National Research Council of Italy

Institute of Geosciences and Earth Resources

EM methods for geothermal exploration

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Geothermal Exploration

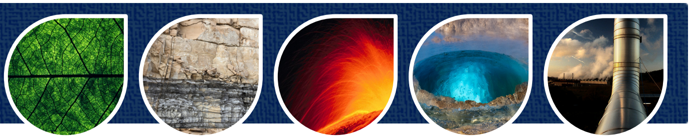
1. EM general
2. Natural Source
3. Controlled Source
4. Direct Current
5. Application

EM geophysical methods

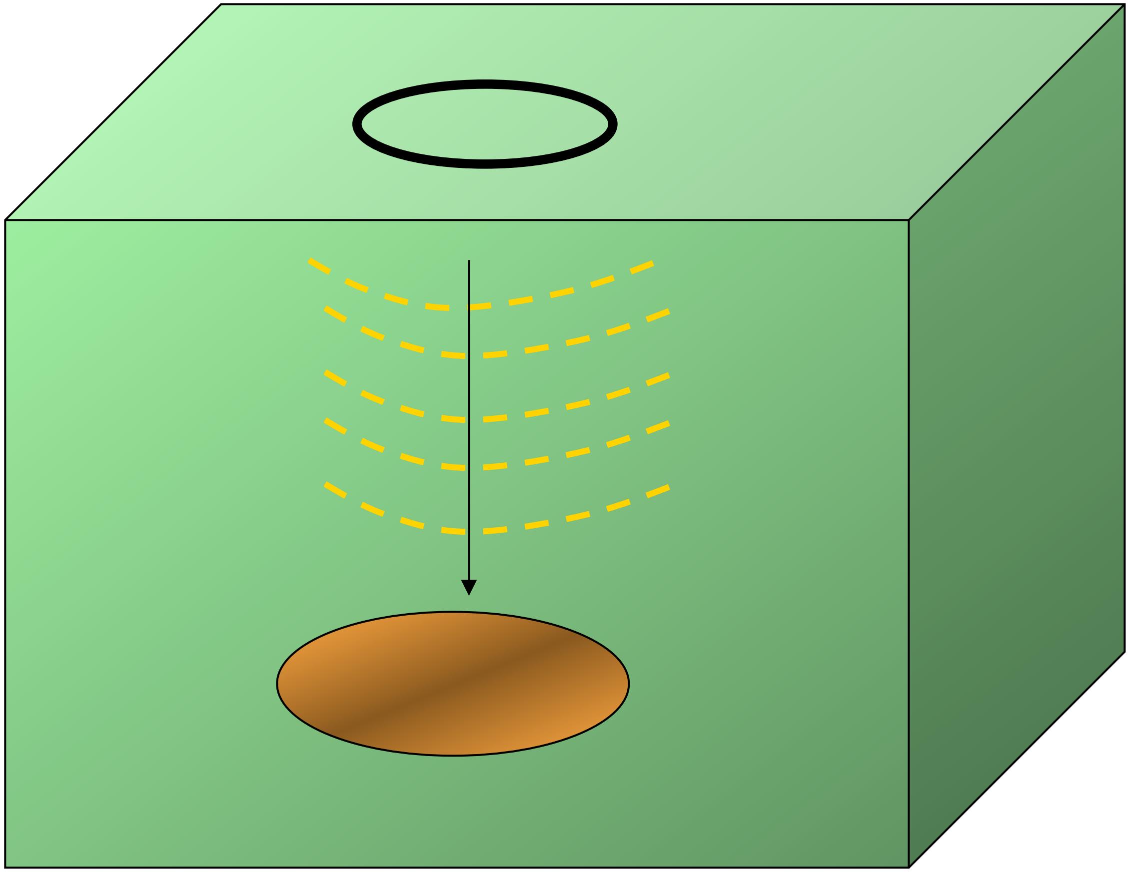
1

Why use them? What do they explore? How?

EM methods generalities



What EM methods are



When an alternate current circulates in an electronic circuit, it induces a magnetic field, and viceversa.

Electromagnetic properties of the medium influences the EM propagation

What EM methods are

natural-source induction methods

(magnetotellurics, audiomagnetotellurics and self-potential)

controlled-source induction methods

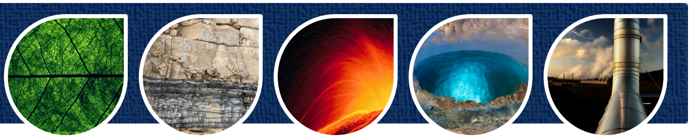
(TDM, VLF)

direct current methods

(SEV, electric tomography)

Their objective is the mapping of **electrical structures** at depths that are meaningful in terms of geothermal exploration.

These depths must be several kilometres at least when looking for the anomaly in conductivity associated with HT geothermal reservoir rocks, and several tens of kilometres when seeking the thermally excited conductive zone associated with the heat source of a geothermal system.

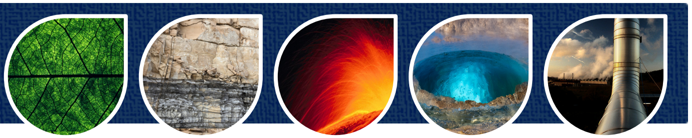


What EM methods are

Inductive methods usually provide information on conductivity-thickness products of conductive layers, whereas they usually provide only thickness information on resistive layers.

On the contrary, **direct current techniques** usually provide information on resistivity-thickness products for resistive layers and thickness products for conductive layers.

For this reason, inductive methods are the most suitable for geothermal exploration, since the target is conductive.

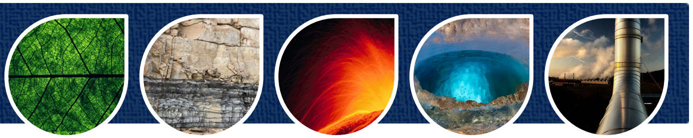


What EM methods are

Electrical current may propagate thanks to the mobility of free charge carriers that allows current **conduction**

Main propagation mechanisms are:

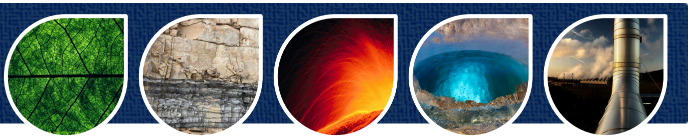
Electronic (<10^{-8} Ωm)	electrons	metals
Semi-conduction $10^{-5} \div 10^{-3}$ Ωm	electrons and ions	Solfurs
Electrolitic	ions	brines, salty water, melts



What EM methods are

Resistivity depends on of both **host rocks** and **pore fluid** properties

Rocks	Temperature & Pressure
	Lithology, Clays (Surface conduction)
	Microstructural properties (e.g., permeability, porosity)
Fluids	Amount
	Nature (liquid or vapor phase, other liquids and gases)
	Salinity

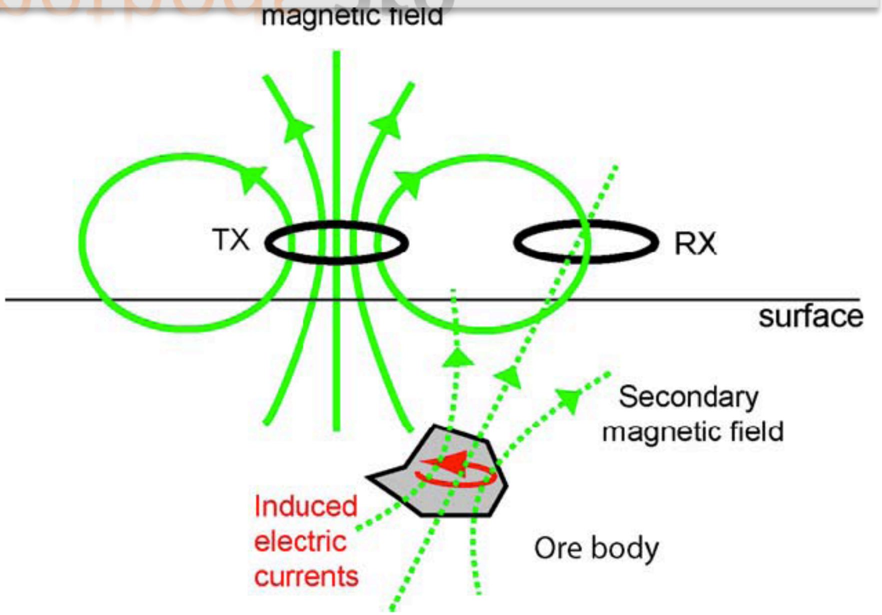


What EM methods are

All **EM inductive methods** have common features:

•a **primary field** (either natural or artificial) incident on the Earth. This can be man made or natural. The geometry can be that of a plane wave or generated by a dipole transmitter (TX). The time variation can be a single harmonic frequency or a pulse;

•a **secondary field** induced by the primary field (eddy currents are induced, amplitude and phase of the wave is changed). To first order, the Earth can be considered a conductor while the air is a resistor.



The **total electromagnetic field that will be measured at the receiver (RX) is** the sum of the primary and secondary fields. Surface (or borehole) measurements of total E and/or H fields are made by placing the RX at a line/grid of points.

These measurements can be made as a function of frequency or time.

2

How does it work? What does it provide?

Magnetotelluric among EM methods

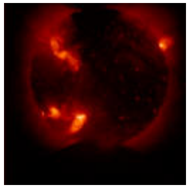
Magnetotelluric among EM methods

Magnetotellurics (MT for short) is a technique which utilizes the earth's naturally occurring electromagnetic field to image the subsurface's electrical resistivity structure.

Natural electromagnetic waves are generated in the earth's atmosphere by a range of physical mechanism:

High frequency signals originate in lightining activity

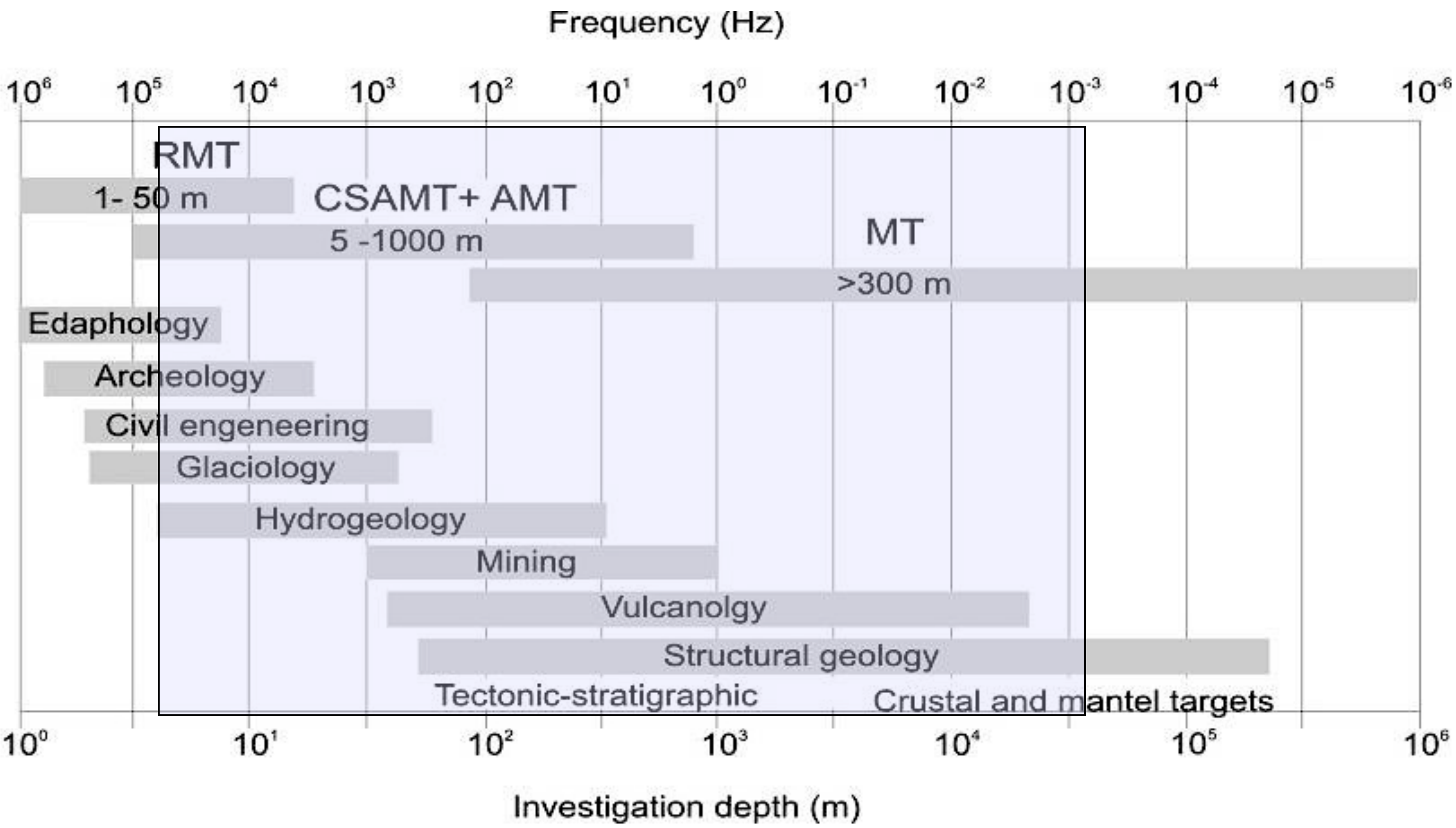
Intermediate frequency signals come from ionospheric resonances

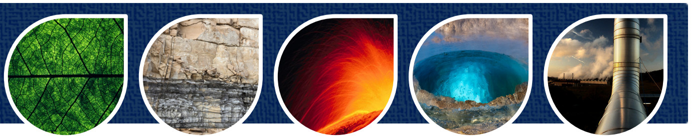


Low frequency signals are generated by sun-spots

Even if the two types of sources create incident EM fields with different features, the almost plane-wave propagates on the vertical inside the ground, due to the large difference of resistivity between atmosphere and earth.

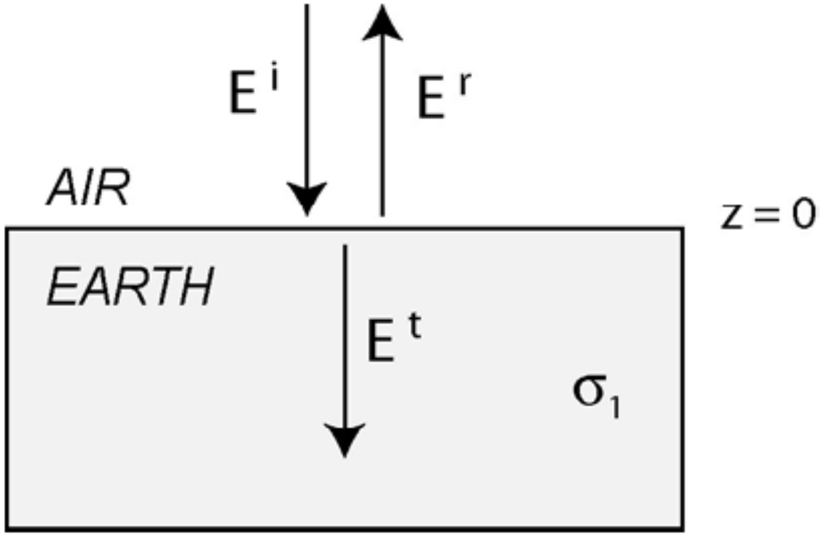
Magnetotelluric among EM methods





Magnetotelluric among EM methods

These electromagnetic waves penetrate the earth and return to the surface bearing information on its electrical resistivity structure.



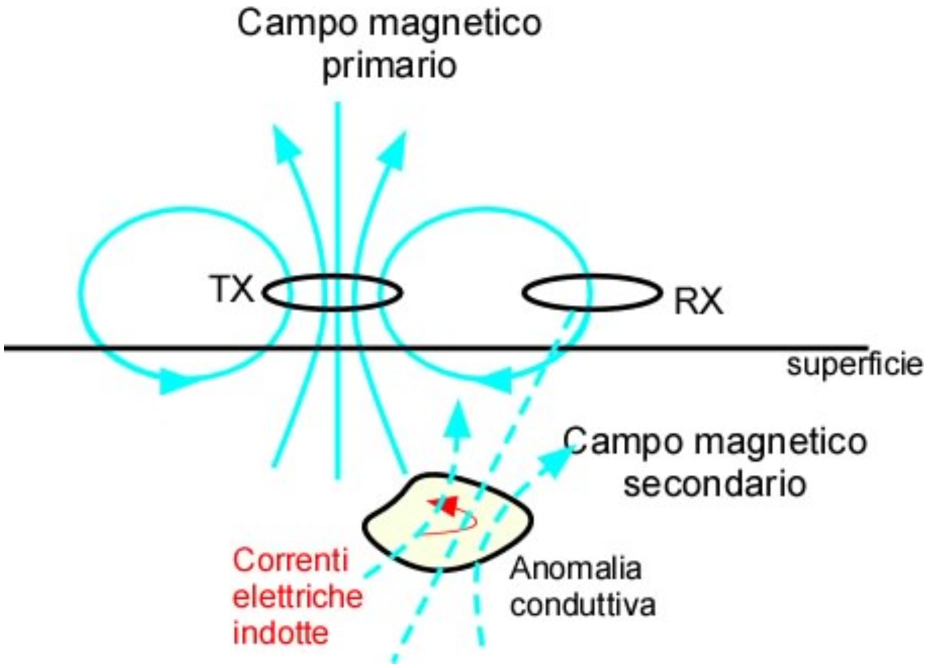
E^i= incident (**primary**) field
E^r= reflected field
E^t= transmitted field

E^t diffuses in the Earth, since Earth's conductivity is large with respect to air's conductivity.
A **secondary** field is generated by eddy currents induced by the transmitted field

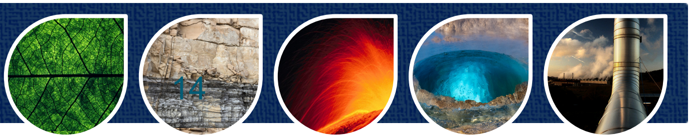
Magnetotelluric among EM methods

The **total** EM field measured at the surface (receiver Rx) is the sum of the primary and the secondary field.

Measuring E and H field at the surface we can retrieve information regarding the underground resistivity structure



Any EM inductive method follows this scheme.
Depending on the method, the fields can be measured as a function of time or of frequency



By some tortuous mathematics it is possible to demonstrate that the ratio between electric (E) and magnetic (H) fields at the earth's surface is independent from the source electromagnetic field, but depends only on the electrical resistivity structure of the subsurface.

This ratio, or transfer function, is called **impedance**

$$|Z_{xy}(\omega)|^2 = \left|\frac{E_x(\omega)}{H_y(\omega)}\right|^2 = \frac{\omega\mu_0}{\sigma_1}$$

E.g., in a uniform earth

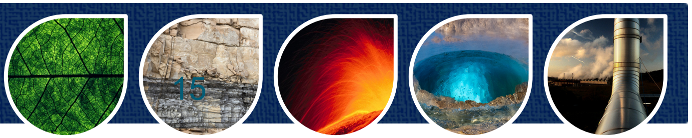
where $\rho = 1/\sigma$

$$Z_{xy} = \omega\mu/k = (1+i)\sqrt{\frac{\rho\omega\mu}{2}}$$

or, solving for ρ,

$$\rho_{xy} = \frac{Z_{xy}Z_{xy}^*}{\mu\omega}$$

where Z^* is the complex conjugate of Z.

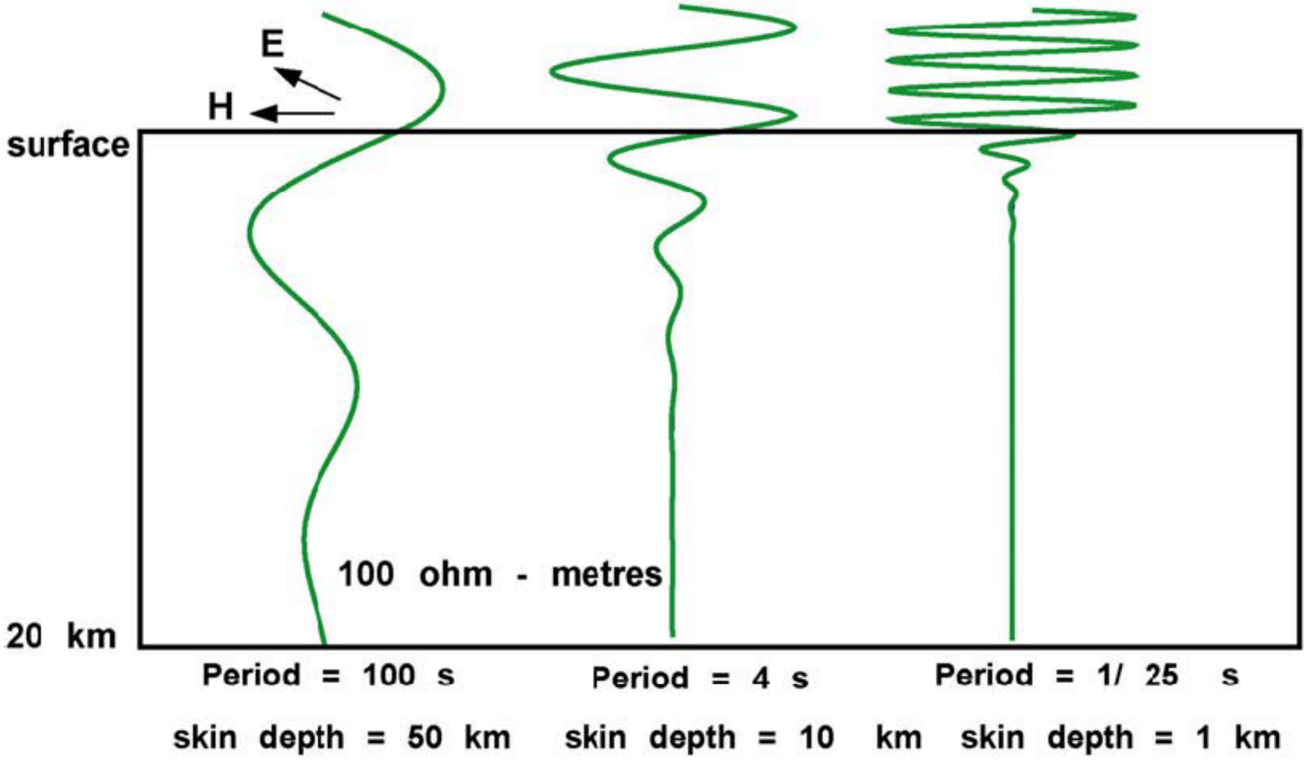


Magnetotelluric among EM methods

As these waves travel into the Earth's interior they decay at a rate dependent upon their wavelengths.

$$\text{Depth of penetration} \;\propto\; \mathrm{sqrt}\,(\text{period} \times \text{resistivity})$$

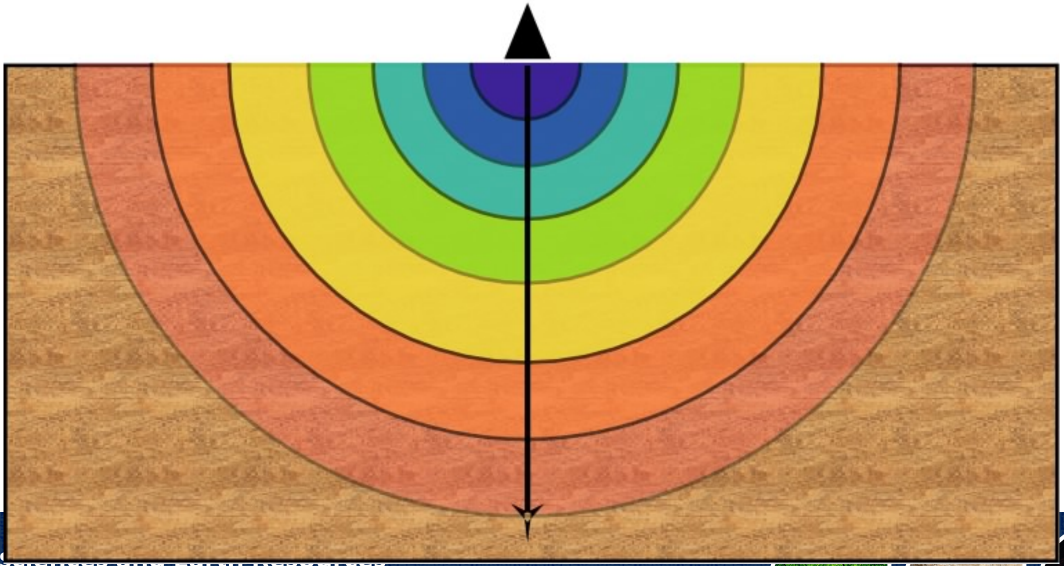
$$\text{Resistivity of ground} \;\propto\; \left(\frac{E}{H}\right)^2$$



Magnetotelluric among EM methods

It is important to consider which part of the Earth is being sampled in such a measurement. Since the EM fields attenuate in the Earth with a length scale of a skin depth (δ), this measurement samples a **hemisphere around the observation site, radius δ.**

Data derive not only from the geometrical-physical features on the vertical of the recording site, but depends also on the later features: this lateral dimension increases with depth (decreases with frequency)



Magnetotelluric among EM methods

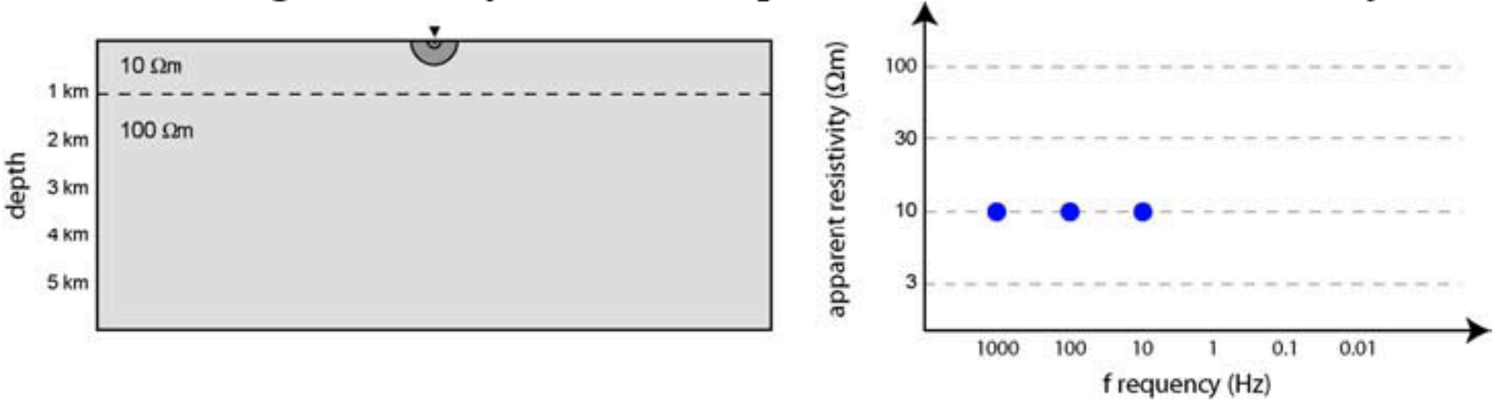
When we pass from homogeneous half-space to 1D, 2D and 3D conditions, the resistivity we compute as a function of impedance is no more the true resistivity of the medium, but an **apparent resistivity**.

Definition: Resistivity of a fictitious homogenous subsurface that would yield the same impedance as the earth over which measurements were actually made

Magnetotelluric among EM methods

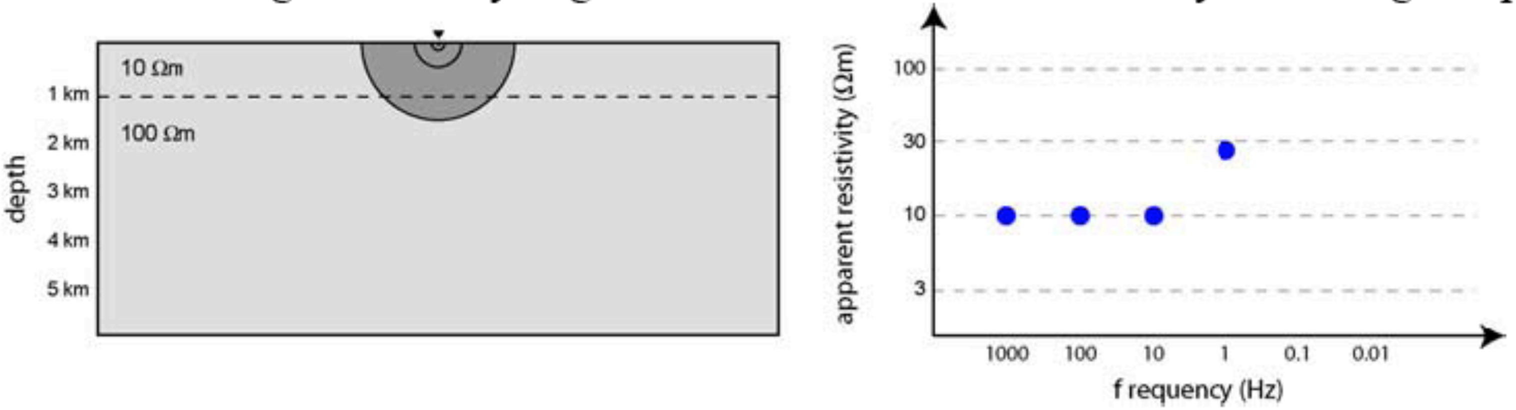
(1) f = 10 Hz δ = 500 m in upper layer

- At high frequency, the skin depth is much less than the thickness of the layer.
- Average resistivity over a hemisphere, radius = δ is the resistivity of the upper layer.



(2) f = 1 Hz δ = 1580 m in upper layer

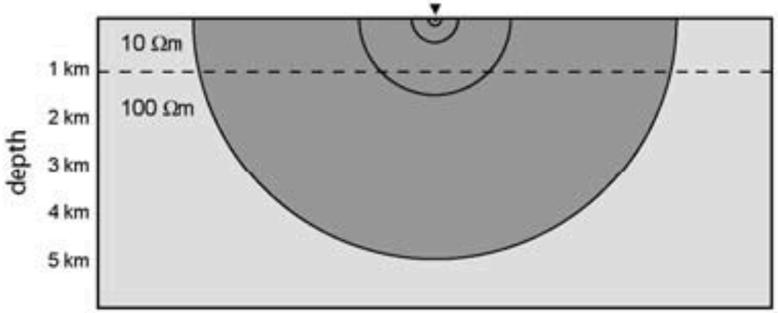
- The EM signals are now just sampling the lower layer,
- Average resistivity begins to increase as the 100 Ωm layer is being sampled.

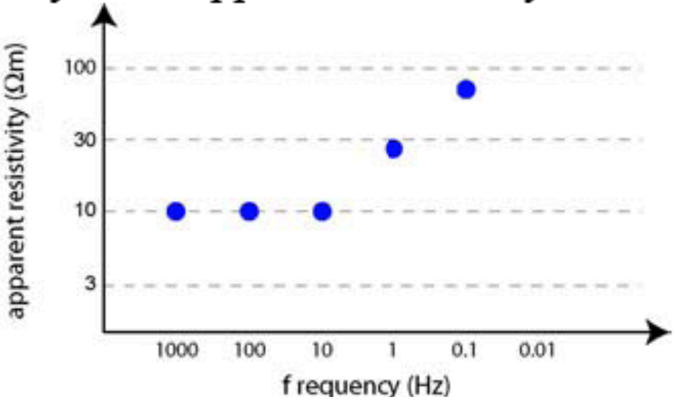


Magnetotelluric among EM methods

(3) f = 0.1 Hz δ = 5000 m in upper layer

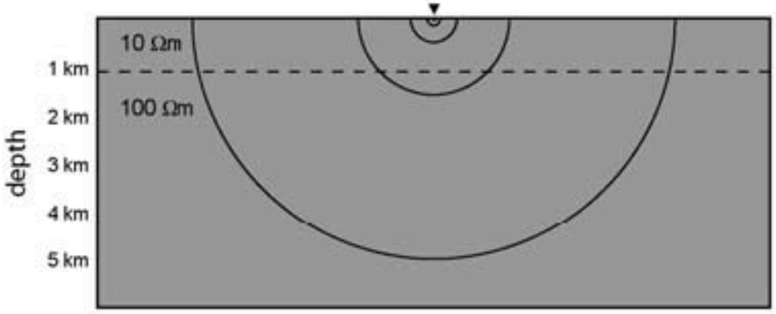
- Hemisphere dominated by lower layer, so apparent resistivity close to 100 Ωm

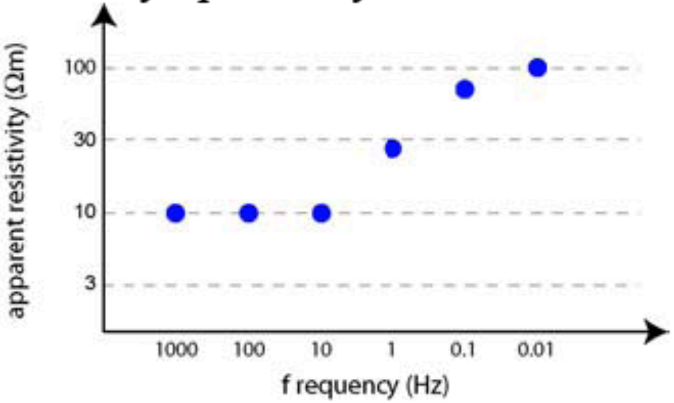




(4) f = 0.01 Hz δ = 15800 m in upper layer

- Apparent resistivity approaches 100 Ωm asymptotically

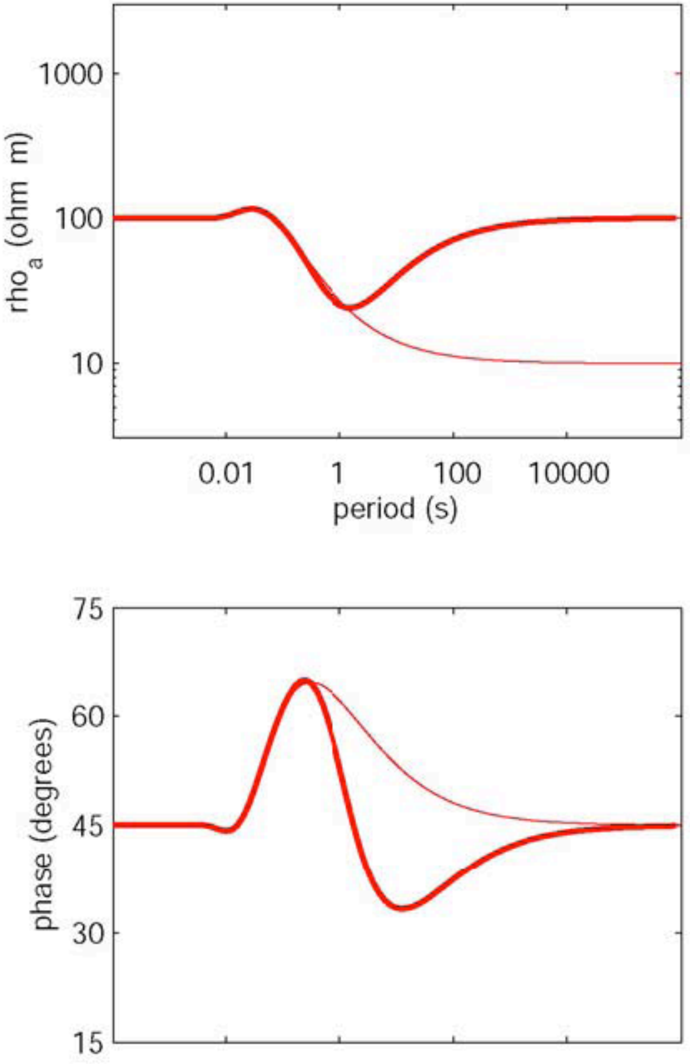


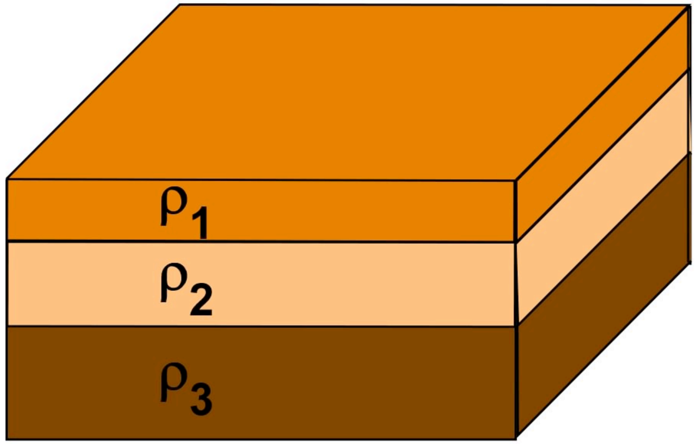


Magnetotelluric among EM methods

In a three layers model, ρ_a is equal to ρ_1 at high frequency and goes exponentially to ρ_3 at low frequency. In the middle it approaches ρ_2.

The shape, and therefore our ability to detect differences, depends very much on the relative resistivity and thickness of the layers, on the measured frequency and data quality.

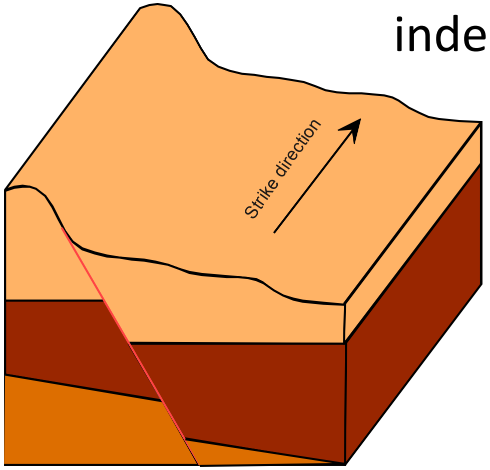


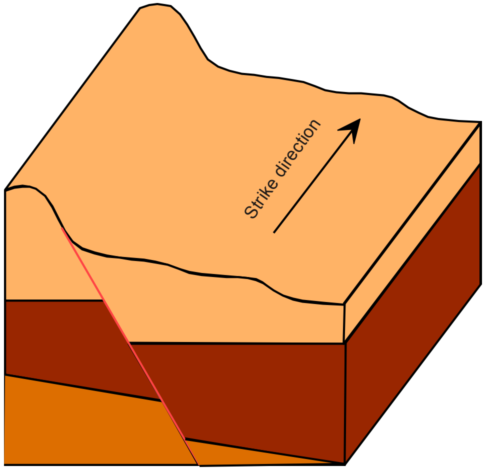


Magnetotelluric among EM methods

2-D case

By Maxwell equation we can derive equations can now be separated into two independent subsets:





Electric field polarized parallel to the strike direction.

Magnetic field components are confined to the *y-z plane.*

This is called the

Transverse Electric (TE) mode or E-polarization

Magnetic field polarized parallel to the strike direction.

Electric field components are confined to the *y-z plane*

This called the

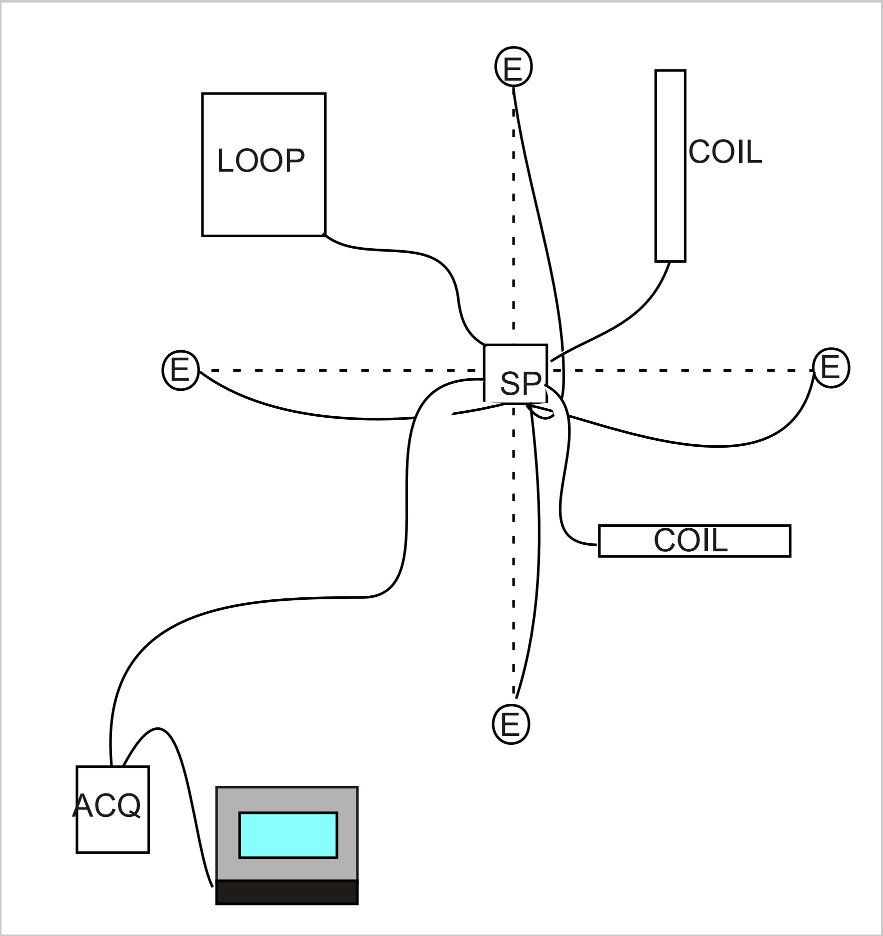
Transverse Magnetic (TM) mode or B-polarization

Magnetotelluric among EM methods

MT data are acquired in the field, as measurements of electric and magnetic fields with time
$E_x(t)$, $E_y(t)$, $H_x(t)$, $H_y(t)$, $H_z(t)$

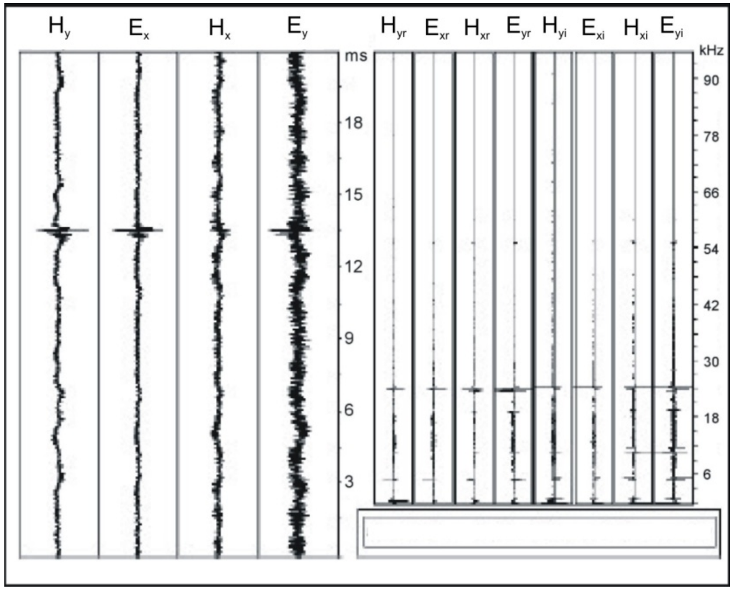
Timing is obtained from GPS time signals.

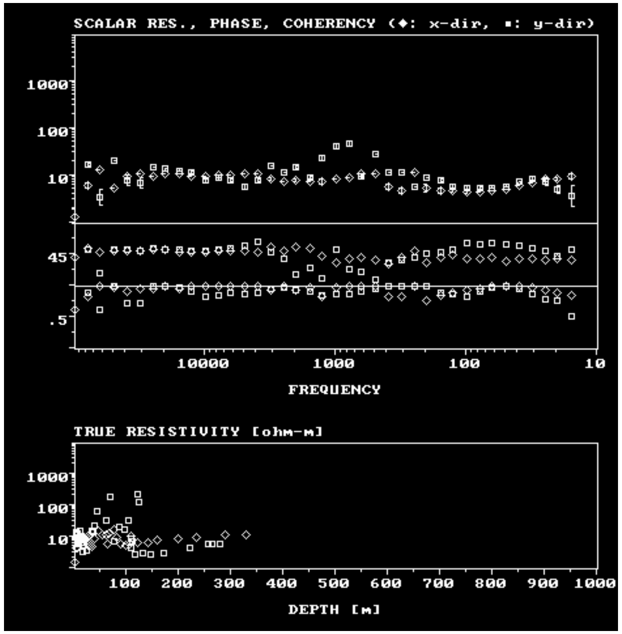
Care must be put on the choice of the site, trying to avoid possible noise sources, such as power lines, electrified railways, pipelines.



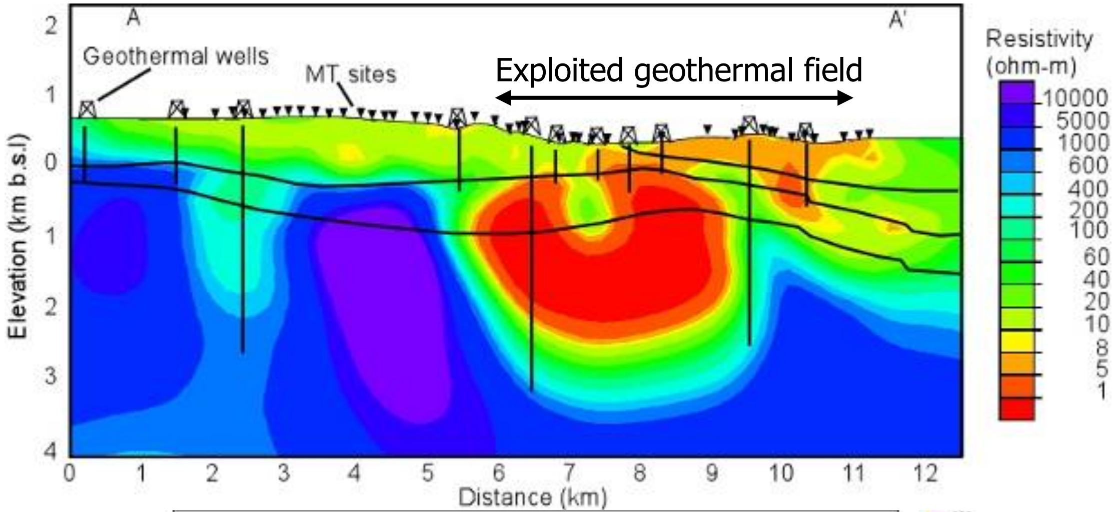
Magnetotelluric among EM methods

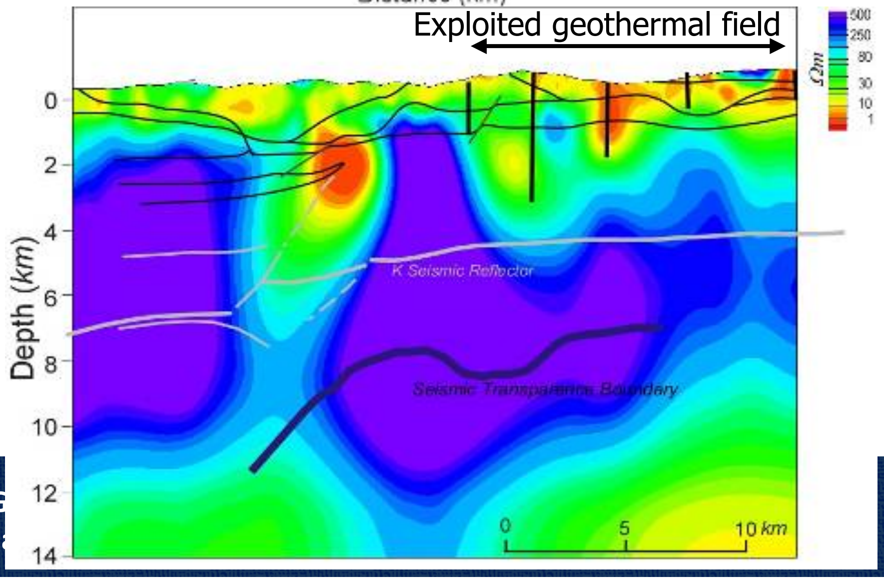
When the analysis is done in the field (real time analysis), decisions about re-occupying stations and siting additional stations can be made in a timely manner that will reduce overall operating costs.





Magnetotelluric among EM methods

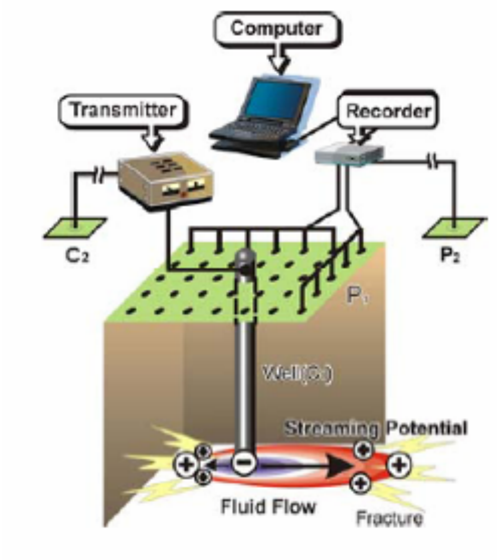




Magnetotelluric data, after processing and modelling, provide the resistivity distribution at depth of various km.

Example: The correspondence between areas of low resistivity inside the resistive basement and geothermal reservoirs was very evident in the Mt. Amiata water-dominated system. See also correspondence with fault position *(from Manzella)*.

Self-potential among EM methods



In **Self-potential** (SP) method only the naturally existing voltage gradients in the earth are measured

Causes of these natural voltages:

- oxidation or reduction of various minerals by reaction with groundwater
- generation of Nernst voltages where there are concentration differences between the waters residing in various rock units
- streaming potentials, occurring when fresh waters are forced to move through a fine pore structure, stripping ions from the walls of the pores

In geothermal areas, very large self-potential anomalies have been observed, and these are apparently caused by a combination of thermoelectric effects and streaming potentials

Self-potential among EM methods

An example of
Self-Potential data acquired
at Hatchobaru geothermal
field, Japan (from Ushijima
et al, WGC2000)

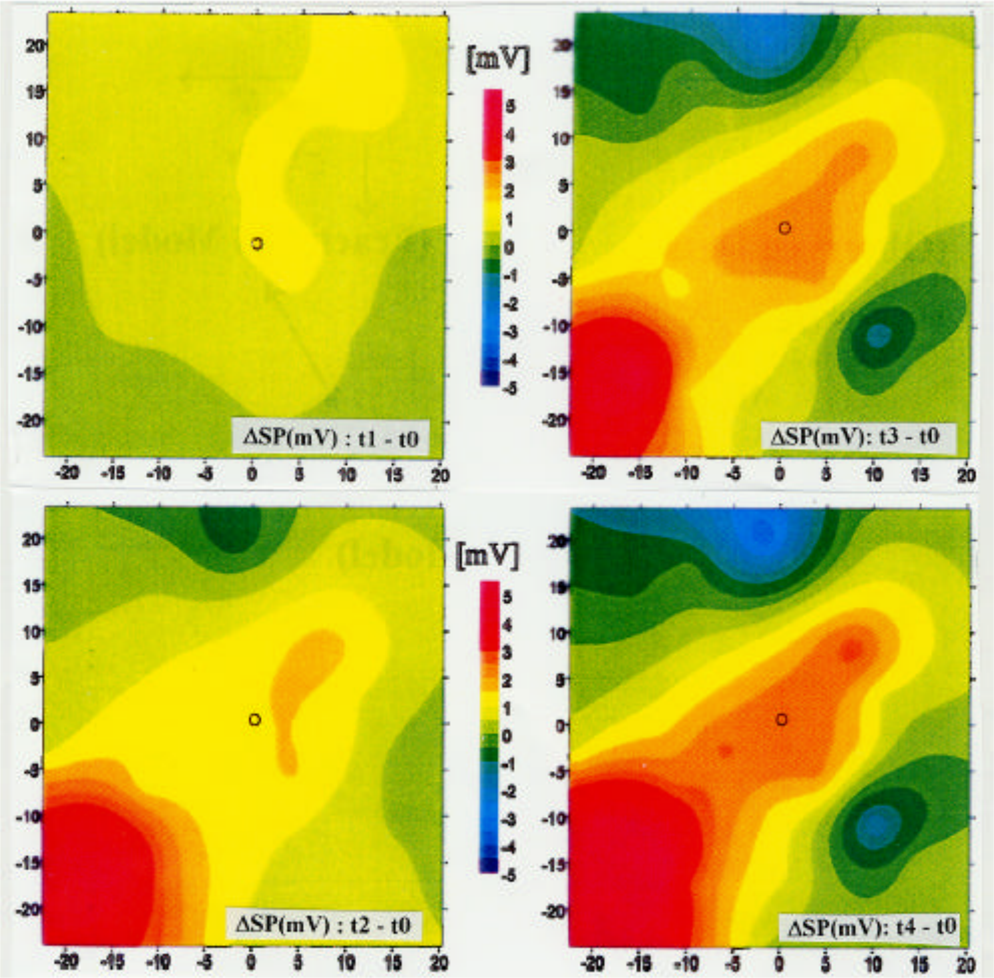
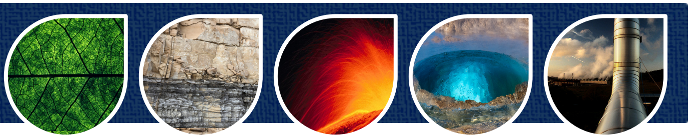


Figure 6. Residual SP variation with a function of time observed during water injection into H-28 exploratory borehole in Hatchobaru geothermal area.

3

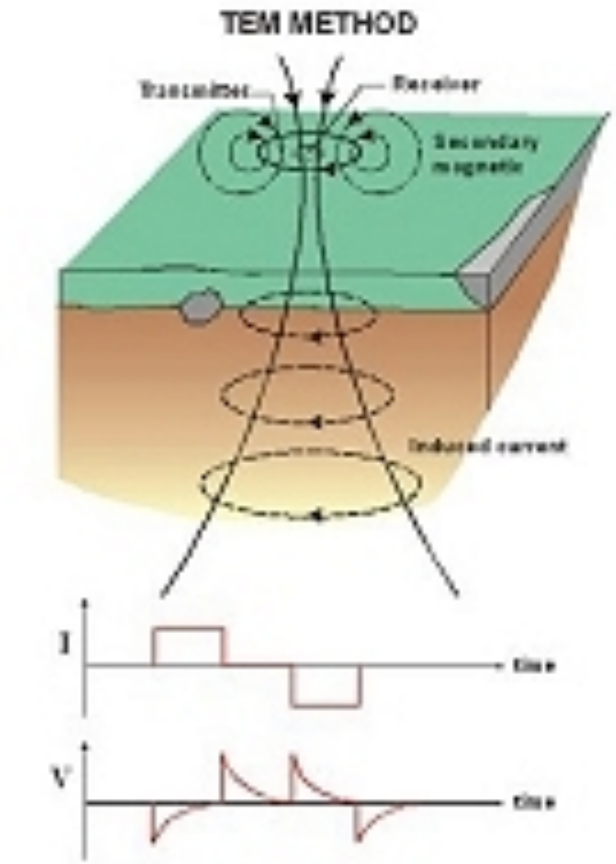
How does it work? What does it provide?

Controlled source among EM methods



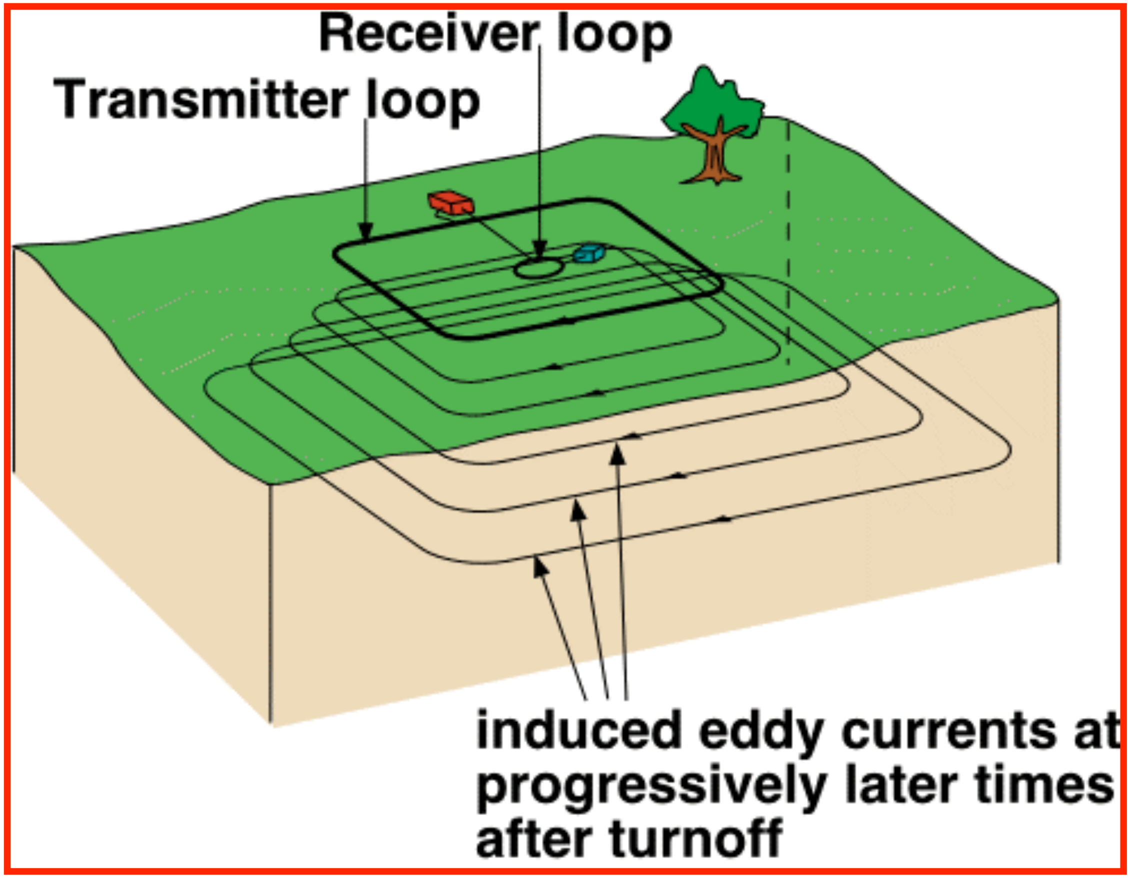
CSEM among EM methods

Active or Controlled Source electromagnetic (CSEM) methods are used mainly for shallow depth resistivity studies and to help with static shift corrections of MT data. Most commonly **central loop TEM** is used, which is based upon inducing currents in the ground electro-magnetically via a loop laid on the surface. The loop has a square shape, each side measuring several hundred meters. A magnetic spool is placed at the centre of the square, after which DC current is applied to the loop.



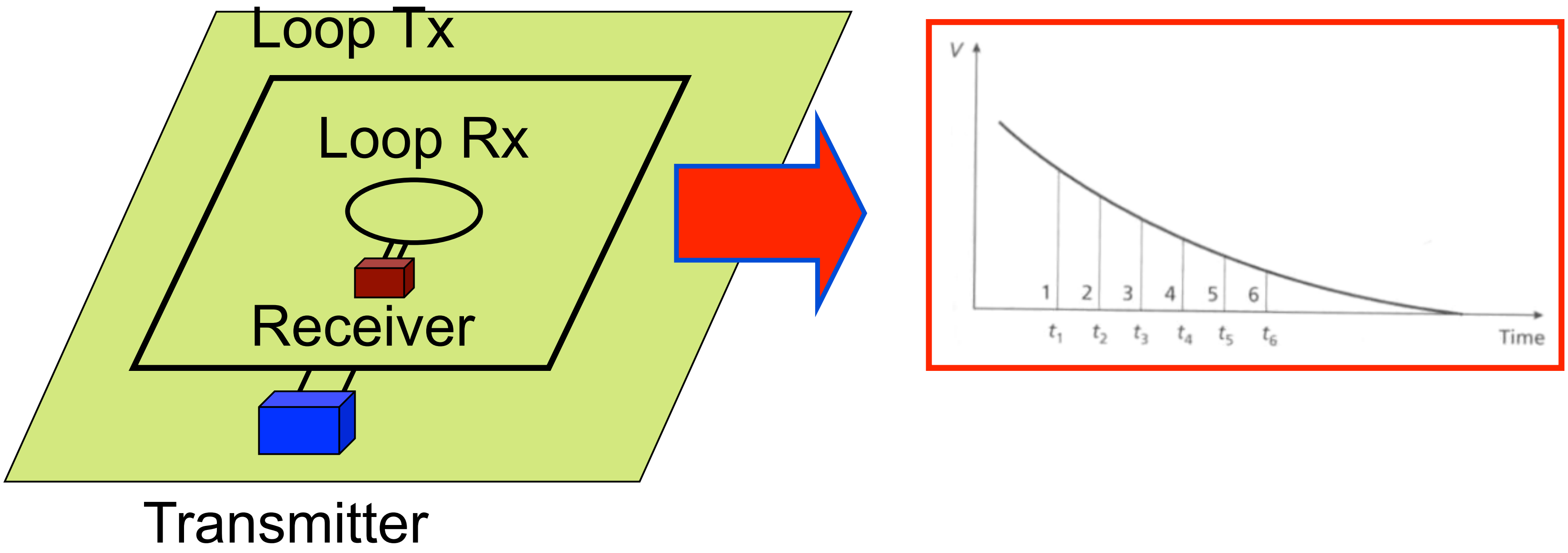
CSEM among EM methods

The produced primary EM field diffuses at depth and interacts with electric conductivity structures.



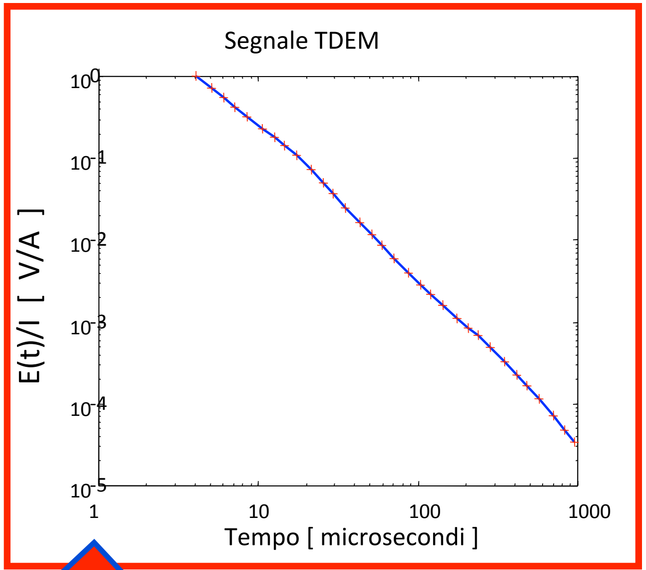
CSEM among EM methods

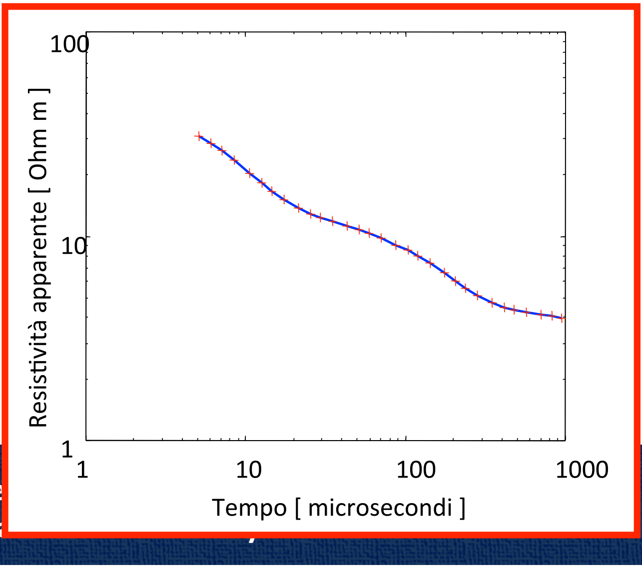
The current is then interrupted and the transient of EM force induced from the secondary magnetic field at the receiver is recorded. Readings are done from the turn-off at fixed intervals during the decay of the secondary magnetic field as it approaches zero, the last ones reaching the deepest structures.



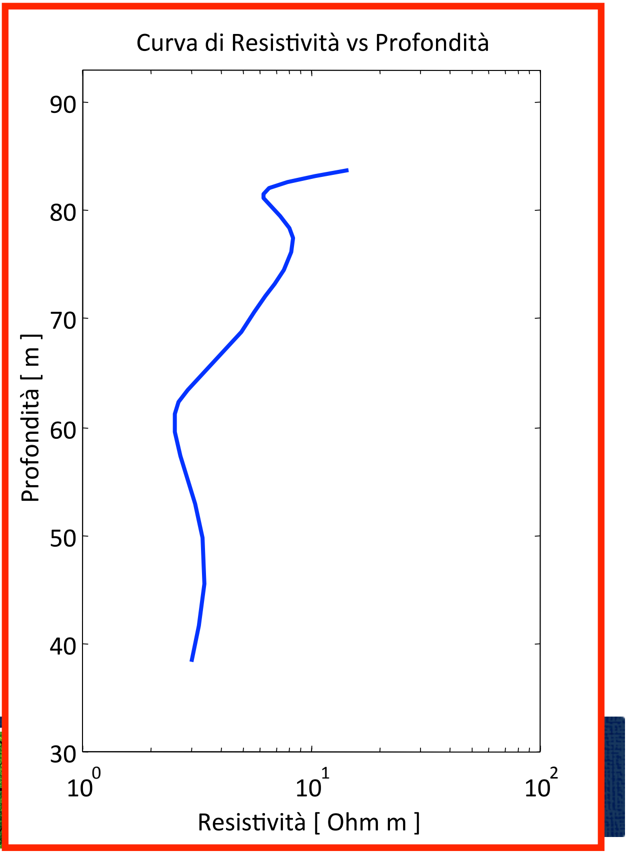
CSEM among EM methods

Apparent resistivity as a function of time is then transformed in an effective resistivity at an approximate depth, producing the resistivity distribution at depth on the vertical of the site

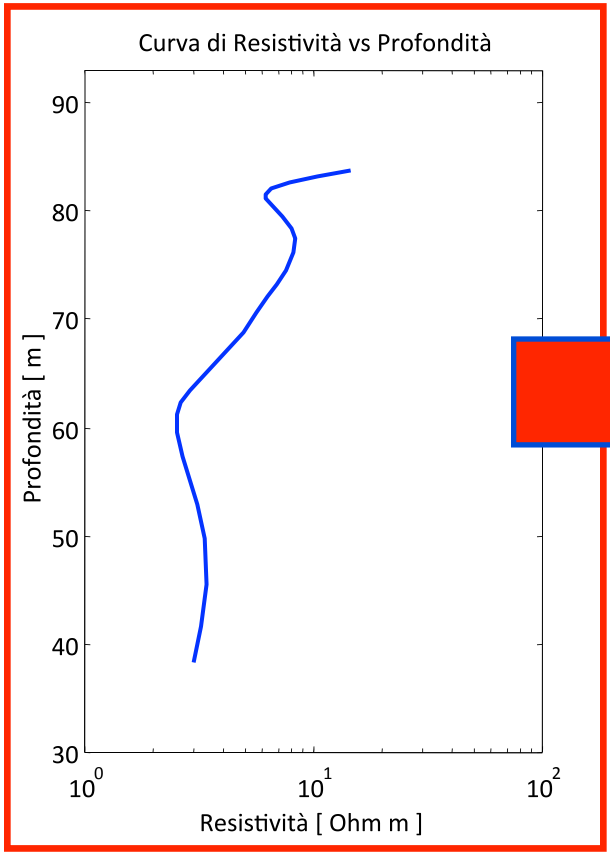


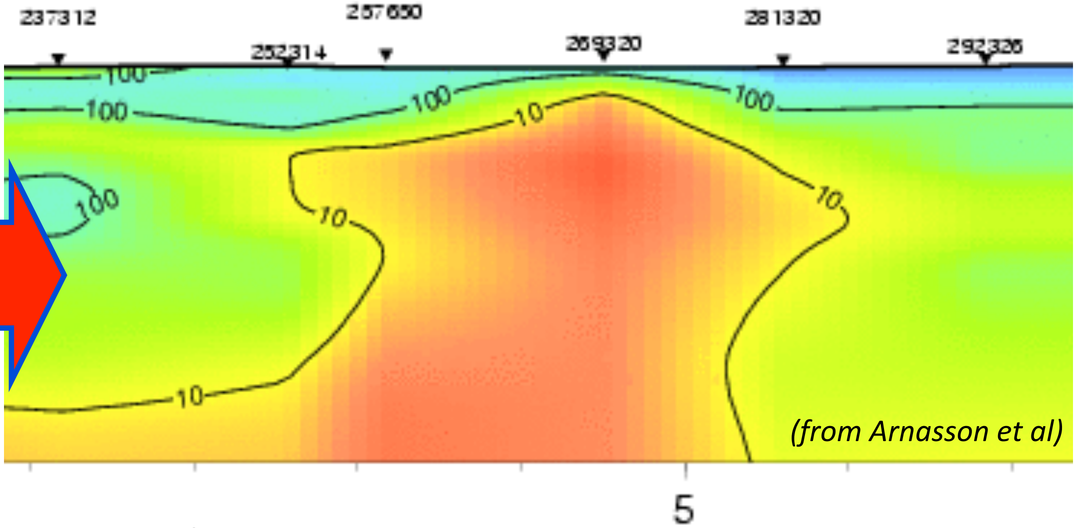


Niblett-Bostick transformation



CSEM among EM methods





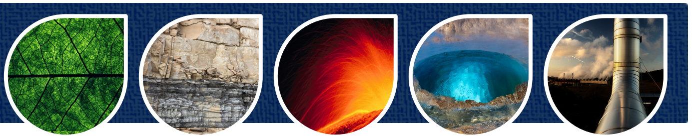
(from Arnasson et al)

As for MT, TEM provide resistivity distribution at depth.
By interpolation of 1D models along profiles, it is possible to obtain 2D and 3D resistivity distribution at depth.

4

How does it work? What does it provide?

Direct current among EM methods



DC among EM methods

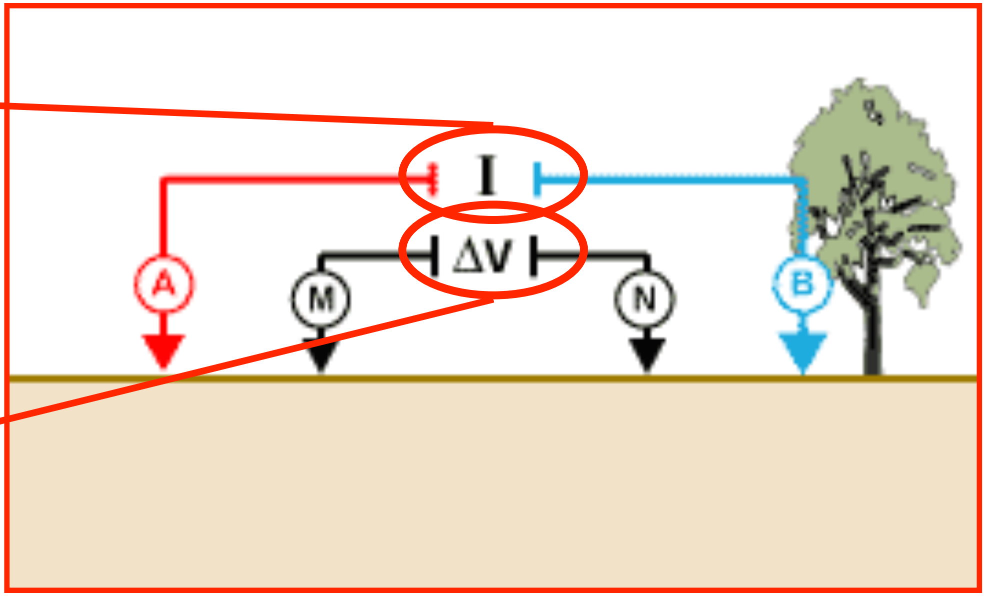
The **direct current resistivity** (DC or electrical) method comprises a set of techniques for measuring earth resistivity that are significantly simpler in concept than the magnetotelluric method.

The magnetotelluric method is an induction method in which the depth of penetration of the field is controlled by the frequency of the signals analysed.

The direct current methods achieve control of the **depth** of the penetration by regulating the **geometry of the array** of equipment (transmitter and receivers relative position).

DC among EM methods

In practice current is injected through a couple of electrodes (usually named A and B) and the voltage difference is measured between two other electrodes (usually named M and N)

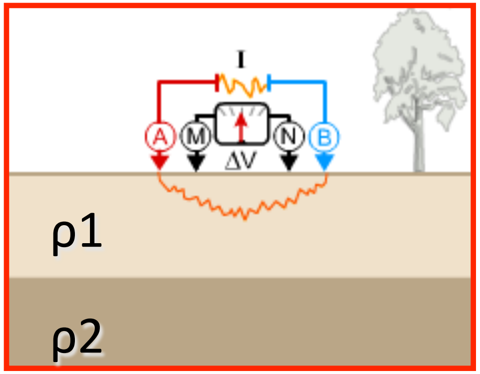


Injected current

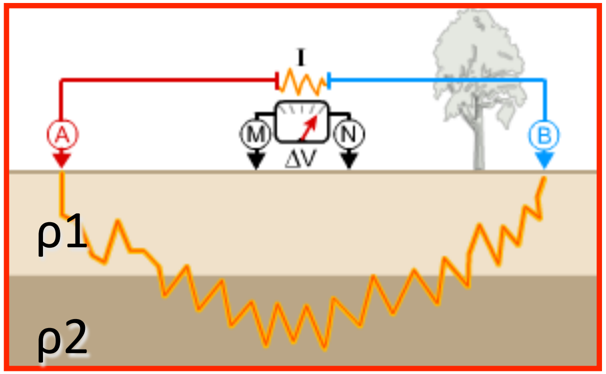
Resulting voltage difference

DC among EM methods

Short electrode distance (AB): the current is essentially confined to the shallow layer (resistivity ρ1) and the apparent resistivity value is related only to ρ1.



For increasing AB distance, a progressively larger current portion flows in the deeper layer with resistivity ρ2 and the apparent resistivity is more and more influenced by ρ2.



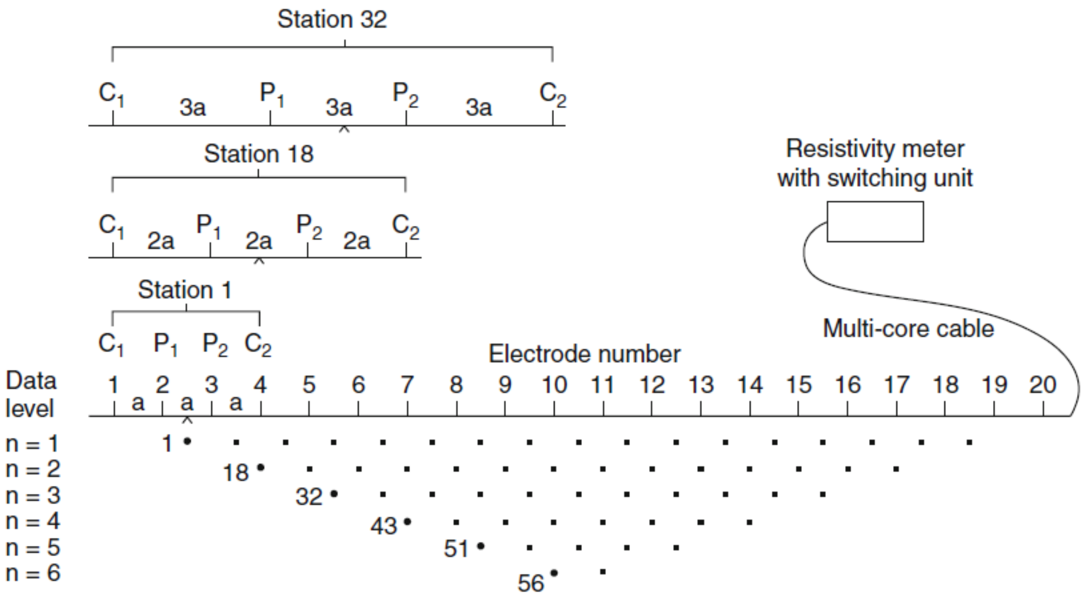
DC among EM methods

Before early 1990s, the electrical resistivity method was mainly used in resistivity sounding, profiling, and mapping surveys and quantitative interpretation was mainly confined to 1-D (one-dimensional) structure of the subsurface consisting of horizontal layers.

DC among EM methods

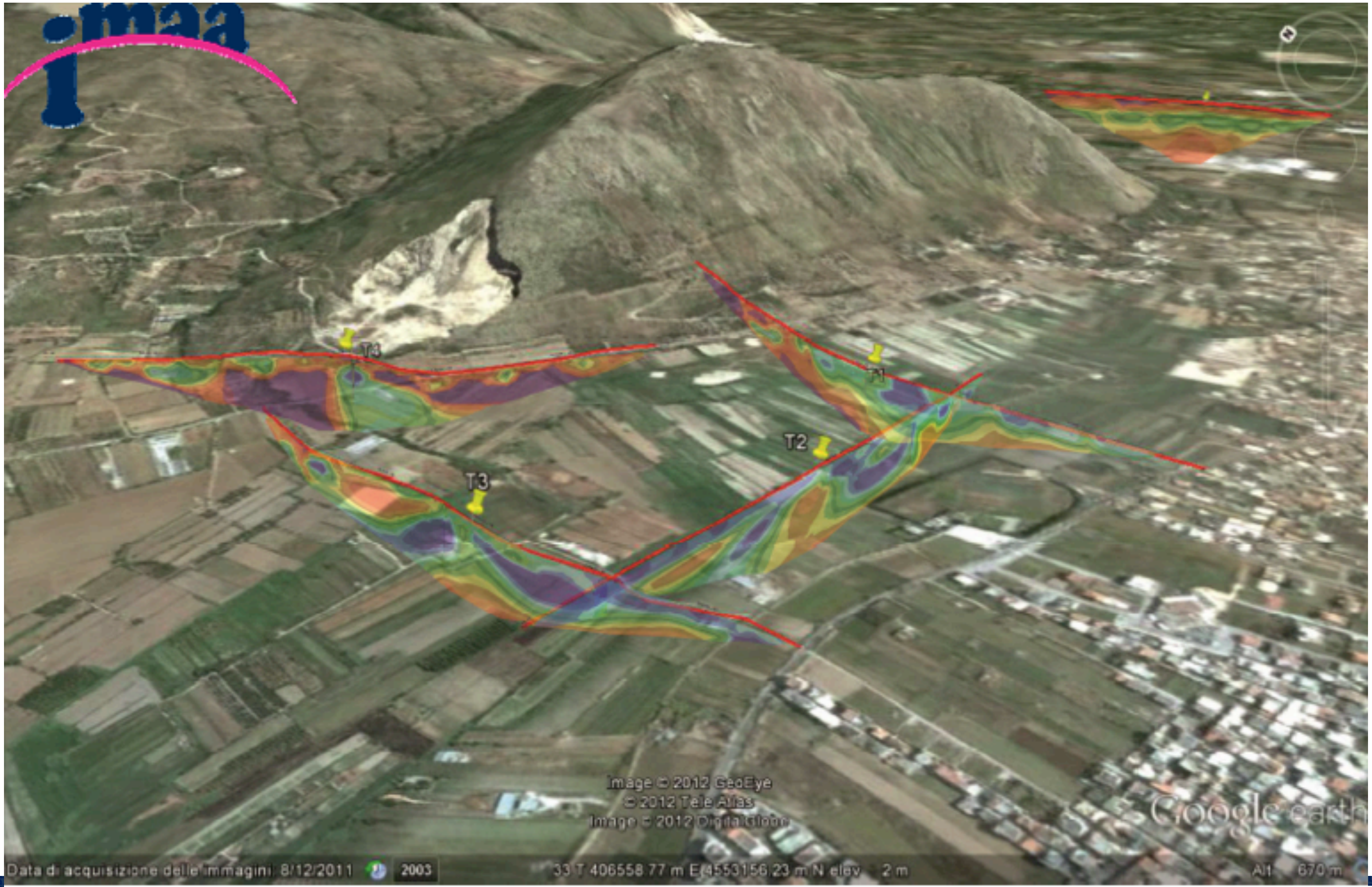
The development of multielectrode and multichannel systems over the past 20 years has sparked a revolution in resistivity surveying. The advent of 2-D and 3-D (three-dimensional) **resistivity tomography** has opened up whole new application areas to electrical methods.

The multielectrode systems made it practical to carry out 2-D imaging surveys that give a more accurate picture of the subsurface in a routine manner.



Instrumentation, Electrical Resistivity, Figure 4 Schematic diagram of a multielectrode system used for a 2-D electrical survey and an example sequence of measurements used to build up a pseudosection using the Wenner array.

DC among EM methods

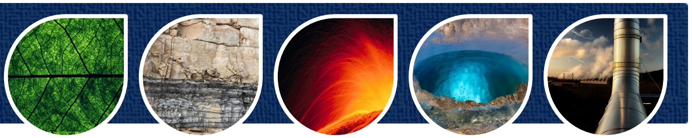


(from Rizzo et al. 3013)

IP among EM methods

Induced polarization (or IP) is a secondary measurement that can be made at the same time as DC resistivity if the correct equipment is included. IP have been known for a long time, but sparsely used, mainly confined to mineral (ore) exploration.

IP measurements respond to variations in the capacity for subsurface materials to retain electric charge. This physical property is referred to as **chargeability**. The principal materials that exhibit this property are clays, graphite, and sulphide minerals. However, small changes in chargeability can be detected when groundwater is contaminated with salt, hydrocarbons, or other materials.

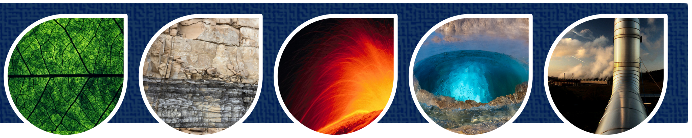


IP among EM methods

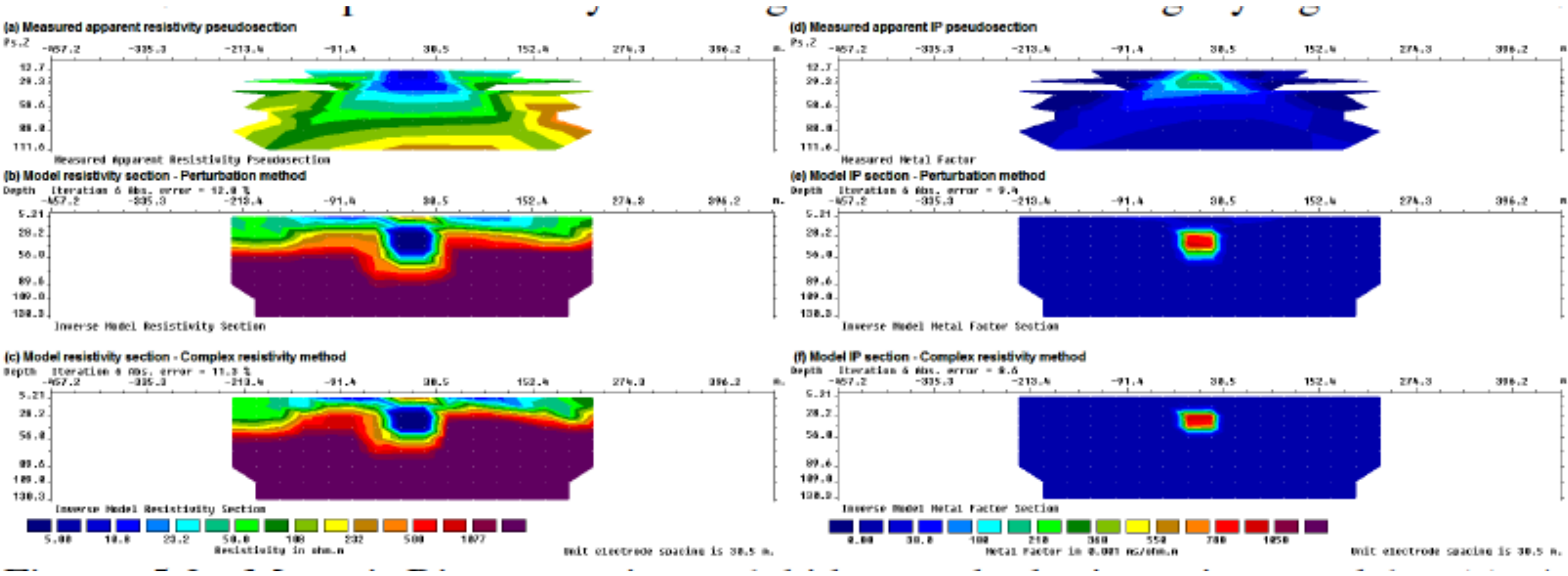
With modern equipment, acquisition and modelling of IP data has been made quick, and IP data are now used to hellp in ERT data interpretation, since it is possible to distinguish the effects due to clays, since the chargeability of clays (in the 10 to 50 mV/V range) is much smaller that that due to conductive minerals.

The IP effect is caused by two main mechanisms, the membrane polarization and the electrode polarization effects. The membrane polarization effect is largely caused by clay minerals present in the rock or sediment.

IP measurements are made in the time-domain or frequency domain.



IP among EM methods



(a) Measured apparent resistivity pseudosection

(d) Measured apparent IP pseudosection

(b) Model resistivity section - Perturbation method

(e) Model IP section - Perturbation method

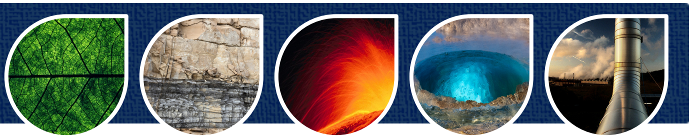
(c) Model resistivity section - Complex resistivity method

(f) Model IP section - Complex resistivity method

5

What can be investigated?

Applications of EM methods



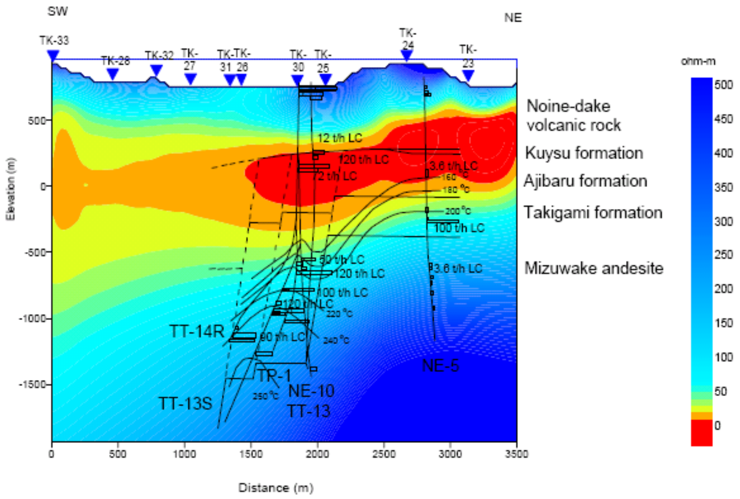
Application of EM methods

Various targets can be imaged by resistivity geophysical methods

- Regional structure (geothermal system)
- Fracture detection
- Monitoring

Application of EM methods at regional scale

Takigami Geothermal Area, Japan



From Ushijima et al., WGC 2005

"the low resistivity zone in the northeastern part is intensive and shallower than that in the southwestern par,t in good agreement with the geological feature"

Advantages

- cheap
- recognize fluid filled volumes

Disadvantages

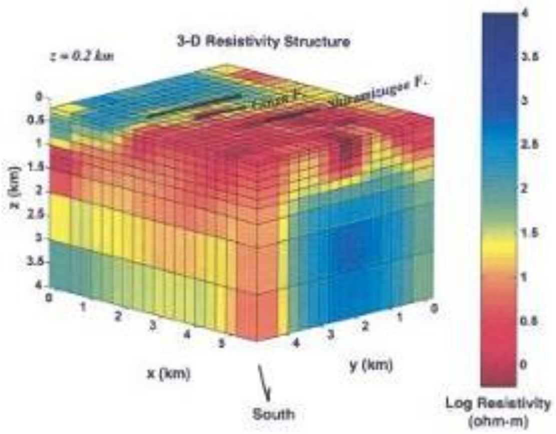
- difficulty to distinguish alteration clays from actual fluid circulation (frozen condition)
- poor geometrical resolution (volume sounding). Improved with dense spacing

Application of EM methods for fracture and fault detection

Advantages
- cheap
- resistivity changes are strong
- EM strike direction may define azimuth

Disadvantages
- low geometrical resolution (lateral resolution improved when using short site spacing)

Ogiri geothermal zone, Japan



From Uchida, 2005

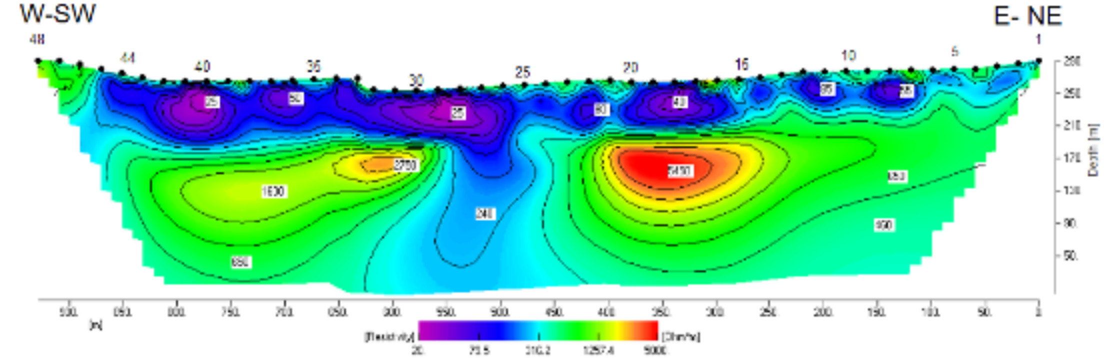
3-D view of the resistivity model, from south. Shallow blocks to a 200m depth are stripped out and approximate locations of three faults are overlaid.

Application of EM methods for monitoring

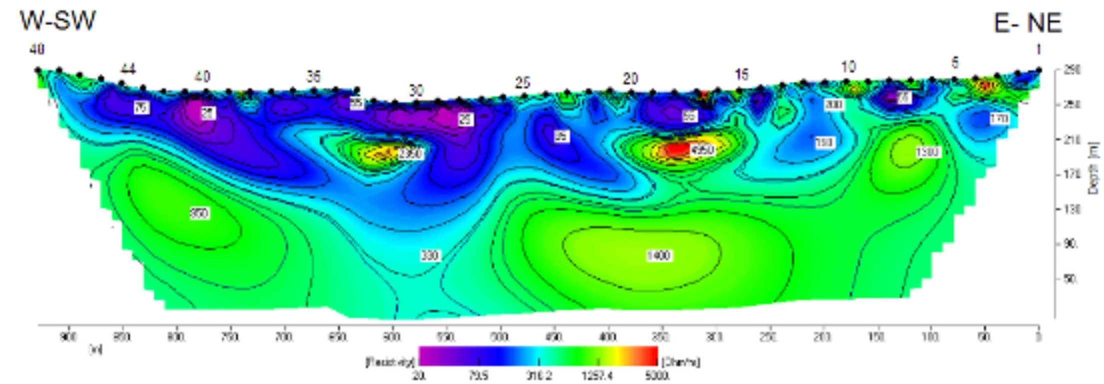
Geothermal exploration at a low enthalpy field using ERT. The low resistivity zone is coincident with a known fault where warm and saline fluids mix with surface and fresh water. An example of monitoring the effect on resistivity change when fresh water is pumped out from a well at the center of profile: the increase of salinity and temperature in the subsurface decreases the resistivity

25 Novembre 2008

Polo-Dipolo-rho



26 Novembre 2008



Thank you for your attention

Jigokudani Hot Springs – Japan

It is not wise to define a particular sequence of geophysical surveys as being applicable to all potential reservoirs

Physical property **Target**	Density	Magnetic susceptibiity	Electrical resistivity	Dielectric permittivity	Seismic velocity
Porosity	🟥	⬜	🟥	🟩	🟩
Permeability	⬜	⬜	🟦	🟦	🟦
Water content	🟩	⬜	🟥	🟥	🟩
Water quality	⬜	⬜	🟥	⬜	⬜
Clay content	🟦	⬜	🟥	🟦	🟩
Magnetic mineral content	🟩	🟥	🟦	⬜	⬜
Metallic mineral content	🟥	⬜	🟥	⬜	🟩
Mechanical properties	🟩	⬜	🟩	🟦	🟥
Subsurface structure	🟩	🟩	🟩	🟥	🟥

🟥	🟩	🟦	⬜
Strong	**Moderate**	**Weak**	**None**

Degree of relationship

A geothermal system generally causes inhomogeneities in the physical properties of the subsurface, which can be observed to varying degrees as anomalies measurable from the surface.

Changing physical parameters:

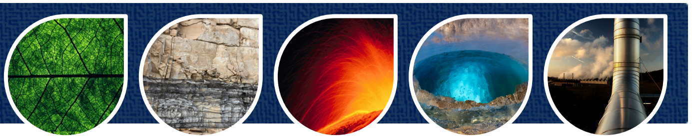
temperature (*thermal survey*)

electrical conductivity (*electrical and electromagnetic survey*)

elastic properties influencing the propagation velocity of elastic waves (*seismic survey*)

density (*gravity survey*)

magnetic susceptibility (*magnetic survey*).



Geophysics: GRAVITY SURVEYS

The earth's gravitational field is usually described by the vertical component of the gravitational acceleration g_z.

Combining two of Newton's law

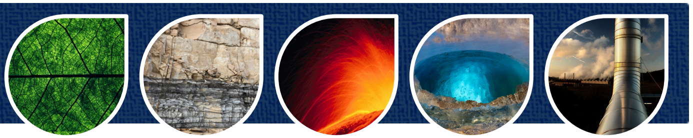
Universal law of gravitation $F=Gm_1m_2/r^2$

G gravitational constant 6.67 x 10^-11 m3 kg-1 s-2

Second law of motion $F=mg$

g gravitational acceleration or "gravity"

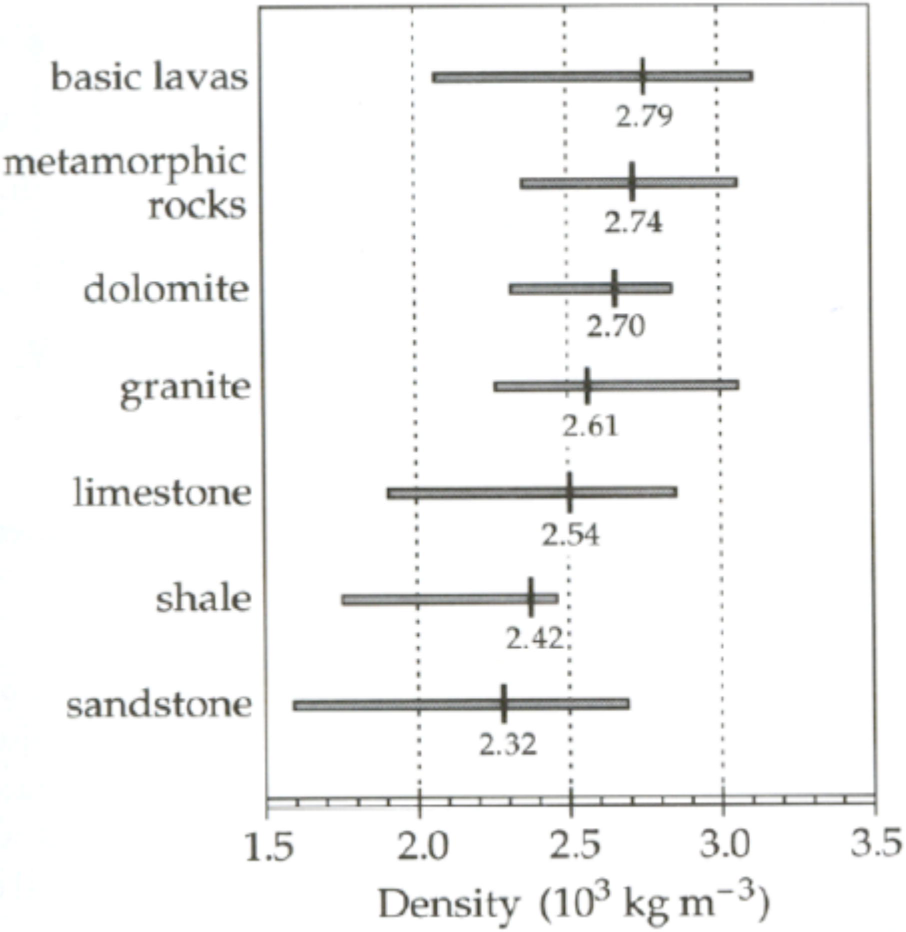
We obtain $g=GM_E/R^2$



Geophysics: GRAVITY SURVEYS

Positive gravity anomalies > higher density
associated with plutonic intrusions and dykes, deposition of silicates from hydrothermal activities during greenschist metamorphism.

Negative gravity anomalies > lower densities
caused by higher porosities or by highly fractured parts of a rock, alteration minerals produced by circulation of hot water



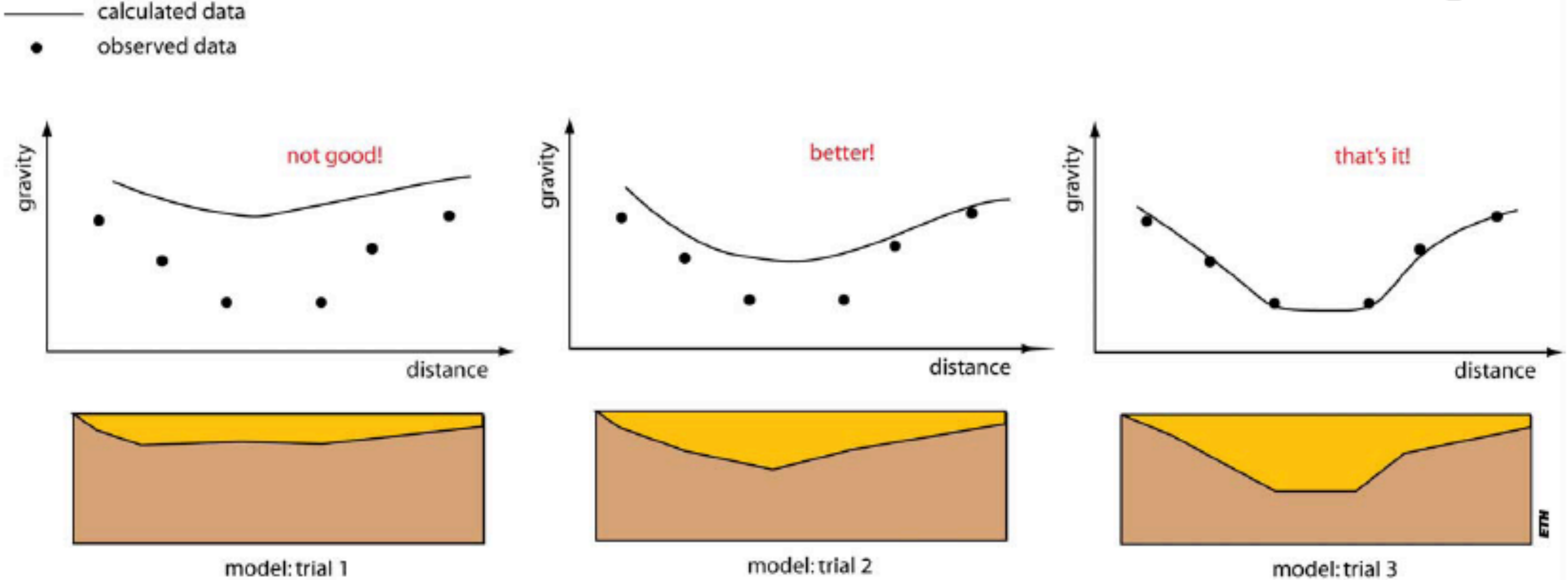
Geophysics: GRAVITY SURVEYS

(1) Construction of a reasonable model

(2) Computation of its gravity anomaly

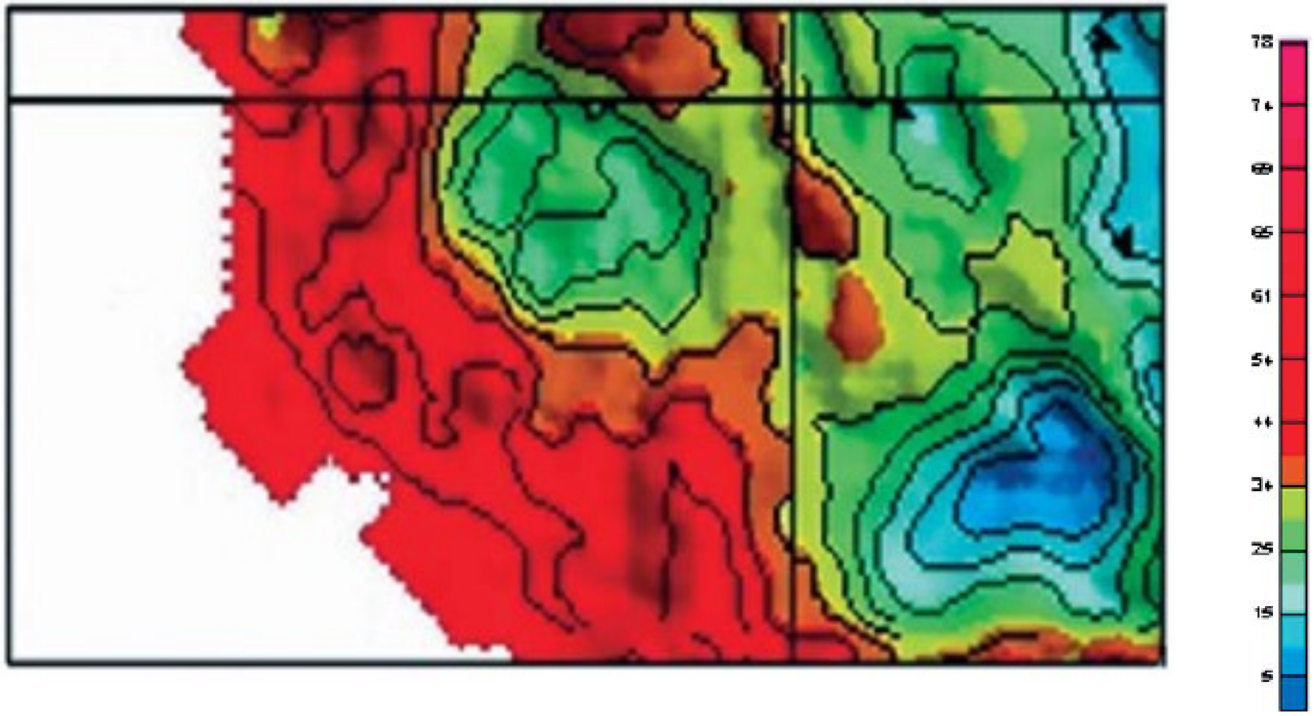
(3) Comparison of computed with observed anomaly

(4) Alteration of the model to improve correspondence of observed and calculated anomalies and return to step (2)



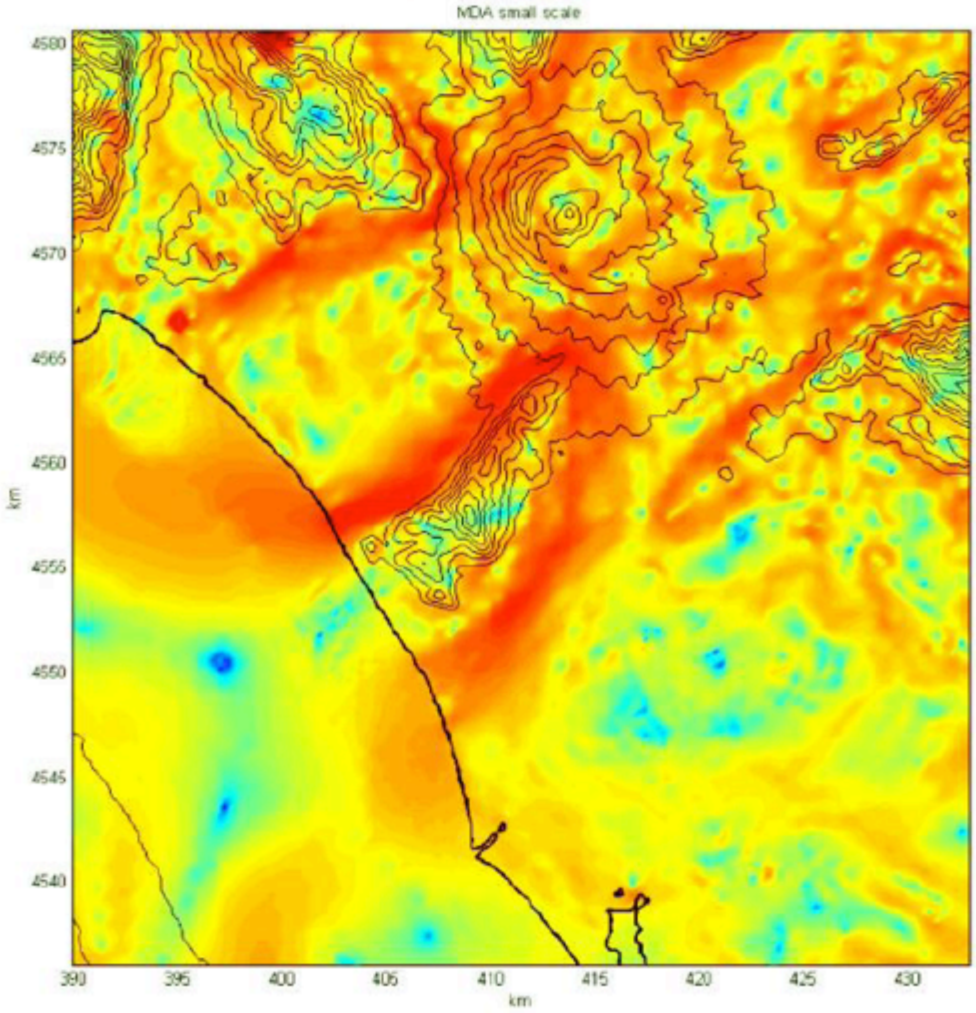
Geophysics: GRAVITY SURVEYS

Contour map of Bouguer anomaly with lines of equal gravity anomaly. These lines are called isogals - gal in memory of Galileo Galilei.

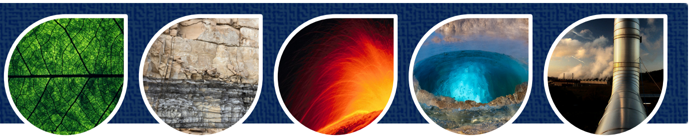


Gravity data in Tuscany. In the figure it is evident how the main geothermal fields of Larderello, Travale and Amiata can be recognized as areas of anomalously low density and high heat flow *(from Orlando, 2005)*.

Geophysics: GRAVITY SURVEYS



After careful data processing and modelling, gravity data may help in delineating structural features. In the figure it is evident how structural lineaments can be recognized in an area of Campania *(from La Manna et al, 2013)*.



Geophysics: GRAVITY SURVEYS

Gravity monitoring surveys are performed also to define the change in groundwater level and for subsidence monitoring.

Fluid extraction from the ground which is not rapidly replaced causes an increase of pore pressure and hence of density. This effect may arrive at surface and produce a subsidence, whose rate depends on the recharge rate of fluid in the extraction area and the rocks interested by compaction.

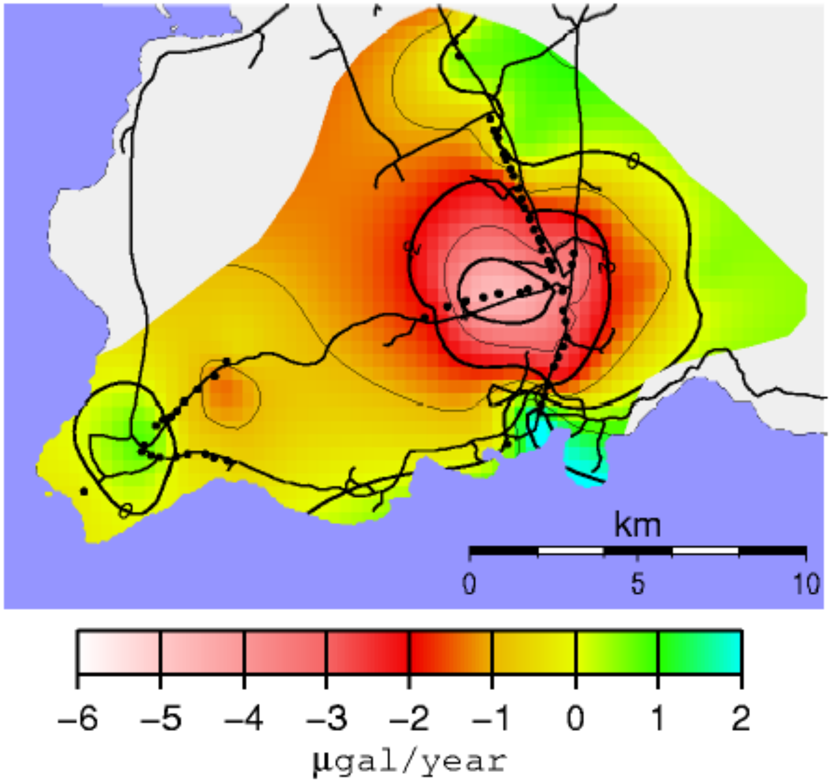
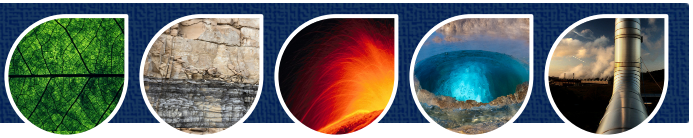


Figure 8. Mean gravity variation (μgal/year) from 1975 to 1999. Only points measured in 1999 and at least two times earlier are used.

from WGC200

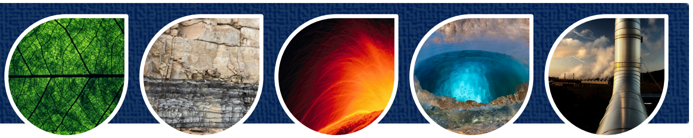


Geophysics: GRAVITY SURVEYS

The advantages of gravimetric methods over other geophysical methods are that they are comparatively easy to use and fairly economical as far as their absolute cost is concerned.

They do provide a good estimate of the extent of bodies with certain density contrasts and can thus help constrain the location and extent of reservoirs.

The resolution and quality of data, however, decrease considerably with depth. Gravimetric studies therefore provide a useful tool to be used for shallow reservoirs in conventional systems and, given their often ambiguous results, *in combination* with other geophysical methods.

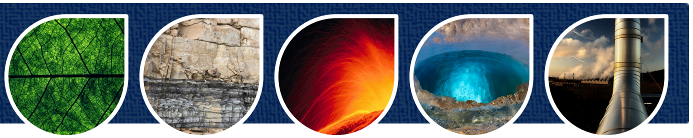


Geophysics: MAGNETIC SURVEYS

Investigation on the basis of anomalies in the Earth’s magnetic field resulting from the magnetic properties of the underlying rocks (magnetic susceptibility and remanance)

Several minerals containing iron and nickel display the property of ferromagnetism. Rocks or soils containing these minerals can have strong magnetization and as a result can produce significant local magnetic fields.

Rock magnetism is acquired when the rock forms, and it reflects the orientation of the magnetic field at the time of formation. But rock magnetism can also change with time, if the rock is subjected to temperatures above a certain point, called the Curie temperature, above which it loses its magnetic properties, and it is remagnetised once it cools down again, now induced by the magnetic field present at that time.



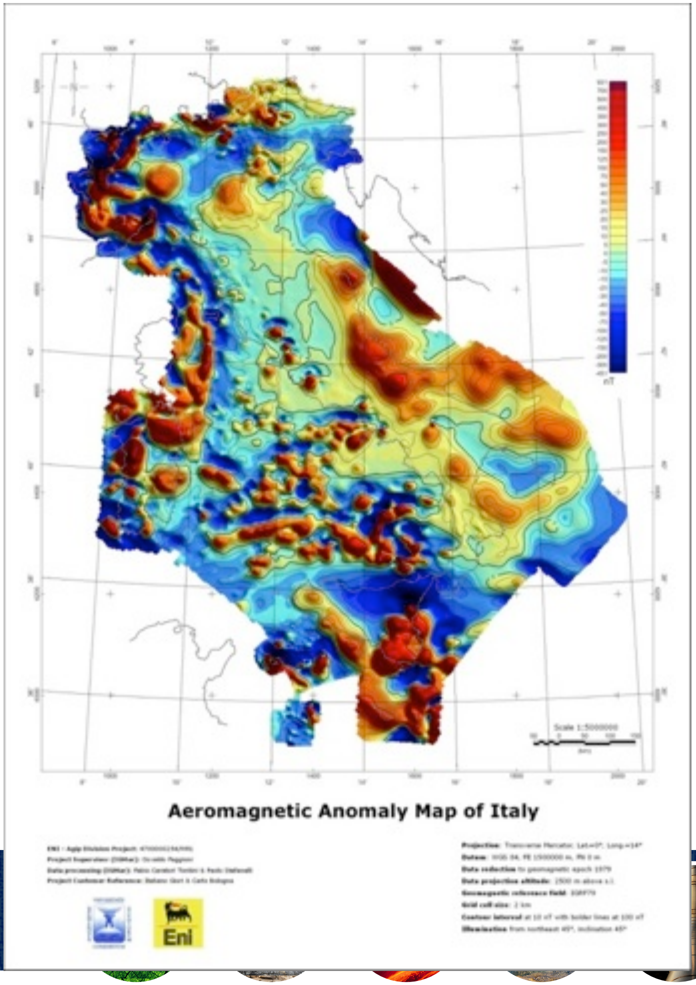
Geophysics: MAGNETIC SURVEYS

Measurements are performed using magnetometers either at the surface or airborne, if the objective is regional mapping.

Silicate minerals, rock salt (halite) and limestones (calcite) have a very low magnetic susceptibility and are therefore not useful for magnetic measurements.

Consequently, sedimentary rocks usually have much lower magnetic susceptibilities than igneous or metamorphic rocks. Thus the magnetic method has traditionally been used for identifying and locating masses of igneous rocks that have relatively high concentrations of magnetite, which is the most common of the magnetic minerals.

Strongly magnetic rocks include basalt and gabbro, while rocks such as granite, granodiorite and rhyolite have only moderately high magnetic susceptibilities.



Geophysics: MAGNETIC SURVEYS

Curie temperature is in the range of a few hundred to 570°C for titano-magnetite, the most common magnetic mineral in igneous rocks

Magnetisation at the top of the magnetic part of the crust

⇓

relatively short spatial wavelengths

Magnetic field from the demagnetisation at the Curie point in depth

⇓

longer wavelength and lower amplitude magnetic anomalies

This difference in frequency characteristics between the magnetic effects from the top and bottom of the magnetised layer in the crust can be used to separate magnetic effects at the two depths and to determine the Curie point depth.