RESOLVING RESISTIVE ANOMALIES DUE TO GAS HYDRATE USING ELECTROMAGNETIC IMAGING METHODS

Carsten Scholl*, R. Mir, E. C. Willoughby, R. N. Edwards
Department of Physics
University of Toronto
60 St. George Street, Toronto, Ontario, M5S 1A7
CANADA

ABSTRACT
Active marine electromagnetic methods have proven to be a powerful tool to detect resistivity anomalies associated with gas hydrate. However, because the propagation of electromagnetic fields for these methods works in the diffusive regime the spatial resolution of the resistivity structure is limited. So far only bulk electrical properties have been estimated from measured data, although hydrate bearing layers are found to be highly heterogeneous. We computed response curves for synthetic one- and two-dimensional models to investigate the resolution capabilities for various measurement geometries with respect to resistive features. Electric dipole transmitters (TXs) are used as sources. In the marine case, the in-line electric dipole-dipole configuration has proven its capabilities to detect the shallow resistive gas-hydrate. Our model study demonstrates that both the depth to a resistive feature can be resolved nicely using data for multiple TX-RX offsets. However, resolving smaller features of the resistive zone, for example if the zone is split in separate resistive layers, is extremely difficult. The resolution of the target can be improved using electrical downhole transmitters. So far there have been no reports of the detection of permafrost gas hydrate deposits with surface electromagnetic methods. Our calculations show that a similar setup to that used in the marine case is capable of detecting gas hydrate on land. The resolution, however, is lower than for the marine case, because of the significantly greater depths to the target.

Keywords: electromagnetic methods, permafrost, gas hydrates

NOMENCLATURE

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INTRODUCTION
In almost all sedimentary environments, ions moving in the fluids within the pore space make the dominating contribution to the electrical resistivity of the formation [1]. In most saturated sedimentary formations, exceptionally high electrical conductivities can be attributed to highly saline brines in the pores, whereas low conductivities are encountered when a resistive material such as oil or gas occupies and blocks pore space.

Being electrical insulators, gas hydrate and free methane act in the same way as oil and therefore exhibit a resistivity signature [2, 3]. Therefore, electromagnetic (EM) methods can be used to remotely sense gas hydrate. The resistivity signature of a gas hydrate filled formation is far more pronounced than in seismic velocities. This produces...
enables EM methods to detect gas hydrate more directly than seismic methods. There are many reports of successful applications of EM methods to gas hydrate mapping in a marine environment [4, 5, 6].

These studies have proven the general sensitivity of EM methods to bulk gas hydrate concentrations in the marine environment. In this paper, we examine the usefulness of different source and receiver configurations for the exploration and monitoring of gas hydrates. In particular, the application of a borehole-to-surface configuration in the marine case is examined, as well as the resolution of a permafrost gas hydrate deposit with surface EM measurements.

**EM METHODS**

In EM methods, the electromagnetic response of the earth to a time varying source field is measured. Many different variations of EM methods are in use [7]. The electromagnetic responses are either interpreted in frequency or time domain. That is, the data is viewed as either the response to a continuous EM excitation at certain frequencies or as time varying response due to an EM excitation at \( t=0 \) s. The numerous methods mainly differ in the type of source (galvanic or inductive, natural or artificial) and the measured field component (electrical or magnetic and orientation), and the frequency band or time range used.

All EM methods - with the exception of ground penetrating radar [8] - work in a frequency band where the propagation of the electromagnetic disturbance produced by the source is a diffusive process, not a wave phenomenon. This limits the spatial resolution of the method. Features appear more blurred than in seismics. On the other hand, EM methods do not suffer from scattering but instead behave according to bulk properties.

For reasons discussed below, we will consider only active methods, where an artificial source is used (the transmitter or TX). From a time domain point of view, the current is changed in the TX at \( t=0 \) s. This means, for example, that the current in the transmitter is switched off or on or the polarity of the current is changed. This change induces current systems in the subsurface according to Lenz' law. With time, the current systems diffuse down and outwards, depending on the electric properties of the surrounding medium. The magnetic and electric field components of this diffusive process are recorded at the receivers (RXs).

The speed of the disturbance depends on the electrical resistivity \( \rho \) (or its reciprocal, the electrical conductivity) and the magnetic permeability \( \mu \). The distance that the maximum of the induced current system traveled in time \( t \) after changing the state of the TX, is

\[
d = \frac{2\rho}{\pi \mu} \approx 1262\sqrt{\rho},
\]

assuming that the magnetic permeability is close to the vacuum magnetic permeability, as is the case for most rocks. Hence, to probe greater depths, later times after a change of state in the TX are required.

Gas hydrate forms only under low temperature/high pressure conditions. In the marine case, gas hydrates have been found right below the seafloor down to a few hundred meters below the seafloor under minimum of 300-500 m of water. On land, hydrate can form only under the pressure of several hundred meters overburden. The order of magnitude for resistivity values are around 1 \( \Omega \)m or 10 \( \Omega \)m for a marine/permafrost environment, respectively. This means that data ideally should be recorded roughly in the time range from 0.1 ms up to a few seconds.

In a strict sense, the exploration depth of a certain setup does not depend on the offset between TX and RX, only on the measured time window. Nevertheless, the signal amplitude and therefore the signal-to-noise ratio (SNR) at a given time \( t \) depends on this offset. In general, the depth to which the data is most sensitive increases with offset. Due to the diffuse nature of the propagation of EM field disturbances, the spatial resolution of EM methods will decrease with increasing distance between the target and the TX or RX.

The main difference between the marine and the land case is that in the latter the upper half-space (air) is an almost perfect insulator. Hence, the information about the change of state in the transmitter reaches all receivers almost instantly, creating a discontinuity in the fields at \( t=0 \) s. The
signals after that originate from the field disturbance diffusing downwards into the subsurface. In the marine case TX and RXs are located in a conductive medium, commonly at or near the seafloor. Unless the water is very shallow, the upper half-space is dominated by the highly conductive seawater. Thus there is no instantaneous response at the receivers at t=0 s. In deep water, the part of the disturbance propagating through the more resistive formations below the seafloor arrives at the RXs first, followed by the part that propagated through the more conductive seawater.

The choice of the transmitter and receiver type is relevant because of the different electric current systems involved. Magnetic dipole transmitters on the surface produce only horizontal current systems in a layered subsurface. The same is true for natural sources used in passive EM methods. Grounded, electric dipole transmitters, however, produce current systems with horizontal and vertical components. Magnetic receivers on the other hand are only sensitive to horizontal current systems at the surface, whereas the electrical components are sensitive to the total current system.

Current systems with a vertical component are affected by resistive layers, since the current has to flow through the resistor. The EM field is distorted in the process. Changes in horizontal current systems propagate inductively to deeper layers and thus can "jump" over resistive layers. Hence, they are less affected by a resistive layer [9]. This means that in order to resolve a resistive target like gas hydrate it is beneficial to deploy grounded dipoles as TX and RXs. For marine cases, a configuration of coaxial, horizontal magnetic dipoles works as well [10]. The rapid decay of the magnetic dipole's source field with offset, however, makes it difficult to get sufficient signal-to-noise ratio for larger offsets, limiting the exploration depth of such a system to the first tens of meters.

The setup used at the UofT Group for marine gas hydrate exploration is shown in Figure 1. The method is commonly referred to as marine controlled-source electromagnetics (CSEM). An electric dipole transmitter is towed on the seafloor. The TX is followed by a chain of receivers at different offsets, recording the in-line component of the electric field parallel to the TX direction [2]. In other experiments, the TX dipole is towed
above an array of stationary ocean-bottom receivers, which typically record the other components of the EM field as well. Depending on the offset, marine CSEM can be used to detect resistive bodies down to more than a kilometer below the seafloor.

Figure 2 shows the setup of the land-based method called long-offset transient electromagnetics (LOTEM) [11]. The setup is similar to marine CSEM in that it also uses a grounded dipole as source. Electric and magnetic field components are measured at different receiver sites. The direction of the TX defines the x-direction.

In most cases both electric and magnetic components are measured simultaneously at each receiver site in order to improve the resolution of the subsurface, as we will show later. Typically, coils are used to measure components of the magnetic field so the measured datum is the time derivative of the magnetic component (dH_x/dt, dH_y/dt, and dH_z/dt). In popular setups the receivers are placed either in-line with the TX (Fig. 2b), where E_y, dH_x/dt and dH_z/dt vanish for a layered medium, or broadside (Fig. 2a), where only E_y and dH_x/dt vanish. Typically, LOTEM can be used to detect resistivity anomalies in the depth range from a few hundred meters down to several kilometers.

**MARINE GAS HYDRATE**

It is well established that marine CSEM can be used to estimate the bulk gas hydrate content of the sediments below the seafloor. Ongoing research focuses on the resolution of the spatial distribution, or temporal changes therein, of gas hydrate deposits.

![Figure 3](image)

**Figure 3:** Evolution of the electric field for a horizontal dipole TX at the surface (a-d) and for a vertical dipole at z=200 m (e-h) for two different models (see text).

Figure 3a and b show electric field magnitude at two different times 0.33 ms (a) and 12.2 ms (b) after switching on a 1 m long unit dipole. The x-directed TX dipole is located at the origin. The model is a 1 Ωm half-space below a sea layer with a thickness of 1 km and an electrical conductivity of 3.2 S/m. Clearly, the electric fields propagate faster in the more resistive material below the seafloor.

Figure 3c and d show the propagation of the electric fields where a resistive layer with 10 Ωm
and a thickness of 50 m is buried at a depth of 80 m. This resistive layer can represent a zone with a higher gas hydrate concentration. At the instant the current system reaches the more resistive zone, it starts to propagate along it far more rapidly than in the surrounding material.

Figure 4: In-line $E_x$-fields measured at five offsets for the two models shown in Figure 3a-b (“Halfspace”) and Figure 3c-d (“Resistive layer”).

Figure 5: Geometry for the marine resolution study; three $E_x$-RXs (triangles) at the seafloor are used with 100 m long TX dipoles.

Figure 4 shows electric field responses for these two models, simulated for seafloor in-line $E_x$-receivers at 50, 100, 200, 500 and 1000 m offset from the TX dipole. The curves are normalized to their direct current value. The most prominent effect is that beyond a certain offset, the initial increase of the fields appears earlier in time when the resistive layer is present.

Edwards [2] used this effect to determine parameters of a one-dimensional (1-D) model of gas-hydrate concentrations. The time to an initial increase of the fields at short offsets basically is a measure of the resistivity of the uppermost layer. The time delay at large offsets indicates the resistivity of the second layer, and thus its gas hydrate concentration. The offset at which this time delay appears is indicative of the depth to the resistor. However, constructing a vertical resistivity model from data recorded at different offsets has the drawback that multiple TX positions are required to separate horizontal from vertical resistivity changes.

Figure 6: Data calculated for the configurations shown in Figure 5 (a: seafloor TX, b: downhole TX). The colors correspond to those of the triangles in Figure 5.

Scholl and Edwards [12] propose to use a vertical electric downhole transmitter with seafloor receivers in order to improve resolution of the resistivity structure. Figure 3e-h show the evolution of the electric field for a vertical, 1 m long unit dipole at a depth of 200 m below the seafloor at $x=0$ m. The fields behave similarly to
those shown in Figure 3a-c with the exception that the field has to propagate through the resistive target layer even at short offsets to reach the seafloor. This means that the data recorded at the seafloor for all offsets is sensitive to this layer. On the downside, this also means that the depth of the resistor is only poorly resolved. [12]. This can be partly countered by using a vertical dipole that extends from a depth above the resistor to one below it. This way, the vertical position of the resistor also defines how much of the TX dipole is above or below the resistor, significantly changing the shape of the measured curves.

In particular, the configuration is useful for monitoring applications, because here the number of transmitters is limited to only a few, at most. Furthermore, in most monitoring applications the depth range in which changes can occur are known so the lower vertical resolution is less of an issue.

The gas hydrate deposits typically exhibit small scale heterogeneities [13]. In this paper we investigate the resolution of the two configurations to finer structures within the layer. The configuration used in this example is shown in Figure 5. The target (“gas hydrate zone” in Figure 5) is not a homogenous layer, but instead consists of a more conductive layer embedded in a 10 Ωm layer (see black line in Figure 7). The data for both configurations is shown in Figure 6. Gaussian noise was added to all data points. The standard deviations of the noise are estimated from previous field measurements to be $2 \times 10^{-10}$ V/m (RX1), $4 \times 10^{-11}$ V/m (RX2) and $5 \times 10^{-12}$ V/m (RX3).

Detailed resolution studies for EM methods are not straightforward because of the complex dependency of the response curves on the model parameters. Simply comparing responses for different models is easy, but unfortunately a measurable difference in the response curves does not mean that a certain model can be resolved uniquely. As in many other methods in geophysics, noisy and incomplete EM data sets suffer a certain degree of ambiguity. That is, different models can be found that fit the data equally well.

In the best case, all models fitting the data show common features that result in the same geological interpretation. We will explore the model space for equivalent solutions by means of a non-linear inverse algorithm with random starting models. First, each parameter (layer thickness or resistivity) of the original five layer model (see Figure 7) is changed in turn in twelve steps over 3 orders of magnitude. The other parameters are changed randomly by 20%. These models then are used as starting models for a Marquardt-type inversion [14, 15], which tries to automatically change the model parameters so that the model response fits the original simulated data.

No 1-D model with only three layers was found to fit the data for either configuration. This means that the data indicates that the layer is not just one
uniform resistor. The results of the equivalency calculations are shown in Figure 7. The gray lines show equivalent models that fit within their error estimates. In the seafloor-TX case (Figure 7a) all models include conductive layers, predominantly in the upper part of the target depth range. Position, thickness and resistivity of this conductor, however, are not resolved. The resolution is far superior for the downhole TX (Figure 7b). The more resistive parts of the target are very well resolved with minor ambiguities of the embedded, conductive layer.

The downhole configuration can also be used in more complex environment. Figure 8 demonstrates the effect of a two-dimensional (2-D) structure. The data for this plot was calculated using a time domain, finite difference algorithm based on spectral Lanczos decomposition [16]. Plotted is the evolution of the magnitude of the electric field for a vertical, electrical dipole extending from 50 m to 200 m below seafloor. In the example, the resistive target layer (zone of high gas hydrate concentration) ends at an offset of 350 m away from the borehole. At early times (8a and c), the change in the current system propagates through the resistive zone quickly. There is no obvious difference between the infinite (a) and the finite layer (c). This changes when the current systems reach the end of the layer at 350 m. The current system now can only propagate with the slower speed of the more conductive background material (d) unlike the infinite layer (b), which significantly distorts the current system.

Interpretation of complex resistivity structures for electromagnetic methods still is difficult [e.g. 17]. Nonetheless, the discontinuities in the layer produce measurable effects in the electric fields at the seafloor which can help to qualitatively identify lateral variations. Panels e) and f) show the component-wise differences between the electric fields for the two models used in this figure. The difference in the $E_x$-component exhibits a sign reversal at the lateral position at which the model discrepancy occurs. The difference in the vertical $E_z$-component exhibits a maximum at the same position. The time at which these patterns first appear is indicative of to the depth to the model discrepancy.
APPLICATION OF LOTEM TO PERMAFROST GAS HYDRATE

Several authors report the application of EM methods in a permafrost environment [18, 19, 20, 21], in which gas hydrate could potentially be found. In most cases, however, the target was the permafrost layer. Gas hydrates were not detected, due to the limited resolution of the methods used with respect to these deep seated resistors. LOTEM potentially has a higher sensitivity as other EM methods to deep, resistive targets.

Figure 9: Location of the Mallik test site.

In order to test the feasibility of LOTEM to the detection of permafrost gas hydrate, we chose a resistivity model derived for the gas hydrate test site at Mallik, NWT, Canada [22]. We chose a 1-D model for the study because there is not sufficient information available to derive a more complex resistivity model. In addition, 1-D forward and inverse calculations are significantly faster than multidimensional computations.

The geological setting at Mallik

The Mallik test site is located east of the Mackenzie Delta, NWT, Canada (Figure 9). The area is covered with a thick sequence of sediments. The gas hydrate was found beneath 300 m to 700 m of permafrost within unconsolidated or poorly consolidated sediments of Tertiary age (Figure 10).

The Mallik research report [22] provides resistivity logs for depths between 810 and 1152 m. Above and below the hydrate-bearing zone (between approximately 890 and 1140 m), the resistivity logs show rather uniform values of a few $\Omega$m. Within the hydrate zone, the resistivity is in general about one order of magnitude higher, with maximal values of about 120 $\Omega$m [23]. The log exhibits a series of spikes within the hydrate layer, indicating a heterogeneous distribution of gas hydrate.

Figure 10: Stratigraphic column in the Mackenzie-Beaufort region; Permafrost and gas hydrate interval are not to scale (modified from [22]).

The shallow resistivity structure is relevant, because the resolution of a specific feature at depth depends greatly on the resistivity structure of the overburden. Unfortunately, the logs do not provide resistivity values at smaller depths. We
will use typical values for the permafrost layer in the area as measured by Todd and Dallimore [21]. Their results and the well-log combined in one model yield the resistivity model shown in Figure 11. Todd and Dallimore interpret the resistive layers as ice-bonded sediments and the conductive parts as non-ice-bonded with the pore space filled with water.

Figure 11: Resistivity model for the Mallik test site compiled from borehole and surface EM data.

Resolution of the 1-D model
Figure 12 shows transients for the $E_x$- and the $dH_z/dt$-component computed for the model shown in Figure 11. The fields are calculated for the broadside ($dH_z/dt$) and in-line ($E_x$) configuration. The TX has a length of 1 km. The receiver locations are at an offset of 3 km from the midpoint of the TX. The current is either switched on to 60 A (Figure 12a,b) or switched off from 60 A (Figure 12c). The resistivity $\rho_5$ of the target layer, labeled “Gas Hydrate” in Figure 11, is varied from in four steps from 3 $\Omega$m (no gas hydrate) to 100 $\Omega$m (high gas hydrate concentration).

Unlike the marine case, the responses instantaneously jump to a finite value because the upper half-space consists of highly resistive air, not conductive sea-water. The switch-on fields exhibit differences for the individual resistivity values starting at around 30 ms. The responses are clearly above the typical noise values for LOTEM data, which are $10^{-10}$ V/m$^2$ and $10^{-8}$ V/m for magnetic and electric components, respectively.

Figure 12: LOTEM responses for the model shown in Figure 11 for different resistivities $\rho_5$ of the hydrate layer.

There is an obvious difference between the $dH_z/dt$ and the electrical components. For $dH_z/dt$, the
curves differ significantly when \( \rho_5 \) is initially increased. The differences are small when \( \rho_5 \) is increased beyond 30 \( \Omega \)m. Due to the sensitivity to vertical current systems, the curves for higher resistivities can be discriminated using the electric field components. This means that the magnetic field components can be used to detect a general increase of the resistivity in the target layer, but the magnitude of the increase can not be resolved easily without electric field components.

Figure 13: LOTEM data with error bars for the model shown in Figure 11. This data is used for the equivalence analysis shown in Figure 14.

Switch on and switch off fields only differ by a constant, which is the respective field component at the direct current (DC) limit. The fields can be converted from switch on to switch off fields by \( \text{Field}(t)_{\text{on}} = \text{Field}_{\text{DC}} - \text{Field}(t)_{\text{off}} \). For \( \text{dH}/\text{dt} \) only the sign changes, as the time derivative of a constant is zero. Even though there is no fundamental difference in switch on and switch off fields, we found that using switch off fields for the electrical components increased the resolution of the target, in part because of the higher relative differences for different models (compare Figure 12b and 12c).

As discussed in the marine CSEM section, measurable differences do not necessarily mean that the target can be resolved uniquely. We use the same approach as for the CSEM data (non-linear inversion with random initial models) to investigate the resolution properties of the LOTEM data sets. Prior to inversion, Gaussian noise was added to the data with a standard deviation of a typical noise floor of either 0.5% or 10\(^{-10}\) V/m\(^2\) (for \( \text{dH}/\text{dt} \)) or 10\(^{-8}\) V/m (for \( \text{E}_x \)), whichever is larger. Error estimates are set accordingly. The data is shown in Figure 13.

The results of the calculations are shown in Figure 14. Again, the gray lines indicate the range in which the different parameters can be changed without degrading the data fit. The depth to the target layer or its thickness and resistivity varies so much that it is not clear whether a resistive layer has to exist at all. To clarify this, a second panel right of the depth/resistivity section shows a histogram of the resistivity at the target depth. The true value is marked with the black line. Note that the result of this histogram is biased towards the true value as most inversions were started with a \( \rho_5 \) value close to the true value. The number of models found thus is less relevant than if they found a model at all with a specific resistivity value.

The results for the \( \text{dH}/\text{dt} \) -component (Figure 14a) show the limitation of the magnetic field components to resolve resistive targets. In all three resistive zones, equivalent models were found with high resistivities of 500 \( \Omega \)m or more. The histogram shows that the range of resistivity values for the target layer (hatched) is between 3 \( \Omega \)m (no hydrate) to around 600 \( \Omega \)m. The boundary at 609 m, however, is well resolved.

The results for the LOTEM \( \text{E}_x \)-component are worse (Figure 14b). The few models that were found fitting the data, are highly ambiguous. This ambiguity can be reduced significantly, when both data sets are combined (Figure 14c). Here, the target layer is well resolved.
As indicated in Figure 10, the hydrate layer at Mallik was found to be subdivided into three zones with higher resistivity and thus higher hydrate concentrations [22]. As a test for the resolution capabilities of LOTEM with respect to these features within the hydrate zone, we recomputed data sets for a model where the target zone is subdivided into three layers (see black line in Figure 15). The results of the equivalence analysis using both the broadside $dH_z/dt$ and the in-line $E_x$ components are shown as gray lines in Figure 15. As initial models, the same sequence of random models was used as was used for Figure 14. The subdivision of the hydrate zone obviously is not sufficiently resolved by the surface data sets, because a number of models were found that fit the data within error estimates, without introducing a further subdivision of the target layer.

**CONCLUSIONS**

The calculations for the marine CSEM show that the data for receivers at multiple offsets can be used to resolve the vertical hydrate distribution. A vertical downhole TX can help to reduce the ambiguity of structures within the hydrate layer.

The LOTEM method seems to be suitable to map permafrost gas hydrates. Due to the greater depth
to the hydrate layer however, the resolution is limited. Resolving smaller scale structures such as subdivisions within the hydrate zone is difficult.

REFERENCES