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The Moho as a Magnetic Boundary

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National Aeronautics and
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THE MOHO AS A MAGNETIC BOUNDARY

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ABSTRACT

We present new magnetic data for mantle derived rocks - peridotites from St. Pauls rocks, dunite xenoliths from the Kaupulehu flow in Hawaii, as well as peridotite, dunite, and eclogite xenoliths from Roberts Victor, Dutoitspan, Kilbourne Hole, and San Carlos diatremes. The rocks are paramagnetic or very weakly ferromagnetic at room temperature. Saturation magnetization values range from 0.013 emu/gm to < 0.001 emu/gm. A review of pertinent literature dealing with analysis of the minerals in mantle xenoliths provides evidence that metals and primary Fe_3O_4 are absent, and that complex Cr, Mg. Al, Fe spinels dominate the oxide mineralogy. These spinels would be non-magnetic at mantle temperatures. The available literature dealing with crystallization of mafic and ultramafic liquids indicates that the refractory spinels rich in Cr, Mg and Al are the first to crystallize and that FeTi spinels would 'partition' to the liquid. The depletion of the mantle via partial melts to the crust ensures the non-magnetic state of the mantle. The crust/mantle boundary can be specified as a magnetic mineralogy discontinuity. Extensive serpentinization of upper mantle rocks seems unlikely. Curie depth estimates from aeromagnetic anomalies do not require a source of magnetization in the mantle. All of the available evidence supports the new magnetic results, indicating that the seismic MOHO is a magnetic boundary. The source of magnetization is in the crust and the maximum Curie isotherm depends on the magnetic mineralogy and is located at depths which vary with the regional geothermal gradient.

INTRODUCTION

One of the expected returns from the magnetometer satellite (Magsat) soon to be launched by NASA is a global magnetic anomaly map. The minimum wavelength of anomalies measured by a satellite are comparable with its elevation. Because the orbital altitudes ($\geq$ 350 km for Magsat) are large with respect to any reasonable estimate of magnetic crustal thickness, the anomalies are in effect an expression of gross regional magnetization and thickness variation in the whole of the "magnetic crust", the layer bounded by the earth's surface above and having as a lower bound some isotherm which corresponds to the Curie point of the magnetic minerals. This isotherm is called the Curie isotherm.

For interpretation, however, we are interested in lateral variation in the form of the magnetization - depth curve as a reflection of vertical lithologic variation. To accomplish this, independent information is required on the magnetization of representative suites of rocks under conditions appropriate to the deeper parts of the magnetic crust. Such a body of data does not presently exist; accordingly appropriate laboratory studies are in progress at the Goddard Space Flight Center. A first step in developing a feeling for the form of the gross magnetization - depth profile is to evaluate the relative importance of the magnetization of the crust and upper mantle. If for the sake of argument a magnetic mineral with a 500° Curie point is responsible for magnetization in crustal rocks, and the 500°C isotherm is in the mantle, but the mantle is non-magnetic, then the effective base of the 'magnetic crust' is the moho.

The Moho, a seismic boundary, which separates the crust from the mantle, exists beneath continents and ocean basins at variable depths. Observed seismic velocities (see Meissner, 1973) geochemical considerations (Ringwood, 1975), and analyses of upper mantle xenolith suites indicate that the upper mantle is peridotitic. Strong arguments can be presented in favor of the Moho as a compositional boundary (Meissner, 1973; Ringwood, 19775; Dawson, 1977). Several authors have presented crustal column stratigraphy based on studies of xenolith suites (Padovani and Carter, 1977; Dawson, 1977; McGetchin and Silver, 1972; Eggler and McCallum, 1974) and the Ivrea zone (Fountain, 1976). These models support the hypothesis that the MOHO is a chemical discontinuity, with peridotites dominantly below the moho and granulite grade metamorphics directly above the moho along with minor amphibolites, pyroxenites, etc.

In recent years, xenolith suites found in Kimberlite diatremes and alkali basalts have been recognized as samples that bracket the moho (Meyer, 1977; McGetchin and Silver, 1972; Dawson, 1977; Boyd and Nixon, 1975; Best, 1975 and others).Xenolith suites contain representative upper mantle material, and a study of them is indicative of upper mantle magnetic properties.

EXPERIMENTAL RESULTS

Several small chips of each of the mantle xenoliths listed in Table 1 were measured on a Princeton Applied Research vibrating sample magnetometer, to determine their paramagnetic susceptibility and the saturation magnetization. These samples were chosen by Dr. W. Melson of the US National Museum for our use. The replicate measurements indicate that the values listed in Table 1 are representative of the bulk sample. The paramagnetic susceptibility is a measure of the total contribution from non-magnetic spinels and iron bound in silicates or glass. The saturation magnetization is a measure of the component responsible for remanent magnetization (i.e., the ferromagnetic component). Examples of the hysteresis loops for peridotites, dunites and eclogites from San Carlos, Kilbourne Hole, Roberts Victor and Dutuoitspan Kimberlite diatremes, peridotites from St. Pauls rocks, and dunites from the Kaupulehu flow in Hawaii are shown in Figure 1. The dominance of paramagnetism indicated by the small ferromagnetic component and the linear extension of the loop, is evident. Saturation magnetization values for the xenoliths range from 0.013 emu/gm to < 0.001 emu/gm. This is equivalent to about .001% to .01% Fe_3O_4. Serpentinized ultramafic rocks studied by Lienert and Wasilewski (1979) have I_s values ranging from 0.048 to 1.39 emu/gm. These results clearly demonstrate that the ultramafic xenoliths are non-magnetic or very slightly magnetic at room temperature. Details of the room temperature, cryogenic temperature, and high temperature magnetic studies and the evaluation of the specific mineral contributions to paramagnetism and ferromagnetism in these rocks will be presented elsewhere (Wasilewski and Melson, in preparation).

In the next two sections we present evidence that Fe_3O_4 and metals are not present in upper mantle rocks, and that the Moho appears to be a magnetic mineralogy discontinuity with complex non-magnetic spinels in upper mantle rocks and the magnetic FeTi spinels in the crust.

MAGNETIC MINERALOGY

The literature has been reviewed, specifically to determine which oxides, sulphides, and metals might be present in ultramafic mantle xenoliths (Meyer, 1977; Smith and Levy, 1974; Best, 1974; Boyd and Nixon, 1975; Haggerty, 1976; and others). Oxides found in these xenoliths are complex spinels rich in Cr, Mg, and Al which plot within the prism (Figure 2) having corners which are specific end members of the various relevant solid solution series, and magnesian ilmenite. The magnetic transition temperatures are indicated. The spinels have significantly reduced magnetization values and Curie points less than $\sim 300^{\circ}C$. Metal has not been described and Fe_3O_4 when found in xenoliths appears to be due to an apparent crustal alteration (Frisch, 1975; Evans and Frost, 1975; Bliss and McLean, 1975).

The reasons for the spinel compositions observed are suggested by experimental petrology studies concerning the crystallization in basaltic melts. Chromite crystallizes early with Mg and Al in solid solution. In essence the early formed spinels are enriched in Cr, Mg and Al relative to any later formed spinel crystals (Fe and Ti enrichment takes place in the liquid) and further cooling produces no spinel since Cr goes to the silicates, whereafter further cooling produces titanomagnetite (Hill and Roeder, 1974) which would be expected to be absent from a partially depleted upper mantle. The above results are consistent with those of (a) Mori and Green (1978) on laboratory duplication of phase equilibria in natural garnet lherzolites at mantle pressures (30-40 kbar) and temperatures (950-1500°C), and (b) by Gibb (1971) and Gibb and Henderson (1971) with ultrabasic dike rocks. They observe magnesian ilmenites and chromites. It follows from the work of Irvine (1967) and Hill and

Roeder (1974), that partial melting of ultramafic rocks will partition Fe and Ti to early liquids with the more refractory Mg, Cr, Al enriched spinels remaining as crystalline phases.

Sulphides are non-magnetic above $\sim$ 320°C. There does not appear to be any desulfurization leading to Fe or FeNi alloys in any of the xenolith suites studied (Meyer and Boctor, 1975; Vakhrushev and Sobolev, 1973; Desborough and Czmanske, 1973).

Moody (1976), summarizing available pressure-temperature-oxygen fugacity conditions for serpentinization concludes that the process will take place in the middle to upper parts of the crust. In none of the mantle xenoliths or granulite facies rocks is there real evidence of extensive serpentinization (Padovani and Carter, 1977; McGetchin and Silver, 1972; Fountain, 1976; Dawson, 1977).

It is apparant that the crust/mantle transition is a magnetic mineralogy transition, with the refractory (non-magnetic) spinels in the mantle and FeTi spinels in the crust. The potential magnetic state of crustal rocks is initially a function of rock composition and thereafter is related to reequilibration conditions at specific temperatures and oxygen fugacity, as will be described below.

MAGNETIC ANOMALIES AND CRUSTAL MAGNETIZATION

Recent interpretations of long wavelength anomalies consider Curie isotherm depth, relation of Curie isotherm depth to Moho depth and source of magnetization within the crust (Shuey et al., 1973, 1977; Hall, 1974, 1968; Byerly and Stolt, 1977; Krutikovskaya and Pashkevich, 1977; Green, 1976; Smith et al., 1977). The Curie isotherm depth has in the past been considered the position of the Fe_3O_4 Curie point (580°C) isotherm whose depth may vary from $\sim$ 20 km (Basin and Range) to $\sim$ 70 km (Sierra Nevada) depending on the geothermal gradient. Most thermal models for stable regions place the 550°C isotherm near to well below the Moho (Hyndnam et al., 1968; Blackwell, 1971; McGetchin and Silver, 1972; Smithson and Decker, 1974; Rao and Jessup, 1975). Recent studies accept the reasonable conclusions in the work of Buddington and Lindsley (1964) and consider that the Curie isotherm varies between $\sim 500^{\circ}C$ and $\sim 550^{\circ}C$ based on the observations that titanomagnetites in plutonic rocks and high grade metamorphic rocks have experienced oxidation exsolution.

Hall's (1974) analyses lead him to conclude that deep crustal magnetization is the source for long wavelength anomalies, though no specific depths are easily assigned. Krutikhovskaya and Pashkevich (1977) consider the lower crust to have a magnetization 5-10 times higher than the upper crust, a result consistent with Hall's (1974) estimate. Byerly and Stolt (1977) examined the northern border of the Basin and Range province in Arizona and correlate anomalous Curie depths with Moho depth and seismic velocities in this transition region.

All of the information based on analyses of medium to long wavelength anomalies suggests that (a) estimates to bottom of deepest sources on continents do not require sources within the mantle, (b) lateral variations in magnetization may be due to petrologic factors or changes in Curie isotherm topology and (c) the lower crust may be more magnetic than the upper crust, and (d) the upper few kilometers make no significant regional contribution. These points will be considered in the next section.

MAGNETIZATION IN THE CRUST

The earth's crust is heterogeneous, both laterally and vertically and at present we cannot locate magnetization sources in the crust with any degree of confidence. If we consider that exposures of high grade metamorphic rocks are similar to rocks in the deep crust (Smithson, 1978) then application of the Buddington-Lindsley (1964) model forces us to accept a maximum Curie isotherm of $\sim 500^{\circ}$ to $\sim 550^{\circ}C$. Based on the work of Padovani and Carter (1977) there is an indication of relatively reducing conditions in some parts of the lower crust. However, in a more hydrous environment such as described by McGetchin and Silver (1972) more oxidizing conditions could prevail in the lower crust. Buddington and Lindsley (1964) indicate that only small amounts of H_2O are required to effectively oxidize a titanomagnetite and that most of Fe_2TiO_4 will be oxidized in many rocks. The Fe_2TiO_4 content of the spinel solid solution is systematically limited for each rock type. The more basic the rock, generally the higher the initial Fe_2TiO_4 content in the spinel solid solution.

Serpentinization has been considered to be an important process in producing metals with Curie points greater than 580°C (Haggerty, 1978). These metals if present would effectively lower the Curie isotherm, making the crust everywhere magnetic, and placing the Curie isotherm everywhere in the upper mantle. Haggerty (1978) on the one hand states that there is evidence that stoichiometric Fe_3O_4 is relatively rare in the crust and then invokes serpentinization as a source of metals. However, magnetic (Wasilewksi and Lienert, 1979; Lienert and Wasilewski, 1979; Cox et al., 1969; Saad, 1965) and mineralogical summaries (Eckstrand, 1975; Moody, 1976; Wenner and Taylor, 1971; Wicks and Whittaker, 1977) clearly document that Fe_3O_4 is the dominant magnetic phase in serpentinites.

We can make generalizations, derived from medium to long wavelength anomaly interpretation, in terms of rock magnetization. Rocks in the upper few kilometers of the crust, including the 'crystalline basement', may contain an 'original' remanent magnetization modified by thermal overprints. These vectors may be highly variable. Hall (1968) noted very little regional significance to surface sample magnetization. As depth increases the conditions for coherent regional magnetization are enhanced. The remanent vector is diminished with increasing temperature, and thermal overprints found near the surface are eliminated. The viscous magnetization is enhanced (Shimizu, 1961). Initial susceptibility increases with increasing temperature particularly in the region within 100°C-150°C of the Curie point. These aspects of rock magnetization can probably account for anomalous lower crust magnetization. The thickness of crust between the 400°C and 550°C isotherms may be 5 to 20 km depending on the steepness of the geothermal gradient. This anomalous layer would be thickest under shield areas, and this is where Hall (1968, 1974) and Krutikhovskaya and Pashkevich (1977) have concluded it might be present. In essence the deeper a magnetic rock is in the crust, the greater is the component of magnetization, due to viscosity and induced magnetization (due to enhanced initial susceptibility), along the present geomagnetic field. The deep crustal magnetization variations are then related to petrologic variations and the topology of the Curie isotherm.

CONCLUSION

Mantle xenoliths have low to negligible levels of magnetization because their magnetization is due to complex Cr, Mg, Al, Fe spinels and magnesian ilmenite and titanomagnelites and metal are absent. The xenoliths are geographically well distributed, and petrographic and microprobe analyses of worldwide mantle xenoliths confirm the presence of the dominantly non-magnetic mineralogy. We refer to the common mantle xenolith mineralogy as non=magnetic because even if they are slightly magnetic at room temperature, they would be non-magnetic at mantle temperatures no matter what the geothermal gradient. This is because substitution of Cr, Mg and Al into the spinel lattice lowers the Curie point. There are good reasons for the existence of these 'refractory' spinels in the upper mantle, and the FeTi spinels are partitioned to partial melts, e.g., crustal gabbroic rocks etc., making the moho a magnetic mineralogy transition.

The initial magnetization of rocks in the crust is related to rock type, which determines the original composition of the titanomagnetite solid solution, and thereafter to the oxygen fugacity conditions at specific temperatures of reequilibration. At the present time it is not immediately apparent where in the crust the magnetization sources are located. We have explained, however, that at depth all magnetization should be directed along the present geomagnetic field, and a 5-20 km thick region of the lower crust may have anomalously high magnetization because of the effect of temperature on rock magnetization. To develop a model for the magnetic structure of the earth's crust which will configure the source locations requires a body of currently non-existent data. We require magnetic data on rocks: from the lower crust, from high grade metamorphic terrain, and from specific crustal sections which expose portions of the lower to middle crust. We need to acquire data on rock magnetization at pressures and temperatures appropriate to crustal conditions. Time is an important consideration as none of the crustal rocks contributing to long wavelength anomalies have been quenched and temperature enhances viscous magnetization.

The magnetic data once acquired must then be appropriately synthesized into a framwork compatible with current 'high resolution' geophysical (e.g., Mueller, 1977; Proedhl, 1977; Smithson, 1978, Fountain, 1976) and geochemical (e.g., Padovani and Carter, 1977; McGetchin and Silver, 1972; Dawson, 1977) results which point to a complex crust with vertical and lateral heterogeneity. These measurements are currently being made at the Goddard Space Flight Center magnetics laboratory.

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Table 1. Saturation Magnetization and Paramagnetic Susceptibility for Upper Mantle Xenoliths

USNM #	LOCATION	ROCK	I_s[1]	xp[2]
110596	Roberts Victor	Phlogopite Eclogite	*	9.7
110608	Roberts Victor	Phlogopite Eclogite	*	9.6
110602	Dutoitspan	Peridotite	*	7.5
110597	Dutoitspan	Garnet Peridotite	*	73.60
114026/40	Kilbourne Hole	Augite-Hernblenoe Nodule	.008	12.5
114026/13	Kilbourne Hole	Peridotite	*	11.17
114027/13	San Carlos	Brown Peridotite	*	25.7
114027/6	San Carlos	Peridotite	.008	12.7
110391/10	St Pauls Rocks	Peridotite Mylonite	.002	7.5
110393/10	St Pauls Rocks	Brown Hornblende Mylonite	.005	8.5
111120-2	Hawaii	Olivine Nodule	*	3.52
111807	Hawaii	Olivine Nodule	.013	1.76
113095-8	Hawaii (Kaupulehu Flow)	Olivine-Pyroxene-Anorthite Nodule	.007	7.23

* < .001 emu/gm

1 emu/gm

2 10^{-6} emu/gm

Table 2. Schematic for Lower Crust/Upper Mantel Magnetic Mineralogy

ROCKS	Magnetic Mineralogy	Magnetic State
Granite gnesis Granulite grade metamorphics	FeTi spinels with composition dependent on rock type	Initial Curie points depends on rock type (i.e., titanomagnetite composition)
Amphibolites pyroxenites (Crust)	If serpentinization then Fe_3O_4	Subsolidus oxidation exsolution gives a Curie point of 500°C -550°C If Fe_3O_4 then Curie point is $\sim 580^\circ$
(Upper Mantle) Peridotite (Dunite Lherzolite Harzburgite) eclogite	Complex spinels (rich in Cr, Mg, Al) Ilmenite (with Mg)	Curie points < 400°C small saturation magnetization, remanence, and initial susceptibility

Crystallization of Basaltic Liquid (Hill and Roeder, 1974)

falling temperature

$$\text{Chromite (with Mg+Al)} \rightarrow \text{Enrichment in Fe+Ti} \rightarrow \text{no spinel} \rightarrow \text{titanomagnetite}$$

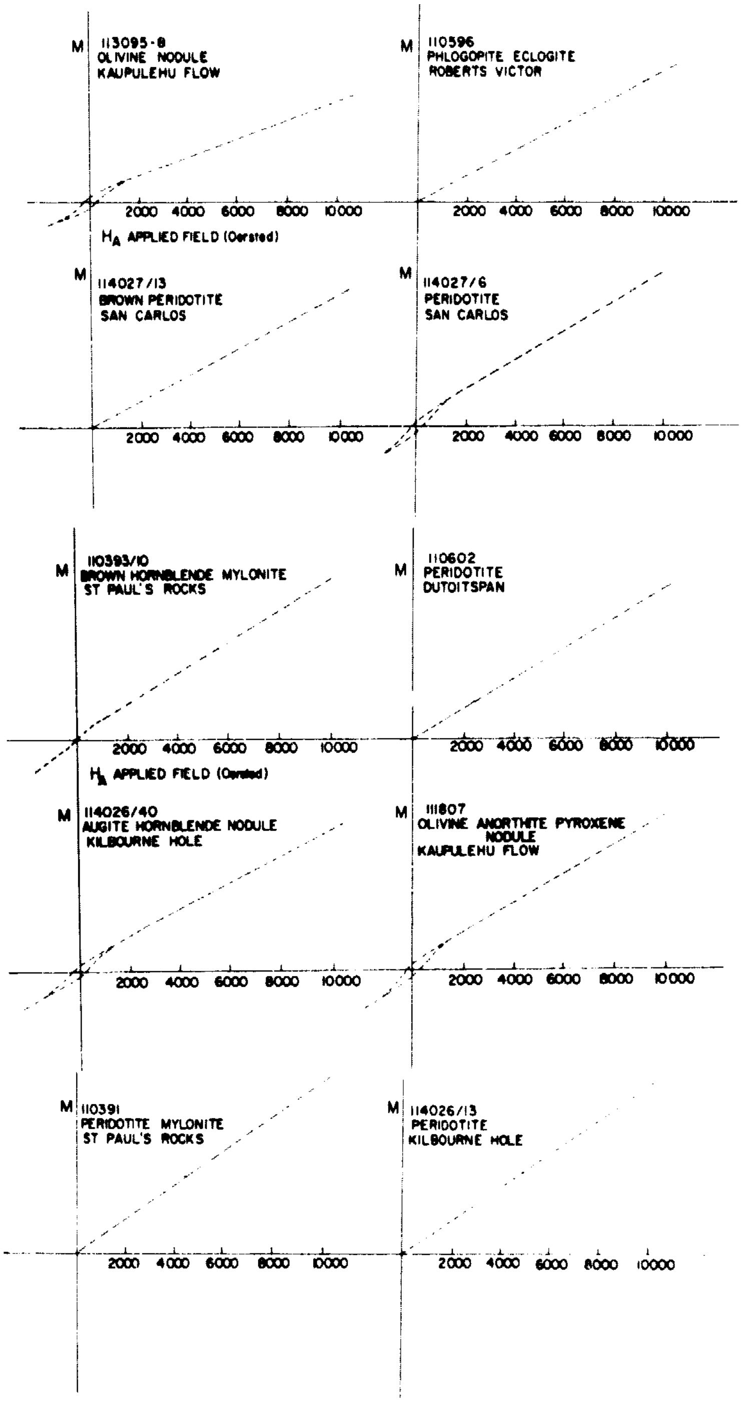

Fig. 1. Magnetization (M) vs field (H) curves (hysteresis loops) for mantle xenoliths and St. Pauls rocks peridotite. Values of saturation magnetization and paramagnetic susceptibility are given in Table 1.

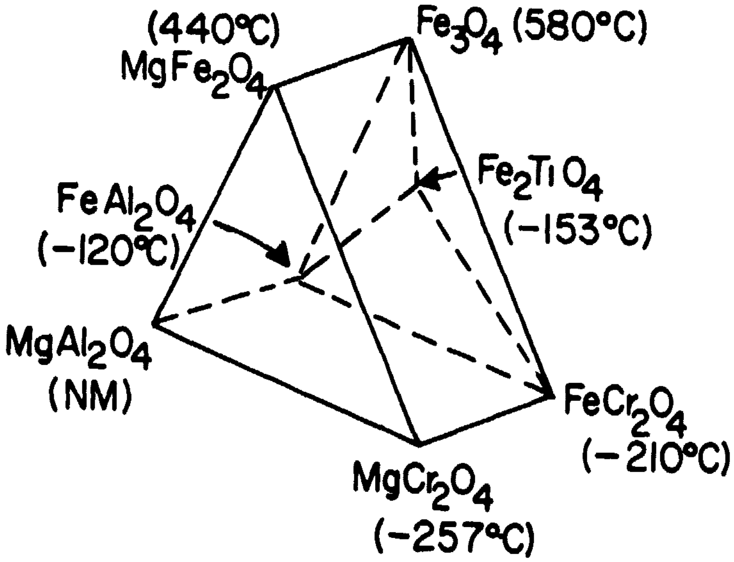

Fig. 2. Multicomponent spinel prism. Magnetic transition temperatures are indicated. Only Fe_3O_4 and $MgFe_2O_4$ are ferromagnetic at room temperature. Spinels in mantle xenoliths fall within this prism.