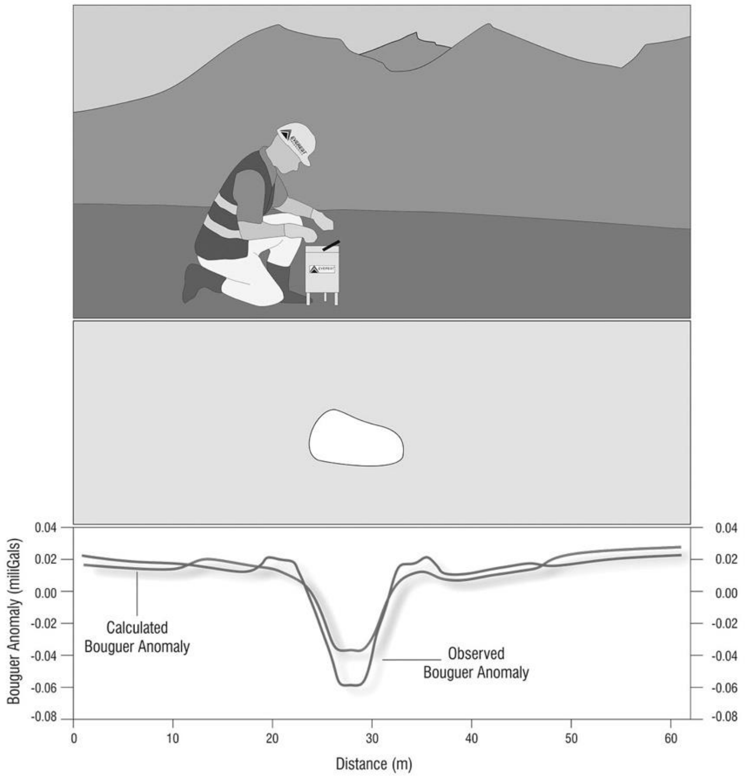
Gravity & Magnetic Methods

What are potential field methods?

Gravity and magnetic surveys are called **potential field methods** because the quantities we measure (gravitational acceleration and magnetic field intensity) are both derived from scalar potentials satisfying Laplace's equation. Unlike seismic surveys, no active source is needed: the signal originates from the physical properties of the Earth itself (density and magnetisation).

Seismic = active source.

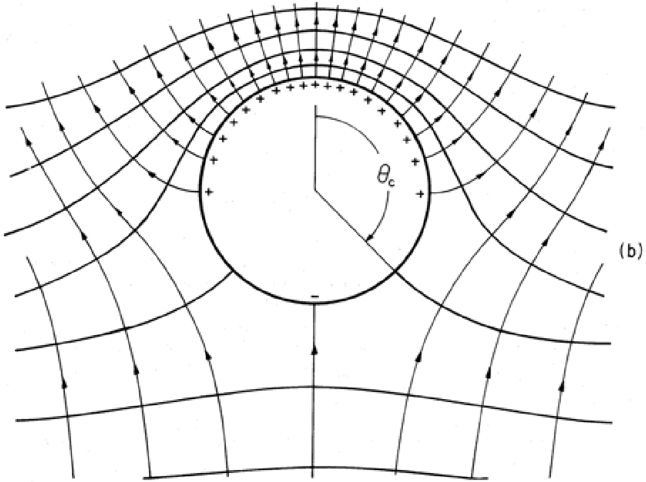
Potential fields = passive (the Earth is the source.)



The Laplace Equation (The “smoothness rule” of physics)

$$\nabla^2\phi = 0$$

$$\nabla^2\phi = \frac{\partial^2\phi}{\partial x^2} + \frac{\partial^2\phi}{\partial y^2}$$



Plain-language explanation

ϕ (phi) = a potential field

(gravity potential or magnetic potential)

The Laplacian (∇^2) measures how curved or uneven the field is.

The equation means: *At any point, the value of the field is the average of the surrounding values.*

Why Gravity and Magnetics follow Laplace's equation

Gravity:

$$g = -\nabla\phi_g$$

Magnetic:

$$B = -\nabla\phi_m$$

Gravity and magnetic signals originate from rocks inside the Earth.

Gravity anomalies are produced by changes in rock density.

Magnetic anomalies are produced by rocks that contain magnetic minerals.

Meaning: The field we measure is just the spatial change of a potential.

Applications: Gravity in exploration geophysics

Small scales in microgravity - variations from e.g. voids, sinkholes, lava tunnels…

Photos credit: Prof. J Neuberg (Leeds University)

Applications: Gravity in exploration geophysics

Gravity surveys measure tiny variations in Earth's gravitational acceleration caused by differences in the density of subsurface rocks. If a region underground contains unusually dense material, the gravitational attraction there is slightly stronger. If the rocks are less dense (for example a sedimentary basin), the gravitational attraction is slightly weaker.

By mapping these small variations in gravity across an area, we can infer how density changes underground and therefore identify geological structures such as basins, faults, intrusions, or mineral deposits.

Reminder: Newton's law of gravitation

Near the Earth's surface we usually measure gravitational acceleration, written as g. The average value of gravity on Earth is:

g =~ 9.81 m/s^2

Exploration geophysics is not concerned with the total value of gravity, but with very small variations in this value caused by geological structures. These variations are typically only a few milligals:

1 mGal = 10^{-5} m/s^2.

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

F is magnitude of gravitational force. m1 and m2 are masses. r is the distance between the two masses. G is the gravitational constant 6.67 × 10^{-11} Nm^{2}kg^{-2}

Newton's law of gravitation states that every mass attracts every other mass with a force proportional to their masses and inversely proportional to the square of the distance.

Why density matters

Gravity depends directly on mass, and mass depends on density. Different rock types have different densities. Typical values are:

Sediments: ~2000–2400 kg/m^3

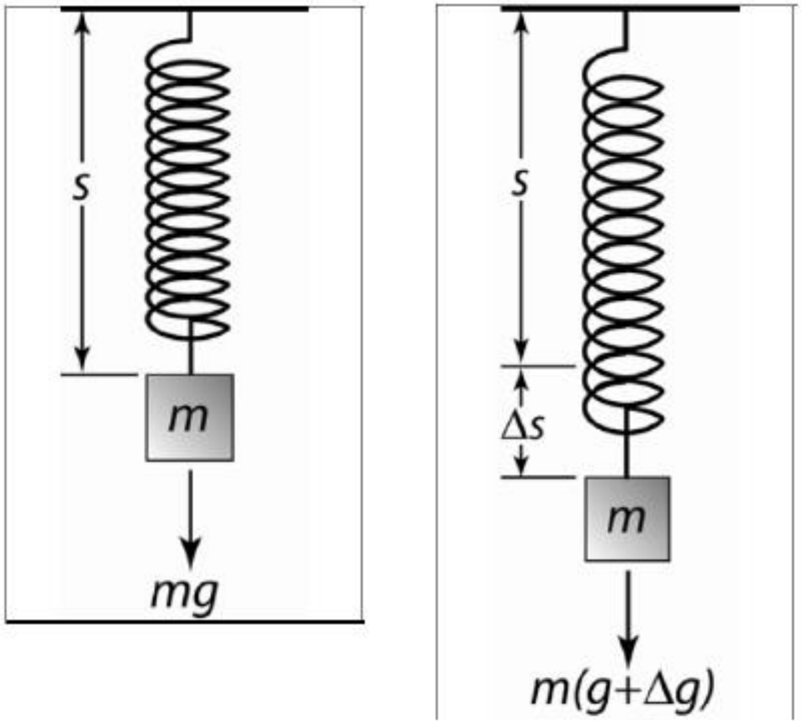
Granite: ~2600–2700 kg/m^3

Basalt: ~2900–3100 kg/m^3

Metallic ore bodies: sometimes >4000 kg/m^3

If a dense body is buried underground, it adds extra mass locally and increases gravity slightly. If low-density material is present, gravity becomes slightly weaker. *Gravity surveys therefore provide information about density contrasts in the subsurface.*

Applications: Gravity in exploration geophysics



If gravity increases:

- Spring gets stretched
- Spring force increases
- Extension stops when forces balance again

$$\Delta s \rightarrow \Delta g$$

What do we actually measure?

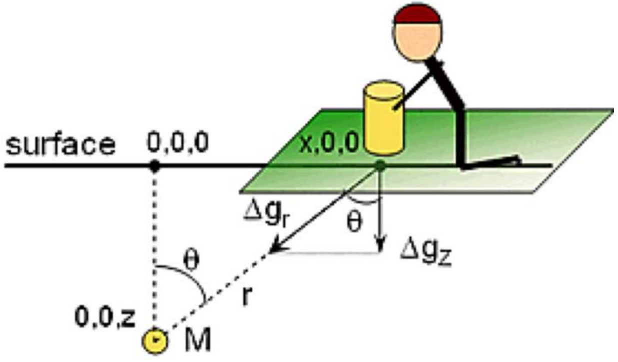
A gravimeter measures the vertical component of gravitational acceleration, usually written as Δg_z . **This means the instrument does not measure the full gravitational force vector, but only the part of that force that points vertically downward.**

To understand this, imagine a small dense body buried underground at depth z. That body attracts the instrument. The gravitational force produced by that mass points toward the mass itself, not straight downward. **So the attraction is along the direction r, the distance between the instrument and the buried mass.**

What do we actually measure?

Newton's law tells us the magnitude of this attraction:

$$g_r = \frac{GM}{r^2}$$



Where:

G is the gravitational constant

M is the mass of the buried object

and *r* is the distance between the instrument and the mass

However, the gravimeter only measures the vertical part of this attraction. To obtain that, we project the force onto the vertical direction.

What do we actually measure?

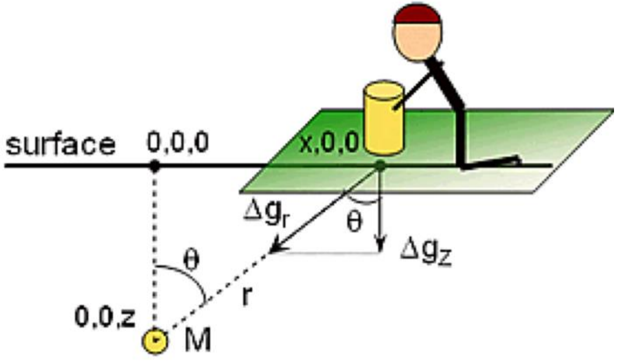
$$\Delta g_z = g_r \cos(\theta)$$

$$\cos(\theta) = z/r$$

$$\Delta g_z = g_r \frac{z}{r}$$

$$\text{Substituting} \quad g_r = \frac{GM}{r^2}$$

$$\boxed{\Delta g_z = \left(\frac{GM}{r^2}\right)\frac{z}{r} = \frac{GMz}{r^3}}$$



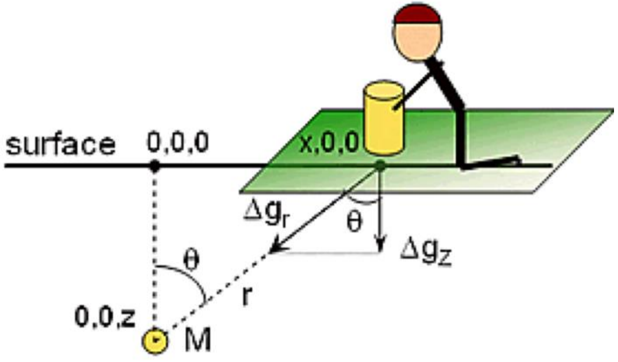
What do we actually measure?

If the horizontal distance from the body is *x*, then:

$$r^2 = x^2 + z^2$$

If the horizontal distance from the body is *x*, then:

$$\Delta g_z = \frac{GMz}{(x^2 + z^2)^{3/2}}$$

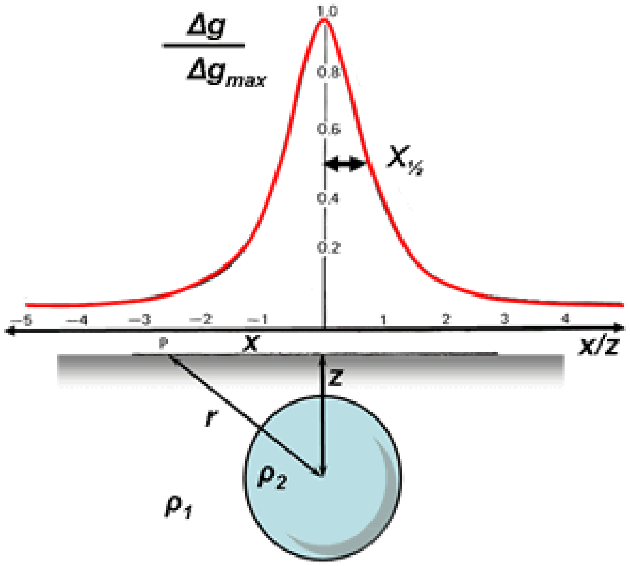


This equation describes the gravity signal measured as we move across the surface above the buried mass.

Why the gravity curve looks like a bell shape?

The previous equation tells us how the signal varies along a profile. When the instrument is directly above the mass (x=0), the distance to the mass is simply z. The gravity anomaly reaches its maximum value. As we move away from the mass, the distance r increases and the measured attraction decreases rapidly.

$$\Delta g_z = \frac{GMz}{(x^2 + z^2)^{3/2}}$$

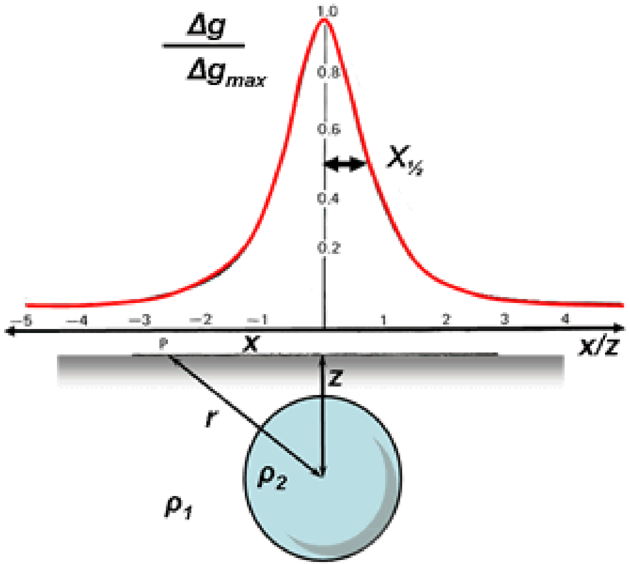


This equation describes the gravity signal measured as we move across the surface above the buried mass.

Why density does not appear in the equation?

The equation contains the mass M, but in geology we usually do not know the exact mass of the underground body. Instead we describe it using density contrast. Density contrast is simply the difference between the density of the anomalous body and the surrounding rocks:

$$\Delta\rho = \rho_{body} - \rho_{host}$$



For a spherical body of radius R:

$$M = \frac{4}{3}\pi R^3 \Delta\rho$$

Corrections of gravity

Raw gravity measurements are not enough! The gravity measured by the instrument is influenced by many factors that are not related to geology.

These include (not all):

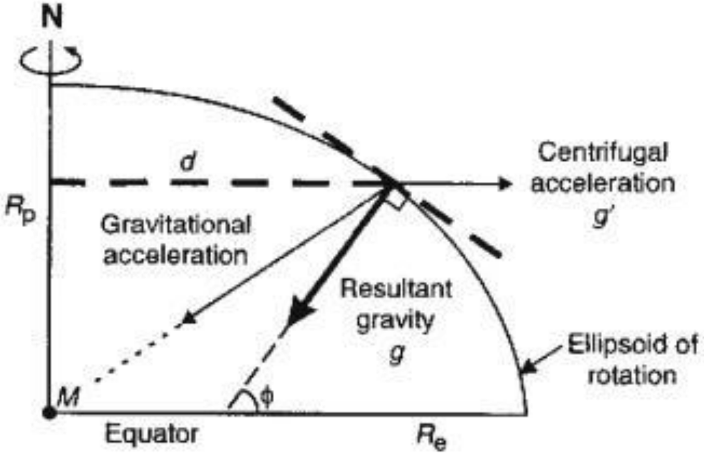
1. Latitude correction
2. Elevation (free-air) correction
3. Bouguer correction
4. Tidal correction

If we want to isolate the geological signal, we must apply a series of corrections to the measured gravity. Only after these corrections can we obtain the quantity used in exploration: **the Bouguer anomaly.**

Latitude correction

Gravity is not the same everywhere on Earth. Because the Earth rotates and is slightly flattened at the poles, gravity varies with latitude. Gravity is weaker near the equator and stronger near the poles.

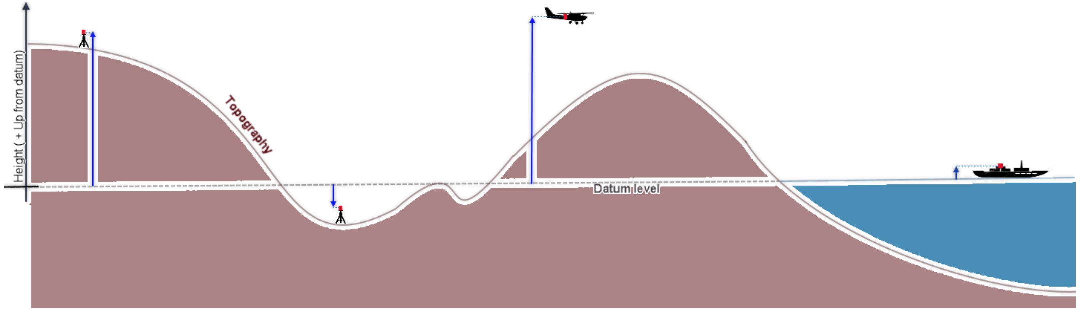
For this reason, we calculate what the normal gravitational field should be at a given latitude and subtract it from the measured value. This removes the large-scale variation caused by Earth's shape and rotation.



Elevation (free-air) correction

Gravity decreases with distance from the center of the Earth. If a measurement is taken at a higher elevation, gravity will naturally be slightly smaller.

To compare measurements taken at different elevations, we correct for this effect using the free-air correction. This correction adjusts the measured value to what gravity would be if the measurement had been taken at sea level.

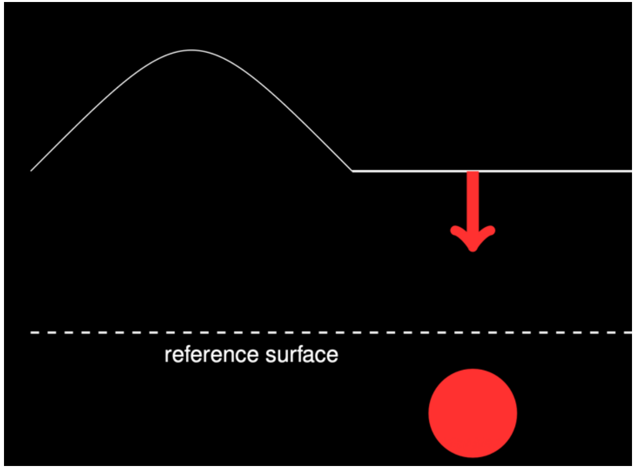


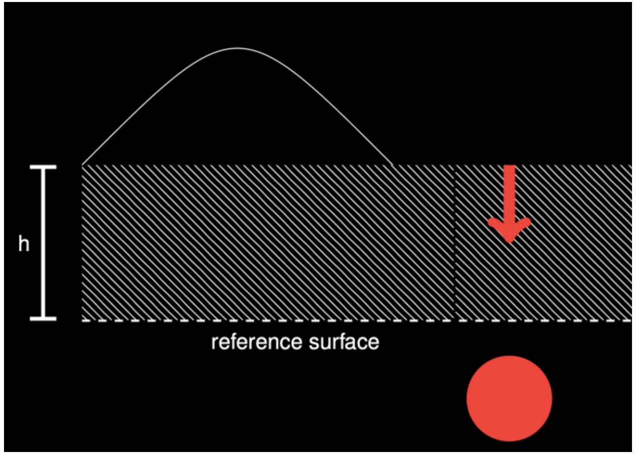
Bouguer Correction

When a measurement is taken at elevation h, there is a column of rock between the station and sea level. That rock has mass and therefore contributes additional gravitational attraction. To remove this effect we subtract the gravitational attraction of this rock slab.

This is called the Bouguer correction. The correction depends on the assumed density of the rocks and is commonly written as

$$\Delta g_B = 2\pi G\rho h$$



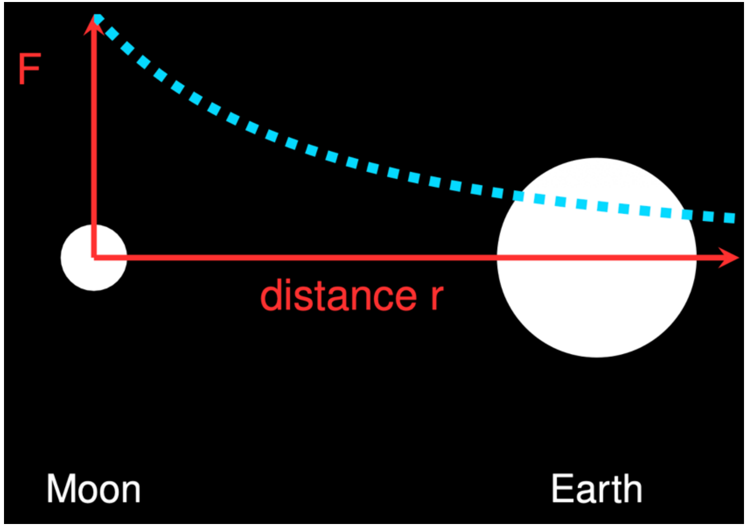


Tidal Correction

The gravity measured at a station is not constant in time. Gravity changes slightly during the day. The Moon pulls the Earth toward it, and the Sun also exerts a gravitational pull.

These forces slightly deform the Earth and change the gravitational acceleration measured at the surface. As a result, gravity at a fixed location can vary by roughly ±0.2 to 0.3 mGal during a day.

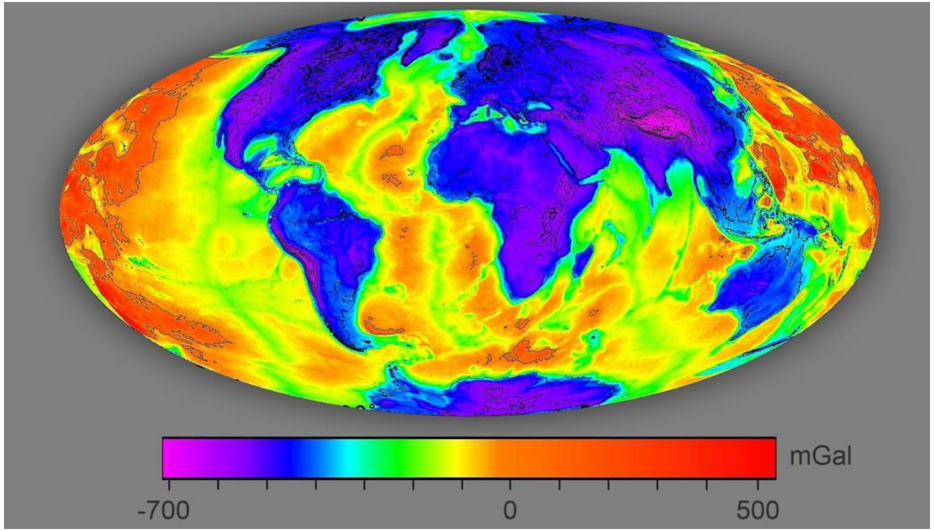
For comparison, many geological anomalies we want to detect are only a few mGal, so tidal effects are significant and must be removed.



Results? Bouguer anomaly!

After applying all corrections, we obtain the so-called Bouguer anomaly! Bouguer anomaly reflects **lateral density contrasts** in the subsurface. *The Bouguer anomaly represents the difference between the observed gravity and the gravity expected from a smooth, uniform Earth.*

- **Positive anomaly** → denser-than-average material below the station (e.g., mafic intrusion, ore body)
- **Negative anomaly** → less-dense-than-average material (e.g., granite batholith, salt dome, sedimentary basin)

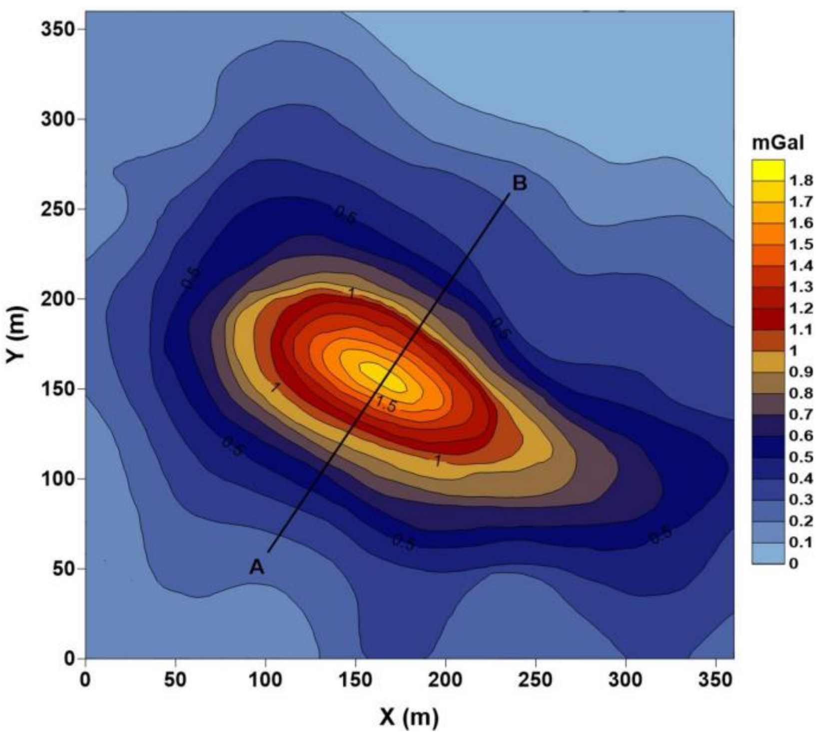


Results? Bouguer anomaly!

The Bouguer anomaly is the geophysicist's "density X-ray" of the crust.

The key challenge is non-uniqueness: multiple density models can explain the same anomaly.

Independent geological or seismic constraints are essential.



Non-uniqueness: the fundamental ambiguity

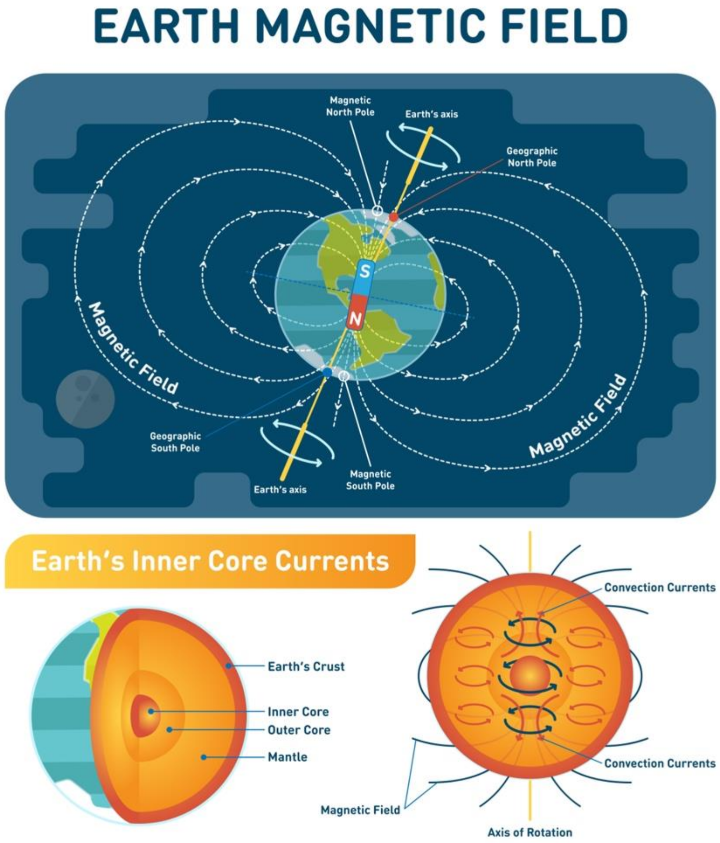
A **key limitation of gravity** (and all **potential field**) methods is non-uniqueness: many different subsurface density distributions can produce exactly the same surface anomaly. Consequences:

- A single gravity profile cannot uniquely determine depth AND density AND geometry simultaneously
- Additional constraints are essential: borehole data, seismic profiles, geological mapping
- Forward modelling produces a model consistent with the data, not the unique model

From gravity to magnetics: a parallel structure

Gravity and magnetic methods share the same mathematical framework but measure different physical properties.

Earth's magnetic field is generated by convection of molten iron in the outer core, a “geodynamo”. To first order, it resembles a magnetic dipole tilted ~11° from the rotation axis.



Earth's magnetic field

Singapore:
F = 42,500 nT
I = 14.07 deg
D = 0.15 deg

The field is characterised by:

- **Total intensity F**: ~25,000–65,000 nanoTesla (nT, varies by location)
- **Inclination I**: angle below horizontal (positive downward in northern hemisphere)
- **Declination D**: angle from geographic north

The **IGRF** (International Geomagnetic Reference Field) is a spherical harmonic model of the main field, updated every 5 years.

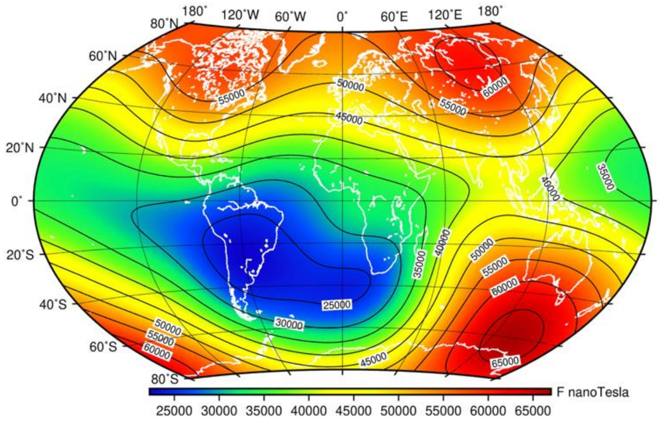


Figure 10: Map of total intensity at 2020.0

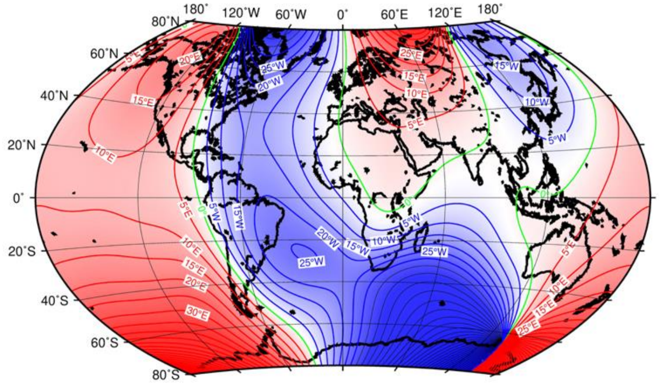


Figure 2: Map of declination (degrees East or West of true north) at 2020.0

Rock magnetism: what creates magnetic <u>anomalies</u>?

Magnetic anomalies arise from **magnetisation contrasts** between rock types. Rocks contain magnetic minerals in varying amounts:

- **Magnetite (Fe_3O_4), Ilmenite, pyrrhotite, Hematite**
- **Most sediments**: very low susceptibility (~10^{-5} SI) (effectively transparent magnetically)

Application: construction of a magnetic anomaly

Btot = Be + Bi + Br

measured → Btot

IGRF → Be

Bi + Br: What we are interested in during surveys = anomaly!

Btot total magnetic field

Be Earth magnetic field

Bi induced magnetisation

Br remanent magnetisation

Application: construction of a magnetic anomaly

$$\mathbf{Btot} = \mathbf{Be} + \mathbf{Bi} + \mathbf{Br}$$

measured (Btot)

IGRF (Be)

What we are interested in during surveys = anomaly! (Bi + Br)

Induced magnetization: $\mathbf{Bi} = \chi\,\mathbf{Be}$

Bi induced magnetisation
χ susceptibility
Be ambient magnetic field

χ can vary over several orders of magnitude:

- small negative for diamagnetic material (e.g., gold, silver)
- small positive for paramagnetic material (e.g., lithium)
- high positive for ferromagnetic material (e.g., iron, nickel, cobalt)

χHematite > χBasalt > χRhyolite > χSandstone > χChalk

χmetal ore > χbasic igneous> χacid igneous > χsediments

Application: construction of a magnetic anomaly

Susceptibility of geological materials

Rock type	Susceptibility	Br/Bi
Sedimentary rocks	0.00005	< 0.1
Metamorphic rocks	0.0003	< 1
Granites and rhyolites	0.0005	~ 1
Gabbros and basalts	0.006	≥ 10
Ultrabasic rocks	0.012	

Application: construction of a magnetic anomaly

Btot = Be + Bi + Br

measured (Btot) — IGRF (Be) — Bi + Br: What we are interested in during surveys = anomaly!

Remanent magnetisation: Br

Remanent magnetisation (or remanence) is the magnetization left behind in a ferromagnetic material, such as iron or magnetic minerals, after an external magnetic field is removed

Induced vs. remanent magnetisation

Imagine placing a piece of iron near a magnet. Even if the iron was not magnetic before, the external magnetic field causes the magnetic domains inside the material to align slightly. The material becomes magnetised only because the field is present.

Rocks behave in the same way in the Earth's magnetic field. Magnetic minerals such as magnetite become weakly magnetised simply because they sit inside the Earth's field.

This is called induced magnetisation.

Its direction is always the same as the present Earth magnetic field, and its strength is proportional to that field.

Induced vs. remanent magnetisation

Some rocks acquired magnetisation when they formed, and they keep that magnetisation even after the external field changes.

A classic example occurs when magma cools. Above a certain temperature (the Curie temperature) magnetic minerals cannot hold a permanent magnetisation. When the rock cools below this temperature, the magnetic minerals lock in the direction of the Earth's magnetic field at that moment. That magnetisation can remain stable for millions of years.

This is called remanent magnetisation.

The key difference is that remanent magnetisation does not depend on the present Earth field.

What do we actually do in a magnetic survey?

A magnetic survey measures small variations in the Earth's magnetic field across an area in order to detect magnetic minerals in the subsurface. In practice, the instrument measures the total magnetic field **Btot** at many locations along survey lines.

$$\mathbf{Btot} = \mathbf{Be} + \underbrace{\mathbf{Bi} + \mathbf{Br}}$$

measured (Btot)

IGRF (Be)

What we are interested in during surveys = anomaly! (Bi + Br)

Be is the regional magnetic field generated inside the Earth's core.
Bi and **Br** are the contributions produced by magnetised rocks underground.

Magnetic survey step-by-step

Step 1: planning the survey – define the survey line(s)

Step 2: Measuring the magnetic field in the field, including the total magnetic field Btot, the geographic coordinates (GPS), and the time of measurement.

Step 3: Removing the regional magnetic field

Step 4: Mapping and interpreting the magnetic anomaly