Gravity 5

Instruments and gravity processing

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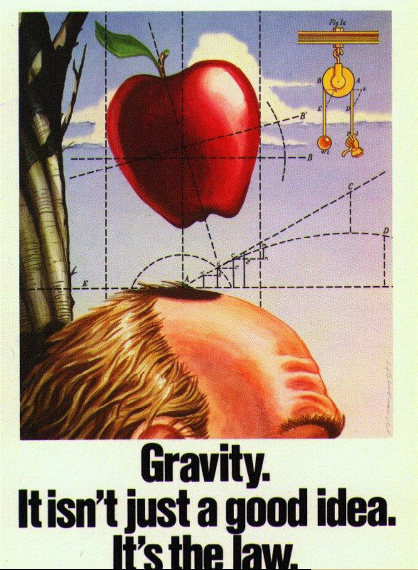
Potential Fields Geophysics: Week 5

Objectives for Week 5



- Learn about gravity instruments
- Learn about processing of gravity data
- Make the corrections to calculate a simple Bouguer anomaly

The pendulum

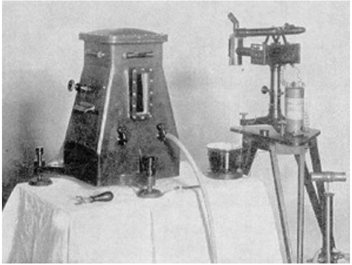
The first gravity data collected in the US were obtained by G. Putman working for the Coast and Geodetic survey, around 1890. These gravity data comprised a set of 26 measurements made along a roughly E-W transect across the entire continental US. The survey took about 6 months to complete and was designed primarily to investigate isostatic compensation across the continent.

Putnam used a pendulum gravity meter, based on the relation between gravity and the period of a pendulum:

$$g = \frac{4\pi^2 l}{T^2}$$

where:
l is the length of the pendulum
T is the pendulum period

Seems easy enough to obtain an absolute gravity reading, but in practice pendulum gravity meters are problematic. The length of the pendulum can change with temperature, the pendulum stand tends to sway, air density effects the measurements, etc. The Mendenhall pendulum (around 1910) handles some of these problems, for instance by using a vacuum, but basically one cannot see subtle variation in gravity with a pendulum apparatus!



Example

Assuming the period of a pendulum is known to be 1 s exactly, how well must the length of the pendulum be known to measure gravity to 10 mGal precision?

$$l = \frac{T^2 \Delta g}{4\pi^2} = 2.5\mu$$

The zero-length spring gravimeter

The first truly portable gravimeter was invented by Lucien LaCoste in the mid-thirties, when he was in his twenties.This gravimeter makes relative measurements rather than absolute measurements. It does not provide any information about the absolute acceleration due to gravity, only the relative change in gravity from place to place. These types of meters are still the work-horses of the gravity world. Most gravity measurements you will use are made using this type of instrument, now manufactured by Burris.

This gravimeter contains a mass attached to a cantilevered beam and suspended with a metal or quartz spring. The tension on the spring can be adjusted to bring the beam to a null position. The force required to move the beam to the null position is proportional to the change in gravity. The apparatus is called a zero-length spring meter because the spring is pre-stressed: if the mass were removed altogether, the spring would contract to “zero” length.

To make accurate measurements, the instrument must be level (aligned with the vector of the Earth’s gravity field), in a place quiet enough to avoid vibrations (trucks rumbling by and earthquakes are a problem!), and given sufficient time to be in thermal and mechanical equilibrium (avoiding sharp changes in temperature that effect the instrument’s metal parts, etc.).

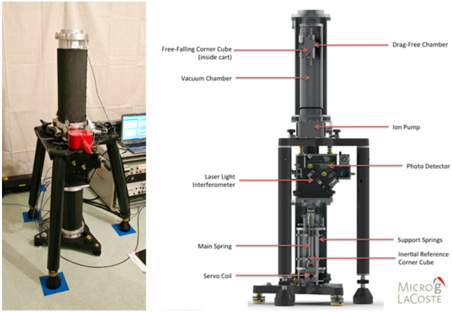
The main features of the gravimeter design have not changed from LaCoste’s early concept. Improvements have been made to electronically determine the null position of the meter, using a servo motor interfaced to a PDA. A number of features of the Burris gravity meter are designed to minimize instrument drift, including a heater to maintain constant temperature, a sealed case to minimize the speed of pressure change inside the gravimeter, and electronic monitoring of the instrument level. In ideal conditions, gravity measurements with this instrument are good to 1–10 μGal.

Absolute gravimeter

An absolute gravimeter, like a pendulum apparatus, measures gravitational acceleration directly, rather then the relative change in gravity from place to place. These instruments are "sort of" portable (weighing around 50 kg), can make accurate determinations of the gravity field within one hour, and are increasingly field-worthy, although still quite expensive (>100K USD).

The current generation of absolute gravity meters are free-fall devices, simply measuring the time it takes an object to fall a given distance in a vacuum. Improvements in the instrument have involved improving the timing of the fall (using an atomic clock) and the distance of the fall (using a high-precision laser), the quality of the vacuum, a dampening system to lessen the effect of vibrations, and the ease with which free-fall measurements are repeated.

The absolute precision on the best absolute gravimeters is about 1 μGal.



Gravity from space

Space-based gravity observations generally rely on accurately measuring a satellite's position and comparing it to its theoretical position given a model of the Earth's (or other planet's) gravity field, assuming no anomalous mass distribution. In early efforts, this involved tracking a single satellite in its orbit and the departure of its true position from an expected position.

Alternatively, the exact altitude of the satellite could be compared to its expected altitude, by measuring altitude using a satellite-mounted laser altimeter. Variations in altitude measured by laser are a particularly good way to track variation in gravity at sea. Because the altimeter is measuring the ocean height variation, this method essentially maps gravity variation on the ocean surface rather than at the height of the satellite, so relatively small anomalies can be detected.

Modern gravity missions, such as GRACE and GRAIL, map variation in the gravity field by placing two satellites in orbit along the same orbital track. Gravity anomalies cause the distance between the two satellites to deviate from their expected distance of separation. The size of the gravity anomaly (or the excess mass that will be detected) depends on the altitude of the satellites and the distance between them.

Gravity corrections

Putman had a big job on his hands after he finished gathering his gravity data. He knew that the gravity varied for many reasons other than lateral variations in mass in the subsurface. We have already seen that gravity varies with latitude due to the shape and rotation of the Earth, the tides, and with variations in elevation ($\frac{dg}{dR}$).

Gravity corrections involve reducing the effects of these features of Earth/planetary gravity in order to isolate the effects of lateral variations in mass. These corrections are not a trivial exercise. Some of the corrections are mathematically complex (e.g., the effect of latitude). By making the corrections in a step-wise fashion, removing one effect at a time systematically, we often can learn more about the interior of the Earth than we could if we simply make the corrections all at once.

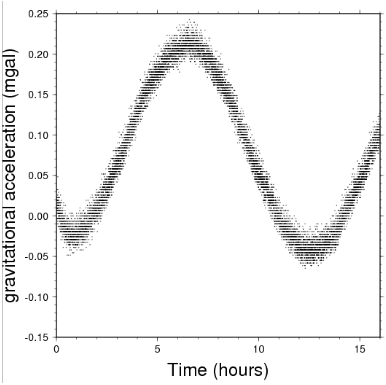
The usual corrections made to gravity data include:

- Instrument drift, associated with variations in gravity only due to the fact that the gravimeter registers different readings with time, due to mechanical, thermal, and electrical changes in the instrument
- Tidal corrections, to account for the time varying gravitational acceleration due to the motion of the Sun and Moon
- Theoretical gravity correction, to account for the shape and rotation of the Earth
- Free air correction, to account for variations in gravitational acceleration with elevation
- Atmospheric correction, to account for varying density of the atmosphere with elevation
- Simple Bouguer correction, to account for the average density of rocks as a function of elevation (sometimes called the Bullard A correction)
- Spherical cap correction, to account for the change in the Bouguer correction due to the roughly spherical shape of the Earth (also known as the Bullard B correction)
- Terrain correction, to account for the exact form of the terrain and its influence on density distribution around the gravity station (also known as the Bullard C correction)
- Isostatic correction, to account for broad (long wavelength) variations in gravity due to isostatic compensation of the crust.

Instrument drift and the tides

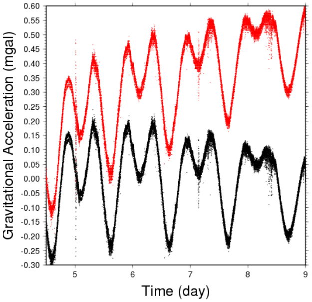
Gravity varies over time at a specific location for several reasons:

1. The instrument reads out different values of gravity over time because of changes in the instrument itself. For example, changes in temperature might cause thermal expansion of the beam, changing the tension on the "zero-length" spring. This type of variation with time is referred to as instrument drift.
2. Gravity varies due to tidal effects. For any location at the surface of the Earth, the distance to the Sun and Moon change continually with time, changing the gravity field.
3. On short time scales, gravity varies due to ground vibrations, microseismic events, or teleseisms (ground motion due to distant earthquakes). The instrument may also go out of level during measurements due to settling on soft ground. These types of effects are usually referred to as noise and are minimized by adapting field procedures.
4. Gravity varies with time for geological reasons. The elevation of the gravimeter might change due to subsidence or inflation of the site. Mass distribution might change with time, through: formation of fractures, intrusion of magma, or dissolution of minerals.



A drift curve measured with a LaCoste "zero-length" spring instrument in the USF lab. The drift is dominated by tidal effects. Measurements are at 1 s intervals. Note the random measurement error due to ground motion in the lab is about 0.05 mGal.

Instrument drift and the tides



A continuous record of gravitational acceleration in the USF lab, gathered at 1 s intervals for several days. The actual drift curve is shown in red, the detrended drift curve s shown in black.The drift curve (red) can be thought of as consisting of two components. The Earth tide varies predominantly on approximately 12 and 24 hr cycles. The tide develops longer period variation (a beat) due to the elliptical motion of the Sun and Moon.

The second component is nearly linear instrument drift, caused by mechanical and/or thermal changes in the instrument. The linear drift is approximately 0.081 mGal per day.

Example

Calculate the tidal acceleration, a_T due to the moon at the Earth's equator at lunar noon (the moon is directly overhead). From Newton's law (Module 1):

$$a_T = GM_m \left(\frac{1}{(r_L - R_E)^2} - \frac{1}{r_L^2} \right)$$

where
M_m is the mass of the moon (7.3547×10^{22} kg)
r_L is the distance from the center of mass of the Earth to the moon (3.84×10^8 m)
R_E is the radius of the Earth at the equator (6.378137×10^6 m)
G is the gravitational constant.

With these values, the tidal acceleration is approximately 0.11 mGal due to the Moon (not including the gravitational effect of the Sun).

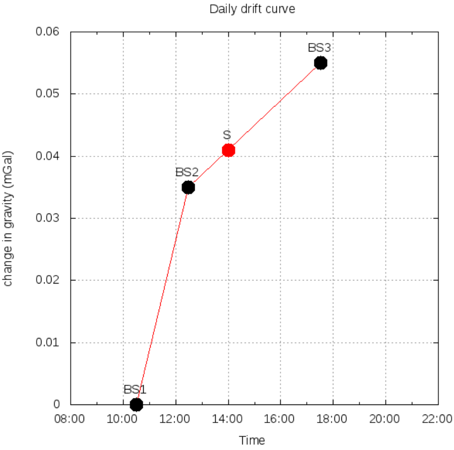
Instrument drift during surveys

During surveys it is not possible to gather data with a meter and continually monitor drift with the same meter. To monitor drift during surveys, repeated measurements are made by returning to a base station ($BS1$, $BS2$, $BS3$) and change in gravity due to drift is recorded. What is the drift correction for station S, measured at time t_S =14:00? The time of this measurement falls between base station readings $BS2$ and $BS3$, so a linear drift correction is:

$$\Delta g_S = m(t_S - t_{BS2}) + \Delta g_{BS2}$$

where:
m is the local slope of the drift, calculated only from successive base station measurements $BS2$ and $BS3$
Δg_{BS2} is the drift since the beginning of the daily measurements measured at time t_{BS2}

If t_{BS2} =12:29, $\Delta g_{BS2} = 0.035$ mGal and t_{BS3} =17:32, $\Delta g_{BS3} = 0.055$ mGal, at t_S =14:00, $\Delta g_S = 0.041$ mGal. This drift correction is subtracted from the observed gravity to obtain the drift corrected gravity reading. That is, instrument drift causes the gravity value to be "too high" at 14:00 at station S and the drift correction accounts for this.

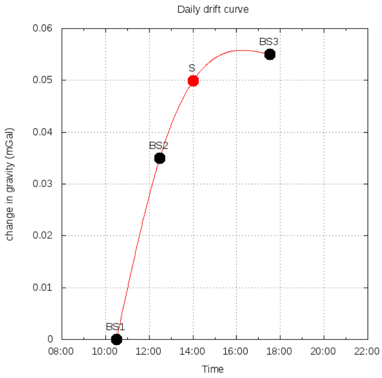


Instrument drift during surveys

There is no *a priori* reason to think that drift between successive base station readings is linear. An alternative model is that the instrument drift varies as a smooth curve in time, and base station readings sample this smooth variation.

The plot at right shows the drift measured at the base station measured three times during the day. These are the same data as shown on the previous slide, where a linear drift model was used to estimate drift at time t_S for drift at station S. Now the drift model is nonlinear, calculated using a cubic spline. The spline passes through each base station reading exactly and interpolates drift in between base station readings with as little curvature as possible.

Comparing the linear and nonlinear drift curves, it is clear that the drift correction for station S is different. In the linear model $\Delta g_S = 0.041\,\text{mGal}$; in the nonlinear model $\Delta g_S = 0.05\,\text{mGal}$. This difference gives a sense of the uncertainty in the drift correction. The more base station readings made, the more likely the linear and nonlinear models will agree, but at the cost of making fewer readings at other gravity stations. If few base station readings are made, there will be greater uncertainty in estimation of instrument drift. If drift readings are only made at the beginning and end of the day, then drift can only be estimated using a linear model, and the uncertainty is not known.



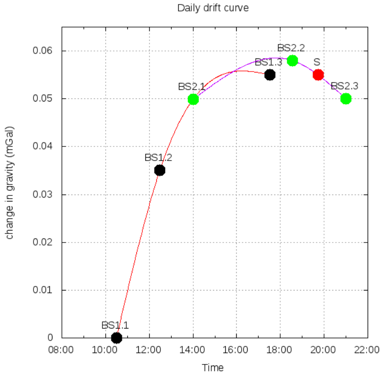
This cubic spline fit was graphed using gnuplot. Gnuplot has great functionality for fitting drift data using a variety of methods. The drift at time S was also determined using Gnuplot. Example codes for estimating the nonlinear drift are provided in the supplementary material.

Instrument drift during surveys with multiple base stations

It is often impractical to return to the base station to measure drift, especially in areas where transport is problematic or the survey area is very large. Usually in such circumstances, multiple base stations are established to measure drift.

The stations must be "tied" by overlapping measurements in time as illustrated on the drift curve plot at right. Three base station readings are shown for the first base station ($BS1.1$ – $BS1.3$). At 14:00, a second base station is established ($BS2.1$ – $BS2.3$). The difference in the drift curves interpolated between the time of measuring $BS2.1$ and $BS1.3$ gives a sense of uncertainty in the drift model. Drift can be estimated at time t_S for gravity station S because the drift curve for base station 2 is tied to the drift curve for base station 1. That is, the total drift when $BS2.1$ is established at 14:00 is known from the first base station drift curve. So, all drift during the survey can be corrected relative to the first reading of the day ($BS1.1$).

If multiple base stations are required and very high precision in the drift correction is needed, then a network is established by returning to both base station 1 and base station 2, and additional stations as needed, repeatedly during the survey.

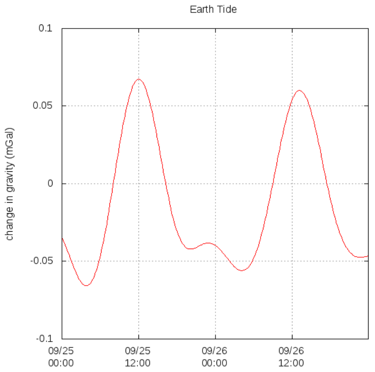


Tides and drift

A large fraction of the change in gravity with time is not due to actual instrument drift, but due to actual changes in gravitational acceleration associated with the Earth tide. The Earth tide, as distinguished from the ocean tide, is the change in acceleration due to the change in position of the Sun and Moon. Unlike the instrument drift discussed on previous slides, it is possible to develop a physical model of the expected change in Earth tide at any location and at any time on the surface of the Earth, as long as an astronomical model of the position of the Sun and Moon is available.

The plot at right shows a model of the expected variation in gravity due to the Earth tide calculated for a specific location and time.

Many modern field gravimeters include software that calculates the Earth Tide. Usually, these instruments will report the gravity reading, the tide correction, and the corrected value. Older field gravimeters do not include this feature, so often the tidal correction is applied and drift calculated on the residual.
Earth tide programs do not account for all possible tidal effects. For example, the ocean tide is usually out of phase with the Earth tide, but the change in ocean height with the tide may have a significant impact on gravity measured near the coast. This effect is not accounted for in the Earth tide correction.



The plot shows calculated upward acceleration due to the Earth tide. In order to correct for the tide the calculated values are added to the the observed gravity.

Summary of Drift and Tidal Corrections

- Drift and tide corrections are a critical part of gravity data collection and processing. The frequency with which base station readings are made depends on the precision required.
- Tidal corrections are routinely made using Earth tide codes and algorithms. These calculations do not account for all aspects of time variation in gravity due to the tide: ocean loading and the local elastic response to the tide also play significant roles.
- Usually the drift correction is made after the tidal correction is made. That is, the instrument drift is the residual time varying gravity after the theoretical Earth tide is removed.
- Instrument drift is not completely known during surveys. Therefore the instrument drift is modeled. Models include linear interpolation and nonlinear interpolation between base station readings.
- For multi-day surveys, it is critical that drift measurements be collected at a single base station each day (best method) or at multiple base stations that have been tied to the original base station.
- Often it is necessary to tie the entire survey to absolute gravity. This can be done if the network of new stations, including base stations, is tied to a point of known absolute gravity.

Theoretical gravity

Putman realized that gravity varies with latitude due to the change in diameter of the Earth with latitude and change in centrifugal acceleration with latitude. He used a method developed by Helmert to correct for this variation based on Clairant's theorem derived from LaPlace's equation. Now we use the Somigliana closed-form solution to estimate theoretical gravity, the expected value of gravity as a function of latitude, given the value of gravity at the equator:

$$g_T = \frac{g_e(1 + k\sin^2\phi)}{\sqrt{(1 - e^2\sin^2\phi)}},$$

where g_T, is the theoretical gravity on the GRS80 reference ellipsoid at latitude ϕ, g_e is normal gravity at the equator equal to 978032.67715 mGal, k is a dimensionless derived constant equal to 0.001931851353, and e is the first numerical eccentricity, with e^2 having a value of 0.0066943800229. Updated ellipsoids have been developed since 1980, of course, but they have a negligible effect (measured in thousandths of a milligal) on the theoretical gravity. Until a new ellipsoid is internationally accepted, the GRS80 should be used.

The latitude correction

To compare two gravity stations at different latitudes, it is necessary to compute theoretical gravity, g_T, for each station latitude. The difference in g_T between the two stations is the latitude correction. For example, if base station BS is south of gravity station S, and both are located in the northern hemisphere, then gravity is higher at S than at BS, and the latitude-corrected value of gravity at S will be less than the observed value at S.

Free air

The difference in elevation between the base station and the measurement point results in a difference in gravitational acceleration. The free air correction is applied to account for difference in gravity due to difference in height. For the GRS80 ellipsoid, the precise free air correction is:

$$\delta g_h = -(0.3087691 - 0.0004398 sin^2 \phi)h + 7.2125 \times 10^{-8} h^2,$$

where the free air correction, δg_h, is calculated in milliGals and h is the elevation of the gravity station at latitude ϕ with respect to the reference ellipsoid, measured in meters. Note that the free-air correction depends on latitude. That is, the vertical gradient in gravity varies with changes in the shape of the Earth.

An approximate formula, $\delta g_h = -0.3086h$, is widely used, especially to compare the free air correction among stations in a local network (where change in latitude is not significant).

Correcting for differences in elevation

To "free air" correct a gravity station it is necessary to compute δg_h, where h is the elevation difference between the gravity station and the reference ellipsoid. If the gravity station, S, is at a higher elevation than the reference ellipsoid, gravity at S will increase after the free-air correction is applied. Suppose a base station is at a higher elevation than station S. Gravity will decrease at station S after the free air correction is made, to "reduce" station S to the base station.

The atmospheric mass correction

For a gravity meter located at the surface of the Earth, the mass of the atmosphere pulls up on the meter, reducing gravity overall. The mass of the atmosphere varies with height and this change affects gravity measurements. The atmospheric correction attempts to account for the average change in the mass of the atmosphere between the base station, or reference ellipsoid, and the gravity measurement point. The formula for the atmospheric correction is:

$$\delta g_{atm} = 0.874 - 9.9 \times 10^{-5} h + 3.56 \times 10^{-9} h^2,$$

where the atmospheric correction, δg_{atm}, is given in milligals and h is the elevation of the gravity station in meters above mean sea level. For a station at sea level, the correction is 0.874 mGal. That is, the mass of the atmosphere reduces gravity at the sea level station compared to gravity at that position if no atmosphere were present. The effect of the atmosphere is less with increasing elevation because less atmosphere is pulling up on the meter. Note that the correction is done with respect to sea level rather than with respect to the ellipsoid. Density of the atmosphere also changes with time, but this change is not accounted for by the atmosphere correction.

Making the atmospheric mass correction

For all gravity stations located at the surface of the Earth, gravity will increase after the atmospheric correction is applied. The magnitude of the increase will be less for gravity stations located at higher elevations than for gravity stations located at lower elevations.

Simple Bouguer (Bullard A)

The Bouguer correction accounts for the mass of average crust between the base station, or reference ellipsoid, and the measurement point, given the height difference between them. If the measurement point located on the Earth's surface is above the reference ellipsoid, then the "extra" mass of rock between the station and reference ellipsoid has the effect of increasing gravity. Similarly, if a measurement station is located at an elevation below the reference ellipsoid then rock is "missing" that would theoretically be pulling up on the meter, with the effect of reducing gravity at that measurement station. The goal of the simple Bouguer correction is to remove this effect of "excess" and "missing" mass, using a formula based on a simple shape – the infinite slab (see Module 4):

$$\delta g_{bc} = 2\pi G \rho h$$

where ρ is the density of rock comprising the Bouguer slab, and h is the height difference between the gravity measurement point and the reference ellipsoid.

Making the Simple Bouguer correction

To make the Bouguer correction, it is necessary to compute δg_{bc}, where h is the elevation difference between the gravity station and the reference ellipsoid. If the gravity station, S, is at a higher elevation than the reference ellipsoid, gravity at S will decrease after the simple Bouguer correction is applied. Suppose a base station is at a higher elevation than station S. Gravity will increase at station S after the Bouguer correction is made, to "reduce" station S to the height of the base station.

Simple Bouguer (Bullard A + B)

The Bouguer correction can account for the spherical cap-shape of this mass of rock, as described in La Fehr (1991). Far from the gravity meter, the topography falls below the horizon because of the Earth's curvature. The spherical-cap correction accounts for this curved-shape of the Bouguer slab.The formula for the spherical-cap Bouguer correction is:

$$g_{sc} = 2\pi G\rho[(1+\mu)h - \lambda(R+h)],$$

where g_{sc} is the gravity correction due to the spherical cap in milligals, ρ is the density of the material making up the spherical cap, μ and λ are dimensionless coefficients that vary as a function of latitude. R is the mean radius of the Earth at the latitude of the gravity station and h is the elevation of the gravity station with respect to the reference ellipsoid. Unfortunately, the constants μ and λ are complex to calculate!

$$\mu = \frac{1}{3}\eta^2 - \eta$$

$$\lambda = \frac{1}{3}\left[(d + f\delta + \delta^2)\left((f-\delta)^2 + k\right)^{1/2} + p + m\ln\left(\frac{n}{(f - \delta + ((f-\delta)^2 + k)^{1/2}}\right)\right]$$

where: $d = 3\cos^2(\alpha) - 2$, $f = \cos(\alpha)$, $k = \sin^2(\alpha)$, $p = -6\cos^2(\alpha)\sin(\alpha/2) + 4\sin^3(\alpha/2)$, $m = -3\sin^2(\alpha)\cos(\alpha)$, $n = 2(\sin(\alpha/2) - \sin^2(\alpha/2))$, $\delta = R/(R+h)$, $\eta = h/(R+h)$, $\alpha = S/R$, $S = 166735\,\text{m}$, which is the standard radius of the spherical cap.

The simple Bouguer correction (Bullard A) is replaced with the spherical cap correction (Bullard A + B) to account for curvature of the slab. Is this curvature important? LaFehr found it can effect gravity up to about 0.1 mGal in comparing stations at greatly different elevations.

Summary

Most gravity surveys are done to identify subtle variations in density within the Earth, with the goal of better understanding the geology on a variety of scales, from the near-surface to the entire lithosphere. Because gravity anomalies are subtle, many factors not related to density variations within the Earth must be accounted for independently. Gravity processing is all about accounting for these factors. One person's trash, however, is another person's treasure. A great deal can be learned from studying gravity anomalies after the free-air correction has been made, and before the Bouguer correction has been made.

In this module we have concentrated on the steps needed to make a simple Bouguer correction.

People often ask, "do I add or subtract the correction?" Please never fall into that trap! The best strategy is to visualize what you are trying to accomplish with the correction and to verify that your correction achieves this goal! This becomes more important as an ever greater variety of platforms are conceived for measuring gravity with a ever wider variety of purposes.

There are additional corrections routinely done in processing gravity data. In airborne and marine surveys, the Estövös correction is used to account for the motion of the vehicle relative to the angular rotation of the Earth. The terrain correction is used to account for deviations of the topography from the ideal Bouguer slab or spherical cap, the isostatic correction is used to account for regional variations in gravity associated with isostatic compensation. When do these corrections need to be made? It is best to think of gravity processing as a process – isolate the anomaly of interest by performing the corrections necessary, and no more!

Further reading

Did you have the impression that gravity processing was simple? Many people struggle to make appropriate gravity corrections and it is not always clear, even to experts, which corrections are most appropriate. A simple paper describing standard corrections and an excel spreadsheet to make them is:

- Holm, D. I., and J. S. Oldow, 2007, Gravity reduction spreadsheet to calculate the Bouguer anomaly using standardized methods and constants, Geosphere, v. 3; no. 2; p. 86–90; doi: 10.1130/GES00060.1

A more complete discussion of the principles and standards for gravity processing is:

- Hinze, W. J., et al., 2005, New standards for reducing gravity data: The North American gravity database, Geophysics, v. 70, no. 4; p. J25–J32, doi: 10.1190/1.1988183.

A thorough development and discussion of the spherical cap correction is found in:

- LaFehr, T. R., 1991, An exact solution for the gravity curvature (Bullard B) correction, Geophysics, v. 56. no. 8, p. 1179–1184.
- Talwani, M., 1998, Errors in the total Bouguer reduction. Geophysics 63, Special Section: Shallow Seismic Reflection Papers, 1125–1130. doi: 10.1190/1.1444412
- LaFehr, T. R., 1998, On Talwani's "Errors in the total Bouguer reduction". Geophysics 63:4, 1131–1136.

Are these procedures set in stone? Not at all. See, for example, the paper on alternative reduction procedures:

- Nozaki, K., 2006, The generalized Bouguer anomaly, Earth Planets Space, 58, 287–303.

EOMA

1. Note that the atmospheric mass correction only accounts for change in the mass of the atmosphere above the gravimeter as a function of height. The atmospheric pressure also changes with time (so there is wind in the atmosphere!). How significant is the change in atmospheric pressure as a factor in drift? Approximate the atmosphere as an infinite slab. Recall that the pressure is related to density, height, and gravity ($P = \rho g_T h$), where g_T is the theoretical gravity at the location of the meter. Substitute this relationship into the infinite slab formula and solve for the change in gravity with a 1 millibar change in pressure, and a 10 millibar change in pressure. Discuss you results. Is these a significant factor for microgravity surveys?
2. Use the code *earthtide.pl* to calculate the upward acceleration due to the Earth tide at 0 N (the equator) and 90W starting on October 1, 2013, for a period of seven days. Make a plot of the output. Repeat the calculation for upward acceleration at 45 N, 90W for the same time period. How does the tide vary with latitude? Explain this result.
3. The Geophysics field camp made a gravity survey across the Borah Peak fault in the summer of 2013. They measured gravity at a base station intermittently during their survey. The base station measurements are located in the supplementary file: *field_camp_drift.dat*. Plot the drift curve using gnuplot (see the file *field_camp_nonlinear.gnu*). Describe an anomalies on this drift curve. Use the gnuplot script: *field_camp_drift_corrections.gnu* to estimate the drift using the cubic spline model. If you do not change the script, the output will be printed to the new file: *field_camp_interpolated_drift.out*. Now, using this model, correct the observed gravity readings found in the file: *field_camp_observed.dat* for drift. Note: these data have already had the tidal correction applied.
4. Develop a tool for processing gravity data. Make sure the tool can process data as described in this Module. You need to be able to calculate the simple Bouguer anomaly (Bullard A). You are welcome to implement the Bullard A+B (spherical cap) correction if you wish. Calculate the simple Bouguer anomaly for the data in *field_camp_observed.dat*, and plot this resulting profile of simple Bouguer gravity anomaly vs. distance from the base station. Note; you are welcome to develop this tool as a script in PERL, Python or Matlab, or as a spreadsheet.