Application of magnetics for geothermal exploration

Geothermal energy is the natural heat of the Earth. The thinned and fractured crust allows the magma to rise to the surface as lava. The greatest part of the magma does not reach the surface but heats large regions of underground rocks. Rainwater can seep down the faults and fractured rocks for miles. After being heated, it can return to the surface as steam or hot water. When hot water and steam reach the surface, they can form fumaroles, hot springs, mud pots and other interesting phenomena. Otherwise, when the rising hot water and steam is trapped in permeable and porous rocks under a layer of impermeable rock, it can form a geothermal reservoir, which could be a powerful source of energy. This clean renewable energy can be used in many areas having accessible geothermal resources. For finding them geological, geochemical and geophysical surveys are necessary to be done. Magnetic can be an important part of the complex of geophysical geothermal exploration methods.

As is known, crustal rocks lose their magnetization at the Curie point temperature. At this temperature, ferromagnetic rocks become paramagnetic, and their ability to generate detectable magnetic anomalies disappears. The Curie temperature for titanomagnetite, the most common magnetic mineral in igneous rocks, is less than approximately 570^{O}C. It has been established that an increase in titanium content of titanomagnetite causes a reduction in Curie temperature and the titanium content generally increases in the more mafic igneous rocks (Byerly and Stolt, 1977). Consequently, it may be possible to locate a point on the isothermal surface by determining the depth to the bottom of a polarized rock mass. Thus the deepest level in the crust containing materials which create discernible signatures in a magnetic anomaly map is generally interpreted as the depth to the Curie isotherm.

One of the important parameters which determine the relative depth of the isotherm with respect to sea level is the heat content in a particular region. The heat content is generally proportional to the local temperature gradient, thermal capacity, and generation of heat. A region with significant geothermal energy near the surface of the earth is characterized by an anomalously high temperature gradient and heat flow. It is therefore to be expected that regardless of the composition of the rocks, the region will be associated with a conspicuously shallow Curie point isotherm relating to the adjoining regions isotherm (Bhattacharyya and Leu, 1977).

Magnetic anomalies are analyzed for estimating the depths to the bottoms of magnetized bodies in the crust. These depths, when contoured for the entire area, should provide a picture of the spatial variation of the Curie isotherm level. This picture should correlate to a significantly high degree with various known indices of geothermal activity in the area under consideration. The practical importance of a study on this correlation lies in the possibility of establishing a useful reconnaissance method, based on geomagnetic data, for rapid, regional geothermal exploration.

The method has been very popular during the last 30 years. Many projects have been made for reconfirmation of prospective geothermal areas, analysis of regional relationships among known geothermal areas, and discovery of new geothermal prospects using the creation of a Curie point depth map as an integral part of the exploration (Yellowstone National Park, Cascade Range of Oregon, Japanese Islands, Northern Red Sea, Trans-Mexican Volcanic Belt, parts of Greece, etc.).

The mathematical model on which the great mass of the analysis is based is a collection of random samples from a uniform distribution of rectangular prisms, each having a constant magnetization. The model was introduced by Spector and Grant (1970), and has proven very successful in estimating average depths to the tops of magnetized bodies. One principal result of their analysis is that the expectation value of the spectrum for the model is the same as that of a single body with the average parameters for the collection. In polar coordinates (s, ψ) in frequency space, this spectrum has the form

$$\begin{aligned} F(s,\psi) = 2\pi JA&\left[N + i(L\cos\psi + M\sin\psi)\right] \\ &\times\left[n + i(l\cos\psi + m\sin\psi)\right] \\ &\times\frac{\sin(\pi sa\cos\psi)}{\pi sa\cos\psi}\cdot\frac{\sin(\pi sb\sin\psi)}{\pi sb\sin\psi} \\ &\times\exp\left(-2\pi is(x_0\cos\psi + y_0\sin\psi)\right) \\ &\times\left[\exp(-2\pi sz_t) - \exp(-2\pi sz_b)\right], \end{aligned} \tag{1}$$

where

J – magnetization per unit volume;
A – average cross-sectional area of the bodies;
L, M, N – direction cosines of geomagnetic field;
l, m, n – direction cosines of the average magnetization vector;
a, b – average body x- and y- dimensions;
x_o , y_o – average body x- and y- centre location;
z_t , z_b – average depths to the top and bottom of the bodies;

Following Bhattacharyya and Leu (1975, 1977), Okubo et al. (1985) approach the estimation of bottom depths in two steps: first, find the centroid depth z_o; and second, determine the depth to the top z_t . The depth to the bottom (inferred Curie point depth) is calculated from these values: $z_b= 2z_o- z_t$. This method recognizes that there is no wavelength range in which the exponential signal from the bottom dominates the one from the top and that a direct calculation of depth to the bottom requires a simultaneous computation of the depth to the top – a much more complex problem. The terms involving z_t and z_b can be recast into a hyperbolic sine function of z_t and z_b plus a centroid term. At very long wavelengths (compared to the body dimensions), the hyperbolic sine tends to unity, leaving a single term containing z_0, the centroid. The terms involving the body parameters (a, b, z_b - z_t) may be approximated by their leading terms, to yield:

$$
\begin{aligned}
F(s,\psi) = 4\pi^2 VJs\big[N + i\big(L\cos\psi + M\sin\psi\big)\big] \\
\times\big[n + i\big(l\cos\psi + m\sin\psi\big)\big] \\
\times\exp\big(-2\pi is\big(x_0\cos\psi + y_0\sin\psi\big)\big) \\
\times\exp\big(-2\pi sz_0\big)\ ,
\end{aligned}
\tag{2}
$$

where V is the average body volume

Equation (2) can be recognized as the spectrum of a dipole. In effect, the ensemble average at these very low frequencies is that of a random distribution of point dipoles. A few methods were found for estimating z_0 from equation (2) (Bhattacharya and Leu, 1975, Spector and Grant, 1970). The most important finding is that the following decision will be independent of the details of the body parameters (prism, cylinders, or whatever), provided that the dimensions in all directions are comparable. Since geothermal areas are often in volcanic regions and are likely to have strong remanent magnetization, the second advantage of the algorithm is that there was not invoked reduction to the pole, or even assumption for induced magnetization was made. Numerical experiments have shown that there were not significant differences between results obtained with and without reduction.

At somewhat shorter wavelengths, the signal from the top dominates the spectrum and an estimate of the depth to the top can be obtained. Returning to equation (1) and assuming that a range of wavelengths can be found for which the following approximations hold:

$$
\frac{\sin(\pi sa\cos\psi)}{\pi sa\cos\psi} \cong 1,
$$

$$
\frac{\sin(\pi sb\sin\psi)}{\pi sb\sin\psi} \cong 1,
$$

and

$$
\exp(-2\pi sz_b) \cong 0
$$

the spectrum reduces to the form

$$
\begin{aligned}
F(s,\psi) = 2\pi JA\big[N + i(L\cos\psi + M\sin\psi)\big] \\
\times\big[n + i(l\cos\psi + m\sin\psi)\big] \\
\times \exp\big(-2\pi is(x_0\cos\psi + y_0\sin\psi)\big) \\
\times \exp\big(-2\pi sz_t\big).
\end{aligned}
\tag{3}
$$

To make sense for the above approximations, the bodies must in general be large in depth compared to their horizontal dimensions. Equation (3) is in fact the spectrum of a monopole. Because of the similarities to equation (2) the same methods to estimating z_t can be used. With the averaged location of the centroid and the mean depth to the tops of magnetized bodies known, it is simple and straightforward to calculate the mean depth to the bottoms of these bodies. The calculated depth is interpreted as the depth to the Curie isotherm for the block. Then taking into account the average elevation of the ground surface and assuming the value of the Curie point temperature the temperature gradient in depth can be estimated.

Similar studies have shown a strong correlation between geothermal hot regions and the thickness of the magnetized crust.

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