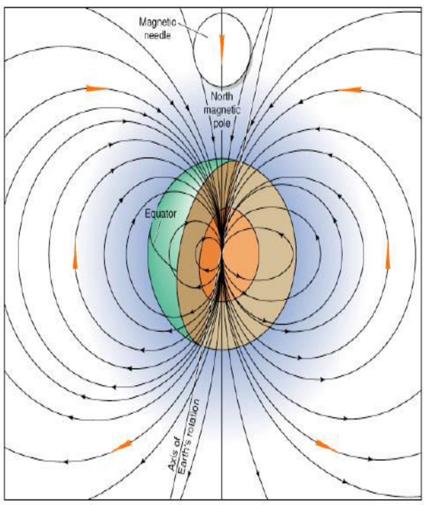
Ministry of Higher Education and Scientific Research

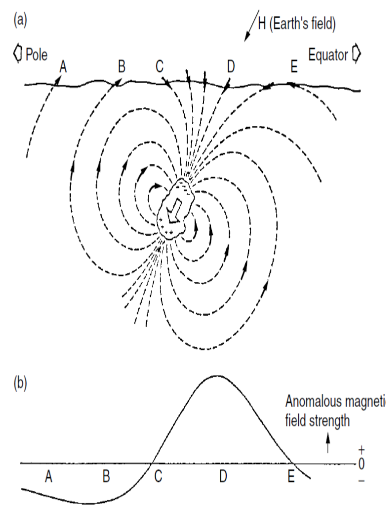
Al-Karkh University of Science

College of Geophysics and Remote Sensing

Department of Geophysics

Magnetic Method





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PhD. In Geophysics

Introduction

Compasses and dip needles were used in the middle Ages to find magnetite ores in Sweden, making the magnetic method the oldest of all applied geophysical techniques. It is still one of the most widely used, even though significant magnetic effects are produced by only a very small number of minerals.

Magnetic field strengths are now usually measured in *nanoTesla* (nT). The pre-SI unit, the gamma, originally defined as 10−5 gauss but numerically equal to the nT, is still often used.

In principle, magnetic surveying is similar to gravity, i.e., we are dealing with potential fields.

There are three fundamental differences, however:

• We are dealing with *vector* fields, not scalar. We cannot always assume that the magnetic field is vertical as we can for gravity.

• Magnetic poles can be *repulsive or attractive*. They are not always attractive as in gravity.

• The magnetic field is dependent on *mineralogy*, not bulk properties. Thus, what may be a trivial (to us) change in composition can have a large effect on the magnetic field.

Magnetic measurements are simple to make and reduce, but very complicated to understand and interpret. Magnetic surveying is one of the oldest methods in geophysical prospecting, but has become relegated to a method of minor importance because of the advent of seismic reflection surveying in the last few decades. In terms of line-km measured each year, however, it is the most widely-used survey method. The great problems involved in interpreting magnetic anomalies greatly limit their use.

Magnetic prospecting is used to search for oil and minerals, for archaeology research and searching for hazardous waste. The prime targets are the depth to basement (i.e., the thicknesses of sedimentary sequences), igneous bodies, kimberlite pipes, hydrothermal alteration (geothermal) and archaeology, *e.g.* fire pits, kilns and disturbed earth. Very recently, there has been a resurgence in interest in magnetic surveying because of the advent of extremely high-resolution surveys that can reveal structure in sedimentary sequences.

Theories on the origin of the Earth's field

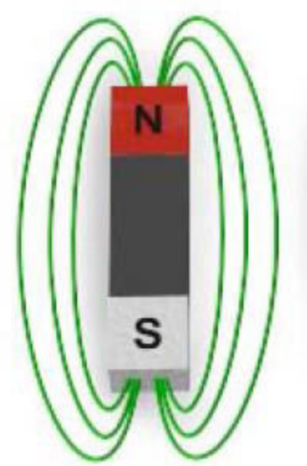
In about 1600 W. Gilbert absolved the pole star of responsibility for the Earth's magnetic field. A century later, Halley rejected magnetized surface rock as the source because the field changes with time. He suggested that there is a magnetized sphere within the Earth. Early in the 20th century, Einstein described the origin of the Earth's magnetic field as one of the fundamental unsolved problems in physics.

It is now believed that the Earth's magnetic field is generated by a kinematic fluid dynamo. Fluid flowing across field lines in the Earth's liquid outer core induces the magnetic field. That the Earth's outer core is liquid was shown by earthquake seismology, which revealed that shear waves are not transmitted by the outer core. A material that does not transmit shear waves is the definition of a fluid. Thus modern earthquake seismology provided the critical information necessary to explain this phenomenon that had puzzled scientists for four centuries.

It has been shown that the Earth's field would decay for realistic Earth properties, and this resulted in several decades of highly complex mathematical debate and the development of Physically-reasonable numerical models that predicted a sustained magnetic field. Two-scale dynamo models suggested that very small-scale fluid motions are important. These models could not be tested numerically until recently. The mathematical arguments were inconclusive and experiment was impractical. In the last few years, numerical modeling work initially suggested that a chaotic field would result. However, the addition of a solid inner core (also known for several decades from earthquake seismology) to the models stabilised the predicted field. After 40,000 years of simulated time, which took 2.5 months of Cray C-90 CPU time, a reversal occurred, which was a big breakthrough.

Principle

Near to a bar magnet, magnetic flux exists. Field lines are revealed by iron filings which will orient parallel to lines of force that follow curved paths from one pole to another. The Earth behaves as a giant magnet.



Field lines around a bar magnet

So-called *north-seeking* poles are +ve, and are south poles. Poles always occur in pairs, but Sometimes one is a very long distance from the other and can be ignored in modeling.

Some basic parameters and variables used in magnetic method are:

Magnetic force F: This is the force between two poles. It is attractive if the poles are opposite and repulsive if the poles are the same.

Intensity of induced magnetization I: If a body is placed in a magnetic field it will acquire magnetisation in the direction of the inducing field. The intensity of induced magnetisation, I, is the induced pole strength per unit area on the body placed in the external field.

Magnetic susceptibility k: This is the degree to which a body is magnetised.

Magnetic units

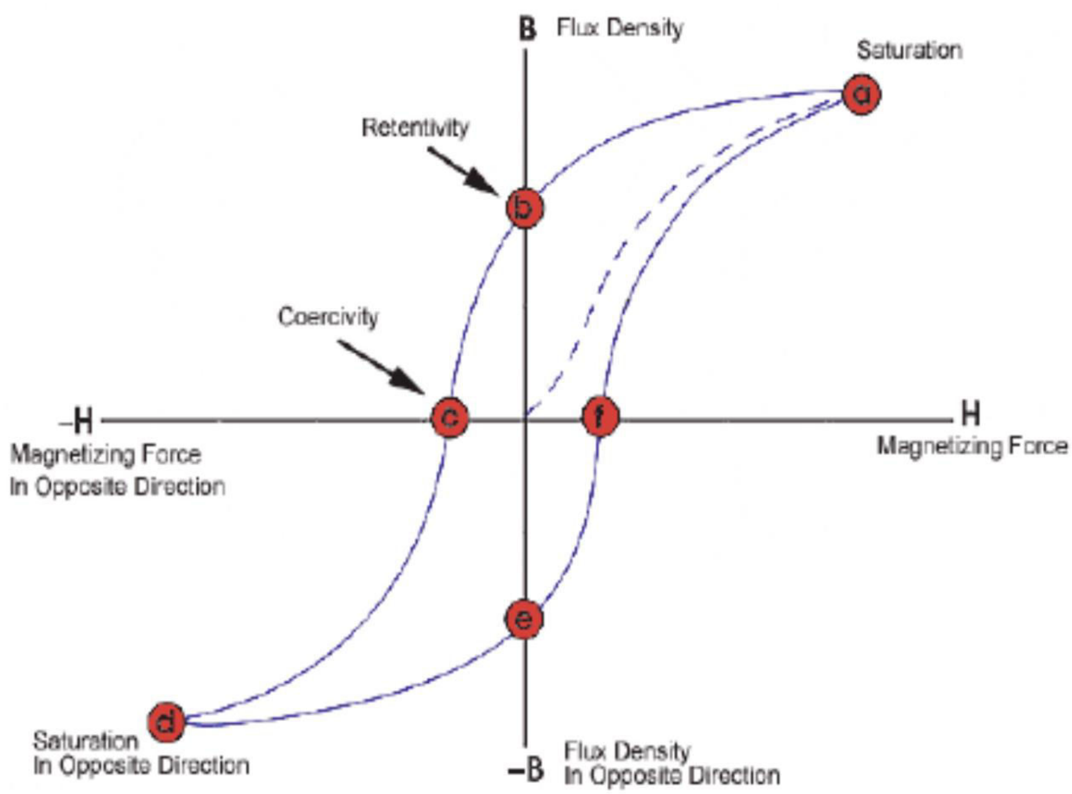
In practical surveying, the gamma (γ), which is the same as the nanoTesla (nT), is used:

$1 \gamma = 10^{-9}$ Tesla (or Weber/m^2 – SI unit)

The strength of the Earth's field is about 50,000 γ. Typical anomalies have amplitudes of a few 100 γ. The desirable field precision is usually 1 γ, or 10-5 of the Earth's total field. This contrasts with gravity, where desirable field precision is 0.01 mGal, or 10-8 of the Earth's field.

Inducing magnetization

The true behavior of induced magnetization may be investigated by placing a sample in a Coil until the sample is *saturated*, after which further increase in H produces no further increase in B. When H is returned to zero, some magnetization still remains. This is called r*emnant magnetization*. This pattern continues and forms a *hysteresis loop*. It shows how a sample can stay magnetized after the magnetizing force is gone.

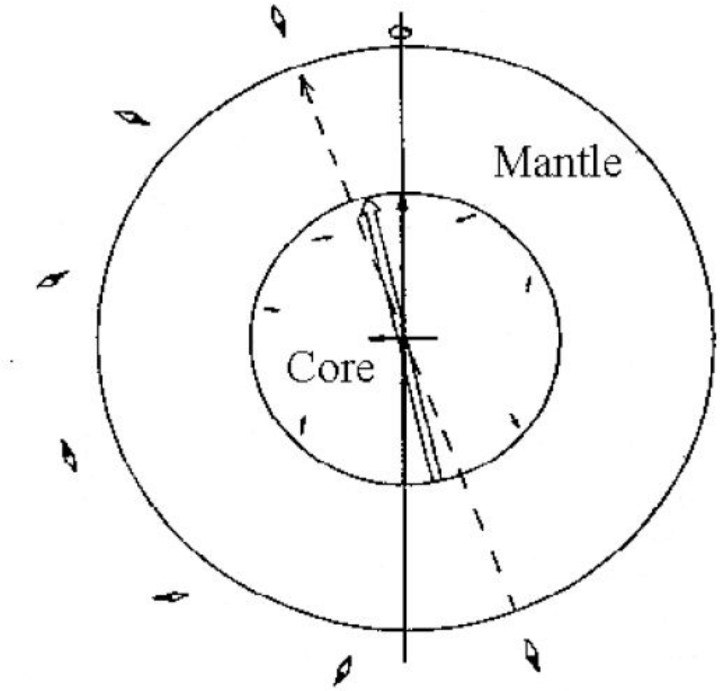


A hysteresis loop

The Earth's geomagnetic field

The Earth's magnetic field is more complicated than a simple dipole. It consists of:

a) The main field: This approximates to a non-geocentric dipole inclined to the Earth's spin axis. It can be modeled as polar and equatorial dipoles. A simple dipole is a good approximation for 80% of the Earth's field. The remainder can be modeled as dipoles distributed around the core/mantle boundary.



Modelling the Earth's magnetic field with dipoles

The origin of the Earth's field is known to be 99% internal and to be generated by convection in the liquid outer core, which drives electric currents. It cannot be due to magnetized rocks because it must be deep, and rocks lose all magnetization above the Curie temperature. The Curie temperature for magnetite is 578°C, whereas the temperature of the core is probably ~ 5,000°C.

b) The external field: This accounts for the other 1% of the Earth's field. It is caused by electric currents in ionized layers of the outer atmosphere. It is very variable, and has an 11-year periodicity which corresponds to sunspot activity. There is a diurnal periodicity of up to

30 γ, which varies with latitude and season because of the effect of the sun on the ionosphere. There is a monthly variation of up to 2 γ which is the effect of the moon on the ionosphere. Superimposed on this are micro pulsations which are seemingly random changes with variable amplitude, typically lasting for short periods of time.

Magnetic storms are random fluctuations caused by solar ionospheric interactions as sunspots are rotated towards and away from the Earth. They may last a few days and have amplitudes of up to 1,000 γ within 60° of the equator. They are more frequent and of higher amplitude closer to the poles, *e.g.*, in the auroral zone. The possibility of magnetic storms must be taken into consideration in exploration near the poles, *e.g.*, in Alaska.

c) Local anomalies: These are caused by magnetic bodies in the crust, where the temperature is higher than the Curie temperature. These bodies are the targets of magnetic surveying.

Mathematical treatment of main field

The terms used are:

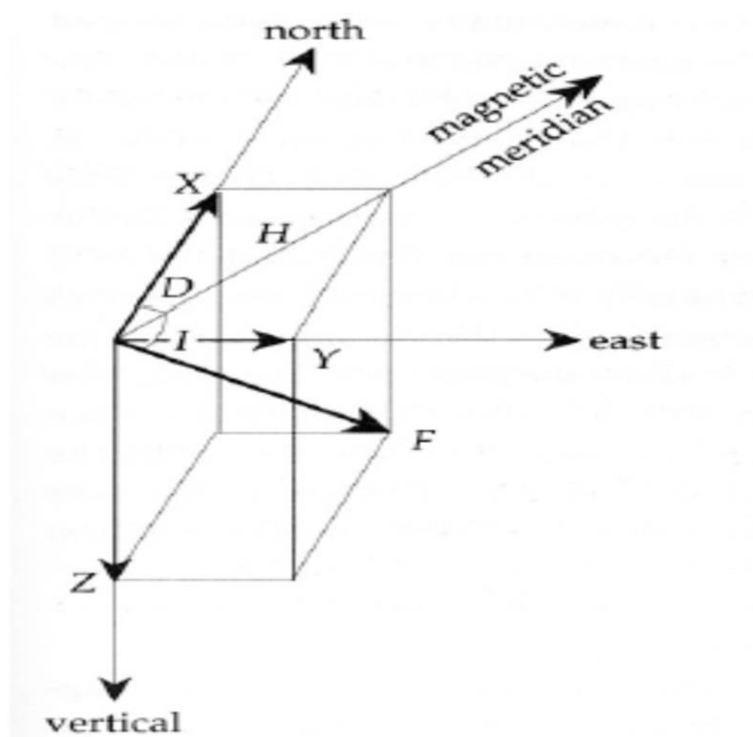
F = total field

H = horizontal component

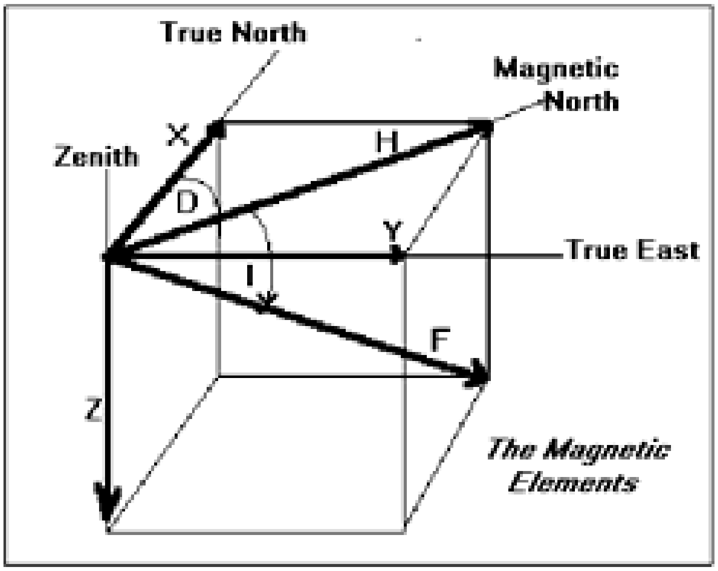
Z = vertical component

I = inclination

D = declination

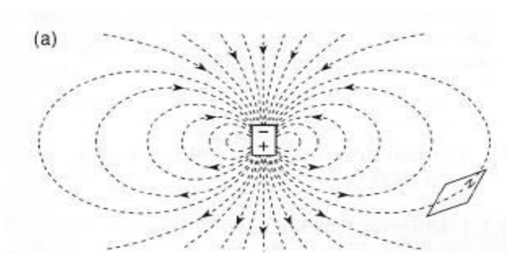


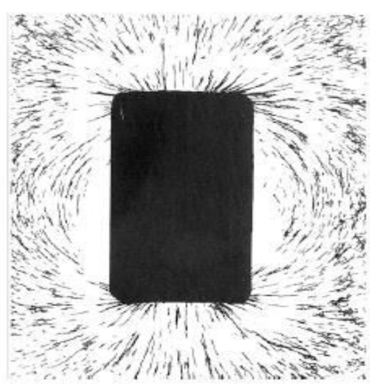
In surveying, ΔH, ΔZ or ΔF can be measured. It is most common to measure ΔF. Measurement of ΔH and ΔZ is now mostly confined to observatories.



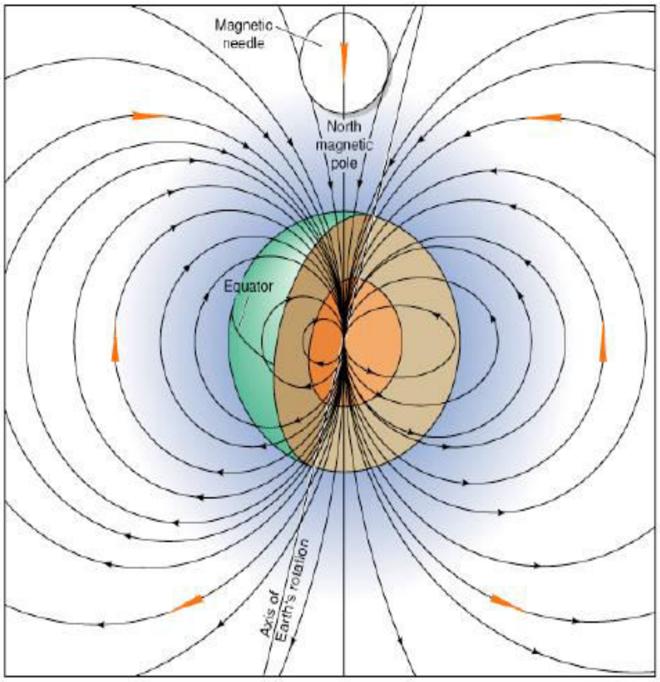
Components of Earths Magnetic Field

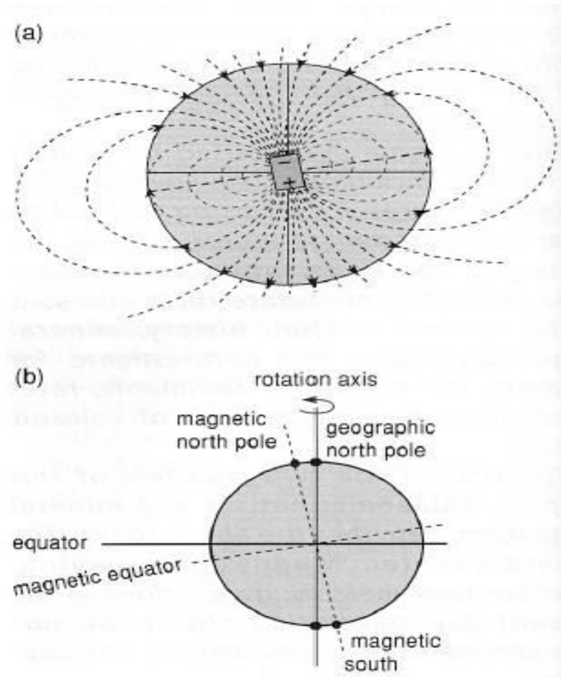
If a compass is placed in the vicinity of a magnet and moved along the direction it is pointing It traces a path from one end of the magnet to the other Many such paths (starting at different points) These paths represent the lines of **magnetic field.**

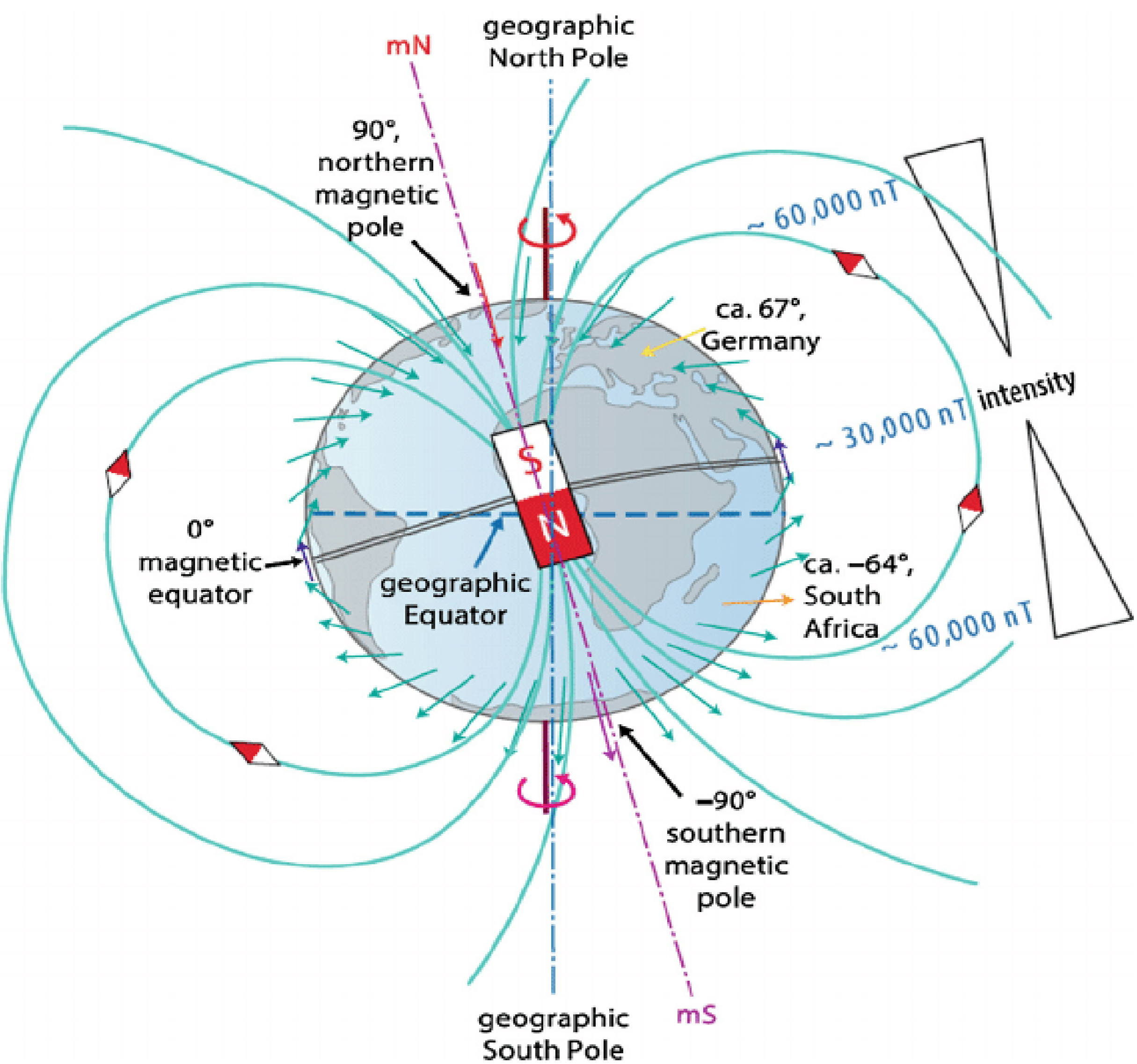




A compass points to the magnetic North because it aligns with the Earth's magnetic field lines The Earth's magnetic field can be approximated by a dipole (magnet) tilted about 11.5 degrees from its rotation axis. Magnetic North differs from true or geographic North.


Magnetic needle
North magnetic pole
Equator
Axis of Earth's rotation


(a)
(b)
rotation axis
magnetic north pole
geographic north pole
equator
magnetic equator
magnetic south


mN
geographic North Pole
90°, northern magnetic pole
~ 60,000 nT
ca. 67°, Germany
~ 30,000 nT
intensity
0° magnetic equator
S
N
geographic Equator
ca. –64°, South Africa
~ 60,000 nT
–90° southern magnetic pole
geographic South Pole
mS

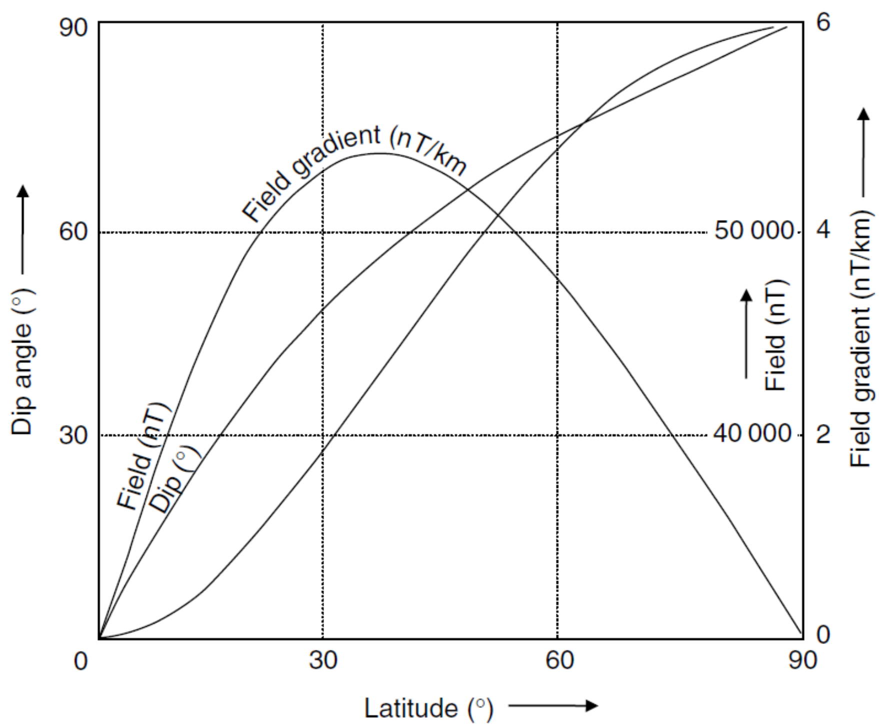
From the previous figure, The Earth's magnetic field (the geomagnetic field). Notice that the southern and northern magnetic poles and the magnetic equator do not coincide with the geographical poles and the geographic equator. Also notice that the magnetic field lines intersect the Earth's surface at different angles depending on the magnetic latitude (blue-green lines and vectors). The intersection angle is called the magnetic inclination. Magnetic inclination is +90° at the Magnetic North Pole (red vector), c. +67° at the latitude of Germany (yellow vector), 0° at the magnetic equator (dark blue vectors), c. −64° at the latitude of South Africa (orange vector), and −90° at the Magnetic South Pole (magenta vector) (Adapted with permission after Wiltschko and Wiltschko (1996) and Mouritsen (2013).) The magnetic intensity varies from c. 60,000 nT near the magnetic poles to c. 30,000 nT along the magnetic equator.

The magnetic fields of geological bodies are superimposed on the background of the Earth's main field. Variations in magnitude and direction of this field influence both the magnitudes and shapes of local anomalies.

In geophysics, the terms *north* and *south* used to describe polarity are replaced by positive and negative. The direction of a magnetic field is conventionally defined as the direction in which a unit positive pole would move but, since all things are relative, geophysicists give little thought to whether it is the north or south magnetic pole that is positive.

The Earth's main magnetic field originates in electric currents circulating in the liquid outer core, but can be largely modeled by a dipole source at the Earth's centre. Distortions in the dipole field extending over regions thousands of kilometers across can be thought of as caused by a relatively small number of subsidiary dipoles at the core–mantle boundary.

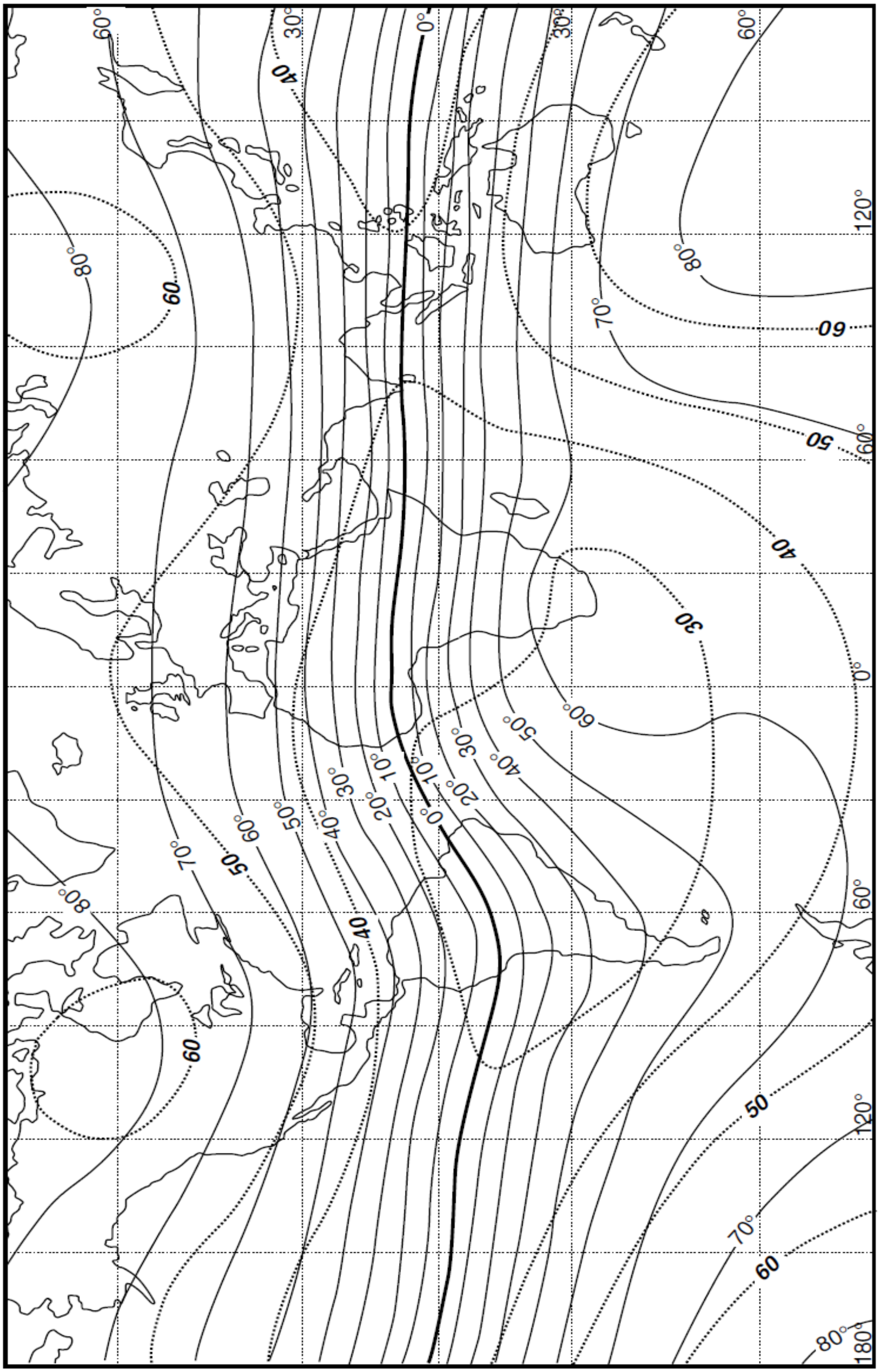
The variations with latitude of the magnitude and direction of an ideal dipole field aligned along the Earth's spin axis are shown in the following figure , Note that near the equator the dip angles change almost twice as fast as the latitude angles.



The figure above shows the Variation in intensity, dip and gradient for an ideal dipole aligned along the Earth's spin axis and producing a polar field of 60 000 nT.

To explain the Earth's actual field, the main dipole would have to be inclined at about 11◦ to the spin axis, and thus neither the magnetic equator, which links points of zero magnetic dip on the Earth's surface, nor the magnetic poles coincide with their geographic equivalents (see the following figure).

The North Magnetic Pole is in northern Canada and the South Magnetic Pole is not even on the Antarctic continent, but in the Southern Ocean at about 65◦S, 138◦E. Differences between the directions of true and magnetic North are known as declinations, presumably because a compass needle *ought* to point north but *declines* to do so.



Magnetic Dip (continuous lines, values in degrees) and magnetic intensity (dotted lines, values in thousands of nT) of the Earth's magnetic field. The thick continuous line is the magnetic equator.

Dip angles estimated from the global map (the previous figure) can be used to obtain rough estimates of magnetic latitudes and hence (using the last previous figure) of regional gradients. This approach is useful in determining whether a regional gradient is likely to be significant but gives only approximate correction factors, because of the existence of very considerable local variations. Gradients are roughly parallel to the local *magnetic* north arrow, so that corrections have E–W as well as N–S components. In ground surveys where anomalies of many tens of nT are being mapped, regional corrections, which generally amount to only a few nT per km, are often neglected.

<u>The International Geomagnetic Reference Field (IGRF)</u>

The variations of the Earth's main field with latitude, longitude and time are described by experimentally determined International Geomagnetic Reference Field (IGRF) equations, defined by 120 spherical harmonic coefficients, to order $N = 10$, supplemented by a predictive secular variation model to order $N = 8$. The shortest wavelength present is about 4000 km. IGRFs provide reasonable representations of the actual regional fields in well-surveyed areas, where they can be used to calculate regional corrections, but discrepancies of as much as 250 nT can occur in areas from which little information was available at the time of formulation.

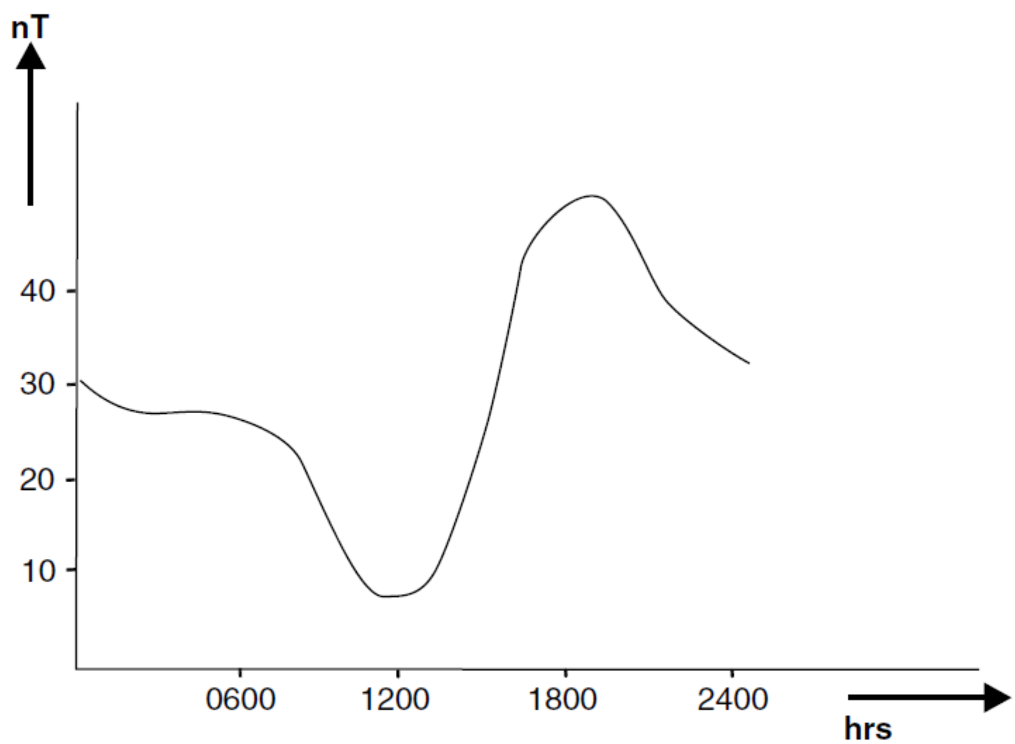
Because the long-term *secular* changes are not predictable except by extrapolation from past observations, the IGRF is updated every five years on the basis of observations at fixed observatories and is also revised retrospectively to give a definitive model (DGRF). GRF corrections are vital when airborne or marine surveys carried out months or years apart are being compared or combined but are less important in ground surveys, where base stations can be reoccupied.

<u>Secular variations in the main field</u>

These are very long period changes that result from convective changes in the core. They are monitored by measuring changes in I, D and F at observatories. The Earth's field is also subject to *reversals*, the last of which occurred at 0.7 Mya (i.e., 0.7 million years ago). They appear to be geologically sudden, not gradual, and their frequency is very variable. They are used for palaeomagnetic dating. This is done by comparing the sequence of reversals in the field area of interest to the known, dated, geological record of reversals.

Diurnal variations

The Earth's magnetic field also varies because of changes in the strength and direction of currents circulating in the ionosphere. In the normal *solarquiet* (Sq) pattern, the background field is almost constant during the night but decreases between dawn and about 11 a.m., increases again until about 4 p.m. and then slowly declines to the overnight value (see the following figure).



Typical 'quiet day' magnetic field variation at mid-latitudes.

Peak to-trough amplitudes in mid-latitudes are of the order of a few tens of nanoTesla. Since upper atmosphere ionization is caused by solar radiation, diurnal curves tend to be directly related to local solar time but amplitude differences of more than 20% due to differences in crustal conductivity may be more important than time dependency for points up to a few hundred kilometers apart. Short period, horizontally polarized and roughly sinusoidal *micro pulsations* are significant only in surveys that are to be contoured at less than 5 nT.

Within about 5◦ of the magnetic equator the diurnal variation is strongly influenced by the *equatorial electro jet*, a band of high conductivity in the ionosphere about 600 km (5◦ of latitude) wide. The amplitudes of the diurnal curves in the affected regions may be well in excess of 100 nT and may differ by 10 to 20 nT at points only a few tens of kilometers apart.

Many of the magnetic phenomena observed in polar regions can be explained by an *auroral electro jet* subject to severe short-period fluctuations. In both equatorial and polar regions it is particularly important that background variations be monitored continuously. Returning to a base station at intervals of one or two hours may be quite insufficient.

Magnetic storms

Short-term auroral effects are special cases of the irregular disturbances (Ds and Dst) known as *magnetic storms.* These are produced by sunspot and solar flare activity and, despite the name, are not meteorological, often occurring on clear, cloudless days. There is usually a sudden onset, during which the field may change by hundreds of nT, followed by a slower, erratic return to normality. Time scales vary widely but the effects can persist for hours and sometimes days. Micro pulsations are generally at their strongest in the days immediately following a storm, when components with periods of a few tens of seconds can have amplitudes of as much as 5 nT.

Ionospheric prediction services in many countries give advance warning of the general probability of storms but not of their detailed patterns, and the field changes in both time and space are too rapid for corrections to be applied. Survey work must stop until a storm is over. Aeromagnetic data are severely affected by quite small irregularities and for contract purposes *technical magnetic storms* may be defined, sometimes as departures from linearity in the diurnal curve of as little as 2 nT in an hour. Similar criteria may have to be applied in archaeological surveys when only a single sensor is being used (rather than a two-sensor gradiometer).

Geological effects

The Curie points for all geologically important magnetic materials are in the range 500–600 ◦C. Such temperatures are reached in the lower part of normal continental crust but below the Moho under the oceans. The upper mantle is only weakly magnetic, so that the effective base of local magnetic sources is the Curie isotherm beneath continents and the Moho beneath the oceans.

Massive magnetite deposits can produce magnetic fields of as much as 200 000 nT, which is several times the magnitude of the Earth's normal field. Because of the dipolar nature of magnetic sources these, and all other, magnetic anomalies have positive and negative parts and in extreme cases directional magnetometers may even record negative fields. Anomalies of this size are unusual, but basalt dykes and flows and some larger basic intrusions can produce fields of thousands and occasionally tens of thousands of nT. Anomalous fields of more than 1000 nT are otherwise rare, even in areas of outcropping crystalline basement. Sedimentary rocks generally produce changes of less than 10 nT, as do the changes in soil magnetization important in archaeology. In some tropical areas, magnetic fields of tens of nT are produced by maghemite formed as nodular growths in laterites. The nodules may later weather out to form ironstone gravels which give rise to high noise levels in ground surveys. The factors that control the formation of maghemite rather than the commoner, non-magnetic form of hematite are not yet fully understood.

Rock magnetism

Kinds of minerals magnetism

1-Diamagnetism: In diamagnetic minerals, all the electron shells are full, and there are no unpaired electrons. The electrons spin in opposite senses and the magnetic effects cancel. When placed in an external field, the electrons rotate to produce a magnetic field in the opposite sense to the applied. Such minerals have negative susceptibilities, k. Examples of such materials are quartzite and salt. Salt domes thus give diamagnetic anomalies, i.e., weak negative anomalies.

2- Paramagnetism: Paramagnetic minerals are ones where the electron shells are incomplete. They generate weak magnetic fields as a result. When placed in an external field, a magnetic field in the same sense is induced, i.e., k is positive. Examples of materials that are paramagnetic are the Ca - Ni element series.

3- Ferromagnetism: Ferromagnetic minerals are minerals that are paramagnetic, but where groups of atoms align to make domains. They have much larger k values than paramagnetic elements. There are only three ferromagnetic elements – Iron, Cobalt and Nickel. Ferromagnetic minerals do not exist in nature.

There are three types of ferromagnetism:

a- ***pure ferromagnetism:*** all the domains align the same way, producing strong magnetism.

b- ***ferrimagnetism:*** In the case of ferrimagnetic minerals, the domains are subdivided into regions that are aligned in opposition to one another. One direction is weaker than the other. Almost all natural magnetic minerals are of this kind, *e.g.*, magnetite (Fe2O3), which is the most common, ilmenite, titanomagnetite and the oxides of iron or iron-and-titanium.

c- ***antiferromagnetism:*** *a*ntiferromagnetic minerals have opposing regions that are equal.

An example is haematite (Fe2O3). Occasionally there are defects in the crystal lattice which cause weak magnetisation, and this is called *parasitic antiferromagnetism*.

The Curie temperature

This is the temperature of demagnetization. Some examples of Curie temperatures are:

- Fe 750°C
- Ni 360°C
- magnetite 578°C

The majority of anomalies in the crust result from magnetite, and so knowledge of the Curie temperature and the geothermal gradient can give information on the depth range of causative bodies.

Types of magnetism

1- Induced magnetism: This is due to induction by the Earth's field, and is in the same direction as the Earth's field. Most magnetization is from this source. It is important to appreciate that since the Earth's field varies from place to place, the magnetic anomaly of a body will vary according to its location.

2- Remnant magnetism: This is due to the previous history of the rock. There are various types:

a- Chemical remnant magnetization (CRM): This is acquired as a result of chemical grain accretion or alteration, and affects sedimentary and metamorphic rocks.

b- ***Detrital remnant magnetisation (DRM):*** This is acquired as particles settle in the presence of Earth's field. The particles tend to orient themselves as they settle.

c- Isothermal remnant magnetism (IRM): This is the residual magnetic field left when an external field is applied and removed, *e.g.*, lightning.

d- Thermoremnant magnetization (TRM): This is acquired when rock cools through the Curie temperature, and characterizes most igneous rocks. It is the most important kind of magnetization for palaeomagnetic dating.

e- Viscous remnant magnetism (VRM): Rocks acquire this after long exposure to an external magnetic field, and it may be important in fine-grained rocks.

Induced and remnant magnetism

The direction and strength of the present Earth's field is known. However, we may know nothing about the remnant magnetization of a rock. For this reason, and because in strongly magnetized rocks the induced field dominates, it is often assumed that all the magnetization is induced. The true magnetization is the vector sum of the induced and remnant components. However, the remnant magnetization be measured using a static or Spinner magnetometer, which measure the magnetism of samples in the absence of the Earth's field.

Rock susceptibility

These are analogous to density in gravity surveying. Most rocks have very low susceptibilities. The susceptibility of a rock is dependent on the quantity of ferrimagnetic minerals. *In situ* measurements of rock susceptibility may be made using special magnetometers but it is more common to measure a sample in the laboratory, *e.g.*, using an induction balance. The sample is placed in a coil, a current is applied and the induced magnetization is measured. It is usual to quote the strength of the field applied along with the result. If the applied field was very much greater than the Earth's field, the value obtained may not be suitable for interpreting magnetic anomalies.

Another method of obtaining the susceptibility of a rock is to assume that all the magnetization is due to magnetite. The volume percent of magnetite is multiplied by the Susceptibility of magnetite. This method has produced good correlation with field measurements.

Susceptibility (k), ranges over 2-3 orders of magnitude in common rock types. Basic igneous rocks have the highest susceptibilities since they contain much magnetite. The proportion of magnetite tends to decrease with increasing rocks acidity, and thus k tends to be low for acid rocks such as granite. The susceptibility of metamorphic rocks depends on the availability of O_2 during their formation, since plentiful O_2 results in magnetite forming.

Sedimentary rocks usually have very low k, and sedimentary structures very rarely give large magnetic anomalies. If a large magnetic anomaly occurs in a sedimentary environment, it is usually due to an igneous body at depth.

The *susceptibility* of a rock usually depends on its magnetite content. Sediments and acid igneous rocks have small susceptibilities whereas basalts, dolerites, gabbros and serpentinites are usually strongly magnetic. Weathering generally reduces susceptibility because magnetite is oxidized to hematite, but some laterites are magnetic because of the presence of maghemite and remanently magnetized hematite. The susceptibilities, in rationalized SI units, of some common rocks and minerals are given in the following table.

Common rocks	
Slate	0–0.002
Dolerite	0.01–0.15
Greenstone	0.0005–0.001
Basalt	0.001–0.1
Granulite	0.0001–0.05
Rhyolite	0.00025–0.01
Salt	0.0–0.001
Gabbro	0.001–0.1
Limestone	0.00001–0.0001
Ores	
Hematite	0.001–0.0001
Magnetite	0.1–20.0
Chromite	0.0075–1.5
Pyrrhotite	0.001–1.0
Pyrite	0.0001–0.005

Common causes of magnetic anomalies

Dykes, folded or faulted sills, lava flows, basic intrusions, metamorphic basement rocks and ore bodies that contain magnetite all generate large-amplitude magnetic anomalies. Other targets suitable for study using magnetic method are disturbed soils at shallow depth, fire pits and kilns, all of which are of interest in archaeological studies. Magnetic anomalies can be used to get the depth to basement rocks, and hence sedimentary thickness, and to study metamorphic thermal aureoles.

Instruments for Measuring Rocks Magnetism

1 -Observatory instruments: These are similar to field instruments, but measure all three components of the field.

2 -Magnetic balance: Magnetometers were originally mechanical, but after World War II they were replaced by flux-gate instruments, rendering the magnetic balance now obsolete. However, a lot of data exist that were collected using magnetic balances. An example is the Schmidt magnetic balance. A magnet pivots about a point that is not the centre of gravity.

The torque of the Earth's magnetic field balances with the gravity effect at the centre of gravity. The angle of pivot is a function of the magnetic field, and is measured by a light beam that is projected onto a scale. This changes between measuring stations and thus the magnetic balance is a *relative* instrument.

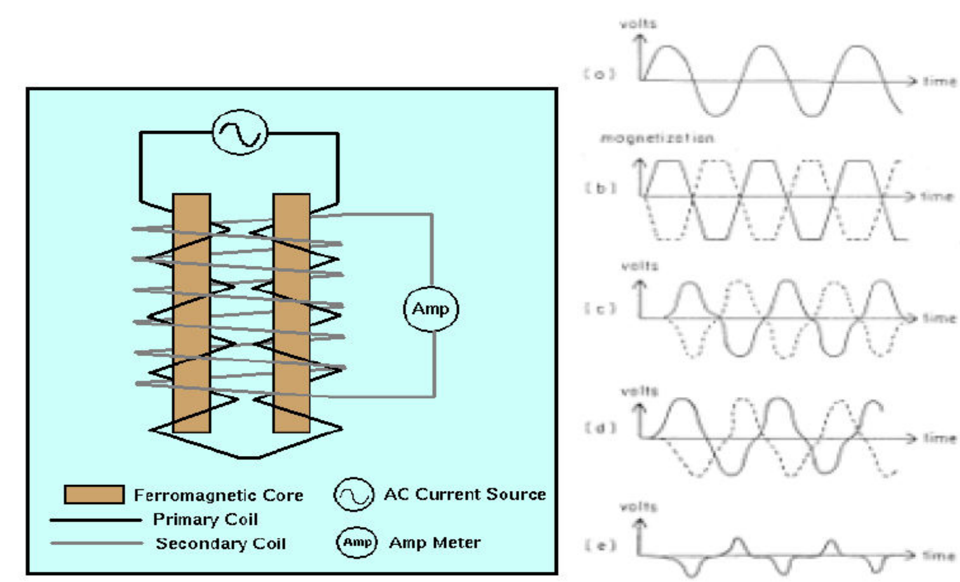
3- Flux-gate magnetometer: This was originally developed for detecting submarines in World War II. It is used a lot for aeromagnetic work because it makes continuous measurements. The construction of this instrument involves two coils wound in opposition. A current is passed through to induce magnetization. A secondary winding measures the voltage induced by the induced magnetization. In the absence of the Earth's field these two cancel out. An AC current is applied that saturates the cores in opposition in the absence of the Earth's field. The Earth's field reinforces one core and opposes the other. This causes the voltages induced in the secondary coils to get out of step. The result is a series of pips whose height is proportional to the ambient field. The precision is better than 0.5 - 1 γ.

This instrument has the following disadvantages:

• It is not an absolute instrument and it is liable to drift,

• It is insufficiently accurate for modern work.

For these reasons it has now been largely superseded by the proton precession and alkali Vapor magnetometers for land and sea work. It is still used in boreholes, however.



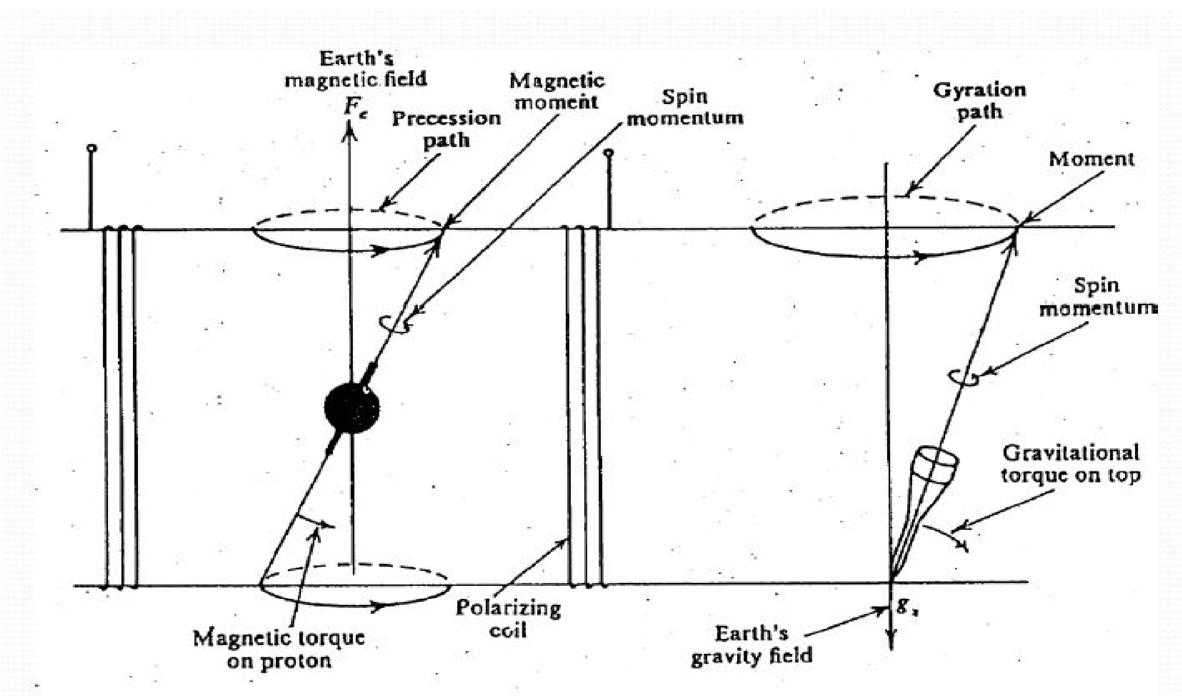
Schematics showing design of flux-gate magnetometer and principles of operation

4 -Proton precession magnetometer (free-precession magnetometer):

This instrument resulted from the discovery of *nuclear magnetic resonance*. Some atomic nuclei have magnetic moment that causes them to precess around the ambient magnetic field like a spinning top precesses around the gravity field. Protons behave in this way. The magnetometer consists of a container of water or oil, which is the source of protons, around which a coil is wound. A current is applied so a field of 50-100 oersteds is produced. The container must be oriented so this field is not parallel to the Earth's field. The current is removed abruptly, and the protons precess as they realign to the Earth's field. The precession frequency is measured.

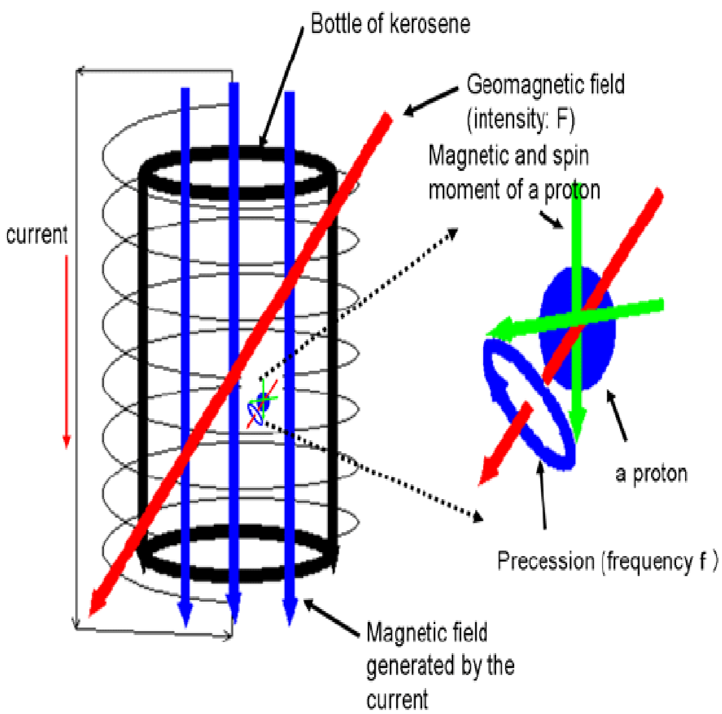
When the current is applied, it takes the protons 2-3 seconds to fully align, and this follows an exponential relationship. The current is cut off abruptly compared with the period of precession (*e.g.*, over 50 ms). It takes time for the precession to build up after the applied field has been removed. This then decays exponentially. About 50,000 cycles are measured and this gives an accuracy of 1 cycle in 50,000, i.e. 1 γ. It takes about 1/2 second to count 50,000 cycles. The strength of the measured signal is about 10 mV (10 microvolts).

Most proton precession magnetometers have a precision of 1 γ but models are available that have precisions as good as 0.1 or 0.01 γ.



Proton precession and spinning-top analogy

The proton precession magnetometer makes use of the small magnetic moment of the hydrogen nucleus (proton). The sensing element consists of a bottle containing a low freezing-point hydrocarbon fluid about which is wound a coil of copper wire. Although many fluids *can* be used, the manufacturer's recommendation, usually for high-purity decane, should always be followed if the bottle has to be topped up. A *polarizing* current of the order of an amp or more is passed through the coil, creating a strong magnetic field, along which the moments of the protons in the hydrogen atoms will tend to become aligned.



Advantages offered by proton precession magnetometers include:

• Great sensitivity,

• They measure the total field,

• They are absolute instruments,

• They do not require orientation or leveling like the flux-gate magnetometer,

• There are no moving parts like the flux-gate magnetometer. This reduces power consumption and breakdowns.

The disadvantages include:

• Each measurement takes several seconds. This is a great disadvantage for aeromagnetic work,

• Field gradients that are so large that they are significant within the bottle will cause inaccurate readings or no sensible reading at all.

• They do not work in the presence of AC power interference, *e.g.*, below power lines.

With the proton magnetometer, surveys are very simple and quick.

5 -Overhauser effect proton magnetometer:

This magnetometer uses a proton-rich fluid with paramagnetic ions added. The paramagnetic ions resonate at a frequency called the *free-electron* resonant frequency, which is in the VHF radio frequency range. A saturating VHF signal is applied. The nuclear spin of the protons is polarized as a result of interaction with the electrons. This is the equivalent to magnetic polarization in the proton-precession magnetometer. In the case of the Overhauser effect, the polarization is continuous and thus the proton precession signal changes continuously with the ambient magnetic field.

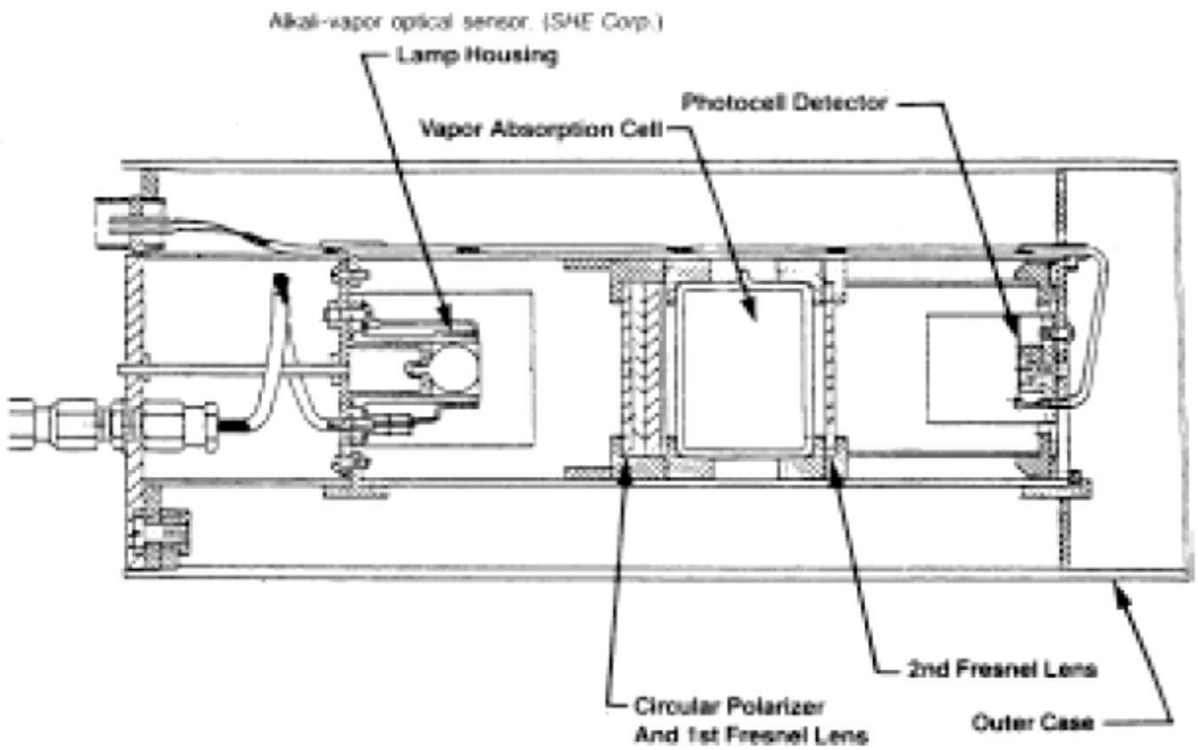
The Overhauser effect proton magnetometer has the following advantages over the proton precession magnetometer:

• It produces a continuous effect. Less time is needed to make a measurement and it can thus sample more rapidly – up to 8-10 readings per second may be made,

• The signal strength is 1-10 mV, so the signal-to-noise ratio is better than for proton precession magnetometers.

6 -Optical pump (or alkali vapour) magnetometer:

This device dates from the 1960s. It is used if sub-γ sensitivity is needed, *e.g.* for sedimentary targets and for measuring magnetic gradients directly. It consists of a cell of helium, cesium, rubidium or some alkali-metal vapor which is excited by light from a source of the same material. The energy states of the electrons of the alkali metal atoms are affected by magnetic fields. In the presence of the light, the depopulation of energy states by light absorption and movement to higher states will be unequal. The repopulation to lower states by emission of energy will be equal for all states and thus unequal populations in the various energy states result. This is the *optically-pumped* state, and the gas is more transparent like this. The transparency is measured, and gives a precision of 0.005 γ in the measurement of the strength of the Earth's field.



Alkali vapour magnetometer

7 - Magnetic gradiometer: It is possible to measure either the vertical or horizontal gradient using two optical pumping devices separated by fixed distances. This may be done on land with an instrument carried by hand, or from a helicopter or ship. The advantages of this type of instrument are:

• Diurnal variation corrections are not necessary.

• Shallow sources with steep gradients are accentuated compared with deep sources with gentle gradients.

8- The SQUID system:

SQUID stands for Superconducting Quantum Interference Device. It has a high sensitivity, and is used to measure both the direction and magnitude of the Earth's field. Thus three components are needed. It has a sensitivity of 10-5 γ. The response is flat at all frequencies, and thus it can measure a changing magnetic field also, from DC to several 1,000 Hz. This instrument operates at liquid helium temperatures. It is physically large and thus not very portable. Some uses include magneto telluric measurements, measurements of drift of the Earth's field and laboratory measurements for remnant and induced magnetization of rock samples.

A SQUID magnetometer

<u>Magnetic surveys</u>

Magnetic surveys either directly seek magnetic bodies or they seek magnetic material associated with an interesting target. For example, magnetic minerals may exist in faults or fractures.

1- Land surveys:

These are usually done with portable proton precession magnetometers. Profiles or networks of points are measured in the same way as for gravity. It is important to survey perpendicular to the strike of an elongate body or two-dimensional modeling may be very difficult.

It is necessary to tie back to the base station at 2-3 hour intervals, or to set up a continually-reading base magnetometer. This will give diurnal drift and detect magnetic storms.

The operator must:

- Record the time, at which readings were taken, for drift correction,
- Stay away from interfering objects, e.g., wire fences, railway lines, roads.
- Not carry metal objects *e.g.*, mobile phones, and
- Take multiple readings at each station to check for repeatability.

Reduction of the observations is much simpler than for gravity:

1. The diurnal correction

This may be up to 100 γ. Observatory data may be used if the observatory is within about 100 km and no major magnetic bodies occur in between, which might cause phase shifts in the temporal magnetic variations. A magnetic storm renders the data useless.

2. Regional trends

These are corrected for in the same way as for gravity, i.e., a linear gradient or polynomial surface is fit to regional values, and subtracted. The UK regional gradient is 2.13 γ/km N and 0.26 γ/km W. Another method is to subtract the predicted IGRF (International Geomagnetic Reference Field) which is a mathematical description of the field due to the Earth's core. There are several such formulae to choose from, all based on empirical fits to observatory or satellite data.

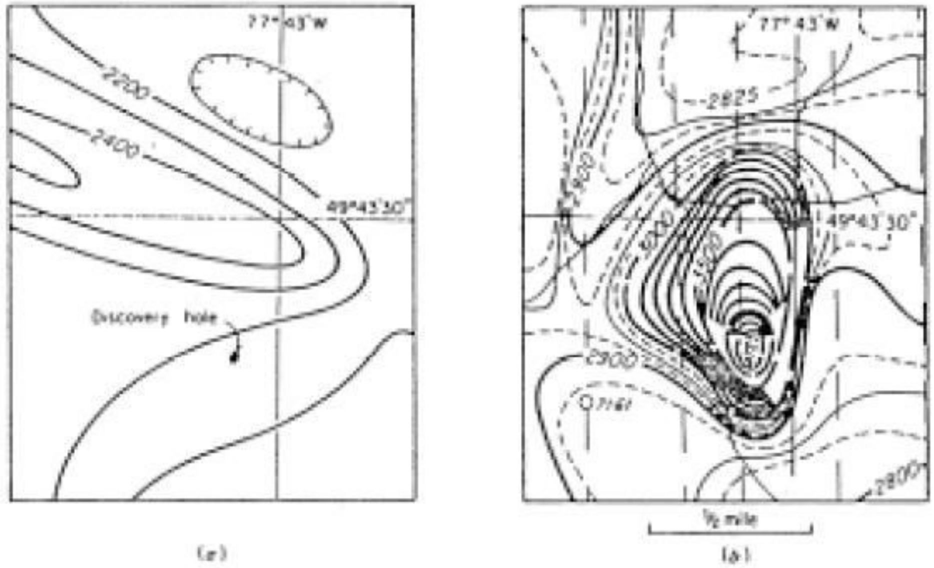
The other corrections made to gravity readings are not necessary in magnetic surveys. In particular, small elevation changes have a negligible effect on magnetic anomalies, and thus elevation-related corrections are not needed for most surveys.

Air Magnetic surveys

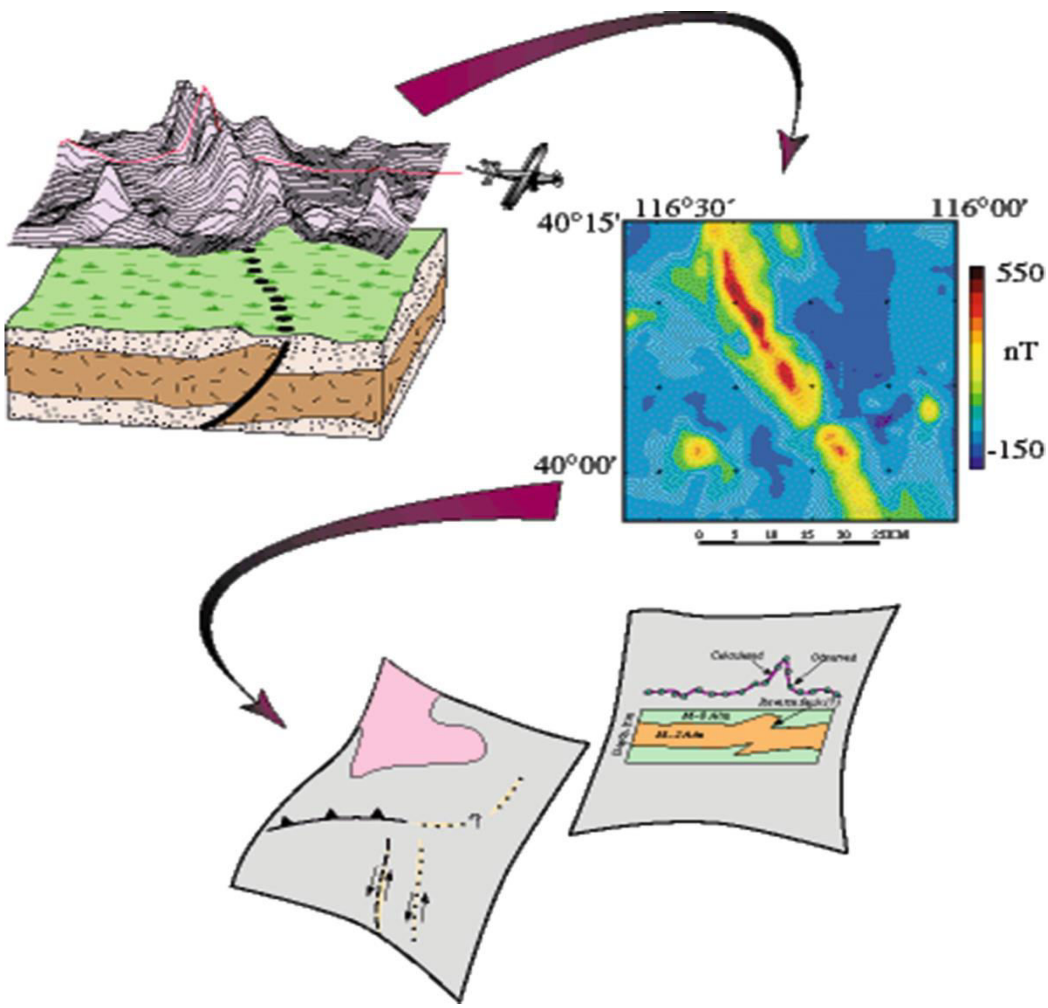
Most magnetic surveying is aeromagnetic surveying, and it may be done with either airplane or helicopter. Helicopters are more suitable for detailed or difficult access areas, though airplanes are cheaper. Usually a proton magnetometer is towed behind the helicopter, and thus discrete measurements are made. The magnetometer is then called the *bird*. It may also be mounted as a tail stinger on planes because of problems with sensor motion and cable vibrations that result from higher speeds. Wingtip mounting is also available. If the instrument is housed inboard, it is necessary to correct for the magnetic effect of airplane. It is also possible to measure vertical and longitudinal gradients and often several magnetometers are flown to maximize the use of the flight. Aeromagnetic surveying is not useful for surveys where great spatial accuracy or very dense measurements are required. The use of aeromagnetic surveying is limited by navigation which is a first order problem for such surveys. Before the widespread use of the GPS this was done by radio beacon *e.g.*, Loran or aerial photography. Photography is not a solution sometimes, however, *e.g.*, over jungles or the sea where there are no landmarks. Doppler navigation was sometimes used. This involves radio beams that are bounced off the ground both before and aft of the aircraft. This gives the speed of the aircraft accurately. The recent advent of the GPS has enabled Aeromagnetic surveys to be used more widely for sea surveys. Prior to the GPS, navigation accuracy was often no better than 100 m.

The layout of the survey depends on target scale and anomaly strike. Usually a criss-cross pattern of perpendicular flight paths is adopted. The lines perpendicular to strike are more closely spaced, and the *tie lines* at right angles to these may be typically at 1/4 or 1/10 the density.

The optimum design of lines has been examined analytically to determine the optimum spacing for a given anomaly width. The difference in anomaly deduced from lines with spacing's of 0.5 and 0.25 miles is illustrated by a study of one of the largest sulphide deposits in Canada:



Results from sparse and dense flight lines



The flight height is typically 200 to 1,000s of feet, and should remain as constant as possible.

For oil reconnaissance, the most interesting feature is generally deep basement structure. In this case, high surveys are flown, typically above 1000', to effectively filter out the signals from small, shallow bodies.

Diurnal drift and other errors, *e.g.*, variable flying height can be averaged out by:

• minimising the RMS of the line crossover measurement differences, or

• fitting a high-order polynomial to each line and eliminating the differences completely.

This procedure is similar to that applied to the SEASAT data and is a technique used in geodetic surveys. A fixed ground magnetometer is used to monitor for magnetic storms. Aeromagnetic surveys have the major advantages that they are very cheap, can cover huge areas, and can filter out shallow, high-frequency anomalies. This latter is also the major disadvantage of aeromagnetic surveying for prospecting for shallow bodies within mining range. Onboard computers for quasi-real time assessment of the data are becoming more common, as it is important to make sure the data are satisfactory before demobilizing.

Sea Magnetic Surveys

The instrument is towed behind the ship at a distance of up to 500 m to avoid the magnetic effect of ship. It is then known as the *fish*. The instrument is made buoyant and a proton magnetometer is usually used. The sampling frequency is typically 4-20 s, giving measurements spaced at intervals of 8-16 m if the ship speed is 4-6 knots. GPS navigation is almost universally used now. Loran navigation was most common in the past.

Sea magnetic surveys are generally conducted at the same time as a seismic survey. The ship's course is optimised for the seismic survey and it is thus usually non-optimal for the magnetic survey. There may also be problems making the diurnal correction if the ship is more than 100 km from land. Under these circumstances the diurnal correction may have to be done by tie-line analysis. Recently, the longitudinal gradient is frequently measured and used, and has caused great improvement in the usefulness of marine magnetic data for oil exploration.

Examples of Global regional magnetic surveys

Good examples to study to gain familiarity with the application of magnetic surveying to studying a range of geological problems include:

• The combination of sea and land survey data and tectonic interpretation done in the Eastern Mediterranean and Near East. For details, see: http://www.gii.co.il/html/ground/GravityNew/1.pdf

• The use of magnetic anomalies to deduce the tectonic evolution of the North Atlantic. For original paper, see: Nunns, A. G. (1983), Plate tectonic evolution of the Greenland-Scotland ridge and surrounding regions, in *Structure and Development of the Greenland-Scotland Ridge*, edited by M. H. P. Bott, Saxov S., Talwani, M. & Thiede, J., pp. 1-30, Plenum Press, New York and London.

• Correlation of volcanic features, *e.g.*, Long Valley caldera, with the regional magnetic field in California.

Magnetic survey Data display

As with all geophysical data, outliers must be removed before turning to modeling and interpretation work. Algorithms are available for this, or the geophysicist can do it by brain. It may be desired to interpolate all the measurements onto a regular grid in the case of aeromagnetic data which are not uniformly measured. It is critical to maintain the integrity of the data when this is done. For example, if contouring at 5 γ intervals, each data point should fit the raw measurements to within 2.5 γ.

Contour maps are the most common, but generating them without degrading the data nor introducing statistically unsupported artifacts is not simple. Decisions that must be taken in designing the final map are:

• contour interval (*e.g.* 0.25 γ),

• sampling interval (*e.g.* 75 m),

• height (*e.g.*, mean-terrain clearance of 150 m),

• spacing of flight lines,

• geomagnetic reference surface to subtract,

• % of data samples to contour,

• interpolation method (*e.g.*, minimum curvature with bicubic spline refinement).

Aeromagnetic data

Displaying the data as offset profiles is common.

High frequencies attenuate out at great height. Very quiet, smooth aeromagnetic maps characterize sedimentary basins with deep basement. A lot of high frequency anomalies indicate shallow anomalous bodies that may be ore or igneous rocks. It is important to find the depth of bodies in sedimentary regimes. An irregular basement is often truncated by erosion, and thus the depth corresponds to the thickness of the sedimentary sequence.

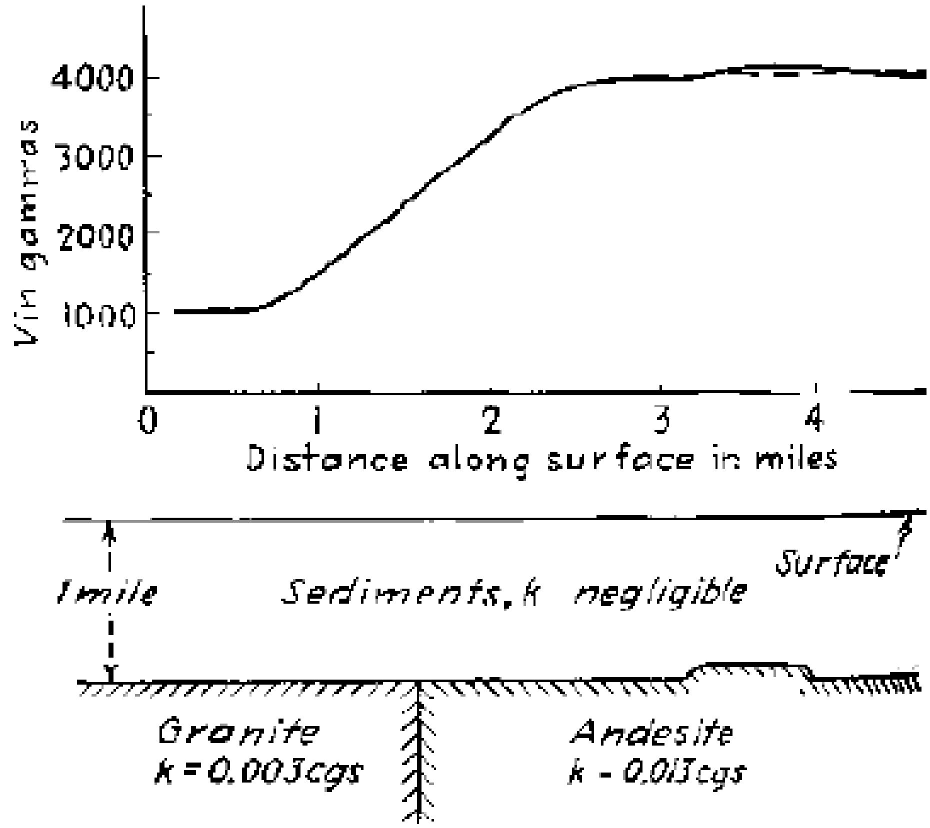
Abrupt changes in magnetic character may reflect boundaries between magnetic provinces, or a basement fault. The strike of the magnetic trend indicates the structural trend of area.

Using processing methods such as upward and downward continuation, the magnetic field can be calculated at any height. It may be helpful to display the data as they would look had the survey been made at a higher elevation, thereby filtering out high-frequency anomalies due to small, shallow bodies.

Data interpretation

1 -Problems and errors related to magnetic data interpretation:

a- Magnetization: This may not be uniform, but the interpreter is usually obliged to assume it is. Large anomalies tend not to reflect major lateral changes in structure, as they do in gravity, rather lateral changes in rock susceptibility. In the example illustrated below, a lateral change in susceptibility gives a 3000 γ anomaly. Elsewhere, a basement ridge 1000' high gives only a 120 γ anomaly. The basement ridge is important for assessing the oil potential of the sedimentary basement, but the change in susceptibility is of no consequence.



Comparison of magnetic effect of lateral susceptibility change in basement with effect of Structural feature on basement surface.

b- Ambiguity: The ambiguity problem is the same for magnetic surveying as it is for gravity. There is ambiguity between size and distance. In the case of magnetic the problem is much worse, as there are many other ambiguities, *e.g.*, between the direction of the body magnetization and the dip of the Earth's field.

<u>Noise in ground magnetic surveys</u>

Magnetic readings in populated areas are usually affected by stray fields from pieces of iron and steel (*cultural noise*). Even if no such materials are visible, profiles obtained along roads are usually very distorted compared to those obtained on parallel traverses through open fields only 10 or 20 m away.

Since the sources are often quite small and may be buried within a metre of the ground surface, the effects are very variable. One approach to the noise problem is to try to take all readings well away from obvious sources, noting in the field books where this has not been possible. Alternatively, the almost universal presence of ferrous noise can be accepted and the data can be filtered. For this method to be successful, many more readings must be taken than would be needed to define purely geological anomalies. The technique is becoming more popular with the increasing use of data loggers, which discourage note-taking but allow vast numbers of readings to be taken and processed with little extra effort, and is most easily used with alkali vapor and fluxgate instruments which read virtually continuously.

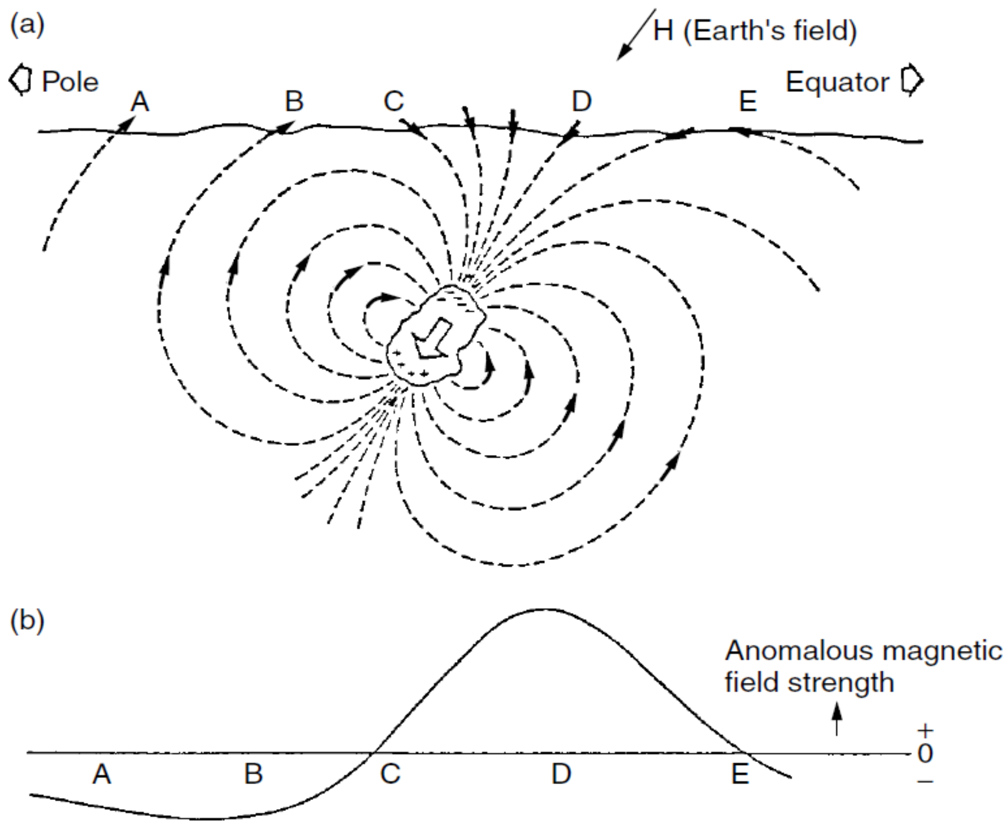
Simple Magnetic Interpretation

Field interpretation of magnetic data allows areas needing infill or checking to be identified and then revisited immediately and at little cost. Good interpretation requires profiles, which preserve all the detail of the original readings, and contour maps, which allow trends and patterns to be identified. Fortunately, the now almost ubiquitous laptop PC has reduced the work involved in contouring (providing the necessary programs have been loaded).

Forms of magnetic anomaly

The shape of a magnetic anomaly varies dramatically with the dip of the Earth's field, as well as with variations in the shape of the source body and its direction of magnetization. Simple sketches can be used to obtain rough visual estimates of the anomaly produced by any magnetized body.

The following figure shows an irregular mass magnetized by induction in a field dipping at about 60◦. Since the field direction defines the direction in which a positive pole would move, the effect of the external field is to produce the distribution of poles shown. The secondary field due to these poles is indicated by the dashed lines of force. Field direction is determined by the simple rule that like poles repel.



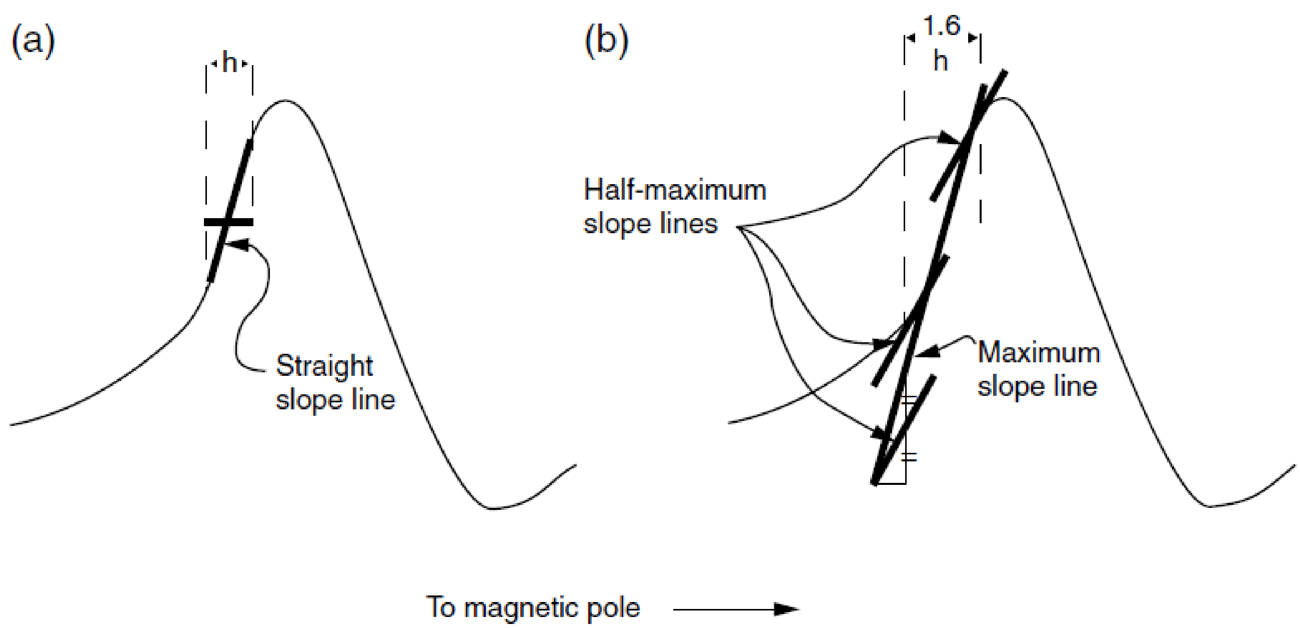
Mid-latitude total field anomaly due to induced magnetization.
(a) The induced field. (b) The anomaly profile, derived as described in the text.

If the secondary field is small, the directions of the total and background fields will be similar and no anomalous field will be detected near C and E. The anomaly will be positive between these points and negative for considerable distances beyond them. The anomaly maximum will be near D, giving a magnetic profile with its peak offset towards the magnetic equator (the previous figure). At the equator the total-field anomaly would be negative and centered over the body and would have positive side lobes to north and south, as can easily be verified by applying the method of the figure above to a situation in which the inducing field is horizontal.

Because each positive magnetic pole is somewhere balanced by a negative pole, the net flux involved in any anomaly is zero. Over the central parts of a uniform magnetized sheet the fields from positive and negative poles cancel out, and only the edges are detected by magnetic surveys. Strongly magnetized but flat-lying bodies thus sometimes produce little or no anomaly.

‘Rule-of-thumb’ in depth estimation of magnetized subsurface bodies

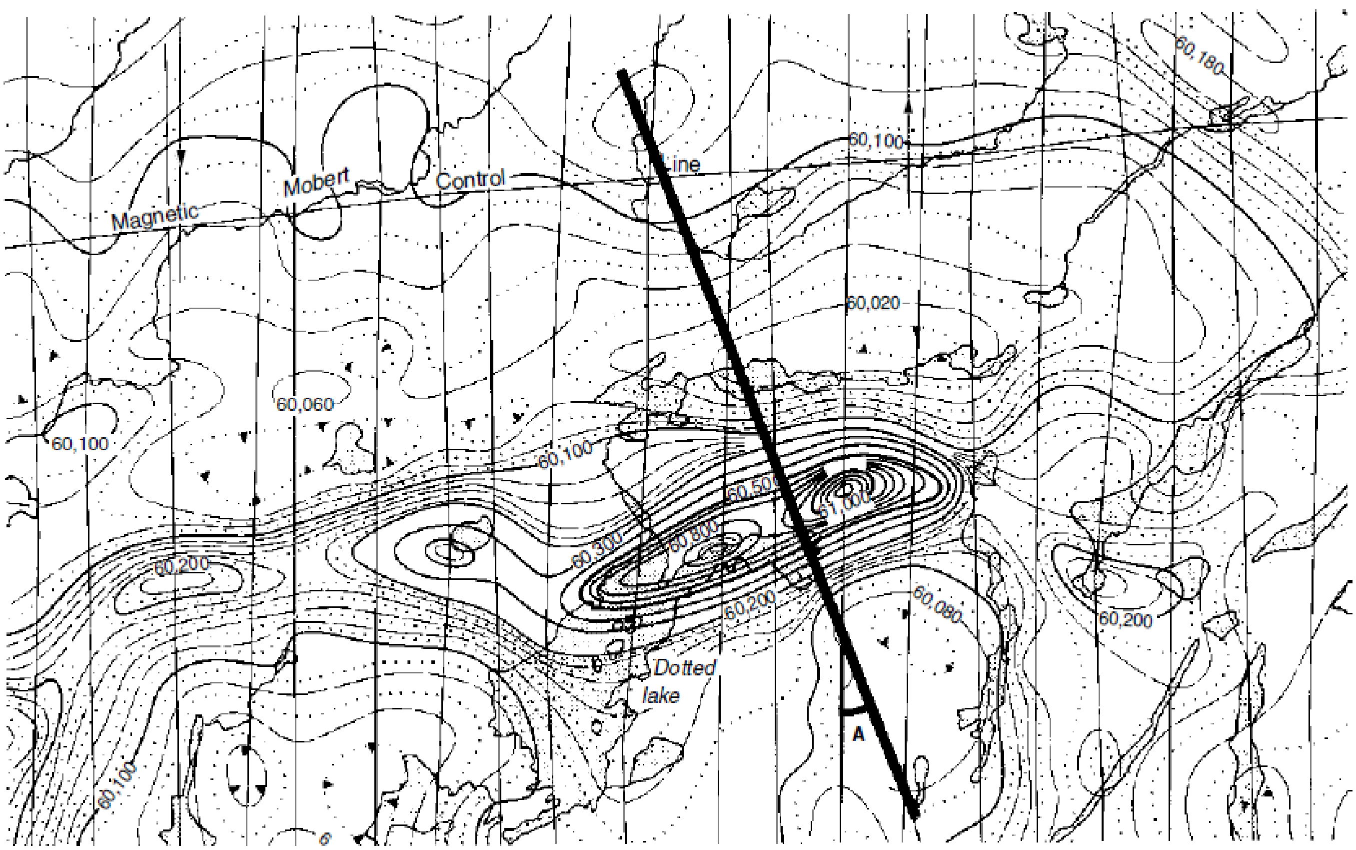
Depth estimation is one of the main objectives of magnetic interpretation. Simple rules give depths to the tops of source bodies that are usually correct to within about 30%, which is adequate for preliminary assessment of field results. In the following Figure(part a) the part of the anomaly profile, on the side nearest the magnetic equator, over which the variation is almost linear is emphasized by a thickened line. The depths to the abruptly truncated tops of bodies of many shapes are approximately equal to the horizontal extent of the corresponding straight-line sections. This method is effective but is hard to justify since there is actually no straight segment of the curve and the interpretation relies on an optical illusion.



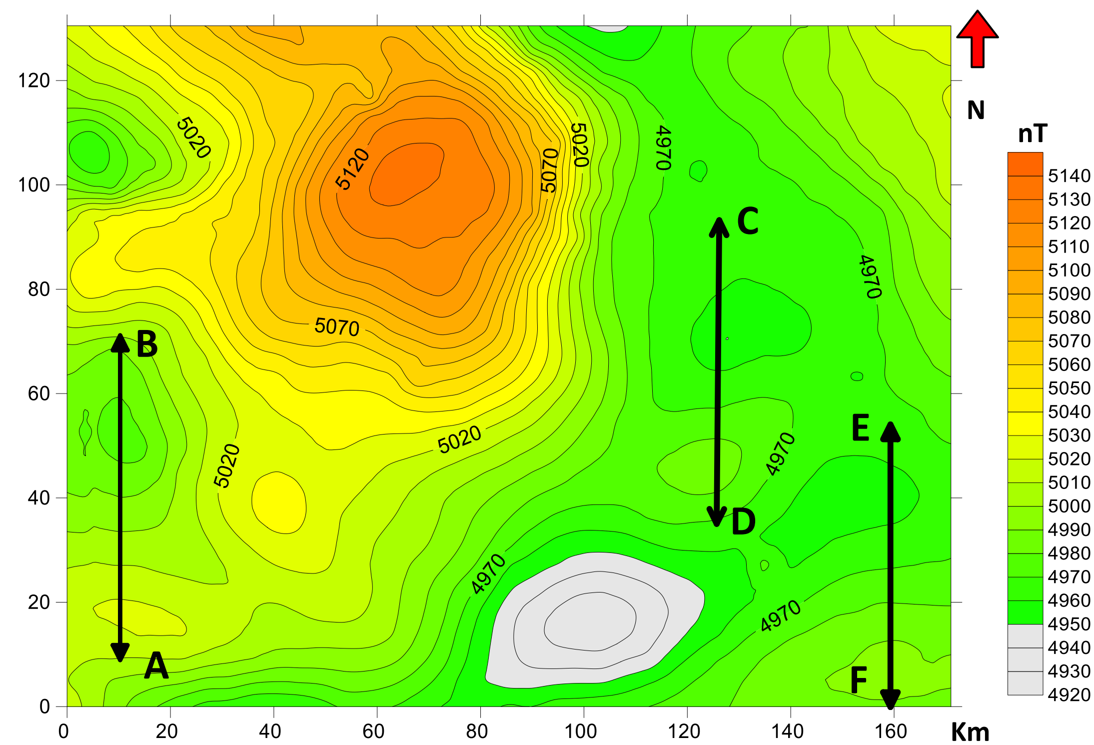
Simple depth estimation: (a) Straight slope method. The distance over which the variation appears linear is (very) roughly equal to the depth to the top of the magnetized body. (b) Peters’ method. The distance between the contact points of the half-slope tangents is (very) roughly equal to 1.6 times the depth to the top of the magnetized body.

In the slightly more complicated *Peters’ method*, a tangent is drawn to the profile at the point of steepest slope, again on the side nearest the equator, and lines with half this slope are drawn using the geometrical construction of the above figure (part b). The two points at which the half-slope lines are tangents to the anomaly curve are found by eye or with a parallel ruler, and the horizontal distance between them is measured. This distance is divided by 1.6 to give a rough depth to the top of the source body.

Peters' method relies on model studies that show that the true factor generally lies between about 1.2 and 2.0, with values close to 1.6 being common for thin, steeply dipping bodies of considerable strike extent. Results are usually very similar to those obtained using the straight slope. In both cases the profile must either be measured along a line at right angles to the strike of the anomaly or else the depth estimate must be multiplied by the cosine of the intersection angle (A in the following figure).



Effect of strike. A depth estimate on a profile recorded along a traverse line (i.e. one of the set of continuous, approximately straight lines) must be multiplied by the cosine of the angle A made with the line drawn at right angles to the magnetic contours. The example is from an aeromagnetic map (from northern Canada) but the same principle applies in ground surveys.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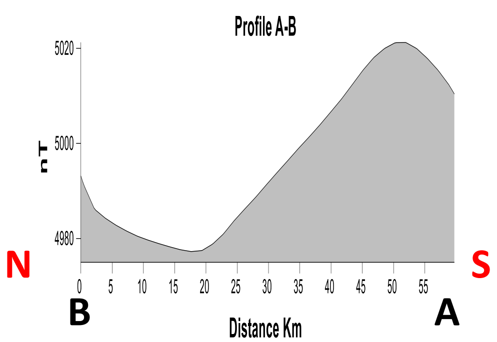
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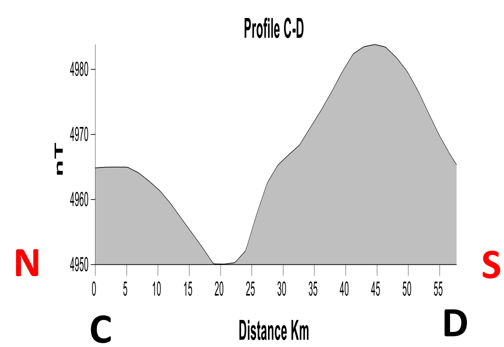
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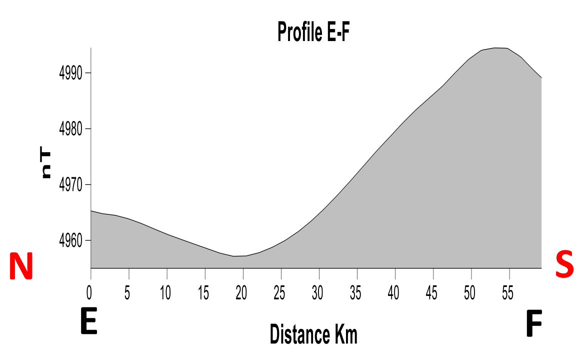
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