Paper 11

Magnetotellurics: Status Quo and Quo Vadimus

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ABSTRACT

The natural source electromagnetic technique magnetotellurics (MT) is still a somewhat esoteric method, principally in the academic domain, that has not reached its enormous potential. It was proposed conceptually initially in Japan in the 1940s and then in France and Russia in the 1950s, and early developments were primarily related to theory and to hardware; the first commercial application of MT was for geothermal studies. Well into the 1970s the electric and magnetic time series were often digitized from paper records, processing was relatively unsophisticated using Fourier transform spectral methods, and interpretation was often based on one-dimensional approaches. From the early 1980s onwards there have been tremendous advances in instrumentation, processing, analysis, modelling and inversion to the extent that MT is now a relatively robust and established mapping tool used for imaging three-dimensional electrical conductivity variations from the near surface (100 m) to deep within the mantle (1,000 km), both on land and in marine environments. Diverse commercial applications include environmental geophysics (using very high frequency radio-MT, RMT), geothermal and mining studies (using high frequency audio-MT, AMT), hydrocarbon studies and deep crustal studies (using conventional MT), and lithospheric mantle studies (using long period MT, LMT). At this time three-dimensional inversion is being applied more and more routinely, even to profile data, thanks to codes generously made freely-available. However, appreciation and understanding of the inherent difficulties in 3D MT inversion is in its infancy, especially of how to treat galvanic distortions.

The three main limitations of MT currently limiting its broader adoption and application are (1) insufficient sensors, (2) inadequate methods for treating heavily noise contaminated data, and (3) inability to model Earth at the scale of conductivity variation. Advances on all three of these should occur over the next decade. For (1), we are typically spatially undersampling our targets of interest, and there really needs to be a huge reduction in cost of especially electric field electrodes, but also magnetic coils, and attendant recorders. For (2), MT is hampered in semi-urban environments, and often in rural environments, due to contamination from anthropogenic electromagnetic noise sources, such as AC power sources, DC trains, cow fences, pumps (especially DC), leaky transformers etc., and also some natural electromagnetic noise sources, such as frequent lightning strokes. Although MT time series processing advanced significantly in the 1980s with the implementation of robust processing schemes, these are inadequate when noise contamination dominates the time series. Finally, the lateral variations in conductivity can be at the electrode line-length scale, or even electrode scale, yet our current capability limits inversion in both lateral directions so our smallest cells are at larger scales. Multi-scale inversion methods need to be developed that mimic the physics of induction.

INTRODUCTION

Electromagnetic (EM) methods offer a number of advantages compared to all other geophysical methods; chiefly far superior resolution to potential field methods and far cheaper than land seismic methods. Of EM methods, the magnetotelluric method (MT) is in a special class that uses natural-occurring sources; lightning storms at frequencies >8 Hz and solar-terrestrial interactions at lower frequencies.

In MT we measure the time-varying horizontal components of the electric (approximated by the voltage differences between ground points) and magnetic fields (measured by a coil magnetometer at mining-scale frequencies). These time series are transformed into the frequency domain, and complex ratios of the electric to magnetic fields are computed. The magnitudes and phases of these ratios contain information about the vertical and lateral variations in electrical resistivity. MT has been, and is being, used for problems related to groundwater, mining, geothermal, and hydrocarbon exploration, and crustal and upper mantle imaging. There are over 3,500 papers published on MT in the academic literature (from Web of Science), and the number published per year is exponentially increasing (Figure 1).

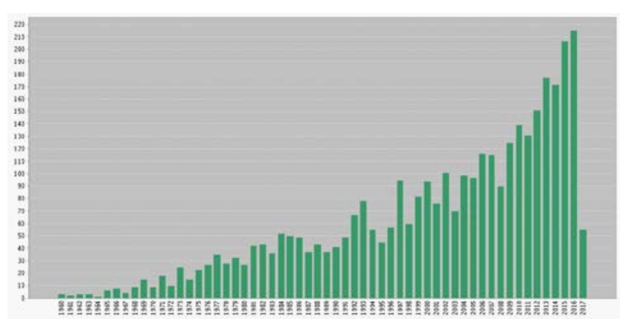

Figure 1: Publications on MT since 1960 from Web of Science (snapshot take in early-May, 2017).

This statistic does not capture the complete story – Google Scholar has almost 40,000 hits for the keyword "magnetotelluric*. Also there is a huge concomitant accelerated growth in use of MT in industry; the Society of Exploration Geophysicists (SEG) alone has over 2,350 articles on MT in its

In "Proceedings of Exploration 17: Sixth Decennial International Conference on Mineral Exploration" edited by V. Tschirhart and M.D. Thomas, 2017, p. 139–158

database – there are as many Technical Abstracts as there are papers published in its journal Geophysics.

Magnetotellurics does not have the resolution of seismology, as it is a diffusive technique and not a wave technique, but it is significantly superior in its resolving power to potential field (PF) methods, primarily as the physical property of interest varies by orders of magnitude. Also, in contrast to PF methods, for MT in one-dimensional (1-D) there exists a uniqueness theorem for continuous, error-free data (Bailey, 1970). Thus, non-uniqueness is inherently a consequence of data inadequacy and insufficiency, which drives us to collect more data with higher accuracy and precision at more and more sampling points. In terms of cost, land MT lies in between active seismic methods and potential field methods, and is comparable to that of passive seismic methods using ambient noise. Acquisition times are far shorter than passive seismic methods however.

The first commercial application of MT was for geothermal purposes in the United States in the early-1960s. Since that time MT has been used for a wide range of commercial investigations, including groundwater, geothermal energy (both high and low enthalpy), mineral exploration, hydrocarbon exploration, and diamond prospectivity mapping. Most recently it has been used for 4-D imaging for geothermal, carbon sequestration and shale gas reasons. The purpose of this paper is to outline the current state-of-the-art of land MT, particularly as applied for mineral exploration, and to present ideas for where advances need to be made to make the method even more worthwhile.

Brief EM Theory

Electromagnetic methods are sensitive to electrical parameters of the subsurface, namely dielectric permittivity ε and electrical conductivity σ. These two are components of the propagation constant *k* within a uniform medium whereby

$$\begin{aligned} k^2 &= i\,\omega\,\mu\sigma - \omega^2\,\mu\,\varepsilon \\ &= \omega\,\mu\,(i\sigma - \omega\,\varepsilon) \end{aligned}$$

where ω is the radian frequency of propagation, μ is the magnetic permeability of the medium, and *i* denotes imaginary unit. The second term in the RHS of the propagation constant equation is known as the Maxwellian term, as it was introduced by James Clark Maxwell [(1)] in 1855 to unify and codify the relationships between electric and magnetic fields in a medium. The relationship between electric and magnetic fields had interested and confounded scientists in the early part of the 19th Century, until Michael Faraday [(2)] in the U.K. in 1831, and independently Joseph Henry in France in 1832, appreciated that an electric field is generated, or "induced", if there is a time-change in the magnetic field, and that the amplitude of the induced electric field is proportional to the rate of change of the magnetic field.

At high frequencies >1 GHz used in GPR then ωε >> σ and dielectric permittivity is the dominant parameter as the Maxwellian displacement currents (second term in propagation constant equation) are greater by orders of magnitude than Ampere's conduction currents (first term). In contrast, at low frequencies <10 kHz then σ >> ωε for Earth materials and electrical conductivity dominates as conduction currents are far larger that displacement currents. There exists an overlap region between 10 kHz and 1 GHz where both phenomena must be considered, and the associated mathematics is less tractable than in either end-member regime where one or the other term can be neglected.

In the general case the solutions to Maxwell's Equations relating the electric and magnetic field behaviour are the standard Helmholtz wave equations. In geophysical exploration for mineral targets low frequencies <10 kHz are of primary interest given their greater depth of penetration, so we principally sense vertical and horizontal variations in electrical conductivity. At such low frequencies the governing Maxwell's equations relating the electric and magnetic fields can be reduced to diffusion equations, the same type of equations that describe, for example, the diffusion of thermal fields through a medium. As discussed by Chave et al. (2017), EM propagation at the microscopic level is a consequence of diffusion of charge-carrying microscopic particles that drift on the application of a source current, and microscopic and macroscopic diffusion were shown to be in direct relationship by Einstein (1905).

Electrical Conductivity of Minerals and Rocks

The electrical conductivity σ of natural rocks varies by many orders of magnitude, from tens to hundreds (10-100) Siemen per metre (S/m) for massive sulphides to one ten thousandth or less (<0.0001) S/m for competent igneous and metamorphic rocks. The bulk conductivity of rocks is usually dictated by the presence of a minor, partially interconnected, conducting phase; either a solid phase, such as sulphides (of interest in mining scale studies), or a liquid phase, such as volatiles or partial melt.

As most rocks have conductivities <1 S/m, it is more convenient to discuss their electrical resistivity, ρ [(3)], which is simply 1/conductivity. Thus, massive sulphides have a resistivity of one tenth to one hundredth (0.1-0.01) Ohm-metre (Ωm), whereas highly resistive rocks have resistivity >10,000 Ωm. A plot of the range of likely values of electrical resistivity of rocks is shown in Figure 1 (redrawn from Palacky (1987)).

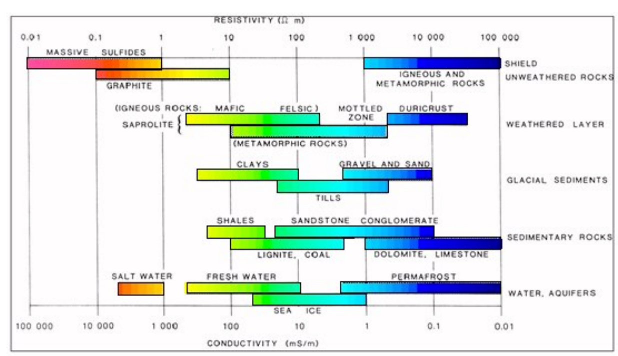

Figure 2: Variation of the electrical resistivity of rocks.

Figure 2 superbly illustrates why EM methods are so successful for mineral exploration for base and precious minerals. The difference in electrical resistivity between a conducting massive sulphide target and competent host rock can be many orders of magnitude. This compares to seismological properties, where massive sulphides can make for a decent acoustic target due to the high compressional velocity and density of pyrite, and the relatively slow but high density of pyrrhotite compared to mafic and felsic rocks (Salisbury et al., 1996; Salisbury et al., 2000). However, the reflection coefficient must exceed 0.06 to be detectable, which does occur and seismic reflection has had some success at imaging massive sulphides. However, the resistivity contrast is so great that it amply compensates for the diffusive nature of EM imaging over wave field nature of seismic reflection imaging. As Alan King said 10 years ago (King, 2007), "*This contrast is of the order of 8-9 orders of magnitude and makes measurement of electrical conductivity by far the most effective single tool in the identification of semi-massive to massive Ni-Cu-S's*."

Interconnection of Minerals

It must be remembered that for EM methods not only is the conductivity of the target important, but also the intrinsic connectivity. In the particular case of massive sulphides, the resistivity of pyrite can range from 0.00002 to 2 Ωm (Abraitis et al., 2004), and pyrite ore can range from 0.01 to 1000 Ωm (Pearce et al., 2006). Undeformed pyrite can often be in rocks as unconnected grains, thus the bulk conductivity is that of the lower Maxwell-Garnet/Hashin-Shrikman bound for a two-phase medium (Maxwell-Garnett, 1904; Hashin and Shtrikman, 1962) and the whole rock resistivity is only marginally reduced. An example of this is the Yellowknife Fault Zone (YFZ), known for its orogenic gold hosted in pyrite. The YFZ has a somewhat reduced resistivity compared to the surrounding crust, but is still resistive at 1,000 Ωm (Jones and Garcia, 2006). In contrast, when there has been significant deformation then the grains can become elongated and there is high interconnectivity, thus the bulk conductivity is that of the upper Maxwell-Garnet/Hashin-Shrikman bound. An example of this behaviour is the Bathurst No, 12 deposit, where the pyrite dominated massive sulphides are very conducting, with resistivity as low as 2 Ωm (Katsube et al., 1997; Queralt et al., 2007).

In contrast to the case for pyrite, the resistivity of pyrrhotite ore ranges from 0.001 – 0.1 Ωm (Pearce et al., 2006). This is because it forms a well interconnected network within the ore. Chalcopyrite lies between pyrite and pyrrhotite, with a resistivity range for chalcopyrite ore of 0.01-10 Ωm (Pearce et al., 2006).

Interconnectivity does not have to be 100%. The conducting material can be partially interconnected and still result in significant resistivity decrease.

Electrical Anisotropy

An unusual and somewhat exotic circumstance can occur when there is anisotropic deformation, as in such that the grains become elongated in one direction only. In this case then electrical current flows more easily along the strike direction than across it, thus electrical conductivity of the bulk rock is anisotropic. An example of this is the gold-bearing pyrite in Trans-Hudson Orogen sulphides, where studies on hand samples show that the pyrite grains are concentrated in the fold axes and connected along strike of the folds, leading to observable anisotropy at the hand sample scale, local scale, regional scale and even continental scale (Katsube, 1996; Jones et al., 1997). Samples from other mining camp localities also exhibit strong electrical anisotropy (Katsube et al., 1996; Connell et al., 2001). Vertical-to-horizontal electrical anisotropy also occurs in sediments due to compaction, with a vertical resistivity that is greater than horizontal resistivity by a factor of 2–10, but the MT method is intrinsically insensitive to such anisotropy in layered media (Yin and Weidelt, 1999).

There are other reasons for intrinsic anisotropic conductivity, such as the result of crystal growth or the difference in proton diffusion along the axes of olivine crystals. To complicate matters, the anisotropy direction may be oblique to regional geoelectrical strike.

Electrical anisotropy has not been that well studied to date, and we have few modelling/inversion tools for dealing with it. It is likely to be far more common than we have appreciated, certainly at the mining scale, and is one of the challenges for the future if we wish to accurately and precisely image the electrical resistivity distribution of mineral deposits to determine likely value.

Value Based on Conductivity

There are very few publicly reported studies on estimating ore grade, and hence deposit value, based on observed anomaly conductivity. One exception is McDowell et al. (2004) who outline Inco's positive experience for their implementation of a conductivity-based nickel grade estimation technique.

MAGNETOTELLURICS BACKGROUND

That there was a relation between natural magnetic field variations and electric field variations was known in the middle of the 19th Century, and Airy (1868) was the first to make simultaneous measurements of the time-varying terrestrial electric and magnetic fields; this he executed at the Royal Observatory, Greenwich. For the First International Polar Year 1882–83 the German team made both electric and magnetic measurements during their campaign at Kingua Fjord (now Clearwater Fjord), off Cumberland Sound, Baffin Island in northern Canada (Jones and Garland, 1986). In order to obtain measurable signal for the electric fields, the electrode lines were each of some 5 km in length. The processing and interpretation of these data by Jones and Garland (1986) represent the earliest MT responses ever derived and interpreted.

The theoretical basis for the magnetotelluric technique (MT) was proposed independently in Japan in the 1940s by Rikitake (1948), in France by Cagniard (1953), and in Russia by Tikhonov (1950) (Wait, 1962; Chave and Jones, 2012b). Of these, the paper by Cagniard (1953), a seismologist working for CGG and who gave the name "*magneto-telluric*" to the method, has rightly been the most cited and was the most influential of the three in the 1950s and 1960s as it described practical methods for the field implementation and interpretation of MT as an exploration tool. [(4)]

The basic concept in MT is that we measure the time variations of the electric and magnetic fields on Earth's surface, typically with orthogonal dipoles for the electric field and orthogonal magnetic sensors (coils for frequencies of interest in mining scale studies), and relate the orthogonal pairs in terms of the scaled magnitude squared of their ratio as an *apparent resistivity* ρ_a, given by

$$\rho_{a,xy}(\omega) = \frac{1}{\omega\mu_0}\left|\frac{E_x(\omega)}{H_y(\omega)}\right|^2$$

in the simple scalar formulation, where ω is radian frequency (cps), E_x (V/m) and H_y (A/m) are the electric and magnetic fields in the x and y directions respectively, and μ_0 is the magnetic permeability of free space. A similar equation exists for the relationship between E_y and H_x. This apparent resistivity is scaled so that it is equal to the true resistivity of a half space. There is also a phase, ϕ_{xy}, which is the phase lead of the E-field over the H-field, and the phase is 45° over a uniform half-space.

Apparent resistivity and phase curves are shown in Figure 3 for a 300 Ωm 3-km-thick layer over a conductive half-space of 3 Ωm, and over a resistive half-space of 30,000 Ωm. Note that for increasing apparent resistivity there is initially decreasing phase followed by increasing phase, and vice-versa for decreasing apparent resistivity. The change in phase slope occurs at the point of inflection in the apparent resistivity curve. This relationship between apparent resistivity and phase was later recognized as Hilbert Transformation.

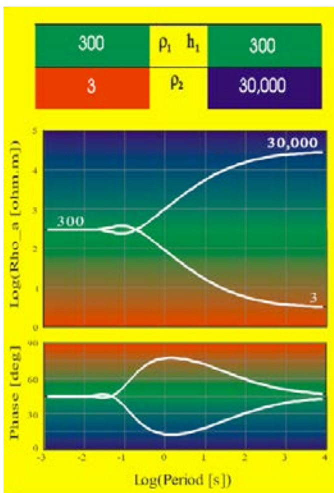

Figure 3: Apparent resistivity and phase curves for two layered Earth models.

There is interesting behaviour in the apparent resistivity curves in that they initially increase before decreasing, or decrease before increasing (see apparent resistivity curves in Figure 3, top). The cross-over to asymptotically increasing (for ρ_2 = 30,000 Ωm) or asymptotically decreasing (for ρ_2 = 3 Ωm) is approximately 5 Hz, which is a skin depth (see below) in the top layer of 3.9 km, approximately the thickness of the top layer. At frequencies higher than the skin depth to the subsurface layer or body, then the fields destructively interfere – this phenomenon is discussed by Jones (1986) in the context of so-called induction arrows—and thus the opposing increase or decrease prior to decrease or increase.

Following Cagniard's ideas, MT was initially performed in this scalar way, relating the electric field in one direction to the magnetic field in the perpendicular direction. Interpretation was in terms of 1-D layered Earths that were found through trial-and-error. Ted Madden and colleagues at the M.I.T. Geophysics Laboratory in particular made a number of significant advances in MT in the mid- to late-1950s, including studying 2-D behaviour of the fields (Neves, 1957) ([5]), introducing a tensor relationship between **E** and **H** [(6)] (Cantwell, 1960), and making the first repeatable and reliable MT measurements using then modern instrumentation and statistical analysis methods (Cantwell and Madden, 1960).

One issue with EM is that propagation into the ground is not simply arithmetic with decreasing period/increasing resistivity. We are often guided as a rule-of-thumb by the skin depth effect, which is the depth at which the EM fields are attenuated by *1/e*. The skin depth (δ, in m) in a uniform half-space is given by the square root of the resistivity (ρ in Ωm) of the half-space divided by the frequency (f, in Hz),

$$\delta = 503\sqrt{(\rho/f)} \quad .$$

However attenuation in the real Earth is very different from simple exponential decay even for a simple layered 1-D Earth. Also, the electric and magnetic fields attenuate differently in heterogeneous media; this is easily appreciated when one considers the end-member cases of a perfect conductor or a perfect resistor at depth. For a perfect conductor the horizontal electric field attenuates to zero, whereas for a perfect resistor it is the horizontal magnetic field that is attenuated to zero.

Figure 4 shows the *1/e* attenuation at 5 Hz for a 300 Ωm half-space (black), and for a 3-km-thick 300 Ωm layer over a conductive (red) and resistive (blue) half-space, where the E-field attenuation is shown in solid lines and the H-field attenuation in dashed lines. Note that in the top layer over a conducting half-space the E-field is attenuated far more severely than for a uniform half-space (solid red) and the H-field far less so. The attenuation into the conducting half-space is simply exponential. The opposite holds for a resistive half-space at depth (blue lines).

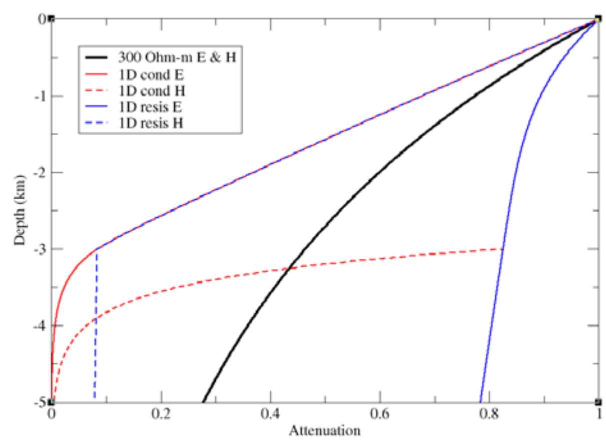

Figure 4: EM attenuation in a half-space (black) and for a conducting (red) and resistive (blue) lower half-space, with E-field attenuation in solid lines and H-field attenuation in dashed lines.

As discussed by Jones (2006), EM fields are "arrested" in conducting bodies and are "accelerated" in resistive ones. Assuming uniformly sampled estimates in frequency at 10 per decade, if we plot the data of Figure 2 in terms of approximate depth (Niblett-Bostick depth (Jones, 1983)) rather than period (Figure 5), we see that in the case of the layer over the conducting half-space (red points) there are far more points sampling the target depth region from 2 km to 4 km where there is significant resistivity change than for the data for a layer over a resistive half-space (blue points). For a conducting half-space there are 18 frequencies (almost 2 decades in frequency) sampling that critical depth range, whereas for the resistive half-space there are only 4 frequencies (less than ½ decade). This demonstrates one reason why in MT there is far poorer resolution of resistive structures than of conductive ones, simply because there are far fewer frequencies sampling them because of this "acceleration". Generally, in MT a major issue is that there is poor resolution of the region directly below a conductive region.

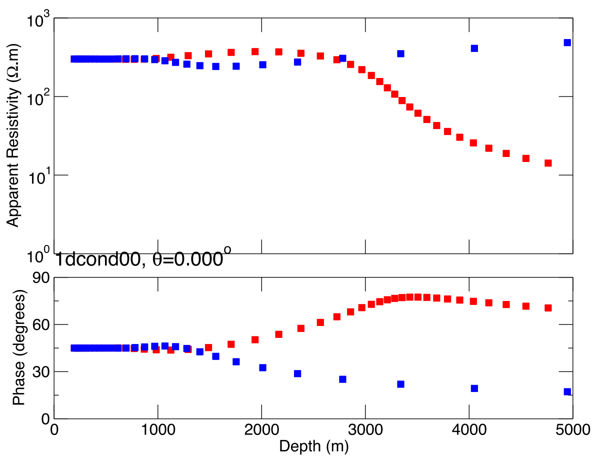

Figure 5: MT data of Figure 3 plotted as approximate depth

Magnetotellurics falls into broad "classes", depending on the depth of interest:

Radio-MT (RMT) – top 100+ m: refers to measurements at frequencies from 1 kHz to 1 MHz using military and civilian radio transmitters as the sources (Pedersen et al., 2006). It is applied for near-surface targets, such as contaminated vadose, buried waste, and archaeological sites, agricultural drainage, fracture zone, hydrogeology, and other targets. Given the nature of the sources the method has often been applied in scalar-MT mode with 1-D modelling as the output. Data acquisition takes less than a minute. Standard MT physics is adopted, i.e., displacement currents are ignored, which is questionable at very high frequencies.

Audio-MT (AMT) – top 2+ km: refers to measurements at the "audio" frequencies of 20 kHz to 8 Hz (the Schumann resonance frequency) where the signals are from global lightning storms (Garcia and Jones, 2002a). AMT is the method usually used for mineral exploration. Data acquisition usually takes around 15 minutes.

Broadband MT (BBMT) - crust: refers to measurements made from around 400 Hz to around 2,000 s. MT is used for geothermal and hydrocarbon exploration as the EM fields have less penetration at AMT frequencies in sedimentary basins. Data acquisition is usually overnight, so around a minimum of 12 hours. If high quality data are required in the MT deadband (see below), then acquisition may take two or even three nights.

Long-period MT (LMT) – upper mantle: refers to acquisition at periods greater than around 20 s. LMT is used for upper mantle penetration (Jones et al., 2009). One major difference to AMT and BBMT is that a different magnetic field sensor is used, namely a ring-core fluxgate magnetometer rather than a coil magnetometer. Acquisition is usually of the order of 2–4 weeks, typically a solar rotation cycle to catch all sunspots, due to the need for sufficient numbers of Fourier estimates for averaging.

Controlled-source AMT (CSAMT): refers to a hybrid AMT acquisition technique where a source, usually a grounded dipole, is used to generate the source signals instead of relying on natural sources. The responses are interpreted as if in the far field of the source, which cannot always be guaranteed, thus the method has all the disadvantages of controlled-source EM (CSEM) and not its main advantage. The method was proposed by Goldstein and Strangway (1975) as the instrumentation and processing methodologies of the day could not yield robust, repeatable AMT response estimates. Unless operating in a culturally noisy area or data in the AMT deadband (see below) are vitally important, AMT is superior to CSAMT for not only logistical cost but also superior modelling and inversion approaches.

In 2-D, where there is an anomaly that has a significant strike length, Maxwell's Equations decouple into two independent modes; one mode for currents travelling along the strike of the anomalies and in termed the TE (transverse electric) mode, and the other mode for currents travelling along the profile perpendicular to strike, and is termed the TM (transverse magnetic) mode. [7] Essentially the TE mode is primarily sensitive to the geometry of current flow in the subsurface, whereas the TM mode is primarily sensitive to the geometry of charge distribution in the subsurface. When undertaking 2-D inversion using both modes, one is formally performing a joint inversion as the two datasets a sensitive to independent information.

The relationship between the vertical magnetic field component (Hz) and the horizontal magnetic field components (Hx, Hy) is also often used in MT. This relationship has various names, but the most common, albeit incorrect, name used today is the "tipper". This name comes from the audio frequency magnetics (AFMAG) technique, and relates to how much of the horizontal field is "tipped" into the vertical direction. The tipper response is far superior at localization, as it reverses sign on either side of an anomaly (Jones, 1986; Jones and McNeice, 2002).

There are two frequency bands that present problems for AMT and BBMT, and these are named "*deadbands*". The AMT deadband is from approx. 1 kHz to 5 kHz, and is a consequence of the different lightning sources at AMT frequencies (Garcia and Jones, 2002a). The MT deadband is from approx. 10 Hz to 0.1 Hz, and again is the consequence of different MT sources, with lightning discharges at high frequencies and solar interactions at low frequencies. The MT deadband is, unfortunately, also the band for maximum ambient seismic noise from wind vibration coupling into the ground through tree roots, so special attention must be paid during fieldwork if the responses in this band are critical.

Although the application at AMT frequencies for direct detection in mineral exploration has received the greatest attention, it should be noted that where there exists conducting cover, such as the regolith in Australia, then the responses are driven to lower frequencies (Komenza et al., 2015) and acquisition through the MT deadband (10 Hz to 0.1 Hz) is required, which necessitates different acquisition methodologies and often different sensors.

Also, the location of mineralized bodies is dictated by regional structures in the crust and mantle (Begg et al., 2010a; Begg et al., 2010b; Griffin et al., 2013), so large-scale reconnaissance using deep-probing BBMT and/or LMT should be an essential component of any regional exploration programme in greenfields areas, especially those with cover (Dentith, 2016). This is the overarching philosophy behind the Australian Lithospheric Architecture Magnetotelluric Project (AusLAMP) (Duan et al., 2016; Robertson et al., 2016) that will, when completed, cover the whole of Australia on a regular grid with 55 km station spacing. In a similar way, EarthScope/USArray (USA) and SINOPROBE (China) have regional MT components that will result in continental lithospheric-scale information upon which regional and local targeted studies can be based.

An example within Canada of regional controls on local structures is the broad region of reduced resistivity observed in the crust and mantle directly below the YFZ that were taken as indicative of lithospheric-scale pathways for fluids that deposited the gold mineralization. A second example is the combined seismic/MT imaging of the Superior Boundary Zone (White et al., 1999) that showed crustal scale faults as pathways for mineralization.

In addition, long period MT data have been used for diamond prospectivity (Jones and Craven, 2004), principally to map the lithosphere-asthenosphere boundary.

In order to limit the scope of the paper to reasonable dimensions, the paper focusses on conventional land-based MT for mineral targets, and results obtained using CSAMT are not included. Also not included are results using airborne magnetics-only methods, such as AFMAG or ZTEM – these are discussed in a companion paper at Exploration 17 by Joel Jensen.

MAGNETOTELLURICS – STATUS QUO

Instrumentation

Instrumentation has come a long way since the early days, when every MT system was different, electronic equipment was individual and hand-built, and the time series fluctuations were recorded on paper charts and had to be manually digitized (this author's unfortunate history in the mid-1970s). Modern, commercially-manufactured 16-bit digital recording systems began to become available in the early-1980s, with Phoenix Geophysics's MT-16 being the first using a HP9845 for real-time processing and display.

High frequency data (384 Hz to 8 Hz) were acquired by the Phoenix system in narrow bands yielding two estimates per band, and this offered the attraction of dynamic acquisition focussing attention on those frequencies where superior estimates were required by performing a Singular Value Decomposition in real-time and inspecting the eigendata (**U**) matrix (Jones and Foster, 1986). Newer systems acquire data in broader frequency bands so this dynamic acquisition approach is no longer an attraction, but there are arguments that it should be re-investigated in order to ensure excellent data in the key frequency band, both in terms of quality and quantity (see Quo Vadimus Instrumentation section).

New recording systems have been 24-bit since the mid-1990s, with Phoenix Geophysics's MTU family of receivers and Metronix Geophysics's ADU-06 and now ADU-07, and Zonge's GDP receivers. Newer generation 32-bit receivers are becoming available, with Phoenix's MTU-7 family.

Coil magnetometers for sensing the magnetic field are conventional in design with windings around a core. There is a trade-off between high sensitivity due to a high number of windings and eddy currents induced in the core of the coil that introduce noise, so that it has been common to use one coil for AMT acquisition (shorter, with fewer turns) and a different coil for MT acquisition (longer, with more turns). This trade-off is shown in the comparison of the noise levels of coils shown in Figure 6 compiled by Ferguson (2012). There have been attempts to construct hybrid coils that cover the whole range of 20 kHz to DC, but in those cases compromises are made and the hybrid coils are "*Jack-of-all trades but master of none*". For mineral exploration purposes, then, as shown in Figure 5,

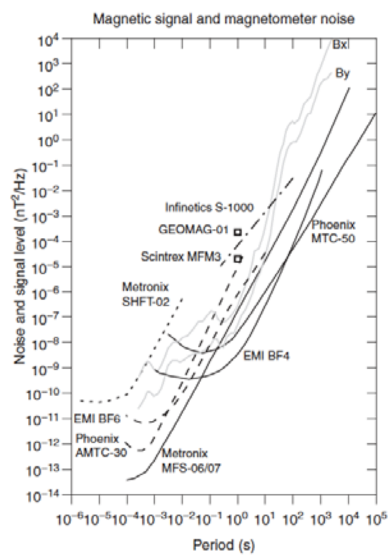

Figure 6: Relative noise level, sensitivity and phase response of some induction coil and fluxgate magnetometer sensors. Results are shown for BBMT induction coil sensors (solid lines), AMT induction coil sensors (long-dashed lines), RMT induction coil sensors (short-dashed line), air-loop induction coil sensors (double dot-dashed line) and fluxgate sensors (single dot-dashed line).

Phoenix's AMTC-30 and Metronix's MFS-06/07 are the best choices.

Appreciation of the noise introduced by polarization charges on metal electrodes resulted in the development and use of non-polarizing electrodes since the mid-1970s. Electrode design is still very primitive, and 19th Century physicists would recognize them. The most common metal and its salt electrodes in use are Pb-$PbCl_2$, with Cu-$CuSO_4$ as a close second. The best by far are Ag-AgCl, having the lowest potential difference between the metal and its salt resulting in lowest intrinsic noise level (see Figure 9.2 in Ferguson, 2012), but they are very expensive. For AMT measurements however, a staked metal electrode is sufficient as polarization charges are not of concern at AMT frequencies.

One issue with electrodes that is insufficiently appreciated is the effect of high electrode contact resistance on MT responses. Contact resistance should be small, typically <1 kΩ, especially compared to the input impedance of the receiver. A high electrode contact resistance that is of order of the input impedance of the receiver results in Earth acting as a low pass RC filter and attenuating high frequency signal. An example of this effect is shown in Figure 7; MT site *baf-001* was located on recently exposed rocks from the retreat of the Barnes Glacier on Baffin Island. Contact resistance was measured at 2 MΩ, which is of order that of the Phoenix MTU-5A input impedance (quoted as ">1 MΩ"). Attenuation of the electric fields at periods <0.1 s (frequencies >10 Hz) is obvious, with dropping apparent resistivities and phases driven to 0° (PhaXY, red) and 90° (PhaYX, blue). The nearby sites all had apparent resistivities of 10,000 Ωm at high frequencies. These high frequency data are completely unusable, meaning that we can get no information on the top 10-15 km of the crust below this site.

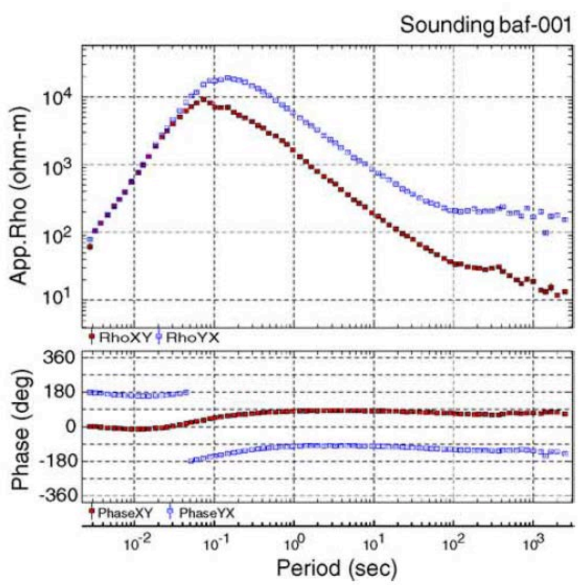

Figure 7: MT responses at site baf-001 of Evans et al. (2005) acquired on recently unglaciated terrane on Baffin Island.

For the e-lines themselves, there has been some discussion related to how to make e-field measurements (Wait, 1989; Wu and Thiel, 1989; Thiel, 2000). Certainly the advantages of using co-axial wire to shield the weak voltage signals from inductive effects has been well established for over three decades and co-ax cable should be the industry standard. [8]

For more detailed discussion and presentation on MT Instrumentation, please see the excellent chapter by Ferguson (2012) in the MT textbook of Chave and Jones (2012a).

Processing

Until the mid- to late-1980s, processing of MT time series followed conventional Fourier methods with estimation of spectral averages then determination of impedance tensor elements and their errors using standard parametric estimators (e.g., Sims et al., 1971). Estimates of impedances improved significantly with the introduction of the remote reference (RR) technique by Gamble et al. (1979) to remove auto-power biases (9), and remote-reference MT has become routine for acquisition.

These standard time series techniques were often found to fail, despite using RR. With the introduction of robust processing methods in the mid- to late-1980s far more stable estimates could be determined—see the comparison of five different conventional approaches against three robust approaches in Jones et al. (1989). The M-regression codes of Chave (Chave and Thomson, 1989; 2004) and of Egbert (Egbert and Booker, 1986; Egbert and Livelybrooks, 1996; Egbert, 1997), and the Least Trimmed Squares (LTS, Rousseeuw and Leroy (1987)) code of Jones (Jones and Jodicke, 1985; method 6 in Jones et al., 1989), are freely available on MTNet (www.mtnet.info), so there is little excuse for not undertaking robust processing of time series.

As important as deriving robust estimates of the response functions is deriving robust estimates of their errors—with poor errors then acceptable model space is poorly defined; it is either conservatively too large or overly-optimistically too small, with severe consequences for the veracity of the conclusions one can draw when modelling/inverting the responses. Jackknife error estimates are robust and make few assumptions about the nature of the signal and the noise (Thomson and Chave, 1991). In contrast, parametric estimators make ill-founded assumptions about the nature of the signal and the noise; independence, stationarity, ergodicity, Gaussian distributed, etc. A comparison between standard parametric error estimates and non-parametric error estimates consistently demonstrates that parametric estimates are far too low when the number of spectral estimates averaged together is high (Chave and Jones, 1997). This is particularly a problem in the MT deadband of 10 Hz to 0.1 Hz, where there may be thousands of spectral estimates so parametric estimators, which universally have a $1/n$ factor (where n is the assumed number of degrees of freedom), lead to ridiculously small error estimates that are inconsistent from frequency to frequency, i.e., there scatter is greater than their errors, and there is no possible Earth model that can fit the response estimates to within their estimated statistical errors. A routine approach to handle this problem is to adopt an *error floor*, which means if the errors are below a set value they are increased to that value, but if they are above it then they are unchanged. However, a far better approach is to derive robust and appropriate errors and be consistent and faithful to them. Error floors are though also adopted to deal with the problem of inadequate gridding of the subsurface.

Most processing codes do not consider the relationships between the series being processed; they could be processing a linear transfer function between apples and oranges. However, some have attempted to process data in a manner that honours the type of permissible response functions of Earth, the most notable being that of Larsen et al. (1996) who select time series sections that yield impedances consistent with Parker's D^+ algorithm (see below). Larsen et al. (1996) showed particular success at dealing with data from a heavily-contaminated culturally noisy location at a geothermal field in Italy (Larderello), admittedly with a remote reference station on an island off the coast. A robust M-regression algorithm that utilizes the Hilbert Transform relationship between the real and imaginary parts was published by Sutarno (2008), which was an advance on work presented 18 years previously (Sutarno and Vozoff, 1991).

A new approach by Chave (2014; 2016) formulates a maximum likelihood estimator (MLE) that exploits the stable nature of MT data, and its two-stage implementation first fits stable parameters to the data and then MT responses are solved for. The MLE is inherently robust, and differs from the conventional robust estimator as it is based on a statistical model derived from the data, whereas robust estimators are *ad hoc*, being based on a robust model that is inconsistent with actual data.

Analysis 1: Distortion Appraisal and Removal

Magnetotellurics underwent a huge leap through the 1980s and early-1990s when the effects of galvanic distortions began to be fully appreciated and understood, and methods presented for their removal. Such distortions occur naturally and are due to the perturbation of regional electric fields by very local, near surface inhomogeneities that are smaller than the scale length of the experiment so cannot be correctly modelled (Jones, 2012). In its simplest form, i.e. in isotropic or anisotropic 1-D or in 2-D in strike directions, distortion of the electric field creates a frequency-independent, multiplicative shift of the apparent resistivity curves either upwards (more rarely) or downwards (more commonly). This phenomenon was given the name *static shift*, after the term *statics* in seismology (Andrieux and Wightman, 1984, and a numerical explanation of its behaviour was presented by Jones (1988). Solutions for the 2-D problem of arbitrary local galvanic distortion were given by Bahr (1984; 1988), Zhang et al. (1987), Groom and Bailey (Bailey and Groom, 1987; Groom and Bailey, 1989), and Chave and Smith (1994), amongst others, and all are based on the concepts outlined in Richards et al. (1982) that advanced the 1-D approach of Larsen (1977) into 2-D. Of all of these, the Groom-Bailey approach has been the most successful and is the most used for removing the determinable parts of distortion, particularly the multi-site, multi-frequency version of McNeice and Jones (1996; 2001). Its advantages over all other techniques are discussed in Jones (2012), particularly that it treats the problem in a statistical manner by fitting a model of distortion to the responses and that it separates and removes the determinable from the indeterminable parts of distortion.

There is still a problem trying to remove galvanic distortion for a 3-D regional Earth prior to modelling/inversion. Attempts were made by Garcia and Jones (1999; 2002b), adapting the Groom-Bailey parameterization, and Utada and Munekane (2000), but these, whilst showing promise, were found not to be useful.

An alternative approach is to actually model the distortion as part of inversion. This was first undertaken in 2-D assuming simple static shift effects by deGroot-Hedlin (1991), and the approach was adopted in other codes (Ogawa and Uchida, 1996), notably that of Rodi and Mackie (2001) which is implemented in Schlumberger's WinGLink package and in CGG's Geotools package. Subsequently, deGroot-Hedlin (1995) inverted for Earth structure and the full 2x2 galvanic distortion matrix at each site. In 3-D, Sasaki (2004) and Sasaki and Meju (2006) adopted the 2-D concept of determining static shifts at each site. However, as Jones (2011) pointed out, this is erroneous in the general 3-D case due to phase mixing effects. The most promising approach to date is that of Avdeeva et al. (2015) who solve for the 3-D resistivity structure of the Earth as well as local galvanic distortion at each site. This approach has also been adopted by Kordy et al. (2016b).

Analysis 2: Response Appraisal

Having derived the optimum responses, that optionally in 1-D, anisotropic 1-D, and in 2-D are free of local distortion effects, the next step is to ensure that the responses are internally consistent. What this means is that the apparent resistivities and phases form a Hilbert Transform pair to within their errors (Jones, 1980). For all causal transfer functions, which means there is no output before there is any input (i.e., the impulse response function is zero for negative lags), the real and imaginary parts form a Hilbert Transform pair. This was discovered independently in different fields, and is known for example as Bode's Relations in servo-mechanics, the Kramers-Kronig relations in atomic scattering theory, and the Kertz operator in geomagnetism (Siebert and Kertz, 1957). See also Jones (1980).

The 1-D MT response function also has the property that it is a *minimum phase response function*, which means that the amplitude and phase of the response function must also form a Hilbert Transform pair (Weidelt, 1972). This property was also shown to be true of the TM mode in 2-D by Weidelt and Kaikkonen (1994). As well as satisfying this Hilbert Transform relationship, there are a series of properties of the 1-D MT impedance that must be obeyed, and these are listed in Weidelt (1972, Eqns. 2-30 - 2-34) and Jones (1980).

Formally, the 2-D TE response does not have this property, nor 3-D responses, and there exist extreme models that demonstrate the breakdown of this amplitude-phase relationship (Parker, 2010). However, in my experience it is exceedingly rare to find either real MT responses or synthetic ones that do not obey this amplitude-phase relationship, certainly to within data errors.

One that does, and an example is presented below, is cause for serious concern.

This Hilbert Transform real-imaginary relationship was exploited by Parker (1980) in his D^+ model solutions to the 1-D problem. These solutions provide the lowest possible misfit to a set of data, and take the form of infinitely thin conductive sheets separated by zero conductivity (infinite resistivity). This was extended to consideration of amplitude-phase consistency by Parker and Booker (1996) in their *Rho*$^+$ algorithm, which has proven to be very worthwhile for identifying inconsistent data. An example is shown below in Figure 8 from Jones and Garcia (2003) where the apparent resistivities (dots with error bars) are inconsistent with their prediction (dashed lines are 95% confidence intervals) using the phases.

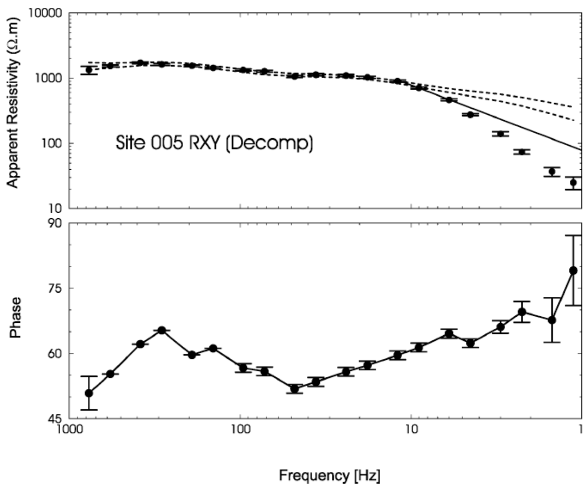

Figure 8: Derived apparent resistivities (top plot, filled circles with errors) and phases (bottom plot, filled circles with errors) of the XY mode from one site (005) on the Okak Bay survey (Jones and Garcia, 2003). The 95% confidence intervals of the of the prediction of the apparent resistivities from the phases are the dashed lines (top plot). The solid line from 10 Hz to 1 Hz on the apparent resistivity plot is the 45° descending line, which marks the steepest possible gradient for a 1-D or 2-D TM curve.

Modelling & Inversion

The MT technique has undergone incredible advances since those early days of first MT fieldwork in the late-1950s. The standard interpretation tool by most MT practitioners until the mid-1970s was 1-D forward modelling. Then 1-D inversion and 2-D forward modelling became the standard tools, and especially in 1-D a vast number of approaches were presented and codes became available. Although there were presentations of 2-D inversion approaches from the mid-1970s onwards (e.g., Jupp and Vozoff, 1977; Coen, 1981; Rodi and Minster, 1983; Zhang et al., 1986), either the approaches were of limited value or the codes were not made publicly available.

MT underwent a huge leap when 2-D inversion codes began to be made freely available, initiating with the *Occam2D* code of deGroot-Hedlin and Constable (1990) and quickly followed by *RRI* (Rapid Relaxation Inversion) of Smith and Booker (1991) and *REBOCC* (REduced Basis OCCam's Inversion) of Siripunvaraporn and Egbert (2000).

A similar history took place over the last decade with 3-D inversion. Although there have been presentations of approaches since the late-1980s (Madden and Mackie, 1989; Mackie and Madden, 1993; Newman and Alumbaugh, 2000; Tan et al., 2003; Sasaki, 2004; Avdeev and Avdeeva, 2009; Grayver, 2015; Kordy et al., 2016b), it is only since two codes have been made freely publicly available to academia that 3-D inversion by practitioners has become more routine. These two codes are *REBOCC3D* of Siripunvaraporn et al. (2005) and *ModEM* of Egbert et al. (2010) and Kelbert et al. (2014). Both of these codes have rectilinear meshes, which is a disadvantage when modelling topography and highly complex 3-D bodies.

It cannot be sufficiently stressed that 3-D inversion is computational time and memory intensive, which are its biggest challenge. However, there is also still much work to be done to understand how to implement these new 3-D inversion tools, and indeed, as demonstrated amply in 2-D, significant advances in understanding occur when codes are made available to the community and are widely used. Codes that remain proprietary serve neither the community nor the code writer well—most code writers are not practical geophysicists.

MAGNETOTELLURICS – MINERAL EXPLORATION APPLICATIONS

Audio-MT has been used for mineral exploration since its early days (Strangway et al., 1973) and has advanced significantly since Goldstein and Strangway (1975) found it necessary to use a grounded dipole source to achieve repeatable results for high frequencies, inventing the CSAMT method. Developments in instrumentation, particularly higher sensitivity coils and lower noise telluric electronics, developments in time series processing methods, and advances in our understanding of the nature of the source fields (Garcia and Jones, 2002a) have made the AMT method a viable choice for mineral exploration.

AMT is being used with increasing frequency for mineral exploration as the targets of interest these days are deeper than conventionally explored previously using CSEM methods. The commodities searched for have been diverse, from base to precious metals to uranium and diamonds, and all of these commodities have been part of AMT exploration programmes in Canada over the last 15–20 years. AMT has been used as a broad reconnaissance tool (Dentith et al., 2012; Dentith et al., 2013), and also for “sterilizing” areas, i.e. blanketing an area with AMT sites to ensure that there is no major deposit to be found prior to allowing the claim to lapse.

A huge amount of AMT work for mineral exploration has and is taking place, but comparatively little is reported in the public domain. Some of the more recent papers describing the results of MT studies over various commodities are:

Ni-Cu-PGE:

<u>Voisey’s Bay:</u> Balch et al. (1998); Zhang et al. (1998); Balch et al. (2000); Watts and Balch (2000)

Sudbury: Stevens and McNeice (1998); Zhang et al. (1998); Spicer (2016).
Thompson: King (2007)
Kevitsa, Finland: Le et al. (2016a)
Noril'sk: Varentsov et al. (2013)

Cu-Zn-Ag:
Kidd Creek: Gordon (2007)

Gold:
NICO deposit, NWT: Hayward et al. (2016)
B.C.: Hubert et al. (2016)
Carlin Trend: Le et al. (2016b)

Uranium:
Athabasca Basin: Craven et al. (2002; 2003); Leppin and Goldak (2005); Tuncer et al. (2006); Farquharson and Craven (2009); Goldak et al. (2010); Hautot et al. (2011)
NICO deposit, NWT: Hayward et al. (2016)
Cariewerloo Basin, South Australia: Crowe et al. (2013)

Diamonds:
Kimberlite pipe imaging: Pettit (2009); La Terra and Menezes (2012)
Diamonds – Regional Studies:
Northern Alberta: Turkoglu et al. (2009)
Sask Craton: Jones et al. (2005)
Slave Craton: Jones et al. (2003); Snyder et al. (2014)
Southern Africa regional study: Jones et al. (2009); Muller et al. (2009)

MAGNETOTELLURICS – QUO VADIMUS?

Magnetotellurics has come a long way since its first implementation 60 years ago in the early-1950s to the extent that it is now a reasonably reliable geophysical exploration method that produces robust results. However, there are still challenges to be faced if it is to become far more widespread and routinely used. Some of these I have alluded to above.

Survey Design

There is insufficient Survey Design in MT studies. Cost-effectiveness should be an important consideration when planning an MT survey. Whereas this is a far more important aspect for CSEM studies, to dictate optimum source-receiver separation, fields to acquire, station separation, and frequencies of importance (Maurer et al., 2000; Maurer et al., 2010), it is critical that in MT we have a good idea prior to a survey what the key frequencies are, what components need to be measured and what the site spacing should be. Acquiring data at sites where there is little of interest is wasted finances, and acquiring data at too few sites where high resolution is required is critical. Indeed, a primary question prior to undertaking an MT survey is whether the target can be imaged at all. Also, it is important to consider the EM fields separately rather than their ratios, as expressed as MT apparent resistivities and phases. This knowledge aids in ensuring the survey performed is optimal in terms of cost-effectiveness.

An example is the numerical examination by Jones and McNeice (2002) of a simplified body to represent the sub-vertical Kidd Creek deposit. In its simplest representation, so that we can examine the best-possible scenario, the deposit was modelled as thin (<100 m), conducting (10 Ωm in 10,000 Ωm host) at depths of 1.4 – 3.1 km, with a dip of 83°.

The TE 2-D response at a site at the surface on top of the anomaly is shown in terms of fields in Figure 8, where the top plot is the amplitude of the parallel to strike e-field ("offline E" for an acquisition profile aligned perfected perpendicular to strike) and the lower plot is the amplitude of the perpendicular to strike (i.e. parallel to profile) h-field ("inline H"). The solid lines are the responses without any overburden, and the dashed lines are the responses with 30 m of 25 Ωm (i.e. 1.25 S) overburden. Without overburden, the maximum response in the offline E component is a 28% reduction at 33 Hz, whereas the maximum response in the inline H component is a 74% enhancement at 12 Hz. Surprisingly, most of the anomalous MT response to the body comes from the magnetic field, not from the electric field. With overburden, these maximum responses are at lower frequencies, namely 15 Hz and 6 Hz respectively for offline E and inline H, and of lower amplitude.

The TM 2-D responses, i.e. inline E and offline H, are not shown, but the amplitude of the response is below noise level. (TM H = offline H, is uniform in 2-D.) Also not shown is the tipper response, but it has very high sensitivity to the anomaly and far superior location resolution than TE.

From this simulation one can draw the conclusion that it is essential to measure all 5 MT components at each site to fully sense the anomaly, and data should be acquired into the MT deadband down to 1 Hz for even moderate overburden. Site spacing conclusions can also be drawn.

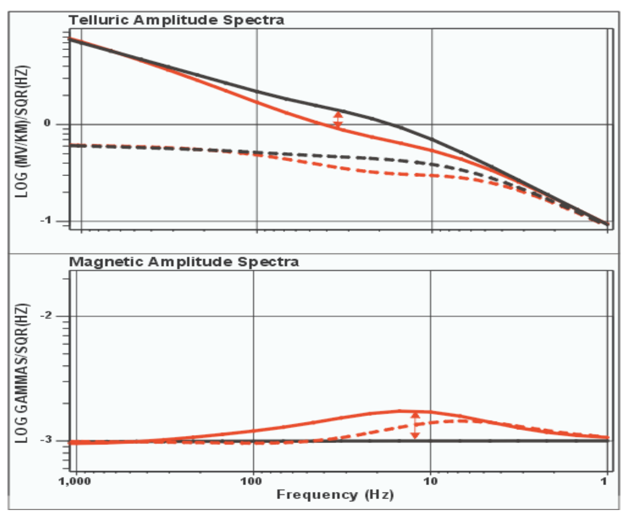

Figure 9: Parallel E-field (Ex) and perpendicular H-field (Hy) amplitudes at a site on top of the Kidd Creek deposit. Solid lines are without overburden, and dashed lines are with overburden. Black lines are the field amplitudes in the absence of the anomaly, and red lines are those with the anomaly.

Instrumentation

The biggest problem with the MT method for exploration is that it is under-capitalized. By that I mean that we do not have enough sensors on the ground so we often suffer from data insufficiency and data inadequacy. Compared to modern seismic reflection where typically some 80% or more of the data are discarded due to oversampling and redundancy, in MT we fight hard for every single bit and byte. Attempts were made in the late-1980s to introduce seismic acquisition technology into MT by AET with their EMAP system, and the network approaches in the 1990s with MIMDAS then in the 2000s with Titan-24 and Orion3D. However, as discussed by Jones and McNeice (2002) for mining scale targets, and by Jones (2006) for crustal scale targets, there exist classes of bodies that hardly respond in the 2-D TM-mode, which is primarily what is acquired by the current network-type systems. It is clear that optimal resolution comes from full 5-component (Ex, Ey, Hx, Hy, Hz) measurements at every site.

To facilitate high density full 5-component MT, we need far cheaper MT equipment and far more efficient and effective means of deployment, so that we can have far more sensors on the ground and redundancy in our data. Rapid AMT acquisition approaches are being explored (Cameron, 2002; Cameron and Wolfgram, 2016).

For e-field measurements, capacitive electrodes, or capacitively-coupled insulated long wire antennas (LWA), are worth considering for AMT acquisition, certainly at high frequencies (>100 Hz) and in some environments. However, the time to install conventional AMT rod electrodes is minimal—it is the time to lay out the e-line with orientation and length precision that takes the time for the e-field measurements. If the area of investigation is 1-D, then high precision in orientation is not required, and error in length simply appears as a static shift effect.

For the LWA design, as pointed out by Wait (1994; 1999) two decades ago, the input impedance of a staked antenna is much lower than the input impedance of the unstaked antenna; the issue related to input impedance of the MT receiver is discussed above. Indeed, electrodes on dry rock with a contact resistance of over 2 MΩ are, to all intents and purposes, ungrounded.

For the h-field measurements, it is time consuming to bury the coils, so tri-axial magnetometer coil supports, if shielded from wind vibration, may be attractive at frequencies above say 10 Hz.

Processing

Despite the superb advances of robust time series processing methods, developed initially in the 1980s, to their current state, as exemplified by the recent review paper of Chave (2016), there are still significant challenges related to signal-to-noise issues in AMT deadband (1kHz to 5 kHz) and the MT deadband (10 Hz to 0.1 Hz), to cultural noise contamination, especially cow-fence noise and DC noise from trains and mining infrastructure, and to continuous lightning strikes, amongst other problems.

Wavelet processing methods have been attempted (see overview in Escalas et al., 2013; and recent paper by Larnier et al., 2016), and were partially successful at recovering data through the AMT deadband (Garcia and Jones, 2008), but these approaches have not become routine. Most authors chose the Morlet wavelet, however there is no reason to assume that this is the appropriate natural wavelet to use for EM, and indeed the wavelet applied to the electric fields should be the time derivative of that applied to the magnetic field given Faraday's Law. However, wavelet methods do hold tantalising promise of extracting a very limited amount of signal buried in a lot of noise. It does though need to be appreciated that Fourier based approaches where the section length is 1 over the frequency of interest are "wavelet" methods that preserve the bias control that wavelet methods lack.

Other unconventional (for MT!) approaches are being advanced, such as Empirical Mode Decomposition (Chen et al., 2012; Neukirch and Garcia, 2014; Mehl, 2016), "morphological filtering" (Li et al., 2017a), and lightning stroke location processing (Hennessy and Macnae, 2015). Particularly the latter is showing excellent promise for obtaining estimates in the AMT deadband.

Finally for processing, all current processing codes currently derive estimates on a uniform sampling of log(frequency), typically 6–8 points per decade. As shown in Figure 5, uniform sampling in log(frequency) certainly does not translate into uniform sampling in depth, or even log(depth). There should be iterative communication between inversion and processing so that anomalous structures are sufficiently well sampled in frequency.

Modelling & Inversion

Computational capability has increased significantly over the last decades, and we can perform a 2-D forward model of a relatively large size (200 horizontal x 100 vertical) in fractions of a second on a laptop. However, 3-D forward modelling of relatively large size is still challenging, making inversion even more so. A 100 x 100 x 50 cell 3-D model took 2 days on a Sun Blade 1000 a decade and a half ago to derive the forward solutions at 11 frequencies (Queralt et al., 2007). Today that is reduced significantly by an order of magnitude, but nevertheless many 3-D inversions are performed on small sizes (50 x 50 x 30) to achieve results within a reasonable timeframe; not many of us have the luxury of access to very large high performance clusters (Newman, 2014). Within a decade we will be able to undertake 3-D forward models on the laptops of the day, making 3-D inversions as routine as 2-D one are today. But notwithstanding this computational difficult, we also still have a long way to go to understand the strengths and limitations of 3-D inversion—we just about understand 2-D inversion.

The greatest challenge to modelling Earth is that we are unable to model our data at the whole range of scale lengths of Earth. The smallest natural scale length of MT data is around each individual electrode, so is of order 1 m or less. Typical lengths that are considered are the electrode line length, which is typically 25-50-100 m for mineral exploration studies, and horizontal lengths of model cells are usually 1/3 of this length in the vicinity of the MT site. However, small-scale, local inhomogeneities exist at scale lengths smaller than the electrode line length, and those are typically absorbed into galvanic distortion effects. Adaptive meshing is essential, and multi-scale modelling approaches are being investigated and are showing promise (Caudillo-Mata et al., 2017).

In addition, all 3-D MT modelling codes assume a point E-field measurement, whereas in reality we make an approximate measurement of the E-field by determining the voltage difference between the two electrodes. This issue was considered in 2-D by Poll et al. (1989), who developed a code that determined voltages rather than E-fields. That code was employed by Jones (1988) to demonstrate the *static shift* effect in MT data.

Inversion codes need to use unstructured dynamically-adaptive meshes. Advances are occurring in this with Grayver (2015), Kordy et al. (2016a; b), and Li et al. (2017b), and a recent comparison of Finite Element (FE) and Finite Volume (FV) schemes, all using unstructured meshes, illustrates their advantages and disadvantages (Jahandari et al., 2017). However, currently all codes have a fixed solution mesh specified at the outset, and in many codes, certainly those widely available in greatest use, this is a rectilinear mesh. Whereas the initial mesh may be appropriate for the adopted uniform half-space start model, during inversion the model changes such that there may be far too many cells in a region, leading to inefficient forward calculations, or far too few cells, leading to inaccurate forward calculations.

Codes need to become widely available that actually mimic precisely, not approximately, the way that MT measurements are made in the field. And they need to become widely available—codes that remain proprietary do not significantly advance the method.

Our measure of the closeness of our model to the observations needs to be seriously addressed. In almost all cases, this closeness is simply measured by the summed misfit, a normalized root mean square (nRMS). As shown by Jones (1993), it is possible to have very dissimilar models that have the same nRMS to a set of data, but their fit based on other higher-order statistical measures are very different. Tarantola (2006) discusses this issue, highlighting that the supremacy of least-squares (L2 norm) over least-absolute-value (L1 norm) is entirely due to its computational simplicity, despite the greater robustness of the L1 norm. Avdeeva et al. (2016) proposed multiple hypothesis tests for verifying the veracity of the fit of (Gallardo et al., 2012) a model to data.

Resolution testing in MT, in even 2-D never mind 3-D, is in its infancy, and, since the advent in MT of Tikhonov-style regularization with most codes yielding Occam-style models that trade-off some measure of misfit (almost universally nRMS) against some measure of roughness (Constable et al., 1987), we have become rather complacent (read "lazy") and present a single model claiming that it has certain properties without fully considering defining the bounds of acceptable model space. We know that we are dealing with a problem that is highly non-linear, and that linearized resolution estimates (e.g., Schwalenberg et al., 2002) can be wildly in error, especially of resistive regions (e.g., Ledo et al., 2004). "Feature testing" is our current best approach (e.g., Solon et al., 2005).

As expounded by Tarantola (2004) we should approach the inverse problem as a statistical one and aim to define the posterior PDF for the resistivity in each cell. 1-D statistical approaches have been used for four decades (Jones and Hutton, 1979). For 3-D this does require very fast forward codes and thousands of CPUs so that stochastic methods can be applied in reasonable timeframes.

Joint inversion of "MT with something" is showing remarkable results (e.g., Moorkamp et al., 2011; Takougang et al., 2015), sometimes with some form of coupling between the parameters, such as cross-gradient constraints (Gallardo and Meju, 2007) or structural constraints (Gallardo et al., 2012). Looser "mutual information" coupling used in imaging approaches may prove to have significant advantages as there are far less assumptions being taken (Mandolesi and Jones, 2014).

Finally, following Popper (1959) we should always remember that we are on safest ground when we undertake falsifiability, i.e., when we reject hypothesis. We can never prove anything, but we can disprove a significant amount. Models that do not fit the data are more powerful and contain more robust information than those that do.

CONCLUSIONS

MT has come a long way since its inception, and, as exemplified by the exponential increase in the number of publications on or using MT (Figure 1), advances are coming along on many fronts at a rapid rate. Those who may have been discouraged by the results from an MT survey undertaken a decade ago should give the method another opportunity to prove its worth. It is not a panacea, and should not be applied to solve all problems (*"If you only own a hammer, all problems become nails!"*), but it certainly should be one of the weapons in the exploration geophysicist's arsenal.

ACKNOWLEDGEMENTS

I would like to thank the organizers of Exploration '17 for the invitation to present this review paper. Discussions with many colleagues over the years are represented in it, and comments on an earlier version from Anna Avdeeva, Alan Chave, Shane Evans and Jessica Spratt are very much appreciated.

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[1] As an interesting and ironic historical footnote, Maxwell applied for the Chair of Natural Philosophy at the University of Edinburgh in 1860, where he was born and had earned his BSc, losing out to Peter Tait, who very appropriately studied, amongst other things, the trajectory of golf balls. The irony is that the Physics Department at Edinburgh is housed in the James Clark Maxwell Building – they named their building after him but didn't give him the Chair. The Geophysics Department at Edinburgh, where I earned my PhD, was also housed in JCMB from the mid-1970s until it merged with Geology in 1989.

[2] Humorously, when Michael Faraday was reporting on his experiments on electricity and magnetism, Mr. Gladstone, then Chancellor of the Exchequer, interrupted him to put the impatient inquiry: '*But, after all, what use is it?*' Like a flash of lightning came the response: *'Why, sir, there is every probability that you will soon be able to tax it!*' And how true that has become.

[3] It is rather unfortunate that the symbol used for electrical resistivity, the Greek letter rho ρ, is the same one used for density. This easily leads to confusion when reading geophysical papers and especially during multi-disciplinary presentations.

[4] I recommend to everyone to read the very interesting discourse between Louis Cagniard and James Wait that is in the appendix to Wait (1954). Wait's opposition was primarily based on the uniform field assumption made by Cagniard, which is a valid concern at longer periods close to the equatorial or auroral zones but not at all an issue for mining scale studies.

[5] For a truly 2D Earth, then Maxwell's Equations decouple into two independent sets. One set contains the Ex, Hy and Hz components (assuming the x-direction is along strike), and this is called the TE-mode (Transverse Electric, meaning transverse to the profile). The other set contains Hx, Ey and Ez components, and is the TM-mode (Transverse Magnetic). Older literature used terms such as E-parallel or E-polarization for the TE-mode, and H-parallel, H-polarization, or E-perpendicular for the TM-mode.

[6] The tensor introduced in Cantwell (1960) was an Admittance Tensor, **Y = H/E**. The more common Impedance Tensor, **Z = E/H**, was apparently introduced first by Pokityanski (1961, see Wait, 1962).

[7] In older literature, you will find the TE-mode referred to as E-parallel, as in parallel to structure, or E-polarization, and the TM-mode as H- or B-parallel or E-perpendicular or H- or B-polarization.

[8] The outer shielding should be grounded at the ground point of the receiver, and left open at the electrode end.

[9] In the remote-reference technique, espoused in MT initially by Gamble et al. (1979), biases in response estimates caused by auto-power spectra are avoided through using time series recorded at a site some distance from the main site. Thus, only cross-spectra are involved in response and error estimation. The approach was first proposed in economic theory by Gini (1921, reviewed in Reiersol, 1950).