Appendix A

A.1. Basic Theory of Geoelectrical Resistivity

The comprehensive theory of geoelectrical resistivity method is given in geophysical textbooks such as Robinson (1988), Telford et al. (1990), Burger (1992) and Reynolds (1997). The electrical resistivity method uses the measurement of potential differences at points on the earth's surface that is produced by directing current flow into the subsurface.

The basic geoelectrical resistivity is started from a steady and continuous electric current flow in a homogenous and isotropic conductive half space. The flows of current in the ground are related through Ohm's Law (Telford et al., 1990):

$$J = \sigma E \qquad \text{Equation 1}$$

where σ is the conductivity of the medium (siemens per meter), J is the current density (amperes per square meter) and E is the electric field intensity (volts per meter). The electric field is the negative gradient of electric scalar potential V

$$E = -\nabla V \qquad \text{Equation 2}$$

Since for a stationary current in a homogeneous medium $\nabla.\mathbf{J} = 0$, combination of Equation 1 and 2, therefore

$$\nabla\sigma.\nabla V + \sigma\nabla^2 V = 0 \qquad \text{Equation 3}$$

In region with σ constant, the Eaquation 3 follow Laplace's equation:

$$\nabla^2 V = 0 \qquad \text{Equation 4}$$

A.2. Single Current Electrode

The Laplace's equation can be considered in spherical coordinate due to the spherical symmetry exists in the earth. From the symmetry of the system, the potential is a function of the distance r from the electrode only. The Laplace's equation will give to

$$\nabla^2 V = \frac{d^2V}{dr^2} + \frac{\left(\frac{2}{r}\right)dV}{dr} = 0 \qquad \text{Equation 5}$$

When the electrode delivering current is located at the surface of the homogeneous isotropic medium, since the air is nonconductive, the boundary condition at the surface requires that $E_z = \frac{\partial V}{\partial z} = 0$ at z = 0. Then, for an isotropic half space of constant conductivity with a single current source located at the surface, the solution is:

$$V = -\frac{A}{r} + B \qquad \text{Equation 6}$$

where A and B are constant. When r -> ∞, so V = 0 and B will be equal to 0. The total current crossing a spherical surface is

$$I = 2\pi r^2 J = -2\pi r^2 \sigma \frac{dV}{dr} = -2\pi\sigma A \qquad \text{Equation 7}$$

Since $J = -\sigma \nabla V$ (combination of Equation 1 and 2) and $\sigma = 1/\rho$, so that

$$A = -\frac{I\rho}{2\pi} \qquad \text{hence, } V = \left(\frac{I\rho}{2\pi}\right)\frac{1}{r} \qquad \text{Equation 8}$$

A.3. Two Current Electrodes

Ideally in practice, two current electrodes are required, C1 (source) and C2 (sink). When the distance between the two current electrodes is finite (Figure 1), the potential at any nearby surface point will be effected by both current electrodes. The potential (V_1) due to C1 at P1 is $V_1 = -\frac{A_1}{r_1}$, $A_1 = -\left(\frac{I\rho}{2\pi}\right)$ and the potential (V_2) due to C2 at P1 is $V_2 = -\frac{A_2}{r_2} =$ where $A_2 = \left(\frac{I\rho}{2\pi}\right) = -A_1$. Thus, total potential at P1 due to C1 and C2 is

$$V_1 + V_2 = \frac{I\rho}{2\pi}\left(\frac{1}{r_1} - \frac{1}{r_2}\right) \qquad \text{Equation 9}$$

Finally, by knowing total potential at P2 due C1 and C2, then the difference in potential between P1 and P2 will be

$$\Delta V = \frac{I\rho}{2\pi}\left\{\left(\frac{1}{r_1} - \frac{1}{r_2}\right) - \left(\frac{1}{r_3} - \frac{1}{r_4}\right)\right\} \qquad \text{Equation 10}$$

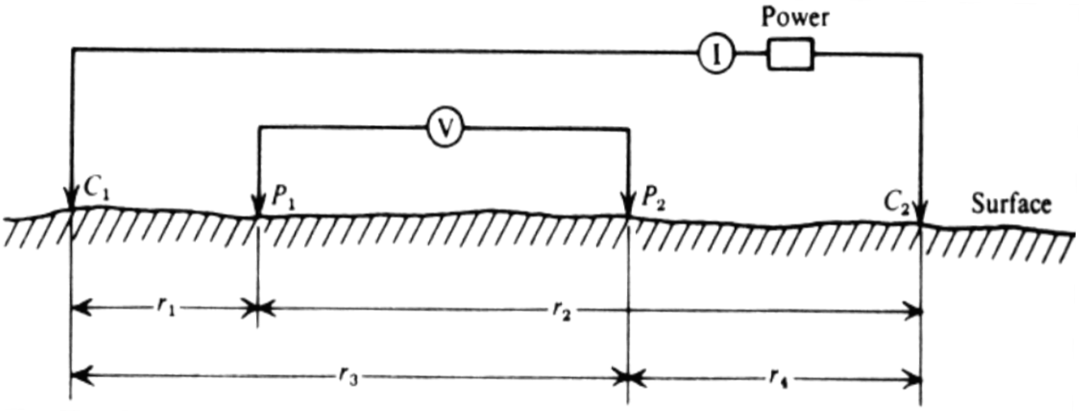


Figure 1. Two current electrodes and two potential electrodes on the surface of homogeneous isotropic ground of resistivity ρ (Telford et al., 1990).

A.4. Apparent Resistivity

The Equation 10 gives the potential that would be observed over a homogeneous half-space with a typical four electrodes configuration (two current electrodes and two potential electrodes). Over homogeneous isotropic ground this resistivity will be constant for any current and electrode arrangement. The subsurface is typically heterogeneous so that the resistivity observed is known as the apparent resistivity (ρ_a).

The apparent resistivity depends on the configuration of the electrodes and is determined by the injected current I and voltage ΔV. Thus, the apparent resistivity is expressed as

$$\rho_a = G\frac{\Delta V}{I} \qquad \text{Equation 11}$$

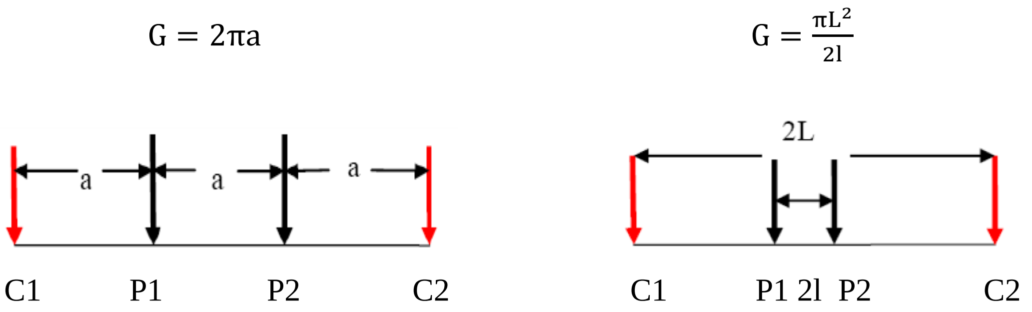
The geometric factor G in the Equation 11, which depends on the electrode configuration, is given as

$$G = \frac{2\pi}{\left[\left(\frac{1}{r_1}-\frac{1}{r_2}\right)-\left(\frac{1}{r_3}-\frac{1}{r_4}\right)\right]} \qquad \text{Equation 12}$$

A.5. Electrodes Configuration

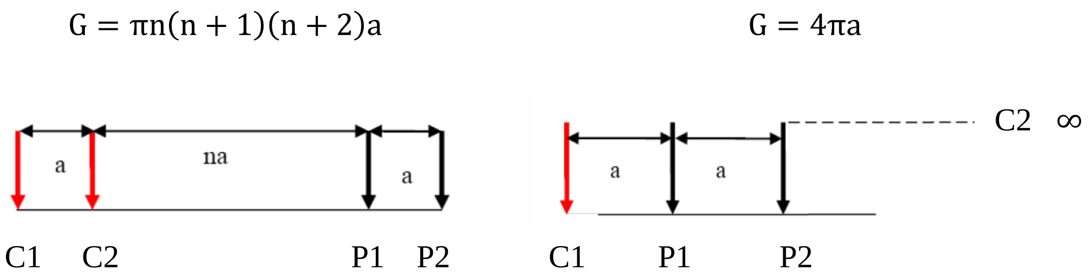
The electrode arrays most commonly used include Wenner (alpha), Schlumberger, dipole-dipole and pole-dipole arrays (Loke, 2004; Telford et al., 1990; Robinson, 1988). Electrode arrays with their corresponding geometric factor (G) are illustrated in Figure 2. The geometric factor is different from one to another due to the difference in the electrode configuration. The apparent resistivity values observed by the different array types over the same structure can be different (Schrott and Sass, 2008; Kneisel, 2006, Loke, 2004). The choice of a particular array depends on a number of factors, which include the sensitivity of the array to vertical and lateral variations in the resistivity of the subsurface, its depth of investigation, the horizontal data coverage and signal strength of the array (Schrott and Sass, 2008; Loke, 2004; Kneisel, 2006; Griffiths and Barker, 1993). Another importance consideration is a survey time taken for certain case.

If the measurement is done for a small electrode spacing, the apparent resistivity becomes the true resistivity of the material assuming that it is homogeneous (Telford, et al. 1990).



Wenner (Wenner Alpha) array

Schlumberger array



Dipole-dipole arrays

Pole-dipole arrays

Figure 2. Electrode configuration in geoelectrical resistivity survey.

In this research the equipment utilized for all geoelectrical resistivity data acquisition uses Terrameter SAS4000 manufactured by ABEM (Abem, 2007) with Wenner configuration. The Wenner array shows a good signal–noise ratio and is favourable for the detection of horizontal layers. The signal strength is inversely proportional to the geometric factor used to calculate the apparent resistivity value for the configuration. For the Wenner configuration, the geometric factor is smaller than the geometric factor for other configurations (see Telford, et al. 1990 and other text books for detail). Among the common configurations, the Wenner configuration has the strongest signal strength. This can be an important factor if the survey is carried in areas with high background resistivity changes (Schrott and Sass, 2008; Kneisel, 2006; Loke, 2004; Abu-Shariah, 2002). In additional, the Wenner configuration has a moderate investigation depth and has a good resolution for horizontal structures with vertical changes of resistivity. Since the total number of measurements required is less than for other configurations the time to complete a survey is comparatively short, however, also the information of the subsurface obtained is less than that derived from other arrays (Kneisel, 2006). Relatively shorter survey time taken is needed when to survey the same location with high resolution.

A.6. Depth of Investigation

A depth of investigation depends on the electrode spacing. Increasing of electrode spacing, the array configuration can reach the deeper target. Loke (2004) used the sensitivity function (Edwards, 1977) to obtain the depth of investigation. In resistivity sounding surveys, the subsurface is assumed to consist of horizontal layers. For horizontal layer, the layer in direction of x and y extends from – infinite to + infinite. The sensitivity function can be obtained by integrating 3D the following sensitivity function

$$F_{1D}(z) = \frac{1}{4\pi^2}\int_{-\infty}^{+\infty}\int_{-\infty}^{+\infty}\frac{x(x-a)+y^2+z^2}{[x^2+y^2+z^2]^{1.5}[(x-a)^2+y^2+z^2]^{1.5}}\,dxdz \qquad \text{Equation 13}$$

The Equation 14 has a simple analytical solution by Roy and Apparao (1971) (Loke, 2004), which is given by

$$F_{1D}(z) = \frac{2}{\pi}\frac{z}{(a^2+4z^2)^{1.5}} \qquad \text{Equation 14}$$

This equation is also known as the depth investigation characteristic and has been used by many authors (Loke, 2004; Merrick 1997; Barker, 1991; Edwards, 1977). Figure 3 shows a plot of this function for Wenner array. Base on Figure 3, some authors have used the the maximum point as the depth of investigation of the array. Edwards (1977) and Barker (1991) has introduced that a more robust estimate is the "median depth of investigation". That means, depth of investigation for Wenner array is 0.519 time electrode spacing.

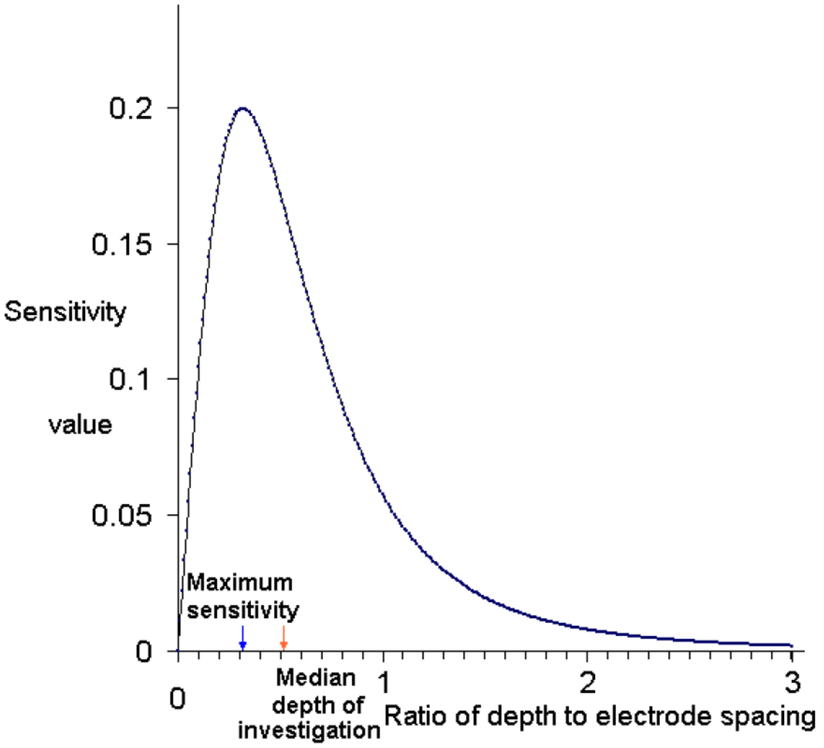


Figure 3 The sensitivity function and median depth of investigation for the Wenner array (Loke, 2004).

A.7. Basic Inversion Concept

The comprehensive theory of “least-squares” inversion method used in Res2DINV programs can be found in Loke (1995, 1996, 2002 and 2007). The basic is aimed at finding a resistivity distribution that gives a response similar to the actual measured values. Res2DINV programs, where the calculated input parameters are the resistivity values of the cell (model blocks), while the data is the calculated apparent resistivity values. It is well known that for the same measured field output data set, there is a wide range of resistivity distribution whose calculated apparent resistivity values agree with the measured values (Loke, 1999, 2004). This inversion software tries to reduce the difference between the calculated and measured apparent resistivity values by adjusting the resistivity of the model blocks. A measure of this difference is given by the root mean square error (RMS). However, the best model from a geomorphological or geological perspective might not be the one with the lowest possible RMS (Loke, 1999, 2004; Kneisel, 2006). Thus, it is essential to perform the interpretation with consideration of the local geomorphological setting.

A.8. Electrical Properties of Earth Materials

The relationship between the resistivity of a porous rock and the fluid saturation factor is given by Archie's Law which gives applicable for certain types of rocks and sediments, particularly those that have a low clay content.

$$\rho = a\rho_w \emptyset^{-m} \qquad \text{Equation 15}$$

where ρ is the rock resistivity, ρ_w is fluid resistivity, ϕ is the fraction of the rock filled with the fluid, while a and m are two empirical parameters (Loke, 2004). For most rocks, a is about 1 while m is about 2. Under certain special conditions, the above equation can be used to determine the change in the fluid saturation or fluid resistivity with time.

According to Robinson (1988), electrical currents can flow from one earth material to another earth material due to the following factors:

1. Free electron movement or the existence of ions in the material. The movement can occur within minerals such as graphite, magnetite, or pyrite
2. Ion movement in water. Ions can form from salt material such as sodium chloride and magnesium chloride.

The electrical properties of minerals and rocks have been a subject of study for almost two centuries. Laboratory measurements of resistivity and dielectric constant have been made for most known varieties of minerals, rocks, soil and unconsolidated sediment, and for groundwater. In addition, many field measurements have been made with these substances in their natural settings. Typical values of resistivity are presented in Table 1. (Robinson, 1988)

There is considerable variation in resistivity between lithologically and mineralogically similar specimens. Natural metals and metallic ore minerals tend to be good conductors of electric current. Otherwise, most minerals are insulators or highly resistive conductors. Rocks consisting of metallic ore minerals have broader but generally lower ranges of resistivity, largely because water is present in the pore space of the rock.

The values measured in a survey of electrical resistivity depend much more on rock porosity and nature of the groundwater than on lithology. For this reason, there are no generally applicable guidelines for interpreting the lithology of the zones of the contrasting resistivity delineated by a survey. Interpretation may be possible on a local basis, however, if there is independent information about the nature of the rocks in the area of the survey (Robinson, 1988).

Table 1. Resistivities of Earth Materials (Robinson, 1988).

Earth material	Resistivity, Average or Range (ohm.m)	Earth material	Resistivity, Average or Range (ohm.m)
Metal		*Other Mineral*	
Copper	1.7×10^{-8}	Calcite	$10^{12} – 10^{13}$
Gold	2.4×10^{-8}	Anhydrite	$10^{9} – 10^{10}$
Silver	1.6×10^{8}	Halite	$10 – 10^{13}$
Graphite	10^{-3}	Coal	$10 – 10^{11}$
Iron	10^{-7}		
Lead	2.2×10^{-7}	*Crystalline Rocks*	
Nickel	7.8×10^{-7}		
Tin	1.1×10^{-7}	Granite	$10^{2} – 10^{6}$
Zinc	5.8×10^{-8}	Diorite	$10^{4} – 10^{5}$
		Gabbro	$10^{3} – 10^{6}$
Sulfide Ore Minerals		Andesite	$10^{2} – 10^{4}$
		Basalt	$10 – 10^{7}$
Chalcocite	10^{-4}	Peridotite	$10^{2} – 10^{3}$
Chalcopyrite	4×10^{-3}	Schist	$10 – 10^{4}$
Pyrite	3×10^{-1}	Gneiss	$10^{4} – 10^{6}$
Pyrrhotite	104	Slate	$10^{2} – 10^{7}$
Molybdenite	10	Marble	$10^{2} – 10^{8}$
Galena	2×10^{-3}	Quartzite	$10 – 10^{8}$
Sphalerite	102		
		Sedimentary Rocks	
Oxide Ore Minerals			
		Shale	$10 – 10^{3}$
Bauxite	$10^{2} – 10^{4}$	Sandstone	$1 – 10^{8}$
Chromite	$1 – 10^{6}$	Limestone	$50 – 10^{7}$
Cuprite	$10^{-3} – 300$	Dolomite	$102 – 10^{4}$
Hematite	$10^{-3} – 10^{7}$		
Magnetite	$10^{-5} – 10^{4}$	*Unconsolidated Sediment*	
Ilmenite	$10^{-3} – 10^{2}$		
Rutile	$10 – 10^{3}$	Sand	$1 – 10^{3}$
		Clay	$1 – 10^{2}$
Silicate minerals		Marl	$1 – 10^{2}$
Quartz	$10^{10} – 10^{15}$	*Groundwater*	
Muscovite	$10^{2} – 10^{14}$		
Biotite	$10^{2} – 10^{6}$	Portable well water	$0.1 – 10^{3}$
Hornblende	$10^{2} – 10^{6}$	Brackish water	0.3 – 6
Feldspar	$10^{2} – 10^{4}$	Seawater	0.2
Olivine	$10^{3} – 10^{4}$	Supersaline brine	0.05 – 0.2

A.9. Limitations

a. Field Limitations

Geoelectrical resistivity methods produce inaccurate representation of the subsurface when the surveyed fields are damp, swampy, or puddles of water after heavy rainfall. Puddles of water will influence the geoelectrical resistivity reading, which affect some of the potential difference measurements. The result of this situation is the slight distortion in the resistivity contours of the inverse model section (Loke, 2004; Abdul Nassir, 1997)

Another the field limitation in the survey was a lack of available space to lay the survey cables. Undulating and sneaking surfaces made it difficult to conduct the 2D survey which requires straight lines to lay the cables. Although the undulating areas can still be surveyed, the topography of the data had to be included in the processing. Other limitation are the safety reason, whereas cables had to be far from moving vehicles to avoid damage caused by the weight of vehicle and induction.

The ABEM Terrameter SAS4000 was that maximum spreading of the four wheel cables are only 400 meters. This spreading can reach around 60 m depth. In the other hand, the electrodes spacing had to be closer to produce a higher resolution but reduces depth penetration.

There are a number of noise sources that can affect the measurements of voltage and current, and are describe below.

Presence of Nearby Conductors

Geoelectrical resistivity surveys cannot be done near conductors that make contact with the ground. For example, the presence of buried pipes (pipe conductor) or chain-linked fences will act as current sinks. The current will preferentially flow along these structures rather than flowing through the earth because of their low resistivity. The presence of these nearby conductors essentially acts as electrical shorts in the system and it should be avoid (Robinson, 1988).

Electrode Polarization

A metallic electrode like a copper or steel rod will generate a measurable contact potential if in contact with an electrolyte other than a saturated solution. These potentials can be a significant fraction of the total potential measured. Knowing the near surface of the field prior doing survey is very important to avoid the noise effect (Robinson, 1988).

Telluric Currents

Naturally existing currents flow within the earth is referred to as telluric currents. The existence 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. In practice, natural electrical currents that must be avoided is when the lightning flash (Robinson, 1988).

b. Limitations in Theoretical

The geoelectrical resistivity technique includes the limitation involved in the interpretation of complex geology and the existence of natural currents and potentials. Geological mapping is necessary to do in the field prior to interpretation stage.

If there is thin layer in between thicker layer, it is very difficult to interpret. This problem can be solved by increasing the density of data point. More density of data point can be obtained with reducing the electrodes spacing. However, the depth investigation will be shallower. This resolution limitation is common in all geophysical methods (Heiland, 1968).

The non-uniqueness of the solution in the inversion scheme can lead to ambiguity or misinterpretation. Thus, a basic knowledge of the medium under study is recommended before interpretation stage (Samouelian, 2005). Turesson (2006), Garambois et al. (2002) and Slater and Reeve (2002) recommend that to reduce the inversion ambiguity and to improve the quality of the interpretation, data from several different prospecting methods should be integrated.

In this research, the direct surface resistivity measurements, borehole geophysics data, hydrogeochemical and soil properties analysis data were used to reduce the ambiguity and improved quality of the interpretation. Unfortunately, the borehole geophysics data is only available in several palaces.

A.10. Borehole Geophysics

Borehole geophysics is the science of recording and analyzing measurements of physical properties made in wells or test holes. The main objective of obtaining borehole data is to get direct information of the subsurface with depth. Probes that measure different properties are lowered into the borehole to collect continuous or point data that is graphically displayed as a geophysical log (Robinson, 1988).

Borehole geophysics is used in groundwater and environmental investigations to obtain information on rock lithology and fractures, permeability and porosity, and water quality. The common borehole geophysics used in the hydrogeology study are gamma ray, self potential (SP), resistivity and neutron porosity. In this research, the existing natural gamma ray data was used to study subsurface litology.

Gamma Ray

Gamma ray measurements detect variations in the natural radioactivity originating from changes in concentrations of the trace elements uranium (U) and thorium (Th) as well as of the major rock forming element potassium (K) (Baker, 2002).

Since the concentrations of these naturally occurring radioelement vary between different rock types, natural gamma-ray logging provides an important tool for lithologic mapping and stratigraphic correlation. Gamma-ray logs are important for detecting alteration zones, and for providing information on rock types. For example, in sedimentary rocks, sandstones can be easily distinguished from shales due to the low potassium content of the sandstones compared to the shales. Potassium in the micas and the clay minerals produced by decomposition of micas and alkali feldspars includes small portion of the radioactive isotope potassium-40. Trace amounts of uranium and thorium also occur in shale. These unstable elements produce measurable levels of gamma radiation (Killeen, 1986).

In sedimentary rocks, potassium is, in general, the principal source of natural gamma radiation, primarily originating from clay minerals such as illite and montmorillonite. In igneous and metamorphic geologic environments, the three sources of natural radiation (U, Th and K), may contribute equally to the total gamma radiation detected by the gamma probe. Because gamma rays can be detected through steel, logging can be done inside drill rod or casing with a slight decrease in sensitivity (Baker, 2002).

By using gamma ray data, it is possible to do quantitative evaluation of shale/clay content. It assumes that radioactive minerals are absent other than in shale and clay. The gamma ray shale index (I_{GR}) is defined as (Baker, 2002):

$$I_{GR} = (GR - GR_{cn})/(GR_{SH} - GR_{cn}) \qquad \text{Equation 16}$$

Where

GR = log response in zone of interest (API units)

GR_{cn} = Log response in a zone considered clean, shale free (API unit)

GR_{SH} = Log response in a shale bed (API units)

I_{GR} has been empirically correlated to fractional volumes of shale in otherwise clean reservoir rock to provide a correction to the linier I_{GR} response in rocks of certain ages. When the formations are Tertiary or Quartenary and especially those that are unconsolidated, the following equation is used to overestimates shale/clay volumetric:

$$V_{SH} = 0.083\left(2^{3.7I_{GR}-1}\right) \qquad \text{Equation 17}$$

While for the true shaliness correlation in pre-Tertiary-age rock (Mesozoic and Paleozoic), the following equation is needed:

$$V_{SH} = 0.33\left(2^{2I_{GR}-1}\right) \qquad \text{Equation 18}$$

Limitations

The limitations of borehole geophysics in the field are almost non-existent. However the problems encountered are the lack of drilling area due to permission and authorization issues with the local residents. Another problem arises as a result of the characteristics of the log equipment. Casing wells made of metal or PVC requires correction of the gamma ray value. Establishing the distinction between sand and limestone is the main objective of the interpretation stage. However the value of natural gamma ray responses may be similar for two separate formations. Understanding of the geology study area gives a great additional value in interpretation stage.

Appendix B

B.1. Hydrogeochemical Method

Geoelectrical resistivity methods provide information in terms of resistivity variations in the subsurface. However, it does not give further details on the exact condition of groundwater in terms of total dissolved solid (TDS), pH, and other cation and anion content. Other methods are needed to support the geoelectrical resistivity method. In this research, a combination of geoelectrical resistivity, hydrogeochemichal, borehole geophysics and soil property analysis methods have been used.

Two conditions of measurement needed to obtain hydrogeochemical data are (i) in-situ parameters measured at the site and (ii) cations and anions concentration in the water sample.

B.2. In-Situ Parameters

There are two types of in-situ parameters measurement. They are (1) water and (2) physical well. The water in-situ parameters include pH, temperature, conductivity, total dissolved solid and salinity. The physical well parameters are well depth, water level, well coordinate and ground level. All the in-situ parameters are measured in the site.

pH

pH is a measure of the acidity or basicity of a solution. It is defined as the activity of dissolved hydrogen ions (H^+). Hydrogen ion activity coefficients cannot be measured experimentally, so they are based on theoretical calculations. The pH scale is not an absolute scale; it is relative to a set of standard solutions whose pH is established by international agreement (Buck, 2001).

Pure water is said to be neutral. The pH for pure water at 25 °C (77 °F) is close to 7.0. In general, water with a low pH (< 6.5) could be acidic, soft, and corrosive. Therefore, the water could contain metal ions such as iron, manganese, copper, lead, and zinc or, on other words, elevated levels of toxic metals (Convington, 1985).

Water with a pH > 8.5 could indicate that the water is hard. Hard water does not pose a health risk, but can cause aesthetic problems. These problems include an alkali taste to the water, formation of a deposit on dishes, utensils, and laundry basins, difficulty in getting soaps and detergents to lather, and formation of insoluble precipitates on clothing. The ideal pH level of drinking water should be between 6-8.5, the human body maintains pH equilibrium on a constant basis and will not be affected by water consumption. For example our stomachs have a naturally low pH level of 2 which is beneficial acid that helps us with food digestion. pH measurements are important in medicine, biology, chemistry, food science, environmental science, oceanography and many other applications (Convington, 1985).

Total Dissolved Solid

Total Dissolved Solids (TDS) is an expression for the combined content of all inorganic and organic substances contained in a liquid which are present in a molecular, ionized or micro-granular (colloidal sol) suspended form. Generally the operational definition is that the solids must be small enough to survive filtration through a sieve size of two micrometres. Total dissolved solids are normally only discussed for freshwater systems, since salinity comprises some of the ions constituting the definition of TDS. The principal application of TDS is in the study of water quality for groundwater, streams, rivers and lakes as an aggregate indicator of presence of a broad array of chemical contaminants (Schofield, 2004).

Primary sources for TDS in receiving waters are agricultural runoff, leaching of soil contamination and point source water pollution discharge from industrial or sewage treatment plants. More exotic and harmful elements of TDS are pesticides arising from surface runoff (Grasby, 1997). Certain naturally occurring total dissolved solids arise from the weathering and dissolution of rocks and soils. The United States has established a secondary water quality standard of 500 mg/L to provide for palatability of drinking water.

Temperature

Ground water temperature is affected by atmospheric temperature where located near the earth surface. Influence of atmospheric temperature to ground temperature become smaller as the depth is deep. The relationship between the ground water temperature and the annual average temperature given by (Isamu et al., 2002):

$$T_e = 0.83\ T_a + 3.7 \qquad \text{Equation 19}$$

Where : Te = constant temperature of ground water (^{0}C)

Ta = annual average temperature (^{0}C)

Salinity

Salinity is the level of soluble salts in soils or waters. In general, this is used to describe the presence of elevated levels of different salts such as sodium chloride, magnesium and calcium sulphates and bicarbonates, in soil and water. It usually results from water tables rising to, or close to, the ground surface (Tao et al., 2009, Carter et al., 2008).

The three main types of salinity are:

a. Dry land salinity, which is caused when saline seeps scald the surface of non-irrigated lands, in turn affecting plant growth and degrading soil structure

b. Irrigation salinity, which results when overuse of water for irrigation causes rising watertables

c. Sea water intrusion, which occurs in coastal aquifer systems where sea water replaces groundwater that has been over-exploited.

Conductivity

The conductivity of a substance is defined as the ability or power to conduct or transmit heat, electricity, or sound. Its units are Siemens per meter [S/m] in SI and millimhos per centimeter [mmho/cm]. Its symbol is k or s. Conductivity is inversely proportional with resistivity (Hounslow, 1995).

Pure water is not a good conductor of electricity. Ordinary distilled water in equilibrium with carbon dioxide of the air has a conductivity of about 20 dS/m. Because the electrical current is transported by the ions in solution, the conductivity increases as the concentration of ions increases. Thus conductivity increases as water dissolved ionic species. The relation between resistivity and conductivity is (Hounslow, 1995)

$$R\ (\text{ohm.cm}) = 10^6 / C\ (\mu\text{S/cm}) \qquad \text{Equation 20}$$

B.3. Physical In-situ Parameters

The physical conditions of the well include the ground level where the water sample is collected, well depth and water level. The X Y coordinate of the well is collected to determine the location. While ground level data is collected to determine the position of the well is installed relatively to the mean sea level. Knowing the water level is useful to define direction of groundwater movement. Another use of water level is to see whether two or more wells with different penetration depths (at the same location) communicate with each other i.e. whether they are hydraulically connected.

B.4. Cation and Anion

Major Cations

1. Calcium (Ca)

Calcium is the element with atomic number 20. Calcium is a soft grey alkaline earth metal, and is the fifth most abundant element by mass in the Earth's crust. Calcium is also the fifth most abundant dissolved ion in seawater by both molarity and mass, after sodium, chloride, magnesium, and sulphate (Dickson, 1994).

Water that contains a lot of calcium and magnesium is said to be hard (LANL, 2003). The hardness of water is expressed in terms of the amount of calcium carbonate, the principal constituent of limestone-or equivalent minerals that would be formed if the water were evaporated (Hounslow, 1995). Water is considered soft if it contains 0 to 60 mg/L of hardness, moderately hard from 61 to 120 mg/L, hard between 121 and 180 mg/L, and very hard if more than 180 mg/L. Very hard water is not desirable for many domestic uses; it will leave a scaly deposit on the inside of pipes, boilers, and tanks.

Hard water can be softened at a fairly reasonable cost, but it is not always desirable to remove all the minerals that make water hard (USGS, 2000).

2. Magnesium (Mg)

Magnesium, with atomic number 12 and common oxidation number +2, an alkaline earth metal, is the ninth most abundant element in the universe by mass. Magnesium constitutes about 2% of the Earth's crust by mass, which makes it the eighth most abundant element in the crust. Magnesium ion's high solubility in water helps ensure that it is the third most abundant element dissolved in seawater (Peter, 1985). Magnesium ions are sour to the taste, and in low concentrations help to impart a natural tartness to fresh mineral waters (Bernath, 1985)

3. Sodium (Na)

Sodium has atomic number 11 and group as a soft metal, reactive and with a low melting point. Sodium reacts quickly with water, and also with snow and ice, to produce sodium hydroxide and hydrogen. All natural waters contain sodium ions as one of the most abundant element on the planet. High concentrations of sodium in inland waters, however, are normally associated with pollution from industrial discharges or sewage effluent or, in coastal areas, sea water intrusion. Normally, sodium concentrations are below 200 mg/L (WHO, 1984). Most sodium in land water results from natural ion exchange, where clay reacts with calcium and magnesium and releases sodium (Hounslow, 1995).

4. Potassium (K)

Potassium is the chemical element with atomic number 19. It is a soft silvery-white metallic alkali metal that oxidizes rapidly in air and is very reactive with water.

Potassium in nature occurs only as ionic salt. As such, it is found dissolved in seawater, and as part of many minerals.

Potassium is commonly found in soils and rocks. In soils containing appreciable amounts of clay, these metals are not mobile. Sodium and potassium are released slowly upon dissolution of rocks. Consequently, potassium concentrations increase as residence time in ground water increases (MPCA, 1999).

Potassium is an important element in chemical fertilizer (Yang et al., 2006). It is strongly held by clay particles in soil. Therefore, leaching of potassium through the soil profile and into groundwater is important only on coarse-textured soils. Potassium is relatively soluble in groundwater and its concentrations increase with time (MPCA, 1999).

5. Iron (Fe)

Iron is a metal that occurs naturally in soils, rocks and minerals. In the aquifer, groundwater comes in contact with these solid materials dissolving them, releasing their constituents, including Fe into the water. At concentrations approaching 0.3 mg/L of Fe, the water's usefulness may become seriously impacted, e.g., there may be a metallic taste to the water. At this concentration, however, the health risk of dissolved Fe in drinking water is insignificant.

Dissolved Fe in groundwater depends on the amount of oxygen in the water and degree of acidity, i.e., its pH. Iron, can occur in two forms: as Fe2+ and as Fe3+. When levels of dissolved oxygen in groundwater are greater than 1- 2 mg/L, iron occurs as Fe3+, while at lower dissolved oxygen levels, the iron occurs as Fe2+. Although Fe2+ is very soluble, Fe3+ will not dissolve appreciably. If the groundwater is oxygen poor, iron will dissolve more readily, particularly if the pH of the water is on the low side (slightly more acidic) (Drever, 1982).

Dissolved oxygen content is typically low in deep aquifers, particularly if the aquifer contains organic matter. Decomposition of the organic matter depletes the oxygen in the water and the iron dissolves as Fe^{2+}. Under these conditions, the dissolved iron is often accompanied by dissolved manganese or hydrogen sulphide. When this water is excavated to the surface, the dissolved iron reacts with the oxygen in the atmosphere, changes to Fe^{3+} (i.e., is oxidized) and forms rust-colored iron minerals (Drever, 1982).

B.5. Major Anions

1. Chloride

The chloride ion is formed when the element chlorine picks up one electron to form an anion (negatively-charged ion) Cl^-. The salts of hydrochloric acid HCl contain chloride ions and can also be called chlorides. The word chloride can also refer to a chemical compound in which one or more chlorine atoms are covalently bonded in the molecule. This means that chlorides can be either inorganic or organic compounds. The simplest example of an inorganic covalently-bonded chloride is hydrogen chloride, HCl. A simple example of an organic covalently-bonded (an organochloride) chloride is chloromethane (CH_3Cl), often called methyl chloride. Common source of chloride are halite (NaCl), sea spyar, brines, and hot springs (Hounslow, 1995).

Chloride is also a useful and reliable chemical indicator of river / groundwater faecal contamination, as chloride is a non-reactive solute and ubiquitous to sewage and potable water.

2. Nitrate

In inorganic chemistry, a nitrate is a salt of nitric acid with an ion composed of one nitrogen and three oxygen atoms (NO_3^-). Nitrate occurs in almost all natural waters (Hounslow, 1995). However, nitrate contamination of groundwater is a worldwide problem (Birkinshaw & Ewen, 2000; Saadi & Maslouhi, 2003). Nitrate is soluble and negatively charged and thus has a high mobility and potential for loss from the unsaturated zone by leaching (Chowdary et al., 2005; Lee, et al., 2006). Elevated nitrate concentrations in drinkingwater can cause methemoglobinemia in infants and stomach cancer in adults (Wolfe and Patz, 2002).

Groundwater pollution is caused mainly by agricultural practices such as the use of chemical or natural fertilizers, pesticides, and herbicides. Localized industrial activities (organic pollutants and heavy metals) and inadequate or improper disposal of wastewater and solid waste including hazardous materials is also to be considered as the main ground water pollution (Almasri and Kaluarachchi, 2004).

Nitrification is a microbial process by which reduced nitrogen compounds (primarily ammonia) are sequentially oxidized to nitrite and nitrate. Ammonia is present in drinking water through either naturally-occurring processes or through ammonia addition during secondary disinfection to form chloramines. The nitrification process is primarily accomplished by two groups of autotrophic nitrifying bacteria that can build organic molecules using energy obtained from inorganic sources, in this case ammonia or nitrite (USEPA, 2002; 1980). In the first step of nitrification, ammonia-oxidizing bacteria oxidize ammonia to nitrite according to the equation in 13.

$$NH_3 + O_2 \longrightarrow NO_2^- + 3H^+ + 2e^- \qquad \text{Equation 21}$$

Nitrosomonas is the most frequently identified genus associated with this step, although other genera, including Nitrosococcus, and Nitrosospira. Some subgenera, Nitrosolobus and Nitrosovibrio, can also autotrophically oxidize ammonia (Simon, 2001). In the second step of the process, nitrite-oxidizing bacteria oxidize nitrite to nitrate according to equation 14.

$$NO_2^- + H_2O \longrightarrow NO_3^- + 2H^+ + 2e^- \qquad \text{Equation 22}$$

Nitrobacter is the most frequently identified genus associated with this second step, although other genera, including Nitrospina, Nitrococcus, and Nitrospira can also autotrophically oxidize nitrite (Simon, 2001).

Nitrate is quite soluble in water and is not significantly adsorbed by clay-rich soils because it is an anion. Nitrate represents the stable end product of the nitrification process. Usually, nitrification occurs mostly in the aerobic unsaturated zone. It is common knowledge that nitrification is an oxic process. Below the groundwater table, nitrate is reduced to nitrogen gas by denitrification. This process occurs primarily in anoxic conditions and involves heterotrophic bacteria (the major genera is Paracoccus, Pseudomonas) (Lee et al., 2006).

3. Sulphate

The sulfate ion is a polyatomic anion with the empirical formula SO_4^{2-} and a molecular mass of 96.06 daltons; it consists of a central sulfur atom surrounded by four equivalent oxygen atoms in a tetrahedral arrangement. The sulfur atom is in the +6 oxidation state while the four oxygen atoms are each in the −2 state. The sulfate ion carries a negative two charge (Drever, 1982).

Appendix C

Soil Properties Analysis

C.1. Grain Size Soil Distribution

Grain-size distribution is commonly used for soil classification. However, the grain-size distribution is also used as a basis for estimating soil behaviour. The size of particles that make up soil vary over a wide range. Soil is generally called gravel, sand, silt or clay, depending on the predominant size of particles within the soil (Das, 2001).

Hamlin (1991) reported that the gravel grain is more than 2 mm in diameter. The sand grains range from 0.0625 mm to 2 mm in diameters. Some authors divide the sand-size grain into fine sand, medium sand and coarse sand. Whilst, silt is ranging from 0.003 mm to 0.0625 mm. Clay is the particles that make up the soil are less than 0.003 mm in diameter.

In this study, grain-size distribution is used to classify the litology of the sampled area. It is also done to correlate soil character with surface resistivity values. This is very significant when interpreting the geoelectrical model.

C.2. Soil Hydraulic Conductivity.

Hydraulic conductivity, most commonly represented as K, is a property of vascular plants, soil or rock that describes the ease with which water can move through pore spaces or fractures. It depends on the intrinsic permeability of the material and on the degree of saturation (Bear, 1972; Nazrul, 2006). The pore-size distribution, the regularity of the pores, and their continuity has a great influence on the soil's hydraulic conductivity. Nevertheless, the study and characterization of the porosity aiming at an assessment of the K is not sufficiently advanced to be practical on a large scale. The saturated hydraulic conductivity is a key parameter when considering flow through porous media and the flow parameter that does possess the largest uncertainty (Petter, 1990).

In this study, the inversed auger method was employed to derive K value. This type of test is also known as the Porchet method (Oosterbaan and Nijland, 1994; van Hoorn, 2007; Van Genuchten, 1980). The inversed auger hole method, consists of boring a hole to a given depth, filling it with water, and measuring the rate of fall of the water level. The surface over which water infiltrates into the soil at time, t, equals (see Figure 21.):

$$A_t = 2\pi r h_t + \pi r^2 \qquad \text{Equation 23}$$

According to the law of Darcy:

$$Q_t = K\,A_t = 2\,K\,\pi\,r\,(h_t + r/2\,) = -\,\pi\,r^2\,dh/dt \qquad \text{Equation 24}$$

Integrating between the limits t = 0, h_0 and t, h_t and rearrangement of the equation yields:

$$K = 1.15\,r\,\left(\frac{log\left(h_0 + {}^{r}/_{2}\right) - log\left(h_t + {}^{r}/_{2}\right)}{t}\right) = 1.15\,r\tan\alpha \qquad \text{Equation 25}$$

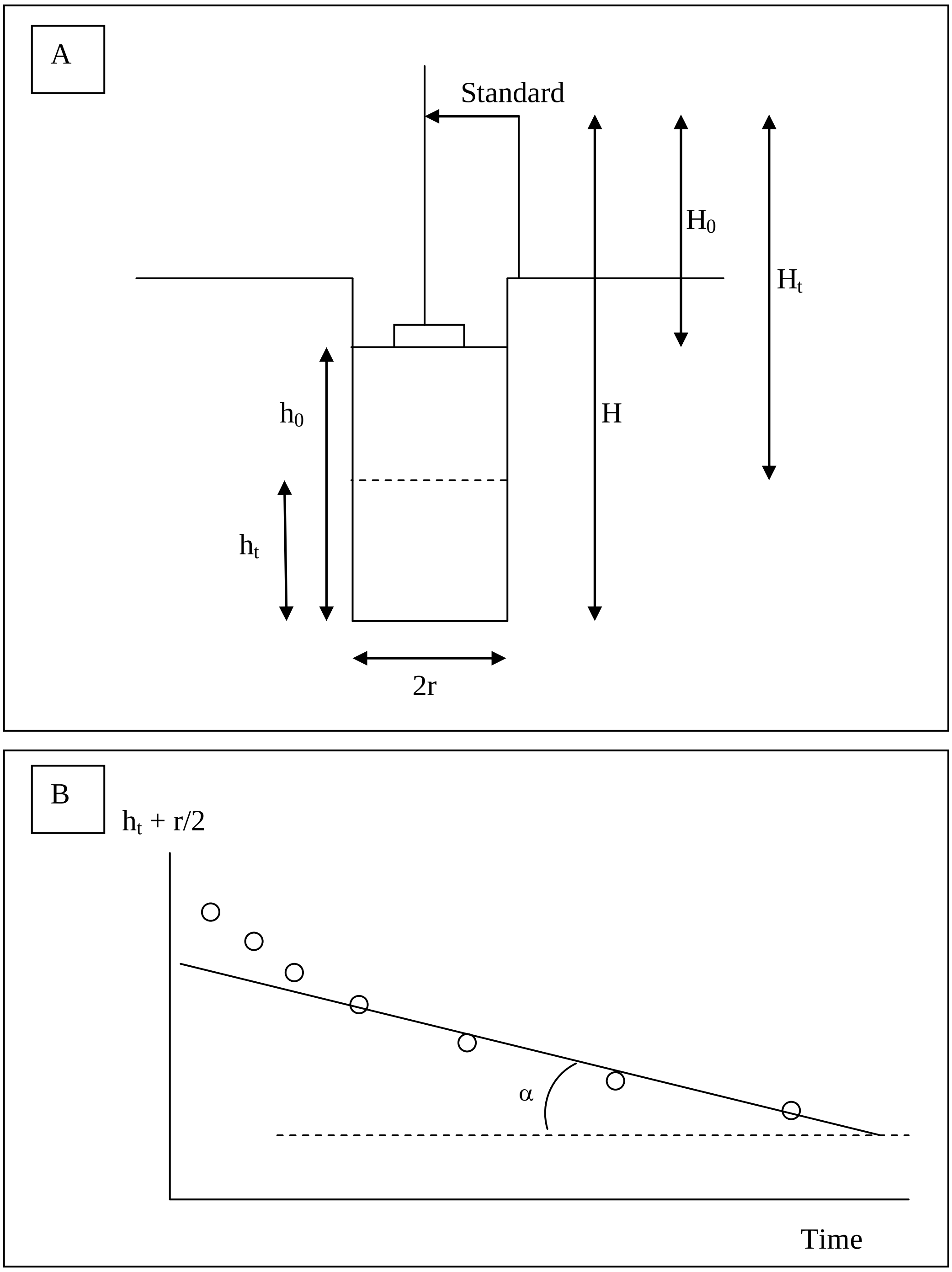


Figure 21. Inversed Auger Hole Method. Field measurement set up (A) and sketch of derived data (B) (after van Hoorn, 2007).

C.3. Moisture Content

Soil water content or more known as soil moisture content is very important role in geoelectrical resistivity survey. The soil surface plays an important as a boundary between the atmosphere and the unsaturated zone. Soil moisture content is of fundamental importance in hydrogeological processes. How much water infiltrates, evapotranspires, and recharges the subsurface depends on the soil moisture content (Black, 1965)

A gravimetric method (Black, 1965) was adopted in measuring moisture content. In this methods, soil need to be weighed before drying stage. Sample and its container were heated in oven at 105^0C for 24 hours to achieve stabilization mass at constant value (without water content). Sample was taken from the oven and left until it becomes cool. Weight reduction as gravimetrical soil moisture content in dry weight basis (θ_d) was calculated using the ratio of the mass of the water divided by the mass of the dry sample.

$$\theta_{\mathrm{d}} = \frac{(\text{weight of wet soil+tare})-(\text{weight of dry soil+tare})}{(\text{weight of dry soil+tare})-(\text{tare})} = \frac{\text{weight of wet soil}}{\text{weight of dry soil}} - 1$$

Equation 26

Finally, the water content in volumetric basis (θ_v) for each soil sample can be calculated using the following equation:

$$\theta_v = \theta_{\mathrm{d}} \mathrm{x}\ \frac{\text{soil bulk density}}{\text{water density}}$$

Equation 27