The ELECTROMAGNETIC methods for geothermal exploration

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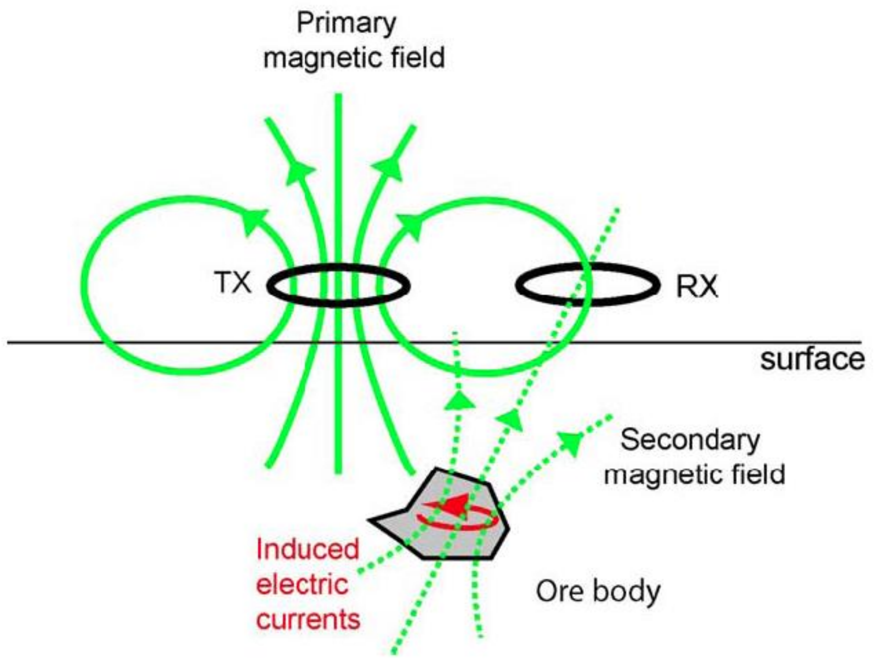
All EM inductive methods have common features:

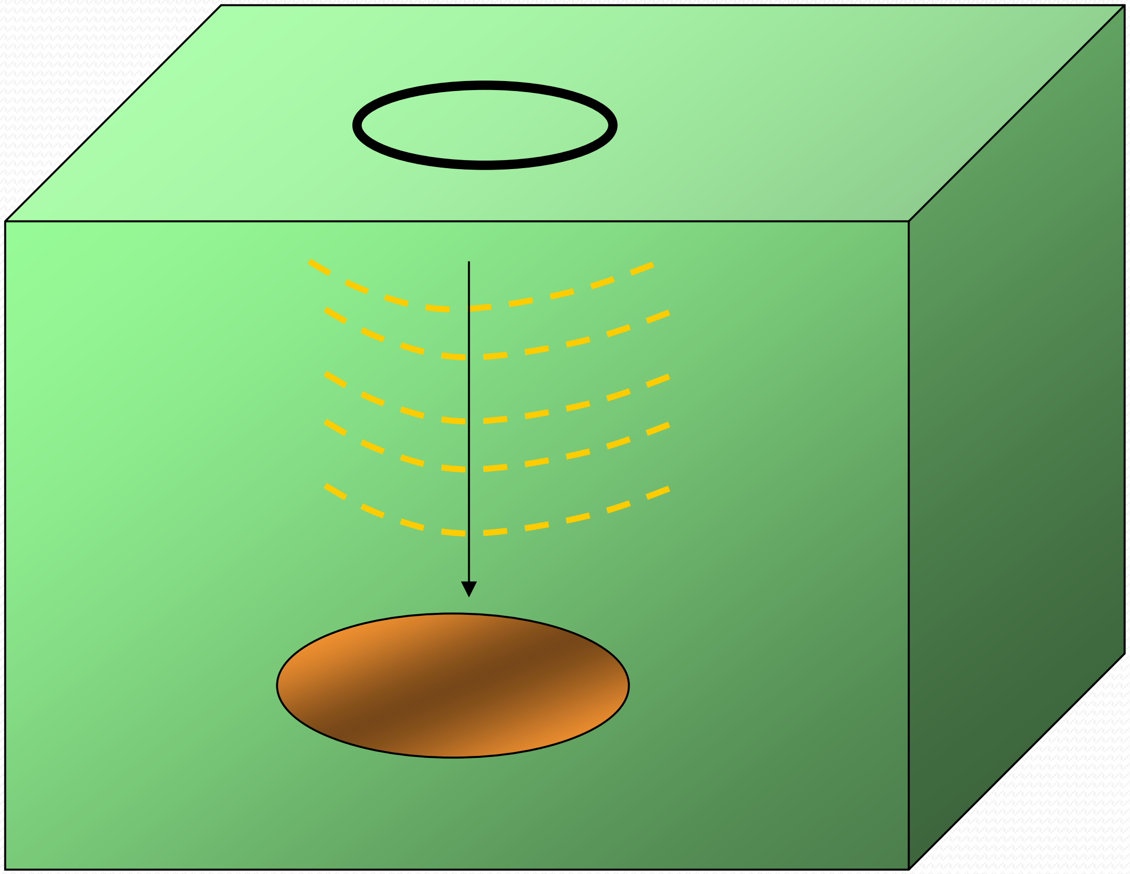
•a **primary field** (either natural or artificial) incident on the Earth. This can be man made or natural. The geometry can be that of a plane wave or generated by a dipole transmitter (TX). The time variation can be a single harmonic frequency or a pulse;

•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.



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

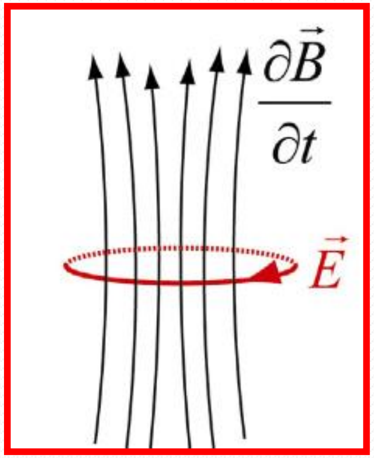
Electromagnetic properties of the medium influences the EM propagation

$$\nabla \times E = -\frac{\partial B}{\partial t}$$

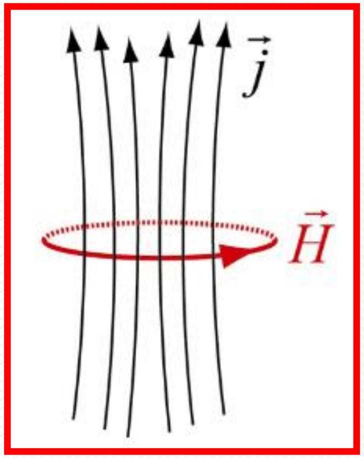
$$\nabla \times H = j$$

Faraday's Law

Ampere – Maxwell Law



Faraday's law describes how a variable magnetic field produces a variation of electric field



Ampere's law describes how an electric current or an electric field varying with time produces a magnetic field

Current density (conduction)

$$\nabla \times H = J_C + J_D$$

$$\nabla \times H = \sigma \cdot E + \varepsilon \frac{\partial E}{\partial t}$$

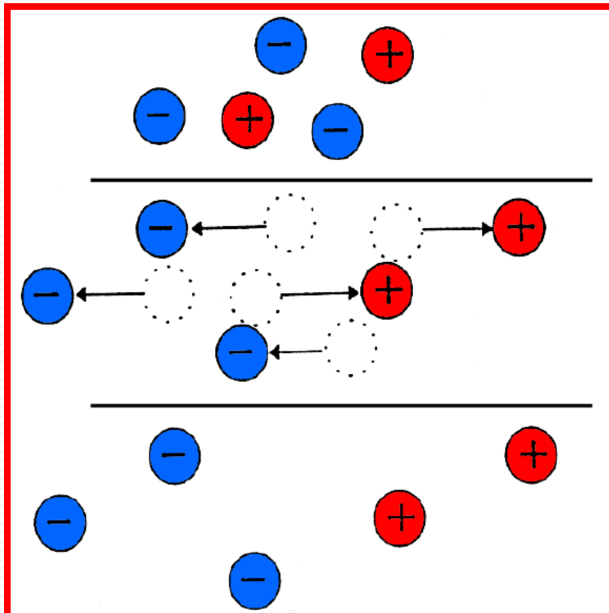
Current density (displacment)

Dipole intensity

$$D = \varepsilon \cdot E$$

$$J_D = \frac{\partial D}{\partial t} = \varepsilon \cdot \frac{\partial E}{\partial t}$$

Displacement current



No E field

E field applied

E field removed

Electric field intensity
(V/m)

Magnetic permeability
[Wb/Am]

$$\nabla \times E = -\mu \frac{\partial H}{\partial t}$$

Faraday’s Law

$$\nabla \times H = \sigma \cdot E + \varepsilon \frac{\partial E}{\partial t}$$

Ampere – Maxwell Law

Magnetic field intensity
(A/m)

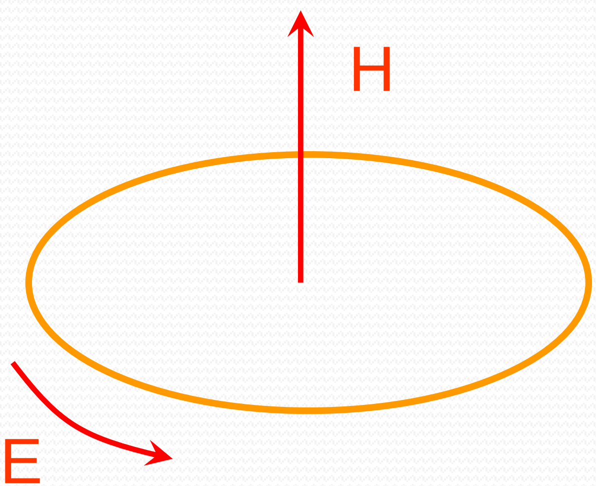
Electrical
permittivity[F/m]

Electrical conductivity [
S/m]

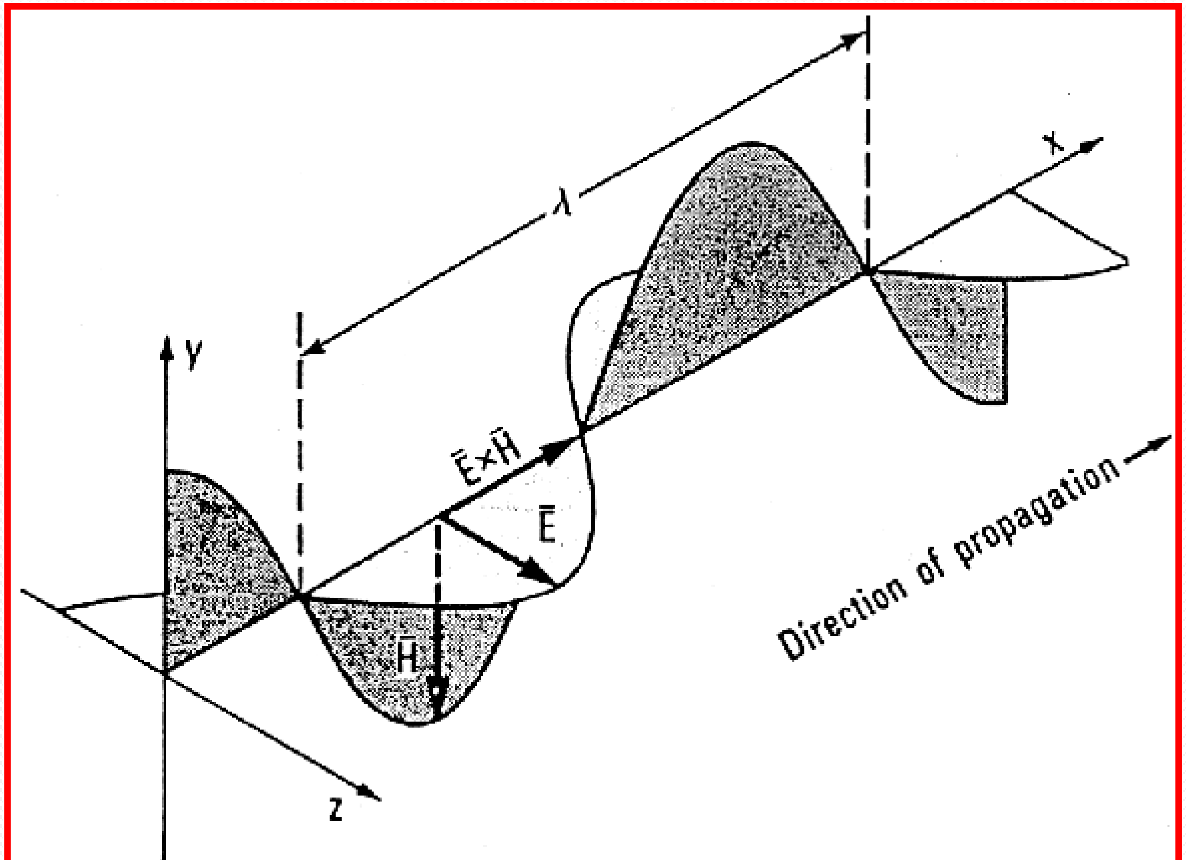
For a field varying with time in a sinusoidal way:

$$E(t) = E_0 \cdot e^{i\omega t}$$
$$H(t) = H_0 \cdot e^{i\omega t}$$

$$\omega = 2\pi f$$



$$\nabla^2 E = i\omega\mu\sigma E - \omega^2 \mu\varepsilon E$$
$$\nabla^2 H = i\omega\mu\sigma H - \omega^2 \mu\varepsilon H$$



Electric and magnetic field vectors orthogonal among themselves and to the wave direction.

Electric field vectors will always align themselves at right angles to perfectly conductive surfaces and a wave can therefore be guided by enclosing conductors.

The extent to which this is possible is governed by the relationship between the wavelength of the radiation and the dimensions of the guide.

Waves at VLF frequencies propagate very efficiently over long distances in the waveguide formed by the ground surface and the ionosphere.

ARTIFICIAL SOURCE ELECTROMAGNETIC METHODS

An artificial source is used to generate the electromagnetic (EM) field and the frequencies are so low that the measurement distance is less than the free-space wavelength. This is the quasi-static range, where conduction currents rather than displacement currents predominate.

The distribution of currents induced in the earth depends on the product of electrical conductivity, magnetic permeability and frequency

ARTIFICIAL SOURCE ELECTROMAGNETIC METHODS

parametric sounding
(similar to the natural-source EM sounding technique):
measurements of the EM response at several frequencies or times, obtaining information on the variation of conductivity with depth

geometric or distance soundings
(resemble geoelectrical soundings):
measures the response at a single frequency
at several source-receiver separations
A set of measurements made at several spacing contains information on the variation of conductivity with depth

The ***source*** is usually an insulated loop.

Small loops can be treated mathematically as time-varying magnetic dipoles, especially if the loop dimensions are less than five times the distance to the nearest receiver.

The moment (source strength) of the dipole depends on the current as well as turns-area product of the loop.

In deep soundings grounded wires are used more than loops since the primary field falls off less rapidly at large distances from grounded wires than from loops.

Grounded-wire antennas are straight lengths of wire laid on the surface and connected to the earth through low resistance electrodes at either end.

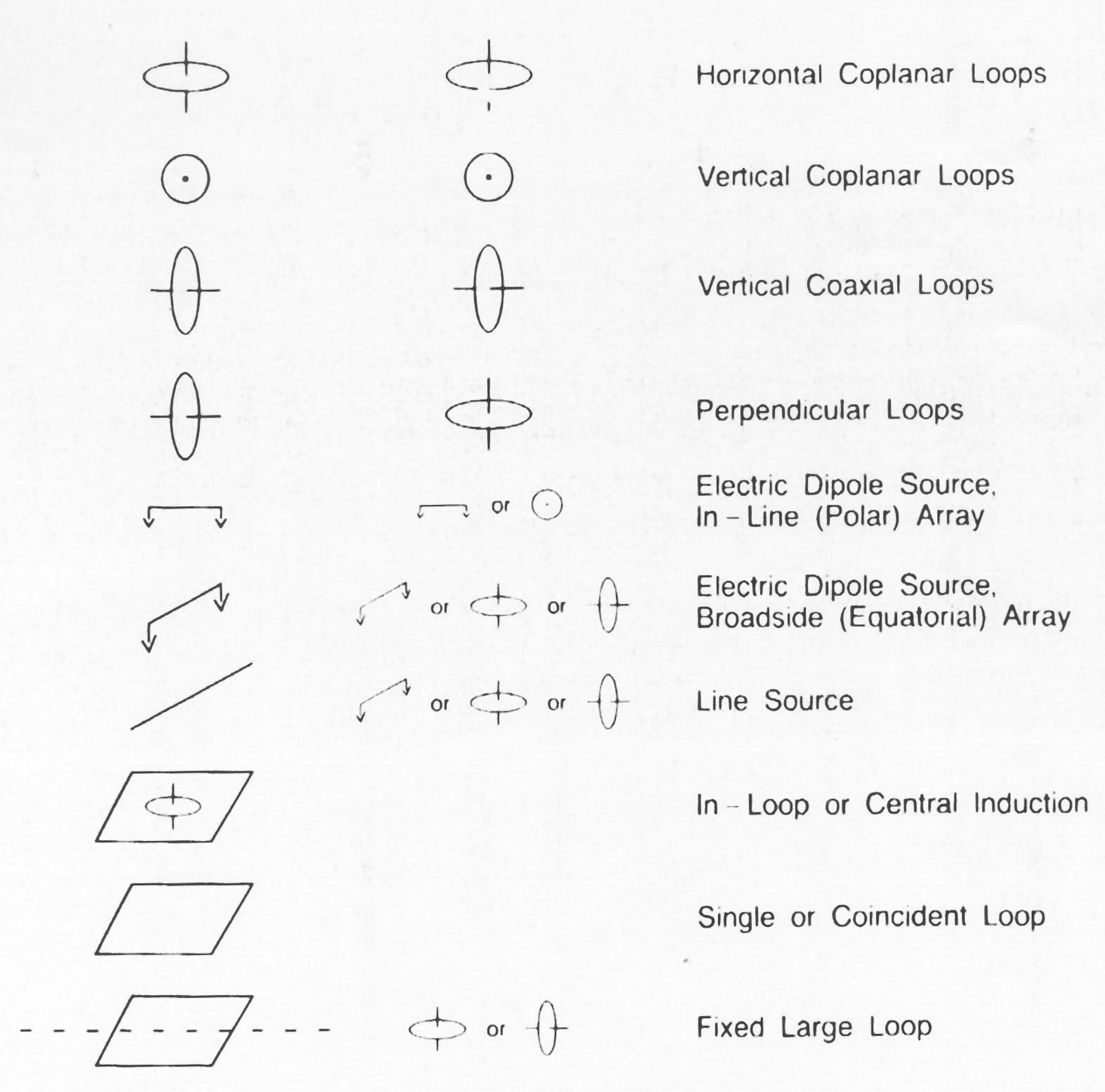
Generally, if the length of wire is greater than five times the distance to the receiver site, wires can be treated mathematically as electric dipoles.
Usually they are sufficiently short and the frequency low enough that the current distribution along the wire can be considered uniform.

In most frequency-domain (**FEM**) systems, a current having approximately a sinusoidal or a square waveform is driven through the antenna by an amplifier or switcher, and the frequency is usually changed in discrete steps.

In time-domain (**TDEM** or **TEM**) systems the most common waveform is a train of approximately square-bipolar pulses with an off-time between pulses.

Three types of ***receivers*** are used for EM soundings: induction coils, magnetometers and grounded wires.

FEM results are often expressed in terms of V/I, B/I, or dB/dt/I, where I is the source current, V is the voltage difference, and B is the magnetic flux density. **TEM** results are expressed as impedances, V/I. Impedance is then transformed to apparent resistivity. Apparent resistivity is defined as the resistivity of the homogeneous earth which would produce the measured response at each frequency or time.



Typical source-receiver geometries:

it is specified by giving the orientation of the source and receiver antennas and the orientation and length of the line joining the center of the two antennas.

A geometric sounding may be made by leaving the source fixed and moving the receiver, or, both source and receiver may be moved with respect to the center of the array.

The range of depths which can be explored with FEM generally depends on the source-receiver separation as well as the frequency used.
In FEM it is always necessary to somehow distinguish the *secondary* field from the *primary* field.

The ***primary field*** is the alternating magnetic field induced by the alternating current flowing in the transmitter in the range of a few hundred Hz to a few kHz.

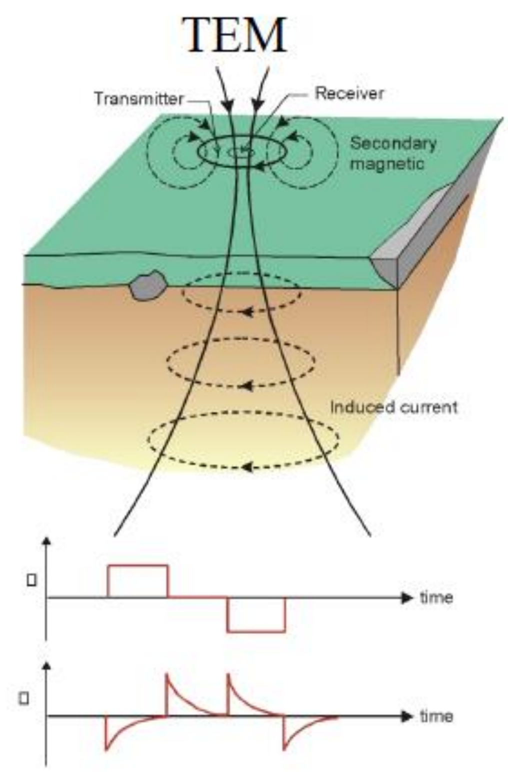
The primary field induces an alternating current to flow in all the conductors present in the earth (eddy currents), which induces an alternating magnetic field, called ***secondary field***, which also extends through the region which includes the receiver.
The secondary field from a particular layer will be very small compared with the primary field unless the separation between the source and receiver is on the order of the depth to the layer or deeper.

Thus it is generally necessary to employ source-receiver separations of the order of one or two times the maximum depth to be sounded; at a smaller separation than this it is not possible to accurately measure the small secondary field in the presence of the much larger primary field.

If the spacing is too large there is difficulty resolving the parameters of thin layers, especially when they are deep.

For measurements in the near zone the spacing should not be larger than five times the thickness of any layers to be resolved.

TEM method



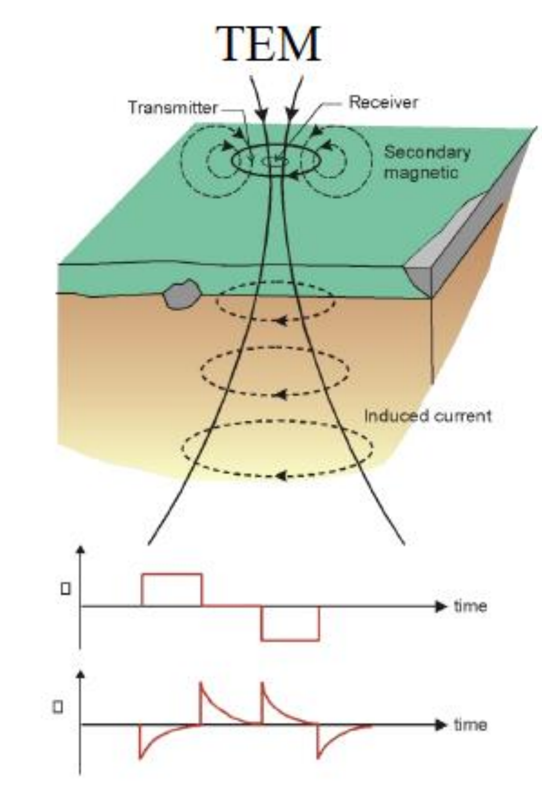
Active electromagnetic (EM) methods are used mainly for shallow depth resistivity studies and to help with static shift corrections of MT data.

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.

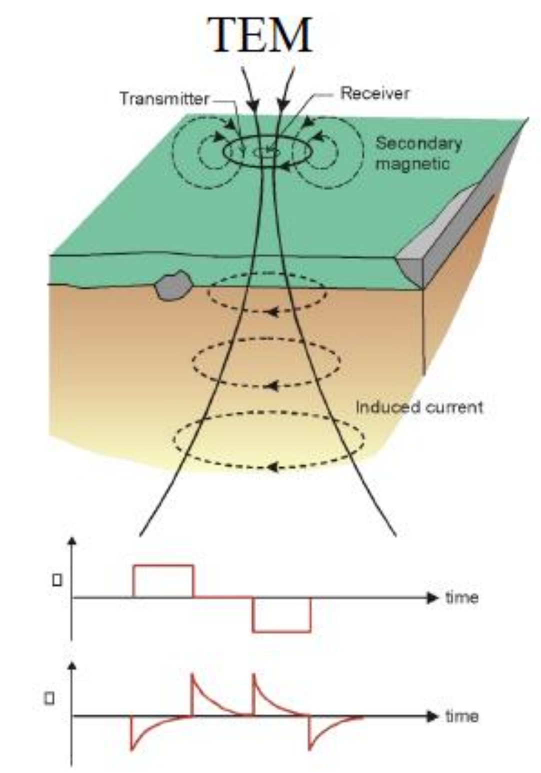
TEM method

At time t=0 the current is abruptly switched off and the decaying magnetism induces eddy currents in the formation that try to counteract the magnetic decay. The spool at the loop's centre measures the magnetic decay at the surface with time elapsed since the current was switched off.

The secondary field due to the induced eddy currents in the ground are then recorded in the absence of the primary field. This allows calculation of the formation resistivity below the loop.



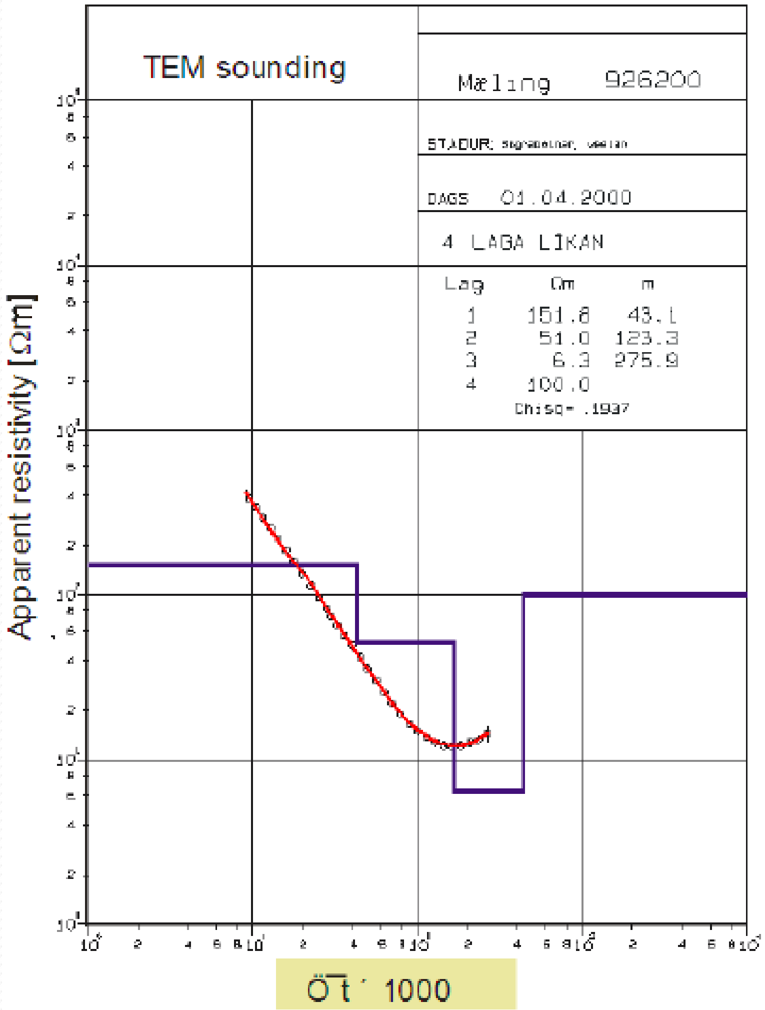
The depth range for TEM (made during the current off-time) depends on the sample time measured and signal-to-noise ratio, and not on the source-receiver separation; in principle one can sound to any required depth using a single small loop.



The product of a time-domain electromagnetic sounding survey is a curve relating apparent resistivity to the time following the beginning of the transient coupling between source and receiver.

It is presumed that, as this time becomes progressively larger, the eddy currents giving rise to the transient coupling occur at greater and greater depths beneath the receiver.

In a complete interpretation, the response observed in the field is modelled for reasonable earth structures in much the same manner as with any of the other electrical methods.



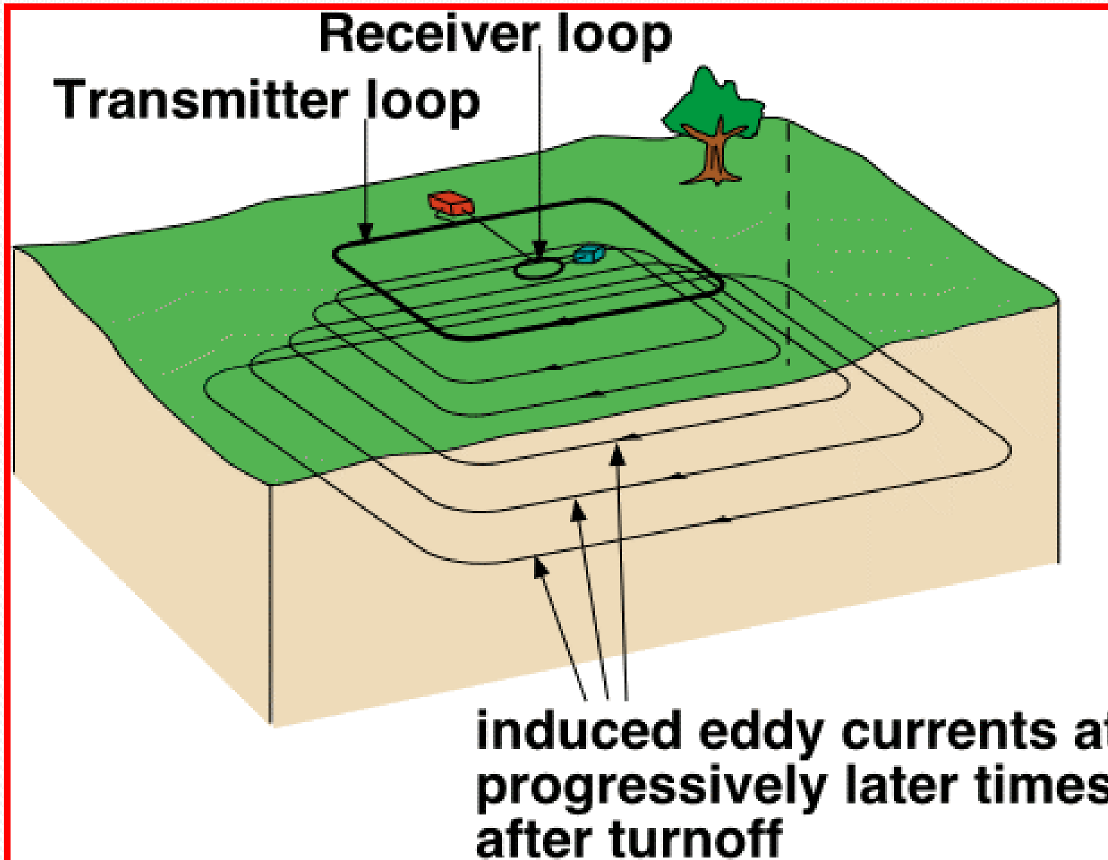
The transmitter is usually a loop with a square shape, the side length varying as a function of investigation depth (usually 100-300 m).

An electric current is injected into the electric wire of the loop (intensity = 3-10 A) in form of pulses of few milliseconds of duration.

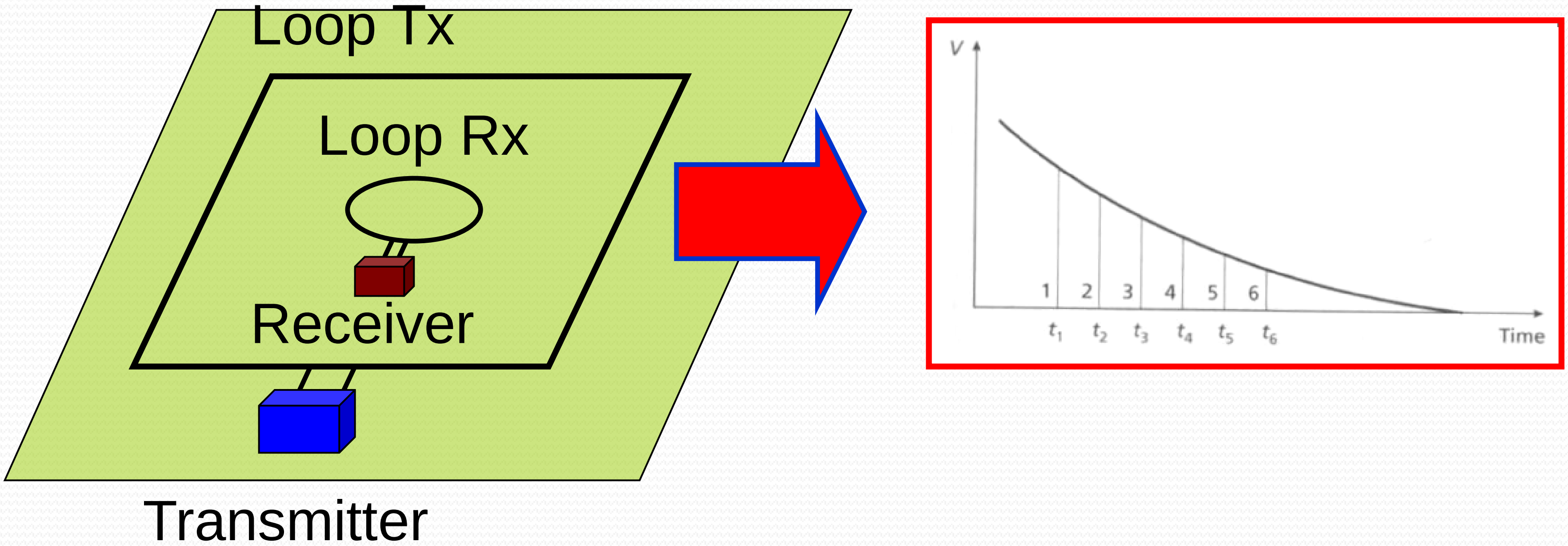
The investigation depth depends on the signal strength (more windings, more energy) and on the loop dimension

A rule of thumb: z= loop diagonal)

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



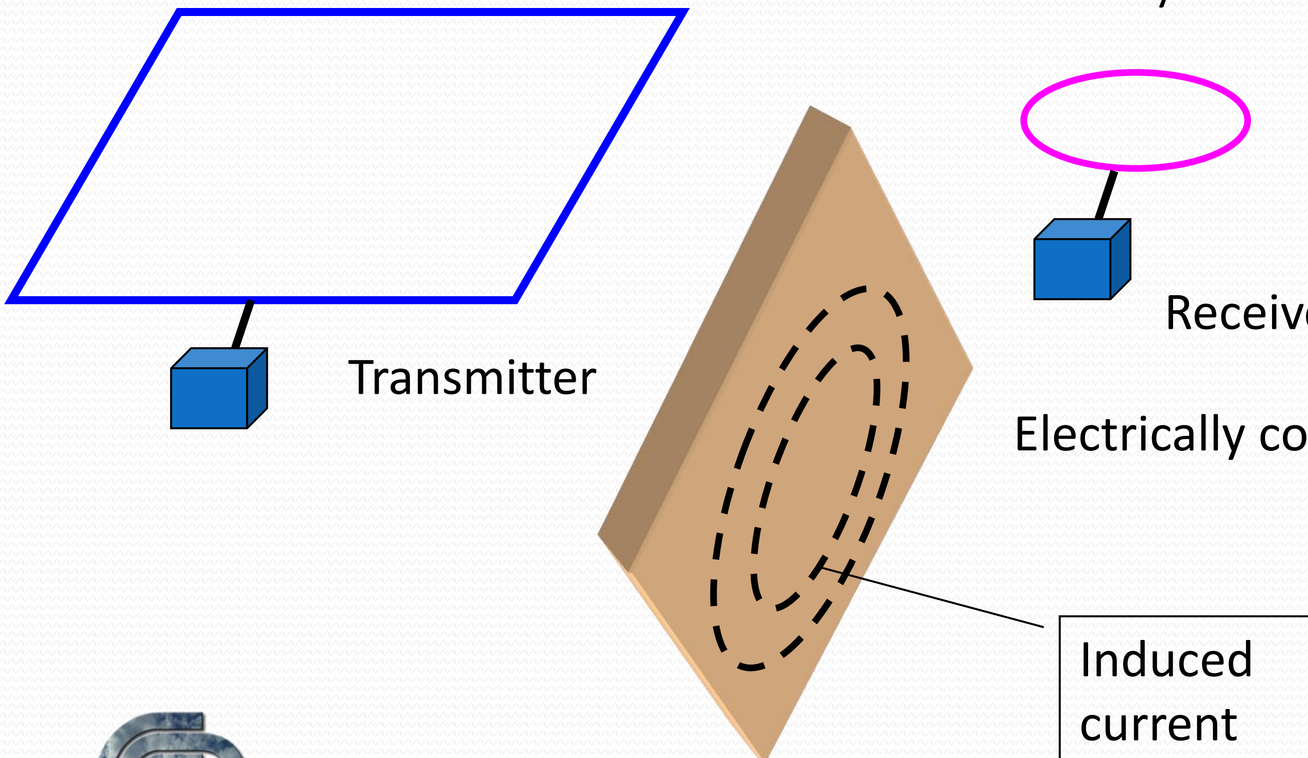
The current is then interrupted and the transient of EM force induced from the secondary magnetic field at the receiver is recorded. Readings are done from the turn-off at fixed intervals during the decay of the secondary magnetic field as it approaches zero, the last ones reaching the deepest structures.

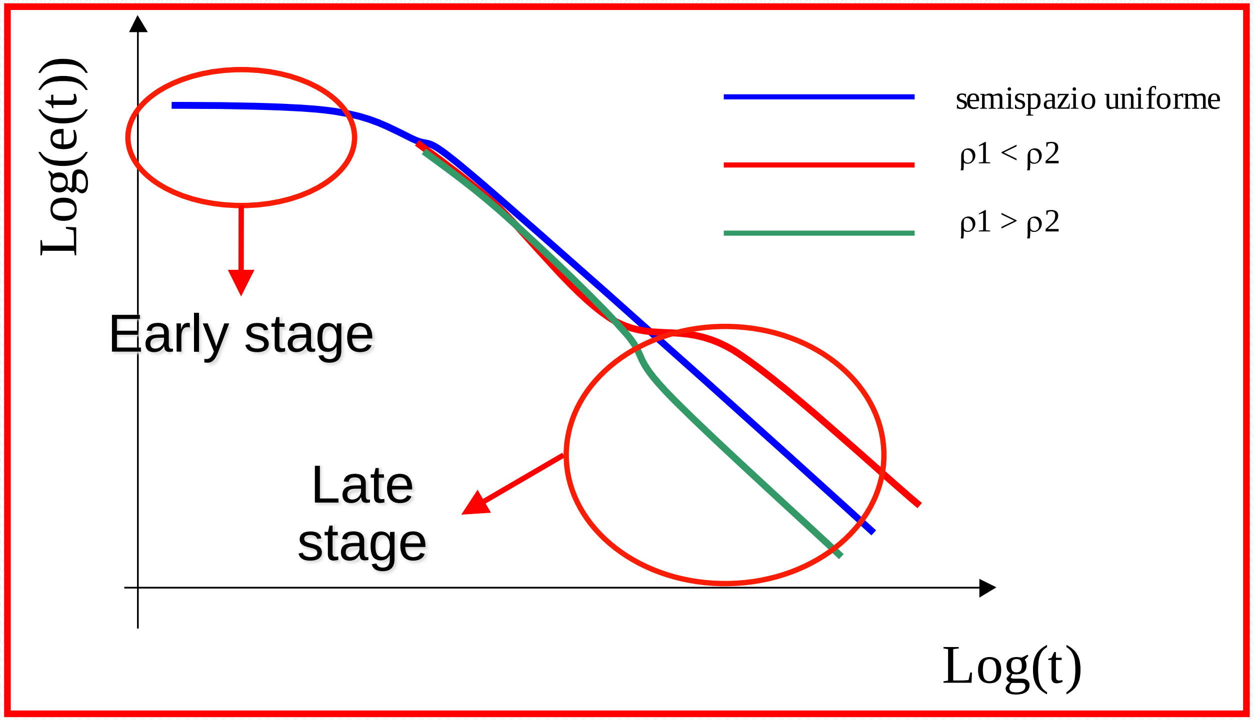


The current due to the primary field flow when the current in the transmitter loop is turned off varies with time and decays from the initial stationary field value

The voltage at the two poles of receiver loop is proportional to the time varying secondary magnetic field. This latter is proportional, at any moment, to the current i(i).

$$\boxed{e(t) \Leftrightarrow \frac{1}{\tau} e^{-t/\tau}}$$





TEM response in an homogeneous and uniform two-layered half-space

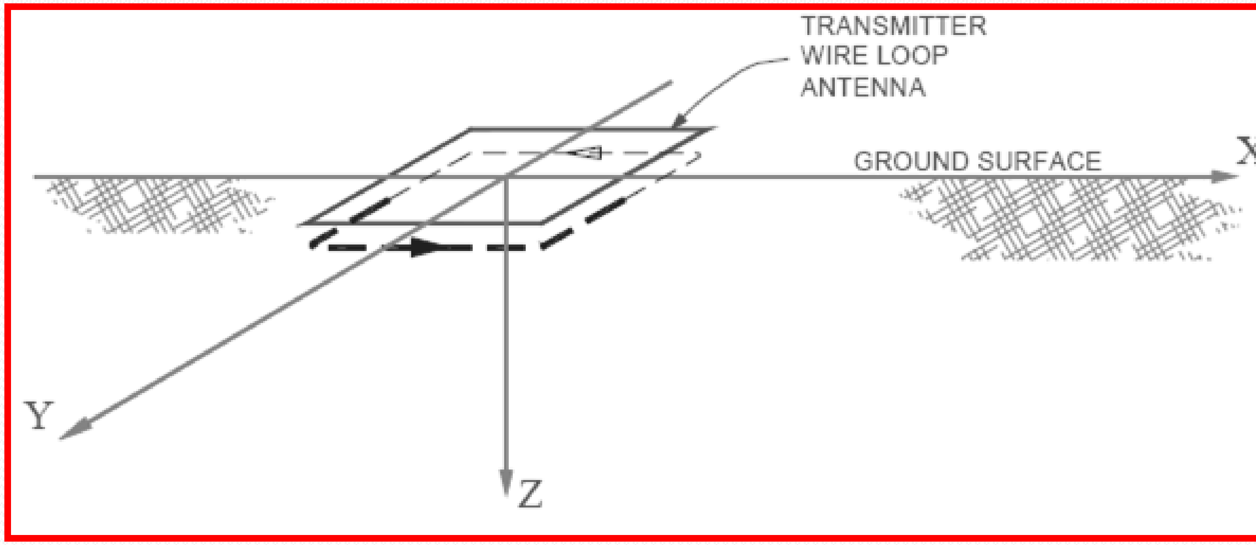
The measured resistivity in the subsurface is expressed as apparent resistivity **ρa**, the „average resistivity" of the structures below the centre of the sounding. It is a function of several variables, including:

- Measured voltage;
- Time elapsed from turn off;
- area of loops/coils;
- number of windings in loops/coils
- Magnetic permeability.

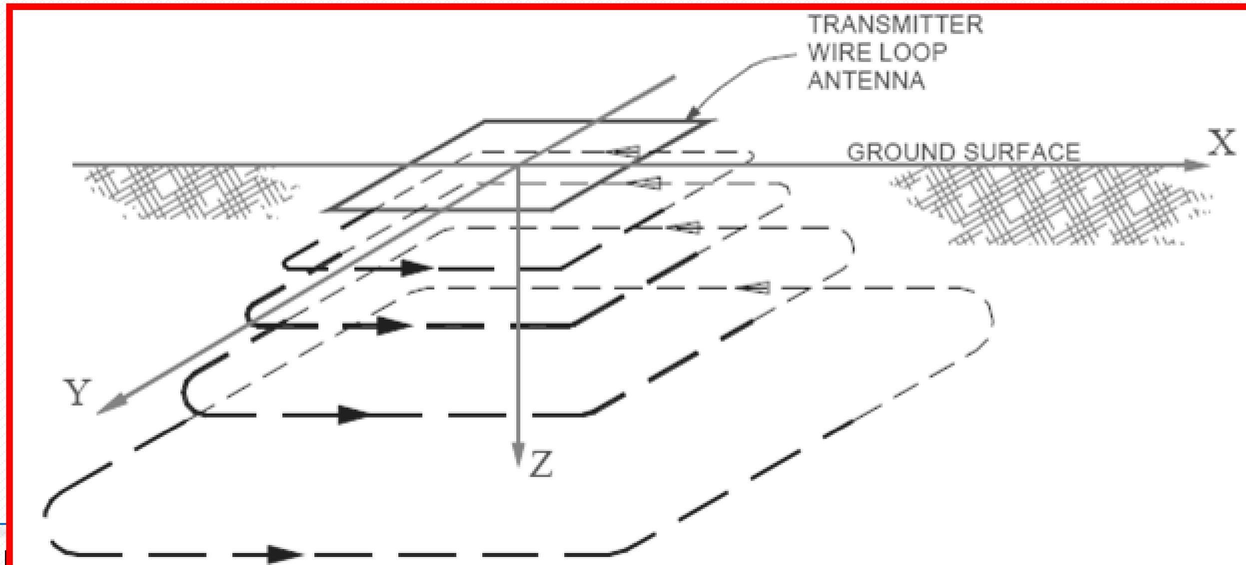
For a homogeneous half-space (for a layered earth the expression is much more complicated), apparent resistivity, **ρa**, expressed in terms of induced voltage **V(t, r)** at late times after the source current is turned off, is given by:

$$\rho_a = \frac{\mu_0}{4\pi}\left[\frac{2\mu_0 A_r n_r A_s n_s I_0}{5t^{5/2}V(t,r)}\right]^{2/3}$$

Early stage



Late stage

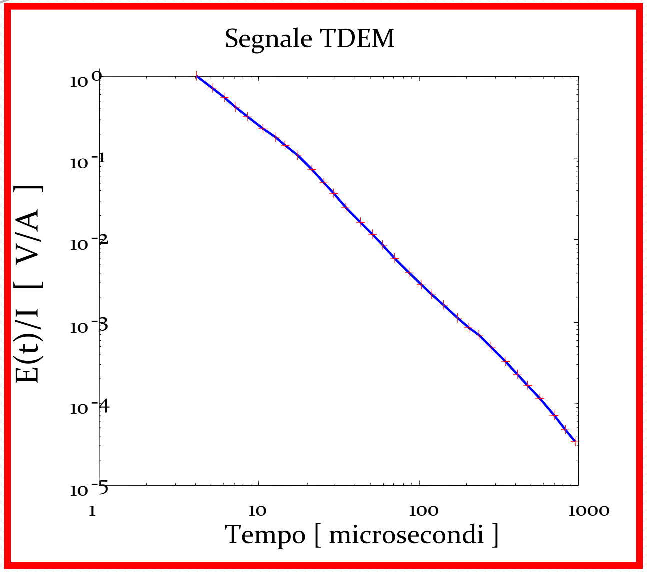


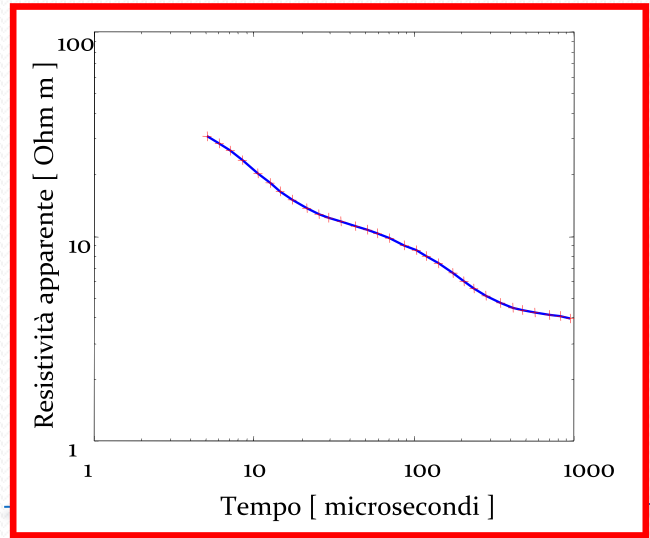
The product of a time-domain electromagnetic sounding survey is a curve relating apparent resistivity to the time following the beginning of the transient coupling between source and receiver.

It is presumed that, as this time becomes progressively larger, the eddy currents giving rise to the transient coupling occur at greater and greater depths beneath the receiver.

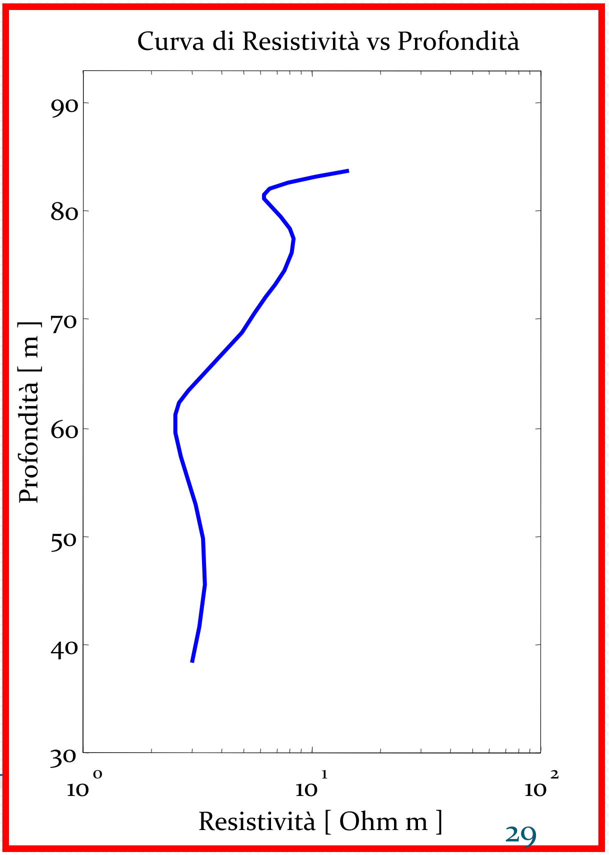
In a complete interpretation, the response observed in the field is modelled for reasonable earth structures in much the same manner as with any of the other electrical methods.

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



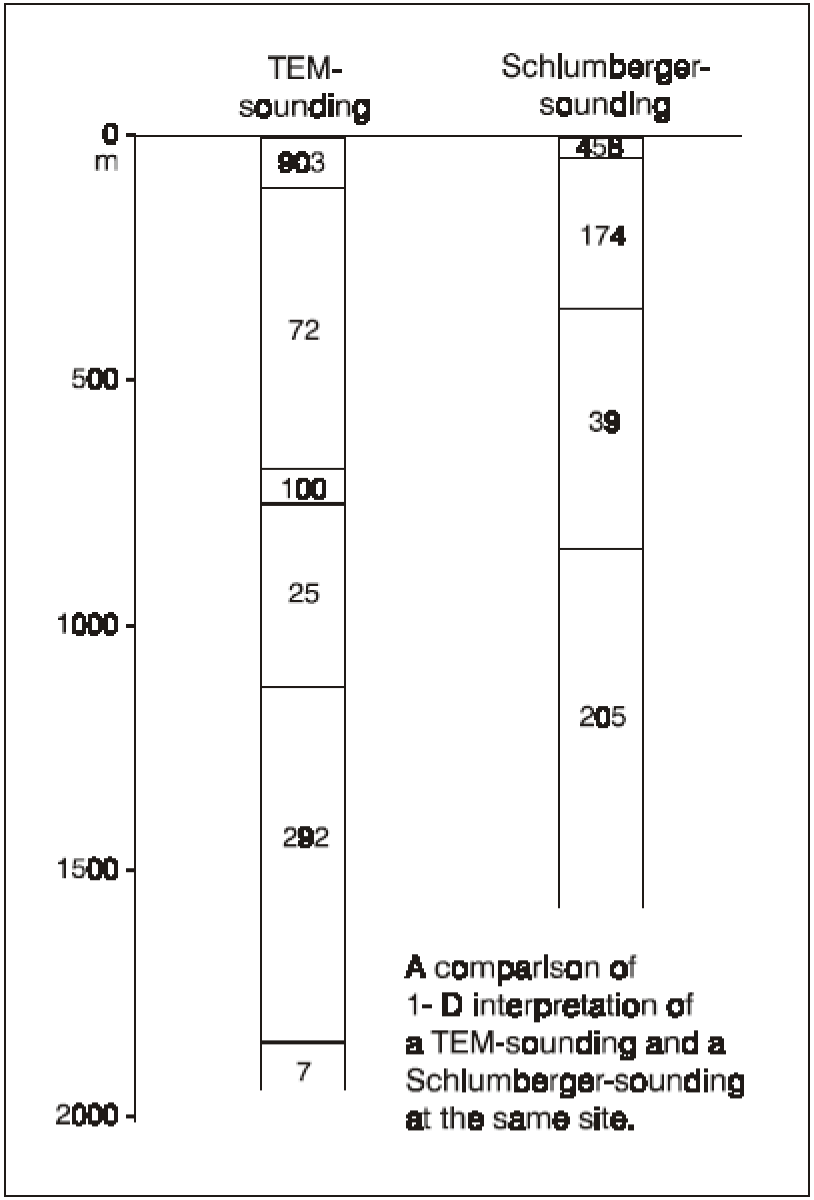
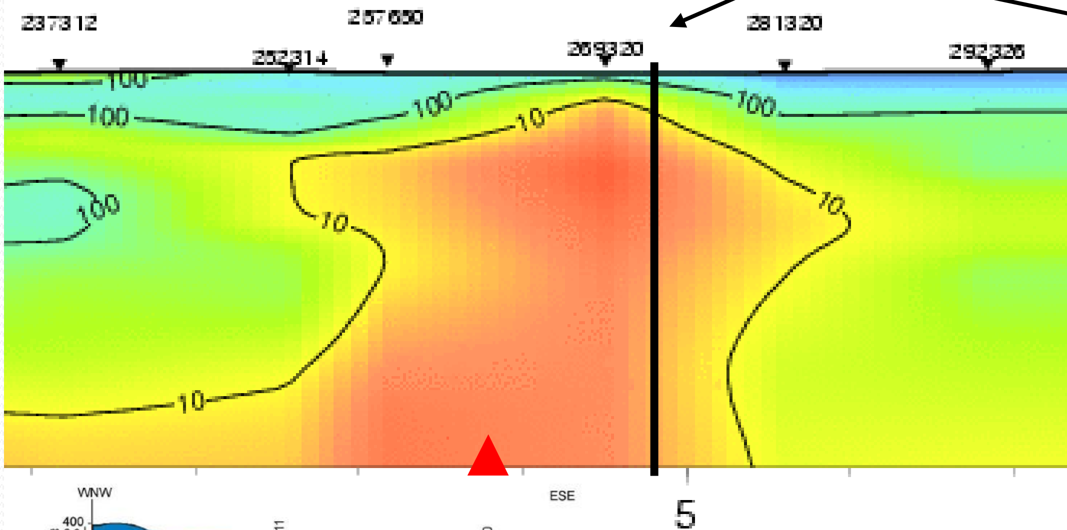


Figure 4. Comparison of 1-D interpretation of TEM and Sclumberger soundings showing greater vertical resolution and depth penetration of the TEM soundings.

BA02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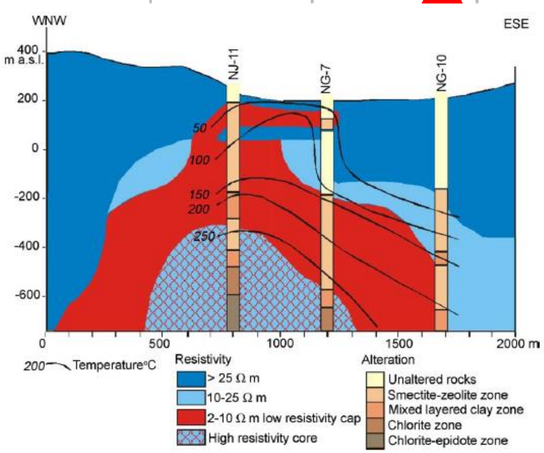
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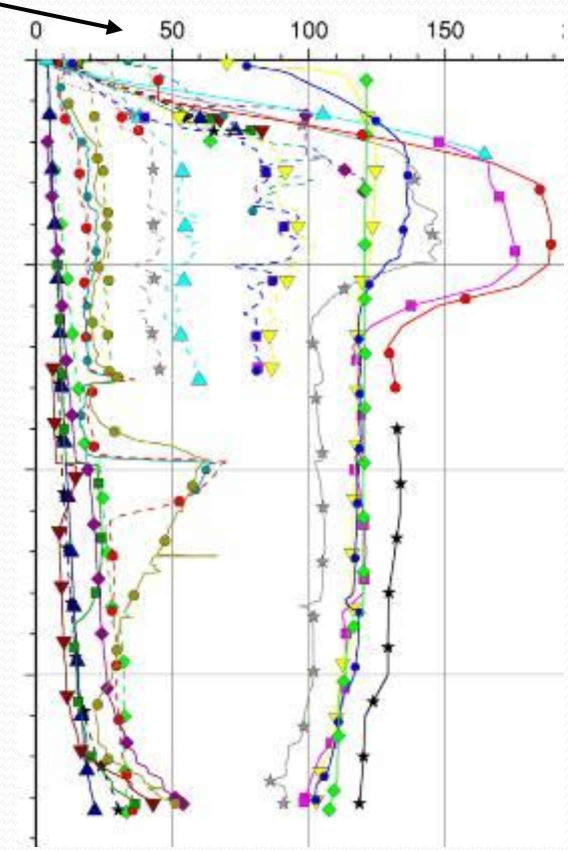
1500
2000 m
200 Temperature°C
Resistivity
> 25 Ω m
10-25 Ω m
2-10 Ω m low resistivity cap
High resistivity core
Alteration
Unaltered rocks
Smectite-zeolite zone
Mixed layered clay zone
Chlorite zone
Chlorite-epidote zone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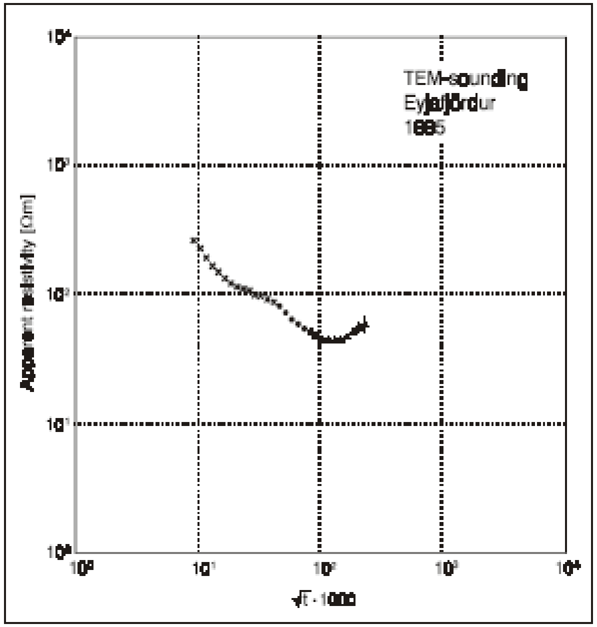


Figure 2. Example on a TEM apparent resistivity cur (resistivity with depth) after pre-processing of data.

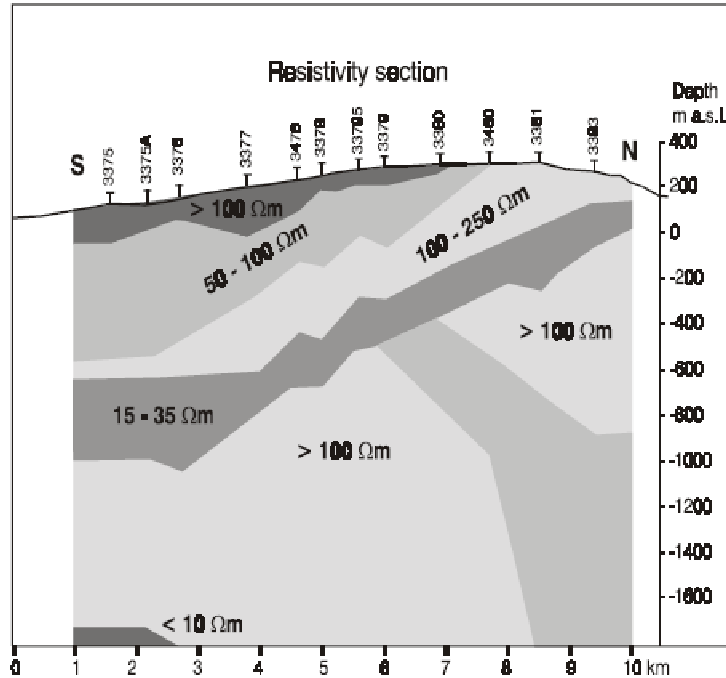


Figure 3. Resistivity cross section based on one dimentional inversion of the TEM data.

An example of TEM data acquired at North Iceland
(from Flovenz and Karlsdottir, WGC2000)

The TEM method has many advantages over the Schlumberger DC sounding method:

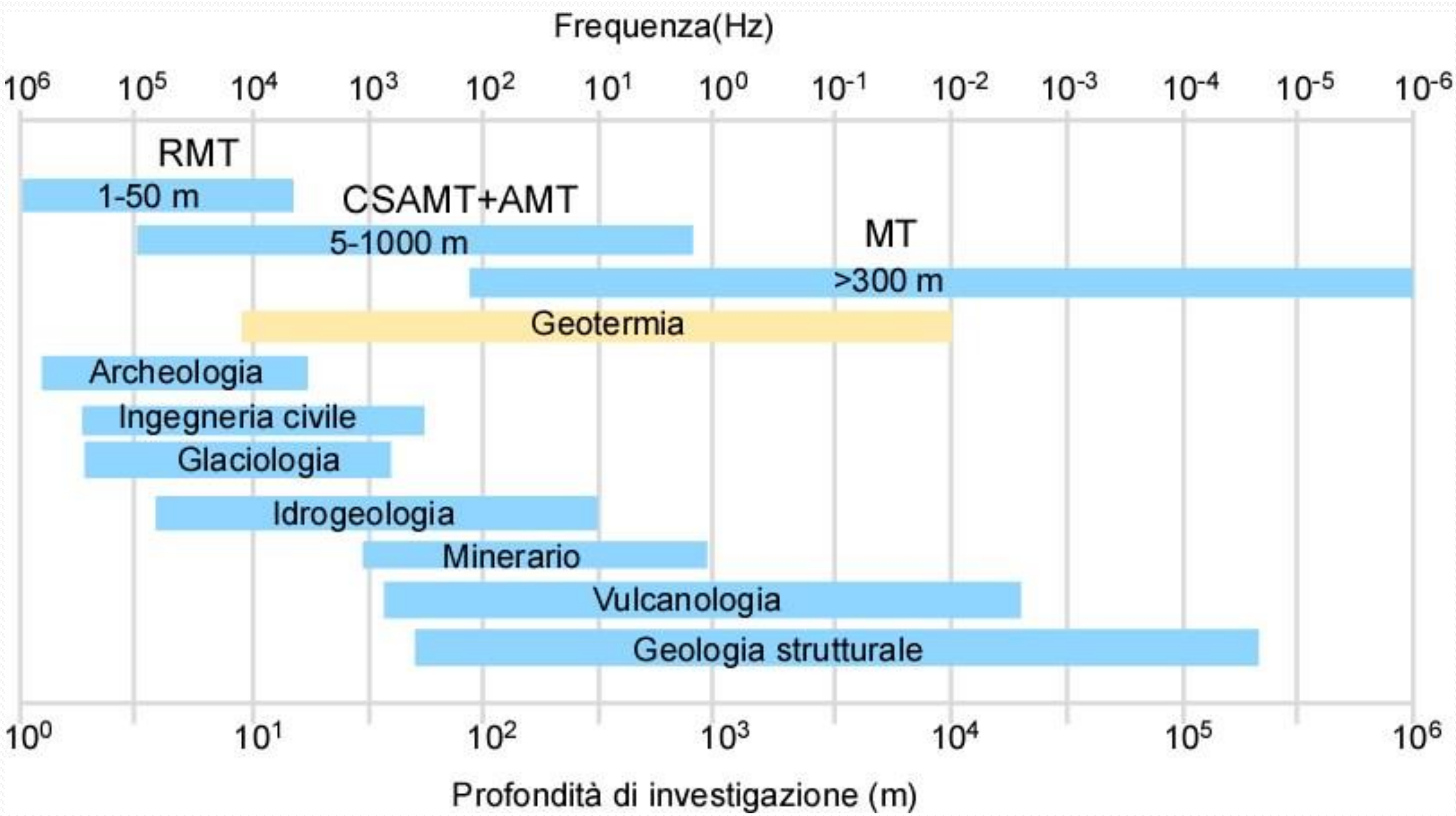
- In TEM, since no current has to be injected into the earth, wires are shorter (though heavier. Main advantage in areas of high contact resistivity at surface and thus current transmission difficult (deserts, lava fields and even possible on snow and ice, or bare rock).
- In TEM, distortions due to local inhomogeneities are small, since the signal (the downward migrating currents) are more downward focussed
- Similarly, TEM is much less sensitive to lateral resistivity variations than DC methods, making 1-D interpretation better justified.
- DC signal is low in low-resistivity structures like in geothermal areas, whereas TEM signal is strong, increasing depth penetration in target areas.
- TEM needs less manpower, both in the field and for interpretation; measurements are considerably faster to carry out. Thus it is more cost effective, or allows collection of data in higher density, and consequently giving a more detailed model of the geothermal system.

THE MAGNETOTELLURIC METHOD

Objectives:

Definition of the shallow structure
Reservoir(s) detection
Individuation of resistivity anomalies due to temperature
Evaluation of underground water mixing
Definition of hydrothermal alteration
Monitoring of possible changes inside the reservoir
Definition of the thermal energy source
(hot and possibly molten deep magmatic bodies)

This method was developed in the ‘50s by Cagniard and Tikhonov, and in the last years have experienced a rapid expansion.

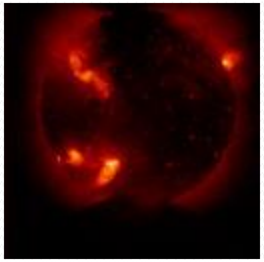

Frequenza(Hz)
10^6 10^5 10^4 10^3 10^2 10^1 10^0 10^-1 10^-2 10^-3 10^-4 10^-5 10^-6
RMT
1-50 m
CSAMT+AMT
5-1000 m
MT
>300 m
Geotermia
Archeologia
Ingegneria civile
Glaciologia
Idrogeologia
Minerario
Vulcanologia
Geologia strutturale
10^0 10^1 10^2 10^3 10^4 10^5 10^6
Profondità di investigazione (m)

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

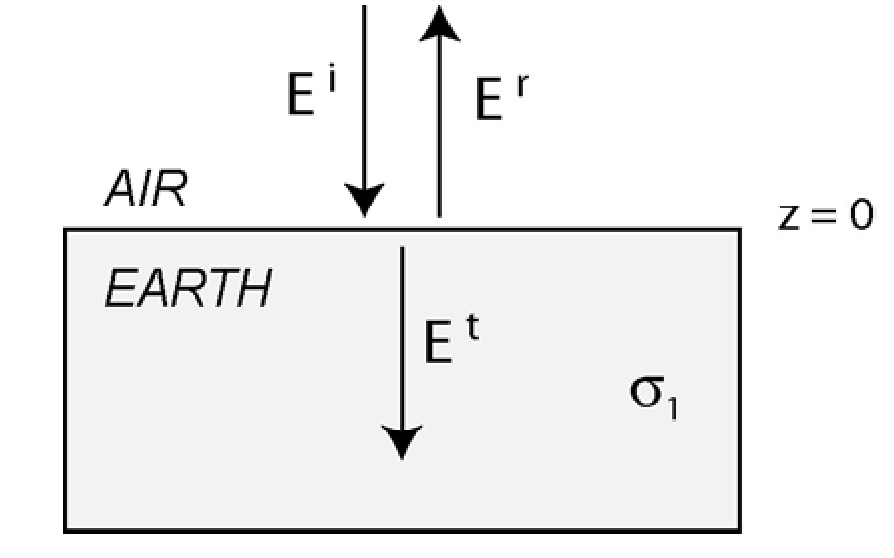
Intermediate frequency signals come from ionospheric resonances



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.

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

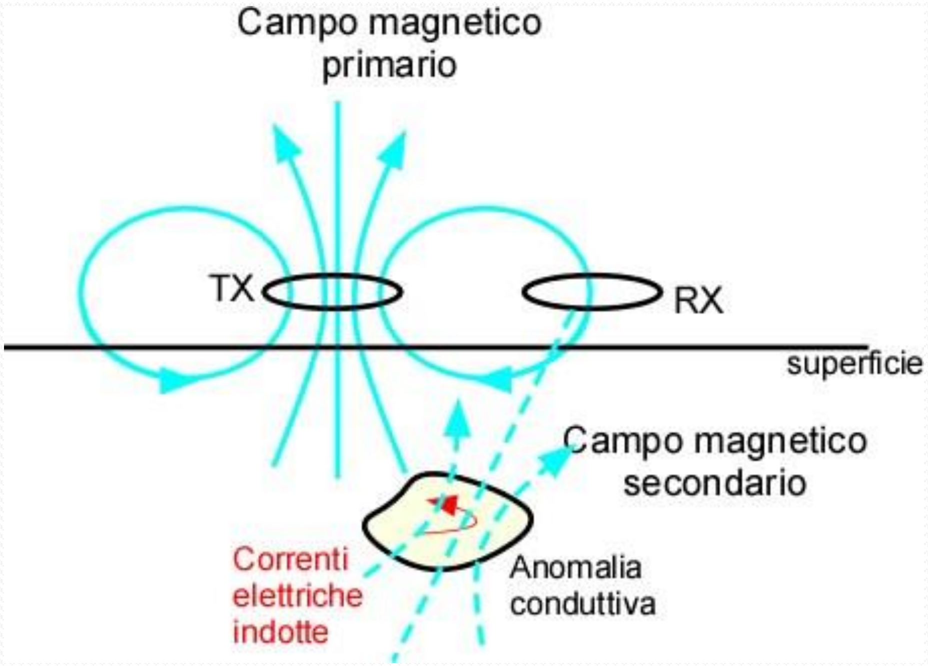


E^i= incident (**primary**) field
E^r= reflected field
E^t= transmitted field

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

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**

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

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.

By measuring E and H at the surface we can generate electrical resistivity models of the earth. Electrical resistivity can then be interpreted, guided by other fields observations, such geological and other geophysical constraints.
Beside resistivity, other parameters can be computed, which provide information of resistivity distribution and dimensionality.

The most important and used is the **phase**, Φ, of the impedance, which is the difference between the phase of **E** and **H**.
In a uniform earth the phase is 45 degrees.

$$\varphi_{ij}(\omega) = \tan^{-1}\left(\frac{Z_{ij}^{I}(\omega)}{Z_{ij}^{R}(\omega)}\right)$$

$$\Phi(\omega) = Z_{xy}(\omega) = \tan^{-1}\left[\frac{(1-i)}{\sqrt{2}}\sqrt{\frac{\omega\mu_0}{\sigma_1}}\right] = -\frac{\pi}{4}$$

In order to satisfy Maxwell's equations, within the earth **E** and **B** vary as

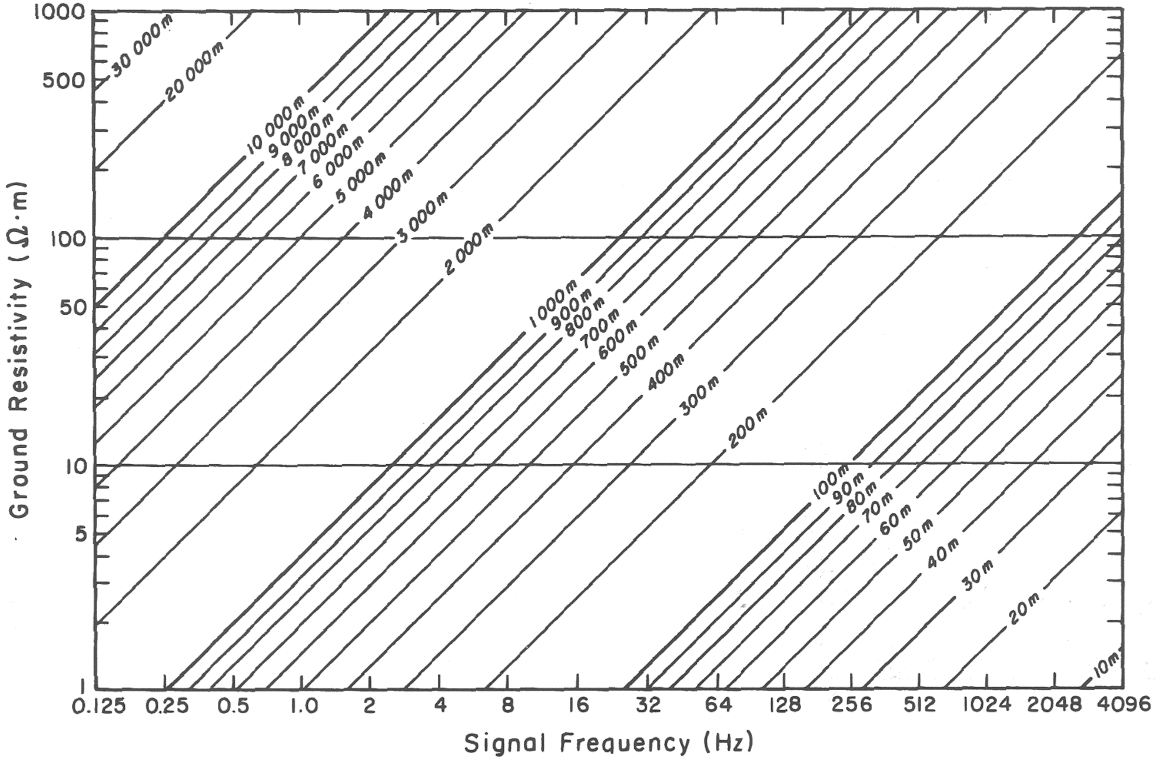
$$A = A_0 e^{-i(kz-\omega t)} = A_0 e^{i\omega t} e^{-i\alpha z} e^{-\alpha z}$$

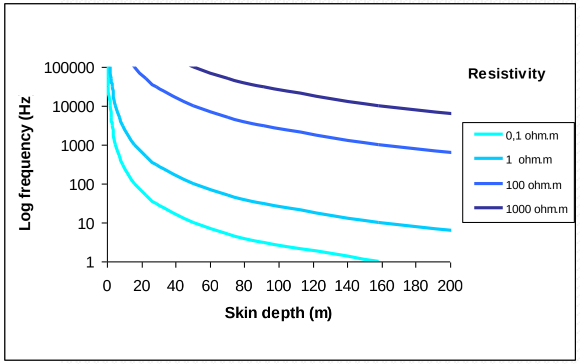
where A_0 is the surface value. That is, the fields vary as the product of four terms:

- $e^{i\omega t}$, a sinusoidal time variation,
- $e^{-i\alpha z}$, a sinusoidal depth variation,
- $e^{-\alpha z}$, an exponential decay with depth, and
- A_0

From the third term, the amplitudes at
$z = \delta = 1/\alpha$ are $1/e$ of their surface values,
which is why δ is called the skin depth.

Plot of skin depth versus frequency and resistivity



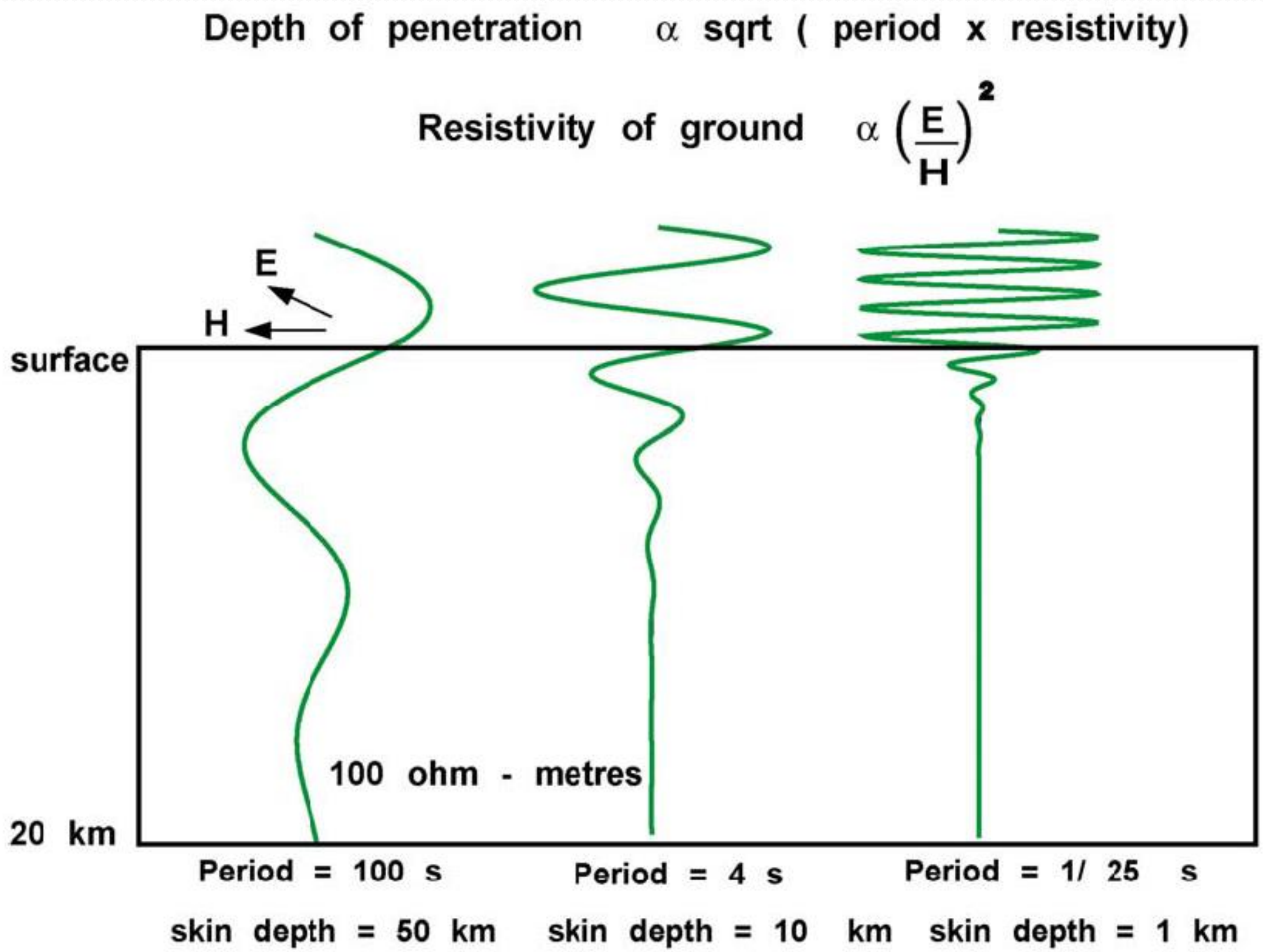


A useful approximation is given by

$$\delta \approx 500 \sqrt{\rho / f}$$

Depth of investigation depends on the resistivity (direct proportionality) and frequency (inverse proportionality)

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



Example: At a frequency of 1 Hz (middle band) a ground of an average resistivity of 1000 ohm-m (many metamorphic rocks and some limestones have this resistivity) can be investigated down to a depth of 15 km

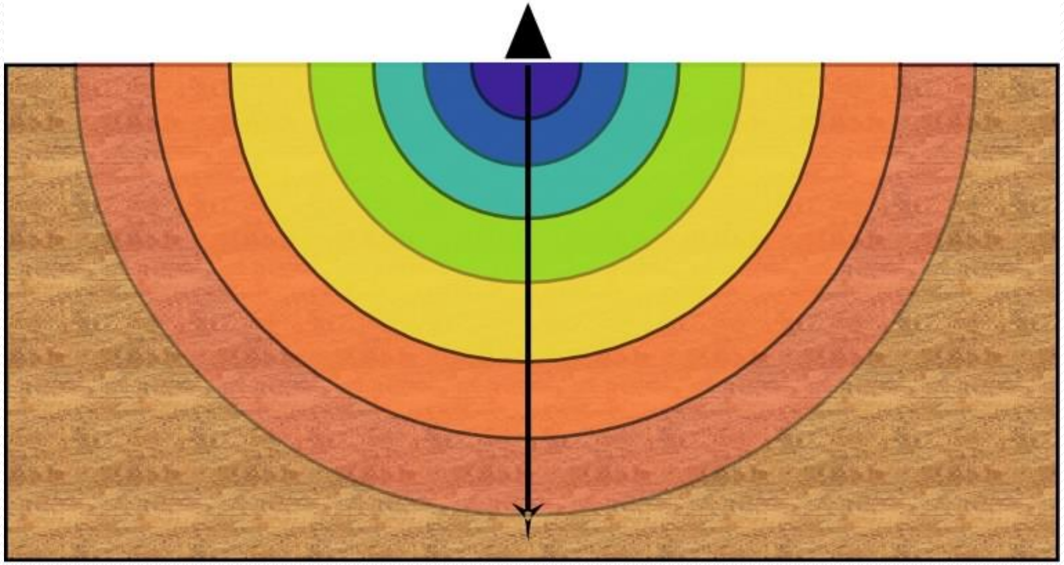
This is the main advantage of MT: large depths can be reached easily, without the need of an artificial source

In the field, we measure the electric and magnetic fields in the frequency band we are interested in

The data are the sum of incident, reflected, transmitted fields: however the ratio E/H, i.e., the impedance, depends only on the geometrical and physical (electric) feature of the ground. Investigation depth depends on the acquired data frequency

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)



In more complicated models, e.g. horizontal layers or 2-D or 3-D structures, the relations between the **E** and **H** fields also become more involved.

In **horizontal layers (1D)** some energy is reflected at each interface, and internal reflection occurs within each layer. The expressions for **E** and **H** include two terms in each layer, of the form:

$$A\,e^{+ikz} + B\,e^{-ikz}$$

one for up-going and the other for down-going energy.

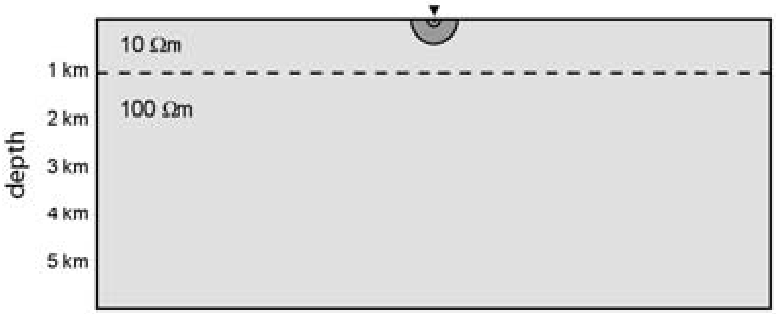
The fields remain horizontal and at right angles to one another unless there is anisotropy in the horizontal plane.

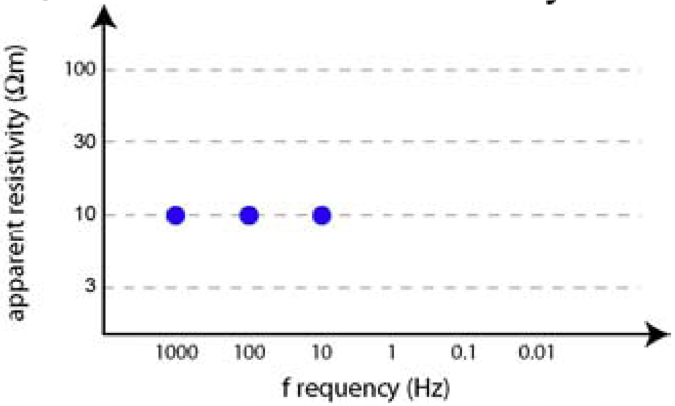
When we pass from homogeneous half-space to 1D, 2D and 3D conditions, the resistivity we compute as a function of impedance is no more the true resistivity of the medium, but an ***apparent resistivity***.

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

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

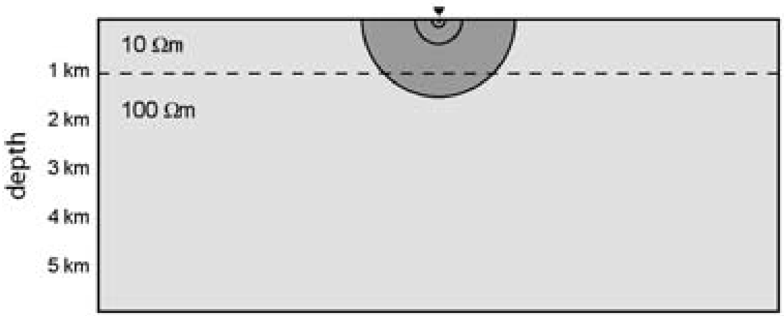
- At high frequency, the skin depth is much less than the thickness of the layer.
- Average resistivity over a hemisphere, radius = δ is the resistivity of the upper layer.

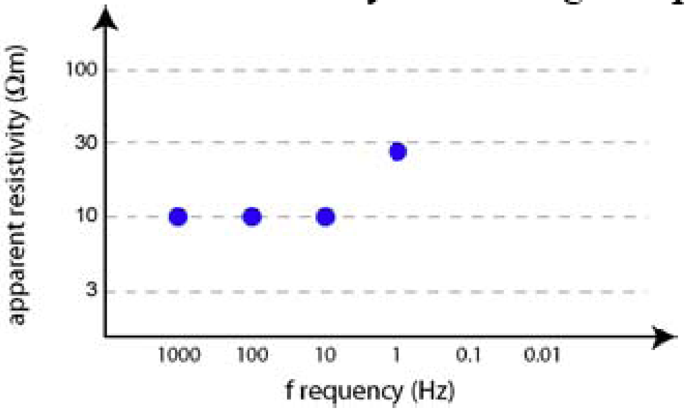




(2) f = 1 Hz δ = 1580 m in upper layer

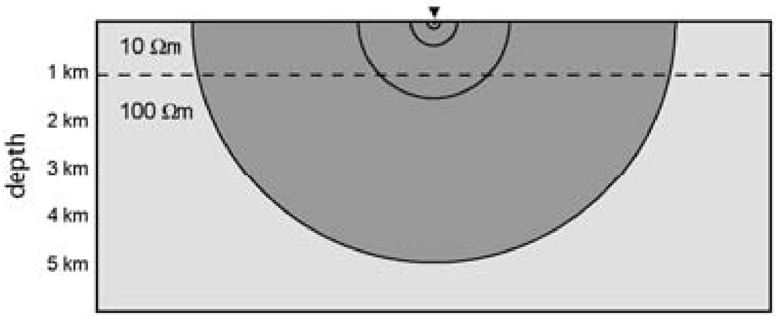
- The EM signals are now just sampling the lower layer,
- Average resistivity begins to increase as the 100 Ωm layer is being sampled.

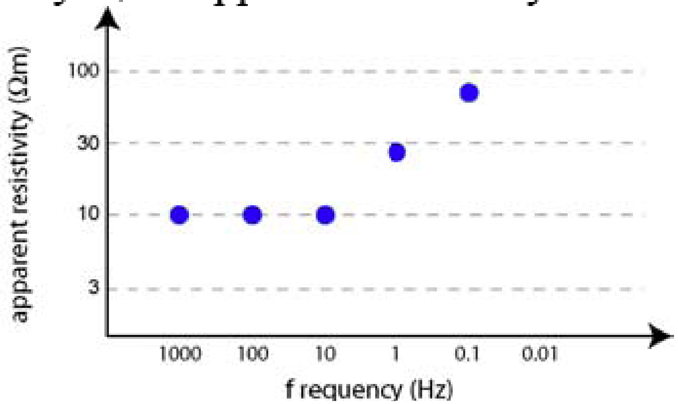




(3) f = 0.1 Hz δ = 5000 m in upper layer

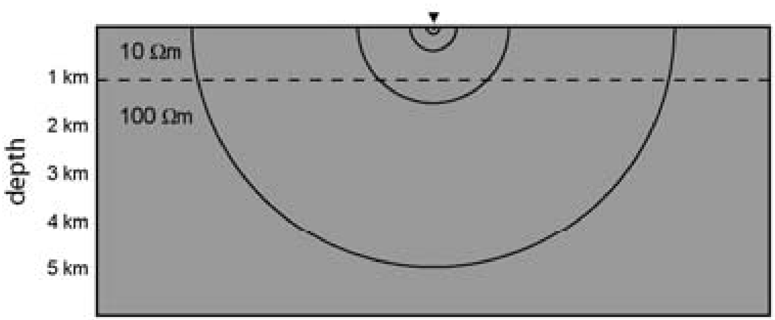
- Hemisphere dominated by lower layer, so apparent resistivity close to 100 Ωm

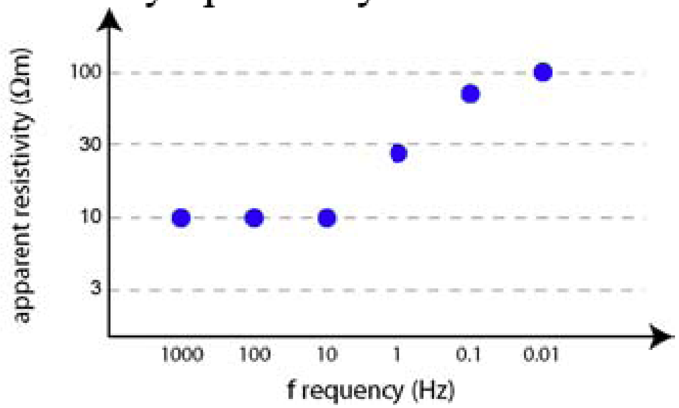




(4) f = 0.01 Hz δ = 15800 m in upper layer

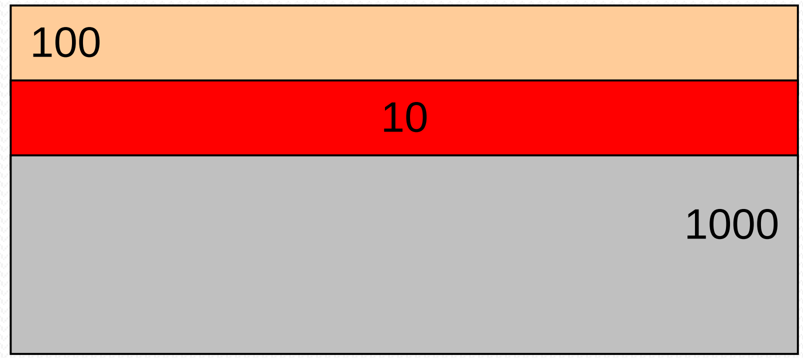
- Apparent resistivity approaches 100 Ωm asymptotically

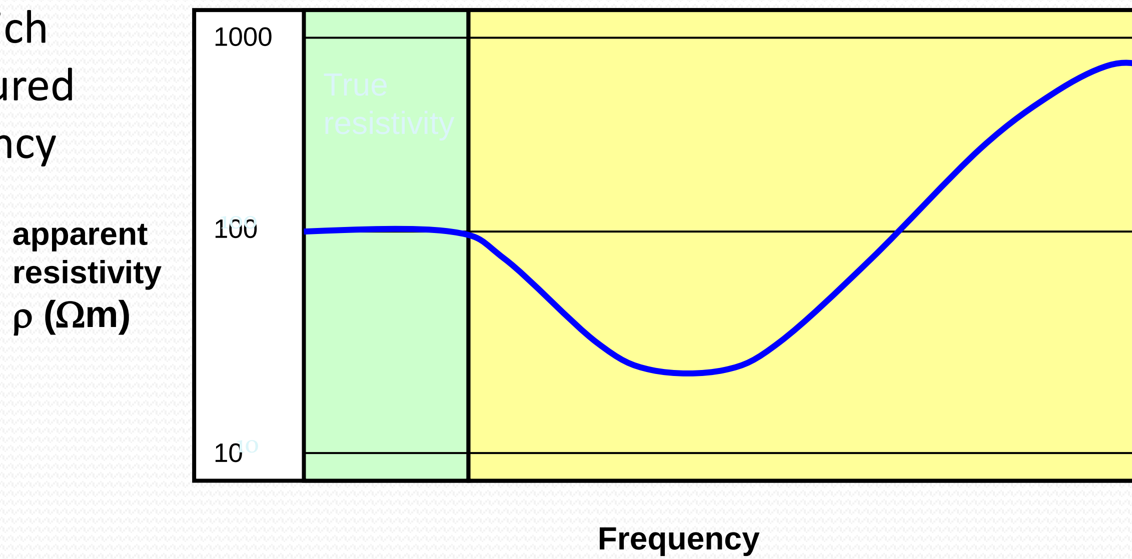




We can still get E_x and H_y (or E_y and H_x) on the surface and do the resistivity calculation, but the result depends on frequency and is now an **apparent resistivity** $\rho_a(f)$.

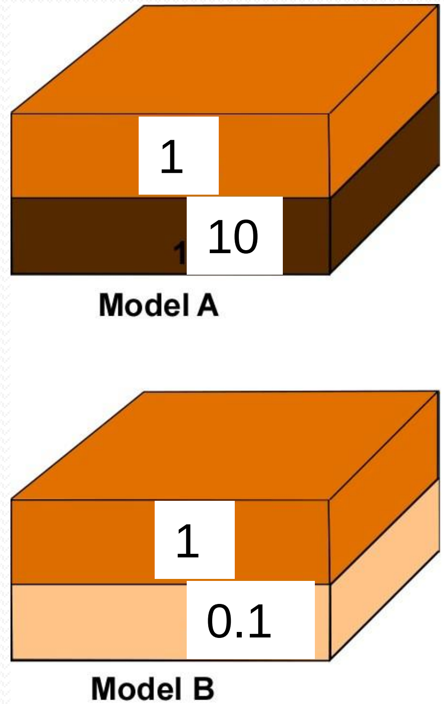
Apparent resistivity is defined as the resistivity of the homogeneous earth which would produce the measured response at each frequency

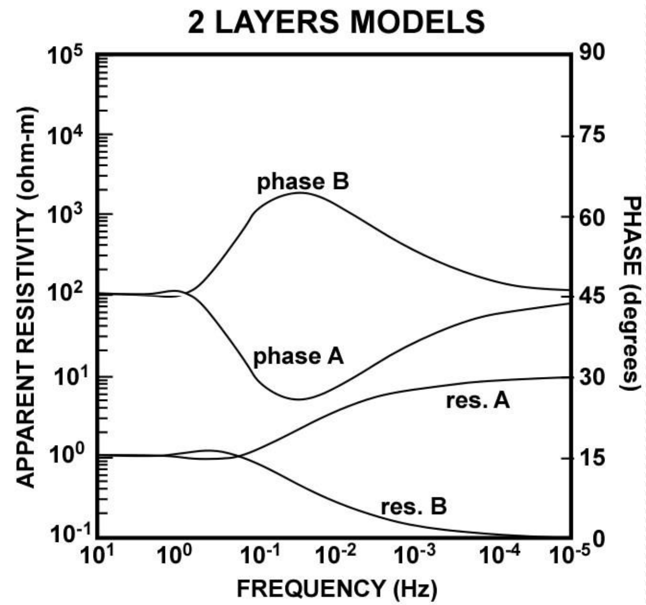




Model A ($\rho_2 = 10\ \rho_1$)

Model B ($\rho_2 = (1/10\ \rho_1)$



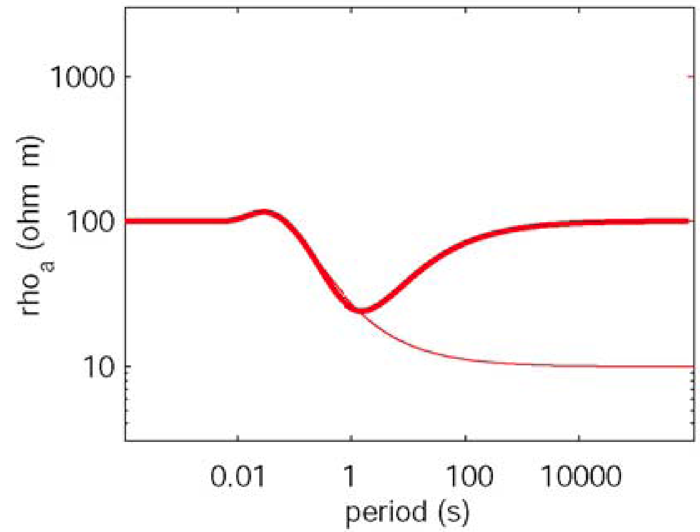


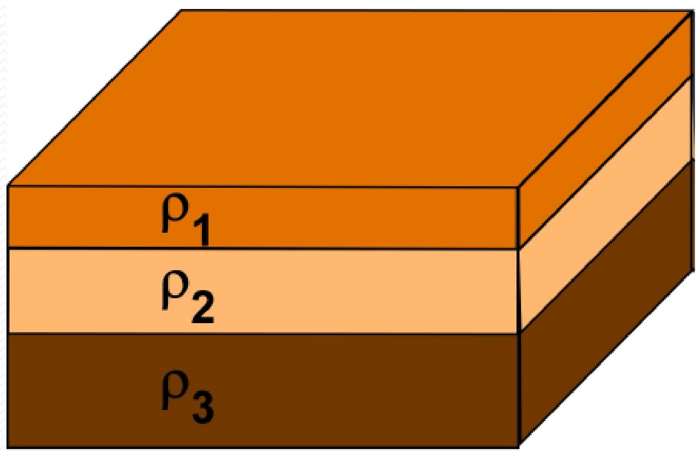
At high frequencies, since the skin depth is much lower than the thickness of the shallow layer, Alle frequenze più alte, $\rho_a = \rho_1$ in both models.

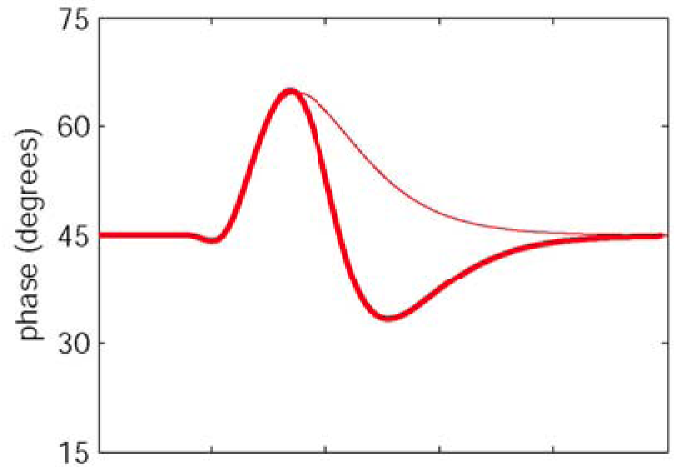
At low frequency the skin depth becomes larger than the thickness of the first layer and ρ_a approaches 0.1 in model B and 10 in model A

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.

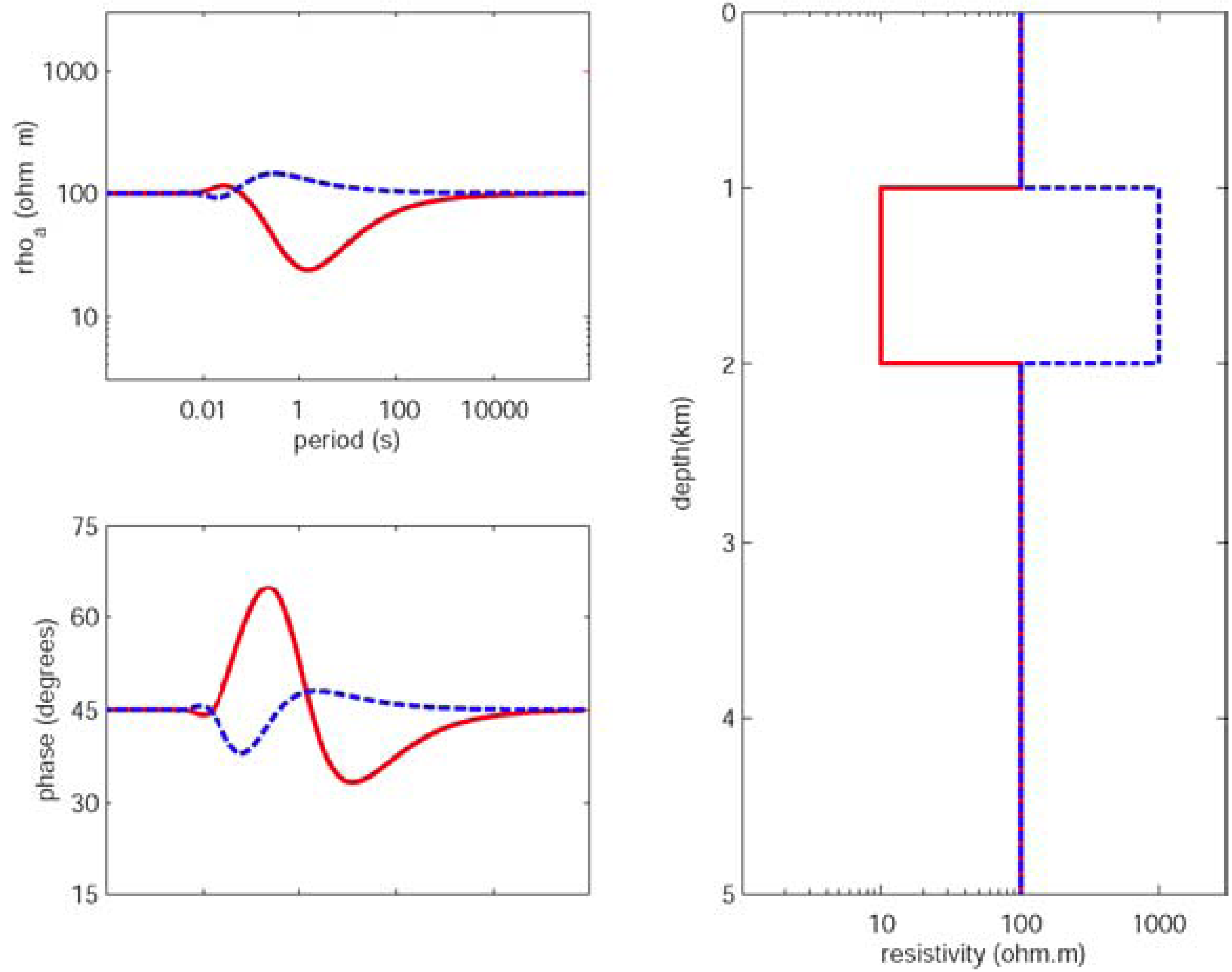
The shape, and therefore our ability to detect differences, depends very much on the relative resistivity and thickness of the layers, on the measured frequency and data quality.



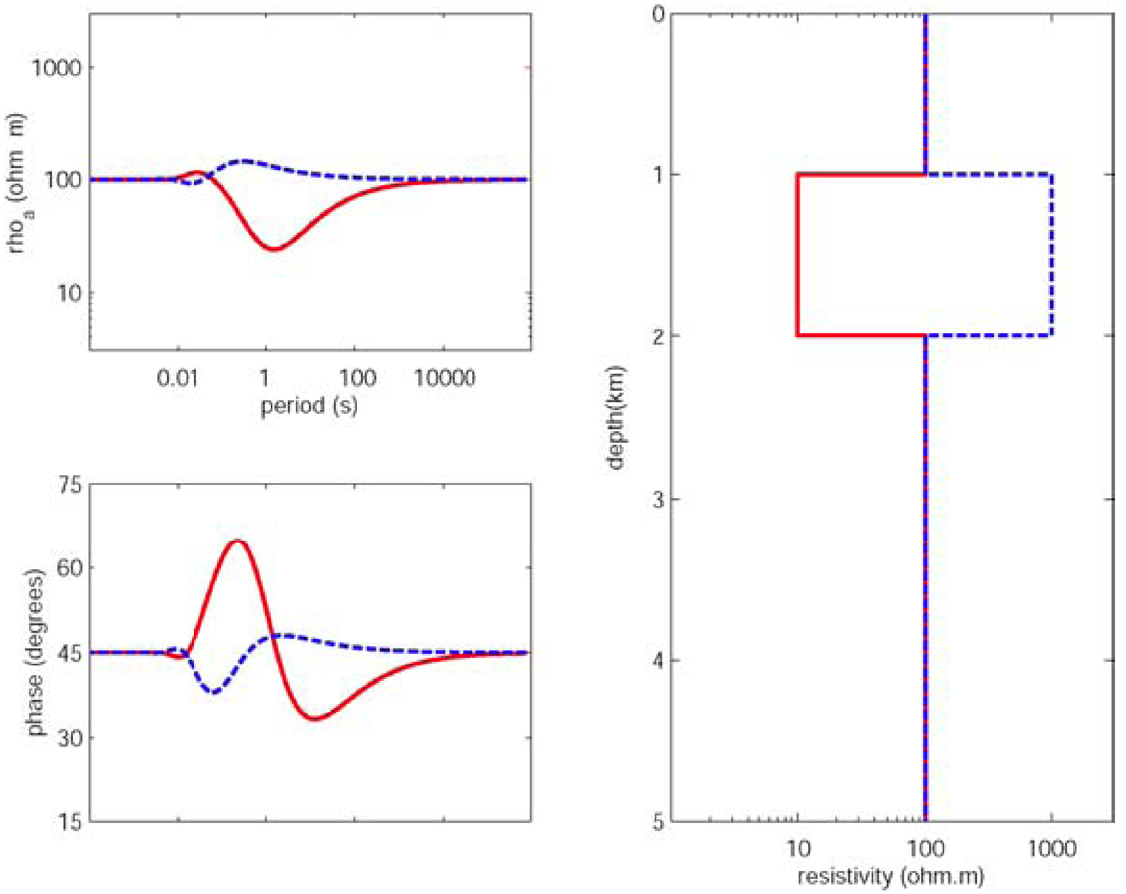




Intermediate conductive or resistive layer

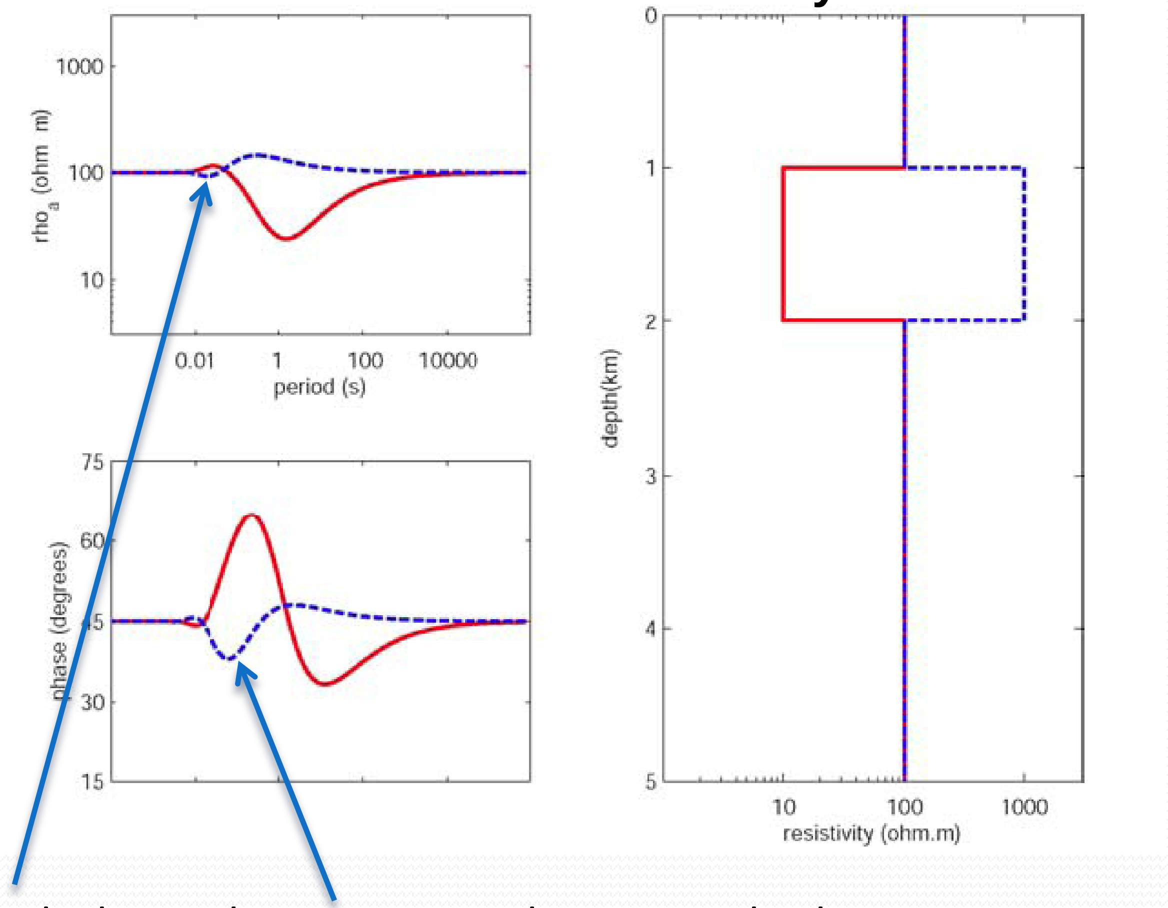


Intermediate conductive or resistive layer



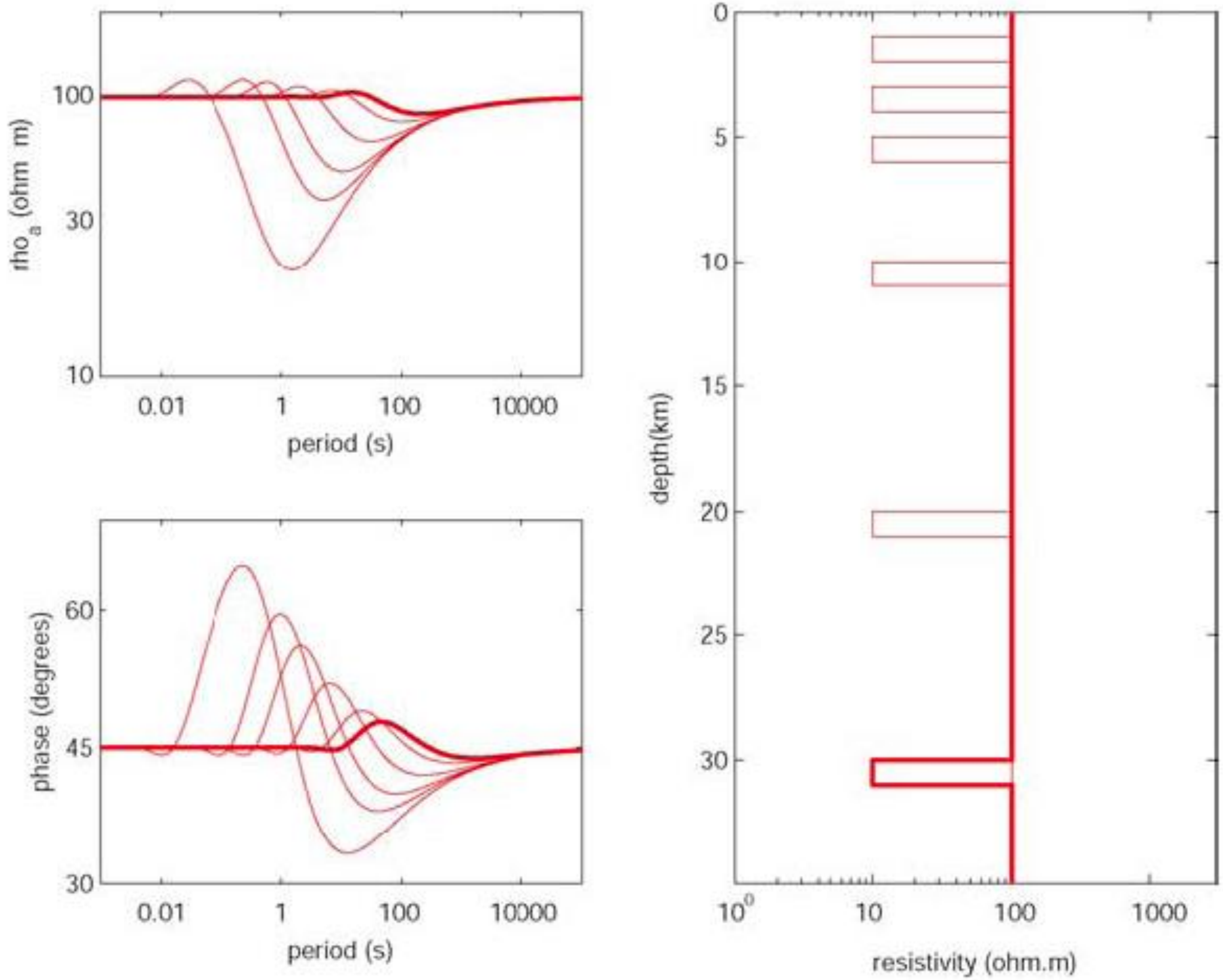
A model with a conductive intermediate layer is more visible since MT field are strongly affected by the conductor

Intermediate conductive or resistive layer

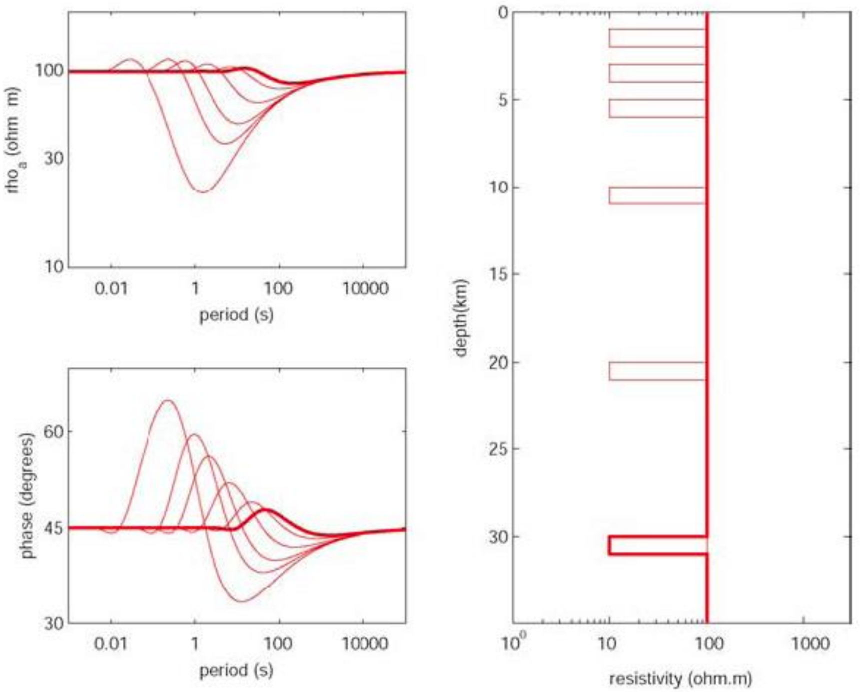


Look also at the resonance phenomena both on apparent resistivity and phase, when skin depth is close to the the first layer thickness (it is seldom observed due to data noise)

Conductive layer at variable depth



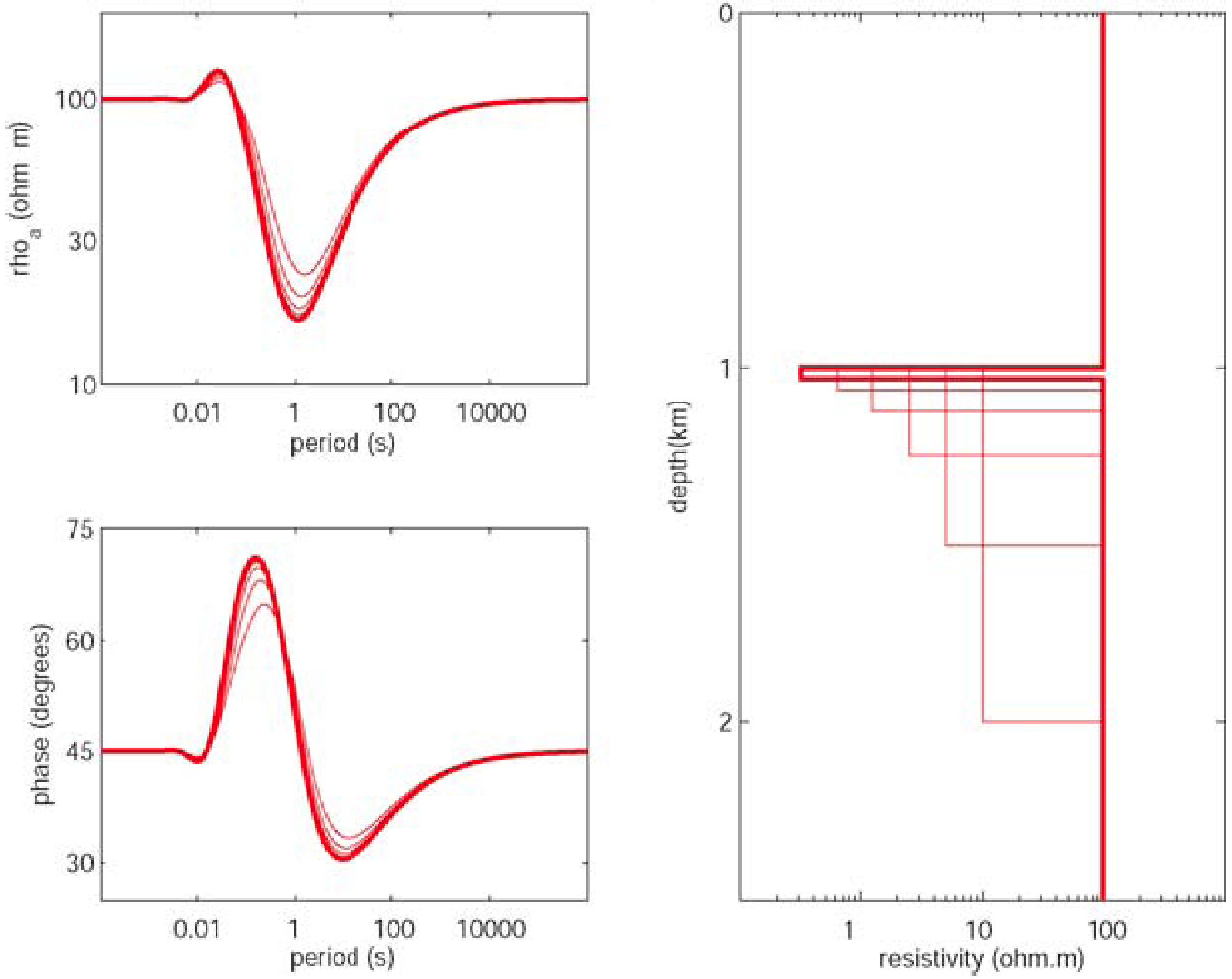
Conductive layer at variable depth



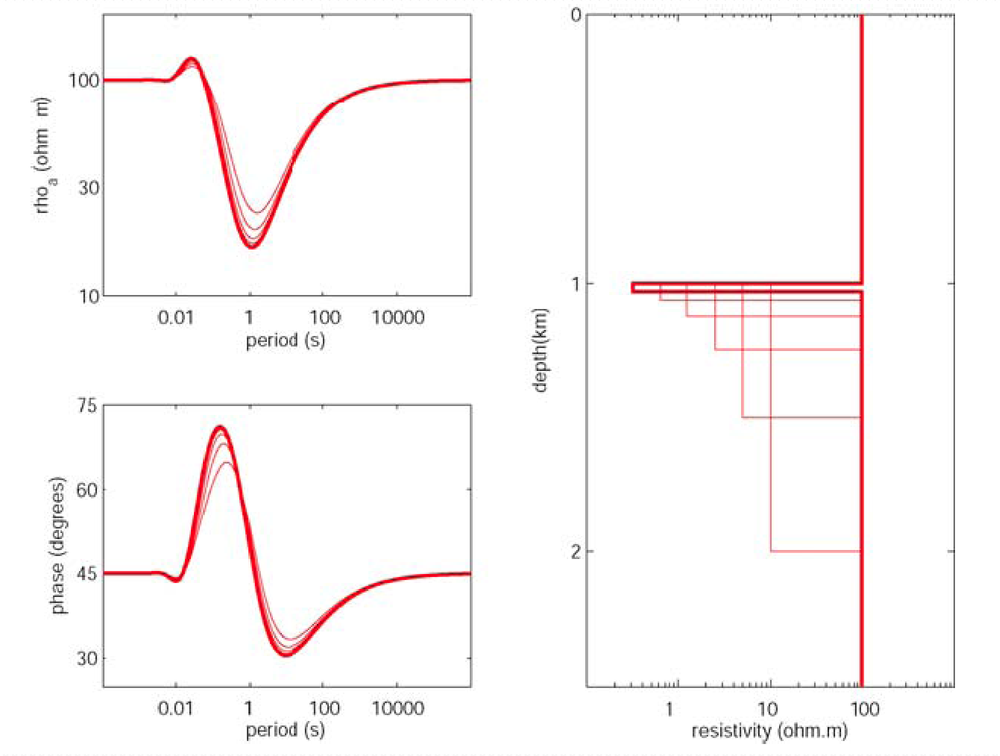
- As foreseen by skin depth equation, increasing depth the period at which we observe the effect of the layer increases (frequency decreases)
- The effect depends on the depth, since it is an apparent resistivity (average from surface down to the layer)
- MT can detect the top of the layer (10% of error)

What about the bottom?

Layer having constant conductance (conductivity x thickness)



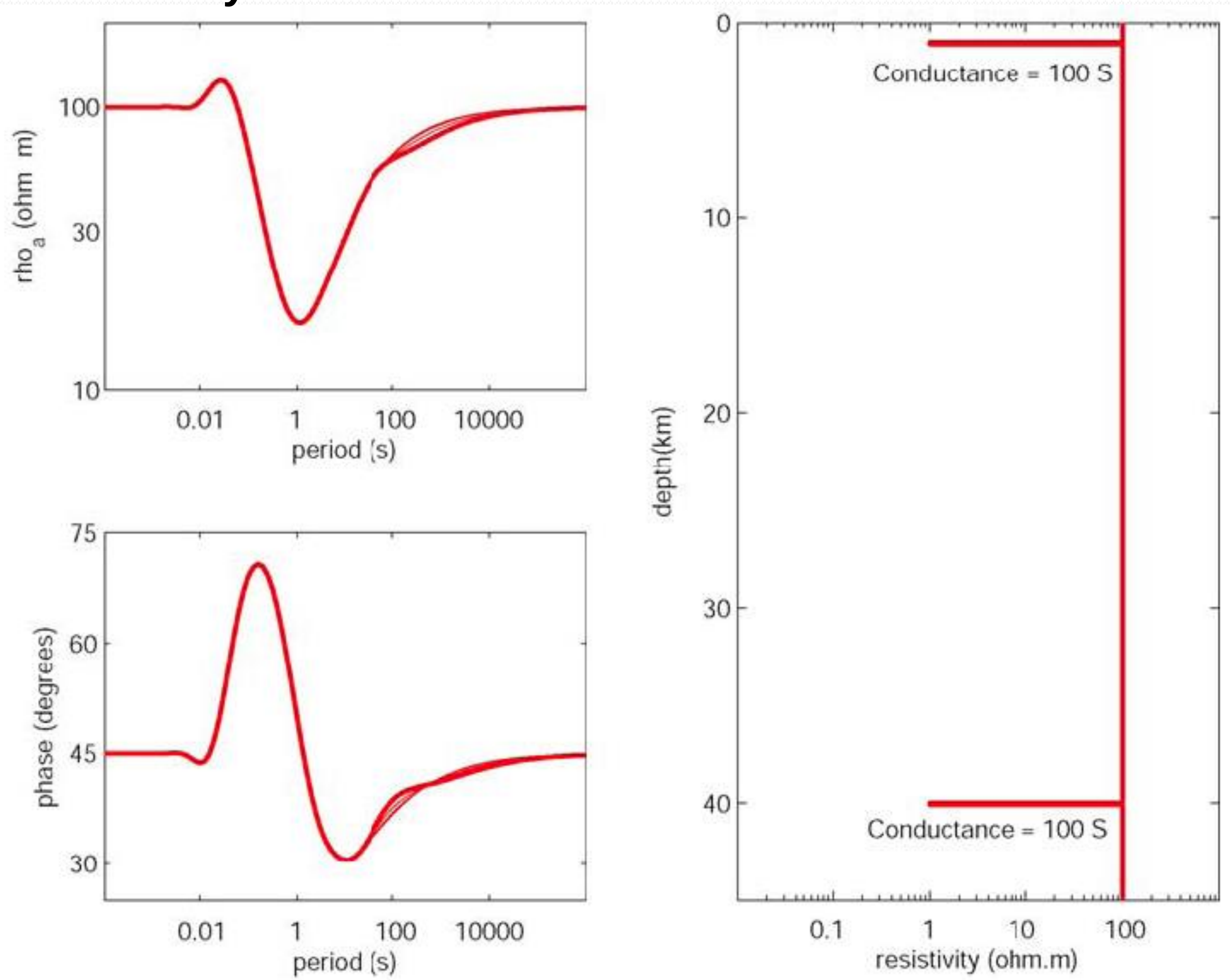
Layer having constant conductance (conductivity x thickness)



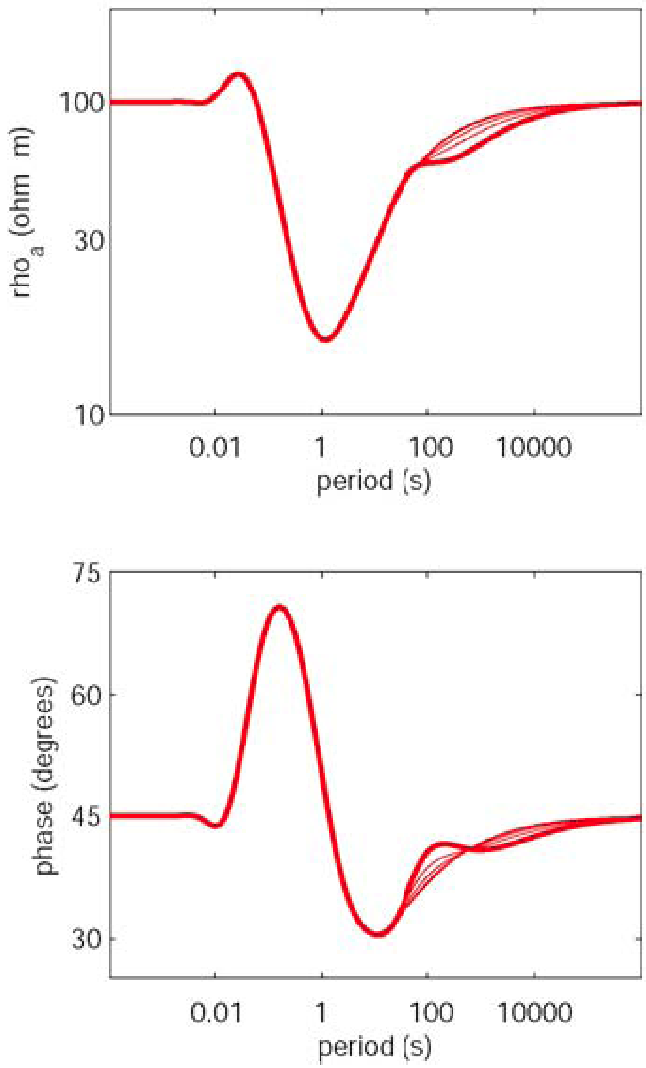
This is an example of **non-unicity**

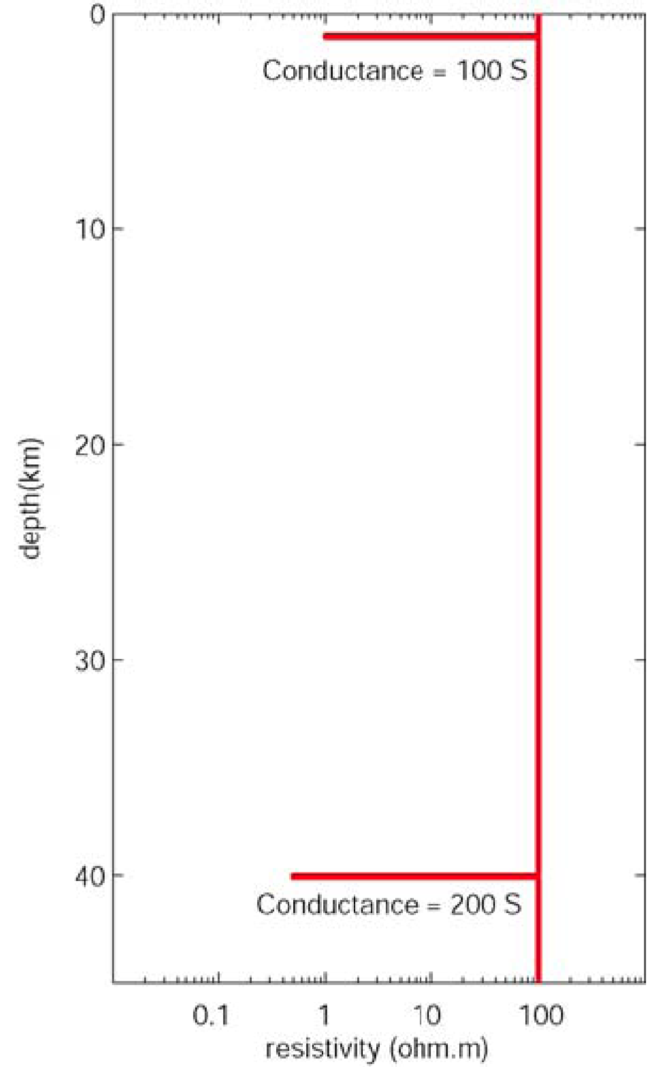
MT can **not** distinguish between layers having the same conductance and needs other information

Two conductive layers

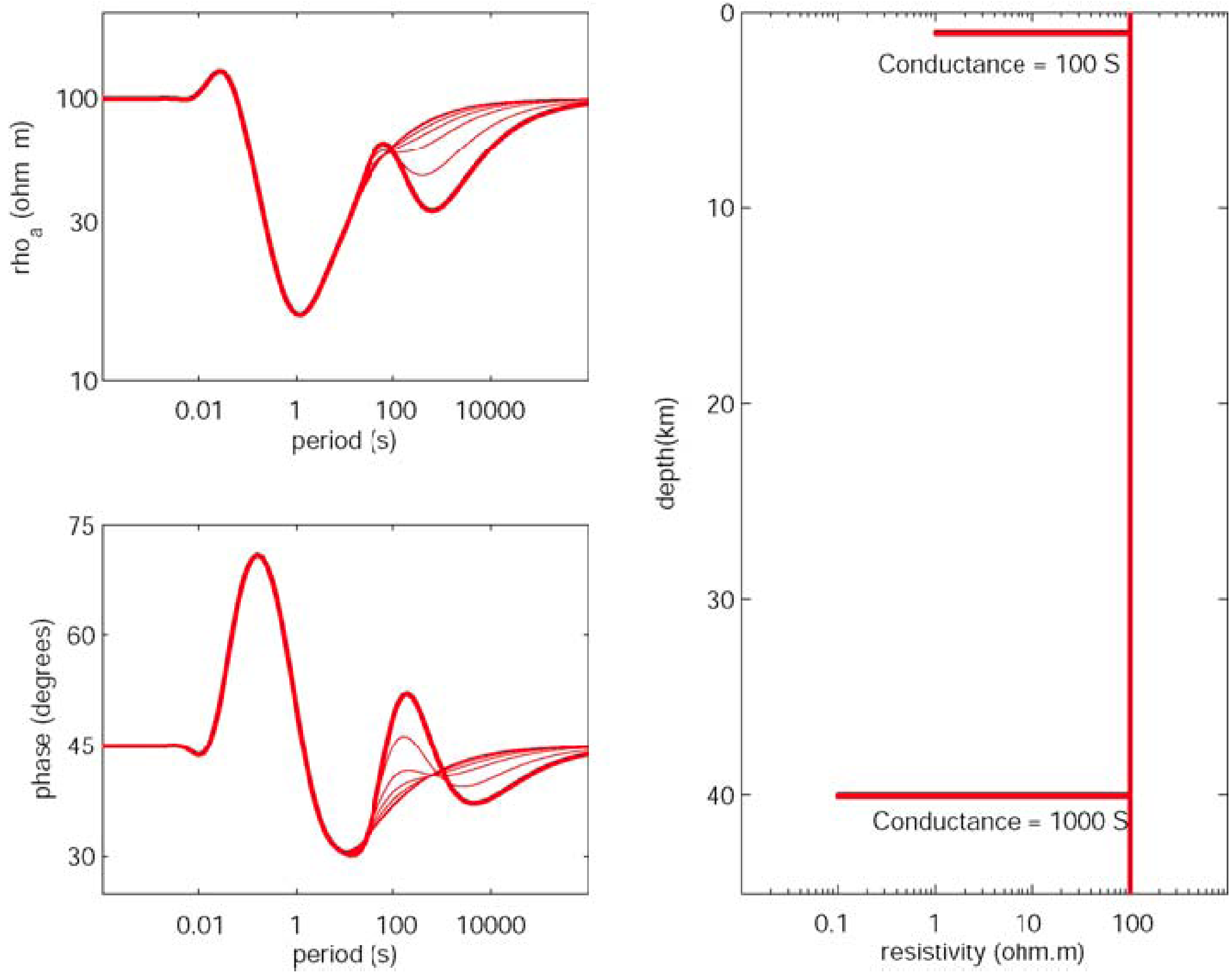


Two conductive layers

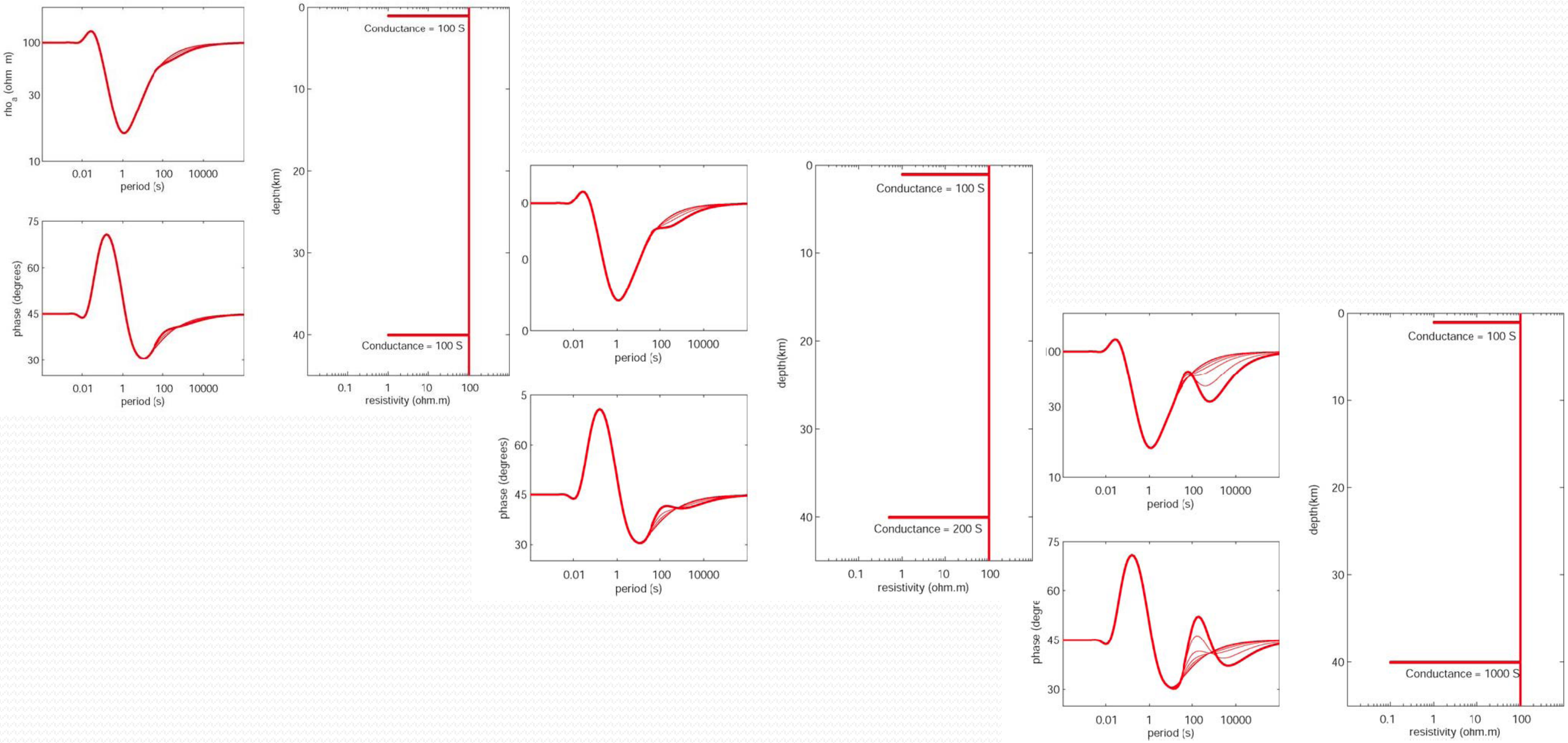




Two conductive layers



Two conductive layers



The deep conductive layer can be defined by MT when its conductance is larger than the one of the shallow conductive layer

The relationships among the field components at a single site are systematically contained in the *impedance* and the *tipper*.
They are the quantities from which conductivity structure is interpreted.

In general, H_x has an associated E_y and some E_x, both of which are proportional to H_x. Likewise, H_y causes an E_x and some E_y, so that at each frequency we would expect a linear system to behave as:

$$E_x = Z_{xy} H_y + Z_{xx} H_x$$
$$E_y = Z_{yx} H_x + Z_{yy} H_y$$

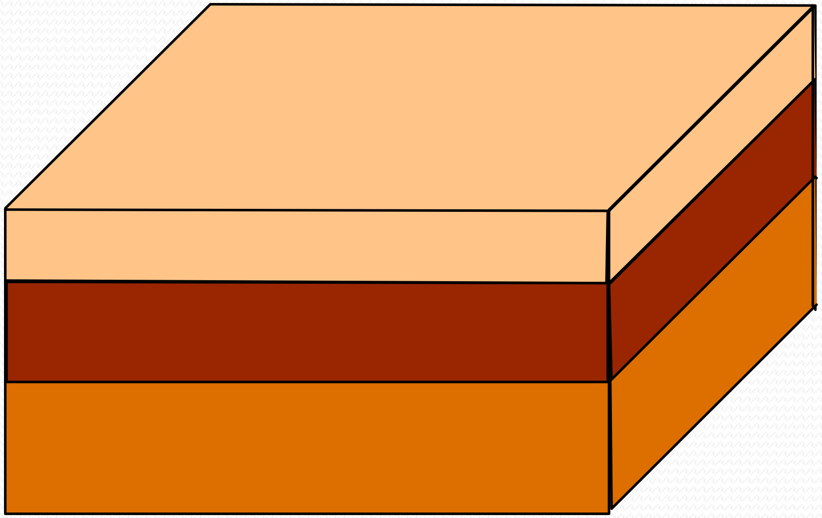
where each term is frequency dependent. This is commonly written as:

$$\begin{pmatrix} E_x \\ E_y \end{pmatrix} = \begin{pmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{pmatrix} \begin{pmatrix} H_x \\ H_y \end{pmatrix}$$

or

$$\mathbf{E} = \mathbf{ZH}$$

Uniform or horizontally layered earth: 1D case



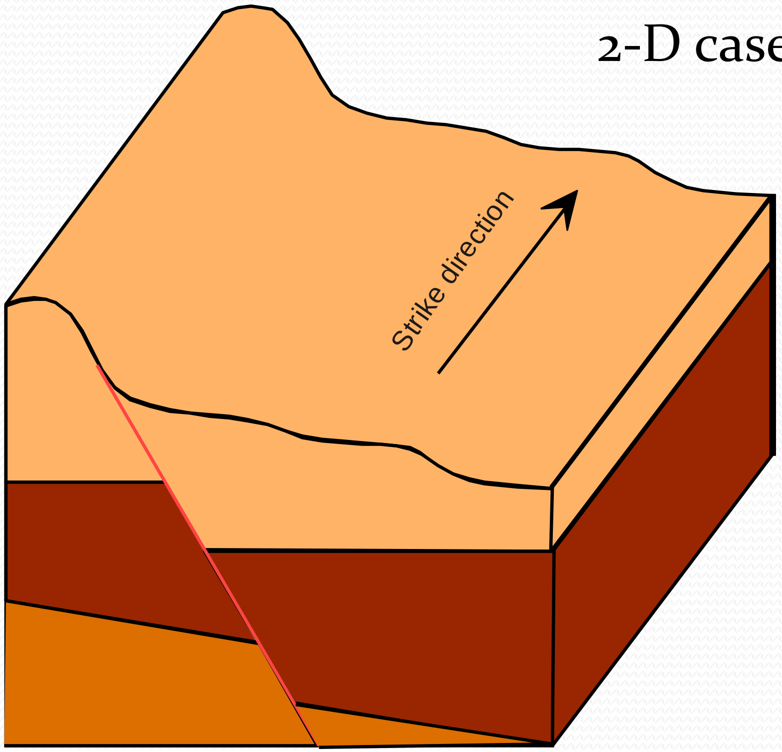
$$Z_{xx} = Z_{yy} = 0$$

$$Z_{yx} = -Z_{xy}$$

and the equations reduce to

$$E_x = Z_{xy} H_y$$

$$E_y = Z_{yx} H_x = -Z_{xy} H_x .$$

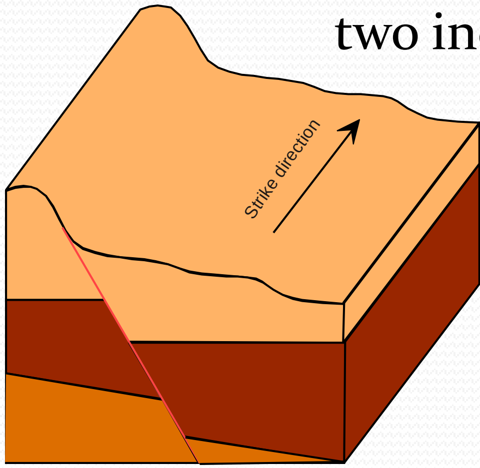


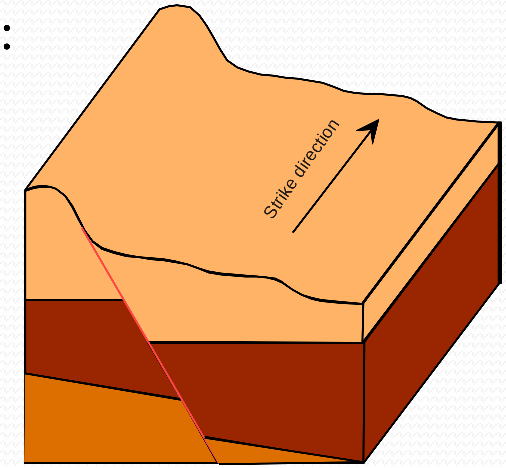
2-D case

Consider a 2-D Earth where the geological strike is parallel to the *x-axis.* Electrical resistivity varies in only the *y and z directions.*
Resistivity does not vary in the *x-direction and all structures extend to x =* $\pm\infty$
In contrast to a 1-D Earth, vertical field components (E_z and H_z) can be non-zero.

2-D case

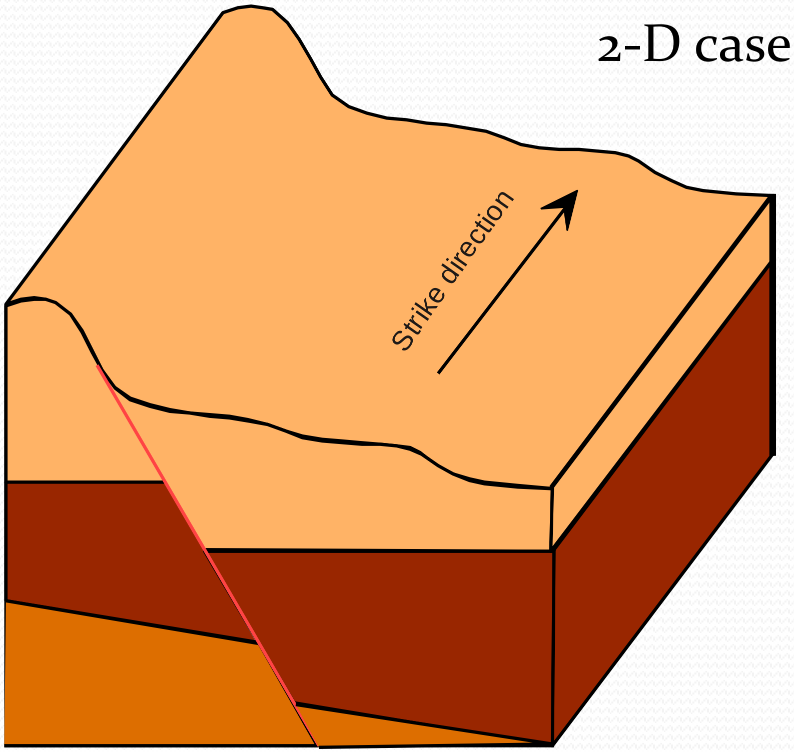
By Maxwell equation we can derive equations can now be separated into two independent subsets:





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



2-D case

If the x or y axis is along the strike then

$$\begin{bmatrix} E_x \\ E_y \end{bmatrix} = \begin{bmatrix} 0 & Z_{xy} \\ Z_{yx} & 0 \end{bmatrix} \begin{bmatrix} H_x \\ H_y \end{bmatrix}$$

$$Z_{xx} = Z_{yy} = 0$$

$$Z_{yx} \neq - Z_{xy}.$$

If neither axis is along the strike

$$Z_{xx} = -Z_{yy} \neq 0$$

The subsurface structure can be studied by making simultaneous measurements of the strength of the magnetic field variations at the surface of the earth and the strength of the electric field component at right angles in the earth.

Because the direction of polarisation of the incident magnetic field is variable and not known beforehand, it is common practice to measure at least two components of the electric field and three components of the magnetic field variation to obtain a fairly complete representation.

For surveys that are intended for the study of electrical structures tens of kilometres in depth, the range of frequencies needed to achieve penetration is from a few tens of Hertz to a few hundred microHertz.

In practice we are often interested in regarding the fields or the tensor elements as if they had been measured in some other set of coordinate directions.
For instance, the dimensionality (1D, 2D or 3d) or the strike direction is seldom known very precisely at the time of a field survey.

For this reason we acquire data as a set of orthogonal measurements, and we then rotate by using some math.

If we rotate the vector **E** field through angle +θ (clockwise as seen from above)
to be **E**', then:

$$E'_x = \cos\theta\ E_x + \cos\theta\ E_y$$
$$E'_y = -\sin\theta\ E_x + \sin\theta\ E_y$$

or

$\mathbf{E}' = \mathbf{R}\mathbf{E}$.

In the same way,

$\mathbf{H}' = \mathbf{R}\mathbf{H}$

and

$\mathbf{Z}' = \mathbf{R}\mathbf{Z}\mathbf{R}^T$,

where $\mathbf{R}^T$, the transpose of **R**, is

$$\mathbf{R}^T = \begin{pmatrix} \cos\theta & -\sin\theta \\ \sin\theta & \cos\theta \end{pmatrix}$$

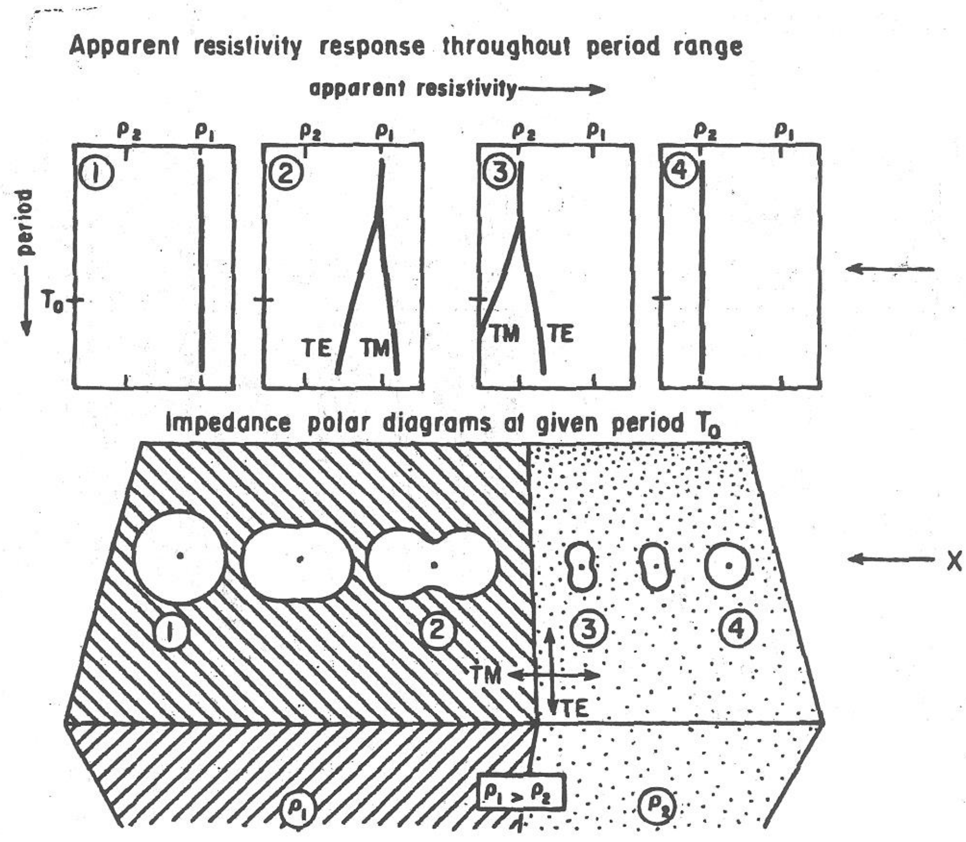
Starting from a tensor impedance **Z** which is derived from measurements, we compute the Z components (Z_{xx}, Z_{xy}, Z_{yx}, Z_{yy}) and we derive the corresponding resistivity a phase values.

If the underground is 1D whatever the rotation we would find just one value of impedance: $Z_{xy}=-Z_{yx}$ and the resistivity and phase xy and yx curves are identical.

For 2-D conditions, several different means have been used to find the rotation angle θ_0 between measurement direction and strike. All of them, however, start from the idea the when Z is measured along strike direction, Zxx and Zyy are equal to zero.

One of these is to rotate the Z_{ij} in steps (say 5 degrees), plot them on a polar diagram, and pick an optimum angle from the plots. An optimum angle maximizes or minimizes some combination of the Z_{ij}.

These interesting diagrams, called *polar figures* or *impedance polar diagrams,* are usually plotted at many frequencies, because in practice the strike direction often changes with depth.



Another way to find θ_o is to differentiate $Z_{xy}(\theta)$ and $Z_{yx}(\theta)$ with respect to θ to give an angle θ_o which optimizes:

$$|Z'_{xy}(\theta_o)|^2 + |Z'_{yx}(\theta_o)|^2$$

at each frequency. His solution

$$4\theta_o = \tan^{-1} \frac{\left[(Z_{xx} - Z_{yy})(Z_{xy} + Z_{yx})^* + (Z_{xx} - Z_{yy})^*(Z_{xy} + Z_{yx})\right]}{|Z_{xx} - Z_{yy}|^2 - |Z_{xy} + Z_{yx}|^2}$$

also maximizes $|Z_{xy}|$ and minimizes $|Z_{xx}|^2 + |Z_{yy}|^2$.

There is no solution in the 1-D case, whereas in a clearly 2-D case it usually has a definite value.
In the 3-D case its meaning is usually questionable and there is considerable research in progress into ways to present and interpret Z_{ij} in structural terms.

Of the four values between 0 and 180 degrees, the ”choice” of strike direction is started by evaluating the mentioned optimization at two adjacent values, one a minimum and the other a maximum. This leaves four possible solutions at 90 degree intervals, or two possible strike directions. The choice between these solutions can only be made from independent information, usually the relations between vertical and horizontal **H** components or from geological constraints.

When the coordinates are rotated, certain combinations of terms are constant even though the individual terms vary. These are:

$$Z_{xx} + Z_{yy} = c_1,$$
$$Z_{xy} - Z_{yx} = c_2,$$

and

$$Z_{xx} Z_{yy} - Z_{xy} Z_{yx} = c_3,$$

where the absolute value is the determinant of the impedance tensor. The ratio c_1/c_2 is the impedance skew.

c_1 will be zero in (noise-free) 1-D and 2-D models, so the skew is used as a measure of three dimensionality. It does not change with rotation of coordinates.

A quantity which *does* vary with setup direction is impedance ellipticity,

$$\beta(\theta) = \frac{Z_{xx}(\theta) - Z_{yy}(\theta)}{Z_{xy}(\theta) + Z_{yx}(\theta)}$$

This is zero (for noise-free data) in the 1-D case, and in the 2-D case when the *x* or *y* axis is along the strike.

Impedance ellipticity, like impedance skew, is used to indicate whether response at a site is 3-D.

It can usually be assumed that $H_z \approx 0$ except near lateral conductivity changes, where $\nabla \times \mathbf{E}$ has a vertical component

$$H_z = T_x H_x + T_y H_y$$

where the elements T_i are complex since they may include phase shifts.

Given a 2-D structure with strike in the x′ direction, in those coordinates the previous equation simplifies to:

$$H_z = T_y' H_y'$$

Here, T′, since it represents a tipping of the **H** vector out of the horizontal plane, is called the tipper.

T' is of course zero for the 1-D case.

The modulus of T' varies between 0.1 and 0.5. The lower part of the range is often blurred by noise, since H_z is so weak.

The required rotation angle φ to x′ can be estimated by the field data by finding the horizontal direction y′ in which H(φ) is most highly coherent with H_z. There is generally a definite solution in the ”almost” 2-D situation. In that case the phases of T_x and T_y are the same, the ratio T_y/T_x is a real number, and

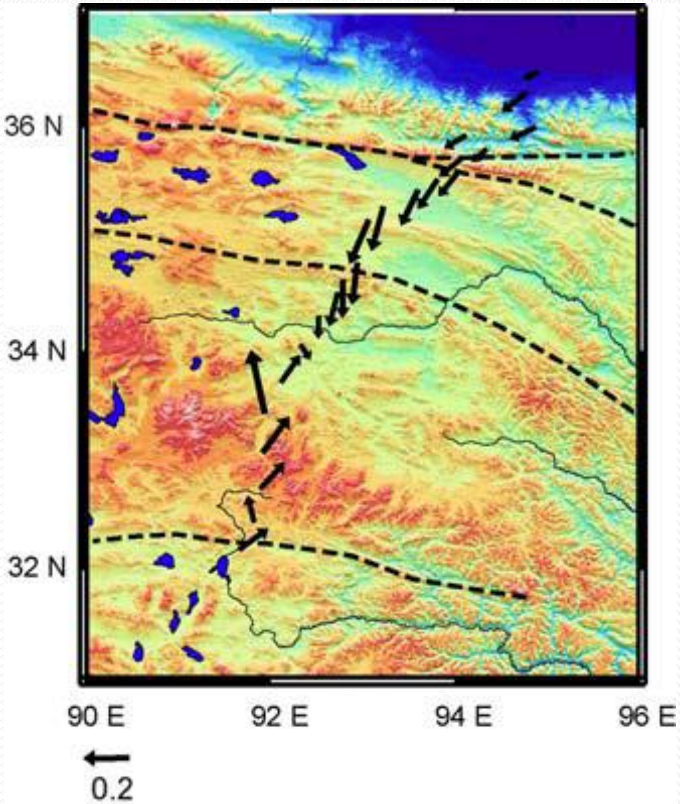
$$\varphi = \arctan (T_y/T_x).$$

When Hz is ≠ 0 a useful tool can be define: the induction vectors, which are computed with components

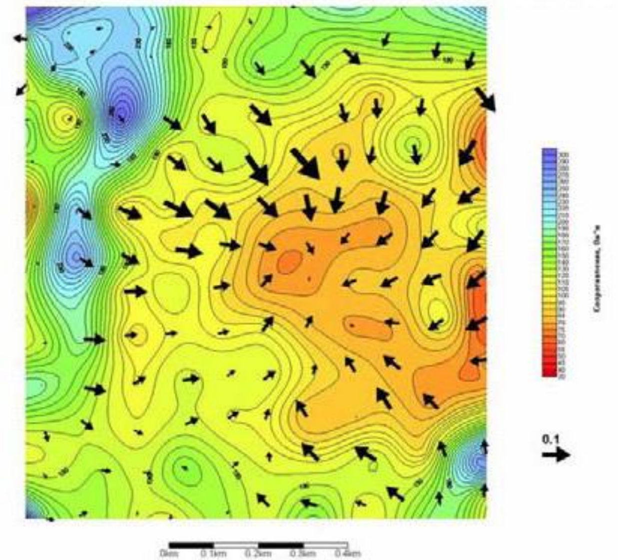
$I_y = H_z/H_y$ and $I_x = H_z/H_x$

In the *Parkinson convention*, induction vectors will point at conductors.
In the *Schmucker convention* they point away from conductors.
Measuring vertical magnetic fields requires a lot of physical effort. Burying an induction coil requires a 1 m hole to be dug vertically downwards.

The location of the reversal denotes the center of the conductor.



Induction vectors (Parkinson Convention) in Tibet
(from Unsworth, 2004)



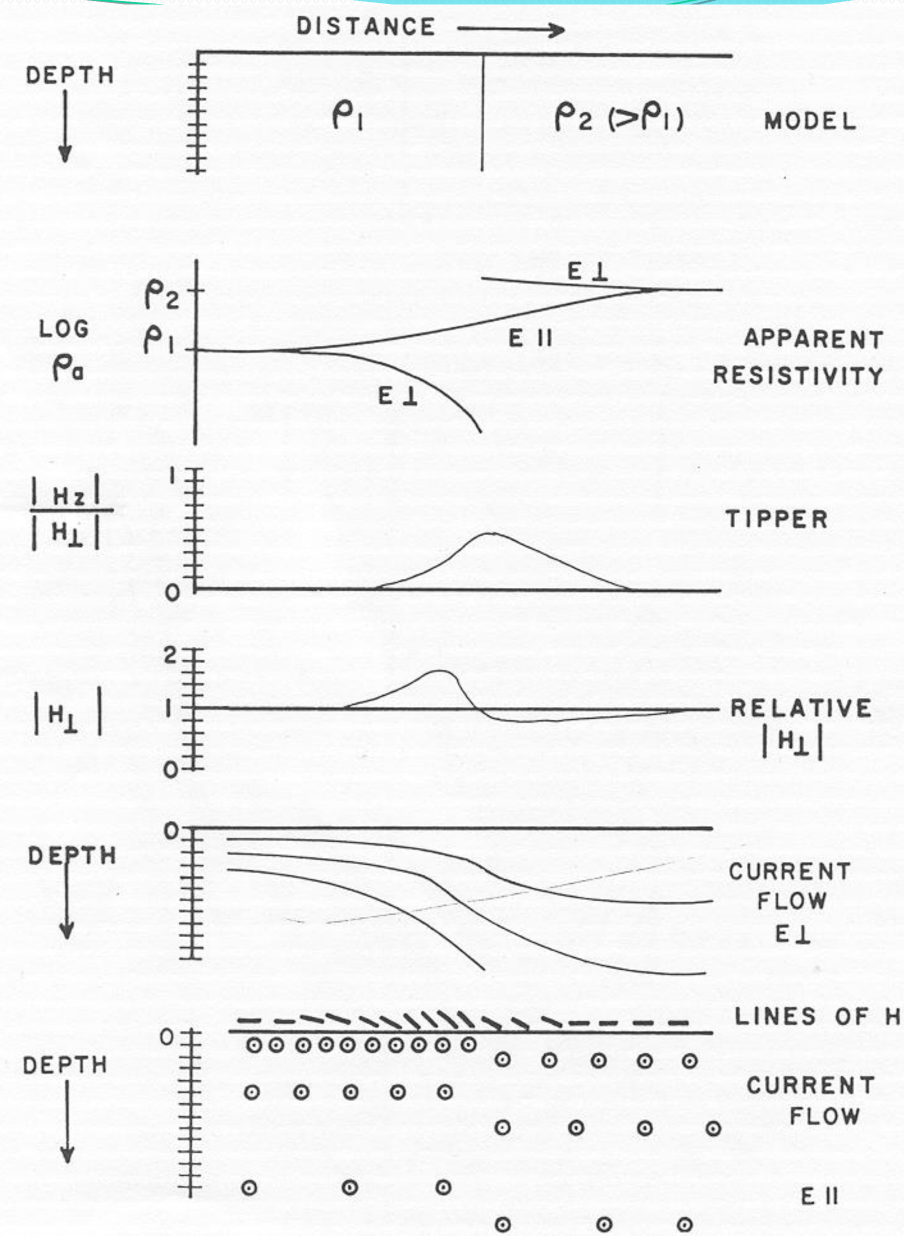
Induction vectors (Parkinson Convention)above a Siberian kimberlite pipe, which forms a low resistivity zone
(courtesy of the Nordwest company)

Definition of a 2D condition

Boundary conditions:
E parallel to the strike ($E_{||}$) is continuous, whereas the current density of
E orthogonal to strike ($E_{\perp}$) is continuous.

$E_{\perp}$ is lower where the field is greater (conductive medium) and vice versa.

The same for impedance and hence for resistivity

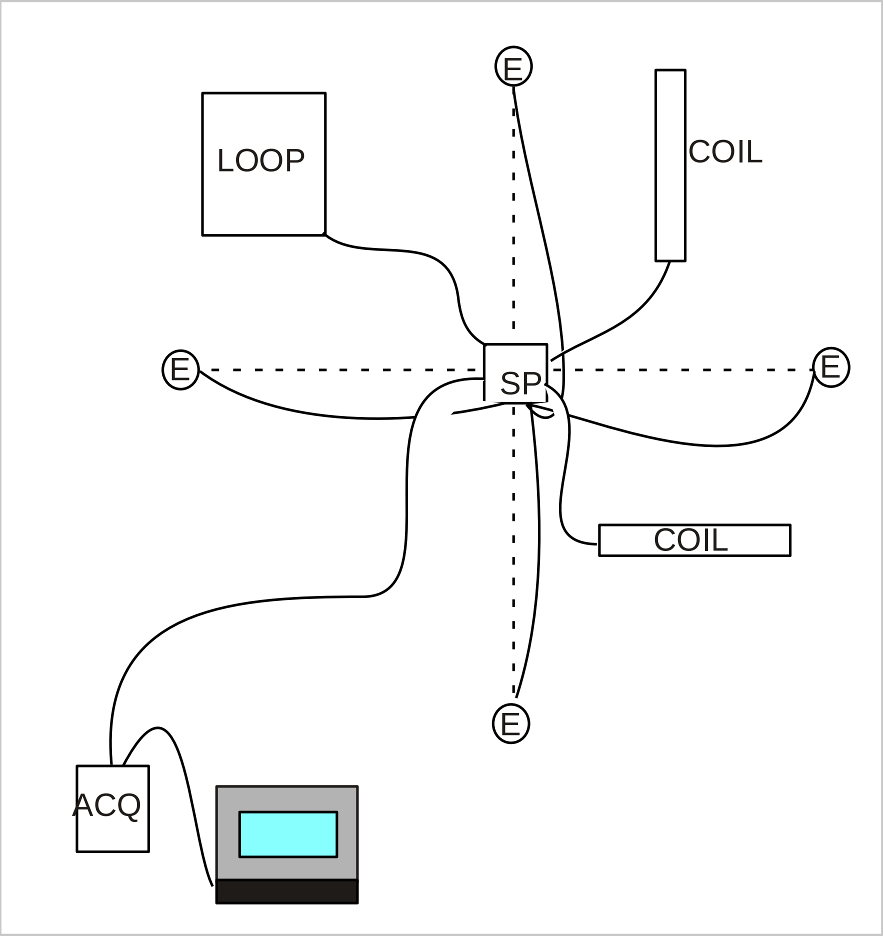


MT data are acquired in the field, as measurements of electric and magnetic fields with time

$E_x(t)$, $E_y(t)$, $H_x(t)$, $H_y(t)$, $H_z(t)$

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.



Electric field measurements

~100 m dipole with non-polarizing electrode at each end. The instrument measures the **voltage** between the electrodes. Usually two dipoles are used, one NS and one EW.

Wire should be placed on the ground so it can’t move

Metal stakes can be used above 50 Hz. For lower frequency as they corrode they generate noise, and can also become polarized. Non-polarizing electrodes (pots containing a metal stake embedded in a porous mixture saturated with a salt of the same metal. Usually Pb-ClPb)

Magnetic field measurements

3 components of the magnetic field are usually measured at each station.

At high frequency (above 0.01 Hz) an induction coil is used. This is a cylinder with millions of turns of copper wire. A change in the magnetic field along the axis of the coil will induce a voltage in the coil. Thus only relative measurements of the magnetic field are made. The coils are usually buried to minimize motion and give thermal stability.

For lower frequencies a 3-component flux gate magnetometer is used. These give absolute measurements of the magnetic field with a precision to 0.01 nT, so there is no lowest frequency at which they work.

Audio magnetotellurics (AMT)

Few hours recording time per site.

Exploration to 1-2 km depth

Data from 10000-1 Hz.

Induction coils for magnetic sensor

Broadband magnetotellurics (BBMT)

Typically 1-2 days recording per site.

Used in exploration to 10 km, often in commercial exploration.

Data from 1000 Hz to 1000 s.

Induction coils for magnetic sensor

Long period magnetotellurics (LMT)

Several weeks recording at each station.

Used for deeper exploration, primarily in academic studies.

Data from 1-10000 s.

Fluxgate magnetometer for magnetic sensor

Inasmuch as the natural noise field is not particularly well structured, but consists of an unpredictable assemblage of impulsive waveforms, it is necessary to analyse the natural field over a time span which is long compared to the period of the lowest frequencies being studied.

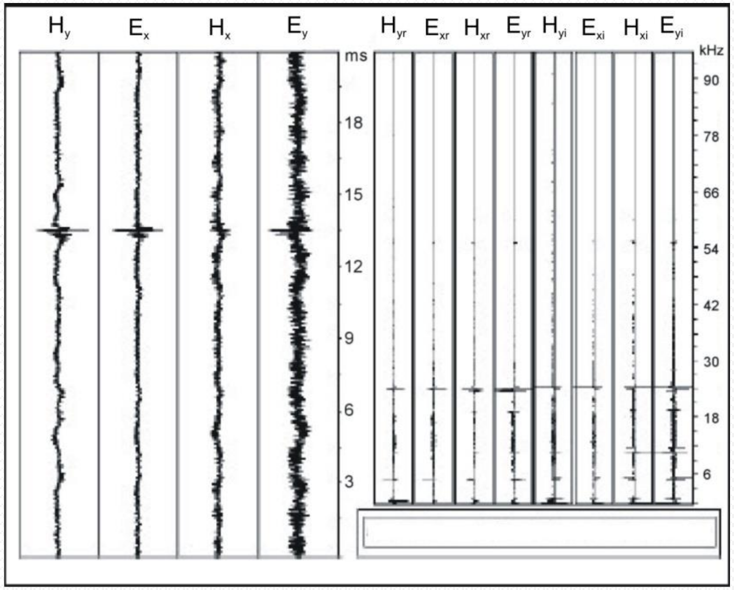
Thus, if the lowest frequency desired in a survey is 500 μHz (period of 2000 s), it is necessary to analyse the field over a time duration at least 10 times as great, or 20 000 s.

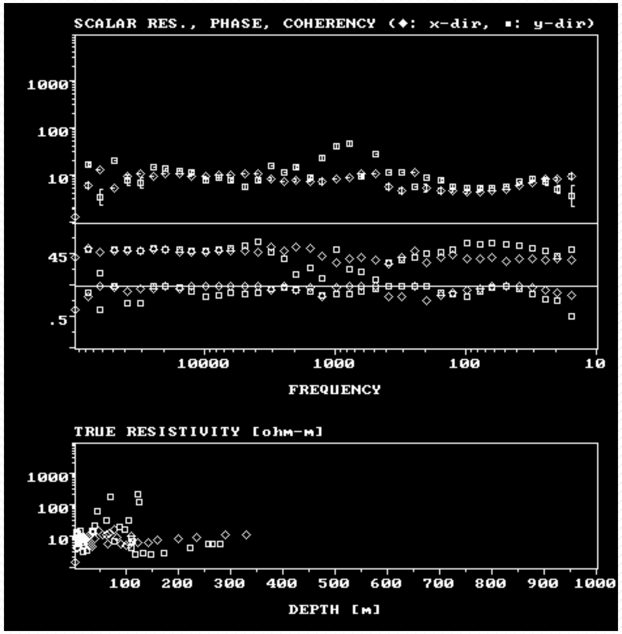
This represents the single largest disadvantage of the magnetotelluric method: it provides slow coverage of a prospect area and is therefore costly.

In most systems for carrying out magnetotelluric surveys today, the five field components are converted to digital form and are either stored for later spectral analysis or converted immediately to spectral form before being stored.

The accuracy with which the data are converted to digital form is important; a dynamic range of 24 bits is desirable so that weak spectral components will be recognisable in the presence of other, stronger spectral components.

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.





In order to obtain spectra at frequencies as high as several tens of Hertz, a maximum sampling rate of 100 Hz is required.
To increase the statistical quality, oversampling and decimation is often performed, increasing the sampling frequency up to 2560 Hz.
The lowest frequency that will be detected in a time series is $f = 1/L$ where L is the length of time in seconds.

Averaging over multiple cycles of the signal is required to get an accurate estimate of a given frequency. This typically requires at least 20 cycles.

Broadband MT example : To measure $\rho_a(\omega)$ and $\Phi(\omega)$ at 20 Hz**, at least 1 second** of data is needed.
Long period MT example : To estimate $\rho_a(\omega)$ and $\Phi(\omega)$ at a period of 20,000 s (5.5 hours), data should be recorded for **at least 110 hours. In reality, longer** recordings are often needed since the MT signal is not always present.

Magnetotelluric time series analysis

The 5 field components are then Fourier transformed into the frequency domain to give estimates of
$E_x(\omega)$, $E_y(\omega)$, $H_x(\omega)$, $H_y(\omega)$, $H_z(\omega)$

Apparent resistivity, ρ_a (ω), and phase, $\Phi(\omega)$, are then computed as a function of frequency.
ρ_{xy} (ω) and $\Phi(\omega)_{xy}$ are computed from E_x *and* H_y
ρ_{yx} (ω) and $\Phi(\omega)_{yx}$ are computed from E_y *and* H_x

If the Earth is 1-D then $\rho_{xy}(\omega) = \rho_{yx}(\omega)$
Generally this is not the case and 2-D modelling and inversion is needed (possibly 3-D).

In the real world MT signal, being so weak, is always affected by some kind of noise

Typical sources of noise include:

•Ground motion (wind, distant ocean, foot traffic, trains …)

•Electric power lines (even after a notch filter is used at 50 or 60 Hz)

•Electric trains (DC is really bad with ground return)

•Cathodically protected pipelines

•Water pumps

•Vehicles and trains

•Electric fences

•Animals eating cables

•Water flowing over electrodes

One important development in MT has been improved signal processing (*i.e.* making sure the Fourier transform only takes account of *E-H field combinations* that are **coherent. This obviously requires that local noise be excluded from the** analysis.

Coherence is computed as the cross-correlation between the electric and magnetic fields. If the fields are linearly related, coherence is unity; if there is noise in any of the field components which produces a spectrum that does not obey the fundamental equations above, the coherence will be reduced.

When coherence drops below reasonable values (0.85 to 0.90), it is common practice to discard the apparent resistivities that are calculated.

In recent years, attempts have been made to eliminate uncorrelated signals that appear on one or more of the field components.

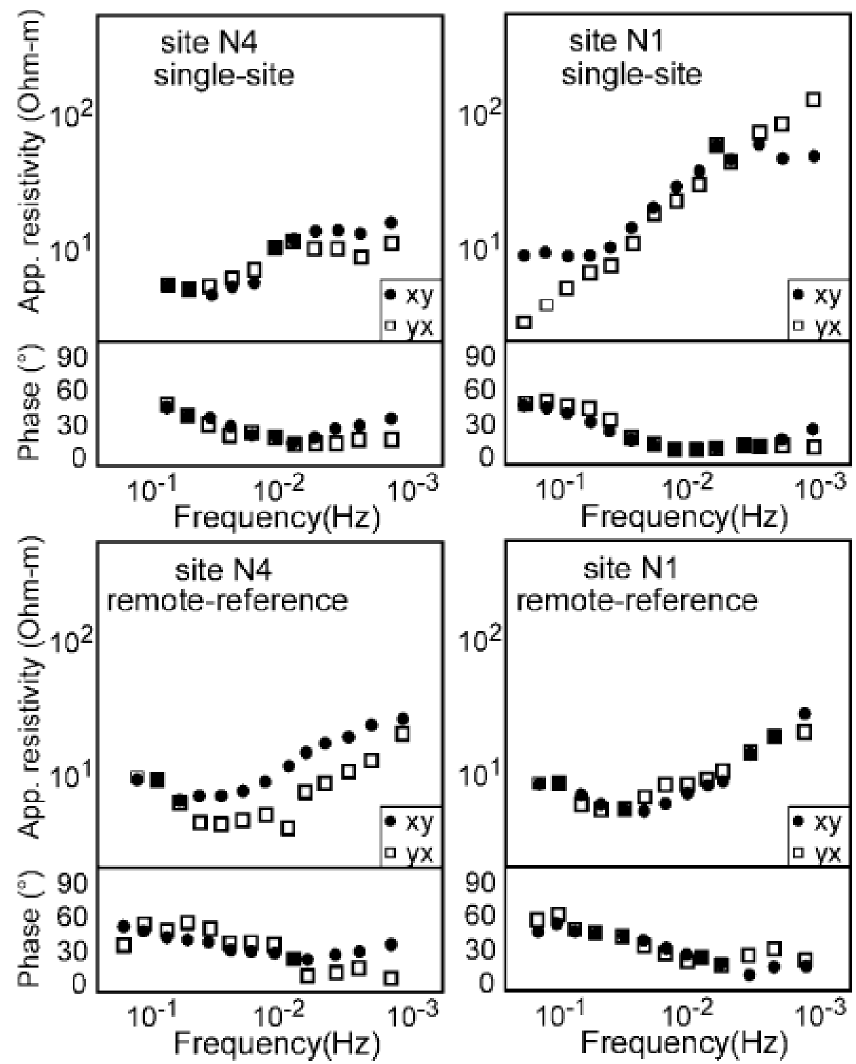
At the present time, the most popular approach is that where two or more magnetotelluric soundings are recorded simultaneously at different sites. This should measure the
same signal but different noise.

Magnetic fields are correlated between the two or more receiver stations, with uncorrelated portions being considered as noise in the magnetic field detection, and removed for later processing.

This method, which requires at least two synchronised MT sites, is called *remote-reference* acquisition

In areas where uncorrelated noise has been a problem in obtaining magnetotelluric soundings, this procedure has resulted in significant improvements in the quality of the data.

Over a portion of the frequency range where noise is a particular problem (from 0.1 to 10 Hz), the multiple-station approach has permitted data to be obtained where previously it had been impossible.



How remote should the remote reference be? A remote MT station should be out of range of the noise, but close enough to measure essentially the same signal.

To eliminate ground motion, 500 m is often enough.

To eliminate the effect of DC electric trains, 500 km might be needed.

MT data should always be collected with a remote reference in operation!

As might be expected, the spectral analysis of a long data series, combined with the need for extensive tensor rotation and testing of the spectral values, result in a volume of processing that is as time-consuming and costly as the original acquisition of data.

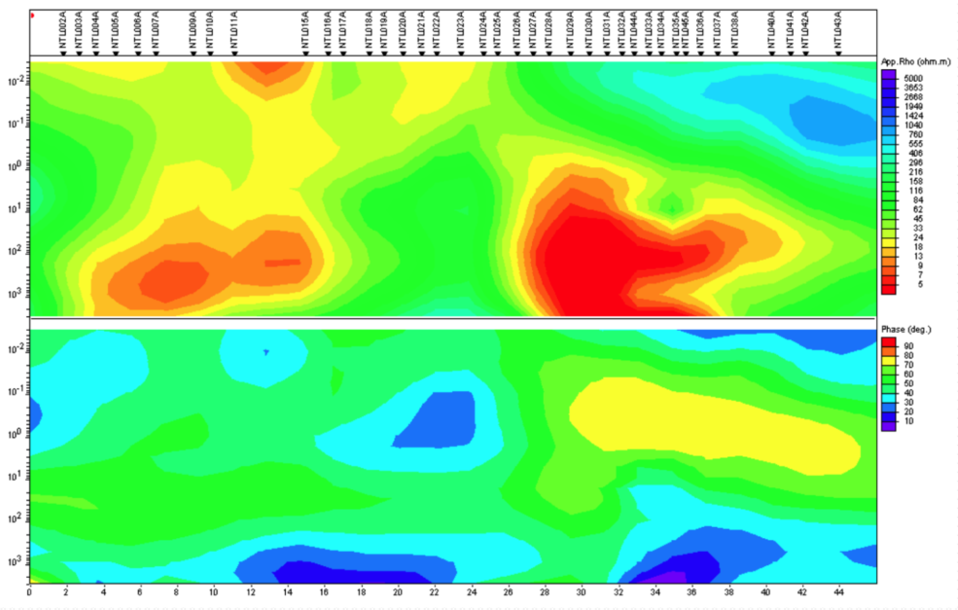
The current trend is to compute the spectral analyses and the rotation of the tensor impedance in the field. This is highly desirable in that the magnetotelluric method does not always provide useful results, even after measurements have been made with reliable equipment.

If the natural electromagnetic field strength is unusually weak during a recording period, or if there is some phenomenon which precludes an effective analysis of the field, it may be necessary to repeat the measurements at a more favourable time.

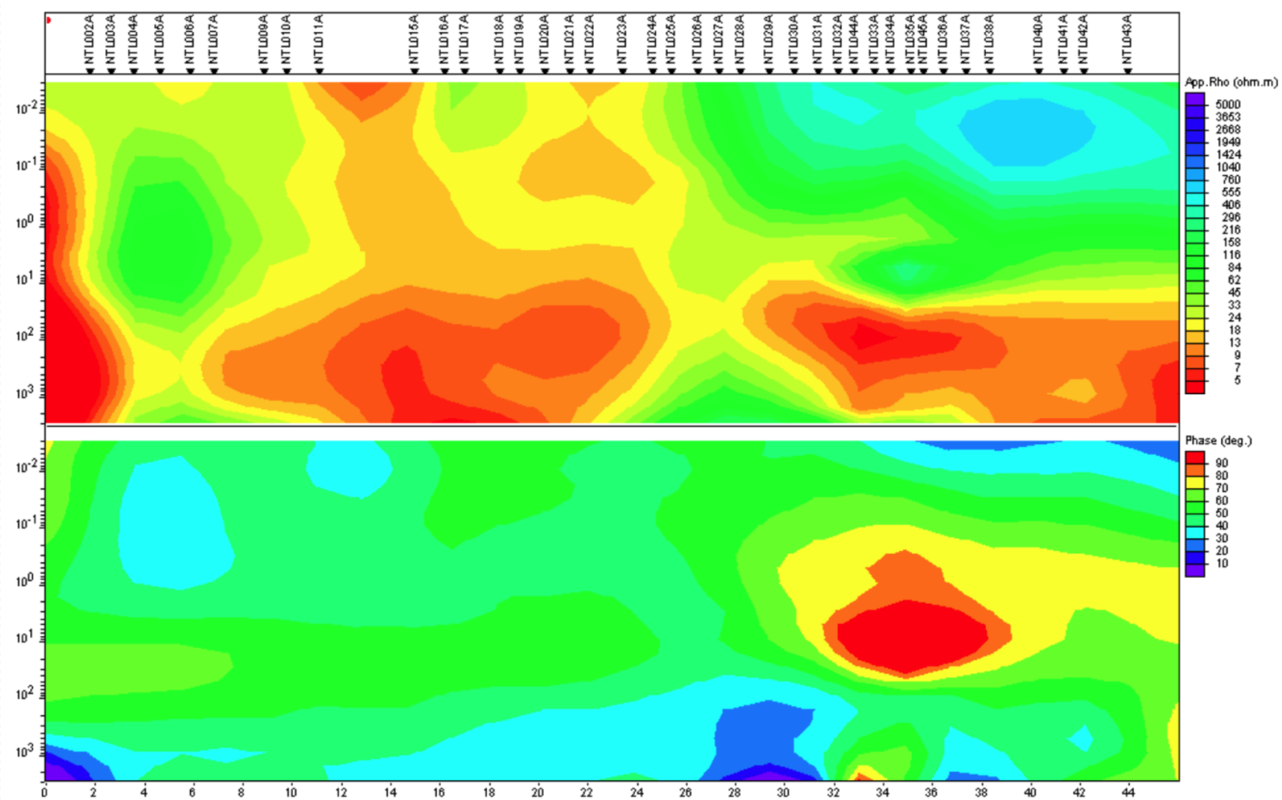
Pseudosection display of MT data

The **pseudosection is a contour plot with distance on the horizontal scale and** frequency on the vertical scale.

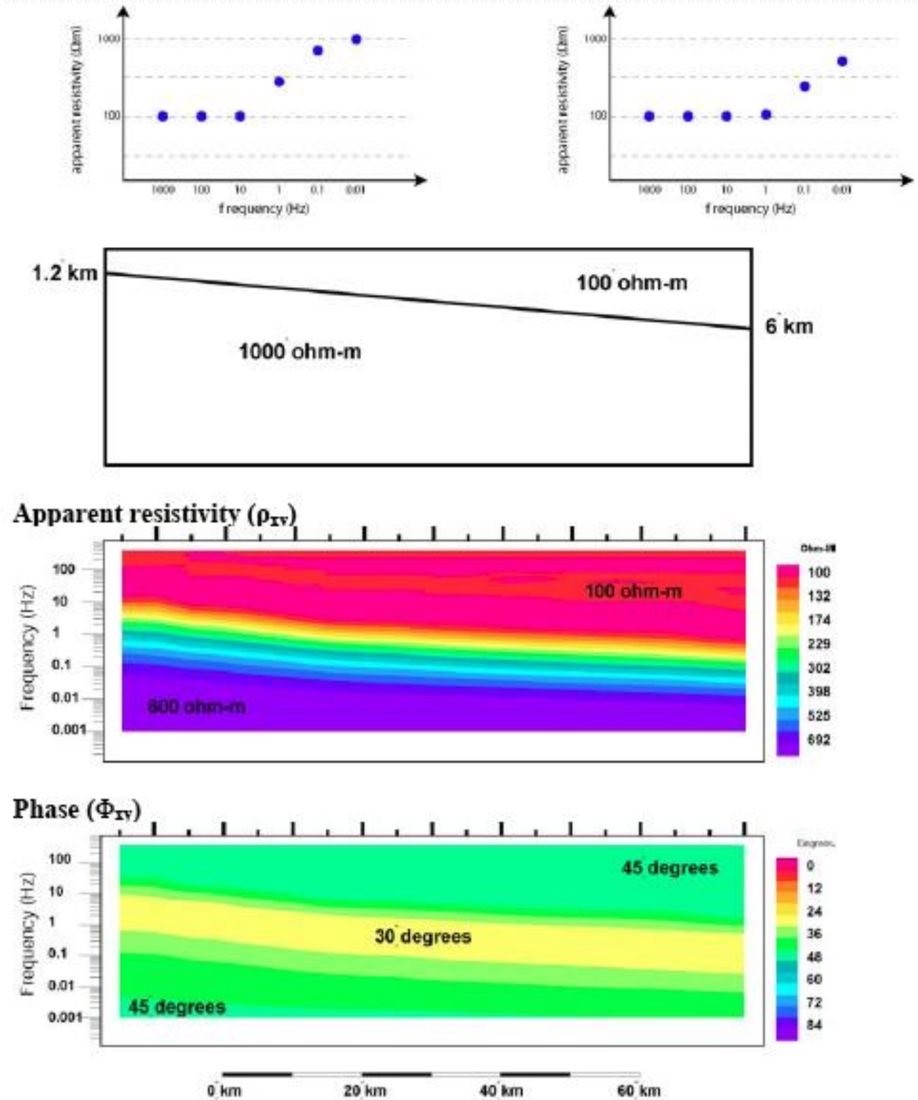
It allows apparent resistivity and phase at many MT stations to be shown in a single figure.



Plotting frequency on the vertical axis gives an impression of how the resistivity varies with depth (lower frequency is equivalent to greater depth).

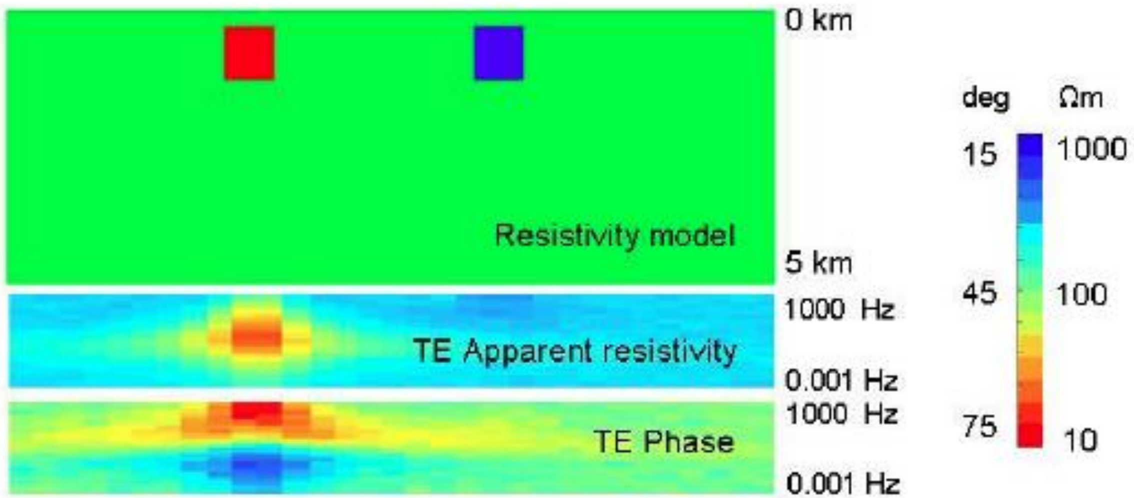


As the layer thickens to the right, the frequency needed to pass through the 100 Ωm layer decreases.



TE-mode

x-direction is along strike (into paper) and y-direction is along MT profile.



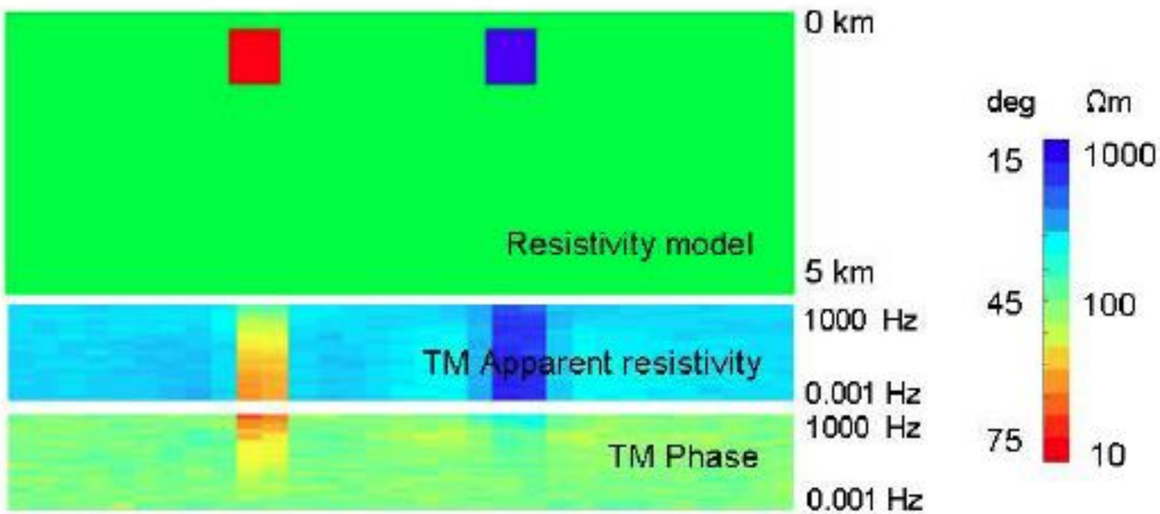
•In the TE mode, electric currents never cross boundaries between regions of differing resistivity. Thus since E_x is spatially continuous with respect to y, ρ_{xy} is also spatially continuous.

•Electric currents are induced in the conductive prism, but not in the resistive prism. Physics is dominated by **inductive effects that will not be observed at** zero frequency. Resistive prism (1000 Ωm) hardly changes the apparent resistivity. It will not be easily detected by the TE mode.

•The conductive prism (10 Ωm) produces a low apparent resistivity, with the maximum response around 1 Hz. Response of conductive prism **disappears at low frequency (0.001 Hz) because induction** is sensitive to rate of change of magnetic fields. Inductive effects disappear at low frequency.

TM mode

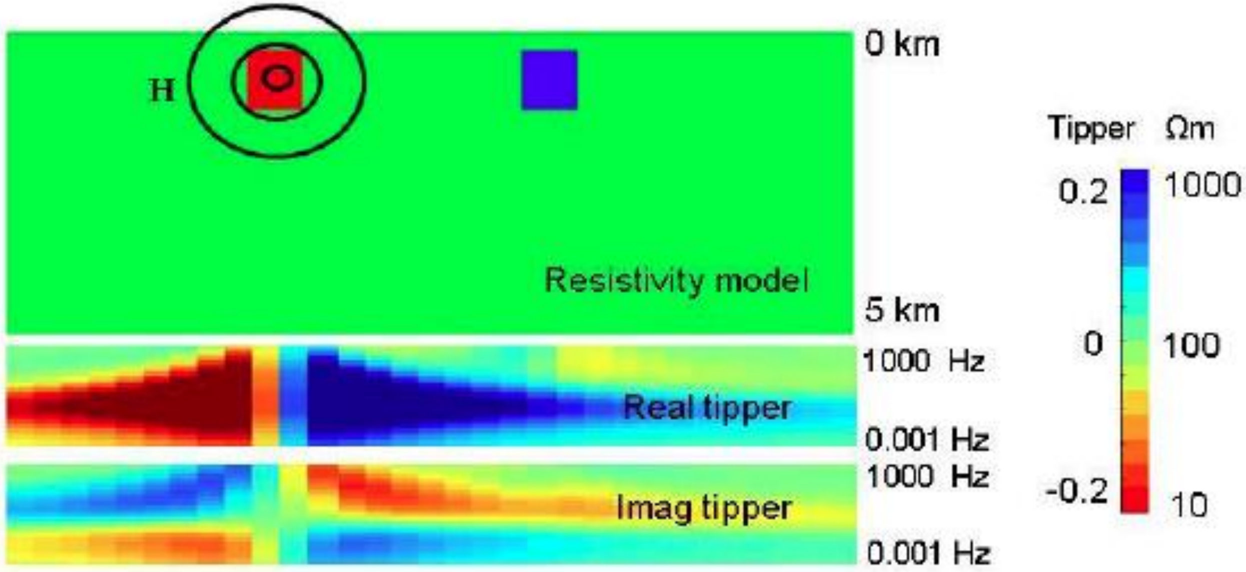
x-direction is along strike (into paper) and y-direction is along MT profile.



•In the TM mode, electric currents **cross the boundaries between regions of differing**
•resistivity. This causes electric charges develop on the boundaries.
•Thus physics of this mode includes both **inductive and galvanic effects. Galvanic** effects, such as charge build up on boundaries, will be observed at all frequencies (including direct current). In contrast, inductive effects decrease at low frequencies and are absent at zero frequency.
•Since E_y is **spatially discontinuous with respect to y, the apparent resistivity (ρ_{yx}) can** also spatially discontinuous at resistivity boundaries.
•Both prisms have an effect on the apparent resistivity pseudosection.
•Effect on apparent resistivity does not disappear at low frequency (galvanic effect).

x-direction is along strike (into paper) and y-direction is along MT profile.

Tipper (vertical magnetic field transfer function)



•Note **reversal in sign above the conductive prism. This is equivalent to a reversal in the** induction vectors.

•Almost no change in tipper is observed above the resistive prism. It will not be detected by tipper data.

MT data are interpreted by the process of **modelling**.
A Model is the representation of the resistivity distribution by a discrete volume element grid or mesh.

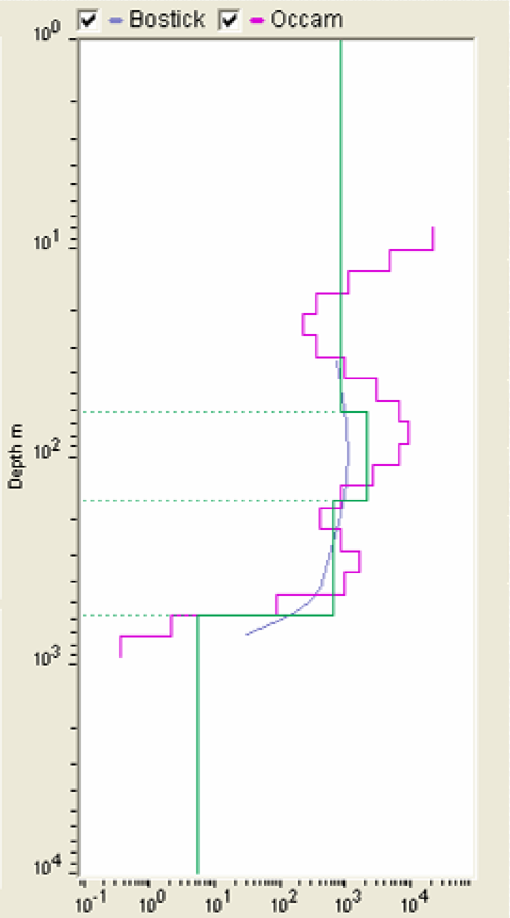
Forward:
A model is chosen to represent the subsurface resistivity distribution and by numerical simulation, the response of the model is computed and synthetica data produced.

Inversion:
1) A model is chosen to represent the subsurface resistivity distribution (**a priori model**).
2) By numerical simulation, the response of the model is computed and synthetica data produced. These are compared to the real data. The parameters of the model are then systematically varied until the numerical ‘data’ match the observed data.
3) The”best” model is obtained and an error (difference between observed and computed data) is obtained

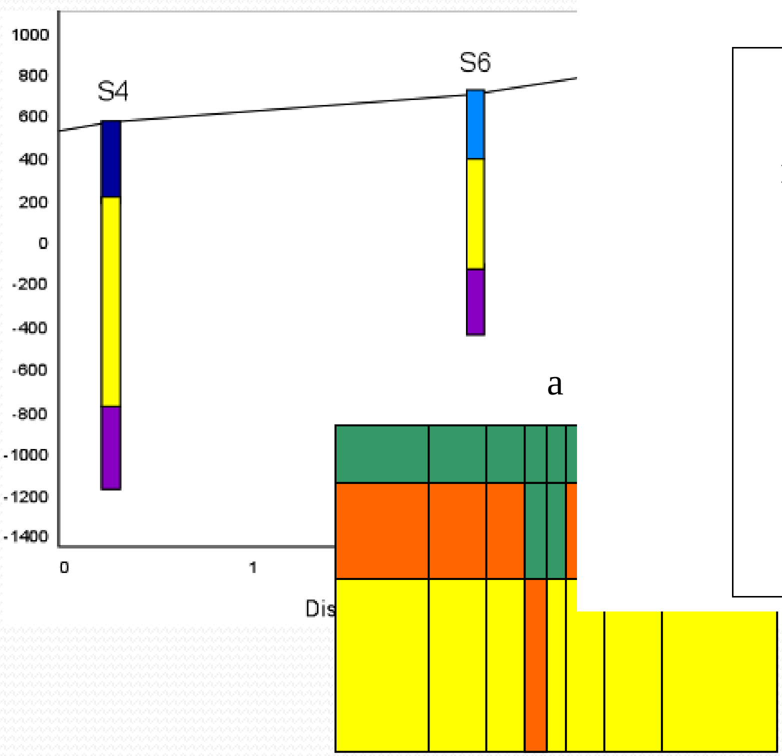
1D modelling

The most common are

- Bostick inversion (the resistivity is a continuous function of depth),
- Layered inversion
- Occam inversion

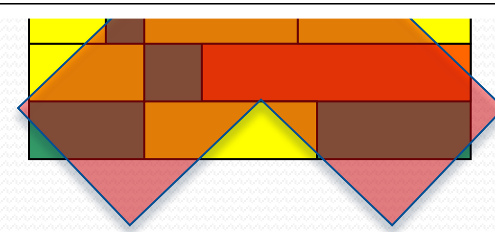


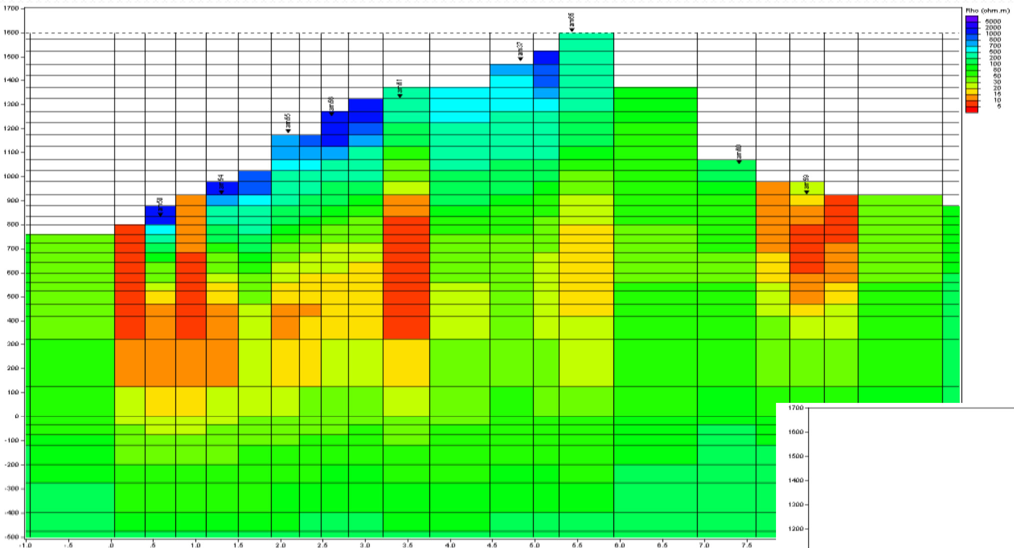
Model files contain the dimensions of prismatic (3D) rectangular (2D) or linear (1D, thickness) elements composing the model mesh.

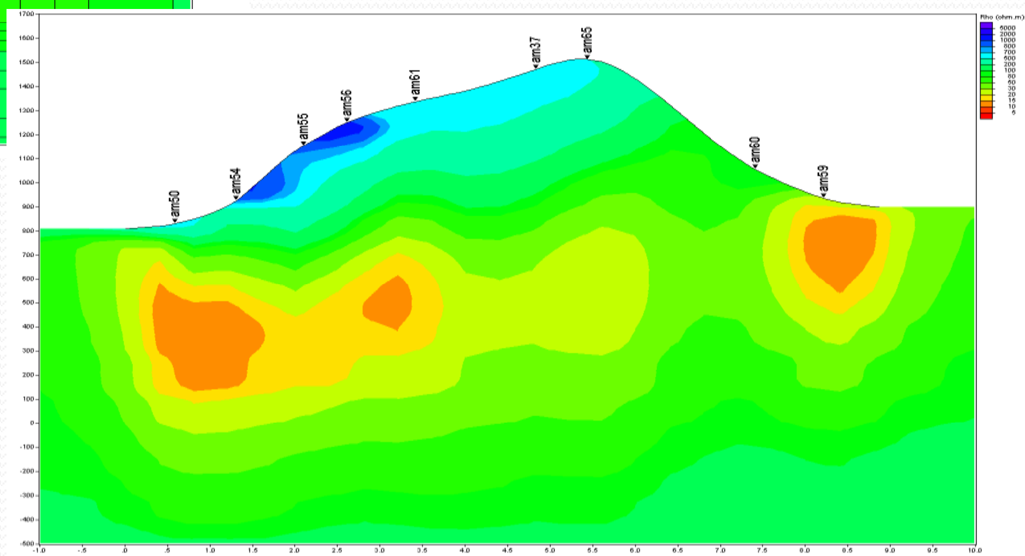


Model file ('north input.example1')

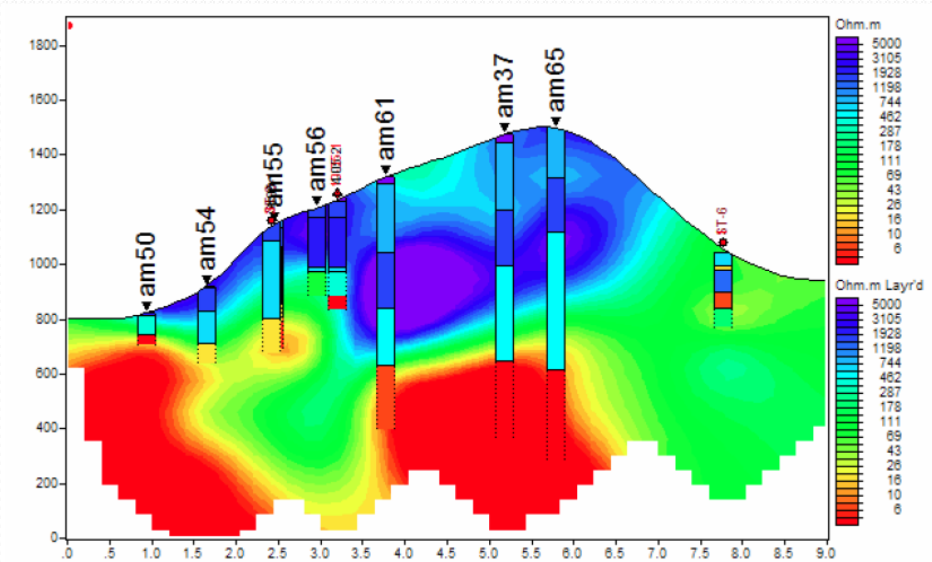
```
10 5
10000 5000 2000 2000 1000 1000 2000 2000 5000
10000
30 100 200 300 500 1000
1 1 1 1 1 1 1 1 1 1
1 1 2 3 3 3 3 2 1 1
2 2 3 3 3 3 3 3 2 2
1 1 2 3 3 3 3 2 1 1
2 2 2 2 2 2 2 2 2 2
1000 100 5
```







The final model may be imaged as a mesh (left) or a smooth distribution obtained by interpolation and gridding (right)


am50
am54
am55
am56
am61
am37
am65
ST-6
Ohm.m
5000
3105
1928
1198
744
462
287
178
111
69
43
26
16
10
6
Ohm.m Layr'd
5000
3105
1928
1198
744
462
287
178
111
69
43
26
16
10
6
1800
1600
1400
1200
1000
800
600
400
200
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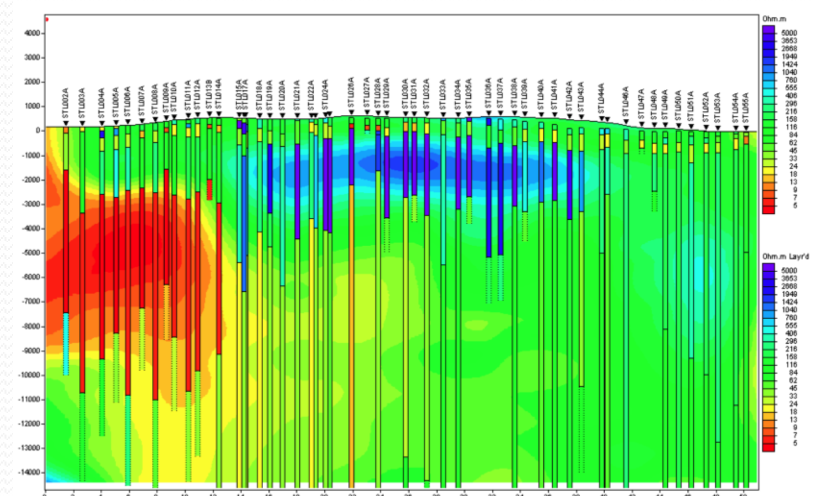
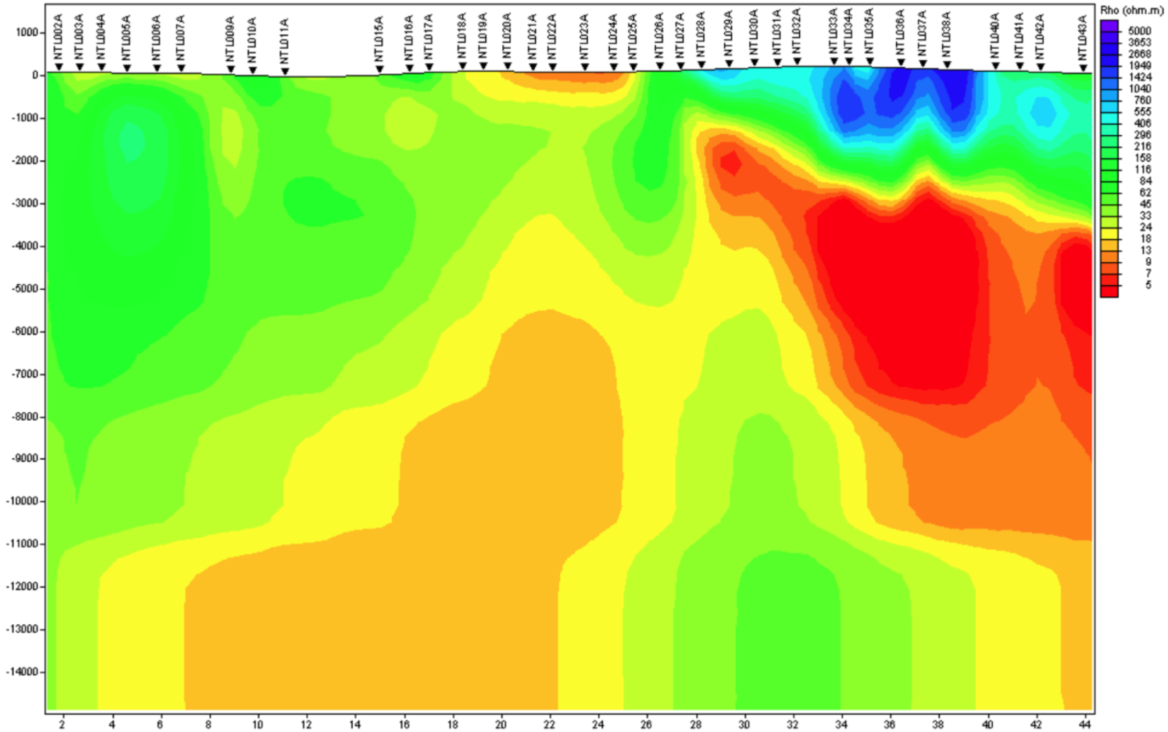
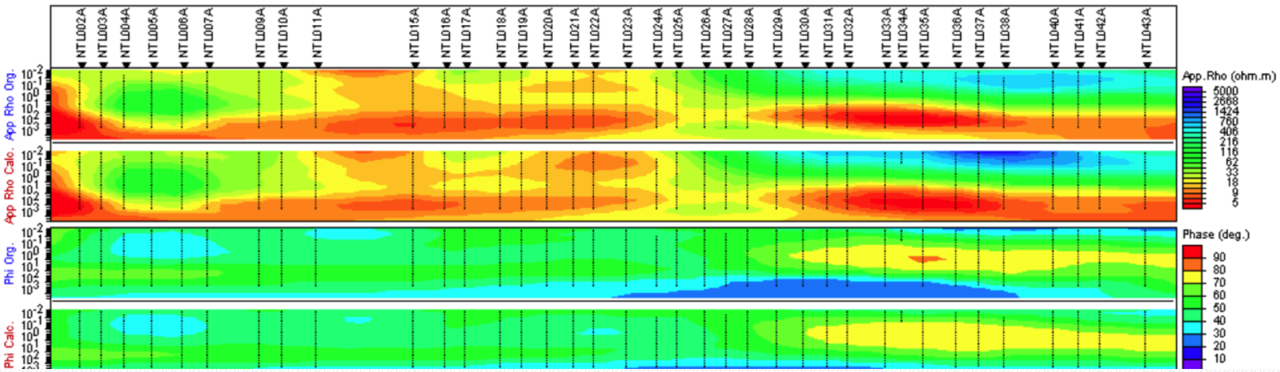
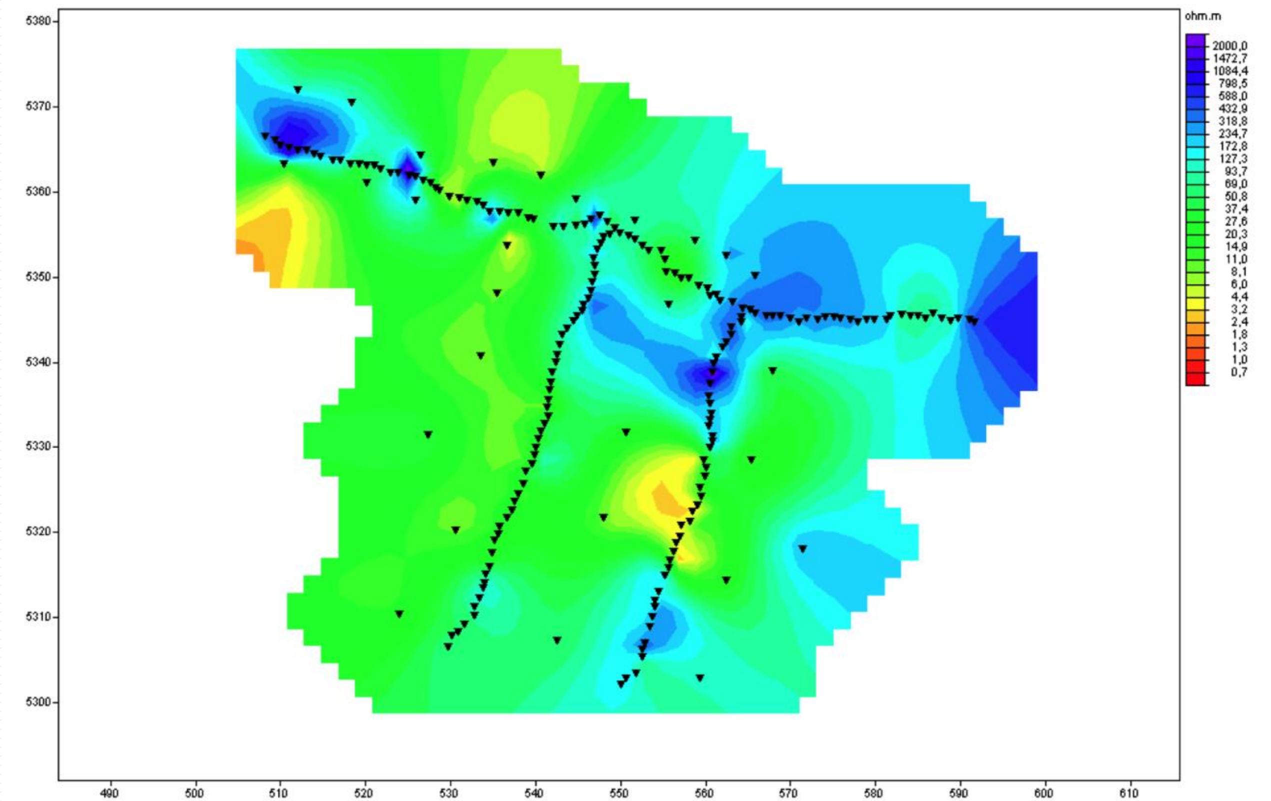

Ohm.m
Ohm.m Layr'd
0 2 4 6 8 10 12 14 16 18 20 22 24 26 28 30 32 34 36 38 40 42 44 46 48 50





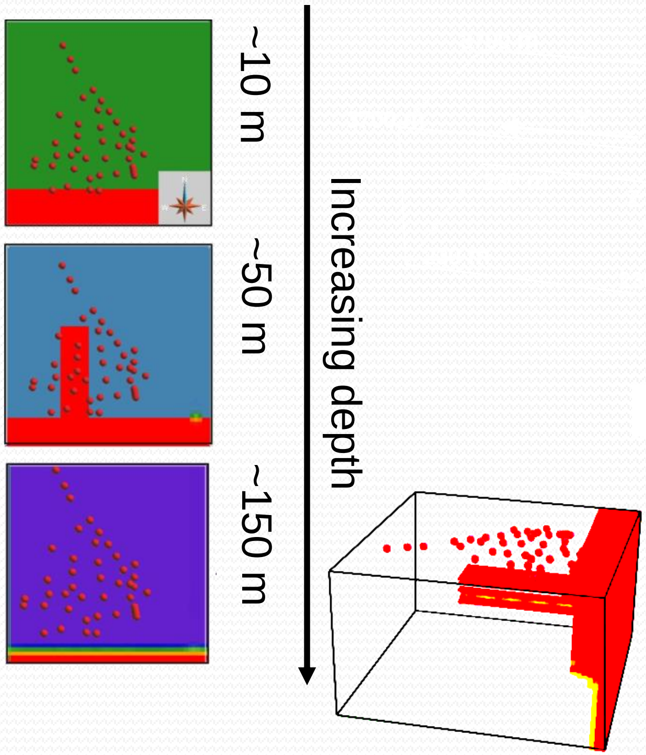
To check model quality we compare real and snthetic data (pseudosectios)

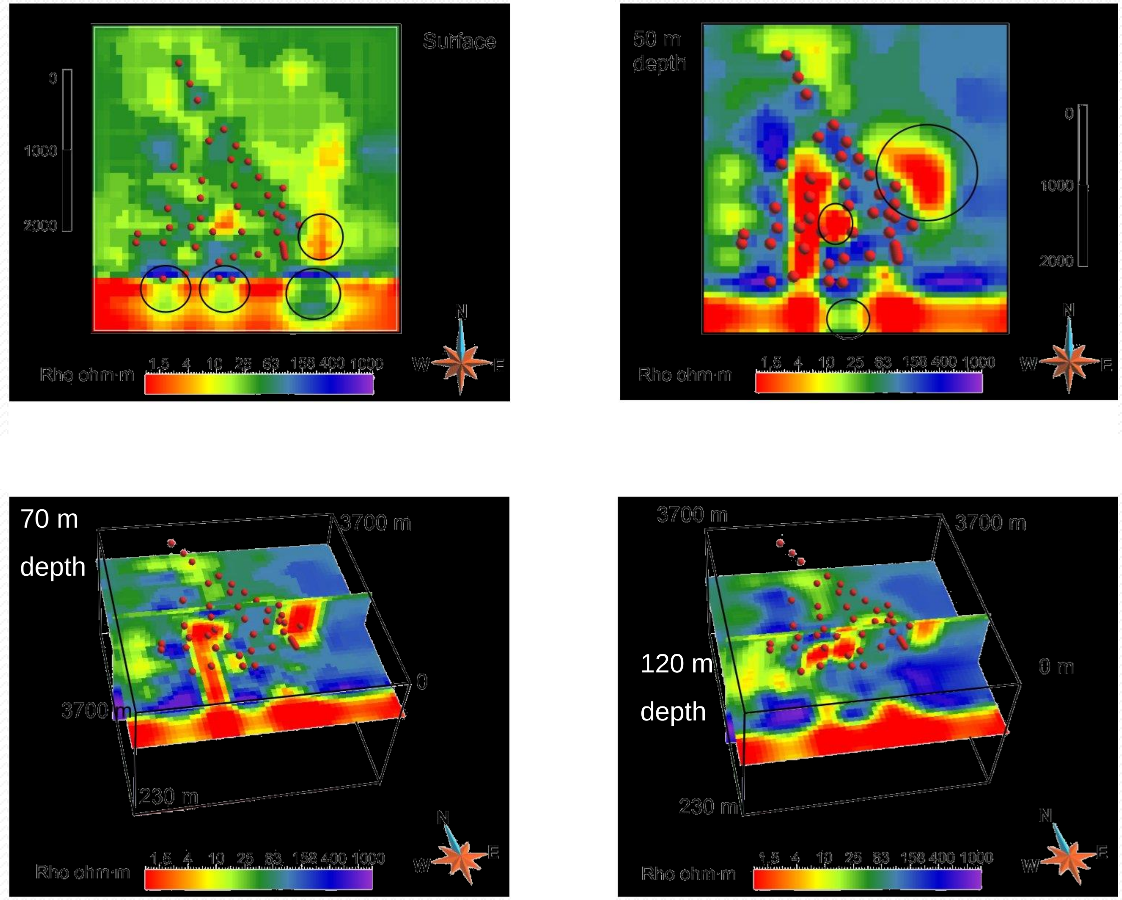


We can image resistivity distribution at depth by interpolating resistivity values obatined in 2D or 1D models section at the same depth.

Plan slice of interpolated resistivity values from 2D models obtained by joint inversion of TE and TM data (-1 km a.s.l.).

Synthetic model



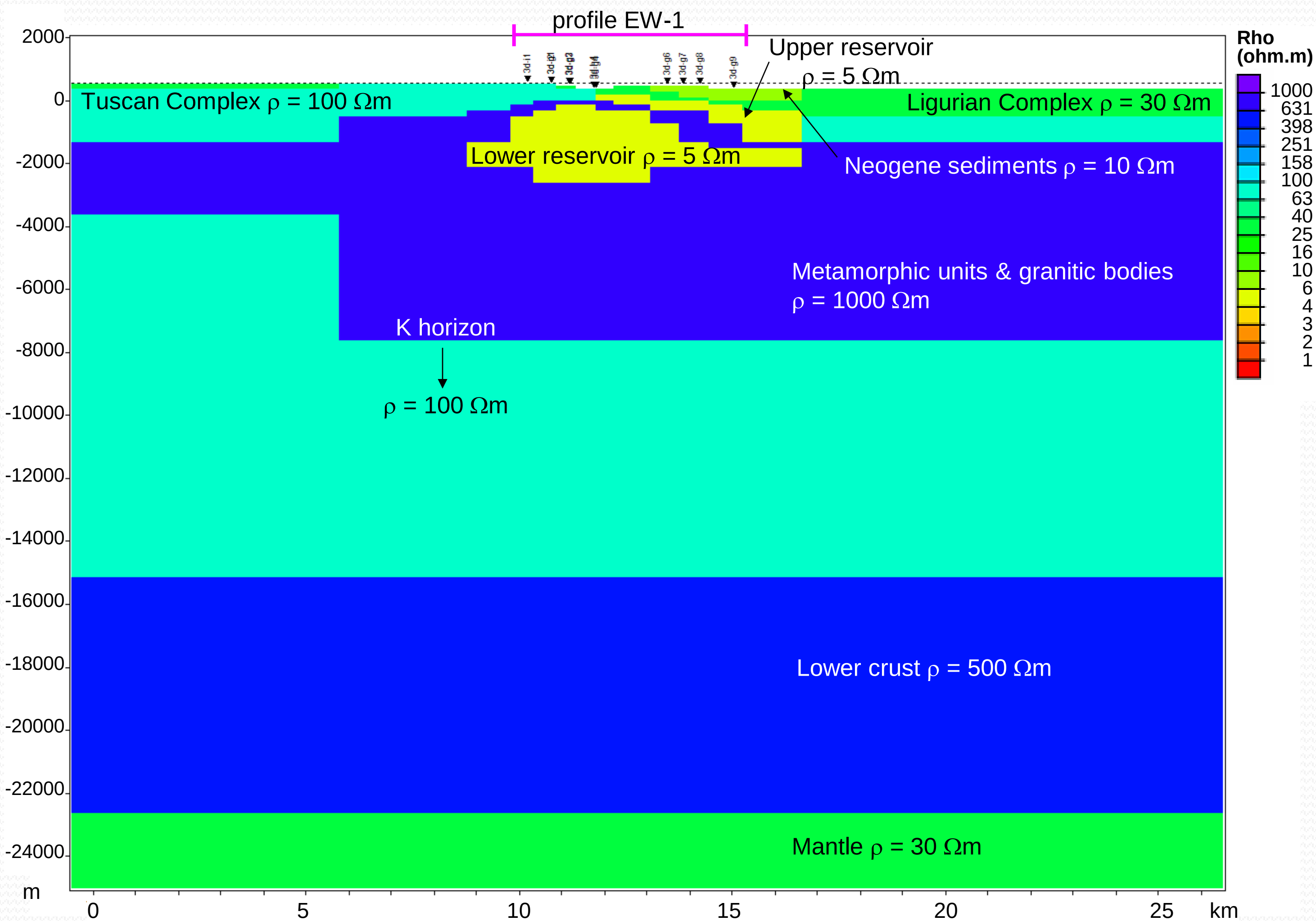


Forward modelling

(Mackie and Madden, 1993)

Inverse modelling WSINV3D

(Siripunvaraporn et al. 2005)



profile EW-1
Upper reservoir
ρ = 5 Ωm
Rho
(ohm.m)
1000
631
398
251
158
100
63
40
25
16
10
6
4
3
2
1
2000
0
-2000
-4000
-6000
-8000
-10000
-12000
-14000
-16000
-18000
-20000
-22000
-24000
m
Tuscan Complex ρ = 100 Ωm
Ligurian Complex ρ = 30 Ωm
Lower reservoir ρ = 5 Ωm
Neogene sediments ρ = 10 Ωm
Metamorphic units & granitic bodies
ρ = 1000 Ωm
K horizon
ρ = 100 Ωm
Lower crust ρ = 500 Ωm
Mantle ρ = 30 Ωm
0
5
10
15
20
25
km

In **Controlled Source Audio-frequency MagnetoTellurics (CSAMT)**, the most common artificial source frequency-domain electromagnetic sounding, a scalar impedance and its phase are computed.

An apparent resistivity can then be plotted against frequency, and data are treated as in MT.

The top 2 to 3 km of the earth's crust can then be mapped.

This method proved useful in detecting the shallower part of geothermal systems, like reservoir resistivity contrasts, but its limited depth of investigation is not suitable to define the deep heat sources.

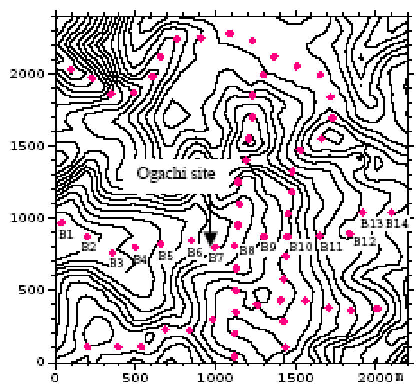


Fig.4 Location map of CSAMT survey points

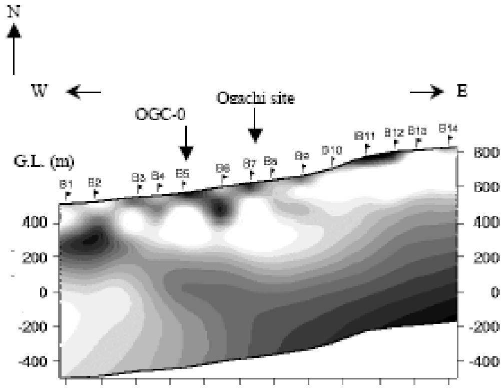


Fig.5 Resistivity profile along B-line in CSAMT method.

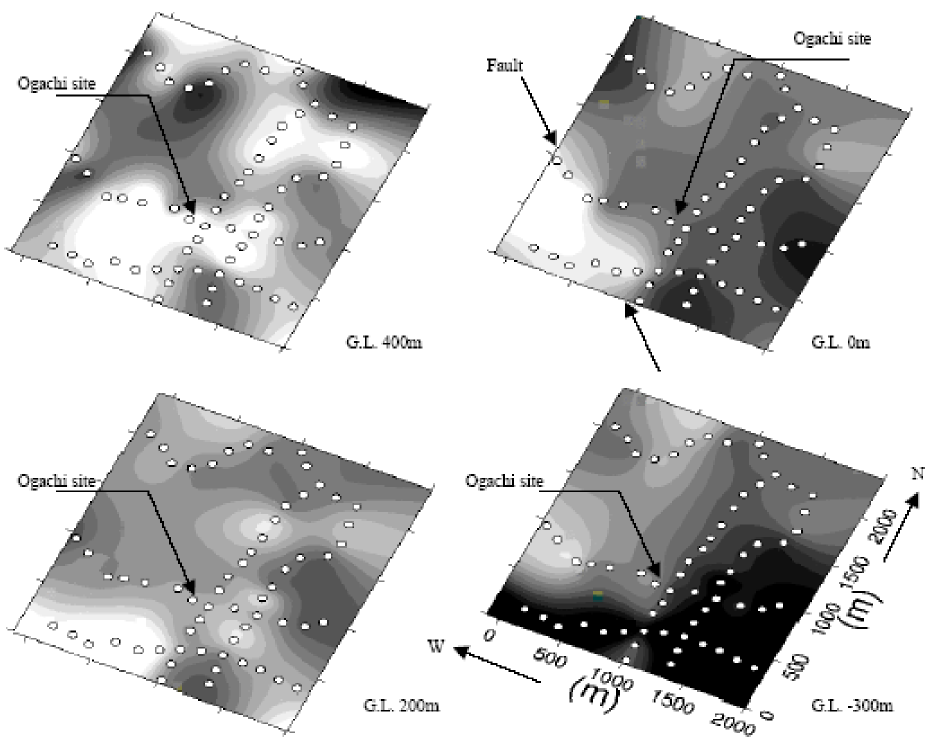
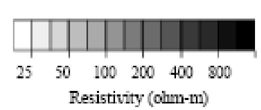
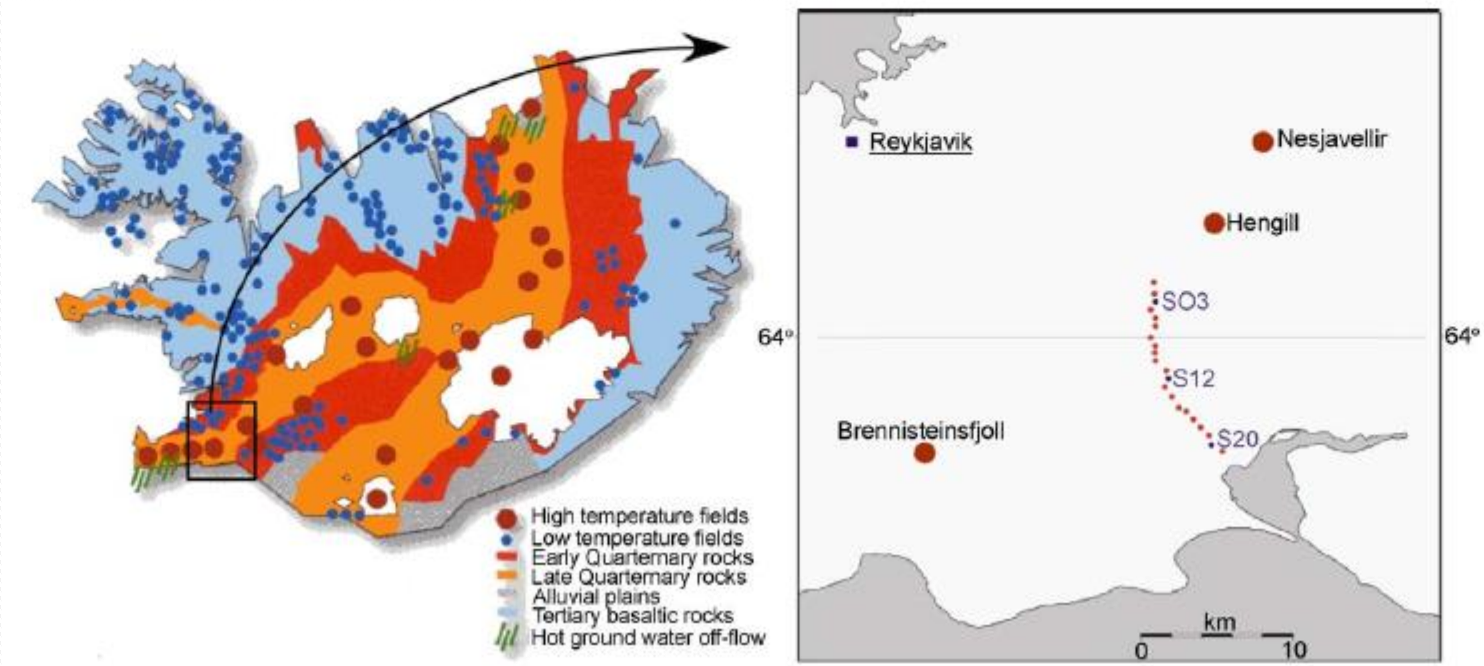


Fig.6 Resistivity horizontal sections by CSAMT method.



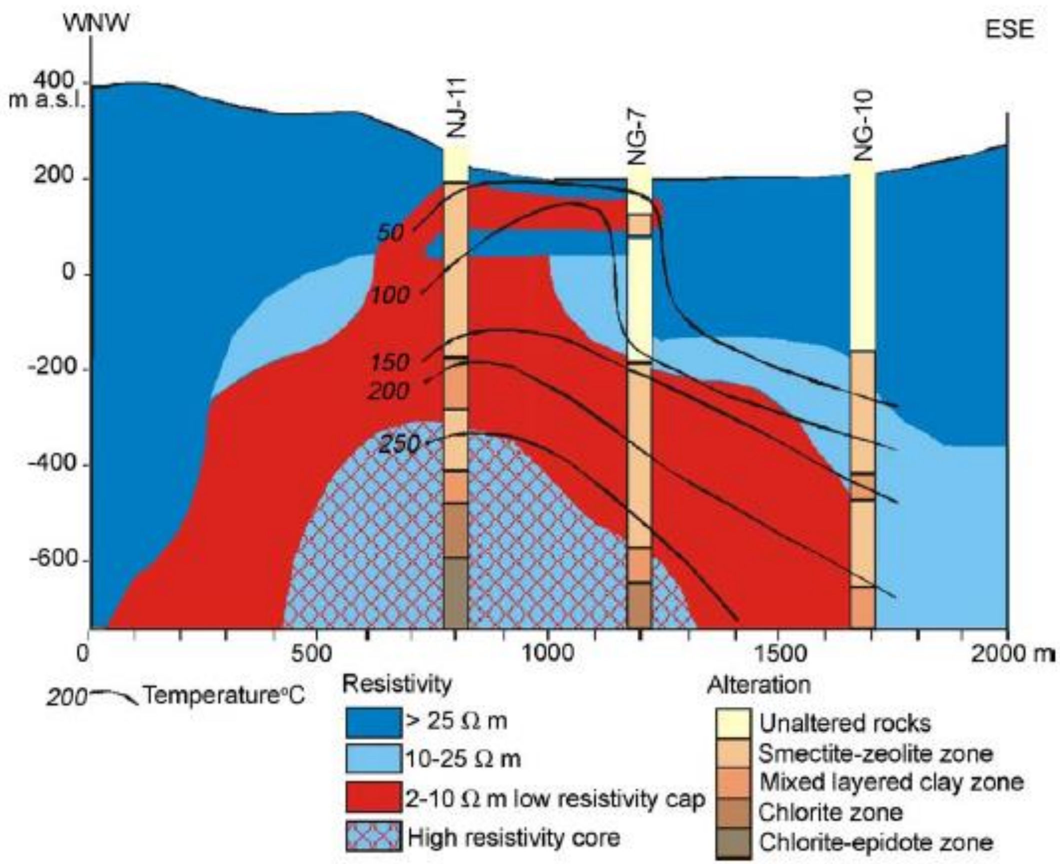
An example of CSAMT data acquired at Ogachi HDR test site, Japan (from Suzuki et al, WGC2000)

Case Study: Iceland



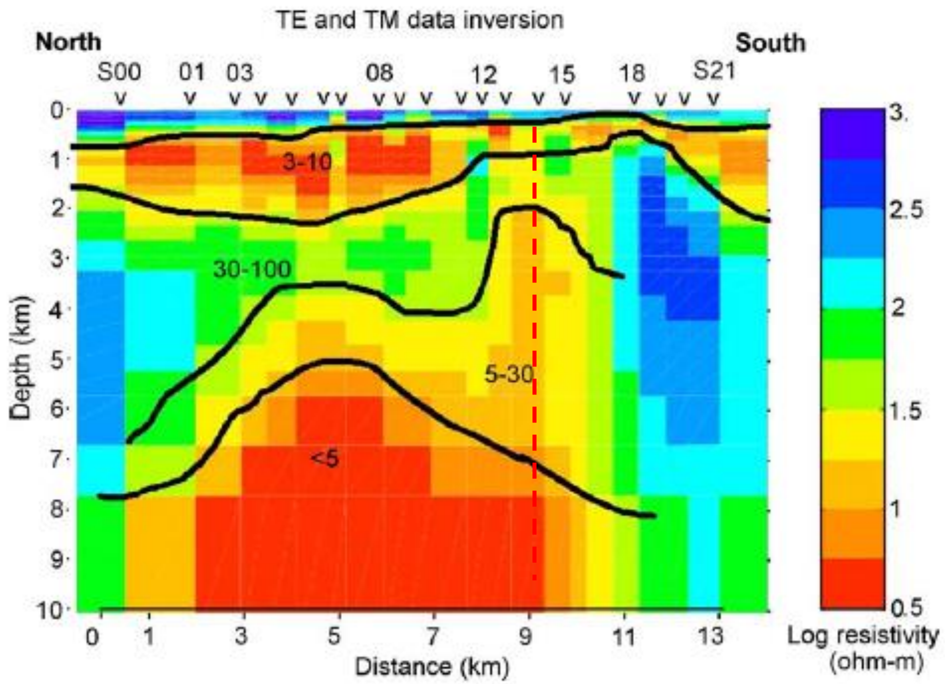
The main objective of the project ‘Deep Geothermal Prospecting (DGP) in Iceland’ organized by Orkustofnun, the National Energy Authority of Iceland, was to identify the best combination of exploration methods for mapping deep geothermal resources outside the known geothermal systems.

Case Study: Iceland



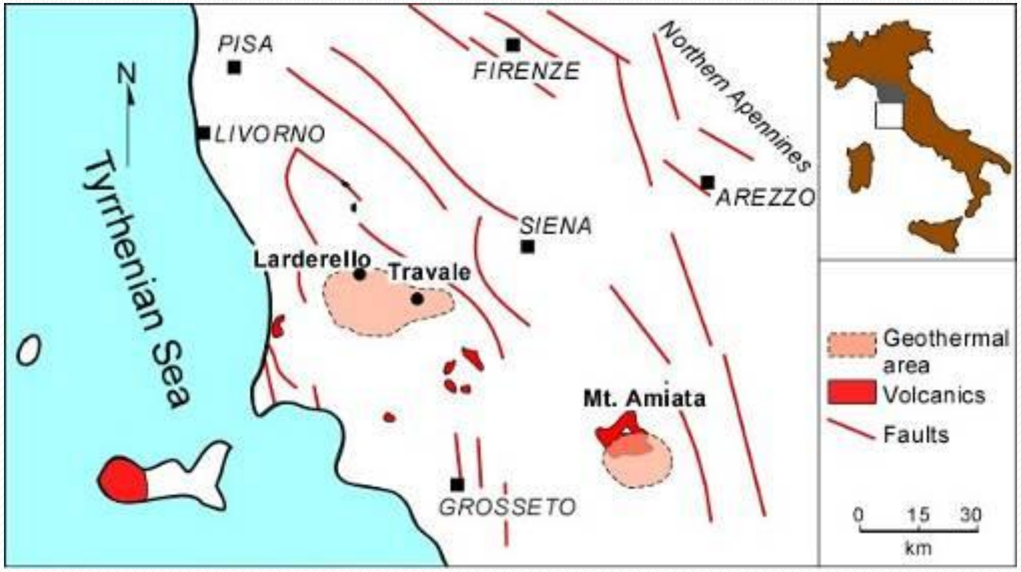
The main purpose was to define the possible deep "geothermal print" between two main geothermal fields.

Case Study: Iceland



An alteration zone is present beneath the surface, where there are no obvious geothermal manifestations; this suggests that a hydrothermal fluid circulation exists or have existed at depth.
The geothermal reservoir might be connected through a fault to a deep conductor representing the heat source of the System.

Case Study: Italy

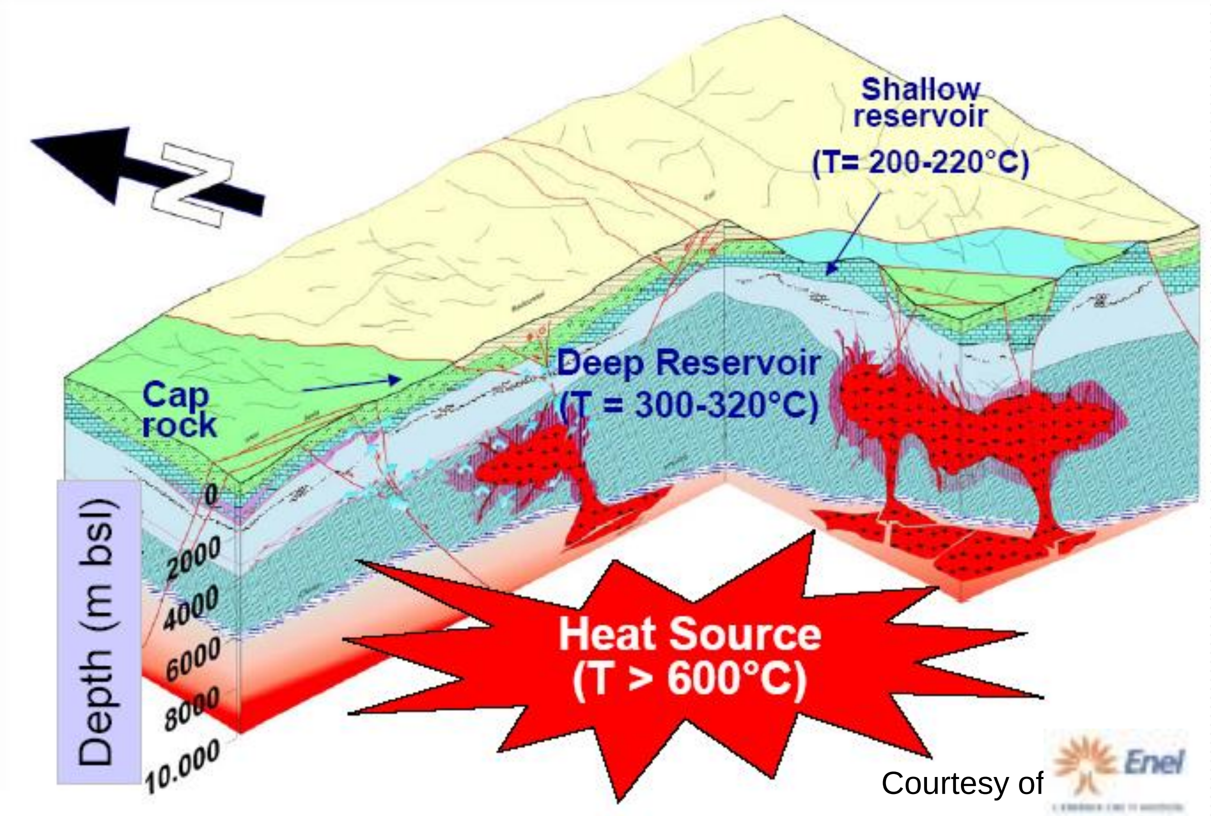


Tuscany is characterised by NNW-SSE faults linked to the Alpine orogeny.

The main extensional tectonic was accompanied by anatectic magmatism whose age decreases in an eastward direction.

Shallow intrusive bodies are the heat sources of the **two most important geothermal areas, Larderello/Travale and Mt. Amiata.**

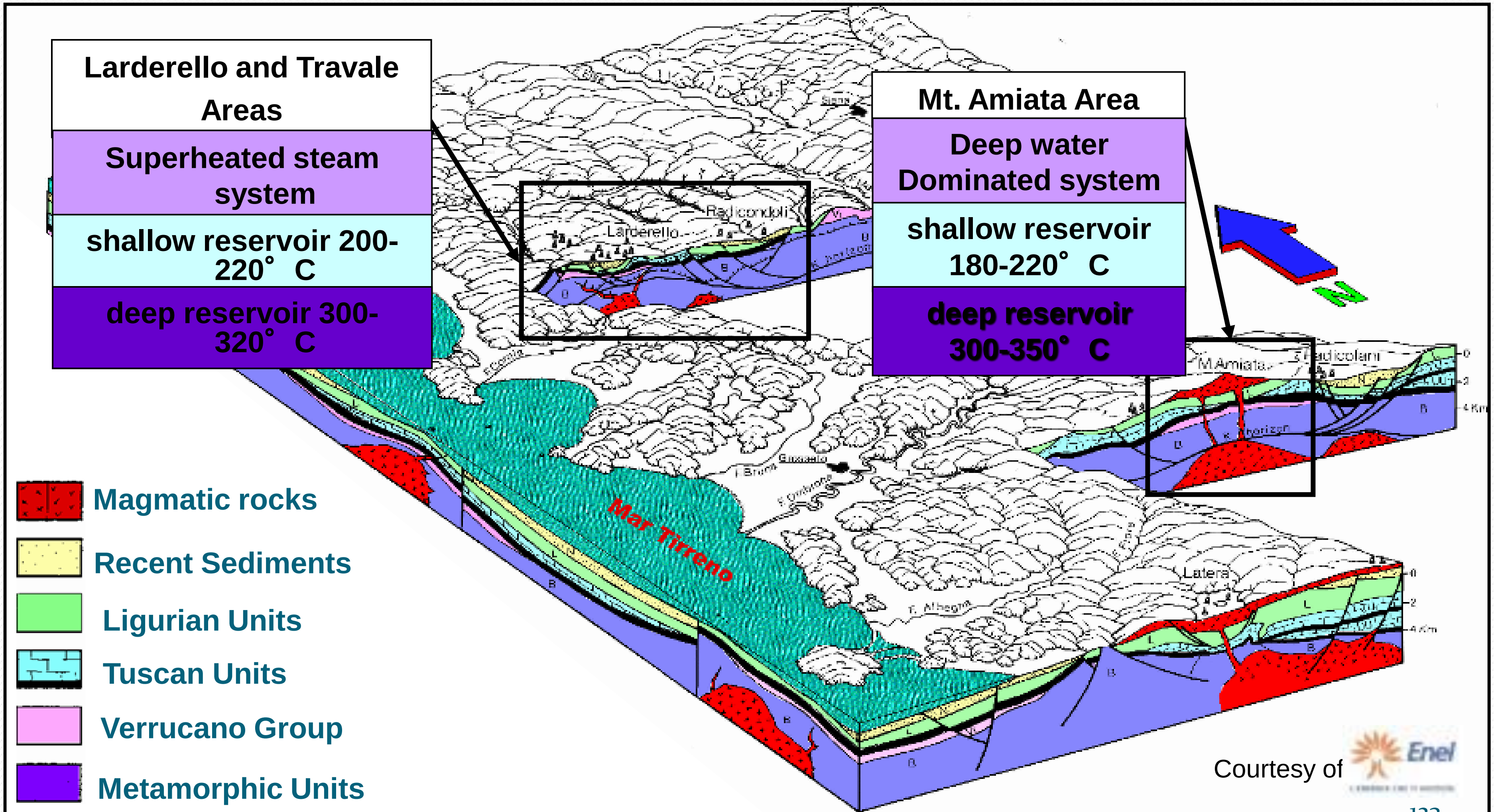
Case Study: Italy



Very high temperatures characterize the two geothermal reservoirs.

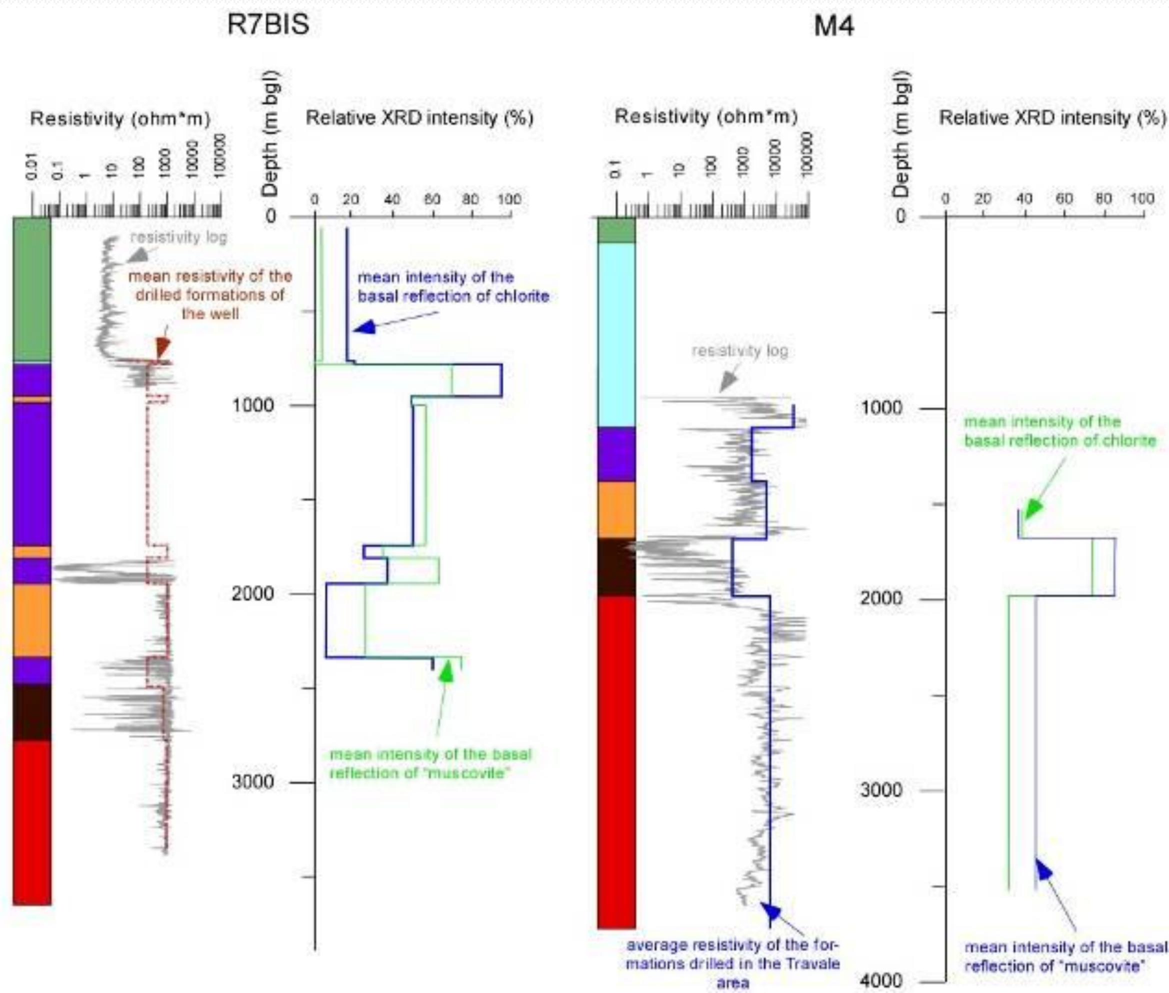
The geothermal exploration targets are mainly located in metamorphic and granitic rocks down to 4700 m depth

Case Study: Italy

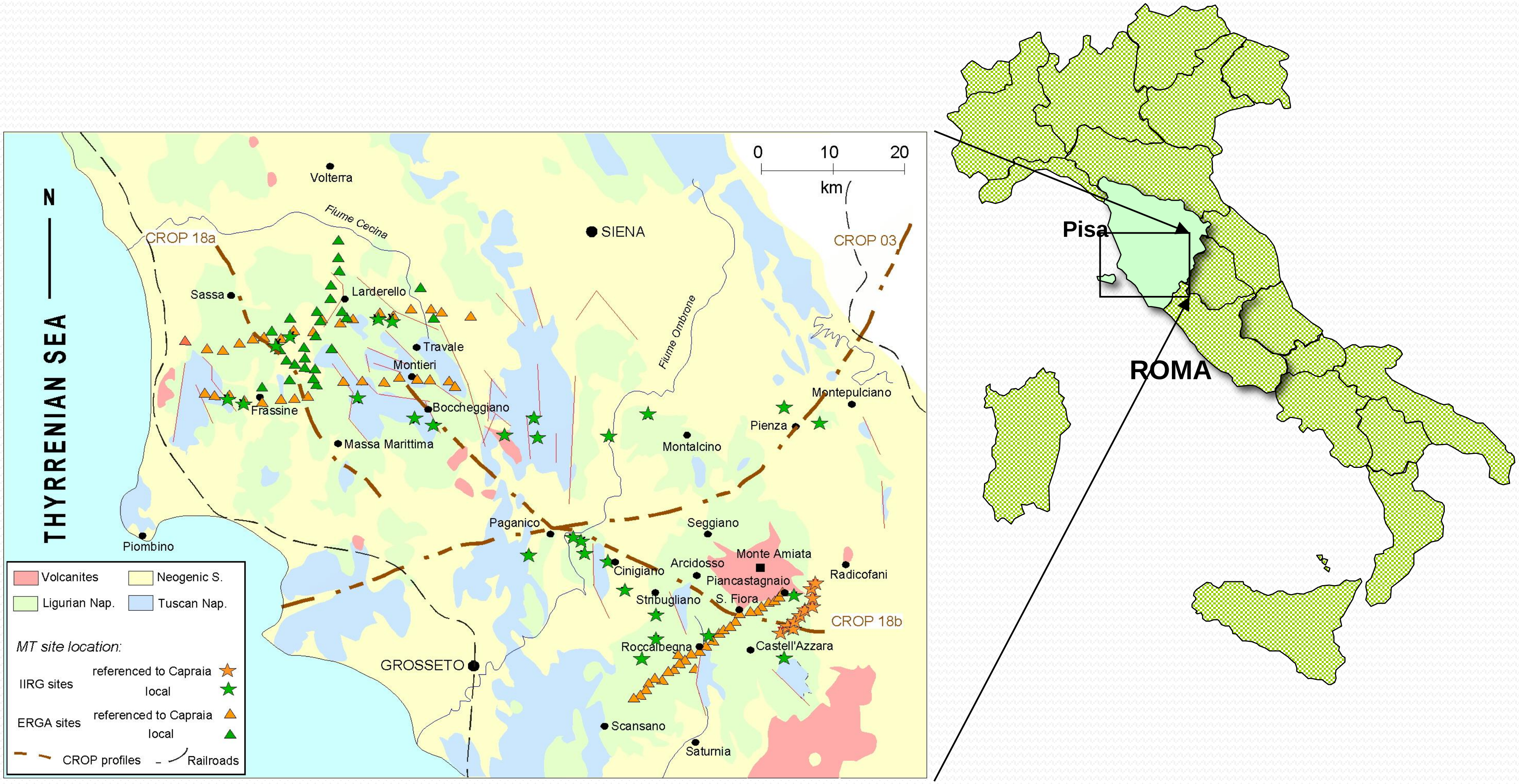


Case Study: Italy

In the Larderello field rock matrix should not provide strong variation since resistivity changes little passing from metamorphic to granitic rocks. Moreover, the exploited geothermal fluid is superheated steam, which by itself should not contribute to a resistivity reduction. Partial melting reduces resistivity, and this effect is very probable in the medium-lower crust, where teleseismic tomography defined low velocity bodies. However, melts are not present at the depths of geothermal reservoir where low resistivity anomalies are located.



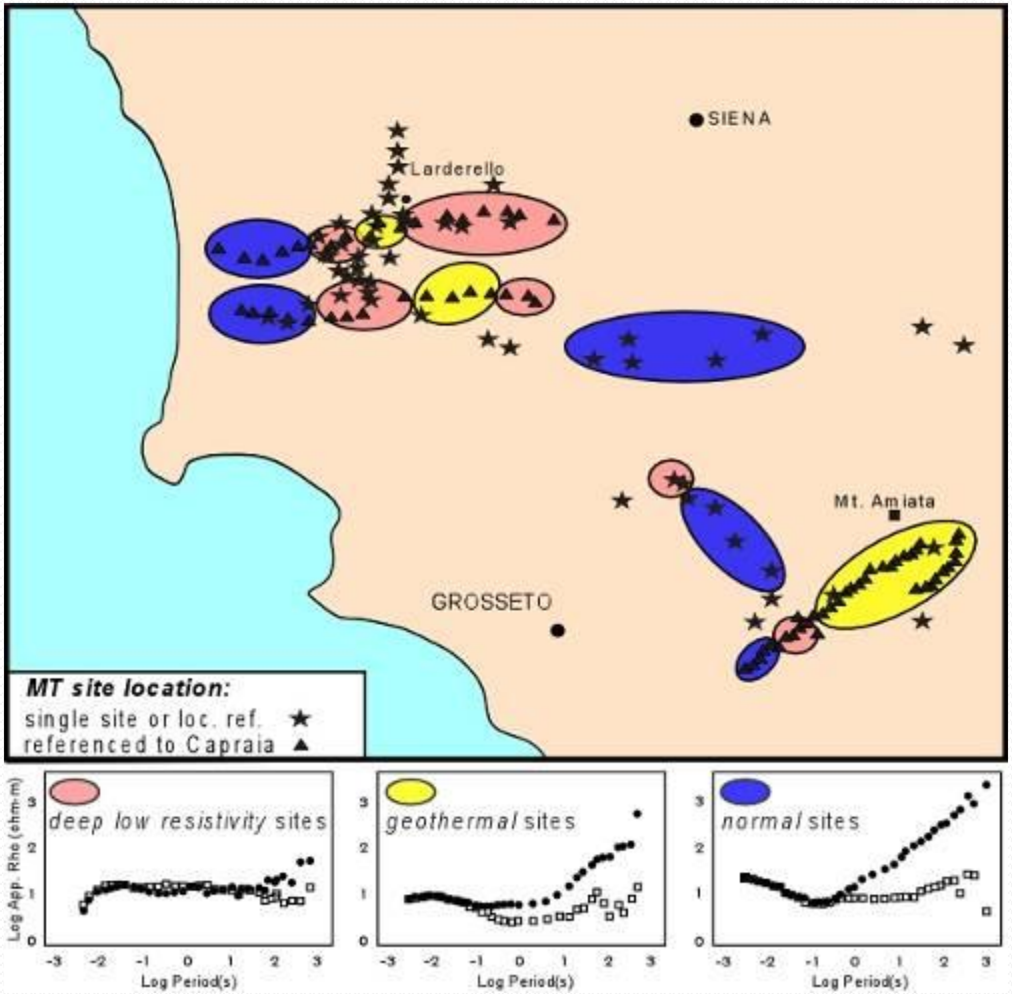
Case Study: Italy



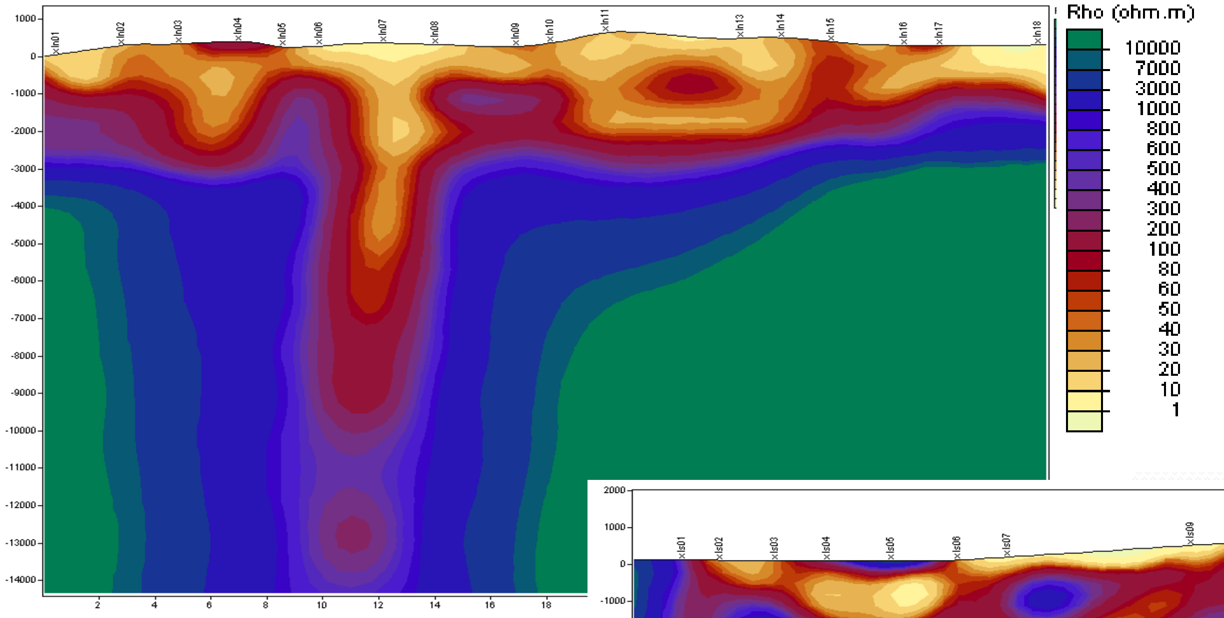
Many MT data have been collected in the last two decades, both for crustal and geothermal investigation, facing a strong EM coherent noise problem (remote reference and robust processing are a “must”)

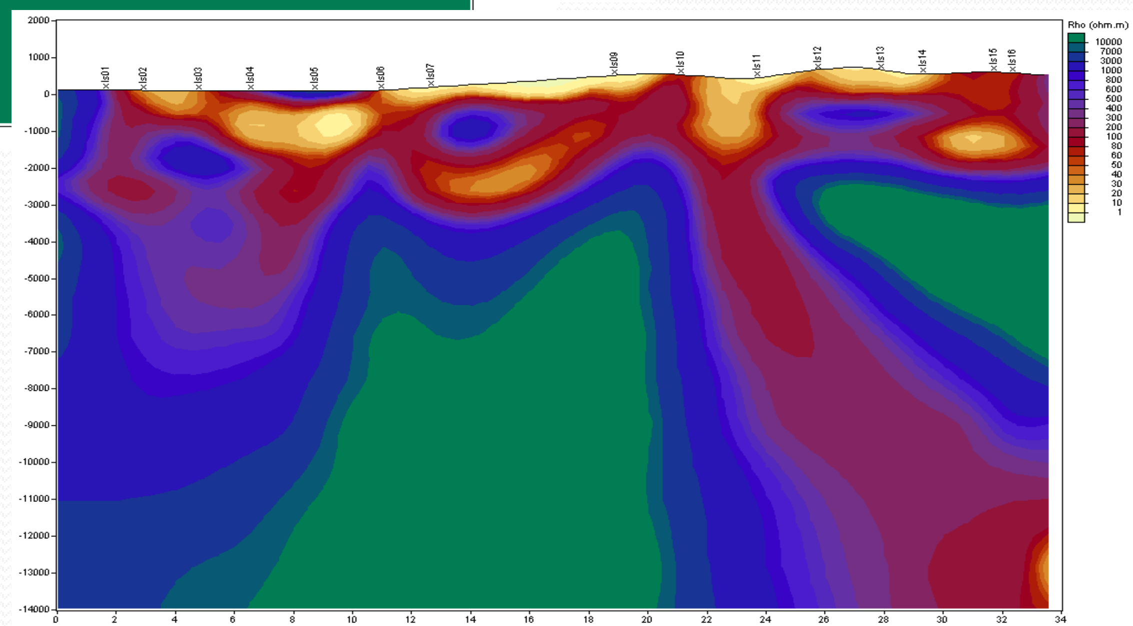
Case Study: Italy

Geothermal areas do not always correspond to the areas with the lowest resistivity



Case Study: Italy

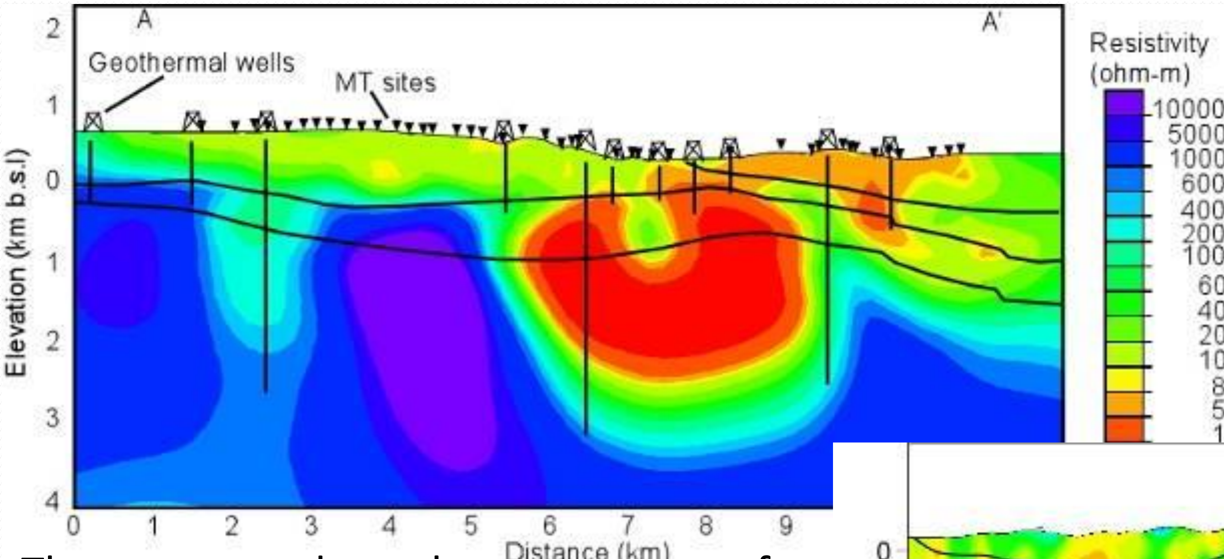




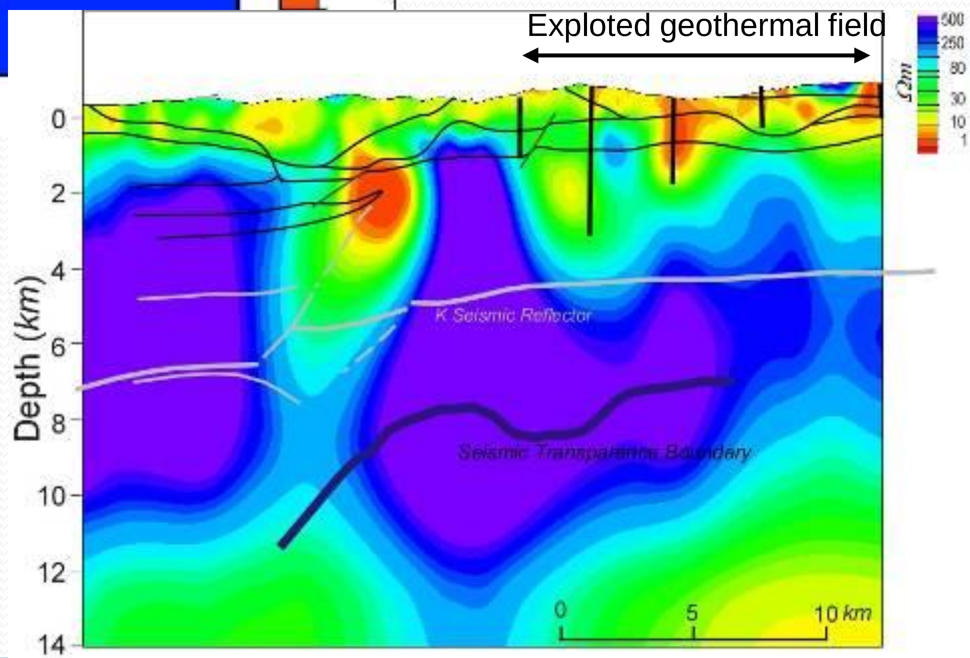
Larderello 1992

Deep conductive features were found in some areas of the field (some of the most productive but in the superheated steam reservoir)

Case Study: Italy

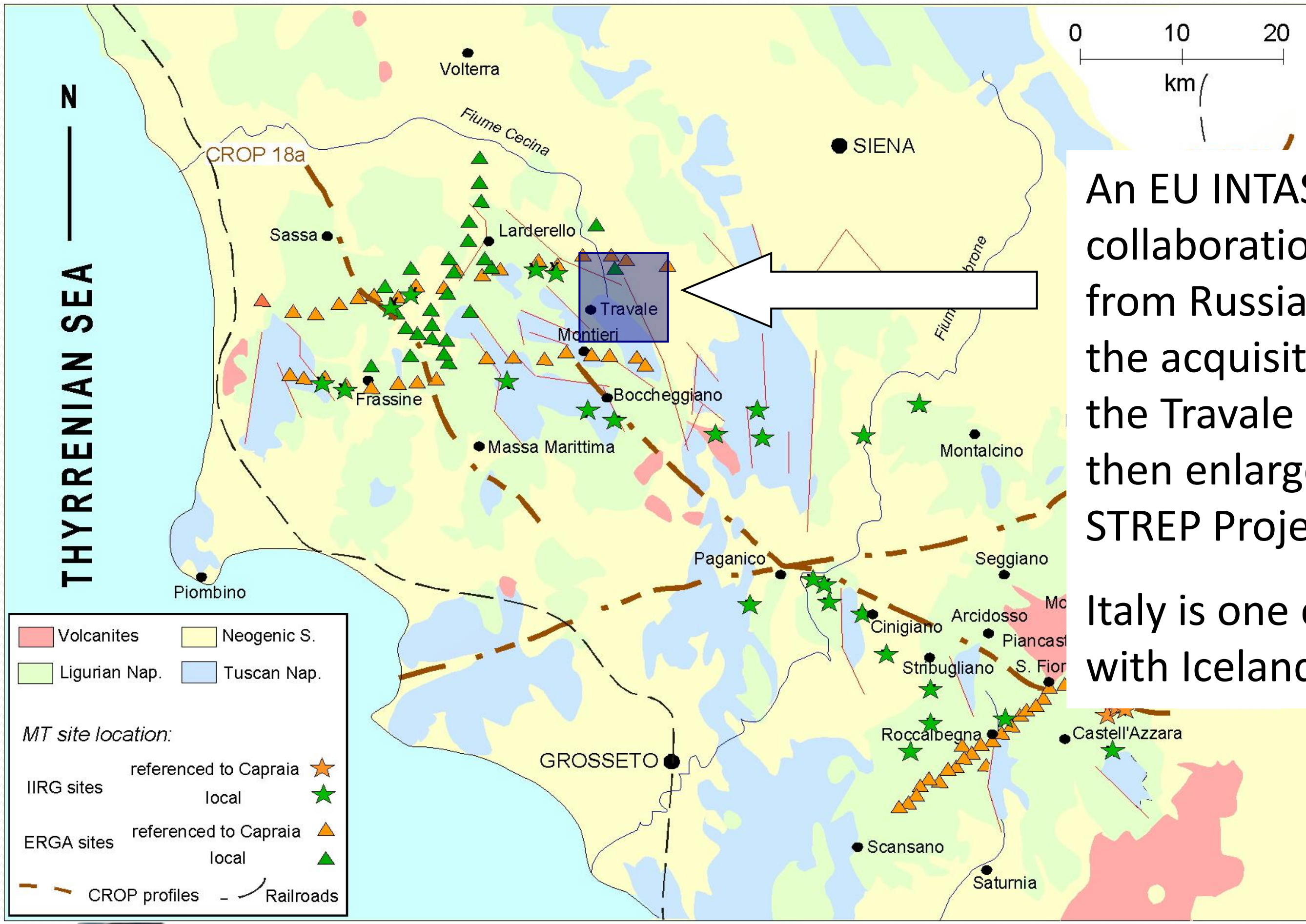


Amiata 1994 and 1998



The correspondence between areas of low resistivity inside the resistive basement and geothermal reservoirs was very evident in the Mt. Amiata water-dominated system.
However, only few sites out of the main exploited geothermal field showed deep conductive features such as in the Larderello area

Case Study: Italy



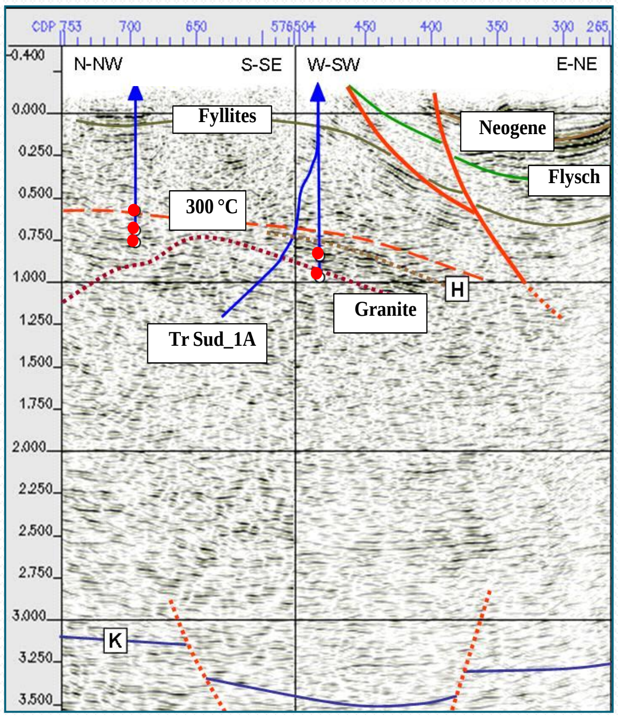
An EU INTAS project in collaboration with colleagues from Russia and Iceland, allowed the acquisition of new MT data in the Travale area. The area was then enlarged thanks to the EU STREP Project (I-GET).

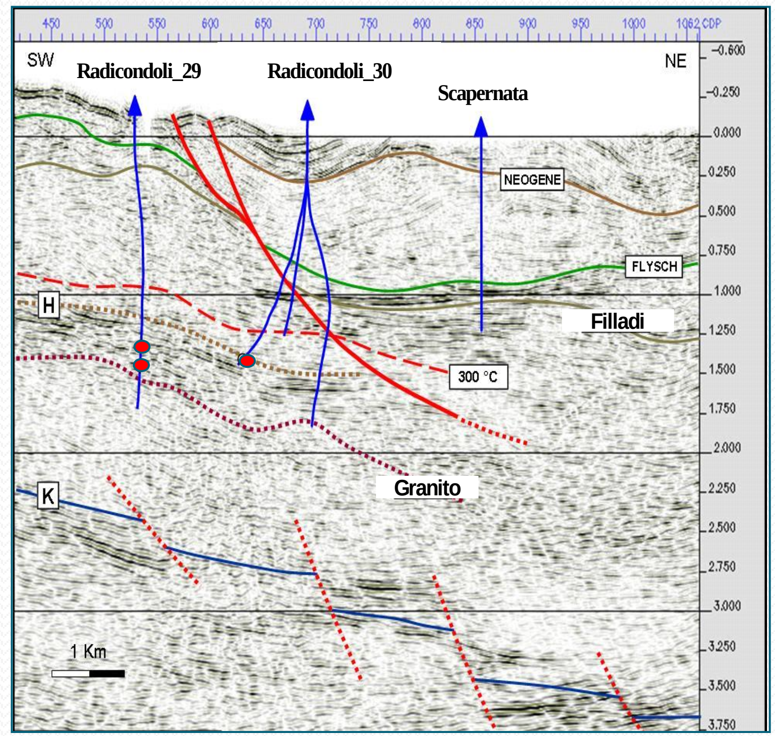
Italy is one case study, together with Iceland, Germany, Poland.

Case Study: Italy

The main targets have been geothermal exploration, correlation between T° and resistivity, reduction of mining risk (productive fracture identification).

3D reflection seismic has shown important achievements in this matters.





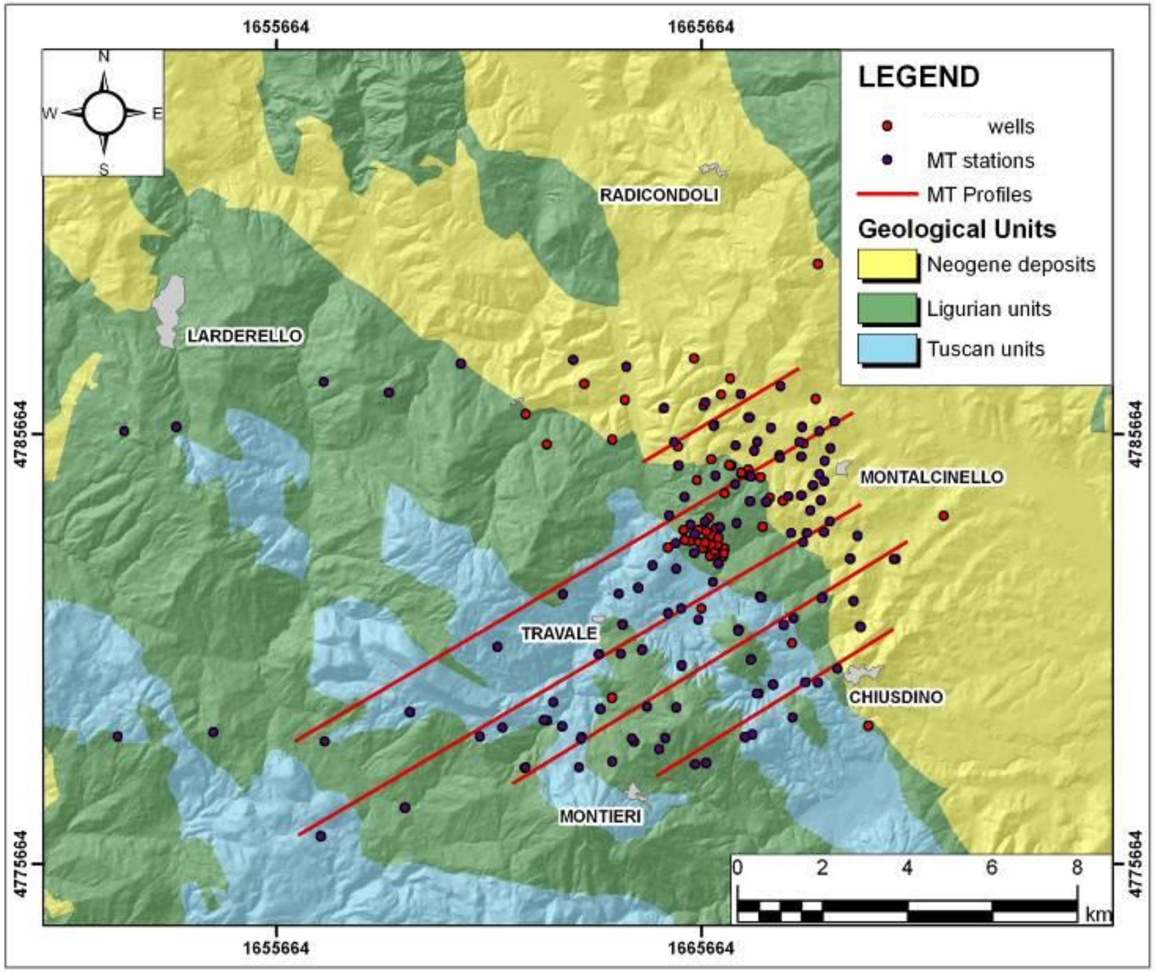
Correlation between seismic reflections (H marker) and fractures (red dots)

Courtesy of Enel

Case Study: Italy

MT surveys were carried on 2004 and a few months ago.

Some sites have been acquired outside the main exploited area along a profile crossing important faults



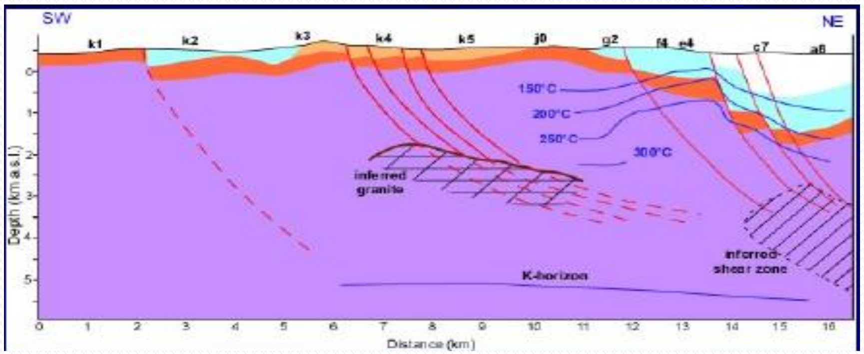
CNR-IGG A. Manzella

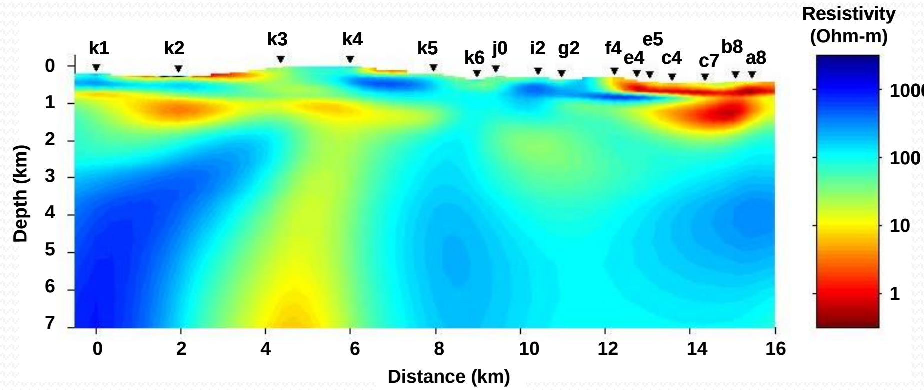
Case Study: Italy

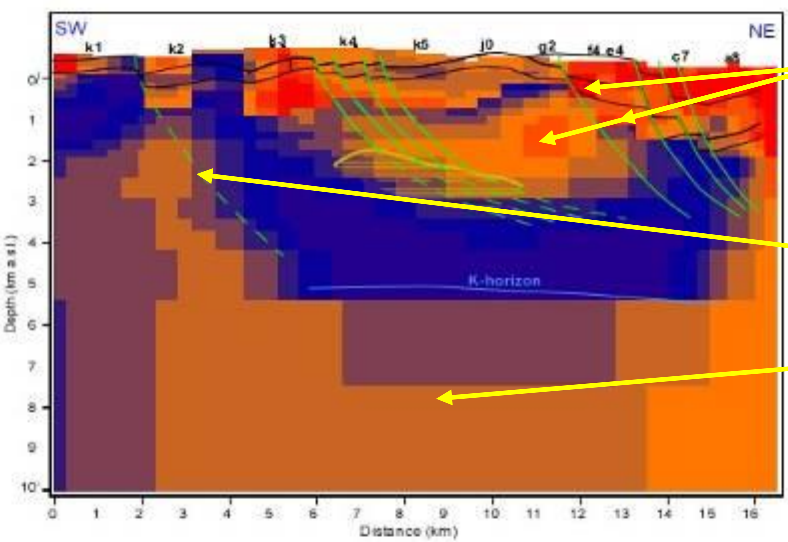
Data were acquired using both a local reference site (3-4 sites were acquired simultaneously) and a very remote reference site (Sardinia island on 2004, Capraia Island on 2006-2007).

Remote site in Sardinia was 500 km off the main area, and remote-reference technique is useful
for frequency < 1 Hz
(ok for coherent low frequency noise created by railways)

Case Study: Italy







low resistivity anomalies
inside the resistive basement

high-angle, conductive fault
zone-like structure close to
Boccheggiano fault

broad, deep crustal conductor
below the seismic K-horizon

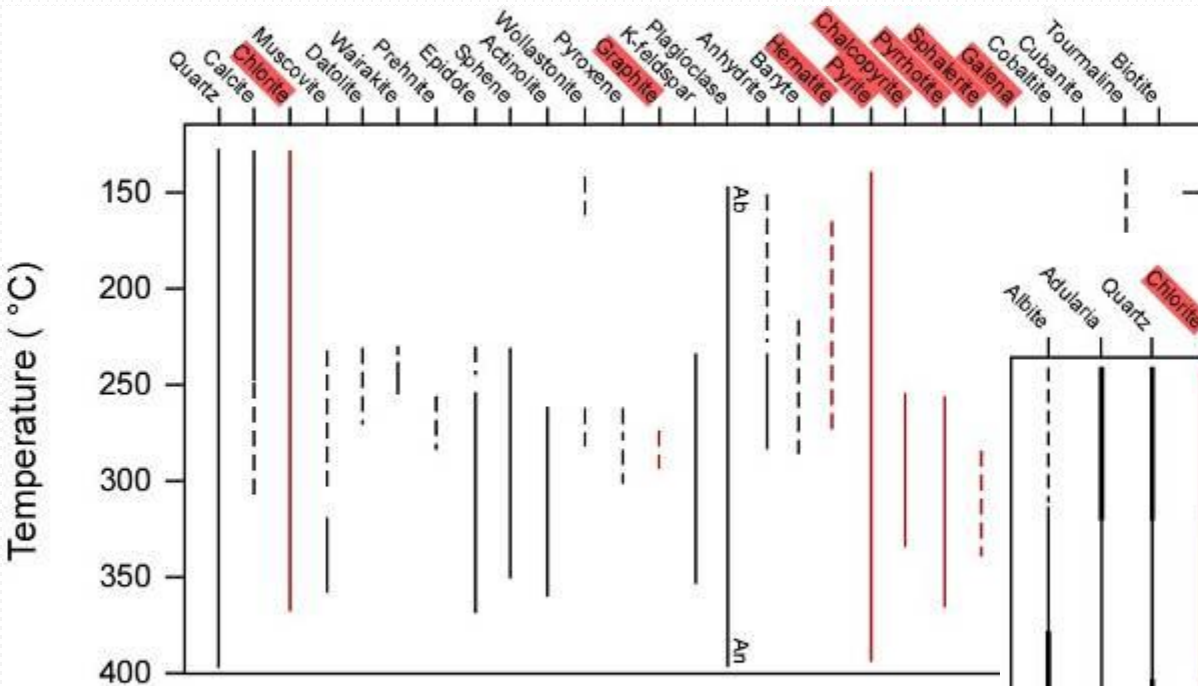
2D TE-TM inversions using
different approaches

Case Study: Italy

Two possible explanations for the resistivity reduction from 10^3 to 10^0 Ωm observed in Travale remain: the effect of alteration minerals and the presence of brines at liquid phase whose interconnection is sufficient to produce electrolytic conduction.

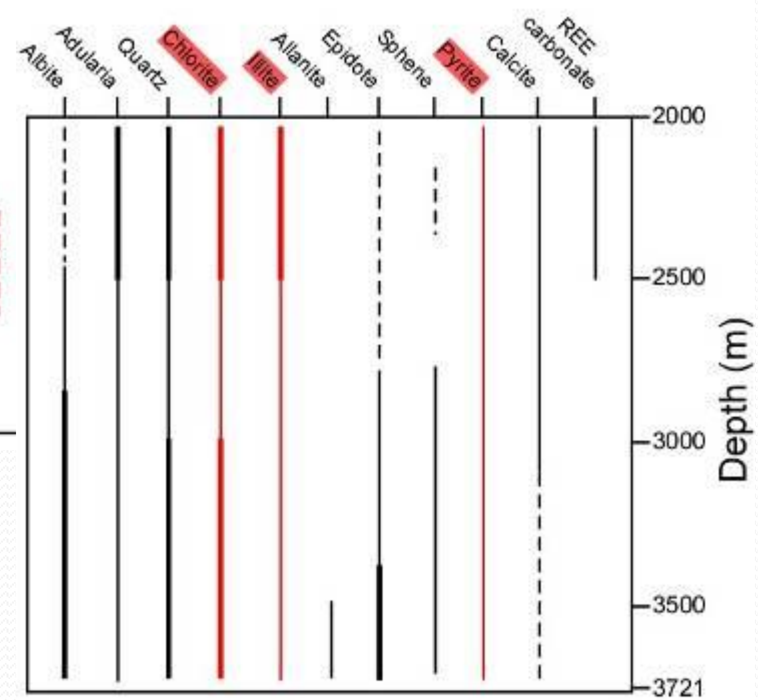
The strong resistivity decrease, at the moment not clearly explained by lithological changes, opens new questions regarding the importance of alteration, mineralization and fluid nature within the geothermal reservoir.

Case Study: Italy



Distribution of the hydrothermal minerals with temperature in the Larderello field.

Conductive minerals are highlighted in red.

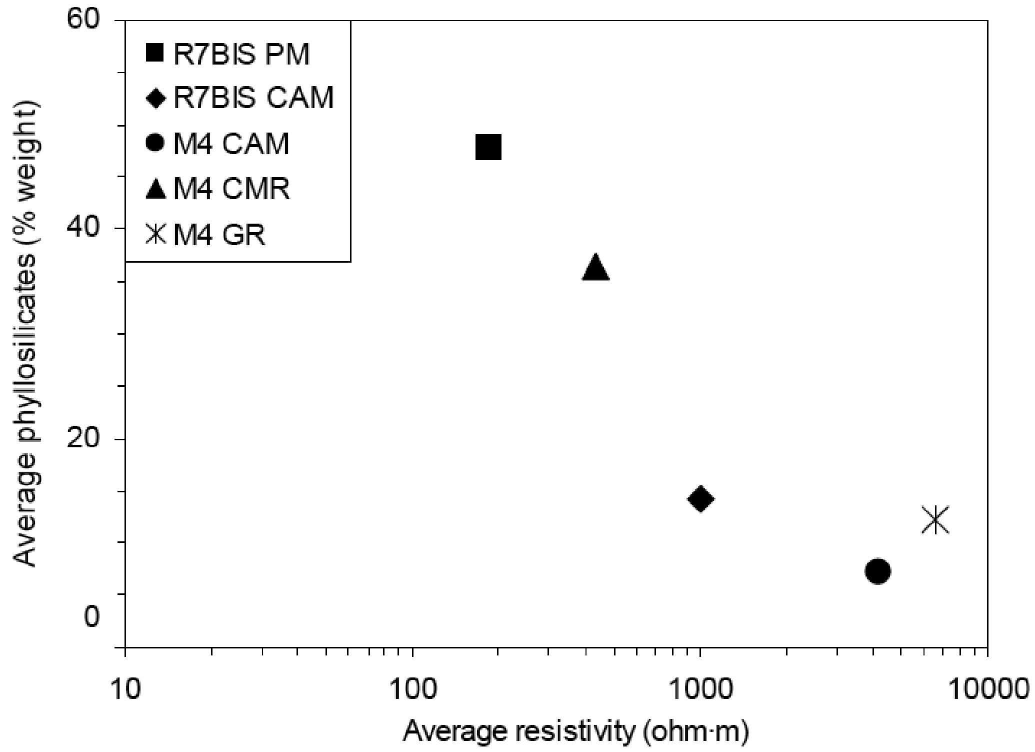


Distribution and abundance of authigenic minerals in the granite crossed by MONT-4 well.

Conductive minerals are highlighted in red.

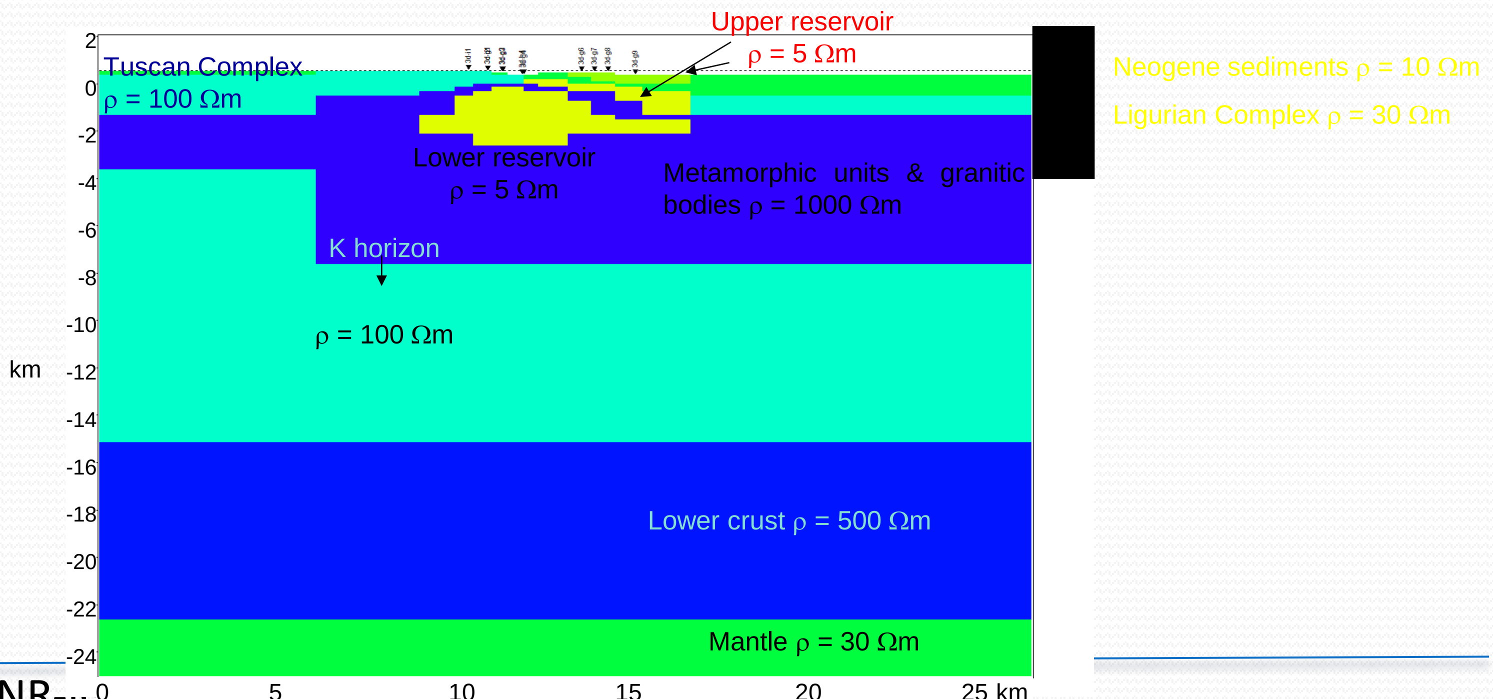
Case Study: Italy

.A mineralogical analysis on samples from two wells determined that lithology and heterogeneities of reservoir rocks and the alterations affecting them may only account for the small variation observed in the resistivity logs, but not for the main anomaly detected by the magnetotelluric data.



Case Study: Italy

Different 3D models of Travale area were developed on the basis of 1) the resistivity of lithological units obtained from a resistivity log; 2) zones of high temperature and productive fractures indicated by drillholes; 3) bright-spot type horizons identified in seismic profiles; and 4) estimated resistivity values for the lower crust and mantle. Responses of forward modelling were calculated and compared with MT data measured in the field.



Case Study: Italy

• Neither the resistivity of the lithological units nor the presence of an upper and a lower reservoir limited to the currently productive fractures can explain the conductive anomaly evidenced by MT data at the Travale geothermal field.

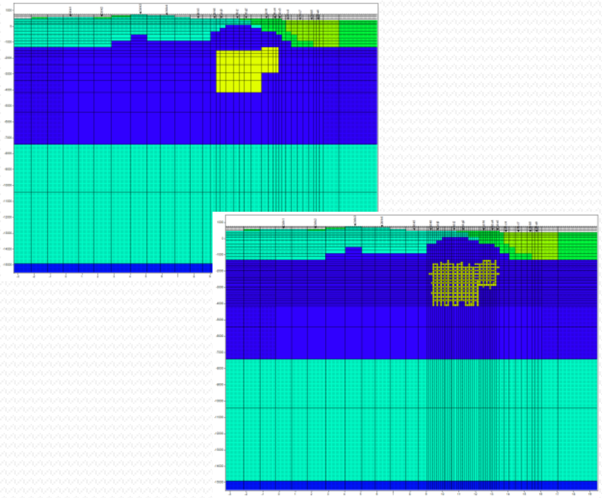
• The closest fit to the measured MT data corresponds to the model comprising a deep conductive body of ~ 60 km^3 volume, *i.e.*, larger than the one enclosed by the currently exploited productive fractures.

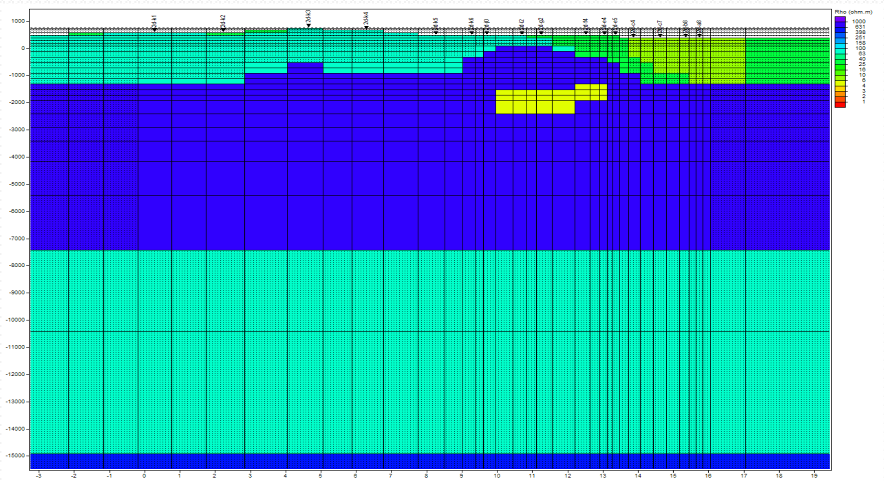
• It is still a matter of discussion whether the conductive anomaly corresponds to a liquid phase within the lower reservoir, the presence of widespread, connected alteration minerals or both.

• When a relatively conductive lower crust and a conductive mantle are both present in the models, we obtain better fits to the measured MT data. This is in agreement with the resistivity decrease expected at these depths due to the higher temperatures and presence of melt phases in an active tectonic region such as southern Tuscany.

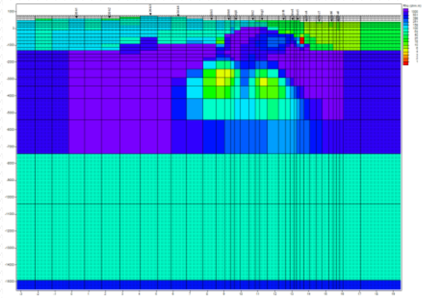
Case Study: Italy

In order to test of MT resolution and optimize site density, we computed synthetic MT data by 3D forward models with different kind of reservoirs (pore-controlled-resistivity and fracture-controlled-resistivity) of various dimensions and inverted these results in 2D using *a priori* models including only the electrical resistivities due to the lithological units.

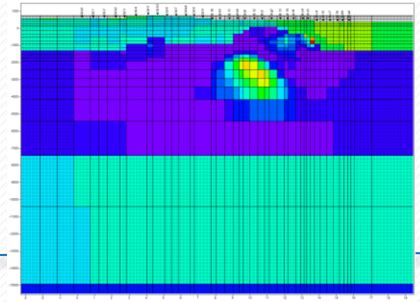




Model simulating a small (2.4 km²) pore-controlled-resistivity reservoir within a real geological section



Sparse sites



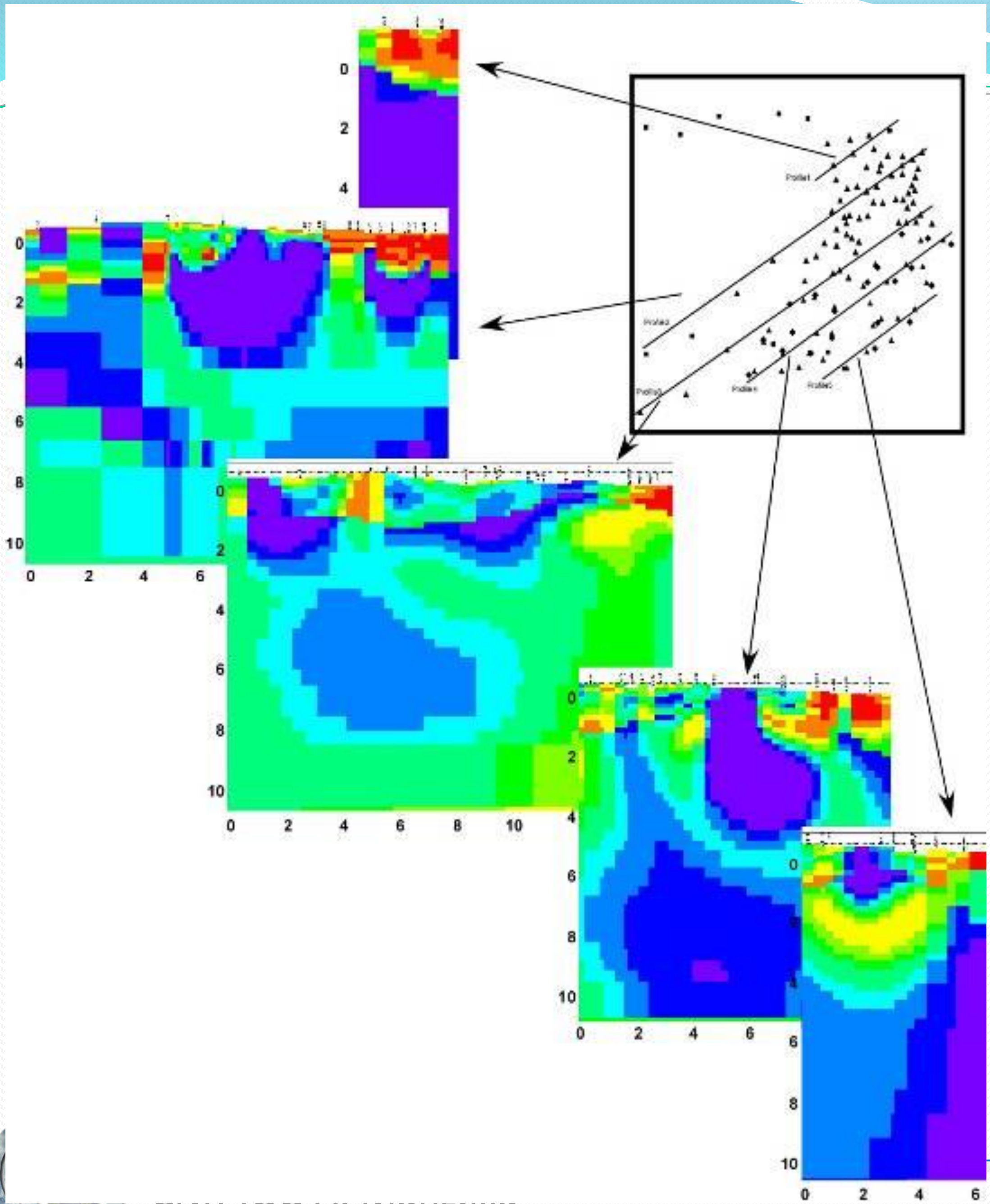
Dense sites

Case Study: Italy

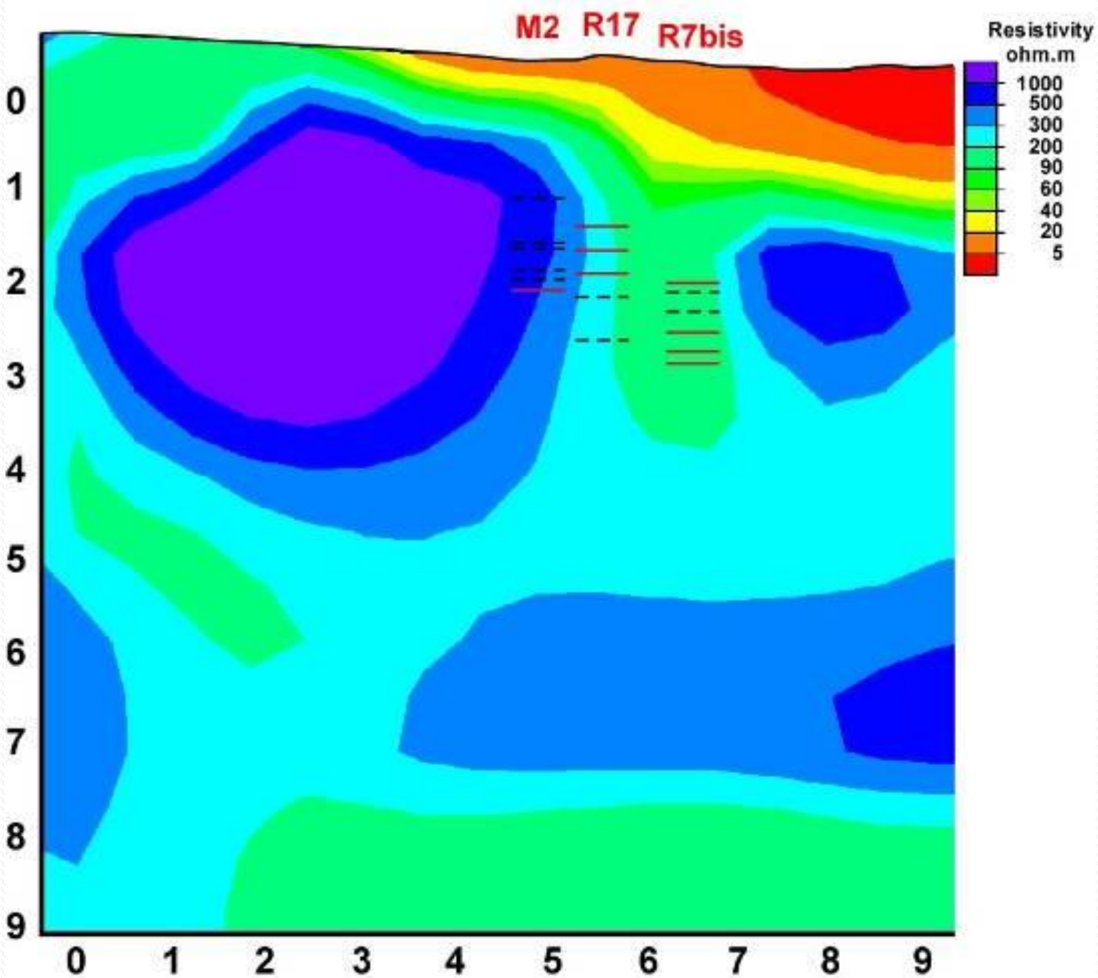
• In the models including the small and large pore-controlled-resistivity reservoirs, 2D inversions are able to define the lateral extent and top of the conductive bodies, but do not identify the bottom boundary. On a dense MT station profile (stations every ~ 500 m) the geometry of the reservoirs is attained with a slightly better resolution.

• The difference in size between the reservoirs is practically unattainable by 2D inversions. The MT method can only calculate the conductance (conductivity*thickness) of an anomalous body.

• In the case of the fracture-controlled-resistivity reservoir, 2D inversions are unable to define the grid of fractures, not even with a high density of MT stations. Furthermore, we obtain an overestimate of the anomaly depth.

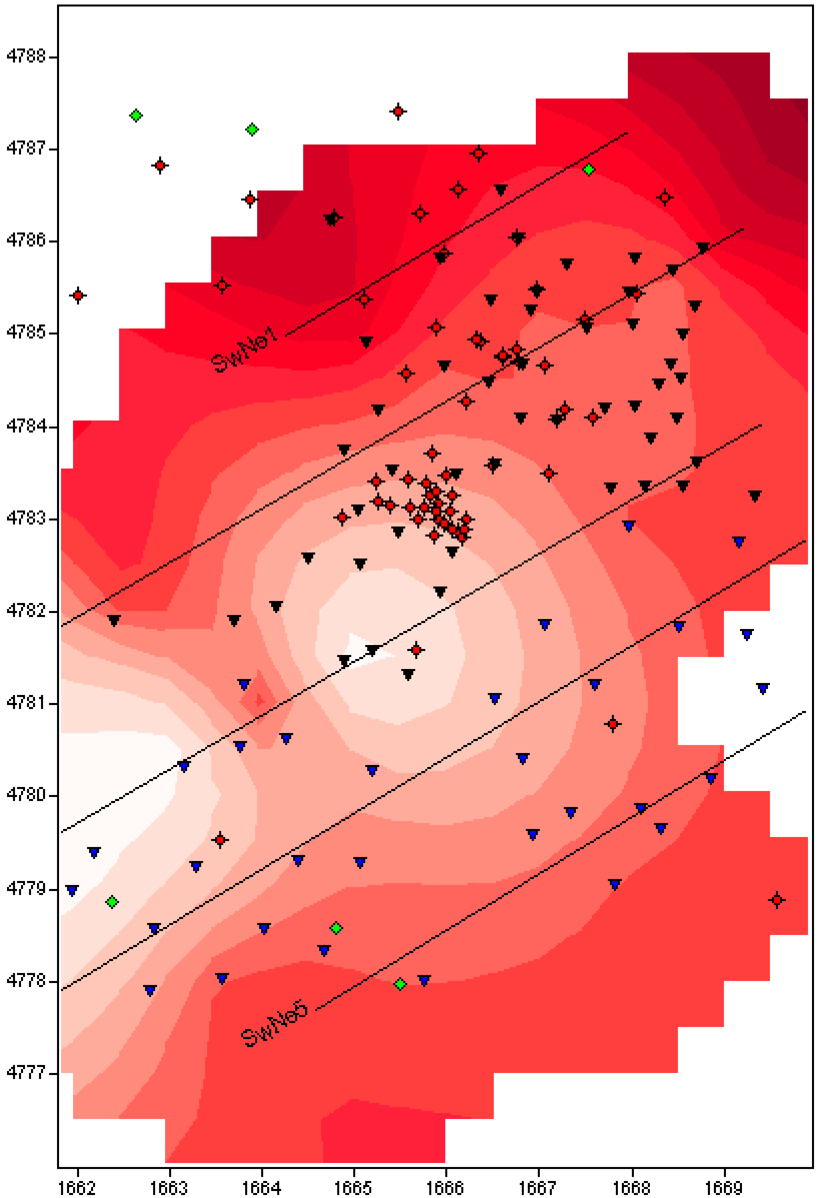


2D TE-TM inversions



Productive fractures can not be defined just on the base of MT data. However, MT might provide info regarding areas of enhanced productivity

0 km b.s.l.



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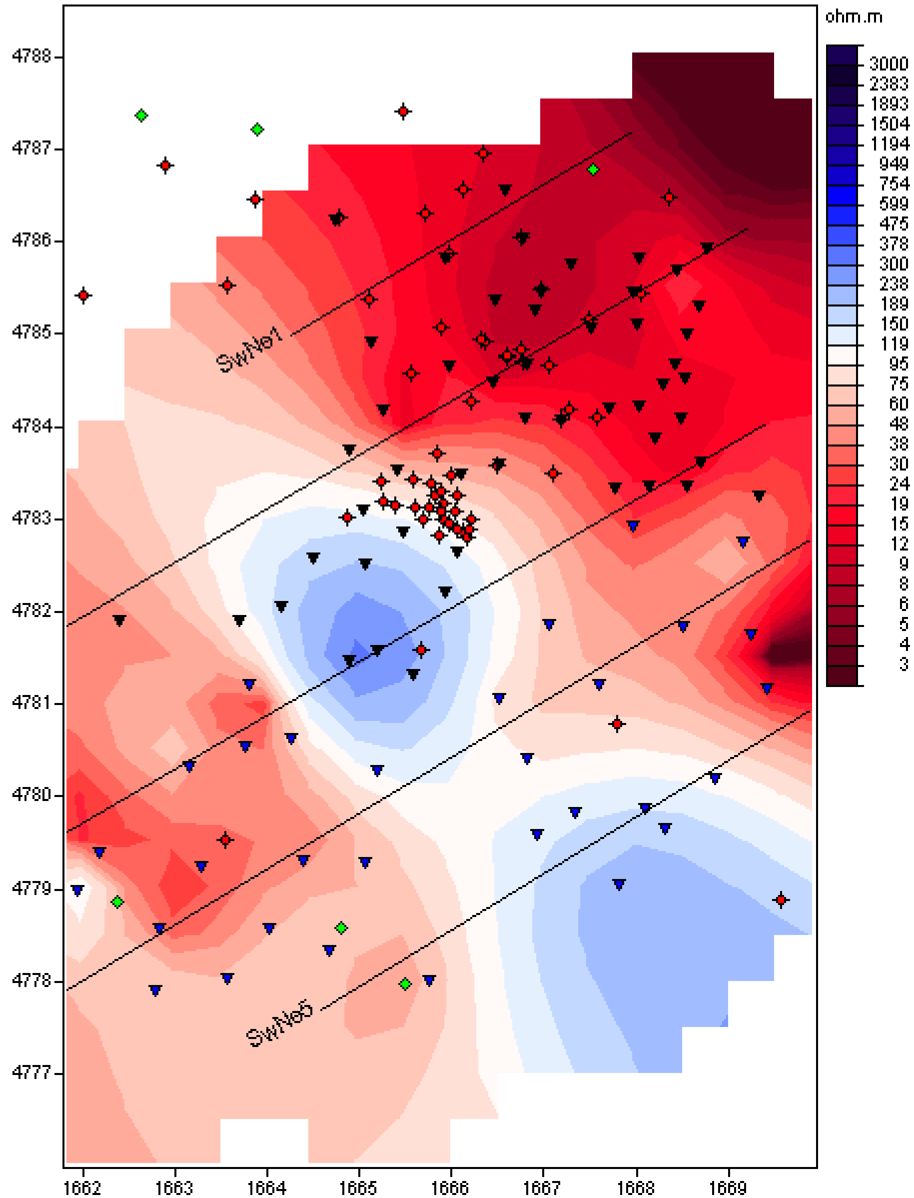

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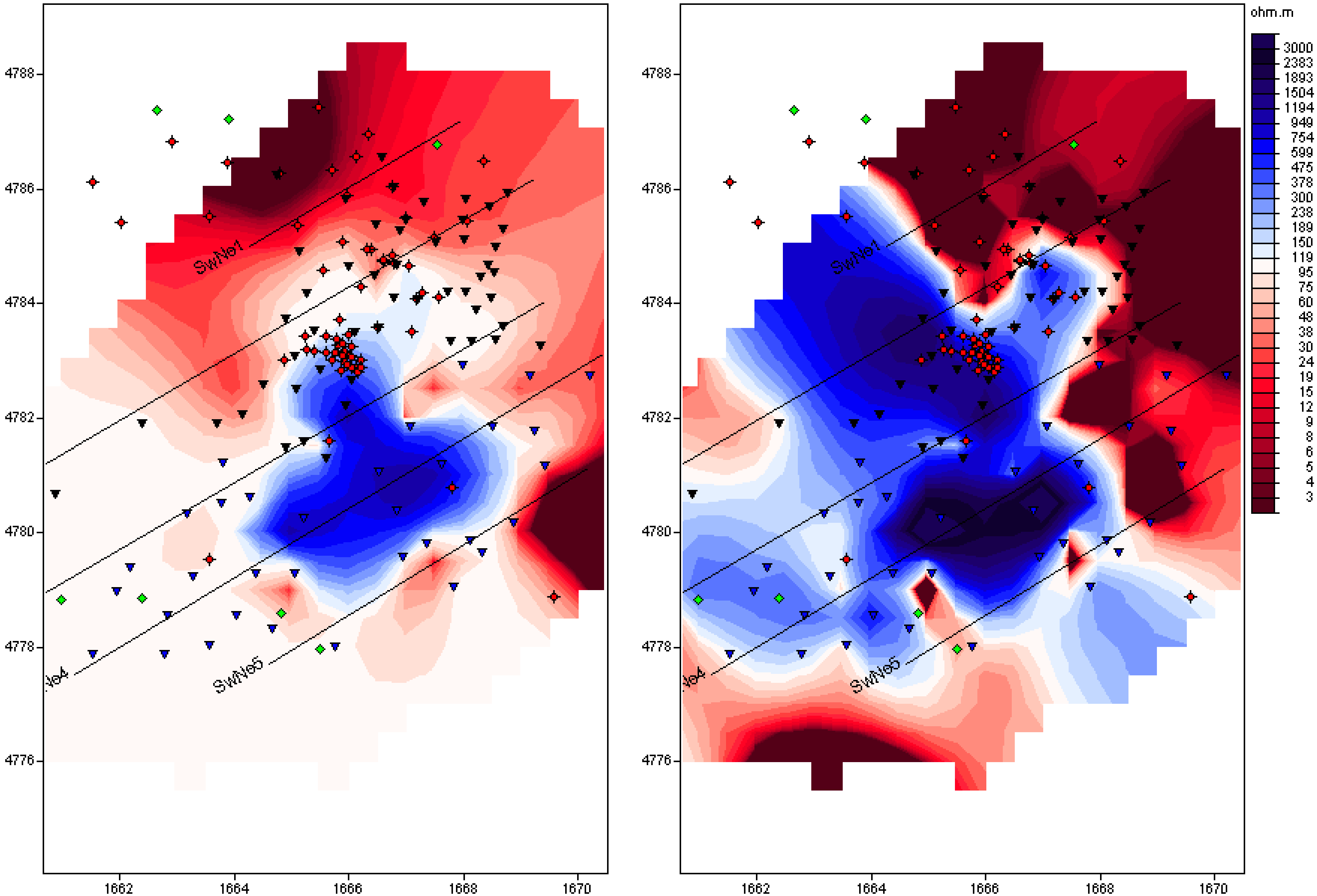
0.5 km b.s.l.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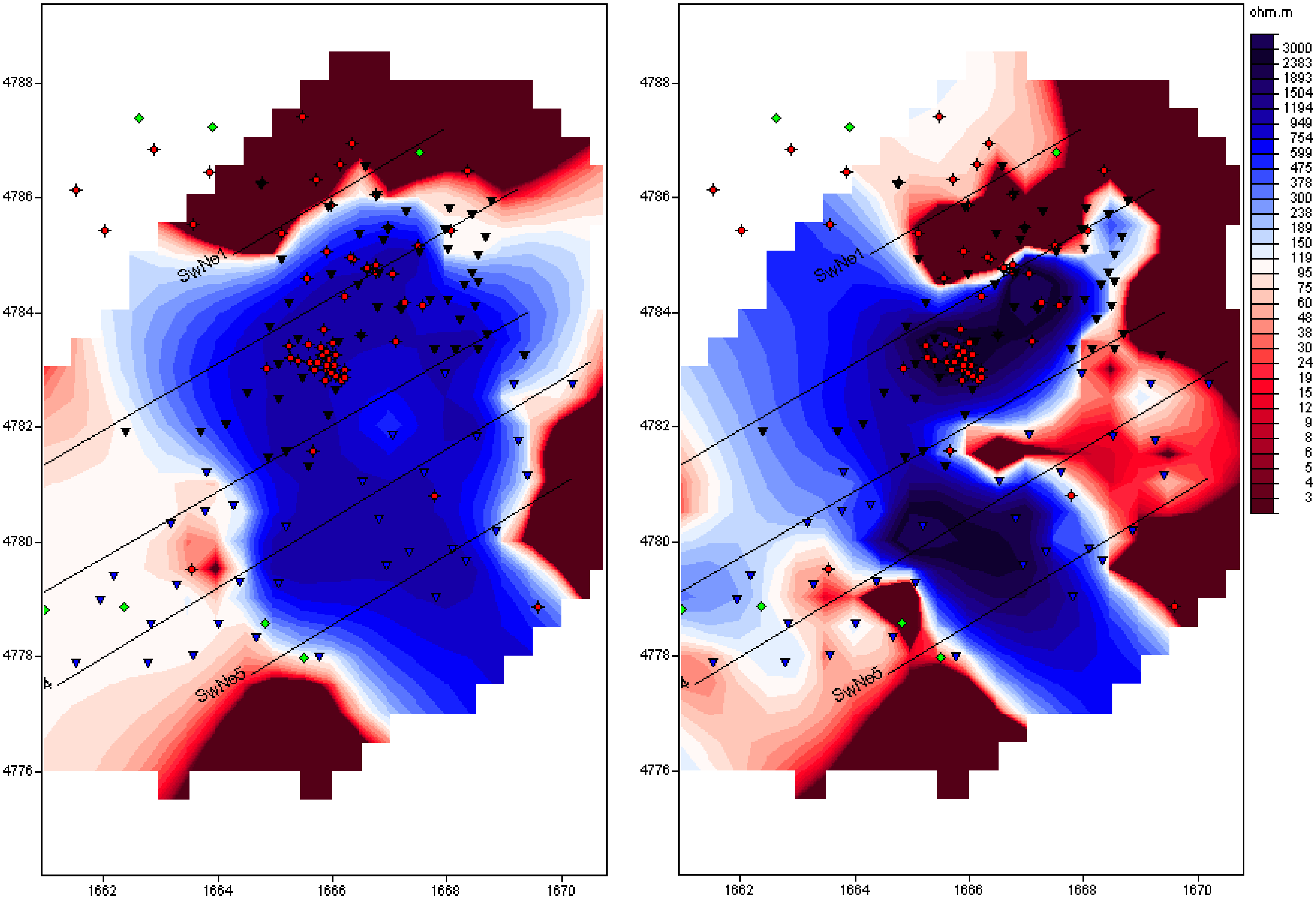

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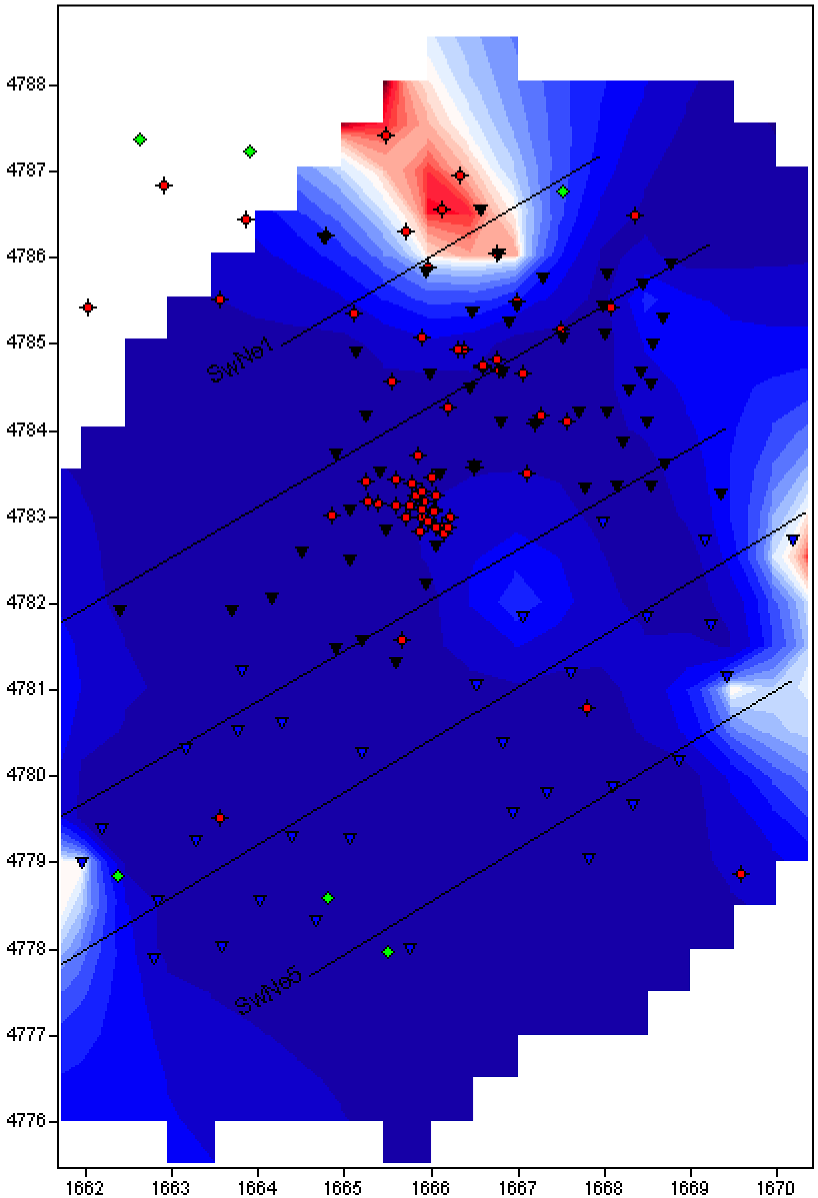

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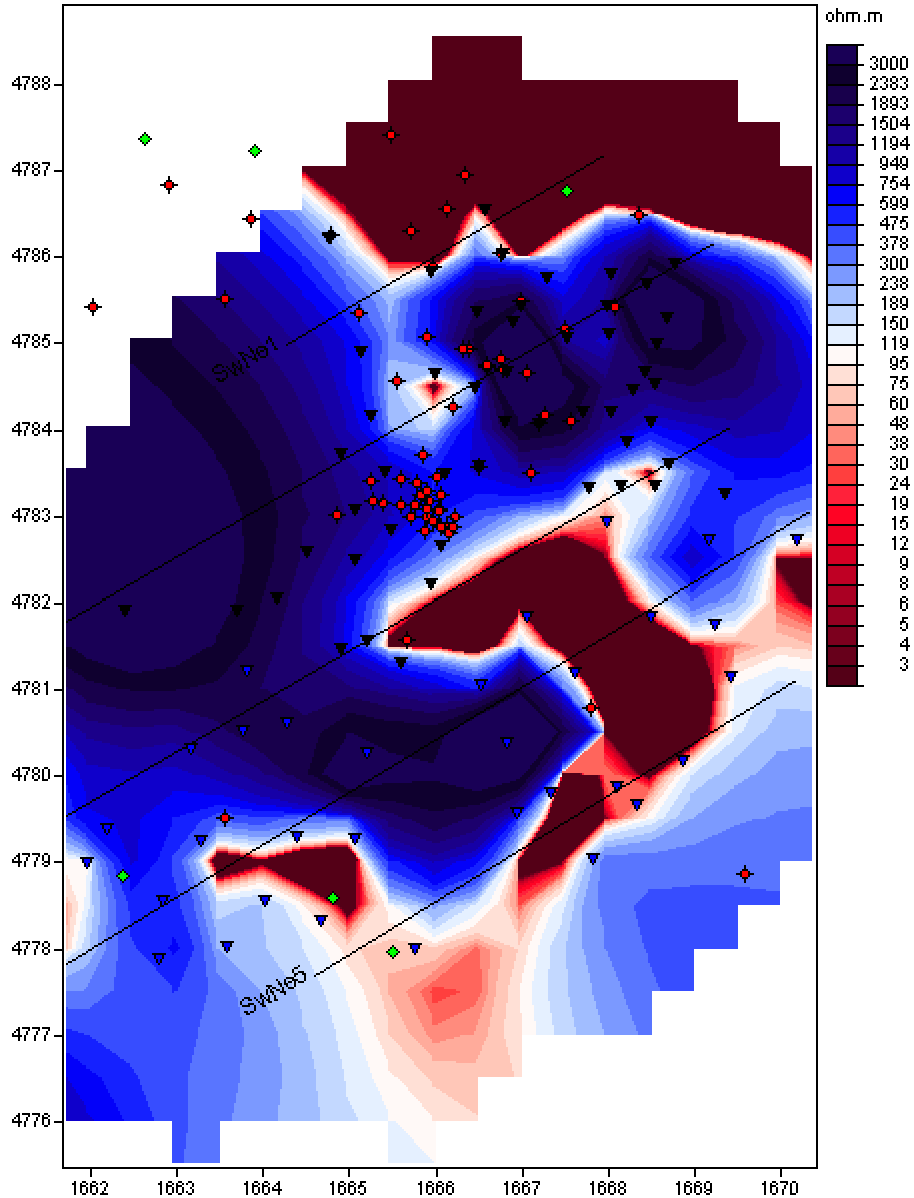

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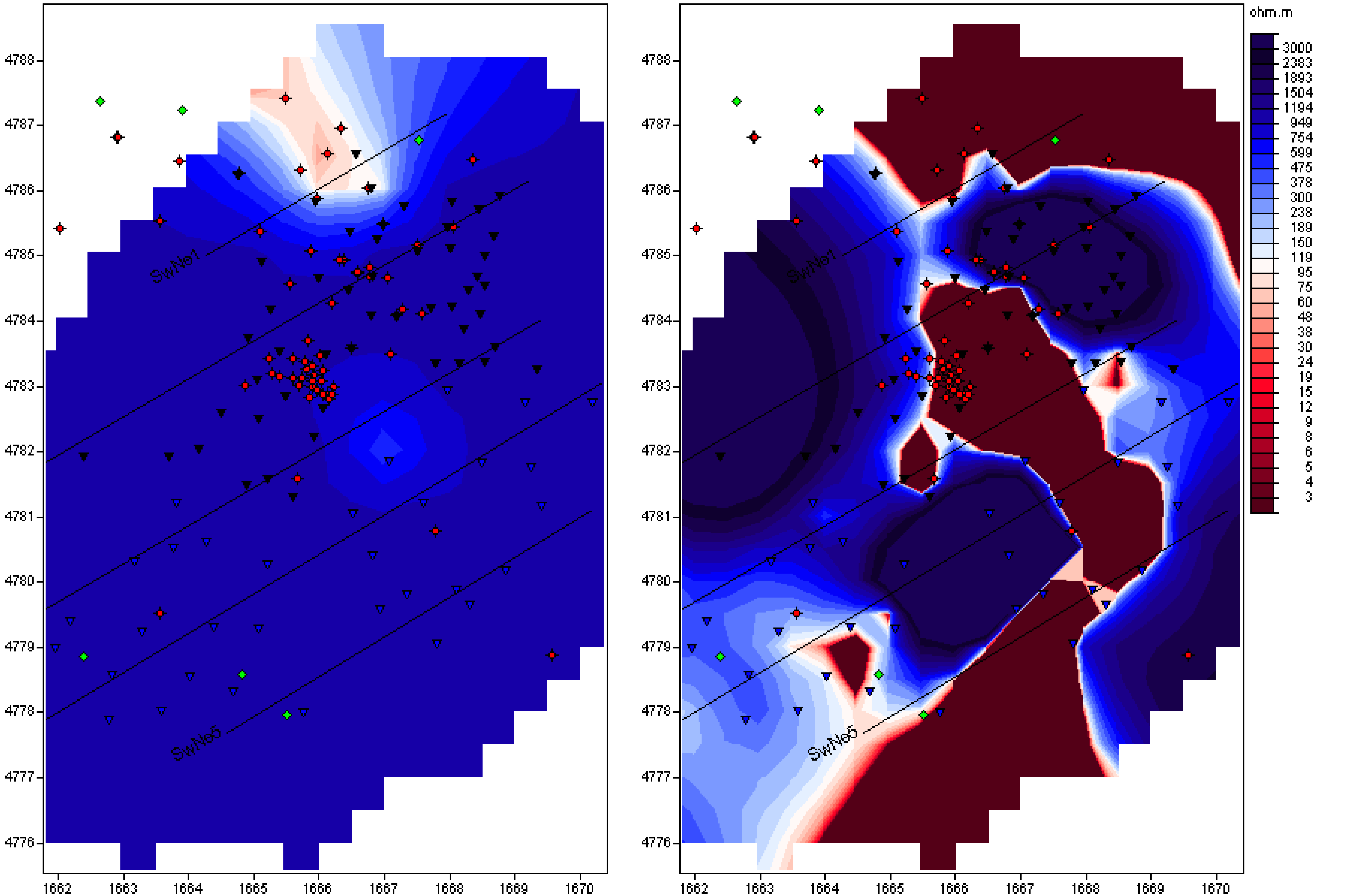

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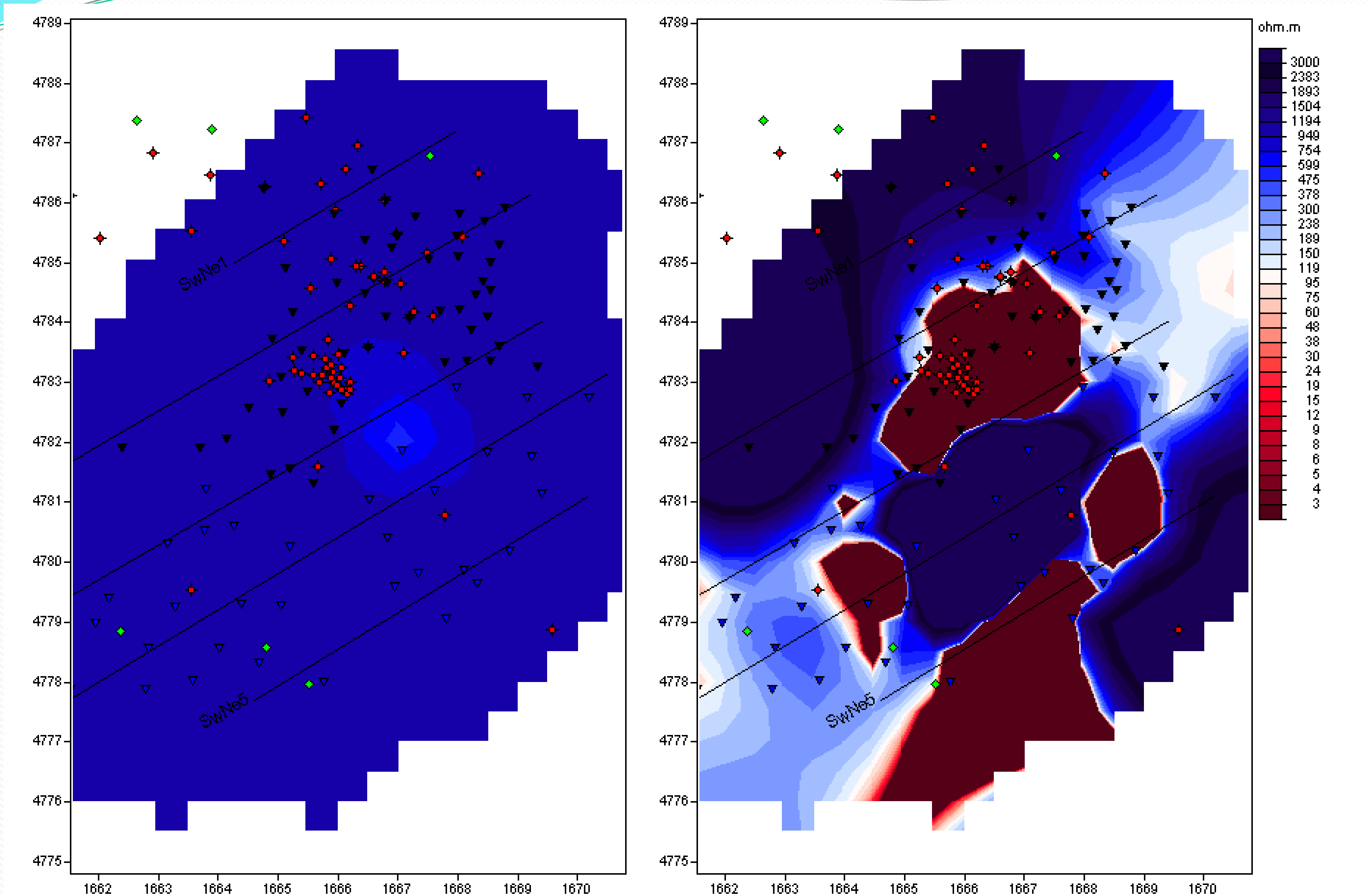
3 km b.s.l.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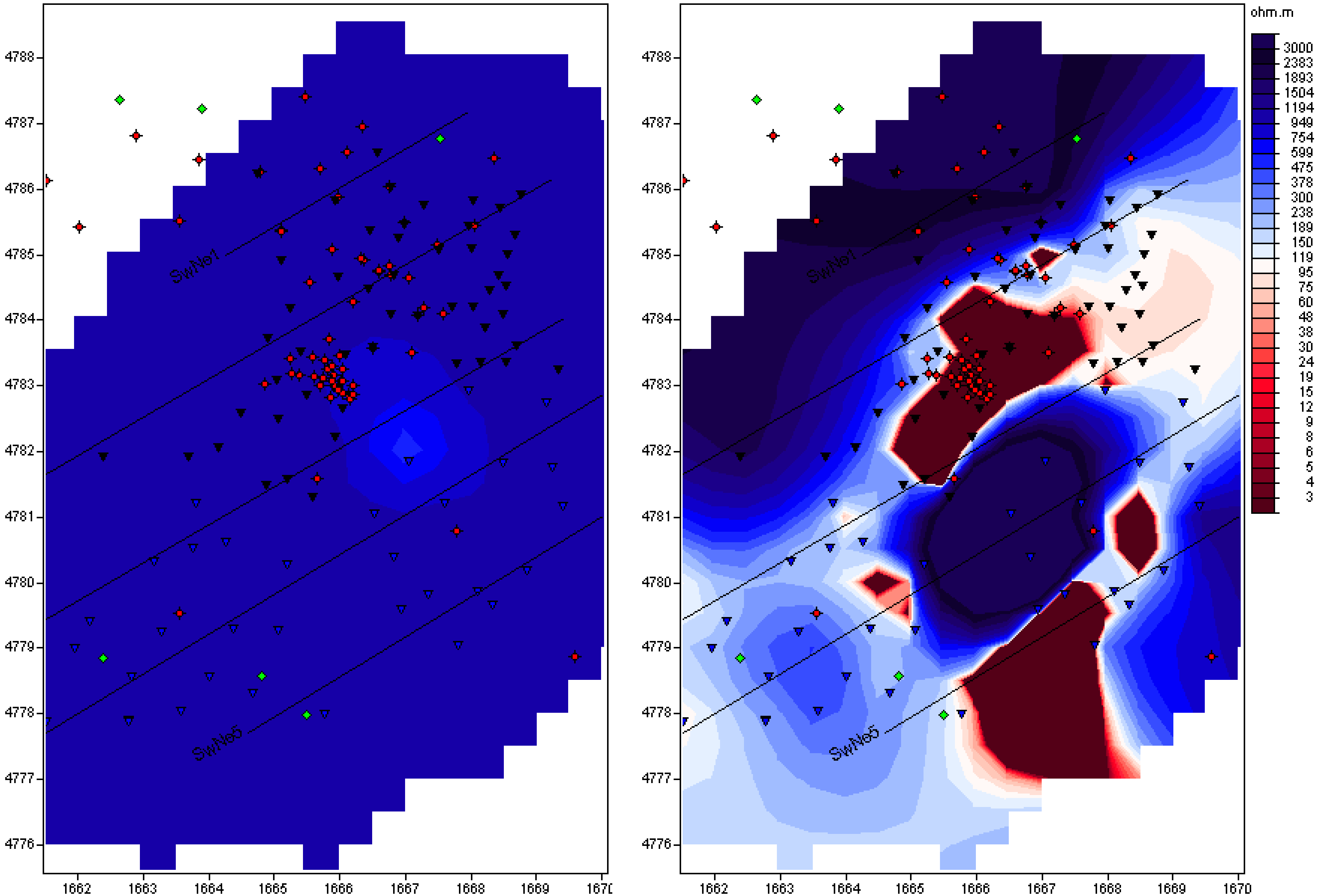

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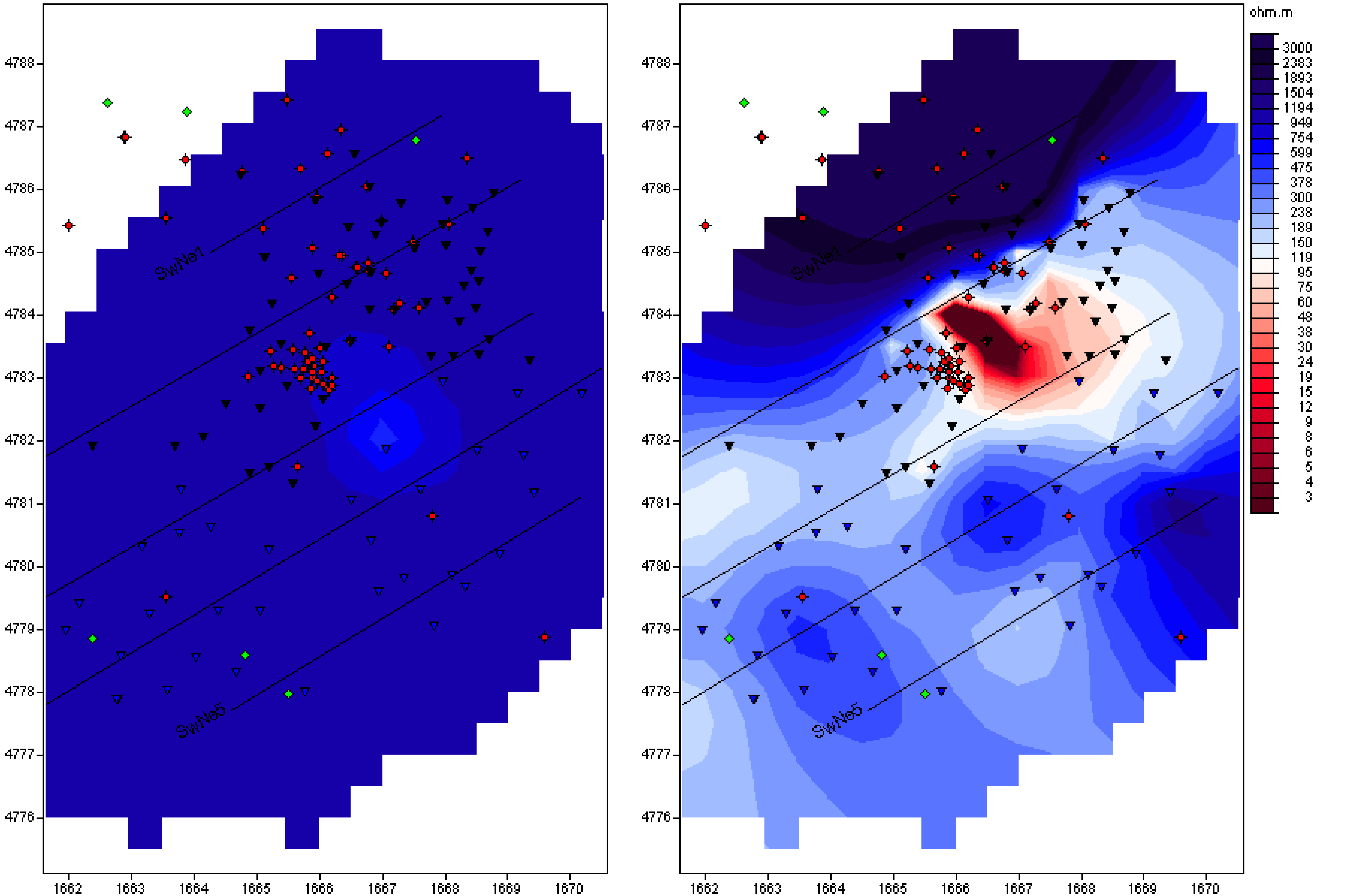
5 km b.s.l.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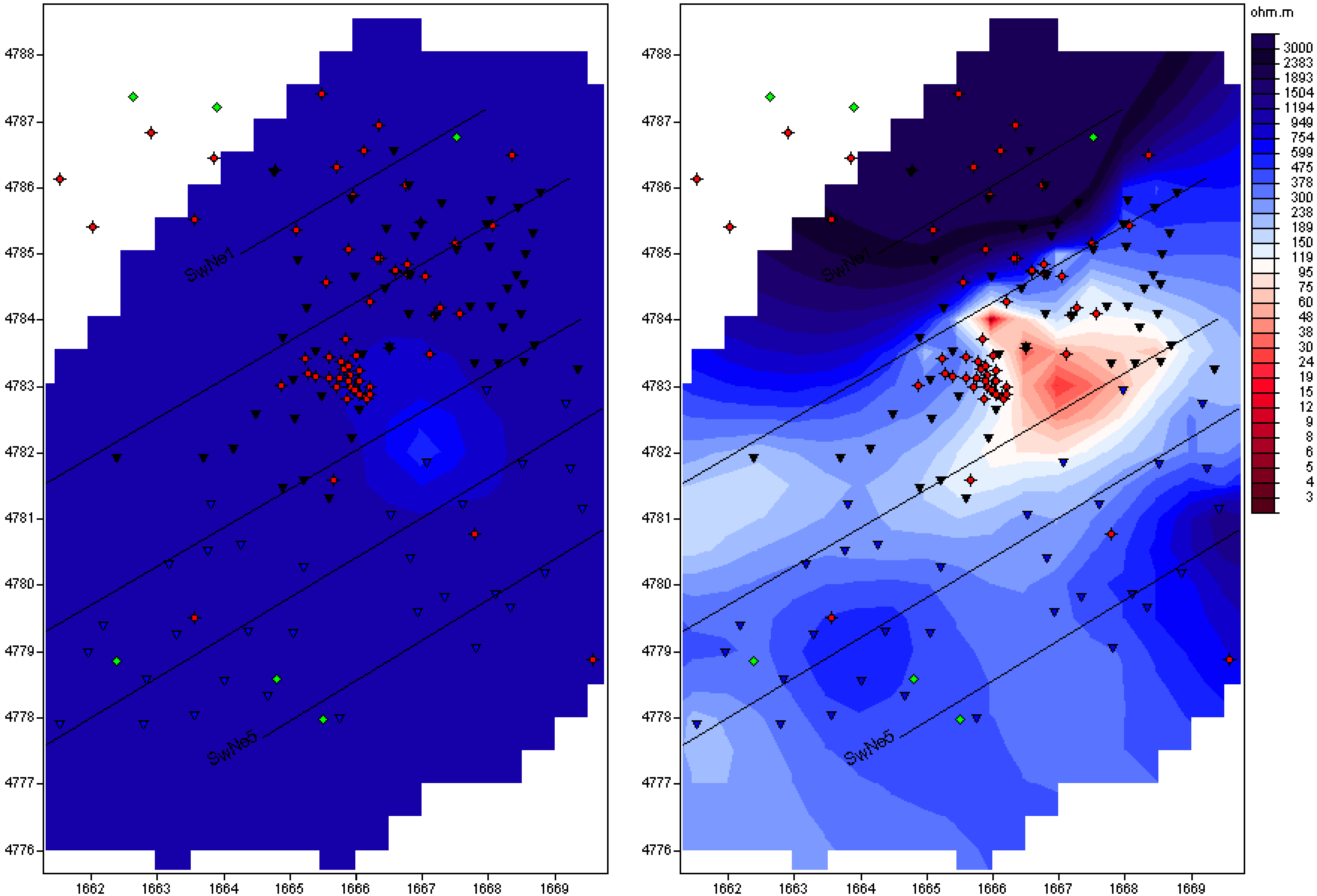
6 km b.s.l.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Advantages and disadvantages of artificial-source EM methods compared with natural-source sounding methods:

- provide better resolution and are less easily distorted by lateral variations in resistivity than soundings obtained with natural-source methods

- may be more expensive than natural-source sounding (shallower penetration, higher number of soundings, transmitter)

- only a few km may be investigated whereas natural methods may sounds several km

- are effective in resolving the parameters of conductive layers but are less effective in determining the resistivity of resistive layers in comparison to direct-current (dc) methods

- highly resistive layers do not screen deeper layers from being resolved by EM soundings as is the case with dc soundings

- depend only on the longitudinal conductivity of a horizontally layered earth whereas resistivity measurements depend on both transverse and longitudinal resistivity. Thus, in principle, EM soundings can yield more accurate depth estimates than dc resistivity soundings over an isotropic earth.

Of course, EM soundings made with a grounded wire source and receiver give dc resistivity results at asymptotically low frequencies.

The question of whether EM or resistivity soundings are least affected by lateral variations in resistivity depends on the specific techniques used in each case and on the scale of the inhomogeneity.
The relative cost of EM and resistivity soundings depends on the technique and the application; either can be the least expensive for a particular purpose.

Applications

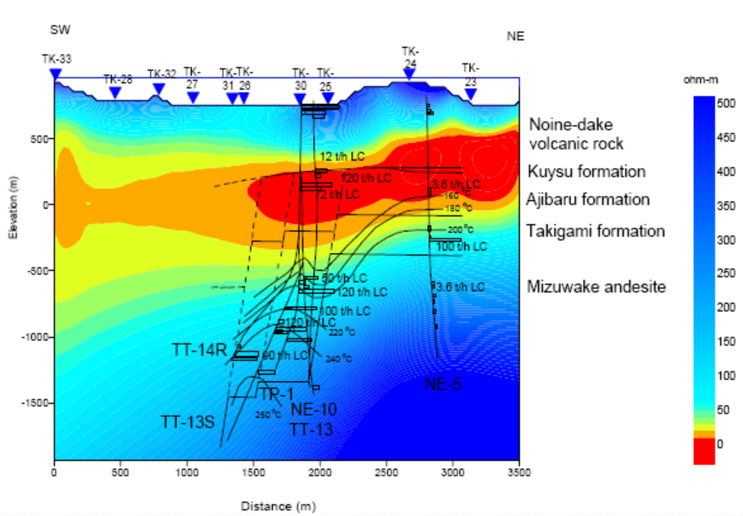
Various targets can be imaged by resistivity geophysical methods

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

Applications

Regional exploration

Takigami Geothermal Area, Japan



From Ushijima et al., WGC 2005

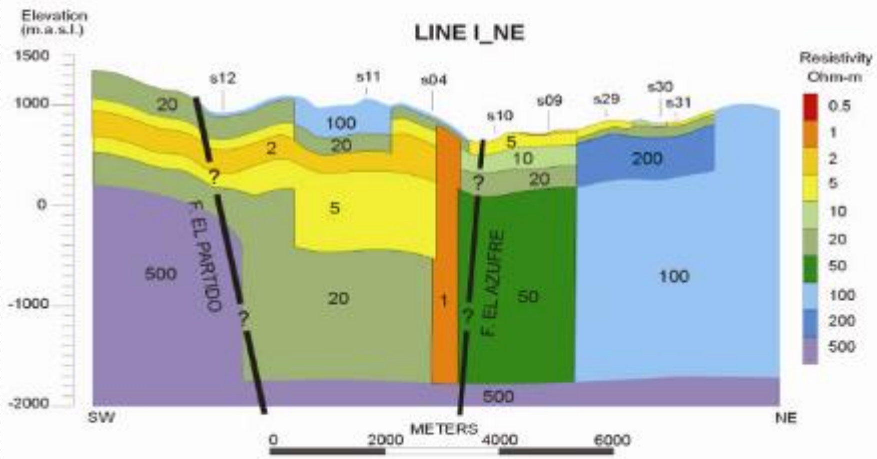
"the low resistivity zone in the northeastern part is intensive and shallower than that in the southwestern par,t in good agreement with the geological feature"

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

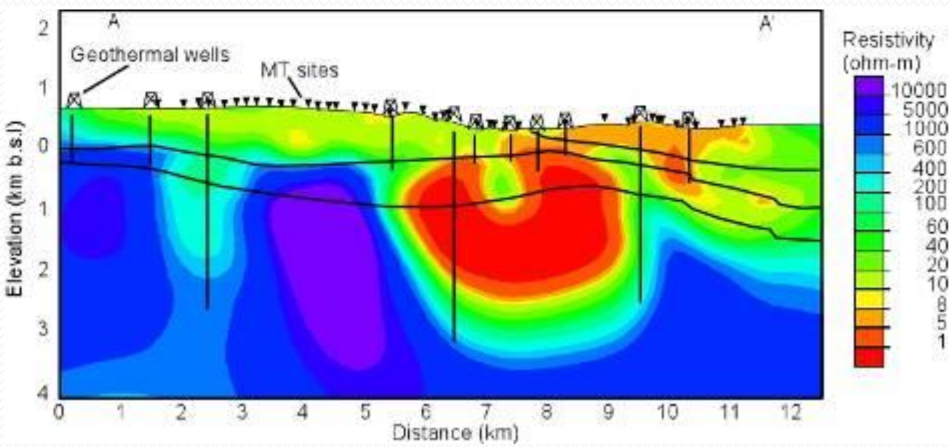


Las Tres Virgenes Geothermal Area, Mexico

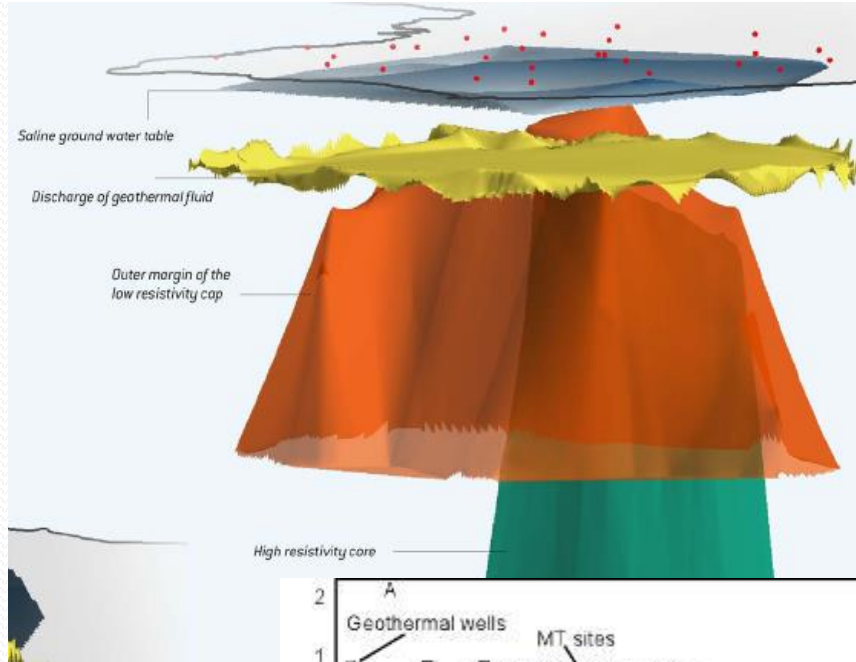
From Romo et al., WGC 2000

The results suggest the presence of a highly attenuating and conductive zone along El Azufre Canyon, which corresponds with the production interval of wells LV-2 and LV-3/4. A graben structure is outlined.

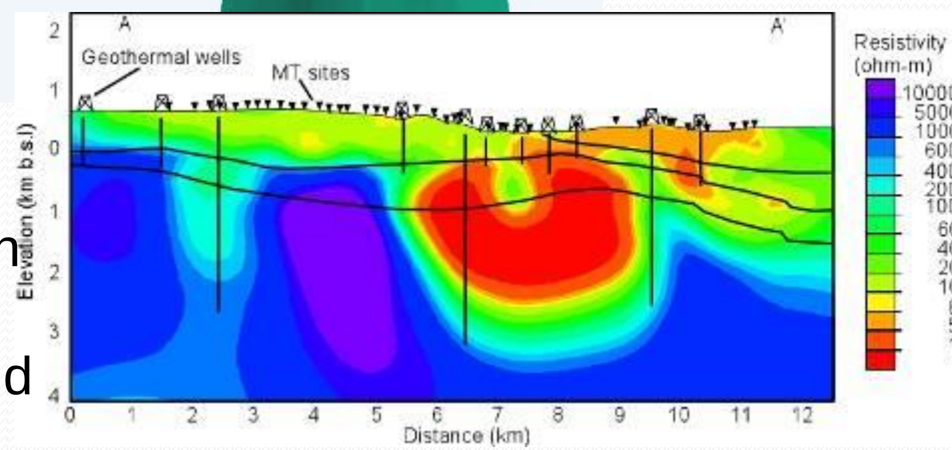
Mt. Amiata Geothermal Area, Italy



In volcanic rocks TDEM and MT have defined the main structure, driven mainly by alteration minerals



From Karlsdottir, ENGINE Worhshop1



MT maps the effect of main litological units.
Combination of seismic and MT data.

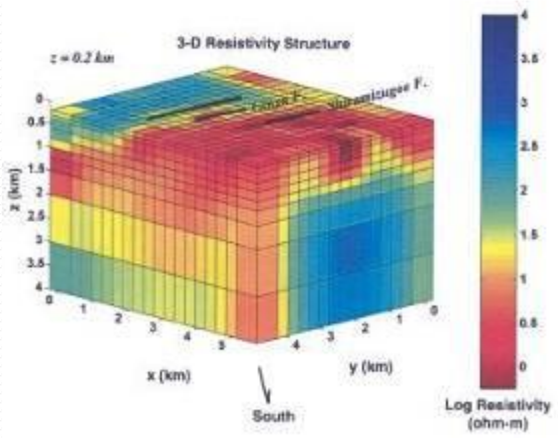
Applications

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)

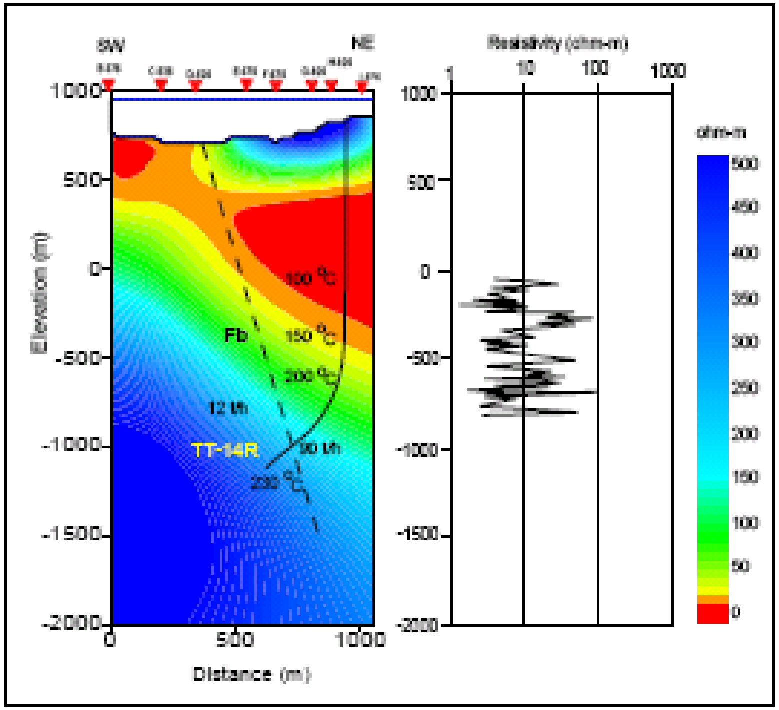
Ogiri geothermal zone, Japan

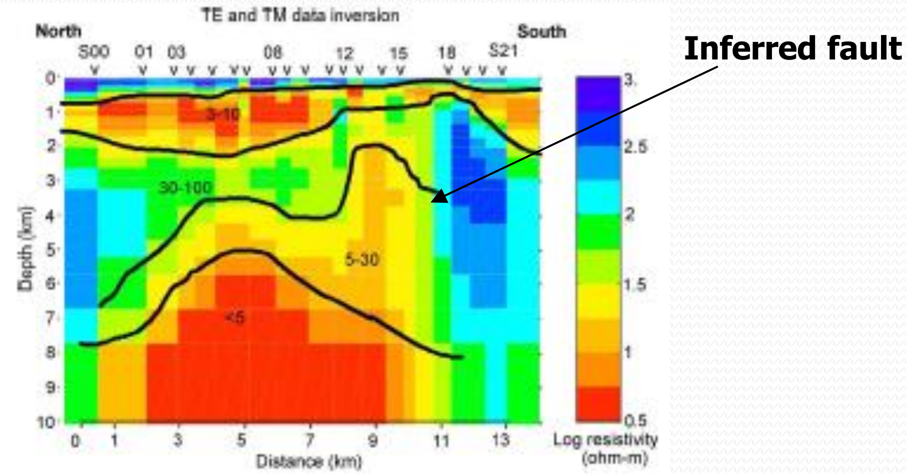


From Uchida, 2005

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.

Area geotermica di Takigami, Giappone



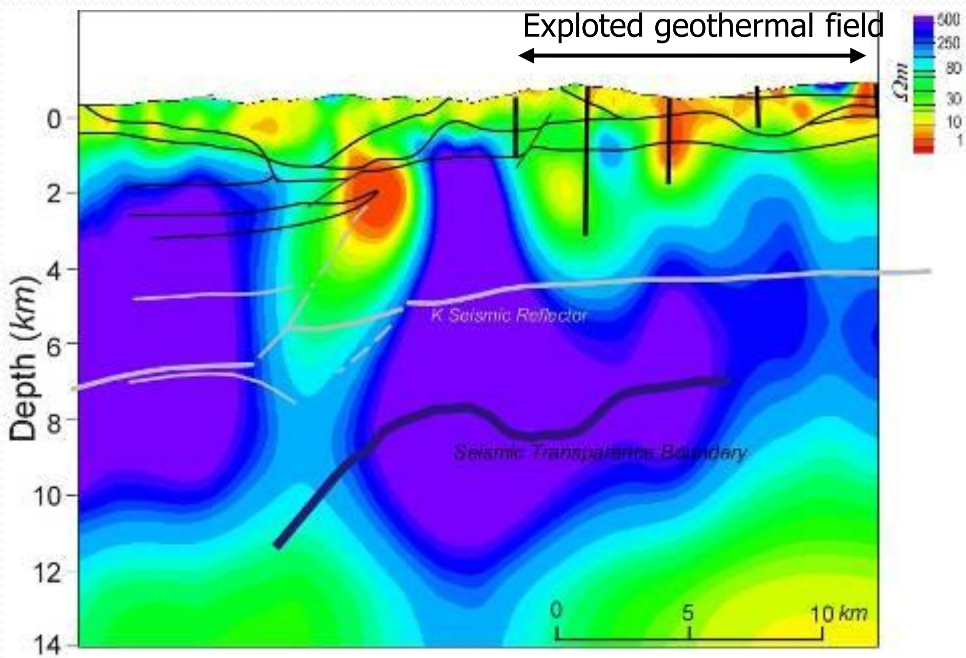


Oskooi et al., 2005

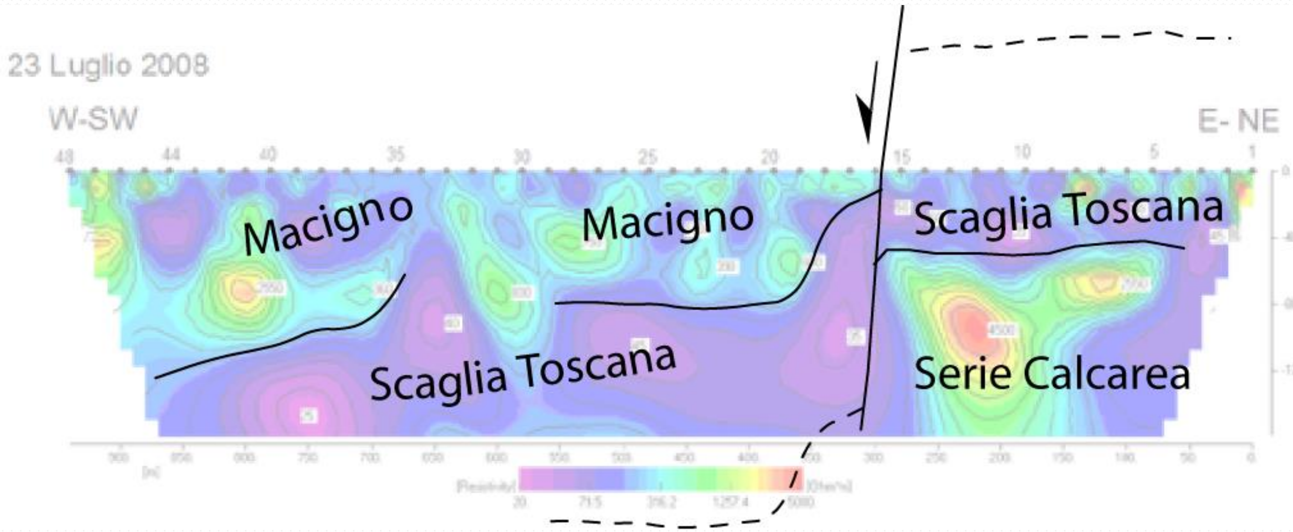
Tratto da Tagomori et al., WGC 2005

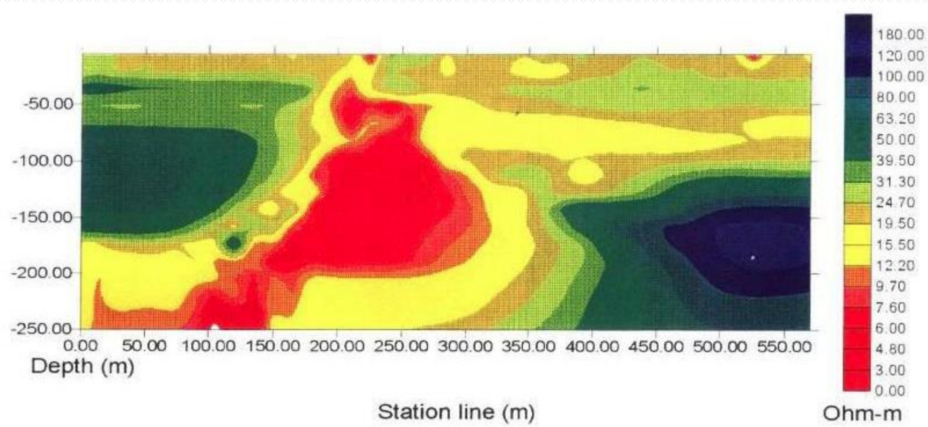
"the large lost circulation occurred at the depth of 1300 m BSL for the well TT-14R (90 t/h) when the well crossed through the electrical discontinuity Fb"

The correspondence between areas of low resistivity inside the resistive basement and geothermal reservoirs was very evident in the Mt. Amiata water-dominated system.
A fault defined by 2D reflection seismic corresponds to a low resistivity anomaly > water and/or clay
Heat flow provide extra data



Geothermal exploration at a low enthalpy field using ERT. The low resistivity zone is coincident with a known fault.





Geothermal exploration at a hot spring area near Beijing using AMT. The low resistivity zone is coincident with a known fault.

Applications

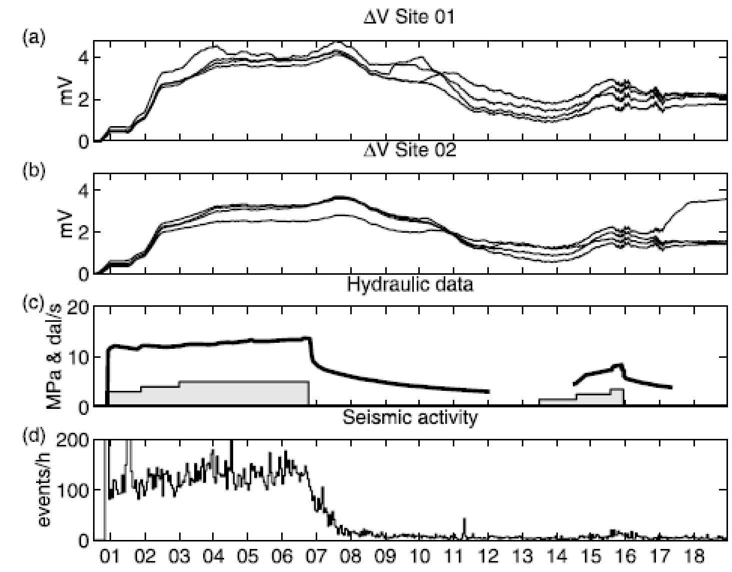
Monitoring

Phase change of pore fluid (boiling/condensing) in fractured rocks can result in resistivity changes that are more than an order of magnitude greater than those measured in intact rocks

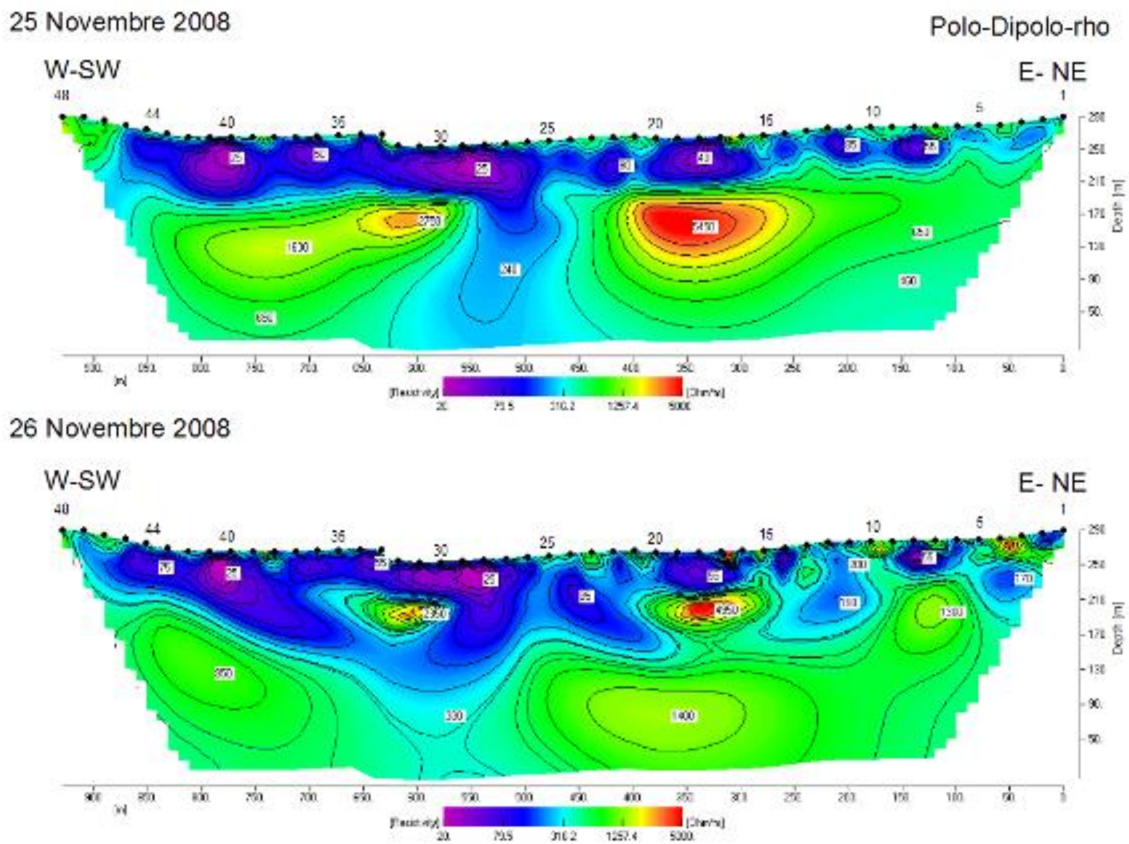
Production-induced changes in resistivity can provide valuable insights into the evolution of the host rock and resident fluids.

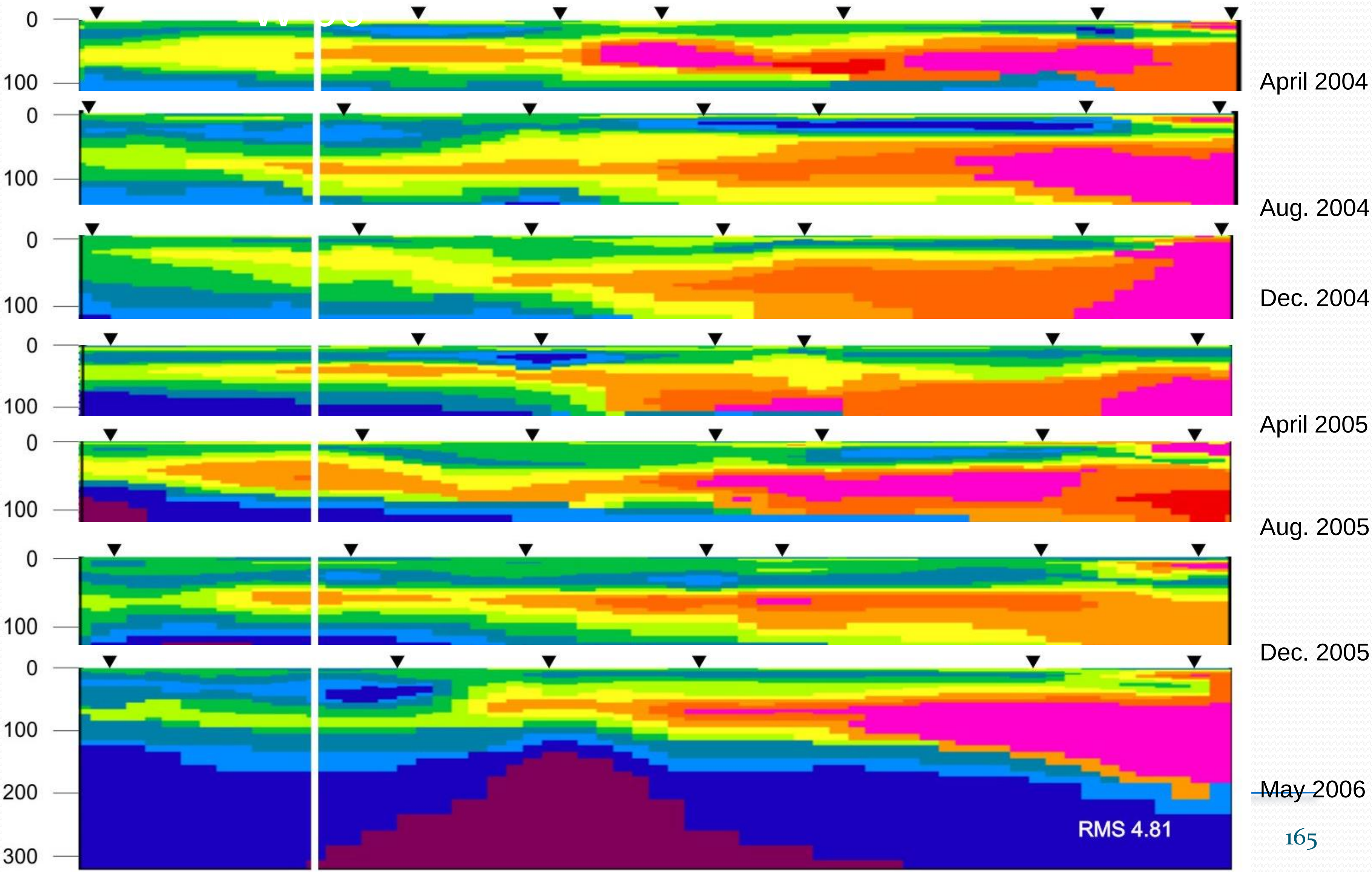
No examples or applications found in literature

Some examples from SP (electric field) showing interesting results: is it possible to use the same kind of information in MT? To be defined



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




0
100
April 2004
Aug. 2004
Dec. 2004
April 2005
Aug. 2005
Dec. 2005
May 2006
200
300
RMS 4.81