



Chap VI: Geoelectrical methods used in Hydrogeology

Geoelectrical methods are a non-invasive (non destructive) geophysical techniques that measure the electrical properties (primarily electrical resistivity or its inverse, conductivity, and chargeability/Induced polarization) of the subsurface **to infer its composition, fluid content, and geological structure**. They provide spatially extensive, high-resolution information about the subsurface in a **cost-effective manner**.

They send electrical current into the ground (or measuring natural electrical fields/passive mode in case of Spontaneous Potential SP) and analyzing how the subsurface conducts electricity. Because **water content, salinity, and geology strongly affect conductivity**, they can reveal hidden groundwater structures and processes better than other hydro-geophysics methods.

We will discuss here in this chapter the Resistivity, Spontaneous Potential or Self Potential (SP) and Vertical electrical soundings (VES).



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Resistivity: It is applied to map the resistivity structure of the underground. Rock resistivity is of special interest for hydrogeological purposes: it allows, e.g., to **discriminate between fresh water and salt water, between soft-rock sandy aquifers and clayey material, between hard rock porous/fractured aquifers and low-permeable claystones and marl stones, and between water-bearing fractured rock and its solid host rock.**

How it works?

Resistivity of the ground is measured by injected currents and the resulting potential differences at the surface. The general field layout is presented in Figure. 17. Two pairs of electrodes are required: electrodes A and B are used for current injections, while electrodes M and N are for potential difference measurements.

For a homogeneous ground and an arbitrary electrode arrangement (Fig. 17A) the resistivity ρ (unit: Ohm*meter, Ωm) as the relevant petrophysical parameter can be calculated from the current I and the potential difference U by:

$$\rho_A = K \cdot \frac{U}{I}$$



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K is the geometric factor (unit: meter) and can be calculated from the electrode spacing.

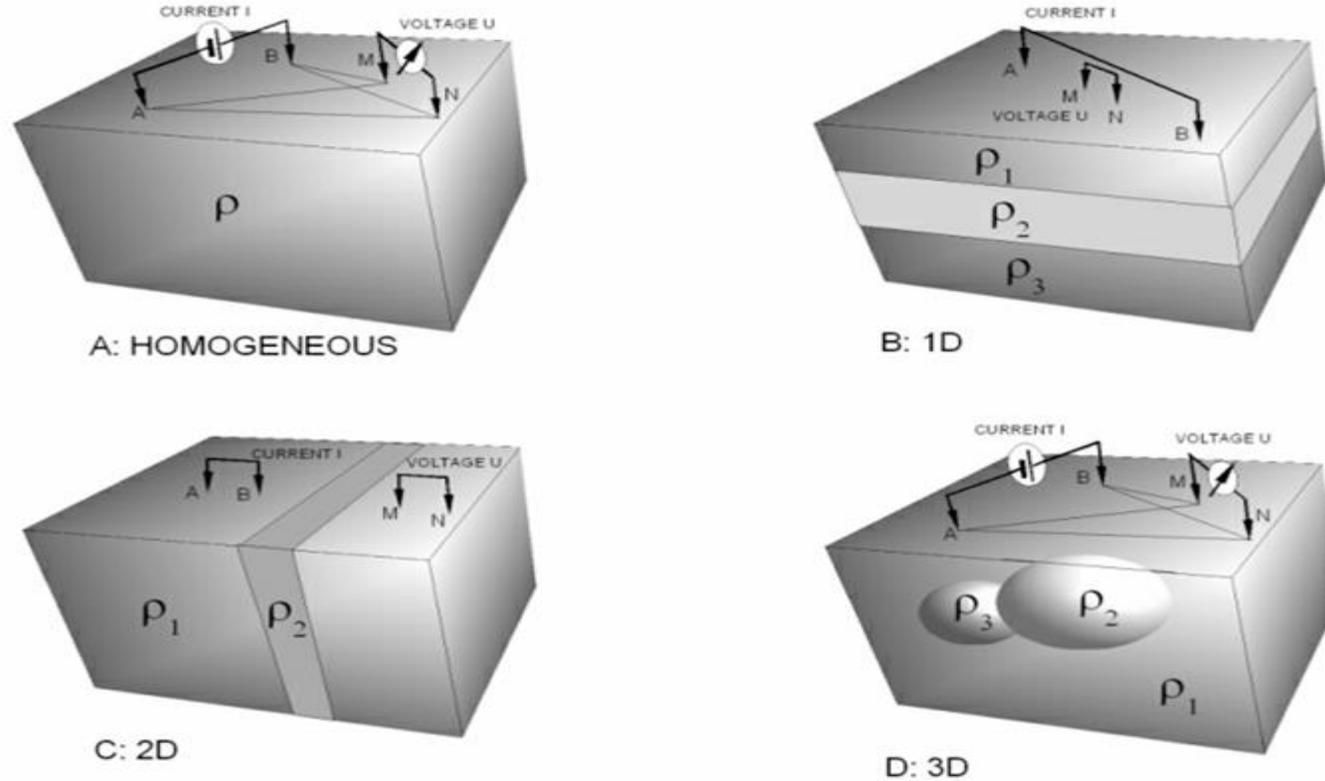


Figure 17: Electrode arrangement for apparent resistivity measurements (Source: Reinhard Kirsch 2009)
D. Layered ground, C. 2D Resistivity distribution in the ground, D. 3D Resistivity distribution in the ground



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by systematically changing and/or moving the electrodes configuration, we can have a huge data set that must be processed by computer to get the underground resistivity distribution that will be translated into geological structures.

The below is the current flows lines and the equipotentials distribution in a homogeneous ground (See figure 18).

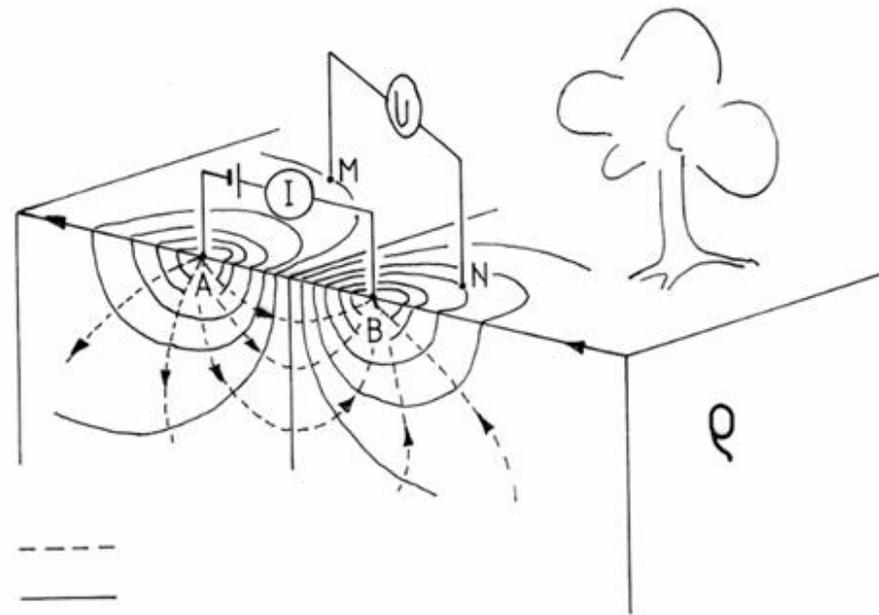


Figure 18: Current flow lines and equipotentials in a homogeneous ground (Source: Reinhard Kirsch 2009)



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The **most common electrodes configurations** used for ground-surface surveys are illustrated in Figure 19

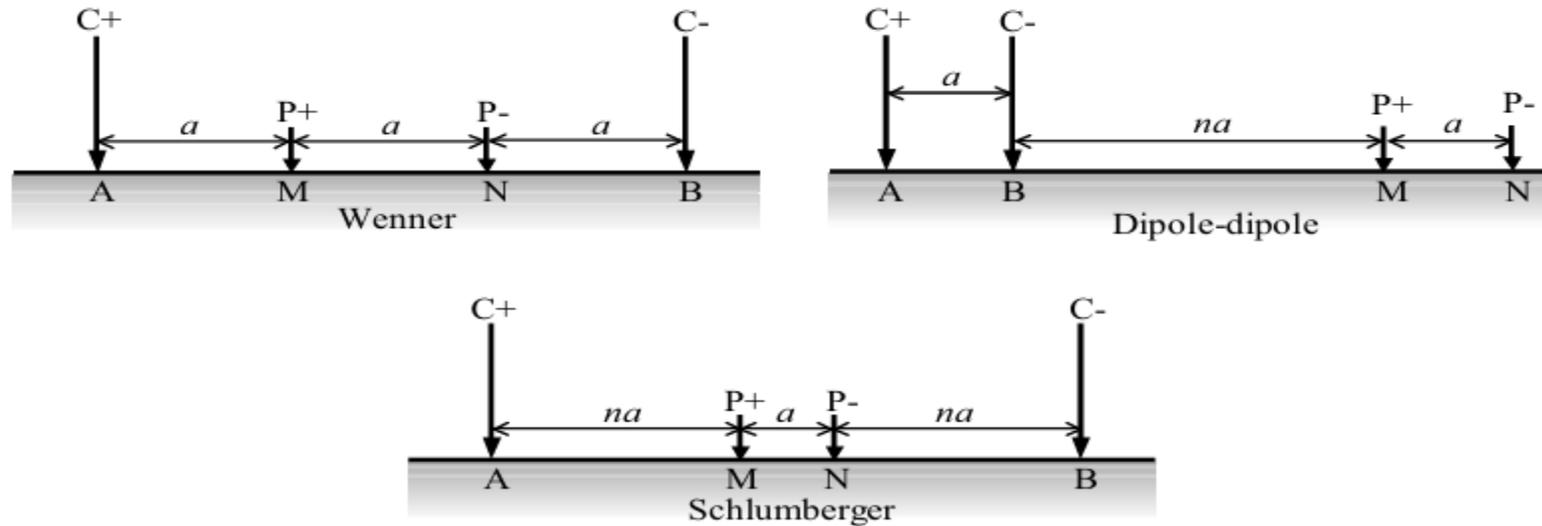


Figure 19: Most common electrode configurations (Source: Rubin & Hubbard 2005)

Electrodes Config methods	Best For	Limitations
Wenner	Good vertical resolution, robust against noise.	Moderate depth of investigation.
Dipole-Dipole	Good horizontal resolution, sensitive to vertical structures.	Weak signal strength → sensitive to noise.
Schlumberger	A good compromise (Ver, HZ) , efficient for profiling.	Lower signal strength than Wenner when $AM \approx MN$.



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The recommended use of the previous configurations is summarized in the below table:

Goal	Recommended configuration
Mapping groundwater table	Wenner or Schlumberger
Imaging contaminant plumes / salinity	Dipole–dipole
Fracture mapping	Dipole–dipole
Deep aquifer mapping	Schlumberger

In a water saturated media the most used relationship is the Archie's Law (Gus Archie 1942) which is a fundamental empirical law in hydro-geophysics that links the **electrical resistivity of a saturated rock or soil** to its **porosity** and **pore-water resistivity**.

- Given by: $\rho = a \cdot \rho_w \cdot \phi^{-m}$ Where:
- ρ = bulk resistivity of the rock/soil
- ρ_w = resistivity of the pore water
- ϕ = porosity
- a = tortuosity factor (≈ 1)
- m = cementation exponent ($\approx 1.3-2.5$)
- This equation shows that resistivity increases as porosity decreases and decreases as water becomes more conductive.



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Archie's Law is useful in Hydro-geophysics for:

- porosity estimation from resistivity especially in unconsolidated sediments or sandstone aquifers.
- Calculate water saturation; useful in mapping vadose zone moisture, contamination plumes, or saltwater intrusion.

Note: The validity of Archie's law is restricted to materials with highly resistive mineral grains and conducting pore fluid. Modified formulations are required for material with surface conductivity like clay.



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Typical Resistivity Values of formation

Material / Formation	Resistivity Range ($\Omega \cdot m$)	Notes / Hydrogeophysical Interpretation
Clay / Shale	1 – 20	Very conductive due to surface conduction; indicates fine-grained, low-permeability materials.
Silty Soil	10 – 100	Moderately conductive; higher than clay but lower than sands.
Sandy Soil (moist)	20 – 200	Influenced strongly by water content and salinity.
Dry Sand	1,000 – 10,000	Very resistive when dry.
Gravel	100 – 600	Coarse and typically resistive when clean.
Limestone	50 – 5,000	Depends heavily on fracturing and water content.
Sandstone	20 – 2,000	Clean, saturated sandstones are moderately resistive; clay-rich much lower.
Dolomite	100 – 10,000	Typically more resistive than limestone.
Basalt (dry)	1,000 – 100,000	Very resistive when fresh and dry.
Basalt (weathered / fractured)	30 – 500	Fractures and clays reduce resistivity.
Granite / Igneous Rocks	1,000 – 1,000,000+	High resistivity; decreases with fracturing or groundwater.
Groundwater (fresh)	5 – 50	Depends on dissolved solids (TDS).
Groundwater (saline)	0.2 – 5	Highly conductive due to salinity.



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Vertical electrical soundings (VES): It is the 1 D Dimensional case of resistivity method applied to a horizontally or approximately horizontally layered earth in order to study mainly changes in the vadose zone (saturation zone) under natural loading. The outcomes of a VES survey are, the number of layers, their thicknesses and resistivities.

How it works?

In VES Survey the **current electrode spacing A-B is progressively increased** while the center **M-N** of the array remains fixed.

This allows investigation to increasing depths, giving a **1D resistivity model** of the subsurface (ρ vs. depth) See Figure 20.

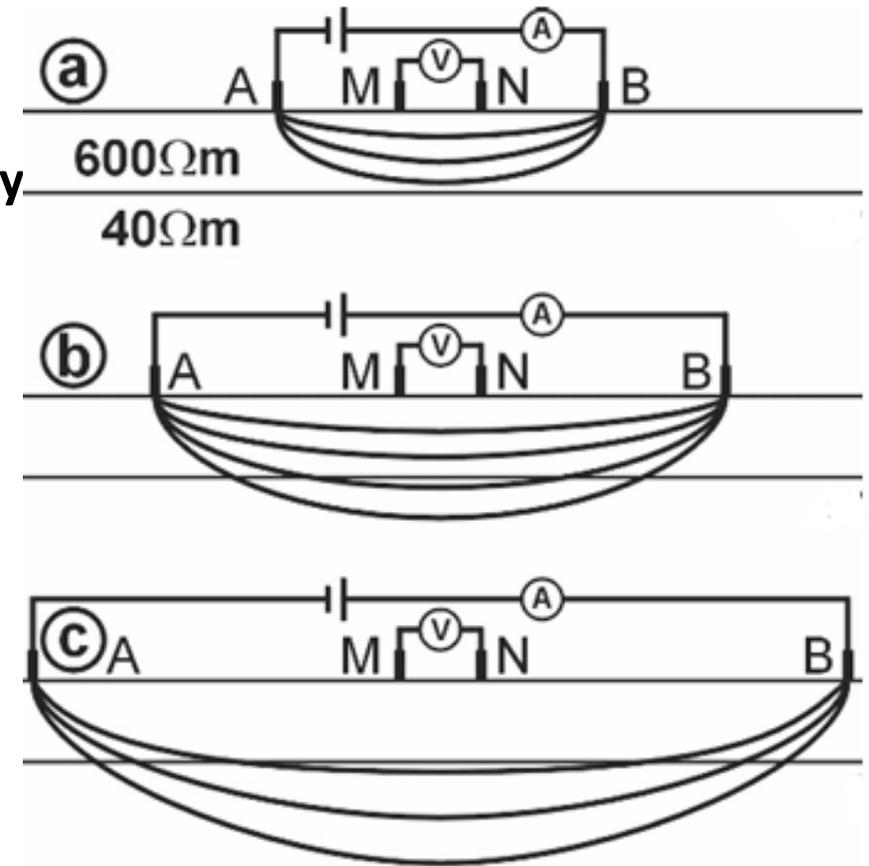


Figure 20: VES Survey configuration (Source: Reinhard Kirsch 2009)



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The gradual increase the current-injecting electrodes AB spacing, leads to an increasing penetration of the current lines and in this way to an increasing influence of the deep-seated layers on the ρ_a apparent resistivity. The step-wise measured apparent resistivities are plotted against the current electrode spacing in a log/log scale and interpolated to a continuous curve. This plot Fig-21 is called sounding curve, that is the base of all data inversion to obtain the resistivity/depth structure of the ground.

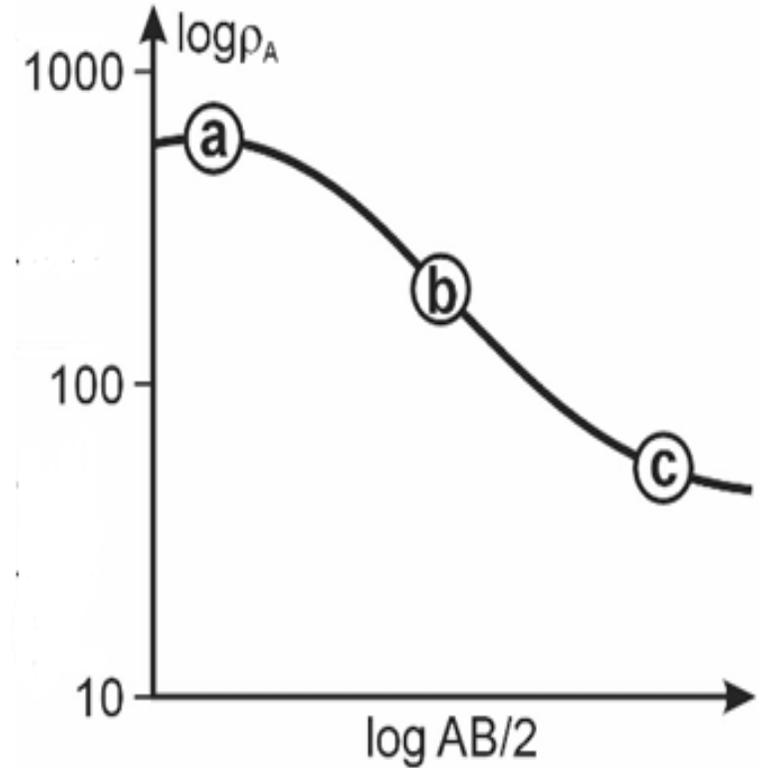


Figure 21: VES measurements plot (Source: Reinhard Kirsch 2009)



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- The best common VES is the **Schlumberger method**:
 - A ——— M — N ——— B
Current electrodes: A, B
Potential electrodes: M, N
 - **AB/2** → half current electrode spacing
 - **MN/2** → half potential spacing
 - **Depth of investigation** ≈ 1/3 to 1/2 of AB/2.

Important: topographic conditions can prevent a full Schlumberger spread. So; alternative method like Wenner can be used for a VES.

The curve shape in log-log plot indicates layering type.

- **H-type:** high–low–high resistivity
- **K-type:** low–high–low resistivity
- **A-type:** increasing resistivity with depth
- **Q-type:** decreasing resistivity with depth
- **HKH, KHK, etc.** for complex geology

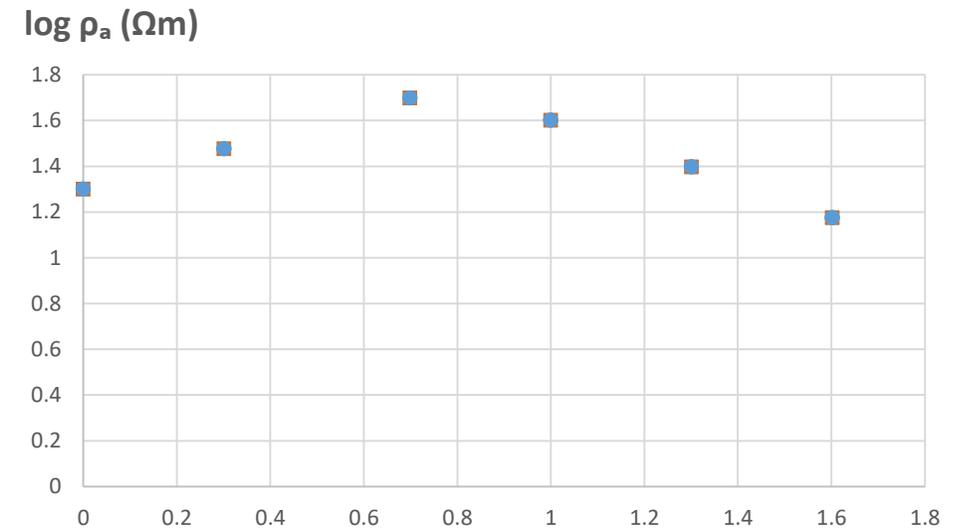


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Exercise: A given VES survey data is in the below table

AB/2 (m)	1	2	5	10	20	40
ρ_a (Ωm)	20	30	50	40	25	15

- **Step 1 – Conceptual Plot (log–log)**
- Plotting **AB/2** (x-axis, log scale) vs **ρ_a** (y-axis, log scale):
- Resistivity rises: **20 \rightarrow 30 \rightarrow 50 Ωm**
- Then decreases: **50 \rightarrow 40 \rightarrow 25 \rightarrow 15 Ωm**
- So the curve shape is:
Low \rightarrow High \rightarrow Low
- **Step 2 – Identify curve type**
- This is characteristic of a:
- **H-type curve**
- (high resistivity layer sandwiched between two lower resistivity layers)





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• Step 3 – Geological Interpretation

- Interpretation of H-type curves:
- **Top layer:** moderately low resistivity (20–30 Ωm) → likely clayey sand or moist soil
- **Second layer:** higher resistivity (50 Ωm) → dry sand/gravel or weathered rock
- **Third layer:** lower resistivity (15–25 Ωm) → saturated layer (possible aquifer), clay-rich, or saline water
- **Hydrogeophysical meaning**
- The downward decrease in resistivity suggests:
- A **saturated zone** or A **clay-rich aquifer**
- Or a **saline/brackish water layer**

Final result: Most likely → An aquifer located beneath a resistive dry layer.

Important: You Cannot Compute ρ_i and h_i by Hand because:

- Apparent resistivity is not a direct average — it's influenced by all deeper layers.
- The forward function $\rho_a(AB/2)$ is **non-linear**.
- Finding ρ_i and h_i requires solving a **non-linear optimization** problem.
- The inversion handles thousands of recursive calculations per iteration.



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Spontaneous Potential or Self Potential: These measurements belong to the earliest methods used in applied geophysics. Originally applied in mining to ore body exploration (اكتشاف الاجسام الخامة), then used widely standard tool with borehole logging and is now of increasing interest in geothermal application, and hydrogeology.

It is the **natural electrical direct currents originating from various electrochemical, electrophysical, and bioelectrical** processes in the ground. The processes of reduction and oxidation above and below the ground water table define **mineralization potentials** of some **100 millivolts** related with highly conductive ore bodies or graphite deposits.

- Electrochemical potentials in the order of **ten millivolts** result from ion flow in connection with variable electrolytic concentrations of the ground water and with clay mineral membrane effects.
- In hydrogeology most relevant are streaming (or electrokinetic or filtration) potentials from ground-water flow in porous rocks.

Basic Principles of Streaming Potential: When a fluid (typically water) flows through a porous medium under a pressure gradient, generating a measurable electric potential difference due to charge separation in the electrical double layer (EDL) at the solid-fluid interface. This double layer usually fixes pore fluid anions while cations remain mobile. On water flow, the cations are transported, synonymous with an electric current and the setting-up of a positive potential in the direction of the flow.



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Field Procedure of SP measurement: it is a simple and easy technique that measure the voltage between two unpolarizable electrodes using two primary configuration; the Fixed Base Station Method and the Gradient Method.

A) Fixed Base Station Method: it is a moving probe, with a high-impedance voltmeter and a cable layout as shown in Figure 22 it is also called Total Potential method. In general the SP stations are arranged in a grid

- One electrode is fixed at a reference (base) station (assumed 0 mV). The second (roving/moving) electrode is moved to each measurement point along a line or grid. The voltmeter measures the absolute potential difference relative to the fixed base.
- **Data output:** Direct potential values (in mV) at each station → easy to contour into SP maps or map of self potential anomalies.

Advantages:

- Higher accuracy for small anomalies.
- Better detection of subtle signals.
- Less cumulative error.



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Disadvantages:

- Requires long insulated wire (up to 1 km wire); station spacing depends on the project conditions and may be of the order of tens of meters.
- Logistically heavy; risk of wire damage/breakage.

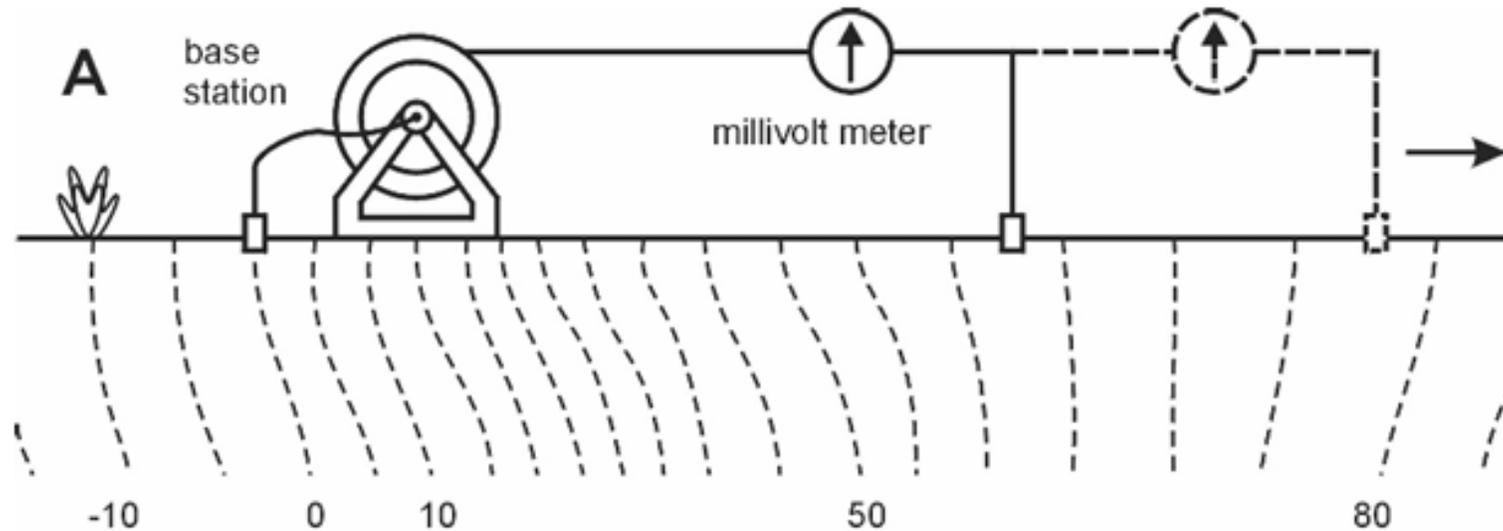


Figure 22 A: Fixed base method layout to measure the self potential current.

Source: Reinhard Kirsch 2009



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B) Gradient method (Dipole method/leapfrog method): Two unpolarizable electrodes with fixed short separation (e.g., 10 m) are used see figure 22 B. Measure potential difference between them, then move both forward, trailing becomes leading. It integrate (it sum) potential gradients along the line to reconstruct total potential. These gradient are the rate of change of electric potential with distance, expressed as $\Delta V / \Delta x$ (in mV/m or V/m).

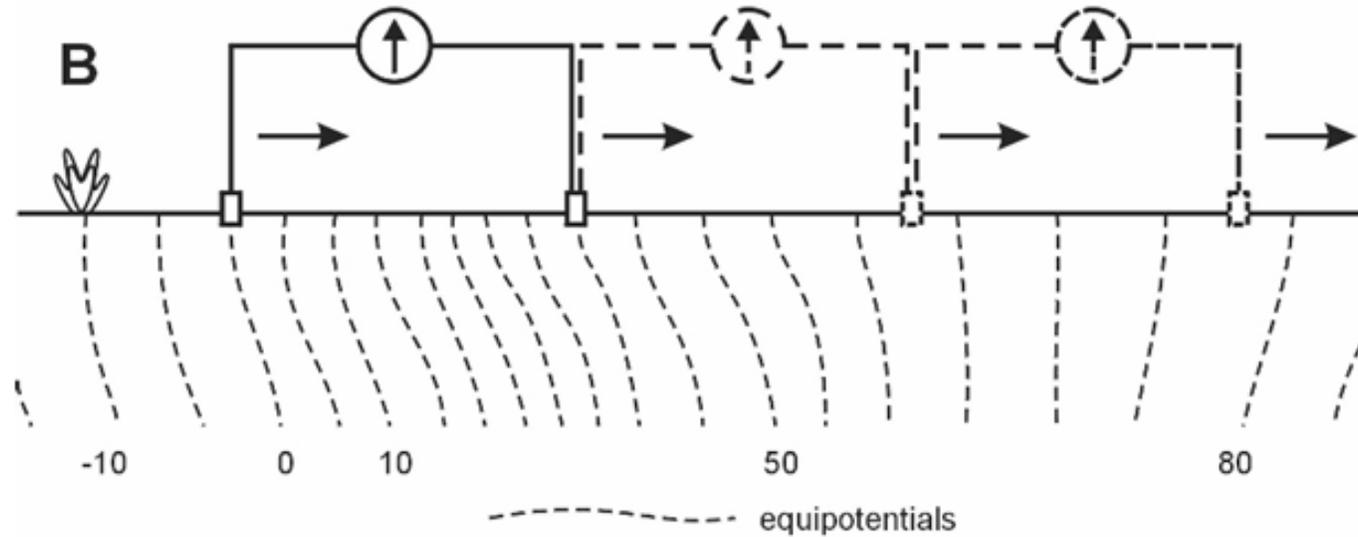


Figure 22 B: Gradient method layout to measure the self potential current.

Source: Reinhard Kirsch 2009



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Data output: Potential gradient (mV/m) between stations → integrated to get total potential map.

Advantages:

- Short wire only (equal to electrode spacing).
- Faster and lighter logistics for long lines.
- less drift error (the error caused by the gradual shift in a gauge's measured values over time) from long cables.
- Suits well for large survey.

Disadvantages:

- Measurement errors accumulate during integration → hides small anomalies.
- Amplifies noise over long distances.



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Comparison Table between Fixed Base Method and Gradient Method for SP measurement

Aspect	Fixed Base Method	Gradient Method
Electrode Setup	One fixed, one roving	Two moving (fixed spacing, leapfrog)
Measured Quantity	Total potential (mV) relative to base	Potential gradient (mV/m)
Wire Length Needed	Long (full survey length)	Short (electrode spacing only)
Accuracy for Small Anomalies	High	Lower (cumulative errors)
Noise Sensitivity	Lower	Higher (amplifies during integration)
Logistics	Cumbersome	Easy & fast
Typical Use	Precise mapping (e.g., mineralization)	Long reconnaissance lines



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Depends on the geographic context the high-amplitude bioelectric potentials from vegetation can seriously overprint the geologic SP signature, and time variations of SP fields are a matter of further complexity (Ernstson and Scherer 1986 in Kirsch 2009).

Self-potential data are plotted on profiles in figure 23 map pf anomalies or in the form of iso-potentials figure 24.

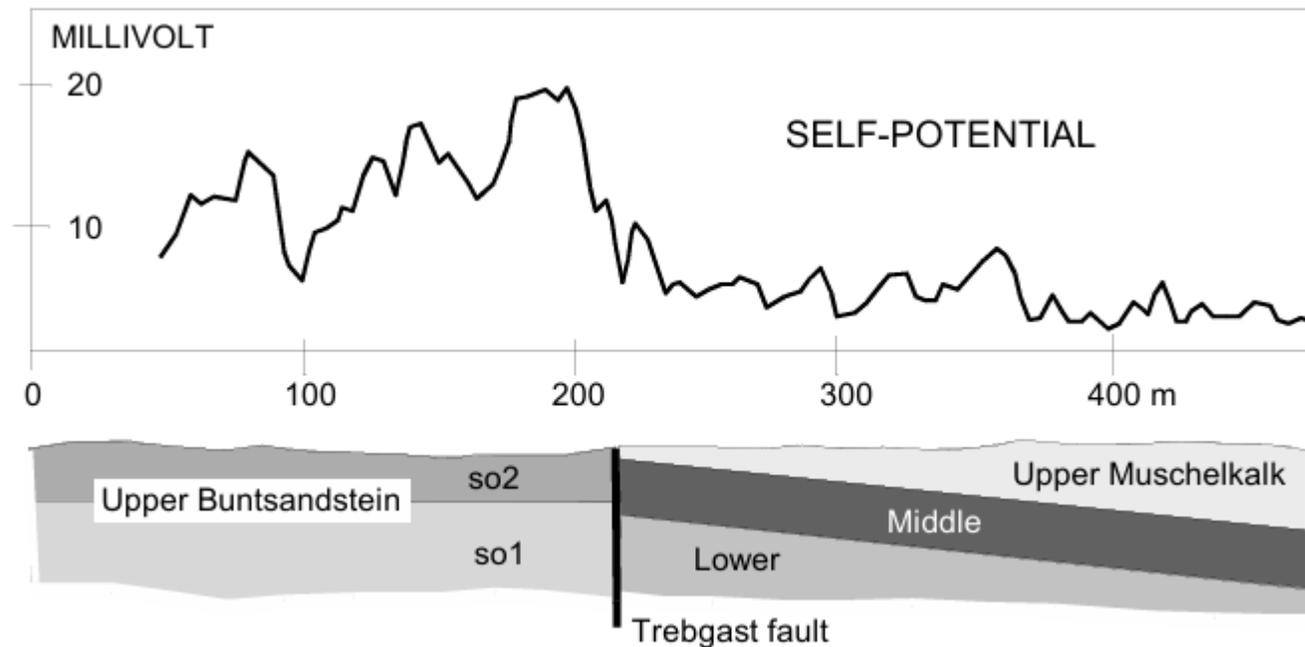


Figure 23: Profile of self-potential anomalies across a fault [Source: Reinhard Kirsch 2009](#)
Salah Eddine BOUHENICHE Jijel university Academic Year 2025_2026



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SP interpretation is mostly done qualitatively with the aim to, e.g., delineate ground-water flow and to locate zones of ground-water infiltration and ascent (going up) see figure 24 to understand what is happening around a well while pumping.

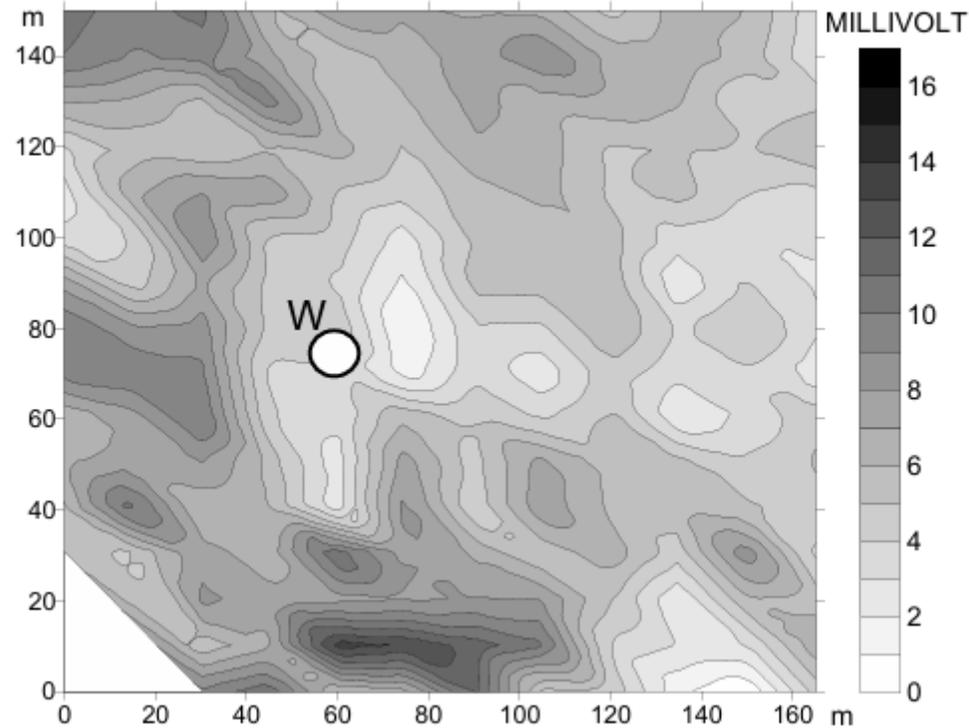


Figure 24: Self-potential anomalies around a well (w) during a pumping test [Source: Reinhard Kirsch 2009](#)



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Important: New approaches use mathematical methodologies to directly relate SP signals with hydraulic conductivity distribution in the ground (Rizzo et al. 2004 in Kirsch 2009) and with the contours of the ground-water table (Birch 1998, Revil et al. 2003 in Kirsch 2009).

SP is a passive surface surveys or borehole (hole of the well); map SP anomalies related to flow it can be applied to study dam seepage.

Its main application are: Groundwater flow mapping, contaminant plumes, permeability proxies to correlate (direct measurement of permeability are impractical at field scale) between geophysical and hydrogeological properties.