

# The MAGNETOTELLURIC method for geothermal exploration

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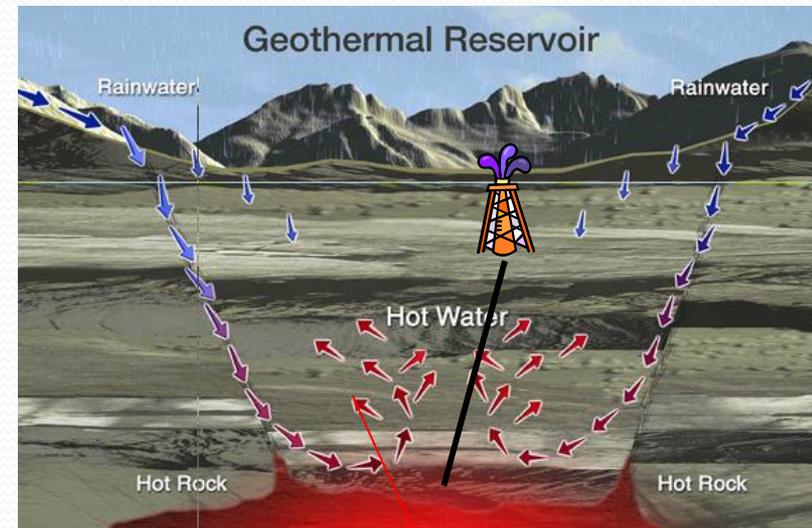


# Goals to be achieved by Geophysics

## 1: before and during production

- Improve methods of identifying prospective reservoirs without drilling
- Define boundaries (lateral and vertical)
- Improve methods to identify drilling targets

Main permeability is driven  
by fracture and faults.  
>30% of wells not economic



# Goals to be achieved by Geophysics

## 2: during and after production

- Continuously characterize the reservoir during energy extraction
- Follow the effect of production and fluid re-distribution, including the formation of steam or gas cap
- Characterize the rock fabric to define fluid flow paths within reservoir
- Track injected fluids
- Characterize formations during deep drilling and stimulation in order to predict reservoir performance/lifetime (effectiveness and sustainability)



# Why resistivity?

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



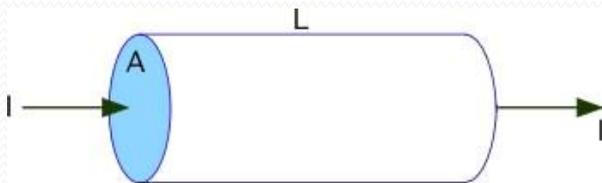
Electrical resistance = the resistance of a medium to electrical current transmission  
Ohm Law

$$R = \frac{V}{I}$$

Where V= potential difference (volts) and I= current flow (ampere)  
R is adimensional.

Current density :  $J = \frac{I}{A}$

Resistance =  $R \propto \frac{L}{A} = \frac{\rho L}{A}$



Resistivity, which is the resistance of a unit volume, has unit ohm-m

The conductivity is

And has unit Siemens/m

$$\sigma = 1/\rho$$



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

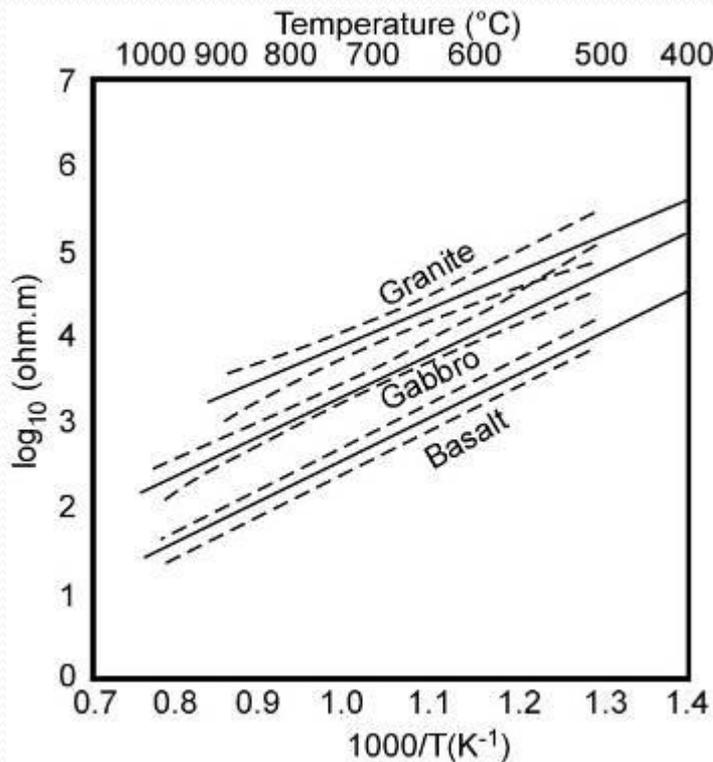
Main propagation mechanisms are:

|  |           |                            |
|--|-----------|----------------------------|
| Electronic ( $<10^{-8} \Omega\text{m}$ )                 | electrons | metals                     |
| Semi-conduction<br>$10^{-5} \div 10^{-3} \Omega\text{m}$ | electrons | Sulfurs<br>and ions        |
| Electrolitic   | ions      | brines, salty water, melts |



# Why resistivity?

## Temperature and Pressure



At normal temperature at the earth surface, silicate minerals have very high resistivity.

The higher the temperature, the lower the resistivity.

Approaching the melting point of a rock the resistivity becomes low enough to become comparable with resistivities in water-saturated rocks



From Karya and Shankland, 1983

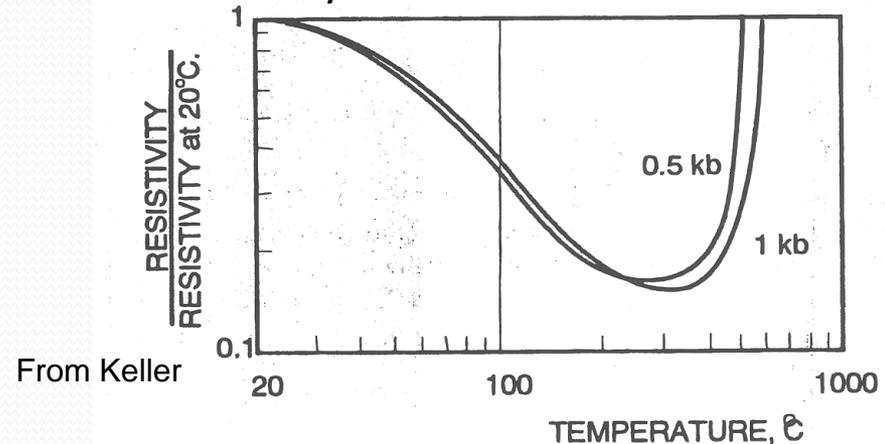
# Why resistivity?

## Temperature and Pressure

When rocks contain fluids, electrical conduction takes place mainly by passage of current through the fluid in the pores, since almost all rock-forming minerals are almost insulators at low-medium temperatures.

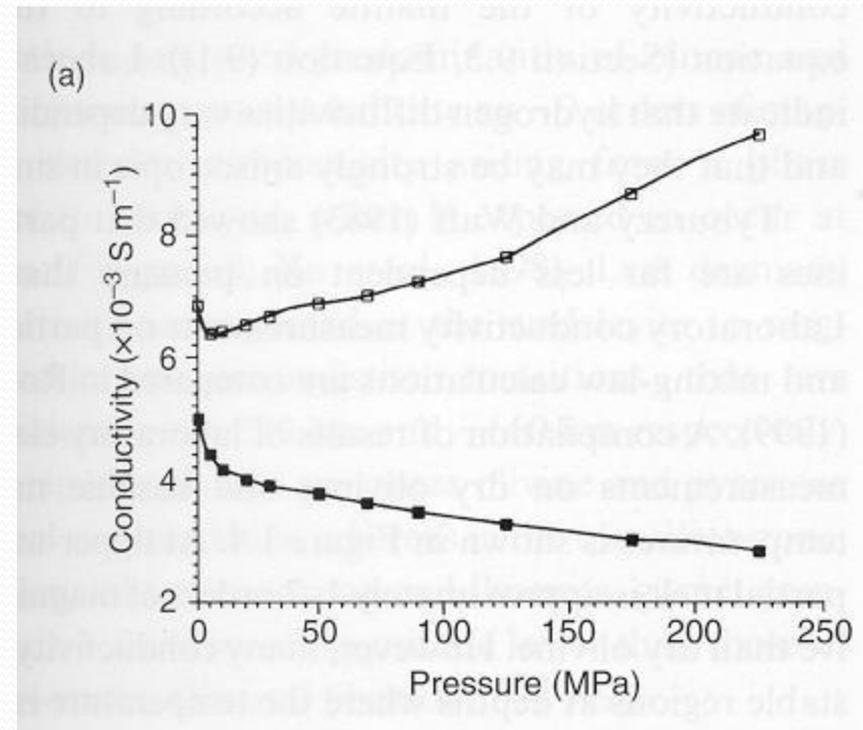
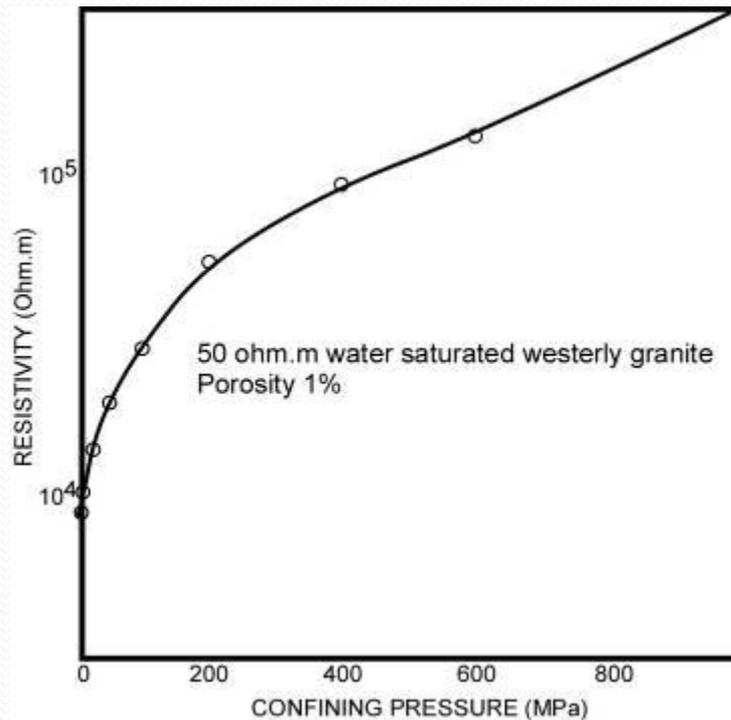
The conductivities of both the electrolytes and the rock matrix are temperature dependent in a manner that causes a large reduction of the bulk resistivity with increasing temperature.

The maximum enhancement in conductivity is approximately sevenfold between 350°C and 20°C for most electrolytes



# Why resistivity?

## Temperature and Pressure

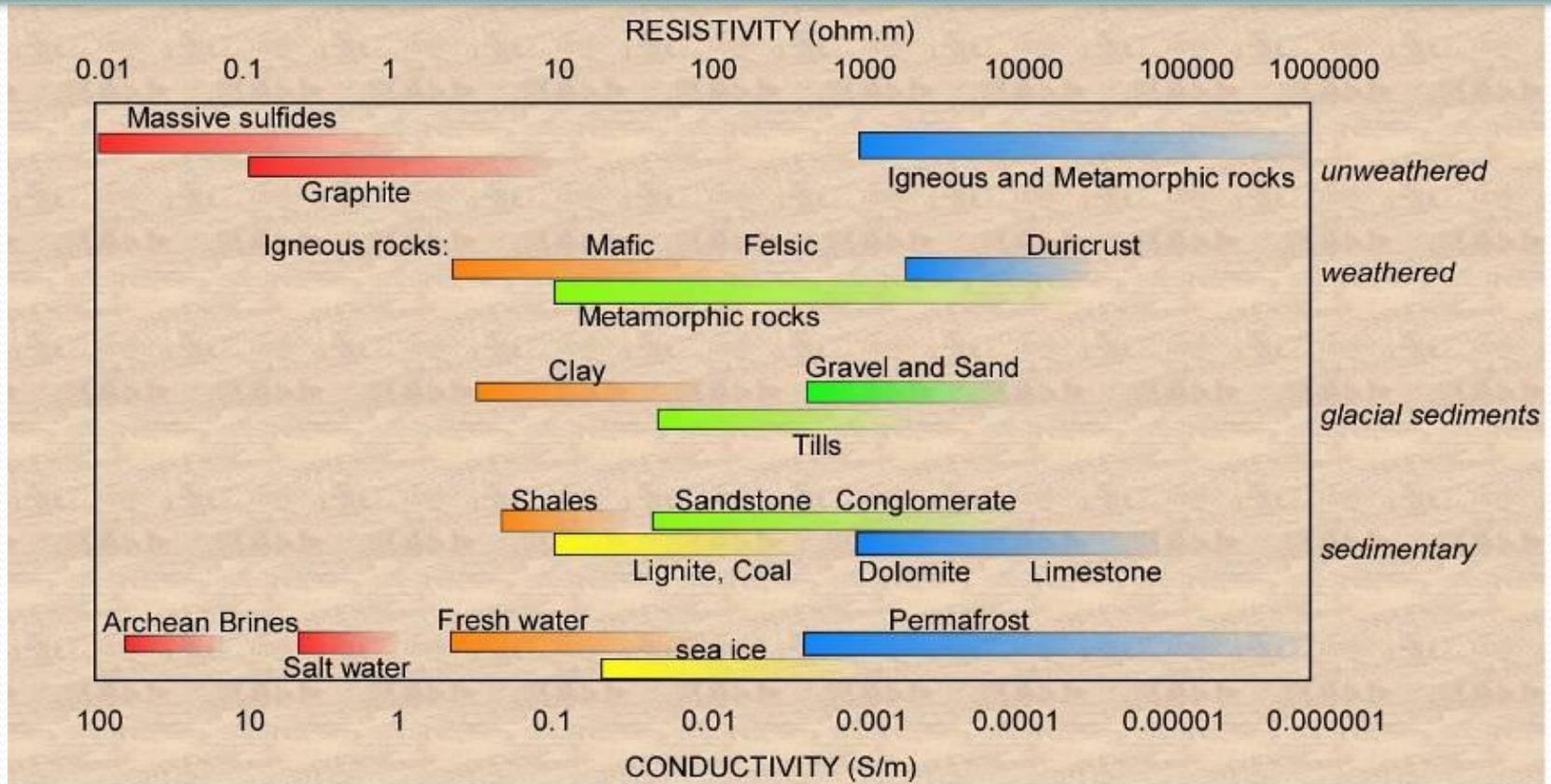


Pressure influence resistivity mainly in connection to fluids and, to a lesser extent, to reconnection of solid phases.



# Why resistivity?

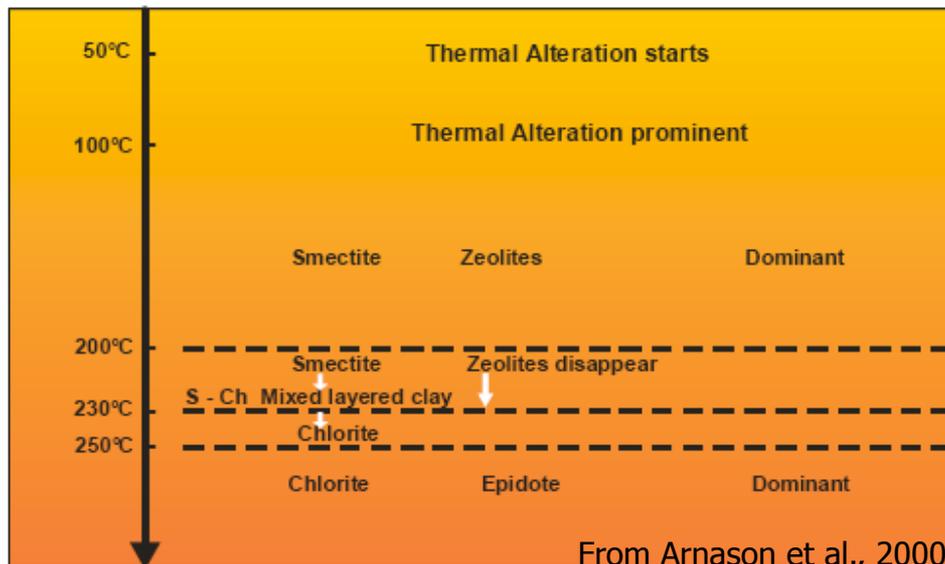
## Lithology



# Why resistivity?

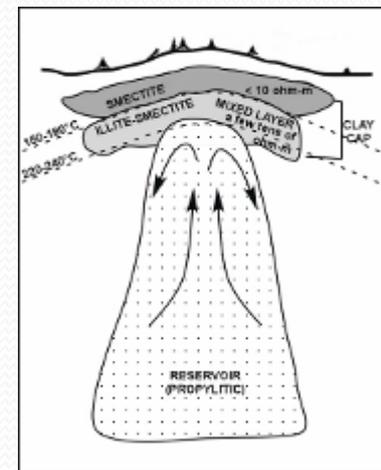
## Clay minerals and surface conduction

Alteration mineralogy at different temperatures



Clays not only decrease the resistivity by themselves, but also increase the surface effect (frequency-dependent IP)

The resulting resistivity is also related to the presence of clay minerals, and can be reduced considerably when the clay minerals are broadly distributed.



From Pellerin et al., 1996

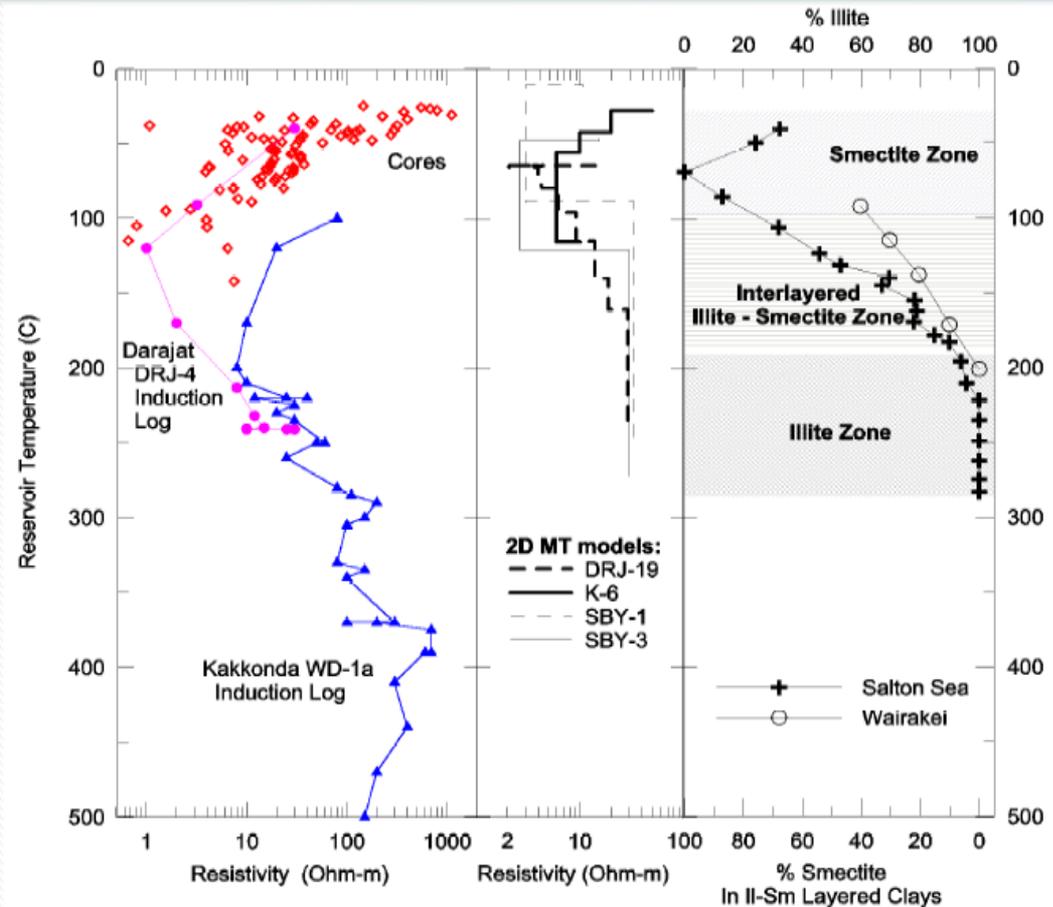


# Why resistivity?

## Clay minerals and surface conduction

The effect of clay over the resistivity is important in geothermal areas since clayey alteration minerals are very frequent, particularly in volcanic rocks.

Comparison of the resistivity structure with geological data in volcanic geothermal fields has shown a good correlation with alteration mineralogy.

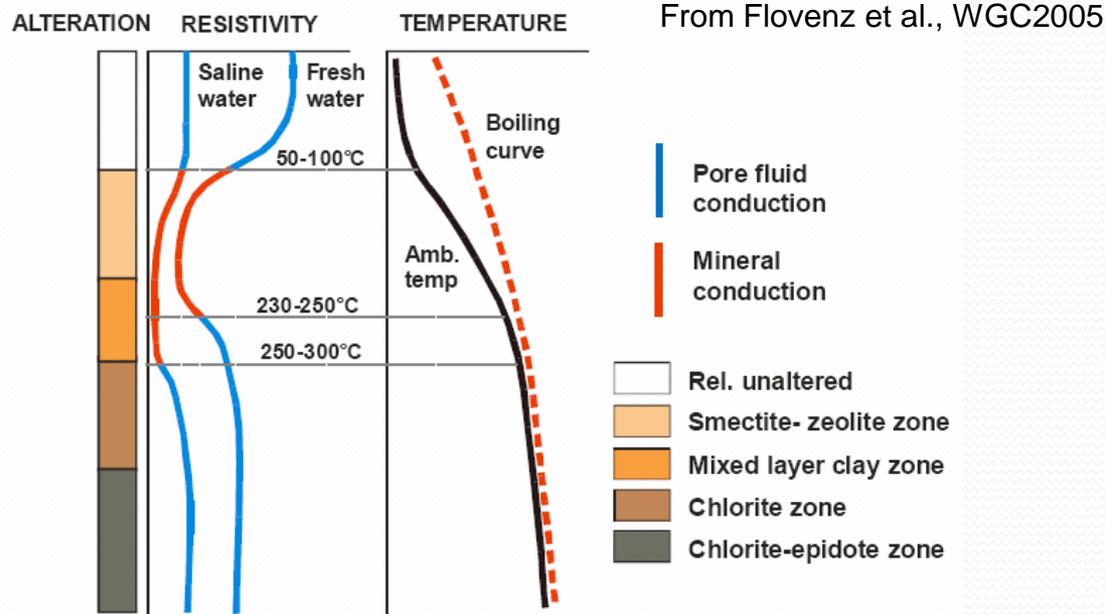


# Why resistivity?

## Clay minerals and surface conduction

Resistivity should be always considered with care. Experience has shown that the apparent one-to-one correlation between low resistivity and the presence of fluids is not correct, since alteration minerals produce comparable, and often higher reduction of resistivity with respect to fluid flow.

### Resistivity Structure summarised



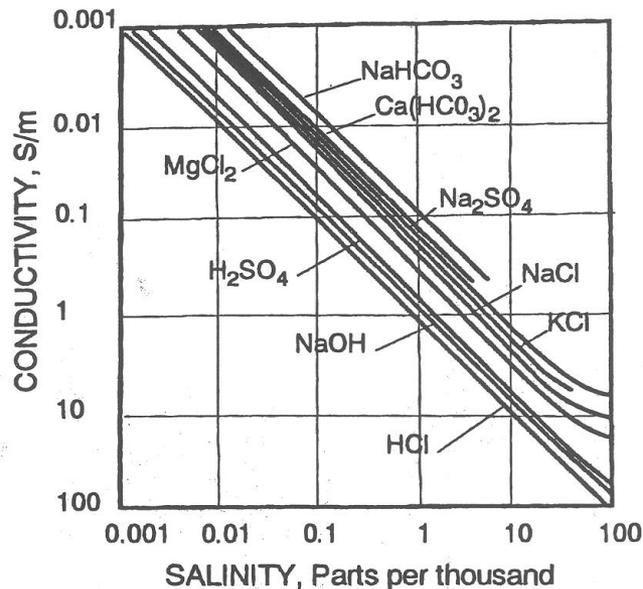
Moreover, although the hydrothermal systems in volcanic rocks have an associated low-resistivity signature, the converse is not true.



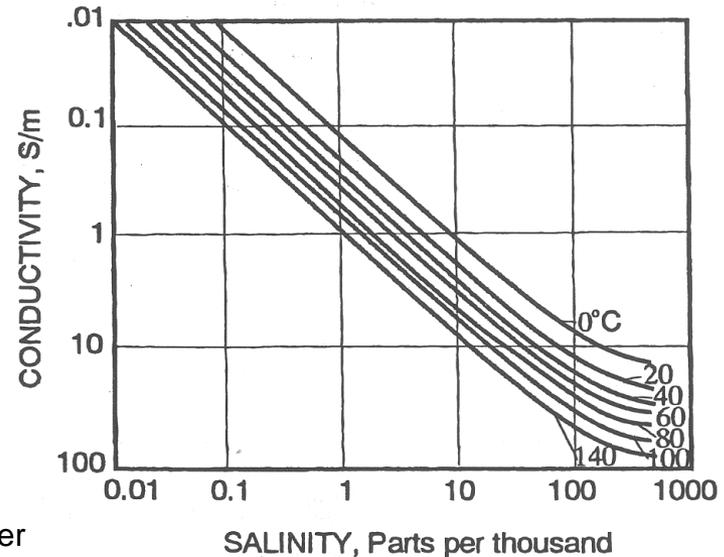
# Why resistivity?

## Salinity

Geothermal waters have high concentrations of dissolved salts which provide conducting electrolytes within a rock matrix



From Keller

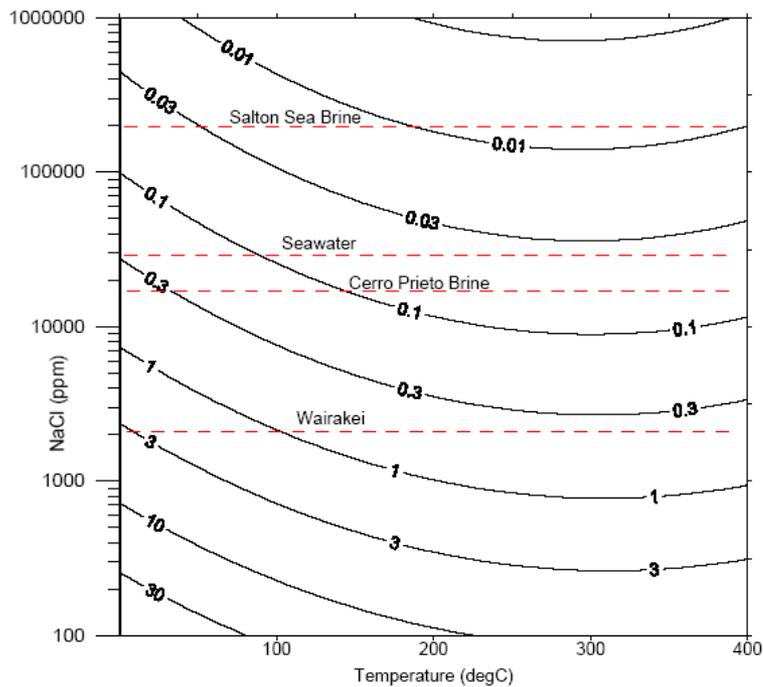


As a result, it is not unusual to see an increase in conductivity by an order of magnitude or more in a geothermal reservoir compared with rocks at normal temperatures removed from the reservoir.



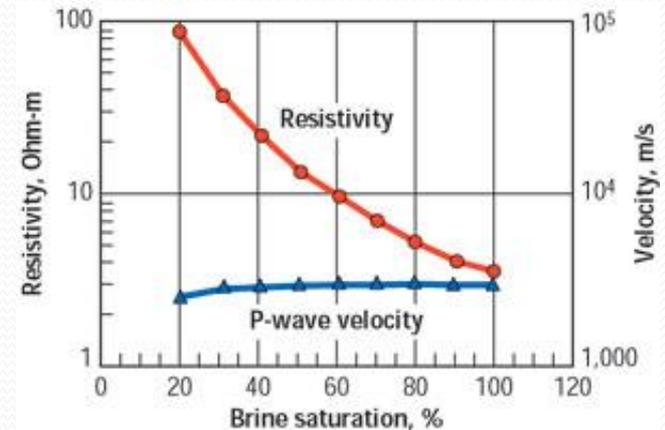
# Why resistivity?

## Salinity



Variation of resistivity (in ohm-m) of NaCl solutions. The salinity of several known geothermal field is also shown.

From Ussher et al., WGC2000.

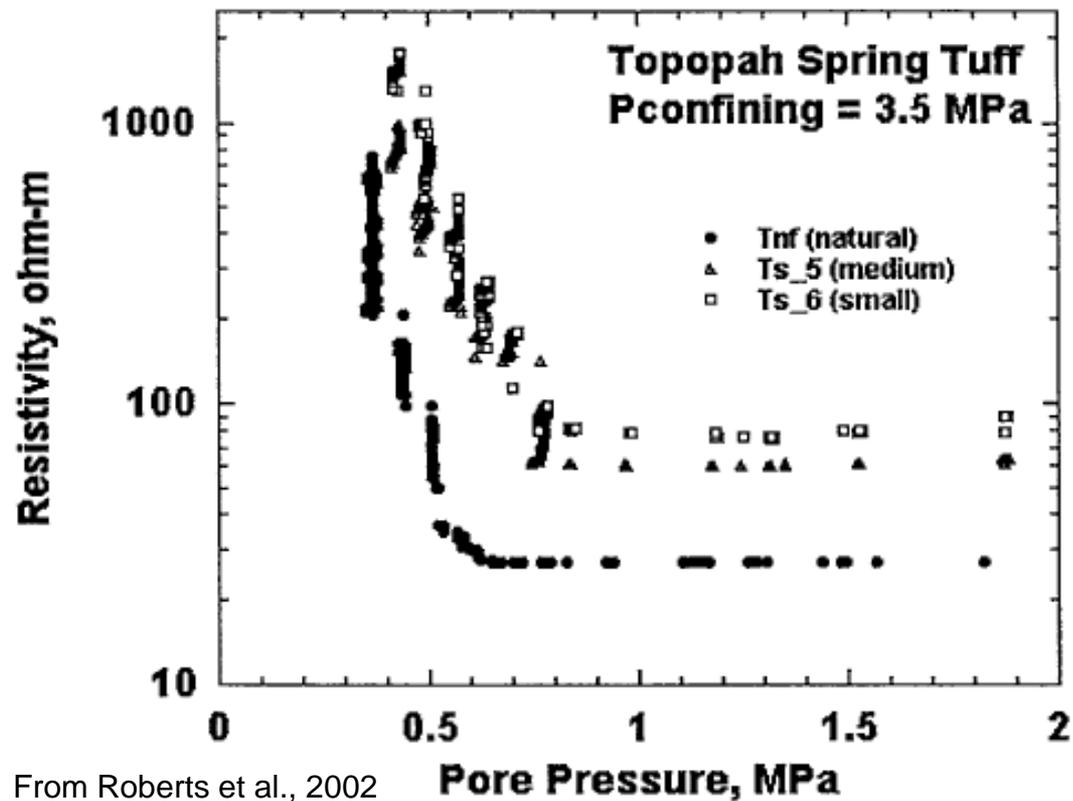


From Wright, Ziolkowski & Hobbs TLE 9/02; after Wilt & Alumbaugh, 1998



# Why resistivity?

Fluid phase

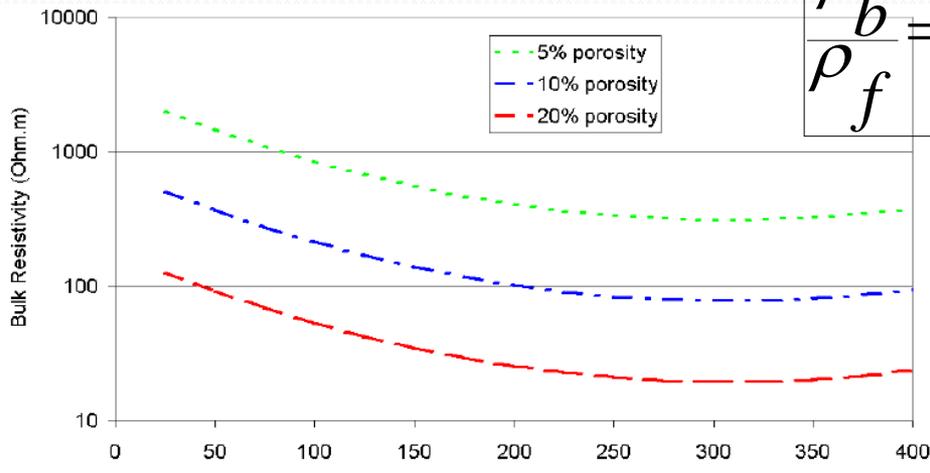


# Why resistivity?

## Hydraulic properties

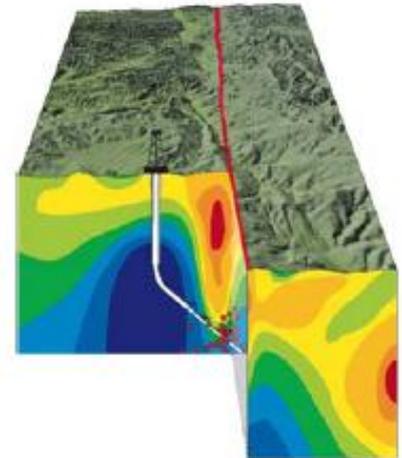
In most rocks there is an empirical relationship, established by *Archie (1942)*, between the ratio of the bulk rock resistivity,  $\rho_b$ , to the pore fluid resistivity,  $\rho_f$ , called the formation factor,  $F$ , and the volume fraction **porosity**,  $\phi$ . The relationship, now called Archie's law, is:

$$\frac{\rho_b}{\rho_f} = F = a\phi^{-m}$$



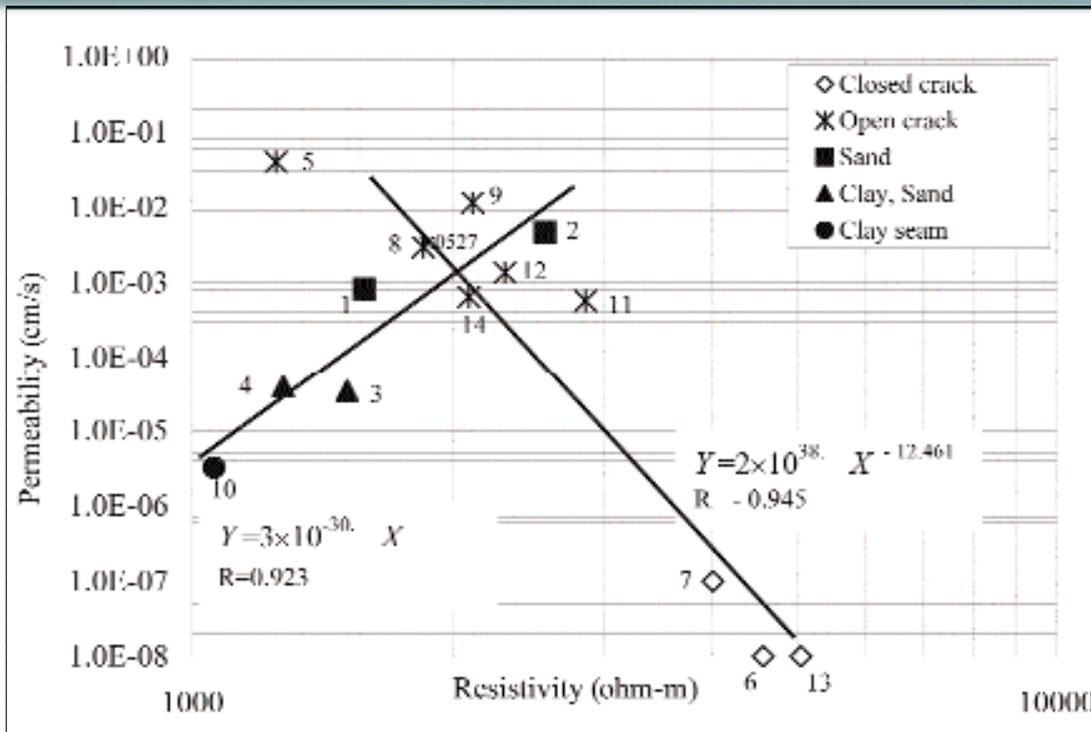
Bulk resistivity for rocks saturated with 1000 ppm NaCl solutions, using Archie's law.

From Ussher et al., WGC2000.



# Why resistivity?

## Hydraulic properties



Some empirical equations relating resistivity and permeability were obtained on the electrical logging and in-situ permeability data from boreholes.

Relationship between resistivity and permeability measured in three boreholes in Mt. Tsukuba in Ibaraki Prefecture, central Japan (*Sudo et al., 2004*).



# Why resistivity?

## Hydraulic properties

Frequency dependent complex property

$$\sigma^* = \frac{1}{\rho^*} = i\omega\epsilon^*$$

$$\sigma_{eff}(\omega) = \sigma'(\omega) + i\sigma''(\omega)$$

- Real component  $\sigma'$  ohmic conduction (energy loose) measured with EM and DC methods
- Imaginary component  $\sigma''$  polarization (energy storage) measured with IP, SP methods



# Why resistivity?

## Hydraulic properties

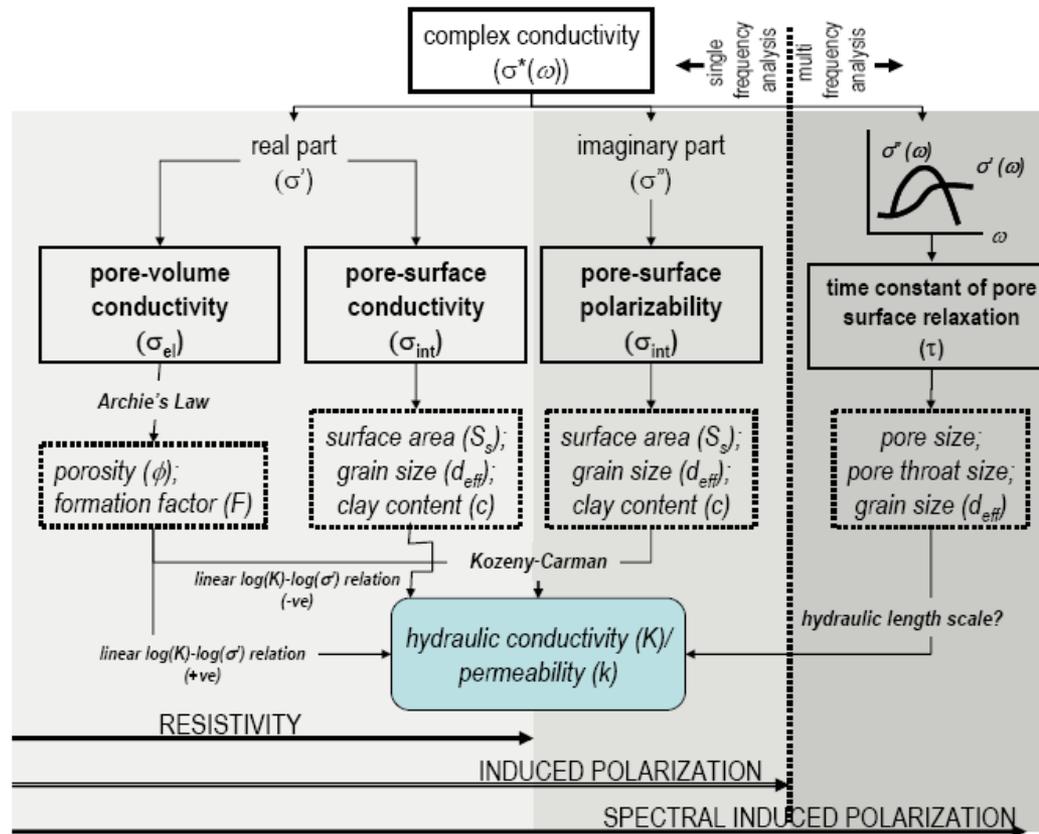


Figure 1. Flowchart summarizing efforts to predict hydraulic conductivity ( $K$ ) from petrophysical relationships established for electrical measurements.

From Slater 2006



# EM Geophysical methods

## ***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 reservoir rocks, and several tens of kilometres when seeking the thermally excited conductive zone associated with the source of a geothermal system.

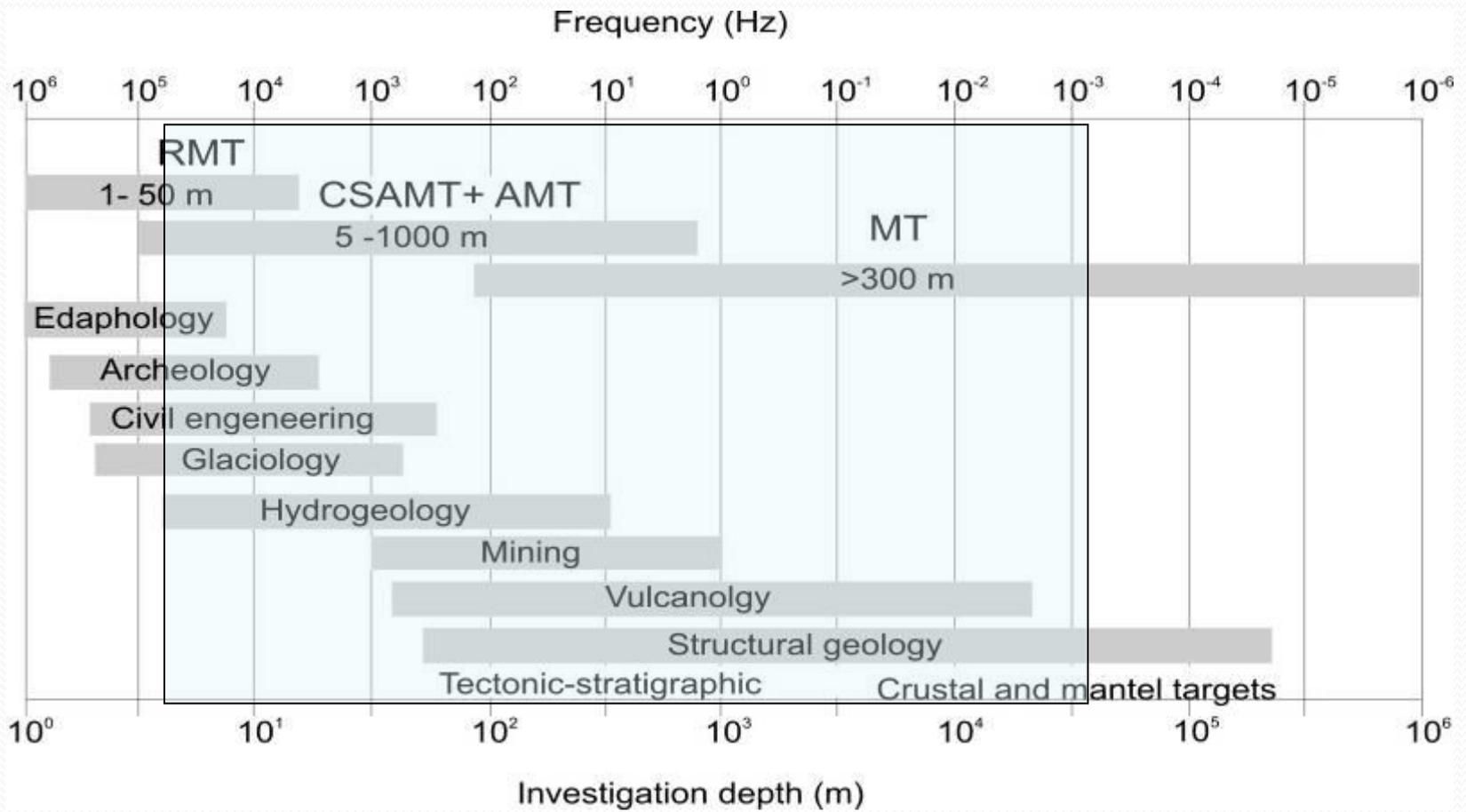


Inductive methods usually provide information on conductivity-thickness products of conductive layers, whereas they usually provide only thickness information on resistive layers.

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

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





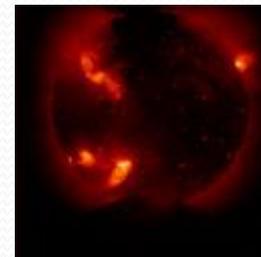
**Magnetotellurics** (MT for short) is a technique which utilizes the earth's naturally occurring electromagnetic field to image the subsurface's electrical resistivity structure.

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

High frequency signals originate in lightning activity



Intermediate frequency signals come from ionospheric resonances

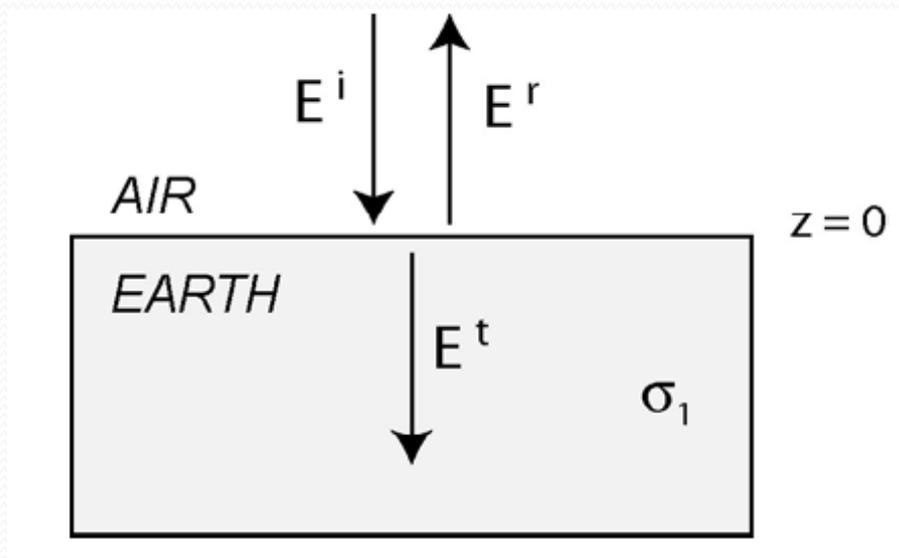


Low frequency signals are generated by sun-spots

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

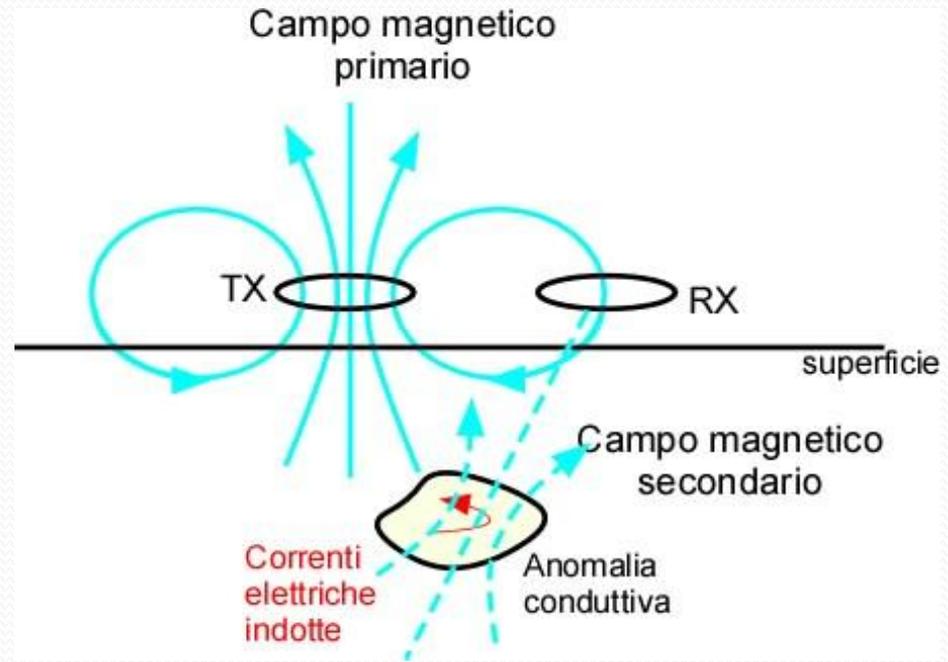
$E^t$  diffuses in the Earth, since Earth's conductivity is large with respect to air's conductivity.

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**

$$\left| Z_{xy}(\omega) \right|^2 = \left| \frac{E_x(\omega)}{H_y(\omega)} \right|^2 = \frac{\omega \mu_0}{\sigma_1} \quad \text{E.g., in a uniform earth}$$

where  $\rho = 1/\sigma$

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

or, solving for  $\rho$ ,

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

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**,  $\Phi$ , 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

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

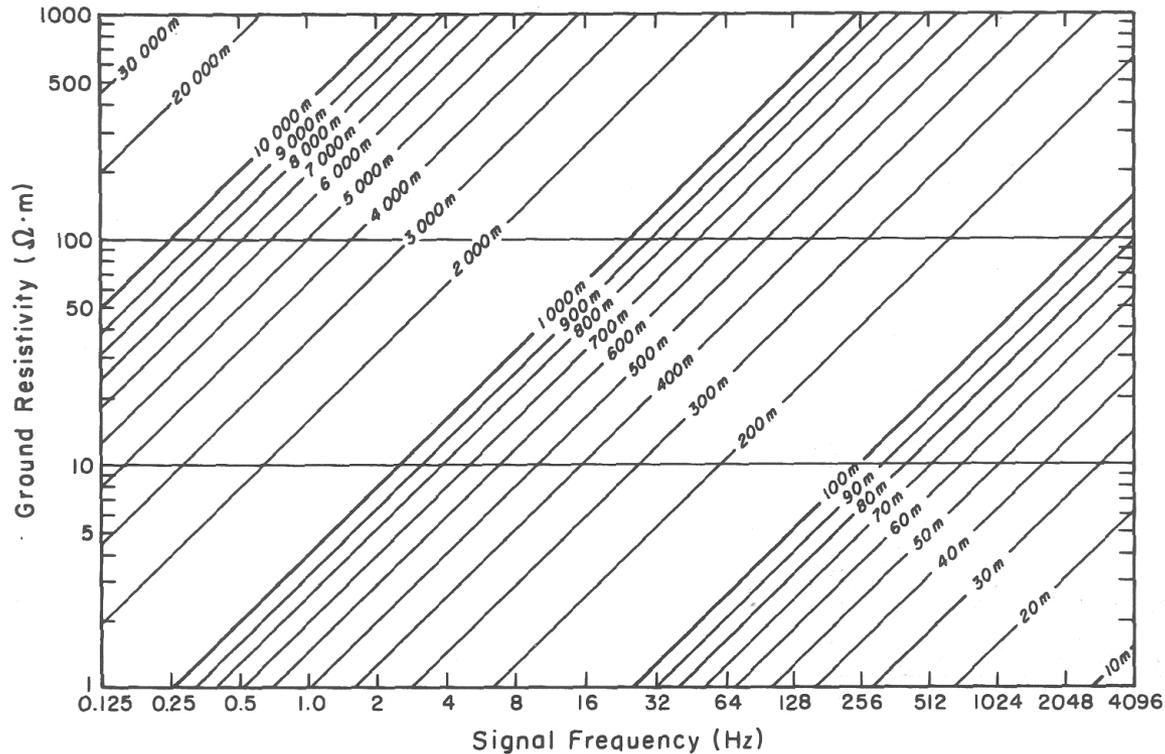
where  $A_0$  is the surface value. That is, the fields vary as the product of four terms:

- $e^{i\omega t}$ , 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  $\delta$  is called the skin depth.



## Plot of skin depth versus frequency and resistivity



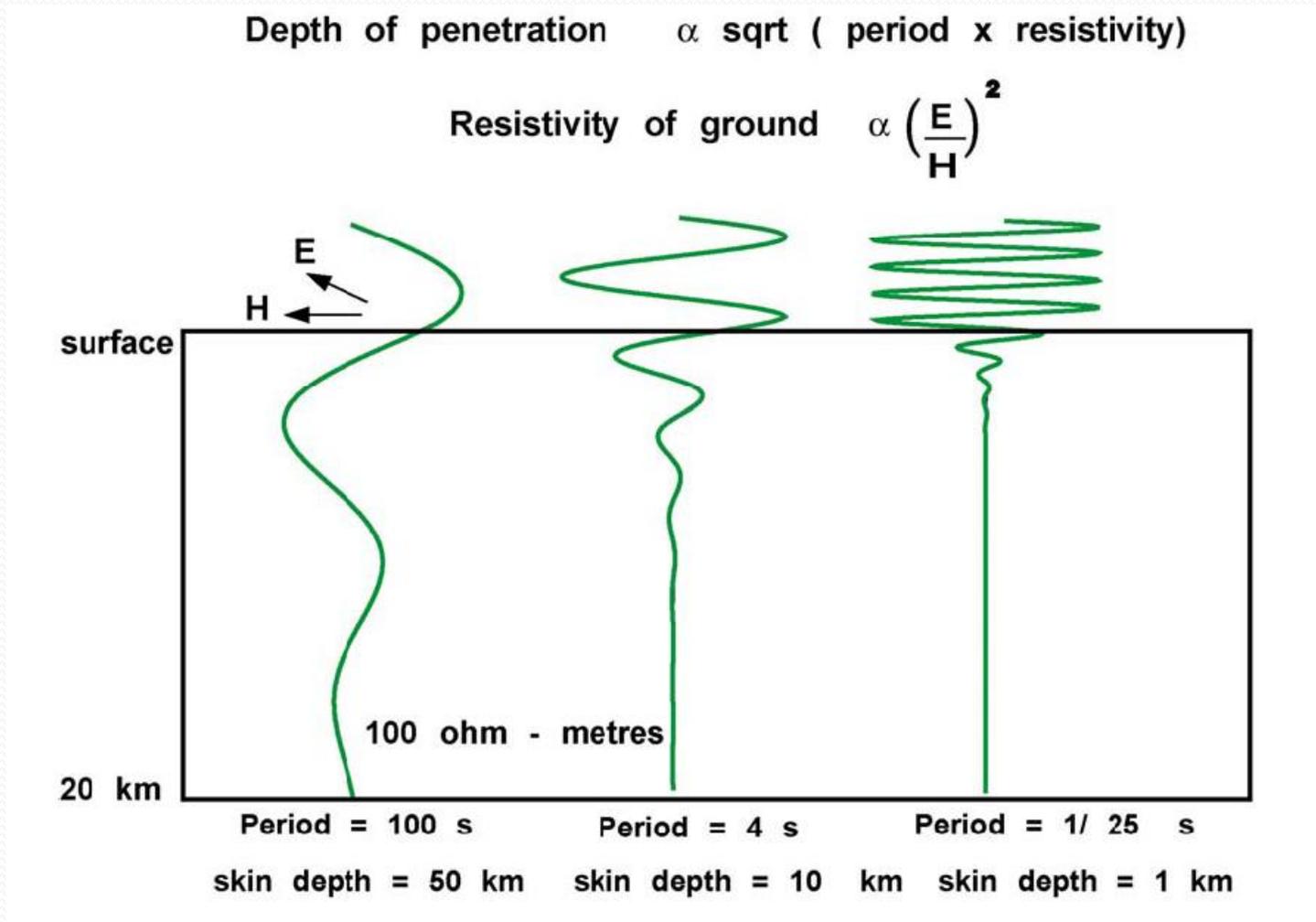
A useful approximation is given by

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

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

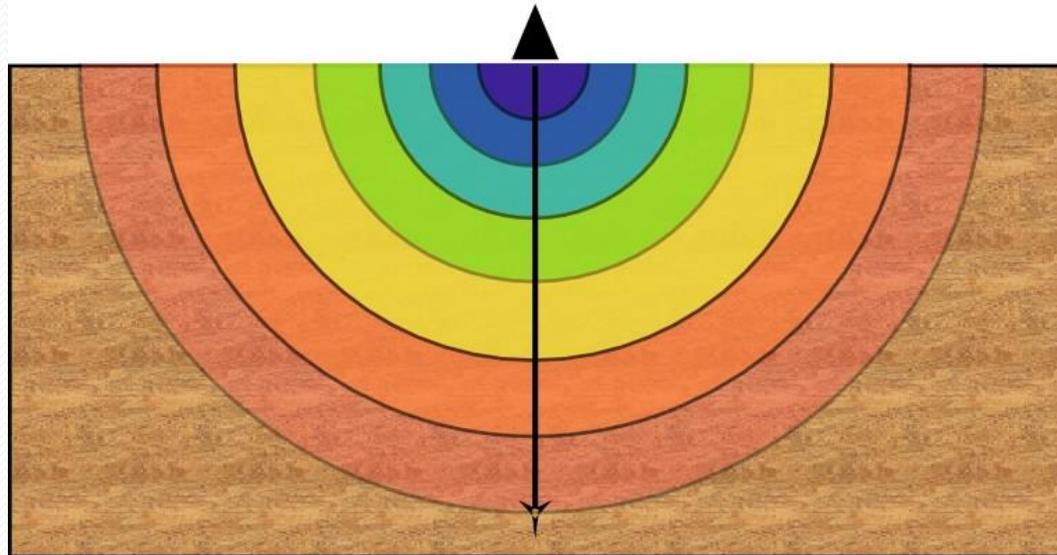
In the field, we measure the electric and magnetic fields in the frequency band we are interested in

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 ( $\delta$ ), this measurement samples a **hemisphere around the observation site, radius  $\delta$ .**

Data derive not only from the geometrical-physical features on the vertical of the recording site, but depends also on the lateral 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:

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

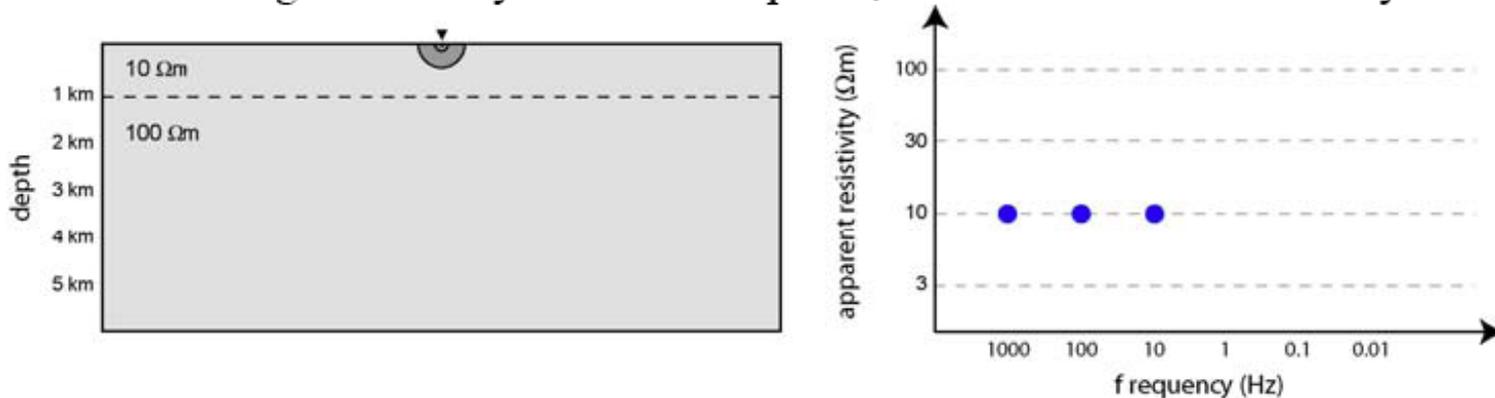
one for up-going and the other for down-going energy.

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



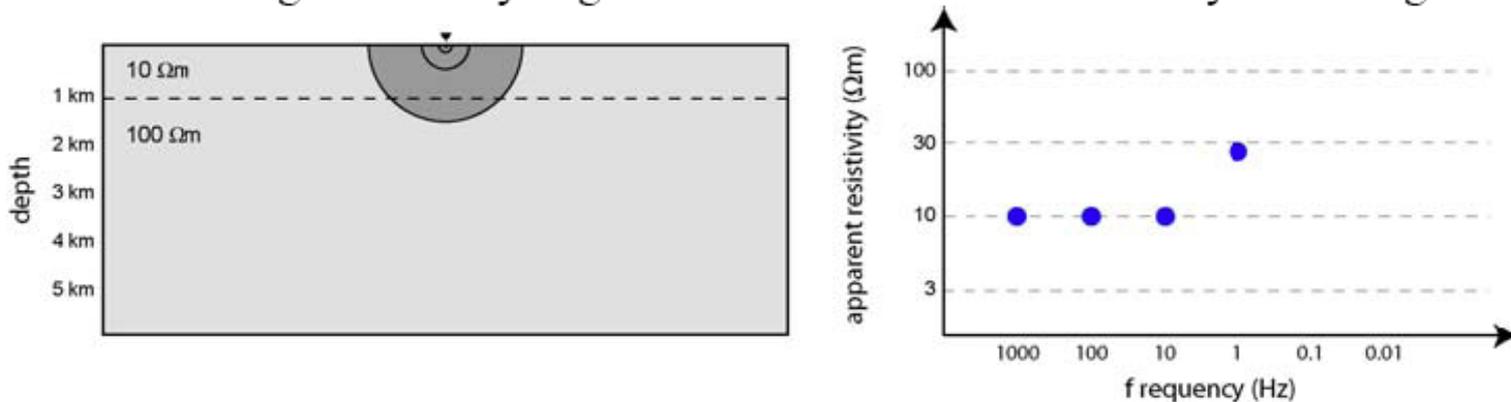
(1)  $f = 10 \text{ Hz}$   $\delta = 500 \text{ m}$  in upper layer

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



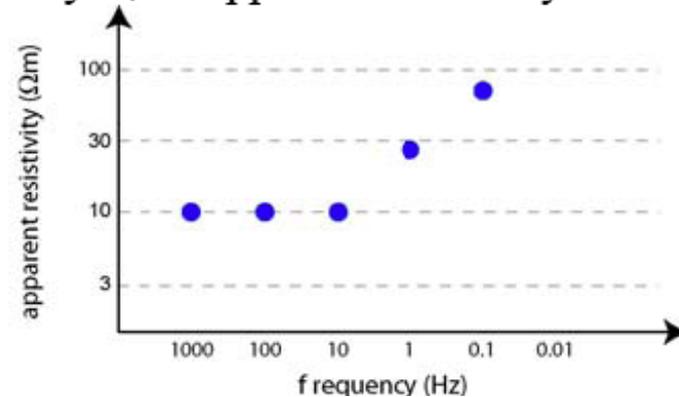
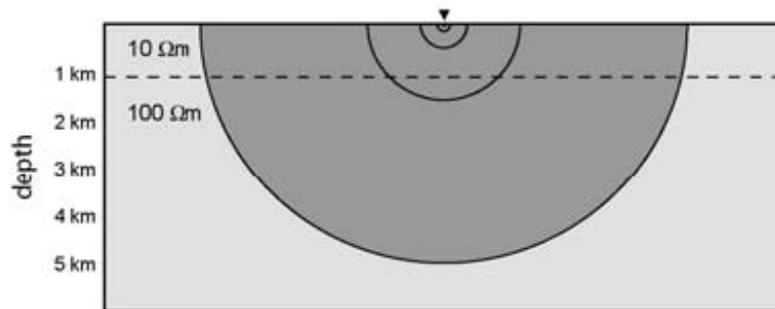
(2)  $f = 1 \text{ Hz}$   $\delta = 1580 \text{ m}$  in upper layer

- The EM signals are now just sampling the lower layer,
- Average resistivity begins to increase as the  $100 \text{ } \Omega\text{m}$  layer is being sampled.



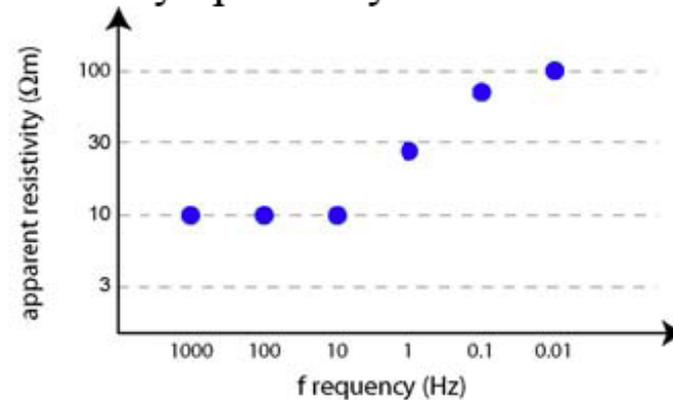
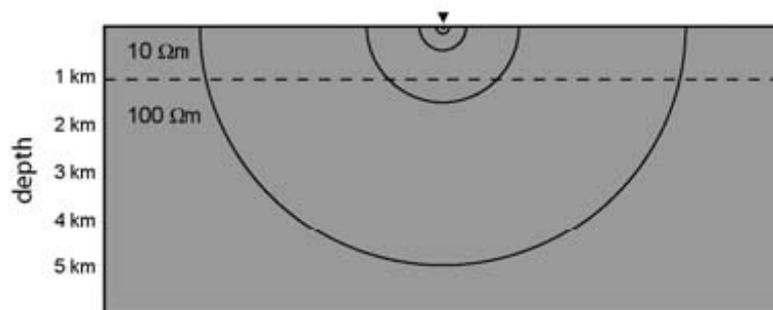
(3)  $f = 0.1 \text{ Hz}$        $\delta = 5000 \text{ m}$  in upper layer

- Hemisphere dominated by lower layer, so apparent resistivity close to  $100 \text{ } \Omega\text{m}$



(4)  $f = 0.01 \text{ Hz}$        $\delta = 15800 \text{ m}$  in upper layer

- Apparent resistivity approaches  $100 \text{ } \Omega\text{m}$  asymptotically

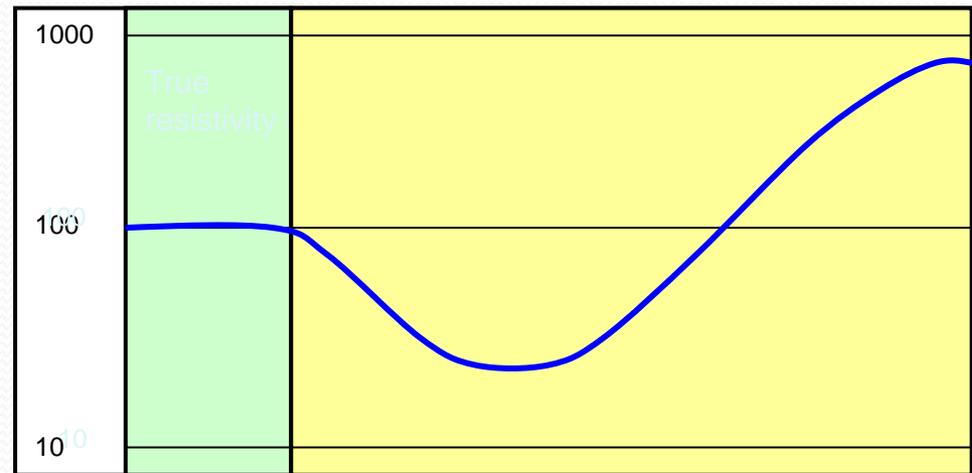


We can still get  $E_x$  and  $H_y$  (or  $E_y$  and  $H_x$ ) on the surface and do the resistivity calculation, but the result depends on frequency and is now an **apparent resistivity**  $\rho_a(f)$ .

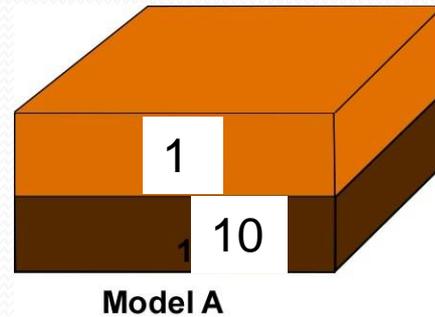
Apparent resistivity is defined as the resistivity of the homogeneous earth which would produce the measured response at each frequency



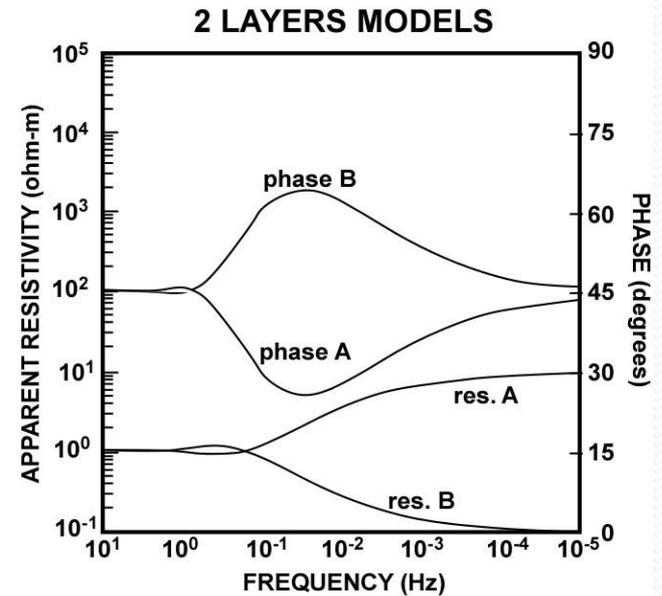
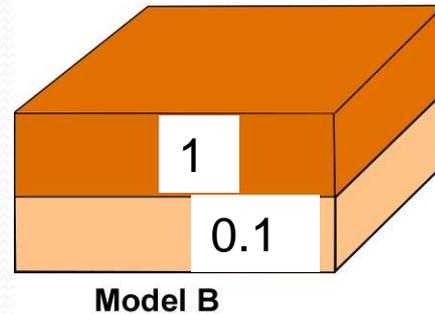
apparent resistivity  $\rho$  ( $\Omega\text{m}$ )



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



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



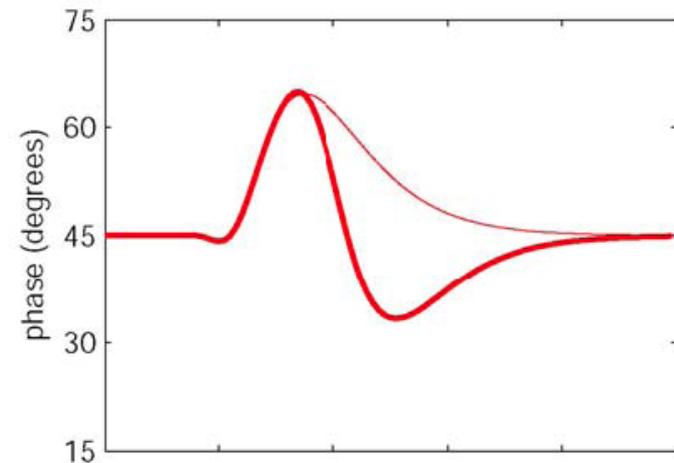
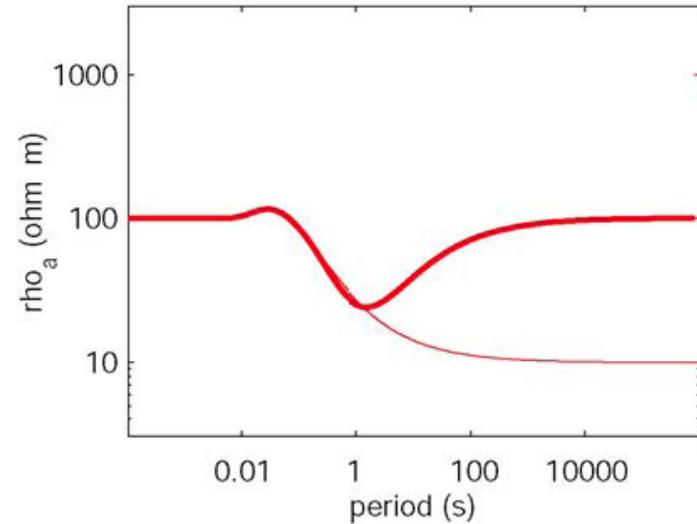
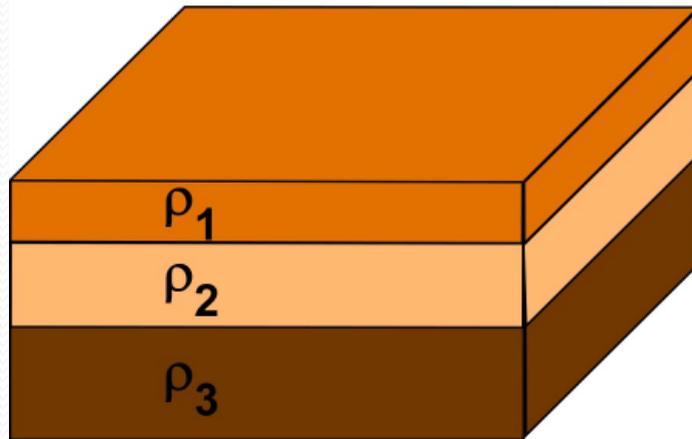
At high frequencies, since the skin depth is much lower than the thickness of the shallow layer, Alle frequenze più alte,  $\rho_a = \rho_1$  in both models.

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

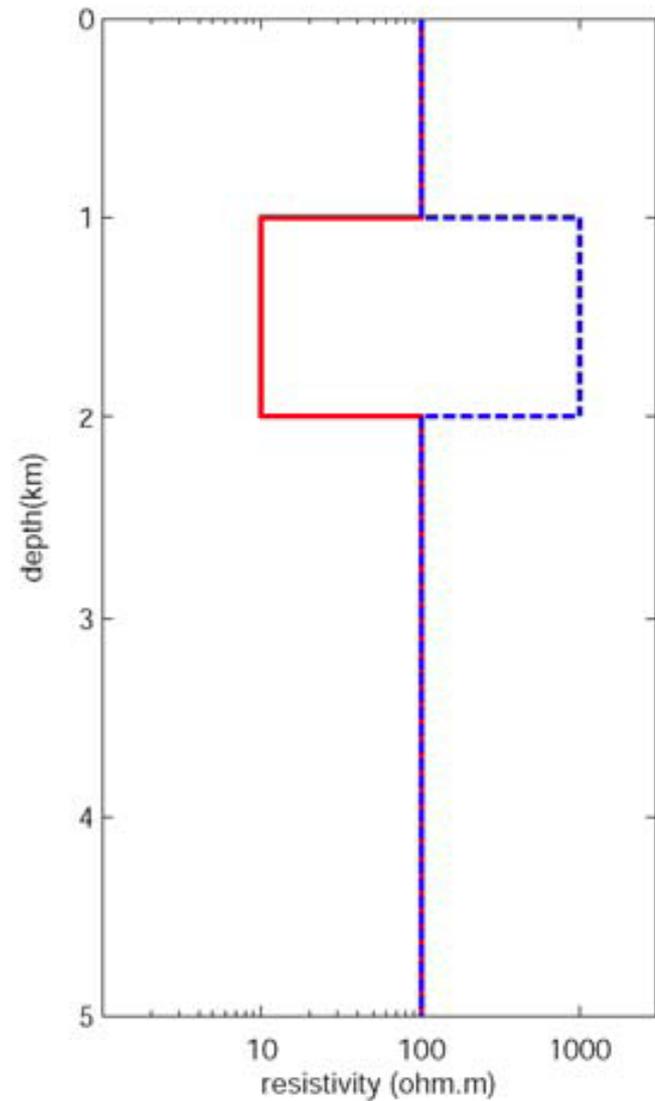
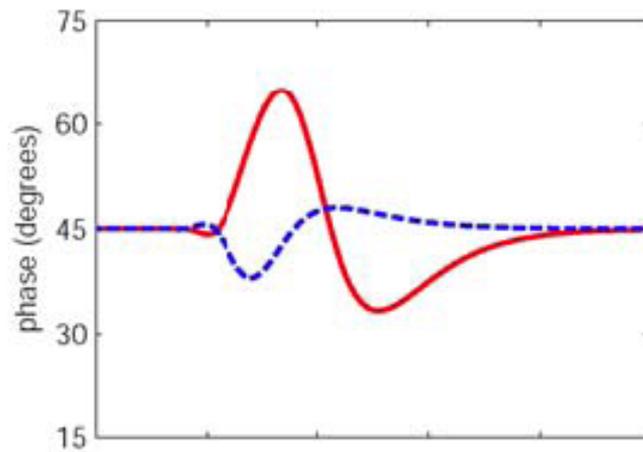
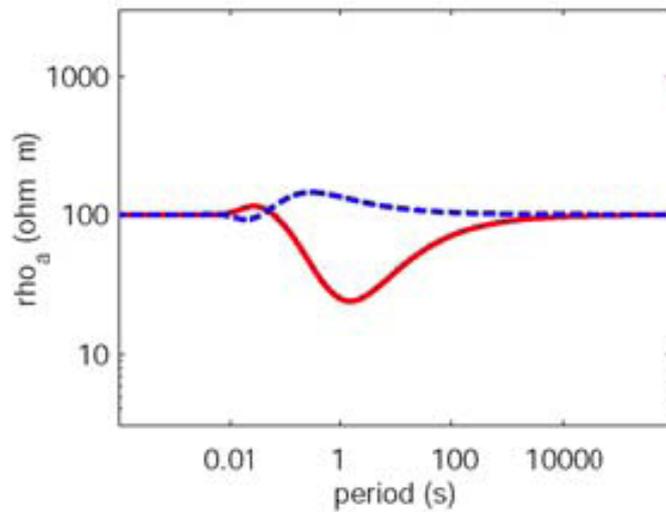


In a three layers model,  $\rho_a$  is equal to  $\rho_1$  at high frequency and goes exponentially to  $\rho_3$  at low frequency. In the middle it approaches  $\rho_2$ .

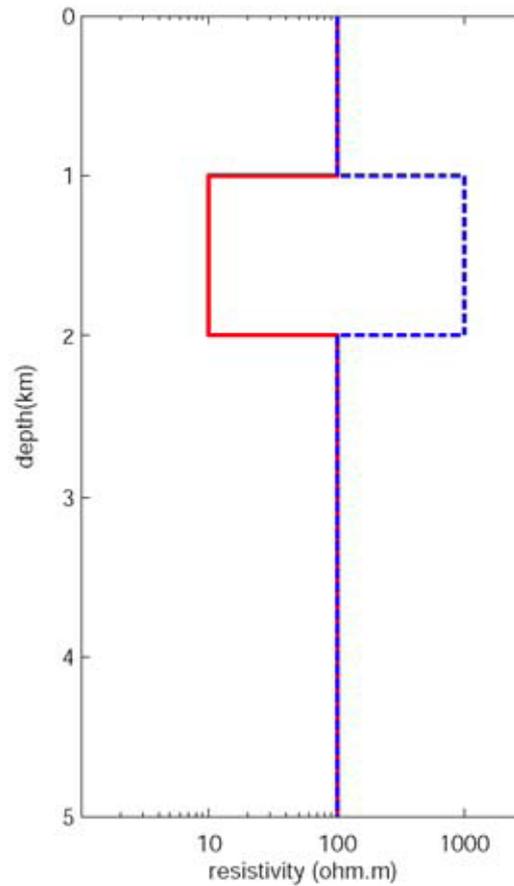
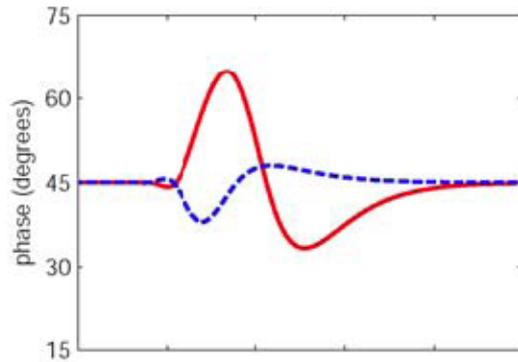
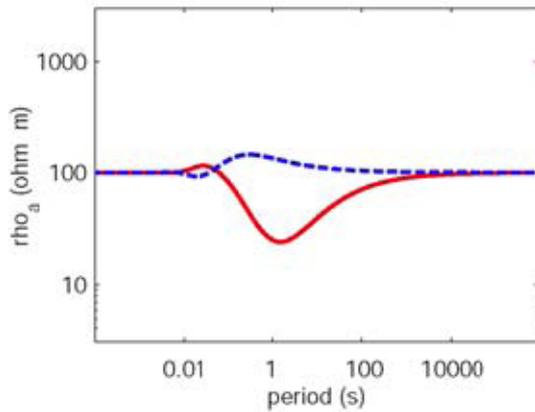
The shape, and therefore our ability to detect differences, depends very much on the relative resistivity and thickness of the layers, on the measured frequency and data quality.



# Intermediate conductive or resistive layer



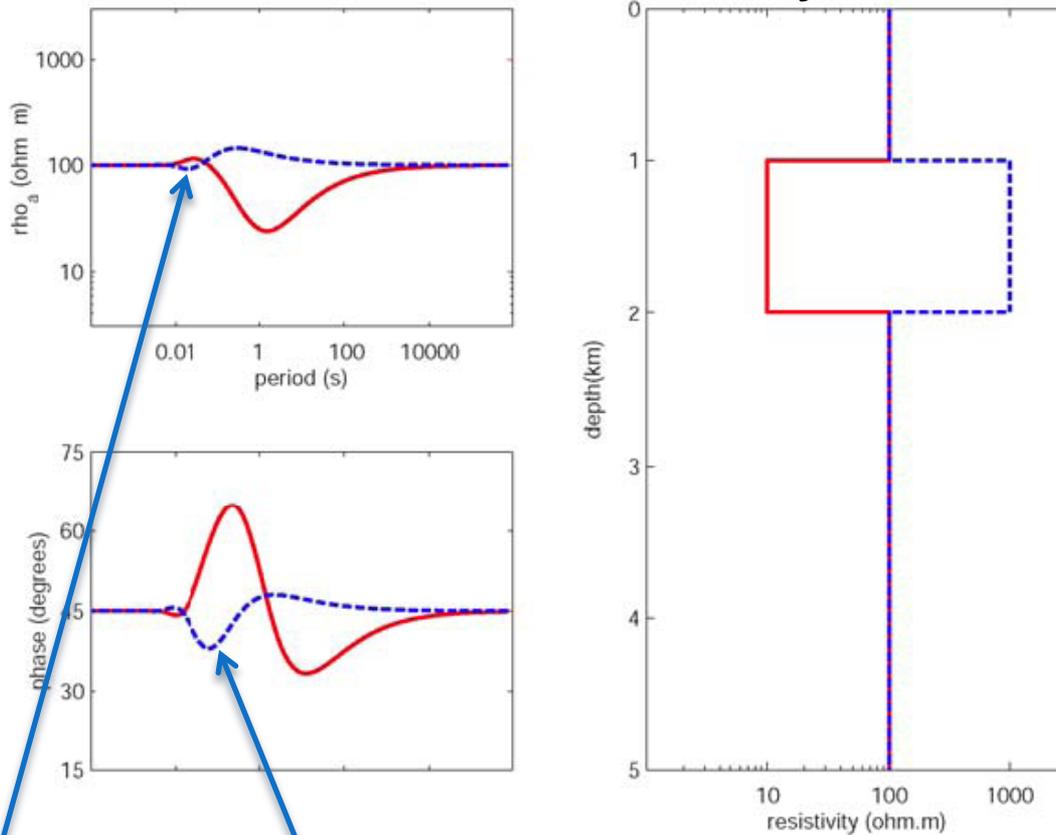
## Intermediate conductive or resistive layer



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



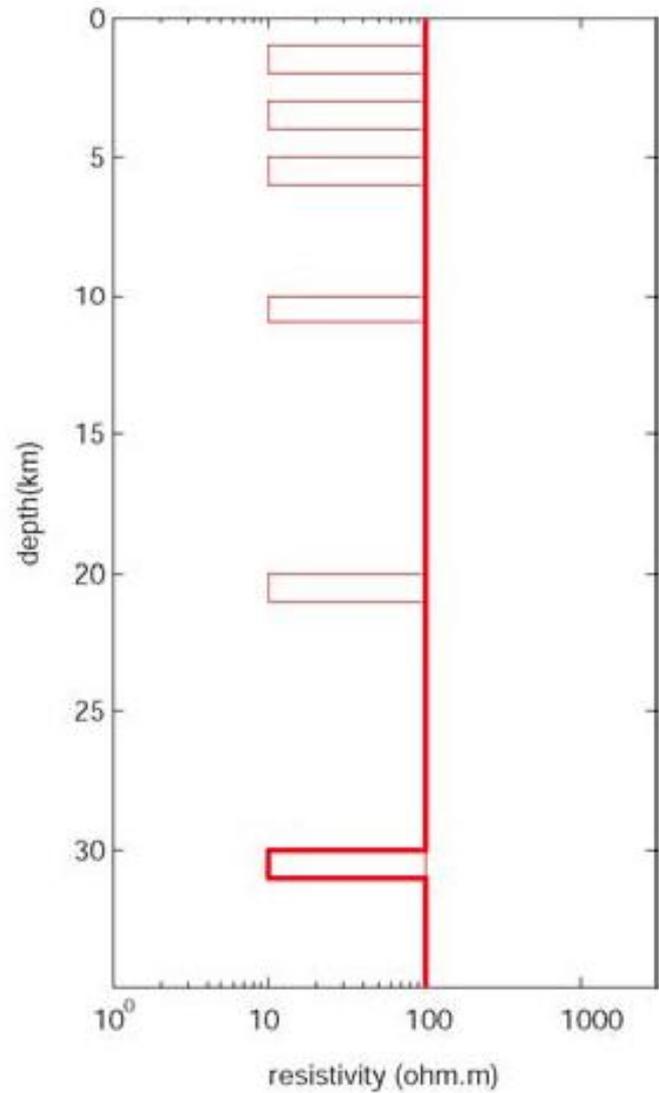
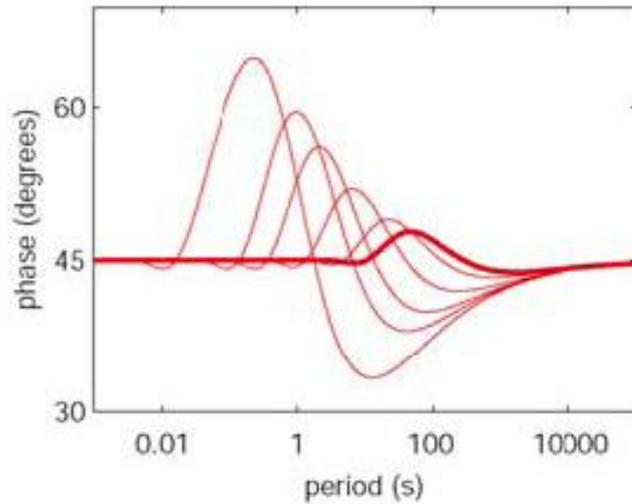
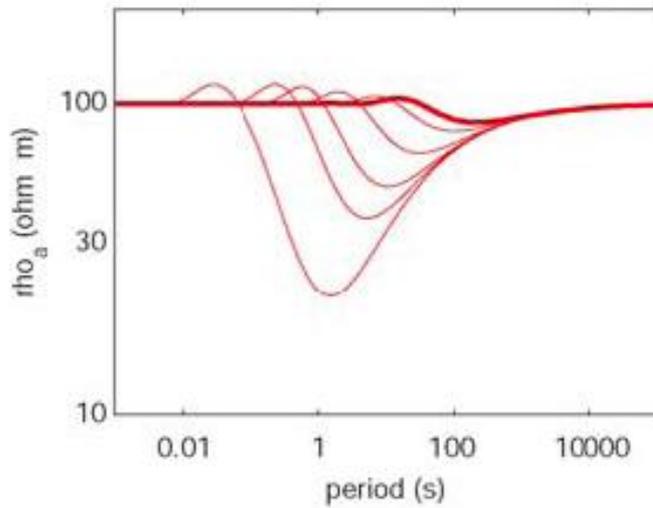
## Intermediate conductive or resistive layer



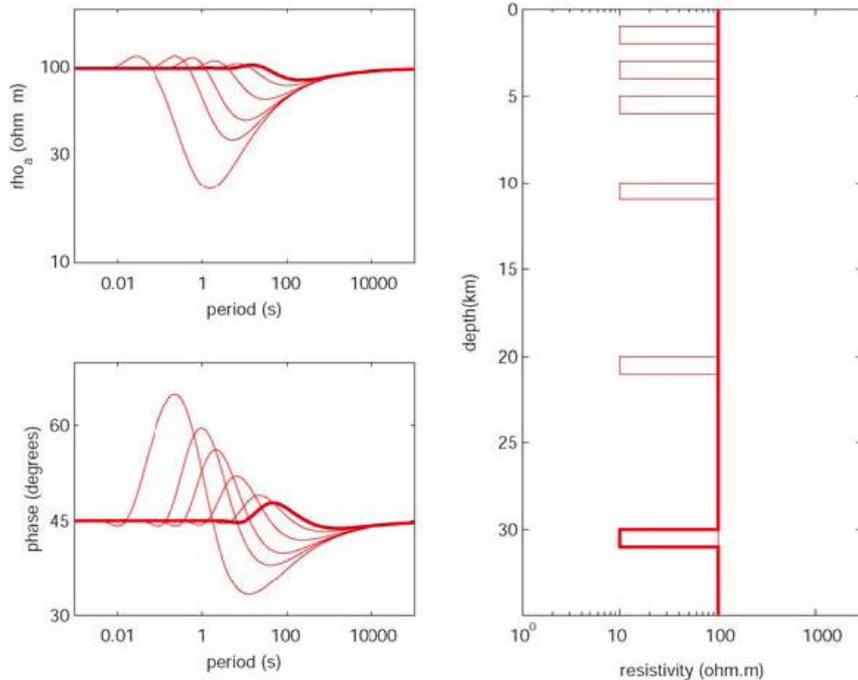
Look also at the resonance phenomena both on apparent resistivity and phase, when skin depth is close to the the first layer thickness (it is seldom observed due to data noise)



# Conductive layer at variable depth



## Conductive layer at variable depth

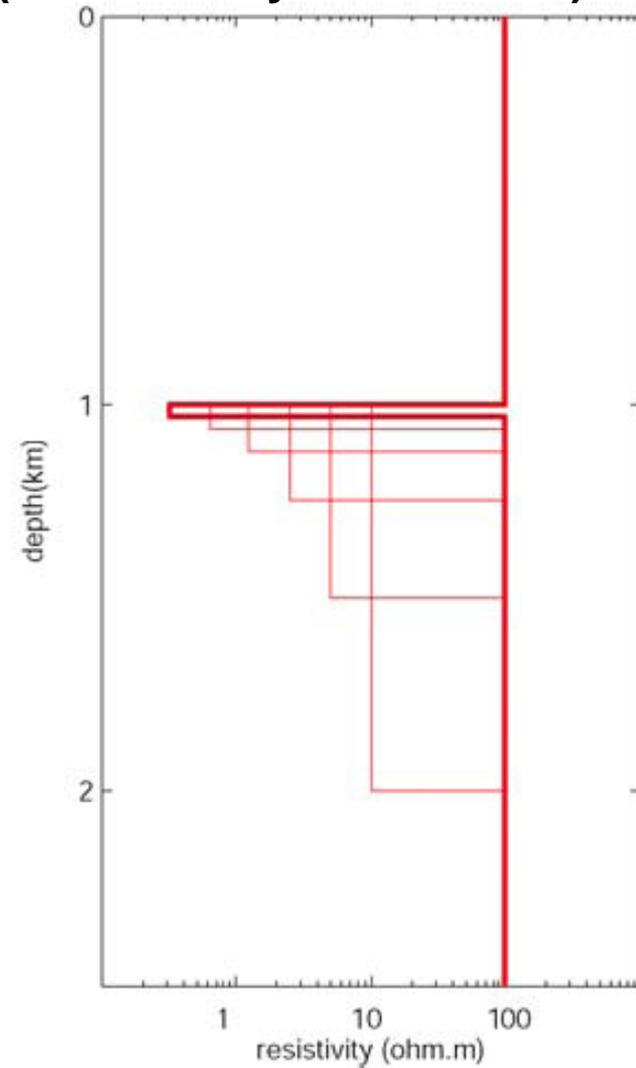
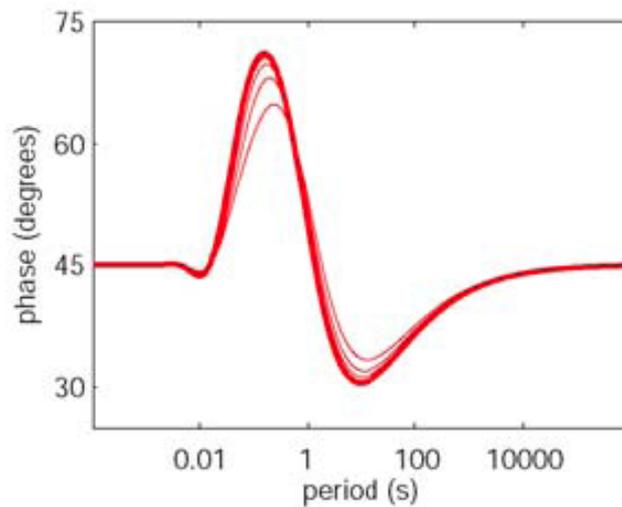
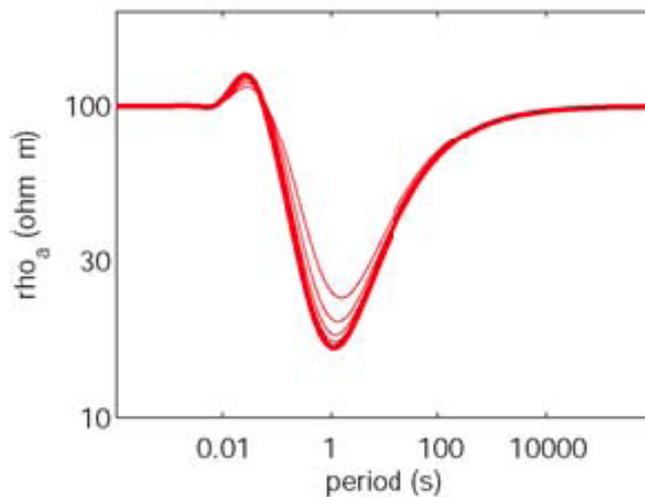


- As foreseen by skin depth equation, increasing depth the period at which we observe the effect of the layer increases (frequency decreases)
- The effect depends on the depth, since it is an apparent resistivity (average from surface down to the layer)
- MT can detect the top of the layer (10% of error)

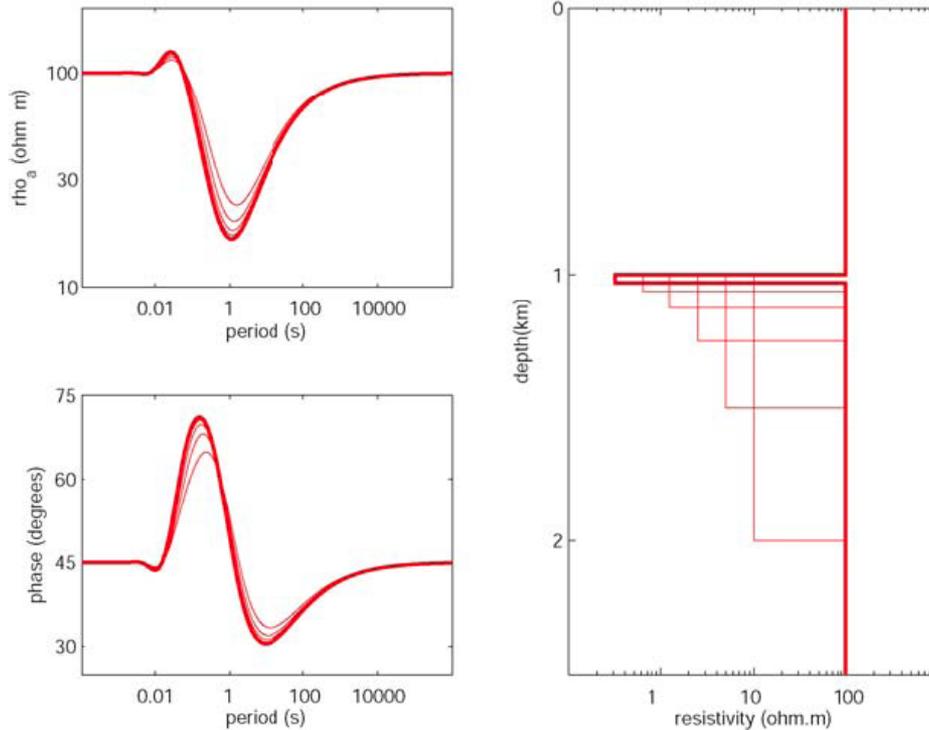
What about the bottom?



## Layer having constant conductance (conductivity x thickness)



## Layer having constant conductance (conductivity x thickness)

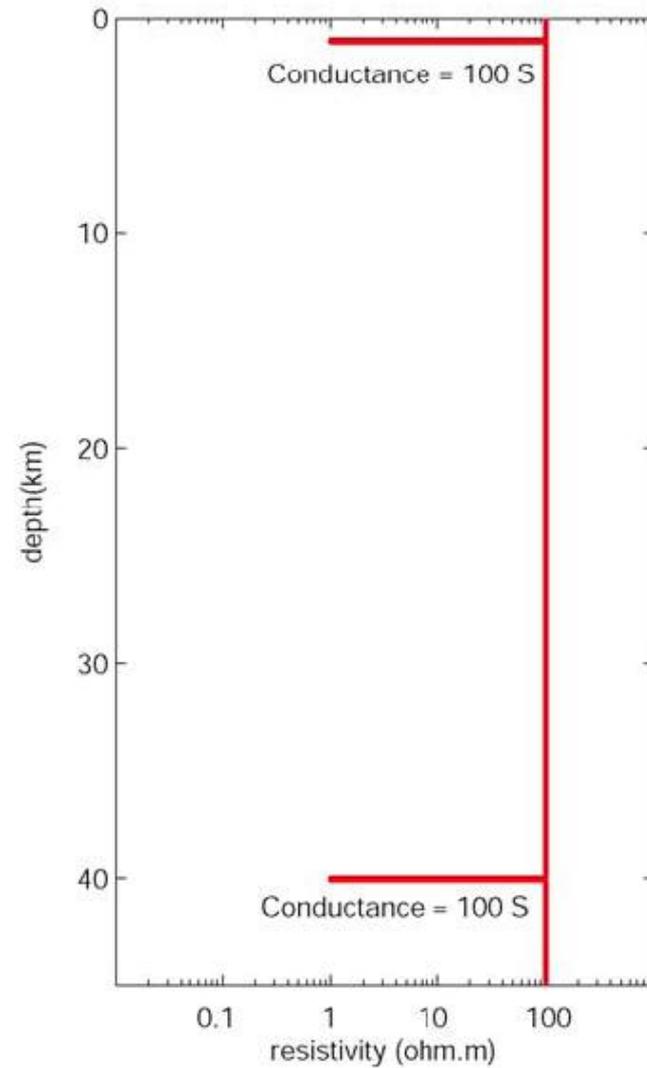
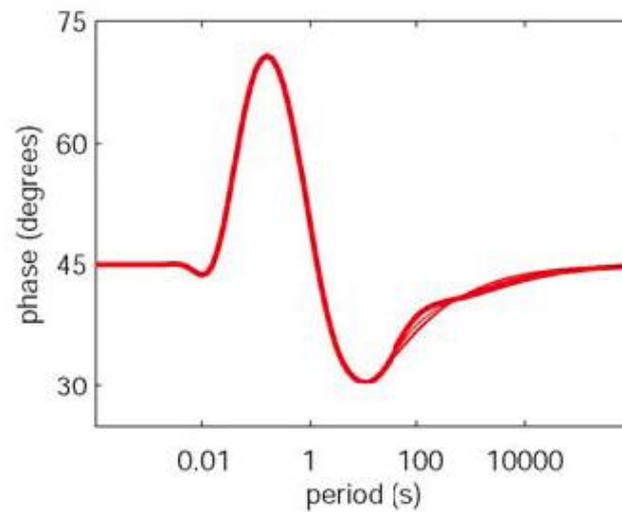
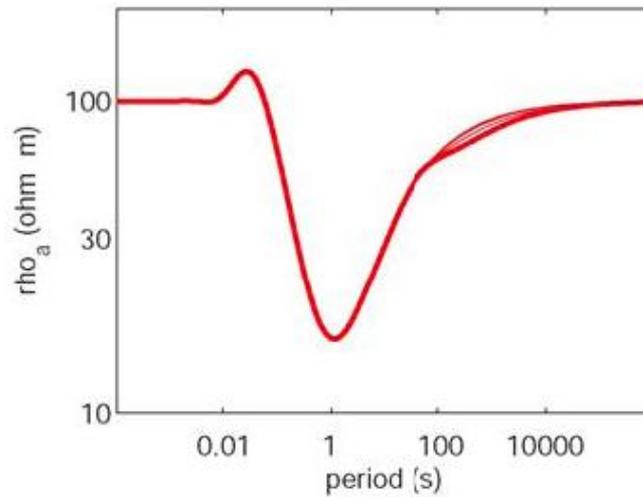


This is an example of **non-unicity**

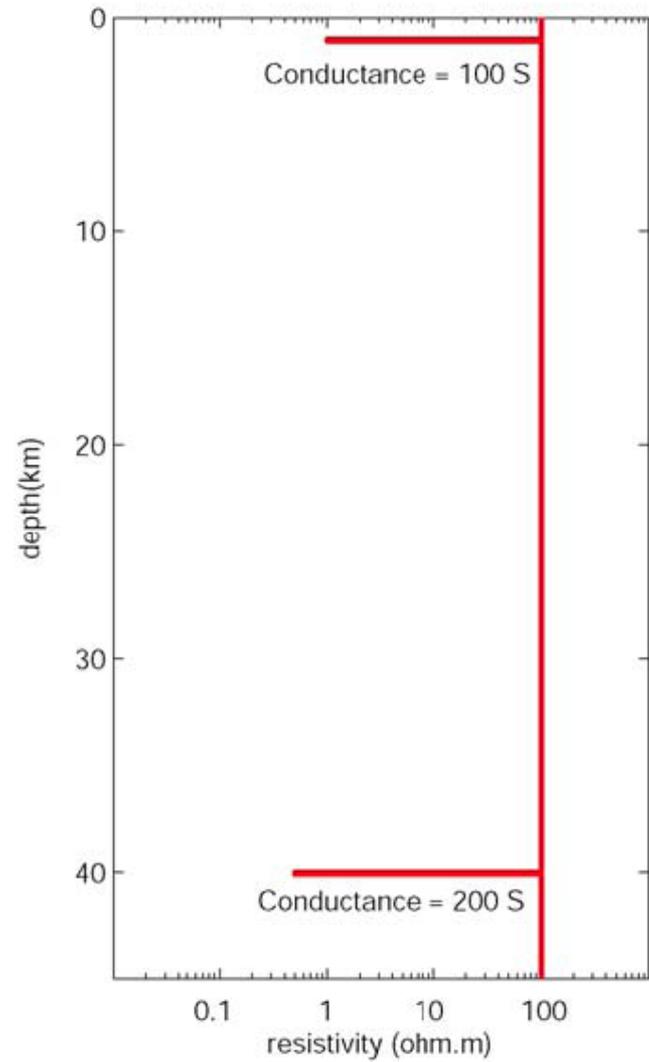
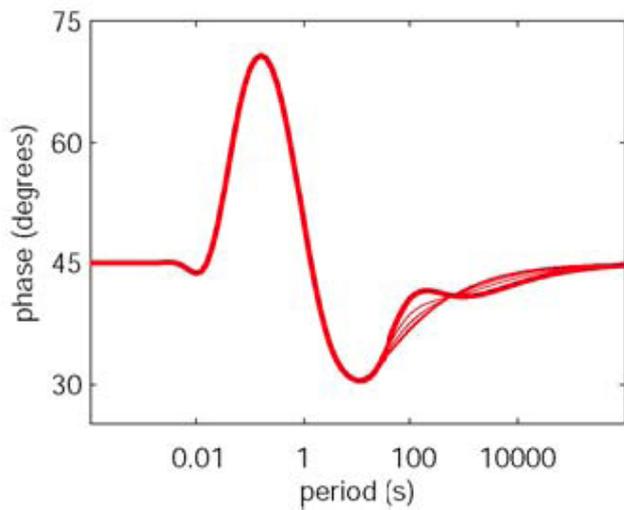
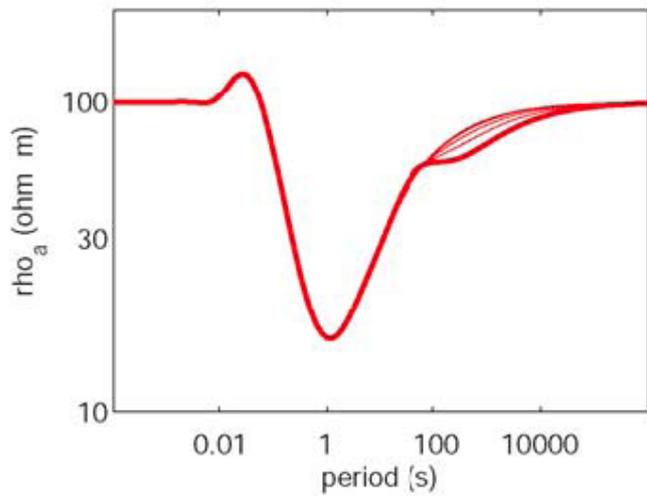
MT can **not** distinguish between layers having the same conductance and needs other information



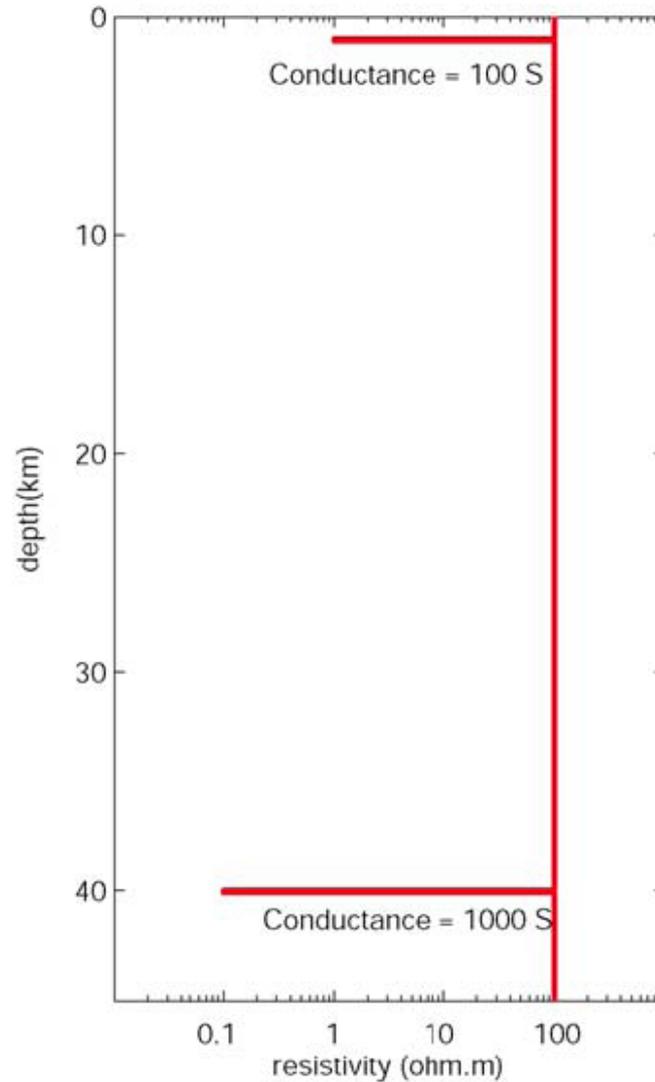
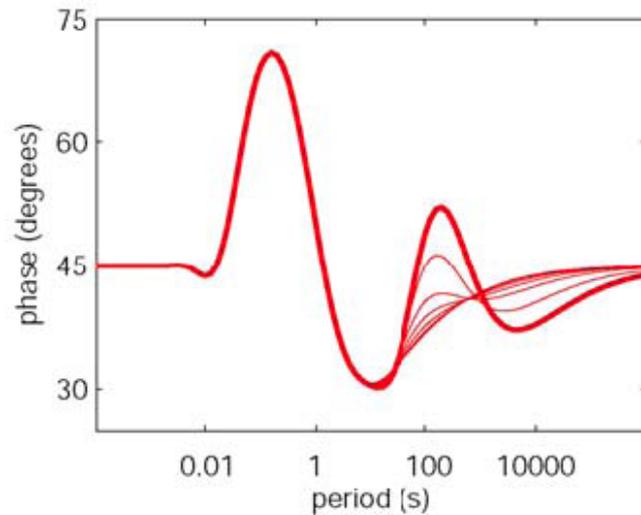
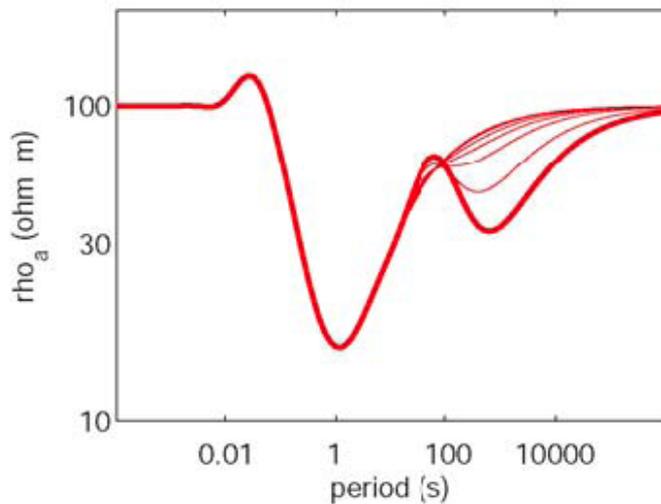
## Two conductive layers



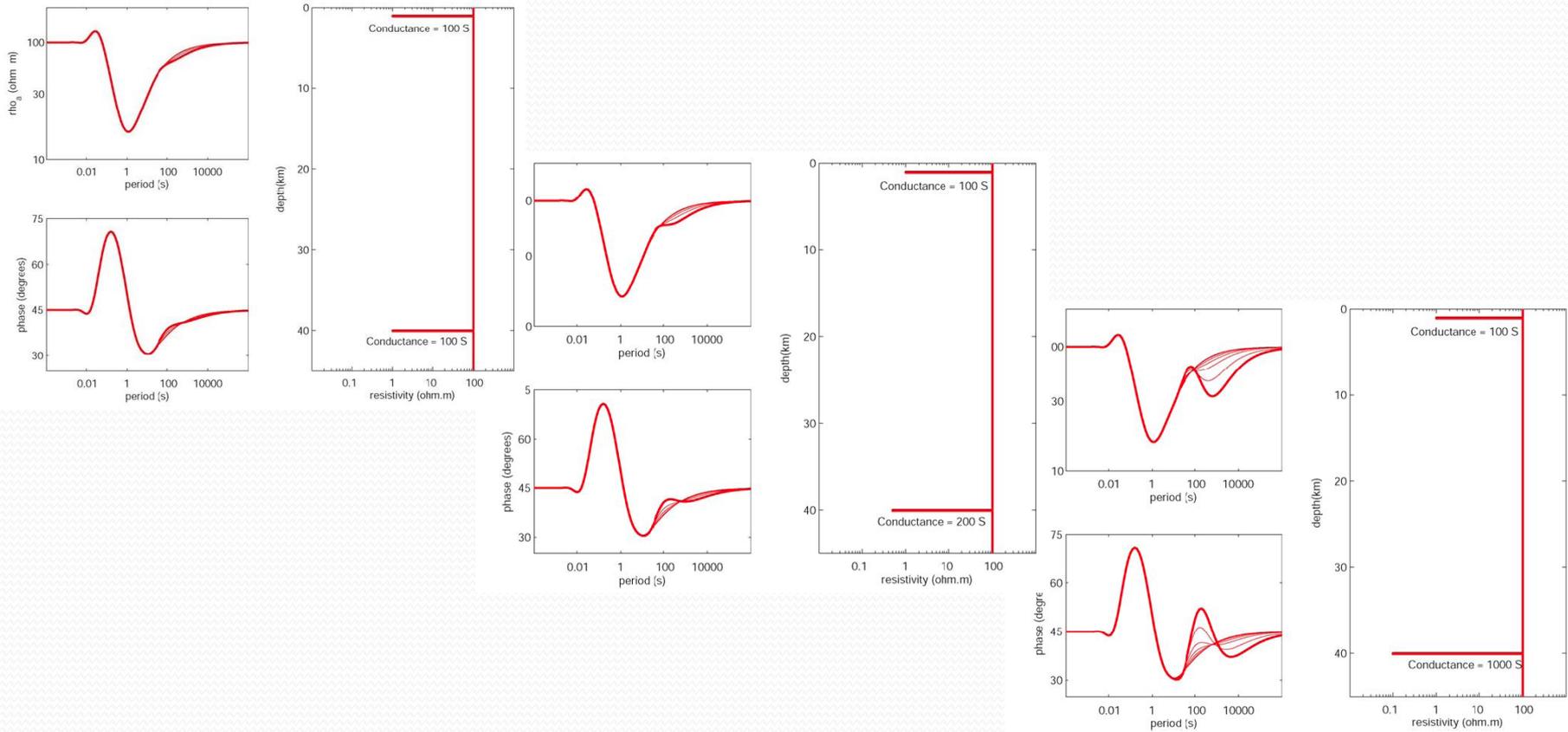
# Two conductive layers



# 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:

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

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

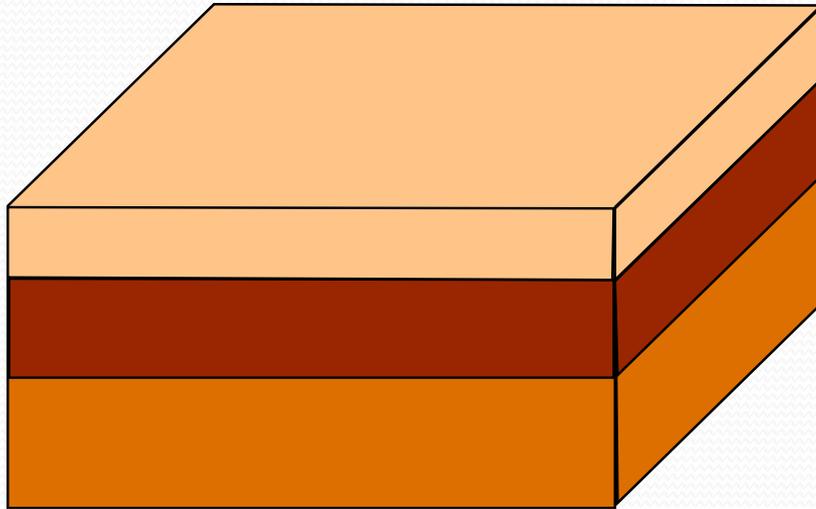
$$\begin{pmatrix} E_x \\ E_y \end{pmatrix} = \begin{pmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{pmatrix} \begin{pmatrix} H_x \\ H_y \end{pmatrix}$$

or

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



## Uniform or horizontally layered earth: 1D case



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

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

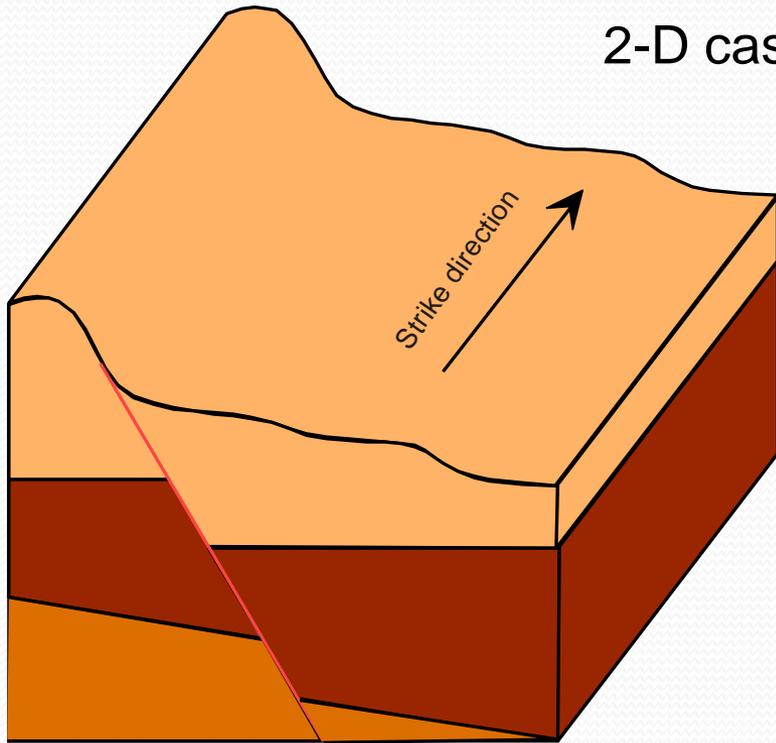
and the equations reduce to

$$E_x = Z_{xy} H_y$$

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



## 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.

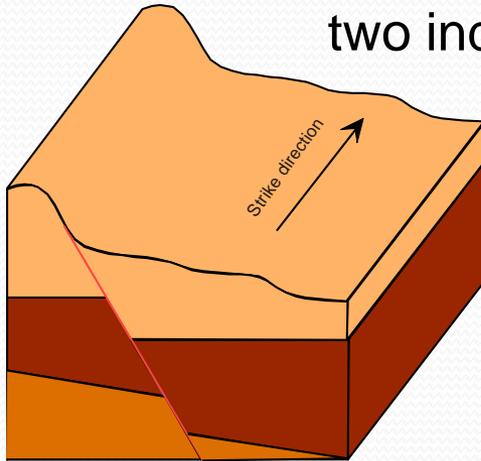
Resistivity does not vary in the  $x$ -direction and all structures extend to  $x = \pm\infty$

In contrast to a 1-D Earth, vertical field components ( $E_z$  and  $H_z$ ) can be non-zero.



## 2-D case

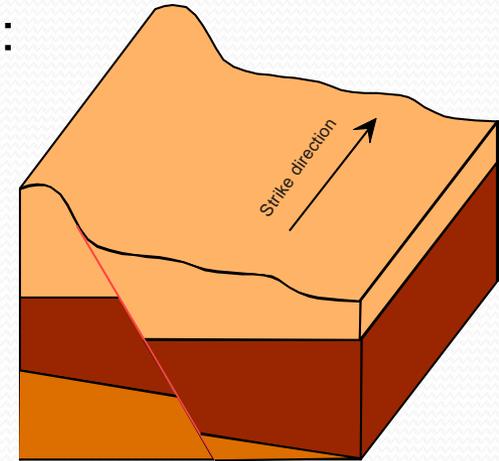
By Maxwell equation we can derive equations can now be separated into two independent subsets:



**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**



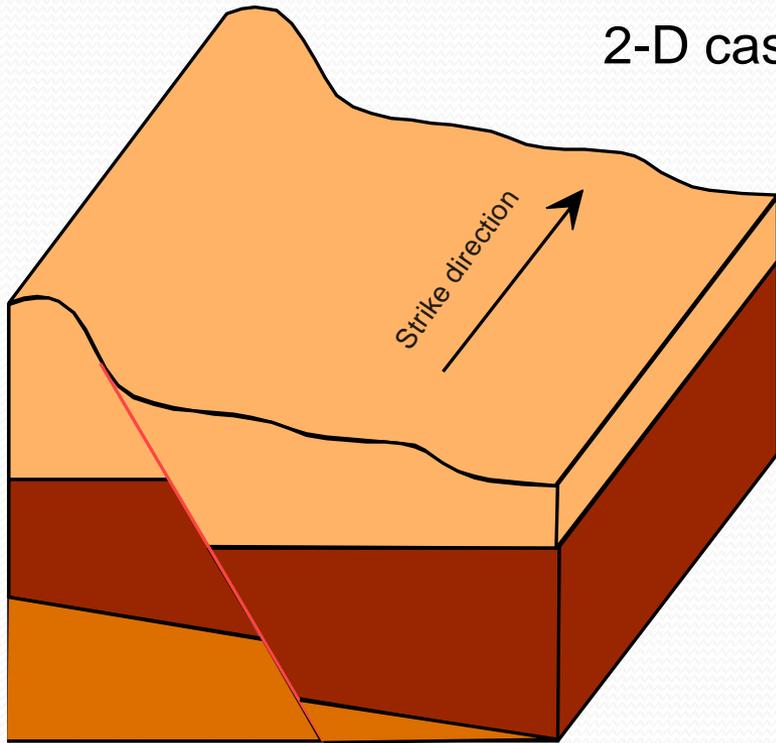
**Magnetic field polarized parallel to the strike direction.**

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  $+\theta$  (clockwise as seen from above)  
to be **E'**, then:

$$\begin{aligned} E'_x &= \cos\theta E_x + \sin\theta E_y \\ E'_y &= -\sin\theta E_x + \cos\theta E_y \end{aligned}$$

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  $\mathbf{R}$ , is

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



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

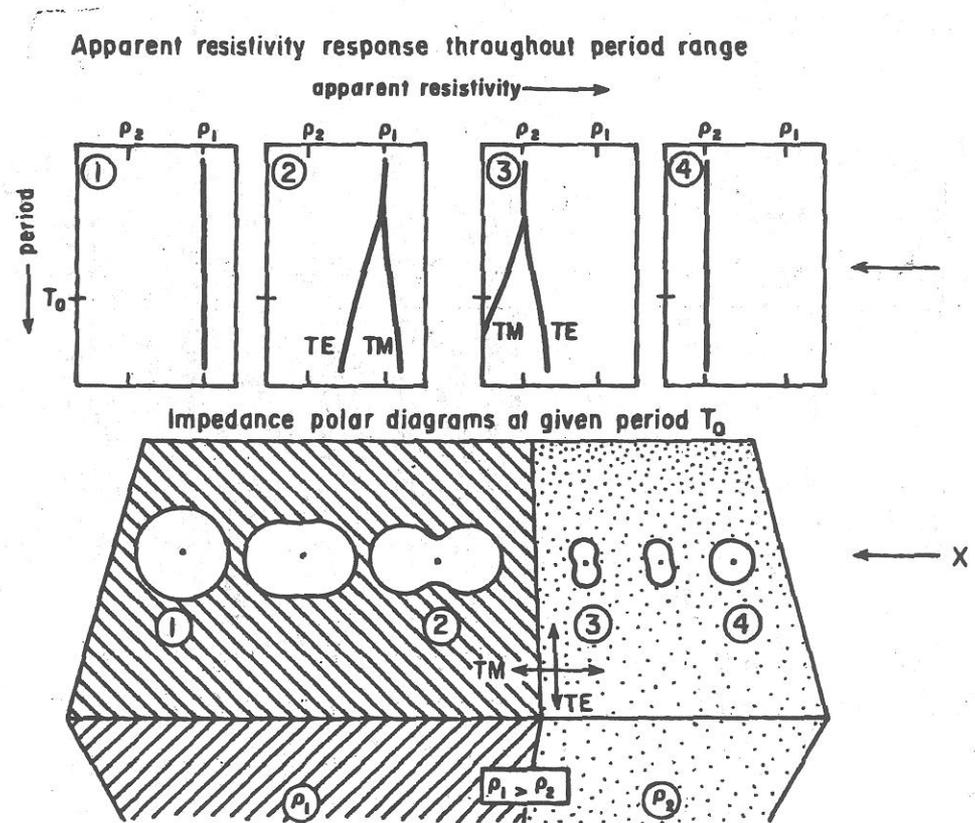
If the underground is 1D whatever the rotation we would find just one value of impedance:  $Z_{xy} = -Z_{yx}$  and the resistivity and phase xy and yx curves are identical.

For 2-D conditions, several different means have been used to find the rotation angle  $\theta_0$  between measurement direction and **strike**. All of them, however, start from the idea that when Z is measured along strike direction,  $Z_{xx}$  and  $Z_{yy}$  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  $\theta_0$  is to differentiate  $Z_{xy}(\theta)$  and  $Z_{yx}(\theta)$  with respect to  $\theta$  to give an angle  $\theta_0$  which optimizes:

$$|Z_{xy}(\theta_0)|^2 + |Z_{yx}(\theta_0)|^2$$

at each frequency. His solution

$$4\theta_0 = \tan^{-1} \frac{[(Z_{xx} - Z_{yy})(Z_{xy} + Z_{yx})^* + (Z_{xx} - Z_{yy})^*(Z_{xy} + Z_{yx})]}{|Z_{xx} - Z_{yy}|^2 - |Z_{xy} + Z_{yx}|^2}$$

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:

$$\begin{aligned}Z_{xx} + Z_{yy} &= C_1, \\Z_{xy} - Z_{yx} &= C_2,\end{aligned}$$

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  $\mathbf{H}$  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  $\varphi$  to  $x'$  can be estimated by the field data by finding the horizontal direction  $y'$  in which  $H(\varphi)$  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  $H_z$  is  $\neq 0$  a useful tool can be define: the **induction vectors**, which are computed with components

$$I_y = H_z/H_y \quad \text{and} \quad I_x = H_z/H_x$$

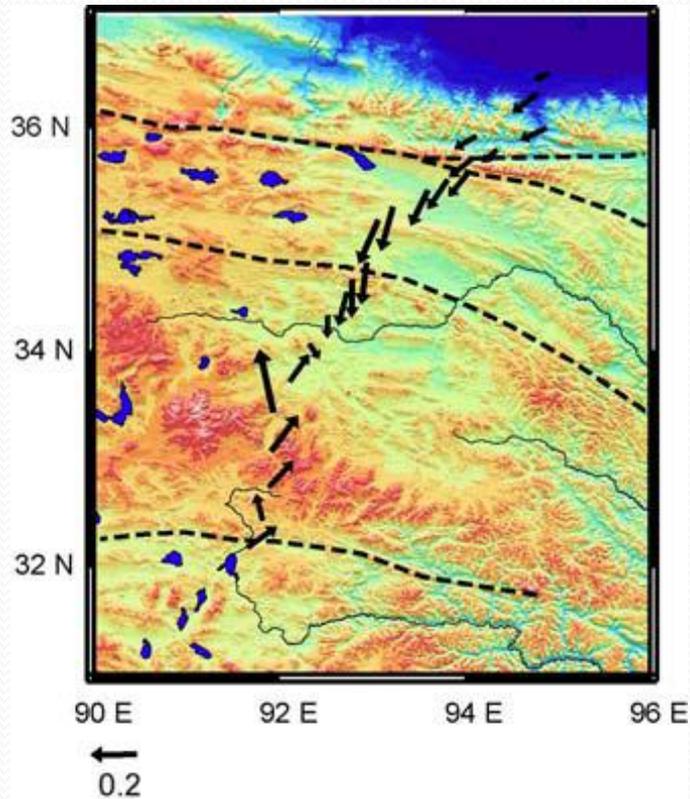
In the *Parkinson convention*, induction vectors will point at conductors.

In the *Schmucker convention* they point away from conductors.

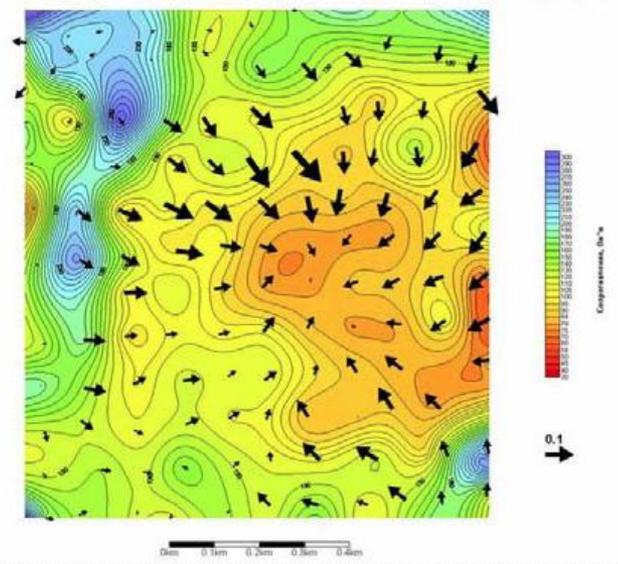
Measuring vertical magnetic fields requires a lot of physical effort. Burying an induction coil requires a 1 m hole to be dug vertically downwards.

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 Northwest company)

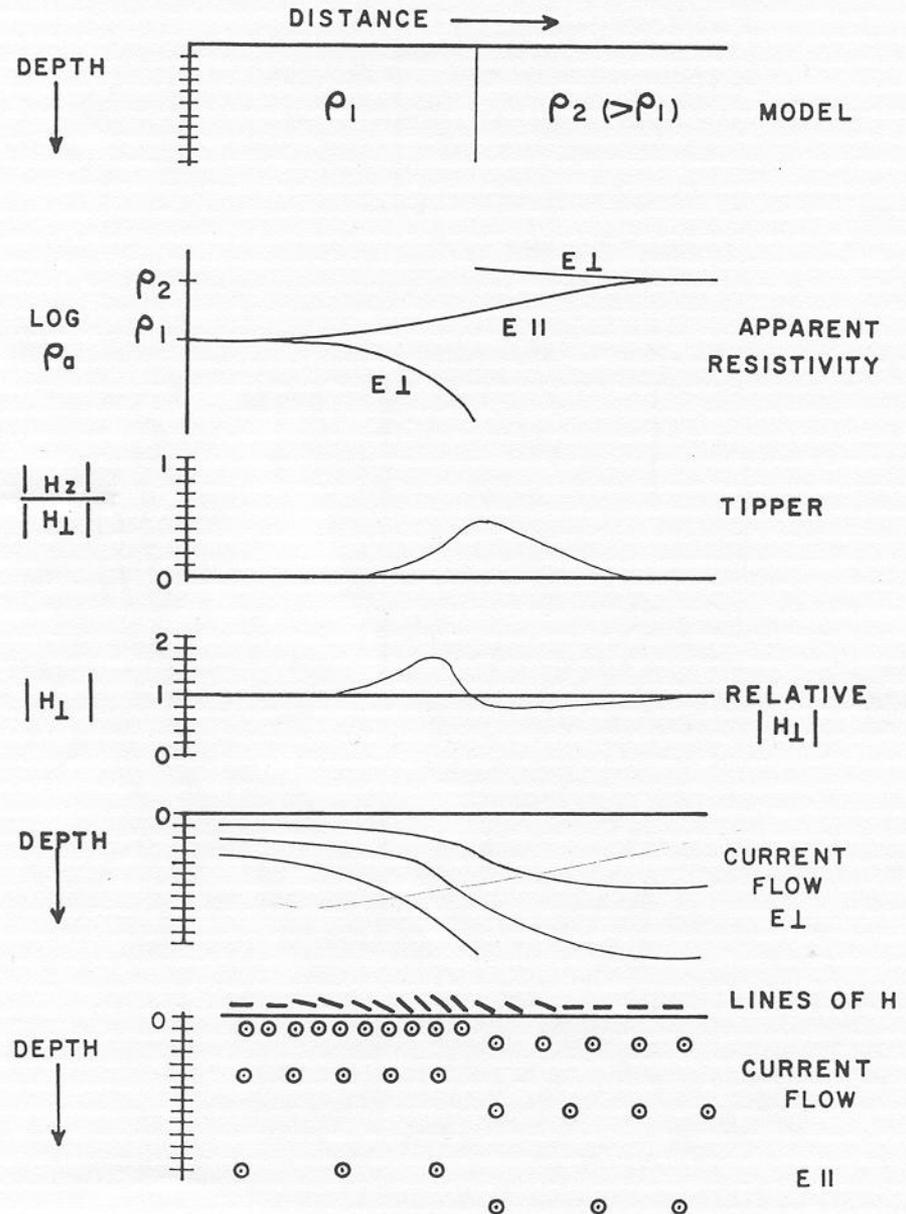


## Definition of a 2D condition

Boundary conditions:  
 $E_{\parallel}$  is continuous,  
 whereas the current density of  
 $E_{\perp}$  is continuous.

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

The same for impedance and hence for resistivity

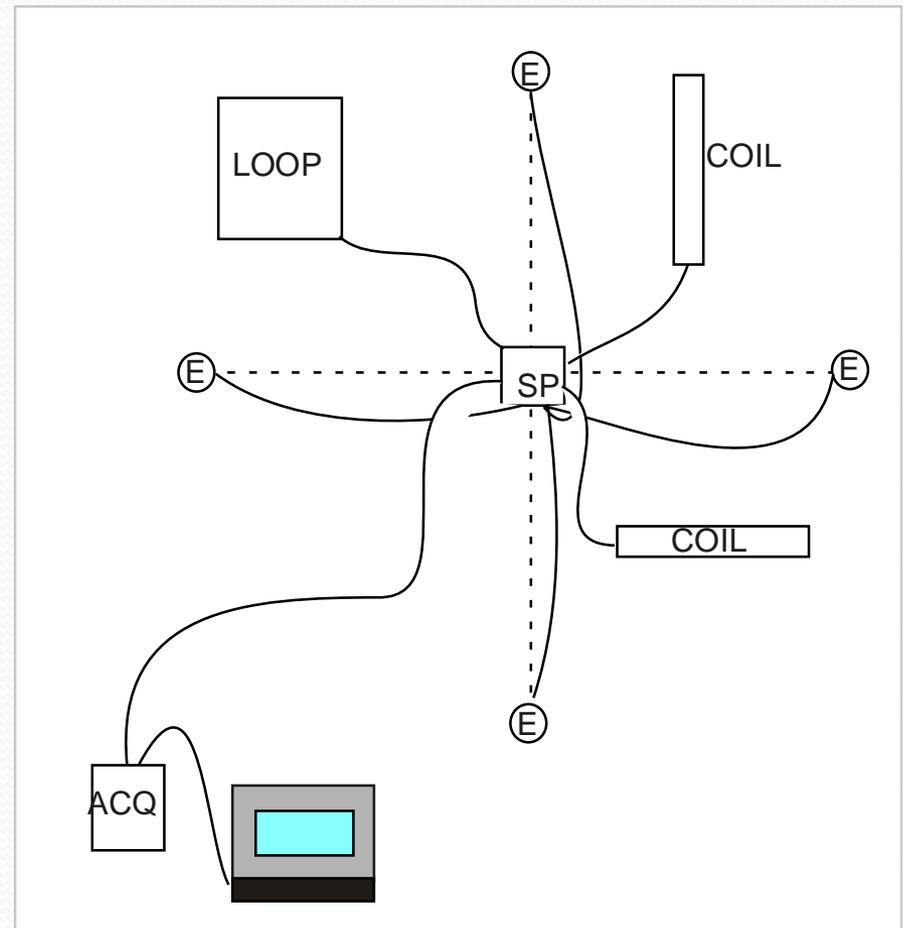


MT data are acquired in the field, as measurements of electric and magnetic fields with time

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

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-CIPb)



## ***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



High Frequency Magnetotelluric Receiver  
[www.westsystems.it](http://www.westsystems.it)



## ***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  $500 \mu\text{Hz}$  (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.

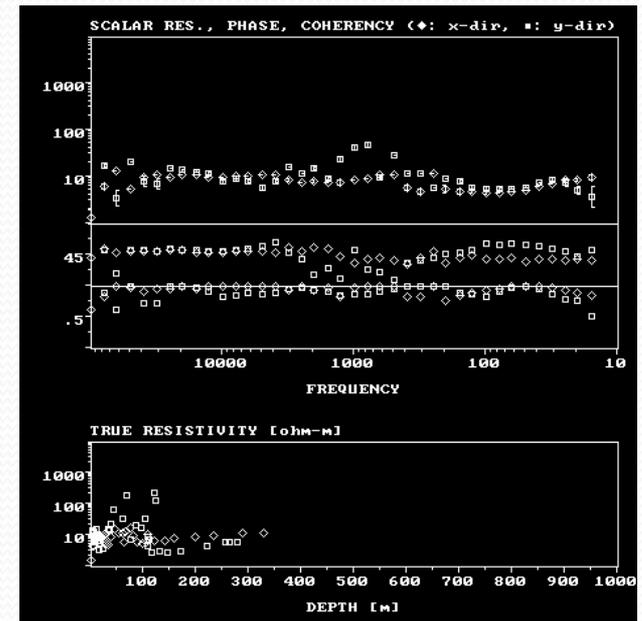
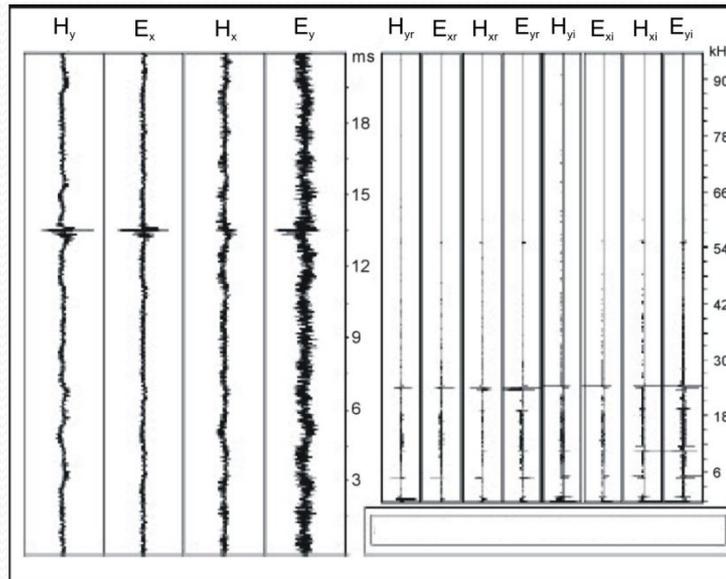


In most systems for carrying out magnetotelluric surveys today, the five field components are converted to digital form and are either stored for later spectral analysis or converted immediately to spectral form before being stored.

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,  $\rho_a(\omega)$ , and phase,  $\Phi(\omega)$ , are then computed as a function of frequency.

$\rho_{xy}(\omega)$  and  $\Phi(\omega)_{xy}$  are computed from  $E_x$  and  $H_y$

$\rho_{yx}(\omega)$  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.**

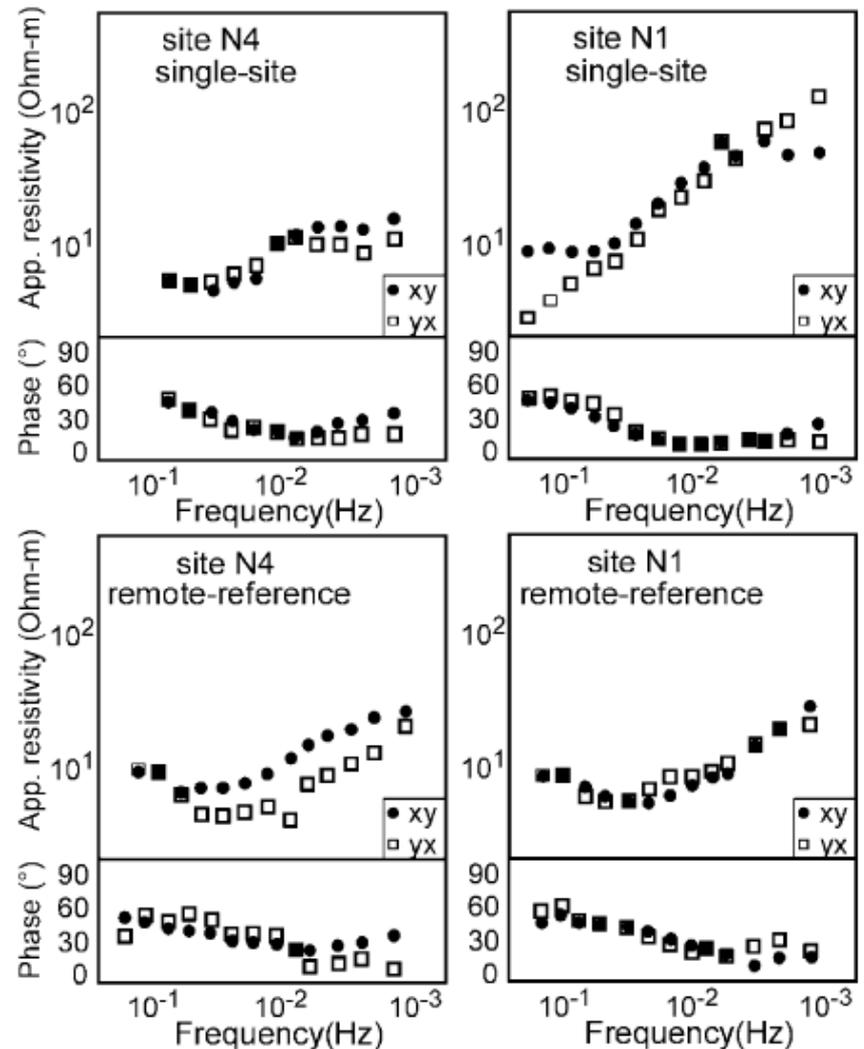
Magnetic fields are correlated between the two or more receiver stations, with uncorrelated portions being considered as noise in the magnetic field detection, and removed for later processing.

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.

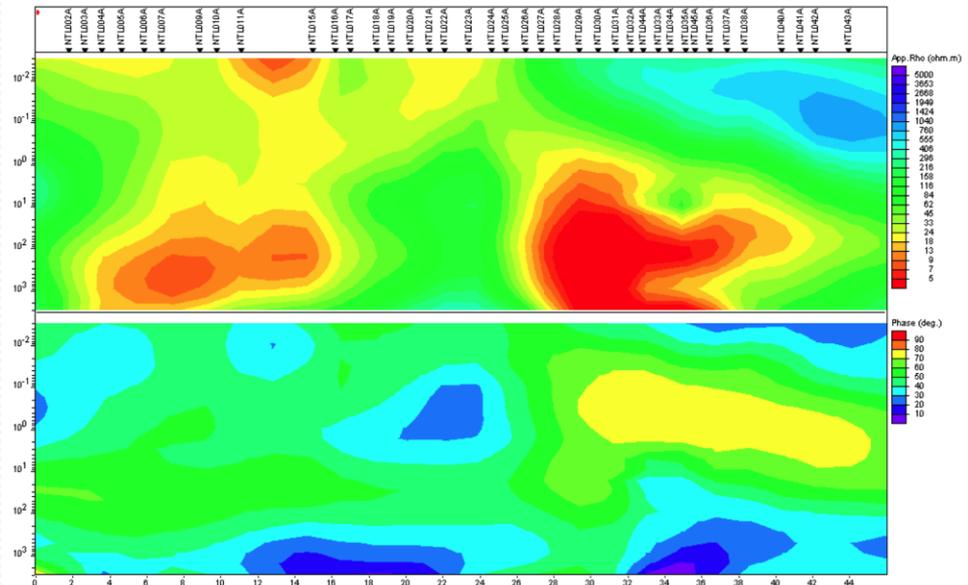
If the natural electromagnetic field strength is unusually weak during a recording period, or if there is some phenomenon which precludes an effective analysis of the field, it may be necessary to repeat the measurements at a more favourable time.



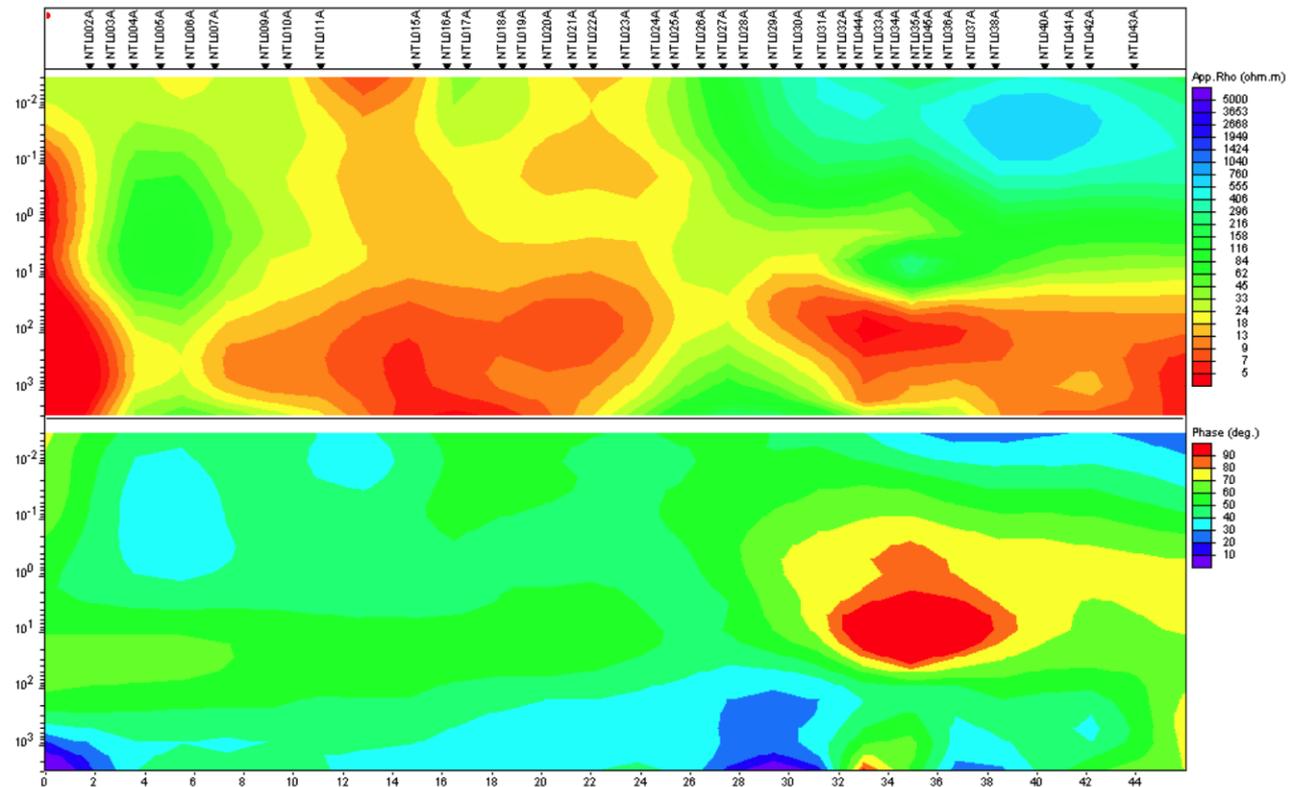
## Pseudosection display of MT data

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

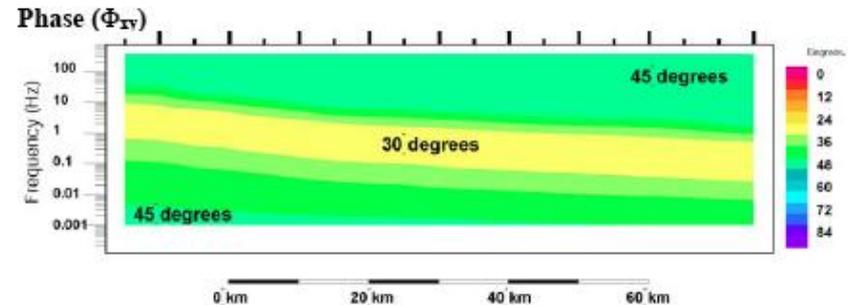
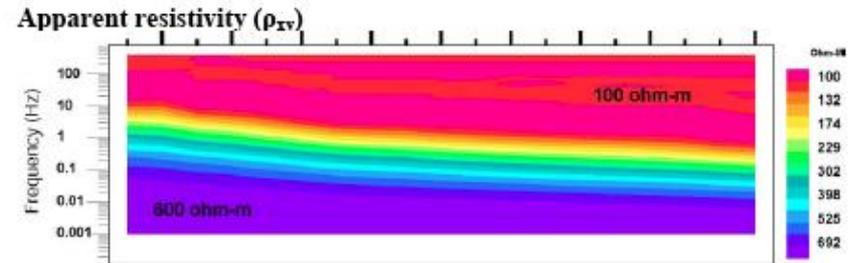
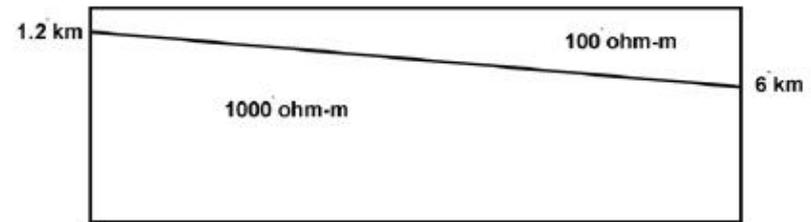
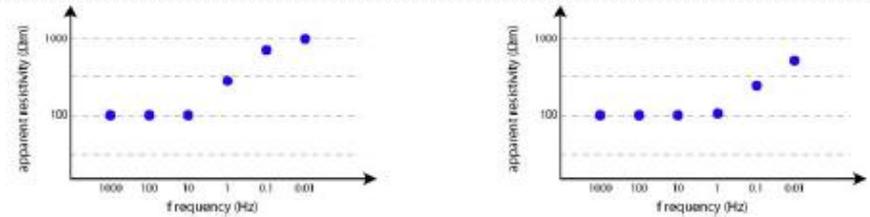
It allows apparent resistivity and phase at many MT stations to be shown in a single figure.



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

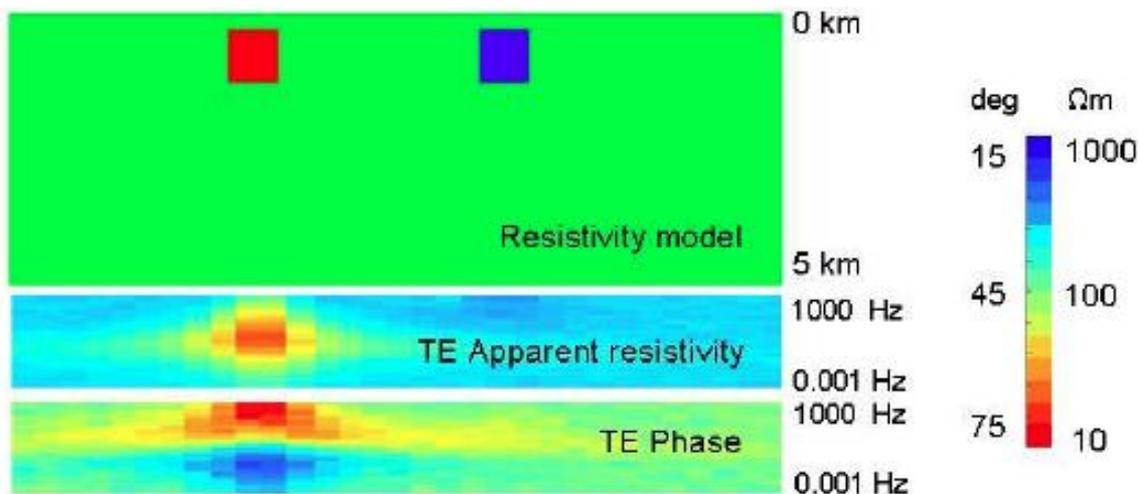


As the layer thickens to the right, the frequency needed to pass through the 100  $\Omega\text{m}$  layer decreases.



## TE-mode

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

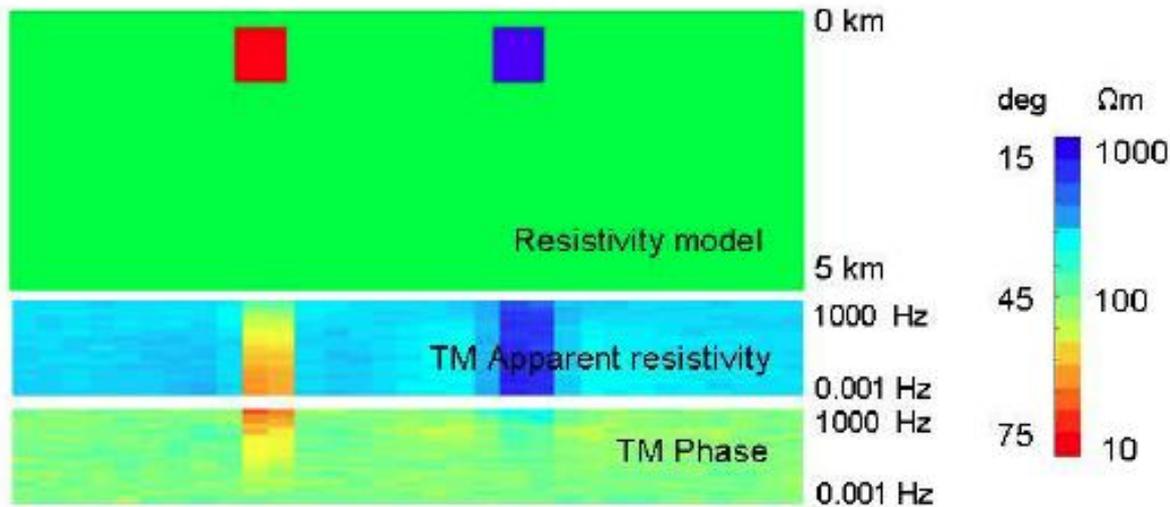


- In the TE mode, electric currents never cross boundaries between regions of differing resistivity. Thus since  $E_x$  is spatially continuous with respect to  $y$ ,  $\rho_{xy}$  is also spatially continuous.
- Electric currents are induced in the conductive prism, but not in the resistive prism. Physics is dominated by **inductive effects that will not be observed at zero frequency**. Resistive prism (1000  $\Omega\text{m}$ ) hardly changes the apparent resistivity. It will not be easily detected by the TE mode.
- The conductive prism (10  $\Omega\text{m}$ ) 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.

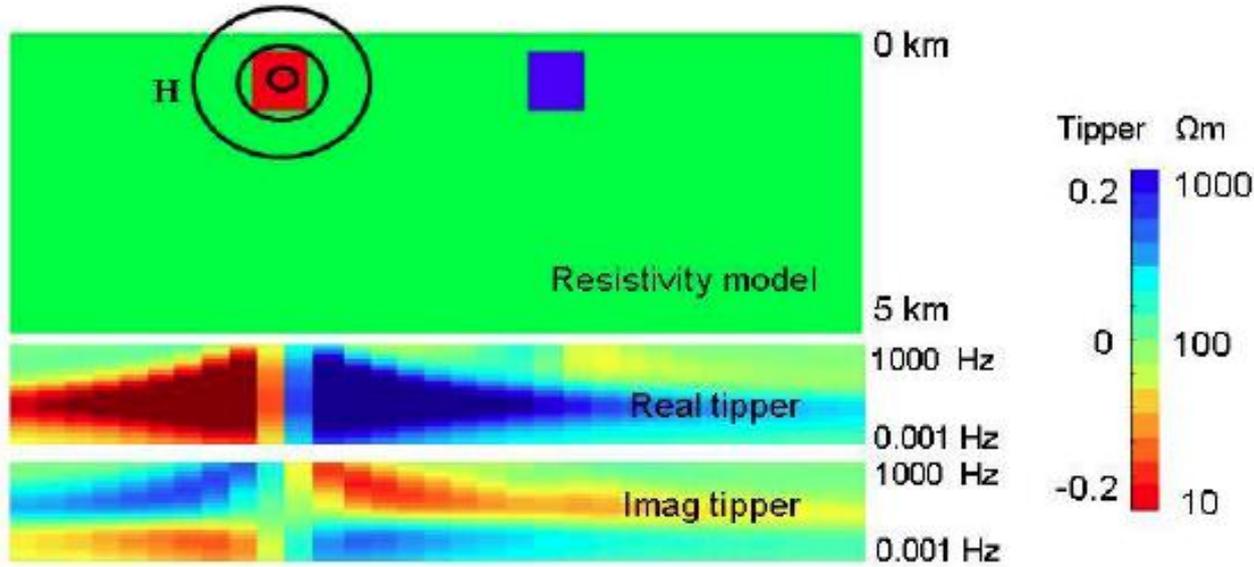


- In the TM mode, electric currents **cross the boundaries between regions of differing resistivity**. This causes electric charges develop on the boundaries.
- Thus physics of this mode includes both **inductive and galvanic effects**. **Galvanic** effects, such as charge build up on boundaries, will 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 ( $\rho_{yx}$ )** 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 synthetic data produced.

### **Inversion:**

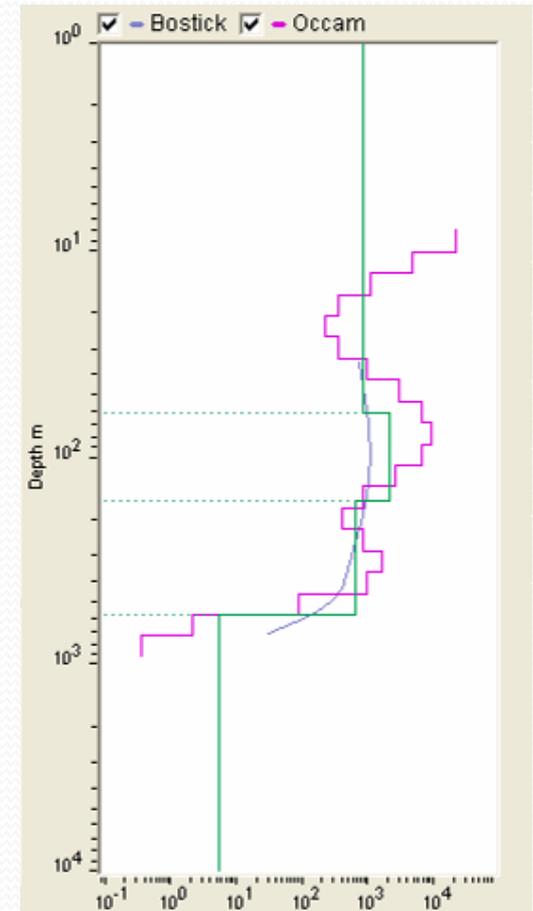
- 1) A model is chosen to represent the subsurface resistivity distribution (**a priori model**).
- 2) By numerical simulation, the response of the model is computed and synthetic data produced. These are compared to the real data. The parameters of the model are then systematically varied until the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



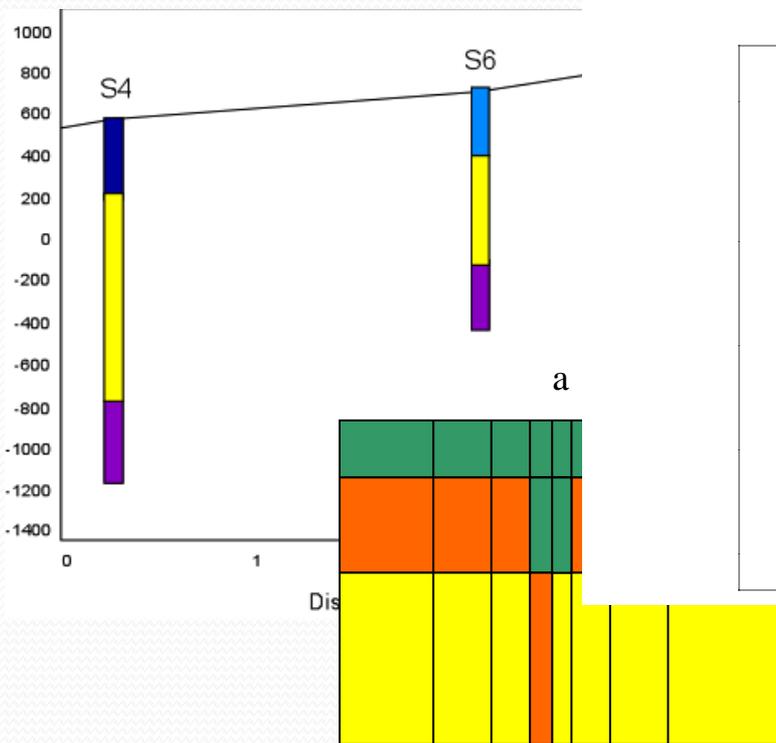
## 1D modelling

The most common are

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

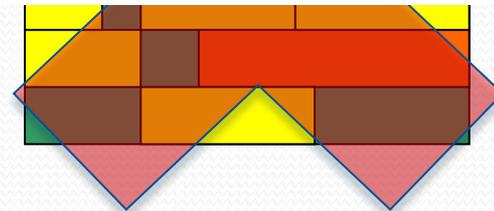


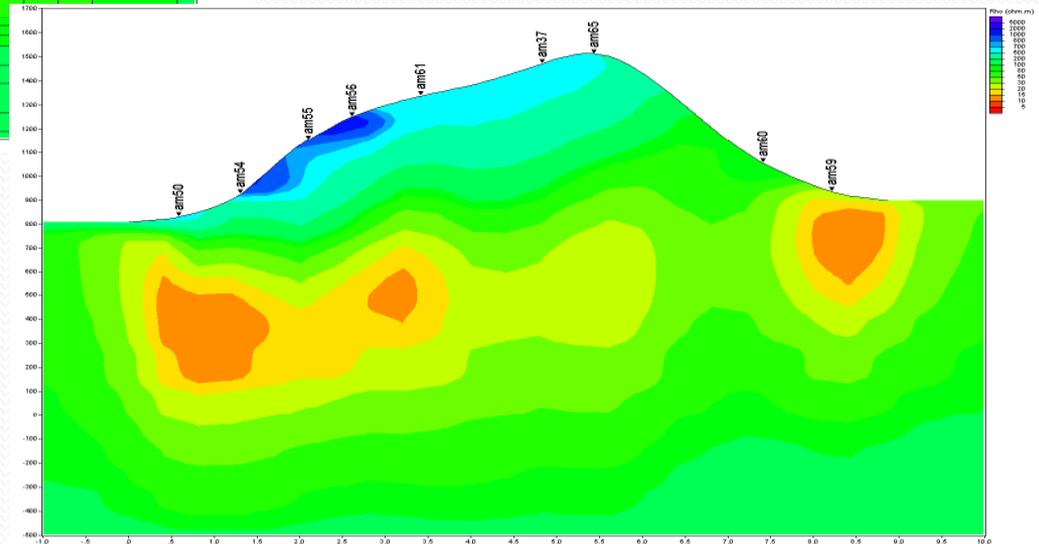
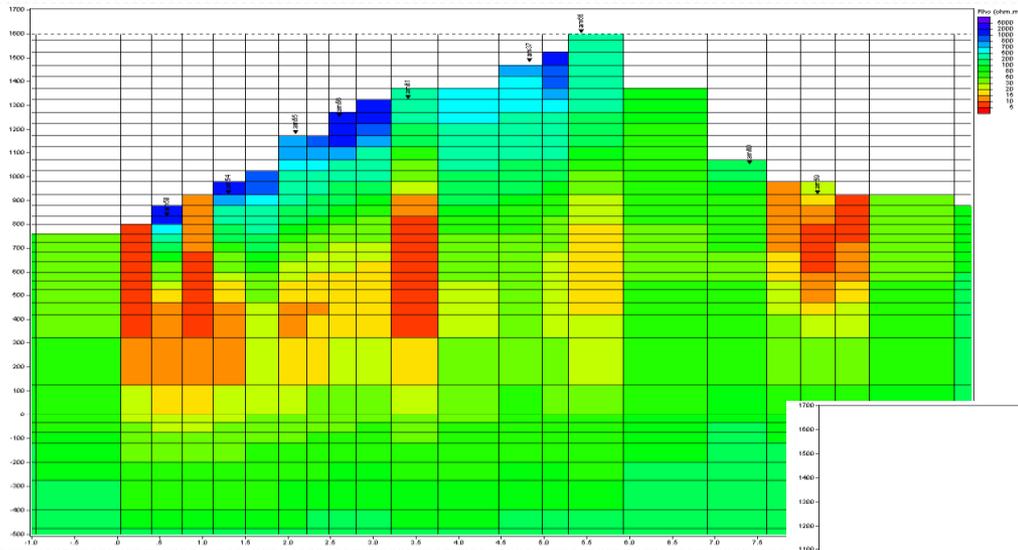
Model files contain the dimensions of prismatic (3D) rectangular (2D) or linear (1D, thickness) elements composing the model mesh.



```

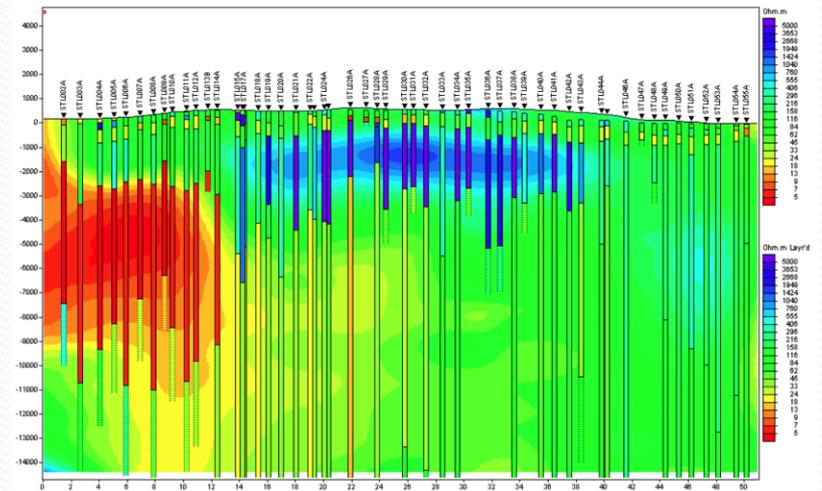
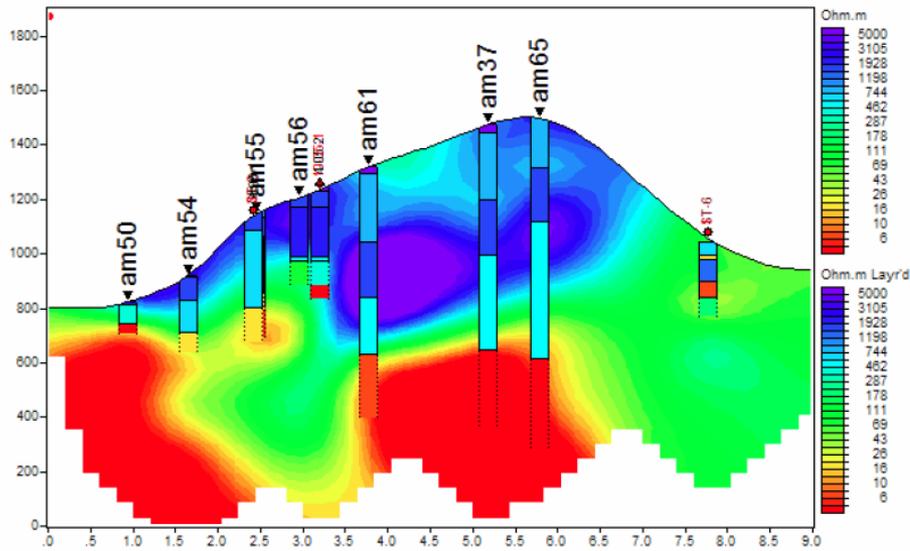
Model file ('north input.example1')
      10 5
10000 5000 2000 2000 1000 1000 2000 2000 5000
      10000
      30 100 200 300 500 1000
      1 1 1 1 1 1 1 1 1
      1 1 2 3 3 3 3 2 1 1
      2 2 3 3 3 3 3 3 2 2
      1 1 2 3 3 3 3 2 1 1
      2 2 2 2 2 2 2 2 2 2
      1000 100 5
    
```

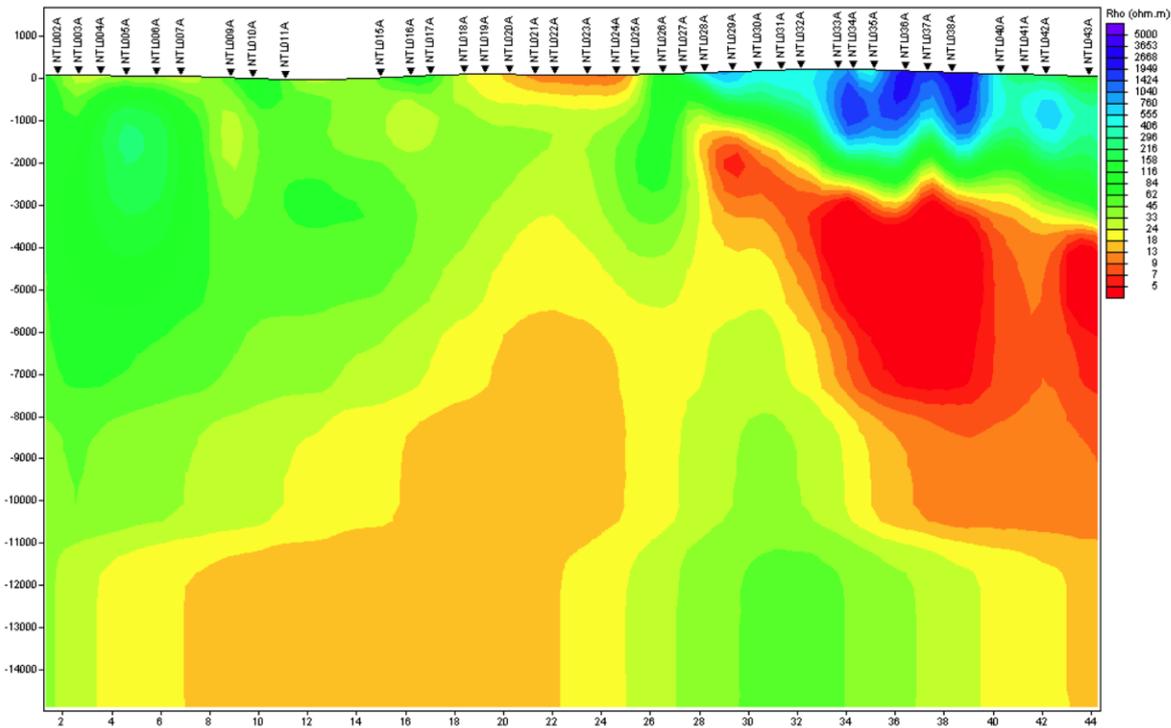




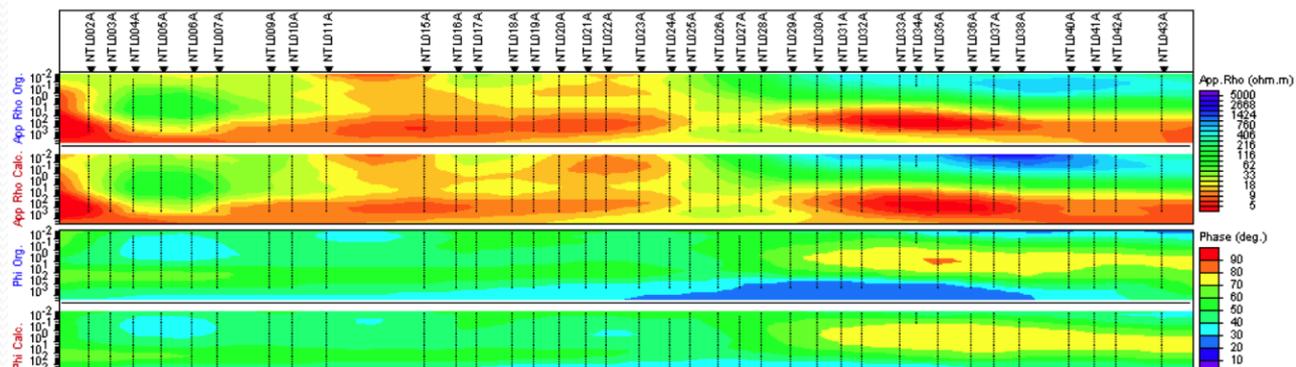
The final model may be imaged as a mesh (left) or a smooth distribution obtained by interpolation and gridding (right)

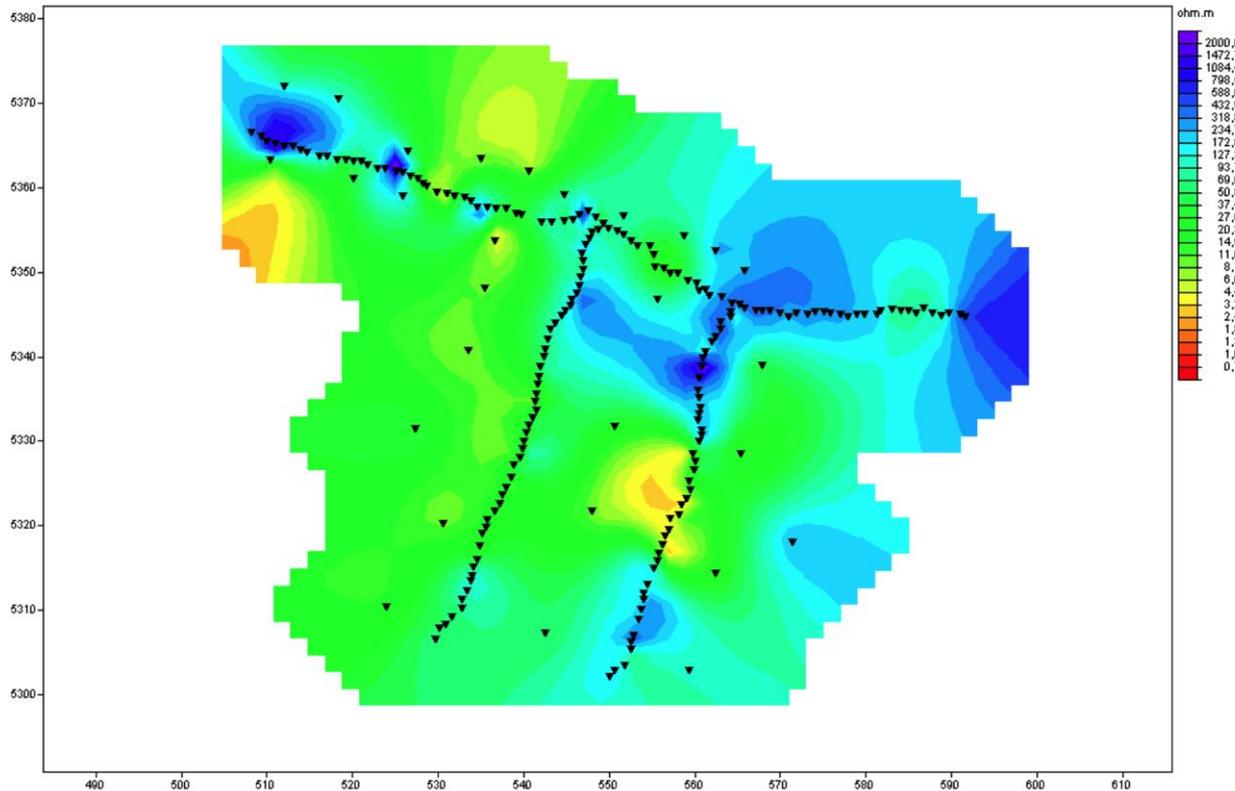






To check model quality we compare real and synthetic data (pseudosections)



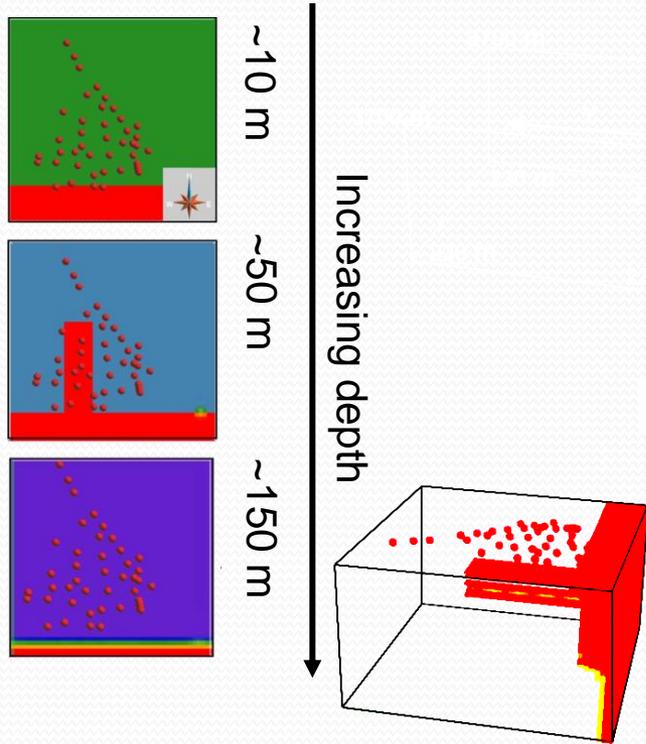


We can image resistivity distribution at depth by interpolating resistivity values obtained in 2D or 1D models section at the same depth.

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

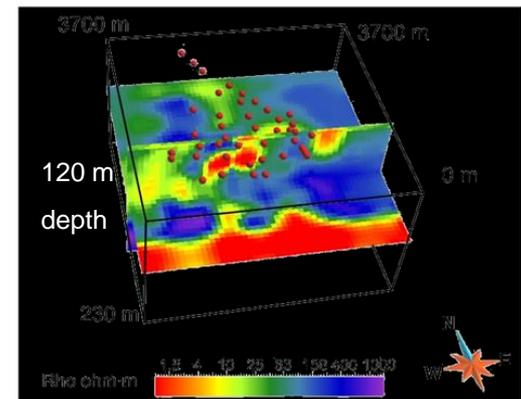
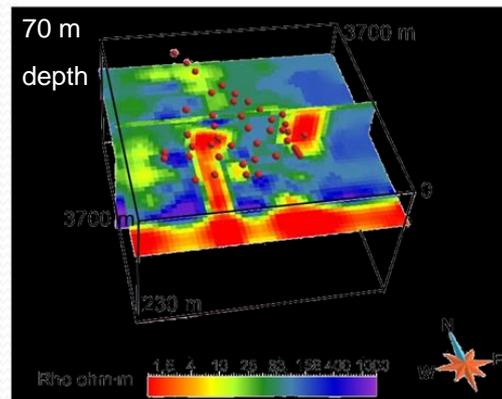
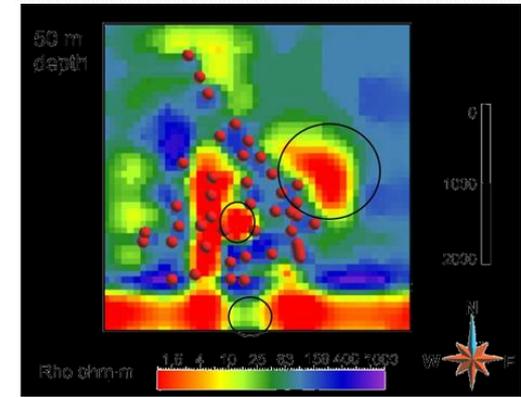
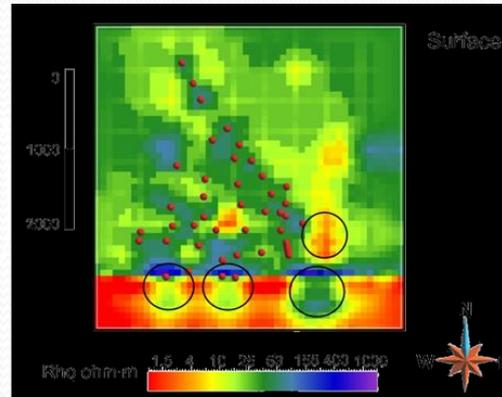


# Synthetic model



Forward modelling

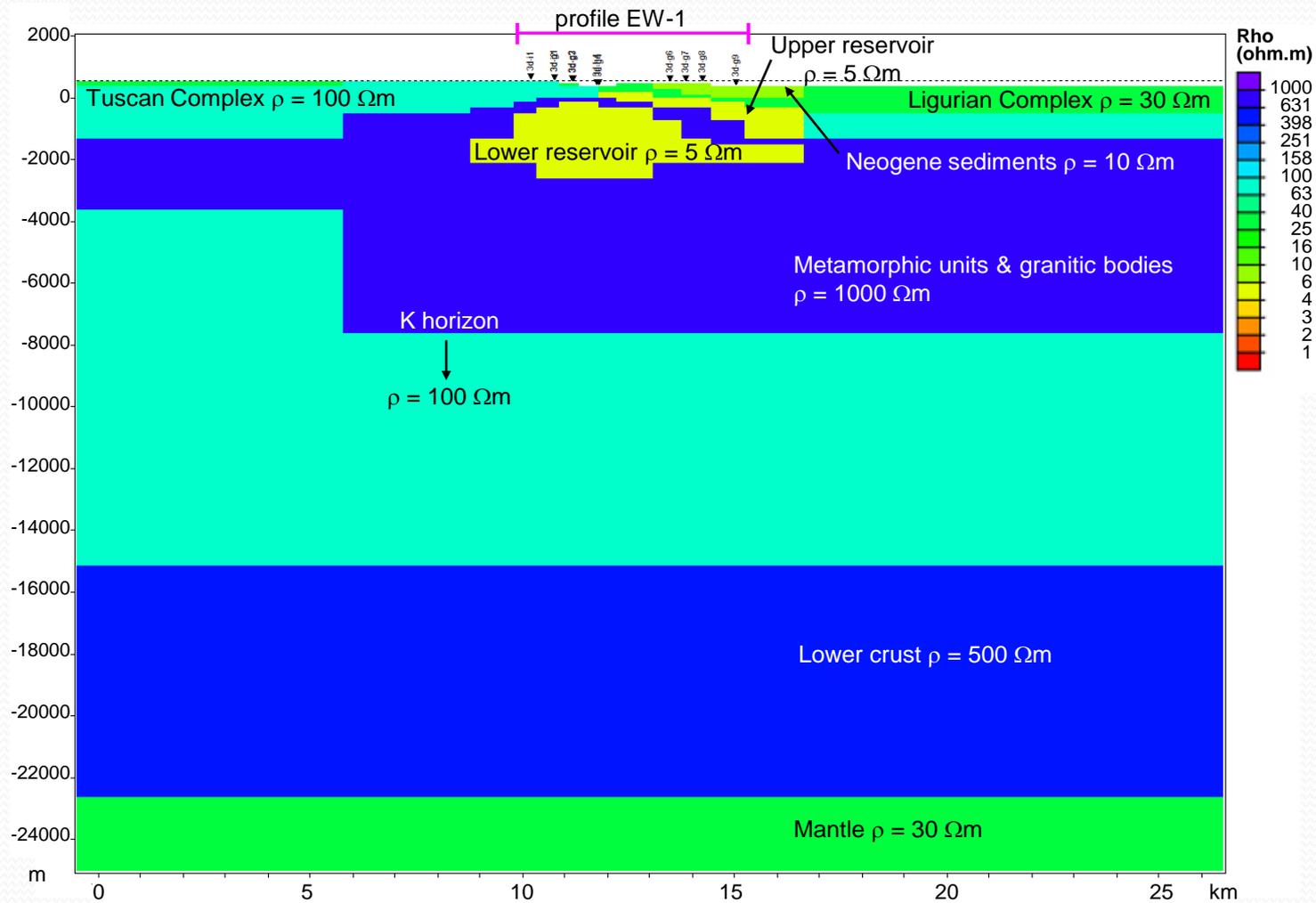
(Mackie and Madden, 1993)



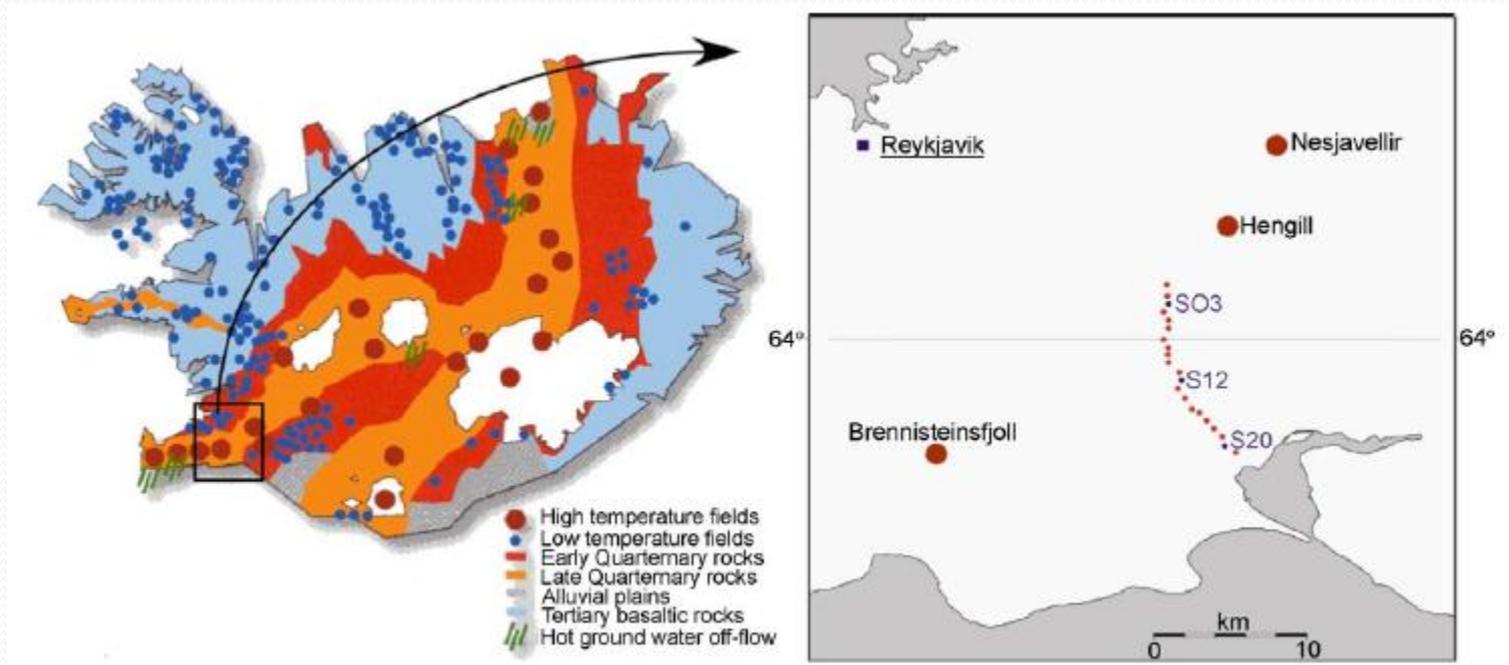
Inverse modelling WSINV3D

(Siripunvaraporn et al. 2005)





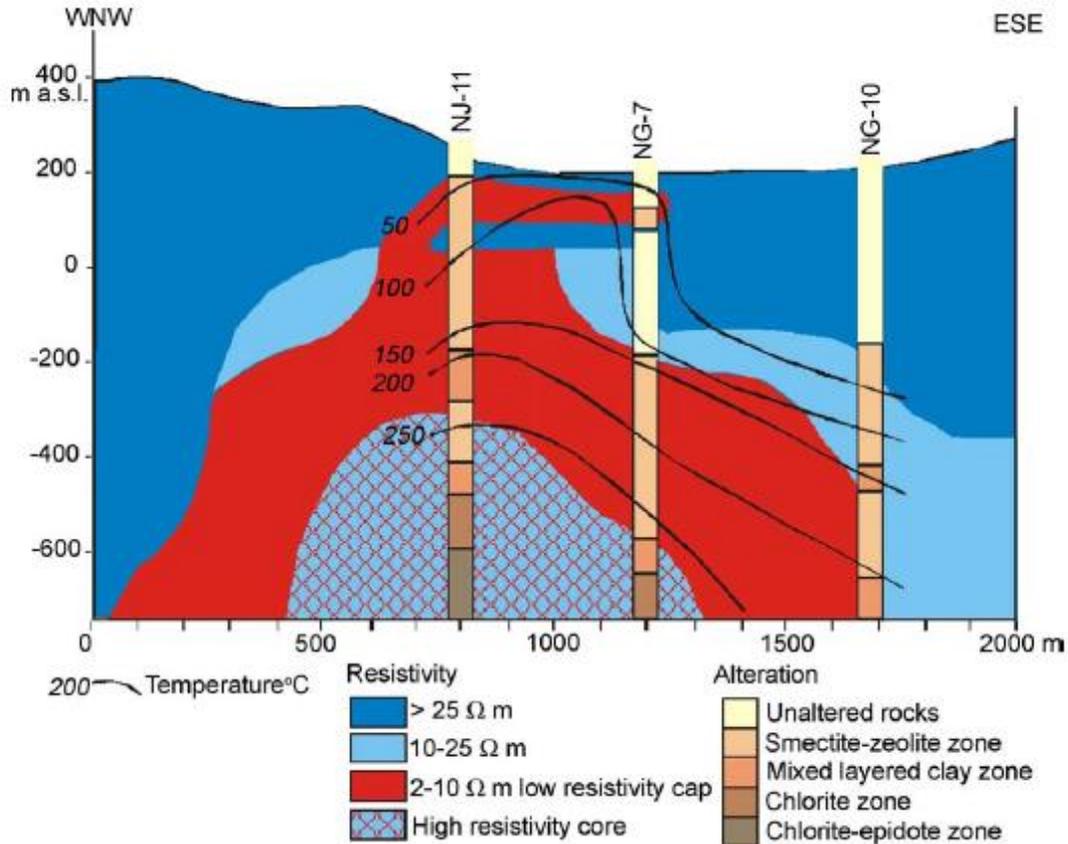
# Case Study: Iceland



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.



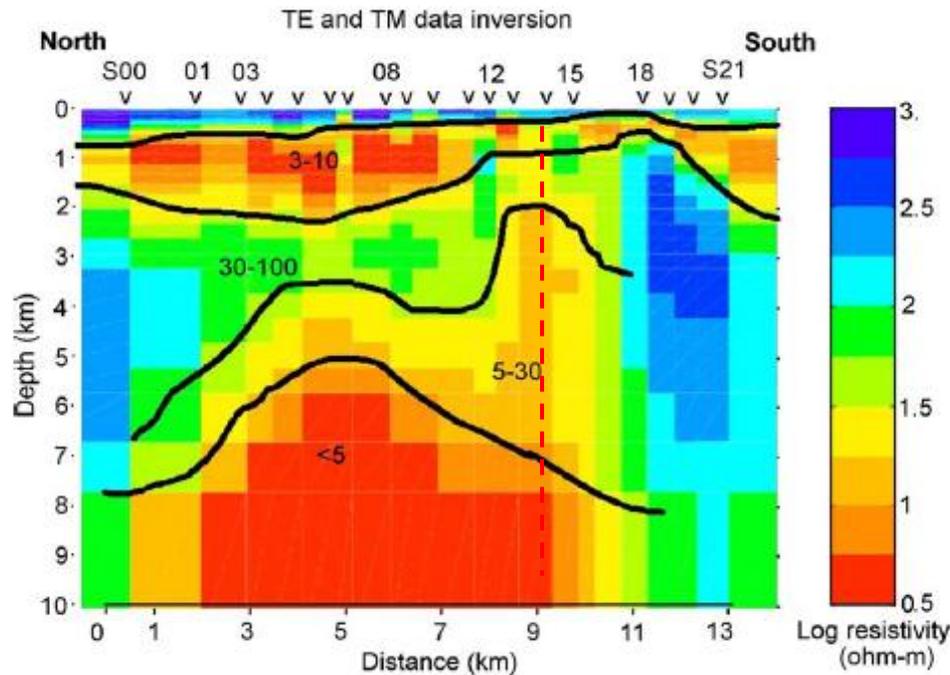
# Case Study: Iceland



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



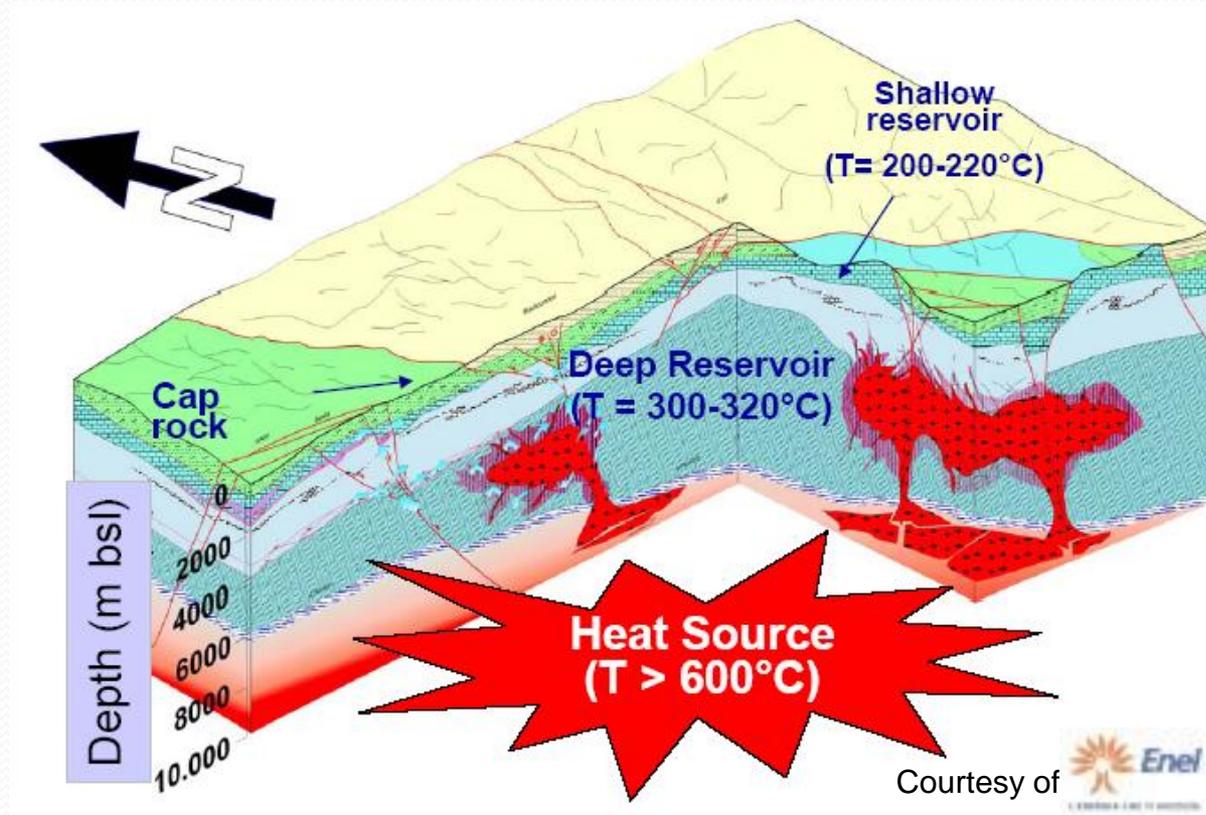
Tuscany is characterised by NNW-SSE faults linked to the Alpine orogeny.

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



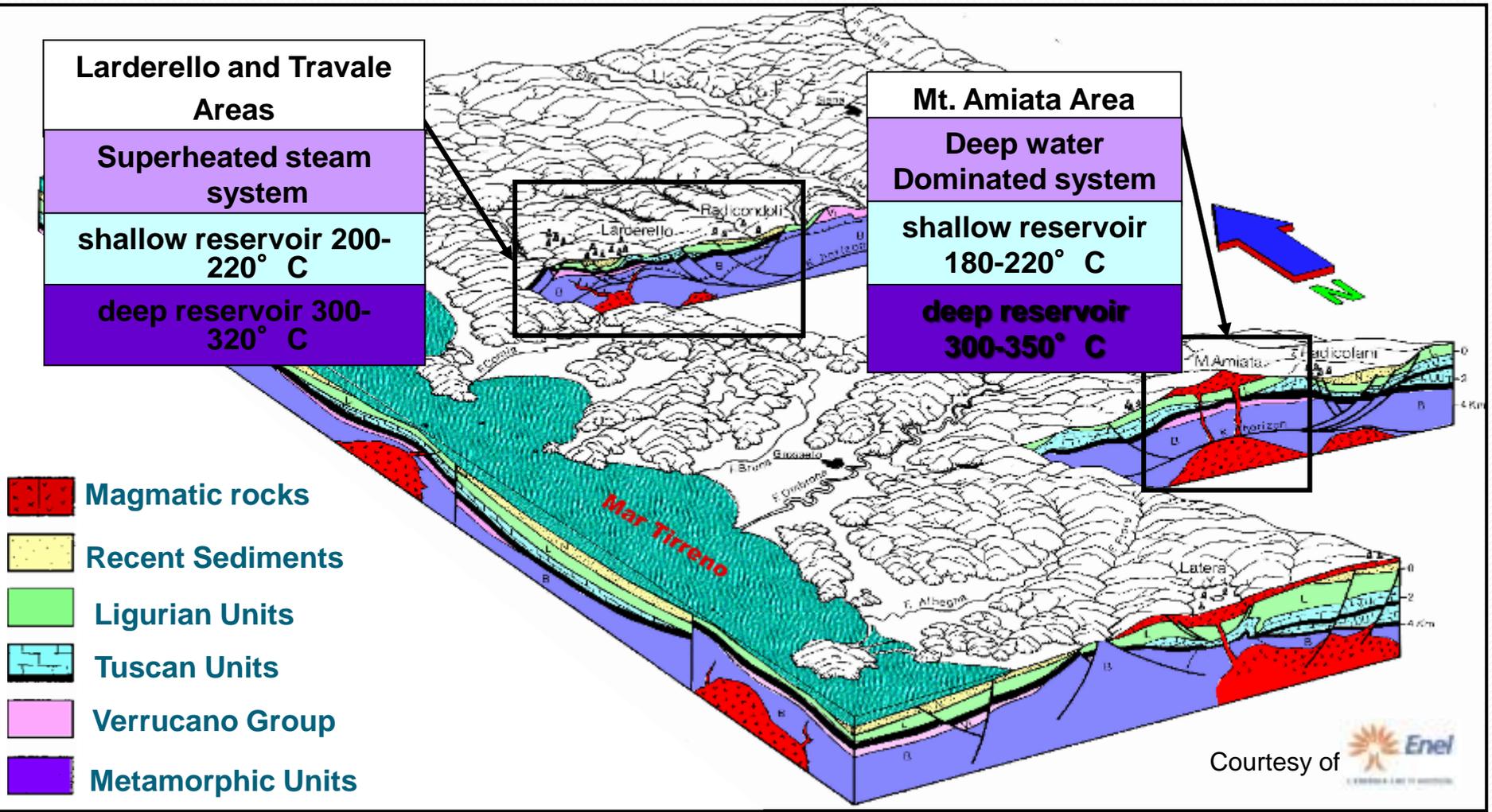
# Case Study: Italy

**Larderello and Travale Areas**

- Superheated steam system
- shallow reservoir 200-220° C
- deep reservoir 300-320° C

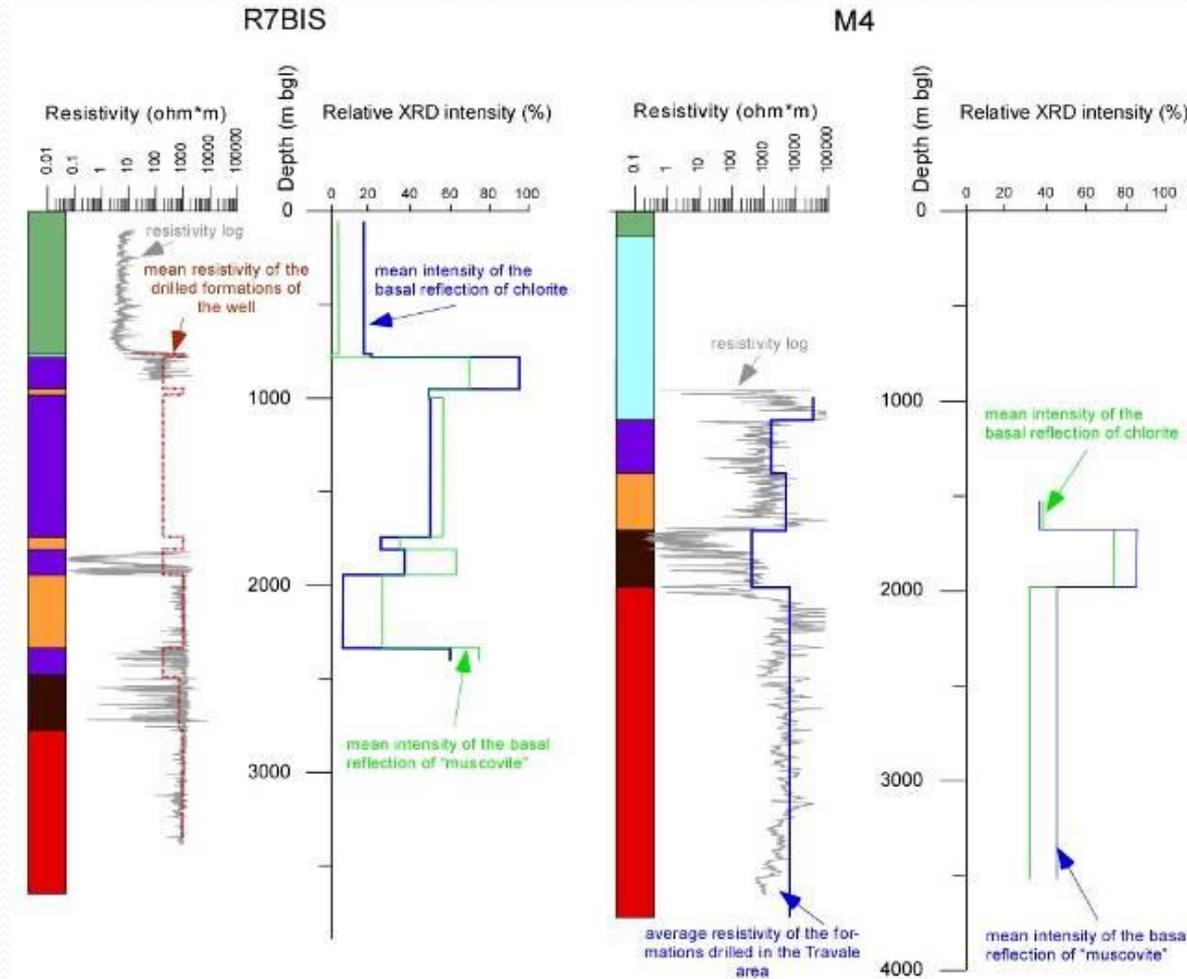
**Mt. Amiata Area**

- Deep water Dominated system
- shallow reservoir 180-220° C
- deep reservoir 300-350° C

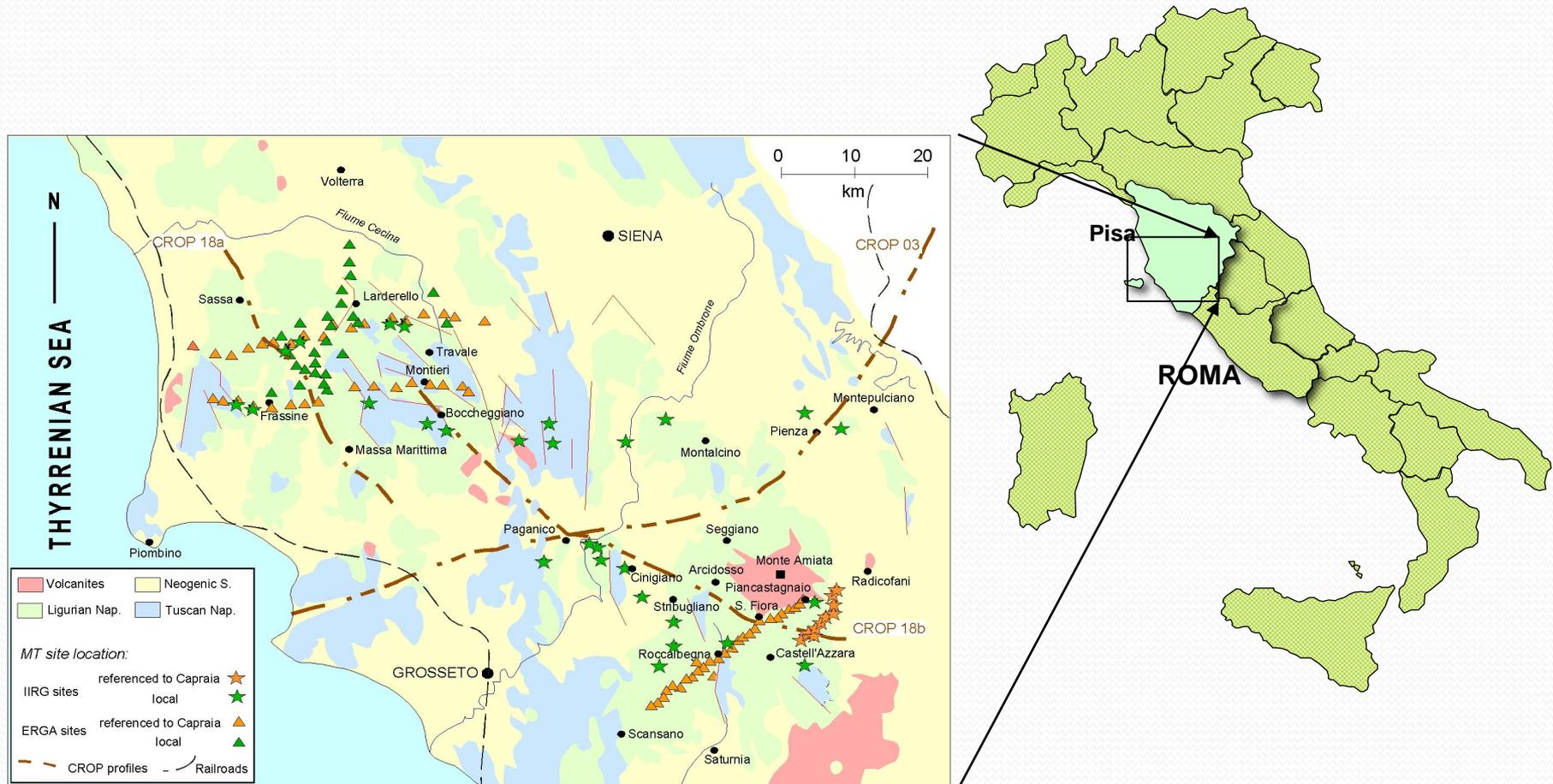


# Case Study: Italy

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

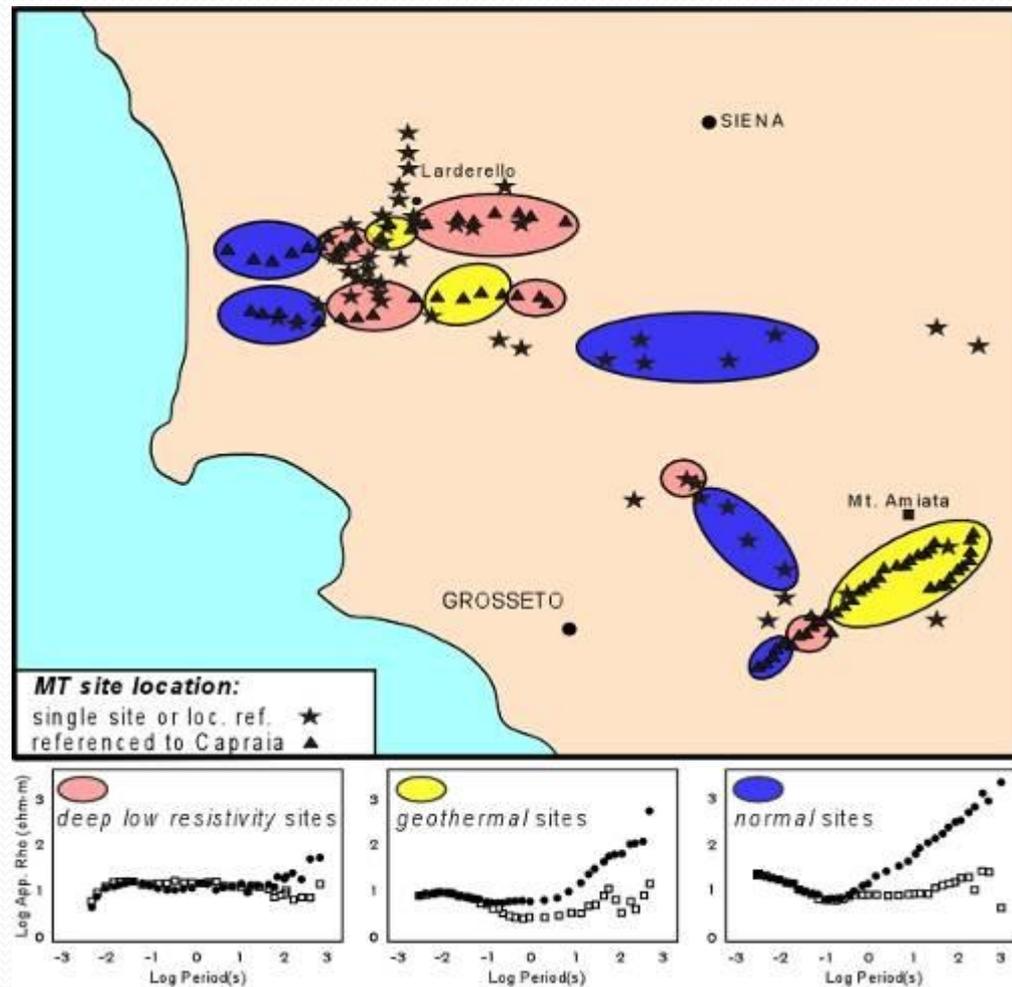


Many MT data have been collected in the last two decades, both for crustal and geothermal investigation, facing a strong EM coherent noise problem (remote reference and robust processing are a “must”)

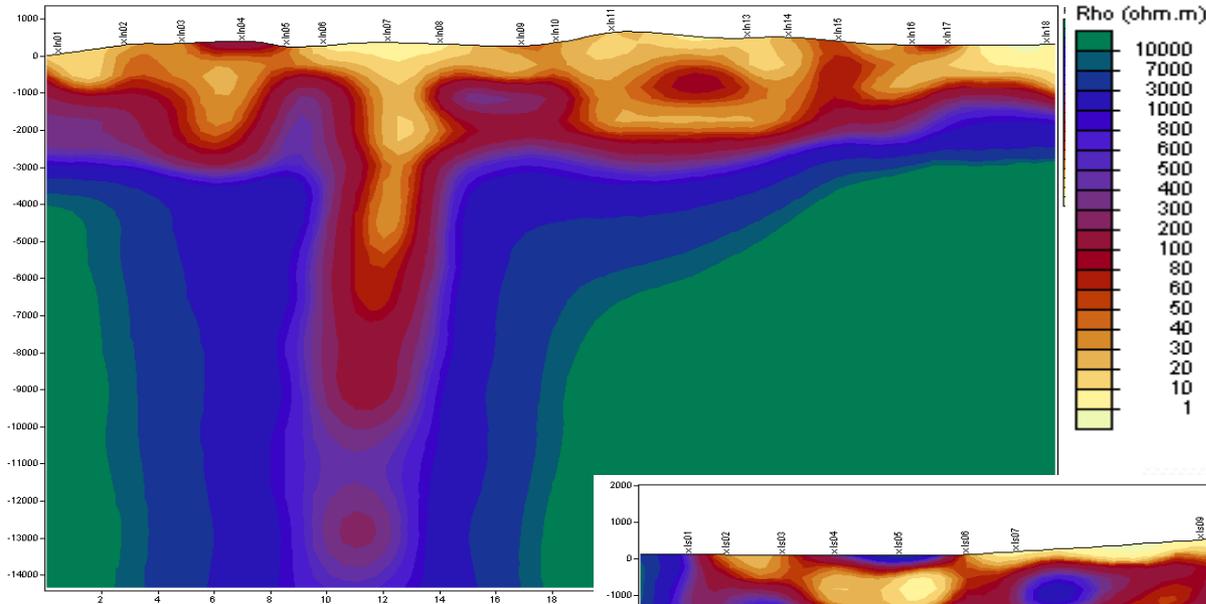


# Case Study: Italy

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

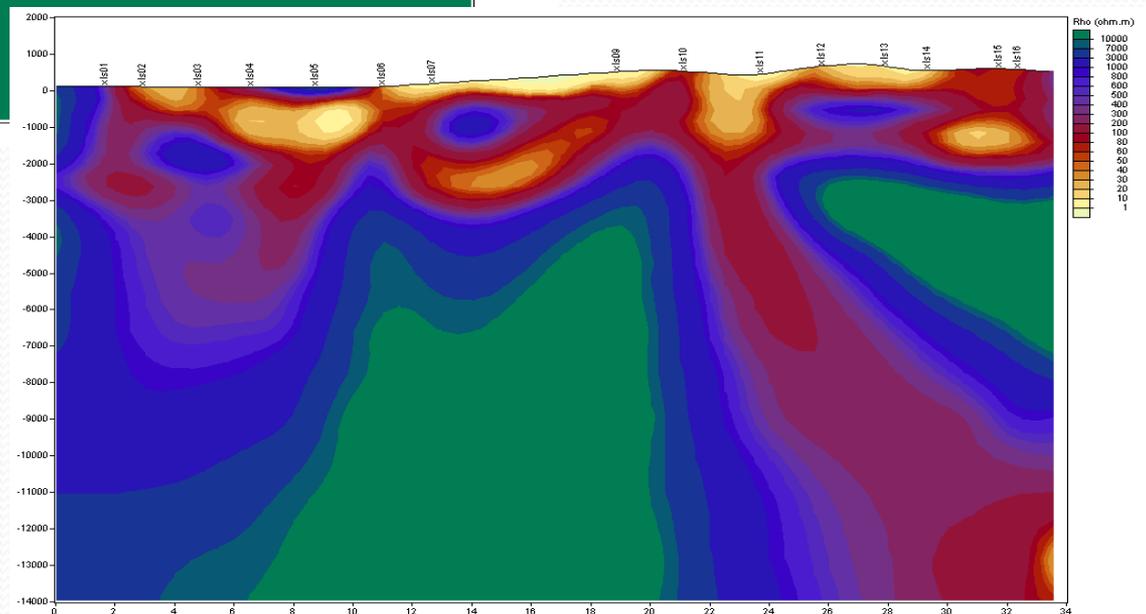


# Case Study: Italy

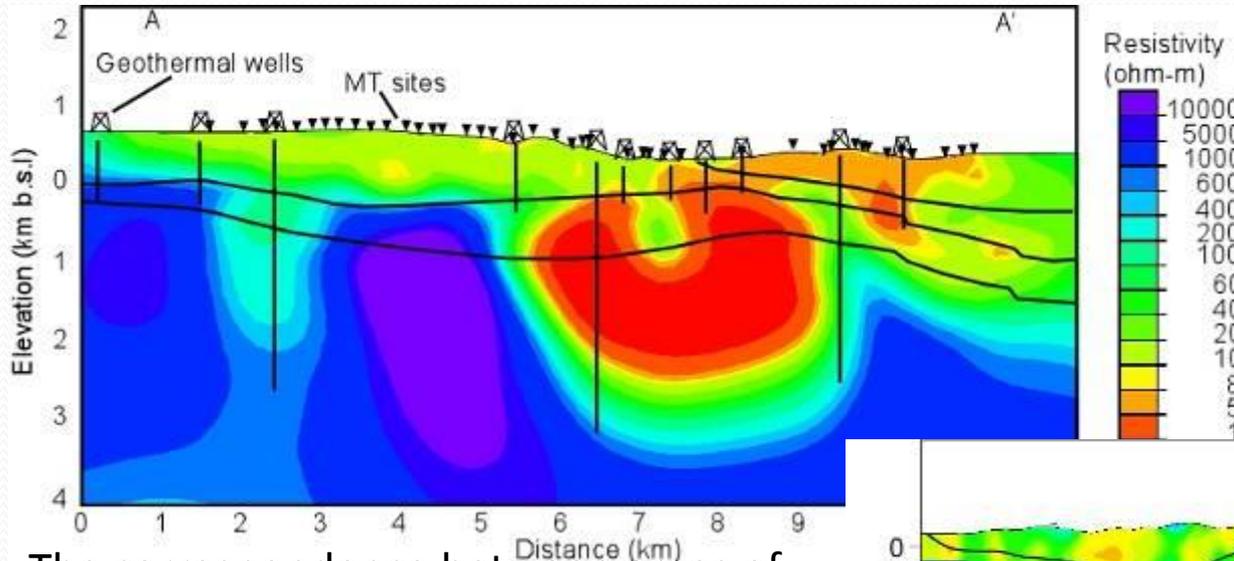


Larderello 1992

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

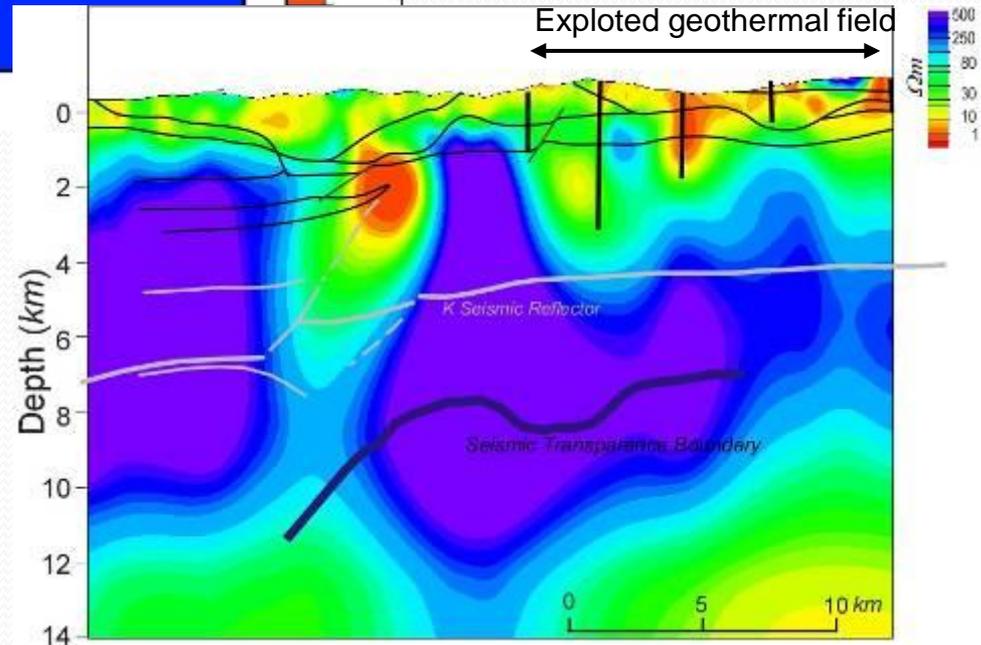


# Case Study: Italy

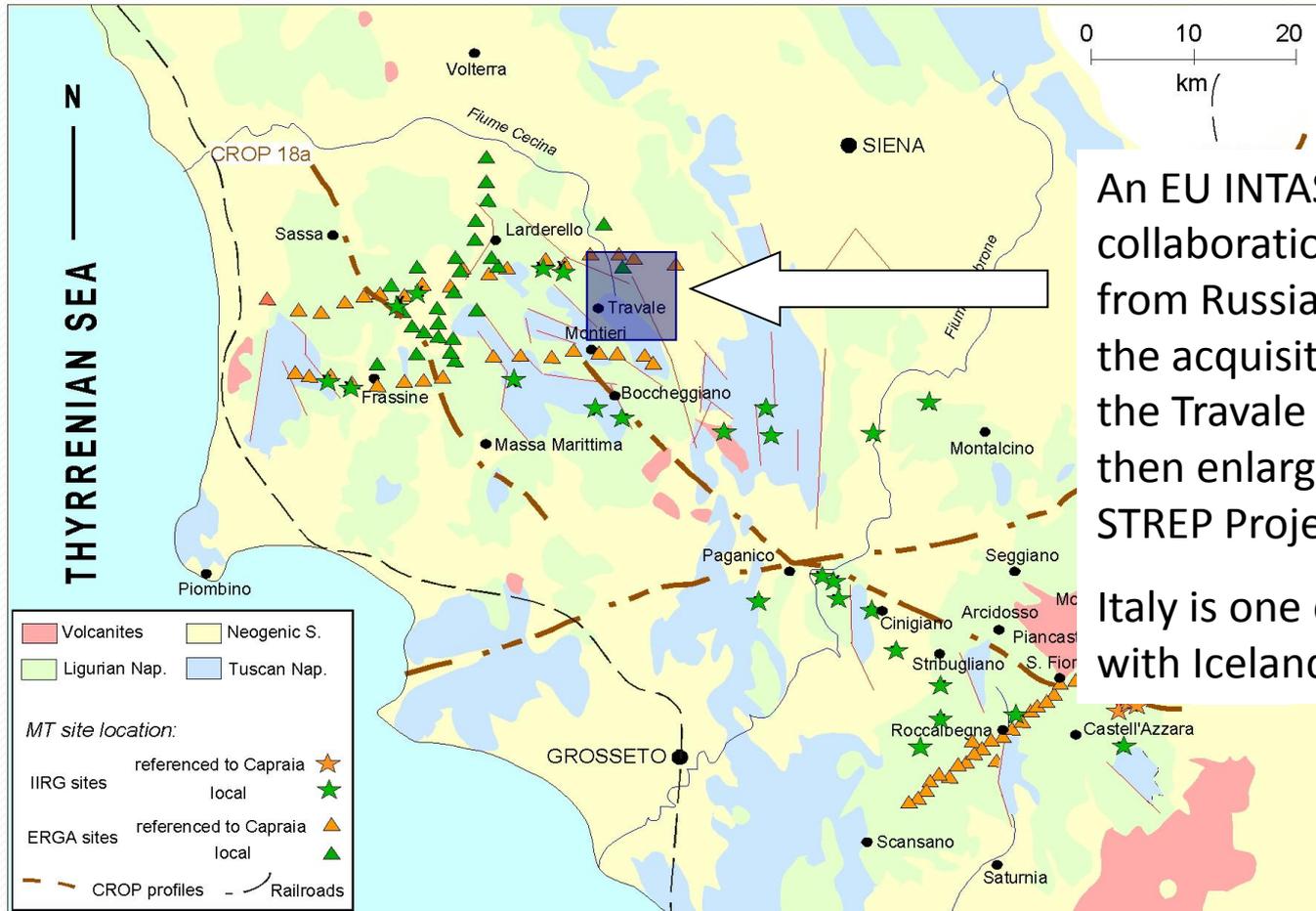


Amiata 1994 and 1998

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).

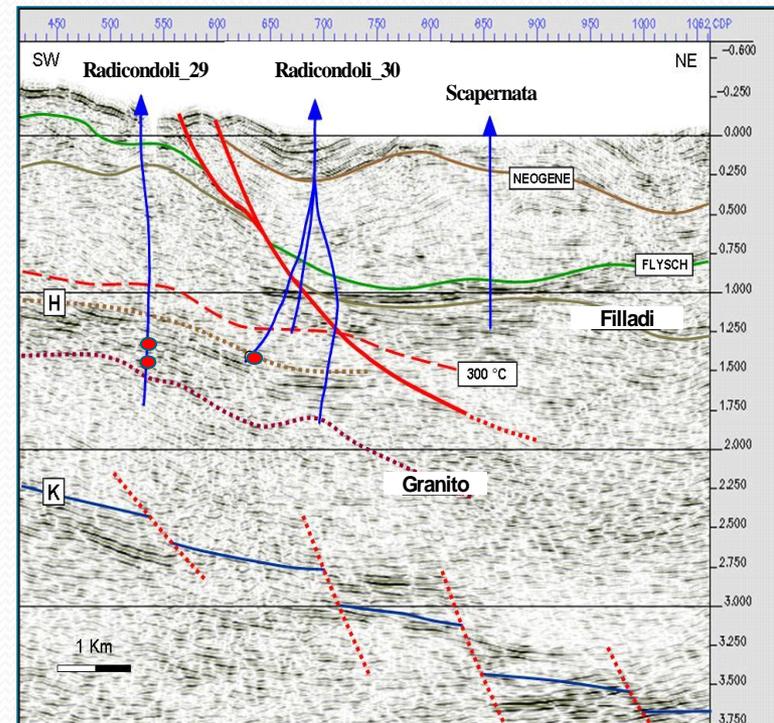
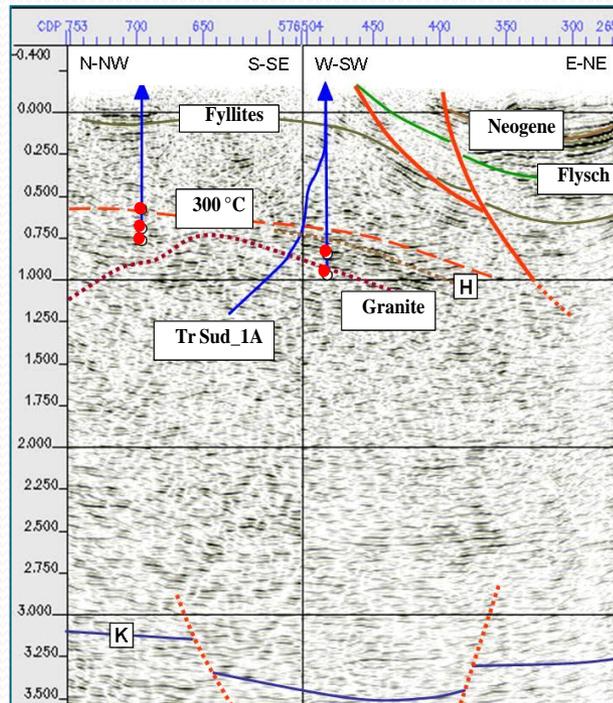
Italy is one case study, together with Iceland, Germany, Poland.



# Case Study: Italy

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

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



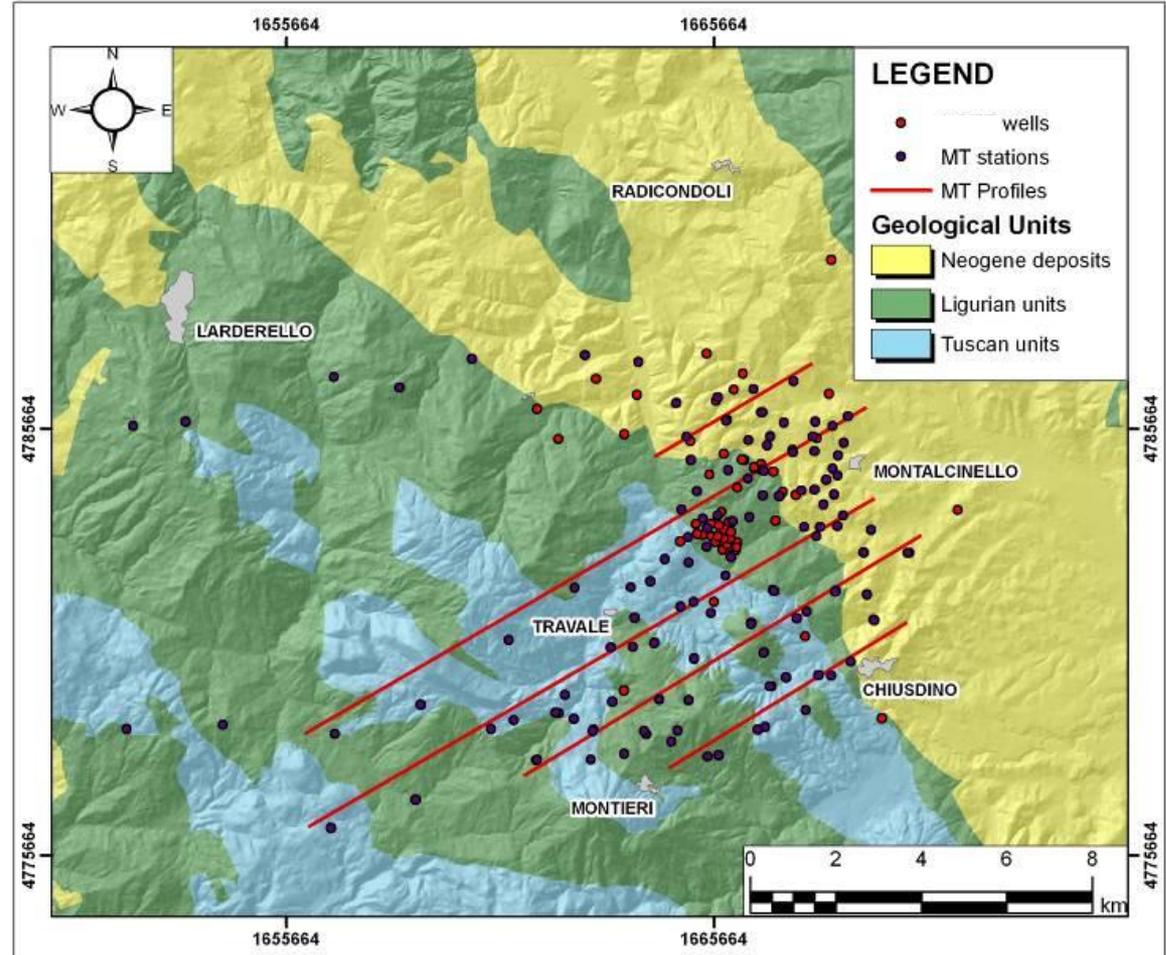
Correlation between seismic reflections (H marker) and fractures (red dots)



# 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



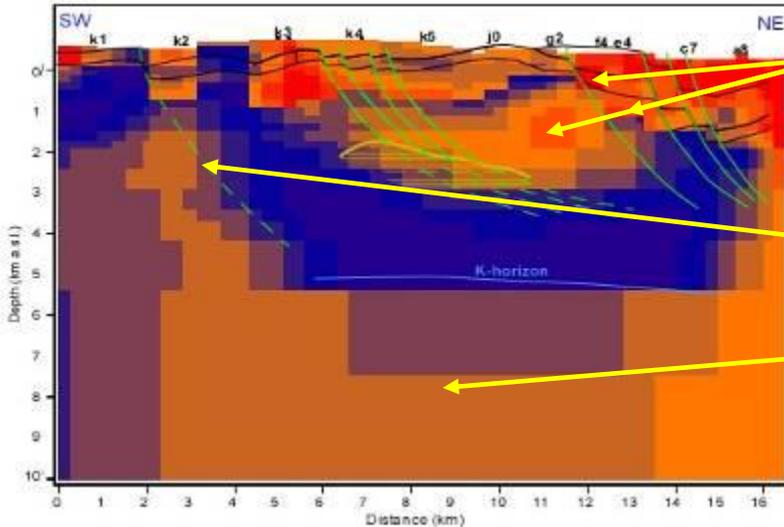
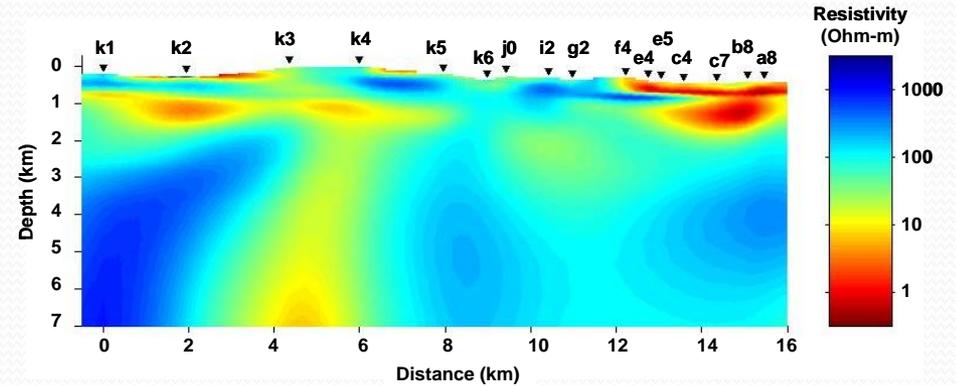
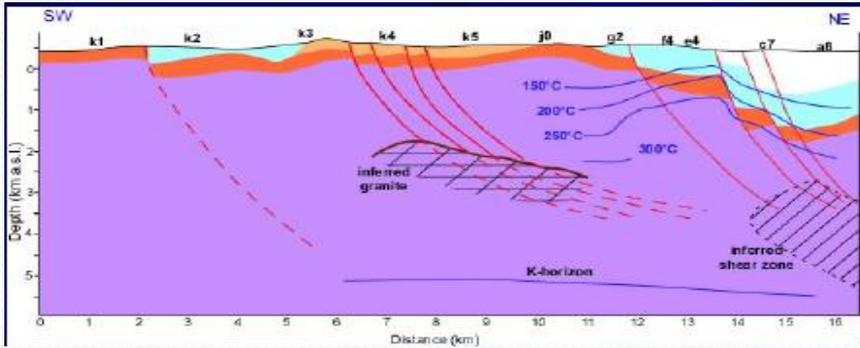
# Case Study: Italy

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



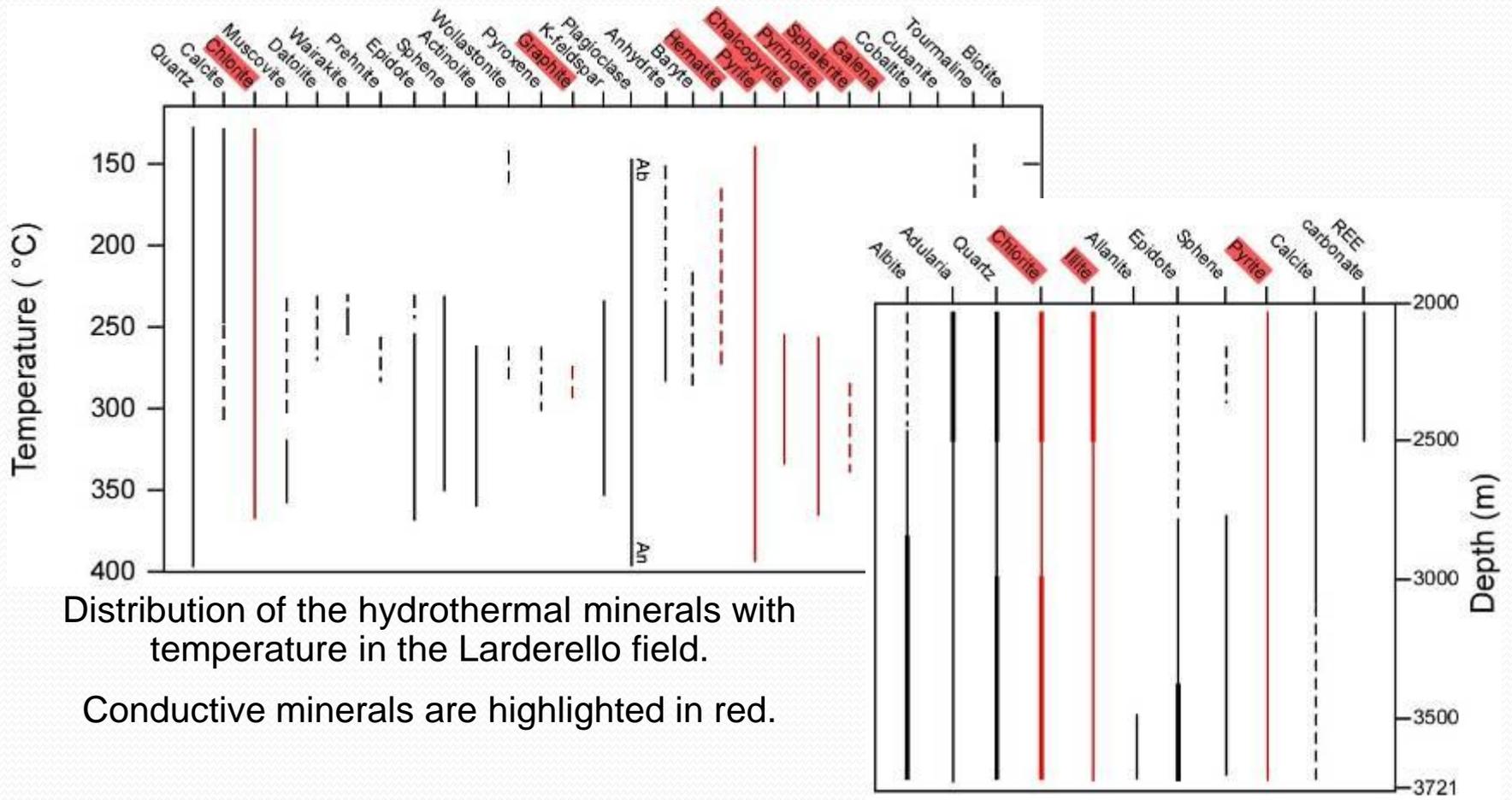
# Case Study: Italy

Two possible explanations for the resistivity reduction from  $10^3$  to  $10^0$   $\Omega\text{m}$  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.

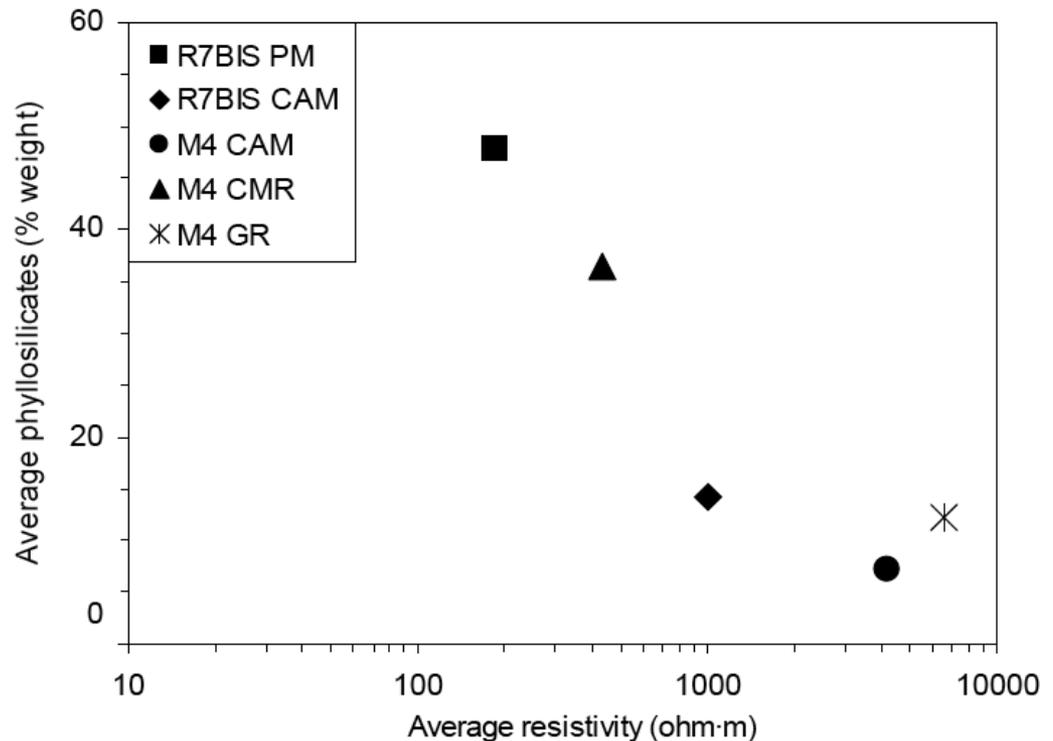
Distribution and abundance of authigenic minerals in the granite crossed by MONT-4 well.

Conductive minerals are highlighted in red.



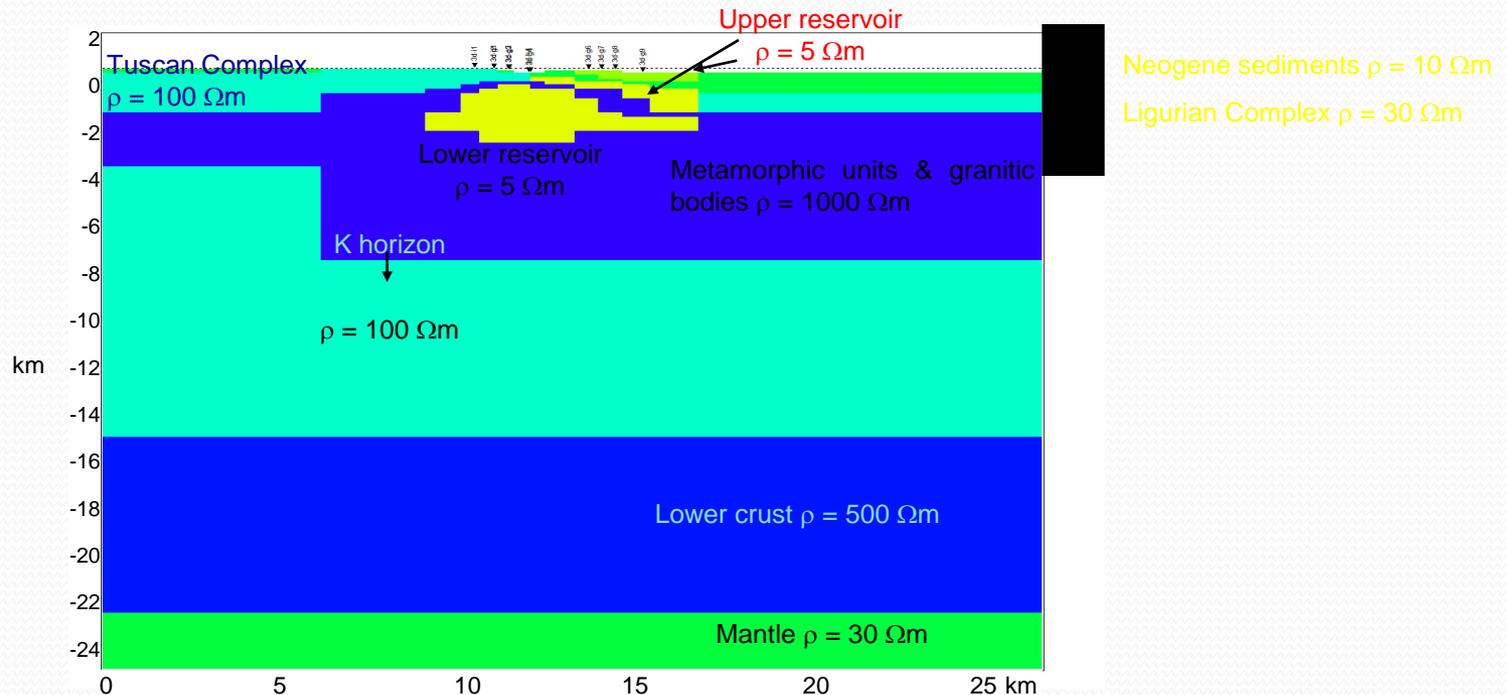
# Case Study: Italy

.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.



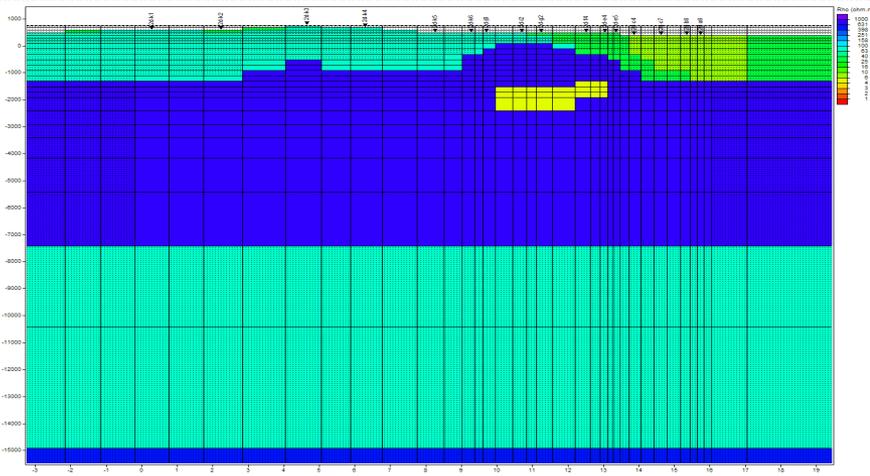
# Case Study: Italy

- 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  $\sim 60 \text{ km}^3$  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.

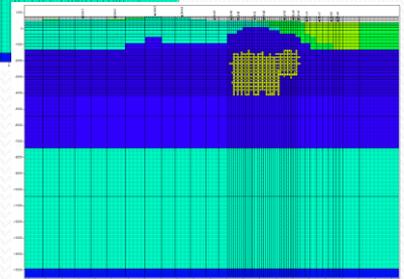
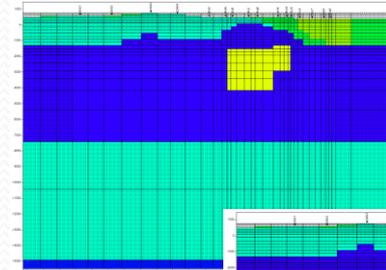


# Case Study: Italy

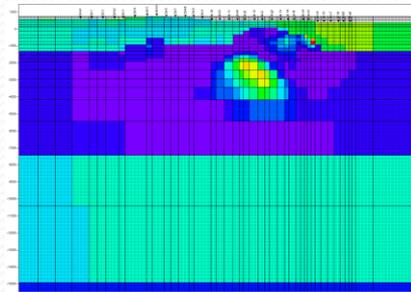
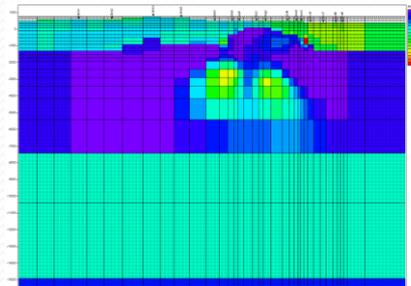
In order to test of MT resolution and optimize site density, we computed synthetic MT data by 3D forward models with different kind of reservoirs (pore-controlled-resistivity and fracture-controlled-resistivity) of various dimensions and inverted these results in 2D using *a priori* models including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



Model simulating a small (2.4 km<sup>2</sup>) pore-controlled-resistivity reservoir within a real geological section



Sparse sites



Dense sites

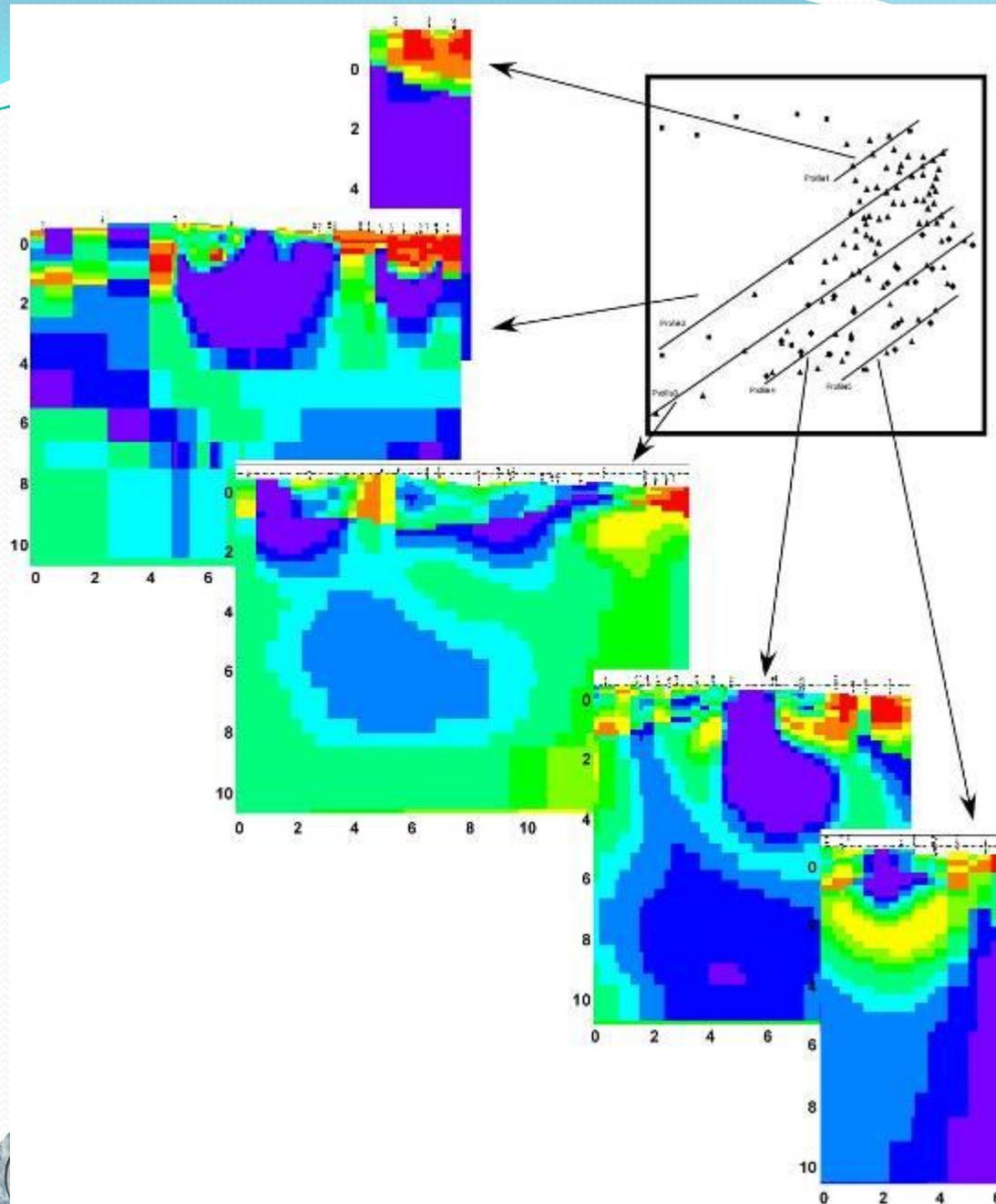


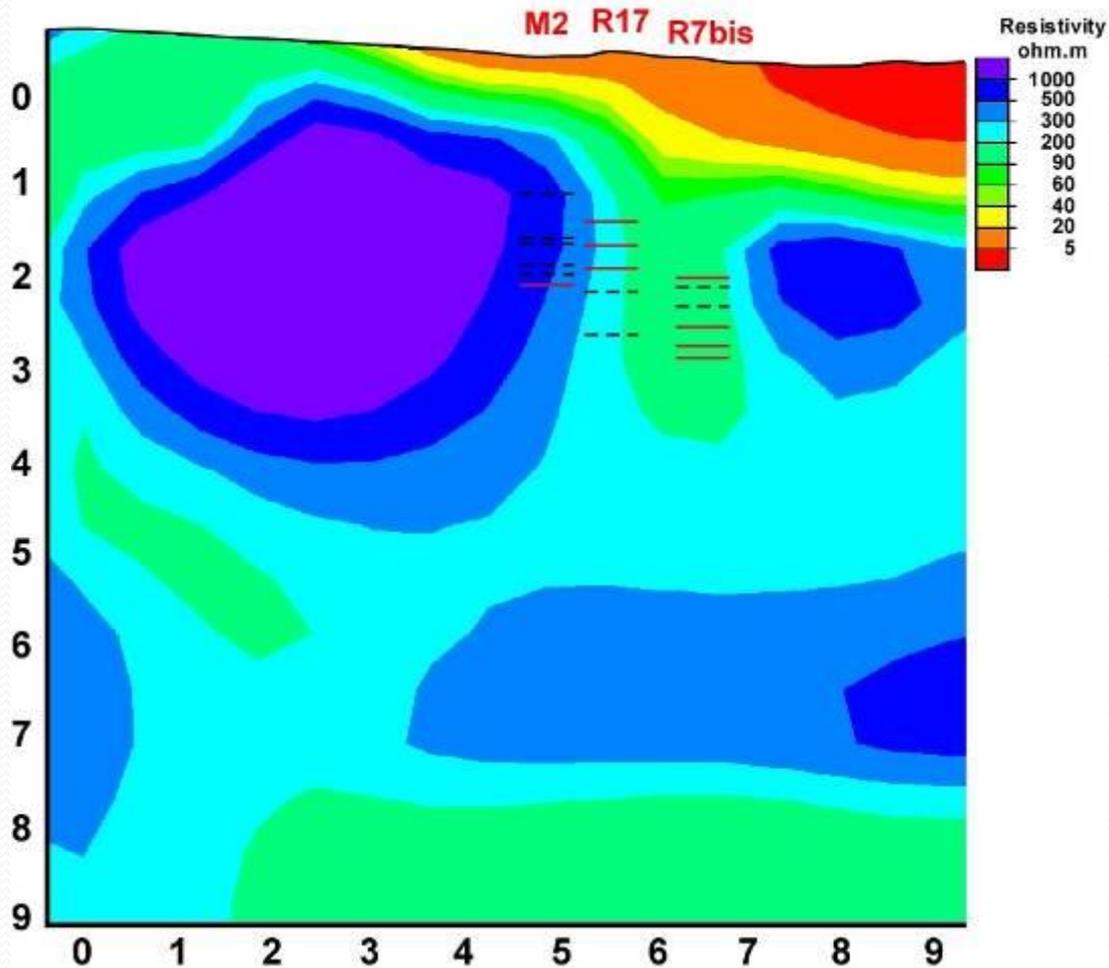
# Case Study: Italy

- 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

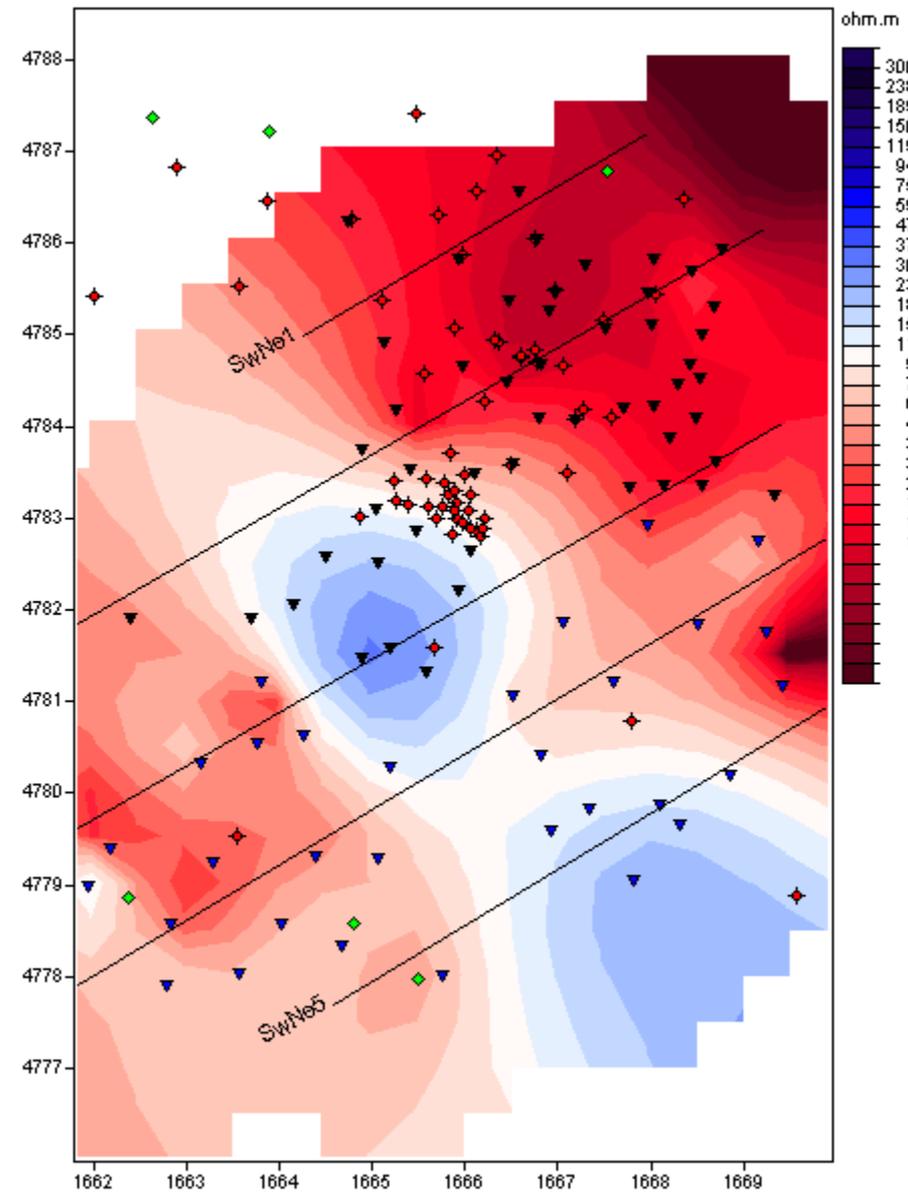
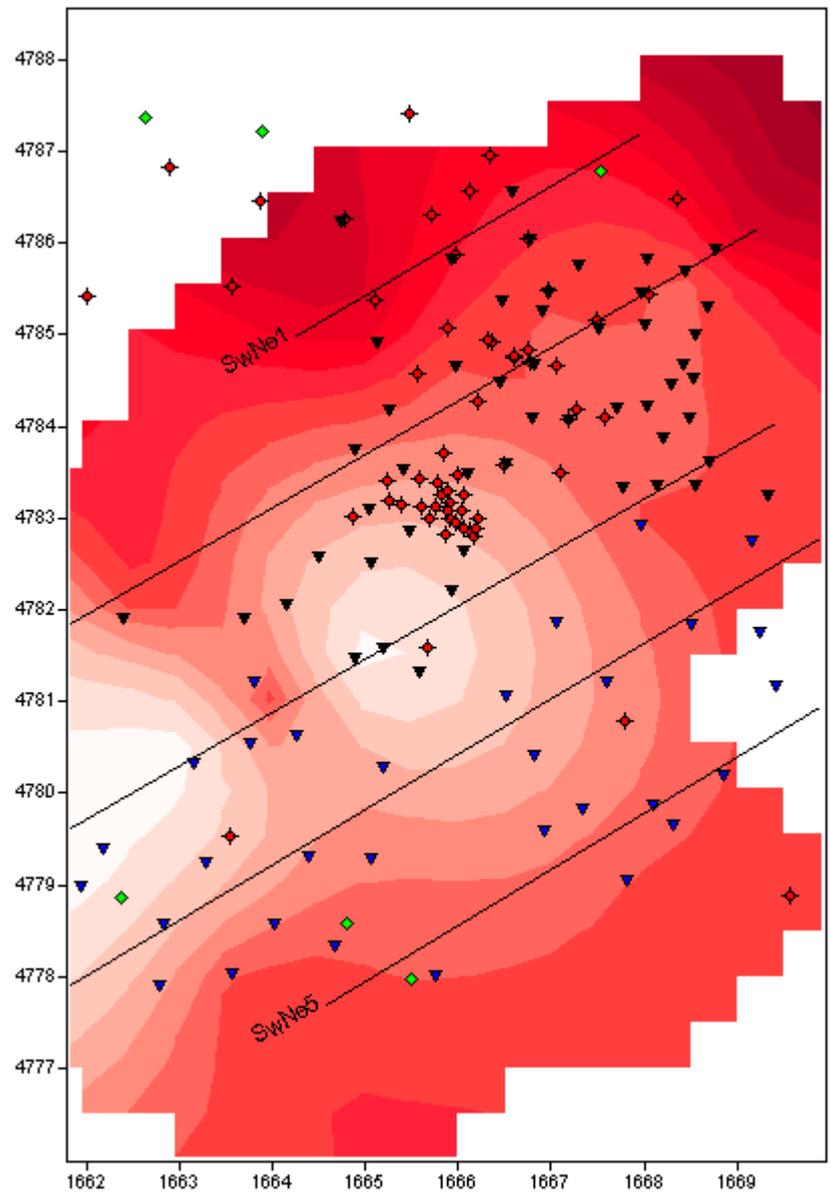




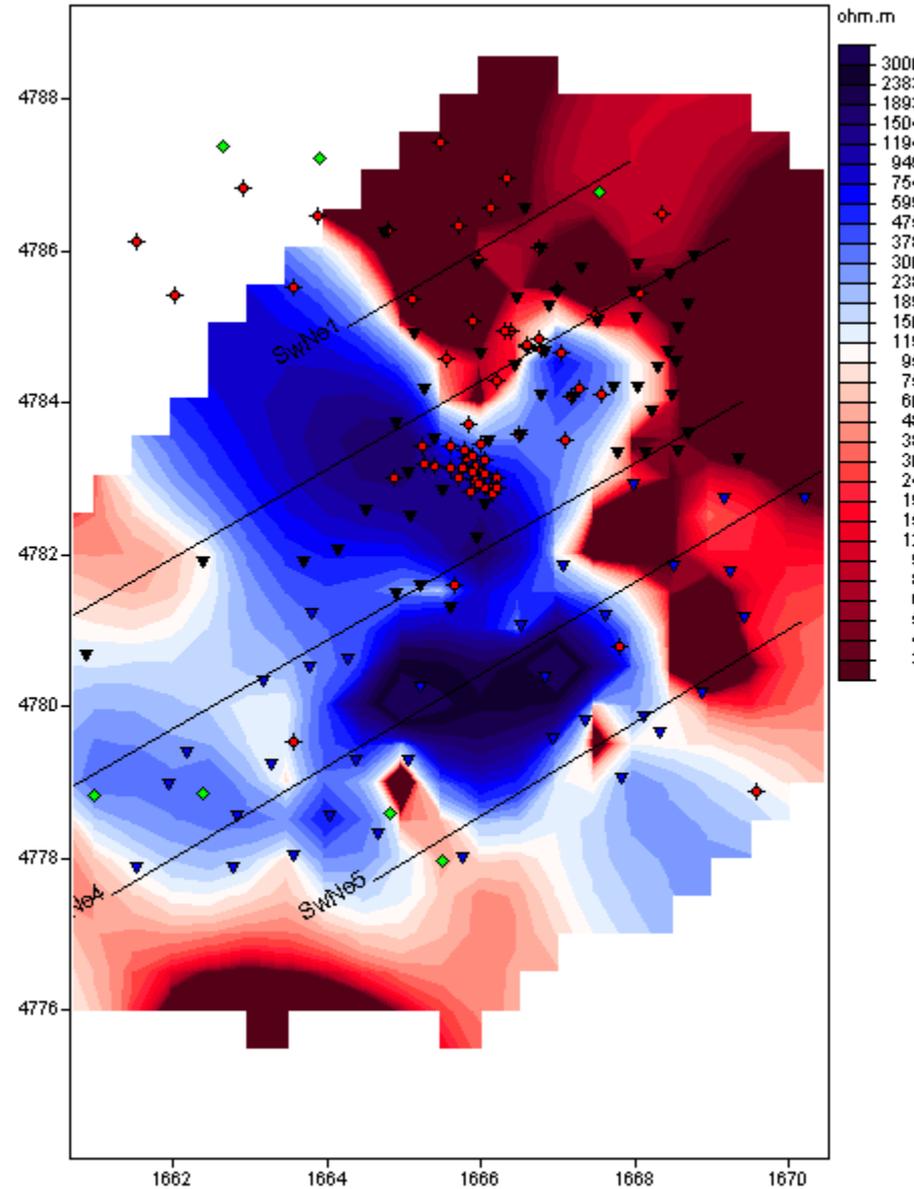
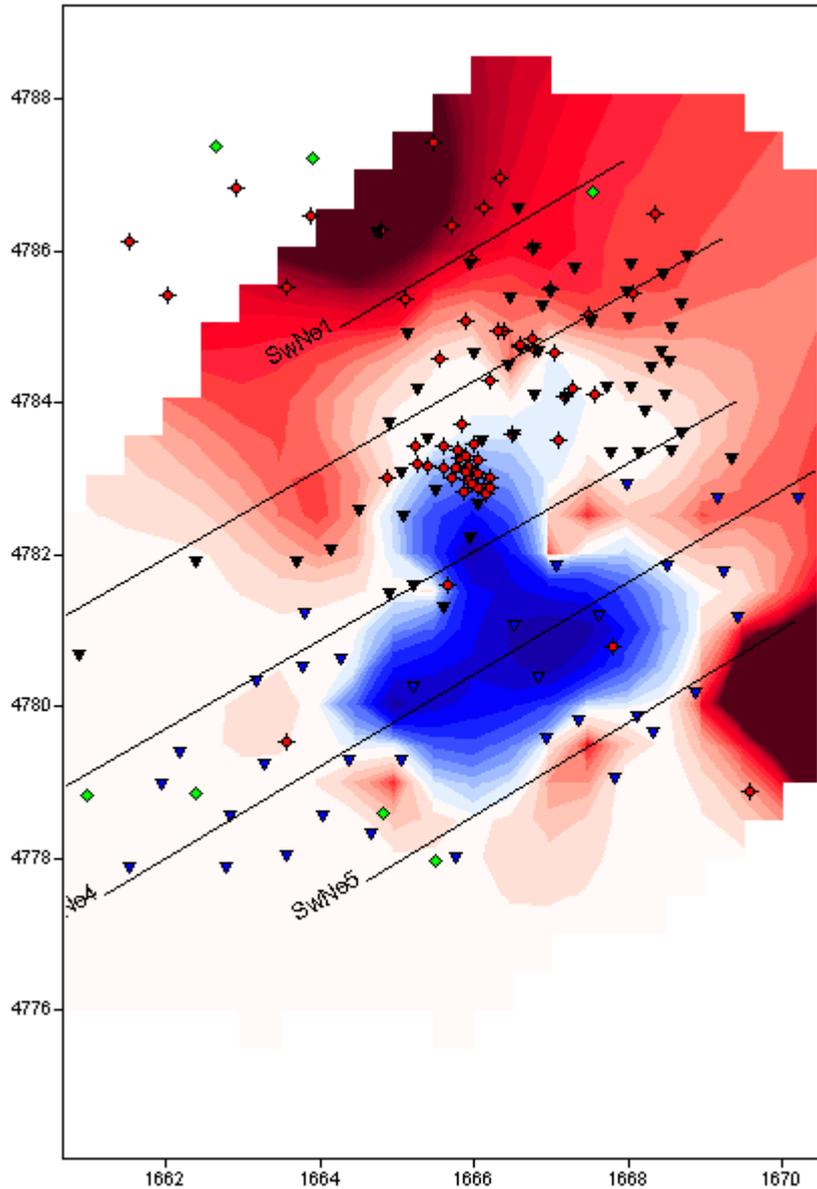
Productive fractures can not be defined just on the base of MT data. However, MT might provide info regarding areas of enhanced productivity



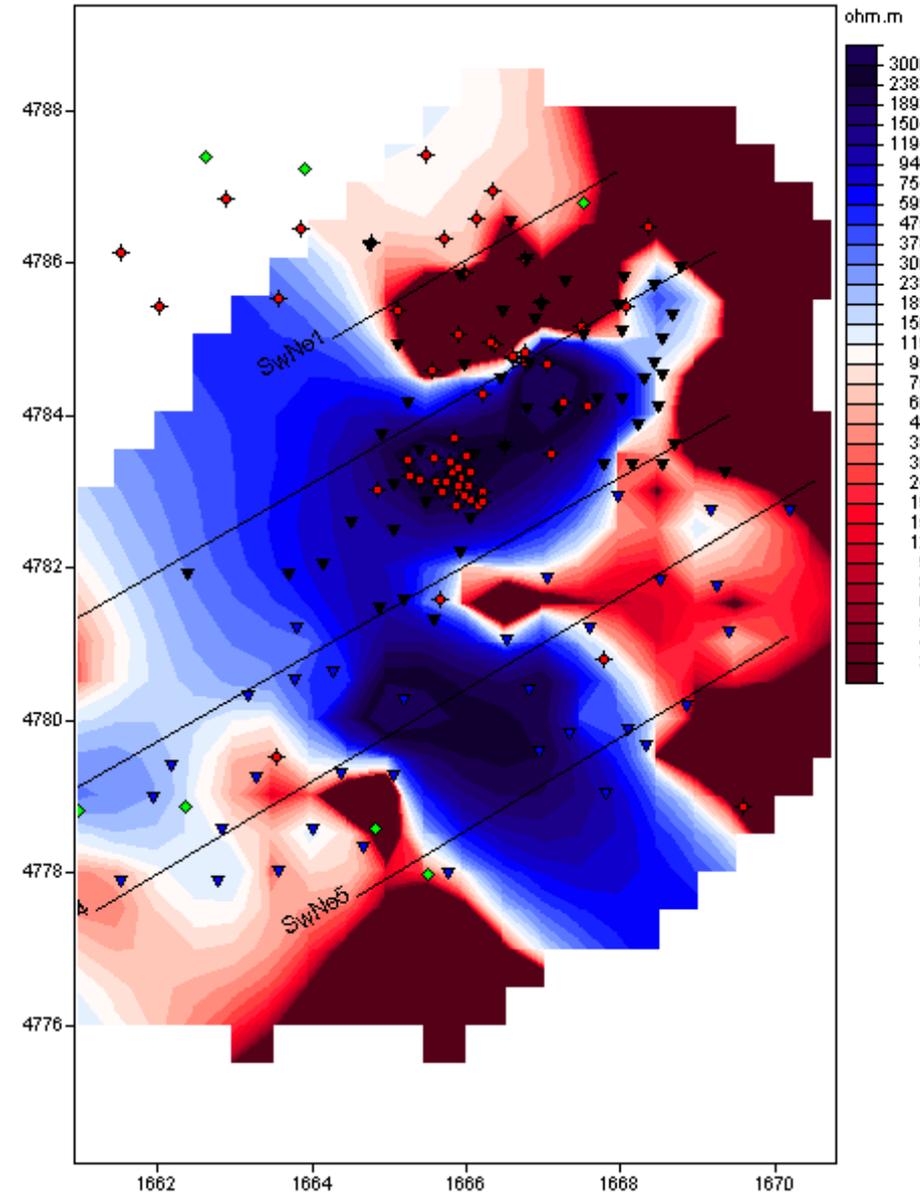
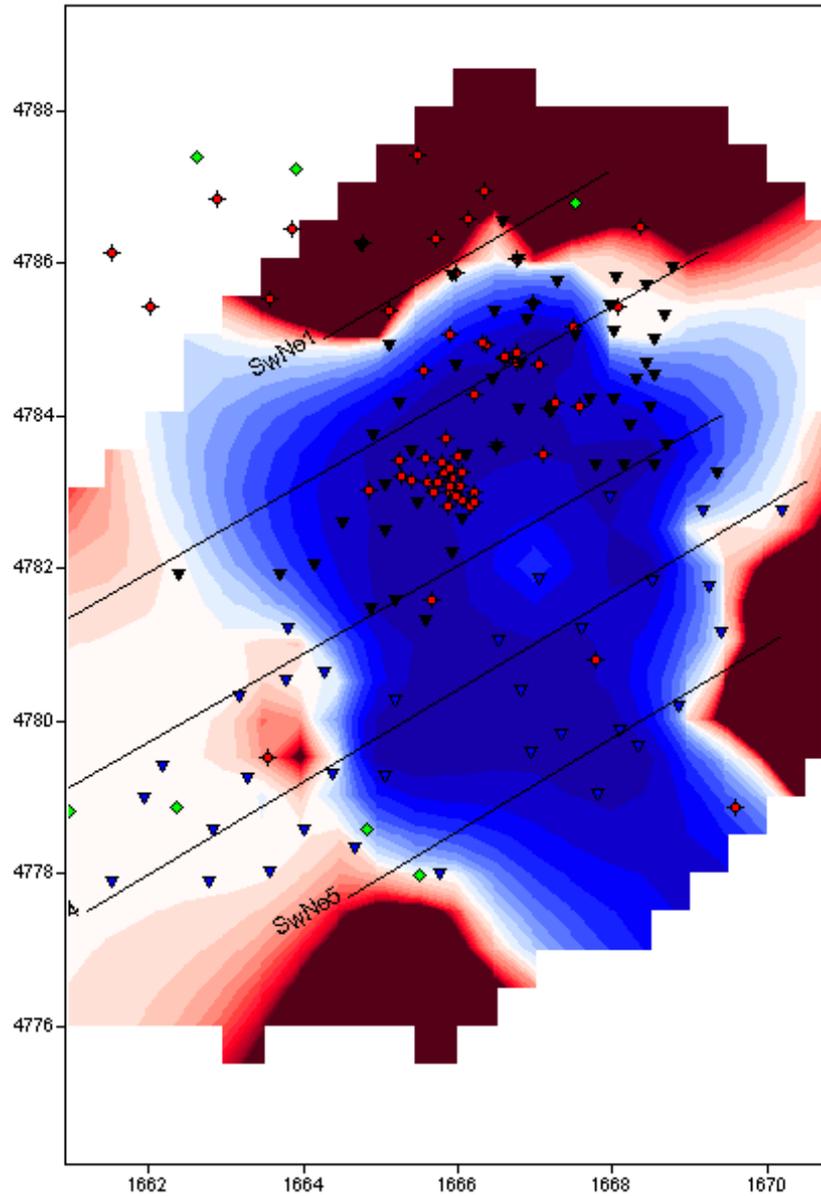
# 0 km b.s.l.



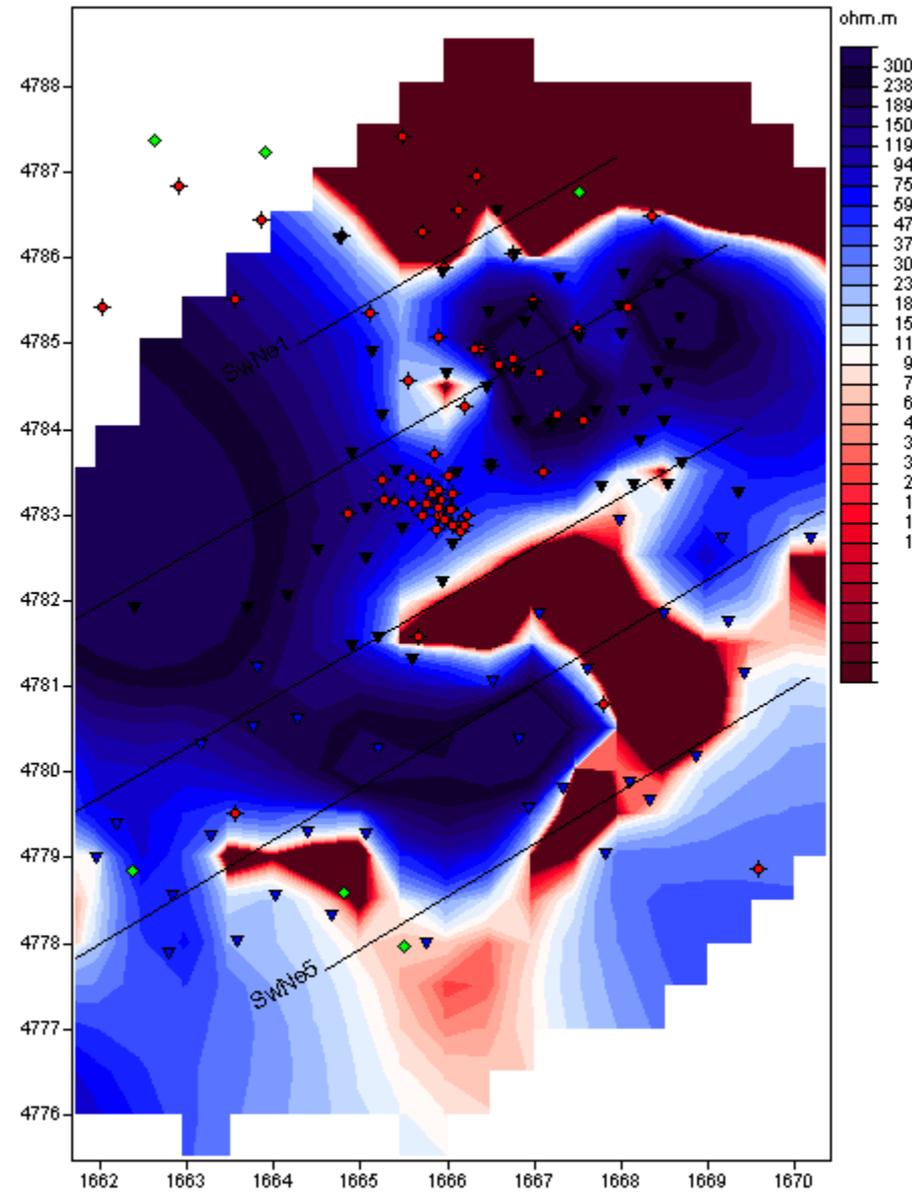
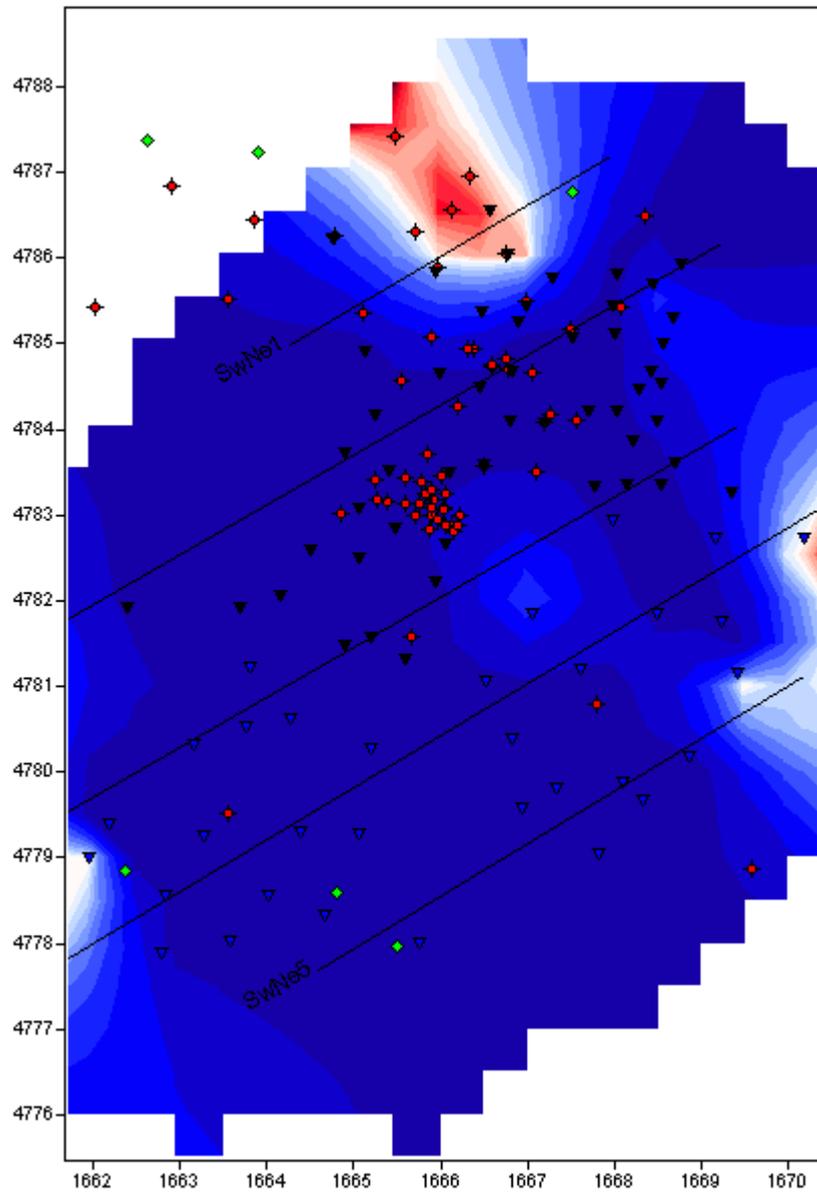
# 0.5 km b.s.l.



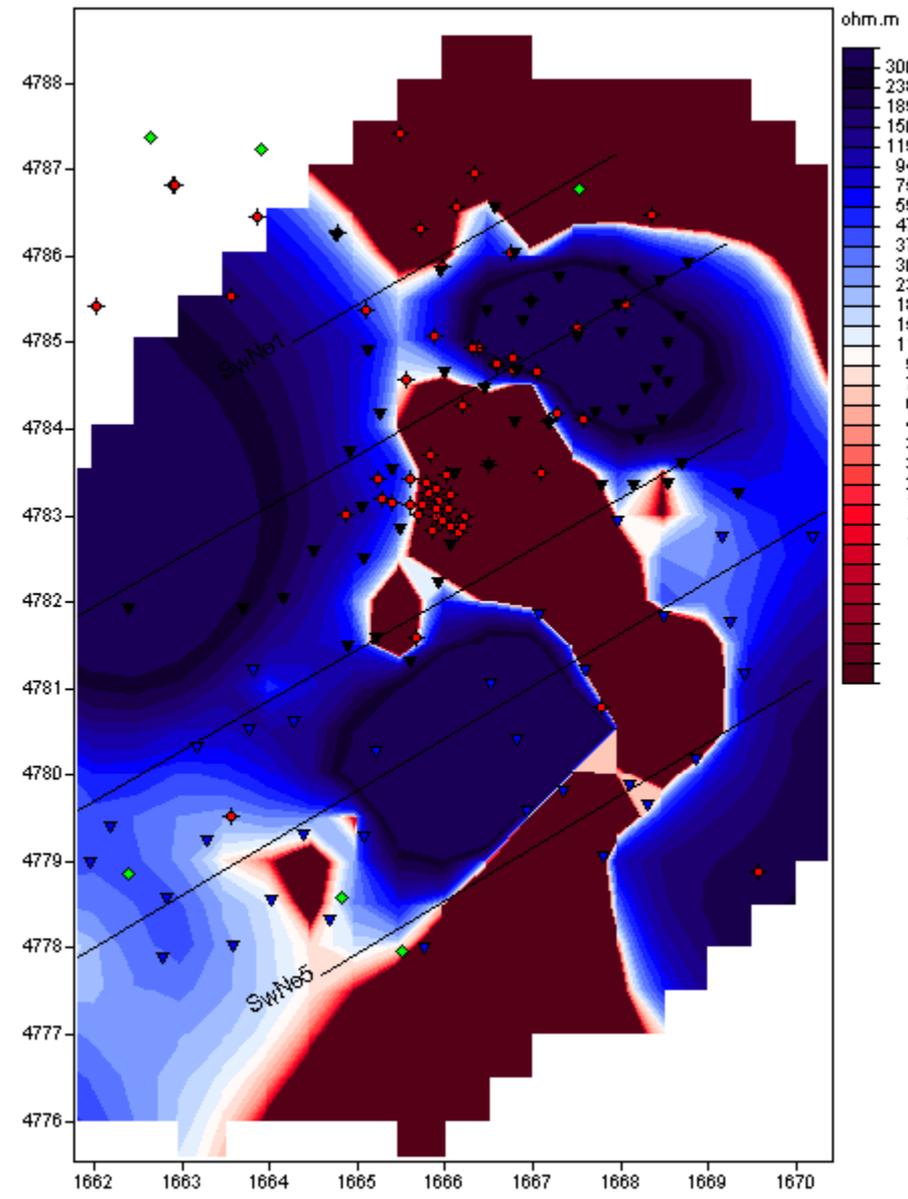
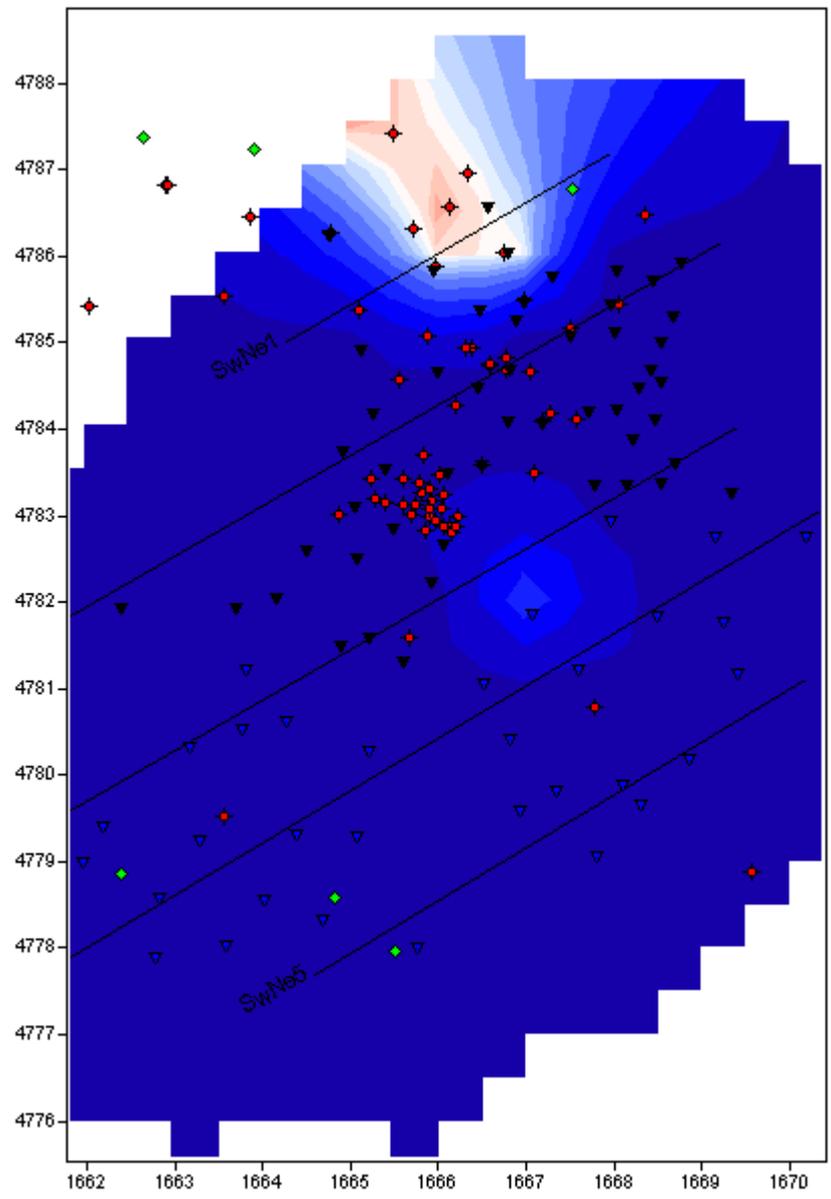
# 1 km b.s.l.



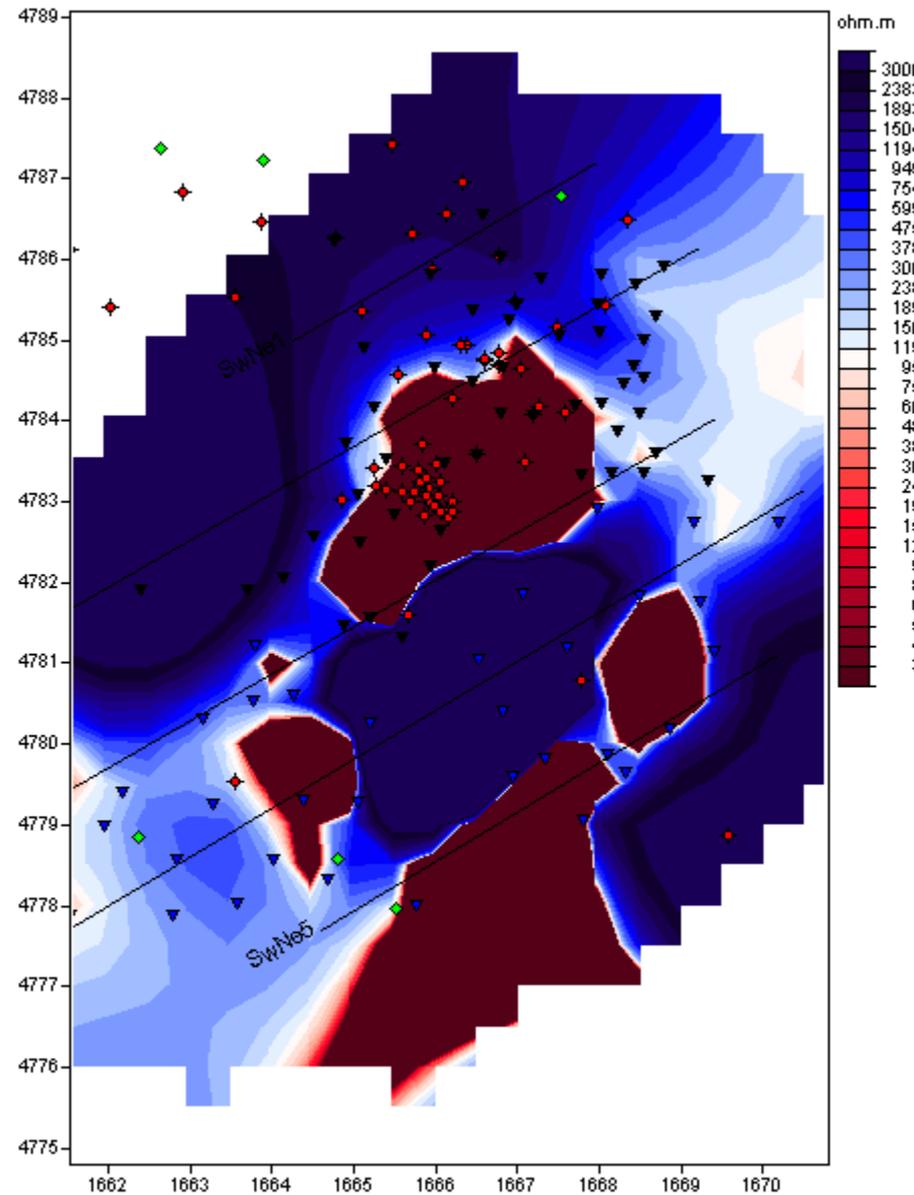
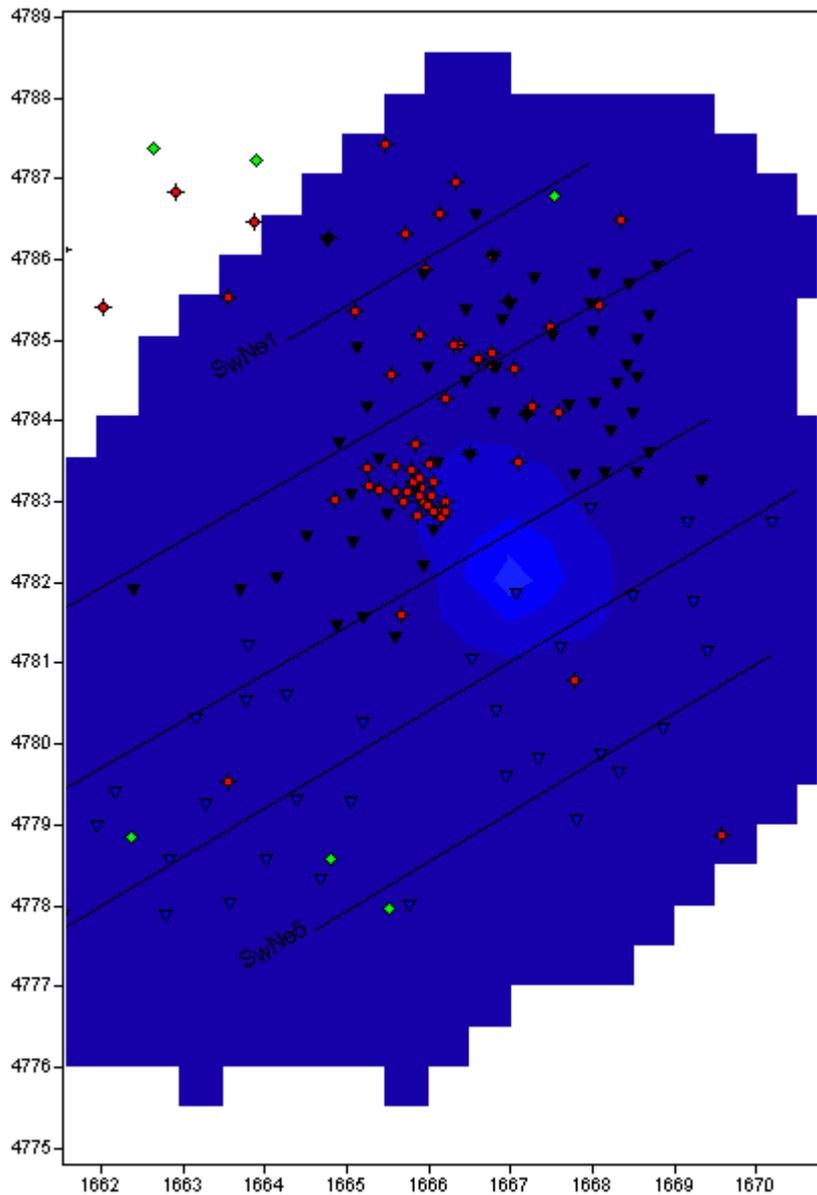
# 1.5 km b.s.l.



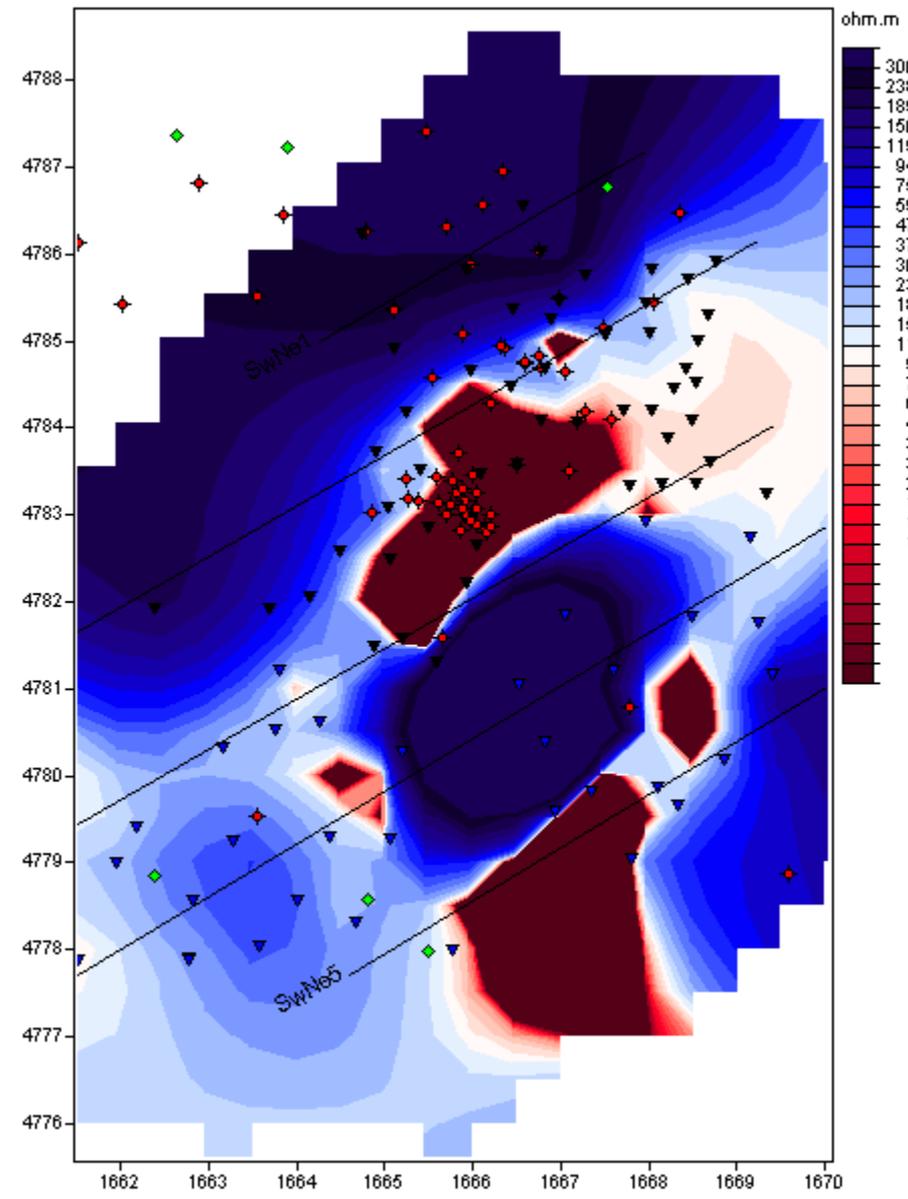
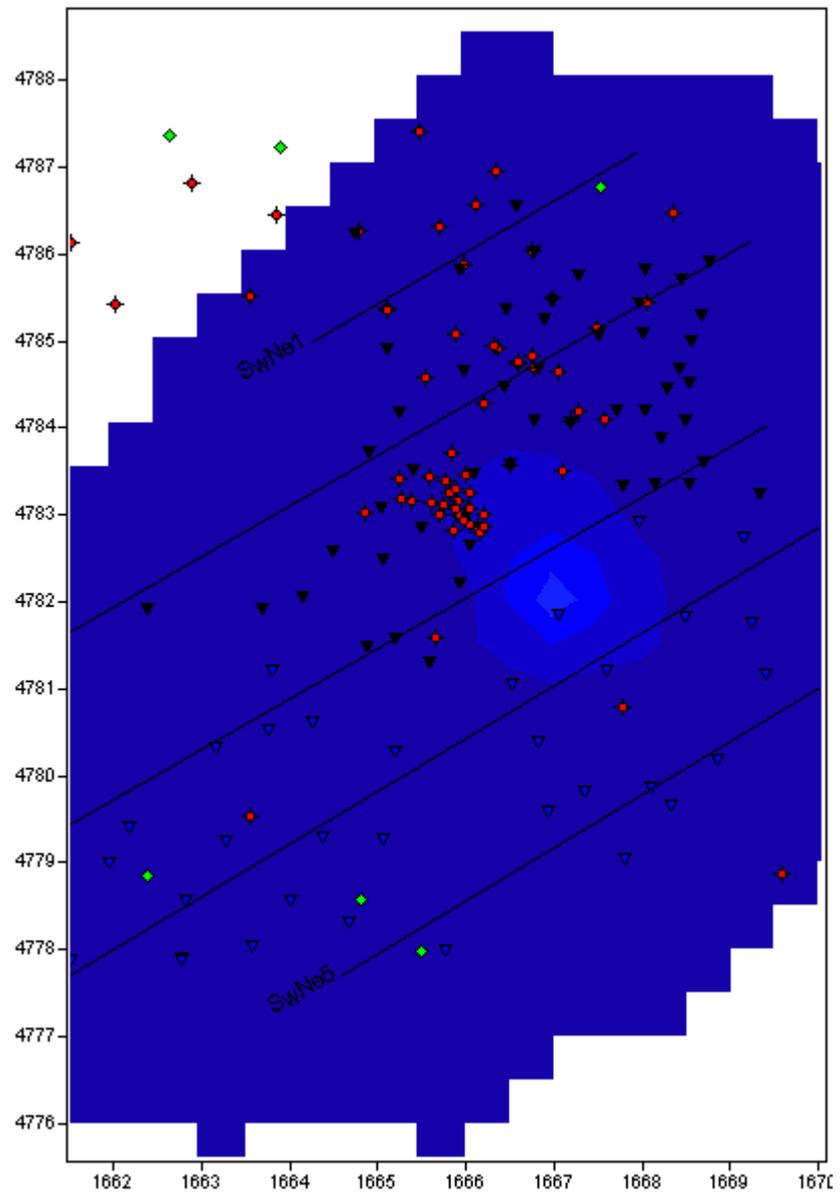
# 2 km b.s.l.



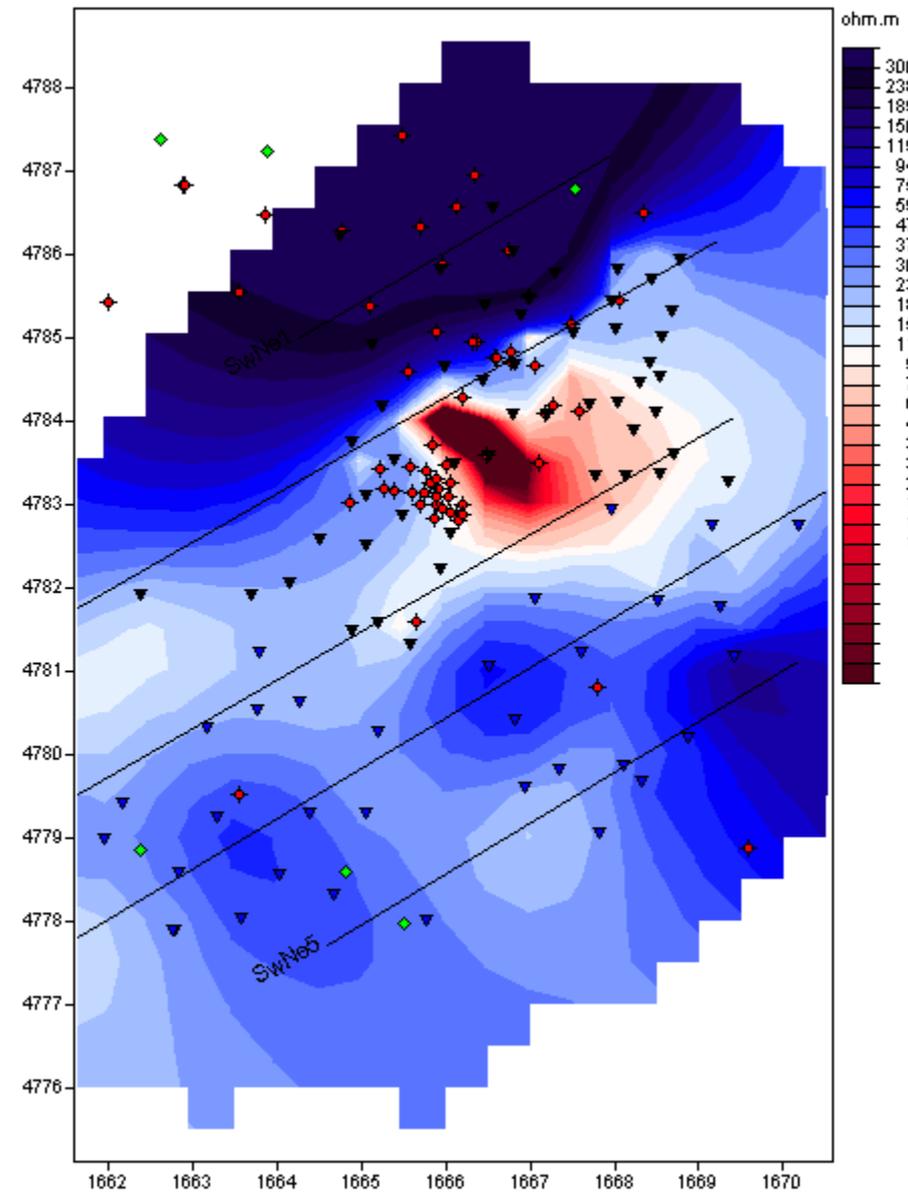
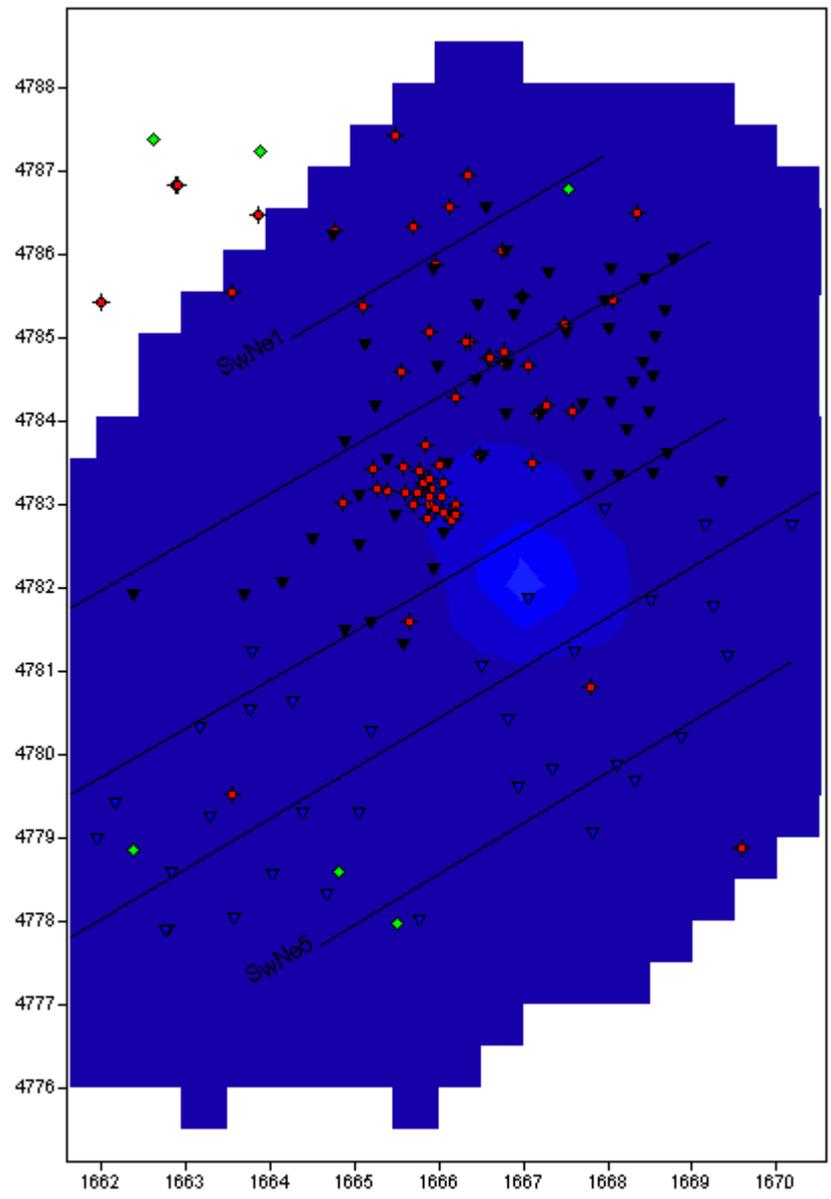
# 3 km b.s.l.



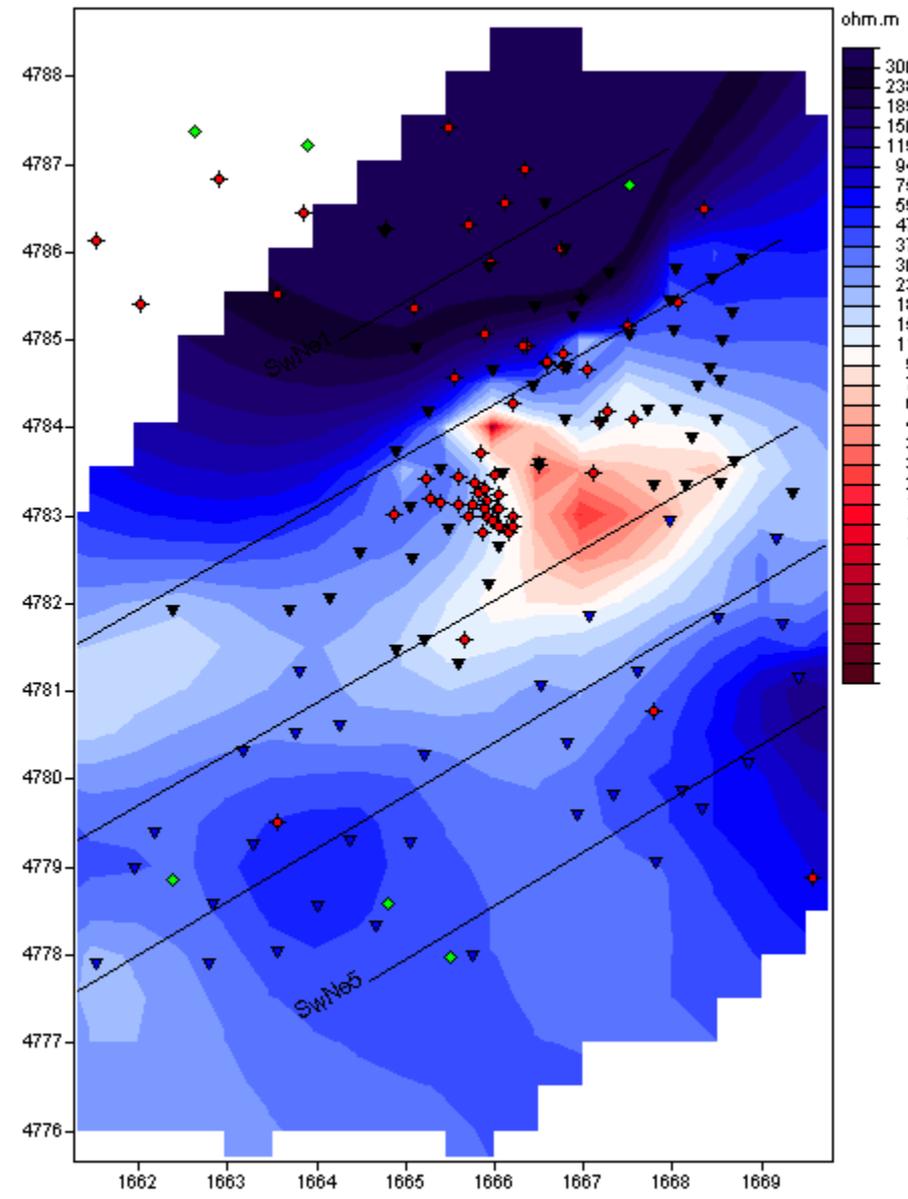
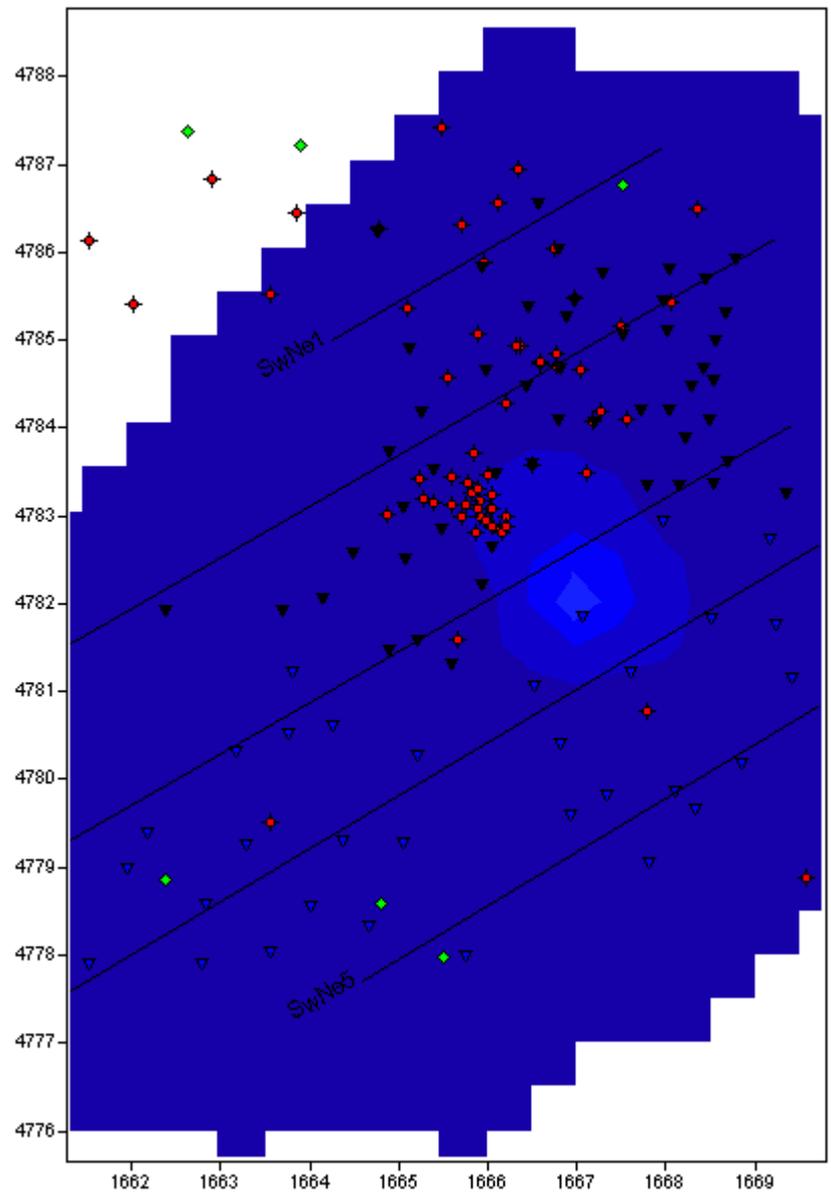
# 4 km b.s.l.



# 5 km b.s.l.



# 6 km b.s.l.



# Applications

Various targets can be imaged by resistivity geophysical methods

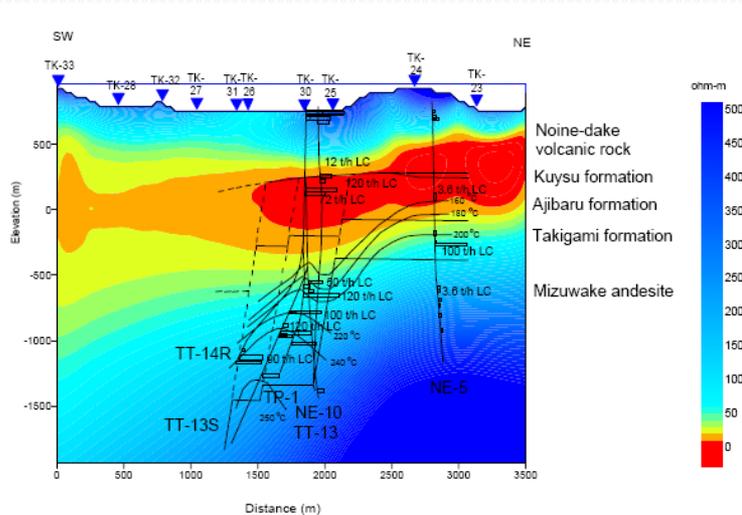
- Regional structure (geothermal system)
- Fracture detection
- Monitoring



# 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 part, in good agreement with the geological feature”

### Advantages

- cheap
- recognize fluid filled volumes

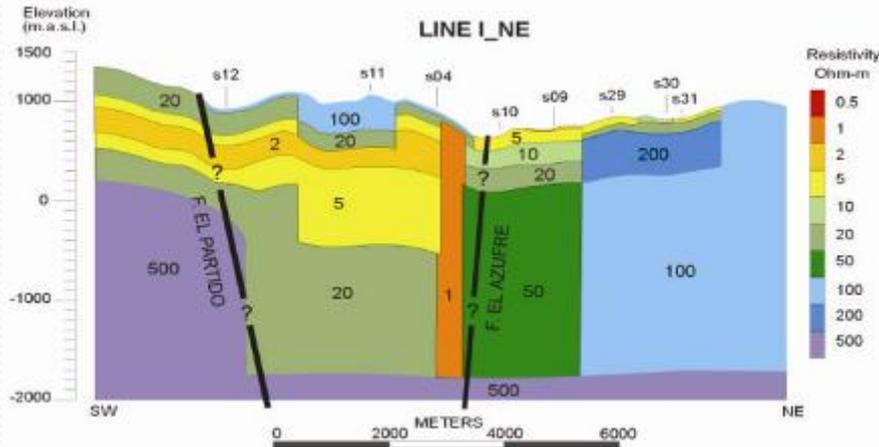
### Disadvantages

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



# Applications

## Regional exploration

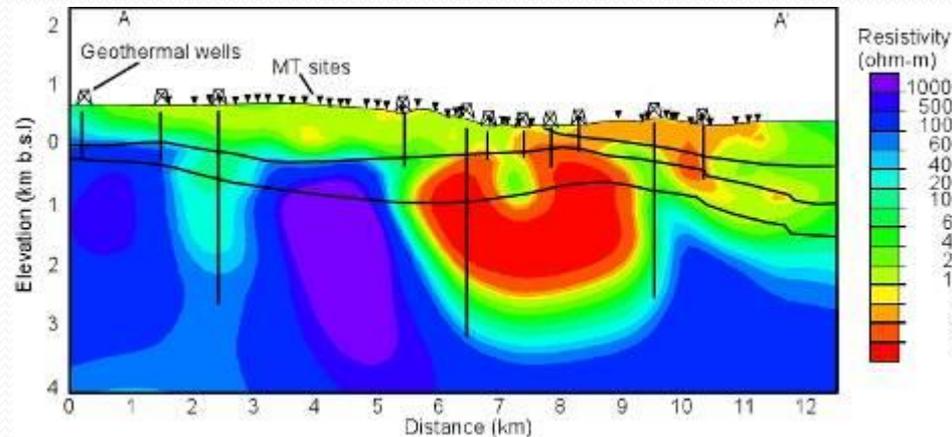


### Las Tres Virgenes Geothermal Area, Mexico

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

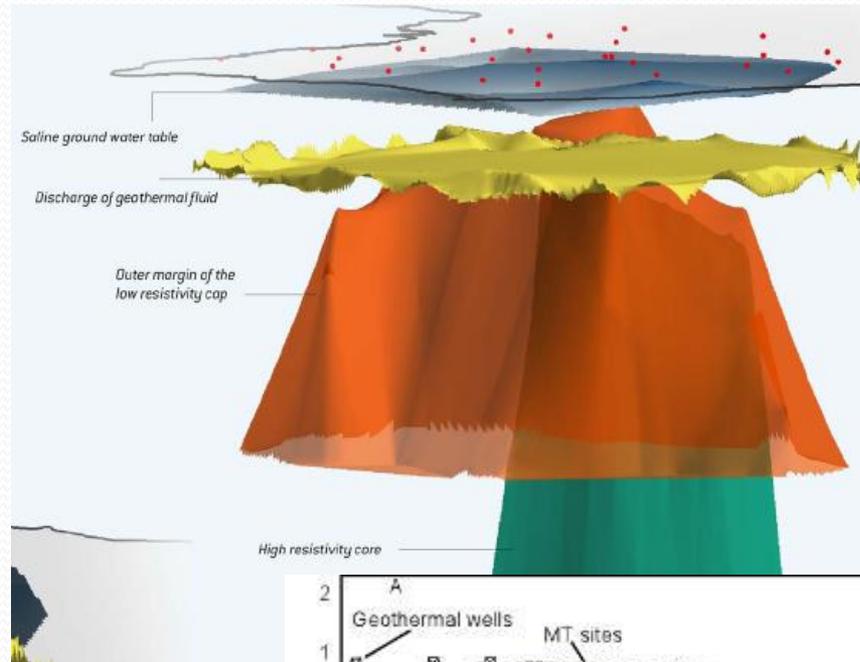


# Applications

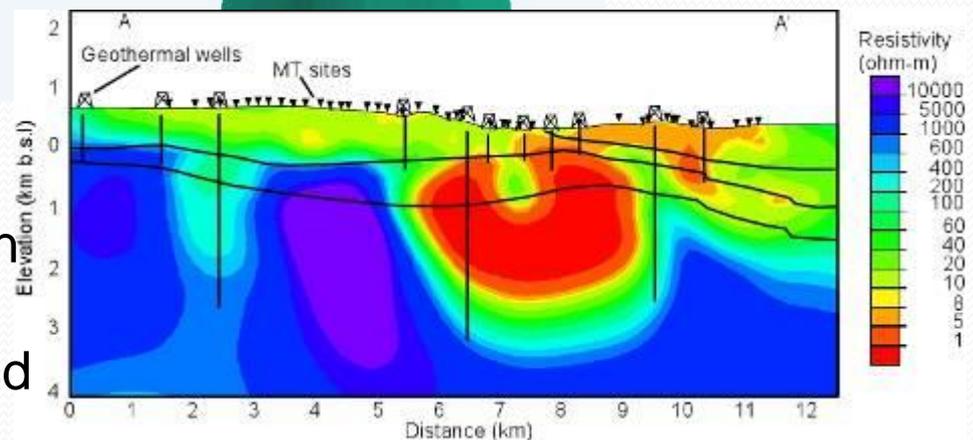
## Regional exploration

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

From Karlsdottir, ENGINE  
Workshop1



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



# Applications

## Fault/fracture detection

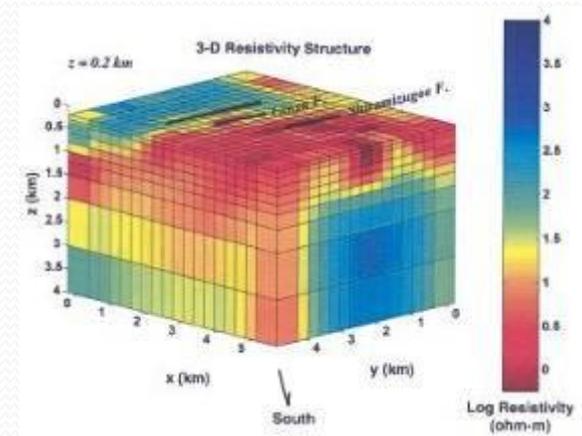
### Advantages

- 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

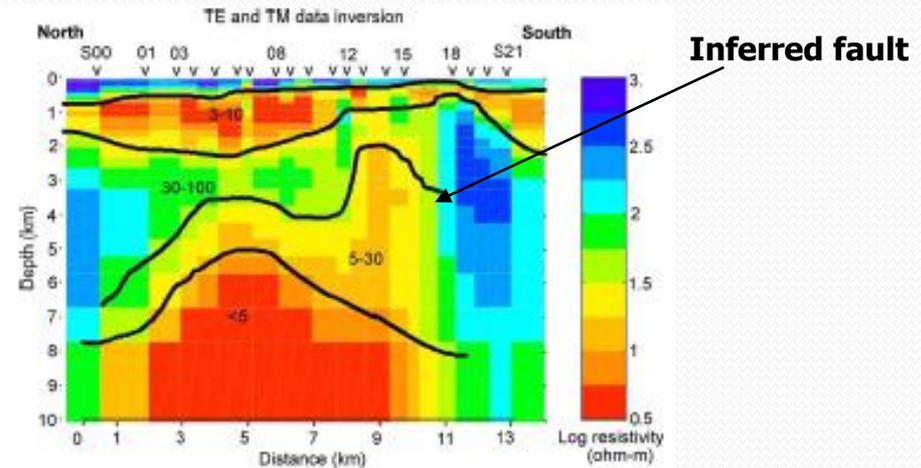
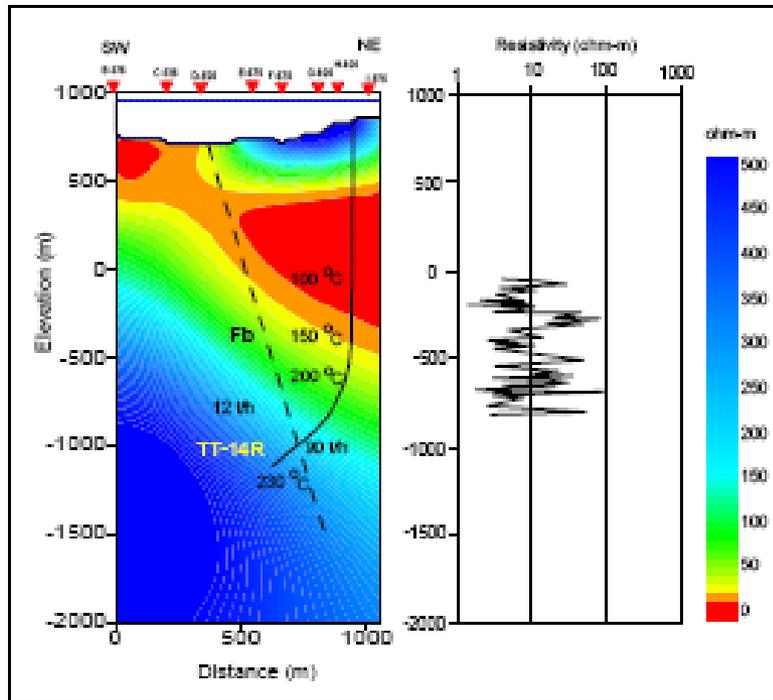
3-D view of the resistivity model, from south. Shallow blocks to a 200m depth are stripped out and approximate locations of three faults are overlaid.



# Applications

## Fault/fracture detection

### Area geotermica di Takigami, Giappone



Oskooi et al., 2005

Tratto da Tagomori et al., WGC 2005

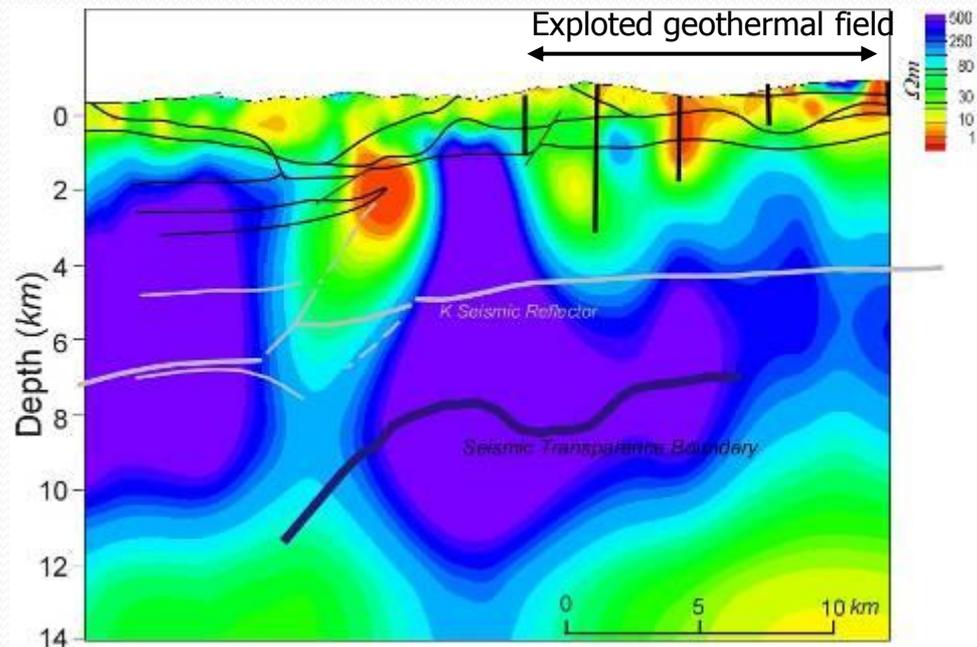
“the large lost circulation occurred at the depth of 1300 m BSL for the well TT-14R (90 t/h) when the well crossed through the electrical discontinuity Fb”



# Applications

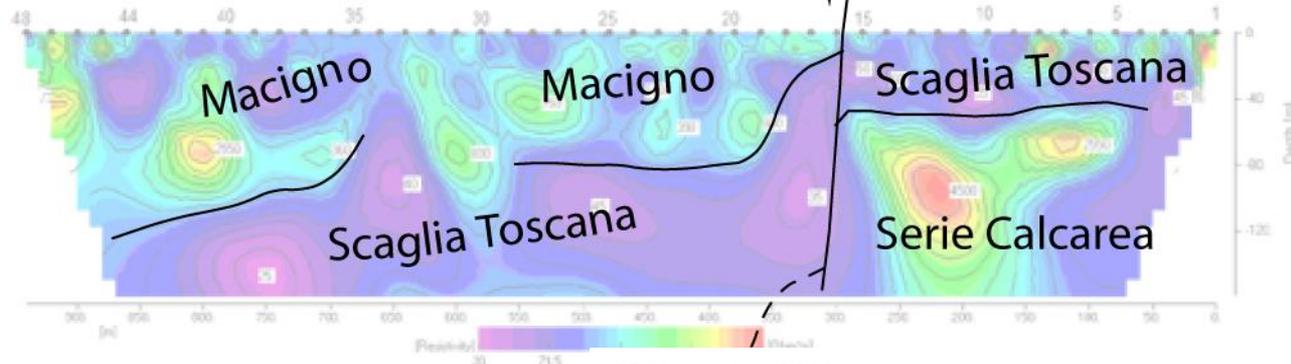
## Fault/fracture detection

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



23 Luglio 2008

W-SW

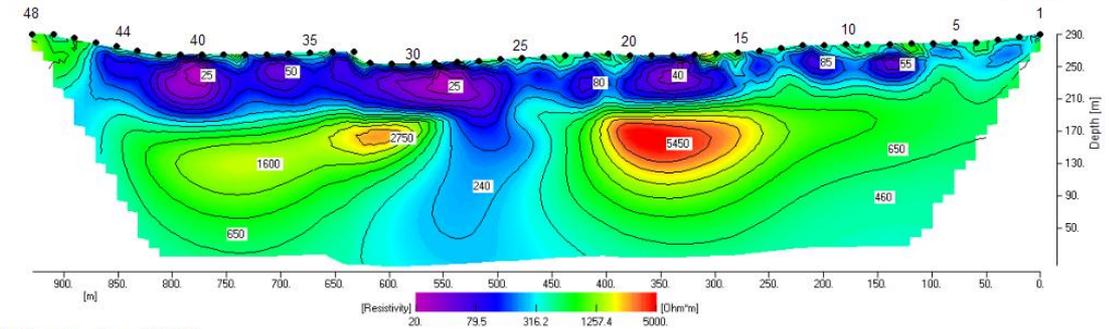


--- 25 Novembre 2008

Polo-Dipolo-rho

W-SW

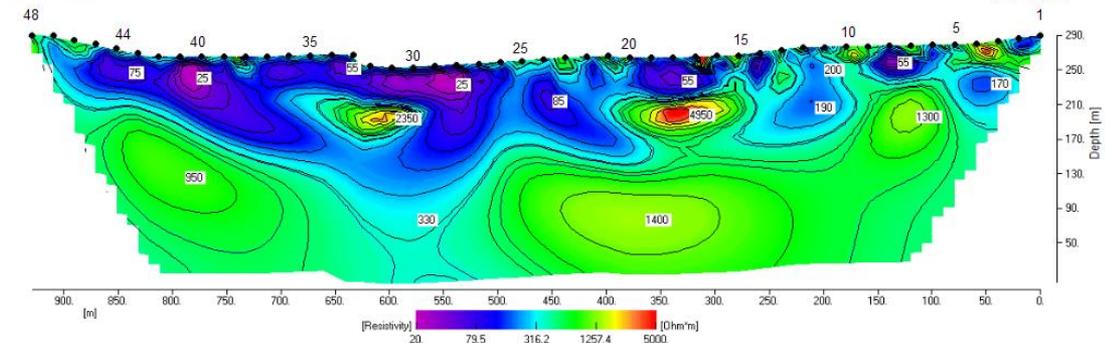
E-NE



26 Novembre 2008

W-SW

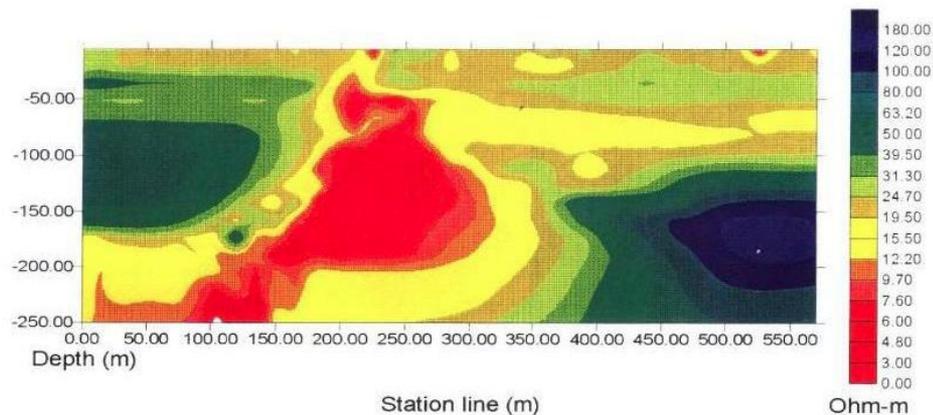
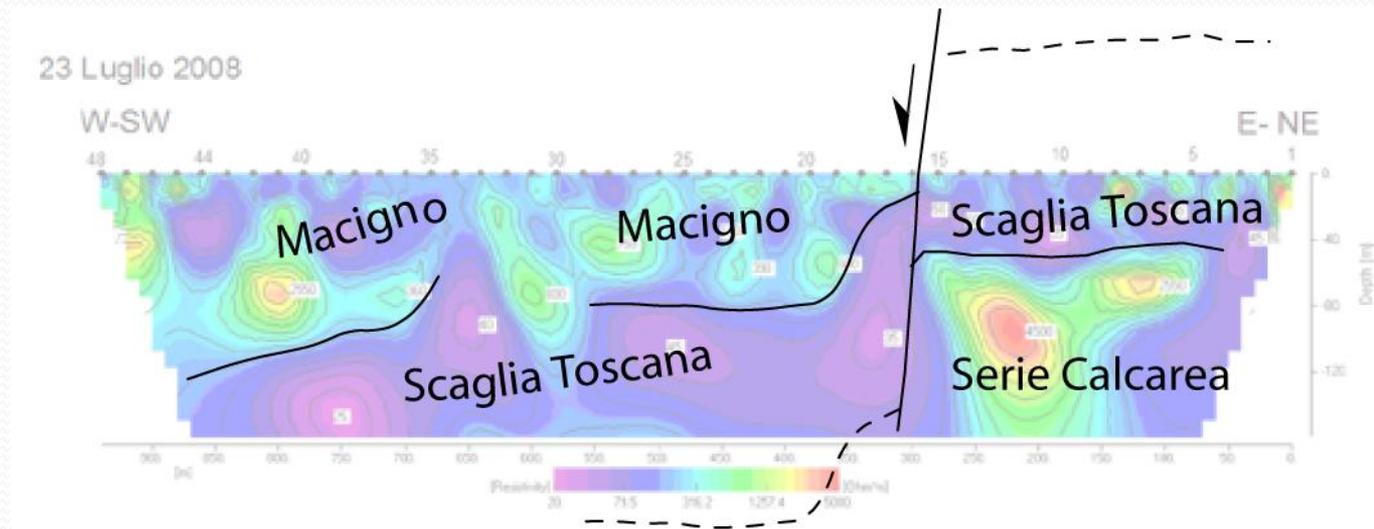
E-NE



# Applications

## fracture detection

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

## fracture detection

The presence of dykes is revealed by high susceptibility. On the other end circulation of hydrothermal fluids causes alterations in the rock which lead to a reduction in susceptibility. This reduction is a consequence of the destruction of the magnetite contained in the rocks. That way, units of volcanic rocks and lava flows can easily be distinguished from hydrothermally altered rock units, which makes geomagnetic surveys a useful tool for geothermal prospecting at high enthalpy volcanic reservoirs coupled with TDEM and MT.

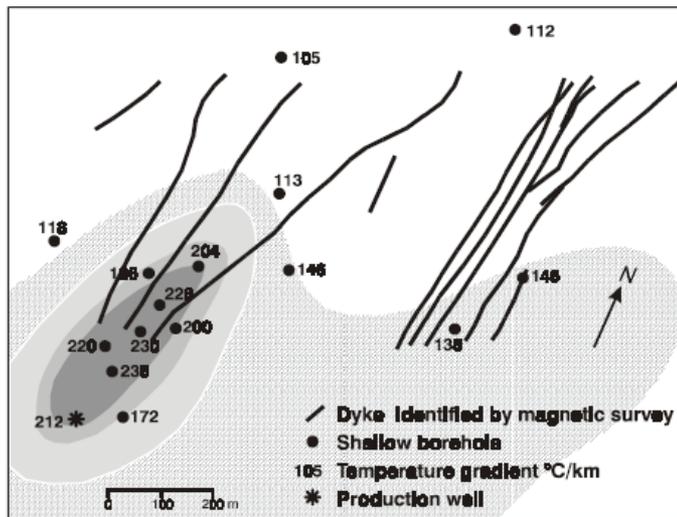


Figure 5. An area outlined in figure 1. Dykes mapped by ground magnetic surveys emplaced on a temperature gradient map. The production well ARS-29 is marked by a star. Note the two dyke swarms and that the main anomaly in the gradient is located, where three dykes are converging.

(from Flovenz et al, WGC2000)



# Applications

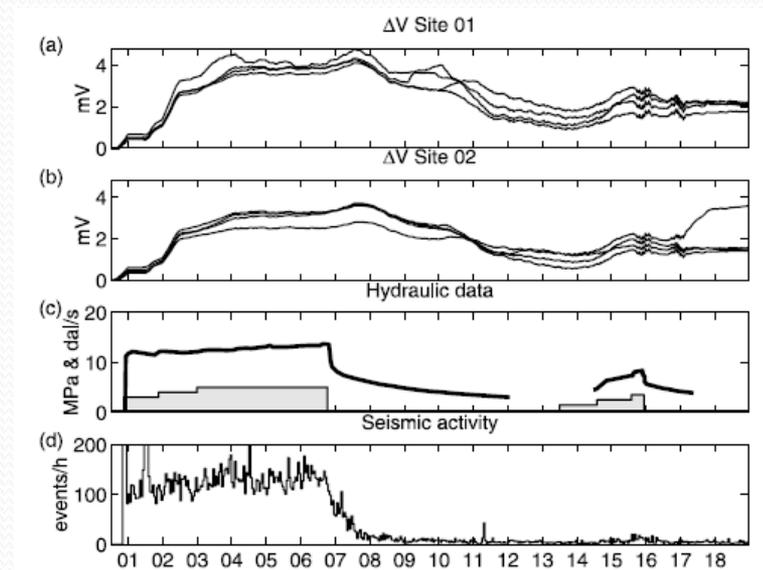
## monitoring

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.

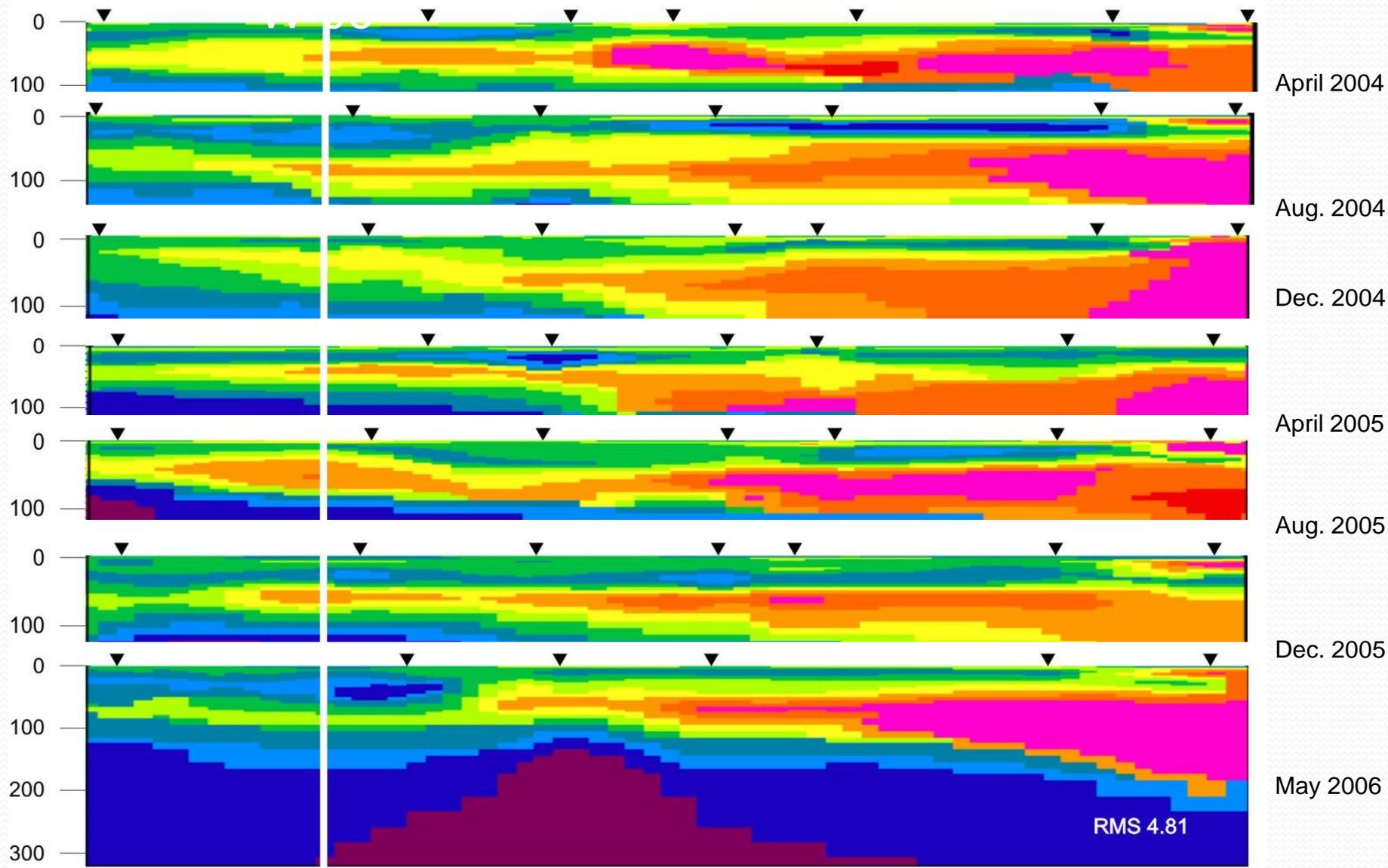
No examples or applications found in literature

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



# Applications

monitoring



# Applications

## 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

