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Application of Geophysics to Acid Mine Drainage Investigations

Volume I

Literature Review and Theoretical Background

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by

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EXECUTIVE SUMMARY

Surface and borehole geophysical methods are an important tool in the investigation and monitoring of acid mine drainage (AMD) pollution. Geophysical methods can be used in AMD investigations to:

- map sites of disposal of mining waste;
- map the extent of AMD pollution in aquifers;
- monitor velocity of AMD ground-water pollution;
- delineate depth and lateral extent of waste fills;
- identify potential source areas of AMD;
- locate geologic structures such as faults and formation contacts;
- identify gravel channels within finer grained materials;
- define water table in unconsolidated sediments;
- identify subsurface voids;
- map higher permeability areas within aquifers;
- define the extent of weathering, fracturing and faulting;
- define the extent of protective clay layers;
- map soil conductivity; and
- provide complimentary data for correlating borehole and monitoring well data.

Geophysical methods that can directly measure AMD pollution rely on the resistivity or conductivity contrast caused by the increase in total dissolved solids associated with the lowering of pH and the dissolution of sulfide minerals. This increase in soil/rock conductivity is due to a large increase in specific conductance of ground water within the soil/rock pores. The relationship between the ground-water geochemistry and the resistivity of the whole soil/rock mass is complex and non-linear. State-of-the-art application of geophysical methods to investigations of environmental and hydrogeologic problems requires that site-specific data be developed to calibrate general empirical relationships. Geophysical techniques best suited to directly detect subsurface AMD pollution include direct current (D.C.) resistivity and electromagnetic methods. Induced polarization, self potential, mise-à-la-masse methods may also be of value in some hydrogeologic settings. Although other geophysical methods do not directly measure conductivity, they can be of value in AMD investigations in that they provide complementary information to aid in the subsurface interpretation. These methods include:

seismic, gravity, magnetometer, ground penetrating radar, and various borehole methods.

Field investigations to evaluate the application of D.C. resistivity, electromagnetic, self potential and magnetometer methods were undertaken at four closed mine sites with known or suspected AMD pollution. Results of the field investigations are presented in the second volume of this report. The following conclusions were drawn from an analysis of the survey data collected in the field investigations.

- Electromagnetic (EM) surface methods were successful in detecting and mapping acidic ground water in mine waste piles.
- D.C. resistivity methods were successful in developing vertical profiles of acidic mine waste material that correlated well with electromagnetic survey data and well logs.
- Self potential surveys were successful in detecting acidic ground water flow from mine waste ponds.
- Magnetometer surveys were proved useful in distinguishing between an increase in subsurface conductivity due to buried man-made iron or steel objects, and higher specific conductance from acidic ground water or high conductivity of soils, rock or mine waste.

Included in this report is a review of the literature related to the reported and potential application of geophysical methods to AMD investigations. The report also briefly discusses:

- The geochemistry of AMD.
- The relationship between ion concentration and specific conductance.
- Empirical relationships that are available to predict the resistivity of soil/rock.
- Formulas for determining the optimum line spacing for geophysical surveys and the associated probabilities.

RECOMMENDATIONS

The following are recommended after an extensive review of literature on geophysical applications to acid mine drainage (AMD) and mapping of fluids with high specific conductance:

- A combination of several geophysical methods such as EM, resistivity, self-potential and seismic provide a cost-effective method for evaluating potential impacts, defining subsurface structures, identifying potential flow paths for contaminants, and defining the extent of ground-water pollution.
- Geophysical surveys should become a standard procedure for preliminary investigation and monitoring of AMD where subsurface pathways are not readily discernable.
- Geophysical surveys should be used to monitor the migration of AMD plumes between points of sample collection, such as monitoring wells and springs, to confirm that the best sample locations have been selected and to help in understanding seasonal changes in ground-water movement.
- Geophysical instruments are expensive and require technical training to use. Many local and state agencies faced with AMD do not have the economic resources and staff to develop an in-house geophysical program. Ready access to geophysical expertise is desirable. Consideration should be given to developing regional groups of geophysical experts that can provide professional services and training to local and state agencies. These regional groups should reside in organizations that have a strong interest in applied geophysical research. Such organizations might include universities, the U.S. Geological Survey, some state geological surveys, or technical committees composed of geophysical and geologic professionals from government and the private sector..

I. INTRODUCTION

PURPOSE AND SCOPE OF STUDY

Acid mine drainage (AMD) contributes to the pollution of both surface and ground water. Increasing demand on limited water resources necessitates identification and monitoring of sources of contamination such as AMD. Traditional methods used for detection, monitoring and delineation of ground-water contamination rely on monitoring wells. Proper location of monitoring wells is critical for obtaining accurate data from which to interpret the extent and concentration of contaminants. A misplaced well can yield erroneous data on the impact of the contaminants and increase long-term monitoring costs.

AMD is generally associated with an increase in concentration of heavy metals and other ionic species which increase the specific conductance of both surface and ground waters. This increase in specific conductance allows mapping of AMD ground-water contamination using geophysical methods such as resistivity, electromagnetics, and self potential. The conductivity of mine waste and underlying material is a function of the type of soil and rock, the porosity, and the specific conductance of the fluids that fill the pores. The specific conductance of the pore fluids is often the dominant source of the electromagnetic response (McNeill, 1980). Electrical geophysical methods can be used to map inorganic contaminants, identify direction and extent of contaminant flow, estimate concentration gradients, develop time-series measurements of plumes, and provide data for contaminant plume modeling. Although geophysical studies should not completely replace the use of monitoring wells to collect subsurface hydraulic and water-quality data, they can provide a low cost alternative to using only monitoring wells for the detection of AMD and help in selecting locations for monitoring wells.

For this report several tasks were undertaken to evaluate the utility of geophysical techniques in detection and monitoring of AMD contamination from mine wastes. The tasks were conducted in three phases:

- o Review and summary of literature on geophysical methods that may be useful in evaluating migration of the high specific conductance contaminants in ground water.

- Selection of several closed mine waste sites in California as candidates for site-specific geophysical study.
- Evaluation of selected sites using electrical resistivity, self potential, and electromagnetic geophysical surveys to determine the applicability of these techniques in detecting and evaluating AMD contamination from mine waste. Sites with existing ground-water-quality data were of particular interest because the data allow a semi-quantitative correlation between geophysical response and ground water quality.

The geophysical and geochemical criteria identified in the literature review, in conjunction with certain non-technical criteria such as site access, were used to develop system to rank mine sites for site-specific investigation.

Reviewed literature focused on geophysical and geochemical conditions that are known to be associated with AMD and can be detected by surface and borehole geophysical techniques. These conditions include:

- The presence of conductive minerals, such as pyrite, chalcopyrite, pyrrhotite, hematite, etc.
- The presence of conductive dissolved electrolytes in surface and ground water, such as H^+, Fe^{2+}, Fe^{3+}, SO_4^{2-}, Cu^{2+}, and Zn^{2+}.
- Porosity of soils and interconnection of rock fractures.
- The extent to which pores are filled with water, i.e., moisture content.

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II. ACID MINE DRAINAGE (AMD)

GENERAL DISCUSSION OF AMD[1]

When iron sulfide minerals are exposed to air and water by mining activities, they oxidize and can produce sulfuric acid. The low pH water in turn leaches metals from other minerals associated with the sulfide mineralization. The problem of AMD from mining is complicated by time delays associated with the formation of AMD. Time delays in AMD production are often due to eventual consumption of gangue minerals with buffering capacity, usually calcite and other carbonates, as well as to changes in the nature of ore with depth (e.g. increasing sulfide content). Often, mines with no problem during early periods of mining develop AMD several years or decades later.

Conclusions regarding AMD reached in a Mine Waste Study commissioned by the State Water Resources Control Board and conducted by the University of California at Berkeley are (CSWRCB, 1988):

- AMD is produced by naturally occurring minerals that often are present in great quantities in mine wastes.
- Once AMD develops, it tends to get worse. Common iron and sulfur oxidizing bacteria, *Thiobaccillus* and *Ferrobacillus ferrooxidans*, accelerate the process when the pH of the mine waters falls below a critical value. Thus, the process can be difficult to stop.
- AMD may not develop during mining operations but after closure, when mitigation measures are difficult to adopt.

GEOCHEMICAL SETTING FOR AMD

AMD is generated from rock containing sulfide minerals, in particular iron sulfides such as pyrite and marcasite, that are exposed to oxygen by air and water during mining. Stumm and Morgan (1981) list the following reactions as characteristic of the oxidation of pyrite when exposed to air and water:

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[1] *References begin on page 100.*

$$FeS_2(s) + 7/2\ O_2 + H_2O ==> Fe^{2+} + SO_4^{2-} + 2\ H^+ \quad (1)$$

$$Fe^{2+} + 1/4\ O_2 + H^+ ==> Fe^{3+} + 1/2\ H_2O \quad (2)$$

$$Fe^{3+} + 3\ H_2O ==> Fe(OH)_3(s) + 3\ H^+ \quad (3)$$

$$FeS_2(s) + 14\ Fe^{3+} + 8\ H_2O ==> 15\ Fe^{2+} + 2\ SO_4^{2-} + 16\ H^+ \quad (4)$$

AMD generation is accelerated by acidophilic bacteria, *Thiobacillus ferrooxidans* and *Thiobacillus thiooxidans*, which derive energy from the oxidation of sulfide minerals and Fe^{2+}, accelerating reaction (2) by 3 to 6 orders of magnitude (CSWRCB, 1988). The production of acid is greatest at about pH 3.0 due to bacterial activity and diminishes between pH 3.0 and 2.0 due to buffering reactions that precipitate sulfate, ferric sulfate, and ferric hydroxides (Singer and Stumm, 1970; CSWRCB, 1988).

The oxidation of sulfides and resulting generation of acid releases heavy metals and sulfosalts from ore-forming minerals and gangue rock. Heavy metals such as silver (Ag), cadmium (Cd), cobalt (Co), copper (Cu), mercury (Hg), manganese (Mn), molybdenum (Mo), nickel (Ni), lead (Pb), zinc (Zn), arsenic, (As), antimony (Sb), and selenium (Se) are commonly associated with AMD. Table 1 lists typical ranges of AMD contaminant levels from the Appalachian coal region of the United States (Onysko, 1985). Data compiled as part of a study of mine waste in California (CSWRCB, 1988) found the concentration of heavy metals rises with a decrease in pH and, except for copper, zinc, and sulfate, concentration of most other heavy metals and sulfosalt anions rarely exceeds 100 ppm (CSWRCB, 1988).

GEOCHEMICAL AND GEOPHYSICAL RELATIONSHIPS

General Discussion

The ability of geophysical methods to detect and map an AMD plume in the subsurface is directly related to the contrast between background and AMD plume conductivity. The degree of contrast is controlled by the signal noise caused by variability of the hydrogeology, cultural features, and by limitations of the instruments in sensing plumes at depth. Benson and others (1985) evaluated the correlation between geophysical signature of high specific conductance ground-water plumes and the quality of ground water. They evaluated sources of high specific conductance plumes caused by chloride, sodium, and sulfate that leaked into shallow ground

Table 1

TYPICAL ACID MINE DRAINAGE CONTAMINANT LEVELS[2]

Parameter	Range of Concentration			
pH	8.3	-	1.3	
Iron				
Fe^{2+}	0	-	5,000	mg/L
Fe^{3+}	0	-	1,000	mg/L
Total Soluble, as Fe	0	-	6,000	mg/L
Sulfate				
SO_4^{2-}	50	-	10,000	mg/L
HSO_4^-	0	-	20,000	mg/L
Total, as SO_4^{2-}	50	-	20,000	mg/L
Aluminum	0	-	100	mg/L
Manganese	0	-	100	mg/L
Acidity to pH 3, as $CaCO_3$				
Hydrogen Ion	0	-	2,500	mg/L
Bisulfate Ion	0	-	10,000	mg/L
Ferric Ion	0	-	1,000	mg/L
Aluminum Ion	0	-	500	mg/L
Ferrous Ion	0	-	5,000	mg/L
Humic Acids	0	-	500	mg/L
Total	0	-	20,000	mg/l
Total Dissolved Solids	250	-	25,000	mg/L
Suspended Solids	5	-	5,000	mg/L
Ionic Strength, molar	0	-	0.7	
Redox Potential	-200	-	+1,000	mV

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[2] *Table after Onyska, 1985*

water. D.C. resistivity and electromagnetic (EM34-3 and EM31) methods were compared. Specific conductance measured in ground-water monitoring wells within the plume and in background areas were found to have a correlation coefficient ranging from 0.77 to 0.90 at a 95 percent confidence interval with the surface geophysical measurements. Increases in hydrogeologic complexity, the number of cultural features, and depth of plume all contributed to a reduction in correlation. Researchers have found that a plume conductivity of at least twice background is necessary to identify a plume boundary (Greenhouse and Slaine, 1986; Grady and Haeni, 1984; and Slaine and Greenhouse, 1982).

Correlation of electrical geophysical instrument readings and ground water quality can be done either by using the bulk material conductivity with empirical equations, such as Archie's law, or by site specific comparison of geophysical readings with water quality data (Benson and others, 1985; and Grady and Haeni, 1984). The latter provides a more reliable correlation while the former can help predict variations in readings.

Electric current can flow through a geologic material in three ways: ohmic, electrolytic, and dielectric conduction (Telford and others, 1990). Ohmic current occurs in materials such as metals where free electrons can easily be made to flow. Electrolytic current occurs with ions in solution and is comparatively slow. Dielectric conduction occurs in poor conductors or insulators when an external electric field causes electrons to displace slightly, resulting in a temporary polarization of the material.

Electrical resistivity is a measure of the ease with which electrical current flows through a substance. Resistivity (ρ) of a material of cross-sectional area (A), length (L) and resistance (R) between two faces is:

$$\rho = R \cdot A / L$$

where:

$R = V / I$ (Ohm's law)

V = voltage, volts

I = current, amperes

Electrical resistivity is a property of the medium only and is not proportional to the dimensions of the medium. The units of resistivity in the MKS system (meter-kilogram-second) are ohm-meters and in the CGS system (centimeter-gram-second) ohm-centimeters. Conductivity (σ) is the reciprocal of resistivity ($\sigma = 1/\rho$), and is commonly measured in milliSiemens per meter (mS/m) or millimhos per meter (mmhos/m), where 1 Siemen ≐ 1 mho. Historically, the conductivity of water, or specific conductance, is given in units of millimhos per meter (mmho/m), micromhos per centimeter (μmhos/cm). More recently unit of Siemen has replaced the mho. Table 2 is a list of conversion factors for resistivity to conductivity.

Table 2

RESISTIVITY-CONDUCTIVITY CONVERSION FACTORS

1 ohm-meter = 100 ohm-centimeter
1 / ohm-meter = 1 mho / meter
1 / ohm = mho
1 mho = 1 Siemen
1 mho / meter = 1000 milliSiemens / meter
1 mho / centimeter = 1×10^6 microSiemens / centimeter
1 / ohm-meter = 1×10^4 microSiemens / centimeter
1 milliSiemen / meter = 10 microSiemens / centimeter

Conductivity of most soil and rock minerals is generally very low because most are electric insulators. The presence of some minerals such as magnetite, specular hematite, carbon, graphite, pyrite and pyrrhotite can greatly increase the conductivity of rock and soils (Keller and Frischknecht, 1966). Table 3 lists resistivity values for selected soil and rock minerals.

Conductivity of electric current through soil and rock is primarily through the pore water electrolytes. Thus, the volume of pores, the interconnection of pore space and the characteristic of the pore water electrolyte are all important factors in determining the overall conductivity of soil or rock. The following are factors that affect conductivity or resistivity of soils and rocks and are discussed below.

Table 3

RESISTIVITIES OF SELECTED ROCKS AND MINERALS[3]

Rock/mineral	Resistivity ($\Omega \cdot m$) Low	 High	 Average
Arsenopyrite, FeAsS	2.0 x 10^{-5}	1.5 x 10^{+1}	1.9 x 10^{-3}
Bornite, Cu_5FeS_4	1.6 x 10^{-6}	6.0 x 10^{-3}	9.3 x 10^{-5}
Chalcocite, Cu_2S	8.0 x 10^{-5}	1.0 x 10^{-4}	
Chalcopyrite, $CuFeS_2$	3.0 x 10^{-5}	8.3 x 10^{+1}	7.3 x 10^{-4}
Copper-native, Cu	1.2 x 10^{-8}	3.0 x 10^{-7}	
Covellite, CuS	3.0 x 10^{-7}	8.0 x 10^{-7}	
Cuprite, Cu_2O	1.0 x 10^{+1}	5.0 x 10^{+1}	
Enargite, Cu_3AsS_4	2.0 x 10^{-4}	9.0 x 10^{-1}	2.0 x 10^{-2}
Galena, PbS	6.8 x 10^{-6}	5.8 x 10^{-1}	1.9 x 10^{-2}
Hematite, Fe_2O_3	2.1 x 10^{-3}	4.0 x 10^{0}	
Ilmenite, $FeTiO_3$	1.0 x 10^{-3}	4.0 x 10^{0}	
Iron-metal, Fe			1.0 x 10^{-8}
Magnetite, Fe_3O_4	1.5 x 10^{-5}	1.0 x 10^{+4}	5.0 x 10^{-4}
Marcasite, FeS_2	1.0 x 10^{-3}	1.5 x 10^{-1}	
Pyrite, FeS_2	6.0 x 10^{-5}	1.2 x 10^{0}	2.2 x 10^{-1}
Pyrrhotite, Fe_7S_8	2.0 x 10^{-6}	1.6 x 10^{-4}	7.7 x 10^{-5}
Sphalerite, ZnS	2.7 x 10^{-3}	4.0 x 10^{+4}	
Sulfur, pure crystals, S			1.0 x 10^{+16}
Tetrahedrite, $Cu_{10}(Fe,Zn)_2Sb_4S_{13}$	3.0 x 10^{-1}	3.0 x 10^{+4}	1.1 x 10^{+2}
Argillites	1.0 x 10^{+1}	8 x 10^{2}	
Basalt	1.0 x 10^{+1}	1.3 x 10^{+7}(dry)	
Diabase	2.0 x 10^{+1}	5.0 x 10^{+7}	
Gabbro	1.0 x 10^{+3}	1.0 x 10^{+6}	
Gneiss	6.8 x 10^{+4}(wet)	3.0 x 10^{+6}(wet)	
Granite	3.0 x 10^{+2}	1.0 x 10^{+6}	
Limestone	5.0 x 10^{+1}	1.0 x 10^{+7}	
Marble	1.0 x 10^{+2}	2.5 x 10^{+8}(dry)	
Quaternary/Tertiary terrestrial sands	1.5 x 10^{+1}	5.0 x 10^{+1}	
Quaternary/Tertiary marine sands	1.0 x 10^{0}	1.0 x 10^{+1}	

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[3] Table after Keller, 1966 and 1989; and Telford and others 1976 and 1990

Table 3 (continued)

RESISTIVITIES OF SELECTED ROCKS AND MINERALS

Rock/mineral	Low	Resistivity (Ω · m) High	Average
Quartizite	1.0 x 10^{+1}	2.0 x 10^{+8}	
Quartz diorite	2.0 x 10^{+4}	2.0 x 10^{+6}(wet)	
Sandstone	1.0 x 100	6.4 x 10+8	
Shale	2.0 x 10^{+1}	2.0 x 10^{+3}	
Shists, calcareous & mica	2.0 x 10^{+1}	1.0 x 10^{+4}	
Skarn	2.5 x 10^{+2}(wet)	2.5 x 10^{+8}(dry)	
Slates	6.0 x 10^{+2}	4.0 x 10^{+7}	
Tuff	2.0 x 10^{+3}(wet)	1.0 x 10^{+5}(dry)	
Unconsolidated alluvium & sands	1.0 x 10^{+1}	8.0 x 10^{+2}	
Unconsolidated clays	1.0 x 10^{0}	1.0 x 10^{+2}	2.0 x 10^{+1}(wet)

- Porosity, permeability and heterogeneity
- Electrolyte concentration
- Moisture content
- Temperature

Porosity, Permeability and Heterogeneity

Conductivity of clean sands can be expressed by the empirical formula of Archie (Telford and others, 1990):

$$\rho_e = a \cdot \phi^{-m} \cdot S^{-n} \cdot \rho_w$$

where: ρ_e = bulk resistivity of the material
ρ_w = resistivity of pore water

ϕ = porosity, expressed as a fraction of pore volume
S = fraction of pores containing water
a,n,m = constants of regression, with the following ranges,

$$0.5 \leq a \leq 2.5$$
$$1.3 \leq m \leq 2.5$$
$$n \approx 2$$

Mazac and others (1987) note that "m", the coefficient of cementation, is somewhat larger than 2 for cemented and well sorted granular rock and somewhat less than 2 for poorly sorted and poorly cemented granular rock. For unconsolidated, weakly cemented materials with a porosity that ranges from 20 to 70 percent, a value of 1.3 can be assumed for "m" (McNeill, 1980; and Keller, 1989). The "a" coefficient varies from slightly less than 1 in rock with intergranular porosity to slightly more than 1 in rock with only joint porosity and as high as 3.5 in vesicular tuffaceous rock (Carmichael, 1989).

A commonly used term in formation evaluation is the "formation factor" (FF) which is equal to ρ_e/ρ_w (Telford and others, 1990). Lynch (1962) provided an alternate formula for formation factor stating that it is a constant of proportionality between the bulk resistivity of the formation and the resistivity of the pore water. It depends only on the tortuosity and porosity of the rock and is independent of the water in the pore space as expressed in the following formula:

$$\rho_e = FF \cdot \rho$$

where:

$FF = (L_e/L)^2 \cdot (1/\phi)$
$(L_e/L)^2$ = tortuosity
L = actual length of a rock block
L_e = length of tortuous path through pores in block, $L < L_e$
ϕ = porosity

For soils, the minerals of sand and silt grains are generally electrically neutral and act as insulators. Clays act either as insulators or conductors depending on their moisture content and cation exchange capacity (CEC). The CEC is a measure of the number of cations required to neutralize the clay particles and is given in milliequivalents adsorbed per 100 grams of soil.

Clays develop from sheet-like or layered silicates which have a negatively charged surface. Cations such as Ca^{2+}, Mg^{2+}, H^+, K^+, Na^+, NH_4^+ are loosely held to clay particle surfaces and are available for exchange with other more strongly attracted cations, or they may go into solution when water is present. The small particle size of clays, <0.002 millimeters in diameter, provides a very high surface area per unit volume of soil. Thus a large number of cations can be adsorbed per unit volume and it is these adsorbed cations that affect the soil conductivity.

Although clay particles are poor conductors, a small amount of moisture will create a thin layer on the particle where adsorbed ions can be dissociated and become available for ionic conductivity. For shaly sandstones, Lynch (1962) shows that an apparent formation factor FF_a can be estimated using a method proposed by Hill and Milburn (1956). Specifically, FF_a can be related to a formation factor for a fluid with a resistivity of 0.01 ohm-meters ($FF_{0.01}$) by the following formula:

$$FF_a = F_{0.01} \cdot (100 \cdot \rho_w)^{b \cdot \log(100 \cdot \rho w)}$$

where:

$b = -0.135 \cdot (CEC/\phi) - 0.0055$

ϕ = volumetric water content, cm^3 water/ cm^3 soil

ρ_w = resistivity of pore water

Although this formula was derived for formations encountered in oil exploration, it indicates that CEC has an exponential effect on the conductivity of rock. For soils and rocks where the pore water is not saline, the conductivity can be strongly influenced by clay content and should always be considered when interpreting geophysical surveys (McNeill, 1980).

McNeill (1980) proposed the following rules of thumb for determining the change in pore water concentration needed to affect bulk soil conductivity:

$$\sigma_a = 0.25 \cdot \sigma_w$$
$$\sigma_w = (1/6) \cdot TDS$$
$$\sigma_a = (1/25) \cdot TDS$$

where:

σ_a = apparent bulk soil conductivity, mS/m

σ_w = pore water specific conductance, mS/m

TDS = total dissolved solids, ppm

Field application of geoelectric methods requires consideration of the way electrical properties are averaged for sections of soil and rock which are generally heterogeneous. Only borehole geophysical logging provides a method for measuring resistivity of individual layers. Surface geophysical methods average weathering, depositional layering, and structural discontinuities. The geologic heterogeneity may also make the earth material electrically anisotropic, that is, the resistivity may vary with orientation. For example, stratified rocks are generally more conductive along bedding planes than perpendicular to them. Keller and Frischknecht (1966) point out that a distinction should be made between the "geoelectric section" and the "geologic section". Their boundaries may differ because the geoelectric boundary is based on resistivity contrasts which may not follow a boundary defined on geologic principles such as the fossil record.

Electrolyte Concentration

For electrical current to flow through soil or rock, ions must move through the electrolyte solution in the pores. The specific conductance of an electrolyte is proportional to both the total number of ions in solution and their velocity or ionic mobility. The term specific conductance is used to express the electrical conductivity of a body of unit length and unit cross section at a specific temperature. Pure water has a very low specific conductance, a few hundred micromhos per centimeter (μmhos/cm) or microSiemens/cm (μS/cm) at 25° C (Hem, 1989). For natural water an approximately linear relationship can be found between the total dissolved solids and the specific conductance.

An estimate of the relationship between specific conductance and total dissolved solids can be expressed by the formula:

$$S = K \cdot A$$

where:

K = specific conductance, μmhos/cm

S = total dissolved solids, mg/L

A = regression coefficient

Figure 1, taken from Hem (Fig. 10, p. 67, 1989), shows a straight line with a slope of 0.59 visually fitted to data for waters of the Gila River. A linear regression fit to the lower concentration data shows that the slope of the line is steeper. A regression line slope ranging from 0.54 and 0.96

can be expected for natural waters with most slopes lying between 0.55 and 0.75. The higher values are associated with waters that have a high concentration of sulfate.

An alternative method for estimating specific conductance and quality of water is to use the Schlumberger NaCl curves in Figure 2 developed for fluid-conductivity logs (Alger, 1966; Keys and MacCary, 1971; and Keys, 1989). To use this method, convert the ionic concentration, in milligrams per liter, to an equivalent NaCl concentration, then sum to find a total equivalent NaCl solution. On Figure 2, use the diagonal line equal to the NaCl concentration to find the temperature of the formation waters and read the conductivity or resistivity of the solution on the top or bottom horizontal scales, respectively. The ion conversion factors for calculating an equivalent NaCl solution are given in Table 4 (Keys, 1989; and Lynch, 1962).

For an electrolyte to conduct electric current the solute ions must move through the solvent (water) to transfer charge. The effectiveness of an ion in transferring charge depends upon its charge, its size, and the way it interacts with the solvent (Hem, 1989). A measure of the potential for an ion to transfer charge can be found in its ionic mobility. The ionic mobility represents the terminal velocity, in centimeters per second, of an ion in a potential gradient of 1 volt per centimeter (Barrow, 1979). Ionic mobility is a function of temperature and concentration of salts in solution. In a highly-concentrated solution an ion's velocity or ionic mobility is reduced by the motion of other ions. Changes in temperature also affect the viscosity of the solution and the ability of ions to move. The effect of temperature changes on ion mobility is discussed in more detail below. Table 5 lists the ionic mobility for selected ions at low concentrations and standard temperature of 25° C (Vanysek, 1993).

Table 4

NaCl CONVERSION COEFFICIENTS[4]

Ion	Conversion Coefficient
Ca^{2+}	0.95
Mg^{2+}	2.00
K^+	1.00
SO_4^-	0.50
HCO_3^-	0.27
CO_3^{2-}	1.26

[4] *Conversion coefficients from Keys and MacCary, 1971*

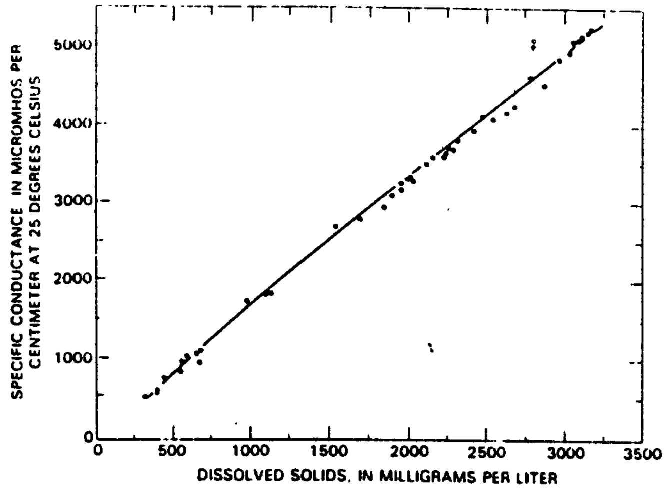

Figure 1. Dissolved solids and specific conductance of composites of daily samples from Gila River, Arizona (from Hem, 1989).

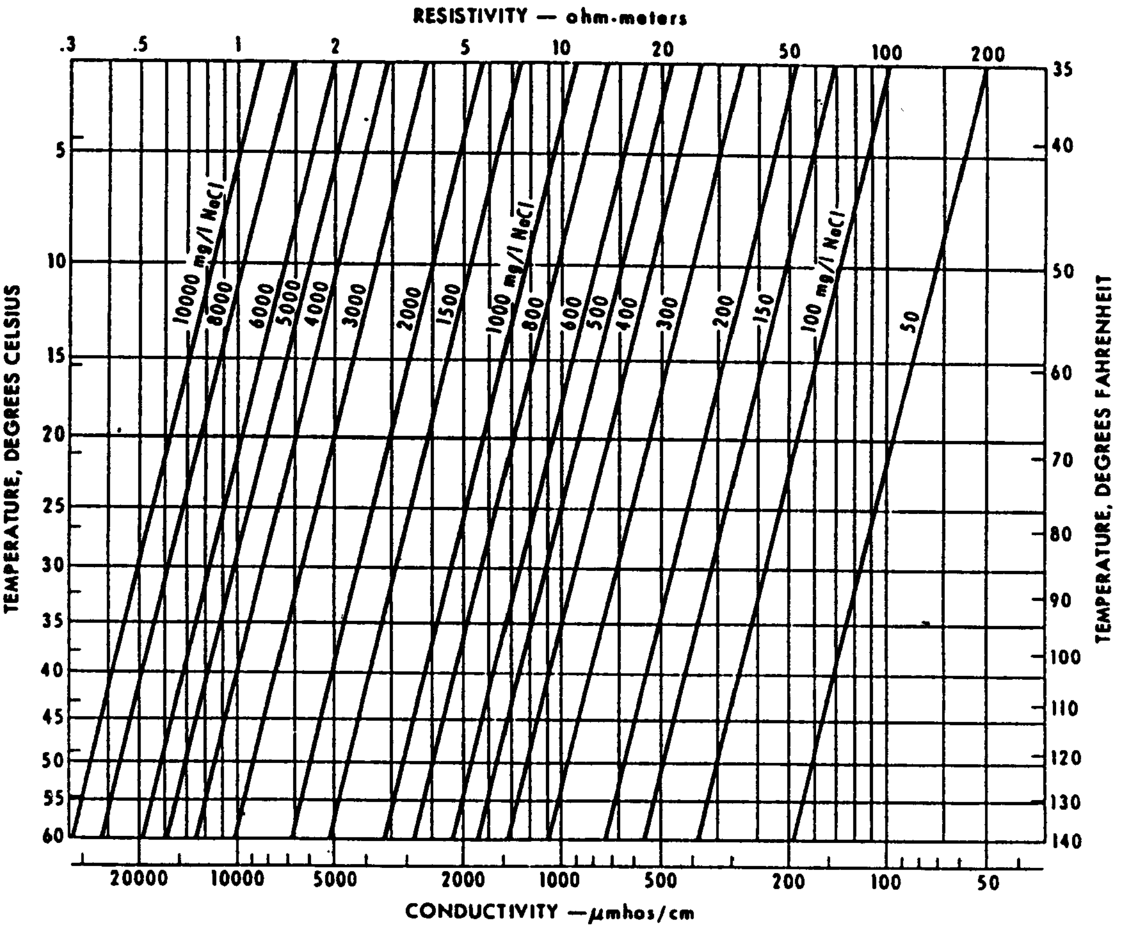

Figure 2. Electrically equivalent concentrations of a sodium chloride solution as a function of resistivity or conductivity and temperature (from Keys and MacCary, 1971).

An approximation of the specific conductance for dilute solutions of natural water can be calculated with the following formula (McNeill, 1980; Barrow, 1979; and Keller and Frischknecht, 1966):

$$\sigma = \mathcal{F} \cdot [C_1M_1 + C_2M_2 + ... + C_nM_n] = 96500 \cdot \Sigma(C_iM_i)$$

where:

σ = specific conductance in mS/m

C_i = gram-equivalent weight of the i^{th} ion per cubic meter of water ($gm \cdot eq \cdot wt \cdot m^{-3}$). A gram-equivalent weight is the formula or molecular weight of an ion divided by its valence.

M_i = mobility of the i^{th} ion in square meters per second per volt ($m^2 \cdot sec^{-1} \cdot V^{-1}$).

$\mathcal{F}$ = Faraday's constant, 96,500 coulombs of charge per gram equivalent weight of salt in solution. 1 coulomb = ampere·sec

For example, one gram of common salt, NaCl, dissolved in 1 cubic meter of pure water (1 mg/L concentration) would have a specific conductance of 0.22 mS/m = 2.2 μS/cm (McNeill, 1980).

$$\sigma = 96500 \cdot \{[(23/58)/23] \cdot 5.2 \times 10^{-8} + [(35/58)/35] \cdot 7.9 \times 10^{-8}\}$$

$$\sigma = 0.22 \text{ mS/m} = 2.2\ \mu\text{S/cm}$$

where: The atomic weight of Na^+ = 23, Cl^- = 35, and the molecular weight of NaCl = 58.

An alternative method for relating specific conductance to solution concentration that accounts for ionic strength is given by the following formula (Laxen, 1977):

$$L = 1000 \cdot \Sigma(\lambda_n \cdot N_n)$$

where:

L = specific conductance, μS/cm

λ_n = equivalent conductance of ion n

N_n = normality of ion n ($millieq \cdot l^{-1} \times 10^{-3}$)

Table 5

IONIC CONDUCTIVITY AND MOBILITY[5]

Ion	Valence	Ionic-λ_n^o Conductivity 10^{-4} m²·mho	Ionic-M_i Mobility $10^{-8}\cdot m\cdot s^{-1}/(V\cdot m^{-1})$	mg/L to Milliequivalents	[x] Coefficient
Ag	1+	61.90	6.42	0.00927	44.6
Al	3+	61.00	6.32	0.11119	217.6
Ba	2+	63.60	6.59	0.01456	119.1
Be	2+	45.00	4.66	0.22190	102.0
Ca	2+	59.47	6.16	0.04990	115.3
Cd	2+	54.00	5.60	0.01779	110.3
Co	2+	55.00	5.70	0.03394	111.2
Cr	3+	67.00	6.94	0.05770	229.2
Cu	2+	53.60	5.55	0.03147	109.9
Fe	2+	54.00	5.60	0.03581	110.3
Fe	3+	68.00	7.05	0.05372	231.7
H	1+	349.65	36.24	0.99216	110.7
Hg_2	2+	68.60	7.11	0.00453	123.7
Hg	2+	63.60	6.59	0.00997	119.1
K	1+	73.48	7.62	0.02558	47.2
Li	1+	38.66	4.00	0.14407	39.2
Mg	2+	53.00	5.49	0.08229	109.2
Mn	2+	53.40	5.53	0.03640	109.8
NH_4	1+	73.50	7.62	0.05544	47.2
N_2H_5	1+	59.00	6.11	0.03025	43.9
Na	1+	50.08	5.19	0.04350	41.8
Ni	2+	50.00	5.18	0.03407	106.6
Pb	2+	71.00	7.36	0.00965	126.0
Ra	2+	66.80	6.92	0.00885	122.1
Rb	1+	77.80	8.06	0.01170	48.2
Sr	2+	59.40	6.16	0.02283	115.3
UO_2	2+	32.00	3.32	0.00741	90.1
Zn	2+	52.80	5.47	0.03059	109.2
$Au(CN)_2$	1-	50.00	5.18	0.00402	41.8
$Au(CN)_4$	1-	36.00	3.73	0.00332	38.6
CN	1-	78.00	8.08	0.03844	48.3
CNO	1-	64.60	6.65	0.02380	45.2
Cl	1-	76.31	7.91	0.02821	47.9
ClO_2	1-	52.00	5.39	0.01483	42.3
ClO_3	1-	64.60	6.69	0.01198	45.2
ClO_4	1-	67.30	6.97	0.01006	45.8
F	1-	55.40	5.74	0.05264	43.1
$Fe(CN)_6$	4-	110.40	11.44	0.01887	527.5
$Fe(CN)_6$	3-	100.90	10.46	0.01415	299.8
H_2AsO_4	1-	34.00	3.52	0.00710	38.1
HCO_3	1-	44.50	4.61	0.01639	40.6
HF_2	1-	75.00	7.77	0.02564	47.6
HS	1-	65.00	6.74	0.03024	45.3
HSO_3	1-	50.00	5.18	0.01234	41.8
H_2SbO_4	1-	31.00	3.21	0.00533	37.4
MnO_4	1-	61.30	6.35	0.00841	44.4
MoO_4	1-	74.50	7.72	0.00625	47.5
$N(CN)_2$	1-	54.50	5.65	0.01514	42.9
NO_3	1-	71.42	7.40	0.01613	46.7
OCN	1-	64.60	6.69	0.02380	45.2
OH	1-	198.00	20.52	0.05880	75.9
PO_4	3-	69.00	7.15	0.03159	233.8
SCN	1-	66.00	6.84	0.01722	45.5
SO_4	2-	80.00	8.29	0.02082	134.2
$Sb(OH)_6$	1-	31.90	3.31	0.00447	37.7
SeCN	1-	64.70	6.71	0.00953	45.2
SeO_4	2-	75.70	7.85	0.01310	130.4

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[5] *Table modified after Vanysek, 1993; and Laxen, 1977. Values given for standard temperature of 25°C.*

The concentration in milligrams per liter can be converted to milliequivalents using conversion factors listed in Table 5. The equivalent conductance for each ion is calculated using the formula (Laxen, 1977):

$$\lambda_n = \lambda_n^0 + [x] \cdot I^{1/2} / (1 + I^{1/2})$$

where:

λ_n^0 = equivalent conductance for ion *n* in an infinite dilute solution, $cm^2/(equivalent \cdot \Omega) = 10^{-4}\ m^2 \cdot mho$

$[x]$ = conversion coefficient for ion *n*

I = ionic strength

Values of λ^0 are listed in Table 5. Values of coefficient [x] incorporates ionic charge and are listed in Table 5. For a detailed discussion of the derivation of coefficient [x] see Laxen (1977).

For a dilute solution, ionic strength can be calculated from the normality using the formula (Laxen, 1977):

$$I = 0.5 \cdot \Sigma (N_n \cdot z_n)$$

where:

N_n = normality of ion *n*

z_n = valence of ion *n*

Using the above formulas, a computed value for the NaCl solution specific conductance of approximately 2.2 μS/cm is obtained. Applying this value to the regression formula given on page 15, a regression coefficient of 0.46 (1/2.2) is found. This coefficient is slightly less than that found for natural waters which usually contain some divalent ions.

Moisture Content

Rhoades and others (1976) evaluated the relationship between soil electrical conductivity (EC), soil water content (θ), soil salinity, and other pertinent soil properties. They developed a two parallel conductor model for bulk soil electrical conductivity (EC_a).. The two conductors are due to a bulk liquid-phase conductivity which depends linearly on the electrical conductivity of the soil water (EC_w), and a bulk surface conductivity (EC_s) from exchangeable ions at the solid-liquid interface. This relationships can be expressed in the formula:

$$EC_a = EC_w \cdot \theta \cdot T + EC_s$$

where: θ = volumetric water content

T = transmission coefficient which accounts for tortuosity of electrical current flow

An empirical formula for T is given by:

$$T = a \cdot \theta + b$$

where: a and b are regression constants taken from a plot of $\{(EC_a - EC_s)/EC_w\}/\theta$ versus θ

Rhoades and others (1976) found the surface conductivity greater for finer-textured soils. When water content was less than a threshold value ($\theta_t = -b/a$), the conductivity of the soil due to the pore water electrolyte was zero. Their experiment in four soils found θ_t to range from 0.05 to 0.12. Rhoades and others (1976) concluded that it may be possible to estimate EC_s, and coefficients a and b based on soil texture and mineralogy. Thus pore water electrical conductivity (EC_w) could be determined by measuring the bulk electrical conductivity (EC_a) and the volumetric water content (θ).

Keller and Frischknecht (1966) discuss the resistivity of rocks and note that pores may be filled with electrolytes (water), gases (air), or non-conductive fluids (oil). They give the following formula for the relationship between the bulk resistivity and the degree of saturation:

$$\rho/\rho_{100} = S_w^{-n_1} \quad ; \quad S_w > S_{wc}$$

where: ρ = bulk resistivity of partially saturated rock

ρ_{100} = bulk resistivity of fully saturated rock

n_1 = experimentally derived parameter ≈ 2

S_w = fraction of pore volume filled with electrolyte

S_{wc} = critical water content

This relationship holds provided the water content is above a critical value (S_{wc}) based on texture. S_{wc} is defined as the point where a continuous film of water still covers the rock surfaces. S_{wc} varies from approximately 25 percent for sandstone to 80 percent for igneous rocks.

When the moisture content drops below the critical saturation, the formula for relating resistivity to degree of saturation is:

$$\rho/\rho_{100} = a \cdot S_w^{-n2} \; ; \; S_w < S_{wc}$$

where: a and n_2 are empirically derived.

Parameter "a" is found to range from 0.05 for sandstones to 0.5 for igneous rocks, and parameter "n_2" from 4 to 5. Sarma and Rao (1962a and 1963b) found a similar effect of a double-gradient resistivity curve for varying moisture content on unconsolidated clean river sands. They noted a trend for increase in the critical saturation with a decrease in grain size. For quartz ground almost to clay size, they found the critical saturation to be at 100 percent. Coefficients for "n_1" at $S_w < S_c$ ranged from 1.0 to 1.9, and "n_2" from 1.7 to 4.0.

The influence of partial saturation of soil and rock leads to the paradox that the conductivity of soils in arid regions tends to be greater than in humid regions. In humid regions the downward leaching of water removes ions from clays thereby lowering conductivity. In contrast in arid regions the upward evaporation of soil moisture tends to cause salt build-up at the surface thereby increasing conductivity (Keller and Frischknecht, 1966).

Temperature

Specific conductance of natural water changes with temperature primarily because of viscosity which affects the ionic mobility (Hem, 1989). Seasonal changes in water temperature or stratification of waters can have a significant effect on specific conductance. Field values of specific conductance are commonly corrected to the 25°C standard temperature. As a general rule, the specific conductance of a dilute solution changes approximately 2 percent for every 1° C change in temperature (Hem, 1989). McNeill (1980) provides the following formula for temperature effects:

$$\sigma(T) = \sigma_{25C} \cdot [1 + \beta \cdot (T - 25°C)]$$

where: β = 2.2×10^{-2} per °C

T = temperature in degrees celsius

III. APPLICATION OF GEOPHYSICAL METHODS TO AMD INVESTIGATIONS

GENERAL DISCUSSION

The following sections present reviews of geophysical methods as they relate to investigation of AMD pollution. Each of the sections discuss the basic geophysical theory, the historic and potential application to AMD investigations, and field procedures. More detailed discussion is provided for those methods that have direct applicability to AMD problems: D.C. resistivity, mise-à-la-masse, induced polarization, electromagnetics, and self potential. Methods that do not have a direct relationship to AMD pollution (seismic, gravity, magnetometer, ground penetrating radar) are discussed briefly and references are provided. The final section on borehole geophysical methods is not intended as a detailed review but as a source of references for more detailed information.

The next section is about optimum sample spacing and the probability of finding an object, given the sample spacing and the size of the object. The formulas are primarily for a two dimensional space, that is, the object to be found is assumed to lie on the plane of the sampling grid. It should be recognized that the sampling space of geophysical methods is really three dimensional and complex. The ability of a geophysical method to detect a buried object is related to the sensitivity of the instrument, the basic physics of the method, the geometry of the target, the depth of the target, and the location of the target with respect to the sampling points. Thus selection of the appropriate geophysical method(s) for an investigation should be based on whether the instrument can measure to the target depth; then the issue of grid spacing should be addressed. Careful consideration of grid spacing and probabilities can provide insight into the density of sampling that is necessary to find a specific size object.

OPTIMUM SAMPLING IN GEOPHYSICAL SURVEYS[6]

Geophysical surveys are generally conducted along a line or grid in order to intersect features of interest. The selection of the proper sampling density is critical to finding the object of interest at minimal cost. The "optimum" distance between survey lines should be such that the geophysical anomaly is defined. However, the concept of "optimum" should also be expressed in terms of the probability of detecting a geophysical anomaly, since "optimum" implies maximization or minimization with respect to a subjective benefit. The "optimum" survey grid would always be the one that has a 100 percent chance of finding the object of interest at the least cost.

The issue of optimum point or line spacing has been generally addressed by Parasnis (1986) and in more detail by Agocs (1955), as well as Kendal and Moran (1963). For geologic structures that are uniform over a long distance, the spacing of geophysical survey lines can be kept large provided the individual observation points along the line are relatively close. As a general rule, geophysical surveys can not be expected to yield information about features whose depth is much smaller than the distance between observations (Parasnis, 1986).

The following formulas can be used to find the probability of intersecting an anomaly (Agocs, 1955):

- For a circular anomaly of diameter D, that is less than the line spacing S, the probability of a line crossing the anomaly would be:

$$P = D / S$$

- For a randomly distributed, elongated anomaly of length L, where the line spacing S is greater than L, the probability of crossing the anomaly would be:

$$P = (2 \cdot L) / (\pi \cdot S)$$

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[6] *References begin on page 100.*

- For a randomly distributed anomaly having a length L, that is greater than the line spacing S, the probability of crossing the anomaly would be:

$$P = [2 \cdot L / (\pi \cdot S)] \cdot \{1 - [(1 - S^2) / L^2]^{1/2}\} + (2 / \pi) \cdot \arccos(S / L)$$

or

$$P = \{[(2 \cdot L) / (\pi \cdot S)] \cdot [1 - \sin(\theta_o)]\} + \{(2 \cdot \theta_o) / \pi\}$$

where: $S / L = \cos(\theta_o)$

θ_o = degrees in radians

- For a randomly distributed rectangular anomaly of length L and width W, whose diagonal is less than the line spacing S, the probability of crossing the anomaly would be:

$$P = 2 \cdot (L + W) / (\pi \cdot S)$$

- For irregular shaped anomaly of perimeter A, the probability of crossing the anomaly would be:

$$P = A / (\pi \cdot S)$$

- For a grid spacing of S and T, where T > S, and where an elongated anomaly of length L, L < S, the probability of crossing the anomaly would be:

$$P = \{2 \cdot L \cdot (S + T) - L^2\} / \{\pi \cdot S \cdot T\}$$

- For a circular anomaly of diameter D where D is smaller than the side of the grid, the probability of crossing a randomly distributed anomaly would be:

$$P = D \cdot \{(S + T) - (\pi \cdot D)\} / \{S \cdot T\}$$

D.C. RESISTIVITY GEOPHYSICAL METHODS[7]

GENERAL DISCUSSION

Basic Principles

Resistivity using direct current is one of the oldest surface geophysical methods. Conrad Schlumberger first applied D.C. resistivity in 1912 (Koefoed, 1979). The D.C. resistivity method investigates changes in subsurface layering geoelectric properties by passing direct current or low frequency alternating current through electrodes placed in the earth and then measuring the drop in potential (voltage) across another set of electrodes.

When resistivity measurements are made in an earth that is a homogeneous, isotropic, semi-infinite half-space, the value measured is the true resistivity of the earth. However, the earth is made of layers of materials that vary laterally and vertically in resistivity. Thus, the values of resistivity measured with resistivity methods are apparent and not true. The measured resistivity is a function of electrode geometry, electrode spacing, the true resistivity and thickness of each layer, layer dip, and lateral variations within layers.

The resistivity of a homogeneous, isotropic material between two electrodes can be found using the formula:

$$\rho = (A / L) \cdot R$$

where:

ρ = electrical resistivity, $\Omega \cdot m$

A = cross-sectional area, m^2

L = length of current flow path, m

R = $\Delta V / I$, Ohm's law

ΔV = difference in voltage potential across two electrodes, volts

I = electrical current, amperes

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[7] *References begin on page 104.*

Geoelectric properties of a layer (*i*) can be described by resistivity (ρ_i) and thickness (h_i). Using these two parameters, other geoelectric properties can be defined (Keller and Frischknecht, 1966; Ward, 1990; and Zohdy and others, 1974).

For a column of homogeneous earth material with a unit cross-sectional area made of layers of differing resistivity and thicknesses, a longitudinal unit conductance (S_i) for layer *i* taken parallel to the layering, and a transverse unit resistance (T_i) taken normal to the layering can be defined. Geoelectric parameters T_i and S_i are also known as "Dar Zarrouk" parameters and are given by the following formulas:

$$S_i = h_i / \rho_i$$
$$T_i = h \cdot \rho_i$$

Electrical properties of a geoelectric layer *i* can also be described by the longitudinal resistivity along bedding planes (ρ_L) and the transverse resistivity perpendicular to bedding planes (ρ_T). These properties are defined by the following formulas:

$$\rho_L = h_i / S_i$$
$$\rho_T = T_i / h_i$$

The total longitudinal conductance (S_T) and transverse resistance (T_T) of a column of geoelectrically different layers can be calculated by summing the unit longitudinal conductances and transverse resistances using the following formulas:

$$S_T = \Sigma(h_i / \rho_i)$$
$$T_T = \Sigma(\rho_i \cdot h_i)$$

For current flowing perpendicular to the layers, the average transverse resistivity (ρ_{Ta}) can be found by dividing the total transverse resistance by the total column thickness using the formula:

$$\rho_{Ta} = \Sigma(\rho_i \cdot h_i) / \Sigma h_i$$

where: $\Sigma(\rho_i \cdot h_i)$ = total transverse resistance of *i* layers

Σh_i = total column thickness of *i* layers

For current flowing parallel to the layers, the average longitudinal resistivity (ρ_{La}) can be found by dividing the total column thickness by the total longitudinal conductance using he using the formula:

$$\rho_{La} = \Sigma(h_i / \rho_i) / \Sigma h_i$$

where: $\Sigma (h_i / \rho_i)$ = total longitudinal conductances of *i* layers

Σh_i = total column thickness of *i* layers

Longitudinal resistivity is always smaller than transverse resistivity with the difference being a measure of the geoelectrical anisotropy (Keller and Frischknecht, 1966). A coefficient of anisotropy can be defined as the square root of the ratio of the total resistivity measured in the two principle directions, along and perpendicular to layering and is given by the formula:

$$\lambda = [\rho_T / \rho_L]^{½} = \{[\Sigma(h_i \cdot \rho_i) \cdot \Sigma(h_i / \rho_i)] / (\Sigma h_i)^2\}^{½}$$

An average or mean square resistivity of an anisotropic block can be given as the square root of the product of the total horizontal and vertical resistivities and is given by the formula:

$$\rho_m = (\rho_T \cdot \rho_L)^{1/2}$$

Equivalence and Suppression

When interpreting multilayer soundings, T and S are sometimes all that can be determined uniquely because various combinations of h_i and ρ_i can give the same or equivalent sounding curve. This is the problem of "equivalence" and is important when interpreting soundings (Maillet, 1947; Parasnis, 1986; Zohdy and others, 1974; and Telford and others, 1990). As an example, a thin resistive layer is characterized by $T = h_i \cdot \rho_i$, whereas a thin conductive layer is characterized by $S = h_i / \rho_i$. Although T and S can be found, h_i and ρ_i cannot. Thus, it is impossible to distinguish between two highly resistive beds of different thickness and resistivity if the product $h_i \cdot \rho_i$ is the same, or between two highly conductive beds if the ratio of h_i / ρ_i is the same (Telford and others, 1990). Also, two or more layers can be combined to provide an equivalent sounding curve. Other methods of geophysical survey or actual subsurface layering information are needed to confine the solution.

Another problem with interpretation of D.C. resistivity surveys is that of suppression. When the thickness of a layer is small compared to its depth, or when the thickness is small compared to the layers above and below, the influence of the layer on the apparent resistivity measured at the surface is small. The presence of the layer is unknown or suppressed, except for layers of extremely high or extremely low resistivity.

Many computer programs that model resistivity sounding curves for a multilayer earth can also produce a correlation matrix that indicates sensitivity of the relationship between h_i and ρ_i. Ward (1990) states that correlation coefficients must be greater than 0.72 to be statistically significant.

Survey Methods

Electrical resistivity surveys can be performed in three general configurations: vertical sounding, horizontal profiling, and a combination of the two. Two other methods, *mise-à-la-masse* and induced polarization, will be discussed separately.

Vertical sounding (VES) is a process by which the separation between the current and potential electrodes is progressively increased over a fixed central point. This causes the array to measure the apparent resistivity at progressively deeper depths. VES is used to investigate vertical changes in subsurface layering and produces a geoelectrical cross section.

Horizontal profiling is a method for measuring the lateral variations in resistivity by progressing the survey along a linear traverse using a constant electrode spacing. A constant spacing implies a constant depth of investigation. Horizontal profiling is primarily used to locate geologic structures such as buried channels, faults, dikes, and anomalous 2-D and 3-D bodies.

The combined method repeats the horizontal profiling method at wider electrode spacings to obtain a series of profiles that can be presented as a geoelectric pseudo-cross-section.

Array Electrode Geometries

The most common D.C. resistivity methods use four electrodes placed in a straight line with various geometries. One pair of electrodes induces current into the subsurface, and the other pair measures the difference in electrical potential or voltage. When two current electrodes are placed into the earth's surface and an external D.C. current (or with more modern instruments a low frequency alternating current) is applied, electrical current flows through the earth from one electrode to the other as shown in Figure 3a for a homogeneous, isotropic earth model. The lines of current flow are perpendicular to the lines of drop in electrical potential, or the equipotential lines. The ideal current lines shown in Figure 3a are marked with the percentage of current carried by each line. The drop in potential between successive equipotential lines is constant.

The electrode configurations commonly used in D.C. resistivity studies (Zohdy and others, 1974) are:

1. Wenner
2. Schlumberger
3. Dipole-Dipole
4. Lee-partitioning

The Wenner array uses four electrodes placed in a straight line and spaced at equal intervals (a) as shown in Figures 3b and 4a. The apparent resistivity (ρ_w) is a function of the distance between electrodes and is given by the formula in Figure 4a.

The Schlumberger array uses four electrodes along a straight line but with an irregular spacing where the distance AB between the current electrodes is equal to or greater than 5 times the distance MN between the potential electrodes as shown in Figure 4b. The apparent resistivity is

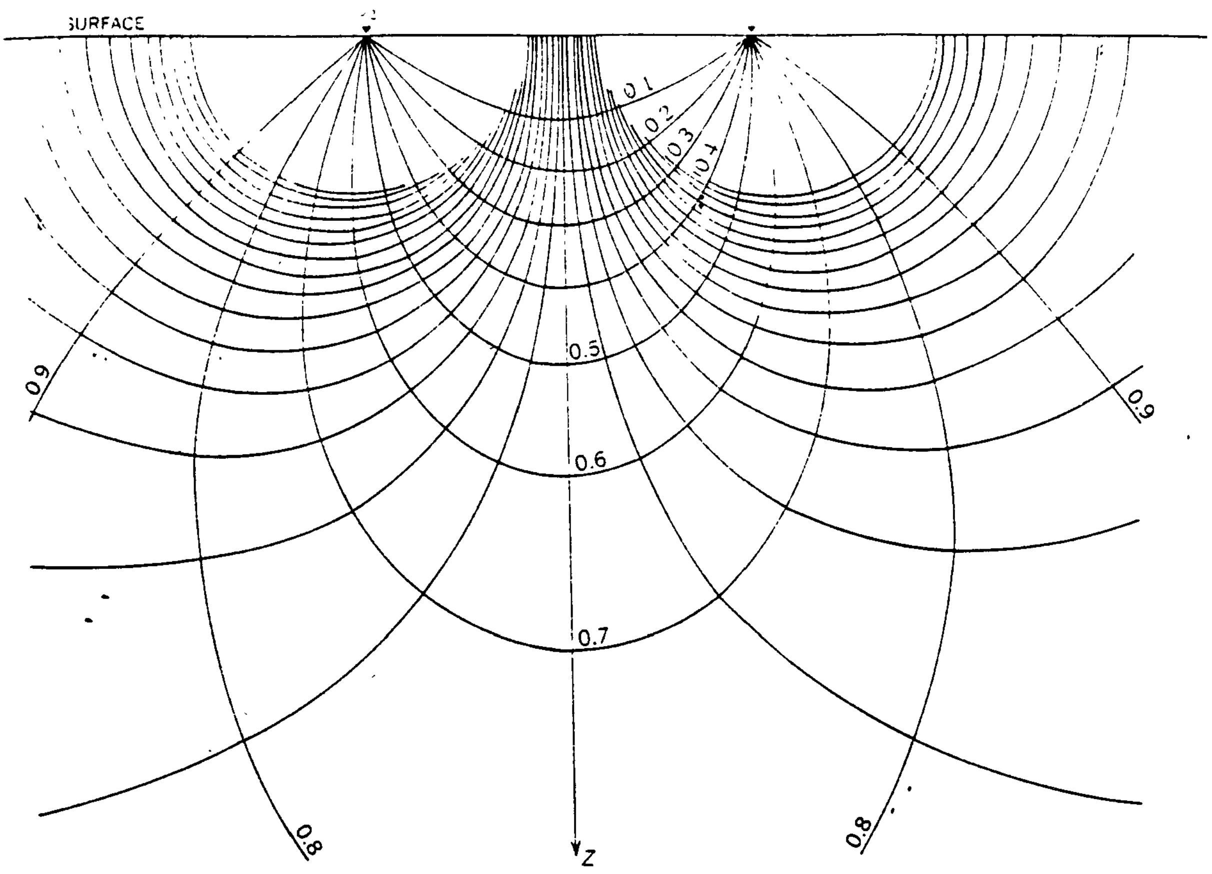


Figure 3a. Potential and current distribution in a vertical plane along the line of electrodes. Current lines of flow each carry one-tenth of the total current. The potential drop between successive equipotential lines is constant (from van Nostrand and Cook, 1966).

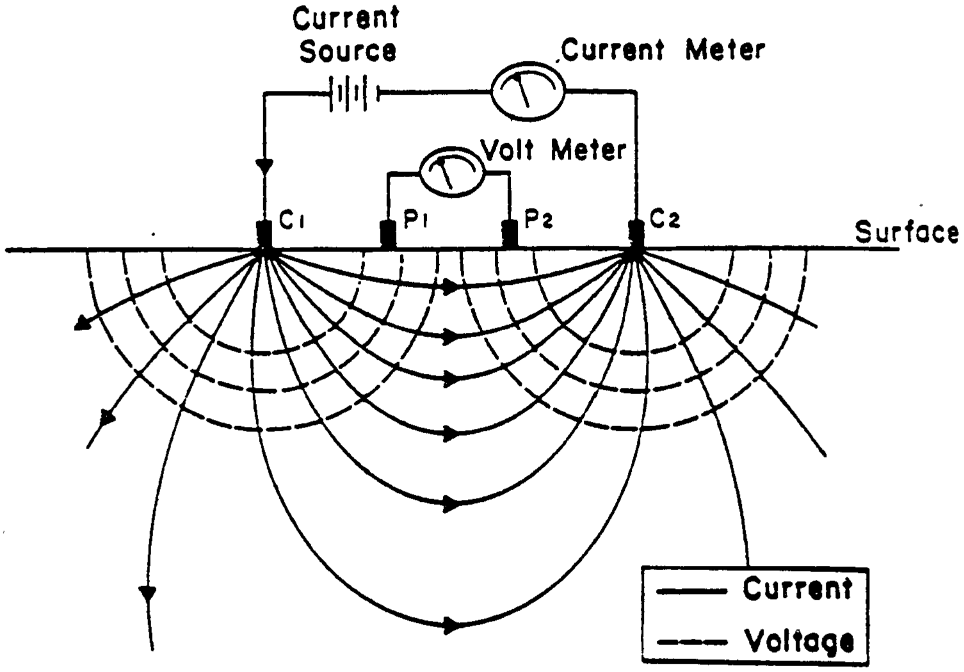


Figure 3b. Wenner array D.C. resistivity equipotential and current flow lines. Current is induces in outer set of electrodes, C_1 and C_2, and voltage drop is measured across inner set of electrodes, P_1 and P_2 (from USEPA, 1986).

a function of the distance AB/2. In practice, MN is kept constant and AB/2 is expanded to obtain deeper soundings. As the distance AB/2 widens, the voltage drop across MN is reduced. At some point, it is necessary to increase the MN distance to obtain deeper soundings. The apparent resistivity (ρ_s) is given by the formula in Figure 4b.

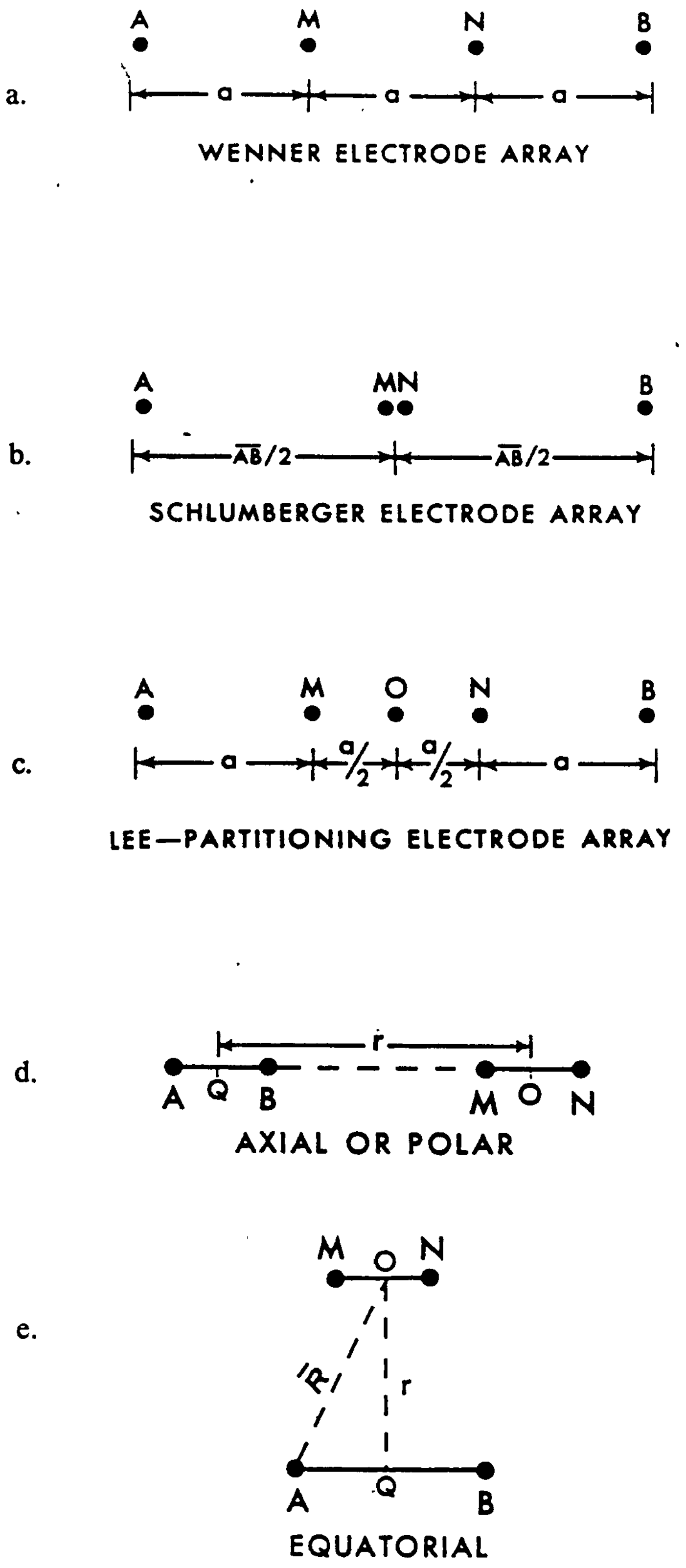
The Lee-partitioning array is similar to the Wenner array except that an additional electrode (O) is placed midway between the M and N potential electrodes. Potential differences are measured between the MO and NO. The formula for calculating apparent resistivity (ρ_l) is given in Figure 4c.

The dipole-dipole array has several configurations but the most common are the axial and equatorial as shown in Figure 4d and e, respectively (Parasnis, 1986; and Zohdy and others, 1974). The dipole-dipole array differs from the others in that the current electrodes are separated from the voltage electrodes. The separation between each electrode within a pair is often kept the same and is significantly smaller than the distance (r) between the current and potential electrodes as shown in Figure 4d. Dipole-dipole has an advantage over Schlumberger, Wenner and Lee-partition because shorter AB and MN spacings are needed for deep penetration. Disadvantages are that the dipole-dipole is harder to interpret and local lateral variations are harder to detect. An alternative to the dipole-dipole method is to enlarge the separation at one or both pairs of electrodes so that they act as bipoles creating a bipole-dipole or bipole-bipole array. Figure 4e shows an equatorial bipole-dipole array and formulas for calculating the apparent resistivity for the axial (ρ_a) and equatorial (ρ_e) arrays.

Depth of Investigation

The Wenner and Schlumberger methods are used most often in environmental and ground-water investigations because they are the easiest to implement and interpret. Interpretation of resistivity measurements assumes that the earth is made of a sequence of geoelectrically distinct layers of finite thicknesses. Contacts between layers are assumed to be horizontal to the array spacing.

When selecting the type of array to use for a survey, it is important to note that the actual depth of investigation of each type of array is governed by a number of factors including signal-to-noise ratio, sensitivity to surficial inhomogeneities, sensitivity to bedrock topography, sensitivity to dip



$$\rho_w = 2\pi a \frac{\Delta V}{I}$$

$$\rho_s = \pi \frac{(\overline{AB}/2)^2 - (\overline{MN}/2)^2}{\overline{MN}} \frac{\Delta V}{I}$$

$$\rho_l = 4\pi a \frac{\Delta V}{I}$$

$$\rho_a = \frac{\pi r^3}{(AB)(MN)} \frac{\Delta V}{I}$$

$$\rho_e = \frac{2\pi r^3}{(AB)(MN)} \frac{\Delta V}{I}$$

Figure 4. Electrode configurations for D.C. resistivity arrays with formulas for interpreting apparent resistivity (after Zohdy and others, 1974).

of layers, and others (Ward, 1990).

There are a number of investigations of the depth to which a surface D.C. resistivity survey can detect a buried target or differentiate layers (Evjen, 1938; Muscat, 1941; Roy and Apparao, 1971; Roy, 1972; Barker, 1989; and Apparao and others, 1992). One of the problems in comparing these studies is that the definition of the depth of investigation or depth of detection is not consistent. The depth of investigation for resistivity methods can be defined as the depth to or below which one-half of the total current penetrates (Evjen, 1938). Evjen found this depth to be approximately equal to half the current electrode spacing for a homogeneous earth. An alternate definition (Roy and Apparao, 1971) of the depth of investigation characteristic is the point at which a thin horizontal layer contributes the maximum amount of the total signal at the surface. Roy and Apparao (1971) and Roy (1972) found for four methods the following values for the depth of investigation based on a distance of L between the extreme electrodes:

Equatorial dipole-dipole	0.125·L
Polar dipole-dipole	0.195·L
Schlumberger	0.125·L
Wenner	0.110·L

Barker (1989) reviewed the issue of the depth of investigation and defined a normalized depth of investigation characteristic curve. Barker (1989) found the depth of investigation to be 0.170·L for the Wenner array, 0.190·L for the Schlumberger, and 0.250·L for the dipole-dipole. Where L is the distance between extreme electrodes.

Apparao and others (1992) defined the depth of detection of an electrode array as the depth below which a target cannot be detected assuming a minimal anomaly of 10 percent. This minimal anomaly is based in part on the findings of van Nostrand (1953) that the limiting depth of detection for an infinitely conducting buried spherical target is approximately equal to the radius of the sphere for the Wenner array. The depth of detection is not easily predicted and is dependent upon the shape, size, and conductivity contrast of the target, and orientation of the sounding array relative to the target.

Ward (1990) presents a detailed discussion of the factors influencing the depth of investigation and sensitivity of each array. In general, the depth of investigation for the Wenner, Schlumberger and Lee-partition methods is predominantly controlled by the distance between current electrodes. Thus equal current electrode spacing produces nearly equal depths of investigation with the Schlumberger array having a slightly greater penetrating depth and resolving power in VES surveys than the Wenner for the same current electrode (AB) spacing. For horizontal profiling they are approximately equal and the dipole-dipole survey is considered less sensitive (Ward, 1990). Barker (1989) notes that for all methods a low resistivity layer at the surface reduces the depth of investigation.

Today a more practical approach for evaluating the depth of investigation is to conduct several forward modeling computer simulations to evaluate the sensitivity of the method in detecting the assumed target(s).

Comparison of Arrays

Each D.C. resistivity method has advantages and disadvantages over the others. Zohdy and others (1974), Pennington (1985), and Ward (1990) provide a detailed discussion of the pros and cons of various resistivity methods. The following is a list of items that can be used to compare the four methods:

- Most D.C. resistivity methods assume the subsurface layers are horizontal and geoelectrically homogeneous. This condition is almost never met, and small lateral variations in near-surface resistivity cause noise and relative error that affects Wenner, Schlumberger and Lee-partition methods more than the dipole-dipole method. However, because the Schlumberger array maintains a constant spacing for the potential electrodes for several current electrode spacings, the error caused by lateral change is smaller and more easily detected than with the Wenner array. As long as the lateral extent of the near-surface inhomogeneity is small compared to the separation of the current electrodes, the relative error produced by the Schlumberger array will be the same for all measurements with the same potential electrode spacing.

- Dipole-dipole arrays are generally applied to deep investigations. The separation of the current and potential electrodes allows for shorter cables than the other methods for a particular depth of investigation. However, the potential difference measured with the dipole-dipole is smaller than with the others and requires more current input and a more sensitive instrument for measuring the voltage potential.

- The potential electrode spacing (MN) for the Schlumberger array is always kept small compared to the current electrode spacing (AB) which should be greater than or equal to at least 5 times the MN spacing. With the Wenner array, the current electrode spacing is always equal to 3 times the potential electrode spacing.

- Schlumberger MN spacing is changed occasionally, as the signal becomes weak, while the AB spacing is changed between each measurement. With the Wenner and Lee-partition arrays MN is always changed along with the AB electrodes for each measurement.

- The AB electrode spacing for the Schlumberger, Wenner and Lee-partition arrays should be at least 3 times and preferably 5 to 10 times the maximum depth of interest.

- The smallest electrode spacing for the Schlumberger, Wenner and Lee-partition arrays should be less than one-half the minimum depth at which a change in material is expected.

- Spacing of electrodes should be logarithmic. That is, data should be taken at roughly equal spacings as plotted along a log scale of AB or AB/2 spacing.

- Dipole-dipole can provide more details on the dip of layering than other methods.

- Stray currents in industrial areas and telluric currents measured with long array spreads affect the Schlumberger array less than Wenner and Lee-partition arrays.

- Drift or unstable potential measurements from electrodes in the ground are less with a Schlumberger array since the potential electrodes tend to stabilize after 5 to 10 minutes.

- Current leakage from cables is less of a problem with the dipole-dipole array. The problem of inductive coupling is minimized.

- Special methods of interpretation are needed for dipole-dipole.

- It is possible to get an indication of the lateral change in the subsurface with the Lee-partition array, but data reduction is not simple.

- The Schlumberger sounding curve is discontinuous whenever the MN spacing is enlarged. The curve is shifted upward or downward to smooth the sounding plot. The discontinuity is an indication of local inhomogeneities near the MN electrodes.

Interpretation of Resistivity Data

Today, VES is interpreted by using a numerical model to find a best-fit match of the data to an ideal geoelectric VES curve (Ward, 1990; Zohdy and others, 1974). Horizontal profiling is generally interpreted qualitatively by plotting the field apparent resistivity values as either a linear trend plotted on an X-Y graph, or by contouring a map of apparent resistivity at a specific electrode spacing. The contour map presents a two-dimensional picture of the apparent resistivity at a constant a-spacing but at an apparent depth.

The combined method of horizontal profiling at different electrode spacings usually presents the data as a contoured geoelectric cross section or pseudo-section (Ward, 1990; Hallof, 1992). The resistivity contours are pseudo-contours since the field apparent resistivity data are arbitrarily assigned to a point below the center of the array at a depth often set equal to one-half the current electrode spacing. Other methods of cross section interpretation use numerical computer models to estimate the distribution, resistivity and thickness of layers (Hohmann, 1982).

Interpretation of D.C. resistivity data can be a complex task when the subsurface is not ideal, that is, not a homogeneous, horizontally-layered and laterally extensive half-space with geometrically-simple-shaped, homogeneous resistivity targets. Prior to the availability of computers, interpretation of VES data was by matching field data to Master Curves for 2, 3, 4 or 5 layer models developed based on various assumptions about the contrast in layer resistivity (Mooney and Wetzel, 1954; Flathe, 1963; Orellana and Mooney, 1966; Orellana and Mooney, 1970; Larzeg, 1973; and Koefoed, 1979). The use of master curves is limited because the infinite combinations of resistivity and layer thickness could not possibly be tabulated. Today computer simulations have replaced curve matching.

The shape of the apparent resistivity curve is influenced by the electrode spacing, the width of the discontinuity, the angle between the profile line and the strike of the discontinuity, and the resistivity of the juxtaposed rock units. There can be abrupt changes in the slope of the apparent resistivity curve where the electrodes of a horizontal profile survey cross a buried structure. A structure such as a fault zone can either increase or decrease conductivity depending on the nature of the surrounding rock and the material that fills the fault zone.

One of the basic assumptions for D.C. resistivity surveys is that they are conducted across a homogeneous, horizontal ground. When variations in slope are greater that 10 degrees or highly irregular, there are noticeable effects that should be corrected (Fox and others, 1980; Telford and others, 1990; and Ward, 1990). Impacts of topography are caused by variations in depth to layering, and moisture content as well as concentration of current flux in valleys and divergence of current beneath hills (Fox and others, 1980; Holcomb and Jiracek, 1984; and Ward, 1990).

Any of the electrode array configurations discussed above can be used for mapping lateral geologic changes, but the profiles of each array differ considerably (Telford and others, 1990). The variation in array profiles is more dramatic with a wide zone of anomalous resistivity, such as a fault zone or dike, than with a contact between two rock types. Telford and others (1990) demonstrate that for zones of anomalous resistivity, the half Schlumberger array best reproduces the shape of the anomalous zone.

The Wenner and Schlumberger resistivity methods are best suited for delineating horizontal layers and vertical contacts, and are less useful for bodies of irregular shape (Ward, 1990; and Telford and others, 1990). These resistivity methods are not particularity sensitive to 3-D anomalies unless the depth of the anomaly is less than the radius of the anomaly. For the Schlumberger array, a maximum response of 12 percent contrast occurs when the depth to the center of a buried sphere is equal to the sphere's diameter. Hallof (1992) indicates that for buried spherical shaped bodies the dipole-dipole array provides the largest anomaly.

Numerous studies have been conducted to determine the response of surface D.C. resistivity to various buried simple-geometric-shaped objects (van Nostrand, 1953; Vozoff, 1958; Unz, 1963; Al-Chalabi, 1969; Zohdy, 1969b and 1970b; Coggon, 1971; Stefanescu and Stefanescu, 1974; Koefoed, 1976a and 1976b; Dey and Morrison, 1979; Speigel and others, 1980; Pridmore and others, 1981; Hohmann, 1982; Mundry, 1984; and Tripp and others, 1984; and Griffiths and Barker, 1993).

Interpretation of resistivity soundings and profiles can now be readily done using forward and/or inverse modeling on personal computers (Koefoed, 1979; Mooney, 1980; Interpex, 1988; Bankey and Anderson, 1989; Orndorff and others, 1989; Zohdy and Bisdorf, 1989; and Bisdorf and Zohdy, 1990). However, the interpretation of data is still limited by the problems of equivalence and suppression as discussed above.

USE OF D.C. RESISTIVITY IN AMD INVESTIGATIONS

General Discussion

Reports on the use of conventional D.C. resistivity for investigation of AMD pollution include: Merkel (1972), Greenfield and Stoyer (1976), Ladwig (1982), Kehew and Groenewold (1983), Dave and others (1986), Ebraheem and others (1990), and Benson (1993). These reports document the application of D.C. resistivity to investigating subsurface distribution of mine wastes, identifying possible high mineralization areas within the waste, identifying areas of ground-water pollution and subsurface discharge, and development of empirical relationships between apparent resistivity and ground-water quality (total dissolved solids and specific conductance). The investigation methods used were VES and horizontal profiling methods with the Wenner and Schlumberger arrays.

D.C. resistivity has been used by geologists for many years in the exploration of mineral, ground-water and petroleum resources. Geophysical textbooks and publications on the general application of D.C. resistivity include: Maillet (1947), Keller and Frischknecht (1966), Kunetz (1966), van Nostrand and Cook (1966), Morley (1967), Soil Test (1968), Bhattacharyya and Patra (1968), Parasnis (1973 and 1986), Koefoed (1979), Telford and others (1976 and 1990), Hallof (1992), and Keller (1993).

Publications that address the general application of surface D.C. resistivity to ground-water, environmental and geotechnical investigations include: Flathe (1955), Breusse (1963), Krulc and Mladenovic (1969), Zohdy and others (1974), Kelly (1976), USEPA (1978), Mooney (1980), Griffiths and King (1981), Bruehl (1983), Dobecki and Romig (1985), Bisdorf (1985), Kelly and others (1988), Ward (1990), Buselli and others (1992), and Henderson (1992).

Although the number of reports on D.C. resistivity investigations of AMD pollution are limited, the literature has numerous D.C. resistivity studies on ground-water pollution. The results of these studies can be readily applied to investigation of AMD since they deal with geologic, geochemical, and hydrogeologic conditions often found at mine sites. Discussion of D.C. resistivity investigations as applied to AMD is divided into the following general categories:

o geologic structure and stratigraphy
o ground-water protection zones
o aquifer properties and ground-water quality
o fracture systems
o landfills and waste impoundments
o open and filled voids
o long-term monitoring

Geologic Structure and Stratigraphy

Studies of the application of D.C. resistivity to investigation of geologic stratigraphy and structures are numerous and include texts and reports by Logn (1954), van Nostrand and Cook (1966), Lennox and Carlson (1967a and 1967b), Zohdy (1970), Telford and others (1976, 1990), Koefoed (1979), Verma (1979), Ayers (1989), Barker (1990); Alfano (1993), and Kelley and Mares (1993). Geologic structures of particular interest in AMD investigations include faults, fracture systems, rapid lateral and vertical changes in stratigraphy, boundaries of buried waste piles, buried alluvial channels, buried tunnels and mine workings, open sink holes or cavities, and filled cavities.

Application of D.C. resistivity to investigation of near vertical discontinuities is discussed in detail by van Nostrand and Cook (1966). Investigation of lateral discontinuities is best done using horizontal profiling, provided the depth of the target is identified. Selection of the proper horizontal array should be done by evaluating existing subsurface data and conducting several preliminary VES surveys.

Direct current resistivity has been used in numerous studies to investigate aquifer properties, the extent of ground-water resources, and mapping of the depth and lateral extent of the fresh-water salt-water interface. These studies include: Hallenback (1953), Breusse (1963), van Dam and Meulenkamp (1967), Flathe (1967 and 1976), Krulc and Mladenovic (1969), Zohdy (1969a), Merkel and Kaminski (1972), Zohdy and Jackson (1973), Frohlich (1973 and 1974), van Dam (1976), Topfer (1976), Worthington and Griffiths (1976), Worthington (1977), van Overmeeren (1981), Urish (1983), Underwood and other (1984), Park and others (1984), Mark and others (1986), Owen and others (1991), and Briz-Kishore (1992).

Ground-Water Protection Zones

Several studies have been conducted to evaluate the use of surface resistivity in delineating ground-water protection zones around production wells (Mazac and others, 1987; Kalinski and others, 1993a). These studies rely on the fact that the apparent resistivity on the surface can be correlated with the thickness of a protective clay or silt layer once sufficient VES surveys have been conducted. Kalinski and others (1993b) discuss the use of D.C. resistivity soundings and profiling to empirically estimate the vertical travel time of ground water through a protective clay layer. Their study found that the vertical travel time can be directly proportional to the square of the longitudinal conductance (S^2) of the protective clay layer when the layer is more conductive than the underlying aquifer, a common occurrence for fresh-water sandy aquifers overlain by a clay layer. Mazac and others (1987) discussed the development of ground-water protection zone using various geophysical methods including D.C. resistivity.

Aquifer Properties and Ground-Water Quality

One of the more interesting applications of D.C. resistivity is its use in estimating physical and chemical properties of aquifers. Studies have demonstrated that empirical relationships can be developed between geoelectrical parameters such as apparent resistivity, transverse resistivity, longitudinal conductance, and apparent formation factor, and between aquifer properties such as hydraulic conductivity, transmissivity, effective porosity, specific capacity, salinity, and total dissolved solids (Page, 1968; Worthington, 1975 and 1976; Griffiths, 1976; Henriet, 1976; Kelly, 1977; Mazac and others, 1979, 1985, 1987, 1990, and 1992; Urish, 1981; Kosinski and Kelly, 1981; Niwas and Singhal, 1981; Kelly and Reiter, 1984; Ponzini and others, 1984; Ringstad and Bugenig, 1984; Taylor and Cherkauer, 1984; Frohlich and Kelly, 1985; Bardossy and others, 1986; Huntley, 1986; Ahmed and others, 1988; Mbonu and others, 1991; Ritzi and Andolsek, 1992; and White (1994).

Typically correlations are developed by conducting VES surveys adjacent to wells where subsurface data are known. Researchers often correlate Dar Zarrouk parameters T and S (transverse resistivity and longitudinal conductance) to aquifer characteristics, since they can be determined more easily and uniquely than layer resistivity (ρ_i) or layer thickness (h_i). Several studies have been conducted in the attempt to remove the effects of changing pore water resistivity by normalizing the apparent transverse resistivity. They employ the use of a correction factor (ρ_{aw}/ρ_w) which is a subjective average water resistivity (ρ_{aw}) divided by the site specific pore water resistivity (ρ_w) (Ponzini and others, 1984; and Ahmed and others, 1988). Geostatistics have also been used to correlate the spatial variability of the data (Bardossy and others, 1986; and Ahmed and others, 1988).

Niwas and Singhal (1981) proposed the combining of Darcy's and Ohm's laws, with Dar Zarrouk parameters to form two new fundamental laws:

$$T = K \cdot \sigma \cdot R \quad \text{and} \quad T = K \;/\; (\sigma \cdot C)$$

where:
T = aquifer transmissivity
K = hydraulic conductivity
σ = aquifer electrical conductivity = $1 / \rho$
ρ = aquifer resistivity
R = transverse resistivity = $h \cdot \rho$
C = longitudinal conductance = h / ρ

Niwas and Singhal propose that if $K \cdot \rho$ is constant for a given area, the hydraulic conductivity can be correlated to aquifer resistivity as measured by surface D.C. resistivity surveys.

In general, these researchers found that for development of empirical correlations a large number of field sites are needed within the area of interest. The data should span the entire range of interest, since extrapolation beyond known data is questionable.

Fracture Systems

Migration of fluids in bedrock fractures and joints is primarily controlled by the orientation, aperture, density and type of in-fill material. Surface D.C. resistivity studies of fracture systems can be used to determine the location, orientation, degree of anisotropy, direction of greatest interconnection, fracture porosity, and transmissivity anisotropy (Merkel and Kaminski, 1972; Habberjam, 1975; Mallik and others, 1983; Smith and Randazzo, 1989; Stewart and Wood, 1990; Ritzi and Andolsek, 1992; Haeni and others, 1993; al Hagrey, 1994; and Lane and other, in press).

Merkel and Kaminski (1972) conducted radial profiles centered on production wells using a modified dipole-dipole array where one of the current electrodes was placed in the well. They found better delineation of the fracture systems when the well electrode was within a fracture system.

Mallik and others (1983) found the axis of the apparent resistivity ellipse developed from radial VES surveys compared favorably with surface-mapped fracture systems. Also, plots of the VES anisotropy (λ) versus electrode spacing indicated variations of fracture density with depth.

Taylor and Fleming (1988) proposed a method for azimuthal resistivity surveying in jointed rock. They used the Wenner array in radial or azimuthal resistivity surveys to determine fracture anisotropy, direction of greatest connectivity, and fracture porosity. They found the major axis of the resistivity ellipse was parallel to the direction of greatest joint interconnection. When the length of the fractures exceeds the array spacing, the major axis resistivity closely parallels the strike of the most prominent set of fractures. When the length of the fractures is less than the array spacing, the major axis of resistivity is oriented in the direction of greatest connectivity for the combined set of fractures.

Ritzi and Andolsek (1992) used the method of Taylor and Fleming to determine the orientation of the anisotropic transmissivity ellipse around a pumping well. The results of the azimuthal resistivity survey compared well with the drawdown ellipse observed during a pumping well test. They concluded that D.C. resistivity can be used to predict hydraulic characteristics of an aquifer and can be used to locate wells.

The basic premise of radial or azimuthal resistivity surveys is that the resistivity ellipse, a polar plot of apparent resistivity by azimuth, should indicate the direction through the fracture system having the least hydraulic resistance. The authors all noted the paradox of resistivity anisotropy described by Keller and Frischknecht (1966): for steeply dipping fractured or layered rock the maximum apparent resistivity is measured along the strike of the fracture and the minimum is measured on the perpendicular.

Lane and others (in press), and Haeni and others (1993) used an azimuthal square-array D.C. resistivity method to map fractures in bedrock. They found that the square-array D.C. resistivity sounding method is more sensitive to rock anisotropy than the Schlumberger or Wenner arrays.

The square-array D.C. resistivity method also requires 65 percent less surface area than the Schlumberger or Wenner arrays (Habberjam and Watkins, 1967; and Habberjam, 1972).

Landfills and Waste Impoundments

Numerous investigation of landfills and waste impoundments using D.C. resistivity methods have been reported in the literature (Cartwright and McComas, 1968; Warner, 1969; Klefstad and others, 1975; Stollar and Roux, 1975; USEPA, 1978; Greenhouse and Harris, 1983; Grady and Haeni, 1984; Ruby, 1984; Stierman, 1984; Rumbaugh and others, 1987; Stierman and Ruedisili, 1988; Robert, 1989; Barker, 1990a; Buselli and others, 1990; Butler and Llopis, 1990; Carpenter and others, 1990 and 1991; Ross and others, 1990; and al Hagrey, 1992).

The general conclusions from these waste impoundment and landfill studies regarding the applicability of D.C. resistivity survey to AMD investigations are: 1) the lateral and vertical extent of landfills can be defined; 2) conductive leachate plumes from landfills and waste impoundments can be mapped provided the geoelectric contrast caused by the pollution is sufficient to overcome natural noise and scatter in formation resistivity; 3) the integrity of landfill cap can sometimes be assessed and fractures identified; 4) the best results are obtained when the ground water is shallow and the geology is relatively homogeneous.

Open and Filled Voids

Direct current resistivity has been used in combination with other geophysical methods such as seismic refraction, microgravity, and ground penetrating radar, to locate subsurface voids that are either filled with air, water or clay (Cook and Nostrand, 1954; Cook and Gray, 1961; Dutta and others, 1970; Bates, 1973; Speigel and others, 1980; Denaham and Smith, 1984; Filler and Kuo, 1989; Smith and Randazzo, 1989; and Nelson and Haigh, 1990). These investigators generally found that relative to surrounding rock, air-filled cavities exhibit very high to infinite resistance and water- or clay-filled , low resistance. Filler and Kuo (1989) found that D.C. resistivity delineated shallow limestone cavities where seismic and ground penetrating radar did not. Nelson and Haigh (1990) found that resistivity surveys over air-filled sinkholes were effective in finding caverns but not useful in locating the associated fracture systems.

Long-Term Monitoring

The use of D.C. resistivity in conjunction with other geophysical methods for long term monitoring of ground-water quality has been discussed by Wilt and Tsang (1985), Benson and others (1988); Barber and others (1991); and Hanson and others, (1993). Long-term monitoring using resistivity requires a reasonable understanding of the subsurface geology and hydrology, knowledge of potential sources and geochemical nature of pollution, and development of background or a baseline resistivity signature. Detection of change is complicated by seasonal variations in soil moisture, ground-water levels, and changes in the direction of ground-water flow. Monitoring highly conductive fluids such as AMD can be successful when the contrast with background or baseline data can be distinguished from random noise and seasonal variations.

D.C. RESISTIVITY FIELD PROCEDURES

Direct current resistivity surveys can be conducted either as vertical soundings (VES) or horizontal profiles. Soundings are conducted by expanding the electrode spacing over a fixed point. Thus VES data are collected at discrete points and provide information on the vertical change in layer resistivity and thickness. Horizontal profiles occur along linear or near-linear trends, or when evaluating anisotropy along radial profiles. The intent of a profile survey is to observe lateral changes or to define the extent of an anomalous buried target. Profile surveys should be oriented to maximize the reading over the anomaly, generally perpendicular to the strike of the target. However, special cases exist, such as azimuthal or radial surveys, where use of various orientations can reveal important information.

Details of field procedures for D.C. resistivity are discussed by Soil Test (1968), Zohdy (1968b), Mooney (1980), Milsom (1989), Telford and others (1990), and Ward (1990). The following is a summary of recommendations for field work::

o For VES surveys, several sites should be selected to develop an understanding of the range of resistivity and thickness. Site selection should be based on prior knowledge of targets, actual subsurface data, estimates of subsurface geology, estimates of geophysical response based on forward modeling, and locations where information would be of maximum benefit. Accuracy of VES interpretations is maximized when some actual subsurface data are available.

- Plot and evaluate VES survey data in the field to make adjustments in survey direction and spacing, or method, and to detect errors in equipment and survey procedures.

- Several VES surveys should be run perpendicular to each other to develop an understanding of the magnitude and impact of anisotropy.

- VES surveys should be conducted initially in areas where horizontal profiling is planned in order to determine the appropriate electrode spacing(s). Mooney (1980) recommends that the profiling array spacing be at least 1½ to 2 times the target depth. Thus for the Wenner array, the overall spread would be 4 to 5 times the target depth.

- To develop correlation between VES data and aquifer properties or ground-water quality, surveys should be run adjacent to the production or monitoring wells. If the well casing is conductive, the survey should be sufficiently distant enough that the well casing does not interfere. Trial surveys may be needed to establish the proper distance.

- Because horizontal profiles are conducted to find lateral changes in the subsurface, surveys should be run over targets. The lateral extent of the survey should be of sufficient distance that the entire array is outside the target's influence. Lateral line spacing and the associated probabilities of finding an target have been discussed above.

- If profiles are repeated to obtain information at various depths, then a reduction in electrode spacing of between 1 to ½ , or an expansion in electrode spacing of 1 to 2 is suggested.

- Data for VES surveys should be collected so that there are at least six data points per decade. One decade is equal to a factor of 10 on the logarithmic scale. To obtain this spacing, each new spacing is found by multiplying the previous value by $10^{1/6} = 1.47$. Increment round-off is done in practice.

- For VES surveys the span of the data should be at least 2 decades with 2½ to 3 decades preferred.

- The small electrode spacings should be at least one-half the minimum depth at which a change in material is expected.

- Resistance at the electrode should be minimized. This can be accomplished by: 1) using non-polarizing electrodes, such as ceramic copper-sulfate pots; 2) placing metal electrodes into moist earth; 3) using two or more electrodes connected in parallel and spaced a meter or two apart perpendicular to the survey line; 4) pouring water or salt water around the electrode.

- Surveys should be 1 to 2 electrode spacings away from sharp changes in topography, such as cliffs or road cuts. Position the survey parallel to the contour of hill or breaks in slope to minimize the impact of topography.

- As the spacing of the survey is widened, there is a tendency for the value of resistance to decrease. However, increases or "backups" can occur when one current electrode is placed in or above material of much higher resistivity, or the dip of the underlying layer is changing rapidly. A new survey should be run at right angles to evaluate the cause of the backup.

- Electrodes should be seated in soils and not driven into gravels or rock. If penetration is not sufficient where originally planned, offset the electrode at a right angle to the survey line.

MISE-À-LA-MASSE GEOPHYSICAL METHODS[8]

The *mise-à-la-masse* method or charged-body potential method is a three point resistivity method. One of the current electrodes (positive pole) is placed into the conductive mass, usually an ore body, and the other electrode is placed sufficiently distant so that its impact is negligible. Two options exist for the configuration of the potential electrodes. One method measures the potential gradient by moving two electrodes placed relatively close together over the area surrounding the buried current electrode. The other method places one potential electrode at a reference point at a distance sufficient to be considered infinite, several hundred meters from the study area. The other potential electrode is then moved about to read the total potential (Telford and others, 1990).

The application of the *mise-à-la-masse* method to subsurface investigations is similar to the self-potential method, discussed below, and is used to obtain a general understanding of the lateral extent of a conductive body. The method is most often used in mineral exploration to determine the extent and interconnection between pyritic ore bodies (Parasnis, 1973). The method works best when the conductivity of the target is 10 to 100 times that of the surrounding country rock (Eloranta, 1984; and Beasley and Ward, 1986). *Mise-à-la-masse* surveys are often done in conjunction with self-potential surveys because they can indicate the dip of the conductive body better than the self-potential method (Bhattacharya and others, 1984).

Papers that document the use of *mise-à-la-masse* in environmental and hydrogeologic studies include: delineation of fracture systems (Jamtlid and others, 1984; and al Hagrey, 1994); estimation of ground-water flow direction and velocity; estimation of dispersion velocity from tracer injection tests and infiltration tests (Mazac and others, 1980, 1987 and 1992; Clasen, 1988; Cahyna, 1990; and Kelly and Mares, 1993).

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[8] *Reference begin on page 118..*

Implementation of the *mise-à-la-masse* method requires that a borehole or well be drilled into the conductive mass or fracture system of interest. Measurements are made radially from the buried electrode. At least two measurement points are needed on each selected azimuth in order to determine the velocity of ground-water or plume. These measurements should be made away from the buried electrode at a distance of at least 1½ times the depth to the aquifer if the well is non-conductive and three times the depth if the well is conductive (Mazac and others, 1992). Maximum depth of exploration for ground water is approximately 10 meters with a radial distance of approximately 150 meters (Jamtlid, 1984; Mazac and others, 1992).

INDUCED POLARIZATION GEOPHYSICAL METHODS[9]

GENERAL DISCUSSION

When direct current induced in the ground through electrodes is turned off, the drop in voltage, as measured at two potential electrodes, is not instantaneous but decays exponentially over several seconds to minutes of relaxation time. This delayed voltage response is due to an induced polarization (IP) of the earth material. The cause of the polarization is not fully understood but is thought to be a combination of electrode polarization and membrane polarization (Sumner, 1976; Ward, 1990).

Electrode polarization is due in part to the presence of a diffuse double layer (DDL) of ions on the surface of clay particles. This DDL is caused by a layer of exchangeable cations on the surface of the negatively charged clay particles, making a double layer known as the Helmholtz double layer (Bohn and others, 1985). In the presence of pore water, the cations are not as tightly held to the surface and diffuse into the aqueous phase. The concentration of the cations decreases exponentially away from the clay surface. The thickness of the DDL is defined as the distance over which the solution concentration is affected by the charge of the clay particles. The thickness (d) of the DDL can be described by the formula (Ward, 1990):

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[9] *References begin on page 119.*

$$d = K_e \cdot \kappa \cdot T / (2 \cdot n \cdot e^2 \cdot \upsilon^2)^{1/2}$$

where:

n = ion concentration of bulk solution

υ = valence of ions

e = unit of electronic charge

K_e = dielectric permittivity of the fluid

κ = Boltzmann's constant, 1.38054×10^{-16} erg/°K

T = Temperature, °K

The thickness of the DDL layer increases with an increase in K_e and T, and decreases with an increase in n and υ.

The movement of electrical current across the DDL can be done in two ways, faradaic and non-faradaic. Faradaic transfer is caused by electrochemical reactions where charge is physically carried across the interface by electron transfer. With non-faradaic transfer, no electrons are transferred but charge is built up across the DDL. Non-faradaic transfer can be described as a simple capacitor whose impedance varies with frequency. Electrode polarization can be represented and predicted by a simple electrical circuit known as the Cole-Cole model of relaxation (Ward, 1990).

A second mechanism for polarizability is termed membrane polarization and is important in rocks and soils with a few percent clay. When an electric current is induced in the earth, membrane polarization occurs within small diameter pores because the cations in the DDL block movement of anions, but allow cations to move easily. This creates an ion-selective membrane (Ward, 1990). The ion concentration gradient that develops opposes the flow of current, reducing ion mobility. This reduction in ion mobility is frequency dependant, the greatest occurring at low frequencies (0.01 Hz). As the frequency increases, the ion-selective membrane effect becomes less important and at approximately 1000 Hz no longer has an affect on the mobility of ions.

Variations in the polarizability of soil and rock can be an important phenomena in environmental and hydrogeologic investigations. Polarizability can be defined as the ratio of the voltage potential at a given time after the current is turned off, to the potential before the current is turned off (Ward, 1990). The value is expressed as a percentage, given the term chargeability, and derived by the formula (Ward, 1990; Kelly and Mares, 1993):

$$\eta_a(t_i) = [\Delta V_{IP}(t_i) / \Delta V_{RA}] \cdot 100\%$$

where: $\eta_a(t_i)$ = apparent polarizability at time t_i
$\Delta V_{IP}(t_i)$ = induced potential at time ti, mV
ΔV_{RA} = primary potential, mV

The rate of discharge of voltage is given by:

$$\alpha_{IP} = V_{IP}(t_1) / V_{IP}(t_2)$$

where: t_1 and t_2 are two different times after the current is turned off

IP is similar to D.C. resistivity in that the ratio of the voltage output to the current input is a measure of the impedance of the earth. IP surveys are conducted with electrode arrays similar to D.C. resistivity surveys and are often done concurrently by using the value of voltage difference at the peak of the current cycle.

IP surveys can be conducted as either vertical soundings (VES-IP), profiling or a combination of the two. Analysis of data is similar to D.C. resistivity. Computer modeling or master curves can be used to determine vertical layering (Seigel, 1959; Dieter and others, 1969; Sadek, 1983; Anderson and Smith, 1986; and Interpex, 1988), pseudo cross-sections can be constructed, and contour maps of polarizability (chargeability) at equal electrode spacing can be used to interpret subsurface features (Ward, 1990).

IP theory and array configurations, mathematical models for buried targets, are discussed by Marshall and Madden (1959), Coggon (1973), Frische and Von Buttler (1957), Hohmann (1975 and 1988), and Fox and others (1980).

IP contrasts with D.C. resistivity in that there are two modes of data collection and interpretation: time domain and frequency domain. In time-domain IP, a square waveform primary current input is turned on and off periodically. Output voltage is measured at various times after the current is turned off, typically between 0.5 seconds and 2 seconds. The decay voltage at a time t_i can be normalized by the primary voltage using the formula given above for chargeability. The resulting units are millivolts/volt. This normalized voltage or chargeability is a fundamental expression of IP polarizability. The decay curve can also be integrated over a time period ($V\Delta t$) and normalized by both V_{RA} and Δt to yield a unit of millivolts-millisecond/volt or milliseconds. A third method of analyzing time domain data is the Newmont method which is a modification of the integration method. The standard Newmont IP cycle is 3 seconds on, 3 seconds off, and 1 second of integration time, and is typically written M331 (Ward, 1990).

The frequency-domain IP method, sometimes called complex resistivity, uses a continuous sine wave of current as input. Output is measured in both real (Re) and imaginary (Im) (quadrature) components. Amplitude (ρ_a) of the apparent resistivity and phase shift (Φ) of the output voltage waveform are given by (Ward, 1990):

$$\rho_a = (Re^2 + Im^2)^{1/2}$$
$$\Phi = \arctan(Im / Re)$$

In frequency-domain IP, the amplitude of the apparent resistivity and the phase shift can be measured over several decades range in frequency (0.03 Hz to 300 Hz). In field applications this frequency dependence of IP polarizability becomes important because it restricts the range of frequencies used to between 0.03 Hz and 3 Hz. The low frequency is restricted by noise caused by interaction of the earth's magnetic field with solar activity, while the high frequency is restricted by electromagnetic coupling of the transmitter and receiver (Sumner, 1976; and Ward, 1990).

Results of frequency-domain IP are apparent resistivity measured at different frequencies. The impact of frequency and polarizability can be quantified by comparing the apparent resistivity at two frequencies. This is known as the percent frequency effect (PFE) and is given by (Ward, 1990):

$$PFE = 100 \cdot (\rho_1 - \rho_2) / \rho_1$$

where:
ρ_1 = apparent resistivity at frequency 1
ρ_2 = apparent resistivity at frequency 2

For normal frequency-domain IP surveys, the PFE is positive, the phase shift lags (negative phase angle), and the secondary decay voltage (output) has the same sign as the primary voltage (input). Negative IP effects are possible and commonly cause a positive phase angle. Negative IP effects can be caused by heterogeneities in the soil, inductive coupling, polarizability of type-K and type-Q layer sequences, and some 2-D and 3-D bodies (Ward, 1990).

Inductive coupling has several causes including capacitive, cultural and electromagnetic (Sumner, 1976). Capacitive coupling is caused by electrical leakage from wire-to-wire, wire-to-ground, or electrode-to-wire. Cultural coupling is caused by artificial, grounded conductors such as pipes, fences, or power lines. Electromagnetic coupling is caused when the IP transmitter and receiver circuits behave like primary and secondary windings of an electrical transformer. That is, the primary circuit induces current into the secondary circuit. Electromagnetic coupling increases with frequency.

USE OF IP IN AMD INVESTIGATIONS

Induced polarization has been used in conjunction with conventional D.C. resistivity studies to explore for ore bodies, evaluate ground-water resources, investigate landfills, define limits of clay layers, identify the fresh-water salt-water interface, and find source areas for saline ground-water contamination (Vacquier and others, 1957; Breusse, 1963; Ogilvi, 1967; Ogilvi and Kuzmina, 1972; Bodmer and others, 1968; Cahyna, 1990, Cahyna and others, 1990; Draskovits and others, 1990; Ward, 1990; Sandberg, 1993; and Kelly and Mares, 1993).

Investigation of AMD using IP methods has not been reported in the literature, but the environmental and hydrogeologic applications reported above suggest that IP methods can be utilized. The following characteristics of IP methods apply to environmental and hydrogeologic investigations:

- Polarizability is related to the particle size distribution. IP decay rate decreases with an increase in the diameter of sand particles. That is, the larger the sand, the shorter the relaxation time.
- Maximum IP effects occur when the clay content ranges from 3 to 10 percent. The lowest polarizability is found in clean quartz sand, and pure clay. In water saturated sandstone and alluvium, IP appears to be caused by clay coatings on the surface of sands and gravels.

- Flocculation and dispersion of clays can effect polarizability. Sodium and potassium causes clays to disperse and swell, closing pores and reducing IP membrane effects. Saturation with calcium causes clays to flocculate, increasing pore interconnection and thereby the polarizability. For flocculated clays, polarizability is at a maximum at a clay content of 5 to 9 percent.

- Polarizability of clay appears to be directly related to the cation exchange capacity. Montmorillonite has a higher polarizability than kaolinite.

- Polarizability of water saturated soils is generally higher than that of for partially saturated soils.

- Specific conductance of pore water can effect the polarizability. The IP decreases with an increase in salt concentration. IP decays more rapidly with an increase in pore water salinity. At a salinity greater than 10 grams per liter, IP is negligible. IP is most strongly affected by cations whereas anions such as Cl^-, SO_4^{2-}, and PO_4^{3-}, cause almost no change. IP decreases with an increase in cation valence. For alkaline soils, the polarizability is greater than neutral soils.

- Generally, polarizability decreases with decrease in resistivity. Thus clay horizons and saline water give smaller anomalies.

- Polarizability is nearly constant at temperatures as high as 40°C. At greater temperatures, IP declines. At temperatures below 0°C polarizability increases in coarse materials, but remains almost constant in clays.

- The signal-to-noise ratio for IP is often better than for D.C. resistivity in shaley sandstone.

Barker (1990) used IP to investigate saline polluted ground water and found it performed better than D.C. resistivity for low levels of pollution in shaley sandstone. Griffiths and others (1981) traced the flow of NaCl pollution from an old foundry spoil pile along fissures caused by collapse of underground mine workings. The low resistivity and high chargeability aligned with the suspected fissures. Chargeability increased when Cl^- concentrations rose from 5 to 500 mg/l, but decreased above 500 mg/l.

Mazac and others (1990) found that there is no relationship between IP and aquifer hydraulic conductivity. Cahyna and others (1990) investigated ground-water contamination associated with a foundry slag pile that contained heavy metals. Their laboratory studies showed the polarizability and resistivity of the slag to be higher than that of the surrounding country rock. However, field studies found that only IP could delineate the slag, and that high IP chargeability was associated with concentrated sources of ground-water pollution.

Although IP methods are most commonly used for mineral exploration, they can be applied to environmental investigations including AMD. Problems with implementation of IP surveys in AMD studies would include: noise caused by heterogeneities of the waste piles, reduction in polarizability due to increase in solution concentration, and the unknown effects of metal hydroxide precipitates on interconnection of pores and polarizability.

ELECTROMAGNETIC METHODS[10]

GENERAL DISCUSSION

Electromagnetic (EM) geophysical surveys involve the use of frequency-domain or time-domain electromagnetic fields to map near-surface geologic features by identifying variations in the conductivity or resistivity of soil and rock. Frequency-domain and time-domain EM surveys can be done on the ground, in fixed-wing aircraft or helicopters (Telford and others, 1990). Frequency-domain EM methods include conventional very low frequency (VLF), VLF resistivity, audio-magnetotelluric, controlled source audio-magnetotelluric, slingram, ground conductivity meters and bore-hole methods. Time-domain EM methods include fixed-loop slingram and central-loop configurations (McNeill, 1990). As with conventional resistivity methods, EM methods can be used to identify and map subsurface geologic features using variable spacing for vertical soundings or fixed spacing for cross-sectional profiles. EM methods can also be divided into two categories based on the nature of the source wave, either planar or loop. The planar wave is considered to propagate across the earth's surface as a uniform, horizontal wave form. The loop wave is generated by a local magnetic dipole. Details on the physics of EM methods are discussed by Telford and others (1976, 1990), Parasnis (1973, 1986), Sharma (1986), Dobrin (1976), and Keller and Frischknecht (1979).

EM methods commonly used in ground water and environmental investigations include VLF, fixed-spacing frequency-domain, ground-conductivity meters, and center-loop time domain. These methods are most often used because of the commercial availability of EM systems, portability, relatively rapid data acquisition, and ability to give relatively good resolution of the subsurface. Table 6 is a summary of EM geophysical survey methods (McNeill, 1990).

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[10] *References begin on page 122.*

Table 6

Summary of Electromagnetic Survey Methods[11]

EM Method	Frequency	Parameter	Principal Application	Advantages	Disadvantages
Conventional VLF	Single	H	Mapping Structures	Very fast, One person operation; Inexpensive; Works well at high resistivity levels	Shallow depth; Depth limited by conductive surface; Provides limited subsurface information; Sensitive to transmitter direction; Areas of little to no signal strength
VLF Resistivity	Single	E, H, ϕ	Soundings	Very fast, One person operation; Inexpensive; Works well at high resistivity levels	Shallow depth; Depth limited by conductive surface; Resolves at most two layer; Areas of little to no signal strength; Static shift
Audio-Magnetotelluric (AMT)	Broadband	E, H	Soundings	Relatively fast; Inexpensive; Under optimum conditions gives good sounding data; Works well at high resistivity levels	Signal strength varies; Sensitive to source orientation; Static shift
Controlled Source Audio-Magnetotelluric (CSAMT)	Broadband	E, H, ϕ	Soundings	Fast; Under optimum conditions gives good sounding data; Works well at high resistivity levels	Expensive; Static shift; Transmitter overprint
Frequency Domain Slingram	Multiple fixed frequency	B	Mapping structure; Soundings	Relatively fast for structural mapping and soundings; Relatively inexpensive; Good measure of bulk conductivity in conductive ground	Limited ability to sound; Zero error; Coil alignment and spacing critical; Surveys difficult in uneven terrain; Poor results in resistive terrain

[11] *Table modified after McNeill, 1990*

Table 6(continued)

Summary of Electromagnetic Survey Methods

EM Method	Frequency	Parameter	Principal Application	Advantages	Disadvantages
Ground Conductivity Meter	Multiple fixed frequency	B	Mapping conductivity and structure; Simple soundings	Fast for structural and conductivity mapping; Relatively fast for soundings; Relatively inexpensive; Good measure of bulk conductivity in conductive ground	Very limited sounding data; Shallow exploration depth (<60m); Non-linear response at high conductivity; Poor results in resistive terrain
Time-Domain EM (TDEM)	Varies current with time	dB/dt	Mapping structure; Soundings	High degree of survey flexibility; Moderately fast; Insensitive to intercoil spacing and alignment; Zero level well known; Good measure of bulk conductivity in conductive ground;	Relatively expensive; Poor results in resistive terrain
			Soundings	Very fast; Good lateral resolution; Good resolution of equivalence; Good measure of bulk conductivity in conductive ground	Relatively expensive; Poor results in resistive terrain

H = Magnetic field strength
E = Electrical field strength
ϕ = Phase difference between E and H

VLF METHODS

Conventional VLF

Two methods of VLF geophysical surveying are employed in environmental studies: conventional VLF and VLF resistivity. Both methods rely on electromagnetic fields being transmitted by a distant VLF transmitter. VLF transmitters are worldwide and transmit at frequencies in the band of 15-25 kHz (kilohertz) for the purpose of military, air and marine communications (Parasnis, 1986; Telford and others, 1990). The VLF transmitter antenna is an oscillating electrical dipole that is effectively a grounded vertical wire several hundred meters long. The electric and magnetic fields radiate as ground, space and ionosphere waves. At distances much greater than the wave length, 20 kHz ≈ 15 km, the field of the VLF dipole can be considered a uniform field within a small area of several kilometers (km) (Telford and others, 1990). These uniform VLF waves consist of three fields, a vertical electrical field, a horizontal magnetic field perpendicular to the direction of the wave propagation, and a small horizontal electrical field oriented with the direction of wave propagation. There is no vertical magnetic field (McNeill, 1990).

Conventional VLF geophysical surveys measure components of the magnetic field. Recent studies by McNeill and Labson (1990) have shown that the VLF anomaly is caused by a subsurface horizontal electrical field that induces electrical charges at the interface between two materials of differing resistivity rather than from induced eddy currents as previously thought. Conventional VLF response is greatest when the buried conductor strikes in the direction of the wave propagation, and the response strength falls off with the cosine of the angle between the conductor strike and the wave direction (McNeill, 1990).

Depth of exploration is limited to 60 to 70 percent of the skin depth, δ, as determined by the following formula:

$$\delta(\mathrm{m}) \approx 500 \cdot (\rho / \upsilon)^{1/2}$$

where: ρ = earth's resistivity, $\Omega{\cdot}\mathrm{m}$

υ = wave frequency, Hz

This depth of exploration can be shallow (McNeill, 1990). For example, for a 25 Ω·m material and a wave frequency of 15 to 25 kHz the skin depth is approximately 10 to 15 m.

Conventional VLF geophysical surveys are performed using the following steps (McNeill, 1990; and Parasnis, 1986):

- The presence of a conductor causes a secondary magnetic field out-of-phase with the primary horizontal magnetic field, creating an elliptically polarized magnetic field.
- The VLF instrument has two solenoids aligned in the same plane, one vertical and one horizontal.
- The vertical solenoid is rotated until the minimum signal is obtained. In this orientation the vertical solenoid is aligned with the minor axis of the secondary ellipse of polarization and is perpendicular to the major axis. The tilt angle, θ, from the vertical is equal to the angle between horizontal and the major axis of the ellipse of polarization.
- The second, horizontal solenoid measures the strength of the major axis field.
- The ratio between the signal strength of the vertical solenoid (minor axis, b) and the horizontal solenoid (major axis, a) is termed the VLF eccentricity.
- The tilt angle measures the in-phase or real component of the secondary vertical magnetic field. While the ratio of b/a is a measure of the quadrature or imaginary component.

Although the field strength is reduced if the conductor is not aligned with the VLF wave, the tilt angle and the b/a ratio are not significantly changed. VLF surveys have several problems that limit their application (McNeill, 1990; and Parasnis, 1986). Conductive overburden and variations in the thickness of overburden can generate significant VLF anomalies. VLF is influenced by topography and readings increase positively going uphill and negatively going downhill (Parasnis, 1986). Adjacent anomalies are superimposed and special filtering is necessary to resolve individual responses. In some areas of the world, VLF signal strength is too weak and a portable transmitter is necessary.

VLF Resistivity

Measurement of the ratio of the amplitudes of the horizontal electric and magnetic VLF fields at the earth's surface can provide an apparent resistivity using the following formula (McNeill, 1990):

$$\rho_a = [1 / (\mu_o \cdot \omega)] \cdot [|E| / |H|]^2$$

where:

ρ_a = apparent resistivity, $\Omega \cdot m$

$|E|$ = amplitude of horizontal electrical field, volt/meters

$|H|$ = amplitude of horizontal magnetic field, ampere/meter

μ_o = $4 \cdot \pi \times 10^{-7}\ \Omega \cdot sec \cdot m^{-1}$ (magnetic permeability of free space)

ω = $2 \cdot \pi \cdot \upsilon$

υ = wave frequency, Hz

In practice, the phase angle by which the horizontal electrical field leads the horizontal magnetic field indicates whether resistivity increases or decreases with depth. For a homogeneous half-space, a phase angle of 45° would be read. A greater angle indicates resistivity decreases with depth, and a lesser angle reflects an increase in resistivity (Brooks and others, 1991). Unfortunately, assumptions need to be made about the upper layering in order to calculate the true resistivity (McNeill, 1990). Thus the method has difficulty resolving more than two layers. A forward modeling computer program is available for VLF interpretation (Grantham and Haeni, 1986).

AUDIO-MAGNETOTELLURIC AND CONTROLLED SOURCE AUDIO-MAGNETOTELLURIC METHODS

Two EM methods that are not usually used for environmental investigations are the audio-magnetotelluric (AMT) and controlled source audio-magnetotelluric (CSAMT). The AMT method relys on naturally occurring electomagnetic fields in a frequency band of 1 Hz to 20 kHz (lightning strikes). The CSAMT method is similar to AMT but uses a transmitter for a source (McNeill, 1990). Both methods measure the ratio between horizontal electrical and horizontal magnetic fields as a function of frequency to estimate an apparent resistivity. The frequency dependent apparent resistivity is modeled using a computer to estimate a simple geoelectric section. Although the AMT method is easy to implement, errors caused by static shift can lead to problems with interpretation. Static shift is caused by electrical charges that are induced at the contact between layers of different resistivity which strongly amplify the local fields. Errors of one order of magnitude in resistivity and thickness are possible. AMT also needs a strong signal and minimal 50/60 Hz power line noise. The CSAMT method can be used to overcome a weak source. A problem arises in the need to separate the transmitter from the site of investigation by a distance of at least three skin depths. For a source operating at 10 Hz in a 1000 Ω·m ground, a minimum separation distance of 15 km is needed. The local geoelectric setting of the transmitter site can also overprint the investigation site readings and introduce significant error.

Syed and others (1985) used CSAMT to locate improperly plugged oil wells that were leaking saline brine into a shallow aquifer. To date, there is insufficient research in using CSAMT at a geologically complex site (Bartel, 1990).

FREQUENCY-DOMAIN SLINGRAM AND GROUND CONDUCTIVITY METER METHODS

The EM method most often used for environmental studies, which is a variation of the slingram method, employs the ground conductivity meter (McNeill, 1990). The slingram method consists of using two magnetic dipole loops connected by a fixed length cable. Loops can be in any orientation but the most common methods require that both the transmitting and receiving loops be aligned in either horizontal (vertical dipole) or vertical (horizontal dipole). The geophysical survey is conducted by moving the loops across the earth at a fixed interval. The coils can be operated at multiple frequencies and separation distances to obtain either vertical soundings or cross-sectional profiles of the conductivity of the earth's layers.

The receiving loop measures the strength and phase shift of the secondary field generated by subsurface conductors. The secondary field induced into a subsurface conductor will generally differ in phase with the transmitted, primary field by a phase difference angle, θ. The vector of the secondary field strength, H_s, can be broken into two components: a component parallel to the primary field, the in-phase or real component ($H_s \cdot \cos\theta$), and a component 90° out-of-phase, the quadrature or imaginary component ($H_s \cdot \sin\theta$) (Parasnis, 1986). To measure the in-phase component, the coil alignment and the intercoil spacing must be carefully maintained. This is often difficult in uneven terrain. The vertical dipole method (horizontal loop) is sensitive to narrow conductive zones or keels, and to conductive overburden whose extent is equal to or greater than the intercoil spacing. The vertical dipole method is also sensitive to steeply dipping, poor conductors such as water filled fractures. The horizontal dipole method (vertical loop) is less sensitive and is often the preferred coil orientation for geologic studies (McNeill, 1990).

Slingram systems are influenced by conductive overburden. The primary EM field decreases by 1/e at the skin depth with a phase change of 1 radian (57.3 degrees) (Parasnis, 1986). For overburden thickness of less than the skin depth, the attenuation and phase change is proportionally smaller. Thus the phase of the field exciting the subsurface conductor is not the same as that of the transmitter. The secondary field from the conductor as measured by the receiving coil is likewise influenced by the overburden. The effect of conductive overburden is to channel the primary field currents into the subsurface conductor where the current density is higher than would be expected. The channeling effect increases with increased frequency as long as the depth of the overburden is much less then the skin depth. At high frequencies, the primary currents in the overburden are at the surface and screen the subsurface, eventually causing the anomaly to disappear.

Ground conductivity meters differ from the conventional slingram system because (McNeill, 1990):

- The low operating frequency, or low induction numbers as defined below, means that the receiver responses are mostly in the quadrature phase which is linearly proportional for low to moderate ground conductivity.
- Ground conductivity meters are one order of magnitude more sensitive than the slingram systems due to operation at low induction numbers, operating frequencies low enough so that the skin depth is always significantly greater than the intercoil spacing.
- The quadrature zero level is set at the factory and remains nearly constant. This results in an accurate and constant measure of the bulk ground conductivity.
- The horizontal and vertical dipole methods give differing depth responses. The horizontal dipole responds to the subsurface within approximately 75 percent of the intercoil spacing, while the vertical dipole responds to the subsurface within 150 percent of the intercoil spacing (McNeill, 1983 and 1990).
- The maximum intercoil spacing is 40 meters, which is less than conventional slingram systems.

Two commonly used EM ground conductivity systems, EM31 and EM34-3, are produced by Geonics Limited. These instruments are operated at fixed intercoil spacings and frequencies and give direct readings of apparent ground conductivity. As with the slingram systems, the receiving coils of these instruments sense the secondary magnetic currents, H_s, and compare them to the primary field, H_p, for specific frequencies so that they operate at low-value induction numbers. The induction number is defined as the ratio of the intercoil spacing to the skin depth (s/δ). When the induction number is much less than unity (McNeill, 1980):

$$H_s / H_p = (i \cdot \omega \cdot \mu_o \cdot \sigma \cdot s^2) / 4$$

where:

H_s = secondary magnetic field at the receiver coil
H_p = primary magnetic field at the receiver coil
ω = $2 \cdot \pi \cdot \upsilon$
υ = frequency, Hz
μ_o = permeability of free space, $\Omega \cdot \mathrm{sec} \cdot \mathrm{m}^{-1}$
σ = ground conductivity, mho/m
s = intercoil spacing, m
i = $(-1)^{1/2}$

Thus the apparent ground conductivity value, in $\mathrm{mho} \cdot \mathrm{m}^{-1}$, given by the instruments is equal to:

$$\sigma_a = 4 / (\omega \cdot \mu_o \cdot s^2 \cdot [H_s / H_p]$$

Figures 5a and 5b show the normalized response curves, $\phi_v(z)$ and $\phi_h(z)$ and cumulative response curves, $R_v(z)$and $R_h(z)$, normalized to a "z" value that is found by dividing the depth to a layer by the intercoil spacing. The $\phi(z)$ function is the relative contribution to the secondary magnetic field from a thin layer at a depth of z. The R(z) function is the relative contribution to the secondary magnetic field from all material below a depth z, or cumulative response curve (McNeill, 1980).

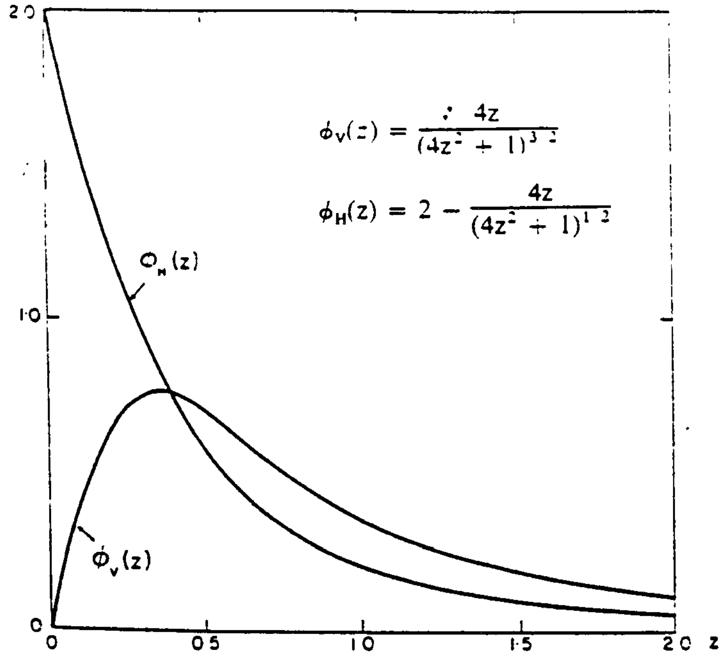

Figure 5a: Normalized depth (z) versus relative instrument response $\phi(z)$ curves for vertical and horizontal dipoles for frequency domain EM (from McNeill, 1980).

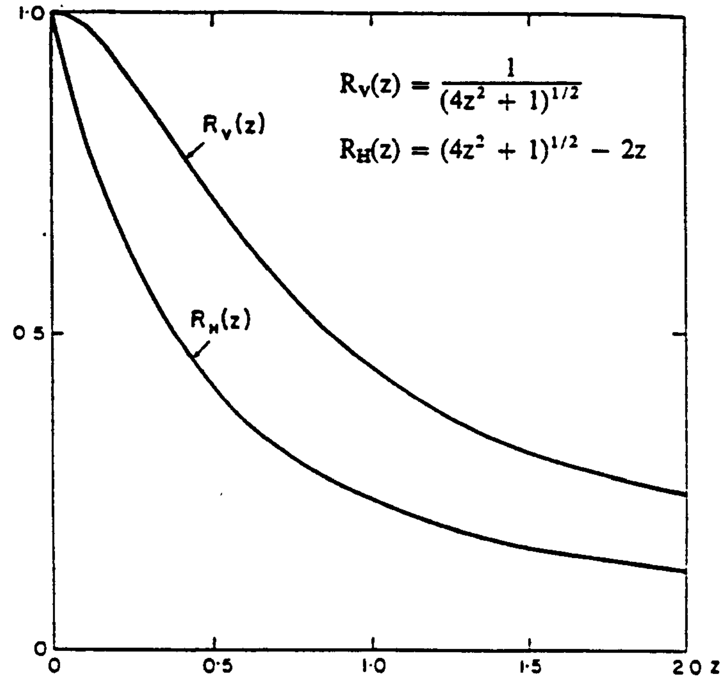

Figure 5b. Normalized depth (z) versus cumulative instrument response R(z) curves for vertical and horizontal dipoles for frequency domain EM (from McNeill, 1980).

TIME-DOMAIN EM METHODS (TDEM)

Another EM method TDEM varies the current in the primary field with time rather than providing a continuous alternating wave (frequency-domain). The transmitting coil is energized with a steady electrical current. A constant magnetic flux is inducted into the subsurface but no secondary currents are created because the flux does not vary with time. When the primary current is abruptly turned-off, the primary flux falls to zero and creates transient secondary currents within subsurface conductors because the flux is varying with time. The time dependent decay of the magnetic field induces transient electromotive forces into the receiving coil (McNeill, 1990; Parasnis, 1986). Environmental studies employ two methods of TDEM: the slingram and the central loop.

The slingram method is similar to the frequency-domain methods. That is, small transmitter and receiver loops are moved along at a fixed separation and a cross-sectional profile of the earth is obtained. Vertical sounding can be made by varying the intercoil separation. The TDEM slingram method has three important differences from the frequency-domain slingram method (McNeill, 1990):

- TDEM is less efficient, therefore the transmitting loop must be large or multi-turn in order to create a large dipole moment. Common dimensions are 5 m square.
- Constraints of the precision of intercoil spacing and alignment are less for TDEM because measurements are made while the transmitter is off.
- Zero level of the receiver is more accurate with TDEM because measurements are made while the transmitter is off.

The second TDEM method uses a large transmitting loop, typically 20 to 150 meters square, with a small receiving loop in the center of the transmitting loop or at the outer edge. Placement of the receiving coil at a distance equal to the side length of the transmitter loop can reduce the magnitude of the inaccuracies caused by induced polarization (IP) effects. When the primary field is turned off, horizontal eddy currents are instantly generated near the transmitting loop that try to maintain the magnetic field at the same strength as before the current was turned off. These eddy currents increase in depth and expand radially with time. The decay of the magnetic field is a measure of the resistivity of the subsurface as a function of depth. TDEM has the following advantages over conventional D.C. resistivity for deep sounding (McNeill,1990):

- TDEM is faster to carry out. The rate of surveys can be increased by laying out several transmitting loops with the transmitter, receiver and receiver loop moved along to collect data.
- Depth of exploration of TDEM sounding is larger than the array dimension unlike D.C. resistivity which must have an array length several times the depth of interest.
- The central loop method is insensitive to overburden conductivity variations and local inhomogeneities.
- The problem of equivalence is less for TDEM than for D.C. resistivity. However, D.C. resistivity is superior to TDEM at resolving the resistivity and thickness of intermediate resistive layers.
- Depth of exploration of the TDEM method is greater than that of the slingram method and can be up to several kilometers.

USE OF EM METHODS IN AMD INVESTIGATIONS

EM methods have been used for the past 20 years for evaluating subsurface geology, subsurface hydrology and water quality. Although EM methods provide information on the subsurface geoelectric layering that is often less precise than D.C. resistivity data, the ease of use and the rapidity of obtaining data make it a preferred method for shallow environmental studies, especially for reconnaissance investigations. In recent years numerous environmental studies have used EM methods. Geonics Ltd. (1992a, 1992b) has published two extensive bibliographies of studies and reports on EM geophysical methods used for:

- agriculture soils mapping;
- archaeology investigations;
- mapping underground voids in limestone;
- locating underground tanks;
- defining depth and extent of permafrost zones;
- predicting the need for cathodic protection;
- landfill siting and monitoring;
- mapping sites of disposal of manufacturing waste, radioactive waste, mining waste, and petrochemical waste;
- mapping higher permeability aquifer channels;
- mapping the extent of contamination in aquifers;

- mapping anisotropy in the subsurface;
- defining the extent of weathering, fracturing and faulting;
- identifying areas of higher ground-water yield;
- mapping the saltwater-fresh water interface;
- defining the extent of protective clay layers;
- exploration for mineral deposits; and
- exploration for petroleum reserves

While the studies listed by Geonics do not all specifically address AMD problems, many of them evaluate geologic and ground-water settings that are similar to those found at mine sites. The two Geonics bibliographies are a valuable resource and can be obtained from Geonics Ltd. in Mississauga, Ontario, Canada, at telephone (905) 670-9580, fax (905) 670-9204.

Published reports of EM investigations of AMD problems are limited and most studies were conducted in coal mine areas rather than the pyritic hard rock mines of the western United States. Greenfield and Stoyer (1976) conducted one of the earliest studies. They ran four profiles in a Pennsylvania strip mine across areas of potentially high AMD ground-water flow using horizontal and vertical dipole EM at 40 and 60 meter spacings. They also ran D.C. resistivity profiles for more detailed layering information. Their study identified a fracture zone that produced a high EM anomaly due to increased saturation by high specific conductance AMD (≈ 330 mmhos/m). They also noted that the same fracture system might produce a low EM anomaly during a drought because of more rapid drainage.

Ladwig (1982) conducted EM studies at three reclaimed coal mines that ranged in surface area from 15 to 37 acres (6 to 15 hectares). Chemical concentrations of the AMD at the three sites ranged from 20 to 4000 mg/L acidity, 18 to 800 mg/L iron and 200 to 1800 mg/L sulfate. EM profiles using 10- and 20-meter intercoil spacing were able to identify areas of buried refuse with high pyrite content, areas of shallow AMD ground water, source areas for surface seeps, a subsurface drain, and delineate the previously unknown extent of the surface mine excavation.

Brooks and others (1991) used a ground conductivity meter (EM34) at a 10-meter intercoil spacing and VLF resistivity (EM16R) over an abandoned Indiana underground and surface coal mine. Specific conductance of the acidic ground water ranged from 3,000 to 18,000 μmhos/cm. Results of the EM surveys were combined with drill hole logs and monitoring well water quality data to infer sources of ground-water contamination, and to delineate the hydrologic connection between waste rock with a high percentage of pyrite (4 to 10 percent) and the surrounding undisturbed materials. A high conductivity anomaly near an adjacent stream was inferred to indicate a connection between the stream alluvium and the underground mine workings.

More recently, studies of geophysics application to AMD problems have been conducted by King and Pesowski (1993, 1991), and King and Sartorelli (1991). These studies found that VLF resistivity and EM ground conductivity meter surveys could delineate conductive plumes leaking from drilling fluid disposal pits, an abandoned uranium mine, and from brine and tailing storage facilities. They also used EM methods to delineate buried channels beneath a proposed tailings storage facility.

To further evaluate the potential application of EM investigations to solving AMD problems, other studies of conductive ground water pollution with EM were reviewed. The studies of interest to AMD investigations can be classified into three general categories: 1) subsurface geologic interpretation; 2) mapping saline brine; and 3) monitoring and mapping leachate from landfills.

The use of EM methods to interpret the subsurface geology relies on a correlation between geologic layering and conductivity contrasts. In natural geologic materials, conductivity contrasts occur due to variations in porosity, permeability, grain size and pore water quality (see discussion of geochemical and geophysical relationships, page 6). At a mine site where AMD pollution is present, the ability to use conductivity contrasts to delineate subsurface geology can be restricted by the random noise caused by heterogeneities of the waste piles, migration of very high specific conductance AMD fluids into different geologic units (which masks the natural differences), a lack of knowledge about man-made subsurface conditions such as the depth to a mine excavation or waste fill, and the lack of knowledge about the location of subsurface openings.

Nevertheless, EM geophysical methods have successfully delineated the general nature of subsurface layering in several studies including: Morgenstern and Syverson (1988), Sanders and Cox (1988), Monier-Williams and others (1990), Mazac and others (1990), Lawrence and Boutwell (1990), and Wrightman and others (1992). The use of EM for investigating saline fluids or fluids with high total dissolved solids is discussed by: Slaine and Greenhouse (1982), Barlow and Ryan (1985), Stewart and Brentnall (1986), Lyverse (1989), Stewart (1990), Sartorelli and others (1990), Goldstein and others (1990), Barker (1990), Street and Engel (1990), Lahti and Hoekstra (1991), Hoekstra and others (1992), and King and Pesowski (1993). The use of EM for investigating and monitoring landfills is discussed by: Slaine and Greenhouse (1982), Greenhouse and Harris (1983), Lasky (1985), Greenhouse and Slaine (1986), Jansen and others (1992), and al Hagrey (1992).

Interpretation of EM data for environmental studies is often done by comparing line plots of EM readings taken at different times or along a traverse, contouring a plan view of data from a surveyed grid, or conducting forward and/or inverse 1-D or 2-D modeling to create geoelectric cross sections. Greenhouse and Slaine (1982, 1986) recommended that EM data be normalized to either an arbitrary constant background value or a value taken at the initial sampling (time = 0), and that the results be expressed in terms of decibels (dB) using the following formula:

$$\sigma_{decibel} = 20 \cdot \log_{10}(\sigma_{x,y} / \sigma_{background})$$

where:

$\sigma_{x,y}$ = apparent ground conductivity at a point x,y

$\sigma_{background}$ = background apparent ground conductivity

Normalization of the conductivity has the following three advantages over the use of actual values:

- o Normalization to a zero background puts all instrument readings on a common format and allows comparisons of data between instruments of differing sensitivity.
- o Logarithmic contour lines do not cluster around contaminant sources to the degree that linear data do.
- o The procedure is objective except for the selection of the constant background value.

By expressing the apparent conductivity in decibels it is often easier to view contrasts and changes over background or initial values. A conductivity value of zero dB indicates no change over background, a value of +6 dB is equal to twice background, and a value of -6 dB is equal to one-half background. Slaine and Greenhouse (1982) suggest that a noise level of approximately 4 dB (or 1.6 times background) should be anticipated. Greenhouse and Slaine (1986) propose the development of a "standard section" on which a 1-D sensitivity model is run to evaluate the impact of variations in layer thickness and layer conductivity. Based on the model results, a normalized background, uncontaminated section can be selected. Greenhouse and Slaine (1986) introduce the concept of a "formation ratio" and a "surface ratio" which are the ratios of the conductivity of individual layers and of the whole section to the background section. They suggest that a formation ratio of at least 2 times background is needed to create a surface ratio of 1.5 times background. Also, a threshold value of as much as 6 times the background is needed to provide reliable evidence of contamination at a single station. Greenhouse and Slaine (1986) noted that contoured data can provide a more reliable indication of contamination over plots of single station data.

Monier-Williams and others (1990) found that topography had a significant influence on EM surveys of landfills due to a reduction in the depths to ground water and an underlying clay layer as the surface elevation decreased. Adopting the decibel normalization method of Greenhouse and Slaine, they suggested an empirical method for correcting the effects of topography. In effect a function, $\sigma(h)$, relates the surface elevation to the apparent conductivity of uncontaminated ground. The function $\sigma(h)$ is then substituted for $\sigma_{background}$ in the Greenhouse and Slaine formula.

An alternative method for removing the effects of topography is to calculate the apparent conductivity at the water table. Using EM data and depth to the water table measured at two wells, Emilsson and Wroblewski (1988) simultaneously solved McNeill's 2-layer case equations to find the conductivity at the water table. For more complex layering, forward computer modeling can be used to predict the impacts of topography and overlying layer thickness. Contouring of the depth, thickness, or the Dar Zarrouk parameters of a geoelectric layer based on forward or inverse computer modeling (Interpex, 1988; and Grantham and others, 1987) can provide a better definition of anomalies because the noise from other layers is removed.

Lawrence and Boutwell (1990) have suggested that when detailed subsurface data are available from drill holes and monitoring wells a multivariate regression analysis can develop a correlation between EM signature and geologic layering. They recommend that a minimum of 5 data points be used to develop such correlations.

Mazac and others (1990) suggest a method for site-specific correlation of hydraulic conductivity and surface geoelectrical measurements for both the saturated and unsaturated zones for porous and fractured rock. The relationship between hydraulic conductivity, K, and resistivity, ρ, is non-linear, but can approach linear for some parameters within a limited range. They propose a method by which observed or interpreted parameters are placed into subclasses, each with a specific weighting factor. The sum of the weights pertinent to all parameters, expressed as a percent, gives an integral indicator value, I. The integral indicator value can be contoured and a linear relationship between saturated K and I can then be found. Once developed for a specific site, the K-I relationship can be used to delineate areas of differing hydraulic conductivity.

EM FIELD PROCEDURES

EM surveys are conducted similarly to other surface geophysical surveys. That is, the assumed target area(s) is (are) identified; the objectives of the survey are defined; based on sensitivity and applicability the proper EM methods are selected; and the survey points and grid are laid out. EM surveys at mines often focus on known or observed sources of AMD seepage rather than the entire site. EM surveys for AMD at mine sites can be complicated by a lack of knowledge about historic operations, by abandoned and buried metal objects, and by the roughness of the terrain. The following are general procedures recommended for EM surveys (McNeill, 1980, 1985b, and 1990; Geonics, 1988). For specifics on instrument calibration and operation the operating manual should be consulted (Geonics, 1985a and 1991).

- Select target area and determine orientation of feature of interest, if possible.
- Optimize sample point and line spacing, intercoil separation based on results of test lines, knowledge of target dimensions and depth.
- Determine whether ground conductivity contrast is sufficient to identify the plume or buried object of interest. Use knowledge of uncontaminated materials and estimates of contaminant concentration and specific conductance for a preliminary model instrument response.
- Target should have an apparent ground conductivity that is at least 150 percent above or below background.
- Lay out grid and identify sample points with non-conductive stakes or markers.
- Calibrate instrument at site as required by manufacturer.
- Orient traverses perpendicular to strike of target, if possible.

- Extend traverse lines into areas of no contamination, if possible. For point monitoring, select stations within and outside of the plume. Select a background site to obtain a normalization value.

- Avoid areas of irregular topography, if possible, because variations from the ideal half-space model can introduce error. Run survey lines along contour of slopes, if necessary.

- To determine the degree of anisotropy, take readings at a station at differing azimuth orientations to find anisotropic ellipse.

- Determine extent of interference from fences, powerlines, and pipes by making a traverse perpendicular any such features until EM readings stabilize.

- EM horizontal dipole (vertical coils) is sensitive to:
  - near surface variation in conductivity;
  - overhead powerlines, keep receiver furthest from powerline;
  - fences, buildings and tanks, experiment to find safe distance; and
  - near surface metal debris.

- EM vertical dipole (horizontal coils) is sensitive to:
  - subsurface variations in conductivity;
  - lithologic changes from faults;
  - pipe lines;
  - vertical dikes; and
  - shallow buried metal debris.

- Data collection using written notes and data loggers can significantly speed up field survey time and data processing.

- Written notes are essential. Data loggers can introduce error when station location or instrument settings are not set properly. Errors are recoverable provided written notes document starting and ending locations and station numbers, line orientations, intercoil spacings, station spacings, field personnel, dipole orientation, and other critical survey information.

SELF POTENTIAL METHODS (SP)[12]

GENERAL DISCUSSION

Self potential (SP) voltages occur in the subsurface due to natural electrochemical, electrokinetic or thermoelectric reactions. These potentials are associated with weathering of sulfide mineral bodies, variations in mineral content along geologic contacts, bioelectric activity, corrosion, and

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[12] *References begin on page 127.*

thermal or pressure gradients in fluids. There are four principal mechanisms that produce self potentials: 1) electrokinetic or streaming; 2) diffusion or liquid junction; 3) Nerst or shale; and 4) mineralization (Telford and others, 1990).

Electrokinetic potential, also called electrofiltration, and streaming or zeta potential, is caused when an electrolyte flows through a capillary or more accurately a porous media. The result is a potential difference between the ends of the passageway that can be expressed by the formula (Parasnis, 1986):

$$E_k = -\{\varepsilon \cdot \rho \cdot \zeta \cdot P\} / \{4 \cdot \pi \cdot \mu\}$$

where: ε = dielectric constant of electrolyte, $F \cdot m^{-1}$
ρ = resistivity of electrolyte, $\Omega \cdot m$
ζ = streaming potential, V
μ = dynamic viscosity, Pa·sec
P = pressure gradient, $Pa \cdot m^{-1}$

Dielectric constant is a measure of the electric polarization that occurs when an electric field is applied and varies inversely with frequency (Telford and other, 1990). The dielectric constant is the ratio of the specific capacity of the material to the specific capacity of a vacuum (8.85×10^{-12} farads per meter; Keller, 1989). Dielectric constant of a rock or soil material changes with water content because water has a high dielectric constant, approximately 80 $F \cdot m^{-1}$ (see Tables 6 and 7 in Keller, 1989; and Table 5.5 in Telford and others, 1990).

Streaming potential can be found associated with flow of water through porous media. The streaming potential is a double layer potential between the solid and solution phase (Telford and others, 1990). Streaming potentials are generally of minor importance in mineral exploration surveys, but can be found associated with large negative anomalies on topographic highs and are important in spontaneous potential borehole logging where drilling fluids flow into porous formations.

The diffusion potential is due to a difference between the concentrations of electrolytes and can be expressed by the formula in milliVolts (mV) (Telford et al., 1990):

$$E_d = -\{R \cdot \theta \cdot (I_a - I_c) \cdot \ln(C_1 / C_2)\} / \{F \cdot \eta \cdot (I_a + I_c)\}$$

where R = the gas constant, $8.314\text{x}10^{-3}$ $\text{kJ}\cdot{}^\circ\text{C}^{-1}\cdot{}^\circ\text{K}$
F = Faraday's constant, 96.445 $\text{kJ}\cdot\text{V}^{-1}\cdot\text{mol}^{-1}$
θ = absolute temperature, °K
η = ion valence
I_a = mobility of anions, $\text{m}\cdot\text{s}^{-1}/\text{V}\cdot\text{m}^{-1}$
I_c = mobility of cations, $\text{m}\cdot\text{s}^{-1}/\text{V}\cdot\text{m}^{-1}$
C_1, C_2 = concentration of electrolytes

The Nerst potential occurs when the electrolyte concentrations around two identical metal electrodes are different. The potential can be expressed in milliVolts (Telford and others, 1990) by:

$$E_n = -\{R \cdot \theta \cdot \ln(C_1 / C_2)\} / \{F \cdot \eta\}$$

Where the coefficients are the same as above.

For example a NaCl solution where $I_a / I_c = 1.49$ produces a diffusion potential at 25 °C of $E_d = -11.6 \cdot \log(C_1 / C_2)$ and a Nerst potential of $E_n = -59.1 \cdot \log(C_1 / C_2)$ (Telford and others, 1990).

The electrochemical self potential is the sum of the diffusion and Nerst potentials and can be expressed in milliVolts as:

$$E_c = -70.7 \cdot [(T + 273)/273) \cdot \log(C_1 / C_2)$$

where: T = temperature, °K

Mineralization potential, or sulfide potential, occurs when two dissimilar metal electrodes are immersed in a homogeneous electrolyte. This potential is commonly found with ores containing pyrite and chalcopyrite. Although the exact mechanism for large negative anomalies over sulfide ores is not fully understood, the use of self-potential surveys to locate ore bodies is a common practice in mineral exploration. The combined effect of the diffusion and Nerst and mineralization potentials is thought to be the principle cause of the large anomalies found with certain mineral ores. Theories on the causes for self potential voltages associated with mineral deposits are discussed by Sato and Mooney (1960), Kilty (1984), Corry (1985), and Furness (1992 and 1993). Theories on causes for self-potential voltages associated with geothermal systems are discussed by Corwin and Hoover (1979) and Sill (1983).

USE OF SP IN AMD INVESTIGATIONS

The use of self potential methods for investigating AMD problems is limited to areas where there is sufficient ground-water flow to develop streaming potential anomalies because the diffusion, Nerst and mineralization potentials are all reduced significantly by high specific conductance electrolytes (Bogoslovsky and Ogilvy 1973, 1972; and Ogilvy, Ayed and Bogoslovsky, 1969). A review of literature found only two published reports on the use of SP for evaluating acidic fluids (Stierman, 1984; and Reznik, 1990). The results of the Stierman study at an acid waste disposal pond were too noisy to provide any definitive results. Reznik's (1990) study at a partially reclaimed coal mine had better results. Reznik was able to identify ground-water flow direction and sources of AMD seepage. Even though the application is limited SP can be useful in AMD investigations because it can: 1) identify zones of increased infiltration; 2) identify areas of leakage or seepage from canals, dams, springs and reservoirs; 3) evaluate effectiveness of ground-water drainage structures; and 4) evaluate effects of ground-water pumping.

Studies of SP phenomena related to infiltration have been reported by Ernstson and Scherer (1986), Bogoslovsky and Ogilvy (1970, 1972, 1973), and Erchul (1986). In general, infiltration of surface water into fractured rock or karst sink holes produces SP minimum when compared to the surrounding lands. This reduction in SP voltage is thought to be linearly related to the pressure gradient as long as the flow is laminar and not turbulent (Ogilvy, Ayed and Bogoslovsky, 1969). Widening of fractures tends to decrease amplitude of the negative anomaly. Infilling of fractures with sand up to approximately 40 percent causes the negative SP anomaly to increase. Infilling with clay decreases the amplitude of the anomaly.

Studies of SP anomalies associated with canal, dam and reservoir leakage have been reported by Ernstson and Scherer (1986), Ogilvy and Bogoslovsky (1979), Bogoslovsky and Ogilvy (1972), and Ogilvy, Ayed and Bogoslovsky (1969). These studies found that a negative SP anomaly is generally associated with areas of seepage with the intensity of the anomaly increasing at higher rates of seepage.

Bogoslovsky and Ogilvy (1973) found that for subsurface drains, positive SP anomaly equipotential lines parallel the discharge into the drain and negative anomalies are often associated with the drain outlet. Bogoslovsky and Ogilvy (1973) used a SP survey to qualitatively evaluate the effectiveness of a drainage system installed to dewater a landslide. The SP survey found an increase in SP values adjacent to a section of the slide that continued to fail, suggesting that the drain was not functioning.

Schiavone and Quarto (1984) found that SP anomalies can be associated with upward flow of fresh water across geologic discontinuities or lithologic boundaries. Asymmetrical anomalies are often associated with vertical boundaries while symmetrical anomalies occur over horizontal boundaries (Schiavone and Quarto, 1984; and Fitterman, 1979). Radial flow to a pumping well can produce a circular positive anomaly and injection of water a negative anomaly (Schiavone and Quarto,1984; and Bogoslovsky and Ogilvy, 1973).

Analysis of SP survey data is generally qualitative rather than quantitative because the SP phenomena involve several different potentials, and because the exact nature of the mineralization potential is still uncertain. Several authors have presented models for SP interpretation of anomalies over vertical boundaries (Fitterman, 1979), vertical dikes (Fitterman, 1983), two-dimensional sheet-like and cylindrical sources (Satyanarayana Murty and Haricharan, 1985), spherical and elliptical ore bodies (Becker and Telford, 1965; and Telford and others, 1990), a thermal point source with vertical contact and overburden (Corwin and Hoover, 1979; and Sill, 1983), current dipoles in half-space (Kilty, 1984), and surface and vertical profiles over vertical and dipping sheet-like bodies (Furness, 1992).

Telford and others (1990) report the depth of an ore body is approximately one-half of the width of the negative anomaly measured at half the amplitude, as a rule of thumb. Based on a limited survey of SP studies, this rule of thumb yields depths that are ±100 percent of the true depth. Data from SP surveys are generally used as qualitative information along with other more quantitative methods, such as D.C. resistivity and IP. Data are presented either as a plan map with contours of SP voltages or as x-y plots of SP voltage versus distance along a traverse.

SP FIELD PROCEDURES

Field procedures for SP surveys use two electrode configurations: 1) two electrodes separated by a constant distance are moved along a survey line, using a voltmeter to measure the gradient between them, or 2) one electrode is placed as a base point and the other electrode is moved from station to station using a long wire. The second method has an arbitrary referenced "absolute" SP as a base point. The base point electrode is connected to the negative terminal of the voltmeter. When the base point needs to be moved, the potential of the new base point is measured relative to the previous base point so that data can be corrected to a fixed reference voltage.

Detailed description of field procedures, precautions and instrument specifications are given by Corwin (1990), Corwin and Hoover (1979), and Corry (1985). A summary of field procedures and sources of noise is given below.

- Electrodes for the SP survey should be non-polarizing. Standard SP electrodes are unglazed, porous porcelain pots with a metal terminal that is submerged in a salt solution of similar composition, such as copper terminals in copper sulfate solution. The permeable pot slowly leaks solution into the soil to make a good electrical contact with the ground.

- Although porcelain electrodes are considered "non-polarizing" they do respond to such factors as temperature, soil moisture, and soil chemistry. For example, for a copper-sulfate electrode, changes in moisture content of 1 percent can cause a +0.3 to +1.0 mV variation. Thus observation of soil moisture conditions can be critical when SP anomalies are a few tens of milliVolts (Corwin, 1990).

- Electrodes should be buried at least 6 inches for detailed surveys where the SP anomaly has a low amplitude, such as over geothermal systems. Random noise of sufficient amplitude to mask the anomaly is caused by moisture differences.

- For detailed low amplitude SP anomaly surveys, the ground surrounding the electrodes should not be wetted and the contact resistance should be less than 50 k·Ω. Voltage should be read before contact resistance because the ohm-meter will temporarily polarize the electrode.

- Temperature variations between the electrodes should be kept to a minimum to reduce voltage drift which can range from +0.5 to +1.0 mV per degree Centigrade.
- A high impedance ($\geq 1 \times 10^7$ ohm-meters) digital voltage meter should be used for field readings. The negative terminal of the voltmeter should be connected to the base station.
- Base station electrode should be placed outside an area of steep SP gradients for maximum anomaly amplitude.
- Base station electrode should be placed outside a chemically reduced environment such as a bog or marsh for maximum anomaly amplitude.
- Power stations, pipelines with cathodic protection, culverts, grounded fences, drill-hole casings and buried metallic objects should be avoided.
- Read differences between the electrodes at least at the beginning and at the end of the survey to allow for correction of drift. At the beginning of the survey the difference should be less than 2 mV. If greater, the electrodes should be cleaned and filled with fresh electrolyte.
- Long period telluric currents caused by temporal variation in the earth's magnetic field can be several hundred mV/km, and can increase the noise for a survey more than 1 km long.
- Vegetated areas should be avoided because bioelectrical activity can generate SP anomalies as high as 150 mV/m.

SEISMIC GEOPHYSICAL METHODS[13]

Seismic geophysical methods for subsurface investigations are divided into two categories, refraction and reflection. These methods differ in the progression of seismic energy through the earth's layers, and in data collection and interpretation. For shallow engineering, environmental and hydrogeologic investigations, the seismic reflection method is not commonly used. A brief discussion of seismic reflection applied to environmental and hydrogeologic investigations and a limited reference list are presented at the end of this section. The following discussion is a brief review of seismic refraction methods as applied to AMD, environmental and hydrogeologic problems. Haeni (1988) provides a more detailed discussion of seismic refraction applications to hydrologic studies including an extensive annotated bibliography.

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[13] *References begin on page 130.*

Seismic energy travels through the ground as four types of waves: 1) a compressional or longitudinal wave (ground motion parallel to the direction of propagation); 2) a shear or transverse wave (ground motion perpendicular to the direction of propagation); 3) a Rayleigh wave (travels along the surface); and 4) a Love wave (travels along the surface). For engineering, environmental, and hydrogeologic seismic surveys, the compressional wave is utilized most often. The shear wave is used for special engineering studies. Surface waves are not used. Compressional waves travel the fastest, thus they are the first to arrive at the geophones and are the easiest to identify on the seismographic record.

Figure 6 shows schematically the ray-paths of energy from a surface source for a two-layer earth model. The two-layer earth model assumes ideal horizontal, homogeneous, and isotropic layers, and velocity increasing with each deeper layer. Seismic waves travel outward from the energy source along four ray-paths:

- as a direct ray-path along the surface;
- as a totally reflected ray-path that strikes the interface between two layers at an angle greater than a critical angle of incidence (i_c), with all of the energy reflected toward the surface at the velocity of the upper layer (V_1);
- at a critical angle of incidence (i_c) where part of the energy is reflected at velocity V_1 and part of the energy is refracted along the interface between the two layers at the velocity of the deeper layer (V_2); and
- as a ray-path that strikes the interface at an angle less than the critical angle where part of the energy is reflected and part is refracted downward into the deeper layer.

The seismic refraction method uses the first and third ray-path properties and the critical angle of incidence to develop formulas for calculating layer velocity and thickness. The critical angle of incidence at the interface of two layers is given by:

$$i_c = \arcsin(V_1 / V_2)$$

where: V_1 = the velocity of the upper layer

V_2 = the velocity of the lower layer

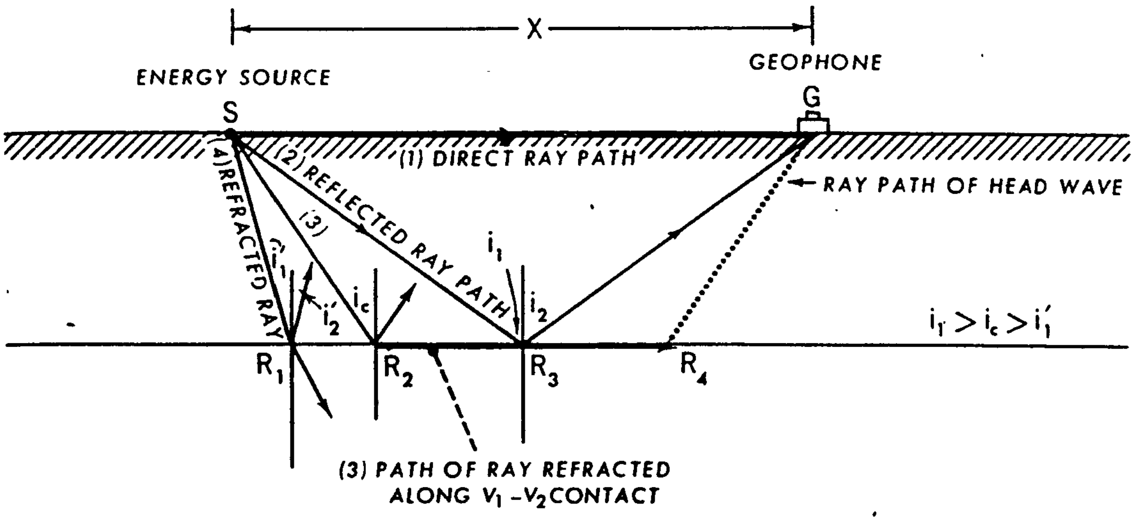

Figure 6. Schematic ray-path diagram for seismic energy generated at source S and received at geophone G (from Zohdy and others, 1974).

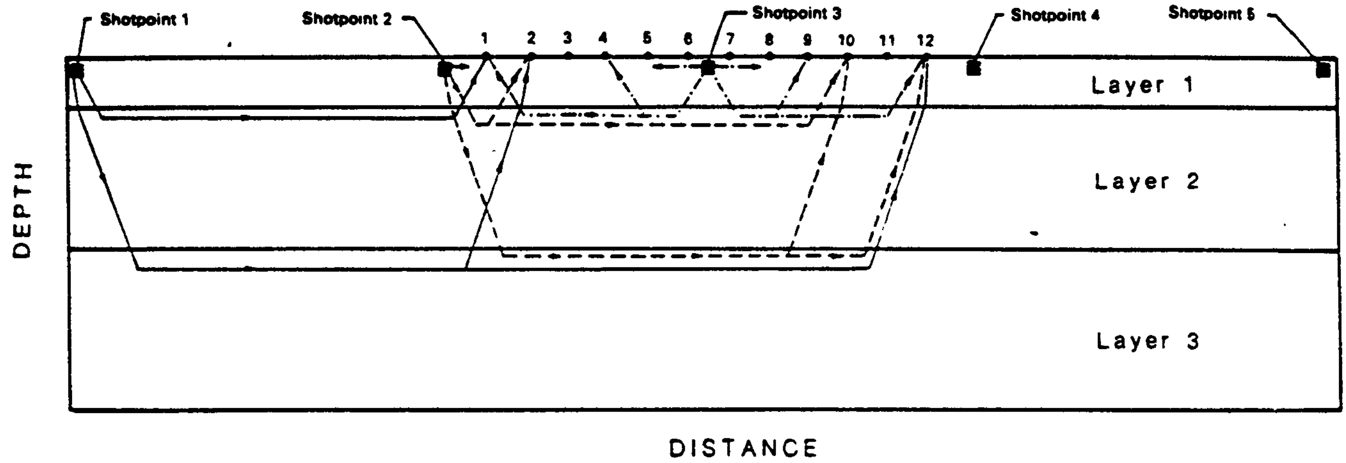

Figure 7. Field setup of shotpoints and geophones for delineation of multiple-refracting horizons. Only selected raypaths for shotpoint 1, 2 and 3 are shown. The raypaths for shotpoints 4 and 5 are the mirror image (with respect to shotpoint 3) of the raypaths for shotpoints 2 and 1, respectively (from Haeni, 1988).

Formulas for solving problems involving more than two layers have been developed from this relationship and other geometric properties (Mooney, 1984; Haeni, 1988; and Crice, 1992). Refraction interpretation formula assumptions are: 1) layer boundaries are planar and have a uniform slope; 2) the land surface is flat; 3) all layers are seismically homogeneous and isotropic; and 4) seismic velocity increases with depth. Formulas for three-layer problems can be readily solved by hand or with hand-held calculators (Ballantyne and others, 1981). More complex formulas are solved using computers (Hunter, 1981; Mooney, 1984; and Orndorff and others, 1989).

Seismic refraction surveys are usually conducted by placing a cable and geophones along a linear trend at regular spacings. To maximize information, energy is sent into the ground from several locations along the geophone spread and off-end. A typical layout of a seismic refraction line is shown in Figure 7. The seismic wave sensed at each geophone is recorded on a seismograph either as an analog or digital signal. The array of geophones is physically or electronically moved along the line to obtain additional seismic profiles. The spacing of the geophones is determined in part by the depth of interest, the thickness of the layers and the size of the target. There should be at least two data points for each layer of interest to avoid ambiguities in interpretation. Refraction cables commonly have takeouts for connecting the geophones at 25-, 50- and 100-foot increments. Actual spacing can be adjusted by leaving slack in the cable between geophones.

For compressional wave seismic refraction surveys, the time of the arrival of the first compression wave is selected for each geophone from the recording. To enhance the signal from a low energy source and to reduce the effect of noise, most modern seismographs allow stacking of repeated shots. Variations in the subsurface geology will cause the arrival times at the geophones to differ and plots of the distance versus first-arrival time will be non-symmetrical. Interpretations of the layer dip, thickness, velocity and material types are based on these variations as seen in the time-distance plots.

Interpretation of refraction data is done by plotting the first arrival times for each geophone on the vertical axis at their distance from the shot point (energy source) on the horizontal axis as shown in Figure 8. Straight lines are drawn through points thought to represent first arrivals at a single layer. The slope of a line is the inverse of the velocity of the layer it represents. Depth to a layer interface is calculated based on either: 1) a crossover distance (x_{cu}, or x_{cd}) at which the refracted wave traveling along the layer interface reaches the geophone faster than the direct wave; or 2) by projecting each segment to zero distance and using the zero intercept time (t_{2u}, or t_{2d}). The crossover distance for each interface is found at the break in slope as shown of Figure 8. The intercept times and the crossover distances are directly dependent on the thickness and the compressional velocity of the each layer. A detailed discussion of seismic refraction theory, field methods, and interpretation is given by Mooney (1984) and Haeni (1988).

An alternative method for interpreting seismic refraction data that give a depth and velocity profile beneath each geophone is the generalized reciprocal method (GRM). GRM interpretation calculates velocity analysis and time-depth functions to define continuous forward and reverse direction profiles using the technique of phantoming (Palmer, 1980 and 1981; Lankston and Lankston, 1986; and Lankston, 1990). The computed values for the velocity analysis and the time-depth functions are referenced to a position, termed the G-position, which lies half-way between two geophones separated by a distance XY. Multiple calculations are done for various XY distances to generate values of the velocity analysis and time-depth functions versus G-position, creating a suite of curves. Based on a subjective criteria of either minimal curve irregularity for the velocity analysis function or maximum curve irregularity for time-depth function, an optimum XY spacing is selected. The time-depth function curve for the optimum XY spacing is then used to calculate the depth to the target refractor surface at each G-position. The closer the geophones are spaced the more detailed the information on the refractor surface.

When an optimal XY distance can be determined with confidence, GRM can determine the depth to the target refractor even when the velocity and thickness of the overlying layers are unknown. GRM can also give an indication that undetected layers are present and indicate whether they are hidden because of a velocity inversion or because they are too thin (Lankston, 1989 and 1990; and Palmer, 1980). The GRM requires many calculations, and computer programs are available (Hatherly, 1976).

Special refraction and downhole survey are performed for additional information about the dynamic properties of the subsurface layers, including shear velocity and shear modulus. The geophones and energy sources differ from normal compressional surveys and digital recording of the wave form is required to process the signal (Mooney, 1974; and Dobecki,1979).

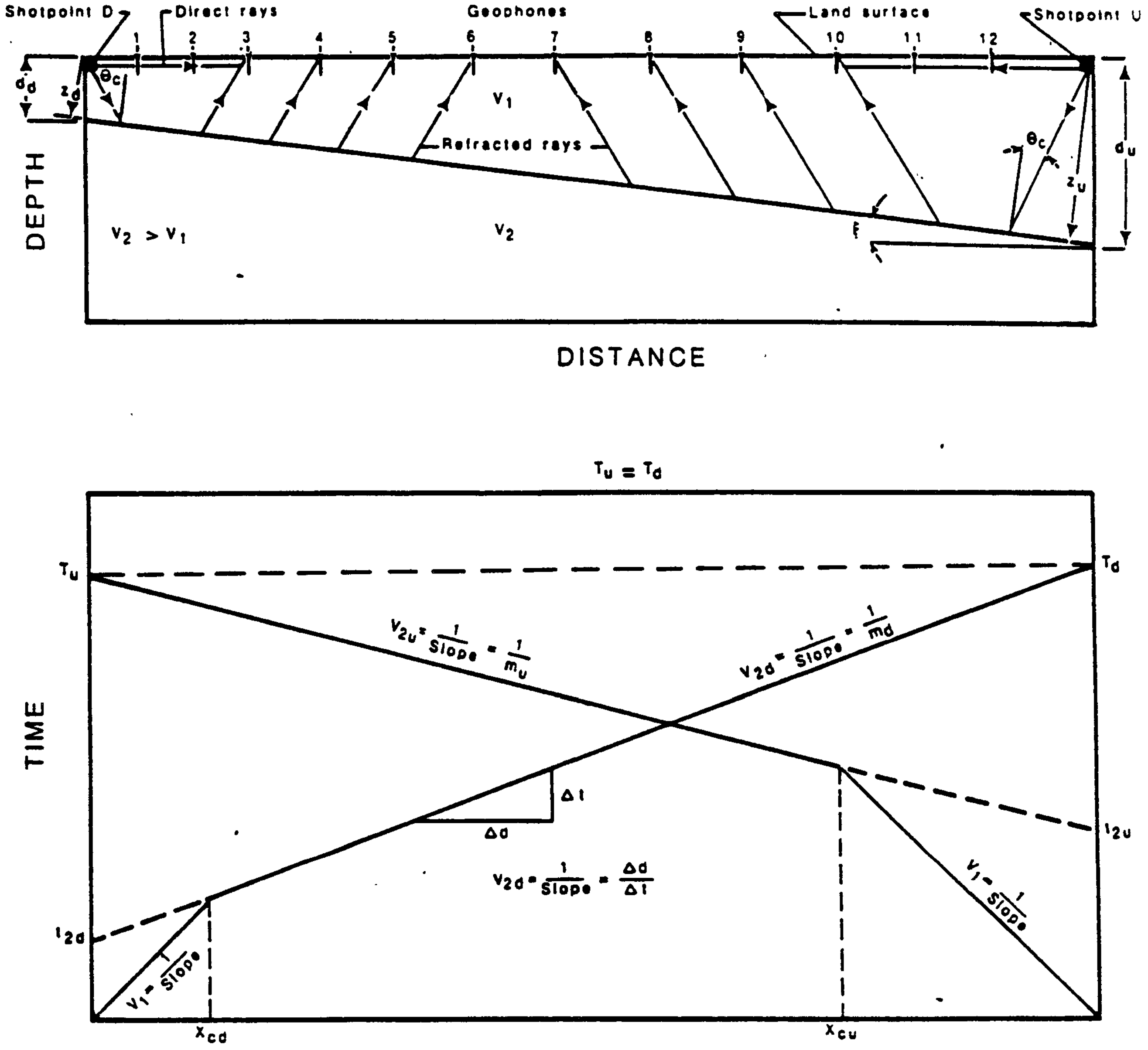

Figure 8. Typical geophone and shotpoint layout with seismic raypaths and time-distance plots for a two-layer model with a dipping boundary (from Haeni, 1988).

Because the material properties measured by seismic methods, primarily velocities of sound waves, are not directly related to AMD geochemistry or conditions that generate AMD, the application of shallow seismic refraction to AMD investigations is limited to general evaluations of subsurface geology and hydrology. Knowledge about subsurface geology and hydrology can complement and help calibrate other geophysical methods such as D.C. resistivity and electromagnetics. Seismic refraction and perhaps reflection surveys can:

- provide complementary data for calibration of other geophysical surveys;
- delineate depth and lateral extent of waste fills;
- define and locate geologic structures such as faults and formation contacts;
- identify gravel channels within finer grained materials;
- define water table in unconsolidated sediments;
- identify subsurface voids; and
- azimuth refraction surveys can be used to measure anisotropy in fractured rock and in some cases determine the main direction of vertical fracturing.

Papers that describe the use of seismic refraction methods in environmental and hydrogeologic investigations include: Burke (1967), Eaton and Watkins (1967), Lennox and Carlson (1967), Bruehl (1983), Kopsick and Sanders (1983), Taylor and Cherkauer (1984), Underwood and others (1984), Haeni (1988), King and others (1989), Barker (1990), Lankston (1990), Steeples and Miller (1990), Carpenter and others (1991), Cooksley (1992), Crice (1992), and Davies and others (1992).

Although reflection methods are generally not applied to shallow environmental or hydrogeologic investigations, some research has been reported by Muir and Higgins (1990), Davies and King (1992), Davies and others (1992), King (1992), Hill (1992), and Brabham and McDonald (1992).

GRAVITY GEOPHYSICAL METHODS[14]

Measurement of the earth's gravitational field at the surface are compared to an arbitrary datum to learn about subsurface structures. Predictable differences in the earth's gravitational field are caused by changes in elevation and latitude, diurnal tidal distortion, and lateral variations in rock bulk density. The standard cgs unit for the earth's gravitational acceleration is approximately 980 cm/sec^2 or 980 Gals (Galileos) = 1g (gravity), and 1 Gal = 1 cm/sec^2 = 0.0010197 g. The units of gravitational acceleration commonly used in land gravity surveys is milligals (mGals). One milligal is one-thousandth of a Gal $\approx 1 \times 10^{-6}$ g. Gravity meters can detect variations in the earth's gravitational field as small as a few hundredths of a milliGal or better. For example, the LaCoste and Romberg Model D microgravity meters are precise to 0.001 mGal (1×10^{-9} g). In order to identify small geologic changes, field data must be corrected to account for variations in elevation, latitude, terrain, tidal effects and instrument drift (Telford and others, 1990).

Free-air and Bouguer elevation corrections must be applied to field data. The free-air correction adjusts for the difference in distance from the center of the earth to the datum plane and to the station. The datum plane is usually sea level but adjustments can be made to any datum. The adjustment is the addition of 0.09406 mGal per foot (0.3086 mGal/m) of elevation above the sea level datum. The adjustment is made regardless of whether or not there is rock material between the sea level datum and the station, hence the term free-air correction. The Bouguer correction removes the effect of the gravitational attraction caused by an assumed infinite slab of material between the horizontal plane of each station and the datum. The Bouguer correction assumes an infinite slab of material of bulk density ρ_b with a base at the datum (sea level) and the top at the station elevation. A Bouguer correction of $0.01278 \cdot \rho_b$ mGal/ft ($0.04192 \cdot \rho_b$ mGal/m) is subtracted from the field data, assuming a sea level datum (Telford and others, 1990).

Field data must be corrected for latitude because the earth is not a true sphere, but an ellipsoid. The latitude correction is subtracted from the free-air and Bouguer corrected data and is given by the formula (Telford and others, 1990):

$$g = 978031.846 \cdot [1 + 0.005278895 \cdot \sin^2(\phi) + 0.000023462 \cdot \sin^4(\phi)]$$

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[14] *Reference begin on page 133.*

where: g = acceleration of gravity, mGals

ϕ = the latitude of station in degrees

The terrain correction adjusts for the effects of topographic variations in the vicinity of the data station. Hills nearby the station give an upward component of gravitational attraction that counteracts a portion of the downward pull of the earth. Valleys nearby have the effect of causing a smaller downward pull at the station than is accounted for by the Bouguer correction. The terrain correction is begun by centering a template over the station and estimating the average elevation within each area delineated by radial lines and concentric circles. The terrain correction factor is the elevation difference between the station and each compartment, and is calculating the gravitational effect represented by each compartment. The total terrain correction factor is added to the gravity value at each station (Dobrin, 1976; and Telford and others, 1990).

If a high precision gravity survey is required, correction for the effects of tidal changes and instrument drift must be made. Periodic readings at one station are necessary to estimate the magnitude of the drift and adjust the data accordingly. Earth's tide can cause as much as 0.3 mGal cyclic changes in gravity at any one point (Telford and others, 1990).

The correction of the field data for free-air, Bouguer, and latitude is termed the Simple Bouguer Anomaly. The inclusion of the terrain correction is called the Complete Bouguer Anomaly.

One of the main problems with interpretation of gravity data is the need to remove the effects of other features, such as regional gradients, that are superimposed onto the signature of the anomaly of interest. Several methods that have been used to extract the anomaly of interest include the following methods (Telford and others, 1990):

- The trend surface method assumes that regional gravitational effects can be modeled by a simple low-order polynomial surface such as a plane. The trend surface is then subtracted from the observed data and a map of gravity residuals results. The concept is that the removal of the regional data will enhance the trends caused by the local anomalies. In some cases where data are noisy, a high-order polynomial surface is fit to the observed data before subtracting out the low-order regional surface. Drawbacks to the method include that the regional effects may not be accurately represented by a simple surface and that the use of high-order polynomials to fit the field data can introduce anomalies that are due solely to the fitting processes.

- A spatial frequency filtering method uses band-pass filters to isolate spatial frequency components caused by shallow or deep density changes. This method requires some knowledge of the wavelength of the feature being investigated. The apparent wavelength of the anomaly is roughly proportional to the depth of the lateral density change causing the anomaly. A shallow mass or near-surface noise causes a short wavelength anomaly while deep regional mass causes a broad wavelength anomaly.

- The forward modeling method computes an ideal gravitational simulation that should be caused by the feature being investigated. Results of the simulated model are then compared with the actual data. Repeat simulations and comparisons are made until an acceptable fit is found.

- The graphical method produces a smoothed surface that does not contain local anomalies and then this surface is subtracted from the original data to produce a map of residuals representing the local anomalies.

- The gridding method is used to predict the regional gradient by averaging the values within a specific radius of a station. The radius is on the same order of magnitude as the depth of the anomaly of interest.

- The second derivative method uses a grid to calculate the second derivative at a station from a weighted average of surrounding stations. The second derivative method measures the curvature of the gravity field and is most influenced by shallow anomalies.

For environmental and engineering studies where the targets are generally small shallow features, high-precision microgravity surveys are done. Microgravity studies of small areas require precise instrument readings and gravity station location. A precision of 10 centimeters in elevation and 30 meters in latitude (0.03 mGal accuracy) with grid spacings of a few meters is commonly needed (Telford and others, 1990). Environmental surveys of small areas generally assume that the regional gradient is a simple sloping plane or curve. Like most geophysical methods, interpretation of gravity data is non-unique and the interpreter must use all available data to limit the solutions.

Application of gravity surveys in AMD investigations would include: delineating waste pile dimensions, especially for deep canyon fills; and identifying voids, tunnels and subsurface mine workings. Reports and papers on the use of gravity surveys for environmental, engineering and hydrogeologic studies include: Dean (1958), Eaton and Watkins (1967); Lennox and Carlson (1967); Arzi (1975), Ibrahim and Hinze (1972), Zohdy and others (1974), Carmichael and Henry (1977), Butler (1984), Dahlstrand (1985), Roberts (1989), Roberts and others (1989), Hinze (1988), Kick (1989), Adams and Hinze (1990), Hart and Muir (1990), Richard and Wolfe (1990), Sandberg and Hall (1990), Wolfe and Richard (1990), West (1992), and Kelly and Mares (1993).

MAGNETIC GEOPHYSICAL METHODS[15]

Land based magnetic surveys measure the variations in the earth's magnetic field caused by near-surface ferrous materials to locate buried targets. The earth's total magnetic field intensity is not uniform because of the dipoler nature of magnetism and because of the distribution in the earth of rocks having differing magnetic properties. The sun and solar flares temporarily distort the earth's magnetic field.

Diurnal variations occur during daylight hours although their timing and magnitude are not totally predictable (Telford and others, 1990). They usually must be removed from total field intensity data because their magnitude of tens of gammas can obscure anomalies of interest. During field surveys that last several hours, a recording base station or periodic recording at the same point is needed in order to record the diurnal variation.

Modern magnetic instruments usually measure the total magnetic field intensity. An alternative method is the vertical gradient magnetometer survey. This method measures the magnetic field at two points vertically separated by approximately 1 meter with the gradient being the change in magnetism between sensors. While the total magnetic field for the earth varies throughout the day, the gradient between these two vertically separated points should be relatively constant. Thus there is no need to correct for diurnal variations, and repeat surveys should give similar data.

The magnetometer survey is generally carried out along a linear traverse or a grid. Data corrected for diurnal effects and sometimes smoothed are generally analyzed qualitatively by contouring or plotting a profile. Quantitative interpretation can be done (Telford and others, 1990), but usually the buried metal objects of interest in environmental work do not lend themselves to simple geometric models.

While there are several types of instruments used for measuring the earth's magnetic field, the proton-precession magnetometer is the most common in environmental and engineering studies (Breiner, 1973 and 1992; and Telford and others, 1990). The proton-precession magnetometer can read the total magnetic field to a sensitivity of 0.1 gamma. The total intensity of the earth's

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[15] *References begin on page 135.*

magnetic field in the contiguous United States ranges from 50,000 to 60,000 nanoteslas (1 nanotesla = 1 gamma = 10^{-5} gauss = 10^{-5} oersted = 10^{-9} webers/m^2 = 10^{-9} tesla).

Magnetometer surveys can be of use in AMD investigations because they allow for rapid identification of shallow buried ferrous objects, common to mine sites. Magnetometer surveys can also aid in distinguishing between a conductive soil/rock mass from buried ferrous metal as being the source of an electromagnetic, low resistivity anomaly. For this reason, magnetometer surveys are commonly done for quality control in electromagnetic and resistivity surveys.

Details on the theory and application of proton-precession magnetometers can be found in Breiner (1973 and 1992) and Telford and others (1990), as well as general geophysical textbooks (see list of general references). Discussions of survey methods and interpretation for environmental problems are given by Zohdy and others (1974), Fowler and Pasicznyk (1985), Hinze (1988), Allen and Rogers (1989), and DeReamer and Pierce (1990).

GROUND PENETRATING RADAR METHOD[16]

In recent years the increased demand in the environmental studies for detailed knowledge about the shallow subsurface and of objects buried within the upper 15 to 30 feet has led to the development of a ground-based, high frequency, electromagnetic geophysical instrument known as ground penetrating radar (GPR). This geophysical instrument employs a short duration, electromagnetic pulse using a broad-bandwidth antenna placed directly on the ground. The depth of penetration and resolution are controlled by the antenna frequency which commonly ranges from 80 Hz to 1000 Hz. GPR detects differences in the dielectric properties of buried materials. Voids and buried objects such as barrels and sewer lines have sufficiently different dielectric properties to be detected.

GPR has advantages over other geophysical methods in that: 1) it can detect changes in the dielectric properties of material that may not be apparent to magnetic or resistivity geophysical methods; 2) it has the highest target resolution of any shallow depth geophysical method; 3) a survey can be done as rapidly as the operator can walk it; 4) output is a real-time plot and does

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[16] *References begin on page 136.*

not require correction or processing; and 5) real-time graphic output allows rapid interpretation and modification of survey.

GPR limitations include: 1) survey results are site dependent; 2) an increase in moisture content of soils can reduce the signal, and ground water absorbs the signal; 3) an increase in clay content reduces the signal; 4) penetration and resolution are frequency dependent; 5) the antenna must pass near the object to detect it.

GPR surveys are typically conducted along a grid or a linear traverse across an area of interest. Grid size varies with the target size (see previous section on line spacing and probability, page 24). Preliminary surveys are needed to calibrate the instrument to site conditions.

GPR can be applied to AMD investigations where shallow buried objects are of interest and contrasts in the dielectric properties of materials or voids are assumed. These types of investigations might include locating buried barrels, pipelines, old disposal trenches, and shallow voids. A drawback to the use of GPR is the radar antenna must be in direct contact with the ground to develop an adequate return signal. An irregular surface common at mine sites will significantly affect the results.

Review of GPR theory and methods for environmental studies is given by: Morley, (1974), Cook (1975), Coon and others (1981), Benson and others (1984), Davis and others (1984), Wright and others (1984), Underwood and Eales (1984), Olhoeft (1988), Boucher and Galinovsky (1989), Davis and Annan (1989), Daniels (1989), Filler and Kuo (1989), Hennon (1990), Beres and Haeni (1991), Fisher and others (1992), and Allen and Seelen (1992).

BOREHOLE GEOPHYSICAL METHODS[17]

Borehole geophysical or well logging includes all methods for lowering a sensing device down a cased or uncased borehole to record physical, chemical, electrical, electromagnetic, or radioactive parameters along the well bore. Borehole logging is used extensively by the petroleum and minerals industries, and to a limited extent by the geotechnical and water resources industries. In

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[17] *References begin on page 138.*

recent years with the increase in the number environmental and hydrogeologic investigations associated with ground-water pollution, borehole geophysics have become an important tool to assist the geologist and hydrogeologist in interpreting subsurface geology and hydrology.

Papers that report extensive use of borehole geophysical methods in the study of AMD problems or in the monitoring of AMD were not found in the literature search. Nevertheless, there are numerous studies of high specific conductance ground-water pollution that utilized borehole geophysics. These studies clearly indicate that borehole geophysics can have direct application to AMD investigations and to long-term monitoring of AMD pollution. There are numerous borehole geophysical methods and many variations on each method developed for special applications. The purpose of this summary is not to cover the topic in detail, but to provide a general overview of the available methods and references for more detailed research. Therefore, the following is a brief discussion of the methods of borehole geophysics as might be applied to the investigation of AMD.

The types of borehole geophysical logging devices are numerous but can be generally divided into the following five groups based on the parameter measured:

- Electrical resistivity and conductivity;
- Nuclear;
- Acoustic;
- Borehole fluid; and
- Well construction.

Electrical resistivity or conductivity logging tools are designed to measure formation and formation water resistance using a variety of designs and electrode configurations. These methods generally are done in an open hole although some of the electromagnetic induction tools can penetrate non-conductive well casing, such as plastic. The electrical methods are accomplished by inducing current into the borehole wall and reading the response at one or more electrodes. The electrical resistivity methods include: point resistivity, short normal, long normal, lateral resistivity, focused resistivity, microlog, dipmeter, induced polarization, and electromagnetic induction logs. This group should also include a special type of electrical log, spontaneous potential, that measures natural potential differences caused by electrochemical reactions between layers.

Nuclear logging commonly uses measurements of either the level of natural radioactive material decaying in formations, or the level of backscatter or adsorption in the formation material caused by a radioactive source within the logging tool. The characteristics of the formations are determined by the amount of natural radioactivity or the response to radiation. An advantage of radioactive logs is that they can be run in open or cased holes. Commonly used radioactive logs include: natural gamma, gamma-gamma, neutron, spaced neutron, density, gamma spectrometry, neutron-activation, pulsed-neutron-decay, and dual-detector-density logs.

Acoustic logging uses sound waves generated by the logging probe to measure formation properties, primarily porosity, fracture density and orientation, and the quality of annular space cementation. Acoustic methods include: sonic, tube wave amplitude, variable density, full waveform sonic, borehole acoustic televiewer, and acoustic cement bond logs.

Borehole fluid logs measure the physical and chemical characteristics of the fluids in the borehole and the adjacent formation and include: fluid conductivity, brine injector-detector, radioactive tracerjector, Eh, pH, temperature and differential-temperature, impeller-type flowmeter, and heat-pulse flowmeter logs. These logs must be run in an open hole that has ground water in equilibrium with the formation water.

The final group are well-construction logs: downhole video camera, casing-collar indicator, directional survey and caliper. These are run as part of the engineering design of the well or following well completion to document the construction.

Interpretation of borehole logs can be done qualitatively or quantitatively. Most quantitative methods were developed by the petroleum and minerals industries. Numerous books are available on well log interpretation and include: Lynch (1962), Schlumberger (1972a, 1972b, 1974, 1977), Dresser Atlas (1975), Merkel (1979), Hearst and Nelson (1985), Labo (1986), and Brock (1986).

Several papers that discuss the principles, applications and interpretation of borehole geophysics as it relates to environmental and fresh ground-water investigations are available and include:

Jones and Buford (1951), Keys (1967 and 1989), Keys and MacCary (1971), MacCary (1978), Engineering Enterprises (1985), Driscoll (1987), Collier and Alger (1988), Stegner and Becker (1988), Stowell (1989), Daniels and Keys (1990), Howard (1990b), Paillet and Saunders (1990), Roscoe Moss Company (1990), Yearsley and Crowder (1990 and 1991), Crowder and others (1991), Yearsley and others (1991), Yearsley and Crowder (1991), Mwenifumbo (1993), and Welenco Inc. (undated).

Several authors address the use of borehole geophysical logs in determining subsurface parameters of shallow formations and fresh-water aquifers. Parameters that are measured include: lithology, porosity, permeability, bulk density, clay and shale content, moisture content, ground water flow direction and velocity, and chemical dispersion. Papers on these subjects include: Croft (1971), Barker and Worthington (1973), Griffiths (1976), Ogbe and Bassiouni (1978), Biella and others (1983), Snelgrove and McNeill (1985), Spencer (1985), Huntley (1986), Taylor and others (1989), Burns (1990), and Hess and Paillet (1990).

Correlation of borehole geophysical measurements and quality of shallow ground water is discussed by numerous authors including: Turcan (1962 and 1966), Alger (1966), Moore (1966), Desai and Moore (1969), Worthington and Barker (1972), Worthington (1976), Kwader (1985 and 1986), Guo (1986), Alger and Harrison (1988), and Jorgensen (1989 and 1990).

Studies on the use of borehole geophysical methods used to investigate fracture zones, fracture density, orientation, and transmissivity are given by: Deluca and Buckley (1985), Morin and others (1988), Merin (1989), and Howard (1990a).

Figure 9 is a matrix listing most of the geophysical logs commonly used in environmental and ground-water investigations along with the applicability of each method to various geologic and hydrologic studies (Keys and MacCary, 1971; Welenco, undated). Figure 9 indicates whether the logging method can be used with open or cased holes.

Application of borehole geophysical logs to the investigation of AMD is similar to the application to other environmental investigations. Applications include: aid in identifying and correlating subsurface geologic units, determining physical parameters such as hydraulic conductivity, determining water quality, evaluating the interconnection of aquifers, and determining the effectiveness of ground-water cleanup programs. An additional use of borehole logging for AMD investigations is the use of conductivity logs to map the extent and change in high specific conductance ground-water pollution. For example, electromagnetic induction tools, such as the Geonic's EM39, can be run down wells with non-metal casings to measure the conductivity of the formation fluids. Thus periodic logging of the well can monitor the appearance and extent of an AMD plume.

	SPONTANEOUS POTENTIAL	SHORT NORMAL	LONG NORMAL	LONG LATERAL	INDUCTION LOG	MICROLOG	E M CONDUCTIVITY	CALIPER LOG	GAMMA RAY LOG	NEUTRON LOG	SPACED NEUTRON LOG	DENSITY LOG	SONIC LOG	TEMPERATURE LOG (DIFFERENTIAL)	DIPMETER	SIDEWALL CORES	FLOWMETER LOG	FLUID CONDUCTIVITY	DOWNHOLE CAMERA	COLLAR LOCATOR	DIRECTIONAL SURVEY
HOLE INCLINATION															○						○
FISH LOCATION																			○●	○	
DIP AND STRIKE OF FORMATIONS															○						
DEPTH CONTROL																			●	●	
HOLE DIAMETER								○●													
CASING LEAKS OR PLUGGING														●					●		
CASING CORROSION								●											●		
WELL CONSTRUCTION								●				●							●		
CEMENTING							●					●	●	●							
SOURCE & MOVEMENT OF WATER IN WELL														○●			●		●		
DISPERSION DILUTION							○	●						●				●			
DIRECTION VELOCITY GROUND WATER														○●			○●				
INFILTRATION										○●	○●			●							
MOISTURE CONTENT										○●	○●										
CHEMICAL & PHYSICAL PROPERTIES	○						○	●								○		●			
WATER TABLE			○							○●									○●		
BULK DENSITY												○●				○					
EFFECTIVE POROSITY			○	○							○					○					
TOTAL POROSITY										○●	○●		○			○					
CLAY & SHALE CONTENT	○	○	○	○	○	○	○	●	○●	○●	○●	○●				○					
BED THICKNESS	○	○			○	○	○	●	○●	○●	○								○		
LITHOLOGY	○	○	○	○	○	○	○	●	○●	○●	○●	○●	○●			○			○		
CORRELATION	○	○	○		○		○	●	○●	○●	○●		○●			○			○		

Figure 9. Matrix of applicability of borehole geophysical methods commonly used in engineering and environmental investigations. Open circle = open hole; filled circle = cased hole. Casing material for EM methods must be non-conductive such as plastic. (modified after Welenco, undated; and Keys and MacCary, 1971)

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V. LIST OF ABBREVIATIONS

A = ampere
B = magnetic field
°C = degrees Celsius
cgs = centimeters-grams-seconds
cm = centimeter
E = electrical field strength
𝔉 = Faraday's constant
H = magnetic field strength
Hz = Hertz = cycles / second
°K = degrees Kelvin
kJ = kilojoules
L = liter
m = meter
mg = milligram
mg/L = milligrams / Liter
mks = meters-kilograms-seconds
mho = 1/ohms = 1 / Ω
mmho/m = millimhos / meter
mm = millimeter
mol. = mole
mS = milliSiemen
mS/m = milliSiemens / meter
mV = milliVolt
Pa = Pascal
s = second
S = longitudinal conductance
T = transverse resistance
v = volts
ρ = resistivity
σ = conductivity
ϕ = phase difference between E and H
Ω = ohm
μ = magnetic permeability
μH = magnetic field
μS = microSiemen
μS/cm = microSiemens / centimeter
μmho/cm = micromhos / centimeter
μmho/m = micromhos / meter