GEOPHYSICS FOR MINERAL EXPLORATION

A Manual for Prospectors

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FOREWORD

This manual was generated as a result of persuasion by Norm Mercer, then Mineral Exploration Consultant with the Newfoundland Department of Natural Resources (NDNR), while the author was with GeoScott Exploration Consultants Inc. Both of these people are now retired. The manual has benefitted greatly from contributions from Gerry Kilfoil and Phil Saunders, both of NDNR, and from Ron Verrall and Susan Scott, consultants.

Any errors that remain, however, are the responsibility of the author.

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GEOPHYSICS FOR MINERAL EXPLORATION

1.0 INTRODUCTION

The purpose of carrying out geophysical surveys is to find out something about the rocks in the survey area. Geophysical methods all depend on measuring a physical property of rocks. There are only a few such properties, shown in Table 1:

Table 1: Geophysical properties useful for surveys, roughly in order of increasing survey cost.

Property	Typical rocks/minerals	Geophysical Methods	Notes
Magnetic susceptibility	Magnetite, ultra-mafics, iron-rich rocks	Magnetic	Mainly used for geologic mapping, sometimes for identifying magnetite bodies or kimberlites.
Electrical Conductivity sigma (σ) or Resistivity rho (ρ) $\sigma = 1/\rho$	Metallic sulphides, serpentinite, graphite, water-filled shears, salt water	Electromagnetic (EM)	VLF EM Frequency-domain (MaxMin), Time-domain (TEM or TDEM, Pulse EM) finds conductive minerals, but not sphalerite
Radioactivity	Potassium (feldspars), Uranium, Thorium	Radiometric, gamma-ray spectrometry	Surveys generally respond only to sources at or very near the surface.
Electrical Polarisability	Metallic sulphides, graphite, serpentinite, magnetite	Induced Polarisation (IP)	Time-domain, frequency domain or phase angle. Finds massive or disseminated metallic sulphides, but not sphalerite.
Density	Sulphides including sphalerite, barite	Gravity	Can find sphalerite and other sulphides, useful for geologic mapping.
Seismic velocity	Massive sulphides have higher velocities than barren rocks.	Seismic reflection or refraction	Generally used to map structure in 3D, sometimes to identify massive sulphide bodies. Can look kilometres deep. Is very expensive.

Geophysical surveys generally look for concentrations of anomalously high values of the property being measured. The results of the survey are used to identify a target of interest, or to correlate the spatial variation of values of the property with variations in the geology. Thus the primary reasons to do geophysics are to get information on geology, and possibly to find targets of economic interest.

2.0 MAGNETIC SURVEYS

The earth has a magnetic field which behaves roughly like a bar magnet not quite lined up with the earth's north-south axis. In 2012, the north magnetic pole lay northwest of Resolute, and is drifting to the north and west. In 2015 it is predicted to be at 80.3N, 72.6W. In central Newfoundland, a compass would point to magnetic north, about 20 degrees west of true north. In Newfoundland the magnetic field of the earth is inclined downwards at about 68 degrees to the horizontal. The field interacts with the rocks through which it passes. If the rocks are magnetic (have high susceptibility) they become magnetised, and their field adds to that of the earth. Thus the total magnetic field is stronger over magnetic rocks. Magnetic fields are measured in nanoteslas (nT), which used to be called gammas (γ).

Magnetic surveys are carried out with magnetometers, which measure the total magnetic field with high precision. Because the earth's magnetic field varies slightly from day to night, and with solar activity, it is important to use a base station magnetometer to record the daily or diurnal changes in the magnetic field at a fixed point. The base station should be established in an area with low spatial magnetic variation. The base-station values can be used to correct the readings taken with a roving magnetometer. Typically, magnetic readings are taken at fixed intervals (12.5 m or even 5 m) on grid lines. Some magnetometers are coupled to GPS receivers. They can be operated in 'walking' mode, with readings taken either at fixed time intervals or at

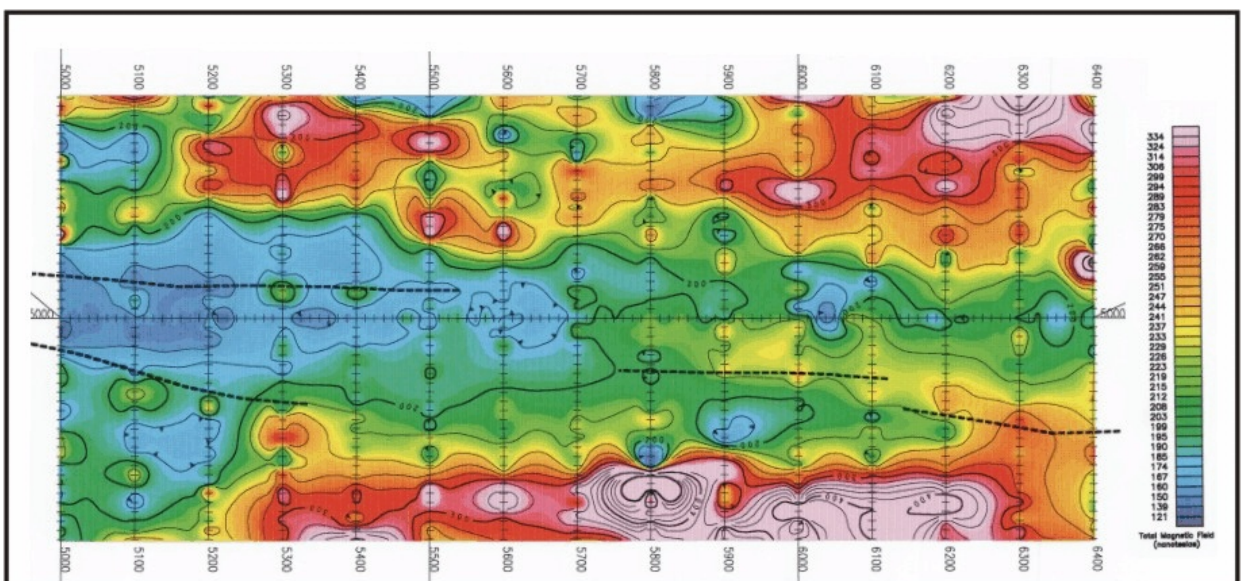


Figure 1: Total field magnetic map, central Newfoundland. The dashed lines are axes of weak HLEM conductors, generally parallel to the structure indicated by the magnetic contours. (NDNR Geofile 012A_1312)

fixed distances taken from the GPS positions. Reconnaissance surveys with a walking magnetometer and GPS do not require a cut grid. Magnetometer operators should ensure that they do not carry any strongly magnetic items such as an axe or geological hammer, which might influence readings.

Magnetic maps are interpreted in terms of geology. Low-susceptibility rocks such as limestones show as areas of low and relatively uniform magnetic fields, while mafic and ultramafic rocks show as areas of higher and more variable magnetic fields. Comparison of magnetic readings in areas of known geology allows extrapolation of the geology into areas where the rocks are covered. Less frequently, magnetic surveys are used to outline areas with high concentrations of magnetite, potentially as sources for iron ore. To help visualise the distribution of values, colour is employed. In geophysical displays, deep blues denote the lowest values, while reds and purples identify the highest readings. A scale bar is usually included to show the range of values.

Figure 1 shows a total field magnetic map from an area in central Newfoundland (NDNR Geofile 012A_1312). The magnetic survey outlined three distinct zones where the magnetic field ranges from a low of 150 nT (blue areas) to a high of 500 nT (red areas). Note that this is a relatively low range of magnetic field values. Although there are three distinct zones, none of the underlying rocks is strongly magnetic. Also shown on this map are axes of weak electromagnetic couductors (see discussion in Part 3.2 below), generally aligned with the edges of the magnetically indicated rocks.

The central part of the grid (blue area) has the lowest magnetic expression, whereas the areas to the grid north and grid south (yellows through reds) are more magnetic. Because of the lack of outcrop information in the area it is difficult definitively to correlate these zones with bedrock geology. An outcrop of visibly altered felsic volcanics appears to be located between the northwest magnetic high and the central low. It is possible that the grid north high is correlative with a plagioclase and blue quartz porphyritic felsic tuff unit, even though it was not noted to be magnetic. On the same basis, the grid south high may be correlated with a unit of mafic volcanics which was noted to be weakly magnetic at one sample location. Without the magnetic map, it would be difficult to make any deductions about the geology.

3.0 ELECTROMAGNETIC SURVEYS

Electromagnetic (EM) surveys use a transmitter to generate a time-varying electromagnetic field in the earth, known as the primary field. This field gives rise to small time-varying voltages in the earth. Where the earth is conductive, the voltages drive small time-varying flows of current, which give rise to electromagnetic fields of their own called secondary fields. The primary and secondary fields add together.

EM surveys measure the earth's willingness to conduct electricity, or conductivity (σ in siemens/m). The higher the conductivity, the more current will flow in the earth for a given electrical field strength. If the earth is not conductive, the unwillingness of the earth to conduct electricity is expressed in resistivity (ρ in ohm-metres). The higher the resistivity, the less current will flow for a given electrical field strength. At low frequencies, conductivity and resistivity are inversely related: $\sigma = 1/\rho$.

3.1 VLF EM Surveys

The simplest EM technique is VLF EM. For VLF (very low frequency)EM, the transmitters are operated for communication purposes by the United States and other countries for communication with their submarines. The frequencies are typically at around 20 kilohertz (KHz). In terms of radio transmission, this is a very low frequency. VLF EM surveys can be carried out either with a Geonics EM16 or with a Crone Radem. In addition, most magnetometers offer the option of including VLF measurements at the same time. Survey lines are usually laid out to be roughly at right angles to the direction to the transmitter station.

Figure 2 shows model VLF EM profiles over a vertical conductive dyke at depths from 10 to 200 metres, calculated after the method of Fraser (1969). The surface projection of the conductor is at the crossover point. If the two halves of the profile are symmetrical, as in Figure 2, then the conductor is vertical. If the profile is not symmetrical, then the larger peak lies over the down-dip side of the conductor. As depth increases, the separation between positive and negative peaks increases, but there is no simple theoretical relationship. The signal from a VLF transmitter generates an alternating sheet of current in the earth, with the flow direction everywhere oriented towards the transmitter. If the resistivity of the rocks is everywhere uniform, then the current sheet is uniform as well. However, if, for example, an appropriately-oriented long shear zone of 1 000 ohm-metres (Ω-m) lies in a host of 10 000 Ω-m, then the current sheet will be deformed and ten times as much current will flow in the shear zone as in the host rocks. The shear zone will give rise to a strong VLF anomaly, even though it does not have a very low resistivity. As a result of this phenomenon, known as current channeling, many VLF anomalies are generated by geological features which do not necessarily carry sulphides. Thus a VLF survey is an aid in mapping geological structure, rather than just a means of finding conductors which could be hosts to economic mineralisation. Current channeling enhances the response of features with significant strike length, generating in them stronger anomalies than in short conductors with the same resistivity contrast.

Consequently, for a conductor to have a VLF anomaly, it must strike generally towards the VLF transmitter. Thus, not all conductors will be detected in a VLF EM survey with a single

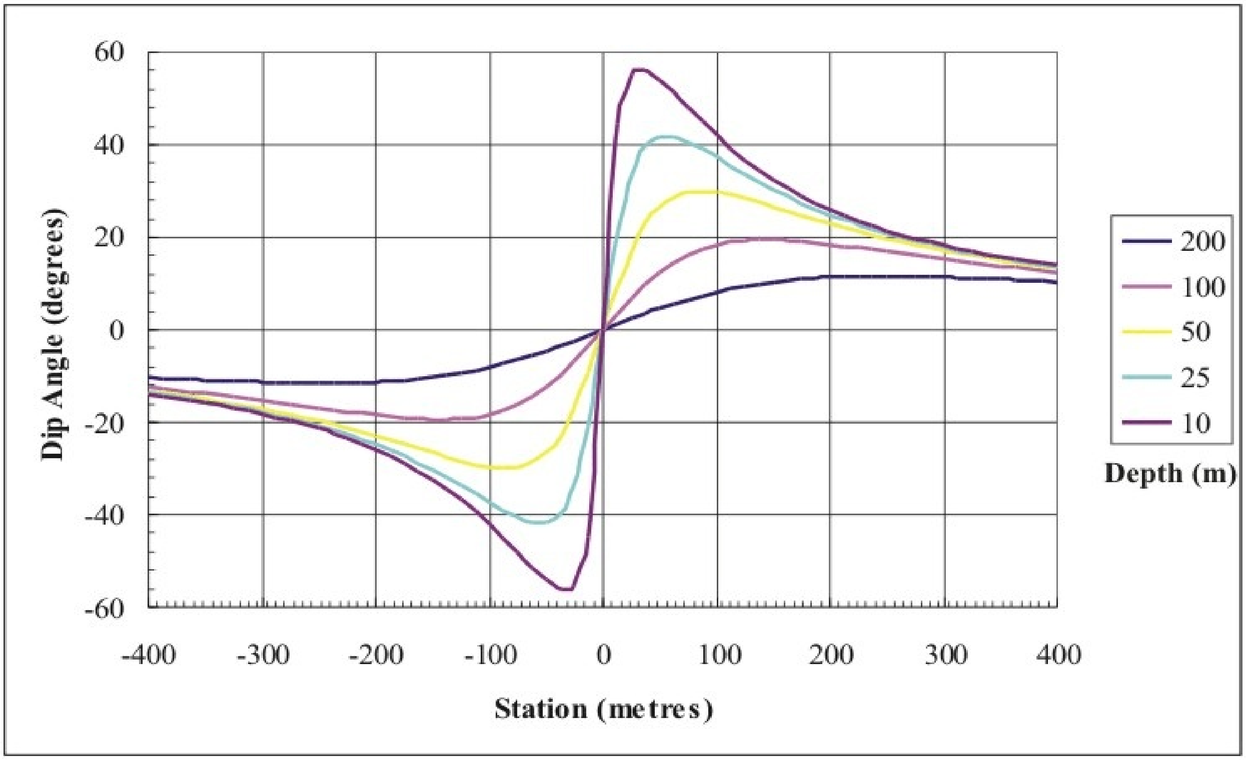


Figure 2: Model VLF EM dip angle profiles for a conductive vertical dyke, calculated by the method of Fraser (1969) for depths from 10 to 200 metres.

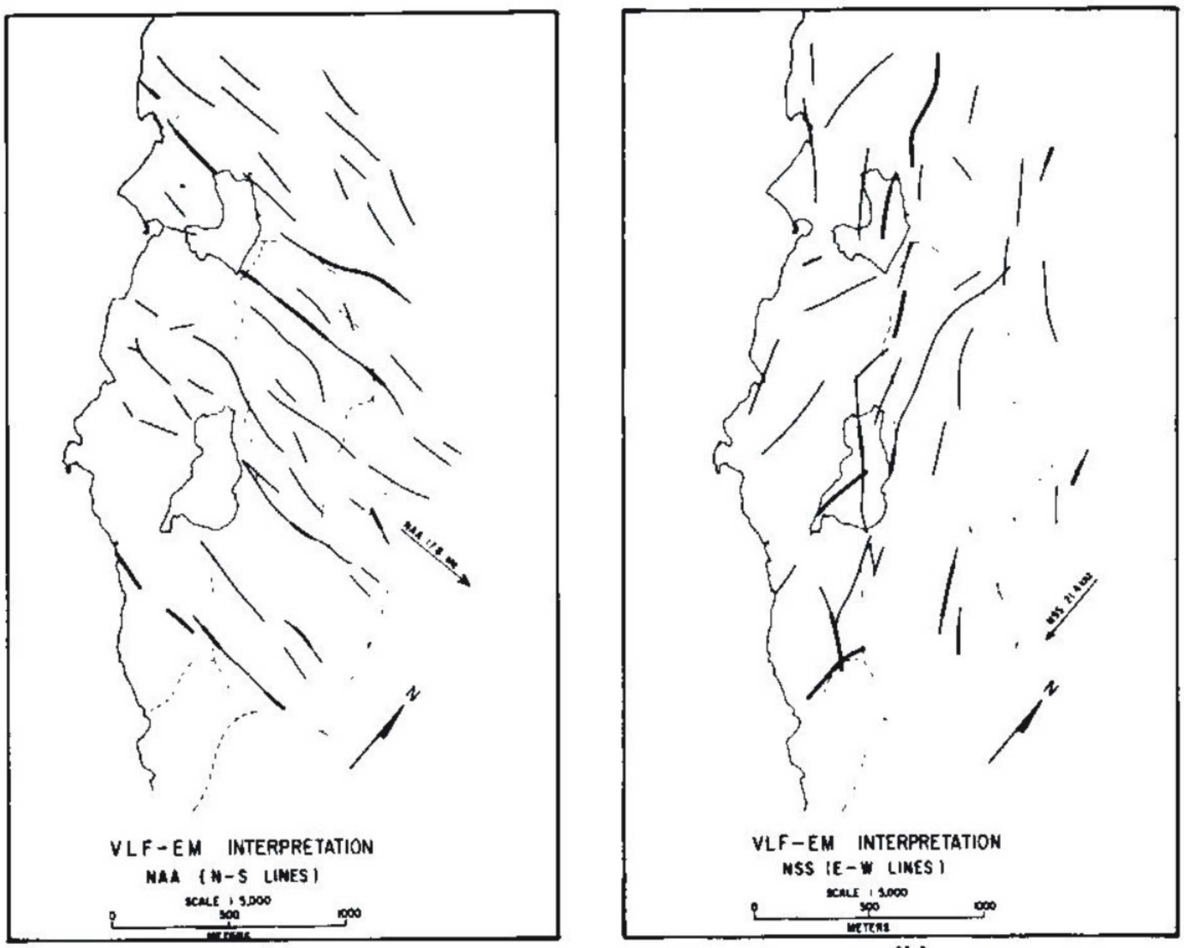
transmitter azimuth. In Newfoundland the useable transmitters are in Cutler (Maine), Seattle (Washington), and possibly Rugby (England). The great-circle bearings (determined by laying a string on a globe to connect the transmitter and the field area) of all these transmitters lie within 15 degrees of each other, at about 105 degrees. Thus a long conductor striking generally south-westerly will give rise to a significant anomaly, while the same conductor striking northwesterly will produce a much weaker VLF anomaly.

For example, Figure 3 shows the impact of local transmitter bearing on conductor detectability. The survey is from an area near Chalk River in Ontario, where fortuitously the azimuths of Cutler and Annapolis (then still in operation) differed by 87 degrees. Figure 3a shows the results of a survey using Cutler as the transmitter, and Figure 3b the results of a survey using Annapolis as the transmitter. Obviously neither survey alone could have identified all the VLF conductors in the area. To be detected in a survey using one transmitter, conductors must be well-coupled to that transmitter, by striking generally towards that transmitter. Conductors striking at right angles to the transmitter direction are poorly coupled to it, and will not be detected in a survey using that transmitter.

One recent problem in performing VLF EM surveys is that the stations do not necessarily transmit at the times desired. Until lately it has been possible to obtain transmitting schedules from the US Naval Observatory in Washington, but this information is no longer available.

Three websites recommended by Geonics that often have information about past and present VLF transmitter activities are:

http://moondog.astro.louisville.edu
http://www.sidstation.loudet.org (SID Monitoring Station A118 Home Page)
http://www.radiotelescopebuilder.com



(a) (b)

Figure 3: VLF EM surveys over an area near Chalk River, Ontario. (a) Survey using NAA Cutler, Maine, due east of site, on survey lines from south to north. (b) Survey using NSS Anapolis, Maryland, due south of site, on survey lines from east to west. (After Dence and Scott, 1979).

There is a solution to the twin problems of poorly-coupled conductors and unpredictable transmission times. It is possible to use a local transmitter, the TX27, which is available from Geonics in Toronto. With a TX27 one can use a long wire grounded at the ends, laid a kilometer away from the survey area, at right angles to the desired conductor orientation. This arrangement will generate a reasonably uniform VLF field over an area of several square kilometres.

3.2 Horizontal Loop EM Surveys

The most common horizontal loop EM (HLEM) system is the Apex MaxMin. In an HLEM survey, there are two coils or loops. One of these loops is the transmitter, which generates an alternating EM field (the primary field) in the ground beneath. If there is a conductor in the ground, then small circulating currents will flow in the conductor, and give rise to their own EM field (the secondary field) at the surface.

The other loop is a receiver. In the MaxMin the receiver loop is wound around two magnetic cores, which form the ‘horns’ on the receiver. Normally the primary field is much stronger than the secondary field. In order to detect the secondary field, a small part of the primary field is sent from the transmitter via cable to the receiver, and is used to cancel the primary field at the receiver, leaving only the secondary field to be detected.

The receiver measures two quantities, the in-phase component and the quadrature component of the secondary field, expressed as a percentage of the primary field at the receiver. There is a phase shift or time delay in generating the secondary field of the conductor. The part of the secondary field that is not delayed is the in-phase component, and the part that is delayed is the quadrature component. Anomalies from good conductors have large in-phase and small quadrature components, while weaker conductors have low in-phase and high quadrature components.

The MaxMin offers a range of frequencies which depends on the model. Typical frequencies increase from 110 hertz (Hz) by factrs of two to 3550 Hz or even higher. In a mineral exploration survey, both lowest and highest frequencies are read, but not all intermediate frequencies are necessarily used.

Interpretation of HLEM data depends on the assumption that the transmitter and receiver coils are coplanar (that is, in the same plane) at a constant separation. In areas of topography, maintaining this assumption is not trivial. The MaxMin system allows the operator to measure the slope to the next station before leaving the current one. If this measurement is done at each station, the computer in the MaxMin receiver will integrate the slope readings to determine the elevation difference between the Tx and Rx coils. It will then display the tilt required for coil coplanarity, and also the amount of extra separation to be used to account for the cable draping over topography.

Figure 4 shows a typical HLEM profile over a conductor. Note that the anomaly amplitude increases with the frequency. In this example, in-phase and quadrature components are roughly equal, implying an intermediate conductor strength. For a discussion of interpretation procedures for HLEM data, see Appendix A.

The separation between the transmitter and receiver coils controls the depth at which conductors can be found. With the MaxMin, spacing between transmitter and receiver can be 25 m, 50 m, 100 m, 200 m or more, depending on the power of the transmitter. A rule of thumb is that good conductors will be found if they are shallower than half the coil separation.

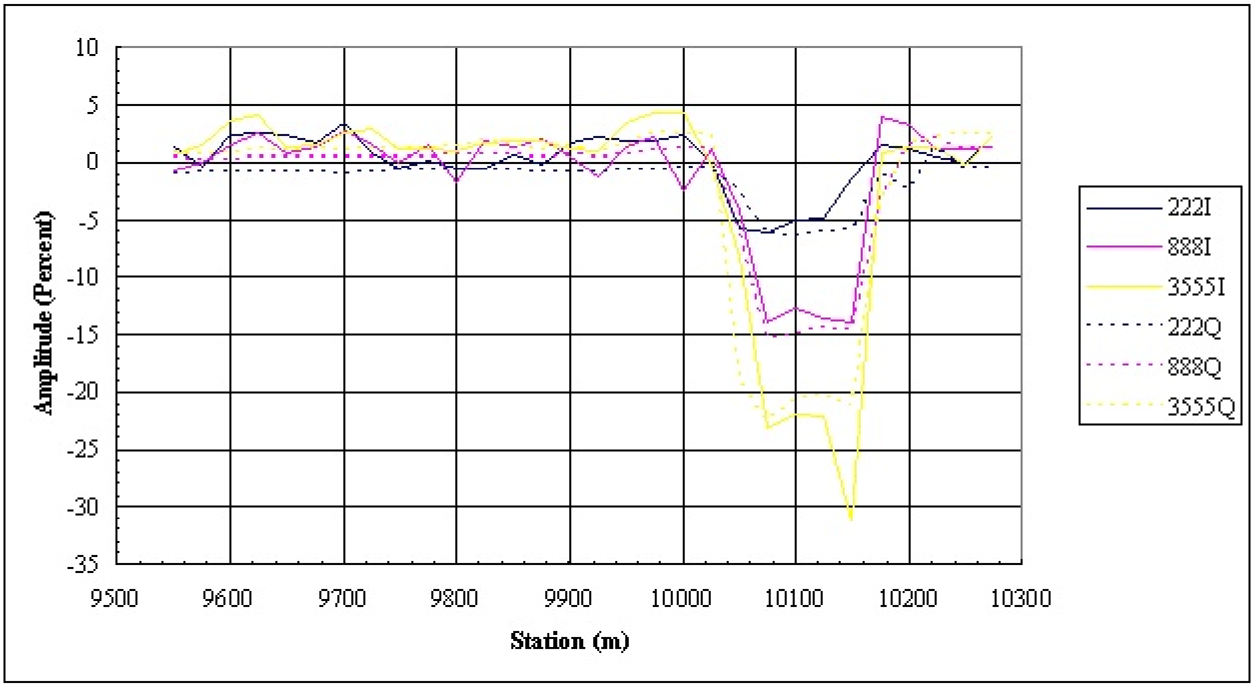


Figure 4: Typical MaxMin profile, 100 m coil separation, 25 m station interval, with three frequencies: 220Hz, 880Hz and 3555Hz. (Unpublished data).

3.3 Time-domain EM Surveys (TEM, TDEM, Geonics EM37, Crone Pulse EM, UTEM, etc)

Transmitters for time-domain EM systems usually use large rectangular loops laid on the ground, typically with sides hundreds of metres long. The waveform of the transmitted signal is designed to produce repeated bursts of energy. Typical waveforms are half-sine (Crone Pulse EM), triangle waveform (UTEM), turn-off ramp function (Protem) and other repetitive signals. Each impulse of energy travels down into the earth, generating time-decaying currents in any conductor within range. By monitoring the signal in the off time after the transmitted pulse, it is possible to detect the fields of these decaying currents.

From the character of the decay fields it is possible to determine depth, shape and approximate conductance of the anomalous bodies. Discussion of interpretation techniques for time domain EM is beyond the scope of this manual. However, an extensive discussion is presented in Chapter 7 of Telford et al. (1990).

In the use of TDEM for exploration for nickel, it has been found that the conductivity of massive nickel sulphide bodies is very high. As a result, some of the energy in the transmitted pulse is absorbed immediately by the sulphides, and the resulting transient is thus reduced. In order to detect such massive nickel sulphides, it is necessary to measure the transmitted pulse, and determine its deformation. At Voisey's Bay, for example, normal off-time transients generated stronger anomalies outside the massively-conductive Ovoid than in it. In order to map the anomaly generated by the Ovoid, it was necessary to measure the field during the on-time and compare the predicted and measured waveforms. UTEM, however, already measures during the transmission time and is not subject to the same limitation.

4.0 RADIOMETRIC SURVEYS.

4.1 Introduction

The discussion in this section is based on Telford et al., (1990). Although there are at least 20 known naturally-occurring radioactive elements, the only ones of significance in geophysical prospecting are uranium (U), thorium (Th) and an isotope of potassium (^{40}K). One other, rubidium, is useful in determining the age of rocks, but the rest are either so weak, so rare, or both, as to be of no significance in applied geophysics. The two elements uranium and thorium are important as sources of fuel for the generation of heat and power in nuclear reactors. Large parts of the world have been surveyed on the ground and particularly from the air, in the search for uranium.

4.2 Radioactive Decay

The radiation from naturally-occurring radioactive elements consists of three distinct types: alpha (α), beta (β) and gamma (γ) radiation. All three types of rays can produce scintillations or phosphorescence in certain minerals and chemical compounds. In addition they ionize gas, making it electrically conducting, and they affect photographic emulsions in much the same way as light or X-rays.

The three rays have very different penetrating powers. Thus, α rays are easily stopped by a piece of paper, and β rays are easily stopped by a few millimetres of aluminum. The attenuation of γ rays requires several centimetres of lead. The range of α and β rays in overburden or rock is virtually zero, while γ radiation can pass through from 50 to 75 mm of rock.

4.3 Instruments for detecting radioactivity

There are two principal instruments, the Geiger counter, and the scintillometer. The γ-ray spectrometer is an extension of the scintillometer.

The Geiger-Müller counter is a simple device that responds primarily to β radiation. The detector is a thin-walled cylindrical tube with a very thin mica window in the end to permit the entry of the β radiation. The tube consists of an axial anode wire and a coaxial cathode cylinder polarised by several hundred volts, and is filled with an inert gas such as argon at low pressure. Radiation entering the tube ionizes gas molecules, and the positive ions and electrons are accelerated to the cathode and anode respectively. These charges also ionize other gas atoms en route, so that the original entering ray produces an avalanche of charged ions that cause a substantial discharge pulse across the anode resistor. This pulse is amplified to produce a click in a speaker or headphones. Additionally, successive pulses charge a capacitor and the charge leaks off through a microammeter, which registers a current proportional to the strength of the charge entering the condenser.

The prospecting Geiger counter is simple and cheap. However it has little else to recommend it. It must be held close to the outcrop to detect the β rays, and is thus a tool of limited application, seldom used in modern prospecting.

The scintillometer works by counting scintillations produced in a detector by γ radiation. The most efficient detector is made by growing natural crystals of sodium iodide (NaI), treated with thallium (Tl). The NaI is transparent to its own fluorescent emission, and all faces but one are coated with light-reflecting material. If the crystal is large enough, then the conversion efficiency for natural γ radiation is close to 100%.

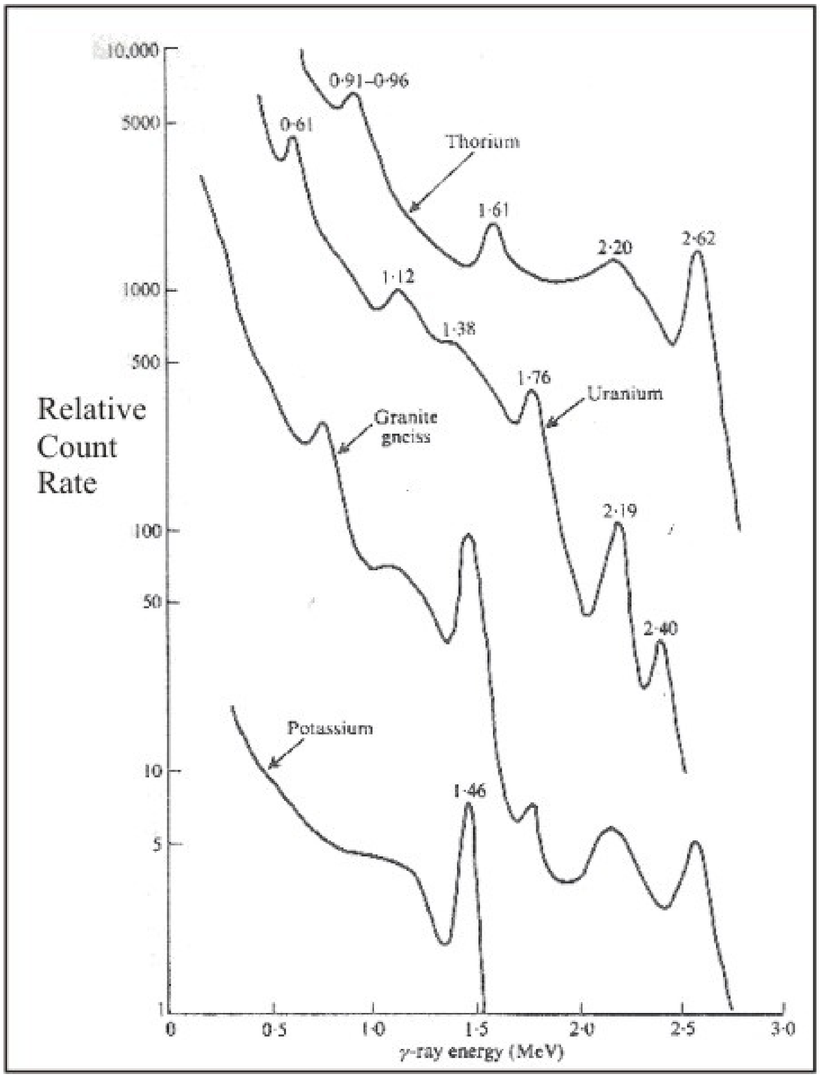


Figure 5: Typical γ ray spectra for Th, U, a granite gneiss and ^{40}K, after Telford et al, 1990, p624.

Light generated in the NaI crystal by γ radiation is amplified through a photomultiplier tube; the output current is applied across a resistor to produce a voltage pulse which is amplified and integrated as in the Geiger counter. The great advantage of this instrument is in the efficiency of γ radiation detection, but it will also detect β radiation. The instrument is somewhat larger and heavier than the Geiger counter, and also more expensive.

A logical extension of the scintillometer is a spectrometer that distinguishes characteristic γ rays from ^{40}K, U and Th. Such instruments are widely used in airborne surveys and some portable units are also available. The intensity of the light pulse, and thus the amplitude of the output voltage pulse, are proportional to the initial γ ray energy. γ radiation from different sources has different energy levels. The output spectrum of the spectrometer is complex, and has peaks which are associated with the energy of radiation from different sources. Figure 5 shows typical pulse-height spectra for different geological sources. To obtain a spectrum, the relative count rate of incoming γ radiation, is plotted against the energy measured in millions of electron volts (MeV). In a field situation, some γ rays lose energy by scattering in escaping from the source, and also during passage through the air to the crystal. In addition, some of the γ rays lose only part of their energy within the NaI crystal. This, coupled with the fact that the U and Th series emit numerous γ rays over a wide energy range, results in a complex pulse-height spectrum, as shown in Figure 5.

All four curves in Figure 5 have characteristic peaks. The pure potassium sample produces a relatively simple curve having only the ^{40}K peak at 1.46 MeV. Thorium is characterised by the strong 2.62 MeV peak of ^{208}Tl (thallium), a constituent of the Th decay series. The uranium spectrum is most complex, although the peak at 1.76 MeV (million electron volts) is reasonably distinctive. Potassium and thorium are clearly evident in the spectrum for the granite gneiss. A small fraction of uranium is also indicated.

A prospecting spectrometer should be able to isolate the K, U and Th peaks at 1.46, 1.76 and 2.62 MeV. It should use at least three separate channels, each sensitive to one of the energy peaks. Generally the channel centres and widths are adjustable.

Normally ground radiometric surveys are used as followup after airborne surveys. Airborne spectrometers can use much larger crystals, measure more channels and correct both for drift and background radiation. Ground surveys can then help to isolate the sources of the airborne anomalies. The interpretation of ground survey results are mainly qualitative. This is partly due to the extremely small depth of penetration possible with the method. It is also the result of the inherent complexity of the γ ray spectra.

For many years, radiometric surveys, both airborne and ground, have been primary tools in the exploration for uranium. In addition, thorium often is associated with Rare Earth Element (REE) minerals, Thus, airborne gamma-ray spectrometric surveys are a key tool in the exploration for REEs. Similarly, K response can be used to identify zones of potassic alteration, and thus contribute to geological mapping based on geophysical survey results.

5.0 INDUCED POLARISATION SURVEYS

5.1 Introduction

Electric current in rocks is usually carried by the movement of charged ions in the water which fills the pores. If a pore is blocked by a metallic substance such as a metallic sulphide, then the current flow must transfer to the sulphide, where it is carried by the movement of electrons. As a result there is a small delay at the interface while the charged ion loses its charge either by taking on an electron from the metal or by releasing one into the metal. Because of this delay, a concentration of ions, waiting to lose or gain an electron, develops at the interface during the flow of current. When the current is interrupted, the ion concentration disperses away from the interface, and this drift is seen as a transient voltage along the pore. Thus induced polarisation (IP) arises at the interfaces between pores and metallic sulphides. The strength of an IP response is governed not by the volume of sulphides in the rock, but by the amount of sulphide surface exposed to the pore water. A certain volume of fine-grained sulphide will generate a larger IP effect than the same volume of coarse-grained sulphide. Consequently the strength of an IP anomaly is not necessarily an indication of the volume of metallic sulphides present. Because a rock showing IP effects can store electric energy, even though for short times, it is said to be chargeable, and the chargeability of a rock is a measure of the strength of its IP effect. See Appendix B, Section B2, for a definition of chargeability.

A few other minerals can also generate IP effects. For instance, graphite carries current by the drift of electrons, and thus gives rise to an IP effect. Similarly some magnetites, and even some serpentinites have been known to generate IP responses. Note that non-metallic sulphides such as sphalerite do not give rise to an IP effect.

Appendix B amplifies the description of the development of IP effects. It also describes the modes of measuring IP effects, and the units of the measurements.

5.2 IP measurements: Electrode arrays

Making IP measurements involves use of a transmitter to generate a current, and a receiver to measure the resulting voltages. The transmitted current is injected through two transmitter electrodes, and the received voltages are measured through two or more receiver electrodes.

Transmitter electrodes are usually metal rods, although in some cases of high-resistivity areas, they may be metal screen or even aluminum foil buried in pits moistened with water. Receiver electrodes are often metal rods. For greater stability and lower noise levels, porous pots should be used. A porous pot is a small container with a porous bottom such as an unglazed ceramic disc, filled with a saturated solution of copper sulphate, with a copper electrode inserted in the pot. Contact with the earth is through the fluid in the porous bottom of the pot, and the connection to the receiver is through the copper electrode. Porous pots give more stable measurements, particularly where the acidity of the ground varies along the measured profile (e.g. when the line varies from bog to dry earth).

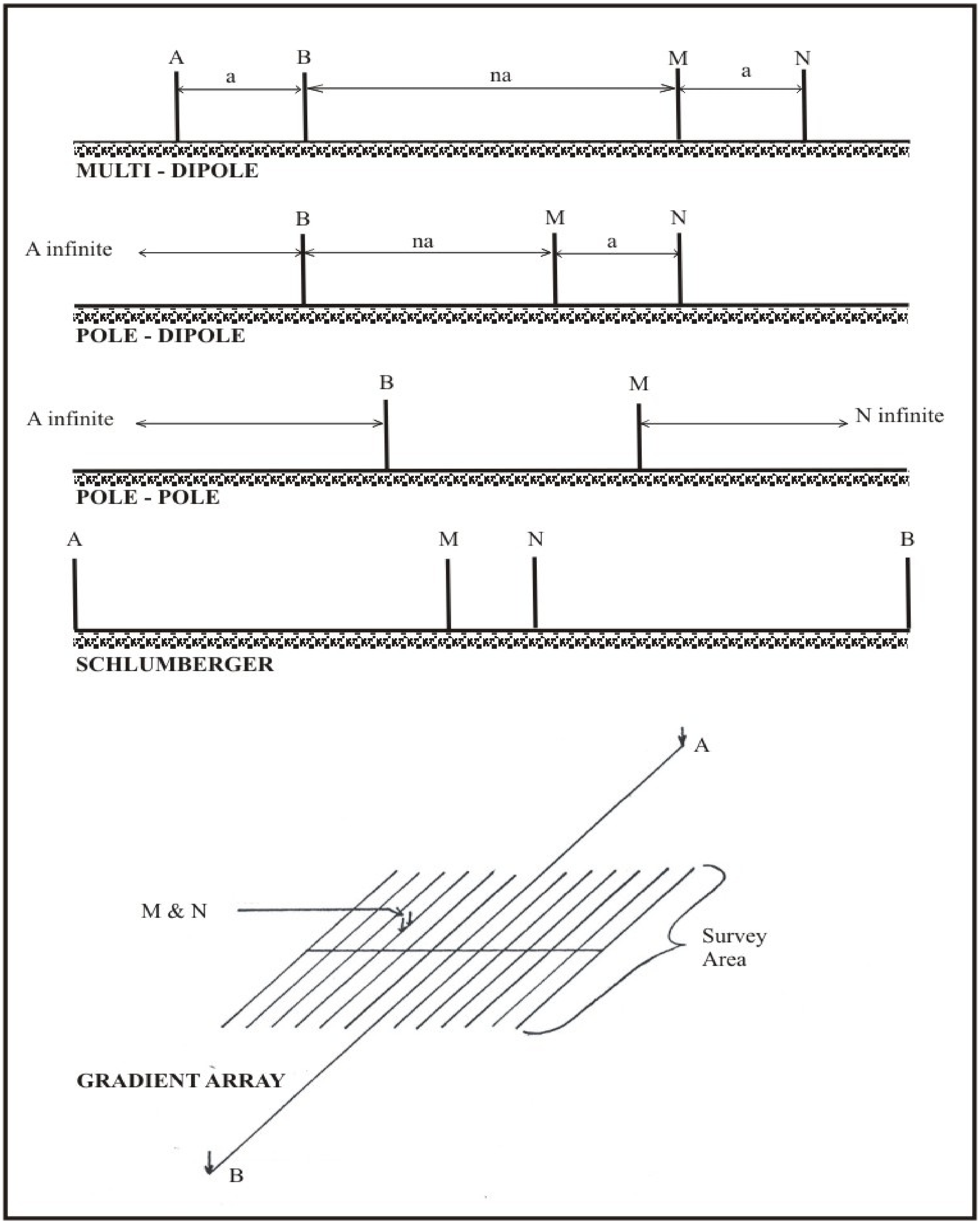


Figure 6: Electrode arrays used for IP surveys. A and B are transmitter electrodes, M and N are receiver electrodes. For the gradient array, co-ordinates are $A(x_A, y_A)$, $B(x_B, y_B)$, $M(x_M, y_M)$ and $N(x_N, y_N)$, where the survey lines run parallel to AB. In multi-dipole and pole-dipole arrays, the dipoles are separated by distances which are multiples (n = 1, 2, 3, etc.) of the dipole size a..

IP measurement electrode arrays include multi-dipole, pole-dipole, pole-pole and Schlumberger and gradient. Figure 6 shows these arrays, and Section B.5 in Appendix B gives the formulae for apparent resistivity for each of the arrays.

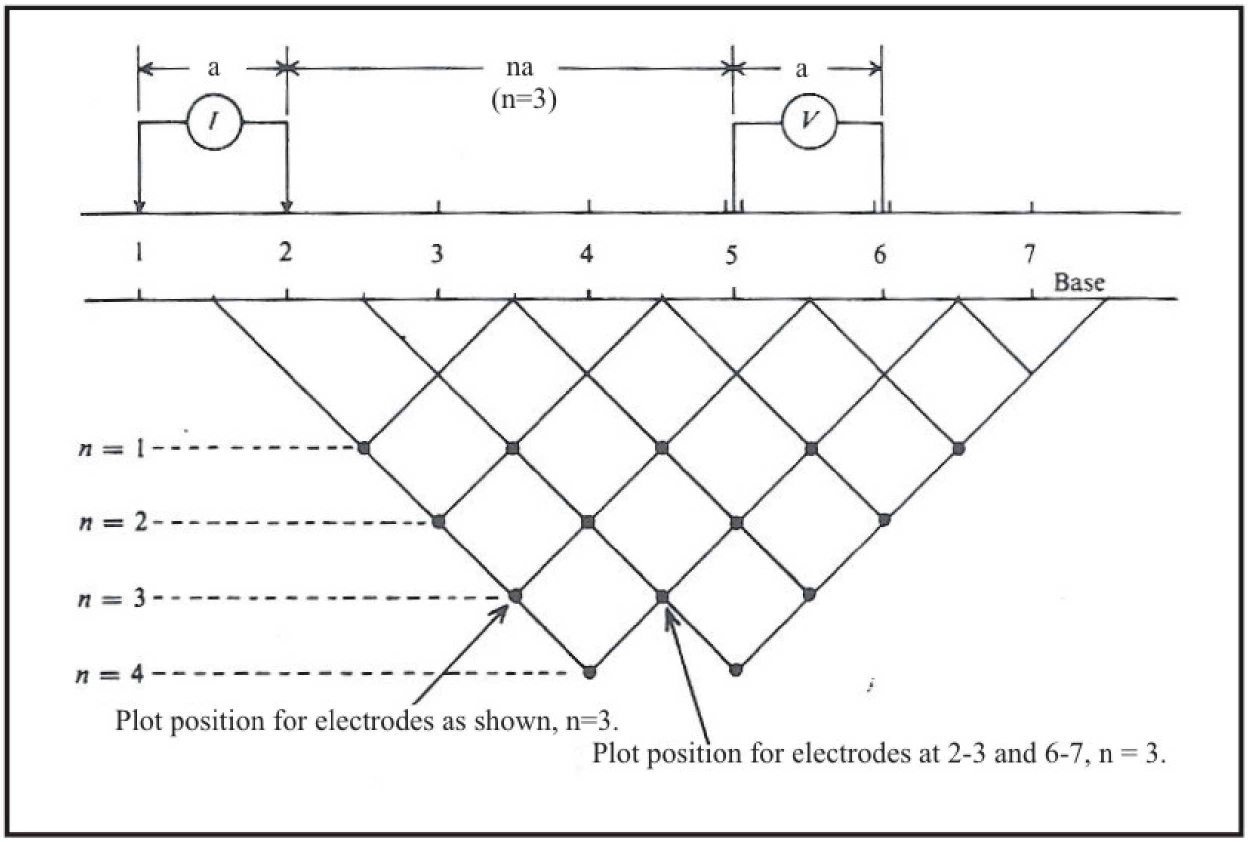


Figure 7:: Method of plotting to generate a pseudosection of IP or resistivity data.

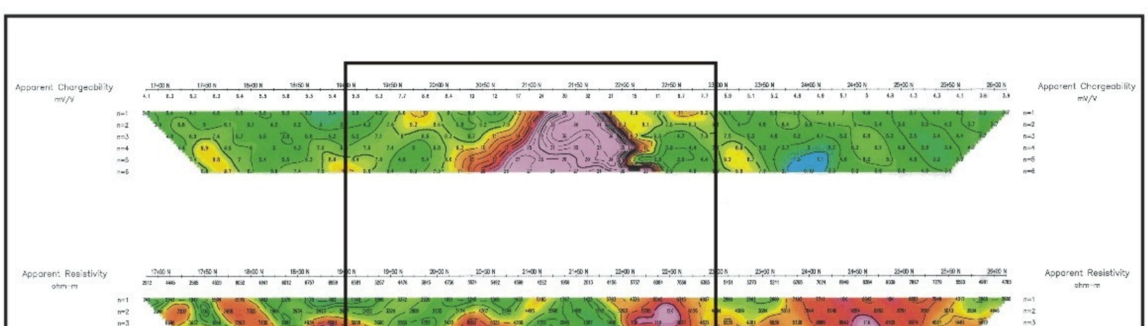


Figure 8: Typical pseudosection, time-domain IP/resistivity, a = 25m, n = 1 to 6, apparent chargeability above, apparent resistivity below. Rectangle encloses data inverted for Fig. 9.

5.3 Multi-dipole Surveys

The most common arrays now are multi-dipole and pole-dipole. Data acquired by these arrays are usually portrayed in pseudosections, which demonstrate the relative depth dependence of the measurements in a schematic way. Figure 7 shows how a pseudosection is plotted from multi-dipole data. For pole-dipole plots the diagonal for the transmitter is drawn from the one transmitter electrode instead of from the centre of the dipole.

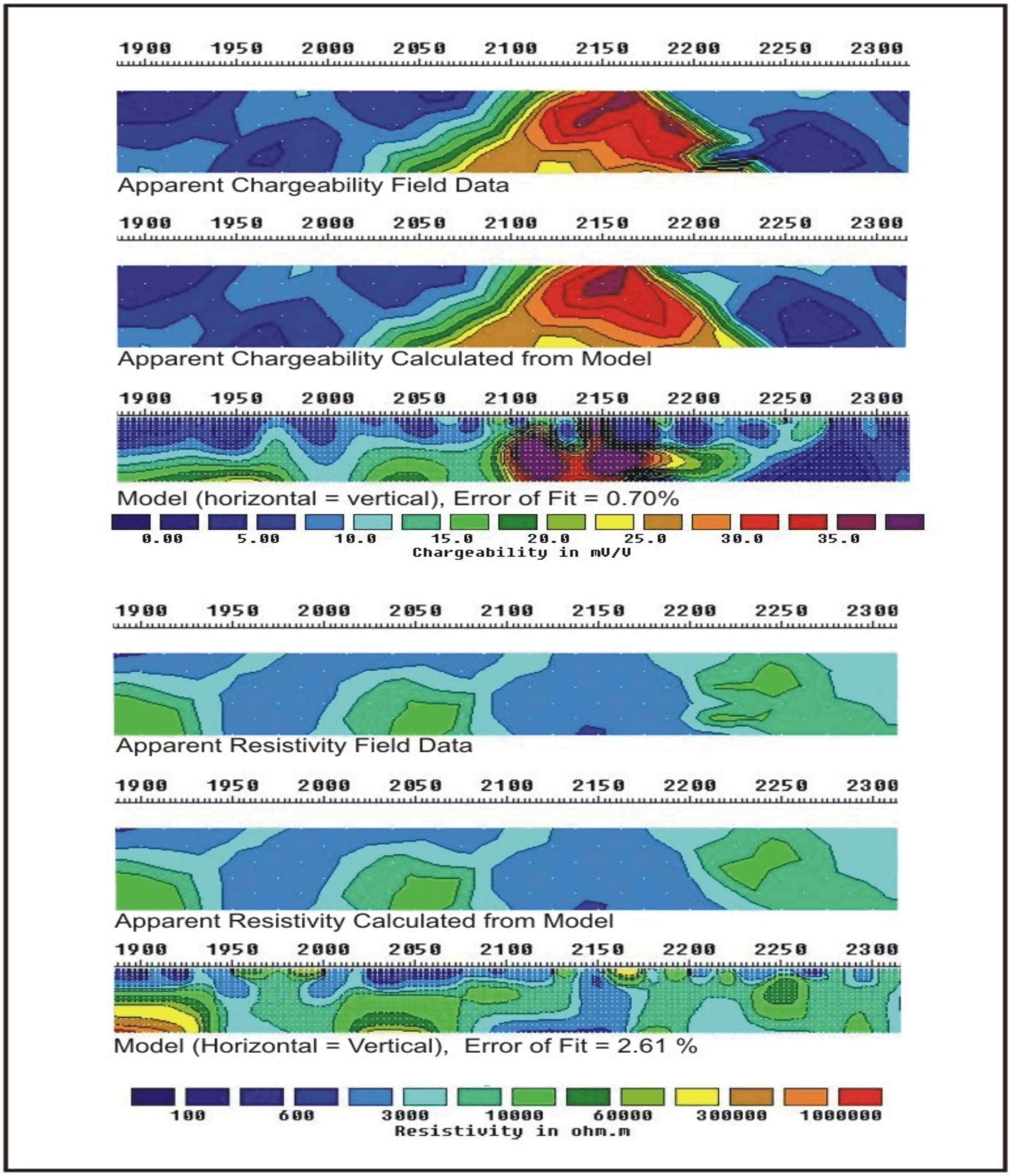


Figure 9: Results of inversion of central part of pseudosection data of Figure 8. . Chargeability values in millivolts per volt (mV/V), resistivity values in ohm-metres (Ω-m). White dots in models show centres of cells used for inversion. (Unpublished data)

However, it should be remembered that a pseudosection is not a geological section. Even simple bodies such as dykes can generate complex anomaly patterns. If the pseudosection is for a multi-dipole array, then an anomaly over any body will be the same regardless of which side the transmitter dipole is of the receiver array. However, with the pole-dipole array, the shape of the anomaly depends on which side the transmitter electrode is of the receiver array.

Figure 8 shows a typical pseudosection of multi-dipole IP data, with chargeability data contoured linearly at intervals of one mV/V, and resistivity data contoured logarithmically at intervals of 1, 3, 5, 10, 30 etc. It is increasingly frequent not just to present the pseudosections of chargeability and resistivity, but also to invert the data and show the inverted sections as well. Figure 9 shows the inversion results for the area within the rectangle in the pseudosection of Figure 8.

The inversion process predicts the field measurements for the model, and compares them to the observed field data. The error of the fit is calculated as the root of the average of the square of the difference between observed value and value predicted from the model. An estimate of the reliability of the model can be made by looking at the fitting errors. In this example the resistivity fit is 2.61 percent. An error this small is quite acceptable. Similarly the chargeability fit is 0.70 per cent, which is remarkably good as well.

5.4 Gradient Surveys

To carry out a gradient survey, the two transmitter electrodes are established at points outside the survey area, along the projection of the centre survey line. The transmitter separation can be up to a few kilometres, provided enough current is transmitted to produce a large enough signal within the survey area. The receiver dipole can be whatever is appropriate for resolving the expected anomalies. Typical receiver dipole separations range from 25 to 100 metres.

The gradient array yields only one set of resistivity and IP measurements for each point. Consequently gradient data are usually portrayed as maps showing the spatial variation of the data. A gradient survey is a useful tool for covering a large area in which the positions of anomalies are not well known. Coverage with a gradient survey is much more rapid than with a pole-dipole or multi-dipole array, because only the receiver electrodes must be moved. Figure 10 shows the results of a gradient survey over an area 400 m east-west and 600 m north-south. Readings were taken with 25 metre dipoles, on lines 25 m apart. Layout of transmitter wire, survey of 6100 metres and pickup of transmitter wire required 14 survey hours.

Once a gradient survey has been carried out, and approximate anomaly positions identified, then suitable lines can be selected to carry out detailed surveys with multi-dipole or pole-dipole arrays. The detailed surveys then yield better definition of the anomalies.

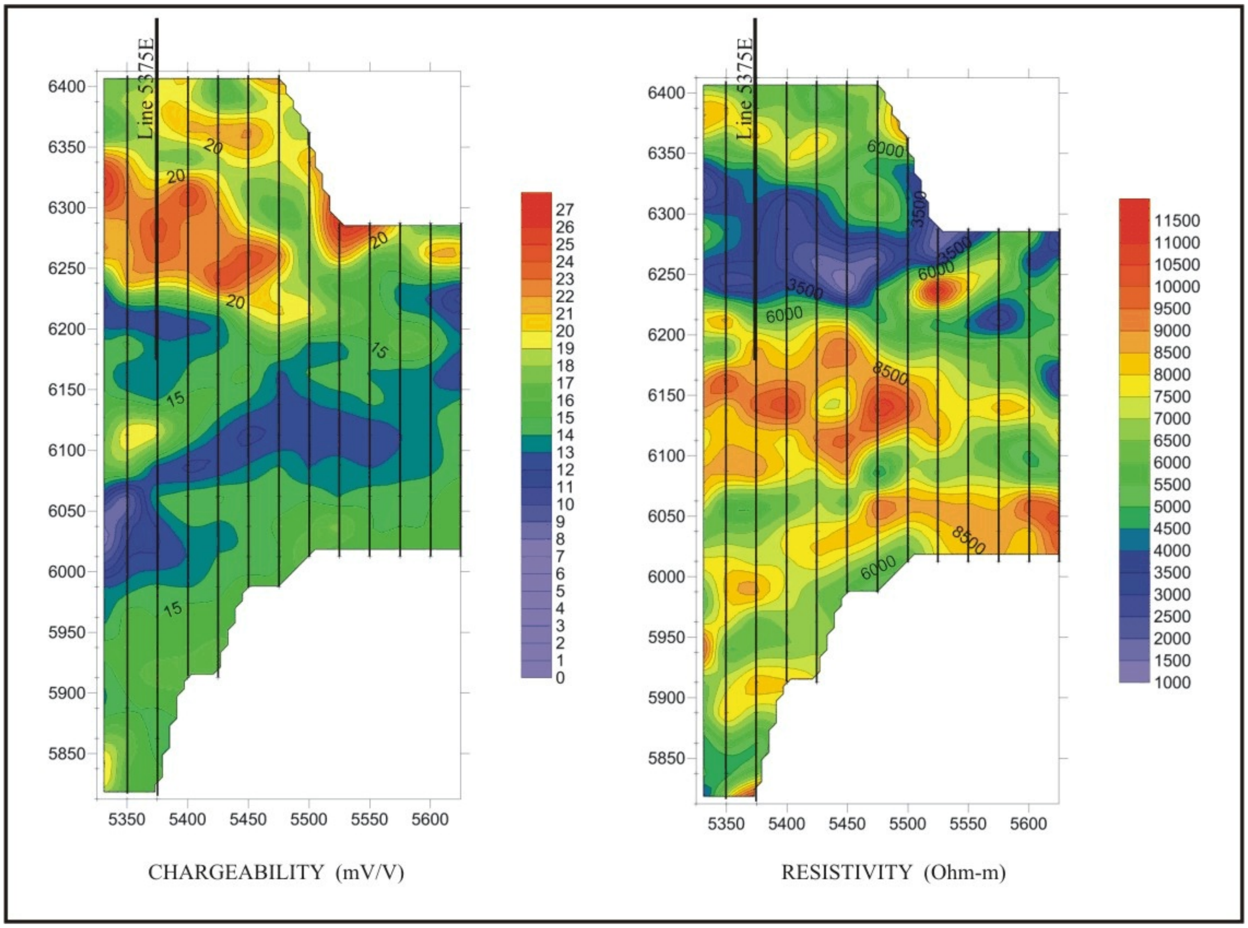


Figure 10: Gradient IP and Resistivity survey near Baie Verte. Data from NDNR 012H/16/1710.

6.0 GRAVITY SURVEYS

6.1 Introduction

The earth exerts an attraction on any mass that is within its gravity field. The attraction varies inversely with the square of the distance between the centre of the earth and the centre of the mass. We recognise this attraction as the “weight” of the mass. In a local situation, a given mass will weigh more if the earth is more dense under it. This relationship is the basis of the gravity method. In gravity surveys the object is to locate, from irregularities in the earth’s gravity field, small variations in density of rocks below the earth’s surface.

A gravimeter, used to measure the earth’s gravitational field, is a very sensitive weigh scale. It has a small reference mass mounted on a hinged arm. The arm is held horizontally in place by a spring which is adjustable. The spring is adjusted to move the mass to be in alignment with a reference mark. The adjustment to the spring is calibrated to read the local value of the earth’s gravity field. The gravimeter is a very delicate instrument which is sensitive to better than one part in ten million of the earth’s gravity field. Gravity measurements can outline local anomalies in the earth’s gravity field which are generated by local high-density bodies in the earth. The earth’s field is about 980 gals (the gal is named after Galileo, who first experimented with the earth’s gravity field), but significant gravity anomalies may be as small as 0.1 milligal, where one milligal (mgal) is about one millionth of the earth’s gravity, or one thousandth of a gal.

6.2 Corrections needed in gravity surveys.

Because the “weight” of a mass varies with the square of the distance to the centre of the earth, small variations in elevation of the survey station can greatly affect the measured value of gravity. As a result, it is necessary to correct for changes in location and elevation of the survey stations.

The first correction is known as the latitude correction. Because the earth rotates, it bulges at the equator, and flattens at the poles. Thus gravity is stronger at the poles than at the equator. In Newfoundland, the latitude correction is about 0.8066 mgal per north-south kilometre, and is subtracted as the survey station moves northward.

The second correction is the free-air correction. In Newfoundland, the earth’s field varies by about 0.197 mgal per vertical metre. If a station is above the reference datum, the correction is added to the reading; if the station is below the reference datum, then the correction is subtracted from the reading. To calculate the latitude and free-air corrections to within 0.01 mgal, the north-south location must be known within about 13 metres, and the elevation to about 3 cm or better.

The third correction is known as the Bouguer correction. It accounts for the attraction of material lying between the station and the reference datum. The Bouguer correction is {0.04192 x (density of the intervening material)} per vertical metre. If the density of the intervening material is the average rock density of 2.67 grams per cubic centimetre (g/cc) then the Bouguer correction is 0.112 mgal/metre. It is applied in the opposite sense to the free air correction, and is subtracted from the field reading when the station is above the reference datum, and added when it is below.

The free-air and Bouguer corrections are often combined into the elevation correction. For a density of 2.67 g/cc, the combined correction is 0.19673 mgal/metre.

In areas of severe topography, a further correction, the terrain correction, is also necessary. When a mountain sits above a survey point, the upward attraction of the part of the mountain above the station reduces the reading. Terrain corrections are complex, and, in pre-computer days, involved enormous amounts of hand calculation. Now, however, digital processing uses digital terrain models to calculate terrain corrections. Surveys in areas of deep lakes must also allow for the reduction in the field readings because of the lower density of water. This correction is also performed digitally, provided care is taken to measure the distribution of water depths.

Carrying out gravity surveys is not difficult, particularly because modern gravimeters are automated in their measurement process. However, correcting raw gravity readings to produce meaningful values involves precise land survey. In open terrain, the usual approach to obtaining precise corrections is to use a precise Global Positioning System (GPS) operated in a real-time kinetic mode (RTK). Note that a hand-held GPS does not offer adequate precision. The precise GPS system uses a base station which tracks temporal changes in its position, and uses the changes to determine positional corrections. In RTK mode, these corrections are broadcast in real time. The survey station (the rover) is set at each survey point, and uses the broadcast corrections to produce a precise position for the point. A GPS survey in RTK mode is capable of horizontal precision of less than a centimetre, with vertical precision about twice the horizontal precision. However, if the sky is obscured by heavy vegetation, the GPS system may not see enough satellites to be able to determine its position. In such areas, recourse must be had to traditional land survey techniques to fill in between GPS points.

In order to preserve the elevation precision needed for adequate corrections, when the gravimeter is set up at each station, the height of the instrument (HI) above the reference elevation point is recorded. The value of HI is added to the ground elevation to obtain the elevation used in calculating corrections.

In performing gravity surveys, it is normal practice to establish a gravity base station, which is read at least at the beginning and end of the day, and sometimes more frequently. If a survey is to be extended at a later date, using the same base station ensures that the two data sets will be seamless.

6.3 Interpreting gravity survey data.

The final gravity data are usually plotted and contoured in the same manner as magnetic data. Figure 11a shows a contour map of gravity data. The data set has had latitude, free air and Bouguer corrections. Often gravity results such as these will show a gradual trend across the

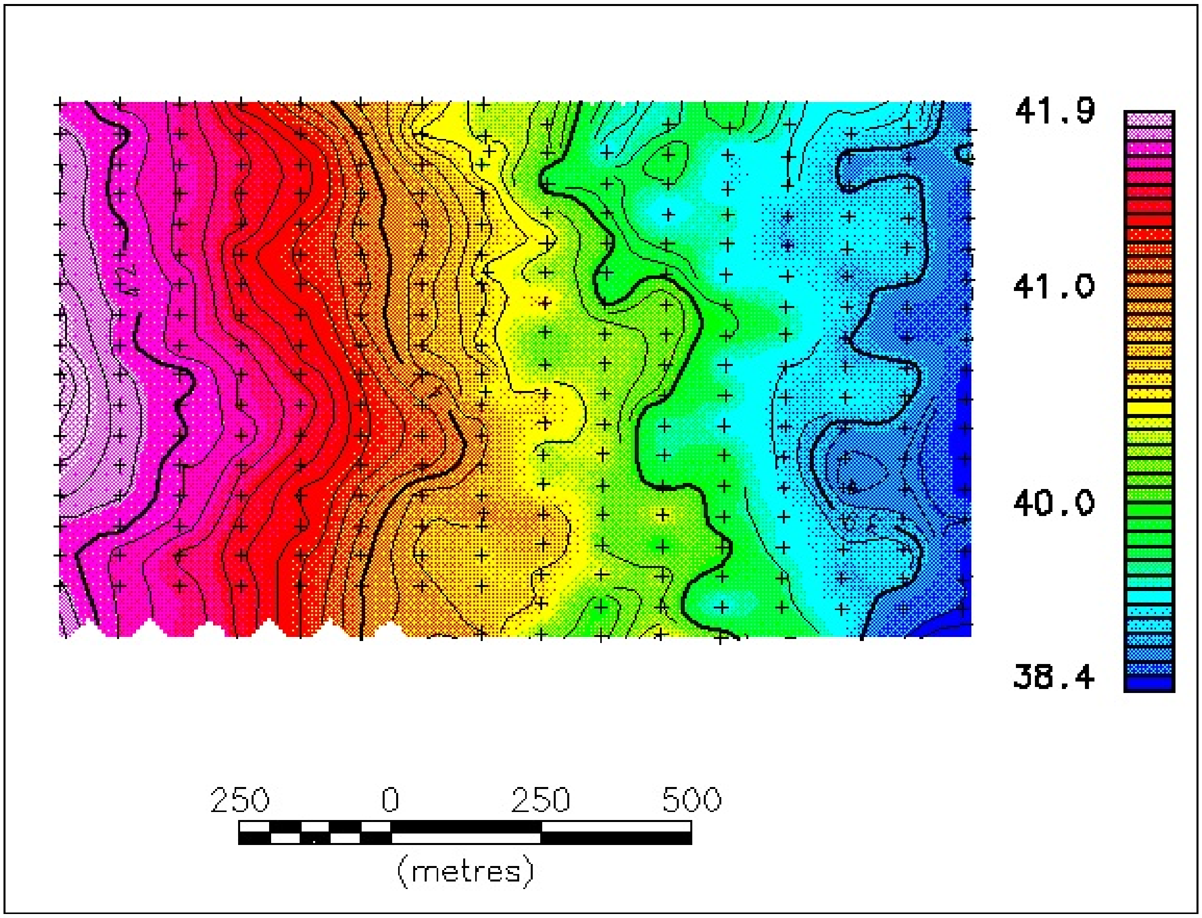


Figure 11a: Gravity map showing contoured data with latitude, free air and Bouguer corrections. Contour interval 0.1 mgal (unpublished data).

area, which will distort local anomalies. The gradual trend in gravity values can arise from deep–seated geological variations, and can be removed from the data. The high gravity at the west end and the low at the east end both disappear when a linear regional trend is removed (Figure 11b). Normally such trend removal is carried out by digital processing.

Gravity surveys are sometimes used as a follow-up to EM surveys, when there is a possibility that the conductor seen by the EM survey could be graphite. Sulphide bodies are usually more dense than the host rocks, and thus yield local high-gravity anomalies. Graphite, on the other hand, is less dense than sulphides, and close in density to typical host rocks. Graphite conductors do not usually give rise to local high-gravity anomalies. Gravity is also sensitive to the presence of sphalerite, which is usually more dense than its host.

If the survey area is underlain by rocks of different densities, then gravity can be used to map the distribution of each rock type.

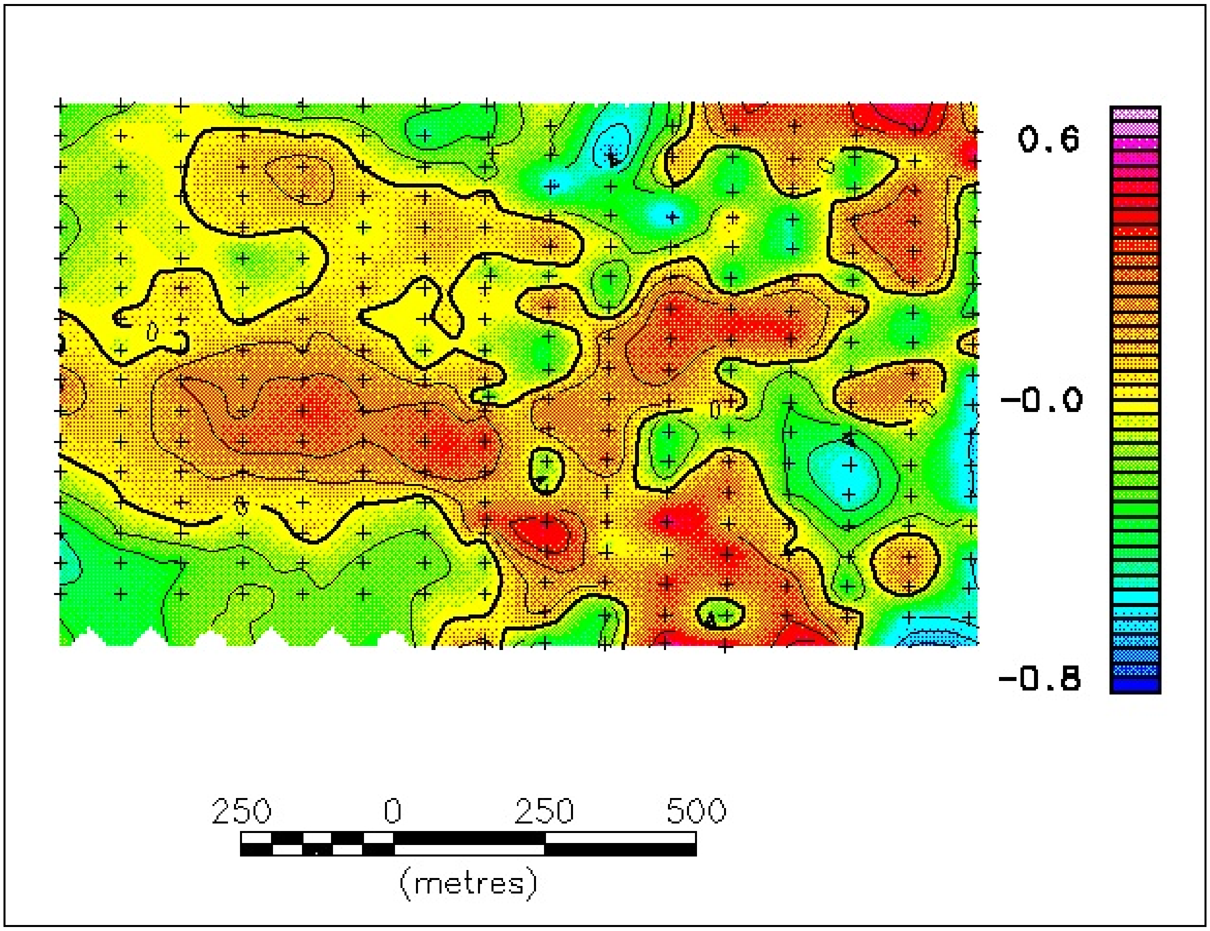


Figure 11b: Gravity data of Figure 11a, with linear regional trend removed from data. Contour interval 0.1 mgal (unpublished data).

Because the strength of a gravity anomaly falls off with the square of its depth, gravity surveys cannot identify bodies buried deeper than a few hundred metres, no matter how high the density contrast.

Gravity has been very useful for identifying low-density accumulations of salt and potash. Near Stephenville, there are large accumulations of salt and potash in the nearby Carboniferous Bay St. George subbasin. The same rocks also host gypsum and anhydrite (higher density materials), which are often in close spatial association, which can make the interpretation of the data a bit tricky.

7.0 SEISMIC SURVEYS

Although seismic surveys are mainly employed for hydrocarbon exploration, there have been some recent applications to mineral exploration. Massive sulphide bodies have higher seismic velocities and densities than their host rocks. If the bodies are big enough, they can give rise to strong reflections. Surveys can be carried out in either two or three dimensions. Two-dimensional (2D) surveys yield single profiles, but are susceptible to errors generated by reflectors that are not directly under the survey line, but off to one side. Three-dimensional (3D) surveys involve observations on a full grid, with sources and receivers on various lines. 3D surveys are an order of magnitude more expensive than 2D ones. However, any major seismic survey is extremely expensive and extremely complex. Treatment of them is beyond the scope of this manual.

Occasionally, shallow seismic refraction or reflection surveys are used to map the thickness of overburden, or to help in the exploration for and evaluation of granular mineral deposits. For example, shallow seismic surveys were used to map the overburden thickness over the Voisey’s Bay Ovoid deposit. The results provided input into the design of the open pit.

8.0 MULTIPLE METHODS AND SURVEY DESIGN

In many cases, more than one geophysical method can be used to help understand the geological context of the geophysical survey results. For example, an electromagnetic survey may have identified a conductor, but graphite is known to occur in the area. A small gravity survey can be carried out to see if the EM conductor is associated with higher density material. If there is no gravity anomaly then the conductor is not likely to arise from metallic sulphides.

The gravity method (including airborne gravity) is currently being used extensively in the Labrador Trough in the search for iron ore (Kilfoil, pers. comm.). It is not difficult to identify surface outcrop of iron formations in this area with a magnetometer, However, these rocks are so magnetic (and variable) that much of the magnetic response (variability) is due to exposed or subcropping ore, and this strong response masks any response of ore at depth. Thus there is no magnetic way to determine how extensive the ore is. The rocks, at least in the southern Trough are also highly deformed and folded, so any realistic modelling of the magnetic data would be next to impossible. The gravity method is useful for distinguishing the wheat from the chaff; it highlights the areas having significant subsurface accumulations. In the northern Trough, near Schefferville, parts of these iron formations have been leached & high-graded to "direct shipping ore", in which the magnetite has been completely replaced by hematite. The geophysical signature here is low magnetic readings associated with high gravity, lying along trend with very magnetic, unaltered iron formations.

Within the limited size of this document, it has not been possible to include enough examples of field survey results. An excellent publication full of field examples is Practical Geophysics III, recently published by the Northwest Mining Association.

When the use of geophysics is being considered, it is useful to go back to Table 1 (Page 1) to help to define a geophysical approach. The first step is to identify the purpose of carrying out the survey. Is it to map geology, trace structure, and/or to search for concentrations of particular minerals? With the purpose identified, the next step is to ask what geophysical properties can be expected for the country rocks, the host rocks and the target body, and what contrasts in the properties will be needed to fulfill the purpose of the survey. At this point, it should be evident which geophysical method or methods offer the best chance of meeting the purpose of the survey.

If there is any doubt whether the best survey design has been achieved, it would be wise to obtain qualified advice about the survey design, to be sure that the resulting survey generates the best possible outcome. An excellent source of qualified advice is the Matty Mitchell Room in the Newfoundland Department of Natural Resources, whose staff can obtain advice from many disinterested experts within the Department.

Before beginning a survey, it is important to ensure that the survey grid is laid out to enhance the response from the bodies of interest. Normally the grid is oriented so that the survey lines cross the strike of the expected target at right angles. If the results of an airborne geophysical survey are available, then the grid can be oriented to cross interesting airborne anomalies at right angles.

Some targets, such as gold in quartz, may not have a geophysical property to help in exploration. However, it may be possible to find an indirect property, such as a property of the host rocks, which will help in exploration. For example, the quartz rocks hosting gold may have very high resistivity values. In this case a resistivity/IP survey may help in mapping the distribution of high-resistivity quartz, which would help to concentrate exploration activity on suitable host rocks.

As time passes, the fashion in mineral exploration can change, and targets which have been dismissed in the past may become attractive in the future. An example is graphite, which for many years has been considered a misleading pest in mineral exploration. Recent discovery of high-grade graphite deposits has turned graphite from hateful villain to desirable target.

A valuable and largely under-used resource in mineral exploration with geophysics is the assessment files maintained by the Department of Natural Resources. Unless the area being considered is truly unexplored, there will almost certainly be geophysical survey reports in the assessment files for the area under consideration. At the least, these reports can indicate what sorts of results can be expected from specific geophysical techniques. At best, the reported results may contain anomalies associated with new or different targets, often neglected in the original report. Before designing a survey on the basis of expected target responses, it very useful to see what results have been obtained in the past.

9.0 REFERENCES

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Appendix A

Interpretation of MaxMin Horizontal Loop Data

Interpretation of MaxMin Horizontal Loop Profiles.

To demonstrate the process of interpreting horizontal-loom EM profiles, the MaxMin profile shown in Figure 4, page 8 of the main text, will be used. It is assumed that the profile was obtained with co-planar coils, and with the coil separation adjusted to allow for topographic variations.

Figure A1 (after Telford et al.,1990) shows some MaxMin profiles for a perfect conductor, derived from model studies. The upper set of curves is for profiles which cross the conductor strike at right angles. The lower set is for profiles which cross the conductor strike at 60 degrees. If the two positive peaks are the same magnitude then the conductor is vertical. For dipping conductors the higher positive peak lies over the down-dip side. Figure A2 shows Argand diagrams or characteristic curves for three different conductor dips.

If the conductor is narrow, then the distance between the two zero points equals the coil separation. The width of a wide conductor can be estimated by the difference between the coil separation and the distance between the two zero points in the profile.

The first step in the interpretation is to estimate the width of the conductor by measuring the distance between the two zero points. The profile in Figure 4 appears to be narrow, because the zero-point separation is equal to the coil separation. The second step is to estimate dip from the asymmetry of the profile, by comparison with the profiles in Figure A1. From profiles on adjacent lines, it is known that the profile line in Figure 4 crossed the conductor strike at right angles. Comparing the profile shape in Figure 3 with the curves in the upper part of Figure A1 suggests that the conductor dips at about 60 degrees to the east. Figure A-2 shows three Argand diagrams for conductor dips of 90^0, 60^0 and 30^0. Choose the second diagram as representing the dip closest to that estimated. From the HLEM profile of Figure 4, pick the values of in-phase and quadrature midway between the zero points. Table A1 shows the values picked for the three frequencies.

On the Argand diagram, draw a vertical line at the value of the in-phase component, and a horizontal line at the value of the quadrature component for each frequency. The intersection of the two lines defines a point on the Argand diagram. The dashed contours on the Argand diagram express depth to the conductor as a multiple of the coil separation (i.e. z/l, where z is the depth and l is the coil separation). Estimate the value of z/l for the point, and multiply the value by the coil separation (100 m for the profile in Figure 4) to get the depth to the conductor. Table A1 shows the values of z/l and z for the three frequencies. The radial lines on the Argand diagram are in terms of the response parameter $p = \mu\omega\sigma lt$, where

p is the response parameter,
mu (μ) is the magnetic permeability of the host (usually assumed to be $4\pi \times 10^{-7}$),
omega (ω) is the angular frequency of the profile being interpreted ($\omega = 2\pi f$, where f is the frequency of the profile measurement in hertz),
sigma (σ) is the conductivity of the conductor in siemens per metre,
l is the separation between the coils in metres and
t is the thickness of the conductor in metres.

For thin conductors the conductance, or strength of the conductor, is given by the conductivity-thickness product σt, in units of siemens.

Table A1: Values for the profiles of Figure 3 for three frequencies.

Frequency f (Hz)	ω = 2πf	In-phase (%)	Quadrature (%)	Colour	z/l	Depth z (metres)	p	Conductivity-thickness σt (siemens)
222	1 395	-5	-7	Blue	.25	25	3.7	21
888	5 580	-13	-15	Red	.19	19	4.9	7
3555	22 337	-22	-20	Yellow	.12	12	7.0	2.5

As the frequency increases, the profile amplitude of the anomaly also increases. In addition, the depth to the conductor appears to decrease, probably because the conductor is weak near the surface, and improves in strength with depth. The values of σt decrease with increasing frequency, which is unusual. More often the value of σt increases with increasing frequency.

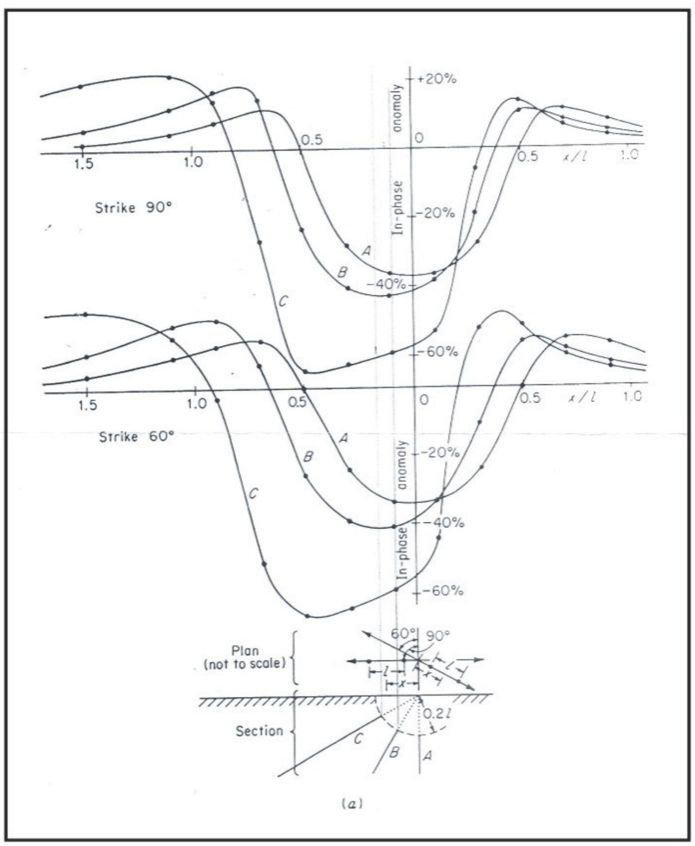

Figure A1: Horizontal loop EM type curves for three different dips and two different profile crossing angles, for a perfectly conducting half-plane (σt approaching infinity). After Telford et al. (1990).

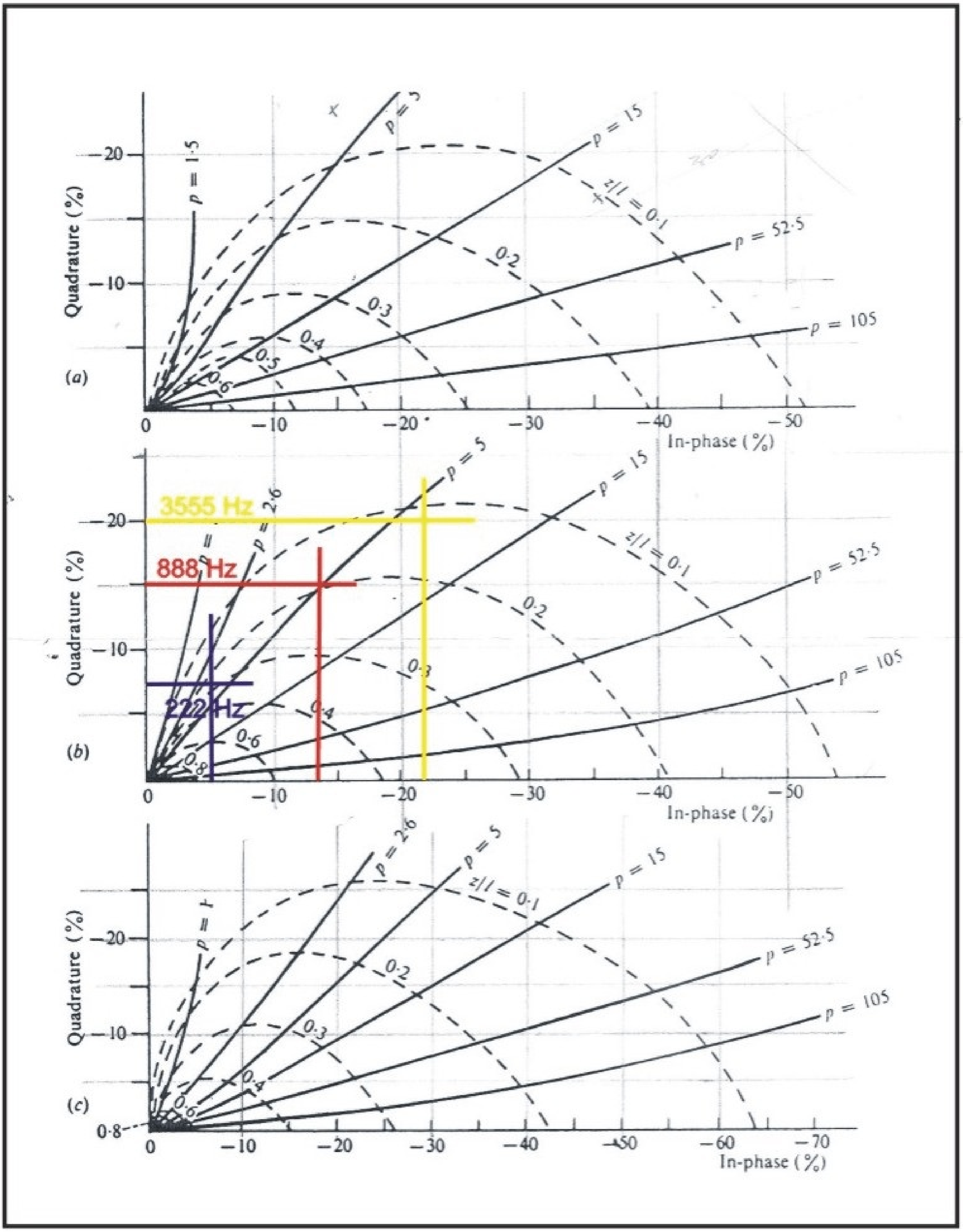

Figure A2: Characteristic curves for horizontal-loop system over a dipping sheet, after Telford et al, 1990. Response parameter p = σμωtl. See text for explanation. Coloured lines represent values from the profile in Figure 4 on page 8.

Appendix B

Induced Polarisation: Basis and Measurement Modes

B.1 Basis of IP phenomena.

All rocks contain at least very small amounts of water in their pores. Pure water is quite resistive electrically. Water in rock pores, however, usually contains many dissolved compounds, which have dissociated into charged ions. For example, if salt (NaCl) is dissolved in water, it dissociates into two types of ions: Na^+, with a positive charge, and Cl^-, with a negative charge. These ions are somewhat larger than the water molecules, and thus move slowly through the water. If a battery were to be connected along the length of a pore with Na^+ and Cl^- ions in the water, then the Na^+ ions will drift towards the negative terminal of the battery, and the Cl^- ions will drift towards the positive battery terminal.

Most rock-forming minerals (quartz, feldspar, hornblende, etc.) are electrical insulators, and cannot carry any appreciable current. The exceptions are conductive minerals, such as metallic sulphides (but not sphalerite) and graphite. In almost all rocks, therefore, electrical conduction is through the movement of charged ions in the water in the pores.

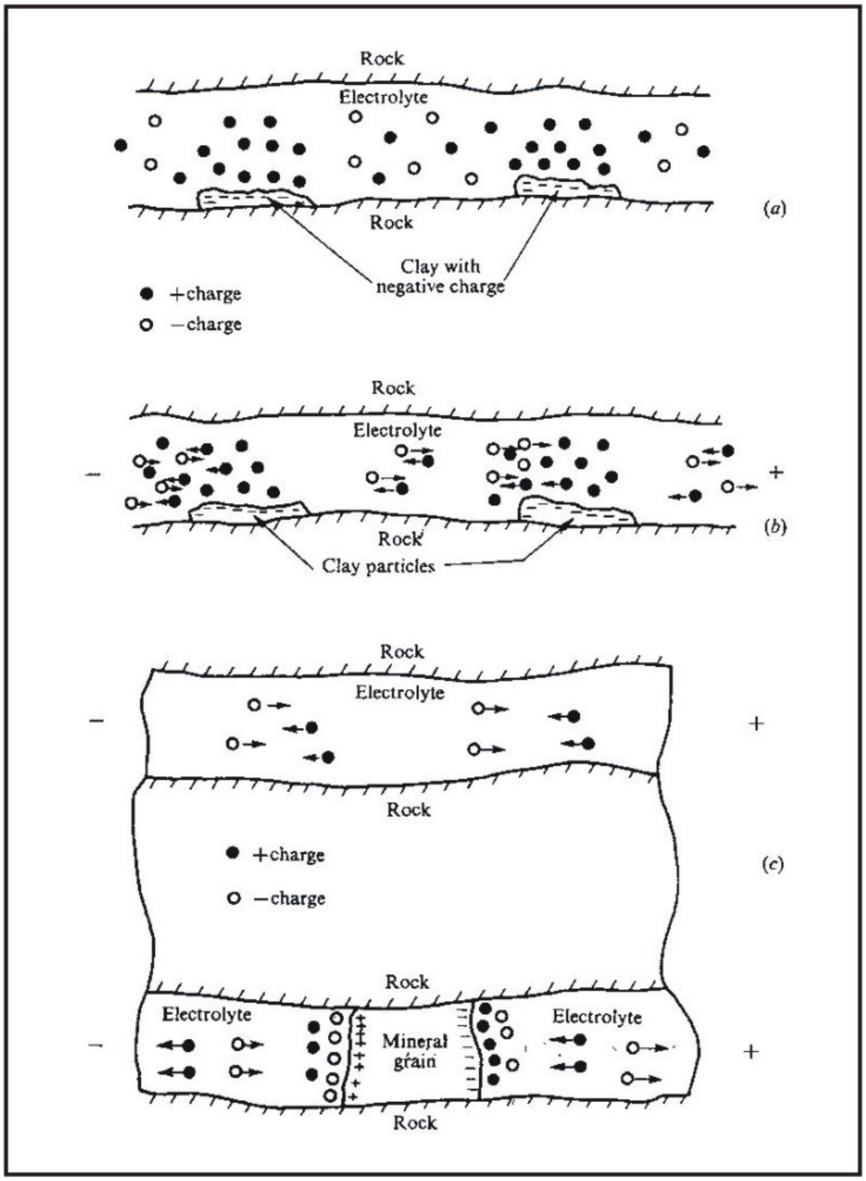

Figure B1: Membrane and electrode polarisation effects (after Telford et al. (1990).

The generation of IP effects can be due to either of two phenomena, membrane polarisation or electrode polarisation. Figure B1 shows the mechanisms involved. In Figure B1, pores in a rock are illustrated in a simplified manner as layers of electrolyte between rock walls. The upper pore in Figure B1(c) illustrates normal conduction of electric current in a plain pore. If, however, there are clay particles exposed in the walls of the pore, then the situation is depicted in Figure B1(a). Because clay particles have unsatisfied negative charges on their ends, they attract concentrations of loosely-bound positive ions. When a voltage is imposed along the pore, as in Figure B1(b), the drift of charged ions is opposed by these bound concentrations, particularly if the pore is narrow. The concentrations of positive ions are deformed during current flow, and return to their equilibrium positions when the voltage is removed. This return is seen as a small transient current, which generates a small but measurable transient voltage which is measured as an IP effect. This phenomenon is known as membrane polarisation. It occurs in almost all rocks, and gives rise to a small background IP effect.

Minerals that conduct electricity by the drift of electrons exhibit another form of polarisation known as electrode polarisation. Such minerals include the metallic sulphides, some oxides such as magnetite, ilmenite, pyrolusite and cassiterite, and, unfortunately for prospecting geophysics, graphite. When a pore is blocked by a metallic sulphide particle, as in the lower pore of Figure B1(c), the sulphide has a net surface charge on its faces as a result of the flow of current. Because the velocity of current flow is slower in the pore than in the sulphide, ions accumulate at the faces of the sulphide, awaiting their turn to give up or acquire an electron. When the current is interrupted the ions drift back to their equilibrium positions, thus generating a measurable transient voltage in the rock. Electrode polarisation can generate much stronger transients than membrane polarisation.

Because the movement of electrons in the sulphide particle is essentially instantaneous, the strength of the electrode polarisation is independent of the length of the sulphide, and depends only on the surface area exposed at each end of the particle, and on the mobilities of the ions in the pore water. It is for this reason that IP effects cannot be used directly to estimate sulphide volume. For a given percent content of sulphides, if the sulphides are very fine-grained, or smeared out in fractures, then the ratio of surface area to volume is much higher than if the sulphides are relatively coarse-grained. Thus fine-grained sulphide distributions tend to give rise to higher IP effects than the equivalent concentration of coarse-grained sulphides.

B.2 IP measurements: Time Domain IP

There are three means of measuring IP effects. The most common is Time-Domain IP, in which current is injected into the ground, interrupted while a transient decay is measured, injected in the opposite direction, and again interrupted while a transient decay is measured. Typically the transmitted current I is an interrupted square wave, with equal on-time and off-time. Most commonly the on-time and off-time are two seconds each, so that a whole cycle of measurement has a period of 8 seconds. The transient decay voltage (the secondary voltage V_s) is integrated over much of the off-time, either as a single value or more commonly as a series of values which allow determination of the dependence of the decay on time. The integrated V_s is called the chargeability.

During the current on-time, the voltage in the earth resulting from the current flow (the primary voltage) is measured. The primary voltage (V_p) is used to calculate the apparent resistivity ρ_a of the ground:

Apparent Resistivity $\rho_a = (V_p/I)\ G$,

where ρ_a = apparent resistivity,

G is a geometric factor which depends on the layout of the transmitter and receiver electrodes, and

ρ_a has units of ohm-metres if the distances in the geometric factor are in metres.

To compensate for variations in the transmitted current and the current timing, the chargeability is usually normalised by dividing by the primary voltage V_p and by the window over which the transient is integrated.

The formula for chargeability is:

$$Ch = \left(\int_{t1}^{t2} V_s(t)\, dt\right) / \left(V_p\,(t_2 - t_1)\right),$$

where: Ch is the chargeability in millivolts/volt or mV/V
t_1 is the time, after current turn-off, at which integration of the secondary voltage began, and
t_2 is the time, after current turn-off, at which integration of the secondary voltage ended, so that
$(t_2 - t_1)$ is the time over which the secondary voltage is integrated, in seconds,
$V_s(t)$ is the secondary voltage as a function of time after turn-off of the current, and
V_p is the primary voltage measured during the current-on time.

B.3 IP measurements: Frequency Domain IP

A second method of measuring IP is by using two different frequencies, one higher than the other, transmitted one after the other. For each frequency, a V_p is measured. Because the ionic drift is more complete at lower frequencies, the low-frequency V_p (V_{LF}) is higher than the high-frequency V_p (V_{HF}). The IP effect is quoted as the percent frequency effect (PFE):

$$\text{Apparent PFE} = 100\ (V_{LF} - V_{HF}) / V_{LF}$$

where V_{LF} is the primary voltage at the lower frequency,
V_{HF} is the primary voltage at the higher frequency, and
PFE has units of percent.

Occasionally PFE values are normalised by dividing by the apparent resistivity measured at one or other of the frequencies (usually, but not always, at the lower frequency). The resulting number is called the Metal Factor:

$$\text{Apparent Metal Factor MF} = 1000\ \text{PFE} / \rho_a$$

where ρ_a is the apparent resistivity measured at the lower frequency,
PFE is the percent frequency effect,
1000 is an arbitrary factor chosen to give MF values appropriate values from 10 upwards, and
MF has no units.

The concept of metal factor arose early in the development of the IP method, because it was felt that low resistivity values would damp the value of PFE more than high resistivity values.

B.4 IP measurements : Phase Angle

Because there is a slight delay in building up the clouds of ions at the interfaces with metallic sulphides, the development of the primary voltage V_p is slightly behind the development of current flow. Comparison of the waveforms of transmitted current and measured primary voltage allows determination of that delay, which is called the phase angle. Although phase angles in electronics are usually measured in degrees, the phase angles associated with IP are usually quite small when quoted in degrees. Thus phase angles in IP are quoted in milliradians, where

π radians = 180 degrees, and

phase angle in milliradians = (Phase angle in degrees) x180 degrees x 1000 / π

B.5 Calculations of apparent resistivity for common IP electrode arrays.

The common electrode arrays used in electrical surveys are shown in Figure 6 on Page 13 of the main part of this work. Apparent resistivity calculations for the common IP electrode arrays are as follows (see Figure 6, p.13, for explanation of the geometric parameters):

Multi-Dipole $\rho_a = \pi\, a\, n\, (n+1)\, (n+2)\, V_p / I,$

Pole-Dipole $\rho_a = 2\, \pi\, a\, n\, (n+1)\, V_p / I,$

Pole-Pole $\rho_a = 2\, \pi\, BM\, V_p / I,$

Schlumberger $\rho_a = \pi\, (V_p / I) \llbracket (AB / 2)^2 - (MN / 2)^2 \rrbracket / MN,$ and

Gradient $\rho_a = 2\, \pi\, (V_p / I) / \llbracket (1 / MA) - (1 / AN) - (1 / MB) + (1 / BN) \rrbracket$

where

$$MA = \llbracket (x_M - x_A)^2 + (y_M - y_A)^2 \rrbracket^{1/2}$$
$$AN = \llbracket (x_N - x_A)^2 + (y_N - y_A)^2 \rrbracket^{1/2}$$
$$MB = \llbracket (x_M - x_B)^2 + (y_M - y_B)^2 \rrbracket^{1/2}$$
$$BN = \llbracket (x_N - x_B)^2 + (y_N - y_B)^2 \rrbracket^{1/2}$$

If all the linear dimensions are in metres, V_P is in volts, and I is in amperes, then the apparent resistivities all have units of ohm-metres.

Note that apparent resistivity values are only equal to the true earth resistivity values if the resistivity values are uniform to much greater depth than the dimensions of the array used to measure them. This situation is extremely rare, so much so that I have rarely experienced it in my professional life.

Appendix C:

List of Acronyms, Symbols and Terms

AB	Distance between transmitter electordes in IP surveys.
a, na	Dipole length, number of dipole lengths.
Chargeability	IP response in time-domain IP.
EM	Electromagnetic.
Frequency domain	Measuring with continuous single-frequency signal.
GPS	Global positioning system.
HI	Height of instrument (in levelling, and in gravity survey).
HLEM	Horizontal loop electromagnetic.
HEM	Helicopter electromagnetic.
Hz	Hertz, measurement of frequency (1 Hz = 1 cycle per second).
IP	Induced polarisation.
K	Potassium
KHz	Kilohertz, =1 000 Hz
MeV	Million electron volts (measure of energy for nuclear radiation).
mrad	Milliradian (unit of measurement of IP response in phase-angle IP)
mS	Millisiemen (1 000 mS = 1 S).
mS/m	Millisiemen per metre (1/(1 mS/m) = 1 000 Ωm)
mgal	Milligal (1 mgal = $1x10^{-6}$ of the earth's gravitational field)
mV/V	Unit of chargeability in time-domain IP.
MN	Distance between receiver electrodes in IP surveys.
NaI	Sodium Iodide, crystal used in γ-ray spectrometers.
nT	Nanotesla, used to be called gamma (γ). Earth's magnetic field in Newfoundland is about 52 000 nT.
NDNR	Newfoundland Department of Natural Resources.
NLDNR	Newfoundland and Labrador Department of Natural Resources.
PFE	Percent frequency effect (unit of measurement of IP response in frequency-domain IP).
Protem	Geonics Time-domain EM system.
Pulse EM	Crone time-domain EM system.
REE	Rare earth elements.
S	Siemen (unit of electrical conductivity).
Th	Thorium
Tl	Thallium
Time domain	Measuring with repeated broad band impulses.
TDEM, TEM	Time-domain electromagnetic.
UTEM	University of Toronto TDEM System, offered by Lamontagne Geophysics. UTEM uses a triangular waveform.
U	Uranium
VLF	Very low frequency (15 to 25 KHz), type of EM survey.
α, β, γ	Alpha, beta, gamma (types of nuclear radiation).
μ	Magnetic permeability.
Ω-m	Ohm metre (unit of resistivity).
ρ	Resistivity (units of Ω-m). See Sect. 3.0 for esplanation.
σ	Conductivity (units of mS/m). See Sect. 3.0 for esplanation.
σt	Conductivity-thickness product (units of mS).
ω	Angular frequency in radians/second (1 Hz = 2π radians/second).