

## Apparent Conductivity (or Resistivity) Revisited

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The concept of apparent conductivity (or resistivity) is simple, yet many, including those in the geophysical profession, often misunderstand it. Such misunderstanding is reflected in this typical question: why is the conductivity (or resistivity) measured by instrument A different from that measured by instrument B?

This question arises when we forget the difference between *resistivity* and *apparent resistivity*. *Resistivity* is an intrinsic property of a microscopic volume of material, such as *density*. *Apparent resistivity* is a volume average of a heterogeneous half-space (Figure 1), except that the averaging is not arithmetic but dependent on each instrument and how it is used. This begs another question: if the *apparent resistivity* of a given ground is different for every instrument and for every measurement configuration, how can you make any sense out of it? The answer: *very carefully*.

By the way, the same idea applies to other parameters -- such as *apparent density* or *apparent susceptibility* -- that attempt to represent a real-world heterogeneous earth by an imaginary homogeneous half-space. The thing to remember, though, is that the resulting homogeneous half-space is not a simple average of the original heterogeneous earth

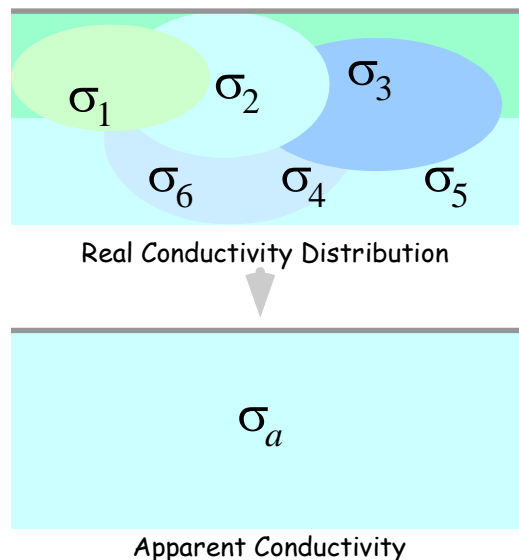


Figure 1. Concept of apparent conductivity is representing a heterogeneous earth as a homogeneous earth having a single conductivity.

### Electrical Method of Determining Apparent Conductivity

The term, *apparent resistivity*, usually first appears in geophysics textbooks in the chapter that deals with the DC electrical method. The unit of resistivity is ohm-m and conductivity is its inverse in siemen/m. The apparent resistivity  $\rho_a$  is defined in DC resistivity as

$$\rho_a = 2\pi G \frac{\Delta V}{I} \quad (1)$$

where  $\Delta V$  is the voltage between a pair of potential electrodes and  $I$  is the current that flows through another pair of source electrodes.  $G$  is called a geometric factor that changes, depending on the geometry of the four electrodes. For a Wenner array that uses four equally-spaced electrodes, for instance,  $G$  is the electrode spacing itself. Even for this simple array, each electrode spacing generates a different apparent resistivity because the spacing controls the volume of the subsurface sampled by the measurement. Only when the earth is a homogeneous half-space is the apparent resistivity the same as the true resistivity.

So, what does it mean when somebody says that a given ground has a resistivity of 30 ohm-m? Not much, unless one specifies how it was measured.

### Electromagnetic Method of Determining Apparent Conductivity

By tradition, the electrical method “measures” resistivity and the EM method “measures” conductivity. The main advantage of EM is that it does not require ground contact, as do the electrodes in the DC electrical survey. An added parameter in EM is the operating frequency. The EM method involves transmitting and receiving the EM fields, commonly using a set of coils. As an example, consider a pair of horizontal coils that are separated by a distance  $r$ . A routinely measured quantity is called the *mutual coupling ratio*, which, for a horizontal coplanar (or vertical dipole) coil configuration over a layered earth, is written as:

$$Q = \frac{H_s}{H_p} = -r^3 \int_0^\infty \lambda^2 R(\lambda) J_0(\lambda r) e^{-2\lambda h} d\lambda. \quad (2)$$

$H_p$  and  $H_s$  are the primary and secondary fields at the receiver coil;  $J_0$  is the 0-th order Bessel function;  $r$  is the coil separation; and  $h$  is the sensor height above the ground.  $Q$  represents the secondary field normalized against the primary field at the receiver coil. Most frequency-domain sensors (ground as well as airborne) measure  $Q$  in a unit of parts-per-million, or ppm. The kernel  $R$  corresponding to a homogeneous half-space is:

$$R(\lambda) = \frac{\lambda - \sqrt{\lambda^2 + i2\pi f \mu \sigma}}{\lambda + \sqrt{\lambda^2 + i2\pi f \mu \sigma}}, \quad (3)$$

where  $f$  is the transmitter frequency in Hz,  $\mu$  the magnetic permeability, and  $\sigma$  the half-space conductivity. Based on measured  $Q$  at a particular frequency over a real (heterogeneous) earth, we can invert (2) to obtain the *apparent* half-space conductivity  $\sigma_a$ . It is obvious from (2) that the resulting  $\sigma$  depends on coil separation, sensor height, and frequency. In addition, each coil configuration (vertical coplanar, coaxial, etc.) has a different formula for  $Q$ .

Figure 2 shows an example where a layered earth (shown in the inset) produces different apparent conductivities depending on transmitter-receiver separations and frequencies. As noted, there is no single apparent conductivity representative of a heterogeneous earth because the apparent conductivity is sensor- and frequency-dependent. As an example, we

should say that the ground has 30 mS/m at 10 kHz with a 2-m coil separation, at 1-m height, and at horizontal coplanar configuration.

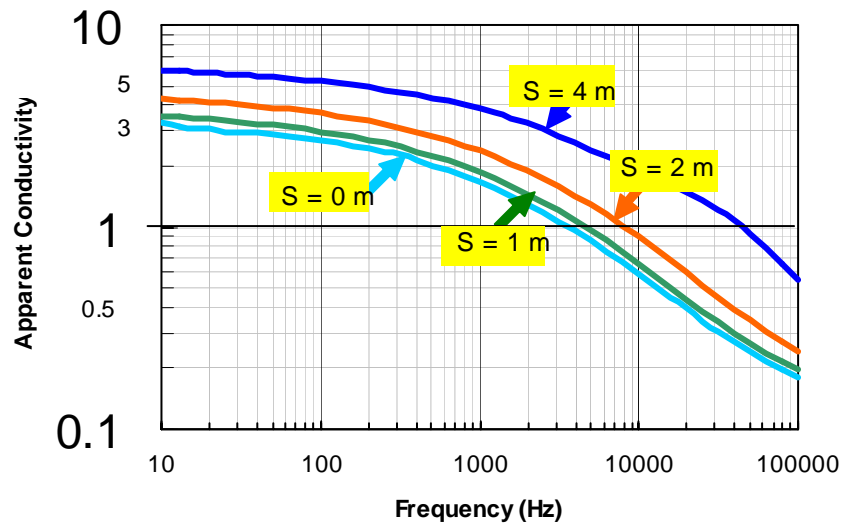
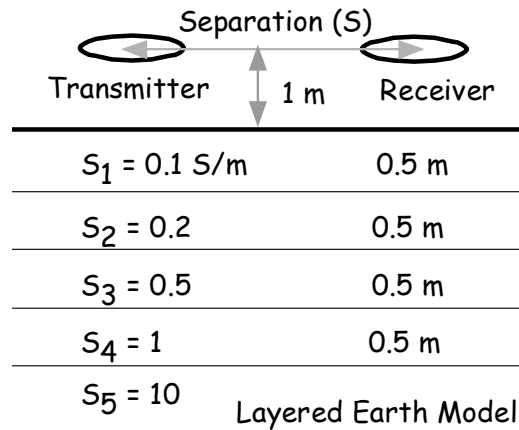


Figure 2. A layered earth shown in the inset produces different apparent conductivities at different transmitter-receiver separations and frequencies.

### Data Interpretation

For fixed sensor geometry, a broadband sensor like the GEM-2 (Won et al., 1996) or the GEM-3 (Won et al., 1997) generates one apparent conductivity map for each frequency at a fixed sensor height. In theory, two maps could be generated – one for the inphase measurement and another one for the quadrature; in practice, the quadrature provides a better signal for geologic applications. Similar results apply to time-domain sensors operating at multiple time gates. It is also possible for a small, shallow, highly conductive feature (such as a man-made metal object) to generate *negative* apparent conductivity with perfectly good data.

The idea of using multiple frequencies stems from the so-called skin-depth that is inversely proportional to frequency: a low-frequency signal travels far through a conductive earth and, thus, "sees" deep structures, while a high-frequency signal can travel only a short distance and thus, "sees" only shallow structures. Therefore, scanning over a range of frequencies is equivalent to depth sounding.

Depth sounding by changing the transmitter frequency is called *frequency sounding*, which measures the target response at many frequencies in order to image the subsurface structure. In contrast, depth sounding by changing the coil separation is called *geometrical sounding*, which usually requires multiple operators tending separate coils connected by wires and measuring consoles. The two sounding concepts are depicted in Figure 3. For shallow surveys, the frequency sounding method offers high spatial resolution, survey speed, ease of logistics, and data precision. *In frequency sounding, there is no obvious relationship between the coil separation and the depth of exploration.* For instance, the co-located coils configuration (zero separation) employed by the GEM-3 does not mean that it has no depth of exploration.

What do we do with all these different apparent conductivity maps at many frequencies? The broadband EM data should be ultimately used to derive a 3-D conductivity structure of the earth through mathematical inversion. Unfortunately, a reliable and cost-effective inversion technique is yet to be discovered. Without rigorous inversion, however, the broadband data are useful for scanning through depths, ensuring coverage of the desired depth range, as long as we clearly understand the dependence of the apparent conductivity (or resistivity) on frequency and other survey parameters.

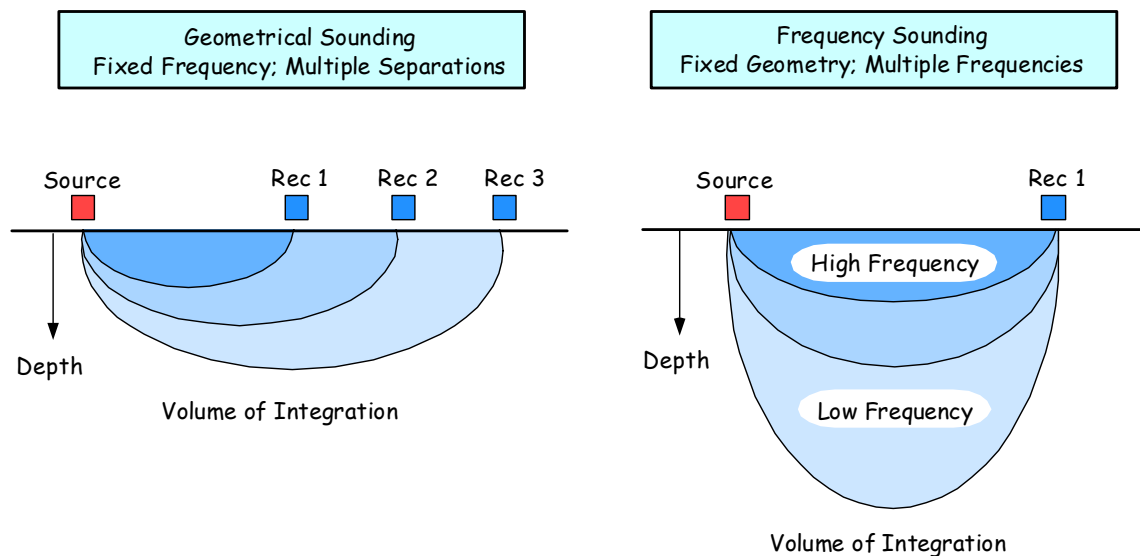


Figure 3. Comparison between geometrical and frequency sounding methods.

## References

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