In this article, we show that the controlled source electromagnetic (CSEM) method is complementary to the seismic method. Land and marine CSEM methods have developed almost independently. Both methods are discussed, but the focus is more on the application of the marine CSEM method, since this has had the most attention in the past decade. Active methods using man-made EM sources are used to investigate subsurface reservoirs and, in principle, are able to distinguish between those that are saturated with electrically resistive hydrocarbons and those that are saturated with electrically conductive brine. Therefore, they have the potential to rank the prospectivity of structures known from seismic data, but before drilling. Novel techniques for the processing of marine CSEM data include removal of the airwave that travels through the air at the speed of light and the suppression of magnetotelluric (MT) noise. For transient pseudorandom binary sequence (PRBS) data, deconvolution is an important part of signal-to-noise ratio enhancement. EM data have much lower resolution than seismic data and therefore need to use the subsurface structure obtained from seismic data plus rock physics relations to constrain resistivities in starting models for inversion.

INTRODUCTION
Rocks are composed of minerals that form a solid matrix containing pores. The fraction of rock volume occupied by pore space is the porosity $\phi$. The pores are full of fluids, sometimes including air. The solid matrix is normally extremely resistive, as there are very few charged objects or ions free to move and conduct electricity. The fluid in the pores, on the other hand, contains ions that can move in the fluid and therefore conduct electricity. Normally, the fluid is saltwater and the conductivity of the rock depends on the concentration of salt in the water, the fraction of the pore space that contains saltwater, and the freedom of movement of the ions between pores.

Sometimes the pores may also contain hydrocarbons as liquid, gas, or both. When hydrocarbons are present, there are generally three fluid phases: saltwater, hydrocarbon liquid, and hydrocarbon gas—normally methane. The hydrocarbons are not ionized and so they are not conductors of electricity. It follows that the presence...
of hydrocarbons increases the resistivity of the rock. The greater the fluid fraction, or saturation, of hydrocarbons, the greater is the resistivity of the rock. As shown in Figure 1, the effect of replacing saltwater by hydrocarbons can increase the resistivity by orders of magnitude, whereas the effect on seismic P-wave velocity is small. This is the reason the petroleum industry is becoming interested in EM methods in addition to very well-established seismic methods.

EM methods can be divided into passive and active methods. Passive methods use Earth’s naturally occurring EM field to investigate its interior. Active methods use a man-made EM source and one or more receivers to investigate Earth’s interior. This distinction is similar to that for seismic and acoustic methods: passive seismic and acoustic methods record and analyze events that create vibrations, such as naturally occurring earthquakes and rock bursts, man-made explosions, including underground nuclear tests, and movement of submarines. Active seismic methods are used in exploration especially in the search for potential traps for oil and gas.

Figure 2 illustrates the setup for conventional marine controlled source EM surveying (in [2]–[5]), which uses receiver nodes on the seafloor that measure all three components of both the electric and magnetic fields, and an electric dipole source, towed by a vessel, that transmits a continuous signal, typically a square wave or a similar periodic signal, with a spectrum containing discrete frequencies. The conventional marine CSEM method has been used mainly in deep water (deeper than 500 m), but in the last few years it has also been used in shallower water (e.g., [6]). If the source is on the x-axis, a receiver also on the x-axis is in line with the source and is known as the in-line component. The in-line component of the electric field is used to detect subsurface resistors.

An alternative CSEM technique, which has also been used on land, also uses a horizontal electric dipole source and dipole electric receivers, but the source signal is transient: after a certain amount of time, the earth response reaches a steady state. After a steady state has been reached, the cycle

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**Figure 1** Resistivity and P-wave velocity as a function of brine saturation for a porous sandstone. Figure redrawn from [1] and used with permission.

**Figure 2** The acquisition setup for the marine CSEM method with a towed electric dipole source and nodal seafloor receivers. Figure used with permission from [2].
may be repeated. We describe the transient approach in detail later. One type of transient response is the response to a reversal in polarity of a direct current (DC) current (e.g., [7]–[11]). Another type of transient response is the response to a single period of a PRBS, demonstrated by [12] for land data and by [13] for marine data.

Since most of the articles in this issue of IEEE Signal Processing Magazine are concerned with seismic data processing, we thought it would be helpful to present the fundamental principles of EM propagation in relation to seismic wave propagation. A plane wave analysis then leads to the concept of skin depth. In the conventional approach to the CSEM method, the source signal is continuous. Transient source signals have also been used and we discuss the issue of source control before discussing the multitransient EM method (MTEM). The discussion of CSEM data processing focuses on noise reduction and removal of the so-called airwave, although some CSEM experts prefer to interpret CSEM data without the use of airwave removal algorithms. The MT signal is a powerful source of noise for CSEM methods, especially in shallow water, and cultural noise associated with electrical equipment is especially important for the land CSEM method.

**FUNDAMENTAL PRINCIPLES OF EM PROPAGATION IN CONDUCTING MEDIA**

**SEISMIC AND EM PROPAGATION: WAVE EQUATION AND DIFFUSION EQUATION**

Seismic waves deform the media—fluids and solids—in which they propagate very slightly: the strains are usually much smaller than $10^{-6}$. After a seismic wave has passed a particular point in the medium, the medium at the point returns to its original state. There is no change in the medium. The energy associated with the wave is contained in the wave. No energy is left behind in the medium after the wave has passed. It follows that there can be no DC component to seismic wave propagation. (This statement does not necessarily apply in a region close to the source, where the displacement may be large, the medium may be permanently deformed, and Hooke’s law itself may not apply.) If the recorded seismic data contain DC, this must be a mistake introduced by the recording electronics and is removed from the data in processing before subsequent processing.

EM waves in conducting media—fluids and solids—have both electric and magnetic fields. The electric fields are associated with currents according to Ohm’s law, the currents generate magnetic fields according to the Biot-Savart law, and changes in the electric and magnetic fields are related by Faraday’s law of EM induction. Maxwell developed his famous theory of electromagnetism by starting with the experimental evidence presented by Faraday. The important point for the exploration geophysicist is that the electric and magnetic fields are related, and whenever there is current, as there must be in a conducting medium, there are losses. These losses are the principal difference between seismic and EM propagation. And because they are permanent, there must be a DC component to EM wave propagation.

Seismic waves propagate according to the wave equation that is derived from two more fundamental equations: Newton’s second law of mechanics (force equals mass times acceleration) and Hooke’s law of elasticity (stress is proportional to strain). In solids, there are two kinds of elastic waves: longitudinal, or P-waves, in which the particle vibration is parallel to the direction of wave propagation; and shear, or S-waves, in which the particle vibration is perpendicular to the direction of propagation. Fluids have no shear strength and therefore do not support shear waves. P-waves propagate in fluids and are known as acoustic waves. The wave equation in a fluid is

$$\nabla^2 p - \frac{1}{c^2} \frac{\partial^2 p}{\partial t^2} = 0, \quad (1)$$

where $p(x, y, z, t)$ is the pressure in the fluid at the point $(x, y, z)$ and time $t$, and $c$ is the velocity of propagation, given by

$$c = \sqrt{\frac{K}{\eta}}, \quad (2)$$

where $K$ is the bulk modulus of the fluid and $\eta$ is its density. For the simple case of a monopole source of acoustic energy in a fluid, the wave equation may be written as

$$\frac{\partial^2 (rp)}{\partial r^2} - \frac{1}{c^2} \frac{\partial^2 (rp)}{\partial t^2} = 0, \quad (3)$$

where $r$ is the distance from the source. The pressure function $p(r, t)$ is then

$$p(r, t) = \frac{1}{r} q \left( t - \frac{r}{c} \right), \quad (4)$$

in which $q(t)$ is the source time function with dimensions of pressure times distance. In exploration geophysics, the conventional units of $q(t)$ are bar-m (1 bar = $10^5$ Pa).

Figure 3(b) shows the response to a monopole source in an infinite fluid. The arrival time at the receiver is proportional to the distance $r$ from the source, while the amplitude decays as $1/r$, as expressed in (4). In all this, we have ignored seismic attenuation, because we are assuming the deformations obey Hooke’s law, which is correct to first order. There is some seismic attenuation, which is a second-order effect, and is caused by rock physics effects that we ignore in this analysis. Figure 3(c) shows the corresponding response to an impulsive electric dipole source in an infinite conducting medium. The peak of the response arrives at a time proportional to $r^2$, while the amplitude decays as $1/r^2$, which
is dramatically different from the purely geometrical decay in the acoustic case.

The electric field at the point \((x, y, z)\) and time \(t\) is the vector \(\mathbf{E}(x, y, z, t)\), which has units of \(\text{V m}^{-1}\) and obeys the following wave equation

\[
\nabla \times \nabla \times \mathbf{E} + \mu \varepsilon \frac{\partial^2 \mathbf{E}}{\partial t^2} + \mu \sigma \frac{\partial \mathbf{E}}{\partial t} = 0, \tag{5}
\]

in which \(\sigma\) is the electrical conductivity; \(\mu\) is the magnetic permeability, which, for most nonmagnetic materials, is normally taken to be equal to the magnetic permeability of free space \(\mu_0 = 4\pi \times 10^{-7} \, \text{H m}^{-1}\); and \(\varepsilon\) is the electrical permittivity, normally taken to be the permittivity of free space \(\varepsilon_0 = 8.85 \times 10^{-12} \, \text{F m}^{-1}\). Every component of the vector \(\mathbf{E}\) field interacts with every other component as the field propagates.

It is important to understand the difference between the EM propagation in air and in the conducting earth. Air is an extremely poor conductor of electricity (3 to \(8 \times 10^{-13} \, \text{S m}^{-1}\)). For geophysical purposes, we may regard the conductivity of air as essentially zero. The last term in (5) is then quite negligible and EM wave propagation in air is therefore described by

\[
\nabla \times \nabla \times \mathbf{E} + \frac{1}{c_0^2} \frac{\partial^2 \mathbf{E}}{\partial t^2} = 0, \tag{6}
\]

in which \(c_0 = 1/\sqrt{\mu_0 \varepsilon_0}\) is the velocity of light. That is, it is the familiar EM wave propagation with all waves throughout the EM spectrum traveling at the speed of light.

To tackle the problem of EM propagation in the conducting earth, we transform (5) to the frequency domain using the derivative theorem to yield

\[
\nabla \times \nabla \times \hat{\mathbf{E}} - \omega^2 \mu \varepsilon_0 \hat{\mathbf{E}} - i\omega \mu \sigma \hat{\mathbf{E}} = 0, \tag{7}
\]

where \(\hat{\mathbf{E}}\) is the Fourier transform of \(\mathbf{E}\) and \(\omega\) is angular frequency. For exploration to depths greater than a few meters, low frequencies must be used and \(\omega \ll \sigma / \varepsilon_0\). Equation (7) then reduces to

\[
\nabla \times \nabla \times \hat{\mathbf{E}} - i\omega \mu \sigma \hat{\mathbf{E}} = 0, \tag{8}
\]

and (5) becomes

\[
\nabla \times \nabla \times \mathbf{E} + \mu \sigma \frac{\partial \mathbf{E}}{\partial t} = 0. \tag{9}
\]

For uniform conducting media, where there are no free charges, (9) simplifies to the diffusion equation

\[
\nabla^2 \mathbf{E} - \mu \sigma \frac{\partial \mathbf{E}}{\partial t} = 0, \tag{10}
\]

in which the vector components are independent. The corresponding magnetic field intensity \(\mathbf{H}(x, y, z, t)\), with units \(\text{A m}^{-1}\), obeys a similar equation,

\[
\nabla^2 \mathbf{H} - \mu \sigma \frac{\partial \mathbf{H}}{\partial t} = 0, \tag{11}
\]

The electric and magnetic fields are related by Faraday’s law

\[
\nabla \times \mathbf{E} = -\frac{\partial \mathbf{H}}{\partial t}. \tag{12}
\]

**PLANE WAVE ANALYSIS AND SKIN DEPTH**

The responses shown in Figure 3(b) and (c) are very different. One important difference is that the shape of the wavelet is independent of the source-receiver distance for the seismic case but is very dependent on distance for the EM case. Much of seismic data processing relies on this invariance of the wavelet with distance. These processes are not applicable to the EM data. To understand the reason for the differences, it is convenient to simplify the propagation to one dimension.

We choose the \(z\)-direction as positive downward and restrict the discussion to waves which propagate vertically. That is, the pressure and the EM fields are functions only of \(z\) and \(t\). The wave equation for the pressure (1) can be written as

\[
\frac{\partial^2 p}{\partial z^2} - \frac{1}{c^2} \frac{\partial^2 p}{\partial t^2} = 0, \tag{13}
\]

while the diffusion equation for the magnetic field (11) can be written as

\[
\]
\[
\frac{\partial^2 H}{\partial z^2} - \mu_0 c^2 \frac{\partial H}{\partial t} = 0, \quad (14)
\]

Taking the Fourier transform of these two equations enables the fields to be expressed as functions of frequency \(\omega\), instead of time \(t\): \(\hat{p}(z, \omega)\) and \(\hat{H}(z, \omega)\). The two equations now become

\[
\frac{\partial^2 \hat{p}}{\partial z^2} + \frac{\omega^2}{c^2} \hat{p} = 0, \quad (15)
\]

\[
\frac{\partial^2 \hat{H}}{\partial z^2} + i \omega \mu_0 \sigma \hat{H} = 0. \quad (16)
\]

Comparing (15) and (16), we recognize that the quantity \(\omega/\mu_0 \sigma\) is velocity-squared, so \(\sqrt{\omega/\mu_0 \sigma}\) is the velocity of propagation of the electric field, which is frequency dependent and complex. Equation (15) has the well-known solution

\[
\hat{p}(z, \omega) = \hat{p}^+ \exp(i \omega z/c) + \hat{p}^- \exp(-i \omega z/c), \quad (17)
\]

which is a wave of amplitude \(\hat{p}^+\) propagating in the z-direction at velocity \(c\), and a wave of amplitude \(\hat{p}^-\) propagating in the negative z-direction at velocity \(c\). The amplitudes of these waves are \(\hat{p}^+\) and \(\hat{p}^-\) and are determined by the boundary conditions. Let’s now consider that there is only a downgoing wave; that is, \(\hat{p}^- = 0\), and

\[
\hat{p}(z, \omega) = \hat{p}^+ \exp(i \omega z/c). \quad (18)
\]

Applying exactly the same reasoning to (16), the downgoing wave solution for the magnetic field intensity is

\[
\hat{H}(z, \omega) = \hat{H}^+ \exp(\sqrt{\omega \mu_0 \sigma} z), \quad (19)
\]

which can be written as

\[
\hat{H}(z, \omega) = \hat{H}^+ \exp(-z \sqrt{\omega \mu_0 \sigma}/2) \exp(i \omega \mu_0 \sigma/2). \quad (20)
\]

This is a wave propagating in the positive z-direction with a frequency-dependent propagation velocity, but its amplitude decays exponentially with increasing \(z\). The amplitude decreases by a factor \(1/e\) for a propagation distance

\[
d = \sqrt{\frac{2}{\omega \mu_0 \sigma}}. \quad (21)
\]

This distance is known as the skin depth. Conductivity of seawater is about 3.2 Sm\(^{-1}\). Conventional CSEM surveys often use a source with fundamental frequency 0.25 Hz for which the skin depth in seawater is 563 m.

The physical significance of skin depth is that it is a serious impediment to the resolution of small targets at depth. The smaller and the deeper the target, the harder it is to detect and resolve. Doubling the thickness of the overburden reduces the highest detectable frequency by a factor of four. Normally, a feasibility study should be done to ascertain the detectability of the target. It is important that this analysis be done with three-dimensional (3-D) not one-dimensional (1-D) modeling to obtain reasonable estimates of the likely EM responses.

**THE EVOLUTION OF THE MARINE CSEM METHOD**

The conventional marine CSEM method evolved from the marine MT method. Both techniques make measurements of orthogonal components of the electric and magnetic fields using autonomous receiver nodes. The key difference is that the MT method [14] uses the naturally occurring source field generated by solar interaction with the earth’s magnetic field, while the conventional CSEM technique uses a man-made or “controlled” horizontal electric dipole source towed close (50 m) to the seafloor. The source transmits a waveform with desired spectral properties into the earth. A continuous square wave with a fundamental frequency of between 0.1 and 1 Hz is often used.

The use of a controlled source was motivated by the need to provide information about the subsurface at frequencies above those available to marine MT in deep water (that is, above about 0.01 Hz). Frequencies in the range 0.01–1 Hz are sensitive to variations in the top few kilometers of the subsurface where hydrocarbon accumulations are found but are also required to constrain near-surface resistivities for the inversion of the lower frequency MT data and deeper earth resistivity. Another advantage of using a controlled source is that a horizontal dipole produces fields both tangential and orthogonal to the target whereas MT fields are purely tangential (in a 1-D earth). The component of the horizontal dipole source that produces a vertical electric field normal to horizontal layer boundaries, however, is particularly sensitive to thin resistive layers characteristic of hydrocarbon reservoirs. The MT source field is sensitive to conductive targets but has poor sensitivity to resistors.

The Scripps Institution of Oceanography developed techniques and instrumentation for marine MT [15] in the 1960s and 1970s as academic tools for investigation of the lithosphere and mantle. The use of a controlled source with the existing MT receivers followed on from the early work in MT [16], [17]. The node-based CSEM system used today is relatively unchanged from the original Scripps system.

A detailed description of the node-based CSEM equipment including the first towed source that was developed at Cambridge is given in [18]. The source is a horizontal electric dipole with current electrodes separated by 100–400 m and towed in-line by the vessel at about 1.5 kn. The data acquisition setup is illustrated in Figure 2. The source used in deep water CSEM applications transmits a high voltage and low current signal down a cable from the vessel to a transformer at the source. This signal is then transformed to a low-voltage, high-current signal, which is rectified, polarity reversed at prechosen times, and transmitted between two electrodes with the return current through the water. The high-voltage transmission from the vessel to the source minimizes losses suffered in the cable. Due to the conductive nature of the seawater, it is not necessary for the source electrodes to be in contact with the seafloor to obtain good coupling. The source is generally towed approximately 50 m above the seafloor. This is close enough that the signal is not attenuated significantly in the water.
column and large enough to avoid small topography variation of
the seafloor.

An example of two different CSEM node receivers is shown in
Figure 4 [19]. The electric field is measured by a pair of
perpendicular dipoles each about
10 m in length and the magnetic
field is measured by three orthog-
onal magnetometers. A new com-
 pact CSEM receiver [Figure 4(b)]
has recently been developed pri-
 vately in collaboration with
Scripps [20] and uses capacitive electrodes and short electric field
receivers (<1 m) within the body of the receiver module, which
eliminates the need for extended electric field arms.

The commercialization of MT and CSEM methods for
hydrocarbon exploration was largely due to exploration mov-
ing into deep water where these techniques were traditionally
applied. The increased noise level and airwave problem, which
we discuss later, prevented these techniques from being suc-
 cessful in shallow water.

The first company to show interest in the CSEM method for
hydrocarbon exploration was Exxon [21]. The recent development
of node-based CSEM method for oil and gas exploration was a
result of work carried out by Statoil, a Norwegian oil company.
The first test survey took place offshore Angola in 2000 [4]. This
was carried out in 1,200 m of water with the target 1,100 m below
the seafloor using the academic equipment described above. From
this initial survey, commercial hardware was developed with three
companies offering node-based CSEM services. The commercial-
ization of the CSEM method coincided with the advent of deep
water exploration in the Gulf of Mexico and offshore West Africa.
With drilling moving into deep water and the associated higher
drilling costs, the CSEM tech-
nique has become a key tool in
derisking such wells [22].

A marine transient CSEM sys-
tem was developed at the
University of Toronto [23], [24]
and first applied to shallow subsea
investigation of hydrothermal structures; source-receiver separ-
ations, or offsets, were typically about 100 m and the source cur-
rent was about 3 A. The system was then applied to the detection
of gas hydrates with a source current of 50 A and offsets of up to
500 m [10], [25]. This system used a source and a two-channel
receiver cable attached to a single boat with both source and
receiver stationary during recording. The commercial develop-
ment of a static transient marine EM system with an ocean bot-
tom cable (OBC) receiver was developed from the land-based
multitransient EM method [11]. A source current of 800 A and up
to 30 in-line receivers each 200 m in length were used to record
the in-line electric field at offsets of up to 8 km [13]. A fully towed
streamer-based EM system is described in [26] and illustrated in
Figure 5. With a cable-based system, data are transmitted to the
acquisition boat in real time allowing real-time quality control.

In summary, there have been two separate marine CSEM
developments. The conventional marine CSEM method uses

![Diagram](image1.png)

**FIG4** (a) The conventional marine CSEM node receiver (modified from [19] and used with permission) and (b) a new compact receiver.
receiver nodes on the seafloor and a towed dipole electric source transmitting a continuous signal, as illustrated in Figure 2. The transient marine CSEM method uses receivers that are connected to a vessel, allowing real-time quality control and a dipole electric source that transmits a broad bandwidth signal with a beginning and an end. It is clear that source control and receiver deployment are separate issues. These two developments may not stay separate, however. In the following sections on data acquisition and processing, we maintain the separation but note that these techniques face the same problems in the subsequent data processing to attenuate MT noise and eliminate the airwave.

ACQUISITION AND PROCESSING OF CONVENTIONAL MARINE CONTROLLED SOURCE EM DATA

POSITIONING
The position of the source above the seafloor is performed through real-time winch control based on altimeter data received from the source. The source is generally towed at about 1.5 kn (compared with 4–5 kn for seismic data), which is the minimum speed to have some control of the deep-towed source. The lower the speed, the longer the available stack time window for a given travel distance. The low towing speed results in a small towing angle to the vertical, which reduces the positional uncertainty of the source. Towing slower than 1.5 kn would result in the source being adversely affected by cross currents. Uncertainties in source positioning are still considered the biggest source of error in CSEM data [27] and result in errors in source-receiver offset and azimuth. The source electrodes may not be in the same horizontal plane (tilt), and may not be in the vertical plane of the sail line (yaw). The tilt and yaw must be measured and included in the modeling. Errors associated with source electrode positioning are frequency dependent, as higher frequencies decay more frequently with offset than lower frequencies. The problems are greatest for high frequencies and short offsets. Short baseline acoustic positioning on both receiver electrodes and the source vessel are used in water depths of less than 3 km.

Positioning of receiver nodes is performed by acoustic surveying as the source vessel sails over the receivers, normally using two perpendicular lines. Seafloor orientation of the receivers is generally obtained from the data. Direct determination of the receiver orientation from gyroscopes is expensive with the additional drawback of added weight from the extra batteries required. Compasses are fitted to all receivers but are often inaccurate and require frequent recalibration [28]. Errors in positioning are currently the main barrier to improving the quality of CSEM data. Reference [29] quote a positional accuracy of 5%, which equates to a maximum signal-to-noise ratio of 26 dB regardless of the source current. For this reason, improvements in positioning accuracy must be achieved before more powerful sources are required.

CLOCK DRIFT
Timing on the source is controlled by the global positioning system (GPS) time from the boat. All the receivers contain internal clocks that are synchronized to GPS time before deployment and following recovery are again compared to GPS time. Drift can be on the order of a few milliseconds per day. Accurate phase requires that the source and receiver clocks are synchronized.

The present receiver clocks are accurate to within 10 ms over a period of a few days. By applying a temperature-dependent drift correction to the timing it is claimed that a phase accuracy of one degree is now possible [30], although the typical inversion misfit threshold for phase data is 5° [31]. The ability to use phase information in inversion is crucial in providing depth sensitivity.
seen that almost 80% of the acquired data is a combination of the two components, which can be exploited only through a full 3-D inversion.

Following acquisition, the receivers release their concrete base weight and float to the sea surface for recovery and data transfer. Data are acquired continuously from the moment the receiver is deployed; during periods when the source is either not active or a long way from the survey site, the receivers acquire MT data that can be utilized in inversion, provided the water depth is not so deep that it filters out the MT source signal.

**DETERMINATION OF AMPLITUDE AND PHASE**

CSEM data require little processing before inversion. Most of the steps simply correct the data for orientation and amplitude to make them comparable with the synthetic data generated during inversion.

The amplitude and phase of frequencies transmitted in a continuous source waveform are determined from time windows of the data known as “stack frames” [32]. A stack frame is defined as the length of time it takes the vessel to travel a distance of one source length. This is typically 120 s, but is restricted to consist of an integral number of fundamental source periods. The stack length can be increased to improve the signal-to-noise ratio, but this is limited by the towing speed. The amplitude and phase may be obtained simply through Fourier transform of the data within the stack window, or a best-fitting sinusoid in a least-squares sense at the various transmission frequencies may be found in the time domain and the amplitude and phase of this sinusoid then determined. Amplitude and phase as a function of offset are the input to inversion.

**SOURCE NORMALIZATION**

Measurement of the source waveform is made by a small data logger attached to the source. Source and receiver data for the same time window are transformed to the frequency domain, becoming complex. The complex receiver signal at a given frequency is then divided by the complex source signal at the same frequency, and finally the amplitude of the resulting complex number is scaled by the receiver electrode length to give an amplitude with units of V/Am². The source signal normally has its energy confined to a small number of discrete frequencies and the frequency domain division is performed only at those frequencies.

**DETERMINATION OF ROTATION ANGLE**

The first step in CSEM data processing is to determine the orientation of the receiver node on the seafloor. When the node lands on the seafloor its receivers are in an arbitrary position, with the Ex and Ey components orientated at unknown angles relative to the source line. In the in-line direction, the amplitude is a maximum; in the cross-line direction, it is a minimum. Reference [28] describes a method for determining the receiver orientation by rotating the Ex and Ey components to find maximum (in-line) and minimum (cross-line) directions. Over a 1-D earth with source and receiver in-line, the cross-line signal amplitude is zero. However, 2-D and 3-D effects introduce a cross-line component and noise is always present. In addition, if the source towing angle is more than a few degrees from in-line, it adversely affects the determination of the rotation angle. Another approach to determining the rotation angle is to find it with the inversion procedure itself; that is, allowing the rotation angle to be another unknown. Reference [33] describe a procedure to invert for the rotation angle and seafloor conductivity simultaneously, which appears to work on both in-line and out of line source-receiver geometries.

**SOURCE CONTROL IN CSEM METHODS**

A variety of periodic continuous signals has been developed for use in marine CSEM methods, including a square wave and various waveforms that concentrate the energy in selected frequencies, as discussed by [34].

An alternative CSEM technique, also discussed by [34], uses a horizontal electric dipole source and dipole electric receivers, but the source signal is transient: after a certain time, the earth response reaches a steady state. After the steady state has been reached, the cycle may be repeated. One type of transient response is the response to a reversal in polarity of a DC current; another type of transient response is the response to a single period of a PRBS. For both cases the full response is measured over the whole cycle time. Reference [12] concluded that using a PRBS for the source current signal rather than a step in DC current allowed data with better signal-to-noise ratio to be obtained in a given time.

Let the source current be \( I(t) \) and the resultant voltage at the receiver be \( V(t) \). Since Maxwell’s equations are linear, the response of the earth can be regarded as a causal linear filter with impulse response \( g(t) \) that depends on the position and...
direction of the injected current at the source and the position and orientation of the receiver electrodes. These three quantities are related by the convolution

\[ V(t) = \int_{0}^{\infty} g(\tau) I(t-\tau) d\tau. \quad (22) \]

The lower limit of the integral is zero because the earth is causal and cannot respond before there is an input. Since the flow of current in a conducting earth is a lossy process, the impulse response \( g(t) \) must decay to zero as \( t \to \infty \). Within the precision of the measurements, therefore,

\[ g(t) = 0, \text{ for } t > T_g, \quad (23) \]

where \( T_g \) is a time greater than which the response is too small to detect. That is, the earth impulse response \( g(t) \) is transient: it has a beginning and an end.

Given that \( g(t) \) is of finite duration, what is the best function for \( I(t) \)? This is the problem of source control. To obtain the response \( g(t) \), or its Fourier transform \( \hat{g}(f) \) without bias to any particular frequency, \( I(t) \) should be a function whose amplitude spectrum is constant over the known frequency range of interest.

Two time functions with flat amplitude spectra that are used extensively in the measurement of impulse responses are swept frequency sine waves (used in radar and exploration seismology in the Vibroseis technique) and PRBSs (used for many decades in electrical and electronic applications). The instantaneous power of a swept frequency signal is time variant, whereas the instantaneous power of a PRBS is constant for its duration. For Vibroseis, the implementation of PRBSs is difficult, because the inertia of the vibrating masses inhibits rapid reversal of the direction of motion. It is not difficult, however, to switch the direction of current flow in resistors, and PRBSs are therefore very suitable for EM applications.

A PRBS is a sequence of \( N = 2^n - 1 \) samples that switches from one level to the other at pseudorandom multiples of a basic time interval \( \Delta t \); \( n \) is known as the order of the sequence. The PRBS has an amplitude spectrum that is flat in the frequency interval

\[ \frac{1}{N\Delta t} \leq f \leq \frac{1}{2\Delta t}. \quad (24) \]

The recorded data need to be sampled at a rate that is greater than, or equal to \( 1/\Delta t \) to obtain the full benefit of the source spectrum.

Using a current signal \( I(t) \) such as a single period of a PRBS, of time duration \( T_s \), which has the full bandwidth of the impulse response \( g(t) \), (22) can be written as

\[ V(t) = \int_{0}^{T_s} g(\tau) I(t-\tau) d\tau. \quad (25) \]

This is a complete convolution of finite duration \( T_g + T_s \). The only issue remaining is to determine \( g(t) \), given \( I(t) \) and \( V(t) \).

\( I(t) \), of duration \( T_s \), must have the full bandwidth of the impulse response \( g(t) \). Because \( I(t) \) has finite duration \( T_s \), it is a transient source signal. The response \( V(t) \) must be measured for the minimum duration \( T_g + T_s \), beginning at the start time of the source signal. Because \( V(t) \) has a finite duration, with a known beginning and end, it is also a transient.

**TRANSIENT CSEM METHOD**

The transient EM method has been used for many years for mineral exploration, but has not yet become a standard tool for hydrocarbon exploration and production. A standard work on the theory of transient EM was written by [35] who summarized 20 years of work done in Russia and North America. Reference [7] presented the state of the art of long offset transient EM (LOTEM) surveying for land applications. The MTEM method was developed initially in the University of Edinburgh [9], and then by MTEM Limited and by PGS. It works both onshore and offshore.

In 1994 and 1996, a time-lapse transient EM data set was obtained along a line over a gas storage site at St. Illiers la Ville in France using a long-period square wave dipole current source, inline and cross-line electric dipole receivers, and horizontal loops to measure the rate of change of the vertical magnetic field [9], [36], as shown in Figure 7.

The transient responses obtained were essentially step responses. As described above, the propagation through the air travels at the speed of light, so the arrival of the step at the receiver is virtually instantaneous. The propagation through the earth is slower and frequency dependent, and arrives later. The step response and its time derivative, the impulse response, were modeled by [11] for a 1-D earth and a source-receiver separation of 1 km in a similar gas storage situation; the result is reproduced in Figure 8. The response without the resistive layer is shown in black. The response with the resistive layer is shown in red. The arrival of the step through the air is known as the *airwave*. When the step is differentiated it becomes an impulse, as shown in Figure 8(b). The impulse response of the earth, traveling more slowly, arrives after the impulsive airwave and is separated from it in time. There is a dramatic increase in amplitude when the resistive hydrocarbon layer is present.

Two different approaches were taken to the processing and inversion of the St. Illiers la Ville EM data. First, [36] used the classic modeling and inversion approach of [7], but with 3-D modeling. Second, [9] recovered the earth impulse responses from the data and processed them, taking special care to correct timing errors, to obtain common-offset stacks, similar to seismic data processing. A comparison of the two approaches to the data processing is shown in Figure 9. Figure 9(a) shows the data input to the 3-D inversion by [36], displayed as common-midpoint gathers; Figure 9(b) shows the same data.
as Figure 9(a), but after careful adjustment for timing errors, and displayed as common-offset gathers. The resistive gas reservoir can be seen clearly in Figure 9(b) at 4 ms on the vertical axis and between 3000 and 5000 m on the horizontal axis.

The MTEM method evolved from the result shown in Figure 9(b). The essence of the method is that both the voltage at the receiver $V_1(t)$ and the input current $I_1(t)$ are measured simultaneously and the earth impulse response is recovered from these two measurements by deconvolution. A diagrammatic plan view of one possible setup is shown in Figure 7(a).

A transient current, typically a step function or a finite-length signal such as a pseudorandom binary sequence (PRBS), is injected between two source electrodes and is measured and recorded. The time-varying voltage response between each pair of receiver electrodes is also measured simultaneously. If the response reaches a steady state before the next change in current is applied at the source, the full response has been measured and is the convolution

$$V(t) = \Delta x_x \Delta x_r I(t) \ast g(t) + n(t),$$  \hspace{1cm} (26)

in which $V(t)$ is the measured voltage response at the receiver, $I(t)$ is the measured input current applied at the source, the asterisk $\ast$ denotes convolution, $g(t)$ is the unknown earth impulse response, and $n(t)$ is uncorrelated noise; $\Delta x_x$ is the source dipole length, $\Delta x_r$ is the receiver dipole length, and $t$ is time. The source current and received voltage are recorded with identical devices, whose effects cancel in the subsequent deconvolution. The total circuit impedance is, in general, complex and therefore the source current is in general out of phase with the applied voltage. The response in (26) depends on the injected current, and that is what we measure: the complex impedance effects are automatically taken into account.

For the land case, a schematic cross section of the setup is shown in Figure 7(a), and for the marine case, one possible configuration is shown in Figure 5. In both of these cases, the source and receiver electrodes are in a straight line. Since it is known from (5) that the different components of the electric field interact with each other in propagation, it is clear that more components and more azimuths should be measured. Onshore, the electrode positions, or pegs, are known from surveying. Offshore, acoustic transponders are attached to the cable at the electrode positions and are positioned using a commercial underwater acoustic positioning system. The whole setup can be moved along to continue the line, very similar to the 2-D seismic reflection method. It is necessary to have offsets up to four times the depth of the target to resolve both its top and bottom. Normally, about 40 receiver channels of equal spacing are used. This choice is somewhat arbitrary, but it has been found to give good lateral resolution equal to about half the receiver spacing. For a target at 1 km depth, the receiver dipole length would be 100 m and the receiver spread would be 4 km long. Onshore, a roll-along system similar to the seismic reflection method is used. Two special features of the method are precise timing and real-time quality control.
DECONVOLUTION TO RECOVER THE IMPULSE RESPONSE

The convolution in (26) applies because the earth system is linear. In the frequency domain, the convolution becomes a multiplication, so deconvolution becomes a division. A typical land data example of the measured current input, measured voltage output at one receiver, and the result of deconvolution to obtain the earth impulse response for the source-receiver pair are shown in Figure 10. Deconvolution in the presence of noise is described in [11].

In land data, the impulse response exhibits an initial impulse at the time break (which is the “airwave”); this is followed by the earth impulse response. The noise consists of random noise, MT noise, and nonrandom cultural noise from power lines and railways that is normally orders of magnitude greater than the MT component. The noise can be reduced by a variety of processes, including stacking. The fundamental frequency and harmonics of the cultural noise are often not constant and the phase of each of these frequencies also varies in an apparently random way.

IMPULSE RESPONSE OF A HALF-SPACE

The response of a half space of resistivity $\rho$ ohm-m to a 1 A-m step applied at an electric dipole source on the surface was derived by [37]. The impulse response is obtained by differentiating the step response:

$$g(\rho, r, t) = \frac{-\rho}{8\pi \sqrt{\pi}} \exp \left(\frac{-r^2}{4c^2t}\right) \frac{r}{t} \Omega/m^2/s,$$

in which $r$ is source-receiver offset in meters, $c^2 = \rho/\mu$, with magnetic permeability $\mu = 4\pi \times 10^{-7}$ Hm$^{-1}$, and $t$ is time. This function has a peak at time

$$t_{\text{peak}} = \frac{\mu r^2}{10\rho} s.$$

Substituting $\tau = t/t_{\text{peak}}$ into the expression in (27) gives the result

$$g(\rho, r, \tau) = 5.65 \times 10^6 \frac{\rho^2}{r} \exp \left(-\frac{5}{2\tau}\right).$$

Equation (28) states that the peak of the earth impulse response arrives at a time proportional to $r^2$, while (29) shows the amplitude decays as $r^{-5}$, as mentioned above in connection with Figure 3(c). The timing and amplitude of the peak can be used to estimate subsurface resistivities [11], [38].

The function described by (29) has the same shape as the black curve in Figure 8(b). It is very similar in shape to the real earth impulse response of Figure 10(c) and thus gives an analytical approximation to the real data.

PROCESSING OF MARINE CSEM DATA

THE AIRWAVE

In shallow marine CSEM methods, the response at the receivers is complicated by the interaction of the seawater/air interface. A large amount of energy that arrives at the receivers propagates through the air at the speed of light. This energy is often referred to as the “airwave.” Reference [39] gives the following analytic frequency-domain expression for the simple case of a double half-space of air and water

$$\hat{E} = \hat{L} + \hat{D} + \hat{I}$$

where,

$$\hat{L} = \frac{M\rho}{2\pi r^3} \exp(-2kr)$$

and

$$\hat{D} = \frac{M\rho}{2\pi r^3} [(1 + kr)\exp(-2kr)]$$

Figure 8: (a) Step response of a 20 $\Omega$-m half-space at an offset of 1,000 m to a 1 A-m step at the source dipole (black curve), and with a 25-m-thick, 500-$\Omega$-m resistive layer at a depth of 500 m (red curve). (b) Normalized impulse response, for same configuration as (a) with normalization factor 3.433E +6. The black vertical arrow represents the pure inductive effect of the impulse at the source. Figure used with permission from [11].
In these equations, $M$ is the source dipole moment, with units A-m, $\rho$ is the water resistivity ($\Omega$-m), $z$ is the source and receiver depth below the sea surface (m), $r$ is the horizontal source-receiver offset (m), $R_1$ (m) is the distance between the receiver and image source given by 

$$R_1 = \sqrt{r^2 + (2z)^2},$$

$k$ is the wavenumber in the water (m$^{-1}$), given by

$$k = \sqrt{\omega \mu_0 \sigma},$$

where $\omega = 2\pi f$ and $f$ is frequency (Hz), $\mu_0$ is magnetic permeability, taken to be $4\pi \times 10^{-7}$ H/m, and $\sigma = 1/\rho$ is conductivity (S/m). The terms $L$, $D$, and $I$ in (30) are known as the primary airwave, the direct wave and the source image, respectively. These three components are illustrated in Figure 11.

The primary airwave travels vertically through the water and then through the air at the speed of light with only geometrical spreading of $1/r^3$. It is a function of seawater conductivity, source and receiver depth, and source-receiver offset only. When a layered earth is considered, a second component of the water layer response is introduced, which is coupled to every resistivity boundary within the subsurface. Reference [40] describe this energy as multiple reverberations at the source and receiver side between every resistivity boundary and the sea surface. This is analogous to multiple reflections in seismic reflection data, though in this EM case the process is diffusive in the water vertically above and below the source and receiver. The reverberations are most sensitive to the shallow resistivity structure and can be considerably larger in amplitude than the primary airwave when the seafloor is resistive.

An important part of CSEM data processing is the removal of the airwave because its presence significantly reduces

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**Figure 9** Data from the St. Illiers la Ville gas storage site. (a) Input to 3-D inversion by [36] and (b) the same data set as (a) after different processing to give the first derivative of the impulse response displayed as 1,000-m common-offset section [9]. Figure used with permission from [9].

**Figure 10** (a) Current measurement for a PRBS binary sequence source time function; (b) receiver response for the input signal shown in (a); and (c) result of deconvolution, showing the impulse response of the earth, plus noise, for three different values of the listening time. (Note that (c) is plotted on a time scale different from (a) and (b) to enable the features of the impulse response to be seen clearly.)
the sensitivity to subsea resistors in shallow water. The effect of a thin conducting layer above the source and receivers provides a short path from the source to the receivers vertically through the water and then through the air at the speed of light with only geometrical spreading of $1/r^3$ where $r$ is the horizontal source-receiver offset.

**The Airwave in Transient EM**

The impulse response of the earth recovered with a transient source exhibits partial temporal separation of the earth response and airwave in shallow water as shown in Figure 12. Reference [41] showed that there is an observation window in offset versus arrival time space which grows as the water depth decreases. As a result the airwave is not considered a great problem in shallow water for transient EM.

A technique to remove the airwave in the time domain was proposed by [42]. It requires a long offset measurement of the airwave on the order of 20 km, which is completely separated from the earth response and water layer response. This estimate of the airwave can then be used to estimate and remove the airwave in the received data. Limitations of this method are that it assumes that the water depth is the same for both measurements and that the near-surface resistivity structure is broadly the same in both locations to record the same multiple reflections.

Another approach to the elimination of the airwave for a transient source is to position the source and receiver close to the sea surface such that the measured airwave is the same as that measured on land; that is, the airwave becomes a delta function band limited only by the acquisition system. The water layer is then included as the first layer in the inversion and becomes part of the subsurface response. This approach is particularly well suited to a towed system and shallow water so that the resolution of the target is not adversely affected by the water layer.

**The Airwave and Conventional Continuous Source EM**

In the frequency domain, airwave contamination is recognized as a decrease in slope of the amplitude versus offset (AVO) curve to $1/r^3$ and a flattening of the phase versus offset (PVO) curve [27].

Many of the approaches to removing the airwave appear to provide an exact theoretical solution that is difficult to realize in practice. Receiver spacing in field data are not small enough to carry out up/down separation. Small deviation from the vertical when measuring $E_z$ results in the airwave being measured. Separation into TE and TM modes assumes a 1-D earth. Other techniques involve taking the difference of two large quantities to find a small quantity. When the airwave is a significant proportion of the total response, the required error on acquisition parameters is very small and the signal-to-noise ratio of the data must be very high. An important requirement of any technique is that it must be possible to model and invert the data after airwave removal. As a result, any technique that also removes the subsurface signal will make it difficult to accurately model the resultant response. For these reasons, the preferred approach of many in the CSEM community is to include the airwave in the inversion process. The effect of the airwave is greatly reduced at frequencies in the range 0.01–0.1 Hz compared with the frequency range 0.1–1 Hz commonly employed in deep water.

**Noise in Seafloor EM Measurements**

The noise floor of seafloor measurements is a crucial factor in marine EM surveying. The noise characteristics are quite different in deep and shallow water environments.
In water several kilometers deep, the main sources of noise are from receiver electronics, electrode noise, and water motion. The movement of electrically conductive water at velocity \( v \) through the earth’s magnetic field \( B \) due to tidal currents and wave motion induces an electric field \( E \) at the receivers given by

\[
E = v \times B,
\]

while microseisms, generated by sea surface gravity waves, displace the electrodes relative to each other at about 0.2 Hz [43]. There is currently no literature on the removal of induction noise; the effect of noise at a particular frequency such as that at 0.2 Hz caused by surface waves can be avoided by choosing source frequencies that avoid this frequency range. MT noise, which originates in the ionosphere, is low-pass filtered by the thick conductive seawater layer with little energy above 0.1 Hz reaching the deep seafloor.

In shallow water, the noise due to wave and tidal motion is far greater than in deep water and the amplitude of MT noise is considerably greater. Figure 13, modified from [44], shows the total noise (the lower solid line), including MT noise, measured in the Gulf of Mexico on the seafloor in 100 m water depth. The spectrum has a peak at about 0.1 Hz attributed to ocean swell noise in the shallow water. The background MT field can be seen to increase in amplitude towards low frequencies at approximately 20 dB per decade.

The noise floor of CSEM receivers is the voltage measured between two electrodes and divided by the electrode separation to give units of volts per meter (V/m). This measurement may also be expressed as the square root of the power spectral density V/m/\( \sqrt{\text{Hz}} \). The quoted instrumental noise floor for deep sea measurements is \( 10^{-3} \text{V/m}/\sqrt{\text{Hz}} \) [44] or \(-190 \text{ dB} \) in Figure 13 at 1 Hz.

A common practice in conventional CSEM literature is to quote a noise floor with units of V/Am\(^2\), where the noise voltage is normalized by both the receiver electrode separation and the source dipole moment (which is the product of the source length and the source current amplitude) as well as the bandwidth of the stacking window. A reduction in the noise floor quoted for conventional CSEM methods in the past decade has in fact largely been due to the increase in the source dipole moment; the true noise levels are unchanged.

For node-based CSEM methods in shallow water during periods of high MT activity, the data can be degraded so much that lines must be retowed. Reference [45] describes a remote reference technique for the removal of MT signals from seafloor nodes. The technique involves deploying two or three nodes outside the survey area and obtaining transfer functions between the remote stations and stations in the survey area during a period when the source is not transmitting. These transfer functions are a function of frequency, subsurface conductivity, and seawater conductivity but are independent of the MT signals. During periods when the source is active, these transfer functions can then be used to calculate the MT signals in data recorded in the survey area with the source signal present, based on the source-free remote measurement of the MT signal.

Reference [46] demonstrated a technique for removing MT noise without a remote reference measurement by exploiting the properties of the earth impulse response and the high spatial correlation of the MT signal. Figure 14(a) shows a raw common-source gather, 250 s in length (vertical axis), with offsets (horizontal axis) increasing from 2,200 m on the left to 7,000 m on the right. The long-period noise is well correlated from trace to trace; the response to the PRBS input decays dramatically from near to far offsets. Figure 14(b) shows the result of deconvolution in a 20 s window containing the impulse response: the long signal responses to the pseudorandom input current have been compressed to impulse responses, but the noise remains. An estimate of the noise is obtained by subtracting the short impulse response from the nearest 250 s trace. This noise estimate is similar to the noise on the other traces. To determine the component of the noise on each subsequent trace that is correlated with this noise estimate, a Wiener filter is found for each trace that best estimates the correlated part—the noise—from this noise estimate. The noise estimated in this way on each
subsequent trace is then subtracted from the trace to reveal the impulse response, as shown in Figure 14(c). The increase in signal-to-noise ratio from Figure 14(b) to (c) is about 20 dB. This is equivalent to increasing the source current from 700 A to 7,000 A.

Removing MT noise brings the noise level much closer to that of deep water EM receivers and significantly increases the operational capabilities of transient CSEM in shallow water. This technique attenuates all types of spatially correlated noise, not just MT noise, and so may help to reduce noise associated with water and wave motion.

MODELING AND INVERSION
One of the main barriers to the widespread use of EM in exploration has been the difficulty of making the data more understandable. Ray theory, which is so useful in understanding seismic data, does not apply to diffusive EM data. The analysis of CSEM data is carried out through iterative forward modeling or inversion using a model of the subsurface resistivity that generates synthetic data that fit the field data to within a specified misfit based on errors present in the data. The forward modeler may discretize the earth in 1-D (plane horizontal layers), 2-D (resistivity variation in the x-z plane), or 3-D, with resistivity variation allowed in all directions. The 2-D and 3-D forward modelers can discretize the subsurface using three different techniques: finite difference [47], finite element [48], and integral equation [49].

Inversion is very sensitive to the starting model; it pays to have the best possible starting model that is consistent with other known a priori information in the survey area, including especially seismic data and well logs. Since there are no common parameters in seismic and EM wave propagation, it is not straightforward to connect this information to a background resistivity model. Normally, rock physics is used to connect seismic velocities to resistivities via porosity, although there is no agreed procedure for doing this and the steps are not always well explained. For example, a 3-D resistivity forward model shown in [2, p. WA9] was “guided by 3-D seismic data, well log data...” In addition to the dimensionality of the earth, another complication (which is of particular importance in EM data) is the effect of resistivity anisotropy, which is seen in induction log measurements in deviated wells. The vertical and horizontal resistivities are different. Normally, the vertical resistivity is greater than the horizontal.

FOR THESE REASONS, THE PREFERRED APPROACH OF MANY IN THE CSEM COMMUNITY IS TO INCLUDE THE AIRWAVE IN THE INVERSION PROCESS.

Figure 15 from [50] shows the improvement that can be obtained by taking anisotropy into account in 3-D inversion. The isotropic inversion is sensitive to both vertical and horizontal resistivities and appears to include both these effects in the same image resulting in low resistivity features above and below the target due to low horizontal resistivity. Although this result used only in-line data and two frequencies, there is still a clear improvement when anisotropy is included.

CONCLUSIONS
The CSEM method of exploration has the potential to discriminate between more conducting brine-saturated and more resistive hydrocarbon-saturated reservoirs. EM data are, however, very different from seismic data and normal seismic data processing cannot be applied. In particular, the
EM wavelet shape varies as it propagates, high frequencies being attenuated faster than low frequencies, with the attenuation characterized by the well-known skin effect. For this reason, seismic stacking techniques cannot be applied.

The conventional marine CSEM method evolved from marine MT and uses very similar multicomponent receiver nodes on the seafloor. Traditionally, conventional CSEM methods use a dipole electric source that transmits a continuous signal with energy concentrated in a few discrete frequencies. The received data are divided into time windows of typically 120 s duration and transformed to the frequency domain for analysis.

Transient CSEM methods have had a separate evolution and use receivers that are connected to the vessel, permitting real-time quality control and data analysis. The source is also a current dipole, but the source signal is transient, having a beginning and an end, and normally containing a broad bandwidth of frequencies, enabling the complete impulse response of the earth to be recovered.

Presently, these two methods are separate. But since source control is unconnected with the measurement at the receiver, we see no technical reason to maintain the distinction between these techniques.

The CSEM method suffers from poor signal-to-noise ratio for targets deeper than about 2 km and a considerable effort is being devoted to noise reduction, especially for the airwave and MT noise. There is considerable scope for development. The three components of the electric field interact with each other during propagation, so it is clear that all three components should, therefore, be measured. It is known that 3-D effects are very important, so it is not sufficient to measure along a line.

At present, the interpretation product of the CSEM method is an inversion, which is essentially a resistivity model that yields synthetic data that match the real CSEM data within some error. The final model is obtained by iterative forward modeling, beginning with one or more starting models that are based on all available geophysical and well data and usually some rock physics. For geophysicists accustomed to seismic data acquisition and processing, it is very unsatisfactory. The positive conclusion is that there is plenty of scope for improvement. The potential of the method to identify hydrocarbons before drilling is a terrific incentive to develop new acquisition and processing methods to make CSEM become a reliable geophysical tool for hydrocarbon exploration, appraisal, and reservoir characterization.

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