1- Measuring the Magnitude of Earthquake

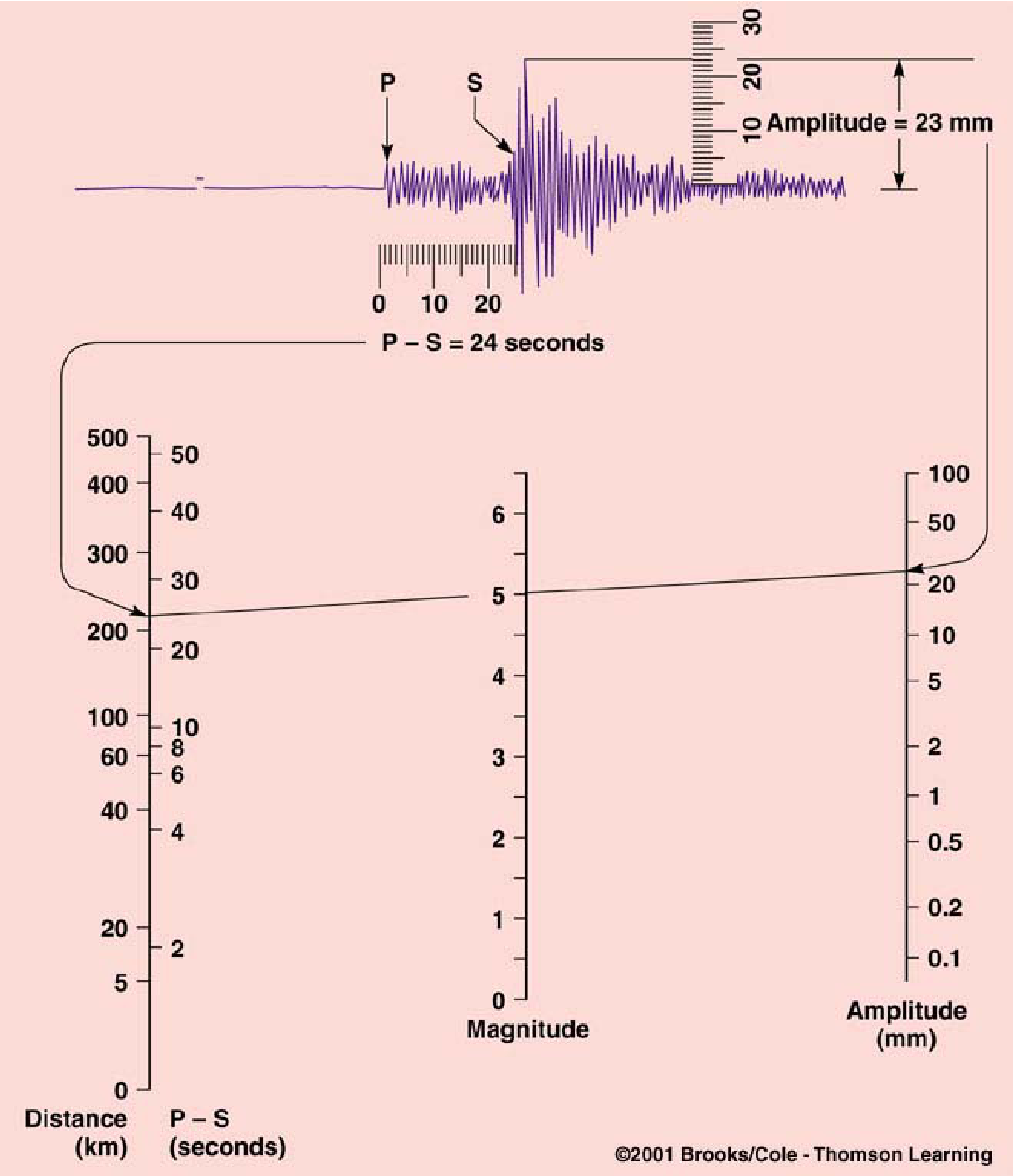
The magnitude of most earthquakes is measured on the **Richter scale**, invented by Charles F. Richter in 1934. The Richter magnitude is calculated from the amplitude of the largest seismic wave recorded for the earthquake, no matter what type of wave was the strongest.

The Richter magnitudes are based on a logarithmic scale (base 10). What this means is that for each whole number you go up on the Richter scale, the amplitude of the ground motion recorded by a seismograph goes up ten times. Using this scale, a magnitude 5 earthquake would result in ten times the level of ground shaking as a magnitude 4 earthquake (and 32 times as much energy would be released). To give you an idea how these numbers can add up, think of it in terms of the energy released by explosives: a magnitude 1 seismic wave releases as much energy as blowing up 6 ounces of TNT. A magnitude 8 earthquake releases as much energy as detonating **6 million tons of TNT**. Fortunately, most of the earthquakes that occur each year are magnitude 2.5 or less, too small to be felt by most people.

The Richter magnitude scale can be used to describe earthquakes so small that they are expressed in negative numbers. The scale also has no upper limit, so it can describe earthquakes of unimaginable and (so far) inexperienced intensity, such as magnitude 10.0 and beyond.

- Magnitude
  - Richter scale measures total amount of energy released by an earthquake; independent of intensity

– Amplitude of the largest wave produced by an event is corrected for distance and assigned a value on an open-ended logarithmic scale.



2- Hazards associated with Quakes

- **Shaking:**

Frequency of shaking differs for different seismic waves.

High frequency body waves shake low buildings more.

Low frequency surface waves shake high buildings more.

Intensity of shaking also depends on type of subsurface material.

Unconsolidated materials amplify shaking more than rocks do.

Fine-grained, sensitive materials can lose strength when shaken. They lose strength by *liquefaction*.

Buildings respond differently to shaking depending on construction styles, materials

Wood -- more flexible, holds up well

Earthen materials -- very vulnerable to shaking.

- **Ground displacement:**

Ground surface may shift during an earthquake (esp. if focus is shallow).

Vertical displacements of surface produce *fault scarps*.

- **Tsunamis (NOT tidal waves)**

Tsunamis are huge waves generated by earthquakes undersea or below coastal areas.

If earthquake displaces sea surface, wave is generated that can grow as it moves over sea surface.

- **Fires**

Usually occurs from shifting of subsurface utilities (gas lines)

3-

a) Types of Body Waves Body waves propagate by a series of compressions and dilatations of the material (compressional wave) or by shearing the material back and forth (shear wave). A *compressional wave* is a primary or P-wave because compressional waves arrive first from earthquakes; they are also called longitudinal and push-pull waves because particles of the material move back and forth, parallel to the direction the wave is moving (Fig. 8a). An example is a sound wave traveling through air. The sound propagates as the air is compressed then dilated; a person's eardrum senses the density changes in the air.

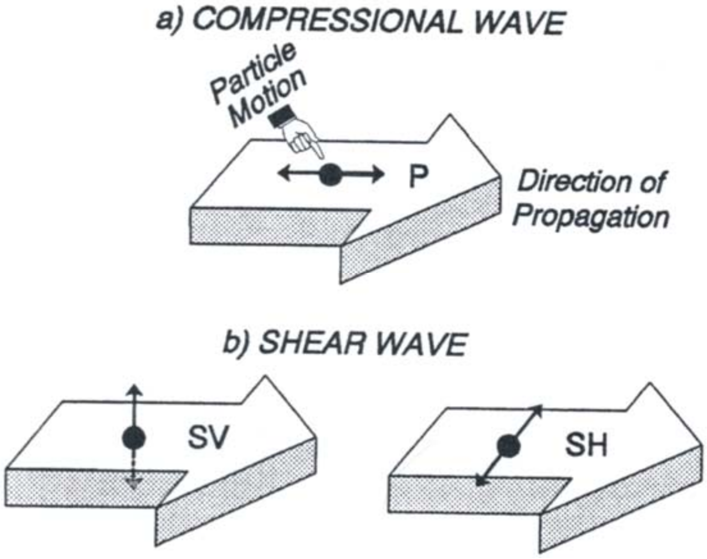


FIGURE 8 *Particle motions for body waves.* a) In a compressional wave, particles of the material move back and forth, parallel to the direction the wave energy moves. b) Particles of the material move perpendicular to the direction of propagation of shear wave energy. For wave energy moving horizontally: SV = shear wave with vertical particle motion; SH = shear wave with horizontal particle motion.

Particle motions for a *shear wave* are perpendicular to the direction of propagation (Fig. 8b). Shear waves are also referred to as *secondary or S-waves* because they arrive from an earthquake after the initial compressional waves, and as *transverse waves* because of their particle motions.

Velocity of Body Waves The term *velocity* refers to a vector, with both magnitude and direction. In seismology *velocity* is commonly used to refer to the magnitude component, or *speed,* with direction not necessarily implied. For an elastic, isotropic, unbounded material, seismic velocities depend on the elastic constants (k, μ, E, υ) and density (ρ) of the material. The more resistant the material is to deformation (that is, the greater the material's incompressibility or rigidity), the faster the waves travel. Body wave velocities are given by:

$$\boxed{V_P = \sqrt{\frac{k + 4/3\mu}{\rho}} = \sqrt{\frac{\lambda + 2\mu}{\rho}}}$$

and

$$\boxed{V_S = \sqrt{\frac{\mu}{\rho}}}$$

Where:

V_P = velocity of the compressional wave

V_S = velocity of the shear wave.

The velocity equations lead to the following generalizations. 1) *For the same material, shear waves will always travel slower than compressional waves.* 2) *The more rigid the material, the higher the P and S wave velocities.* For example, seismic waves speed up as they travel from the asthenosphere into the mantle part of the lithosphere, suggesting that the mantle lithosphere is more rigid (higher μ) than the asthenosphere. 3) *Fluids (liquid or gasses) have no shear strength* (μ = 0). This lack of rigidity means two things: a) shear waves cannot travel through fluids, like air, water and earth's outer core; and b) compressional waves travel slower through the liquid state than through the solid state of the same material (the fluid outer core, for example, has lower V_P than the solid inner core).

Observations suggest that the following factors generally lead to an *increase in seismic velocity.* 1) *Increasing depth within the Earth.* As materials are better compacted and cemented, they become more rigid and incompressible, increasing both k and μ. 2) *Increasing in density.* This generalization is commonly true for rocks, even though the equations show an inverse relationship between velocity and density. As rocks become more dense, they generally become incompressible and more rigid; the corresponding elastic constants (k and μ) commonly increase more than the density (ρ). 3) *Decreasing porosity.* With decreasing pore space, a rock becomes more dense. At the same time, it is generally more lithified and has less fluid content, thereby increasing both k and μ . 4) *Change from liquid (or partial melt) to solid.* The rigidity (μ) increases, raising both V_P and V_S. Seismic waves therefore speed up while traveling from the asthenosphere to the mantle lithosphere, and from the outer core to the inner core.

	ELASTIC CONSTANTS 10⁹ N / m²		kg / m³ / g / cm³	SEISMIC VELOCITIES km / s	
	Bulk Modulus (k)	Shear Modulus (μ)	Density (ρ)	Compres. Wave (V_p)	Shear Wave (V_s)
Air	0.0001	0	1.0 / 0.001	0.32	0
Water	2.2	0	1000 / 1.0	1.5	0
Ice	3.0	4.9	920 / 0.92	3.2	2.3
Shale	8.8	17	2400 / 2.4	3.6	2.6
Sandstone	24	17	2500 / 2.5	4.3	2.6
Salt	24	18	2200 / 2.2	4.7	2.9
Limestone	38	22	2700 / 2.7	5.0	2.9
Quartz	33	39	2700 / 2.7	5.7	3.8
Granite	88	22	2600 / 2.6	6.7	2.9
Peridotite	139	58	3300 / 3.3	8.1	4.2

FIGURE 9 Typical values for elastic constants, density, and seismic velocities for selected materials, listed according to increasing compressional wave velocity (V_p). Compiled from Kinsler et al. (1982) and other sources. SI units for density are kg/m³; the literature, however, commonly gives densities in g/cm³.

Fig. 9 gives typical physical properties for various substances, listed downward according to increasing P-wave velocity. The fluids (air and water) have some resistance to compression ($k \neq 0$), supporting P-wave travel. Lake of rigidity ($\mu = 0$) in the fluids contributes to low P-wave velocities and results in no propagation of S-waves ($V_s = 0$). Going down the list the materials generally become stronger (higher k and μ) and more dense with increasing velocity.

The graphs of P- and S-wave velocity vs. Density (Fig. 10) were compiled from the information in Fig. 9.

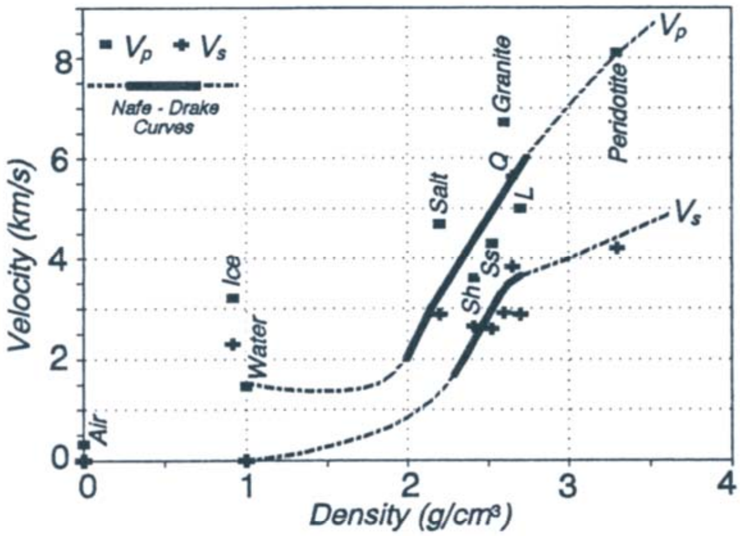


FIGURE 10 Graph of seismic velocity vs. density for materials presented in Fig. 9. L = limestone; Q = quartz; Sh = shale; Ss = sandstone. Solid rectangles with labels are compressional wave velocities (V_P); corresponding shear wave velocities (V_S) shown directly beneath by plus's. Empirical ("Nafe-Drake") curves, developed through analysis of numerous rock and sediment samples, are superimposed on the graph (Ludwig, Nafe, and Drake, 1971). Portions of these curves, highlighted by the solid lines, show roughly linear relationship between seismic velocities and densities for crustal rocks. 1 g/cm^3 = 10^3 kg/m^3.

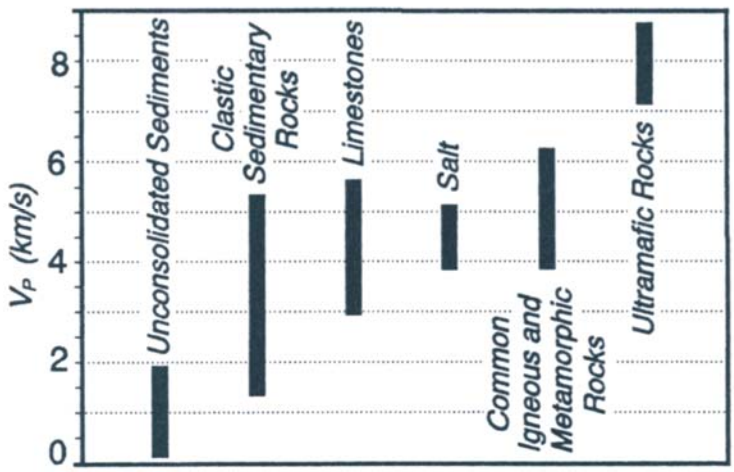


FIGURE 11 Approximate ranges of compressional wave velocity (V_P) for some materials encountered at Earth's surface (from Griffiths and King, 1981).

The solid portions of the superimposed curves high-light the crudely linear relationship between density and seismic velocity for crustal rocks and minerals (shale, sandstone, salt, limestone, quartz, granite).

Fig. 11 presents common ranges of compressional wave velocities for materials encountered at Earth's surface. The values on the chart illustrate that it is difficult to identify rocks based only on velocity information, because the wide ranges result in sizeable overlaps. It is clear, however, that seismic velocities are generally lowest for unconsolidated sediments, higher for sedimentary rocks, still higher for crystalline rocks of the crust, and highest for ultramafic rocks that formed in the mantle.

b) Critically Refracted Arrival

When seismic energy encounters material of different velocity, some of the energy may be transmitted into the second material (Fig. 19). An increase in velocity speeds up wave fronts in the second material; to remain perpendicular to wave fronts, raypaths bend across the interface (Fig. 22a). *Refraction* describes the bending of raypaths as seismic waves travel from one material to another. For a wave traveling from material of velocity V_1 into velocity V_2 material (Fig. 22b), raypaths are refracted according to *Snell's Law.*

$$\boxed{\frac{Sin\,\theta_1}{V_1} = \frac{Sin\,\theta_2}{V_2}}$$

where:

θ_1 = angle of incidence

θ_2 = angle of refraction

V_1 = seismic velocity of incident medium

V_2 = seismic velocity of refracting medium.

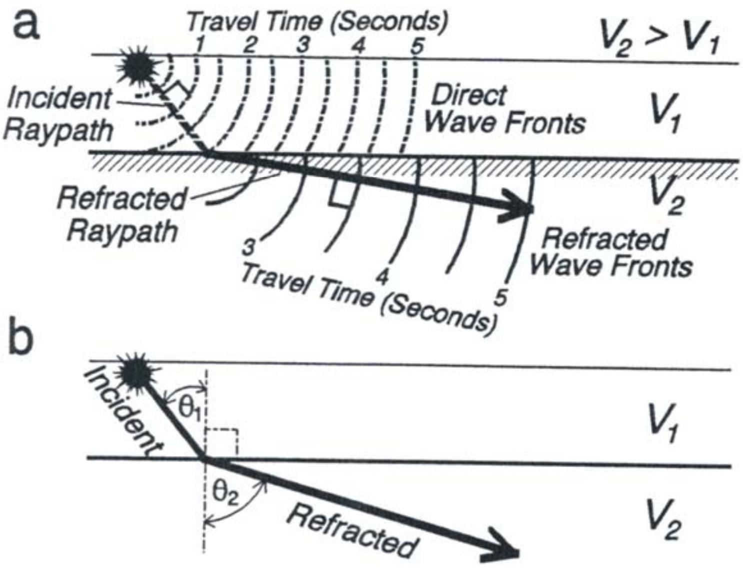


FIGURE 22 Refraction from layer of velocity (V_1) to one of velocity (V_2). a) Wave fronts are distorted from perfect spheres as energy is transmitted into material of different velocity (Fig. 15 b). Raypaths thus bend ("refract") across an interface where velocity changes. b) The incident (θ_1) and refracted (θ_2) angles are measured from a line drawn perpendicular to the interface between the two layers.

Fig. 23 illustrates how Snell's Law describes three situations: I) if the velocity *decreases* across the interface, the ray is refracted *away* from the interface; 2) if the velocity remains the *same,* the ray is not bent; 3) if the velocity *increases* across the interface, the ray is bent toward the interface.

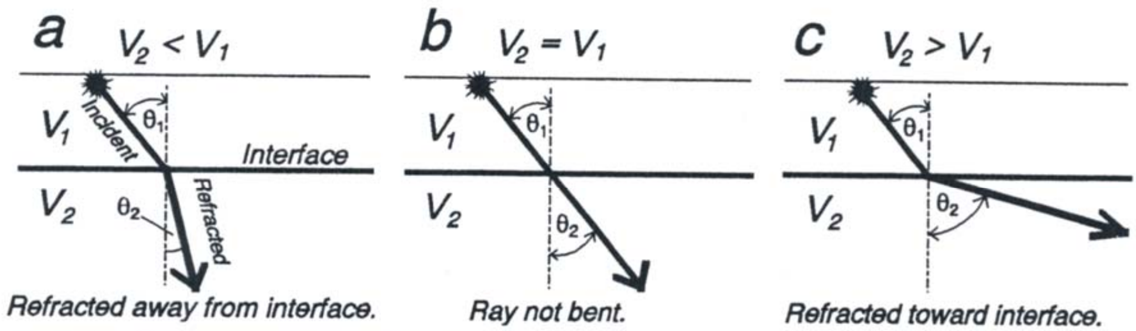


FIGURE 23 Behavior of refracted ray when velocity (a) decreases, (b) remains the same, and (c) increases across an interface.

In all cases, as the angle of incidence increases, so does the angle of refraction (Fig. 24a). A special situation, known as *critical refraction,* occurs when the angle of refraction (θ_2) reaches 90° (Fig. 24b). Note that critical refraction can occur only when $V_2 > V_1$, as in Fig. 23c.

Electric

c) Sources of Noise

There are a number of sources of noise that can affect our measurements of voltage and current from which we will compute apparent resistivities..

- *Electrode Polarization* - A metallic electrode, like a copper or steel rod, in contact with an electrolyte other than a saturated solution of one of its own salts, like ground water, will generate a measurable contact potential. In applications such as SP, these contact potentials can be larger than the natural potential that you are trying to record. Even for the DC methods described here, these potentials can be a significant fraction of the total potential measured.

For DC work, there are two possible solutions.

1. Use nonpolarizing electrodes. These are electrodes that contain a metallic conducting rod in contact with a saturated solution of its own salt. Copper and copper sulfate solution are commonly used. The rod and solution are placed in a porous ceramic container that allows the saturated solution to slowly leak out and make contact with the ground. Because these solutions are rather environmentally unfriendly, and because the method described below is easy to employ, these so-called *porous pot* electrodes are rarely used in DC work. They are, however, commonly used in SP and IP surveys.
2. A simple method to avoid the influence of these contact potentials is to periodically reverse the current flow in the current electrodes or use a slowly varying, a few cycles per second, AC current. As the current reverses, the polarizations at each electrode break down and begin to reverse. By measuring over several cycles, robust current and voltage measurements can be made with negligible polarization effects.

- *Telluric Currents* -As described previously, naturally existing currents flow within the earth. These currents are referred to as telluric currents. The existance of these currents can generate a measurable voltage across the potential electrodes even when no current is flowing through the current electrodes. By periodically reversing the current from the current electrodes, or by employing a slowly varying AC current, the effects of telluric currents on the measured voltage can be cancelled.

- *Presence of Nearby Conductors* -Electrical surveys can not be performed around conductors that make contact with the ground. For example, the presence of buried pipes or chain-linked fences will act as current sinks. Because of their low resistivity, current will preferentially flow along these structures rather than flowing through the earth. The presence of these nearby conductors essentially acts as electrical shorts in the system.

- *Low Resistivity at the Near Surface* -Just as nearby conductors can act as current sinks that short out an electrical resistivity experiment, if the very near surface has a low resistivity, it is difficult to get current to flow more deeply within the earth. Thus, a highly conductive* near-surface layer such as a perched water table can prevent current from flowing more deeply within the earth.

- *Near-Electrode Geology and Topography* - Any variations in geology or water content localized around an electrode that produce near-surface variations in resistivity can greatly influence resistivity measurements. In addition, rugged topography will act to concentrate current flow in valleys and disperse current flow on hills.

- *Current Induction in Measurement Cables* - Current flowing through the cables connecting the current source to the current electrodes can produce an induced current in the cables connecting the voltmeter to the voltage electrodes, thereby generating a spurious voltage reading. This source of noise can be minimized by keeping the current cables physically away from, a meter or two, the voltage cables.

d) Soundings and Profiles survey types

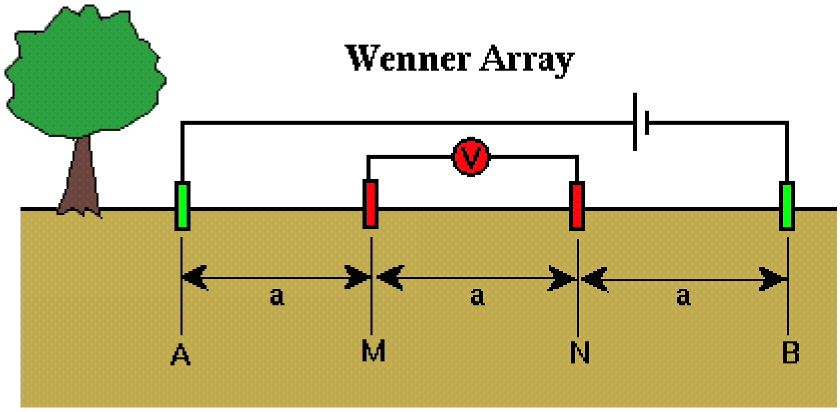
- *Resistivity Soundings* - the resistivity method can detect variations in resistivity that occur solely with depth. In fact, this method is most commonly applied to look for variations in resistivity with depth. Surveys that are designed to determine resistivity variations with depth above some fixed surface location are referred to as *resistivity soundings*. In principle, the two-electrode experiments described previously are examples of soundings. In these experiments, electrode spacing is varied for each measurement. The center of the electrode array, where the electrical potential is measured, however, remains fixed. An example of a problem for which one might employ resistivity soundings is the determination of depth to the water table.

- *Resistivity Profiles* - Like the gravity and magnetic methods, resistivity surveys can also be employed to detect lateral variations in resistivity. Unlike soundings, profiles employ fixed electrode spacings, and the center of the electrode spread is moved for each reading. These experiments thus provide estimates of the spatial variation in resistivity at some fixed electrode spacing. Surveys that are designed to locate lateral variations in resistivity are referred to as *resistivity profiles*. An example of a problem for which one might employ resistivity profiles is the location of a vertical fault.

Resistivity Soundings

When doing resistivity sounding surveys, one of two survey types is most commonly used. For both of these survey types, electrodes are distributed along a line, centered about a midpoint that is considered the location of the sounding. The simplest in terms of the geometry of electrode placement is referred to as a *Wenner* survey. The most time effective in terms of field work is referred to as a *Schlumberger* survey.

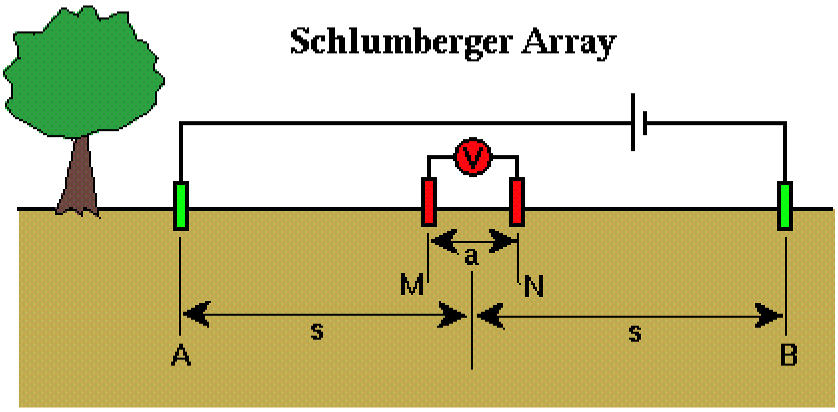
For a Wenner survey, the two current electrodes (green) and the two potential electrodes (red) are placed in line with each other, equidistant from one another, and centered on some location as shown below.



$$\boxed{\rho_a = 2\pi a \frac{\Delta V}{i}}$$

The apparent resistivity computed from measurements of voltage, . □□□□□*V*, and current, *i*, is given by the relatively simple equation shown above. This equation is nothing more than the apparent resistivity expression shown previously with the electrode distances fixed to *a*. To generate a plot of apparent resistivity versus electrode spacing, from which we could interpret the resistivity variation with depth, we would have to compute apparent resistivity for a variety of electrode spacings, *a*. That is, after making a measurement we would have to move all four electrodes to new positions.

For a Schlumberger survey, the two current electrodes (green) and the two potential electrodes (red) are still placed in line with one another and centered on some location, but the potential and current electrodes are not placed equidistant from one another.



$$\rho_a = \frac{\pi(s^2 - a^2/4)}{a}\frac{\Delta V}{i}$$

The current electrodes are at equal distances from the center of the sounding, *s*. The potential electrodes are also at equal distances from the center of the sounding, but this distance, *a/2*, is much less than the distance *s*. Most of the interpretational software available assumes that the potential electrode spacing is negligible compared to the current electrode spacing. In practice, this is usually interpreted to mean that *a* must be less than *2s/5*.

In principle, this implies that we could set *a* to be less than *2s/5* for the smallest value of *s* that we will use in the survey and never move the potential electrodes again. In practice, however, as the current electrodes are moved outward, the potential difference between the two potential electrodes gets smaller. Eventually, this difference becomes smaller than our voltmeter is capable of reading, and we will need to increase *a* to increase the potential difference we are attempting to measure.