Chapter 1

Introduction

Looking into the Earth's interior has always been a challenge because the deepest boreholes penetrate not more than 10 km and that depth has been reached in only few locations e.g. in Germany (Baum et al., 1995; Emmermann and Lauterjung, 1997). All sorts of measuring techniques and methods have been employed to reveal the mystery of deep inside. Geology, structural properties of rocks, surface geophysics, borehole geophysics, petrophysics, laboratory experiments, geochemistry, geochronology, fluid analysis, heat flow measurements have regularly been employed in all possible geometry to infer the rock types at various depths. Deep seismic sounding, gravity interpretation, geophysical well-logging, electrical methods and magnetic methods have contributed to the geophysical interpretation of the rock types lying at shallow and great depths. Not only these geophysical methods have benefited immensely to the exploration geophysicists to locate minerals and oil and gas reserves in the Earth but have also contributed to the understanding of the geophysical, geochemical and geological processes deep inside the Earth.

One of the geophysical methods, which have been regularly used for both exploration and understanding of deep interior of the Earth, is the magnetic method. Magnetic methods have been known now for more than 400 years since William Gilbert published *De Magnete* in 1600. Magnetic methods gained popularity and importance once it was established through the study of geomagnetism that Earth is a great magnet. Geomagnetism deals with the study of Earth's magnetic field. Earth's magnetic field not only changes with the latitude but also with time. To find the sources for these variations in the magnetic field recorded at the Earth's surface, more detailed studies were conducted. It is now known that the magnetic field at any location near the Earth can be attributed to a combination of four sources located respectively in the Earth's core, in the mantle, in the Earth's crust and in the Earth's ionosphere and beyond.

The largest in magnitude is the field from the core, or the 'main field'. Near dipolar in nature, the strength of the main field is approximately 60000 nT (nanotesla) at the poles and approximately 30000 nT at the equator. The variation of the main field with time is called its 'secular variation'. External current systems are time-varying on a scale of seconds to decades, which can cause near-Earth fields to vary in magnitude from fractions of a nT to thousands of nT. The external current systems are located in and on the boundary of a cavity-like region surrounding the Earth, known as the magnetosphere, and in the Earth's ionosphere. The magnetosphere is a region which protects the earth from bombardment of charged particles from the Sun. Although always present, the strength and location of these currents varies considerably between magnetically quiet and disturbed times. Because they vary with time, the external fields cause induced currents in the Earth resulting in an 'induced field'. The ionosphere is a region above the Earth's surface, where the atmosphere is ionized due to the Sun's ultraviolet radiation. The resulting high conductivity in the ionosphere causes electric currents. These currents cause field variations in the ionosphere, which vary not only with respect to sun lit regions but also with latitude. Fields from the Earth's crust again range from fractions of a nT to thousands of nT. In general, the crustal source dimensions are small compared to the external field sources. Also called 'anomaly fields', crustal sources are associated with variations in the geologic and or geophysical properties of the Earth's crust. A portion of this field is from induced magnetisation and will follow variations of the main field. Another portion is remanent magnetisation, i.e., it is frozen in the rocks and varies on a geologic time scale. The contribution of magnetisation in the uppermost mantle to the geomagnetic field is still debated but is generally considered to be minor because of the high temperature of the rocks of the mantle. The temperature within the Earth increases with depth, and at a certain depth it is above the Curie temperature of the dominant magnetic minerals, where they lose their ferromagnetic properties. Curie temperature for magnetite is 578^0 C, and this isotherm marks the boundary below which magnetisation from the rocks of the upper mantle does not contribute significantly to the anomaly fields measured at the Earth's surface.

1.1 Some pertinent definitions

The magnetic field defined above as the 'main field' is the Earth's magnetic field, which starts from the liquid outer core of the Earth. It has a direction that is usually neither horizontal nor-in-line with the geographic meridian. The magnitude of this field 'F', the inclination (or dip) of the needle from the horizontal, I, and the angle it makes with geographic north (the declination), D, completely define the main magnetic field.

Any magnetizable body placed in an external field (here it is the main field of the earth) becomes magnetized by induction. Magnetisation is due to the reorientation of atoms and uncoupled electron spins. Magnetisation is measured by the magnetic polarization **M**, also called magnetic dipole moment per unit volume. The degree to which a body acquires induced magnetisation is determined by its magnetic susceptibility, *k*. Susceptibility is a fundamental rock parameter in magnetic exploration. The magnetic response of rocks and minerals is determined by the susceptibility of magnetic materials (mainly magnetite and pyrrhotite) contained in the rocks. They generally persist both over small and large distances; thus magnetic maps generally exhibit both small-scale and large-scale regional features. These features are called the 'crustal field' or the 'anomaly field' as defined earlier. The vertical component of this field is the vertical field anomaly and is positive downward. The total field anomaly or the scalar anomaly is the component of this field in the direction of the main field.

The magnetic field measured at any point near the surface of the earth is the sum of the main field, the field due to crustal magnetisation and the external field. The field from the core and the crust comprises the internal part of the total field. Our effort here is to study only the crustal component of this internal magnetic field. The sources for these crustal anomalies as defined above are primarily due to strong magnetisation exhibited by some specific minerals, which are discussed next.

Magnetism in rocks can be divided on the basis of their behaviour when placed in an external field (Telford, 1990; Blakely, 1996). *Diamagnetism* is an inherent property of all matter. In diamagnetism, an applied magnetic field disturbs the orbital motion of electrons in such a way as to induce a small magnetisation in the opposite sense to the applied field. Consequently, diamagnetic susceptibility is negative. *Paramagnetism* is a property of those solids that have atomic magnetic moments. Application of a magnetic field causes the atomic moments to partially align parallel to the applied field thereby producing a net magnetisation in the direction of the applied field. The susceptibility is positive. Both diamagnetic and paramagnetic materials are insignificant contributors to the geomagnetic field.

Certain materials have atomic moments where neighbouring moments interact strongly with each other. This interaction is a result of quantum mechanical effect called exchange energy. Certain paramagnetic elements, namely iron, cobalt and nickel, have such strong magnetic interaction that the moments align within fairly large regions called domains. This effect is called *ferromagnetism*, and it is ~ 10^6 times the effects of diamagnetism and paramagnetism. Ferromagnetism decreases with increasing temperature and disappear completely at the Curie temperature. The domains in some materials are subdivided into subdomains that align in opposite directions, so that their moments nearly cancel. Such substances are called *antiferromagnetic*. The common example is hematite. In some materials, the magnetic subdomains align in opposition but their net moment is not zero, either because one set of subdomains has a stronger magnetic alignment than the other or because there are more subdomains of one type than of the other. These substances are *ferrimagnetic*. Examples of the first type are magnetite and titanomagnetite, oxides of iron and titanium. Pyrrhotite is a magnetic mineral of the second type. Practically, all relevant magnetic minerals in the Earth's crust are ferrimagnetic.

1.2 Need for the present study

The crust primarily consists of igneous, sedimentary and metamorphic rocks. Sedimentary rocks generally have low magnetisations because of the lack of magnetite whereas the underlying crystalline basement of igneous and metamorphic rocks can be highly magnetic. Exceptions to the sedimentary rocks are the banded iron formations (BIF), which can be highly magnetic due to large amounts of magnetite they contain. The magnetic method has provided a primary tool for mapping the structure of the crystalline rocks of the continental crust. These mapping abilities can be crucial where rocks have limited exposure or no exposure at the surface because they are generally covered by sediments or water. Anomaly maps derived from magnetic data taken with airborne and shipborne instruments are routinely used in the preparation of geologic and geophysical models of the crust. These surveys concentrate on geologic features less than about 50 km in lateral dimension. However, in past decades there has been increasing interest in studying large-scale anomalies, hundreds of kilometers in extent, that appear in regional compilations of data from airborne surveys. These large-scale anomalies are referred to as the long-wavelength anomalies. Spatial variations in these anomalies reflect variations in crust and upper-mantle magnetisation. Crustal magnetic field variations are caused not only by variation in the amount of magnetic minerals but also reflect variations in the crustal thickness, the depth to Curie isotherm, and geochemical nature of the rock source. Shive et al. (1992) claimed that long wavelength anomalies are generally caused by sources that lie deep in the crust and that lower crustal rocks are much more magnetic than the rocks exposed at the surface. Studies of such long wavelength anomalies or 'regional anomalies' have spurred renewed laboratory investigations of rock magnetisation especially for rocks present in the crust (Rochette, 1992).

Mapping global crustal magnetic fields started when the first series of satellites mapping only the scalar field anomaly revealed one of the strongest anomaly at the satellite altitude, the 'Bangui anomaly', centered at Bangui, capital of Central African Republic (Regan et al., 1975). This anomaly is located 6^0 N of the equator. Later satellite missions mapped both scalar and vector components of the magnetic anomalies to locate the sources more accurately. These satellite anomaly maps provide a distinct advantage over the conventional aeromagnetic surveys. Mapping global scale anomalies can be easily accomplished without respect for any political boundary. They are of uniform precision and spatial distribution. The data can be acquired within a short span of time without actually getting interfered by time varying Earth's main field. This possibility also helps in studying the regional anomalies without having to knit together data from surveys taken at different times and reduced under differing models of the core field (Hinze and Zietz, 1985). One of the principal contributions could be to extract a geographical subset from the global distribution of data for further detailed analysis. Here, we define long wavelength anomalies as those which are resolvable by satellite measurements (half-wavelength > 200 km). Having pointed out the advantages that these satellite data have over data acquired by conventional magnetic methods, let us look at the developments in the mapping of crustal field anomalies using satellite data and their impact over the last three decades.

1.3 Lithospheric field models

The POGO (1965-1971) satellite missions, Magsat (1979-1980) and, more recently, Ørsted (February 1999 onward), CHAMP (July 2000 onward) and SAC-C (November 2001 onward) in near-Earth orbits have helped us to understand not only the Earth's internal field but also the external fields with sources in the ionosphere and the magnetosphere. Based on the available satellite data from the last three decades, several lithospheric field anomaly maps have been computed by various workers. The POGO missions resulted in the first consistent map of crustal magnetic fields on a global (Regan et al., 1975) and regional scale (Langel and Thorning, 1982). These maps were primarily scalar magnetic anomaly maps. The satellites orbited within an altitude range of 400 to 1510 km and, hence, due to poor lateral resolution and limited accuracies of their magnetometers were not suited to map crustal fields accurately. Magsat orbited at altitudes of 550 km down to 300 km from October 1979 to June 1980. Maps produced from Magsat data resolve magnetic features with wavelengths down to about 500 km (Langel et al., 1982a; Yanagisawa and Kono, 1985; Cain et al., 1989a; Cohen and Achache, 1990; Arkani-Hamed et al., 1994; Ravat et al., 1995; Sabaka et al., 2000). However, inaccuracies of the star cameras affected the quality of the vector data. The launch of the Ørsted satellite in February 1999, after a gap of nearly two decades, led to the derivation of highly accurate main field models (Olsen et al., 2000; Olsen, 2002), but resolution of the crustal field was affected by the higher orbital altitude of the satellite (650–865 km). In July-2000, CHAMP (http://op.gfz-potsdam.de/champ/main_CHAMP.shtml) was launched into a low Earth orbit of 450 km. The present altitude of its almost circular orbit is 400 km. Its magnetometers resolve the crustal magnetic anomalies more accurately than any of its predecessors. Maus et al. (2002), using the CHAMP scalar and vector data, derived a first spherical harmonic expansion of the crustal magnetic field. The first revision MF2 (http://www.gfz-potsdam.de/pb2/pb23/SatMag/model.html) of the model of the earlier map prepared by Maus et al. (2002) includes more CHAMP satellite vector and scalar data and, hence, is a more accurate crustal field model than its predecessor. We therefore use the MF2 model for all our subsequent studies.

One way to characterise the improvements of different crustal field models is to compare their power spectra. Magnetic anomalies of the crust cover a broad band of wavelengths (or wavenumbers). A quantitative summary of the importance of different wavelengths at the Earth's surface can be obtained by considering separately the fields for various spherical harmonic degrees n by computing the power spectrum. The strength of the internal part of the field at wavelengths corresponding to n, can be measured by the square of the average over the Earth's surface. The expression of power spectrum used here is shown in Equation (1.3) and is divided by a factor $2n+1$. Figure (1.1) shows the power spectra (Mauersberger, 1956; Lowes, 1966) of the total intensity anomaly at 400 km altitude of several lithospheric field models, together with the spectrum of our CHAMP model (Maus et al., 2002). The comparison of power spectra of different lithospheric field models shows a discrepancy in Figure (1.1) by a factor 10 in magnitude. Figure (1.2) shows the power spectra of field models computed by our group at GFZ using the CHAMP and Ørsted data. Interestingly, all the spectra exhibit similar power, stating that the field models derived using these data are more reliable and accurate than the models derived using Magsat data. This is an exciting result for the larger geomagnetist community because the field models derived using POGO and Magsat satellites were partly erroneous due to the error in their raw data (Ravat et al., 1995) and the new satellite data is now providing increasingly accurate field models. There were obvious discrepancies in interpreting different field models derived using POGO and Magsat satellite data. Therefore, they were not able to make much of an impact to the contribution of solid earth sciences. On the contrary, after a gap of nearly two decades, the lithospheric field models computed using the Ørsted and CHAMP data provided a very good estimate of magnetic anomaly features. These field models are compared in the Figure (1.5) for CHAMP and in Figure (1.6), for Ørsted derived vertical field anomaly maps. Even the small-scale anomaly features are delineated (half wavelength ~ 250 km) accurately and the patterns match well. This shows that the anomalies observed by these satellites are genuine and real.

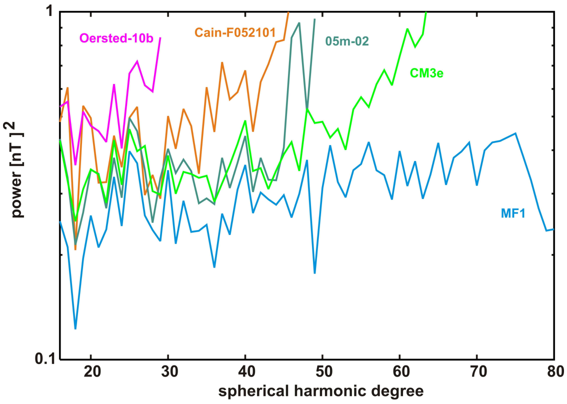


Fig. 1.1. Power Spectra of the total intensity anomaly (Mauersberger-Lowes spectra/2n+1) at 400 km altitude of lithospheric field models derived by various workers from Magsat, POGO, Ørsted and CHAMP data.

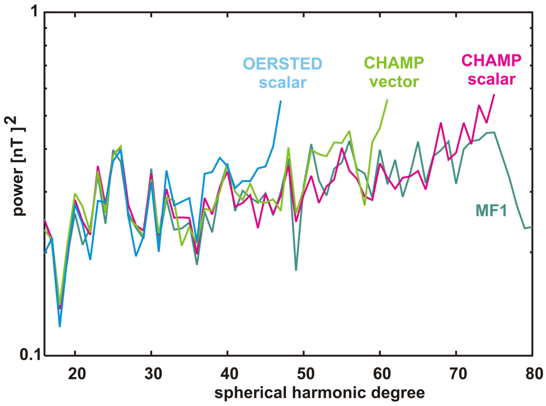


Fig. 1.2. Power Spectra of the total intensity anomaly (Mauersberger-Lowes spectra/2n+1) at 400 km altitude of Ørsted and CHAMP lithospheric field models derived by our group.

1.4 Preparation of field models for interpretation

The vertical field as seen by CHAMP is shown in Figure (1.3). It shows the dominance of magnetic field from the Earth's core (main field) and masking of the magnetic field from the crust and, hence, no meaningful information about the crust can be extracted from this map. In order to study the Earth's magnetic field of the crust it is necessary to remove the field of the core and interpret the residual map. It would seem that to differentiate and eventually separate the field originating from the sources within the crust and that from the core of the Earth would be an easy task. On the contrary, it is not so. From the Figure (1.3), it is evident that the long wavelengths caused by the Earth's core or the main field masks the signals from the crust and hence we cannot see the signals from the crust. To find a solution to the problem, Langel and Estes (1982), worked out an easy way by analyzing the power spectra of the internal field model and subsequently removing the long wavelength features. The procedure is described below.

As the coverage of the satellite survey is in a spherical shell, the most common technique to construct field models from satellite data employs spherical harmonic analysis. Gauss showed as early as 1839 that the magnetic field due to sources internal to the Earth could be represented at any point (r, ϕ, θ) outside the Earth:

$$B = -\nabla V \tag{1.1}$$

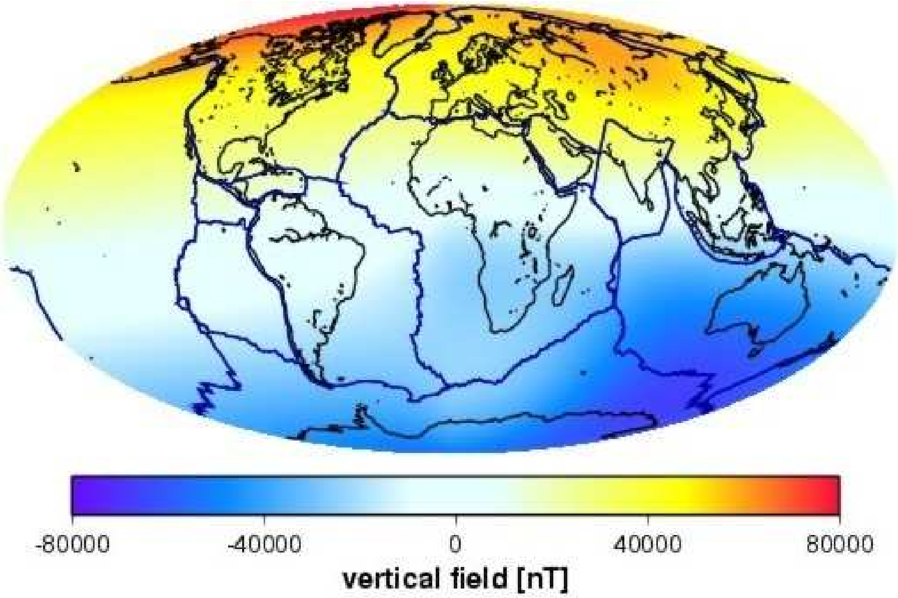


Fig. 1.3. Internal field model for the vertical component at an altitude of 400 km.

where V can be replaced by a spherical harmonic expansion of the form

$$V = a\sum_{n=1}^{N}\sum_{m=0}^{n}\left(\frac{a}{r}\right)^{(n+1)}\left[g_n^m \cos m\phi + h_n^m \sin m\phi\right]P_n^m(\cos\theta) \tag{1.2}$$

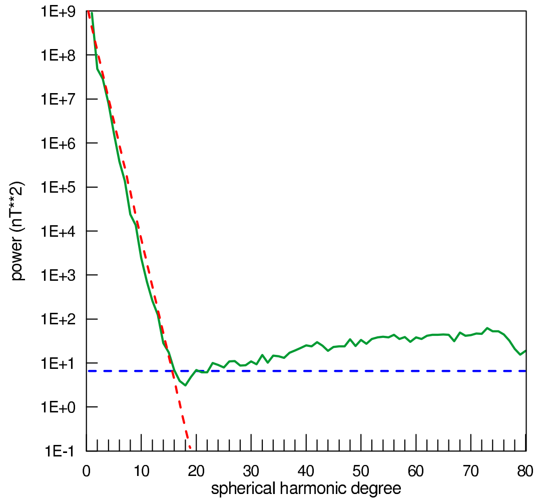
where, a is the mean radius of the Earth, and P_n^m are the Schmidt normalized associated Legendre polynomials. The terms g_n^m and h_n^m are called the Gauss coefficients and are identified by their degree n and order m.

Langel and Estes (1982) used these spherical harmonic coefficients to estimate the Mauersberger-Lowes power spectra (Mauersberger, 1956; Lowes, 1966) of the global magnetic field model given by:

$$\text{Power} = R_n = (n+1)\sum_{m=0}^{N}\left[\left(g_n^m\right)^2 + \left(h_n^m\right)^2\right] \tag{1.3}$$

They plotted the power spectrum for their global magnetic field model. Langel and Estes (1982), interpreted the steep part of the spectrum (n<14) to be caused by the sources in the Earth's core i.e., dominated by the main field or the core field and the flatter part of the spectrum to be dominated by the crustal sources. They removed the degrees less than 14 to obtain the crustal magnetic anomaly map by attributing the higher degrees (n>14) to be mainly from the crust. The transitional degrees 13-15 were attributed to signals from both the core and the crustal sources.

Fig. 1.4. Mauersberger-Lowes power spectra (Mauersberger, 1956; Lowes, 1966) of an internal field model from CHAMP and Ørsted data.



Following a similar approach, using the internal field model derived using Ørsted and CHAMP data, the power spectrum is computed for all the degrees 1-80 and is plotted in Figure (1.4). MF2 model is calculated up to a maximum spherical harmonic degree of 80 (http://www.gfz-potsdam.de/pb2/pb23/SatMag/model.html). The figure shows a change in the slope of the curve between spherical harmonic degrees 13-15. This means the degrees 1-13 are dominated by the long wavelengths of the main field. The degrees 14-15 are transitional as both the wavelengths from the main field and crustal field are present. However, the higher degrees from 16-80 are now dominantly crustal in origin (Fig. 1.4). The present work retains this distinction based on the power spectrum studies and spherical harmonic degrees 1-15 are removed from the internal magnetic field model to produce crustal field anomaly map for degrees 16-80. In the process, some part of long wavelength features of the crust is also removed and is evident from the Figure (1.4). Arkani-Hamed and Strangway (1985) and Harrison et al. (1986) based on a similar study independently concluded that the crustal portion of the magnetic field is limited to degrees 19 and greater. However, removing degrees 1-18 would further reduce the long wavelength component from the crust and, hence, it is not appropriate to interpret only degrees 19 and beyond.

The crustal field model at an altitude of 400 km computed using CHAMP data for degrees 16-80 for the vertical component is shown in Figure (1.5). The anomaly patterns now reveal features having sources in the crust. This map can now be used to find the sources causing the anomalies. Just to confirm the consistency and genuine existence of these anomaly patterns the degrees 16-80 using only Ørsted data is also shown in downward continued map at the same altitude (Fig. 1.6). A comparison of the two maps clearly points that the anomaly patterns match well. Even the comparable power spectra of the two maps already discussed in section (1.3) (Fig. 1.2), show that the anomaly patterns are real and genuinely of crustal in origin. It is hoped that interpretation of such maps should reveal new information and also confirm previous results about the magnetic sources within the crust. It should be quoted that even though Ørsted satellite is in higher orbit the data is utilized to produce such good lithospheric map and it matches extremely well with the one produced by much lower orbital altitude of CHAMP. We now discuss the usefulness of these maps and the modelling techniques used till now to retrieve information from them.

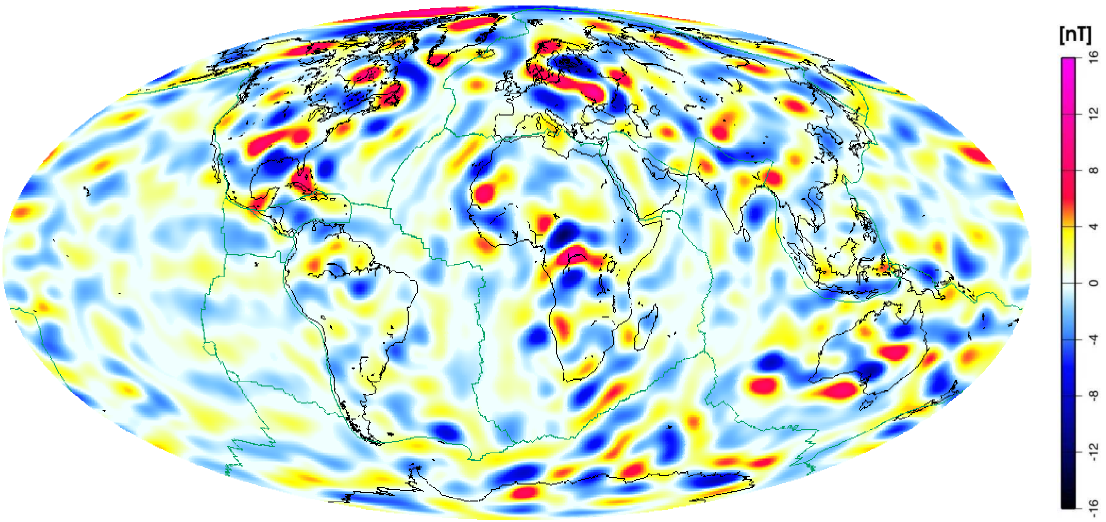


Fig. 1.5. Lithospheric field model for vertical component derived from CHAMP scalar data at 400 km for degrees 16-80.

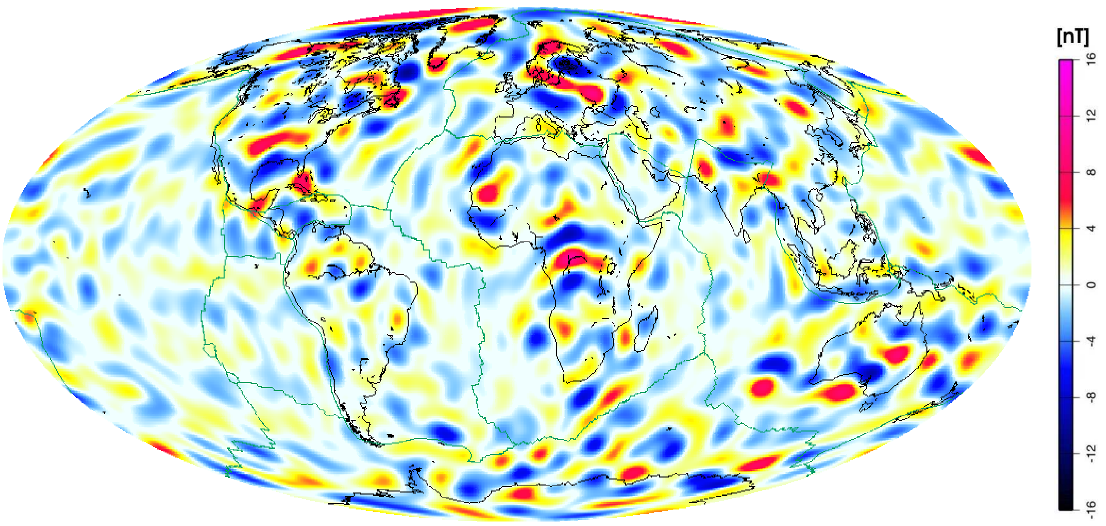


Fig. 1.6. Lithospheric field model for vertical component derived from Ørsted scalar data at 400 km for degrees 16-80.

1.5 The objective of deriving the lithospheric field models

One objective of deriving these crustal field maps is to characterize the sources causing these anomalies and to interpret these maps in terms of known geology. The interpretation methods commonly employed often consist of visual comparison of an anomaly map with geologic or tectonic maps. Many of the earliest interpretive studies of satellite anomaly maps were of this type. Hastings (1982) compared the features of scalar

Magsat anomaly map with that of the tectonic map of Africa. Induced magnetisation was assumed as the source for the anomalies in their work. Frey (1982a) using Magsat data compared the crustal scalar magnetic anomaly map with the global geological map and showed that large-scale features such as shields, cratons and subduction zones are associated with positive anomalies and basins and abyssal plains with negative anomalies. Similar studies for Asia were illustrated by Frey (1982b), for South America by Hinze et al. (1982), and for India by Achache et al. (1987). They attributed the negative anomalies to crustal thinning associated with elevated Curie isotherm in the region and vice-versa for the positive anomalies. These interpretations are oversimplified. In the above context, the Figures (1.7a), (1.7b) and (1.7c) illustrate the response for the total field intensity and the vertical field of a magnetizable body placed in an inducing main field of the earth. The positive and negative anomalies mentioned above are the total intensity (scalar) anomaly measured at various latitudes, as shown in Figures (1.7a), (1.7b) and (1.7c). Thus, a magnetizable body placed in the inducing main field of the Earth can be replaced by a dipole of comparable magnitude and direction. The large geological regions mentioned above were modelled by those authors accordingly, to find the direction and magnitude of magnetisation.

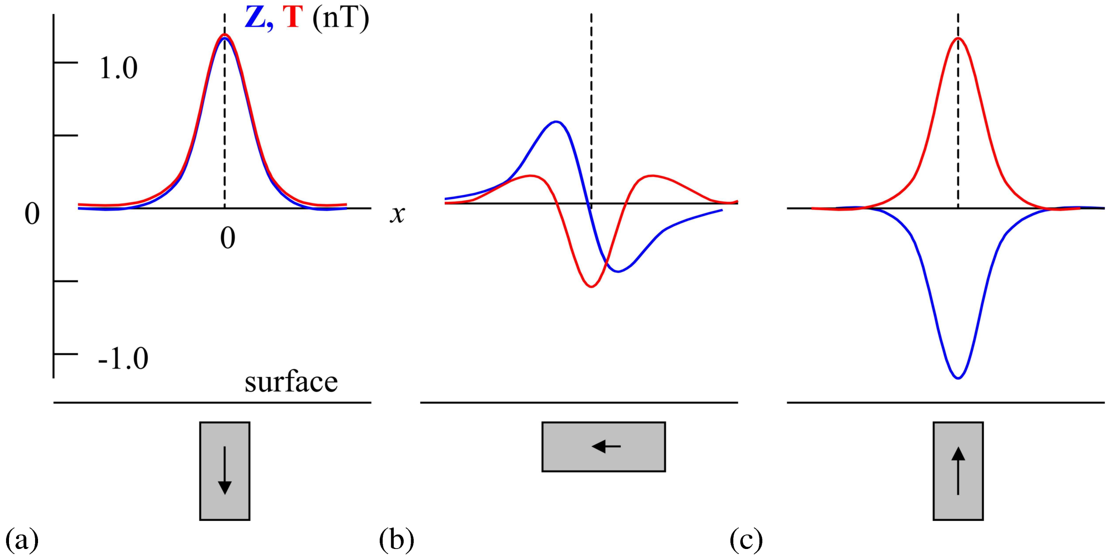


Fig. 1.7. Profiles for (a) vertical dipole located at north pole, (b) horizontal dipole located at equator, and, (c) vertical dipole located at south pole. Lines in red are the total field anomaly and in blue the vertical field anomaly for a dipole (marked in arrows) induced and directed along the main field of the earth.

More quantitative interpretations involve direct comparisons between geophysical quantities such as gravity, heat flow and depth to the Moho. The ultimate aim, however, is to extract geological and tectonic information from the crustal anomaly maps and to explain the inferred geology in terms of tectonic and geologic processes. The distribution of magnetisation is not uniquely determined by the magnetic field and, hence, is often estimated using constraints from other geoscience data (e.g., surface geology, gravity, and heat flow). Mayhew (1979), von Frese, Hinze and Braile (1981) and Whaler and

Langel (1996) followed such methods to find the susceptibility and magnetisation distributions.

Forward modelling methods based on initial model information are those given by Meyer et al. (1983) and Hahn et al. (1984). These are global models of the crust composed of two- or three-layers. The crustal model defined by Meyer and Hahn was global grid representing the various geological regions. Assuming only induced magnetisation, each block-layer was represented by a dipole directed along the main field, with susceptibility assigned based on a classification into one of 16 crustal types. The classification was based on surface geology and seismic investigation. Layers classifications, with magnetisation in amperes per meter in a 50,000nT field in brackets, were as follows: sea water (0), sediments (0), upper oceanic layer (1.0), upper continental layer (1.0-2.0), lower crustal layer (0.5), intermediate crustal layer (1.2), and sub-Moho (0). Comparison of the modelled magnetic field showed some local agreement with measured anomaly maps, but in general, their models did not reproduce the observed anomaly features. This initial model proved to be too general and hence the results were largely in disagreement with the Magsat derived magnetic anomaly map.

Purucker et al. (1998) derived a simple *a priori* model magnetisation distribution for both continents and oceans, termed “Standard Earth Magnetisation Model”, or SEMM-0. The continent-ocean margin was placed at the 1-km bathymetric contour. Values of susceptibility were assigned to continental crust (0.025 SI units) of thickness 40 km, and oceanic crust (0.04 SI) of thickness 7 km, and remanent magnetisation (0.1 SI) to Cretaceous Quiet Zones (KQZ) within the oceanic crust. The model takes the form of 11,562 equivalent point dipoles, each representing a region equal to a 1.89^0 X 1.89^0 area at the equator. Except in the KQZs, the magnetisation direction was assumed to be along the direction of the ambient main field. Using a random search inverse method, an equivalent source magnetisation model (SEMM), was derived by modifying the initial *a priori* model SEMM-0 until the resulting anomaly pattern matched a Magsat-derived magnetic anomaly map in a least-squares sense. SEMM models were presented in terms of the product of susceptibility times thickness. The inherent drawback of the SEMM-0 model is the simplification of magnetic properties of the crust, particularly for the continents where lateral variations in susceptibility are large. Recently, Purucker et al. (2002) modified the continental crustal thickness of their *a priori* model based on heat flow provinces of the world. Cratons are older crust and hence were considered to be associated with low heat flow. Basins are made of younger crust and comparatively have a higher heat flow (Taylor and McLennan, 1985). Based on the heat flow value in the cratons and in the younger crust the thicknesses of the cratons was estimated to be 81 km while in the basins it was only 20 km. The basic methodology to arrive at the final model remained the same. This model also lacked lateral variation in magnetic properties of the crust, though the incorporation of heat flow measurements improved the results marginally.

In addition to the drawbacks of the above models, the interpretation of magnetic anomalies is further complicated by certain uncertainties, which are discussed next.

1.6 Ambiguities in interpretation

The interpretation of anomaly maps is not straightforward but is complicated by ambiguities inherent in the interpretation of magnetic anomaly maps. One of the first ambiguities is that it is not possible to distinguish whether the anomaly pattern is due to remanence or induced magnetisation. This means all the anomaly patterns observed can be explained just by induced magnetisation (Maus and Haak, 2003). Making the interpretation more complex is the lack of vertical resolution. Satellite and airborne magnetic data can only infer height integrated magnetisation, i.e., thickness multiplied by susceptibility. So, it is immaterial to place the source body at the surface of a region or deep inside, as the satellite data cannot distinguish its location in the vertical column. However, *a priori* information like stratigraphy can partially alleviate the problem. One of the inherent problems is the unability to remove the short wavelength component of the main field completely from the lithospheric anomaly maps. Removal of the long wavelength part of the main field also removes the long wavelength component of crustal sources. A further ambiguity is caused by the presence of magnetic field annihilators. Some susceptibility distributions across the magnetic equator in the presence of a dominantly dipolar inducing field produces no observable magnetic field outside the earth (Maus and Haak, 2003). Another annihilator derived by Runcorn (1975), states that a homogeneous spherical shell magnetized from the inside does not produce a magnetic field outside of the Earth. The presence of anomalies due to remaining external fields aggravates the problem, especially in the polar and equatorial regions where the external sources show most affect. In order to reduce the affects of these ambiguities, a forward modelling technique to interpret the anomalies is followed. The modelling method employs some simplifying assumptions, which are described in the following paragraph.

1.7 The present work

The work presented here is a Geographical Information System (GIS) based forward modelling technique developed to model the various geological units of the continents and oceans using a standard geological map (CGMW, 2000). Unlike the global models of Hahn et al. (1984) and Purucker et al. (1998), which assumed a single susceptibility value for all the continents in their initial model, the initial global model derived here is highly variable and each geological region of the world has a unique susceptibility value. The derivation of this susceptibility value depends upon the known rock types exposed in a geological region, and a standard susceptibility value is derived by mathematically averaging the susceptibility values for all the identified rock types. Multiplying the average susceptibility value for the geological region of the upper crust by a constant factor derives the susceptibility value for the lower crust. This factor is based on the work by Taylor and McLennan (1985) who proposed an average composition of the upper and lower continental crust. Using the global seismic crustal structure, a vertically integrated susceptibility (VIS) model is computed for a geological region by integrating the contribution of average susceptibility times thickness of the upper crust and susceptibility times thickness of the lower crust below that region. A global VIS model is derived by knitting the VIS value for all the geological units of the world. Starting with this initial global VIS model, the vertical field anomaly is computed at an altitude of 400 km and compared with the CHAMP vertical field anomaly map. The model derived first is the *initial model*. In the next iteration, some specific regions are selected for detailed analysis where the *initial model* is in disagreement with the observed magnetic anomaly map. The parameters, which are to be defined later in chapter 2, are changed based on the information from other published results, to improve the fit between the predicted and observed magnetic anomaly map. This refined model, called the *first iteration* model, is then used to draw inferences about the tectonic and geologic features underlying the younger crust and the nature of the less-understood lower crust.

Assumptions in the present work:

The present work incorporates some simplifying assumptions because of the inherent ambiguity of the magnetic inverse problems.

1. Moho is considered a magnetic boundary. This is supported by the work of Wasilewski and Mayhew (1992), who find the Moho to be the lower boundary of the magnetic crust above a non-magnetic upper mantle below. This implies that the region above the Moho discontinuity is the sole source for the crustal anomaly. The assertion of Haggerty (1978) that the serpentinisation of upper mantle may produce magnetites is still debated, as the work of Wasilewski et al. (1979) and Wasilewski and Mayhew (1982) provided no evidence of magnetites in the xenoliths from the upper mantle. A apart from the evidence from mantle xenoliths, the assumption of the Moho as a magnetic boundary greatly simplifies the modelling because the depth to the Moho is readily available from seismic crustal model.

2. The present work assumes the upper and the lower crust not to contain remanence. The inability of magnetic anomaly maps derived using satellite and airborne data to distinguish the anomalies due to induced or remanent magnetisation allowed us to assume the induced magnetisation as the only source for the anomaly.

It is further added that the present geothermal models (Artemieva and Mooney, 2001) are not accurate enough with the present resolution being 10^0 X 10^0, in addition to large gaps in the data especially in Africa and South America. Hence, they have not been included in the present work.

Our basic assumption that Moho is a magnetic boundary is likely to be a gross simplification in many regions of the world (Chapman and Furlong, 1992). Our efforts through the present work are to detect such anomalous regions, which would be helpful in understanding the composition of lower crust in those regions and possibly modify our *initial model*.

Aim:

The present work concentrates on the interpretation of anomalies observed over the continents. Our efforts here are not only to understand these anomalies but, in particular, to infer geological information from these maps.