RUSSIAN JOURNAL OF EARTH SCIENCES, VOL. 3, NO. 2, PAGES 97–124, MAY 2001

Petromagnetism of the continental lithosphere and the origin of regional magnetic anomalies: A review

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Abstract. The paper discusses the current ideas of regional magnetic anomaly sources and experimental evidence for the formation and behavior of various magnetic minerals within a wide range of pressure, temperature and oxygen fugacity. Four thermodynamic zones of formation conditions of magnetic minerals are shown to exist; these are the hematite, magnetite, silicate and metal-Fe zones successively changing, as the oxygen fugacity decreases (from strongly oxidizing to strongly reducing conditions). The effects of pressure and diffusion processes on titanomagnetite alterations, as well as the oxygen fugacity and fluid composition implications for the composition and concentration of magnetic minerals, are considered. Experimental studies show that ferromagnetic minerals do not form from silicates under "dry" conditions or in the presence of water vapor. The paper presents results derived from our studies of petromagnetic characteristics of rocks that formed under near-surface conditions (basalts and gabbroids) and in the lowermost continental crust(xenoliths in igneous rocks of Afar, Mongolia, the Lesser Caucasus, Kurile Islands, and Yakutia), as well as rocks from Archean-Proterozoic metamorphic sequences (Aldan and Anabar shields and Voronezh crystalline massif). The implications of secondary processes, such as chloritization and amphibolization, for the alterations in the ferromagnetic fraction and magnetic properties of these rocks are considered. Our results and literature data reviewed indicate that, since the Archean, igneous rocks formed in extension zones under surface and near-surface conditions have made a major contribution to the crust magnetism and regional magnetic anomalies. This situation is presently preserved in spite of metamorphism and substantial recrystallization at various depths.

1. Introduction

The problems of the geological origin of regional magnetic anomalies and the petromagnetism of the lower continental crust have been studied for many years. However, the origin of rock magnetization in the lower crust is still debatable: some researchers relate this magnetism to the granulite-facies metamorphism, whereas others believe that the primary magmatic conditions of the formation of magnetic minerals are the main controlling factor and, although deep metamorphism alters these minerals, the primary magnetism or nonmagnetism of rocks remain, as a rule, preserved.[1] On the other hand, the works of Russian authors are virtually unknown abroad. It is for these two reasons that we undertook this work.

The magnetic survey is the least expensive and most accessible geophysical method of studying the lithosphere. For this reason, nearly the entire surface of the Earth is covered by the surface, airborne and satellite magnetic surveys. However, the spatial distribution of magnetic masses in the lithosphere and their effective magnetizations are difficult to estimate due to the ambiguity of magnetic data inversion. Incorporation of other geophysical data reduces the ambiguity of the problem but does not remove it. Also, the prob-

Paper number TJE01059.
CCC: 0000–0000/2001/0302–00059$18.00

The online version of this paper was published June 26, 2001.
URL: http://rjes.agu.org/v03/tje01059/tje01059.htm
Print companion will be issued.

[1] Hereinafter we accept that rocks are magnetic or nonmagnetic if they, respectively, can or cannot produce magnetic anomalies; their respective magnetizations are more than 0.5–1.0 A/m or less than 0.1 A/m.

lems of the rock magnetism origin, sources, preservation and evolution remain relevant. Their solution is largely assisted by magnetomineralogical and petromagnetic studies of rocks and by experiments reproducing the $T-P-fO_2$ conditions of the rock and magnetic mineral formation.

Many magnetic and petromagnetic studies focus on near-surface rocks in areas where magnetic anomalies can be directly compared with concrete exposed geological bodies; rarer studies are devoted to the origin of regional magnetic anomalies produced by deep-seated magnetic bodies and only in a few cases attempts were made to elucidate general petromagnetic characteristics of the lithosphere.

As is evident from numerous publications on the origin of magnetic anomalies and on the magnetism of exposed rocks, the majority of magnetic rocks are represented by volcanics and near-surface intrusives. The petromagnetic study of these rocks provides constraints on the origin of their magnetization which should also be valid for deep-seated bodies.

A petromagnetic model of the oceanic lithosphere has been constructed on the basis of the hypothesis of *Vine and Metthews* [1963] and synthesis of petromagnetic data on rocks composing the lithosphere under present and extinct oceans [*Dunlop and Prevo,* 1982; *Gordin et al.,* 1993; *Johnson,* 1979; *Kent et al.,* 1978; *Kidd,* 1977; *Pechersky,* 1994; *Pechersky and Didenko,* 1995; *Pechersky et al.,* 1994; and others]. A basic feature of the model is the primary magmatic formation of the oceanic crust; the model is stratified as follows: an upper magnetic layer including lavas (2A), parallel dikes that are feeding lava channels (2B), and gabbro (3A); a lower nonmagnetic layer of accumulative gabbro and pyroxenites (3B); and a nonmagnetic upper mantle. The latter is underlain by a secondary magnetic layer formed due to the serpentinization of the uppermost mantle peridotites. The magnetic polarity distribution in this layer is shown to be rather chaotic [*Nguen and Pechersky,* 1989]. This scheme does not account for the primary heterogeneity of layers 2 and 3A (their magnetization varies from early nonmagnetic and weakly magnetic generations of dikes and lavas to strongly magnetic late differentiates [*Pechersky and Didenko,* 1995]) and for the role of secondary nonmagnetic rocks related to greenstone alterations etc.

A much more complicated problem is the distribution of magnetic masses in the continental crust. General geological evidence, data on real deep rocks and experimental data indicate that magnetic anomalies on various scales are mainly produced by magnetite-bearing igneous surface and near-surface rocks subsequently descended to great depths. Thus, metasedimentary rocks are present in most sections of Archean rocks attributed to the lowermost continental crust, i.e. a considerable part of these sequences formed at the Earth's surface. Therefore, regular features of the near-surface igneous rock magnetism should also be characteristic of magnetism of the lower continental crust. In this context, it is important to assess the effect of deep metamorphism on the magnetism of lower continental crust rocks.

Iron-rich fluids are the second probable source of enrichment of rocks with magnetic minerals.

A synthesis of petromagnetic, petrochemical and mineralogical evidence, as well as data of experiments reproducing the formation and alteration conditions of rocks, indicates that the diversity of the formation conditions of lithosphere rocks can be described in terms of four petromagnetic types (Table 1) [*Pechersky,* 1994].

2. Sources of Regional Magnetic Anomalies

Without ranking priorities, we want to specially mention the works of *Krutikhovskaya* [1976, 1986], who for many years was engaged in solving the problem of multidisciplinary interpretation of regional magnetic anomalies and studying petromagnetic properties to gain constraints on the deep structure of the continental lithosphere.

As seen from the maps of the anomalous magnetic field, the lithosphere is largely composed of nonmagnetic rocks, and the magnetic mass distribution is strongly inhomogeneous both laterally and vertically, which is reflected in the differentiation, intensity and morphology of the magnetic anomalies. The morphology of both local and regional anomalies is dominated by linear and isometric types. This typification persists at all hierarchical levels, from local anomalies to those derived from satellite data, and has a genetic (primarily tectonic) significance. Regional magnetic anomalies (a few tens of kilometers across) are usually attributed to sources at depths greater than 10–15 km.

We present an example demonstrating the correlation between magnetization estimated from the magnetic anomaly intensity and tectonic position of magmatic bodies on the territory of North Eurasia (Figure 1). Thus, zones of rift, islandarc and intraplate volcanism are dominated high-magnetization volcanics, virtually regardless of the rock age; even among acid rocks, more than 60% have a magnetization greater than 0.3 A/m. Fold zone volcanics of collision and folding epochs contain a large amount of even basic nonmagnetic rocks (more than 70% have a magnetization of less than 0.1 A/m). Marked magnetic and nonmagnetic "tails" in the first and second groups of volcanics, respectively, may be related to inaccurate delineation of volcanics occurrence zones and/or with deuteric alterations. A similar pattern is observed in intrusive rocks, but the group of nonmagnetic basic cumulates is most pronounced here (see below).

Due to the ambiguity of the magnetic data inversion, even a multidisciplinary interpretation of data does not remove inversion divergences: the lower boundary of regional magnetic anomaly sources ranges in various models from 15–20 to more than 40 km [*Belusso et al.,* 1990; *Bulina,* 1986; *Karataev and Pashkevich,* 1985, 1986; *Krutikhovskaya,* 1986; *Lugovenko et al.,* 1984; *Mayhew et al.,* 1985; *Pashkevich et al.,* 1986; *Pechersky,* 1991; *Pechersky et al.,* 1975; *Piskarev and Pavlenkin,* 1988; *Schlinger,* 1985; *Toft and Haggerty,* 1988; *Wagner,* 1984; *Warner and Wasilewski,* 1995; *Wasilewski and Mayhew,* 1982], reaching Moho or not exceeding the 580°C depth (Curie point of magnetite). Variations in both depth and shape of deep magnetized bodies responsible for regional magnetic anomalies are naturally related to specific features of the geological regional structure and evolution. Modeling results indicate that the regional field cannot be accounted for by Moho undulations with a uniform magnetization of the lower crust.

Table 1. Characteristic of petromagnetic rocks types of the lithosphere

Petromagnetic type	Ultramafic-mafic	Femic	Sialic-mafic	Sialic
Average magnetization, A/m	<0.5	1.0–10.0	1.0–5.0	<0.5
Oxidation conditions (buffer)	QMF and below; nearly close system	close to Ni-NiO, typical of basaltic magma; nearly close system	from Ni-NiO to MH; open system	close to QMF; nearly open system
Magnetic minerals	rare grains of ilmenite, titanomagnetite (TiO_2>20%), and secondary magnetite	titanomagnetite (TiO_2>20%) and ilmenite; secondary titanomagnetites (TiO_2<10%) and magnetite	titanomagnetites of various compositions and ilmenite; secondary titanomagnetites (TiO_2<10%) and magnetite	ilmenite; secondary titanomagnetites (TiO_2<10%) and magnetite
Rocks	pyroxenites, layered gabbro-pyroxenite complex, accumulative gabbros; other (rarer) basic rocks	basalts, ferrogabbros and other products of Fenner-type differentiation, rarer intermediate and acidic rocks	basalts, andesites, diorites, granitoids and other products of calc-alkaline (Bower-type) differentiation of magma	granitoids, rarer other rocks
Tectonic types	oceanic crustal layer 3B, crust of "dry" rift and sedimentary basins, greenstone belts	extension structures: oceanic rifts, "wet" rifts of continents, intraplate magmatism regions	active margins, island arcs, junction zones, tectonic-magmatic activation zones	compression structures, collisional and folding areas

Multidisciplinary studies [*Krutikhovskaya,* 1986; *Pechersky,* 1994] show that the belts of regional magnetic anomalies are mostly confined to tectonically and magmatically active suture zones separating the largest crustal blocks; magnetic mineral enrichment and depletion episodes took place at stages of extension (femic blocks) and compression (sialic blocks). Overall, regional magnetic anomalies are of a polygenetic and diachronous origin and are primarily associated with early consolidation zones composed of the oldest complexes of basic granulites and, occasionally, other metamorphic rocks [*Belusso et al.,* 1990; *Krutikhovskaya,* 1986; *Krutikhovskaya et al.,* 1984; *Liu,* 1998; *Liu and Gao,* 1992; *Liu et al.,* 1994; *Mayhew et al.,* 1985; *Pechersky,* 1994; *Wagner,* 1984; *Wasilewski and Mayhew,* 1982; *Wasilewski and Warner,* 1988; *Yakovlev and Markovskii,* 1987; and others]. Some authors emphasize that amphibolization leads to the enrichment of rocks with magnetite [*Ermakov and Pechersky,* 1989; *Genshaft et al.,* 1985; *Krutikhovskaya,* 1986; *Lutz,* 1974; *Williams et al.,* 1986; *Yakovlev and Markovskii,* 1987], whereas others note an opposite effect, namely, a sharp drop in rock magnetization associated with the granulite-to-amphibolite facies transition [*Afanas'ev,* 1978; *Golovin and Petrov,* 1984; *Pashkevich et al.,* 1986; *Schlinger,* 1985; *Wasilewski and Warner,* 1988]. There are known examples when acidic rocks from crustal sections were magnetic, and basic rocks, nonmagnetic [*Liu and Gao,* 1996; *Pilkington and Percival,* 1999; *Williams et al.,* 1985]. Deep rocks (xenoliths) of Mongolia and Central Asia range in magnetization from nonmagnetic mantle hyperbasites and weakly magnetic pyroxenites to magnetic intermediate granulites [*Genshaft and Pechersky,* 1986; *Lykov et al.,* 1981; *Pechersky,* 1991, 1994]. Anomalously high magnetite concentrations (to 10% and more) are common in high activity zones such as the Lesser Caucasus, Kamchatka and Ivrea [*Belusso et al.,* 1990; *Ermakov and Pechersky,* 1989; *Genshaft and Pechersky,* 1986; *Genshaft et al.,* 1985; *Lykov and Pechersky,* 1984; *Wasilewski and Warner,* 1988; and others]. Moreover, xenoliths in many regions include highly magnetic pyroxenites of the "black series", which are rocks of the lowermost crust and uppermost mantle, characterized by signatures of polymetamorphism and partial melting [*Genshaft and Pechersky,* 1986; *Genshaft and Saltykovsky,* 1987; *Mayhew et al.,* 1985;

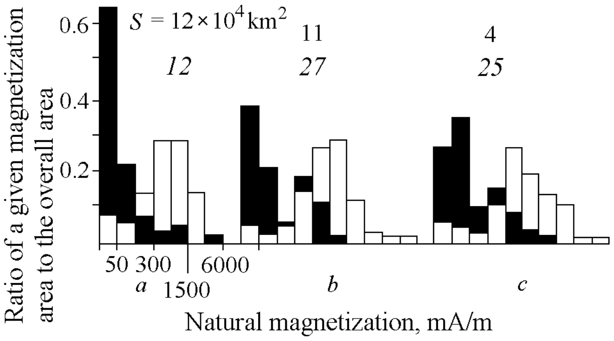

Figure 1. Histograms of natural magnetization in volcanics of various ages and compositions in North Eurasia (former USSR). Solid columns refer to times of folding and collisions, and open columns refer to times of the rigid crust magmatism associated with extension. S is the overall area of volcanics occurrence, involved in a given histogram (the upper and lower numbers characterize the solid- and open-column histograms respectively). The area-averaged magnetization is estimated from magnetic anomalies (since the sizes of magma bodies substantially exceed the magnetic survey height, the magnetization was calculated from the formula $J = \nabla T_{\mathrm{a}}/4\pi$ for a half-space).

Pechersky, 1994; *Saltykovsky and Genshaft,* 1985; *Semenova et al.,* 1984; *Wasilewski and Mayhew,* 1982]; however, these anomalously high magnetizations are not sources of regional magnetic anomalies, as is evident from the absence of the latter in the Lesser Caucasus, Kamchatka, Kurile and other regions. Less than 10% samples studied represent magnetic black pyroxenite xenoliths. This local enrichment in magnetic minerals is related to magmas entrapping the xenoliths.

Since multidomain grains of magnetite prevailing in the deep continental crust are mostly in an equilibrium state, the deep rock magnetism is mainly controlled by the magnetite concentration and induced magnetization irrespective of the $P-T$ conditions up to temperatures of 550–580°C (Curie point of magnetite) [*Markovskii and Tarashchan,* 1987; *Zavoiskii and Markovskii,* 1983]. However, single-domain and pseudosingle-domain grains can make a certain contribution to the magnetization of deep rocks due to their stress state, multiphase alteration of ilmenite and titanomagnetite, and decomposition of pyroxenes with the formation related minerals; therefore, magnetic anomalies may be in part associated with the remanent magnetization, as is the case with Proterozoic anorthosites of Lithuania, Ukraine and Norway [*Bogatikov et al.,* 1975; *McEnroe et al.,* 1996], granulites of central Australia [*Kelso et al.,* 1993] and Labrador [*Kletetschka and Stout,* 1998], and others. However, the uniformity of ancient natural remanent magnetization (NRM) directions (if $Q_{\mathrm{n}} > 1$, i.e. if the remanent magnetization dominates the induced one) is improbable with magnetic bodies of 10–20 km in thickness and about 100 km in lateral size, their slow and irregular cooling, and long-term complex metamorphism against the background of varying geomagnetic polarity; in view of these factors, any noticeable contribution of remanence to regional magnetic anomalies is hardly to be expected. Moreover, the relative contribution of induced versus remanent magnetization increases in the lowermost high-temperature crust, characterized by conditions favorable for the formation of the recent high-temperature viscous remanent magnetization (e.g. see [*Schlinger,* 1985; *Williams et al.,* 1986]).

Airborne and satellite magnetic survey data gave lower crust average magnetizations of 5 A/m in central Canada [*Hall,* 1974], 2 A/m in northwestern Germany [*Hahn et al.,* 1976], 2–4 A/m in the Ukrainian Shield [*Krutikhovskaya and Pashkevich,* 1979] and -3.5 ± 1 A/m in the United States [*Schnetzler,* 1985]. These values are consistent with data of direct measurements of the deep rock magnetization (see below).

All researchers note a significant role of granitization often decreasing the rock magnetization.

The magnetization of all types of primary igneous rocks and sedimentary-volcanic sequences drastically decreases beyond the regional magnetic anomaly zones in areas of regional high-grade metamorphism, which has been established for the Baltic Shield [*Golovin and Petrov,* 1984; *Schlinger,* 1985] including the Kola overdeep drillhole section [*Brodskaya et al.,* 1992; *Kozlovskii,* 1984] and for the Canada Shield [*Pilkington and Percival,* 1999; *Williams et al.,* 1986].

Upper mantle rocks are nonmagnetic in all regions, with and without regional magnetic anomalies.

Thus, the comparison of regional magnetic anomalies with geological settings and deep rock magnetization suggests that their sources are located within the crust and are mostly represented by basic granulites. The above evidence provides no keys to the problem of accumulation of magnetic minerals in the crust, and its solution requires mineralogical and petrological information.

3. Experimental Data

The formation of magnetic minerals in the crystallization medium requires, in addition to necessary iron, the presence of Ti, Mg and other cations of the Earth's most widespread magnetic minerals: magnetite, titanomagnetites, hemoilmenites and pyrrhotite. As is evident from statistical data, formation of magnetic minerals requires more than 1% Fe present in rock [*Pechersky et al.,* 1975]. This is a necessary, but not sufficient, condition because there are known numerous examples when rocks of similar composition and with similar Fe concentrations range in the concentration of magnetic minerals from <0.01% to 5% and more.

The formation and composition of magnetic minerals are controlled by the general pressure P, temperature T, oxygen fugacity fO_2, pH value and other, less significant parameters. According to data of experiments with basalt systems of normal Fe concentration (e.g. see [*Lykov and Pechersky,* 1976, 1977; *Ringwood,* 1975]), titanomagnetites crystallize at $T<1100$°C and $P<13$ kbar. With increasing pressure, titanomagnetites are first replaced by weakly magnetic Mg-Al-ferrospinel (Figure 2) and then by garnet. An increase in

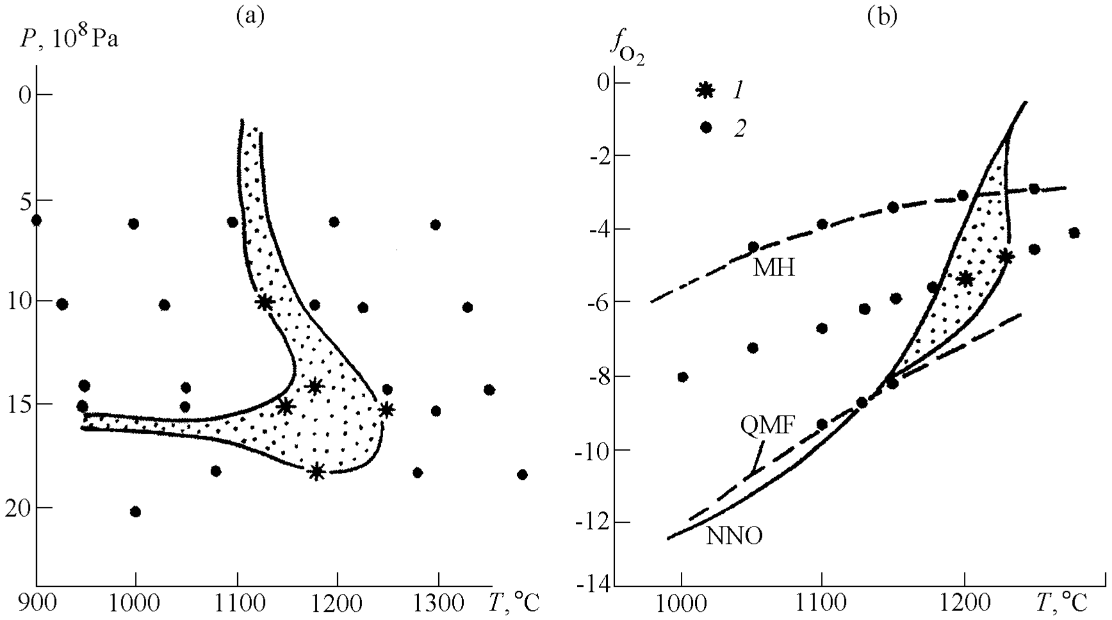

Figure 2. (a) $T-P$ and (b) $T-fO_2$ fields of magnetic mineral (titanomagnetite) formation in basalts (after [*Lykov and Pechersky,* 1976, 1977]). Broken lines delineate the crystallization regions of weakly magnetic Mg-Al ferrospinels (also shown by the symbols 1); 2 – bold dots are the points of titanomagnetite crystallization.

the concentration of alkaline elements in basalts leads to earlier crystallization of ore phases and increases the Ti concentration in titanomagnetites and the Mg+Al concentration in Mg-Al-ferrospinels [*Pechersky,* 1994; *Pechersky et al.,* 1975]. The direct effect of pressure on the ratio Fe^{3+}/Fe^{2+} in melts was confirmed experimentally: the ratio decreases with increasing P [*Borisov et al.,* 1991]; accordingly, the Ti concentration in crystallizing titanomagnetite increases [*Genshaft and Sattarov,* 1983; *Osborn et al.,* 1979].

Four following thermodynamic zones of the magnetic mineral formation have been recognized within the $P-T-fO_2$ region [*Pechersky,* 1985; *Pechersky et al.,* 1975].

- *Hematite* zone: strongly oxidizing conditions at the Earth's surface giving rise to minerals containing solely Fe^{3+} (hematite, maghemite, iron hydroxides and Fe^{3+}-silicates).

- *Magnetite* zone: weakly oxidizing conditions giving rise to minerals containing Fe^{2+} and Fe^{3+} (titanomagnetites and other ferrospinels, hemoilmenites).

- *Silicate* zone: relatively reducing conditions with virtually absent Fe^{3+}, giving rise to ilmenite, ulvospinel, hercynite and other Fe^{2+} ferrospinels, pyrrhotite, pyrite, and Fe^{2+} silicates.

- *Metal-Fe* zone: strongly reducing conditions: free metal iron appears in addition to minerals of the silicate zone. In lithosphere these cases are exotic; apparently, the metal-Fe zone is located at the mantle base and in the core. This zone is typical of lunar rocks and meteorites. The boundaries between these zones roughly coincide with the hematite-magnetite (HM), quartz-magnetite-fayalite (QMF) and iron-fayalite (IF) buffers.

The admixture of silicon dramatically complicates the simple formation scheme of magnetic minerals in the Fe-Ti-O system. In this case, in addition to the standard thermodynamic factors, the bond strength (the covalent bond in silicates is stronger than the ionic bond in Fe-Ti oxides) and the related solubility play a significant role in this case. The Fe partition coefficient is largest during the solid-to-fluid and fluid-to-melt transitions and is more than 10 times smaller in the opposite direction [*Kadik et al.,* 1990]. Therefore, the Fe enrichment process is most intense in a melt, and iron is transported mainly by the melt and, to a lesser degree, by the fluid. Then, other conditions being equal, the formation of magnetic minerals from melts is most preferable [*Kadik et al.,* 1990; *Marakushev and Bezmen,* 1983; *Mueller and Saxena,* 1977; *Pechersky et al.,* 1975]. Iron easily enters a low-pH fluid and is transported by it. Only the fluids enriched in Fe are potential sources of higher crystallization of magnetic minerals, which takes place as soon as such a fluid enters a relatively reductive ($pH > 7$) environment, which was demonstrated experimentally [*Gantimurov,* 1982; *Kadik et al.,* 1990; *Korzhinskii,* 1967; *Letnikov et al.,* 1977; and others].

Crystallization of high-Ti ferrospinels under $P-T$ conditions of the silicate zone was observed at high pressures. Thus, titanomagnetite segregations in kimberlite were discovered at 55 kbar [*Girnis et al.,* 1995]. The titanomagnetite formation is related to the presence of water in the system

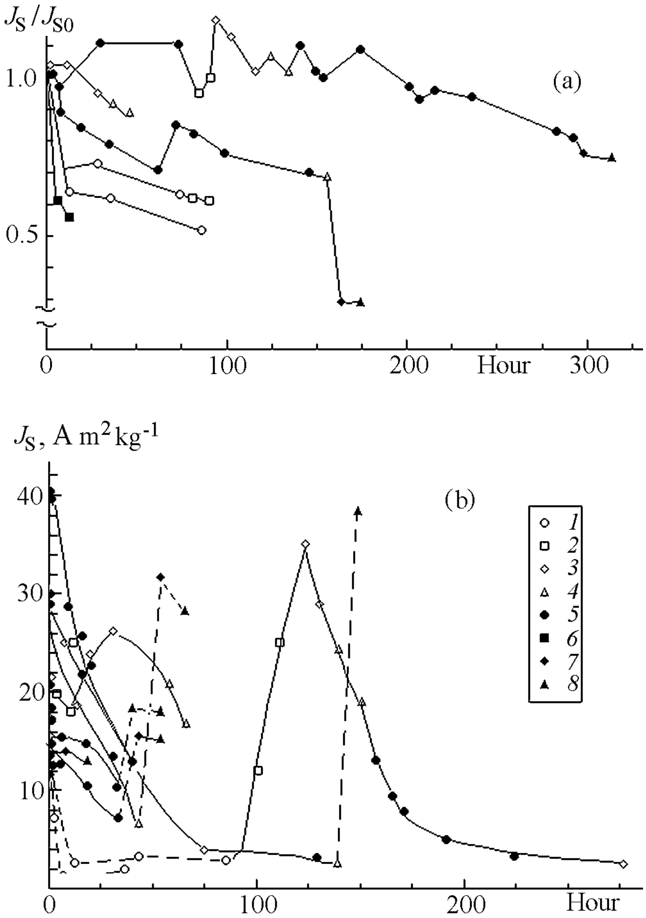

Figure 3. Variations in the saturation magnetization J_s during heating of gabbro and pyroxene samples: (a) nonmagnetic gabbro; (b) magnetic ferrogabbro. (1) – 800°C, CO_2+20%CO; (2) – 800°C, CO_2; (3) – 950°C, CO_2; (4) – 950°C, CO_2+5%CO; (5) – 950°C, H_2O; (6) – 950°C, H_2O+3%CO; (7) – 950°C, H_2O+3%NH_4OH; (8) – 950°C, H_2O+1%NH_4OH.

and to a higher oxygen fugacity. High-Ti ferrospinel crystallized from a melt and replaced ilmenite relicts at pressures of 15 to 50 kbar in mixtures of peridotite and basalts with ilmenite and metal iron [*Genshaft et al.,* 2000]. Silicate and carbonate (calcite) melts mixed in these experiments. The spinel composition corresponds to the solid solution of ulvospinel and magnesioferrite (or chlorospinel). Such spinel is not a ferromagnet but can give rise to secondary ferromagnets that form, replacing it, under lower-temperature and higher-oxidation conditions. The reality of natural processes involved in the formation conditions of this type of high-Ti ferrospinel in the upper mantle is corroborated by the existence, under mantle conditions, of silicate melts enriched in Fe and Ti with a composition similar to the melts experimentally studied [*Bell et al.,* 1997; *Gibson et al.,* 2000; *Girnis,* 1998; *Grachev,* 2000; *Green and Wallace,* 1988; *Hauri et al.,* 1993; *Schiano et al.,* 1994; and others].

In order to assess the influence of deuteric alterations on the petromagnetic information, *Pechersky et al.* [1989] conducted three series of experiments with the thermal treatment of samples of natural and synthetic rocks under "dry" and "wet" conditions.

Pressure effect. Oceanic basalt samples were held in sealed ampoules for 0.5 to 24 hours at 1000°C under a pressure of 1.5 to 7 kbar. According to the Lindsley thermobarometer [*Spencer and Lindsley,* 1981], the oxygen fugacity is higher than the Ni-NiO buffer. During thermal treatment, titanomagnetite experiences multiphase oxidation with the formation of ilmenite lamellae and low-Ti titanomagnetite cells between them. With increasing pressure and/or thermal treatment time, the titanomagnetite in cells decreases its Ti concentration up to the appearance of magnetite. Moreover, some iron escapes from titanomagnetite grains, which is evident from an increase in the mean composition of grains from $x = 0.615$ to $x = 0.65$–0.695 and from a decrease in the bulk concentration of titanomagnetite by 20–25% (judging by the values of J_s and T_c). Pressure and time strengthen the effect of alteration of titanomagnetite grains and removal of iron from them. A pressure increase noticeably raises the magnetic hardness (J_{rs}/J_s rises from 0.04–0.08 to 0.15–0.21 and H_{cr}, from 5–10 to 20–25 mT), which is related to an increase in the stress state of titanomagnetite grains due to the increasing number of their structural defects.

Role of diffusion in the titanomagnetite alteration process. A mixture of synthetic titanomagnetite and natural pure olivine was used. The samples were heated to 800°C, 1000°C and 1150°C in vacuum. This relatively simple system allows one to trace the process of the fluid-free reworking of titanomagnetite grains. During the thermal treatment of the mixtures, the titanomagnetite composition changed from $x = 0.1$–0.4 to $x = 0.66$–0.9 and ilmenite grains appeared. At the contacts of large grains of olivine and titanomagnetite, iron diffuses from titanomagnetite into olivine, and magnesium, from olivine into titanomagnetite, which is reflected by an increase in the concentrations of MgO to 8% and TiO_2 to 30% and by 2 to 27% Fe enrichment and 5 to 7% MgO depletion in olivine grains. *The introduction of additional iron into olivine does not result in the formation of magnetic minerals in the rock.*

Effect of the oxygen fugacity and fluid composition on the composition and concentration of magnetic minerals. The experiments were conducted on three types of samples chosen in such a way as to allow one to observe, in the process of thermal treatment, the diffusion-induced appearance of newly formed minerals (large crystals of pyroxene), transport of material by the fluid (nonmagnetic porous pyroxene gabbro sample) and transformations of original magnetic minerals (ferrogabbro samples containing up to 40% decomposed titanomagnetite and magnetite). The fluid composition and oxygen fugacity were adjusted by the blowing of gas mixes of various compositions through the kiln (Figure 3). The temperatures of the experiments were 800°C and 950°C.

Thermal treatment of the pyroxene crystal during 200 hours did not change the concentration (J_s), composition (T_c) and structural state (J_{rs}/J_s) of magnetic minerals regardless of gas environment fO_2 and pH. The gas easily penetrates into the nonmagnetic porous gabbro sample, and some iron is removed from the sample in thermal treatment process, which noticeably decreases J_s (Figure 3a).

A similar result was obtained from an experimental study of bimetasomatic processes in the nonmagnetic granodiorite-limestone/dolomite system [*Zaraiskii et al.,* 1986]. As was observed in all variants of the experiments widely differing in temperature (400 to 900°C), pressure (0.7 to 5 kbar) and composition, pH (1 to 13) and fO_2 of the fluid, magnetite and other magnetic minerals form not *in situ* but beyond granodiorite, where pH sharply increases. In these experiments the 600°C 0.1-kbar diffusion coefficient was 3.2×10^{-4} cm^2/s, which is about 10^{14} faster than the diffusion of iron in titanomagnetite during its multiphase oxidation [*Pechersky et al.,* 1975; *Petersen,* 1970].

In the experiments with the highly magnetic ferrogabbro samples under dry, relatively reducing conditions (CO_2+CO) (Figure 3b, runs 1, 2 and 4), the initial decomposed titanomagnetite first homogenizes, ilmenite lamellae gradually dissolve and J_s drops. After 12 hours titanomagnetite is nearly homogeneous. After 90 hours the reduction process is still active; the microprobe analysis, as well as T_c = 760–770°C and a J_s increase (Figure 3b), indicate the appearance of high-Ti titanomagnetite with T_c = 100°C and metal iron in the form of inclusions of less than 1 μm in size within the titanomagnetite grains. The thermal treatment in CO_2 (dry oxidation conditions above the Ni-NiO buffer) again leads to the heterogeneous oxidation of titanomagnetite; accordingly, J_s increases (Figure 3b) and a phase with $T_c \geq 560$°C appears. Longer duration of the thermal treatment in CO_2 results in a J_s decrease, largely due to partial removal of iron from titanomagnetite grains.

Wet conditions enhance the process of the titanomagnetite destruction (Figure 3b). Both diffusive withdrawal of iron and intense corrosion of titanomagnetite grains (with noticeable participation of silicates) are in progress. The microprobe data indicate that the average composition of the silicates remain nearly unchanged. The removed iron "settles" within the corroded grains. Introduction of 3%NH_4OH into the water vapor creates conditions reductive for titanomagnetite and leads to the appearance of metal iron accompanied by an abrupt increase in J_s (Figure 3b). The metal iron represented by small grains and dendrites is located within the contours of large grains of titanomagnetite and magnetite, i.e. most iron moves insignificantly.

Generalization of the Experimental Results

(1) *Rock-forming silicates produce no newly formed magnetic minerals* at high temperatures both under dry conditions and with the participation of water vapor. New magnetic minerals form during the *in situ* recrystallization of Fe-Ti ores under new $T-fO_2$ conditions. (2) Under high temperatures and under both dry and water vapor conditions, magnetite and titanomagnetite are destroyed in two ways: (a) diffusive removal of iron from grains; (b) fluid corrosion of grains. *No variant of the experiments showed an increase in the total concentration of magnetic minerals in the sample*, and the original nonmagnetic minerals remained nonmagnetic. (3) The process of corrosion and destruction of titanomagnetite and magnetite is independent of the oxygen fugacity within its wide variation range. (4) During the destruction of titanomagnetite and magnetite most iron remains within grain boundaries. Iron bound in silicates is nearly immobile and is not affected by the fluid. (5) The high pressure only accelerates the destruction process of titanomagnetite and magnetite but does not change its essential features. (6) Dry experiments model, to an extent, conditions of the granulite metamorphism which is nearly isochemical and proceeds with insignificant participation of fluids and accordingly with inert behavior of most elements including iron [*Lutz,* 1974; *Mueller and Saxena,* 1977; *Perchuk,* 1973; *Yakovlev and Markovskii,* 1987; and others]. Consequently, the granulite metamorphism cannot release any substantial amount of iron from silicates to form, at its expense, new magnetic minerals.

4. Petromagnetic Characteristic of Near-Surface Igneous Rocks

Most of the studied sections of Archean rocks contain former sedimentary rocks, i.e. considerable portions of rocks composing the lower continental crust formed at the surface. Moreover, spreading extension structures were directly involved in the formation of the basaltic crust of the primary ocean [*Marakushev,* 1992; *York,* 1993]. For example, such Archean complexes as the gray gneisses and paragneisses and greenstone belts are metamorphosed volcanic-sedimentary sequences, dikes, sills and layered gabbro-pyroxenite complexes; i.e. they represent an assemblage close to ophiolites and are regarded as sections of the paleo-oceanic crust [*Condie,* 1983; *Taylor and McLennan,* 1988; *Zonenshain et al.,* 1990]. In the post-Archean time the crust was accreted from above due to later magmatism and sedimentation, collisions, block overthrusts and similar processes; newly formed foldbelts experienced orogenic uplifts accompanied by deep-seated granitic magmatism [*Kropotkin et al.,* 1987; *Marakushev,* 1992; *Taylor and McLennan,* 1988; *Zonenshain et al.,* 1990]. Accordingly, we address the magnetism of near-surface, primarily oceanic, igneous rocks as a key to understanding the magnetism of the lower continental crust. Also, we assess the influence of deep metamorphism on the rock magnetism of the latter.

Numerous data indicate that the differentiation of the basaltic magma formed and accumulated under spreading centers is of key importance for the formation of the oceanic crust. The magma differentiation results in the division of rocks into early nonmagnetic cumulates and magnetic products of the crystallization differentiation. The differentiation degree of a melt controls the amount of iron in the melt and thereby the amount of titanomagnetite, the main carrier of the crustal magnetism. This process is clearly demonstrated with the example of intrusive gabbros of Iceland, Cape Verde, Kamchatka, Kuriles, Lesser Caucasus, Afar and Patynski intrusive [*Bogatikov et al.,* 1971; *Ermakov*

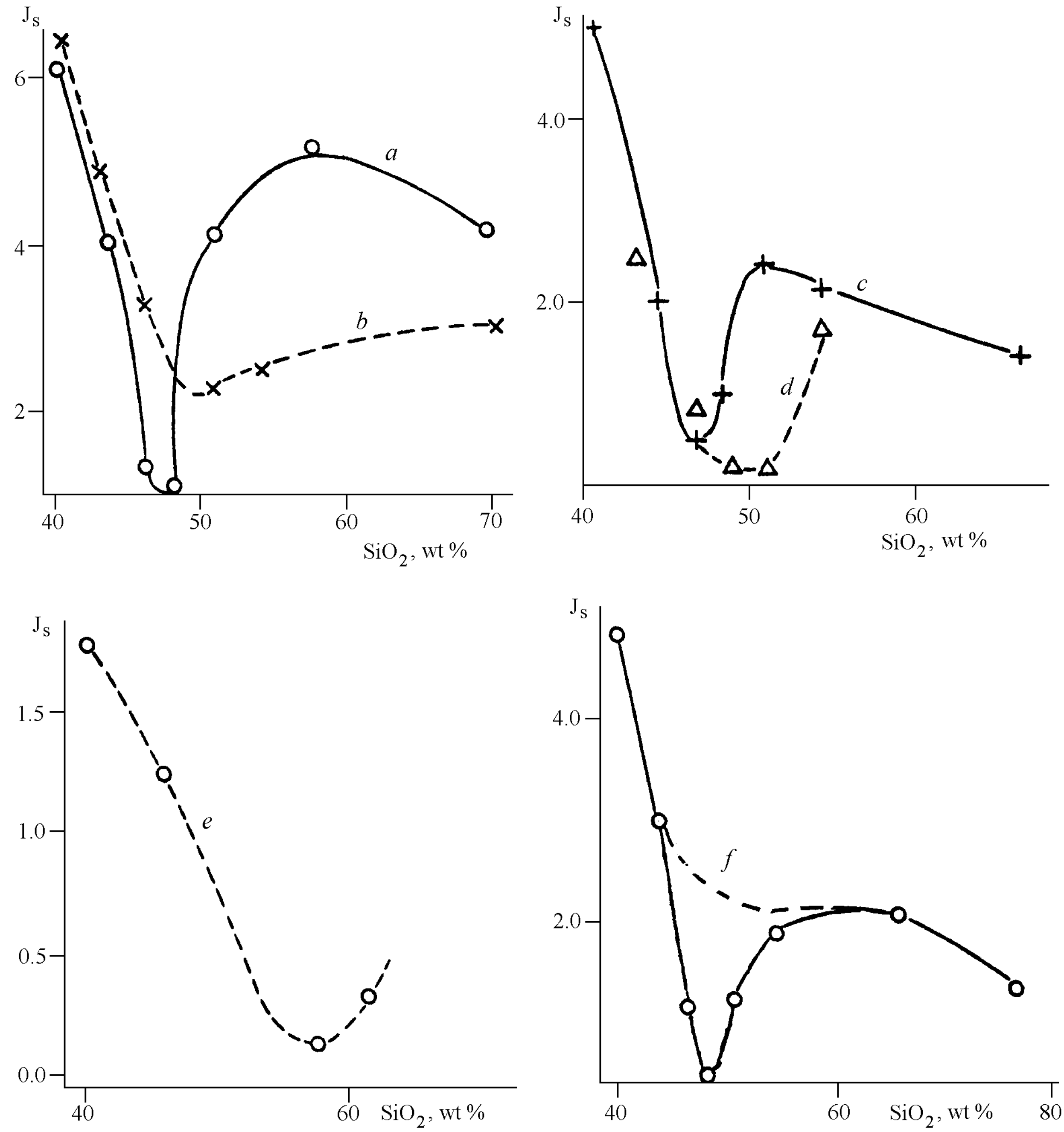

Figure 4. Saturation magnetization J_s averaged over SiO_2 intervals. Minimums in the curves near 50%SiO_2 is related to accumulative rock series. (a) Iceland; (b) Kurile Islands; (c) Lesser Caucasus; (d) Kamchatka Peninsula; (e) Cape Verde fault; (f) all data.

and Pechersky, 1989; *Genshaft et al.,* 1985; *Kashintsev and Pechersky,* 1983; *Lykov et al.,* 1994; *Pechersky,* 1994; *Pechersky and Didenko,* 1995; *Zolotarev et al.,* 1988]. In accordance with both petrochemical characteristics and magnetic mineral concentration (J_s), all rocks form two groups corresponding to two magmatic trends: accumulative trend and magma differentiation (Figures 4 to 7), resulting in the formation of nonmagnetic and magnetic rock groups. The process proceeds under conditions of a nearly anoxic system, which increases the ferruginosity of the melt and the concentration of magnetic minerals in late cumulates and particularly in residual melts. Even in relatively shallow sources the basaltic magma preserves small values of the oxygen fugacity which are lower than the QMF buffer by at least 1–2 orders of magnitude [*Kadik et al.,* 1990; *Sato and Valenza,* 1980]. High-magnesium and calcium minerals crystallizing

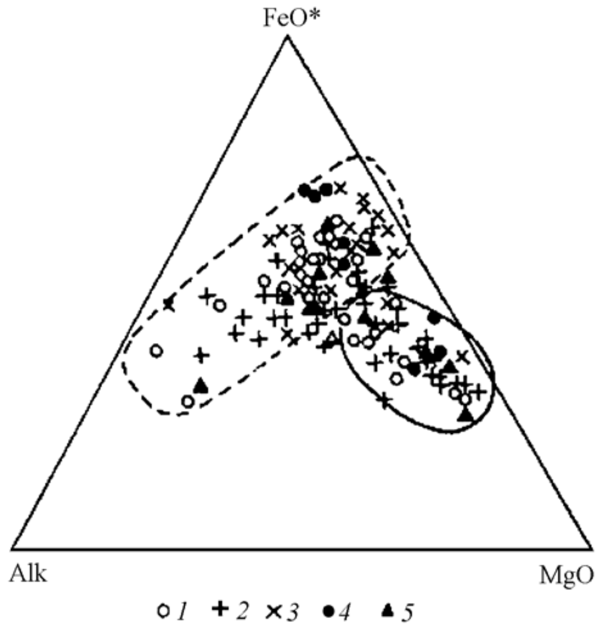

Figure 5. The gabbro AFM diagram (see the text). The broken line delineates rocks associated with the differentiation trend, and the solid line delineated the region of accumulative rocks. (1) – Iceland; (2) – Lesser Caucasus; (3) – Kurile Islands; (4) – Cape Verde Islands; (5) – Kamchatka.

from magma under these conditions are olivine, plagioclase and pyroxene (± chromite), which form nonmagnetic cumulates. Detailed studies of intrusive massifs and inclusions in extrusive bodies in Iceland also indicate prevailing occurrence of accumulative and heteroaccumulative rocks [*Genshaft and Saltykovsky,* 1999]. The chemistry of accumulative rocks is characterized by a narrow range of variations in the SiO_2 concentration (46–48 wt %) and bulk iron (FeO^* 5–10 wt %) with large variations in the MgO concentration (up to 20 wt %) (Figures 5 to 7). The products of residual melt crystallization (magma trend of differentiation) are distinguished by higher concentrations of TiO_2 and FeO^* (Figures 5 to 7), the presence of modal titanomagnetite and hemoilmenite, and a high magnetization.

Strong degree of the melt differentiation can lead to the appearance of primarily nonmagnetic dikes and lavas. Thus, the study of the Shuldak section of parallel dikes [*Pechersky and Didenko,* 1995; *Pechersky et al.,* 1983] showed that the older dikes developed in spreading minicenters are mostly weakly magnetic and nonmagnetic, whereas the younger dikes are most magnetic. Another example is given by lavas and dikes of the Alai Range [*Pechersky and Didenko,* 1995; *Pechersky and Tikhonov,* 1988]. Here the following two stages differing in time are recognizable: (a) formation of primarily nonmagnetic parallel dike complex and lavas and (b) a complex of dispersed dikes, cutting the rocks of the first stage, and erupted magnetic pillow lavas. Thick dikes occur which have primarily nonmagnetic central parts and primarily magnetic peripheral parts; according to petrochemical characteristics, their central parts are similar to the first stage dikes, and their peripheral parts, to the second stage dikes.

The crystallization differentiation of magmas at various depths of intermediate chambers (formation of rocks of accumulative and magmatic types) also plays an important role in the development of island-arc structures [*Kadik et al.,* 1990].

The differentiation in Archean complexes should largely biased toward primarily nonmagnetic rocks due to more reductive magma conditions resulting in the preferable crystallization of ilmenite in igneous rocks (see section 5).

Thus, the concentration of magnetic minerals as a source of regional magnetic anomalies (i.e. "magnetism-nonmagnetism" of rocks) is primarily controlled by the magmatic stage. The division of igneous rocks into magnetic and nonmagnetic types is relevant to both basic and acidic varieties and is largely controlled by the tectonic factor: extension zones are dominated by magnetic rocks, and compression zones, by nonmagnetic rocks (Figure 1).

Now, we address the effect of deuteric alterations on the magnetism of magma formations.

The main magnetization carriers in igneous rocks are titanomagnetites which are unstable under surface conditions and their multiphase oxidation often starts at the cooling stage of magma products resulting in the formation of an aggregate of ilmenite and magnetite. Slow multiphase titanomagnetite oxidation occurs even at relatively low temperatures of the Earth's surface [*Nguen and Pechersky,* 1982]; thus, the average relative amount of magnetite amounts to 30–50% in young subaerial basalts, increases in older basalts, and is close to 100% at a 200-Ma age; i.e. nearly 50% magnetite in old subaerial basalts formed as a result of low-temperature multiphase alterations of titanomagnetite (Figure 8). Thus, the main regular feature largely persists at this stage: magnetic/nonmagnetic igneous rocks remain magnetic/nonmagnetic.

The role of such deuteric alterations of rocks as serpentinization, amphibolization, chloritization etc. cannot be assessed unambiguously (see section 2). It is established that the scales of secondary processes in rocks of the magmatic differentiation trend are much larger compared to cumulates.

Magnetic minerals in altered rocks are often secondary, being formed as a result of solid-state reactions; thus, ore grains in altered rocks usually do not coincide in composition with the primary magmatic titanomagnetite and are corroded and impregnated with silicates; their rounded smooth amoeba-like shapes are evidence for the formation during solid-state reactions [*Ermakov and Pechersky,* 1989; *Genshaft et al.,* 1985; *Lykov et al.,* 1994; *Zolotarev et al.,* 1988]. Gabbros of Iceland, South Mugodzhary etc. preserved decomposed grains of primary titanomagnetites similar in average composition to titanomagnetites of rift basalts ($x_{\mathrm{av}} \geq 0.65$). Secondary alteration features in primary titanomagnetites were discovered, in particular, in gabbros of Iceland, Caucasus, Kurile Islands, South Mugodzhary, Alai Range etc. [*Ermakov and Pechersky,* 1989; *Lykov et al.,* 1992; *Pechersky and Didenko,* 1995; *Pechersky and Tikhonov,* 1988; *Pechersky et al.,* 1983, 1994; *Zolotarev et al.,* 1988]. For example, dike diabases of the parallel complex near the contact with a gabbro body (South Mugodzhary) experienced

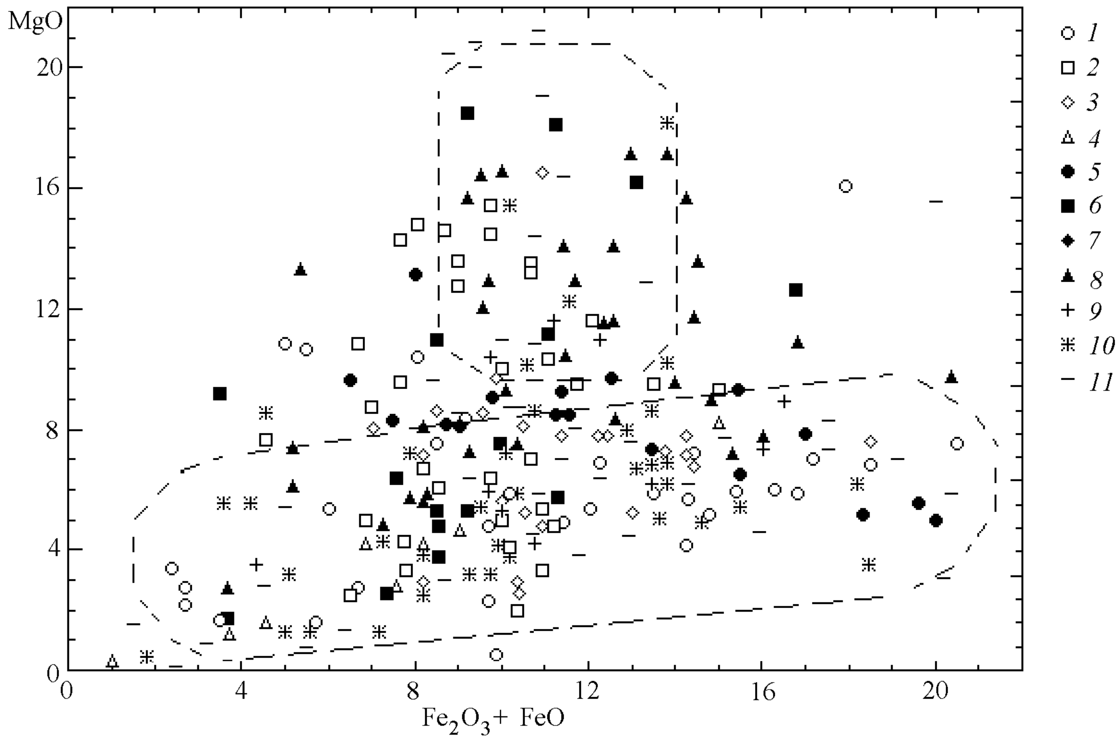

Figure 6. Relationship between MgO and FeO+Fe_2O_3 in rocks from various regions: (1) – Iceland; (2) – Lesser Caucasus; (3) – Kurile Islands; (4) – Tien Shan; (5) – Cape Verde fault; (6) – Kamchatka; (7) – Trans-Baikal region; (8) – xenoliths from Yakutia kimberlite pipes; (9) – magnetic xenoliths of Yakutia ($J_s > 1$ A m^2/kg); (10) – Aldan shield; (11) – Voronezh crystalline massif. The broken lines delineate tentative areas of the accumulative trend (vertical area) and differentiation trend (horizontal area).

metasomatic reworking, as is evident from their intense amphibolization accompanied by silicate-induced corrosion and decomposition of high-Ti titanomagnetite grains and formation of new grains of secondary low-Ti titanomagnetite ($x_{av} < 0.3$) having a fresh habit [*Pechersky and Didenko,* 1995; *Pechersky et al.,* 1987]. The process of amphibolization and intense Fe enrichment in the gabbroids composing most xenoliths in young lavas of the Kurile Islands is accompanied by enrichment in secondary low-Ti titanomagnetite [*Ermakov and Pechersky,* 1989]; this process preserves primary magmatic patterns of petrochemical characteristics and magnetization (Figures 4 to 6), i.e. the enrichment in titanomagnetite is not a direct result of the gabbro amphibolization with which the destruction and reworking of primary magmatic magnetic minerals are associated. Temperatures estimated from titanomagnetite-ilmenite intergrowths in amphibolized and other diabases and gabbroids of the Alai Range, Lesser Caucasus, South Mugodzhary, Kamchatka, Simushir Island, and Iceland, gneisses of Yakutia, Ivrea granulites, "black" pyroxenites of Mongolia and so on vary from 110 to 450°C [*Ermakov and Pechersky,* 1989; *Genshaft et al.,* 1985, 1995; *Lykov et al.,* 1994; *Pechersky,* 1991; *Pechersky and Tikhonov,* 1988; *Pechersky et al.,* 1983; *Wasilewski and Warner,* 1988] (also see section 5). The solid-state crystallization of magnetic minerals in the most of the above examples is corroborated by a stronger magnetic anisotropy. Rare titanomagnetite grains similar in average composition and habit to primary magmatic varieties are found in rocks without traces of deuteric alterations; the crystallization temperature of intergrowths of such titanomagnetite and ilmenite agrees with a melt existence range of 1100 to 1400°C [*Genshaft and Pechersky,* 1986; *Lykov et al.,* 1994].

Comparison of the aforementioned facts suggests that magnetic minerals crystallizing due to secondary processes replace ores existing in rocks, but these processes do not change the primary magnetic/nonmagnetic nature of the rocks. As is evident from the examples of Iceland and the Lesser Caucasus [*Genshaft et al.,* 1985; *Lykov et al.,* 1994; *Pechersky,* 1994], magnetization in the rocks studied is virtually independent of the amphibole concentration in rocks, and J_s noticeably drops with the increasing chlorite concentration.

A magnetization decrease in rocks associated with their amphibolization and chloritization may be due to the concurrent process of corrosion of ore grains by silicates [*Brodskaya et al.,* 1992; *Kozlovskii,* 1984; *Olesen et al.,* 1991; *Schlinger and Veblen,* 1989; and others]. The appearance of secondary magnetite due to deuteric hydrothermal alterations in ultrabasic rocks and allivalites was examined in detail in a case study of the Eastern layered intrusion, Isle Rum, Scotland [*Housden et al.,* 1996]. Both peridotites and allivalites are nonmagnetic and have a maximum susceptibility of 10^{-5} SI units and $J_s < 0.3$ A m^2/kg. Very fine magnetite grains fixed solely by a Curie point of about 575°C appear in peridotites due to oxidation processes in a 500–800°C interval, the magnetite concentration reaching 0.3%. The fluid penetration and formation of magnetite along with typically hydrothermal minerals such as amphibole, biotite etc. occur along small cracks as a result of brittle deformations of the peridotites. The secondary magnetite concentration is markedly lower in plagioclase-rich allivalites because they remain quasi-ductile, i.e. impermeable for the fluid. The magnetite formation at the expense of the fluid-transported iron, rather than olivine, is supported by the absence of a correlation between the ferruginosity of olivine and magnetite concentration in peridotites.

Such are the regular patterns of the formation and transformation of magnetic minerals in igneous rocks of upper layers of the crust.

5. Petromagnetic Constraints on the Lower Continental Crust

Petromagnetic information on the deep continental crust and upper mantle is recovered from two types of rocks: (1) Precambrian rock masses that experienced Granulitic metamorphism under conditions of the lower continental crust and whose plates were subsequently expelled to the Earth's surface; and (2) xenoliths of deep rocks entrained to the surface by basaltic magmas. The first type allows one to directly study the lithosphere section and to observe relationships between rocks, their spatial distribution etc.; however, these rocks were subjected to significant deuteric alterations during their subsequent existence (for example, secondary magnetite and ilmenite in granulites formed, at least in part, at temperatures <500°C, see section 4). Rocks of the second type are free from deuteric alterations associated with the stages of their uplift and subsequent existence, but they represent a set of accidental materials uncorrelated with the lithosphere section; they are transported from near-chamber zones characterized by specific processes of crystallization and recrystallization.

The aforesaid relates to the *objects of study.* The correct *methodological* approach is by no means less important. For example, petrochemical and other characteristics of rocks involved in petromagnetic measurements must be examined and usually this is not done. Petrochemical data allow one to assess the preservation degree of the material balance characteristic of the magmatic process; often the metamorphic control of magnetic properties of rocks is actually not present, which is established exactly from the comparison of petrochemical and magnetic characteristics (section 4) confirmed by experimental data (section 3). Also important are data on magnetic anisotropy, measurements of which are useful for correlating the formation of magnetic minerals and pre-, syn- and postmetamorphic deformations, but this evidence is used quite rarely.

Below we discuss results obtained from various regions.

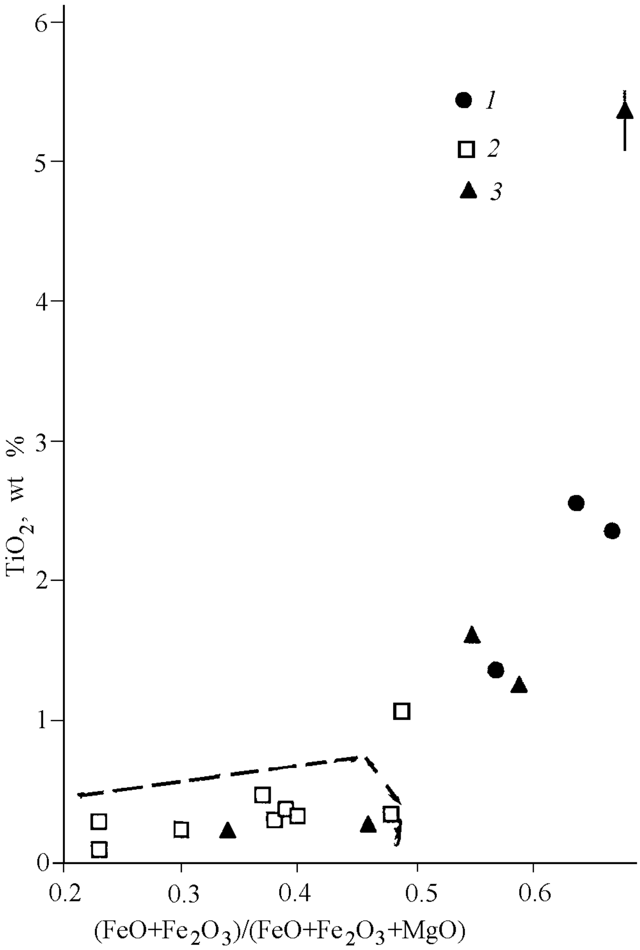

Figure 7. Diagram of $FeO+Fe_2O_3/FeO+Fe_2O_3+MgO$ versus TiO_2 averages reflecting the differentiation trend of basaltic magma (see the text for the regions included). The solid circles are oceanic basalts and basalts containing xenoliths, open squares are gabbros from the ocean floor and xenoliths, and solid triangles are Patynskii intrusive rocks ranging from primary nonmagnetic olivine gabbros to highly magnetic ferrogabbros. The broken line bounds the region of primary nonmagnetic gabbros. Points close to ($FeO+Fe_2O_3/FeO+Fe_2O_3+MgO \approx 0.57$, $TiO_2 \approx 1.5$) correspond to the original basaltic magma. The crystallization differentiation generally trends from cumulates (bottom) to ferrogabbros (top).

A. Study of Xenoliths

Afar (Ethiopia) [*Kashintsev and Pechersky,* 1983]. Numerous inclusions of deep mantle and crustal rocks (harzburgites, lherzolites, wehrlites, pyroxenites, gabbros and anorthosites) in young alkaline basalts of Ethiopia have been studies. The majority of the xenoliths are nonmagnetic.

Mongolia [*Lykov and Pechersky,* 1984; *Lykov et al.,* 1981]. A large group of xenoliths of deep rocks from

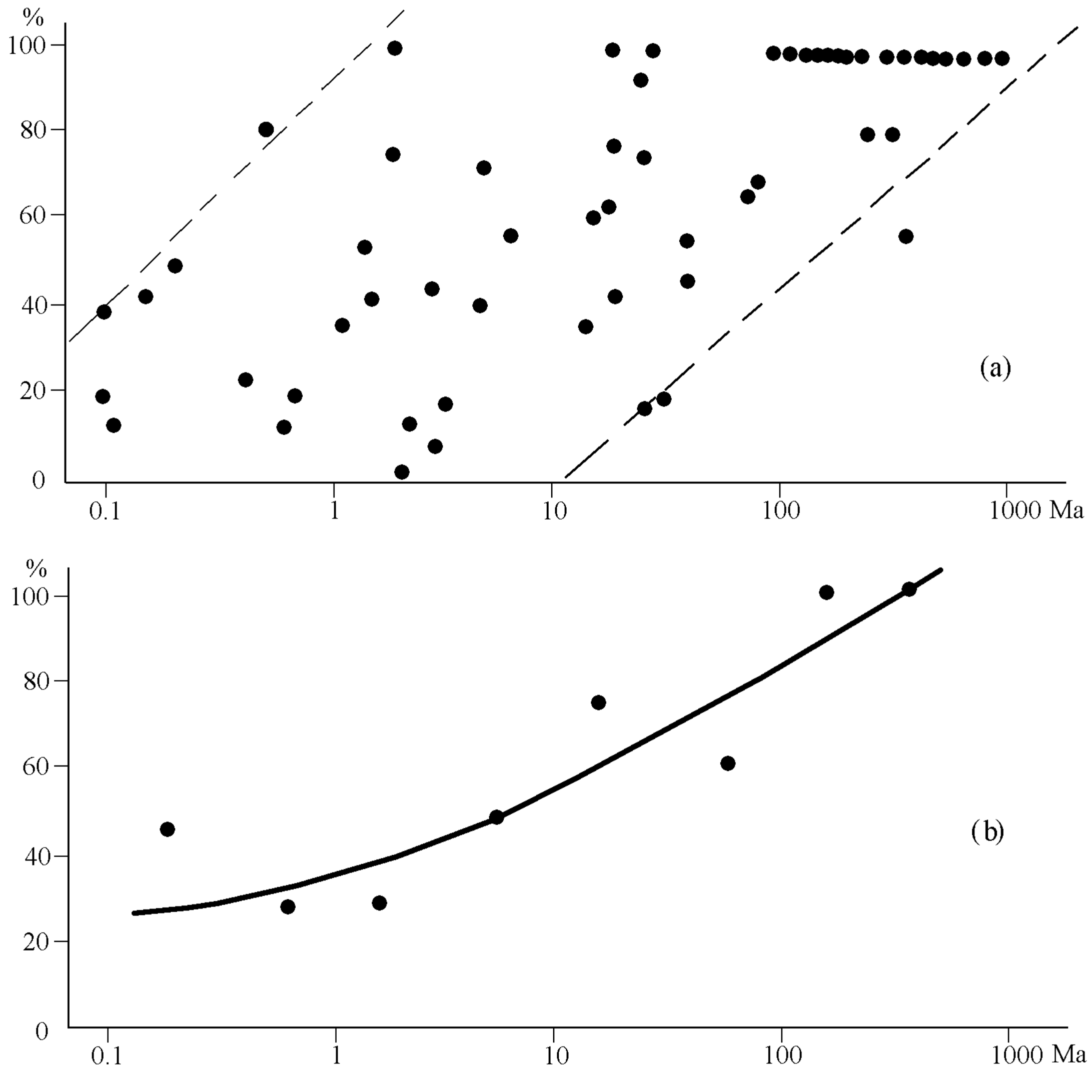

Figure 8. Distribution of relative amount of magnetite (in %%) in Phanerozoic subaerial basalts as a function of their age. The relative amount of magnetite was determined from curves of thermomagnetic analysis: (a) data averaged over collections; (b) all data averaged over equal intervals of the logarithm of time.

Pliocene-Quaternary basalts of central Mongolia was studied. According to petrographic characteristics they form the following two groups. (1) Ultrabasic mantle rocks mostly represented by lherzolites and eclogites; the majority of samples are nonmagnetic; the absence of ores is established from electron microscopy and microprobe studies; the rocks include rare grains of secondary Mg-Al-Fe spinel with T_c = 320–380°C and small grains of secondary magnetite in cracks and at edges of silicate grains. (2) Crustal rocks represented by pyroxenites, gabbros and more acidic varieties. Main varieties with SiO_2 = 45–55% are nonmagnetic ($J_s < 0.2$ A m^2/kg) and the more acidic rocks are magnetic (J_s = 0.7–1.5 A m^2/kg), which is consistent with the accumulative and differentiation trends (Figures 4 to 7). Measured Curie points of the crustal rocks are close to magnetite, whereas the T_C values estimated from microprobe measurements titanomagnetite grains of average compositions range from 190 to 480°C. Along with the presence of decomposition structures, this fact suggests that magnetite of magnetic crustal inclusions is a product of primary titanomagnetite decomposition.

Lesser Caucasus [*Genshaft et al.,* 1985; *Lykov and Pechersky,* 1984]. Inclusions in Pliocene-Quaternary volcanics and in the Tertiary Kayalu-Koyarchinskii diorite intrusive were studied. Xenoliths (gabbros, pyroxenites, gabbro-amphibolites and amphibolites) are everywhere similar in mineral composition. The epigenetic origin and an insignificant control of the transporting magma are supported by the fact that inclusions similar in composition and mineralogy occur in host rocks of different petrochemical types and by the absence of correlation between petrochemical characteristics of the inclusions and host rocks ($r<0.1$). As transformations progressively change from amphibolization to partial melting, the ores component intensely develops and magnetization of the samples increases. The main ore mineral is low-Ti titanomagnetite (TiO_2 <10%) which is usually decomposed and grains of which are often corroded and range in size from a few micrometers to 1 mm; ilmenite is less widespread. Rocks unaffected by deuteric alterations are weakly magnetic ($J_s \approx 0.1$ A m^2/kg; $\kappa \approx 10^{-2}$ SI units). Magnetization is clearly increases with extent of partial melting (J_s up to 10 A m^2/kg; κ up to 16×10^{-2} SI units). Abundant titanomagnetite crystallizes during the partial melting process. The amphibolization control of magnetization is weaker and apparently has a maximum in the intermediate region; strongly amphibolized rocks without melting traces are weakly magnetic. The fact that the enrichment in secondary magnetic minerals is unrelated to deuteric alterations [*Genshaft et al.,* 1985] is supported by a distinct tendency of increase in magnetization (concentration of magnetic minerals) with the increasing ferruginosity (FeO+ Fe_2O_3)/(FeO+Fe_2O_3+ MgO) (r=0.81) and iron oxidation extent Fe_2O_3/FeO+Fe_2O_3 (r=0.85). Furthermore, the J_s versus SiO_2 dependence in Lesser Caucasus xenoliths is similar to the left branch of differentiation (Figure 4); the points in the AFM (Figure 5) and MgO versus FeO+Fe_2O_3 (Figure 6) diagrams lie in the region of primary magmatic accumulative and differentiation trends. Fe and Ti (Figure 7), as well as Mg and Ca, are closely correlated. It is unlikely that Fe/Ti and Mg/Ca in the fluid composition are the same as in the magma [*Pechersky,* 1994].

Thus, the Lesser Caucasus data clearly demonstrate that the involvement of deuteric alterations in the magnetite enrichment of rocks is unlikely. An increase in the concentration of magnetic minerals from cumulates to later differentiates is a primarily magnetic feature affected by transformations of magnetic minerals up to the appearance of secondary magnetite after primary magnetic minerals in the process of metamorphism.

Kurile Islands [*Ermakov and Pechersky,* 1989]. The xenoliths of gabbroids from young lavas of the Kurile Islands (Paramushir, Simushir, Kunashir and Shikotan) were studied as an example of the crustal section under an island arc. The gabbroids formed in two stages: coarse-grained gabbro-allivalites first formed, and afterward they reworked, amphibolized and partially melted under conditions close to the granulite facies of metamorphism ($P \leq 9$ kbar and $T \leq 900°$C) with the formation of secondary titanomagnetite (possibly, under the action of the host magma in the near-chamber zone). In view of a significant role of compression, which is evident from the strong magnetic anisotropy, the reworking depth was smaller than 30 km. According to the composition (x) of titanomagnetites, they can be subdivided into four groups: (1) gabbro-allivalites with x = 0.12–0.29 and the admixture concentrations Al_2O_3 = 4.2% and MgO = 3.8%; the titanomagnetite grains are homogeneous, often rounded and usually large; (2) pyroxene gabbros with x = 0.12–0.29 and admixtures Al_2O_3 = 0.8% and MgO = 1.8%; fine decomposition is often observed; (3) leucocratic gabbros and gabbro-diorites with x_{av} = 0.16 and MgO = 1.4%; titanomagnetite grains are often decomposed and are corroded by sphene; (4) amphibole gabbros and amphibolites with x_{av} = 0.14. Al_2O_3 = 7.1% and MgO = 3.9%; titanomagnetite grains are homogeneous and fresh. The homogeneity of titanomagnetites in the first and fourth groups is confirmed by the agreement between the Curie points estimated from titanomagnetite compositions and measurements. The above composition of titanomagnetites is typical of island arc volcanics [*Pechersky,* 1994; *Pechersky and Didenko,* 1995; and others]. The measured Curie points in the second and third groups are close magnetite values and markedly diverge from theoretical estimates. The average titanomagnetite compositions are very close to those of secondary titanomagnetites from gabbros of both continental and oceanic crust that formed as a results of destruction of primary titanomagnetites (see section 4). Magnetization of the rocks vary within wide limits: κ from 2 to 63×10^{-3} SI units, J_s from 0.2 to 10 A m^2/kg and Q_n from 0.3 to 8.6. Most titanomagnetite grains are decomposed and corroded and include isolated large grains of homogeneous titanomagnetite having the shape typical of magmatic crystallization. Late grains of secondary titanomagnetite are fresher and have a rounded shape consistent with solid-state high-temperature crystallization. The least altered rocks are isotropic (average anisotropy of susceptibility of 1.03); reworking of the rocks enhances anisotropy (1.10–1.33, the average 1.18). Therefore, reworking and particularly amphibolization of the rocks occurred under conditions of a high stress.

Petrochemical characteristics show the gabbroids studied to be consistent with the process of magmatic differentiation: the minimum in the J_s versus SiO_2 curve is close to that typical of cumulates; gabbros of the differentiation trend prevail (Figures 4 to 7) [*Pechersky,* 1994]. The rocks altered under high-oxidation conditions as is indicated by the value of Fe_2O_3/(FeO+Fe_2O_3): it is 0.19 in unaltered wehrlite, 0.27 in allivalites, about 0.4 in gabbros, 0.45 in leucocratic gabbros and 0.5–0.6 in recrystallized amphibolized varieties.

Yakutia [*Genshaft et al.,* 1995; *Pechersky,* 1994]. Xenoliths from kimberlite pipes of Yakutia are rocks of metamorphic sequences of the Anabar shield basement and represent an example of the "titanomagnetite-free", primarily magmatic ore mineralization. After formation of the kimberlite bodies, the rocks experienced near-surface hydrothermal metasomatic transformations represented by carbonatization, serpentinization and chloritization, which disturbed the material balance (removal of silicon and iron, influx of potassium and calcium, relative enrichment in titanium and magnesium, and a ferruginosity decrease). According to the relationship between mineral phases, the studied xenoliths form three groups: (1) rocks of a distinctly magmatic genesis: gabbros (serpentinized and carbonatized), pyroxenites (serpentinized), diorites (weakly affected by deuteric alterations) and hornblendites; (2) garnet-free rocks of the granulite facies metamorphism, pyroxene and amphibole plagiogneisses, biotite-amphibole schists and amphibolites, and pyroxene-amphibole schists; (3) eclogitized rocks of the granulite facies metamorphism, plagioclase rocks with garnet and pyroxene, amphibole-pyroxene-garnet schists, and pyroxene-garnet schists. Rocks of the second and third groups virtu-

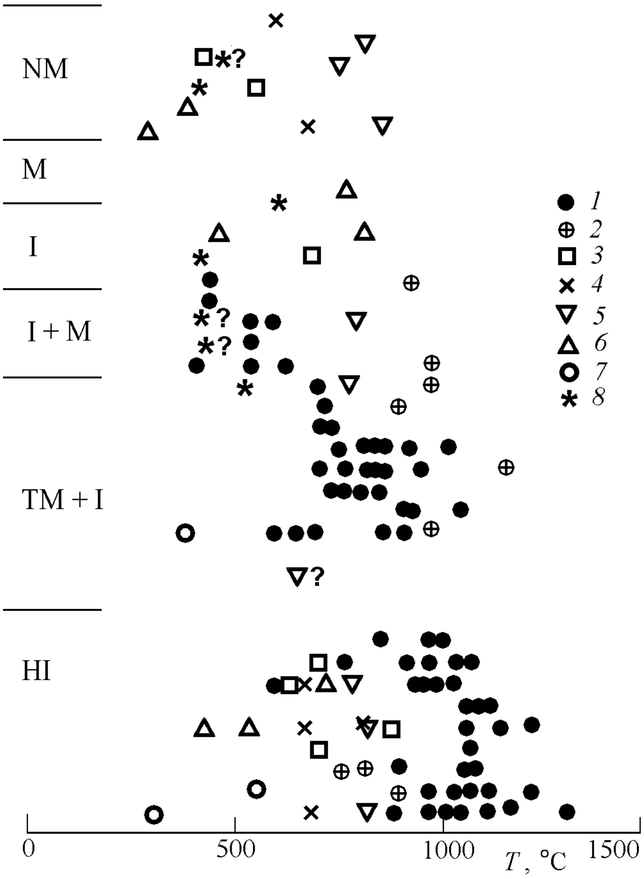

Figure 9. Equilibrium temperatures of various mineral pairs grouped according to types of minerals present in Yakutia xenoliths: NM, primary nonmagnetic minerals with low-temperature secondary magnetite; M, secondary magnetite; I, ilmenite; HI, hemoilmenites; TM, titanomagnetites. (1)–(7) Mineral pairs shown: (1) titanomagnetite(magnetite)-hemoilmenite, (2) clinopyroxene-orthopyroxene, (3) clinopyroxene-garnet, (4) amphibole-garnet, (5) amphibole-plagioclase, (6) chlorite-garnet, (7) plagioclase-alkali feldspar; (8) data of the Sholpo-Luzyanina magnetic thermometer.

ally coincide in Fe and Ti concentrations, which reflects the "memory" of their primarily magmatic origin. Overall, the Yakutia xenoliths are mostly represented by differentiation products of the basaltic magma (Figure 6). A bias toward accumulative trends is associated with the aforementioned deuteric alterations in the rock composition. According to the composition of ore minerals, the studied samples are divided into the following four groups. (1) Hemoilmenite. Preserved homogeneous grains are present but most grains are decomposed into ilmenite+high-Ti titanomagnetite+magnetite; based on average compositions of intergrowths of hemoilmenite and titanomagnetite, the Lindsley thermometer temperature of their formation is about 1400°C, fO_2 is close to the QMF buffer, which is evidently consistent with the early crystallization magma conditions, and the temperature of the hemoilmenite and titanomagnetite decomposition products is 800 to 1200°C. (2) Titanomagnetite+ilmenite. Most samples represent products of the group-1 hemoilmenite decomposition and recrystallization; the formation temperature of this grain association is 700 to 1000°C. (3) Ilmenite+magnetite. The main ore mineral is ilmenite ($x = 0.93$); isolated large grains of magnetite appear; the formation temperature of ilmenite+magnetite intergrowths is less than 600°C. (4) Primarily nonmagnetic group of samples. Undoubtedly secondary, often fine-grained magnetite is only present.

Magnetite is the main magnetic phase in the xenoliths. Also recognized are hemoilmenite (T_C = 100–200°C) and titanomagnetites (T_C = 200–450°C) which are products of the multiphase oxidation of hemoilmenite. The J_s value ranges from ~0.01 to 15 A m^2/kg, the mode being observed in the interval 0.1–1.0 A m^2/kg. Metamorphic rocks are more magnetic than rocks that preserved primary structures and their respective averages are J_s = 1.16 A m^2/kg (0.05–15 A m^2/kg) and J_s = 0.59 A m^2/kg (0.02–5.8 A m^2/kg). This is related to the appearance of secondary magnetite. The magnetization markedly drops in eclogitized rocks, averaging J_s = 0.5 A m^2/kg. Grains are relatively large, whereas most magnetite grains are relatively small. According to Sholpo-Luzyanina magnetic thermometer [*Sholpo*, 1977], the majority of grains of the hemoilmenite sample group formed at temperatures not lower than 600°C; magnetically soft grains of the titanomagnetite+magnetite group formed below the Curie point of magnetite, and magnetically harder grains, above it. Nearly all magnetite of the ilmenite+ magnetite and primarily nonmagnetic groups formed below its Curie point. Magnetite is more abundant in anisotropic amphibole-bearing gneisses and schists; the magnetic susceptibility anisotropy in schistose rocks averages 1.22, whereas in rock of magmatic genesis it is 1.08. In the process of carbonatization and serpentinization, earlier magnetite is either destroyed or substantially reworked, so that later (post-stress) magnetite prevails and the resulting average anisotropy of such samples is 1.07.

Compositions of various mineral assemblages were used for estimating the $P-T$ conditions of the formation of various parageneses (Figure 9). According to these data, several stages can be recognized in the evolution of deep rocks. They formed primarily as igneous rocks of shallow crystallization differentiation under oxygen fugacity conditions close to the QMF buffer (Figure 10). A temperature decrease from 1300 to 950°C leads to the multiphase oxidation of primary hemoilmenite and titanomagnetite with the oxygen fugacity increasing up to the Ni-NiO buffer (Figure 10). Mineral assemblages indicate substantial reworking of primary magmatic rocks in the granulite facies of metamorphism. Current mineralogical geothermobarometers yield temperatures of 650–870°C and pressures of 5–10 kbar for the formation conditions of the granulite assemblages. Since the pressure was not isotropic (at least at the crystallization stage of anisotropic magnetite), the actual depth of the metamorphism is likely to have been smaller than 25 km. Under these conditions, the decomposition of hemoilmenite and titanomagnetite continued, with the formation of the assemblage of titanomagnetites of various compositions and ilmenite. The cooling of the rocks under higher oxygen fugacity conditions led to the formation of the assemblage of ilmenite and magnetite.

Thus, this example on the one hand clearly demonstrates the secondary nature of the regional magnetic anomaly sources formed as a result of the deep recrystallization of hemoilmenite and ilmenite typical of Archean volcanics into magnetic minerals and on the other hand is consistent with the main concept of the formation of magnetic rocks due to crystallization or subsequent recrystallization of primary magmatic Fe-Ti ores.

Ross Island, the Antarctic [*Warner and Wasilewski,* 1995]. This is a region of continental rifting, high heat flow and crustal thinning. The studied xenoliths from Cenozoic island volcanoes are dunites, pyroxene granulites and hornblendite. Virtually nonmagnetic granites and granite gneisses prevail among the regional xenoliths from the upper crust [*Behrendt et al.,* 1991].

Pyroxene granulites represent the lower crust and consist of primary plagioclase, pyroxene, olivine and ilmenite (to 3%). Besides large grains of primary ilmenite, segregations of fine ilmenite after amphibole (decomposition) are observed. The crystallization temperatures of ortho- and clinopyroxene are 736 to 994°C. Judging from average compositions of coexisting ilmenite and titanomagnetite (Lindsley thermometer) their crystallization temperatures are 720 to 830°C, which is consistent with the initial multiphase oxidation stage of ilmenite. According to mineral equilibria of olivine and pyroxenes, the xenoliths of pyroxene granulites originated at depths of 12 to 20 km. Secondary mineralization is observed in all granulites, with amphibole being a main secondary mineral. Another secondary mineral is biotite usually associated with amphibole. Secondary titanomagnetite develops after ilmenite grains. Partial melting features are often observed, and enrichment in Fe-Ti oxides (mostly high-Ti titanomagnetite) is associated with the partial melting areas in many granulites; this titanomagnetite is in turn subjected to multiphase oxidation. Importantly, *the partial melting zones enriched in ores are typical of ilmenite-rich granulites, whereas no or minor ores are present in the partial melting zones in granulites that do not contain primary Fe-Ti oxides,* with olivine crystallizing in such zones. Magnetization of the granulites vary within wide limits (κ from 0.28 to 36.7×10^{-3} SI units and J_n from 0.23×10^{-4} to 90.2×10^{-4} A m^2/kg) and granulites subjected to partial melting are most magnetic. Overall, the magnetization of the granulites is markedly lower than that of their host lavas, and in some samples it is even lower than in upper crust granites and granite gneisses. The ores concentration and accordingly magnetic susceptibility were found to correlate with the iron concentration in pyroxenes of all xenoliths and primarily in granulites. This fact is *unfavorable* for associating this correlation with metamorphism but corroborates a primary magmatic distribution of iron in the process of melt differentiation with formation of low-Fe cumulates and high-Fe differentiates.

The dunite samples are mostly composed of olivine grains of various sizes with occasional chromite grains the majority of which are associated with partial melting. Isolated grains of magnesioferrite were found in a sample of partially melted dunite. The crystallization temperatures of olivine-chromite are 1012 to 1106°C and correspond to the uppermost mantle according to the geothermal gradient in the region. Geophysical data fix the Moho at depths of 20 to 23 km. The presence chromite and absence of garnet are evidence that the dunites have come from depths not greater than 45 km. The dunites are weakly magnetic ($\kappa < 5\times10^{-3}$ SI units and $J_n < 7\times10^{-4}$ A m^2/kg).

The amphiboles of hornblendite differ those of granulites by a high Ti concentration because hornblendite is a derivate of the alkaline magma of the Cenozoic MacMurdo volcano group. Besides amphibole, hornblendite contains clinopyroxene and ores (3%) that are mostly represented by ilmenite, rarer heterogeneously oxidized titanomagnetite and magnetite, and minor pyrrhotite. The sample experienced substantial partial melting, with which numerous fine grains of titanomagnetite are associated; accordingly, the its magnetization is high.

Remanence carriers in all of the magnetic samples are

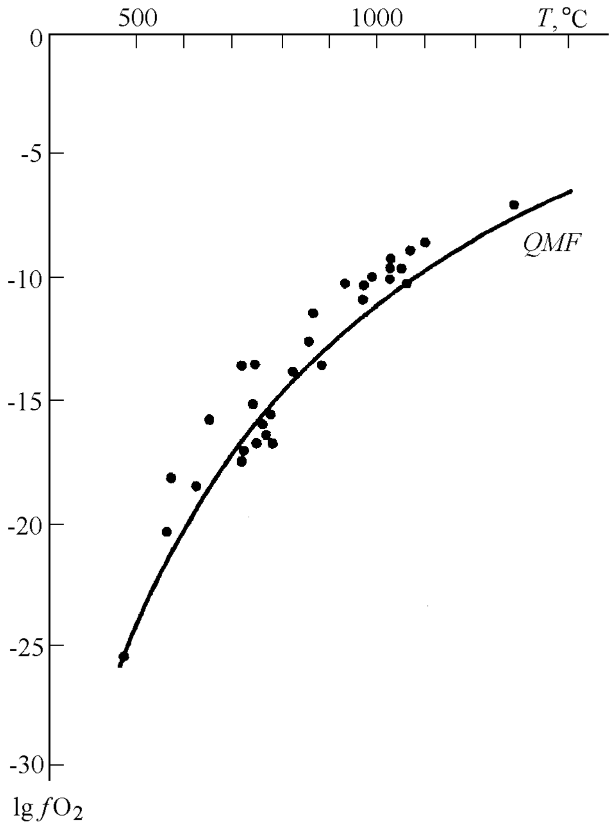

Figure 10. The T versus fO_2 diagram of xenoliths from Yakutia kimberlite pipes obtained from intergrowths of titanomagnetite and hemoilmenite (Lindsley geothermometer [*Spencer and Lindsley,* 1981]).

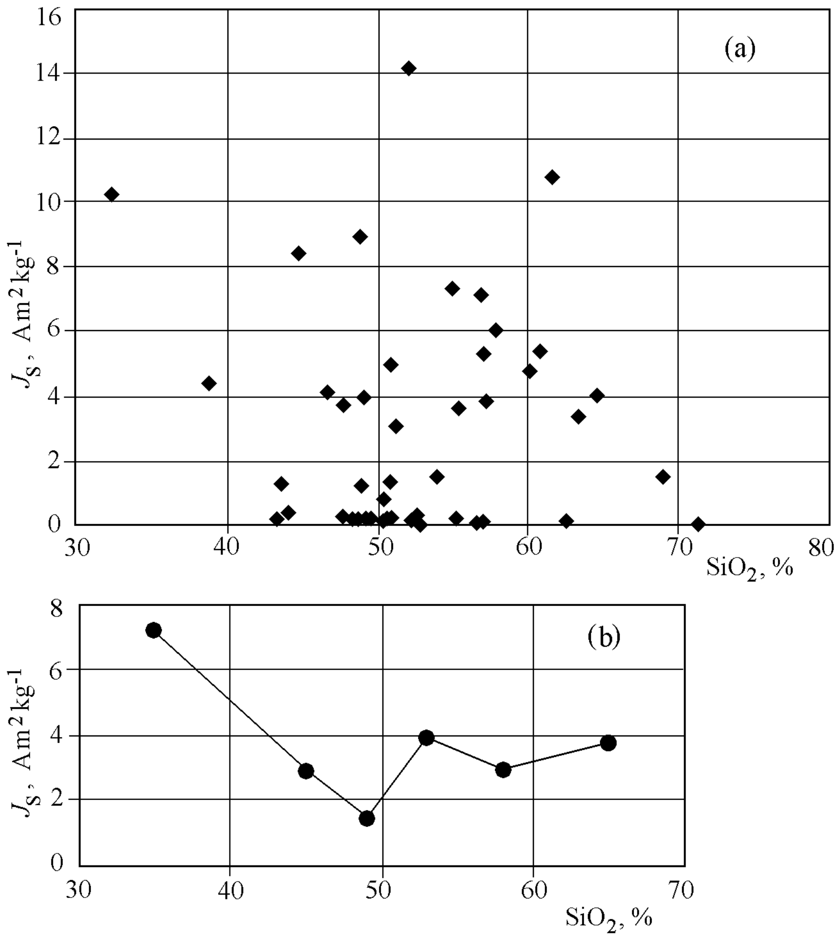

Figure 11. The SiO_2 versus J_s diagram for Archean metamorphic rocks of the Aldan Shield: (a) data from samples; (b) the same data averaged over SiO_2 intervals.

pseudosingle-domain grains with $T_C \geq 550°C$.

Nearly all xenoliths were variously oxidized at the cooling stage. The oxygen fugacity as constrained from heterogeneously oxidized titanomagnetites varies within 2 to 3 orders of magnitudes around the QMF buffer. Judging from the appearance of pseudobrucite after ilmenite, its oxidation started above 800°C, but in most cases the oxidation temperature was lower and pseudobrucite lamellae did not appear, but rutile lamellae after ilmenite were segregated. Thus, the authors argue for weak magnetism of the lower crust in the Ross Island region; its primary ore mineral is ilmenite, and a higher magnetization of rocks is related to their partial melting and formation of secondary titanomagnetite during or after the time at which the xenoliths were trapped by the magma of the Cenozoic volcanoes. Moreover, due to a high geothermal gradient, only the upper 12 km of the crust, where the temperature does not exceed 550°C, can be magnetic. Therefore, the lower crust and upper mantle underlying the present rift are *in situ* nonmagnetic, which is consistent with a lower anomalous field above the region determined from Magsat data.

B. The Study of Archean-Proterozoic Metamorphic Sequences

Aldan Shield [*Bazhenova et al.*, 1998, 2001]. The Aldan Shield is a complex structure with long-term development of magmatism and prograde and retrograde metamorphism involving its Archean and Proterozoic rocks. The oldest formations are arch structures inner parts of which are composed of variously amphibolized and granitized metabasites (mostly pyroxenites) and enderbites subjected to the granulite facies metamorphism; the age of the enderbites is 3.6 Ga. The outer parts of the arches form linearly extended belts composed of younger rocks, namely, granite gneisses, enderbites, metabasites (metagabbros, amphibolites, pyroxene-amphibole and biotite-amphibole schists) and aluminous and carbonate rocks subjected to the amphibolite facies metamorphism. The age of this complex is 3.1 to 3.3 Ga [*Glukhovskii et al.*, 1993]. The Archean basement is superimposed by trough structures mainly composed of amphibolites, metabasites and granite gneisses. The trough structures range in age from 3.1 to 2.9 Ga.

Geological and petromagnetic characteristics were studied in samples from central, peripheral and outer parts of Central Aldan, Chara, Sunnaginski and other arches and from superimposed troughs. Magnetic metabasites were studied in more detail.

Petrochemical characteristics refer all studied metabasites to primary magmatic rocks of the differentiated tholeiitic and calc-alkaline series similar to recent geodynamic settings of continental margins and island arcs. The Aldan Shield rocks occupy a position in the MgO-(FeO+Fe_2O_3) diagram (Figure 6) very close to data from igneous rocks and are divided into two groups respectively represented by the differentiation trend and by less distinct accumulative trend (see section 4). A certain rightward bias of the accumulative points is similar to data from Yakutia kimberlites, i.e. the Aldan cumulates are more ferruginous. The primary magmatic situation is also reflected by a close positive correlation of Fe and Ti in the rocks: the trend of the Aldan Shield points in the FeO+Fe_2O_3/FeO+Fe_2O_3+MgO-TiO_2 diagram is similar to the trend observed in Figure 7 but is shifted somewhat to the right, i.e. closer to island-arc magmatism. The SiO_2 versus J_s diagram (Figure 11) is particularly important in this case; although the data give a large scatter, and chemical analyses were mostly made on magnetic rocks (i.e. their role in Figure 11 is overestimated), the points in Figure 11a can be divided into magnetic ($J_s > 2$ A m^2/kg) and nonmagnetic ($J_s < 0.5$ A m^2/kg) groups; the first encompasses a wide SiO_2 interval, from 33 to 70%, and corresponds to the magmatic differentiation trend; the points of the second group lie within a narrower SiO_2 interval, from 45 to 57%. The averaging of data over the SiO_2 intervals yields a J_s minimum near 50% SiO_2 corresponding to the accumulative trend (Figure 11b; cf. Figure 4).

The weakly magnetic and nonmagnetic rocks include chrome spinels and small ores crystals within pyroxene grains. Magnetic samples commonly contain, in parageneses with pyroxene, large (hundreds of micrometers) grains of ilmenite and magnetite, often as intergrowths; some grains exhibit clear features of subsolidus decomposition. Magnetite in some samples contains inclusions of rutile and/or sphene. The compositions of coexisting ilmenites and magnetites suggest that most formation temperatures did not exceed 500°C. Furthermore, small ores segregations are present along fine cracks in silicates and, as rims around silicates, on boundaries of silicate grains; they arose at retrograde metamorphism stages.

The recrystallization temperature pressure estimated from mineralogical geothermobarometers range from ∼950 to 300°C and from 9 to ∼5 kbar respectively. In this temperature interval, crystallization conditions of coexisting magnetite-ilmenite pairs were close to the Ni-NiO buffer, and the transition from granulite to amphibole facies of metamorphism is characterized by redox conditions complicating from those close to Ni-NiO buffer to conditions close to MH buffer.

The magnetic susceptibility κ varies from 10^{-5} to 10^{-1} SI units and has a bimodal distribution (Figure 12): 62% samples are virtually nonmagnetic and their mode lies in the interval $(0.035–0.1)\times10^{-3}$ SI units; only 18% samples are magnetic and their mode lies in the interval $(1–2)\times10^{-2}$ SI units. The structure-sensitive characteristics J_{rs}/J_s and Q_n vary from 0.002 to 0.2 and from 0.08 to 2.7 respectively, implying the predominance of large multidomain grains in all studied rocks. Rare, mostly nonmagnetic, samples have $J_{rs}/J_s > 0.1$ and $Q_n > 1$ (occasionally greater than 10), which is evidently related to the presence fine magnetite grains with a concentration not higher than 0.05%. According to the thermomagnetic analysis, magnetic minerals are predominantly represented by magnetite (T_C close to 580°C). Pyrrhotite (T_C = 340°C) is present in two samples. No correlation is present between κ and ores concentration determined from thin sections (Figure 13); overall, the thin-section constrained ores concentration is a few times higher than the magnetite concentration determined from J_s or κ. In conjunction with microprobe data this indicates that ores are dominated by ilmenite.

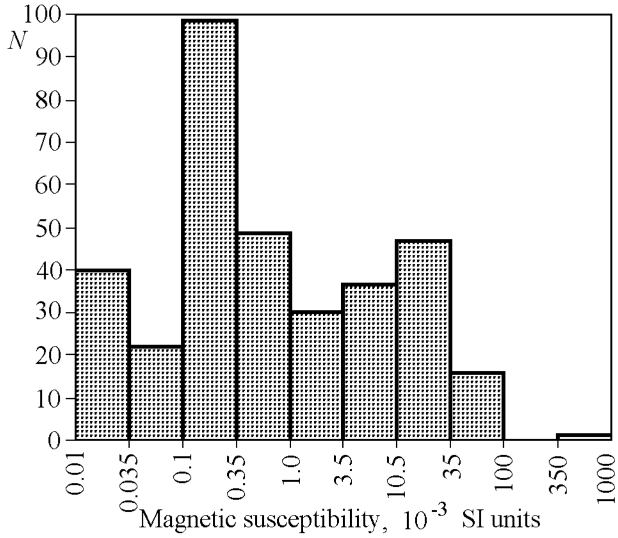

Figure 12. Magnetic susceptibility histogram for the Aldan Shield rocks.

The nonmagnetic rocks are mostly represented by metasedimentary acidic rocks (most of which appear to have a collisional origin) such as granites, granodiorites, syenites, granite gneisses, biotite-bearing gneisses, and schists; on the whole, sedimentary and "acidic" rocks account for 70% samples, whereas basic rocks such as pyroxenites, gabbros, pyroxene and amphibole gneisses and schists, and amphibolites account only for 30% samples; most magnetic rocks are basic (gabbros, amphibolites and pyroxene-amphibole gneisses) account for 64% thin sections, and 36% represent acidic rocks. This is reflected in the behavior of the quartz+alkali feldspar concentration sum. Thus, in a 0 to 25–30% interval of this sum, the scatter in κ values ranges from nonmagnetic values to $\kappa = 4\times10^{-2}$ SI units and the susceptibility abruptly drops at higher concentrations (Figure 14). Apparently, a 25–30% quartz+alkali feldspar concentration marks the transition from partially granitized rocks to acidic igneous rocks proper.

The above partitioning of magnetic and nonmagnetic varieties among petrographic rock groups evidently reflects the

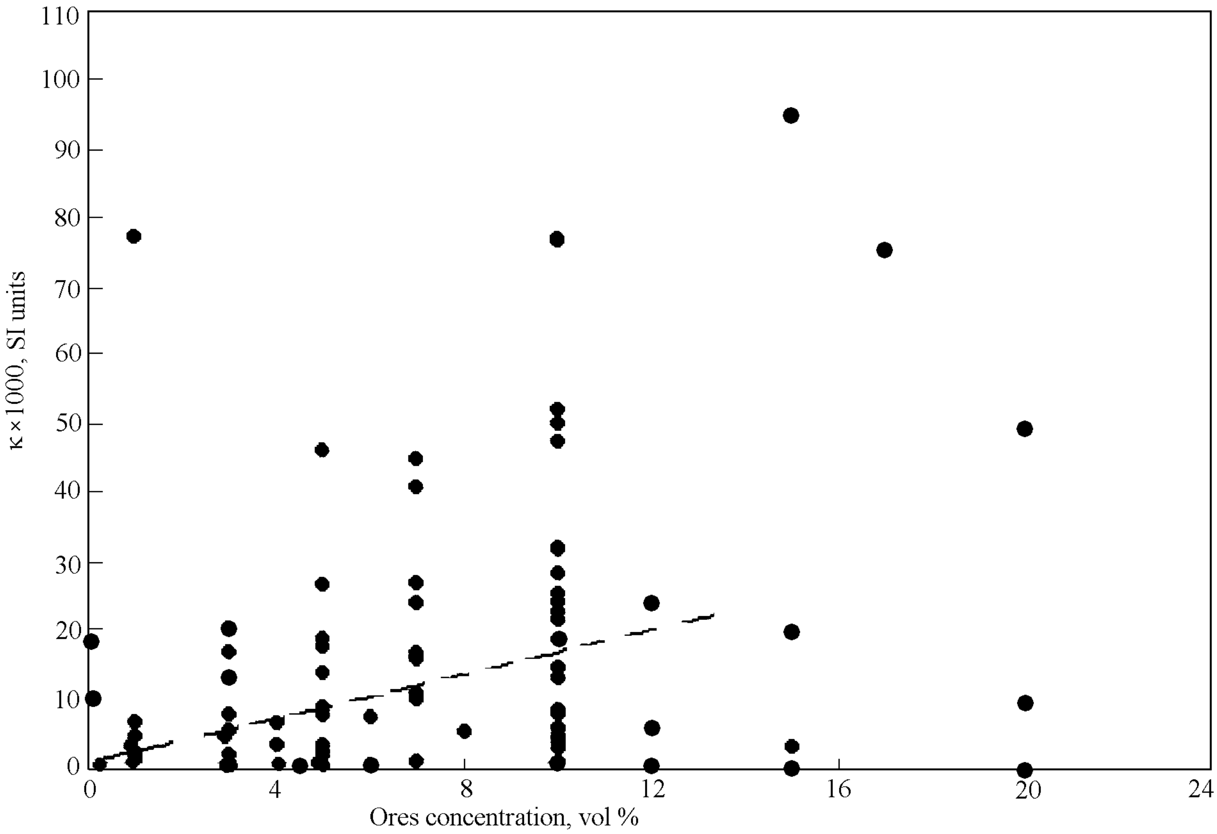

Figure 13. Magnetic susceptibility versus ores concentration from petrographic data on thin sections of the Aldan Shield rocks.

primary partitioning of magnetic minerals in these groups that has not been appreciably affected by the subsequent metamorphism accompanied by stress. The primary magmatic partition of magnetic minerals is supported by petrochemical characteristics (see above). The majority of magnetite grains appear as intergrowths with ilmenite, i.e. these intergrowths are likely to be destruction products of primary titanomagnetites and possibly hemoilmenites. Intergrowths and lamellae of ilmenite in magnetite are observed even in samples in which ores were derived from silicates at various stages of retrograde metamorphism.

Such a high percentage of nonmagnetic minerals in the collection is due to, first, a significant number of primary nonmagnetic acidic igneous rocks (as was mentioned above), second, the presence of primary nonmagnetic sedimentary rocks, third, accumulative basic rocks (Figure 11), fourth, the fact that *the Archean magmatic conditions were more reductive than in the later time,* and fifth, ores are more often destroyed under the retrograde metamorphism conditions. *A higher magnetization formed due to subsequent processes of ores (primarily ilmenite) transformations.*

As regards the dependence on the magnetic susceptibility (Table 2), i.e. on the concentration of magnetic minerals, the rocks are virtually indistinguishable in their schistosity and/or granitization degrees (Figure 14); they are quite uniform, which reflects the general similarity in the extent of regional metamorphism accompanied by stress. The uniformity of the stress metamorphism resulted in similar patterns of the magnetic anisotropy behavior patterns (Table 2). The major anisotropy control is effected by paramagnetic minerals in nonmagnetic rocks and by magnetite in magnetic rocks. Since the $\kappa_{\max}/\kappa_{\min}$ and E averages of both groups of rocks are very similar (Table 2), on can state that their anisotropy has a general origin.

As seen from Table 2, the κ value weakly correlates with the extent of deuteric alterations (retrograde metamorphism etc.) which gradually rises from 1.4 in nonmagnetic varieties to 1.6 in rocks with $\kappa = 10^{-2}$ SI units and abruptly drops to 1.2 in magnetic rocks. Therefore, most magnetite formed before the stress metamorphism, and its small part formed in the process of deuteric alterations. Secondary magnetite appeared *after* the stress, as is evident from the absence of correlation between the extent of deuteric alterations and anisotropy (Table 2).

Voronezh crystalline massif [*Genshaft et al.,* 1997]. The Voronezh crystalline massif is an inlier of the Precambrian basement and includes a series of Archean blocks separated by linear graben-synclines. The linear zones are composed of variously metamorphosed volcanic-sedimentary rocks and intrusive bodies. Structural elements of the massif have signatures in geophysical fields, in particular, in the anomalous magnetic field. Magnetic anomalies (usually linearly extended) are commonly associated with

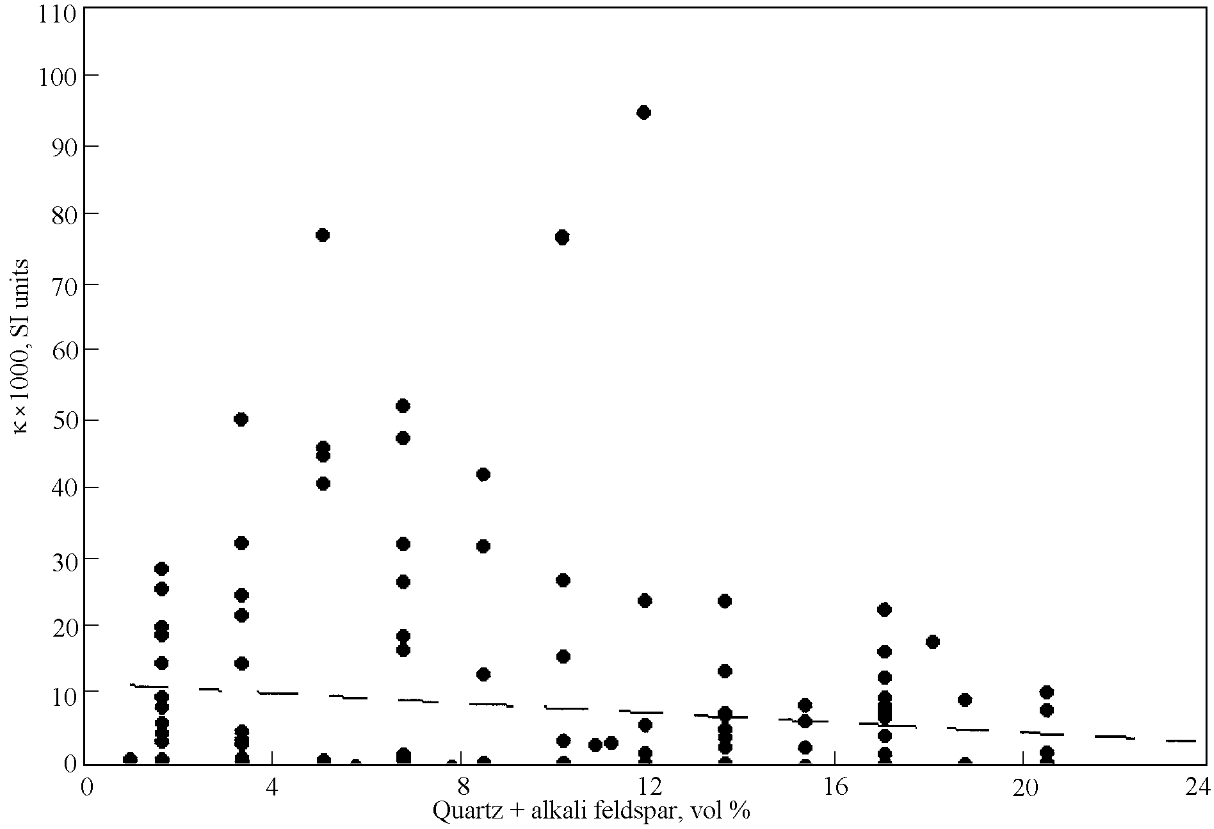

Figure 14. Magnetic susceptibility versus quartz+alkali feldspar concentration in the Aldan Shield rocks.

ultrametamorphic and magmatic bodies [*Nadezhka et al.*, 1989].

The studied samples are dominated by ortho-rocks: granitoids, gabbro-diorites, gabbroids and pyroxenites; para-rocks are represented by plagiogneisses, quartzites and various schists. The rocks experienced granulite-facies metamorphism and are amphibolized and often biotitized. The ores present in all rocks are coarse-grained magnetite and ilmenite (often their intergrowths) and large grains of decomposed (into gabbro) titanomagnetite. Small grains of magnetite and sulfides are observed in pores and at boundaries of silicates. According to the Lindsley geothermometer, ilmenite-magnetite intergrowths formed within a 1000–460°C interval at fO_2 close to the QMF buffer.

The thermomagnetic analysis either resolves solely magnetite ($T_C \approx 580°C$) or the curves $J_s(T)$ have a paramagnetic

Table 2. Some characteristics of Aldan Shield basement rocks

κ, 10^{-3} SI units	$\kappa_{max}/\kappa_{min}$	$=\kappa_2\kappa_2/\kappa_1\kappa_3$	Texture	Deut. alter.
0.01–0.11	1.26 (1)	1.01 (1)	20, 53	1.4
0.12–0.2	1.20 (10) [1.07–1.32]	0.94 (1)	17, 48	1.32
0.2–04	1.13 (10) [1.02–1.32]	1.02 (3) [1.0–1.05]	42, 67	1.53
0.4–0.7	1.17 (18) [1.05–1.27]	0.99 (5) [0.95–1.06]	40, 50	1.48
0.75–2.0	1.20 (11) [1.05–1.33]	1.01 (7) [0.96–1.08]	53, 80	1.59
2.0–10.0	1.26 (25) [1.03–1.85]	1.05 (15) [0.88–1.22]	37, 60	1.62
10.0–35.0	1.27 (24) [1.07–2.02]	1.04 (29) [0.84–1.28]	22, 45	1.17
≥40	1.24 (10) [1.05–1.38]	1.03 (12) [0.84–1.31]	29, 50	1.29

Note: κ is the magnetic susceptibility; $\kappa_{max}/\kappa_{min}$ is the κ anisotropy (parenthesized is the number of samples); $E = \kappa_2\kappa_2/\kappa_1\kappa_3$ is the magnetic fabric (planar if $E > 1$ and linear if $E < 1$): $\kappa_1 = \kappa_{min}$, κ_2 is the intermediate axis of ellipsoid and $\kappa_3 = \kappa_{min}$. Texture is the percentage of samples (thin sections) of foliated rocks (the first number refers to markedly foliated rocks and the second number refers to markedly+weakly foliated rocks). "Deut. alter." is the average characteristic of deuteric and later alterations superimposed on regional metamorphism, including unaltered (0), weakly altered (1) and appreciably altered (2) rocks (as assessed from thin sections).

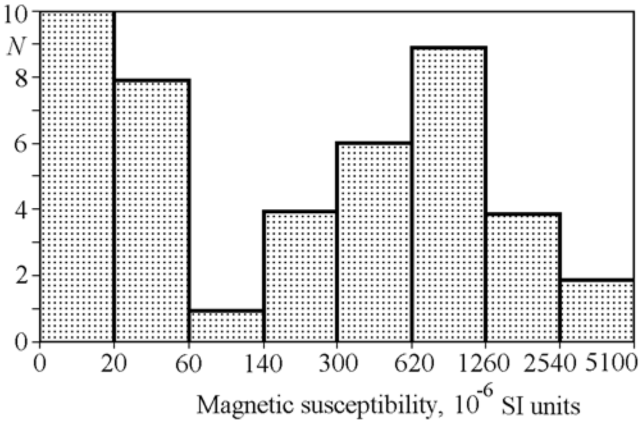

Figure 15. Magnetic susceptibility histogram for rocks of the Voronezh crystalline massif.

hyperbolic shape. Most samples exhibit weak magnetic anisotropy (less than 1.10), but the anisotropy in gabbroid samples is appreciable and averages 1.22.

The magnetic susceptibility distribution of the rocks is bimodal (Figure 15). The group of nonmagnetic rocks ($\kappa \leq 15 \times 10^{-5}$ SI units) is mainly represented by acidic igneous and basic accumulative rocks, whereas the group of magnetic rocks ($\kappa \geq 10^{-3}$ SI units) mainly includes basic and intermediate igneous rocks (differentiation trend). The J_s versus SiO_2 dependence refers most of the studied rocks to two groups of the primary magmatic type: accumulative group ($J_s < 0.5$ A m^2/kg and SiO_2 of about 50%) and magma differentiation trend ($J_s > 1$ A m^2/kg and SiO_2 of 35% to 65%) (see also Figure 6). For example, rocks of the Shiryaevskaya intrusion, rich in MgO, FeO and CaO and depleted in Fe_2O_3 and TiO_2, are nearly an order of magnitude less magnetic than rocks of the Smorodinskaya intrusion [*Skryabina and Afanas'ev,* 1981].

One can draw a conclusion that the magnetization of the studied samples is controlled by the amount of secondary magnetite whose concentrations are consistent with the primary magmatic stage of rock formation. The superimposed (retrograde) metamorphism only leads to recrystallization of magnetic minerals and their partial destruction; secondary magnetic minerals can form during the superimposed processes of metasomatism, granitization and serpentinization of ultrabasic rocks [*Afanas'ev,* 1978], but magmatic processes made the major contribution to the formation of magnetic minerals.

Anabar Shield. In addition to xenoliths from kimberlite pipes (see above), the magnetic susceptibility was measured on a large series of samples from ancient beds of the Anabar Shield (collection of V. L. Zlobin). They include rocks from the Maganskii, Daldynskii and Khapchanskii terranes separated by collisional zones. The structure of the terranes involves volcanic-sedimentary complexes of Archean (3.2–3.0 Ga) and Proterozoic (2.3–2.0 Ga) ages metamorphosed under the granulite facies conditions (P=6–11 kbar, T=700–900°C); collisional zones of deep faults are specified as zones of superimposed metamorphism (mostly in the amphibolite facies), granitization and migmatization [*Lutz and Oksman,* 1990]. Predominant rocks of the terranes are bipyroxene and hypersthene plagiogneisses and schists. Lenticular bodies of wehrlites are occasionally found. Ilmenite is the main ore mineral. In several cases, secondary magnetite replaces dark-colored minerals during retrograde transformations of granulitic rocks: in the process of amphibolization and biotitization of schists and gneisses, the magnetite develops as opacite rims along joints on boundaries between pyroxene and amphibole. Magnetite typically appears during the serpentinization of olivine in wehrlites. Rocks subjected to retrograde alterations occur in the vicinity of the collisional zones. These processes are likely to be responsible for the anomalous saw-toothed magnetic field.

The majority of measured samples are virtually nonmagnetic. The magnetic susceptibility of ortho-rocks have a basically bimodal distribution (Figure 16a): the first (nonmagnetic) group of rocks has a magnetic susceptibility less than 1.2×10^{-4} SI units, the second (weakly magnetic) group has a mode in the interval (4–36)$\times10^{-4}$ SI units, and less than 10 samples are magnetic ($\kappa \geq 10^{-2}$ SI units). 90% samples

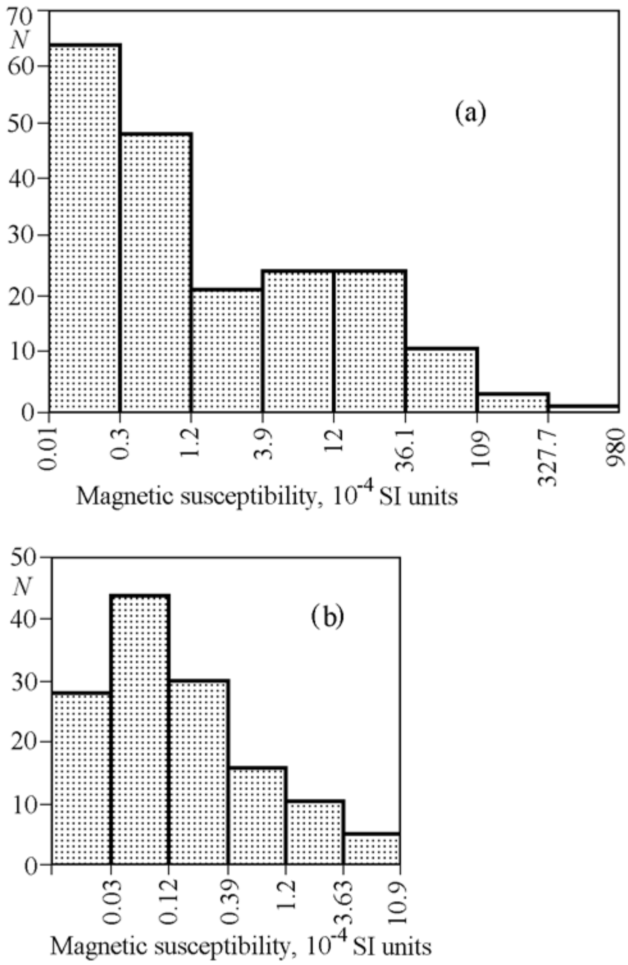

Figure 16. Magnetic susceptibility histograms for (a) ortho- and (b) para-rocks of the Anabar Shield.

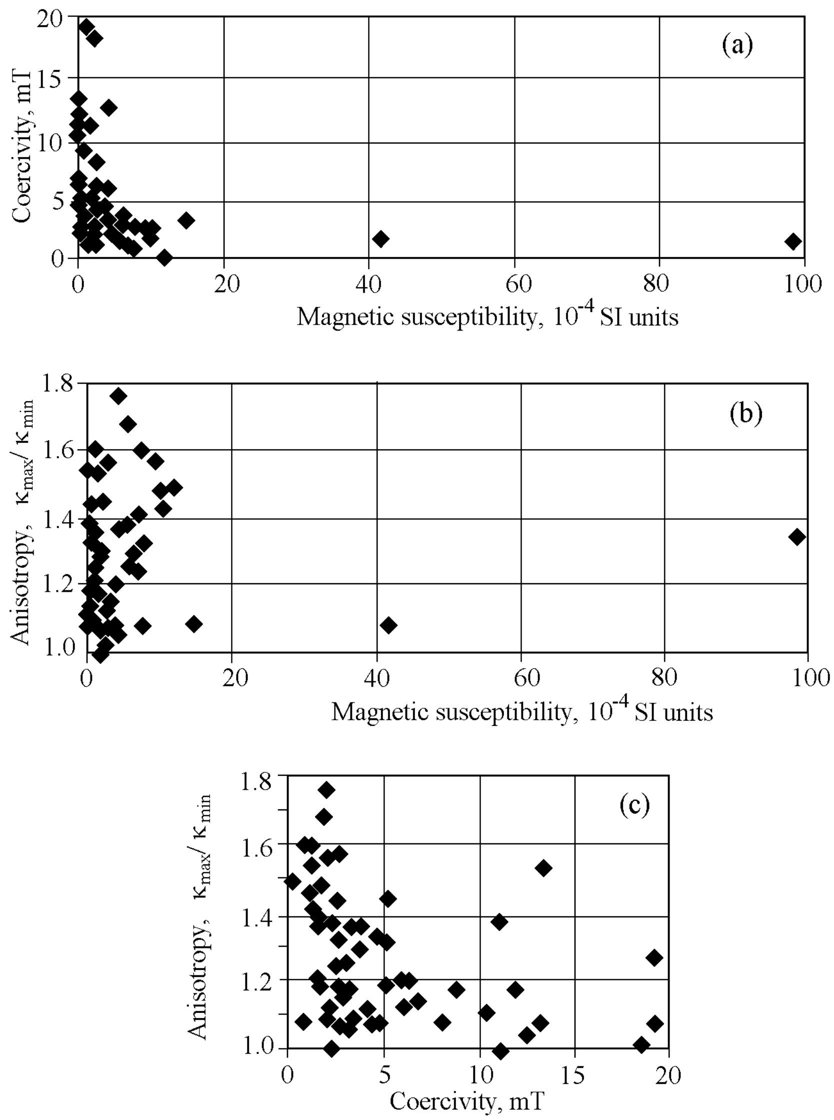

Figure 17. Plots of (a) the magnetic susceptibility versus coercivity, (b) magnetic susceptibility versus its anisotropy and (c) magnetic susceptibility anisotropy versus coercivity.

of para-rocks have $\kappa \leq 10^{-4}$ SI units (Figure 16b). The majority of magnetic minerals are magnetically soft (coercivity less than 6 mT) irrespective of the susceptibility value, i.e. magnetite concentration (Figure 17a); fine-grained magnetite with $H_c > 10$ mT is present in a part of the virtually nonmagnetic samples and its occurrence is characteristic of the retrograde metamorphism stage. The majority of the rocks are transformed into gneisses and schists and have accordingly high magnetic anisotropy (Figure 17b): the anisotropy varies from 1.0 to 1.44 (average 1.2) in former sedimentary rocks and from 1.02 to 1.76 (average 1.29) in igneous rocks. Some nearly isotropic samples represent either igneous rocks less affected by the stress metamorphism or appreciably serpentinized and carbonatized rocks in which magnetite formed *after* the stress. This is demonstrated in Figure 17c, where most magnetically soft rocks with $H_c < 6$ mT (i.e. containing coarse-grained multidomain magnetite) are highly anisotropic, whereas the anisotropy is markedly weaker in magnetically hard rocks ($H_c > 10$ mT).

Thus, data on magnetic properties of the Anabar Shield rocks provide an example when nonmagnetic varieties are predominant among primary magmatic and sedimentary rocks, and this relationship has largely been preserved notwithstanding substantial metamorphism. Appreciable serpentinization of olivine-bearing rocks is not a deep process and does not characterize the lower crust magnetism but it is an important source of linear magnetic anomalies above fault zones.

North China Shield [*Zhang and Piper,* 1994]. The belt of Archean-Proterozoic granulites extends along the northern margin of the North China Shield; the belt is represented in 75–80% by tonalites, trondhjemites and granodiorites transformed into gray gneisses and in 15–20% by metasedimentary rocks. These rocks are 3.8–3.3 Ga old and were subjected to the granulite metamorphism at a time of about 2.5 Ga, after which they experienced the amphibolite facies stage of metamorphism. Both types of the rocks comprise the same set of rock-forming minerals: pyroxene, biotite, feldspar and quartz. The percentage of feldspar and quartz is 40–45% in gray gneisses and 80–85% in the para-rocks. The following two cycles of the granulite metamorphism were recognized: (1) 12–14 kbar and 800–900°C, metamorphism and deformations of igneous rocks; and (2) 8 kbar and 800°C, peak of metamorphism of overlying metasediments; overall, this is evidence for the crustal thickening in a collision process. Most of the measured ortho-rock samples are weakly magnetic with $\kappa = (1.8–6.9)\times10^{-3}$ SI units and rare samples are magnetic with $\kappa = (10–65)\times10^{-3}$ SI units; the para-rocks are nonmagnetic with $\kappa = (0.5–0.9)\times10^{-3}$ SI units. The following four ores generations were recognized. (1) The earliest large (mostly 100–500 μm) grains of ilmenite and titanomagnetite decomposed into magnetite and ilmenite; the grains usually range in concentration from 1 to 10%. The authors report a Fe concentration in "titanomagnetite" (48–52% FeO) inconsistent with actual titanomagnetite (even pure ulvospinel contains 64% FeO); given a relatively low susceptibility, one may state that large ore mineral grains are dominated by ilmenite. The authors assign the crystallization of coarse-grained ilmenite and titanomagnetite to the first cycle of granulite metamorphism. (2) Fine inclusions (less than 1%) of ore mineral in pyroxene and amphibole. They appear to have formed near the peak of granulite metamorphism. (3) Rims and pseudomorphs of ores around garnet grains at the boundary with primary titanomagnetite. These formations are related to the stage of the pressure drop from 14 to 8 kbar between two cycles of metamorphism during a rapid uplift of the block. (4) Filling of cracks between and within silicate grains with fine grains of magnetite and hematite at the last stage of the block uplift.

The value of the magnetic susceptibility anisotropy varies from 1.2 to 1.4 with an average of 1.28.

Ivrea zone, Italy. The Ivrea zone is an example of a favorable combination of geological and geophysical information of the lower crust origin of a zone above which a regional magnetic anomaly is observed. This arcuate zone is represented by alternating igneous and sedimentary rocks originated under surface conditions and afterward subjected to deep metamorphism progressively intensifying in the northward direction (across the zone) from the amphibolite to granulite facies. As a result, the sequence was transformed into basic granulites and granite-bearing rocks interbedded with para-rocks (sillimanite-quartz-feldspar gneisses and amphibole-pyroxene-plagioclase granofelsites). Large bodies of ultrabasic rocks are present in the northern part of the sequence. Geological and geophysical evidence shows that the Ivrea zone is a "plate" of the lower continental crust tectonically driven to the surface (e.g. see [*Mehnert,* 1975; *Wasilewski and Fountain,* 1982]). According to data of a detailed magnetic survey, a regional magnetic anomaly fixed above the zone consists of a series of local anomalies of 0.2 to 2.0 km across [*Schwendener,* 1984]. The most intense anomalies are associated with exposures of basic granulites, amphibolites, metagabbros etc. and with faults developed in ultrabasic rocks in areas of widespread serpentinization. Exposures of ultrabasic and para-rocks are characterized by a lower anomalous field [*Schwendener,* 1984; *Wagner,* 1984]. Measurements on samples [*Wagner,* 1984] revealed a bimodal distribution of magnetic susceptibility in the basic rocks, with modes lying (1) between 10^{-4} and 10^{-3} SI units and (2) at $\kappa > 10^{-2}$ SI units. This is typical of the primary magmatic trend reflecting the magma differentiation into nonmagnetic cumulates and magnetic differentiates, which was preserved in spite of the substantial deep metamorphism of the rocks. The para-rocks are weakly magnetic.

Main ore minerals are ilmenite and magnetite [*Wasilewski and Warner,* 1988]; all of the rocks have similar ilmenite compositions ranging from zero in peridotites to 0.2–1% in metasediments and 0.1–6% in basic granulites. *Metasedimentary rocks do not contain magnetite* (we emphasize this because both basic and sedimentary rocks are transformed into granulites). Cr-Al spinel is typical of the ultrabasic rocks. All rocks include small amounts of sulfides, primarily, pyrite and pyrrhotite. Whereas ilmenite is represented by nearly regular crystals, magnetite occurs as veinlets, fine-grained inclusions between silicates, and rims around Cr-Al spinel; intergrowths of ilmenite and magnetite crystals are occasionally observed, their Lindsley-thermometer temperature of formation being below 500°C.

According to data of thermomagnetic analysis, the main (and often the only) carrier of magnetization in the basic and ultrabasic rocks is magnetite ($T_C = 565–580$°C) and rarer pyrrhotite, which is the only carrier in metasedimentary rocks. As seen from hysteresis characteristics, single-domain grains prevail in the sedimentary rocks; pseudosingle-domain and multidomain grains, in the ultrabasic rocks; and multidomain grains, in the basic rocks. The NRM origin is complicated because of the complex nature of metamorphism.

Thus, the bulk of magnetite responsible for the lower crust magnetization in the Ivrea zone weakly depends on specific features of deep metamorphism but is largely controlled by the composition and formation conditions of initially magmatic rocks (differentiation resulting in the formation of primarily nonmagnetic and primarily magnetic groups of rocks) and primarily nonmagnetic sedimentary rocks (their hydrothermal alterations giving rise to secondary, relatively low-temperature magnetite due to the recrystallization of primary ores in basic and ultrabasic rocks, whereas similar alterations did not produce magnetite in sedimentary rocks.

Superior Province, Canada [*Pilkington and Percival,* 1999; *Williams et al.,* 1986]. Studies were conducted within the Pikwitonei and Minto blocks of Archean rocks. The first area includes greenstone belts composed of norite and amphibolite gneisses, metagabbros, amphibolites, metasediments and metavolcanics, surrounded by gabbroid gneisses. Overall, acidic rocks account for 80% section, and basic and sedimentary rocks compose the remaining 20%. The rocks range in age from 2.5 to 3.1 Ga. The Late Archean metamorphism occurred at 780–889°C and 9±1 kbar, i.e. in the lowermost crust. The uplift to the Earth's surface took place 2.5–2.3 Gyr ago. Two thirds of the studied samples are nonmagnetic; the magnetic samples mostly represent acidic intrusive rocks with $\kappa = 2.3\times10^{-3} - 0.1$ SI units. According to data of thermomagnetic analysis, they contain magnetite (T_C = 540–580°C). Pyrrhotite was found in three weakly magnetic samples. The following two important results were obtained. (a) No correlation is present between the magnetic susceptibility and degree of metamorphism. (b) The susceptibility is clearly dependent on the rock composition: thus, the susceptibility of most samples of acidic rocks is an order of magnitude higher than in the intrusive, volcanic and sedimentary rocks and does not depend on bulk iron in the rocks (1.0–3.0% Fe in acidic intrusives, 6.0–11.0% Fe in basic intrusives and volcanics and 0.5–5.0% Fe in sedimentary rocks). The authors suggest that magnetite in the study rocks has a retrograde origin, formed during their cooling and possibly did not exist in the lowermost crust. However, such an interpretation is inconsistent with numerous data yielding evidence that magnetite is the main magnetic component in xenoliths of deep rocks, and the xenoliths were transported to the surface in a very short time, i.e. magnetite formed in the lowermost crust, where they were trapped. (The authors do not answer the question, why magnetite formed solely in acidic intrusive rocks, whereas all rocks of the province were involved in the process of the retrograde metamorphism). Some basic rocks preserved ilmenite lamellae from the primary decomposed titanomagnetite in which magnetite cells are replaced by amphibole, and Fe-Ti oxides are observed to be replaced by epidote and sphene. Supposedly, basic rocks contained ilmenite and titanomagnetite that were decomposed during the retrograde metamorphism. The average induced magnetization for the ancient crust section of the Pikwitonei block does not exceed 1 A/m, which is insufficient for producing regional magnetic anomalies by the lower crust (see section 2). Supposedly, the magnetization deficit is due to viscous magnetization that forms in the lower crust.

The most part of the Minto block is composed of charnockites (magmatic orthopyroxene granitoids) ranging in age from 3.0–2.9 (early tonalites) to 2.7 Ga (volcanic-sedimentary rocks). This complex is interpreted as continental margin arcs similar to belts of the Pikwitonei block. The distribution of the Minto rocks as a function of magnetic susceptibility is bimodal, with virtually nonmagnetic ($\kappa = 2 \times 10^{-4} - 10^{-3}$ SI units) and magnetic (κ = 0.01–0.2 SI units) rock groups. The nonmagnetic group includes metasediments, metavolcanics and small amounts of granites and granodiorites, and the magnetic group is represented by pyroxene granitoids, granites, granodiorites, tonalites, diorites and minor metasediments and metavolcanics. The main ore mineral of magnetic rocks is magnetite (1–5%); ilmenite is also present in most rocks. Early high-temperature large (30–100 μm) crystals of magnetite formed together with pyroxene, whereas later magnetite grains crystallized in interstitions; magnetite lamellae in pyroxenes are found. The formation temperature of coexisting magnetite-ilmenite intergrowths (400–600°C) is markedly lower than the crystallization temperature of the adjacent pyroxenes (>700°C), which is explained in term of recrystallization during cooling or during subsequent low-temperature metamorphism. The magnetite concentration in the rocks does not correlate with their bulk iron.

Variations in the magnetic susceptibility correlate with the intensity of regional magnetic anomalies, and the most intense anomalies are associated with calc-alkaline arcs. A nearly normal field is observed in areas composed of the volcanic-sedimentary sequences. The observed magnetic anomalies can be reasonably accounted for by the induced magnetization produced by upper crust sources having the susceptibility of surface rocks. The dominating role of induced magnetization is also supported by a value of the Koenigsberger ratio smaller than 1.0 in 92% samples. A satellite positive magnetic anomaly (8 nT) is observed above the Minto block [*Arkani-Hamed et al.,* 1994]. The authors do not exclude a possible contribution of viscous remanence to the anomalous field. Mineralogical geothermobarometers indicate that the granitoids crystallized 700–1000°C and 5–6 kbar, i.e. at depths of 15 to 18 km. Their petrology and petrochemistry demonstrate differentiation trends in a calc-alkaline magma from pyroxene-biotite-magnetite diorites and granodiorites to amphibole-biotite granites transformed into charnockites.

Thus, the sources of regional magnetic anomalies in the Superior Province distinguish from the majority of other regions in which the anomaly sources are associated with basic granulites of the lower crust. Magmatic charnockites usually occur as components of granulite terranes dated at the Archean [*Percival,* 1994; *Ridley,* 1992] and Proterozoic [*Newton,* 1992; *Young and Ellis,* 1991].

Lofoten and Vesteralen, Norway [*Griffin et al.,* 1978; *Schlinger,* 1985]. The region exemplifies a province of deep origin with which regional magnetic anomalies of up to 700 nT are associated. It is composed of migmatite gneisses of volcanic origin, metavolcanics and metasediments (2.8–2.7 Ga), Proterozoic metasediments and metavolcanics (2.1–1.8 Ga), and numerous intrusions (1.8–1.7 Ga). All rocks are mostly basic, and only southwestern Lofoten migmatites are more depleted in SiO_2. The rocks experienced granulite facies metamorphism and later (1.2–1.1 Ga) retrograde metamorphism in the amphibolite facies. The crust of the region is 20–25 km thick.

The magnetic susceptibility and NRM of the studied samples have the following values: $\kappa = 10^{-2} \pm 1.1 \times 10^{-2}$ SI units and NRM = 0.58±1.27 A/m in gneisses of the amphibolite facies; $\kappa = 4.8 \times 10^{-2} \pm 2.6 \times 10^{-2}$ SI units and NRM = 2.9±3.0 A/m in gneisses of the granulite facies; $\kappa = 6.7\times10^{-2} \pm 4.0\times10^{-2}$ SI units and NRM = 10.2±10.4 A/m in basic and ultrabasic rocks; $\kappa = 2.0 \times 10^{-2} \pm 1.6 \times 10^{-2}$ SI units and NRM = 1.87±1.7 A/m in intermediate intrusive rocks (mangerites), with the minimum value $\kappa = 0.5 \times 10^{-2} \pm 0.4 \times 10^{-2}$ SI units (NRM = 0.22±0.27 A/m) observed in mangerites subjected to retrograde metamorphism; and $\kappa = 5.8 \times 10^{-2} \pm 3.0 \times 10^{-2}$ SI units and NRM = 5.1±6.7 A/m in southwestern Lofoten basic rocks. All determinations average $\kappa = 3.5 \times 10^{-2}$ SI units, which agrees with an average of 3.8×10^{-2} SI units derived from regional magnetic anomalies. The Q_n average is 0.3–0.5. The major contribution to the regional magnetic anomalies is therefore made by the induced magnetization. Moreover, NRM in the majority of samples is unstable and is dominated by a viscous component.

Two distinct variation patterns are recognizable in the magnetic susceptibility: (1) it increases from intermediate to basic rocks and (2) markedly drops under the effect of retrograde metamorphism.

Ore minerals in the rocks are represented by predominant ilmenite and by magnetite; their crystals range in size from 100 to 1000 μm. Data of the thermomagnetic analysis of J_s and χ (T_C = 560–575°C) indicate that, in all of the studied rocks, magnetite is the main carrier of magnetization responsible for the magnetic field in the region. Pyrrhotite is occasionally present. Ilmenite occurs as isolated grains, intergrowths with ilmenite, and lamellae in magnetite; it is abundant in fresh mangerites, basic rocks and anorthosites but rare in gneisses. The presence of ilmenite-hematite intergrowths is interpreted as evidence for original (primary) hemoilmenite ranging in Fe_2O_3 concentration from 3 to 23%. Magnetite usually occurs as isolated grains and occasionally contains ilmenite or hematite lamellae. Magnetite, ilmenite and hematite are present as inclusions in clinopyroxene. Secondary garnet rims are usually observed around ores grains and around pyroxene, olivine etc.; this is attributed to the cooling of the rocks at high pressure after granulite metamorphism. This pattern was significantly complicated by the subsequent widespread oxidation and retrograde metamorphism. The conditions of the major stage of granulite metamorphism were P = 9–12 kbar and T = 850–950°C. Metamorphic transformations resulted in partial or even complete replacement of Fe-Ti oxides by silicates. The retrograde metamorphism proceeded with the participation of a fluid (water) and was accompanied by the replacement of ilmenite and titanohematite by sphene and by the formation of hematite, biotite, amphibole and epidote.

Synthesis of Petromagnetic Data

(A) The sections of Precambrian rock masses presently assigned to the lower continental crust are mostly represented by volcanic-sedimentary sequences and near-surface intrusive bodies of Archean and Proterozoic ages well consistent with the formation conditions of the oceanic crust. The subsequent sedimentation and magmatism accreted the crust from above, and collisional and other folding processes led to subsidence, intense deformations and deep metamorphism of the rocks. Primary magmas coeval with the crust formation in the Archean were dominated by relatively reductive conditions, close to the "silicate" zone; accordingly, ore minerals are mostly represented by ilmenites and by rarer high-Ti titanomagnetites close to ulvospinel; at early cooling stages they were heterogeneously altered, producing magnetite, the main carrier of the Archean crust magnetism. The subsidence of the Archean sequences and their metamorphism under the silicate zone conditions preserved (at least partially) the primary magmatic ilmenites and titanomagnetites.

The Archean crust was mostly nonmagnetic and relatively thin, so that no *regional* magnetic anomalies are likely to have existed in the Archean.

(B) The main patterns in the distribution of magnetic minerals in the Archean crust are as follows. (1) *Geological* pattern: sedimentary rocks are commonly nonmagnetic, whereas igneous rocks are either magnetic or nonmagnetic, depending on the tectonic setting and differentiation processes; mantle rocks are nonmagnetic. (2) *Tectonic* pattern: igneous magnetic rocks occur in extension zones (spreading etc.), and igneous nonmagnetic rocks occur in compression zones (collisional and co-folding magmatism). (3) *Magmatic* pattern: magma crystallization differentiation process proceeds in the extension zones resulting in the formation of two rock groups: nonmagnetic and weakly magnetic cumulates and magnetic products of the differentiation. This subdivision correlates well with the concentration of magnetic minerals and with some petrochemical characteristics (Figures 4 to 7). The magnetic rocks are solely primary igneous rocks of mainly basic and, more rarely, acidic and intermediate compositions. Accumulations of magnetic minerals of a different origin are very rare.

The following striking features should be noted. (a) *Invariability* of the $TiO_2/(FeO+Fe_2O_3)$ ratio patterns in rocks and titanomagnetites as subdivided into two levels: the lower level is represented by oceanic and continental rift basalts (0.2–0.1 in rocks and 0.28–0.31 in titanomagnetite) and the upper level includes *all* magnetic gabbros (0.02–0.06 in rocks and 0.06–0.12 in titanomagnetite). The first level suggests narrow variation limits of oxidation conditions in an equilibrium basaltic magma of rift zones; these conditions appear to be consistent with the thermodynamic equilibrium of the basaltic magma at depths of primary chambers (50–60 km). The second level corresponds to the conditions of island-arc magmatism and high-temperature reworking of crustal rocks under uniform oxidation conditions at depths of 5 to 25 km [*Pechersky,* 1994; *Pechersky et al.,* 1975]. (b) *Recurrence* of the formation and partition patterns of magnetic and nonmagnetic rocks from the Archean to the present time, indicating the uniformity of magmatic processes throughout the Earth's geological history.

(C) The role of metamorphism. The bulk of rocks preserves their primary division into magnetic and nonmagnetic groups in spite of deep metamorphism: sedimentary rocks usually remain nonmagnetic or weakly magnetic, although their iron content is quite sufficient for producing magnetite in appreciable concentrations; most igneous rocks remain, as before, divided into magnetic differentiates and nonmagnetic cumulates. Of course, the metamorphism make a certain contribution to the rock magnetization, but it is small compared to the primary magmatic contribution. As the lower crust cooled, the oxidation conditions were enhanced, leading to the decomposition and recrystallization of Fe-Ti oxides with the formation of magnetite, which enriched the granulites in relatively low-temperature magnetite. The *in situ* recrystallization process reflects the primary distribution of Fe-Ti oxides in Archean beds and magmatic bodies within them.

Many researchers note that the magnetization of Archean rocks increases from amphibolite to granulite facies. Actually, the retrograde metamorphism is more often *superimposed on* granulites, i.e. the magnetization *decreases* from granulites to amphibolites. Granulites are involved in a "dry", nearly isochemical process, when the mobility of iron in silicates is low. Consequently, the main process is the recrystallization of primary Fe-Ti ore minerals. Magnetic minerals in granulites are in part secondary products of decomposition of nonmagnetic ores such as ilmenites and Mg-Al-Cr ferrospinels relatively abundant in the lower crust and upper mantle. All mantle rocks, including those containing ilmenite, are nonmagnetic; i.e. ilmenite and nonmagnetic ferrospinels are transformed into magnetic minerals under more oxidative conditions than those existing in the upper mantle.

Magnetite usually forms during metamorphism with *the participation of a fluid.* Experiments show that the presence of a fluid is a necessary, but not sufficient, condition for the formation of isolated magnetite grains because the fluid should be enriched in iron. Many authors emphasize that the amount of reduced gases (H, CO, CH and others) increases with depth, the acidity of such fluid increases and they become good solvents and iron carriers. This is one of the most probable ways for the formation of a Fe-rich fluid. As a result, conditions are created that are favorable for the iron settlement in the form of magnetite and related ferrospinels.

(D) A specific feature of xenoliths is their supply from near-chamber zones with a specific regime in which a large amount of fluids is accumulated and rocks are partially melted, producing high-Ti titanomagnetites consistent with the conditions in the chamber at the entrapment time of xenoliths. Hence, the similarity between the titanomagnetite compositions in "black" pyroxenites and host basalts (x=0.6–0.65). Such enrichment of deep rocks in magnetic minerals in near-chamber zones is a local phenomenon, as is evident, for example, from the absence of magnetic anomalies along the Kurile Islands and in the volcanic areas of the Lesser Caucasus and Mongolia, where xenoliths of deep rocks are quite often magnetic.

Conclusion

Our studies and the review of worldwide data indicate that, from the Archean until the present, igneous rocks formed in extension zones under surface or near-surface conditions have been the main source of the magnetism of the crust and regional magnetic anomalies. This situation has been preserved notwithstanding metamorphism and recrystallization of magnetic minerals.

Under lower continental crust conditions, silicates do not produce new magnetic minerals (at least in amounts appreciably affecting the anomalous magnetic field). Under favorable $T-fO_2$ conditions, new magnetic minerals can form in three ways: crystallization of primary minerals from (a) melt and (b) Fe-rich fluids and (c) *in situ* recrystallization of Fe-Ti oxides in accordance with changing $T-fO_2$ conditions.

The region of stable existence of primary igneous magnetic minerals (first of all, titanomagnetites) extends to depths of 40–50 km and the region of their most favorable crystallization is no deeper than 30 km. The Curie points of such primary titanomagnetites are usually lower than 300°C; i.e. under conditions of the lower continental crust, they are nonmagnetic and cannot be a source of regional magnetic anomalies. However, as a result of their *in situ* recrystallization the primary titanomagnetites and ilmenites can produce magnetic minerals close to magnetite. In this case, rocks containing primary titanomagnetites and ilmenites are main potential sources of regional magnetic anomalies.

Acknowledgments. This work used results of the studies supported by the Russian Foundation for Basic Research, project nos. 99-05-64890 and 00-05-64431.

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(Received June 15, 2001)