Geomagnetic Depth Sounding

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1 INTRODUCTION

The contribution of induced currents to geomagnetic field variations has not been investigated thoroughly for the Indian region. I would like to impress that their effect is quite significant and at least for those investigations where the data from Annamalainagar, Kodaikanal and Trivandrum are used, a proper account of the induced part must be included to arrive at any meaningful conclusion. The data of these three observatories are used extensively for their unique location by being under equatorial electrojet. But their this position complicates the analysis since the electrojet is well known to introduce inhomogeneties of a relatively smaller scale in the inducing field. This non-uniformity compounded with the complex land-sea boundary near the tip of the Indian peninsula makes the analysis complex, which in the long run can be rewarding too.

As everywhere else, the study of an effect becomes an investigation of the cause itself. This has rather been the history of development of science. In geomagnetism too the analysis of transient variations have ultimately ended in delineating the features of their twin causes - the internal conductivity and the nature of source fields. These studies called "geomagnetic depth sounding" are now a discipline in themselves. It has been long known that the observed variations are partly of external and partly of internal origin. The internal part arising from currents induced inside naturally will depend heavily on the conductivity distribution there. If we can separate the internal part from the external part and can further separate its various frequency components, the conductivity profiles as a function of depth can then be drawn. The external part obtained after separation, is more meaningful for the study of the morphology and features of the inducing field.

The above remarks should not be construed to mean that variations must be partly of external and partly of internal origin for geomagnetic depth sounding. It works also when the whole field is of internal origin and under this class come the studies which calculate the conductivity of materials deep inside the earth (Banks 1972). This in itself is a big subject and will not be discussed here. I will limit myself only to the upper layers by covering the period range of 10 minutes to 6 hours. For this period range the depth of penetration varies from a few kilometres to 500 kilometres. A common feature of these variations is that they are partly of external and partly of internal origin.

The problem of calculating the induced internal currents for a given source field is prohibitive when the conductivity (σ) is a function of r, θ, ϕ - the spherical coordinates of a place. The algorithm is highly time consuming even for the fast computers. To simplify the problem, the internal current is divided into two parts: one called the normal part takes into account the average of the current induced over a large region and the perturbations introduced in this average by regional features are associated with the anomalous part. The division may seem rather arbitrary, but it has proved useful in defining and isolating lateral inhomogeneties in conductivities.

Before we proceed further, let me say a few words on the importance of conductivity profiles. No doubt it has a great geophysical importance for it tells us the nature of the crust and its structure, it has also a very big physical significance. The conduction in the upper mantle is primarily a semiconduction process. Hence, it is strongly dependent upon temperature through the relation.

$$\sigma = \sigma_0 e^{-T_0/T}$$

where T is the temperature and σ_0 and T_0 are constants depending upon pressure and composition of the crust. The above relation makes the conductivity variation a very good indicator of the temperature gradient of a region. The temperature variation can lead one to conclude the source of heat. The subject discussed here is thus not important only for geomagnetism but also for seismic and heat flow studies. As a matter of fact all three compliment each other to find the proper geology of a place, for by themselves rarely they give an unique answer.

2 OCEAN EFFECT AND ISLAND EFFECT

First let me take an example to show the effect of induced currents on observed variations. I will discuss the results from a floating ice station (Zhigalov 1960). Currents induced by a uniform field in a homogeneous region are such that they reduce the vertical component and enhance the horizontal component (Price 1965). The ideal way to see this effect would be to record the variations over the water of an ocean. It is important that the instrument be stationary, which is difficult on a ship. One may ask, why not do the same on a small island. The islands introduce perturbations in the current flow because these are upwellings which make a hole in the whole current system. A good alternative is to choose a floating ice island. A beautiful result following the approach has been reported by Zhigalov (1960) and his results are shown in Fig. 1. The records were obtained on the drifting ice station "North Pole 6". In the figure curve 1 shows the ocean depth, curve 2 the ratio γ_Z/γ_H for all periods and curve 3 shows the same ratio for variations having periods less than or equal to 10 mts. It is found that for depths greater than 2.0 kms., the variations follow exactly the depth profile. Secondly, when the floating ice has come to a region where the water is sufficiently deep the vertical component gets highly suppressed. This is called "Ocean effect". The ocean effect is not so obvious on a shallow sea. The Z-amplitude is reduced but it does not vanish.

Studies of Zhigalov interested others to look for similar effect on islands. As mentioned before the islands rising from the bed of the ocean interrupt currents flowing in the ocean. The currents usually divide and stream past the island flowing clockwise at one edge and anticlockwise at the opposite edge. Near the island edge the magnetic field is mainly controlled by the currents flowing adjacent to it, the Z-variations on opposite sides of the island will tend to be in opposite. (The effect is not so strongly seen in horizontal components for their normal part is much larger than the usual strength of the anomalous part). This effect is called "island effect". Mason (1963) has reported such effects through his investigations on the Hawaiian island of Oahu.

3 COASTAL EFFECT

After knowing that the land-sea boundary introduces perturbations in the induced currents; it would be of interest to see the effect of these perturbations at coastal stations. As mentioned in the last section, the Z-variations show the effect most conspicuously and their amplitude is strongly dependent on the distance of the station from the coast, or more precisely on the distance from the continental shelf. The coastal effect has been examined in great detail by many workers notable among them being Parkinson (1959, 1962, 1964), Rikitake (1964) and Schmucker (1964). In all these the variations showed a marked dependence on distance and at observatories separated even by a few tens of kilometres the Z-traces differed substantially. This is not surprising in view of the difference in conductivity of land and sea mass. The conductivity of sea water is $1\,\Omega^{-1}m^{-1}$ and that of the land mass is in the range of $0.001\,\Omega^{-1}m^{-1}$. Because of this huge contrast the current on two sides of the boundary may differ substantially for certain preferred direction of polarization of the inducing field.

As an example of the coastal effect the results of Rikitake and his collaborators will be discussed. The anomalies in Z-variations in Japan have been reported in a series of papers. Fig. 2 shows one of their results recorded during a bay disturbance that occurred on April 18, 1958 (Rikitake 1964). No significant variation is layered earth $Z_N = 0$, but the assumption that $H_a = D_a = 0$ is rather strong. Near complex conductivity contrasts the last assumption may not be true and values of H_T and D_T recorded at a nearby normal station should be used for H_N and D_N for the station sitting on the anomalous region. However, with the above assumptions, we get

$$Z_T = Z_H \cdot H_T + Z_D D_T$$

The above equation is commonly written as:

$$\Delta Z = A \Delta H + B \Delta D \qquad (1)$$

by replacing Z_T, H_T and D_T with ΔZ, ΔH and ΔD respectively. Also the transfer functions Z_H and Z_D have been replaced with A and B.

The linear relationship (Eqn. 1) within the components of the change vector of a geomagnetic disturbance implies that the vector should lie on a preferred plane and this plane is commonly called Parkinson's plane. The existence of this plane is taken as a good signature of a conductivity contrast. The plane is usually drawn in polar coordinates by plotting angles θ and ϕ of the change vector. Here θ denotes the angle the vector makes with the upward vertical and ϕ the angle between the horizontal projection of the vector and magnetic north. The two angles can be calculated through the relations:

$$\tan\theta = \frac{\left[(\Delta H)^2 + (K\Delta D)^2\right]^{1/2}}{\Delta Z}$$

and $$\tan\phi = \frac{K\Delta D}{\Delta H},$$

with $$k = H/3440$$

In the above equations H, ΔH and ΔZ are expressed in gammas and ΔD in minutes of arc.

As an example the polar diagram of Parkinson's plane, the diagram showing the change vectors at Darwin, Australia is shown in Fig. 3. The diagram is taken from Parkinson (1959). The orientation of the plane is given by the direction θ_S which is the direction of the maximum slope upward of the preferred plane and by the ratio S_R. S_R is the ratio of the vertical change to the component of the horizontal change in the direction of tilt (θ_S). In terms of A & B, the two are expressed as:

noticed in H traces from one end of the island to the other. In D too there is not much of a difference. But when we compare Z at various stations, we notice a remarkable enhancement in certain regions. As a word of caution, it may be mentioned that in this particular example, enhancement in Z is not totally due to the coastal effect - rather a major part is associated with the complicated subsurface structure of Japan. It is interesting to note that the regions showing geomagnetic anomalies are the areas with high seismic and volcanic activities. Similar correlation has also been found in the U.S.A. between the three effects. This makes geomagnetic depth sounding a potential tool for studying subsurface structures.

4 TRANSFER FUNCTIONS AND PARKINSON'S PLANES

The method of analysis, especially when data from only a few stations are available, centres on the concept of transfer functions as used by Schmucker (1964, 1970) and Everett and Hyndman (1967). Electromagnetic induction phenomena are governed by Maxwell's equations. These equations are linear and hence certain linear relationship between inducing and the induced field must be operative. Schmucker considers a hypothetical spatially uniform normal time varying field (H_N, D_N, Z_N) over a laterally uniform structure. The conductivity is considered to be a function of depth only. Effect of any lateral gradient is accounted through perturbations in this otherwise normal field. It should be emphasized that the normal part comprises the inducing field and the part of the induced field excluding the perturbations. The variations associated with these perturbations are called the anomalous part of the field (H_a, D_a, Z_a). The perturbations are related to the normal field by a nine-term transfer function:

$$\begin{pmatrix} H_a \\ D_a \\ Z_a \end{pmatrix} = \begin{pmatrix} h_H & h_D & h_Z \\ d_H & d_D & d_Z \\ z_H & z_D & z_Z \end{pmatrix} \begin{pmatrix} H_N \\ D_N \\ Z_N \end{pmatrix}$$

Each element of the matrix is complex and frequency dependent. It is common to seek a best fit of the data to the equation.

$$Z_a = Z_H H_N + Z_D D_N + Z_Z Z_n$$

Furthermore, it is assumed that $Z_T = Z_a$ and $H_T = H_N$ and $D_T = D_N$. The subscript T stands for the total field. The assumption $Z_T = Z_a$ is not so serious since over a parallel

$$\tan \theta_S = B/A \qquad \text{and} \qquad S_R = \sqrt{A^2 + B^2} \tag{2}$$

The above equation defines both the direction and magnitude of the vector. If we draw the vector by moving clockwise from magnetic south the direction of the induction vector, conforms to the convention followed by Parkinson. The vector in this case points towards the source of current concentration causing the anomaly. Figure 4 shows a map of Australian region with the Parkinson's vectors for magnetic variations. This again is from a work of Parkinson (1964). It is interesting to note that at coastal stations the vectors point towards the nearest deep part of the ocean. A long vector (Eqn.2) means that even a small horizontal variation in its direction can create a large variation in the vertical component. A correlation of the type given by Eqn. (1) occurs not only near coasts but even at inland stations situated near conductivity contrasts.

Assumptions leading to Parkinson's relation (Eqn. 1) are valid to a great extent near two dimensional structures - which usually exists near straight coasts. Cox et al. (1970) discuss in detail the perturbations in induced telluric currents near the shore-line of a large ocean and also their dependence on the direction of ionospheric electric currents. Interested readers are referred to this article for understanding the physics of the phenomena. Here it is suffice to say that two unique features of a two-dimensional structure are: (i) that θ_S must be real, i.e. A & B must have same phase and (ii) that θ_S must be frequency independent. Furthermore θ_S is normal to the conductivity contrast. All these conditions are approximately satisfied in the results of the analysis of California data (Cox et al. 1970).

Calculations with two-dimensional structure were performed for the California coast and the features that best fit the data are shown in Fig. 5. The results show that the conductivity structure of the continental crust and the upper mantle differs considerably from its oceanic counterpart. The conducting layers of the mantle are much shallower under the ocean. In California the anomaly in Z-variations is partly due to the contrast of the land-ocean conductivity and partly because of the mantle structure. Whether or not the widespread manifestations of the coastal anomaly are of this dual origin is yet to be investigated.

5 CONDUCTIVITY ANOMALY IN THE ANDES OF PERU AND BOLIVIA

The results of this study are unique for it has been the first thorough investigation of a low-latitude region. They are discussed in great detail in Schmucker (1969). Some aspects are given below.

A seismic expedition of Carnegie Institution, Washington, had discovered high attenuation of seismic waves travelling across the Andes mountain range. Schmucker in 1963 set up a net of nine Askania Variographs to list whether this zone of high seismic attenuation would appear as a zone of high conductivity and show up in geomagnetic variations.

At low latitudes the presence of equatorial electrojet introduces complexities in analysis of data. The presence of the electrojet is known to introduce non-uniformities in the inducing field which affect the analysis in two ways. First, the inducing field is no longer uniform over a large region and the scale length of the field may even become comparable to the conductivity contrast. Secondly, when the scale length of the source field is much smaller in comparison to the skin depth, the intensity of the induced current is reduced to zero. Absence of the induced current removes the signature of the anomaly. However, if only night time records are used, both these limitations disappear.

Variations during a typical night event are shown in Fig. 6 (The location of the recording stations are shown in Fig. 7). The figure shows the pattern of an equatorial bay field during night hours. The field has remarkable uniformity in the horizontal component. The H amplitudes hardly change over a latitude of 37°. There is reversal of Z-amplitudes between CAS-HUC, CAT-HU and ARE-COC. These suggest presence of induction currents flowing in the highly conducting channel beneath the Andes. Some irregular differences are noticeable in D. Nevertheless, the overall D amplitude is small and the horizontal disturbance vector points northwards to the centre of the bay vortex.

For comparison, day-time fluctuations are shown in Fig. 8. Maximum H amplitudes are noted near the dip equator (CAT, ABA). The D fluctuations at DEA, SIS and COC are anomalous. The Z amplitudes at COC and CAM are quite large. The last two effects are due to internal conductivity of Andes. The depth of the zone has to be small in comparison to the half-width of the jet field - say of the order of 50 km or less to have significant induced currents. The results postulate a conductivity of at least 0.1 $\Omega^{-1}\,m^{-1}$. Conductivities of this order are usually expected deep within the mantle. Their presence under Andes at 50 km shows a presence of unusual thermal state.

6 MEASUREMENTS WITH ARRAY OF MAGNETOMETERS

Most of the array studies carried out so far have used portable low-cost magnetic variometers designed by Gough and Reitzel (1967). The instrument has one small magnet for each of the components H, D and Z, suspended on taut torsion wires. The orientations of these magnets are recorded every 10 sec. on a 35 mm film. The instruments run on battery power and can operate unattended for three weeks at a stretch. They are usually buried in ground.

Many examples of data collected from a chain of these variometers are given in a review article by Frazer (1974). These have been extremely useful in locating anomalies and investigating spatial and temporal structure of the inducing fields. Plots of Fourier amplitudes and phases of the recorded variations for selected periods have often made the analysis easier and the anomalies identified through these studies have agreed well with geological structure found through other parallel methods. For quantitative work a formal separation of the field into components of internal and external origin is necessary. The difficulties involved are critically reviewed by Frazer (1974). The attempts so far have not borne encouraging results. The problem gets worst since the internal field has to be further separated into its normal part and anomalous part. With all these in view, it has often been suggested that the accuracy provided by formal separation is not worth the effort. However, the conclusion is based on the analysis of results of only one array. Frazer suggests that formal separation of the field of an array over a complicated structure involving curved or multiple anomalies would be of considerable interest. This can positively decide whether formal separation is in fact better or not than the method of approximate separation. Nityananda et al. (a contribution to this symposium) through the analysis of isolated events have found the tip of the Indian peninsula to exhibit a very complex conductivity structure. An array study over this region may prove useful in delineating a structure of geomagnetic data analysis if not the geophysical structure of the peninsula.

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