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Direct Estimation of Depth to Geological Causative Bodies Using the Half-Width Concept of Gravity and Magnetic Methods

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Abstract
Direct estimation of depth to geological causative bodies using half-width gravity and magnetic methods was carried out. This paper aims to estimate the vertical depth to the centre of a causative body in gravity and magnetic profiles. 3D body (sphere in case of geometrical body), and cylindrical body were considered in the estimation. The depth Z to the centre of the homogeneous sphere and cylinder from the gravity effect was estimated to be $1.305X_{1/2}$ and $0.577X_{1/2}$ respectively. The depth Z to the centre of a homogeneous sphere from the magnetic effect was estimated to be $1.5X_{1/2}$. This paper has shown that it is possible to use gravity results to determine magnetic effects for the equivalent geometrical shapes, through the use of Poisson's relations. Only 2D features in this study are generally amenable to derivation. To ensure a high degree of accuracy, I recommend among other methods the use of source parameter imaging techniques for aeromagnetic and gravity data for the estimation of causative depth in any area of interest.

Keywords: Geological, Causative body, Half-width, Gravity, Magnetic

Introduction

Gravitational methods entail the measurement of minute fluctuations in the Earth's gravitational field across its surface, stemming from lateral variations in mass distribution within the Earth's crust. These variations, influenced by rocks of differing densities, lead to irregularities in gravity intensity (Al-Banna, 1992). Ground-based gravimeters are commonly utilized to precisely measure these gravity field variations at various points. By subtracting a regional field from the measured field, anomalies in gravity are determined, indicating variations in source body density. Positive anomalies suggest shallow high-density bodies, while negative anomalies are linked with shallow low-density bodies. Gravity surveys involve measurements of the gravitational field across designated areas, aiming to correlate gravity variations with density distribution differences, particularly in the horizontal direction, and consequently, rock types (Smith et al., 2008). In certain instances, the entire gravitational field (absolute field) is measured using a pendulum. However, practical applications typically involve measuring the difference between gravitational fields at any two points using a gravimeter. Gravity anomalies represent differences between the actual gravity field and a reference field. Magnetic prospecting methods, akin to gravity methods, are rooted in potential field theory. These methods rely on dipolar fields with varying, time-dependent directions. Magnetic measurements assess variations in the geomagnetic field arising from differences in rock magnetization or the presence of soils rich in magnetic oxides. The unit commonly used for magnetic field intensity in geophysics is the gamma, equivalent to one nanotesla (1.0×10^{-9} tesla) or 1.0×10^{-5} gauss, convenient for expressing fluctuations in the Earth's field due to solar influence and geomagnetic storm phenomena (Fairhead, 2011). The Gauss serves as the unit of magnetic induction in the CGS system, with 1 Gauss ≡ 10^{-4} Tesla.

While land-based magnetic exploration techniques resemble gravity surveys, contemporary magnetic prospecting is predominantly conducted from the air, with aeromagnetic maps proving highly advantageous. The aeromagnetic method is effective for identifying subsurface structures and determining basement depth due to significant susceptibility contrasts between the basement and sedimentary rock (Ozebo et al., 2014; Nwosu, 2014). Though sharing commonalities with gravity methods, the magnetic method is more complex both theoretically and practically. Unlike gravity, magnetic force involves both attraction and repulsion, while gravitational force is solely attractive.

A magnetic anomaly indicates a local departure of the geomagnetic field from the overall field produced by Earth's core dynamo and external sources, attributed to the magnetization of nearby crustal rocks (Ofoha et al., 2016). This anomaly is determined by measuring the local magnetic field and subtracting the prediction of a global field model, such as the International Geomagnetic Reference Field (IGRF). Crustal rocks may acquire magnetization in situ after being heated and then cooled through their Curie temperature in Earth's ambient magnetic field. Alternatively, magnetized particles may align themselves during or after deposition in a sedimentary environment, resulting in the retention of net magnetization in the rock (Al-Mufarji et al., 2017). Normally, the small atomic magnets within a material are randomly oriented, nullifying their magnetic effects. However, the presence of an external magnetic field slightly alters the electron orbits, causing them to oppose the external field, thus weakening the magnetic field—a phenomenon termed diamagnetism, which is characteristic of all substances (Al-Mufarji et al., 2017).

Materials and Methods

Gravity surveys are typically conducted to identify variations in the subsurface density of geological materials. Changes in lateral density within the subsurface can influence the gravitational force at the surface. The force of gravity attributed to a buried mass difference, such as concentration or void, is added to the overall gravitational force exerted by the entire mass of the Earth (Ritchard et al., 2000). Gravitational prospecting is grounded in Newton's law of gravitation, which mathematically describes the force of mutual attraction between two particles based on their masses and the distance that separates them.

$$F = G\frac{m_1 m_2}{r^2} \qquad (1)$$

where F = force of attraction, m_1 *and* m_2 = masses of the two bodies, r = distance between the two bodies and G = the universal gravitational constant = $6.664 \times 10^{-11}\ m/Kgs^2$.

Instruments measure the vertical component of the gravitational effect of the target g_z. However, the formula:

$$g = \frac{F}{m_2} = G\frac{m_1}{r^2} \qquad (2)$$

gives the acceleration experienced in line with two masses where r is the distance between centre of masses (sensor and target mass). This law as stated mathematically in equation (2) is valid if we assume *spherical masses, homogeneous and isotropic bodies and medium, uniform density and non-rotating earth or stationary earth* (Shehu et al., 2017). Measurement in gravity is always in a vertical direction i.e. g_z. Since we measure g_z the surface over the target, we must resolve the geometry. Then g_z at the location (x,0,0) is

$$g_z = g\cos\theta$$
$$But, BO = (z^2 + x^2)^{1/2}$$
And
$$\cos\theta = \frac{z}{\sqrt{z^2 + x^2}}$$
$$\therefore, g_z = \frac{gz}{\sqrt{z^2 + x^2}} \qquad (3)$$
At $\theta = 0$, $x = 0$ *and* g_z *becomes* g

Magnetic Method

The Earth's magnetic field, produced by the outer core and shielding the planet from solar energy, mimics the operation of a colossal magnet aligned with the Earth's axis. Characterized by positive and negative poles, a bar magnet exhibits dipolarity, forming a complete field loop. Similar poles repel each other, while opposite poles attract, generating magnetic fields applicable at various scales. All substances contain mobile electrical charges, leading to the formation of magnetic dipole moments that interact with and are influenced by other magnetic fields. The magnetic susceptibility (Xm) of a material serves as an indicator of its reaction to an applied magnetic field, correlating with its atomic structure and consequent magnetic characteristics (Thurston & Smit, 2005). This susceptibility, defined as the ratio of magnetization to magnetizing-field strength, is contingent upon the type(s) and concentration(s) of magnetic materials present, such as ferromagnetic, diamagnetic, or paramagnetic substances.

When two poles of strength P0 and P1 are separated by a distance r, the force between them is determined by Coulomb's law for magnetic poles, resembling Newton's law of gravitational force to a significant extent. Hence,

$$F = \frac{1}{\mu} \cdot \frac{P_0 P_1}{r^2} \qquad (4)$$

Where μ is the permeability of the medium in which the poles are situated and depends on the properties of the medium. μ is precisely 1 in vacuum and in air.
A more pragmatic measure than force is the magnetic field strength present at a specific point in space. Magnetic field strength is characterized as the force per unit pole strength exerted on a small pole with strength P0 positioned at that particular point. Field strength H due to a pole of strength P_0 at a distance r away is

$$H = \frac{F}{P_0} = \frac{1}{\mu} \cdot \frac{P_0 P_1}{r^2} \cdot \frac{1}{P_0} = \frac{P_1}{\mu r^2} \qquad (5)$$

Isolated poles do not exist. The fundamental magnetic entity is therefore the magnetic dipole. A dipole of equal pole strength $+P$ and $-P$ separated by a short distance L will produce a dipole PL of moment M given as

$$M = PL \qquad (6)$$

Any magnetic material placed in an external field will have magnetic poles induced upon its surfaces by the magnetic field. It therefore becomes magnetized by induction. The intensity of magnetization (I) is defined as the magnetic moment per unit volume of the magnetized material or body. That is

$$I = \frac{M}{V} = \frac{PL}{V} \qquad (7)$$

The intensity of magnetization is proportional to the strength of the field. In geophysical work, we are primarily concerned with moderately magnetic materials and weak fields. In this case, the involved magnetization or polarization is in the direction of the applied field. This form of magnetization can be viewed as the alignment of elementary magnets or dipoles, which initially had a random orientation. It is postulated that the quantity of aligned magnets is contingent upon the intensity of the magnetizing field. If the magnetizing field (I) remains constant and maintains a consistent direction throughout, the substance is referred to as uniformly magnetized.

In the case of a homogeneous external field (H) making an angle θ with the normal to the magnetizable body, the induced pole strength per unit area is determined as follows:

$$I = KH\cos\theta \qquad (8)$$

For the field normal to the surface,

$$I = KH\cos\theta = KH\cos(90) = KH \qquad (9)$$

K is the proportionality constant known as susceptibility, representing the extent to which a body undergoes magnetization. It serves as a gauge of the ease of magnetization, reflecting the density of elementary magnets per unit volume of materials and their mobility, indicating how readily they can be aligned. This parameter holds paramount importance in magnetic prospecting as the magnetic response of rocks and minerals is influenced by the quantity of magnetic materials present.

For vacuum and non-magnetic substances, K equals 0. Materials exhibiting positive susceptibilities are categorized as paramagnetic, while those with exceptionally high susceptibility may be termed ferromagnetic. Some substances, such as rock salt and anhydrite, exhibit negative susceptibilities and fall into the category of diamagnetic materials.
When a magnetic body is subjected to an external field H, its internal poles align with the field H, generating its own field H', which is correlated with the intensity of magnetization I through the following formula:

$$H' = 4\pi I$$

This extra field H' increases the total field within the body. Magnetic induction B is defined as the total field within the body. Thus

$$B = H + H' = H + 4\pi I = H + 4\pi(KH) = H(1 + 4\pi K)$$

The ratio of the induction B to the magnetizing force H is the magnetic permeability μ. Thus

$$\mu = \frac{B}{H} = 1 + 4\pi K \qquad (10)$$

Permeability is a measure of the modification of the force of attraction or repulsion between two magnetic poles.

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Depth Estimation
Half-Width Method
The half-width (X1/2) technique is employed in the field for promptly estimating the depth of a specific source. It involves measuring the horizontal distance from the peak of the anomaly to the point where the anomaly's value has decreased to half of its maximum (Fig. 1).

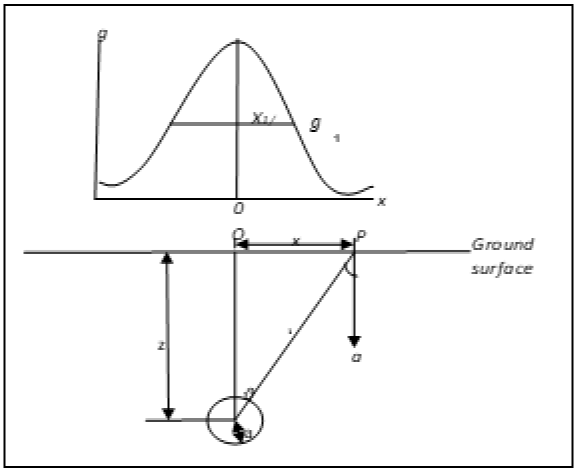

Figure 1: The half-width method (Kearey et al., 2002)

Half-width refers to half the width of an anomaly, measured between points where the amplitude is reduced to half of the peak amplitude. This method is commonly applied in determining the depths to the centres of causative bodies in gravity and magnetic profiles. To estimate the vertical depth to the centre of mass, one can equate the amplitude at x = 0 to twice the amplitude at x. This leads to a half-width rule that provides the depth to the centre of a sphere, expressed as Z = 1.3X1/2, where X1/2 represents half the width of the measured gravity or magnetic anomaly at half of its amplitude

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Discussion
The Concept of Half-width ($X_{1/2}$)
Gravity Effect
The gravity effect of a homogeneous sphere is the same as it would be if its entire of mass were concentrated at a point at its centre.
For this sphere of mass, m, at the point P at a distance, r, from its centre, the vertical component of gravity becomes

$$g_z = g\cos\theta = G\frac{mz}{r^3};\ \ \cos\theta = \frac{z}{r} = \frac{z}{(x^2+z^2)^{1/2}}$$

$$= G\frac{V\rho z}{r^3} = \frac{4}{3}G\pi a^3\rho\cdot\frac{z}{r^3} = K\frac{\rho a^3 z}{(x^2+z^2)^{\frac{3}{2}}} \quad (11)$$

where $K = \frac{4\pi G}{3} = 27.9\ \times 10^{-3}\ in\ meters\ = 8.53\ in\ mGal; m = V\rho\ and\ V = \frac{4}{3}\pi a^3$

If a, z and x are given and ρ is the density, then we can write G in milligals and g_z becomes

$$g_z = \frac{8.53\rho a^3 z}{(x^2+z^2)^{3/2}} = \frac{8.53\rho a^3}{z^2}\left(\left(\frac{x}{z}\right)^2+1\right)^{-\frac{3}{2}} \quad (12)$$

The half-width is half the width at half amplitude (g_{max}).
At the point $x = 0$,

$$g = g_{max} = \frac{8.53\rho a^3}{z^2} \quad (13)$$

At the point $x = X_{1/2}$

$$g = \frac{1}{2}g_{max}$$

$$\frac{8.53\rho a^3}{z^2}\left[\left(\frac{x_{\frac{1}{2}}}{z}\right)^2+1\right]^{-\frac{3}{2}}=\frac{1}{2}\left(\frac{8.53\rho a^3}{z^2}\right)$$

$$i.e,\left[\left(\frac{x_{1/2}}{z}\right)^2+1\right]^{-\frac{3}{2}}=\frac{1}{2} \quad (14)$$

Solving this equation gives

$$z = 1.305X_{1/2} \quad (15)$$

The gravity effect of a homogeneous cylinder is the same as if the mass were concentrated on a line along its axis. The vertical component of gravity from a horizontal line element of infinite length can be shown by interpretation of the effect for a point mass to be

$$g_z = \frac{2Gmz}{r^2};\ m\ being\ the\ mass\ per\ unit\ length \quad (16)$$

If we replace the horizontal line element with a cylinder of density ρ, the mass per unit length becomes

$$m = \rho V = \pi\rho R^2 \quad (17)$$

and g_z becomes

$$g_z = \frac{2Gz}{r^2}\cdot\pi\rho R^2 = 2\pi G\rho R^2\cdot\frac{z}{r^2} = 2\pi G\rho R^2\cdot\frac{z}{(x^2+z^2)} \quad (18)$$

Expressing the above equation in terms of x/z, we have

$$g_z = 2\pi G\rho R^2\cdot\frac{1}{z^2}\cdot\frac{z}{\left(\frac{x^2}{z^2}+1\right)} = 2\pi G\frac{\rho R^2}{z}\cdot\left(\frac{x^2}{z^2}+1\right)^{-1} \quad (19)$$

Substituting numerical values for the various constants gives

$$g_z = 12.77\frac{\rho R^2}{z}\cdot\left(\frac{x^2}{z^2}+1\right)^{-1} \quad (20)$$

At the point $x = 0$,

$$g = g_{max} = 12.77\frac{\rho R^2}{z} \quad (21)$$

At the point $x = X_{1/2}$

$$g = \frac{1}{2}g_{max}$$

That is

$$12.77\frac{\rho R^2}{z}\cdot\left(\frac{{X_{1/2}}^2}{z^2}+1\right)^{-1} = \frac{1}{2}\left(12.77\frac{\rho R^2}{z}\right)$$

Therefore, by the half-width concept, the depth to centre of a cylinder $Z = X_{1/2}$

For infinite vertical cylinder,

$$g_z = 6.39\frac{\rho R^2}{z}\cdot\left(\frac{x^2}{z^2}+1\right)^{-\frac{1}{2}} \quad (22)$$

and

$$Z = 0.577\ X_{1/2} \quad (23)$$

Magnetic Effects

Similar to the gravitational field, the magnetic field possesses a potential. This potential represents the work required to move a unit magnetic pole from infinity to a point situated at a distance r from another source of magnetic polarity with a strength of P. The magnetic potential can be expressed as:

$$U = \frac{1}{\mu}\cdot\frac{P}{r} \quad (24)$$

The magnetic field is determined by taking the spatial derivative of the magnetic potential. The direction of the magnetic field corresponds to the direction associated with these derivatives. For example, the x – x-component of the field is $\partial/\partial x$. The magnetic potential and hence the field strength U at any point can be expressed in terms of the gravitational potential V by use of Poisson's equation written as

$$U = -\frac{I}{G\rho} \cdot \frac{\partial V}{\partial i} \tag{25}$$

where i = direction of magnetization, I = magnetization or polarization, ρ = density and G = universal gravitational constant.

The magnetic field component in any direction is given as

$$H_s = -\frac{\partial U}{\partial s} = \frac{I}{G\rho} \cdot \frac{\partial}{\partial s}\left(\frac{\partial V}{\partial i}\right) \tag{26}$$

Therefore, if the object is polarized vertically along the z-axis, and if one seeks the horizontal component Hx of the magnetic field, it can be derived from the following equation:

$$H_x = -\frac{\partial U}{\partial x} = \frac{I}{G\rho} \cdot \frac{\partial}{\partial x}\left(\frac{\partial V}{\partial z}\right) \tag{27}$$

and the vertical component is written as

$$H_z = -\frac{\partial U}{\partial z} = \frac{I}{G\rho} \cdot \frac{\partial}{\partial z}\left(\frac{\partial V}{\partial z}\right) = \frac{I}{G\rho}\left(\frac{\partial^2 V}{\partial z^2}\right) \tag{28}$$

A homogeneous sphere buried at a large depth compared to its diameter and polarized in the earth's field could be said to have a distribution of dipoles which can be approximated by a single vertical magnet with the negative pole near the centre of the upper half and the positive pole near the centre of the lower half. Describing these equivalent poles, each with a strength denoted as P and separated by a distance represented by L within the sphere of radius R and susceptibility K, we can express the relationship as follows:

$$I = \frac{M}{V} = \frac{3}{4} \cdot \frac{PL}{\pi R^3} \tag{29}$$

Therefore,

$$P = \frac{4}{3}\pi R^3 \cdot \frac{I}{L} = \frac{4}{3}\pi R^3 \cdot \frac{K}{L} H_z \tag{30}$$

The above approximation suggests that the field of the sphere is the same as that of a vertical dipole illustrated below. In practice, the representation of the field of a sphere by that of a single dipole is not very useful because there is no way one can determine the effective separation of the poles L. The most satisfactory approach is to use the Poisson's relation stated earlier as

$$H_z = \frac{I}{G\rho}\left(\frac{\partial^2 V}{\partial z^2}\right) \tag{31}$$

The gravitational potential due to a sphere of radius R and known depth to the centre of mass can be determined. For example, for a sphere of radius R, density ρ and mass m, the gravitational potential at distance r from the centre is

$$V = \frac{Gm}{r} = \frac{4}{3} \cdot \frac{G\pi R^3 \rho}{r} = \frac{4}{3} \cdot \frac{G\pi R^3 \rho}{(x^2 + z^2)^{\frac{1}{2}}} \tag{32}$$

Then,

$$\frac{\partial V}{\partial z} = -\frac{4}{3} \cdot G\pi R^3 \rho \cdot \frac{z}{(x^2 + z^2)^{\frac{3}{2}}} \tag{33}$$

and

$$\frac{\partial^2 V}{\partial z^2} = \frac{4}{3} \cdot G\pi R^3 \rho \cdot \frac{(2z^2 - x^2)}{(x^2 + z^2)^{\frac{5}{2}}} \tag{34}$$

One can derive H_z for the sphere by using the Poisson's equation (31) and (34). Thus

$$\begin{aligned} H_z &= \frac{I}{G\rho}\left(\frac{\partial^2 V}{\partial z^2}\right) = \frac{I}{G\rho}\left(\frac{4}{3} \cdot G\pi R^3 \rho \cdot \frac{(2z^2 - x^2)}{(x^2 + z^2)^{\frac{5}{2}}}\right) \\ &= \frac{4\pi R^3 I}{3}\left(\frac{(2z^2 - x^2)}{(x^2 + z^2)^{\frac{5}{2}}}\right) \\ &= \frac{4}{3}\pi R^3 I \left(\frac{z^2\left(2 - \left(\frac{x}{z}\right)^2\right)}{z^5\left[1 + \left(\frac{x}{z}\right)^2\right]^{\frac{5}{2}}}\right) \end{aligned} \tag{35}$$

$$= \frac{4\pi R^3 I}{3z^3}\left(\frac{\left[2-\left(\frac{x}{z}\right)^2\right]}{\left[1+\left(\frac{x}{z}\right)^2\right]^{\frac{5}{2}}}\right) \quad (36)$$

At the point $x = 0$,

$$H = H_{max} = \frac{4\pi R^3 I}{3z^3}$$

At the point $x = X_{1/2}$

$$g = \frac{1}{2} g_{max}$$

That is

$$\frac{4\pi R^3 I}{3z^3}\left(\frac{\left[2-\left(\frac{X_{1/2}}{z}\right)^2\right]}{\left[1+\left(\frac{X_{1/2}}{z}\right)^2\right]^{5/2}}\right) = \frac{1}{2}\left(\frac{4\pi R^3 I}{3z^3}\right)$$

$$\left(\frac{\left[2-\left(\frac{X_{1/2}}{z}\right)^2\right]}{\left[1+\left(\frac{X_{1/2}}{z}\right)^2\right]^{5/2}}\right) = \frac{1}{2} \quad (37)$$

$$i.e.\, 4 - \frac{2X_{1/2}^2}{z^2} = \left(1 + \frac{X_{1/2}^2}{z^2}\right)^{\frac{5}{2}} \quad (38)$$

Applying the binomial series and ignoring powers of $X_{1/2}$ and z more than 2, the righthand side of the equation (38) becomes

$$\left(1 + \frac{X_{1/2}^2}{z^2}\right)^{5/2} = 1 + \frac{5}{2}\left(\frac{X_{1/2}^2}{z^2}\right) = \frac{2z^2 + 5X_{1/2}^2}{2z^2}$$

$$\Rightarrow \frac{4z^2 - 2X_{1/2}^2}{z^2} = \frac{2z^2 + 5X_{1/2}^2}{2z^2} \quad (39)$$

$$8z^2 - 4X_{1/2}^2 = 2z^2 + 5X_{1/2}^2$$

$$z^2 = \frac{9}{6} X_{1/2}^2$$

$$Z = 1.5 X_{1/2} \quad (40)$$

Conclusion

Direct estimation of depth to geological causative bodies using the half-width method of gravity and magnetic was carried out. This paper aims to estimate the vertical depth to the centre of a causative body in gravity and magnetic profiles. 3D body (sphere in case of geometrical body), and cylindrical body were considered in the estimation. The depth (Z) to the centre of the homogeneous sphere and cylinder from the gravity effect was estimated to be $1.305X_{1/2}$ and $0.577X_{1/2}$ respectively. The depth to the upper limiting depth is less than $1.305X_{1/2}$ (Kearey et al., 2002). The depth Z to the centre of a homogeneous sphere from the magnetic effect was estimated to be $1.5X_{1/2}$. This paper has shown that it is possible to use gravity results to determine magnetic effects for the equivalent geometrical shapes, through the use of Poisson's relations. Only 2D features in this study are generally amenable to derivation. The generalized forms for which vertical magnetic fields and gravity fields have been derived can be used to approximate several geologic features. An up warp on the basement surface might be represented by a sphere, a basement ridge by a horizontal cylinder, or intrusive of different types by vertical cylinders or sheets. A fault affecting the basement surface could correspond to the edge of a rectangular slab.

Recommendation

To ensure a high degree of accuracy, the study recommends among other methods the use of source parameter imaging techniques for aeromagnetic and gravity data to the estimation of causative depth in any study area of interest.

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