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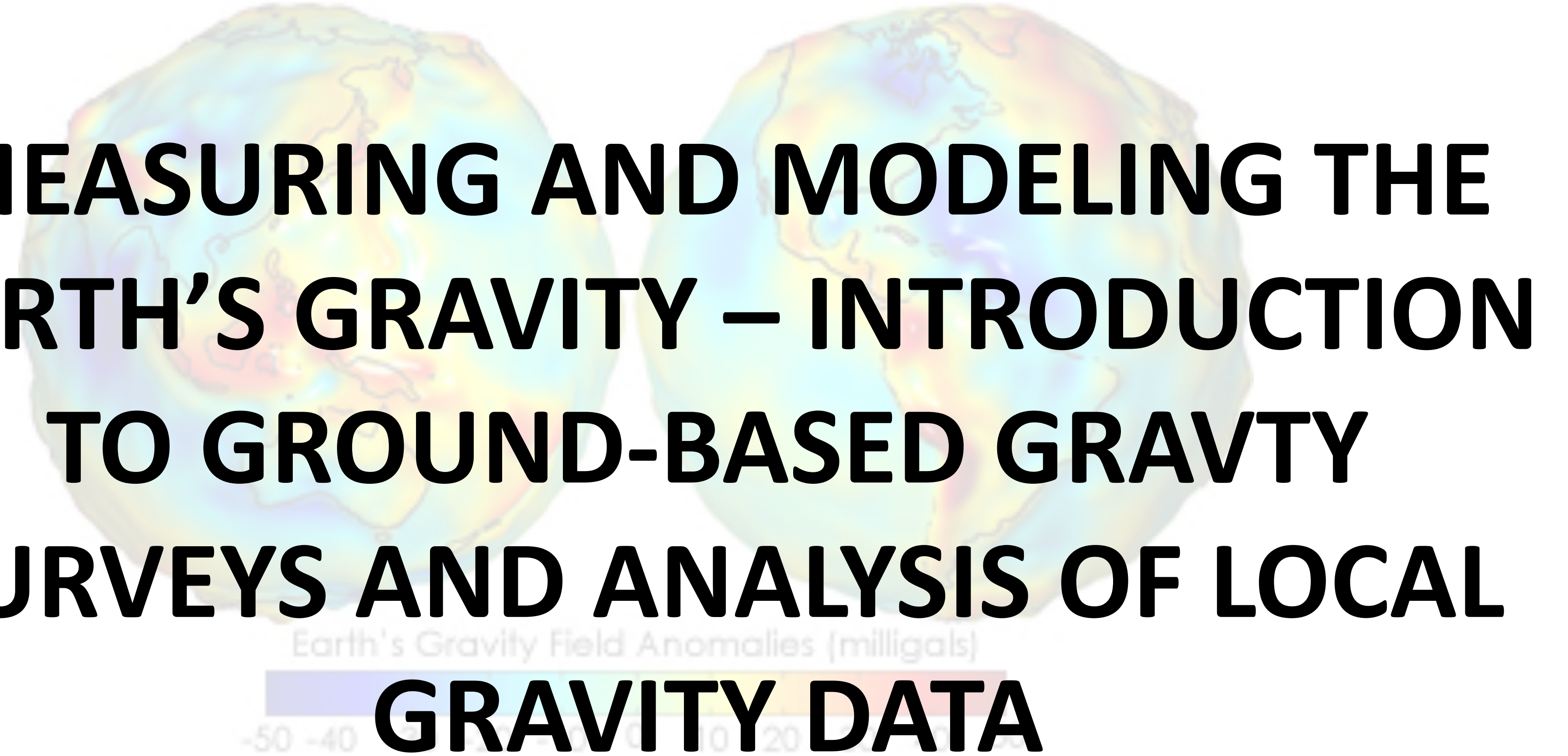
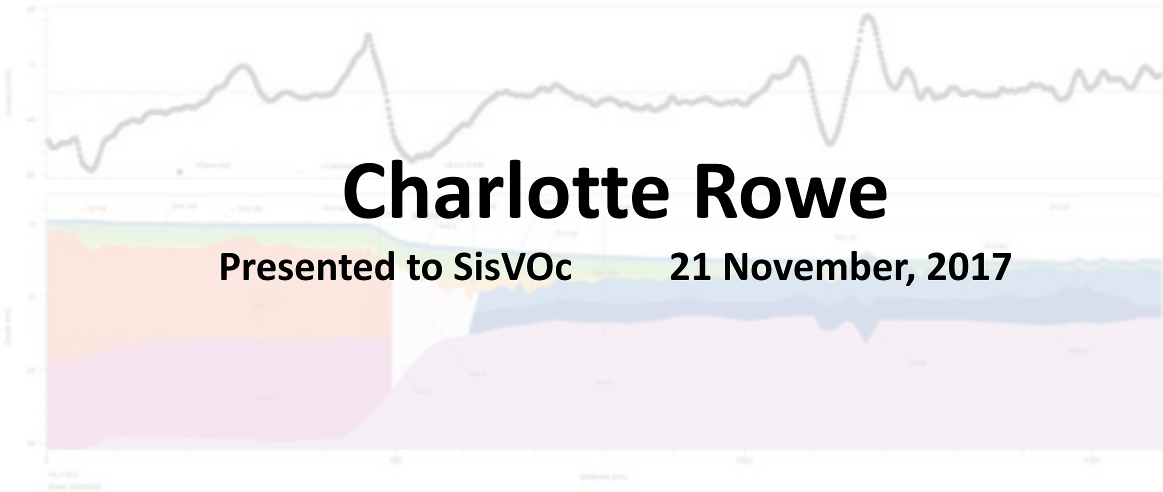
MEASURING AND MODELING THE EARTH'S GRAVITY – INTRODUCTION TO GROUND-BASED GRAVTY SURVEYS AND ANALYSIS OF LOCAL GRAVITY DATA



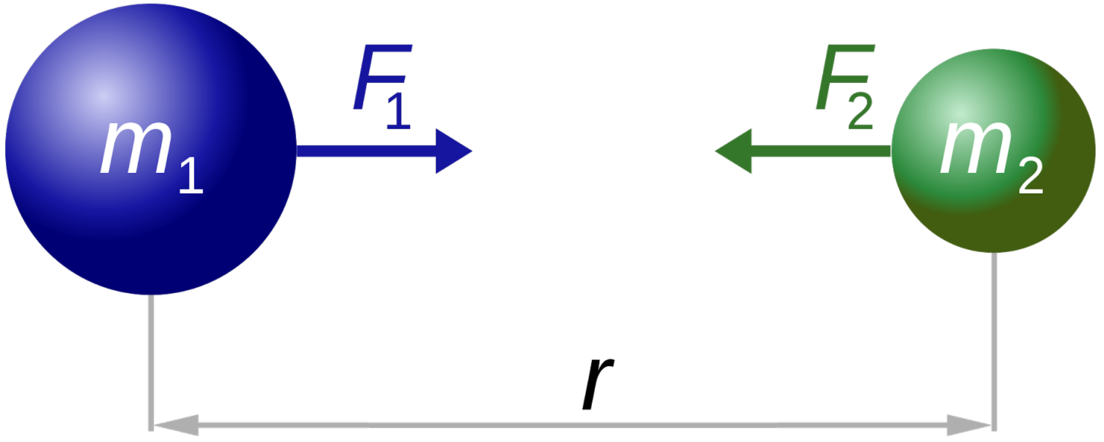


Charlotte Rowe

Presented to SisVOc **21 November, 2017**

Gravitational Force and Measuring Gravity

Gravitational Force



Gravitational attractive force between two masses, m_1 and m_2 at a distance r from one another:

$$F_1 = F_2 = G\frac{m_1 \times m_2}{r^2}$$

Where the Universal Gravity Constant:
$G = 6.674\times10^{-11}\ m^3 \cdot kg^{-1} \cdot s^{-2}$

Acceleration due to gravity measured on the Earth:

$\boldsymbol{g} = GM_E/r_E^2$, where M_E is that mass of the Earth and r_E^2 is the distance from the center of the Earth

On average this is ~ 9.81 m/s^2 (or 981 Gals). But we know that the gravitational acceleration varies from place to place on the earth.

Measurement of Absolute Gravity:

Pendulum Method: Measure the period $T = 2\pi\sqrt{\frac{I}{mgh}} = 2\pi\sqrt{\frac{L}{g}}$

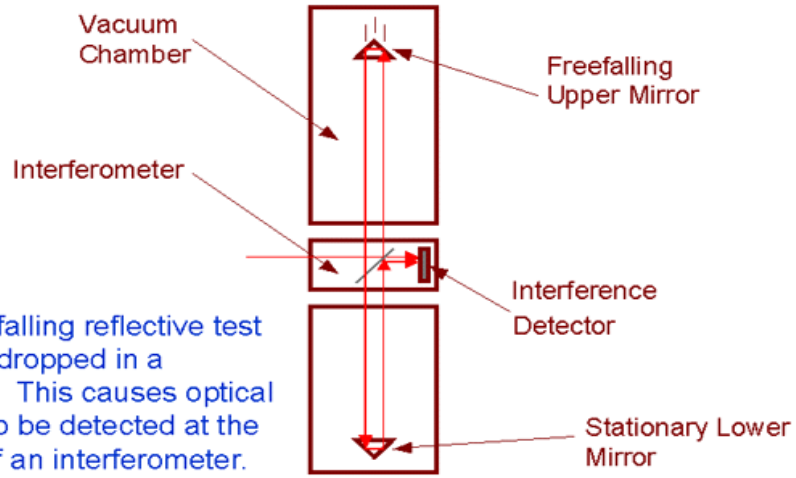
To measure 1 mgal variation, the period must be measured to within 1μs.

Free-fall Method: Measure the fall of a mass: $z = z_0 + ut + gt^2/2$

To measure 1 μgal variation, time must be measured to within 1ns.

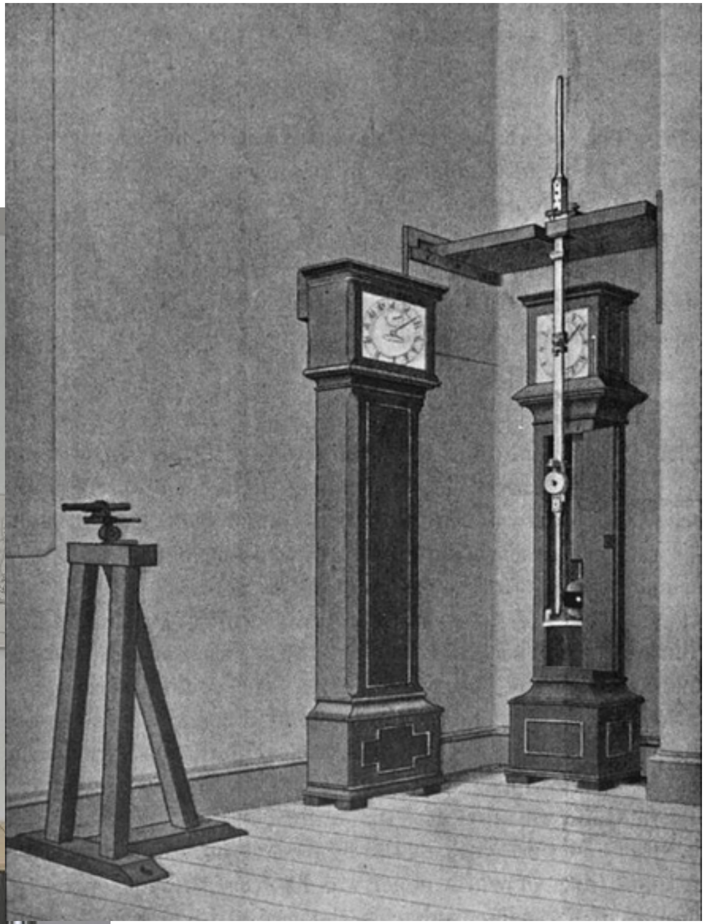
Rise-and-fall Method: Measure time T for a thrown ball to rise and fall a height z: $z = g(T/2)^2/2$. Then $g = \frac{8(z_1 - z_2)}{(T_1^2 - T_2^2)}$. μgal precision; not portable.

FG-5 Principle of Operation



A freely falling reflective test mass is dropped in a vacuum. This causes optical fringes to be detected at the output of an interferometer. This signal is used to determine the local gravitational acceleration.

Micro-g-LaCoste free fall absolute gravimeter. These devices are becoming sufficiently stable and portable to be used in the field.



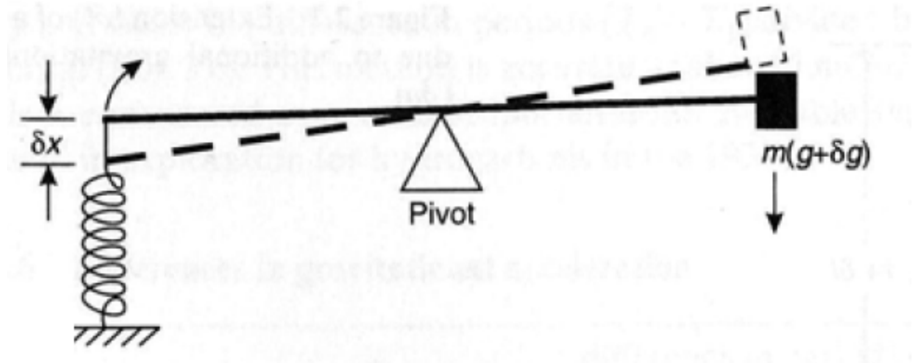
Example pendulum absolute gravity meters.

Relative Gravimeters

Principle operation of a stable gravimeter

These devices are less sensitive than the unstable types. Typical examples of manufacturers are

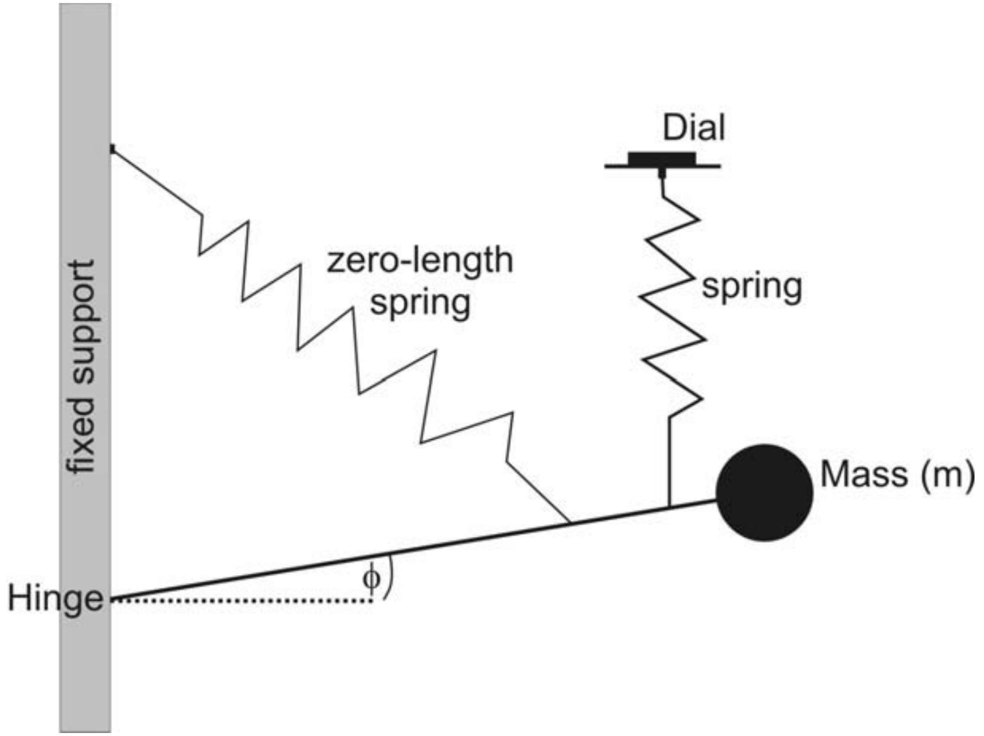
Askania, Boliden, Gulf



Principle operation of an unstable gravimeter

This is the type of gravimeter in use for most gravity surveying these days. Typical makers are

Thyssen, LaCoste-Romberg, Worden

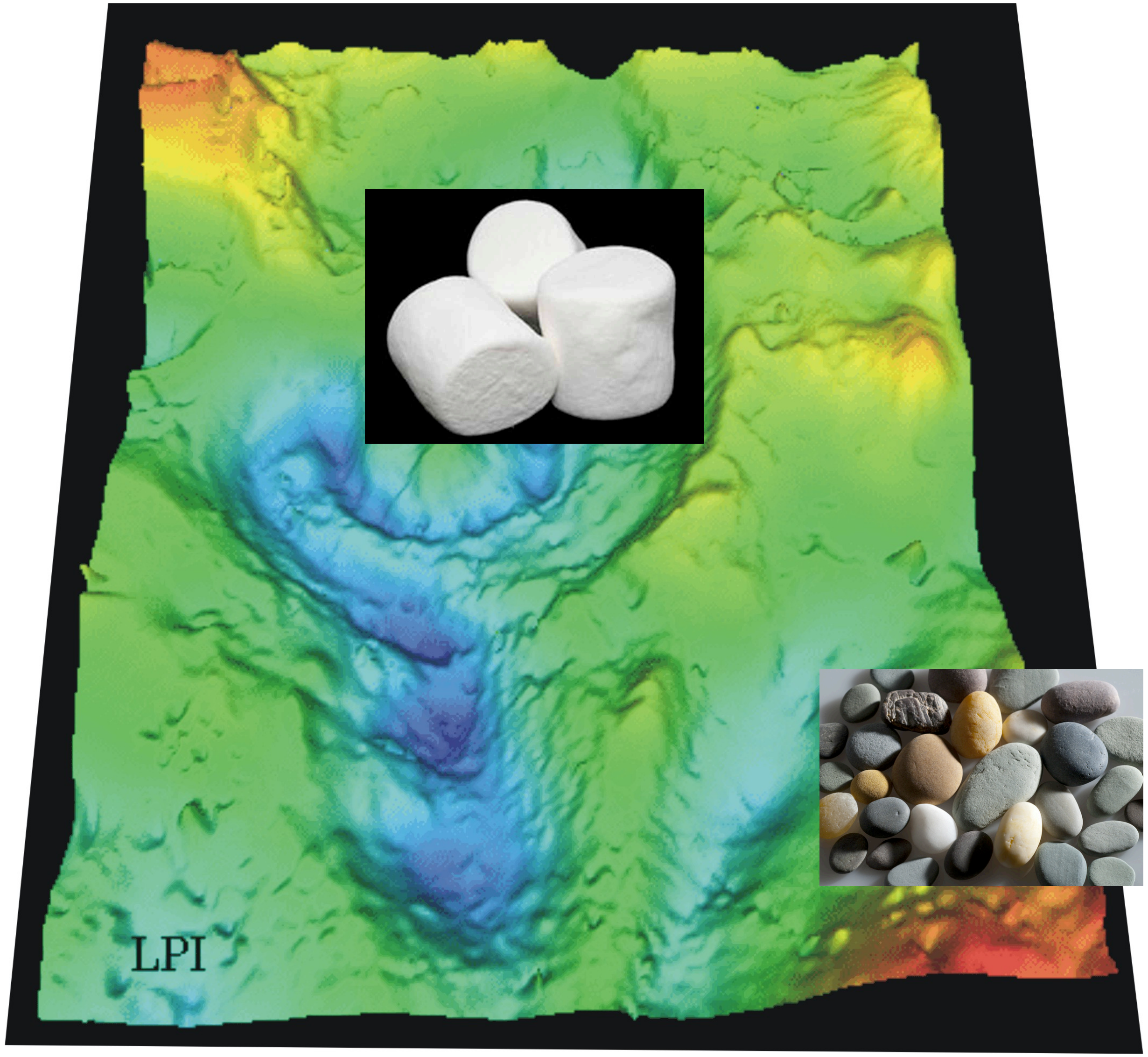


Gravity surveys seek to determine variations in the **density** of the subsurface.

In gravity surveys we measure ***g***, the acceleration due to gravity. ***g*** varies with elevation, latitude, topography, tides, and near-surface density. We make a number of corrections to our observations to produce a map of gravity anomalies focusing on local variations in near-surface density

Salt domes, sedimentary basins,
mine shafts, caves, molten rock, marshmallows
= gravity low

Metalic ore bodies, geologic anticlines, dense rocks
= gravity high



Conducting a Gravity Survey

Gravity anomalies are very small in amplitude compared to the main field

Gravity is usually measured in mgal or gu (gravity unit)
1mgal = 1×10^{-5} ms^{-2}
1 gu = 1×10^{-6} ms^{-2}

Accurate gravity surveying is a slow process, but it is not difficult.

- Level gravimeter carefully
- Measure height accurately
- About 20 mins per reading
- Return to base every 1-2 hours

Station spacing and size of the survey area depends on the estimated size of the anomalous body whose gravity signature you wish to characterize. There may be hundreds of stations in a small area if you are looking or small features.

Required Corrections to Gravity Measurements

Measuring the gravity does not give us the whole story.

Measurements in the field are influenced by many external factors, masking the view of the real target until the effect of each factor is removed from our readings:

(a) Instrumental drift
(b) Tide
(c) Latitude
(d) Free Air
(e) Bouguer
(f) Terrain
(g) Isostatic

These are the corrections made to the data.

Additionally for gravimeters mounted on ships or aircraft we have to apply an another correction, called the Eötvös correction, which addresses the effect of the Coriolis force of the rotating Earth and our velocity increasing, or decreasing relative to it when we travel East or West.

Many things perturb gravity readings and mask the signature of a targeted anomaly.

Gravity surveys can be done on the ground, from drones, from airplanes and using orbiting satellites. We will not address airborne or satellite surveying methods here, other than to point out a few key considerations:

- The amplitudes of gravity anomalies are lower when they are measured from aircraft or from space. Moreover, the farther the measurement is from the source, the more spread the anomaly is. This makes short-period anomalies from small targets nearly impossible to separate, although methods such as "upward continuation" can be used to try to model how such anomalies might appear at altitude.
- Continuous data collection in gravity surveys in the air or at sea requires accounting for the Earth's rotational spin, and any Eastward or Westward velocity of the vessel, which will either increase or decrease the gravitational influence of the Coriolis effect. This is the Eötvös correction.

Eötvös Correction: Moving eastward at v_E, your angular velocity increases by: $\Delta\omega = v_E / (R_E \cos\lambda)$. This change increases the centrigugal acceleration:

$$\Delta a_C = \left(\frac{da_C}{d\omega}\right)\Delta\omega = \left(2\omega R_E \cos\lambda\right)\left(\frac{v_E}{R_E \cos\lambda}\right) = 2\omega v_E.$$

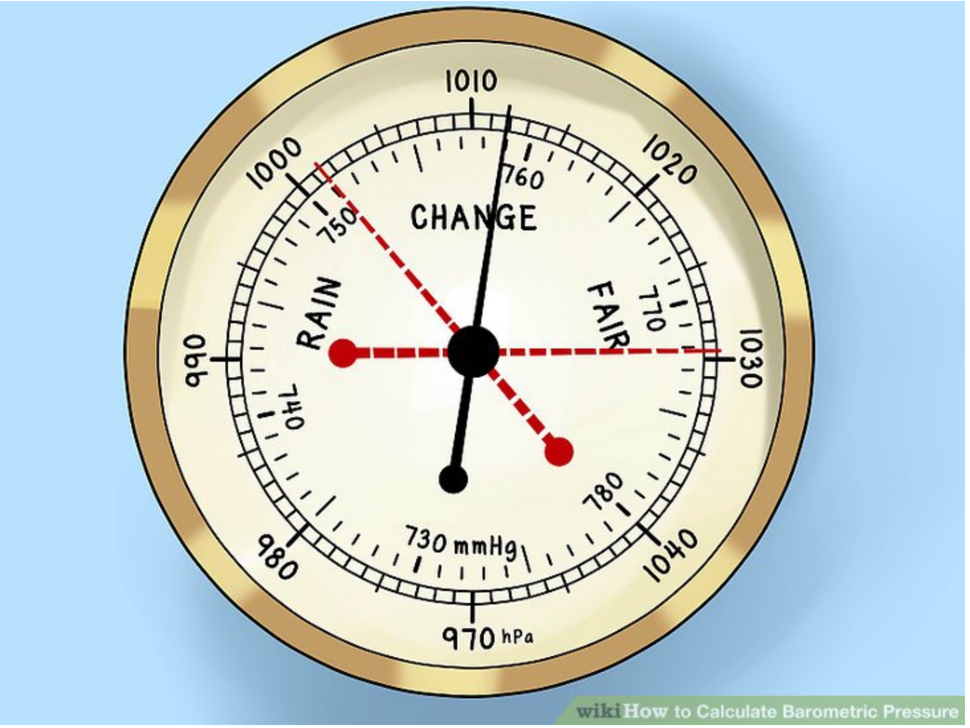
Downward gravity changes by:

$\Delta g = -2\omega v_E \cos\lambda$. *The Eötvös effect decreases gravity when moving east.*

Barometric Pressure Corrections

On days with a normal weather pattern, barometric pressure variations are in the range of 0.3 – 1 kPa (i.e. 1 – 3 μGals) per day. There will be times, however, when a major pressure front (e.g. a thunderstorm) moves rapidly through the survey area. Such a weather system can give rise to pressure changes totalling 5 kPa (i.e. 18 μGals) in amplitude, with temporal gradients of the order of 0.5 kPa/hr (1.8 μGal/hr) and spatial gradients of the order of 0.2 kPa (0.7 μGal) in 10 km distances.

For most surveys, the effects of barometric pressure can be neglected, but if microgravity measurements are being conducted (for example, tracking underground fluid migration such as groundwater or magma), it may be necessary to remove the barometric pressure effects from gravity readings.



Latitude (North or South)

Because the Earth is not a sphere, but instead is a flattened ellipsoid, gravitational pull increases as we move toward the poles from the equator – because we are closer to the center of the Earth..

A latitude-correction might become important if your survey covers large areas and crosses significant distances in a N-S orientation.

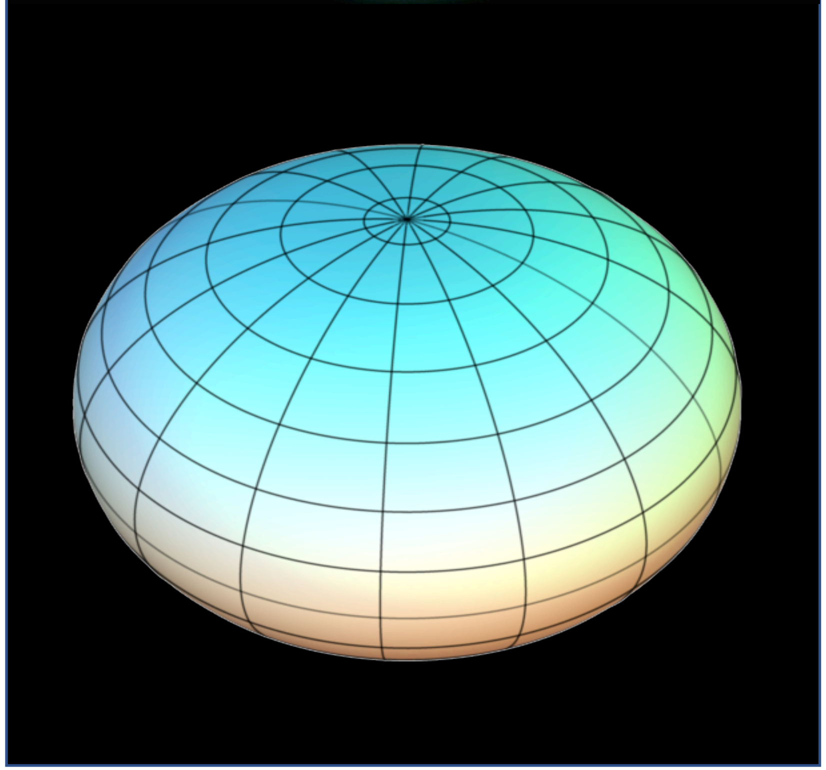
The international standard of gravitational acceleration as a function of latitude φ [NOTE: φ is in rad, not degrees. Multiply with π/180 to use degrees]:

$$g0(\varphi) = 9.78031846\ [1+0.005278895 \sin^2\varphi - 0.0000023462 \sin^4(\varphi)]\ \mathrm{m/s^2}$$

For a small, local survey, it is possible to simply apply a linear gravity gradient as a function of the distance north or south of the base station,

$\delta gL = -8.108 \sin 2\phi$. (units per km N)

This correction is subtracted from readings as you move away from the Equator.

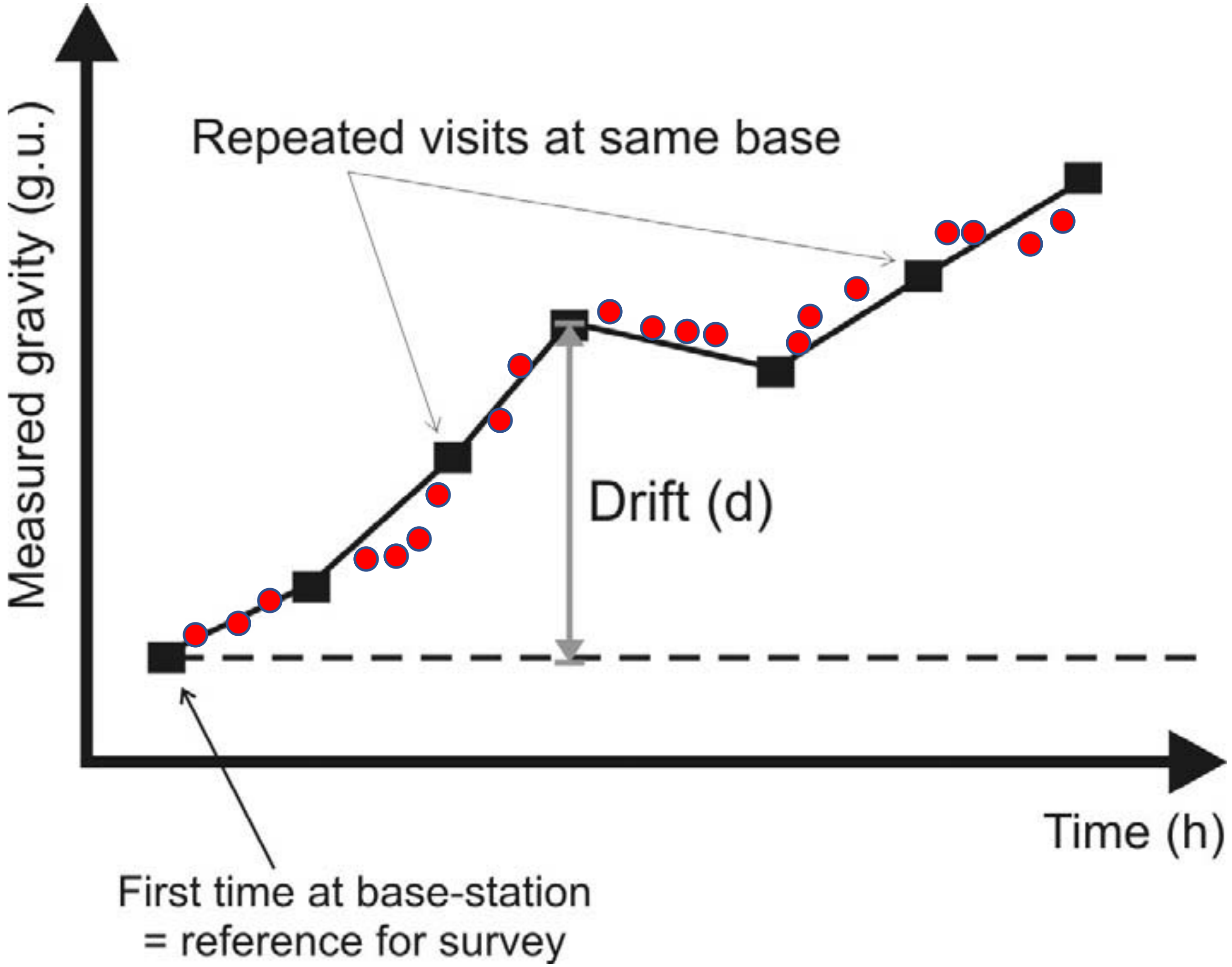


Drift

One problem that affects all gravity readings is instrumental “drift.” This is a slow change in the *apparent* gravitational acceleration that comes from physical changes in the instrument itself, largely as a result of temperature* changes. These instruments are very sensitive and a slight change in the temperature will cause changes (expansion or contraction) of components within the instrument, causing it to give us a different reading.

We usually address instrumental drift by taking repeated measurements at one “base” station. Except for the changes in gravity due to the tides (which we will discuss shortly), gravity at this location should not change.

* Modern gravimeters contain internal temperature controls to reduce this effect.



Removing the (assumed) linear drift between the base station readings (black boxes) allows us to correct readings at other stations (red circles) to remove instrumental drift.

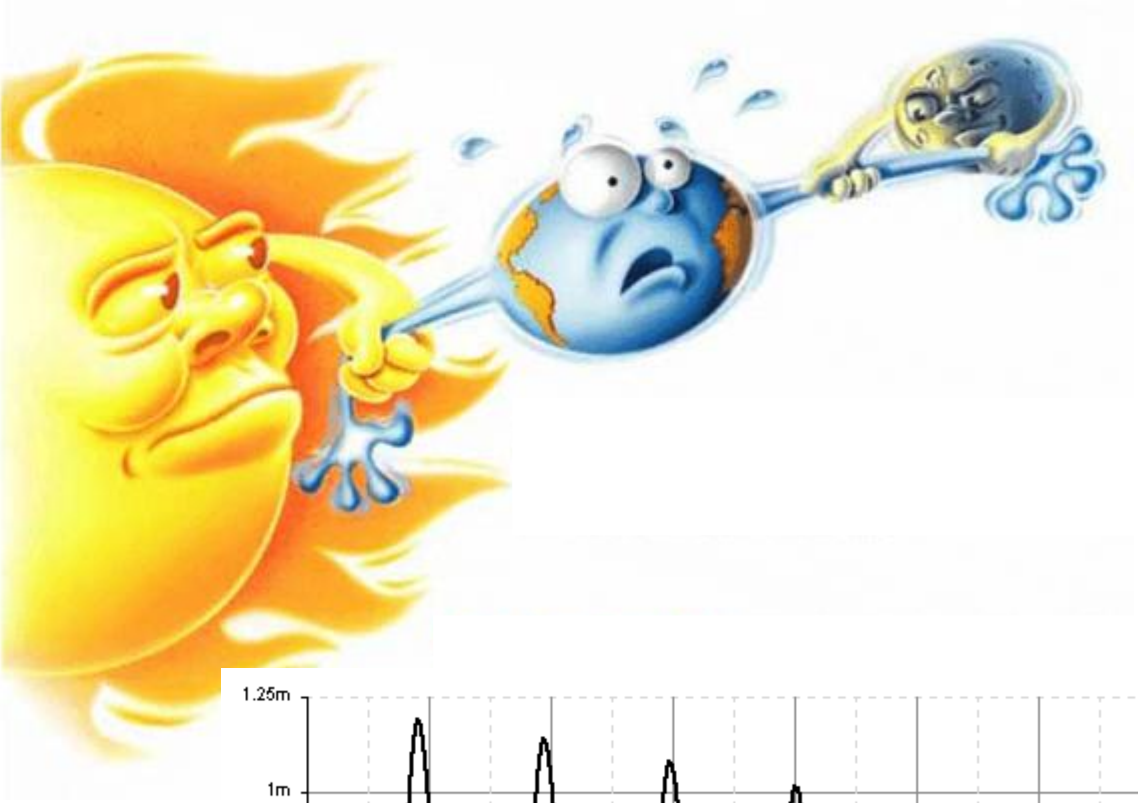
Tidal deformation of Earth due to Sun and Moon

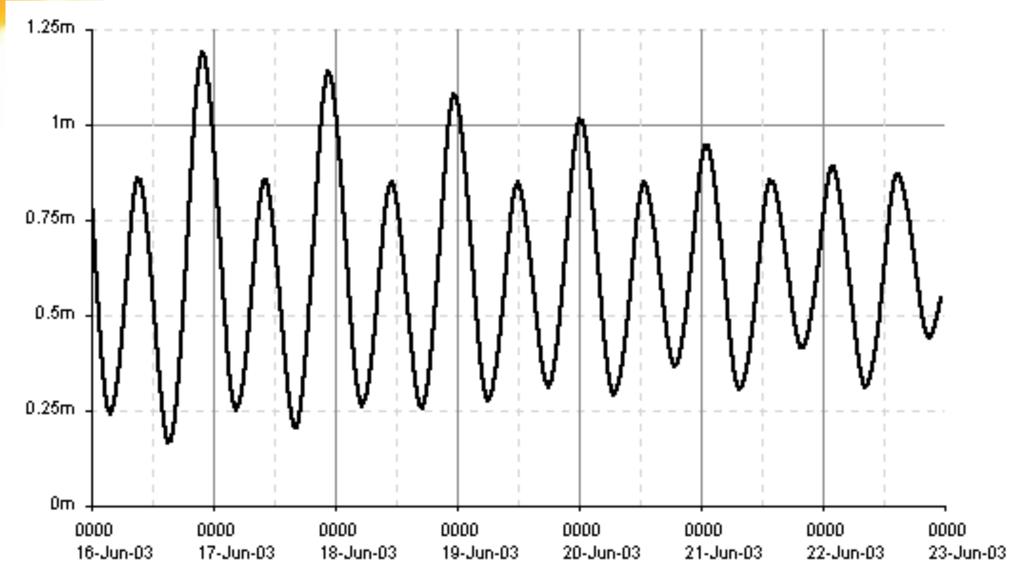
The shape of the Earth changes frequently due to *tidal deformation* caused by the gravitational attraction of the Moon and Sun.

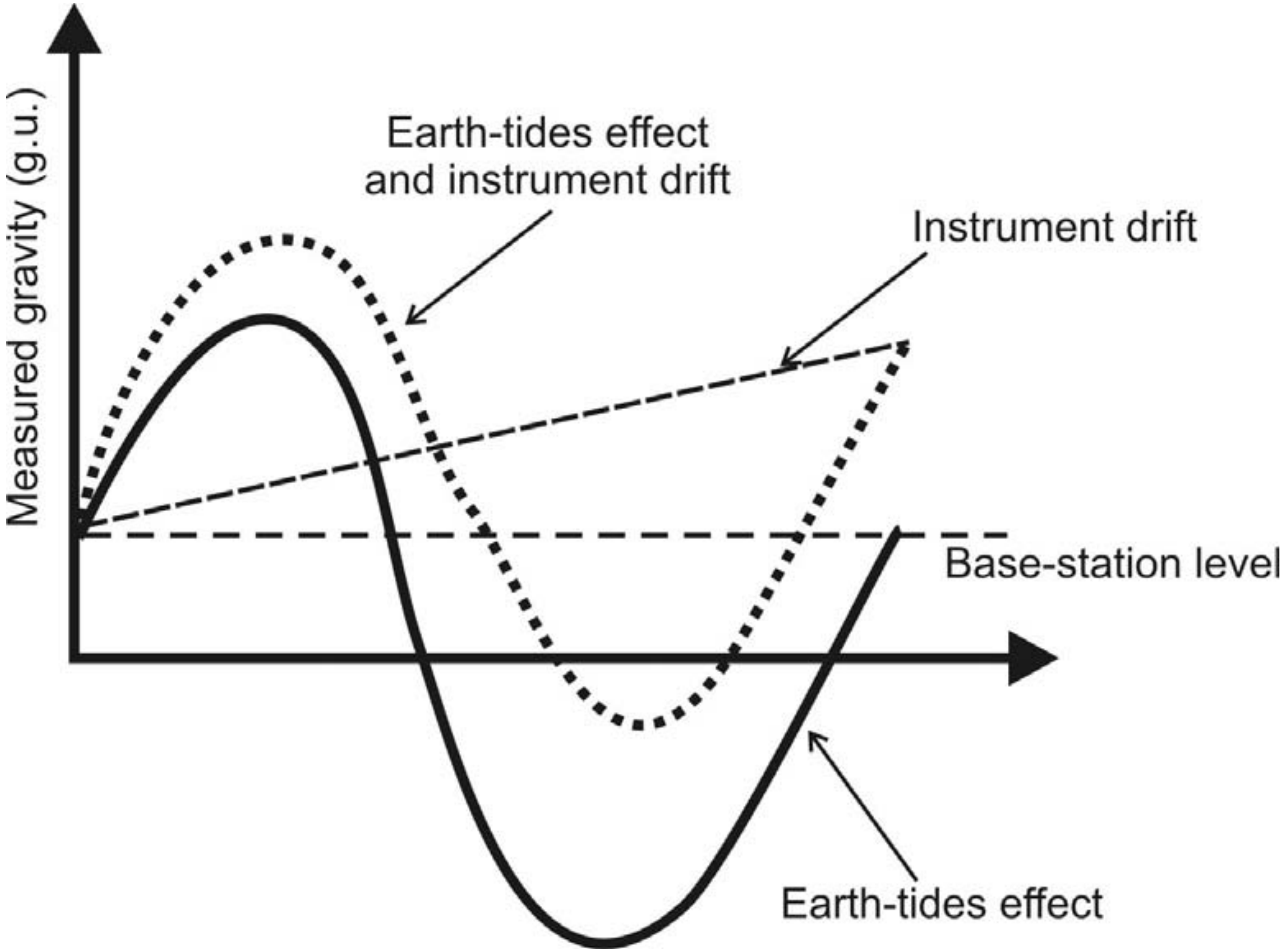
The sea-tides are a well-known phenomenon; there are two high and two low tides every day.

The gravitational effect of Sun and Moon is not always aligned, but if they act in the same plane, the tides are much larger. We call those “spring tides.” If Sun and Moon are at a right angle (90^{o}) to one another, and their effects partially cancel one another, we get “neap tides.”

Tidal maxamia cause the gravimeter to be farther from the center of the earth, which means the gravity reading at “high tide” is lower. These effects, too, have to be removed from the data.



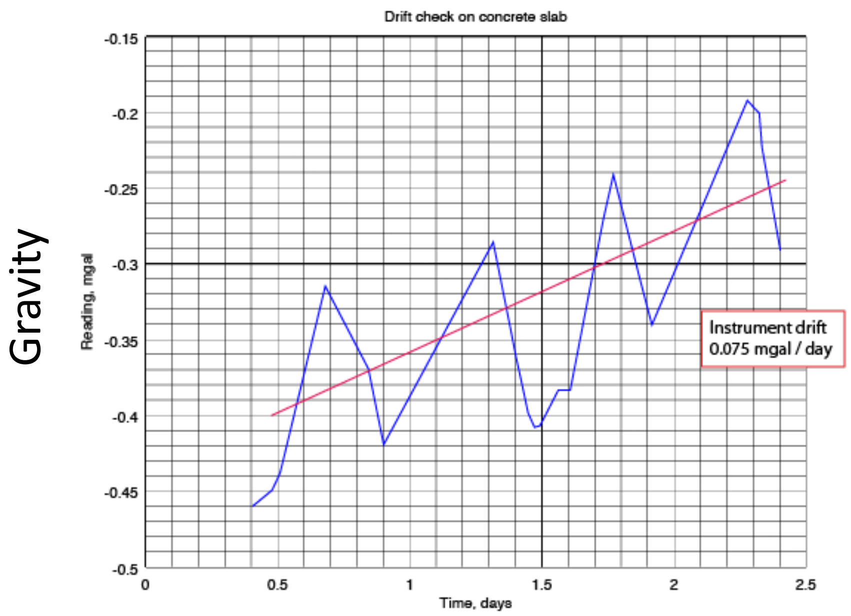


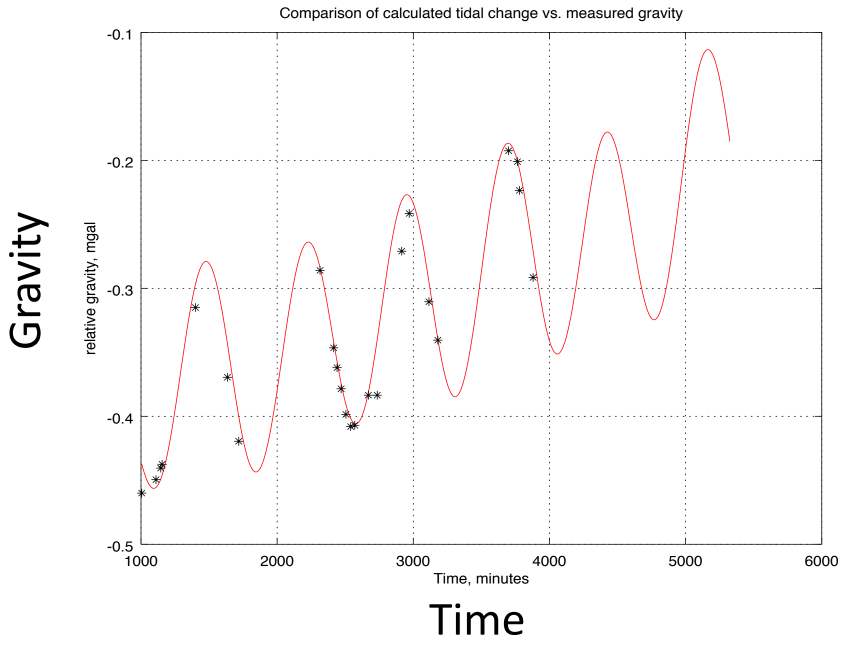


The base station can move (depending on latitude) up to 30 cm up and down with respect to the center of the earth, twice a day. Thus, the resulting gravity readings would look like a sinusoid.

Of course we also have to deal with the instrumental drift. An example of combined drift and tide is shown as the dotted line.

Shown here are several gravity readings I made with a gravimeter at my house.





The readings define a “sawtooth” pattern (with a slope). These readings were made in exactly the same place, over and over again over several days.

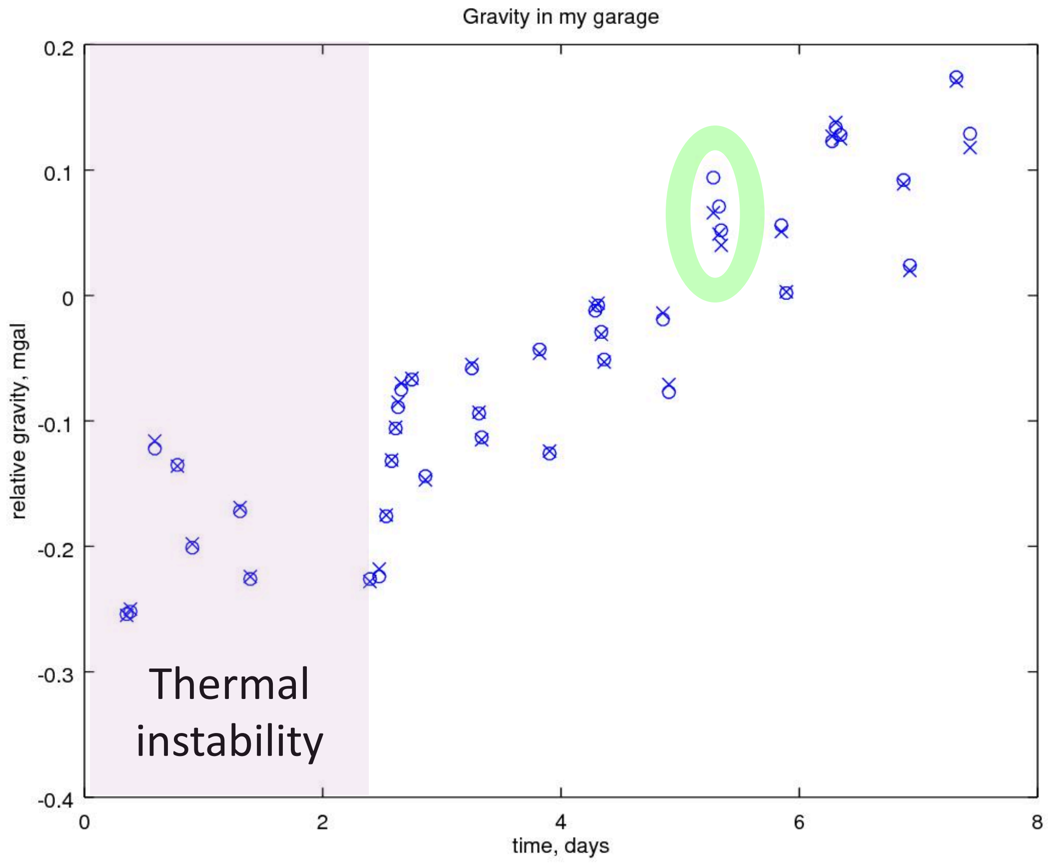
The slope (red line) is an estimate of the slope of this sawtooth function, which provides a first order approximation of the instrumental drift

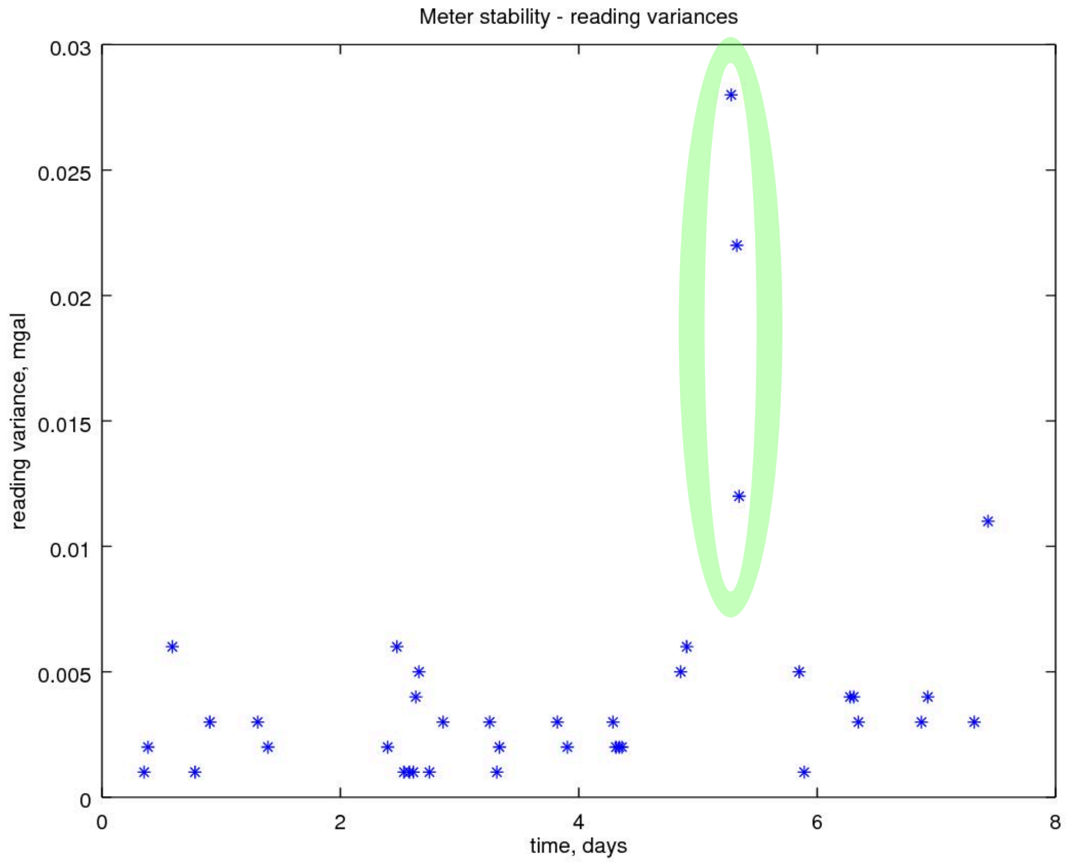
Here I plot the data again, with the estimated tidal gravity function as well, adjusted by the linear drift function I determined above. The tidal function is easily and accurately calculated, and can be predicted for any point on earth at any time in the past or future.

It appears the gravimeter is stable and is measuring the gravity inside my garage consistently.

Other Problems from Afar

Here is another drift test (left) taken with the gravimeter in a single location for eight days. You can see the quasi-linear drift as well as the sinusoidal function due to the tides. Each reading was estimated twice, and these numbers were recorded. Note at 4.5 days, the pairs of readings become inconsistent for awhile. If we plot (right) the variance of the readings we see a spike at about 4.5 days, where the two readings vary by as much as 30 μgal – while at other times the variation is less than 5 μgal. What could cause this?



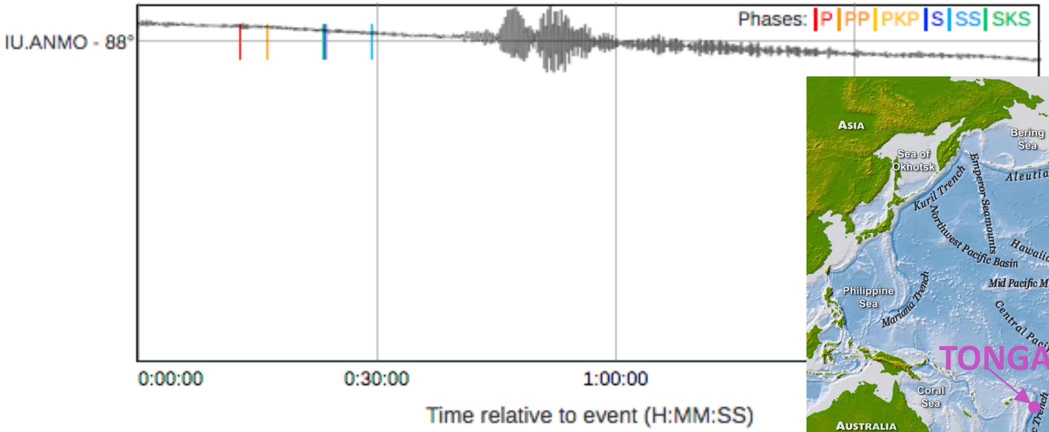


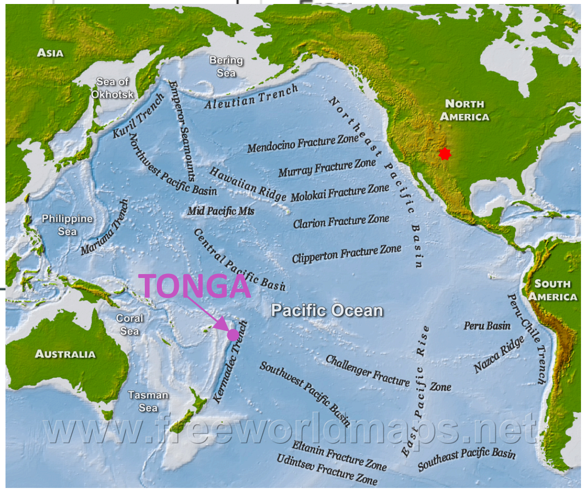
Puzzle solved!

A quick visit to the USGS global earthquakes website shows the problem: An earthquake of M 5.1 in Tonga occurred earlier that morning. The waveform for the nearest GSN broadband seismic station to the gravimeter (ANMO) was found at IRIS. It shows that the surface waves for this earthquake arrived at just the right time to cause trouble for the gravity readings.

DATE and TIME (UTC)	LAT	LON	MAG	DEPTH km	DIST km	LOCATION (Shows interactive map)	IRIS ID (Other info)
19-SEP-2017 15:34:23	-24.11	-179.80	4.2	500	439	SOUTH OF FIJI ISLANDS	10412638
20-SEP-2017 17:24:15	-17.76	-178.55	4.3	547	762	FIJI ISLANDS REGION	10412240
21-SEP-2017 00:05:20	-18.33	-175.34	5.0	223	628	TONGA ISLANDS	10406908
21-SEP-2017 00:43:43	-18.32	-177.95	4.5	525	679	FIJI ISLANDS REGION	10406921
21-SEP-2017 07:09:28	-16.62	-174.58	4.3	235	824	TONGA ISLANDS	10412836
21-SEP-2017 22:49:27	-23.73	-179.93	4.3	551	453	SOUTH OF FIJI ISLANDS	10411540
23-SEP-2017 00:28:53	-17.90	-178.49	4.4	580	745	FIJI ISLANDS REGION	10407544
23-SEP-2017 06:02:17	-24.28	-179.86	4.5	526	446	SOUTH OF FIJI ISLANDS	10408872
24-SEP-2017 12:59:00	-21.41	-176.68	4.5	7	311	FIJI ISLANDS REGION	10411580
25-SEP-2017 11:47:57	-18.00	-178.50	4.6	540	735	FIJI ISLANDS REGION	10408096
25-SEP-2017 11:49:22	-17.86	-178.46	4.2	596	748	FIJI ISLANDS REGION	10411581
25-SEP-2017 22:00:02	-19.64	-177.47	4.3	546	524	FIJI ISLANDS REGION	10411411
26-SEP-2017 01:42:42	-18.06	-178.29	5.1	579	720	FIJI ISLANDS REGION	10408478
26-SEP-2017 04:20:00	-23.71	-176.94	6.4	98	151	SOUTH OF FIJI ISLANDS	10408504
26-SEP-2017 21:29:06	-23.59	-179.95	5.4	538	457	SOUTH OF FIJI ISLANDS	10408743
27-SEP-2017 07:55:44	-21.27	-179.18	4.3	610	485	FIJI ISLANDS REGION	10408910
28-SEP-2017 23:11:15	-24.87	-179.79	5.4	441	447	SOUTH OF FIJI ISLANDS	10409462
29-SEP-2017 05:30:30	-24.08	-179.52	4.5	546	410	SOUTH OF FIJI ISLANDS	10409556
01-OCT-2017 10:06:55	-24.48	-176.69	5.1	88	135	SOUTH OF FIJI ISLANDS	10410070
01-OCT-2017 14:08:00	-22.80	-174.53	5.0	10	163	TONGA ISLANDS REGION	10410117
01-OCT-2017 15:10:04	-18.28	-177.82	4.2	600	679	FIJI ISLANDS REGION	10410143
02-OCT-2017 23:44:05	-15.45	-175.13	4.3	256	949	TONGA ISLANDS	10413057
07-OCT-2017 17:54:38	-18.98	-175.74	4.6	195	557	TONGA ISLANDS	10411835
07-OCT-2017 23:59:44	-20.29	-178.46	4.6	614	512	FIJI ISLANDS REGION	10411864
08-OCT-2017 14:04:38	-19.01	-175.56	6.1	10	553	TONGA ISLANDS	10411942
11-OCT-2017 13:15:03	-18.34	-172.14	5.0	10	716	TONGA ISLANDS REGION	10412569
17-OCT-2017 09:53:42	-20.56	-175.55	5.1	128	380	TONGA ISLANDS	10413291
17-OCT-2017 11:27:59	*-23.98*	*-175.48*	*5.1*	*10*	*0*	*TONGA ISLANDS REGION*	10413306

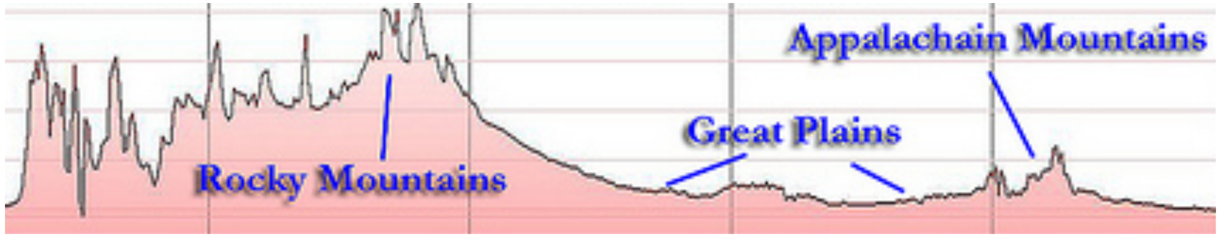
Latitude	Longitude	Date	Depth	Magnitude
23.9803° S	175.4848° W	2017-10-17 11:27:59 UTC	10.0 km	mb5.1



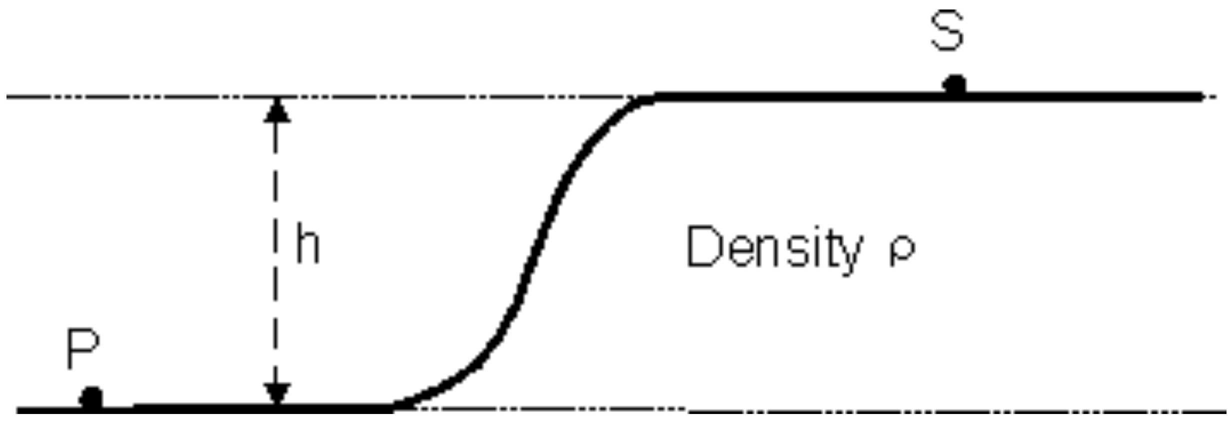


A gravimeter works by balancing a mass on a spring, and is a very sensitive long-period seismometer. It's easily disturbed by long-period surface waves from a large, distant earthquake. This seismic record shows that for a time period of about an hour, there would be no point trying to get high-precision gravity readings on that morning. A larger earthquake would last longer.

Elevation Effects



Free Air correction

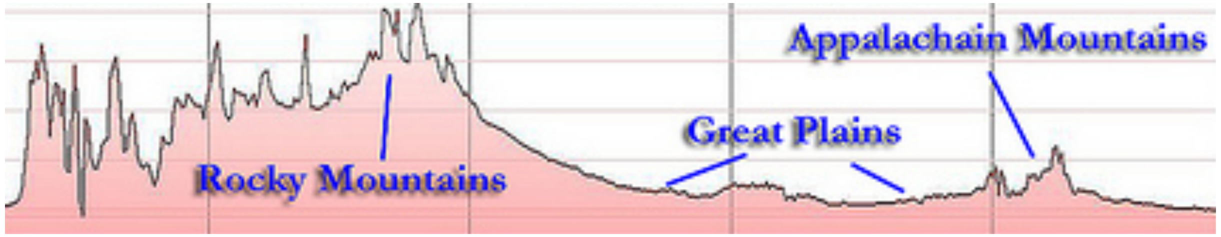


The basis of this correction is the reduction in magnitude of gravity with height, irrespective of the kind of rocks beneath. The free-air correction is the difference between gravity measured at sea level (or other datum) and at an elevation *h*.

Commonly a value of 0.3086 mGal per meter is acceptable for most practical applications. The free-air correction term varies with latitude from 0.3083 mGal per meter at the equator to 0.3088 mGal per meter at the poles. With a normal measuring precision of 0.1 mGal or less for most modern gravimeters, the station elevation needs to be known within 3 –5 cm.

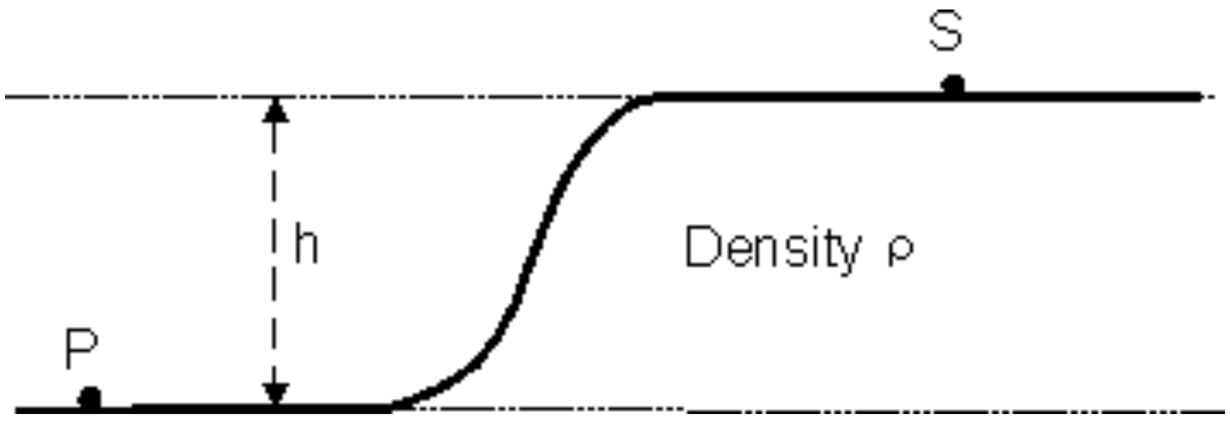
This means that when a ground based gravity survey is performed, the topography must be known very precisely.

Elevation Effects



Bouguer correction

As with the free air correction, the Bouguer correction targets the change in height of the gravity reading position compared to some selected datum, but this correction is opposite in polatiry. The Bouguer correction accounts for the added mass beneath the gravimeter as a result of being on top of a hill instead of in a valley.



This correction therefore is subtracted from the gravity reading, whereas the Free Air correction is always added to the reading.

The Bouguer correction can vary depending on the density of the rocks that are being corrected for. It assumes an infinite slab of material of the chosen thickness and density, and reduces the theoretical downward pull of that slab.

BC = -.00004191 ρh in gravity units, or .0004191 ρh in mGal, where ρ is the density of the slab.

Finding the right density for Bouguer Correction

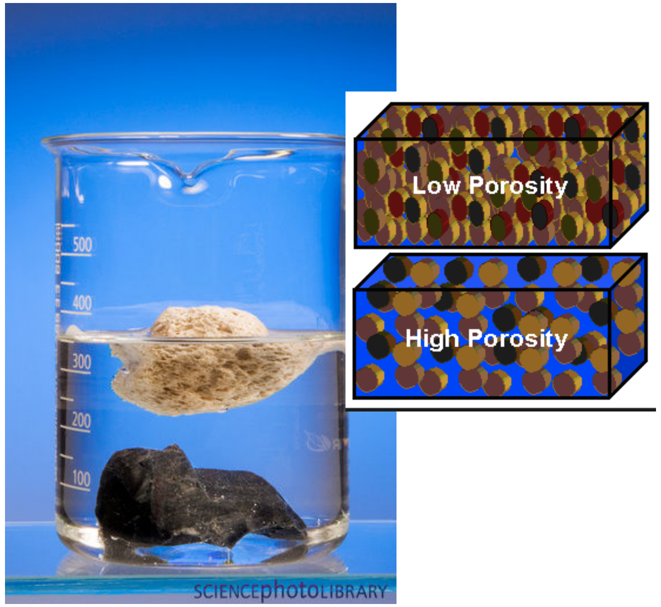
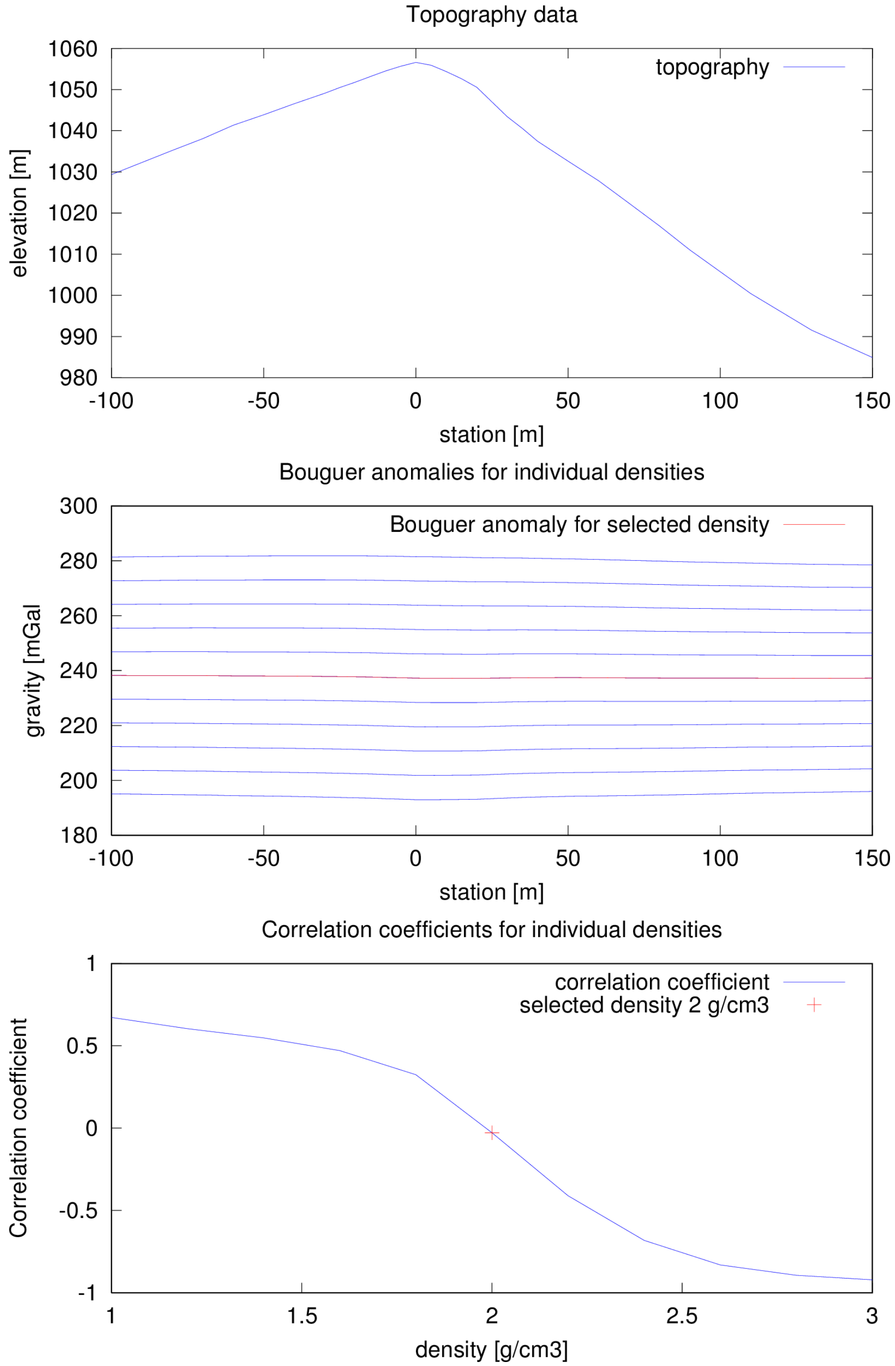


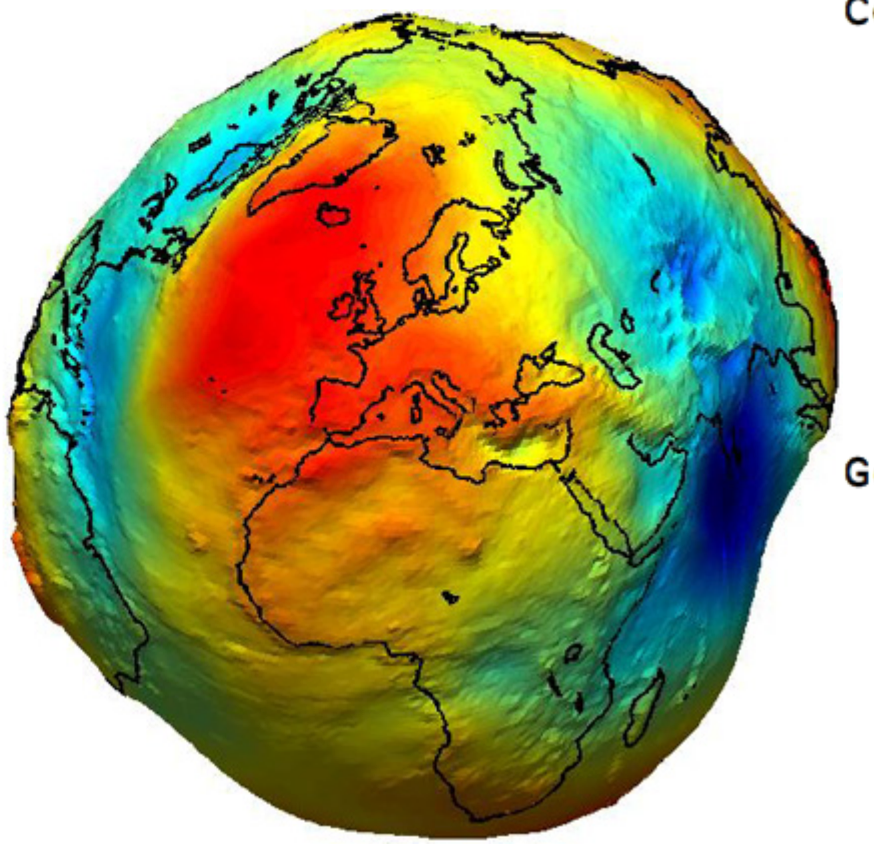
Illustration of the Nettleton's method for density estimate.

Alluvium (wet)	1.96–2.00
Clay	1.63–2.60
Shale	2.06–2.66
Sandstone	
Cretaceous	2.05–2.35
Triassic	2.25–2.30
Carboniferous	2.35–2.55
Limestone	2.60–2.80
Chalk	1.94–2.23
Dolomite	2.28–2.90
Halite	2.10–2.40
Granite	2.52–2.75
Granodiorite	2.67–2.79
Anorthosite	2.61–2.75
Basalt	2.70–3.20
Gabbro	2.85–3.12
Gneiss	2.61–2.99
Quartzite	2.60–2.70
Amphibolite	2.79–3.14
Chromite	4.30–4.60
Pyrrhotite	4.50–4.80
Magnetite	4.90–5.20
Pyrite	4.90–5.20
Cassiterite	6.80–7.10
Galena	7.40–7.60

The Bouguer curve that correlates *least* with the topography (either positively or negatively) is chosen as the correct density estimate.



Top: topography along a profile. Middle: Bouguer curves for different densities. The best density estimate is plotted in red. Bottom: topography and Bouguer curve correlation for different densitites.



Combined Correction: Free air and Bouguer corrections are often combined:

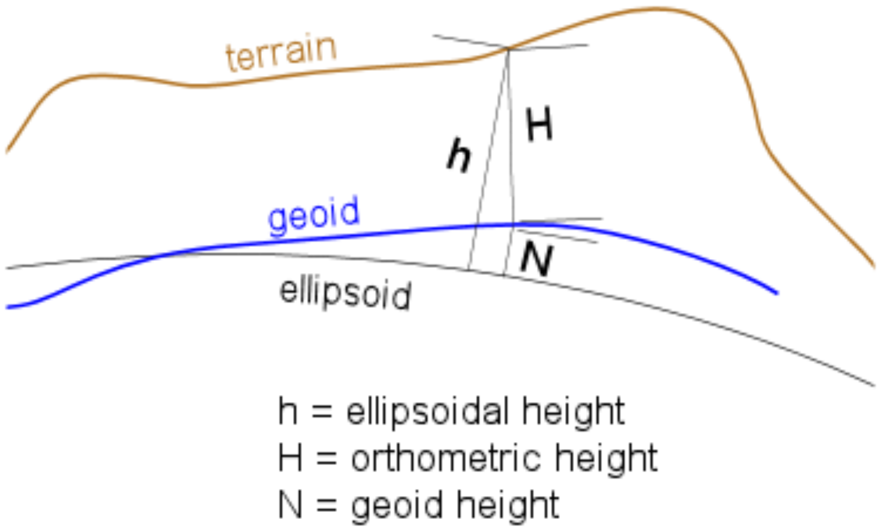
$$\Delta g_{FA} + \Delta g_{BP} = \left(0.3086 - 0.0419\rho \times 10^{-3}\right) \text{ mgal/m} = 0.197 \text{ mgal/m}$$

assuming a crustal density of 2670 kg/m^3. To obtain 0.01 mgal accuracy:

-- location must be known to within 10 m (for latitude correction)

-- elevation must be known to within 5 cm (for combined correction)

Geoid Correction: For long wavelength surveys, station heights must be corrected for the difference in gravity between the geoid height and the reference ellipsoid, which can vary spatially.

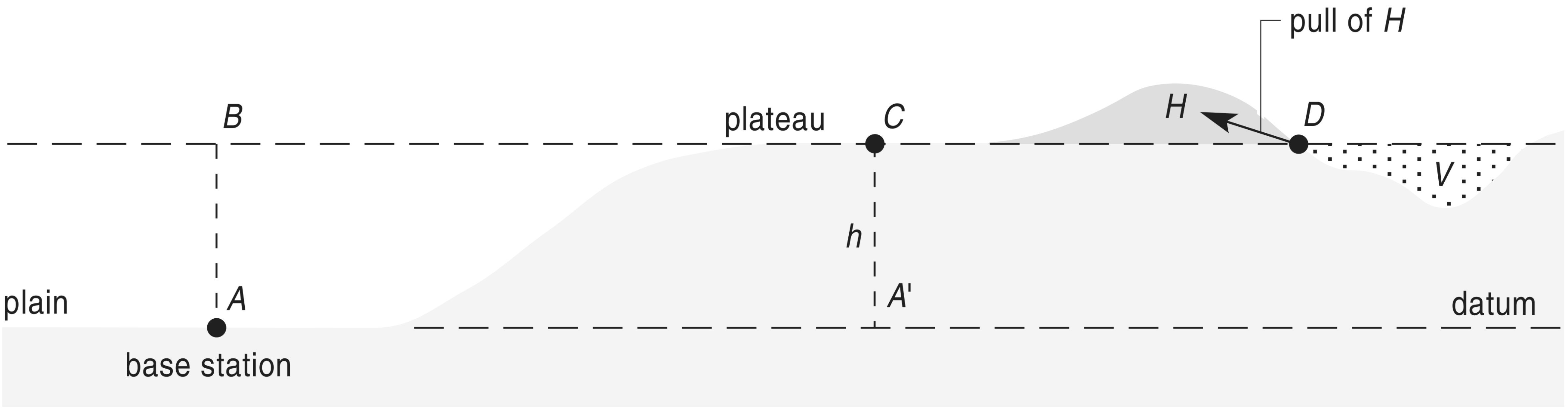


The geoid is defined as the surface of the earth's gravity field, which approximates mean sea level. It is perpendicular to the direction of gravity pull. Since the mass of the Earth is not uniform at all points, the magnitude of gravity varies, and the shape of the geoid is irregular.

Terrain correction

The effect of topography (valleys, hills) can be quite substantial. The hill with excess mass has its center of mass above the level of the valley, where the gravimeter station is located. The resulting attractive force by the hill slightly reduces the measured gravity at the station. Similarly the effect of a valley on gravity measurements with a station on top of a hill can be visualized by defining the missing mass of the valley relative to the hill as –M. Thus, the measurement of gravity on top of the hill is again underestimated.

We therefore must correct each gravity station's readings to account for the topography around the station. This correction needs to be made to account for the topography out to several kilometers, although the amplitude of the correction decreases with distance.

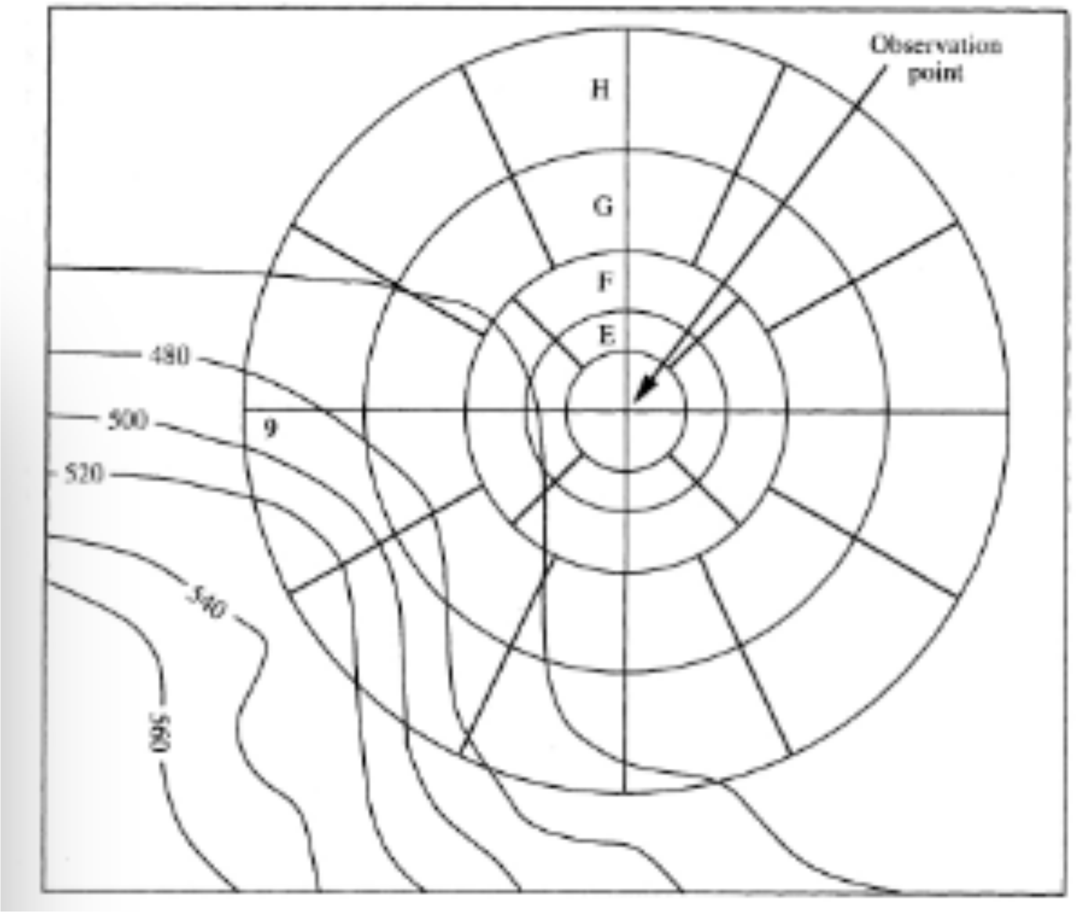


Terrain correction

is a complicated task. It requires knowledge of detailed topography as well as a good understanding of the density of the surrounding rocks (which may also be the target of the survey).

Historically a **Hammer Terrain correction** has been used, named after Sigmund Hammer (1939). This Hammer-corrections template consists of a series of segmented concentric rings, which are superimposed over a topographic map. The average elevation of each segment is estimated and each segment is given a constant background density, identical to the density used in the Bouguer correction. Today, GIS (Geographic Information-system) can help digitize topography and calculate average masses of the topography surrounding the stations. Modern gravity interpretation software usually comes with a tool for terrain correction, that can be linked to standard GIS data formats, or can import a Digital Elevation Map (DEM) to facilitate the corrections.

Hammer template



"Life's a drag, and then you die. But first, you have to do terrain corrections."

--- anonymous Geophysics Student, University of Alaska summer field camp, 1986

Each zone is a circular ring of given radii (in feet) divided into 4, 6, 8, 12, or 16 compartments of arbitrary azimuth. "h" is the mean topographic elevation in feet (without regard to sign) in each compartment with respect to the elevation of the station. The tables give the correction "T" for each compartment due to undulations of the terrain in units of 1/100 mg. for density $(\sigma) = 2.0$. This correction, when applied to Bouguer anomaly values which have been calculated with the simple Bouguer correction, is always positive.

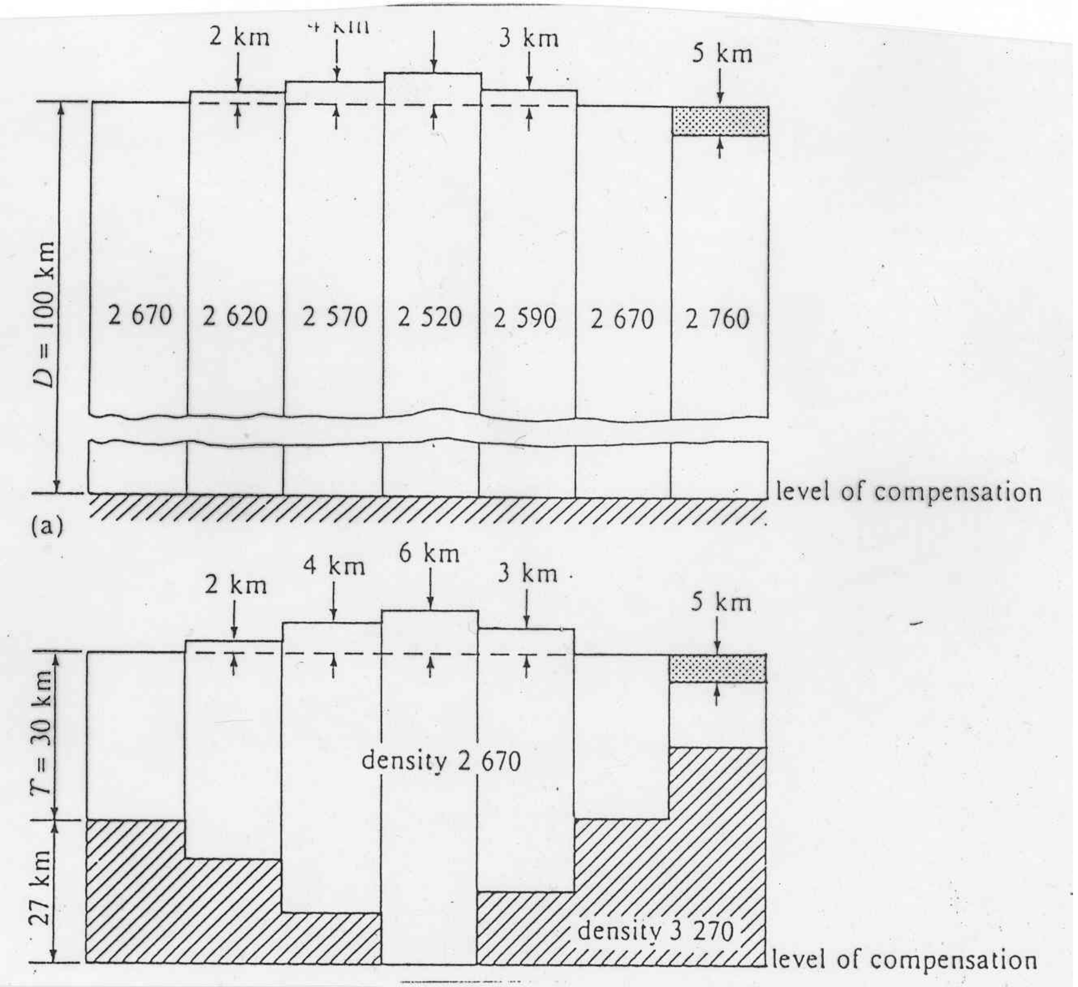
Zone B 4 compartments 6.56 to 54.6*		Zone C 6 compartments 54.6 to 175		Zone D 6 compartments 175 to 558		Zone E 8 compartments 558 to 1280		Zone F 8 compartments 1280 to 2936		Zone G 12 compartments 2936 to 5018	
$\pm h$ (ft.)	T	$\pm h$ (ft.)	T	$\pm h$ (ft.)	T	$\pm h$ (ft.)	T	$\pm h$ (ft.)	T	$\pm h$ (ft.)	T
0 to 1.1	0	0 to 4.3	0	0 to 7.7	0	0 to 18	0	0 to 27	0	0 to 58	0
1.1–1.9	0.1	4.3–7.5	0.1	7.7–13.4	0.1	18–30	0.1	27–46	0.1	58–100	0.1
1.9–2.5	0.2	7.5–9.7	0.2	13.4–17.3	0.2	30–39	0.2	46–60	0.2	100–129	0.2
2.5–2.9	0.3	9.7–11.5	0.3	17.3–20.5	0.3	39–47	0.3	60–71	0.3	129–153	0.3
2.9–3.4	0.4	11.5–13.1	0.4	20.5–23.2	0.4	47–53	0.4	71–80	0.4	153–173	0.4
3.4–3.7	0.5	13.1–14.5	0.5	23.2–25.7	0.5	53–58	0.5	80–88	0.5	173–191	0.5
3.7–7	1	14.5–24	1	25.7–43	1	58–97	1	88–146	1	191–317	1
7–9	2	24–32	2	43–56	2	97–126	2	146–189	2	317–410	2
9–12	3	32–39	3	56–66	3	126–148	3	189–224	3	410–486	3
12–14	4	39–45	4	66–76	4	148–170	4	224–255	4	486–552	4
14–16	5	45–51	5	76–84	5	170–189	5	255–282	5	552–611	5
16–19	6	51–57	6	84–92	6	189–206	6	282–308	6	611–666	6
19–21	7	57–63	7	92–100	7	206–222	7	308–331	7	666–716	7
21–24	8	63–68	8	100–107	8	222–238	8	331–353	8	716–764	8
24–27	9	68–74	9	107–114	9	238–252	9	353–374	9	764–809	9
27–30	10	74–80	10	114–120	10	252–266	10	374–394	10	809–852	10
		80–86	11	120–127	11	266–280	11	394–413	11	852–894	11
		86–91	12	127–133	12	280–293	12	413–431	12	894–933	12
		91–97	13	133–140	13	293–306	13	431–449	13	933–972	13
		97–104	14	140–146	14	306–318	14	449–466	14	972–1009	14
		104–110	15	146–152	15	318–331	15	466–483	15	1009–1046	15

* Radii of the zone in feet.

Isostasy

Get isostatic anomalies at foreland basins, oceanic ridges and post-glacial basins

and for all small scale features (these are not isostatically compensated)



Isostatic anomaly = observed Bouguer anomaly - expected Bouguer anomaly

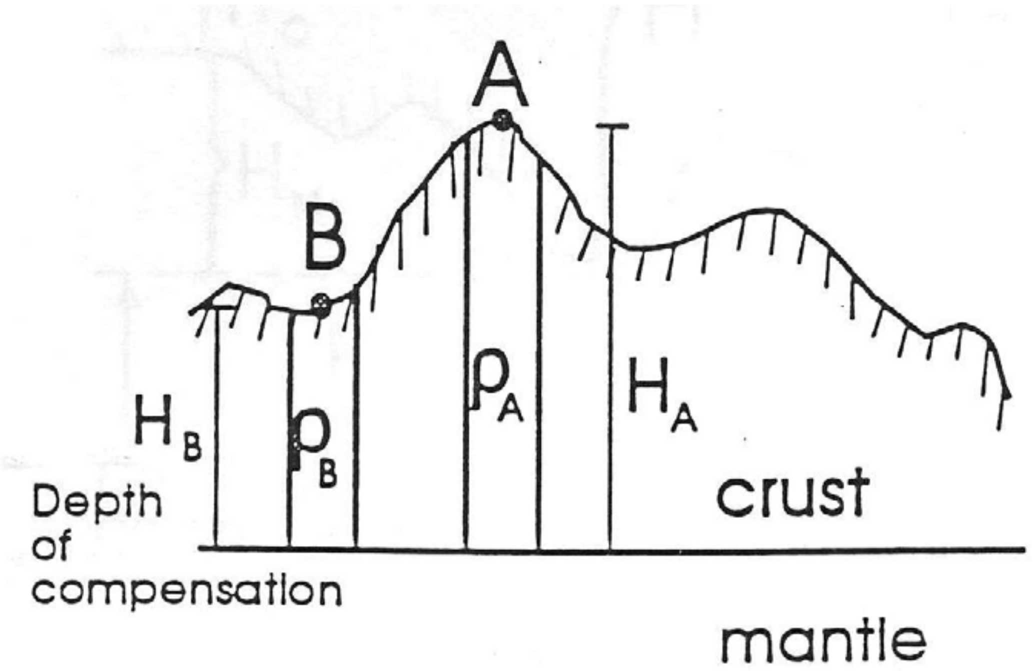
Pratt isostatic compensation model

Pratt suggested that the density of the rocks forming the mountains is less than those forming the lowlands, such that the total mass of a column of mountain to a given depth (called depth of compensation) is equal to the total mass of a column in the lowland to the same depth .

In this model the Earth's crust is approximated as blocks of equal masses floating on the mantle. The elevation-density relationship is given as:

$\rho_A H_A = \rho_B H_B$

Using the diagram to the right as an example



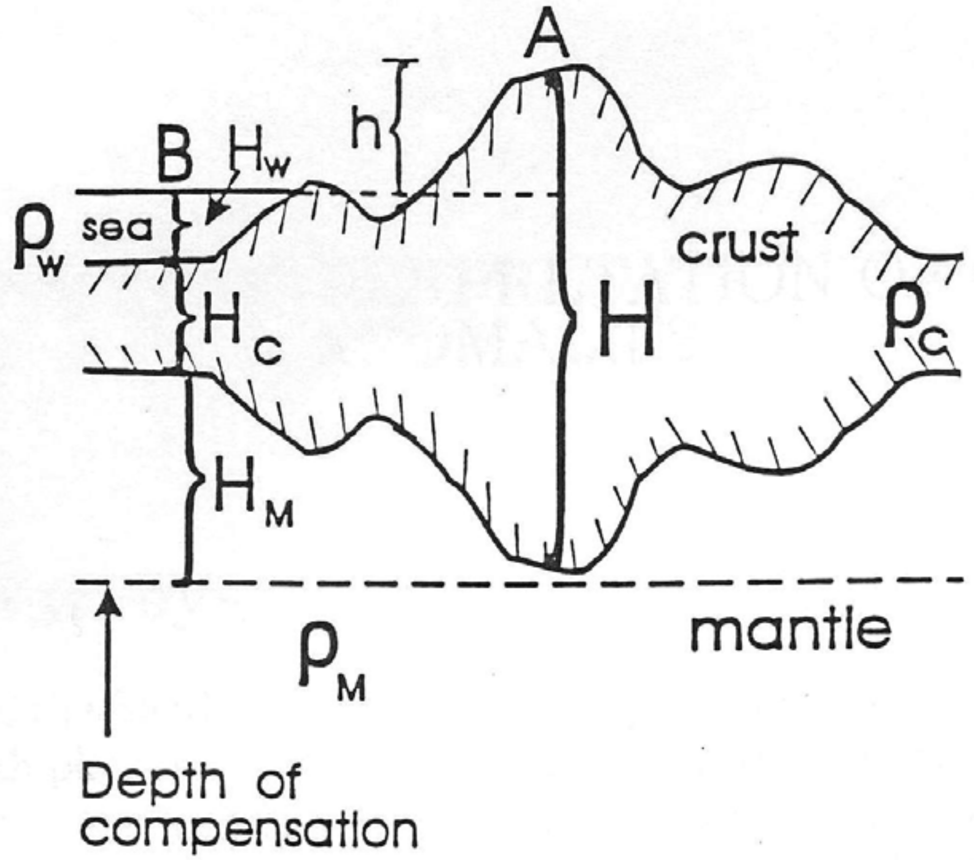
Airy isostatic compensation model

Airy suggested that the density of the Earth's crust does not change strongly enough to give rise to observed Bouguer anomalies, rather the thickness of the crust (roots) changes remarkably. The low density crust is floating in the high density mantle. The higher the mountain, the bigger the crustal root beneath. The elevation-thickness relationship is determined with respect to the sea level:

$$H_W\rho_W + H_C\rho_C + H_M\rho_M = H\rho_C,$$

Where ρ_W, ρ_C, and ρ_M are densities of the water, crust and mantle.

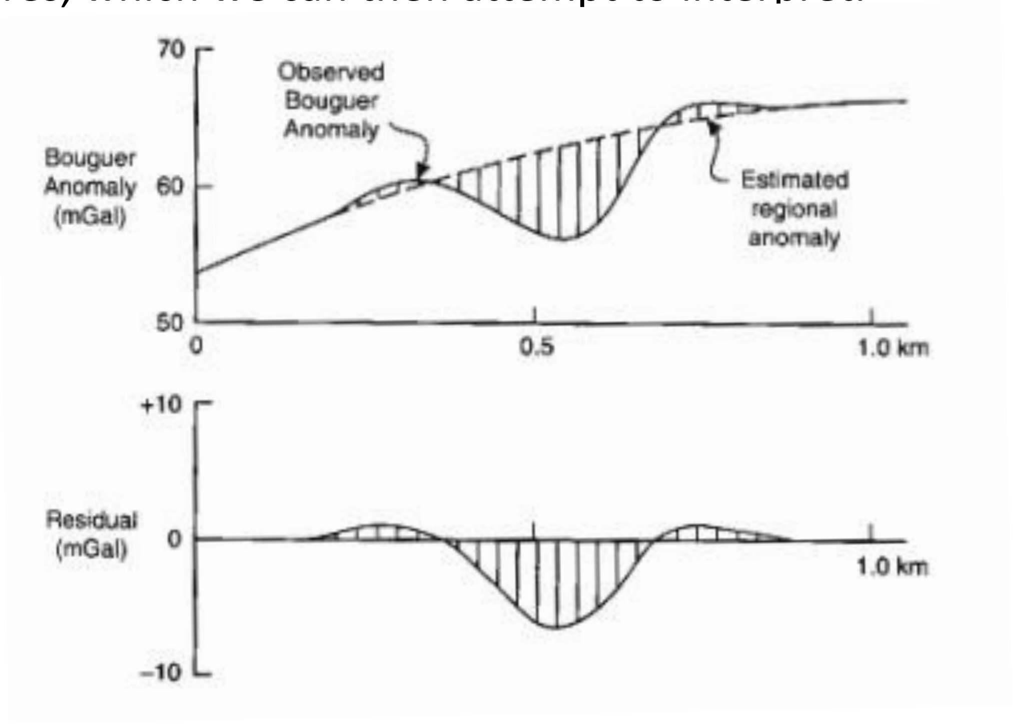
Airy's model is preferred geologically and seismically, whereas Pratt's model is easier to use to calculate the isostatic correction, but the results are similar. The aim of the isostatic correction is that effects of the large-scale changes in density should be removed, thereby isolating the Bouguer anomaly.



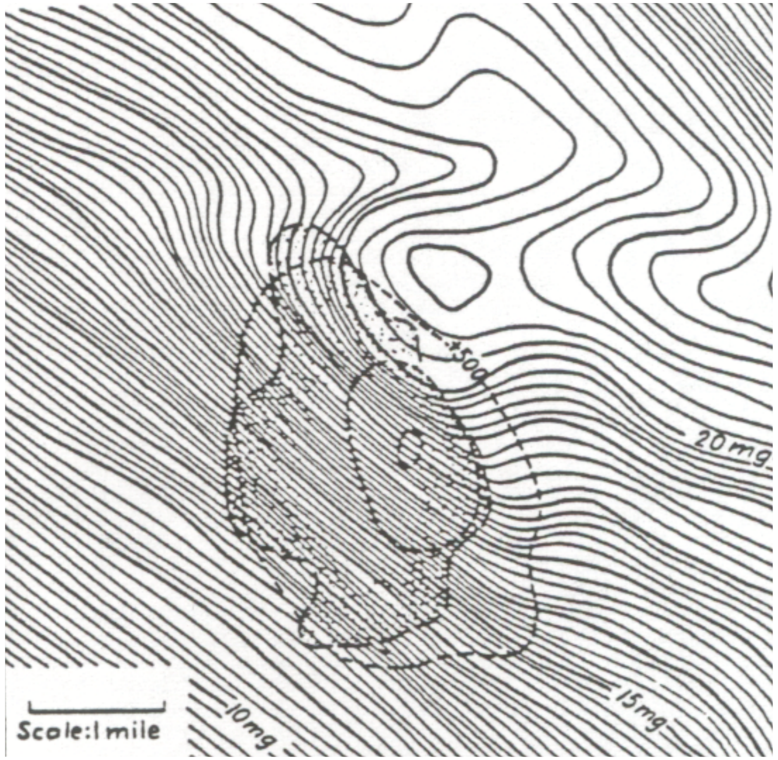
Removing the regional trend

To model local gravity anomalies, it may also be necessary to remove the local gravity trend – assumed to derive from deeper structures such as mantle density changes or Moho topography, from our data. This can be done by fitting a trend surface – either linear (a plane) or a long-wavelength polynomial surface.

This leaves the gravity anomaly due only to shallow, local structures, which we can then attempt to interpret.



Strong regional dip, deflected by oil-filled anticline

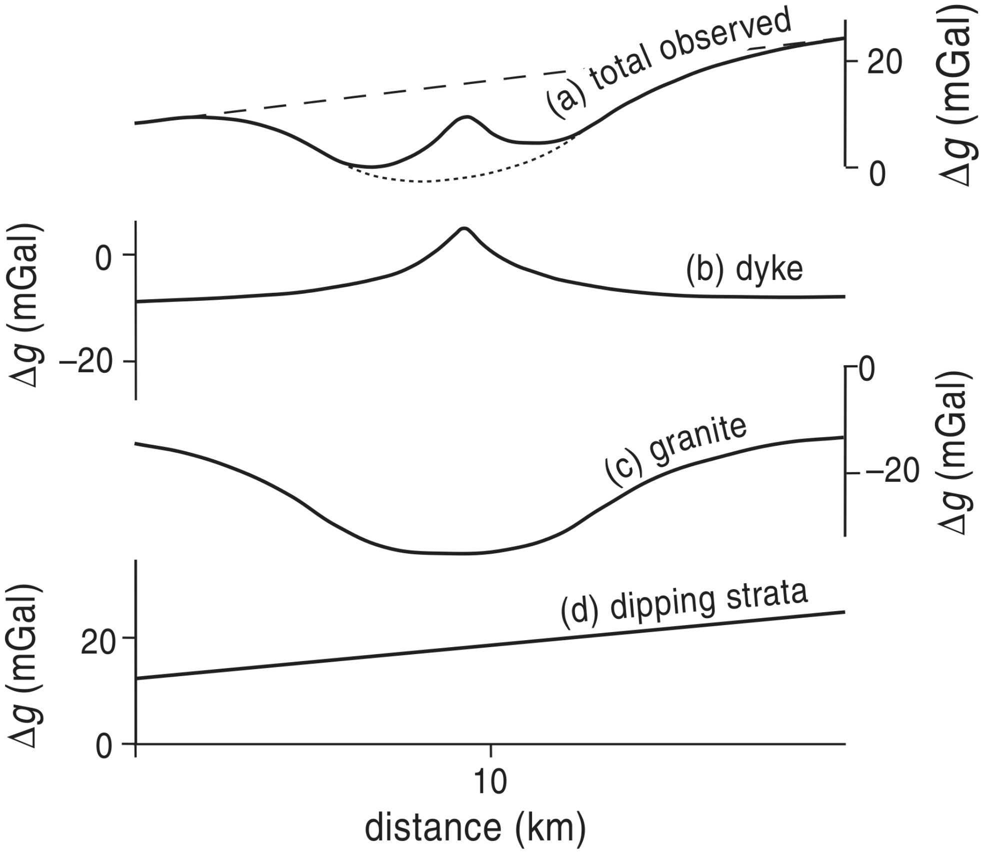


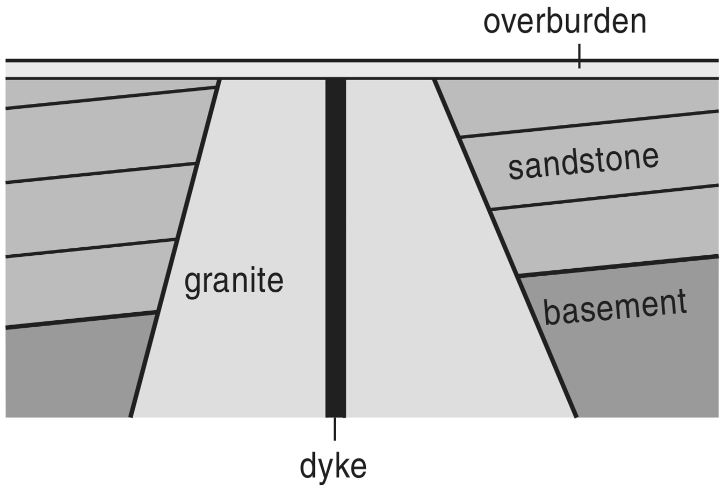
Removing the regional trend

Here is an example of a total gravity anomaly observed in a linear transect (top), along with the contributions of three subsurface features:

- Central basaltic dike
- Larger granitic pluton
- Regional dipping strata into which these have been injected.

By having some a priori understanding of the geology in the area, we can go beyond the visual approimation method for removing a regional trend and support our adjustments to the observed gravity using geologic information.

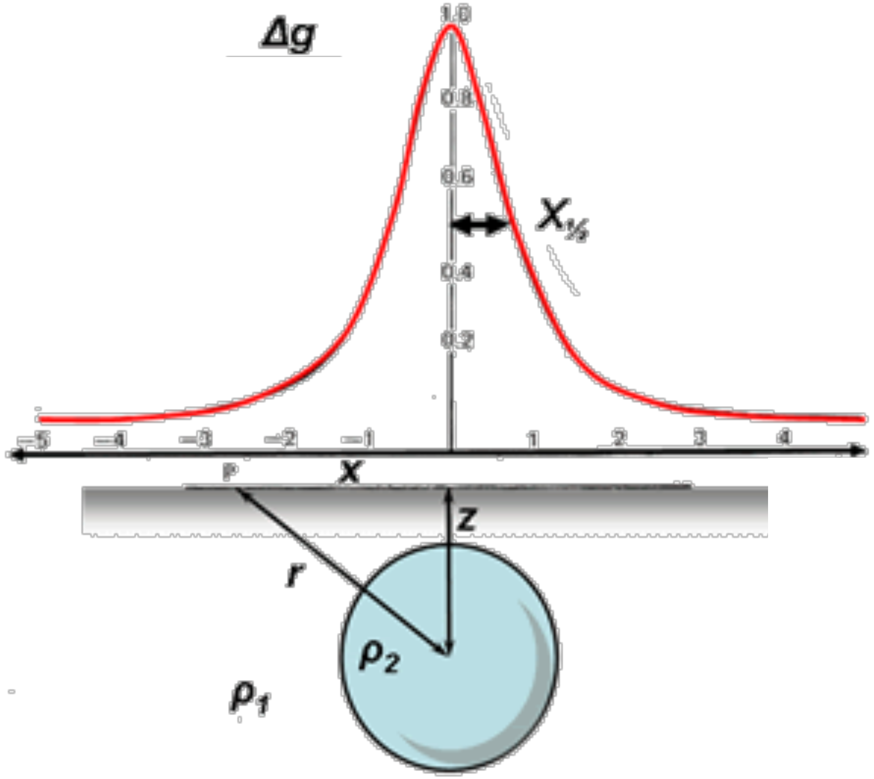


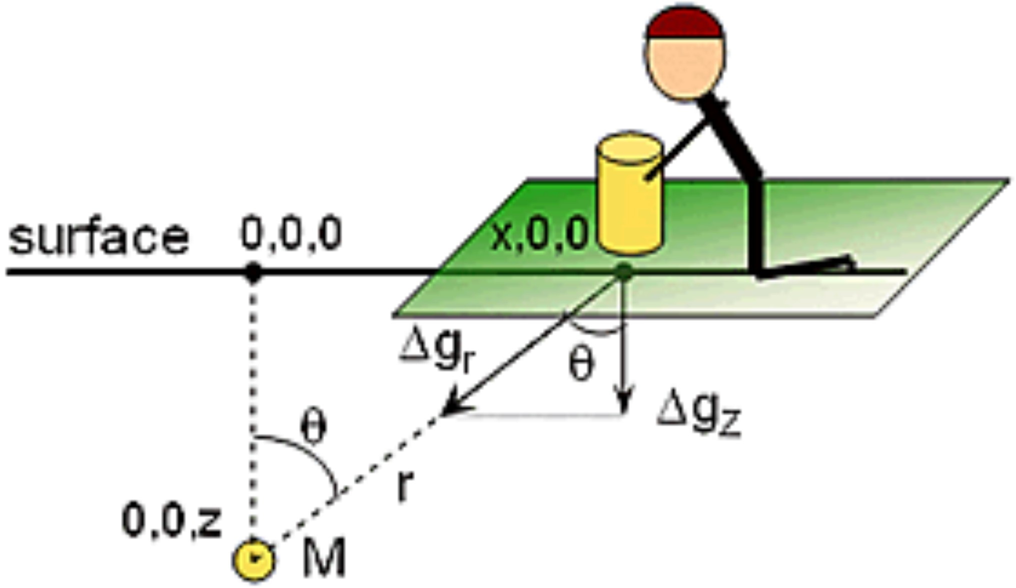


Gravity Effects of Simple Shapes

Modeling Anomalies using simple shapes

We often interpret gravity observations based on well-known signatures for simple geometric shapes. For instance, based on the calculated gravitational effect for a buried point source (below), we can estimate the effect of a buried sphere of a specific depth and specific radius, with a known density contrast to the material in which it is embedded (right), for a profile of gravity measurements that cross directly over the sphere.

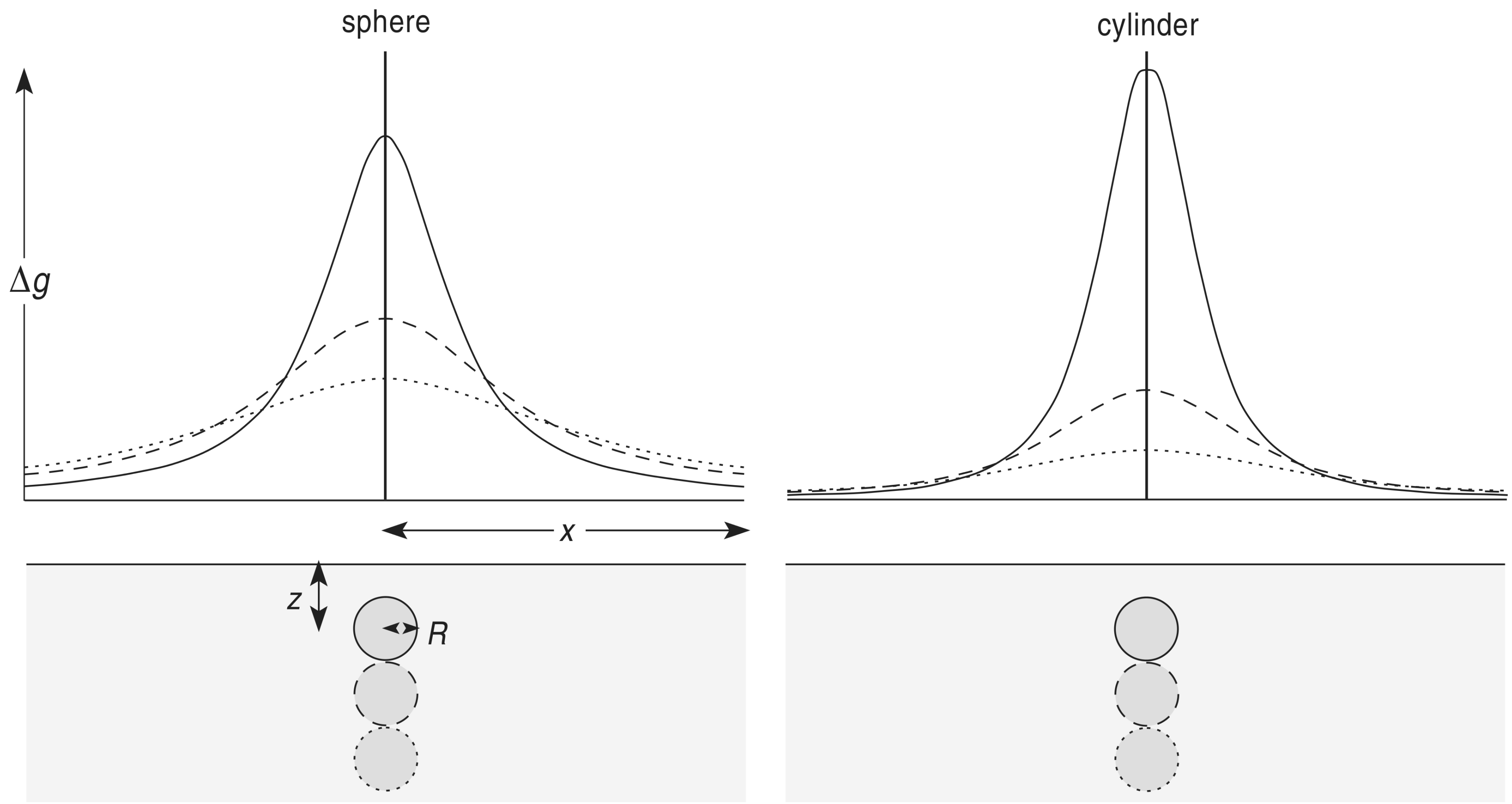




$$\Delta g = \frac{4G\Delta\rho\pi b^3 z}{3\left(x^2 + z^2\right)^{3/2}}$$

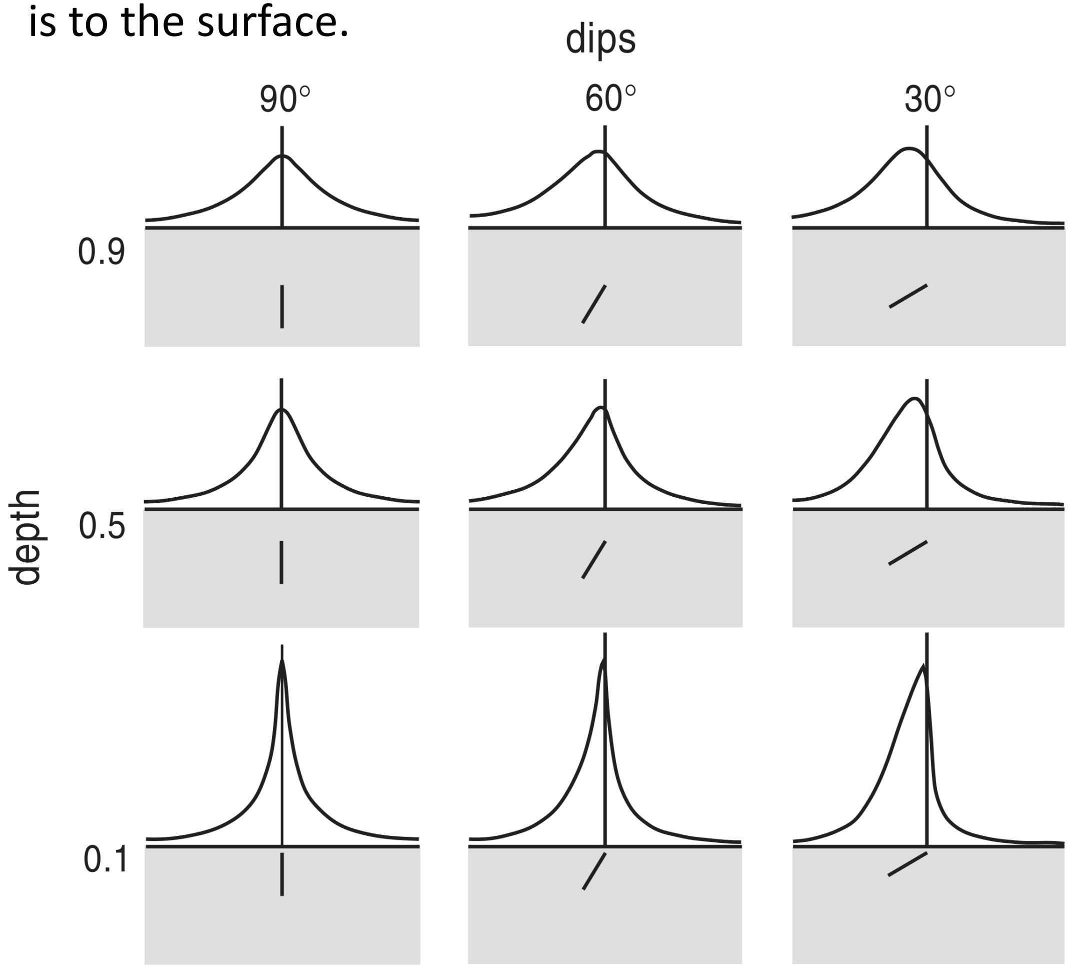
where *b* is the radius of the sphere. The maximum depth of the body is 1.3 times $x_{1/2}$

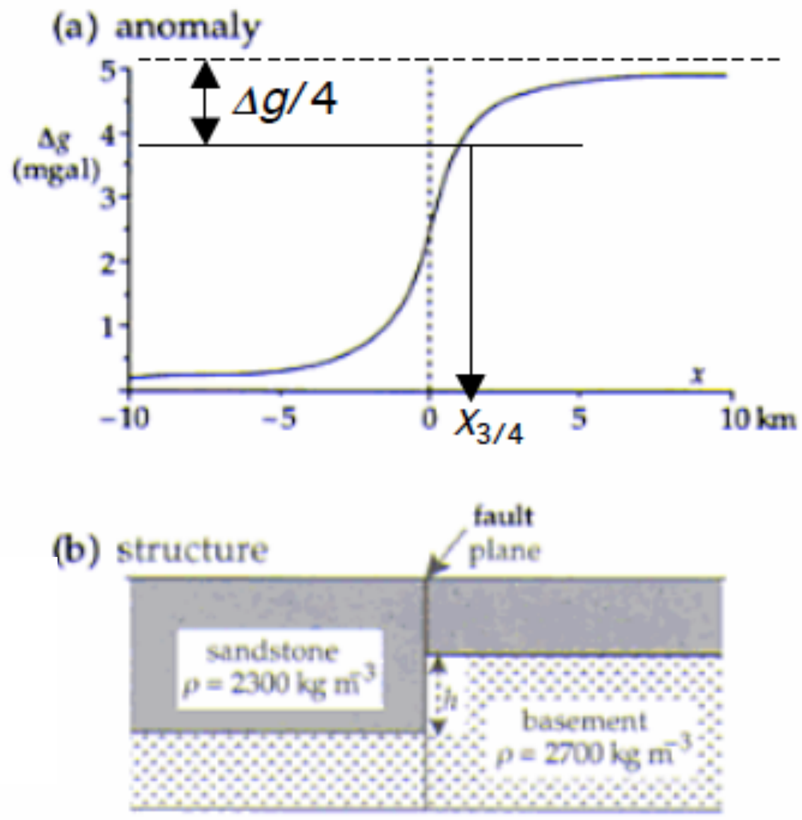
The use of simple shapes to predict anomalies can provide initial insights into observed gravity; however, variations in size, depth and density contrasts for these shapes can produce similar anomalies and can complicate the interpretation of our observations.

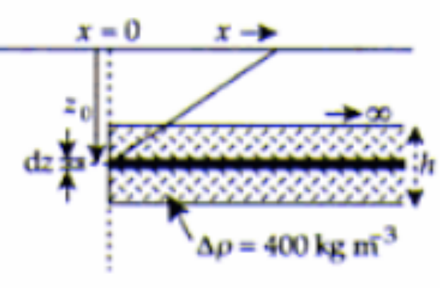


A Dipping Dike

Here we have qualitative anomaly predictions for a variety of depths and attitudes (tilts) of a buried high-density dike. Details of the anomaly curvature and symmetry / asymmetry can provide clues regarding the angle of this feature and how near it is to the surface.



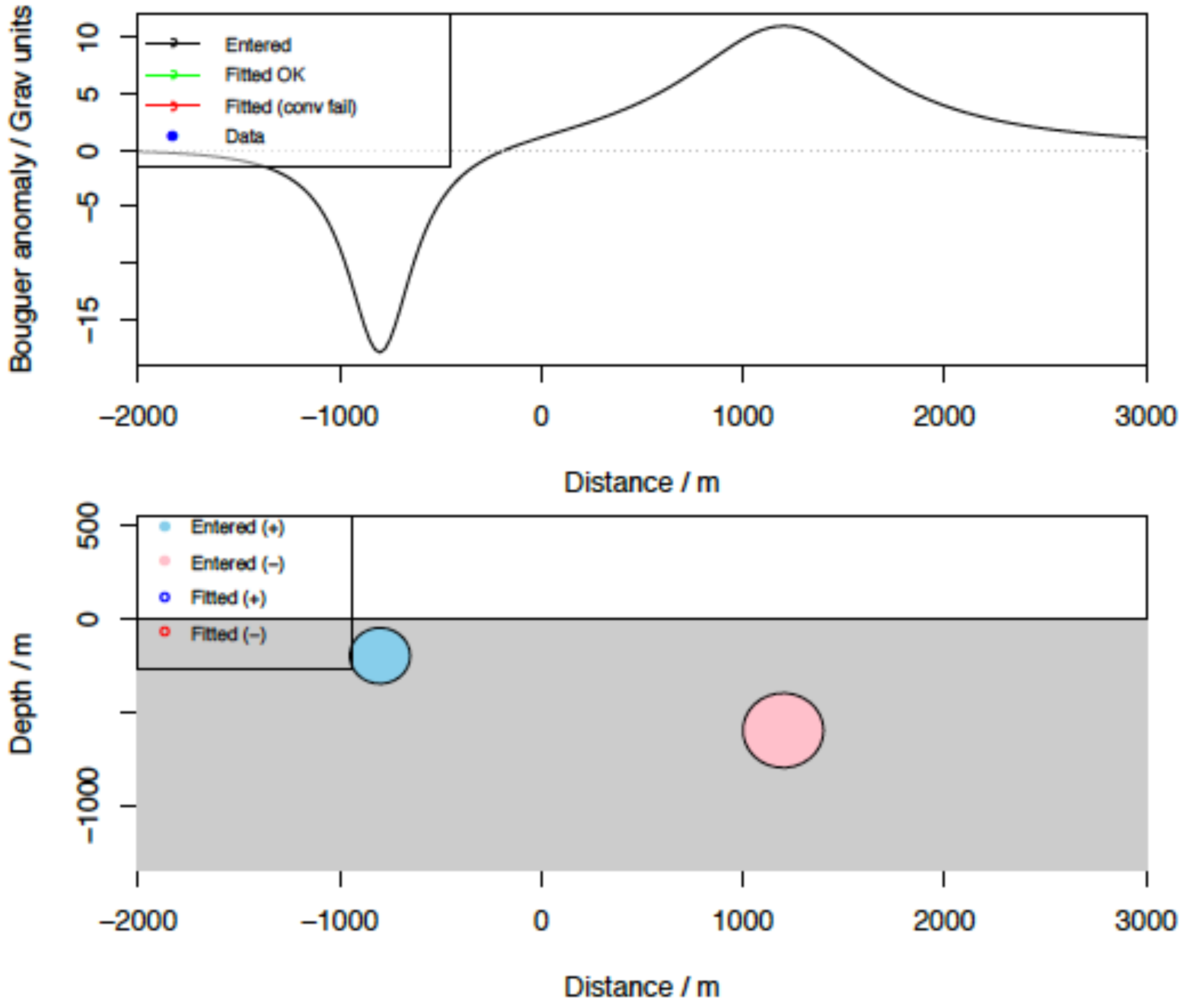




A buried fault with vertical offset can be modeled using an infinite half slab

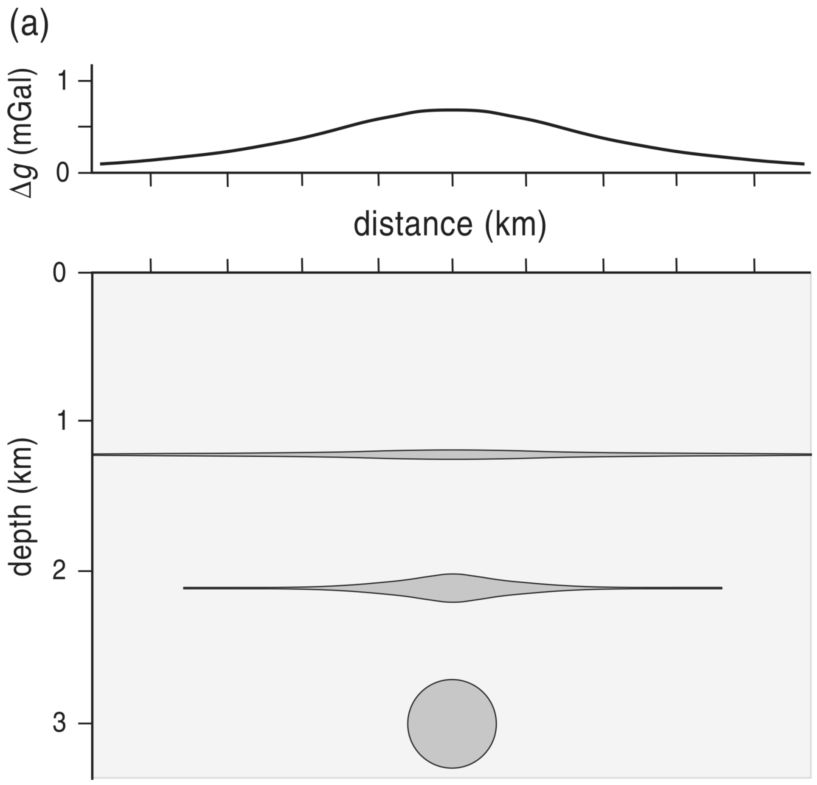
Predicting Gravity by Building Models

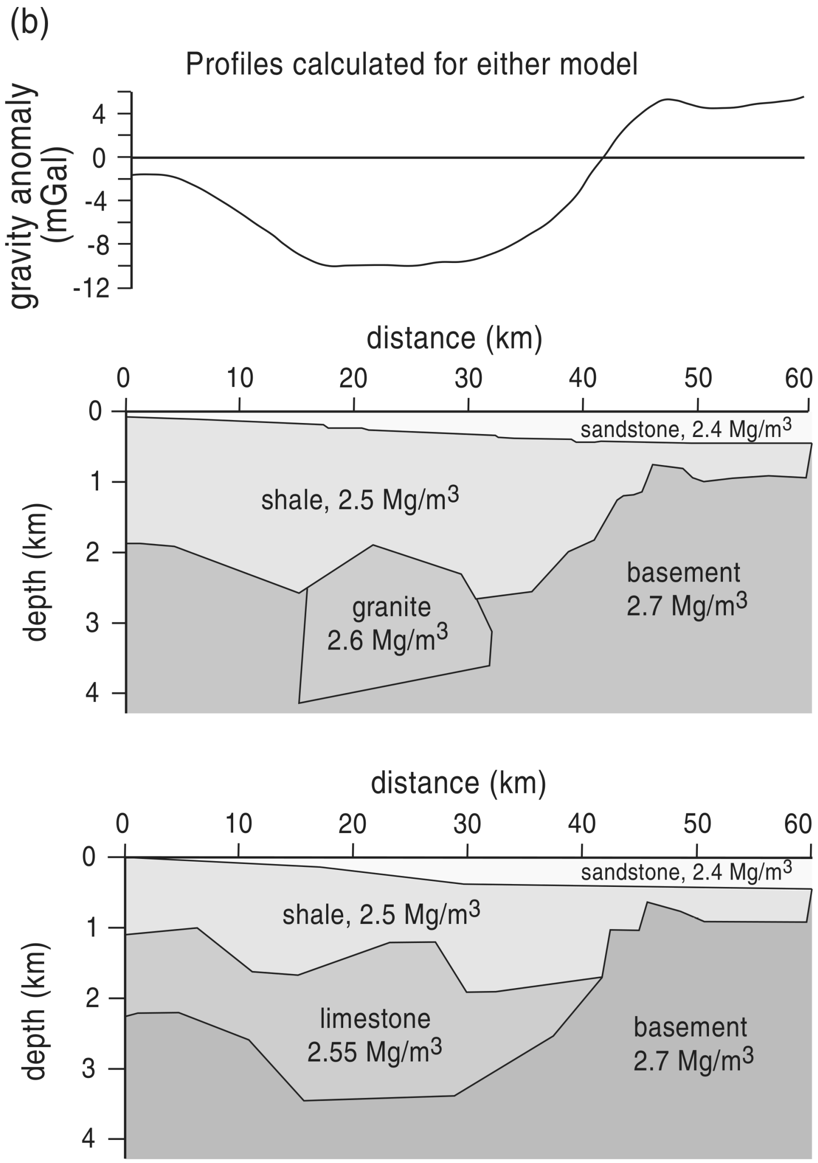
Both negative and positive density contrasts can be modeled for any gravity survey target. The lower panel shows a cross-section through the ground. The circles represent denser (right) and less dense (left) regions. The upper panel shows the gravity that might be measured at the surface.



WARNING!

Gravity is a *Potential Method*, meaning that we try to interpret the sources that contribute to a total *potential force* (this is also true for magnetic surveying). As such, we can always find a variety of physical models that can produce the same observations. This means that no model based solely on gravity observations can be considered to be uniquely correct. Always, additional information is needed before confident interpretation of the gravitational data is possible.



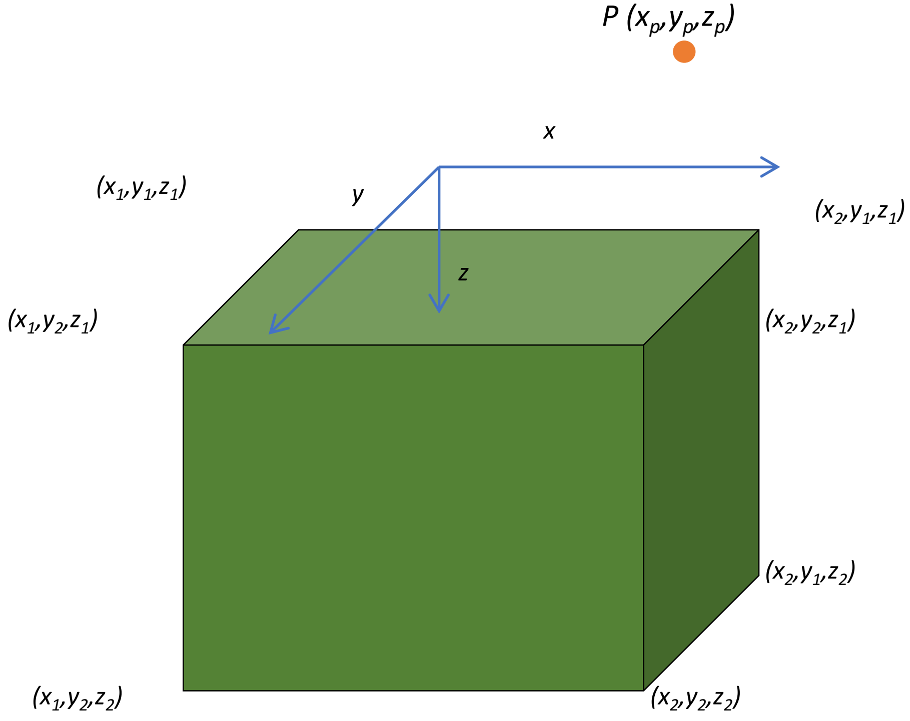


Building a simple density model of prisms to estimate theroretical gravitational attraction for various anomalies

The contribution of each model cell at point P may be given as:

$$g_z(mgal) = G * \Delta\rho * \sum_{i=1}^{2}\sum_{j=1}^{2}\sum_{k=1}^{2}\mu_{ijk} *$$

$$\left[\Delta z_k * \arctan\left(\frac{\Delta x_i \Delta y_j}{\Delta z_k R_{ijk}}\right) - \Delta x_i \log\left(R_{ijk} + \Delta y_j\right) - \Delta y_j \log\left(R_{ijk} + \Delta x_i\right)\right]$$



Where:

g_z vertical component of gravitational attraction

G universal gravity constant = $6.67e^{-11} m^3 Kg^{-1} s^{-2}$

$\Delta\rho$ (kg/m^3) density contrast of prism

$\mu_{ijk} = (-1)^i(-1)^j(-1)^k$

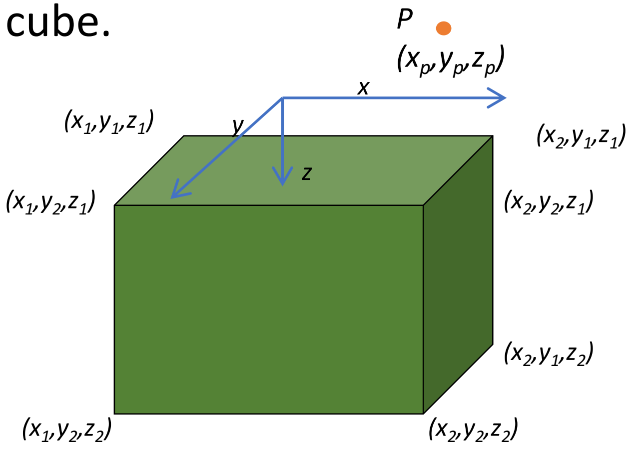
$\Delta x_i = (x_i - x_p)$, $\Delta y_j = (y_j - y_p)$, $\Delta z_k = (z_k - z_p)$

R_{ijk} (distances from each corner to point p) =

$sqrt(\Delta x_i^2 + \Delta y_j^2 + \Delta z_k^2)$

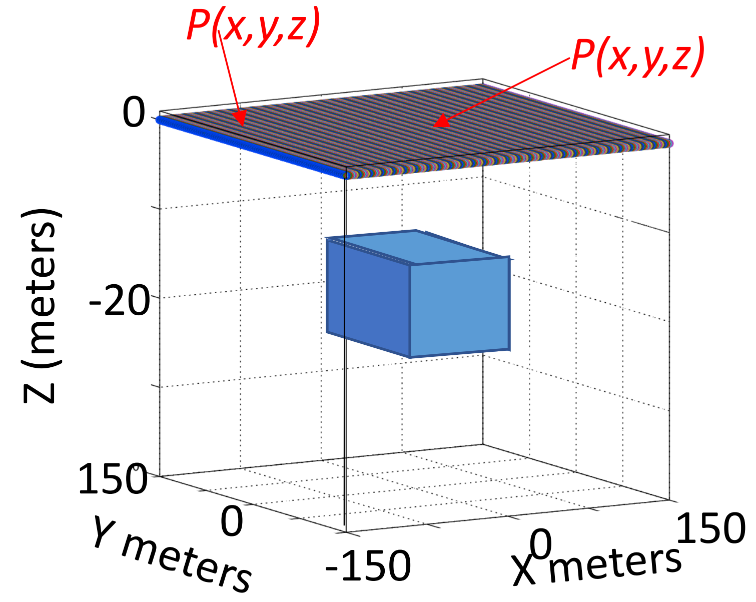
Building a Computer Model to Predict Gravity

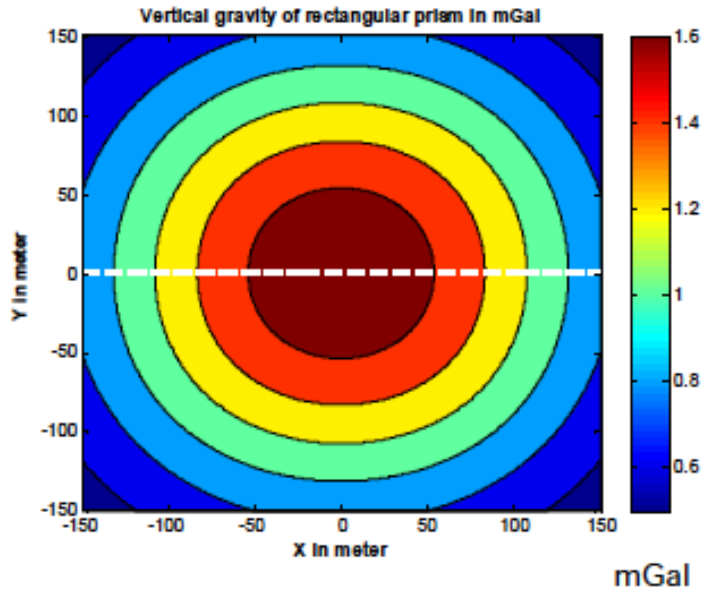
Here we take the imaginary cube and position it at some depth beneath the surface of our model. For each X,Y position of our sensor (grid of points at the top of the model) we calculate the gravitational pull of the cube.



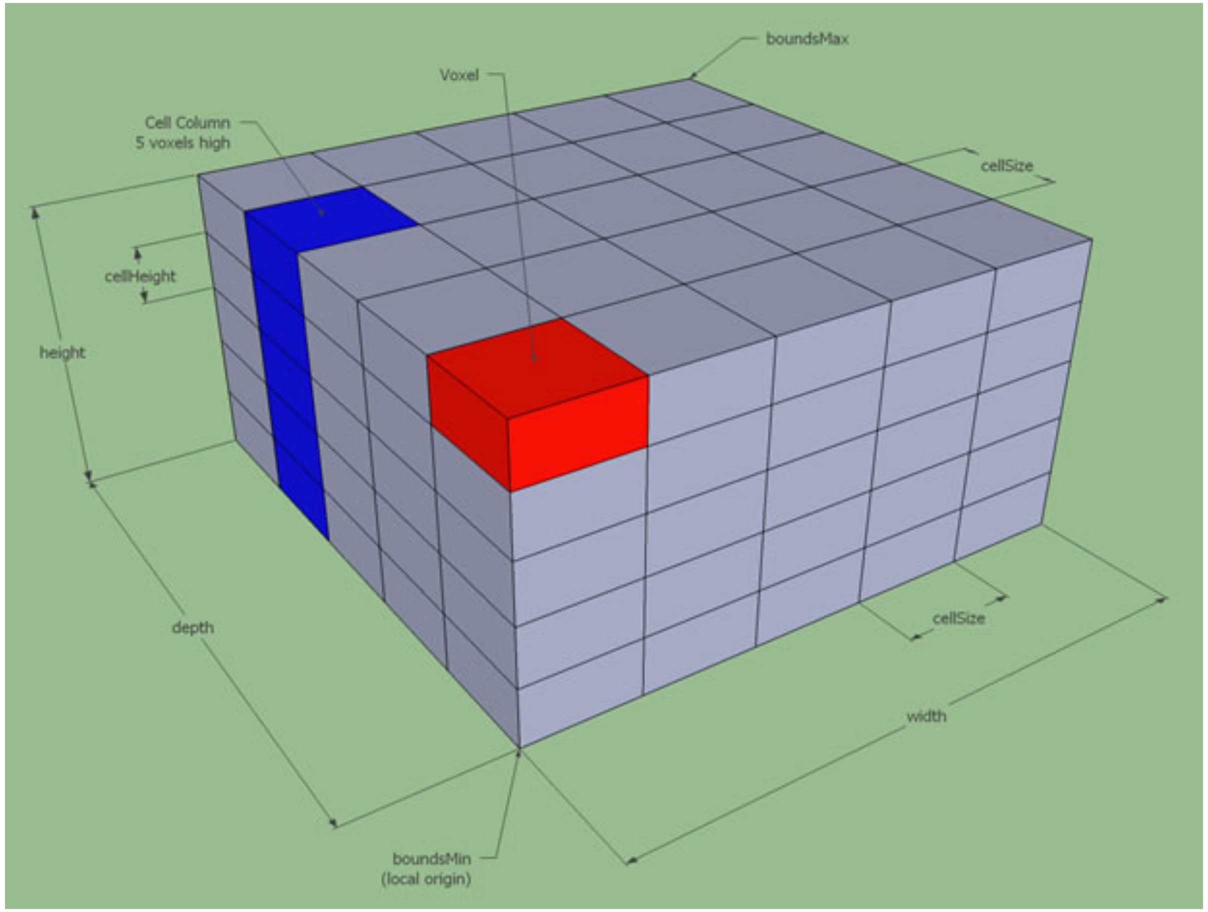
For example using the following parameters:
X_1 = -100m, x_2= 100m, Y_1= -100m, Y_2= 100m, Z_1= -100m, Z_2 = -200 m, $\delta\rho$ = 2000 kg/m^3

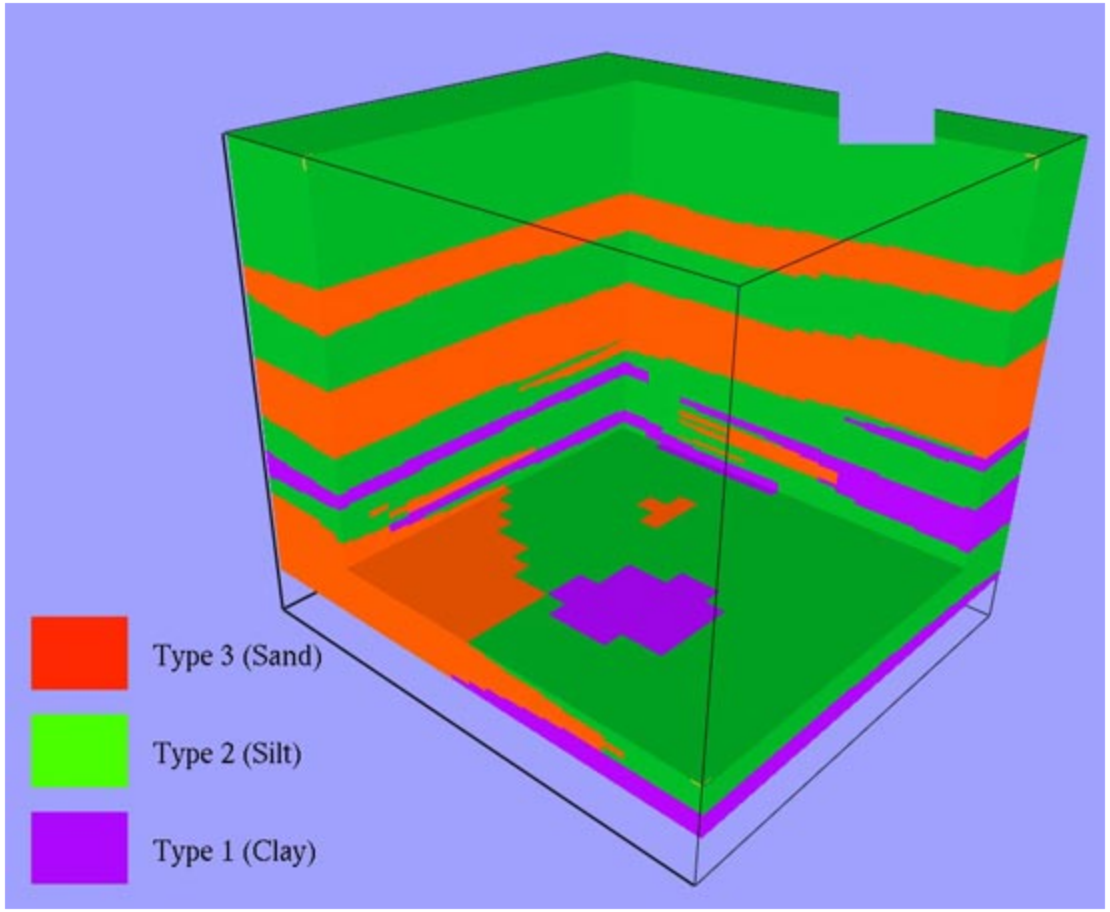
for one buried prism, we can predict an anomaly as shown here, for a dense rectangular network of readings taken over a grid of 150 m^2.





We construct our model of prisms, assigning different densities to different prisms (cells / voxels) to approximate any subsurface density structure we like. These models can become very complex, depending on the local geology and the resolution of our survey.

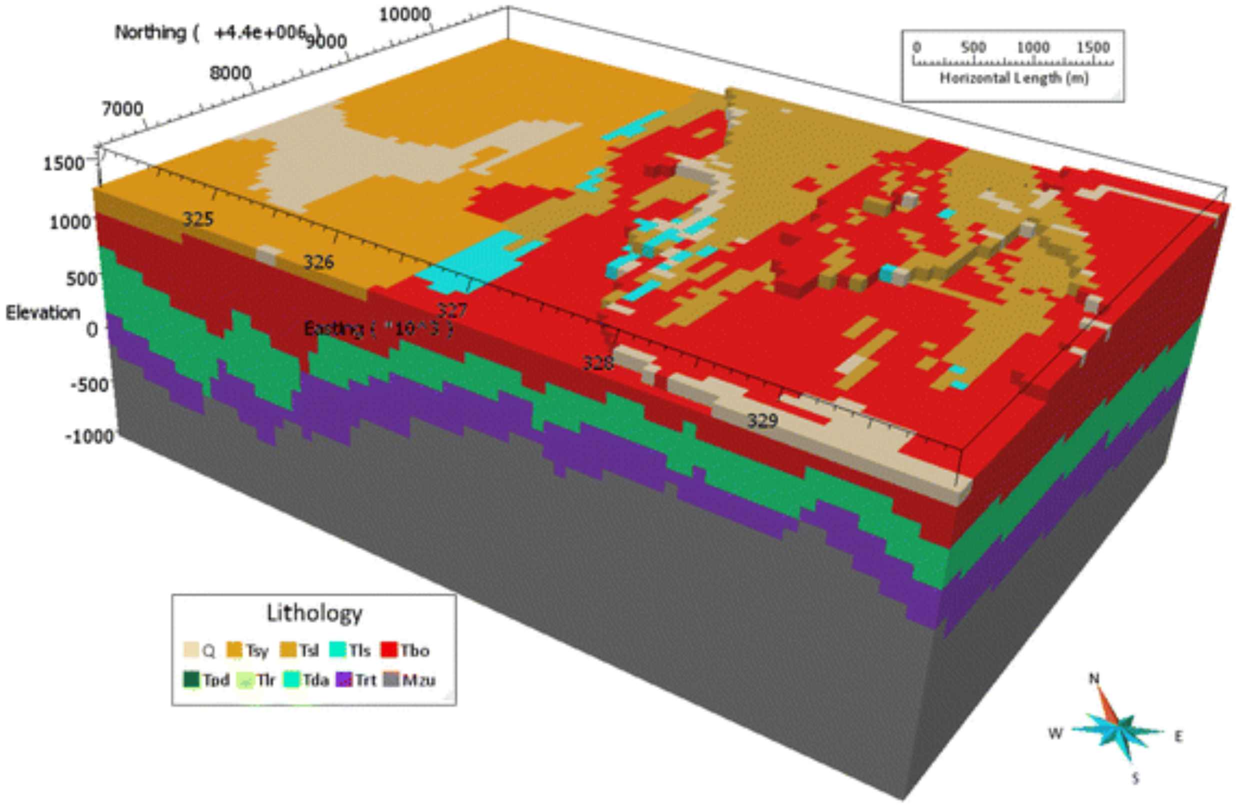


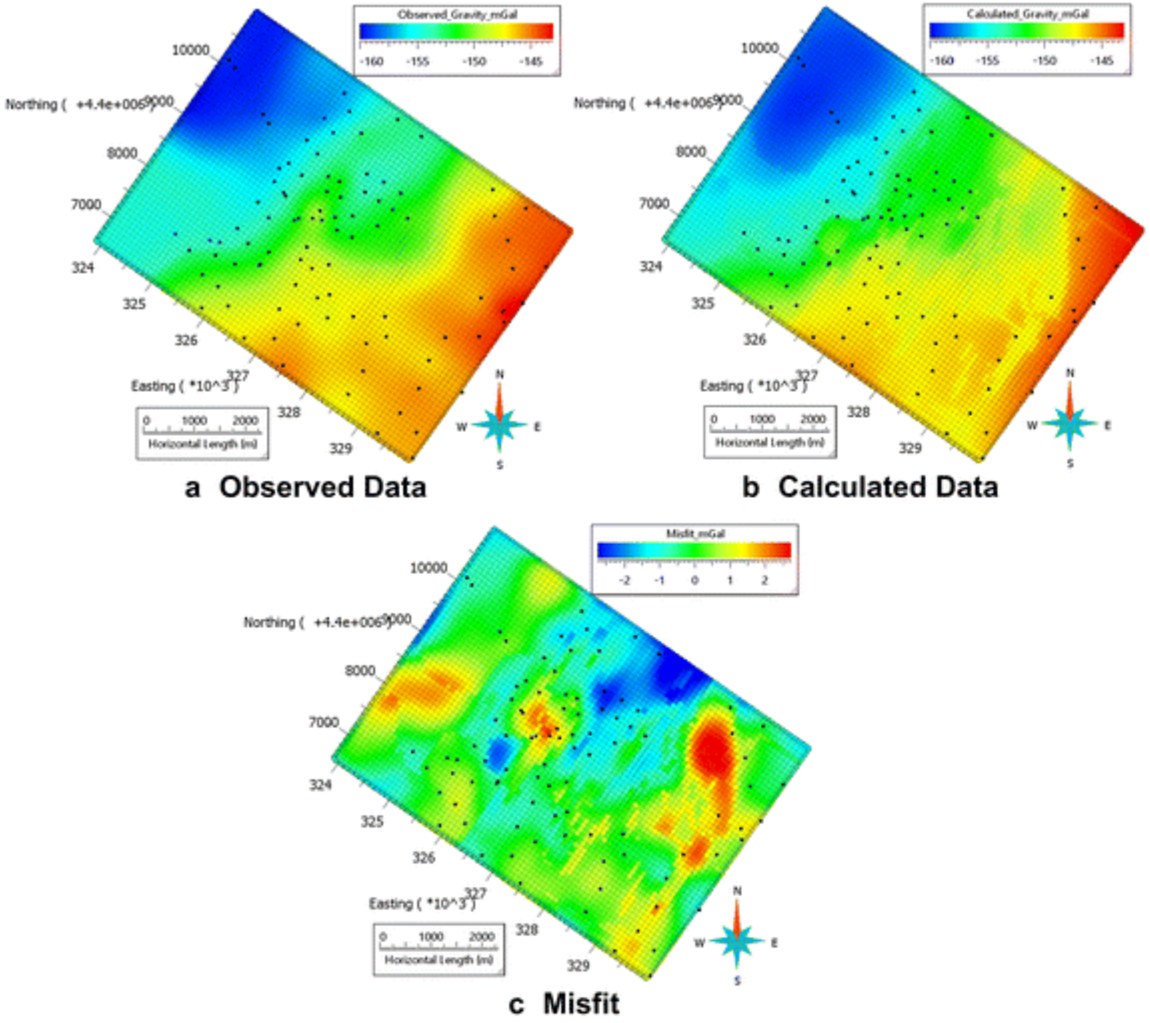


Using Models to Interpret Reality

Comparing the Forward Model to the Observations

Here we show a very complex geologic model for the Bradys Geothermal Field in western Nevada, U.S.A. (on the left) consisting of many prisms of different densities. The calculated gravity field (panel **b** on the right) can be compared to the actual, measured gravity field (panel **a** on the right) to determine the misfit (panel c on the right). The 3D model can be adjusted to improve the fit between observed and calculated gravity values.

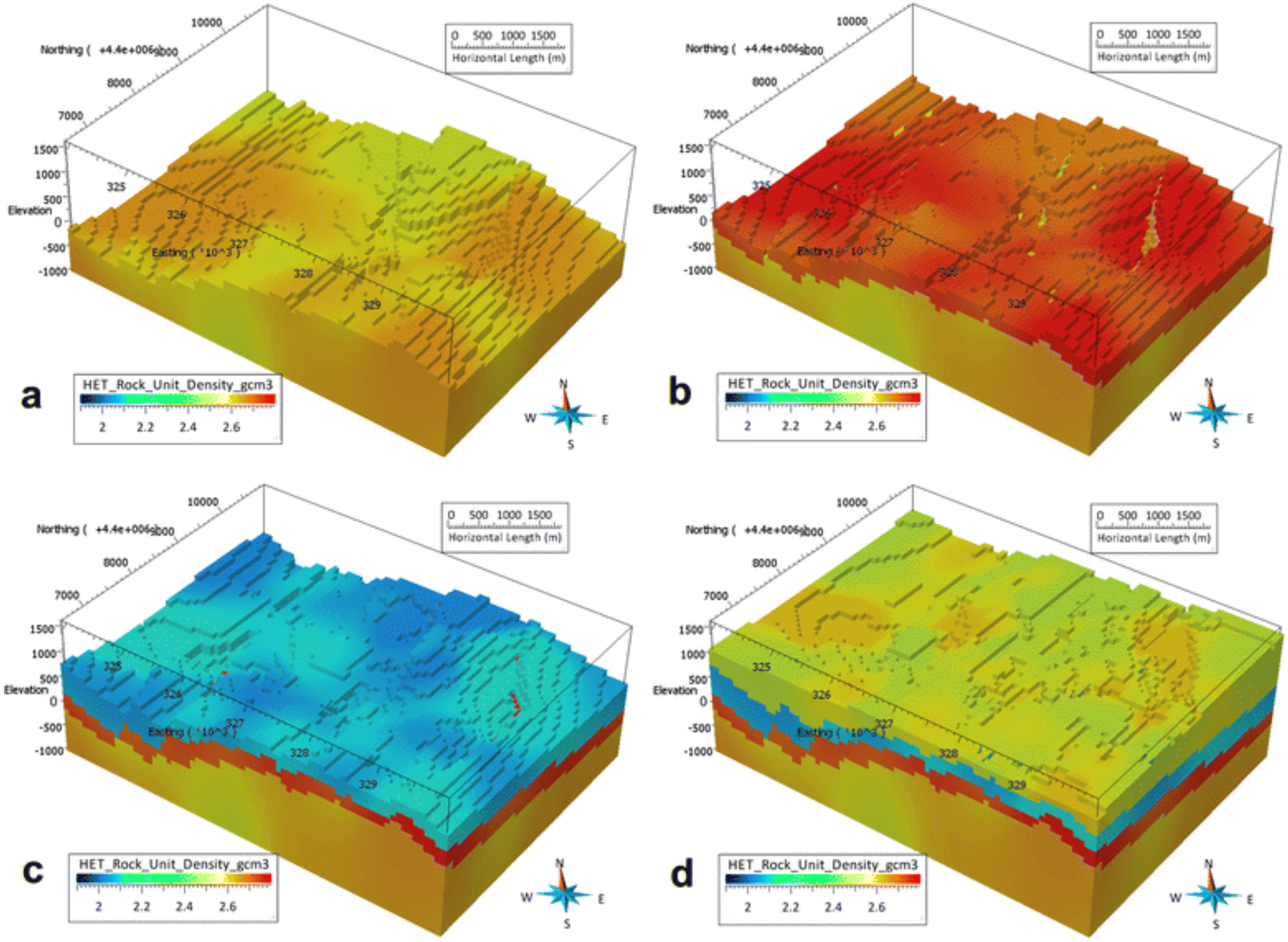




Witter et al. (2016), Geothermal Energy 4:14

Varying the Rock Density

If the geologic structure is well-known through seismic imaging or borehole logging, then the geometry of the rock units cannot simply be adjusted to reduce the gravity residuals observed from the modeling. But the densities of the different rock units may be adjusted, within reasonable bounds, and important information about the thermal, porosity, permeability or saturation of the target area may be obtained.



Summary:

- We can measure changes in gravity from place to place on the earth.
- These measurements require careful recording of location, elevation and time for each reading.
- These readings must be adjusted for known effects (such as elevation, latitude, tides) that can bias our data and mask the signal of interest.
- After making corrections to our data, we can remove regional trends to obtain local Bouguer anomalies.
- The Bouguer anomalies arise from variations in the subsurface density structure.
- We can build models to explain our observations, but these models must be consistent with what is known about the local geology.
- Combining gravity models with other information – geologic, seismic, electromagnetic, will improve confidence in the results.

Gravity is a potential method, meaning that its results are ambiguous in isolation. Other information is always needed to interpret gravity anomalies with confidence.