Engineering and Environmental Geophysics

Gravity, magnetic, electrical and electromagnetic surveying

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INTRODUCTION

Warning: This script is only a complement to the PowerPoint presentation. The related illustrations and mathematical expressions can be found directly in the slides.

Objectives of geophysical surveys

Geophysics applies the principles of physics to the study of the earth. Three classes of objectives are addressed by geophysical surveys: the measurement of geologic features, the in situ determination of engineering properties, and the detection of hidden cultural features. Geologic features may include faults, bedrock topography, discontinuities, voids, and groundwater. Engineering properties that can be determined in situ include elastic moduli, electrical resistivity and, to a lesser degree, magnetic and density properties. Hidden cultural features available for geophysical detection and characterization include buried underground tanks and pipes, contaminant plumes, archaeological structures and landfill boundaries.

Geophysical methods

Geophysical methods can be classified as **active or passive techniques**. Active techniques impart some energy or effect into the earth and measure the earth materials' response. Passive measurements record the strengths of various natural fields which are often continuous in existence. Active techniques generally produce more accurate results or more detailed solutions due to the ability to control the size and location of the active source. **Classified by physical effect measured**, the following surface techniques (with their operative physical properties in italic) are considered herein:

- Gravitational field method (*density*)
- Magnetic field method (*magnetic susceptibility* and *remanence*)
- Electrical and electromagnetic methods with natural electrical fields (self-potential: *electrical conductivity*), resistivity (*electrical conductivity)* and induced polarization (*electrical capacitance*) and electromagnetic (*electrical conductivity* and *inductance*)
- Wave field techniques: seismic (*density* and *elastic moduli*) method and *dielectric constant* (radar) techniques.

Geophysical measures for environmental studies can also be applied in the subsurface. Down-hole application of geophysics provides in situ measurements adjacent to the borehole (well-logging), between boreholes (crosshole) or across the medium to the surface (hole-to-surface).

General observations

Several general observations should be kept in mind when considering applications of geophysical methods. First, problems in geological, geotechnical or environmental projects require some **basic geological information** prior to use of geophysical techniques. A well-designed survey will provide the best solution at the lowest cost. The vast majority of objectives are inferred from the known geologic data and the measured geophysical **contrast** (e.g. a difference in resistivity between the medium and a target). If no sufficient contrast is present, the geophysical survey cannot be successful. The interpretation of geophysical contrasts is based on geologic assumptions or direct measurement of physical rock properties (e.g. on an outcrop).

The aim of the geophysical techniques is the creation of a **model**, i.e. a simplified and ideal view of a physical reality. The geophysical model often does not correspond exactly to the geological model. The correlation of measured geophysical contrasts with geologic inferences most often is empirical and certainly is dependent on the quality of both the results and the hypotheses. Preparation of geophysical models almost always assumes the following:

- Earth materials have distinct subsurface boundaries.
- A material is homogeneous (having the same properties throughout).
- The unit is isotropic (properties are independent of direction).

These assumptions are, in many cases, at variance with the reality of geologic occurrences. Units may grade from one material type to another with no distinct surface between two materials. At some scale, inhomogeneities exist in practically all units. Properties may occasionally vary greatly in magnitude with direction (anisotropy), such as in shales.

Forward solutions proceed from cause to effect (i.e. measure data) and are unique determinations. But most geophysical methods do not directly measure the parameter desired by the geologist or engineer. Usually a model is obtained using an **inverse solution** (i.e. creating a model from the measured data). Inversion implies that a cause was inferred from an effect. The physical property, the cause, is inferred from the field survey readings, the effects. **Inverse resolutions are not unique** conclusions, and provide a most likely solution selected from numerous possibilities. Ambiguity, however, can be summarized as an equivalence of geometry/size and a material’s properties. Structure may be reevaluated by changing physical parameters. Ambiguity applies to all geophysical methods, and is most conveniently resolved by understanding geologic reality in the interpretation. The extent to which these presumptions are valid or the magnitude that the assumptions are in error will have a direct bearing on the conclusions.

Interpretation is a continuous process throughout geophysical investigations. Implementation of a geophysical model, which satisfactorily accounts for the geophysical observations, and can be correlated with available ground truth (e.g. boreholes) can be a laborious interpretative process, especially since iterations of both the geophysical models and the geologic model are usually required. Production of the final product in a form (image) useful to the customer (engineer or geologist) is the most necessary interpretative step.

Applications and limitations

The number of geologic issues considered are limited to the problems most commonly encountered in an engineering or environmental context, since the number of geologic problems is vastly larger than the number of geophysical methods. One cannot rely blindly on the applicability of the table of possible application shown in the slides, because geology is the most important ingredient of the selection of a method. This table will suggest potential geophysical techniques for particular needs. Geologic input, rock property estimates, modeling, interference effects, and budgetary constraints are co-determining factors of method selection. The geophysicist(s) must have access to all relevant information concerning the site. This data includes: site geology, site maps, boreholes logs, sources and contaminant types that are known or presumed, hazards and safety conditions impacting field work, **geophysical noise** (often related to industrial or urban location), etc. Choosing the appropriate technique is often not obvious and the **combination of several techniques** frequently required.

GRAVITY SURVEYING

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Introduction

Lateral **density changes** in the subsurface cause a change in the force of gravity at the surface. The intensity of the force of gravity due to a buried mass difference (concentration or void) is superimposed on the larger force of gravity due to the total mass of the earth. Thus, two components of gravity forces are measured at the earth's surface: first, a general and relatively uniform component due to the total earth, and second, a component of much smaller size which varies due to lateral density changes (the **gravity anomaly**).

Applications

For engineering and environmental applications (**microgravity**), the scale of the problem is generally small (targets are often from 1-10 m in size: cavities, paleo-valleys). Therefore, conventional gravity measurements, such as those made in petroleum exploration, are inadequate. Station spacings are typically in the range of 1-10 m. Microgravity requires preserving all of the precision possible in the measurements and analysis so that small objects can be detected. By very precise measurement of gravity and by careful corrections for variations in the larger component due to the whole earth, a gravity survey can detect natural or man-made voids, variations in the depth to bedrock, and geologic structures of engineering interest.

The first question to ask when considering a gravity survey is our ability to detect a resultant gravity anomaly. Inputs required are the probable geometry of the anomalous region, its depth of burial, and its density contrast. A generalized rule of thumb is that a body must be almost as big as it is deep.

Density of rocks

Values for the density of shallow materials are determined from laboratory tests (samples). Density estimates may also be obtained from geophysical well logging. Densities of a specific rock type on a specific site will not have more than a few percent variability as a rule. However, unconsolidated materials such as alluvium and stream channel materials may have significant variation in density, due to compaction and cementation.

Gravity of the Earth and units of gravity

In 1687 Newton published his work on the universal law of gravity in his book "Philosophiae Naturalis Principia Mathematica". **Newton's law of gravitation** states that every particle in the universe attracts every other particle with a force that is directly proportional to the product of their masses and inversely proportional to the square of the distance between them. Every planetary body, including the earth, is surrounded by its own gravitational field, which exerts an attractive force on any object that comes under its influence. This field is proportional to the body's mass and varies inversely with the square of distance from the body.

In physics, the units for gravity (**acceleration**) is m/s^2. In gravity surveying the unit is the **milligal** (1 mgal = 10^{-5} m/s^2). The gravitational field is numerically equal to the acceleration of objects under its influence, and its value at the earth's surface (for an homogeneous, spherical, non-rotating earth), denoted g_N, is approximately 9.81 m/s². This means that, ignoring air resistance, an object falling freely near the earth's surface increases in speed by 9.81 m/s for each second of its descent. Thus, an object starting from rest will attain a speed of 9.81 m/s after one second, 19.62 m/s after two seconds, and so on. According to Newton's Law, the earth itself experiences an equal and opposite force to that acting on the falling object, meaning that the earth also accelerates towards the object. However, because the mass of the earth is huge, the acceleration produced on the earth by this same force is negligible.

The **spheroid** is an ideal representation of the earth as an ellipse of rotation. The gravity (g) is the sume of the gravitational acceleration (g_N) and the centrifugal force (g_c). The **geoid** is essentially the figure of the earth abstracted from its topographic features. It is an idealized equilibrium surface of sea water, the mean sea level surface in the absence of currents or air pressure variations and continued under the continental masses. The geoid, unlike the ellipsoid (spheroid), is irregular and too complicated to serve as the computational surface on which to solve geophysical problems. A **reference ellipsoid**, customarily chosen to be the same size (volume) as the geoid, is described as an oblate ellipsoid that approximates the mean sea-level surface (geoid) with the land above removed The reference ellipsoid is defined in the **Gravity Formula 1967** and is the model used in gravity surveying. The geometrical separation between it and the reference ellipsoid varies globally between ±110 m.

Measurement of gravity

In gravity surveying, we only measure the **vertical component of the gravity field**. **Absolute measurements** give directly the value of g but are tedious to get. In practice, **relative measurements** are used. Modern instruments (stable or astable gravimeters) have a least-significant precision of 0.01 to 0.001 mgal. Factors which may significantly increase the error are soft footing for the gravimeter plate (surveys have been done on snow), wind, and ground movement (trucks, etc.) in the frequency range of the meter. Special versions of these meters which filter the data appropriately (stack measurements during a certain time period, CG5 Autograv) can compensate for some of these effects.

Gravity surveying, data correction

Up to 50 percent of the work in a microgravity survey is consumed in the surveying. For the very precise work described above, **relative elevations** for all stations need to be established to ±1 cm. A firmly fixed stake or mark should be used to allow the gravity meter reader to recover the exact elevation. **Position in plan** is less critical, but relative position to 50-100 cm or 10 % of the station spacing (whichever is smaller) is usually sufficient. Differential GPS, can achieve the required accuracy, especially the vertical accuracy, only with the best equipment under ideal conditions.

High station densities are often required. It is not unusual for intervals of 1-3 m to be required to map anomalous masses whose maximum dimension is 10 m. Because the number of stations in a grid goes up as the square of the number of stations on one side, profiles are often used where the geometry of the longest dimension of the sought body can be established before the survey begins. After elevation and position surveying, actual measurement of the gravity readings is often accomplished by one person in areas where solo work is allowed.

Gravity measurements, even at a single location, change with time due to **earth tides**, **meter drift**, and **human inattention errors**. These time variations can be dealt with by good field procedure. The earth tide may cause changes of 0.24 mgal in the worst case but it has period of about 12.5 hours; it is well approximated as a straight line over intervals of 1-2 hours or it can be calculated and removed. Drift, due to short-term variations in gravimeter readings caused by less than perfect elasticity of the moving parts of the suspension, is also well eliminated by linear correction. Detection of human inattention errors is also a matter of good field technique and repeated readings.

The most commonly used survey improvement technique is to choose one of the stations as a **base** and to reoccupy that base periodically throughout the working day (**loops**). The observed base station gravity readings are then plotted versus time, and a line is fitted to them to provide time rates of drift for the correction of the remainder of the observations. Typically eight to ten measurements can be made between base station occupations; the time between base readings should be on the order of 1-2 hours. If base reoccupations are done approximately every hour, known errors such as the earth tide are well approximated by the removal of a drift correction that is linear with time. Even if the theoretical earth tide is calculated and removed, any residual effects are removed along with instrumental drift by frequent base station reoccupation. Where higher precision is required, each station must be reoccupied at least once during the survey. For example, if the stations are in an order given by a,b,c,d,e,f,... then the station occupations might be in the sequence ab, abc, abcd, bcde, cdef, defg, etc. In this way, each station would be occupied four times. Numerical adjustments, such as least squares minimization of deviations, may be applied to reoccupation data sets. This procedure allows data quality to be monitored and controlled and distributes any systematic errors throughout the grid.

This base can also be used to link the relative gravity difference to an **absolute value of gravity** using the Swiss network.

After correcting the data, the **observed gravity** g_{obs} is obtained.

Data processing: gravity reduction

The **Bouguer anomaly** is defined as the difference between the observed gravity and the modelled gravity. The **modelled gravity** is obtained using a series of 4 corrections. This operation is called **gravity reduction**.

The smooth latitude dependence of gravity (**latitude correction**) is given by the equation based on the reference spheroid. Calculations will show that if the stations are located with less than 0.5 to 1.0 m of error, the error in the latitude variation is negligible. In fact, for surveys of less than 100 m north to south, this variation is often considered part of the regional and removed in the interpretation step. Where larger surveys are considered, the Gravity Formula 1967 gives the appropriate correction.

The **free-air** and the **Bouguer-slab** are corrected to a datum (the datum is an arbitrary plane of constant elevation to which all the measurements are referenced). The formula for free air and Bouguer corrections contains a surface density value in the formulas. If this value is uncertain, its multiplication times the altitude change necessary to bring all stations to a common level can lead to error. Obviously, the size of this altitude change for each station is dependent on surface topography. All of these estimates are based on a mistaken estimate of the near surface-density, not the point-to-point variability in density, which also may add to the error term.

The terrain correction (basically due to undulations in the surface near the station) has two components of error. One error is based on the estimate of the amount of material present above and absent below an assumed flat surface through the station. This estimate must be made quite accurately near the station; farther away some approximation is possible. In addition to the creation of the geometric model, a density estimate is also necessary for terrain correction. This estimate does not include terrain density variations. Even if known, such variations are difficult to apply as corrections.

Once the **Bouguer anomaly** is obtained, an important step in the analysis remains. This step, called **regional-residual separation**, is one of the most critical. In most surveys, and in particular those engineering applications in which very small anomalies are of greatest interest, there are gravity anomaly trends of many sizes. The larger sized anomalies will tend to behave as regional variations and the desired smaller magnitude local anomalies will be superimposed on them. A simple method of separating residual anomalies from microregional variations is simply to visually smooth contour lines or profiles and subtract this smoother representation from the reduced data. The remainder will be a **residual anomaly** representation. However, this method can sometimes produce misleading or erroneous results. Several automatic versions of this smoothing procedure are available including polynomial surface fitting and band-pass filtering. The process requires considerable judgement and whichever method is used should be applied by an experienced interpreter. Note that certain unavoidable errors in the reduction steps may be removed by this process. Any error which is slowly varying over the entire site, such a distant terrain or erroneous density estimates, may be partially compensated by this step. The objective is to isolate the anomalies of interest. Small wavelength (about 10 m) anomalies may be riding on equally anomalous measurements with wavelengths of 100 or 1,000 m. The scale of the problem guides the regional-residual separation.

Intrepretation

Different techniques can be used to interpret the residual anomaly. Using the typical response of simple shapes, some parameters about the size of the object or the depth to the target can be obtained using a **direct interpretation** technique. Unless you are very close to the body its exact shape is not important. Thus, a rectangular-shaped horizontal tunnel can be modeled by a horizontal circular cylinder and a horizontal cylinder sometimes can be modeled by a horizontal line source.

Software packages for the **indirect interpretation** of gravity data are numerous. The usual inputs are the residual gravity values along a profile or traverse. The traverse may be selected from a grid of values along an orientation perpendicular to the strike of the suspected anomalous body. The interpreter then constructs a subsurface polygonal-shaped model and assigns a density contrast to it. The gravity response is the sum of all the basic polygons that constitutes the model. When the trial body has been drawn, the computer calculates the gravity effect of the trial body and shows a graphical display of the observed data (residual anomaly), the calculated data due to the trial body and often the difference. The geophysicist can then begin varying parameters in order to bring the calculated and observed values closer together. Parameters usually available for variation are the vertices of the polygon, the length of the body perpendicular to the traverse, and the density contrast. Most programs also allow multiple bodies. Because recalculation is done quickly, the interpreter can rapidly vary the parameters until an acceptable fit is achieved. Additionally, the effects of density variations can be explored and the impact of ambiguity evaluated (see below). Alternatively, this adjustment can be automatically done via an **inversion algorithm**, possibly using predefined constraints. In that case, the algorithm will automatically minimize the error between the observed data and the calculated data.

Gravity surveys are limited by **ambiguity** (non-unicity) and the assumption of homogeneity. A distribution of small masses at a shallow depth can produce the same effect as a large mass at depth. External control of the density contrast or the specific geometry (form of geologic plausibility, drill-hole information, or measured densities) is required to resolve ambiguity questions. A well designed interpration/inversion program should always allow to include **constrains** (e.g. borehole information) to reduce the ambiguity.

MAGNETIC SURVEYING

Warning: This script is only a complement to the PowerPoint presentation. The related illustrations and mathematical expressions can be found directly in the slides.

Introduction

Magnetic surveys are based on the investigation on the basis of anomalies in the Earth´s magnetic field resulting from the magnetic properties of the underlying rocks (**magnetic susceptibility and remanence**).

Applications

In most engineering and environmental scale investigations, the sedimentary and alluvial sections will not show sufficient contrast such that magnetic measurements will be of use in mapping the geology. However, the presence of ferrous materials in ordinary municipal trash and in most industrial waste does allow this technique to be effective in direct detection of landfills. Other ferrous objects which may be detected include pipelines, underground storage tanks, archaeological features and unexploded ordnance.

Magnetic rock properties

Many rocks and minerals are weakly magnetic or are magnetized by induction in the earth’s field, and cause spatial perturbations or “anomalies” in the earth’s main field. Man-made objects containing iron or steel are often highly magnetized. The most important exception is magnetite. From a geologic standpoint, magnetite and its distribution determine the magnetic properties of most rocks. There are other important magnetic minerals in mining prospecting, but the amount and form of magnetite within a rock determines how most rocks respond to an inducing field. Iron, steel, and other ferromagnetic alloys have susceptibilities one to several orders of magnitude larger than magnetite. The exception is stainless steel, which has a small susceptibility. The importance of magnetite cannot be exaggerated. The table in the slides provides some typical values for the susceptibility of rock materials. Note that the range of values given for each sample generally depends on the amount of magnetite in the rock.

Theory: Earth’s magnetic field and units

The earth possesses a magnetic field caused primarily by sources in the core. The form of the field is roughly the same as would be caused by a dipole or bar magnet located near the earth’s center and aligned subparallel to the geographic axis. Secondary dipoles fields can explained local variations in this model. A small fraction only of the earth’s magnetic field comes from the ionosphere. The earth’s magnetic field dominates most measurements on the surface of the earth. The earth’s total field intensity varies considerably by location over the surface of the earth and also in time. (e.g. secular variations, periodic changes detected in paleomagnetic studies). To describe the geomagnetic field, two elements are used: the **inclination** and **declination**. The intensity of the earth’s field is customarily expressed in S.I. units as **nanoteslas** (nT). This field has a value of about 47,000 nT in Switzerland.

The slides exposes in a logical approach the main basic concepts and definitions in physics applied to the study of magnetic phenomena (and the reader is encouraged to refer to these slides to refresh these notions!). To summarize, we can say that the **induced magnetization** refers to the action of the **magnetic induction field** on the material causing the material itself to act as a magnet. The field caused by such a material is directly proportional to the intensity of the ambient field and to the ability of the material to enhance the local field, a property called **magnetic susceptibility**. The induced magnetization is equal to the product of the volume magnetic susceptibility and the inducing field of the earth.

Magnetic measurements

Ground magnetic measurements are usually made with portable instruments called **magnetometers,** at regular intervals along more or less straight and parallel lines which cover the survey area. Often the interval between measurement locations (stations, located using a differential GPS) along the lines is less than the spacing between lines. In the proton magnetometer, a magnetic field which is not parallel to the earth's field is applied to a fluid rich in protons causing them to partly align with this artificial field. When the controlled field is removed, the protons precess toward realignment with the earth's field at a frequency which depends on the intensity of the earth's field. By measuring this precession frequency, the total intensity of the field can be determined. The physical basis for several other magnetometers, such as the cesium-vapor magnetometers, is similarly founded in a fundamental physical constant. The optically pumped magnetometers have increased sensitivity and shorter cycle times making them particularly useful in airborne applications. The incorporation of computers and built-in memory in magnetometers has greatly increased the ease of use and data handling capability of magnetometers. The instruments typically will keep track of position, prompt for inputs, and internally store the data for an entire day of work.

Magnetic surveying, data corrections

The magnetometer is operated by a single person. Data recording methods (direct readings or averaging) will vary with the purpose of the survey and the amount of noise present. In either case, the time of the reading is also recorded unless the magnetometer stores the readings and times internally (because of temporal changes in the magnetic field). To obtain a representative reading, the sensor should be operated well above the ground to avoid picking small near surface features (e.g. cans). One obvious exception to this is ordnance detection where the objective is to detect near-surface objects. Ordnance detection requires not only training in the recognition of dangerous objects, but experience in separating small, intense, and interesting anomalies from more dispersed geologic noise. For some purposes a close approximation of the **gradient** of the field is determined by measuring the difference in the total field between two closely spaced sensors. The quantity measured most commonly is the vertical gradient of the total field.

Intense fields from man-made electromagnetic sources can be a **problem in magnetic surveys**. Most magnetometers are designed to operate in fairly intense 60-Hz and radio frequency fields. However extremely low frequency fields caused by equipment using direct current or the switching of large alternating currents can be a problem. Pipelines carrying direct current for cathodic protection can be particularly troublesome. Although some modern ground magnetometers have a sensitivity of less than 0.1 nT, sources of cultural and geologic noise usually prevent full use of this sensitivity in ground measurements. Steel and other ferrous metals in the vicinity of a magnetometer can distort the data. Nearby metal objects may cause interference. Some items, such as automobiles, are obvious, but some subtle interference will be recognized only by an operator with large expertise. Old buried curbs and foundations, buried cans and bottles, power lines, fences, and other hidden factors can greatly affect magnetic readings. Moreover, large belt buckles, zippers, watches, eyeglass, bra frames, keys, and pencils, can all contain steel or iron and must be removed when operating the unit. A compass should be more than 3 m away from the magnetometer when measuring the field! A final test is to immobilize the magnetometer and take readings while the operator moves around the sensor. If the readings do not change by more than 1 or 2 nT, the operator is "magnetically clean."

To make accurate anomaly maps, temporal changes in the earth's field during the period of the survey must be considered. Normal changes during a day, sometimes called **diurnal drift**, are a few tens of nT but changes of hundreds or thousands of nT may occur over a few hours during **magnetic storms** due to solar activity. During severe magnetic storms, which occur infrequently, magnetic surveys should not be made. The correction for diurnal drift can be made by repeat measurements of a base station at frequent intervals. The measurements at field stations are then corrected for temporal variations by assuming a linear change of the field between repeat base station readings. Continuously recording magnetometers can also be used at fixed base sites to monitor the temporal changes. If time is accurately recorded at both base site and field location, the field data can be corrected by subtraction of the variations at the base site. The base-station memory magnetometer, when used, is set up every day prior to collection of the magnetic data. The base station ideally is placed at least 100 m from any large metal objects or travelled roads and at least 500 m from any power lines when feasible. The base station location must be very well described in a field book as others users may have to locate it based on the previous written description.

The **International Geomagnetic Reference Field** (IGRF) defines the theoretical undisturbed magnetic field at any point on the earth's surface in simulating the observed geomagnetic field by a series of dipoles. This formula is used to remove from the magnetic data those magnetic variations attributable to this theoretical field (latitude correction).

It can be emphasize that **no elevation correction** is applied for ground surveys. The terrain correction is very difficult to applied (generally rarely applied) since we need to know about the magnetic properties of the topographic features.

After all corrections have been made, magnetic survey data are usually displayed as individual profiles or as contour maps. Identification of anomalies caused by cultural features, such as railroads, pipelines, and bridges is commonly made using field observations and maps showing such features. A regional gradient allows for the description of a **residual magnetic anomaly**.

Interpretation

Total magnetic disturbances or anomalies (difference between the regional gradient and the observed magnetic data) are highly variable in shape and amplitude; they are almost always asymmetrical (unlike in gravity!), sometimes appear complex even from simple sources, and usually show the combined effects of several sources. An infinite number of possible sources can produce a given anomaly, giving rise to the term **ambiguity**. One confusing issue is the fact that most magnetometers used measure the total field of the earth: no oriented system is recorded for the total field amplitude. The consequence of this fact is that only the component of an anomalous field in the direction of earth's main field is measured. Anomalous fields that are nearly perpendicular to the earth's field are undetectable. As for gravity, **direct and indirect interpretation** is available.

Comparison with gravity

Despite evident similarities in the two techniques, several very important differences can be pointed out:

- Magnetic properties of the rocks disappear at about 20 to 40 km depth (due to Curie temperature)
- Variations of magnetic permeability range over several orders of magnitude, density over only a range of 20-30%
- Density is a scalar, intensity of magnetization is a vector
- 2:1 length-width ratio sufficient to validate 2D approximation in gravity, but 10:1 for magnetics!
- Survey faster and simpler than gravity, since no leveling required
- The magnetic anomalies are asymmetric depending on the latitude! The magnetic anomalies are more complex than the gravity anomalies

Electrical Surveying: Resistivity methods

Warning: This script is only a complement to the PowerPoint presentation. The related illustrations and mathematical expressions can be found directly in the slides.

Introduction

Surface electrical resistivity surveying is based on the principle that the distribution of electrical potential in the ground around a current-carrying electrode depends on the **electrical resistivities** distribution of the surrounding soils and rocks. The usual practice in the field is to apply an electrical direct current (DC) between two electrodes implanted in the ground and to measure the difference of potential between two additional electrodes. The current used is either direct current, commutated direct current (i.e., a square-wave alternating current) or alternating current (AC) of low frequency (typically about 20 Hz). All analysis and interpretation are done on the basis of direct currents.

Applications

In civil-engineering applications, resistivity surveys can be useful in detecting bodies of anomalous materials or in estimating the depths of bedrock surfaces. In coarse granular soils, the groundwater surface is generally marked by an abrupt change in water saturation and thus by a change of resistivity. Therefore, this technique is very used in hydrogeophysics. Generally, since the resistivity of a soil or rock is controlled primarily by the pore water conditions, there are wide ranges in resistivity for any particular soil or rock type, and resistivity values cannot be directly interpreted in terms of soil type or lithology. Commonly, however, zones of distinctive resistivity can be associated with specific soil or rock units on the basis of local outcrops or borehole information.

Electrical properties of rocks and units

All materials, including soil and rock, have an intrinsic property, called **electrical resistivity** (or simply **resistivity**), that governs the relation between the current density and the gradient of the electrical potential. Variations in the resistivity of subsurface materials, either vertically or laterally, produce variations in the relations between the applied current and the potential distribution as measured on the surface, and thereby gives information about the composition, extent, and physical properties of the subsurface materials. The various electrical geophysical techniques distinguish materials when a **contrast** exists in their electrical properties. No resistivity contrast means no resistivity anomaly!

Note that if the **resistance** (R) is expressed in ohms (Ω), the **resistivity** (ρ) has the dimension of ohm-meter (Ωm). The **conductivity** (σ) of a material is defined as the reciprocal of its resistivity and is expressed in Siemens per meters (S/m). Resistivity is thus seen to be an intrinsic property of a material, in the same sense that density and elastic moduli are intrinsic properties, and does not depend on the shape of the sample, contrary to resistance!

In most earth materials, the conduction of electric current takes place almost entirely in the water occupying the pore spaces or joint openings, since most soil- and rock-forming minerals are essentially non-conductive. Since the conduction of current in soil and rock is through the **electrolyte** (the **ions** in water carry the current) contained in the pores, resistivity is governed largely by the porosity of the material and the geometry of the pores. Pore space may be in the form of intergranular voids, joint or fracture openings, and closed pores, such as bubbles or vugs (in lavas). Only the interconnected pores (effective porosity) effectively contribute to conductivity, and the geometry of the interconnections, or the tortuosity of current pathways, further affects it. The resistivity of a saturated porous material can be linked to the resistivity of the pore water using the formation factor in the empirical **Archie's Law**. The formation factor is a function only of the properties of the porous medium, primarily the porosity and pore geometry. Another parameter that influences the resistivity of a rock is the saturation. Moreover, as water forms a conductive electrolyte with the presence of chemical salts in solution, the conductivity is proportional to the salinity. Finally, the effect of increasing temperature is to increase the conductivity of the electrolyte because the viscosity of the fluid decreases. There is no simple link between resistivity and permeability.

Bodies of **clay** or shale generally have lower resistivity than soils or rocks composed of bulky mineral grains. Although the clay particles themselves are nonconductive when dry, the conductivity of pore water in clays is increased by the desorption of exchangeable cations from the clay particle surfaces. Clays and a few other minerals, notably magnetite, carbon, pyrite, and other metallic sulfides, may be found in sufficient concentration in the soil or rock to make it conductive. In massive metallic ores, when the metallic grains are connected, the current flows through the **electrons** contained in the metal.

The ranges of values for resistivity is very large. The values given in the slides are only informative: the particular conditions of a site may change the resistivity ranges. For example, coarse sand or gravel, if it is dry, may have a resistivity like that of igneous rocks, while a layer of weathered rock may be more conductive than the soil overlying it. In any attempt to interpret resistivities in terms of soil types or lithology, consideration should be given to the various factors that affect resistivity, and in-situ measurements on samples, boreholes or outcrops should be done (parametric measurements). It is the enormous variation in electrical resistivity found in different rocks and minerals which makes resistivity techniques attractive.

Basic theory

Consider a single point electrode, located on the surface of a semi-infinite, electrically homogeneous medium, which represents a homogeneous earth. The **equipotential surfaces** represent shells surrounding the current electrodes, and on any one of which the electrical potential is everywhere equal. The current lines represent a sampling of the infinitely many paths followed by the **current**, paths that are defined by the condition that they must be everywhere normal to the equipotential surfaces. The effect of an electrode pair (or any other combination) can be found by superposition, i.e. the added effect of individual current electrodes gives the final value for the potential field. In addition to two current electrodes, another pair of electrode is also used between which the potential difference ΔV may be measured. The potential field decreases rapidly far from the electrodes. The current and potential electrodes can be interchanged without affecting the results. This property is called **reciprocity**. Different types of **electrode arrays** or **spreads** are commonly used (e.g. Schlumberger, Wenner, or dipole-dipole).

The resistivity of the medium can be found from measured values of V (in Volts), the current I (in Amperes), and the geometric factor K (in m). K is a function only of the geometry of the electrode arrangement and the geometry of the investigated structure (e.g. a half space for measurements collected on the earth surface). Wherever these measurements are made over a real heterogeneous earth, the data from resistivity surveys are interpreted in the form of values of apparent resistivity. **Apparent resistivity** is defined as the resistivity of an equivalent electrically homogeneous and isotropic half-space that would yield the potential measured on the heterogeneous earth using the same applied current for the same arrangement and spacing of electrodes. The apparent resistivity is equal to the true resistivity of the ground only if the earth is homogeneous. The resistivity surveying problem is then the use of apparent resistivity values from field observations at various locations and with various electrode configurations to estimate the true resistivities of the earth materials present at a site.

Depth of investigation

For a same electrode spacing and for a two-layer earth, the current mainly flows in the first layer if this layer is more conductive than the basement, or in the basement if it is more conductive than the surface layer. Moreover, for small electrode spacings, the apparent resistivity is close to the surface layer resistivity, while at large electrode spacings, it approaches the resistivity of the basement layer. But this asymptotic behaviour of variations in apparent resistivity is different if the basement is more conductive or more resistive than the first layer. There is therefore no simple relationship between the electrode spacing and the depths to the interfaces between layers and the depth of investigation depends on the resistivity contrast. Typically, a maximum electrode spacing of three or more times (sometimes 10) the depth of interest is necessary to assure that sufficient data have been obtained, depending on the resistivity contrast in the ground.

Instruments and measurements

The theory and field methods used for resistivity surveys are based on the use of direct current, because it allows greater depth of investigation than alternating current (large **skin depth**) and because it avoids the complexities caused by effects of ground inductance and capacitance and resulting frequency dependence of resistivity. However, in practice, actual direct current is not frequently used for two reasons. First direct current electrodes produce **polarized ionization fields** in the electrolytes around them, and these fields produce additional electromotive forces that cause the current and potentials in the ground to be different from those in the electrodes. Second **natural earth currents** (**telluric currents**) and spontaneous potentials, which are essentially unidirectional or slowly time-varying, induce potentials in addition to those caused by the applied current. The effects of these phenomena, are reduced by the use of alternating current.

Current injection

In concept, a direct current, or an alternating current of low frequency, is applied to the current electrodes, and the current is measured with an ammeter. Current electrodes used are generally stainless steel stakes. They must be driven into the ground far enough to make good electrical contact. If the contact is bad and the injected current is too small, the quality of measurements is degraded (sensitive to noise). One common difficulty is the high **contact resistance** between current electrodes and soil. It can sometimes be alleviated by pouring salt water around the current electrodes or adding electrodes in parallel. However, if the problem is due to a combination of high earth resistivity and large electrode spacing, the remedy is to increase the input voltage across the current electrodes. Power is usually supplied by dry cell batteries in series in the smaller instruments and motor generators in the larger instruments. From 90 V up to several hundred volts may be used across the current electrodes in surveys for engineering purposes. On the current electrodes, also, the actual value of contact resistance does not affect the measurement, so long as it is small enough that a satisfactory current is obtained, and so long as there is no huge difference between the two electrodes. Contact resistance affects the relationship between the current and the potentials on the electrodes, but because only the measured value of current is used, the potentials on these electrodes do not figure in the theory or interpretation. Typical currents in instruments used for engineering applications range from 2 mA to 500 mA.

Potential measurement

Independently, a potential difference ΔV is measured across the potential electrodes, and ideally there should be no current flowing between the potential electrodes. This is accomplished with a very high input impedance operational amplifier. One advantage of the four-electrode method is that measurements are not sensitive to contact resistance at the potential electrodes so long as it is low enough that a measurement can be made, because observations are made with the system adjusted so that there is no current in the potential electrodes. With zero current, the actual value of contact resistance is immaterial, since it does not affect the potential.

External influences on measurements

Telluric currents are naturally occurring electric fields that are widespread, some being of global scale. They are usually of small magnitude, but may be very large during solar flares or if supplemented by currents of artificial origin. Spontaneous potentials in the earth may be generated by galvanic phenomena around electrochemically active materials, such as pipes, conduits, buried scrap materials, cinders, and ore deposits. They may also occur as streaming potentials generated by groundwater movement. Electric fields associated with groundwater movement will have the greatest amplitude where groundwater flow rates are high, such as through subsurface open-channel flow. The effect of telluric currents can be cancelled in applying a polarity reversing switch to make readings with alternately reversed current directions in the current electrodes. The average values of V and I for the forward and reverse current directions are then used to compute the apparent resistivity. This strategy can be also used to eliminate the effects of potential **electrode polarization** because the polarized ionization fields do not have sufficient time to develop in a half-cycle, and the alternating component of the response can be measured independently of any superimposed direct currents. The frequencies used are very low, typically below 20 Hz, so that the measured resistivity is essentially the same as the direct current resistivity. An alternative technique is to use nonpolarizing electrode to measure the potential (see lecture on Spontaneous Potential).

Resistivity measurements can also be affected by metallic fences, rails, pipes, or other conductors, which may provide short-circuit paths for the current. The effects of such linear conductors can be minimized, but not eliminated, by laying out the electrode array on a line perpendicular to the conductor. Also, electrical noise from power lines, cables, or other sources may interfere with measurements. Rejection filters for defined frequencies (16-20 Hz, 50-60 Hz) are now common in modern instruments. Sometimes electrical noise comes from temporary sources, so better measurements can be obtained by waiting until conditions improve (during the night). Modern resistivity instruments have capability for data averaging or stacking; this allows resistivity surveys to proceed in spite of noisy site conditions and to improve signal-to-noise ratio for weak signals.

Survey strategies and interpretation

An electrode array with constant spacing is used to investigate lateral changes in apparent resistivity reflecting lateral geologic variability (**Constant Separation Traversing** (CST) or **resistivity mapping** or **profiling**). To investigate changes in resistivity with depth, the size of the electrode array is varied (**Vertical Electrical Sounding**, VES). The apparent resistivity is affected by material at increasingly greater depths as the electrode spacing is increased. Because of this effect, a plot of apparent resistivity against electrode spacing can be used to indicate vertical variations in resistivity. The combination of CST and VES gives a high resolution technique (**Electrical Resistivity Imaging** or **Tomography**, ERT)), that is able to image 2D or 3D structures in the subsurface.

Constant Separation Traversing

Surveys of lateral variations in resistivity can be useful for the investigation of any geological features that can be expected to offer resistivity contrasts with their surroundings. Steeply dipping faults may be located by resistivity traverses crossing the suspected fault line for example. Resistivity mapping geology is made with a fixed electrode spacing, by moving the array between successive measurements along a line of traverse or a grid. If a symmetrical array, such as the Schlumberger or Wenner array, is used, the resistivity value obtained is associated with the location of the center of the array. To plan a mapping survey, any available a priori geological information, such as the depth of the features of interest, should be considered in making this decision, which governs the effective depth of investigation. One way to help plan the survey is to construct model VES sounding curves (see below) for the expected models, vary each model parameter separately and then choose electrode separations that will best resolve the expected resistivity/depth variations. The spacing of adjacent resistivity stations, or the fineness of the grid, governs the resolution of detail that may be obtained. As a general rule, the spacing between resistivity stations should be smaller than the width of the smallest feature to be detected. Mobiles arrays on automated vehicles can be used for a rapid near-surface mapping (in agricultural or archaeological studies for example).

Data obtained from horizontal profiling are normally **interpreted qualitatively**. Apparent resistivity values are plotted and contoured on maps, or plotted as profiles, and areas displaying anomalously high or low values, or anomalous patterns, are identified. Interpretation of the data, as well as the planning of the survey, must be guided by the available knowledge of the local geology. The interpreter normally knows what he is looking for in terms of geological features and their expected influence on apparent resistivity. The construction of theoretical profiles is feasible for certain kinds of idealized models (e.g. faults, dikes, filled sinks, and cavities), but the usefulness of the study of such model is limited in practice.

Vertical Electrical Sounding

In the VES technique, a 1D model where the resistivity varies only with depth, is sought. The Schlumberger array is commonly used for sounding. In this method, the center point of the array is kept at a fixed location, while the electrode locations are varied around it. The apparent resistivity values, and layer depths interpreted from them, are referred to the center point. Minimum and maximum spacings to use are governed by the need to define the needed depth of investigation. Frequently, the maximum useful electrode spacing is limited by available time, site topography, or lateral variations in resistivity. This vertical VES curve is plotted on a log-log graphic of the array length versus apparent resistivity. The apparent resistivity curve should be plotted as the survey progresses in order to judge whether sufficient data have been obtained. Also, the progressive plot can be used to detect errors in readings or spurious resistivity values due to local effects.

Data collection strategy is as follows: an initial spacing (the distance from the center of the array to either of the current electrodes) is chosen, and the current electrodes are moved outward with the potential electrodes fixed. As the distance between the current electrodes is increased, the sensitivity of the potential measurement decreases. Therefore, at some point it will be necessary to increase the potential electrode spacing. Changing the spacing of the potential electrodes may produce an offset in the apparent resistivity curve as a result of lateral inhomogeneity or changes in the depth of investigation. Under such conditions, the cause of the offset can often be determined by repeating portions of the sounding with different potential electrode spacing.

The plot of apparent resistivity versus spacing is always a smooth curve where it is governed only by vertical variation in resistivity. Reversals in resistance and irregularities in the apparent resistivity curve, if not due to measurements errors, both indicate lateral changes and should be further investigated. Comparison with theoretical multilayer curves is helpful in detecting such distortion. Excessive dip of subsurface strata along the survey line (more than about 10%), unfavourable topography, or known high lateral variability in soil or rock properties may be reasons to reject field data as unsuitable for interpretation in terms of simple vertical variation of resistivity.

The **interpretation** of VES data (**inversion**) is to use the curve of apparent resistivity versus electrode spacing, plotted from field measurements, to obtain the parameters of the geoelectrical section, i.e. the layer resistivities and thicknesses. From a given set of layer parameters, it is always possible to compute the apparent resistivity as a function of electrode spacing (the VES curve). This represents the forward problem. But unfortunately, for the inverse problem, it is not generally possible to obtain a unique solution. To deal with the problem of **ambiguity**, the interpreter should check all possible interpretations in a certain error range by computing the theoretical VES curve for the interpreted section and comparing it with the field curve. The test of geological reasonableness should also be applied and information from direct measurement on outcrops (**parametric sounding**) should be included.

The simplest multilayer case is that of a single layer of finite thickness overlying a homogeneous halfspace of different resistivity. The VES curves for this two-layer case vary in a relatively simple way, and a complete set of reference curves (master curves) can be plotted on a single sheet of paper. The curves are plotted on a logarithmic scale, both horizontally and vertically, and are normalized by plotting the ratio of apparent resistivity to the first layer resistivity against the ratio of electrode spacing to the first layer thickness. Each curve of the family represents one value of resistivity contrast between the two layers $k = (\rho2-\rho1)/(\rho2+\rho1)$.

Because the apparent resistivity for small electrode spacings approaches the resistivity of $\rho1$ and for large spacings approaches the resistivity of $\rho2$, these curves begin at $\rho a/\rho1 = 1$, and asymptotically approach $\rho a/\rho1 = \rho2/\rho1$. Any two-layer curve for a particular value of k, or for a particular ratio of layer resistivities, must have the same shape on the logarithmic plot as the corresponding standard curve. It differs only by horizontal and vertical shifts, which are equal to the logarithms of the thickness and resistivity of the first layer.

Where three or more strata of contrasting resistivity are present, the VES curves are more complex than the two-layer curves. For three layers, there are four possible types of VES curves, depending on the nature of the successive resistivity contrasts. The classification of these curves is found in the literature with the notations H, K, A, and Q. These symbols correspond respectively to bowl-type curves, which occur with an intermediate layer of lower resistivity than layers 1 or 3; bell-type curves, where the intermediate layer is of higher resistivity; ascending curves, where resistivities successively increase; and descending curves, where resistivities successively decrease. With these methods, the use of standard curves is cumbersome and interpretation of VES curves can be carried out with a computer. Thus, trial-and-error interpretation of VES data is feasible. Trial values of the layer parameters (resistivities and thicknesses) can be guessed, checked with a computed apparent resistivity curve, and adjusted to make the field and computed curves agree. Several commercial software allow for the use of this method to obtain the layer parameters automatically by iterative automatic inversion, starting with an initial estimate obtained by an approximate method or given by the user (based on a priori information, e.g. parametric sounding).

Two problems are frequent in the interpretation of resistivity soundings. The **principle of equivalence** concerns a layer for which the resistivity is either greater than, or less than, both the layers above and below (sounding types K and H). A conductive layer between two resistive beds (type H) is essentially defined by its "longitudinal conductance" i.e. the thickness of the layer divided by its resistivity (h/ρ). As long as this ratio stays the same the curve is unchanged. The limits of validity of this principle depend on the characteristics of the whole suit of beds. A resistant layer, between two or more conductive beds (type K) is essentially defined by its "transverse resistance" ($h*\rho$) i.e. the product of its resistivity and its thickness. As long as this product remains constant and in some limits, the electrical sounding curve remains unchanged. The solution to this problem is to get additional a priori information either on the resistivity or the thickness of the layer. The **principle of suppression** is also very significant during the interpretation of the electrical soundings. This principle relates to layers whose resistivities are intermediate between that of the enclosing beds (sounding types A and Q). Such layers have practically no influence on the VES curve as long as they do not have a large enough thickness and remains invisible!

Electrical Resistivity Tomography

A two dimensional imaging of the ground (ERT)can be seen as a combination of the vertical electrical sounding and lateral profiling. With help of 2D images, changes of resistivity can not only be detected in the vertical direction but also in the horizontal. Two dimensional imaging are usually carried out using a large number of electrodes which are typically set up along a straight line. In this case it is assumed that resistivity does not change in the direction that is perpendicular to the survey line (2D models). Normally a constant spacing (named electrode spacing) between the electrodes is used. The electrodes are linked by a multi-core cable and connected to a switching unit. The whole survey can be controlled with a laptop computer where the user can program a sequence of resistivity measurements. This sequence is based on a series (several hundreds) of CST measurements along the profile with different electrode spacings and based on some predefined electrode arrangements (e.g. Wenner, Schlumberger). With increasing the distance between the electrodes used, the depth of investigation is also increasing.

When the survey starts, the laptop computer automatically selects the appropriate electrodes for each measurement in the programmed sequence. The measured apparent resistivities are recorded in the laptop memory and often plotted in a **pseudosection** form. In a pseudosection, each point is plotted at a distance along the profile that corresponds to the middle of the array and to a depth proportional to the electrode spacing. This pseudo section is only a convenient way of plotting the data. This is not an image of the true subsurface resistivities!

Once the data are collected, a special program is used to solve the **inverse problem** (i.e. getting an image of the subsurface resistivities from the pseudosection).

For 3D resistivity, the same approach is used but with some differences. First a grid of electrodes is used to investigate changes in the resistivity properties in all directions. Second electrodes arrays such as pole-pole or pole-dipole are preferred. Finally this type of survey is generally more time consuming than 2D survey and special 3D resistivity inversion software are used.

Limitations in interpretation

Apart the problem of electrical noise (telluric currents) or short-cut from buried pipes (see above), the electrical resistivity method has some inherent limitations that affect the resolution and accuracy that may be expected from it. Like all methods using measurements of a potential field (gravity), the value of a measurement obtained at any location represents a weighted average of the effects produced over a large volume of material, with the nearby portions contributing most heavily. The collected information does not lend themselves to high resolution for interpretations. There is another feature common to all potential field geophysical methods: a particular distribution of potential at the ground surface does not generally have a unique interpretation. While these limitations should be recognized, the non-uniqueness or ambiguity of the resistivity method is scarcely less than with the other geophysical methods (e.g. gravity or magnetic) since we have a direct control on the source in that case. For these reasons, it is always advisable to use several complementary geophysical methods in an integrated exploration program rather than relying on a single exploration method.

Electrical Surveying: Induced Polarization method

Warning: This script is only a complement to the PowerPoint presentation. The related illustrations and mathematical expressions can be found directly in the slides.

Introduction

In certain conventional resistivity surveys, we can note that the potential difference, measured between the potential electrodes did not drop instantaneously to zero when the current is turned off. Instead, the potential difference drops sharply at first, then gradually decays to zero after a given interval of time. This means that certain bodies in the ground can become electrically polarized, forming a battery when energized with an electric current. Upon turning off the polarizing current, the ground gradually discharges and returns to equilibrium. This phenomenon is the foundation of a geophysical survey technique called Induced Polarization (IP).

Applications

The IP method is widely used in exploration for ore bodies, principally of disseminated sulfides. The IP method is also used in groundwater exploration, to discriminate between salt water layers and clay, and for the study of pollutants in the ground.

Basic theory of the IP effect and units

The origin of IP phenomena is complex because numerous physiochemical process and conditions are likely responsible for its occurrence. Only a simple description will be given here. The IP phenomena can be of two main origins. The first origin can be found in metallic ore bodies. When a current passes across a metal grain (electronic conductor) dipped in an electrolyte, charges can pile up continuously at the interface when all the processes in the electrolytic reaction are not equally rapid. This produces the **electrode polarization**. The extra piled-up charge diffuses back into the electrolyte when the current is stopped. The time constant of buildup and decay is typically several tenths of a second. The second origin, and the most important source of nonmetallic IP in rocks, can be found in certain types of clay minerals. The surfaces of clay particles, the edges of fibrous materials normally have unbalanced negative charges that attract a cloud of positive ions from the surrounding electrolyte. When an electric current is forced trough a clay-electrolyte system, positive ions can readily pass trough this cloud but negative ions are blocked, forming zones of ion concentration. The return to the equilibrium distribution after the current is switched off constitutes a residual current that appears as IP effect. Discriminating between the two origins of IP phenomenon (electrode or membrane polarization) is generally not possible in the field.

Two different techniques of IP surveys can be envisaged. The first technique is the study of the decaying potential difference as a function of time and is known as **time-domain IP**. In this method the geophysicist looks for portions of the earth where current flow due to polarization is maintained for a short time after the applied current is terminated. In time-domain IP, the polarizability of the medium is defined by the **Chargeability** (in seconds) that is the ratio of the area under the decay curve (in millivolt-seconds) to the potential difference (in millivolts) measured before switching off the current. The second technique is to study the effect of alternating currents on the measured value of resistivity, which is called **frequency-domain IP**. The geophysicist tries to locate portions of the earth where resistivity decreases as the frequency of applied current is increased. In frequency-domain IP, we defined the **Frequency Effect** (in %)

or the **Metal Factor** (in Siemens per meters, S/m) to compare resistivity collected at one low frequency (0.05-0.5 Hz) to the one collected at one higher frequency (1-10 Hz). When a whole range of frequencies is investigated (from 10^{-2} to 10^{4} Hz), we talk about **Spectral Induced Polarization** (SIP) and the shift between the injected current and measured potential is recorded. SIP data are **amplitudes** (amperes and volts) and **phase shift** (milliradians).

IP effect of rocks

Typical values of IP for rocks are not easy to give since these values depend on the current frequency used. We can nevertheless say that the IP effect is higher for disseminated than massive clay and metallic particles, depends on the concentration of clay and metallic particles, increases if water in the ground has a low conductivity, increases with decreasing porosity and varies with the amount of water in the ground. The IP effect is large for clay and almost all the metallic sulfides (except sphalerite, because it can be very resistive!) such as pyrite, graphite, some coals, magnetite, pyrolusite, native metals and other minerals with a metallic luster, if they are **disseminated** (the IP effect is a surface phenomenon!). Pure quartz or water show no IP effect!

Instruments and measurements

The same electrode arrays than in conventional resistivity are used with a preference for the dipole-dipole array, to decrease electromagnetic coupling between adjacent wires. Nevertheless, different measurement recording devices are used for time-domain IP and frequency-domain IP. Frequency-domain IP is difficult to use in the field and necessitates specific instrumentation. Oppositely, modern conventional resistivimeters frequently include also measurement in time-domain IP. Stainless steel electrodes are commonly used for IP surveys but stability of potential measurements can be a problem (non polarizable electrodes can be sometimes used, see lecture on SP surveying).

Survey strategies and interpretation

The techniques of **sounding** and **profiling** (mapping), used in resistivity measurements, are also employed for the IP method. The apparent chargeability (with linear scale) versus the electrode spacing (with log scale) is plotted on a graph. The IP sounding curve can be interpreted using a computer in the same way than resistivity. 2D IP imaging (**tomography**) can be also collected in the same manner than 2D resistivity imaging techniques. Again, it must be emphasized that the apparent IP pseudosection is not an image of the IP properties of the subsurface. As for resistivity, the IP data must be **inverted** using specific algorithms to get a reliable image of the ground. The limitations are the same than for conventional resistivity surveys, but IP method is more sensitive to telluric and industrial electromagnetic noise.

IP information can be of significant value in complementing a resistivity survey. IP data are sometimes more effective to discriminate between various geological units than resistivity. In fact, each rock in the subsurface has a resistivity value that is influenced by several parameters such as the water content or water resistivity. Oppositely, only sediments containing clay, graphite or metallic particles have an IP signal. This binary behaviour (i.e. either an IP signal or not) can be very fruitful when a specific target is searched (e.g. a graphite vein). For SIP, more advanced models (**Cole-Cole models**) can provide information on material structures (e.g. size of pores) from the spectral resistivities.

Electrical Surveying: Self Potential method

Warning: This script is only a complement to the PowerPoint presentation. The related illustrations and mathematical expressions can be found directly in the slides.

Introduction

Natural potentials (or **self potentials**, SP) can occur in the earth without any artificial (i.e. man-made) source. These natural potentials are considered as "noise" in conventional resistivity surveys but can also be used to determine information about the subsurface.

Application

Some engineering and environmental occurrences may be mapped by contouring surficial voltages of natural origin between a base (or reference) electrode and a mobile electrode. Sulfide ore bodies have been sought by SP generated by ore bodies acting as batteries. Flow of gasses and fluids in pipes, leakage of a reservoir within the foundation or abutment of a dam, movement of ionic fluids to or within the groundwater, flow of geothermal fluids, and movement of water into or through a karst system can be the origin of natural potentials. These potentials may exceed the background voltage variation of a site, creating significant anomalies.

Basic theory of SP effects and units

Two main natural potential are of interests in surface engineering geophysics. The first natural potential is the **electrokinetic**, or **streaming potential**. It is due to the flow of a fluid with certain electrical properties (conductivity, viscosity) passing through a pipe or porous medium with different electrical properties. The movement of the fluid can be of gravity origin for example. The relation between the pressure difference between the extremities of the pipe and the generated voltage is expressed with the coupling coefficient that varies with temperature, grain size etc... The SP anomaly is positive where the hydrostatic pressure decreases (i.e. in the direction of the water flux). The second natural potential of interest is the **mineralization**, or **electrolytic contact potential**. It is produced at the surface of a conductor (massive ore body) with another medium. A general theory (but this is still not really clear…) for this natural potential involves an oxidation-reduction process in connection with the water table and the metallic conductor. This dependence to the water table limits the depth of investigation of the mineralization potential survey to the depth of the water table. The measured potentials are expressed in **Volts** or millivolts (mV).

Instruments and measurements

A simple SP survey consists of a base electrode and a roving electrode to determine potential differences on a grid or along profile lines. The required equipment merely includes electrodes, wire and a precise millivolt meter.

The electrodes in contact with the ground surface should be **nonpolarizing electrodes**, also called porous pots. Porous pots are metal electrodes suspended in a supersaturated solution of their own salts (such as a copper electrode suspended in copper sulfate) within a porous container (often made of porcelain). These pots produce very low electrolytic contact potential, such that the background voltage is as small as possible. The porous pots are sealed to avoid evaporation of the salt solution. Sealed pots can keep their supersaturated solutions for more than a week, even in arid regions. Refilling of the pots' solution must occur before a day's work due to the possible contact potential change while performing a measurement set. It is important that all pots contain the same solution mix. The **wire** used in SP surveys must be strong, abrasion resistant and of low resistance. Small section copper wire with an insulated cover has been found to be suitable. A **high-input-impedance voltmeter** (possibly digital) is used to read the potential in the millivolt range. The input impedance of voltmeters should exceed 50 MΩ. Higher input impedances are even desirable due to the impedance reduction of air's moisture. Filters about 60 Hz will reduce stray alternating current (AC) potentials in industrial areas or near power lines. The resolution of the meter should be 0.1 or 1.0 mV because background potentials for these surveys may be at a level of a few tens of millivolts. Note that potentials exceeding 1.0 V have occurred for ore bodies. To be useful interpreted, self-potential anomalies must exceed the background to be apparent.

Survey strategies and interpretation

In a SP survey, the potential are measured along profiles or a grid relatively to a **base electrode**. Previously measured locations may need to be remeasured on a systematic or periodic basis. Reoccupation of stations is necessary when very accurate surveys are being conducted and for sites with temporal potential changes or spatial variations of electrode potential. Electrodes may have contact differences due to varying soil types, chemical variations, or soil moisture. To measure SP data, a contact with moist soil must be assured. Power lines, metal fences, and underground utilities are cultural features that affect the potential field extraneous to the normal sources of interest. Variation in the flow of fluid due to rainfall, reservoir elevation changes, channelization of flow, or change of surface elevation are also sources of variation of streaming potential. Self potentials may have temporal or spatial changes due to thunderstorm. High telluric potential variations can make an SP survey impossible!

Most SP investigations use a **qualitative interpretation** of the profile amplitudes or grid contours to evaluate SP anomalies. Flow sources produce potentials in the direction of flow: fluid inflow produces negative relative potentials; outflow of the fluid results in positive potentials. Interpretations for a dam embankment with possible underseepage would be determined from the profiles across the crest. Negative anomalies may be indicative of flow from the reservoir at some depth. Outflow at the toe of an embankment or at shallow depths beneath the toe would produce positive anomalies. Mineral or cultural utilities produce varying surface potentials depending on the source. Semiquantitative forward solutions may be estimated by analytical equations for sphere, line, and plate potential configurations. These solutions aid in interpretation of the field readings, but their use is limited in practice. A common way for a more effective modeling of SP effects is simulating the source using a series of dipoles in the ground.

ELECTROMAGNETIC SURVEYING

Warning: This script is only a complement to the PowerPoint presentation. The related illustrations and mathematical expressions can be found directly in the slides.

Introduction

Electromagnetic (EM) surveying methods make use of the response of the ground to the propagation of **electromagnetic fields**. This response varies according to the electrical **conductivity** of the ground.

Applications

EM techniques are used in exploration of massive ore bodies and exploration for fossil fuels (oil, gas, coal), because this technique has a large depth of penetration). Note that if ore bodies are disseminated, the IP technique is certainly more suitable! In near-surface geophysics, EM is used in engineering/construction site investigations, glaciology or permafrost studies and also in archaeological surveys. The advantages of EM over DC resistivity are significant, including improved speed of operation, improved lateral resolution, improved resolution of conductive bodies and no problems injecting current into a resistive surface layer. The disadvantages are that the EM method does not work well in very resistive material, only basic interpretation schemes are available and equipment tends to be somewhat more costly due to its greater complexity. Moreover, EM methods are more sensitive to electromagnetic noise than resistivity methods.

Conductivity of rocks

Conductivity (in Siemens per meter, **S/m**) is the inverse of resistivity. The electrical properties of rocks follow the same description as given in the lecture on Resistivity techniques.

The electromagnetic induction process

An EM transmitter outputs a time-varying electric current into a transmitter coil. The current in the transmitter coil generates a primary magnetic field of the same frequency and phase (**Ampere's law**). Lines of force of this primary magnetic field penetrate the earth and may penetrate a conductive body. When this occurs, an electromotive force (EMF) or voltage is set up within the conductor, according to **Faraday's Law**. Current will flow in the conductor in response to the induced electromotive force. These currents will usually flow through the conductor in planes perpendicular to lines of magnetic field of force from the transmitter and are restricted by the conductor's geometry. Current flow within the conductor generates a secondary magnetic field whose lines of force, at the conductor, are such that they oppose those of the primary magnetic field. The receiver coil, at some distance from the transmitter coil, is therefore energized by two fields: the primary field from the transmitter and the secondary field from the induced currents in the ground. The EMF induced by these two fields in the receiver can be measured and further analysed (see next paragraph). Note that the induced currents occur throughout the subsurface, and that the magnitude and distribution are functions of the transmitter frequency, power, and geometry as well as the distribution of all electrical properties of the subsurface. Note that the primary field comes not necessarily from a transmitter coil but can be of natural origin (see below).

Phase's shifts

To describe the resulting field, **in phase** (or **real**) and **out-of-phase** (or **imaginary** or **quadrature**) components can be used. Compared to the primary magnetic field, the EMF induction creates a first phase lag of $\pi/2$ for the secondary magnetic field. Moreover, the target creates an additional phase lag, depending of its conductivity. If the conductivity is very high, the imaginary component is small and the total phase lag is π. If the conductivity is very low, the real component is small and the total phase lag is $\pi/2$. Therefore, the analysis of the real and imaginary components gives information on the conductivity of the subsurface. Plots of real and imaginary amplitudes can be done to interpret the conductive/resistive nature of a target.

To electronically separate the two components, some kind of phase link between transmitter and receiver to establish a time or phase reference is required. This is commonly done with a direct cable link, sometimes with a radio link. But note that the simpler frequency domain EM systems are **tilt angle systems** which have no reference link between the transmitter and receiver coils. The receiver simply measures the total field irrespective of phase, and the receiver coil is tilted to find the direction of maximum or minimum magnetic field strength. In fact, if the receiver coil is parallel to the magnetic field, no induction in the coil is measured and the orientation (tilt) of the magnetic field can be deduced. With tilt angle systems, therefore, the objective is to measure deviations from the normal infield direction and to interpret these in terms of geological conductors.

Penetration of electromagnetic waves in the ground

For the investigation of the subsurface, we need to describe how deep the EM field penetrates in the ground. We define the **skin depth** as the depth at which the amplitude of the field reaches 1/e of its original value at the source. This depth increases with decreasing conductivity and decreasing frequencies. The **depth of penetration** is the maximum depth at which a conductor may still produce a recognizable EM anomaly. This empirical relationship is usually a multiple of the skin depth.

Measurement strategies and interpretation

We can define three main classes of EM methods: the uniform field methods, the dipolar field methods and the time-domain EM.

Uniform field methods

Natural or artificial fields can be produced at a large distance (hundreds of kilometers) from the receiver and can be seen as being uniform. The induced currents produce secondary magnetic fields which can be detected at the surface through deviation of the normal radiated field. The different techniques often measure the components of both magnetic and electrical fields. Magnetic fields are measured with coils whereas electrical fields are measured with dipoles. Natural field have the disadvantage that we have a poor control on the source direction and intensity of the primary field.

The **Magneto-telluric** (MT) technique is the only electrical technique capable of penetrating to the depths of interest to the oil industry (mapping salt domes and anticlines). The source are fields of natural origin (magneto-telluric fields) resulting from flows of charged particles in the ionosphere. This flow is correlated with diurnal variations in the geomagnetic field caused by solar emissions. Two components of the electric field and three components of the magnetic field are measured. Frequencies range from 10^{-5} Hz to 20 kHz.

The **Audio Magneto Telluric** (AMT) method uses equatorial thunderstorms as sources (1 to 20 kHz). These natural EM fields are called sferics. Sferics propagated around the earth between the ground and the ionosphere. The very broad frequency spectrum available can be filtered to select a desired frequency and obtain a depth of penetration up to 1 km. AMT soundings can be carried out in selecting different frequency ranges. Unfortunately, it is more and more difficult to use this technique in our regions because of the increasing electromagnetic noise in urban areas.

The **Very Low-Frequency Tilt** (VLF-tilt) method uses powerful remote radio transmitters set up in different parts of the world for military communications (mainly for submarines). VLF means very low-frequency, about 10 to 30 kHz, but relative to frequencies generally used in geophysical exploration, these are actually high frequencies! These radio transmitters are very powerful and induce electric currents in conductive bodies thousands of kilometers away. VLF response is maximal when the target strike is in the direction of the transmitter. These can be sometime a limitation but as there are a number of transmitters worldwide, the selection of an appropriate transmitter is seldom a problem. Because of the rudimentary nature of VLF measurements, simple interpretational techniques suffice for most practical purposes. The conductor is located horizontally at the inflection point marking the crossover from positive tilt to negative tilt and the maximum in field strength. A variant is the **VLF-R** method that measures the electric field with a pair of electrodes simultaneously with the magnetic field. Frequencies can be chosen among different local radio transmitters to evaluate apparent resistivity values.

The **Controlled Source Audio Magneto Telluric** (CSAMT) is similar to MT but using a remote (2 to 8 km) electrical dipole as source (1 Hz to 10 kHz) and a series of electrical dipoles and magnetic coils as receivers. The advantage is that the source frequency and location is known (artificial source). This allows for a better understanding of the geological features in the ground.

Dipolar field methods

In this kind of survey, one person carries a small transmitter coil, while a second person carries a second coil which receives the primary and secondary magnetic fields. Perhaps the most important is that the operating frequency is low enough at each of the intercoil spacings that the electrical skin depth in the ground is always significantly greater than the intercoil spacing. Under this condition (known as "**operating at low induction numbers**", or **LIN**), the ratio of the secondary magnetic field to the primary field can be represented as a function of the bulk conductivity of the ground, if this conductivity is not too high (should be less conductive than 0.1 S/m). We call these systems "terrain conductivity meters". In principle, either intercoil spacing or frequency can be varied to determine variation of conductivity with depth. In devices such as the EM-38 (14.6 kHz), EM-31 (9.8 kHz) and EM-34(three operating frequencies: 6.4 kHz, 1.6 kHz and 0.4 kHz) systems, frequency is varied as the intercoil spacing is varied to respect the LIN condition (EM-38: 1m; EM-31: 3.6 m; EM-34: respectively 10, 20, and 40 m). Phase and other information are obtained in real time by linking the transmitter and receiver with a connecting cable. The correct intercoil spacing is adjusted by measuring the primary magnetic field with the receiver coil so that it is the correct compensation value for the appropriate distance. Terrain conductivity meters are operated in both the horizontal and vertical dipole modes. These terms describe the orientation of the transmitter and receiver coils to each other and the ground, and each mode gives a significantly different response with depth. In the vertical dipole mode, the instruments are relatively sensitive to intercoil alignment, but much less so in the horizontal dipole mode.

Terrain conductivity is displayed on the instrument in mS/m (milliSiemens per meter). Because terrain conductivity meters read directly in apparent conductivity most surveys using the instrument are done in the profile mode, with a station spacing dictated by the required resolution and time/economics consideration. Limited information about the variation of conductivity with depth can be obtained by measuring two or more coil orientations and/or intercoil separations since this changes the depth of penetration. Therefore, **1D soundings** can be carried out in using different intercoil distances and dipole orientations around the same central location. Commercial EM software allow for an interpretation of theses soundings in the same way than for vertical electrical sounding (VES). The maximum number of geoelectric parameters, such as layer thicknesses and conductivities, which can be determined, is less than the number of independent observations. This means that EM sounding have a lower resolution than VES. The most common instrument (EM-34) uses three standard intercoil distances and two intercoil orientations, which results in a maximum of six observations. Without other constraints, a two-layer model is the optimum. These soundings are commonly merged to obtain a section of the ground conductivities.

Time-domain EM

The principles of time-domain EM (TDEM) resistivity sounding are relatively easily understood. A transmitter is connected to a square loop of wire laid on the ground. The side length of the loop is generally equal to the desired depth of exploration. Then we abruptly reduce the transmitter current to zero, that induces, in accord with Faraday's law, a short-duration voltage pulse in the ground, which causes a loop of current to flow in the immediate vicinity of the transmitter wire. In fact, immediately after transmitter current is turned off, the current loop can be thought of as an image in the ground of the transmitter loop. However, because of finite ground resistivity, the amplitude of the current starts to decay immediately. This decaying current similarly induces a voltage pulse which causes more current to flow, but now at a larger distance from the transmitter loop, and also at greater depth. This deeper current flow also decays due to finite resistivity of the ground, inducing even deeper current flow and so on. The amplitude of the current flow as a function of time is measured by measuring its decaying magnetic field using a small multi-turn receiver coil usually located at the center of the transmitter loop. From the above it is evident that, by making measurement of the voltage out of the receiver coil at successively later times, measurement is made of the current flow and thus also of the conductivity of the earth at successively greater depths.

To accurately measure the decay characteristics of this voltage the receiver contains a series of narrow time windows, each opening sequentially to measure (and record) the amplitude of the decaying voltage at the successive times. Note that, to minimize distortion in measurement of the transient voltage, the early time windows, which are located where the transient voltage is changing rapidly with time, are very narrow, whereas the later windows, situated where the transient is varying more slowly, are much broader. This technique is desirable since wider windows enhance the signal-to-noise ratio, which becomes smaller as the amplitude of the transient decays at later times. For shallower sounding, where it is not necessary to measure the transient characteristics out to very late times, the period is typically of the order of 1 msec or less, which means that in a total measurement time of a few seconds, measurement can be made and stacked (repeated) on several thousand transient responses. This is important since the response from one pulse is exceedingly small and it is necessary to improve the signal-to-noise ratio by adding the responses from a large number of pulses.

Fast **computer inversion programs** allows TDEM data to be automatically inverted to a layered earth geometry in a matter of a few seconds. An inversion program offers an additional significant advantage. All electrical sounding techniques (VES, TDEM) suffer to a greater or less extent from **equivalence**, which states that, to within a given signal-to-noise ratio in the measured data, more than one specific geoelectric model will fit the measured data. This problem can be addressed by inverting different types of data together (i.e. VES + TDEM), since these techniques are prone to different types of equivalences. This is called **joint inversion**. The results of joint inversion are much more accurate than the results of individual inversions.