2.5D inversion of CSEM data in a vertically anisotropic earth

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Abstract. The marine Controlled-Source Electromagnetic (CSEM) method is a low frequency (diffusive) electromagnetic subsurface imaging technique aimed at mapping the electric resistivity of the earth by measuring the response to a source dipole emitting an electromagnetic field in a marine environment. Although assuming isotropy for the inversion is the most straightforward approach, in many situations horizontal layering of the earth strata and grain alignment within earth materials creates electric anisotropy. Ignoring this during interpretation may create artifacts in the inversion results. Accounting for this effect therefore requires adequate forward modelling and inversion procedures. We present here an inversion algorithm for vertically anisotropic media based on finite element modelling, the use of Fréchet derivatives, and different types of regularisation. Comparisons between isotropic and anisotropic inversion results are given for the characterisation of an anisotropic earth from data measured in line with the source dipole for both synthetic and real data examples.

1. Introduction
The controlled-source electromagnetic method (CSEM) has proven to be an important addition to traditional seismic imaging techniques in hydrocarbon exploration. Other electromagnetic subsurface imaging methods, such as magnetotellurics and electrical tomography, have been widely used in geophysical exploration. In comparison, CSEM uses a horizontal electric dipole and has the advantage that it is sensitive to both galvanic and inductive effects at the same time. Although assuming isotropy can be sufficient in a number of situations, the presence of electric anisotropy in the earth’s crust, due to thin layer interbedding or grain alignments in the sediments can significantly alter the response measured at the seafloor [1]. Isotropic inversions of CSEM data characterising such structures often provide oscillatory artificial results [2] or results which are not consistent with other subsurface imaging techniques (well logging or seismic). As a consequence, accounting for anisotropic effects becomes a necessary challenge to explaining results that isotropic models cannot.

In this paper, we focus on the modelling and inversion of the 3D electromagnetic field measured over a 2D vertically anisotropic earth, also called vertically transverse isotropic earth, and present case study examples of inversions using synthetic as well as real survey data.

2. Modelling
Modelling anisotropy of the earth for CSEM problems has attracted more attention in recent years [3, 4], and methods for interpreting CSEM data using vertically anisotropic models were
also recently introduced [5, 2, 6, 7, 8].

We present here a numerical model based on nodal finite elements, which is an extension to include vertical anisotropy of the isotropic model developed by Unsworth et al. [9]. The electromagnetic field is decomposed into a primary field $E^p$ and secondary fields $E^s$ and $B^s$, the former being the semi-analytic solution of an isotropic half-space model of conductivity $\sigma_0$ and the latter a finite elements solution. The conductivity tensor is defined in an orthonormal coordinate system $(O, x, y, z)$, where $O$ is the origin of the coordinate system, $x$ the invariant (i.e. strike) horizontal direction and $z$ is the vertical direction pointing upwards:

$$\vec{\sigma} = \begin{pmatrix} \sigma_h & 0 & 0 \\ 0 & \sigma_h & 0 \\ 0 & 0 & \sigma_v \end{pmatrix} = \begin{pmatrix} \sigma_0 & 0 & 0 \\ 0 & \sigma_0 & 0 \\ 0 & 0 & \sigma_0 \end{pmatrix} + \begin{pmatrix} \Delta \sigma_h & 0 & 0 \\ 0 & \Delta \sigma_h & 0 \\ 0 & 0 & \Delta \sigma_v \end{pmatrix}$$.

The $(Ox)$ direction can be considered invariant if the earth properties can be approximated by infinite bodies in that direction. The secondary electromagnetic field is described by the Maxwell-Faraday and Maxwell-Ampère equations using the quasi-static approximation:

$$\nabla \times E^s = -\frac{\partial B^s}{\partial t},$$

$$\nabla \times B^s = \mu_0 \vec{\sigma} E^s + \mu_0 \Delta \vec{\sigma} E^p.$$  

We introduce the anisotropy ratio $\lambda = \sqrt{\frac{\sigma_v}{\sigma_h}}$, and apply a spatio-temporal Fourier transform to Maxwell’s equations in the diffusive regime to obtain two PDEs for $E^s$ and $B^s$:

$$\mu_0 \nabla \left[ \left( \begin{array}{ccc} \frac{2\gamma}{\gamma_h} & 0 & 0 \\ 0 & \frac{1}{\gamma_h} & 0 \\ 0 & 0 & \frac{1}{\gamma_h} \end{array} \right) \nabla \hat{E}_x^s \right] - \mu_0 \sigma_h \hat{E}_x^s \nabla \sigma_h \hat{E}_x^p + ik_x \mu_0 \nabla. \left( \begin{array}{c} \frac{\Delta \sigma_h}{\gamma_h} \\ 0 \\ \frac{\Delta \sigma_v}{\gamma_v} \end{array} \right) \hat{E}_y^p \right]$$

$$- ik_x \nabla \times \left[ \left( \begin{array}{ccc} \frac{1}{\gamma_z} & 0 & 0 \\ 0 & \frac{1}{\gamma_z} & 0 \\ 0 & 0 & \frac{1}{\gamma_z} \end{array} \right) \nabla \hat{B}_x^s \right]. u_x \right.$$  

and

$$i\omega \nabla. \left[ \left( \begin{array}{ccc} \frac{1}{\gamma_z} & 0 & 0 \\ 0 & \frac{1}{\gamma_z} & 0 \\ 0 & 0 & \frac{1}{\gamma_z} \end{array} \right) \nabla \hat{B}_x^s \right] - i\omega \hat{B}_x^s = -i\omega \nabla \times \left[ \left( \begin{array}{c} \frac{\Delta \sigma_h}{\gamma_h} \\ 0 \\ \frac{\Delta \sigma_v}{\gamma_v} \end{array} \right) \hat{E}_y^p \right]. u_x \right.$$  

$$- ik_x \nabla \times \left[ \left( \begin{array}{ccc} \frac{1}{\gamma_z} & 0 & 0 \\ 0 & \frac{1}{\gamma_z} & 0 \\ 0 & 0 & \frac{1}{\gamma_z} \end{array} \right) \nabla \hat{E}_x^s \right]. u_x \right.$$  

where $\mu_0$ is the magnetic permeability of vacuum, $\omega$ the angular frequency, $k_x$ the spatial wavenumber of the Fourier domain, $\gamma^2 = k^2_x - i\mu_0 \omega \sigma_h$, $\gamma^2 = k^2_x - i\mu_0 \omega \sigma_v$, and $i = \sqrt{-1}$. Equations (1) and (2) are two coupled Sturm-Liouville equations. They admit a unique solution and are simultaneously solved using a finite element discretisation of the two-dimensional domain $D \subset (O, y, z)$ and homogeneous Dirichlet boundary conditions at the edge of the discretisation domain. Then, the different components of the electromagnetic field can be derived from $\hat{E}_x^s$ and $\hat{B}_x^s$ using projections of Maxwell equations onto the coordinate axes [9]. The finite element solution can be computed on an unstructured discretisation mesh, which also enables inversions models including varying seawater depth [10] or complex sub-surface stratigraphy. Due to the high difficulty of estimating the accuracy of the result via functional analysis [11], the accuracy of the code is validated for a 1D model with a flat seafloor against a semi-analytic modelling code using integral equations.

A illustration of 2.5D forward modelling validation is given for the synthetic 1D model in Figure 1. The model consists of a 5000 m deep seawater column of resistivity 0.3125 $\Omega$m
Figure 1. Synthetic 1D vertically anisotropic model with two sources used to validate the finite element code.

and three layers of resistivity 2 Ωm, 50 Ωm and (3, 5) Ωm with thicknesses of 1500 m, 100 m and 1400 m respectively. Two sources are transmitting a signal with a frequency of 0.25 Hz. The response and its difference compared to that calculated using a semi-analytic 1D code are displayed on Figure 2. From the lower panel of Figure 2, we observe that there is less than 1% difference at all source-receiver offsets, which indicates that the 2.5D forward modelling code is extremely accurate.

In typical CSEM survey modelling, there are a lot more source positions than receiver positions, since the source dipole continuously emits along the survey tow line, therefore we use the reciprocity theorem of the electromagnetic fields to model sources where the actual receivers are and sample receiver positions along the dipole tow line in order to save computational time.

3. Inversion

The iterative inversion algorithm which is used is an adaptation of the Gauss-Newton method developed by Constable et al. [12], also known as the Occam inversion. It uses the Fréchet derivatives of a misfit functional characterising the discrepancy between the data and the
reconstructed model with respect to the variation of resistivity in a particular cell of the parametrisation domain to reconstruct the two-dimensional resistivity tensor $\bar{\rho} = \bar{\sigma}^{-1}$. The misfit functional $U$ is given by

$$U = \frac{1}{N} \|Wd - WF(m)\|^2 + \mu (\alpha \|\partial m\|^2 + \beta \|m - m^*\|^2) - X^*, \quad (3)$$

where $d$ is the data vector, $W$ its associated diagonal weighting matrix (for each datum $d_i \pm \Delta d_i$, $W_i = \frac{1}{\Delta d_i}$), $N$ the number of data, $m$ the vector of model parameters, $F(m)$ the response vector of the model, $m^*$ the a priori model, $\partial$ the smoothing operator, $\mu$ its associated Lagrange multiplier, $\alpha$ and $\beta$ the relative weights of the regularisation terms, and $X^*$ the misfit objective. Here $\|\cdot\|$ denotes the $L^2$ norm of the measurement space. Typically $d_i = \log_{10} E_{\|,i}$, where $E_{\|,i}$ is the $i$th measurement of the electric field parallel to the source dipole, so that short and long source-receiver offsets have equal weight in the misfit functional. For the inversions, we use the root mean square (RMS) value of the residuals as indicator of misfit: if $X_i = W_i(d_i - F_i(m))$ is the residual associated to the datum $d_i$, we have an $RMS = \sqrt{\langle X^2 \rangle}$, where $\langle \cdot \rangle$ denotes the mean. Ideally, the residuals $X_i$ are normally distributed with mean $\langle X \rangle \approx 0$ and standard deviation $\sqrt{\langle X^2 \rangle - \langle X \rangle^2} \approx \sqrt{\langle X^2 \rangle} = RMS$ at the end of the inversion process and therefore the final value of the RMS misfit functional should be close to 1, which should then be the theoretical chosen value for the misfit objective $X^*$. $m$ is a vector of horizontal and vertical log-resistivities $\log_{10} \rho_h$ and $\log_{10} \rho_v$ and the smoothing term is decomposed as follows:

$$\|\partial m\|^2 = \|\partial_h \rho_h\|^2 + \|\partial_v \rho_v\|^2 + \|\partial_h \rho_h\|^2 + \|\partial_v \rho_v\|^2,$$

where $\partial_h$ and $\partial_v$ respectively represent the horizontal and vertical smoothing matrices.

The misfit functional $U$ with respect to the parameter vector $m$ is locally linearised, so that the search for its minimum provides the update for the parameter vector $m$ at iteration $k + 1$ via the equation

$$m_{k+1} = [\mu \partial^T \partial + (WJ_k)^T(WJ_k)]^{-1} (WJ_k)^T(Wd_k), \quad (4)$$

with $\hat{d}_k = d - F(m_k) + J_k m_k$ and $J_k$ the Jacobian matrix of derivatives at iteration $k$ defined by the elements $J_{ij,k} = \frac{\partial E_{(m_k)}}{\partial m_{j,k}}$ [12]. The Fréchet derivatives of the component electric field parallel to the source dipole in the Fourier domain with respect to the parametric conductivities of a cell $l$ in vertically anisotropic media are given by [13]:

$$\frac{\partial \hat{E}_{\|}}{\partial \sigma_{h,l}} = \int_{D_l} (\hat{E}_z \hat{E}_x^* + \hat{E}_y \hat{E}_y^*) \, dydz, \quad (5)$$

and

$$\frac{\partial \hat{E}_{\|}}{\partial \sigma_{v,l}} = \int_{D_l} \hat{E}_z \hat{E}_z^* \, dydz, \quad (6)$$

where $\hat{E}^*$ is an adjoint field generated by reciprocal source dipoles, parallel to the real one located at all modelled receivers positions. The adjoint method is proven to be more efficient for computing the derivatives when the number of observation sites is smaller than the number of parameters [14]. The derivatives are then computed at the seafloor in the $(x, y, z)$ domain using an inverse Fourier transform (in practice, a sine or cosine transform). If the source dipole is across strike (i.e. along $(O_y)$) then the field measured at the seafloor is $E_{\|} = E_y$, whereas if it is along strike $(O_x)$ it is $E_{\|} = E_z$. The Fréchet derivative of the component of the electric field perpendicular to the dipole is similarly computed using an adjoint source dipole perpendicular to the real one.
Although the problem is ill-posed, it can be approximately solved using appropriate regularisation, so that unrealistic results from local minima of the misfit functional are avoided. Possible regularisations are the minimisation of roughness of the model, the minimum norm distance to an a priori model [15] or resistivity smoothing breaks at horizons derived from seismic imaging [16].

The electromagnetic field created by a dipole in a conducting medium is characterised by two modes: the Transverse Magnetic (TM) mode, whose current lines are vertical loops, and the Transverse Electric (TE) mode, whose current lines are horizontal loops [13]. Due to the prevalence of the TM mode component of the electric field in the inline configuration, CSEM data acquired in this geometry are primarily sensitive to the vertical resistivity of an anisotropic structure (Figure 3 right panel). The TM mode also provides the primary sensitivity to thin resistive structures in the earth, which are virtually invisible to the inductive TE mode. As a result there is little or no sensitivity to the horizontal resistivity of thin resistivity layers [2], as illustrated in the left panel of Figure 3. As a consequence, inline CSEM measurements will only allow recovery of the vertical resistivity of thin resistive layers, and will provide the best constraint on the vertical resistivity of the overburden. Wide azimuth acquisition including broadside data, where the TE mode is predominant, increases the sensitivity to the horizontal resistivity of thick overburden structures, however the horizontal resistivity of thin resistive layers is still poorly constrained.

4. Examples

4.1. Synthetic data

We present the inversion of synthetic horizontal electric field amplitude and phase data calculated from the model shown in Figure 4. This contains a 2 km wide and 50 m thick hydrocarbon layer of anisotropic resistivity \((50, 60) \, \Omega m\) buried at 900 m below a sloping seafloor. A \((1, 3) \, \Omega m\) anisotropic layer lies above the reservoir and the rest of the model is isotropic. Vertical 1D profiles are given at the abscissa \(x = -1000 \, m\) and \(x = 1000 \, m\) along the line in Figure 5. The signal is transmitted at 0.125 Hz, 0.375 Hz and 0.625 Hz (corresponding to the fundamental, third and fifth harmonics of a square wave transmission) every 500 m at 40 m above the seafloor, and is measured at ten receivers along a 10 km long line. The initial model for the inversion is an isotropic 2 \, \Omega m half-space for both the isotropic and anisotropic inversions.
Figure 4. Resistivity structure used in the synthetic inversion study.

Figure 5. Vertical 1D profiles with their squared anisotropy ratio $\lambda^2$.

The 2D isotropic resistivity section obtained from the isotropic inversion is presented in Figure 6. The residual associated with each data point is plotted as a function of range and distance along the line in Figure 7. The final RMS misfit is 1.081 after 39 iterations. The distribution of residuals is reasonably normally distributed around zero and has no spatial trend: their mean is -0.0159 and their standard deviation is 1.0809. More precisely, the amplitude residuals of the first two frequencies are well distributed around zero, whereas the ones at the third frequency have more negative bias at the shorter ranges. From Figure 6, we observe that the hydrocarbon layer is recovered at the correct lateral location. However its horizontal extent is too large (most likely the result of the isotropic inversion merging the effect of the target and the anisotropic overburden layer) and extra oscillations are included in the overburden layer and below the target. These artifacts are the result of the breakdown of the isotropic assumption.

Conversely, the anisotropic inversion result in Figure 8 does not include such oscillatory resistivity values, and its final RMS misfit is equal to 0.98 after just 9 iterations. The mean of the residuals is -0.0146 and their standard deviation 0.9799. The distribution of residuals is displayed in Figure 9. Vertical 1D profiles through the section are shown in Figure 10. As expected, the horizontal resistivity of the thin hydrocarbon layer is not recovered. However
Figure 6. Result of isotropic inversion of synthetic data, iteration 39.

Figure 7. Residuals obtained from the isotropic inversion of synthetic data, RMS=1.081.

Figure 8. Anisotropic inversion of synthetic data, iteration 9.

the effect of the resistive hydrocarbon layer is clear in the vertical resistivity, although it has been smeared over a range of depths by the smooth regularisation applied. As a result although the resistivity of the hydrocarbon layer as recovered by the inversion is lower than the input value of 60 Ωm, its thickness is greater resulting in the same integrated vertical resistance. The resistivity of the basement of the background model is recovered by both the isotropic and
anisotropic inversions, but its extent is slightly masked by the presence of the target, especially in the isotropic case. The map of the squared anisotropy ratio \( \lambda^2 \) is displayed in Figure 11. Since

**Figure 9.** Residuals obtained from the anisotropic inversion, RMS=0.980.

**Figure 10.** Vertical 1D profiles with their squared anisotropy ratio \( \lambda^2 \).

**Figure 11.** Squared anisotropy ratio for the anisotropic inversion result, iteration 9.
the hydrocarbon reservoir is not recovered in the horizontal resistivity, the ratio is extremely high at this location and obviously does not reflect reality: rather it is caused by the relative sensitivity to the horizontal and vertical resistivity within the layer. However the squared anisotropy ratio of the layer above the target is rather well recovered at ranges where there is sufficient source-receiver coverage: one observes a squared anisotropy ratio above 2, whereas the rest of the reconstructed model is largely isotropic. Away from the resistive reservoir, the maximal squared anisotropy ratio is 2.3 in Figure 10.

4.2. Real data

We now present the result from the inversion of survey data acquired in 2005 offshore West Africa. The survey consisted of two lines crossing a prospective hydrocarbon reservoir lying against a salt diapir (see Figure 12). The structure of the salt diapir is constrained using coincident seismic data. Its top lies at approximately 1.1 km below the seafloor, and is expected to be a resistive feature in the earth. In this study we present the results from the East-West trending line, which crosses the salt diapir (labelled line 1). The data consists of the horizontal electric field collected along the 25 km long line at source transmission frequencies of 0.125, 0.375 and 0.625 Hz. An example of the amplitude and phase data collected at receiver f11 (located at the crossing point of the two lines) is shown in Figure 13. Figure 13 also shows the resistivity log from a well located close receiver f11. A standard induction log from a vertical well measures the horizontal component of the resistivity. If the well log resistivity is block averaged, and the CSEM response of this block averaged well log model is compared to the acquired data then the misfit between the two is large. This is because in-line CSEM data, as shown earlier, are primarily sensitive to the vertical resistivity of the earth which is considerably higher. In this case an anisotropic model in which the horizontal resistivity equals that measured in the well, and the vertical anisotropy is one to five times higher provides a much better fit to the data (see Figure 13). The simple 1D modelling of the well log demonstrated that an anisotropic earth is required to explain the data and motivated the need to perform an anisotropic inversion. As a starting point an isotropic inversion of the data was performed, to provide a comparison with the anisotropic results.

The 2D isotropic resistivity section obtained at iteration 10 from the isotropic inversion of the amplitude data from line 1 is presented in Figure 14. The RMS misfit is equal to 0.82, and the spatial distribution of residuals is presented in Figure 15. In this case the misfit is well below 1, because the size of the error bars has been set to a minimum value of 5% of the amplitude, and the inversion succeeded in producing a model which reasonably fits the data to this degree.

![Figure 12](image-url) Survey map with the delineation of the salt diapir and the two CSEM lines with the receiver locations.
Figure 13. Electric field amplitude and phase data from receiver f11 (left panel). The right panel shows the resistivity log from a well close to this receiver. If a model is constructed from the well log data, which primarily measures the horizontal resistivity, then this does not fit the data (green curve). A much better fit is achieved by an anisotropic model in which the vertical resistivity is considerably higher than the horizontal (red curve).

Figure 14. Isotropic inversion result, survey data from line 1, iteration 10.

The absolute values of the error bars have however a minimal impact on the final inversion result when they are modified for the whole dataset. Only a modification of the error bars relating to a part of the dataset can significantly alter the inversion result by altering the data balance between different offsets. The fundamental frequency residuals along the central part of the line are highest, with positive bias at short ranges and negative bias at long range. Their mean is 0.1812. The residuals for the fifth harmonic are also biased, although they are smaller in magnitude with a mean of 0.2553. The resistivity model recovered is shown in Figure 14 and is rather oscillatory. As observed in the synthetic case this roughness is an artifact of the inappropriate isotropic assumption. The increase in resistivity towards of the East of the line may be related to the presence of the salt diapir at depth, however neither the position nor the
resistivity of this are well constrained. In addition, a regional resistive layer is observed at about 3 km below sea surface.

The 2D anisotropic resistivity section obtained at iteration 13 from the anisotropic inversion of the same data is presented in Figure 16. The RMS misfit is equal to 0.38, and the spatial distribution of residuals is presented in Figure 17. The final RMS is very low, again due the values assigned to the error bars. The residuals more closely follow a normal distribution this
time (their mean is -0.0049 and standard deviation 0.3800), and the RMS misfit is clearly lower than with the isotropic inversion. From Figure 16, we can observe that the salt diapir is clearly resolved in the anisotropic result as an area of elevated horizontal and vertical resistivity. A higher vertical resistivity is also recovered to the West of the diapir in the vicinity of the well. The high horizontal and vertical resistivity below 5000 m indicates electrical basement. The result presented in Figure 16 has a low structural resolution since it results from an unconstrained inversion of the data. One way to improve the resolution would be to include seismically derived structural information, for example breaks in the smoothing constraint at the top basement and diapir interfaces.

The result corendered with the seismic section shows a very good match between the seismic delineation of the salt diapir and the feature in the horizontal resistivity (Figure 18). It can also be noted that the top western part of the diapir is surprisingly more conductive, suggesting the presence of highly saline fluids associated with the diapir in the sub-surface in this area.

![Figure 18. Anisotropic inversion result corendered with the seismic section, iteration 13.](image)

Finally, the 1D profile the anisotropic inversion result is compared with the well resistivity log data measured near receiver f11, i.e. at 18.2 km along the line in Figure 19. The horizontal resistivity recovered by the anisotropic inversion agrees with the well log. The reconstructed

![Figure 19. 1D resistivity profile compared to the well log data.](image)
horizontal resistivity is slightly higher than the well log at depths below 1800 m, also due to the smoothing which spreads the resistivity from highly resistive regions to low resistive ones. The vertical resistivity is higher and follows a trend similar to the horizontal resistivity, with an average squared anisotropy ratio equal to 3.22, consistent with earlier forward modelling.

5. Conclusion
When the earth is anisotropic, an isotropic inversion of inline CSEM data will lead to artifacts in the results which are misleading if interpreted in terms of geology. Comparatively, the anisotropic inversion provides a result which is very close to the true model in synthetic inversions and prevents excessive oscillations in the retrieved resistivity profiles. Another benefit of the anisotropic inversion is that the horizontal resistivity profile recovered from data collected in line with the source dipole will fit a well log (which only measures the horizontal resistivity of the earth), whereas the isotropic inversion will provide a profile which is more likely to look like the vertical resistivity.

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