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The Geophysical Expression of Selected Mineral Deposit Models

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PREFACE

A continuous, dependable supply of mineral resources is essential to the economic strength of the United States. In our industrial society these non-renewable resources are the physical base from which most goods are manufactured. Recently the domestic discovery and production of new mineral deposits has slowed due to many factors, including high cost and environmental constraints. Most of the obvious near-surface ore deposits have been found, which has led integrated exploration programs to look deeper or in areas that are under cover. Geophysical methods provide an important advantage in this search for undiscovered mineral deposits. Effective use of integrated geophysical data allows a three dimensional picture of the subsurface yet data acquisition usually leaves the surface undisturbed.

Geophysical methods are based on the measurement of natural and artificial fields that are influenced by the distribution of rocks that have varying physical properties. Knowledge of the physical properties of various rock types and minerals is a prerequisite to successful interpretation using geophysical techniques. A wide variety of valuable information may be acquired by the selection and application of the appropriate geophysical techniques, along with an understanding of the regional and deposit-scale geophysical characteristics of mineral deposits.

The application of geophysics begins at the reconnaissance stage or regional scale, where remote sensing and airborne methods serve to outline broad geologic features or favorable terrain. In the detailed follow-up stage a variety of ground methods are directed at finding targets. The final stage might utilize down-hole techniques or underground surveys to define an orebody or additional reserves. Several geophysical methods can be applied and results integrated for direct detection of ore bodies, indirect detection of characteristic geologic features, or as an aid to geologic mapping.

Assessments of federal lands emphasize the evaluation of large tracts of land for potential resources of all commodities that might occur. This process is interdisciplinary and geologic, geochemical, and geophysical data must be integrated to ascertain if there is evidence of mineralization within the area of study. Geophysical data are integrated into the assessment process at various levels depending on the scale or desired resolution. Regional geophysical data sets such as aeromagnetic, gamma-ray, and gravity are readily available but may be sparse in coverage. Deposit scale geophysical surveys on certain deposit types are numerous, but acquisition of such data has waned (domestically) in recent years, and many data sets are contained in proprietary company files.

A descriptive mineral deposit model, such as given by Cox and Singer (1986), is a list of regional and local characteristics covering geology, mineralogy and geochemistry. The geophysical characteristics compiled here are an important component of the continuously evolving deposit model and therefore complement these previously published characteristics. The purpose of the geophysical model is to provide, where possible, quantitative values of physical properties and their ranges, in order to permit quantitative modeling of the geophysical response. The ultimate function of ore deposit models is to use the geologic, geochemical and geophysical characteristics to unravel the genesis and to better predict the location of new deposits. This then leads to more accurate mineral resource assessments and successful exploration programs.

W. D. Heran

THE GEOPHYSICAL EXPRESSION OF SELECTED MINERAL DEPOSIT MODELS

Introduction to Geophysical Deposit Models

by

D.B. Hoover, W.D. Heran, and P.L. Hill

INTRODUCTION

The use of formal mineral deposit models in the assessment of mineral resources on public lands has been established for almost 10 years within the U.S. Geological Survey. A catalogue of deposit models developed for assessment purposes was published in 1986 (Cox and Singer, 1986) and a supplemental catalogue appeared in 1991 (Orris and Bliss, 1991). Both of these catalogues succinctly summarize the geologic and to a lesser extent the geochemical signatures of the deposits, but give virtually nothing regarding the geophysical expression of the deposits. Thus the geophysicist assigned to an assessment team had to rely on his experience in order to interpret the significance of available geophysical data to the potential for various types of deposits in the area of study. This procedure presented problems in making full use of available data because of inexperience of some staff, lack of familiarity with all deposit types under consideration, and incomplete understanding of the varieties of geophysical data being used. It was also recognized that geophysical data needed to play a greater roll in the assessment process where relevant geologic data were obscured by barren cover rocks.

Information used to assess covered areas is obtained by extrapolation from outcrop, from secondary effects such as dispersion haloes that may be identified by geochemical or geophysical techniques, or by direct measurement of some physical property or property contrast at depth by geophysical methods. Thus, the applicability of geophysical data to assessment and exploration becomes increasingly important as the focus changes to covered deposits.

To better meet the needs of USGS staff for basic information on the geophysical signatures of the various deposit models of Cox and Singer (1986) an effort was initiated to compile a preliminary description of the geophysical characteristics of their 85 original models. The geophysical models that follow are interim compilations intended to be descriptive in nature, as the Cox and Singer models are, and to be relatively free from genetic constraints. We hope that this compilation, by being descriptive in nature, will be found useful even if current ideas on the genesis of some deposit types may change.

This paper is divided into two main parts, an extensive introduction, and a catalogue of geophysical models. The introduction explains the rational for, and format of the models, provides a brief review of geophysical methods, and gives numerous tables and graphs showing values and ranges of physical properties of host and cover rocks. By summarizing host and cover rock properties in the introduction, model compilers do not have to address host rock properties or property contrasts between various host rocks and the deposit when preparing a model. A catalogue of models follows the introductory material, each model being prepared by different staff of the Branch of Geophysics.

This compilation is, of necessity, preliminary because most deposit types have not had complete geophysical descriptions given in published literature, or relatively little public information of any kind is available on which to base a geophysical description. When trying to define the averages and ranges of physical properties of individual deposit types, the limitations of public information become even clearer. However, a start needs to be made, and if it contributes nothing else, it will identify areas of weakness in our data base. This we hope will be a challenge to other users, to make corrections where errors occur, but more importantly to augment the data base with their own hard data.

In looking over the geophysical literature we find that there are numerous papers that review the geophysical characteristics of a particular deposit, but very few that try to summarize results for a particular deposit type. But, it is the summary papers that provide the synoptic view on which to base a model description. Excellent examples of such papers are those of Kamara (1981) or Macnae (1979) on diamond-bearing kimberlites. Papers such as these on all the various deposit types would be desirable, but are not likely to appear in the near future. The compilation presented here is intended to provide interim guidance on the geophysical characteristics of deposits until adequate review papers are prepared, as well as to provide sufficient literature references to ease the search for the person needing further information.

Descriptions of deposit geophysical characteristics tend to focus on large scale, deposit-size, property variations, especially in Western literature, with progressively lesser emphasis given to district or regional characteristics. In part this approach reflects the territorial divisions between government and the private sector, the government generally having responsibility for providing basic regional data, and the private sector having responsibility for resource development. This dichotomy in scale has tended to place emphasis in the Western literature on direct deposit expression, with regional or district scale attributes generally passed over in discussions of deposit signatures. Yet it is the regional and district scale attributes of deposits that are important in most government resource assessments. In the former USSR, on the other hand, assessment and exploration have been done by the state, and regional geophysical investigations of the entire crust play a more direct part in regionalization of areas favorable for mineralization (Brodovoi and others, 1970; Zietz and others, 1976; Kužvart and Böhmer, 1986) than is evident from Western literature. Brodovoi and others (1970) note that in Kazakhstan the use of deep seismic, gravity, aeromagnetic and deep electrical data for mapping the depth to and thickness of major crustal units; depth to the Mohorovicic discontinuity; location, size and type of intrusive rocks; and major crustal structures are used as aids in defining metallogenic regions. Zietz and others (1976) state that in the southeast part of the USSR, tin is associated with a thick crust, lead and zinc deposits correlate with intermediate crustal thickness, while copper and gold are found in areas of thin crust. Our compilations attempt to include regional characteristics, but in many cases information is not directly available.

In attempting to compile the geophysical characteristics of a wide variety of ore deposits, we find that two distinct approaches are possible. One is to focus on individual geophysical techniques and the types of geological problems that may be addressed by each. Deposit types are then related to geophysical methods by identification with particular geologic attributes of a deposit, i.e., are there magnetic minerals in the deposit, or is the deposit fault-controlled? In this approach, it is hard for the compiler to specify, or the user to know, what types of methods may have been used on a particular type of deposit, or what methods are more commonly used.

Another approach is to focus on the deposit type, and identify geophysical characteristics known for that deposit and methods that have been applied. A somewhat similar method has been given by Vakhromeyev and Baryshev

(1984). In this approach, the user has all the attributes of a particular model conveniently at hand. This latter method has been used in this compilation, because assessment and exploration address one, or few, deposit types at a time making it desirable to have a summary of all characteristics of each deposit type in one place rather than scattered throughout a text. This method also ties geophysical models directly to Cox and Singer (1986), focusing on the models rather than the geophysical technique, and as such this compilation is intended to be used as a companion text.

Many authors have unselfishly contributed to this compilation of geophysical signatures of ore deposit models. For each model, the compilers are identified adjacent to the title of the model. Whenever practical, we request that when reference is made to particular models, the individuals who compiled the models be cited, rather than referencing this entire report.

ACKNOWLEDGMENTS

A compilation such as this could not be made without the support and encouragement of co-workers and colleagues both within and outside Government service. We especially want to thank Dave Campbell, Mike Foose, Andy Griscom, Bill Hanna, Bill Hasbrouck, Dan Knepper, and Jim Pitkin, all with the USGS, for their support and assistance in review of this work. From outside government Jack Corbett and Frank Fritz reviewed the material and contributed many useful suggestions and much information. The earlier work of Ed Ballantyne on his Doctoral thesis also needs to be acknowledged as providing a substantial data base from which to start building the models.

FORMAT OF THE MODELS

The model descriptions give first the title, compilers, and geophysically similar models followed by nine principal headings: A, geologic setting; B, geophysical definition of the geologic environment; C, geophysical definition of the deposit; D, shape and size of deposit, and any alteration halo and/or cap; E, a physical property table for the deposit, alteration halo, cap, and host rock if appropriate; F, remote sensing characteristics; G, general comments; H, reference list; I, selected illustrations. A few comments are necessary on each of these divisions.

The title section identifies the Cox and Singer (1986) model or models, model number, the compilers, and geophysically similar models. Identification of geophysically similar models is important because it calls attention to models with similar characteristics that the geophysicist should be aware of for assessment or exploration work. By identifying models that are geophysically similar, the compiler does not imply that there is any genetic similarity. He only means to identify other models that he believes have a sufficient number of similar attributes that they need to be considered when evaluating geophysical data.

In some cases compilers have lumped several related Cox and Singer models into one geophysical model because of the similarity of geophysical signatures and/or because of a lack of information on which to separate the deposits geophysically.

The geologic setting, heading A, is intended to be a succinct statement to remind the user of the nature of the Cox and Singer (1986) model. The geophysical models are intended to complement those of Cox and Singer. Users are referred to Cox and Singer (1988) for more details of the geologic setting.

The geophysical definition of the geologic environment, heading B, briefly states the regional- or district-scale geophysical characteristics associated with the deposit. These are features that have been suggested in the literature as important for localizing the particular deposit type. Most relate to small scale structural and lithologic features that define permissible terrains but are rarely deposit specific.

Deposit definition, heading C, briefly states the geophysical attributes of the deposit as described in the literature, and the geophysical methods most used. This section is quite variable in content. Some deposit types provide direct geophysical evidence of mineralization, but many only provide indirect evidence. The compilers provide a summary of exploration experience from the literature which can be highly variable in quality and amount of data. For example, the geophysical literature on porphyry copper deposits (model 17) is extensive, but that for Olympic Dam (Cu-U-Au, model 29b) or Kipushi (Cu-Pb-Zn) deposits (model 32c) is quite scarce. The compilers may comment on the potential for a particular geophysical method that, from the literature, had not been tried. A geophysical method not referred to may imply that the method was never tried. However, it also could be due to there being little chance for success of the method, that it would probably not be cost effective, or that it was tried and not found useful. Too often, only successful efforts are reported and the unsuccessful ignored. The user of this model compilation needs to keep in mind the caution flag raised by absence of some methods, but also needs to keep an open mind for the overlooked opportunity.

The next two headings (D size and shape, and E physical properties) provide, to the extent possible, hard data so that modeling of a deposit may be done using a variety of host rock and overburden. These are the quantitative parameters of what Vakhromeyev and Baryshev (1984) call the physico-geological model of an ore deposit. The size of the deposit, its alteration halo and cap, if important, are given. Where grade/tonnage data are available from Cox and Singer (1986) the deposit volume is given for the 90th, 50th and 10th percentiles of deposits, using the average deposit density from heading E. The generalized shape for the deposit, halo and cap are also given for input to a modeling program as appropriate.

Specific physical properties listed in the table (division E) include density, porosity, magnetic susceptibility, magnetic remanence, electrical resistivity, induced polarization (IP) effect, seismic velocity, radioactive element (radioelement) (K, U, Th) content, and an "others" category. These properties are listed separately for the deposit, any alteration halo, secondary cap, and host rock if appropriate. By breaking down the deposit and its host environment in this way, the geophysicist is able to calculate the response of a deposit in almost any setting with or without alteration products, and for any kind of cover, at least for those properties where specific property values can be assigned. For many of the geophysical responses of ore bodies it is the physical property contrast that is important, rather than absolute values of properties. However, since host and cover rocks may vary significantly it is not practical to list physical property contrasts in this table. This has tended to limit our ability to identify quantitative values for a number of properties. However, for those wishing to compute model responses, the reference list should provide supplemental information.

Where a numeric value is assigned in the table or numeric ranges are given, superscript numbers refer to the references from which the data were obtained. Units for the various physical properties may vary among models reflecting what was available in the literature. This is a particular problem for electrical induced polarization (IP) measurements which are reported in various ways that are not dimensionally consistent. A problem also exists for gamma-ray spectrometry for radioelement concentrations, as too few systems are calibrated so that only counts-per-second are often reported. For such cases the compiler decides how best to present the results.

For many entries in the table reliable quantitative values are not available. For these cases, when sufficient literature information is available to make an informed qualitative estimate, the compiler will insert high, medium, low, variable, etc., as a best estimate. If this qualitative estimate is suspect, the qualitative term will have a question mark following it such as (high?). If the compiler feels there is insufficient information on which to even hazard a guess, then the entry will be a question mark (?). Properties of the host rock are given in the table only if a particular host rock is unique to the deposit, or as for Olympic Dam (29b) for which there is only one example. Where the deposit may be hosted by a variety of rock types an asterisk (*) is shown, indicating that the properties for any particular host should be obtained from tables that follow in this introduction. Properties for overburden will also be found in these tables. In some cases the property headings of deposit, alteration halo, cap, and host rock, have been changed because of the way that geophysics is applied to particular deposit types, and because of limitations of literature information.

For example, in the case of carbonatites the deposit, alteration halo, cap, and host categories were changed by substituting alkaline complex for cap. This was done because of the wide variety of commodities found in carbonatites, their variable geophysical expression, and little use, yet, of geophysics in exploration for the specific deposits. The principal use of geophysics in this case has been in definition of the entire alkaline complex and some individual lithologies rather than in deposit definition.

Because of difficulties in fitting the specific physical properties measured by remote sensing methods into the physical properties listing of heading E, and the way that remote sensing methods are applied to minerals deposit exploration and assessment, a separate division, F, was created for this group of geophysical techniques. Under the remote sensing division, descriptions of characteristic features are given.

Following the above headings detailing the geophysical attributes of the deposit type is a heading for comments (G). In heading G the compiler gives general comments about the deposit, attributes that do not fit into other headings, and suggestions.

A list of references (heading H) follows that may include cited and uncited references. This list is not exhaustive. However, it contains many of the more comprehensive and significant references. An effort was made to include references to a wide variety of geophysical methods. In many cases compilers made a literature search of the American Geological Institute's GeoRef data base CD-ROM (DeFelice, 1991) particularly to identify foreign language literature, and to find quantitative physical property data. The reference lists are intended to provide a firm basis for those wishing to further review the geophysical literature on a particular deposit type.

The final heading (I) presents a few selected geophysical maps, or profiles, or cartoons from the literature illustrating typical responses for the deposit type. These have been redrafted from the originals for clarity.

GEOPHYSICAL METHODS

In this section a very brief review of the various geophysical methods mentioned in the models is given. This review indicates typical applications or problems that each technique can address, and points out some limitations in minerals assessment and exploration. Details of each geophysical technique cannot be given; these are adequately covered in standard texts. However, most English texts provide few practical examples or clues as to what techniques are most applicable to various types of mineral deposits. Texts that partially address these concerns are Van Blaricom (1980) published by the Northwest Mining Association, Kužvart and Böhmes (1986) who provide an eastern

European view, and the older encyclopedia texts of Heiland (1940) and Jakosky (1950) both of which have sections devoted to geophysical methods in mining worth occasional review. Present day geophysical journals provide limited help, devoting most pages to techniques or theory, and few to case histories.

Although limited to precious metal exploration in Nevada, USA, Corbett (1991) gives an excellent overview of geophysical methods currently being used, and addresses costs and survey design. He notes the need for a geologic model of what is sought, and need for more physical property information so that the geophysicist may better determine if the target is detectable. The models presented in this paper are a start at meeting the needs of the exploration geophysicist as given in the article by Corbett.

Table 1 is a chart showing the various geophysical methods for each of which the physical property, measured parameters, anomaly source and depth of investigation are given, along with examples of application in direct and indirect minerals exploration. This is an adaptation of a chart compiled by Companie General de Geophysique, Massy, France and published with modification by Van Blaricom (1980). The table also shows whether the method may be used in airborne, ground or borehole applications and the relative importance of each of these applications for minerals exploration.

The left half of the chart relates to the physical principles and geophysical aspects of each method, and identifies the basic causes of the possible geophysical anomalies. If an ore deposit does not provide, directly or indirectly, a measurable property (generally a change in a physical property between host rock and ore body) then geophysics will be of no help. Depth of burial by cover rocks is also extremely important in assessing the potential for geophysical methods to identify favorable lithologies, host structures, or the deposit itself. As anomaly sources are buried deeper, their response decreases in amplitude and their spatial wavelength increases until at some point they disappear into the geologic noise. The physical properties of cover, host rock, and deposits provided in this compilation permit modeling so that the user may estimate the possibility of detection for various deposit types of varying depth.

Some geophysical methods, such as gamma-ray and remote sensing measure only surface attributes, and others such as thermal, and some electrical are limited to relatively shallow measurements. While this is a restriction, it does not necessarily imply that these methods are useless for deeper deposits. Secondary and subtle effects, as from geochemical haloes, can often be identified by these methods as indirect measures of the presence of mineralization or structures.

The right half of Table 1 shows applications to minerals exploration both for direct detection, and for indirect detection. For each geophysical method examples are cited from the literature. This table provides an overview of the way that geophysical methods can be applied to minerals assessment and exploration and a feeling for the type of contribution to be made by each technique. Comments on each of the methods follow.

<u>Gravity</u>

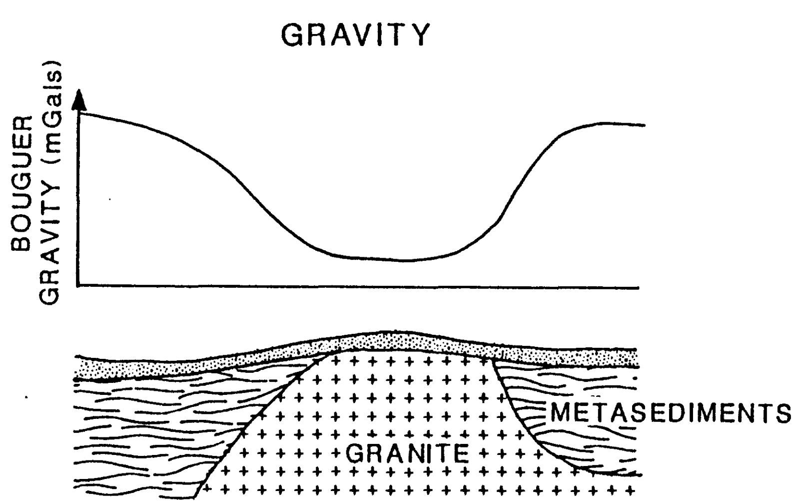


Table 1

GEOPHYSICAL METHOD						APPLICATIONS TO MINERAL COMMODITIES			
A=airborne, B=borehole, G=ground	Physical Parameter measured	Typical units	Relevant physical property	Typical source of anomaly	Depth of investigation	Direct detection	Example	Indirect detection	Example
1. Gravity G,B,A	Total attraction of earth's gravity field (essentially the vertical attraction of anomalous masses)	Milligals or gravity unit (0.1 mGal)	density	contrast in rock density	all	deposits of dense ores	chromite, Cuba Davis et al., (1957)	gold in volcanics	U.S.A., Kleinkopf et al. (1970)
								bauxite in karst cavities	USSR, Babayants et al. (1970)
							Massive sulfides Portugal Richard et al. (1984)	niobium in alkaline ring complexes	Canada Gold et al. (1966)
								sulfur in caprock	U.S.A. Nettleton (1956)
	gradient of earth's gravity field	Eötvös unit (10^{-9} gal/cm)				low density evaporite diapirs	salt domes Peters and Dugan (1949)		
2. Magnetic A,G,B	vertical, or other vector component, or total attraction of earth's magnetic field	nanotesla, or gammas	magnetic susceptibility and remanent magnetization	contrast in magnetic susceptibility; or remanent magnetization, or both	Surface to Curie isotherm	magnetite ore	Peru, Gay (1966)	Gold bearing Paleoplacers	S. Africa, Roux (1970
							Canada, Gaucher (1965)	chromite	Turkey, Yungul (1956)
	gradient of earth's magnetic field	nanotesla/m	"	"		banded iron formation	Mauritania, Gross and Strangway (1966)	diamond-bearing kimberlites	Gerryts (1967)
								base metals related to banded-iron formation	S. Africa, Campbell and Mason (1979)

Table 1 (continued)

3. Gamma-ray a. scintillometry G,A,B	rate of gamma-ray photons received	counts per second	total quantity of K+U+Th and daughters or	contrast in total K+U+Th in upper 50 cm of earth	upper 50 cm	Uranium	Australia, Montgomery (1972)	Apatite geologic mapping	Brazil, Barreto (1974) U.S.A., Pitkin (1968)
b. spectrometry A,G,B	rate of gamma-ray photons received and their energy	counts per second in various energy bands. If calibrated then %K, and PPM equivalent U and Th	quantity of K,U,Th and daughters	contrasts in K,U, and Th contents in upper 50 cm of earth		Uranium, Thorium "	Greenland, Løvborg et al. (1972) Australia, Tipper and Lawrence (1972)	tin gold greenstone belt mapping	Australia, Webster (1984) various countries Hoover & Pierce (1990) Brazil, Pires and Harthill (1989)
4. Seismic a. refraction G,B	source-receiver position and travel time of seismic energy	meters, milliseconds	Velocity of P or S waves	contrasts in layer velocities or presence of structures	all	Low velocity massive sulfides	U.S.A., Elliot (1967)	uranium lithologic mapping related to sulfides	U.S.A., Pakiser and Black (1957) USSR Vostretsov (1968)
b. reflection G,B	"	"	"	Presence of structures or velocity variations		coal	Ziolkowski (1979)	tin placers geologic province definition	Indonesia Singh (1984) Germany, Trappe et al. (1988)

Table 1 (continued)

5. Thermal a. bore-hole or (shallow hole)	thermal gradient or temperature	degrees C/m degrees C	thermal conductivity	varying heat flux or contrast in thermal conductivity	that of hole	pyrite	USSR, Lakhtionov (1968)	geologic province definition	USSR, Neprimerov et al. (1989)
						Uranium	Australia, Houseman et al. (1989)		
b. remote sensing	surface temperature day and night	degrees C	thermal inertia	contrast in thermal inertia	on order of 5 cm			silicic alteration gold district	U.S.A., Watson et al. (1990)
6. Electrical (see comments in text)									
a. Self potential G,B	natural DC field	millivolts	Eh/pH nd presence of electronic conductor; streaming potential coef.; Thermal coupling coef	vertical change in Eh/pH with presence of electronic conductor; groundwater flow; thermal flux	that of source	massive sulfides	India, Sengupta et al. (1969)		
						Cu, pyrite veins	Japan, Yamada (1967)		
b. mise-a-la-masse G,B	applied DC or low frequency AC field	millivolts	resistivity	presence of conductive ore body	that of conductive body	massive sulfide	Cu-Zn-Ag, Canada, Reed (1979	conductive phyllite containing Cu sulfides	Finland, Ketola (1979)

Table 1 (continued)

c. galvanic resistivity G,B	electrode position, applied current, and electric field	meter, amps, millivolts	resistivity	lateral or vertical contrast in resistivity	in practice to about 2 km	massive sulfide	Cu, India, Sengupta et al. (1969)	silicified zone, Au deposit lithologic structure, and alteration mapping of polymetallic replacement deposit	U.S.A., Ehni (1991) U.S.A., Keller et al. (1975)
d. induced polarization G,B	change of resistivity with frequency (PFE) or Phase angle between transmitted and received signal(ϕ) or normalized area of part of received voltage decay curve (chargeability M)	% change milliradians milliseconds	interface ionic polarization phenomena	presence of metallic luster minerals within pore space, presence of surface active clay and zeolite minerals	in practice to about 2 km	disseminated sulfides replacement Zn-Pb karst bauxite	Canada, Hansen and Bass (1966) Albania, Langore et al. (1989) Canada, Lajoie and Klein (1979) Yugoslavia, Sumi (1965)		

Table 1 (continued)

e. electromagnetic methods (see text for comments) G,A,B many variations available	dependent on method	dependent on method	resistivity	changes in earth resistivity	dependent on method; may be restricted to shallow exploration i.e., VLF, or go to mantle depths i.e., MT	massive sulfides airborne magnetite mapping	Canada, Telford and Becker (1979) Canada, Fraser (1981)	alteration, lithology, and structure mapping Au deposit Kimberlite mapping	New Zealand, Wilds and MacInnes (1991) various locals, Gerryts (1967)
f. optical range remote sensing imaging	intensity of reflected light (UV, VIS, IR)	normally recorded as optical or digital intensity image	spectral reflectance, Albedo	changes in spectral reflectance and Albedo	surface only			Au-quartz veins in shear zones uranium alteration in Goldfield District	Australia, Longman (1984) U.S.A., Vincent (1977) U.S.A., Rowan et al. (1977)

The gravity method has been used in exploration for nearly 80 years and makes use of gravity anomalies computed from gravity measurements. In current exploration practice, these measurements usually are made by using ground-based gravimeters that measure variations in the gravity field from one point to another but with amazing accuracy and precision. The gravimeter is not an absolute instrument, but is the only geophysical instrument--and one of the few instruments known to science--that can measure a change in a targeted quantity to about one part in a billion. Because the gravitational effects of shallow bodies targeted in exploration are orders of magnitude smaller than the gravitational effect of the mass of the earth (that also defines the "vertical" direction), it is essentially the "vertical" component of the anomalous mass that is measured. For subsurface exploration, special types of gravimeters are used in boreholes to measure underground densities over a larger volume and with more accuracy than other borehole density-sensing devices.

In exploration of an earlier time, pendulums that measured the absolute value of gravity and Eotvos torsion balances that measured the horizontal gradients of gravity were used, more commonly in searches for hydrocarbons than those for minerals. More recently, special types of commercially developed gravimeters, a gravity gradiometer developed by the Department of Defense, and an experimental gyrostabilized array of accelerometers developed jointly by the Charles Stark Draper Laboratory and U.S. Geological Survey are being evaluated. The last being a technique for extracting vector gravity information in contrast to the non-directional scalar information obtained by all other measuring devices. While the only airborne systems that are commercially available use gravimeters, these systems are used primarily by oil and mineral companies for regional exploration over areas that are relatively inaccessible.

The gravity anomalies used in exploration are computed by subtracting from the measured local field an assumed regional field predicted on the basis of previously assigned densities and geometrical factors for the earth and its topography. It is fortunate that this subtraction process also eliminates the earth-rotation part of the measured gravity, because the resulting gravitational part can be used directly to correlate anomaly with the density of the body that causes it. These gravity (now simply gravitational) anomalies are highs--relatively positive--over shallow bodies that are high in density but are lows--relatively negative--over shallow bodies that are low in density. Thus, high-density bodies of chromite, hematite, and barite generate gravity highs but low-density bodies of halite, weathered kimberlite, and diatomaceous earth generate gravity lows. Apart from these correlations, the gravity method offers another feature unique to it and of exceptional value in prospecting--namely, the capability of predicting the total anomalous mass that causes a given anomaly by analysis only of the anomaly itself. This capability, beyond offering estimates of ore tonnages, translates into predictions of ore volume, given estimates of ore density. It may be noted that, while the gravity method (and magnetic method--to be discussed in the following section) detect only lateral contrasts of physical property (density or magnetization), electrical methods also detect vertical contrasts of physical property (resistivity or conductivity).

The gravity method is generally used in an indirect detection mode for identification of structures or lithologies controlling ore deposition. However, the method is applicable to direct exploration for very high or low density ores such as chromite or halite. In some cases it can be effectively used to provide a measure of overburden thickness. In other cases, where the size of the ore body and its density contrast with the host are sufficiently large, gravity methods can provide a better estimate of reserves than limited drilling.

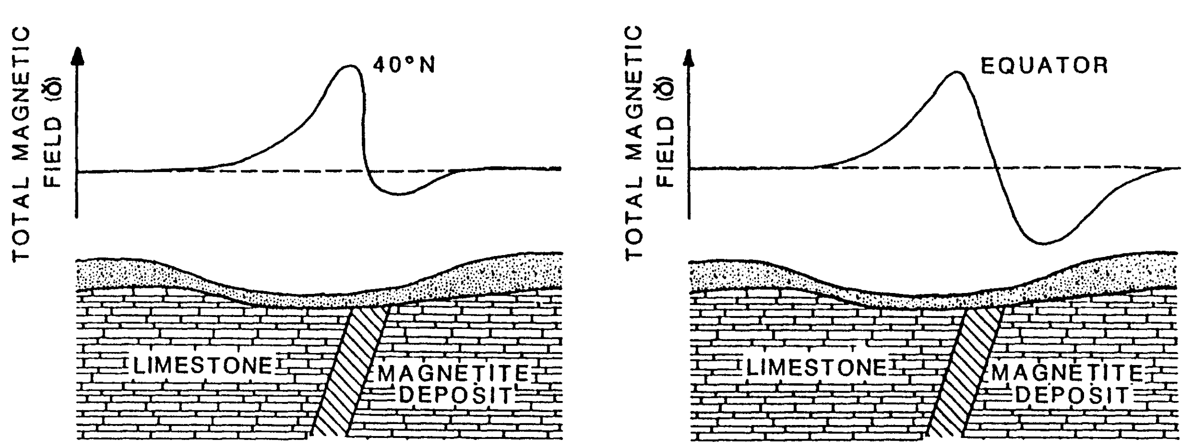
Direct detection of ore by gravity methods is well illustrated by the work of Yungul (1956) in Turkey, and Davis and others (1957) in Cuba in the exploration for podiform chromite. Yungul presented a series of curves that define the maximum magnitude of the anomaly to be expected as a function of deposit size and depth of burial. Using grade-tonnage data from Cox and Singer (1986), we note that the complete range of values to be expected for an economic deposit may be calculated in a similar way. Figure 1 presents an example for major podiform chromite deposits using grade-tonnage data of Singer and others, 1986. Following Yungul, a spherical chromite body is assumed with a density of 4.0 gm/cm^3. Host density is assumed at 2.67 gm/cm^3, a little larger than Yungul used. Three curves are shown on figure 1 giving the maximum value of the gravity anomaly for deposits of 0.0022 million tonnes, 0.02M tonnes and 0.2M tonnes. These values represent the 10th, 50th, and 90th percentiles of known deposit sizes. The area bounded by the 0.0022 and 0.2M tonne curves, the line defining the top of the spherical ore body at the surface, and a horizontal line representing geologic noise gives the range of maximum gravity anomalies as a function of depth of burial to be expected for this type of deposit. Figure 1 clearly shows that geologic noise needs to be minimal if the smaller economic bodies are to be found.

These curves are dependent on the density contrast between host and the chromite ore which can vary due to uncertainties in both host and ore densities. From much of the published literature a density of 4.0 gm/cm^3 appears reasonable for chromite (for example Mironov 1972) but measurements by Segalovich (Solovov and others, 1970) on 565 samples of podiform chromite from the Kempirsoi massif, Kazakhstan give a median density of 3.57 gm/cm^3. A density as low as this would significantly affect the detectability of chromite bodies from that shown in figure 1. This serves to emphasize the need for caution when using average rock property values from published compilations.

Figure 2 is a similar illustration, but for karst bauxite deposits. Again the body is assumed spherical but with a density of 2.45 gm/cm^3, and in a 2.55 gm/cm^3 host, an average value for limestone. From Mosier (1986) the 10th, 50th and 90th percentile of deposit sizes are 3.1M, 23M, and 170M tonnes. The maximum gravity anomaly for the karst bauxite model is seen to be slightly less than that for major podiform chromite deposits, even though the sizes of bauxite deposits are much larger. This again points out the difficulty of identifying the smaller bauxite deposits with gravity methods.

MAGNETIC

<u>Magnetic</u>



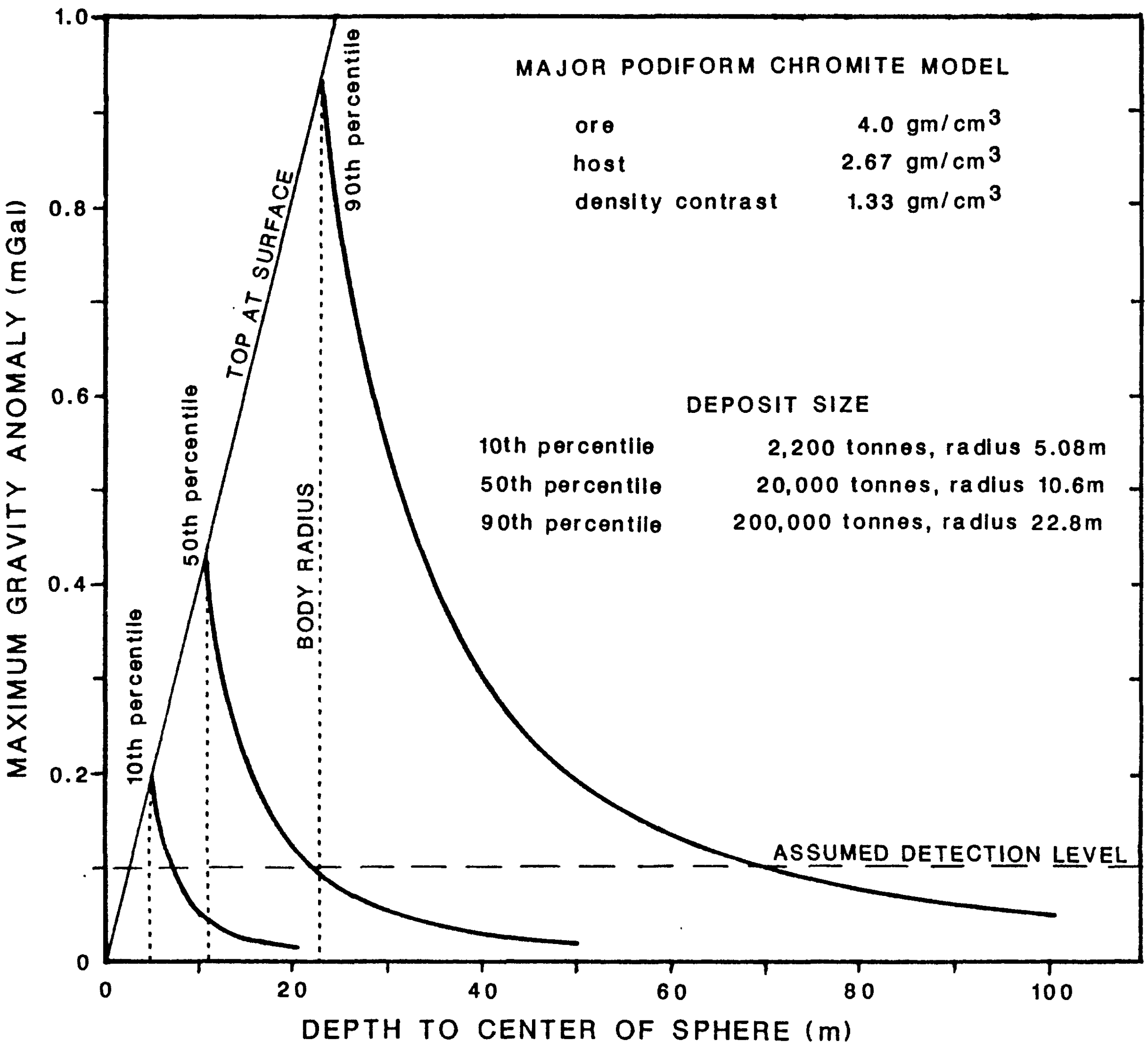


Figure 1. Graph showing the maximum gravity anomaly due to a spherical body of chromite, 4.0 gm/cm^3 in a 2.67 gm/cm^3 host as a function of depth of burial for bodies of 0.0022 M, 0.02M, and 0.2M tonnes. Size range of ore bodies represent the 10th, 50th and 90th percentiles of major podiform chromite deposits from Singer and others (1986).

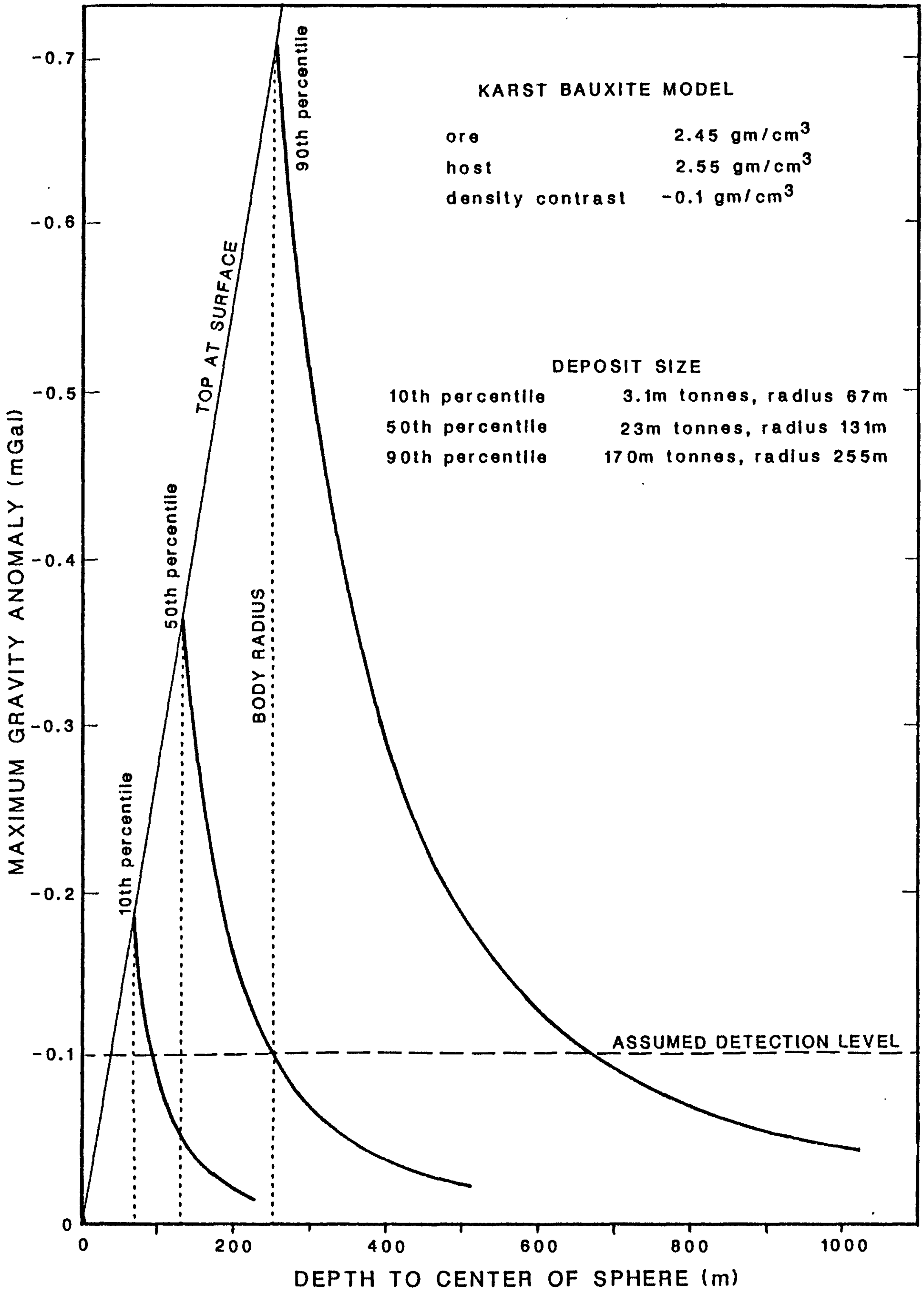


Figure 2. Graph showing the maximum gravity anomaly due to a spherical body of bauxite, 2.45 gm/cm^3 in a 2.55 gm/cm^3 host as a function of depth of burial for bodies of 31M, 23M, and 170M tonnes. Size range of ore bodies represent the 10th, 50th and 90th percentiles of karst bauxite deposits from Mosier (1986).

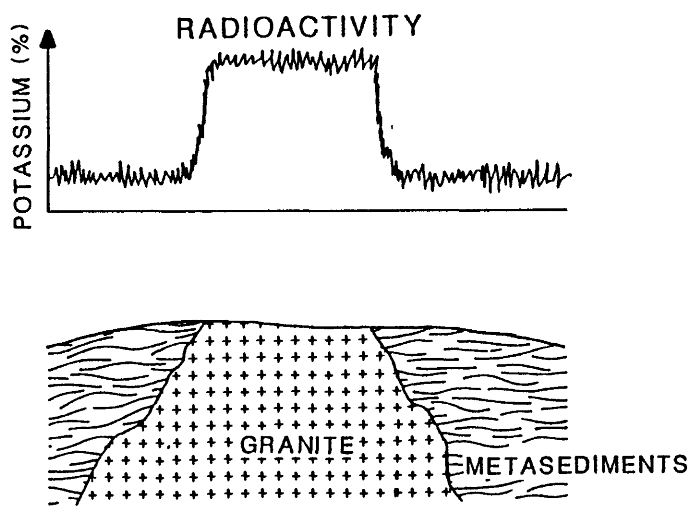
The magnetic method has been in use for more than one hundred years and makes use of magnetic anomalies computed from magnetic measurements. Although exploration programs included measurements made by using dip-needles and vertical or horizontal magnetic balances prior to about 1950, more recent programs almost exclusively restrict measurements made by using fluxgate, proton-precession, Overhauser, and optical absorption magnetometers. Normally total-field data are acquired; occasionally, vector measurements are made.

At exploration depths it is the presence of magnetic iron oxide (magnetite), iron-titanium oxides (titanomagnetite, titanomaghemite, and titanohematite), and iron sulfides (pyrrhotite and greigite) containing various combinations of induced and remanent magnetization (which added together vectorially comprise the total magnetization) that perturb the earth's primary field (Reynolds and others, 1990). The magnitudes of both induced and remanent magnetization depend on the quantity, composition, and size of the magnetic-mineral grains. The induced magnetization, which is the product of the magnetic susceptibility and the earth's ambient field, can be expressed by the magnetic susceptibility because the ambient field is relatively constant in magnitude and direction. The direction of induced magnetization approximately coincides with the direction of the ambient field, except for bodies exhibiting a strong anisotropy of magnetic susceptibility, such as magnetite and iron formation. The magnitude and direction of remanence further depends strongly on the various physico-chemical environments and various directions, polarities, and strengths of magnetic fields to which magnetic minerals have been subjected during their existence. A particularly striking contrast between induced and remanent magnetization relates to magnetic-mineral grain size: In general, relatively small grains exhibit a small susceptibility, and thus a weak induced magnetization, whereas they produce a relatively strong and stable remanent magnetization. While large grains usually exhibit a large magnetic susceptibility, and thus strong induced magnetization, they may produce either a weak or strong remanent magnetization. The relationship between the two kinds of magnetization is often expressed by the Koenigsberger ratio of remanent magnetization magnitude to induced magnetization magnitude. It should be noted that induced magnetization can profoundly affect the results of some electromagnetic measurements over electrically conductive, magnetite-rich bodies, especially those measurements made by using a controlled source in the frequency domain, as discussed in a later section.

In contrast to gravity anomalies, which occur directly over their sources, magnetic anomalies usually are shifted or displaced laterally relative to their sources, depending upon magnetization direction. Fortunately, it is often possible to re-position magnetic anomalies directly over their sources by judicious application of filtering techniques.

Magnetic anomalies also may be associated with alteration of magnetic minerals in rocks that host ore deposits related to hydrothermal systems (Hanna, 1969; Criss and Champion, 1984) and thus may outline zones of fossil hydrothermal activity. Because the rock alteration can effect a change in bulk density as well as magnetization, the magnetic anomalies, when corrected for magnetization direction, sometimes coincide with gravity anomalies. This association of a contrast of both magnetization and density in a homogeneous body implies an association of magnetic and gravitational anomalies that is expressed by Poisson's relation. In exploration geophysics Poisson's relation may be used to predict the ratio of magnetization magnitude to density, given the corresponding magnetic and gravity anomalies; further, if either magnetization or density is already known, the other can be computed. Especially interesting to explorationists is the occasional "coupling" of magnetic highs to gravity lows; this "coupling" is sometimes observed over relatively highly magnetic, low-density glassy volcanic rocks containing single-domain magnetic-mineral grains; highly vesicular basalt; serpentinite and weathered kimberlite; and felsic-to-intermediate plutons emplaced into relatively nonmagnetic gneissic terrain.

Although direct magnetic exploration is essentially limited to iron ore deposits such as those of magnetite or banded iron formation, magnetic methods often are an essential tool for deducing subsurface lithology and structure. These methods also may be used for placer identification by mapping of magnetite concentration, exploration for chromite due to associated magnetite, base-metal exploration by identification of associated magnetite or pyrrhotite content, and identification of zones favorable for deposition on regional or district scales.



Gamma-ray methods

Gamma-ray methods may use scintillometry to identify, indiscriminately, the presence of the radioelements potassium (K), uranium (U) and thorium (Th), or by the use of multi-channel spectrometers provide qualitative or quantitative measures of the individual radioelements. Spectrometers may be calibrated to give quantitative measures of radioelement concentrations if readings are made over test areas of known concentrations. It is unfortunate that many published gamma-ray data were obtained without the use of calibrated systems.

Gamma-ray methods have had wide application in uranium exploration because they provide direct detection. However, until recently in the West, these methods have not had as much application to other commodities as the authors believe they deserve. The former Soviet Union appears to have made the most use of this technique (for example see Hoover and Pierce, 1990 or Vavilin and others, 1982) in minerals exploration. For those looking at applications, the Russian literature needs to be examined.

When looking at uranium or thorium values derived from gamma-ray spectrometry the user needs to remember that the values are expressed in equivalent uranium or thorium content based on equilibrium of the decay series. This condition is often not met by uranium in the near surface (Durrance, 1986), because of uranium's mobility in an oxidizing environment. However, it may be relatively immobile in near surface units high in phosphates, clay, or organic materials.

Thorium is generally the most immobile of the three radioelements, behaving geochemically in a way similar to zirconium. It is often found in association with the rare earths. Thorium content, like uranium content, tends to increase in felsic rocks and generally increase with alkalinity. The

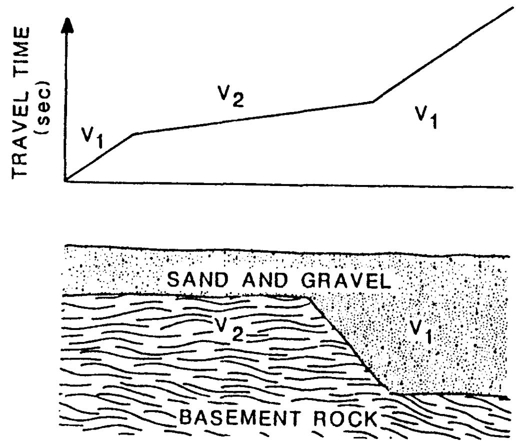
K/Th ratio in igneous rocks is generally on the order of 3 x 10^3, and the Th/U ratio typically 3.5 to 4.0 (Durrance 1986).

Radon and radiogenic helium soil gas methods are used more often by the geochemist than by the geophysicist. They will not, therefore, be considered to a major extent in the compilation.

Indirect applications of gamma-ray methods include exploration for coal and lignite, radioactive heavy minerals, and phosphates. The identification of lithologic differentiation in igneous bodies, and identification of radioelement haloes, primarily potassium, around hydrothermal ore deposits are other important uses. Hansen (1980) provides an excellent review of gamma-ray methods for the explorationist.

<u>Seismic methods</u>

SEISMIC



Seismic techniques have had relatively little use in minerals assessment and exploration at the deposit scale, in part due to their relatively high cost. However, they can provide better structural detail and better estimates of depth to lithologies of differing acoustic impedance than other geophysical techniques. The refraction method is most used in minerals work principally for mapping of low-velocity alluvial deposits such as those of gold, tin, or sand and gravel. The more expensive reflection method is not commonly used except for exploration for salt domes. However, most of the salt dome exploration is for associated petroleum and not for the salt or sulfur content of the dome. The reflection method is also used for offshore placer exploration where data acquisition becomes less expensive.

In this compilation only controlled-source (active) seismic techniques are considered. Large scale regional studies such as used by the Russians for regionalization of metallogenic districts may make use of both active and passive seismic (earthquake, or microseismic sources) methods. Because of difficulties in evaluating these regional data, and assigning characteristics to particular model types passive seismic methods are not generally considered in this preliminary model compilation.

<u>Thermal methods</u>

Two quite distinct techniques are included under thermal methods on table 1. Under (a) are the borehole or shallow probe methods for measuring thermal gradient, which with a knowledge of the thermal conductivity provides a measure of heat flow. These are essentially in-hole techniques. The second technique (b) is an airborne or satellite based method, one in which the earth's surface temperature is determined by measuring the thermal infrared radiation emitted by the surface. By measuring day and night temperatures the thermal inertia of the surficial materials may be calculated.

Borehole thermal methods have direct application to geothermal resources, but are seldom used in minerals exploration. However, there appears to be some potential for this method in exploration. Sources of heat that can produce heat flux anomalies relevant to minerals exploration are oxidizing sulfides, and high concentrations of radioelements. On the regional scale, Brown and others (1980) have shown a correlation between high heat flow provinces and mineralization in Northern-Central England and Southwest England. They suggest that heat production due to radioelements in the Hercynian and post-Hercynian granites was responsible for generating hydrothermal systems long after the granites had cooled, and that these late hydrothermal systems then produced the numerous epithermal mineral deposits of the region. Ovnatanov and Tamrazyan (1970) and Neprimerov and others (1989) also comment on the applicability of thermal methods for identifying structures on a regional scale.

On the deposit scale, a number of papers indicate the potential for thermal studies. High heat flow has been observed over the Olympic Dam Cu-U-Au deposit, Australia (Houseman and others, 1989); over a carbonatite in Nebraska (Gosnold and others, 1981); and over a small mineralized Tertiary intrusive in New Mexico (Zielinski and DeCoursey, 1983). Temperature anomalies over sulfide bodies of about $1^{o}c$ are shown in Lakhtionov (1968) who notes that thermal methods have been used in Russia since 1935. Bose (1983) notes its increased use in India where 2 to $5^{o}c$ anomalies over sulfide bodies are used to help discriminate ore from graphite, but no details are given. Logn and Evensen (1973), based on measurements of thermal conductivity on ore and county rock from the Joma pyrite deposit, also suggest the possibility of thermal measurements to distinguish between sulfide ores and graphite.

Structures such as salt domes, basement highs, sand lenses, and faults also can be identified by thermal methods (Van Ostrand, 1934; van den Bouwhuijsen, 1934; Jakosky, 1950; Ovnatanov and Tamrazyan, 1970). Where boreholes are available, particularly in covered terrain, the explorationist needs to be aware of the potential for thermal methods.

Thermal infrared imaging methods belong to the broader remote sensing techniques. Images obtained in this wavelength range may be used as other photographic or digital images for photogeologic type interpretation or, if day and night images are available, further processed to provide an image of the thermal inertia of the surface. Unconsolidated or glassy materials can be distinguished by their low thermal inertia. This airborne method can also distinguish limestone from dolomite for lithologic mapping.

Electrical methods

In contrast to other geophysical methods, electrical methods comprise a multiplicity of separate techniques that measure distinct geophysical attributes of the earth, with differing instruments and procedures, having variable exploration depth and lateral resolution, and with a large and confusing list of names and acronyms describing techniques and variants of techniques. We have divided the electrical methods into five distinct classes: (A) the self potential, (B) the induced polarization method, (C) the mise-a-la-masse, (D) the galvanic resistivity, and (E) the electromagnetic resistivity. These are shown in Figure 3, where the three distinct source phenomena are identified, and some variations of each method listed. In the

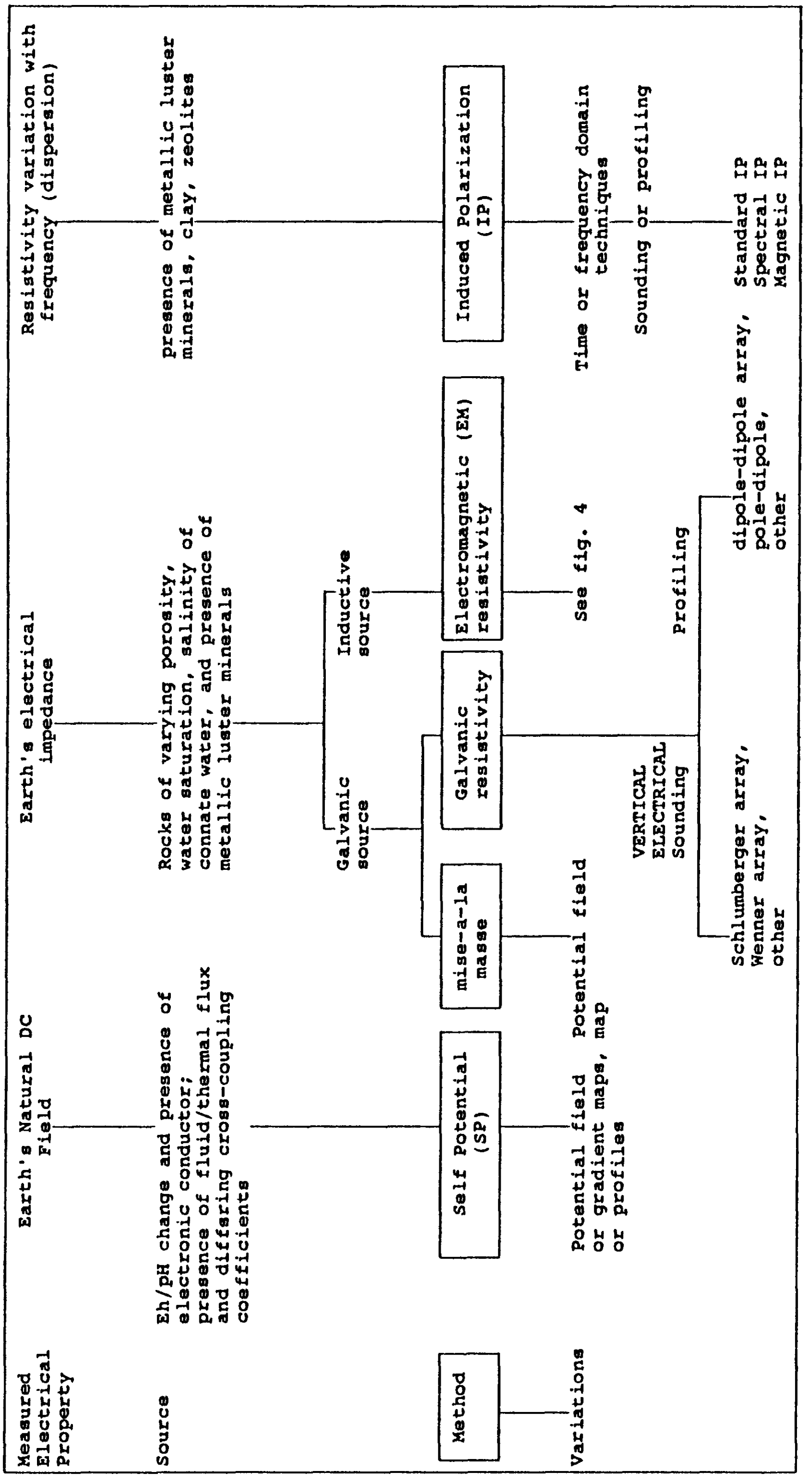
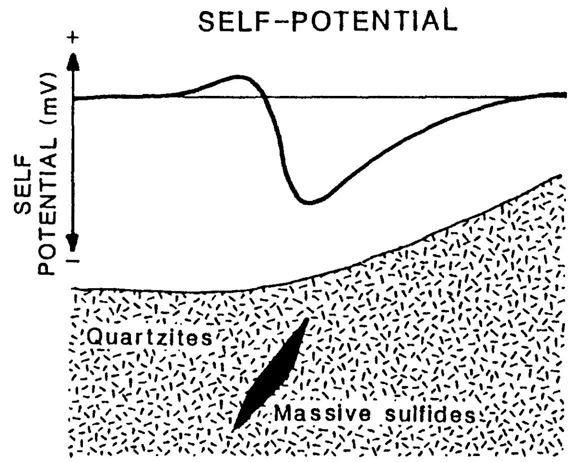


Figure 3. Diagram showing the five principal electrical methods and their source phenomena.

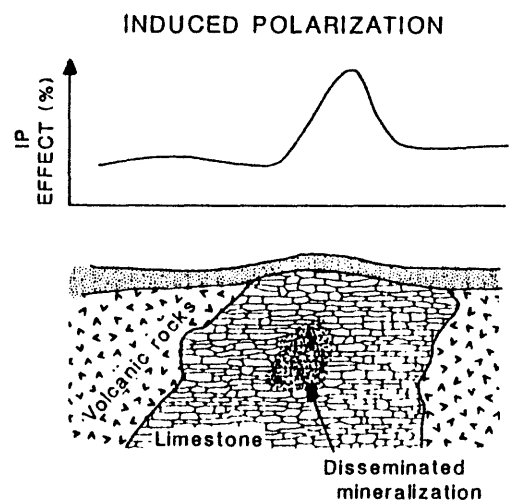
case of electromagnetic methods there are so many variations, and differing acronyms and trade names that the variations are detailed in Figure 4. In spite of all the variants of the electromagnetic method, measurements fundamentally are of the earth's electrical impedance or relates to changes in impedance. Some of the electromagnetic methods listed in figure 4 are really hybrid techniques because source fields may be generated through galvanic contact to the earth (TURAIR, CSAMT, etc.), or receiver electric fields may be measured through galvanic contact to the earth (CSAMT, AMT-MT, VLF, telluric, etc.). However, for convenience these have been classified with the electromagnetic methods.

A. Self Potential

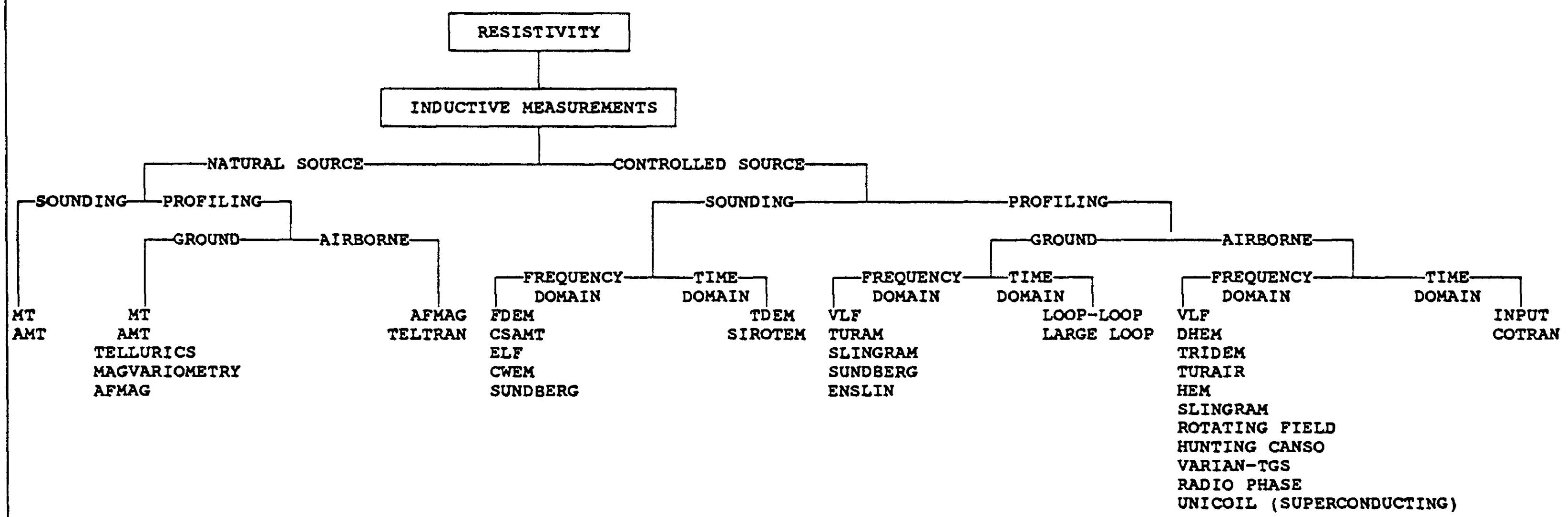


For the self potential method there are several possible sources giving rise to a dc or quasi-dc. natural electrical field. For mineral deposits the most important is the Sato and Mooney (1960) type source established when an electronic conductor, such as a massive sulfide or graphite body, extends between an oxidizing and reducing zone or over a range in pH. Other self potential sources are due to fluxes of water or heat through the earth.

B. Induced polarization



The induced polarization method provides a measure of polarizable minerals within the water bearing pore spaces of rocks. These minerals are



AFMAG - audio frequency magnetics
AMT - audio-magnetotellurics
CSAMT - controlled source audio-magnetotellurics
DHEM - digital helicopter EM
ELF - extra low frequency
EM - Electromagnetics
FDEM - frequency domain EM

INPUT - integral pulsed time domain EM
MT - magnetotellurics
TDEM - time domain EM
TRIDEM - three loop airborne EM
TURAIR - airborne TURAM
TURAM - two loop EM
VES - vertical electric sounding
VLF - very low frequency

TELTRAN - telluric transient
CWEM - continuous wave EM
TURAIR - airborne TURAM
HEM - helicopter EM
COTRAN - correlation of transients

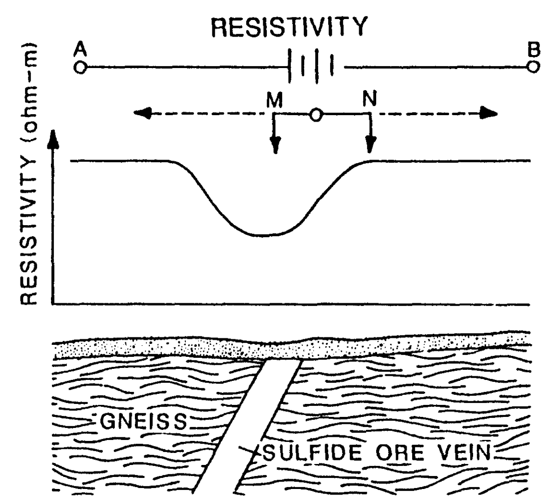
Figure 4. Tree diagram showing a classification of electromagnetic methods, and some of the techniques belonging to each branch.

metallic-luster sulfides, clays, and zeolites. The above mineral groups, in order to be detected, must present an active surface to the water in the pore space. Sulfide mineral grains completely enclosed by a nonconducting matrix such as silica will not be detected by the IP method. Since the IP response relates to the presence of active surface areas within the rock, disseminated sulfides provide a much better target than massive sulfides for this technique. This method has found its greatest application in exploration for disseminated sulfide ores where its good sensitivity (as low as 0.5% total metallic luster sulfide may be detected, according to Sumner, 1976) makes it a primary tool.

C. Mise-a-la-masse

The mise-a-la-masse method is a little used technique that is applied to conductive ore deposits that have a large resistivity contrast with the host rock. Under these conditions, electrical contact is made to the ore body, either at the surface or through a drillhole, with a source of direct or low frequency current. The other electrical pole is placed some distance away. When energized, the ore body becomes essentially an equipotential surface. The field from this body can then be mapped at the surface revealing the position of ore below the surface. An excellent example of this method is given by Mansinha and Mwenifumbo (1983). Application of this method is principally for massive sulfides.

D. Galvanic resistivity

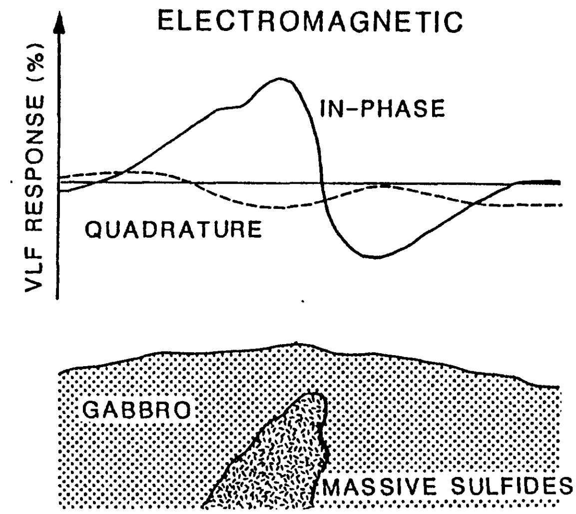


Galvanic resistivity methods, often referred to as "dc" resistivity methods, provide a measure of earth resistivity using a dc or low frequency ac current source. Source current is introduced into the earth, and the electric field is measured, through electrodes in galvanic contact with the earth. Resistivity in earth materials is primarily a function of porosity and water content, high porosity giving low resistivity in water saturated rocks. Resistivity values may range over five orders of magnitude in normal near-surface environments. Electrical conduction in rocks at dc and low frequencies occurs through ionic migration in the water of the pore spaces and more rarely, partially by electronic conduction through metallic luster minerals. Because metallic luster minerals typically do not provide long continuous circuit paths for conduction in the host rock, bulk rock resistivity almost always is controlled by the water content and dissolved ionic species present.

In contrast to potential field methods dealing with natural fields such as gravity, magnetic, and self potential methods, the galvanic resistivity techniques use an applied field and are thus able to control the depth of exploration by the spacing of the current and potential electrodes. If one is looking for lateral resistivity changes within a given depth range, then a fixed electrode array may be used to profile across an area of interest. On the other hand, if information on variations of resistivity with depth are desired, then an array may be expanded about a fixed point (a vertical electrical sounding, VES). The variations between profiling and sounding and between electrode arrays leads to differing names being applied to each variant, i.e., Schlumberger (array), vertical electrical sounding (VES), etc.

The galvanic techniques have application to a wide variety of ore deposit exploration. Massive sulfides can provide a direct very low resistivity target, or alteration products within and around hydrothermal deposits often provide a clear low resistivity target. The wide range of resistivities of earth materials also makes the method applicable to identification of lithologies and structures that may control mineralization.

E. Electromagnetic



Electromagnetic methods are probably the most confusing to the non-practitioner because of the many variants, and acronyms, or trade names used to describe them. Figure 4 presents one scheme for classification of EM methods in a tree form. The first branch is based on whether the energy source is natural or artificial. For each of these the next level of branching is based on whether the method is a profiling technique or a sounding technique. The third level of branching is based on whether it is an airborne or ground method, and the last branch based on time domain or frequency domain techniques. At the ends of these 9 resultant branches are given the names and acronyms of some of the electromagnetic methods that apply. In all, thirty-one different terms are shown, and this is not an exhaustive list.

The practical exploration depth of each system is quite variable and depends on the operating frequencies, the rock resistivity, structure, and the source-to-receiver distance. For controlled source airborne methods the maximum exploration depth is on the order of 100 meters. The natural source airborne methods have greater depth potential, but unfortunately none have been used for many years. As in galvanic resistivity techniques, soundings can be made by changing the source-to-receiver separation. In practice such soundings are normally used in shallow exploration. However, electromagnetic methods also permit sounding by variation of the operating frequency or time, for time domain systems, and this procedure is becoming of greater importance in exploration, especially where definition of deep features are desired. In the compilation of deposit characteristics no attempt is made to distinguish among the numerous EM methods, nor in many cases between EM and galvanic resistivity methods. For all of these various techniques, they either provide a measure of resistivity or impedance or respond to changes in resistivity or impedance, and this is the important attribute for the model.

The most common application of EM methods to minerals exploration has probably been in the search for massive sulfides. Normally airborne methods are used to screen large areas providing a multitude of targets for further screening by ground methods. Airborne EM methods are now beginning to find increasing use in mapping applications where lithologic and structural features can be identified in areas of difficult access or where cover exists (Palacky, 1986; Hoover and others, 1991).

Hohmann and Ward (1981) provide an excellent review of electrical methods that are used in minerals exploration.

Remote Sensing

In table 1 the remote sensing category includes only those methods making use of images obtained in the ultra-violet (UV), visible (VIS) and near infrared (IR) bands of the electromagnetic spectrum. Data in this range are treated in image format, often in digital form, so that they can be conveniently processed. Where single images are used, interpretation of lithologies and structures is based on photogeologic methods. However, recent airborne and satellite multispectral digital systems now permit extraction of much more information from the images. By comparison with known spectral responses of minerals or mineral groups, the presence of iron hydroxides, silica, clay alteration, etc., can be defined over broad areas.

In the compilations of deposit models remote sensing attributes from UV to near IR methods are most often mentioned. However, where information is available, the remote sensing category will include thermal IR characteristics and side-looking airborne radar (SLAR).

Other methods

Like SLAR, there are a number of other geophysical or quasi-geophysical methods that have been applied to mineral deposits or have potential application, but have a very limited history particularly in the western literature. Techniques such as the piezoelectric method for quartz veins (Volarovich and Sobolev, 1969), UV laser induced fluorescence to find scheelite, hydrozincite and other fluorescent minerals (the Luminex method Seigel and Robbins, 1983), airborne gas sniffing such as for mercury, the Russian CHIM (partial extraction of metals) electrogeochemical sampling technique, radon caps, etc., are examples. These are not covered in the model compilation in general. If the compiler finds a reference, and feels that one of these uncommon methods is or may be important then it would be mentioned in the comments section of the model.

Ground penetrating radar also is not covered, although it has had some limited applications in mineral exploration. Hammond and Sprenke (1991) identified sulfides below glacial ice, Davis and others (1984) describe its applications for placer exploration, and Annan and others (1988) show its application in determining stratigraphic relationships in potash mining.

PHYSICAL PROPERTIES OF HOST AND COVER ROCKS

This section summarizes the physical properties of host or cover rocks important for modeling of geophysical responses, or evaluation of geophysical data. The properties considered are density, porosity, magnetic susceptibility and remanence, seismic velocity, resistivity, IP effect, electrokinetic coupling coefficients, thermal conductivity, inertia and heat sources, and radioelement content. This is not a critical summary that tries to evaluate the adequacy of published data. It is simply a summary for which critical evaluation is left to the user. However, the limitations of many published catalogues of physical properties need to be pointed out, because most property measurements are of intrinsic properties derived from measurements on laboratory samples. Properties measured on hand-sized laboratory samples may or may not be representative of those properties in-situ. Particular care needs to be taken if laboratory measurements of density, seismic velocity or resistivity are extended to represent in-situ bulk properties because of macro-scale fractures that may be present in the earth and the amount of interstitial water.

In choosing host or cover rock properties from lists such as presented here the effects of alteration processes on those properties need to be considered. Many of the values represent measurements on relatively fresh samples of rock. Processes such as weathering, digenesis, metasomatism and hydrothermal alteration can significantly affect all of the physical properties, in some cases causing an increase in value and in others a decrease. Processes attendant to mineralization often work to the advantage of the explorationist by providing a larger and more easily identified indirect geophysical target. Alteration processes generally increase the range of possible property values, add to the geologic noise and may make interpretation more difficult.

Density

A number of factors affect the in-situ density of earth materials, including porosity, water content, depth of burial, age, crystallinity and chemistry. Awareness of all these factors is important in evaluating the applicability of the gravity method to a particular exploration problem, especially in regard to the magnitude of "geologic noise" to be expected. Sediments, both unconsolidated and consolidated, may vary considerably in density depending on the degree of saturation. Density contrasts can exceed 0.5 gm/cm^3 between wet and dry sand or gravel. Figure 5 shows ranges of densities, wet and dry, and porosities for sandstones, shales, limestones, dolomites, and unconsolidated material of various types from three literature sources (Telford and others, 1976; Jakosky, 1950; and Fedynskiy, 1967). Most apparent from figure 5 is the wide range of densities exhibited by these sedimentary units. This presents obvious problems to the modeler in knowing what densities to choose where sediments are present. Note also the high densities shown by some shales.

Based on literature values, the density variation for individual igneous rock types is not nearly as large, as shown for sediments. Figure 6 shows the ranges of bulk density from four literature sources (Daly and others, 1966, Telford and others, 1976; Johnson and Olhoeft, 1984; and Mironov, 1972), for 13 igneous rock types. Because of the low porosity of most igneous rocks, there are only minor differences between wet and dry densities. Many references do not indicate whether wet or dry densities were measured, and in some cases original sources are not indicated. Similarities in values between the references shown and other sources suggest that many of the data date back to Daly's 1933 work and Reich's 1914 work (Heiland, 1940).

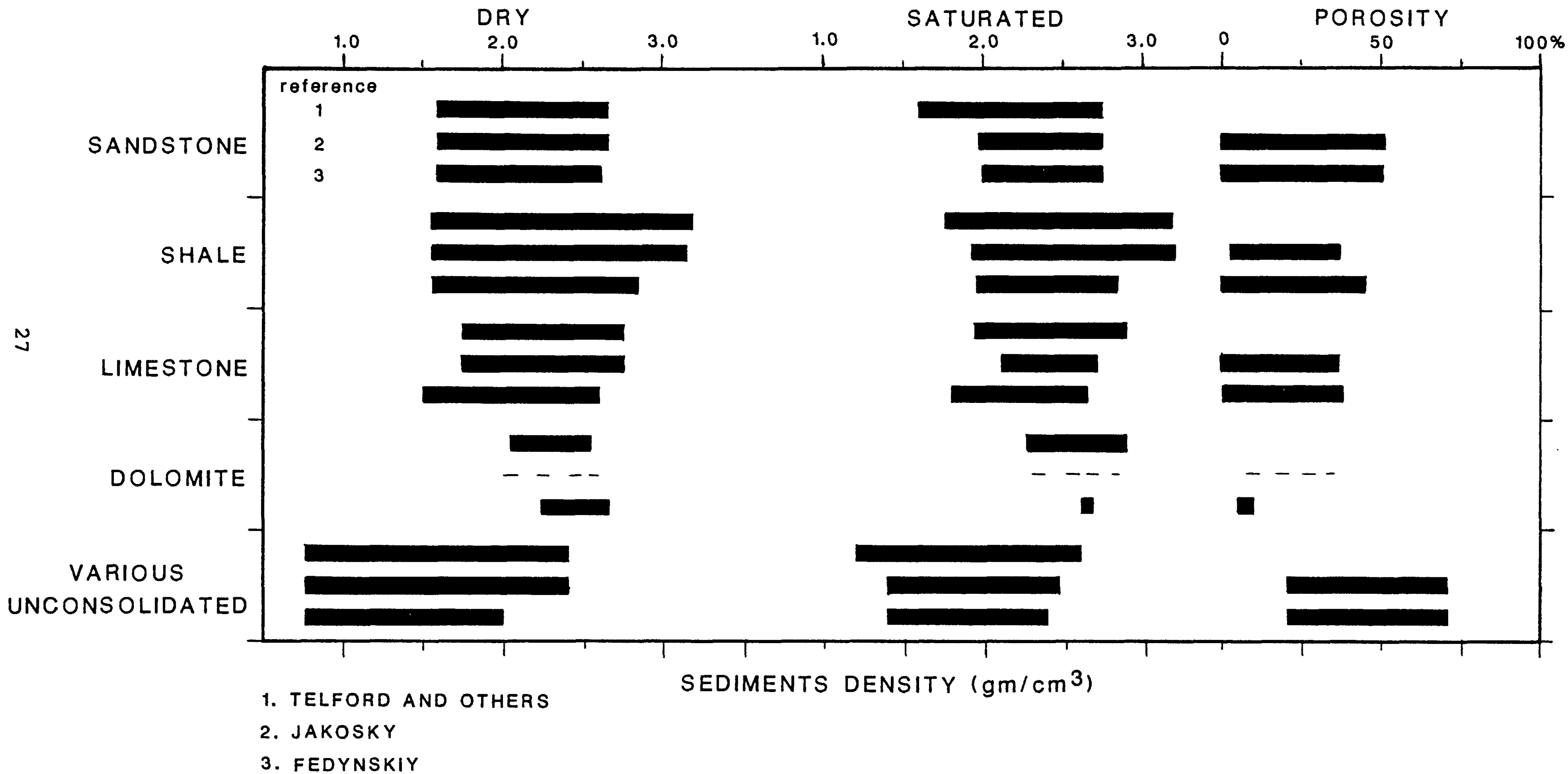


Figure 5. Diagram showing ranges of wet and dry bulk densities and porosity for various sedimentary rocks. Reference sources are 1. Telford and others, 1976, 2. Jakosky, 1950, and 3. Fedynskiy, 1967.

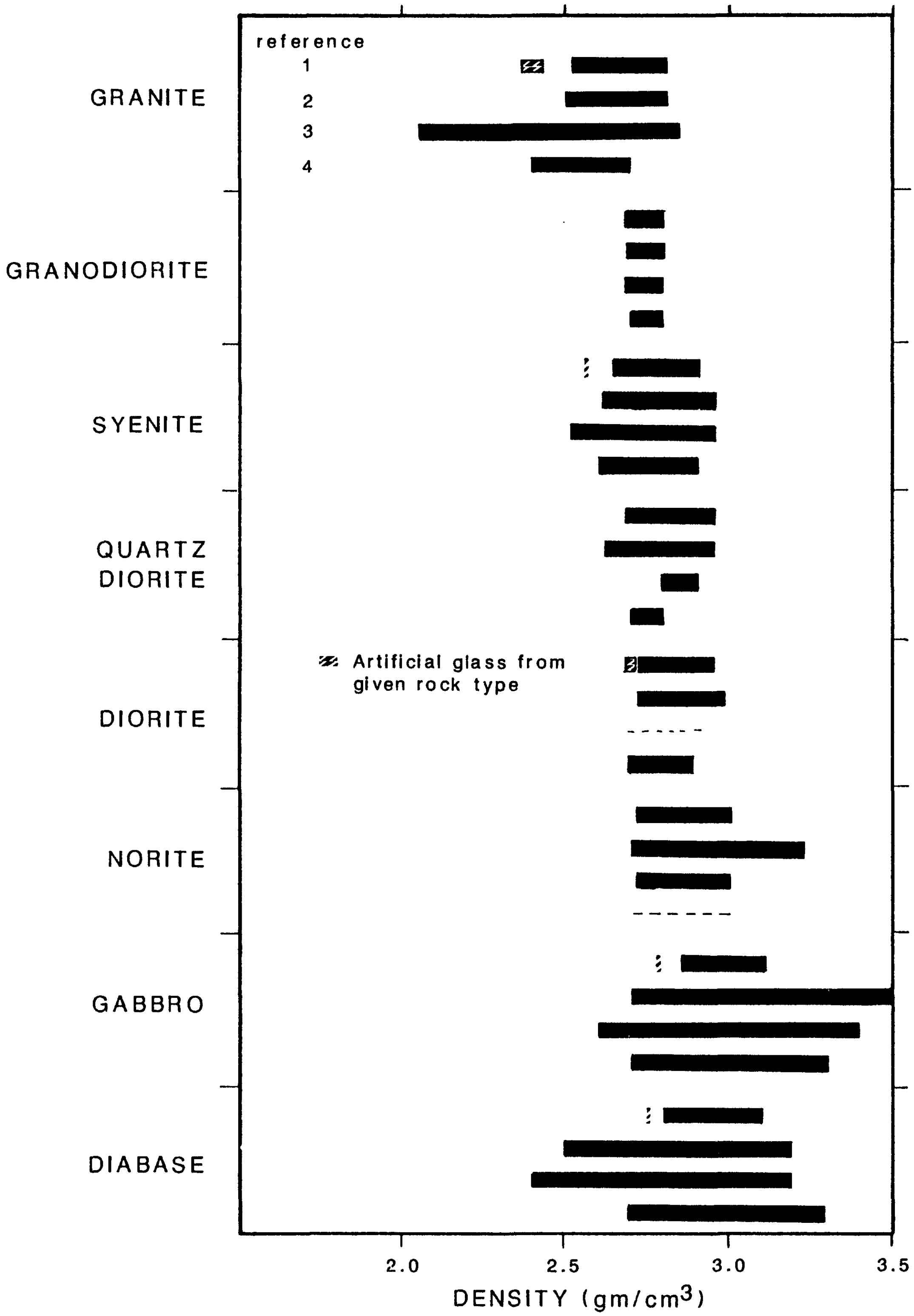


Figure 6. Diagram showing ranges of bulk densities for 13 different igneous rocks. Reference sources are 1. Daly and others, 1966, 2. Telford and others, 1976, 3. Johnson and Olhoeft, 1984, and 4. Mironov, 1972.

The values shown in figure 6 appear to represent relatively fresh samples. Alteration from weathering and some but not all hydrothermal processes would reduce the density. The wide range shown for granite by Johnson and Olhoeft (1984) in figure 6 probably results from inclusion of altered samples among the 334 granites measured.

In figure 6, density values are also given for artificial and natural glasses from Daly and others (1966). Five artificial glasses derived from granites, syenites, diorites, gabbro and diabase are shown next to their counterparts. The glasses are all significantly less dense than their igneous equivalent. Extrusive igneous rocks are typically less dense than their intrusive equivalent in part because of their glass content, and often because of increased porosity.

Figure 7 presents the ranges of densities for 9 metamorphic rocks as reported from the same sources used for igneous rocks. Note the relatively low range reported for slate. As with the igneous rocks many reference sources do not indicate whether the measurements are of wet or dry samples. In most cases these metamorphic rocks also would be of low porosity, if fresh, causing little change between the wet and dry state. Johnson (1983) has tabulated dry bulk density and porosities for 182 different rock samples. Table 2 provides a summary for several types of metamorphic rocks from his work. The porosities do not exceed 3.49% on his samples. Further details on rock densities can be found in the references cited.

Table 2. Dry bulk density and water available porosity of selected metamorphic rocks from Johnson (1983).

Rock type	***Density range***	***Porosity range***	***No. of samples***
Quartzite	2.645-2.733	0.11-0.36	3
Hornfels	2.688-2.709	0.07-0.29	2
Schist	2.634-3.018	0.29-3.49	9
Marble	2.648-2.991	0.07-0.7	9
Slate	2.762-2.79	0.25-0.43	4
Gneiss	2.617-3.137	0.22-1.73	7
Eclogite	3.251-3.359	0.09-0.36	2

Porosity

Porosity is a property that is not measured directly by one of the geophysical techniques, but one that can dramatically affect in-situ density, resistivity and P-wave seismic velocities. It is because of the effect on these other properties that porosity is important in interpretation and modeling of ore deposits. Literature on the geophysical attributes of ore deposits infrequently provides measured porosity data, and this is reflected in the presented models by the few quantitative values for this parameter.

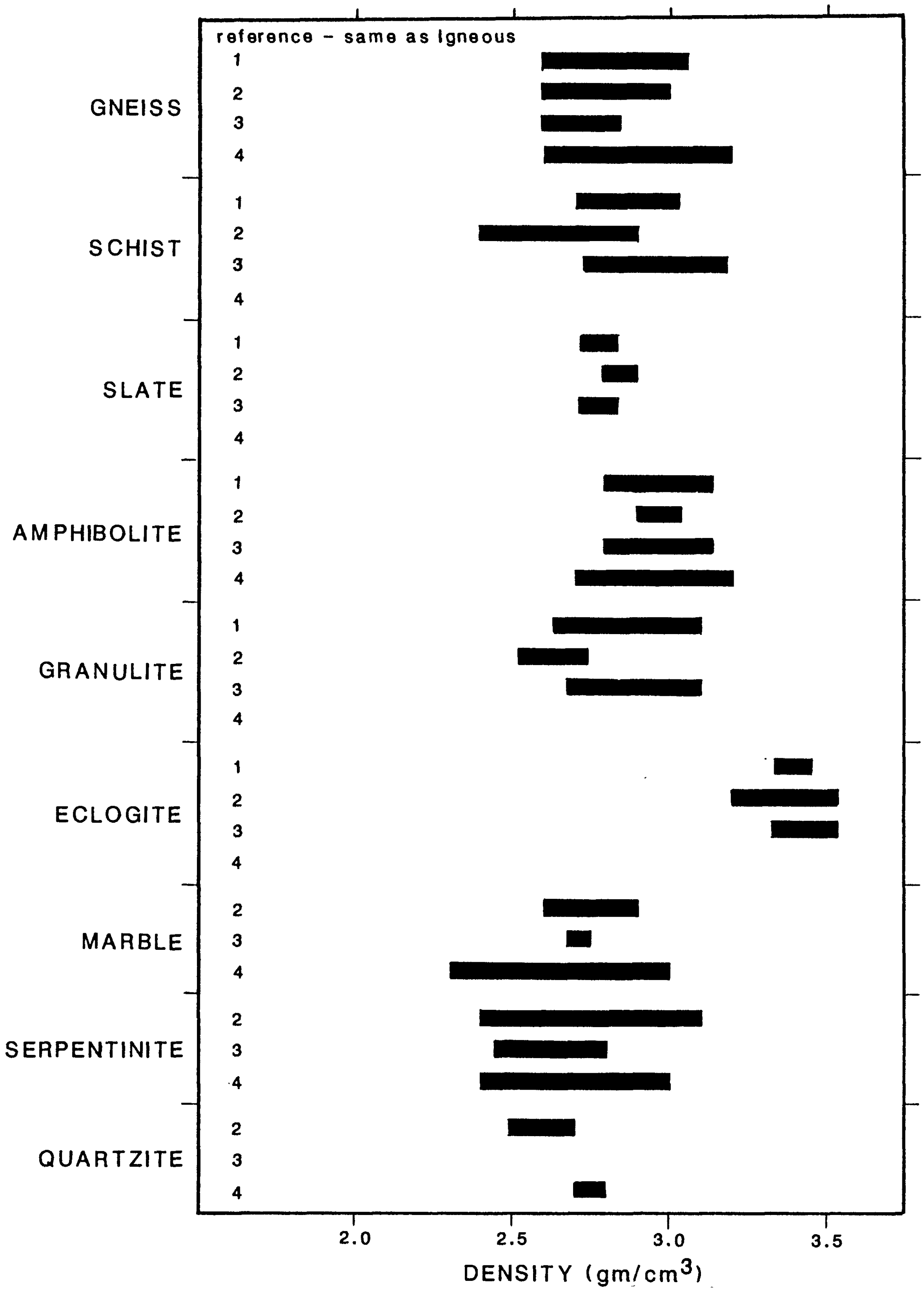


Figure 7. Diagram showing ranges of bulk densities for 9 different metamorphic rock types. References sources are 1. Daly and others, 1966, 2. Telford and others, 1976, 3. Johnson and Olhoeft, 1984, and 4. Mironov, 1972.

There are two types of porosity of primary concern for their effect on rock properties, intergranular, and joint porosity. With the exception of chemical sediments, sedimentary rocks typically have high intergranular porosity. Joint or fracture porosity in all typical rock types is small in magnitude but can be important in its effect on rock properties. Table 3 compares ranges of intergranular and joint porosity for several groups of rocks as given by Keller and Frischknecht (1966).

Borehole logging methods, such as neutron, resistivity and seismic velocity, provide estimates of porosity, but these are primarily used in petroleum work. Direct knowledge of porosity may become increasingly important as the process of in-situ leaching of metallic ores is increasingly utilized.

Porosity is important to the exploration geophysicist principally for the part it plays in affecting density, electrical resistivity, and seismic velocity. Changing from water saturated to dry conditions may decrease densities by 0.5 gm/cm^3, may increase resistivities several orders of magnitude, and may decrease seismic P velocities by a factor of 2 or more. Even low porosity rocks may show significant changes in resistivity and seismic velocities between wet and dry conditions. Christensen (1982) lists a granite with 1.1% porosity whose P-wave velocity decreased by 40% from wet to dry conditions. The effects of water saturation on these other physical properties is covered in the individual sections dealing with these properties.

Table 3. Ranges of intergranular and joint porosity for several rock types. Data from Keller and Frischknecht (1966).

Rock type	*Intergranular porosity %*	*Joint porosity %*
Precambrian igneous and higher grade metamorphic	0-2	0-2
Paleozoic and younger igneous	0-10	0-2
Precambrian sediments and low-grade metamorphic	1-8	0-2
Paleozoic sandstone and shale	5-30	0-1
Paleozoic limestone	2-10	0-2
Paleozoic clastic volcanic	5-30	0-2
Younger sandstone and shale	10-40	0
Younger limestone	4-20	0-2
Younger clastic volcanic	10-60	0

Porosity values for different rock types can be found scattered throughout the physical property literature often in conjunction with density, resistivity, or seismic velocity tabulations. Daly and others (1966) provide porosity ranges and averages for a variety of sediments and Christensen (1982) gives density, porosity, and seismic velocity for a variety of marine sediments and igneous rocks. The data of Daly and others (1966) are summarized in figure 8. These data can be compared with porosity values for sediments given in figure 5. Intrusive and high grade metamorphic rocks typically have porosities of a few percent or less when fresh and unfractured. Fracturing and alteration, however, can increase in-situ porosity significantly, especially where clay minerals are formed. Silicification, on the other hand, often will decrease porosity. As such processes are common during mineralization from hydrothermal systems, their effects on ore and adjacent host rock need to be considered for modeling. Porosities of extrusive rocks can vary widely, but compilations of representative values are not given in the standard references. Johnson (1983) provides some values for these types of rock, ranging from 0% for obsidian to 62.4% for pumice.

It should be remembered that high porosity does not necessarily imply a high permeability for a rock unit. Clays are excellent examples of a lithologic type with high porosity, but very low permeability. There are no generalities that can be made regarding the relationship between porosity and permeability, other than that for rocks of a given porosity the permeability will, in general, decrease with decreasing grain size. Johnson (1983) gives measurements of both porosities and permeabilities made on the same samples of a wide variety of rock types. His data for water available porosity and permeability have been plotted in figures 9a for igneous rocks, and 9b for sedimentary rocks. The wide scatter shown is indicative of the lack of correlation between these two properties. The graphs, figures 9a and 9b, are useful for showing ranges of these two properties as measured on hand-sized specimens.

Magnetic Susceptibility and Remanence

The magnetic properties of rocks depend on the quantity, composition, grain size, and physico-chemical history of magnetic minerals normally present as minor constituents of a rock unit.

Iron ores are the major exception where magnetic minerals can form the bulk of the rock and thus provide a strong target for geophysical exploration. Various authors have provided formulae relating the magnetite content of iron ores and rocks to susceptibility. Werner (Hansen, 1966) derived two expressions, one for Swedish magnetite ore and the other for hematite ore and other rocks. These are, for magnetite ore;

$$k = \frac{Kv}{1+cKv}, \tag{1}$$

where

$$c = \frac{4\pi}{3}\,\frac{1-v^{1/6}}{v} \tag{2}$$

k = ore susceptibility
K = susceptibility of magnetite present
v = volume fraction of magnetite present

and for hematite ore, igneous or sedimentary rocks;

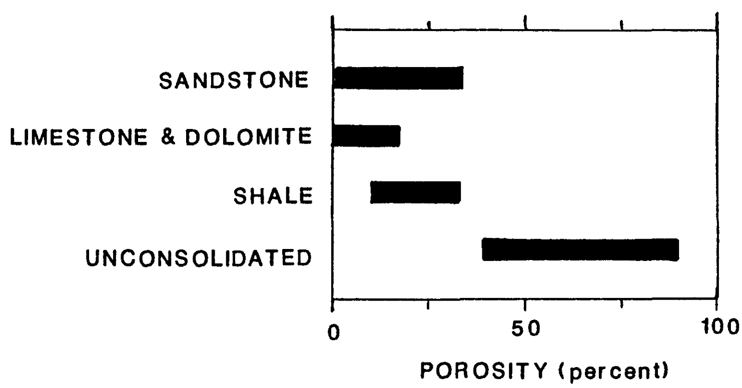
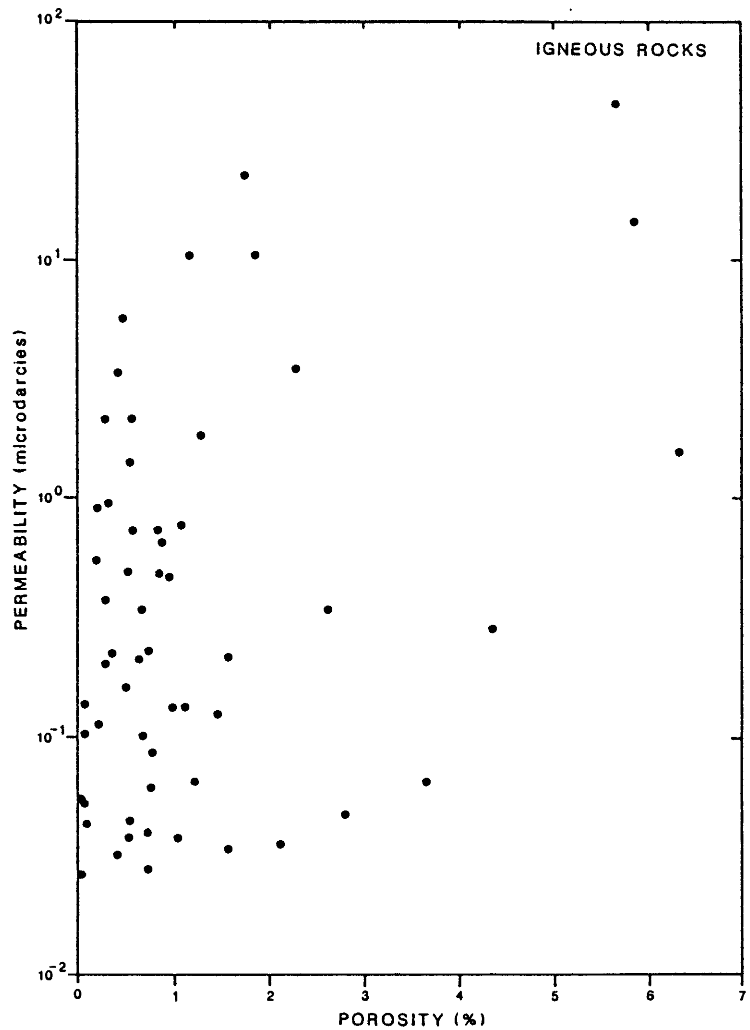
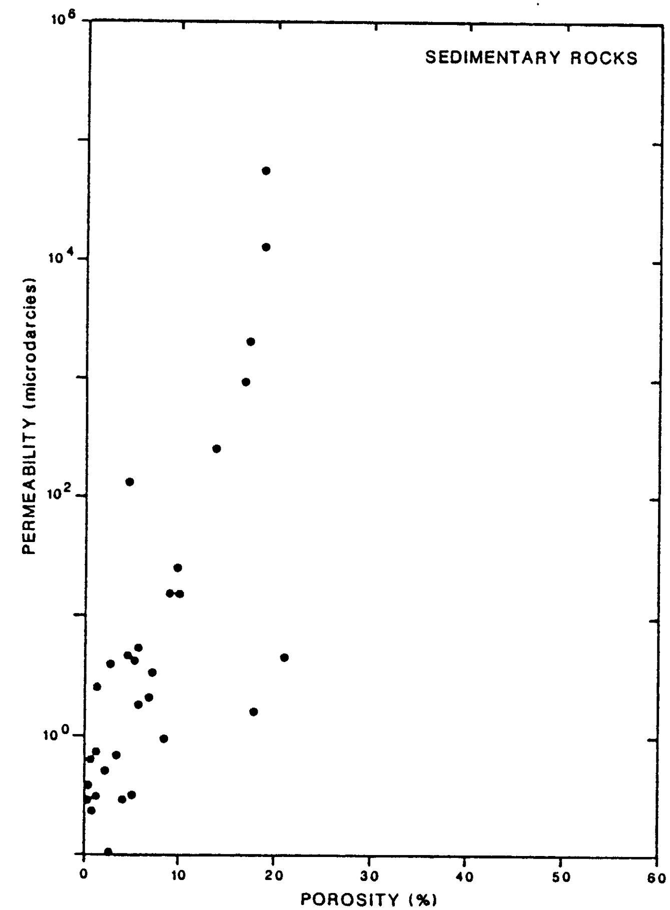


Figure 8. Diagram showing ranges of porosities for sedimentary rocks from Daly and others (1966).





Figures 9a and b. Measured porosity and permeability on a suit of a, igneous rocks and b, sedimentary rocks adopted from Johnson (1983).

$$k = \frac{Kv}{1+ckv} + 120(S-S_1)\times 10^{-6} \qquad (3)$$

where

S = specific gravity of the non-magnetite part of rock
S_1 = specific gravity of the quartz, feldspar, or limestone fraction of the rock.

Grant and West (1965) summarize results from Mooney and Bleifuss (1953) on Precambrian rocks from Minnesota, and from Balsley and Buddington (1958) on Adirondack rocks. Bath (1962) and Jahren (1963) give results on Minnesota iron formation ores, and Dukhovaki (Klichnikov and Benevolenskiy (1970) on granites from the Akchatausk complex of the USSR. The relationships given by these authors are given below and shown in figure 10.

1. Mooney and Bleifuss (1953) $k = 2.89 \times 10^{-3}\ V^{1.01}$ (4)
2. Balsley and Buddington (1958) $k = 2.6 \times 10^{-3}\ V^{1.33}$ (5)
3. Bath (1962) or Jahren (1963) $k = 1.16 \times 10^{-3}\ V^{1.39}$ (6)
4. Klichnikov and Benevolenskiy (1970) $k = 2.38 \times 10^{-3}\ V$, $V<9.6\%$ (7)

These relationships are much simpler than those given by Werner and should provide adequate approximate estimates of susceptibility where magnetite content, V, of a rock is known or can be estimated.

A well known rule-of-thumb is that the more mafic a rock is, the greater is its associated susceptibility (Carmichael, 1982; Grant and West, 1965; Slichter, 1942). This has been expressed as (Carmichael, 1982) basic extrusive > basic intrusive > acid igneous > sedimentary. This relates to the greater magnetic-mineral content of mafic rocks. Where this content is relatively coarse-grained the mafic rock will have a large magnetic susceptibility and if fine-grained it will have a relatively large remanent magnetization. This relationship is documented in various handbooks and texts to which the reader can refer for details. Figure 11 shows the ranges of susceptibilities given by three of these references for igneous rocks. Average values for each group are indicated by a mark on the range bar, and show the increased susceptibility of mafic rocks. The wide range of susceptibilities for a given rock type is also apparent.

Figure 11 shows a wide range of susceptibilities for granites, but does not distinguish between the various varieties of granites. Chappell and White (1974) have defined I-type, for igneous, and S-type for sedimentary, granites based on their chemistry and inferred source rock. Ishihara (1977) has defined ilmenite series and magnetite series granites based on the opaque mineral content (ilmenite-series granites by definition contain less than 0.1% magnetite). These classifications are important in exploration because I-type or magnetite-series granites are generally associated with gold and base-metals while the S-type or ilmenite-series granites relate to tin and tungsten mineralization. A compilation of the variation of susceptibility and other physical properties of these varieties of granites would be desirable, but none is known to the authors. Both the S-type and ilmenite-series granites have low susceptibilities. Based on the definition of the ilmenite series granites, figure 10 can be used to estimate the maximum susceptibility.

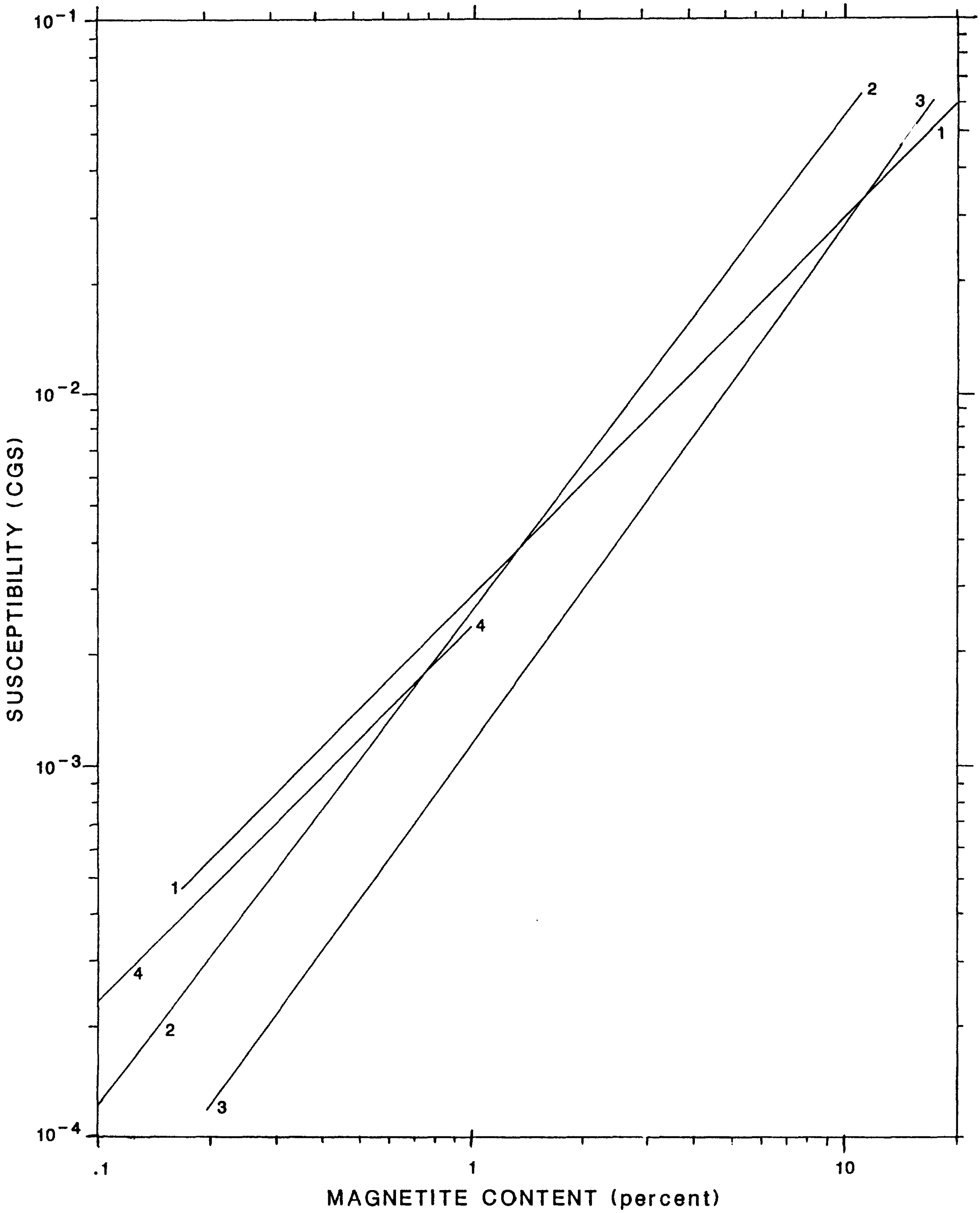


Figure 10. Curves showing empirically derived relationship between magnetic susceptibility and magnetite content from 1. Mooney and Bleifuss (1953), 2. Balsley and Buddington (1958), 3. Bath (1962) or Jahren (1963), and 4. Klichnikov and Benevolenskiy (1970).

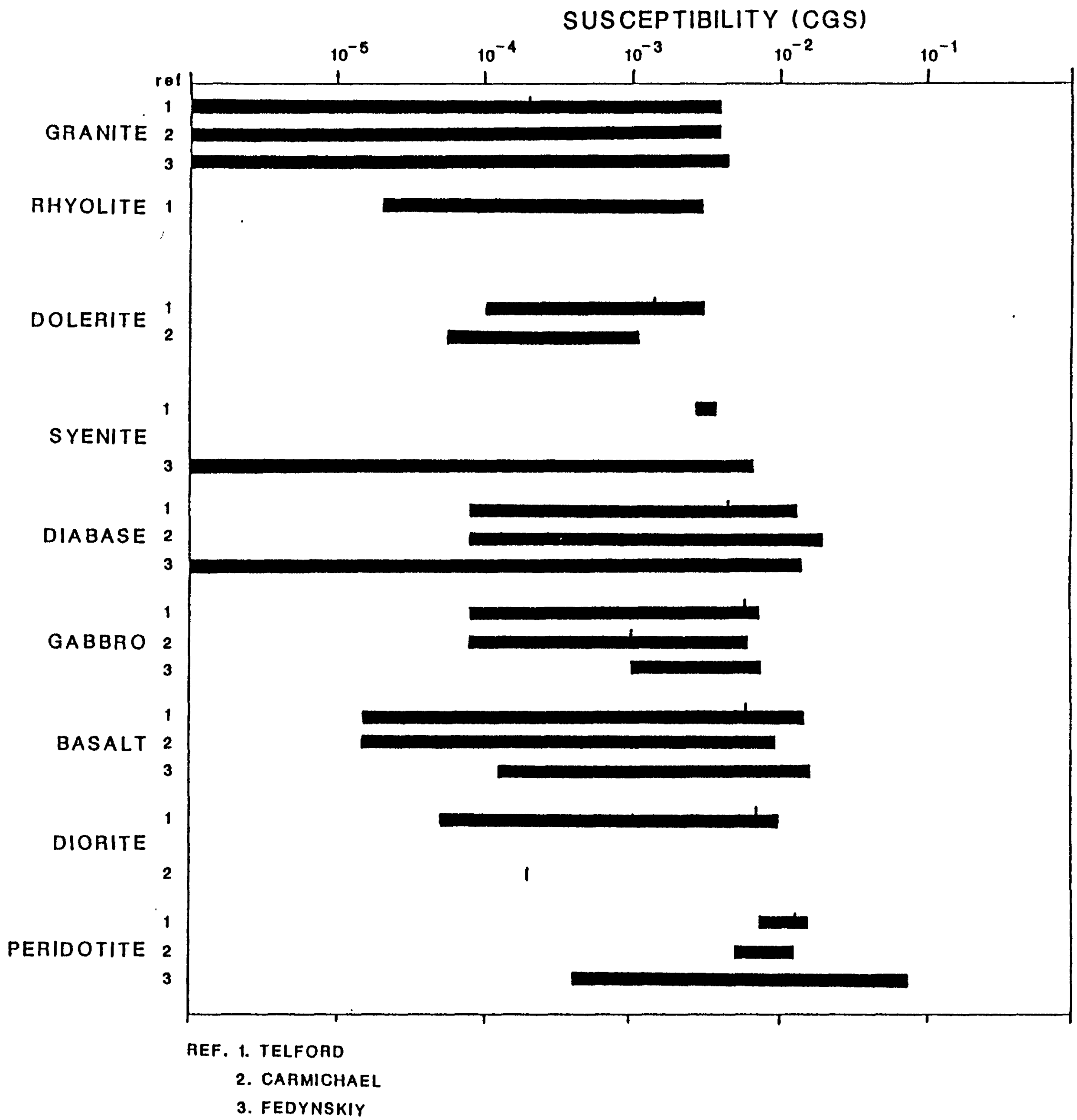


Figure 11. Diagram showing ranges of susceptibilities for various igneous rocks. Reference sources are 1.) Telford and others, 1976, 2.) Carmichael, 1982, and 3.) Fedynskiy, 1967. A bar on the data of Telford and others (1976) indicates the average value.

Figure 12 illustrates the ranges of susceptibilities of sedimentary units taken from four sources identified in the figure. This figure shows the generally lower susceptibilities associated with these rock types. In this figure soils are seen to have a wide range of susceptibilities reaching as high as 10^{-3} cgs/cm^3. Near surface soil zones have been found in many cases to have higher values than lower zones, this is attributed to the formation of maghemite (see Grant and West, 1965 for a summary). Such soils could be a significant source of "geologic" noise in ground surveys.

Besides magnetite ore, several other ores have moderate to high susceptibilities that may permit direct detection by magnetic methods. Figure 13 shows the ranges of susceptibilities reported for eight ores by Carmichael (1982) and Parasnis (1966).

Remanent magnetization, or the related Koénigsberger ratio Q, which is the ratio of remanent to induced magnetization, is not often determined as part of an exploration program for minerals. Thus relatively few data of this important property are available for ore deposits. This is reflected in the compilations of deposit models by the lack of quantitative data. This property is important for interpretation of magnetic data because some anomalies are due in greatest part to remanence. Extrusive volcanic rocks typically show strong remanence or high Q values, but other rocks often show it as well. Hawes (1952) describes studies of the Spavinaw granite in Oklahoma that exhibits a large aeromagnetic anomaly for a granite, and which shows Q values above 100.

Studies of remanence, and values for various rock types have principally been produced from paleomagnetic studies. Such studies have potential for mineral deposit exploration and investigation but are not commonly used in exploration. Hood (1961) has presented a study of the Sudbury basin where he showed that the norite in the mafic complex can be identified by its consistently high Q. Another study by Gross and Strangway (1966) showed how remanent magnetization studies can help address the genesis of iron formation ore.

Details of the various types of remanence, the stability of each, and effects important to the explorationist are beyond the scope of this study. For those wishing to look further into the subject, Grant and West (1965) provide a good overview.

Table 4 lists Q values for selected rock types from Carmichael, 1982 and Kuz'micheva and Diomidova (1968).

Seismic velocity

Investigations of seismic velocities as a function of rock type have been most extensively studied for sedimentary rocks because of the importance of seismic methods in oil and gas exploration. The geophysical literature is replete with details on velocity distribution as functions of porosity, depth of burial, and fracturing in the sedimentary section. There are few in-situ studies of the velocities of ores and host rock environments that are important in minerals investigations. However, laboratory measurements and empirical relationships developed from them do permit reasonable estimates to be made of the velocities to be expected in the minerals environment.

This summary will give ranges of values for seismic P-wave velocities only. Although S-wave techniques are finding increasing applications, they do not appear to have had any significant application yet in minerals

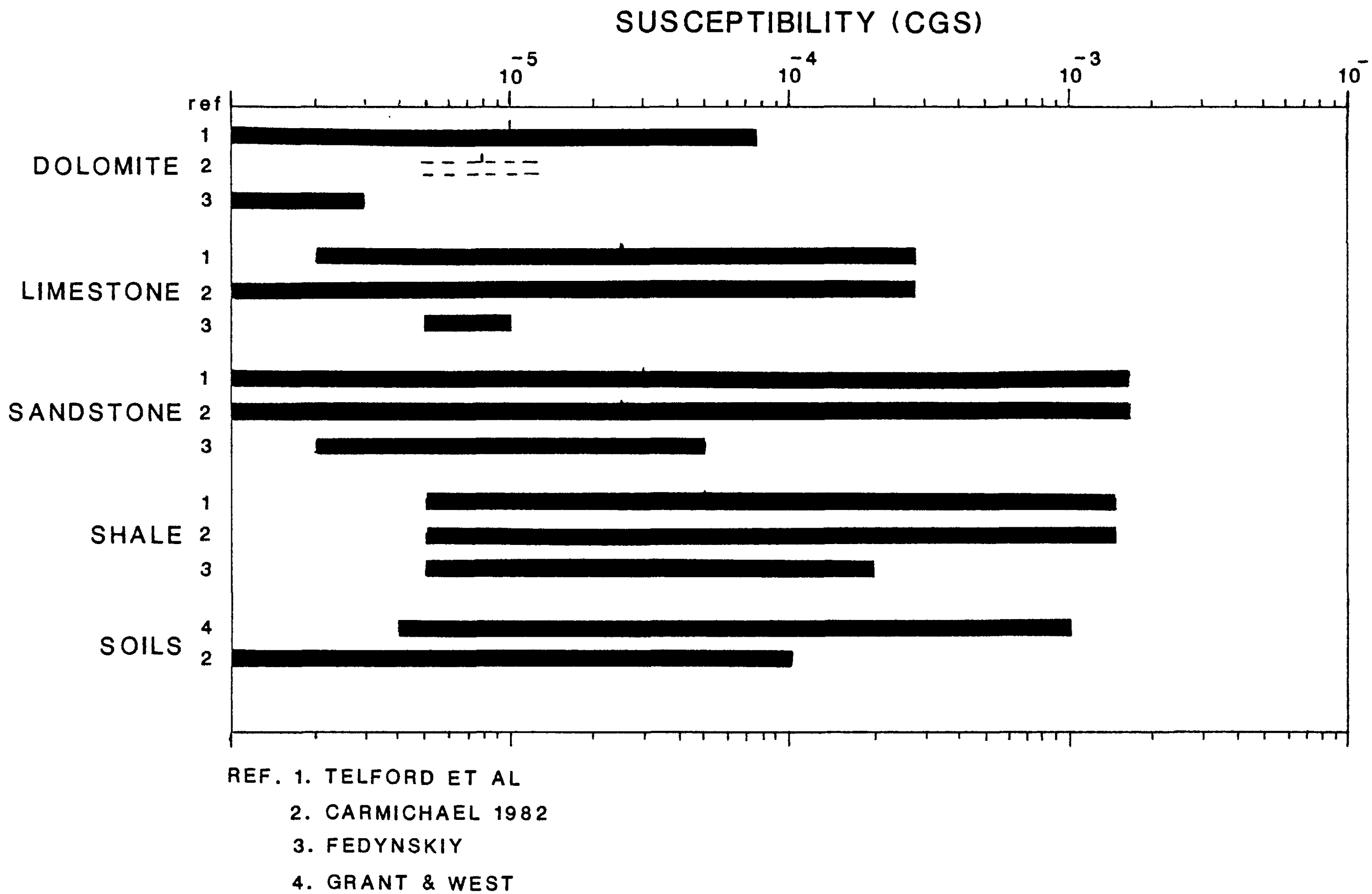


Figure 12. Diagram showing ranges of susceptibilities for various sedimentary rocks. Reference sources are 1. Telford and others (1976), 2.) Carmichael (1982), 3.) Fedynskiy (1967), and 4.) Grant and West (1965).

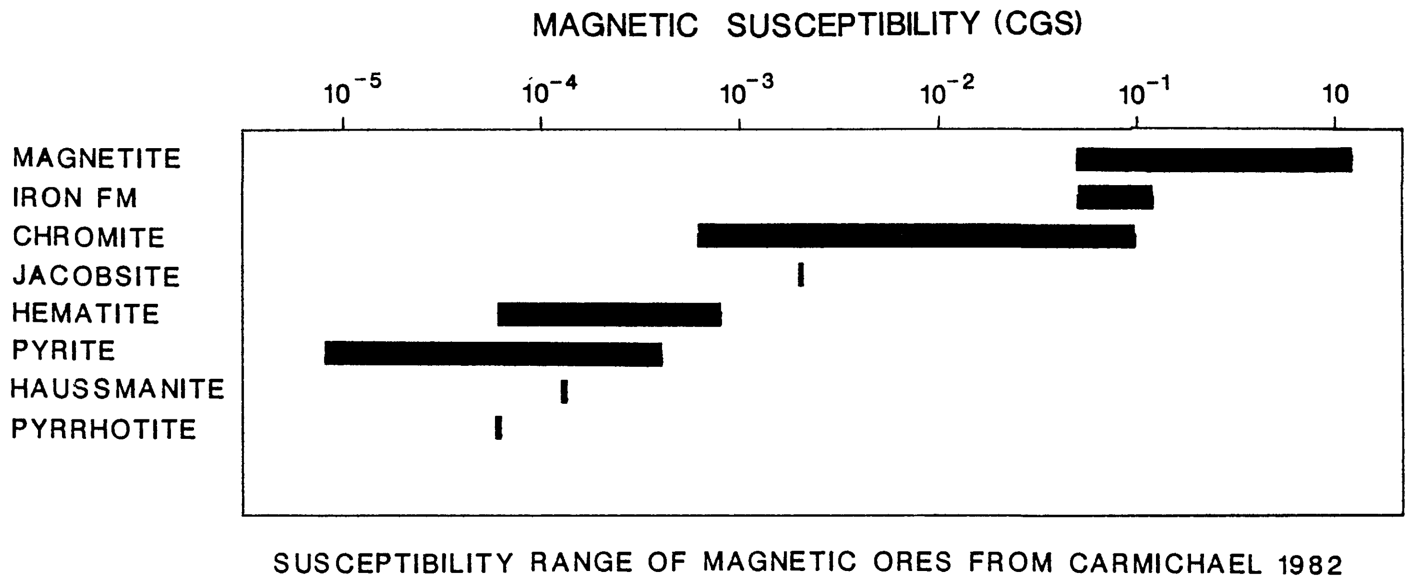


Figure 13. Diagram showing ranges of susceptibilities for eight different ores. Reference sources are Carmichael (1982) and Parasnis (1966).

Table 4. Selected values of the Koenigsberger ratio Q for various rock types from Carmichael (1982) and Kuz'micheva and Diomidova (1968).			
Rock type	***Carmichael range of Q***	***Kuz'micheva and Diomidova mean value***	***range of Q***
Marine sedimentary and shale		5.	
Siltstone	0.02 - 2.0		
Sandstone	1.0 - 4.4		
Granite	0.1 - 1.0		0.3 - 3.4
Granite (Oklahoma)		28.	
Granodiorite	0.1 - 0.2		0.3 - 3.2
Dolerite	2.0 - 3.5		
Diobase	0.2 - 3.5		
Gabbro	1.0 - 9.5		0.3 - 3.05
Volcanics (unspec.)	30. - 50.		
Basalt	1.0 - 160.		
Magnetite ore	1.0 - 94.		
Quartz-diorite			0.3
Granosyenite			0.6 - 1.1

exploration. Domenico (1984) has suggested that the Vp/Vs ratio (γ) and Poisson's ratio (σ) given by

$$\sigma = \frac{0.5(Vp/Vs)^2-1}{(Vp/Vs)^2-1} \tag{8}$$

is characteristic of some sedimentary lithologies with sandstones having Poisson's ratio in the range of 0.17-0.26, dolomites 0.27-0.29 and limestones 0.29-0.33. Thus seismic methods may have potential for lithologic discrimination between sedimentary rocks. Tatham (1982), however, concludes that crack and pore geometry are the prime determinants of the Vp/Vs ratio. Fracture porosity and degree of water saturation in hard rocks also are primary causes for a lowering of both P-wave and S-wave velocity in the near surface. It is not until pressures of about 1 kilobar are obtained that fractures close sufficiently for the seismic velocities to be representative of the true matrix velocity of the rock. Thus, in-situ velocities of ore and host rocks at most mining depths are dependent on the extent of fracturing.

Several authors have related seismic P velocity to density (ρ) (see Grant and West, 1965; Gardner and others, 1974; or Fedynskiy, 1967). The relationships between P velocity and density given in the above references are shown in figure 14. The relationship given by Gardner and others (1974) is

$$V_p = 357.4\rho^4 \tag{9}$$

while Fedynskiy (1967) gives

$$V_p = a\rho - b \tag{10}$$

where a and b are constants given as a = 6, b = -11 (Puzyrev) or a = 7.5, b = 14.8 (Urupov). Within the density range 2.4 to 3.0 gm/cm^3 the velocity functions are all approximately linear.

Woeber and others (1963) provide additional information on the P and S velocity anisotropy and density of a variety of mostly igneous rocks.

A number of papers (Costagna and others, 1985; Domenico, 1984; and Pickett, 1963) empirically relate velocity (V) and porosity (ϕ) for clastic rocks. Pickett proposed the rather simple relationship

$$1/V = A+B\phi \tag{11}$$

where A+B are constants.

Faust (1953) obtained an empirical relationship dependent on the depth of burial D and the formation resistivity (ρ) in the form:

$$V = C\,(D\rho)^{1/6} \tag{12}$$

where C is a constant. However, there is large scatter in the data fitted.

It is doubtful that these empirically determined relationships can be extended to low-porosity igneous or metamorphic rocks.

A theoretical relationship between thermal properties and velocity has also been developed. This will be covered in the section on thermal properties.

In minerals exploration some or all of the host and cover rocks may be in the vadose zone. Under these conditions the degree of water saturation will affect the sonic velocity even of low porosity rocks. Table 5 gives the variation between saturated and air dry samples of igneous and metamorphic rocks and their porosity summarized from Christensen (1982). As can be seen, the variations can be significant for these low porosity examples.

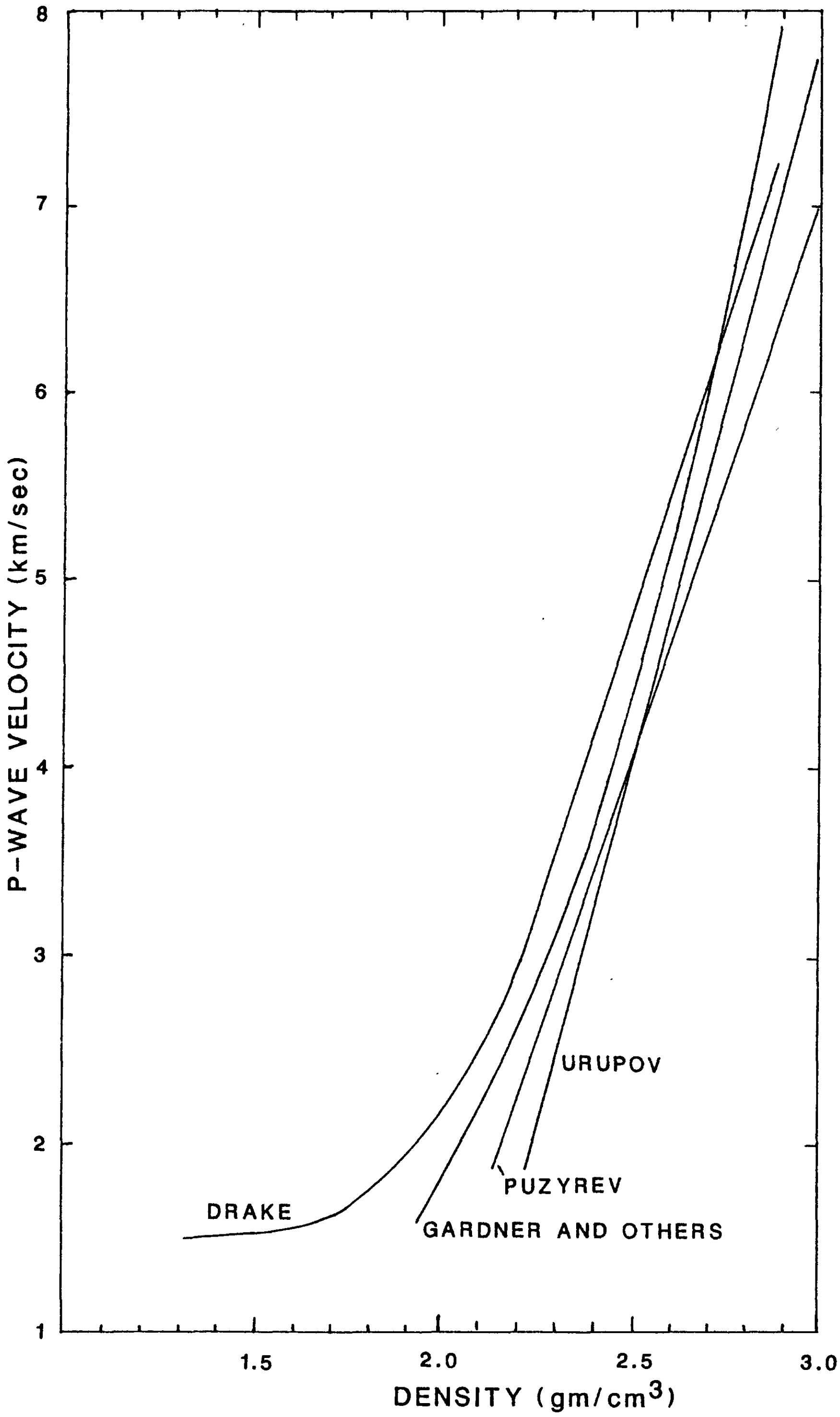


Figure 14. Graph showing empirical relationships between seismic P-wave velocity and rock bulk density from 1. Drake (Grant and West, 1965); 2. Gardner and others (1974); 3. Puzyrev (Fedynskiy, 1967); and Urupov (Fedynskiy, 1967).

Figure 15 illustrates the ranges of P-wave velocities for igneous and metamorphic rocks from several sources. Values shown from Press (1966) are at a confining pressure of 10 bars. Christensen (1982) gives values for a wide variety of igneous and metamorphic rocks, but has not provided a summary. However, in figure 15 the range for peridotite has been extracted, with the low values representative of serpentinized peridotite and higher values correspond to less altered rock. Alteration of other igneous or metamorphic rocks would be expected to similarly lower the seismic velocity.

Figure 16 presents the velocity range associated with sediments and sedimentary rocks. Unconsolidated sediments are seen to have very low velocities.

Table 6 lists seismic velocity ranges of ores or ore minerals from several sources. Note the low velocity of graphite. Jadeite has the highest velocity in the listing, not surprisingly being similar to that of pyroxenite in table 5.

Table 5. Seismic P-wave velocities for selected igneous rocks as a function of water saturation (adopted from Christensen, 1982).

Rock type	*porosity range %*	*Vp range air dry km/sec*	*Vp range saturated km/sec*
Granite	0.6 - 1.8	3.25 - 5.40	5.1 - 6.3
Syenite	0.7 - 1.1	4.10 - 5.45	5.40 - 6.45
Gneiss	0.3 - 0.9	3.65 - 5.20	5.45 - 6.00
Granulite	0.2 - 0.4	4.90 - 5.60	5.45 - 6.00
Diorite	0.6 - 1.2	4.65 - 5.90	5.95 - 6.6
Norite	0.2	6.15 - 7.00	6.70 - 7.10
Pyroxenite	0.2	7.10 - 8.15	7.70 - 8.40

ELECTRICAL PROPERTIES

Resistivity

In the upper kilometer or so where most minerals exploration takes place the resistivity of earth materials is even less representative of lithologic type than are seismic velocities. In this environment earth resistivity is generally determined entirely by rock porosity and the conductivity of the fluid filling the pore space. Because porosity and fluid conductivity are not parameters closely related to lithologic type, especially for igneous and metamorphic rocks, empirical relationships between resistivity and lithologic type show a wide variation. Further complicating the picture are problems related to how representative are laboratory measurements on hand or core samples of in-situ resistivities. Samples may dry prior to measurement requiring rehydration with water that may not be representative of the original connate water, and small samples may not be representative of the effect of fractures in the bulk. In part because of these problems, several authorities do not even provide summary results of resistivities of wet igneous or metamorphic rocks as a function of specific lithologic types (Keller, 1966, Keller and Frischknecht, 1966, Keller, 1982, and Olhoeft,

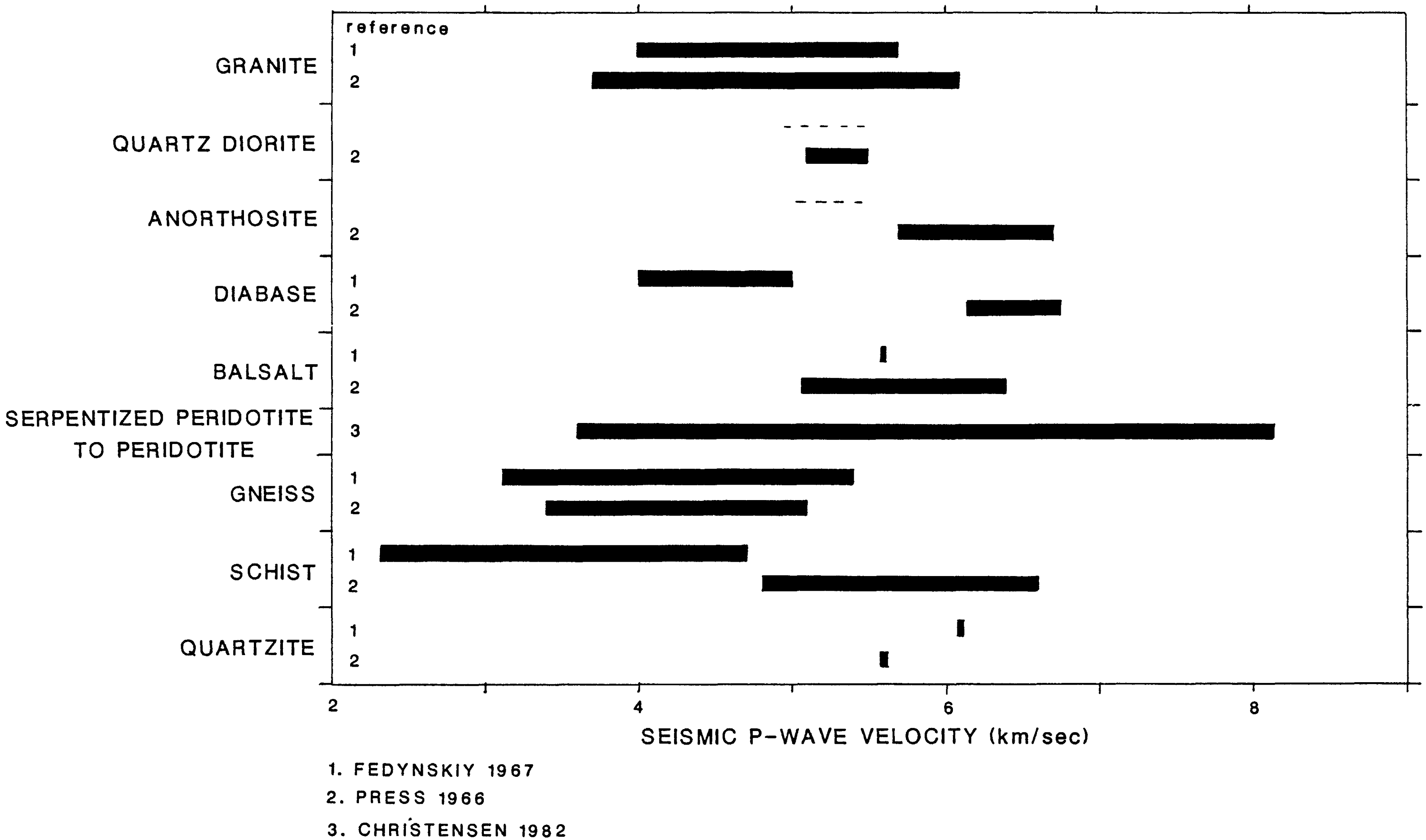


Figure 15. Range of P-wave velocities for selected igneous and metamorphic rocks. Reference sources are 1. Press (1966), 2. Fedynskiy (1967); and 3. Christensen (1982).

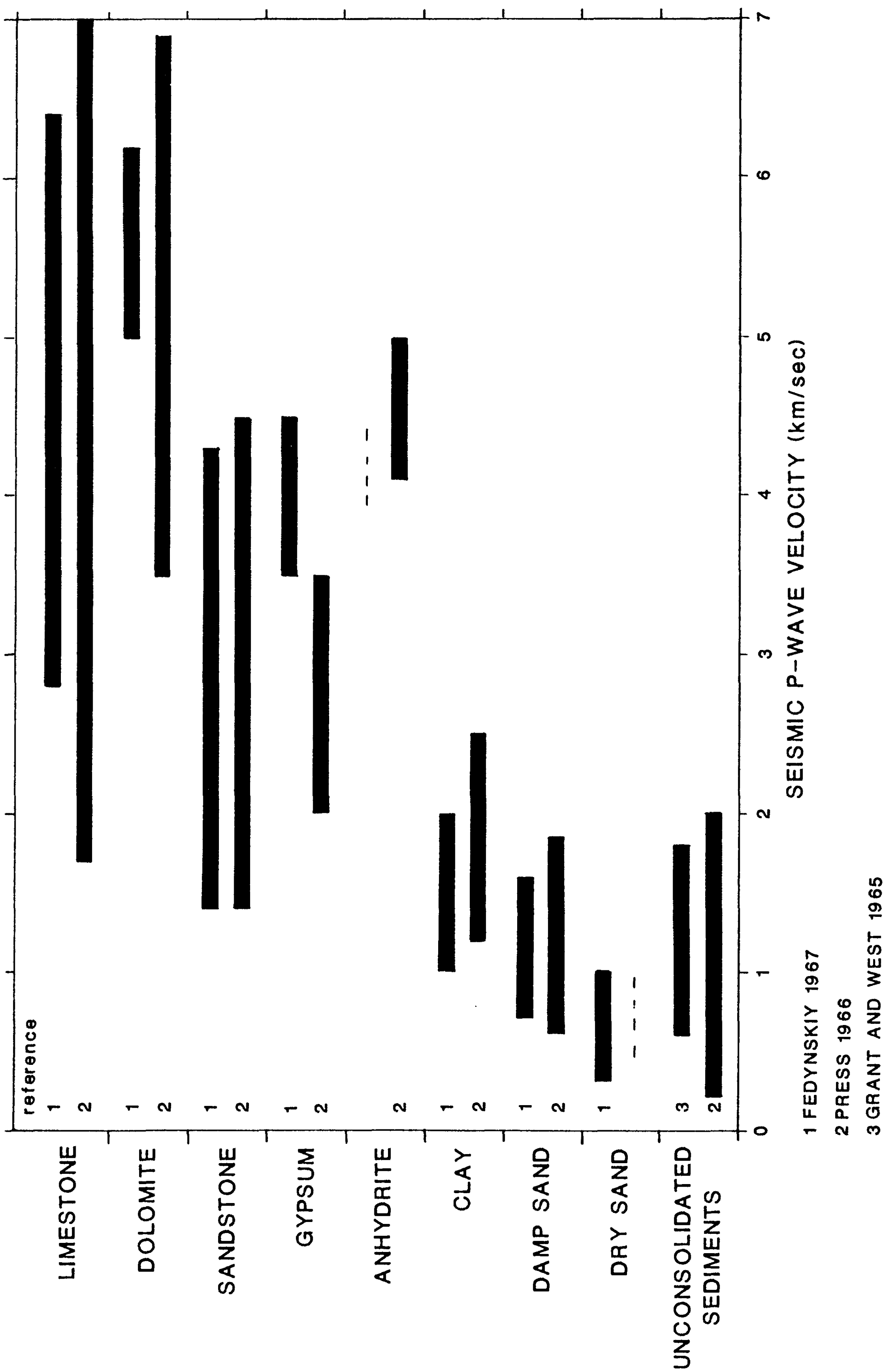


Figure 16. Range of P-wave velocities for selected sedimentary rocks. Reference sources are 1. Press (1966); 2. Fedynskiy (1967) and 3. Grant and West (1965).

1981). Sedimentary rocks because of their importance to the petroleum industry are more extensively studied.

Table 6. Seismic P- and S-wave velocities for selected ores and ore minerals from Woeber and others (1963) and Christensen (1982).

Ore mineral	*density gm/cm³*	*Average P velocity km/sec*	*Average S velocity km/sec*
graphite	2.16	3.06	1.86
limonite	3.55	5.36	2.97
pyrhotite	4.55	4.69	2.76
pyrite	4.81	7.69	4.78
magnetite	4.81 - 6.77	4.18 - 4.87	1.97
hematite	4.93 - 5.00	6.82 - 7.72	3.84
siderite	3.57	7.01	
magnesite	2.80 - 2.97	7.11 - 8.12	
jadeite	3.18 - 3.33	8.21 - 8.67	
anhydrite	2.63 - 2.93	4.90 - 2.29	
gypsum	2.29	4.95	
halite	2.16	4.13	

Illustrative of the broad range of resistivities measured on rocks and the lumping of major lithologic types are figures given by Grant and West (1965) and similar illustrations from Sumner (1976), that were reproduced by Hallof (1980). These are shown in figure 17 redrafted to the same scale for comparison. The figure shows resistivities covering a span of ten orders of magnitude, 10^{-1} to 10^{9} ohm-meters. The upper four decades represent measurements on laboratory samples of low porosity or very dry samples, and are not representative of in-situ bulk resistivities which will rarely exceed 10^{5} ohm-m. Again, this emphasizes the need for caution when applying laboratory results to the field situation.

Below the water table, and in rocks where electronic conduction processes due to metallic luster sulfide minerals can be neglected, conduction is controlled by migration of charged ions in the pore water. The charged ions can be considered to be particles moving under the force of an applied electric field in a viscous fluid. The current then is proportional to the number of ions, ion size, water viscosity, and electric field. Rock resistivity then is related principally to the concentration of dissolved and ionized species in solution, and water viscosity since ion size does not vary appreciably. Viscosity is a factor when the pore water temperature changes. Increasing temperature decreases the viscosity and, decreases the resistivity.

Although the connate water resistivity and porosity control the resistivity of most rocks, connate water resistivity typically does not vary

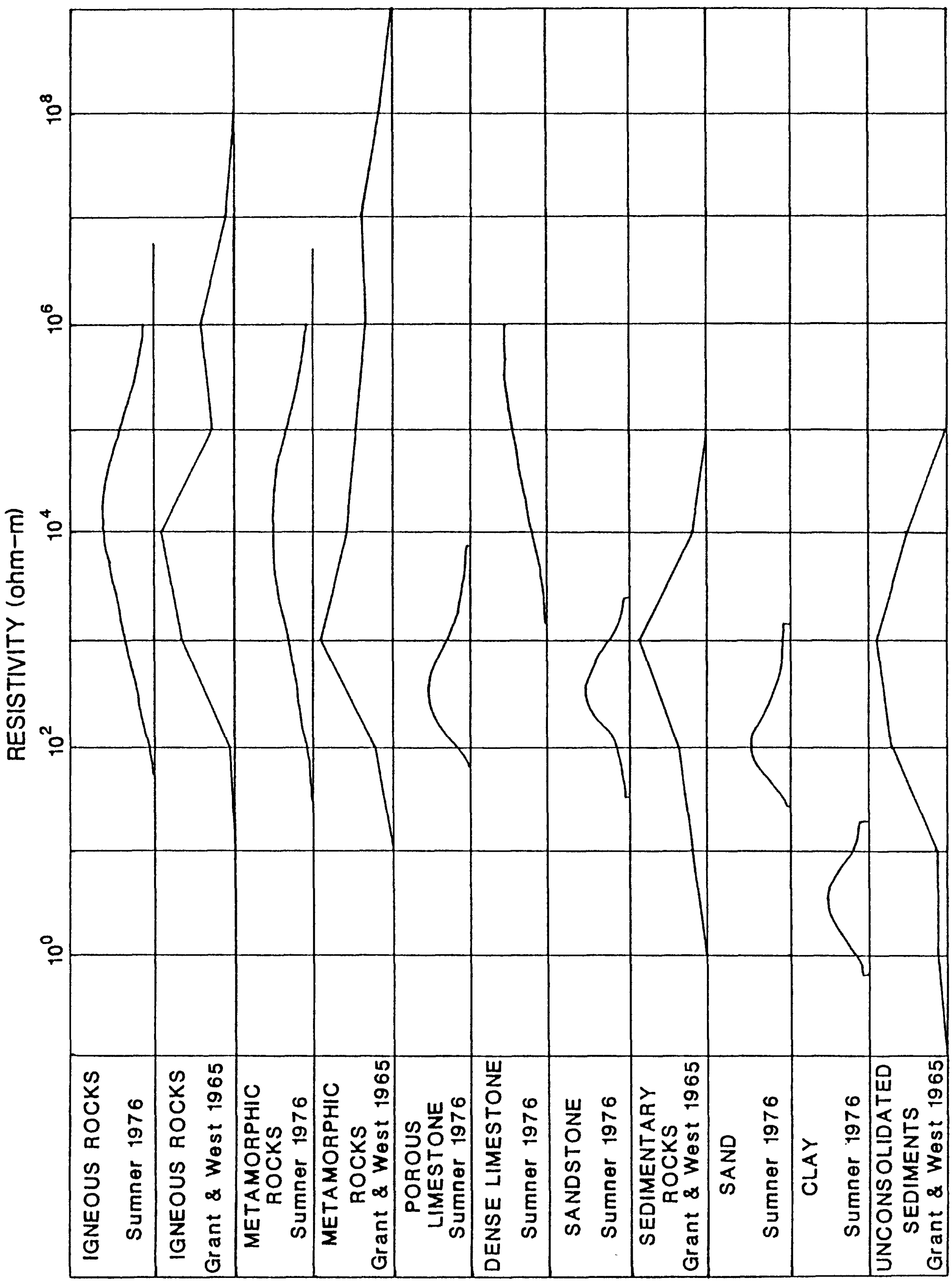


Figure 17. Distribution diagrams of resistivity values for several types of rocks from Sumner (1976) and Grant and West (1965).

over a wide range. Table 7 from Keller and Frischknecht (1966), gives average values of connate water resistivity observed for a number of regions and lithologies. In general, shallow waters and waters from crystalline rocks have the highest resistivities, while deeper waters, or those from sediments, show the lowest resistivities. The average resistivity range is seen to be a little over two orders of magnitude.

The more extensive studies of sedimentary rocks have resulted in an empirical relationship for sediments called Archie's law that relates resistivity (ρ) to porosity (ϕ) and the connate water resistivity (ρ_w). The relationship is:

$$\rho = A\rho_w\phi^{-m} \tag{13}$$

where A and m are empirical constants dependant on the type of sediment. Keller and Frischknecht (1966) give values of A and m for a number of sediments in which A varies from 0.47 to 2.3 and m from 1.64 to 0.23. Where reasonable estimates of sedimentary host or cover rock porosity and connate water resistivity can be made, Archie's law can be used to provide resistivities for modeling.

In the vadose zone, rock resistivities do not increase in direct proportion to the degree of undersaturation. Nonlinear effects result from the presence of a continuous film of water between grains at larger values of partial saturation, and from surface conduction effects. Keller and Frischknecht (1966) provide further details on these effects.

Desert soils and alluvium exhibit an interesting paradox in the vadose zone where the dryness might be expected to result in increased resistivity over that of soils in moister regions. However, most desert soils show resistivities in the range of 10-100 ohm-m, about an order of magnitude lower than soils from regions with average rainfall. This is due to the compensating effect of increased salt content in the residual moisture held by the soils.

Table 7. Connate water average resistivities from various regions and lithologies (from Keller and Frischknecht, 1966).

Water Source	***Average Resistivity (ohm-m)***
European igneous rocks	7.6
South African igneous rocks	11.0
South African metamorphic rocks	7.6
Precambrian Australian metamorphic rocks	3.6
Pleistocene to recent European sedimentary rocks	3.9
Pleistocene to recent Australian sedimentary rocks	3.2
Tertiary European sedimentary rocks	1.4
Tertiary Australian sedimentary rocks	3.2
Mesozoic European sedimentary rocks	2.5

Table 7. Connate water average resistivities from various regions and lithologies (from Keller and Frischknecht, 1966).	
Paleozoic European sedimentary rocks	0.93
Oil field chloride waters	0.16
Oil field sulfate waters	1.2
Oil field bicarbonate waters	0.98
Jurassic Colorado-Utah USA waters	1.8
Cisco series Texas, USA. waters	0.061
Pennsylvanian, Oklahoma, USA waters	0.062

For a given rock type the resistivity normally increases with age and depth of burial, principally because of the associated decrease in porosity. Keller and Frischknecht (1966) give a table of resistivity ranges for rocks of various ages and types that shows a typical change of 1.5 to 2 orders of magnitude increase from Quaternary to Precambrian for the same type of rock. These results are shown graphically in figure 18. The five orders of magnitude range shown here is more representative of in-situ values measured in the earth.

The presence of electronic conductors, such as graphite and the metallic luster sulfides, do not normally contribute to the lowering of host and cover rock resistivities. Their presence as ore or in association with ore is often what makes an ore body detectable by electrical methods. These situations are covered in the individual models under properties of the deposit itself. Occasionally, a situation is encountered where sulfides or graphite may be deposited as thin continuous films in a network along the fracture planes of rocks. Stockwork deposits are a good example of where this may occur. When this happens, a few percent of these electronic conductors can cause a dramatic lowering of resistivity suggestive of much greater mineral content. Graphite is particularly notable for its ability to reduce rock resistivity when present in small quantity along fracture planes (Grant and West, 1965).

Although the above discussion identifies many problems that appear to make the application of resistivity methods difficult in minerals exploration, in practice there are many straightforward uses. Mafic rocks are found to weather to lower resistivity colluvial material than felsic rocks. This permits mapping of such units by airborne electromagnetic methods as demonstrated by Palacky (1986) in Brazil, and to exploration for kimberlite pipes (Gerryts, 1967). Hydrothermal ore deposits often have large alteration haloes extending for some distance into various host rocks. Alteration processes often increase the porosity of the host rock during argillization and propylitization, or if silicification occurs, the porosity may decrease. Either of these processes can provide a clear resistivity contrast that can be detected. Deposits often are controlled by fracturing and faulting which causes increased host rock porosity that can be detected by the associated lower resistivity. However, in practice, within given districts, lithologic units often will have uniform electrical properties that do permit them to be mapped.

Sedimentary and metasedimentary rocks, especially those that show a strong orientation of pore spaces, or that contain interbedded units, may exhibit significant resistivity anistropy. Keller and Frischknecht (1966)

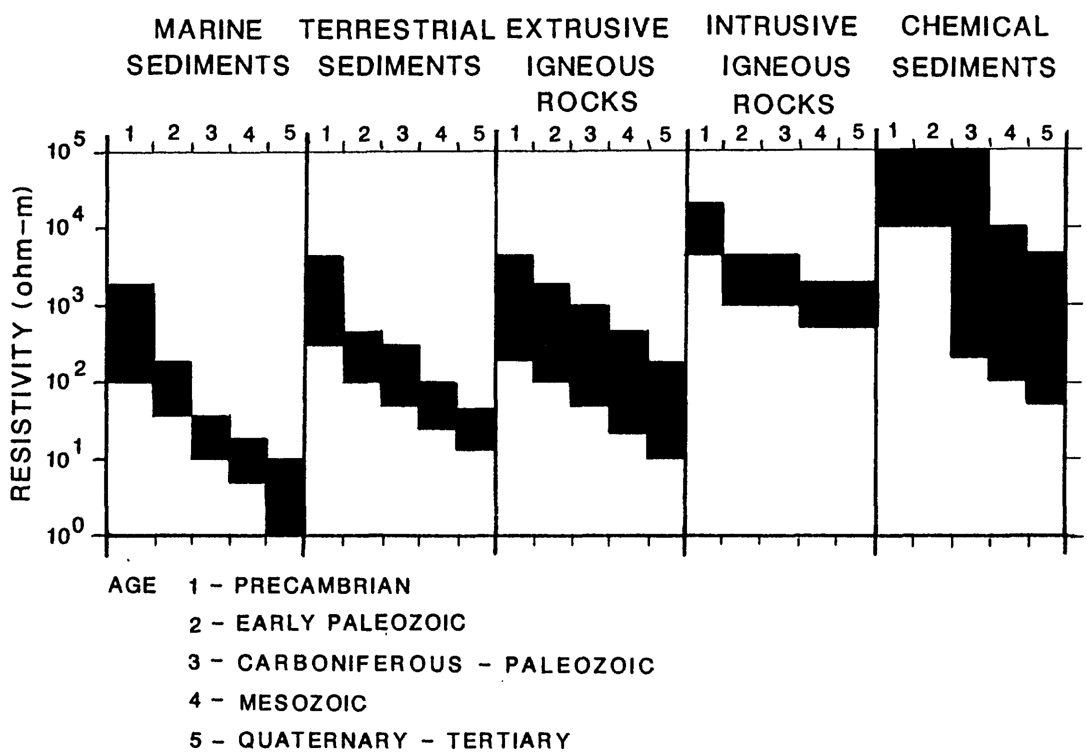


Figure 18. Diagram showing variation of resistivity as a function of age for marine and terrestrial sediments, extrusive and intrusive igneous rocks and for chemical sediments (from Keller and Frischknecht, 1966).

give values of coefficients of anisotropy for various rock types, some of which show resistivity differences exceeding a factor of 10. Crystalline rocks in which a preferred fracture orientation exists also may show significant anisotropy.

Figure 19 shows ranges of resistivities for nine types of sedimentary rocks given by Fedynskiy (1967) and Telford and others (1976). This figure can be compared with results for similar sedimentary rocks given in figure 17.

Figure 20 shows resistivity ranges for twelve types of igneous and metamorphic rocks also taken from Fedynskiy (1967) and Telford and others (1976). On average, these rock types show a wide range of overlap in values.

Telford and others (1976) also give a listing of resistivities observed on a variety of metallic ores, to which the interested reader is referred.

IP Effect

The presence of polarizable minerals, principally metallic-luster sulfides, graphite, clays, and zeolites, within a rock mass are measured by the induced polarization method. Several electrochemical processes, occurring within the earth during passage of electrical current, have been proposed as the cause for the IP phenomenon (Keller and Frischknecht, 1966; Sumner, 1976; Telford and others, 1976; Hallof, 1980). However, some aspects of the conventional theory have recently been questioned (Fink, 1979). But, no matter what the details of the source mechanisms, the IP method is a very effective tool in exploration where sulfides are present. This is especially the case for disseminated sulfides because the polarization phenomenon is proportional to the active surface area of the sulfides within the pore spaces of the host rock.

The polarization response for sulfides is a complex function of a number of parameters including, host rock porosity, grain size, degree of water saturation, current density, and sulfide content. Sumner (1976) states that sulfide content as low as 0.5% may be detected. Membrane polarization response, related to clay or zeolite content, is dependent on the type of clay or zeolite, the cation exchange capacity, and how the minerals are arranged within the pore spaces of the host rock. Keller and Frischknecht (1966) give a qualitative explanation for this type of polarization phenomenon, and show that maximum polarization occurs for clay content generally less than 10% of pore volume. In the case of montmorillonite, the maximum occurs near 0.5% of pore volume. Readers should consult the references cited for details of these complex phenomena.

Equipment used to measure the polarization response of rocks operates in either the time or frequency domain, with several variations of these two methods in use. No measurement standards are in place, and due to instrumentation differences a variety of parameters have been used to measure and report the magnitude of polarization response. These various parameters are not dimensionally equivalent. The more common parameters are frequency effect (FE) or percent frequency effect (PFE), chargeability (M), phase angle (ϕ), and metal factor (MF). To add to the confusion, the above terms are not defined consistently. For example, the frequency effect (FE) which is the normalized difference in resistivity measured at two separate frequencies has been defined as:

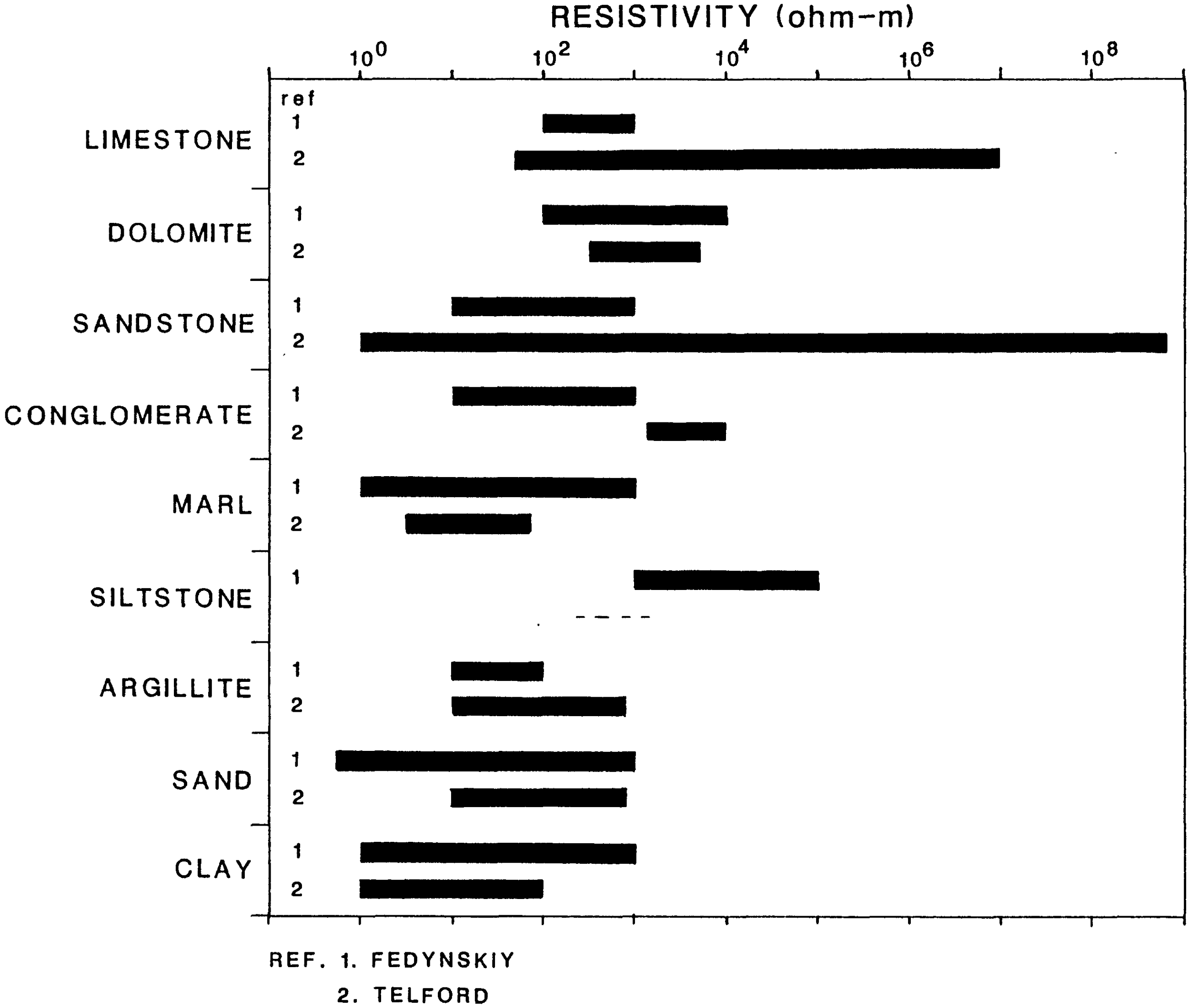


Figure 19. Ranges of resistivities for selected sedimentary rocks from Fedynskiy (1967) and Telford and others (1976).

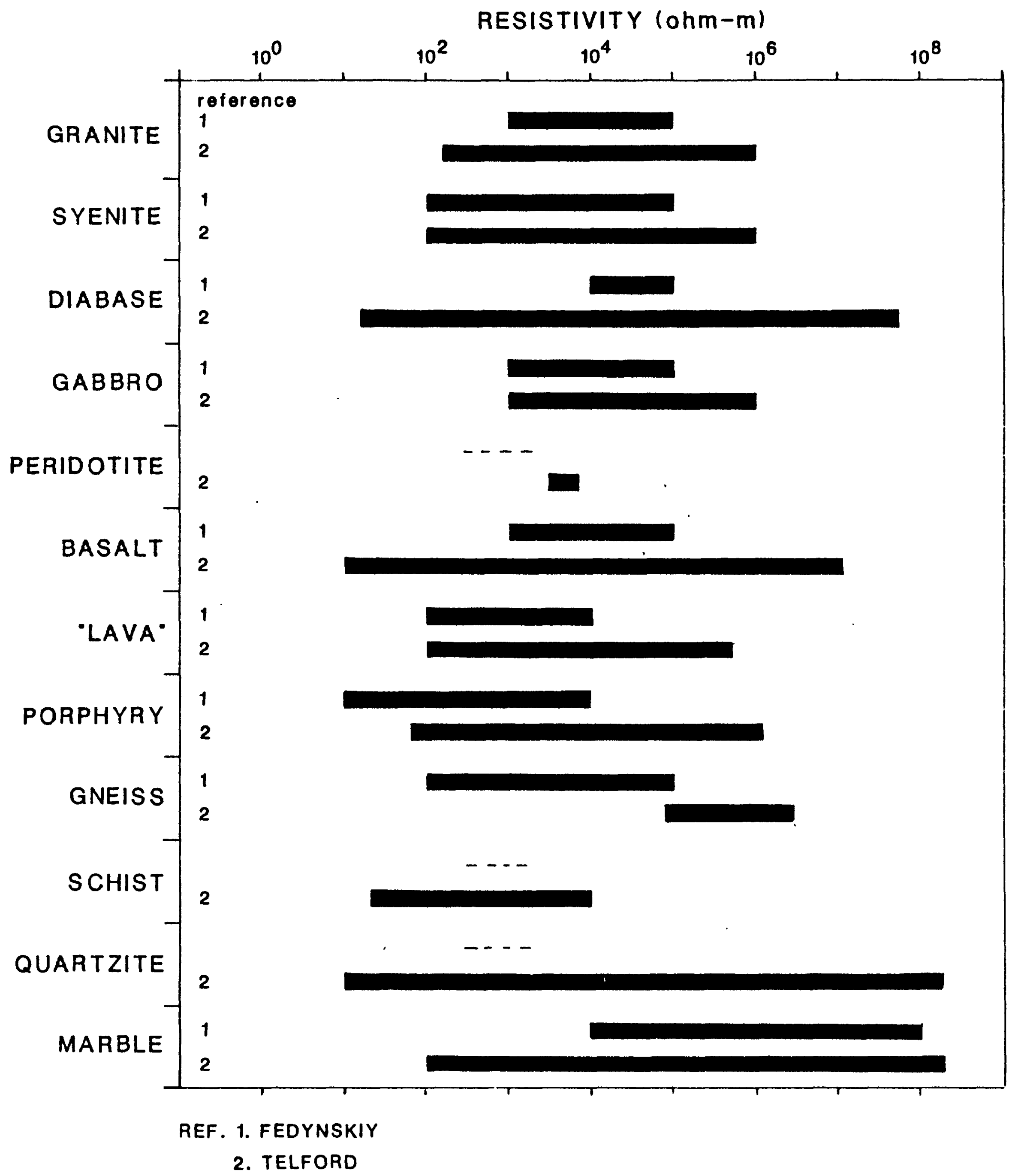


Figure 20. Ranges of resistivities for selected crystalline rocks from Fedynskiy (1967) and Telford and others (1976).

1. $FE = \frac{Q_{f1} - Q_{f2}}{Q_{f2}}$ Telford and others, 1976; Sheriff, 1991; Rogers, 1966; Madden and Marshall, 1959. (14)

2. $FE = \frac{Q_{f1} - Q_{f2}}{Q_{f1}}$ Brant and others, 1966 (15)

3. $FE = \frac{Q_{f1} - Q_{f2}}{\sqrt{Q_{f1}\, Q_{f2}}}$ Keller and Frischknecht, 1966. (16)

where frequency $f_2 > f_1$

Two or more frequencies may be measured, but should always be specified.

The metal factor (MF) or metal conduction factor has been defined as

4. $MF = (2\pi\ 10^5)\ \frac{Q_{f1} - Q_{f2}}{Q_{f1}\, Q_{f2}}$ Telford and others, 1976; Brant and others, 1966; Madden and Marshall, 1959. (17)

5. $MF = \frac{10^5\ FE}{Q_{f1}}$ Rogers, 1966. (18)

6. $MF = \frac{2000\ PFE}{Q_a}$ (Q_a not specified at f_1 or f_2) Sumner, 1979. (19)

7. $MF = 10^3 \times \frac{M(\text{chargeability})}{Q_a}$ Witherly and Vyselaar, 1990. (20)

In the time domain, chargeability is commonly measured, and this has been defined as simply the ratio of the primary voltage (V_p) to the secondary (decay) voltage (V_s) measured at a specific time after current shut-off and is dimensionless.

$$8.\quad M = \frac{V_s}{V_p} \quad \text{Brant, 1966; Telford and others, 1976; Sheriff, 1991; Sumner, 1976.} \tag{21}$$

The secondary voltage may be measured at one or several times during the decay.

Chargeability has also been defined as the time integral of the voltage decay curve between two specified times normalized to the primary voltage

$$9.\quad M = \frac{1}{V_p}\int_{\tau_1}^{\tau_2} V_\tau \, dt \quad \text{Sheriff, 1991; Sumner, 1976.} \tag{22}$$

Historically, frequency domain measurements were reported in FE or PFE, but more recently instrumentation has been developed to give the amplitude and phase angle ϕ between transmitted and received voltage as a measure of polarization response. Typically phase response is reported in multiradians at a particular frequency. When such instrumentation is used over a wide frequency range the method is referred to as spectral IP or the complex-resistivity method. This method provides far more information on electrical relaxation phenomena (information on electrical relaxation phenomena may be found in Pelton and others (1978) and Olhoeft (1981)) in the earth than more convention IP techniques in which measurements are taken at only two or three frequencies or time delays. The increased information permits, in some cases, discrimination between polarizable minerals present in the rock. Sumner (1976), Pelton and others (1978) and a series of papers in Fink and others (1990) provide details on the complex resistivity method.

Although theoretical and empirical relationships between the various polarization measures have been given, comparisons between published results are often impossible because the frequencies or times over which measurements were made are often not stated. Thus, compilations of physical properties such as presented here become impractical because reported IP parameters cannot be made commensurate. Part of the problem is that IP investigators have been reporting the response measured by individual instruments, rather than a direct measure of a rock property. A step towards reporting of a direct rock property is calculating of the Cole-Cole parameters derived from IP measurements (Pelton and others, 1978). However, a field test comparing Cole-Cole parameters derived from frequency domain and time domain systems (Johnson, 1990) showed correspondence between areas of high and low-values, but relatively poor quantitative agreement.

Thus, for cover rock polarizability properties we do not present comparisons between tables derived from various sources. In figure 21 we give ranges of IP response presented as metal factor for several types of lithologies and ores from Madden and Marshall (1959) which were derived from both laboratory and field work. The definition of MF is as given in equation 17 at frequencies of 10 Hz and "dc". This figure serves to show the difference between lithologies and relative range of values of this particular

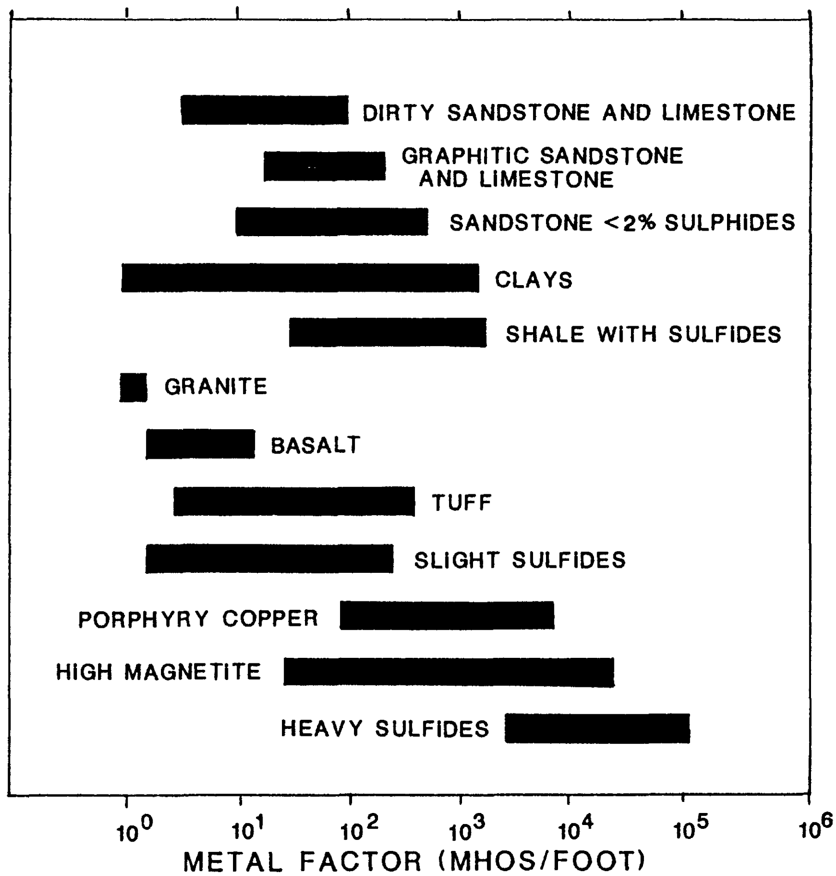


Figure 21. Range of IP response shown as metal factor for several types of rocks and ores from laboratory and field measurements from Madden and Marshall (1959).

polarization measure. Telford and others (1976) also give several tables showing typical polarization responses of geologic materials. When using the accompanying models descriptions, the problems of measures of polarization response need to be kept in mind. Where quantitative values are quoted, the reader wishing to use these should go to the original reference for clarification.

Electrokinetic coupling coefficient

Self potentials arising from current flow through a natural electrochemical cell in the earth, related to oxidation-reduction reactions are not directly related to a specific rock property. Thus, no listing of properties can be given relevant to this type of self potential source. Also, water moving through the earth will generate a self potential signal by an electrokinetic mechanism. Normally, voltages generated by the process would be noise that perturbs the signal from a sulfide deposit. However, water flow controlled by fractures and faults may generate a self potential field than will define the structure. The self potential generated is determined by the electrokinetic coupling coefficient of the rocks. Johnson (1983) gives laboratory values for a variety of rock types, and summarizes results for various groups of rocks. His summary results are shown in figure 22 for the mean of each group and one standard deviation to each side.

Johnson (1983) in his summary did not include rocks of high permeabilities which also showed very high coupling coefficients. Values of 34 to 226 mv/atm. were observed on samples of pegmatite, pumice, tuff, scoria, tuffa, and coquina. Only two marble samples, and one limestone sample gave measurable negative values.

OPTICAL PROPERTIES

Spectral Reflectance

Spectral reflectance is a physical property of materials that describes how light in a continuous electromagnetic spectrum interacts with the material. In simple terms, incident light in the optical portion of the spectrum (visible and near-infrared) can be: (1) transmitted, (2) absorbed, and (3) reflected such that,

$$\tau + \alpha + \rho = 1, \tag{23}$$

where

τ = transmissivity,
α = absorptivity, and
ρ = reflectivity.

For nearly all naturally occurring materials at the earth's surface, the transmisivity is 0, so that,

$$\alpha + \rho = 1. \tag{24}$$

Absorption and reflection are sensitive to the wavelength of the incident light, that is, reflectivity and absorptivity are spectral properties. The amount of light reflected at any given wavelength ($\rho\lambda$) is a function of the elemental content and molecular structure of the materials, and measurement of reflected light across the visible and near-infrared

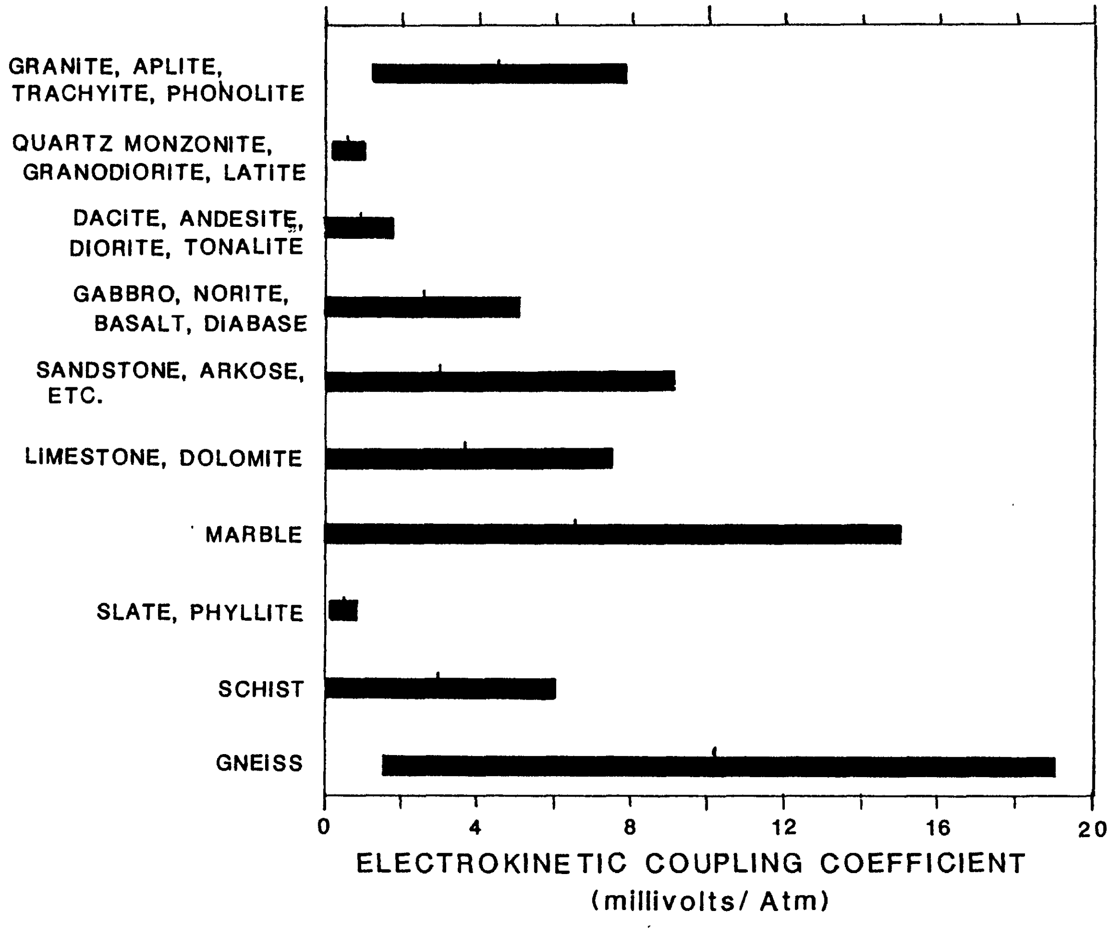


Figure 22. Electrokinetic coupling coefficients for selected lithologies from Johnson (1983). The tick indicates the mean for each lithology, and the bar extends one standard deviation on each side.

spectrum can provide information diagnostic of minerals species. Spectral curves (figure 23) show the proportion of reflected to incident light as a function of wavelength. The wavelength of reflection minima and their size and shape provides the key information for identifying minerals from their spectral reflectance characteristics. An excellent detailed discussion of electromagnetic radiation and its interaction with matter is contained in Hunt (1980).

Broad band remote sensing systems, such as the experimental airborne Thematic Mapper Simulator (TMS) and the commercially available Landsat Thematic Mapper (TM) satellite system, provide local to regional geographic coverage. But the relatively broad bands of these systems only permit the detection of broad mineral groups with similar basic spectral reflectance characteristics, such as iron oxides, hydroxyl-bearing silicates (clays and micas), and carbonates (Hunt and Salisbury, 1970, 1971; Hunt, Salisbury, and Lenhoff, 1971a, 1971b, 1972, 1973). Many of the minerals in these mineral groups are found in hydrothermally altered rocks or are weathering products derived from altered rocks (Hunt and Ashley, 1979; also see Selected Bibliography in Knepper, 1989). Broad band, multispectral imaging systems provide digital data suitable for detecting and mapping the presence and gross geometry of hydrothermal systems, as well as the distribution of selected industrial minerals, such as carbonates, clays, and zeolites.

High resolution, narrow band multispectral imaging systems, such as NASA's experimental AVIRIS (Airborne Visible and Infrared Imaging System) and the commercially available GER (Geophysical Environmental Research) instrument, obtain digital image data for each ground picture element (pixel) with a spectral resolution nearly matching the resolution of laboratory instruments. These data permit the identification of the presence of many mineral species, providing detailed information on possible mineral zoning and the occurrence of specific industrial mineral products.

THERMAL PROPERTIES

Thermal conductivity and inertia

The thermal properties of primary concern to the geophysicist are thermal conductivity, K, and thermal inertia, I. These are not independent properties, but are related by the following equation:

$$K = I^2/\rho c \qquad (25)$$

where c is the specific heat and ρ is the density.
Thus, rocks with high thermal conductivity will also have high thermal inertia. In this section, thermal conductivities only will be given. Those wishing values of thermal inertia may calculate them from the equation given above using densities given previously. The specific heat for rock forming minerals is nearly constant at 0.2 cal/gm ^{o}C. The principal factor complicating the relation given in equation 25 is the presence of pore water in the more porous sediments where specific heat may approach that of water at 1.0 and conductivity may increase as much as eight times. The effect of pore water on these thermal properties is evident in table 8 which gives the ranges of thermal conductivities observed for various sedimentary rocks and unconsolidated materials. Figure 24 gives thermal conductivities measured for various types of rocks. This figure shows the distinct difference between dolomite and limestone, and the high values and large range characteristic of quartzite.

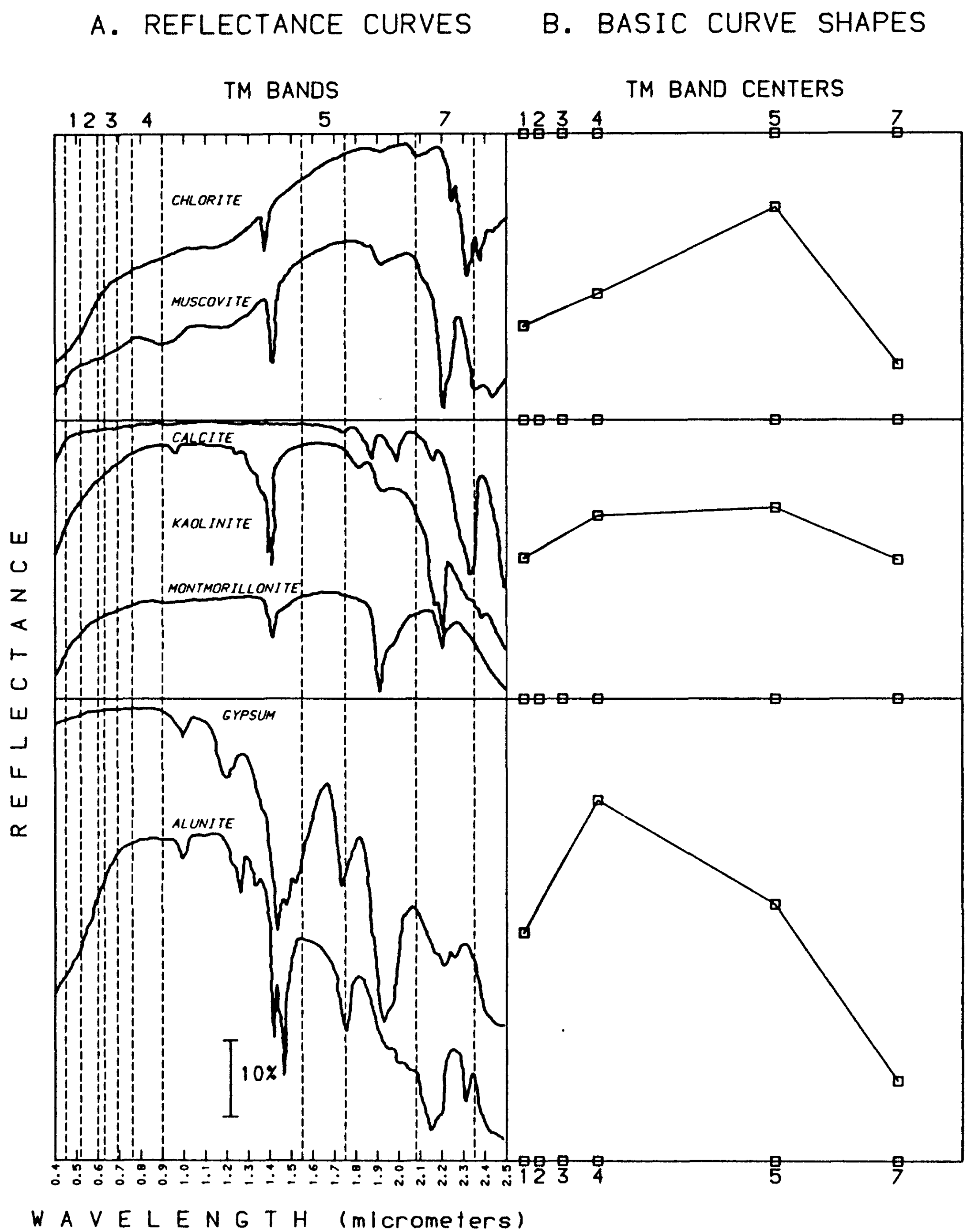


Figure 23. A. Spectral curves of common minerals often associated with hydrothermally altered rocks, showing the locations of the Landsat Thematic Mapper spectral bands. The curves are offset vertically to allow curve stacking. From R. Clark (U.S. Geological Survey, unpublished data).

B. A method of grouping the minerals based on the basic shape of their reflectance curves.

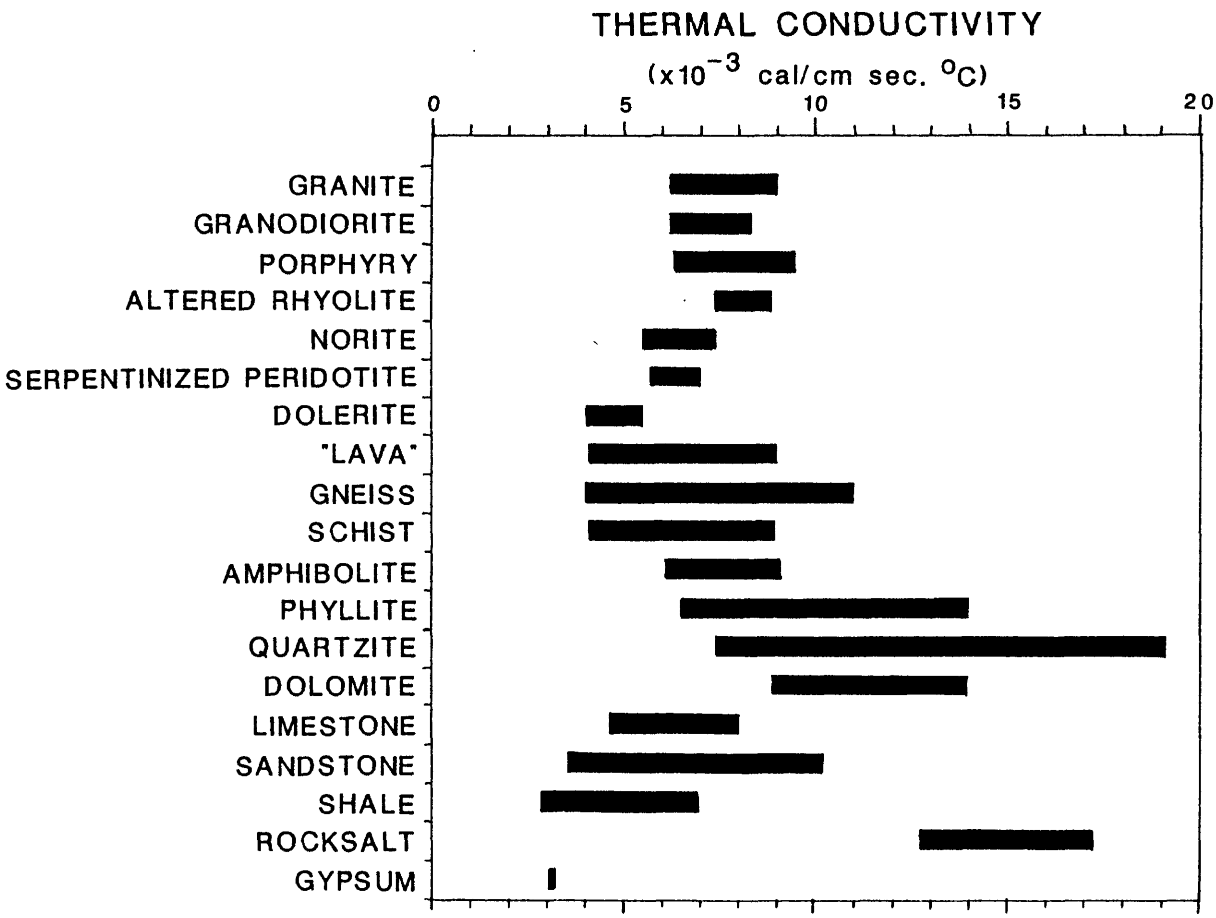


Figure 24. Thermal conductivities for selected sedimentary rocks from Clark (1966).

Many sedimentary and metamorphic rocks show distinct anisotropy, with thermal conductivities varying by as much as three times transverse or parallel to the grain (Heiland, 1940 and Clark, 1966).

The DeBye relationship given by solid state theory provides a relationship among density, seismic velocity and thermal conductivity for crystalline solids (Kittel, 1956). This relationship is

$$K = 1/3Cv\Omega = 1/3c\rho v\Omega \tag{26}$$

where C = specific heat per unit volume
v = seismic P velocity
Ω = phonon mean free path
c = specific heat per unit mass
ρ = density

Table 8. Thermal conductivities (K) for various sedimentary rocks as a function of water content, values in 10^{-3} calories/cm. sec. ^{o}c adapted from Clark (1966).

Rock Type	*Porosity*	*% moisture*	*K*	*% moisture*	*K*
Limestone	3.4 13.2	air dry	5.03 4.40	saturated	6.19 4.97
Dolomite	1.7	"	7.10	"	8.02
Sandstone	0.5 3.0 15.5 22. 29. 59.	" " " " " "	9.20 15.5 7.26 4.43 3.69 1.26	" " " " " "	10.4 17.7 14.01 6.02 9.67 4.86
Pumice	-	"	0.60	"	1.2
Sand	-	0.2	0.65	30.	3.94
Loam	-	0.3	0.79	27.	5.50
Clay	-	1.4	0.57	67.	3.6
Frozen soil	-	20.	0.83	60.	2.71

The Debye relationship has been used by Russian investigators (Neprimerov and others, 1989) to give estimates of thermal conductivities of crustal rocks based on their seismic velocities. Heiland (1940) gives a modified form for this relationship as follows:

$$K = Av^2\rho^2 \quad \text{where A is a constant} \tag{27}$$

Heat sources

At present, radioelement decay constitutes the principal source of heat production for the earth as a whole, being approximately equal to that lost (Durrance, 1986). Quantitatively, the radioelement heat sources are much more significant than sources related to oxidizing sulfides. Radiogenic heat production from the Carnmenellis tin granite of SW England has been sufficient to account for all the heat necessary to drive the hydrothermal systems responsible for mineralization in the district (Durrance, 1986).

Only one locality where a natural fission reaction has occurred is known, that being the Oklo deposit in Gabon (Cowan, 1976). The estimated energy output of this natural reactor was 15,000 megawatt-years at a rate of 10 to 100 kilowatt per sec. Six individual reactor zones are known at the deposit. These zones are estimated to have gone critical about 1800 Ma ago, operating for about 0.5 Ma (Durrance, 1986).

Assuming standard isotopic ratios, the heat production from the three radioelements are, uranium 0.73, thorium 0.20, and potassium 2.7 x 10^{-5} calories per gram-year (Wetherill, 1966). Thus, for most rocks, which have a thorium to uranium ratio of 3.5:1 to 4:1, the heat contribution due to the thorium and uranium content is about equal. The contribution due to potassium for most rocks is typically several times smaller than that due to uranium or thorium. Table 9 gives values of heat production for various rock types whose radioelement composition is give by Vavilin and others (1982).

Locally, heat produced by oxidizing sulfides can be important. The quantity of heat generated is dependant on the particular sulfide, and the end products of oxidation. The rate of heat production is directly related to the rate of oxidation. For particular oxidation reactions, the heat generated can be calculated from values of heats of formation given in thermochemical compilations. The oxidation of pyrite to ferrous or ferric sulfate plus sulfuric acid provides an example to show the order of magnitude of heat generated. The reactions are:

1. $2\ FeS_2 \quad + \quad 7.5\ O_2 + H_2O = Fe_2(SO_4)_3 + H_2SO_4$ (28)

   2 x -42.52   0   -68.32   -641.77   -210.3

2. $2\ FeS_2 \quad + \quad 3.5\ O_2 + H_2O = Fe(SO_4) + H_2SO_4$ (29)

   -42.52   0   -68.32   -217   -210.3

Heats of formation in kilo-calories per mole are indicated below each reactant or product, with the products assumed in dilute solution. For reaction 1 the heat produced is 699 kcal or 2.91 kcal/gm of pyrite, and for reaction 2, 316 kcal or 2.64 kcal/gm of pyrite.

Table 9. Average radiogenic heat production for various types of rocks. Compositional data from Vavilin and others (1982).

Lithology	*Composition*			*Heat Production cal/gm 10^6 yr.*			
	K%	*U ppm*	*Th ppm*	*K*	*U*	*Th*	*Total*
Peridotite pyroxenite	0.012	0.03	0.08	0.0032	.022	.016	0.041
Gabbro-diabase	0.56	0.6	1.8	0.15	.44	.36	0.95
Quartz diorite-granodiorite	2.05	2.1	8.3	0.55	1.53	1.66	3.74
Plagiogranite	1.26	2.7	9.6	0.34	1.97	1.92	4.23
Granite	4.43	4.5	18.0	1.20	3.28	3.60	8.08
Alkali-granite	3.90	6	25.	1.05	4.38	5.00	10.43
basalt-diabase	0.55	0.7	2.3	0.15	0.51	0.46	1.12
andesite	1.66	1.2	4.0	0.45	0.88	0.80	2.13
dacite	2.20	2.5	10.0	0.59	1.82	2.00	4.41
rhyolite	3.87	4.7	19.0	1.04	3.43	3.80	8.27
gravel-conglomerate	-	2.4	9.0	-	1.75	1.80	3.55
sandstone-siltstone	1.07	2.9	10.4	0.29	2.12	2.08	4.49
clay-argillite	2.66	4.0	11.5	0.72	2.92	2.30	5.94
limestone	0.27	1.6	1.8	0.073	1.17	0.36	1.60
dolomite	0.27	3.7	2.8	0.073	2.70	0.56	3.33
chert-quartzite	trace	1.7	2.2	trace	1.24	0.44	1.68
evaporites	trace	1.0	1.0	trace	0.73	0.20	0.93

RADIOELEMENT CONTENT

The radioelements K, U, and Th in rocks are present either as essential elements in a variety of minerals present in a rock, or substituting, often as trace elements, in minerals. Potassium, for most rocks, is present in the potash feldspars, microcline and orthoclase or in micas such as muscovite and biotite. In evaporites, the minerals sylvite and carnallite can be important sources of potassium. In epithermal deposits, hydrothermal alteration assemblages may include alunite, sericite, and adularia as the principal potassium-containing species of mineral.

Uranium and thorium in igneous and metamorphic rocks are typically present in the accessory minerals apatite, sphene, and zircon, and in the rarer species allanite, monazite, pyrochlore, thorite, uraninite, and xenotime. In the near surface oxidizing environment, uranium tends to be very mobile, dissolving in the ground water and then being deposited where conditions are more reducing. Carnotite, uranophane, and other secondary uranium minerals are of this type. Thorium, on the other hand, is relatively stable, remaining with the resistate minerals.

Compilations of the radioelement contents of rocks are rather sparse in the literature. Data from Vavilin and others (1982) were given in table 9 where radiogenic heat production was discussed. The compositional data given there can be compared with table 10 which shows some values obtained on various U.S. Geological Survey standard rocks and several average rock types. A prominent feature seen in table 10 is the abrupt decrease in radioelement content between mafic and ultramafic rocks.

Another feature evident in table 10 is the general correspondence between potassium and thorium. Portnov (1987) discusses this relationship in detail. As the ratio K(%)/Th(ppm) is typically 0.17 to 0.20, he calls rocks having ratios significantly different from this range either thorium or potassium specialized. Igneous rocks or derived metasomatites showing potassium specialization are identified with gold-silver, silver-polymetallic, molybdenum, and bismuth deposits. Thorium specialized rocks and metasomatites

Table 10. Radioelements content of selected USGS rock standards and of other rocks. Data from Van Schmus (1984).

Rock Type	***Composition***		
	K%	U ppm	Th ppm
GSP-1 granodiorite	4.50	2.0	104
G-1 granite	4.45	3.4	50
Ave. Low-Ca granite	4.20	4.7	20
G-2 granite	3.67	2.0	24
STM-1 nephyline syenite	3.54	9.1	27
RGM-1 rhyolite	3.49	5.8	13
MAG-1 marine mud	2.96	2.8	12.2
QLO-1 quartz latite	2.90	5.8	13
SDC-1 mica schist	2.71	3.1	11.4

Table 10. Radioelements content of selected USGS rock standards and of other rocks. Data from Van Schmus (1984).

Rock Type	***Composition***		
	K%	U ppm	Th ppm
Ave. shale	2.66	3.7	12.
AGV-1 andesite	2.35	1.9	6.4
SCo-1 shale	2.20	3.1	9.5
BCR-1 basalt	1.38	1,7	6.0
Ave. sandstone	1.07	1.7	5.5
Ave. basalt	0.83	0.9	2.7
W-1	0.52	0.6	2.4
BHVO-1 basalt	0.43	0.5	0.9
Ave. carbonate	0.27	2.2	1.7
Ave. ultramafic	0.003	0.001	0.004
DTS-1 dunite	0.001	0.004	0.01
PCC-1 peridotite	0.001	0.005	0.01

are identified with tin, tungsten, rare-earth, and rare-metal deposits. These characteristics thus can be used to assess the favorability of regions for one or the other group of mineral deposits.

Hansen (1980) and Wollenberg and Smith (1987) have presented average values and ranges of radioelement content for a variety of rocks. These data are shown in figure 25. These data indicate the potential of gamma-ray spectrometry in lithologic discrimination where element concentration and element ratios may be combined to give a powerful mapping tool.

Soils may be depleted in radioelement content, particularly in uranium and potassium, dependent on the weathering environment, and host mineralogy of the radioelements. Soils in tropical regions are particularly prone to leaching of most soluble components. However, as Guillemot (1988) has shown good quality radioelement data, capable of mapping lithologies, can be obtained even in tropical areas.

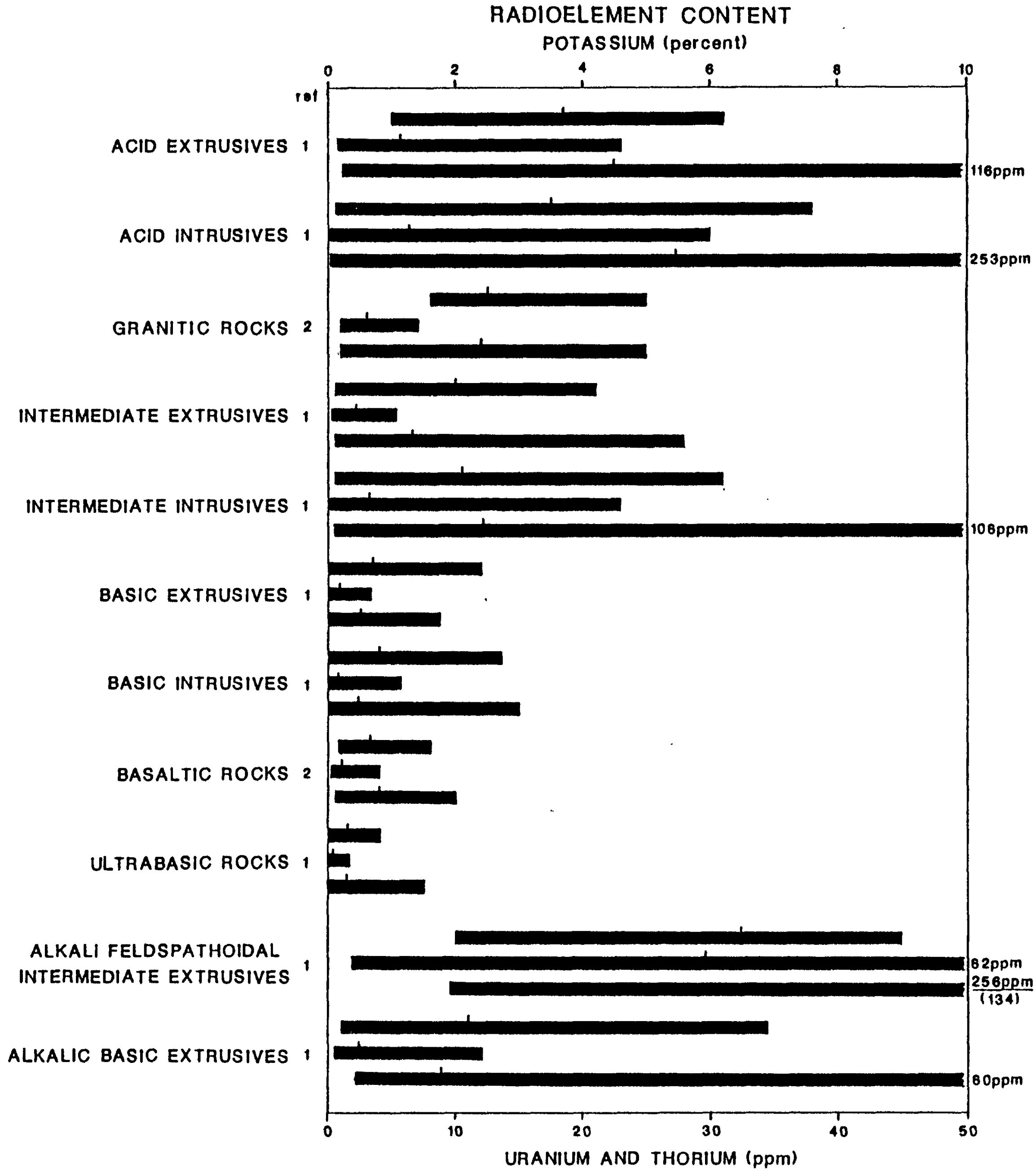


Figure 25. Radioelement contents reported for a variety of lithologies from 1. Wallenberg and Smith (1982) and 2. Hansen (1980). For each type of lithology the elements are in the order top to bottom, K, U, Th. The small vertical bar indicates the mean value.

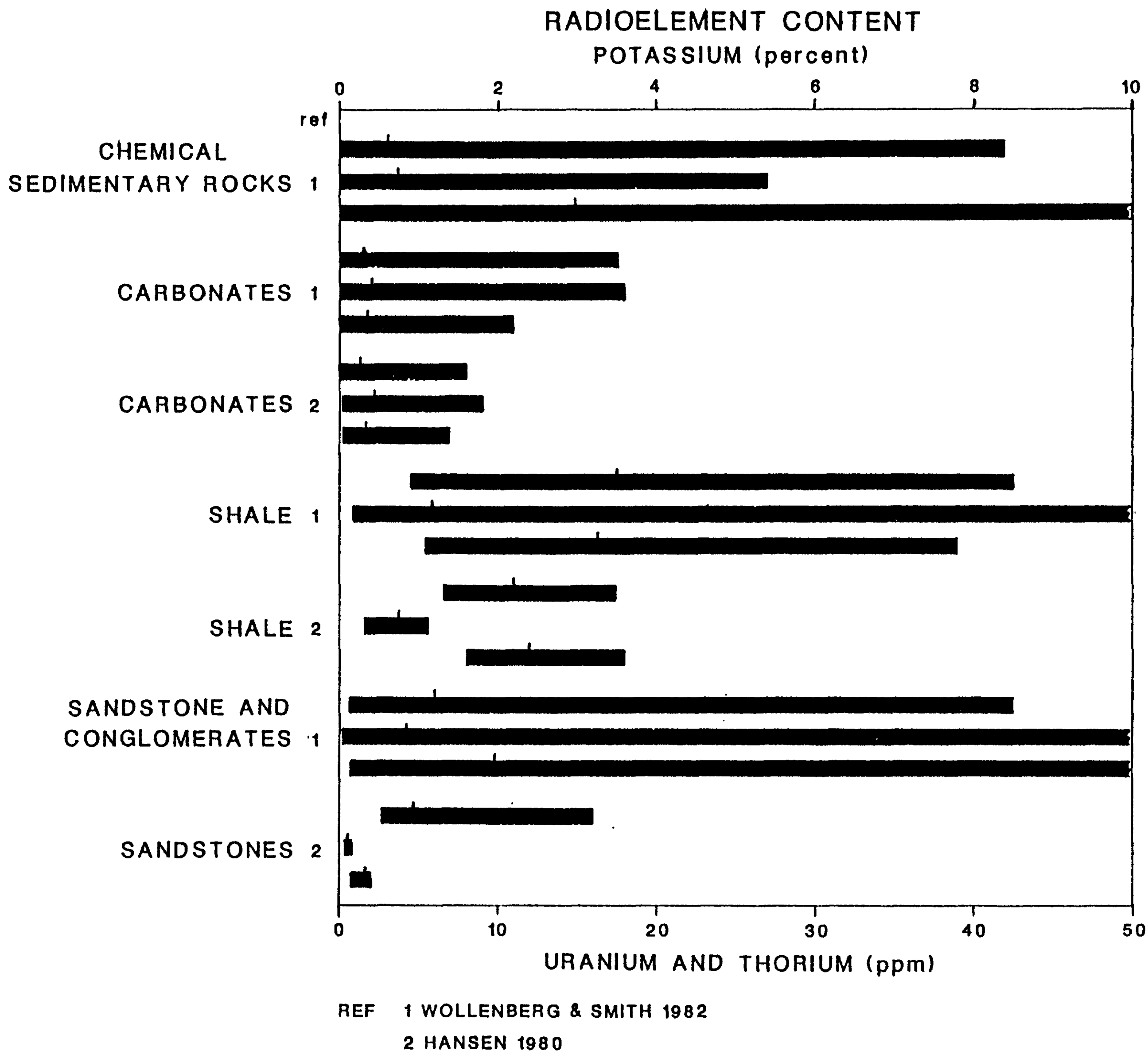


Figure 25 (continued)

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A CATALOGUE OF SELECTED GEOPHYSICAL

DEPOSIT MODELS

The following 10 geophysical models are interim descriptive models that cover 17 of the 85 geologic models of mineral deposits given by Cox and Singer (1986). Users of these models are requested to reference the author, or authors, when referring directly to a particular model, and not to the editors of this publication.

The models that follow are listed, in order, below.

Deposits related to alkaline intrusions

1. Geophysical model of carbonatites, Cox and Singer Model 10
2. Geophysical model of diamond pipes, Cox and Singer model 12

Deposits related to felsic phanerocrystalline intrusive rocks

3. Geophysical model of tin skarn and related deposits, Cox and Singer models 14b tin skarn, 14c replacement tin, 15b tin veins and 15c tin greisens

Deposits related to subaerial felsic to mafic extrusive rocks

4. Geophysical model of hot springs Au-Ag, Cox and Singer model 25a
5. Geophysical model of Creede, Comstock, Sado, Goldfield and related deposits, Cox and Singer models 25b Creede, 25c Comstock, 25d Sado, and 25e quartz-alamite Au
6. Geophysical model of carbonate-hosted Au-Ag, Cox and Singer model 26a

Deposits in clastic sedimentary rocks

7. Geophysical model of Olympic Dam Cu-U-Au, Cox and Singer model 29b

Deposits related to regionally metamorphosed rocks

8. Geophysical model of low-sulfide Au-quartz veins, Cox and Singer model 36a
9. Geophysical model of Homestake Au, Cox and Singer model 36b

Deposits related to surficial processes and unconformities

10. Geophysical model of placer Au-PGE, and PGE-Au, Cox and Singer models 39a placer Au-PGE, and 39b placer PGE-Au

GEOPHYSICAL MODEL OF CARBONATITES

COX AND SINGER MODEL NO. 10 Compiler - **D.B. Hoover**

Geophysically related models-No. 12, Diamond pipes; No. 29b, Olympic Dam

A. Geologic Setting

- Alkaline volcanic and sub-volcanic complexes emplaced along major zones of crustal weakness within continental, commonly Precambrian, crust (Garson, 1984).
- Syenite and carbonate magmas emplaced as intrusions or ring dikes, cone sheets, dikes and plugs within complex alkaline volcanic centers. Often associated with mafic and ultramafic units.
- Commodities include REE, apatite, magnetite, Nb, Cu, Th, U, Zr, Ti, Ni, Sr, V, fluorite, lime, and vermiculite.

B. Geologic Environment Definition

Regional magnetic, gravity, radioelement, and remote sensing surveys may identify deep-seated fault systems, expressed as lineaments, within stable continental crust. The East African Rift, and St. Lawrence River fault system are examples (Rao, 1986; Paarma, and Talvitie, 1972).

Individual intrusive or volcanic centers show as circular to elliptical bodies in remote sensing images, and on magnetic, gravity, and radioelement maps. On magnetic, gravity, and radioelement maps the centers typically show as large amplitude, very complex highs, reflecting the variety of lithologies and alteration effects present. Fenitization of country rock up to 1-2 km from intrusive, is often expressed as a high-potassium halo. Phosphate and potassium enrichment in complexes can be expressed in remote sensing images by enhanced vegetation (Cole, 1982; Dawson, 1074; Ramberg, 1973; Saterly, 1970; Vorob'yev, and others, 1977).

C. Deposit Definition

Geophysical expression is highly variable dependent on the commodity, carbonatite type, extent of post-emplacement alteration, and adjacent lithologies both within the alkaline complex and of the host rock. No general rules can be given. In some cases there can be direct expression as for magnetite, niobium, uranium, and apatite mineralization expressed as magnetic or radioelement highs. In other cases magnetic, gravity, and radioelement data are used to define individual lithologies or structures. Ground geophysics appears to have had little use in direct deposit exploration. This probably reflects the variety of commodities, and lithologies present, rather than any inherent problem with methods. In some cases electrical methods should be relevant to deposit definition, such as where hydrothermal alteration is present giving lower resistivity. Where magnetite and disseminated sulfides are present, then IP methods would be useful (Gold and others, 1967; Secher, and Thorning, 1982).

D. **Size and Shape of**	**Shape**	**Average Size/Range**
Alkaline center	cylinder or cone, oval to circular in cross section	<10 km diam.
Deposit	highly variable, may be tabular, cylindrical or irregular	$21x10^6/5.7x10^6-7999x10^6$ m^3
Alteration haloe	irregular, about deposit concentric to alkaline center	Fenitization may extend 2 km beyond alkaline center[2,20]

E. Physical Properties (units)	Deposit	Alteration	Alkaline Complex	Host
	Calcitic, dolomitic or ankeritic carbonatites	Fenitization	Intrusive/volcanic center, often with mafic and ultramafic units	Continental crust
1. Density (gm/cc)	2.79-3.41[5]	?	2.79-3.41[5]	*
2. Porosity	variable	low?	variable	*
3. Susceptibility	highly variable	medium < volcanic complex	variable, high on average	*
4. Remanence	variable, gen. low	?	variable,	*
5. Resistivity	high in massive units? medium in breccia?	?	medium to high, low cap in tropics	*
6. IP Effect	medium?, variable 1-2% sulfide[20] + magnetite give target	low-medium?	highly variable, reflects magnetite and disseminated sulfide distribution	*
7. Seismic Velocity	medium-high?	high?	high?	*
8. Radioelements				
K (%)	highly variable 0-10	high, generally > background	high to very high 3-10x background[1,18] internally highly variable	*
U (ppm)	highly variable to 100's[13]	low to medium?	"	*
Th (ppm)	"	"	"	*
9. Other				
heat-flow	?	?	may be double regional due to radioelement content	*

F. Remote Sensing Characteristics

Visible and near IR--Lineaments reflecting major zones of crustal weakness (Garson, 1984); arcuate patterns reflecting volcanic centers (Theilen-Willige, 1981), radial or annular drainage patterns. Enhanced vegetation over K and P-rich carbonatites (Cole, 1982). Spectral identification of CO_3, REE, and ferrous iron in combination unique to carbonatites (Rowan and others, 1986). The close spacial association of carbonate and alkalic rocks may be indicative.

Thermal IR--Airborne thermal imaging can identify carbonatite complex reflected in variation of fracture pattern and residual soils (Talvitie and others, 1981). The paupacity of quartz and dominant feldspar and mafic minerals may be identifiable.

G. Comments

High density mafic and carbonate rock units generally produce a gravity anomaly of a few to 10's of mgals over alkaline complex. Magnetite in complexes can exceed 50% in some carbonatites, and is often Ti-rich. There is very little public literature relating to electrical or seismic methods applied to these deposits. However, the presence of magnetite, and commonly 1-2% disseminated sulfide (Stockford, 1972) in many deposits suggests that I.P. could be a useful tool. No IP surveys are known in such deposits. In tropical regions the weathering of mafic units within an alkaline complex should produce a low resistivity cap (Palacky, 1986) detectable by airborne EM techniques. Most radioelement literature does not give results from a calibrated system, thus providing few quantitative estimates. In tropical regions potassium and especially uranium will be leached from the surface, but thorium can be concentrated (Issler, 1976). In these regions thorium provides the best radioelement indicator.

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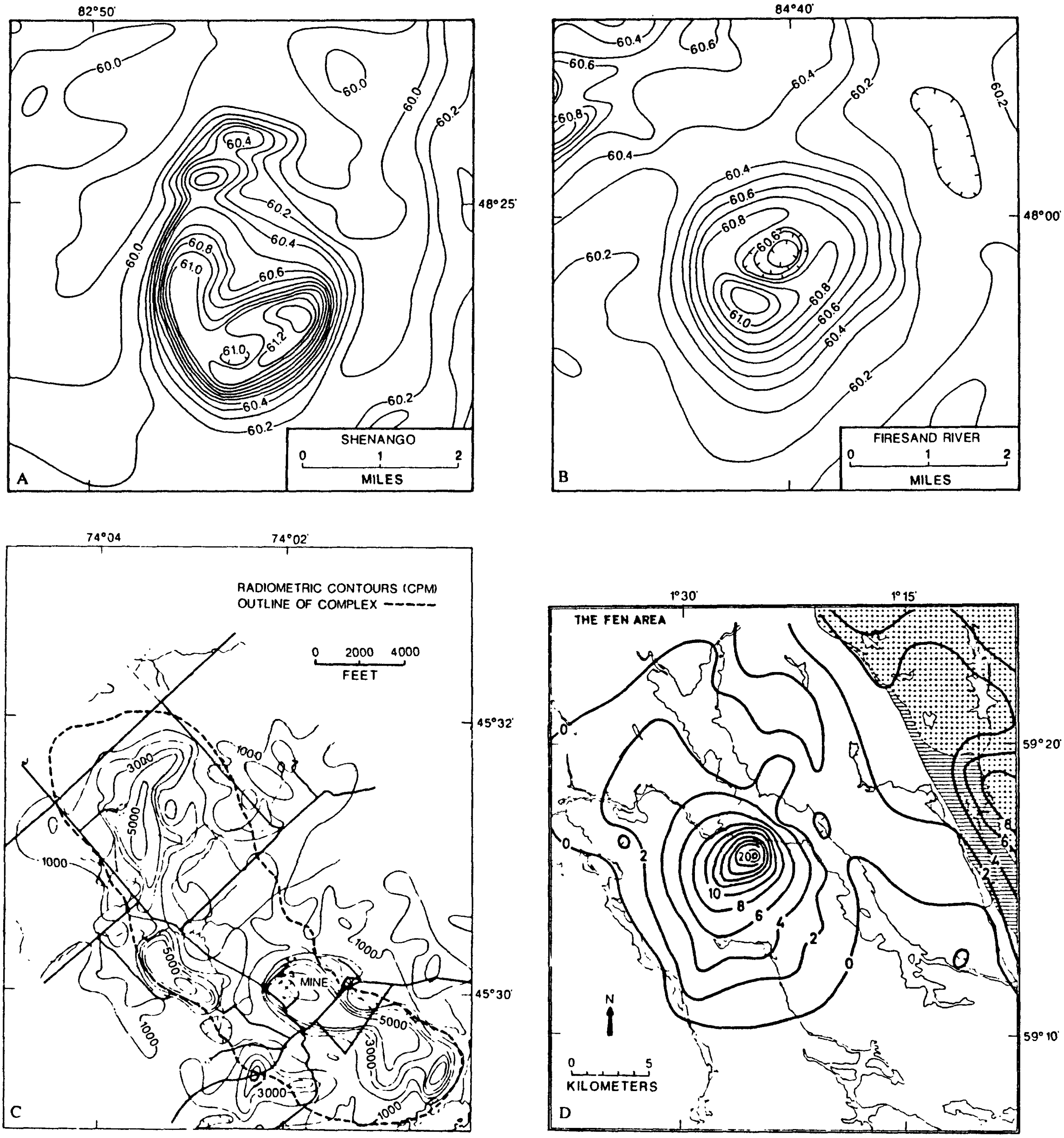

Figures; A and B. Aeromagnetic maps of the Shenango and Firesand River carbonatite complexes, Ontario, Canada, adapted from Satterly (1970). The magnetic features are about three miles across. Note the central low in the Firesand River map that Satterly (1970) reports is seen in simple carbonatite complexes. C. A total-count airborne gamma-ray map of the Oka complex, Quebec, Canada, adapted from Gold and others (1966). Areas of high counts are 4 to 5 times background. D. Residual gravity map of the Fen complex, Norway, adapted from Ramberg (1973). The 23 mgal-anomaly can be modeled by a 15 km high vertical cylinder having a density contrast of 0.48 gm/cm^3.

GEOPHYSICAL MODEL OF DIAMOND PIPES

COX AND SINGER Model No. 12 Compilers - **D.B. Hoover**
Geophysically similar models-No. 10 Carbonatites; **D.L. Campbell**
No. 29b, Olympic Dam

A. Geologic Setting
- Kimberlite or lamproite diatremes emplaced along zones of basement weakness within or on the margins of stable cratons; often in groups of three or more (Dawson, 1971).
- Often spatially related to carbonatites, but not normally occurring along same zones of crustal weakness (Dawson, 1967; Garson, 1984). A genetic relationship is open to question.

B. Geologic Environment Definition

Regional magnetic, gravity, and remote sensing surveys may identify deep-seated fracture systems and related anteclises or syneclises that define zones of weak crust favorable for emplacement (de Boarder, 1982; Tsyganov, and others, 1988).

C. Deposit Definition

Individual diatremes generally appear as circular to elliptical bodies in remote sensing images, and on magnetic, gravity, or resistivity maps. The diatremes may show as distinct magnetic highs (Yakutia, West Africa) of hundreds to a few thousand nT, but high remanence or magnetic host rocks can result in negative or no anomalies. Gravity (order of 1 mgal), resistivity, and seismic velocity anomalies generally show as lows over the diatremes related to serpentization and weathering of the mafic rocks. Radioelement surveys have generally not been effective, although in Yakutia Fedynsky and others (1967) report that they have been used to differentiate between diamond-bearing basaltic kimberlite from barren micaceous kimberlite and carbonatites (da Costa, 1989; Kamara, 1981; Gerryts, 1970; Macnae, 1979; Guptasarma and others, 1989).

D. **Size and Shape of**	**Shape**	**Average Size/Range**
Deposit	Vertical cone, carrot-like	0.1 to 5 km diameter; generally 0.4 to 1 km depth to about 2 km
Alteration haloe	Irregular about pipe	thin, not geophy. significant
Cap	Elliptical cylinder	0.1 to 5 km, 0-10's m thick

E. Physical Properties (units)	Deposit	Alteration	Cap	Host
	kimberlite or lamproite pipe	Si, CO_2, K metasomatism	clay-rich weathering zone-blue and yellow ground	any cratonic unit
1. Density (gm/cm^3)	2.75[5] 2.64-3.12[2,5,11] 2.35-2.55[8]	?	2.35?[13] 2.5-2.62[2]	*
2. Porosity	low-moderate	low?	high[2]	*
3. Susceptibility (cgs)	$1x10^{-4}$-$1x10^{-2}$[8] to $2.3x10^{-3}$[6]	?	$1x10^{-5}$-$1x10^{-3}$[8] to $2x10^{-5}$[6]	*
4. Remanence	variable 0-0.8-2.0[13]	?	variable	*
5. Resistivity (ohm-m)	100-2000[2,8,11,13]	medium-high	2-100[2,8,11,13]	*
6. IP Effect (msec.)	low	low?	low,0-4[10]	*
7. Seismic Velocity (km/sec)	2.6-3.3[2]	high?	1.5[2]	*
8. Radioelements				
K (%)	2.6 average 0.07-6.7[1]	medium	medium?	*
U (ppm)	0.26, average 0.07-0.8[1]	low	very low	*
Th (ppm)	0.44, average 0.17-0.9[1]	low	low	*

F. Remote Sensing Characteristics

Visible and near IR-lineaments may reflect zones of crustal weakness along which pipes were emplaced. Lineament intersections may be favored locations (Tsyganov and others, 1988). Vegetation anomalies related to drainage and lithologies can be used for location. Alteration products of kimberlite, such as serpentine, chlorite, and vermiculite show distinct spectral absorption features that can be detected (Kingston, 1989).

G. Comments

The relatively small size, 0.4-1.km, of most pipes requires detailed coverage for identification. Signature differs from carbonatites in reduced amplitude of magnetic anomaly, and by a small negative gravity anomaly in contrast to the large positive anomaly of carbonatites. A combination of magnetic, gravity, and resistivity methods is most used in exploration. No single method is universally applicable. Radioelement methods have had relatively little use, although they should have some application in differentiating varieties of kimberlites and lamproite. Some Russian literature (Ratnikov, 1970) gives very low values of density for kimberlites. These probably refer to serpentinized or weathered samples and are not representative of unaltered rock. Gerryts (1967) gives a rule-of-thumb of 1 mgal/183 meters (200 yards) of pipe diameter for the gravity low. A broad gravity high ring about the central low, and due to dense, deeper, kimberlite has not been observed. Guptasarma and others (1989) report both positive and negative gravity and magnetic responses over kimberlites in India.

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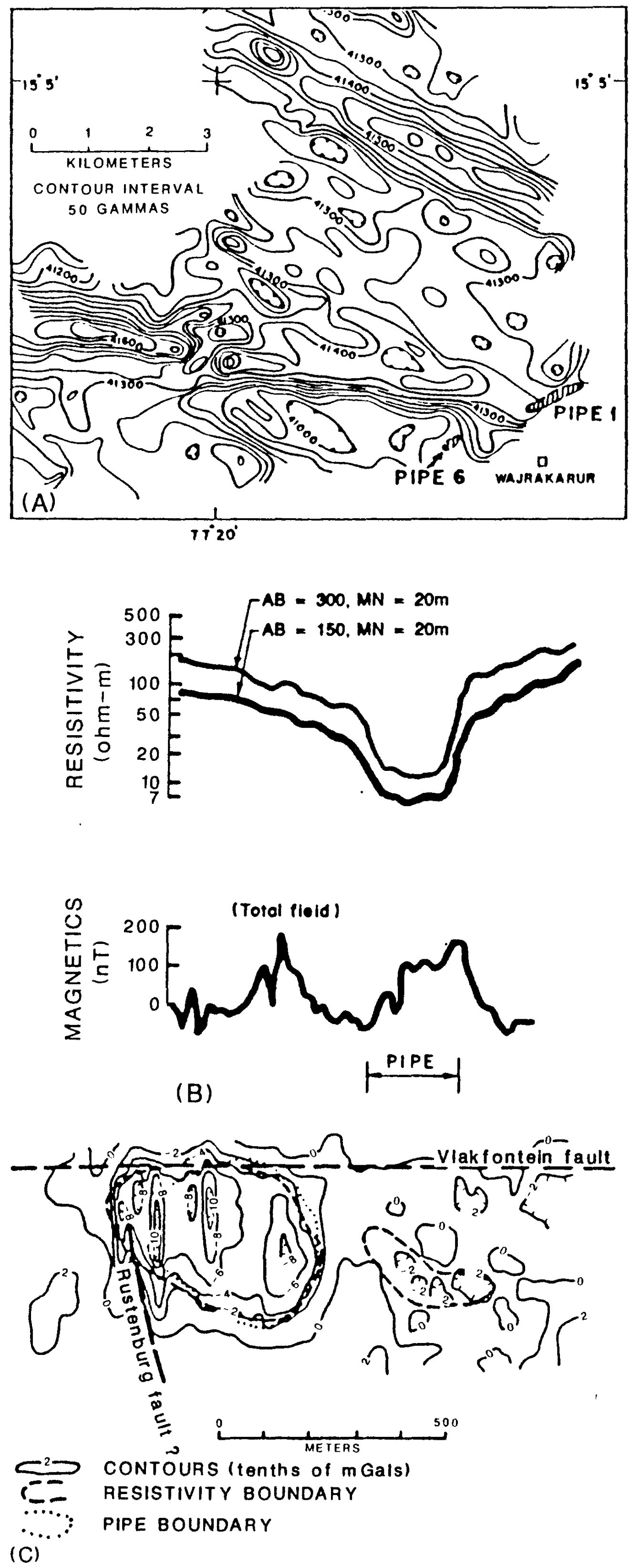

Figures A. Strong regional magnetic linear adjacent to two kimberlite pipes in the Wajrakarur area, Andhra Pradesh, India adapted from Guptasarma and others (1989). Contour interval is 50 gamma. B. Resistivity and ground magnetic traverse across the Palmietfontein pipe South Africa adapted from da Costa (1989). C. A residual gravity map of the Palmietfontein pipe also showing its emplacement at the junction of the Vlakfontein and Rustenburg faults, after da Costa (1989).

GEOPHYSICAL MODEL OF TIN SKARN AND RELATED DEPOSITS

COX AND SINGER combined models; 14b, tin skarn; 14c, replacement tin; 15b, tin veins; 15c, tin greisens; and tin pegmatites-(no geological model yet)

Compilers-D.B. **Hoover**
D.H. **Knepper**

Geophysically related models; 20a, porphyry tin; 20b, tin-polymetallic, 25b rhyolitic tin; 39e, placer tin

A. Geologic Setting

- Pegmatite, greisen, vein, skarn, replacement, and fissure lodes developed within or external to specialized leucogranites.
- Generally associated with apical parts of the specialized leucogranites.
- Plutons are generally of S or A type, ilmenite series granitoids (Hosking, 1988)

B. Geologic Environment Definition

Geophysics is used primarily to define permissive areas where leucocratic source granites are present, especially highly differentiated, late stage phases of these source granites.

Magnetic response of parent granitoid weak to non existant, typical of S or A type, ilmenite series granites. Where skarn, replacement, or vein mineralization is present in host rock an irregular magnetic haloe may be present (Bishop and Lewis, 1988; Asanov, 1978; Xianguang, 1988; Webster, 1984a).

Parent granitoid generally shows as a gravity low due to typical 0.1 gm/cm^3 density contrast (Asanov, 1978; Gongjian and Rui, 1988; Bishop and Lewis, 1988).

Radioelement content of parent granitoid is high to very high, giving definition by airborne methods where exposed. High uranium (Yeates and others, 1982) or high thorium (Chatterjee and Muecke, 1980; Towsey and Patterson, 1984) often are keys to enriched parts of pluton.

Electrical techniques are used to map the upper surface of buried plutons to locate apices (Alonso and Corral, 1983; Xianguang, 1988).

Exposed parts of felsic plutons may have distinctive remote sensing signatures, and separate intrusive phases may be identified (Amaral, 1982; Keighley and others, 1980, Guerra, 1978, Moore and Camm, 1983). Covered tin granites have been identified from Landsat data (Marconnet, 1984). Tin granites have been reported confined to areas of relatively shallow depths to the Mohorovičić discontinuity (Gongjian and Rui, 1988), but Zeitz and others (1976) note that in southeast Russia tin is related to a thick crust.

C. Deposit Definition

Geophysical expression is variable, in part due to different deposit types covered. Magnetic and electrical methods are most often applied because of presence of pyrrhotite, and accessory sulfides present in ore, giving rise to local magnetic, resistivity and IP anomalies. Locally, elevated radioelement content may identify favorable structure, especially where late-state alteration has decreased potassium content (Towsey and Patterson, 1984). Gravity may reveal local highs related to metamorphism of surrounding host (Asanov, 1978; Rulski, 1982).

D. **Size and Shape of**	**Shape**	**Average Size/Range**
apices of pluton	irregular blister to vertical cylinder	0.25 - 2 km diam.
deposit	variable, irregular tabular	skarn, replacement 2.6x10^{6}m^3 greisen .3x10^6-24x10^{6}m^3 vein 8.9x10^{4}m^3
alteration haloe	donut shape around apices, irregular around deposit	may extend to several kms from pluton

E. Physical Properties (units)	Deposit	Alteration	Source Granitoid	Host
			Leucocratic Granitoid	
1. Density (gm/cm^3)	3.9[26]; 2.52-2.84[22] $\Delta\sigma$=0.10[3]	2.8[26] 2.67[3]	low, 2.6[26]; 2.54-2.68[22] 2.55[3]	*
2. Porosity	variable	variable	variable-low	*
3. Susceptibility S.I.x10^{-3}	variable, low to moderate, 6.43[22]	variable	low, 6-87[22]	*
4. Remanence	?	?	?	*
5. Resistivity (ohm-m)	10's-100[26] 2000-6000[3] 17-30K[22]	2000-8000[26]	high-very high 400-1900[26] 26-56K[22]	*
6. IP Effect %	5-12[9] 8-12[3]	moderate	low	*
7. Seismic Velocity (km/sec)	P-2.7-4.1[22] S-1.1-2.6[22]	high?	high P-4.6-6.3[22] S-2.8-4.0[22]	*
8. Radioelements				
K (%)	variable, high or low	?	moderate to high	*
U (ppm)	high?	must be >4[25]	moderate to high	
Th (ppm)	variable, high to low	>6[7]	moderate to high	
9. Other				
self-potential	moderate to high	?	low-very low	*

F. Remote Sensing Characteristics

Tin-bearing ore minerals are not spectrally distinctive in the visible and near-infrared parts of the spectrum. Remote sensing expression is based on indirect indicators including spectral, albedo, and textural differences between granitic parent rocks, the contact metamorphic aureole, and the country rocks (James and Moore, 1985; Keighley and others, 1980), vegetation differences between granitic parent rocks and the country rocks (Moore and Camm, 1982), and structural features mappable on various types of images and photos (James and Moore, 1985). Hydroxyl-bearing alteration minerals (clays, micas), iron oxides, and carbonate host rocks can be uniquely identified with high-spectral resolution instruments in the visible and near-infrared (Rowan and others, 1983; Clark and others, 1990), and broad-band data in the visible and near-infrared, such as Landsat Thematic Mapper, are effective for separating the clay-carbonate and iron oxide mineral groups from unaltered, non-iron oxide bearing and non-carbonate parent and country rocks (Knepper, 1989).

G. Comments

A distinct gravity low, to several 10's of mgal, and magnetic low surrounded by an annular ring of local magnetic highs often is a key to presence of source granite and mineralization in surrounding host. The high resistivities of source granites and metamorphosed host give good penetration by EM methods. AEM methods may be particularly effective in this environment. The radioelement pattern is apt to be complex due to late stage metasomatism that may produce local K and Th depletion. Passing mention in the literature is made to piezoelectric methods for tin vein deposits.

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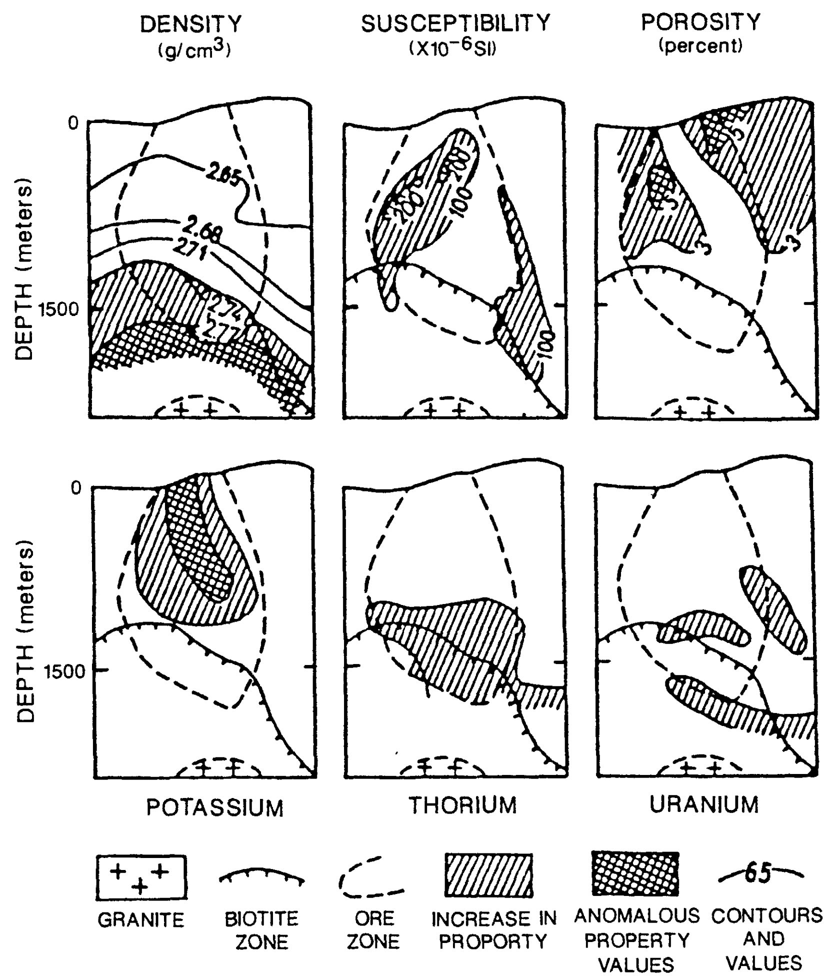

Figure A. Simplified diagram showing distribution of rock density, suscep-tiblity, percent frequency effect (IP), and radioelement content around a tin vein and stockwork deposit. Adapted from Rulski, 1982.

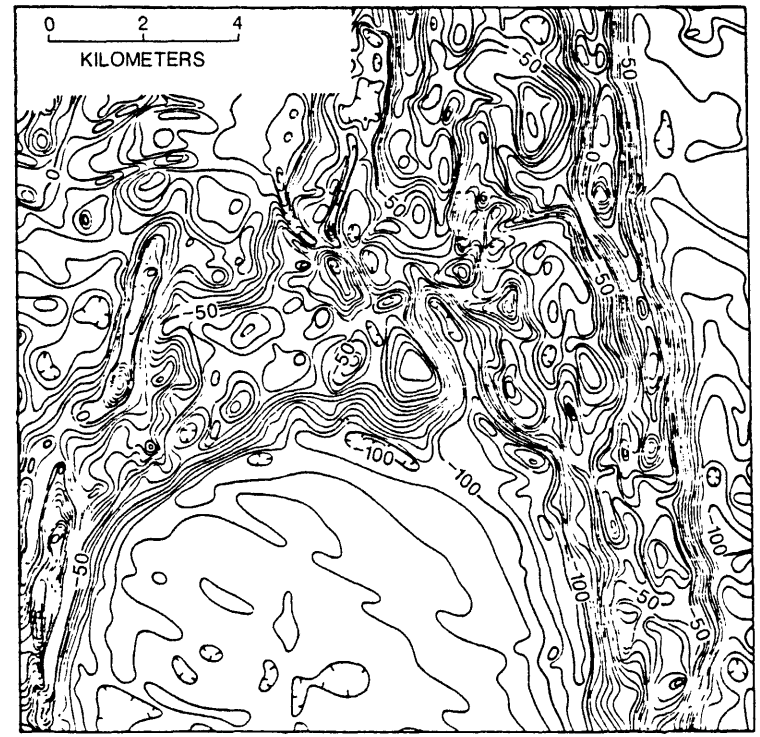

Figure B. Magnetic expression of the Weethalle tin granite and surrounding area, New South Wales, Australia, flown at 85 m above ground and 250 m line spacing. Adapted from Webster, 1984a.

GEOPHYSICAL MODEL OF HOT SPRINGS Au-Ag

COX AND SINGER MODEL NO. 25a Compiler - **W.D. Heran**

Geophysically similar models-No. 25b-25g

A. Geologic Setting

- Located along regional fracture systems associated with felsic intrusive and extrusive volcanism above subduction zones, rifted continental margins, and transform faults.
- Deposits occur in subaerial rhyolitic volcanic centers, rhyolite domes, and the shallow parts of related geothermal systems; along caldera rim fracture zones; and high angle basin and range faults.
- Deposit is usually shallow, disseminated and/or stockwork veins, containing fine grain silica and quartz in silicified volcanic breccia with gold-silver, pyrite, adularia, antimony and arsenic sulfides.

B. Geologic Environment Definition

Airborne magnetics or electromagnetics may define rhyolite volcanic centers, domes and calderas. Regional gravity low or arcuate magnetic high indicate caldera or volcanic dome (Tooker, 1985). Linear trends or breaks in aeromagnetic or gravity data, may define major structures (Tingler and Berger, 1985). Remote sensing can identify deep-seated regional fractures, lithologic changes and alteration. Airborne gamma ray data will show host rhyolite high in all radioelements.

C. Deposit Definition

Widespread alteration may be defined by a magnetic low (Williams and Abrams, 1987; Allis, 1990). Areas of brecciation will show a gravity low (Tingley and Berger, 1985; Allis, 1990). The silica cap will be indicated by an electrical resistivity high surrounded by a low. Disseminated sulfide and clays can be mapped as an IP high (Allis, 1990). Local faults and altered areas may be defined by electromagnetic methods (Kawasaki, and others, 1986). Radioelement surveys may define alteration.

D. **Size and Shape of**	**Shape**	**Average Size/Range**
Deposit	Irregular Cone Tabular	$5.2x10^6m^3$, $.68-40x10^6m^3$
Alteration Haloe/s	Concentric	
Cap	Lens	

E. **Physical Properties** (units)	**Deposit**	**Alteration**	**Cap**	**Host**
1. density	2.0-2.9,2.5[2]	?	?	2.35-2.7,2.52[2]
2. porosity	?	?	?	?
3. susceptibility (10^{-6}cgs)	?	?	?	20-3000[2]
4. remanence	?	?	?	?
5. resistivity (ohm-m)	10-2500,300[2]	?	?	50-3000[2]
6. chargeability (mv-sec/v)	5-150,30[2]	?	?	100[2]
7. seismic vel. (km/sec)	(low)	(low)	?	
8. radiometric				
K-%	?	(high)	?	(high)
U-ppm	?	?	?	(high)
Th-ppm	?	?	?	(high)
9. Other (specific)				

F. Remote Sensing Characteristics

Visible and near IR - lineaments reflecting major crustal weak zones; arcuate patterns reflecting volcanic centers; color anomaly due to limonite or quartz may be spectrally detectible from TIMS data. Thermal IR-use unknown.

G. Comments

Remote sensing, regional magnetics and gravity are often used in reconnaissance. Closely spaced airborne magnetic and electromagnetic surveys may further define favorable areas. A combination of detailed gravity, ground magnetics, electromagnetics and electrical surveys can define target dimensions. The Bell Springs deposit, Nevada, U.S.A. was discovered by Noranda Mining in 1979 as a result of a gamma-ray survey during exploration for uranium (Bussey and others, 1991).

Geophysical methods have been employed in recent years to identify modern hot spring or geothermal systems and the literature is full of examples. Hot spring Au-Ag deposits are fossil geothermal systems and many geophysical methods used to explore for modern geothermal systems will be applicable Allis (1990).

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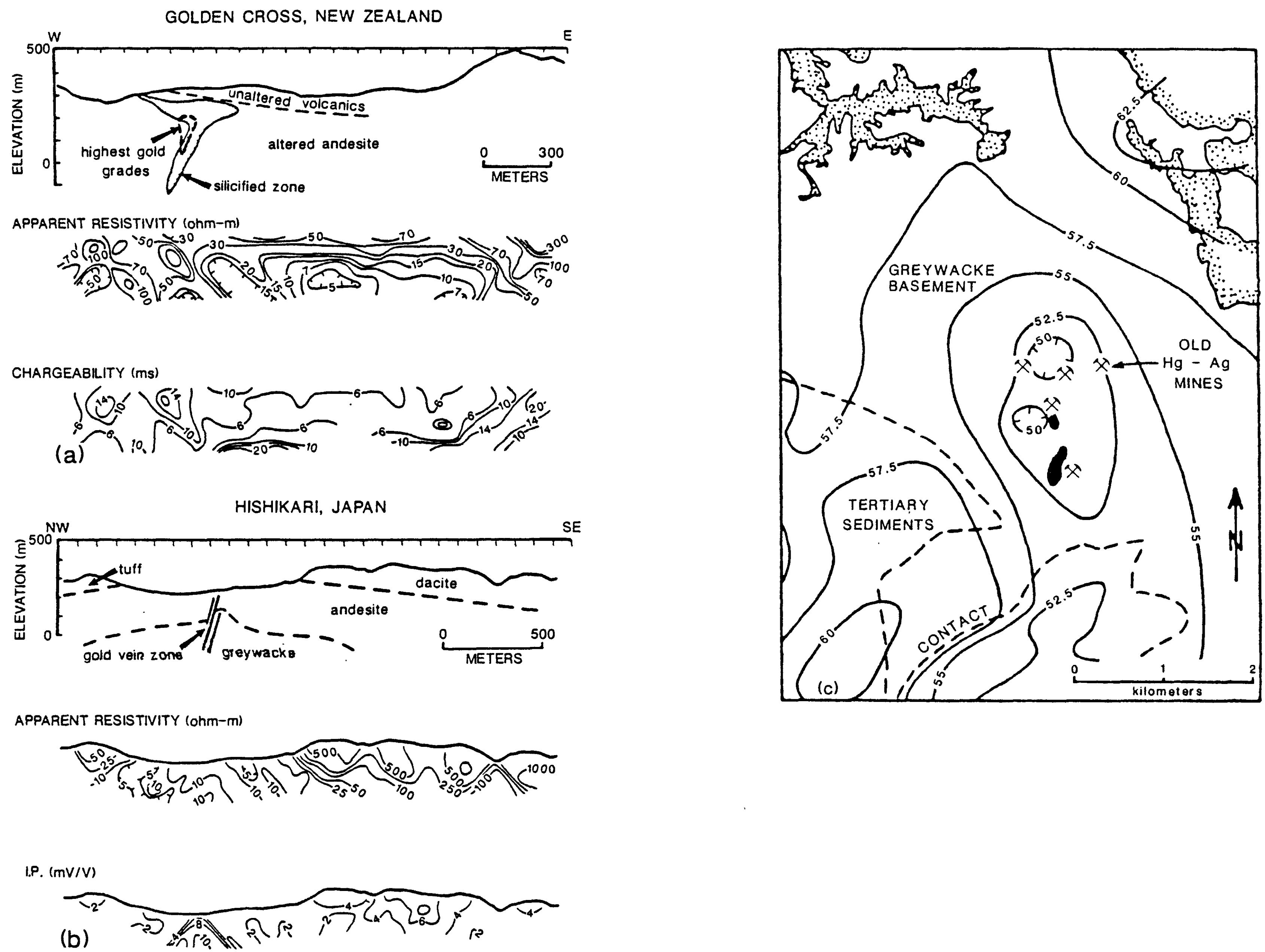

Hot Springs Au model illustrations showing (a) IP data across the Golden Cross Au deposit, New Zealand, (b) IP data across the Hishikari gold deposit, Japan, and (c) gravity data at the Puhipuhi deposit New Zealand. The Puhipuhi deposit occurs in relatively high density host rocks.

GEOPHYSICAL MODEL OF CREEDE, COMSTOCK, SADO, GOLDFIELD AND RELATED EPITHERMAL PRECIOUS METAL DEPOSITS

COX AND SINGER MODELS:
Creede epithermal vein (25b),
Comstock epithermal vein (25c),
Sado epithermal veins (25d),
and quartz-alunite Au-Ag (25e).

Compilers - **D.P. Klein**
V. Bankey

•Models with related geophysical characteristics (Cox and Singer, 1986): Au-Ag-Te veins (22b), Polymetallic veins (22c), Hot springs Au-Ag (25a).

A. Geologic setting (Cox and Singer, 1986)
- •Continental, usually mid-Tertiary, felsic volcanic centers.
- •Faulted, fractured, and brecciated, andesitic to rhyolitic lavas and tuffs, hypabyssal, porphyritic dacite to quartz monzonite intrusions.
- •Deposits occur in the edifice of volcanic morphologic features, often near edge of volcanic center, or above or peripheral to intrusions.
- •Commonly associated with resurgent caldera structural boundaries.
- •Commodities: Au, Ag, Cu, Pb, Zn

B. Geologic Environment Definition

Gravity lows are common over thick silicic volcanic rock sequences and caldrons. The presence of a deep, low-density granitic batholith within the basement rocks may contribute to the gravity low (Ratté and others, 1979; Plouff and Pakiser, 1972, Steven and Eaton, 1975).

Short-wavelength magnetic anomalies are common over volcanic terranes because of variable magnetizations and polarizations. This pattern may contrast with an area of moderate to intense alteration that will display a longer-wavelength low, often linear in the case of vein systems, caused by destruction of magnetite. Local highs may be associated with hypabyssal intrusions (Ratté and others, 1979; Wynn and Bhattacharyya, 1977; Irvine and Smith, 1990, Doyle, 1990).

Radiometric highs may occur from regional potassic enrichment associated with volcanism. Regional alteration patterns may also be apparent in multi-spectral remote sensing (Marsh and McKeon, 1983; Podwysocki and others, 1983; Duval, 1989; Watson, 1985).

Regional seismic sound velocity for volcanic sequences are low compared to basement rock. Seismic reflections are generally incoherent and noisy because of signal scatter by volcanic layers and structure (Hoffman and Mooney, 1984; McGovern, 1983).

Regional resistivity is generally low for weathered and altered andesitic to rhyolitic volcanic rocks as compared to high resistivity typical of buried intrusions or uplifted basement or carbonate sedimentary rocks (Frischknecht and others, 1986; Long, 1985; Senterfit and Klein, 1991).

C. Deposit Definition

There are no consistent geophysical signatures to directly identify epithermal vein mineralization. However, several geophysical characteristics are diagnostic of favorable structures and alteration. These characteristics are best measured using closely spaced ground measurements (Irvine and Smith, 1990; Allis, 1990; Doyle, 1990; Johnson and Fujita, 1985; Middleton and Campbell, 1979; Senterfit and Klein, 1992; Zonge and Hughes, 1991).

Gravity highs will be caused by felsic intrusions within flow or tuff sequences, by structural highs of basement or carbonate rocks within the volcanic sequence, or by silicification of otherwise relatively low-density volcanic rock. Weak, local lows may exist over zones of brecciation or fracturing. Weak, local highs may be found over dense silicic vein systems hosted by more porous volcanic rocks. On deposit-scale investigations, high-precision gravity to resolve anomalies of the order of 1 mgal (to 0.5 rarely) would be required (Irvine and Smith, 1990; Allis, 1990; Criss and others, 1985; Kleinhampl and others, 1975; Ratté and others, 1979; Locke and De Ronde, 1987).

Magnetic lows will be associated with alteration; however, discriminating such lows from the background may be difficult on a deposit scale.

Radiometric anomalies are expected across epithermal veins because of potassic alteration, which is common in the upper levels of veins (Marsh and McKeonn, 1983; Pitkin and Long, 1977).

Resistivity highs flanked by resistivity lows are characteristic of a simple and idealized quartz-adularia vein system with associated argillic to propylitic alteration. However, there may be geologic structures and petrologic complications that distort this ideal picture. More generally, resistivity lows will be associated with: 1) sulfides when concentrated and connected at about 5-percent volume or more, 2) argillic alteration, and 3) increased porosity related to wet, open fractures and brecciation. Resistivity highs will be associated with zones of silicification, intrusion, or basement uplifts (Senterfit and Klein, 1992; Zonge and Hughes, 1990; Irvine and Smith, 1989; Allis, 1990; Doyle, 1990, Frischknecht and others, 1986).

High induced polarization (IP) will be associated where pyritization has developed (Zonge and Hughes, 1991).

D. Shape and size of deposit (Buchanan, 1981; Heald and others, 1987):

Element	**Shape**	**Average Size (Range)**
Vein system, or district	lenticular, interlaced	3 km (1-9 km) width, 7 km (2-21 km) length, probably 2-4(?) km depth extent.
Ore deposit	lenticular, interlaced, discontinuous	width and length is highly variable, but a fraction (0.2?) of vein system; vertical extent of ore averages 500 m; paleodepths to initiation of ore (relative to original surface) is about 400 m (200 to 700 m).
Alteration halo	symmetric with the vein system; siphon shape in cross-section, narrowing with depth; capped with siliceous sinter that is often missing because of erosion.	width is of the order of 2 or 3 x the width of the vein system, roughly centered linearly on the vein system, wider on hanging wall if appreciable dip is present.

E. Physical properties

Bracketed values are averages. Source references are indexed with trailing letters. Queried values are guesses.

Property [units]	**Deposit** [silicic - potassic vein]	**Alteration** A: argillic (illite-kaolin) P: propylitic (chlorite-minor kaolin)	**Volcanic host**

1. Density (gm/cm^3)	quartzite 2.6[29]	?	rhyolite [2.5][29] andesite [2.7][29] tuffs 1.5-2.5[14]
2. Porosity (percent)	2-5?	5-20?	3-50[6,18] (fig.22 in)[18]
3. Suscepti- bility (cgs)	negligible?	negligible?	rhyolite [$.3 \times 10^{-3}$][23, 27]; undifferentiate d Tertiary volcanic rocks: (Arizona) $.05 \times 10^{-3}$ - 5×10^{-3}, (northern Montana), [0.7×10^{-3}][37,3].
4. Remanence (cgs-emu/cc)	negligible?	negligible?	undifferentiate d Tertiary volcanic rocks: (Arizona): $.005 \times 10^{-3}$ - 100×10^{-3}, (northern Montana): [11.1×10^{-3}][37,3].
5. Resistivity (ohm-m)	high; greater than 1,000?	A: low; less than 10? P: low; less than 100?	Tertiary volcanic rocks (Arizona): 20-2,000 [38].
6. Induced polarization (IP) (percent-frequency effect:PFE)		?	PFE > 10 with about 2 or 3-% disseminated sulfides[4]. Tertiary volcanic rocks (Arizona): <5[4].
7. Seismic sound (Vp) velocity (km/s)	quartzite 5.37-5.63[6]	lower?	wet tuffs: 2.61-3.92[6]; rhyolite: [3.27] 2.94-4.90[6]; volcanic breccia: 4.22[6]; (measurements at 0.1 kb; anisotropy is high (17-26% in some rhyolites with >10 (measured 18-32%) porosity.

8. Radio-elements K (%) U, Th (ppm)	K: high? U, Th ?	K: high? U, Th ?K:	moderate to low? U,Th ?

F. Remote sensing characteristics

In areas of low to moderate cover, remote sensing images in the visible and infrared bands can be processed to identify exposures of oxidized argillic alteration, and anomalous silicification, although there is considerable room for non-unique discriminations for moderate to low-spectral-resolution systems (Rowan and others, 1974, 1977; Podwysocki and others, 1983; Watson, 1985, Watson, 1990, Watson and Raines, 1990). The basis of such discrimination are the following:

a) In the visible through infrared wave-spectrum, ferric iron, water, and hydroxyl complexes have narrow (about 0.1 μm) and characteristic reflective minimums between 0.4 and 2.4 μm (Rowan and others, 1977).

b) Silica-rich assemblages have emissivity minimums near 8-10 μm (Watson and others, 1990).

c) These reflective and emissivity minimums can distinguished with moderate resolution (0.1 μm) airborne- or spacecraft spectral scanners.

There are often distinctive spectral waveforms for other minerals and mineral assemblages (Hunt, 1989), that require high-resolution (.01 to .001 μm), spectral scanners, currently available only on airborne systems (Watson and Raines, 1989).

G. General Comments

H. References

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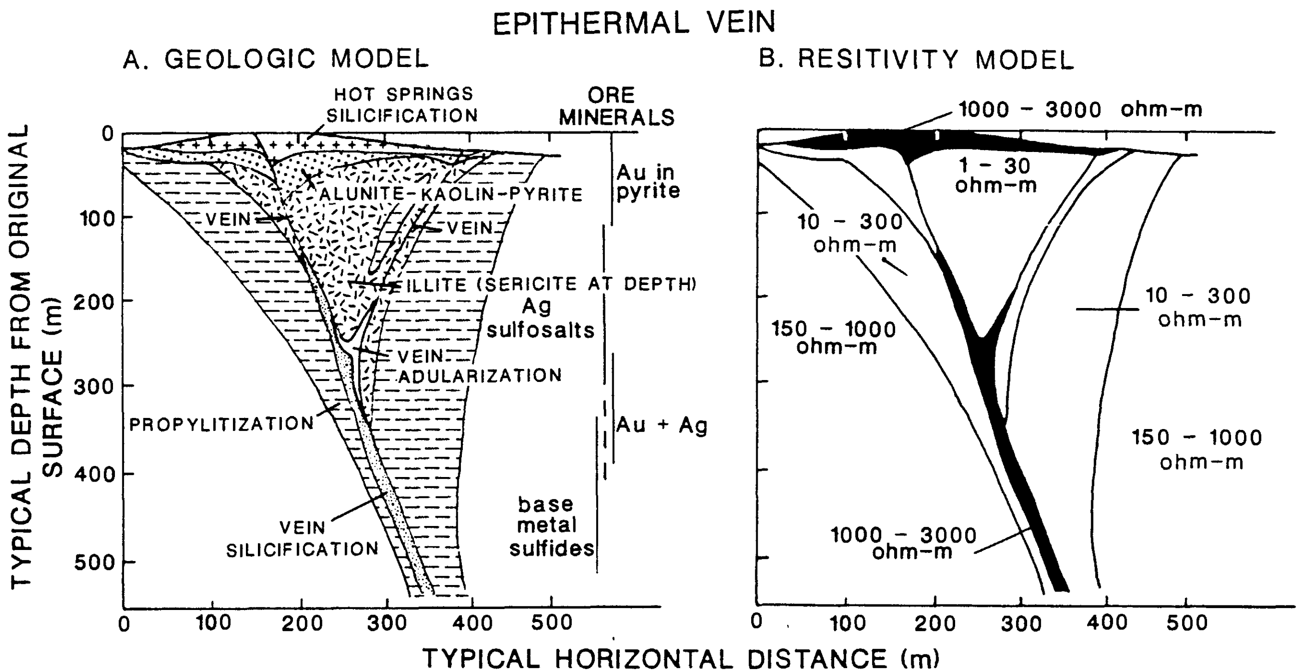


Fig. 1 -- A) Conceptual model of alteration for an epithermal vein (adapted from Buchanan, 1981; Irvine and Smith, 1990); B) Inferred Electrical resistivity for the model shown in (A).

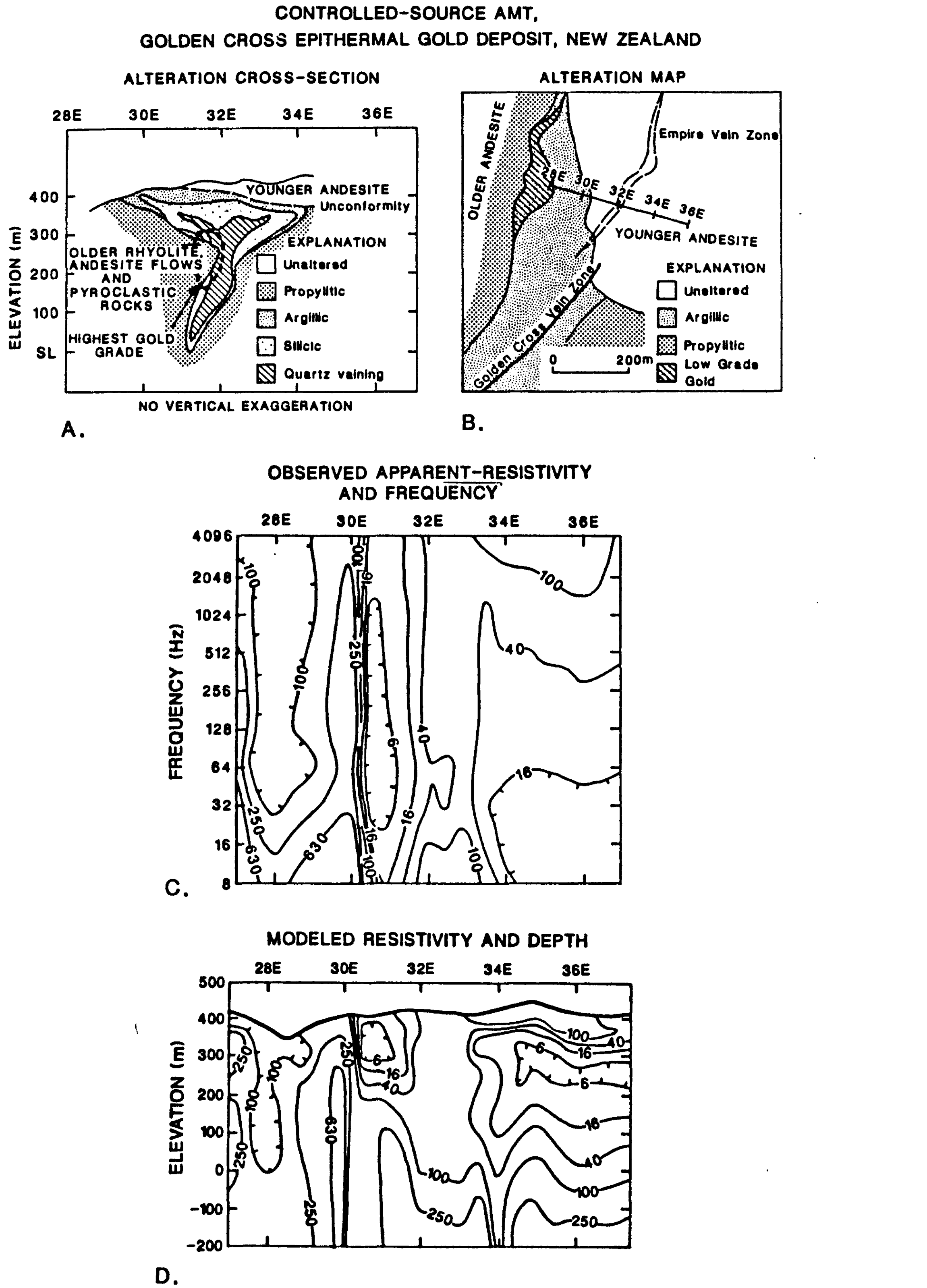
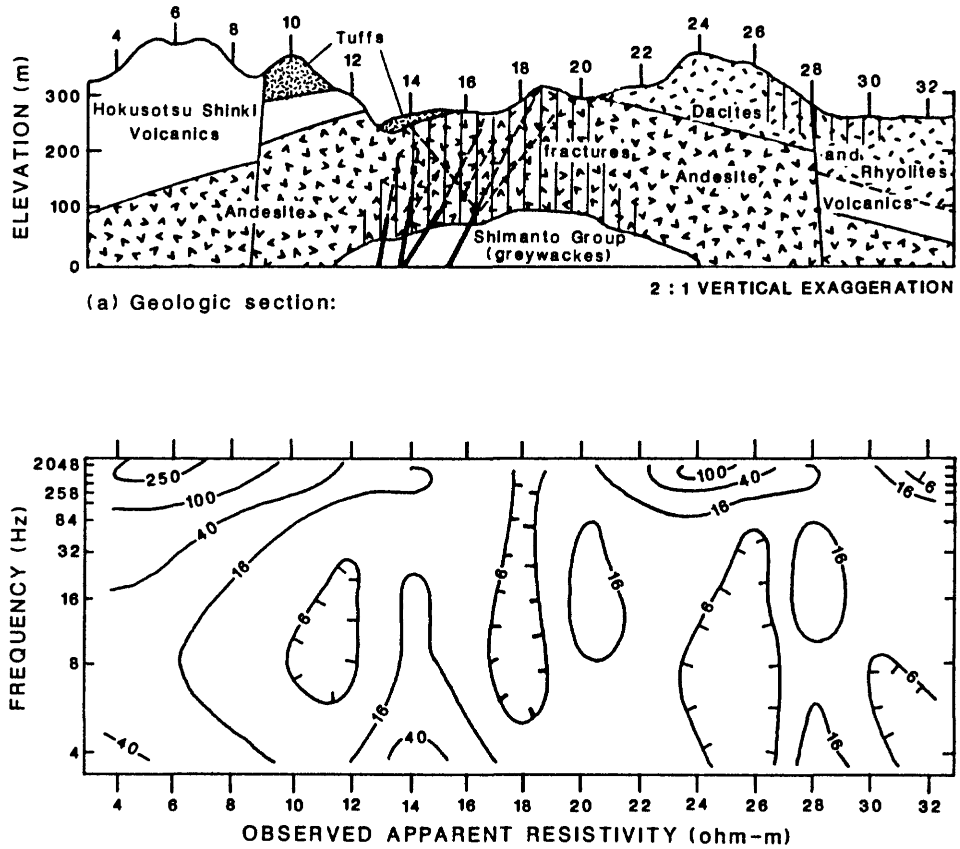


Fig. 2 -- Electrical resistivity data from controlled source audio-magnetotelluric (CSAMT) data over an epithermal vein system on the northern coast of North Island, New Zealand (adapted from Zonge and Hughes, 1991). A) Simplified alteration map of survey area showing the location of electrical traverse. B) Simplified interpretation of the alteration and veining along the cross-section traversed. C) Electrical pseudo-section of apparent resistivity vs. frequency. D) Electrical section of resistivity vs. depth resulting from inversion of data. On (C), the high-resistivity (> 250 ohm-m) silicified zone, bordered by anomalous lows (< 100 ohm-m) forms a prominent vertical electrical structure. The andesitic to rhyolitic host rocks show resistivities varying from about 5 to 500 ohm-m.

HISHIKARI EPITHERMAL GOLD DEPOSIT, KYUSHU, JAPAN



(a) Geologic section:

(b) Resistivity pseudo-section from controlled-source audiomagnetotelluric traverse

Fig. 3 -- Electrical resistivity data from controlled source audio-magnetotelluric (CSAMT) data over the Hishikari epithermal vein system in Kyushu, Japan (adapted form Zonge and Hughes, 1991). (A) Simplified geologic section across the deposit. B) Electric pseudo-section of apparent-resistivity versus frequency across the deposit. The Hishikari deposit is within an active geothermal system. Gold mineralization was found in veins in the Shimanto Group. The overall low resistivity is related to rock saturated with hot-water. Resistivity variations are associated with variable porosity and temperature, with the low resistivity under sounding 18 interpreted as a primary fracture system that continues into the mineralized vein system in the Shimanto Group (Zonge and Hughes, 1991, p. 797).

GEOPHYSICAL MODEL OF CARBONATE-HOSTED Au-Ag

COX AND SINGER MODEL No. 26a

Compilers - **W.D. Heran**
D.B. Hoover

A. Geologic Setting

- Regionally adjacent to or along high-angle normal fault zones related to continental margin rifting, or regional thrust faults or bedding.
- Selective hydrothermal replacement of carbonaceous limestones or dolomite where these are intruded by igneous rocks.
- Very fine grain native gold and/or silver, pyrite and arsenic sulfide disseminated in host rocks and associated silica replacement.

B. Geologic Environment Definition

Remote sensing data can define major lineaments and tectonic structural zones and their intersection with major fault systems (Rowan and Wetlaufer, 1981). Remotely sensed data can define major lithologic boundaries and areas of alteration (Kruse and others, 1988). Aeromagnetic and gravity methods have been used to delineate the margins of intrusions in the near subsurface and determine major faults beneath sedimentary cover (Grauch, 1988, Grauch and Bankey, 1991). Radioelement data have possibilities for defining zones of hydrothermal alteration associated with faulting (Pitkin, 1991). Airborne electromagnetic resistivity data have been used to map lithology and detect alteration in addition to delineating structure at the surface and under shallow cover (Taylor, 1990; Pierce and Hoover, 1991; Hoover and others, 1991; and Wojniak and Hoover, 1991).

C. Deposit Definition

High angle fault zones and shear zones can be mapped by a variety of electromagnetic methods as conductive anomalies within sediments or crystalline rocks (Hoekstra and others, 1989; Hoover and others, 1984; Heran and Smith, 1984; Heran and McCafferty, 1986). Detailed magnetic and gravity surveys can be employed to delineate pluton margins, map major fault zones, lithologic boundaries and determine depth of alluvial cover (Grauch, 1988). Electrical resistivity methods are able to map hydrothermal alteration and faulting as a resistivity low and silicification caps as a high (Hallof, 1989; Corbett, 1990; and Hoekstra and others, 1989). Seismic methods have been used to delineate high angle faults, lithologic contacts and hydrothermal alteration (Cooksley and Kendrick, 1990). Gold bearing structures containing clay or carbonaceous (graphite) material can be mapped using the induced polarization method. Radiometric surveys have possibilities for mapping alteration along faults (Porter, 1984).

D. Size and Shape of	**Shape**	**Average Size/Range**
Deposit	Tabular to highly irregular	3.9,0.4-20x10^6 m^3
Alteration Haloe/s	Variable, irregular	?
Cap	Irregular blanket	?

E. Physical Properties	**Deposit**	**Alteration**	**Cap**	**Host**
Description	Hydrothermal replacement of calcareous rock	-	Siliceous replacement	calcareous rocks
1. density (gm/cc)	2.6	-	?	2.65-1.9-2.9[1]
2. porosity	?	?	?	
3. susceptibility (10^{-6}cgs)	0.0-?	-	low	20-0-280[1]
4. remanence	low	-	low	?
5. resistivity (ohm-m)	20,10-50	-	variable	1500-350-6000[1]
6. chargeability (mv-sec/v)	30,20-40[1]	-	?	5.2-20[1]
7. seismic vel. (km/sec)	2.5[2]	-	?	4.3-6[2]
8. radioelements				
K%	variable up to 4.5	-	?	0.27
U-ppm	variable up to 10	-	?	2.2,0.1-9.0
Th-ppm	variable up to 16	-	?	1.7,0.1-7
9. Other (specific)				

F. Remote Sensing Characteristics

Visible and Near IR - Regionally, landsat images have been used to delineate lineaments and major structural zones in Nevada (Rowan and Wetlaufer, 1981). Distinctive signitures can be detected from hydrothermal alteration products exposed at the surface (ie., illite, kaolinite, montmorillonite, jarosite and alunite).

Thermal IR - Emissivity-ratio images prepared from data acquired by the Thermal Infrared Mapping System (TIMS) have been used to detect and map previously unrecognized silicified carbonate host rocks at the Carlin deposit, Nevada (Watson and others, 1990).

G. Comments

Intense exploration for this deposit type in the 80's has generated a variety of geophysical applications many of which are helpful in areas of cover. Regional exploration can be greatly aided by remote sensing and airborne EM. Ground EM profiling is the best bet for locating faults. Electrical or EM techniques can define areas of alteration.

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GETCHELL MINE, NEVADA
RESISTIVITY AND RADIOELEMENT DATA

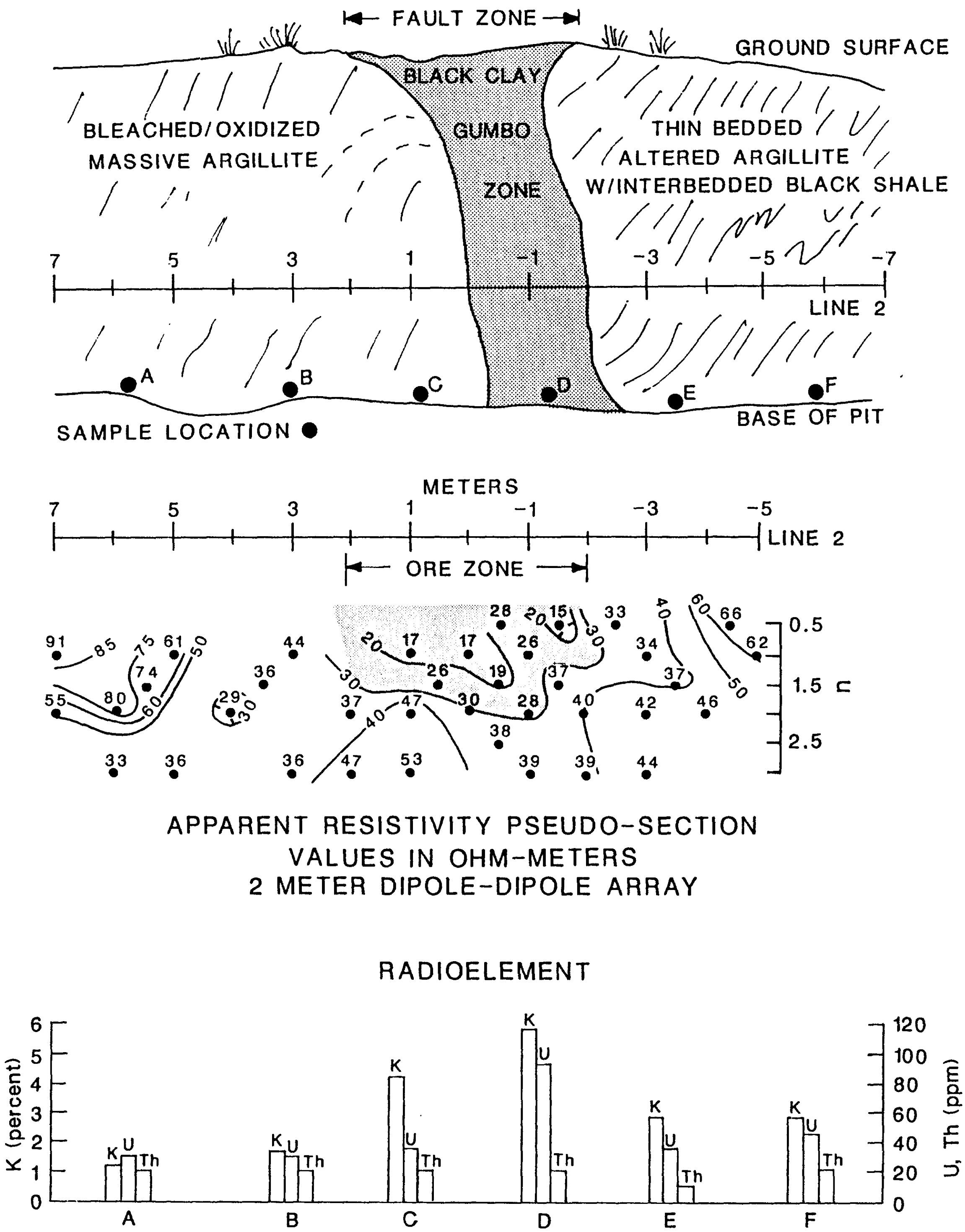

Figure 1. Apparent resistivity and radioelement data obtained in the South Pit at the Getchell Mine, Humbolt County, Nevada. Short-dipole in-situ electrical measurements taken across a mineralized structure, show low resistivities correlated with the ore zone. Radioelement data show significant increases in K and U in the ore zone (Heran and Smith, 1984).

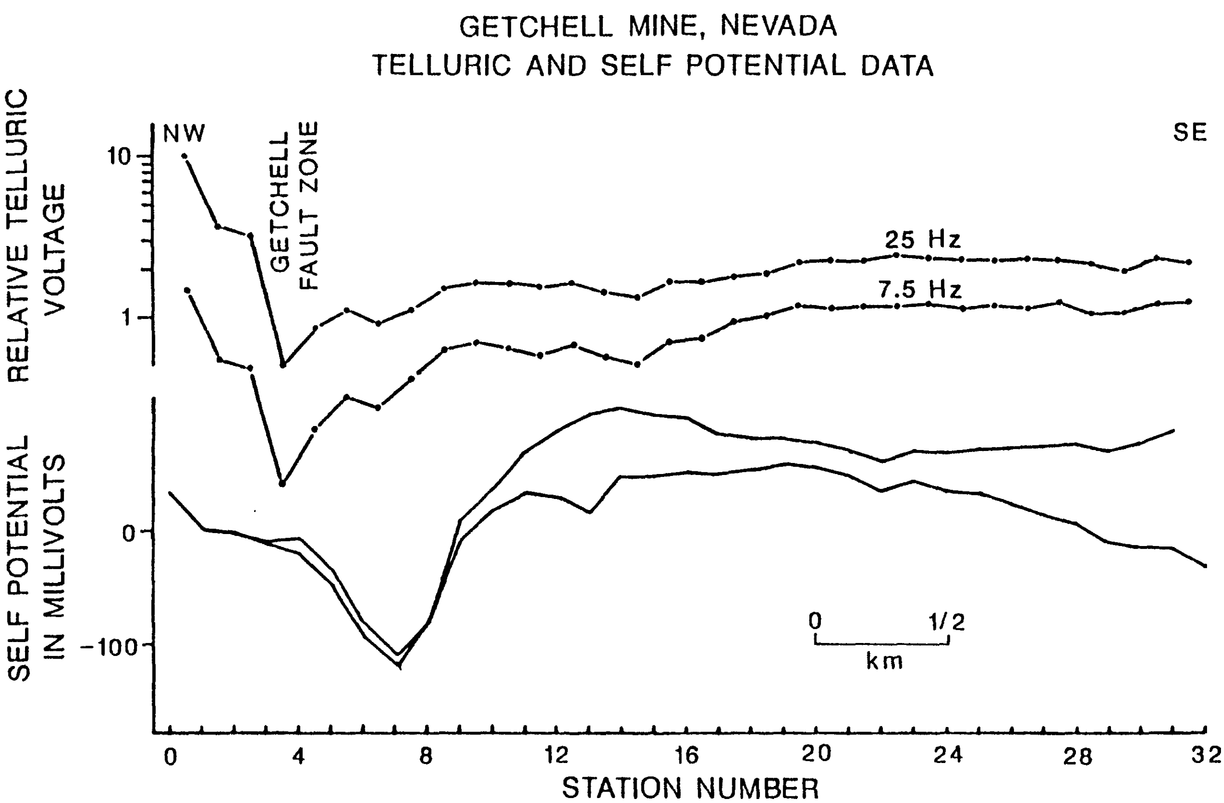

Figure 2. Telluric and self potential data across the northern end of First Miss Gold's Summer Camp deposit prior to mining. The Getchell fault zone shows as a pronounced low in relative telluric voltage, indicating that the fault zone at 25 and 7.5 Hz has much lower resistivity than the surrounding rocks. A broad self potential low is observed east of the main fault zone which is inferred to be related to graphitic material along a parallel fault (from Hoover and others, 1986).

GEOPHYSICAL MODEL OF OLYMPIC DAM

COX AND SINGER MODEL NO. 29B

Geophysically similar models-No. 10 Carbonatites; No. 12, Diamond pipes

Compilers - D.B. Hoover
L.E. Cordell

A. Geologic Setting

- Pipe-like structure emplaced within Proterozoic anorogenic alkali-granite basement, along a regional basement fracture system.
- Deposit has 350 m of unmineralized late Proterozoic and Cambrian sedimentary cover.
- Model covers only one deposit, has been subject to significant modification since discovery, and is probably still subject to change.
- Commodities are Cu, U, and Au.

B. Geologic Environment Definition

Deposit is on the Stuart shelf in the extreme northwest part of the Adelaid geosyncline, emplaced within the Gawler craton. The Adelaid geosyncline is inferred to be a failed rift, that partially opened in the south (White, 1983). The syncline is defined geophysically by a central gravity high with flanking lows, and bounded by sub-parallel lineaments seen in gravity, magnetic, and remote sensing data. The deposit is located at the intersection of a major west-northwest trending photolineament, and a north-northwest trending gravity lineament (Roberts and Hudson, 1983).

C. Deposit Definition

Originally defined by coincident gravity (-18 mgal) and broad magnetic (+1000 nT) anomalies. Discovery site selection was based on lineament analysis, and coincidental gravity and magnetic anomalies from sources shallow enough to test by drilling. Original exploration model was based on a basaltic Cu model (Cox and Singer model no. 23), where the magnetic high was believed due to extensive basalts, and the gravity high due to a basement horst block within the volcanics (Rutter and Esdale, 1985). However, predicted depths to the anomalous magnetic and gravity sources were 2000 m, and 1150 m respectively, raising some initial questions about the model. Seismic reflection data identified a strong reflector at 350 m, that suggested the source might be shallow. After discovery, gamma-ray logging showed that uranium content was very high, to 600 ppm, thus the deposit if not covered would be detectable by its radioelement signature. The deposit is also characterized by low resistivity and increased polarization relative to the host granite.

D. **Size and Shape of**	**Shape**	**Average Size/Range**
Deposit	Vertical cylinder	Diam. 3 km; height >800 m
Alteration haloe	Irregular	Dolerite in pipe least affected, not geophysically significant?
Cap	Not present	-----------------

E. Physical Properties (units)	Deposit Alkalic-granite + hematite breccia pipe	Alteration chlorite, hematite quartz	Cap none	Host anorogenic alkalic-granite
1. Density (gm/cm^3)	3.5 average 3.0-4.5[1]	?	-	2.67?
2. Porosity	medium high?	?	-	low
3. Susceptibility (cgs)	$8x10^{-3}$ average $2x10^{-4}$-$3x10^{-3}$[1]	?	-	?
4. Remanence	?	?	-	?
5. Resistivity (ohm-m)	high variable 0.1-100's[1,3]	?	-	?
6. IP Effect (mv-sec/V) (mradians)	 60-average[1] 20-120[3]	?	-	low
7. Seismic Velocity (km/sec)	low	?	-	high
8. Radioelements				
K (%)	high?	?	-	?
U (ppm)	440 to 640[5]	?	-	?
Th (ppm)	?	?	-	?
9. Other heat-flow (mw/m^2)	120-275[2]	-	-	66-82[2]

F. Remote Sensing Characteristics

Visible and near IR--Presence of a major, broad (up to 48 km wide), continental lineament important in original area selection (O'Driscoll and Keenihan, 1980). If not covered, hematite, chlorite, sericite, and silica alteration should be definable.

G. Comments

The gravity anomaly is explained by the presence of the hematite-rich breccia. The source of most of the magnetic anomaly is deeper than presently explored (1150 m), but generally assumed directly related to the deposit. The high heat flow is due to the highly elevated uranium content. The geophysical signature of this single deposit is indistinguishable from that of carbonatites and similar to that of diamond pipes.

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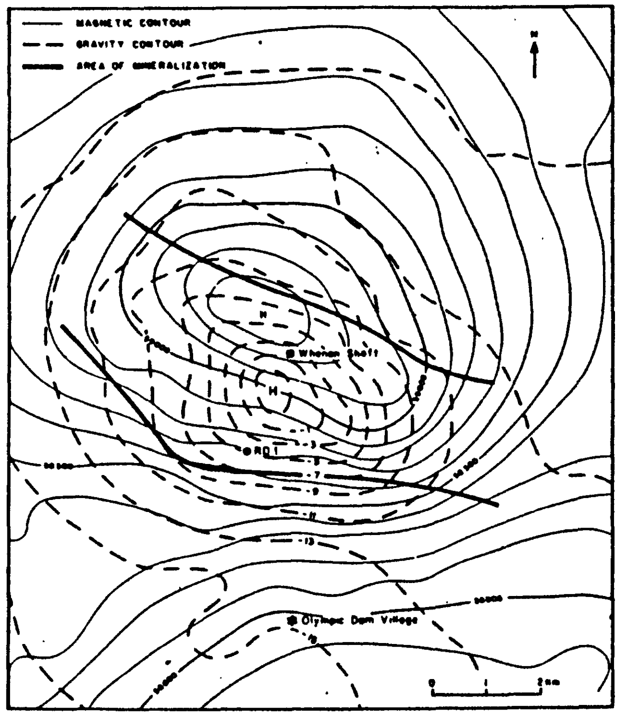

Figure A. Gravity and magnetic anomalies at Olympic Dam, Australia area of mineralization is located between the dotted lines. Adapted from? Roberts and Hudson (1983).

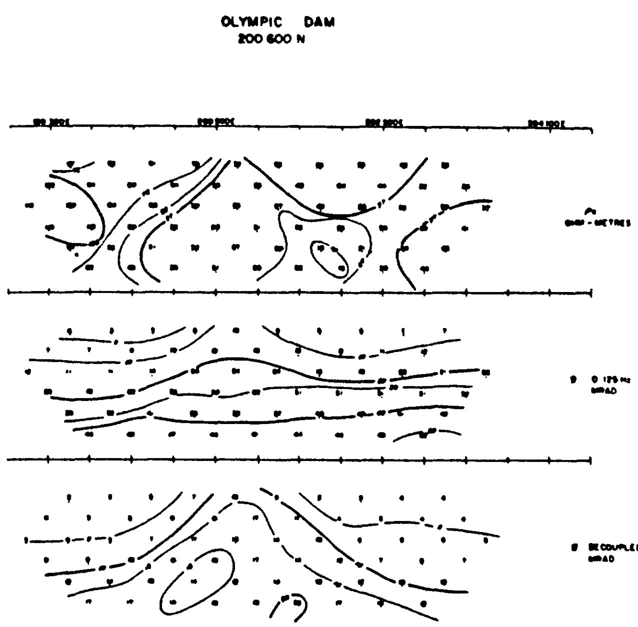

Figure B. Induced polarization section across the Olympic Dam deposit. Adapted from Esdale and others (1987).

GEOPHYSICAL MODEL OF LOW SULFIDE Au-QUARTZ VEINS

Cox and Singer Model No. 36a Compiler - **W.D. Heran**

Geophysically similar models - no. 36b - Homestake Au

A. Geologic Setting

- Within greenstone belts, oceanic metasediments, regionally metamorphosed volcanic rocks, alpine gabbro and serpentine, graywacke, chert, shale and quartzite. Also late granitic batholiths.
- Fault and joint systems produced by regional compression.
- Gold hosted in massive quartz veins that generally post date regional metamorphism, and are persistent along regional high angle faults and joint sets. Disseminated ore occurs where veins cut granitic bodies.

B. Geologic Environment Definition

Remote sensing can be utilized to map regional lineaments, faults and joint systems, and lithologic boundaries (Longman, 1984; Clark and others, 1990; Crosta and Moore, 1989). Aeromagnetic data can delineate greenstone belts, as a high or a low depending if the belt is magnetic-rich or magnetic-deficient (Grant, 1984). Aeromagnetic surveys have been used to define regional faults and shears, and map the distribution of mafics, ultramafic and sediments within greenstone terrains (Lindeman, 1984; Boyd, 1984; Whitaker and others, 1987; Isles and others, 1989). Combined magnetic and gravity surveys have been used to map faults, shears, and other structures where basement rocks are covered (Whitaker and others, 1987; Suppel, and others, 1986). Gravity data and electrical soundings have been used to determine depths of greenstone sequences (DeBeer, 1982).

C. Deposit Definition

Aeromagnetic data can be useful at the deposit scale in identifying altered ultramafics, shown as flat areas or shallow magnetic lows, and delineating local structures, stratigraphy and intrusives (Pickette, 1982). Ground follow-up magnetic surveys would help further delineation of lithology and structure. Au-quartz veins hosted in mafic rocks are often characterized by linear magnetic lows related to magnetite destruction (Patterson and Hallof, 1991). Electromagnetic methods can be used to map faults or shear zones (conductive response if open), veins (resistive response), geologic contacts and alteration (Subrahmanyam and Jagannadham, 1984). Massive quartz veins can be mapped by detailed electrical resistivity or IP surveys as resistivity highs (Macnae, 1988; King, 1957; Murdoch, 1984). Radioelement surveys can be used to detect alteration (Costa and Byron, 1988). Remote sensing methods can detect minerals such as quartz, siderite, ankerite, sericite, and carbonates if they are present in great enough quantities (Knepper, 1989).

D. **Size and Shape**	**Shape**	**Average Size/Range**
Deposit	Thin tabular	$1.\times10^4m^3/3.8\times10^2m^3-3.5\times10^5m^3$
Alteration Haloe/s	Irregular	
Cap		

E. **Physical Properties** (Units)	**Deposit**	**Alteration**	**Host**
1. density	2.6[1] average	?	*
(gm/cc)	2.4-2.8	?	*
2. porosity	?		
3. susceptibility	100[1] average	?	*
(10^{-6} cgs)	10-8000	?	*
4. remanence	?		
5. resistivity	3000[1] average	?	*
(ohm-m)	1000-12000	?	*
6. chargeability	25[1]	?	*
(mv-sec/V)	10-80	?	*
7. seismic vel.	(4-6)	?	*
(Km/sec)			
8a. radiometric K	(low-high)	?	*
b. U	(low)	?	*
c. Th	(low)	?	*

F. Remote Sensing Characteristics

Remote sensing applications to exploration are based on identifying indirect indicators of potential host rocks including spectral, albedo, and textural characteristics. Potential host rocks composed of iron oxides and carbonate minerals can be uniquely identified with high spectral resolution instruments (imaging spectrometers) in the visible and near-infrared (Rowan and others, 1983; Clark and others, 1990). More importantly, imaging spectrometer data can be used to identify and map the distribution of specific iron oxide species (Taranik and others, 1991). Broad-band data in the visible and near-infrared, such as Landsat Thematic Mapper, are effective for separating carbonate- and iron oxide-bearing potential host rocks from other lithologies on regional and local scales (Knepper, 1989). Enhanced Landsat data have been used to define lineaments, fracture patterns and major structures (Longman, 1984). Airborne MSS data can delineate faults, joints and stratigraphic units (Honey and Daniels, 1985).

Thermal IR - Chlorite may be detectable in host rocks; siderite, ankerite, sericite, and carbonates at the deposit scale.

G. Comments

Geophysical data in the literature are very sparse for this deposit type, at the deposit scale. Regionally data are fairly abundant within greenstone terrains.

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GEOPHYSICAL MODEL OF HOMESTAKE Au

COX AND SINGER Model No. 36b Compiler - **W.D. Heran**

Geophysically similar models - No. 36a Low Sulfide Au-Quartz Veins

A. Geologic Setting

- Mainly within Archean age regionally metamorphosed (greenschist-facies) mafic and felsic metavolcanic rocks, komatiites, and volcaniclastic sediments interlayered with banded iron-formation. Greenstone units typically intruded by felsic plutons and locally by quartz and/or syenite porphyry.
- Deposits are common near regional division between predominantly metavolcanic and metasedimentary rocks in greenstone belt.
- Stratabound to stratiform deposit consisting of bedded ores of native gold with various sulfides in Fe-rich siliceous or carbonate-rich chemical sediments overlying vein and stockwork feeder zones, often interlayered with flow rocks. Beds may be cut by quartz-carbonate veins containing gold. Deposits are commonly structurally controlled.

B. Geologic Environment Definition

Remote sensing data can delineate regional lineaments, major structural zones, lithologic boundaries and areas of hydrothermal alteration (Honey and Daniels, 1985; Crosta and Moore, 1989; Yatabe and others, 1984; Longman, 1984). Greenstone belts can be outlined by aeromagnetic surveys, which may reflect a regional magnetic low if the belt is magnetite-deficient, in other cases a high if it is magnetite-rich (Grant, 1985). Aeromagnetic surveys are used to define regional structures and locate iron rich metasediments and mafic and ultramafic volcanic rock, within the greenstone belt (Lindeman, 1984; Boyd, 1984). Airborne magnetic data may also define intrusives at the edges or within greenstone belts which may be magnetite deficient compared to normal granitoid rocks (Grant, 1984). Combined airborne EM/magnetic surveys have been used in mapping structure within greenstone belts (Boa Hora, 1986). Airborne radioelement surveys can delineate high potassium zones related to sericite alteration and help define lithologic boundaries (Cunneen and Wellman, 1987). Gravity can be utilized to help determine the depth of belt rocks, define shear zones and folded structures or locate buried intrusives (Costa and Byron, 1988). Electrical soundings and gravity data have been used to model maximum depths of greenstone sequences (DeBeer, 1982).

C. Deposit Definition

Detailed magnetic surveys have been used to map banded iron formations; predict strike extensions, bedding thickness and dip of magnetic zones within the stratigraphic sequence (Lindeman, 1984) and help unravel structure that controls mineralization (Pemberton and others, 1985). Also, detailed magnetic data are employed to map intrusives and dikes associated with ore zones (Koulomzine and Brossard, 1947) and identify alteration which involves both the formation and destruction of magnetic minerals (Fuchter and others, 1991). The strong association of gold with sulfides has permitted the use of a variety of electromagnetic methods to map these zones as conductors (Lindeman, 1984; Valliant, 1985; Costa and Byron, 1988. EM techniques are also used to help map stratigraphy and structure (Pemberton and Carriere, 1985). The induced polarization method is effective in mapping sulfides as resistivity lows and as positive zones of increased polarization (Mathisrud and Sumner, 1967; Sheehan and Valliant, 1985; Hallof, 1985). The IP method can be used to distinguish between mineralized and non-mineralized conductive (EM) anomalies (Costa, and Byron, 1988). IP has been used successfully underground to map pencil-like ore shoots (Mathisrud, and Sumner, 1967). The Mise-a-la-masse electrical technique has been used to delineate the size, shape, and position of individual mineralized units within a sequence (Polomé, 1989). Radiometric surveys can also be used to define areas of hydrothermal alteration (Costa and Byron, 1988).

D. Size and Shape of	Shape	Average Size/Range
Deposit	layered sheet or lens	$0.3 \times 10^6 m^3$, $.03-3.9 \times 10^6 m^3$
Alteration	irregular	

E. Physical Properties	Deposit	Alteration	Host
1. Density (gm/cc)	3.1[1]; average 2.9-3.4[1]	?	*
2. Porosity	?	?	*
3. Susceptibility (10^{-6} cgs)	500[1] average 0-5000[1]		*
4. Remanence	?	?	*
5. Resistivity (ohm-m)	1[1] average .1-10[1]	?	*
6. IP Effect			
chargeability (mv-sec/v)	50[1] average 20-200[1]	?	*
percent freq. effect (PFE)	12.5[1] ave 5-50[1]	?	*
7. Seismic Velocity (km/sec)	?	?	*
8. Radiometric			
K (%)	moderate-high	moderate-high	*
U (ppm)	moderate-very high	variable	*
Th (ppm)	variable	variable	*

F. Remote Sensing Characteristics

Remote sensing applications to exploration are based on identifying indirect indicators of potential host rocks including spectral, albedo, and textural characteristics. Potential host rocks composed of iron oxides and carbonate minerals can be uniquely identified with high spectral resolution instruments (imaging spectrometers) in the visible and near-infrared (Rowan and others, 1983; Clark and others, 1990). More importantly, imaging spectrometer data can be used to identify and map the distribution of specific iron oxide species (Taranik and others, 1991). Broad-band data in the visible and near-infrared, such as Landsat Thematic Mapper, are effective for separating carbonate- and iron oxide-bearing potential host rocks from other lithologies on regional and local scales (Knepper, 1989). Enhanced Landsat data have been used to define lineaments, fracture patterns and major structures (Longman, 1984). Airborne MSS data can delineate faults, joints and stratigraphic units (Honey and Daniels, 1985).

G. Comments

Regional exploration for and within greenstone terranes has commonly employed aeromagnetic data and more recenty radioelement and remotely sensed data, in Australia, Canada, and Brazil. In general, greenstone terranes have a *low and rough* magnetic character, meaning a low background level with numerous intense short-wavelength anomalies (Grant, 1985).

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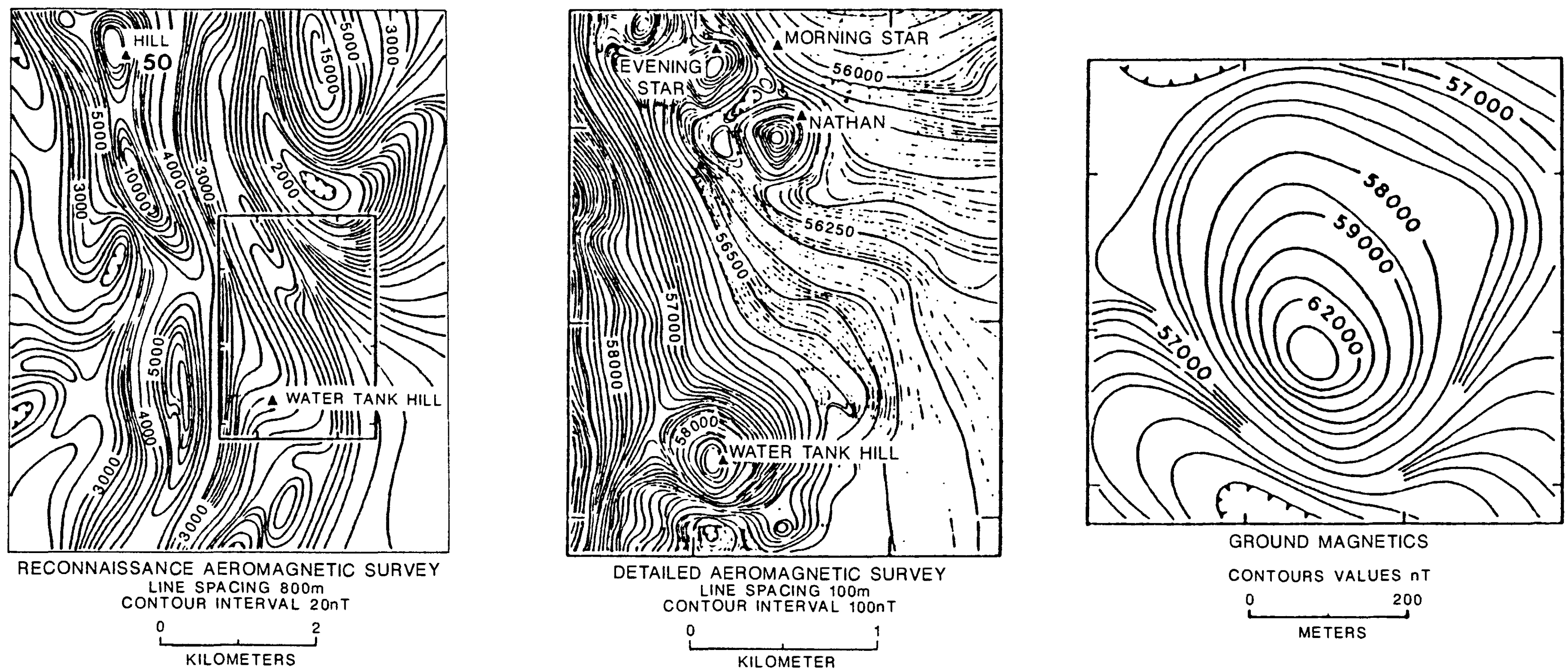

Figure 1. Magnetic data from Water Tank Hill, Mt. Magnet area, Western Australia. The recon aeromagnetic data (800 m) does not define the deposit well, but the detailed (100 m) data show an obvious oval response. Detailed ground magnetics further define its signiture. (peak response is 7000 nT above background) (after Lindeman, 1984)

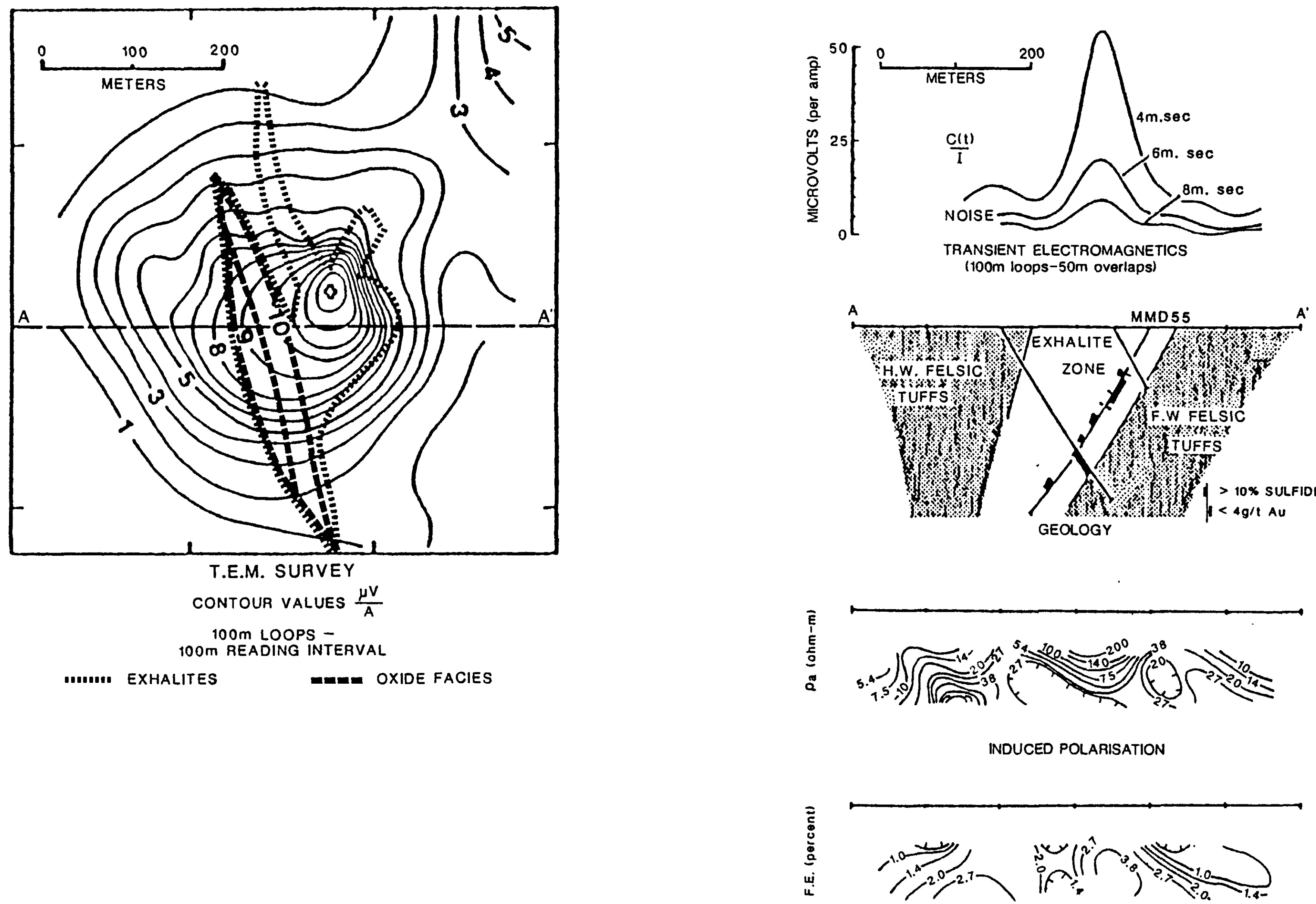

Figure 2. Geology, Transient EM and IP data from Water Tank Hill, Mt. Magnet area, Western Australia. The discovery hole MMD 55 was drilled to test the TEM and IP anomalies. The apparent resistivity high appears to relate to BIF (exhalite zone). (after Lindeman, 1984)

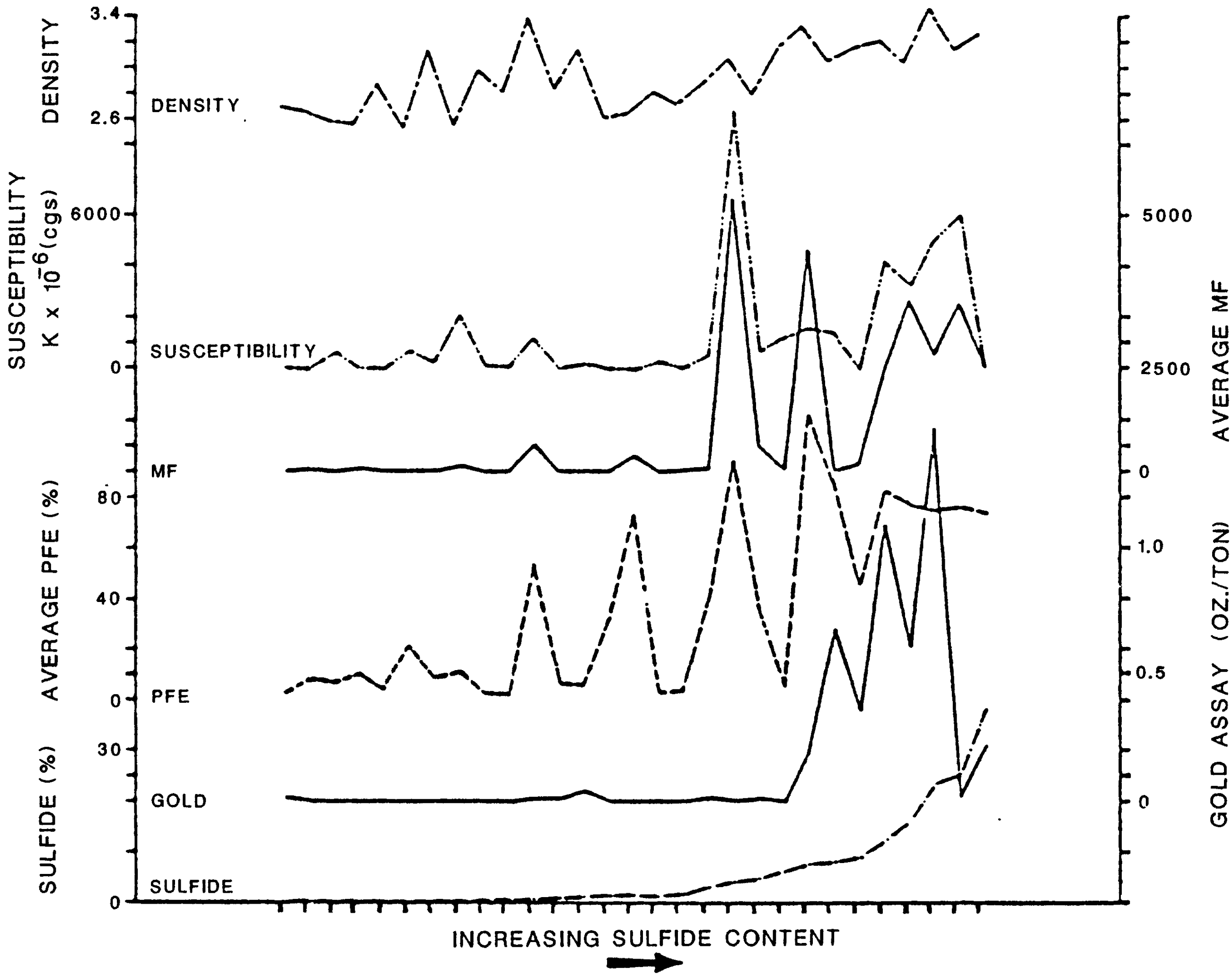

FIGURE 3
LABORATORY PHYSICAL PROPERTY MEASUREMENTS ON CORE SAMPLES FROM THE HOMESTAKE MINE (FROM MATHISRUD AND SUMNER 1967).

GEOPHYSICAL MODEL OF PLACER Au-PGE AND PGE-Au

COX AND SINGER MODEL NO. 39 A&B Compilers - **W.D. Heran and W. Wojniak**

Geophysically similar models - No. 39c Shoreline Placer Ti
No. 39d Diamond Placers
No. 39e Alluvial Placer Sn

A. Geologic Setting

- Shield areas where erosion has produced multicycle sediments; Paleozoic to Mesozoic accreted terranes; Tertiary conglomerate along major fault zones; low terrace deposits; high-level terrace gravels.
- Alluvial gravels and conglomerate with white quartz clasts and heavy minerals; sand and sandstone of secondary important. Cenozoic to Holocene in age.
- Elemental gold and platinum-group alloys in grains and nuggets in gravel, sand, and clay, and their consolidated equivalents, in alluvial, beach, eolian and (rarely) glacial deposits.

B. Geologic Environment Definition

Placers need a source deposit and high to moderate energy environment for their formation, so the reader is referred to the various gold and PGE models that would serve as sources for various placer deposits, to obtain the geophysical definition for those areas.

Remote sensing techniques used in conjunction with geologic, topographic and drainage pattern maps would be most useful for large scale reconnaissance (Macdonald, 1983). Multi-spectral imagery, infra red photography and side-scanning radar can be used to map unconsolidated material (Wilson, 1986).

C. Deposit Definition

Alluvial deposits are almost always less dense than the underlying bedrock. Detailed gravity measurements can be used to delineate modern or ancient drainage patterns and determine bedrock configurations beneath exposed gravels (Macdonald, 1983; Whiteley, 1971). Magnetite or ilmenite black sands are usually associated with the pay streak of a placer deposit so ground magnetic measurements may be utilized in prospecting or for ore zone definition, if the surrounding bedrock is less magnetic (Jakosky, 1950; Adler and Adler, 1985; Schwarz and Wright, 1988). Aeromagnetic data may be useful in exploration if it is of high resolution. Shallow seismic refraction and reflection methods can be used to delineate channels and measure the depth of gravels due to lower seismic velocity of placer materials (Whiteley, 1971). Electromagnetic and electrical sounding techniques can all help in defining the extent and thickness of alluvial deposits and to define structures, faults and lithologic boundary in the bedrock that may have served as natural traps. A combination of the seismic, gravity and resistivity methods have been used successfully for deposit definition (Daly, 1965; Tibbets and Scott, 1972). IP may be used to map variations in clay or sulfide content that maybe correlated to ore grade (Peterson and others, 1968). Other methods that are applicable to exploration and deposit definition are ground penetrating radar (Davis and others, 1984) and airborne and ground radioelement surveys.

D. **Size and Shape**

	Shape	Average size/range
Deposit Au/PGE	Variable, typically elongate tabular	$.55x10^6m^3/.011-25x10^6m^3$
Deposit PGE/Au	Variable, typically elongate tabular	$.05x10^6m^3/.0055-.9x10^6m^3$

E. **Physical Properties**

(units)	Deposit alluvium	Alteration na	Cap na	Host bedrock
1. density (gm/cc)	1.96-2.0 (wet), 1.98 1.5-1.6 (dry), 1.54			*
2. porosity	high			*
3. susceptibility (10^{-6}cgs)	10-500 100K-1Mill, 500K (magnetite)			*
4. remanence	?			*
5. resistivity (ohm-m)	10-800			*
6. chargeability (mv-sec)	1-4			*
7. seismic vel. (km/sec)	.1-2.4			*
8. radiometric				
K-%	1-5			*
U-ppm	2-5			*
Th-ppm	2-30			*

F. Remote Sensing Characteristics

Visible and Near IR - Alluvial deposits containing placers, not alluvial deposits in general, do not have unique spectral reflectance or thermal properties that are diagnostic. Remote sensing techniques for exploring for placers are based on the recognition of indirect indicators of alluvial deposits such as areas of high albedo (Abdel-Gawad and Tubbesing, 1974) and textural characteristics. Aerial photography and aircraft and satellite imaging data in the visible and near-infrared are useful for mapping landforms generally associated with alluvial deposits. In addition, soil moisture differences between alluvium and bedrock often promote selective vegetation growth that can be easily detected. These soil moisture differences may also be detected in sparsely vegetated terrain by thermal infrared and radar imaging techniques.

Thermal IR - Comparison of day and night (thermal infrared) satellite imagery have been used to assist mapping structurally controlled paleodrainage networks (Tapley and Wilson, 1985).

G. Comments

Placers can be highly variable in shape and lithology. The pay streak or ore zone makes up a small part of an alluvial deposit, typically 1-30 m wide by 10-1000 m long. Geophysical methods that work in one area may not in another. Regional exploration typically starts with remote sensing combined with photogeology which may be followed by airborne magnetics and spectrometry. The principal ground methods are seismic, gravity, electrical, magnetics, radar, and radiometrics. These techniques may be used singly, or in combination for direct detection of alluvium, or as an aid in mapping associated features. Magnetics may be used for direct detection of ore bearing zones if enough magnetite is associated with the pay streak.

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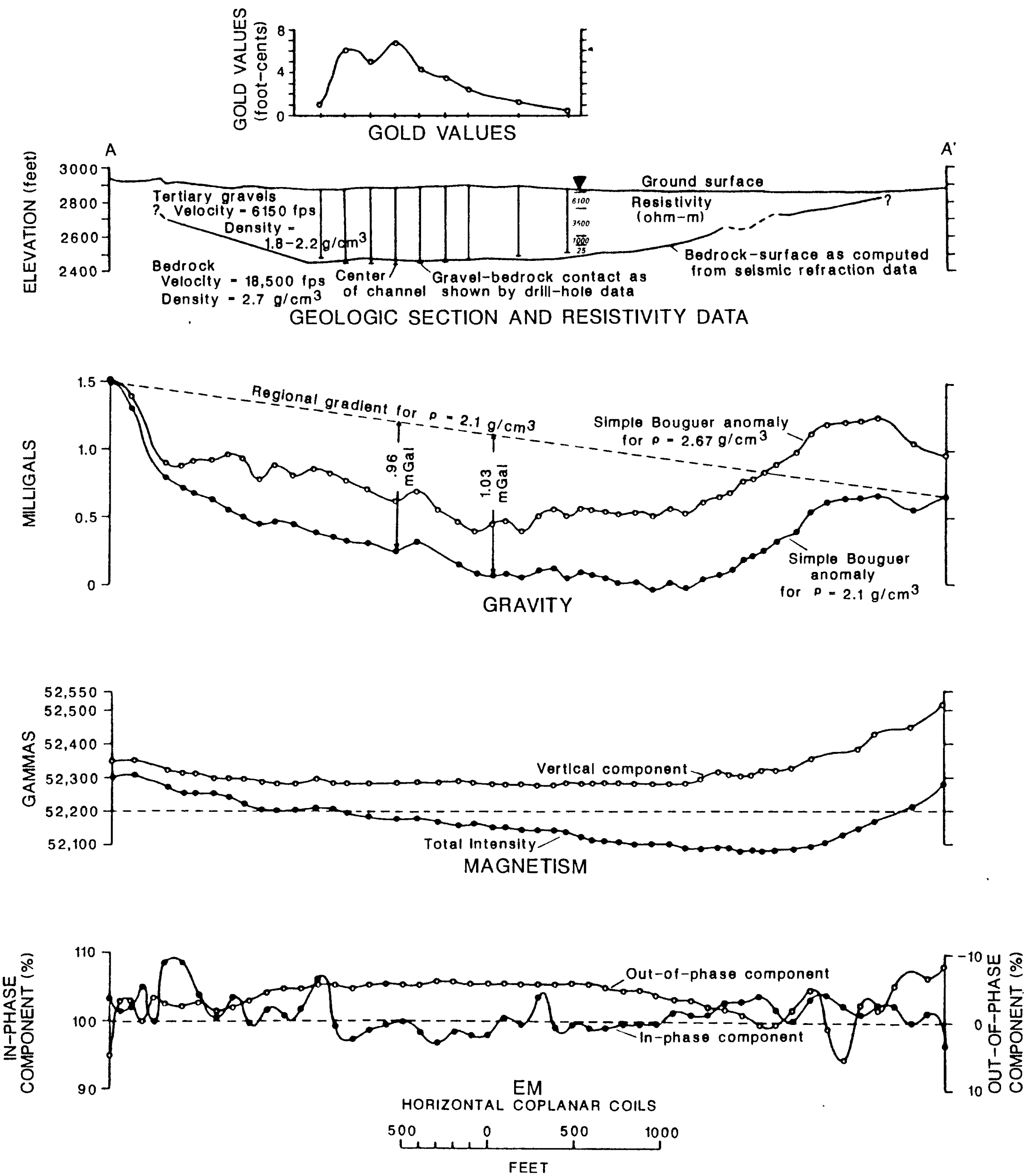

Illustrations showing seismic, magnetic, gravity, and electrical data across a channel in Tertiary gravels near North Columbia, California, adapted from Peterson and others, 1968.