

As to the knowledge of the author and the attempts he have made in reviewing previous works , the only available 3-D gravity inversion work that provided complex 3-D crustal and upper mantle density model of the Northern Ethiopian plateau, Afar rift and the Northern main Ethiopian rift is the work of MAMMO, (2013). The constrained 3-D gravity inversion of (MAMMO, 2013) was employed to study the crustal structure of the flood basalt province of Ethiopia. Yet, there are also few of 3-D gravity inversion investigations performed in the Ethiopian rift system though they lack providing 3-D subsurface density models in their result.

On the other hand, 3-D magnetic inversion work implemented in the Ethiopian rift system is found to be totally absent. Below is a short summary of gravity studies and interpretations that have been have performed in the Ethiopian rift system.

I). The work of MAMMO, (2013) is the most prominent 3-D constrained gravity inversion investigation conducted in the flood basalt province of Ethiopia, Afar rift and the NMER for crustal structural study. The 3-D gravity inversion result of this study had revealed the lateral and vertical extents of the underplate underlying the NW plateau (Fig 1.3) and provided comprehensive 3-D picture of the crust and upper mantle structures, which is useful in understanding the complex structural trends of the geodynamic process in the study area (MAMMO , 2013). In addition, the Author (MAMMO) has implemented 3-D inversion codes for the analysis of the gravity field in the Omo basin in SW Ethiopia for the implications of Hydrocarbon potential (MAMMO, 2009) and for Geophysical modellings of Copper Mineral exploration in the Katta district, western Ethiopia (MAMMO, 2013).

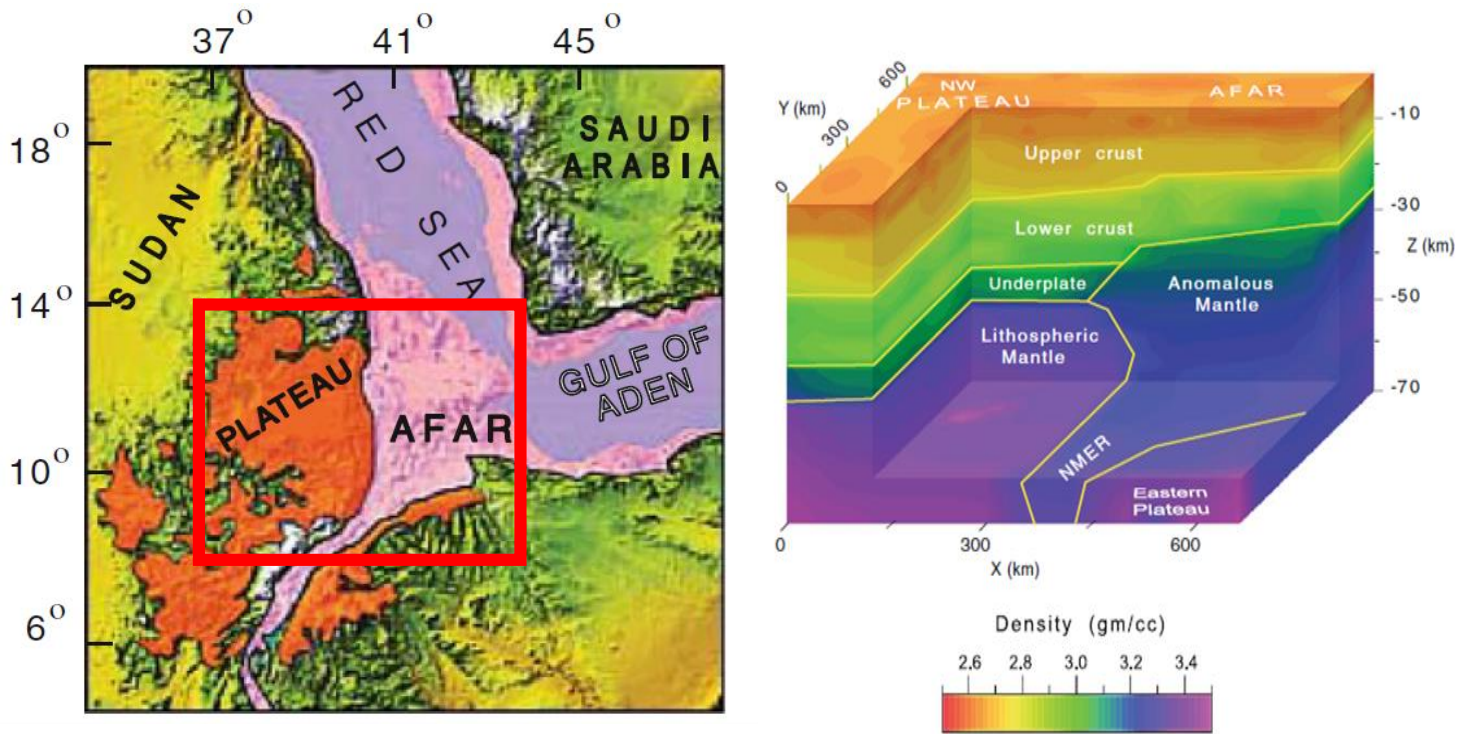


Figure 1.3 a) Location map of the study area (red rectangle) b) 3-D density model recovered from 3-D inversion of gravity data (MAMMO, 2013).

II). 3-D gravity inversion conducted on the crust and upper mantle structure within the East Africa rift system including the Afar depression implemented by ALEMAYO, (2017) showed the existence of thinnest crust (~20 km) beneath northern Afar, thickest crust (~40 km) beneath Ethiopian plateaus and a ~30-34 km thick crust which thins out to 20 km towards the Indian Ocean beneath the Somalia plate.

III). 3-D gravity analysis on the crustal structure of the Main Ethiopian Rift performed by MAHTSENTE et al., (1999) revealed the existence of hot upper mantle material, and hence, intrusion beneath the rift floor. The intrusion, as deduced from this gravity analysis (MAHTSENTE et al., 1999) was possibly of mantle or crust–mantle origin, and mostly follows the main Quaternary rhyolitic centers of the rift floor along the Wonji Fault Belt and Silti Debrezeit Fault Zone. Moreover, MAHTSENTE et al (1999) had only provided 2-D cross sections and advised that interpretation would be more plausible if one considers a 3-D model rather than a section along the model of his inversion results.

IV). ALEMU, (1983) has conducted crustal modelling in the Ethiopian rift system as well as investigated the gravity field and crustal structure of the Main Ethiopian rift using gravity data (ALEMU, 1992). This gravity investigation correlated the positive gravity anomaly trends occurring along the MER with the prevailing structural features. Accordingly, the rift floor and belts of synclinal structures have been correlated with the axes of the positive gravity anomalies while the negative gravity anomalies have been correlated to the uplifted regions and belts of anticlinal structures.

V) MAGURIE et al., (2006) have investigated the crustal structure of the northern Main Ethiopian Rift from the Ethiopia Afar Geoscientific Lithospheric Experiment (EAGLE) controlled-source survey.

The EAGLE controlled-source project had provided an intriguing new image of crustal and uppermost mantle structure beneath a 400 x 400 km region centered on the Boset magmatic segment of the transitional northern Main Ethiopian Rift (Fig 1.4). The EAGLE project performed travel-time tomography and forward modelling and inversion with the RAYINVR code of ZELT & SMITH (1992) and recovered final 2-D ray-trace P-wave velocity models for (a) Line 1 and (b) Line 2 (Fig 1.5).

The layer thicknesses and velocities derived from the EAGLE project confirmed that the crust beneath the MER is still predominantly continental in nature, with the principal modifications being: high-velocity intrusions segmented along the axis of the rift; alteration to a massively intruded, thinned crust in the southwest corner of Afar; and the presence of an anomalous ‘underplate’ layer beneath the plateau northwest of the Afar rift (MAGURIE et al 2006).

However, evidence provided by the EAGLE project on the crustal structure and thickness of the CMER and the SMER is not rich as compared to those of the NMER and the Afar rift.

The results of the controlled source tomography in the CMER show that the CMER have a thicker crust, lower seismic velocity and lack of well-defined surficial magmatic segments that indicate the CMER is less extended and less magmatically modified than the NMER.

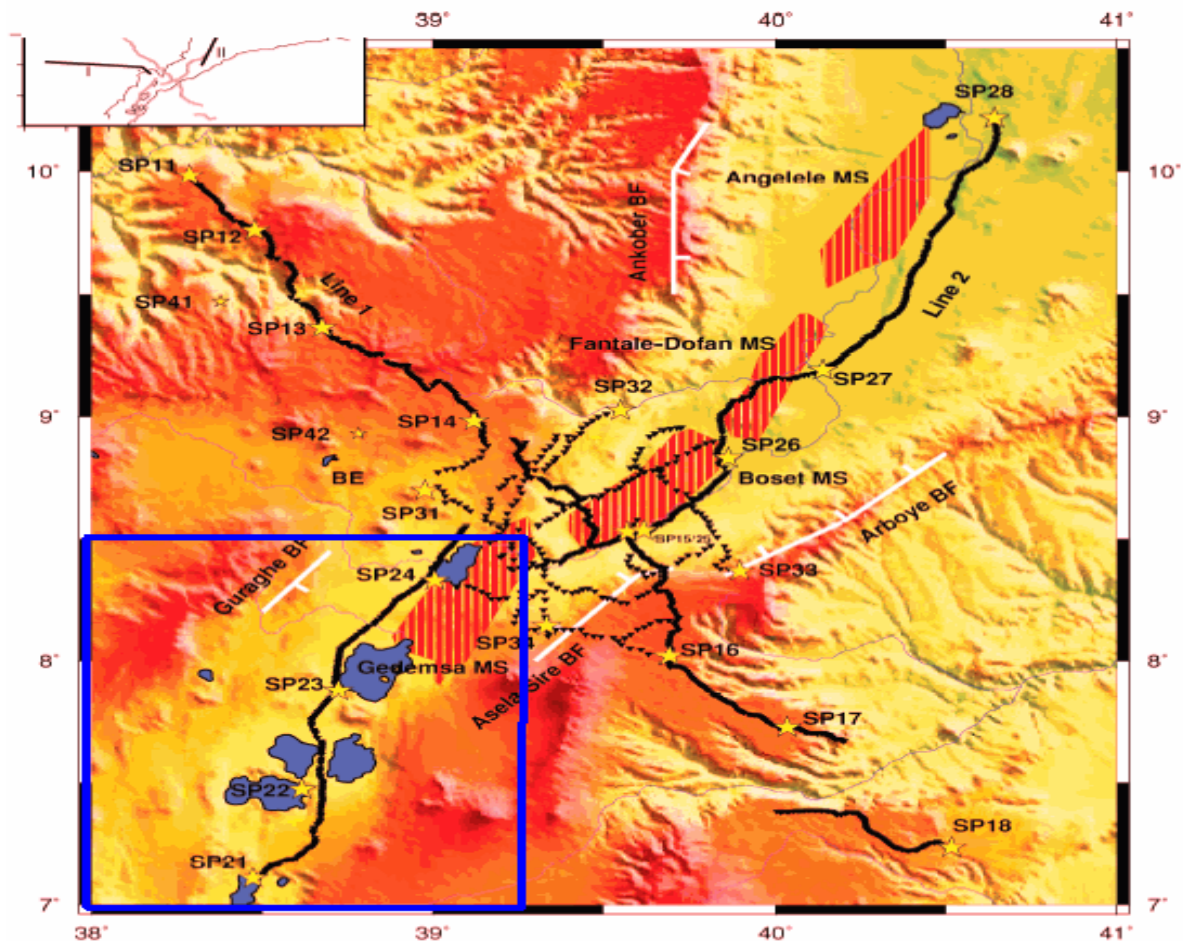


Figure 1.4 Location map of the EAGLE controlled-source project. Yellow stars, shot points (e.g. SP11); black small inverted triangles, controlled source recorders; red hatched areas, magmatic segments (e.g. Boset MS) (modified after WOLFENDEN et al. 2004); brown lines, border faults (e.g. Arboye BF). Blue polygon; study area (CMER).

According to the EAGLE controlled source project results, the lower seismicity in the CMER is an indication for this sector is currently less active. This abrupt change in rift structure, rift character, volcanic activity and seismicity along the axis of the rift represents a segmentation that continues into the mantle lithosphere. The thicker crust in the CMER reflects the lower amount of extension, and the lower magnitude upper-crustal velocities in the CMER reflect the lower degree of magmatic addition to the crust.

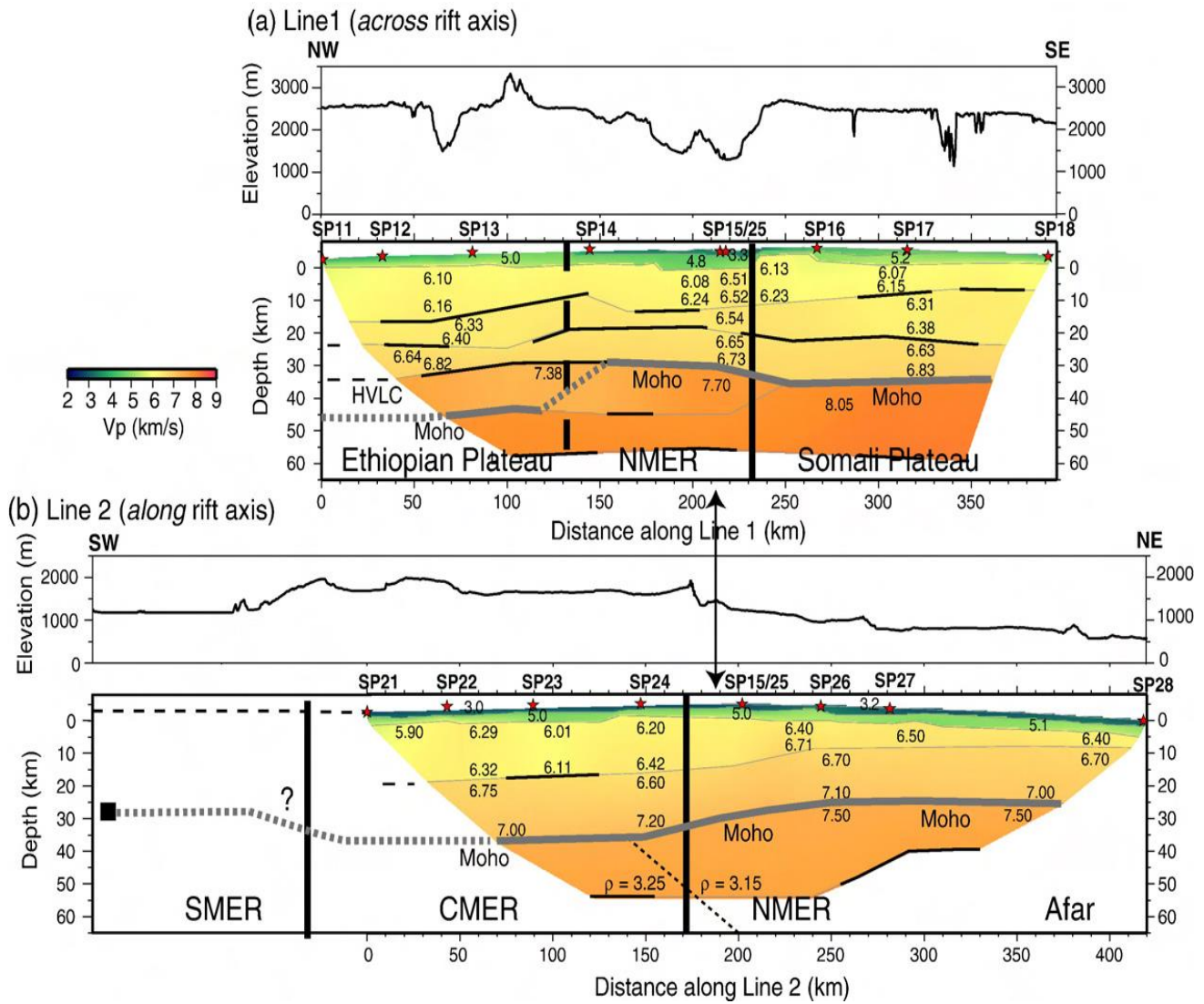


Figure 1.5. Ray trace P-wave velocity model result along L1 & L2 (Fig 1.4).

According to MAGURIE et al., (2006) the high- velocity, V_p (red) bodies in Figure 1.6(a) are interpreted as magmatic intrusions and the low velocity, $-V_p$ (red) bodies in Figure 1.6 (b) as reflecting partial melt and higher upper mantle temperature.

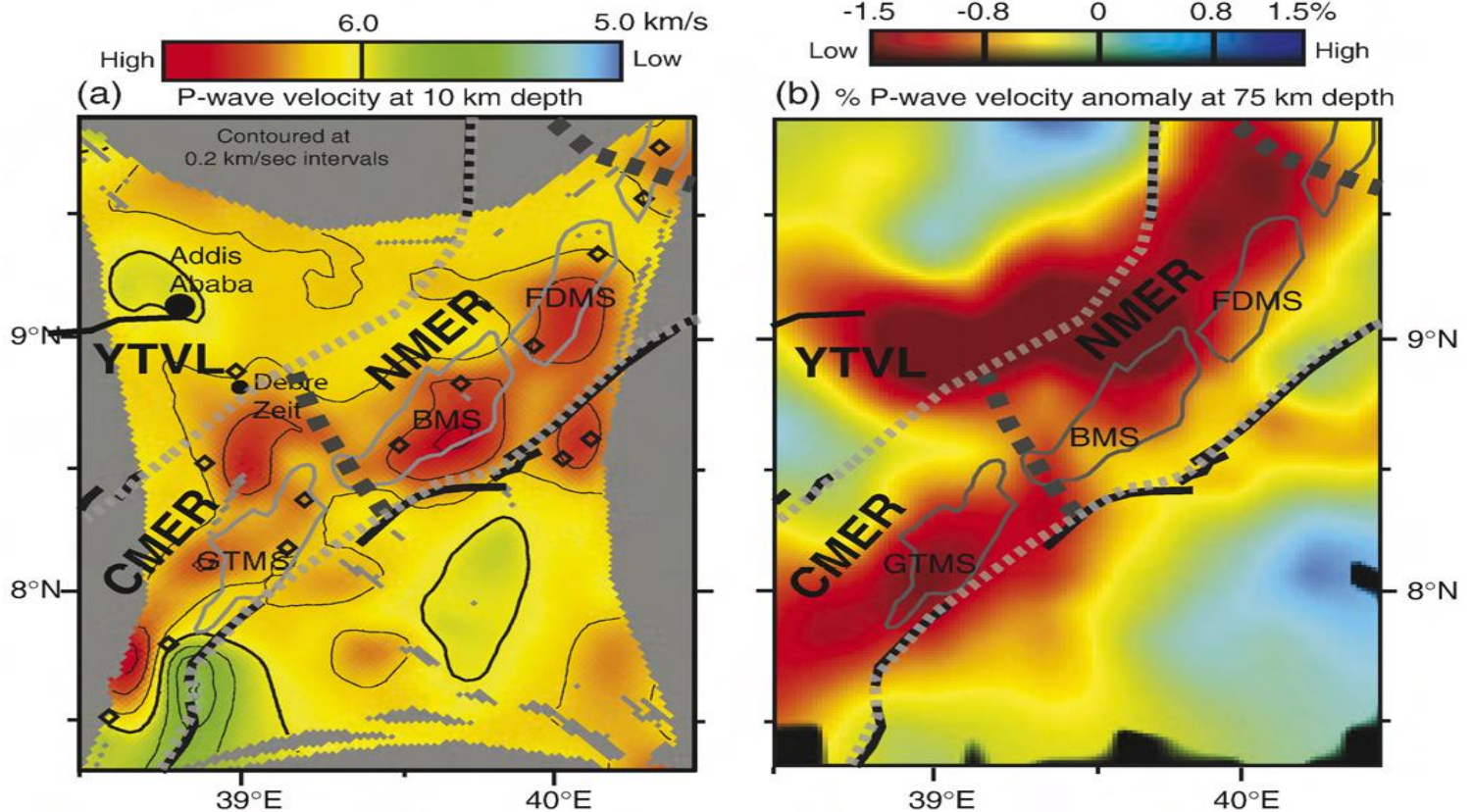


Figure 1.6 a. Depth slices from (a) controlled-source tomography at 10 km below the rift valley floor (modified from KERANEN et al., 2004). 1.6 b) passive source upper mantle P-wave tomography at 75 km depth (modified from BASTOW et al., 2005). (Labels: FDMS: Fentale–Dofen Magmatic Segment, BMS: Boset Magmatic Segment, GTMS: Gedeemssa–Tullu Moye Magmatic Segment, Border faults are shown in black lines, magmatic segments are outlined in thin gray lines; the locations of rift-valley volcanoes shown by black diamonds; the boundary between CMER and is shown in bold black dashed line).

All the previous geophysical studies mentioned have provided a noble knowledge about the crustal structure of the Ethiopian rift system. However, knowledge of the subsurface structure in the Central MER remains poor.

Hence, this MSc thesis research work has carried out 3-D gravity and magnetic inversion investigations in an attempt to fill the missing gaps in the application of 3-D gravity and magnetic inversion codes to study the crustal structure and thickness of the CMER.

1.9 Methodology and Instrumentation

1.9.1 Methodology

In order to achieve the general and specific objectives of this MSc thesis research work three distinct steps were implemented. These are pre –field work, field work and post-field work.

In the pre-field work stage identification and visualization of latest Landsat 8 satellite images, topographic maps, Digital Elevation Model (DEMs), geological maps, literature reviews on geophysical, geological and tectonic setting of the study area and the MER in general, and obtaining the available secondary data (gravity and magnetic) were performed.

In the field work stage primary magnetic data were collected using proton precession magnetometer (GSM-19T) along accessible roads. The Magnetic survey on the southwestern plateau part of the study area with poor accessibility was the most challenging part of the field survey. The major structural trends and volcanic complexes have been noticed in the field.

In the post-field work step after all the necessary primary, secondary, a priori information and data are compiled different types of data reduction, filtering, data extraction and data enhancement techniques were followed to the observed data sets to remove any an unwanted and non-geologic effects prior to the interpretation of the methods.

Airborne gravity data collected for geodetic purposes (2006-2007 GC) obtained from the Ethiopian Geospatial Institute have been processed analyzed and interpreted.

The methodology of this MSc thesis research work requires the application of geophysical inverse theory on gravity and magnetic methods, by performing 3-D Inversions of the two methods.

The 3-D gravity and magnetic inversion methodology presented in this MSc thesis work uses the algorithms of LI and OLDENBURG (1996, 1998). Each gravity or magnetic anomaly datum observed above the surface can be evaluated by calculating the projection of the anomalous gravity field or magnetic field onto a given direction. The anomaly source region will be divided into a set of rectangular prismatic cells by an orthogonal 3-D mesh and assumed a constant density and or magnetic susceptibility value within each cell. Finally, the observed sets of gravity and magnetic data can be related to the subsurface density and or magnetic susceptibility by using a linear inverse

theory. An iterative optimization process is used to recover 3-D subsurface density and magnetic susceptibility models of the source region by solving the formulated inverse problem.

1.9.2 Instrumentation

This thesis is a final output of an integrative use of different instruments, materials, software packages and inversion codes. Garmin 72S GPS and topo maps were used to determine the location (latitude, longitude and elevation) of magnetic data as well as to analyze the topographical variation of the area. Magnetic survey has been collected using Proton precession magnetometer (GSM-19T). Geosoft Oasis Montaj (version 6.4.2) desktop application software has been used to manipulate the raw data and produce a scientific information. Moreover, other complementary softwares including surfer (version 16.0), Map info (version 10.0), Microsoft Office (2016), Global mapper (version 12.0) and ArcGIS version 10.8 has been used.

The University of British Columbia - Geophysical Inversion Facility (UBC-GIF) 3-D inversion codes were used for the 3-D gravity and magnetic inversion in this MSc thesis research work.

1.10 Thesis Structure

This MSc thesis work is organized in seven chapters of distinct contents. The first chapter gives an overview on the general framework of the research including description of the study area, rationale, objectives, and methodological approach of the study, geological and geophysical reviews and the materials utilized in the research work. The second chapter is about the over view of tectonic and geologic setting of the East African rift system, Ethiopian rift system, Main Ethiopian rift and Central main Ethiopian rift system . The Third chapter is about the theoretical foundations of the methods of the study including the fundamentals of inversion in applied geophysics. The fourth chapter outlines the data acquisition and data processing section of the thesis. The data presentation and interpretation part is discussed in chapter five. Chapter six is all about the background, inversion process, inversion results and interpretations of the 3-D gravity and magnetic data inversions. The last seventh chapter of the MSc thesis outlines the conclusions and recommendations driven from the results of the research work.

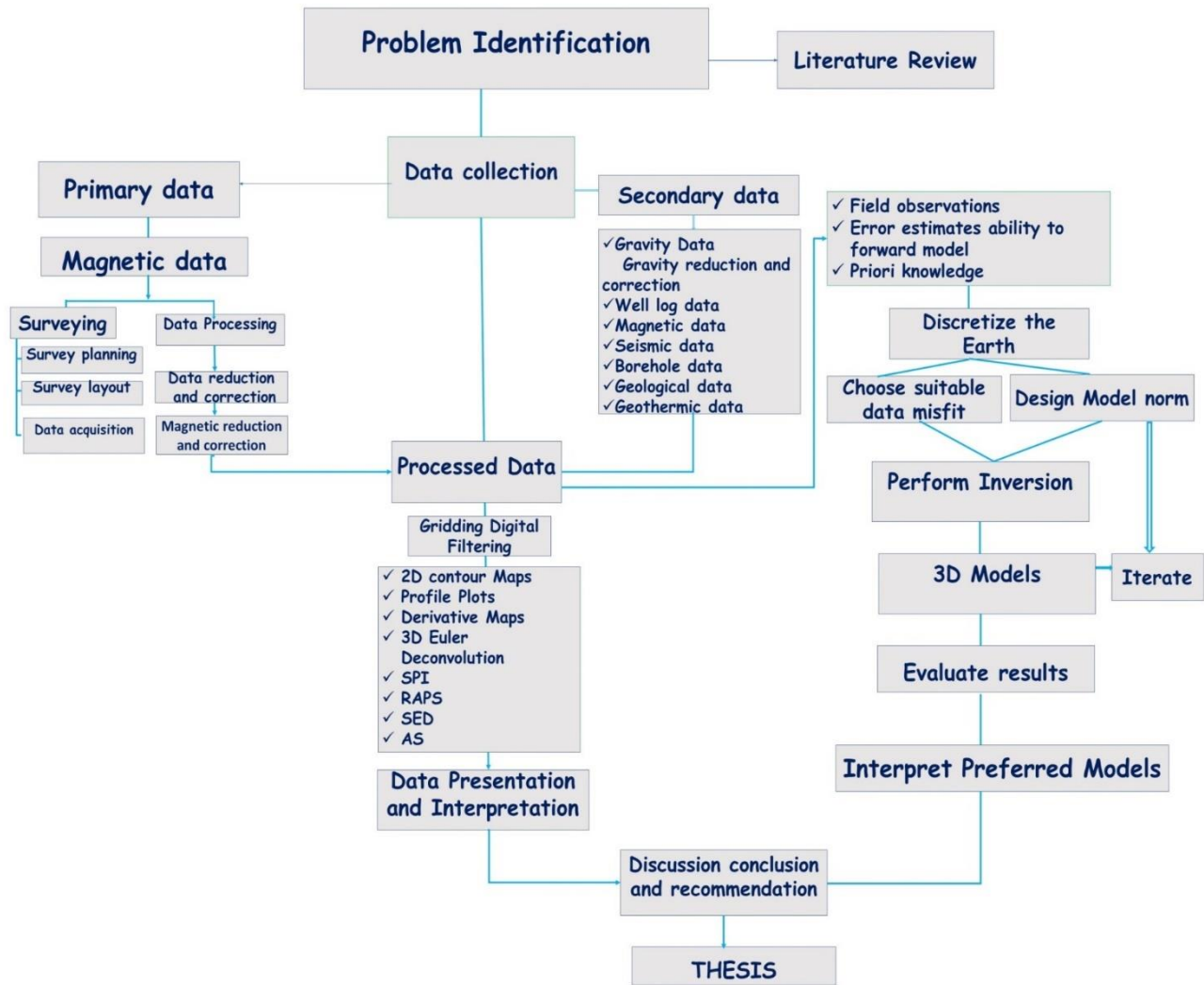


Figure 1.7 general work procedure and inversion flowchart.

CHAPTER TWO

2.0 OVERVIEW OF TECTONIC AND GEOLOGIC SETTING

2.1 Tectonic Setting of East African Rift System

The East African rift system (EARS), an intra-continental ridge system, is segmented in to two branches known as the Eastern branch and the Western branch, comprises a series of rift zones (CORTI, 2009, ROSENDAHL et al., 1986) and extends from the Afar triple junction at the Red Sea - Gulf of Aden intersection to the Zambezi River in Mozambique. A third, southeastern branch is in the Mozambique Channel. The Eastern branch consists of the Ethiopian Rift system and the Kenyan Rift system (Fig. 2.1). The Eastern branch runs over a distance of 2200 km, from the Afar triangle in the north, through the main Ethiopian rift, the Omo- Turkana lows, the Kenyan (Gregory) rifts, and ends in the basins of the North-Tanzanian divergence in the south (Fig 2.1).

The western branch runs over a distance of 2100 km from Lake Albert (Mobutu) in the north, to Lake Malawi (Nyasa) in the south. It comprises several segments: the northern segment includes Lake Albert (Mobutu), Lake Edward (Idi Amin) and Lake Kivu basins, turning progressively in trend from NNE to N-S; the central segment trends NW-SE and includes the basins of lakes Tanganyika and Rukwa; the southern segment mainly corresponds to Lake Malawi (Nyasa) and small basins more to the south. The southeastern branch comprises North striking undersea basins located west of the Davie ridge.

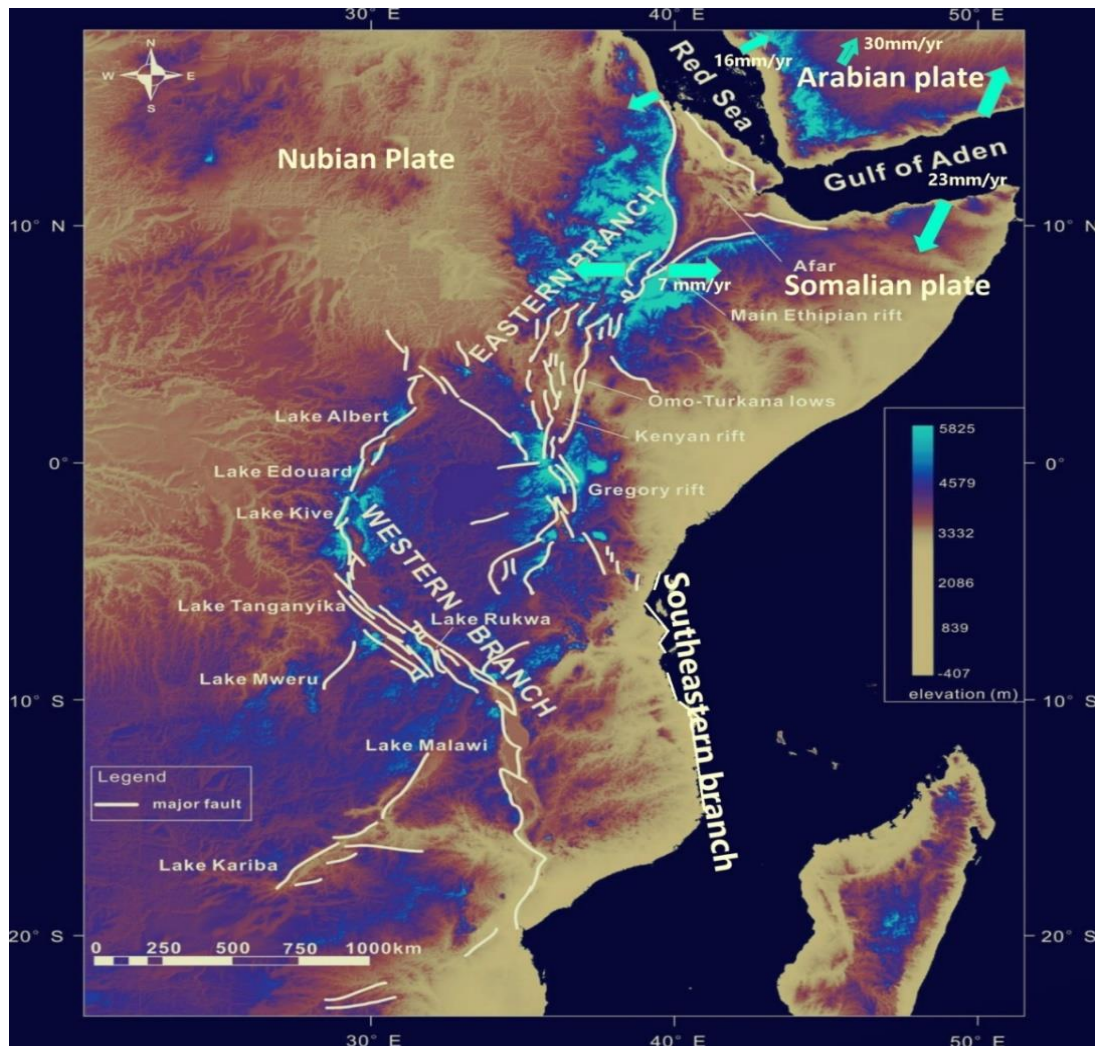


Figure 2.1 Tectonic setting of the East African Rift system (modified after JEAN CHOROWICZ)

2.2 Tectonic Setting of Ethiopian Rift System

The Ethiopian Rift System (ERS), part of the EARS, is divided into at least three major rift basins that consist of the Afar Depression, the Main Ethiopian Rift, and the broadly rifted zone of southwestern Ethiopia (Fig 2.2). Distinct tectonic, volcanic, and geomorphic features characterize these rift basins.

The Afar Rift at the northern end of the ERS is a proto-oceanic rift that is older and currently at more advanced tectonic and volcanic stages than the other two rift segments. The funnel-shaped geomorphic feature is a product of the tectonic interactions of the boundary faults of the two oceanic rifts of the NW-SE trending Red Sea and the E-W-trending Gulf of Aden with the NE-SW-trending continental Main Ethiopian Rift (WOLDEGABRIEL, 2009).

The proto-oceanic Afar Depression transits to the continental Main Ethiopian Rift (MER), which is characterized by a narrow but long basin that is more than 600 km long and about 60 to 80 km wide. The MER floor is marked by a right-laterally displaced active axial rift zone that is about 5 to 10 km wide. At the southern end, the MER is bifurcated into two rift basins before it disappears into uplifted crystalline basement rocks.

The broadly rift zone of southern and southwestern Ethiopia is the northern continuation of the Kenya Rift (Fig 2. 2). The Chew Bahir rift, which is located along the eastern part of the broadly rift zone, is characterized by well-defined rift margins and a narrow rift system (10-20 km) that is more than 300 km long (Fig 2.2). To the west of the Chew Bahir rift, the Omo, Usno, Kibish half grabens occupy the rest of the broadly rifted zone. Singular, well-defined rift margins and tilted rift floors characterize these rift basins.

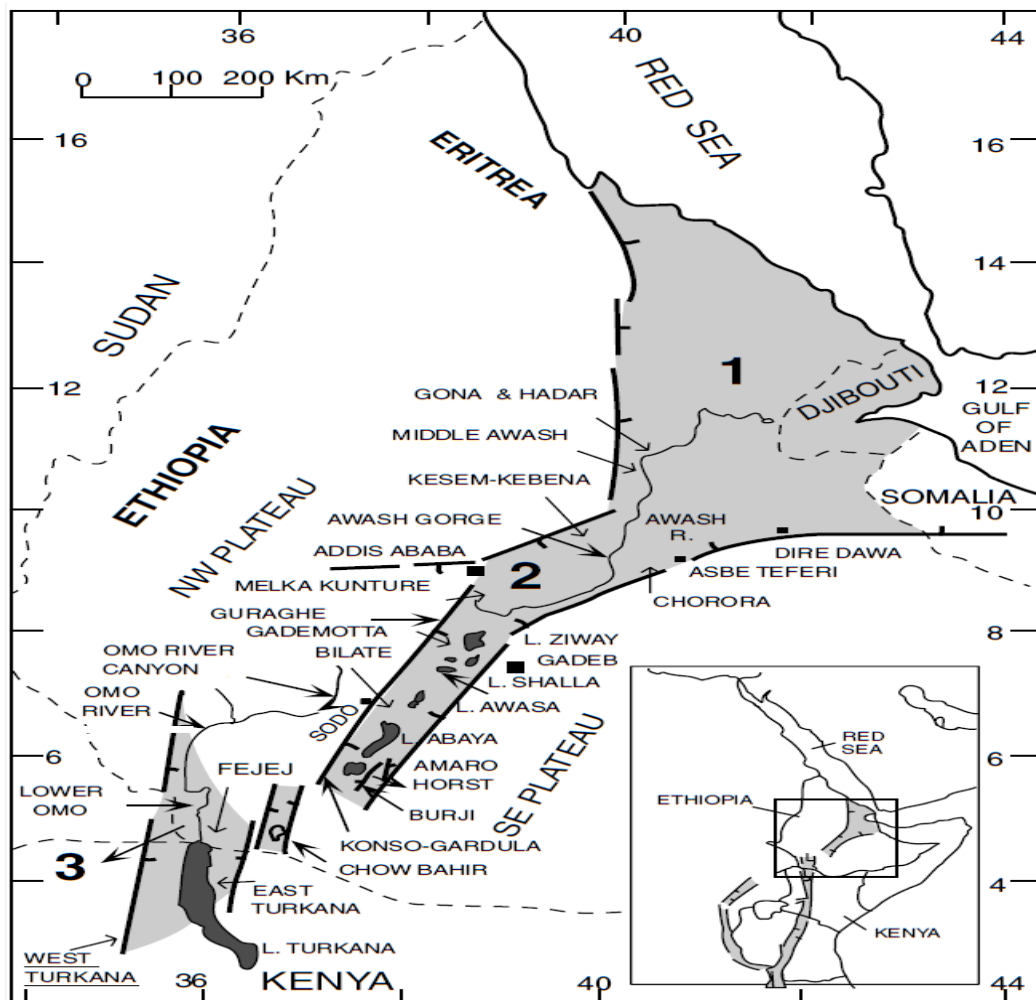


Figure 2.2 Ethiopian rift system (modified after WOLDEGABRIEL, 2000). (1 - Afar rift, 2 - Main Ethiopian rift. 3 - Omo rift).

2.3 Tectonic Setting of the Main Ethiopian Rift

The Main Ethiopian Rift (MER) is a roughly NE trending sector of the East African Rift system that includes a series of rift segments extending from the Afar Triple Junction at the Red Sea-Gulf of Aden intersection to the Kenya Rift (Figs 2.1 and 2.2).

The MER (Fig 2.3) is characterized by active extensional tectonics accommodating 6 –7 mm/yr (Fig 2.1) relative movement between the African and Somalian plates (CHU and GORDON, 1999; FERNANDES et al., 2004).

The Main Ethiopian Rift (MER), is traditionally differentiated into three main sectors (northern MER, central MER and southern MER) differing in terms of rift trend, fault patterns and lithospheric characteristics (MOHR, 1983; WOLDEGABRIEL et al., 1990; HAYWARD and EBINGER, 1996; BONINI et al., 2005). According to HAYWARD and EBINGER (1996) the along-axis variations of rift-related deformation (e.g., fault, basin length and rift flank uplift reflecting differences in the effective elastic thickness of the lithosphere) in these different MER sectors is interpreted as reflecting a variation in rift evolution from mature rifting in the northern MER to early continental rifting in the southern MER.

The MER (Fig 2.3) shows two well-developed fault systems, (1) the border faults and (2) a set of faults affecting the rift floor, usually referred to as Wonji Fault Belt (WFB) (MOHR, 1962; CHOROWICZ et al., 1994; BOCCALETTI et al., 1998; ACOCCELLA et al., 2003). The border faults are normally long, widely spaced and characterized by large vertical offset and are interpreted to have accommodated the tectonic deformation during the initial stages of rifting (HAYWARD and EBINGER, 1996; BOCCALETTI et al., 1998). On the basis of seismicity (KEIR et al., 2006) and local geological data (WOLFENDEN et al., 2004; CASEY et al., 2006) the border fault set is thought to have been deactivated and mostly eroded in the northern MER. The WFB is a tectono-volcanic system characterized by short, closely spaced, active faults that exhibit minor vertical throw, which developed essentially in the last two Ma (BOCCALETTI et al., 1999; EBINGER and CASEY, 2001).

The WFB structures form clearly defined right-stepping en echelon segments obliquely cutting the rift floor that are believed to accommodate the majority of current deformation, as indicated by GPS measurements (BILHAM et al., 1999; BENDICK et al., 2006) and analysis of the current seismicity (KEIR et al., 2006).

In the central and southern MER sectors boundary faults are well developed, and analysis of historical seismicity (GOUIN, 1979; KEIR et al., 2006) and scattered geological data (PIZZI et al., 2006) suggest that at least part of these faults may be active.

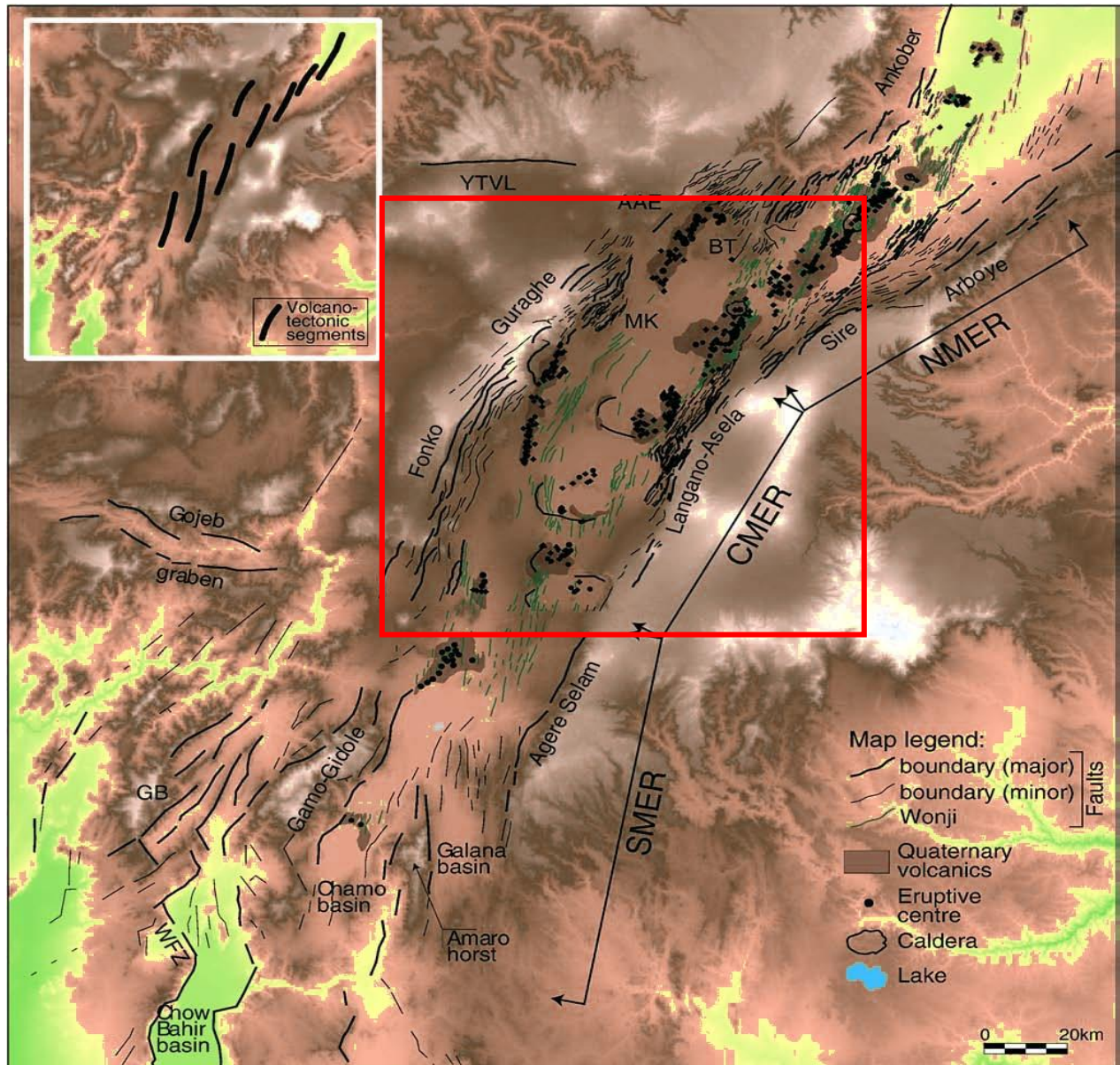


Figure 2.3 Tectonic sketch map of the Main Ethiopian Rift (modified from BOCCALETTI et al., 1998). Inset map shows the en-echelon, right-stepping arrangement of the volcano-tectonic segments of the Wonji Fault Belt. AAE: Addis Ababa Embayment; BT: Boru Toru structural high; MK: Midre Kebd structural high; YTVL: Yerer-Tullu Wellel volcano-tectonic lineament. Red polygon-study area.

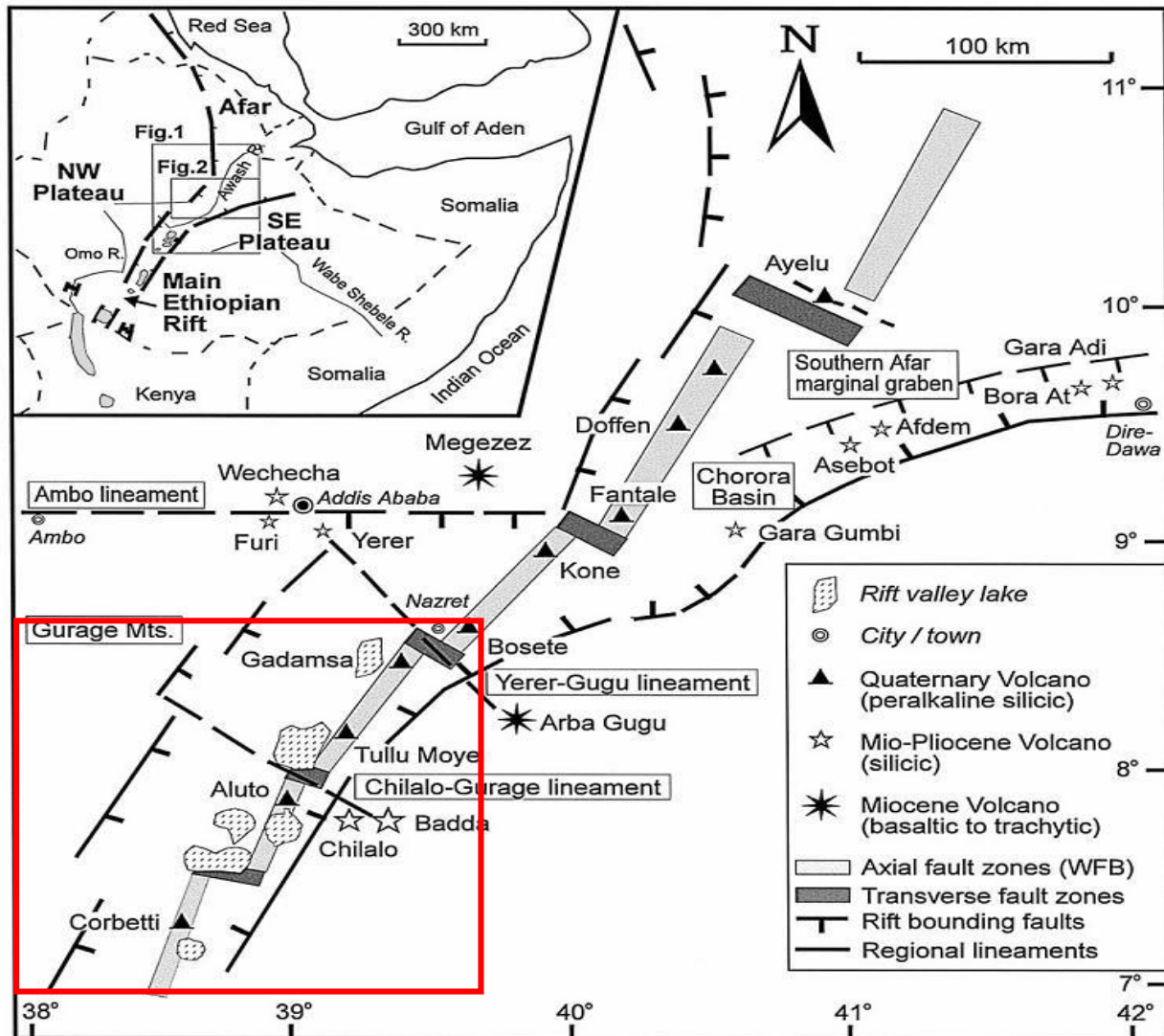


Figure 2.4 Structural sketch of the MER (WOLFENDEN 2004); red polygon -study area (CMER).

2.4 Geology of the Main Ethiopian Rift (MER)

The Main Ethiopian Rift (MER) is a roughly NE-oriented segment of the East African Rift system, which extends from the Afar to the Kenya rift. The MER is composed of northern, central and southern segments. It has commonly been suggested that rift propagation in the MER progressed northwards (WOLFENDEN et al., 2004; MACKENZIE et al., 2005; KEIR et al., 2006), although its evolution is not well understood. An alternative hypothesis implies southward propagation (BUCK, 2006; ROGERS, 2006; BONINI et al., 2005) suggested a heterogeneous time-space evolution, with initial extension in the Southern MER at 20–21 Ma, followed by extension in the Northern MER

at 11 Ma and finally formation of the Central MER at about 5–6 Ma. KERANEN and KLEMPERER (2008) have also proposed irregular rift propagation due to the presence of preexisting structures.

The MER is bounded by discontinuous boundary faults that give rise to major fault escarpments separating the rift depression from the SW and SE Plateaus. These faults are widely spaced and characterized by large vertical offsets (>1 km; BOCCALETTI et al., 1998). In the Central MER, they are mainly oriented 30° – 40° N. The rift floor is affected by dense NNE–SSW-trending fault swarms, with relatively small throws (<100 m) which comprise the Wonji Fault Belt (Fig. 2.5; CORTI, 2009). The Wonji Fault Belt started to develop at the beginning of the (1.8–1.6 Ma, MEYER et al., 1975), and probably reflects the present-day stress field (CHOROWICZ et al., 1994; BOCCALETTI et al., 1998; EBINGER and CASEY, 2001).

Tertiary to Quaternary volcanic are the merely rocks exposed in the MER, apart from minor fluvio-lacustrine sediments, mostly of Quaternary age, that were deposited on the rift floor (ABEBE et al 2010). One exception is an outcrop of Precambrian crystalline rocks covered by Mesozoic sediments at the base of the Guraghe escarpment, near the village of Kella (Figs 2.5) about 80 km south-west of Addis Ababa. Apart from Kella, basement rocks in the MER are exposed only in its extreme south and in the northern Afar.

Recent geological and geochronological investigations (ABEBE et al., 2005; BONINI et al., 2005 and references therein) distinguished four major volcanic episodes in the central MER;

32–29 Ma: The main part of the Ethiopian plateau volcanics consists of the Trap Series (HOFMANN, 1997; PIK et al., 1998; UKSTINS et al., 2002). They cover an area of more than 600,000 km² and locally reaches thicknesses exceeding 3000 m (MOHR and ZANETTIN, 1988). This episode has been attributed to melting caused by the activity of one or two mantle plumes impinging on the base of the continental lithosphere under the Afar or Afar–Northern Kenya (EBINGER and SLEEP, 1998; GEORGE et al., 1998). These volcanic successions are missing at the Kella block in the CMER.

11–8 Ma: After a long period of less voluminous and more local volcanic activity related to the development of shield volcanoes (CORTI, 2009), basaltic–trachytic lava flows and associated pyroclastic were formed. They constitute the largest part of the Guraghe escarpment and extend for several tens of kilometers west of the escarpment. Their average thickness at the Guraghe margin is 600–700 m. Rocks of this volcanic succession are also absent at the Kella block. For

both volcanic events, widespread time-correlative basaltic units occur on the western and eastern margins of the central MER, constraining the lateral extent of these events.

5–3 Ma: Several layers of pyroclastic rocks associated with trachytic and rhyolitic lava domes and flows together with some important central volcanoes were formed in this episode, and cover the MER shoulders and rift floor. Radiometric ages from both the eastern and the western margins of the MER range between 5.2 and 2.6 Ma (CORTI, 2009). These highly differentiated central and/or fissural effusive and explosive eruption products indicate that volcanic activity during this period was very intensive and vigorous.

<3 Ma: The products of this episode consist of uncompacted pumiceous fall and flow deposits, rhyolitic–trachytic lava flows forming central volcanic edifices, fissural basaltic lava flows with associated scoria and phreatomagmatic cones, and interbedded lacustrine deposits (DI PAOLA, 1976; BERHE, 1978; KAZMIN et al., 1980). These deposits are generally confined to the MER floor.

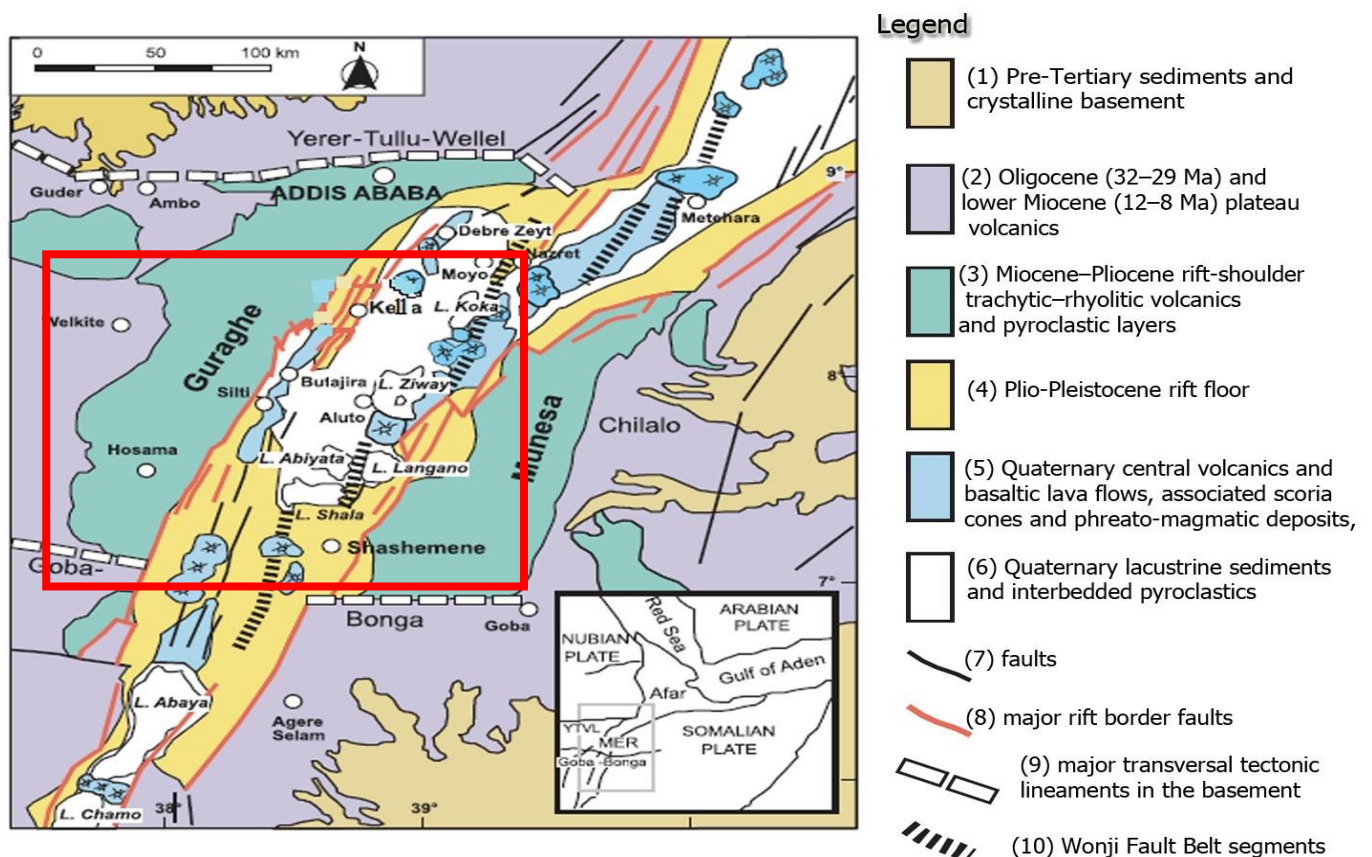


Figure 2.5 Simplified geological map of central Ethiopia (modified after JEPSEN and ATHEARN 1997). Red polygon-study area YTVL, Yerer-Tullu-Wellel volcano-tectonic lineament, MER, Main Ethiopian Rift, red polygon study area (CMER).

2.5 Tectonics, Geology and Stratigraphy of the Study Area (CMER)

2.5.1 Tectonic Setting of the Study Area (CMER)

The study area, which constitutes the CMER, is formed within the Oligocene Ethiopian flood basalt province and is thought to have initiated at between ~20 and ~10 Ma (WOLFENDEN et al., 2004; BONINI et al., 2005; CORTI, 2009). It is bounded by the Yerer-Tullu-Wellel volcano-tectonic lineament to the north, the Goba–Bonga lineament to the south, the Munesa border fault to the east and the Guraghe escarpments to the west (Fig 2.5). The extension was initially localized to several ~60-km long, NE-SW striking, high-angle ($>60^\circ$) border faults that bound the rift, such as the Boru-Toru and the Goba-Bonga structural lineament on the western side of the rift, and the Asela-Sire Border Fault on the eastern side of the rift (BONINI et al., 2005; CORTI et al., 2020). Since the Quaternary, the locus of tectonic and magmatic activity within the CMER is thought to have become focused to a ~20-km-wide zone of small offset faults, aligned cones, and active volcanic centers within the rift floor known as the Wonji Fault Belt (WFB), and also at a few rift marginal magmatic systems, such as the Silti-Debrezeit Fault Zone (SDFZ) towards the western side of the rift (WOLDEGABRIEL et al., 1990; ROONEY et al., 2014; CHIASERA et al., 2018).

Structural and stratigraphic relations of volcanic rocks along both escarpments of the CMER indicate a two-stage rift development (WOLDEGABRIEL et al., 1990). The early phase started during late Oligocene–early Miocene time, was characterized by a series of alternating, and opposed half grabens. The half-grabens evolved into asymmetrical rift during the late Miocene. In the second phase of Quaternary evolution, a shift in deformation along NNE and volcanism produced an echelon WFB segments (CORTI, 2009). This second phase of volcano tectonic evolution resulted an abandonment of boundary faults and localization of deformation and volcanism along the Wonji segments, obliquely affecting the rift floor (CORTI, 2009).

Pre-rift Tertiary volcanic succession covers the escarpment on both sides of the CMER. Basaltic, Salic lava flows and pyroclastic rocks occur in varying proportions and reach several hundred meters thick. The rock types are mainly basalt, trachyte, ignimbrite and rhyolite (Fig 2.5).

The main rifting phase in the CMER took place after 8 Ma (ABEBE et al 2010). As deformation in the NMER started at c. 11 Ma (WOLFENDEN et al., 2004), postponing the formation of the CMER to after 8 Ma means that the different MER segments are actually formed at different times

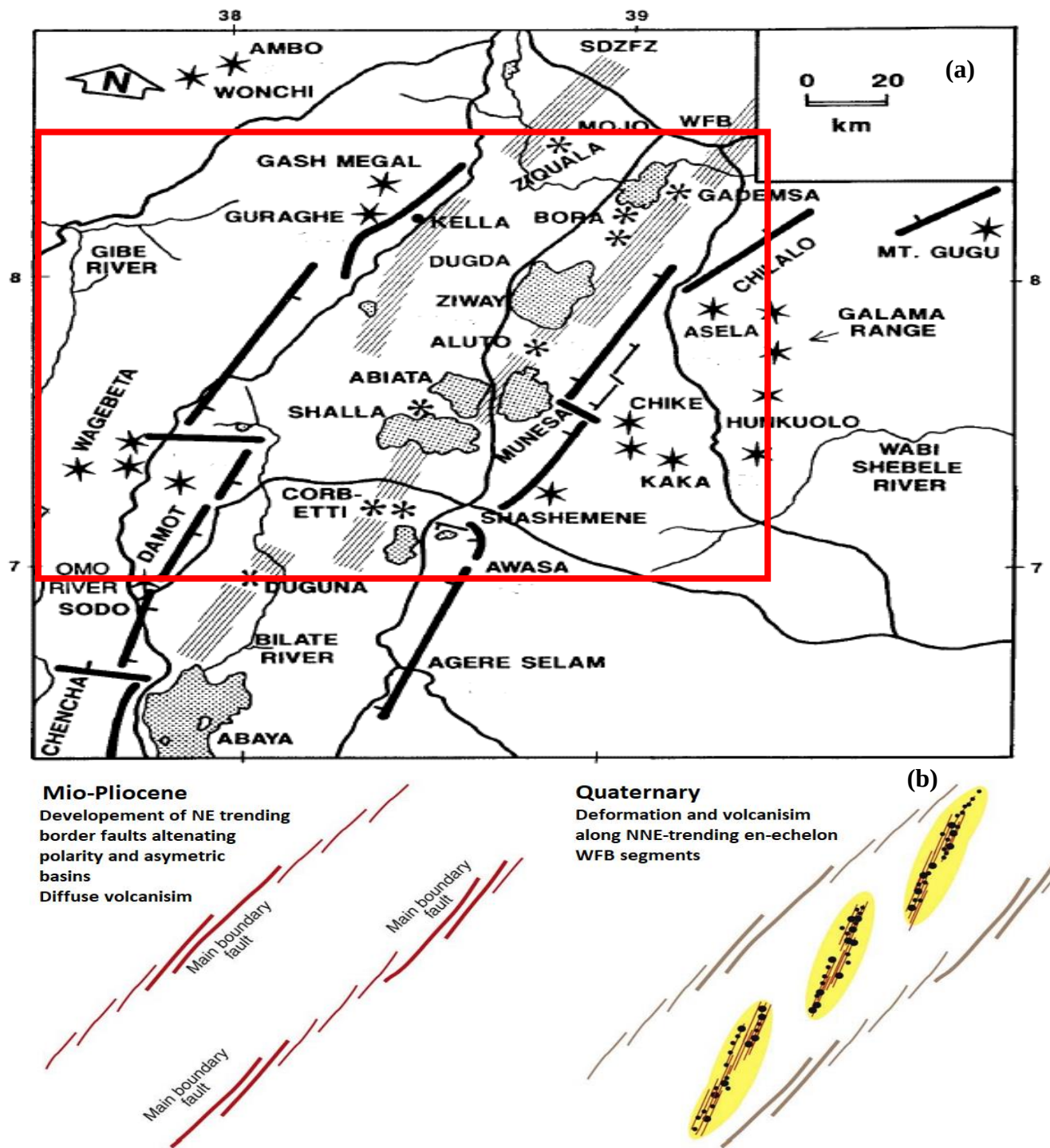


Fig 2.6 (a) Tectonic sketch map of the CMER. (Thick line segments- rift margins faults, pointed stars- rift shoulder central volcanoes, Asterisks-Quaternary per alkaline rhyolite centers). (Modified after WOLDEGABRIEL 1990). (b) Two-phase volcano-tectonic evolution of the CMER (after CORTI, 2008). Red-polygon-study area.

2.5.2 Geology of the Study Area (CMER)

The author of this MSc thesis has compiled a detailed geological map of the CMER (Fig 2.7) by harmonizing the existing geological maps of four adjacent map sheets (Akaki map sheet: by ALEMU et al: 2009 GC, Asela map sheet: by HARO et al 2016 GC, Hosaina map sheet: by ZENEBE et al 2016 GC, and Nazareth map sheet: by V. KAZMIN, M. BERHANE et al 1976 GC, all references there in). All the geological maps of the four adjacent mapsheets were obtained from the Geological Survey of Ethiopia (GSE).

The pre-rift Tertiary volcanic succession shields the escarpment on both sides of the rift. Basalt, salic lava flows and pyroclastic rocks occur in varying proportions and reach several hundred meters thick.

With the formation of the Wonji Fault Belt, tectonic movements and volcanic activity created step-like structures and associated volcanic activity, represented by ignimbrites, basalts and unwedded pyroclastics. The fault zone is linked by central volcanoes disposed along the axial zone of the Wonji Fault Belt. The main yields were rhyolite, trachyte lava flows, pumice, unwedded tuffs, obsidians and pitchstones. The products of the Wonji fault are referred as Wonji Group (KAZMIN et al., 1978 and others). Another type of volcanic activity in Wonji Fault Belt was the eruption from fissures of Pleistocene to Recent basalt lava flows. The basalts are controlled by extensional fractures and commonly characterized by fresh aa surface. Chains of scorraceous cones follow the lines of fractures. These basalts are mostly found in the rift floor; south of lake Ziway and lake Shalla. Recent flows in many cases follow pre-existing topographic low relief areas. Although the development of the rift was controlled by volcanic activity, sedimentation also occurred. Wonji Group rocks are intimately associated with lacustrine sediments related to the ancestral lake in the rift floor in the Pleistocene -Holocene times. Most parts of the floor of the CMER especially those surrounding Lake Langano, Ziway and Hawassa are covered by Pleistocene-Holocene age lacustrine sediments. They were deposited in a huge lake whose level thousands of years before was much higher than at the present. The sediments are mainly characterized by sand and silt.

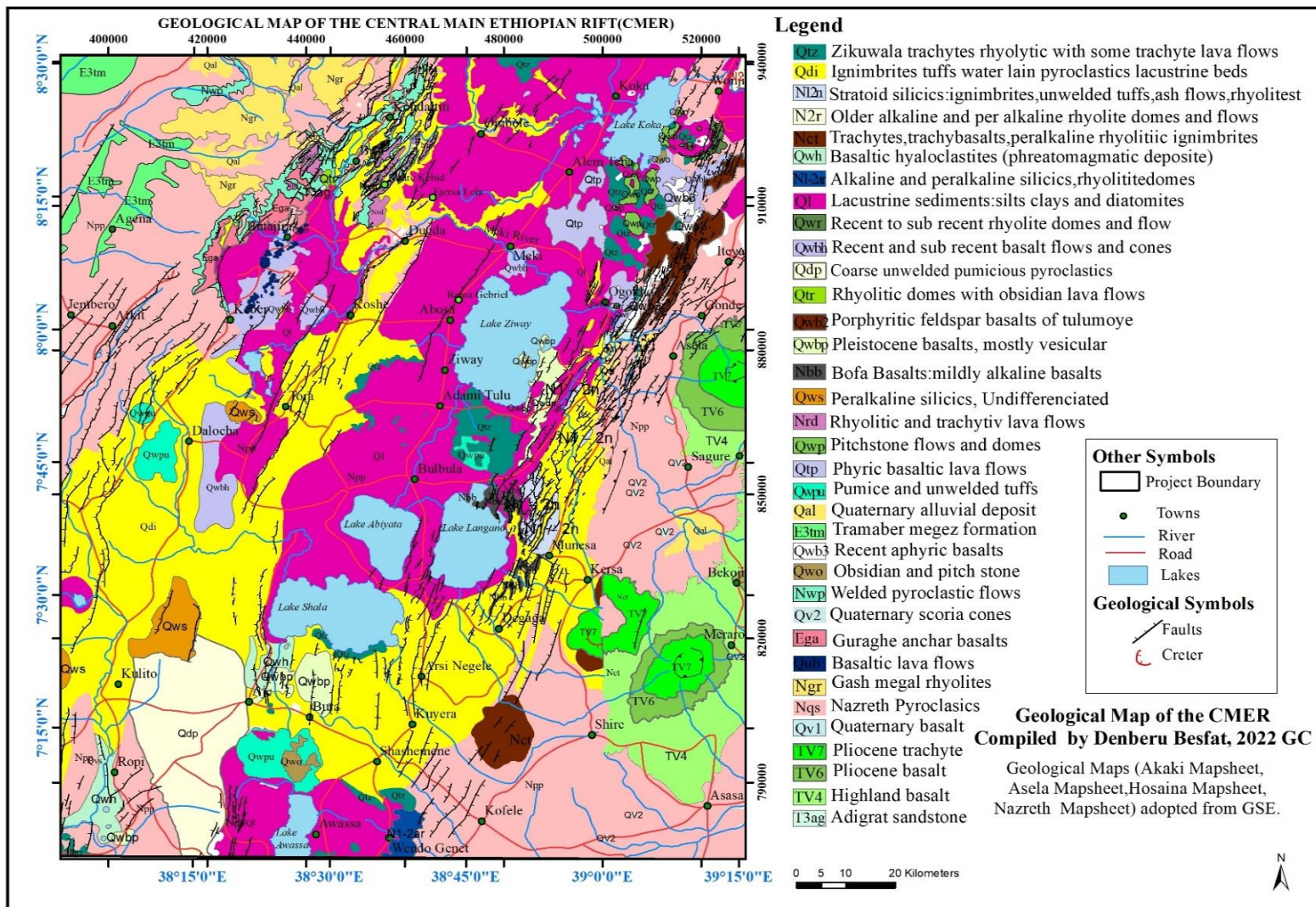


Figure 2.7 Geological and structural map of the CMER

2.5.3 Lithology

Most of the lithological descriptions of the geological map (Fig 2.7) are adopted from Akaki map sheet: by ALEMU et al: 2009 GC, Asela map sheet: by HARO et al 2016 GC, Hosaina map sheet: by ZENEBE et al 2016 GC, and Nazareth map sheet: by V. KAZMIN, M. BERHANE et al 1976 GC) obtained Geological Survey of Ethiopia (GSE). The main lithological units observed on the compiled map are discussed as follows.

2.5.3.1 Pre rift volcanic units

2.5.3.1.1 Adigrat Sandstone (T3ag)

This unit is exposed the base of Kella horst, near Kella town and elsewhere in Serkole river, 5 kilometers west of Ambo town (ALEMU et al: 2009).

2.5.3.1.2 Tarmaber-Megezez Formation (E3tm)

This lithological unit is found in the northwestern portion of the study area (Fig 2.7). It is dark grey in fresh samples, but reddish brown in weathered specimens. It is fine-grained volcanic rock composed of plagioclase, pyroxene, and opaque minerals (ALEMU et al: 2009).

2.5.3.1.3 Guraghe-Anchar Basalts (Ega)

Exposure of this unit is found in the northwestern part of the study area. It is dark color fine-grained rock mainly composed of plagioclase, pyroxene and opaque. Generally, it is fresh, compact fine-grained rock, but in places, it shows vesicular cavity filled with or without secondary minerals of zeolite and /or calcite. It shows aphanitic texture under the microscope (ALEMU et al: 2009).

2.5.3.2 Late Tertiary Volcanics

2.5.3.2.1 Nazareth Group: (N1-2n), and (N1-2ar)

Lithologies of stratoid silicics, ignimbrites, unwelded tuffs, ash flows, rhyolites and trachytes (N1-2n) constitute a considerable portion of the rift escarpments especially in the western part near Munesa town, while in the floor they are unconformably overlapped by younger volcanic of the Dino formation. The time of formation of the Nazareth group is between 9.5 and 3 Ma years (KAZMIN and SEIFE M., 1978). More than 400 m of Nazareth pyroclastic rocks have been encountered in deep geothermal exploration well at Aluto volcano (ABEBE, 1984; EIGS-ELC, 1985 and others). Furthermore, a thickness of more than 600 m for the Pliocene pyroclastic units in the eastern margin of the rift in the Munesa area stated by WOLDEGABRIEL et al., (1990). Another formation of the Nazareth group is the alkaline and per alkaline silicics and ignimbrite (N1-2ar) formed together and found East of Wondo Genet forming the escarp of Hawassa Caldera (ZENEBE et al 2016).

2.5.3.2.2 Bofa Basalts (Nbb)

Bofa basalt is considered to represent the upper part of the stratoid basaltic succession of the Afar Group younger than 4.5 Ma (KAZMIN et.al., 1980). The Bofa Basalts form a wedge between the

Nazareth and Dino silicics and occur as separate exposures north and east of lake Langano where they are recognized as ‘old Rift Basalts’ (UNDP, 1973), and where they finally wedge out. They represent an event of fissural eruption that immediately followed a major faulting event (KAZMIN and SEIFE, 1978, KAZMIN et al., 1980). The fissural basalts of this unit are aphyric locally vesicular and generally form several well-defined flows with scorraceous surfaces, in places separated by paleo soils. The rocks are generally weakly porphyritic, with holocrystalline ophitic groundmass with olivine andesitic labradorite and augite. Andesine is common in the intermediate rocks. K-Ar ages of 2.5 and 2.8 Ma were reported from these basalts, with possible lower age 3.5 Ma and upper 1.5 Ma (KAZMIN et al, 1980), (ZENEBE et al 2016).

2.5.3.2.3 Chilalo Volcanics (Nct, Ncb)

Chilalo, Bada, Kubsa, Kaka, and Enquolo are shield volcanoes located on the eastern rift shoulder, marking the water shade of the rift valley lake basin on the east together with Wechacha, Gash Megal and others; they belong to a numerous group of complex volcanoes developed in the upper Pliocene on both sides of the rift (HALCROW GROUP and GIRD 2008). Lava flows of these volcanoes are contemporaneous with the younger ignimbrites of the Nazareth group. The lower complex is composed of intercalating per alkaline ignimbrites and trachytes (Nct) with clear flow structures in the latter. Very fresh alkaline basalts (Ncb) commonly with porphyritic structure (ZENEBE et al 2016) represent the upper complex flow invariably.

2.5.3.2.4 Nazareth Group (Npp)

The silicic rocks of the Dino Formation are not only confined to the rift floor but are extensively developed in escarpments and on the adjacent plateau where they could not be separated from the silicics of the Nazareth Group they were mapped as Nazareth Group. In the rift escarpments where 500 meters and more of silicic rocks is locally exposed these units occurred. It was noted that on both sides of the rift the upper part of the section normally bears evidence of lacustrine sedimentation in the form of bedded water lain tuffs, sediments, etc. (ZENEBE et al 2016).

2.5.3.3 Quaternary Volcanics /Wonji Group

The latest volcanism in the Ethiopian Rift is associated with the axial extensional zone, referred as the Wonji Fault Belt (MOHR, 1967., and MEYER et al. 1975) introduced the Wonji series for the volcanic rocks of the rift younger than 1.6 Ma that are related to the Wonji fault belt. The

Pleistocene volcanism, however, is not confined to the Wonji Fault Belt only, but that it also occurred in other parts of the rift (KAZMIN and SEIFE, 1978).

On this account KAZMIN and SEIFE M, (1978) changed Wonji series to Wonji group. The Wonji Group includes the entire rift volcanic formed after the last major event of rift faulting, following the accumulation of the Bofa basalts (Nbb) described before. Three main complexes are identified within the group (KAZMIN and SEIFE M, 1978; KAZMIN et. al., 1980). These are (1) stratoid silicic volcanics of the rift floor and shoulders, named as Dino Formation, (2) pantelleritic products of central volcanoes, and (3) fissural basaltic eruptions (basalts of the rift floor). (ZENEBE et al 2016).

2.5.3.3.1 Dino Formation (Qdi, Qdp)

These units cover a considerable portion of the rift floor and it comprises a number of flows of compacted fiamme ignimbrites in place intercalated with aphyric basalt and unwelded pyroclastics. Thickness varies from a few meters to up to 30m (KAZMIN and SEIFE M.1978; KAZMIN et al., 1980).

North-West of Ropi locality are presumably synchronous rocks represented by very coarse light colored unwelded pumiceous pyroclastics (Qdp) reaching thickness of 250 m. The eruption of these pyroclastics was possibly related to the formation of gigantic calderas such as Hawassa, Shalla and others.

2.5.3.3.2 Basalts of the Rift Floor (Qwbp, Qwh, Qwbh)

Another type of volcanic activity in the Wonji Group was the eruption from fissures of Pleistocene basalts (Qwbp) and recent basalts (Qwbh) the bulk of these late basalts are concentrated in the Wonji fault belt, but there are also eruptions in the western marginal part of the rift. The basalts are clearly controlled by extensional fractures and generally display fresh aa surfaces. Chains of scorraceous cones follow the lines of fractures. These basalts are exposed at East of lake Ziway, and South of lake Shalla. Recent flows in many cases follow depressions of relief or flow over fault escarpment. The eruption of basalts usually followed the formation of silicic pantelleritic volcanoes, but some basalt might be contemporaneous with the earliest stages of their development. Other manifestations of recent and sub recent basaltic volcanism are explosion craters and basaltic hyaloclastites (Qwh) were mapped in several localities south of lake Shalla and at the southern part of the study area around Bilate river. The basalts are alkaline or

transitional, mostly mildly alkaline. The absolute age determinations available place most of the basalts in the Pleistocene (Morton et al., 1979). (ZENEBE et al 2016).

2.5.3.3.3 Central Volcanic Complexes (Qws, Qwpu, Qwo)

Most of the central volcanoes of the Wonji group are disposed along the axial zone of the rift, the Wonji fault belt. They are either huge conical mountains or calderas formed in the place of older volcanoes. The main volcanic centers in the rift include Aluto, Shalla and Corbeti, Gademota, Bora, Bericha, Gudemssa and Tulumoye. The position of some volcanoes and calderas is not so clearly defined. Some occur on the rift floor outside the tensional belts, such as Gademota caldera, or even on the rift shoulders Tembero, Wagebeta and Ambericho, show discrete east–west alignment, and may follow transverse faults. The main rock types are rhyolitic and trachytic lava flows. In some centers alkaline and per-alkaline products are mainly represented by pumice and unwelded tuffs (Qwpu), often associated with obsidians and pitchstones (Qwo), the latter usually marking the latest stages of volcanic activity. The sequence of eruption of various rocks differs from one volcano to another. Most of the products of central volcanoes are per alkaline rhyolites and trachytes of pantelleritic or locally of comendetic affinity (GIBSON, 1970; DI PAOLA, 1976; BROTZU et al., 1980; TSEGAYE et al., 2005). Radiometric (mainly K-Ar) and fission track ages conducted on obsidian, alkali feldspar separated minerals gave generally Quaternary age, 1.6-0.1 Ma (MORBIDELLI et al., 1975; KAZMIN et al., 1980; WOLDEGEBRIEL et al., 1990) (ZENEBE et al 2016).

2.5.3.4 Quaternary Sediments

2.5.3.4.1 Lacustrine Sediments (Ql)

In the rift, Quaternary sediments and mostly lacustrine origin are intercalated with Pliocene to Pleistocene ignimbrites both in the rift floor and rift shoulders. The older sediments are lacustrine diatomites, tuffaceous clays and silts interbedded with basal ignimbrites of the Nazareth Group. At the down of Quaternary period a huge ancestral lake, stretching from the area around Chamo-Abaya lakes in the south up to Awash basin in the north existed until it separated in to the present lakes by Late Pleistocene tectonic movements. Pleistocene-Holocene lacustrine sediments cover a significant tract of ground and were deposited in a huge lake whose level before 3500 to 2100 years used to be 100m higher than its current level (KAZMIN et al., 1980). The lacustrine sediments are intercalated with redeposited volcanic ash and tuffs. The lacustrine sediments are mainly represented by sand and silt. The major components of the sediments are volcanic origin, such as

pumice and volcanic ash, obsidian, rhyolite and basaltic rock fragments (TSEGAYE et al., 2005; BEVENUTI et al., 2002). In the rift floor, interlayering of primary air fall pyroclastic deposits and lacustrine deposits are common. The subaqueous pyroclastic deposits can be identified by reverse graded bedding of pumice layers. Rounded rock fragments indicating abrasion during transportation by rivers and waves of lakes, poor consolidation, and presence of evaporites distinguish subaqueous pyroclastic materials from reworked sedimentary deposits. Evaporites such as soda or trona (NaHCO_3) and micro-organic (diatomite) deposits occur around Lake Ziway, North of lake Abiyata; the sediments attain a thickness of 600m. (ZENEBE et al. 2016).

Table 2.1 Stratigraphy of the CMER (adopted from WOLDEGEBREL et al., 1990).

Absolute age			Central MER Lithology
Era	Age		
	Quternary (1.6)		Quaternary sediment Wonji Group
Miocene	Pliocene (5.3)		Chilalo Trachyte
			Butajira Ignimbrite
		Upper 11.2	-----
			Guraghe basalt

		Middle 16.6	-----
		Late 23.7	Shebele Trachyte

		Upper 30	
		Lower 36.6	Kella Basalt

	Eocene		-----
Jurassic		Upper 30	
		Middle	Antalo limestone
		Lower	Adigrat sanstone
			Crystalline basement

2.6 Geological Structure of the Study Area (CMER)

The study area (CMER) is bounded by discontinuous boundary faults that give rise to major fault-escarpments separating the rift depression from the SW and SE plateaus. The rift axis orients between N25° and N45° and is characterized by major rift escarpments on both western and eastern margins; boundary faults show an average trend around N30°E (CORTI, 2009).

The western margin is well expressed by the N25°E–N35°E-trending and ESE-dipping Guraghe and Fonko faults, whereas the eastern margin is well represented by the N30°E-trending and WNW-dipping Munesa-Asela fault system (Fig 2.8). The western rift margin is characterized by the presence of roughly N–S trending structural highs (such as the Boru Toru and Midre Kebd structural highs (Fig 2.3). Both systems are characterized by high-angle (N60°) normal faults, with large cumulative vertical throw (CORTI, 2009).

Geological, Morphostructural and seismicity studies suggests that at least some segments of the boundary faults in the CMER (e.g. Munesa-Asela) still accommodate some extensional deformation (PIZZI et al., 2006), although most of active strain seems to be accommodated by the internal Wonji faults (e.g., BOCCALETTI et al., 1998; BONINI et al., 2005; ABEBE et al., 2005; PIZZI et al., 2006; KEIR et al., 2006a).

In the CMER, quaternary volcanic centers (Corbeti caldera, Shalla caldera, Aluto caldera Gedemssa Caldera) are located following the Wonji fault belt whereas the off rift volcanic centers (Kaka, Bada, Chilalo, Zikuala) are located on the eastern and western shoulder of the rift (Fig 2.8). The superimposed active faults (Fig 2.7, 2.8) appear to terminate at the location of the volcanic centers (or start shifting their propagation direction away from these volcanic centers) indicating that these volcanic centers are in a ductile state and do not accommodate brittle deformation.

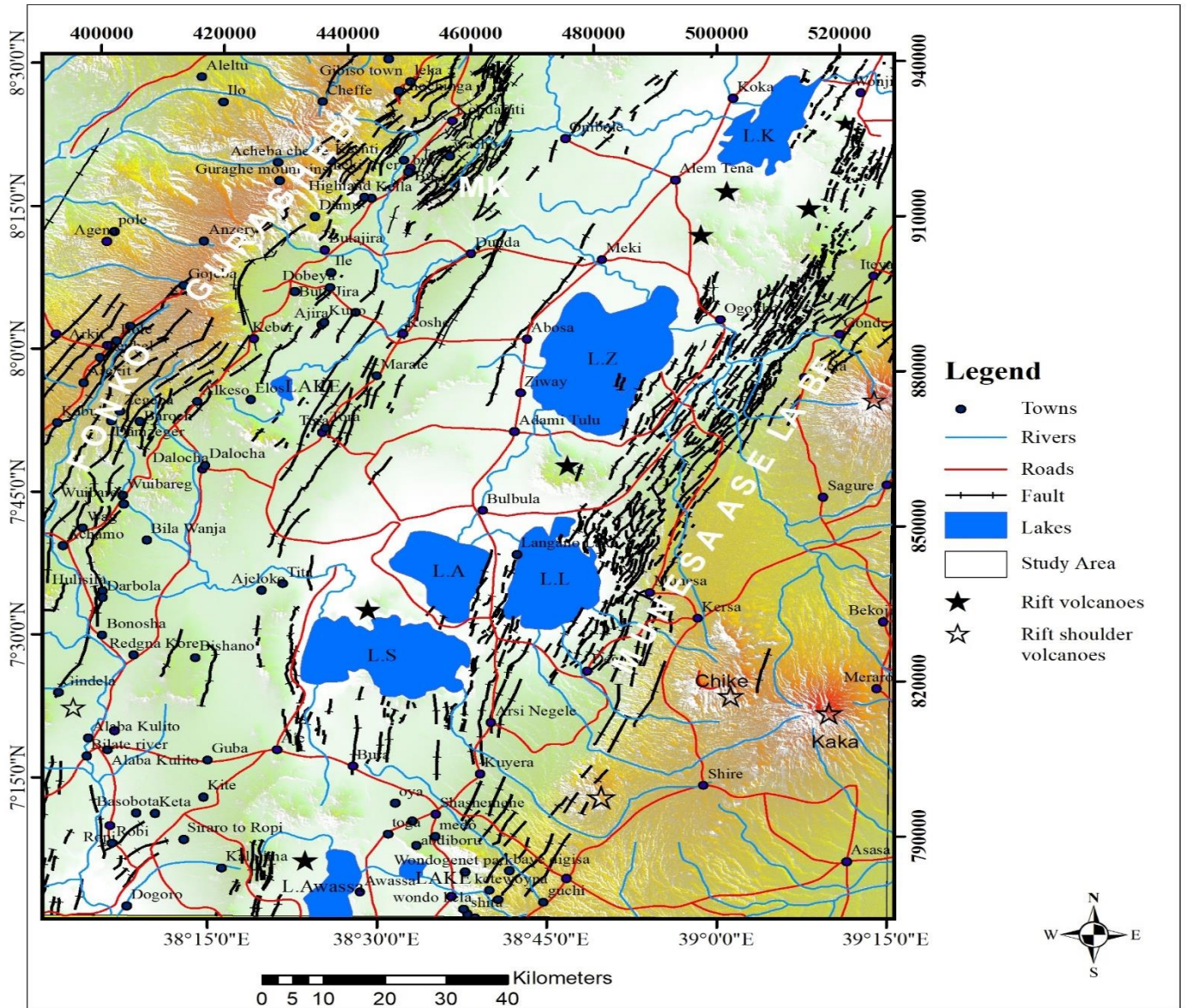


Figure 2.8 Structural map of the CMER

CHAPTER THREE

THEORETICAL FOUNDATIONS OF THE METHODS OF STUDY

3.0 INTRODUCTION

Geophysics, the science of the physics of the Earth from its magnetosphere to the deep interior, is useful in characterizing the subsurface of the Earth. It employs techniques involving the measurement of force fields to study subsurface features and the processes that act upon them. Thus, geophysical studies serve a broad variety of geologic, natural resource, engineering, and environmental purposes (WILLIAM, 2013). Gravity and magnetic methods, which measure very small spatial and temporal changes in the terrestrial gravity and magnetic force fields, have a wide range of uses from submeter to global scales. Although these methods in most cases fail to match the resolution and precision of direct observations, they are rapid, cost-effective, and non-invasive procedures of studying the inaccessible Earth and optimizing the location of drill holes for direct studies and other remote sensing studies that have higher resolution capabilities.

Gravity and magnetic methods are commonly referred to as potential field methods because the measurements involve a function of the potential of the observed field of force, either the terrestrial (airborne) gravity or magnetic field, at the observation site. These methods are widely used at a variety of scales to investigate the Earth because in comparison to most other geophysical methods, the acquisition of data is inexpensive and rapid, and for many applications, the reduction and interpretation of the observations are relatively simple. Furthermore, gravity and magnetic methods always provide information about the subsurface. In addition, there is a large reservoir of these data covering the entire Earth in varying detail that are publicly available at minimal cost to the user.

3.1 Basics of Gravity Method

3.1.1 Principles of Gravity Method

Gravity surveying is based on ISAAC NEWTON's Universal Law of Gravitation, described in *Principia Mathematica* in 1687, which states that the force of attraction between two bodies of

known mass is directly proportional to the product of the two masses and inversely proportional to the square of the distance between them. Mathematically given by:

$$\mathbf{F} = G \frac{Mm}{r^2} \quad 3.1$$

Where M and m are the masses of the two bodies, r the distance separating them, and $G=6.67 \times 10^{-11} \text{ Nm}^2\text{kg}^{-2}$ is the universal gravitational constant.

For a mass m on the surface of a uniform spherical Earth of mass M and radius R , the gravitational attraction on the small mass m is given by:

$$\mathbf{F} = \frac{GM}{R^2} m = gm \quad 3.2$$

The right hand side of equation 3.2, gm , is the weight of the mass and the term GM/R^2 is the acceleration due to gravity, or just gravity.

$$g = \frac{GM}{R^2} \quad 3.3$$

The gravitational acceleration g on the surface of the earth is 9.80 ms^{-2} on average.

If the Earth were, a uniform sphere g would be constant. However, the Earth's ellipsoidal shape, rotation, irregular surface relief and internal mass distribution cause gravity to vary over its surface. The gravitational field of the earth is usually defined in terms of its scalar gravitational potential U as;

$$U = \frac{GM}{R} \quad 3.4$$

The gravitational potential U is a scalar, having magnitude only whereas the gravitational acceleration g is a vector quantity, having both magnitude and direction (vertically downwards). The first derivative of U in any direction gives the component of gravity in that direction. Consequently, a potential field approach provides computational flexibility. Equipotential surfaces can be defined on which U is constant. The sea-level surface, or geoid, is the most easily recognized equipotential surface, which is everywhere horizontal, that is, at right angles to the direction of gravity.

3.1.2 Units of Gravity

In honour of GALILEO, the c.g.s. unit of gravity is called the Gal. 1 Gal= 1 cm/s². Modern gravity meters are extremely sensitive and can measure g to within 1 part in 10⁹. (Equivalent to measuring the distance from the Earth to the Moon to within 1 meter). So the c.g.s unit commonly used in gravity measurement is the milliGal: 1 mGal = 10⁻³ Gal = 10⁻³ cm s⁻². In m.k.s. SI units, gravity is commonly measured in μm s⁻² or g.u. (gravity unit). 10 g.u. = 1 mGal. Both mGal and g.u. are commonly utilized in gravity surveying.

3.1.3 Measuring Gravity

Measurable geophysical gravity anomalies generally range between 0.1 and 0.00001 Gal. This means we must measure accelerations of the order of 1 part in 10⁸ or 10⁹. Generally, in gravity method of exploration the measured parameter is the force on a mass, m, due to the presence of another mass, M, the recorded parameter is acceleration (g), with units of milliGals (compared to m/s²) and the interpreted parameter is usually density of causative buried materials and structures. Gravimeters actually measure the vertical component of the gravitational effect of the target, Δg_z. However, the formula Gm/r² gives acceleration experienced in line with two masses, where r is the distance between center of masses (sensor and target mass). Since we measure Δg_z along the surface over the target, we must resolve this geometry.

Using the Figure (3.1) below, Δg_z at location (x, 0, 0) is

$$\Delta g_z = \Delta g_r \cos(\theta) = \Delta g_r \frac{z}{r}$$

and since $\Delta g_r = G \frac{Mm}{r^2}$, then

$$\Delta g_z = G \frac{Mm}{r^3} = G \frac{Mz}{(x^2+z^2)^{3/2}}$$

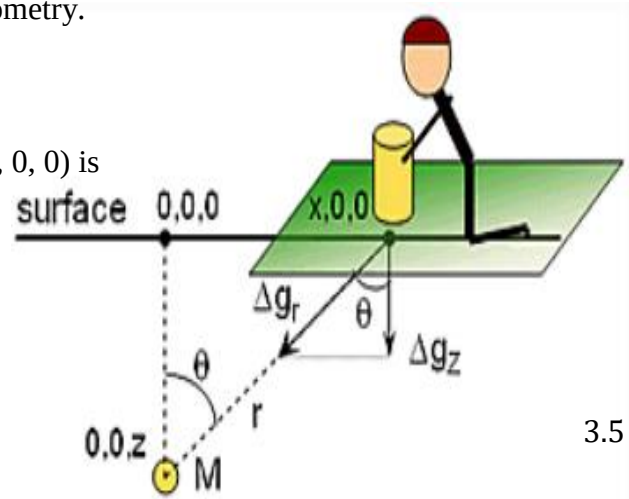


Figure 3.1 gravity measurement

3.1.4 The Physical Property: Density

Gravity surveys are usually conducted to find subsurface variations in density (kg/m^3 or g/cm^3). Densities of geologic materials vary from 880 kg/m^3 (ice) (or 0 for air) to over 8000 kg/m^3 for some rare minerals. Rocks are generally between 1600 kg/m^3 (sediments) and 3500 kg/m^3 (gabbro).

Table 3.1 Approximate densities and density ranges (g/cm^3) of some common rock types and minerals (after ALEMU, 2020).

Rock type	Range (g/ cm^3)	Average (g/cm^3)	Rock type	Range (g/ cm^3)	Average (g/cm^3)
Sediments (wet)			Metamorphic rocks		
Over burden		1.92	Quartzite	2.50-2.70	2.60
Soil	1.2-2.4	1.92	Schists	2.39-2.90	2.64
Clay	1.63-2.6	2.21	Graywacke	2.60-2.70	2.65
Gravel	1.70-2.40	2.0	Marble	2.60-2.90	2.75
Sand	1.70-2.30	2.0	Serpentine	2.40-3.10	2.78
Sandstone	1.61-2.76	2.35	Slate	2.70-2.90	2.79
Shale	1.77-3.20	2.40	Gneiss	2.59-3.00	2.80
Limestone	1.93-2.90	2.55	Amphibolite	2.90-3.04	2.96
Dolomite	2.28-2.90	2.70	Eclogite	3.20-3.54	3.37
			Metamorphic	2.40-3.10	2.74
Igneous rocks			Non-metallic minerals		
Rhyolite	2.35-2.70	2.52	Petroleum	0.6-0.9	-
Andesite	2.40-2.80	2.61	Ice	0.88-0.92	-
Granite	2.50-2.81	2.64	Sea water	1.01-1.05	-
Granodite	2.67-2.79	2.73	Lignite	1.1-1.25	1.19
Prphyry	2.60-2.89	2.74	Softcoal	1.2-1.5	1.32
Quartzdiorite	2.62-2.96	2.79	Anthracite	1.34-1.8	1.50
Diorite	2.72-2.99	2.85	Chalk	1.53-2.6	2.01
Lavas	2.80-3.00	2.90	Graphite	1.9-2.3	2.15
Diabase	2.50-3.20	2.91	Rocksalt	2.1-2.6	2.22
Basalt	2.70-3.30	2.99	Gypsum	2.2-2.6	2.35
Gabbro	2.70-3.50	3.03	Kaolinite	2.2-2.63	2.53
Peridotite	2.78-3.37	3.15	Orthoclase	2.5-2.6	-
Acid igneous	2.30-3.11	2.61	Quartz	2.5-2.7	2.65
Basic igneous	2.09-3.17	2.79	Calcite	2.6-2.7	-
Metallic minerals			Anhydrite	2.29-3.0	2.93
Oxides, Carbonates	2.3-2.55	2.45	Blotite	2.7-3.2	2.92
Bauxite	3.5-4.0	3.78	Magnesite	2.9-3.12	3.03
Limonite	3.7-3.9	3.83	Fluorite	3.01-3.25	3.14
Siderite	4.18-4.3	4.25	Barite	4.3-4.7	4.47
Rutile	4.2-4.4	4.32			
Manganite	4.3-4.6	4.36			
Chromite	4.3-5.0	4.67			
Ilmenite	4.7-5.0	4.82			
Pyrolusite	4.9-5.2	5.12			
Magnetite	5.0-5.22	5.12			
Franklinite	4.9-5.3	5.18			
Hematite	7.4-7.6	7.5			
Galena					

3.1.5 Gravity Field and Figure of the Earth

For a small mass, m , moving with a velocity on the surface of the earth rotating with angular velocity, ω , as shown in the Fig 3.2, it is possible to develop the theoretical gravity or the normal gravity value of the earth by decomposing different forces are acting on m .

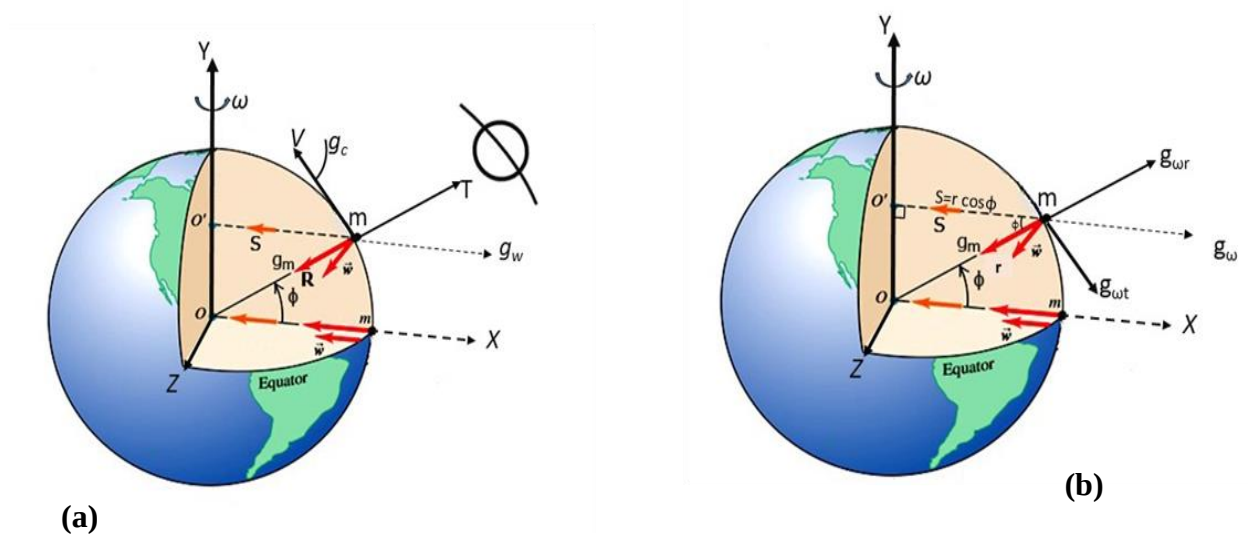


Figure 3.2. a) Gravity force per unit mass acting on a mass, m , (b) components of forces acting on mass, m on earth's surface

In Fig 3.2, g_m - mass gravitational acceleration acting on m due to the Earth's mass M ,

g_w - Centrifugal acceleration due to the Earth's rotation with angular velocity ω ,

T - Tidal force per unit mass acting on due to mass attraction of heavenly bodies,

g_c - Coriolis acceleration acting on m due to its linear motion with velocity v on the surface.

In Figure 3.2 a, the centrifugal acceleration g_w resulting from the centrifugal force per unit mass is given by:

$$\vec{g}_w = \vec{\omega}(\vec{\omega} \times \vec{R}) = \omega^2 R \cos \phi \quad 3.6$$

Similarly, the Coriolis acceleration resulting from the Coriolis force per unit mass is given by:

$$\vec{g}_c = 2(\vec{\omega} \times \vec{v}) \quad 3.7$$

If the mass m is at rest ($v=0$) on the surface of the earth, then $\vec{g}_c = 0$. In actual work the effect of $\vec{g}_c$ and T on g is usually considered negligible, hence, the gravity g refers to the combined effect of both mass gravitational acceleration and rotational (centrifugal) acceleration that is

$$\vec{g} = \vec{g}_m + \vec{g}_\omega \quad \text{or} \quad \vec{g} = \frac{GM_e}{R^2} + \vec{\omega}(\vec{\omega} \times \vec{R}) \quad 3.8$$

Considering Figure 3.2 (b) the centrifugal acceleration g_ω is directed radially outwards along the line $s = r \cos \phi$ is $g_\omega = \omega(\omega r) = \omega^2 r \cos \phi$. Further decomposing g_ω into its tangential ($g_{\omega t}$) and radial centrifugal acceleration ($g_{\omega r}$), $g_{\omega t} = g_\omega \sin \phi$, $g_{\omega r} = g_\omega \cos \phi$ and substituting g_ω in $g_{\omega r} = g_\omega \cos \phi$ yields:

$$g_{\omega r} = g_\omega \cos \phi = (\omega^2 r \cos \phi) \cos \phi = \omega^2 r \cos^2 \phi \quad 3.9$$

The mass gravitational acceleration g_m and $g_{\omega r}$ on m combine to yield the resultant gravitational acceleration $g = g_m + g_{\omega r}$. Conventionally, $g_{\omega r}$ is negative as it is directed radially outwards and g_m is positive as it directed towards the center of the earth. The resultant acceleration g becomes

$$g = g_m - g_{\omega r} = \frac{GM}{r^2} - \omega^2 r \cos^2 \phi \quad 3.10$$

Equation 3.10 shows that the effect of $g_{\omega r}$ is reducing the magnitude of g by $\omega^2 r \cos^2 \phi$. The equation also tells that the value of g varies with the variations in latitude ϕ .

The minimum and maximum values of g can be calculated using equation 3.10 by varying the latitude ϕ from $\phi=0^\circ$ (equator) and to $\phi=90^\circ$ (poles) respectively.

The minimum gravitational acceleration g occurs at the equator where $\phi=0^\circ$ and becomes

$$g_e = g_m - g_{\omega r} = \frac{GM}{r^2} - \omega^2 r \cos^2 \phi = \frac{GM}{r^2} - \omega^2 r \cos^2 0 = \frac{GM}{r^2} - \omega^2 r \quad \text{and}$$

The maximum gravitational acceleration g occurs at the poles where $\phi=\pm 90^\circ$ and becomes

$$g_p = g_m - g_{\omega r} = \frac{GM}{r^2} - \omega^2 r \cos^2 \phi = \frac{GM}{r^2} - \omega^2 r \cos^2 90 = \frac{GM}{r^2}$$

By substituting the mass of the earth, M_E (5.97×10^{24} kg), polar radius R_P (6356.75 km), equatorial radius R_E (6378.14 km) and angular velocity ω (7.29×10^{-5} rad/s) of the earth the magnitude of earth's gravitational acceleration is at the equator and poles is given by;

$$g_E = 978.032 \text{ gal} \quad \text{and} \quad g_P = 983.218 \text{ gal}.$$

For a spherical earth model of radius R , the difference between the calculated polar gravity and g_p and equatorial gravity g_E is determined to be 3.4 gal. i.e. $g_p - g_E = \text{gal}$ where $g_p > g_E$.

Using gravity measuring instruments on the real earth, polar gravity value g_p of 983.218 gal and equatorial gravity value g_E of 978.032 are obtained. The difference between these two observed gravity values is 5.2 gal. The disagreement between the differences in observed and theoretically computed gravity values at polar and equatorial regions is given by

$$(g_p - g_E = 5.2)_{\text{POLES}} \neq (g_p - g_E = 3.4)_{\text{EQUATOR}}.$$

These differences in the value of gravity can be explained by I). The shape of the real earth is not purely spherical rather oblate spheroid, rotationally distorted, flattened at the poles, and bulged at the equator and II) gravity value varies as a function of latitude (ϕ).

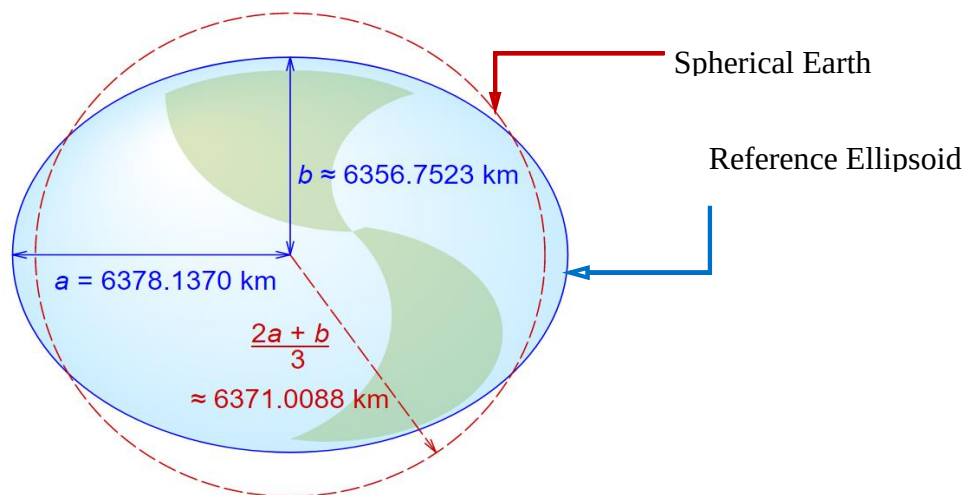


Figure 3.3 Equatorial (a), polar (b) and arithmetic mean Earth radii as defined in the 1984 World Geodetic System revision (not to scale).

3.1.6 The Geoids and Reference Ellipsoid

The equipotential surface of the earth's gravity field that coincides with the Mean Sea Level (MSL) in the absence of currents, air pressure variation, etc., is known as the geoid. Most of the efforts in physical geodesy are concentrated on the geoid determination with steadily increasing accuracy (HEISKANEN and MORITZ, 1967). The geoid is important for understanding more about ocean currents and is used as a reference for traditional height systems and monitoring sea-level change. The mathematical surface that best fits the geoid is the reference ellipsoid. The relation between the

geoid and the ellipsoid is shown in Figure 3.4. The ellipsoid lies below the geoidal surface in elevated regions (continental areas) and above the MSL over the oceans. The geometrical separation between the geoid and the reference ellipsoid is called the geoidal undulation, N , which generally varies globally between ± 110 m.

The earth's gravity field on the ellipsoidal surface varies only with the geographic latitude, ϕ , and is defined by the International Gravity Formula (WILCOX, 1989).

$$g(\phi) = 978.0327(1 + 0.0053024 \sin^2 \phi - 0.0000058 \sin^2 2\phi). \quad 3.11$$

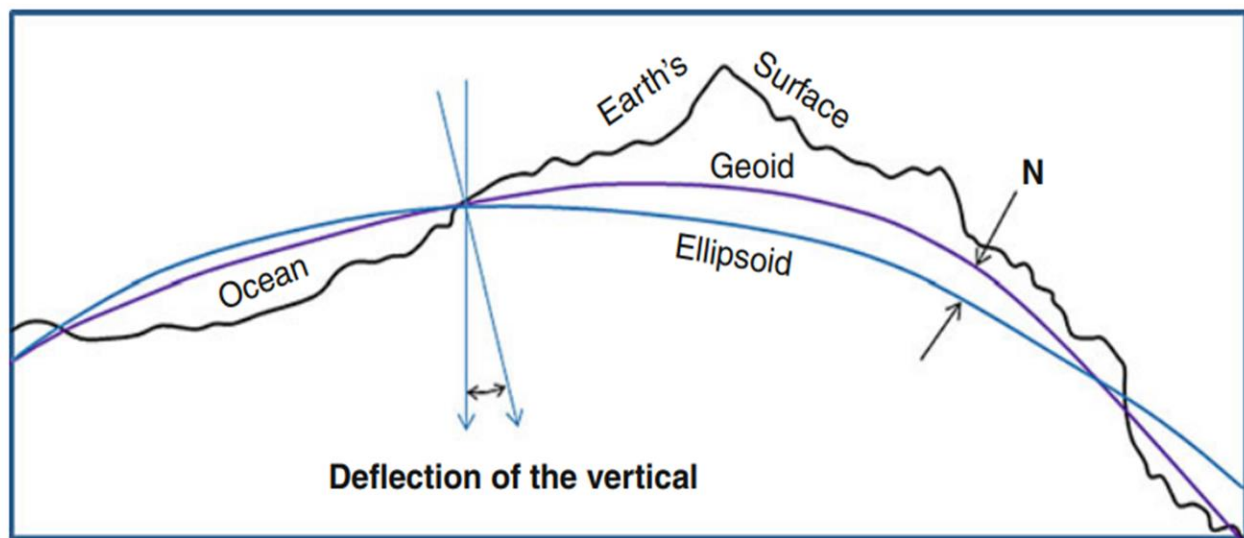


Figure 3.4 Relation between earth's topography, geoid and reference ellipsoid.

3.1.7 Principle Idea of Airborne Gravity

Airborne gravity has become a practical and accurate method to measure the variations of the Earth's gravitational field over regional areas. In geodesy, fixed wing aircraft equipped with scalar gravity meters are preferred. In geophysics, the same configuration is often used for reconnaissance surveys. For aero geophysical prospection, low- and slow-flying aircrafts are favored combined with vector or gradient gravity meters. Airborne gravity can either be used to determine the regional geoid more precisely or for geophysical reconnaissance and prospection. For geodetic purposes rather high altitude straight and level fixed wing aircraft flights utilizing scalar gravity meters are most common (FORSBERG et al., 1996). In geophysics, two aspects occur: large to mid-scale reconnaissance surveys to get an overview on tectonic or geological settings and small-scale prospection in search for mineral deposits or similar tasks (VERYASKIN and MCRAE,

2008). For reconnaissance, fixed wing aircraft with long-range capabilities are preferred. In prospection, high-resolution vector or gradient gravity meters are used on small fixed wing aircraft, helicopters, or airships.

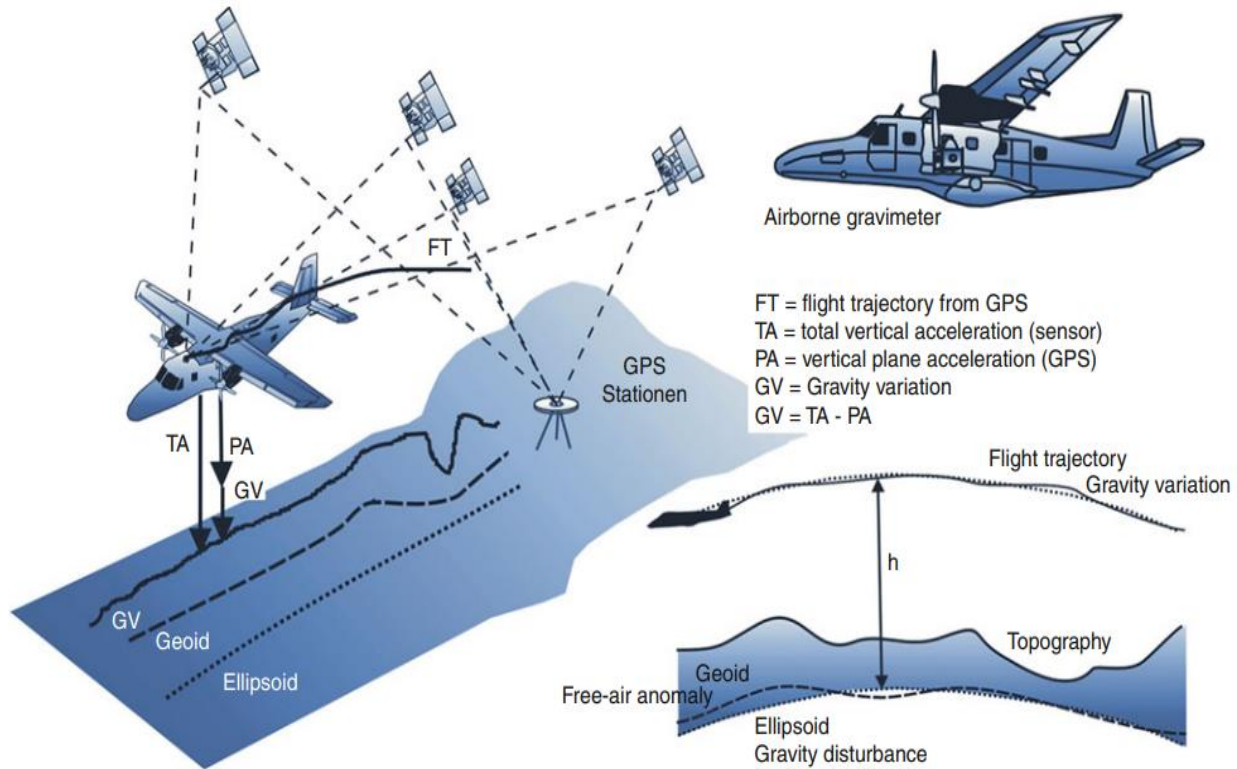


Figure 3.5 Principles of Airborne gravity (After UWE, 2014).

3.1.8 Gravity Data Reductions and Corrections

Raw or field gravity measurements are influenced by a wide variety of terrain, surface, geological, instrumental, and planetary sources. These sources have different amplitudes, periods in the case of temporal variations, and wavelengths in spatial variations that generally mask or at a minimum distort anomalies that are the objective of the gravity study. As a result, gravity observations must be modified to eliminate these unwanted, geologically extraneous effects. This process is commonly referred to as the correction or reduction of gravity data. Generally, variations in gravity field of the earth arises from either in temporal based variations which is time dependent changes in the observed acceleration (e.g. instrument Drift and Tidal effects and or Spatial Based variations which is space or position dependent changes in the observed acceleration (e.g., Latitude variation, Elevation Variations, Slab Effects and Topographic Effects). The reductions and corrections to be done on a raw gravity filed data are outlined as follows.

3.1.8.1 Instrument Scale Factor

The instrument scale factor corrects a reading to a relative milliGal value based on an instrument calibration. The correction can either be constant throughout the instrument range, or it can be derived from a user supplied calibration table.

$$R_c = R * S(R) \quad 3.12$$

where, R_c corrected reading in milliGals, R instrument reading in dial units and $S(R)$ is scale factor (dial units/milliGal), which will be a function of the reading if a calibration table is provided.

3.1.8.2 Tidal Correction

Tidal correction is a time-varying correction applied to the gravity survey data to compensate for Earth Tides. Often, tidal correction is included as a part of Drift Correction and referred to as Tidal Correction. This correction is necessary because the observed gravity field values at any station are affected by positions of moon, sun and heavenly bodies at the time of reading. This is done using the following relation,

$$R_t = R_c + G_t \quad 3.13$$

Where, R_t - tide corrected reading; R_c - scale corrected reading; G_t - tide correction.

3.1.8.3 Instrumental Drift Correction

Drift correction is time-varying correction applied to the observed gravity data to compensate for the gravimeter's drift with time. Instrumental drift takes place because gravimeter springs or torsion fibers are not perfectly elastic but subject to slow creep over long periods because of the thermal expansion, elastic aging or excessive movement. Often, drift is calculated as the linear approximation between Base Station observations recorded at the start and at the end of each Gravity Traverse based on the closure error between the first and last base reading in each loop:

$$d = \frac{(r_{B2} - r_{B1}) - (g_{B2} - g_{B1})}{(t_{B2} - t_{B1})} \quad 3.14$$

where d is the drift in milliGals/hour, r_{B1} base 1 reading, g_{B1} base 1 absolute G in milliGal, g_{B2} base 2 absolute G in miligals, t_{B1} base 1 time, r_{B2} base 2 reading, t_{B2} base 2 time.

3.1.8.4 Absolute Gravity

Absolute value for the gravity over a survey area is obtained from the local Base Station, which is tied to International Gravity Standardization Network (IGSN). The absolute gravity is the earth's vertical acceleration gravitational attraction at the observed station

$$g_a = g_{B1} + (r_h - r_{B1}) - (t - t_{B1})d \quad 3.15$$

Where, g_a absolute gravity in milliGal, g_{B1} base 1 absolute gravity in milliGal r_h instrument height corrected station reading, r_{B1} base 1 reading, t reading time, d drift in milliGal/hour.

3.1.8.5 Latitude Correction

Latitude correction is applied to the observed gravity data to compensate for the increase in gravitational acceleration (attraction) from the value of about 978000 mGal at the Equator to about 983000 mGal to the Poles due to a) variation of the Earth's radius because of Polar Flattening and b) variation of the centrifugal force (which opposes the gravitational acceleration) resulting from the Earth's rotation, as the distance to the Earth's axis varies with latitude.

Latitude correction can be obtained by differentiating International Gravity Formula (often, based on IGRS67). The latitude correction requires the theoretical gravity at the station location on the earth's spheroid.

There are three optional theoretical formulas for gravity latitude correction.

$$1930 \text{ formula: } g_\phi = 978049 (1 + 0.0052884 \sin^2 (\phi) - 0.0000059 \sin^2 (2\phi)) \quad 3.16$$

(BLAKELY, 1995, Cambridge University Press, p 135.).

$$1967 \text{ formula: } g_\phi = 978031.846(1 + 0.005278895 \sin^2 (\phi) + 0.000023462 \sin^4 (\phi)) \quad 3.17$$

(SHERIFF, 1984).

$$1980 \text{ formula: } g_\phi = 978032.7(1 + 0.0053024(\sin^2 (\phi)) - 0.0000058(\sin^2 (2\phi)) \quad 3.18$$

where g_ϕ Theoretical gravity in milliGals (latitude correction) and ϕ is the latitude of the station.

For local surveys, only the gradient due to latitude is important:

$g_\phi = 0.000812132 \sin (2\phi) \cdot y_s$ where g_ϕ is latitude correction, ϕ is latitude of station y_s station distance north of the grid origin in meters.

3.1.8.6 Free Air Correction

Free air corrections corrects for reduction in gravity with height above geoid, irrespective of the nature of the rock below. It is difference between gravity measured at sea level and at an elevation, h , with no rock mass in between. Free air correction is positive above sea level (i.e. added) and negative below sea level. Moreover, it varies slightly with latitude from 3.083 g.u./m at equator to 3.088 g.u./m at poles. With usual gravimeter measurement, having an accuracy of 0.1 g.u elevation must be known to within 5 cm range. Hence, free air correction is highly sensitive to elevations.

Free air correction is computed by

$$FAC = 0.3086 \text{ g.u./m} \quad 3.19$$

3.1.8.7 Free Air Anomaly

The gravity field anomaly at sea level after applying Free-Air Correction and Latitude Correction to the observed gravity data is known as free air anomaly. Because of a direct dependence on the elevation, free air anomaly maps are very reflective of topographic relief.

Free air anomaly is computed by

$$g_{fa} = g_a - g_\phi + 0.3086 * h_s, \text{ (SHERIFF, 1991)}. \quad 3.20$$

Where, g_{fa} -Free air anomaly in milliGals, g_a -absolute gravity, g_ϕ -Latitude correction, h_s -Station elevation in meters, ϕ -Latitude of the station.

3.1.8.8 Bouguer Correction

The free-air correction takes no account for the rock mass between the measurement station and the mean sea level. Therefore, in order to remove this effect Bouguer correction which accounts the effect of rock mass by calculating extra gravitational pull exerted by rock slab of thickness h and mean density ρ should be implemented.

$$BC = 2\pi G \rho h = 0.4191 \rho h \text{ g.u} \quad 3.21$$

Where h is the Bouguer plate thickness in meters, and ρ is the density in Kg m^{-3} . Bouguer correction quite sensitive to value of density used.

3.1.8.9 Simple Bouguer Anomaly

The simple Bouguer anomaly corrects the free air anomaly for the mass of rock that exists between the station elevation and the spheroid. It includes a correction for water depth and ice thickness.

For ground survey (including on lake surface survey) the simple Bouguer anomaly correction formula would be,

$$g_b = g_{fa} - 0.0419088 * [D * H_s + (D_w - D) * H_w + (D_i - D_w) * H_i] - GC \quad 3.22$$

For shipborne (marine) survey

$$g_b = g_{fa} - 0.0419088 * [D * H_s + (D_w - D) * H_w + (D_i - D_w) * H_i] - GC \quad 3.23$$

For airborne survey the Bouguer anomaly correction

$$g_b = g_{fa} - 0.0419088 * [D * H_g + D_w * H_w + (D_i - D_w) * H_i] - GC \quad 3.24$$

Where, g_b - Bouguer anomaly in milliGals, g_{fa} - Free air anomaly, D - Bouguer density of the earth in g/cc., D_w - Bouguer density of water g/cc, D_i - Bouguer density of ice in g/cc, H_s - Station elevation in meters, H_w - Water depth in meters (including ice), H_i - ice thickness in meters, GC - Curvature (Bullard B) correction., H_g - Ground (DEM) elevation at survey station location in meters. The purpose of the curvature correction (GC) as a step in producing the Bouguer anomaly is to convert the geometry for the Bouguer correction from an infinite slab to a spherical cap whose thickness is the elevation of the station and whose radius (arc length) from the station is 166.735 km. Curvature correction depends on the station elevation (e.g. 1.1 mGal at 1000 m, 1.5 mGal at 2000 m, 1.2 mGal at 3000 m, 0.2 mGal at 4000 m, -1.5 at 5000 m and -3.9 at 6000 m), and latitude (ranging in dozens micro Gal) (T.R.LAFEHR, 1991).

3.1.8.10 Terrain Correction

Terrain correction is a correction applied to the gravity data to compensate for the effects of local topographic features when the whole area (terrain) which cannot be approximated by Bouguer Slab, as not all the surroundings are at the same elevation as an observation point. Generally, terrain correction is applied when Bouguer Correction did not account for: a) the absence of mass below an observation point (i.e., Station); b) the absence of mass below and the presence of mass

above an observation point; c) the presence of mass above an observation point. Since these effects are close to the point of measurements, terrain correction can be important in high relief areas. For land and underwater stations, terrain correction is always added whether the feature is a hill or valley. For sea surface stations over relatively deep water, it may be negative. Terrain correction computation is made by dividing the topography into compartments or radial segment zones and totaling the individual effects at each station. Terrain corrections are calculated using a combination of the method described by NAGY (1966) and KANE (1962). To calculate corrections, Digital Elevation Model (DEM) data is sampled to a grid mesh centered on the station to be calculated.

3.1.8.11 Complete Bouguer Anomaly

The Complete Bouguer anomaly corrects the Bouguer anomaly for irregularities of the earth due to terrain in the vicinity of the observation point.

$$g_{CB} = g_b + g_t \quad 3.25$$

Where, g_{CB} - Complete Bouguer Anomaly in milliGals, g_b - Simple Bouguer Anomaly, g_t - Supplied terrain correction in milliGals.

3.1.8.12 Isostatic Correction (IC)

The worldwide average of Bouguer anomalies on land near sea level is approximately zero. In regions of large elevation, they are generally negative, while in oceanic regions mainly positive (TELFORD et al., 1990). These large-scale effects are due to density variations in the crust, indicating denser material beneath oceans and less dense material in elevated areas. As a result, two hypotheses were formulated in the 1850's to account for this large-scale variation in density. The first hypotheses given by Airy proposed a crust of uniform density but variable thickness floating on a liquid substratum of higher density, whereas Pratt suggested a crust where the density varies with topography, being lower in mountain regions and higher beneath oceans (TELFORD et al., 1990). Hence, isostatic correction occasionally is necessary in large-scale surveys to compensate for crustal variations.

3.1.8.13 Isostatic anomalies (IA)

Correspond to gravity anomalies resulting from a homogeneous crust having a flat surface, a standard reference density and constant thickness for the isostatic model (AIRY-HEISKANEN or PRATT-HEYFORD model). Isostatic Anomaly (IA) is given by the following equation,

$$IA = \text{Complete Bouguer anomaly (BA)} + \text{Isostatic correction (IC)}$$

$$IA = g_o - g_\phi - FAC \pm BC + T_c + IC. \quad 3.26$$

3.2. Basics of Magnetic Method

3.2.1 Principles and Quantities of Magnetic Method

Magnetic method of prospecting is the oldest branch of applied geophysics, which utilizes the Earth's natural magnetic field. Magnetic surveys record the spatial variation in the Earth's magnetic field. Every kind of material has unique magnetic properties, and different materials below the ground can cause local disturbances in the Earth's magnetic field that are detectable with sensitive magnetometers. CHARLES AUGUSTIN DE COULOMB, in 1785, showed that the force of attraction or repulsion between electrically charged bodies and between magnetic poles obey an inverse square law like that of the gravity force derived by ISSAC NEWTON.

3.2.1.1 Magnetic Force ($\vec{F}$)

Coulomb's law of Magnetic force states the magnetic force ($\vec{F}$) between two magnetic poles of strength m_1 and m_2 separated by a distance r varies inversely as the square of the distance between them and directly proportional to the product of their strengths. Mathematically:

$$\vec{F} = \frac{\mu_0 m_1 m_2}{4\pi \mu_r r^2} \vec{r}, \text{ where } \mu_0 = 4\pi \times 10^{-7} \text{ is magnetic permeability of vacuum: Vs/Am.}$$

$$\mu_r \text{ is relative magnetic permeability: } \mu_r = \frac{\mu}{\mu_0}$$

$$|r| = \sqrt{(x_2 - x_1)^2 + (y_2 - y_1)^2 + (z_2 - z_1)^2}$$

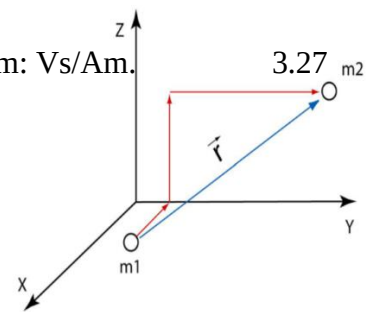


Figure 3.6 magnetic force between two poles.

3.2.1.2 Magnetic Induction Field ($\vec{B}$)

Magnetic induction is vector quantity designated by the symbol ($\vec{B}$) which defines the total magnetic field including the field within a body. Magnetic induction is a vector sum of the external

and internal field strengths. ($\vec{B}$) is expressed as the force created by a pole m and applied on a unitary pole m_1 .

$$\vec{B} = \frac{\vec{F}}{m_1} = \frac{\mu_0 m}{4\pi\mu_r r^2} \vec{r}, \text{ in Vs/m}^2 \text{ or Tesla (T).} \quad 3.28$$

3.2.1.3 Magnetic Moment ($\vec{M}$)

Magnetic moment is a vector parameter of Magnetic Dipole. For magnetic dipoles consisting of two poles of equal strength “ m ” and opposite signs that are separated by a distance “ l ”, the magnetic moment is defined as a vector joining the two poles,

$$\vec{M} = ml\vec{r}, \text{ in Am}^2 \quad 3.29$$

3.2.1.4 Intensity of Induced Magnetization ($\vec{J}_i$)

It is vector quantity that defines Magnetic Moment per unit volume. ($\vec{J}_i$) is a measure of a rock’s ability to be magnetized by the Earth’s magnetic field. Originally random-oriented elementary dipoles within a rock body are lined up in the direction of the Earth’s magnetic field creating a magnetic field of their own, i.e., induced magnetic field. The extent of this lining up is directly proportional to the strength of the Earth’s field and/or any other applied magnetic field at the date of magnetization.

Generally, rocks are magnetized by two mechanisms: 1) Remanent Magnetization, which may date from any time and may point in any direction and 2) Induced Magnetization, which points in the direction of the present Earth’s magnetic field. As a result, intensity of magnetization of any rock is a vector sum of the two-magnetization types – Remanent Magnetization and induced magnetization. Sometimes, intensity of magnetization is referred to as Magnetic Polarization or Intensity of Magnetization. ($\vec{J}_i$) is the ratio of the magnetic moment per unit volume v of the material.

$$\vec{J}_i = \frac{\vec{M}}{v} \text{ in A/m.} \quad 3.30$$

3.2.1.5 Magnetic Susceptibility (k)

It is magnetic property of rocks defining the degree to which the rock may be magnetized by an external (usually the Earth's) magnetic field. For each specific rock type, k is the proportionality constant between Intensity of Magnetization $\vec{J}_i$ and the external Magnetic Field Strength ("H"):

$$\vec{J}_i = kH, \quad k = \frac{\vec{J}}{H}, \quad \text{unit less.} \quad 3.31$$

Magnetic Susceptibility value is directly proportional to the volume percentage of highly magnetized minerals, like Magnetite, contained in a rock rather than a bulk rock property only. The general range of magnetic susceptibility values is about 10^5 . The main causes of Magnetic Susceptibility at atomic scale include the rotation of electrons in various shells around the nucleus and the spin of the electrons and number of electrons in each shell.

Table 3.2 Susceptibilities of rocks and minerals in SI units, (d) = diamagnetic material; (o) = ore; (m) = with magnetic $K \times 10^6$ SI units to the c.g.s. units divided by 4π . (ALEMU, 2020).

Mineral or rock type	Susceptibility*	Mineral or rock type	Susceptibility*
Ingeneous	10 to 65	Metamorphic	315 to 3000
Granite	20 to 50000	Schist	0 to 38000
Grinite (M)	250 to 37700	Slate	125 to 25000
Rhyolite	3000 to 75000	Geneiss	3100 to 75000
Pegmatite		Serpentenite	0 to 73000
Gabbro	800 to 76000	Average for various	-9
Basalts	500 to 182000	Minerals	
Oceanic basalts	300 to 36000	Ice (d)	-10
Pendotite	95500 to 196000	Rocksalt (d)	-13
Average for acid	40 to 82000	Gypsum (d)	-15
ingenous	550 to 122000	Quartz (d)	-80 to -200
Average for basic		Graphite (d)	400
ingenous		Chalcopyrite	50 to 5000
Sedimentaty	-12.5 to +44	Pyrite (o)	420 to 38000
Dolomite(Pure)	20000	Hyrrhotite (o)	1250 to 6.3×10^6
Dolomite(Imoure)	10 to 25000	Ilemenite (o)	314000 to 3.8×10^6
Limestone	0 to 21000	Magnetite (o)	
Sandstone	60 to 18600		
Shales	0 to 360		
Average for various			

3.2.1.6 Magnetic Field Strength ($\vec{H}$)

It is a quantity which describes how magnetic induction ($\vec{B}$) is modified by the magnetic polarization (or magnetization M). In a non-polarizable body, $\vec{H}$ can be regarded as simply a computational parameter proportional to $\vec{B}$.

$$\vec{B} = \mu_0 \vec{H} \text{ , where } \vec{H} \text{ is Magnetic field strength} \quad 3.32$$

When a magnetic material is placed in a magnetic field, the resulting magnetization gives rise to an additional magnetic field in the region occupied by the material. Within the body, the total magnetic induction is given by:

$$\vec{B} = \mu_0 \vec{H} + \mu_0 \vec{J}_i = \mu_0 \vec{H} + \mu_0 k \vec{H} = (1 + k) \mu_0 \vec{H} = \mu_0 \mu_r \mu_r \vec{H} \quad 3.33$$

$\vec{B}$ in Vs/m² or Tesla (T), $\mu_0 = 4\pi \times 10^{-7}$ Vs/Am. The unit used in geomagnetic surveys is the Tesla (T).

1 Tesla = 1 T = 1 N/Am, 1 nT = 10^{-9} T = 1 γ = 10^{-5} Oersted. The C.G.S units of magnetization include gauss (G) and gamma (γ). 1 gauss (G) = 10^{-4} T and 1 gamma (γ) = 10^{-5} G.

3.2.2 Origins of Earth's Magnetic Field

Total magnetic field of the Earth originates largely from within and around the Earth's core. About 99 % (94 % dipole field + 5% non-dipole field) of Earth's magnetic field has an internal origin. The remaining 1 % of the field originates from the current in the ionosphere (diurnal variations, magnetic storms). Earth's geomagnetic composite field is the vector sum of the anomalous field (B_A) and Earth's field (B_0) as shown in figure 3.7 below.

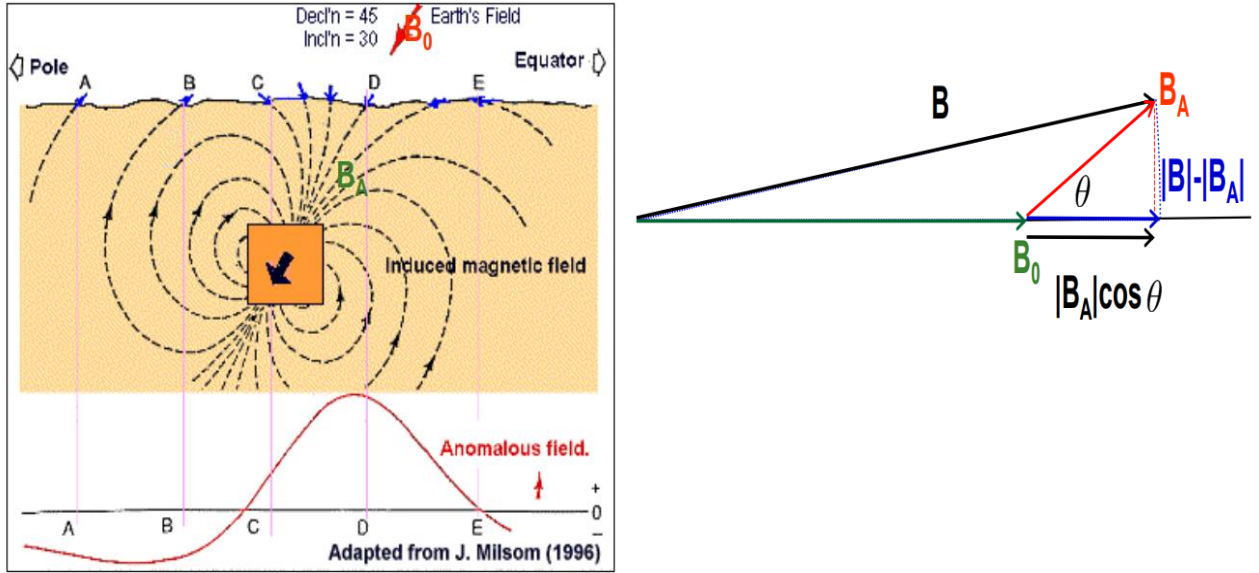


Figure 3.7 (a) Earth's geomagnetic field (b) vector diagram of anomalous and total field.

The composite field $B = B_0 + B_A$ is a vector field which can be decomposed into its three components (B_x , B_y , B_z). The measured total field is $|B| = |B_0 + B_A|$, and the total field anomaly is $\Delta B = |B| - |B_0|$. If $|B_A| \ll |B_0|$, then the total field anomaly ΔB is the projection of the anomalous field onto the direction of the inducing field. From figure 3.7 (b),

$$\Delta B = |B_0 + B_A| - |B_0| \cong B_A \cdot B_0 = |B_A| \cos \theta \quad 3.34$$

Generally, the earth's total magnetic field (B_T), having three origins is expressed as

$$B_T = B_{ext} + B_{int}, \quad 3.35$$

Where B_{ext} is the magnetic field, which originates from the flow of ionized particles (ions) emitted by the sun within the ionosphere towards the magnetic poles and B_{int} is the sum of the dipole field (B_D) originating from the fluid outer core and the magnetism (B_{rm}) which originates from magnetized crustal rocks. In geophysical exploration, our main objective is determining this anomalous rock magnetism.

$$B_T = B_{ext} + B_{in} = B_{ext} + B_D + B_{rm} \quad 3.36$$

3.2.3 Elements of the Geomagnetic Field

The Earth's magnetic field (geomagnetic field) is a vector quantity that spatially varies in magnitude and direction over the surface of the spherical Earth (Fig. 3.8a) as well as in time. Components of the *B-field* can be expressed as Cartesian components (*X*, *Y*, and *Z*) running parallel to the three orthogonal axes. These include, the *X*-component running parallel to the geographic north axis, the *Y*- component running parallel to the geographic east axis and the *Z*-component running downwards along the vertical axis (Fig. 3.8a). The direction of the *B*-field is specified by two angles known as declination *D* and inclination *I*. Declination *D* is the angle between the magnetic north axis and the geographic north axis; inclination *I* is the angle at which the geomagnetic vector *B* dips below its horizontal component *H* acting along the magnetic north axis (Fig. 3.8a). The Earth's magnetic field *B* at any observation site such as (Fig. 3.8a) is described by a set of seven quantities (*B*, *X*, *Y*, *Z*, *H*, *D* & *I*) known as geomagnetic elements. Using trigonometric identities geomagnetic elements can be derived as,

$$\mathbf{B}^2 = \mathbf{Z}^2 + \mathbf{H}^2 = \mathbf{Z}^2 + \mathbf{X}^2 + \mathbf{Y}^2, \mathbf{H}^2 = \mathbf{X}^2 + \mathbf{Y}^2, \mathbf{H} = \mathbf{B} \cos I, \mathbf{Z} = \mathbf{B} \sin I, \mathbf{X} = \mathbf{H} \cos D, \mathbf{Y} = \mathbf{H} \sin D,$$

$$I = \tan^{-1}\left(\frac{Z}{H}\right), D = \tan^{-1}\left(\frac{Y}{X}\right). \quad |\mathbf{B}| \text{ is total magnetic field, } \mathbf{D} \text{ declination, } \mathbf{I} \text{ inclination, } \mathbf{Z} \text{ vertical field,}$$

H-horizontal field.

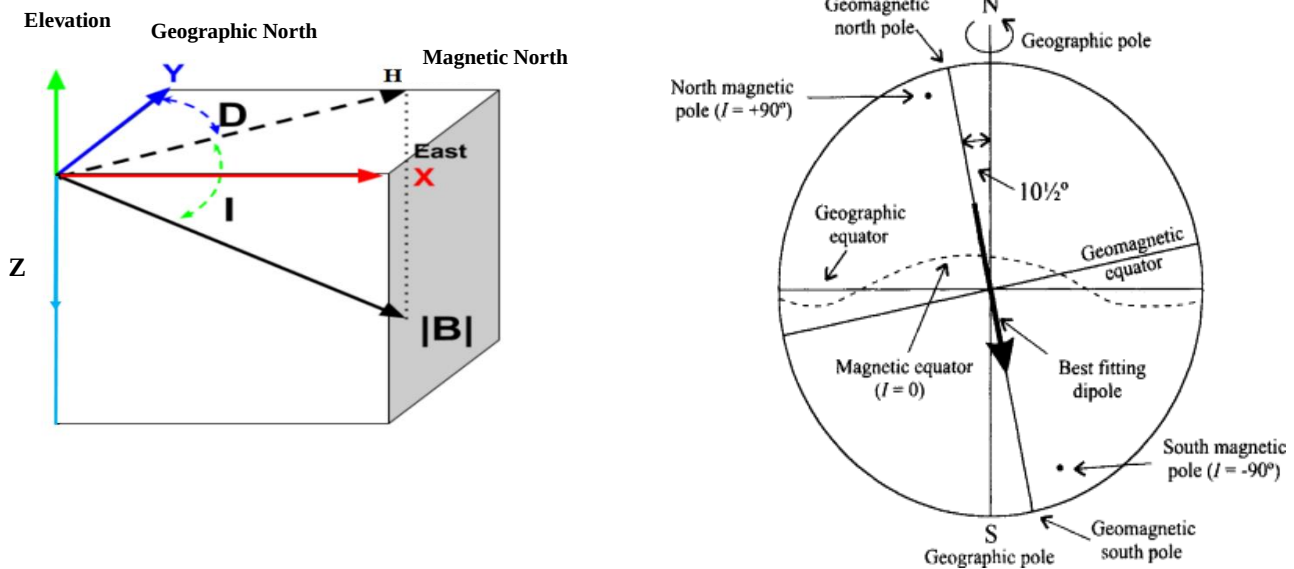


Figure 3.8 (a) Elements of earth's magnetic field. (b) Geographic and geomagnetic poles of the earth.

3.2.4 Temporal and Spatial Variations of the Earth's Total Magnetic Field

3.2.4.1 Temporal Variations of Earth's Total Magnetic Field

The variation of Earth's magnetic field with time known as temporal variation. These variations are classified into three types depending on their rate of occurrence and source. These include:

Diurnal Variation: These are variations in the Earth's magnetic field, which occur over the course of a day and are related to variations in the Earth's external magnetic field. Diurnal variation can amount approximately 20 γ - 30 γ per day and should be accounted for when conducting exploration magnetic surveys.

Magnetic Storms: The abrupt increase in the magnetic activity in the ionosphere occurring occasionally are called magnetic storms. The occurrence of magnetic storms correlates with enhanced sunspot activity. The magnetic field observed during such times is highly irregular and unpredictable, having amplitudes as large as 1000 γ . Exploration magnetic surveys should not be conducted during times of magnetic storms. Because, it is difficult to correct for magnetic storms in the acquired field data. Occurrence of magnetic storms can be detected from continuous records made at nearby magnetic observatories or from repeated measurements made at magnetic base stations during the days a magnetic survey is performed.

Secular Variations: Long-term changes in the geomagnetic field (presumably produced by fluid motion in the Earth's Outer Core) that occur over years are called secular variations. Since secular variations occur slowly with respect to the time of completion of a typical exploration magnetic survey, secular variations will not complicate data reduction efforts of the acquired magnetic field data.

3.2.4.2 Spatial Variations of Earth's Total Magnetic Field

As does the gravity field, the Earth's magnetic field also varies spatially. The International Geomagnetic Reference Field (IGRF) represents a set of reference models to describe the Earth's magnetic field. It is a standard mathematical model for describing the Earth's main magnetic field (dipole field) including its spatial and secular variation using spherical harmonics. The IGRF model is a function of latitude, longitude, time, and elevation.

3.2.5 Magnetic Instruments

Instruments for measuring aspects of the Earth's magnetic field are among some of the oldest scientific instruments in existence. In exploration magnetic surveys the Earth's magnetic field is measured by using instruments called magnetometers. Magnetometers are instruments that are capable of measuring strength of the total field B , the horizontal component H and the vertical component Z of the Earth's magnetic field. Two types of magnetometers are frequently used in magnetic surveying. These are proton precession magnetometer and optically pumped (cesium vapor) magnetometers. There are also other groups of magnetometers known as fluxgate magnetometers. Magnetic gradiometers are also instruments used in magnetic surveying to measure either horizontal or vertical magnetic gradients. These instruments are useful in shallow geophysics to resolve complex anomalies.

3.2.6 Reduction of Magnetic Data

As compared to the reduction of gravity data, magnetic data require very few corrections. One effect that must be compensated for is the variation in the intensity of the geomagnetic field at the Earth's surface during the course of a day known as diurnal variation. The observed total field value (raw data) as measured by the magnetometer has to be corrected for the effect of diurnal variation known as diurnal correction δB_D or diurnal drift correction.

3.2.6.1 Diurnal Correction

The correction applied to the observed total field B_T to remove the time varying external field B_{ext} over the course of a day is known as diurnal correction δB_D ($\delta B_D = B_{ext}$). Diurnal variation seldom reaches as much as 100 γ over 24 hours. During occasional magnetic storms, it can exceed 1000 γ . In ground-based magnetic surveys, a local base station is established within the survey area at a site free from a suspected magnetic noise for computing the diurnal correction δB_D .

Repeated readings in a given station (base station) with a constant interval of time enables us to determine diurnal correction as:

$$\delta B_D = B_i \pm \left(\frac{\Delta B}{\Delta t} \right) \Delta t_i \quad 3.37$$

Where, δB_D is diurnally corrected data measured at time t_i , B_i is the magnetic field measured at time t_i . $\Delta B = B_2 - B_1$; where B_1 and B_2 are base station reading at the beginning and end of the

survey respectively. $\Delta t = t_2 - t_1$, where, t_1 and t_2 are base station reading times at the beginning and end of the survey respectively and $\Delta t_i = t_i - t_1$ is the time difference of any station and base station respectively.

3.2.6.2 Latitude (Geomagnetic) (IGRF Removal) Correction

This correction is equivalent to the latitude correction in gravimetry (reference ellipsoid). We can use the International Geomagnetic Reference Field (IGRF) data, updated every 5 years, which defines the theoretical undisturbed magnetic field at any point of the earth's surface to calculate IGRF correction.

$$B_{IGRF-corrected} = \delta B_D - IGRF \quad 3.38$$

Where, $B_{IGRF-corrected}$ is the IGRF corrected total magnetic intensity, δB_D is diurnally corrected magnetic intensity value and IGRF is the International Geomagnetic Reference Field value defined by its latitude.

3.2.6.3 Anomalous Magnetic Field

In magnetic surveys, we measure the earth's total magnetic field and our main interest might be the anomalous field generated by anomalous bodies. Hence, we have to estimate and subtract the regional magnetic gradient and need to isolate with the residual magnetic anomaly. The anomalous magnetic field can be found by subtracting the regional anomaly from the total magnetic anomaly data.

$$\Delta B = B_i - \delta B_D - IGRF \quad 3.39$$

ΔB – anomalous field, B_i is the observed TMI at each station, δB_D is the diurnal correction and IGRF is the International Geomagnetic Reference Field value defined by its latitude.

3.3 Fundamentals of Inversion in Geophysics

There is an increasing realization that geophysics, which is the only remote sensing discipline available for subsurface problems, must be an integral part of environmental, engineering and exploration problems. There is also another understanding that extracting the information about the subsurface from the geophysical data is not a turn-key operation.

Inversion (or inverse modelling) is the process of mathematically estimating one or more models of subsurface physical property distributions that could explain a data set that was gathered in the field.

Inversion in Applied geophysics contributes excellently towards the solution of a particular problem. Geophysical inverse theory is an exceedingly large topic and not its entire aspects can be covered in depth in this MSc thesis. Generally, there are two types of problems that are regularly referred to as “inverse” problems (OLDENBURG, 2005).

The most common of these is concerned with obtaining estimates for a few parameters (models) when there are many more data than the unknowns are; we refer to this as the parameter estimation problem or overdetermined problem. These problems are generally solved with least squares approaches and there are a substantial number of books and papers available on this topic (e.g., LAWSON and HANSON, 1974).

The second type of inverse problem is known as generic inverse problem where our goal is to find a physical property distribution (density, susceptibility) when the number of data is less than the parameters to be estimated. This problem is also known as an underdetermined problem. There are also other sub-classes of inverse problems called mixed determined inverse problems.

A geophysical remote sensing experiment is typified by the illustration in Figure 3.9. It consists of an energy source, an earth model, and a response (D.W.OLDENBURG, 2014). The energy source can be natural, as is the case for gravity, magnetics, and magneto telluric surveys, or it can be a man-made device, such as a vibrator or explosive source used in seismic studies, or an electric current used in electromagnetic surveys. The earth model is characterized by 3-D distributions of physical properties and the responses can be fields that are measured on or above the surface of the earth or in boreholes. The propagation of energy through the earth depends upon the 3-D distribution of one or more physical properties, i.e., each datum is sensitive to a property variation in a volume.



Figure 3.9 The main components of a geophysical remote sensing experiment

Before solving any geophysical inverse problem, we have to first tackle the forward problem (D.W.OLDENBURG, 2014). With reference to Figure 3.9, the forward problem involves calculating the responses under the assumption that the sources and the earth model are known.

Generally, data from any geophysical experiment can be generically written as,

$$F_j[m] = d_j + n_j \quad j = 1, 2 \dots N \quad 3.40$$

Where F is a forward modelling operator that incorporates details of the survey design (the physics or the geometry) and the relevant physical equations, m is a generic symbol for a physical property distribution (density or susceptibility), and the right hand side represents the observed data, d^{obs} , which consist of the true data d plus additive noise n . F comes in a variety of forms, but more often it is an integral or differential operator, and the resultant equations must be solved numerically.

In a geophysical survey, we acquire N number of data, d^{obs} , and some knowledge about their uncertainties. The goal of the inverse problem is to find the model m (density in gravity and susceptibility in magnetics) that produced the noisy observations. This is a much more difficult process than forward modelling. The primary reason is that each geophysical datum depends upon a volumetric distribution of a physical property and information about the property is encoded in the data in a complex way. Secondly, it is unrealistic to expect that we can determine a 3-D physical property distribution uniquely when we have only a few data. Even if each datum provided the precise value of a physical property at some location, then we have only N such values. That is not enough to provide the solution everywhere. It follows that any solution we obtain from an inverse algorithm must be non-unique, and if we find one solution that acceptably fit the data, then there are an infinitely many others that will just fit as well. Selection of a single (best guess) answer will require additional information.

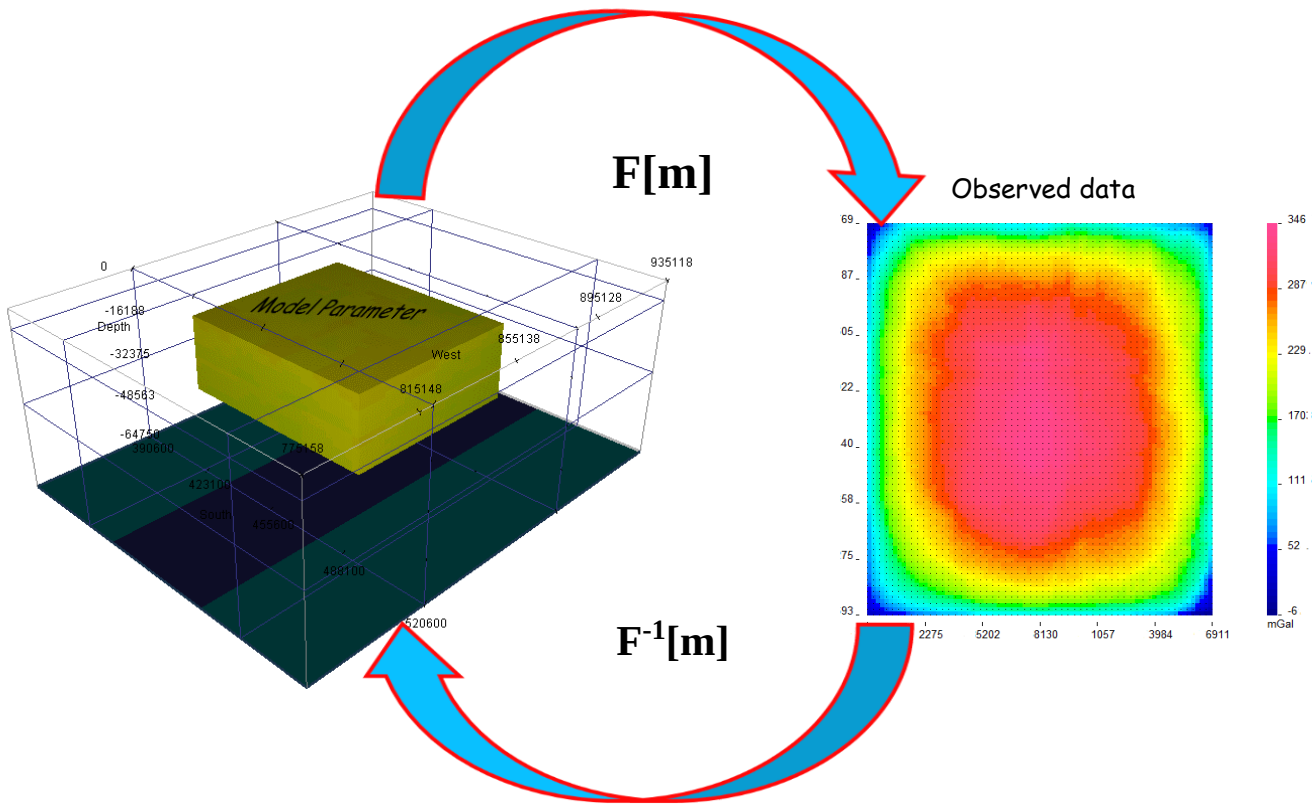


Figure 3.10 Forward and inverse modelling in Applied Geophysics

Forward modelling $F[m]$ is calculating data based up on known earth model parameters while Inversion $F^{-1}[m]$ is estimating a model based up on measured data (ground, airborne, borehole) and some understanding of the setting.

Inversion of potential field data is impeded by several numerical difficulties (OLDENBURG, 2005).

1. Non-existence: Due to the ubiquitous presence of noise in geophysical and geological observations, there may not exist a single model capable of fitting the measured data.
2. Instability: Small changes in the data, such as noise, can result in large changes in the recovered model since the inverse problem is ill-conditioned.
3. Non-uniqueness: There are two sources of non-uniqueness. (i) a finite number of any noise-free data can be reproduced in an infinite number of ways and (ii) Any potential field response can be reproduced by an equivalent layer of sources so there are an infinite number of physical property models that could reproduce gravity or magnetic datasets.

In order to overcome the non-uniqueness of potential field data inversions we have to construct a model objective and data misfit function.

3.4 The UBC-GIF 3-D Gravity and Magnetic Inversion Methodology

The 3-D gravity and magnetic inversion work in this MSc Thesis research uses the latest inversion codes of (YAOGUO LI and DOUGLAS W. OLDENBURG, 1996, 1998) developed at the University of British Columbia Geophysical Inversion Facility (UBC-GIF), Vancouver, Canada. The UBC-GIF 3-D inversion codes are internationally recognized and best known for solving the vast majority of inverse problems in geophysics from small scale to regional scales and are excessively applied in many mineral exploration works.

3.4.1 UBC-GIF 3-D Gravity Inversion Algorithms

The Gravity 3-D suite of algorithms are used to invert gravimetric responses over a three-dimensional distribution of density contrast, or anomalous density (LI and OLDENBURG, 1998).

The 3-D inversion is solved as an optimization problem with the simultaneous goals of (I) minimizing a model objective function and (II) generating synthetic data that match observations to within a degree of misfit consistent with the statistics of those data (LI and OLDENBURG, 1998). To counteract the inherent lack of information about the distance between source and measurement, the formulation incorporates depth or distance weighting.

By minimizing the model objective function, distributions of subsurface density contrast are found that are both close to a reference model and smooth in three dimensions. The user controls the degree to which either of these two goals dominates by incorporating a priori geophysical or geological information into the inversion. Explicit prior information may also take the form of upper and lower bounds on the density contrast in any cell (OLDEBURG, 2005).

The regularization parameter (controlling relative importance of objective function and misfit terms) is determined depending upon how much is known about the errors in the measured data.

3.4.1.1 Background Theory

A gravity experiment involves measuring the vertical components of the gravity field produced by anomalous (either excess or deficient) mass beneath the surface. A distribution of anomalous mass, characterized by anomalous density $\rho(x, y, z)$, produces its own gravity field, g_s , which is superimposed on the ambient gravity field. By measuring the resultant field and removing the ambient field from the measurements through numerical processing, one obtains the field due to the anomalous mass.

The vertical component of the gravity field produced by the density $\rho(x, y, z)$ is given by:

$$g_z(r_o) = \gamma \int_V \rho(\mathbf{r}) \frac{z - z_o}{|r - r_o|^3} dv \quad 3.41$$

Where $r_o = (x_o, y_o, z_o)$ is the vector denoting the observation location and $r = (x, y, z)$ is the source location. V represents the volume of the anomalous mass, and is γ the gravitational constant (OLDEBURG, 1998).

3.4.1.2 Forward Modelling of Gravity Data

Forward modelling of gravity data is a linear problem and can be carried out by performing the integration in equation 3.41. The region of interest is first discretized into a set of 3-D prismatic cells by using a 3-D orthogonal mesh and assume a constant density contrast within each cell. This is the typical inversion methodology of UBC-GIF.

Given such a discretization, the gravity field at the i^{th} location can be written as:

$$\begin{aligned} d_i &\equiv (r_{oi}) \\ &= \sum_{j=1}^M \rho_j \left\{ \gamma \int_{\Delta V_j} \frac{z - z_o}{|r - r_{oi}|^3} dv \right\}, \\ &\equiv \sum_{j=0}^M \rho_j G_{ij} \end{aligned} \quad 3.42$$

Where ρ_j is the anomalous density, ΔV_j is volume of the j^{th} cell, d_i is introduced as the generic symbol for the i^{th} datum, G_{ij} quantifies the contribution of the j^{th} cell to the i^{th} datum.

3.4.1.3 Inversion Methodology of Gravity Data

Let the set of extracted anomaly data be $\mathbf{d} = (d_1, d_2, \dots, d_N)$ and the density contrast of cells in the model be $\rho = (\rho_1, \rho_2, \dots, \rho_M)^T$. The two are related by the sensitivity matrix

$$\mathbf{d} = \mathbf{G}\rho \quad 3.43$$

The matrix has elements, which quantify the contribution to the i^{th} datum due to a unit density in the j^{th} cell.

The inverse problem will now be formulated as an optimization problem where a global objective function, Φ , is minimized subjected to the constraints in equation (3.43).

The global objective function consists of two components: a model objective function, Φ_m , and a data misfit function, Φ_d such that:

$$\Phi = \Phi_d + \beta \Phi_m \quad 3.44$$

Where, Φ is the global objective function, Φ_d is the data misfit, β is the tradeoff parameter and Φ_m is the model objective function (model norm).

3.4.1.3.1 Data misfit (Φ_d)

This is the first part of the global objective function (Φ) which is responsible for ensuring the recovered model predicts data that fits the set of field observations. The data misfit can also be expressed as a finite sum:

$$\Phi_d(\rho) = \sum_{i=1}^N \Phi_{d,i}^2 = \sum_{i=1}^N \left(\frac{F[\rho]_i - d_i}{\varepsilon_i} \right)^2 \quad 3.45$$

Where N is the total number of observations (datum), $\Phi_{d,i}$ is the data misfit for each datum, d_i is the observed datum, $F[\rho]_i$ is the predicted datum for model ρ and ε_i is the uncertainty on datum i .

If we assume that, the noise on each datum is independent and Gaussian, then the uncertainties ε_i can be thought of as estimations of the standard deviation of the noise on each datum. Under this assumption, two things should occur if the recovered model fits the data properly: i) The data misfit $\Phi_d \approx N$, indicating the inversion has fit the signal without fitting the noise. ii) The individual data misfits when plotted are incoherent; that is, the recovered model does not fit the data well in some areas and poorly in others.

3.4.1.3.2 Model Objective Function

The second part of the global objective function (Φ) is the model objective function $\Phi_m(\rho)$, which ensures that the recovered model contains plausible geological structures. $\Phi_m(\rho)$ is where we control and impose structures on the recovered model. The model objective function gives the flexibility to incorporate as little or as much information as possible. At the very minimum, this function drives the solution towards a reference model ρ_o and requires that the model be relatively smooth in the three spatial directions (OLDEBURG, 1998).

The model objective function $\Phi_m(\rho)$ is constructed as;

$$\begin{aligned} \Phi_m(\rho) = & \alpha_s \int_V w_s \{w(z)[\rho(r) - \rho_0]\}^2 dv + \alpha_x \int_V w_x \left\{ \frac{\partial w(z)[\rho(r) - \rho_0]}{\partial x} \right\}^2 dv \\ & + \alpha_y \int_V w_y \left\{ \frac{\partial w(z)[\rho(r) - \rho_0]}{\partial y} \right\}^2 dv + \alpha_z \int_V w_z \left\{ \frac{\partial w(z)[\rho(r) - \rho_0]}{\partial z} \right\}^2 dv, \quad 3.46 \end{aligned}$$

Where the alpha parameters (coefficients) α_s , α_x , α_y and α_z control how important each of the four terms are relative to each other and are used to balance the contributions of the smallness and smoothness components, the functions w_s , w_x , w_y and w_z are spatially dependent, The function w_s is used to force the physical property of cells to be closer to the supplied reference model. The parameters w_x , w_y and w_z , are used to make the model-difference more or less smooth across cell boundaries to reproduce geological continuity or boundaries. The function $w(z)$ is a depth weighting function, ρ_0 is the reference model, and $\rho(r)$ is the recovered model (OLDEBURG, 1998).

The model objective function $\Phi_m(\rho)$ in equation 3.46 consists of two parts. The first component is the smallness term (measures smallness) and the last three components (measures smoothness) are the smoothness terms.

$$\Phi_m(\rho) = \Phi_{small}(\rho) + \Phi_{smooth}(\rho)$$

Where $\Phi_{small}(\rho)$ is the smallness term and $\Phi_{smooth}(\rho)$ is the smoothness term. $\Phi_{small}(\rho)$, the smallness part, defines how the model can vary from the reference model $\rho(r)$ in equation 3.46. The $\Phi_{smooth}(\rho)$, the smoothness term, measures the *smoothness* of the difference between the recovered model and the reference model in each of the three axes and ensure that any discrepancies between the recovered model and the reference model are spread over a region rather than concentrated in individual cells.

The Depth Weighting function $w_r(z)$ is designed to counteract the natural decay of the potential field response with distance from the source so that all cells have an equal likelihood of containing sources. This is necessary as there is no inherent depth information contained in the observed potential field response and a default solution to the inverse problem would result in a model with sources clustered near the surface.

The depth weighting function has the form (LI and OLDENBURG, 1998):

$$W_r(z) = \frac{1}{(z+z_0)^\beta} \quad 3.47$$

Where z is the depth to the center of the cell, z_0 and β are adjustable parameters used to match the weighting function to the kernel's decay with depth, usually $\beta = 2$ for gravity data and $\beta = 3$ for magnetic data (LI, OLDENBURG, 1996 & 1998).

3.4.1.3.3 The Beta (Trade off) Parameter (β)

Beta (β) (trade-off parameter) is part of the global objective function is equation 3.44 which controls the relative importance of the model smoothness through the model objective function and data misfit function. β is scaled to the total number of input data, and is usually automatically recovered through an optimization routine that ensures the target misfit is reached.

In the equation 3.44, when β is too large ($\beta \rightarrow \infty$) small model norm solution and large misfits will be produced and structural information is lost. While when β is too small ($\beta \rightarrow 0$) the data misfit will be small and the model norm will be large but this in turn will produce an overfitting of the data and noise will be imaged as structure.

The best optimal fit between Φ_d and Φ_m will be reached when the noise level is Gaussian and the correct standard deviation is used. This optimum value is achieved at the $\Phi_d^* = N$, where Φ_d^* is the target data misfit and N is total number of data. There are four types of choosing the regularization parameter β . These are Chi fact, generalized cross validation (GCV), L-Curve (Tikhonov curve) and fixed β (UBC-GIF 2002).

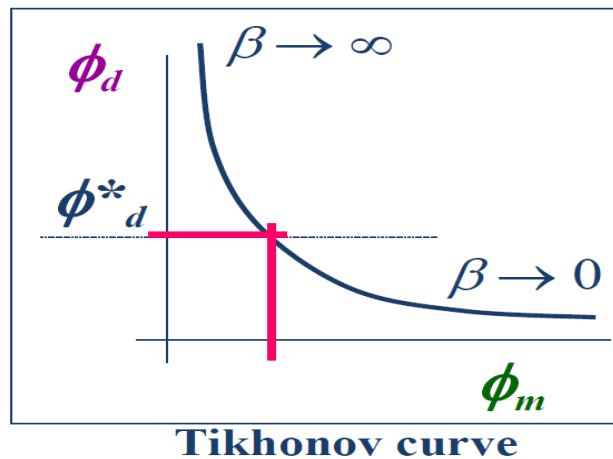


Figure. 3.11 Tikhonov curve (L-curve).

Finally, the solution to the inverse problem is obtained through a mathematical optimization process by minimizing the global objective function (Equation 3.44) as:

$$\text{Min } \Phi = \Phi_d + \beta \Phi_m,$$

Subjected to $\rho^l \leq \rho \leq \rho^u$, where β is a tradeoff parameter, ρ^l is the lower bound density and ρ^u is the upper bound density.

To find the numerical solution the model objective function in equation (3.46) is discretized onto the 3-D orthogonal mesh defining the density contrast model using a finite difference approximation.

3.4.2 UBC-GIF 3-D Magnetic Inversion Algorithms

A magnetic experiment involves measuring the anomalous magnetic field produced by magnetically susceptible materials beneath the surface, which have been magnetized by the earth's main magnetic field. The material with susceptibility $\kappa(x, y, z)$ is magnetized when the earth's main field impinges upon the subsurface formation with flux intensity B_0 . The magnetized material gives rise to a magnetic field, B_a , which is superimposed on the inducing field to produce a total, or resultant, field. By measuring the resultant field and removing the inducing field from the measurements through numerical processing, one obtains the distribution of the anomalous field due to the susceptible material (OLDENBURG, 1996).

3.4.2.1 Forward Modelling of Magnetic Data

For a given inducing field B_0 , the magnetization $\mathbf{J}$ depends upon the susceptibility through a differential equation. In most mineral exploration cases, the actual susceptibility is very small. Thus, we use the first order approximation, where the magnetization is proportional to the susceptibility and is given by the product of susceptibility and the inducing magnetic field $\mathbf{H}_0$.

$$\mathbf{J} = \kappa \mathbf{H}_0 \tag{3.48}$$

Where $\mathbf{H}_0 = \mathbf{B}_0 / \mu_0$ and μ_0 is the free space magnetic permeability. This ignores self-magnetization effect by which the secondary field reduces the total inducing field within the susceptible region and results in a weaker magnetization than that is given by equation (3.48).

The anomalous field produced by the distribution of magnetization $\mathbf{J}$ given by the following integral equation with a dyadic green's function:

$$\mathbf{B}_a(\mathbf{r}) = \frac{\mu_0}{4\pi} \int_V \nabla \nabla \frac{1}{|\mathbf{r} - \mathbf{r}_{oi}|} \cdot \mathbf{J} \cdot d\mathbf{v} \quad 3.49$$

where $\mathbf{r}$ is the position of the observation point and V represents the volume of magnetization at its position, $\mathbf{r}_o$. The above equation is valid for observation locations above the earth's surface (OLDENBURG, 1996).

3.4.2.2 Inversion Methodology of Magnetic Data

Let the set of extracted anomaly data be $\mathbf{d} = (d_1, d_2 \dots d_N)$ and the susceptibility of cells in the model be $\mathbf{k} = (k_1, k_2 \dots k_M)$. The two are related by the sensitivity matrix;

$$\mathbf{d} = \mathbf{G}\mathbf{k} \quad 3.50$$

The matrix has elements G_{ij} , which quantify the contribution to the i^{th} datum due to a unit susceptibility (k) in the j^{th} cell.

The inverse problem will now be formulated as an optimization problem where a global objective function, Φ , is minimized subjected to the constraints in equation (3.50). The global objective (Φ) function consists of two components: a model objective function, Φ_m , and a data misfit function, Φ_d such that:

$$\Phi = \Phi_d + \beta \Phi_m \quad 3.51$$

$$\text{Min } \Phi = \Phi_d + \beta \Phi_m, \text{ s. t. } \kappa^l \leq \kappa \leq \kappa^u,$$

Where β is a tradeoff parameter that controls the relative importance of the model smoothness through the model objective function and data misfit function, (κ^l) is the lower susceptibility bound and (κ^u) is the upper susceptibility upper bounds.

The next step will be the construction of a model objective function which, when minimized, produces a model that is geophysically interpretable. This function gives the flexibility to incorporate as little or as much information as possible. At the minimum, it drives the solution towards a reference model κ_o and requires that the model be relatively smooth in the three spatial directions.

Let the model objective function be,

$$\begin{aligned}\Phi_m(k) = & \alpha_S \int_V w_S \{w(r)[k(r) - k_0]\}^2 dv + \alpha_x \int_V w_x \left\{ \frac{\partial w(r)[k(r) - k_0]}{\partial x} \right\}^2 dv \\ & + \alpha_y \int_V w_y \left\{ \frac{\partial w(r)[k(r) - k_0]}{\partial y} \right\}^2 dv + \alpha_z \int_V w_z \left\{ \frac{\partial w(r)[k(r) - k_0]}{\partial z} \right\}^2 dv, \quad 3.52\end{aligned}$$

Where the alpha parameters (coefficients) α_S , α_x , α_y and α_z control how important each of the four terms are relative to each other and are used to balance the contributions of the smallness and smoothness components, the functions w_S , w_x , w_y and w_z are spatially dependent, The function w_S is used to force the physical property of cells to be closer to the supplied reference model. The parameters w_x , w_y and w_z , are used to make the model-difference more or less smooth across cell boundaries to reproduce geological continuity or boundaries. The function $w(r)$ is a depth weighting function, k_0 is the reference model, and $k(r)$ is the recovered model (OLDENBURG, 2005).

To find a numerical solution the model objective function in equation 3.52 is discretized onto the mesh defining the susceptibility model using a finite difference approximation.

As of the gravity inversion, static magnetic data have also no inherent depth resolution. That is the constructed susceptibility is concentrated close to the observation locations. This is a direct manifestation of the kernel's decay with the distance between the cell and observation locations. Because of the rapidly diminishing amplitude, the kernels of magnetic data are not sufficient to generate a function that possess significant structure at locations that are far away from observations. In order to overcome this, the inversion requires a weighting to counteract this natural decay. Intuitively, such a weighting will be the inverse of the approximate geometrical decay. This gives cells at all locations equal probability to enter into the solution with a non-zero susceptibility. Distance weighting for surface or airborne data and depth weighting for borehole data should be applied in the inversion process. (OLDENBURG, 1998).

CHAPTER FOUR

DATA ACQUISITION AND PROCESSING

4.1 Data Acquisition

4.1.1 Gravity Data Acquisition and Distribution

Airborne gravity has become a more practical and accurate method to measure the variations of the Earth's gravitational field over regional areas. This MSc thesis research work have used airborne gravity data collected covering the study area (Fig 4.1) with 18 km line interval to the study the crustal structure and Moho depth variation beneath the CMER. The gravity data is acquired from Ethiopian Geospatial Institute by following the formal procedures. The airborne survey was conducted using ET-ALD aircraft, a Cessna Caravan from Abyssinian Flight Service, from Oct 07– Nov 15 2006 GC. A total of 92,434 gravity data were collected during the survey time. The mean flight height of the aircraft above the ground was 2307m.

Gravity at the aircraft parking spots was connected to the IGNS71 network station located at the Geophysical Observatory in Addis Ababa University. The relative gravity measurements were conducted using LaCoste and Romberg land gravimeter (G-466). The gravity crossover analysis (RMS) analysis indicated a noise level of 2.7 mGal for the airborne data. (OLESEN, A. V et al., 2007).

The gravity data station spacing was about 833 meters and the maximum profile distance (line spacing) was found 18 km while the minimum profile spacing was 1.7 kms. 2049 of the airborne gravity data have been extracted and used for the investigation in this MSc thesis.

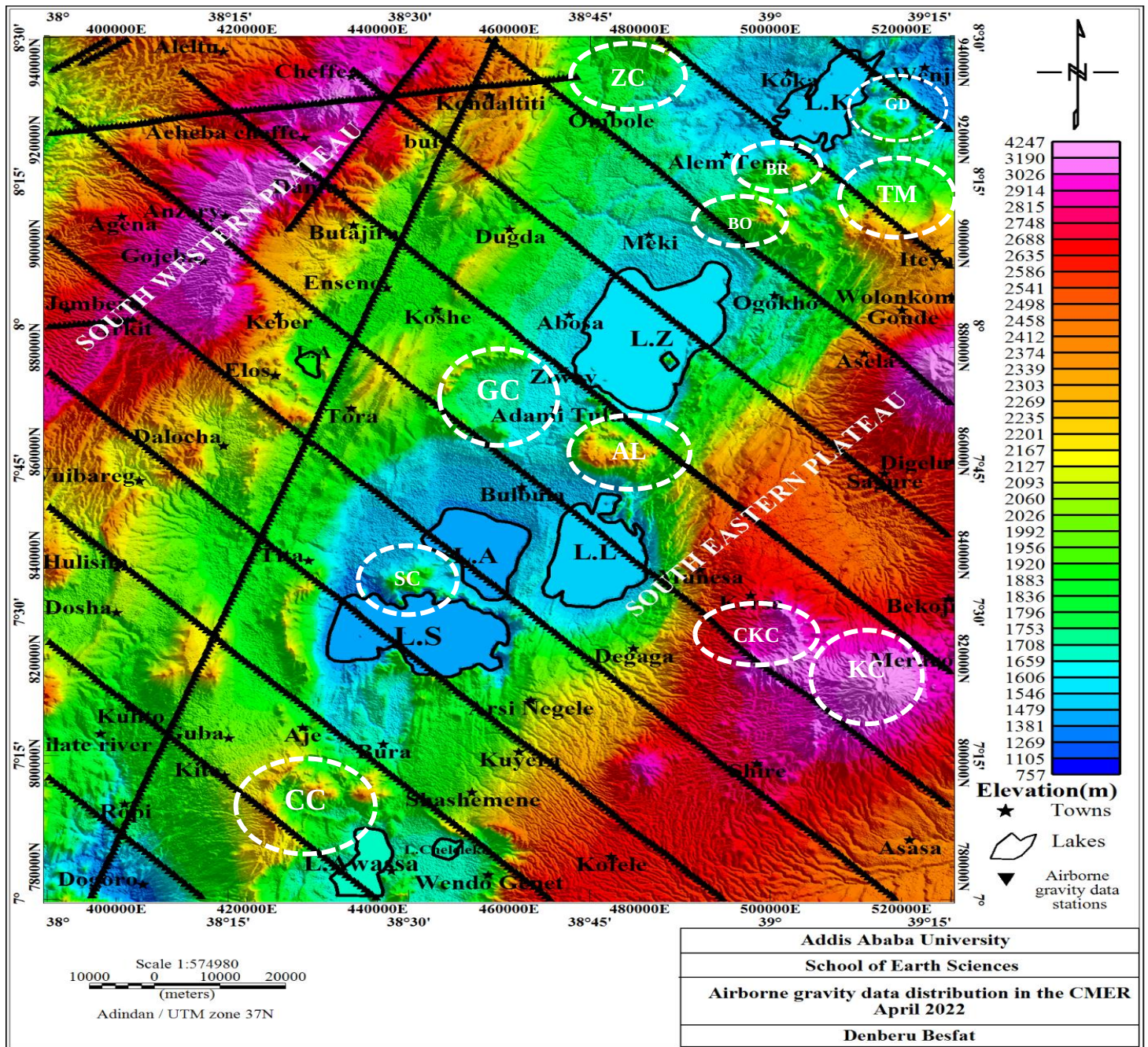


Figure 4.1 Airborne gravity data distribution superimposed on SRTM30m data in the CMER. (Labels: CC – Corbeti caldera, SC - Shalla caldera, KC - Kaka Caldera, CK - Chike Caldera, GC- Gademota caldera, AL -Aluto complex, ZC - Zikuwala Caldera, BO - Bora caldera, BR - Bericha caldera, TM - Tulumoye caldera, GD - Gedemssa caldera, L.Z – Lake Ziway, LA - Lake Abiyata, LK – Lake Koka, LS - Lake Shalla, LA - Lake Langano).

4.2 Gravity Data Processing

4.2.1 Gravity Data Reduction

The airborne gravity data obtained from the Ethiopian Geospatial Institute constitutes measured gravity at the level of the aircraft height, latitude and longitude in degrees, ellipsoidal height of the aircraft and height of the aircraft above the ground.

To reduce and correct the gravity data and compute the CBA before the necessary interpretation is performed Digital Elevation Model (SRTM30m) data is determined for each gravity station.

All the airborne gravity data have been processed by applying all the standard gravity corrections to the measured gravity data in order to reduce it to the mean sea level datum (reference geoid). These corrections include free air correction, simple Bouguer correction, complete Bouguer correction and terrain correction. The mean crustal density of 2.67 gm/cc, water density of 1.0 g/cc, and ice density of 0.97 g/cc are employed for computing topographic correction.

The 1967 International Gravity Formula have been used to compute the theoretical gravity value at each station. In addition, Curvature (Bullard) correction was also included. The purpose of the curvature is to convert the geometry for the Bouguer correction from an infinite slab to a spherical cap whose thickness is the elevation of the station and whose radius (arc length) from the station is 166.735 km. Finally, the above standard corrections are applied to compute the complete Bouguer anomaly (CBA) values for the CMER based on schemes outlined in the theoretical part of the thesis (Chapter 3, Section 3.1.8).

The results of the survey data include the free-air anomaly map, the complete Bouguer anomaly map and its derivative maps (residual anomaly map (Fig 5.5), regional anomaly map (Fig 5.4), isostatic regional anomaly map (Fig 5.6), isostatic residual anomaly map (Fig 5.7) are considered in Chapter 5. The computations are performed using Geosoft Oasis Montaj (version 6.4.2) standard gravity and magnetic data processing and mapping software.

4.2.2 Gravity Data Gridding and Contouring

After all the necessary data reductions and corrections have been done the reduced gravity data was gridded using Geosoft Oasis Montaj Software. Gridding (a process of spatial interpolation) takes point data and interpolates to determine the values at the nodes of a grid in between the data points. Bi-directional Line gridding, which is best suited to rapidly interpolate data that is collected along roughly parallel lines, is used to grid the airborne gravity data. The different final gridded maps produced are presented in Chapter 5. Moreover, contouring (a process of drawing lines on a map at specified intervals or multiples of intervals, which are derived from a gridded data set) is also applied when necessary to the different gravity anomaly maps. The results of the contoured gridded maps are also presented in Chapter 5.

4.2.3 Regional – Residual Gravity Anomaly Separation

Regional – residual separation is a process of isolating the long-wavelength large amplitude smoothly varying regional anomalies, typically caused by deep-seated sources from the short-wavelength low amplitude sharply varying anomalies associated with shallower sources in the upper portion of the basement and within a section.

Since the complete Bouguer anomaly consists effects of both deeper sources (regional anomaly) and shallower sources (residual anomaly), it is necessary to separate the CBA into its regional and residual components. Among the different regional-residual anomaly separation techniques (graphical method, polynomial method, spectral domain filtering_method, upward continuation method, wavelet Analysis) used in the gravity and magnetic exploration, the polynomial trend surface analysis method of regional-residual separation techniques was employed in this thesis work. This technique removes the preferred n^{th} order polynomial trend from the observed gravity anomaly data. In this work, the regional gravity anomaly is estimated by removing the third order polynomial trend from the observed CBA so that the effects of the deep sources will be fully retained.

4.2.4 Gravity Data Enhancement

Data enhancement techniques of potential field are analytical methods that are employed for highlighting the contrasts of gravity anomalies generated by deep-seated and shallow origin anomalous geologic bodies.

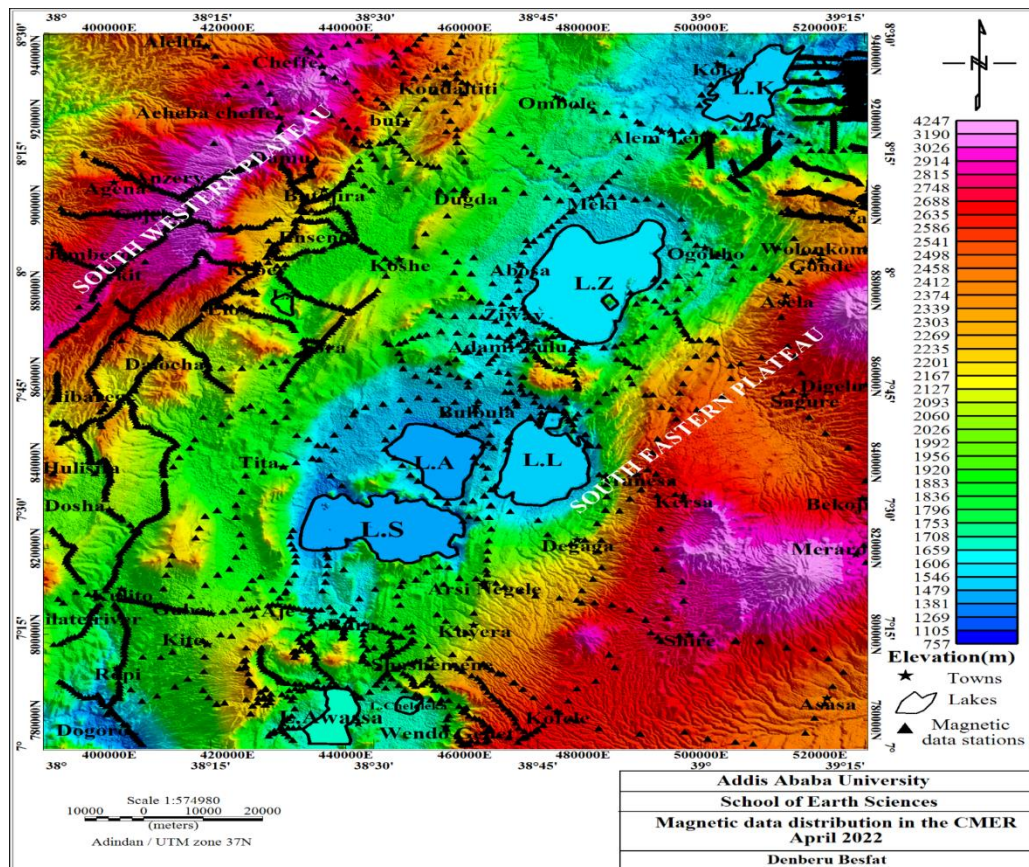
In this MSc thesis, different types of data enhancement techniques are applied to gravity data. These include total horizontal derivative method, analytical signal method, tilt derivative method, source edge detection, and Euler 3-D Deconvolution. The presentations and interpretation results of the gravity data enhancement techniques are considered in chapter 5 (Sec 5.2).

4.3 Magnetic Data Acquisition and Distribution

The magnetic survey was conducted using GSM-19T proton precession magnetometer. It is a standard instrument for many magnetic exploration works and measures the earth's total magnetic intensity. The instrument has a resolution and absolute accuracy of 0.01 and 0.02 nT respectively over its full range temperature. Total field magnetic data was collected following any accessible roads in the study area with a station spacing ranging from 2 km-5 km depending on the exposed outcrops. I.e. when the observed lithology is the same along the traverse line the station spacing was increased up to 5km and when there are frequent variations in lithology then station spacing was reduced to 2 km. The study area is the Central Main Ethiopian Rift (CMER) that covers large amounts of area; hence, attentions have been given to take measurements in all accessible roads including the most highly elevated Guraghe Mountains in which no magnetic data have been previously collected. A number of magnetic Base stations were established in the study area to correct diurnal variations of the raw magnetic data. A total of 1908 primary magnetic data with 0.5 km station spacing previously collected by the principal investigator were merged with 1601 secondary magnetic data collected by previous MSc students of AAU. A total of 3509 magnetic data points is reduced and corrected for diurnal and secular variations of the Earth's magnetic field. The final Total magnetic anomaly data was prepared for producing different 2-D anomaly maps and it was also structured for the 3-D Magnetic inversion work.



Figure.4.2 Magnetic Data acquisition: Left - around Corbeti caldera (exposed outcrop- folded volcanic obsidian). Right - around Zebider Mountains.



4.4 Magnetic Data Processing

4.4.1 Magnetic Data Reduction

As compared to the reduction of gravity data, magnetic data requires very few corrections. These are diurnal correction and IGRF removal correction. The necessary corrections to be performed on the raw magnetic data were discussed on Chapter 3. Section 3.2.6.

The latitude (geomagnetic) or IGRF removal correction was computed using International Geomagnetic Reference Field (IGRF), 2015 data, updated every 5 years, which defines the theoretical undisturbed magnetic field at any point of the earth's surface.

Finally, the total magnetic anomaly (ΔB) corresponding to each observation station is computed by the formula,

$$\Delta B = B_i - \delta B_D - IGRF \quad 4.1$$

B_i is the observed TMI at each station, δB_D is the diurnal correction and IGRF is the International Geomagnetic Reference Field value defined by its latitude. The presentation and interpretation of the computed anomalies is outlined in Chapter 5.

4.4.2 Magnetic Data Gridding and Contouring

After all the necessary data corrections have been done the final computed total magnetic anomaly ($TMA = \Delta B$) data is gridded using minimum curvature technique using Geosoft Oasis Montaj software. Following this gridding technique, different types of maps derived from the TMA including the regional TMA, residual TMA, RTE TMA, are produced. Presentation and interpretation of the TMA map and its derivative maps is outlined in chapter 5.

4.4.3 Regional - Residual Magnetic Anomaly Separation

Like the complete Bouguer gravity anomaly, the total magnetic field anomaly is the total effect of both shallow (short wavelength) and regional (long wavelength) anomaly sources. Therefore, with the objective of suppressing, the response from shallow sources and enhance the response of the deeper sources or vice versa, regional - residual anomaly separation is performed. In this MSc thesis, the polynomial trend surface analysis method of regional-residual separation technique was

used to estimate the regional magnetic anomaly component using the Geosoft Oasis Montaj software. This technique requires removing an n^{th} order polynomial trend from the observed magnetic data. The regional magnetic anomaly was obtained by removing the third order polynomial trend from the observed TMA data.

4.4.4. Magnetic Data Enhancement

Data enhancement techniques of potential field are analytical methods that are employed for highlighting the contrasts of gravity or magnetic anomalies generated by deep-seated and shallow origin anomalous geologic bodies.

In this MSc thesis, different types of data enhancement techniques are applied to the magnetic field data. These include total horizontal derivative method, analytical signal method, tilt derivative method, source edge detection, and Euler 3-D Deconvolution. The presentations and interpretations results of the magnetic data enhancement techniques are considered in chapter 5.

CHAPTER FIVE

DATA PRESENTATION AND INTERPRETATION

5.1 Gravity Data Presentation and Interpretation

Interpretation of gravity data in terms of geology involves the art of translating the compiled gravity anomaly maps of the survey area in terms of geology in order to obtain information about nature of the major geologic units and structures that give rise to the major anomalies revealed by the gravity anomaly maps of the study area. Understanding of the surface geology, the tectonic setting and dimensions of the survey area are important constraints for making reliable interpretations of the compiled gravity anomaly maps in terms of the prominent geological, lithological and structural units of the study area. Therefore, the geological map (Fig. 2.7) and the structural map (Fig. 2.8) of the CMER have been used for the interpretation of the various gravity anomaly maps compiled for the study area.

The final results of the gravity survey data (free-air anomaly values and complete Bouguer anomaly values) computed at all the stations considered in the study area are plotted at their specific locations to compile the free-air anomaly map (Fig. 5.2) and the complete Bouguer anomaly map (Fig. 5.3) of the study area using the Geosoft Oasis Montaj.

The complete Bouguer anomaly map has been used to compile the regional gravity anomaly map (Fig. 5.4), the residual gravity anomaly map (Fig. 5.5), the regional isostatic gravity anomaly map (Fig 5.6) in harmony with the procedures outlined in section 4.2.3 for a further enhancement and interpretation in terms of geology of the gravity data.

More over the presentation of gravity data in spatial 3-D dimension is known in providing an irreplaceable wealthy information about the subsurface. With this regard, the 3-D inversion results of gravity data are presented in this MSc thesis (Chapter 6).

5.1.1 Topographic map

The study area (CMER) is bounded by highly elevated Arsi Bale mountain ranges (including Chilalo, Kaka, etc. mountains) to the east and the Guraghe mountain ranges to the west. The area is characterized by highly elevated topography. In contrast, the central part of the study area is

dominated by topographically low areas, which constitute the rift floor. There are several volcanic complexes or centers (e.g. Aluto, Corbeti, Gademota) and lakes with relatively high and circular topography within the rift floor. Specifically, the topographic map (Fig. 5.1) which is compiled using SRTM30 data reveals minimum values in the floor of the rift and the maximum elevations at the western and eastern escarpments of the rift. The maximum elevation above m.s.l is found on the Chilalo Mountain south west of Asela, which is about 4247 meters and the minimum value of elevation, is observed around Lake Koka, which is about 757 meters above m.s.l.

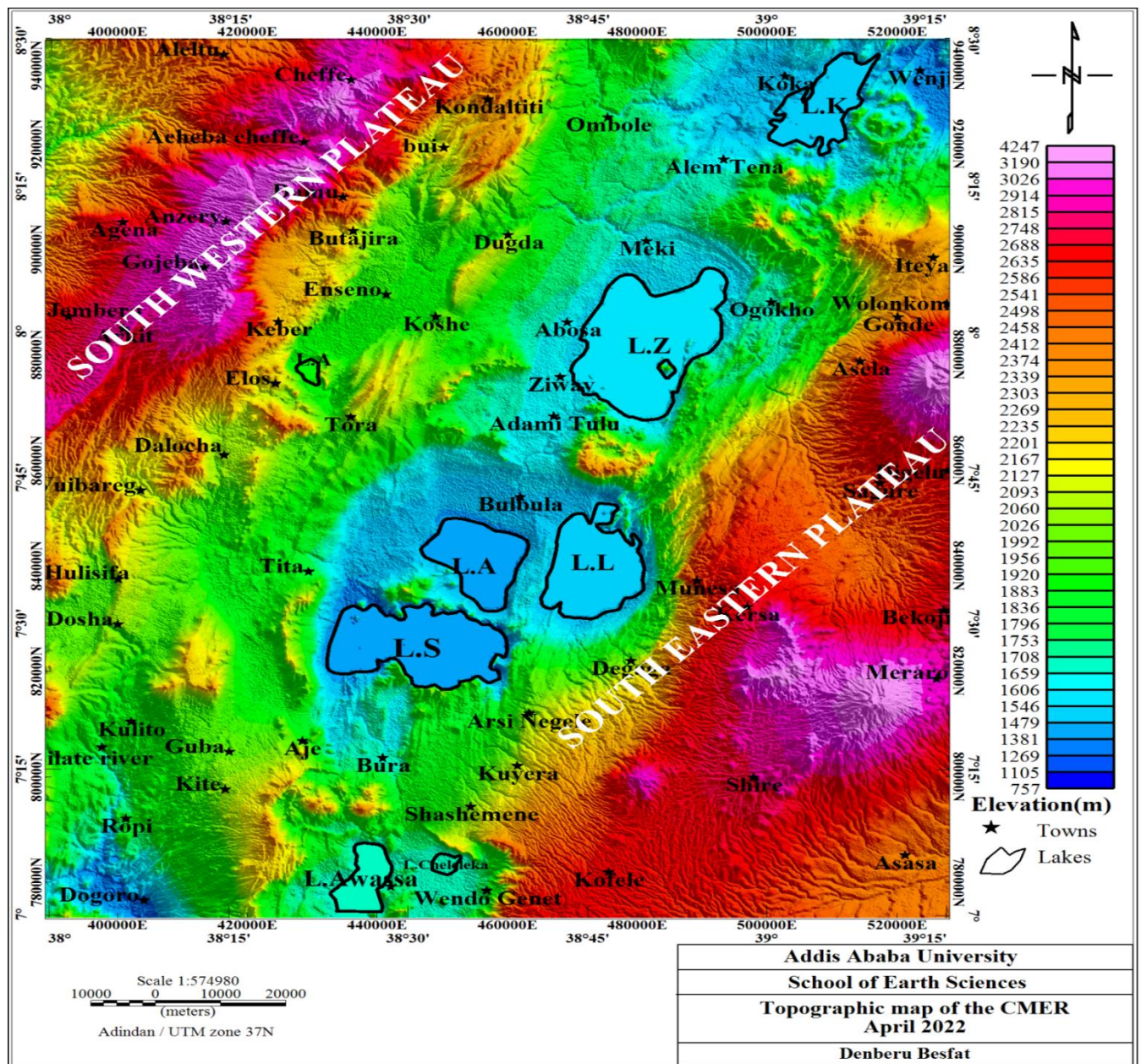


Figure 5.1 Topographic map of the CMER

5.1.2 Free-air anomaly map

The basis of the free-air correction is the inverse square dependence on the distance from the center of mass of the earth to the observation point in Newton's universal law of gravitation. The correction makes an allowance for the decrease in magnitude of gravity with elevation. However, it does not consider the gravitational attraction of landmasses rising above the m.s.l.

The free air anomaly map (Fig. 5.2) compiled from the computed point free air anomalies and the topographic map (Fig. 5.1) compiled from SRTM points are produced using the Geosoft Oasis Montaj software. The anomaly map (Fig. 5.2) reveals a minimum value of about -31 mGal around the rift valley lakes and anomalies greater than 100 mGal on the eastern and western shoulders of plateaus.

The pattern of the major high and low elevation values revealed by the topographic map and the corresponding high and low free air anomaly values revealed by the free-air anomaly map appear to exhibit positive correlation (i.e. as elevation increases the corresponding free air anomaly also increases and vice versa).

The rift components (the SE plateau, the rift floor and the SW plateau) are clearly demarcated on both maps and mark the major structural units (the SE plateau, the rift floor and the SW plateau) of the CMER.

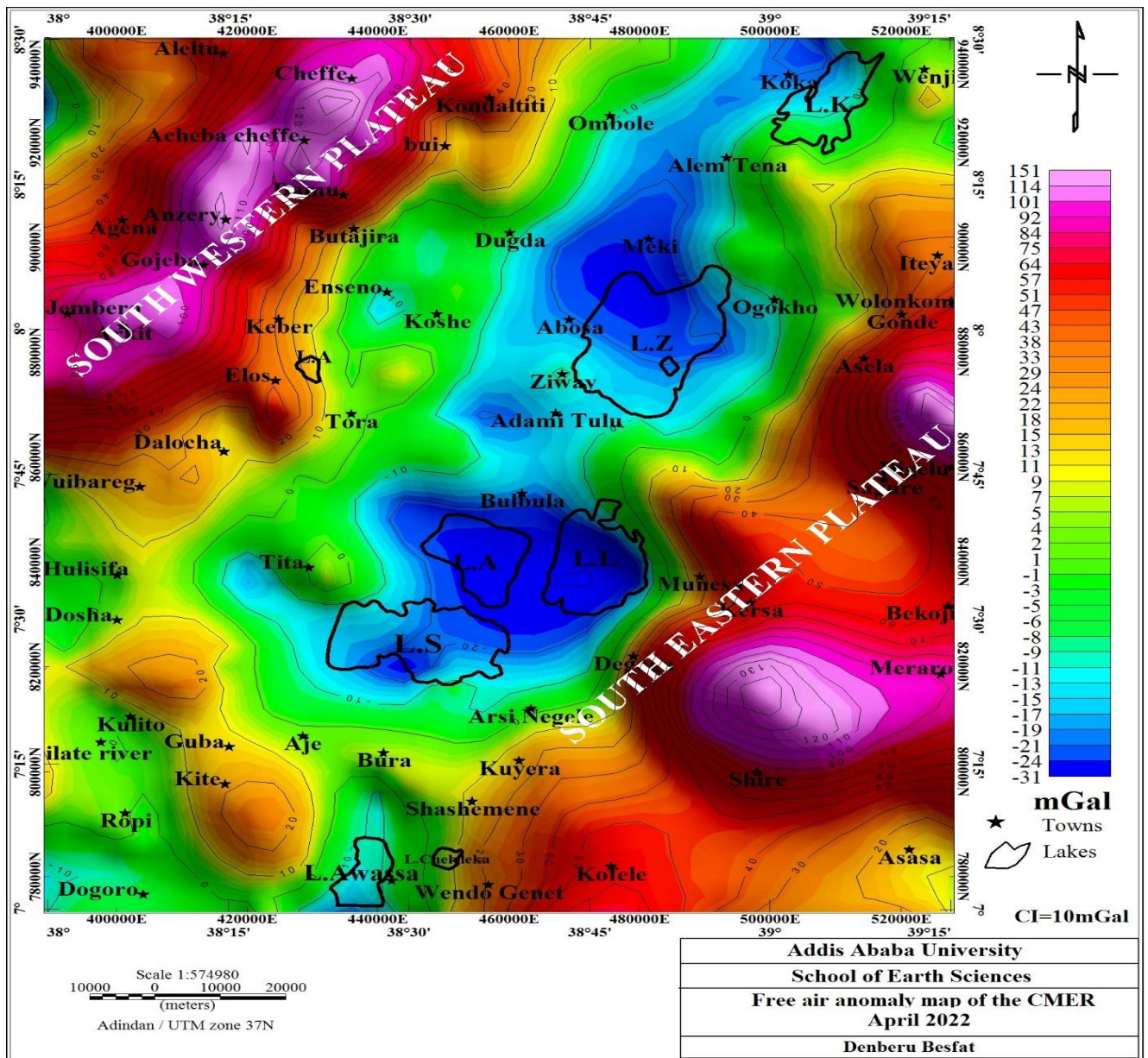


Figure 5.2 Free Air anomaly map of the CMER

5.1.3 Complete Bouguer Gravity Anomaly Map

The complete Bouguer anomaly map compiled for the CMER using the parameters outlined in the data processing part (section 4.2) is given in Figure 5.3. In order to interpret the gross crustal structure of the CMER, the Bouguer gravity anomaly has been contoured with a constant interval of 10 mGal.

The CBA map (Fig 5.3) reveals a minimum value of -291 mGal, a maximum value of -178 mGal and a mean value of -216.5 mGal. The CBA map (Fig 5.3) shows a broad negative anomaly over the SE Plateau with gravity values ranging from -240 mGal to -291 mGal. The gravity values

increase significantly towards the rift floor and reach to a maximum value of -178 mGal near Lake Koka at the northeastern part of the study area.

The rift floor is generally associated with gravity maxima trending in a SW-NE direction. A close examination of the axis of the maximum anomaly in the rift floor suggests that its general trend follows the inferred zone of magmatic intrusions (Corbeti, Shalla, Gademota, Aluto, Bora-Bericha, Tulu Moye and Gudemssa volcanic centers) along the Wonji Fault Belt, forming a patchwork in the north and broadening towards the south.

The maximum Bouguer anomaly (-178 mGal) occurring around Lake Koka and its surrounding could be interpreted as the effect of magmatic intrusions forming the Tulu Moye (TM), Bora (BO), Bericha (BR) and Gudemssa (GD) tectono magmatic centers (Fig 5.3). Geologically this part of the study area is mainly filled with lacustrine sediments and Nazareth series pyroclastic rocks (Fig 2.7). The high gravity anomalies observed above Lake Koka and its surrounding areas are also thought to be partly caused by sediments and intercalated pyroclastic rocks (Fig 2.7) that contain denser material derived from the magmatic intrusions.

Furthermore, the CBA map shows an increase in amplitude when moving northwards following the main protoridge axis of the CMER. This increase in amplitude of the CBA has been attributed to a decrease in depth of the crust mantle interface or an increase in crust/upper mantle density (TIBERI et al, 2005,). This will be further elaborated based on the 3-D gravity inversion results considered in chapter 6.

The CBA is the combined effect of both deeper (low frequency and long wavelength) and shallower (short wave length high frequency) anomaly sources. Therefore, in order to better interpret the gravity anomaly in terms of the effects of deeper and shallow sources independently, it determined that the CBA needs to be decomposed into its regional and residual components.

Generally, the Bouguer anomaly map reveals a broad negative anomaly over the Ethiopian Dome with superimposed short wave length (narrow) positive anomalies over the rift axis (ALEMU A, 1992, GIRDLER et al (1969), GIRDLER and SOWERBUTTS (1970), and BAKER and WOHLBERG 1971).

The broad (long wavelength) negative anomaly observed over the plateau regions are interpreted in terms of an upward thinning of the lithosphere and replacement by anomalously lower density mantle material beneath uplifted regions where as the narrow (short wave length) positive

anomalies occurring over the rift axis are interpreted as being due to intrusive zones in continuity with the lower density mantle material intruding the lower crust and thereby giving a positive density contrast with respect to density of the lower crust.

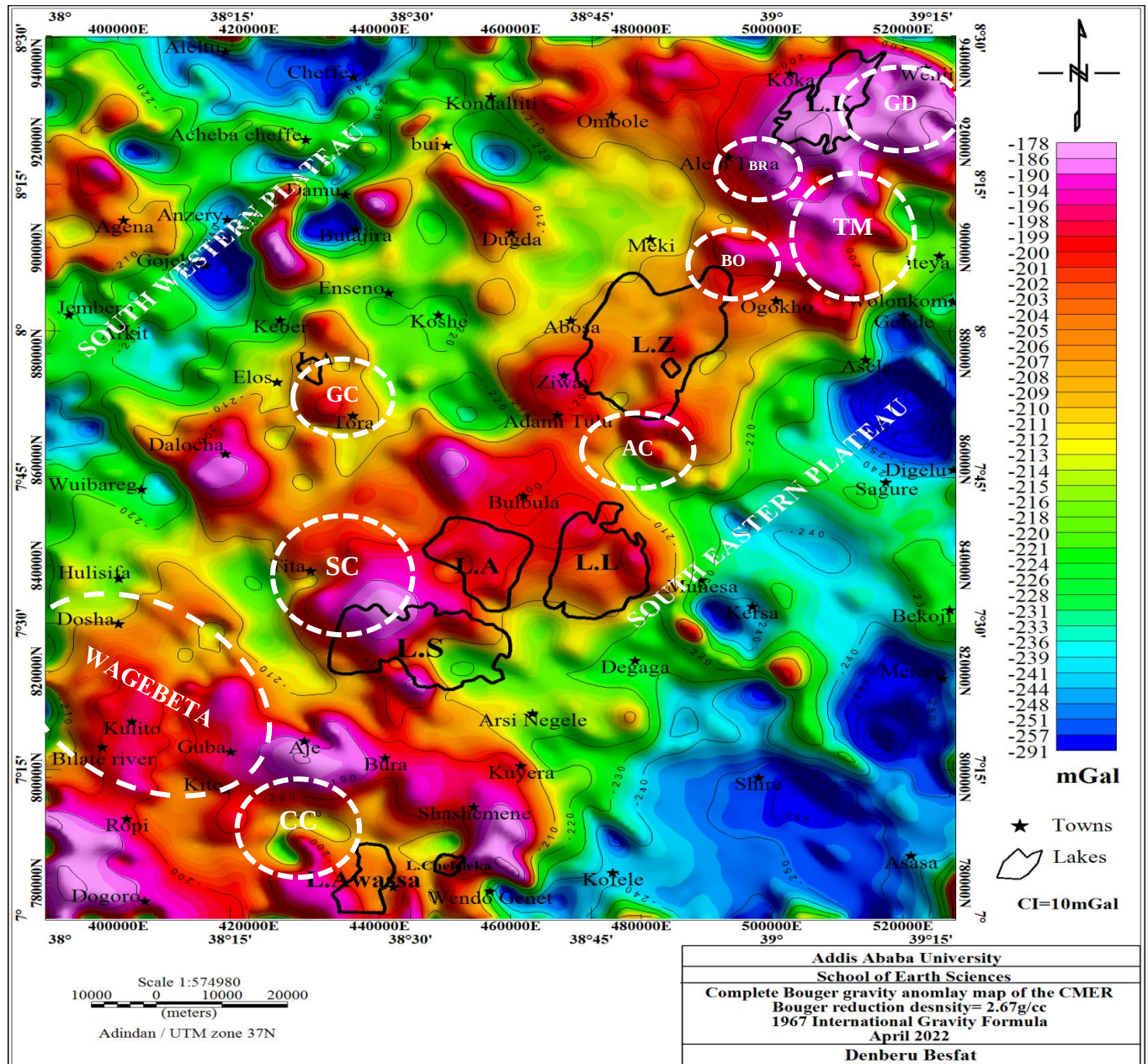


Figure 5.3 Complete Bouguer gravity anomaly map of the CMER. (Labels: CC – Corbeti caldera, GD- Gademota caldera, SC-Shalla caldera, AC-Aluto complex, BO-Bora caldera, BR-Bericha caldera, TM-Tulu Moye caldera, GD-Gedemssa caldera, L.Z –Lake Ziway, LA-Lake Abiyata, LK – Lake Koka, LS-Lake Shalla, LA-Lake Langano).

5.1.4 Regional Gravity Anomaly Map

The regional anomaly map (Fig 5.4) of the study area is compiled by removing the third order polynomial trend from the observed CBA to retain effects of the deep sources. The regional gravity anomaly field has a minimum value of -250 mGal and a maximum value of -191 mGal.

The Regional gravity anomaly map (Fig 5.4) reveals the long wavelength components of the observed gravity field generated by large-scale crustal density variations or undulations of the crust/ mantle in deep subsurface. The minimum anomalies occur over the plateaus and the maximum anomalies occur over rift floor, which appear to indicate a strong correlation with the topography.

As it is clearly seen on the regional gravity anomaly map (Fig 5.4) the orientation of the rift axis, which is dominated by maximum anomalies, appears to follow a SW-NE direction. The maximum regional gravity anomaly value (-191mGal) occur over the locations of Dogoro and Aje in the southwest, Lake Shalla locality in the center and Lake Koka in the NE parts of the study area. Generally, the maximum values of the regional anomaly are concentrated on the main axis of the CMER following the Wonji fault belt. This could be thought to be the effect of denser magmatic materials coming to the upper levels of the crust (result in a thinned crust) beneath the rift floor

Similarly, the minimum regional gravity anomaly value (-250 mGal) occurs over the SE plateau including Asela, Digelu, Bekoji, Shire and Meraro localities of the study area (Fig 5.4). The trend of the minimum anomalies appears to be associated with uplifted platforms of the plateaus bordering the rift floor indicating the existence of a thick crust beneath the SE plateau (Fig 5.4).

A high gradient trend of intermediate regional anomalies occurs over the escarpment bordering the rift floor on the eastern part of the study area. The high gradient indicates a transition from a thick crust over the SE plateau to a thinned crust over the rift floor.

The regional anomaly map (Fig 5.4) also reveals that the northwestern part of the study area corresponding to the SW plateau is dominated by intermediate regional anomalies. These intermediate regional anomalies could be thought to represent a broad transition zone from a thick crust associated with the SW plateau to a thin crust associated with the rift floor.

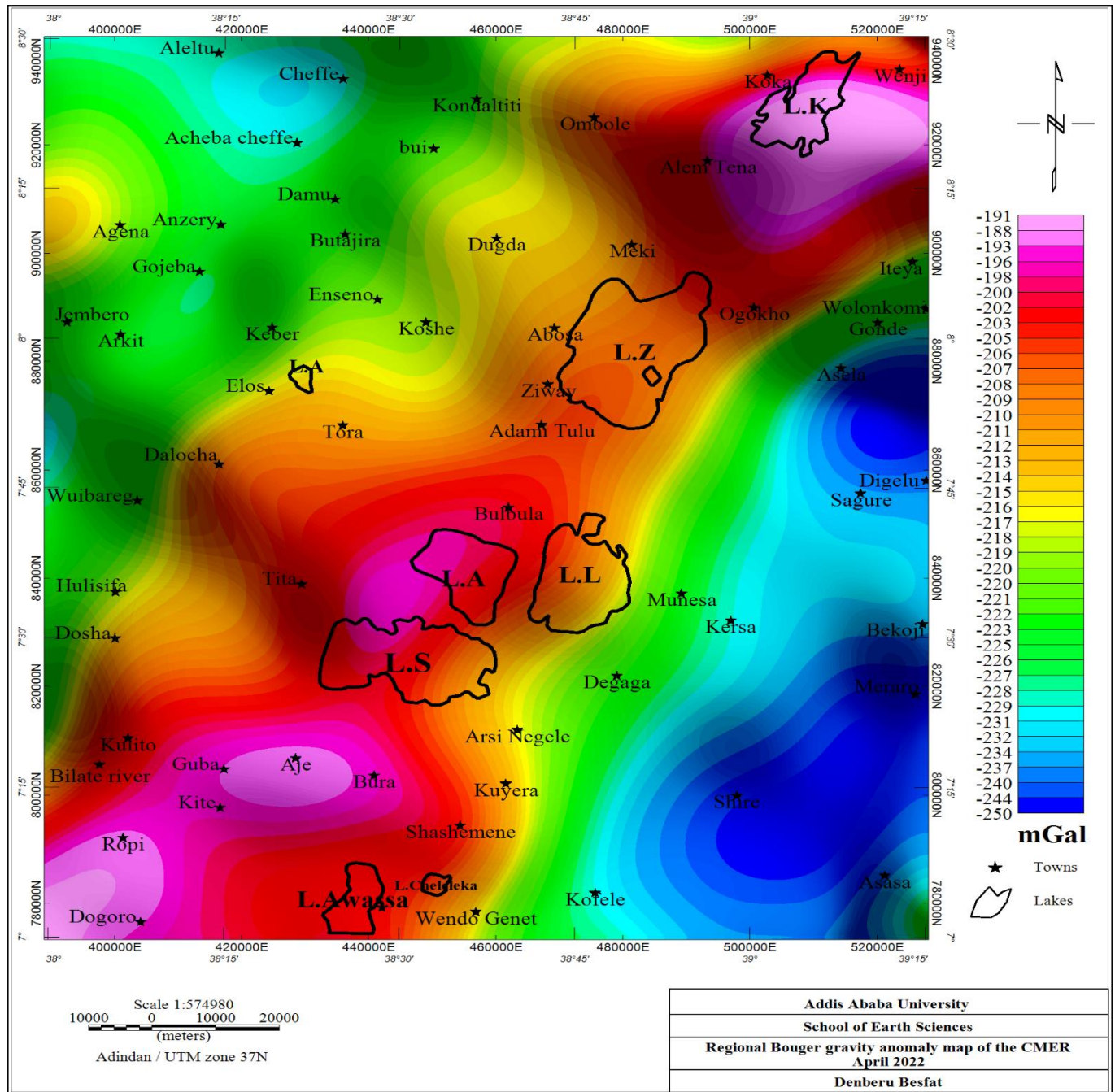


Figure 5.4 Regional Bouguer gravity anomaly map the CMER.

5.1.5 Residual Bouguer Gravity Anomaly Map

The residual Bouguer gravity anomaly (Fig 5.5) is compiled by subtracting the gridded regional anomaly from the gridded complete Bouguer anomaly to enhance the short wavelength (high frequency) components of the gravity signal emanating from shallow and intermediate depths.

The residual short wavelength gravity anomaly is mainly attributed to density contrasts within the shallow subsurface,

The compiled residual Bouguer gravity anomaly map (Fig 5.5) of the CMER has revealed a maximum and minimum values distributed at different locations of the study area. The observed maximum value reaches to 45.2 mGal and the minimum value to -91.5 mGal.

Generally, the maximum values occur over the NW part of the study area dominated by relatively high density pyroclastic rocks of the Nazareth group. The central part of the study area corresponding to the rift floor is dominated by denser quaternary volcanic complexes and the SW part of the study area corresponding to Bilate river basin is dominated by the Wagebeta volcanic complex composed of denser pyroclastic rocks (Fig 5.5 and Fig 2.7). Similarly, minimum residual values also occur over the SE part of the study area corresponding to the SE plateau covered by lighter to intermediate density tertiary basalts, Pliocene trachyte and pyroclastic rocks (Fig 5.5 and Fig 2.7).

The superimposed active faults (Fig 5.5) running along the rift axis following the WFB are associated with maximum residual gravity anomalies generated by denser and active volcanic centers including Corbeti, Shalla, Aluto, Bora, Bericha, Tulu Moye and Gedemssa.

A summary of the specific residual anomalies (high, intermediate and low anomalies) that appeared on the residual anomaly map (Fig 5.5) and their correlation with the geological map (Fig 2.7) of the study area is outlined as follows.

The intermediate to low residual gravity anomalies corresponding to the Munesa-Asela major border faults (MAB) are thought to result from the low-density rocks exposed in the area (Fig 2.7 and 2.8).

The high residual gravity anomalies observed in the southwestern part of the study area are associated with an E-W major geological structure known as the Goba-Bonga lineament.

This E-W trending lineament, which is also characterized by strong seismicity (ASFAW, 1992), cuts the southward continuation of the Guraghe escarpment (Fig 5.5, Fig 2.7 and Fig 2.7).

West of the Corbeti volcanic complex and Lake Hawassa, an elongated intermediate residual gravity anomaly oriented in N-S direction is observed being coincident with the rift controlled Bilate River basin (Fig 5.5 and Fig 2.7).

The high residual anomalies running along the rift floor could be thought as surface manifestations of denser upper-mantle-derived materials (intrusions) mostly associated with the Quaternary peralkaline rhyolite centers of the rift floor. These maxima are observed to dominantly occur along the Wonji Fault Belt and the Silti Debrezeit Fault Zone (Fig 2.7, Fig 2.8).

The patchy high residual anomalies observed at the northeastern corner of the study area are associated with locations of the Bora, Bericha, Tulumoye and Gudemssa magmatic centers that are composed of denser to intermediate pyroclastic, rhyolite and trachyte lava flows (Fig 2.7).

The positive residual anomaly occurring along the Silti (Butajira) - Debrezeit Fault Zone is caused by with the peralkaline rhyolite center of Mt. Duguna.

Generally, the residual highs and low values in observed at different parts of the study area. The high values of residual anomalies can be interpreted to the anomaly source is located at a shallow depth and the lower residual anomalies are regions where the source of the anomaly is deep.

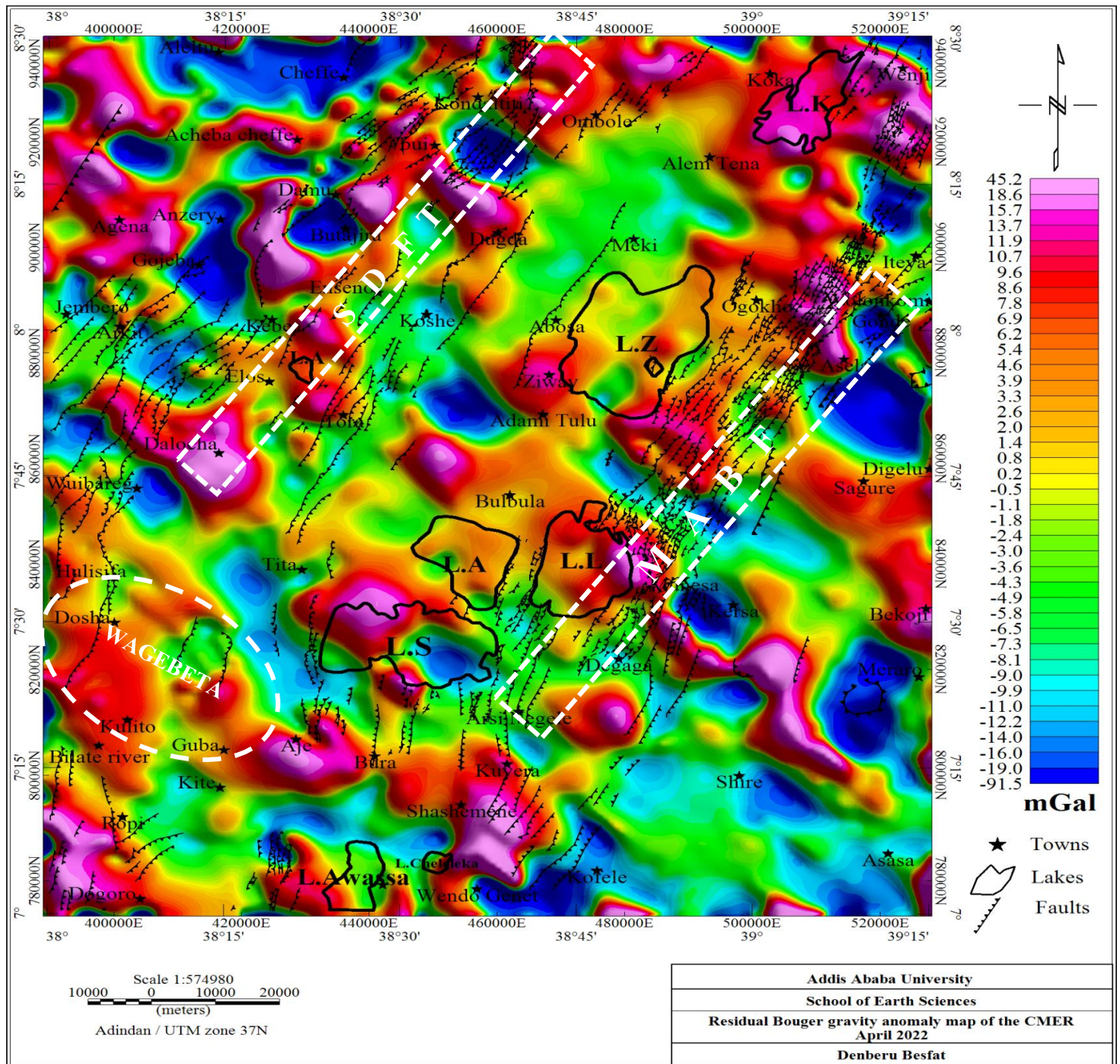


Figure 5.5 Residual Bouguer gravity anomaly map of the CMER (SDFT-Silti Debrezeit Fault Zone, MABF-Munesa Asela Border Fault).

5.1.6 Isostatic Gravity Anomaly

Although the Bouguer reduction process removes the gravitational effect of topography to sea level, the effect of the compensating mass at depth is not accounted for and hence a correlation with topography still exists. To enhance the signal due to intracrustal density contrasts, it is desirable to make a correction to remove this effect. The long-wavelength anomalies are assumed to result from isostatic compensation of topographic loads.

The long wavelengths of the Bouguer gravity field correlate inversely with the long wavelengths of topography. Masses or roots at the base of the crust supporting the topography cause this correlation according to the theory of isostasy.

Isostatic correction is preferable because it accounts for the effects of the low-density topographic (crustal) roots beneath continental areas and high-density mantle material (antiroots) beneath oceanic areas. Continental areas associated mass deficit caused by low-density root materials immersed in the mantle whereas Oceanic areas are associated with mass excess caused by high-density antiroots. The theoretical basis of the isostatic correction including the Isostatic anomaly (IA) is documented in Chapter 3 Section 3.1.8.12 of the thesis.

5.1.6.1 Isostatic Regional Gravity Anomaly Map

In order to study the deep crustal structure of the CMER, the Airy Isostasy model was used to calculate the isostatic gravity anomaly of the study area. The Isostatic anomaly is computed by applying the isostatic correction to the CBA.

The isostatic correction values of the study area are computed by using SRTM30 topographic grid, Airy Isostasy model (model assumes that mountains have a varying crustal root but constant density that compensates for the relief) with a standard depth of compensation 30 km and a Crust-Mantle density contrast of 0.6 g/cc. The Isostatic regional gravity anomaly map (Fig 5.6) of the study area is then compiled by applying the computed isostatic correction values to the Bouguer anomaly map (Fig 5.3) using Geosoft Oasis Montaj (v. 6.4.2).

Since isostasy is a regional phenomenon, the compiled isostatic regional anomaly map (Fig 5.6) is anticipated to reveal the effects of deep sources (roots and anti-roots). The anomaly map shows maximum value of -177.7 mGal, a mean value of -185.9 mGal and a minimum value of -196.5 mGal of isostatic anomaly filed of the study area.

The minimum values occur over the SE plateau and the maximum values over the rift axis particularly the isostatic maximum peak values occur over the surroundings of Lake Shalla and Lake Abiyata.

The computed isostatic regional gravity anomaly (Fig 5.6) better reveals the existence of density variations in central part of the CMER, the SW plateau and the SE plateau.

Low isostatic anomalies, which are effects of low-density crustal root material embedded in high-density mantle material, occur over the SW and SE plateaus. These low isostatic anomalies could be interpreted as the effects of low anomaly sources (low-density crustal root materials) extending to greater depths. Moreover, this implies the presence of a thick crust beneath the plateaus.

On the other hand, the high isostatic anomalies observed over the rift floor are effects of high anomaly sources (high-density mantle material) intruding into the crust. This implies that the high isostatic anomalies indicate the presence of a thin crust (a crust where a high-density mantle materials or magmatic intrusions have replaced the crust) beneath the rift floor.

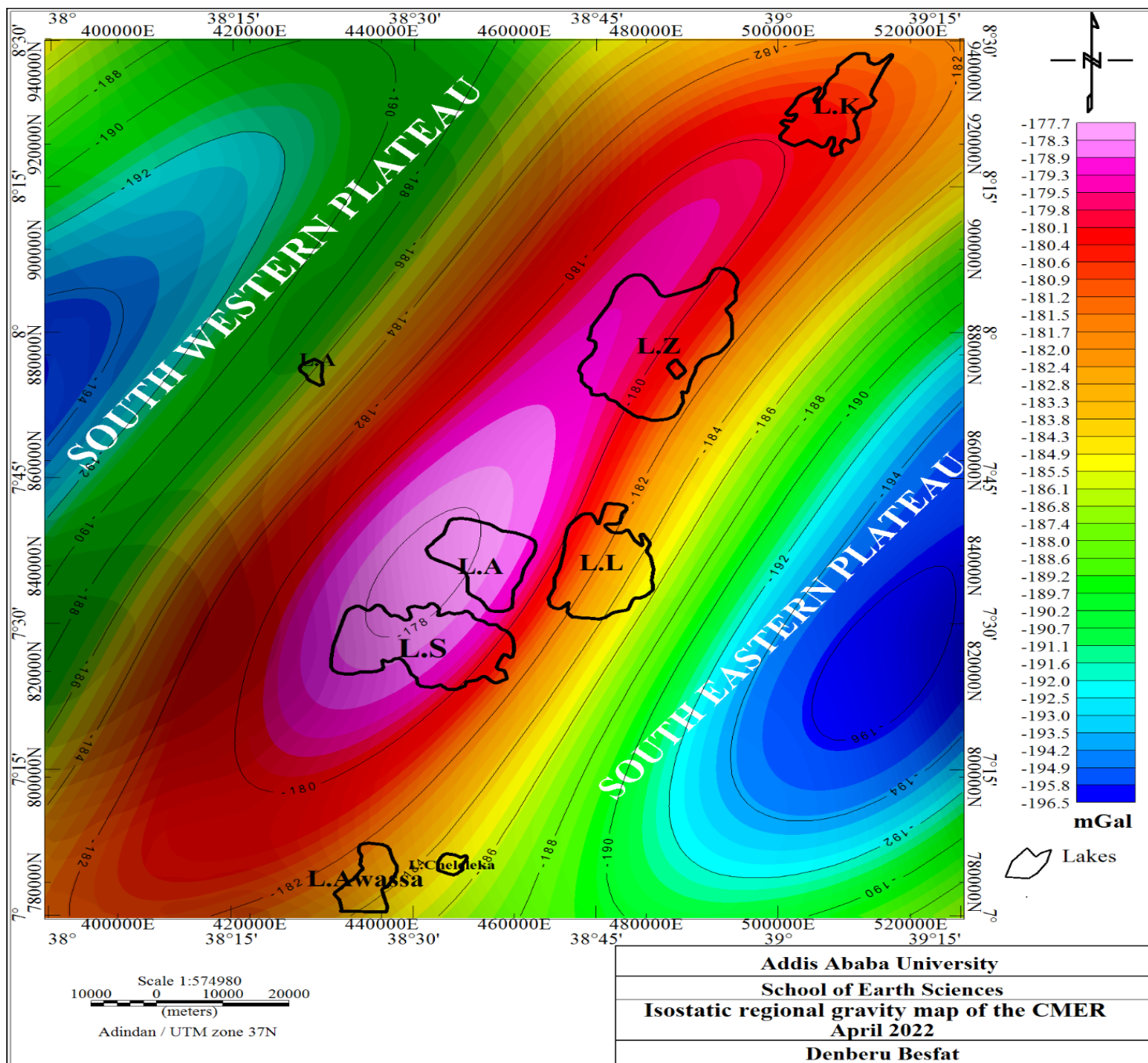


Figure 5.6 Isostatic regional gravity anomaly map of the CMER

5.1.6.2 Isostatic Residual Gravity Anomaly Map

Isostatic residual gravity anomaly maps are produced by subtracting the long-wavelength anomalies produced by masses deep within the crust or mantle from the complete Bouguer anomaly grid. The long-wavelength anomalies are assumed to result from isostatic compensation of topographic loads. Isostatic residual gravity anomaly maps therefore reveal more clearly the density distributions within the upper crust that are of interest in many geologic and tectonic studies than Bouguer gravity anomaly maps.

The Isostatic residual gravity anomaly map (Fig 5.7) is computed by subtracting the Isostatic regional gravity anomaly from the complete Bouguer anomaly grid using Geosoft Oasis Montaj software.

The isostatic residual anomaly map (Fig 5.7) closely represents lateral variations in density of the middle and upper Crust. Generally, it is the sum of the anomalous effects due to variations in the crustal density and anomalous effects due to the deeper masses that support these crustal variations.

The isostatic residual gravity anomaly map (Fig 5.7) reveals a maximum 12 mGal over Lake Koka, mean value of -30 mGal and the minimum values -148.9 mGal around the SE plateau.

The isostatic residual gravity anomaly map of CMER reveals the gravity responses after the long wavelength isostatic roots are removed. The isostatic residual anomaly map shows areas of differing crustal densities, areas that are not in isostatic equilibrium and Moho variation that does not correspond with variation in topography.

Generally, isostatic anomaly values are zero over a region where the crust below the area is in isostatic equilibrium (compensated) and negative over a mountain range which was originally in isostatic equilibrium (has formed a root in the mantle) and drained down by erosion faster than the rise or disappearance of the root. That is, there is a relic root (remnant root) of the mountain that is still in the mantle. The relic root having a negative density contrast with respect to the mantle gives rise to the negative isostatic anomaly while the mountain that formed the root has already completely drained down by erosion in relatively short geologic time. This phenomenon is known as over compensation (ALEMU, 2020).

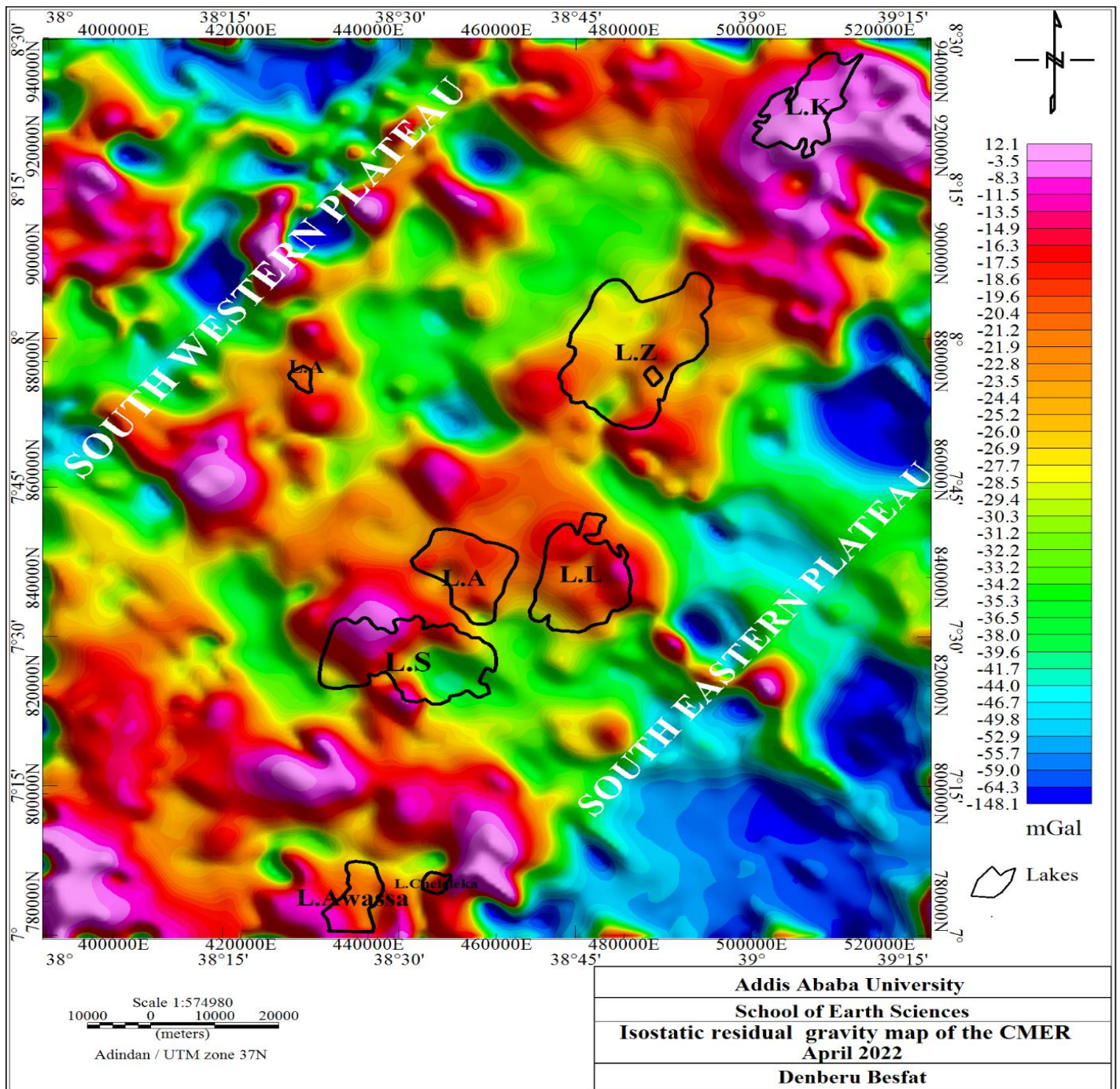


Figure 5.7 Isostatic residual gravity anomaly map of the CMER.

5.1.7 Moho Depth and Moho Undulations in the CMER

Moho depth estimation in the Ethiopian rift system have long been a topic for several Geoscientific researchers. Accordingly, a number of gravity and seismic investigations have been conducted in Ethiopian rift particularly in the MER to determine the depth to the Moho. A summary of the findings of previous gravity and seismic investigations implemented to determine the Moho depths in the MER, SE plateau, MER, Ethiopian Plateau and CMER is provided in Table 5.1.

Table 5.1 Summary of the findings of previous gravity and seismic methods studied in the SE plateau and the CMER for the determination of Moho depth.

Geophysical methods	Moho depth(km)	Study areas	Author/s
Seismic refraction/wide angle reflection	~39–40	SE MER	MACKENZIE <i>et.al</i> (2005)
3-D constrained gravity inversion	~34–40	Ethiopian plateau	MAMMO (2013)
Gravity and topography data	~40	SE MER	TIBER <i>et al.</i> (2005)
High-precision gravity data	~38–51	MER flanks	CORNWELL <i>et al.</i> (2006)
RFs and Rayleigh wave group velocities	~38	CMER	KERANEN <i>et al.</i> (2009)
RFs and Rayleigh wave group velocities	~38–41	Easter shoulder	KERANEN <i>et al.</i> (2009)
A 2-D forward gravity model	~38	CMER	EMISHAW <i>et al.</i> (2017)
Receiver function analysis	~33–44	Ethiopian plateau	DUGDA <i>et al.</i> (2005)
Receiver function analysis	~27–38	MER	DUGDA <i>et al.</i> (2005)
3-D gravity modelling	~30–50	MER & Adjoining plateau	MAHATSENTE <i>et al.</i> (1999)
Gravity and topography data (This MSc Thesis)	~36–44	CMER & Adjacent plateau	BESFAT (2022)

In this MSc Thesis, the Moho depth map (Fig 5.8) of the study area (SW plateau, SE Plateau and the CMER) is computed using high-resolution airborne gravity data and SRTM30m topographic grid data. The values in the topography grid were expressed in meters above sea level.

The isostatic computing algorithm calculates the root-depth grid by assuming Airy Isostasy model and applying a one-dimensional calculation at every grid node.

The Moho-depth map of the central Main Ethiopian Rift is shown in Fig. 5.8. The contour lines represent the crustal thickness distribution beneath the study area. The computed depths to the crust–upper mantle interface resemble the general shape of the isostatic gravity anomaly (Fig. 5.6). The computed Moho depth (Fig 5.8) varies between 38 km (38,000m) and 44 km (44,000m) beneath the SE and SW plateaus and between 36 km (36,000m) to 38 km (38,000m) beneath the rift floor. The thinnest part of the crust (≤ 38 km) which is shown as a morphological low on the contour lines in Fig. 5.8 separates the thickest crust beneath the western and southeastern plateaus. The low trend of the Moho-depth map (~ 36 km) corresponds with the maximum gravity anomaly of the rift floor (Fig. 5.6), whereas the high (~ 44 km) trend of the Moho-depth map, underlying the two adjoining plateaus, corresponds with the minimum gravity anomaly of the rift (Fig. 5.6). The thickest crust is approximately 44 km and is located under the southeastern plateau. Fig. 5.8 shows that the crust–upper mantle interface beneath the two opposite facing escarpments steepens towards the graben, and eventually attains its minimum thickness (~ 36 km) along the rift floor over the Lake Shalla region. The estimated Moho-depth, based on the results of RFs and Rayleigh wave group velocities is 38 km under the CMER (KERANEN et al., 2009). Similarly, the estimated Moho-depth values, based on High-precision gravity data in the MER flanks (CORNWELL et al., 2006) is between 38-51 km. The estimated Moho-depth results of the 3-D gravity modelling 38 km under the CMER (MAHATSENTE et al., 1999) were found to be between 38 and 51 km.

The computed Moho depth values in this Thesis are in a good agreement with the values determined by previous studies (KERANEN et al., 2009, MAHATSENTE et al., 1999, EMISHAW et al., 2017 and DUGDA et al 2005).

However, the Moho-depth results in this MSc thesis need to be considered as holistic approximations. Uncertainties due to inaccurate assumptions in the assigned density contrast, compensation depth used and the topography grid itself could affect and bias the results. Moreover, the depth to the Moho computation is found to be very sensitive to any small changes in the density contrast and compensation depths used. Hence, it should be known that varying these two parameters leads to a different Moho depth result.

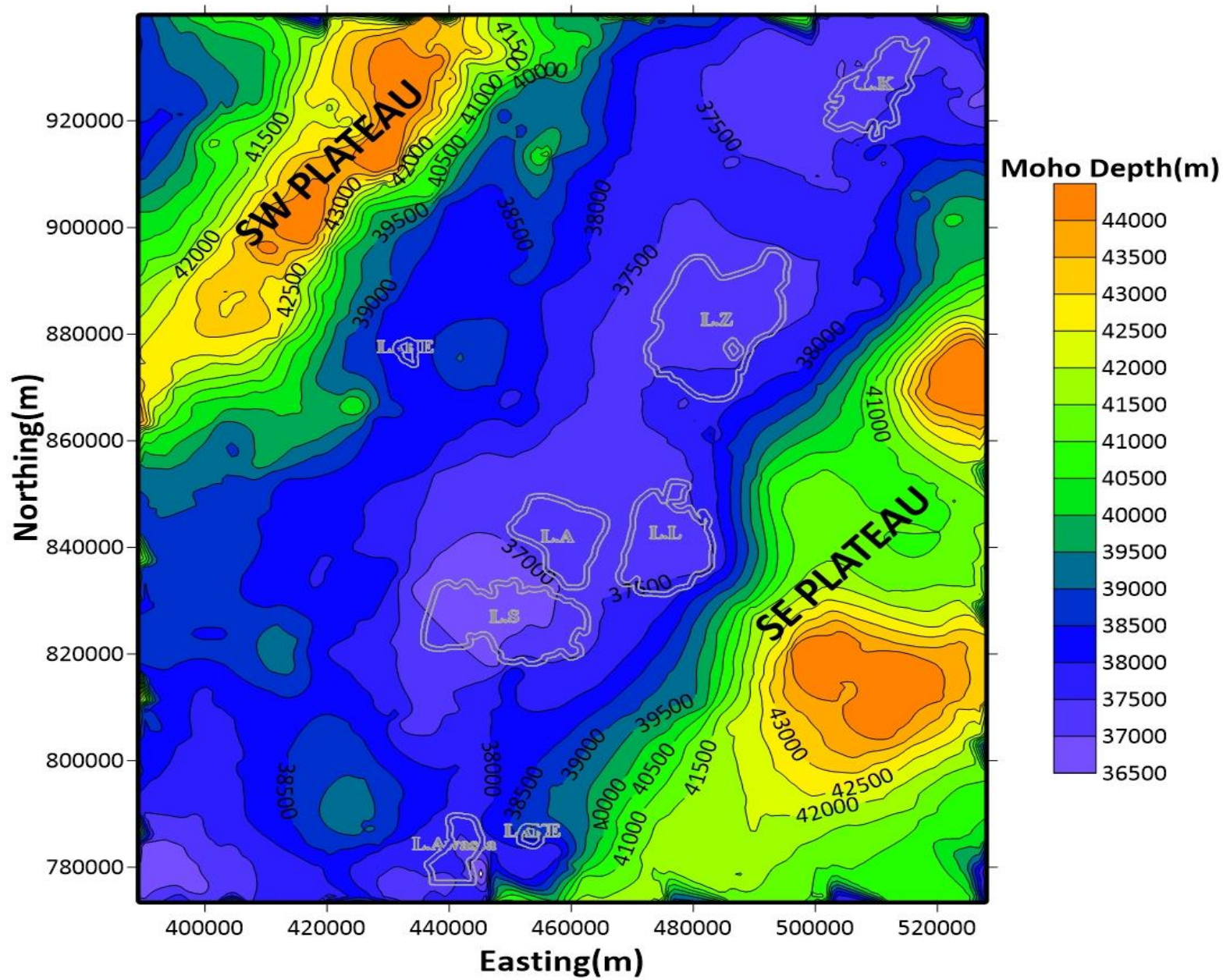


Figure 5.8 Moho depth map of the CMER

Alternatively, the Moho interface undulation $W(k)$ can be mapped using regional isostatic gravity field from the downward continuation formula given by (PARKER, 1973; MAMMO., 2004; MAMMO., 2013., BANKS et al., 1977., FORSYTH, 1985):

$$W(k) = \frac{e^{kz}}{2\pi G(\rho_m - \rho_c)} Q(k)H(k) \quad 5.1$$

where, (k) is undulation of moho interface, G is the universal gravitational constant; ρ_m and ρ_c are densities of the mantle and crust, respectively and $Q(k)H(k)$ is regional isostatic anomaly, Z is average depth to the Moho.

The Figures 5.9 (a) and (b) show the 3-D depth to the Moho and Moho undulation maps of the study area. The Figure 5.9 (a) is displayed here in order to show the true depth to the Moho in the SW and SE plateau regions is greater than the depth to the Moho of the rift floor while Figure 5.9 (b) shows the Moho undulation (crust-mantle interface) compiled by using the methods of (PARKER, 1973., MAMMO., 2004., MAMMO, 2013., BANKS et al., 1977., FORSYTH, 1985).

Accordingly, the compiled Moho undulation map (Fig 5.9b) shows an inverse relationship with the topography. Masses or roots at the base of the crust supporting the topography cause this relationship according to the theory of isostasy.

The relief map shows the existence of mass deficit (thick crust) in the southwestern and southeastern plateaus regions caused by low-density root materials immersed in to the mantle and shows the existence of mass excess (thin crust) beneath the rift caused by high-density antiroots intruding into the crust.

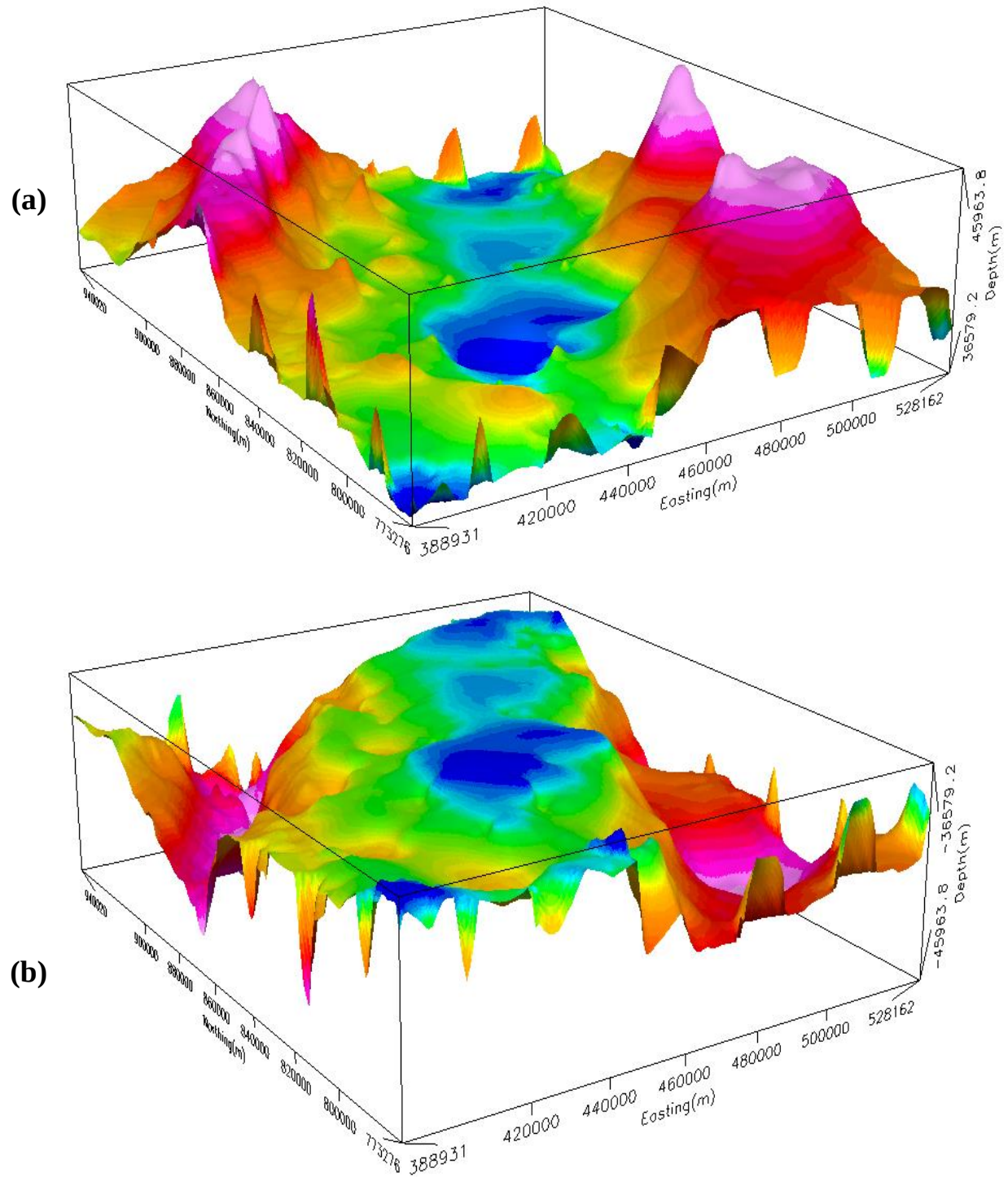


Figure 5.9. (a) 3-D view of the Moho Depth map of the CMER (b) the Moho undulation (crust-mantle interface) map of the CMER.

5.2 Enhancement of Gravity Data

In order to interpret the compiled gravity anomalies in terms of deeper and shallower sources different data enhancement techniques including radially averaged power spectrum (RAPS), Total horizontal derivative (THDR), gravity analytical signal (AS), tilt derivative (TDR), Euler 3-D Deconvolution and source edge detection (SED) are employed to the gravity data. The results of these data enhancement techniques are presented as follows.

5.2.1 Radially Averaged Power Spectrum (RAPS)

SPECTOR and GRANT (1970) have developed the general mathematical basis for the application of power spectrum analysis to potential field data interpretation. They studied the relationship between the power spectrum of potential field anomalies and the average depth of source bodies using some statistical assumption. This provides a foundation for parameter estimation of depth to gravity/magnetic anomalous sources (MAMMO 2011, LAWAL T. O. et al, 2015). The plot of log power spectrum against the radial wave number considered in this study gives results of different spectrum segments. The slopes obtained from the plot are determined to indicate the average depth to the source bodies.

In this study, 2-D fast Fourier transformation (2-DFFT, Oasis Montaj 6.4.2) is applied to the complete Bouguer anomaly grid to calculate the radially averaged energy power spectrum file. The calculated energy spectrum file was logarithmically plotted. As a result, a two-dimensional power spectrum curve is obtained (Fig. 5.10). Based on the appearance of the spectrum, (i.e., change in the slope of the spectrum curve), the slopes of the segments were used to calculate the average depths to anomaly sources (Fig 5.10). The depth of each source ensemble responsible for each segment is calculated by introducing the slope of this segment using the formula

$$h = \frac{-Slope}{4\pi} \quad 5.2$$

Fig (5.10) shows the plot of radially averaged power spectrum versus the local wave number produced using Geosoft Oasis Montaj. Based on the obtained slopes of the plotted segments (Fig 5.10) the average depth estimates for gravity anomalous sources is calculated to amount to be 12.58 km for deeper sources and 6.92 km for shallow sources.

The depth values determined using RAPS analysis indicate that most of the anomaly sources in the study area appear to occur at shallow depth which is estimated to be less than 6.92 km.

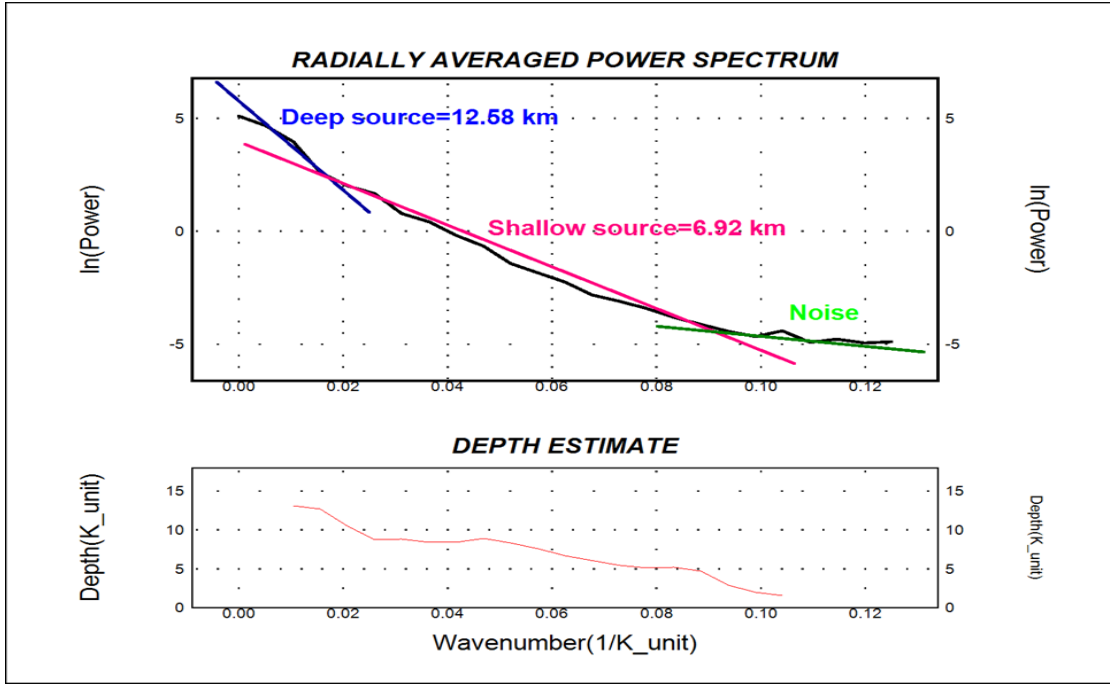


Figure 5.10 Radially Average power spectrum depth estimate plot

5.2.2 Total Horizontal Derivative Gravity Map

According to BLAKLEY, et al (1986), the total horizontal derivative (THDR) or gradient method is used to enhance and locate anomaly source edges in potential field methods. The method has long been used extensively to map the boundaries of density or susceptibility contrasts. It exploits the fact that the horizontal derivative of the residual gravity field generated by a shallow tabular body tends to have maximum values over the edges of the anomalous body in case the edges are vertical and well separated from each other. THDR is not only less sensitive to the noise in the data, but also robust in detection of shallow gravity sources (PHILLIPS, J.D. 2002).

In general, the magnitude of the total horizontal derivative for gravity field can be calculated as:

$$|THDR(x, y)| = \sqrt{\left(\frac{\partial g}{\partial x}\right)^2 + \left(\frac{\partial g}{\partial y}\right)^2} \quad 5.3$$

Where g is the observed gravity field, and $\left(\frac{\partial g}{\partial x}\right)^2$ and $\left(\frac{\partial g}{\partial y}\right)^2$ are the two horizontal derivatives of the observed gravity field in the x and y directions, respectively.

In this study the THDR gravity map (Fig 5.11) of the study area is computed by applying the THDR filter to the gridded residual gravity anomaly map (Fig 5.5) using the source edge detection (SED) algorithm of BLAKLEY, et al (1986) which is an extension in Geosoft Oasis Montaj. The compiled THDR map is anticipated to locate edges of shallow gravity anomaly sources.

The computed THDR gravity map (Fig 5.11) shows high amplitude values over the SW and SE parts of the study area which indicate presence of structural discontinuity (mainly boundary faults) between the rift floor and the flanks of the rift. Moreover, this maximum amplitude is an indication for the presence of strong slopes (i.e. transition from the plateaus to the rift) or maximum lateral density variations.

The amplitude of the THDR along the CMER is relatively minimum. This can be correlated with a decrease in structural discontinuity and lateral variations along the rift axis, which is mainly filled with lacustrine deposits (Fig 2.7). A closer observation on the patterns of THDR amplitudes along the rift axis one can observe differences in density between the volcanic centers and their surroundings.

In order to better, detect the anomaly source edges a peak value of 3 by 4 (grid values in three directions) level of peak detection is applied to the THDR grid using the source edge detection (SED) algorithm of BLAKLEY, et al (1986). A peak detection in three directions usually works best and is recommended in locating anomaly source edges (BLAKLEY, et al, 1986). Subsequently the peak detections applied were automatically generated in a database containing the peak level and peak direction. The generated peak value solutions were then plotted on the THDR map (Fig 5.11) in order to better visualize the detected anomaly source edges. The detected source edges superimposed on the THDR map (Fig 5.12) shows the strike direction of the edge (contact) by the direction of the long axis, and the ‘down-gradient’ direction by the dip indicator.

The plotted SED symbols over the total horizontal derivative grid trace the local peaks. The ‘down-gradient’ direction points away from the source that is locally denser or magnetized.

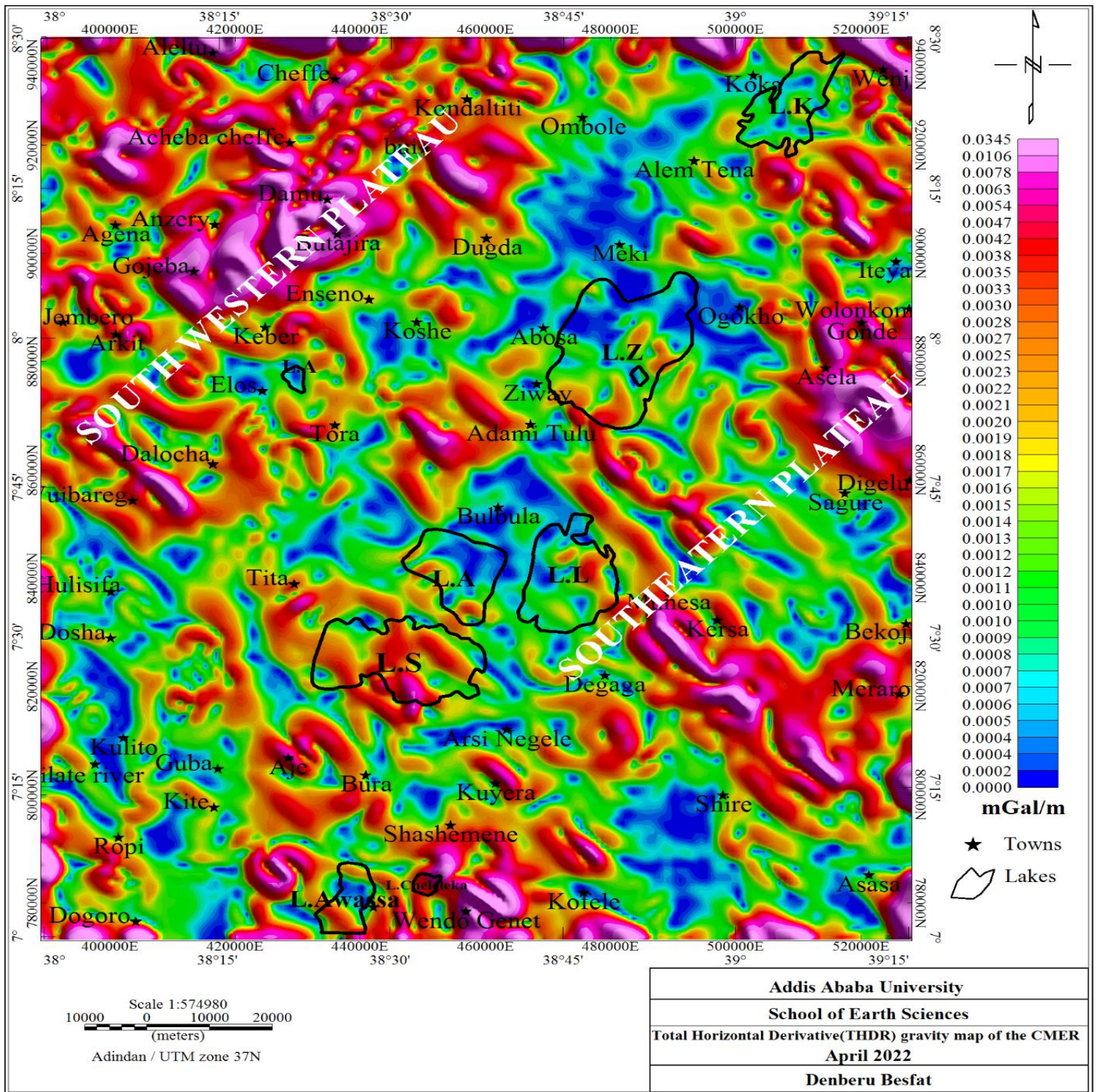


Figure 5.11 Total horizontal derivative gravity map

The SED plot map (Fig 5.12) shows many denser source edges (contacts or faults) detected over the SW, SE plateaus. Moreover, the SED map also clearly identifies the volcanic centers (Corbeti caldera, Gademota caldera, Shalla caldera, Aluto caldera) which are associated with maximum

SED values superimposed on low SED values occurring over most of the rift floor. The detected edges (contacts) traced by the SED method are well correlated with the geological map of the CMER (Fig 2.6). The SW and SE parts of the study area which constitute older pre rift rocks are associated with a number of faults (lateral density variations) traced by SED map (Fig 5.12).

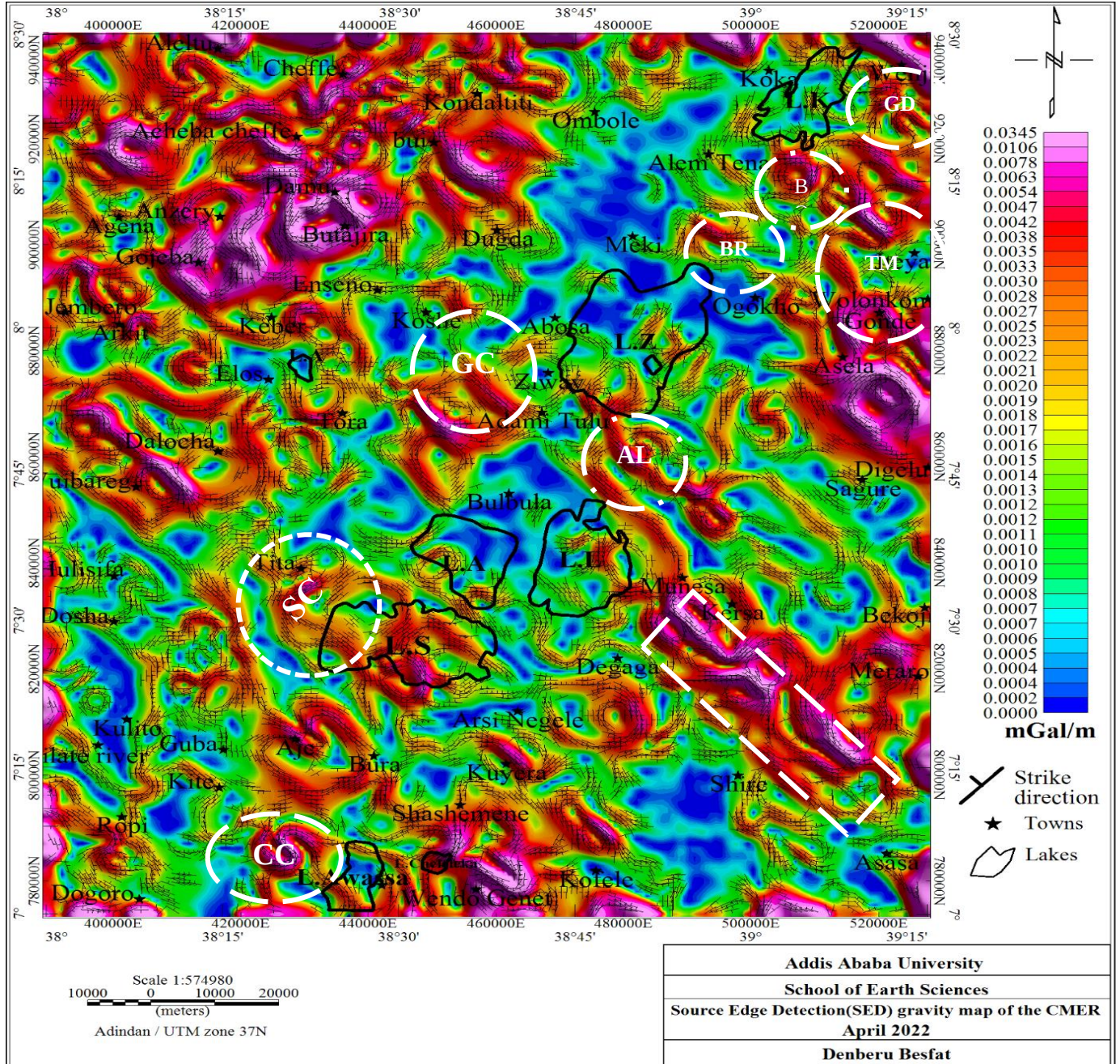


Figure 5.12 Source edge detection (SED) gravity map of the CMER. (Labels: CC- Corbeti caldera, GD- Gademota caldera, AL- Aluto Langano, SC-Shalla Caldera, BR - Bericha caldera, BO-Bora caldera, TM-Tulumoye, GD- Gedemssa caldera).

5.2.3 Gravity Analytical Signal Map

The analytic signal (AS) method is yet another powerful data enhancement technique for delineating geological boundaries (anomaly source geometry) in the subsurface (ROSET, 1992). AS is used for delineation of maxima values of anomalous density distributions in the subsurface. Analytic Signal amplitude values reach their maxima over their corresponding source bodies and, more often, over their edges. Analytic-signal map helps to reveal the anomaly texture and highlight discontinuities by enhancing short-wavelength anomalies. The shallow sources become dominant if there is more than one source. However, the AS acts poorly in enhancing anomalies generated by the existence of shallow and deep bodies at the same time due to its weakness in balancing the amplitudes of different anomalies (COOPER, G.R 2009). The magnitude of a gravity AS is calculated by taking the square root of the sum of square derivatives in all three directions.

$$|AS(x,y)| = \sqrt{\left(\frac{\partial g}{\partial x}\right)^2 + \left(\frac{\partial g}{\partial y}\right)^2 + \left(\frac{\partial g}{\partial z}\right)^2} \quad 5.4$$

where AS (x, y) is the amplitude of the analytic signal at point (x, y), g is the observed gravity at anomaly of preference at point (x, y) and $\partial g/\partial x$, $\partial g/\partial y$ are the two horizontal derivatives and $\partial g/\partial z$ is the first vertical derivative of the observed gravity anomaly.

The analytical signal gravity map of the study area (Fig. 5.13) is generated from the gridded residual anomaly map (Fig. 5.5).

Interpretation of AS map is correlated with the geological map of the CMER (Fig 2.7) in order to better understand boundaries of the shallower anomaly sources.

Accordingly, NW, SE and NE parts of the study are marked by maximum AS values on the analytical map of the study area. In the SW part of the CMER welded pyroclastic rocks, the Gash Megal rhyolite, Tarmaber-Megezez formation, and the tertiary highland basalts in the SE margins of the rift are reflected by maximum gradients in the AS map (Fig 5.13). These maximum AS values can be correlated with the residual gravity anomaly map (Fig 5.5) which shows higher gravity anomaly values over these regions. On the contrary, the lacustrine sediments in the central portion of the geological map are characterized by low values in the AS (Fig. 5.13). The rift floor volcanic centers (Corbeti, Gademota, Aluto Langano, Bora, Bericha, Gademssa and Tulumoye) and off rift regions (Zikuwala and Chilalo volcano) are clearly reflected by circular shaped gradient

maxima in the AS. Hence, the AS can be considered as a powerful technique in enhancing the geometry of shallow seated anomaly sources.

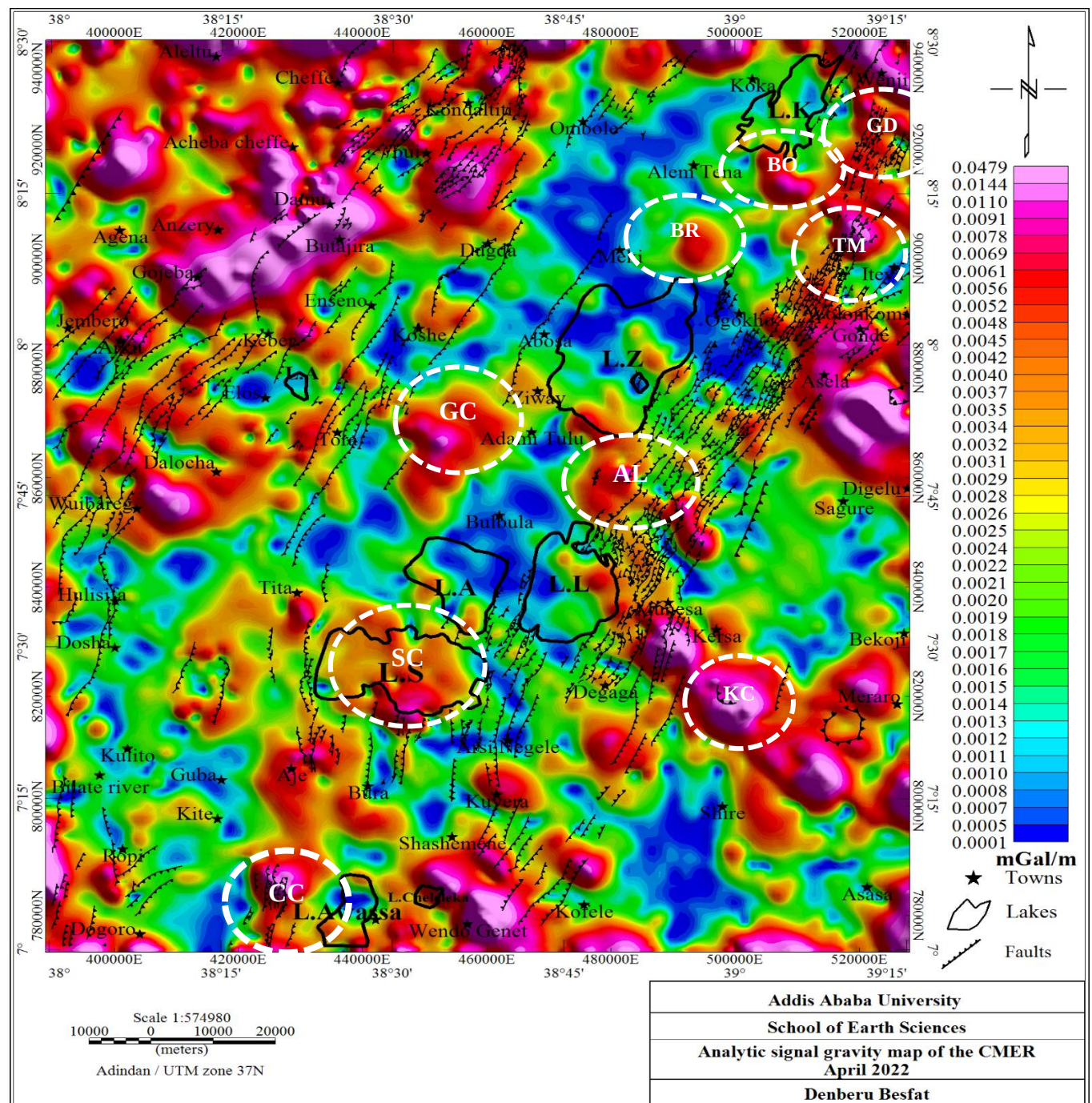


Figure 5.13 Analytical signal map. (Labels: CC- Corbeti caldera, SC- Shalla Caldera, GD- Gademota caldera, AL- Aluto Langano, BR - Bericha caldera, BO-Bora caldera, TM-Tulumoye, GD-Gedemssa caldera, KC-Kaka Caldera).

5.2.4 Tilt Angle (Tilt Derivative) Gravity Map

The tilt angle derivative (TDR) is another edge detection method used for mapping shallow mapping shallow basement structures and mineral exploration targets (MILLER and SINGH, 1994; VERDUZCO et al. (2004). The local phase or tilt angle, or tilt derivative (TDR), is expressed as:

$$TDR = \theta = \tan^{-1} \left[\left(\frac{\partial g}{\partial z} \right) / \left(\frac{\partial g}{\partial h} \right) \right] = \tan^{-1} \left[\frac{VDR}{THDR} \right] \quad 5.5$$

Where, $\frac{\partial g}{\partial x}$, $\frac{\partial g}{\partial y}$, $\frac{\partial g}{\partial z}$, are the derivatives of the observed gravity anomaly of preference in the x, y, z direction respectively.

$\frac{\partial g}{\partial h} = \sqrt{\left(\frac{\partial g}{\partial x} \right)^2 + \left(\frac{\partial g}{\partial y} \right)^2}$ is the total horizontal derivative (THDR) of the observed gravity

anomaly of preference in x, y directions, $\frac{\partial g}{\partial z}$ is the first vertical derivative (VDR). (WILLIAM, et al 2013).

Tilt derivative amplitude variations have a certain range, between $-\pi/2$ and $\pi/2$ according to the nature of the arctangent trigonometric function. Its amplitude has three rates: positive over the source, zero at/near the edge of the source, and negative outside the source (ROSET, 1992). In the presence of noise, this technique acts as an effective signal discriminator for both shallow and intermediate sources but becomes blurred for sources at considerable depths, where it can not reveal deep-level geologic boundaries. Although its efficiency reduces with depth, this method produces relatively sharp anomalies and generates better definition of edges over the body (VERDUZCO, B 2004).

The TDR map (Fig. 5.14) of the study area is produced from the gridded THDR gravity map (Fig 5.11) and first VDR grids using the source edge detection (SED) algorithm. The map is compiled in order to identify the location of faults, contacts, edges and geological boundaries in the study area. Moreover, it is also expected to reveal structures/contacts and boundaries better than the analytic signal map.

From a geological perspective, the study area is characterized by a set of numerous faults oriented along N-S, NE-SW and NW-SE directions (Fig 2.8).

The positive peak values (pink colors) in the TDR map (Fig 5.14) indicate the anomaly source; the zero values (yellow lines) represent edges of the anomaly sources and the negative values (blue colors) represent region outside of the anomaly source.

Source edge detection (SED) symbols (marked by black lines with small hatches) are plotted on the TDR map (Fig 5.14). The plotted symbols show the strike direction of the edge (contact) by the direction of the long axis, and the ‘down-gradient’ direction by the dip indicator. The ‘down-gradient’ direction points away from the source, which is locally denser. The main advantage of the TDR is that, its zero contour lines (zero crossing lines or yellow contour lines) or lie on or close to the fault/contact locations.

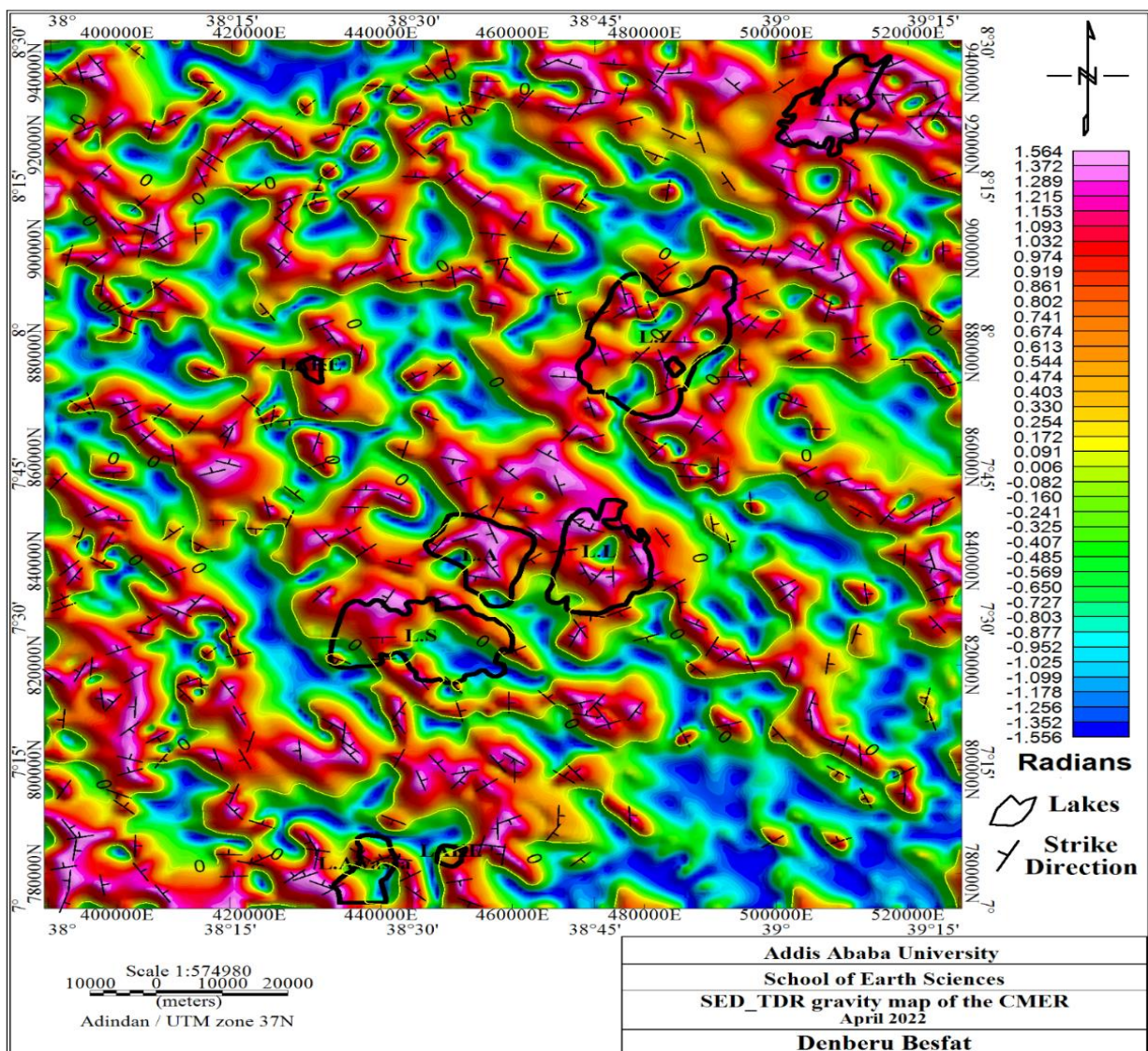


Fig 5.14 Tilt angle derivative (TDR) gravity map of the CMER

5.2.5 Euler 3-D Deconvolution Gravity Map

Euler 3-D Deconvolution (Euler 3-D Modelling) is an automated inverse mathematical modelling algorithm, which is applied to line or gridded gravity data to solve Euler's Homogeneity Equation for source depths and locations of causative bodies of a particular structural type. The method is based on the concept that anomalous magnetic or gravity fields of localized structures are homogeneous functions of the source coordinates and, therefore, satisfy Euler's Homogeneity Equation. Euler 3-D modelling results are proved more effective in mapping structural boundaries such as faults. The standard Euler 3-D equation relates the field (magnetic or gravity) and its components to the location of the source, with the degree of homogeneity N , which may be interpreted as a structural index. (THOMSON, 1982). The structural index is a measure of the rate of change in the field with distance. For a subsurface gravity anomalous source of 3-D nature buried at any point $P (x_0, y_0, z_0)$, the Euler 3-D Deconvolution (Euler's homogeneity equation) at the surface point $Q (x, y, z)$ is given by:

$$(x - x_0) \left(\frac{\partial g}{\partial x} \right) + (y - y_0) \left(\frac{\partial g}{\partial y} \right) + (z - z_0) \left(\frac{\partial g}{\partial z} \right) = N(\Delta g - \beta) \quad 5.6$$

where β =regional gravity value, N =structural index (SI) which is the degree of homogeneity of the anomalous source known as measure of the rate change with distance of a potential field.

Table 5.2. Summary of structural indices for simple models in a magnetic field and for a gravity field.

Geologic model	Number of infinite dimensions	Magnetic SI	Gravity SI
Sphere	0	3	2
Pipe	1(z)	2	1
Horizontal cylinder	1(x-y)	2	1
Dyke	2(z and x-y)	1	0
Sill	2(x and y)	1	0
Contact	3(x, y and z)	0	NA

The standard Euler 3-D Deconvolution plot (Fig.5.15) of the study area is compiled by applying the standard 3-D Euler deconvolution filter to the gridded Bouguer gravity anomaly map (Fig 5.3) using Oasis Montaj 6.4.2. The optimal structural index is selected by a visual inspection of maps compiled using various structural indices until the best clustering of solutions is obtained. Based on this procedure, a structural index of two was determined and applied to compile the Euler 3-D deconvolution map (Fig 5.15) in order to determine the depth, spatial distribution of 3-D anomaly sources (magmatic intrusions) occurring beneath the study area. The map reveals 3-D anomaly sources occurring at different depths and locations marked by colored symbols plotted on the map. These include; a grey color circle indicating gravity sources with depth greater than 12 km, yellow circle gravity sources that range from 10 km to 12 km, the pink circle from 8 km to 10 km, tourmaline green circle representing sources that range from 6 to 8 km depth, blue circles from 4 km to 6 km, the green circles from 2 km to 4 km, red circles from 500 m to 2 km and the small black circle showing depths less than 500 m.

The Euler 3-D symbol plot map reveals that the deepest (10-12 km depth) anomaly sources (marked by yellow colored circles) are randomly exposed (do not follow a specific pattern) over study area.

The anomaly sources occurring at intermediate depths (8-10 km) marked by pink circles are mainly exposed beneath the NE, SE and central part of the study area corresponding to the locations of Chilalo mountains.

Similarly, anomaly sources with a depth range between (6-8 km) marked by tourmaline green depth are revealed beneath SE of Lake Ziway, Lake Langano and Lake Shalla regions.

Generally, the distribution of the sources (shallow, intermediate and deep sources) over the study area revealed by the Euler 3-D deconvolution map appear to be consistent with the depth estimates obtained using RAPS method (Fig 5.10). Moreover, the Euler 3-D Deconvolution map results could be compared and justified to correlate with the locations of the major anomalies (positive and negative) observed on the residual gravity anomaly map (Fig. 5.5).

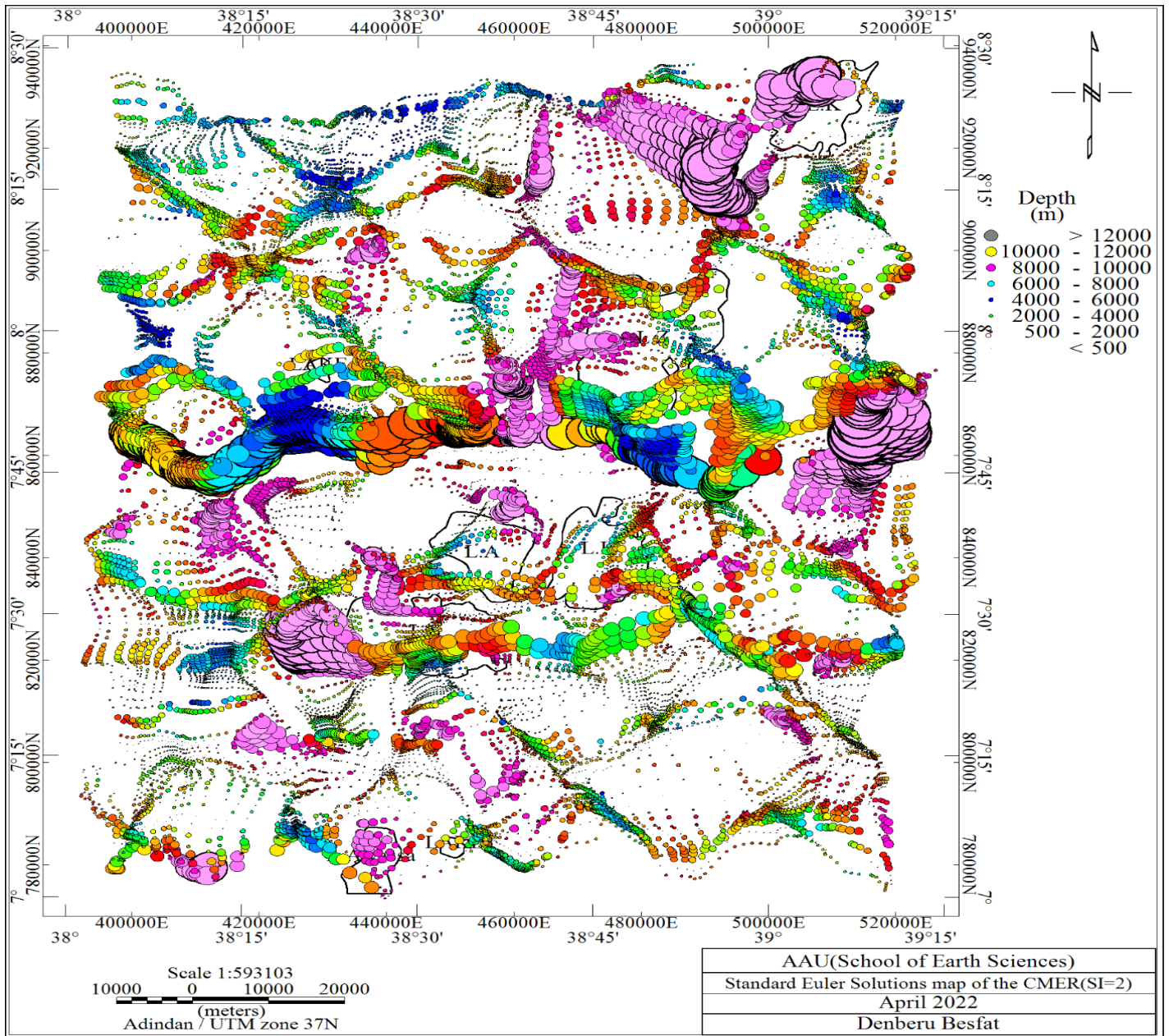


Figure 5.15 Standard Euler 3-D Deconvolution gravity depth plot of the CMER

The Euler 3-D depth solutions were gridded to produce the Euler-3-D depth map (Fig 5.16) map of the study area. Accordingly, the Euler 3-D depth map depicts the depth and location of anomaly sources as displayed by the Euler 3-D depth plots (Fig 5.15) over the survey area. The Euler 3-D depth map attains its maximum values over the regions indicated by red to pink colors and attains its intermediate to low depth values over the rest of the regions.

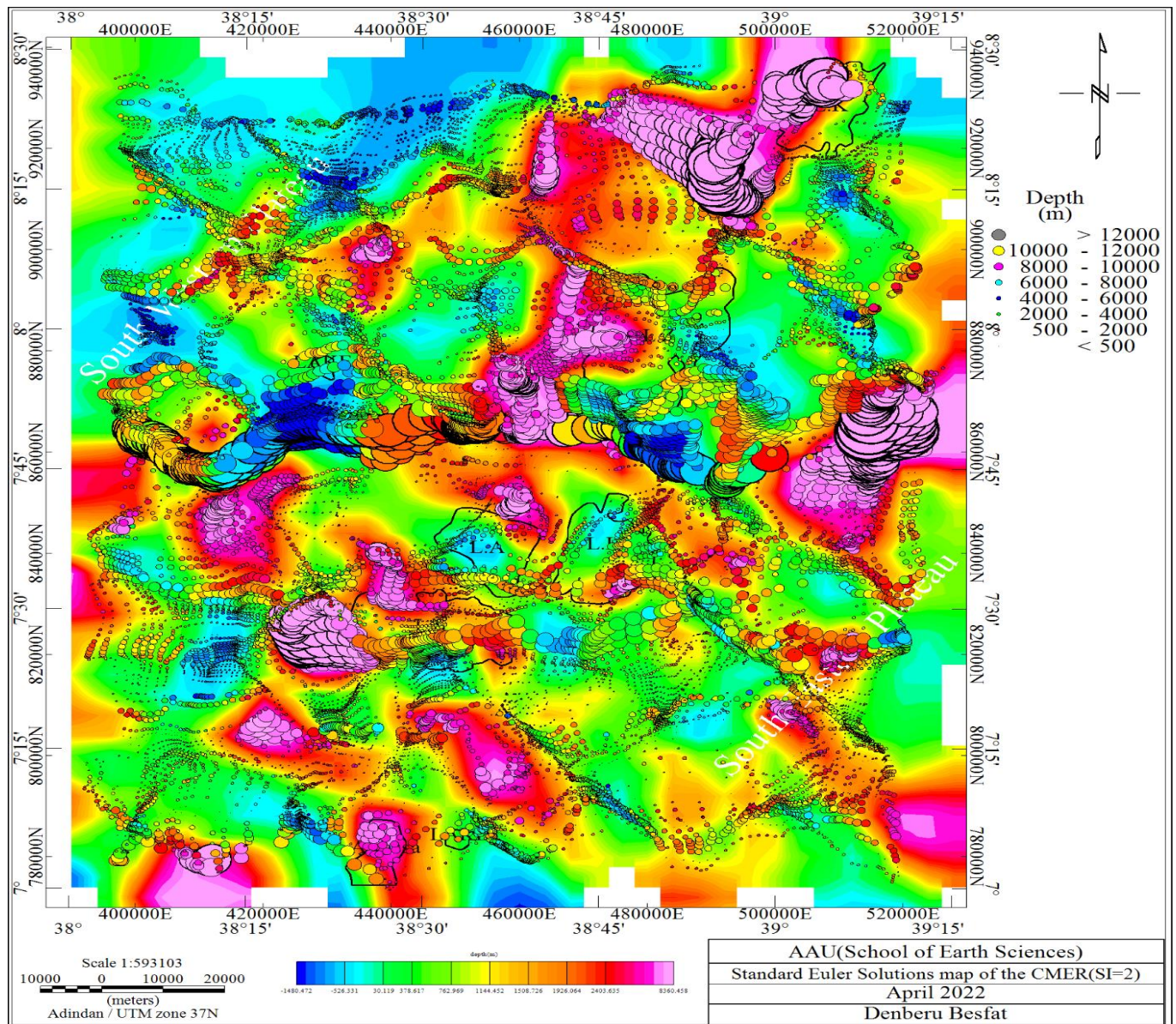


Figure 5.16 Standard Euler 3-D depth solutions superimposed on the Depth map

5.3 Magnetic Data Presentation and Interpretation

Magnetic interpretation, as in the gravity interpretation, operates at several levels of complexity. It can range from simple identification and location of anomalous magnetic bodies in the subsurface to three-dimensional modeling leading to complete characterization of anomaly sources.

The objective of the magnetic survey in this study is to identify and characterize the Earth's crust that is associated with anomalous regions beneath the study area. Consequently, different magnetic anomaly maps that are assumed relevant for this thesis work are produced following the appropriate reduction procedures using Geosoft Oasis Montaj (Version 6.4.2) software.

The final results of the magnetic survey (total magnetic field intensity values and total magnetic field anomaly values) computed at all the stations considered in the study area are plotted at their locations to compile the total magnetic field intensity map (Fig. 5.17) and the total magnetic field anomaly map (Fig. 5.18) of the study area. The total magnetic field anomaly map has been used to compile the regional magnetic anomaly map (Fig. 5.19) and the residual magnetic anomaly map (Fig. 5.20) for a further enhancement and interpretation of the magnetic data in terms of geology. Moreover, presentation of magnetic data in spatial 3-D dimension provides a wealthy information about the subsurface. With this regard, the 3-D inversion results of magnetic data in the form of 3-D susceptibility models are also presented in this MSc thesis work. (Chapter 6).

5.3.1 Total Magnetic Field Intensity Map

The total magnetic field intensity map (Fig.5.17) of the study area is compiled by gridding all the magnetic data points that are corrected for diurnal variation at their respective locations using minimum curvature gridding technique. The map is supposed to show the variation in magnitude of the combined effect rock magnetization and the dipole field over the CMER

The total field intensity map (Fig 5.17) reveals a minimum intensity value of 34348 nT, a maximum intensity value of 36336 nT and a mean intensity value of 35357 nT. The total intensity shows low magnetic intensity (up to 35121nT) over the southern part, intermediate magnetic intensity responses (up to 35324nT) over the central part and high magnetic intensity responses (36336nT) over the northern and southeastern part of the CMER.

The total magnetic intensity map (Fig 5.17) generally shows the regional variations of the earth's dipole field with no significant effect of rock magnetization observed over the study area.

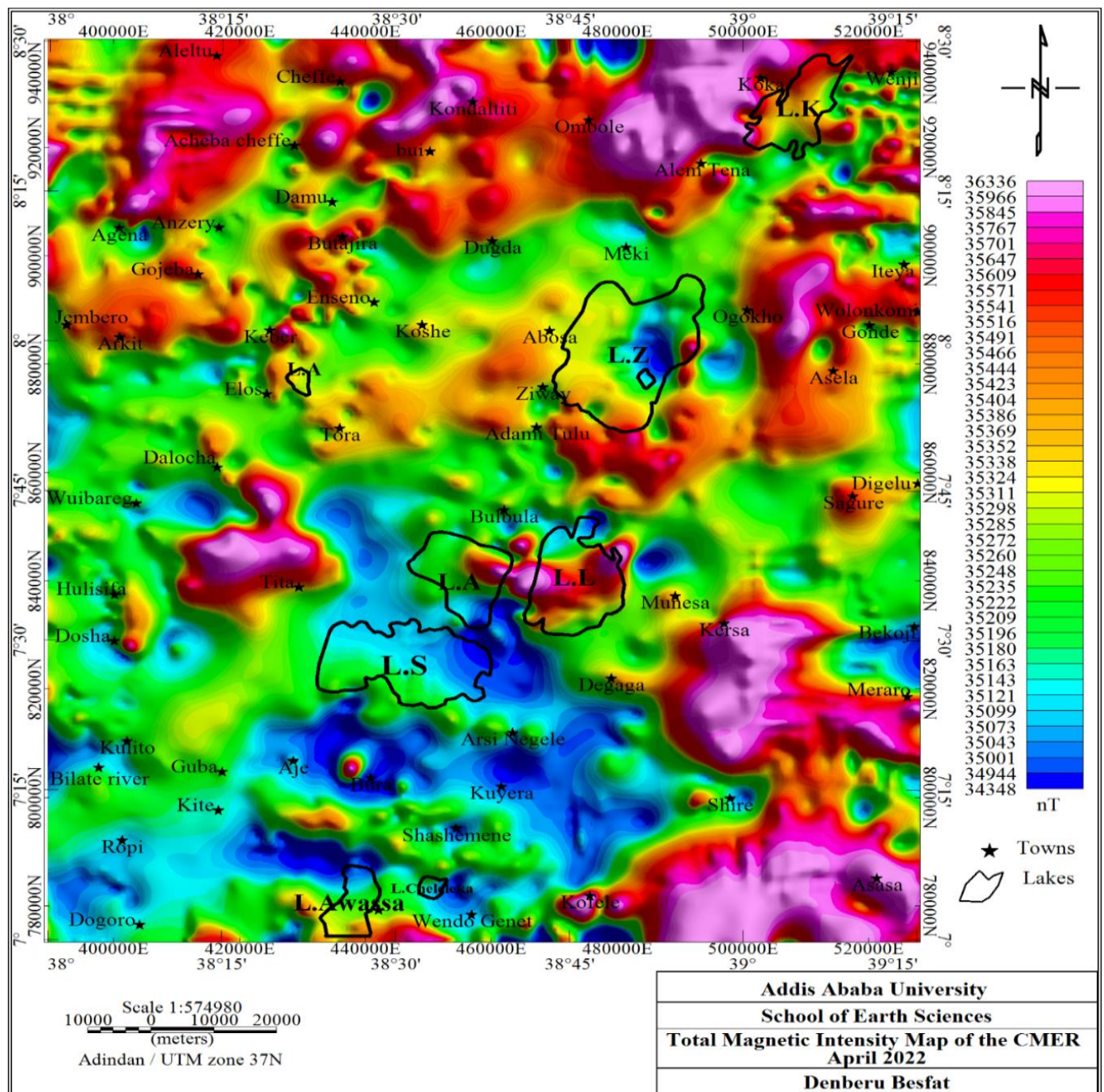


Figure 5.17 Total magnetic intensity map of the CMER

5.3.2 Total Magnetic Anomaly Map

Magnetic anomaly generally is positive when the field of a buried source reinforces the Earth's magnetic field, and it is negative when the source's field opposes the Earth's magnetic field (such as strong Remanent Magnetization) or has insignificant susceptibility values (SERGUEI et al, 2000).

The equator reduced total magnetic anomaly (RTE_TMA) map (Fig 5.18) of the CMER is compiled by removing the dipole field from the total magnetic intensity by applying IGRF correction and reduction to the equator (RTE) filtering using Geosoft Oasis Montaj. RTE filtering is applied to the TMA since it is a stable filter at low latitudes between $+20^{\circ}\text{N}$ - 20°N in opposition to the reduction to pole (RTP) filter (RUDER et al, 2016). The RTE filter helps imaging only the southern and northern boundaries of the TMA anomaly this is because the core field's magnetic lines of forces are trending parallel to the surface of the earth at the magnetic equator and there is no tilt or deep on the vector. However, it can have difficulty in mapping the eastern and western edges of the anomalies (RUDER et al, 2016). At low magnetic latitudes, anomalies over magnetically susceptible bodies show negative values instead of positive and the anomalies largely depend on azimuthal direction (BEARD et al., 2000). Moreover, the study area CMER is located at low latitudes.

The compiled RTE _TMA map (Fig 5.18) shows the responses in rock magnetization of the study area and reflects the occurrence of prominent magnetic lows (as low as -672 nT) and magnetic highs (as high as 1150 nT) occurring at different locations.

High (480 nT) to very high (1150 nT) total magnetic anomalies are observed over the Ombole , Kondaliti, Meraro, Asasa, Kersa, Kofele, and Tita localities and over the Aluto Langano, Adami Tulu, Lake Hawassa, and Gademota volcanic centers. These high to very high total magnetic anomalies are thought to be attributed to either low magnetic susceptibility of magmatic intrusion below the volcanic centers or due to the low magnetic susceptibility of rocks. The high TMA values in the SW plateau region over the Acheba cheffe and Jembero towns are an indication of the presence of weak zones and/ or contacts of rocks with prominent susceptibility contrasts.

The eastern of Lake Koka, Alemtena town, north of Munesa town, Lake Ziway, Lake Shalla and Aje town are characterized by negative magnetic anomalies, which is thought to be attributed to the thick accumulation of sediments. The geological map (Fig.2.7) show these areas are totally

covered by sediments although the thickness of the sediment varies from place to place. As mentioned earlier these low to very low anomalies are possibly due to the high magnetically susceptible sediments that are possibly derived from the neighboring volcanic materials that have high magnetic contents.

Generally, the discontinuities of anomalies over the volcanic centers can be associated with the presence of tectonic structures such as faults, fracture zone or similar structures and the isometric (close to circular) shape of anomalies observed are mostly related to magmatic (intrusive) bodies.

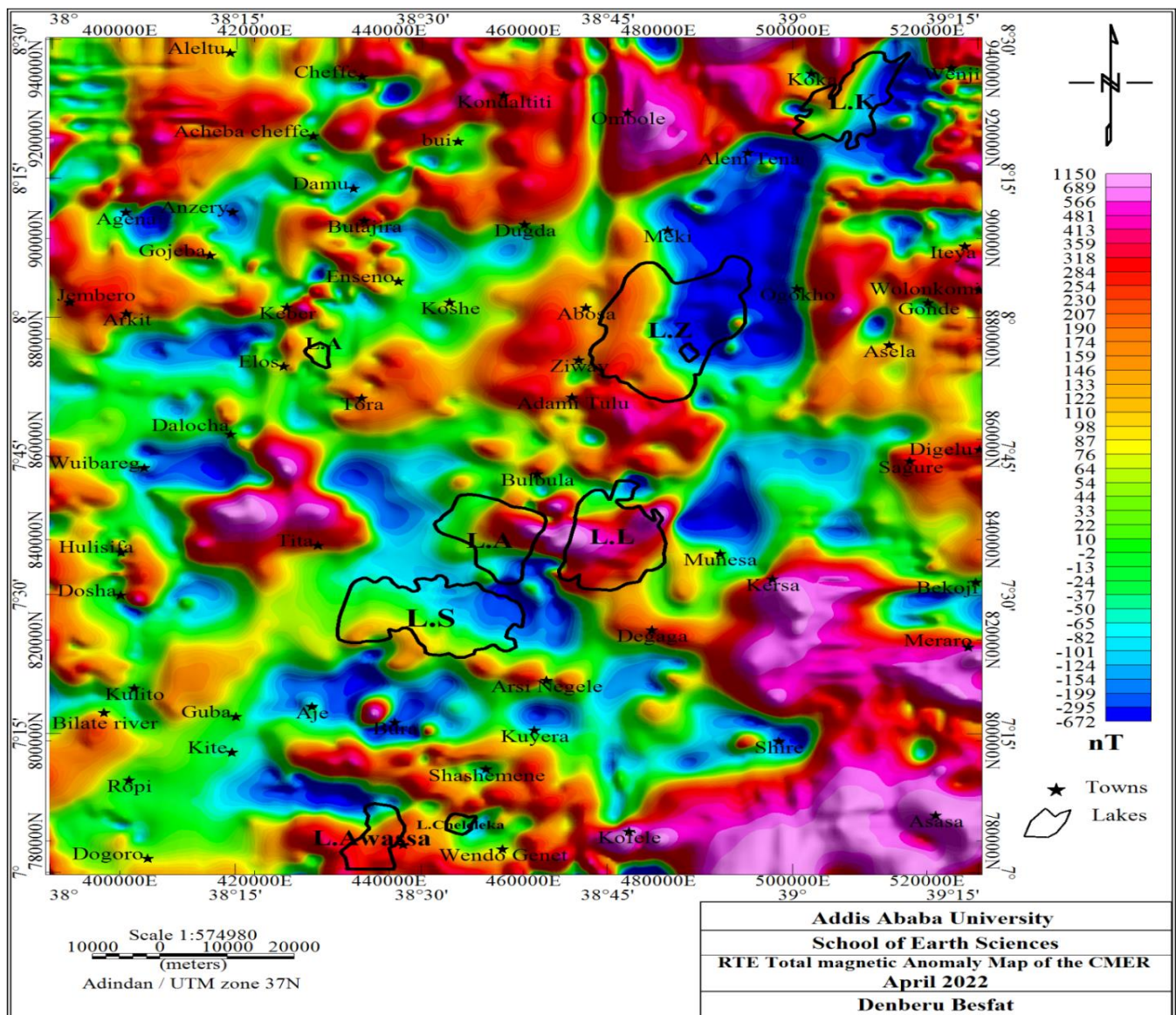


Figure 5.18 RTE_TMA map of the CMER.

5.3.3 Regional Magnetic Field Anomaly Map

In order to provide relevant information associated with the regional (deep seated) and residual (shallow origin) effects of anomaly sources isolating these two components of the RTE_TMA is a crucial procedure.

Accordingly, Polynomial trend surface analysis method is used to separate the regional and residual components of the RTE_TMA anomaly using Geosoft Oasis Montaj. The third order polynomial trend has been removed from the RTE_TMA grid to estimate the regional anomaly and compile the regional magnetic anomaly map (Fig 5.19) of the study area.

The regional magnetic anomaly map (Fig 5.19) shows distribution of the major regional anomalies (maximum and minimum) that are generated by sources occurring at deeper depths beneath the study area.

The maximum anomalies observed over the northern and southeastern parts of the study area are thought to be associated with the occurrence of low susceptibility rock units at deeper depths (possibly below 14 km depth as determined from RAPS, Fig 5.21).

The minimum anomalies are observed over the southern part of Lake Koka, NE part of Lake Ziway, surroundings of Lake Shalla, Dalocha and Aje localities.

These minimum anomalies are thought to be attributed to the presence of high susceptibility rock units occurring at deeper levels (possibly below 14 km depth as determined from the RAPS, Fig 5.21).

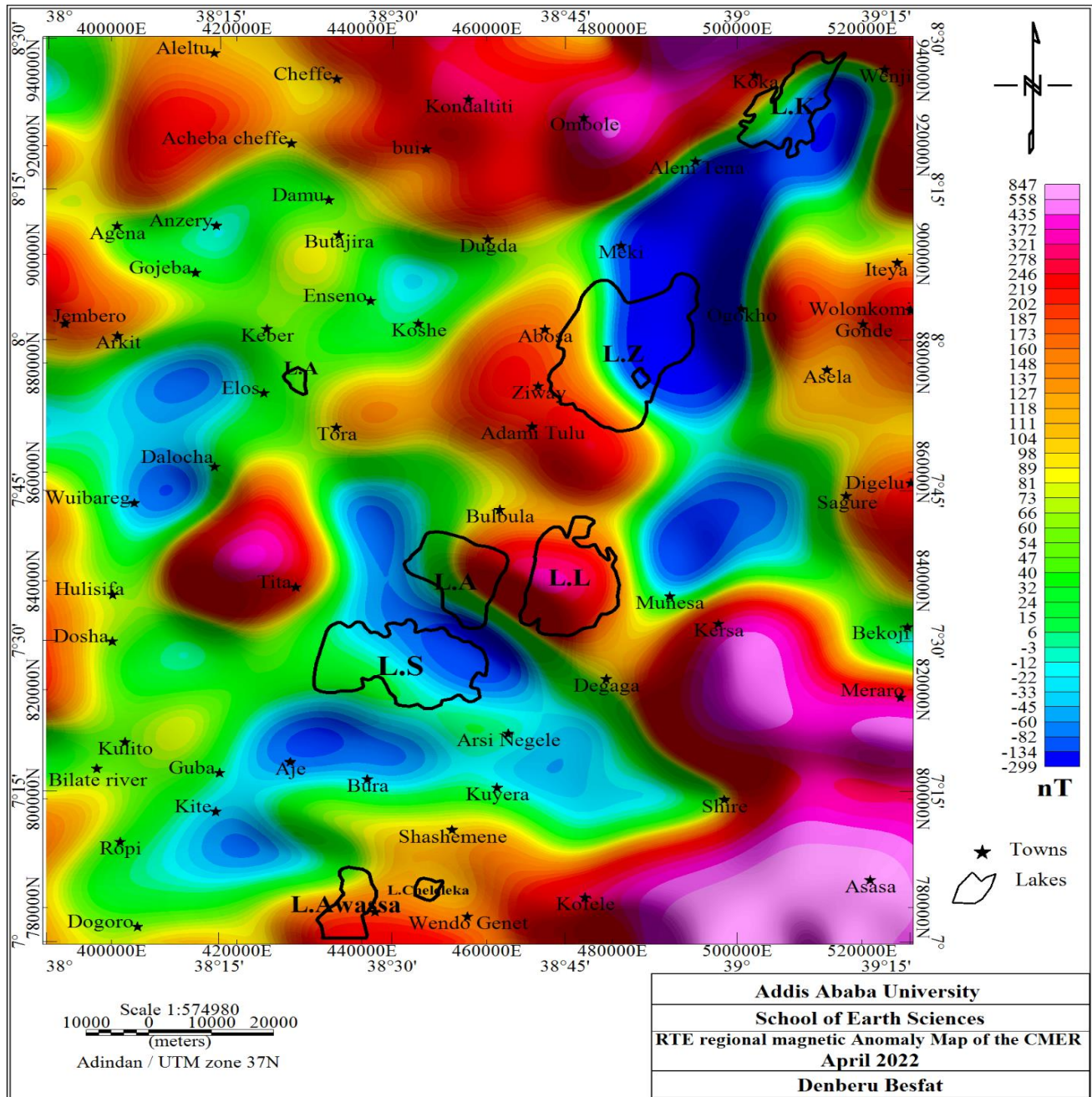


Figure 5.19 Regional magnetic anomaly map of the CMER

5.3.4 Residual Magnetic Field Anomaly Map

The residual magnetic anomaly map (Fig 2.20) of the study area is compiled by subtracting the regional magnetic anomaly grid from the RTE_TMA grid using Geosoft Oasis Montaj Software. The map is anticipated to enhance short wavelength (high frequency) anomalies that are generated by magnetic sources occurring at shallower and intermediate depths. The intermediate depth

amounts to 4.29 km and the shallower depth amounts to 1.35 km as determined using the RAPS method.

The residual anomaly map (Fig 5.20) generally shows the presence of high and low anomaly values distributed in different parts of the CMER. The map (Fig 5.20) shows a maximum anomaly of 872 nT and a minimum anomaly of -570 nT in the residual anomaly filed over the study area.

The residual magnetic anomaly map delineates areas like Aluto volcanic center, Adami Tulu, and Tita with very high magnetic anomalies. The Munesa-Asela border faults, Lake Langano, Corbeti volcanic center and Hawassa caldera complexes are also characterized by high magnetic anomaly values. All these high anomalous zones are concentrated within the WFB that runs parallel and close to the Munesa-Asela border faults.

The area around Tita and Tora, which is close to the SDZFFZ also shows high residual magnetic anomaly. This indicates that the main sources of these magnetically anomalous zones found within or close to the WFB and SDZFFZ are probably found at shallow depths.

The anomaly variations shown in the residual magnetic anomaly generally are associated with the geologic materials and structural features in the study area. With this regard, the geological map of the study area (Fig 2.7) assists in the interpretation of the residual magnetic anomaly map.

The northeastern part of Lake Langano, which is covered by Bofa basalt, indicates positive magnetic anomalies, which may result from high remanent magnetization. Similarly, the western part of Meraro and the northern part of Shire localities are demarcated by positive residual anomalies, which are thought to be associated with the Pliocene trachyte and Pliocene basalts.

On the other hand, the central part of Lake Ziway, NE of Lake Ziway, areas associated with Lakes Abiyata and Langano, west of Lake Abiyata, north of Corbeti are dominated by low to very low magnetic anomalies. Thick Quaternary sediments (Fig. 2.7) cover these zones.

Since the study area is located at low magnetic latitude (magnetic equatorial region), there exists an inverse relationship between nature of magnetic susceptibility of the geologic bodies/sources and their associated magnetic responses (anomalies). Therefore, the high to very high magnetic anomalies depicted over the volcanic centers are attributed to low magnetic susceptibility geologic materials beneath these silicic volcanic centers.

On the other hand, the negative magnetic anomaly zone, which covers most of the Lakes region and associated depressions and part of the highland, are attributed to high magnetic susceptibility sediments derived from the neighboring volcanic rocks.

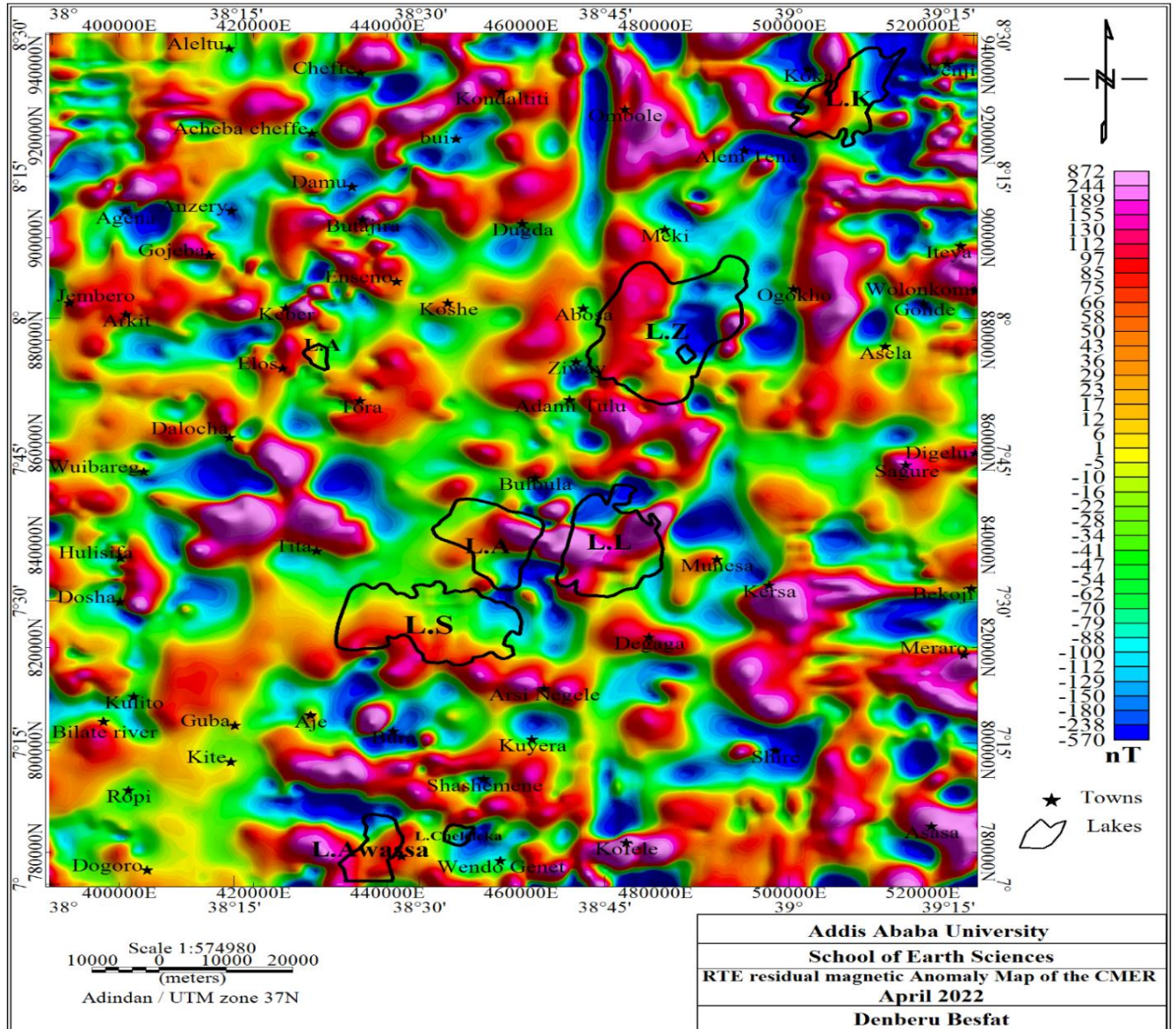


Figure 5.20 Residual magnetic anomaly map of the CMER.

5.4 Enhancement of Magnetic Data

Similar to the data enhancement techniques applied to the gravity filed discussed in section 5.2 of this MSc thesis work different data enhancement techniques are also applied to the magnetic field data in order to boost the interpretation process. Enhancement techniques including radially

averaged power spectrum (RAPS), Total horizontal derivative (THDR), magnetic analytical signal (AS), tilt derivative (TDR), Euler 3-D Deconvolution and source edge detection (SED) are employed to the magnetic data. The results of these data enhancement techniques are presented as follows.

5.4.1 Radially Averaged Power Spectrum Analysis (RAPS)

As it is discussed on section 5.2.1 of this MSc thesis, radially averaged power spectrum analysis (RAPS) is based on the principle that the potential field data measured at the surface is the resultant effect of the sources occurring at various depths. These depths can be determined from the radially averaged power spectrum of the potential fields (BLAKELY 1995, MAMMO 2011).

In this research, 2-D fast Fourier transformation (2-DFFT, Oasis Montaj 6.4.2) is applied to the RTE_TMA grid to compute the radially averaged power spectrum depth estimates.

Accordingly, the average depth to the magnetic anomaly sources are computed from the slopes of the power spectrum plot (Fig 5.21) using equation 5.2. From the calculated depth result, the deeper anomaly sources occur below 14.01 km depth, intermediate anomaly sources at depths less than 4.29 km and shallower sources at less than 1.35 km depth. From these results, we can see that most of the magnetic anomaly sources are located at intermediate and shallow depths. Moreover, these anomalies are clearly demarcated in the residual magnetic anomaly map (Fig 5.20).

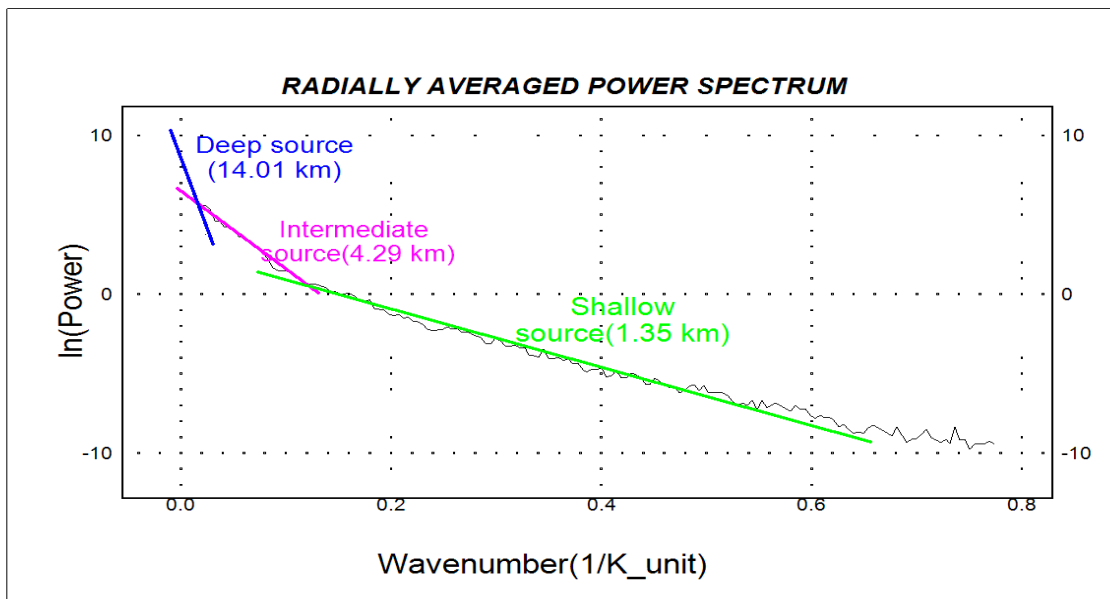


Figure 5.21 Radially Average power spectrum depth estimate plot

5.4.2 Total Horizontal Derivative (THDR) Magnetic Map

The total horizontal derivative (THDR) or gradient method has been used extensively to map the boundaries of susceptibility contrasts (BLAKLEY, et al 1986). It because the horizontal derivative of the residual magnetic field generated by a tabular body tends to have maximum values over the edges of the anomalous body in case the edges are vertical and well separated from each other. It has high amplitude over the edge of the anomaly.

Total horizontal derivative for magnetic field can be formulated as,

$$|THDR(x,y)| = \sqrt{\left(\frac{\partial M}{\partial x}\right)^2 + \left(\frac{\partial M}{\partial y}\right)^2} \quad 5.7$$

Where M is the observed magnetic field, and $\left(\frac{\partial M}{\partial x}\right)^2$ and $\left(\frac{\partial M}{\partial y}\right)^2$ are the two horizontal derivatives of the observed magnetic field in the x and y directions, respectively.

The THDR magnetic map (Fig 5.22) of the study area is compiled by applying THDR filter to the gridded residual magnetic anomaly map (Fig 5.20) using the source edge detection (SED) algorithm of BLAKLEY, et al (1986) an extension in Geosoft Oasis Montaj. The compiled THDR map expected to locate edges of shallow magnetic anomaly sources.

The compiled THDR magnetic map (Fig 5.22) shows high amplitude values over the northern, northeastern and SE parts of the study area which indicate presence of structural discontinuity (mainly boundary faults) between the rift floor and the flanks of the rift. Moreover, this maximum amplitude is an indication for the presence of strong slopes (i.e. transition from the plateaus to the rift) or maximum lateral susceptibility variations. Likewise, the amplitude of the THDR map is also high over towns like Tita, Bura and the Lake regions of Ziway, Langano, Abiyata and Hawassa, which can be correlated with either structural discontinuities and or abrupt variations in rock susceptibilities.

The amplitude of the THDR along the CMER is relatively minimum. This can be related with a decrease in structural discontinuity and lateral variations along the rift axis, which is mainly filled with identical lacustrine deposits (Fig 2.7). A close observation on the patterns of the THDR amplitudes along the rift axis one can observe differences in susceptibility between the volcanic

centers and their surroundings. However, the general pattern of the THDR follows the major boundary faults.

With the aim of better detecting the anomaly source edges a peak value of 3 by 4 (grid values in three directions) level of peak detection is applied to the THDR grid using source edge detection (SED) tool. Afterwards the peak level detections applied are automatically generated in a database containing the peak level and peak direction. The produced peak value solutions are then plotted on the THDR map (Fig 5.23) in order to better visualize the detected anomaly source edges.

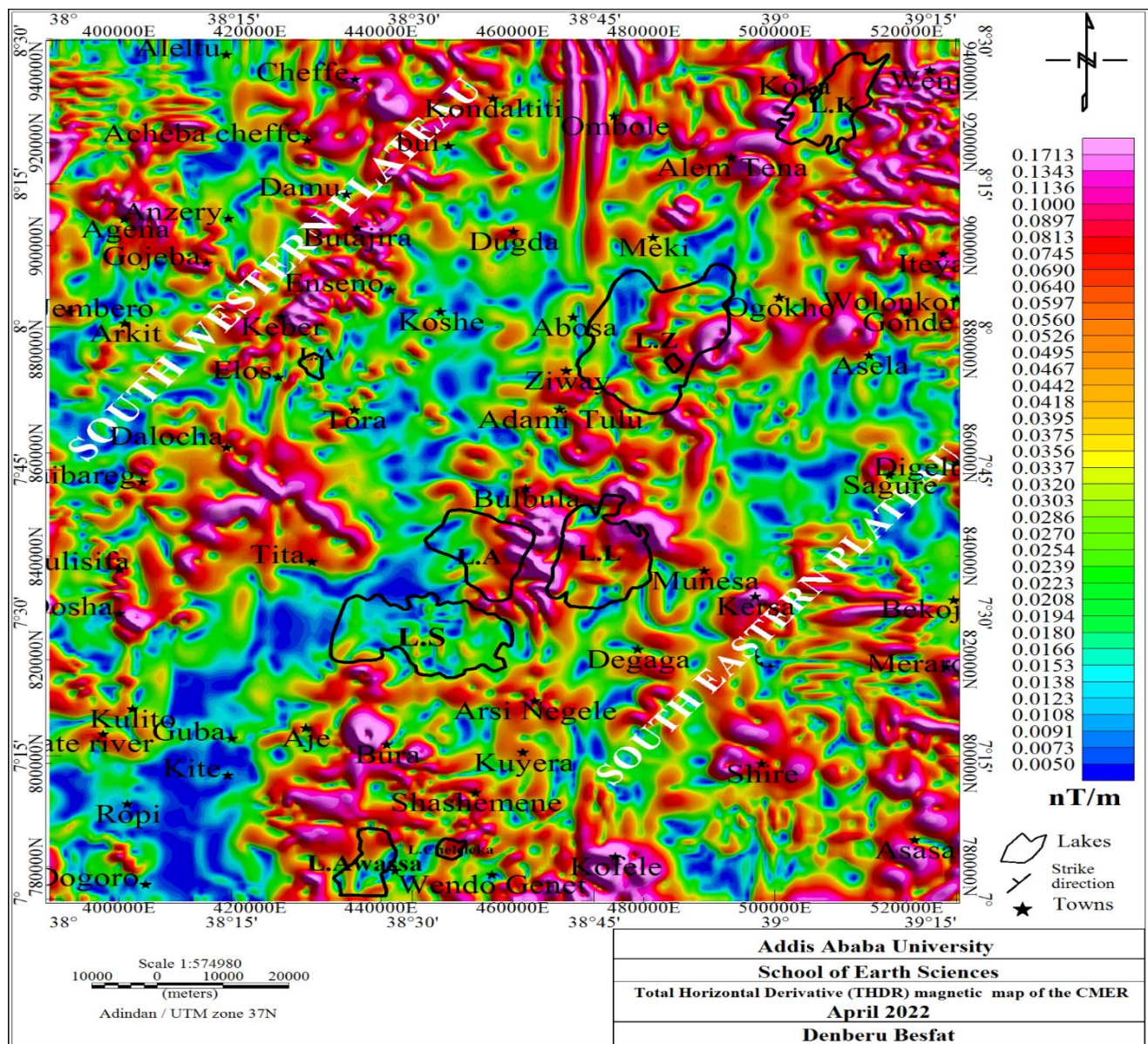


Fig 5.22 Total horizontal derivative (THDR) magnetic map of the CMER

The detected source edges superimposed on the THDR map (Fig 5.23) show the strike direction of the edge (contact) by the direction of the long axis, and the ‘down-gradient’ direction by the dip indicator. The plotted SED symbols over the total horizontal derivative grid traced the local peaks. The ‘down-gradient’ direction points away from the source, which is locally magnetized.

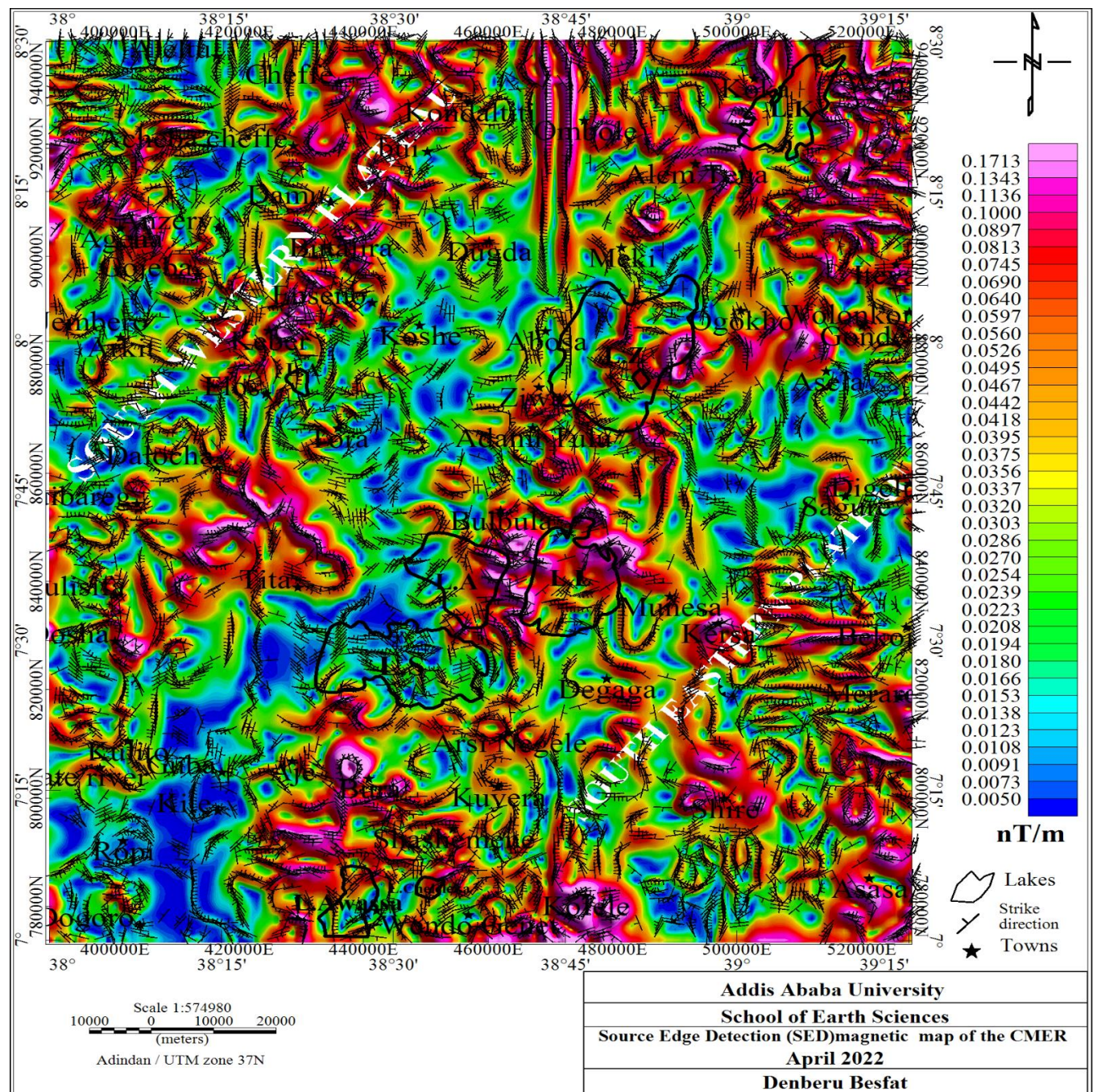


Figure 5.23 Source edge detection (SED) magnetic map of the CMER.

5.4.3 Magnetic Analytical Signal Map

Analytical signal (AS) method is another powerful data enhancement technique for locating the edges of magnetic source bodies, particularly where remanence and/or low magnetic latitude complicates interpretation. (Geosoft Oasis Montaj Modified Topic v6.3 (21-Dec-05). AS is used for delineation of maxima values of anomalous susceptibility distributions in the subsurface. Analytic Signal amplitude values attain their maxima over their corresponding source bodies more often, over their edges. Magnitude of AS of magnetic anomalous field is calculated by taking the square root of the sum of square derivatives in all three directions:

$$|AS(x,y)| = \sqrt{\left(\frac{\partial M}{\partial x}\right)^2 + \left(\frac{\partial M}{\partial y}\right)^2 + \left(\frac{\partial M}{\partial z}\right)^2} \quad 5.8$$

Where AS (x, y) is the amplitude of the analytic signal at point (x, y), M is the observed magnetic at anomaly of preference point (x, y) and $\partial M/\partial x$, $\partial M/\partial y$ are the two horizontal derivatives and $\partial M/\partial z$ is the first vertical derivative of the observed magnetic anomaly.

The AS magnetic map of the CMER (Fig. 5.24) is compiled from the gridded residual magnetic anomaly map (Fig 5.20). The geological map of the study area (Fig 2.7) is used for the interpretation of the Analytical signal magnetic map in order to better understand boundaries of shallow anomaly sources.

Consequently, SW, SE and NE parts of the study are manifested by maximum AS values on the analytical map of the study area. The SW part of the study area, which constitute the Tarmaber-Megezez formation, welded pyroclastic rocks, basalt flows and cones are demarcated by maximum AS values. The tertiary basalts over the SE margins of the rift are also reflected by maximum AS magnetic values. These maximum AS values can be correlated with the residual magnetic anomaly map (Fig 5.20) which shows higher magnetic anomaly values over these regions.

On the contrary, the lacustrine sediments in the central portion of the geological map are characterized by low values in the AS (Fig. 5.24) map. The rift floor volcanic centers (Corbeti, Aluto Langano, Gudemssa and Tulu Moye) and off rift regions (Zikuwala and Chilalo volcano) are clearly reflected by circular shaped gradient maxima in the AS. Hence, the AS can be considered as a powerful technique in enhancing the geometry of shallow seated anomaly sources.

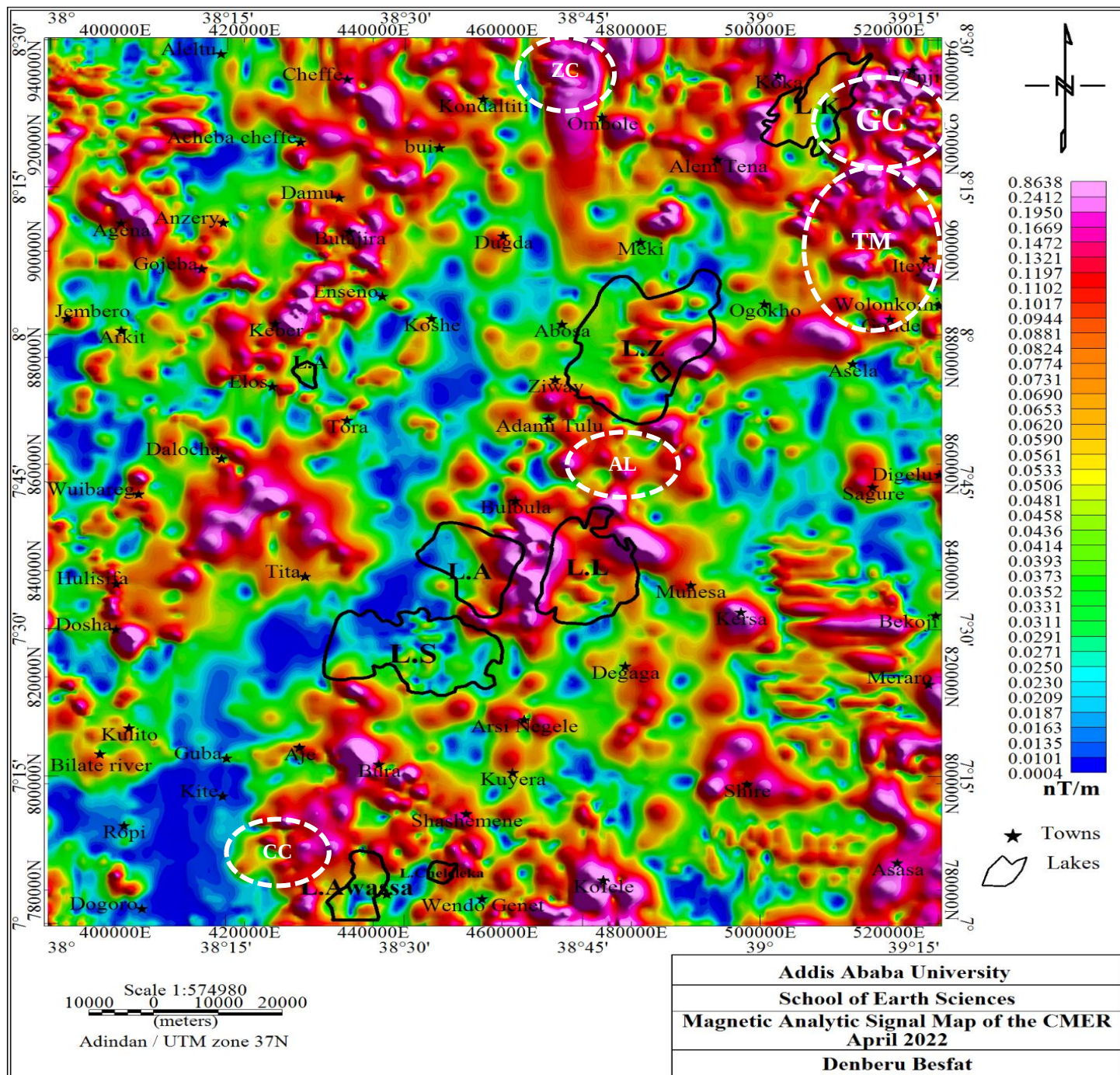


Figure 5.24 Magnetic analytical signal map of the CMER. (Labels: CC- Corbetti caldera, AL- Aluto Langano, TM-Tulu Moye, GC-Gedemssa caldera)

5.4.4 Tilt Angle (Tilt Derivative) Magnetic Map

The tilt angle derivative (TDR) method is another anomaly source edge detection method useful for mapping shallow basement structures and mineral exploration targets. (MILLER et al, 1994). The local phase or tilt angle, or tilt derivative (TDR), is formulated as:

$$TDR = \theta = \tan^{-1} \left[\left(\frac{\partial M}{\partial z} \right) / \left(\frac{\partial M}{\partial h} \right) \right] = \tan^{-1} \left[\frac{VDR}{THDR} \right] \quad 5.9$$

Where, $\frac{\partial M}{\partial x}$, $\frac{\partial M}{\partial y}$, $\frac{\partial M}{\partial z}$, are the derivatives of the observed magnetic anomaly of preference in the x, y, z direction respectively.

$\frac{\partial M}{\partial h} = \sqrt{\left(\frac{\partial M}{\partial x} \right)^2 + \left(\frac{\partial M}{\partial y} \right)^2}$ is the total horizontal derivative (THDR) of the observed magnetic anomaly of preference in x, y directions, $\frac{\partial M}{\partial z}$ is the first vertical derivative (VDR). (WILLIAM J, et al 2013).

Tilt angle derivative amplitude variations have a certain range, between $-\pi/2$ and $\pi/2$. Its amplitude has three phases: positive over the source, zero at/near the edge of the source, and negative outside the source (ROSET, et al 1992). In the existence of noise, this technique acts as an effective signal discriminator for both shallow and intermediate sources but becomes blurry for sources at considerable depths, where it can not reveal deep-level geologic boundaries. Although its efficiency reduces with depth, this method produces relatively sharp anomalies and generates better definition of edges over the body (VERDUZCO, et al., 2004).

The TDR map (Fig. 5.25) of the study area is compiled from the gridded THDR map (Fig 5.22) and first VDR grid using the source edge detection (SED) algorithm. The map is generated in order to identify the location of faults, contacts, edges and geological boundaries in the study area. Furthermore, it is also expected to reveal structures/contacts and boundaries better than the analytic signal map. From geological perspective, the study area is categorized by a set of numerous faults in the N-S, NE-SW and NW-SE directions (Fig 2.8).

The reorientations of anomalies' axes revealed over many parts of the study area is attributed to the presence of folded/faulted structures or when a change of geological strike is occurred. The discontinuities in anomalies observed are rather an indication for the presence of tectonic structures

such as faults, fracture zone or other similar structures. The narrow elongated chain observed on the northern part of the CMER around Ombole town is an indication for the presence of faults.

The tilt derivative map (Fig 5.25) reveals the presence of abundant E-W trending anomaly zones, which are possibly comparable with fault offsets. Besides, there are some other trends in the study area with NW-SE orientation. The peak values in TDR grid are extracted in order to locate the source edges (faults/contacts) and are plotted over the TDR map (Fig 5.25).

The positive values (pink colors) in the TDR map (Fig 5.25) designate the anomaly source, the zero crossing values (yellow colors) indicate the edge of the anomaly source and the negative values (blue colors) characterize region outside of the anomaly source. The detected source edges superimposed on the TDR map (Fig 5.25) shows the strike direction of the edge (contact) by the direction of the long axis, and the ‘down-gradient’ direction by the dip indicator. The plotted SED symbols over the tilt derivative grid trace the local peaks. The edge of the traced local peaks (contacts) superimposed on the TDR map (Fig 5.25) is indicated by longer axis while the dip indicator (small hatches) indicated by the down gradient direction. The ‘down-gradient’ direction points away from the source, which is locally magnetized.

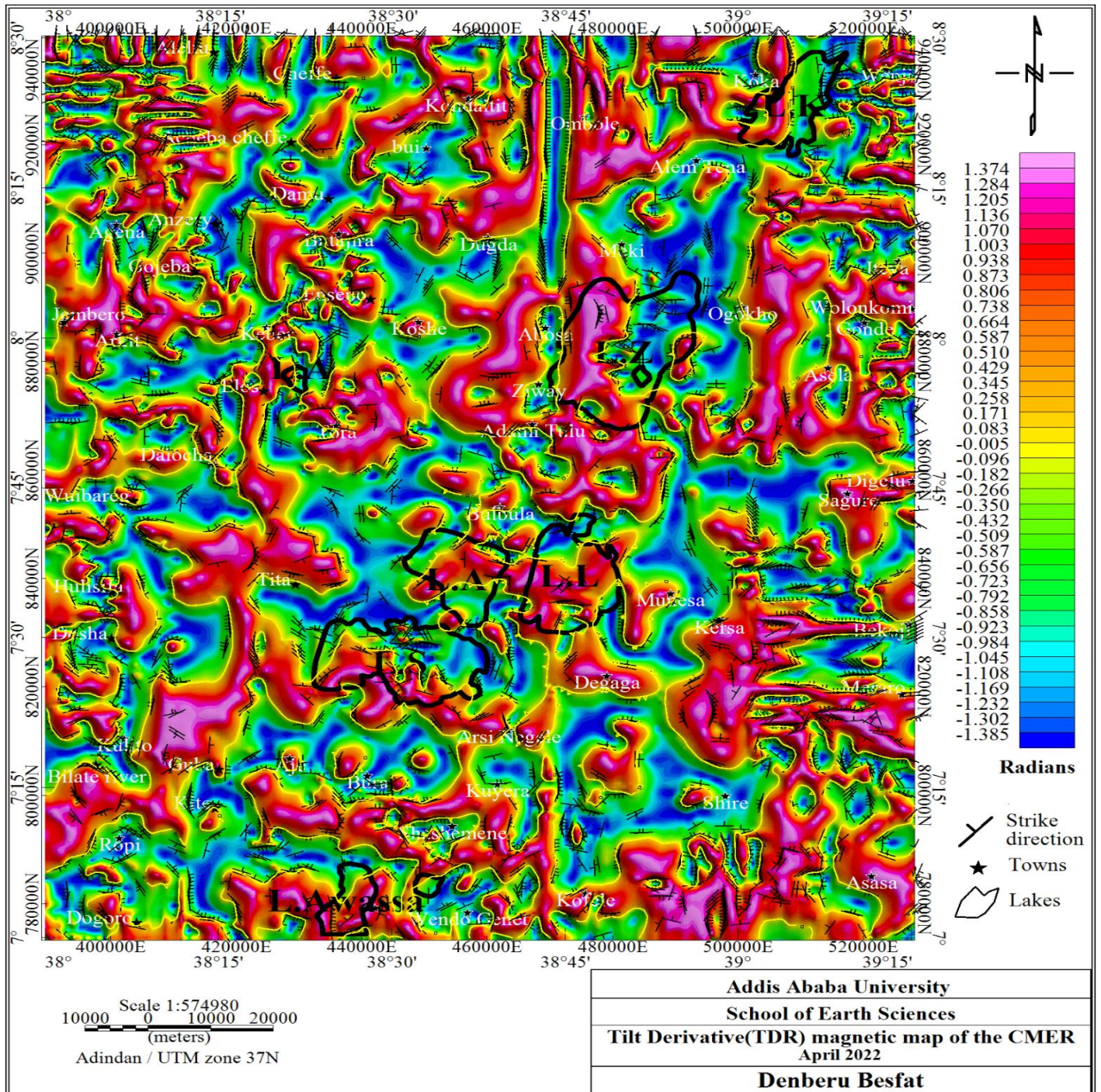


Figure 5.25 Tilt derivative magnetic map of the CMER

5.4.5 Euler 3-D Deconvolution magnetic map

In this study, the standard Euler 3-D Deconvolution is applied to the RTE_TMA magnetic anomaly grid using Geosoft Oasis Montaj software in order to estimate the depth and location of the

magnetic anomaly source of a particular structural type defined by the structural parameter. From the principle of Euler 3-D deconvolution (SERGUEI et al, 2000), the structural index values 0, 1, 2, 3 represent geological features of contact, sill/dike, vertical pipe/horizontal cylinder and sphere respectively. The Euler 3-D optimal structural index is selected by a visual inspection of maps that are compiled using various structural indices until the best clustering of solutions is obtained.

A structural index of two was determined and applied to the residual magnetic anomaly map to derive the Euler 3-D deconvolution depth results. The map reveals magnetic sources of different depths marked by different colored symbols plotted on the map. These include a violet circle indicating magnetic sources with depth greater than 9.5 km, a light red circle indicating magnetic sources that range from 8 km to 9.5 km depth, a salmon circle represents sources that range from 6.5 km to 8 km depth, a cyan circle representing a depth range between 5 km to 6.5 km, a blue circle from 3.5 km to 5 km, a red circle from 2 km to 3.5 km and a small yellow representing sources that range from 0.5 km to 2 km depth.

The blue and red circles, which represent relatively shallow sources, are distributed along the WFB. In particular, they are densely concentrated over the locations of the Aluto, Shalla and Ziway volcanic complexes indicating that the magnetic sources there are of shallow origin. The violet circle, which represents relatively deeper (> 9.5 km), is mostly distributed over the shoulders of the SW and SE plateaus indicating the magnetic sources below the shoulders of the plateaus are relatively deeper. The cyan and salmon circles which represent relatively intermediate magnetic sources are concentrated over the areas dominated by thick sediments of the rift lakes and substantial parts of the SW and SE plateaus indicating that the sources there are at relatively intermediate depths. The distribution of the sources (shallow, intermediate and deep sources) over the study area revealed by the Euler deconvolution magnetic map appear to be consistent with the distribution of the sources revealed by the regional magnetic anomaly map (Fig. 5.19) and the residual magnetic anomaly map (Fig. 5.20).

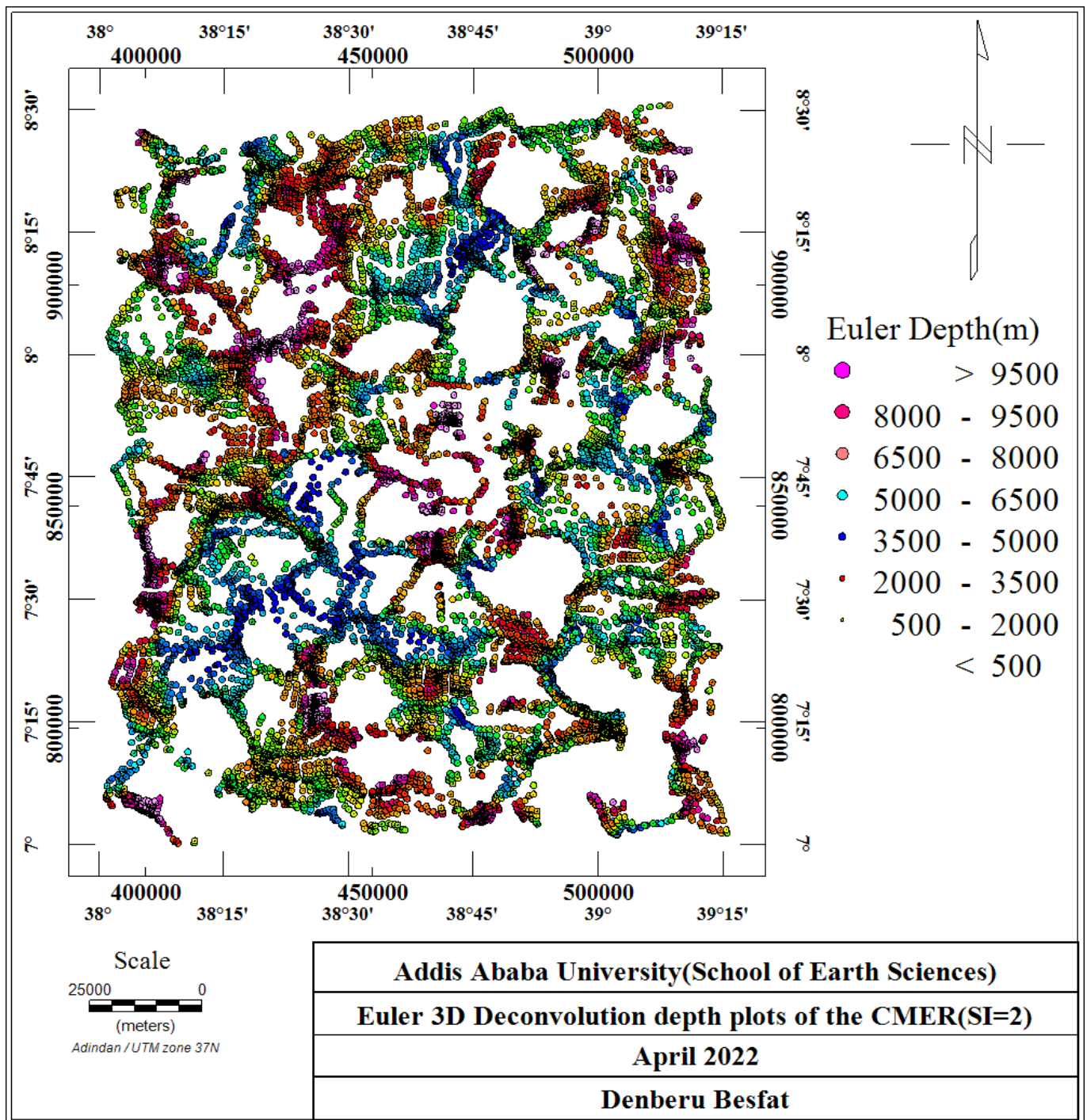


Figure 5.26 Standard Euler 3-D Deconvolution magnetic depth plot of the CMER

CHAPTER SIX

3-D INVERSION OF GRAVITY AND MAGNETIC DATA

6.1 Background

Inversion is a mathematical procedure for deriving a set of parameters describing a model that can explain a set of observations (WILLIAMS, 2008).

This MSc Thesis research work illustrates 3-D inversion of airborne gravity and ground magnetic data acquired over the central part of the main Ethiopian rift (CMER). The Results of the 3-D inversion are presented as 3-D of subsurface density and magnetic susceptibility models.

The 3-D gravity and magnetic inversion work in this research uses the codes of LI and OLDENBURG (1996, 1998) developed at the University of British Columbia - Geophysical Inversion Facility (UBC-GIF). The theoretical and mathematical foundations for 3-D gravity and magnetic inversion methodology are outlined in Chapter 3 (Section 3.5).

6.2. 3-D Gravity and Magnetic Inversion Procedures

6.2.1 Data preparation

3-D Inversion of gravity and magnetic data requires a number of procedures. Data preparation is the most crucial aspect of the process. For 3-D inversion of gravity and magnetic data, the data must only contain extracted anomaly data with no contributions from fields caused by materials outside the volume under the survey area (LI and OLDENBURG, 1998).

For the 3-D gravity and magnetic inversion work in this MSc thesis, the extracted gravity and magnetic anomaly data with station locations that are at a regular spacing were prepared prior to the inversion process.

Upward continuation of the extracted anomaly data is a recommended procedure in order to ensure fast convergence (MAMMO, 2013) with a reasonable time during an inversion. Moreover, upward continuation also helps in order to remove near surface (shorter wavelength) effects that introduce errors and artifacts. Usually upward continuation to a height of half the thickness of the shallowest cells is indorsed by UBC-GIF. Accordingly, the extracted gravity and magnetic anomaly data were

upward continued to 10 kms in order to ensure fast convergence. Specifying standard deviations as a percentage of the datum plus a minimum value in units of nT or milliGals is also another important procedure that helps to achieve a suitable degree of misfit in the inversion.

In this study, the gravity data was assigned 2 % plus 0.01mGal error and the magnetic data was assigned 5% plus 0.01nT of error. Figure 6.1 shows the extracted gravity and magnetic anomaly data prepared for the 3-D gravity inversion.

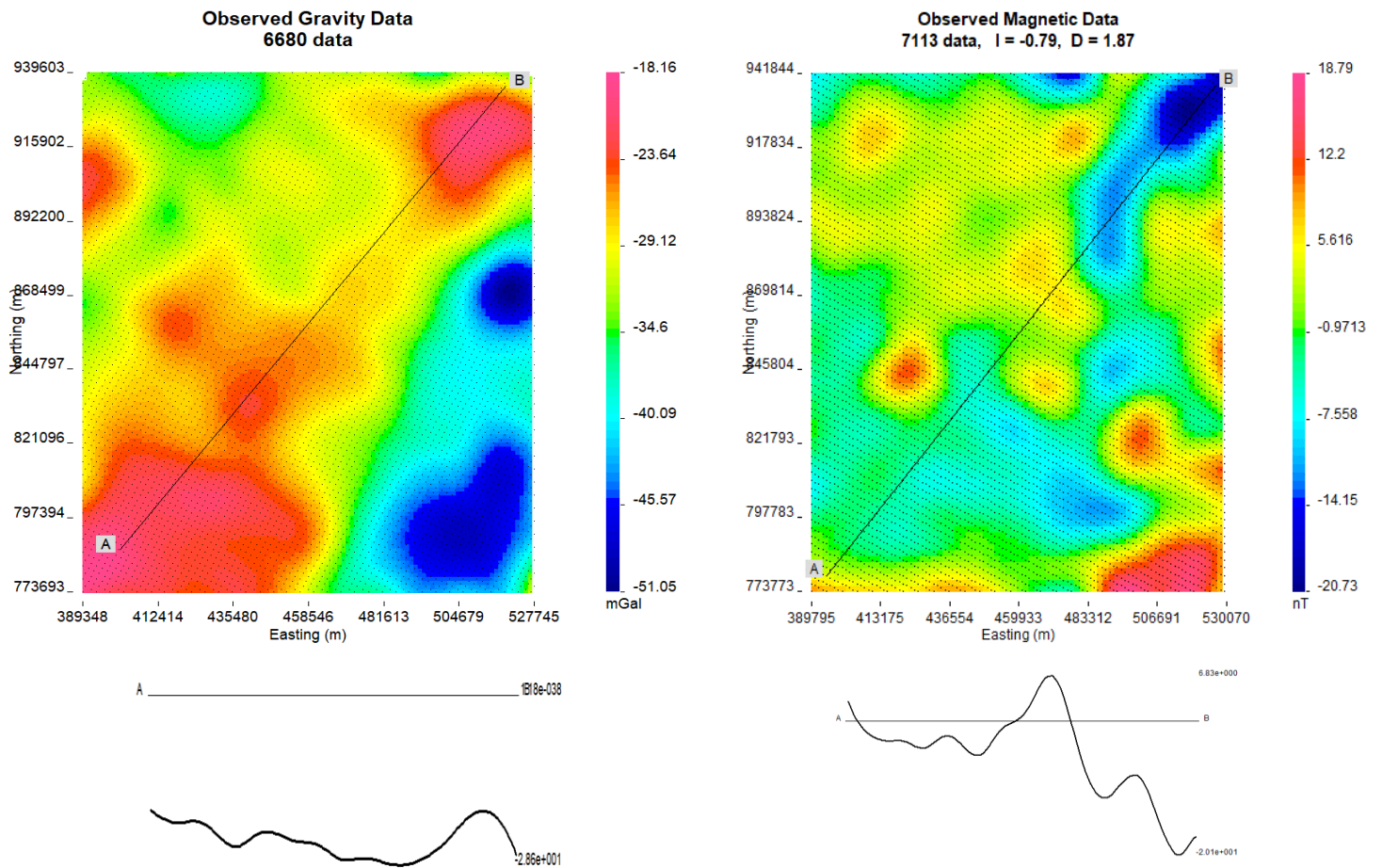


Figure 6.1 Extracted gravity and magnetic anomaly data sets

6.2.2. 3-D Mesh Design

The next important step for an inversion work is designing a 3-D orthogonal mesh. Since the subsurface is a continuous medium, it must be discretized into a finite number of cells before calculations can be performed to generate data or to estimate models by inversion of measurements. This process of discretizing the earth is called Mesh design.

The mesh file is used for 3-D forward modelling and inversion. The size of the cells should reflect the resolving power of the survey. If the cells are too large, important geologic features may not be adequately modeled. If they are too small, it should not adversely affect the inversion outcome but it may slow the process down due to an increase in the size of the system of equations to be solved.

3-D orthogonal mesh has two parts namely the core (local) mesh and the padding (regional) cells. The core mesh is a 3-D grid cells (along X, Y, and Z) that cover an area from where the data is collected whereas Padding cells are extra 3-D grid cells that surround the core mesh and are used to impose boundary conditions (to avoid the effects of any anomalous structures outside of the core mesh). Padding cells are included in the inversion process only. When the final inversion is achieved, the padding cells will be cut off from the inversion result.

Figure 6.2 shows the designed 3-D orthogonal mesh for the 3-D gravity and magnetic data inversion in this MSc Thesis. The central part of the mesh shows the core mesh (CMER) and the extra cells in four directions are the padding cells.

The study area (CMER) is discretized in to 168,000 cells (60 cells to the east, 70 cells to the north and 70 cells to the down vertical direction) including padding cells. This is by far a very big mesh having high resolution (many small cells) which is often desirable. However, it requires more computer memories and processing time during the inversion. This is due to the number of observed data multiplied by the number of mesh cells ($6680 \times 168,000 = 1,122,240,000$, for gravity) is performed in calculating the sensitivity matrix and stored in the RAM, which is found more than tens of giga bytes (GB) in the inversion process in this work. However, the resolution of the recovered models is expected to be improved. Moreover, the current UBC-GIF codes support a maximum number of 200,000 cells in a Mesh and 10,000 sets of observed gravity or magnetic data. The top southwest corner of the mesh is placed at 229400 (easting), 615389 (northing) and zero meters depth.

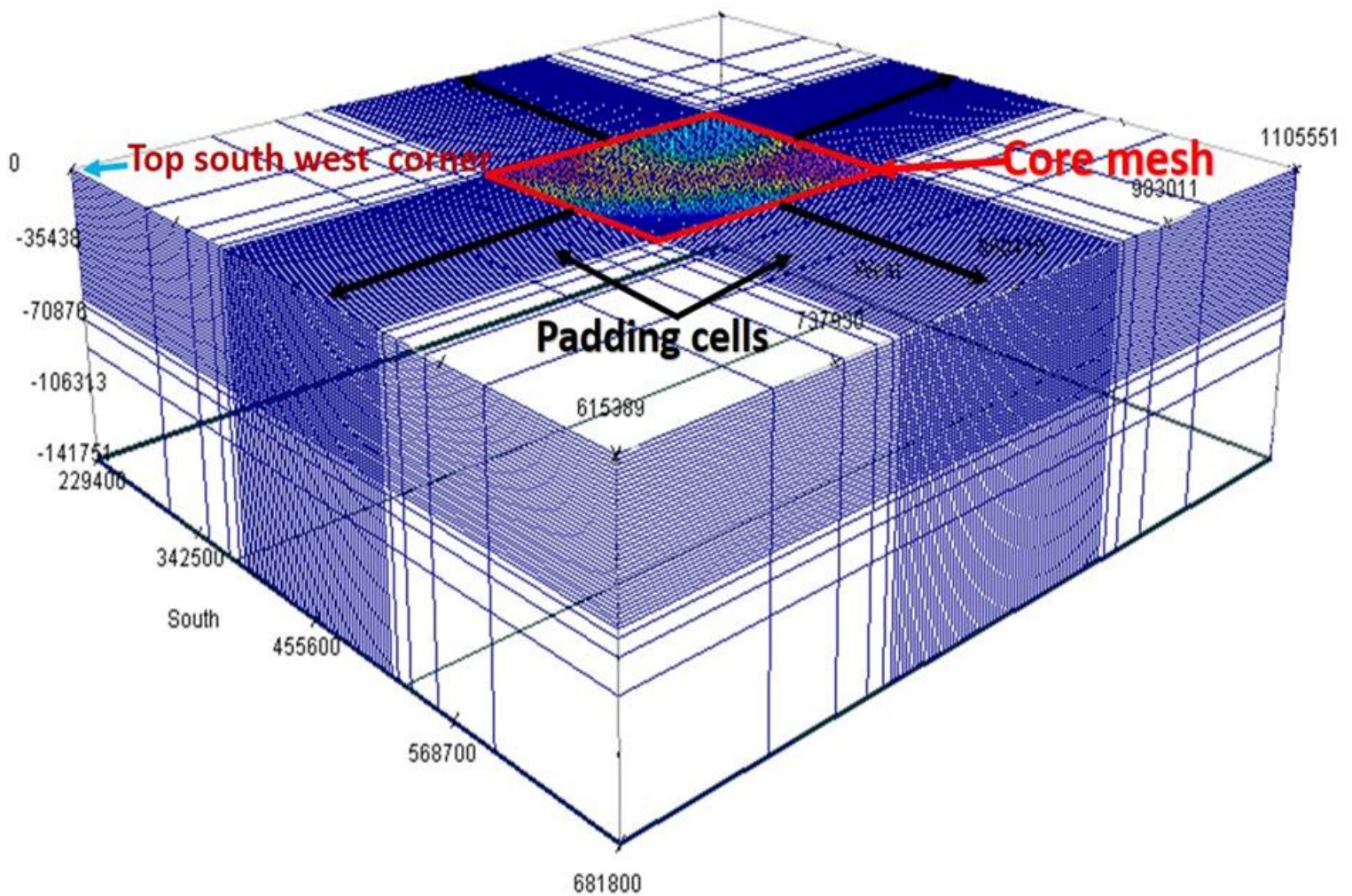


Figure 6.2 Designed 3-D orthogonal mesh showing the core mesh and the padding cells.

6.2.3. 3-D Inversion of Gravity data

Gravity inversion involves recovering subsurface density distributions from the observed gravity data. However, the inverse calculation, deriving density distributions from the observed data, is hampered by numerical problems and non-uniqueness. There may be an infinite number of models that explain the observed gravity data equally well. One approach is to seek a model that fulfills certain criteria based on the geological character expected in the model (LI and OLDENBURG, 1998).

The UBC–GIF gravity inversion codes seek a model that is as close as possible to some physical property reference model, which can be based on the expected geology, with a measure of

smoothness between cells (LI and OLDENBURG, 1996, 1998). The recovered model may be viewed as the simplest model that can satisfy the available geological information and the geophysical data.

However, the inversion can be also run without any geological information using default settings (a reference model that is zero everywhere) to recover a physical property model that captures the key characteristics needed to reproduce the geophysical data (WILLIAMS, 2008). This has been the strength and the main use of the UBC–GIF inversion codes for mineral exploration problems (WILLIAMS, 2008). On the other hand, when detailed geological information is included, the recovered physical property models (density contrast models) can contain as much detail as necessary to satisfy all the available information, and will also provide much more reliable predictions of the subsurface physical property distribution (WILLIAMS, 2008).

Inversion of gravity data highly suffers from an inherent non-uniqueness problem. In order to address the non-uniqueness of the solutions and obtain a single optimal solution (density model), a model objective function (Φ_m) and a data misfit functions (Φ_d) are defined. Then the inversion process is then posed as a mathematical optimization problem.

The model objective function $\Phi_m(\rho)$ is constructed as;

$$\begin{aligned} \Phi_m(\rho) = & \alpha_s \int_V ws \{w(r)[\rho(r) - \rho_0]\}^2 dv + \alpha_x \int_V wx \left\{ \frac{\partial w(r)[\rho(r) - \rho_0]}{\partial x} \right\}^2 dv \\ & + \alpha_y \int_V wy \left\{ \frac{\partial w(r)[\rho(r) - \rho_0]}{\partial y} \right\}^2 dv + \alpha_z \int_V wz \left\{ \frac{\partial w(r)[\rho(r) - \rho_0]}{\partial z} \right\}^2 dv, \quad 6.1 \end{aligned}$$

where α_s , α_x , α_y and α_z control how important each of the four terms are relative to each other and are used to balance the contributions of the smallness (closeness) and smoothness components, the functions ws , wx , wy and wz are spatially dependent, ws is used to force the physical property of cells to be closer to the supplied reference model. wx , wy and wz , are used to make the model-difference more or less smooth across cell boundaries to reproduce geological continuity or boundaries. $w(r)$ is a depth weighting function, ρ_0 is the reference model, and $\rho(r)$ is the recovered model.

Similarly, the data misfit function is constructed as;

$$\Phi_d(\rho) = \sum_{i=1}^N \Phi_{d,i}^2 = \sum_{i=1}^N \left(\frac{F[\rho]_i - d_i}{\varepsilon_i} \right)^2 \quad 6.2$$

Where N is the total number of observations (datum), $\Phi_{d,i}$ is the data misfit for each datum, d_i is the observed datum, $F[\rho]_i$ is the predicted datum for model ρ and ε_i is the uncertainty on datum i .

The inversion is formulated using Tikhonov regularization parameter (β) which balances between $\Phi_d(\rho)$ and $\Phi_m(\rho)$. The density value within each prism is determined by solving the optimization problem:

$$\text{Minimize } \Phi = \Phi_d(\rho) + \beta \Phi_m(\rho)$$

$$\text{Subjected to } \Phi_d = \Phi_d^* \quad 6.3$$

where Φ is the global objective function and Φ_d^* is the target misfit, which, assuming that noise is Gaussian with zero mean and standard deviation ε_i , will be equal to the number of data points, N .

A principal step in 3-D inversion of gravity data is construction of a reference model with all available geophysical and geological a priori information that is used to constrain the inversion process (MAMMO, 2013). In this study, attempts are made to gather any a priori information available in the study area (CMER) to constrain the gravity inversion. With this regard, the Ethiopia Afar Geoscientific Lithospheric Experiment (EAGLE, MACKENZIE et al., 2005; MAGUIRE et al., 2006) conducted along and across the Ethiopian rift is considered. However, only the along axis seismic profile (Fig 1.4) was found to occur in the study area (CMER). EAGLE's investigation in the CMER and SMER is limited as compared to that of the Ethiopian plateau, NMER and Afar rift and hence the required available data for the study area is limited. In fact, only four along axis shot points (SP21, SP22, SP23, and SP24) (Fig 1.4) are located in the CMER.

Constructing a full 3-D complicated reference model by incorporating all the available geophysical and geological information (if any) is an exceedingly advanced type of inversion, which requires an extensive amount of time. However, in this research work attempts are made to construct a non-complicated reference model (Fig 6.3) using the geophysical information obtained from the four

seismic shot points in the study area (Fig 1.4). The subsurface velocity information provided by EAGLE's Ray Trace P-wave velocity model result along L2 (Fig 1.5) from shot points SP21- SP24 is first converted into density values using NAFE-DRAKE density velocity relationship given by,

$$\rho = -168.03 + 1765vp - 481.72vp^2 + 60.973vp^3 - 2.6861vp^4 \quad 6.4$$

Where, ρ is the density of subsurface in kg/m^3 and vp is p - wave velocity in kms^{-1} (NAFE, DRAKE, 1957). The estimated density values obtained from the corresponding p-wave velocities beneath each seismic shot point is used to construct the reference model by assigning the density contrasts to each layer beneath each shot point in the study. Figure 6.3 shows the constructed reference model with the top of the model overlaid by airborne gravity data.

Table 6.1-estimated density values obtained from seismic p-wave velocities.

Layers	SP21		SP22		SP23		SP24	
	Depth(m)	$\rho(\text{gm/cc})$	Depth(m)	$\rho(\text{gm/cc})$	Depth(m)	$\rho(\text{gm/cc})$	Depth(m)	$\rho(\text{gm/cc})$
Sediment	0-4.5	2.55	0-4.5	2.55	0-4	2.55	0-3	2.55
UC	4.5-27.5	2.84	4.5-26.5	2.84	4.0-24	2.84	3.0-21	2.84
LC	27.5-45.5	2.96	26.5-43.5	2.96	24-42	2.96	21-37	2.96
M	45.5-70	3.2	43.5-70	3.2	42-70	3.2	37-70	3.2

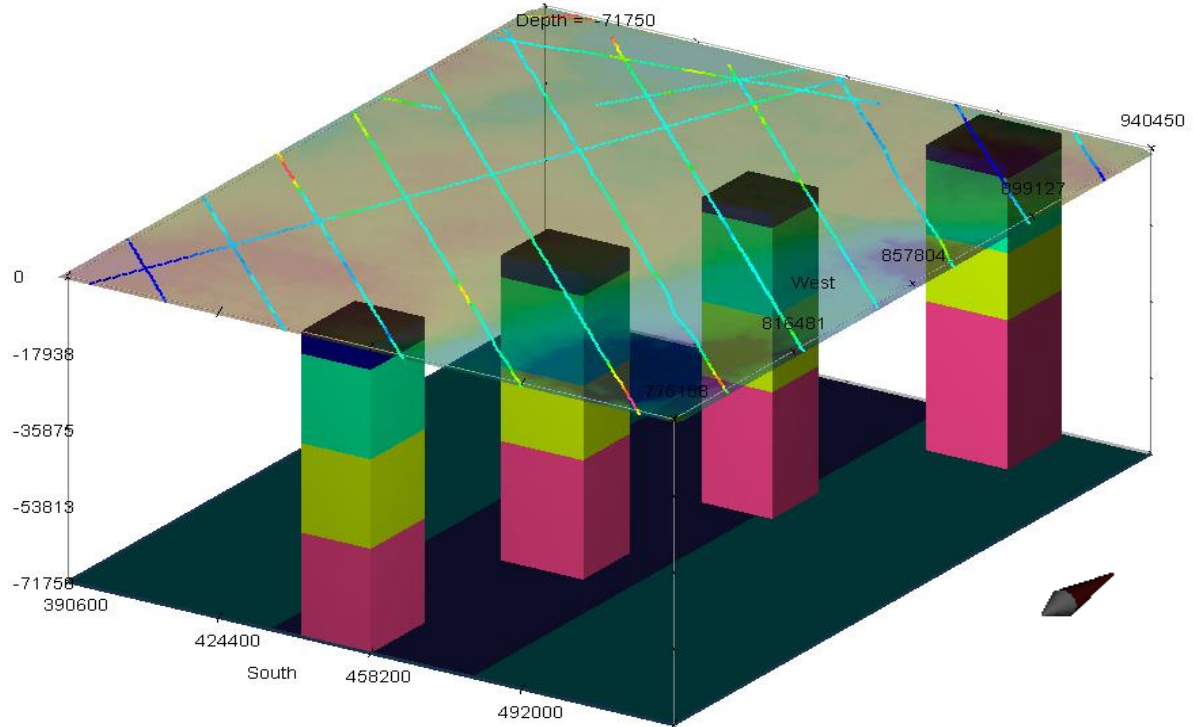


Figure 6.3. The constructed reference model based on the EAGLE project seismic data

Prior to the actual subsequent inversion is run, 3-D gravity forward modelling was executed in order to generate synthetic gravity data from the constructed reference model (Fig 6.3). The generated synthetic data (Fig 6.4 a) was further analyzed and prepared for 3-D inversion in order to recover 3-D subsurface density contrast model (Fig 6.6) for the CMER from the synthetic data itself.

Figures 6.4 a and b show the observed gravity (synthetic data generated from 3-D forward modeling) and the predicted gravity data generated by the inversion of the synthetic data. Moreover, data predicted from the inversion was compared with the synthetic data produced from 3-D forward modelling. The misfit between the two data sets is less than 24 mGal. Hence, inversion has reproduced an acceptable data.

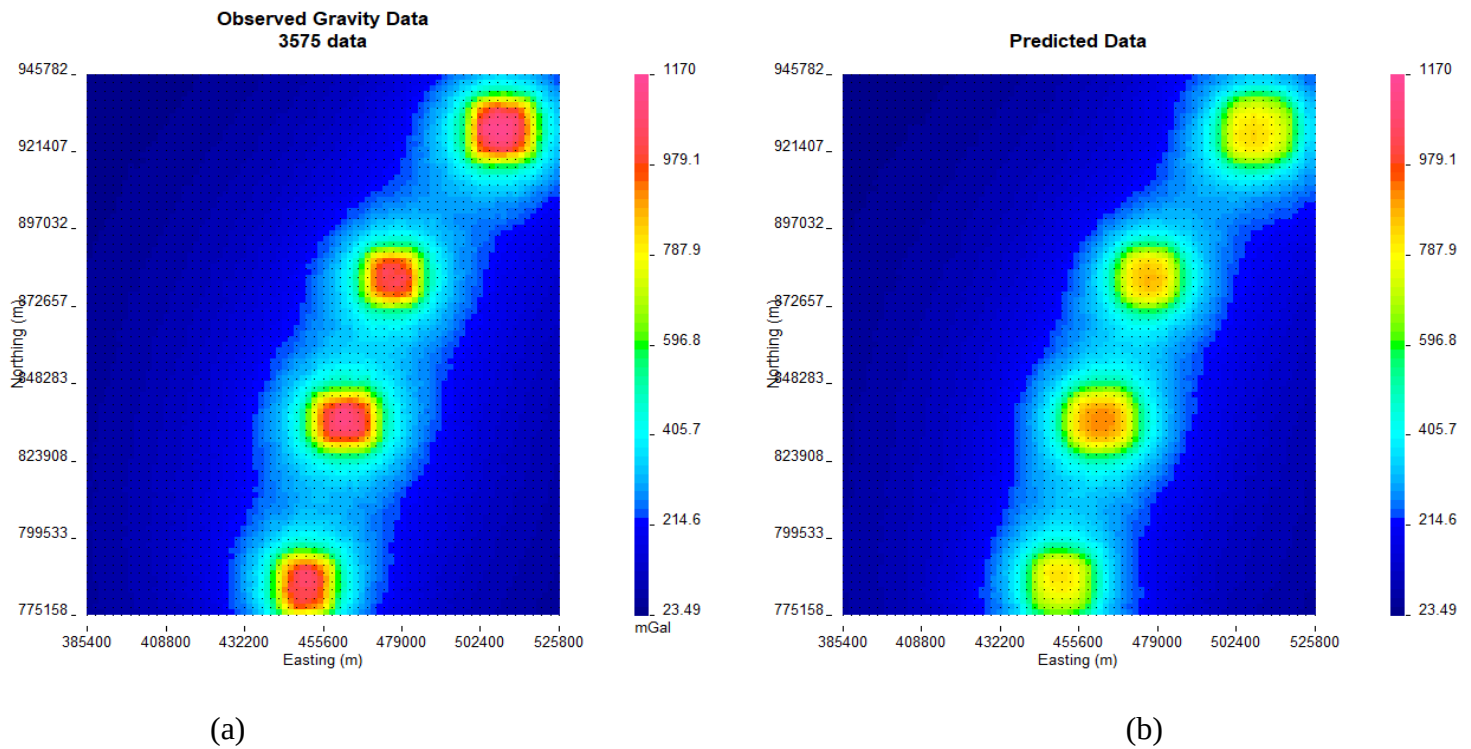


Figure 6.4 (a) Synthetic data obtained from 3-D forward modelling and (b) Predicted data from the 3-D inversion.

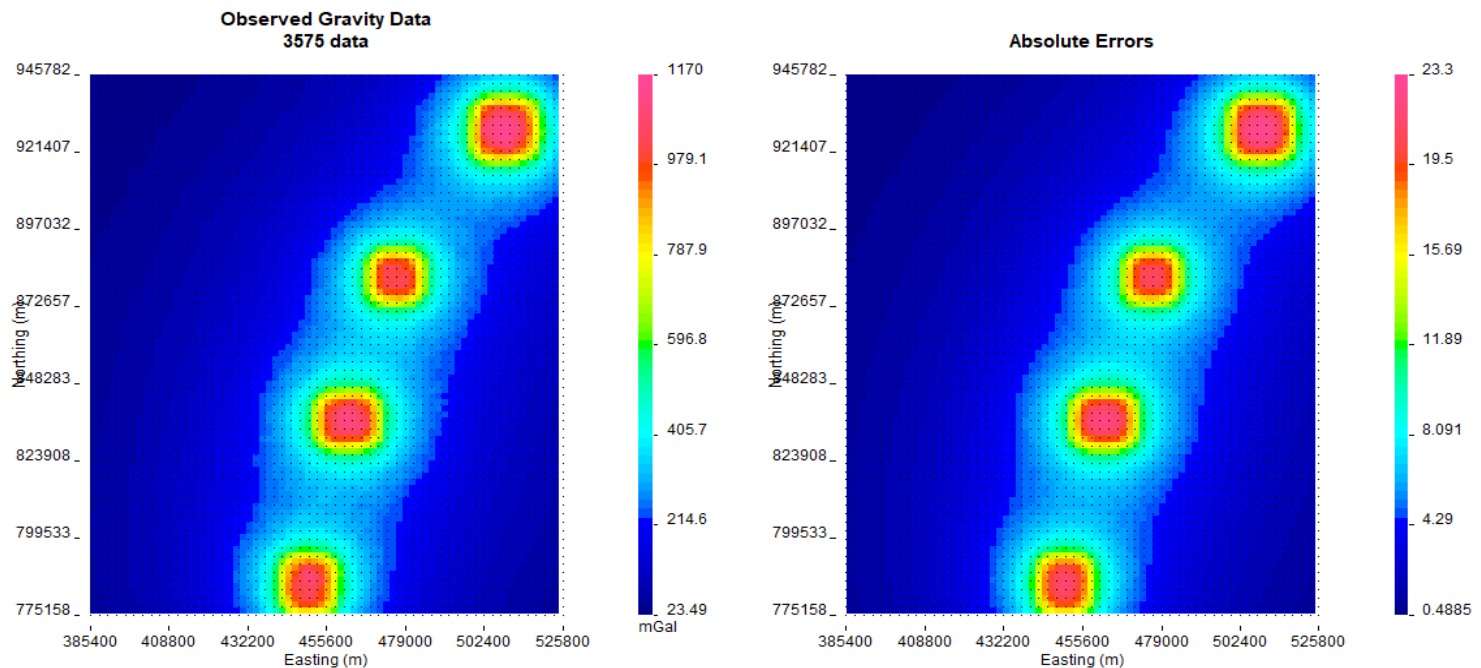


Figure 6.5. Absolute error between the observed (synthetic) and the predicted data

The recovered 3-D density model (Fig 6.6) produced from the inversion of the synthetic data provides the lateral and vertical variations of density in each cell in terms of density contrast. The recovered density contrast had a maximum value of 0.98 g/cc and a minimum value of -0.4205 g/cc.

The variable for the inversion is density contrast and the recovered model is shown in two different ways. The first option for displaying the recovered model is by slices through the density model and the second option is by using iso surfaces. An isosurface is essentially a 3-D contour surface that displays an isosurface by not displaying cells with values that are outside specified upper and lower bounds. By applying a cutoff option to define upper and lower bounds the programme will hide values outside these bounds, leaving a 3-D body that shows the shape of material whose physical properties are within the bounds.

Generally, the UBC-GIF objective function constructs smooth models, so discrete geological boundaries may appear as gradational changes, unless a particular value is used in the image, as in the iso surface renderings. High-density values are red, and low background values are red (OLDENBURG, 1998).

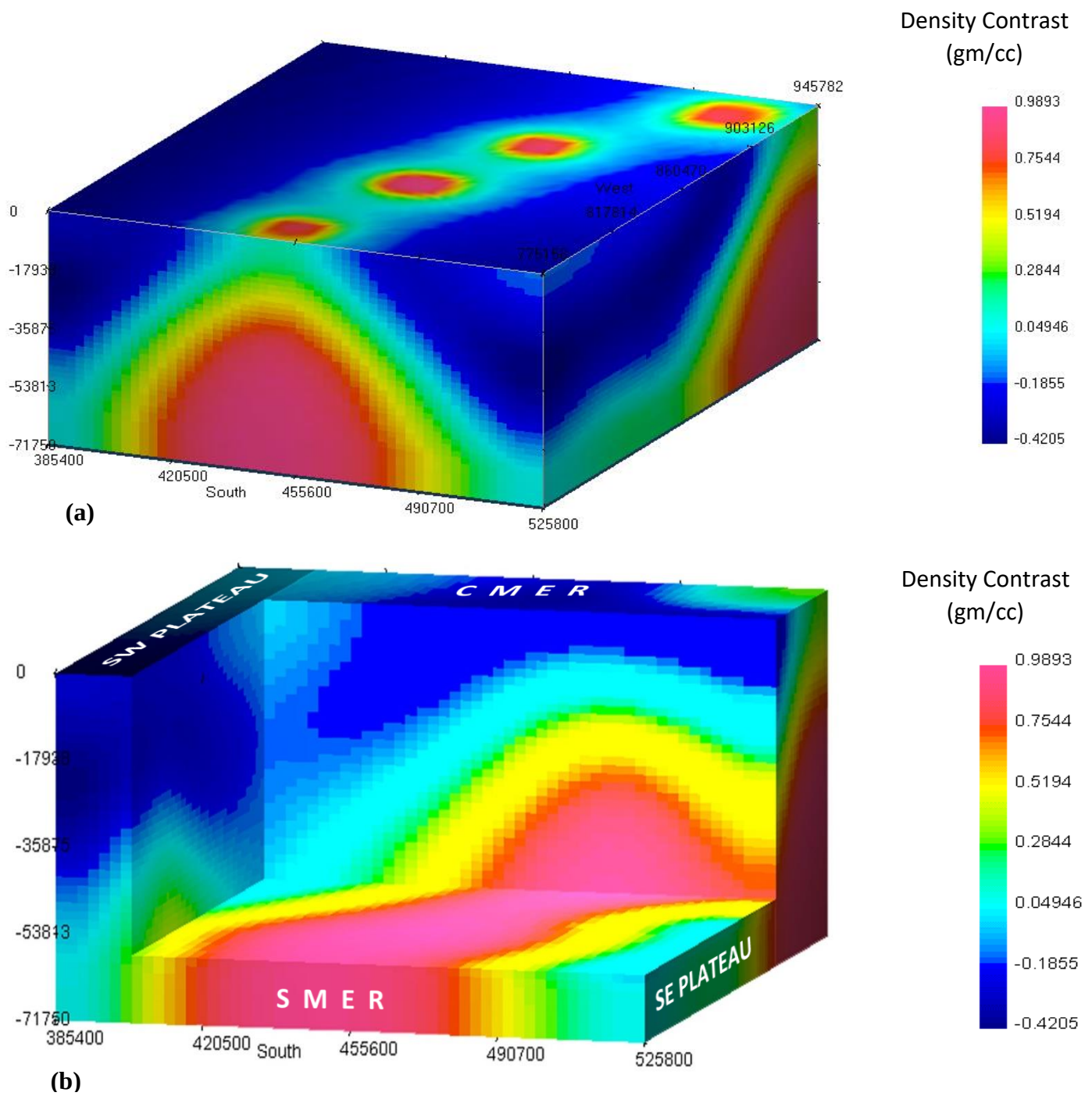


Figure 6.6. (a) 3-D Density model recovered from 3-D inversion of synthetic gravity data and
(b) 3-D volume rendered image of the recovered density model.

The recovered 3-D density model is displayed by applying iso cutoff values (Fig 6.7) of 0.793 for minimum density contrast and 0.981 for maximum density contrasts. The displayed density model is just a total sum of discrete cells whose values are within the specified range mentioned above.

Moreover, as expected the iso surface (Fig 6.7) displays maximum density contrast that is revealed along the rift axis and is located at greater depths. These positive density contrasts indicate that the background density is lower than the density of the cells. The positive (maximum) regional isostatic anomaly is more or less concentrated on the floor of the rift axis (Fig 5.6). This could be due to the presence of a thinned crust beneath the floor of the rift, excess mass and magmatic intrusions from upper mantle are causing this positive density contrast.

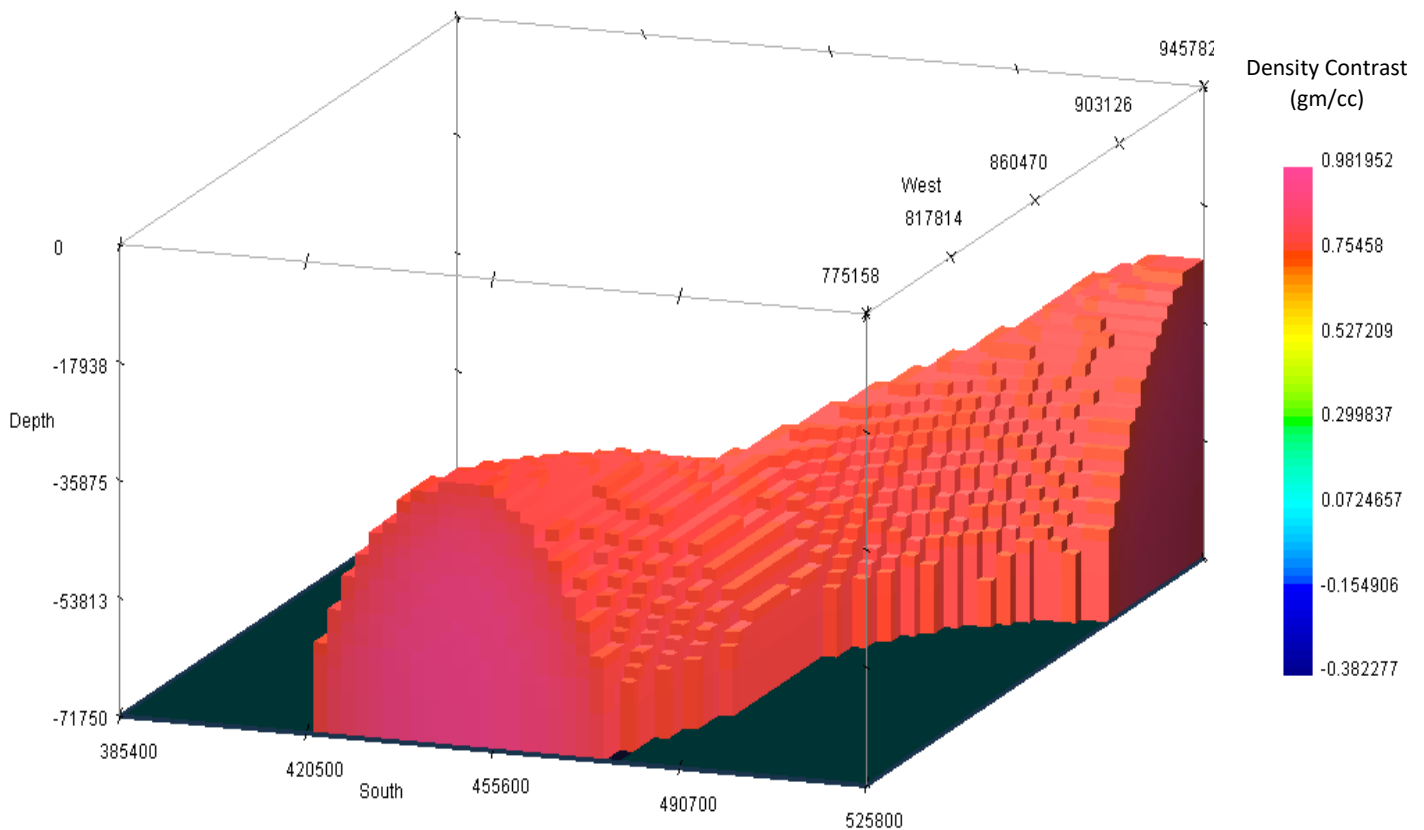


Figure 6.7. 3-D density model displayed as ISO surfaces (min 0.793, max 0.981).

The 3-D density model is further displayed by applying iso cutoff values (Fig 6.8) of 0.471 for maximum density contrast and for -0.199 for minimum density contrasts. The negative density contrasts (blue shaped density models) indicate that the background density values are higher than the density values in the cells recovered by inversion. These regions are mainly the shoulders of the rift floor. Whereas the positive density contrasts (red to yellow shapes) indicate, the background density value is low and these regions are revealed over the axis of the rift floor.

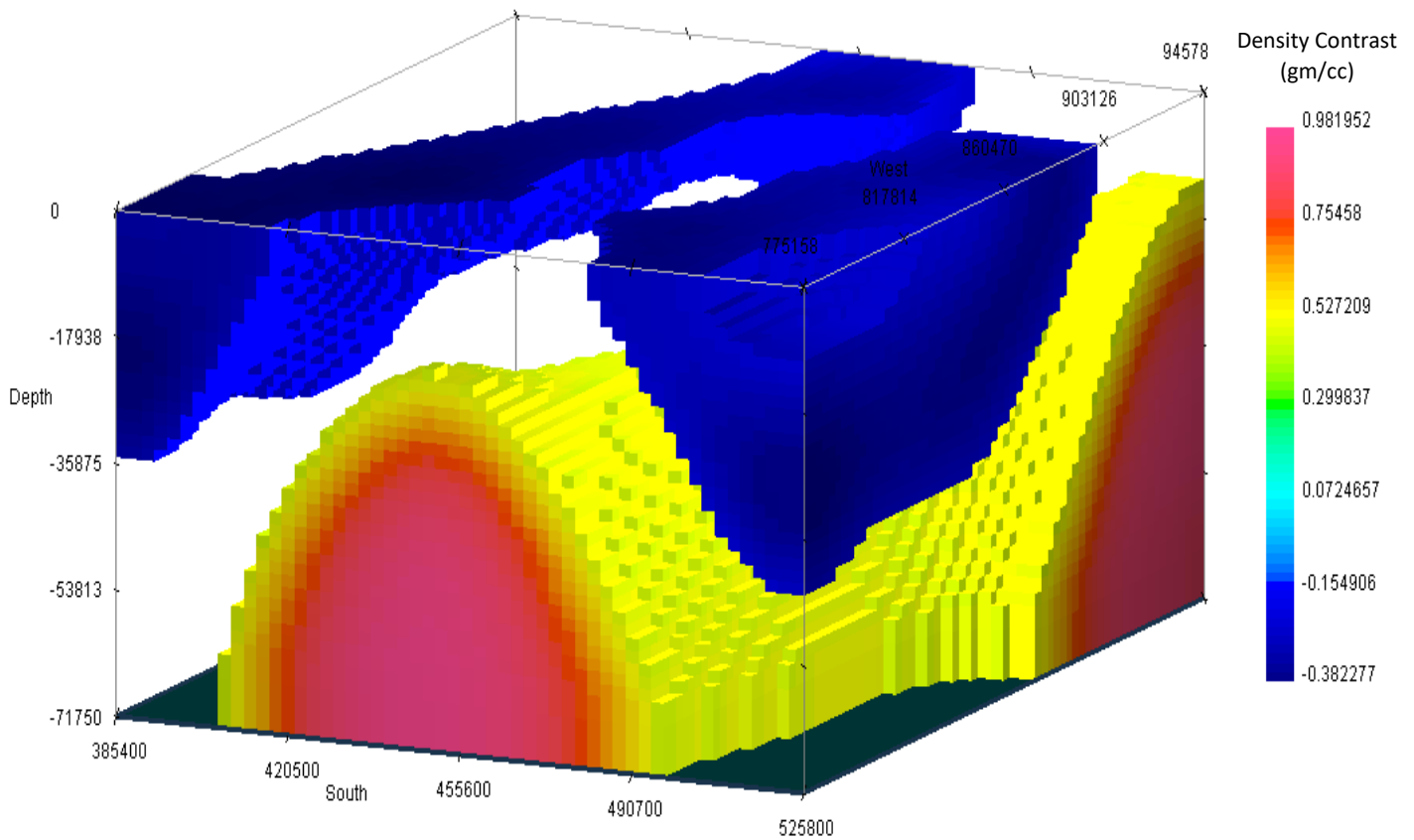


Figure 6.8 .3-D density model displayed as ISO surfaces (max0.471, min -0.199).

Once the ability to forward model and 3-D synthetic gravity data inversion is tested by recovering 3-D density contrast model from the synthetic data the next step is performing the actual 3-D inversion of real gravity data. For the inversion of the gravity data the necessary input parameters, control files and inversion codes were prepared.

The real gravity data 3-D inversion first begins by setting up the model objective function (Equation 6.1) and data misfit function (Equation 6.2). Moreover, the solution to the inverse problem is obtained by mathematical optimization process. Accordingly, the inversion is first run by using the default alpha and length parameters. Additionally, the default distance weighting was used during the first inversion. However, for the subsequent inversions it was changed to depth weighting in order to counter act the natural decay of potential field gravity data with depth.

After a number of inversion iterations are performed by changing the parameters in the model objective function an acceptable data misfit is finally reached and 3-D subsurface density model was recovered which is fairly thought to satisfy and legitimately reproduce the observed geophysical data.

The final parameters set in the model objective function were the following values:

$\alpha_s = 0.0004$, $\alpha_x = \alpha_y = \alpha_z = 1$, $\beta = 0.4$, corresponding length scales $L_e = L_n = L_z = 100$, lower and upper density bounds (-2.0, 2.0 gm/cc). Moreover, topography data was also included in the inversion.

6.2.4. 3-D Gravity Inversion Results

The final 3-D gravity inversion result constrained by the density values obtained from seismic study (Table 6.1) recovered 3-D subsurface density model (Fig 6.9) is displayed in 3-D volume and volume rendered views. The recovered 3-D density model shows the lateral and vertical distribution of density in the central main Ethiopian Rift (CMER). In order to better observe the subsurface density variations, the recovered model is also displayed in slices. The recovered density model shows the range of density values in (gm/cc).

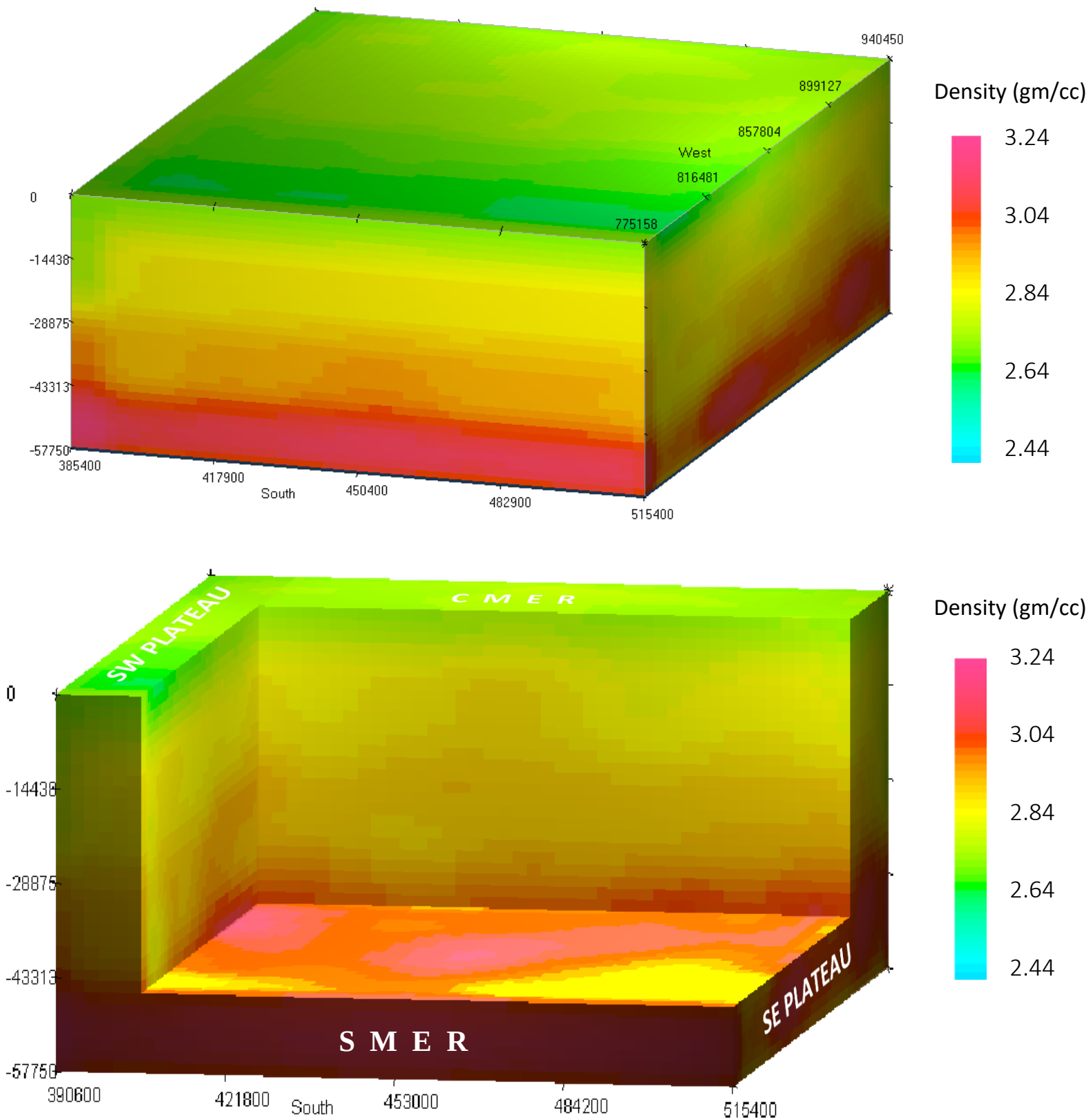


Figure 6.9. 3-D volume and 3-D volume rendered Density model of the CMER

UBC-GIF inversion algorithms can construct 3-D models of density contrasts, so interpretation requires an appropriate display for studying the resulting structures. A 3-D cube of densities can be displayed in sectional or planar cuts or as a 3-D image after volume rendering (Fig 6.9). Diagonal and vertical planar slices applied to the recovered model (Fig 6.9) are displayed in Figures 6.10 (a) and (b). The recovered model diagonally sliced from the bottom SW margin to the NE end of the study area (Fig 6. 10 a) shows the lateral and vertical subsurface density variations beneath the CMER following the protoridge axis of the rift. The second sliced (Fig 6.10 b) part of the recovered model is a superposition of different slices sliced at different vertical depths superimposed on top of each other.

Similarly, the recovered density model sliced cut out from the east and south sides of the model in order to produce the 3-D volume planar image are also shown in Figures 6.11 and 6.12.

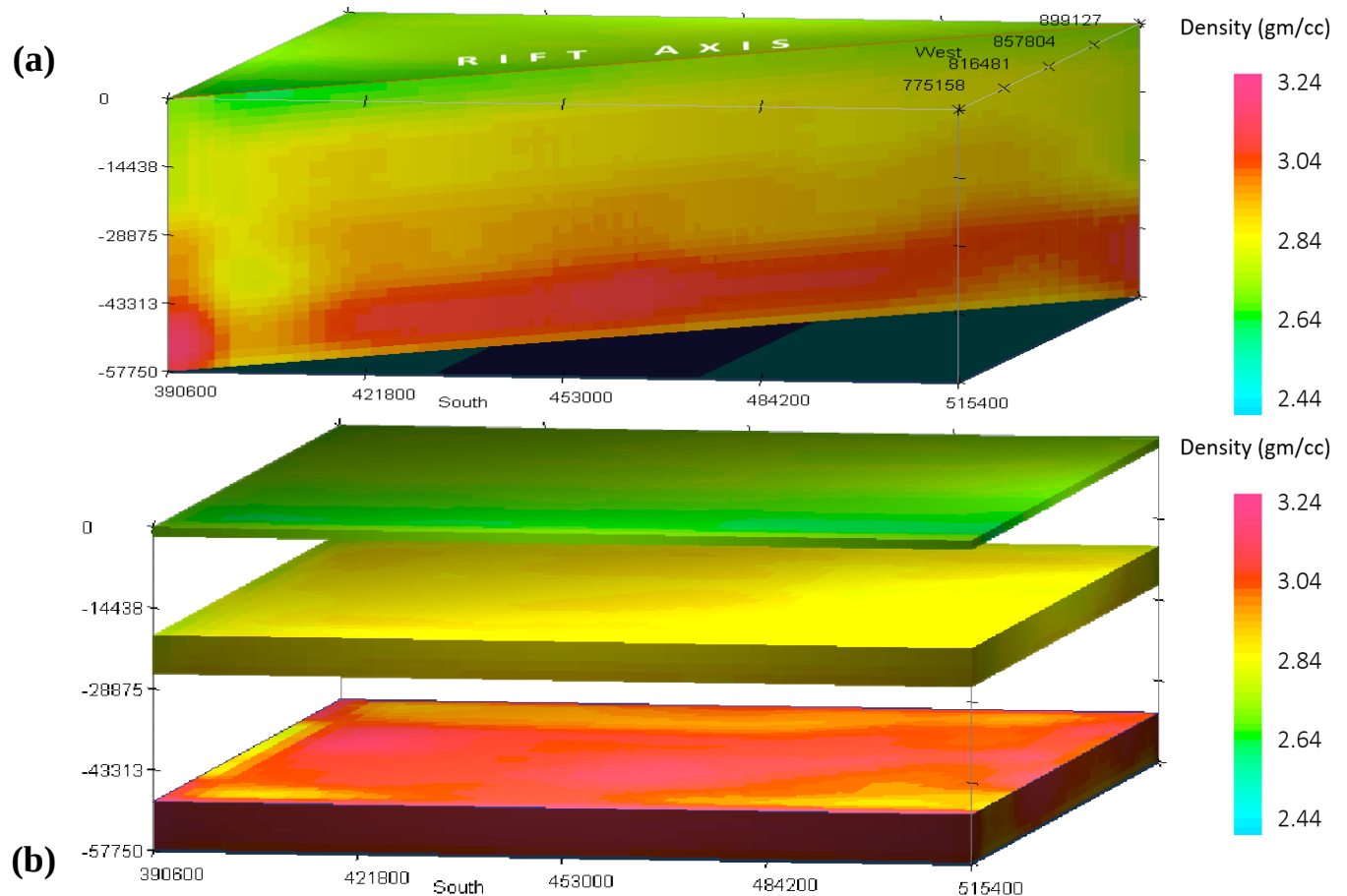


Figure 6.10 (a) Diagonal Slice of the recovered density model (b) Vertical slices shown at different depths.

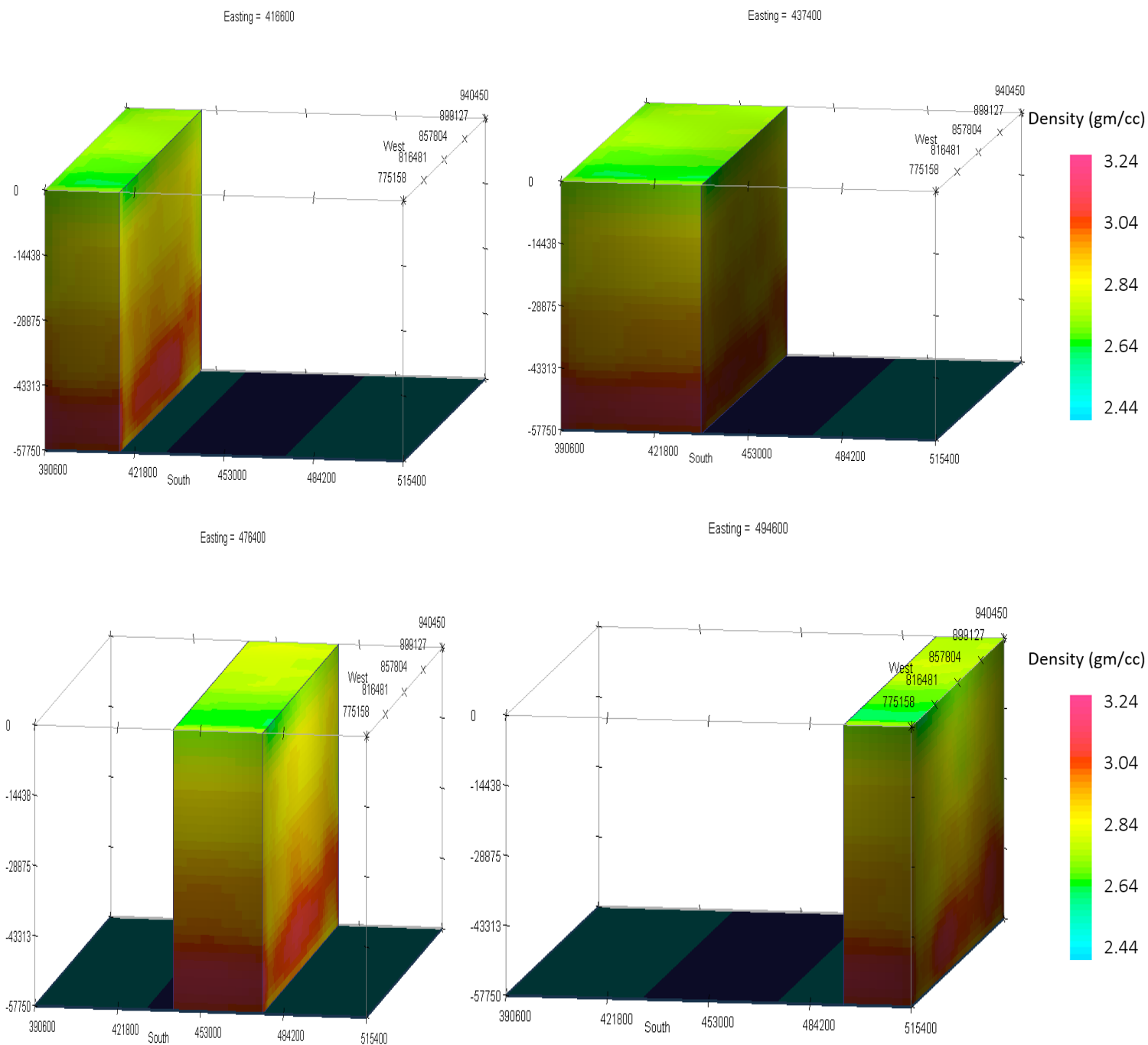


Figure 6.11. Slices applied to the 3-D Density model at lines 416600E, 437400E, 476400E and 494600E.

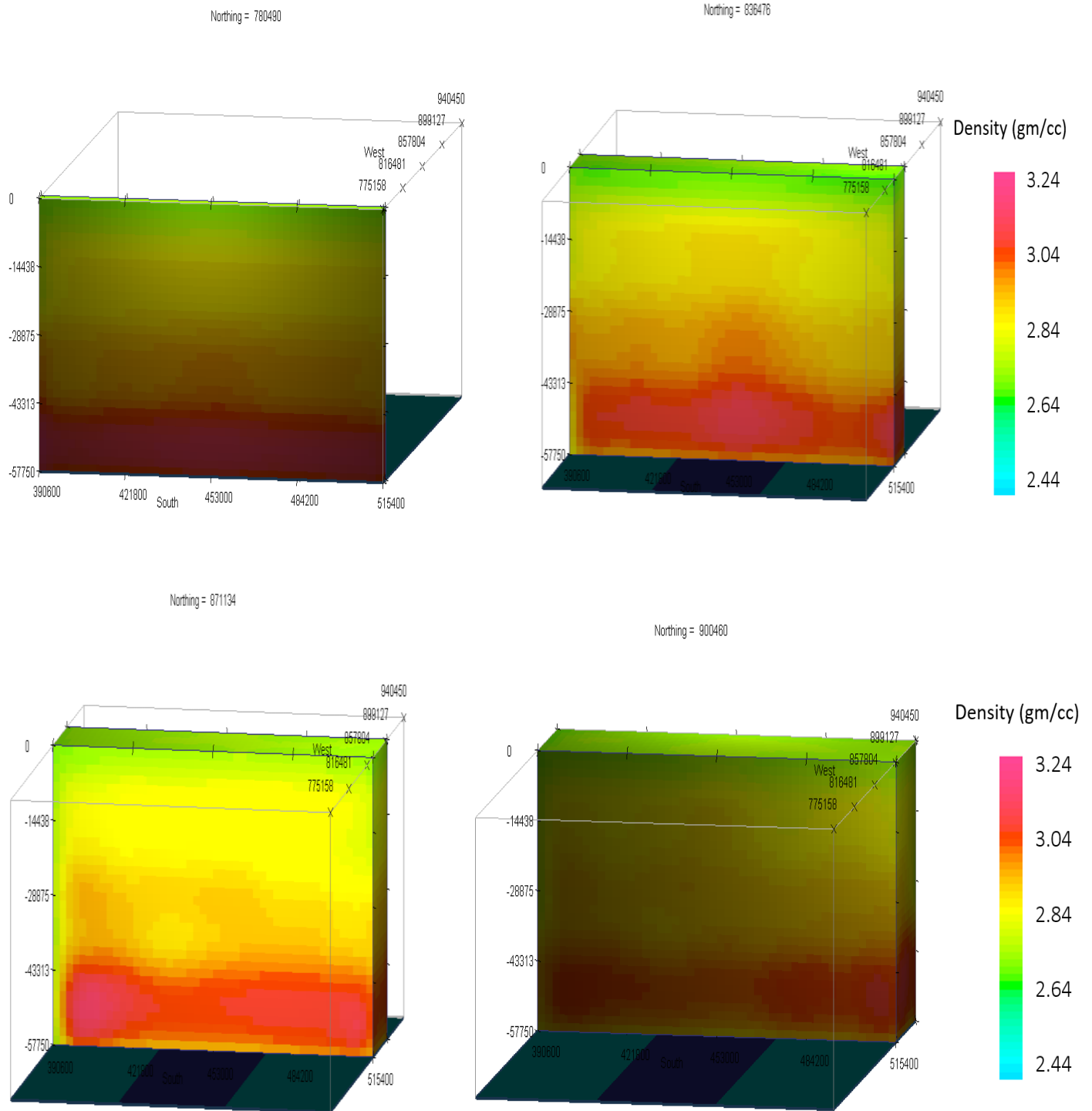


Figure 6.12. Slices applied to the 3-D Density model at lines 780490S, 836476S, 871134S and 900460S.

The computed misfit between the observed and predicted data obtained from the 3- D constrained gravity inversion is less than 4 mGal and the absolute error in is also determined to be less than 1.2.

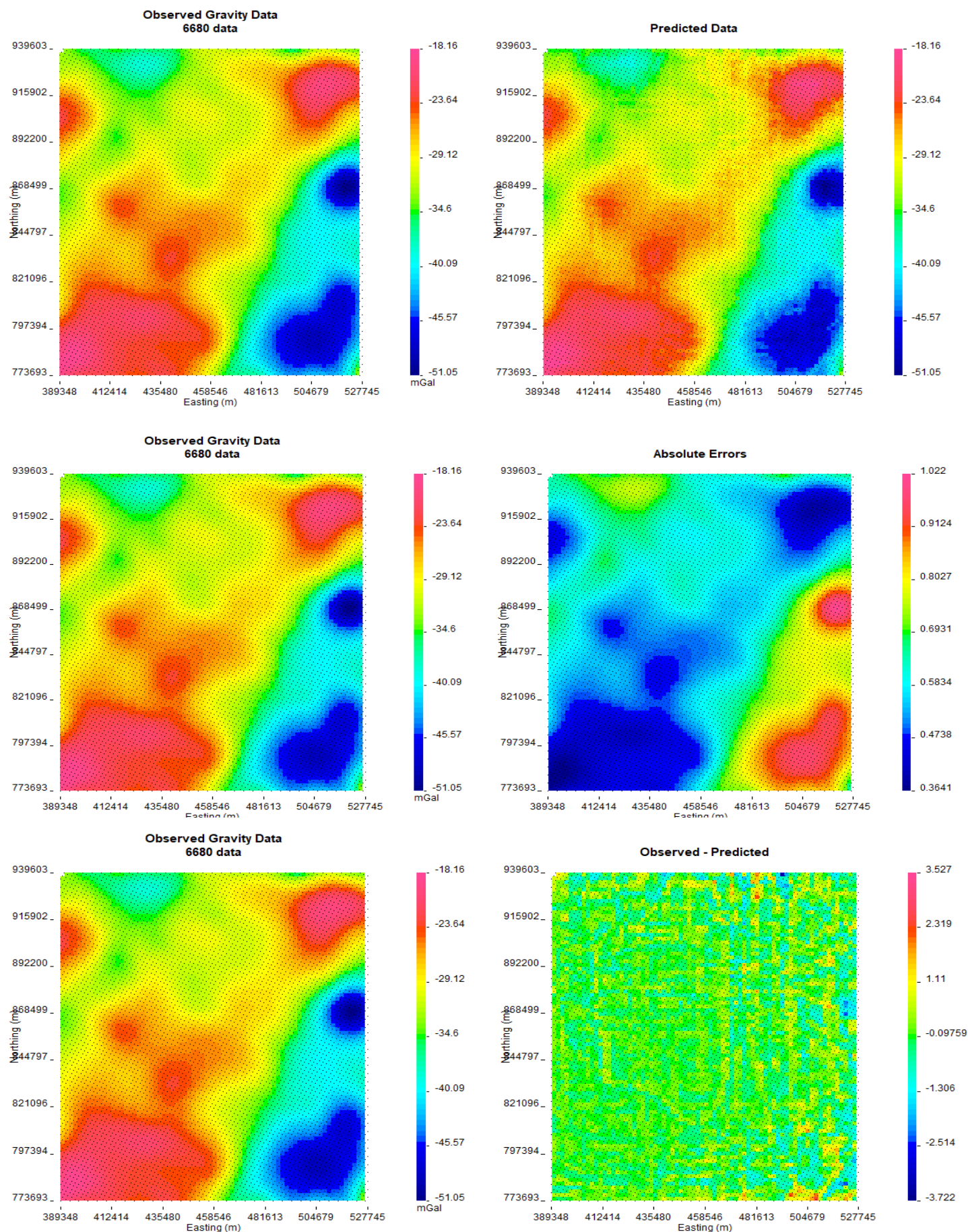


Figure 6.13. The misfit and absolute error between observed and predicted gravity data.

6.2.5. 3-D Inversion of Magnetic Data

Magnetic data inversion involves recovering subsurface susceptibility distributions from the observed magnetic data. The 3-D inversion generally follows the same procedure as gravity data inversion. The process begins by preparing the data, error estimations, setting up the model objective function, the data misfit function and finally solving the inverse problem by minimizing objective function shown in equation 3.51. Accordingly the 7113 extracted magnetic anomaly data (Fig 6.1) were inverted using the algorithms of LI and OLDEBBURG , 1996,. The mesh designed (Fig 6.2) was used for the inversion. After a number of iterations are performed by changing the parameters in the model objective function an acceptable data misfit was finally obtained and 3-D subsurface susceptibility model was recovered. The misfit between the observed magnetic data and the predicted data from the 3-D inversion result is considerably small.

The model objective function Parameters used for the 3-D magnetic inversion were $(k_0, \alpha_s, \alpha_x, \alpha_y, \alpha_z) = (0, 0.0004, 1.0, 1.0, 1.0)$, the reference susceptibility was assumed to be zero, the initial model was set to zero, upper and lower bounds were imposed between one and zero, recovered susceptibilities were assumed to be positive, topography data over the area was included in the modeling. The 3-D magnetic inversion process took a relatively higher CPU run time as compared to the gravity inversion.

Moreover, the model objective function was standard with the addition of a depth weighting function which is based upon the natural decay of dipolar magnetic fields. This is a crucial aspect of the objective function since it causes the recovered susceptibilities to be distributed with depth. Without it, the susceptibilities tend to accumulate near the surface rather than being distributed throughout the model's volume.

6.2.6. 3-D Magnetic Inversion Results

The recovered 3-D susceptibility model (Fig 6.14) is displayed as a 3-D volume of susceptibility. The question of how to examine the result is not trivial. Slices show values of susceptibility at all cells on the slice. Isosurface images show the shape of the susceptibility structure as a surface of constant susceptibility.

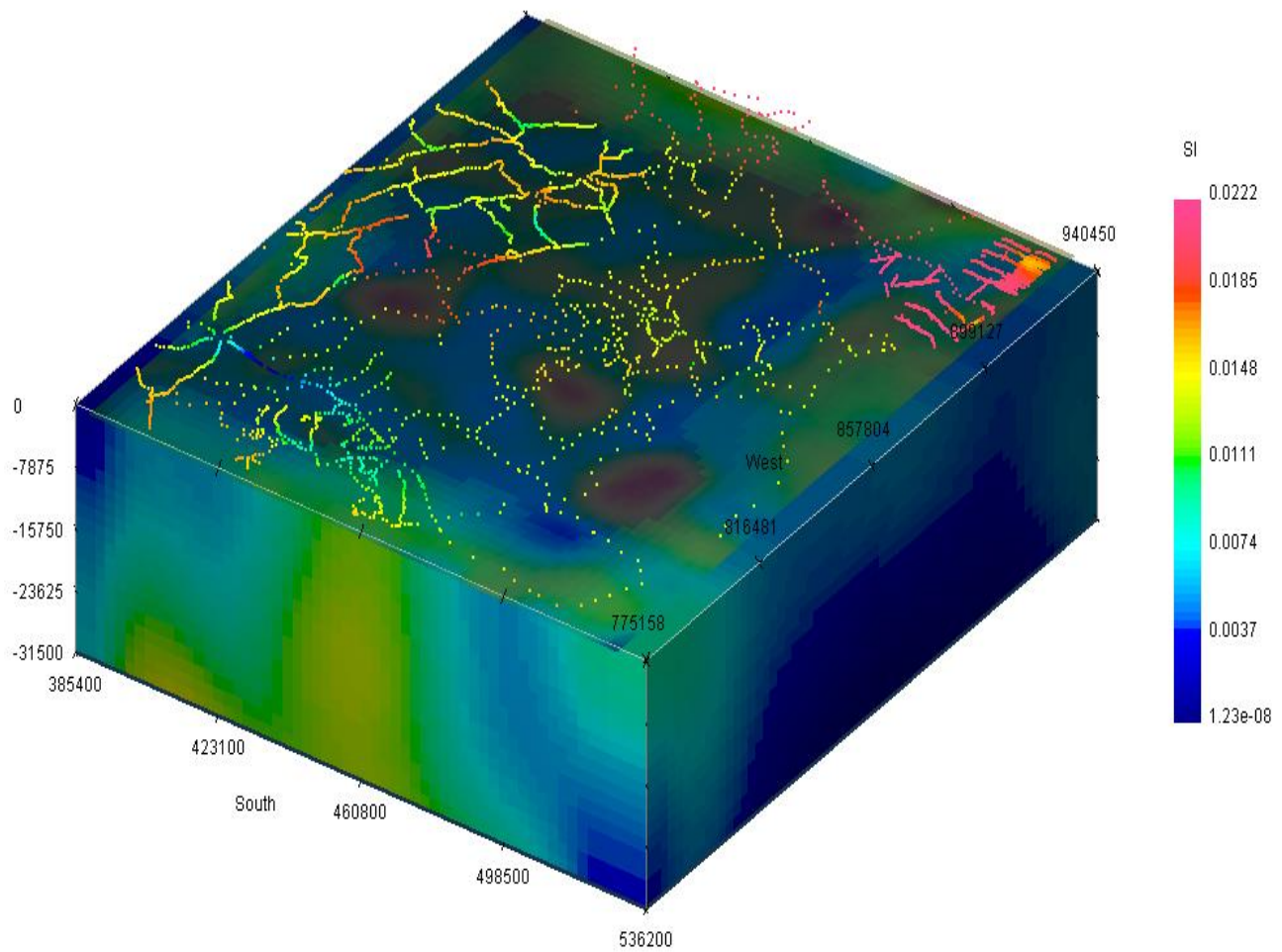
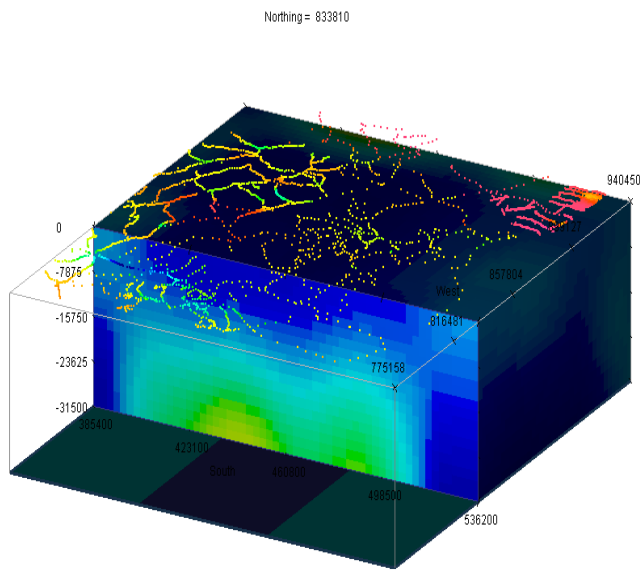
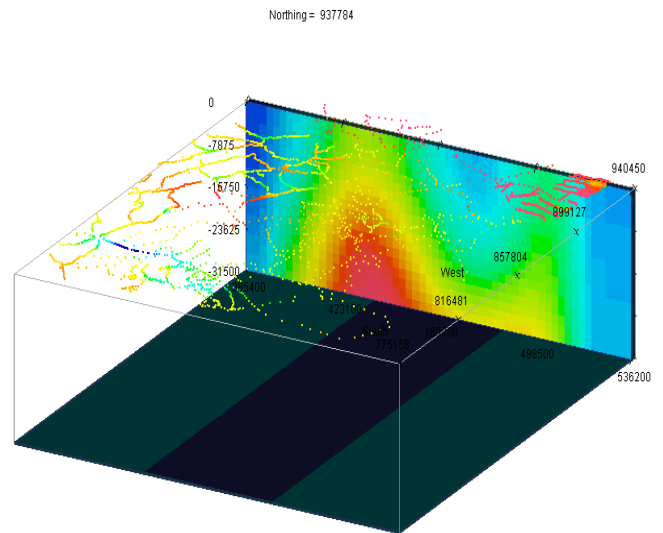


Figure 6.14. Recovered 3-D susceptibility model (points on the top the 3-D susceptibility model show the magnetic data stations).

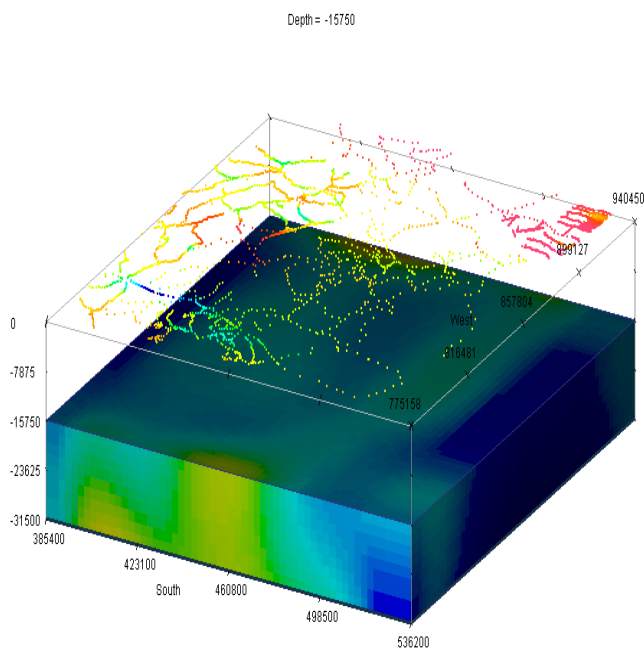
The recovered 3-D susceptibility model (Fig 6.14) is sliced at different lines in order to observe the susceptibility regions at different locations.



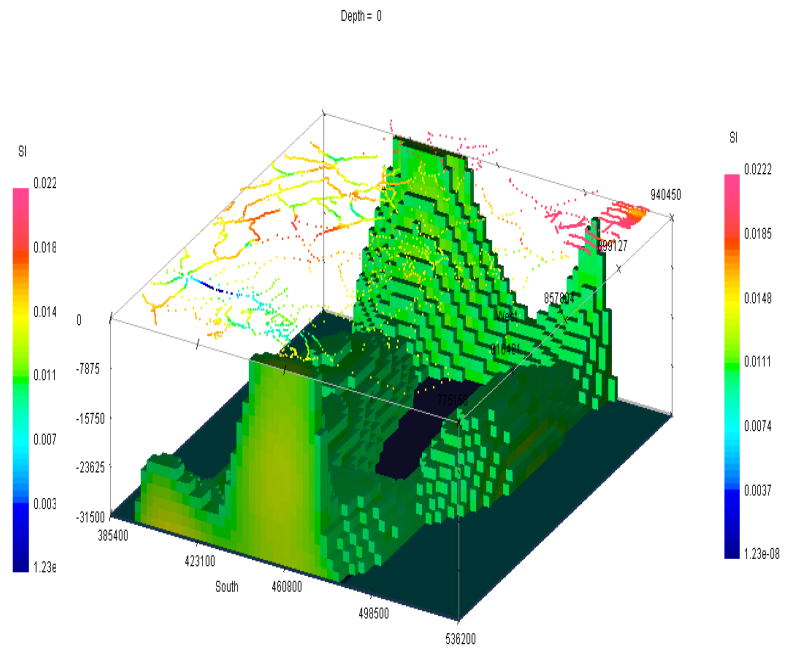
(a) Slice at line 833810N



(b) Slice at line 937784N

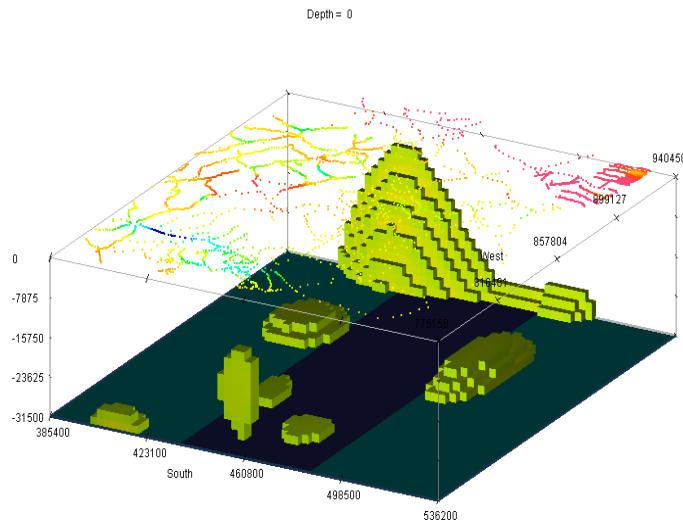


(c) Slice at at 17500 km depth.

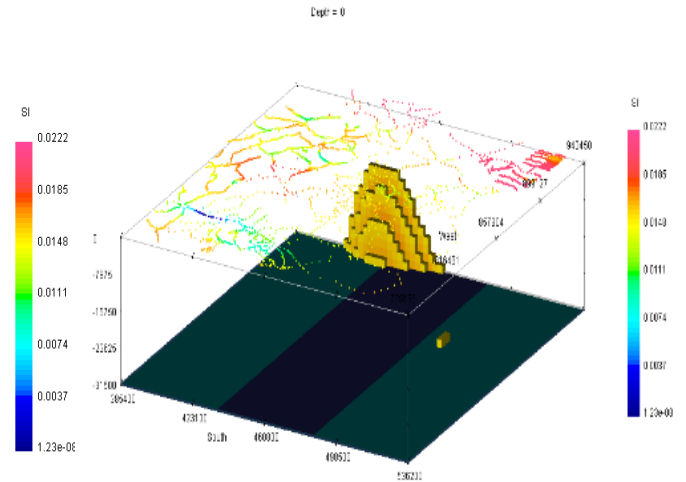


(d) ISO surface 0.01 (SI units)

Figure 6.15. 3-D susceptibility model presented at different slice lines and ISO surfaces



(d) ISO surface 0.013 (SI) units



(e) ISO surface 0.015 (SI) units

Figure 6.16.3-D susceptibility model presented at different ISO surfaces

The predicted magnetic data from the 3-D magnetic inversion and the observed magnetic data are found to be in a good match with 5% Gaussian error contamination of 0.01 nT of floor value. The misfit between the observed and predicted data is generally less than 2.31 nT.

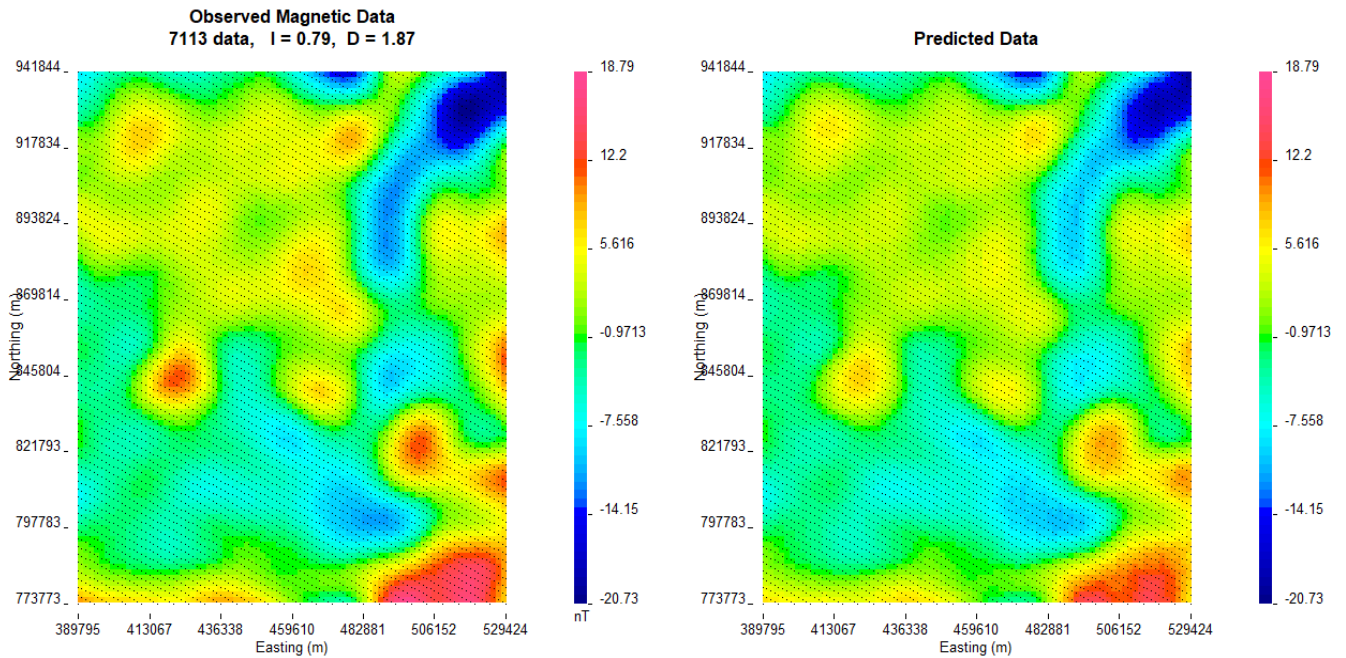


Figure 6.17. Comparison between the observed and the predicted magnetic data from the 3-D magnetic inversion

6.3 Interpretation of 3-D Gravity and Magnetic Inversion Results

6.3.1 Interpretation of Gravity Inversion Results

Generally successful inversions provide useful information about the geologic or geotechnical questions (OLDENBUG, 2005). However, it is evident that inversion tools are by no means simple "black box" techniques. The data supplied, the details of implementation, and making appropriate use of results, all require a certain degree of understanding about all the steps of the process. Like all technically complicated procedures, the likelihood of obtaining useful results from the investment in time, technology and expertise will depend upon how skillfully and professionally the whole process is managed. With this regard maximum effort and considerable time is spent on adapting the inversion codes and implement them to crustal study investigations.

Accordingly, with all the limitations faced from different dimensions which include lacks of detailed a priori information in order to construct a complicated and 3-D geological reference model, this MSc thesis research work has maximally tried to capture the key characteristics needed to reproduce the geophysical gravity data beneath the study area through 3-D inversions using the constraints obtained from seismic studies in the CMER.

According to MAMMO, (2013) a 3-D geological reference model is expected to include geological facts including orientations (strike, dip, plunge and depth) in each discrete cell within a mesh (MAMMO, 2013).

The 3-D density model recovered from the 3-D inversion is shown in Fig. 6.9. The 3-D density model, presented there as volume, diagonal slice, planar slices, 3-D volume rendered image and ISO surfaces are thought to permit visual inspection of the 3-D volume. Moreover, the lateral and vertical density values in the crust are displayed in gm/cc. The recovered 3-D density model is thought to have legitimately reproduced the observed gravity with minimum absolute errors. The misfit between the observed gravity values and those predicted from the inverted density anomaly model is shown in Fig 6.13. The misfit between the observed and predicted data is generally less than 4 mGal and the absolute error in is also determined to be less than 1.2.

The ISO surfaces produced by applying cutoff values (Fig 6.7 and 6.8) show the range of density contrast values in the specified range superimposed each other. The displayed ISO surface density reveals that the maximum density contrasts is concentrated along the axis of CMER.

Closer examination of the recovered density models (eg. Figs 6.9, 6.10 and 6.11) in combination with the regional isostatic anomaly map (5.6) highly assists the interpretation process.

The regional gravity isostatic anomaly map (Fig 5.6) reveals the rift floor mainly the rift axis is associated with positive(maximum) anomaly values where as the shoulders of the rift are associated with negative (minimum) isostatic anomalies.

The positive isostatic anomaly values (Fig 5.6) are correlated with the effects of deep sources with the existence of positive density variations over the rift floor which is associated with effects of high anomaly sources (high-density mantle material) intruding into the crust. Moreover, this phenomenon implies the presence of a thinned crust (a crust where a high-density mantle materials or magmatic intrusions have replaced the crust) beneath the rift floor. Furthermore, the Moho thickness of the rift floor is determined to be smaller than the thickness of the rift shoulders (Fig 5.8).

The 3-D gravity inversion results (e.g Figs 6.6, 6.7, 6.8) also reveal the presence of maximum positive density contrasts beneath the rift floor. The positive density contrasts indicate that the background density value is low and the high density magmatic intrusions intruding from the upper mantle are causing the positive density contrast.

The regional gravity isostatic anomaly map (Fig 5.6) also reveals that the shoulders of the rift floor are associated with negative minimum anomaly values. The negative regional isostatic anomaly values are associated with the effects of low-density crustal root material embedded in high-density mantle material and hence the density contrast over these regions is found to be negative. Besides, the low isostatic anomalies revealed are an indication for the presence of a thick crust beneath the plateau regions. The depth to the Moho of the shoulders of the rift floor is also determined to be deeper than the thickness of the rift floor (Fig 5.8).

The 3-D gravity inversion results (Fig 6.6, 6.9, 6.10) also reveal negative density contrasts in the shoulders of the rift floor. The negative density contrasts indicate that the background density value is high which is thought because of the low density crust subducted (embedded) in to the high density upper mantle causing a negative density contrast.

The 3-D constrained gravity inversion result (Fig 6.9) shows the existence of hot upper mantle material beneath the rift floor. The 3-D inversion results obtained in this MSc thesis are in a good agreement with the conclusions given by MAHTSENTE et al (1999).

6.3.2 Interpretation of Magnetic Inversion Results

Inversion of magnetic data highly suffers from the presence of remanent magnetization (or remanence) which is a permanent magnetization obtained in the past that can often be oriented in a direction different from the Earth's field today. Hence, the induced and remanent components can be oriented in different directions. Typical magnetic inversion methodologies such as that of LI and OLDENBURG (1996) assume no remanent component exists. Hence, the magnetization is assumed to lie in the direction of the Earth's field and erroneous results can be obtained if this assumption is made incorrectly (LELIEVRE, 2009).

As compared to the 3-D gravity inversion presented in this study the a priori information obtained to constrain the 3-D magnetic inversion were almost null and hence the inversion was performed by setting the reference model to zero. However, with the expense of the powers of the UBC-GIF algorithms, the recovered 3-D susceptibility model is thought to capture regions of high and low susceptibilities. Moreover, in order to tackle the non-uniqueness of the inversion process all recommended and available methods were included while setting up the model objective function and defining the error in the data. With this regard depth weighting was included in the inversion process. Another uncertainty imposed on the magnetic inversion is the quality and resolution of the magnetic data itself. This is because some of the magnetic data (1601 points) obtained from previous students of AAU were collected at different scales. Moreover, these data were also found to be collected little earlier years ago. Therefore, the nature of the variation of the earth's magnetic field with time will complicate the 3-D magnetic inversion process.

The results of the 3-D inversion of surface magnetic data reproduced in the form of 3-D susceptibility distribution, slice lines and ISO surfaces are shown in Figs. 6.14, 6.15 and 6.16.

The recovered model is 3-D volume of susceptibility and is sliced at different lines and cutoff values to display iso surfaces. The Isosurface susceptibility images show the shape of the susceptibility structure as a surface of constant susceptibility. Moreover, the lateral and vertical susceptibility variations in the crust are displayed in SI units. The recovered 3-D density model

has fairly reproduced the observed magnetic with minimum absolute errors. The misfit between the observed magnetic data and predicted data from the 3-D magnetic data inversion is found to be less than 2.31 nT. Moreover, Figure 6.17 shows the comparison between observed and predicted data.

The 3-D susceptibility model slices and Iso surfaces at different lines show concentrated susceptibility highs observed in the central northern and eastern regions.

The average depth to the magnetic anomaly sources as computed from RAPS (Fig 5.21) are in a good agreement with the depth revealed by 3-D magnetic inversion (Fig 6.15 c). Hence, the major regional anomalies (maximum and minimum) are generated by sources occurring at deeper depths beneath the study area.

The maximum anomalies observed over the northern and southeastern parts of the study area are thought to be associated with the occurrence of low susceptibility rock units at deeper depths possibly below 14 km depth as determined from RAPS, Fig 5.21 and the 3-D magnetic inversion (Fig 6.15 c).

CHAPTER SEVEN

CONCLUSIONS AND RECOMMENDATIONS

7.1 Conclusions

3-D inversion of airborne gravity and ground magnetic data in conjunction with the analysis of regional isostatic gravity field studies have been investigated over the central part of Main Ethiopian Rift (CMER).

The investigation has determined the depth to the Moho of the CMER and adjacent plateaus from the analysis of the regional isostatic gravity field. The 3-D gravity and magnetic data inversions have provided a comprehensive 3-D crustal density and susceptibility models of the CMER from the inversions of observed gravity and magnetic data. In order to tackle the inherent non-uniqueness problems of potential field data inversion, the 3-D gravity inversion was attempted to be constrained by using the seismic constraints obtained from the EAGLE controlled source project.

The recovered models are thought to have provided a comprehensive picture of the crust and upper mantle structures of the CMER and adjacent plateaus. Moreover, the results are thought to be useful in understanding the complex structural trends and geodynamic processes occurring over the study area. The 3-D inversion results are also thought to be useful in giving insights into regions lacking seismic data coverage like that of the SMER.

Both the 3-D inversions of gravity data and the analysis of the isostatic gravity field have revealed that the rift floor is associated with positive density contrasts while the shoulders of the rift floor are associated with negative density contrasts. The positive density contrasts (maximum isostatic anomalies) revealed over the rift floor are attributed to the high density mantle material intruding the rift floor and while the negative density contrasts over the shoulders of the rift are associated with the low density crust being embedded in to the high density upper mantle material and hence producing negative density contrasts (negative isostatic anomalies).

The constrained 3-D gravity inversion result had revealed the existence of hot upper mantle materials (intrusions) beneath the rift floor. The intrusions certainly of mantle or crust–mantle origin monitor the main Quaternary rhyolite centers of the rift floor.

The computed depth to the Moho of the CMER revealed that the crust beneath rift floor is thinner and shallower than the shoulders of the rift (SW and SE plateaus). Moreover, the thickness of the crust imaged beneath rift floor ranges between 36 and 38 kilometers while it was between 38 and 44 kilometers beneath the shoulders of the rift. The computed Moho depth values obtained in this MSc Thesis were found to be in a good agreement with the Moho depth values determined by previous studies (KERANEN et al., 2009, MAHATSENTE et al., 1999, EMISHAW et al., 2017 and DUGDA et al 2005).

Moreover, the 3-D gravity inversion result had also depicted the presence of crustal thickness variations occurring over the study area. The crust- mantle interface values as inferred from the 3-D gravity inversion showed that the thickness of the crust beneath the plateau regions is slightly higher than the values determined from the analysis of the isostatic gravity field. The 3-D gravity inversion had also illustrated a general crustal thickness variation (decrement) in moving northwards from the southern part of the study area (CMER). This crustal thickness variation is also revealed in the Depth to the Moho map of the study area as well as the constructed reference model using the seismic constraints.

The analysis and 3-D gravity and magnetic inversion investigation results in this MSc thesis are closely associated with the main structural features. Different gravity and magnetic anomaly enhancement techniques including Source Edge Detection (SED), Radially Averaged Power Spectral Analysis (RAPS) and Euler 3-D Deconvolution were employed to detect anomaly source edges and estimate the depth to the anomaly sources over the study area.

Both synthetic and real world gravity data were tested in full 3-D forward modelling and inversions in order to relate test the significances of the two techniques . The results were compared and were found in correlated with a small misfit.

The ultimate goal of this research was to invert geophysical gravity and magnetic data to recover 3-D distributions of subsurface density and magnetic susceptibility of any magnitude and geometric complexity. The work in this thesis was successful in reaching this goal. The inversion

methods implemented were successful in inverting both synthetic and real world survey data collected over the CMER.

7.2 Recommendations

This MSc thesis research work have implemented 3-D gravity and magnetic inversion codes (LI and OLDENBURG, 1996, 1998) to study the crustal structure over the CMER and adjacent plateaus by recovering 3-D density and magnetic susceptibility model parameters. Accordingly, the following recommendations are given.

Besides their default inherent nonuniqueness problems faced during inversions of potential field gravity and magnetic data, 3-D inversion of the two methods increases the amount of subsurface information that can be extracted from the observed gravity and magnetic data.

With this regard I highly recommend to include 3-D gravity and magnetic inversion investigations in any geophysical exploration task especially in mineral exploration works. However, it should also be remembered that incorporating any available geophysical, geological and geochemical a priori information through constructing a full 3-D reference model should be considered to tackle the non uniqueness of the inverse problem and provide a comprehensive and plausible physical property models.

Successful inversion of geophysical field data is highly judged by two ingredients high quality survey design and data acquisition, as well as high quality inversion of the filed data through setting up clear goals. Hence I recommend that any geophysical data to be collected in the field should be performed with the maximum possible accuracy and quality survey design in order to end up with successful inversion results.

The study area (CMER) was found to be small in areal coverage to provide a more regional crustal image of the MER and hence if future 3-D gravity inversion investigations are to be performed then the areal coverage should be large enough to cover the SMER and SBRZ of Ethiopia so that a more comprehensive 3-D crustal structural density model of the Ethiopian rift system will be generated by compiling with the previous 3-D gravity inversion conducted in the flood basalt province of Ethiopia and the results of this MSc thesis in the CMER.

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