MAGNETIZATION MODELS DERIVED FROM HIGH RESOLUTION SATELLITE LITHOSPHERIC ANOMALY FIELDS

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ABSTRACT

Three-component magnetic anomaly measurements simulated from the MF6 model [1] were inverted to find a minimum norm, continuously varying magnetization model using the method of [2]. Spherical harmonic degrees 16 to 120 were used to define the anomaly field, and the dataset was simulated at 300km altitude on a regular grid of 64621 points. A damping parameter controlling the relative importance assigned to goodness-of-fit and root-mean-square magnetization amplitude provides a means to overcome the inherent non-uniqueness in magnetization modelling. We set the magnetized layer thickness to 40km, taken to be the best global compromise between much greater depths to Curie temperature beneath old, stable cratonic regions and the thinner magnetized oceanic crust. Both induced and remanent magnetization are important in all areas – there is no strong support from our models for the assumption sometimes made that induced magnetization dominates over continental areas and remanent magnetization dominates in the oceanic crust. However, we use a variety of techniques to demonstrate that there are subtle, but significant, differences between the magnetization of continental and oceanic regions.

INTRODUCTION

The magnetic field originating in the Earth's lithosphere is known as the crustal anomaly field, since it is the small component remaining when the external, induced, and main (originating from geodynamo action in the core) fields have been removed. Magnetic field measurements from near-Earth orbiting satellites have provided information on the crustal anomaly field to increasingly high spherical harmonic degree as the volume of data has increased and methods for performing the separation of the various field sources have improved. The latest model, MF6 [1], has reliably determined coefficients up to spherical harmonic degree 120. We have used degrees 16 to 120 of MF6 to simulate 3-component vector measurements of the anomaly field at 300 km above the Earth's surface on a 1°x1° grid and inverted them for a global, continuously varying model of crustal magnetization using the method of [2], which finds the distribution of magnetization with minimum root-mean-square (RMS) amplitude subject to a given fit to the data. The relative importance attached to fitting the data and minimising the magnetization amplitude is controlled by a damping parameter. Since the data were simulated from spherical harmonic coefficients, they were assigned equal standard deviations. We make no assumptions about the magnetization direction, so both induced and remanent magnetization can result.

Our method requires a fixed, uniform magnetized layer thickness, whereas it is well known that the depth to the Curie isotherm (where rocks lose their magnetization) varies, being less beneath the oceans and greater beneath the continents. However, orbiting satellites at altitudes of a few hundred kilometres above the Earth's surface essentially 'see' the magnetized crust as a thin sheet, and provide no resolution of depth variation. Our assumed uniform thickness of 40 km is typical of continental crust thickness, and a compromise between thinner oceanic crust and thicker continental crust beneath old, stable cratons. The main effect of changing the layer thickness is to alter the average magnetization, such that the vertically integrated magnetization at a given point is constant. We also find that within a broad range of thicknesses, there is little overall change in misfit as thickness varies, for a given damping parameter. Thus our results are only weakly dependent on choice of layer thickness.

METHOD

Details of the method are given in [2] and references therein. Magnetization is expressed as a linear combination of the kernels relating the data to the model through integration over the volume of the magnetized crust. The multipliers are determined from the data by solving a data-by-data matrix system. Direct solution is impractical for such a large problem (3-components at each of 64621 points, i.e. 193863 data), but the satellite 'footprint' is small – only a small volume of the crust directly beneath contributes significantly to the observation. Thus most of the matrix elements appearing in the equations relating data to multipliers are negligibly small. We impose a threshold below which (in absolute value) elements are treated as if they were zero, and obtain a solution using a sparse iterative conjugate gradient method. The threshold is chosen such that altering it has a no discernible effect on the solution; we find that retaining (i.e. treating as non-zero) just 1% or so (a few hundred million) of the matrix elements is sufficient. Jacobi pre-conditioning improves convergence. The code to calculate the matrix elements parallelizes effectively.

Over a broad range of damping parameters, the pattern of magnetization is robust, but the RMS amplitude increases as damping parameter decreases. Beyond a certain point, stability of the pattern is lost, and the solution takes more iterations to converge, or does not converge at all. We take this to indicate we are attempting to over-fit, or fit noise in, the data. The damping parameter in our preferred solution is chosen to be larger than the value at which stability is lost; however, it is not so large that the amplitude of the solution fails to satisfy the Parker [3] bound, which for a maximum satellite field amplitude of 42 nT at 300 km altitude requires that magnetization must exceed 0.39 A/m somewhere in our 40 km thick magnetized layer. Since the RMS amplitude is less well constrained than the pattern of magnetizations, we must take care in its interpretation; in what follows, we concentrate on the relative magnetization amplitudes.

RESULTS

Our preferred model is presented in Fig. 1, both as orthogonal components and amplitude and direction, in projections centred on 0° and 180° longitudes for greater clarity. The RMS misfit between the data and their predictions by the model is 1.29 nT, and the RMS magnetization amplitude is 0.081 A/m. Fig. 1 shows magnetization at 20 km depth, the mid-point of the magnetized layer. The only significant variation with depth is in amplitude; at 20 km, the average magnetization is 0.113 A/m, calculated from the grid from which Fig. 1 was generated. Shallower magnetizations are stronger and deeper ones weaker to give the volume average of 0.081 A/m. Well-known magnetic anomalies such as Bangui and Kursk show up clearly as high amplitude magnetization features. The magnetization amplitude plots (top right of Fig. 1) make it particularly clear that magnetization is lower over the oceans than continents, but as remarked earlier this may reflect that the oceanic magnetization in our model is spread over a thicker layer than is appropriate. Magnetization is also high towards the poles, reflecting the stronger anomaly field values in the MF6 model used to synthesise the data. Angles defining the direction of magnetization in Fig. 1 are only shown where magnetization amplitude exceeds a certain threshold – they are determined from ratios of the field components, so are poorly constrained when the numerator and denominator are small, i.e. magnetization is weak.

Induced magnetization is in the direction of the main field, so by plotting the component of magnetization in the main field direction, and in two orthogonal directions, here chosen to be perpendicular to the main field in the meridian plane (the vertical plane containing the main field direction), and perpendicular to the meridian plane, we can investigate the extent to which our model is consistent with induced magnetization. Fig. 2 shows the result, indicating that the component in the direction of the main field (top panel) is of a similar magnitude to the component orthogonal to it in the meridian plane, and there is no obvious difference between oceanic and continental magnetization compared to the standard orthogonal components shown in Fig. 1. We cannot conclude that the magnetization component parallel to the main field is all induced, since it could be remanent magnetization acquired when the main field was anti-parallel to its current direction (or even remanent magnetization acquired in the current main field). However, the fact that the magnetization component perpendicular to the main field is equally strong shows that remanent magnetization is important, in both oceanic and continental crust. The component of magnetization perpendicular to the meridian plane is noticeably smaller than the other two, but the fact that it is non-zero indicates that magnetized material has been translated and/or rotated as expected from plate tectonic theory, assuming its magnetization was acquired in a field similar to the present (i.e. predominantly axial dipole).

We have undertaken a comparison between magnetization strengths and seismically-determined crustal thickness over continental regions from point values provided by W.D. Mooney (pers. comm., 2005), using a magnetization model deduced from an earlier, lower resolution, model in the MF series. Comparing point values is preferable to using values interpolated onto a regular grid or synthesised from spherical harmonic analysis. Statistically significant correlations between magnetization strength and crustal thickness were obtained, and the correlation coefficients increased if unusual magnetic features were excluded (e.g. the Bangui anomaly when undertaking a correlation for Africa). This

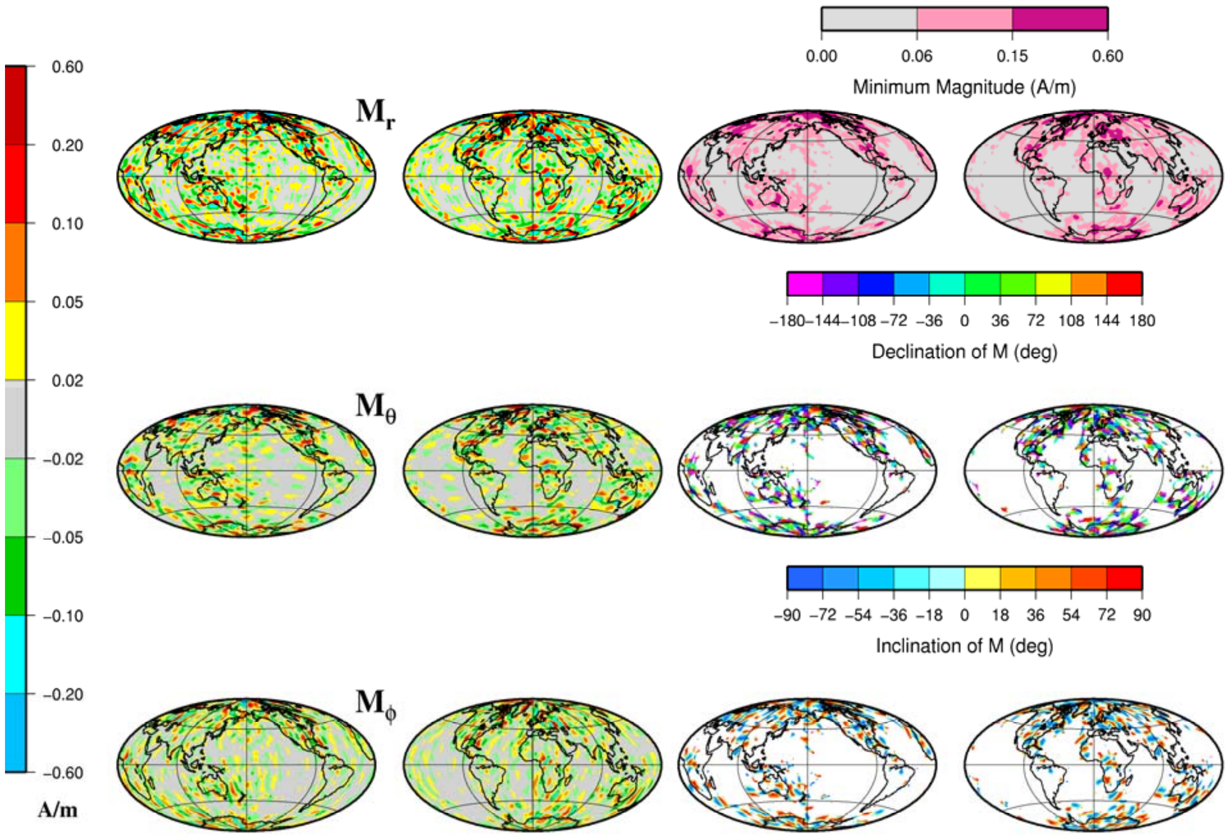

Fig. 1. Magnetization of our preferred model at 20 km depth. Left: radially outwards, north and east components (from top to bottom) using the scale bar to the left. Right: amplitude (top) and direction using scale bars above; declination (middle) is the angle from north and inclination (bottom) the angle from horizontal.

suggests that satellite-derived magnetization can be used as a proxy for crustal thickness when no other information is available.

CONTINENT-OCEAN CONTRAST

Many researchers have proposed that compositional, thermal, thickness and other differences between oceanic and continental crust mean that their magnetization signatures should be distinguishably different. We have seen (Figs. 1 and 2) that magnetization amplitudes are generally lower over the oceans, but noted that this could be because our assumption about the thickness of the magnetized oceanic layer is wrong. Thus we have undertaken some other analyses to address this issue. First we use a lower magnetization strength cut-off (0.3 A/m) for the directional plots, and display them separately for oceanic and continental areas (Fig. 3). The dark grey regions are continents on the oceanic plots, and vice versa for the continental plots, and the light grey areas are where the magnetization is less than the cut-off. Now the oceanic model shows patterns of up and down magnetization inclination symmetric about and parallel to the mid-ocean ridge in parts of the Atlantic Ocean. However, these are considerably thicker than their equivalents in the data we synthesised for inversion from the MF6 model, indicating that some resolution is lost in our modelling.

Fig. 4 plots median magnetization amplitudes as a function of main field amplitude (to account for the increase polewards) over the oceans and continents from the 1° grids of magnetization values. The error bars are one standard deviation, and are larger over the continents because their surface area is smaller so there are fewer values contributing to the estimates. Values are distinguishably different for the continents and oceans (with the latter higher), despite the large error bars for the continental points.

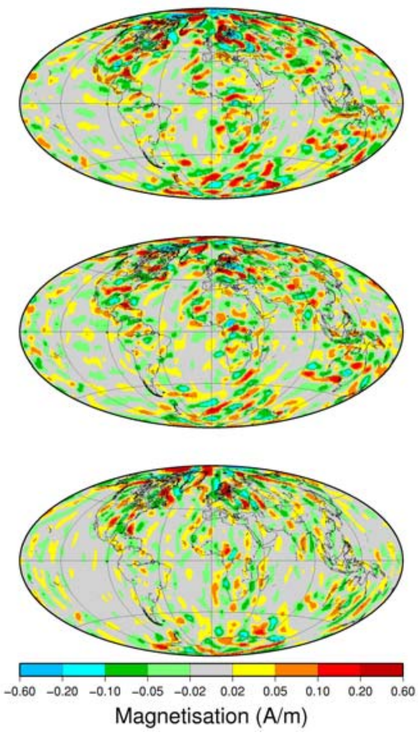

Fig. 2. Components of magnetization for our preferred model in the direction of the main field (top panel), orthogonal to the main field in the meridian plane (middle panel) and orthogonal to the meridian plane (lower panel).

Magnetization correlation length has been determined both using the covariance formula of [4] (Fig. 4), from the spherical harmonic field coefficients of the anomaly field, and by direct calculation from our model, giving a correlation length (in units of angle subtended at the centre of the Earth) of around 5°, regardless of which magnetization

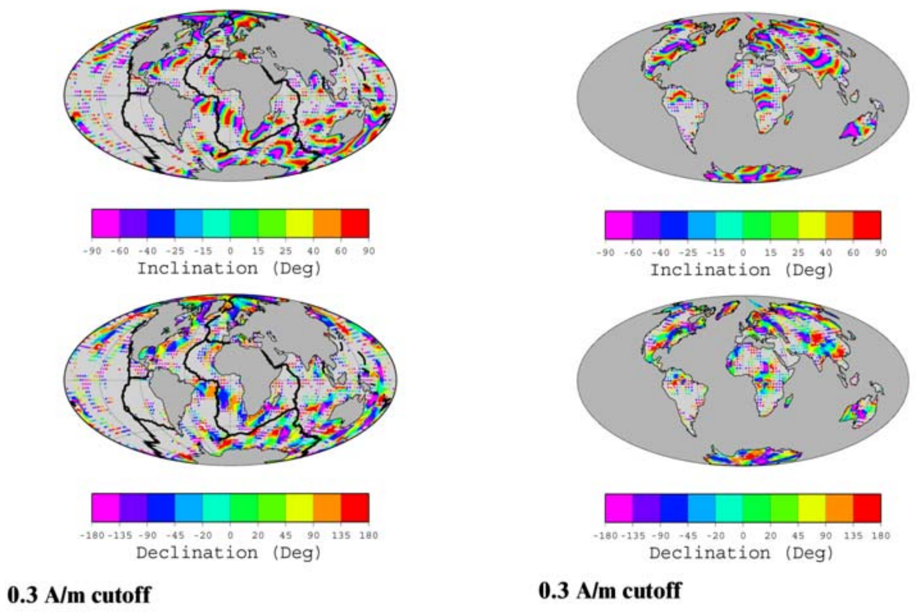

Fig. 3. Magnetization directions for the oceans (left), with ocean ridges superimposed, and continents (right). Dark grey regions are the continents (oceans) on the oceanic (continental) direction plots; light grey regions are where the magnetization is less than 0.3 A/m.

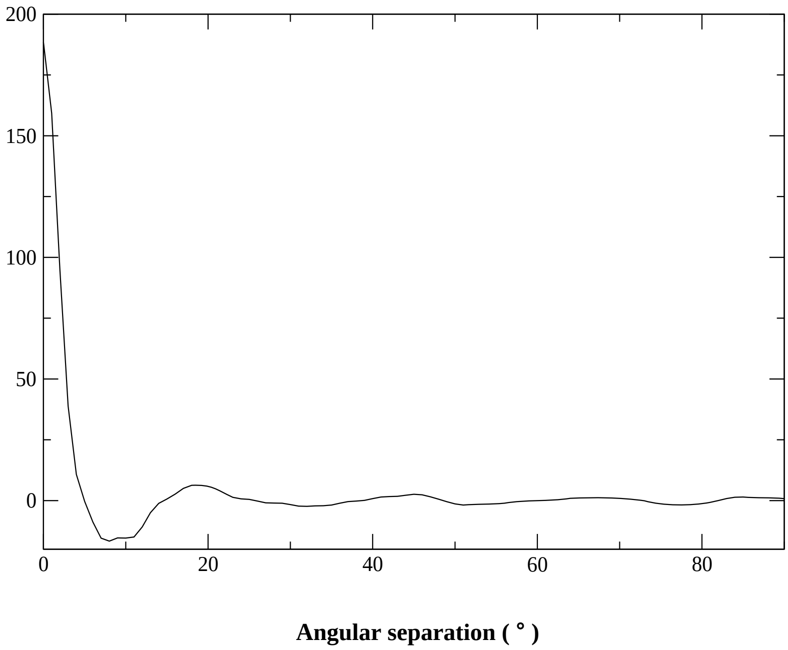

Fig. 4. Magnetization covariance as a function of angular separation in degrees using a formula given by [4] on MF6 coefficients between degrees 16 and 120, indicating a correlation length of about 5° (where covariance first vanishes). Covariance is in units of nT^2 (see [4] for details).

components are correlated. We have also analysed the directly calculated correlation length separately for the continents and oceans, which give values of 5° and 6° respectively, again regardless of which magnetization components are correlated. However, although the difference is consistent, the standard deviations on the correlation lengths are larger than the differences between the continental and oceanic values for the number of point pairs we used. Increasing the number of pairs to obtain smaller error bars would mean we would effectively be using the same point in multiple correlations, and so the estimates would not be independent.

FUTURE WORK

We plan to repeat the pointwise comparison of magnetization strength and seismically-determined crustal thickness using the higher resolution model presented here. Directional correlation of the anomaly field in the oceans shows a difference in correlation length along and perpendicular to isochrones (see [1]), demonstrating that MF6 resolves ocean stripes. We will undertake a similar comparison for our magnetization model to confirm (or otherwise) the suggestion of ocean stripes in Fig. 3. Data from SWARM, especially gradient data from satellites flying side-by-side, will provide even greater detail of the anomaly field which we hope to represent in magnetization models.

ACKNOWLEDGEMENT

KAW was partially funded by NERC grant NER/O/S/2003/00674.

REFERENCES

[1] S. Maus, et al., “Resolution of direction of oceanic magnetic lineations by the sixth-generation lithospheric magnetic field model from CHAMP satellite magnetic measurements”, *Geochem. Geophys. Geosyst.*, vol. 9, Q07021, doi:10.1029/2008GC001949, 2008

[2] K. Whaler and M. Purucker, “A spatially continuous magnetization model for Mars”, *J. Geophys. Res.*, vol. 110, E9, E09001, doi:10.1029/2004JE002393, 2005

[3] R.L. Parker, “Ideal bodies for Mars magnetics”, *J. Geophys. Res.*, vol. 108, E1 5006, doi: 10.1029/2001JE001760, 2003

[4] C. Rygaard-Hjalsted, C.G. Constable and R.L. Parker, “The influence of correlated crustal signals in modelling the main geomagnetic field”, *Geophys. J. Int.*, vol. 130, pp. 717-726, 1997