MAGNETIC MINERALOGY OF THE LOWER CONTINENTAL CRUST

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Abstract. Rocks that were once present in the deep continental crust can be found in surface exposure as xenoliths, as parts of high-grade metamorphic and plutonic terrains, and as obducted lower crustal cross sections. A review of magnetic and mineralogic studies of samples representing all such occurrences shows that magnetite is the only magnetic phase present in significant amounts in these rocks. Thermodynamic considerations indicate that Ti-poor magnetite is the expected magnetic oxide under equilibrium conditions that exist in the lower crust, regardless of the primary magmatic composition of the spinel phase. Furthermore, it seems unlikely that serpentinites with metal alloys contribute significantly to the magnetization of the lower crust or mantle. Thus magnetite in the deep crust is the most likely candidate for the source of the long-wavelength magnetic anomalies observed in regional aeromagnetic and satellite magnetometer surveys. If so, the entire crust should be magnetic in most continental areas, including shields. In areas of high heat flow the bottom of the magnetic crust may lie considerably above the Moho.

Introduction

During the last decade much attention has been focussed on long-wavelength (or regional) magnetic anomalies observed in regional aeromagnetic surveys [e.g., Hall, 1974; Krutikhovskaya et al., 1979] and satellite magnetometer studies [e.g., Regan et al., 1975; Schnetzler and Allenby, 1983]. Presuming that these anomalies result from magnetic sources in the crust of the earth, an obvious application of the above observations is the development of crustal models that are consistent with these anomalies. However, attempts to develop such models have been handicapped by a surprising lack of knowledge about the magnetic properties of the lower crust. For example, there has been considerable debate about the fundamental question of the magnetic carrier of lower crustal magnetization [Haggerty, 1978, 1979; Wasilewski et al., 1979].

Identification of magnetic source minerals in the lower crust is critical. Any model must include information about the intensity and distribution of magnetization. The intensity of magnetization is controlled by the magnetization of the magnetic phases present and also by their volume fraction in the crust. Finally, knowledge of the magnetic mineralogy provides information

about the thermal and barometric conditions obtaining at the time of their equilibration. Different minerals have different Curie temperatures and thus require different locations of the Curie temperature isotherm, which limits the thickness of the magnetic crust.

The primary purpose of this paper is to support the contention that nearly pure magnetite is the only important magnetic source in the deep continental crust. We will review the observational evidence resulting from rock magnetic studies and provide petrologic justification in support of this claim. We will then conclude with some comments about the implications for the distribution of magnetization in different continental crustal sections.

Magnetic Measurements of Crustal Rocks

One obvious approach to identifying the magnetic carrier in the deep crust is to study rocks that are presently exposed but were once at lower crustal depths. These rocks represent three categories: xenoliths, general plutonic and high-grade metamorphic terrains, and obducted cross sections of lower continental crust. The first two categories may include some samples that did not come from the lower crust. Nevertheless, any clear trends in magnetic mineralogy are useful because most of the samples probably do have lower crustal origin. The third category will include rocks that were undoubtedly once part of the lower crust.

The only magnetic study of crustal xenoliths is by Wasilewski and Mayhew [1982]. They studied seven xenoliths of probable lower crustal origin. Thermomagnetic data were measured for six samples. Four show thermomagnetic curves characteristic of magnetite with at most a few percent ulvospinel in solid solution. The other two, which were from unusually hot environments (e.g., rift valleys), contain an ulvospinel-rich titanomagnetite.

Magnetic studies of plutonic rocks and rocks from high-grade metamorphic terrains are too numerous to compare in detail. However, relatively few of these address directly the identification of the magnetic phases. To minimize the possibility of misidentification, we will restrict our consideration to studies in which magnetic mineral identification was made by J_s versus T analysis supported by optical microscopy [e.g., Larson et al., 1969; Lidiak, 1974; Hall et al., 1979; Piper, 1979; Shive and Morel, 1980]. All such studies indicate that low-Ti titanomagnetite is the dominant, often the only, significant magnetic phase present, although pyrrhotite is occasionally found in mafic supracrustal rocks. Titanomagnetite Curie temperatures are usually between 550° and 580°C. The Lewisian rocks studied by Piper

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Paper number 5B5615.

0148-0227/86/005B-5615\$05.00

[1979] had somewhat lower Curie temperatures, but only four of 39 samples with reversible curves had T_c values below 520°C.

There are a number of places on the surface of the earth where the upper passive plate of a convergent continental plate margin has been uplifted and eroded to expose its lower portions [Fountain and Salisbury, 1981]. The informal type sections for such occurrences is the Ivrea zone in northern Italy [Berckhemer, 1969]. Systematic magnetic studies in which magnetic phases were identified have been published for only three of these lower crustal sections: the Ivrea zone, the Lofoten-Vesterålen terrain of Norway, and the Pikwitonei-Cross Lake subprovinces of Manitoba.

Wasilewski and Fountain [1982] found that magnetite was the only important magnetic phase in all four samples from the Ivrea zone for which identification was attempted. Identification was made on the basis of J_s - T curves (T_c between 560° and 580°C) and polished section analysis. Schlinger [1985] performed thermomagnetic analysis, reflected light microscopy, and microprobe analysis of 40 samples from the Lofoten-Vesterålen terrain representing diverse lithologies mostly at amphibolite and granulite metamorphic grade. Nearly pure magnetite (T_c between 550° and 575°C) was the only magnetic phase in all 40 samples. M.L. Williams et al. [1985] studied 52 samples collected from a traverse across the Pikwitonei-Cross Lake subprovinces. Metamorphic grade ranges from greenschist to granulite; silicic plutonic rocks dominate in the section. Twenty-seven samples are dominantly paramagnetic, and three mafic supracrustal samples contain pyrrhotite. The remaining 22 samples are from rocks that make up about 80% of the section. They have reversible thermomagnetic curves with Curie temperatures between 560°C and 580°C. This is diagnostic of magnetite with less than 2% ulvospinel in solid solution, a composition which is confirmed by X ray diffraction of magnetic separates from five samples.

The single occurrence of native iron in granulite [Haggerty and Toft, 1985] should be addressed in some detail. The specimen in question occurs as an inclusion in a kimberlite pipe. It is clear from the description that it has been subjected to significant late stage metasomatism some of which may be related to the processes associated with the eruption of the kimberlite. Texturally, the native iron is part of this late event and not from the main stage assemblage (see, for example, Figure 1 of Haggerty and Toft [1985]). If we accept Haggerty and Toft's estimate that the main-stage Fe-Ti oxides equilibrated between the IW and MW buffers, then by definition they will not be stable with pure native iron. It is only reasonable to assume that the native iron formed along with the other late-stage phases as a result of metasomatism of a fluid rich in CH_4 . That such a fluid could have been associated with early stages of kimberlite development is not impossible. Pasteris [1981] describes graphite and serpentine from kimberlite, and Frost [1985] argues that such an assemblage is indicative of CH_4 -bearing fluids. We therefore conclude that although the occurrence of native iron described

by Haggerty and Toft [1985] poses fascinating questions both to kimberlite genesis and to the character of the lower continental crust, it is by no means unequivocal evidence that metals are present in most mantle or crustal environments.

From the above review it appears that magnetite is the only important magnetic mineral in lower crustal rocks presently exposed at the earth's surface. It could be argued, however, that the magnetic minerals now present in these rocks were not present when they were actually in the lower crust but are retrograde phases formed during the subsequent tectonic history. In the following section we will present some geochemical arguments that suggest that this is not so; that is, that magnetite would have been the stable phase in these rocks under lower crustal conditions. In addition, we will offer some explanations why magnetite, rather than some other phase, is the only stable magnetic mineral in lower crustal conditions.

Petrologic Arguments

Factors Controlling the Composition of Magnetite

The numerous reports of nearly pure magnetite from crustal rocks detailed above seem in contradiction to the wide range of magnetite compositions found in igneous rocks. As Haggerty [1979] points out, the composition and hence Curie temperature of titanomagnetite in igneous and metamorphic rocks is a complex function of temperature, oxygen fugacity, and rock composition. The temperature and oxygen fugacity dependence of titanomagnetite composition when saturated with respect to ilmenite is shown in Figure 1 as calculated from data of Andersen and Lindsley [1985], following the technique of Spencer and Lindsley [1981]. After the convention of Frost et al. [1986] oxygen fugacity in this figure has been normalized to that of the FMQ (fayalite, magnetite, quartz) buffer at 1 bar.

Since magnetite has its maximum Ti content when it coexists with ilmenite, the isopleths on Figure 1 give the maximum ulvospinel content possible in magnetite at given T and f_{O_2} . Also shown in Figure 1 is the Curie temperature for magnetite when saturated with respect to ilmenite. The exact locations of the isopleths for very low ulvospinel contents are subject to error because they lie well outside the range of the experimental data on which Figure 1 is based. This places some uncertainty on the exact location of the Curie isotherm. It is possible that it may lie as much as 30° below what is shown on Figure 1. This uncertainty, however, does not alter the conclusion derived from Figure 1 that if the oxides in the lower crust have equilibrated, the Curie temperature for most rocks should lie above 500°C. From this it is clear that knowledge of the behavior of oxides during cooling is key to understanding the magnetic properties of magnetite in the lower crust.

The only rocks to which these arguments do not apply are those which contain magnetite but not ilmenite. These are quite rare in nature and are represented mostly by metamorphosed iron formations and alkalic volcanic and plutonic

Fig. 1. Diagram showing effect of oxygen fugacity on the Curie temperature of titanomagnetite when it coexists with ilmenite. Isopleths give the ulvospinel content of magnetite. $\Delta \log f_{O_2}$ refers to the deviation of f_{O_2} from the FMQ buffer. HM, hematite-magnetite buffer. Figure is calculated from the expressions of Spencer and Lindsley 1981 using the solution models of Anderson and Lindsley 1985.

rocks. In the former this lack presents no problem because the magnetite in iron formations is nearly devoid of Ti [Florán and Papike, 1978] and hence can be treated as pure Fe_3O_4 . In the latter, magnetite may be quite rich in the Fe_2TiO_4 molecule and yet show no signs of retrogressive reequilibration [Powell, 1978]. Furthermore, alkalic rocks make up only a small portion of the earth's crust.

Most igneous rocks form at oxygen fugacities within three log units of the FMQ buffer [see Haggerty, 1976]. For those rocks containing two Fe-Ti oxides the composition of the spinel phase present can be obtained from Figure 1. At igneous temperatures the titanomagnetite in these rocks would have a wide range of composition. However, because the ulvospinel isopleths have a progressively steeper slope in f_{O_2} -T space as temperature falls, the composition range for oxides in igneous rocks will become progressively narrower as the rocks reequilibrate to lower temperatures.

The exact details of how the oxide compositions in rocks change on cooling is controlled by the individual f_{O_2} -T path followed. At present the f_{O_2} -T paths followed during cooling of rocks are not well constrained. In general, we can consider each path to consist of three stages. At high temperatures, f_{O_2} will be controlled by oxide-silicate equilibria. As temperature falls, the silicate composition will eventually become locked in. At this point the oxygen fugacity will be controlled by interoxide equilibria. Finally, at some sufficiently low temperature the

interoxide exchange equilibration will cease and oxygen fugacity will become fixed, probably on a very local basis, by the exsolution and oxidation processes occurring within the individual titanomagnetite grains. Graphite should not be important because graphite and magnetite rarely coexist and because graphite controls f_{O_2} only in the presence of a free fluid phase which² is quickly consumed during cooling by retrograde reactions.

The exact T- f_{O_2} trends followed by each process and the temperature at which each process becomes "locked in" is poorly known. For oxide-silicate equilibration the path followed will depend on the mineral assemblage present. For example, in a rock with fayalite and no ilmenite, the silicate-oxide cooling trend will follow the equilibrium:

or FMQ, while a rock with biotite and k-feldspar without ilmenite will follow a path along the equilibrium:

or KMA. In rocks with two Fe-Ti oxides the oxide-silicate cooling trend is somewhat more difficult to determine because of the variable substitution of titanium into magnetite. In fayalite-bearing rocks with two oxides, equilibrium (1) becomes

Fig. 2. $\Delta \log f_{O_2}$ -T diagram showing the location of important oxide-silicate equilibria in Mg-free systems (heavy lines) for $P_{H_2O} = 1$ kbar (possible hydration of fayalite at low temperatures is ignored). FMQ, fayalite-magnetite-quartz; KMA, Kspars-magnetite-annite; QUILF, quartz-ulvöspinel-ilmenite-fayalite. Light lines indicate the location of common buffers: HM, hematite-magnetite; MW, magnetite-wüstite; WI, wüstite-iron; QIF, quartz-iron-fayalite. See text for discussion.

or QUILF and reaction (2) becomes

or KUILA.

The experimentally determined positions of FMQ [Myers and Eugster, 1983], QUILF [Frost et al., 1986], and KMA [Eugster and Wones, 1962] for Mg-free systems ($P_{H_2O} = 1000$ bars) are shown on Figure 2. The location of KUILA is not shown on Figure 2 because it has not been determined experimentally. Since biotite can dissolve Ti but Ti cannot dissolve biotite, it is difficult to predict where KUILA will lie relative to KMA. Chances are good that KUILA will lie close to KMA but that it will have a more oxidizing slope, because at progressively lower f_{O_2} titanomagnetite will become increasingly enriched in Ti. This oxidizing cooling trend along KMA may account for the relatively oxidized nature of some biotite granites [Czamanske and Mihalik, 1972; Czamanske and Wones, 1973]. In natural rocks the presence of Mg will displace the curves to higher T and f_{O_2} . Sufficient Mg will also stabilize orthopyroxene at the expense of olivine + quartz and give rise to the buffering

assemblages: olivine-orthopyroxene-magnetite-ilmenite and orthopyroxene-quartz-magnetite-ilmenite. Despite this effect the displaced curves should behave similarly to the Mg-free curves shown on Figure 2. On cooling, therefore, magnetite-bearing anhydrous rocks, such as those with pyroxenes and olivine, will follow oxide-silicate equilibration paths that are parallel to FMQ or QUILF, that is, paths that have constant relative oxygen fugacity or that are only weakly oxidizing in relative f_{O_2} . As long as P_{H_2O} remains close to P_{total} , however, rocks with biotite could follow strongly oxidizing trends, parallel to KMA. An indication that such buffering does take place is seen in the data of Carmichael [1967], who shows that fayalite-bearing rhyolites equilibrated at oxygen fugacities near FMQ. Rhyolites with orthopyroxene equilibrated along a curve roughly parallel to FMQ but at oxygen fugacities about one log unit higher and those with biotite equilibrated at oxygen fugacities up to three log units above FMQ.

At present the temperature at which oxide-silicate equilibria become frozen can be determined only for rocks containing fayalite and quartz. This is done by calculating the conditions where the compositions of the oxides are mutually compatible with that of fayalite according to the QUILF equilibrium [see Frost et al., in press]. When done for five samples from the Upper Border Group of the Skaergaard intrusion (data from Naslund [1980]), it can be

Fig. 3. $\log f_{O_2}$ diagram showing paths followed during oxide-silicate, oxide-oxide, and intraoxide reequilibration in rocks with anhydrous silicates. See text for discussion.

seen that these rocks, which had magmatic temperatures in excess of 985°C [Lindsley et al., 1969], reequilibrated along QUILF to temperatures between 900°C and 730°C (see Figure 3). There is no reason to assume that rocks with orthopyroxene as the major silicate rather than fayalite would have very different blocking temperatures for the oxide-silicate equilibria, but rocks with hydrous silicates would almost certainly equilibrate to lower temperatures.

Once the silicates cease reacting with the oxides, the T - f_{O_2} trend becomes a function of the relative abundances of the oxides present. Rocks in which ilmenite is the major oxide will follow trends close to those of the ilmenite isopleths in Figure 3, while those in which magnetite is dominant will more closely follow the ulvospinel isopleths. By plotting the T - f_{O_2} results obtained from reintegrated oxide² compositions, Morse [1980] documented three cooling trends for the Kiglapait complex, with each trend relating to rocks with differing relative abundances of the two oxides. These trends, as reproduced in Figure 3, are moderately reducing and seem to represent trends for rocks in which magnetite was the dominant oxide. The data from Morse [1980] indicate that there is a wide blocking temperature for interoxide equilibration, for his two-oxide temperatures cover a range of 550° to 900°C .

The best source of data for the cooling trend followed during "oxyexsolution" of magnetite (intragrain reequilibration) is that of Bohlen and Essene [1977] who provide analyses of host magnetite, "exsolved" ilmenite, reintegrated magnetite, and groundmass ilmenite from granulites in the Adirondacks. The temperatures of equilibration calculated for the host magnetite-ilmenite pairs lie well outside of the

calibration range of the Fe-Ti oxide geothermometer and hence are subject to considerable error. Even so, two important points are evident. First is that no consistent trend is followed during intragrain equilibration. Of the seven host-lamellae pairs given by Bohlen and Essene [1977], five underwent slight to strong reduction on cooling, while two underwent relative oxidation. Second, reequilibration locked in at temperatures well below the blocking temperature of magnetite.

Although the equilibration trends on Figure 3 come from three highly divergent rock types, they probably reflect the general behavior of oxides in rocks that are either devoid of water or where water ceased to move through the rock at temperatures above those required to hydrate the olivines and pyroxenes. In relatively wet rocks, where hydrothermal activity may occur at temperatures considerably below the solidus, oxide-silicate equilibration may persist to temperatures well below 600°C , as witnessed by the low Ti content and homogeneous nature of magnetite from biotite and hornblende granites [cf. Czamanske et al., 1977].

Regardless of the cooling trends followed, it is clear that for the purpose of determining the magnetic signature of deep-seated rocks the primary magmatic composition of the spinel phase is unimportant. On cooling the spinel will clear itself of titanium, either through intergrain diffusion or by "oxyexsolution" and for most oxygen fugacities within 3 log units of the FMQ buffer in most crustal rocks the spinel will be a magnetite that is pure enough to be magnetically ordered by the time the rock has cooled to between 550° and 525°C . The magnetic signature of the crust is therefore dependent on only two factors: the abundance of magnetite in the rocks

present and the depth to the blocking temperature isotherm. The latter of course is determined by the geothermal gradient.

Factors Controlling the Magnetic Signature of Mantle Rocks

Haggerty [1978, 1979] argues that sources for the aeromagnetic anomalies are in the mantle. Wasilewski et al. [1979], however, found that mantle nodules were only weakly magnetic and argued that the Moho is a magnetic boundary. One might contend that the samples studied by Wasilewski et al. [1979] represent only a miniscule portion of the earth's mantle and that native metal alloys nonuniformly scattered within the mantle could produce the observed magnetic anomalies. Recent work [Frost, 1985], however, shows that the metal alloys will be stable in serpentinites only at or near the temperature for the reaction

Even at 10 kbar this reaction extends only to temperatures of 480°C, well below the Curie temperature for magnetite.

The fact that serpentinites can be highly magnetic [Carmichael, 1982] might seem to belie the contention that the Moho is the magnetic basement. However, petrologic studies of spinels in ultramafic rocks [Evans and Frost, 1975] show that magnetite is present only at low temperatures, where the components such as Cr_2O_3 and Al_2O_3 , which normally dissolve into the spinel phase, are taken up by hydrous silicates such as serpentine and chlorite instead. At temperatures above the stability of antigorite, spinel in ultramafic rocks is relatively enriched in Al_2O_3 and Cr_2O_3 and is nonmagnetic.

It is theoretically possible for serpentinites and their associated magnetite and metal alloys to occur in the upper mantle in oceanic areas or even in continental regions of extremely low heat flow. In fact, such rocks might make a major contribution to the magnetic signature from oceanic areas where peridotite diapirs have risen to high crustal levels, for example St. Paul's Rocks [Melson et al., 1972]. The western Tethys ophiolites are often considered to be exposures of such serpentinites [Abbate et al., 1984]. However, geologic evidence strongly indicates that the majority of the suboceanic mantle and the mantle below the continents is serpentine-free. This is supported by the observation that many ophiolites were hot enough during emplacement to produce contact metamorphism (see Karamata [1980] for a review of occurrences). The serpentine in such ophiolites is, therefore, interpreted as having formed after emplacement of the peridotites in the continental crust. As a result, the primary mineral assemblages in the lower portions of ophiolites are generally considered to reflect granulite facies of metamorphism [Coleman, 1977]. That mantle xenoliths from continental regimes also have granulitic mineralogies [Forbes and Kuno, 1967] is evidence that the mantle below continents is also serpentine-free.

Thus on the basis of petrologic arguments and limited measurements of magnetic susceptibility

[Wasilewski et al., 1979] it seems safe to conclude that the Moho is, indeed, magnetic basement. This would be true even if the upper mantle were composed of eclogite rather than peridotite. The few magnetic susceptibility measurements available show eclogites to be paramagnetic [Wasilewski et al., 1979]. This is consistent with petrologic observations that magnetite is generally lacking in eclogites, with rutile + ilmenite stable instead (see, for example, Medaris [1980] and Ridley [1984]). Apparently at high pressures magnetite is destabilized by the tendency for Fe^{3+} to dissolve readily into garnet and omphacitic pyroxene. This is indicated by the fact that magnetite seems to be formed by retrogressive or decompressive reactions in eclogites [see Mysen and Heier, 1972; Cortesogno et al., 1977].

Discussion

Assuming that nearly pure magnetite provides the magnetic source in deep crustal rocks, we now consider the question "what is a reasonable model for the distribution of magnetization in the crust?" To answer this question completely we need information about 1) the magnetic properties of magnetite in deep crustal rocks, 2) the volume percent magnetite as a function of depth in the crust, and 3) the depth at which magnetite ceases to be strongly magnetic. None of these factors can be thoroughly evaluated at this time, but it is appropriate to briefly review some of the considerations relevant to each.

The magnetic properties of interest are induced and remanent magnetization. High-temperature susceptibility measurements on samples from two separate lower crustal sections [Schlinger, 1985; M. Williams et al., 1985] show that induced magnetization does not vary by more than about 10% between room temperature and 540°C. At that point the susceptibilities drop sharply to comparatively insignificant values. No Hopkinson effect rise was observed for any sample.

Much less is known about the remanent magnetization of lower crustal rocks. Many authors have attempted to measure the importance of lower crustal remanence through comparisons of natural remanent magnetization (NRM) intensities with induced magnetization. This is misleading. Lower crustal remanence is almost certainly a high-temperature viscous magnetization (VRM). NRM consists of thermoremanent magnetization (TRM) acquired during cooling plus secondary components of magnetization. Furthermore, NRM is measured at room temperature. For these reasons its intensity is a poor indicator of the strength of VRM in the lower crust. The best way to characterize the remanent contribution is to measure VRM acquisition in lower crustal samples under lower crustal conditions. No such measurements have yet been made.

The distribution of magnetite in crustal rocks is extremely variable. The prevailing view has been that magnetite content should increase with iron content and (perhaps) with increasing metamorphic grade. The Pikwitonei study [M. Williams et al., 1985], however, showed that

TABLE 1. Thermal Parameters and Depth to the 600°C Isotherm for Several Continental Heat Flow Provinces

Province	Q_s	Q_r	D	Z_{600}
Basin and Range	92	59	9.4	20
Sierra Nevada	39	17	10.1	74*
Eastern Australia	72	58	11.1	23
Eastern United States	57	33	7.5	39*
United Kingdom	59	24	16.0	40*
Central Australia	83	27	11.1	33
Ukrainian Shield	37	25	7.1	55*
Western Australia	39	26	4.5	55*
Canadian Shield	39	21	14.4	59*
Baltic Shield	37	22	8.5	61*

Q_s is the mean surface heat flow in mW/m^2 , Q_r is the reduced heat flow in mW/m^2 , D is the characteristic depth for crustal heat production, and Z_{600} is the depth to the 600°C isotherm.

* Depths that exceed the seismically determined thickness of the crust in that province.

silicic rocks were about an order of magnitude more magnetic than mafic rocks and that magnetization did not depend on metamorphic grade. More systematic magnetic studies of lower crustal sections, together with petrologic analysis of the phase assemblages, should provide reasonable limits for this important variable.

The bottom of the magnetic source layer in the crust will be the depth at which magnetite becomes effectively paramagnetic. The high-temperature susceptibility measurements suggest that this will occur at 540°–550°C, although remanent contributions could be important at slightly higher temperatures. The Curie temperature of magnetite increases with pressure at the rate of 1.85°C/kbar [Schult, 1970]. Taking this into account, it is reasonable to say that lower crustal rocks should lose their magnetization at about 600°C.

The depth of the 600° isotherm depends on the geothermal conditions of the crustal section. To obtain estimates for different continental heat flow provinces, we have assumed that the distribution of radioactive heat production decays exponentially with depth and evaluated the equation

$$T = T_o + \frac{Q_r Z}{K} + \frac{(Q_s - Q_r) D (1 - e^{-Z/D})}{K}$$

that describes temperature T versus depth Z for this situation [Turcotte and Schubert, 1982]. Here Q_s is surface heat flow, Q_r is reduced heat flow, D is the characteristic depth for heat production, and K is thermal conductivity. Values of Q_s , Q_r , and D for different provinces are taken from Sclater et al. [1980]; K is assumed to be 2.5 $\text{W/m}^2\text{K}$ [Pollack and Chapman, 1977].

Table 1 shows the results of these calculations. The depth to the 600°C isotherm varies from 20 km in the Basin and Range province to 74 km in the Sierra Nevada. In only three provinces (Basin and Range, Eastern Australia, and Central Australia) do the depths lie within the crust. The lower part of the crust should be nonmagnetic in those three areas. The Z_{600} values for the other seven provinces are greater than their seismically determined crustal

thickness [Soller et al., 1982]. Here the entire crust should be magnetic, and, assuming that the mantle is relatively nonmagnetic [Wasilewski et al., 1979], the crust-mantle boundary would define the bottom of the magnetic source layer.

We note that considerable uncertainty attends the calculations for the 600° isotherms. These arise from uncertainties in the thermal parameters and the assumption that heat production decreases exponentially with depth [Hart et al., 1981; Nicolaysen et al., 1981]. The depths should therefore not be taken literally. Our intent is to point out that variations in continental geotherms and crustal thicknesses will cause considerable differences in the depth at which magnetite loses its magnetization. In some places, particularly shields and areas of low heat flow, the entire crust should be magnetic. In areas of high heat flow the bottom of the magnetic crust may lie considerably above the Moho.

Acknowledgements. David Fountain introduced us to lower crustal studies through his Pikwitonei project, and Martin Williams provided much of the magnetic and petrologic work on that study. Don Lindsley produced much of the petrologic data on which this paper depends and gave us helpful advice during the development of this paper. The petrologic part of this study was partially supported by National Science Foundation grant EAR - 8409663 (R.B.F.).

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(Received April 23, 1985;
revised January 6, 1986;
accepted January 5, 1986.)