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ESTIMATION OF LOWER CRUST MAGNETIZATION FROM SATELLITE DERIVED ANOMALY FIELD

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ABSTRACT

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Various lines of evidence point to the lower crust as the source of the long-wavelength magnetic anomaly field measured by the POGO and Magsat satellites. Using seismically determined lower crust thicknesses and equivalent source inversion of the satellite anomaly data, magnetization for the lower crust for much of the United States has been calculated. The average magnetization for two hundred sixty-six 150 × 150 km areas is 3.5 A/m with a standard deviation of 1.1 A/m. These values are consistent with laboratory measurements of mafic-ultramafic rocks expected in the lower crust, and in agreement with previous estimates of lower crust magnetization based on long-wavelength aeromagnetic data. Average lower crust thickness for the same areas is 18.2 km ($\sigma = 6.4$ km). Thus, over large regions, it appears that variation in magnetization and variation in magnetic layer thickness contribute almost equally in causing the anomaly field variation at satellite altitude.

INTRODUCTION

Long-wavelength geomagnetic anomalies ($\gtrsim 250$ km wavelength) have been observed by magnetometers flown on Cosmos 49 (Zietz et al., 1970), the POGO series (Regan et al., 1975), and the recently completed Magsat mission (Langel, 1981). The anomaly field exhibits a range of about ± 10 nT. It is extracted from the total signal measured by the satellite by removal of the 30,000–50,000 nT main field through a spherical harmonic analysis, after both minimizing variations (by picking data from magnetically quiet days) and removing an external field of up to ± 100 nT (Langel et al., 1980). Due to the size of the signal compared to the "noise" and the ad hoc nature of some of the signal extraction techniques, much effort to date has been directed toward verification of the physical reality of the crustal anomalies. Verification has included comparisons between anomalies derived by the different satellites (e.g., Langel, 1981), comparisons between anomalies and known tectonic areas (e.g., Mayhew et al., 1980), comparisons between anomalies and upward continued

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aeromagnetic data (e.g., Langel et al., 1980; Phillips and Hildenbrand, 1981), and sensitivity studies on the effect of possible perturbing factors (e.g., Taylor et al., 1981). Because of this concentration on the validity of the anomalies, comparatively little work has been directed toward an understanding of the sources of the anomalies (Regan and Marsh, 1982; Mayhew et al., 1982b). Beyond an understanding that the signal extraction technique separates a small anomaly field produced in the upper part of the earth (called "crustal" for simplicity) from the much larger main field produced in the core, there is little quantitative information on what properties in the upper part of the earth are contributing to variations in the anomaly field, and to what degree.

Variations in the crustal field can be ascribed to: (a) variations in thickness of a magnetized layer; (b) lateral variations in concentration of magnetic material as evidence by variation in magnetization within the layer; and (c) variations in the intensity and/or direction of remanent magnetism. This paper concentrates on the first two of these possibilities. We take the magnetic layer to be the lower crust, and, for those areas in the United States where lower crust thickness can be inferred from seismic records, examine the variation in magnetization required to produce the satellite measured anomaly field. Initial examination of the Magsat vector magnetometer data indicates that most of the scalar anomaly field is consistent with an

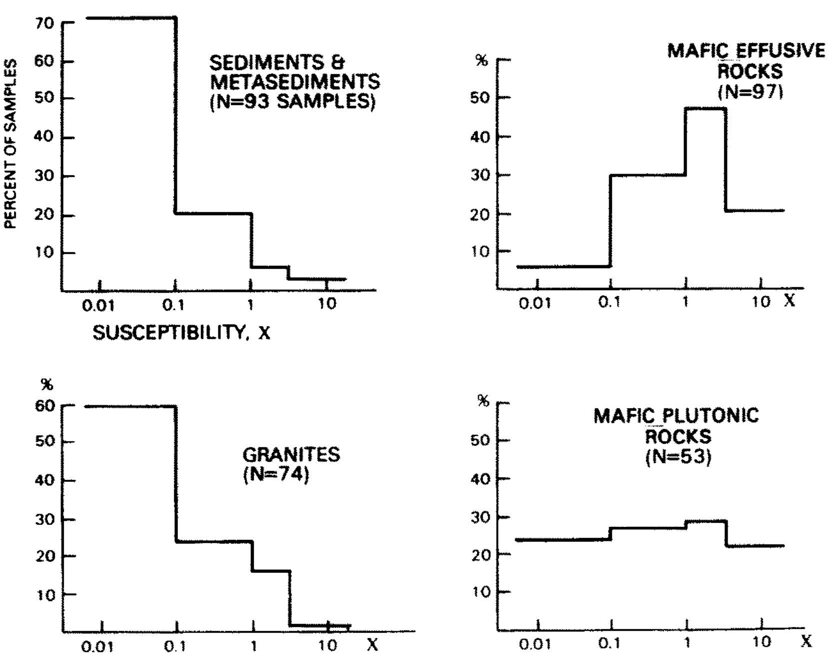

Fig. 1. Histograms of laboratory measured magnetic susceptibilities in various rock types, from Lindsley et al. (1966). Susceptibility in International System (SI) units.

induced field model only; however, some regions may require remanent magnetization (Langel et al., 1982). At this early stage in the analysis of the component data we have assumed that the magnetization is in the direction of the main field.

Laboratory measurements of rock magnetism, together with general models of crustal structure give some insight on the magnetic layer. Lindsley et al. (1966) summarized the range of magnetic susceptibilities in major rock types, compiling over 300 laboratory measurements. These analyses show that only a small percent of sediments (8%), meta-sediments (7%) and granites (17%) analyzed had susceptibilities greater than 1 (International System units), in contrast to the high percent of mafic effusive (66%) and plutonic (49%) rocks which had susceptibilities greater than 1. The distribution of ranges is shown in Fig. 1. Additionally, Wasilewski and Fountain (1982) have reported magnetic susceptibility measurements of rocks from the Ivrea zone, Italy, considered to be a cross-section of the crust and upper mantle turned on it's side and exposed at the surface. They report high magnetic susceptibilities (> 0.5, and typically > 1) for the mafic–ultramafic rocks, all of which are found in the lower crust. Rocks from the upper crust had low susceptibilities, less than 0.005. In another laboratory study, Wasilewski et al. (1979) measured magnetic properties of a number of xenoliths presumably derived from the upper mantle and found very low magnetization. This, coupled with an analysis of the xenolith mineralogy, lead the authors to conclude "that the crust–mantle transition is a magnetic mineralogy transition, with the refractory (non-magnetic) spinels in the mantle and Fe–Ti spinels in the crust".

General models of continental crustal structure usually invoke a silicic upper crust, approximating granitic or granodioritic in composition and a lower crust more mafic in composition. The traditional seismic boundary between these two layers, the Conrad discontinuity occurs at a P-wave velocity of about 6.5 km/sec. Thus, although it is recognized that both layers are probably quite inhomogeneous, the laboratory rock measurements point to a lower crust with magnetic susceptibilities at least an order of magnitude greater than rocks of the sedimentary cover, upper crust or mantle. It is assumed in this paper that "the lower crust appears to be the most geologically realistic source of large-scale magnetization required to satisfy modeling of long-wavelength magnetic anomalies" (Wasilewski and Mayhew, 1982). We will investigate the consequences of this assumption, namely:

(1) What values of lower crust magnetic susceptibility are required to produce the satellite measured long-wavelength magnetic anomalies, given some knowledge of lower crustal thickness?

(2) Are these values reasonable for expected lower-crust rocks? That is, are the values consistent with the lower-crust magnetic layer model?

(3) Are the long-wavelength anomalies primarily produced by variation in magnetic layer thickness or lateral variation in magnetic susceptibility?

Previous work along these lines has been done using long-wavelength aeromagnetic data. Hall (1974) compared magnetic-anomaly data with upper, lower and total crust thickness for twelve selected regions of Manitoba and northwestern Ontario, Canada. He found a linear correlation with all three layers, but concluded that the most likely source region was the lower crust, and estimated a magnetization for this region of 5.3 A/m. Krutikhovskaya and Pashkevich (1979) examined the relation between magnetization and total crustal thickness for the Ukrainian Shield and found a strong linear relationship. They concluded that the magnetic anomalies primarily reflect crustal thickness, with the magnetization of the lower crust five to ten times greater than that of the upper crust, but inhomogeneous. These studies, however, differ from this paper in that they used a long-wavelength portion of aeromagnetic, rather than satellite data, with bandpass filters of 60–4000 km (Hall) and 60–300 km (Krutikhovskaya and Pashkevich), while wavelengths of greater than approximately 300 km from the POGO data was used here. Also the scale of the test sites are quite different. The test sites of Hall and Krutikhovskaya and Pshkevich were much smaller than used in this study—150–300 km by 1000 km in size, while this study encompasses an area approximately 2500 km by 5000 km. Lastly, both previous studies used a few (~ 12) points within their regions to test correlations; the present study contains 266 points, each representing an area of 150×150 km.

APPROACH

The conterminous United States was picked as the study region as a map exists which expresses the variation in dipole strength at the surface necessary to develop the magnetic anomaly field measured by POGO (Mayhew, 1982a), and considerable seismic data on Conrad and Moho discontinuity depths are available (Allenby and Schnetzler, 1983). Mayhew (1982a) developed a technique to invert magnetic anomaly field data, extracted from the total field measured by the satellite, to dipole sources placed at the earth's surface in an equal area grid and magnetized in the direction of the earth's main field. Magnetic moments of the dipoles are varied in order to minimize the differences, for each pass, between the satellite measured magnetic anomaly field and the field generated by the dipoles, at the appropriate altitude. The resulting dipole strengths are then contoured. Such a dipole map will exhibit more detail than the block averaged anomaly map. Figure 2 has been modified from Mayhew's (1982a) "equivalent layer magnetization model" and represents the dipole moments required to produce the measured anomaly field. The method of data reduction used to produce the anomaly field, and from this field the dipole moment map, produce relative, not absolute, values. Mayhew's original "equivalent source magnetization model" map (1982a), has positive and negative values. Assuming negative values of dipole moment are due to ambiguity in the zero level, and that the points with the lowest moment (highest negative value) should have near zero, positive values we added $1 \cdot 10^{18}$ Am^2 to Mayhew's values. Thus the range of the map runs from $+2$ to 24 ($\times 10^{17}$ Am^2), rather than from -8 to $+14$. The zero level is still undefined (but hopefully more realistic); thus Fig. 2 is called an *apparent* dipole moment map.

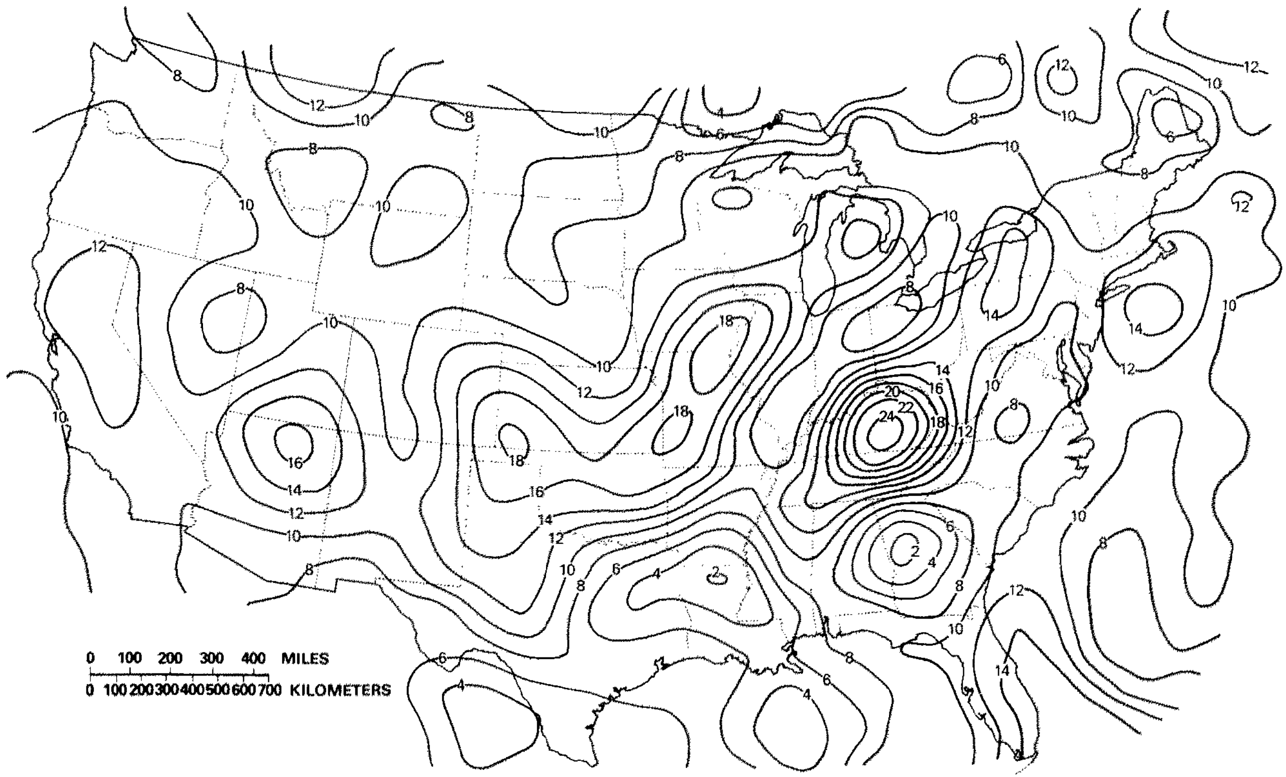

Fig. 2. Map of dipole moments, distributed in an equal area (150 km × 150 km) array at the earth's surface and aligned in the earth's main field direction, required to produce the POGO derived magnetic anomaly field. Modified from Mayhew (1982a). Units: 10^{17} Am^2.

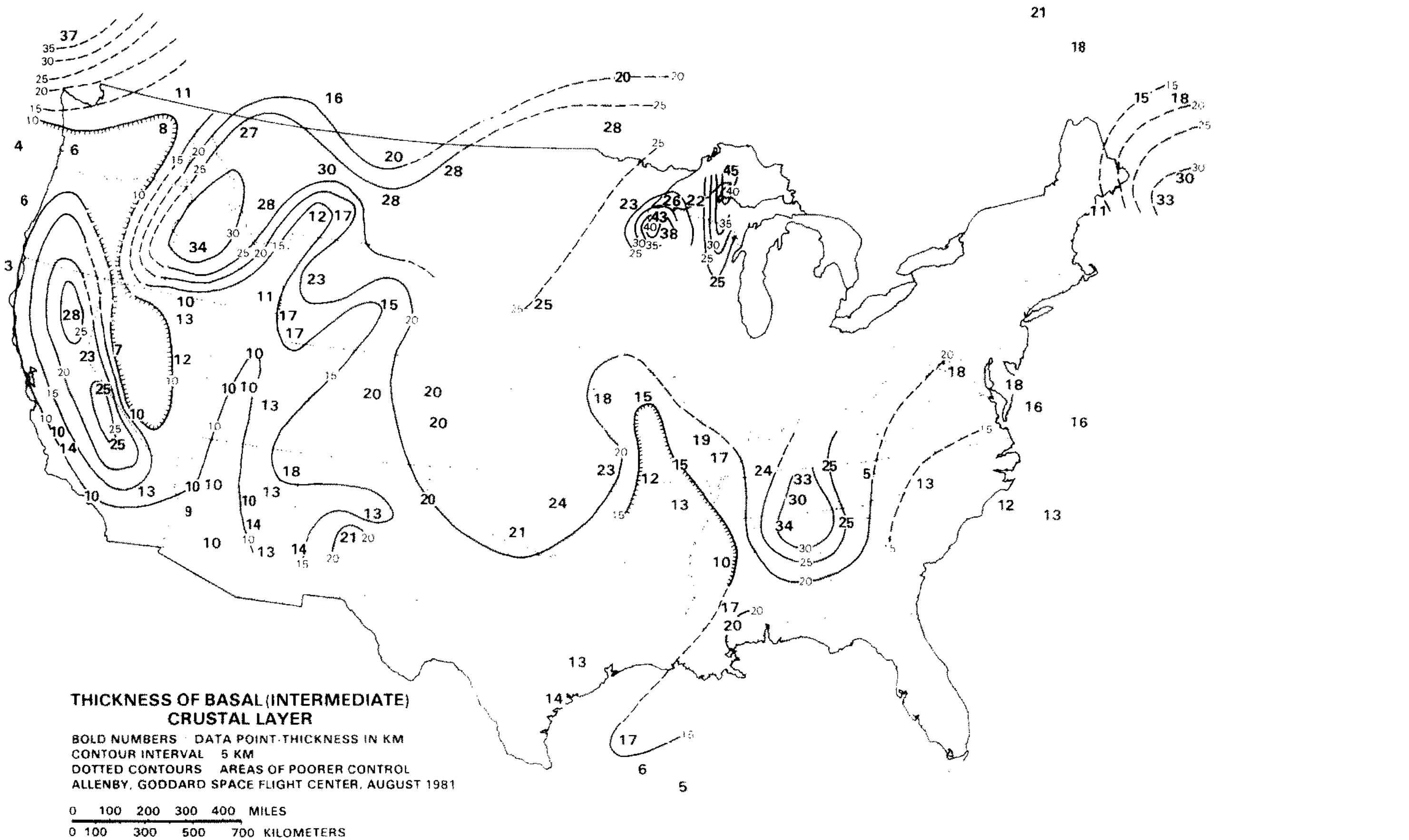

Fig. 3. Thickness of the lower crust in kilometers, from Allenby and Schnetzler (1983). Light figures are contour line values; bold figures are seismically determined data points.

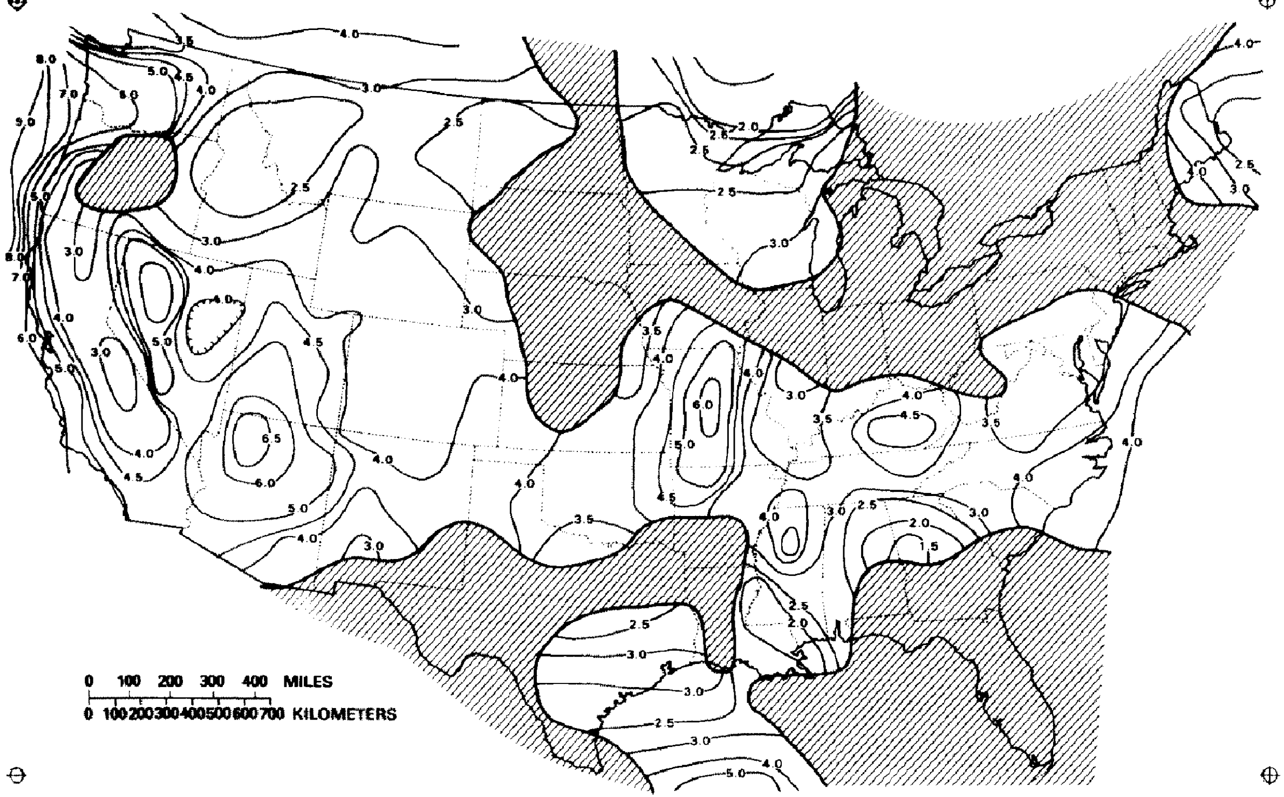

Fig. 4. Magnetization contrast in the seismically determined lower crust. Units are A/m. Shaded region includes areas greater than 225 km from either Moho or Conrad depth measurements.

Depth to the Conrad and Moho discontinuities were taken from comprehensive compilation of explosive seismic refraction lines in the United States by Allenby and Schnetzler (1983). Figure 3, from their paper, shows lower crustal thickness. The quality of the data, as well as its spatial density is quite variable. From a comparison of results from different seismic lines in the same regions we estimate an accuracy of $\pm 10\%$. Some areas of the United States, such as the northern mid-west and New England, have no deep seismic refraction lines and lower crustal thicknesses in these areas are not determined.

The apparent dipole moment representation of the source (Fig. 2) was then converted to a magnetic moment per unit volume, an apparent magnetization, based on magnetic layer thickness from Fig. 3. Both maps were gridded at Mayhew's (1982a) original 150 km dipole spacing and values of dipole moment and lower crust thickness were linearly interpolated at each grid point.

The apparent magnetization was then converted to absolute magnetization values according to a method described by Mayhew (1982b), which assumes that the magnetic dipole strength must go to zero if the magnetic layer goes to zero. The absolute magnetization (m) differs from the apparent magnetization (Δm) by a zero level ambiguity or off-set (M), i.e., $m = \Delta m + M$. The zero level off-set and thus the absolute magnetization can be calculated using two (or more) blocks of differing thickness and the same absolute magnetization. We assume the off-set is relatively constant, so blocks of equal apparent magnetization will have approximately the same absolute magnetization. As any two blocks could have rather large errors in thickness estimation, we have used all block thicknesses within a small range of apparent magnetization. The 246 blocks with apparent magnetization between 1.0 and 5.0 A/m were divided in eight groups at 0.5 A/m intervals; dipole moment vs. thickness for all blocks within each interval were plotted and a least squares line calculated. The y-intercept, where thickness went to zero, was the off-set. Calculated off-sets for the eight intervals averaged 0.94 ± 0.04 A/m. This value was added to all apparent magnetizations to arrive at absolute values. The contoured grid values are shown in Fig. 4. On this figure we have arbitrarily excluded areas where the grid points are more than $1\frac{1}{2}$ grid spaces (225 km) from the nearest seismically measured thickness as being of undetermined thickness.

DISCUSSION

Comparison of Figs. 2, 3 and 4 (i.e., dipole moment, lower crust thickness, and magnetization) indicate that both variation in lower-crustal thickness and magnetization play significant roles in producing long wavelength magnetic anomalies. The large positive dipole moment anomalies centered in Kentucky and through northern Oklahoma are primarily the result of increased lower crustal thickness; high magnetization is not required. However, the large dipole moments centered in Missouri and northern Arizona are in areas of relatively thin lower crust of high magnetization.

Two other areas show high magnetization on Fig. 4—the areas centered on northwestern Nevada and southwestern Washington. Unlike the regions just discussed, these two areas do not have anomalously high magnetic moment values; the very thin lower crust in these areas, however, requires high magnetization to produce normal dipole moments on Fig. 2. Thus the four areas which show high magnetization (> 4.5 A/m) on Fig. 4 are all regions of thin lower crust–less than 15 km (and less than 10 km in three of the four regions). Two of the four regions have expressions (highs) on the magnetic moment maps, while two do not. In all four regions the lower crust thicknesses are defined by several seismic records.

Regions of anomalously low magnetic moments ($< 6 \cdot 10^{16}$ Am^2) include central Georgia, Louisiana and east Texas, northeastern Mexico and southern Texas, and a small area in northern Maine. In all these areas seismic data which define lower crust thickness are very sparse to nonexistent. Lower crust thicknesses in the first two areas *appear* normal, but the only seismic data are on the periphery of the dipole moment anomalies. Thus, the lower magnetizations for these areas, shown in Fig. 4, are not well determined. Clearly, however, these regions have very low magnetic dipole moments and either the magnetic layer is very thin or the rocks comprising the layer have low magnetizations. Central Idaho–western Montana and western North Dakota are regions of rather normal dipole moments and thick lower crustal thicknesses; therefore magnetization is low.

The above discussion assumes the source of the magnetization is the region between the Conrad and Mohorovičić discontinuities. We have not quantitatively considered the role of elevated temperature in the crust. Clearly, where the Curie isotherm is above the Moho, it will define the lower boundary of the magnetic layer. However, an evaluation of the influence of temperature on the magnetization model presented here would require more information than presently available on Curie points of lower crustal rocks and the distribution of temperature with depth over large areas (considering the spatial resolution of the satellite magnetic measurements). Most, but not all, laboratory measurements of lower crust xenoliths show approximately 560°C Curie points. In hydrous, oxidizing conditions, metamorphism leads to magnetite dominated, c. 560°C, Curie points; anhydrous, reducing conditions (e.g., Rio Grande rift) lead to ilmenite dominated Curie points of ~ 300°C (Wasilewski and Mayhew, 1982). However, only a very limited number of rocks have been analyzed. Likewise, although temperature profiles have been modeled for several specific locations in the U.S., they have large uncertainties and the spatial extent for which the model is appropriate is not specified. Of the localities modeled, depth to the 560° isotherm in the Colorado Plateau (McGetchin and Silver, 1972) and the Sierra Nevada (Blackwell, 1971) are below the Moho. However, modeled temperature distributions at Battle Mountain, Nevada (Lachenbruch and Sass, 1977), Yellowstone (Smith et al., 1974) and the Basin and Range in southern Nevada (Blackwell, 1971) place the 560°C isotherm close to or above the Conrad discontinuity. If these thermal models are accurate, and applicable for a large region, the lower crust could not be the source of long-wavelength crustal magnetic fields in these areas. Until more data is available on the spatial variation of Curie points and Curie depths it is not possible to quantitatively evaluate the relative role of temperature on satellite magnetic anomalies.

In a qualitative sense, however, in regions where the Curie isotherm is in the lower crust the magnetizations shown in Fig. 4 must increase as there is a smaller magnetic layer thickness to accomodate the required dipole moment. Heat flow and geothermal gradient determinations can be used to estimate, in this qualitative sense, the effect of undulations of the Curie isotherm on magnetization. The crust of the eastern part of the U.S. (approximately east of longitude 100°) is basically cool with the Curie isotherm well into the mantle, but high heat-flow regions (> 1.5 HFU) occupy much of the western U.S. (Blackwell, 1971; Lachenbruch and Sass, 1977). Much of the Basin and Range has high heat flow but also has high magnetization, even if the full lower crust thickness is considered. If the Curie isotherm is into the crust, as is modeled in both northwestern and southern Nevada, then the magnetization of the rocks above the Curie isotherm must be even higher. Thus, in high heat flow areas, the magnetization values on Fig. 4 should be considered minimum values.

Mayhew (1982b) noted an inverse relation between dipole moment (designated vertical integral of magnetization) with regional heat-flow in the western U.S. Using heat-flow data from Lachenbruch and Sass (1977) we find a small negative correla-

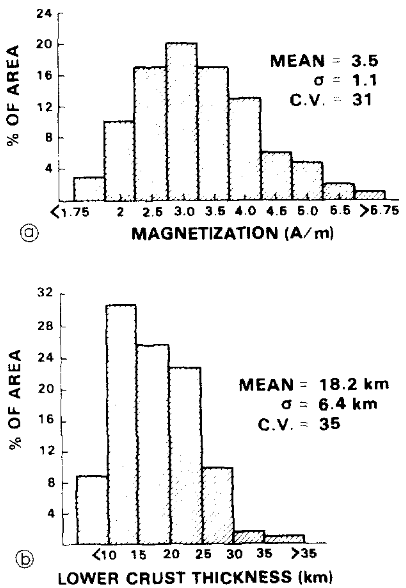

Fig. 5. Histograms of magnetization in the lower crust (a) and thickness of the lower crust (b). Number of grid points in both histograms is 266. *C.V.* is coefficient of variation.

tion ($r = -0.3$) between heat flow and dipole moment for the 161 grid points west of longitude 100°. Obviously other factors are also operative.

Figure 5a shows the distribution of absolute magnetization of the 266 grid points where magnetization was calculated from the dipole and lower crust thickness data. The mean is 3.5 A/m with a standard deviation of 1.1 A/m. Estimates of magnetization of the lower crust in various regions, made from modelied sources of long wavelenght aeromagnetic data range from 2 to 10 A/m (Caner, 1969; Hall, 1974; Elming and Thorne, 1976; Hahn et al., 1976; Krutikhovskaya and Pashkevich, 1979; summarized by Wasilewski and Mayhew, 1982). Excluding the earliest estimate, the four remaining estimates range from 2 to 5 A/m. Much of the Unites States' lower crust falls into the 2–5 A/m range; only about 15% of the area appears to have lower crustal magnetizations of less than 2 A/m or more than 5 A/m.

The main magnetic field is almost constant over the United States (0.54 ± 0.02 gauss). Therefore bulk magnetic susceptibilities of lower crust rocks are approximately twice the magnetization values of Fig. 4 or 5a. The mean magnetic susceptibility is 6.5 (SI units); a range of susceptibilities of 2.5–10.5 encompass 95% of the area. These susceptibilities are compatible with laboratory measurements of mafic rocks expected in the lower crust (Fig. 1).

Figure 5b is a histogram of lower crust thicknesses for the same 266 grid points. Comparing the variation of magnetization and thickness, it can be seen that the coefficient of variation of magnetization is slightly lower than of thickness (31 vs 35), but, to a first order, these two parameters have almost the same variation over the very large test region. At satellite altitude, the effect of variation of either is the same; that is, an anomaly can be caused by increasing one and decreasing the other in the same proportion. The histograms in Fig. 5 indicate that on a continental scale variations in lower crust thickness and magnetization contribute almost equally to variations in the crustal anomaly field.

SUMMARY

The source of long-wavelength magnetic anomalies, as measured by satellite, is most probably the mafic and ultramafic rocks of the lower crust. Using an inversion of the POGO magnetic anomaly field and a compilation of published deep-probing seismic data, we have calculated the magnetization of the lower crust, assuming negligible contribution from the sedimentary cover, upper crust, and mantle, and assuming the Curie isotherm is near or below the Moho at the spatial resolution of the satellite. Examination of lower crust thickness and magnetization shows that:

(1) Increased lower crust thickness account for the two largest highs observed in the satellite magnetic anomaly field–one centered over Kentucky and the other over northern Oklahoma; neither regions require high magnetization. However, two other, smaller, anomaly field highs over Missouri and the Colorado Plateau are in regions of relatively thin lower crust, necessitating high magnetization in the lower

crust of these regions. Regions which appear to have unusually low magnetization, such as Georgia, are areas of sparse lower crust thickness data.

(2) Magnetization for the lower crust, over those portions of the U.S. where lower crust thickness can be estimated, average 3.5 A/m, with standard deviation of 1.1 A/m. This magnetization is consistent with laboratory measurements of mafic and ultramafic rocks expected in the lower crust. It is also consistent with the 2–5 A/m range of lower crust magnetization recently estimated from aeromagnetic data. This self consistency supports the model of the lower crust as the primary source of the anomaly field.

(3) Lower crust thickness over the same region averages 18.2 km, with standard deviation of 6.4 km. The coefficients of variation of lower crust thickness and lower crust magnetization (35 and 31 respectively) indicate that, to a first approximation, variation in these parameters contribute about equally to the magnetic anomaly field variation.

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