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Remote determination of magnetic properties and improved drill targeting of magnetic anomaly sources by Differential Vector Magnetometry (DVM)

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ABSTRACT

The induced magnetisation of a magnetic source is proportional to the ambient magnetic field and varies in response to natural geomagnetic variations, such as diurnal changes, storm fields and pulsations. In contrast, the remanent magnetisation is independent of changes in the ambient field. The local perturbation of the geomagnetic variations arising from a subsurface magnetic body can be determined by simultaneous monitoring of geomagnetic variations at two sites: one within the static magnetic anomaly associated with the body and another at a remote base station. Total field measurements can only provide a qualitative indication of the relative contributions of remanent and induced magnetisation to the anomaly. Monitoring of all three field components at the on-anomaly and base stations, however, allows the components of the second order gradient tensor of the anomalous pseudo-gravitational potential to be determined. This tensor depends only on the source geometry and the measurement location and is independent of the nature (remanent or induced), magnitude or direction of the source magnetisation.

Without making any assumptions about source geometry or location, the Koenigsberger ratio (Q), the direction of remanence and the direction of total magnetisation can be obtained from the components of this tensor. This information can constrain magnetic modelling prior to drilling and remove a major source of ambiguity in magnetic interpretation. The direction to the centre of a compact source can be determined directly from diagonalisation of the tensor. Values of Q constrain the magnetic mineralogy of the source and the remanence direction can discriminate sources of different ages or geological histories. Thus the method can also alleviate the geological ambiguity that afflicts magnetic interpretation.

Field trials of differential vector magnetometry (DVM) at several sites, including the Tallawang magnetite deposit, New South Wales, have demonstrated the validity of the proposed *in situ* method. However, a number of technical difficulties must be resolved before this method can be used routinely. Accurate characterisation of departures from mutual orthogonality of the components measured by each vector sensor, and the relative orientation of the anomaly and base station sensors, are crucial to the successful implementation of the method.

Keywords: vector magnetometers, geomagnetic variations, magnetic variometers, magnetic properties, remanent magnetisation, Koenigsberger ratio, magnetic modelling, drill targeting, Tallawang magnetite deposit

INTRODUCTION

Magnetic surveying is one of the most cost-effective exploration tools, with applications to geological mapping at a variety of scales (Jaques et al., 1997; Gunn et al. ,1997), as a delineator of mineralised environments and in many cases as a direct detector of orebodies (Gunn and Dentith, 1997). However, interpretation of magnetic survey data is afflicted by ambiguity, which reduces the effectiveness of the magnetic method and leads to increased exploration costs (in cases where multiple drill holes are required to test a magnetic source, for example). Information on magnetic properties of causative bodies constrains interpretation of their magnetic anomalies and can greatly reduce both geometric and geologic ambiguity.

This paper describes a method, Differential Vector Magnetometry or DVM, for obtaining information on magnetic properties of subsurface sources, using local perturbation of natural geomagnetic variations. The geomagnetic field varies due to diurnal variation, pulsations and magnetic storm activity (Parkinson, 1983). Magnetic storms produce geomagnetic variations of 100 nT or more, diurnal variation typically produces changes of ~50 nT and pulsations have typical amplitudes of 1-10 nT with periods in the range 1-300s. In conventional magnetic surveys these variations are monitored at a base station, so that a first-order removal of their effects on the survey data can be carried out. However, the induced magnetisation of magnetic bodies varies in phase with the ambient field, producing small local perturbations of the regional geomagnetic variations. These local perturbations can be measured by monitoring geomagnetic variations at two locations simultaneously, one location within the magnetic anomaly associated with a magnetic body and the other location away from the influence of the body.

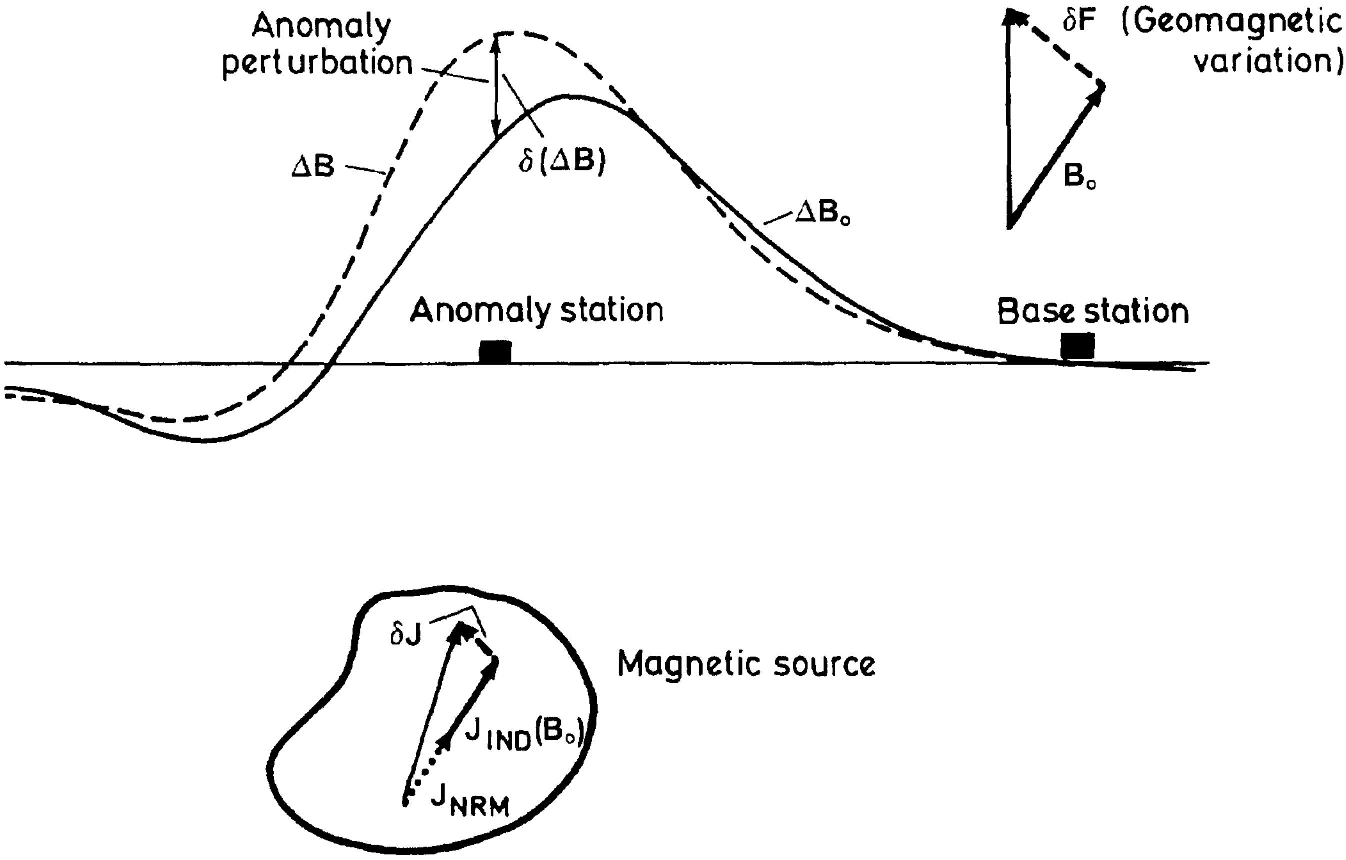

Figure 1. Effect of geomagnetic variation δF on induced magnetisation and total magnetisation of a magnetic body and its associated magnetic anomaly ΔB. DVM involves monitoring δF at a base station, whilst simultaneously recording changes in ΔB at a station within the magnetic anomaly. The solid line represents an anomaly profile for a single component of the time-averaged anomaly ΔB_0, while the dashed line indicates the corresponding anomaly while δF is acting.

Figure 1 illustrates the principle of the DVM method. Although the remanent magnetisation of the subsurface body in Figure 1 is shown as parallel to the time-averaged induced magnetisation, it may have any direction. A change $\delta\tilde{F}$ in the ambient magnetic field produces a corresponding change in the induced magnetisation and hence changes the magnitude and direction of the total magnetisation. The amplitude and form of the magnetic anomaly produced by the body changes slightly in response to this change in magnetisation. The base station monitors the regional geomagnetic variations, while the on-anomaly station records changes in the anomalous field vector **ΔB**.

A key assumption of the DVM method is that the "primary" geomagnetic variations at the two measurement locations are essentially identical, i.e. the observed differences reflect only the influence of the local magnetic source. External sources of geomagnetic variations, the ionosphere and magnetosphere, are very distant compared to the typical baselines used in DVM, for which separations between on-anomaly and base stations are generally in the range 100 m-1000 m. It is therefore reasonable to assume that the time-varying fields of external origin are identical over the DVM set-up. Secondary geomagnetic variations, which arise from currents induced in the Earth by external time-varying fields, contribute about half as much as the external variations to the total time-varying field and may adversely affect DVM in some circumstances. Provided the wavelength of the anomalous variation is much greater than the DVM base line, the geomagnetic variations of internal origin simply provide part of the driving signal on which the method is based. Simple amplitude offsets of the geomagnetic variations between stations are not serious, but phase differences at the two locations significantly affect the results. Wanliss and Antoine (1995) examined the effects of magnetic induction over a baseline of 17 km. Their results suggest that 1-2 nT differences may be fairly commonplace over such distances, which suggests that 0.1 nT differences may be common over a 1 km baseline. Where such gradients in induced geomagnetic variations affect the DVM surveys, it may be necessary to filter out high frequency pulsation-related variations and rely on the relatively large amplitude, low frequency effects, such as diurnal variation and magnetic storms, for which the secondary fields arise predominantly from deeper levels of the crust and are therefore essentially uniform over the DVM baseline.

The local amplification of geomagnetic variations has been observed over magnetic anomalies using rubidium-vapour total field sensors initially (Ward and Ruddock, 1962; Goldstein and Ward, 1966), and later using fluxgate vector magnetometers (Parkinson and Barnes, 1985). The former studies only provided qualitative indications of relative contributions of remanent and induced magnetisations to anomalies. The latter vector study permitted the estimation of the Koenigsberger ratio, Q, of the source, provided that the remanence direction could be assumed to be parallel to the induced magnetisation. With that assumption, those authors concluded that $Q \approx 0.4$ for the Savage River orebody, which is a reasonable value for a massive magnetite deposit.

This conclusion required a theorem, due to David Clark, that parallel anomalous field vectors, resulting from remanent and induced magnetisation, imply parallelism of these magnetisation components. This theorem was included as an appendix in the paper by Parkinson and

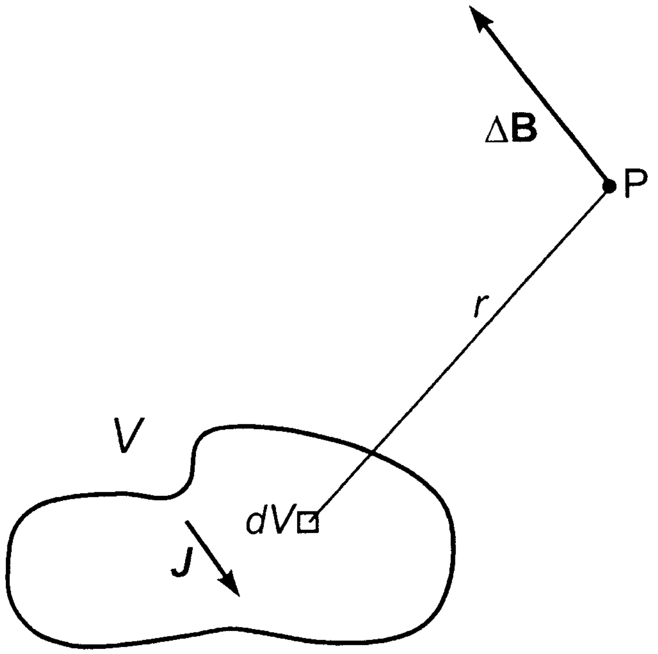

Figure 2. Geometry of source V, producing pseudogravitational potential U and anomalous magnetic field $\Delta\mathbf{B}$ at observation point P.

Barnes (1985). However, by analysing vector variometer data more fully it is shown here that not only Q, but also the direction of magnetic remanence, the direction of the total magnetisation and the direction to the centre of a compact source can be determined, for any remanence direction.

THEORY

Relationships between magnetisation and anomalous field

The DVM method is a new application of some standard results of magnetostatic pole theory. Brown (1962) gives a particularly lucid and comprehensive account of magnetostatic theory, which contains all the necessary background.

Consider a finite, homogeneous body of volume V and magnetisation $\mathbf{J}$. The magnetisation produces a magnetic anomaly $\Delta\mathbf{B}$ at point P outside the body. In appropriate units, the pseudogravitational potential U of the body V is equivalent to the magnetic scalar potential that would be produced by a distribution of unit magnetic pole density throughout V. U may be expressed formally as a volume integral:

$$U(\mathrm{P}) = \int_V \left(\frac{\mathrm{d}V}{r}\right), \tag{1}$$

where r is the distance from the volume element $\mathrm{d}V$ to the observation point P (Figure 2). The magnetic scalar potential Ω due to the uniform distribution of magnetisation, $\mathbf{J}$, throughout V is given by Poisson's relationship:

$$\Omega = -\nabla U.\mathbf{J}. \tag{2}$$

The anomalous magnetic field due to the body V is therefore:

$$\Delta\mathbf{B} = -\nabla\Omega = \nabla\nabla U.\mathbf{J}. \tag{3}$$

Then

$$\Delta\mathbf{B} = \mathbf{A}.\mathbf{J}, \tag{4}$$

where

$$\mathbf{A} = \nabla\nabla U. \tag{5}$$

$\mathbf{A}(\mathbf{r})$ is a second order tensor field, which depends only on source geometry and is independent of the nature (remanent or induced), magnitude or direction of the magnetisation. The components of $\mathbf{A}$ are:

$$a_{ij} = \frac{\partial^2 U}{\partial x_i \partial x_j} \tag{6}$$

These components have a simple physical interpretation (see Figure 3). For any specified direction of magnetisation, the anomalous magnetic field $\Delta\mathbf{B}$ at P has three components, each of which is proportional to the magnitude of the magnetisation. For example, the z-component of $\Delta\mathbf{B}$ that is produced by the x-component of magnetisation, $\Delta B_z(J_x)$, is proportional to J_x. Then, from equation (4), a_{zx} is given by

$$a_{zx} = \Delta\mathbf{B}_z(\mathbf{J}_x) / \mathbf{J}_x. \tag{7}$$

The other tensor components are defined analogously. Two important properties of $\mathbf{A}$ follow from (6):

$$a_{ij} = \frac{\partial^2 U}{\partial x_i \partial x_j} = \frac{\partial^2 U}{\partial x_j \partial x_i} = a_{ji}, \tag{8}$$

$$a_{xx} + a_{yy} + a_{zz} = \frac{\partial^2 U}{\partial x^2} + \frac{\partial^2 U}{\partial y^2} + \frac{\partial^2 U}{\partial z^2} = \nabla^2 U = 0. \tag{9}$$

Equation (9) follows from the fact that U is a potential field, which obeys Laplace's equation ($\nabla^2 U = 0$) outside V. Equations (8) and (9) respectively state that $\mathbf{A}$ is symmetric and traceless. It follows from (8) and (9) that only five of the nine components of $\mathbf{A}$ are independent.

The same analysis may be applied to a point within the magnetised body, with one significant difference. Inside the body U obeys Poisson's equation ($\nabla^2 U = -4\pi$), rather than Laplace's equation, implying that the trace of the tensor $\mathbf{N} = -\nabla\nabla U$ is 4π, rather than zero. $\mathbf{N}$ is in fact the demagnetising tensor for the internal point. Thus (-$\mathbf{A}$) is the external analogue of the point-function demagnetising tensor.

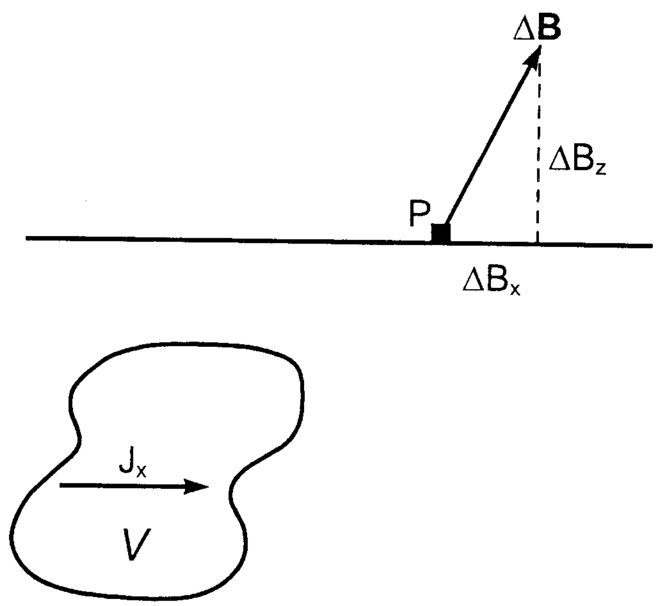

Figure 3. Physical interpretation of the tensor components a_{ij}. If the magnetisation is directed along the x-axis, a_{zx} is given by $a_{zx} = \Delta B_z(J_x)/J_x$.

Because the matrix $[a_{ij}]$ is symmetric, it can be diagonalised by suitable rotation of coordinate axes. The eigenvectors $\hat{\mathbf{u}}_i$ ($i=1,2,3$) define three mutually orthogonal axes. With respect to this new set of Cartesian axes $[a_{ij}]$ is diagonal:

$$\begin{bmatrix}\Delta B_1\\ \Delta B_2\\ \Delta B_3\end{bmatrix}=\begin{bmatrix}a_{11} & 0 & 0\\ 0 & a_{22} & 0\\ 0 & 0 & a_{33}\end{bmatrix}\begin{bmatrix}\mathbf{J}_1\\ \mathbf{J}_2\\ \mathbf{J}_3\end{bmatrix}. \tag{10}$$

Because the trace of a symmetric matrix is invariant under rotation of axes, the matrix in (10) is also traceless, i.e.

$$a_1+a_2+a_3=a_{11}+a_{22}+a_{33}=0. \tag{11}$$

In equation (11), the $a_i=a_{ii}$ ($i=1,2,3$) are the eigenvalues of the matrix $[a_{ij}]$, corresponding to the eigenvectors $\hat{u}_i$.

Thus for every point external to the magnetic body V, there exist three mutually orthogonal directions, for each of which the anomalous field is coaxial with, and proportional to, the magnetisation. With respect to this coordinate system, the relationship between magnetisation and the anomalous field is particularly simple:

$$\Delta\mathbf{B}_1=a_1\mathbf{J}_1;\Delta\mathbf{B}_2=a_2\mathbf{J}_2;\Delta\mathbf{B}_3=a_3\mathbf{J}_3 \tag{12}$$

As a specific example, consider a uniformly magnetised sphere (Figure 4). The pseudogravitational potential is equivalent to that of a point source at the centre, i.e. $U=V/r$, where r is the distance from the centre to the observation point. Applying (2), the corresponding magnetic scalar potential is:

$$\Omega=\frac{\mathbf{m}.\hat{\mathbf{r}}}{r^2}, \tag{13}$$

where $\mathbf{m}=\mathbf{J}V$ is the magnetic moment of the sphere.

By (3) and (13), the anomalous field at P is given by:

$$\Delta\mathbf{B}=\frac{-\mathbf{m}+3(\mathbf{m}.\hat{\mathbf{r}})\hat{\mathbf{r}}}{r^3}. \tag{14}$$

From (14), the components of $\mathbf{A}$ may be written explicitly as:

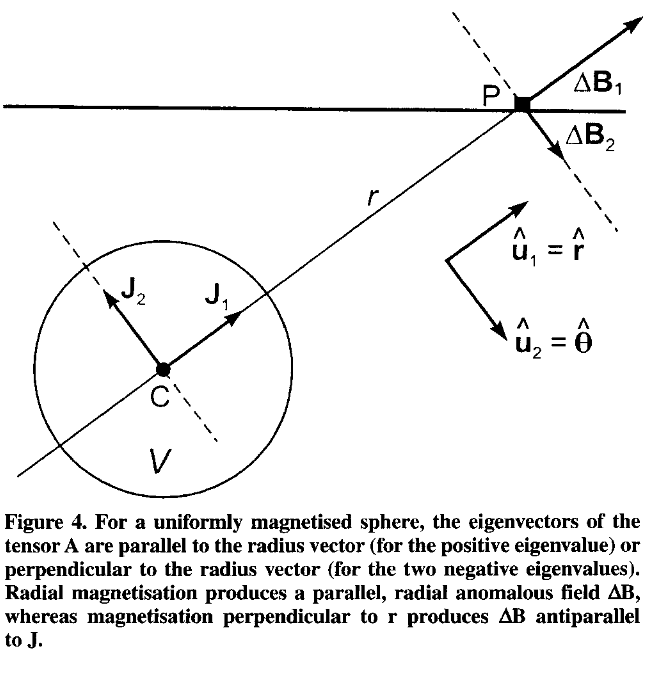

Figure 4. For a uniformly magnetised sphere, the eigenvectors of the tensor A are parallel to the radius vector (for the positive eigenvalue) or perpendicular to the radius vector (for the two negative eigenvalues). Radial magnetisation produces a parallel, radial anomalous field ΔB, whereas magnetisation perpendicular to r produces ΔB antiparallel to J.

$$\left[a_{ij}\right]=V\begin{bmatrix}\frac{2x^2-y^2-z^2}{r^5} & \frac{3xy}{r^5} & \frac{3xz}{r^5}\\ \frac{3xy}{r^5} & \frac{2y^2-x^2-z^2}{r^5} & \frac{3yz}{r^5}\\ \frac{3xz}{r^5} & \frac{3yz}{r^5} & \frac{2z^2-x^2-y^2}{r^5}\end{bmatrix}, \tag{15}$$

or more succinctly:

$$a_{ij}=\frac{3x_ix_j-r^2\delta_{ij}}{r^5}, \tag{16}$$

where δ_{ij} is the Kronecker delta, which has diagonal elements equal to 1 and off-diagonal elements equal to zero.

It is evident that $\mathbf{A}$ is symmetric and traceless, as required. Referring to Figure 4, it is obvious that if $\mathbf{J}$ is parallel to the radius vector $\mathbf{r}$, then $\Delta\mathbf{B}$ is parallel to $\mathbf{J}$, and if $\mathbf{J}$ is perpendicular to $\mathbf{r}$, then $\Delta\mathbf{B}$ is antiparallel to $\mathbf{J}$. Thus the eigenvectors of the matrix in (15) are parallel or perpendicular to $\mathbf{r}$. In terms of the spherical polar coordinates $(\mathbf{r},\theta,\phi)$:

$$\hat{\mathbf{u}}_1=\hat{\mathbf{r}},\hat{\mathbf{u}}_2=\hat{\theta},\hat{\mathbf{u}}_3=\hat{\phi} \tag{17}$$

The corresponding eigenvalues are:

$$a_1=\frac{2V}{r^3},\ a_2=-\frac{V}{r^3},\ a_3=-\frac{V}{r^3}. \tag{18}$$

Equation (17) implies that the direction to the centre of the sphere can be determined from the tensor $\mathbf{A}$ by diagonalisation.

Determination of the tensor A using DVM

The total magnetisation of subsurface magnetic sources is the vector sum of the time-varying induced magnetisation and the unvarying remanent magnetisation. Thus the magnetisation of a source may be written as:

$$\mathbf{J}(t)=\mathbf{J}_I(t)+\mathbf{J}_R=k\mathbf{F}(t)+\mathbf{J}_R, \tag{19}$$

where t is time, k is the effective susceptibility, $\mathbf{F}$ is the ambient geomagnetic field, and the subscripts I and R indicate induced and remanent magnetisations respectively. Equation (19) assumes that the susceptibility is isotropic, which is a reasonable approximation for most rocks.

In the vicinity of a magnetic body the geomagnetic field is perturbed by the anomalous field, $\Delta\mathbf{B}$, which is a function of the total magnetisation $\mathbf{J}$. Thus the magnetic anomaly itself is a function of time. At each point, $\Delta\mathbf{B}$ can be expressed as the sum of a constant field, which arises from remanent magnetisation plus the induced magnetisation in the time-averaged geomagnetic field, and a time-varying component, which corresponds to the magnetisation induced by geomagnetic variations, i.e.

$$\Delta\mathbf{B}(t)=\Delta\mathbf{B}_o+\delta(\Delta\mathbf{B})=\Delta\mathbf{B}(\mathbf{J}_o)+\Delta\mathbf{B}(k\delta\mathbf{F}) \tag{20}$$

$$\therefore \Delta\mathbf{B}(t)=\mathbf{A}.\mathbf{J}_o+\mathbf{A}.k\delta\mathbf{F}=\mathbf{A}.(\mathbf{J}_R+k\mathbf{F}_o)+\mathbf{A}.k\delta\mathbf{F}, \tag{21}$$

$$\delta(\Delta\mathbf{B})=k\mathbf{A}.\delta\mathbf{F}, \tag{22}$$

where δ is used to indicate temporal variations, while Δ indicates spatial differences. The zero subscript indicates time-averaged values or, to a good approximation, the initial values when measurements start. $\Delta\mathbf{B}$ is determined by subtraction of the magnetic field vector at the base station from the field vector at the on-anomaly station.

Remote determination of *in situ* magnetic properties of an anomaly source

In terms of the components $(\Delta X, \Delta Y, \Delta Z)$ of $\Delta\mathbf{B}$ and the components (X, Y, Z) of $\mathbf{F}$, equation (22) gives:

$$\begin{bmatrix} \delta(\Delta X) \\ \delta(\Delta Y) \\ \delta(\Delta Z) \end{bmatrix} = k \begin{bmatrix} a_{xx} & a_{xy} & a_{xz} \\ a_{yx} & a_{yy} & a_{yz} \\ a_{zx} & a_{zy} & a_{zz} \end{bmatrix} \begin{bmatrix} \delta X \\ \delta Y \\ \delta Z \end{bmatrix}. \quad (23)$$

Using the two time series (simultaneous measurements of $\Delta\mathbf{B}$ and $\mathbf{F}$), equation (23) can be solved in the least squares sense for the elements ka_{ij}. Data collection must continue until sufficient variation in all three components of $\mathbf{F}$ occurs (a requirement that is generally not difficult to achieve). The relations (8) and (9) (the symmetry and tracelessness properties respectively) constrain the solutions for the tensor elements. Because k is a scalar, the elements of the tensor $\mathbf{A}$ can be determined at P, within a multiplicative constant. The eigenvectors of $\mathbf{A}$ can therefore be uniquely defined, as can the ratios of the eigenvalues. In particular, for a quasi-spherical body the eigenvector corresponding to the positive eigenvalue of $k\mathbf{A}$ points away from the centre of the body.

The eigenvectors of $k[a_{ij}]$ define a set of axes for which $\Delta\mathbf{B}_o$ and $\mathbf{J}_o$ are coaxial. With respect to these axes:

$$(\Delta\mathbf{B}_o)_i = a_i(\mathbf{J}_o)_i = \frac{ka_i(\mathbf{J}_o)_i}{k}, \quad (24)$$

$$\therefore \frac{\mathbf{J}_o}{k} = \left[\frac{(\Delta\mathbf{B}_o)_1}{k\mathbf{a}_1}, \frac{(\Delta\mathbf{B}_o)_2}{k\mathbf{a}_2}, \frac{(\Delta\mathbf{B}_o)_3}{k\mathbf{a}_3} \right]. \quad (25)$$

Thus the direction of the time-averaged total magnetisation $\mathbf{J}_o$ can be determined from the time-averaged anomalous field $\Delta\mathbf{B}_o$ and the eigenvalues of the observed tensor $k\mathbf{A}$. The magnitude of $\mathbf{J}_o/k$ is also defined by (25), but the intensity of magnetisation is indeterminate if k is unknown.

From (19) we have:

$$\frac{\mathbf{J_R}}{k} = \frac{\mathbf{J}_o}{k} - \mathbf{F}_o. \quad (26)$$

Equations (25) and (26) determine the direction of $\mathbf{J_R}$ and the magnitude of $\mathbf{J_R}/k$. The Koenigsberger ratio Q can also be determined using (26) because:

$$Q = \frac{|\mathbf{J}_\mathrm{R}|}{k|\mathbf{F}|} = \frac{|\mathbf{J}_\mathrm{R}/k|}{|\mathbf{F}|}. \quad (27)$$

Therefore the following properties of the source can be determined from measurements of $\delta(\Delta\mathbf{B})$ and $\delta\mathbf{F}$, without making any assumptions about the source geometry:

- the direction of the total magnetisation,
- the direction of the remanent magnetisation,
- the Koenigsberger ratio Q.

For the special case of two-dimensional sources, the along-strike component (the y-component, say) of magnetisation produces no anomaly. Thus no information on J_y is obtainable from remote observations. The y-component of the anomalous field is zero, irrespective of the magnetisation direction, and therefore $a_{iy} = 0$ $(i = x,y,z)$. For 2D sources therefore, only the across-strike components of total and remanent magnetisations are derivable. For a source in the form of a horizontal cylinder, the eigenvector corresponding to the positive eigenvalue of the 2×2 matrix of non-zero tensor components points away from the cylinder axis.

The theory presented above can also be applied to gradiometric measurements. It is straightforward to show that simultaneous monitoring of the magnetic field gradient, $\nabla(\Delta\mathbf{B}(t))$, and $\mathbf{F}(t)$ at a single station within the anomaly allows $k\nabla\mathbf{A}$ to be determined. It suffices to monitor time variations for one row of the field gradient tensor, e.g. $\partial(\Delta\mathbf{B})/\partial z$, together with $\delta\mathbf{F}$ to determine $\partial\mathbf{A}/\partial z$, which is a symmetric traceless tensor. Diagonalising $\partial\mathbf{A}/\partial z$ enables $\mathbf{J}_o/k$ to be determined, in an analogous manner to that given above. Determination of the other properties follows directly.

Application of DVM to drill targeting

Figure 5 illustrates the application of measurements of the tensor $k\mathbf{A}$ to targeting a compact magnetic source. In this context, a compact source is one for which the dipole field is the dominant component of the multipole field arising from the source. In practice, this requires that the longest axis of the body is smaller than the distance from the centre to the observation point P. At P the eigenvector corresponding to the positive eigenvalue of $k\mathbf{A}$ points away from the "centre of magnetisation". Thus the drilling direction from P to intersect the source is determined. If the measurements are repeated at a different point P' that is also within the anomaly, the location of the centre of the source is uniquely defined.

INSTRUMENTATION

There are three main aspects to the instrumentation requirements of applying differential vector magnetometry to *in situ* determination of magnetic properties. An obvious factor is the necessary sensor sensitivity which is dictated by the anomaly magnitude, the geomagnetic variation magnitude and the Q of the source. The second aspect is the ability to determine components with high precision, i.e. vector measurements, while the third is the measurement of orientation. This last factor involves both intra-sensor orientation, i.e. orthogonality, and inter-sensor orientation, i.e. the accurate alignment of the component measurements. Table 1 summarises, for various circumstances, the magnitude of the signal that must be resolved. A regional geomagnetic field intensity of 50,000 nT is assumed.

The DVM system used in this study is based on Scintrex Caesium vapour sensors, which are total field instruments,

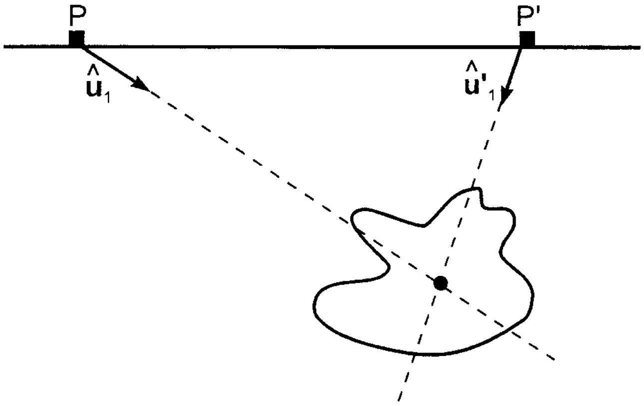

Figure 5. Use of DVM to locate "centre of magnetisation" of compact source, by determining directions to source $\hat{\mathbf{u}}_1$ at station P and $\hat{\mathbf{u}}_1'$ at station P'.

Table 1. Signal strength for differential vector magnetometry.

Anomaly Magnitude (nT)	Geomagnetic Variation (nT)	Local Perturbation (nT) $Q = 0$	$Q = 1$	$Q = 9$
10,000	100	20	10	1
10,000	10	2	1	0.1
10,000	1	0.2	0.1	0.01
1000	100	2	1	0.1
1000	10	0.2	0.1	0.01
100	100	0.2	0.1	0.01
100	10	0.02	0.01	0.001

and can measure only a single field component. However, in conjunction with two orthogonal alternating bias fields generated by mutually perpendicular Helmholtz coils, these total field sensors are capable of determining three components of the field. The principles of the vector magnetometers and details of early versions of the instrumentation were discussed by Schmidt et al. (1993). The Larmor frequency counters were developed by the Geophysical Research Institute, Armidale. The system allows measurement of field vectors several times per second. Computer simulations and bench-testing of the hardware show that errors in total field measurement produces errors in the vector components that are amplified about an order of magnitude, i.e. a 0.01 nT error in |**F**| produces about 0.1 nT error in a component.

The crucial limitation on measurement accuracy for magnetic field components is orientation. In a 50,000 nT ambient field, an orientation error of 1° can produce an error of up to 870 nT in components orthogonal to the field. Absolute measurement of components to < 1 nT requires determination of orientation within a few seconds of arc. It should be noted that for the application of differential vector magnetometry, alignment or relative orientation of the two sensors is the important issue, rather than absolute orientation. Non-orthogonality of magnetometer axes causes similar problems.

The extreme sensitivity of magnetic component measurements to orientation arises because the anomalous local fields are generally small compared to the regional geomagnetic field. Components of the field gradient tensor, on the other hand, are not nearly as sensitive to orientation. This is because the background IGRF gradient is relatively weak compared to the local gradients arising from nearby magnetic sources. Thus sufficiently sensitive magnetic gradiometers could overcome most of the difficulties associated with alignment of sensors. However, sensitivity becomes an issue for gradiometry, because gradient measurements are inherently noisier than component measurements.

Table 2 . Solution stability to simulated magnetometer noise.

Noise (nT)	Source azimuth	Source inclination	Strength (quality of solution)	NRM dec (°)	NRM inc (°)	Q
10^{-6}	225	90	3.88×10^{6}	359	-80	0.235
10^{-5}	228	90	3.89×10^{5}	359	-80	0.235
10^{-4}	265	90	3.89×10^{4}	359	-80	0.235
10^{-3}	355	90	3.89×10^{3}	359	-80	0.235
10^{-2}	2	90	3.90×10^{2}	359	-83	0.239
10^{-1}	2	88	40.3	181	-75	0.292
1	1	71	8.2	182	-5	1.136

We employed a calibration method that enables correction for non-orthogonality of bias coils, by recording measurements of one component, while switching the nominally orthogonal bias field through a sequence of forward, reverse and off. The procedure is repeated for all combinations of bias fields and calibration coefficients that characterise departures from orthogonality are derived and used for reducing subsequent three component data to a truly orthogonal coordinate system. Orientation of the two magnetometers with respect to each other is effected using a theodolite and accurately located grid pegs. Each magnetometer is fitted with three reflector-type targets, which are accurately aligned with the bias coils. The theodolite includes an electronic distance measurement device accurate to 1 mm. Angular measurements are to within 1″.

Forward modelling software was written to test that tensor elements $k\mathbf{a}_{ij}$ and source parameters could be correctly recovered from realistic simulations of geomagnetic variations. This software was also used to characterise the

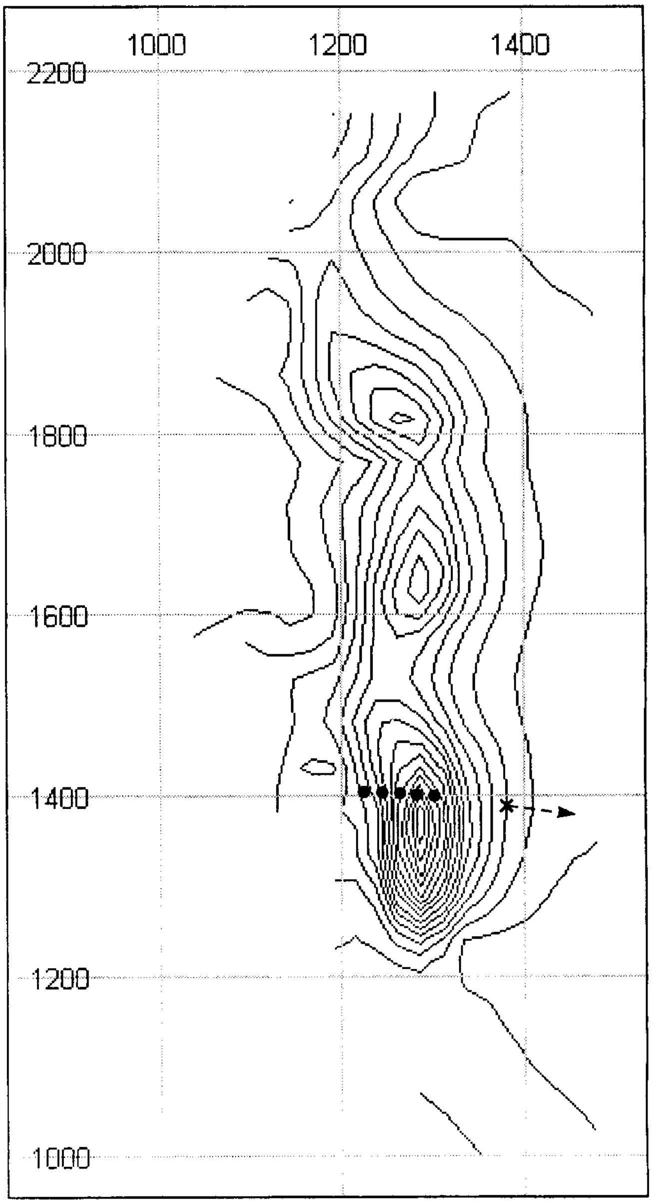

Figure 6. Ground magnetic anomaly over the Tallawang magnetite deposit, with DVM stations indicated by dots. Contour interval is 500 nT. The cross shows the theodolite location, with the direction to the base station indicated. The base station was located about 200 m to the east, upon the granite host.

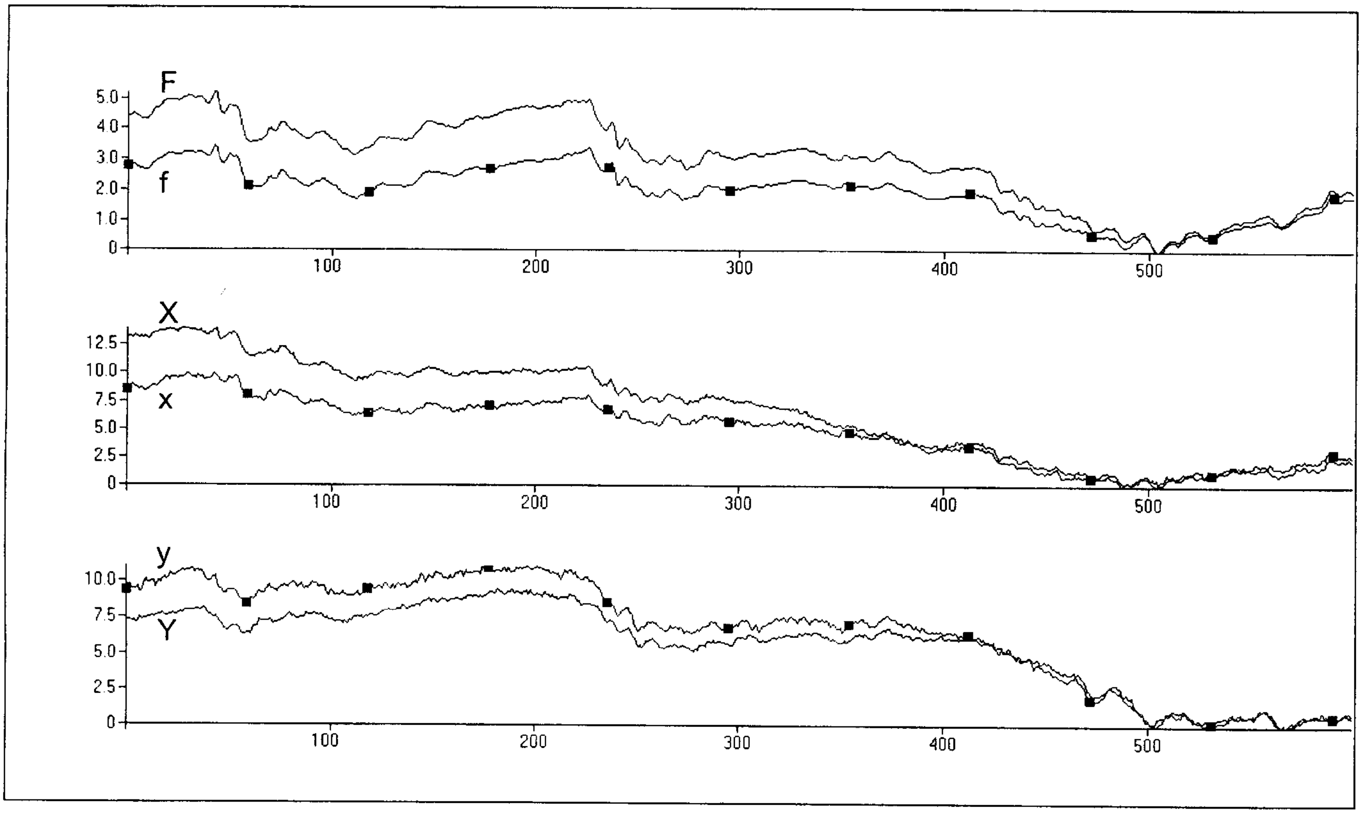

Figure 7. Typical records from Tallawang of the components of the field at an on-anomaly station and the base station. F and f refer to components along the time-averaged field direction, the *x*- and *y*-axes are perpendicular to f, with the *y*-axis horizontal. *X*, *Y* and *F* refer to the anomaly station while *x*, *y* and f refer to the (remote) base station.

sensitivity of the method to magnetometer noise and misorientation. The effects of random noise, magnetometer misorientation and systematic noise in the form of data offsets, amplification and attenuation, have been investigated through simulation.

In one test a spherical source was modelled with a susceptibility of 1.88 SI (0.15 G/Oe) and a remanence of 20 Am^{-1} (0.02 G) with a declination of 0°, an inclination of -80° at a depth of 214 m and a radius of 118 m. The Koenigsberger ratio corresponding to the remanent intensity and susceptibility is 0.234. This source is a crude model of a magnetite-rich ironstone source on one of the field testing areas (Normandy Exploration's White Hill prospect, South Australia: 29°40′S, 135°30′E). The spherical source was located directly below the anomaly station. The regional geomagnetic field had a declination of 6°, an inclination of -62° and an intensity of 57000 nT.

For this model the results for simulated short (time-series of 50 measurements) records are summarised in Table 2. For noise levels of 0.1 nT and 1 nT the solutions for source direction are fair but the determinations of the remanence and *Q* become increasingly poor. In fact the remanence appears to be most sensitive to noise and for noise levels of 1 nT the remanence is grossly in error.

Effects of misorientation were also simulated for this model. Determinations of *Q* are not seriously degraded until quite large rotations, greater than 30′(0.5°). However, for anomalies such as the one at White Hill, adequate orientation accuracy should not pose a problem. With the theodolite used for the field surveys orientations were estimated to be better than ±5′ of arc.

Simulations also show that absolute accuracy in determination of field components is not a critical aspect for the method. Thus constant offsets of the time series from one sensor have very little effect on the results, for offsets up to ~0.1% of the ambient field (50 nT error in a field of 50,000 nT). However, very small offsets of components over part of the time series from one magnetometer, or slow relative drift between the magnetometers, have a very deleterious effect on the calculated body parameters. Such offsets could arise from undetected changes in relative alignment of the sensors, for instance. It is therefore important to monitor relative alignment of the sensors regularly during collection of the time series, if absolute stability of the set-up cannot be guaranteed.

Generalisations on data quality requirements can be made by taking these effects into account and referring to Table 1. Clearly, exceptionally high quality data is needed to employ DVM over magnetic anomalies less than 1000 nT, particularly if the total geomagnetic variation is less than 100 nT and the Koenigsberger ratio is much greater than unity.

RESULTS FROM THE TALLAWANG MAGNETITE DEPOSIT

Situated 17 km north of Gulgong, NSW, the Tallawang magnetite deposit (32°12′S, 149°27′E) occurs as a skarn along the western margin of the Gulgong Granite. The associated ground magnetic anomaly is a strong magnetic high, elongated north-south, with a maximum amplitude of 7500 nT. The magnetite body is tabular, striking NNE and dipping steeply to the west. Oriented block samples from the Tallawang pit and drill core samples taken from selected drill holes at Tallawang South indicate remanence that is directed WNW and steeply up, with representative *Q* in the range 0.2-0.5.

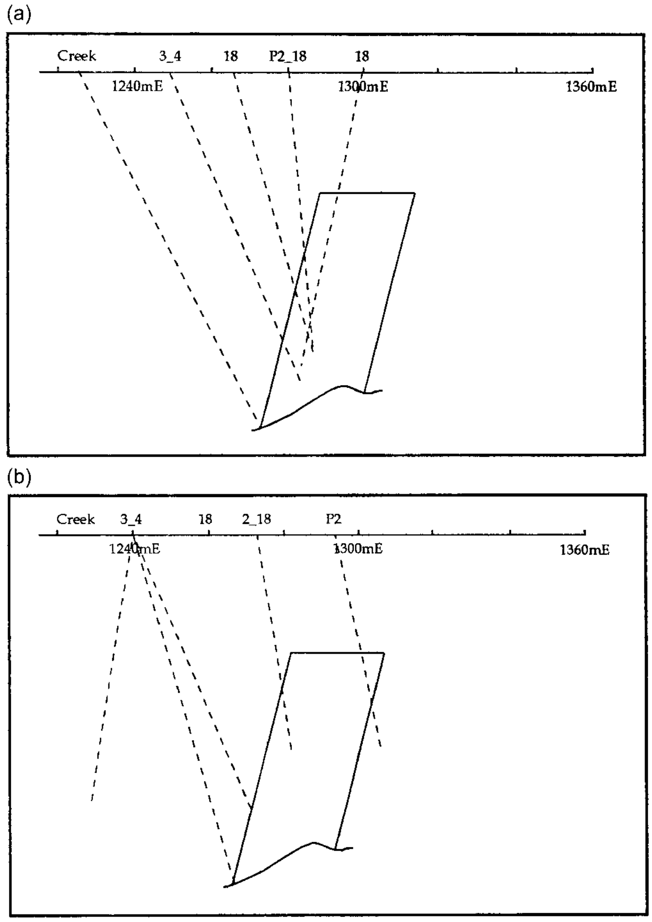

Figure 8. Magnetic model of the Tallawang magnetite deposit, with the DVM sites (a) the theoretical solutions for direction-to-source for a 2D semi-infinite dyke model, and (b) the corresponding calculated solutions from the DVM stations.

Five sites were occupied successively within the anomaly, while the base station was located about 200 m to the east, upon the granite host (Figure 6). Figure 7 shows some typical records of the components of the field at an on-anomaly station and the base station. A magnetic model of the body, a cross-section showing the sites, the theoretical solutions for an 2D semi-infinite dyke model and the calculated solutions are also shown (Figure 8). Therefore the solutions of the source directions from the westernmost sites would be expected to be somewhat shallower than those shown.

There is reasonable agreement between directions to source calculated from the data, particularly from the least noisy records, and the theoretical directions. The calculated across-strike components of remanence were somewhat scattered but overall were directed west with average inclination -72° ± 30°, which is consistent with the measured remanence direction. The calculated Koenigsberger ratios ranged from 0.35 to 0.81 and are reasonable estimates of the *in situ Q*.

CONCLUSIONS

Theoretical analysis, simulations and some encouraging field results indicate that, if components can be measured with a precision of 0.1 nT, and if the geomagnetic variation is of the order of magnitude of 100 nT, then it should be feasible to use the DVM method to study anomalies of about 1000 nT or greater.

It has been shown that while orientation is extremely important, in itself orientation should not pose an insurmountable problem. With the aid of a modern theodolite the necessary accuracy can be achieved. However, the stability of the orientation is critical. Incorporation of sensitive tiltmeters into sensor mounts might greatly ameliorate problems arising from instability of orientation.

Development of an instrument that combines a vector magnetometer and a gradiometer into a single sensor package could overcome most of the orientation problems. Such an instrument would eliminate the need for a separate base station and would minimise adverse effects from magnetotelluric currents. High-temperature SQUID technology may be the best option for further development of the method.

ACKNOWLEDGMENTS

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