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DEPARTMENT OF GEOLOGY

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GRAVITY METHOD OF EXPLORATION

The gravity method of geophysical exploration is based on the measurement of variations in the gravity field caused by horizontal variations of density within the subsurface. Differences in rock density produce small changes in the Earth's gravity field that can be measured using portable instruments known as gravity meters or gravimeters.

Gravity surveys are used either alone or in combination with magnetotelluric, magnetic, and induced polarization and resistivity surveys to determine the location and size of the major source structures containing accumulations of hydrocarbons, massive base metal deposits, iron ore, salt domes, and hydrogeological aquifers. It is an important technique for many problems that involve subsurface mapping, and it is the principal method in a number of specific types of geological studies.

Principle

The gravity field of the Earth is caused by the fundamental phenomenon of gravitation which causes bodies to be attracted toward each other. The basis of the gravity survey method is Newton's Law of Gravitation, which states that the force of attraction F between two masses m_1 and m_2, whose dimensions are small with respect to the distance r between them, is given by where G is the Gravitational Constant (6.67x10-11 m3 kg^{-1} s^{-2}).

$$F = G\,\frac{m_1 m_2}{r^2}$$

Gravity units

The mean value of gravity at the Earth's surface is about 9.8ms^{-2}.Variations in gravity caused by density variations in the subsurface are of the order of 100 mms^{-2}.This unit of the micrometre per second per second is referred to as the gravity unit (gu).

In gravity surveys on land an accuracy of ±0.1 gu is readily attainable, corresponding to about one hundred millionth of the normal gravitational field. At sea the accuracy obtainable is considerably less, about ±10 gu. The c.g.s. unit of gravity is the milligal (1 mgal = 10^{-3} gal = $10^{-3}$$cms^{-2}$), equivalent to 10 gu.

Density of rocks

The SI unit of density is the kgm^{-3} but the Mgm^{-3} is widely used since the values are, numerically, the same as those in the old cgs system in which water has unit density. Most crustal rocks have densities of between 2.0 and 2.9 Mgm^{-3}. In the early days of gravity work a density of 2.67 Mgm^{-3} was adopted as standard for the upper crust and is still widely used in modelling and in calculating elevation corrections for standardized gravity maps. Density ranges for some common rocks are shown in Table -1.

Table -1

Rock	Density (Mgm−3)
Gneiss	2.65–2.75
Granite	2.5–2.7
Basalt	2.7–3.1
Limestone	2.6–2.7
Quartzite	2.6–2.7
Gabbro	2.7–3.3

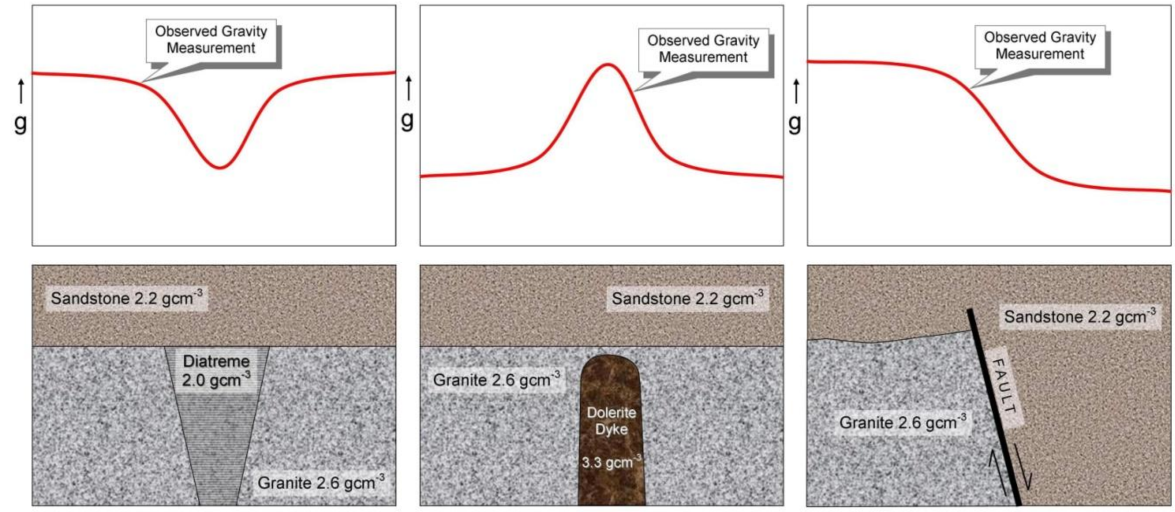

Figure 1- Observed gravity anomalies along different rock subsurfaces.

Instruments

There are two types of gravimeters:

1) Absolute 2) Relative

Absolute gravimeters measure the local gravity in absolute units, gals. Relative gravimeters compare the value of gravity at one point with another. They must be calibrated at a location where the gravity is known accurately, and then transported to the location where the gravity is to be measured. They measure the ratio of the gravity at the two points.

The measurement of an absolute value of gravity is difficult and requires complex apparatus and a lengthy period of observation. Such measurement is classically made using large pendulums or falling body techniques (see e.g. Nettleton 1976, Whitcomb 1987), which can be made with a precision of

0.01 gu. The measurement of relative values of gravity, that is, the differences of gravity between locations, is simpler and is the standard procedure in gravity surveying.

Previous generations of relative reading instruments were based on small pendulums or the oscillation of torsion fibres and, although portable, took considerable time to read. Modern instruments capable of rapid gravity measurements are known as gravity meters or gravimeters. Gravimeters are basically spring balances carrying a constant mass. Variations in the weight of the mass caused by variations in gravity cause the length of the spring to vary and give a measure of the change in gravity. The gravity meters used for the geophysical prospection are the **astatic gravity meters** (Figure 2).

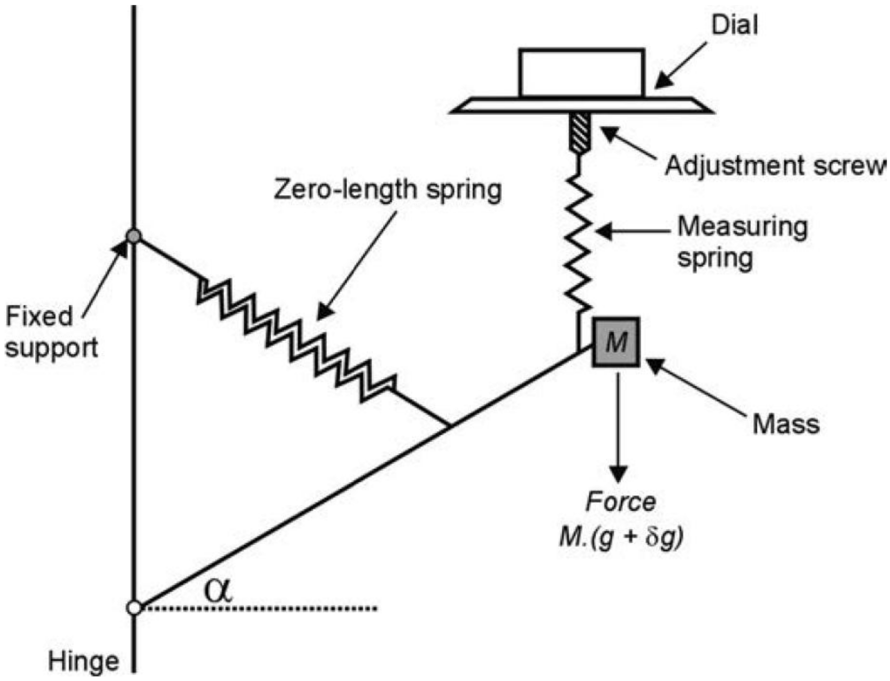

Figure 2- Schematic sketch of the astatic gravity meter. The zero-length spring supports the mass M and keeps it in balance in a selected gravity field g. Measurements are done by rotating the dial, which raises or lowers the measuring spring and provides additional force M .δg to return the mass to the standard position (Milsom, 2011).

Data reduction

The removal of unwanted components is often referred to as *reduction*. Because the Earth is not a perfect homogeneous sphere, the gravitational acceleration is not constant over the whole Earth's surface. Its magnitude depends on five following factors: latitude, elevation, topography of the surrounding terrain, earth tides and density variations in the subsurface.

a) Latitude correction

Latitude corrections are usually made by subtracting the normal gravity, calculated from the International Gravity Formula, from the observed or absolute gravity.

b) Free-air correction

The remainder left after subtracting the normal from the observed gravity will be due in part to the height of the gravity station above the sea-level reference surface. An increase in height implies an increase in distance from the Earth's centre of mass and the effect is negative for stations above sea level. The *free-air correction* is thus positive, and for all practical purposes is equal to 3.086

g.u./metre. The quantity obtained after applying both the latitude and free-air corrections is termed the *free-air anomaly* or *free-air gravity.*

c) Bouguer anomaly

Since topographic masses are irregularly distributed, their effects are difficult to calculate precisely and approximation is necessary. The simplest approach assumes that topography can be represented by a flat plate extending to infinity in all directions, with constant density and a thickness equal to the height of the gravity station above the reference surface. This *Bouguer plate* produces a gravity field equal to $2\pi\rho Gh$, where h is the plate thickness and ρ the density (1.1119 g.u./metre for the standard 2.67 Mgm^{-3} density). The Bouguer effect is positive and the correction is therefore negative.

d) Terrain corrections

In areas of high relief, detailed topographic corrections must be made. Although it would be possible to correct directly for the entire topography above the reference surface in one pass without first making the Bouguer correction, it is simpler to calculate the Bouguer gravity and then correct for deviations from the Bouguer plate. Adding terrain corrections to the simple Bouguer gravity produces a quantity often known as the *extended* or *complete Bouguer gravity*.

e) Tidal correction.

The tidal correction accounts for the gravity effect of Sun, Moon and large planets. Modern gravity meters compute the tide effects automatically.

f) Drift correction

This correction is intended to remove the changes caused by the instrument itself. If the gravimeter would be at one place and take periodical readings, the readings would not be the same. These are partly due to the creep of the measuring system of the gravimeter, but partly also from the real variations – tidal distortion of the solid Earth, changes of the ground water level, etc. The drift is usually estimated from repeated readings on the base station. The measured data are then interpolated, e.g. by a third order polynomial, and a corrections for profile readings are found.

Gravity data processing

Once the gravity data are measured the more demanding task is to be carried out – the data processing and interpretation.

The first step in the data processing is deleting of wrong gravity readings. During the the field measurements there is usually several gravity readings taken at every station. Now, the outliers are removed and the rest of gravity readings from every station are averaged. Next, the readings from the base station are taken to determine the drift of the instrument. First, these data need to be corrected for the different heights of the tripod, the free-air correction.

Second, the drift should be estimated – usually the data are interpolated using the second or third-order polynomial. Third, the readings at individual stations are corrected from the drift. The drift is estimated from the fitted polynomial according to the time of the gravity reading.

Fourth, the drift corrected data are reduced again, now using the latitude, free-air and Bouguer reductions. If necessary, the density for the Bouguer slab is estimated (e.g. Nettleton's method).

There are also additional steps, which depends on the type of the survey and target structures. However, usually we want to suppress regional anomalies and enhance the local ones or vice versa. The regional anomalies is a general term depending on the size of target structures. These anomalies are caused by large and deep structures, often larger than our survey area. In the data they usually represent the long-wavelength high-amplitude anomalies. Sometimes they are also referred to as a trend (Fig.3). There are numerous techniques to remove the trend, the easiest are based on approximation by a polynomial. In this case we take the part of the data without our target anomaly and fit a polynomial through them. This polynomial approximates the effect of large-scale regional structures and we can subtract it from our data leaving us with residual anomalies. The residual anomalies are, in an ideal world, anomalies caused only by our target structures.

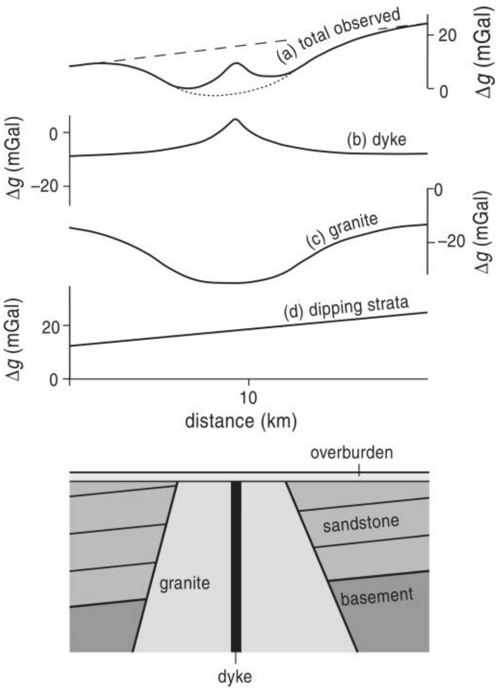

Figure 3-Illustration of regional and residual anomalies (Musset and Khan 2000).The observed gravity curve contains information about all geological structures (topmost curve). If we are looking for the dyke, then the anomalies due to the dipping strata and graniticpluton are not relevant to our research and we would like to remove them from the data to ease the interpretation.

Gravity data interpretation

The interpretation of gravity data could be only a simple qualitative analysis in a way or a more complex quantitative analysis, where, based on the qualitative assignment, we try to somehow model the subsurface. In this respect we have to bear in mind that the interpretation (inversion) of geophysical data is non-unique. In gravity prospection not only that different bodies could have similar anomalies, they can also produce exactly the same anomaly (Figure 4). The non-uniqueness is inherent to gravity data and could not be overcome e.g. by adding more gravity data. The only way how to get sound and reliable interpretation is to include an a priori geological knowledge and, if possible, also data from another geophysical methods.

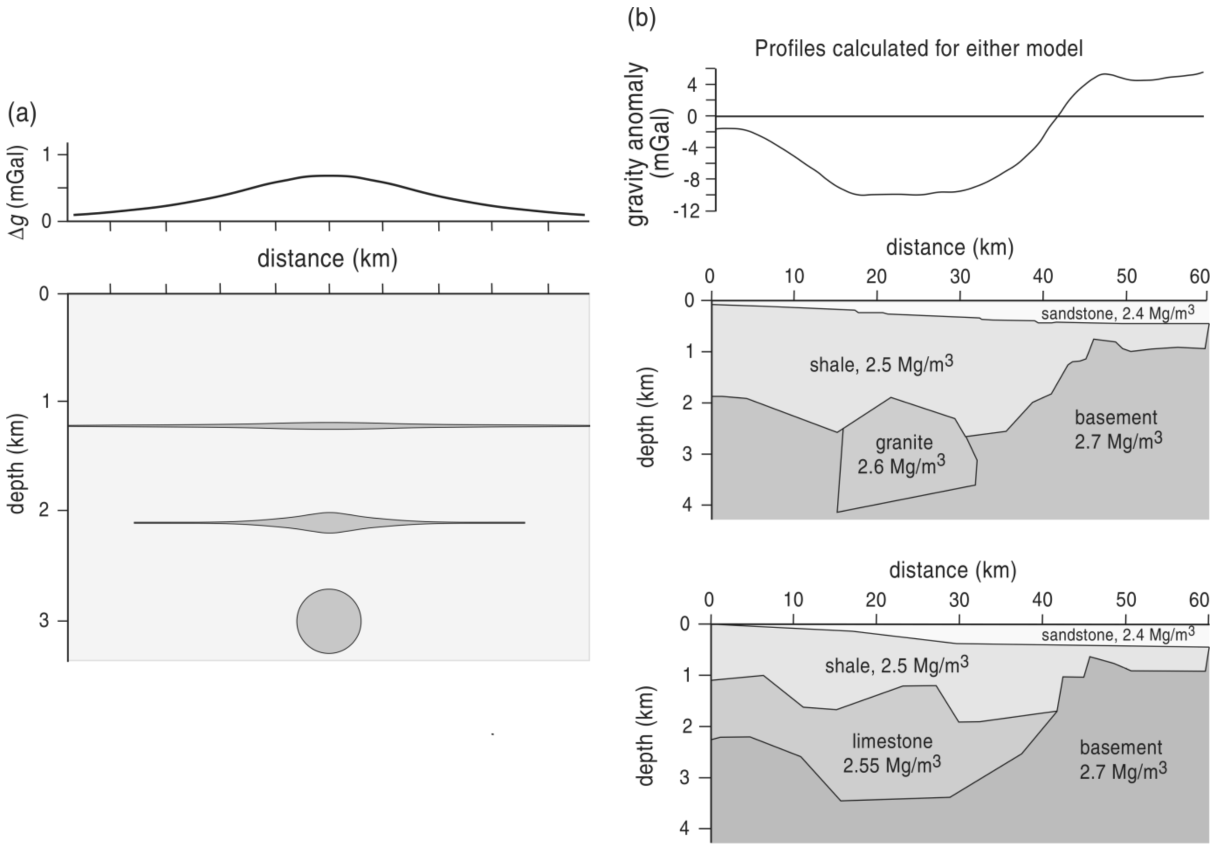

Figure 4- Non-uniqueness of the gravity interpretation. The plotted models produce exactly the same gravity anomalies (Musset and Khan, 2000).