THE MAGNETIC PROPERTIES OF METAPERIDOTITIC ROCKS AS A FUNCTION OF
METAMORPHIC GRADE: IMPLICATIONS FOR CRUSTAL MAGNETIC ANOMALIES

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Abstract. Seventeen samples from the Malenco serpentinite in the Swiss Alps, representing systematic prograde metamorphic conditions from prehnite-pumpellyite to upper amphibolite, are moderately to strongly magnetic (average susceptibility of 6.61×10^{-2} SI units) due predominantly to the presence of a magnetite-rich spinel. Some greenschist facies samples contain the metal alloy awaruite (Ni_3Fe); these contain in addition much greater amounts of magnetite. Small amounts of pyrrhotite in about half the samples complete the contributions to the total magnetization. Susceptibilities generally decrease with increasing metamorphic grade because of the production of increasing amounts of chrome-rich spinel which dilutes the magnetite component. These results indicate that significant magnetic anomalies should be generated in suture zones containing metamorphosed serpentinite and in areas where serpentinite is being actively subducted.

Introduction

In addition to being the main constituent of the upper mantle, peridotitic rocks and their metamorphic equivalents make up a significant portion of the Earth's crust. On the continents, such rocks are found as ophiolites in the major tectonic belts; in oceanic environments they are found in upper crustal levels in transform faults. Hydrothermal alteration of peridotites may produce zeolite to amphibolite facies mineralogies. Thus understanding the way that the magnetic properties of metaperidotitic rocks vary as a function of metamorphism is important in interpreting crustal magnetic anomalies. This is particularly true because there has been speculation that Fe-Ni alloys, which do occur frequently in serpentinites, can be a source of magnetic anomalies in the upper mantle [Haggerty, 1978]. For these reasons we have undertaken a study of how the magnetic properties of metaperidotites vary as a function of metamorphic grade.

Existing magnetic studies of serpentinitized ultramafics [e.g., Burch, 1968, Komarov, 1965; Saad, 1969a, b; Beske-Diehl and Banerjee, 1979; Leinert and Wasilewski, 1979; Wynn and Hasbrouk,

1984] show that serpentinites can be strongly magnetic and that susceptibility increases and the Konigsburger ratio Q decreases in a fairly linear relationship with degree of serpentinization. None of these studies was concerned with the systematic effects of prograde metamorphism on the magnetic properties of an already serpentinized body; thus they are most useful in upper crustal applications such as magnetic modeling of exposed ophiolite complexes and of the oceanic crust. Our concern is to characterize the petrologic and magnetic changes accompanying metamorphism of serpentinites deep in the crust.

The way in which the oxide mineralogy of peridotitic rocks varies as a function of metamorphic grade is fairly well-known [see Evans and Frost, 1975]. In primary mantle peridotites the sole oxide is usually a chrome spinel in which Fe^{3+} makes up only around 10% of the trivalent ions [Irvine, 1965]. Such spinels commonly contain substantial amounts of MgO and Al_2O_3 along with chrome and iron. A major product of serpentinization is magnetite. Page [1967] reports up to 15% magnetite in some serpentinites. The magnetite in serpentinite is formed both by oxidation of FeO in olivine and by alteration of the primary spinels. Much of the primary spinel commonly survives serpentinization and is found in serpentinites as brown grains with an opaque rim [see Beeson and Jackson, 1969; Onyeagocha, 1974; Bliss and MacLean, 1975]. The opaque rims on such grains are termed ferritchromite and are essentially solid solutions of FeFe_2O_4 and FeCr_2O_4 . Unlike primary spinels they contain very little Al or Mg.

The changes in oxide composition during metamorphism of serpentinites are summarized below [after Evans and Frost, 1975]. In antigorite serpentinites the oxides are ferritchromites in which the magnetite percentage ranges from 100% to 30%. Even within individual samples the spinels characteristically show wide ranges in composition. With the breakdown of antigorite the spinels become markedly more chromiferous; pure magnetite has not been recorded as a stable spinel in rocks which do not have serpentine. Finally, at the highest grades corresponding to middle or upper amphibolite facies, the ferritchromites become progressively more aluminous as chlorite begins to break down. The disappearance of chlorite in uppermost amphibolite facies leads to the production of Mg and Al bearing spinels that are indistinguishable from those of a primary mantle peridotite.

Metaperidotites also may contain two other potential magnetic minerals, pyrrhotite and iron-

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nickel alloys. The factors stabilizing pyrrhotite are not well constrained. It is recorded primarily in lower grade portions of the Malenco serpentinite [Trommsdorff and Evans, 1972] and in greenschist or lower grade serpentinites of the Oberhalbstein [Dietrich, 1972]. Native metals, including both native iron and iron-nickel alloys are primarily found in partially serpentinized peridotites (see summary by Frost, [1985]) but recent work shows that FeNi alloys are also stable during prograde metamorphism in rocks with the assemblages chrysotile-antigorite-olivine and olivine-antigorite-brucite [Peretti et al., 1987].

Sample Localities and Mineral Assemblages

The rocks studied came from the Oberhalbstein (1 and 2), the Val Forno (15) and Malenco (3-14 and 16) serpentinites (Figure 1) in the Swiss and Italian Alps. These rocks are part of the ophiolitic belt that is sandwiched between the Austroalpine and Penninic Nappes. It is evident from the presence of ophicarbonates (limestone-serpentine breccias) that these rocks were once exposed as part of the ocean floor in a transform fault regime [Bernoulli and Weissert, 1985]. Circulation of seawater into these peridotitic rocks completely transformed them into chrysotile/lizardite serpentinites prior to alpine metamorphism. Alpine metamorphism associated with the emplacement of the nappes led to an increase in grade from prehnite-pumpellyite in the north to uppermost greenschist in the deepest exposed nappes of the Malenco to the south [Trommsdorff and Evans, 1974; Peretti et al., 1987]. Later contact metamorphism associated with the emplacement of the Bergell tonalite increased metamorphic grade in the eastern portion of the Malenco to upper amphibolite facies [Trommsdorff and Evans, 1972]. The highest grade rocks in the aureole have the same mineralogy as primary mantle peridotites but with metamorphic fabrics.

The samples studied cover a range of mineral assemblages from lizardite/chrysotile serpentinite to forsterite-anthophyllite (Table 1), which represents a profile in metamorphic grade from prehnite-pumpellyite to upper amphibolite facies [Evans, 1977]. A disproportionately larger number of samples was chosen from the antigorite-forsterite-brucite serpentinites in the Malenco because these rocks also contain the largest amounts of iron-nickel alloys. The lizardite/chrysotile serpentinites contain pure magnetite rims on relict primary chrome spinel. In the higher-grade rocks primary chrome spinel is absent, although relicts of the primary spinel are marked by the occurrence of ferritchromites with high Cr contents. These spinels are generally zoned with ferrite-rich rims and Cr-rich cores. Microprobe analyses of the oxides in these rocks show that compositional ranges reflect the changes documented by Evans and Frost [1975], with Cr and Al content of the spinels increasing with increasing metamorphic grade (Figure 2). This change in spinel composition is reflected in

changes in the $\text{Fe}_2\text{O}_3/\text{FeO}$ ratio of the rocks (Table 2). This ratio is greater than 1.0 for rocks with the assemblage lizardite/chrysotile or antigorite-diopside +/- brucite and is less than 1.0 for rocks containing olivine. This marked decrease in the $\text{Fe}_2\text{O}_3/\text{FeO}$ ratio is apparently associated with the reduction accompanying the formation of native metals.

Along with the oxides, the sulfide assemblages also show distinct changes with increasing grade (Table 1). The lowest grade rocks (1 and 2) contain the assemblage pentlandite-magnetite. Pyrrhotite is markedly missing, despite the fact that pyrrhotite has been reported from some samples from the Oberhalbstein [Dietrich, 1972]. We interpret this discrepancy to be due to the fact that the pyrrhotite-bearing serpentinites of Dietrich [1972] were closely related to ophicarbonates, and thus were not strictly isochemically metamorphosed serpentinites. At slightly higher grades hazelwoodite appears and then native metals. We interpret these as forming due to partial breakdown of pentlandite in response to the reducing conditions that accompany the formation of olivine [Peretti et al., 1987]. At higher metamorphic grades, within the contact aureole of the Bergell tonalite, the major opaque assemblage coexisting with a spinel of a variable range in Fe_3O_4 component is pentlandite-troilite. On the basis of textural observations we interpret that as representing high-temperature monosulfide solid solution that exsolved on cooling.

Magnetic Studies

Procedure

Susceptibility measurements were made on core specimens from all seventeen samples. The cores were 2.5 cm in diameter and 2.5 cm long. The instrument, a Bison MS-3, is a standard commercial bridge capable of resolving $\pm 2 \times 10^{-5}$ SI units for samples of this size. In addition, saturation magnetization (J_s) versus temperature determinations were performed for all samples. The thermomagnetic curves were obtained using a Curie balance similar to that described by Doell and Cox [1967]. The balance is capable of resolving a magnetic moment of about 1.5×10^{-6} Am², which corresponds to the contribution of 0.016 mg of magnetite (less than 0.01% magnetite by weight in a 300-mg sample).

All thermomagnetic runs were made under a vacuum of 1.3×10^{-2} Pa (10^{-4} torr) or less in a field of 0.290 T. Samples, consisting of chips from the cores used for susceptibility measurements, weighed between 275 and 350 mg. Samples were heated to 600°C; heating and cooling rates did not exceed 20°C per minute. Temperatures, from a Pt-Pt 13% Rh thermocouple, are accurate to $\pm 5^\circ\text{C}$. The temperature error in the thermomagnetic curve, which includes thermal inertia mismatch between the thermocouple and the sample, should be less than $\pm 10^\circ\text{C}$. Natural

Fig. 1. Index map showing the location of ophiolitic nappes including the Malenco (southeast). Numbers denote sample locations with prograde regional metamorphism (1-10) and prograde contact metamorphism (11-16).

remanent magnetizations (NRM) were measured with a Schonstedt spinner magnetometer. We report NRM values although they are not likely to be representative of remanence values of these rocks at depth [Williams et al., 1985].

SI units are used throughout. In keeping with current standard practice magnetic "fields" (actually inductions) are expressed in teslas (T), where a SI induction of 1 T corresponds to a cgs field (induction) of 10^4 oersted (gauss). Dipole

TABLE 1. Mineralogy and Locations for Malenco Samples

Sample No.	Liz	Chr	An	Bru	Ol	Dio	Cte	Tr	Tc	Ant	Pen	Hez	Aw	Cu	Cpy	Po	Swiss	Coordinates
1	D194	x	x								x						767.22	147.44
2	D645	x	x								x						765.94	153.00
3	M120	(x)	x (x)				x				x	x					786.3	134.26
4	M114	(x)	x			x	x				x	x					785.42	134.07
5a	Sp3	(x)	x		(x)	x	x				x	x	x	x			799.9	131.1
5b	Sp4	(x)	x (x)	(x)	(x)	(x)	(x)				x		x				Same	
6	To3		x	x	x	x	x				x		x	x			789.725	128.5
7	F8		x (x)	x	x	x					x		x				789.625	129.25
8	Mg159		x	x	x	(x)					x		x				790.0	129.4
9	M140		x		x	x					x	x					783.00	126.37
10	Mg123a		x		x	x	x				x	x	x				777.25	125.5
11	Ma95		x		x		x	x	x		x		(x)				778.02	126.85
12	Mg73a		x		x		x	x			x					x	775.9	122.2
13	Mg71a				x		x		x		x					x	775.3	122.1
14	Mg101				x		x	x	x		x				x	x	778.9	129.7
15	160T				x	x	x	x		x	x				x	x	773.85	136.56
16	Ma9b				x		x	x		x	x				x	x	774.3	121.64

An x indicates mineral present as a major phase, and (x) indicates mineral present as a trace phase. Mineral abbreviations: an, antigorite, ant, anthophyllite, aw, awaruite, bru, brucite, chr, chrysotile, cpy, chalcopyrite, cte, chlorite, cu, copper, dio, diopside, hez, heazlewoodite, liz, lizardite, ol, olivine, pen, pentlandite, po, pyrrhotite, tc, talc, tr, tremolite. Secondary sample designations refer to ETH catalogs: D samples are from dissertation of V. Dietrich, Ma samples are from diploma thesis of M. Engi, Mg samples are from research collection of V. Trommsdorff, and all others are from collection of A. Peretti. All samples are stored at the Institut für Mineralogy and Petrology, ETH.

Fig. 2. Variation of spinel composition as a function of metamorphic grade for samples studied in this paper. Metamorphic grade increases upward. In low grade serpentinites, P is the primary spinel, and FC is the ferritchromite margin of primary spinels. Other abbreviations are ant, antigorite, anth, anthophyllite, bru, brucite, chry, chrysotile, di, diopside, ol, olivine, tc, talc, and tr, tremolite. Facies designation from Evans [1977].

moments per unit mass are expressed in units of Am^2/kg , where $1 \text{ Am}^2/\text{kg}$ corresponds to $1 \text{ gauss-cm}^3/\text{g}$. A dipole moment per unit volume of 1 A/m corresponds to a cgs magnetization of 10^{-3} emu/cm^3 . Finally, a susceptibility of 1 (SI) is equivalent to $1/4\pi$ cgs units.

Results

The results of the magnetic measurements are shown in Table 2. Susceptibility measurements range between 1.38×10^{-3} and 234×10^{-3} SI units. The average is 66.1×10^{-3} with a standard deviation of 14.8×10^{-3} . Sample susceptibility values represent averages of values from at least two, sometimes three separate cores. Measurements from different cores showed typical variations of about 10%, indicating that the magnetic properties are fairly homogeneous for core-sized samples. Natural remanent magnetization values range from 2.6 to 341 A/m. They show no consistent relationship to susceptibility and, like the susceptibility values, are generally smaller for the most highly metamorphosed samples (12-16).

Saturation magnetization values and Curie

temperatures were determined from the thermomagnetic work. Saturation magnetization values range between 0.18 and $4.86 \text{ Am}^2/\text{kg}$. The average is 1.89, with a standard deviation of 0.29. Duplicate measurements were made on three samples. In two cases the values differed by about 20%, in the third by a factor of two. Within-sample variation of saturation magnetization is larger than that of susceptibility because of the smaller sample size (300 mg versus about 30).

Examples of thermomagnetic curves are shown in Figures 3a-d. All samples have at least one Curie temperature above 500°C . The maximum Curie temperature is at or above 560°C for most samples, but 11, 12, and 15 have values of 530°C or below. Samples 14 and 16 have a clearly defined second (lower) Curie temperature in the range 520° to 525°C . Most of the samples show evidence of magnetization from a phase with a Curie temperature of about 320°C ; usually this disappears during cooling.

Most samples lost a few percent of their magnetization during the run. This is attributable in most cases to the disappearance of the phase with $T_C \sim 320^\circ\text{C}$. Except for that, the runs were quite reversible. Differences in heating and cooling behavior below 350°C are due partly to thermal inertia mismatch between sample and thermocouple.

Discussion

As a group, the samples from the Malenco Serpentinite may be described as strongly magnetic. Their susceptibilities are several times higher than average values for lower crustal cross sections from Italy [Wasilewski and Fountain, 1982], from Norway [Schlinger, 1985], and from Canada [Williams et al., 1985; Shive and Fountain, 1988]. Their NRM values, although quite variable, are also high. In contrast to typical lower crustal rocks, they are thoroughly capable of meeting the source requirements for lower crustal magnetization deduced from aeromagnetic and satellite magnetometer studies [Hall, 1974; Coles and Currie, 1977; Schnetzler and Allenby, 1983]. They are, in fact, the source of a strong magnetic anomaly [Fischer et al., 1979].

How would these magnetizations be affected if these rocks were in the lower crust, at temperatures of several hundred degrees? To answer this question we must identify the phases responsible for the magnetization. Electron microprobe studies identified magnetite in all samples, but three other magnetic phases were also observed in some samples. These are pyrrhotite (Fe_{1-x}S), awaruite (Ni_3Fe), and chrome ferrite ($\text{Fe}_{3-x}\text{Cr}_x\text{O}_4$). Nickel ferrite (NiFe_2O_4) is another phase that might be expected, but it is not found in metaperidotites due to their reducing nature. NiO has not been found in peridotitic spinels in abundances greater than one percent [Evans and Frost, 1975]. Pentlandite and hazelwoodite are not magnetic at room temperature.

TABLE 2. Magnetic and Petrologic Data for Malenco Samples

Sample NO.	K	J _{NRM}	J _S	T _C	Fe	Fe ₂ O ₃ /FeO	Ni	Cr
1	6.31	113	1.82	565	5.64	1.85	4025	1720
2	1.33	3.3	3.45	(570)	5.42	1.10	3950	1770
3	8.12	39	2.27	(560)	5.16	1.66	4750	1240
4	7.80	11	.87	560	6.28	1.12	3450	2350
5a	19.2	29	2.01	565	6.87	1.09	1100	1890
5b	23.5	25	0.94	575	7.38	1.48	1236	1935
6	4.32	341	4.21	565	5.44	0.48	4650	2040
7	5.62	27	1.66	(575)	5.27	0.90	4400	1990
8	5.01	7.2	4.86	570	4.74	0.34	4750	1085
9	6.13	24	1.48	530	5.64	0.58	4025	2170
10	8.29	84	2.09	565	4.35	0.93	3980	1940
11	5.34	24	1.22	(525)	5.23	0.62	4400	1950
12	1.08	4.7	1.74	(530)	5.86	0.32	3980	1660
13	0.14	59	0.18	(575)	5.00	0.29	4630	1860
14	9.32	8.2	2.37	570, (520)	5.24	0.52	4400	2310
15	0.69	8.4	2.60	525	5.42		3900	1980
16	0.57	2.6	0.18	575, 525	5.85	0.27	4400	2330

Samples are arranged in order of increasing metamorphic grade. J_S is given in Am²/kg; J_{NRM} in A/m; K in SI x 10⁻²; T_C in °C (parens denote inflection at 320°C on heating), 14 and 16 have two distinct Curie points above 500°C; Fe in wt %; Ni and Cr in ppm.

We use a combination of thermomagnetic and petrologic information to assess the relative magnetic contributions of the different phases. Much of this information is shown in Table 2, which lists Curie temperatures as well as contents of iron, nickel, and chrome.

Pyrrhotite has a Curie temperature of 320°C [Nagata, 1961, p. 120]. Small amounts of pyrrhotite are responsible for the small inflections on the heating cycle of the thermomagnetic curves of about half the samples (e.g., Figure 3d). The magnetic contribution of pyrrhotite is not significant. At room temperature it amounts to less than 10% of the total magnetization; this contribution will decrease rapidly and disappear at the 320°C isotherm in the crust. The existence of pyrrhotite is surprising, as hazelwoodite and troilite are the only iron sulfides detected by the microprobe. A small amount of iron sulfide is nonstoichiometric. Because its appearance is apparently random in samples of all metamorphic grade, it may be a weathering or late stage product.

Awaruite is a member of the solid solution of nickel-iron alloys which has been observed in other ultramafic bodies [Dick, 1974; Leinert and Wasilewski, 1979]. It has a saturation magnetization of about 115 Am²/kg, and its Curie temperature is reported by various authors [Leech and Sykes, 1939; Bozorth, 1951; Kneller, 1962; Haggerty, 1978] between 590°C and 620°C. Bozorth [1951] claims that the Curie temperature can vary by about 10°C depending on the thermal history of the sample.

We see no evidence from any thermomagnetic curve for a magnetic contribution from a phase with Curie temperature between 590°C and 620°C. We note, however, that 620°C is the maximum Curie temperature in the nickel rich portion of the Fe-Ni series. Because of this, nonstoichiometric departures from Ni₃Fe could lower the Curie temperature and allow the alloy to thermomagnetically resemble magnetite. To evaluate this possibility, we consider the six samples (5a, 5b, 6, 7, 8, and 10) in which awaruite was detected by the microprobe (and sometimes in hand sample). If we assume that all the nickel in these samples is in the form of awaruite, the resulting saturation magnetizations would be between 0.16 and 0.68 emu/g. This amounts to a maximum possible awaruite magnetization contribution between 8 and 42% of the total. Microprobe analyses also suggest that at least half the nickel in any sample is in silicate minerals, which reduces the Ni₃Fe contribution to 4-21% of the total. A more specific determination, in the form of a microprobe point count, was made for the sample which had the highest metal alloy content in thin section (8). In this sample magnetite was more than ten times as abundant as awaruite, a relationship that is consistent with optical estimates. Taken together these analyses indicate that magnetite should account for at least 90% of the magnetization in samples that contain metal alloys.

Finally, we made a separate thermomagnetic run from sample 8, using a much higher sensitivity scale and concentrating on the region above

Fig. 3. Representative thermomagnetic runs, (a) No. 8, showing single Curie point at 570°C. (b) No. 15 showing Curie point at 525°C. (c) No. 16, showing Curie points at 525°C and 575°C. (d) No. 7, showing disappearance of phase with Curie temperature of 320°C on heating.

570°C. We did detect another thermomagnetic phase with $T_C \approx 600^\circ\text{C}$. Its contribution to the magnetization of sample 8 is only about 5% as great as that of the magnetite. Thus all lines of evidence indicate that although awaruite does occur in several of our samples, it is always accompanied by far greater amounts of magnetite.

It is possible that awaruite could have a significantly greater influence on the remanent magnetic properties of these samples. If its microscopic coercivity were far less than that of the magnetite, it could carry a soft remanence prone to the acquisition of viscous remanence in the laboratory. Even more serious, the presence of two phases with similar Curie temperatures could lead to self reversal by Néel's [1951] two-constituent model. Paleomagnetic studies on rocks of this type should proceed with caution.

The chrome ferrites ($\text{Fe}_{3-x}\text{Cr}_x\text{O}_x$) have Curie points below room temperature for $x > 1.3$ [Francombe, 1957; Robbins et al., 1971]. In the range of interest to us ($0 < x < 1.3$) the Curie temperature varies linearly from 575°C to room temperature, and the saturation magnetization from 90 to about 45 Am^2/kg [Derbyshire and Yearian, 1958; Robbins et al., 1971; Schmidbauer, 1969; 1971]. Low temperature oxidation of chrome ferrites can lower the magnetization and raise the Curie temperature [Schmidbauer, 1971] but this is not likely in rocks containing metal alloys.

Chrome ferrites would produce lower Curie temperatures. Five samples (11, 12, 14, 15, and 16) have Curie temperatures of 530°C and below.

This is consistent with mineralogical analyses, which shows that these samples are higher grade and contain ferrites that are richer in chrome (Figure 2). However, individual samples show inconsistencies when thermomagnetic curves are compared with microprobe analyses. For example, the chrome ferrite content of 13 is in the range $0.4 < x < 0.85$, which should have Curie temperatures between 400°C and 160°C rather than the observed 575°C. In every case the inconsistency has a similar sense; the microprobe analysis shows a composition richer in chrome than that based on Curie temperature (which is always close to magnetite).

We suggest two possible explanations. One is that the samples contain fine grained magnetite with some impurities which escapes the probe analysis. Alternately, compositions that appear to represent homogeneous intermediate chrome ferrites to the probe may in fact be exsolved on an extremely fine scale into magnetite- and chromite-rich intergrowths. Since there is no experimental work that substantiates a chromite-ferrite solubility break we suppose that the former explanation is correct. Genetically, the presence of magnetite in these high grade samples could result from the survival of small magnetite grains that were armored by olivine and protected from reaction with antigorite or chlorite. Alternately, it may have formed from late stage reserpentinization of samples during weathering.

In any case, the magnetic properties of these samples appear to result from contributions of

minerals with compositions close to magnetite. This is consistent with conclusions about magnetic sources in more typical lower crustal rocks [Schlinger, 1985; Williams et al., 1985; Frost and Shive, 1986; Shive and Fountain, 1988]. It is noteworthy for two reasons. First, the Malenco samples are far more efficient at making magnetite than typical mafic rocks from crustal sections. This is shown by Figure 4, which is a plot of susceptibility versus iron content. Susceptibilities fall much closer to the upper magnetite curve than for typical crustal sections (compare with Figure 5 of Williams et al., [1985] and Fig. 4 of Shive and Fountain, [1988]). Secondly, the lack of strong magnetic contributions from metallic phases in these samples strengthens earlier conclusions that magnetite-rich compositions govern magnetic behavior in the deep crust.

Conclusions

This work provides the background for a magnetic model of ultramafic rocks exposed to metamorphism. Initially, peridotites should be paramagnetic because they contain no metals and because the stable spinel is chrome-rich. This is consistent with data of Wasilewski and Mayhew [1982], who found fresh mantle nodules to be paramagnetic. Serpentinization produces large amounts of magnetite [Page, 1967]; Saad [1969a] showed that the susceptibility of serpentinite is directly related to the degree of serpentinization. We have shown that magnetite remains the stable phase at metamorphic grades below that of amphibolite facies, although minor amounts of metal alloys may contribute to magnetization in this range. Higher grades of metamorphism tend to deserpentinize the rock. Metals disappear, and chromite is again the stable spinel phase.

We would thus predict that during increasing metamorphism the magnetic susceptibility of metaperidotites would follow the trend shown by the arrow in Figure 4. Susceptibility should decrease with increasing grade, with the highest-grade metaperidotites being paramagnetic. In a rough manner the rocks measured in this study follow this pattern, with the exception that the highest-grade rocks are somewhat more strongly magnetic than we would predict. Part of this "excess" magnetization may have been produced by late stage alteration of olivine to serpentine, since all high grade samples contain at least a trace of late serpentine. If this is true then equivalent rocks present at depth where they are protected from interaction with surface waters would be paramagnetic. Alternately, the magnetite in these rocks may be present as fine-grained dust included in olivine. If this is true then metaperidotites would retain a strong magnetic signature even when subjected to high-temperature metamorphic conditions where magnetite is no longer the stable spinel phase.

Regardless of which explanation is true, it is

Fig. 4. Susceptibility versus weight percent iron. The arrow shows the expected trend of susceptibility with prograde metamorphism.

evident from this work that serpentinites and their low- and medium-grade metamorphic equivalents will have a strong magnetic signature and therefore can be a source for significant magnetic anomalies. Thus we would expect that suture zones containing significant quantities of serpentinite and metamorphic equivalents would produce distinct magnetic signatures. Such anomalies are present in the Swiss Alps along the belt of ophiolites that connects the Malenco serpentinite with the ophiolites of the Zermatt - Saas area [Fischer et al., 1979] and in the Proterozoic suture of the Nelson Front in Canada [Morely et al., 1967].

The results presented here warrant reconsideration of the contention [Wasilewski et al., 1979; Frost and Shive, 1986] that the Moho is magnetic basement. Much of the observations of Frost and Shive [1986] still pertain, for in most environments serpentinite incorporated into the lower crust will encounter temperatures sufficient to remove magnetite and serpentine from the assemblage. However, there are two instances we can envision wherein serpentinites may form deep crustal anomalies. One may occur in areas underlain by old oceanic crust. Here fluid convection cells may extend deep enough so that they interact with upper mantle that is cold enough for serpentine to be stable. Such upper

mantle may undergo serpentinization with the concurrent development of magnetite. Data from Arkani-Hamed [1988] suggests that precisely this may have happened beneath the oceanic crust, where magnetic anomaly skewness analysis suggests a strong magnetic source in the upper mantle.

Another place where serpentinite may be present in the lower crust is in active subduction zones or in areas where serpentinites have been thrust under the continent and have not been subjected to later heating. Recent studies show that serpentinite from low-temperature eclogite facies of the Italian Alps ($P = 15$ kbar., $T = 550 - 600^{\circ}\text{C}$) contains opaque assemblages similar to those of the low-pressure Malenco serpentinite, including both magnetite and native metals [Rossetti, et al. 1987]. Thus in areas where serpentinite is being actively subducted magnetite may be stable and magnetic down to depths of 40 km or more. Consequently, such areas should be marked by distinctive, deep magnetic anomalies.

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