Rock magnetism, the distribution of magnetic minerals in the Earth's crust, and aeromagnetic anomalies

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By Richard L. Reynolds, Joseph G. Rosenbaum, Mark R. Hudson, and Neil S. Fishman

Abstract

A review of current understanding of relations among rock magnetic properties and petrology of igneous, metamorphic, and sedimentary rocks, as well as of the distribution of magnetic minerals in the Earth's crust, shows that rock magnetic and petrologic studies can provide important constraints on the interpretation of aeromagnetic anomalies. The development of three-dimensional geologic maps based on aeromagnetic data requires an improved understanding of: (1) petrologic and petrochemical controls on magnetic properties of rocks in the lower oceanic and continental crust and of plutons in the middle and upper continental crust; (2) effects of lower and middle crustal temperatures and pressures on total magnetizations; (3) relations among metamorphic facies, premetamorphic lithostratigraphy, and magnetization; and (4) relations among types and origins of ore deposits, host-rock composition and mineralogy, and magnetic signatures that may be diagnostic for specific mineral and energy habitats. Magnetic contrasts in sedimentary rocks are controlled by depositional factors and by geochemical and consequent mineralogic alterations that either enhance or suppress magnetizations. Examples are drawn from the formation of uranium deposits in sandstones and from the effects of hydrocarbon seepage.

INTRODUCTION

The purpose of this report is to present a broad overview of the magnetic properties, petrologic controls, and magnetic mineral composition of crustal sources of magnetic anomalies. This overview is preceded by a brief review of the factors that control magnetic properties of minerals and rocks. We also attempt to identify avenues of rock magnetic and petrologic investigations that will provide important constraints on the interpretation of aeromagnetic data and thus contribute to the development of three-dimensional geologic models. The discussion is intended to serve as a general guide for interpreters of aeromagnetic data to current knowledge and some recognized problems of crustal magnetization.

Special attention is given to the causes of magnetic contrasts in sedimentary rocks. Recently improved techniques for the acquisition and analysis of aeromagnetic data, as well as newly developed models and guides for the exploration of energy and mineral resources, have expanded interest in aeromagnetic anomalies that arise from sedimentary rocks. The potential of aeromagnetic methods for detecting the magnetic-diagenetic effects of vertically directed alteration plumes, such as may be caused by hydrocarbon seepage (Donovan and others, 1979), is an outstanding example. The understanding of such features is currently very limited, but where such an understanding exists, it is based on detailed rock magnetic, mineralogical, and geochemical study.

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MAGNETIC PROPERTIES OF ANOMALY SOURCES

The physical property that relates magnetic anomaly to its source is total magnetization. The magnitude and direction of total magnetization (J_t) are given by the vector sum of two components: remanent magnetization (J_r) and induced magnetization (J_i). J_t can be computed as:

$$J_t = J_r + J_i = J_r + kB,$$

where k is magnetic susceptibility (SI, dimensionless), and vector B is the magnetic induction of the Earth at the location of the body. The measurements of J_r and k of samples at ambient temperature and pressure are easily obtained using laboratory instruments. Another widely used quantity is the Koenigsberger ratio of remanent magnetization magnitude to induced magnetization magnitude.

The magnetic properties of rocks and minerals have been discussed in detail from an aeromagnetics perspective by Haggerty (1979), McIntyre (1980), Grant (1984/1985a, b), and Clark (1983). Clark (1983) and Carmichael (1982) present summaries of representative values of susceptibility, remanent magnetization, and Koeingsberger ratio corresponding to many magnetic minerals and some rock types.

The magnitude of J_i depends on the quantity, composition, and size of the magnetic-mineral grains, as well as on the magnitude of the magnetic induction (the Earth's field direction). The direction of J_i is usually parallel to the local magnetic induction, but it may differ if anisotropy of magnetic susceptibility is great. In general, k, and thus J_i, is relatively low for small magnetic grains and higher for larger grains.

The magnitude of J_r is also a function of quantity, composition, and sizes of the magnetic grains. In addition, naturally occurring J_r (natural remanent magnetization, NRM) is strongly influenced by the physico-chemical environment (which determines the type of remanence that is acquired) and the strength of the induction field at the time of acquisition. Other types of remanence include thermoremanent magnetization-(TRM), chemical remanent magnetization (CRM), isothermal remanent magnetization (IRM), and depositional (or detrital) remanent magnetization (DRM). The in-situ direction of J_r is usually parallel to the ambient magnetic induction (the Earth's field direction) during the acquisition of remanence, but it may be rotated due to later tectonism. In contrast to J_i, small magnetic grains produce relatively strong and stable J_r.

The types of remanent magnetizations that are acquired by different rock types are listed in table 1. Viscous remanent magnetization (VRM), in the direction of the present Earth's field, dominates the remanence of rocks in the middle and lower crust as well as that of coarse-grained plutonic rocks. The direction of NRM for other types of rocks must, in general, be determined experimentally. In some cases NRM magnitudes can be estimated for known or expected rock types (Clark, 1953). However, NRM directions are usually difficult to estimate because the NRM may be the result of both primary and secondary components, which may have been affected by tectonic rotations. An excellent review of paleomagnetic methods, including problems that involve multiple components of magnetizations and tectonic rotations is given by Hillhouse (in press). Not only is knowledge of the direction and magnitude of remanent magnetization commonly important in constraining the total magnetization vector, but knowledge of the type of remanence may be crucial to the correct geologic interpretation of magnetic anomalies. For example, the implications of anomalies over sedimentary rocks are very different if the magnetization is dominated by detrital remanent magnetization rather than by a chemical remanent magnetization related to the presence of hydrocarbons.

MAGNETIC MINERALS AS ANOMALY SOURCES

The most important magnetic minerals for aeromagnetic measurements are those of the magnetite-ulvöspinel (Mt-Usp_{ss}) solid-solution series, the titanomagnetites [$xFe_2TiO_4*(1-x)Fe_3O_4$ $(0 \leq x \leq 1)$, fig 13]. A number of magnetic properties of titanomagnetites vary as a function of composition. For example, Curie temperatures decrease systematically from about 580 °C for ferrimagnetic magnetite to –153 °C for antiferromagnetic ulvöspinel (fig 13). A Curie temperature of 25 °C corresponds to titanomagnetite that has about 75 percent ulvöspinel in solid solution. Saturation magnetization, which is the maximum possible magnetization (92 A m^2/kg for Fe_3O_4), similarly decreases with increasing content of titanium. Although the Earth's magnetic field is too weak to saturate magnetically the common magnetic minerals, saturation magnetization is a useful indicator of the capacities of different magnetic minerals to become magnetized. The susceptibility of Mt-Usp_{ss} is effectively zero for Usp_{ss} contents greater than about 70 percent. For Mt-Usp_{ss} that is more enriched in Mt, the variation of susceptibility with composition depends on grain size. Discussions of relations among grain sizes, domain structures, and magnetic properties are given by Dunlop (1981) and Clark (1983).

Minerals of the ilmenite-hematite (Ilm-Ht_{ss}) solid-solution series [$xFeTiO_3*(1-x)Fe_2O_3$ $(0 \leq x \leq 1)$, and referred to here as titanohematites], vary complexly in magnetic structure and properties as a function of composition. In the compositional range $0.5 \leq x \leq 0.8$, between weakly magnetic hematite (antiferromagnetic) and ilmenite (paramagnetic), the titanohematites are ferrimagnetic, have Curie temperatures in the range of 235 to –26 °C, and attain a saturation magnetization of as much as a third that of magnetite.

Table 1. Types of remanent magnetizations that may contribute to aeromagnetic anomalies

[T, temperature; T_c, Curie temperature. Based partly on table 4.1 in Tarling (1983)]

Remanent magnetization	Abbreviation	Rock types	Definition or characteristic
Natural	NRM	All	Summation of all components of remanence.
Thermal	TRM	Igneous, metamorphic	Acquired during cooling from the maximum T_c to room T (Total TRM).
Partial thermal	pTRM	Igneous, metamorphic.	The TRM acquired during cooling in a T interval having maximum $T<T_c$. The total TRM equals the sum of the pTRM's.
Partial thermochemical.	pTCRM	Igneous, metamorphic.	Remanence acquired by products of exsolution-oxidation during cooling.
Viscous	VRM	All	Secondary remanence related to thermal agitation of domain walls that causes decay of primary remanence and changes domain assemblage to lowest energy state.
Viscous partial thermal.	VpTRM	All	Distinguished from VRM on basis of higher temperature conditions below T_c .
Detrital, depositional.	DRM	Sedimentary	Primary remanence acquired by the physical rotation of detrital grains during deposition.
Postdepositional	PDRM	Sedimentary	Acquired during postdepositional rotation of interstitial grains. Some investigators include effects of early diagenetic chemical-magnetic alterations.
Chemical	CRM	All	Secondary remanence acquired during growth of magnetic minerals in presence of magnetic field. Includes growth by nucleation or replacement.

Titanomaghemites (maghemite) form by low-temperature oxidation of titanomagnetites (magnetite). Maghemites are nonstoichiometric spinels with variable amounts of cation site vacancies caused by either addition of oxygen or removal of metals (shaded pattern in fig. 13). The Curie temperature increases with progressive degree of oxidation, and the saturation magnetization decreases with increased oxidation of titanomagnetites that originally have less than 50 mol percent ulvöspinel. The saturation magnetization of the titanomaghemites is 85 A m²/kg for maghemite (γFe_2O_3). The oxidation of magnetite also commonly produces hematite (saturation magnetization is 0.5 A m²/kg). The factors that dictate the formation of either hematite or maghemite are not well understood but may be related to presence or absence of water; the presence of water favors maghemite.

Other magnetic minerals, which are not illustrated on figure 13, that are or may be responsible for magnetic anomalies include metallic iron, metal alloys of Fe-Ni-Co-Cu, and the magnetic sulfides, monoclinic pyrrhotite and greigite. In addition, as described later, nonmagnetic iron sulfide minerals, particularly pyrite, can play a significant role in changing magnetization in certain sedimentary rocks that have been altered at low and high temperatures.

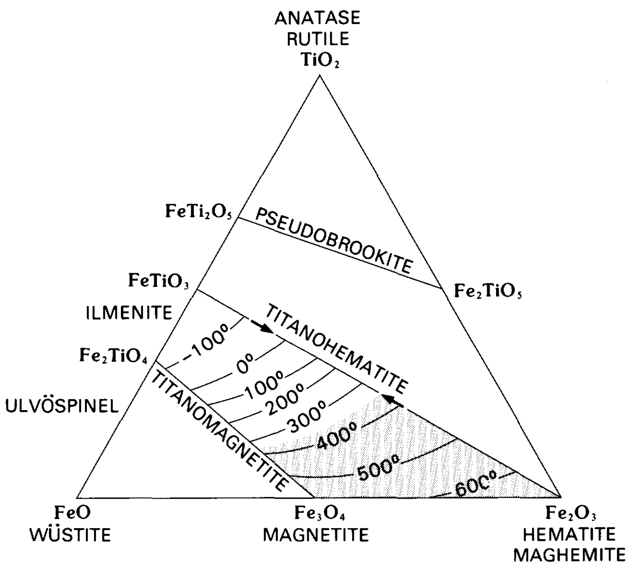

Figure 13. The FeO-Fe_2O_3-TiO_2 ternary system showing the titanomagnetite, titanohematite, and pseudobrookite solid-solution joins, as well as the Curie temperature contours (in degrees Celsius) between the titanomagnetite and titanohematite joins. In the compositional range (50–80 mol percent ilmenite) indicated between the arrows, the titanohematites are ferrimagnetic. The shaded area represents the field of titanomaghemite.

PETROLOGIC CONTROLS ON MAGNETIC PROPERTIES

The magnitudes of remanent and induced magnetization in rocks depend primarily on the abundance, composition, and grain size of magnetic minerals in the rocks, as well as on the nature of the remanence. In igneous rocks, these factors are controlled mainly by bulk chemistry, by initial temperatures of formation, and by cooling conditions, including especially cooling rate and oxygen fugacity (Haggerty, 1979). In metamorphic rocks, such factors are controlled similarly by rock and mineral chemistry and by the thermal history (McIntyre, 1980; Grant, 1984/1985a). In clastic sedimentary rocks the factors are controlled primarily by source area, depositional process, depositional environment, and diagenetic alteration. Although most chemical and biogenic sedimentary rocks have negligible magnetizations, chemical precipitation of abundant magnetite from waters, has occurred under appropriate conditions of pH, amount and valences of iron and sulfur, oxygen fugacity, and temperature.

The following sections focus on factors that determine the types and characteristics of magnetic minerals, as well as the growth or depletion of these minerals in rocks, and that thus control rock magnetic properties. Although the emphasis is not on links between rock magnetization and specific features [such as polarity (positive or negative), magnitude, and shape] of aeromagnetic anomalies, some general statements on these links can be made. In the simplest cases, for example, positive magnetic anomalies can result from strong induced magnetizations, from a dominant VRM, and (or) from stable remanence of normal polarity. Negative magnetic anomalies can be caused by rocks with dominant stable remanence of reversed polarity. However, the shape of the magnetic anomaly depends not only on the direction of total magnetization but also on the shape of the magnetic body and on the spatial relations among different bodies that differ in their magnetizations (either magnitude or direction, or both). An excellent overview of aeromagnetic methods (including data acquisition and processing, and techniques for interpretation) is given by Blakely and Connard (in press); reports by Hinze (1985) and Paterson and Reeves (1985) also cover this subject.

Igneous Rocks

In general, concentrations of Mt-Usp$_{ss}$ grains tend to be higher in basic rocks (about 5–10 percent by volume) than in intermediate and acidic rocks (about 1–3 percent). The difference in magnetization between basic and acidic rocks is not as great as one might expect from differences in the Fe-Ti oxide contents. In basic rocks the titanomagnetites usually have high titanium content, weak magnetizations, and low Curie temperatures. Conversely, in intermediate and acidic rocks the titanomagnetites have low titanium content, strong magnetizations, and high Curie temperatures. In the latter, increasing SiO_2 content is associated with lower titanium content and higher oxidation states.

The aeromagnetic signatures of igneous rocks are strongly influenced by their cooling history, primarily because of the effects of cooling and oxygen fugacity on the solubility of the magnetic oxide solid solutions (Haggerty, 1979). The effects of mineral exsolution and of high- and low-temperature oxidation on mineral type, and consequently on magnetization, Curie temperature, and magnetic stability of the magnetite-ulvöspinel solid-solution series minerals are summarized in figure 14. Both exsolution and high-temperature oxidation favor the isolation of low-temperature titanomagnetite and therefore lead to increases in magnetization, Curie temperature, and magnetic stability. Advanced high-temperature oxidation of magnetite + ilmenite, however, creates weakly magnetic hematite (Curie temperature of 680 °C) and nonmagnetic titanium-rich phases. Low-temperature oxidation of magnetite-ulvospinel to titanomaghemite causes a decrease in magnetization, an increase in Curie temperature, and little change (perhaps a small increase) in magnetic stability. As with high-temperature oxidation, low-temperature oxidation of low-titanium magnetite to hematite greatly diminishes magnetizations.

Exsolution and oxidation also strongly influence rock magnetic properties by decreasing magnetic grain sizes. For example, the separation of original grains of Mt-Usp$_{ss}$ by exsolution or by high-temperature oxidation into segmented particles of magnetite + ulvöspinel or of magnetite + ilmenite, respectively (fig. 14), creates

REACTIONS	SATURATION MAGNETIZATION	CURIE TEMPERATURE	MAGNETIC STABILITY
MINERAL EXSOLUTION Mt–Usp$_{SS}$→Usp+Mt	Increase	Increase	Increase
HIGH-TEMPERATURE OXIDATION Mt–Usp$_{SS}$→Mt+Ilm	Increase	Increase	Increase
ADVANCED OXIDATION Mt+Ilm→Ht+Pb+R	Decrease	Increase	Increase
LOW-TEMPERATURE OXIDATION Mt–Usp$_{SS}$→(Ti) Maghemite Mt→Ht	Decrease	Increase	Increase

Figure 14. Reactions that involve exsolution and oxidation and their effects on saturation magnetization, Curie temperature, and magnetic stability. Mt-Usp$_{ss}$, magnetite-ulvöspinel solid solution; Ilm, ilmenite; Ht, hematite; Pb, pseudobrookite; R, rutile.

smaller intergrowths of ferrimagnetic magnetite with antiferromagnetic or paramagnetic phases. In general, fine-grained titanomagnetites and magnetite have higher remanent and lower induced magnetizations than coarse-grained particles of identical composition. High Koenigsberger ratios ($Q > 1$, remanent magnetizations dominant) are thus characteristic of volcanic rocks, which typically have undergone substantial high-temperature oxidation and contain small magnetic particles, whereas low ratios ($Q < 1$, induced magnetizations dominant) are characteristic of many coarse-grained intrusive rocks.

Metamorphic Rocks

Magnetic minerals may be produced or destroyed during metamorphism. In metasedimentary rocks the production of magnetite depends mainly on two factors that are largely inherited from the original sediment: (1) the total iron content, and (2) the oxidation state of the iron (the relative amounts of Fe^{3+} and Fe^{2+}). These factors control the amounts and types of iron oxide minerals that may form by limiting potential production of iron oxide and by controlling the partitioning of iron between oxides and silicates (McIntyre, 1980; Grant, 1984/1985a). Regional metamorphism of sedimentary rocks under conditions of lower greenschist to granulite facies drive many reactions that are capable of producing secondary magnetite (Grant, 1984/1985a). In general, formation of magnetite is favored in iron-bearing rocks under low-grade conditions or under high-grade conditions that promote the dehydration and breakdown of hydrous minerals such as biotite and amphiboles. Extreme metamorphism, which involves differential melting, strongly reduces magnetizations, possibly because of recombination of iron and titanium oxides to $Mt\text{-}Usp_{ss}$, if titanium is available (McIntyre, 1980; Grant, 1984/1985a).

Other intrinsic properties that influence the production of magnetite in metamorphic rocks include the contents of silica, carbon, and aluminum. McIntyre (1980) made the following generalizations: (1) Magnetite formation is favored in rocks that are undersaturated in silica; in the iron-oxygen system, silica saturation expands the stability field of fayalite (iron silicate) at the expense of magnetite. (2) The presence of carbon may restrict the formation of magnetite. Gas phases CO and CO_2 produced from carbon can lower oxygen fugacities below the magnetite stability field at high metamorphic temperatures (more than 630 °C). (3) In micaceous rocks, excess aluminum favors muscovite over biotite, and as a result the silicate minerals do not compete effectively with the oxides for available iron. Therefore, high aluminium content favors the production of magnetite.

Strong magnetic contrasts may develop in rocks of similar composition that have undergone different metamorphic reactions due to different metamorphic conditions. Moreover, metaigneous rocks of similar bulk composition may have different magnetizations that reflect differences in the original magmatic conditions of oxygen fugacity and temperature.

Magnetizations of igneous rocks are usually diminished by metamorphism and metasomatism due to destruction of preexisting magnetic minerals (Haggerty, 1979; McIntosh, 1983). Magnetite, however, can be produced by alteration of olivine and orthopyroxene during serpentinization of mafic and ultramafic rocks. Evidence for both depletion of primary magnetite and creation of secondary magnetite by hydrothermal alteration in a plutonic terrane (Criss and Champion, 1984) will be further discussed.

Mineralization in Igneous and Metamorphic Rocks

In certain ore-forming environments magnetic contrasts are useful as pathfinders to mineral deposits (Grant, 1984/1985b; McIntyre, 1980; Clark, 1983; Wright, 1981). Destruction of primary magnetic minerals or addition of secondary magnetic minerals, either by conversion of preexisting nonmagnetic phases or by growth from ore-related fluids, can develop such magnetic contrasts. Ferrimagnetic magnetite and pyrrhotite, as well as antiferromagnetic hematite if present in sufficiently large quantity, may be sources of magnetic anomalies in mineralized rocks.

Sedimentary Rocks

The amount, type, and grain sizes of detrital magnetic minerals in epiclastic rocks depend on the nature of and proximity to the source areas, as well as on the depositional environment. Postdepositional diagenetic and authigenic alterations strongly influence the magnetic character of sedimentary rocks and can either enhance or diminish the original magnetization. In general, the degree of oxidation of detrital titanomagnetite, if initially present, to hematite and other ferric oxide minerals increases with age (Van Houten, 1968), but such a relation has not been quantified in terms of magnetization. Rocks that contain detrital and (or) secondary hematite from usual depositional and (or) diagenetic processes are abundant in the sedimentary column, but they have low total magnetizations and are of little interest in aeromagnetic studies. However, important changes in magnetization of sedimentary rocks may be produced by reducing conditions that are related to mineralization or to hydrocarbons. Under sulfidic reducing conditions the original magnetization may be either suppressed by the replacement of the detrital iron oxides by iron sulfides or enhanced by the production of magnetic sulfide minerals, such as pyrrhotite or greigite. The formation of diagenetic magnetite related to hydrocarbons has been proposed as another process by which the magnetization of sedimentary rocks may be increased under reducing conditions (Donovan and others, 1979; McCabe, 1986). Geochemically reducing fluids, especially those with organic acids, are also known to dissolve iron-titanium oxide minerals and thus to diminish magnetizations.

Sedimentary rocks of chemical or biogenic origins, such as evaporites and carbonates, usually have low to negligible magnetizations. In certain settings, such as salt domes, these rocks may be juxtaposed with more magnetic rocks and thereby indirectly associated with subtle but important negative magnetic anomalies. In some strata, magnetic minerals, principally magnetite, have precipitated chemically from sea water and may comprise or be closely associated with mineral deposits. For example, magnetite that formed in this manner produces high magnetizations in Lower Proterozoic banded iron-formations and in distal sediments related to volcanogenic base-metal massive sulfide deposits (McIntyre, 1980; Large, 1977).

CRUSTAL DISTRIBUTION OF SOME MAGNETIC MINERALS

A simplified summary of the distribution of magnetic minerals that may be important for magnetic surveys is presented in figure 15. This figure serves as an outline for the remainder of the paper. The shapes of the symbols denote either a primary (circle) or secondary (square) magnetic mineral. Depletion (destruction) of a mineral is indicated by a circle-slash (⊘) symbol. The sizes of the symbols have not been quantitatively derived and are only crude indications of relative abundance of the different minerals. The symbol sizes have meaning only within a single column (crustal setting); a comparison of sizes between columns has no significance. The summary is not comprehensive and is intended simply as a general guide.

Oceanic Crust

The dominant magnetic mineral in the uppermost layer of newly formed oceanic crust, which is mostly composed of pillow lavas erupted from spreading centers, is titanomagnetite that has a nearly uniform content of ulvöspinel (Usp about 60 percent, which corresponds to a Curie temperature of about 175 °C) (Johnson, 1979). Hydrothermal alteration, however, readily converts the primary titanomagnetite to titanomaghemite, so that a CRM largely replaces the original TRM. The magnitude and direction of the CRM depend on those of the original TRM, the time and degree of alteration, and the polarity history during alteration. A model of such CRM acquisition (Raymond and LaBrecque 1987) helps explain a number of magnetic features of the oceanic crust, among them the decrease in amplitude of anomalies away from spreading ridges, the enhanced magnetization of the Cretaceous Quiet Zones, and amplitude and skewness discrepancies in intermediate- and short-wavelength magnetic anomalies. The alteration also results in increases in Curie temperature and in remanent stability. Nevertheless, the pattern of positive and negative anomalies, which reflect the normal/reversed polarity rock sequence, will be preserved if polarity intervals are randomly distributed during CRM acquisition (Raymond and LaBrecque, 1987).

Magnetite is rarely described as being present in samples of oceanic crust, and thus has been considered uncommon in these rocks. The paucity of reported magnetite occurrences, however, may stem partly from the lack of extensive deep sampling below the pillow-basalt layer. Magnetite (or titanomagnetite that has higher Curie temperatures and magnetizations than that produced near or on the sea floor) may form in the oceanic crust in at least two different ways. These magnetic minerals may form from Mt-Usp$_{ss}$ by oxidation/exsolution during initial cooling of intrusions. This mechanism is represented by the circle for magnetite in figure 15. Magnetite may also be produced by later high-temperature alteration (for example, by hydrothermal reheating) of titanomagnetite under oxidizing conditions or by thermal breakdown of titanomaghemite, regardless of redox conditions. Kristjansson and Watkins (1977), in fact, suggested that the Curie temperature of magnetite (580 °C) determines the base of the magnetic oceanic crust. Other magnetic minerals, which may be responsible for deep-seated magnetic anomalies over oceanic crust but for which direct evidence is lacking, are alloys of Fe-Ni-Co-Cu and metallic iron that might result from the partial serpentinization of ultramafic bodies (Haggerty, 1978). Such magnetic phases have Curie temperatures in the range 620–1,100 °C and thus, if present, would greatly deepen the Curie isotherm.

The 0.5-km-thick layer of pillow basalts makes a substantial contribution to the observed marine magnetic anomalies, but it cannot solely account for the magnitude of observed anomalies. Recent reviews of the sources of the magnetization of the oceanic crust (Johnson, 1979; Lowrie, 1979) favor an additional contribution from a magnetic layer of intrusive rocks below the pillow basalts.

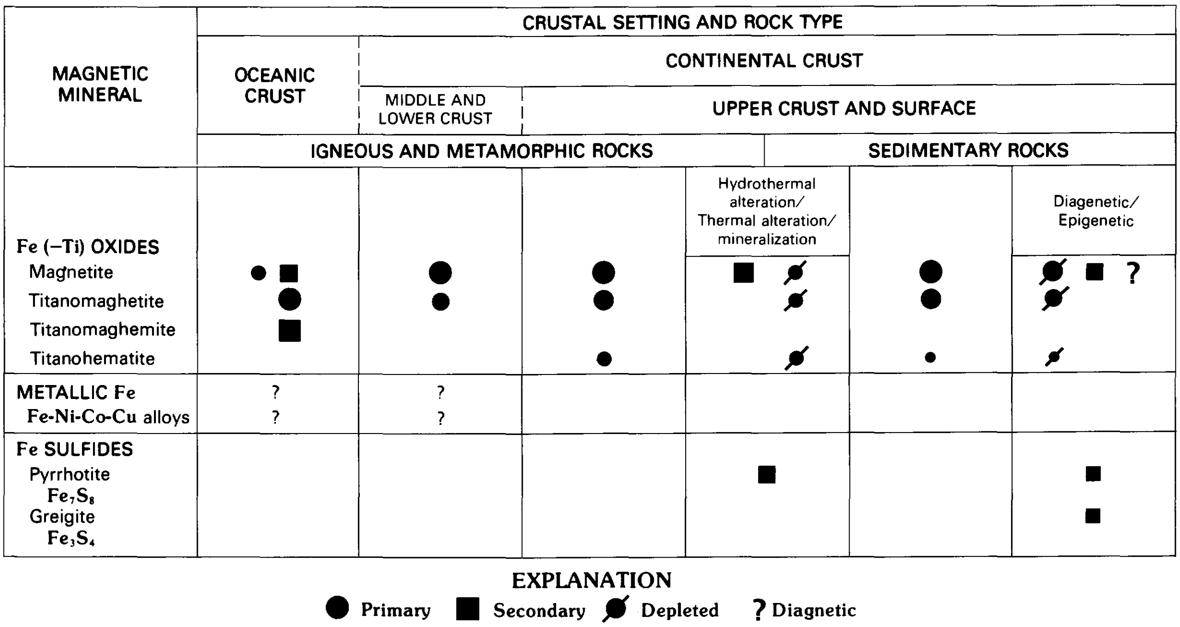

Figure 15. Summary of the crustal distribution of the aeromagnetically important minerals. As explained in the text, the minerals are divided according to primary (circle) and secondary (square) origin. The primary minerals are considered to be (l) minerals crystallized in magma, (2) deuteric alteration products in igneous rocks or metamorphic products in metamorphic rocks, or (3) detrital minerals in sedimentary rocks (including chemically precipitated magnetite, such as that deposited in banded iron-formations). The circle-slash symbol denotes settings or conditions under which minerals may be depleted. Diagenetic magnetite is represented by a bold query. Secondary minerals include those formed by replacement of earlier magnetic precursors (such as titanomaghemite from titanomagnetite in the oceanic crust) and those formed by nucleation or from a nonmagnetic precursor. For example, magnetite in the oceanic crust is considered to be both primary, formed during initial cooling by the exsolution of original $Mt\text{-}Usp_{88}$ (titanomagnetite), and secondary, formed by the hydrothermal alteration of either titanomagnetite or titanomaghemite. Under conditions of hydrothermal alteration, local thermal alteration (as in contact metamorphism), and mineralization in the continental crust (shown in the figure as hydrothermal alteration, thermal alteration, and mineralization, respectively), primary magnetite may be destroyed or new magnetite may be formed. Secondary magnetite may be destroyed later in these settings. The diagenetic and epigenetic alterations are represented in the column on the far right. The sizes of the symbols have not been derived quantitatively and are only crude indications of relative abundance of the different minerals. The symbol sizes have meaning only within a single column (crustal setting); a comparison of sizes between columns has no significance.

A recent iteration of this two-layer model consists of an upper, 0.5-km-thick layer with $M = 5$ A/m (where M is the magnetic dipole moment per unit volume, taking into account partial oxidation of original titanomagnetite) and a lower, 3.5-km-thick, dike and gabbro layer with $M = 0.5$ A/m (Banerjee, 1984). Magnetization in the lower layer may be carried by both primary titanomagnetite and magnetite of either deuteric or hydrothermal origin (Banerjee, 1984).

The magnetic properties of the first drill-core samples from beneath the pillow-basalt layer suggest a more complicated picture than that portrayed by two-layer models and provide direct evidence for a role for magnetite in oceanic-crust anomalies (Smith and Banerjee, 1986). The drill hole penetrated 1,076 m of crust in the Pacific Ocean about 200 km south of the Costa Rica Rift and intersected three distinct petrologic and magnetic layers. The upper layer of pillow basalts (about 0.5 km thick) has an average NRM moment of 5.5 A/m that resides in titanomaghemite. Below this layer is a mixed zone (270 m thick) of pillow lavas and dikes which have been hydrothermally altered to greenschist facies. The alteration produced some magnetite by the oxidation of primary titanomagnetite but more importantly promoted the replacement of titanomagnetite by silicate minerals. Consequently, this layer has a low NRM moment (average of 0.74 A/m) and contributes little to magnetic anomalies. The bottom 300 of the core consists of a sheeted dike complex, in which magnetite production has occurred as in the overlying layer but in which replacement by silicate minerals is minor. As a result NRM moment in the bottom layer is sufficiently large (average of 1.4 A/m) to contribute significantly to the anomalies.

Many factors cause vertical and lateral variations in the magnetization of the oceanic crust and thereby produce irregularities in magnetic anomaly patterns. These factors include the geometry, composition, and cooling history of intrusions below the upper extrusive rocks; the compositions and temperatures of hydrothermal fluids; the geometries of hydrothermal convection cells; and the positions of the cells with respect to the spreading ridge or to the accreting plate (Johnson, 1979; Smith and Banerjee, 1986; Cathles and Fehn, 1985). Better understanding of the influences of these factors on magnetic contrasts is important in mapping and interpreting the geology of the oceanic crust.

Lower Continental Crust

Long-wavelength magnetic anomalies sensed primarily from satellites are potentially useful for mapping large, deep crustal bodies that may reflect the tectonic and geochemical evolution of the continents (Mayhew and others, 1985; Mayhew and LaBrecque, 1987; Wasilewski and Mayhew, 1982; Wasilewski and others, 1979). These deep-seated anomalies require sources with large vertical (about 10–20 km) and lateral (about 100 km or more) dimensions and magnetizations of about 3–6 A/m, approximately 10–100 times greater than that of the average value of upper crustal rocks at the surface and even greater than those obtained from rare deep-crustal sections now at the surface (Shive and Fountain, 1988; Shive, 1989). Induced magnetization is an important (probably dominant) contributor to total magnetization; VRM may contribute to the magnetization of deep crustal sources.

Nearly pure magnetite is probably the major source of high magnetization in the lower crust. This conclusion comes from studies of xenoliths and uplifted sections of lower crustal rocks and from thermodynamic considerations (Wasilewski and Mayhew, 1982; Schlinger, 1985; Williams and others, 1985; Frost and Shive, 1986). The rock magnetic and compositional studies of lower crustal rocks are consistent with thermodynamic predictions of low titanium content in the magnetite; typical measured Curie temperatures range from about 550 to 580 °C. The few exceptions noted by Wasilewski and Mayhew (1982) are xenoliths from rift zones with Curie temperatures less than 300 °C and were interpreted by them to indicate the presence of deep crustal titanomagnetite in anhydrous, perhaps relatively reducing, zones in a steep geothermal gradient. Thus in most areas the Curie isotherm is controlled by nearly pure magnetite. At deeper crustal depths the isotherm is elevated from 580 to about 600 °C, because of the effect of pressure on Curie temperatures (Schult, 1970; Frost and Shive, 1986).

A discontinuity in magnetic mineralogy occurs at the mantle-crust boundary (Mohorovičić discontinuity, commonly called the "the Moho"); below the Moho nonmagnetic spinels dominate (Frost and Shive, 1986; Mayhew and others, 1985). Thus the Moho is considered to be the bottom of the magnetic crust, except in areas of high heat flow where elevated temperatures may raise the base of the magnetic crust far above the Moho (Mayhew and others, 1985). Nevertheless, certain satellite anomalies have dimensions and amplitudes that indicate very deep sources (Taylor and Frawley, 1987), possibly in the upper mantle (P. Taylor, oral commun., 1987); if this is the case, minerals other than magnetite apparently cause the magnetic contrasts. Haggerty (1978, 1979) and Haggerty and Toft (1985) have argued that metallic iron and iron alloys formed in altered serpentinites may account for deep-seated anomalies, but Frost and Shive (1986) dismissed a major contribution from these phases on thermodynamic, petrologic, and geologic grounds. The locations and causes of deep-seated magnetic anomalies are topics that deserve more study.

Other important uncertainties in the distribution of magnetization in the lower and middle crust include changes in the quantity of magnetite as a function of depth and the effects of temperature and pressure on magnitudes of susceptibility and VRM. Enhancement of susceptibility in magnetite with increasing temperature within about 150 °C of the Curie temperature (the Hopkinson effect) has been considered an important contributor to high magnetizations in the lower crust (Wasilewski and others, 1979). Recent experimental work by Schlinger (1985), however, suggests that this contribution may not be as large as previously thought. Another critical factor is the contribution of VRM to total remanence. Although the rate of VRM acquisition increases with increasing temperature (Dunlop, 1983; Shimazu, 1960), data are lacking on the variations in VRM magnitude as a function of temperature and pressure.

Upper Crust and Surface

Igneous and Metamorphic Rocks

At depths shallower than about 5 km in the continental crust, minerals of the magnetite-ulvöspinel series dominate magnetizations of most rocks that are the sources of aeromagnetic anomalies. In certain plutonic, volcanic, and uplifted high-grade metamorphic settings, however, ferrimagnetic titanohematite may also produce anomalies. Many rock types have broad and overlapping ranges of magnetic susceptibilities and magnitudes of NRM, and thus, correlations among aeromagnetic signatures and lithologies are commonly not possible (Clark, 1983). Nevertheless, examples of useful relations between the magnetizations and petrology exist.

An instructive example comes from an aeromagnetic and rock magnetic study of Precambrian igneous and metamorphic rocks of the Adirondack Mountains, New York, in which regional anomalies were closely correlated with mineralogy and lithology (Balsley and Buddington, 1958). Magnetic highs are caused by NRM in titanomagnetites, whereas magnetic lows from gneisses are produced by reversed remanent magnetizations in titanohematite. These reversed directions resulted from a self-reversing process (the acquisition of a remanent direction opposite to that of the applied field) that is peculiar to a narrow range of mineralogic compositions of ferrimagnetic titanohematite. Although potentially important in gneissic terranes, ferrimagnetic titanohematites are not the dominant magnetic species in most other rock types, based on abundant paleomagnetic and rock magnetic data acquired over the past 25 years. This conclusion differs from those of Grant (1984/1985a), who, from our perspective, overemphasized the importance of titanohematite solid-solution series minerals to the NRM of rocks and the role of self-reversed magnetizations in producing magnetic anomalies.

An example that illustrates the potential for identification of rock types by magnetic contrasts concerns granites of different origins. Petrochemical and isotopic distinction can be made between granites derived primarily from melting of igneous crust (I-type) and those from melting of sedimentary crust (S-type). Such a distinction, which reflects regional tectonic and magmatic evolution, may, in some settings, be discerned aeromagnetically. This is because magnetite tends to be sparse in granites derived from carbonaceous metasedimentary rocks (S-type) and abundant in granites generated by melting of rocks that have low contents of carbonaceous matter (I-type). Clark (1983) stated that similar magnetic distinctions exist for amphibolites derived from igneous or sedimentary rocks. In a study of granitoids of the Piedmont Province in the southern Appalachian Mountains, Wenner (1981) and Ellwood and Wenner (1981) found that plutons that have low initial ^{18}O content correspond to I-type granite and have relatively high magnetic susceptibilities (greater than 1.2 $\times 10^{-2}$ SI). In contrast, granites that have high initial ^{18}O content correspond to S-type granite and have susceptibilities less than about 1.2×10^{-2} SI. Such petrologic-magnetic relations are similar to those described by Ishihara (1977, 1981) for "magnetite-series" and "ilmenite-series" granites. The former contain more than 0.1 volume percent iron-titanium oxide (dominantly magnetite) and are considered important sources for magnetic anomalies, whereas the latter contain less than 0.1 percent iron-titanium oxides, mostly ilmenite, and hence usually have magnetizations (magnetic susceptibility less than about 0.3×10^{-2} SI) too low to produce important aeromagnetic anomalies.

Detailed oxygen-isotopic and magnetic susceptibility studies of aeromagnetic anomalies over the southern half of the Idaho batholith (Criss and Champion, 1984) demonstrate the usefulness of, and limits to, the interpretation of the extent and origin of magma types from magnetic data. Many unaltered samples exhibited relations between $\delta^{18}O$ values and magnetic susceptibility similar to those described for the Appalachian Piedmont granitoids (Ellwood and Wenner, 1981). However, because hydrothermal activity strongly modified the original $\delta^{18}O$ and susceptibility magnitudes in many exposed plutons, as discussed later, simple characterization of magma type by magnetic properties is not possible for most plutons (Criss and Champion, 1984). It is worth noting also that some plutons in the Idaho batholith defy the standard geochemical classification as I or S type, or as magnetite or ilmenite series, regardless of hydrothermal alteration.

Volcanic rocks exhibit many characteristics of their intrusive equivalents. However, combinations of exsolution and high-temperature oxidation commonly cause enormous variations in the magnetic properties of volcanic rocks, even within individual flows (Wilson and others, 1968; Haggerty, 1976). Moreover, in some extrusive rocks, notably welded ash-flow tuffs, emplacement and cooling phenomena, such as initial thickness and the effects of compaction, are capable of locally causing large, systematic changes in primary directions and magnitudes of NRM (Rosenbaum and Snyder, 1985; Rosenbaum, 1986; Rosenbaum and Spengler, 1986).

Remanence commonly dominates total magnetizations in volcanic rocks, mainly because of the typically small initial grain sizes and (or) the subdivision of magnetic phenocrysts into effectively smaller portions by exsolution and oxidation; such grains consequently produce relatively high TRM and low susceptibility. For example, an aeromagnetic low within the 23.1-Ma Lake City caldera, San Juan Mountains, Colorado, is produced by the reversely magnetized intracaldera ash-flow tuff (Grauch, 1987), the eruption of which caused the caldera collapse. Understanding of the magnetic sources was aided by modeling that used rock magnetic and paleomagnetic data from the caldera units (Grauch, 1987); the conclusions of this study also provide an instructive example of the interplay between remanent and induced magnetizations. Both types of magnetization exhibit large variations vertically and areally in the intracaldera tuff. The reversed NRM dominates magnetizations in parts of the intracaldera tuff to produce the observed low. Elsewhere, reversed remanent components are effectively cancelled by induced components to yield weak total magnetizations that have little aeromagnetic expression.

Recent contributions have shed light on petrologic controls of magnetite production in metamorphic rocks. Krutikhovskaya and others (1979) showed that premetamorphic compositions and the degree of metamorphism of rocks in the Ukrainian Shield are the main determinates of magnetite content. The quantities of magnetite and of iron + magnesium + manganese + titanium oxides, as well as the magnitudes of the rock magnetization, all increase with a decrease in $Si0_2$ content and with an increase in metamorphic grade. The results imply that metamorphic magnetite was derived from the breakdown of iron and magnesium silicate minerals (Grant, 1984/1985a).

Even in rocks that have similar original compositions, different metamorphic reactions and conditions may either produce or consume magnetite. For example, Robinson and others (1985) proposed that many of the magnetic anomalies associated with chemically similar metavolcanic rocks of the Carolina slate belt in North Carolina are caused by magnetic contrasts produced by different activities of water and oxygen during metamorphism. Apparently, iron was incorporated into magnetite during greenschist-facies metamorphism and into silicate minerals during amphibolite-facies metamorphism of equivalent rocks.

The control on magnetite content by metamorphic reactions is also demonstrated by a detailed study of a contact metamorphic aureole in pelitic rocks at the margins of an Upper Carboniferous granitoid pluton in South Carolina (Speer, 1981). Bulk rock compositions in the aureole hornfelses do not vary greatly. Controlled mainly by temperature but also by changes in fluid pressure and composition, many metamorphic reactions occurred over a short distance. These reactions produced magnetite in the outer and middle parts of the aureole but destroyed magnetite near the pluton under conditions of relatively high temperature and low oxygen fugacity. The metamorphic magnetite produced a ring-shaped magnetic high that is concentric with the contact between the pluton and the country rock.

Magnetite is not the only cause of annular magnetic anomalies around plutons. Pyrrhotite is also reported as the source of magnetic lows from contact metamorphic aureoles in Jurassic argillites around Late Cretaceous and early Tertiary plutons (mainly granodiorites) in the Alaska Range, south-central Alaska (Griscom, 1979; A. Griscom, oral commun., 1987). The magnetic lows are caused by reversed remanent magnetizations carried by the ferrimagnetic (monoclinic) pyrrhotite.

Metavolcanic rocks of virtually the same lithology may have different magnetization because of different original conditions in the source magmas. Urquhart and Strangway (1985) found distinct magnetization differences in a suite of metamorphosed iron-rich tholeiites, which had essentially the same major-element chemistry. They ascribed the differences to contrasts in the original temperature and oxygen fugacity conditions of the source magmas.

We next discuss some effects of hydrothermal fluids and of high-temperature, ore-related alteration on the original magnetizations of igneous, metamorphic, and sedimentary rocks. Although the magnetizations of some rocks exposed to such fluids and conditions may change little and remain dominated by the original magnetic particles, the chemical conditions involved in such alteration usually destroy or create magnetic phases (fig 15), and (or) may reset earlier remanence. The temperatures of such alterations are too low to allow the formation of $Mt\text{-}Usp_{ss}$ or $Ilm\text{-}Ht_{ss}$. Of the magnetic oxides, therefore, only magnetite is normally an aeromagnetically important secondary mineral in these settings. Many ores consist of, or are associated with, metal sulfide minerals, including magnetic pyrrhotite (Fe_7S_8). More common nonmagnetic sulfides, particularly pyrite (FeS_2), also play important roles in decreasing magnetization by replacing magnetic oxides or in indirectly enchancing magnetization by providing concentrations of iron that are easily converted to magnetite by oxidation at elevated temperature.

Hydrothermal Alteration

The geochemical and magnetic study of the Idaho batholith by Criss and Champion (1984) documented both enhancement and depletion of magnetization by hydrothermal activity (at temperatures between about 150 and 400 °C) related to the emplacement of Tertiary intrusions into Mesozoic plutons of the Idaho batholith. Rocks that underwent strong hydrothermal alteration in the interior of the batholith have lower magnetic susceptibilities than their unaltered equivalents, presumably because of oxidation of primary magnetite. Hanna (1969) reached a similar conclusion to explain low magnetizations of mineralized plutons of the Boulder batholith, Montana. In contrast, similar alteration, but probably involving less water and hence lower oxidation potential, increased magnetic susceptibility in parts of some leucocratic plutons in the Idaho batholith. This increase was ascribed to the growth of "hydrothermal" magnetite.

Little is known about the effects of hydrothermally driven metasomatism on the magnetizations of silicic volcanic rocks. It appears, nevertheless, that metasomatic conditions that involve oxygen-rich fluids destroy titanomagnetite. Metasomatic alteration, which resulted in enrichment of potassium, decreased magnetic susceptibility and remanent intensities of intracaldera welded tuffs relative to unaltered material of some ash-flow tuffs of the Mogollon-Datil volcanic field, New

Mexico (McIntosh, 1983). Oxidizing and perhaps slightly acidic conditions were thought to be responsible for the formation of hematite from magnetite and for the dissolution of magnetite. Preliminary study of the Carpenter Ridge Tuff (27.61 m.y.), the major host for precious- and base-metal vein deposits of the Creede mineral district in the Bachelor caldera of the central San Juan volcanic field, Colorado, shows that the potassium-metasomatized intracaldera facies (Bachelor Mountain Tuff) have much lower magnetization than that of non-metasomatized outflow facies (D. Sweetkind, written commun., 1989). Reduction of magnetization by potassium-metasomatism or other pervasive alteration may explain, at least in part, why some calderas, such as the Bachelor caldera, have indefinite aeromagnetic expression, whereas other calderas, which are filled with largely unaltered tuffs of nearly identical initial compositions, are the sources of large-amplitude magnetic anomalies (Williams and others, 1987)

Magnetization and Mineral Deposits

The practical value of magnetic exploration for mineral deposits has long been recognized (Grant, 1984/1985b; Clark, 1983; and Wright, 1981 for recent summaries) and is based on the experience that high or low magnetizations characterize some ore-forming environments. Magnetic contrasts may be related to formation or destruction of magnetic minerals in direct response to mineralization or related to a generally high or low magnetic-mineral content of a particular host rock or, on a larger scale, of a favorable rock assemblage. For example, certain types of mineral deposits were thought by Ishihara (1981) to be related to magnetite-series or to ilmenite-series magmatism with their respective high and low magnetizations. As a specific case, Hattori (1987) described an association between magnetite-bearing, anomaly-producing felsic intrusions and gold deposits in Archean rocks of the Superior Province of the Canadian shield. The association of magnetic intrusions and mineralization suggests either that ore constituents were derived from oxidized magmas or that the intrusions were emplaced in zones of deep flow of gold-bearing fluids. Other associations among magnetic patterns and mineral deposits can be attributed to tectonic setting, in particular to the extent of faulting and its controls on the chemistry and flow paths of altering fluids (McIntyre, 1980).

Effective use of magnetic patterns in mineral exploration depends on understanding relations among petrology and structure of the host terrane, mineralization, and magnetization. Many such relations are discussed in the review by Grant (1984/1985b). Yet much remains to be learned about links among ore petrology and geochemistry, the genesis of magnetic minerals, the structure of the host terrain, and magnetic anomalies, as well as about the applications of these links to effective magnetic exploration in frontier areas

Although a few mineral-deposit settings have a characteristic magnetic signature, most settings do not. This lack of uniformity has many causes, including: (1) variations in chemistry of mineralizing fluids; (2) differences in host-rock composition, structure, and crustal level; and (3) variations in the type and degree of post-ore alterations. We will not dwell at length on the associations among mineral deposits, magnetite, and magnetic signatures that are reported by other authors. However, we will examine some examples that provide typical reasons for different magnetic signatures that arise from similar mineral deposits. These examples are drawn from skarn and porphyry copper deposits and from the magnetite-pyrite association that is common to many types of deposits.

Skarn Deposits

Skarns are metal-bearing deposits associated with the hydrothermal replacement of carbonate rocks by a complex assemblage of Ca-Fe-Mg-Al silicate minerals. The classification, description, tectonic-magmatic settings, and origins of skarn deposits are presented in Einaudi and others (1981). Magnetite is common in many skarns and is the ore mineral in iron skarns. Magnetite enrichment is associated with many zinc-lead, molybdenum, magnesian tin-bearing skarn deposits and with some tungsten and calcic tin-bearing skarns. The magnetite content in skarns depends on many factors. For tungsten skarns, differences in magnetite content reflect different depths of emplacement of related intrusions and different oxidation states of the altered rocks (Einaudi and others, 1981). Tungsten skarns that were formed at great depth in a carbonaceous-rich, low-oxidation, and low-sulfidation environment contain abundant magnetite and only a small quantity of sulfide minerals, mainly pyrrhotite. In contrast, tungsten skarns that were formed at higher levels in carbonaceous-poor, intermediate-oxidation and intermediate-sulfidation environments may lack magnetite but tend to have relatively greater amounts of sulfide, principally pyrite. Copper skarn deposits are of special interest because of their common association with porphyry copper stocks (Einaudi and others, 1981; Einaudi, 1982a, 1982b; Titley, 1982; Beane, 1982; Titley and Beane, 1981; Beane and Titley, 1981). Magnetite is common in both calcic and magnesian copper skarns but is apparently more abundant in the magnesian skarns.

Porphyry Copper Deposits

The association of magnetite with many porphyry copper deposits makes these deposits attractive targets for magnetic exploration. Magnetic anomalies may arise from both the igneous intrusion and from associated peripheral skarns. Jerome (1966) presented a model of a copper porphyry that portrays secondary magnetite in skarn at the margins of a mineralized intrusion as the cause of ground magnetic and aeromagnetic highs that might define an annular anomaly around an orebody. Grant (1984/1985b) discussed porphyries (based largely on the work of Lowell and Guilbert, 1970) and concluded that recognition of favorable environments for porphyry copper deposits may be based on the magnetic identification of felsic-intermediate plutons of appropriate size (about 1–10 km in diameter) and on the possible existence of a "magnetite halo" within the pluton. The Ely, Nevada, area provides an example of positive aeromagnetic anomalies associated with a porphyry copper deposit. Magnetic anomalies reflect the high magnetizations both of the quartz-monzonite intrusion and of the magnetite-bearing skarn in nearby carbonate strata (Wright, 1981).

Other important magnetic characteristics of the porphyry copper environment have been discerned from a regional aeromagnetic map of Arizona which was designed to show large-wavelength anomalies from deeper and broader sources than those at the ore-deposit scale (Sumner, 1985). On this map most of the copper districts are associated with regional lows, with arcuate magnetic lows at the district level, or with a 400-km-long magnetic lineament that is interpreted to reflect a late Precambrian transform fault. Sumner (1985) suggested that the magnetic lows are caused by the destruction of presumable primary magnetite in hydrothermal fluids that circulated in deep fractures. Conversely, the regional magnetic data also revealed a relation between a few deposits and magnetic highs that may be at least partly explained by relatively mafic compositions of the mineralized intrusions.

Magnetite-Pyrite Association

The association between magnetite and nonmagnetic iron sulfide minerals is common to many types of mineral deposits. We discuss nonmagnetic iron sulfides (referred to here as pyrite, the most common variety) because they may play important but not fully recognized roles in the magnetic contrasts of some mineral deposits.

The relationship between magnetite and pyrite in some types of mineral deposits can be explained on the basis of thermodynamically controlled sequences of precipitation (McIntyre, 1980; Grant 1984/1985b; Beane, 1982). Nevertheless, other explanations for the magnetite-pyrite association should be considered, especially in view of the complex thermal, tectonic, and geochemical histories of many mineral districts. For example, pyrite can be converted to magnetite by the loss of sulfur at elevated temperatures. In addition, magnetization can be reduced by the replacement of primary or secondary magnetic iron oxide minerals by nonmagnetic sulfide minerals (Wright, 1981). Such distinctions, which can usually be made on petrographic criteria, are obviously important for genetic models that link mineral assemblages and paragenesis to magnetic anomalies.

The anomaly over the Calico Hills in southwestern Nevada is an example of magnetic high that is caused by the high-temperature conversion of pyrite to magnetite. The anomaly is centered over an area underlain by a thick Paleozoic section which is capped by a thin veneer of hydrothermally altered Tertiary ash-flow tuffs. Bath and Jahren (1984) attributed the anomaly to strongly magnetized, thermally altered Paleozoic strata. Measurement of magnetic properties of drill-core samples indicates that some of the Paleozoic sedimentary rocks are nonmagnetic, whereas others are strongly magnetic (Baldwin and Jahren, 1982); the strongly magnetic rocks have an average NRM of about 4 A/m. Petrographic observations clearly demonstrate that the magnetic rocks contain magnetite which formed by the oxidation of pyrite and the resultant loss of sulfur. The magnetite occurs as replacements of pyrite cubes and pyritohedrons, pyrite veinlets, pyrite that filled cell lumens in detrital plant fragments, and framboidal pyrite. In contrast, nonmagnetic rocks either contain pyrite in the above forms (and little or no magnetite) or show no evidence of having ever contained pyrite.

Sedimentary Rocks

The magnetizations of epiclastic sedimentary rock depend directly on the abundance of detrital magnetic minerals and on the geochemical history of the rocks that either preserved or destroyed detrital magnetic minerals or that created diagenetic-authigenic magnetic oxide or sulfide minerals. Postdepositional effects on rock magnetization are especially important because of their potential as indicators of mineral and hydrocarbon deposits. Although most sedimentary materials contain magnetizations that are useable for paleomagnetism, the relatively thin sedimentary crust is normally not an important contributor to aeromagnetic anomalies. Even the most magnetic of the epiclastic sedimentary rocks generate low-amplitude, high-wavelength anomalies. Because flat-lying or shallowly dipping strata, even if strongly and uniformly magnetic, do not generate strong magnetic contrasts except locally at their edges, most anomalies from sedimentary rocks are those associated with steeply dipping or faulted units or with alterations controlled by fluid migration paths that either are vertically directed or are limited in horizontal extent.

Ironically, the most magnetic rocks, on the average, are sedimentary—the Early Proterozoic quartzmagnetite banded iron-formations (Clark, 1983). These rocks, which contain iron ores, can be easily detected by aeromagnetic methods. Magnetite in these rocks was chemically precipitated from sea water that was rich in dissolved ferrous iron before the buildup of free oxygen in the oceans. The magnetics and economic importance of banded iron-formations are described by McIntyre (1980) and Grant (1984/1985b), and we will not discuss further chemically precipitated, primary magnetite.

Strongly magnetic detrital minerals in the sedimentary rocks are primarily magnetite and titanomagnetite derived from crystalline or other sedimentary source areas. Ferrimagnetic titanohematite is generally rare or absent, but it may be the dominant magnetic oxide in some sedimentary rocks (Reynolds, 1977; Reynolds, 1982; Butler, 1982). Concentrated detrital magnetic grains can, of course, be responsible for anomalies, as shown below; for the most part, however, ferrimagnetic oxides are destroyed under most conditions of oxidation and reduction in sediments and sedimentary rocks. The extent of the destruction depends primarily on the duration and nature of altering conditions (such as Eh, pH, temperature, bacterial activity, concentrations of Fe^{2+} and Fe^{3+}, types and abundances of sulfur species, and organic matter). Such destruction includes dissolution and the oxidative or reductive replacement of the detrital oxides by essentially nonmagnetic minerals, such as of magnetite by hematite or by iron sulfide, respectively. Uncommon, but important in certain sulfide reducing environments, is the authigenesis of ferrimagnetic sulfide minerals, monoclinic pyrrhotite (Fe_7S_8) and cubic greigite (Fe_3S_4). The postdepositional growth of magnetite is currently a topic of growing interest among aeromagnetists and paleomagnetists. Such growth may be via inorganic geochemical pathways (Murray, 1979) or may be the result of bacterial metabolism. Two different types of magnetite-producing bacteria have been identified. Magnetotactic bateria produce small quantities of intracellular magnetite (less than 0.2-μm diameter) and operate in the presence of oxygen in marine, brackish and fresh-water environments; thus they deposit magnetite in surficial sediment layers that contain dissolved oxygen (Blakemore, 1982; Kirschvink and Chang, 1984). In contrast, dissimilatory iron reduction by anaerobic bacteria appears to be capable of producing large amounts of fine-grained (less than 0.2-μm diameter) magnetite outside the microbial cell (Lovley and others, 1987). The only known strain of dissimilatory iron-reducing bacteria was isolated in 1986 from fresh-water anaerobic sediments of the Potomac River (Lovley and others, 1987; Lovley and Reynolds, 1987). Possible roles of anaerobic, magnetite-producing bacteria in generating magnetic anomalies in sedimentary rocks (especially in fault zones above hydrocarbon deposits) are topics immediately worthy of intensive research.

RESEARCH FRONTIERS: PROBLEMS INVOLVING SEDIMENTARY ROCKS

Alteration Associated with Sandstone Uranium Deposits

The following examples from uranium-bearing sandstones in the San Juan basin, New Mexico, and in the south Texas coastal plain illustrate the destruction of detrital magnetic grains. The two areas have either not been surveyed or not been assessed in sufficient detail to determine whether ore-related alteration has caused magnetic anomalies; moreover, rock magnetic studies are incomplete or lacking in these examples. However, on the basis of the originally high contents of detrital magnetic minerals and on the spatial distribution of zones that contain or lack the magnetic grains, it seems likely that current aeromagnetic techniques are capable of detecting the effects of near-surface, uranium-related alterations on sandstones that originally contained abundant detrital magnetic grains.

Most uranium deposits that occur in Phanerozoic sandstones are classed either as tabular or as roll-type deposits (Nash and others, 1981). The largest uranium reserves in the United States are found in the tabular ore deposits of the Upper Jurassic Morrison Formation on the Colorado Plateau. Many of these deposits differ in some important genetic aspects, but all share a common association with organic matter. The tabular orebodies in sandstones of the Westwater Canyon and Brushy Basin Members of the Morrison in the Grants uranium region in west-central New Mexico consist of uranium intermixed with epigenetically introduced organic matter that was expelled by compaction from overlying mud-flat facies of the Brushy Basin Member (Turner-Peterson and Fishman, 1986; Fishman and Turner-Peterson, in press). The organic matter, which was derived from detrital carbonaceous debris, was apparently transported as humic and perhaps as other organic acids in alkaline, mildly reducing fluids. Such acids leach iron from the Fe-Ti oxide minerals (Lynd, 1960; Schnitzer and Skinner, 1965; Baker, 1973; Adams and others, 1974; Schnitzer and Kodama, 1977).

Petrographic, geochemical, field and mine mapping, and laboratory and borehole magnetic susceptibility studies document important relations among the genesis of the uranium deposits (including the nature and timing of ore formation and paths of mineralizing fluids), the alteration of Fe-Ti oxides, and the magnetic properties (Adams and others, 1974; Adams and Saucier, 1981; Fishman and others, 1985; Fishman and Reynolds, 1986; Reynolds, Fishman, Scott, and Hudson, 1986; Fishman and Turner-Peterson, in press). The Fe-Ti oxide minerals (principally titaniferous magnetite and ilmenite) were abundant detrital components of the Morrison sandstones that are now exposed at the margins of the San Juan basin and are still well preserved in places, where they compose nearly 1 weight percent of the rock. In these sandstones, susceptibilities and remanent moments are typically greater than 5×10^{-4} SI (maximum 8×10^{-3} SI) and 10^{-4} A/m, respectively. In sandstone beds that encompass the uranium deposits (the altered zone), however, the detrital oxides are highly altered by iron dissolution and consist typically of relict titanium dioxide. The abundance of such relicts, as well as the spatial and paragenetic relations, indicate postdepositional alteration of detrital magnetic grains during ore formation, which was most likely caused by humic-rich solutions that originated in the Brushy Basin Member (Adams and others, 1974; Adams and Saucier, 1981; Fishman and others, 1985; Turner-Peterson and Fishman, 1986). In the altered zone, magnetic susceptibilities and remanent moments are uniformly less than 5×10^{-4} SI and 10^{-4} A/m, respectively. Based on borehole core studies, the altered zone is as much as 100 m thick and extends from the base of the Brushy Basin Member to below the ore lenses, which are typically only 2–10 m thick. Outcrop studies reveal that this pattern of alteration extends essentially uninterrupted many tens of kilometers along the southern and south-western margins of the basin. The studies also show that the degree of alteration of Fe-Ti oxide minerals decreases away from the Brushy Basin Member. Where the mud-flat facies of the Brushy Basin Member (the source of the humic acids) dies out to the north, the Westwater Canyon Member lacks uranium mineralization and contains abundant, unaltered Fe-Ti oxides through its entire stratigraphic extent (Fishman and Turner-Peterson, in press). Thus the presence of uranium deposits and the related destruction of detrital magnetic grains in the Westwater Canyon Member has been controlled largely by distribution of the mud-flat facies of the overlying Brushy Basin Member (Turner-Peterson and Fishman, 1986).

Roll-type uranium deposits form by invasion of geochemically reduced, iron disulfide (FeS_2) bearing sandstone by oxygenated, uranium-bearing ground waters, followed by the reduction-precipitation of the uranium. Encroachment of such waters into reduced strata creates a tongue-shaped zone of oxidized rock (the altered or oxidized tongue) and a redox (reduction-oxidation) boundary that separates the tongue from reduced ground. Uranium is concentrated in a crescentric envelope or roll on the reduced side of the redox boundary, with the convex side of the roll front pointing in the downdip direction into reduced rock.

Magnetic contrasts may develop in host beds of some roll-type deposits as a result of the replacement of detrital magnetic grains by iron sulfide minerals. In the south Texas coastal plain uranium deposits in the Miocene Catahoula Sandstone (for example, the Benavides deposit) and in the overlying Miocene Oakville Sandstone (for example, the Felder-Lamprecht deposit, Ray Point district), pre-ore sulfide (pyrite and marcasite) created a reduced geochemical and mineralogic zone that controlled the concentration of uranium. The FeS_2 minerals extensively replaced detrital magnetic grains (Reynolds and Goldhaber, 1978; Reynolds, 1982; Goldhaber and others, 1978, 1983). In the Benavides deposit, the sulfidization resulted from the invasion of the shallow (30–45 m deep) aquifers by sulfidic brines from deep sour-gas reservoirs along a normal growth fault that lies 1.5 km downdip from the roll front. The FeS_2 partly to completely replaced titaniferous magnetite grains as much as 2 km updip from the fault. Ferrimagnetic titanohematite, which was initially about half as abundant as magnetite, was more resistant to sulfidization, and consequently it is the dominant magnetic oxide in the reduced zone and in the altered tongue. More than 2 km updip from the fault, the sandstone was never sulfidized; in these beds the Fe-Ti oxide population has only been affected by partial hematite replacement of magnetite (Reynolds, 1982).

Although the magnetic susceptibility of disaggregated core samples was not measured, the abundances and distribution of the magnetic oxides were examined using separates of magnetic minerals. The magnetic fraction, as a percentage of the whole rock, ranges from nearly 1 weight percent in the oxidized beds farthest (2.5 km) updip from the fault to less than 0.1 weight percent in reduced rock closer to the fault. The magnetic fraction of the samples in between decreases systematically within this range with proximity to the fault. This pattern and petrographic-geochemical studies strongly suggest that the magnetic minerals were nearly uniformly distributed in the host beds by depositional processes and that the degree of replacement by sulfide minerals decreased systematically updip from the fault (Reynolds and Goldhaber, 1978; Goldhaber and others, 1978).

A similar pattern, but presumably a more subdued contrast, appears to be present in the Lamprecht-Felder deposit based on laboratory magnetic susceptibility measurements of core samples (from about 60–80-m depth) from the Lamprecht part of the deposit. Values of susceptibility are highest ($3–4 \times 10^{-4}$ SI) in the host beds that lie updip from the roll front and the sulfide-producing Oakville fault (1 and 1.5 km distant, respectively) and decrease systematically (to about

2×10^{-4} SI) in the roll front and closer to the fault (Scott and Daniels, 1976; Scott and others, 1983). Again, the increase in magnetization updip from the fault was likely caused by a decrease in intensity of sulfidization and the corresponding decrease in replacement of detrital magnetic grains.

Observations from the two roll-type uranium deposits suggest that the delineation of growth faults by their magnetic expressions may be a general guide for sulfidically reduced zones with potential for uranium mineralization elsewhere along the Texas coastal plain. Diminished magnetizations are predicted on the updip (northwest) sides of faults. Detailed aeromagnetic surveys directly over known faults, many of which tap the hydrocarbon-producing deep Edwards reef trend, would be desirable, because the faults themselves may have a distinctive signature. Such a possibility is based on observations that faults and fractures that tap hydrocarbon deposits elsewhere can be recognized as magnetic features on aeromagnetic maps. An example is the association between fractures and magnetic contrasts over oil fields in the Williston basin, Montana (R. Wold, oral commun., 1985). Geochemical and mineralogic causes that change magnetization along faults, especially those that may be associated with oil and gas fields, are topics of obvious importance to studies of aeromagnetism and rock magnetism.

In addition to the roll-type uranium deposits in sandstones that contain extrinsic (postdepositionally introduced) sulfide and that lack detrital organic matter, an equally important subclass of roll-type deposits is found in sandstones that contain detrital plant debris. Examples include the roll-type deposits in the Wyoming intermontane basins and in the Eocene Whitsett Formation of the south Texas coastal plain. In these "biogenic" deposits, the pre-ore sulfide is related to the metabolic activity of sulfate-reducing bacteria that feed on carbonaceous matter (Rackley, 1972; Granger and Warren, 1969; Reynolds and Goldhaber, 1983). In biogenic deposits that we have studied, magnetic oxides have been little affected by sulfidization; magnetite may be overgrown by pyrite, but replacement by pyrite is rare. These associations may be related to the typically low concentrations of sulfide in these ore-forming systems. Sulfide sulfur averages between 1 and 4 weight percent of the host rock in the Benavides, Lamprecht, and Felder deposits, but less than 1 percent in a deposit in the nearby Eocene Whitsett Formation. These observations would appear to contraindicate aeromagnetic exploration for roll-type deposits in areas that are remote from faults along which sulfidic brines have passed. Nevertheless, the fluvial sandstones in which most roll-type deposits occur may be aeromagnetic targets, based on the likelihood of greater concentrations of heavy minerals in these relatively coarse-grained, high-energy facies than in enclosing mudstone.

Alteration Associated with Hydrocarbon Seepage

Recently proposed models that link hydrocarbon seepage, authigenesis of magnetic minerals in near-surface strata, and high-frequency magnetic anomalies have encouraged magnetic exploration for oil and gas deposits (Donovan and others, 1979, 1984; Foote, 1984; Saunders and Terry, 1985; McCabe and Sassen, 1986). Interest in this approach stemmed mainly from an aeromagnetic survey of the Cement oil field, Oklahoma, in which total-field anomalies, which have short wavelengths and low amplitudes (less than 40 nT), were detected over an area of past oil seepage and current production (Donovan and others, 1979). Donovan and others (1979) reported as much as 1.2 weight percent magnetite in well cuttings from bleached Permian red beds over the producing reservoirs. They proposed that the magnetite formed by reduction of ferric oxide in the presence of upward migrating hydrocarbons and that the magnetite was the source of the anomalies.

Similarly, aeromagnetic surveys over vast regions of Alaska's North Slope have revealed several areas of anomalous magnetizations over known oil fields and over structures that appear favorable for hydrocarbon deposits based on other geologic and geophysical criteria (Donovan and others, 1984). These anomalies were likewise attributed to formation of magnetite or other magnetic minerals under reducing conditions that were related to hydrocarbon seepage.

Support for an association between magnetite and hydrocarbons comes from recent reports of magnetite with diagenetic morphologies in bitumen and other solid hydrocarbons from surface seeps (McCabe, 1986; McCabe and others, 1987; Elmore and others, 1987). In addition, magnetite in similar forms has been reported in Paleozoic carbonate rocks; in these occurrences diagenetic processes that are related to hydrocarbon migration have been the favored interpretation for the origin of the magnetite (McCabe and others, 1983; Elmore and others, 1986).

Our investigations of the magnetic anomaly-hydrocarbon seepage problem have involved rock magnetic, paleomagnetic, petrologic, and geochemical studies of samples from the Cement oil field, the Simpson oil field on the North Slope of Alaska, and the Jurassic Preuss Sandstone in the Wyoming-Idaho-Utah thrust belt. This work has not found a connection among magnetite, hydrocarbons, and aeromagnetic anomalies.

Instead, the work demonstrates an important association among magnetic iron sulfide minerals, leaked hydrocarbons, and bacterial activity at Cement, and suggests a similar association at the Simpson field (Reynolds and others, 1984; Reynolds, Fishman, Hudson, and others, 1986). Studies of the Preuss illustrate the case in which magnetic anomalies in an area of hydrocarbon potential have been caused by detrital magnetite and are unrelated to seepage (Hudson and others, 1985; Fishman and others, 1989; Reynolds, Fishman, Hudson, and others, 1986). Diagenetic magnetite is represented on figure 3 with a bold query, because, to our knowledge, there are as yet no confirmed cases in which secondary (authigenic) magnetite has produced aeromagnetic anomalies.

Cement Oil Field

Studies of well cuttings, including those examined by Donovan and others (1979), of shallow (surface to about 36 m deep) core and of surface outcrop and quarry exposures have not turned up any evidence for diagenetic magnetite at Cement. Magnetite is abundant in most well cutting samples, but it exhibits synthetic, metallographic textures and (or) contains inclusions of industrial steel (Reynolds, Fishman, Hudson, and others, 1986; Reynolds and others, 1988; Reynolds, Fishman and others, 1990). The magnetite is interpreted to be a contaminant related to drilling; magnetite is a common product of corrosion-oxidation of drill stems (Gray and others, 1980). Furthermore, we have found magnetite in the rusty scale of a used oil-field drill collar, and this magnetite exhibits many textural features similar to those of the Cement magnetites.

Many samples from the cuttings, cores, and surface, however, contain authigenic ferrimagnetic pyrrhotite, which is commonly intergrown with and (or) replaced by pyrite and marcasite. Two lines of evidence—the distribution of the pyrrhotite and the sulfur isotopic composition of the sulfide minerals—strongly suggest that the pyrrhotite is related to migrated hydrocarbons. Pyrrhotite is limited to strata above the oil-producing structures, where it appears to be concentrated in a shallow (200–500 m deep) zone that cuts across lithologic and formational boundaries. Away from the main field and the area of hydrocarbon alteration at the surface, well cuttings do not contain pyrrhotite. Variations in the sulfur isotopic compositions of the sulfide minerals and comparison with published isotopic values of crude oils at Cement (Lilburn and Al-Shaieb, 1984) suggest that the pyrrhotite and other sulfide minerals formed from at least two different sources of sulfide and by a combination of inorganic and organic mechanisms (Reynolds, Fishman, and others, 1990). Aqueous sulfide in hydrocarbon-related fluids that leaked upward from depth along numerous pre-Permian faults in the area was probably the dominant sulfide source for the sulfide minerals at depths greater than about 300 m. There the sulfide minerals apparently formed by the inorganic reaction of aqueous sulfide with available iron. Fault-derived sulfide also probably contributed to the iron sulfide minerals at shallower depths, but this contribution waned toward the surface. With decreasing depth, increasing amounts of sulfide produced by sulfate-reducing bacteria were also incorporated into the iron sulfide minerals. Because these Permian beds contain only small amounts of organic carbon (typically less than 0.1 percent) and lack detrital debris, the sulfate-reducing bacteria must have derived energy from some food source other than plant matter. Organic compounds in or derived from hydrocarbons were likely sources for bacterial consumption in the hydrocarbon seepage plume at Cement.

The contribution of pyrrhotite to the anomaly at Cement is difficult to determine with certainty (Reynolds, Webring, and others, 1990). The magnetic properties of the pyrrhotite-bearing samples (borehole cuttings) necessary for magnetic modelling either cannot be measured, such as NRM direction and magnitude, or cannot be determined with confidence (magnetic susceptibility) because of contamination by magnetite. Moreover, uncertainty regarding the nature of the Cement anomaly and the likely effects of cultural interference have clouded the interpretation picture (Donovan and others, 1986; Boardman, 1985; Foote, 1984). We may conclude, nevertheless, that pyrrhotite is related to hydrocarbon seepage and that is the only possible natural source for anomalous magnetization at Cement.

Simpson Oil Field

At the Simpson oil field, ferrimagnetic greigite (Fe_3S_4) is concentrated locally in Upper Cretaceous sandstone, siltstone, and mudstone. Detrital Fe-Ti oxide minerals occur with greigite in some samples from borehole cores but rarely compose the majority of magnetic grains. Among the detrital oxides, ferrimagnetic titanohematite is more common than titanomagnetite; petrographic observations suggest that this is due to the preferential dissolution of magnetite after deposition. Intensities of NRM of the greigite-bearing samples range from 3×10^{-3} to 2×10^{-1} A/m, whereas those of the samples dominated by the magnetic oxides range from only 6×10^{-4} to 10^{-2} A/m. Magnetic susceptibilities (4×10^{-5} to 3×10^{-3} SI units) are unimportant contributors to the total magnetizations. The remanence that resides in greigite is therefore probably responsible for the anomalies over the Simpson field.

A connection between the greigite and hydrocarbon seepage, however, has not been unambiguously established. The greigite probably formed by processes that involve sulfate reduction by bacteria (Berner, 1981) that used either organic compounds derived from leaked hydrocarbons or detrital organic matter, or both, as food sources. Available evidence points to a combination of factors. The greigite that is found on the surfaces and within cell lumens of abundant detrital plant fragments in many samples from the Simpson field may have formed during early diagenesis in the absence of hydrocarbons. However, greigite is present in some beds that are devoid of plant matter and absent in others that contain such matter. Geochemical studies that involve detailed sulfur isotopic analysis of the organic sulfur and mineral sulfur species may help to resolve the problem.

Preuss Sandstone

The Preuss Sandstone in the Wyoming-Idaho-Utah thrust belt was chosen as one of several targets for testing possible connections between hydrocarbon seepage and magnetic anomalies for the following reasons. An aeromagnetic survey (U.S. Geological Survey, 1981) showed positive magnetic anomalies west of the Absaroka thrust fault in the central thrust belt on trend with hydrocarbon production to the south. A subsequent ground magnetometer survey (S. Oriel and D. Mabey, written commun., 1982) and a preliminary magnetic study (Fishman and others, 1989; Reynolds, Fishman, Hudson, and others, 1986) confirmed the Preuss as the source of the positive anomalies. The magnetic study also revealed that the Preuss possesses a stable remanent magnetization carried by magnetite.

Paleomagnetic analysis indicated that the stable magnetization in the Preuss was acquired after deposition and at least partly during folding (Hudson and others, 1985, 1989). These stable synfolding directions are interpreted to be a viscous partial thermoremanent magnetization (VpTRM), which was acquired at low temperatures (less than 150 °C) over several tens of millions of years as thrusting migrated from west to east. Similar synfolding behavior in some other sedimentary units (principally limestones) elsewhere has been attributed to diagenetic magnetite formed from hydrocarbon-bearing fluids that were mobilized during deformation. In some of these carbonate units, magnetite has been identified in magnetic extracts that include grains with secondary (authigenic or diagenetic) morphologies (McCabe and others, 1983; Elmore and others, 1986). In contrast to these rocks, the Preuss lacks diagenetic magnetite. Rather, the magnetization in the Preuss resides in large (10–80-μm diameter), detrital titaniferous magnetite (contains as much as 14 mol percent ulvöspinel) grains, which are commonly concentrated in heavy-mineral laminations (Hudson and others, 1985; Fishman and others, 1989). The abundance of the detrital magnetite, which can be estimated from magnetic susceptibility, decreases systematically from west (nearer the source area for the magnetite) to east across the depositional basin. The gradient in the susceptibility values has been accentuated by thrusting and folding, which shortened the area by about 50 percent.

Thus, no evidence was found in the Preuss to link the magnetic anomalies with diagenetic magnetite formed under the influence of hydrocarbon-bearing fluids. Instead, the location of the magnetic anomalies is related to the distribution of abundant detrital magnetite. The distribution was controlled by the original depositional setting, by later thrusting and folding, and by the preservation of the titaniferous magnetite by calcite and by a favorable ground-water chemistry.

These results confirm that investigations of sources of magnetic anomalies over oil fields should encompass a wide variety of stratigraphic, geochemical, and structural settings to cover the wide range of diagenetic conditions that may affect magnetization in strata above oil deposits. Even the results of the limited and largely preliminary studies described above indicate that different styles of diagenesis can accompany hydrocarbon seepage and that aeromagnetic data should be cautiously interpreted in the absence of rock magnetic studies.

CONCLUSIONS

Magnetite (Curie temperature about 580 °C) is the dominant source of magnetic anomalies in continental crust. Magnetite is thought by most investigators to be the main source of anomalies from metamorphic rocks within the lower continental crust and thus to control the Curie isotherm in this part of the Earth. The proposed existence of metallic iron and iron alloys in the lower crust and possibly the upper mantle has fostered controversy that will certainly generate more study of this topic. Magnetite also carries the induced and remanent magnetizations in metamorphic rocks and in many acidic plutonic rocks at shallower depths. Titaniferous magnetite, which has lower Curie temperatures, is abundant primarily in basic igneous rocks at mid-crustal and shallower depths. Magnetic titanohematite may dominate the magnetizations of some acidic plutonic, gneissic, and dacitic volcanic rocks, but it is not an important source of anomalies in other settings.

In contrast, titanomagnetite (Curie temperature about 175 °C) and its alteration product, titanomaghemite, dominate magnetizations in upper layers of the oceanic crust, and these minerals probably are the major source of linear magnetic anomalies from the sea floor. At depth, however, magnetite may form by deuteric and hydrothermal alteration of the oceanic crust. Thus the depth to the Curie isotherm of the oceanic crust is controlled by the intrusive, cooling, hydrothermal history of the crust.

Magnetic contrasts in sedimentary rocks are caused by differences in the abundances of detrital magnetic minerals or by diagenetic and epigenetic destruction or addition of magnetic minerals. The types and abundances of detrital magnetic minerals in epiclastic sedimentary rocks reflect the relative and absolute abundances of these minerals in the source areas. Thus the magnetizations of epiclastic sedimentary rocks reside mainly in magnetite and titanomagnetite. Although titanohematite is more resistant to oxidation and reduction in the sedimentary environment than are the Mt-Usp minerals, it is insufficiently common to contribute significantly to magnetic signals. Even where titanohematite is abundant, the distinction between the detrital Mt-Usp and Ilm-Ht magnetic grains is not important for aeromagnetic interpretations.

Destruction of detrital magnetic minerals by post-depositional alteration may produce magnetic contrasts that are diagnostic for structural features such as faults or for facies changes. Such destruction arises from oxidative replacement of magnetite by hematite or from dissolution of magnetite or its replacement by sulfide minerals. Anomalies related to such magnetic contrasts may be useful exploration guides for stratabound mineral deposits.

Magnetic anomalies in areas of oil and gas potential are of current interest as possible indicators of secondary magnetic minerals that are related to hydrocarbon seepage. In such settings rock magnetic, mineralogic and geochemical studies can contribute to making crucial distinctions between primary and secondary magnetic minerals and between primary (DRM) and secondary (CRM, or other) remanence. In addition, the identification of the type of any secondary magnetic mineral (for example, iron oxide or iron sulfide) is important in developing geochemical models for seepage-related magnetizations that can be used for predicting its occurrence elsewhere.

Finally, we emphasize again the importance of alteration on the magnetization of rocks and hence on their aeromagnetic signatures. This conclusion has been stated recently by Mayhew and LaBrecque (1987), and it bears repeating. On a global scale, the types and degrees of metamorphism and hydrothermal activity in both continental and oceanic crust exert fundamental, widespread controls on crustal magnetic properties. On regional and local scales, rock alterations can cause magnetic contrasts that may be diagnostic for certain mineral and energy habitats.

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