Introduction to GEOL384/334 - Key points

- General remarks about this course
- Summary of geophysical methods
  - Why doing geophysical exploration
  - Survey objectives and practice
  - Absolute and relative measurements
  - Natural and controlled-source
- Physical properties of earth materials
- Concept and key properties of Potential Fields
- "Regional background" and "anomalies"
  - Target depth and resolution
- Signal and noise
- Measurement errors

- Reading:
  - Reynolds, Chapter 1
  - Dentith and Mudge, Chapters 1 and 2

About this course

- Years ago, I have come across the "dictionary" definition of a geophysicist shown on the right. I do not know the author, but this description summarized well some of the perception of geophysics by geologists.
- However, this definition is of course a joke. In this class, I intend to show that geophysics is actually simple. I try using few formulae, which should be understood intuitively and without lengthy derivations. Memorizing formulas will not be needed for the exam, but their understanding will be useful.
- We will also see that many figures in geophysics are indeed "debatable", and we will devote significant attention to evaluating these uncertainties.
- The following lecture notes are intended as complementary and not really repeating the material in the texts by Reynolds and Dentith and Mudge. The texts focus on practical introduction to geophysical applications, whereas my goal here is to explain the physical principles of these methods. I will also emphasize the relations of different methods to each other.

ge-o-phys-i-cist, *n.*

A geophysicist is a person who passes as an exacting expert on the basis of being able to turn out, with prolific fortitude, infinite strings of incomprehensible formulae calculated with micrometric precision from vague assumptions, which are based on debatable figures taken from inconclusive experiments, carried out with instruments of problematic accuracy by persons of doubtful reliability and questionable morality for the avowed purpose of annoying and confounding a hopeless chimerical group of fanatics known as geologists who are themselves the lunatic fringe of the scientific community.

Two groups of geophysical methods

- In this class, we will consider two broad groups of geophysical methods:
  - Methods using various forms of "quasi-static" or "stationary" oscillating fields (gravity, magnetic, low-frequency EM)
  - Methods using traveling and reflected waves (seismic and Ground Penetrating Radar; the latter is discussed in GEOL335)
- Both of these groups of methods are applied in a variety of environments (see table in the next slide):
  - Surface recording
    - At all scales from < 1 m to whole Earth
  - Boreholes (well logging)
  - Airborne, satellite
  - Laboratory samples (rock cores)
  - Natural and controlled-source (for electrical and EM)
- We will focus on the principles of the most common surface studies of gravity/mag/EM fields
  - In the last two weeks, we will briefly overview the wave-based seismic method
  - These wave methods are the focus of GEOL335

Summary of geophysical methods

Method	*Property*	*Value Measured*
Gravity	Density	Spatial variations in natural gravity field
Magnetic	Magnetic susceptibility	Spatial variations in natural magnetic field
Radioactive	Abundance of radionucleides	Gamma radiation
Heat flow	Thermal conductivity, radioactive heat production	Heat flow
Electrical	Electrical conductivity	Apparent resistivity
Telluric current	Electrical conductivity	relative apparent resistivity
Spontaneous polarization	Oxidation potential, ion concentrations	Natural electrochemical potentials
Induced polarization	Electronic conductivity	Polarization voltages
Electromagnetic	Electrical conductivity+ magnetic susceptibility	Alternating electromagnetic field, phase and intensity
Seismic	Natural ground motion, velocity, heterogeneities	Ambient seismic noise, travel times, polarization
	Velocity, impedance contrasts	Seismic travel times, amplitudes, reflection patterns
Remote sensing	Natural radiation	Refraction intensity
	Reflectivity (albedo)	Reflected radiation
Borehole	Natural radioactivity	Natural voltages, natural gamma radiation
	Electrical conductivity, seismic velocity, nuclear reactions	Apparent resistivity, travel times, amplitudes, induced and back-scattered radiation

When it comes to imaging and quantitative analysis of the subsurface structure, there is no alternative to

geophysical methods

Yellow background in this table shows the "controlled source" methods using artificial sources The rest are "passive-source" methods using physical effects produced by natural phenomena or (for example) urban noise

Why doing Gravity, Magnetic, or Electrical prospecting?

- Basically, we want to use all possible ways for determining the structure and properties of the subsurface
- Gravity, magnetic and electrical (and also their variants called EM, MT, IP, VLF, etc. discussed later) assess multiple physical properties (and this is good):
  - Density, total mass, magnetization, shapes of bodies (important for mining)
  - Basin shape (for oil/gas)
  - Near-surface stratigraphy and heterogeneity (for geotechnical characterization)
- An advantage of gravity/magnetic/electric work (compared, say, to seismology) is the presence of "zero-frequency" signal in the data
  - This allows seeing the effects of large-scale (regional, deep, or "background") structures as well as small (local, "anomalies") in the same data
  - This allows reconnaissance mapping of large areas by using wide station spacing
  - However, resolution is also much lower and quickly decreases with depth
- This work is relatively inexpensive
  - Cost ~$100 per km compared to $1000's for seismic
  - Practical with 1-3 person crews
  - Requires (comparatively) little data processing
- There is no alternative to geophysical methods if we need to know the structure
  - These are non-invasive methods that allow seeing how the structure continues, for example, from one borehole to the other
  - Seismic method is the best about the structure, but it is also the most expensive

Uses of gravity

- Specifically, gravity surveys are useful (in exploration) for:
  - Reconnaissance for oil and gas
  - Estimates of ore mass
  - Outline of the shape of ore mass
  - Locations of voids, tunnels, etc.

Uses of magnetics

- Magnetic surveys are useful for:
  - Reconnaissance for base metals
  - Location of alluvial gold and heavy minerals
  - Archaeology
  - Reconnaissance for oil and gas
  - Environmental geophysics
  - Identification of the basement and tectonic structure

Uses of electrical methods

- Electrical (resistivity) surveys are useful for:
  - Reconnaissance for metals
  - Location of metals
  - Groundwater
    - In particular, contaminated water is usually high in salinity
  - Engineering
    - Electrical resistivity depends on porosity, weathering, pore water, salinity, saturation, mineralization – all factors important for structure, stiffness

Depth and resolution

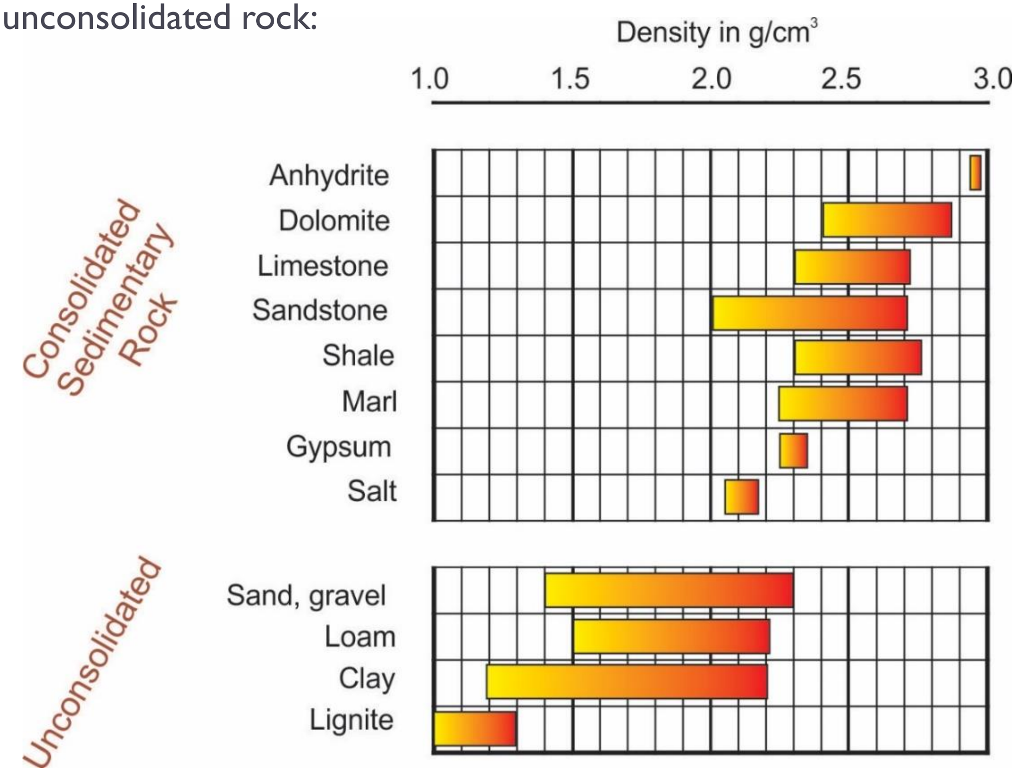
- For all methods, there are two important parameters determining the quality of imaging: imaging depth and resolution (ability to discern detail at that depth)
  - The depth and resolution are usually conflicting requirements (deeper penetration usually comes with poorer resolution of detail)
- With increasing characteristic scale of the anomaly (wave length in seismics):
  - Imaging depth increases
  - Resolution decreases
- For example, for gravity, the horizontal half-height size $w_{1/2}$ of an anomaly at depth h is approximately $w_{1/2} = h/0.65$
  - This is about the distance at which the anomalies can be separated laterally
- In addition, gravity/magnetic imaging is prone to an inherent uncertainty of depth estimation
  - For example, we will see that the source of a gravity or magnetic anomaly can in principle be always located shallower, even at zero depth, right beneath the observation surface
- It is important to realize these uncertainties when interpreting geophysical images

Reference levels, regional averages and “anomalies”

- Most methods that we will study employ “relative” measurements obtained by comparing current readings to adjacent areas or some “reference” values
  - The reason for this is that the typical absolute signal (for example, acceleration of gravity caused by the whole Earth) is much larger than its variation due to the structure of the near surface that we are trying to investigate. However, this absolute signal is generally constant within the span of the survey and can be subtracted.
- To obtain the reference level, two general strategies are used:
  - Repeated (for gravity) or continuous (magnetic) readings at a fixed “base station”. For example, analysis base station data gives the “drift correction” often necessary in gravity measurements (your lab #1)
  - Average regional trend estimated from the entire dataset.
- Interpretation of the average regional trend is known as the “regional/residual separation” (discussed in gravity lecture)
- After subtracting the strong regional trend, the remaining weak signal is reduced to localized “anomalies”
  - Based on locations and spatial shapes of these anomalies, various subsurface structures are characterized

Physical properties - density

- In the following slides, we briefly summarize the physical properties of subsurface rock used in geophysical exploration
  - See section 3.8 in Dentith and Mudge for more detail
- In gravity surveys, we measure variations of mass densities
  - Density can be precisely measured in rock samples
  - Density is lower and quite variable for unconsolidated rock:

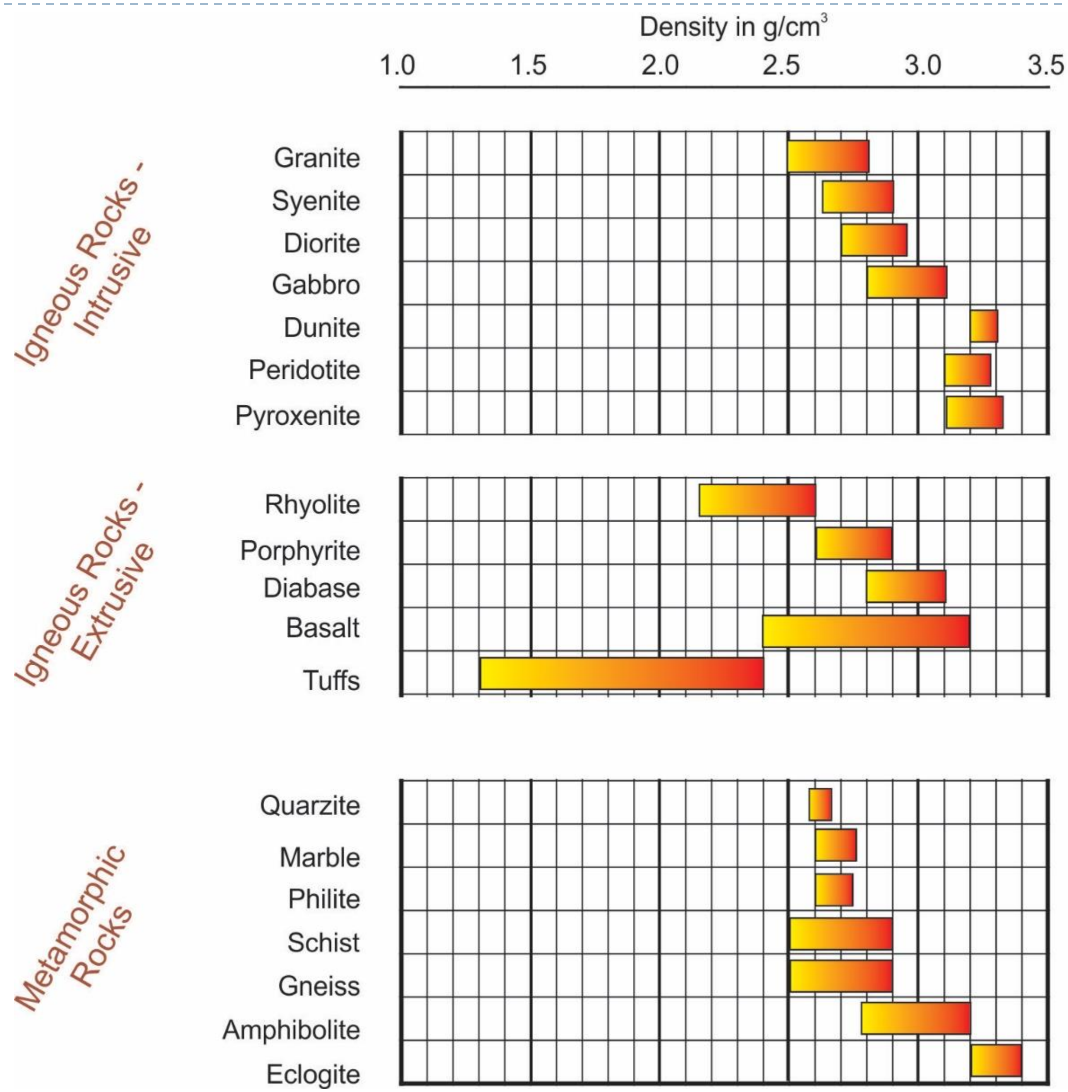


Densities of crystalline rocks

- Note the broader range of densities for deep-crustal and mantle rocks
- Hence, long-scale gravity signatures are dominated by the basement and mantle

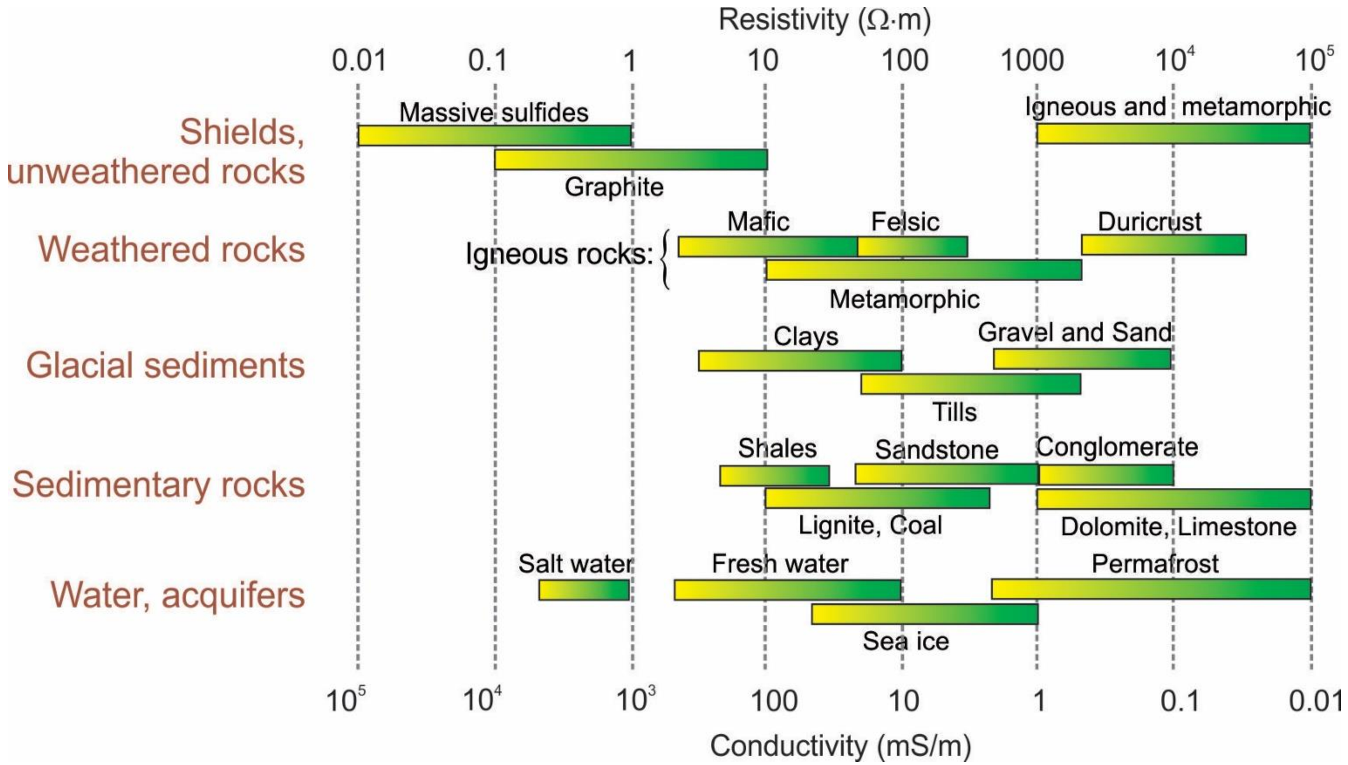
A commonly used estimate of the mean crustal density:

$\rho \approx 2.67\ \mathrm{g/cm^3}$



Physical properties - resistivities

- In electrical resistivity surveys, rock resistivity (or its inverse, conductivity) is measured
  - This quantity are much broader variable (by seven orders of magnitude)
  - Also easily measurable in rock samples or boreholes



Other physical properties – magnetic, dielectric, IP, SP

- In magnetic, EM, IP, SP, and seismic surveys, the properties affecting the resulting images are usually more complex than density and resistivity. We will discuss these properties later.
  - Because of such complexities, these surveys are usually interpreted less quantitatively, with emphasis on characterizing the spatial shapes of the anomalies
  - However, see section 3.9 in Dentith and Mudge for descriptions of magnetic susceptibility for rocks
- EM surveys are significantly affected by resistivity, or even the resistivity is their primary goal. However, the resistivity at finite frequencies is generally different from the one measured in static experiments (we will discuss this in Weeks 4-6)
  - Thus, it is most important that all of the different geophysical surveys sample different (albeit not always well understood) physical properties of the subsurface

Fields

- What is a field? In physics, field is simply any quantity continuously distributed in space and (possibly) variable with time: $f = f\left(x, y, z, t\right)$
  - Examples: temperature, intensity of light, pressure
  - More specific examples for us: mass density, electrical resistivity, or magnetization distributed in space
  - Depending on the type of quantity f, fields can be scalar, vector, or tensor
- In gravity work, we utilize the vector field **g** representing the force of gravity per unit mass:

$$\mathbf{g} = \frac{\mathbf{F}_g}{m}$$

- In magnetic surveys, we use a different type of vector field (denoted $\mathbf{B}$ or $\mathbf{H}$) describing the magnetic force $\mathbf{F}_\mathrm{m}$ by a somewhat different equation (discussed later)
- In seismic work, we simply use the vector field of ground displacement $\mathbf{u}$. In addition, useful tensor strain (denoted ε) and stress (σ) fields are derived from $\mathbf{u}$.

Potential fields

- The concept of potential is key for describing fields like gravity and magnetic, and is also useful in seismics
- A field is called potential if there exists a scalar field U ("the potential") such that the vector field strength ($\mathbf{g}$) represents the negative gradient of U:

$$\mathbf{g} = -\vec{\nabla} U \equiv -\frac{\partial U}{\partial \mathbf{x}}$$

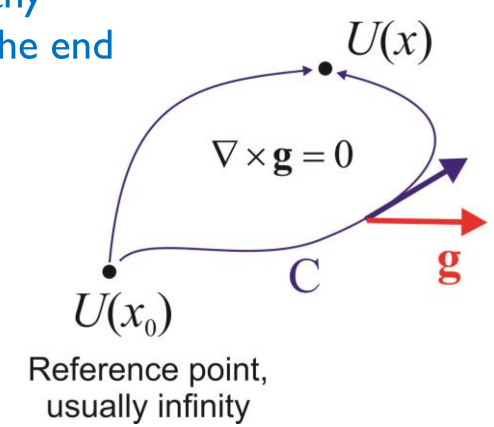
Here and below, bold $\mathbf{x}$ denotes 3-D coordinate vector (x,y,z)

  - This means that the three components of vector field can be derived from a single scalar function $U(\mathbf{x},t)$. This is a powerful physical principle and great simplification of the problem.

- The mentioned physical principle is that the work of $\mathbf{g}$ along any contour C connecting points $\mathbf{x}_0$ and $\mathbf{x}$ is only determined by the end points:

$$\int_C \mathbf{g} d\mathbf{x}' = U(\mathbf{x}_0) - U(\mathbf{x})$$

The work equals the reduction of the potential energy from the starting point to the ending point



Potential character of gravity field

- Consider the gravity strength field **g** (we will define it better later)
- Note that the above property means that **g** is curl-free (curl of a gradient is always zero):

$$\mathbf{curl}(\mathbf{g}) = \nabla \times \mathbf{g} = 0$$

- In addition, for gravity field above ground, the divergence of **g** is zero (no sources or sinks for **g**):

$$\mathrm{div}(\mathbf{g}) = \nabla \cdot \mathbf{g} = \frac{\partial g_x}{\partial x} + \frac{\partial g_y}{\partial y} + \frac{\partial g_z}{\partial z} = 0$$

Note: this is the definition of the divergence operation

- For gravity potential U, this is called the Laplace equation:

$$\nabla^2 U \equiv \frac{\partial^2 U}{\partial x^2} + \frac{\partial^2 U}{\partial y^2} + \frac{\partial^2 U}{\partial z^2} = 0$$

This upside down symbol "Δ" is called "nabla"

  - This equation relates the second derivatives of U
  - From field data, we measure the second derivatives with respect to x and y, and by this equation, we can then determine the variation of U with z. This is the principle of depth continuation for all potential fields.

Source of the field and Poisson's equation

- Physical fields are produced by sources
- For gravity, the mass density ρ is the source
  - For electrostatics and magnetics, the "source" density ρ is replaced with "charge" or "dipole moment", which can be negative
- The effect of the source consists in replacing zero in the right-hand side of eq.: $\operatorname{div}(\mathbf{g}) = 0$ with:

$$\operatorname{div}(\mathbf{g}) = -4\pi G\rho$$

$G \approx 6.674\times10^{-11}\ \mathrm{m^3 \cdot kg^{-1} \cdot s^{-2}}$ is the gravitational constant

- For the potential function U, this gives the Poisson's equation:

$$\nabla^2 U = 4\pi G\rho$$

- Thus, the essence of gravity, electrical, and magnetic methods can be summarized like this:
  - In the free space above ground (where $\rho = 0$), we have the Laplace equation for U
  - Below the surface, we have the Poisson's equation containing ρ
  - $U(x,y,z)$ is continuous across the surface
  - By inverting these equations, from $U(x,y,z)$ measured on the surface, we determine $\rho(x,y,z)$ below the surface

Gauss's law

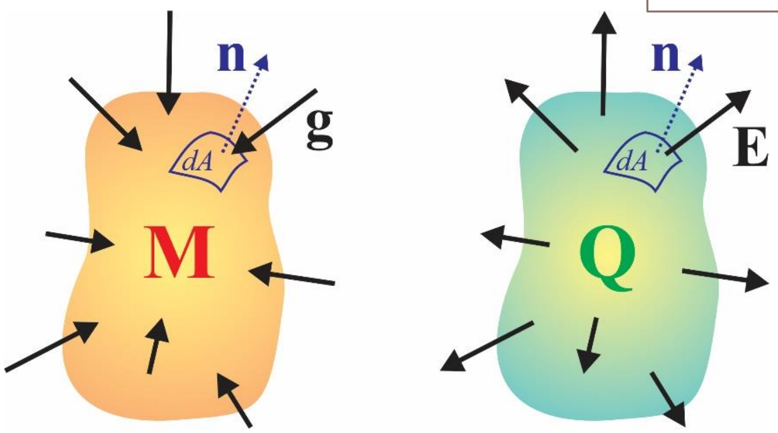
- A simple and useful understanding of most problems in gravity and electrostatics can be obtained without differential equations, by utilizing a theorem called the Gauss's law
- For gravity, this law states that the flux of the field **g** out of a closed volume is proportional to the total mass M within the volume:

$$\oiint_{\text{closed surface}} \mathbf{g} \cdot \mathbf{n} dA = -4\pi G \int_{\text{volume}} \rho dV = -4\pi G M$$

This is the flux of **g** out of the volume

n denotes the outward-directed unit vector orthogonal to the surface, and dA is an element of surface area

M is the total mass enclosed within the volume, regardless of its shape



In gravity

In electrostatics

Finding U from $\mathbf{g}$

- Thus, another way to explain a potential field is this: a potential field is a field with zero curl:

$$\nabla \times \mathbf{g} = 0$$

- For a field with zero curl, the potential can be calculated as:

$$U(\mathbf{x}) = U(\mathbf{x}_0) - \int_C \mathbf{g}\, d\mathbf{x}'$$

where C is an *arbitrary* path from $\mathbf{x}_0$ to $\mathbf{x}$

  - with an arbitrary reference point $\mathbf{x}_0$.
  - U($\mathbf{x}_0$) is some arbitrary level of U. It always cancels out in practical problems.
  - For gravity, this formula gives the well-known relation for potential energy of mass m elevated to some '*height*':

$$E_{\text{potential}} = mU = mg \times height + const$$

- Note that the potential always contains an inherent arbitrary constant (the selected value of $U(\mathbf{x}_0)$)
- The reference point $\mathbf{x}_0$ is usually placed at the infinity, and its potential is assigned $U(\mathbf{x}_0) = 0$

Signal and noise in data

- Geophysical data always contain some SIGNAL and some NOISE
  - Signal is the 'deterministic' part of the record that we want to know. What does this mean?
    - "Signal" is consistent with the interpretation method employed. Examples: regional average or gravity anomaly that we will interpret later.
  - Noise - anything else mixed into the measurement
- Sources of noise:
  - Instrumental and operator errors
  - Geologic sources (structure that we do not consider). For example, dipping structure in cross-line directions are often a major and almost unsurmountable source of "noise" in 2-D imaging
  - Too simple theory (e.g., 2-D sounding in a 3-D Earth).
    - This noise is called "methodological" by Dentith and Mudge
    - This is not really noise but limitation of interpretation. It is unavoidable and must be realized during interpretation.

Signal and noise in data

- Types of noise
  - Coherent (caused by the source and the earth themselves). This noise is worst of all and usually intractable. However, in gravity/magnetic imaging, there exist some "corrections" for several simple types of these noises.
    - This includes the "footprint" of the acquisition system
  - Incoherent (random, coming from unrelated sources). Such noise can be reduced by redundant recording and filtering
- Main task of initial data processing is to separate the signal from noise and increase the signal/noise (S/N) ratio. This is done by:
  - Acquiring redundant data (repeating gravity or magnetic readings, etc.)
  - Merging multiple datasets (with "leveling" on their edges)
  - Removing instrumental drifts
  - Separating the regional trends (longer-wavelength) and residuals (more local)
  - Other types of filtering (e.g., 2-D)
  - Plotting the data in various ways emphasizing the signal

Statistical measure of random noise - variance

- Random noise is commonly measured by its "variance" (denoted σ^2)
  - Variance is the *squared* mean statistical error, and therefore σ measures the mean error
- For example, for gravity $\boldsymbol{g}$, if we imagine an infinite number of measurements for g, each occurring with "probability density" (p.d.f.) $p(g)$, then the variance is:

$$\sigma^2 = \left\langle \left(g - \langle g \rangle\right)^2 \right\rangle \equiv \left\langle (\delta g)^2 \right\rangle \equiv \int \left(g - \langle g \rangle\right)^2 p(g)\, dg$$

where the mean of any quantity y is denoted by: $\langle y \rangle \equiv \int y p(y)\, dy$

(Also note that $\int p(y)\, dy = 1$)

This is the key property of "probability density function" (p.d.f.) $p(y)$

Standard deviation

- However, we always have only a finite number of measurements, and so need to *estimate* $<g>$ and σ^2 from them
- For N measurements, these estimates are:

$$\langle g \rangle \approx \bar{g}_N \equiv \frac{1}{N}\sum_{i=1}^{N} g_i$$

Arithmetic mean, or "sample mean"

$$\sigma^2 \approx s_{N-1}^2 \equiv \frac{1}{N-1}\sum_{i=1}^{N}\left(g_i - \bar{g}_N\right)^2$$

$s_{N\text{-}1}$ is the "standard deviation",
s_{N-1}^2 is called "sample variance"

- Thus, from N measurements, the expected mean absolute error is the standard deviation:

$$\sigma \approx s_{N-1} \equiv \sqrt{\frac{1}{N-1}\sum_{i=1}^{N}\left(g_i - \bar{g}_N\right)^2}$$

Another estimate of random data scatter (σ)

- Thus, from N measurements, the expected mean absolute error is the standard deviation:

$$\sigma \approx s_{N-1} \equiv \sqrt{\frac{1}{N-1}\sum_{i=1}^{N}\left(g_i - \bar{g}_N\right)^2}$$

- Sometimes, the scatter in the data is measured by averaging the squared differences of consecutive observations taken at the same location:

$$\tilde{s}_{N-1} \equiv \sqrt{\frac{\sum_{j=1}^{N-1}\left(g_{j+1} - g_j\right)^2}{N-1}}$$

  - note that N-1 here is the number of pairs of repeated measurements
- This is an approximate standard deviation of the drift (drift is the variation of gravity recordings with time occurring due to Earth's tides and changes within the gravity meter)
- This estimate of error is called REPEATABILITY in your Lab #1