INVERSION OF EMAP DATA WITH STRONG STATIC DISTORTIONS

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Abstract

A stable solution for the inverse problem of the EMAP method is found by applying approximate equality constraints to an iterative inversion procedure. The parametric inversion is made with simple models. The random distribution of static distortions enters the inversion as a layer of small bodies with fixed sizes whose resistivities are free to assume any possible value. The resistivity of this layer tends to converge to the values of the features that generate the static shift, where they occur, and to the values of the upper layer, in those positions that are free of static shifts.

INTRODUCTION

The EMAP method is intended to eliminate or reduce the effects of static distortions that corrupt MT data (Torres-Verdín and Bostick, 1992). It accomplishes this by collecting the data in a continuous line of electric dipoles, which are equally spaced, and applying a spatial low pass filter dependent on the frequency. The method yields a bi-dimensional picture of the geo-electric structures under the line of dipoles. The blurred image in a section affected by static distortions is cleaned after the filter is applied, allowing a safer interpretation.

We present a method for inverting EMAP data which consists of first building an interpretative model from the filtered data, then applying approximate equality constrains (Medeiros & Silva, 1996) to achieve a stable solution. Those constrains allow us to introduce a priori information, in a least squares sense, that is geologically meaningful and not simply some mathematical constraint with little regard to the geology of the area in study. To take in account the static distortions in the data we build our interpretative models with a layer formed by small bodies whose resistivities are free to assume any value, so that we can invert isolated sources of static shift.

THE INTERPRETATIVE MODEL

When dealing with data contaminated with static distortions, it is often difficult to recognize the true features of the earth. In Figure 1 we have a model formed by one layer, termed here the *host layer*, overlying an infinite basement and a target conductive body inside the host layer. The resistivity of the target body is 10 Ωm, it is 1km wide and 100m thick, centered at the center of the dipole line, at x=0, and its top is at a depth of 300 m. Static distortions are created by small bodies with different resistivities scattered in the area bellow the dipole line. Our line is formed by 20 dipoles, each with 200 m, in 4km. When we plot the apparent resistivity section for this model without filtering, we get the result shown in Figure 2.

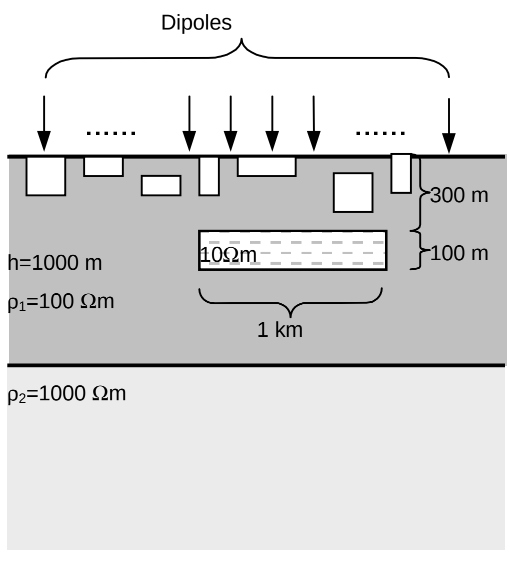

Figure 1 - Model used to generate the data to be inverted. Small bodies create static distortions on the data**.**

After the filtering process, we have a clearer picture of the structures, as shown in Figure 3. In addition, a good indication of the position of the static distortion sources below the dipole line is taken from that section. We start our inversion process by using just such picture to establish the interpretative model to be used.

One characteristic of our interpretative model is the presence of a layer formed by small outcropping bodies of fixed size, one for each dipole in the line, which are intended to simulate the effects of the static distortions in the inversion. We will term those outcropping bodies the "static shift" layer. (See the top layer of the model of Figure 04).

The parameters we are inverting, then, are those determined by the interpretative model plus the resistivities of the bodies in the static shift layer. In our example, we can determine, from the filtered data shown in Figure 3, the presence of the host layer and the target body.

For the example shown here, the parameters are: the thickness and resistivity for the host layer and the resistivity of the basement, the position, size and resistivity of the target body and one value for the resistivity of each body in the static shift layer, which is formed by 20 such bodies, as we have 20 dipoles in the line. We are inverting, then, 28 parameters for this model.

If the inversion is to be stable, it must be independent of the initial guess for the parameters. In our many tests with such models we have always fixed the first values for the thicknesses of the layers in 500 meters, regardless of what models we are inverting. From the apparent resistivity section, we can determine that the layer has smaller resistivity than the basement. We start the inversion with values which are representative of the layers in the section, the same is valid for the target body. The initial guesses for the resistivities in the static shift layer are the values of resistivities as given by each dipole in the highest frequency. In Figure 4, we have the values used for the initial model for our example. Note that the position of the target body in our initial model is such that it is confined in the host layer, as suggested by the filtered apparent resistivity section.

Using those values, we start a process to invert the data with a least squares procedure using approximate equality constraints.

THE CONSTRAINTS

We represent the observations ($\mathbf{y}$) of the method as a function of the parameters ($\mathbf{p}$) and the frequency ($\mathbf{w}$):

$$\mathbf{y} = f(\mathbf{p}, \mathbf{w}) .$$

We try to estimate the parameters from the observations with an iterative process applied to the linear approximation of that function, in which we estimate the variation in the parameters in each step and try to achieve a convenient convergence (Jupp and Vozoff, 1985). This process is usually very unstable so we often have to add a priori information to get useful results. If we have reliable information for any of the parameters from the geology or from some other geophysical method, we can introduce that information in the inversion by using constraint equations of the form

$$\begin{array}{c} \\ \\ \end{array}\begin{matrix} & & i & & j & & \end{matrix}$$

$$\begin{bmatrix} \vdots & & \vdots & & \vdots & & \vdots \\ 0 & \cdots & 1 & \cdots & 0 & \cdots & 0 \\ \vdots & & \vdots & & \vdots & & \vdots \\ 0 & \cdots & 0 & \cdots & 1 & \cdots & 0 \\ \vdots & & \vdots & & \vdots & & \vdots \end{bmatrix} \begin{bmatrix} \vdots \\ p_i \\ \vdots \\ p_j \\ \vdots \end{bmatrix} = \begin{bmatrix} \vdots \\ v_i \\ \vdots \\ v_j \\ \vdots \end{bmatrix},$$

$$\mathbf{Ap} = \mathbf{v} .$$

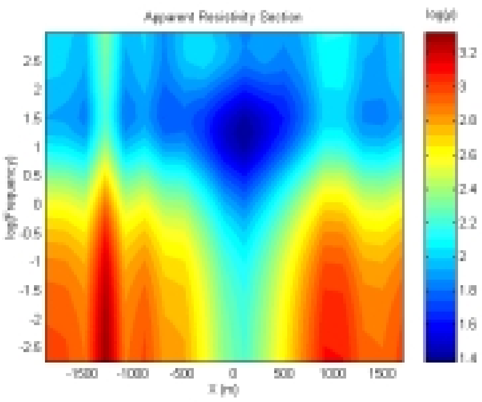

Figure 2 - Apparent resistivity section for the non-filtered data of the model shown in figure 1

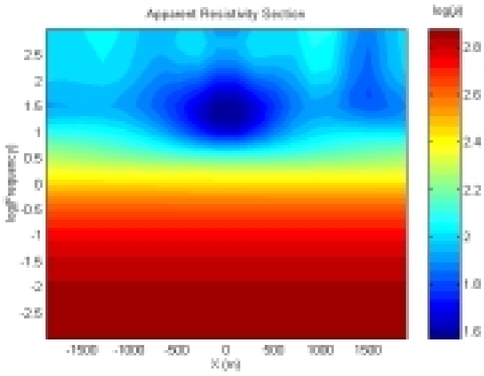

Figure 3 - Apparent resistivity section for the filtered data.

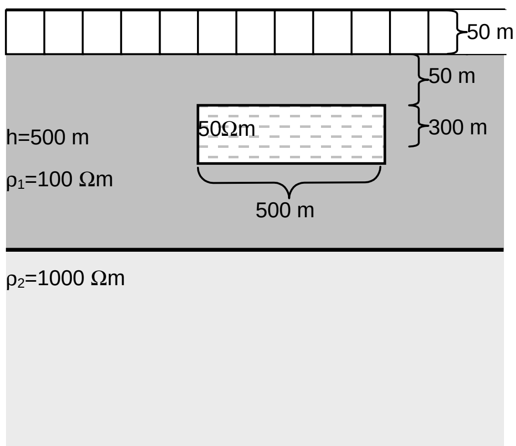

Figure 4 - Initial guess for the inversion. The resistivities of the small bodies in the static shift layer are the values of the apparent resistivity for each dipole in the highest frequency.

Where **A** is the absolute equality constraints matrix, its lines have zero elements everywhere except in the positions corresponding to the parameters that we want to constrain, where the value is 1. **v** is the vector with the values of the constraints.

We solve these equations together with that for the observations, in the same iterative process, so that we are really adding the constraints in a least squares sense, hence the term "approximate equality constraints".

The estimator for each iteration is

$$\delta\mathbf{p} = (\mathbf{J}^t\mathbf{J} + \mathbf{A}^t\mathbf{A} + \lambda\mathbf{I})^{-1}[\mathbf{J}^t\delta\mathbf{y} + \mathbf{A}^t(\mathbf{v} - \mathbf{A}\mathbf{p}_0)],$$

where **J** is the jacobian for the linearized function, λ is the so called Marquardt parameter, used to control the size of the step that is given in each iteration and also to ensure that the inverse exists, $\lambda\mathbf{y}$ is the difference between the observations and the values calculated in the previous iteration, and $\mathbf{p}_0$ is the parameter vector calculated in the previous iteration or the first guess in the first iteration. The convergence is evaluated by the euclidean norm of the vector **y** in relation to the observations $\mathbf{y}_0$.

RESULTS

When we perform the inversion for our example, with the initial guess shown in Figure 4 but with no constraint, we end up with unstable solutions, which are likely to assume very different values for small variations in the geological noise or in the values of the initial guess.

Now, suppose we have a good indication about the value of the resistivity of the body (10 Ωm). Introducing this value as the constraint for that parameter, we are able to stabilize the solutions. The results shown in table 1 are achieved after 8 iterations. It is clear that the geoelectric structure is very well resolved by the inversion. Different initial models yield values that are very close to the ones presented here, ascertaining the stability of the solution.

	True value	**Initial guess**	**Final value**
Thickness of the first layer	1 km	500 m	956.8470
Resistivity of the first layer	100 Ωm	100 Ωm	91.09444
Resistivity of the basement	1000 Ωm	1000 Ωm	907.6140
Resistivity of the target	10 Ωm	50 Ωm	10.00017
Depth of the target	300 m	100 Ωm	266,3729
Length of the target	1000 m	500 m	1031.646
Thickness of the target	100 m	400 m	105.6545
Central position of the target in the X direction	0	0	3.262544

CONCLUSIONS

The result presented here was accomplished for a very simple model, composed of only one layer, but this example is very illustrative of the process and of the kind of gain we can get with the constraints. Notice that we have constrained only one of the parameters of the model, but we achieved the desired stabilization of the solution. For more complex environments, the use of the static shift layer together with the approximate equality constraints have resulted in stable and geologically meaningful solutions.

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