Research article

Depth estimates of anomalous subsurface sources using 2D/3D modeling of potential field data: implications for groundwater dynamics in the Ziway-Shala Lakes Basin, Central Main Ethiopian Rift

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ABSTRACT

Quantitative analysis of potential field data are made in the Ziway-Shala lakes basin over an area bounded by 38°00' E - 39°30' E and 7°00' N - 8°30' N. Most previous geophysical studies in the region under consideration focus on mapping the deep crustal structures and undulation of the Moho depth. Only few studies are targeted at mapping the shallow subsurface structures. The main focus of this paper is mapping geometries of the major lithological and structural units of the shallow subsurface using gravity and magnetic data. The ultimate objective of the research is to understand the hydrogeological dynamics of the region through mapping interfaces geometries. Automatic inversions, 2D joint forward modeling and 3D inversion are the major techniques employed. The 2D Werner de-convolution based on both gravity and magnetic data along the rift axis showed source depths tending to deepen northwards. Source depths estimates determined by Source Parameter Imaging also showed similar tendency. This is further strengthened by the joint 2D forward modeling of gravity and magnetic data which showed the top of the basement is sloping northwards. The result of the 3D gravity interface inversion agrees with results of the above mentioned depth estimation techniques. Finally, the gravity power spectral analysis resulted in two depth estimates, 1.53 km and 2.87 km which approximate the positions of two density interfaces. The shallow depth interface is thought to presumably delineate the low density Fluvio-lacustrine sediments including the rift floor volcanic units and crystalline basement. Our investigation results agree with the results of previous seismic studies which identified low velocity (“sediment-volcanic”) horizon in the rift floor with low resolution. The information obtained with regard to water balance of the basin, salinity level of the lakes and the conceptual hydrological flow model appears to reveal that the groundwater flow in the study region is controlled by subsurface structures, particularly, the mapped interface topographies.

1. Introduction

The nature of heterogeneities of the solid Earth leads to uncertain location of its resources for exploration and exploitation activities to be performed. One way of studying these hidden resources is through characterizing the geology and geologic structures of an area. Geophysical methods are determined to be the well-known and well-developed tools for mapping subsurface physical properties (density and magnetic susceptibility) distributions and geologic structures (Okiwelu et al., 2010; Grauch et al., 2006; Skilbrei, 1991). Subsurface deformational structures could be folds, faults, contacts, fractures, joints and subsurface relief. These structures could be mapped using gravity and magnetic anomalies through the application of different mathematical filtering techniques. The subsurface relief for example could be mapped by defining density/susceptibility interfaces (source-depths) using filters like upward continuation (Kebede et al., 2020), tilt depth (Chen et al., 2016; Salem et al., 2007), 3D Euler Deconvolution (Mammo, 2010;
Keating and Pilkington, 2004), power spectral analysis (Mammo, 2012), 2D Werner Deconvolution (Mammo, 2012), 2D forward modeling (Mickus et al., 2007), 3D structural inversion (Tiberi et al., 2005), genetic algorithm (Montesinos et al., 2016) and artificial neural network approach (Alimoradi et al., 2011). The above mentioned source depth estimations methods are some of the various methods and are chosen differently by different researchers implying that there is no single approach. It is therefore recommended to use more than one method for acceptable estimates of depths and density/susceptibility contrast interface geometries.

A prominent previous geophysical study conducted in study region along the rift axis over the rift floor includes the controlled-source seismic survey conducted by Ethiopia Afar Geo-scientific Lithospheric Experiment (EAGLE) project (Maguire et al., 2006). It includes a 400 km long 2D velocity model along the axis of the Main Ethiopian rift (MER). The main findings of this study includes crustal and sub-Moho P-wave velocity model (Maguire et al., 2006). Though, the study focused on the deep structure, it also mapped the surface layer velocities along the rift axis with low resolution (Maguire et al., 2006). The low layer velocity recorded along this profile (Figure 1, blue color) approximately ranges from 2 km/s to 5 km/s (Maguire et al., 2006). These low layer velocities as compared to layers below were interpreted by Maguire et al. (2006) as “sediment and volcanic”. This low velocity (sediment–volcanic) layer along the profile marked by blue line in the Ziway-Shala lakes basin tends to thicken northwards (Maguire et al., 2006). However, compared to the depth of investigation (> 50 km) and complexity of shallow subsurface structures (presence of several volcanic centers) the result is beyond the resolution of data (Maguire et al., 2006). In conjunction with the EAGLE seismic survey, 2.5 D gravity model were generated to map a preliminary crustal structure (Mickus et al., 2007). The maximum depth of investigation obtained based on these models is about 120 km (Mickus et al., 2007). We consider that these models have mapped the shallow structures of the MER with a low resolution.

This study is devoted to examine the shallow to intermediate lithostructural features of the study area using gravity and magnetic data through mapping source depth location and tracing the pattern of source depth undulations. To carry out this task, the standard quantitative techniques including Power Spectral Analysis, 2D Werner Deconvolution, source parameter imaging (SPI), joint 2D forward modeling and 3D gravity interface inversion are employed.

Figure 1. Controlled-source survey line (zigzag line, blue color) conducted by Ethiopia Afar Geo-scientific Lithospheric Experiment (EAGLE seismic line) project and gravity/magnetic profile line (black line) for source depth estimation along the rift axis in the Ziway-Shala lakes basin.
Summary of the overall activities performed in mapping the layers depths and density/susceptibility contrast interface undulations (topography) is described as a flowchart depicted in Figure 2. It starts by defining the geological problem to be solved; describes the physical principle/equations that govern the measurement of gravity and magnetic data; acquisition of gravity and magnetic data; acquired gravity and magnetic data reduction/correction procedures; gridding and interpolation of the corrected gravity/magnetic data; pole reduction of total magnetic field data with the consideration of low latitude problem; gravity/magnetic anomalies separation into regional and residual components; analyses of residual gravity/magnetic anomalies using automatic Inversion techniques, joint 2D gravity and magnetic modeling and 3D gravity structural inversions are made to estimate depth and geometries of the source interfaces.

1.1. Location of the study area

The Ziway-Shala Lakes Basin is found in the Main Ethiopian Rift (MER) being bounded by 38°00’E-39°30’E and 7°00’N-8°30’N (Figure 3a, b). The rift floor, the border faults escarpments and the uplifted plateaus characterize the physiographic features of the study area (Ayenew, 2005). The main lakes situating within the rift floor of the Ziway-Shala lakes basin are Ziway, Abiyata, Langano and Shala (Figure 3b). These lakes are remnants of a large fresh-water lake that existed during the Early-Mid Holocene and the Late Pleistocene (Legesse et al., 2004). Except the Shala Lake and its catchments the remaining three lakes are connected by a surface water network (Figure 3d). Ketar River from eastern escarpments and Meki River from western escarpments are the two major rivers that discharge water to Lake Ziway (Figure 3d). Lake Ziway delivers a majority of its surface water to Lake Abiyata through the Bulbula River (Figure 3d). Five major rivers (Figure 3d) supply their water to Lake Langano and significantly less water is discharged from this lake to Lake Abiyata. Lake Abiyata is the terminal lake (Ayenew, 2005) and has been undergoing significant lake level changes (Ayenew, 2002).

Lake Shala gets its surface inflow from two main sources that enter from the southeastern and western rift escarpments (Le Turdu et al., 1999). The Ziway-Shala Lakes Basin is delimited in the north and north-east by its adjacent Awash River basin, in the south and south-west by Hawasa...
basin, in NW by Ethiopian plateau and in SE by Somalian plateau. The total area of the basin is about 14,640 km² hydrologically closed (Le Turdu et al., 1999; Chernet et al., 2001) and the altitude varies from 1633 m to 3447 m above mean sea level as read from Digital Elevation Model (DEM) map (Figure 3b).

1.2. Geological and structural settings

The geology and geologic structures (Figure 3c) underlying the basin have been determined to be due to volcanic, tectonic and sedimentation processes (Le Turdu et al., 1999). These processes are thought to result in structural depressions forming the Ziway-Shala Lakes Basin (bounded within pink color) (Figure 3c) which constitutes four major lakes, Ziway, Abiyata, Langano and Shalla lakes (Chernet et al., 2001; Le Turdu et al., 1999).

The surface structures generally have N–S to NNE-SSW, NE-SW, E-W and NW-SE (Korme et al., 2004) orientation and are collectively called Wonji Fault Belt (WFB) (Mohor, 1962), Silti Debre Zeyete Fault Zone (SDFZ) and corresponding boundary faults (Boccaletti et al., 1998). The WFB trending N–S to NNE-SSW is youngest and most active fault system cross-cut by the pre-existing NW-SE Mesozoic Ogaden rift fault (Korme et al., 2004).

2. Datasets and methodology

2.1. Datasets

2.1.1. Gravity data

About 3013 ground based gravity data are obtained from the Geological Survey of Ethiopia and from Alemu (1992). These data were reprocessed, homogenized to the IGSN71, reduced using the Geodetic Reference System 1967 (GRS67), corrected and gridded to generate the Bouguer gravity anomaly map of the study area (Figure 4a).

The Bouguer gravity anomalies are the sum of the effect of shallow (high frequency) and deep (low frequency) causative sources which need to be separated into their respective components. The low frequency
(regional) component is approximated by 6 km upward continuation of the Bouguer gravity anomalies (Kebede et al., 2020) (Figure 4b) and the corresponding residual gravity anomaly component (Figure 4c) is calculated by subtracting the estimated regional (Figure 4b) from the gridded Bouguer gravity anomalies map (Figure 4a). The estimated residual gravity anomalies are subjected to source depths estimation for the region under study.

The residual anomalies map (Figure 4c) contains short, intermediate and long wavelength anomalies and it is this anomaly that is subjected to different filtering techniques and modeling.

2.1.2. Magnetic data

About 263 total intensity primary magnetic field measurements are collected by the principal author using Proton Magentometer. These data are merged with 592 secondary total intensity magnetic field data obtained from MSC thesis works of various researchers (Kelemework, 2016; Berhane, 2015; Kebede, 2014). The total of these (855) data are collected in years between 2014 and 2018 and the distribution of the data are shown in (Figure 5a).

The standard magnetic data correction including correction for main field is made for the IGRF 2005 epoch. This data is gridded and mapped to generate total magnetic field anomaly map of the study area (Figure 5b).

The survey area is located at low magnetic latitude where the inducing field direction is horizontal and smaller in strength. These complicates interpretation of the magnetic anomaly data (Hansen and Pawlowski, 1989) and resulting in source and anomaly mismatched. This is solved using differential reduction to pole (DRTP) (Gupta and Ramani, 1980) operator acting on total magnetic anomaly. The differentially pole reduced total magnetic anomaly is shown in (Figure 4c).

The differentially pole reduced total magnetic anomaly need to be decomposed into regional and residual anomalies which are respectively related to deep and shallow subsurface geology. The regional anomaly is approximated using upward continuation of the pole reduced total magnetic anomaly to height 6 km (Figure 5d). The final differentially pole reduced residual magnetic field is shown in Figure 5e.

2.1.3. Well-log data

The well log dataset consists of core-samples obtained from eight boreholes drilled for the purpose of studying the geothermal resources of the Aluto-Langano geothermal field. These wells depth ranges from 1300 m to 2500 m below surface ground level. Based on these drilling data information the geologic section was constructed along west-east oriented profile by different researchers (Cherkose and Mizunaga, 2018; Wilks et al., 2017; Hutchison et al., 2016; Electroconsult (ELC), 1986). The section stratigraphy from top to bottom given as Silicic products;

Figure 4. Gravity station location distribution overlay Bouguer gravity anomaly of the Central Main Ethiopian Rift and NNE-SSW extending profile line (a) 6km upward continued Bouguer anomaly map which approximate the regional anomaly (b) residual gravity anomaly map (c).
Figure 5. Magnetic data stations distribution map (a) Total field magnetic anomaly map (b) Total magnetic field anomaly reduced to magnetic pole using DRTP (c) Regional magnetic anomaly estimated using upward-continuation filter to height of 6 km (d) the residual magnetic anomaly map (e) of the Ziway-Shala Lakes Basin, Central Main Ethiopian Rift.
Lacustrine sediments; Bofa Basalt, tuff and breccias and Tertiary Ignimbrite layers (Table 1). This geological cross-section information is used for constraining and cross-checking analysis of the gravity and magnetic data. Moreover, the geologic section is used to constrain (used as an initial model) the joint 2D forward gravity/magnetic model and 3D gravity interface based inversion (detail of this stratigraphy in 3D is shown in Figure 6 and Table 1, section 2.2.6).

2.2. Methodology

An inversion process generally includes an automatic (Werner, 1953), Fourier Transform method which includes spectral and 3D interface based inversion (Oldenburg, 1974; Parker, 1972; Spector and Grant, 1970) and optimization (voxel) methods (Williams, 2008; Li and Oldenburg, 1998). These inversion methods are used to invert potential field data into a 2D/3D densities/susceptibilities subsurface spatial distribution or subsurface topographic undulation that generate observed gravity/magnetic anomalies. Except for the voxel based inversion method, all the inversion methods mentioned above are employed in this research work. The governing equations for each method are described in sections 2.2.1 to 2.2.5 below.

### 2.2.1. Power spectral analysis

The application of power spectral analysis to potential field anomalies in one dimension can be performed by transforming the digitized gravity/magnetic data from space domain in to frequency domain. The transformation is used to compute the power or amplitude spectrum in accordance with Eqs. (1) and (2) below

\[
F_w = \sum_{k=0}^{N-1} f(k)e^{-i\omega k}
\]  

(1)

\[
F_w \text{ is the discrete amplitude spectrum which could be respectively written as a sum of real and imaginary (Eq. 2)}
\]

\[
F(\omega) = a(\omega) + ib(\omega)
\]

(2)

Where

\[
|F(\omega)| = \sqrt{a^2(\omega) + b^2(\omega)} \text{ is an amplitude spectrum and } \omega = 2\pi f \text{ is the angular frequency}
\]
Spectral analysis is thus describe the variation of the energy as a function of frequency (wave length) which help to estimate subsurface density contrast depth (Spector and Grant, 1970) given by (Eq. 3)

$$h = \frac{1}{4\pi} \left( \frac{\log E_1 - \log E_2}{K_1 - K_2} \right)$$

(3)

Where

$E_1$ and $E_2$ are power spectra of the gravity field, $\log E_1$ and $\log E_2$ are logarithms of the power spectra $K_1$ and $K_2$ are wave numbers 

his depth to interfaces (layer boundaries)

Eq. (3) gives the depth, $h$, which are obtained from difference of the power spectrum curve slopes divided by $-4\pi$.

The method is a statistical estimation (Spector and Grant, 1970) based on low and high frequency anomalous bodies. These anomalies respectively categorized as sources of low, intermediate and high frequencies characteristics (Kebede et al., 2020).

In this study, the profile data along the rift axis (Figure 1), black line, is extracted from the residual gravity anomaly and is inverted for sources depths at various density interfaces using power spectral analysis mentioned. The depths can be determined along the line in accordance with Eqs. (1), (2), and (3). This method was used by various researchers to estimate depth to density interfaces (Mammo, 2012). The profile data and the result of the analysis (depths of density contrasts) using this depth estimation method are documented in result section of this manuscript.

2.2.2. Werner De-convolution

The mathematical formulation of 2D Werner De-convolution was developed by (Werner, 1953) and followed and extended by (Hartman et al., 1971). Accordingly, the governing equation of a total magnetic field for simple arbitrarily oriented thin sheets (dykes) or interfaces is given by (Eq. 4)

$$F(x) = \frac{Ah + B(x - x_0)}{(x - x_0)^2 + h^2}$$

(4)

Where,

$x$ is the distance along a line perpendicular to the strike of the thin sheet, relative to an arbitrary origin; 
$F(x)$ is the total magnetic field intensity at $x$; 
$h$ is the depth to the top of the thin sheet; 
$x_0$ is the position of the top, projected vertically to intersect the line; 
$A$ and $B$ are parameters related to the magnetic properties and the thickness of the thin sheet as well as its orientation relative to the direction of the Earth’s field.

For the more complex model, the anomalous magnetic field is considered to be the combination of the fields due to two thin sheets and a quadratic background magnetic interference which can be expressed as (Eq. 5)

$$F(x) = \frac{A_1h_1 + B_1(x - x_1)}{(x - x_1)^2 + h_1^2} + \frac{A_2h_2 + B_2(x - x_2)}{(x - x_2)^2 + h_2^2} + c_0 + c_1x + c_2x^2$$

(5)

Eq. (5) has 11 unknowns $A_1, h_1, B_1, x_1, A_2, B_2, h_2, x_2, c_0, c_1$ and $c_2$ of which the first eight variables are important to the interpretation and it is solved analytically. The equations defining the model are solved using observed data values which yield the position, depth, and magnetic/ gravity contrast of buried interfaces. To solve for these unknowns at one point on the profile a ‘window’ of 11 equi-spaced observed data values are required. Solving these systems of equations, results in estimates of these coefficients. This intern allows one to estimate the position and depth of the thin-sheet bodies and geological interference.

This method is an automatic inversion method which does not require prior information. The method generates subsurface point source distribution in 2D.

In this study, profile residual gravity and magnetic anomalies along rift axis (in parallel to EAGLE seismic line) are first extracted respectively from gridded gravity and magnetic anomaly maps. Werner method is applied to these profile dataset to estimate corresponding depths to gravity and magnetic source interfaces. The method was used by different researchers in different study area to estimate depths to gravity and magnetic sources (Mammo, 2012; Thakur et al., 2000).

2.2.3. Source parameter imaging (SPI)

The method is based on complex analytical signal (Eq. 6) which is defined as either in terms of total field or its Hilbert transform (Nabighian, 1972)

$$A(x, z) = \frac{\partial B(x, z)}{\partial x} - j \frac{\partial B(x, z)}{\partial z}$$

(6)

Where,

$$|A| = \sqrt{\left(\frac{\partial B}{\partial x}\right)^2 + \left(\frac{\partial B}{\partial z}\right)^2}$$

is analytical signal amplitude and

$$\theta = \tan^{-1} \frac{\partial B}{\partial x} / \frac{\partial B}{\partial z}$$

is the local phase.

-x and z are Cartesian coordinates for the vertical direction and the direction perpendicular to strike, 
-B($x, z$) the magnitude of total magnetic field and 
-j is the imaginary number

The method according to Thurston and Smith (1997) can also be expressed as the local frequency $f$, given by (Eq. 7)

$$f = \frac{1}{2\pi} \tan^{-1} \left( \frac{\partial B}{\partial x} / \frac{\partial B}{\partial z} \right)$$

(7)

In the analysis of potential fields, it is often more convenient to use local wave number, denoted by, $k$, rather than $f$ which is given by (Eq. 8)

$$k = 2nf = \frac{1}{|A|^2} \left( \frac{\partial^2 B}{\partial x \partial z} + \frac{\partial^2 B}{\partial y \partial z} \right)$$

(8)

Thus, the peaks outline source edges (local depth) can finally derived at these location; $x = 0$, is (Eq. 9)

$$h = 1/k$$

(9)

The depths to magnetic sources in the region under consideration are automatically estimated using Source Parameter Imaging (SPI) based on governing equations (Eq. 7, Eq. 8 and Eq. 9). Oasis Montaj Geosoft software version 7.1 SPI tool is used and the solutions are saved in a database.

The work is started with calculating derivative grids ($\frac{\partial B}{\partial x}, \frac{\partial B}{\partial z}$ and $\frac{\partial^2 B}{\partial x \partial z}$) of differentially pole reduced total magnetic field anomalies which are used as an input to SPI operator. The calculated SPI values are then gridded and mapped to show the magnetic source depths of the study area. This method has been used as magnetic source depths estimates by different researchers in different study locations (Tsepav, 2018; Salako, 2014).

2.2.4. Forward modeling

It describes the process of generating synthetic data (Menke, 1989) from a given Earth model with geometric elements and physical properties as an initial model.
The gravitational attraction \( g \) at a point \( P = (x, y, z) \) due to a volume \( V \) can be expressed in line integral (Telford et al., 1990) as (Eq. 10)

\[
g(P) = -G \int \rho \frac{\hat{r}}{r^3} \, dv
\]  

(10)

Where,

- \( \rho \) is density as a function of space
- \( r \) is the distance from the observation point \( P \) to an element of the body \( dv \),
- \( \hat{r} \) is a unit vector pointing from an element of the mass to \( P \), and
- \( G \) is the universal gravitational constant

The vertical component of the gravity anomaly simply expressed (Talwani et al., 1959) as (Eq. 11)

\[
\Delta g_i = 2G \rho \sum_{i=1}^{n} \Delta g_i
\]  

(11)

Talwani et al. (1959) considered the case of an n-sided polygon and broke the line integral up into n-contributions, each associated with a side of the polygon. He derived expressions for \( z_i \) that make extensive references to trigonometric functions which is clearly stated in his paper (Talwani et al., 1959). The model involves creating a hypothetical geologic model and calculating the gravity response to that earth model.

In this study, Oasis Montaj Geosoft modeling program called GM-SYS is used for intuitive, interactive and real time calculation of the profile residual gravity/magnetic response with change of density, susceptibilities and undulating geometries of layers. This modeling process is non-unique and thus need to be constrained with the present and earlier geological and geophysical knowledge. This method is an established procedures of estimating profile geometries of the interfaces since Talwani et al. (1959).

2.2.5. 3D layered based inversion of gravity data

This is another spectral inversion method based on the spectral content of the data (Parker, 1972) and (Oldenburg, 1974). It helps to map the depth to an interface separating two homogenous media using the sum of Fourier transform of gravity data and the sum of the interface topography's Fourier transform of the interface topography

\[
F[h(r)] = \frac{F[\Delta g_i(x)]e^{-ikr}}{2\pi G \Delta \rho} \sum_{n=1}^{\infty} \frac{k^{n-1}}{n!} F[\hat{h}^n(r)]
\]  

(13)

Where:

- \( F[ g(r) ] \) is the Fourier transform of the gravity anomaly,
- \( G \) is Newton's gravitational constant,
- \( r \) is a horizontal plane
- \( \Delta \rho \) is the density contrast across the interface (two media),
- \( k \) is the wave number,
- \( n \) is an integer
- \( h(r) \) is the topography of the interface
- \( z_0 \) is the mean (reference) depth of the horizontal interface.
- \( F[\hat{h}^n(r)] \) is Fourier transform of the interface topography

Parker-Oldenburg iterative expression (Eq. 13) allows us to calculate the topography of the density interface (layer) iteratively using \( \Delta \rho \) and \( z_0 \) as an input. The first term of Eq. (13) is computed by assigning \( h(r) = 0 \). This value, \( h(r) \), is then used in the equation to evaluate a new estimate of \( h(r) \). This process is continued until a reasonable solution is achieved and the convergence criterion is guaranteed.

In this study the joint algorithm of Parker-Oldenburg inversion (Eq. 12 and Eq. 13) is applied on residual gravity anomaly grid (with constraining initial model compiled in section 2.2.6 and previous knowledge) in the region considered to map the geometries of the crystalline basement layer that is responsible for the observed residual gravity anomalies. The method has been used for crustal structures studies (Tiberi et al., 2005) and estimate of bedrock topographic undulations (Salimi and Motlagh, 2012).

2.2.6. Initial model

Both joint gravity/magnetic 2D forward modeling and 3D gravity structural inversions require a priori constraining information for a final model to be reasonable and acceptable with least tolerable error. These constraints are compiled from well-log data information, seismic velocity model (Maguire et al., 2006) and the depths determined from the spectral analysis, Werner deconvolution and SPI (sections 2.2.1 to 2.2.3). The parameters obtained from these data analysis are density, susceptibility, interfaces and corresponding causative source depths. In geological section described in section 2.1.3, there is no Mesozoic and crystalline basement surface till depth of 2.5 km (Cherkose and Mizunaga, 2018). The gravity data analysis in the study area show that the crystalline basement horizon is found at approximate mean depth of 3.0 km (Kebede et al., 2020). In-between Tertiary ignimbrite and crystalline basement layers, however, there is Mesozoic layer (Woldegabriel et al., 2000). To determine the depth to the top of the Mesozoic layer, we performed a Multi-Layer 3D Gravity forward calculation with all the other interface depths set at predetermined location, varying the depth to the top of the Mesozoic layer (i.e., 2.6 km, 2.7 km, 2.8 km and 2.9 km) and recording how sensitive the misfit is as we vary the depth values considered. The final result which generates the least root mean square error (RMS) is taken as the feasible depth to top of Mesozoic layer. The depth to top of this layer is at 2.9 km as the RMS is smaller than others. The interfaces lithology, density and susceptibility value identified come from various sources. The summary of these parameters which are used as an initial model is described in Table 1. Graphical representation of this initial model is shown in Figure 6.

3. Results

The mathematical formulation of Power Spectral Analysis, 2D Werner De-convolution, Source Parameter Imaging (SPI), 2D joint forward modeling and 3D structural inversion are described in the methodology sections. The first three methods are automatic methods that do not require a priori geological information as compared to the last two. The gravity and magnetic source depths estimated along the rift axis and over rift floor. Figure 1 shows a profile line (black color) where the residual gravity and magnetic anomalies are extracted and analyzed. The zigzag line (blue color) (Figure 1) is a seismic line analyzed and interpreted by the EAGLE project for velocity stratigraphic sections (Maguire et al., 2006).

3.1. Spectral analysis

The profile residual gravity anomaly is extracted along the rift axis and rift floor (Figure 1, along black line) and is shown in Figure 7A. The one dimensional power spectrum technique is applied to residual gravity anomaly profile consisting of 166 data points (Figure 7A). The calculated power spectrum curve is represented as log of spectral energy versus wave number (Figure 7A). The depth to top of the gravity source (density interfaces) is calculated from the slope (gradient) of power spectrum...
Figure 7. Residual gravity profile anomaly along the rift axis that was used in the power spectrum analysis (a) plot of logarithm of the power spectral energy versus wave number decay curve (energy spectrum of the profile) (b) series of trend line segments fitted to energy decay curves to determine slopes (gradients) which could help to estimate average depths to the various density interfaces representing depth to deep (B1), intermediate (B2) and noise (B3) gravity sources.

Table 2. Log spectral depth estimates and category of the anomalous source as deep, Intermediate and noise components.

| Profile Name | Layers | Depth (m) to tops of layers | Causative source categories | Geology approximation from well-log | $r^2$ |
|--------------|--------|-----------------------------|-----------------------------|-----------------------------------|-------|
| South-North  | -36094 | -2872.27                    | Deep                        | Top of crystalline basement layer  | 0.848 |
|              | -19318 | -1537.28                    | Intermediate                | Top of boa basalt layer           | 0.5617|
|              | -297.39| -23.6655                    | Noise                       | -                                 | 0.0002|
curve. The linear curves (from which slope values are read) are fitted to power spectral data based on a piece-wise least-squares linear curve-fitting approach (Figure 7B1, 7B2 and 7B3). The gradients (slopes) are read from the fitted lines which show weighted energy of the source origin from low, intermediate and high frequencies values. Much of this energy comes from the deep sources (low frequency) (Figure 7B1).

The results of the power spectral analysis are summarized in Table 2. Source depths interfaces are categorized as shallow, intermediate and deep depending on the wave number (frequencies) and slope (gradient) variations. The depth to intermediate source along the rift axis and rift floor approximated from residual gravity anomaly is about 1.5 km while the depth to deep gravity sources is 2.87 km. The third slope is nearly representing the noise in the data (Figure 7B3).

Three gradient values are read from the fitted trend lines (Figure 7B1, 7B2 and 7B3) (Table 2, column 2) and depths are calculated based on Eq. (3) and shown in column 3 of Table 2. As the method is used to determine mean depths to the various interfaces of density contrasts, two clearly identified interfaces are read and categorized as deep and intermediate source depths (Table 2, column 4). The depth estimate for the third interface is about 23 m which is estimated with $r^2$ value of 0.0002. Accepting 23 m as a source depth with $r$-square value could lead us to wrong interpretation. Thus, it would be better to consider it as noise component. Validity of the estimated top of density contrasts (Table 2, column 4) could be guaranteed by comparing the gravity source depth results against the geologic stratigraphy (Table 1, Figure 6) produced based on well-log data (Cherkose and Mizunaga, 2018). This source depth estimation method was used to estimate density horizons in northern part of Ethiopia (Mammo, 2010), which lies close to our study area.

Figure 8. Elevation map along the study profile and its trend line fit (a) residual gravity anomaly (b) Werner depth showing Z-contact calculated using 2D Werner de-convolution and trend line fitted to Z-contact along the rift axis (c).
3.2. Werner De-convolution

The gravity/magnetic source depth and location along the rift axis and rift floor are calculated based on an automated depth-estimation method called Werner Deconvolution. The governing equation which helps to make this depth estimate is described in section 2.2.2. This method is an iterative 2D inversion technique which estimates the vertical gravity/magnetic sources depths and locations. The residual gravity anomaly along the rift axis (Figure 8b) is first extracted from residual gravity anomaly map (Figure 4c). This profile anomaly (Figure 8b) is subjected to Werner algorithm to recover gravity source vertical contact depth locations. This is shown in square symbols plotted in Figure 8c). Similarly, the differentially pole reduced residual magnetic anomaly along the same profile line as gravity profile is extracted and subjected to Werner magnetic profile depth algorithm. The result of magnetic source depth is shown in Figure 9b.

The elevation profile along the rift axis was extracted from the DEM (Figure 9a). The trend line fitted to this elevation profile show a negative slope signifying that the elevation tends to decrease northwards. The gravity source depth estimated based on Werner automated inversion algorithm appears to deepen northwards as depicted by the fitted trend line to the gravity source depths (Figure 9c). Similar to the trends from gravity source and elevation, the magnetic source depths also appear to deepen from south to north (Figure 9b).

3.3. Source parameter imaging (SPI)

This is a technique used to estimate the depth to the anomalous magnetic sources which is equivalent to mapping thickness of the sedimentary-volcanic section overlying crystalline basement (magnetic basement). SPI is generally an automatic calculation of source depths from gridded magnetic data (Thurston and Smith, 1997). The method is independent of magnetic inclination, declination and pole reduction, however, the SPI operator is acted on differentially pole reduced magnetic residual anomaly. The target of the filter is to estimate depth to magnetic sources in the Ziway-Shala Lakes Basin. Oasis Montaj Geosoft software version 7.1 is used to generate a SPI value. These values are then gridded and mapped for the region under study (Figure 10a) to show magnetic source interfaces.

Density plot of SPI depth values are made to show the depth distribution (Figure 10b) of the magnetic sources. The depth to magnetic source body after gridding and mapping is depicted as 2D image map (Figure 10a). The estimated depths lie in between 1.1 km and 4.7 km. The gridded SPI value along the rift axis is categorized as “Region A” and “Region B” (Figure 10a) respectively showing sources of magnetic bodies from deep and shallow Earth’s. Crystalline basement rocks are believed to be the source of magnetic anomalies. Along the rift floor, the maximum source depth of about 4.7 km is found in northern water divide, between the Ziway-Shala Lakes Basin and Awash River Basin. The rift floor generally shown to indicate magnetic sources situated at shallow depths (Figure 10a). The reason for existence of magnetic sources at shallow depth could be because of rocks thermal alteration with depths. As the study area lies in volcanically and tectonically active area, the heat sources increases with depths which decreases rocks magnetic susceptibilities.

3.4. Joint 2D gravity/magnetic forward modeling

The joint gravity/magnetic model can be considered to better reflect geology of the study area as compared to the independently generated 2D
gravity and magnetic models. The 2D joint gravity and magnetic model is computed using the residual gravity/magnetic profile based on the initial constraining information listed in Table 1. The initial model consists of six Lithologic horizons namely Pyroclastic-volcanic formation (2.477 g/cc), followed by lacustrine sediment formation layer (2.34 g/cc), Bofa basalt, tuff and breccias (2.81 g/cc), Tertiary ignimbrite layer (2.58 g/cc), Mesozoic sediment formation layer (2.5 g/cc) and crystalline basement formation layer (2.74 g/cc) (Table 1, Figure 6).

The final joint gravity and magnetic model response is taken to be acceptable with a root mean square error (RMS) of 1.61 mGal and 25.568 nT respectively (Figure 11). This joint model gives a good fit between observed and calculated gravity and magnetic anomalies along the profile.

The thickness of sediment-volcanic overlying crystalline basement increases northward as it is shown in Figure 11. This result agree well with the seismic velocity model result though "sediment-volcanic" layer was mapped with low-resolution (Maguire et al., 2006). These depth estimates recovered from the joint gravity/magnetic 2D modeling are consistent with the automated depth-to-source estimates made using Spectral Analysis, 2D Werner De-convolution and Source Parameter Imaging methods.

3.5. 3D gravity structural inversion

The 3D forward and inverse modeling approach is defined by a number of surfaces or geologic horizons inverted for mapping the anticipated depth to layers (Figure 12). The multi layer, surface based, frequency domain forward and inverse modeling of Parker–Oldenburg algorithm is applied to transform the residual gravity anomaly to crystalline basement depths. An apriori information derived from existing geological and geophysical studies (Table 1) are required for carrying out this inversion. It is performed for layers geometries (litho-stratigraphies) in general and crystalline basement layer in particular. Accordingly, the inversion is carried out on crystalline basement surface with the reference depth set at -3 km and layer density set to 2.74 g/cm³. Top of this layer is considered to constitute Mesozoic sediments with a density value of 2.5 g/cm³. The mean density contrast at the interface between crystalline basement and Mesozoic sediment is thus 0.24 g/cm³. The smallest and greatest cut-off

Figure 10. Magnetic source locations calculated using the SPI method (a) point magnetic source depth distributions within the Ziway-Shala Lakes Basin (b).
frequency parameters are chosen as 0.5 and 0.7 km$^{-1}$, respectively. The maximum iteration number is set to 10 with a convergence limit is 0.01mGal. The residual gravity anomaly is inverted for layers undulations and thicknesses. The structural inversion carried on crystalline basement layer is found to converge after four (4) iterations. For the topographic relief calculated (Figure 13d), from the structural inversion the corresponding calculated and misfit gravity anomaly are generated (Figure 13b, c).

The shallow depth to crystalline basement top layer (Figure 13d) is 2.8 km and a deeper top depth to this layer is 4.7 km. These are 200 m up and 1.7 km down from a reference depth 3 km. Along the rift axis particularly starting north of Abiyata Lake, floor of Gademota and extending far north to water divide area (Meki area), the gravity source depth appear to be found at deep depth. Furthermore, it is observed that gravity basement found to deepen northwards, towards Awash Basin.

The region has no previous geophysical studies at shallow/intermediate depth level for comparison of the present result. However, the result is in a fair agreement with 2D Werner De-convolution, SPI and joint 2D gravity/magnetic modeling result performed in this research work. Though mapped with low resolution, the seismic study conducted in the area showing low velocity sediment-volcanic seems to deepen northwards (Maguire et al., 2006), in agreement to the result of 3D inversion.

4. Discussion

4.1. Subsurface structures

The different inversion approaches employed in this research provided depth to various density and susceptibility interfaces. These density/susceptibility layers geometries are closely associated with the boundaries of geological structures (Feng et al., 2016). As documented in Section 3, the sources associated to the gravity and magnetic anomalies are determined using power spectral analysis, 2D Werner Deconvolution, source parameter imaging, joint 2D forward modeling and 3D gravity interface inversion methods.

In this research work, the power spectral analysis is the first inversion method used to map sources resulting in to the gravity anomalies occurring along the rift axis and over the rift floor. The analysis includes determination of slopes of the lines fitted to plots of log of power (energy) versus wave number (Figure 7B1, 7B2 and 7B3). These slope values

Figure 11. 2D joint gravity/magnetic model of the profile running along the rift axis and rift floor, considered to show the gravity/magnetic horizons. The deeper crystalline basement sources overlying shallower sedimentary/volcanic sources.

Figure 12. Residual gravity anomaly (top) inverted for geometrical modification of stacked layers (bottom) particularly for crystalline basement layer.
which are calculated in accord with Eq. (3) and section 2.2.1 are employed to infer depths of the gravity source along the profile considered. The gravity sources that occur at the intermediate depth (1.5 km) are found to possibly correspond to the top of the Bofa Basalt, tuff and breccias layer (Table 1). The gravity source having deeper origin along the profile is estimated to occur at a depth of 2.87 km and this depth is found nearby top of the crystalline basement depth (3 km) identified by Kebede et al. (2020).

Secondly, depth to top of the gravity and magnetic source interfaces along the rift axis are estimated using 2D Werner de-convolution of residual gravity anomaly and differentially pole reduced residual magnetic anomaly profile data (Figure 7a and Figure 8b). Trend lines fitted to Werner estimated gravity and magnetic source depths thought to indicate the deepening of the low velocity “sediment-volcanic” layer from south to north, along the rift axis.

The third filter used to map source depths is SPI. The result obtained from analysis of SPI of magnetic data shows, the “sediment-volcanic” layer overlaying the crystalline basement layer along the rift axis over rift floor is thickening northwards. This means that, thickness of the low magnetic source bodies (sediments) increases northwards. In other words, the crystalline basement rock layer is far deeper in the northern region of the study area than the southern ones.

Since the study area lies in rift environment where volcanic activities are active, the magnetic sources generally shown to be shallow Earth origin (Figure 10a, b). The shallow magnetic sources are more easily detectable than the deep-seated targets as revealed by the SPI symbol depths density map (Figure 10b). Most of these shallow magnetic sources are depicted as category “B” (Figure 10a) with the depths approximately laying in between 1.1 km and 2.6 km. The deepest depth region categorized by “A” (Figure 10a) has an approximate maximum depth of 4.7 km. From depth distribution symbols map (Figure 10b) we observe that, most frequent (concentrated) depth location for magnetic source bodies in the region lies in the depth range of 1 km–3 km.

The magnetic source depth estimated from SPI along the rift axis/rift floor mostly agreed with low velocity “sediment-volcanic” layer identified by Maguire et al. (2006). The result from analysis of SPI agrees with the 2D Werner De-convolution results obtained from both gravity and magnetic data.
The joint gravity/magnetic forward model result shows that the gravity/magnetic basement geometry tends to deepen northwards in the study area. The crystalline basement rocks appear to be the major sources of the anomalies as compared to overlying layers. For example, at the northern end of the profile, low gravity minimum indicate sediment accumulation (Figure 4c) and the corresponding high magnetic value about this location (Figure 5e) is considered to be due to the basement crystalline rocks occurring at relatively deeper depths. Results of the joint 2D forward model fairly agree with the results obtained from the 2D Werner deconvolution and Source Parameter Imaging (Figure 10a). Furthermore, the result of the joint 2D model is in a fair agreement with results of the seismic low velocity model layer of an EAGLE project.

Finally, 3D gravity interface inversion is conducted to map depth and geometries of crystalline basement layer. The inversion results (Figure 13d) have shown that the crystalline basement shallowest depth is found at -2.8 km and extends down to 4.662 km. The crystalline basement topography gets deeper towards north along the rift axis especially near water divide (north of the Meki town) between the Ziway-Shala Lakes Basin and Awash River basin (Figure 13d). The overall result estimated based on automatic inversion, joint 2D modeling and 3D structural inversion shows undulating layers deepening northwards along the rift axis over rift floor. In other words, the “Sediment-volcanic” layer resting over crystalline basement layer is thickening northwards.

### Groundwater flow implications of the mapped structures

The Ziway–Shala Lakes basin is hydrologically closed (Le Turdu et al., 1999) (Chernet et al., 2001) and there is no evidence of significant groundwater outflow (Legesse et al., 2004) from this basin. Using isotopic evidences (Darling et al., 1996) and groundwater flow modeling results the groundwater flow direction is determined to be from Lake Awassa basin (southern basin) towards the low-lying and deep Ziway-Shala lakes basin (Ayenew, 2001). This groundwater flow direction is thought to increase both the groundwater and surface water resource in the basin.

To have a better understanding of the hydrology-hydrogeology, the water balance of the basin (Table 3) is determined with computations.
made based on the hydrological data taken from the work of Ayenew (2002).

According to Le Turdu et al. (1999) and Ayenew (2002) the mean annual rainfall ranges from about 1150 mm in the eastern and western highlands to around 600 mm in the rift floor. The annually weighted actual evapotranspiration values of the rift, the escarpments and the highlands are 656, 892 and 917 mm respectively with an averaged value of 821 mm/year. The annual evaporation values from Abiyata, Langano, Shala and Ziway Lakes are estimated to be 2060, 2010, 2112 and 2022 mm, respectively, with an average value of 2051 mm. These annual evaporation and evapotranspiration estimates are deemed to help calculation of the water balance of the basin.

In the case, the principle of conservation of mass is used to govern the flow of water in and out of a system. Precipitation (P), Evaporation (E), Evapo-transpiration (ET), Surface runoff (SRO) and groundwater flow (GF) are the various components of hydrologic cycle which are related mathematically as (Eq. 14)

$$\Delta S = P - E - ET \pm SRO \pm GF$$

(14)

Where, $\Delta S$ is the change in storage.

On annual basis and under steady state condition where the lakes volume is stable the total inflowing water (rainfall) on the basin should equal to the total water loss (evaporation) ($\Delta S = 0$).

The estimate of water balance of the closed basin (eg. Ziway-Shala Lakes Basin) restated as in the following formula (Eq. 15)

$$\text{Flux}(GF) = P - AET - E_L$$

(15)

Where, $P$ is precipitation, AET is the actual evapotranspiration and $E_L$ is evaporation from lake.

The net total flux of the basin is calculated to be 5265.58 Mm$^3$ which is positive. The positive sign implies that the total precipitation is greater than total evaporation. This positive flux should be reflected on the increase of water capacity in the basin which would be manifested in rise of water level of the lakes. However, observation shows that the actual levels of the Lakes in the basin have been declining from time to time (eg. Abiyata Lake). The conclusion would be made based on the analysis of the water balance of the basin and the nature of the basin is a likely existence of geological structures acting as conduits for outflow of groundwater from the basin.

The second important parameter that characterizes the relationship between geologic structures and groundwater flow is Total Dissolved Solute (TDS). The TDS of the Lakes in the MER depends on inflow and outflow conditions (Gizaw, 1996). Table 4 shows the TDS estimates determined for the six lakes that reside in the MER including Afar.

The salinity levels of Abiyata, Langano, Shala and Ziway Lakes in the MER remain relatively low and stable as compared to those of the closed Lakes in the Afar depression such as Lake Afera and Lakes in Dallol (Table 4). It is confirmed that, salinity increases from catchment divides to the valley floors and in the direction of groundwater flow (Salama et al., 1999). The observed relatively low salinity levels of the Lakes in the Ziway-Shala Lakes basin may thus possibly been thought to indicate a regional groundwater leakage enhanced by subsurface geologic structures.

The third information that support structural control of groundwater flow could be traced from investigation of the general groundwater flow system. Alley et al. (1999) have shown that the groundwater flow in any region can be categorized as shallow (local), intermediate (sub-regional) and deep flow system. According to Alley et al. (1999), the local ground-water subsystem occurs in the upper water-table aquifer that discharge to the nearest surface-water bodies (lakes or streams) and are separated by ground-water divides beneath topographically high area. The intermediate ground-water subsystem in the water-table aquifer in which flow paths originating at the water table discharge into a more distant one. A deep (regional) ground-water-flow subsystem lies beneath the water-table subsystem and is hydraulically connected to local and sub-regional one (Figure 14).

The gravity and magnetic data analyzed in this research work are anticipated to map the subsurface topography over rift floor all along the rift axis in the study area. The mapped subsurface interface geometries resemble the adopted conceptual model shown in Figure 14. The inferred subsurface structures are likely to control the sub-regional to regional groundwater flow of the area considered as they are deemed to satisfy the requirements necessary for groundwater flow on a regional scale. Moreover, groundwater flow through aquifer types associated with rifting environments like that of the study area is likely to be governed by secondary porosity (Morgan et al., 2006) for reasons that active volcanic and tectonic activities induces fracturing.

Comparison of the basin’s water balance, salinity level of the lakes in the basin and the adopted conceptual groundwater flow model indicate that geologic structures control the groundwater dynamics in the study area. Generally, the groundwater dynamics among the inter-basins in the MER are likely to be governed by the subsurface geological structures, in particular layers topography mapped in this work. Therefore, it would be possible to safely conclude that groundwater outflows from the Ziway-Shala Lakes Basin to the Awash Basin. This conclusion supports the previous findings of Abiyu Kebede (2007) who used integrated methods and approaches to map the groundwater flow in the Ziway-Koka corridor indicating a likely groundwater migration from south to north in the corridor.

5. Conclusion

This study has made use of potential field data to recover depth to gravity and magnetic source bodies in the Ziway-Shala Lakes Basin. The automatic inversions, joint 2D gravity/magnetic forward modeling and 3D gravity structural inversion are employed to have a better understanding of lithological layer geometries. Horizons depicting gravity/magnetic source depths are estimated using the mentioned mathematical methods. Profile power spectral gravity depth analysis resulted in two source depths of 1.53 km and 2.87 km. 2D Werner Deconvolution analysis of both gravity and magnetic data shows a source depths deepening northwards, as revealed by the fitted trend line to the geometries in vertical contact. Similarly, the depths to magnetic sources estimated using SPI method range from 1.1 km to 4.7 km and is thought to show deepening (thickening) of the low velocity overburden. Furthermore, the 2D joint gravity/magnetic forward modeling show similar results to those of the 2D Werner Deconvolution and SPI results. The low velocity layer named as “sediment-volcanic” in the seismic study is categorized as six layers earth model in this study. The depth estimates resulting from the automatic methods and gravity/magnetic joint 2D modeling are used to constrain the 3D structural gravity inversion. Accordingly, the 3D gravity interface inversion results obtained based on the Parker-Oldenburg algorithm fairly agree with the results of all the quantitative analysis methods used in this study. The combined interfaces geometries mapped with the various filtering techniques are in close agreement with each other and with the existing apriori geological and geophysical knowledge of the region. Furthermore, the shallow “sediment-volcanic” layers mapped using gravity and magnetic data appear to deepen northwards in consistent with those of the EAGLE seismic study results. The structures (layers topography) mapped in this research are believed to govern the inter-basins groundwater dynamics of the study region.

Declarations

Author contribution statement

Hailu Michael Kebede: Conceived and designed the experiments; Performed the experiments; Analyzed and interpreted the data; Contributed reagents, materials, analysis tools or data; Wrote the paper.
Abera Alemu, Dessie Nedaw, Shimeles Fisseha: Analyzed and interpreted the data; Contributed reagents, materials, analysis tools or data; Wrote the paper.

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Data availability statement

Data will be made available on request.

Appendix

Sample raw gravity Data used

| Lon     | Lat    | Elev  | Grav  | Simple Boug. A | CB(IGSN) |
|---------|--------|-------|-------|----------------|----------|
| 38.0001 | 7.67012| 2110.97| 977489.5 | -219.064 | -217.914 |
| 38.0019 | 8.2205 | 2077.45| 977524.8 | -203.969 | -201.859 |
| 38.0023 | 7.8429 | 2915   | 977316.1 | -238.434 | -232.594 |
| 38.0034 | 7.5376 | 1952   | 977531.9 | -208.719 | -207.339 |
| 38.0038 | 7.2131 | 1879   | 977560.2 | -183.4   | -182.7   |
| 38.0058 | 8.06506| 2523.4 | 977430   | -207.07  | -204.97  |
| 38.0070 | 7.672  | 2071   | 977495   | -221.452 | -220.342 |
| 38.0096 | 7.9688 | 2734   | 977367.9 | -225.779 | -223.859 |
| 38.0136 | 7.30787| 1759.86| 977571   | -198.179 | -195.859 |
| 38.0161 | 7.86139| 2885.44| 977321.2 | -239.59  | -233.94  |
| 38.0170 | 7.2146 | 1790   | 977567   | -194.142 | -193.452 |
| 38.0195 | 8.25724| 2105.49| 977514.4 | -209.712 | -207.702 |
| 38.0215 | 8.12664| 2362.24| 977460.7 | -209.632 | -207.782 |
| 38.0235 | 7.6824 | 2017   | 977493.6 | -233.723 | -232.613 |
| 38.0251 | 8.341  | 1978   | 977532   | -219.359 | -217.929 |
| 38.0272 | 8.03352| 2736.34| 977373.3 | -221.091 | -218.961 |
| 38.0300 | 7.67318| 2098.14| 977486   | -225.191 | -224.341 |
| 38.0300 | 7.09   | 1930   | 977540.7 | -190.126 | -187.896 |
| 38.0321 | 7.8751 | 2880   | 977326.7 | -235.506 | -231.376 |
| 38.0333 | 7.1333 | 2012.86| 977521.5 | -193.984 | -191.904 |
| 38.0340 | 7.676  | 2070   | 977486.2 | -230.544 | -229.684 |
| 38.0383 | 7.9558 | 2820   | 977348.8 | -227.193 | -225.123 |
| 38.0392 | 7.5562 | 1910   | 977528.5 | -216.884 | -215.244 |
| 38.0400 | 7.13   | 1790   | 977567   | -192.253 | -190.363 |
| 38.0404 | 7.3296 | 1813   | 977568.1 | -191.126 | -189.726 |
| 38.0419 | 7.2307 | 1782   | 977565.5 | -197.578 | -197.028 |
| 38.0434 | 7.2312 | 1815   | 977567.2 | -189.397 | -188.837 |
| 38.0454 | 7.92438| 2900.05| 977325.4 | -234.121 | -231.361 |
| 38.0461 | 7.8912 | 2887.05| 977323.7 | -237.473 | -233.463 |

Sample raw magnetic Data used

| Easting | Northing | Elevation | Total Mag, Intensity | Date       |
|---------|----------|-----------|----------------------|------------|
| 467750  | 867000   | 1647      | 35240                | 2014-04-03 |
| 464700  | 863900   | 1651      | 35160                | 2014-04-03 |
| 462200  | 859900   | 1656      | 35091                | 2014-04-03 |
| 460600  | 857450   | 1600      | 35147                | 2014-04-03 |
| 471100  | 856500   | 1586      | 35022                | 2014-04-03 |
| 478714  | 860342   | 2058      | 35468                | 2015-04-02 |
| 478818  | 858846   | 2106      | 35283                | 2015-04-02 |
| 479147  | 857350   | 2031      | 35375                | 2015-04-02 |

(continued on next column)
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