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# GRAVITY AND MAGNETIC METHODS

by

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## Abstract

Spatial variations in the earth's gravity field are caused by lateral variations in rock denstiy. Field surveys routinely measure the gravity field to 1 part in $10^8$ and recent improvements in gravity meters have resulted in instruments capable of one or two orders of magnitude better than this. Because variations in measured gravity that are due to latitude and elevation differences among stations are usually much larger in magnitude than anomalies caused by geologic variations of interest in prospecting, corrections to remove latitude and elevation effects must be based on precise location and elevation information. Positive gravity anomalies are found over some massive sulfide and iron deposits facilitating direct detection of the orebody. Perhaps the most common applications of the gravity method are in aiding geologic mapping. For example, negative gravity anomalies are commonly associated with intrusive complexes and with relatively low-density alluvial basin fill, thus providing ways to map these and other features of interest in prospecting.

Spatial variations in the earth's magnetic field, of interest in exploration, are most commonly due to lateral variations in the distribution of the mineral magnetite. Continuing improvement in magnetometers has resulted in instruments capable of measuring the magnetic field to 1 part in $10^5$, or better, in routine survey applications. This is within the geological noise level for most applications.

The most common prospecting use of the magnetic method is in aiding geologic mapping through detection of anomalies caused by structure or rock - type changes. Direct detection of iron deposits and of magnetic skarn deposits is possible.

No interpretation of gravity or magnetic data above is unique, but ambiguity can generally be reduced through use of geological or other geophysical data. Modern interpretation techniques for both gravity and magnetic data are based around calculating the effects of an assumed model using a digital computer, comparing the model results with the field data, and modifying the model until a satisfactory match is attained. Interactive modeling programs using computer graphics greatly facilitate this process. Advances in field techniques, instrumentation and interpretation continue to be made and hold promise for even more useful applications of gravity and magnetic techniques.

## Introduction

Although many differences exist between them, gravity and magnetic methods of prospecting are often discussed together because of similarities in data display and interpretation techniques. In this section we will consider the principles, instrumentation, data collection, data reduction and application separately, and then review interpretation methods together. Good general references include Grant and West (1965), Dobrin (1976), Rao and Murthy (1978), and Parasnis (1979). Excellent current reviews are given by Tanner and Gibb (1979) and Hood et al. (1979).

## The Gravity Method

### *Principles*

In gravity prospecting we often speak about the acceleration of gravity, which is the acceleration that a freely falling body would experience in the earth's gravitational field. This acceleration is given by $G\,M_e/r_e^2$, where $M_e$ and $r_e$ are the mass and radius of the earth, respectively. It is found by measurement that the earth's gravitational acceleration is about 983 gals ($cm/sec^2$) at the poles and about 978 gals at the equator. The *gal* and the *milligal* are common units, named after Galileo, used in gravity prospecting. Gravity is less at the equator than at the poles because the equatorial radius is greater than the polar radius and because of the variation, with latitude, of centrifugal force due to the earth's rotation.

Modern gravity meters routinely measure spatial variations in the earth's gravity field to 0.01 milligals (1 part in $10^8$) or better in field application, and the newest generation of instruments is capable of ± 0.002 milligals under ideal field conditions. These spatial variations in gravity are caused by lateral variations in rock density when measurements are restricted to the earth's surface. Near-surface density variations affect the gravimeter more than do deep variations, in accordance with the inverse square nature of Newton's law, and most gravity variations of interest in mining exploration result from changes in density within shallow crustal rocks. Because the gravimeter detects lateral variations in rock density, a *density contrast* must exist between the rock body under investigation and its cavity rock if an anomaly is to be found.

Rock density depends upon mineral composition, degree of induration, porosity, and compressibility. Shales display marked variations of density with depth because of their relatively high compressibility. As a general rule, older sedimentary rocks are higher in density than younger sedimentary rocks. Acid igneous rocks are less dense than basic igneous rocks. Most plutonic and metamorphic rocks display smaller ranges in density than do sedimentary and volcanic rocks. Volcanic rocks generally display rapid density variations due to porosity changes from place to place. Table 1 lists typical values of density for a variety of rock types. Note that density variations greater than 25 percent of the average crustal density, 2.67 $gm/cm^3$, are rare in near-surface rocks; in sharp contrast to electrical and magnetic properties of rocks, which can vary over several orders of magnitude.

*Surveying and data reduction*

Field surveys are performed by reading the gravimeter at selected station sites, either on a regular grid or in an irregular pattern as station access and optimum survey design dictate. Repeated readings are commonly made at one- to four-hour intervals at one or more previously established base stations in order to determine instrument drift and local gravity tidal variations.

TABLE 1

DENSITIES OF ROCKS AND MINERALS
(Modified from Dobrin, 1976, with additions)

| NAME | DENSITY, gm/$cm^3$ Range | Average |
|---|---|---|
| Alluvium and Soil | 1.6-2.2 | 1.90 |
| Sandstone | 1.6-2.6 | 2.32 |
| Limestone | 1.9-2.8 | 2.54 |
| Dolomite | 2.4-2.9 | 2.70 |
| Shale | 1.8-2.5 | 2.42 |
| Granite | 2.5-2.8 | 2.67 |
| Diorite | 2.6-3.0 | 2.84 |
| Gabbro | 2.8-3.1 | 2.98 |
| Diabase | 2.8-3.1 | 2.97 |
| Dunite | 3.2-3.3 | 3.28 |
| Quartzite | 2.6-2.7 | 2.65 |
| Gneiss | 2.6-3.1 | 2.75 |
| Schist | 2.6-3.0 | 2.82 |
| Slate | 2.6-2.8 | 2.81 |
| Amphibolite | 2.7-3.2 | 2.99 |
| Eclogite | 3.3-3.5 | 3.39 |
| Salt | 1.9-2.2 | 2.15 |
| Pyrite | 4.9-5.2 | 5.00 |
| Pyrrhotite | 4.5-4.7 | 4.60 |
| Sphalerite | 3.9-4.1 | 4.00 |
| Magnetite | 5.0-5.2 | 5.10 |
| Water | --- | 1.00 |

Information in addition to the gravimeter reading must be known at each site in order to reduce the raw field data. The instrument must be carefully calibrated. Corrections must be made for differences in elevation and latitude among the stations. The latitude correction removes the effects of the northward increase in the earth's field. There are two elevation effects that are usually combined into one correction (Figure 1). A reference elevation is selected to which all elevation corrections are made. For simplicity, in the following discussion the reference elevation is assumed to be the elevation of the survey base station, although any elevation could be selected. The *free air* correction accounts for the decrease in the gravity field with increasing distance from the earth's center, but this correction ignores the mass of material that lies between the ground surface and the reference elevation. The *Bouguer* correction accounts for this mass by assuming it to be an infinite slab of uniform thickness and specified density. Variations from this slab assumption are accounted for by a *topographic correction* which is commonly applied only in areas of rugged topography. Both the Bouguer and the topographic correction require an assumption for the density of near-surface rocks. This density is often assumed to be 2.67 $gm/cm^3$.

The anomalous gravity value in milligals, $G_{sta}$, at the field station relative to the base value, $G_{base}$, is given by

$$G_{sta} = G_{base} + g_{obs} - 0.8121d \, \mathrm{Sin}2\phi \text{ mgal/km (north)} + 0.3086h \text{ mgal/m} - 0.04186 \, \rho h \text{ mgal/m} + \text{terrain correction,}$$

where

$g_{obs}$ = observed gravity reading at the field station

$d$ = distance in km the field station lies north of the base station

$\phi$ = geographic latitude of the base station

h = elevation difference between field station and reference elevation.

Any convenient value for $G_{base}$ may be taken. If the gravity anomaly relative to the *International Ellipsoid* is known for the base, then that value is generally used because the field station then becomes tied to other similar stations elsewhere on earth.

From the above formulas we see that north-south station location must be known to about 10 m and elevation must be known to about 0.05 m in order to make the latitude and elevation corrections of the same order as the 0.01 milligal specification of many surveys. Variations in measured gravity due to latitude and elevation effects will usually be much larger than the anomaly sought.

*Applications*

In some cases, orebodies have been directly detected by gravity surveys. Copper ore associated with massive pyrite bodies was discovered by underground gravity surveying at Bisbee, Arizona (Rogers, 1952; Sumner and Schnepfe, 1966). Gravity data were acquired along mining levels and were then contoured for interpretation both on levels and on vertical sections. Figure 2 shows such a vertical section at Bisbee (Sumner and Schnepfe, 1966; Fig. 4). The cross-hatched areas are the interpreted positions of dense sulfide bodies required to explain the observed gravity anomalies. Note the existence of gravity highs above the interpreted bodies and gravity lows below. The authors state that, of the recommended drill holes, 80 percent encountered sufficient sulfides to account for the gravity results.

Orebodies are often studied after discovery but prior to mining by gravity surveying to determine orebody dimensions, ore grade, and tonnage. Hinze (1966) gives examples of gravity studies to determine location and grade of iron orebodies in Minnesota, Wisconsin and Ontario, and concludes that gravity techniques can be superior to magnetic techniques in certain cases. Tonnage calculations can be made for some orebodies by calculating the excess mass needed to account for the gravity anomaly (Hammer, 1945; Grant and West, 1965, p. 269).

Acidic intrusions, commonly associated with mineralization, sometimes have an associated gravity low. U. S. Geological Survey open-file data show this effect at the Questa district, New Mexico, where the low extends several miles east of known economic mineralization and presumably outlines prospective intrusive rocks at depth. Gravity lows are also observed in many intrusive complexes in the Basin and Range province. Stacy (1976) has documented a correlation between negative gravity anomalies of about 30 milligals and exposed quartz monzonite plutons in British Columbia, and has used this correlation to postulate locations for other plutons beneath volcanic cover. Ager et al. (1973) used results of a gravity survey to propose a model for the subsurface configuration of the Guichon Creek batholith in British Columbia. From the model, a relationship between the occurrence of known disseminated mineralization and batholith geometry was postulated and this relationship forms a valuable guide to further prospecting.

Plouff and Pakiser (1972) show a good example of the use of gravity data to model the geometry of a rather large intrusive complex in southwest Colorado. Figure 3 shows the salient features of the area and the gravity data. The large gravity low is postulated to be due to a concealed batholith that underlies a caldera complex in the San Juan Mountains.

Gravity surveys have been done in the Basin and Range Province and in many other areas of similar structure for the purpose of determining the thickness of basin fill. Gravity lows generally correlate with areas of thicker, low density alluvial material. Kane and Pakiser (1961) give a good example of this application in the Owens Valley of California. The method works well except in areas where intercalated volcanic rocks occur or where the alluvium is well consolidated. In both instances the density contrast between bedrock and basin fill becomes small and can approach zero, rendering the method ineffective. Gravity interpretation for alluvial-filled basins often yields minimum alluvial thickness.

In massive sulfide exploration, the gravity method has been used as detailed follow-up to ground EM surveys to help differentiate sulfide and graphite conductors. Higher priorities for drilling can be given to areas that show coincident positive gravity and EM anomalies, but care must be taken not to drop EM anomalies from consideration simply for lack of a gravity response. If an orebody is narrow or pipelike and is more than a few tens of feet below the surface, the gravity anomaly can be so small as to be lost in geologic noise. For example a vertical tabular orebody, 60 percent sulfide minerals, of 12 million tons that is 10 m wide and 30 m deep will give a maximum gravity anomaly of only 0.5 milligals. Nevertheless, gravity surveys have been found useful in massive sulfide exploration by Seigel et al. (1968) at Pine Point, by Brock (1973) at Faro, Yukon territories, by Schwenk (1976) at Flambeau, Wisconsin, and by many other investigators. A Pine Point example is shown in Figure 4.

## The Magnetic Method

### *Principles*

The earth's magnetic field is believed to originate in the core, although time-varying perturbations to this field originate outside the earth, principally in the ionosphere. Although many theories have been advanced to explain the earth's magnetism, the favored one is that fluid motions in the electrically conducting iron-nickel core cause a self-perpetuating dynamo that generates and sustains the field.

To a good approximation, the field at the earth's surface is dipolar and thus resembles the field that would occur if a powerful bar magnet were placed at the earth's center. The dipolar axis does not correspond with the earth's rotational axis but is displaced slightly in direction. Thus the north and south magnetic poles, where the field becomes vertical, do not correspond with the geographic poles.

The earth's field varies in intensity from about 25,000 gammas (1 gamma = 1 nanotelsa=$10^{-5}$ oersted) at the magnetic equator to about 70,000 gammas at the poles. In direction the field is horizontal at the equator and vertical at the poles. Over most of the United States the field dips 60 to 70 degrees northward.

Rock magnetism is a complex topic whose details are still being studied. Strangway (1967a and b; 1970) and Doell and Cox (1967) give good summaries of this and related topics. Rock magnetism has been treated in detail by Nagata (1961). For our purposes there are three main points to note. First, magnetic minerals and rocks have a component of magnetization, often the chief component, due to induction in the earth's field. This induced component is the response of magnetic minerals to the earth's field, is proportional in intensity to the earth's field strength, and is in a direction parallel to the earth's field. The constant of proportionality is termed the *magnetic susceptibility*. Second, another form of magnetization called *remanent* or *permanent* magnetization often exists and is superimposed on induced magnetization. Remanent magnetization can form as a result of cooling of an igneous rock from a molten state, as a result of metamorphism, as a result of chemical changes, or from other causes. The remanent component of magnetization can be either weaker or stronger than the induced component, and it is often not in the same direction as the induced component. Rocks having small mineral grains, commonly have a larger remanent component than those having larger mineral grains because the stability of remanent magnetization is related to grain size. Third, above a temperature known as the Curie temperature, magnetization changes and, for exploration purposes, rocks cease to be magnetic. The Curie temperature of pure magnetite is 580°C, but impurities can alter this value. This temperature is attained in the earth's crust at a nominal depth of 25 km, although the Curie point isotherm is believed to be much shallower in areas of high heat flow.

Only a few minerals are sufficiently magnetic to cause measurable changes in the earth's field. These are listed together with their magnetic susceptibility and ranges for the susceptibility of common rocks in Table 2. Magnetite is usually the magnetic mineral under consideration in exploration. It is both highly magnetic and widely distributed, principally as an accessory mineral. Empirical relations have been established between magnetite content and magnetic susceptibility of rocks (for example, see Mooney and Bleifuss, 1953). One commonly used rule of thumb is that 1 volume percent magnetite results in a magnetic susceptibility of about $3000\times10^{-6}$ cgs, but this can be highly variable.

*Instrumentation*

Magnetometers in field use commonly measure variations in the intensity of the earth's field to about one gamma, although instruments that detect changes as small as $10^{-5}$ gammas are available. Hood et al. (1979) give a very valuable summary of instruments available today, including manufacturers and specifications. Only a few words will be written here about the most important instrument types.

The flux-gate magnetometer uses an element whose magnetic saturation value is only slightly larger than the earth's field. Variations in the earth's field are detected by measuring the variation in the additional field that must be applied to the element to cause saturation. This instrument is used to measure the total magnetic field in airborne installations and the vertical field component in ground equipment. Most survey installations are capable of about one gamma resolution.

The proton-precession magnetometer measures the precession frequency of protons in the earth's field. This frequency is proportional to the field strength. The sensor is a wire coil wrapped around a bottle containing a hydrogen-rich source such as kerosene. In both airborne and ground instruments the total field is sensed. Most instruments are capable of about 1 gamma resolution.

Higher sensitivity can be attained by use of optically pumped magnetometers. These instruments measure the difference in energy levels for electron orbits developed in a suitable alkali metal vapor (cesium or rubidium) by the earth's field. These magnetometers have a sensitivity of about 0.005 gammas, which is sufficient to allow two sensors separated by a suitable distance to measure magnetic gradient. Total field and vertical gradient airborne surveys are available to facilitate difficult interpretation

TABLE 2

MAGNETIC SUSCEPTIBILITY FOR COMMON MINERALS AND ROCKS

| ROCK OR MINERAL | MAGNETIC SUSCEPTIBILITY X$10^6$(cgs) | |
|---|---|---|
| | Approx. Range | Typical Value |
| Sedimentary Rocks | 0-2,000 | 200 |
| Acidic Igneous Rocks | 600-6,000 | 2,500 |
| Basic Igneous Rocks | 1,000-20,000 | 5,000 |
| Magnetite | 300,000-800,000 | 500,000 |
| Pyrrhotite | --- | 125,000 |
| Ilmenite | --- | 135,000 |
| Franklinite | --- | 36,000 |

problems. Both horizontal and vertical gradient equipment is available for ground use.

Cryogenic magnetometers, using low-temperature physics, are of recent development. The Josephson junction effect is exploited by a device called a Squid, which stands for *superconducting quantum interference device*, and which is maintained at 4.2°K, the temperature of liquid helium. Squid magnetometers have a sensitivity of about $10^{-5}$ gammas facilitating measurement of magnetic gradients in three orthogonal directions, in a small package suitable for ground surveys or aircraft installation. This installation would, of course, also record total field, and such a comprehensive set of data would facilitate better interpretation.

Instruments are also available for measuring the magnetic susceptibility and remanent magnetization component in rocks either *in situ* or in the laboratory. Such rock property measurements are of great value to the interpreter because they help him understand the variations of these important properties over the survey area, and because they facilitate correlation of interpreted results with actual rock types.

*Surveying and data reduction*

Field surveys are performed on the ground, from the air, and by towed sensors under water. On the ground, stations can be occupied either on a regular grid or along available access. Repeated readings are usually made at a base station or, alternatively, a recording base station is operated to facilitate removal of normal diurnal variations and to determine whether or not a magnetic storm is in progress. Latitude corrections are not usually necessary except for extensive surveys because most anomalies of interest will be little affected.

Aeromagnetic surveys provide most of the magnetic data collected for mineral exploration. They have a number of advantages over ground surveys, including generally better coverage, speed, and cost-effectiveness. Modern aeromagnetic systems often incorporate in-flight digital magnetic recording of data, recording barometric and radar altimeters, and Doppler radar navigation. Fixed-wing aircraft can generally drape-survey moderately rugged terrain at 150 to 300 m terrain clearance, along lines as dense as 4 per km and are used for higher altitude, constant elevation surveys as well. Helicopters facilitate closer terrain clearance in rugged terrain and closer line spacing. Good data reduction and plotting is a non-trivial task that requires care and experience. Hood et al. (1979) give a current summary of techniques and pitfalls.

*Applications*

The magnetic method has found very broad application in exploration. Because this method usually maps the distribution of magnetite, it can be used in any application where knowing that distribution might help.

One of the most useful applications of magnetic data is to facilitate geologic mapping. Outcrop geology often can be extended under soil, vegetative, or alluvial cover by observing correlations between magnetic response and geology. Structural and magnetic data trends are commonly parallel. Figure 5 shows an example of aeromagnetic data and an interpretation in terms of sub-till geology in Wisconsin.

In exploration for disseminated copper or molybdenum, the magnetic method is useful in locating and mapping hidden intrusive complexes that can then be surveyed with induced polarization or prospected by other methods to locate sulfide mineralization. Basic portions of these intrusive complexes are commonly more magnetic than acidic bodies. Because acidic rocks commonly have a lower density than basic rocks, gravity and magnetic studies together can help differentiate these.

Magnetic surveying can be very helpful in locating magnetic skarn deposits that are often associated with disseminated and other mineralization in carbonate rocks and are often orebodies themselves. These features are shown by aeromagnetic data from the Ely porphyry copper deposit in eastern Nevada (Figure 6). This deposit is underlain by a large quartz-monzonite intrusion whose upper surface dips steeply northward, but dips at a gentle angle to the south. The country rocks are sedimentary rocks of Paleozoic and Mesozoic age. Mineralization is both disseminated in igneous and sedimentary rocks and magnetite-copper skarn bodies in carbonate rocks. The magnetic anomaly consists of a large, high-amplitude positive anomaly caused mainly by the intrusive rocks with contributions from each skarn deposit. The broad magnetic low immediately north of the positive anomaly is simply the normal effect caused by induction in the earth's field and is an integral part of the whole anomaly. Destruction of magnetite by the process of sulfide mineralization can and does cause magnetic lows over some mineralized areas elsewhere, however.

A rather obvious application of the magnetic method is in prospecting for iron ore directly. Successful surveys have been performed in the Mesabi Iron Range and in Nevada, U.S.A. (Riddell, 1966), in Australia (Webb, 1966), and in many other places (Gay, 1966). Hematite ores often contain enough magnetite to be highly magnetic, and taconite ores are often accompanied by magnetite ores. Once an iron orebody has been discovered, magnetic methods can be applied to determine details.

In massive sulfide exploration the magnetic method can be very useful as a follow-up to the EM method in locating copper-nickel deposits, which often contain magnetic pyrrhotite. Many copper-zinc and other massive sulfide deposits are non-magnetic, however, and it is therefore unwise to use magnetic data to eliminate orebody occurrence. Massive sulfide deposits characteristically occur in greenstone belts, typically in Precambrian rocks. Greenstones are usually more magnetic and more magnetically variable from place to place than are the granites that commonly surround them. Thus aeromagnetic reconnaissance can be used to define greenstone belts that are then prospected by airborne and/or ground EM (Figure 5).

## Gravity and Magnetic Interpretation

Although gravity and magnetic interpretation techniques are generally better developed than are electrical interpretation techniques, much remains to be done. New instrumentation, particularly for precise measurements, will continue to inspire corresponding advances in interpretation methods. The interpreter is obliged to know how to choose and to apply the best techniques.

Complete interpretation requires both geophysical and geological considerations. It is the primary goal of the geophysicist to turn the gravity or magnetic map into one or more geologically reasonable subsurface illustrations showing the depths, lateral boundaries, locations, and density or magnetic susceptibility contrasts of the various bodies detected. The geologist then takes this information on physical property distribution and makes the most reasonable geologic interpretation in terms of rock type distribution. These tasks are far from trivial.

No interpretation of gravity or magnetic data alone is unique. It can be shown that an infinite number of different mass or magnetization distributions can be contrived to explain any given anomaly. Figure 7 illustrates this in one particular case for a gravity profile. Each of the alternative basement reliefs explains the observed anomaly equally well. Ambiguity of interpretation can generally be reduced through use of geological or other geophysical data.

Interpretation usually begins with an attempt to isolate individual anomalies from background or regional values. Definition of the regional effects is subjective. Techniques vary from visual hand smoothing of contours or profiles, to manual averaging of values at specific grid points, to complex computer-assisted filtering. Once a regional field is determined, it is subtracted from the total field and the residual represents effects due to anomalous bodies.

Interpretation of magnetic data is considerably more complicated than is interpretation of gravity data although both represent applications of potential field theory. One complicating factor in magnetic interpretation is that the inclination of the earth's magnetic field varies from horizontal at the magnetic equator to vertical at the magnetic poles. Therefore, the direction of induced magnetization in rock bodies varies in the same way. By contrast, the gravity field is always vertical. The result is that the gravity anomaly due to a certain body is the same no matter what its latitude or longitude on the earth, but a given magnetic body has an anomaly that is much different at the poles than at the equator.

Yet another complicating factor in magnetic interpretation is the possibility of remanent magnetization. The remanent component can align in any direction and can be stronger or weaker than the induced component. Reliable location of magnetic bodies and determination of susceptibility are difficult in the presence of remanent magnetization. Experience and study of computer models aids the interpreter in separating the effects of varying body shape, depth, and physical property contrasts for gravity interpretation, and, in addition, body dip and strike, and relative magnetic field inclination for magnetic interpretation. These effects can be complicated, as Fig. 8 illustrates.

Interpretation methods can be divided into four classes: a) rule of thumb, b) characteristic curve-matching, c) forward modeling, and d) inverse modeling. Progress in development of techniques in each class has led to better interpretation, especially since the advent of the digital computer. Rules of thumb can be used to get a preliminary overview of location and depth of anomalous bodies before more sophisticated techniques are applied. Peters (1949), Smellie (1967) and Dobrin (1976) give useful summaries of a few of these techniques. Many curve-matching techniques are available, generally for interpretation in terms of specific bodies or models (Grant and West, 1965). These techniques are pursued if no computer modeling capability is available or if only a few profiles or anomalies are to be interpreted.

In more complex situations forward modeling is beneficial. In forward modeling a preliminary estimate (i.e., a model of the subsurface configuration of anomalous masses or magnetic bodies) is formed, perhaps by application of rules of thumb. Then the anomalies to be expected are calculated from the model. The calculated results are compared with the observed anomalies, and the model is modified to start the cycle again. This iterative process is continued until a satisfactory match between computed and observed results is obtained. Any geologic control available can be used to constrain the model so that the results, while not unambiguous, are geologically sound. Computer graphics and user-interactive programs facilitate this approach greatly. At the present time comprehensive 2-D and 3-D computer programs are available from several sources, including Snow (1978) and Nutter (1980).

In the inverse approach, mathematical techniques are used to calculate a model directly from the data. Inversion does not yield a unique model either, however. The promise that inversion offers is for rapid and inexpensive interpretation of large amounts of data by letting the computer do most of the work. The challenge is to assure appropriate model constraints and to allow input of geologic knowledge so that the final result is geologically sound. Techniques for 2-D inversion have been developed and successfully applied by Hartman et al. (1971) and by O'Brien (1972). Such modeling is currently at the forefront of development. These techniques are more reliably applied to rather simple geologic situations such as basement studies for petroleum exploration. Interpretation in the much more complex mining environment still relies heavily on experience in spite of increases in the level of sophistication of interpretational aids.

Gravity and magnetic data can be *continued* both upward and downward to determine the map or profile as it would be observed at a higher or lower level. Upward continuation is straightforward and reliable, but care must be taken with downward continuation because small errors in the data are amplified. Potential field data can be continued downward only to the top of the uppermost anomaly-producing body. Continuation operations can be of assistance in matching aeromagnetic surveys at different elevations (Bhattacharyya, et al., 1979).

Sometimes magnetic data are *reduced to the pole* i.e., a computer technique is applied to transform the data to appear as they would if the survey had been performed at the magnetic pole where the inducing field direction is vertical (Baranov, 1957).

Advances both in measurement techniques and in interpretation hold promise for continued and even more useful applications for gravity and magnetic data. These methods have contributed much to exploration geophysics, and will do so in the future.

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