

Elimination of edge effects in potential-field data processing by equivalent source technique

GM1.5

Jianghai Xia *, Richard D. Miller; Kansas Geological Survey; and Don W. Steeples, University of Kansas

SUMMARY

Edge effects are distortions at the edge of a domain which are artifacts of the implicit assumptions of a numerical algorithm or of the limited span of data used to solve the problem (Sheriff, 1991, p. 96). During wavenumber domain filtering of potential-field data, (e.g., reduction to the pole, pseudogravity, and directional derivative, etc.), edge effect distortions are usually parallel to the boundaries of the data set. Amplitudes of these artifacts could be several times the maximum amplitude of the real anomalies in the data. A conventional processing method results in distortion of up to twenty-five percent of the data points. We use an approach developed by Xia et al. (1993) for determining equivalent sources and then calculating specific anomalies based upon these equivalent sources. A synthetic example shows that edge effects are reduced to a negligible level when the reduced-to-the-pole operator is applied to magnetic anomalies. We also apply the reduced-to-the-pole operator to aeromagnetic data in Kansas, which gives a satisfactory result,

INTRODUCTION

In the two-dimensional (2-D) wavenumber domain, the conventional formula to reduce a magnetic total-field anomaly to the pole (Gunn, 1975) is

$$T_{\perp}(\vec{K}) = \frac{|\vec{K}|^2}{(\vec{K} \cdot \vec{f})(\vec{K} \cdot \vec{m})} T(\vec{K}), \quad (1)$$

where $T(\vec{K})$ is the magnetic total-field anomaly in the wavenumber domain with the inclination and the declination of the magnetization defined by the unit vector \vec{m} ; $T_{\perp}(\vec{K})$ is the magnetic total-field anomaly in the wavenumber domain with vertical magnetization (the reduced-to-the-pole anomaly); $\vec{K} (= k_x \vec{e}_x + k_y \vec{e}_y)$ is the wavevector, and \vec{e}_x and \vec{e}_y are unit vectors in x- and y-directions, respectively; \vec{f} is the unit vector of the Earth's field; $\vec{K} = i(k_x \vec{e}_x + k_y \vec{e}_y) + \sqrt{k_x^2 + k_y^2} \vec{e}_z$; and $i = \sqrt{-1}$ is the imaginary unit. Because of data truncation in the spatial domain, edge effects will always exist if equation (1) is used to obtain the reduced-to-the-pole anomaly. To reduce the edge effects, augmentation of original grids in both the x- and y-directions are usually applied during processing (e.g., Blakely,

1981). Augmentation of original grids alone cannot reduce edge effects to acceptable levels for certain anomalies, as can be noted in the following example.

A rectangular solid with a size of 800 km x 800 km x 3 km is buried at the center of the data field. The depth to the top and the bottom of the solid are 3 km and 6 km, respectively. Magnetization contrast of the solid is $1/\pi$ A/m (400 nT). The inclination and declination of the magnetization are 60 degrees and 45 degrees, respectively (assuming $\vec{f} = \vec{m}$). Anomaly values were calculated at $z = 0$ km on a 100 x 100 grid of points, spaced every 1.6 km (1 mi) (Xia and Sprowl, 1992). The synthetic magnetic total-field anomaly (Figure 1) is calculated with the formula given by Bhattacharyya (1964). Figure 2 shows the reduced-to-the-pole anomaly calculated using equation 1 by adding 10 rows to each side of the original grids during the processing. Severe edge effects can be seen along all sides of the map. For this particular example, the amplitude of the edge effects is sometimes four times the maximum amplitude of the real anomaly with about twenty-five percent of the data severely distorted by edge effects.

Silva (1986) introduced an equivalent source method for stabilizing the reduced-to-the-pole operation which gave excellent results and eliminated edge effects. The equivalent source was found by solving an inverse problem. Unfortunately this approach was extremely computation-intensive. Hansen and Pawlowski (1989) presented a regulated wavenumber-domain operator for reduction to the pole at low magnetic latitudes, which appeared to eliminate edge effects. This method required the assumption that the data were on (or reduced to) a horizontal plane. In this paper, we will apply a technique developed by Xia et al. (1993) for determining equivalent sources from data measured on a variable topographic surface and, based upon these equivalent sources, calculate the reduced-to-the-pole anomaly. We find that the edge effects are reduced to an unnoticeable level with minimal computing time necessary to calculate equivalent sources.

METHOD

The unsatisfactory results obtained using the conventional method (equation 1) to reduce the entire aeromagnetic data in Kansas to the pole was the real motivation to develop a

different approach to calculating the reduced-to-the-pole anomaly. The main advantages of processing data in the wavenumber domain are speed of calculation and capacity of handling a large set of data. This is why we were looking for approaches in the wavenumber domain. The approach discussed below keeps these two advantages.

Xia et al. (1993) developed the technique to determine equivalent sources from anomalies measured on a topographic surface. After equivalent sources are determined, the reduced-to-the-pole anomaly can be found by equation 6 in Xia et al. (1993).

$$T_{\perp}(\vec{K}) = 2\pi \frac{(\vec{K} \cdot \vec{f}_z)(\vec{K} \cdot \vec{m}_z)}{|\vec{K}|^2} J(\vec{K}) \times \exp[-|\vec{K}|Z_0] \\ = 2\pi J(\vec{K}) \times \exp(-|\vec{K}|Z_0) \quad (2)$$

where $\vec{f}_z = (0, 0, 1)$ and $\vec{m}_z = (0, 0, 1)$ are the unit vectors of the both Earth's field and the magnetization with the vertical inclination, respectively; Z_0 is the distance from the plane, on which equivalent sources are located, to a higher plane where the reduced-to-the-pole anomaly will be calculated; $J(\vec{K})$ is equivalent sources in the wavenumber domain which are determined by the technique developed by Xia et al. (1993).

The procedure to reduce magnetic anomalies to the pole will consist two steps: 1) determination of equivalent sources $J(\vec{K})$ by Xia et al. (1993); 2) calculation of the reduced-to-the-pole anomaly from $J(\vec{K})$ by equation (2).

RESULTS

By means of this approach, we recalculated the reduced-to-the-pole anomaly from the synthetic problem discussed in the introduction. During the calculation 10 rows are added by a polynomial extrapolation on each side of the original grids. The initial equivalent magnetization is 0 A/m and the inclination and declination are chosen as 60 and 45 degrees for the Earth's field and the magnetization, respectively. Equivalent sources are on the plane of $z = 1,000$ m. The initial root-mean-square error and the maximum deviation are 132 nT and 626 nT, respectively. These are reduced to 7 nT and 9 nT, respectively, after 29 iterations. Determination of equivalent sources took about 6 minutes on a Data General MV 20000 minicomputer, CPU speed of which is about the same as a PC with a 486 DX 33 MHz processor. Figure 3 shows the reduced-to-the-pole anomaly calculated from the equivalent sources. The edge effects are truly negligible. Figure 4 is calculated by the formula of Bhattacharyya (1964) with a vertical inclination. When we visually compare Figure 3 with Figure 4, only the 50-nT-

contour line appears slightly distorted by the edge effects. The maximum amplitude difference between Figures 3 and 4 is 15 nT, about two percent of the amplitude of the real anomaly.

In the following example we apply both equation 1 and the approach discussed in last section to aeromagnetic data in Kansas. There are about 72,000 line-km (8-11 measurements/km) of aeromagnetic data in the Kansas Geological Survey data base. The average distance between flight lines is 3.2 km. The data were measured at three different elevations: 760 m (2,500 A) above sea level over the eastern two quarter of Kansas, 910 m (3,000 ft) over the west-center quarter, and 1,370 m (4,500 ft) above sea level over the western quarter of the state. There is a transition zone about 5-15 km wide in western Kansas, over which the plane changed elevation from 910 m to 1,370 m. The elevations in this zone are linearly interpolated (Yarger, 1985; 1989). The kriging of SURFACE III (Sampson, 1988) is used to grid these data at 1.6 x 1.6 km. The final gridded data set is 205 x 408 points. The original aeromagnetic map was published by Yarger et al. (1981).

Because the aeromagnetic data were collected at different datum levels, the conventional method (equation 1) cannot be directly employed to the data set. We first reduced the data to a plane of 910 m above sea level (Xia et al., 1993) then used equation 1 to calculate the reduced-to-the-pole anomaly (Figure 5). The grids of data were augmented during data processing to 228 x 440. The edge effects are obviously at an unacceptable level. Up to twenty-five percent of the total data are destroyed by severe edge effects.

Now we apply the two step approach (1) determination of $J(\vec{K})$, 2) calculation $T_{\perp}(\vec{K})$ discussed in the last section to the same data. The initial equivalent magnetization is 0 A/m and the inclination and declination are chosen as 65 and 7 degrees for the Earth's field and the magnetization, respectively. The equivalent sources are on the plane of $z = -760$ m (760 m above sea level), which is just below the lowest measurement elevation of the survey. The initial root-mean-square error and the maximum deviation are 190 nT and 1,160 nT, respectively. They are reduced to 4 nT and 20 nT, respectively, after 12 iterations. The calculations took about 20 minutes on a Data General MV 20000 minicomputer. Then, equation 2 was used to calculate the reduced-to-the-pole anomaly, which is Figure 6. Only minimal edge effects are detectable in this figure. For details of this map refer to Xia et al. (in review).

CONCLUSIONS

Edge effects can be severe on the reduced-to-the-pole data if calculated by the conventional method. We have shown

the equivalent source technique (Xia et al., 1993) to be successful in eliminating the edge effects in processed potential-field data (e.g., the reduced-to-the-pole operation). The synthetic and real examples show computational efficiency. This approach can clearly be employed to execute other potential-field transformations, such as continuation, pseudogravity, etc. The approach employs only forward calculations so that the process is very stable.

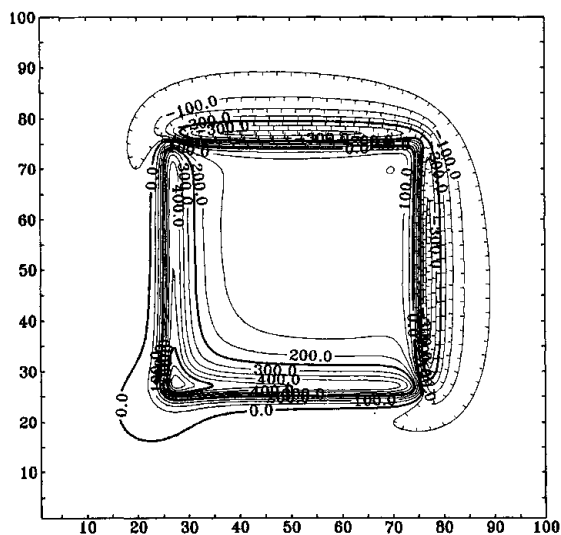


Figure 1. Synthetic magnetic anomaly of a rectangular solid with the inclination of 60 degrees and declination of 45 degrees calculated by Bhattacharyya (1964). The contour interval is 50 nT. The units in the x and y directions are station-spacing, 1.6 km (1 mi).

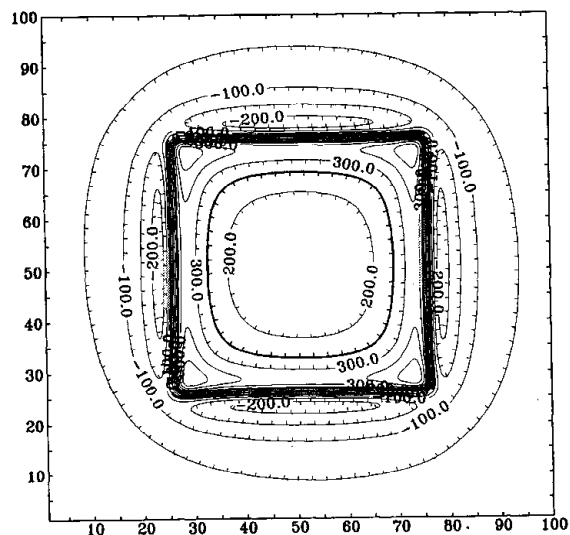


Figure 3. The reduced-to-the-pole anomaly. The approach discussed in the paper is applied to Figure 1. The contour interval is 50 nT. The units in the x and y directions are station-spacing, 1.6 km (1 mi).

ACKNOWLEDGMENTS

The authors appreciate the efforts of Mary Brohammer in manuscript preparation.

REFERENCES

- Bhattacharyya, B. K., 1964, Magnetic anomalies due to prism-shaped bodies with arbitrary polarization: *Geophysics*, 29, 517-531.
 Blakely, R. J., 1981, A program for rapidly computing the magnetic anomaly over digital topography: U. S. Geological Survey Open-File Report 81-0298.

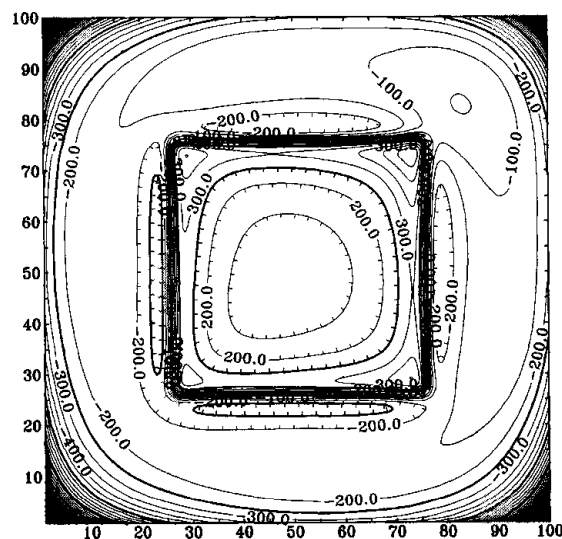


Figure 2. The reduced-to-the-pole anomaly. The reduced-to-the-pole operator (equation 1) is applied to Figure 1. The contour interval is 50 nT. The units in the x and y directions are station-spacing, 1.6 km (1 mi).

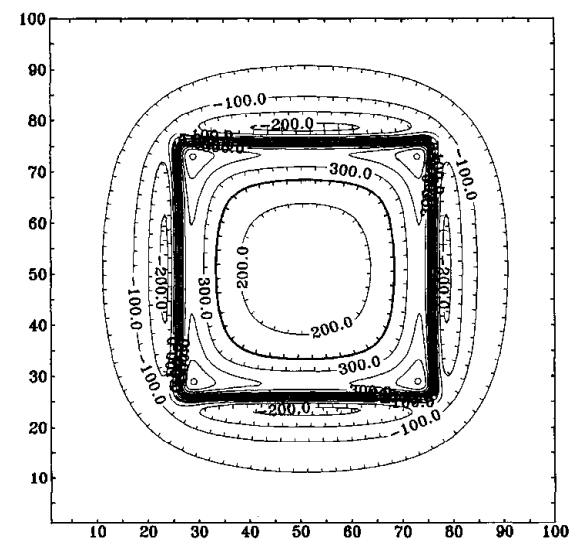


Figure 4. Synthetic reduced-to-the-pole anomaly of a rectangular solid (the inclination equal to 90 degrees) calculated by Bhattacharyya (1964). The contour interval is 50 nT. The units in the x and y directions are station-spacing, 1.6 km (1 mi).

- Gunn, P. J., 1975, Linear transformations of gravity and magnetic fields: *Geophys. Prosp.*, 23, 300-321.
- Hansen, R. O., and Pawlowski, R. S., 1989, Reduction to the pole at low latitudes by Wiener filtering: *Geophysics*, 54, 1607-1613.
- Silva, J. B. C., 1986, Reduction to the pole as an inverse problem and its application to low-latitude anomalies: *Geophysics*, 51, 369-382.
- Sampson, R., 1988, SURFACE III: Interactive Concepts Inc.
- Sheriff, R. E., 1991, *Encyclopedic dictionary of exploration geophysics* (3rd edition): Society of Exploration Geophysicists.
- Xia, J., and Sprowl, D. R., 1992, Inversion of potential-field data by iterative forward modeling in the wavenumber domain: *Geophysics*, 57, 126-130.
- Xia, J., Sprowl, D. R., and Adkins-Heljeson, D., 1993, Correction of topographic distortions in potential-field data: a fast and accurate approach: *Geophysics*, 58, 515-523.

Xia, J., Miller, R. D., and Steeples, D. W., (in review) Aeromagnetic map of Kansas, reduced to a horizontal plane and reduced to the pole (scale 1:1,000,000): submitted to Kansas Geological Survey for publication in Map M-series.

Yarger, H. L., Robertson, R. R., Martin, J. A., Ng, K., Sooby, R. L., and Wentland, R. L., 1981, Aeromagnetic map of Kansas: Kansas Geological Survey, Map M-16, scale 1:500,000, color interval 200 nT, contour interval 50 nT.

Yarger, H. L., 1985, Kansas basement study using spectrally filtered aeromagnetic data; in, *The Utility of Regional Gravity and Magnetic Anomaly Maps*, W. J. Hinze, ed.: Society of Exploration Geophysicists, Special Volume, 2 13-232; Kansas Geological Survey, Open-file Report 83-9, (1983).

Yarger, H. L., 1989, Major magnetic features in Kansas and their possible geologic significance: Kansas Geological Survey, Bulletin 226, 197-213.

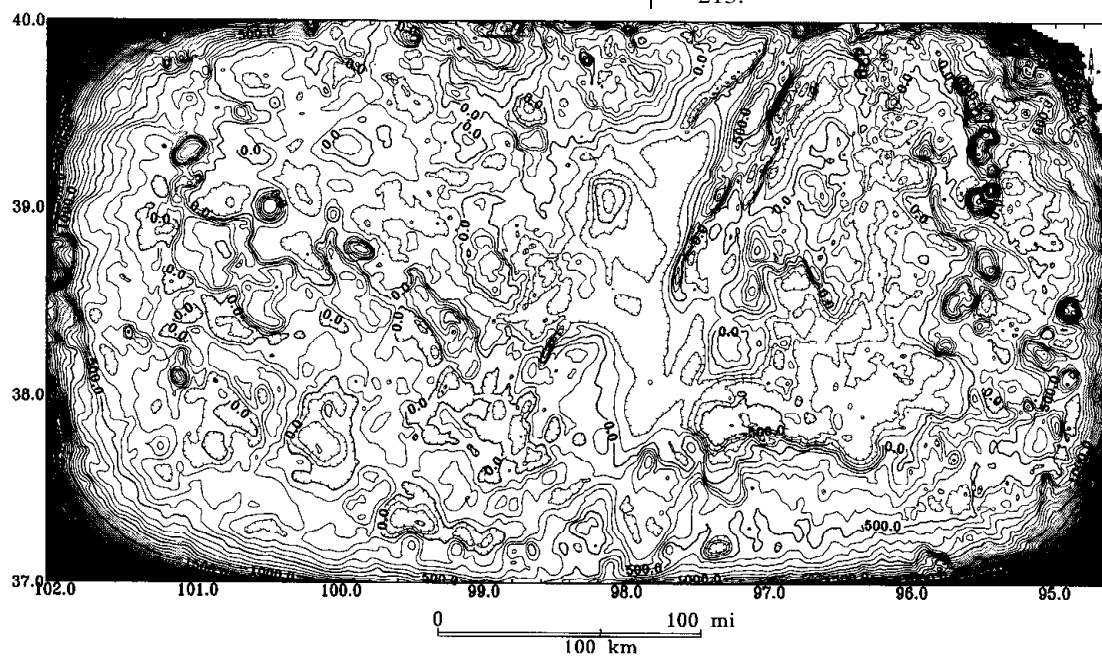


Figure 5. The aeromagnetic reduced-to-the-pole anomaly of Kansas. The reduced-to-the-pole operator (equation 1) is applied to aeromagnetic total-field anomalies (Yarger et al., 1981). The contour interval is 100 nT. Coordinates in the x and y directions are degrees of longitude and latitude, respectively.

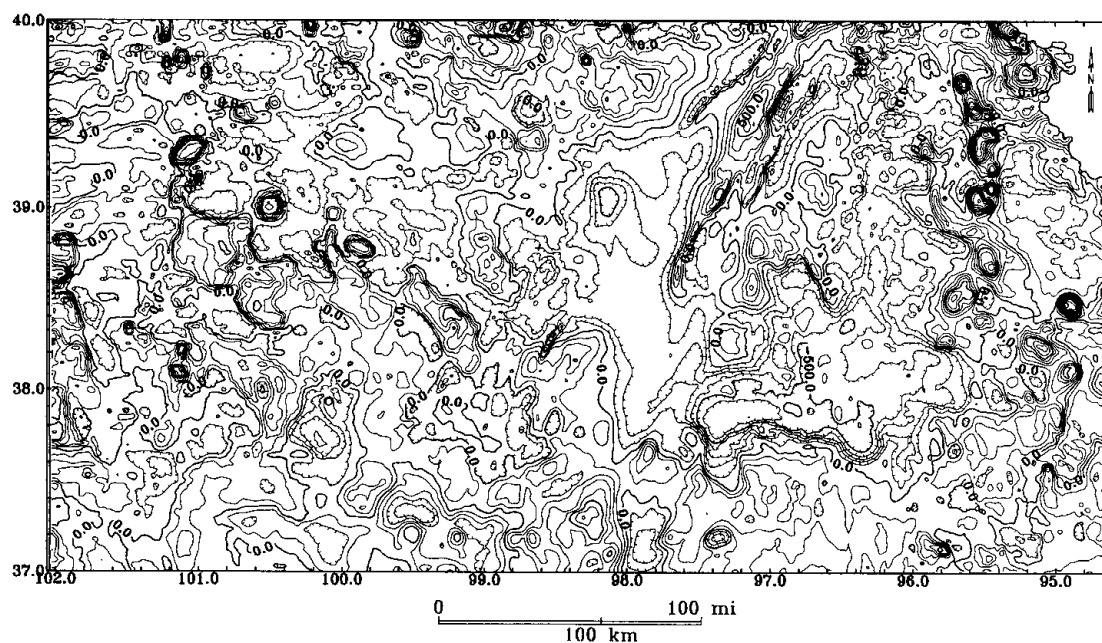


Figure 6. The aeromagnetic reduced-to-the-pole anomaly of Kansas. The approach discussed in this paper is applied to aeromagnetic total-field anomalies (Yarger et al., 1981). The contour interval is 100 nT. Coordinates in the x and y directions are degrees of longitude and latitude, respectively.

Euler deconvolution: Past, present, and future-A review

GMI.7

Alan B. Reid, University of Leeds, UK

Introduction

Euler's homogeneity relation has attracted sporadic interest from geophysicists over the years. It may be stated succinctly in the form

$$(x-x_0) \partial T / \partial x + (y-y_0) \partial T / \partial y + (z-z_0) \partial T / \partial z = N(B-T),$$

where (x_0, y_0, z_0) is the position of a source whose total field T is detected at (x, y, z) . The total field has a regional or background value B . N is the degree of homogeneity, interpreted geophysically as a structural index (SI - Thompson, 1982).

Early work

Hood (1963) pointed out that Euler's relation could be used to calculate depth to point pole or point dipole, given a measured vertical gradient. The possibility was also recognised in a patent application by Varian Associates (S. Breiner, pers. comm).

The relation was subsequently employed to estimate source type, given position and depth known or estimated by other methods (Slack et al, 1967; Barongo, 1984)

Thompson (1982) developed the technique quite fully, as applied to profile data, and suggested that SIs between 0.5 and 3 were useful on pole reduced magnetic data. The fault model (SI 0.5) required some empirically based corrections to obtain correct depth. He showed useful results in a variety of situations. His code found considerable use within Gulf and subsequently, Chevron (N. Gant, pers. comm). Apart from this use, the method appeared to languish until revived by Durrheim (1983). It found application to profile data from the Witwatersrand Basin (Wilsher, 1987, Corner & Wilsher 1989). Wilsher also showed, by application of Poisson's relation, that the vertical gradient of gravity (i.e. pseudomagnetic field) could be expected to behave like magnetic field and could therefore be subjected to profile Euler deconvolution.

Gridded data sets

Reid et al. (1990) followed up a suggestion in Thompson's paper and developed the equivalent method operating on gridded magnetic data. They also extended the basic equation to a modified form which handled the case of the fault model and introduced the concept of the zero SI. Finally, they suggested that the technique could be expected to work on gravity data by showing that Euler's equation was approximately obeyed by the gravity anomaly over a

finite step using an SI of 1.0. They applied the method to a variety of models and to a real data set from southern England, which showed that good correspondence could be obtained between structure and depth derived from Euler deconvolution of gridded magnetic data and that obtained by other (usually more expensive) means.

Enhancement and visualisation

Useful work was done on visualising the results on a graphics workstation by Allsop et al (1991) and was applied to surveys over Wales (McDonald et al, 1992).

The choice of structural index remains a vexed problem, because structures are poorly imaged and depths are biased if the wrong index is used for any given feature (Reid et al, 1990). In any real geological situation, features representing more than one structural index are likely to be present. Neil (1990) and Neil et al (1991) had some success in deriving structural indices as well as positions from the data themselves, using statistical methods which normalised the uncertainties on each position and depth solution. This remains an important long-term goal of anyone wishing to get optimum results from the method.

Marson & Klingele (1993) showed excellent examples of the use of gravity vertical gradient to deconvolve gridded data, but our own experience shows that many real regional data sets do not respond well to this approach. Their models also display the "sprays" of spurious solutions to which the method seems to be prone.

The clean-up of such "sprays" has been addressed with some success by Fairhead et al (1994) who applied a Laplacian filter to obtain those portions of the grid which showed significant curvature and restricted the deconvolution to those areas, so eliminating most spurious solutions.

The future

1. Work remains to develop a reliable method of estimating structural index.
2. Further means of identifying spurious or ill-focused solutions need to be sought.
3. The availability of high-powered graphics workstations at modest cost opens up possibilities in visualisation and on-screen development of interpretation.

We are presently working on these topics and hope to have results to report before long.