CMP inversion and post-inversion modelling for marine CSEM data

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Abstract
Inversion of marine controlled-source electromagnetic (CSEM) data that have been sorted into common mid-point gathers can give an estimate of the resistivity section below a towline. The method is fast and requires relatively small computer resources even for a Levenberg-Marquardt optimization scheme. However, it is not a standalone method. Proper 3D modelling with quantitative error analysis is used in a post-inversion step to refine the resistivity model. We demonstrate how inversion of common mid-point gathers can be used in combination with post-inversion modelling to improve the interpretation of a marine CSEM profile from South-East Asia. Two resistors are identified on the section, and the deeper one is a confirmed gas accumulation in Upper Miocene turbidites.

Introduction
Marine controlled-source electromagnetics (CSEM) or seabed logging (SBL) is now an established technique for hydrocarbon exploration (Eidesmo et al., 2002; Ellingsrud et al., 2002; Srnka et al., 2006). Marine CSEM methods use an electric dipole to probe the subsurface. The technique has proven particularly useful for detecting thin highly resistive layers that are typical for hydrocarbon reservoirs.

Marine CSEM is a low frequency method, which limits the resolution on depth sections obtained with processing methods such as depth migration and inversion. Furthermore, inversion of marine CSEM data gives non-unique results, i.e., there is always more than one resistivity model that can explain the data. The non-uniqueness can take several forms and we discuss two types in this article. One effect is that the measured response from a resistor at a certain depth may be similar to the response from a deeper resistor of greater resistivity. A second effect has to do with the resistor thickness. Marine CSEM data are mainly sensitive to the transverse resistance of a resistor, which is the product of resistivity and thickness. The transverse resistance of a resistor is retrieved by inversion, but in such a way that the thickness is overestimated and the resistivity is underestimated unless constraints are imposed. Thus, the inverted resistivity section usually displays an unrealistically thick resistor. The implication is that the resistivity section requires further interpretation. We propose the use of post-inversion 3D modelling with resistors of realistic size to check which models explain the observed data.

Here we introduce a common mid-point (CMP) inversion algorithm for marine CSEM data. The advantage is that the 1D earth assumption is a better approximation for a CMP gather than for a common source or common receiver gather. There is still room for lateral variation, but as a function only of the CMP coordinate. Very fast modelling tools can be used for plane-layer geometries and implemented in inversion schemes that are numerically highly efficient. The numerical efficiency is the main motivation for working with this CMP inversion method. Initial estimates of the resistivity as a function of distance and depth can be available in 5–10 minutes. The results for CMP intervals of 500 m and a depth sampling interval of 25 m may be available within less than one hour given a good initial model, but we emphasize that CMP inversion is not a standalone method due to the simplifying assumptions that are made. Proper 3D modelling is required for resistivity model verification in a post-inversion step.

We use a Levenberg-Marquardt optimization scheme as described by Mittet et al. (2004). This scheme requires an estimate of the Jacobian with respect to the residual. The calculation of the Jacobian is the main numerical cost in

Figure 1 Seismic section along the towline. The flat spot is at a distance of 10–12 km and depth of 2100 m.

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this inversion scheme. We use plane-layer modelling (Loseth and Ursin, 2007) as the workhorse in the inversion scheme. With plane-layer modelling we obtain electromagnetic fields that have realistic 3D near-field effects and 3D geometrical spreading. We have implemented anisotropic CMP inversion as an option, but it is the isotropic mode that is used for the work discussed here. CMP inversion is one of the schemes that can be used to obtain a 2D resistivity section where marine CSEM data are acquired along a profile. Two other methods are 2.5D inversion and 3D inversion.

In 2.5D modelling and inversion, we work directly in the source-receiver domain. The earth is assumed to be invariant in the direction normal to the toinline (2D earth), but realistic 3D near-field effects and geometrical spreading are taken into account. One important benefit is that the processing time is reduced significantly compared to 3D inversion. For the real earth, it is never the case that the resistivity anomaly is infinite in the crossline direction. However, it may be a good approximation, depending on frequency, if its crossline width is sufficiently large (more than approximately 5 km at 0.25 Hz). If the crossline width of the resistivity anomaly is small, good lateral definition is still possible but there is an increased risk of incorrect positioning in depth, as with CMP inversion. An additional problem is that strong seabed bathymetry effects normal to the toinline direction may introduce errors. Used correctly, 2.5D inversion of marine CSEM line data is a valuable and important processing tool.

Ideally, inversion of marine CSEM data should be performed with a 3D algorithm. All 3D effects are naturally included and the resistivity model can have 3D variations. However, 3D inversion is not only a question of methodology; it depends on the coverage of measured data. The full potential of 3D inversion can only be realized if data are available on a 2D surface with sufficiently dense sampling. Standard marine CSEM survey data are normally available only along single profiles, but we anticipate the development of wide-azimuth data acquisition, as we currently see in the seismic industry. For marine CSEM surveying, wide-azimuth acquisition can be achieved with receivers covering a 2D grid on the seabed. Processing with a full 3D scheme is possible, but numerically very costly.

For electromagnetic data acquired along a profile, the resistivity section derived by inversion requires further interpretation. This requirement applies to CMP inversion, 2.5D inversion, and 3D inversion and is due to limitations in the dataset and possibly also due to the approximations used in the forward model for the inversion scheme.

CMP inversion represents a fast and viable method in situations where we have data on a single profile in combination with moderate lateral resistivity variations. Our strategy is to view any type of inversion as the first step in a processing and interpretation scheme. The second step is to do a set of 3D modelling operations (Maasø, 2007) to obtain realistic interpretations for the 3D resistivity distribution in the survey area. Results from inversion of the marine CSEM data plus seismic data and also MT and gravity data, if available, should be used to build these interpreted resistivity models. For each of these models it is important to measure and quantify the fit between the synthetic data and the observed data. This second step is not always performed, but it is essential in our view. The final interpreted resistivity volume should be able to give modelled data that fit the observed data.

**Theory**

We assume a coordinate system where the receivers are oriented so that the \( x \)-axis points in the towline direction (Mittet et al., 2007). The CMP coordinate \( x_\text{c} \) and the offset coordinate \( x_\text{o} \) can be derived from the source coordinate \( x_\text{s} \) and the receiver coordinate \( x_\text{r} \) as,

\[
x_\text{c} = \frac{1}{2} (x_\text{s} + x_\text{r}),
\]

\[
x_\text{o} = (x_\text{r} - x_\text{s}).
\]

The data are recorded in common receiver gathers and we need to be sorted into CMP gathers. The source coordinate is continuous for time domain data and can be made densely sampled for frequency domain data. The much coarser receiver interval, usually of the order of 1 km, is a challenge. Interpolation is obviously needed to produce CMP gathers data that are densely sampled in the offset coordinate. To obtain the field at an arbitrary receiver coordinate, we obtain the best results if we use simple linear interpolation in that domain. Thus, if \( x_\text{s}^{\text{e}} \) and \( x_\text{r}^{\text{e}+1} \) are the receiver coordinates for two neighbouring receivers and we need data at the arbitrary location \( x_\text{r} \) somewhere between \( x_\text{s}^{\text{e}} \) and \( x_\text{r}^{\text{e}+1} \), we use

\[
E_\text{os}^\text{obs}(x_\text{s}, x_\text{r}, \theta_0) = \frac{E_\text{os}^\text{obs}(x_\text{s}, x_\text{r}^{\text{e}+1}, \theta_0)(x_\text{s} - x_\text{r}^{\text{e}+1}) + E_\text{os}^\text{obs}(x_\text{s}, x_\text{r}^{\text{e}}, \theta_0)(x_\text{s}^{\text{e}} - x_\text{r})}{x_\text{s}^{\text{e}} - x_\text{r}^{\text{e}+1}},
\]

where \( E_\text{os}^\text{obs} \) is the observed inline electric field expressed in terms of source and receiver coordinates. Sorting gives the
observed inline electric field expressed in terms of CMP and offset coordinates, $E^\text{Obs}_x(x_1, x_2, \omega)$, related to $E^\text{Ob}_x$ by

$$E^\text{Obs}_x(x_1, x_2, \omega) = E^\text{Ob}_x(x_1 + \frac{1}{2} x_2, x_2 - \frac{1}{2} x_1, \omega).$$

(3)

Our scheme uses the absolute value of the offset. Then we may get two contributions to a given CMP-offset location. The source and receiver positions can be switched without altering the CMP coordinate and the absolute value of the offset. We use this symmetry property for the electromagnetic fields and average the two contributions.

We need the electric and magnetic difference fields $\Delta E_h(x_1, x_2, \omega)$ and $\Delta H_h(x_1, x_2, \omega)$ for the definition of the data residuals used in the inversion,

$$\Delta E_h(x_1, x_2, \omega) = E^\text{Obs}_h(x_1, x_2, \omega) - E^\text{Mod}_h(x_1, x_2, \omega),$$

$$\Delta H_h(x_1, x_2, \omega) = H^\text{Obs}_h(x_1, x_2, \omega) - H^\text{Mod}_h(x_1, x_2, \omega),$$

(4)

where $E^\text{Mod}_h(x_1, x_2, \omega)$ and $H^\text{Mod}_h(x_1, x_2, \omega)$ are modelled fields using the resistivity model at the present iteration. $H$ is the crossline magnetic field. The data-space residual, $R^\text{D}(x_1, x_2, \omega)$, is

$$R^\text{D}(x_1, x_2, \omega) = W^E(x_1, x_2, \omega) \delta E_h(x_1, x_2, \omega) \Delta E_h(x_1, x_2, \omega)$$

$$+ W^H(x_1, x_2, \omega) \delta H_h(x_1, x_2, \omega) \Delta H_h(x_1, x_2, \omega).$$

(5)

Here $W^E(x_1, x_2, \omega)$ and $W^H(x_1, x_2, \omega)$ are weight functions which ensure that we include contributions to the residual from intermediate and large offsets. Typically,

$$W^E(x_1, x_2, \omega) = \left( \frac{\omega(x_1, x_2, \omega)}{|E^\text{Ob}_x(x_1, x_2, \omega)|} \right)^\gamma + N_{E,h}(x_1, x_2, \omega),$$

$$W^H(x_1, x_2, \omega) = \left( \frac{\omega(x_1, x_2, \omega)}{|H^\text{Ob}_x(x_1, x_2, \omega)|} \right)^\gamma + N_{H,h}(x_1, x_2, \omega),$$

(6)

where $N_{E,h}(x_1, x_2, \omega)$ and $N_{H,h}(x_1, x_2, \omega)$ are estimates of the noise in the observed data, $\omega_1$ is the first harmonic and $\gamma = -0.5$. Both inner and outer mutes are easily implemented with the weight function by setting the weights to zero for the mute intervals.

We use the dimensionless residual of Equation (5) for the inversion, but we also define a similar dimensionless residual as a function of source and receiver coordinates,

$$\tilde{R}^\text{D}(x_1, x_2, \omega) = \tilde{W}^E(x_1, x_2, \omega) \delta E_h(x_1, x_2, \omega) \Delta E_h(x_1, x_2, \omega)$$

$$+ \tilde{W}^H(x_1, x_2, \omega) \delta H_h(x_1, x_2, \omega) \Delta H_h(x_1, x_2, \omega),$$

(7)

which we use for error estimation in the post-inversion modelling step. In the following we use Equation (7) to calculate $\tilde{R}^\text{D}(x_1, x_2, \omega)$, but plot this residual using CMP location and offset as the coordinates so that we can easily see where along the towline we have a good or bad fit between the observed and modelled data.

In order to stabilize the inversion we also define a model-space residual, $R^\text{M}(x_1, x_2, \omega)$. The model-space residual is a slight modification of the minimal gradient support functional (Portniaguine and Zhdanov, 1999). We allow for an arbitrary even exponent $k$, and different weights for horizontal and vertical model variations. Normally we adjust $\beta_{\text{hor}}$ and $\beta_{\text{ver}}$ to give some more weight to the horizontal part, but seldom more than 50 percent. The model-space residual is given by

$$R^\text{M}(x_1, x_2, \omega) = \beta_{\text{hor}} \sum_{x_1 \omega} (\rho(x_1, x_2, \omega) - \rho(x_1, x_2, \omega))^k$$

$$+ \beta_{\text{ver}} \sum_{x_1 \omega} (\rho(x_1, x_2, \omega) - \rho(x_1, x_2, \omega))^k,$$

where $\rho(x_1, x_2, \omega)$ is the resistivity at depth $z$ for CMP position $x_1$. The value of $k$ is still subject to trial and error. For this particular scheme a value in the range 4–6 seems to give good results.

The dimensionless model-space residual gives a coupling between the CMP locations. In this scheme we do not invert for resistivity at each CMP location sequentially. Due to the coupling introduced by the model-space residual, we obtain a simultaneous update for all CMP positions at each iteration.

**Results**

We demonstrate the combination of CMP inversion and post-inversion modelling for a marine CSEM dataset acquired over a prospect in South-East Asia. A total of 18 receivers were deployed, with a separation of 1 km between each receiver. Good data were recorded on 17 of these receivers. One receiver was dropped from further processing. This receiver was located 4 km from the start of the line. A seismic section covering the distance between the first and last receiver is shown in Figure 1. The seismic section has a clear flat spot between 10 km and 12 km along the profile at a depth of 2.1 km (Figure 1). Based on the seismic data, the prospect was interpreted as an Upper Miocene deepwater-slope turbidite reservoir sequence bounded at its base by an unconformity. A regional study of the area interpreted the formation beneath the unconformity to be mobile shale originating from pro-delta facies of Late Middle Miocene age. The prospect is overlain by a graphic interval was thought to be equivalent to intervals water shales that form the top seal. This reservoir stratigraphic interval was thought to be equivalent to intervals found elsewhere in the area where nearby wells had penetrated 100–150 m of net sand with good reservoir quality. The prospect displays a significant seismic amplitude and AVO anomaly at the top reservoir horizon. The termina-
The top of the mobile shale layer is visible in the seismic data as the undulating surface at 2000–2400 m, below which there is a clear degradation in seismic resolution. No wells penetrate the mobile shale layer in the area, but a low resistivity for this layer was predicted due to the expected high content of brine.

The decision to drill was taken based on the seismic data alone. However, the CSEM data were acquired before drilling. The result of the CMP inversion of the CSEM data is shown in Figure 2. The resistivity model after 37 iterations is superimposed on the seismic data. Similar CMP inversion results were available before the reservoir section was penetrated. All of them gave a clear indication of hydrocarbons in the reservoir and indeed this turned out to be a discovery. The result shown in Figure 2 can be obtained in less than 1 hour on a small Linux cluster. For this particular example we used 35 nodes. The separation between each CMP location was 500 m, which gives 35 CMP positions over 17 km. One node is used to calculate the part of the Jacobian related to a given CMP location.

The resistivity model in Figure 2 gives a good estimate of the background resistivity. In particular we recover low resistivity below the undulating surface at 2000–2400 m. This background resistivity model is used for the post-inversion modelling. We find it encouraging that the reservoir appears with such a strong response. However, we also note that there is a difference in size between the seismic anomaly (flat spot) and the electromagnetic anomaly (red blob). The flat spot is only 2 km in length in the towline direction, whereas the electromagnetic anomaly is 4 km long on the resistivity image. The second part of this section shows how we use post-inversion modelling to resolve this problem, but first...

**Figure 3** Observed data and modelled data for the initial model: (a) amplitude, (b) phase. Observed data are plotted with thick lines and modelled data with thin lines. Red curves are for 0.25 Hz; green curves are for 0.75 Hz; and blue curves are for 1.25 Hz.

**Figure 4** Residuals for the initial model used in the CMP inversion.
we need to discuss the fit between observed and synthetic data. Resistivity models from inversion procedures should not be used for interpretation unless error estimates are also available.

The initial model used here had a resistivity of 2 $\Omega$ m in the subsurface. The water-layer resistivity was measured during acquisition and set to an average value of 0.305 $\Omega$ m. Observed data and the response from the initial model are shown in Figure 3. The data are for a single CMP position at the location of the flat spot. The thick lines are for observed data and the thin lines are for modelled data. We used three frequencies for the inversion, 0.25 Hz, 0.75 Hz, and 1.25 Hz. Figure 3a shows amplitude as a function of offset and Figure 3b shows phase as a function of offset. The largest offset used is 10 km. We observe that the initial model does not explain the observed data. The data-space residual, $R_D(x_h, x_c, \omega)$ from Equation (5) can be used for analysis of the error along the towline. This quantity is shown in Figure 4 for the three frequencies. Again it is clear that the initial model does not explain the observed data. We have used an inner mute from 0–1500 m for all frequencies in the calculation of residuals. The calculation of the modelled data is performed with a point-like transmitter. The real data are for a transmitter with a length of 270 m. For offsets above approximately 1500 m the difference between a point-like transmitter and a realistic size transmitter can be neglected for the given frequencies. The outer mutes are frequency-dependent and determined by the offset where the field amplitudes on average are comparable with the noise floor. There is no outer mute for 0.25 Hz. The outer mute for 0.75 Hz starts at the offset of 7000 m and the outer mute for 1.25 Hz starts at the offset of 6000 m.

After 37 iterations we obtain the results for amplitude and phase shown in Figure 5. The CMP location is the same as for Figure 3 so the observed data are identical in these two figures. However, the synthetic data in Figure 5 are very close to the observed data. In Figure 6 we show $R_D(x_h, x_c, \omega)$ for the resistivity model obtained after 37 iterations. Compared to Figure 4 we observe that the residual is much reduced for all frequencies, at all offsets and for all CMP positions along the towline. This is only possible if the difference between observed and synthetic data is small everywhere. The small residual in Figure 6 is due to a resistivity model that does a very good job in explaining the observed data in the CMP-offset domain. However, it is important to bear in mind the non-uniqueness problem which is inherent in the inversion of CSEM data. There are always multiple resistivity models that give approximately the same final error or residual. We need additional data to distinguish between such models. Here we also need to address the effect of the transformation from source-receiver domain to CMP-offset domain. This transformation opens up for the use of a numerically very efficient inversion scheme, but we expect that one of the undesired side effects is lateral smearing of the subsurface response.

For the post-inversion modelling we start with the resistivity model shown in Figure 2, except that we modify the high resistivity part of the model and use typical background resistivities of 2–3.5 $\Omega$ m here. We assume that the 3D background resistivity model has no variations normal to the towline. The data-space residual, $\tilde{R}_D(x_h, x_c, \omega)$ from Equation (7) can be used for analysis of the error along the towline. This quantity is shown in Figure 4 for the three frequencies. Again it is clear that the initial model does not explain the observed data. We have used an inner mute from 0–1500 m for all frequencies in the calculation of residuals. The calculation of the modelled data is performed with a point-like transmitter. The real data are for a transmitter with a length of 270 m. For offsets above approximately 1500 m the difference between a point-like transmitter and a realistic size transmitter can be neglected for the given frequencies. The outer mutes are frequency-dependent and determined by the offset where the field amplitudes on average are comparable with the noise floor. There is no outer mute for 0.25 Hz. The outer mute for 0.75 Hz starts at the offset of 7000 m and the outer mute for 1.25 Hz starts at the offset of 6000 m.

Figure 5 Observed data and modelled data for final model (37 iterations): (a) amplitude, (b) phase. Observed data are plotted with thick lines and modelled data with thin lines. Red curves are for 0.25 Hz; green curves for 0.75 Hz; and blue curves for 1.25 Hz.
ly dependent on the resistor thickness alone, or resistivity alone, for a fixed transverse resistance. Hence, we add resistors to the background model with a 50 m thickness. The transverse resistance of the resistor is then varied by altering the resistivity. We used additional seismic data to estimate the width of the assumed resistor in the crossline direction at the flat spot location to 3–4 km. Our best estimate from comparing observed and modelled electromagnetic fields is 3 km. All results shown in the following are with this width.

We have already noted the discrepancy between the length of the flat spot (2 km) and the EM anomaly (4 km). In order to try to resolve that problem we have generated resistivity models with resistors that are 2 km and 4 km in length. The locations of the resistors are shown in Figure 8. We vary the depth of these objects from 1300–2300 m. The resistivity is varied from 10 to 100 $\Omega$ m. The CPU time required for a modelling job increases with the highest resistivity in the model and will typically be between 4 minutes and 30 minutes. All frequencies can be extracted from one modelling job. For the 3D post-inversion modelling we display results for 0.75 Hz only. The reservoir is small and for this dataset we have found that we have the best sensitivity at this frequency.

poses we transform this residual to the CMP-offset domain (Figure 7). It is easier to see where along the towline we have a large discrepancy between observed and modelled data in the CMP-offset domain as compared to the source-receiver domain. However, the calculation of the residual is more accurate in the source-receiver domain which is also the experimental domain. The residual in Figure 7 can be compared with the residuals in Figures 4 and 6, but as can be seen from the colour scales we have an increased sensitivity to small errors in Figure 7. We will use this scale in the following analysis. The reason is that we assume that most of the large-scale features are included in the background model but the details still need to be resolved and we use this increased sensitivity to resolve them.

In the post-inversion modelling we use the background resistivity model derived from the CMP-inversion and add resistors to this model. In our first series of experiments we added resistors at locations near the flat spot. The amplitude and phase responses from a resistor depend mainly on the transverse resistance, which is the product of resistor thickness and resistivity. This is a very good approximation for the amplitude response and the phase response is only weakly dependent on the resistor thickness alone, or resistivity alone, for a fixed transverse resistance. Hence, we add resistors to the background model with a 50 m thickness. The transverse resistance of the resistor is then varied by altering the resistivity. We used additional seismic data to estimate the width of the assumed resistor in the crossline direction at the flat spot location to 3–4 km. Our best estimate from comparing observed and modelled electromagnetic fields is 3 km. All results shown in the following are with this width.

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In Figure 9a we show the residual for a resistivity model where the red resistor of 4 km length (see Figure 8) is assigned a resistivity of 70 $\Omega$m at a depth of 1700 m. The difference between observed and modelled data is better than for the background model, as can be seen by comparing with Figure 7. The result of increasing the burial depth of this resistor to 2100 m is shown in Figure 9b. This gives an even better fit between the observed and modelled data, but we see that we still have contributions to the residual at three intervals along the towline: from 4–7 km; from 8–10 km; and at the far right of the plot (16 km and above). The red resistor starts at 8 km and ends at 12 km. The result of replacing it by the shorter green resistor of 2 km length (see Figure 8) at 2100 m depth is shown in Figure 10. This resistor starts at 10 km and ends at 12 km. The change of resistor length has two effects. Compared with Figure 9b we see that the residual is reduced between 8–10 km, so we have a better fit to the data here. On the other hand we see that the residual increases in the interval from 4–7 km, so the difference between the observed and modelled data is increased here.

There is now consistency between the length of the flat spot and the length of the resistor used to represent the reservoir. However, there is still a fairly large part of the data that is not explained so we included a second resistor, the blue resistor in Figure 8, with a length of 3 km between 4–7 km. The seismic data were used to estimate that the most probable crossline width of this resistor is only 1.8 km. Because of its limited crossline width, the response of this resistor can only be calculated accurately with 3D modelling.

In Figure 11a we see the result of placing this second resistor at depth 1300 m with resistivity 30 $\Omega$m. Comparing Figures 10 and 11a, it is not yet obvious that inclusion of the second resistor improves the fit between the observed and modelled data. However, when the second resistor is placed at a depth of 1700 m we have the result shown in Figure 11b, and the improvement in fit is dramatic.

It is interesting to have a closer look at the sensitivity of the modelled data to the inclusion of each resistor. We can also get an impression of the uncertainty in transverse resistance versus burial depth. Figure 12 shows the data misfit, $\varepsilon^D$, as a function of resistivity and burial depth. The resistor
thickness is kept fixed at 50 m, so the resistivity axis in this figure could just as well have been converted to transverse resistance. The data misfit is

\[ e^D = \sum_{s_i, s_j, \omega} \tilde{R}^D(s_i, s_j, \omega). \] (9)

In this case we have kept the resistor between 10–12 km at a fixed depth of 2100 m but varied the properties of the resistor between 4–7 km. The data misfit is small for a resistor with a transverse resistance of 1500 Ωm², and is minimum for burial depths between 1600 m–1800 m. There is a trend in Figure 12 indicating that the data misfit stays small if both resistivity and burial depth increase simultaneously.

Figure 13 shows \( e^D \) as a function of resistivity and burial depth when the resistor between 4–7 km is fixed at 1700 m depth with 30 Ωm resistivity and the resistivity and burial depth of the resistor between 10–12 km are varied. The trend of small data misfit when both the resistivity and burial depth increase simultaneously is more pronounced here than in Figure 12. The absolute minimum is for a transverse resistivity of 3500 Ωm² at a depth of 2100 m, but we observe a good fit between the observed and modelled data for depths from 1900–2300 m. We also observe that the transverse resistivity can vary between 2000–5000 Ωm² and still give a small data misfit. This is one manifestation of the non-uniqueness problems that we have in marine CSEM data: there is obviously more than one resistivity model that can explain the observed data.

The two resistors included up to now do a fair job in explaining the observed data, except for the misfit observed at 16 km and above. Figure 14 is a plot of the same residual as that in Figure 11b except that the polarity of the residual is retained. The sign is positive if the amplitude of the observed data is larger than the amplitude of the modelled data. The residual above 16 km has a negative sign, so the amplitude of the modelled data is larger than the observed amplitude. The most likely explanation is that the background model is too resistive in this area. A reduction in the resistivity of the background model at this location would naturally lead to reduced amplitudes for the modelled data due to increased absorption. Hence, we do not expect this part of the residual to be related to an unidentified resistor.

Figure 15 summarizes the final interpretation. The marine CSEM data can be explained by background resistivities in the range 2–3.5 Ωm. The resistor between 10-12 km has a depth and length that is consistent with the flat spot on the seismic section. Its crossline width is 3 km and its transverse resistance is approximately 3500 Ωm². The post-inversion modelling leads to the inclusion of a second resistor between 4–7 km at a depth of approximately 1700 m with a transverse resistance of 1500 Ωm². There are also some strong seismic reflections at the location of this resistor although there is not a perfect overlap. We tested the effect of increasing the size of this resistor by 500 m to the left, to let it overlap more of the strong seismic reflection. The fit between the observed and modelled data was just as good as for the resistor indicated in Figure 15. The interpretation of the seismic data alone was that these strong seismic reflections were most likely due to gas, but lithological effects had not been ruled out. The post-inversion EM results indicate...
higher resistivity than the background formation. The location is not drilled.

**Conclusion**

We have demonstrated a CMP-inversion algorithm for marine CSEM data. The scheme gives a clear indication of a resistor at the location of a flat spot in a seismic section. The seismic and electromagnetic responses are due to a gas reservoir. The main advantage of the CMP inversion is that it is numerically very efficient. Good results can be available within less than one hour given a good initial resistivity model. This establishes the method as a first-look tool. However, this is not a standalone method due to the simplifying assumptions that are made.

Proper 3D modelling is required for model verification in a post-inversion step. We have demonstrated how the 3D modelling solved a problem where the CMP inversion gave a different size for the electromagnetic anomaly compared to the size of a seismic flat spot. The 3D modelling also helped to identify a second resistor. It is our opinion that 3D modelling with a quantitative analysis of the misfit between observed and modelled data for the final interpreted resistivity model should always be performed. In particular, we consider this to be important for data from single profiles where 3D seismic data are available to constrain the crossline widths of resistors.

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**References**


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