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Widmung

Wir wollen unser Kolloquium Ulrich Schmucker (1930 -2008) und Peter Weidelt (1938 – 2009) widmen und ihm den Namen „Schmucker-Weidelt-Kolloquium für Elektromagnetische Tiefenforschung“ geben.

Unser Kolloquium soll Gespräche zwischen allen fördern, die mit elektromagnetischen Methoden das Innere der Erde erforschen wollen.

Ulrich Schmucker und Peter Weidelt haben in diesem Kolloquium seit seiner Gründung neue und anspruchsvolle Methoden entwickelt, dort vorgestellt, oft sogar nur in diesem Kolloquium. Sie haben damit den internationalen Ruf dieses Kolloquiums begründet und das hohe intellektuelle Niveau dieser Forschungsrichtung weltweit bestimmt.

Wenn wir dieses Kolloquium, das eigentlich auch ungenannt schon immer das Schmucker-Weidelt-Kolloquium war, nun auch so nennen wollen, dann hat das drei tiefe Gründe:

Es sollen diese beiden Wurzeln unseres Kolloquiums und unserer Forschungsrichtung im Gedächtnis bleiben.

Das Kolloquium soll uns Teilnehmer erinnern, auch in Zukunft die methodenorientierte Grundlagenforschung und ihre praktische Erprobung auf hohem Niveau als Ideal und Ziel dieses Kolloquiums anzustreben.

Eine besondere Botschaft von Ulrich Schmucker und Peter Weidelt war es, gerade junge Studierende der elektromagnetischen Tiefenforschung von Anfang an in einen ernsthaften und lebhaften Kontakt mit allen Teilnehmern einzubinden, ihnen zuzuhören und zu helfen, sie zu kritisieren und Neues von ihnen zu lernen. Wenn wir diese Botschaft auch in Zukunft zum Arbeitsstil unseres Kolloquiums machen werden, wird die Entwicklung der elektromagnetischen Tiefenforschung so voller Dynamik bleiben, wie sie es bisher war.

I

Vorwort



Mit Erscheinen dieses Bandes und im Einvernehmen mit den Hinterbliebenen wollen wir von nun an das Kolloquium Prof. Ulrich Schmucker (1930 -2008) und Prof. Peter Weidelt (1938 – 2009) widmen. Den Wortlaut der Widmung haben wir an den Beginn des Bandes gestellt. Beide waren für die Entwicklung der elektromagnetischen Tiefenforschung weltweit von herausragender Bedeutung. Ihre Persönlichkeit und Schaffenskraft hatten aber vor allem in Deutschland einen maßgeblichen Anteil an der überaus positiven Entwicklung, die die Elektromagnetische Tiefenforschung genommen hat.

Veranstalter des 23. Kolloquiums war das Deutsche GeoForschungsZentrum - GFZ. Der Tagungsort, die Heimvolkshochschule am Seddiner See, befand sich in unmittelbarer Nähe zur brandenburgischen Landeshauptstadt Potsdam, direkt am Großen Seddiner See, in einer für Brandenburg typischen Umgebung mit vielen Seen, Laub- und Kiefernwäldern. Am Kolloquium hatten sich 88 Teilnehmer angemeldet, größtenteils aus dem deutschsprachigen Raum, aber wie auch schon in den Jahren zuvor gab es auch diesmal eine ganze Reihe von Anmeldungen aus dem europäischen Ausland. Insgesamt waren 15 deutsche und 8 europäische Institutionen vertreten.

Im Gedenken an Ulrich Schmucker hatten seine ehemaligen Schüler bereits ein Kolloquium vom 26 - 28 Juni 2009 im Herz-Jesu Kloster in Neustadt ausgerichtet. Peter Weidelt wurde in einer Reihe von Beiträgen im Rahmen des 23. Kolloquiums gedacht und besonders gefreut hat uns, dass auch Heidi, Anne und Jan Weidelt anwesend waren. Im vorliegenden Tagungsband sind die meisten dieser Beiträge protokolliert. Allerdings haben wir uns entschlossen, sie nicht besonders hervorzuheben, sondern sie wie immer thematisch einzuordnen.

Ulrich Schmucker und Peter Weidelt hatten von Beginn an in diesem Kolloquiumsband publiziert, oft sogar nur in diesem. Der diesjährige Band enthält 32 Beiträge. Es gab die übliche große Bandbreite an

Beiträgen die theoretisch- methodische Aspekte und experimentelle Untersuchungen im Hinblick auf geodynamische und angewandte Fragestellungen abdeckten. Generell war vielleicht eine Zunahme der mehr angewandten Arbeiten festzustellen, was sicherlich auch darauf zurückzuführen ist, dass die elektromagnetischen Verfahren mittlerweile fester Bestandteil der Exploration im off-shore aber auch on-shore Bereich geworden sind.

Quicklebendige Debatten gab es auch im Rahmen unserer neuen Rubrik „Was Sie schon immer über die Elektromagnetik wissen wollten, sich bisher aber nicht zu fragen trauten.“, die Ute Weckmann angeregt hatte. Insbesondere sollten damit die jüngeren Teilnehmer angesprochen werden, jedoch hatten offensichtlich besonders viele der Ü40 Teilnehmer noch jahrelang unbeantwortet gebliebene Fragen. Manchmal war für die Beantwortung der Fragen auch etwas mehr Zeit zum Nachdenken nötig, weshalb wir eine nach dem Kolloquium entstandene Antwort in diesen Band ebenfalls aufgenommen haben.

Für die finanzielle Unterstützung danken wir auch unseren Sponsoren: den Firmen KMS und Metronix. Herzlich bedanken möchten wir uns bei Frau Hannelore Gendt (GFZ) für die Hilfe bei administrativen und finanziellen Aufgaben. Herrn Palm und dem Daten- und Rechenzentrum sei für die Einrichtung des Internetportals gedankt. Roxana Barth und Gregor Willkommen haben für den reibungslosen Ablauf während des Kolloquiums gesorgt und auch tatkräftig bei der Zusammenstellung dieses Bandes geholfen. Natürlich ließ sich ein solches Kolloquium nur mit der Unterstützung der gesamten MT Arbeitsgruppe des GFZ durchführen. Ein herzliches Dankeschön dafür; ebenso an Herrn Bertelmann von der Bibliothek für das Hosting der Protokollbände.

Oliver Ritter und Ute Weckmann

Potsdam, 6. Juni 2010

II

Inhaltsverzeichnis / *table of contents*

Rita Streich & Michael Becken, <i>EM fields generated by finite-length wire sources in 1D media: comparison with point dipole solutions</i>	1
M. Afanasjew, R.-U. Börner, M. Eiermann, O. G. Ernst, S. Güttel & K. Spitzer, <i>2D Time Domain TEM Simulation Using Finite Elements, an Exact Boundary Condition, and Krylov Subspace Methods</i>	16
Tilman Hanstein, <i>TEM with anomalous diffusion in fractal conductive media</i>	25
K. M. Bhatt, A. Hördt & T. Hanstein, <i>Analysis of seafloor marine EM data with respect to motion-induced noise</i>	33
K. M. Bhatt, A. Hördt, P. Weidelt & T. Hanstein, <i>Motionally Induced Electromagnetic Field within the Ocean</i>	46
S. Kütter, A. Franke-Börner, R.-U. Börner & K. Spitzer, <i>Three-dimensional finite element simulation of magnetotelluric fields incorporating digital elevation models</i>	60
M. Becken, R. Streich & O. Ritter, <i>Establishing Controlled Source MT at GFZ</i>	71
Johannes Kenkel, Andreas Hördt & Andreas Kemna, <i>2D-SIP-Modellierung mit anisotropen Widerständen</i>	77
Xiaoming Chen, Ute Weckmann, Kristina Tietze, <i>Towards a 2D anisotropic inversion</i>	88
Kristina Tietze, Oliver Ritter & Ute Weckmann, <i>Substitute models for static shift in 2D</i>	97
Andreas Hördt, Peter Weidelt & Anita Przyklenk, <i>Die Übergangsimpedanz einer kapazitiv angekoppelten Elektrode</i>	102
Häuserer & Junge, <i>Eine Methode gegen auslaufende Ag/AgCl//KCl(H₂O) Elektroden</i>	115
Johannes B. Stoll, Celle, Christopher Virgil, <i>Attitude Algorithm Utilised in Mobile Geophysical Measuring Systems</i>	120
G. Muñoz, K. Bauer, I. Moeck, O. Ritter, <i>Joint Interpretation of Magnetotelluric and Seismic Models for Exploration of the Gross Schoenebeck Geothermal Site</i>	126
Ulrich Kalberkamp, <i>Magnetotelluric measurements to explore for deeper structures of the Tendaho geothermal field, Afar, NE Ethiopia</i>	131
J. H. Börner, V. Herdegen, R.-U. Börner, K. Spitzer, <i>Electromagnetic Monitoring of CO₂ Storage in Deep Saline Aquifers - Numerical Simulations and Laboratory Experiments</i>	137
K. Lippert, B. Tezkan, R. Bergers, M. Gurk, M. v. Papen, P. Yogeshwar, <i>Erkundung eines Aquifers unter dem Mittelmeer vor der israelischen Küste mit LOTEM</i>	143
M. von Papen, B. Tezkan, <i>On the analysis of LOTEM time series from Israel and the preliminary 1D inversion of data</i>	149
P. Yogeshwar, B. Tezkan, M. Israil, <i>Grundwasserkontamination bei Roorkee/Indien: 2D Joint Inversion von Radiomagnetotellurik und Gleichstromgeoelektrik Daten</i>	153
Widodo, Marcus Gurk, Bülent Tezkan, <i>Site Effect Assessment in the Mygdonian Basin (EUROSEISTEST area, Northern Greece) using RMT and TEM Soundings</i>	164
Gerlinde Schaumann, Annika Steuer, Bernhard Siemon, Helga Wiederhold & Franz Binot, <i>Die deutsche Nordseeküste im Fokus von aeroelektromagnetischen Untersuchungen Teilgebiete Langeoog mit Wattenmeer und Elbemündung</i>	177

Annika Steuer, Bernhard Siemon and Michael Grinat, <i>The German North Sea Coast in Focus of Airborne Electromagnetic Investigations: The Freshwater Lenses of Borkum</i>	188
Gerhard Kapinos & Heinrich Brasse, <i>Some notes on bathymetric effects in marine magnetotellurics, motivated by an amphibious experiment at the South Chilean margin</i>	198
Dirk Brändlein, Oliver Ritter, Ute Weckmann, <i>A permanent array of magnetotelluric stations located at the South American subduction zone in Northern Chile.</i>	210
D. Eydam & H. Brasse, <i>Discussion on backarc mantle melting in the central Andean subduction zone, based on results of magnetotelluric studies</i>	216
Václav Červ, Světlana Kováčiková, Michel Menvielle and Josef Pek, <i>Thin Sheet Conductance Models from Geomagnetic Induction Data: Application to Induction Anomalies at the Transition from the Bohemian Massif to the West Carpathians</i>	232
Anne Neska, <i>Subsurface Conductivity Obtained from DC Railway Signal Propagation with a Dipole Model</i>	244
Anja Schäfer, Heinrich Brasse, Norbert Hoffmann and EMTEZ working group, <i>Magnetotelluric investigation of the Sorgenfrei-Tornquist Zone and the NE German Basin</i>	252
P. Sass, O. Ritter, A. Rybin, G. Muñoz, V. Batalev and M. Gil, <i>Magnetotelluric data from the Tien Shan and Pamir continental collision zones, Central Asia</i>	263
Alexander Löwer, <i>Audiomagnetotellurik im Hohen Vogelsberg</i>	269
Ute Weckmann, Carsten Scholl, <i>Dumme Fragen</i>	277

III

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EM fields generated by finite-length wire sources in 1D media: comparison with point dipole solutions

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ABSTRACT

In present-day land and marine controlled-source electromagnetic (CSEM) surveys, EM fields are commonly generated using wires that are hundreds of meters long. Nevertheless, simulations of CSEM data often approximate these sources as point dipoles. Although this is justified for sufficiently large source-receiver distances, many real surveys include frequencies and distances at which the dipole approximation is inaccurate. For 1D layered media, EM fields for point dipole sources can be computed using well-known quasi-analytical solutions, and fields for sources of finite lengths can be synthesized by superposing point dipole fields. However, the calculation of numerous point dipole fields is computationally expensive, requiring a large number of numerical integral evaluations. We combine a more efficient representation of finite-length sources in terms of components related to the wire and its end points with very general expressions for EM fields in 1D layered media. We thus obtain a formulation that requires fewer numerical integrations than the superposition of dipole fields and permits source and receiver placement at any depth within the layer stack. Complex source geometries, such as wires bent due to surface obstructions, can be simulated by segmenting the wire and computing the responses for each segment separately. We first describe our finite-length wire expressions, and then present examples of EM fields due to finite-length wires for typical land and marine survey geometries and discuss differences to point dipole fields.

INTRODUCTION

Controlled-source electromagnetic (CSEM) surveys are a useful exploration tool, applicable, e.g., for exploring hydrocarbon reservoirs, geothermal reservoirs, or for characterizing and potentially monitoring sites considered for carbon sequestration. A multitude of active electromagnetic sources are available, including magnetic loops (Frischknecht et al., 1991; Spies and Frischknecht, 1991) and long wires, grounded in land-based surveys (Strack, 1992; Wright et al., 2002; Ziolkowski et al., 2007) and deployed at the seafloor (Edwards, 2005) or towed through the water (e.g., Constable and Srnka, 2007) in marine surveys. Wire sources with lengths of several 100 m are most commonly used in commercial hydrocarbon exploration because of their capability to generate three-dimensional electrical current systems sensitive to both resistive and conductive, relatively deep targets (Spies and Frischknecht, 1991; Constable and Srnka, 2007).

In processing, inversion and interpretation of CSEM data, the sources are commonly approximated as point dipoles (e.g., Edwards, 1997; Johansen et al., 2005; Ziolkowski et al., 2007). This is adequate for sufficiently large source-receiver distances. However, real surveys targeting increasingly deep and small structures may include distances and frequencies at which the inaccuracy of the dipole approximation is on the order of (or larger than) the target-related anomalies. In such cases, it is vital to consider the actual source geometry.

For horizontally layered media, EM fields can be computed using well-known quasi-analytical solutions that involve numerical evaluations of Bessel function integrals (e.g., Weidelt, 2007; Løseth and Ursin, 2007). Using such point dipole solutions, EM fields due to finite-length wire sources can be synthesized by representing the wire as a line of point dipoles and summing the dipole fields. However, this procedure is computationally expensive, requiring many numerical integrations to calculate all of the point dipole contributions.

To improve the efficiency of long-wire source simulations, Soerensen and Christensen (1994) derived integrated Hankel integral expressions for finite-length wires that barely require more integral evaluations than the computation of point dipole fields. However, their approach includes elaborate computations of filter coefficients for every source-receiver pair. Another approach that reduces the number of integrations from that required for summing dipole fields is the separation of EM fields due to long wires into contributions from the wire and its end points (Ward and Hohmann, 1987). This technique works with standard Hankel filters. Whereas previous application of this approach considered subsurface receivers and sources located at the air-ground interface (Ward and Hohmann, 1987), we have applied it to a more general 1D field formulation that permits source and receiver positions at arbitrary depths within the layer stack (Løseth and Ursin, 2007). This allows simulations of finite marine sources in addition to land sources.

In this contribution, we first describe our finite-length wire representation. We then show examples of EM fields due to finite-length wires in representative land and marine settings over the frequency and distance ranges of typical CSEM surveys, and discuss differences to the fields due to infinitesimal dipoles.

EM FIELD EXPRESSIONS FOR FINITE-LENGTH WIRES

Expressions for the electromagnetic field due to a finite-length wire can be obtained by representing the wire as a line of infinitesimal dipoles and integrating the dipole expressions along the length of the wire. Using the nomenclature of Løseth and Ursin (2007), the EM fields for an x -directed point dipole embedded in a 1D layered medium can be expressed in

a form convenient for later integration as (see Appendix A)

$$E_x = \frac{Idx}{4\pi} \left(\int_0^\infty \frac{-\mu_0}{\sqrt{p_z^r p_z^s}} \mathcal{R}_{11}^A \kappa J_0(\kappa r) d\kappa + \partial_x \left\{ \frac{x^r - x^s}{r} \int_0^\infty \left[\frac{\mu_0}{\sqrt{p_z^r p_z^s}} \mathcal{R}_{11}^A - \sqrt{\frac{p_z^r p_z^s}{\tilde{\varepsilon}^r \tilde{\varepsilon}^s}} \mathcal{R}_{22}^A \right] J_1(\kappa r) d\kappa \right\} \right), \quad (1a)$$

$$E_y = \frac{Idx}{4\pi} \partial_x \left\{ \frac{y^r - y^s}{r} \int_0^\infty \left[\frac{\mu_0}{\sqrt{p_z^r p_z^s}} \mathcal{R}_{11}^A - \sqrt{\frac{p_z^r p_z^s}{\tilde{\varepsilon}^r \tilde{\varepsilon}^s}} \mathcal{R}_{22}^A \right] J_1(\kappa r) d\kappa \right\}, \quad (1b)$$

$$E_z = \frac{jIdx}{4\pi\omega\tilde{\varepsilon}^r} \partial_x \left\{ \int_0^\infty \sqrt{\frac{\tilde{\varepsilon}^r \tilde{\varepsilon}^s}{p_z^r p_z^s}} \mathcal{R}_{22}^D \kappa J_0(\kappa r) d\kappa \right\}, \quad (1c)$$

$$H_x = \frac{Idx}{4\pi} \partial_x \left\{ \frac{y^r - y^s}{r} \int_0^\infty \left[\sqrt{\frac{p_z^r}{p_z^s}} \mathcal{R}_{11}^D - \sqrt{\frac{\tilde{\varepsilon}^r p_z^s}{\tilde{\varepsilon}^s p_z^r}} \mathcal{R}_{22}^D \right] J_1(\kappa r) d\kappa \right\}, \quad (1d)$$

$$H_y = \frac{Idx}{4\pi} \left(\int_0^\infty \sqrt{\frac{p_z^r}{p_z^s}} \mathcal{R}_{11}^D \kappa J_0(\kappa r) d\kappa - \partial_x \left\{ \frac{x^r - x^s}{r} \int_0^\infty \left[\sqrt{\frac{p_z^r}{p_z^s}} \mathcal{R}_{11}^D - \sqrt{\frac{\tilde{\varepsilon}^r p_z^s}{\tilde{\varepsilon}^s p_z^r}} \mathcal{R}_{22}^D \right] J_1(\kappa r) d\kappa \right\} \right), \quad (1e)$$

$$H_z = \frac{jIdx}{4\pi\omega\mu_0} \frac{y^r - y^s}{r} \int_0^\infty \frac{\mu_0}{\sqrt{p_z^r p_z^s}} \mathcal{R}_{11}^A \kappa^2 J_1(\kappa r) d\kappa, \quad (1f)$$

where I is the source current, dx is the length of the source dipole, κ is the horizontal wavenumber, ω is the angular frequency, μ_0 is the vacuum magnetic permeability, $\tilde{\varepsilon}^{\{s,r\}} = \varepsilon^{\{s,r\}} + j\sigma^{\{s,r\}}/\omega$ are the dielectric permittivity and electric conductivity of the source and receiver layer, respectively, $p_z^{\{s,r\}} = \sqrt{\mu_0 \tilde{\varepsilon}^{\{s,r\}} - \kappa^2/\omega^2}$ are frequency-normalized vertical wavenumbers, $r = \sqrt{(x^r - x^s)^2 + (y^r - y^s)^2}$ is the horizontal source-receiver distance, and J_0 and J_1 are the zero- and first-order Bessel functions. $\mathcal{R}_{11}^A$, $\mathcal{R}_{22}^A$, $\mathcal{R}_{11}^D$ and $\mathcal{R}_{22}^D$ are the reflection responses of the layered medium as given by Løseth and Ursin (2007, their Equation 134). These quantities are computed recursively using the properties and thicknesses of all layers and the source and receiver depths. Subscripts 11 denote the TE-mode and subscripts 22 the TM-mode responses. Slightly different expressions apply for receivers below (Equations 134a and b) and above the source (Equations 134c and d of Løseth and Ursin, 2007). In Equations (1), a spatial derivative ∂_x has been retained wherever possible.

To obtain electromagnetic fields for a finite-length wire, we integrate Equations (1) over the wire length. Assuming that the wire is parallel to the x -axis, we express the integrals as discrete sums over N wire elements of length Δx , located at (x_n, y^s, z^s) . The derivatives ∂_x are replaced by $-\partial/\partial_{\Delta x}$ (Ward and Hohmann, 1987). Then the fields at receiver location

(x^r, y^r, z^r) are

$$E_x = \frac{I}{4\pi} \sum_{n=1}^N \Delta x \int_0^\infty \frac{-\kappa \mu_0}{\sqrt{p_z^r p_z^s}} \mathcal{R}_{11}^A J_0(\kappa r_n) d\kappa - \frac{I}{4\pi} \sum_{m=1}^2 (-1)^m \frac{x^r - x_m}{r_m} \int_0^\infty \left[\frac{\mu_0}{\sqrt{p_z^r p_z^s}} \mathcal{R}_{11}^A - \sqrt{\frac{p_z^r p_z^s}{\tilde{\epsilon}^r \tilde{\epsilon}^r}} \mathcal{R}_{22}^A \right] J_1(\kappa r_m) d\kappa, \quad (2a)$$

$$E_y = -\frac{I}{4\pi} \sum_{m=1}^2 (-1)^m \frac{y^r - y^s}{r_m} \int_0^\infty \left[\frac{\mu_0}{\sqrt{p_z^r p_z^s}} \mathcal{R}_{11}^A - \sqrt{\frac{p_z^r p_z^s}{\tilde{\epsilon}^r \tilde{\epsilon}^r}} \mathcal{R}_{22}^A \right] J_1(\kappa r_m) d\kappa, \quad (2b)$$

$$E_z = -\frac{jI}{4\pi\omega\tilde{\epsilon}^r} \sum_{m=1}^2 (-1)^m \int_0^\infty \sqrt{\frac{\tilde{\epsilon}^r p_z^s}{\tilde{\epsilon}^s p_z^r}} \mathcal{R}_{22}^D \kappa J_0(\kappa r_m) d\kappa, \quad (2c)$$

$$H_x = -\frac{I}{4\pi} \sum_{m=1}^2 (-1)^m \frac{y^r - y^s}{r_m} \int_0^\infty \left[\sqrt{\frac{p_z^r}{p_z^s}} \mathcal{R}_{11}^D - \sqrt{\frac{\tilde{\epsilon}^r p_z^s}{\tilde{\epsilon}^s p_z^r}} \mathcal{R}_{22}^D \right] J_1(\kappa r_m) d\kappa, \quad (2d)$$

$$H_y = \frac{I}{4\pi} \sum_{n=1}^N \Delta x \int_0^\infty \sqrt{\frac{p_z^r}{p_z^s}} \mathcal{R}_{11}^D \kappa J_0(\kappa r_n) d\kappa + \frac{I}{4\pi} \sum_{m=1}^2 (-1)^m \frac{x^r - x_m}{r_m} \int_0^\infty \left[\sqrt{\frac{p_z^r}{p_z^s}} \mathcal{R}_{11}^D - \sqrt{\frac{\tilde{\epsilon}^r p_z^s}{\tilde{\epsilon}^s p_z^r}} \mathcal{R}_{22}^D \right] J_1(\kappa r_m) d\kappa, \quad (2e)$$

$$H_z = \frac{jI(y^r - y^s)}{4\pi\mu_0\omega} \sum_{n=1}^N \frac{\Delta x}{r_n} \int_0^\infty \frac{\mu_0}{\sqrt{p_z^r p_z^s}} \mathcal{R}_{11}^A \kappa^2 J_1(\kappa r_n) d\kappa. \quad (2f)$$

For E_x , H_y and H_z , we obtain a contribution from each wire element. This contribution is described by the first sum in Equations 2a, 2e and 2f, with the distance between the receiver and the n^{th} wire element given by $r_n = \sqrt{(x^r - x_n)^2 + (y^r - y^s)^2}$. Upon integrating those terms of Equations (1) that contain derivatives ∂_x , the derivatives disappear, and we obtain explicit contributions from the integration limits, i.e., the end points of the wire. In Equations (2), the end point contributions are given by the summation over m ($m \in \{1, 2\}$), with the distances between the receiver and the wire ends denoted r_m .

To compute all EM field components for a finite-length wire using Equation (2), we have to evaluate three different integrals over the entire wire length. Accordingly, for a wire discretized into N elements, the number of numerical integral evaluations is approximately $3N$. In contrast, Equation (A-7) shows that two different integrals are required for computing only E_x for an infinitesimal dipole. The computation of all electromagnetic field components for a horizontal electric point dipole source requires numerical evaluations of eight different integrals (Løseth and Ursin, 2007). This would result in $8N$ numerical integrations when calculating finite-length wire fields from the contributions of N dipole elements. Compared to the simple summation of dipole fields, the separate computation of wire and end point contributions thus reduces the computational effort by more than 60%.

EXAMPLES

We use the expressions of EM fields due to finite-length wires given in Equation (2) to assess differences between the fields of finite sources and point dipoles for different representative experimental settings of land and marine surveys.

Land survey: straight grounded wire

We have simulated a land survey using the setup depicted in Figure 1. The electric conductivity model is based on the situation at the CO₂ sequestration pilot site in Ketzin, Germany, using measured conductivity values and the actual depth and thickness of the layer into which CO₂ is being injected (Giese et al., 2009). To simulate a realistic survey, in which sensors would typically be buried just below the surface, we placed electric and magnetic field receivers at a depth of 0.15 m. EM fields were computed for an infinitesimal dipole source at $(x, y) = (0, 0)$ and a 1-km long finite-length wire centered at $(x, y) = (0, 0)$. Both sources were located at a depth of 0.1 m.

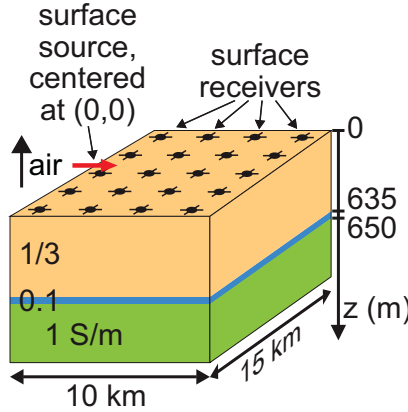


Figure 1: The 1D conductivity model and survey geometry used for simulating a land CSEM survey. The red arrow indicates the point dipole or 1000-m long wire source, centered at $(x, y) = (0, 0)$ and located 0.1 m below the surface. Receivers are located 0.15 m below the surface.

For convenience in the numerical simulations, we place the entire wire at a constant depth, either above or below the air-ground interface, although in real field surveys, grounded wires would typically be used, with the wire laid out on the surface and the end points coupled into the ground, e.g., via metal electrodes. The contribution from the wire body is typically smaller than that from the grounding points, and varies slowly as the wire depth crosses interfaces. Therefore, this simplification does not cause noticeable errors.

In Figure 2, we display the electric field component E_x at a frequency of 0.1 Hz for the model depicted in Figure 1. The finite-wire and point dipole fields differ significantly within a radius of ~ 4 km from the source. Large relative differences also occur in the lowest-amplitude regions at oblique angles to the source; however, these are insignificant, because in these regions, E_x would not be measurable.

For comparison, we show in Figure 3 the electric field for the reservoir model of Fig-

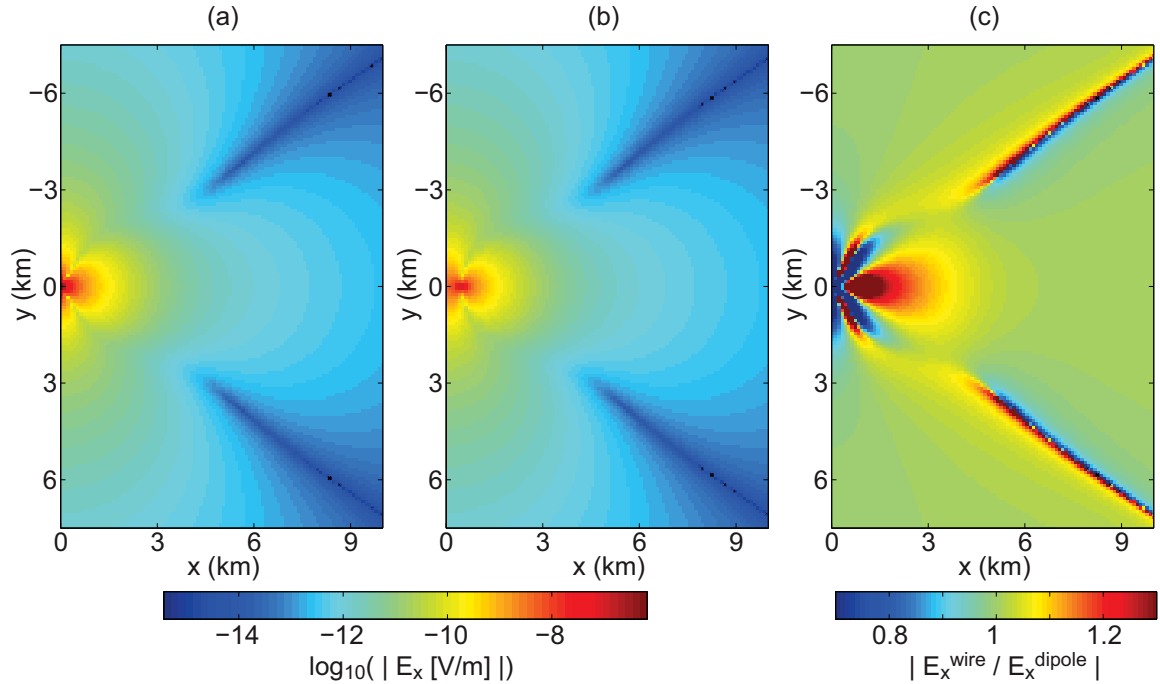


Figure 2: Electric field E_x for the configuration shown in Figure 1, a frequency of 0.1 Hz and (a) a point dipole source at (0, 0, 0.1) m and (b) a 1000-m long grounded wire extending from (-500, 0, 0.1) to (500, 0, 0.1) m. (c) shows the ratio between the finite-length wire and point dipole fields.

Figure 1 relative to the electric field for a model that does not contain a resistive layer. This demonstrates the size of the anomaly we would attempt to detect in the CSEM survey. The reservoir-related anomaly is smaller than the relative differences between finite-length wire and dipole fields, and overlaps spatially with the region in which the response is significantly influenced by the source geometry. This clearly indicates the importance of considering the true source geometry.

Similar observations can be made over a wide frequency range. In Figure 4, we compare the electric field E_x for finite-length and point dipole sources, and for the reservoir and background models, at frequencies of 0.001 Hz and 1 Hz. Significant differences between finite-length and point dipole sources occur in a similar region as for $f = 0.1$ Hz (compare Figures 4a and b to Figure 2c). The reservoir-related anomalies are somewhat smaller than for $f = 0.1$ Hz, underlining again that it is vital to consider the actual source length when searching for such relatively small anomalies.

In Figure 5, we display the finite-length wire fields for the E_y and E_z components, and their ratios to the respective point dipole fields. Here, the relative differences between the finite-length wire and point dipole fields are of similar size as for E_x . The regions in which the responses of finite-length wires and point dipoles differ significantly are similar, or slightly larger, in extent than for E_x . However, the amplitudes of E_z are considerably smaller than those of E_x and E_y , such that nearly the entire region in which E_z would be measurable (assuming instrument detection thresholds of $\sim 10^{-14} - 10^{-15}$, and possible measurement of E_z in shallow boreholes) is strongly influenced by the source geometry.

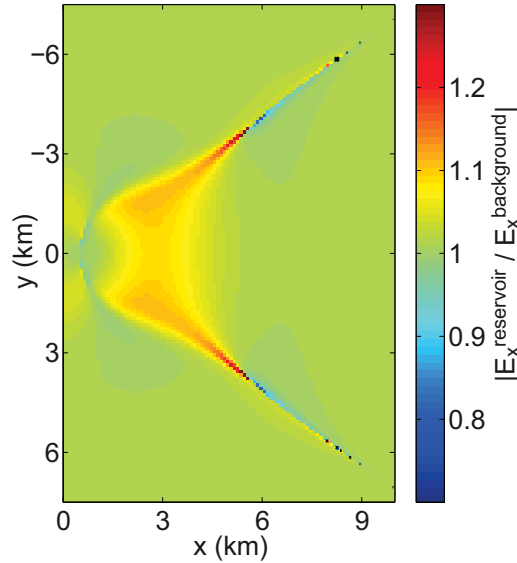


Figure 3: Electric field E_x for the configuration shown in Figure 1, a 1000-m long wire source and $f = 0.1$ Hz, normalized by the electric field for a background model not containing a resistive layer.

Required wire discretization

To minimize the computational effort, we have empirically investigated the effects of wire discretization on the resulting electromagnetic field values. In Figure 6, we display differences between a reference field E_x computed for a 1000-m long wire discretized using an element length of 0.1 m, and more coarsely discretized wires with element lengths varying between 1 m and 50 m. As expected, differences between the reference field and fields computed using coarser discretizations increase as the element length increases. Nevertheless, at the lower frequency of 0.1 Hz (Figure 6a), differences are small for all tested element lengths. As expected, differences are larger at $f = 100$ Hz, with maximum differences occurring in the vicinity of the end point of the wire at $x = 500$ m.

From this test, we conclude that for the lower frequency, a coarse discretization of ~ 20 – 50 m would be sufficient, whereas for the higher frequency, the wire should be discretized using elements no longer than ~ 2 – 5 m. These results may serve as rough guidelines for further simulation studies, but actual required discretizations are likely to depend on the total wire length and the resistivity model. To exclude any discretization-related error, all results presented here were computed using somewhat too careful discretizations with element lengths of 1 m.

Non-straight grounded wire

In real field surveys, surface obstacles may preclude laying out the source wire in a straight line. We have therefore studied differences between the EM fields for straight and non-straight wire sources. EM fields for non-straight wires are calculated by segmenting the wire into straight sections, and computing the responses for each segment separately using

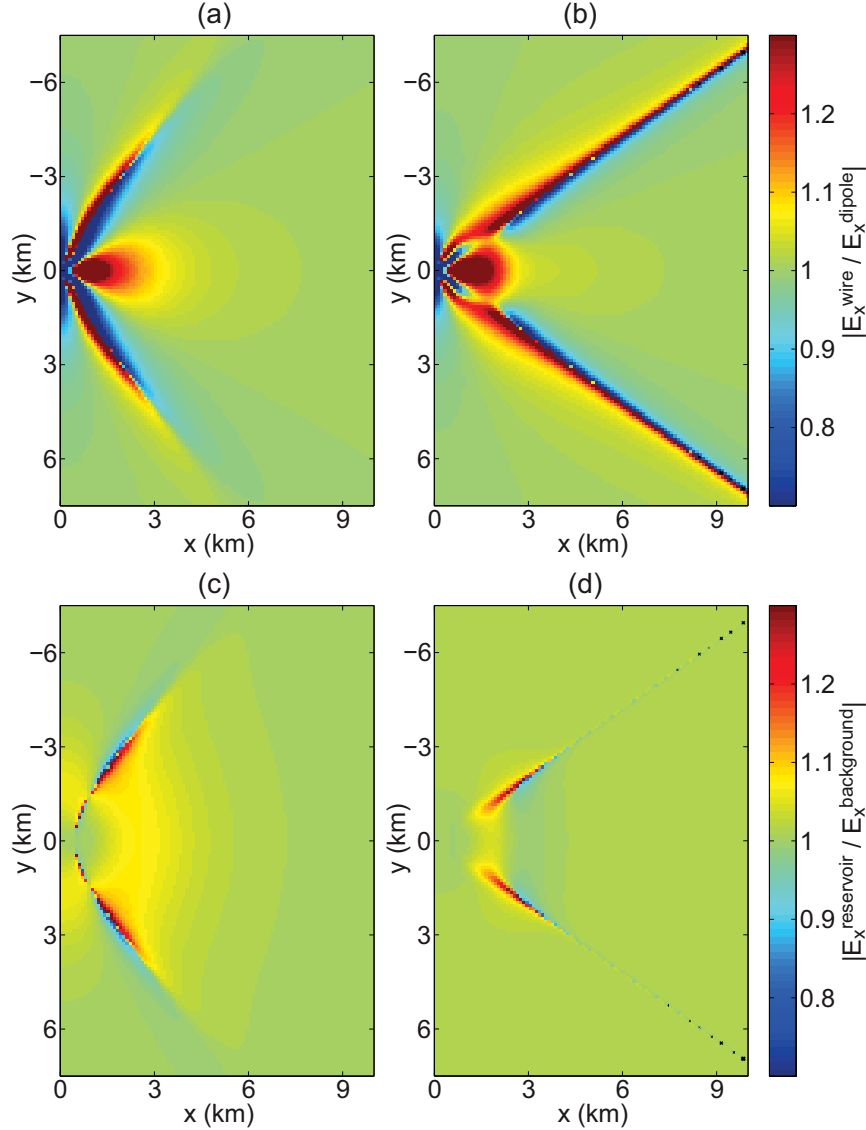


Figure 4: Ratio of E_x for a 1000-m long wire source to E_x for a point dipole for frequencies of (a) 0.001 Hz and (b) 1 Hz, and ratio of E_x for a 1000-m long wire source and the reservoir model shown in Figure 1 relative to E_x for a background model not containing the resistive layer for frequencies of (c) 0.001 Hz and (d) 1 Hz.

Equations (2) with appropriate rotations. As an example, we have computed the responses for a wire that consists of two segments arranged at a 120° angle. This wire has the same grounding points as the 1000-m long straight wire used previously. Figure 7 shows the wire geometry and the electric field E_x for the bent wire relative to E_x for a straight wire at frequencies of 0.001 Hz, 0.1 Hz and 1 Hz.

As expected, the E_x amplitudes for the bent wire are increased relative to the straight wire field at the side to which the wire is deviated ($y > 0$), and decreased at $y < 0$. Furthermore, relative differences between the bent and straight wire fields increase with increasing frequency, reaching a size similar to the reservoir-related anomaly for $f = 0.1$ Hz (com-

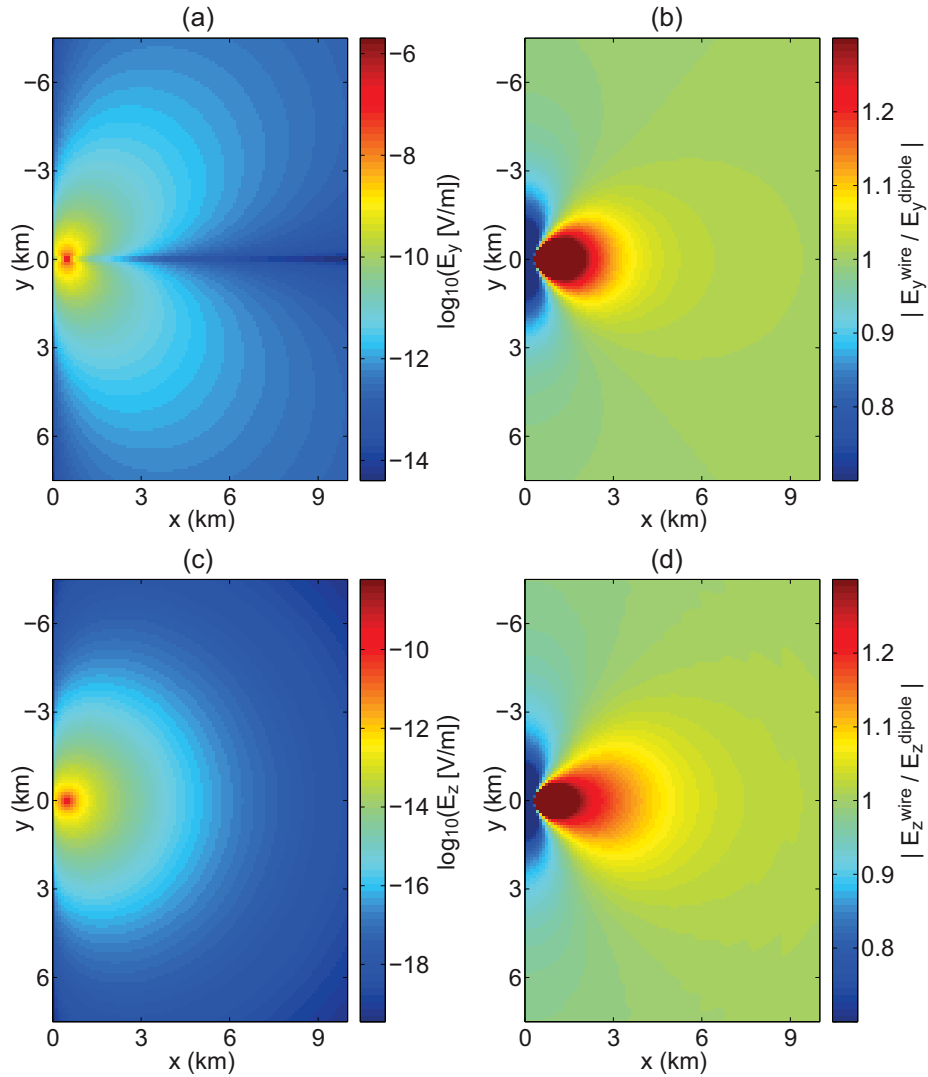


Figure 5: Electric field components (a) E_y and (c) E_z for the model shown in Figure 1 and a 1000-m long wire source at $f = 0.1$ Hz, and (b, d) the ratios of the finite-source E_y and E_z to the respective point dipole fields.

pare Figure 7b to 3), and larger than the reservoir-related anomaly for $f = 1$ Hz (compare Figure 7c to 4d). These results demonstrate that at these frequencies, the EM fields are significantly influenced not only by the grounding point locations, but also by the entire wire layout. In contrast, at $f = 0.001$ Hz, differences between bent and straight wire fields are barely visible (Figure 7a). At this low frequency, the electric field is similar to the potential field occurring in the DC limit (i.e., for $f = 0$). Here, the electric field is practically determined by the grounding point locations alone.

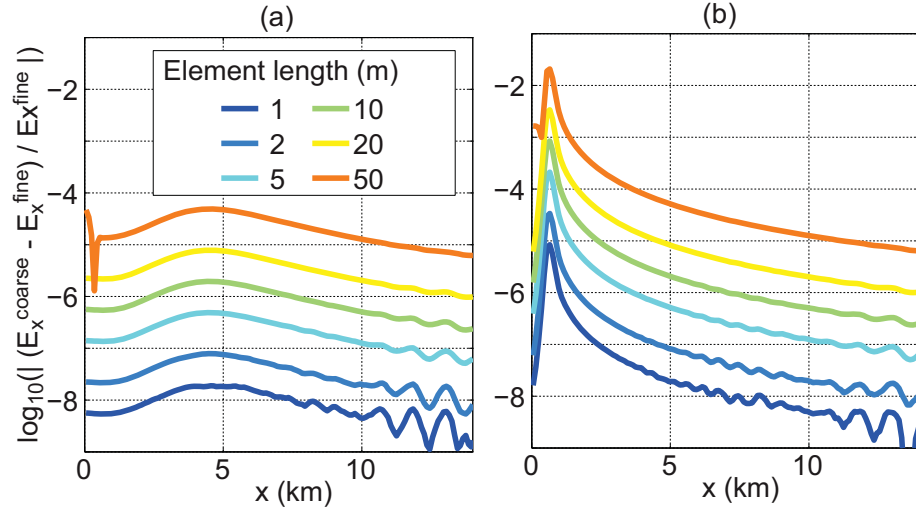


Figure 6: Relative differences between E_x^{coarse} computed for a 1000-m long wire using wire element lengths between 1 m and 50 m, and a reference field E_x^{fine} for a wire discretized using 0.1-m long elements, for frequencies of (a) 0.1 Hz and (b) 100 Hz. The resistivity model shown in Figure 1 was used, and field values were extracted along a line parallel to the source wire at a lateral offset of 50 m.

Marine survey: floating wire

Finite electric dipole sources used in marine CSEM surveys for commercial hydrocarbon exploration are typically $\sim 100 - 300$ m long (Constable and Srnka, 2007). Accordingly, differences between the EM fields due to such finite-length sources and point dipole fields are expected to be somewhat smaller than those observed for the 1-km long wire considered in the land survey examples.

We have computed marine EM responses for the model shown in Figure 8, which contains a 200-m water column and a stack of sedimentary layers, into which a 100-m thick resistive layer, representing a hydrocarbon reservoir, is embedded at a depth of 800–900 m below the seafloor. We simulated a point dipole source located at $(x, y) = (0, 0)$ and a 300-m long wire source, also centered at $(0, 0)$. Both sources were located 50 m above the seafloor, and receivers were placed 0.01 m above the seafloor.

Electric field E_x data for this configuration due to the finite-length wire and point dipole sources are shown in Figures 9a and b, and finite-source E_x data for the reservoir model relative to E_x for a model not containing the resistive layer are displayed in Figure 9c. The finite-length wire and point dipole fields are nearly identical at radii larger than ~ 1 km from the source center. The anomaly due to the resistive layer is several times larger than the relative differences between the finite-length wire and point dipole fields, and is largest at distances well beyond the region significantly influenced by the source geometry. Therefore, the point dipole approximation may be adequate in this case. However, taking into account the exact source geometry may again become important when searching for smaller anomalies caused, e.g., by thinner reservoirs or relatively small 3D structures.

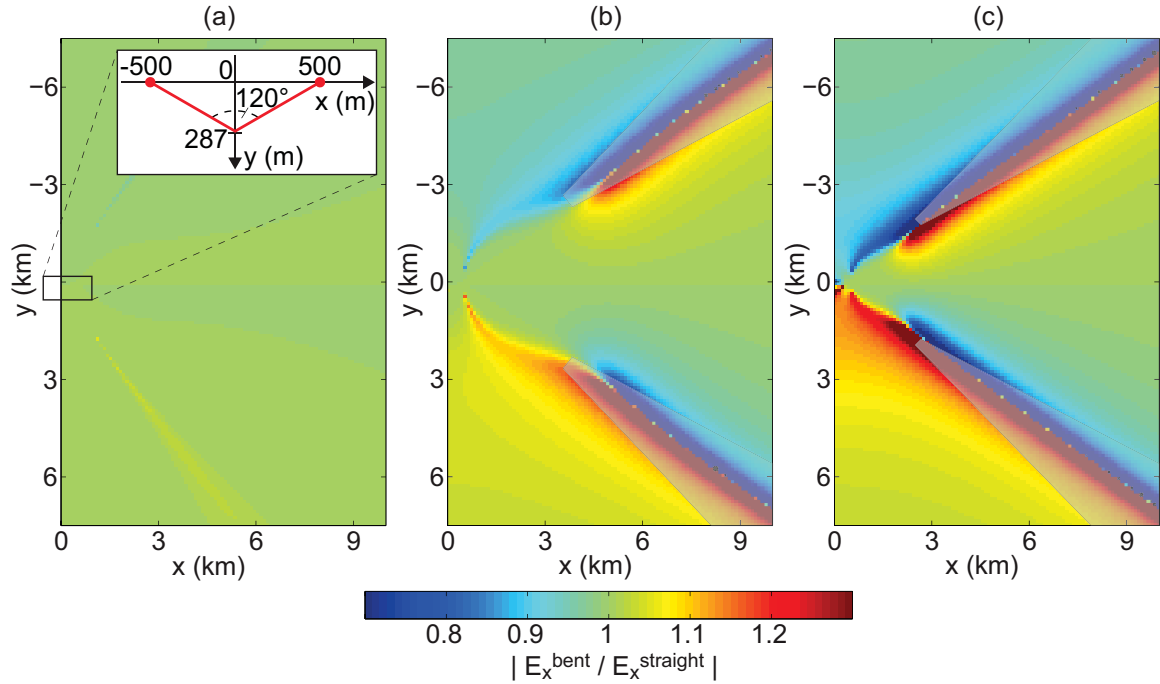


Figure 7: Electric field E_x for a bent wire. The inset in (a) shows the geometry of a wire that consists of two segments and has the same grounding points as the straight wire used previously. Shown is the ratio of E_x^{bent} for the bent wire to $E_x^{straight}$ for the straight wire at depth $z = 0.15$ m and frequencies of (a) 0.001 Hz, (b) 0.1 Hz and (c) 1 Hz. Gray shades in (b) and (c) roughly mark very low-amplitude regions in which E_x would not be measurable.

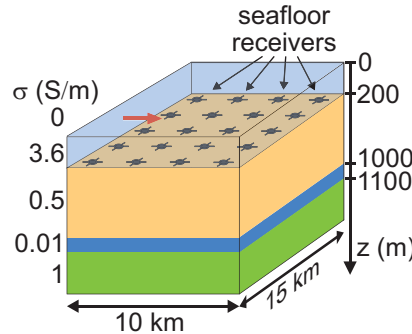


Figure 8: The 1D conductivity model and survey geometry used for simulating a marine CSEM survey. The red arrow indicates the point dipole or 300-m long wire source, centered at $(x, y) = (0, 0)$ and located 50 m above the seafloor. Receivers are located 0.01 m above the seafloor.

CONCLUSIONS

We have presented a formulation for computing electromagnetic fields due to electric dipole sources of finite extent by splitting the responses into contributions from the wire body and its end points. Being derived from a quite general representation of point dipole fields, our finite-length wire formulation allows us to compute EM fields in 1D layered media for

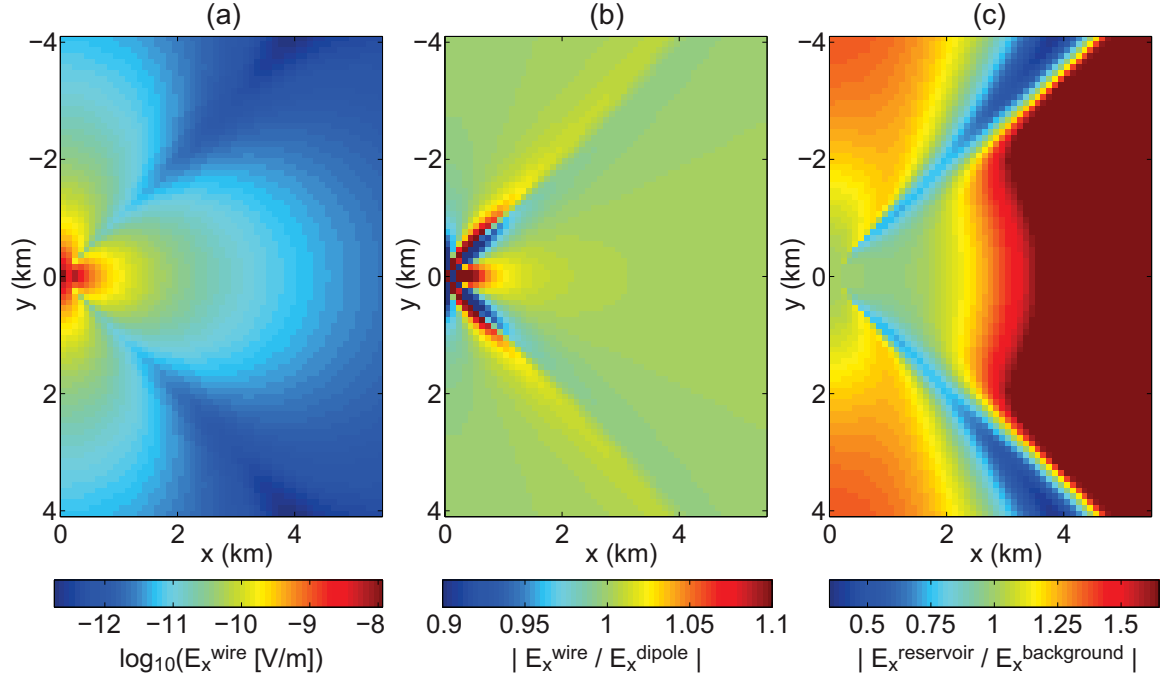


Figure 9: (a) Electric field E_x for a 300-m long wire source and the marine survey configuration shown in Figure 8 at $f = 0.1$ Hz. (b) The ratio between the finite-length wire field shown in (a) and the field due to a point dipole source. (c) The finite-length wire field for the reservoir model of Figure 8 relative to the background field for a model not containing the 100-m thick resistive layer. Note the different color scales in (b) and (c).

sources and receivers located at any depth. Compared to direct summation of dipole fields, we gain $\sim 60\%$ efficiency. Implementation is straightforward, as we only require integral evaluations via standard fast Hankel transforms, for which precomputed sets of coefficients are available.

The utility of the finite-length wire formulation has been demonstrated by presenting EM fields for several representative CSEM survey configurations. Our tests confirm that finite-length wire and point dipole fields can differ significantly over distance ranges reaching several times the wire length. For a simulated land CSEM survey targeting a thin resistive anomaly, comparison of the responses for a 1000-m long grounded wire source to point dipole responses shows that in this case, it is crucial to consider the actual source geometry. The anomalous response of the target structure is smaller than the relative differences between finite-length wire and point dipole fields, and overlaps spatially with the region in which the response is strongly influenced by the source geometry. Qualitatively similar deviations between finite-length wire and point dipole fields are observed over relatively wide frequency ranges and for different EM field components. At frequencies above the ‘effective’ DC limit, it is also important to consider the actual wire layout; knowledge of the grounding point positions is not sufficient for computing accurate responses.

In contrast, for the marine survey simulated, using a shorter wire and thicker target layer, the wire geometry only had a relatively small impact on the responses, and anomalies due to the target layer were spatially well separated from the region significantly influenced by

the source geometry. This indicates that the actual source geometry is of minor importance for the survey considered. Nevertheless, this cannot be generalized to other surveys aiming at detecting smaller target structures that generate weaker EM field anomalies.

Our 1D solution for finite-length wire sources can easily be incorporated into higher-dimensional CSEM modeling and inversion algorithms. Most 2D and 3D modeling schemes use a secondary field approach, in which primary fields for a simple (e.g., homogeneous or 1D) model are computed analytically, and secondary fields arising from deviations of the 2D or 3D resistivity model from the background model are computed numerically. Here, finite-length sources can be included into the background field computations.

ACKNOWLEDGMENTS

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APPENDIX A

EM FIELD EXPRESSIONS FOR POINT DIPOLES

We show the derivation of expressions for dipole EM fields in a form convenient for integration over the length of a source wire using the example of the E_x component. Analogous considerations apply for the other EM field components.

Using the nomenclature of Løseth and Ursin (2007), the space-frequency domain electric field E_x for an isotropic layered medium, with source and receiver embedded at arbitrary depth, is given by the double Fourier integral

$$E_x = -\frac{Idx}{8\pi^2} \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \left[\left(1 - \frac{k_x^2}{\kappa^2}\right) \frac{\mu_0}{\sqrt{p_z^r p_z^s}} \mathcal{R}_{11}^A + \sqrt{\frac{p_z^r p_z^s}{\tilde{\varepsilon}^r \tilde{\varepsilon}^s}} \frac{k_x^2}{\kappa^2} \mathcal{R}_{22}^A \right] \exp\{j(k_x x + k_y y)\} dk_x dk_y, \quad (\text{A-1})$$

where k_x and k_y are horizontal wavenumbers with $\kappa^2 = k_x^2 + k_y^2$, and the other symbols are the same as those explained for Equation (1).

Transformation to cylindrical coordinates using $k_x = \kappa \cos \alpha$, $k_y = \kappa \sin \alpha$, $x = r \cos \beta$, $y = r \sin \beta$, $\xi = \alpha - \beta + \pi/2$, $k_x^2 \rightarrow -\partial_x^2$, and substituting the zero-order Bessel function,

$$J_0(\kappa r) = \frac{1}{2\pi} \int_0^{2\pi} e^{j\kappa r \sin \xi} d\xi, \quad (\text{A-2})$$

results in

$$E_x = \frac{Idx}{4\pi} \left(\int_0^{\infty} \frac{-\mu_0}{\sqrt{p_z^r p_z^s}} \mathcal{R}_{11}^A \kappa J_0(\kappa r) d\kappa - \partial_x^2 \left\{ \int_0^{\infty} \left[\frac{\mu_0}{\sqrt{p_z^r p_z^s}} \mathcal{R}_{11}^A - \sqrt{\frac{p_z^r p_z^s}{\tilde{\varepsilon}^r \tilde{\varepsilon}^s}} \mathcal{R}_{22}^A \right] \frac{1}{\kappa} J_0(\kappa r) d\kappa \right\} \right). \quad (\text{A-3})$$

After evaluating *one* of the spatial derivatives using (Ward and Hohmann, 1987)

$$\partial_x J_0(\kappa r) = -\frac{\kappa x}{r} J_1(\kappa r), \quad (\text{A-4})$$

we obtain

$$E_x = \frac{Idx}{4\pi} \left(\int_0^\infty \frac{-\mu_0}{\sqrt{p_z^r p_z^s}} \mathcal{R}_{11}^A \kappa J_0(\kappa r) d\kappa + \partial_x \left\{ \frac{x}{r} \int_0^\infty \left[\frac{\mu_0}{\sqrt{p_z^r p_z^s}} \mathcal{R}_{11}^A - \sqrt{\frac{p_z^r p_z^s}{\tilde{\epsilon}^r \tilde{\epsilon}^s}} \mathcal{R}_{22}^A \right] d\kappa \right\} \right). \quad (\text{A-5})$$

Equation (A-5) is used as the basis for deriving the expressions for finite-length wire fields. Further evaluation of the second spatial derivative using

$$\partial_x J_1(\kappa r) = -\frac{x}{r^2} J_1(\kappa r) + \frac{\kappa x}{r} J_0(\kappa r) \quad (\text{A-6})$$

results in an explicit expression for the point dipole field in terms of two different Hankel integrals:

$$\begin{aligned} E_x = & \frac{Idx}{4\pi} \left(\int_0^\infty \left[\left(\frac{x^2}{r^2} - 1 \right) \frac{\mu_0}{\sqrt{p_z^r p_z^s}} \mathcal{R}_{11}^A - \frac{x^2}{r^2} \sqrt{\frac{p_z^r p_z^s}{\tilde{\epsilon}^r \tilde{\epsilon}^s}} \mathcal{R}_{22}^A \right] \kappa J_0(\kappa r) d\kappa \right. \\ & \left. + \left(\frac{1}{r} - \frac{2x^2}{r^3} \right) \int_0^\infty \left[\frac{\mu_0}{\sqrt{p_z^r p_z^s}} \mathcal{R}_{11}^A - \sqrt{\frac{p_z^r p_z^s}{\tilde{\epsilon}^r \tilde{\epsilon}^s}} \mathcal{R}_{22}^A \right] J_1(\kappa r) d\kappa \right). \quad (\text{A-7}) \end{aligned}$$

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2D Time Domain TEM Simulation Using Finite Elements, an Exact Boundary Condition, and Krylov Subspace Methods

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1 Summary

In this report we present a numerical method for solving Maxwell's equations in the time domain assuming an arbitrary two-dimensional conductivity distribution including an isolating air half-space. The method allows to carry out the computations for the subsurface only, because the air-earth interface is handled by an exact boundary condition. The spatial discretization is done with the finite element method, leading to a linear system of ordinary differential equations (ODE). We use state-of-the-art Krylov subspace methods for this ODE to advance a given initial electric field to selected times of interest. The presented theory is tested with some standard models and compared to a traditional finite difference time stepping implementation with respect to accuracy and efficiency. The results clearly demonstrate the superiority of the presented method in terms of run time given a comparable accuracy.

Keywords: Analytic boundary condition, finite element method, Krylov subspace, time domain, transient EM

2 Introduction

The transient electromagnetic (TEM) method has become a standard technique in geophysical prospecting during the past years. It is already in wide use, e. g., for the exploration of important resources like hydrocarbons, groundwater and minerals. One important aspect here is a reliable and computationally efficient simulation of the decaying electromagnetic field, which can be leveraged to get a better understanding of field behavior in complicated real-world settings as well as a building block in inversion schemes, that ultimately aim at resolving arbitrary conductivity structures from only a few well-placed measurements.

The predominant forward modeling technique in the literature is the finite difference time domain (FDTD) method, that was already introduced by Yee (1966). An explicit time-stepping technique, that already dealt with an isolating air half-space, was developed by Oristaglio and Hohmann (1984) for the two-dimensional case and later refined by Wang and Hohmann (1993) for three dimensions. Like for other explicit time-stepping methods, the size of the time steps, that the described Du Fort-Frankel scheme can stably perform, depends on the grid spacing and the lowest conductivity. Although the resistive air is already eliminated, thousands of time steps have to be performed although only a few dozen solutions are necessary to describe the decaying field.

The approach taken here is based on a finite element discretization, which allows for greater flexibility when modeling complicated conductivity structures. High accuracy is obtained with less effort compared to graded tensor product grids used with finite differences. It also helps in the construction of an analytic boundary condition, avoiding a few drawbacks the implementation by Oristaglio and Hohmann (1984) has. Contrary to the finite difference approach, the matrices resulting from the discretization are symmetric and, thus, allow for a wide range of efficient and state-of-the-art time integration techniques, which can exploit this property. One such family of time integrators is based on building a Krylov

subspace and extracting approximations to the matrix exponential function from said space, thus evaluating the sought electric field directly at a given time.

3 Theory

3.1 Governing Equations

Our governing equation derives from Maxwell's equations in the diffusive limit, with the constitutive relations used and the magnetic field already eliminated. Thus, we write

$$\nabla \times (\mu^{-1} \nabla \times \mathbf{e}) + \partial_t \sigma \mathbf{e} = -\partial_t \mathbf{j}^e \quad (1)$$

with

$\mathbf{e} = \mathbf{e}(\mathbf{x}, t)$	the electric field,
$\mu = \mu(\mathbf{x})$	the magnetic permeability,
$\sigma = \sigma(\mathbf{x})$	the electric conductivity, and
$\mathbf{j}^e = \mathbf{j}^e(\mathbf{x}, t)$	the impressed source current density.

We now restrict ourselves to the two-dimensional case (xz -plane) with infinitely long line sources perpendicular to this plane. Given these assumptions, we can express the electric field as

$$\mathbf{e}(x, y, z, t) = e(x, z, t) \mathbf{y} \quad (2)$$

where $\mathbf{y}$ is the unit vector along the y -axis and e a scalar function. Equation (1) then reduces to

$$-\nabla^2 e + \sigma \mu \partial_t e = -\mu \partial_t j^e. \quad (3)$$

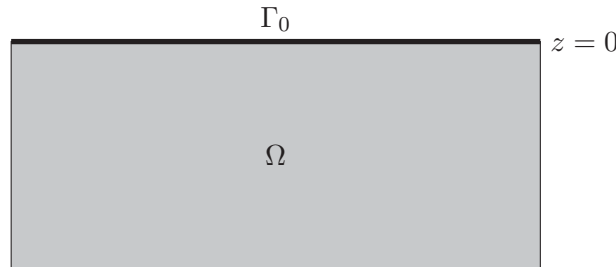


Figure 1: Computational domain with an arbitrary conductivity structure. On top is the air-earth interface (bold), the remaining boundaries are subsurface boundaries.

Our computational domain Ω (cf. Figure 1) is a rectangle and its top edge is aligned with the air-earth interface $\Gamma_0 = \{(x, z) : z = 0\}$ on which we impose an explicit boundary condition and perfect conductor boundary conditions on all other domain boundaries. It is important for these boundaries to be sufficiently far away from the sources, so that they don't distort the propagating field.

Our objective will be to compute the configuration of the electric field e_i at times t_i for $i \in \{1, 2, \dots, n\}$ given an initial field e_0 at t_0 . We use sources that are switched off at $t = 0$ and therefore the right hand side of (3) vanishes for $t > 0$.

3.2 Air-Earth Interface

We assume that e satisfies Laplace's equation in the isolating air half-space ($z < 0$)

$$-\nabla^2 e = 0. \quad (4)$$

Applying the Fourier transform in x to this equation yields

$$-\partial_{zz}\hat{e} + k^2\hat{e} = 0, \quad (5)$$

where $\hat{e}$ is the Fourier transform of e . After a few transformations (cf. Goldman et al. 1986) one can derive the exact boundary condition at $z = 0^+$

$$\partial_z e = T e + \mu \partial_t j^e \quad (6)$$

with the linear operator T being a linear convolution operator in x . This approach is also extensible to three dimensions and is described in detail in Goldman et al. (1989).

3.3 Spatial Discretization

Our next step towards a computable model is the discretization of the spatial domain. We focus on the finite element method, but also briefly mention finite differences because those are also used in the numerical examples.

Finite Elements To be able to apply the finite element method we need a variational formulation of our problem in (3), which is given by

$$\partial_t a(e, \psi)_\sigma + c(e, \psi) + \int_{-\infty}^{+\infty} \partial_z e(x, z = 0^+, t) \psi(x, 0) dx = 0 \quad (7)$$

with the bilinear forms a and c defined as

$$a(u, v)_\sigma = \mu \int_{\mathbb{R}_+^2} \sigma(\mathbf{x}) u(\mathbf{x}) v(\mathbf{x}) d\mathbf{x}, \quad (8)$$

$$c(u, v) = \int_{\mathbb{R}_+^2} \nabla u(\mathbf{x}) \cdot \nabla v(\mathbf{x}) d\mathbf{x} \quad (9)$$

where $\mathbb{R}_+^2 = \{(x, z) : z > 0\}$ denotes the lower half-space.

By using the exact boundary condition (6) we can write

$$\partial_t a(e, \psi)_\sigma + c(e, \psi) + \int_{-\infty}^{+\infty} T e(x, t) \psi(x, 0) dx = -\mu \int_{-\infty}^{+\infty} \partial_t j^e(x, t) \psi(x, 0) dx. \quad (10)$$

To be able to discretize this equation we need a computable expression for the following bilinear form

$$b(\phi, \psi) = \int_{-\infty}^{+\infty} T \phi(x) \psi(x) dx \quad (11)$$

and indeed, after some rearrangement in the Fourier domain, one obtains (Goldman et al. 1986)

$$b(\phi, \psi) = -\frac{1}{\pi} \int_{-\infty}^{+\infty} \int_{-\infty}^{+\infty} \log(|x - y|) \phi'(x) \psi'(y) dx dy. \quad (12)$$

The discretization of this integral on the boundary of a triangulation can now be easily computed.

The continuity of e allows us to use standard linear Lagrange elements on a triangulation of the computational domain. After discretizing the variational formulation we arrive at

$$M \partial_t \mathbf{e} = K \mathbf{e}. \quad (13)$$

Finite Differences For comparison, we also perform a finite difference discretization and get the ODE $\partial_t \mathbf{e} = A\mathbf{e}$ where A is a large sparse matrix representing the discrete curl-curl operator.

The following procedure, that has to be repeated in every time step, outlines the implementation of an exact boundary condition in this setting: Take the field data at the air-earth interface, interpolate it to an equidistant grid, perform an FFT to transform the data into the wave number domain, multiply every wave number with a constant (this would be a convolution in the space domain), revert the FFT and finally interpolate back to the graded grid to get the field values at a negative z , thus, inside the air half-space. Use the data resulting from this *classical upward continuation* in the usual finite difference stencil to compute the top layer in the next step.

A variation of this approach was also implemented for the numerical examples. It combines all of the above steps into a single (dense) matrix that has the same effect but can be evaluated more efficiently. This is called *matrix upward continuation* later on.

Details about the finite difference discretization can be found in Oristaglio and Hohmann (1984).

3.4 Time Integration Techniques

Krylov Subspace Methods Krylov subspace methods cover a big range of applications and are in use for several decades. What we want to focus on are Krylov subspace methods for the evaluation of matrix functions, like we can see in (13), where the function is the matrix exponential and the matrices come from the spatial discretization using finite elements. A nice side effect of the origin of those matrices is, that they are usually symmetric which many algorithms can benefit from.

Given a square matrix A of size $n \times n$, a vector $\mathbf{b}$ of length n and a suitable scalar function f , we can write

$$f(A)\mathbf{b} = p(A)\mathbf{b}. \quad (14)$$

with p a polynomial that Hermite-interpolates in the eigenvalues of A . We will be focusing on the exponential function and on rational Krylov subspaces, that are defined as follows

$$\mathcal{Q}(A, \mathbf{b}) := q_{m-1}(A)^{-1} \mathcal{K}_m(A, \mathbf{b}) \quad (15)$$

with $\mathcal{K}_m(A, \mathbf{b}) = \{\mathbf{b}, A\mathbf{b}, A^2\mathbf{b}, \dots, A^{m-1}\mathbf{b}\}$ and

$$q_{m-1}(z) = \prod_{\substack{j=1 \\ \xi_j \neq \infty}}^{m-1} (z - \xi_j). \quad (16)$$

Such subspaces are constructed, e. g., with the rational Arnoldi method to obtain an orthonormal basis of $\mathcal{Q}_m$ from which approximations to $f(A)\mathbf{b}$ can be computed with several procedures. We have to choose the poles ξ_j of the rational Krylov subspace. Their number and choice highly impact the quality of the approximations that can be extracted from the subspace. Luckily, there is a heuristic to determine good poles for the approximation of the exponential function, given a certain parameter interval (in our case the times we want to advance to) and accuracy requirements. These, and many more things concerning rational Krylov subspaces, are tackled in Güttel (2010), which is also a good starting place to dig deeper into the literature.

Looking more closely at (13) we see, that in order to apply a Krylov subspace method we need to remove the matrix M in front of the time derivative, e. g. by moving it to the right

$$\partial_t \mathbf{e} = M^{-1} K \mathbf{e}. \quad (17)$$

The inverse of the matrix doesn't have to be explicitly computed, but we have to solve a system of linear equations in every iteration step. This might be considered too expensive, so there is a technique

termed *mass lumping* which converts M into a diagonal matrix that can be easily inverted without distorting the solution too much.

However, even if we don't have to solve linear systems with the matrix M , we do have to solve shifted systems with the matrix K during the construction of the rational Krylov subspace. Therefore, an efficient solver is also an important ingredient. We will see in the numerical examples what options we have.

Time Stepping On the other hand there are simple, yet relatively efficient explicit time stepping techniques. One of them is the well-known Du Fort-Frankel method which is implemented here to perform the integration in time for the finite difference discretization.

4 Numerical Examples

All computations were carried out on a machine with four AMD Opteron 8380 quad cores, running at 2.5 GHz with a total of 128 GiB of RAM. The algorithms were implemented in pure MATLAB code running under MATLAB 2008b, with some finite element related tasks (nothing time-critical) performed by COMSOL 3.5a, both running in a 64-bit Linux environment. To generate reproducible timing we restricted ourselves to one computational thread that was explicitly pinned to one of the cores.

4.1 Model

The considered models are basically those from Oristaglio and Hohmann (1984), all based on graded tensor product grids. We extended the mesh slightly so as not to get a distortion from the boundaries for late times. The resulting grid has 236×88 cells, is 12800 m wide and 5255 m deep, with grid spacings ranging between 10 m and 240 m. To make the computations comparable with the finite difference code we decided to use exactly the same grid for the finite element code. We simply subdivided each cell into two triangles, however, the number of unknowns or degrees of freedom and their locations are identical. Technically, this is of course not necessary and a properly adapted unstructured grid would certainly yield even better results at lower cost.

The initial field e_0 at $t_0 = 10^{-6}$ s is due to two line sources at positions $(x, z) = (-500 \text{ m}, 0 \text{ m})$ and $(x, z) = (0 \text{ m}, 0 \text{ m})$, the former being negative while the latter is positive with a unit source current strength. It is computed analytically for a homogeneous half-space that corresponds to the background conductivity of the model.

We have looked at two models with different conductivity structures. See Figure 2 to get an approximate idea of their location inside the grid. The background has a resistivity of $300 \Omega\text{m}$ and the conductor (red) $0.3 \Omega\text{m}$.

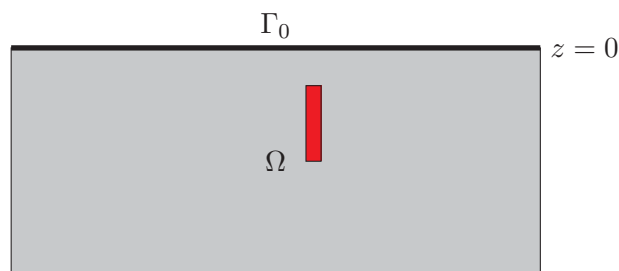


Figure 2: Schematic view of the computed models. The models are the homogeneous half-space and a conductor (red) inside a homogeneous half-space. The conductor is located at the position $290 \text{ m} \leq x \leq 310 \text{ m}$, $100 \text{ m} \leq z \leq 400 \text{ m}$ in the grid.

4.2 Methods and Setup

Independently of the method of choice we start at the time $t_0 = 10^{-6}$ s and integrate from there till we reach $t_n = 10^{-2}$ s. We seek to compute the configuration of the electric field at 10 logarithmically equidistant times per decade.

For the finite difference code we mostly stick to the implementation by Oristaglio and Hohmann (1984), with the only option to perform the upward continuation in the traditional way, which is a significant expense per time step, or we can decide to precompute the matrix formulation of this upward continuation and to apply it in every time step.

For the finite element method we have no choice regarding the exact boundary condition. However, we can choose between different variants of the rational Krylov method. For all of them we have a total of 25 poles, which should be close to optimal for this application. For the first 24 poles we alternate between $2.7826 \cdot 10^5$ and $0.0242 \cdot 10^5$. The last pole is set to infinity.

We have four implementations, that we included in this test. Two of them use mass lumping to reduce the mass matrix to a diagonal matrix. Furthermore, for each group we can either call the standard MATLAB solver or we can exploit the fact that we have multiple identical poles for which we have to solve with different right hand sides, but identical matrices. We do this by computing a sparse LU factorization (we use UMFPACK for this) for every unique pole and then leverage that factorization to speed up the solves during the construction of the rational Krylov subspace.

4.3 Results

We first look at the computation times that are pictured in Figure 3 and listed in Table 1. The times shown here are for the homogeneous half-space model, but aren't significantly different for the other models since the grid and the minimal conductivity are identical. Thus, we can use these numbers to judge the acceleration we can get by using our method.

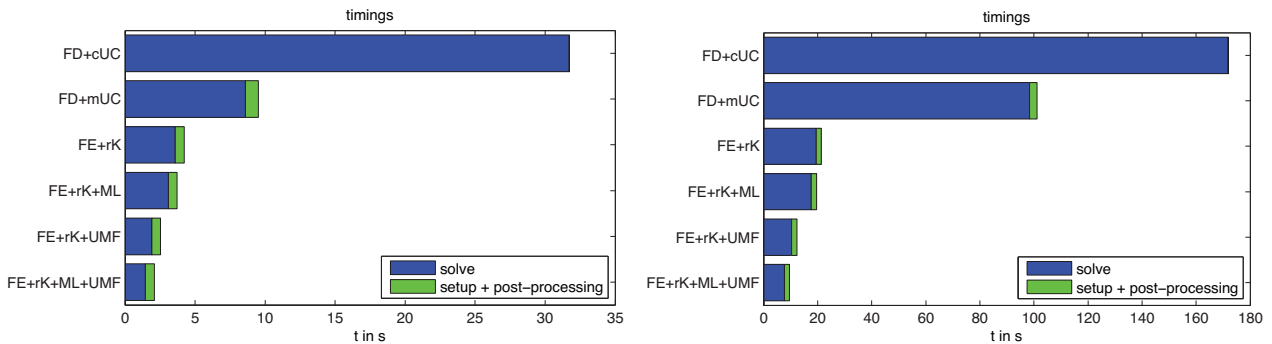


Figure 3: Computation times split into solve and setup/post-processing step, cf. also Table 1. On the *left* are the computation times for the default grid and on the *right* for the regularly refined version of this grid with approximately four times as many nodes and degrees of freedom. The following abbreviations were used: classical upward continuation (cUC), matrix upward continuation (mUC), rational Krylov (rK), mass lumping (ML), UMFPACK solver (UMF).

As is clearly visible, all FE-based methods are significantly faster than their finite difference counterparts, sometimes even 22 times faster.

Looking at the transients (cf. Figure 4) at some select points inside the mesh, we see that although these newer methods are so much faster, we don't really sacrifice accuracy. In fact, in many cases the finite element solution is more accurate than the one obtained by finite differences.

We conclude our numerical examples with showing some cross-sections (cf. Figure 5), that are nothing but a few snapshots of an animation that is just not suitable for this medium, but nevertheless gives a coarse idea of how the electric field propagates with time.

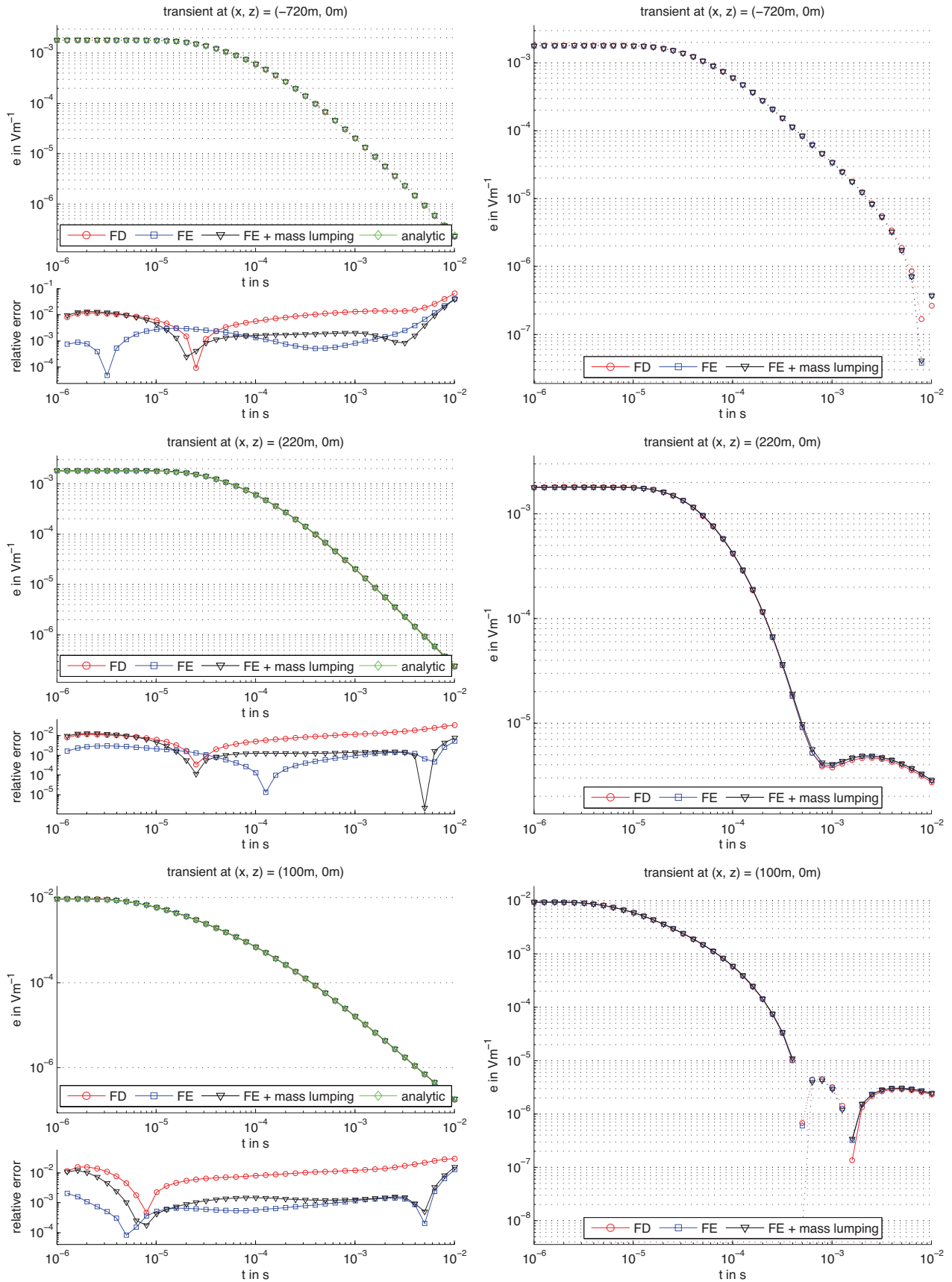


Figure 4: Transients of the homogeneous half-space model with relative errors (left) and the model with a conductor (right). Negative field values are denoted by dashed lines while positive values are drawn with a solid line. We can see that we have a very good agreement between the FD and FE solution. When comparing against an analytical solution in the homogeneous half-space, we see that the relative errors are often lower than those of the FD solution, despite the lower numerical effort.

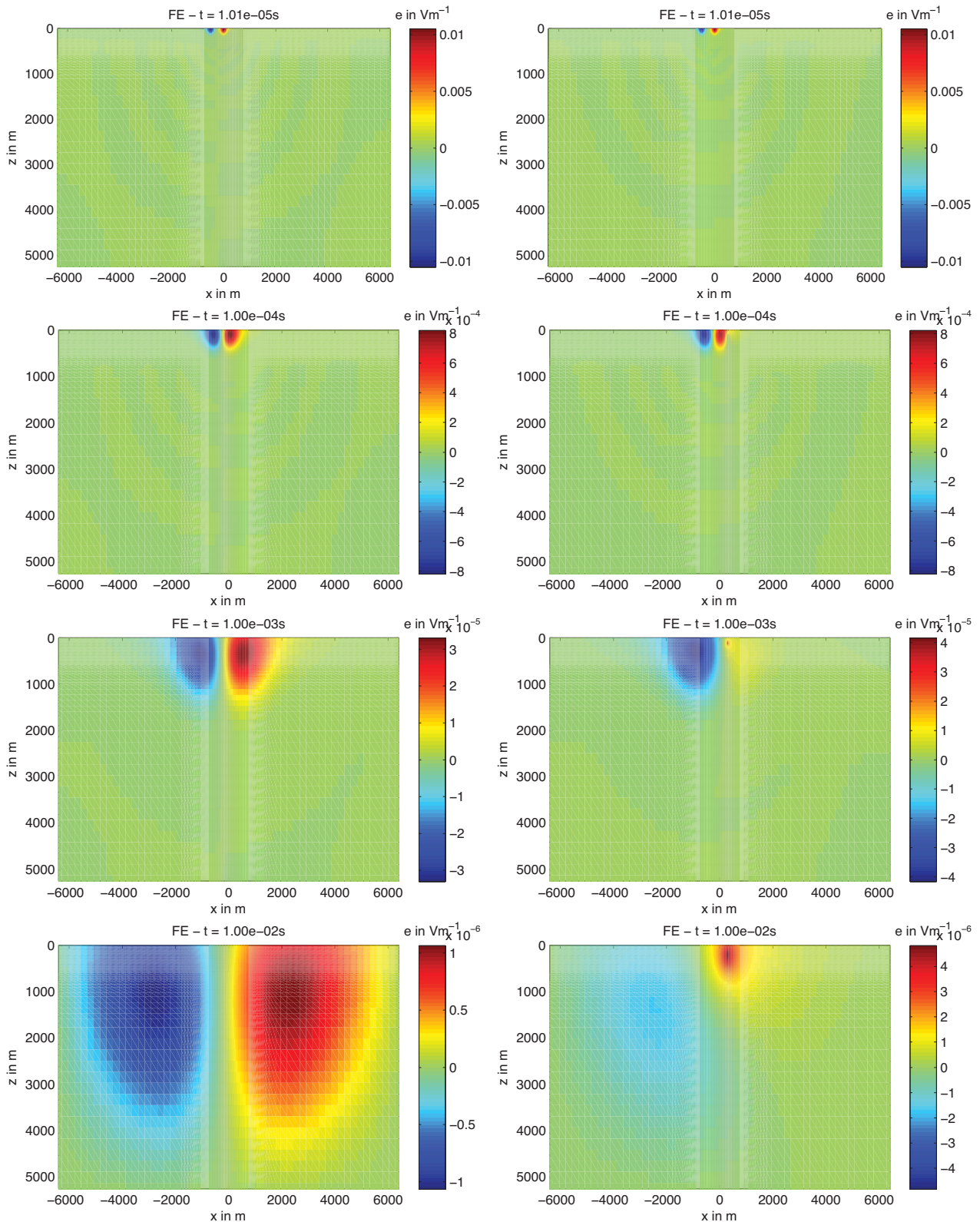


Figure 5: Cross-sections of a homogeneous half-space model (left) and the model with a conductor (right). These cross-sections were plotted to give a visual idea of how the fields advance in time. The common initial configuration was omitted. We decided to plot the results of the FE simulation, but there was hardly any visible difference compared to the FD solution. On the right we can nicely see how the field gets *captured* inside the high conductivity structure and creates an anomaly compared to the homogeneous case.

Method	Default grid		Refined grid	
	Solve	Other	Solve	Other
FD, time stepping, classical upward continuation	31.7 s	< 0.1 s	171.8 s	< 0.1 s
FD, time stepping, matrix upward continuation	8.6 s	0.9 s	98.4 s	2.7 s
FE, rational Krylov	3.6 s	0.6 s	19.4 s	1.9 s
FE, rational Krylov, mass lumping	3.1 s	0.6 s	17.7 s	1.9 s
FE, rational Krylov, UMFPACK	1.9 s	0.6 s	10.4 s	1.9 s
FE, rational Krylov, mass lumping, UMFPACK	1.4 s	0.6 s	7.7 s	1.9 s

Table 1: Computation times for the default grid and one, that was obtained from this with regular refinement. The column *Solve* denotes the time spent for performing the integration in time, while the column *Other* lists the time spent in setting up the boundary condition and the computation of field values from the solution vectors in case of the FE method. Other setup times are not shown. The advantage of FE-based methods can be clearly seen.

5 Conclusions

We were able to leverage several state-of-the-art techniques to create a combined efficient forward modeling code that is significantly faster than traditional codes. We have also seen that we don't have to make sacrifices regarding the accuracy of those computations. This was already achieved by using the mesh from the finite difference discretization, which was helpful for comparison, but which also has many limitations due to its regular structure. An unstructured grid—properly adapted to the problem—could be used to further increase accuracy or speed.

We are working on creating a similar framework for the three-dimensional time domain TEM problem. All of the methods and tools are already existing or have a straightforward extension to 3D. Given the results from 2D we expect an even greater speed-up when applying this to three dimensions.

6 Acknowledgments

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TEM with anomalous diffusion in fractal conductive media

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Abstract

The transient electromagnetic response of an inductive loop source over a half-space with fractal characteristics is simulated. The conductivity of the ground has a spatial distribution, which is described by a roughness parameter. The roughness can be related to the fractal dimension and controls the transient decay of the signal. Asymptotic limits are observed for the early and late time behaviour.

The work is based on a paper by Everett (2009), which has been reviewed and the aspect of asymptotic limits in time has been further investigated. Contrary to that paper, here a relationship between roughness and the electromagnetic decay at early and late time has been deduced. The transient decay for normal diffusion is $t^{-5/2}$ and for anomalous diffusion the decay is slower. The new power law for anomalous diffusion has been proven by theoretical analysis and has been verified by numerical experiments.

The numerical evaluation of the inverse Laplace transform with the method of the fast Hankel transform are excellent in numerical accuracy and the method with the Gaver-Stehfest algorithm is not sufficient to estimate the power law decay at late times.

Anomalous Diffusion

The concept of anomalous diffusion is a useful approach for the description of diffusion process and transport dynamics in complex systems. The fractional equations are derived asymptotically from basic random walk models and become a complementary tool for handling non-exponential relaxation patterns.

For transient electromagnetic diffusion Everett starts this concept with a generalized Ohm's law (Everett 2009, Weiss and Everett 2007)

$$j = \sigma E \rightarrow j = \sigma_{\beta} * E \equiv \int_0^t \frac{\sigma_{\beta} E(\tau)}{(t - \tau)^{1-\beta}} d\tau$$

the parameter $\sigma_{\beta} \sim \sigma t^{\beta}$ describes the generalized electrical conductivity and is appropriate for the anomalous diffusion coefficient. The Ohm's law becomes a convolution between the generalized conductivity and the electric field E . The roughness parameter can vary $0 \leq \beta \leq 1$.

After applying Ampere's and Faraday's law to the generalized current density, we get the fractional diffusion equation

$$\nabla^2 E = \mu_0 \frac{\partial}{\partial t} \sigma_\beta * E = \mu_0 \sigma_\beta {}_0 D_t^{1-\beta} E(t)$$

where now the fractional derivative or Riemann-Liouville operator is introduced (Metzler and Klafter 2000)

$${}_0 D_t^{1-\beta} E(t) = \frac{1}{\Gamma(\beta)} \frac{\partial}{\partial t} \int_0^t \frac{E(\tau)}{(t-\tau)^{1-\beta}} d\tau$$

and Γ is the Gamma function which serves as normalization constant.

The fractional diffusion equation will be solved in the Laplace domain and the transformation of the operator yield to

$$L\{{}_0 D_t^{1-\beta} E(t)\} = s^{1-\beta} \tilde{E}(s)$$

with the complex Laplace variable $s = i\omega$ (Abramowitz & Stegun 1964).

The fractional diffusion equation becomes now a simple expression

$$\nabla^2 \tilde{E} = s^{1-\beta} \mu_0 \sigma_\beta \tilde{E}(s)$$

This differential equation can be solved straightforward with the standard methods as used for the electromagnetic 1-D formulation in a layered medium.

Electromagnetic response of a loop over rough half-space

Everett (2009) has used a separate horizontal loop configuration for transmitter and receiver as it is typically used for TEM. The transmitter loops usually a square loop can be represented by a circular loop with equivalent area. For the separate loop configuration the time derivative of the vertical magnetic field is measured outside the transmitter loop.

In frequency domain the vertical magnetic field is presented as Hankel Integrals

$$\tilde{h}_z(s) = Ia \int_0^\infty \frac{\lambda^2}{\lambda + \sqrt{k_\beta^2 + \lambda^2}} J_1(\lambda a) J_0(\lambda r) d\lambda.$$

The integration is over the spatial wavenumber λ and I is the current in the transmitter, a the transmitter loop radius, r the distance to the receiver point, J_0 and J_1 are Bessel functions of order 0 and 1 and k_β the fractional wavenumber. The integral can be solved analytically for an infinitesimally magnetic dipole, therefore the limit of the first order Bessel function for small radius is taken into account (Abramowitz & Stegun 9.1.10)

$$J_1(\lambda a) \xrightarrow{a \rightarrow 0} \frac{\lambda a}{2}.$$

The response of a vertical magnetic dipole in frequency domain over a rough conductive media is

$$\tilde{H}_z(s) = \frac{m}{2\pi r^3} \frac{1}{u^2} \left[-9 + (9 + 9u + 4u^2 + u^3) e^{-u} \right]$$

with the fractional induction number $u = k_\beta r = (s^{1-\beta} \mu_0 \sigma)^{1/2} r$.

Asymptotic limits in frequency and time domain

For normal diffusion with roughness equal zero the transient response can be also given in analytical expression but for a general fractional induction number with the roughness parameter β the transformation to time domain has to be done by numerical techniques. Only the asymptotic limits for the high and low frequency limit can be calculated in closed form. For high frequency and early time we get

$$\tilde{H}_z^{h.f.}(s) = -\frac{m}{2\pi r^2} \frac{9}{u^2} \quad \leftrightarrow \quad H_z^{e.t.}(t) = -\frac{m}{2\pi r^3} \frac{9}{\mu_0 \sigma r^2} \frac{1}{\Gamma(1-\beta)} t^{-\beta}.$$

The low frequency can be developed in a series

$$\tilde{H}_z(s) = \frac{m}{2\pi r^3} \frac{1}{2} \left[-1 - 2 \sum_{n=4}^{\infty} \frac{(n-1)(n-3)^2}{n!} (-u)^{n-2} \right].$$

The first 3 terms are important for the late time behavior

$$\tilde{H}_z(s) = \frac{m}{2\pi r^3} \frac{1}{2} \left[-1 - \frac{1}{2} u^2 + \frac{4}{15} u^3 \right] - + \dots$$

In time domain the first term is a δ -function without influence on late time decay. The second term is responsible for the late time behavior in a rough medium

$$L^{-1}(s^{1-\beta}) = \begin{cases} \delta'(t) & \text{for } \beta = 0 \\ -\frac{\Gamma(\beta)}{\beta-1} t^{-(2-\beta)} & \text{for } \beta > 0 \end{cases}$$

The third term describes the classical -5/2 decay response in a non-fractional medium and for a fractional medium this term decays faster than the previous second term.

The late time response for the magnetic field can be summarized

$$\dot{H}_z^{l,t}(t) = \frac{m}{2\pi r^3} \begin{cases} \frac{(\mu_0 \sigma)^{3/2}}{10\sqrt{\pi}} t^{-5/2} & \text{for } \beta = 0 \\ \frac{\mu_0 \sigma_\beta}{2} \frac{\Gamma(\beta)}{\beta-1} t^{-(2-\beta)} & \text{for } \beta > 0 \end{cases}$$

The cause for the heavy tailed decay response in rough geological media can be explained by the additive second term which yield to a continuous transition from non-fractional to fractional diffusion as is shown in figure 1 and 2.

Numerical inverse Laplace transform

Numerical methods are applied for the transformation to time domain. Since the transient is a real and causal function, the inverse Fourier or Laplace transform can be calculated by a sine-transform.

$$\dot{h}_z(t) = -\frac{2}{\pi} \int_0^\infty \text{Im}\{\tilde{h}_z(\omega)\} \sin(\omega t) d\omega$$

In this expression I have already considered that the current step in the transmitter described by $1/i\omega$ and the time derivative of the receiver coil by a multiplication with $i\omega$ cancel each other. Since for diffusion processes the kernel function is a smooth function, the technique of the Fast Hankel Transform can be applied. The sine function is expressed by Bessel function with fractional order $1/2$

$$J_{1/2}(x) = \sqrt{\frac{2}{\pi x}} \sin(x).$$

The fast Hankel transform is a well know technique for calculating the transient response, I used 250 filter coefficients, 15 /decade, calculated with the program by Christensen (1990).

Everett (2009) has chosen the Gaver-Stehfest algorithm for his investigation. This method is also good for diffusion processes and has been successfully applied. It is a favoured method in hydrology, because it needs only real arithmetic and the Laplace variable s is considered as a real variable. Knight and Raiche (1982) introduced this technique to the electromagnetic community. Stehfest (1970) published an Algol routine, which can be straight forward translated to other computer languages and Everett (2009) has tabled the coefficients for several total number of coefficients.

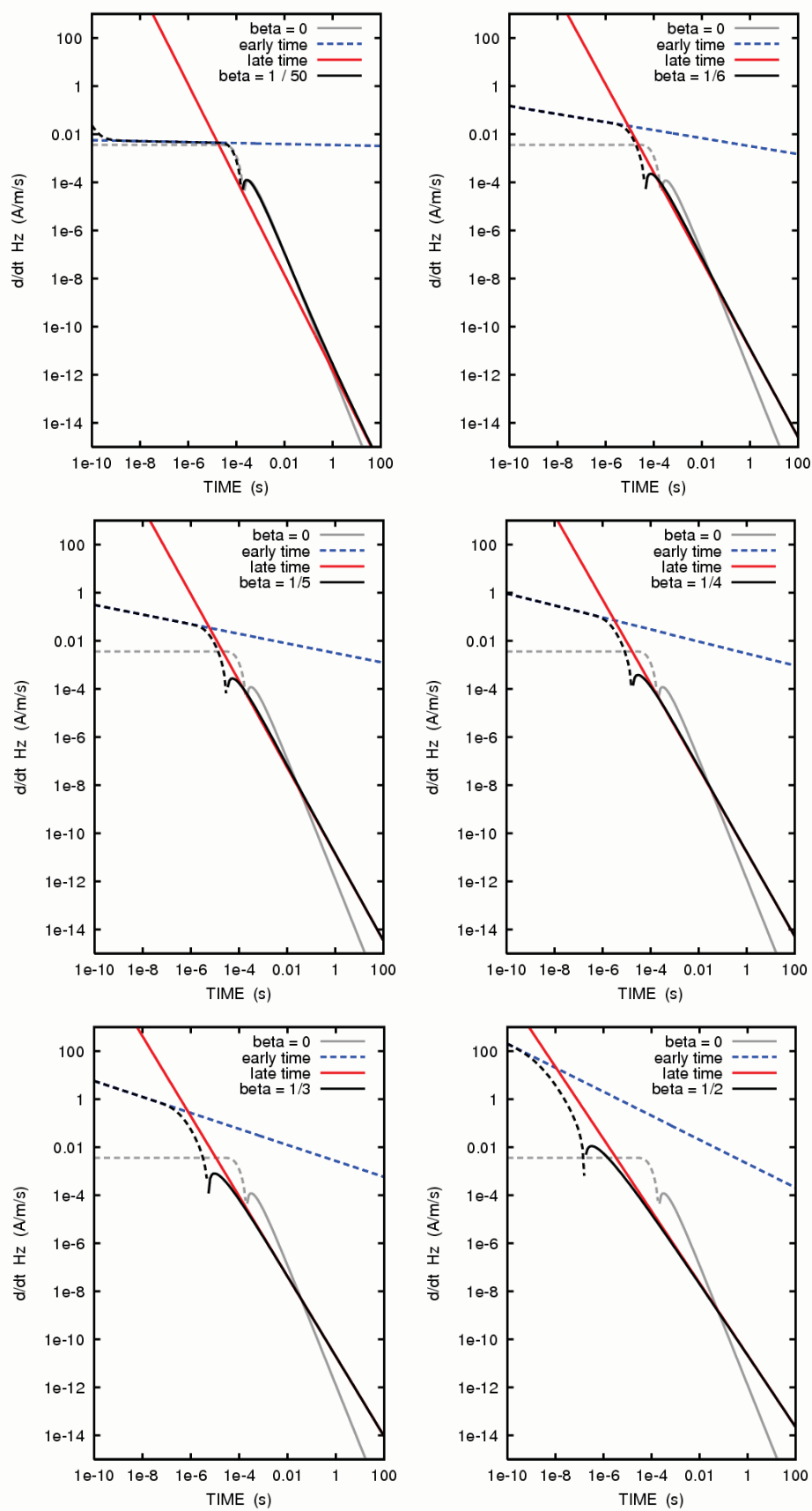


Figure 1: Transient loop-loop response with early and late time approximation over half-space with different roughness in conductivity, the offset is 100 m , $\sigma = 0.1 \text{ S/m}$.

The disadvantage of the Gaver-Stehfest algorithm is that the numerical accuracy cannot generally be increased by increasing the number of coefficients. The accuracy is depending on the accuracy in number of digits of the kernel function and the number of digits of the machine precision. Stehfest recommended 8 coefficients for single precision and 18 coefficients for double precision. Everett used 18 coefficients. In my experiments I have found out that 12 is an optimal number for electromagnetic application.

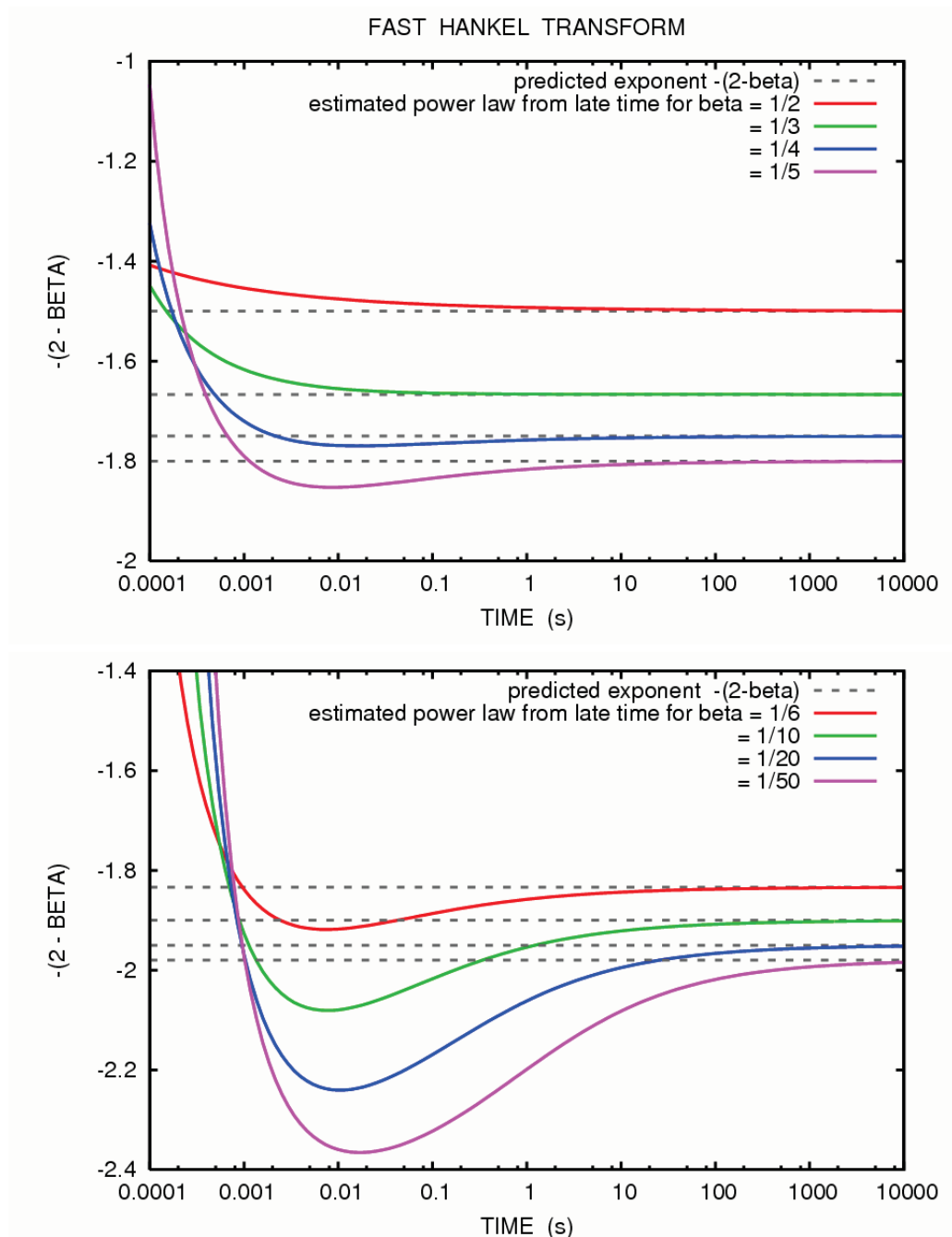


Figure 2: Numerical evaluation of the exponent in the transient decay and theoretical asymptotic limit for different roughness

Results

For the numerical experiment I used the same model as Everett, beside that here a vertical magnetic dipole instead of a horizontal circular loop is used. The base conductivity of the half-space is 0.1 S/m with combination of various roughness parameter β has been applied.

Figure 1 shows the transient as induced voltage due to an step response of the transmitter current for different roughness paramters. The large time range has been chosen to demonstrate the early and late time behavior in comparison with the asymptotic limits. For real measurements the time range will be much smaller. All transients - shown here - are calculated with the fast Hankel transform. As reference the transient of homogenous half-space with normal diffusion is also shown with grey lines. For separate loop configuration the transient shows a sign change. The negative values are shown with dashed lines and positive with solid lines. The asymptotic limits agree very well with theoretical prediced limits at eary and late times shown here as straight lines. The early time behaviour is for practical pupose of minor relevance, because in real meausrements the early time is influenced by system response of the system as the ramp for the current step off. So that in the data the straight line will not be visible. But the late time behaviour will be visible if the late time data can be measured and the data quatilty is sufficient.

To analyze how accurate the power law decay can be estimated from the calculated transients especiale for low roughness numbers, the transients are calculated up to extrem late times and the power law is determined by taking the numerical derivative $d \ln H(t) / d \ln t$. The result is shown in figure 2.

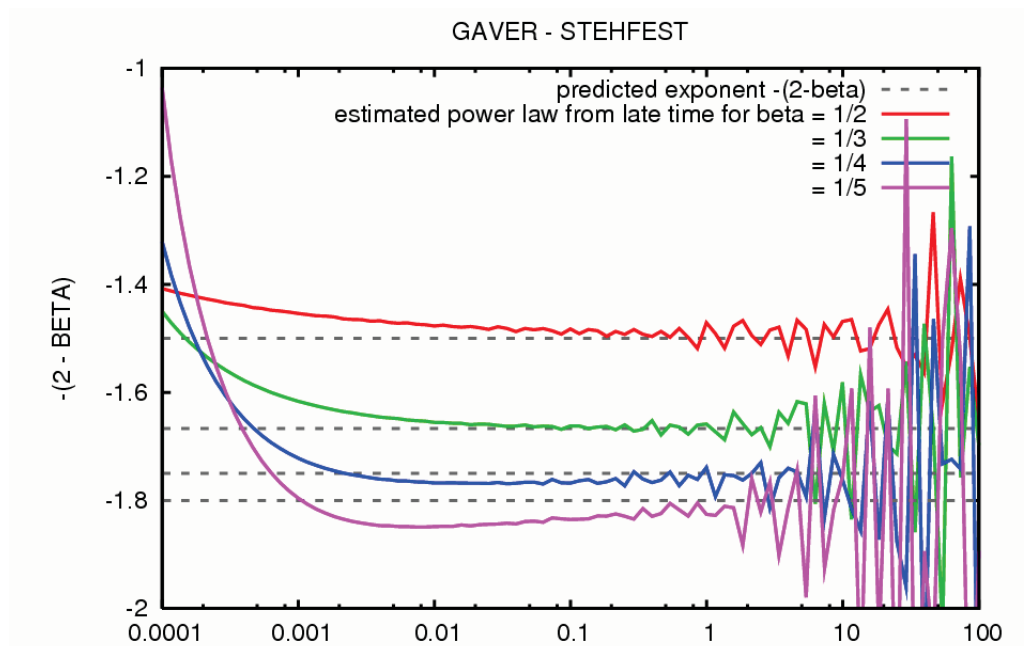


Figure 3: Numerical evaluation of the exponent in the transient decay and theoretical asymptotic limit for different roughness calculated with Gaver-Stehfest alorighm

The graphs show the excellent accuracy of the fast Hankel transform, even for small roughness numbers, and indicates the smooth transition, when the roughness approaches zero. The influence of the anomalous diffusion moves to later times for decreasing roughness.

For very low roughness there will be a time range showing almost normal decay $t^{-5/2}$ and then goint to the predicted power law decay at extrem late time – approaching infinity. Responsible for this behavior is the second term of the low frequency approximation which is added.

The Gaver-Stehfest algorithm is not sufficient to estimate the power law decay. I tried different ways of programming, e.g to consider numerical accuracy the Laplace transform is done before the Hankel transform over the spatial wave number as recommended by Knight and Raiche (1982). The best results are shown in figure 3, achieved with 12 coefficients and using the analytical response for a vertical magnetic dipole in Lapace domain. Notice that the time range is shorter but still 3 decades more than in Everett's paper.

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Analysis of seafloor marine EM data with respect to motion-induced noise

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Abstract

Any appreciable movement of sea water induces an electromagnetic field, which acts as noise for marine controlled source electromagnetic (mCSEM) data. Noise in seafloor mCSEM data is considered small, but since the characteristic reservoir signal is also small, the understanding and possible removal of noise may be essential to increase the number of possible target reservoirs.

A power spectral density plot of the time-series extracts the power of each frequency contribution, but the shortcoming is that the time information is lost. To get time information corresponding to each frequency, we plotted a spectrogram which shows how the spectral density of a signal varies with time. The collective use of these three plots, i.e. time series, power spectral density and spectrogram, helps to analyse the information concealed in the time-series. In a measured signal-free mCSEM data, we are able to identify various oceanic features like microseisms, swell etc, which play a significant role in inducing an electric field.

Key words: mCSEM, Swell, Microseisms

1. Introduction

Marine controlled source electromagnetic (mCSEM) data is generally contaminated by some unwanted electromagnetic (EM) signals ambient at the seafloor. Understanding and possible removal of the noise is essential as technology is advancing from deep ocean to shallow ocean, where the noise is much more effective.

In general, there could be two possible sources of oceanic noise production. Internal oceanic processes or any other external influences like ionospheric and magnetospheric current systems. Faraday (1832) reported that any appreciable movement of seafloor and sea water by induction generates electromagnetic signals with in the ocean. Motional induction got more attention after world war II, with the detailed oceanic induction study by many authors like Sanford (1971), Podney (1975), Chave and Cox (1982), Chave and Luther (1990). Similarity in all these studies is that they all

studied the theoretical aspect of motional induction occurring due to various movements in the ocean. Experimental evidence was provided by Webb and Cox (1986), who made a nice comprehensive study about the seafloor microseisms by observing pressure and electric field spectra at the seafloor.

Wave progressions and oceanic processes are, usually, a function of the regional and local conditions. A variety of motional processes like swell, ripples, internal waves, microseisms, hum etc generate a range of EM noise at the seafloor, which varies from place to place. External, ionospheric and magnetospheric, current systems also contribute EM noise, but they are limited to the lower frequency range due to the conductive filtering of the ocean. Characteristics of the noise are poorly known till now, as only few studies are made. The present study is dedicated to the understanding of the noise characteristics at the sea floor.

For the study, a marine EM data set recorded at the seafloor in the absence of a transmitter, is used. Recording is made for two horizontal components of electric fields i.e. E_x and E_y , which are perpendicular to each other. Visibly, E_x and E_y time series data appears different in their pattern and amplitudes despite the same oceanic environment. This suggests that a comprehensive study on the directional characteristics of the events/sources of noise could play a decisive role in identifying the noise sources.

The oscillatory occurrence, due to a disturbance, can be comfortably visualised in frequency domain. The power spectral density (PSD) of each frequency contribution helps in marking the power of the individual sources. But the shortcoming of the PSD calculation is that if the source of the characteristic frequency is not known a priori, it is difficult to conclude about the source environment with the frequency information only. In this case, time information together with the PSD may help at least to distinguish between localised and ambient sources, which itself is of significant use. To overcome the PSD shortcoming to retain time information corresponding to each frequency, spectrograms are plotted, which show how the spectral density of a signal varies with time.

In the signal-free mCSEM data, we have observed many features which are new for mCSEM studies. The spectrogram presents clear prints of oceanic features like microseisms and swell, which play a significant role in inducing an electric field at the seafloor.

2. Marine Controlled Source Electromagnetism (mCSEM)

A mCSEM data acquisition methodology is shown in Figure 1. In practice, an EM transmitter is towed close to the seafloor to maximize the coupling of electric and

magnetic fields with seafloor rocks. These fields are recorded by instruments deployed on the seafloor at some distance from the transmitter. Details about the data acquisition practice in both time and frequency domain of mCSEM is well documented in the article by Constable and Srnka, (2007).

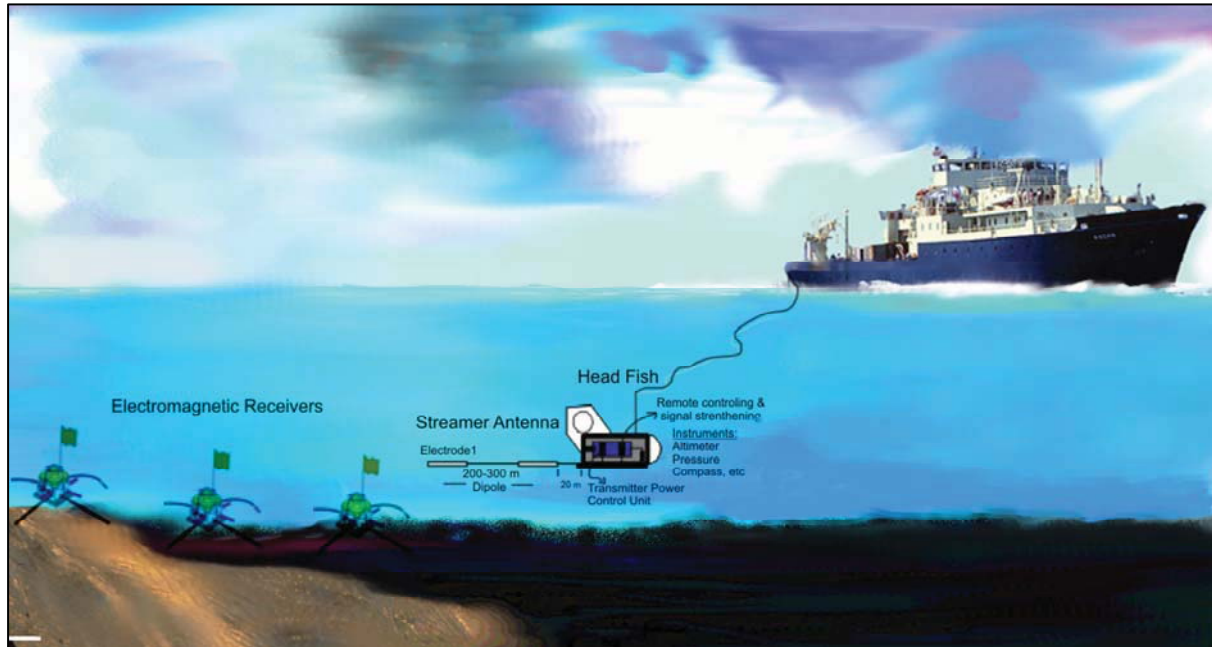


Figure 1: Schematic diagram showing field layout for mCSEM sounding.

3. Sources of mCSEM noise

The noise convolved in mCSEM data is capable of masking significant marine information. Broadly, two origins are expected as noise sources in mCSEM data:-

- a) **External origin:** The magnetospheric and ionospheric current system induces natural electric and magnetic (EM) fields in the conductive formations. The induced fields are signal for the magnetotelluric (MT) technique but noise for mCSEM data. The conductive ocean filters the higher frequencies. The skin depth (δ) is given by

$$\delta \approx 503 \sqrt{\frac{1}{\sigma f}}$$

This implies,

$$f \approx 76,000 / \delta^2 \quad \text{for } \sigma = 3.3 \text{ S/m}$$

Therefore, at the ocean floor the external origin fields effectively contribute frequencies less than or equal to $f = (76,000 / D^2)$, where D is the depth of ocean floor in meters and f is given in (1/s). As example for an ocean of depth 500 m, the external field will contribute frequencies less than 0.3 Hz at the sea floor. This

suggests that the frequencies higher than 0.3 Hz could be by internal oceanic origin.

- b) **Internal origin:** The dynamics of oceanic water with in the ambient geomagnetic field induces electric and magnetic field in the ocean. A frequency range of the fields is generally $0.3 < f \text{ (Hz)} < 16$ (Dahm et. al., 2006) excluding long period waves like tidal waves, tsunami, seiches and storm surge etc.

Other than these noise sources, there are processes like earthquakes, volcanoes, manmade explosions etc occurring at the vibrant ocean floor, which add noise in mCSEM data.

4. Motional induction

The ocean is an electrically conducting fluid containing the ionic charge particles. Moving charge particles in the ambient geomagnetic field experience a deflective Lorentz force. If $\vec{v}$ is the velocity of charge particle q moving in the geomagnetic field $\vec{B}$, then the Lorentz force is-

$$\vec{F}_L = q(\vec{v} \times \vec{B}) \quad (1)$$

The charge q experiences the deflecting force $\vec{F}_L$ because of the action of an electric field which we call Lorentz electric field ($\vec{E}_L$),

$$\vec{F}_L = q\vec{E}_L \quad (1-a)$$

$$\vec{E}_L = \vec{v} \times \vec{B} \quad (2)$$

The field $\vec{E}_L$ generates a secondary electric field $\vec{E}$, mainly by two processes: i) Galvanic process, ii) Inductive process (Bhatt et al, 2010). For a stationary frame of reference, the current density in the Ohms law is given by

$$\vec{J} = \sigma(\vec{E} + \vec{E}_L) = \sigma(\vec{E} + \vec{v} \times \vec{B}) \quad (3)$$

Here, σ is the conductivity of the ocean, $\sigma\vec{E}$ is the current term generated by both a galvanic and an inductive process and $\sigma(\vec{v} \times \vec{B})$ is the source current term for the motional induction case. Under the quasi-static approximation, the set of Maxwell's equation to be solved for the motional induction case is

$$\begin{aligned} \nabla \times \vec{H} &= \sigma(\vec{E} + \vec{v} \times \vec{B}_0) \\ \nabla \times \vec{E} &= -\mu_0 \partial_t \vec{H} \end{aligned} \quad (4-a,b)$$

Simplification of the source $\sigma(\vec{v} \times \vec{B})$ in terms of component J_x, J_y and J_z can be written as,

$$\begin{aligned} J_x &= \sigma(E_x + v_y B_z - v_z B_y) \\ J_y &= \sigma(E_y + v_z B_x - v_x B_z) \\ J_z &= \sigma(E_z + v_x B_y - v_y B_x) \end{aligned} \quad (5)$$

The non-divergent requirement condition of the current (i.e. $\nabla \cdot \vec{J} = 0$) suggests that the current will close its loop with in the earth rather than closing in the ocean. Therefore, the horizontal electric currents (i.e. J_x and J_y) in the ocean are more effective than the vertical currents (i.e. J_z) as the horizontal extent of ocean is much larger than the vertical depth. In mCSEM, the data is recorded at the seafloor, where normally the vertical velocity component (i.e. v_z) is very small. Therefore, terms $v_z B_y$ and $v_z B_x$ is negligible. Above argument simplifies (4),

$$\begin{aligned} J_x &= \sigma(E_x + v_y B_z) \\ J_y &= \sigma(E_y - v_x B_z) \end{aligned} \quad (6)$$

Using (6), simplification of (4-a) gives

$$\begin{aligned} E_x &= -\left(\frac{1}{\sigma} (\partial_z H_y - \partial_y H_z) + v_y B_z \right) \\ E_y &= +\left(\frac{1}{\sigma} (\partial_z H_x - \partial_x H_z) + v_x B_z \right) \end{aligned} \quad (7)$$

Where the first term is the induced field and second term is the source term. Normally, the area used for mCSEM data acquisition is very small to observe horizontal variation in the vertical magnetic field H_z . Therefore, term $\partial_y H_z$ and $\partial_x H_z$ are negligible. Finally, we have a simplified equation,

$$\begin{aligned} E_x &= -\left(\frac{1}{\sigma} \partial_z H_y + v_y B_z \right) \\ E_y &= +\left(\frac{1}{\sigma} \partial_z H_x + v_x B_z \right) \end{aligned} \quad (8)$$

Noticeably, the horizontal particle motion (i.e. v_x and v_y) is source for the induction of the horizontal electric field (i.e. E_x and E_y).

5. The Data

We have analysed horizontal component electric field (i.e. E_x and E_y) data recorded at 500 m depth on the ocean floor. The recording is made in the absence of transmitter current to understand the oceanic background noise. Time series is shown in Figure 2. It is evident that the strength and pattern of E_x and E_y are different.

In equation (8), the first term is much smaller than the second term (i.e. $(1/\sigma)\partial_z H_{y|x} \ll v_{y|x} B_z$), suggesting x-and y- component of velocity principally controls the strength of the E_y and E_x component of the electric field respectively. Evidently, the strength of E_x is higher than E_y (Figure 2). This suggests that the wave velocity in the y-direction will be higher than the x-direction (i.e. $v_y > v_x$). Further, as normally the surface wave moves towards the coast. Accordingly, the velocity component pointing towards the coast has higher velocity than the other horizontal component. The present data is recorded with a setting that the y-component of the receiver points towards the coast and constructs an angle of approx. 55° with it. Thus, the data acquisition setting as well favours for $v_y > v_x$. In general, electric field measurements characterize an average motion over the length of the antenna. Therefore electric field components can be used to derive information about the average velocity of the movements.

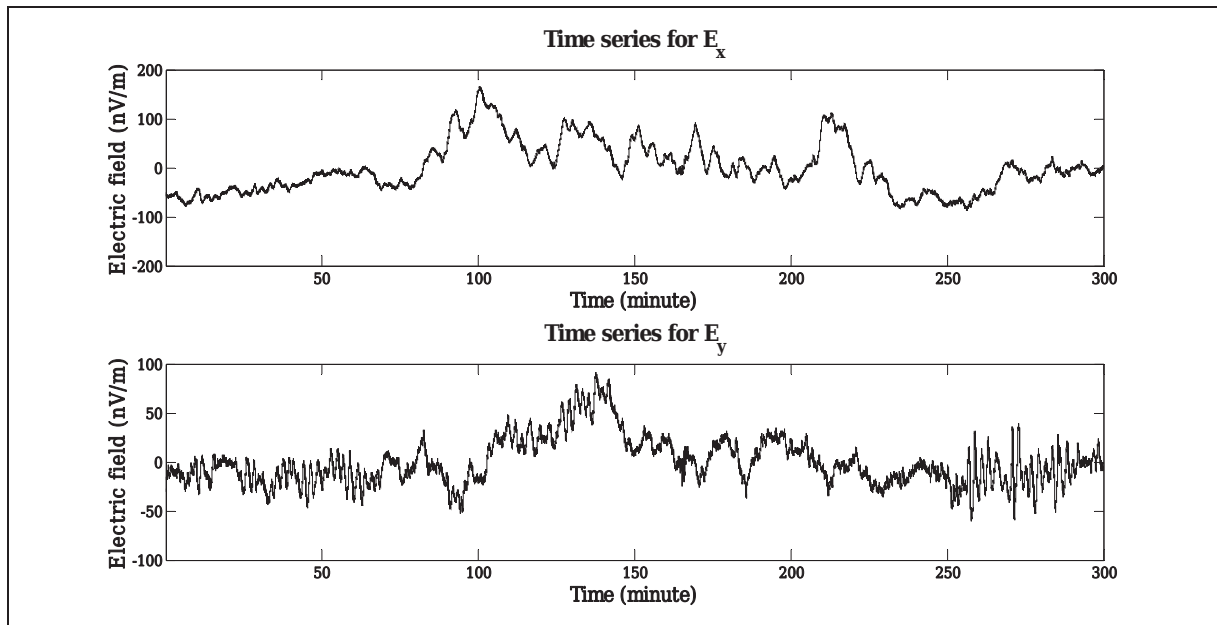


Figure 2: Five hour times-series of two horizontal components of electric fields. E_x (top) and E_y (bottom). The recording is made at the seafloor, 500 m below the sea surface.

6. Power Spectral Density (PSD) and Spectrogram

Power spectral density is calculated to study the power content of the frequencies. For the PSD calculation, steps followed are as follows:

- I. *Inspection of time series (Figure 2):* Inspection is done to recognize glitches or other outliers in the data that are not consistent with the rest of the time series.

- II. *Detrending*: Detrending ensures that the first Fourier coefficient does not dominate the PSD.
- III. *Windowing*: A Hanning window is applied to minimise the leakage of spectral density.
- IV. *Calculation of discrete Fourier transform (DFT)*:
- V. *Calculation of PSD*: PSD is calculated by segment averaging method (Welch, 1967).

The PSD for Ex and Ey is shown in Figure 3. Over all spectral power is decreasing with increase in frequency. Four different slopes can be seen in the both PSD's (Ex and Ey). Let the slope containing frequencies between 0.001 - 0.1 Hz, 0.1 - 2 Hz, 2 - 10 Hz and 10 - 25 Hz represent respectively low, intermediate, sub-high and high frequency range. Evidently, in the low frequency range the power of the electric field rises sharply. Generally, external origin field mainly affects the low frequency range, as higher frequencies are filtered out by the ocean resulting in passage for low frequencies only (i.e. less than 0.3 Hz for a 500 m deep ocean) to the ocean bottom due to skin depth. The oceanic eddies (very long wavelength oceanic feature) are another low frequency source (Chave and Filloux, 1984). Here, these two sources are presumed for the sharp rise in power in the low frequency range (0.001 - 0.1 Hz). In the range of intermediate frequency, four peaks are evident, at 0.2, 0.3, 0.4 and 1 Hz, in the PSD representing that this frequency range is receptive for the various oceanic flows close to seafloor. The sub-high frequency range is nearly flat. A flat PSD, generally, corresponds to a field which contains equal power within a fixed bandwidth which resembles noise. In the high frequency range a sharp rise is evidenced which is presumed due to digitisation noise.

The frequency and power information of a PSD is insufficient to characterise the source process corresponding to the frequencies. Time preservation may help in this subject. Together with the frequency and power, information about time may help in identifying and characterising the source nature of process. The localised time process may represent a source progressing in an interval of time while an ambient time process reflects a continuous process. For the purpose therefore spectrograms are plotted which preserve the time information together with frequency and power. The spectrogram is the discrete-time Fourier transform for a sequence, computed using a sliding window. For a spectrogram calculation, the time series is divided in to segments equal to the length of the hamming window. Each segment overlaps 50% of the samples with the adjacent segment and then PSD is calculated for a defined frequency length. Again and again PSD is calculated by sliding the window to build a spectrogram. The Spectrograms corresponding to time series shown in Figure 2 is shown in Figure 5. There are three

distinct anomalous features evident in spectrograms. One feature is corresponding to time approx. 165 min (like spike) and other two are at frequencies 1, 0.3 Hz. It is difficult to confidently propose a source nature for the spiky feature at approx. 165 min. This could be due to either by a regional earthquake or by a volcano like feature. In general, regional earthquakes are found in band frequencies of 0.1 to 1 Hz and here the spiky feature correspond the same range. The other two features at frequencies 1 and 0.3 Hz are respectively by the microseisms and swell, that will be discussed in the next section.

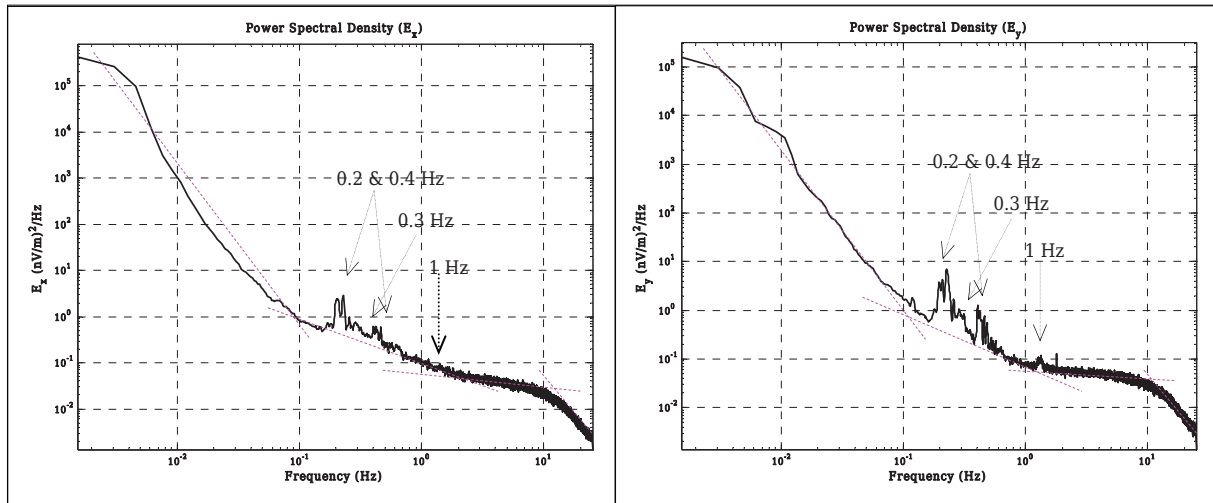


Figure 3: Spectra of the two horizontal electric components E_x and E_y corresponding to time series shown in Figure 2. The graphs show the PSD vs. frequency. The anomalous peaks are visible at 0.2, 0.4 and 1 Hz. There is also a weaker peak at 0.3 Hz. Four slopes can be observed in the spectra (pink dashed lines).

7. Microseisms and swell

Microseisms are like a soft earth tremor originating with in the ocean by non linear interaction of oceanic waves, which causes a continuing oscillation of the ocean floor. The broad frequency range for microseisms is between 0.05 to 1 Hz, which mainly depends on the ocean depth and oceanic conditions.

Languet-Higgins (1950) proposed a mechanism for microseisms (0.05 -2.0 Hz) and showed that if two identical progressive waves travelling in opposite directions interact with each other, there is a second order pressure term effect which does not vanish with depth and can thus reach the deep ocean bottom (Figure 4). Consider two surface waves of frequencies f_1 and f_2 ($f_1 \approx f_2$), moving with approx. same velocity in opposite direction. Let the wave number of frequencies are respectively k_1 and $-k_2$, and then $k_1 \approx -k_2$. Interaction of these waves will leave behind a wave with very small wave number {i.e. $k_1 + (-k_2) = \text{diminutive}$ } and very large wavelength. The large wavelength is capable of creating a pressure disturbance effectively at the ocean floor. The amplitude of the

pressure disturbance is proportional to the product of the interacting wave heights and the frequency. These pressure fluctuations in the water column might then excite Rayleigh waves in the solid earth and be observed as microseisms.

In short, the favourable oceanic conditions for microseisms generation are:

1. Shoreline geological setting creating grounds for nonlinear interaction of surface waves.
2. Near shore reflection of high frequency surface waves and thereafter head on interaction.
3. A fast moving storm creating a sequence of wave in different directions.
4. High frequency wave interaction may generate whitecaps (a wind blown wave whose crest is broken and appears white), which leads in acoustic energy transmission to ocean bottom.

The depth of the ocean is another important factor in microseisms generation. The wave generation in the ocean depends on the wind velocity, which become efficient when the

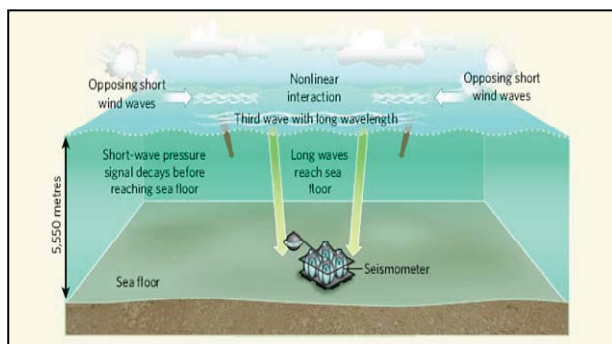


Figure 4: Microseisms mechanism (from Elgar (2008)). Nonlinear interaction of short opposing waves leads to a long wavelength wave generation, which reaches the seafloor and generates

the wave velocity is close to the phase velocity of the ocean waves. Velocity of ocean wave (V) is, roughly speaking, $\sqrt{(gD)}$, where D is ocean depth and g is acceleration due to gravity. Therefore, for a 4 km deep ocean and a 500 m shallow ocean, the velocity V is approximately 200 m/s and 70 m/s respectively. A typical wind velocity rarely exceeds few tens of m/s. This suggests wind velocities are quite close

to the oceanic velocities especially in the shallow oceans. Therefore the generation of microseisms is likely to be efficient in shallow oceans (Tanimoto, 2005).

Swells are a kind of oceanic surface waves. They are often created by the breaking of storms thousands of kilometres away from the seashore. The distance allows the waves comprising the swells to become more stable, clean, and continuous as they travel toward the coast. For microseisms generation, in general, higher frequency waves are more efficient than the lower frequency waves like swells. Swells are more directional and therefore chances for non linear interaction is not as much as of higher frequency gravity waves (Webb and Cox, 1986), which are generated with in the ocean by the influence of gravity at the interface involving the density contrast. In Figure 3 and 5,

the anomalous feature close to 1 Hz is likely due to a microseism. A microseism is a time localised feature, the duration of which depends on the time of effective nonlinear wave interaction. It is evident from Figure 5 that close to 1 Hz, there are four time localised features (E_y component has more clear appearance than E_x), corresponding to time 8, 40, 90 and 250 min approx. Further evidence in support of microseisms is the observed frequency (i.e. 1 Hz).

The acquired electric field data was recorded in a shallow ocean (500 m depth). Normally, a shallow ocean may contribute the microseisms in high frequency range (i.e. close to 1 Hz) as analogous to high frequency even a comparable wavelength (not necessary a huge wavelength) may reach ocean bottom to produce microseisms. The mechanism of microseisms generation suggests that maximum electric field power will be experienced in the component perpendicular to the direction of wave propagation. Here, the velocity component $V_y > V_x$ as amplitude of $E_x > E_y$. The strong power of microseisms in the E_y component compared to the E_x component further supports to interpret 1 Hz frequency as microseisms.

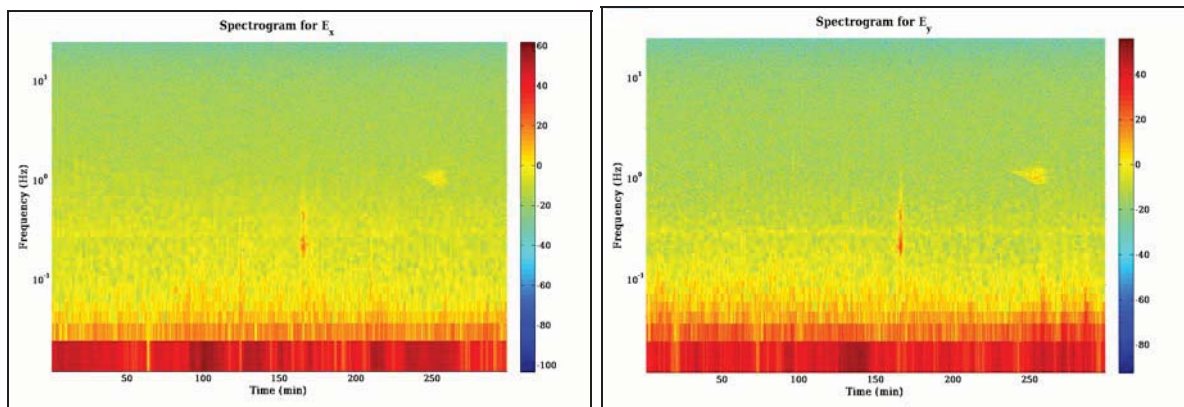


Figure 5: Spectrogram for E_x & E_y . Electric field power is colour coded in dB and displayed as function of time and frequency. Two anomalous features are visible, one at 1 Hz and the other covering a broad frequency range, having max. power at 0.2 & 0.4 Hz.

The individual feature ambient in time at 0.3 Hz (Figure 3 & 5) is interpreted here as a swell. A swell consists of long-wavelength surface waves which are more stable in their directions and frequency than normal wind waves. Swells are dispersive in nature and their frequency is given by $\omega^2 = gs \tanh(sD)$, where g is acceleration of gravity; $s = 2\pi/\lambda$ is wave number, λ is the wavelength of the swell and D is the depth of ocean. A calculation suggests that a swell of frequency 0.3 Hz corresponds to a wavelength of 18 m for 500 m deep oceanic water. The obtained wavelength is well in range for the

wavelengths of swells. As 0.3 Hz frequency is evident throughout the time in spectrogram (Figure 5), this is another strong support for 0.3 Hz to stand for a swell.

7. Microseisms during mCSEM data recording

Microseisms are a powerful feature to effectively generate EM signal at the ocean floor. They are quite capable of masking the mCSEM signals. We have an example comparing the power of microseism and mCSEM transmitter current. In Figure 5, we have seen the spectrograms of mCSEM the time-series recorded in the absence of transmitter current. On the other hand, Figure 6 represents spectrograms of the mCSEM data when transmitter was transmitting signals. Systematic decay in the power is evident in Figure 6, representing transmitter is moving away from the receiver. A gathering of high power (yellow colour) patch is evident corresponding to frequency 0.1 Hz and time 200 min, may be representing a microseism. Clearly, the power of a microseism is significant enough to contaminate the mCSEM data. It is as well evident that the effect of swell is feeble in the presence of a transmitter current. For a larger transmitter receiver distance this effect may be significant.

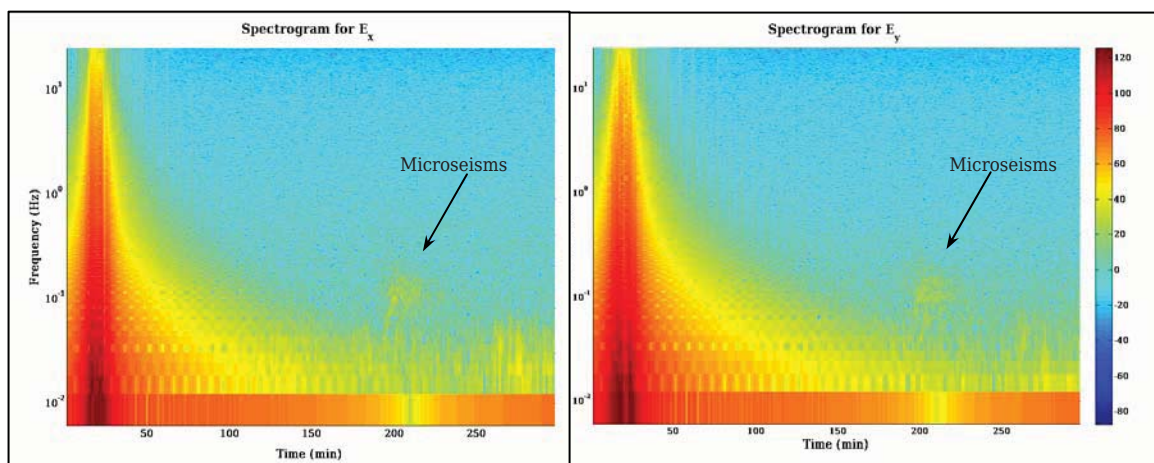


Figure 6: Presence of a microseism during mCSEM data recording. Electric field power is colour coded in dB and displayed as function of time and frequency. Corresponding to frequency 0.1 Hz and time 200 min (approx), a yellow coloured high power patch is evident, which may be by a microseism. The other efficient high power patches in the spectrogram is by the transmitter current

8. Conclusions

By the analysis seafloor electric field data, we provide evidence for the observation of microseisms and swell in mCSEM data. Power spectral density (Figure 3) and

spectrogram (Figure 5) clearly present the evidence for the possible presence of microseism and swell corresponding to the frequencies 1 Hz and 0.3 Hz respectively. Swells are time ambient oceanic feature, which can be clearly marked in the spectrograms. The stability of 0.3 Hz in spectrograms indicates that oceanic conditions were quite stable during the recording of present mCSEM data.

Figure 6 shows that the power of a microseism is significant even when the transmitter current is 'on'. Overall, the time series analysis suggests that features like microseism and swell induce a significant electric field at the seafloor to contaminate the mCSEM data. Proper measures for removal of the noise are essential for a better interpretation of mCSEM data.

Acknowledgement

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Motionally Induced Electromagnetic Field within the Ocean

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Abstract

The contribution of motionally induced electromagnetic (EM) fields at the seafloor is generally considered small, but since the characteristic reservoir signal in marine controlled source electromagnetic (mCSEM) data is also small, the inclusion of the motional induction contribution in modelling the signal will enhance the probability of reservoir detection. Here, we have studied the electromagnetic induction caused by ocean water flow with in earth's magnetic field.

When a charge particle moves with certain velocity in earth's magnetic field, it experiences a Lorentz force. The action of Lorentz force generates a secondary electric field through galvanic and inductive processes. For the mathematical formulation, we considered Lorentz electric field as a source in the corresponding set of Maxwell's equations. We solved these Maxwell's equations for a 1D model and velocity structure using two different Green's function i.e. a two half space Green's function and a layered Green's function. The layered Green's function is especially useful in studying the sensitivity of electric and magnetic field for different conductivity structures in the earth. Further, the signal variation with the conductivity of ocean, depth of ocean and wave velocity is studied to profoundly understand the effect of these parameters.

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1. Introduction

Oceanic water movements in the earth's magnetic field induce electric and magnetic fields in the ocean. The induced fields are signal for oceanographic and seismological applications but noise for magnetotelluric (MT) and marine controlled source electromagnetic (mCSEM) applications. Oceanographers and seismologists use the fields respectively to study the velocity structure and the seismic background noise. The MT and mCSEM signals, which are generally used for lithospheric and hydrocarbon exploration studies, are contaminated by the induced field and act as noise. Our prime focus of motional induction study here is on mCSEM problems; nevertheless the results are also applicable for other scientific applications.

Generally, motionally induced noise in seafloor mCSEM data is considered small, but since the characteristic reservoir signal is also small, the understanding and possible removal of noise may be essential to increase the number of possible target reservoirs.

The electromagnetic induction investigation has a long history. Initially, by an observation of deflection in the galvanometer by stream waves Faraday (1832) concluded that flow of water will induce electric currents, which was afterwards measured experimentally by Young et al (1920). Later it was almost neglected and got re-attention after World War II. Longuet-Higgins et al., (1954) initiated the investigation of electric field induction by surface waves. Crews and Futterman (1962) investigated the magnetic field induction by oceanic movement. Further, Sanford, (1971) extended the theory of motional induction by considering three-dimensional water flows. A comprehensive study of the theory has been made by Podney (1975), who generalised and extended the previous work by removing the restrictive assumptions imposed on ocean-water velocity field. Chave (1983) generalised the EM induction process by considering driving electric field term (i.e. $\vec{v} \times \vec{B}$) and further, Chave and Luther (1990) re-examined the motional induction problem.

A primary purpose of this paper is to present a generalised simple illustrative theory for the problems of motional induction. For the purpose, we have formulated a set of Maxwell's equation for our problem. These equations are simplified for electric and magnetic field components by considering a horizontally progressing ocean wave and then solved by using the Green's function, with appropriate boundary conditions. Here, we solved the problem with two different Green's functions. For a uniformly conductive earth, a two half space Green's function is utilised and for a layered earth, a layered Green's function is utilised. A two half space Green's function doesn't offer reflections because of homogeneous conductivity consideration and thus expresses only the case of downward progressive diffusive waves. The layered Green's function includes both downward and upward propagating diffusive waves offered by the layered boundaries.

2. Problem formulation in terms of Maxwell's equation

2.1 Basics

An electrically conducting fluid like ocean consists of charged particles. The particles in the ambient geomagnetic field experience a deflective Lorentz force. If ' $\vec{v}$ ' is the velocity of charge particle ' q ' moving in the geomagnetic field ' $\vec{B}_0$ ', then the Lorentz force is:

$$\vec{F}_L = q(\vec{v} \times \vec{B}_0) \quad (1)$$

The charge ' q ' experiences the deflecting force ' $\vec{F}_L$ ' because of the action of an electric field which we call as Lorentz electric field ($\vec{E}_L$),

$$\vec{F}_L = q\vec{E}_L \quad (1-a)$$

Therefore, $\vec{E}_L = \vec{v} \times \vec{B}_0 \quad (2)$

The field ' $\vec{E}_L$ ' generates a secondary electric field ' $\vec{E}$ ', mainly by two processes-

(1) **Galvanic process:** At locations, where ' $\vec{E}_L$ ' has a component parallel to the conductivity gradient ' $\nabla\sigma$ ' (σ is conductivity in S/m), space/surface charges are accumulated. The accumulated charges galvanically create a secondary field ' $\vec{E}$ ', even if the wave velocity is constant.

(2) **Inductive process:** The magnetic field ' $\vec{H}_L$ ' is created by the current density ' $\vec{J}_L$ ' i.e.

$$\begin{aligned} \vec{J}_L &= \sigma(\vec{v} \times \vec{B}_0) \\ \nabla \times \vec{H}_L &= \vec{J}_L \end{aligned}$$

when the velocity ' $\vec{v}$ ' changes with time. This induces a secondary electric field ' $\vec{E}$ ' via the law of induction.

We assume that all the hydrodynamics, including Coriolis force, is included in the given ' $\vec{v}$ '. We do not care about the sources of forces. Finally, for a stationary frame of reference, the current density in the Ohms law is given by

$$\vec{J} = \sigma(\vec{E} + \vec{E}_L) = \sigma(\vec{E} + \vec{v} \times \vec{B}_0) \quad (3)$$

Here, ' σ ' is the conductivity of the fluid, ' $\sigma\vec{E}$ ' is the current term generated by both a galvanic and an inductive process and ' $\sigma(\vec{v} \times \vec{B}_0)$ ' is the source current term for motional induction case. Under the quasi-static approximation, the set of Maxwell's equation to be solved for the motional induction case is

$$\begin{aligned} \nabla \times \vec{H} &= \sigma(\vec{E} + \vec{v} \times \vec{B}_0) \\ \nabla \times \vec{E} &= -\mu_0 \partial_t \vec{H} \end{aligned} \quad (4- a, b)$$

In an exact formulation, the ambient magnetic field is the sum of geomagnetic field ' $\vec{B}_0$ ', the external magnetic field by ionospheric currents ' $\vec{B}_{\text{ext}}$ ', the field generated by small local anomalies ' $\vec{B}_{\text{l-ano}}$ ' and the motionally induced field ' $\vec{B}_{\text{mi}}$ ' i.e. $\vec{B} = (\vec{B}_0 + \vec{B}_{\text{ext}} + \vec{B}_{\text{l-ano}} + \vec{B}_{\text{mi}})$. In general, the last three terms are orders of magnitude smaller than the geomagnetic field ' $\vec{B}_0$ '. Therefore, for computations, the local value of earth's magnetic field would be a good choice.

3. A halfspace model

Let the north, east and positive downward directional depth in Cartesian coordinate system be represented by $\hat{x}$, $\hat{y}$ and $\hat{z}$. Consider a two halfspace model with the boundary at the earth surface (i.e. $z = 0$). Let the insulating air halfspace extend in the negative vertical upward direction while the earth halfspace extends in positive vertical downward direction from the boundary surface (Fig 1). The conductivity of the earth's half space is $\sigma = 3.33$ S/m, which contains an ocean and sediments below. In general, the conductivity of the ocean depends on dissolved ions content and their mobility, which is primarily a function of temperature and pressure. The conductivity in warm, top shallow water is normally high (≈ 5 S/m) compared to cold, deep water (≈ 2.5 S/m). For simplicity, oceans can be considered homogeneous as the range of conductivity difference (i.e. 2.5–5 S/m) is small. Consider with depth $d = 1000$ m ocean moving with a velocity $\vec{v}$ in the x-direction (Fig 1). Below the ocean, a subsurface of stationary sediments (i.e. $\vec{v} = 0$, motionless) exists.

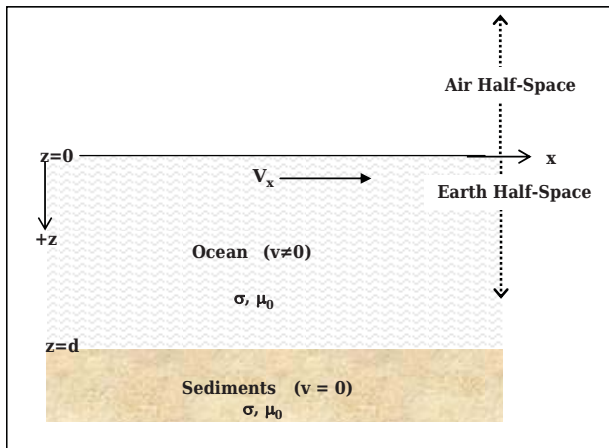


Fig 1. The two halfspace model. $z = 0$ is the boundary between earth and air halfspace. Depth is considered positive in the downward direction. The earth halfspace consists of two layers. The top layer represents the ocean having non-zero velocity in the x-direction. The depth of the ocean is d . Below the ocean, a second layer of static (i.e. velocity zero) sediments extends. The conductivity and magnetic permeability of earth halfspace is σ and μ_0 respectively.

Let $\vec{B}_0 = B_0 \hat{z}$ and $\vec{v} = v_x(z, t) \hat{x}$

This will generate a horizontal electric field perpendicular to the velocity direction and a horizontal magnetic field perpendicular to the electric field, and thus we can write:

$$\vec{E} = E_y(z, t) \hat{y} \text{ and } \vec{H} = H_x(z, t) \hat{x}$$

From eq. (4), we have

$$\partial_z^2 E_y = \mu_0 \sigma \left(\partial_t E_y - B_0 \partial_t v_x \right) \quad (5)$$

Assume a harmonic time dependence for simplicity, $v_x(z, t) = \tilde{v}_x(z) e^{+i\omega t}$,

Then $E_y(z, t) = \tilde{E}_y(z) e^{+i\omega t}$, $H_x(z, t) = \tilde{H}_x(z) e^{+i\omega t}$ and eq. (5) reduces to

$$\partial_z^2 \tilde{E}_y - k^2(z) \tilde{E}_y = -g(z) \quad (6)$$

where, $g(z) = +k^2(z)\tilde{v}_x(z)B_0$ is the source term, $k(z) = \sqrt{i\omega\mu_0\sigma(z)}$ represents the electromagnetic damping phenomenon. It is complex such that amplitude decay is associated with a phase shift with respect to the field at the surface. The Green's function $G(z|z_0)$ corresponding to equation (6) is defined by

$$\partial_z^2 G(z|z_0) - k^2(z)G(z|z_0) = -\delta(z-z_0) \quad (7)$$

Crosswise multiplication with G and $\tilde{E}_y$ and subtraction yields

$$\partial_z(G\partial_z\tilde{E}_y - \tilde{E}_y\partial_z G) = -g(z)G + \tilde{E}_y\delta(z-z_0) \quad (8)$$

The integration of (8) over depths z yields the solution

$$\tilde{E}_y(z_0) = \int_0^d g(z)G(z|z_0)dz, \text{ for } -\infty < z_0 < \infty \quad (9)$$

i.e. a convolution of the source term with the Green's function. The integration, which formally must be carried out from minus to plus infinity, reduces to 0 to d , because that is where the source is nonzero. For the horizontal electric field $\tilde{E}_y(z_0)$ calculation at any desired depth, knowledge of the Green's function is required.

3.1 Green's function

3.1.1. Half space

The Green's function is commonly used to solve inhomogeneous boundary value problems. The solution by means of Green's function gives a special advantage because of its reciprocity property, which states 'relationship between a oscillating source and the resulting field at some point of observation is unchanged even if the observation and source points are interchanged'. Using the boundary conditions, the Green's function is calculated for the halfspace. Calculation is as follows:

For an arbitrary small ε , equation (7) follows,

$$\partial_z G(z_0 + \varepsilon|z_0) - \partial_z G(z_0 - \varepsilon|z_0) = -1 \quad (10)$$

The second boundary condition is that $G(z|z_0)$ is continuous at $z = z_0$ i.e.

$$G(z_0 + \varepsilon|z_0) - G(z_0 - \varepsilon|z_0) = 0 \quad (11)$$

For the uniform halfspace,

$$G(z|z_0) = \begin{cases} A e^{+kz} & \text{for } z < z_0 \\ B e^{-kz} & \text{for } z > z_0 \end{cases} \quad (12-a,b)$$

A and B are determined from (10) and (11) which yields the Green's function for the half space

$$G(z | z_0) = \frac{1}{2k} e^{-k|z-z_0|} \quad (13)$$

3.1.2 Two halfspaces

In order to consider the diffusive reflection and transmission effects due to boundaries, let us consider two uniform half spaces with $k(z) = k_0$ in the air (i.e. $z < 0$) and $k(z) = k$ in earth (i.e. $z > 0$). The Green's function, for $z_0 > 0$, can be written as

$$G(z | z_0) = \begin{cases} \frac{1}{2k} e^{-k|z-z_0|} + R e^{-kz} & \text{for } z \geq 0 \\ T e^{-k_0 z} & \text{for } z \leq 0 \end{cases} \quad (14)$$

where, T and R are respectively transmission and reflection coefficients. The continuity of G and $\partial_z G$ at $z = 0$ yields

$$G(z | z_0) = \begin{cases} \frac{1}{2k} \left[e^{-k|z-z_0|} + \frac{k-k_0}{k+k_0} e^{-k(z+z_0)} \right] & \text{for } z \geq 0, z_0 \geq 0 \\ \frac{1}{k+k_0} e^{k_0 z - k z_0} & \text{for } z \leq 0, z_0 \geq 0 \end{cases} \quad (15)$$

In particular, for the insulating air halfspace $k_0 = 0$, the two halfspace Green's functions take the form

$$G(z | z_0) = \begin{cases} \frac{1}{2k} \left[e^{-k|z-z_0|} + e^{-k(z+z_0)} \right] & \text{for } z \geq 0, z_0 \geq 0 \\ \frac{1}{k} e^{-k z_0} & \text{for } z \leq 0, z_0 \geq 0 \end{cases} \quad (16)$$

3.1.3. A simple model and field expression

Equation (9) includes the source term $g(z)$, which depends on conductivity and velocity, both are depth dependent. For simplicity, consider a uniform horizontal velocity i.e. $v(z) = v_0, 0 \leq z \leq d$. Let the depth dependent conductivity be zero in the air and $\sigma = 3.33 \text{ S/m}$ for the earth i.e.

$$\sigma(z) = \begin{cases} \sigma, & z \geq 0 \\ 0, & z < 0 \end{cases}$$

The equation (9), (16) and (4), yields the horizontal electric and magnetic field expression for the full space.

Electric field in full space:

$$\begin{aligned}\tilde{E}_y(z_0) &= v_0 B_0 (1 - e^{-kd}) & , z_0 \leq 0 \\ \tilde{E}_y(z_0) &= v_0 B_0 (1 - e^{-kd} \cosh(kz_0)) & , 0 \leq z_0 \leq d \\ \tilde{E}_y(z_0) &= v_0 B_0 e^{-kz_0} \sinh(kd) & , z_0 \geq d\end{aligned}\quad (17-a, b, c)$$

Magnetic field in full space:

$$\begin{aligned}\tilde{H}_x(z_0) &= 0 & , z_0 \leq 0 \\ \tilde{H}_x(z_0) &= -\frac{kv_0 B_0}{i\omega\mu_0} e^{-kd} \sinh(kz_0) & , 0 \leq z_0 \leq d \\ \tilde{H}_x(z_0) &= -\frac{kv_0 B_0}{i\omega\mu_0} e^{-kz_0} \sinh(kd) & , z_0 \geq d\end{aligned}\quad (18-a, b, c)$$

The graphical response of (17) and (18) is shown in Fig (2). The response is calculated for a halfspace model (Fig 1) of a conductivity of 3.33 S/m containing 1000 m thick layers of ocean and sediments. The magnetic permeability is kept constant for both halfspace and is equal to free space permeability i.e. $\mu_0 = 4\pi \times 10^{-7}$ Vs/Am. The ocean has a homogeneous velocity of 10 cm/s in an ambient geomagnetic field of 5×10^{-5} T. The responses are studied for frequencies 0.001, 0.01, 0.1, and 1 Hz. In general, the electric field becomes gradually weaker with depth (Fig 2). For 1 Hz, rather than weakening gradually, the electric field amplitude increases and becomes strongest with in the ocean (at 400 m depth). Further, the field B_x is zero at the surface and progressively becomes stronger with respect to depth. At the ocean bottom, it offers strongest amplitude. The zilch of B_x at the earth surface is for the reason that in air halfspace field $\tilde{E}_y$ is constant and therefore $\partial_z \tilde{E}_y = 0$ in particular, which is B_x (eq. 4-b). Further, as far as frequency based variation are concerned, the smallest frequency B_x produces the strongest amplitude at the ocean bottom. The field E_y shows a contrary behaviour. Here the smallest frequency offers the weakest amplitude at the ocean floor. At large frequencies the field E_y can not reach to the ocean floor, because of the shallow skin depth, offers constant amplitude (Fig 3).

The important result in Fig 2 & 3 is that the field E_y is smooth over the boundary between the ocean and sediment. On the other hand, the field B_x offers a sharp change in pattern there (at boundary). Note that the conductivity of the ocean and subsurface (sediments) are identical. They differ only in their velocity state (subsurface is static and

ocean is dynamic). Therefore the sensitivity of the field B_x at the boundary is interpreted as result of velocity change.

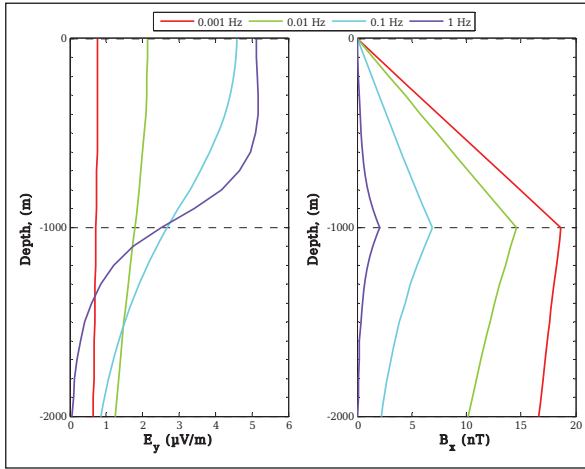


Fig 2. Variation of the horizontal comp. electric (E_y) and magnetic (B_x) field with respect to depth. Variation is shown for four frequencies viz. 0.001, 0.01, 0.1 and 1 Hz. An ocean of conductivity $\sigma=3.33$ S/m extends from surface to 1000 m depth (i.e. $0 \leq z \leq 1000$ m). Below the ocean (i.e. below 1000 m depth) there is a sediment layer with the same conductivity as the ocean (i.e. $\sigma=3.33$ S/m). The horizontal line at 1000 m is to show the separation of these two boundaries.

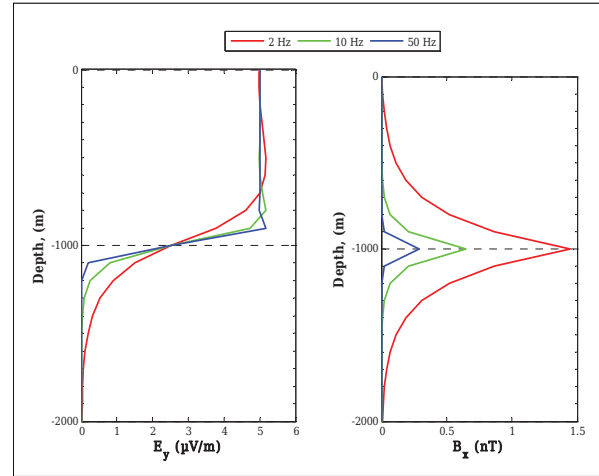


Fig 3. Variation of the horizontal component electric (E_y) and magnetic (B_x) field with respect to depth. Variation is shown for three frequencies viz. 2, 10 and 50 Hz. Other information is the same as in Fig 2. It is evident that at the ocean floor, the field E_y offers almost same amplitude for all the three frequencies. B_x is strongest and weakest respectively for the smallest and largest frequency.

The field E_y (17-b) within the ocean (i.e. $0 \leq z_0 \leq d$) consists of two terms. The first term is the kernel for the motional induction, while the second governs the depth dependent frequency based EM damping. From the expressions (17-b & 18-b), it is clear that the strength of both the fields E_y and B_x depends on the ocean depth (i.e. ocean deepening) because of the factor e^{-kd} . Effect of the ocean deepening is studied for oceans of thickness (depth) varying from 1000 m to 9000 m. Observation depth is constant for all practical cases and is 1000 m (i.e. $z_0 = 1000$ m). Results are shown in Fig (4). E_y is strongest for the thickest ocean and weakens down gradually as ocean shallows up. The reverse is observed for the B_x case with weakest strength in the deepest ocean which progressively becomes stronger as the ocean gradually shallows up. The sensitivity for the deepening effect depends on frequency and therefore on skin depths. For that reason, the deepening effect is more effective at smaller frequencies. The skin depths, for a halfspace of 3.33 S/m, at frequencies 0.001, 0.01, 0.1, 1 and 10 Hz are approximately 8664, 2739, 866, 273 and 86 m, respectively.

The factor which may significantly influence the motional induction is the conductivity of the ocean. The results of the conductivity variation are shown in Fig 5. The field strength is observed at 1000 m depth for conductivities varying from 3 to 5 S/m. Observations are made for five different frequencies (i.e. 0.001, 0.01, 0.1, 1 and 10 Hz). The results suggest that the field B_x is more sensitive to conductivity variation than the field E_y . Further, lower frequencies are more responsive to conductivity variation than high ones because of the thicker skin-depth.

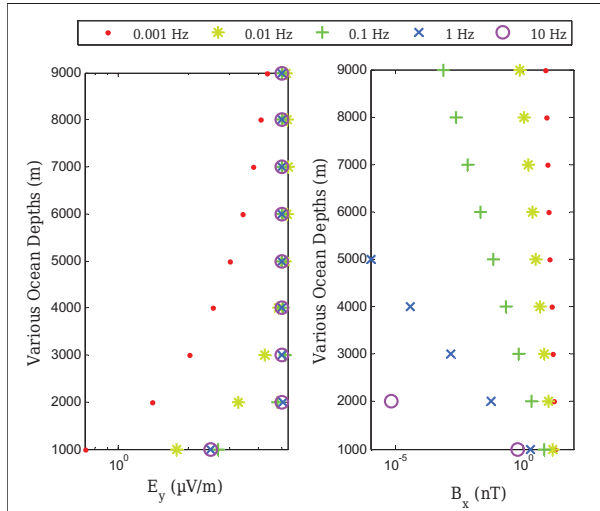


Fig 4. Deepening effect of an ocean: Conductivity of ocean kept constant (i.e. 3.33 S/m). For 1000 to 9000 m deep ocean the field strengths are calculated at frequencies 0.001, 0.01, 0.1, 1 & 10 Hz. The y- and x-axis represents ocean depths and field strength respectively. The observation depth is constant and is 1000 m. The magnetic component for 1 and 10 Hz is truncated for depths greater than 5000 and 2000 m respectively to save the Figure from masking and cluttering.

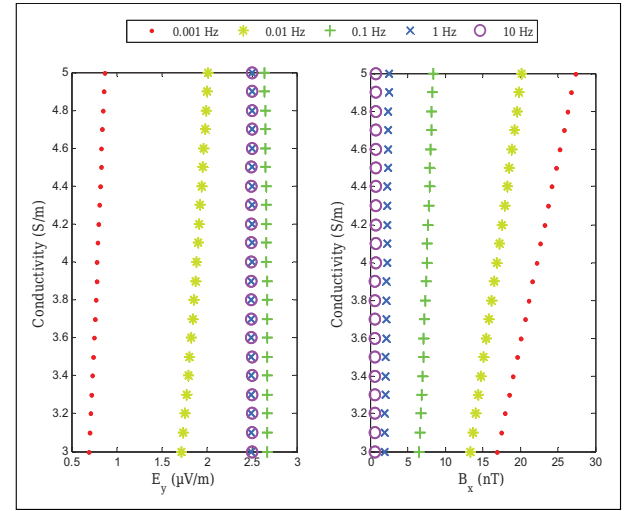


Fig 5. Conductivity variation effect: The conductivity of the ocean is varied from 3 to 5 S/m and the effect is studied for frequencies 0.001, 0.01, 0.1, 1 & 10 Hz. The horizontal component electric and magnetic field strength, observed at 1000 m ocean depth is shown w.r.t. conductivities for the chosen frequencies. Electric and magnetic components both are sensitive to the conductivity of the ocean. Sensitivity increases with the decreasing frequency.

Equation (17) and (18) indicate that the field B_x is more sensitive to conductivity variation in vertical direction, as conductivity is effectively convolved, than the field E_y . But it would not be appropriate to discuss conductivity variation in vertical direction using a halfspace model. Therefore, we will be back on this issue with a justified layered model and a layered Green's function.

Moreover, the equation (17) and (18) implies that the increase in the wave velocity v_0 will cause a constant shift in the E_y and B_x field. Three experiments are conducted for wave velocities of 10, 1 and 0.1 cm/s to study the effects. The results are shown in the Fig 5. A log scale is used for x-axis plotting for clarity reasons. As at the

surface $B_x=0$, therefore the log scale plot starts from 100 m depth as 100 m is the 2nd discretized layer of the ocean. The selection of velocities is such that the 1st selection exceeds the 2nd by factor of 10 and same is true for 3rd and 2nd. Evidently, response for the field E_y and B_x also exceeds by a static shift of factors of 10. The graph clearly illustrates that the velocity is an important parameter for practical simulation, any bias in velocity leads to the same bias in EM field.

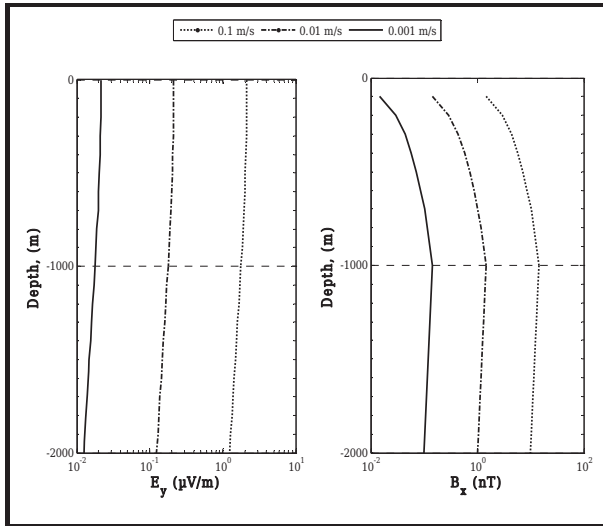


Fig 6. Wave Velocity effect. Semi-log plot of horizontal electric and magnetic field component illustrating the effect of velocity change. The strongest velocity generates the strongest EM field. The B_x plots starts from 100 m depth as the strength of B_x at $z=0$ is zero and at $z=100$ m the next layer starts.

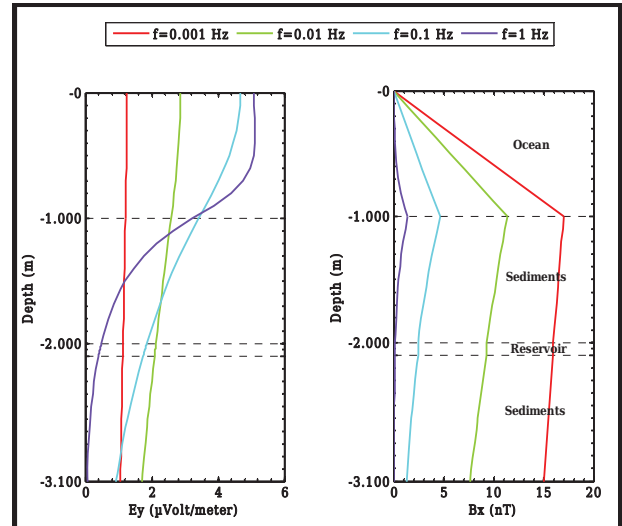


Fig 7. Variation of the horizontal component electric (E_y) and magnetic (B_x) field with respect to depth for a layered model. Variation is shown for four frequencies viz. 0.001, 0.01, 0.1 and 1 Hz. An ocean of conductivity $\sigma=3.33$ S/m extends from the surface to 1000 m depth (i.e. $0 \leq z \leq 1000$ m). Below the ocean there, is a 1000 m thick sediments layer with conductivity $\sigma=1$ S/m. A 100 m thick reservoir of conductivity 0.01 S/m is embedded at 2000 m depth. The horizontal line marks the boundary of different formation.

4. Layered model

4.1. Layered Green's function

Let us consider a layered Green's function defined as

$$G(z | z_0) = \frac{1}{2k_1} \left[e^{-k_1 |z - z_0|} + R_0 e^{-k_1 z} + R_d e^{+k_1 z} \right]; \quad 0 \leq z \leq d, 0 \leq z_0 \leq d$$

where, $k_1 = \sqrt{i\omega\mu_0\sigma_1}$ is the electromagnetic damping in the ocean, R_0 describes the field diffusing downward by reflection from the air-earth interface and R_d describes the field diffusing upward by reflection from the seafloor (i.e. $z = d$). In this particular case the constants R_0 and R_d are determined from boundary conditions:-

a) $\partial_z G(0 | z_0) = 0$, Since $G(z | z_0)$ is constant in the air halfspace

b) $\partial_z G(d | z_0) + \mu_0 b_2(d) G(d | z_0) = 0$, where, $b_2(d)$ is determined via continuous transfer function.

They yield,

$$R_0 = \frac{1 - R_c e^{-2k_1(d_1 - z_0)}}{1 + R_c e^{-2k_1 d_1}} e^{-k_1 z_0} \quad ; \quad R_d = -\frac{R_c (1 + e^{-2k_1 z_0})}{1 + R_c e^{-2k_1 d_1}} e^{-k_1 (2d_1 - z_0)}$$

with $R_c = \frac{\mu_0 b_2 - k_1}{\mu_0 b_2 + k_1}$

4.2 Layered model and Response

EM field responses calculated for a layered model are shown in Fig (6). The parameters of the model are tabled in table 1. The magnetic permeability of the each layer is equal to free space permeability (μ_0). The response is computed for four frequencies viz. 0.001, 0.01, 0.1 and 1 Hz. It is evident that the field E_y varies smoothly over layer boundaries, indicating its sensitivity to the vertically averaged conductivity rather than conductivity variation at layered boundaries. However, the field B_x senses each layer and offers a change in field strength at each layer boundary. Evidently, the magnetic field and its sensitivity for conductivity variation is frequency dependent. At a small frequency the field is strong and vice versa.

Table 1: Layered Model

Layers	Thickness (m)	Conductivity (S/m)	Other Parameters
Ocean	1000	3.33	Wave Velocity=0.1 m/s Ext. Magnetic Field= 5×10^{-5} T Magnetic Permeability of each layer $=4\pi \times 10^{-7}$
Sediments	1000	1	
Reservoir	100	0.01	
Below	1000	1	

5. Discussion and Conclusion

No doubt, the actual earth is 3D and therefore the electromagnetic response of the earth will always be more complex than the response of any simplified model. However, a simplified model, roughly representing actual earth layers can be used to validate results. The choice of the model dimension (i.e. 1D, 2D or 3D) sometimes has a large impact on the result. In particular, for mCSEM surveys, generally 2D data acquisition methodology is followed and therefore a 2D/3D modelling may lead to a better interpretation. However, 1D model formulations are simple and give good insight into the physics of the problems.

Field strength depends on their components (depending on the physical process involved), whose production mainly depends on the involvement of the source components. For example, a 1D model in motional induction study results in E_y and B_x (two) components when a 1D source (i.e. wave-velocity in x-direction) is considered. However, a 2D source (i.e. wave-velocity in x-direction and a wave wavelength in y-direction) consideration results in E_y , B_x and E_z (three) components. This suggests, even a 1D model, with a proper source formulation offers significant insight to the problems. Further, the simplicity of a 1D model is an important advantage. In this paper we thoroughly looked for the sources of the motional induction in the ocean, which is incorporated in 1D model for theoretical development. The horizontal (x-direction) wave motion and a vertical (z-direction) geomagnetic field consideration lead to the excitation of the field components ' E_y ' and ' B_x '. These fields illustrate some of important results and effects of the ocean dynamics in varying oceanic conditions.

In general, for a uniform halfspace, there is gradual reduction and increase respectively in the strength of E_y and B_x w.r.t. ocean depth (Fig 2). This conclusion is valid for frequencies with skin depth ' δ ' greater than the ocean depth ' d '. For the case when $\delta < d$, E_y may show maximum amplitude somewhere in the ocean, rather than at the ocean surface. On the other hand in the ocean the field B_x always offers maximum strength at the floor and minimum (i.e $B_x = 0$) at the surface. Since B_x vanishes at the surface, the mCSEM surface measurement may offer significant noise-free signals there, if MT field is avoided. At the ocean floor as B_x always has maximum strength therefore it is necessary to correct mCSEM data for motionally induced field. It is evident from Fig (2), even in a halfspace where the conductivity of ocean and subsurface is same, B_x senses the boundary of the ocean and subsurface (i.e. sediments) though E_y does not see it. Still, an important question left to answer is 'which field is/are sensitive to layer demarcation, either E_y or B_x or both?' This issue will be conferred below later, in the paragraph with layered earth discussion.

The fields B_x and E_y both are sensitive (Fig 5) to the conductivity variation of the oceans. The observations with several oceans differing in their conductivity suggests that at frequencies with $\delta < d$, in general, the variation of ocean conductivity is negligible but at frequencies with $\delta > d$, the more conductive ocean increases the strength of E_y and B_x field.

The average depth of Atlantic, Pacific and Indian Ocean is respectively 3926 m, 4282 m and 3963 m. Moreover, from place to place the oceanic depths are quite diverse and therefore the significance of deepening of the ocean is studied (Fig 4). For a fixed observation depth in ocean, the gradual increase in the strength of E_y and decrease in the strength of B_x is the consequence of deepening of the ocean. The component E_y and B_x show strongest and weakest strength respectively for the deepest ocean and vice-versa. For frequencies for which the skin depth is smaller than the ocean depth (i.e. $\delta < d$), the further deepening of the ocean (greater than skin depth) is ineffective and therefore the field value may saturate with further deepening.

Another important parameter is the ocean wave velocity. The study suggests (Fig 5) that any bias in velocity value may cause same bias in field strength. The change in velocity does not modify the pattern with depth, but causes a static shift in the field.

Expression for layered Green's function has allowed us to study the EM field variation for different layers with in the earth. The layered model involves both resistive and conductive layers. Clearly, electric field E_y does not see the layered boundaries (Fig 6) although the magnetic field clearly sees it by offering a change in the slope of B_x at the boundaries.

Acknowledgement

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Three-dimensional finite element simulation of magnetotelluric fields incorporating digital elevation models

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1 Introduction

Pronounced topography may have an important impact on the magnetotelluric (MT) response and its neglect may lead to severe misinterpretations. Common simulation techniques based on rectangular grids are generally not well suited to deal with arbitrary geometry. We therefore use vector finite element (FE) schemes formulated on unstructured grids to cope with realistic topography and/or bathymetry. As an example, we have chosen the volcanic island of Stromboli which is located in the Mediterranean Sea off the west coast of Italy. Stromboli is an extreme electric environment with very conductive sea water surrounding the steep topography of the resistive volcanic edifice. Moreover, the varying bathymetry and the topography of the other islands of the Liparian archipelago cause distinct effects on the magnetotelluric response. Due to the large conductivity contrast and these extraordinary features it can be expected that the magnetotelluric apparent resistivities as well as the phases show a very complex behaviour. The objective of this work is to incorporate digital elevation models of this area into our numerical MT simulations allowing for a realistic look at the electromagnetic (EM) induction phenomena in such a complicated environment. On top of its challenge for numerical simulation methods this volcano has always fascinated geoscientists (Fig. 1) and future interdisciplinary studies aim at investigating the inner structure and the processes that lead to the continuous mild eruptions recorded over the last 2 000 years (the so called 'Strombolian activity').



Figure 1: Map of the Aeolian Islands (left, Wikipedia (2010)) and Stromboli (right, SwissEduc (2010))

2 Physical and numerical basics

Based on Maxwell's equations and assuming a harmonic time dependency $e^{i\omega t}$ of the incoming plane wave, the equation of induction in terms of the vector potential $\mathbf{A}$ reads as

$$\nabla \times \mu^{-1}(\nabla \times \mathbf{A}) + (i\omega\sigma - \omega^2\epsilon)\mathbf{A} = 0, \quad (1)$$

where $\mu, \sigma, \epsilon, \omega$ and i are the magnetic permeability, the electric conductivity, the permittivity, the angular frequency and the imaginary unit, respectively. The magnetic field $\mathbf{H}$ and the electric field $\mathbf{E}$ are obtained by

$$\mathbf{H} = \mu^{-1}(\nabla \times \mathbf{A}) \quad \text{and} \quad \mathbf{E} = -i\omega\mathbf{A} - \nabla V. \quad (2)$$

The scalar potential V was eliminated from eq. (1) by the gauge condition $\psi = -iV/\omega$ and substituting $\mathbf{A}$ for $\mathbf{A} - \nabla\psi$ which leaves $\mathbf{H}$ and $\mathbf{E}$ unchanged.

To calculate $\mathbf{A}$ in a bounded domain $\Omega \subset \mathbb{R}^3$, electric and magnetic insulation are required for all outer boundaries parallel ($\Gamma_{||}$) and perpendicular ($\Gamma_{\perp}$) to the current flow, respectively:

$$\begin{aligned} \mathbf{n} \times \mathbf{H} &= 0 \quad \text{on} \quad \Gamma_{||}, \\ \mathbf{n} \times \mathbf{A} &= 0 \quad \text{on} \quad \Gamma_{\perp}. \end{aligned} \quad (3)$$

Depending on whether the condition of electric insulation is applied to the boundaries parallel to the x - or y -direction, xy - or yx -polarisation data are obtained, respectively. At the top (Γ_{top}) and bottom boundaries (Γ_{bottom}) we impose boundary values in terms of the magnetic field induced in a horizontally layered half-space (Wait (1953)):

$$\begin{aligned} H_{\perp} &= 1 \text{ Am}^{-1} \quad \text{on} \quad \Gamma_{top}, \\ H_{\perp} &= H_j(z), \quad j = x, y \quad \text{on} \quad \Gamma_{bottom}. \end{aligned} \quad (4)$$

For the construction of the full 3-D MT impedance tensor, both xy - and yx -polarization data are required. Arranging the x - and y -components of the electric and magnetic fields obtained for xy - and yx -polarization such that

$$\mathbf{E} = \begin{pmatrix} E_x(xy) & E_x(yx) \\ E_y(xy) & E_y(yx) \end{pmatrix} \quad (5)$$

$$\mathbf{H} = \begin{pmatrix} H_x(xy) & H_x(yx) \\ H_y(xy) & H_y(yx) \end{pmatrix}, \quad (6)$$

the impedance tensor $\mathbf{Z}$ can be obtained by $\mathbf{Z} = \mathbf{E} \mathbf{H}^{-1}$. The off-diagonal elements of the impedance tensor are therefore

$$Z_{xy} = \frac{E_x(yx)H_x(xy) - E_x(xy)H_x(yx)}{H_x(xy)H_y(yx) - H_x(yx)H_y(xy)}, \quad Z_{yx} = \frac{E_y(xy)H_y(yx) - E_y(yx)H_y(xy)}{H_x(xy)H_y(yx) - H_x(yx)H_y(xy)}. \quad (7)$$

For any period $T = 2\pi/\omega$, the apparent resistivity ρ_a and phase ϕ can then be calculated by

$$\rho_{xy} = \frac{1}{\omega\mu} |Z_{xy}|^2, \quad \rho_{yx} = \frac{1}{\omega\mu} |Z_{yx}|^2 \quad \text{and} \quad \phi_{xy} = \arg(Z_{xy}), \quad \phi_{yx} = \arg(Z_{yx}). \quad (8)$$

In the following, we use the terms *Zxy mode* or *Zyx mode* to distinguish xy - and yx -polarization from quantities derived by a combination of both (Nam et al. (2007)). An important property of time-harmonic EM field is the skin depth $\delta \approx 503\sqrt{\rho/f}$, where δ , the resistivity ρ and the frequency f are given in [m], [Ω m] and [Hz], respectively. is determined, whereas δ , the resistivity ρ and the frequency f are given in [m], [Ω m] and [Hz], respectively.

To solve the boundary value problem (1)-(4), the finite element method (FEM) is applied on unstructured tetrahedral meshes (see Börner (2010); Schwarzbach (2009)). Tetrahedral meshes are well suited for the spatial discretization of arbitrary 3-D geometries which occur when digital elevation models have to be incorporated into numerical simulations. In the 3-D simulations presented here, curl-conforming Nédélec elements with second-order basis functions are employed. In all experiments the FE discretization has been carried out using the Electromagnetics Module of the COMSOL Multiphysics package.

3 Description of the digital elevation model

The area used in the simulations extends from 38.4° to 39.2° N and 14.7° to 15.7° E. At 39° N, the distance between two degrees of latitude is 110.95 km, whereas the distance between two degrees of longitude is 86.51 km. Hence, the model area has a length of 86.51 km in the east-west (x) and 88.76 km in the north-south (y) direction.

We have used two sets of digital terrain data. The ETOPO1 data set is a 1 arc-minute model and provides elevation values for both land and sea. It is available online at the National Geophysical Data Center (<http://www.ngdc.noaa.gov>) and gives a good approximation for the regional bathymetry. Since,

however, its spatial data density is not sufficient to describe Stromboli's topography, a second data set has been used. These data are available from the Shuttle Radar Topography Mission (SRTM), a project to obtain high-resolution topographic data (<http://srtm.csi.cgiar.org/>). The horizontal resolution of the two data sets are listed in the following table. The digital elevation model for the area considered here has been obtained by blending both ETOPO1 and SRTM data sets.

	resolution x -direction	resolution y -direction
SRTM	0.0721 km	0.0925 km
ETOPO1	1.4419 km	1.8491 km

Interpolating the bathymetry data and fitting the two data sets along the coastlines leads to a digital elevation model as depicted in Fig. 2.

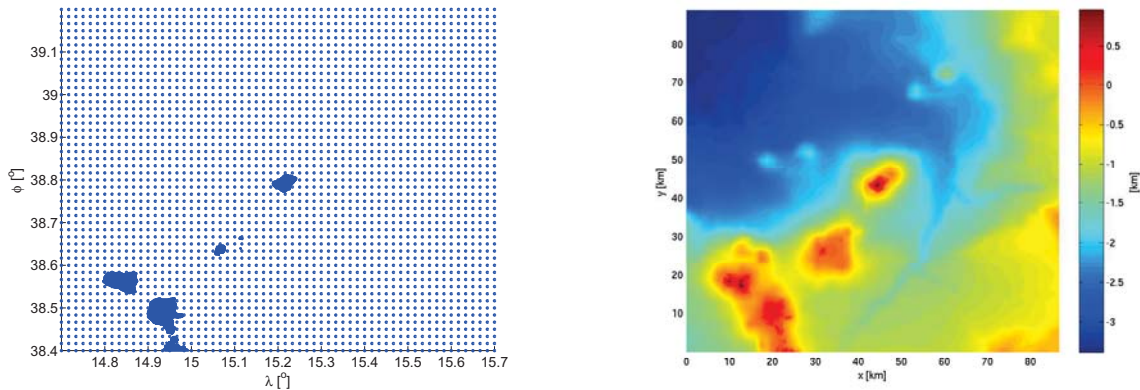


Figure 2: Data points obtained from ETOPO1 (coarse grid) and SRTM (fine grid resembling the islands) for bathymetry and topography (left), digital elevation model obtained by smooth interpolation onto the SRTM grid (right). Stromboli island is located at the centre of both images.

4 Numerical studies

The geometry of the volcanic island of Stromboli is incorporated into the simulation models in three different levels of increasing complexity. This approach enables us to better differentiate which part of the model has an influence on the electromagnetic fields and how distinct this influence is. The first model uses a frustum as the volcanic edifice embedded in a layered background consisting of an air layer, a sea layer and a substratum. In the second model the frustum is replaced with the topography of Stromboli. The edges of this topographic surface have been adjusted to the sea-floor. The third, most complex model uses both topography and bathymetry data.

For the numerical computations three different computer architectures have been used:

computer name	architecture	RAM
<i>klio</i>	4 Intel(R) Xeon(R) @ 3 GHz	16 GB
<i>erato</i>	4 Quad-Core AMD Opteron @ 2.5 GHz	128 GB
<i>RM</i>	56 nodes based on Intel Xeon (E5450 @ 3 GHz)	16 GB per 8-core blade

For each of the three models, apparent resistivities ρ_a and phases ϕ were calculated along profiles running along the sea-floor and over the slopes of the volcano for a period of $T = 10^3$ s. Furthermore, sounding curves were determined for selected sites on the sea-floor and on the volcano to demonstrate the dependency of ρ_a and ϕ on frequency.

4.1 Frustum model

The geometry and appropriate conductivities of the frustum model as well as a map of the locations of the sounding sites are depicted in Fig. 3. Since there are large conductivity contrasts between the

frustum, the sea and the air layer, a non-trivial behaviour of the electromagnetic fields is to be expected. The electric currents are mainly induced in the conductive sea layer and are compressed above the slopes. Moreover, the currents tend to concentrate below the sea surface due to the skin effect. Due to horizontal conductivity contrasts, electric charges accumulate at the interfaces between the volcano and the sea layer. Their surface density is proportional to $-\mathbf{E} \cdot \nabla \sigma$. Therefore, an additional electric field is generated which leads to an increase of the total electric field towards the top of the volcano (around $x = \pm 0.5$ km in the profile, Fig. 4) and to a decrease towards the sea-floor (around $x = \pm 5.6$ km in the profile). This static shift effect results in discontinuities of the apparent resistivity at the base points of the frustum on the sea-floor, at the coastline points and at the top of the volcano. Unlike the apparent resistivity, the phase has a more continuous behaviour and appears to be a rather robust parameter even for pronounced topography. 40 km away from the center of the volcano ($x, y = \pm 40$ km), both phase and apparent resistivity reach their values expected for the half-space (45° , $100 \Omega \text{ m}$).

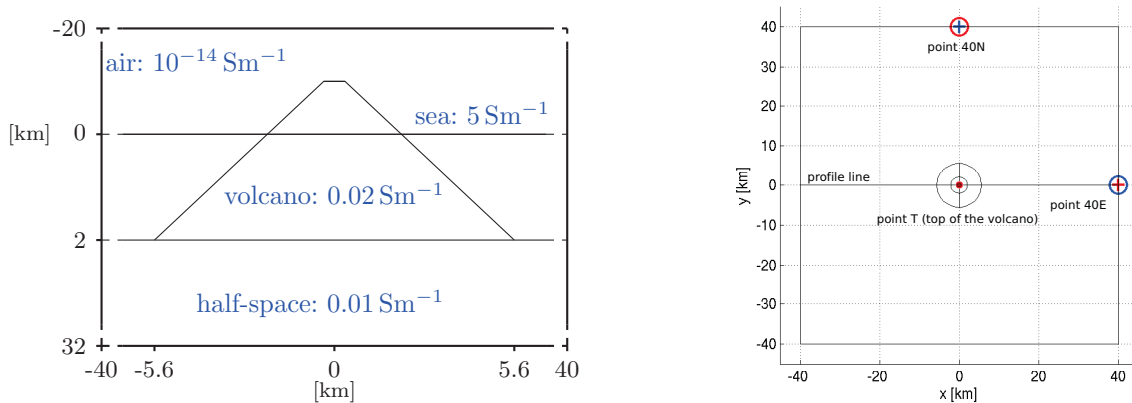


Figure 3: Vertical cross-section through the frustum model (left) and plan view of the model domain (right) showing the locations of the sounding curves (T, N40, E40).

To verify our results, two independent numerical codes have been used to provide reference datasets for the frustum model. More precisely, data have been obtained using a finite element package (FEP) by Schwarzbach (2009), and a finite difference package (FD) by Mackie et al. (1994).

A comparison of the apparent resistivities with those obtained by FEP reveals that they slightly differ near the boundaries of the model and at the lower part of the volcano slopes (Fig. 4, left), where ρ_a is lower for FEP. Moreover, the peaks at the base of the frustum, at the coastline points and at the edges of the plateau are even more pronounced for FEP. Here, we assume an influence of the different spatial discretization of the model. The curves of the phase are nearly congruent for both FE codes.

Apparent resistivities and phases obtained by FD (Fig. 4, right) show similar features in general. However, there are significant differences along the slopes of the volcano. We attribute this to the fact that in the FD code considered here electric field components are located perpendicular to cell faces of a rectangular cartesian grid. Problems in interpolating the discrete electric field components at the desired positions along the volcano slopes are thus to be expected. Finite element techniques on unstructured grids clearly show their superiority in this respect.

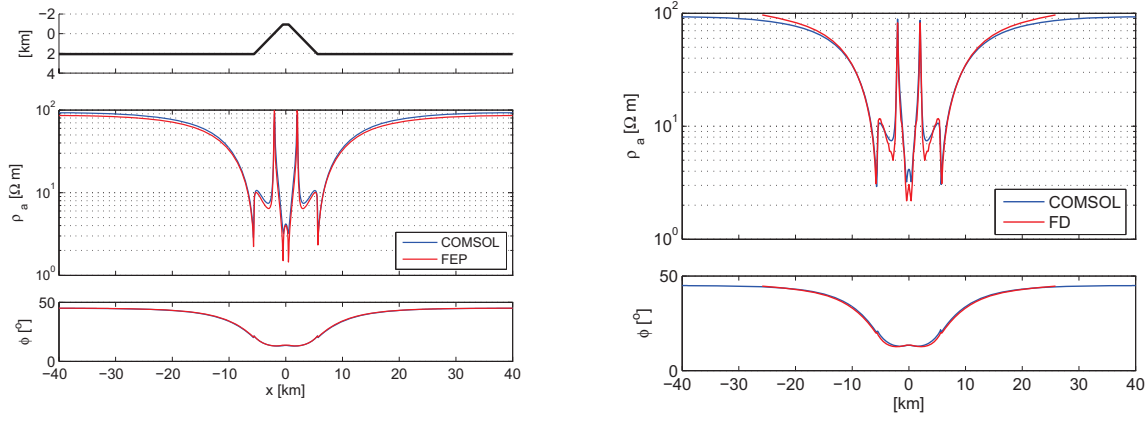


Figure 4: Comparison of the profile curves of the frustum model, $T = 10^3$ s.

Fig. 5 depicts three sounding curves for the sites on top (point T in Fig. 3) and 40 km away from the volcano (E40, N40 in Fig. 3).

Sounding curves for the volcano site T are displayed for periods between 10^{-3} and 10^3 s (Fig. 5, left). For short periods, where the skin depth is small ($\delta \approx 355$ m for $50 \Omega \text{ m}$ and 10^{-2} s), the resistivity of the volcano is reproduced and the phase reaches values around 45° . For periods between 10^{-1} s and 10^2 s, the influence of the conductive sea layer is dominant and therefore, the apparent resistivity decreases whereas the phase increases. For longer periods the resistive half-space dominates the curves ($\delta \approx 16$ km for $100 \Omega \text{ m}$ and 10^3 s). The sounding curves for the two sites on the sea-floor are displayed for a subset of the full period range (between 10^1 and 10^3 s) because of the strong attenuation within the very conductive sea water (Fig. 5). High frequencies generally require a much finer spatial discretization in regions where the skin depth is small. However, the computational experiments have been carried out on a mesh well suited for mid to low frequencies. Results for high frequencies were not satisfactory for the coarse meshes used here, and thus have not been included in the comparison.

The long period ends of the sounding curves nicely approach the values of the lower half-space, i.e. $100 \Omega \text{ m}$ for ρ_a and about 45° for ϕ . For short periods there are deviations from these values, which are larger for the points on the profiles oriented perpendicular to the component of the driving electric field, i.e. for point E40 at $y = 40$ km for ρ_{xy} and the point at $x = 40$ km for ρ_{yx} (cross markers in Fig. 5, cf. Fig. 3, right). As mentioned above, these deviations are due to the insufficient spatial discretization for short periods in regions of strong attenuation.

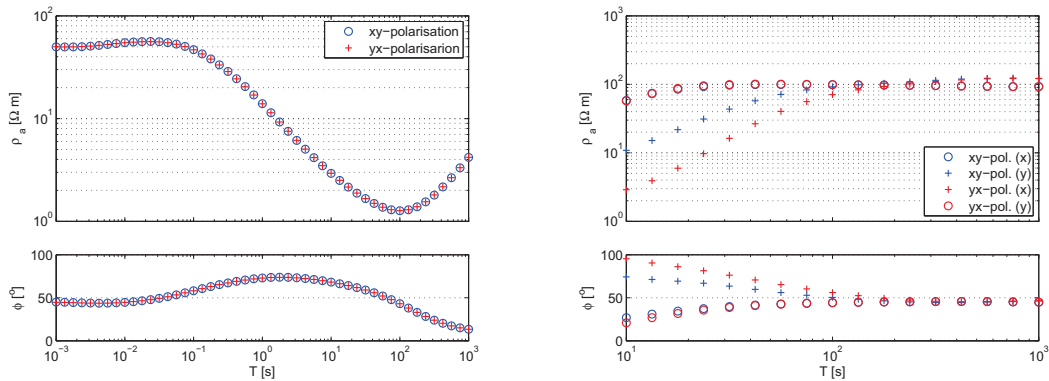


Figure 5: Sounding curves for the volcano site T (left) and the sea-floor sites E40 and N40 (right) for the frustum model. Here, xy -pol. and yx -pol. refer to ρ_{xy} and ρ_{yx} , resp.

The following table summarizes important computational parameters for the three different codes, such as polynomial degree of the basis functions (BFD) for the FE method and degrees of freedom (DOF). The overall CPU time required to obtain the result for one frequency was in the order of two minutes. The total time required to compute all field components for one polarization sums up to

about two hours for all 49 frequencies for the FE codes and 30 minutes for the FD code. Note that the 3-D impedance requires the calculation of both polarizations (see eq. (7)).

	computer	CPUs	BFD	elements	DOFs	time [min]
COMSOL Multiphysics [®]						
<i>xy</i> -pol.	<i>klio</i>	4	quadratic	82 661	523 580	119,56
FEP						
<i>xy</i> -pol.	<i>erato</i>	8	quadratic	57 666	434 456	119,07
FD code						
	FD code	131 x 131 x 154 grid cells + 15 air layers				≈ 30

4.2 Stromboli-Topography model

In the second level of complexity, a realistic topography of Stromboli is incorporated. Since this topography is smoothly adjusted to the flat sea-floor, the edifice is larger in this model than in reality. Furthermore, the conductivity of the volcano has been reduced to 0.01 Sm^{-1} .

Figs 6 and 7 illustrate the model with the positions of the profiles and the locations for which sounding curves have been calculated. Sounding curves are shown for sites *a*, *N* and *E* only.

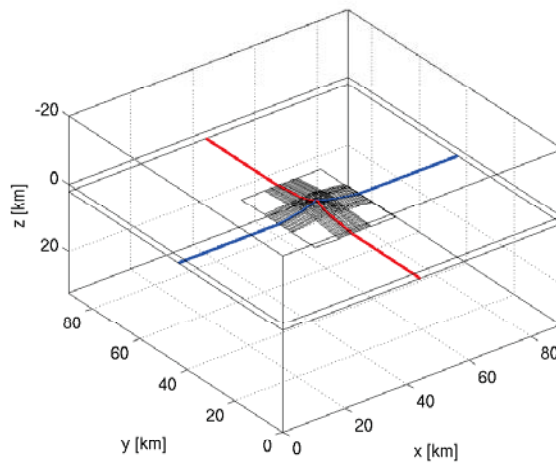


Figure 6: Flat sea-floor model including the real topography of Stromboli island. Red and blue lines indicate the *x*- and *y*-profiles, respectively.

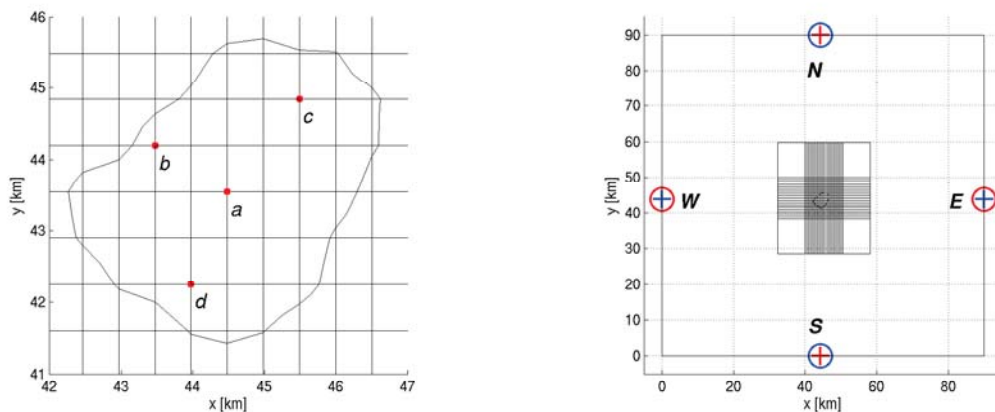


Figure 7: Locations for which sounding curves are computed are denoted by *a*, *b*, *c*, *d* on the island (left) and *N*, *E*, *S*, *W* on the sea-floor (right).

Fig. 8 displays the apparent resistivities ρ_{xy} and phase Φ_{xy} for $T = 10^3 \text{ s}$ along two profiles aligned with the *x*-axis (*x*-profile) and the *y*-axis (*y*-profile). The apparent resistivity for the *x*-profile is associated

with the xy -impedance tensor elements (Z_{xy} mode) while the apparent resistivity for the y -profile is derived from the yx -elements (Z_{yx} mode). The curves for both Z_{xy} and Z_{yx} modes show similar features. This can be explained by the similar topography along the two different profiles. Nevertheless, the extent of the volcano's edifice in y -direction is larger than in x -direction. Hence, the corresponding minima and maxima of the curves are shifted relative to each other.

Due to the smoother transition of the topography from the sea-floor to the volcano slopes the peaks at the base points of the edifice are not as pronounced as for the frustum model. Since there is no volcano plateau in the x -profile, there is just one minimum at $x \approx 44$ km for ρ_{xy} .

As for the frustum model, the sounding curves for the volcano (e.g. for site a in Fig. 9, left) reflect the layering of the earth with regard to the electrical conductivity whereas long periods correspond to large skin depths. Since the same conductivity is assigned to the volcano and the half-space the sounding curves reach values around $100 \Omega \text{ m}$ for short periods.

Due to the asymmetry of the model, offsets occur between the curves derived from Z_{xy} and Z_{yx} that are higher for ρ_a than for ϕ . Hence, it can be assumed that the phase is less affected by these two aspects.

The sounding curves for the sea-floor are displayed in Fig. 9 (right). Since the edifice is larger than the one represented by the frustum the long period apparent resistivities deviate more significantly from the undisturbed half-space values.

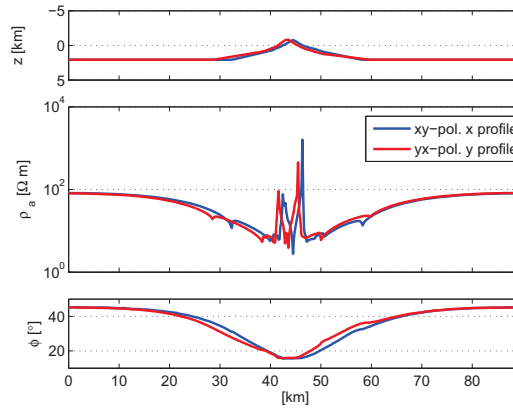


Figure 8: Topography (top), apparent resistivity (center), phase (bottom), derived from Z_{xy} and Z_{yx} (xy - and yx -pol., resp.) along the x - and the y -profile for $T = 10^3$ s.

The differences in the short period range may again be attributed to the insufficient spatial discretization.

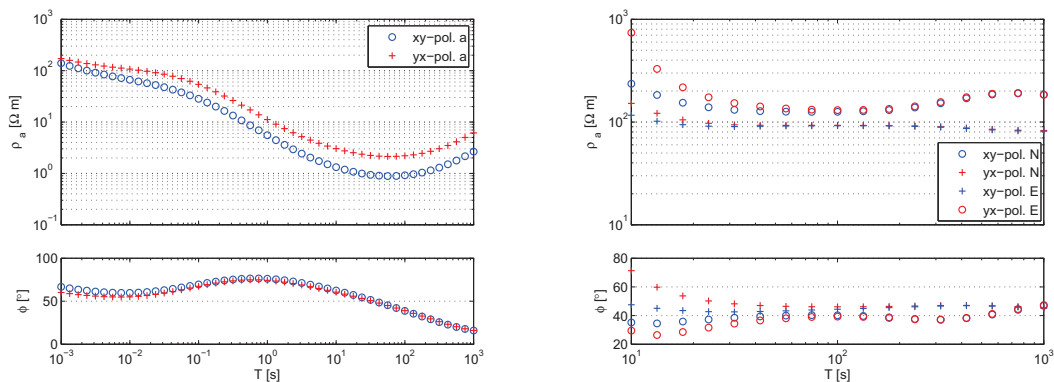


Figure 9: Sounding curves for the island site a (left) and for the sea-floor sites N and E (right). xy -pol. and yx -pol. refer to Z_{xy} and Z_{yx} , resp.

The incorporation of the detailed Strombolian topography leads to a massively increasing number of mesh points. In order to keep computational costs low the meshes for computing the full sounding

curves were coarsened.

	Computer	CPUs	BFD	No. of elements	DOFs	Time [min]
profile curves						
<i>xy-pol.</i>	<i>erato</i>	8	quadratic	443 563	2 823 488	114.78
sounding curves						
<i>xy-pol.</i>	<i>erato</i>	8	quadratic	164 662	1 048 116	≈ 721

4.3 Bathymetry-Topography model

In the third level of complexity, the topography of the island is incorporated as well as the regional bathymetry including the other Liparian islands. This allows to analyse the effect on the electromagnetic fields not only due to the volcano itself but also due to the variable thickness of the sea layer.

Figs 10 and 11 illustrate the model and the locations of the profiles and points for which sounding curves have been calculated. As an example, apparent resistivities and phases for a period of 1000 s are shown for the xN - and xS -profiles as well as the full sounding curves for points a, c, N and S.

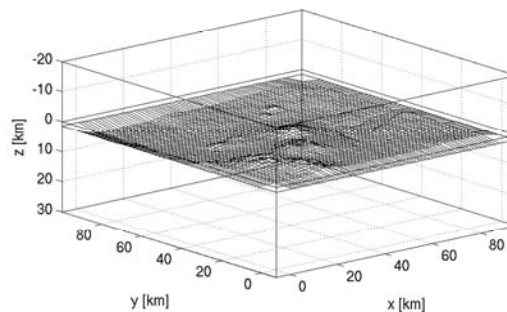


Figure 10: Model including the regional bathymetry and the topography of Stromboli and the Liparian Islands.

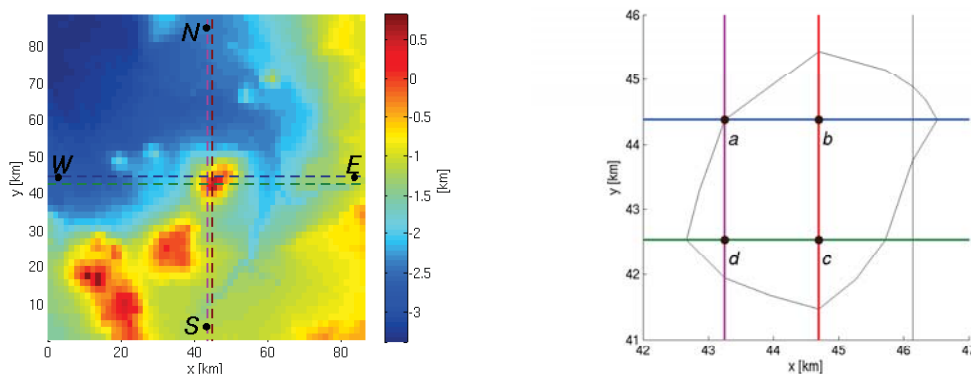


Figure 11: Left: Elevation map inferred from digital terrain data showing the locations of seafloor sounding points N, E, S, W and the profile lines xN , xS in the EW-direction and yE , yW in the NS-direction. Right: Coast line of Stromboli and locations of sounding points a, b, c, d.

The xN and xS profiles running East-West in parallel (Fig. 12) are close to each other. Hence, the features of the respective apparent resistivity and phase curves are similar. As for the previous models, peaks at the coastline and at the top of the volcano can be observed. However, serious perturbations are provoked by the regional bathymetry. Induced currents are vertically compressed in the shallow parts of the ocean yielding higher apparent resistivities. Vice versa, apparent resistivities are decreased in deep water areas. The soundings at site c (Fig. 13, right) again show the influence of the volcano,

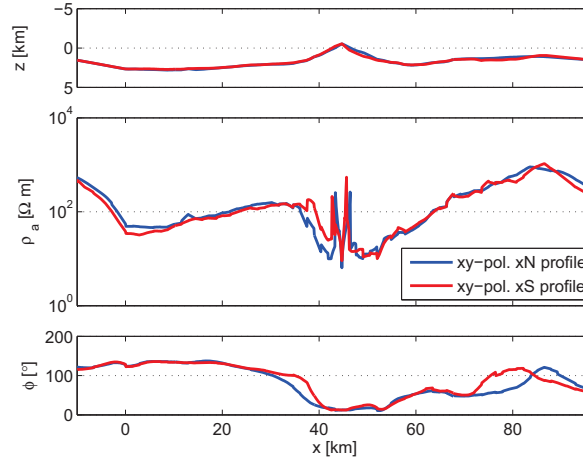


Figure 12: Topography (top), apparent resistivity (center) and phase (bottom) for $T = 10^3$ s at profiles xS and xS . xy -pol. and yx -pol. refer to Z_{xy} and Z_{yx} , resp.

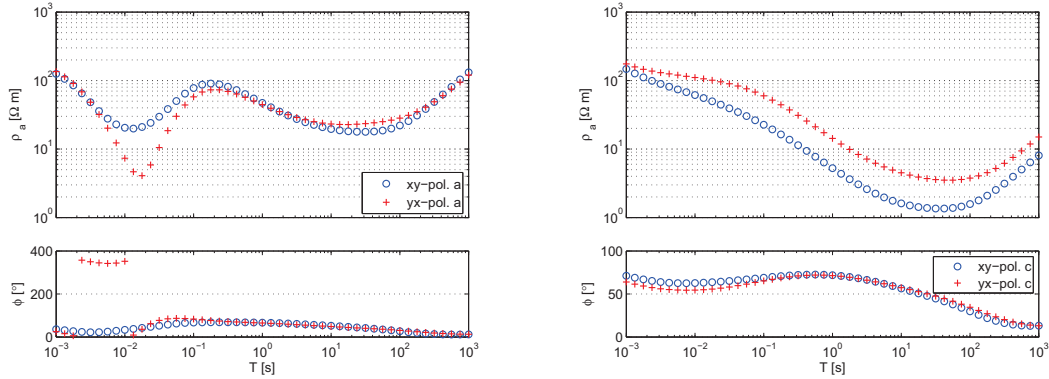


Figure 13: Sounding curves for the island sites a (left) and c (right). xy -pol. and yx -pol. refer to Z_{xy} and Z_{yx} , resp.

the sea layer and the half-space for the appropriate period ranges. However, at site a , which is located directly at the coastline, the sea layer affects the curves already for shorter periods, especially for the Z_{yx} mode. Still, for periods longer than $T = 10^{-1}$ s the shapes of the curves are similar to those at point c (although shifted to higher ρ_a -values).

All curves derived from Z_{xy} and Z_{yx} show an offset which can be attributed to the asymmetric elevation data used in the MT simulations.

For the two sea-floor sites N and S the topography effect varies for the two cases considered (Fig. 14). Site N (Fig. 14, left) is located at greatest distance from the islands. Therefore, the curves are similar to those obtained using the simpler models and for longer periods they are almost assume the corresponding half-space values.

The sounding curves at site S (Fig. 14, right) seem to be affected by two different facts. For ρ_{xy} the graben-like structure in this area (cf. Fig. 11, left) leads to lower apparent resistivities. On the other hand, the more resistive islands affect the ρ_{yx} -curves.

We note that the out-of-quadrant phases in Fig. 14 occur mainly at the high frequencies, which indicates that the mesh used for the numerical computations is still not fine enough to obtain results with the desired accuracy.

The following table summarizes the mesh properties and CPU times associated with the bathymetry-topography model.

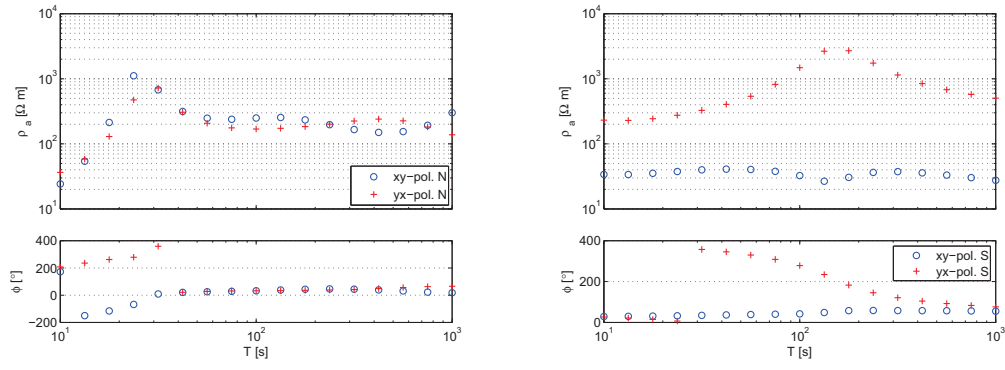


Figure 14: Sounding curves for the sea-floor sites *N* (left) and *S* (right). *xy-pol.* and *yx-pol.* refer to Z_{xy} and Z_{yx} , resp.

	Computer	CPUs	BFD	No. of elements	DOFs	Time [min]
profile curves						
<i>xy-pol.</i>	<i>erato</i>	16	quadratic	546 759	3 479 830	44.41
sounding curves						
<i>xy-pol.</i>	<i>erato</i>	16	quadratic	312 721	1 987 514	≈ 1103

5 Conclusions

In this work, distortion effects on MT data caused by topography and bathymetry are examined for the area around Stromboli. The geometry of Stromboli has been incorporated into FE simulations applying three different levels of increasing complexity.

In the first model, a frustum representing the volcano has been embedded in a layered background. This model yields a non-trivial apparent resistivity profile curve which is mainly dominated by peaks at the base points of the frustum, at the coastline points and at the edges of the volcano plateau. These are caused by static shift effects and the high conductivity contrast between the volcano and the sea layer. The results of the frustum model have been verified by a finite element code by Schwarzbach (2009) and a finite difference code by Mackie et al. (1994).

For the second model of Stromboli digital elevation data of the island have been included in the model geometry. Compared to the frustum model, the apparent resistivity and phase curves along the profiles and the sounding curves for the volcano show similar features.

Finally, digital bathymetry data have been added to the third, most complex Stromboli model to take into account realistic topography and bathymetry features of the region. The apparent resistivity still shows the characteristic peaks at the coastline points for most of the profiles. The effects of the bathymetry and topography of the surrounding islands are superposed, which results in very complicated profile and sounding curves.

These results clearly point out that topography and bathymetry can heavily distort MT data and thus need to be considered for an accurate numerical interpretation of MT measurements.

Finally, we emphasize that FE methods on unstructured grids provide a sophisticated tool for treating problems associated with complex geometry.

Acknowledgements

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Establishing Controlled Source MT at GFZ

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Summary

The multidisciplinary GeoEn project (the Brandenburg pilot project in the BMBF program „Spitzenforschung und Innovation in den Neuen Ländern“) integrates research in the fields of geothermal energy, carbon dioxide capture, transport and storage (CCTS) as well as the exploration of unconventional gas reserves (shale gas). The electrical conductivity is one key parameter to characterize reservoirs and to monitor changes due to circulation or injection of fluids at reservoir depth. Bulk electrical conductivity is highly sensitive to fluids within interconnected pores, and therefore EM techniques are powerful tools for exploring and monitoring geothermal reservoirs, CO₂ storage sites and shale gas reservoirs. In the framework of the GeoEn project, we establish Controlled Source MT (CSMT) at GFZ Potsdam. We aim at combining active and passive MT to image the electrical conductivity structure within the Earth, ultimately in 3D.

For CSMT, we intend to use grounded electrodes to inject a frequency-dependent current into the Earth and measure the induced electric and magnetic fields at near-field to far-field distances. We will use novel transmitter systems from Metronix (Braunschweig) which are presently under development. Standard MT receivers will be utilized to measure the induced electric and magnetic fields. In the scope of the project, we will (i) assemble the transmitter system and the source dipole, (ii) optimize and design CSMT field procedures for geothermal exploration, carbon dioxide reservoir characterization and shale gas exploration (Streich et al., 2010a), (iii) develop and implement time-series processing, (iv) develop and implement 1D modeling (Streich and Becken, 2009) and inversion software and 3D modeling codes (Streich, 2009; Streich et al., 2010b).

In this contribution, we describe a test of long steel electrodes as the current electrodes of the source dipole, and we examine the resolution power of CSMT using 1D inversion of synthetic data.

Current electrodes of CSMT source dipole

Low grounding resistances of the current electrodes are crucial for injecting strong currents into the subsurface. The Metronix 22 kVA transmitter generates currents of max. 40 A (560 V), which can, however, only be achieved if the total resistance of the dipole (1-km long cable and grounding electrodes) is less than 14Ω. We use a thick cable that has a resistance of 2Ω/km. Accordingly, to achieve maximum currents, the electrode grounding resistances should be less than 12Ω. At high frequencies, the maximum current will be further limited by the inductance of the cable.

The grounding resistance depends primarily on the surface of the electrode that is in contact with the ground, and the surrounding resistivity within the Earth. For a homogenous earth, the grounding resistance R of a steel rod is

$$R = \frac{\rho}{2\pi L} \ln \left(\frac{4L}{a} - 1 \right), \quad (1)$$

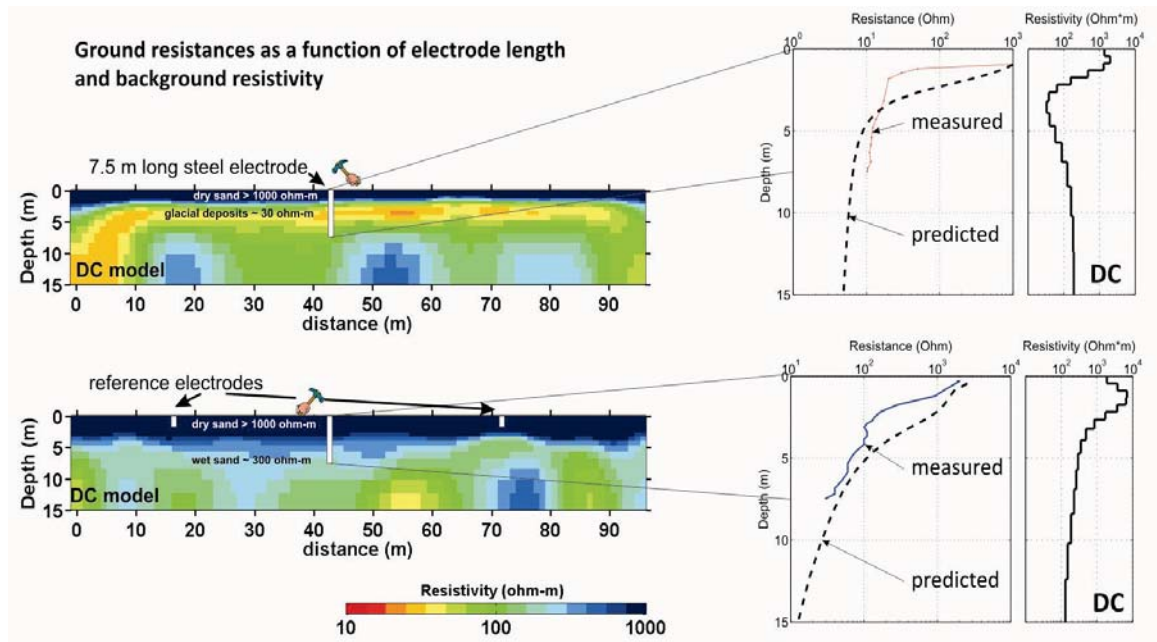


Figure 1: Electrode test. We measured the ground resistance of a steel electrode (diameter 32 mm) vs. depth of the rod at two different locations, which may represent typical conditions in northern Germany. Ground resistance measurements were taken against two reference electrodes at ~30 m distance to the left and right. DC Wenner spreads were measured at the same locations (left panels). The panels to the right show the measured resistances (red and blue lines), compared to theoretical predictions (dashed black lines) based on the resistivity-depth profiles at the electrode locations shown in the rightmost plots. The sudden drop in resistance at <2m depth is mainly attributed to the penetration of the groundwater table. Further resistance reduction is due to increased surface of the electrode with increased penetration depth.

where L is the length of the rod in the ground, a is the diameter, and ρ is the earth resistivity. Note that the resistance decreases slowly for increasingly large electrode lengths L . In practice, we plan to use up to 15-m long steel electrodes with $a = 3.2$ cm, which would result in a grounding resistance

$$R \approx \frac{\rho}{12.5 m} (\Omega), \quad (2)$$

assuming homogeneous ground resistivity. This suggests that $\rho < 75 \Omega\text{m}$ would represent suitable conditions at the grounding point to achieve sufficiently small resistances $R < 6 \Omega$ with 15 m long electrodes. However, we hope that the steel rods penetrate conductive material at shallower depth (e.g. groundwater), which would further reduce the resistance or the required electrode length.

We have tested the grounding resistance of long steel electrodes in a geoelectrically known environment near Nauen, Germany (Figure 1). We rammed two electrodes into the ground and measured the resistances every ~0.3 m against two 2-m long reference electrodes. As expected, the resistances were observed to decrease with increasing penetration depth, starting at >3 k Ω at <1m and decaying approximately exponentially with depth. At ~7.5 m depth, we achieved ~10 Ω at one test location (upper panels in Figure 1), and ~30 Ω at the other test location (lower panels).

The approximate subsurface resistivities at the test locations were determined from 2D DC resistivity sections. At both sites, high resistivities within the upper 2-3 m correspond to a dry sand layer. At

greater depths, we penetrate conductive glacial till at test location 1 (upper panel in Figure 1) and a groundwater layer at test location 2 (lower panel). The measured grounding resistance values are roughly consistent with predictions (dashed lines) based on the geometry and the subsurface resistivity distribution. For resistance predictions, we approximated the resistance R from a parallel circuit of steel rod segments of length ∂L embedded into a medium with piecewise constant resistivity ρ_i . Approximate ρ_i values were extracted from the 2D DC resistivity-depth sections. Let the individual rod pieces have resistance

$$R_i = \frac{\rho_i}{2\pi\partial L} \ln\left(\frac{4\partial L}{a} - 1\right), \quad (3)$$

then the resistance of the entire rod to a depth L is given by

$$\frac{1}{R} = \sum_i \frac{1}{R_i}. \quad (4)$$

Equations 3 and 4 were used to estimate grounding resistances from the 2D resistivity models (Figure 1).

This test shows that long steel rods may be suitable electrodes for CSMT source dipoles. At one test location with favorable geology (glacial till), we achieved a ground resistance of $\sim 10 \Omega$ with electrodes penetrating 7.5 m deep; the second test location, a sandy soil with a freshwater layer at ~ 3 m depth, may require deeper electrodes. Our predictions suggest that the ground resistance of the steel rod may drop at this location to values of $\sim 10 \Omega$ at ~ 15 m depth. The simple model used to predict grounding resistance has proven to be useful for practical purposes. Before installing electrodes, it may be advisable to investigate potential locations with DC resistivity soundings in order to predict expectable resistances and required electrode lengths.

1D CSMT Inversion

Vertical currents, galvanically injected into the subsurface with a grounded dipole source, exhibit sensitivity to buried resistors at depth. This makes the CSMT technique suitable for imaging resistive layers, whereas both passive MT and CSMT are sensitive to conductive layers. The effect of anomalously resistive subsurface structures on surface CSMT data is typically smaller in land applications than in deep-water marine applications. Land applications of frequency-domain CSMT suffer from energy travelling through the air even more than marine applications, which obscures the response from deeper targets. Nevertheless, forward modeling studies suggest that in many cases, the e.m. response of thin resistive layers is above noise level (see Streich and Becken, this issue).

A practical question is to what extent the resistivity models giving rise to these anomalies can be recovered from measured data. We ran 1D Occam-type inversions to investigate the resolution power of CSMT data. To infer the resistivity structure from 1D inversion, we can exploit the spatial decay of CSMT fields with increasing distance to the source, and the frequency dependency of the fields. In land-based applications, we will typically have only few source locations combined with a relatively large number of receivers (say, 100), and potentially long transmitting times that allow us to cover a broad frequency range. In contrast, marine applications use a source towed continuously over a number of receivers, effectively yielding many source point locations, but only short transmitting times and thus a narrow frequency band for a given source location. Hence, land applications will primarily utilize the frequency

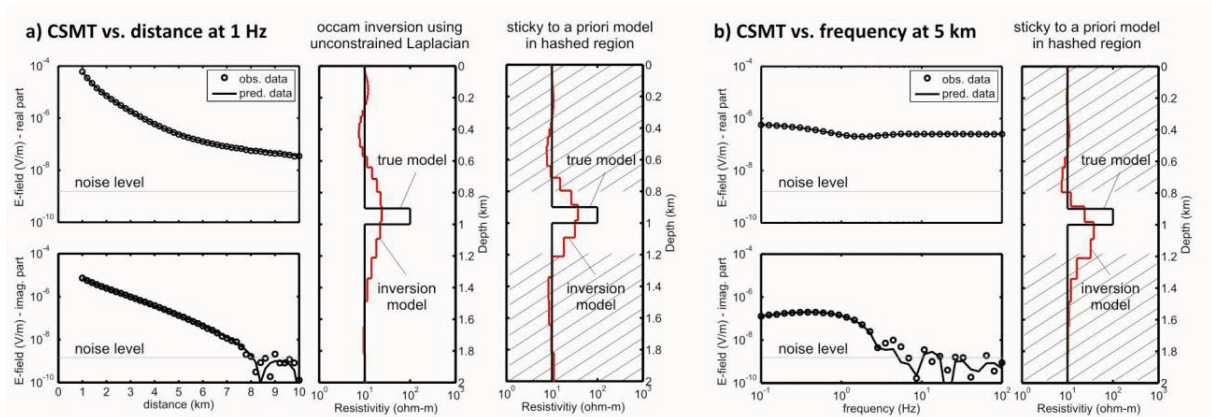


Figure 2: 1D Occam inversion of synthetic CSMT data. The data are the real and imaginary parts of the inline electric field, contaminated with white noise with 1% of the signal amplitude. In addition, white noise with a standard deviation of 10^{-9} V/m was added. **a)** Inversion of the spatial decay of the field for a single frequency (1 Hz), using an unconstrained Occam inversion and using an Occam variant where the deviation from an a priori model is penalized outside of the depth of interest. **b)** Inversion of the frequency spectrum recorded at 5 km distance from the transmitter.

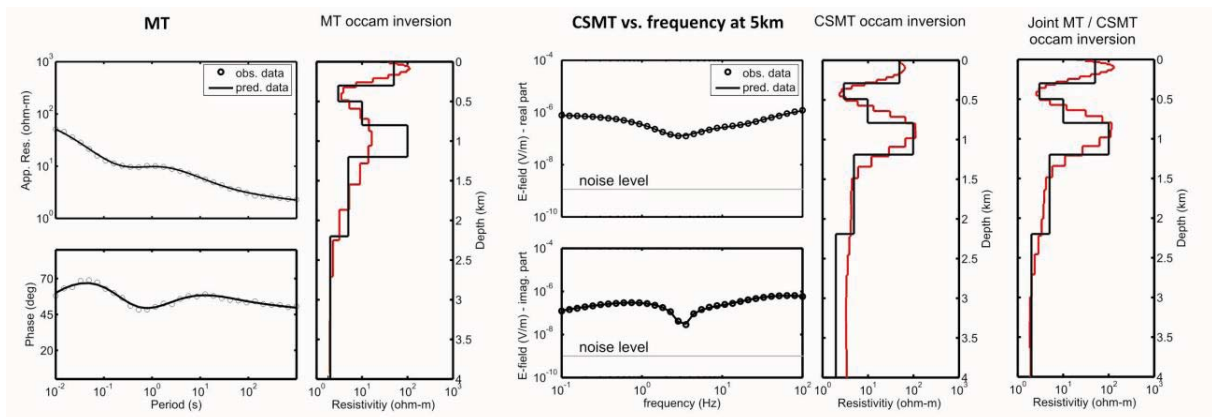


Figure 3: Comparison of single and joint 1D inversion models for passive MT and CSMT data. In (a), the left panel shows passive MT data (apparent resistivity and phase) for the true model and MT inversion result displayed in the right panel. In (b), CSMT data (real and imaginary part of the inline electric field) are displayed for the true resistivity model and CSMT inversion result. (c) shows the resistivity model resulting from joint MT and CSMT inversion. In this example, the MT data add information only at greater depths, where the sensitivity of CSMT sounding is low. In practice, we anticipate that MT data will help construct a (2D/3D) regional background model, and CSMT data will help refine the model at reservoir scale (1D/3D).

dependency of the fields, whereas marine applications will primarily exploit spatial variations of the fields.

We have implemented a 1D CSMT inversion based on Weidelt's formulation of the forward problem (Weidelt, 2007), and analytically derived expressions for the Jacobian matrix required in the inversion. Finite source dipoles (Streich and Becken, this issue) have been incorporated into the inversion; however, in the present form of the inversion scheme, the source and receivers are placed at the surface. Here, we only consider horizontal electric point dipole sources. The inversion is similar to

Occam inversion (Constable et al., 1987); however, Occam's second phase (i.e., the determination of the regularization parameter that yields optimal trade-off between data residuals and model norm) has not been implemented. Instead, we fix the regularization parameter for one entire inversion run.

Using this inversion scheme, we found that both the inversion of multi-offset data at a single frequency and multi-frequency data at a single receiver location have the power to yield comparable inversion models, if the frequency range and the distance to the transmitter are appropriate for imaging the target depth (Figure 2). The left panels in Figure 2a and b depict the inline electric field as a function of distance from the transmitter for the frequency $f = 1$ Hz and as a function of frequency at 5 km distance, respectively, for a model containing a resistive layer embedded into a homogeneous half-space. Random noise of 1% of the field amplitudes and a noise floor of 10^{-9} V/m were added to the data prior to inversion. The central panel in Figure 2a shows the inversion model obtained from smooth inversion using a standard Laplacian penalty on the logarithmic model resistivities (red line). A similar result (not shown) was obtained from inversion of the multi-frequency electric field data shown in Figure 2b. In both cases, the recovered 1D model contains a resistive zone that is smeared over a wide depth range, because regularization penalizes thin model layers.

For monitoring applications (e.g., CO₂ injection), the depth of interest is known. A focused inversion that searches for deviations from an a priori model primarily at reservoir depth may help improve the model. We have therefore incorporated a term into the Occam inversion that allows us to weight the penalty depending on the difference of every inversion model parameter to an a priori model (Key, 2009). The right panels in Figure 2a and b show the result of this variant of regularized inversion. The hashed areas correspond to depth levels where the inversion model is constrained to deviate minimally from the a priori model (which is the true model in this case), whereas the rest of the model was permitted to vary freely (in the sense of a Laplacian regularization). This approach clearly helps focusing the resistive anomaly at the true depth and may thus be adequate for a focused model search.

The penetration depth of CSMT data is limited by the lowest frequencies used and the greatest offsets providing reasonable signal quality. Combining CSMT with passive MT may prove helpful to expand the model scale. Furthermore, both techniques exhibit different sensitivities to resistive and conductive structures (and to deviations from 1D media) and thus provide complementary information. We have therefore implemented a joint 1D inversion for both data types. In Figure 3, we compare the 1D inversion models for a multi-layered structure obtained from MT data and CSMT data alone and from joint CSMT and MT inversion. Here, the CSMT data are the frequency-dependent inline electric field data, measured at 5 km offset from the transmitter. The individual inversion models show that (i) both techniques resolve shallow conductive layers, (ii) the CSMT data resolve a resistive layer at 1 km depth, and (iii) the MT data resolve the conductivity of the basal layer that is too deep to be sensed with CSMT. All of these model features are revealed by joint inversion, suggesting increased resolution capabilities from combined applications of both techniques.

Acknowledgements

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2D-SIP-Modellierung mit anisotropen Widerständen

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1 Einleitung

Die Modellierung von Messdaten der spektralen induzierten Polarisation (SIP) geschieht üblicherweise unter der Annahme makroskopisch, also in der Größenordnung einer Zelle der Modellrechnung, isotroper komplexer Widerstände im Untergrund. Für diese Art der Modellierung von Daten der SIP wurde von A. Kemna das Finite-Elemente-Programm CRMod entwickelt ([1], [2]). Mit der Annahme isotroper Widerstände lassen sich viele Messdaten hinreichend erklären. Nguyen et al. ([3]) zeigten jedoch für den reellwertigen gleichstromgeoelektrischen Fall, dass, wenn der Untergrund aus vielen abwechselnd gelagerten Schichten besteht, die eine geringe räumliche Ausdehnung gegenüber der bei der Inversion benutzen Gitterzellen haben, die Annahme isotroper Leitfähigkeiten zu Fehlinterpretationen führen kann. Hier kann die Erweiterung auf anisotrope Widerstände Abhilfe schaffen, weil diese auf makroskopischer Skala, also Zellenebene, Schichtlagerungen beschreiben können. Das Auftreten von Anisotropieeffekten auch bei SIP wurde von Winchen et al. ([4]) demonstriert. Das Programm CRMod wurde zum Zweck der Modellierung anisotroper komplexer Widerstände erweitert. In diesem Artikel sollen anhand von verschiedenen modellierten Beispielen verschiedene Effekte eines Untergrund mit anisotropen komplexen Widerständen gezeigt werden.

2 Methodik

2.1 Spektrale Induzierte Polarisation (SIP)

Die spektrale induzierte Polarisation (SIP) ist ein elektrisches Verfahren, das wie die Geoelektrik aus dem Messen von Potentialverläufen künstlicher elektrischer Quellen Rückschlüsse auf die Leitfähigkeitsverteilung im Untergrund zieht.

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Das Aufnehmen der Messwerte geschieht in einem wiederum der Geoelektrik sehr ähnlichen Aufbau, jedoch werden zusätzlich Phasenbeziehungen zwischen Sende- und Empfangssignal aufgezeichnet. Als Quellsignal werden Wechselströme unterschiedlicher Frequenzen gewählt. Die Messdaten sind dann Spannungswerte und deren Phasenverschiebungen zum Quellsignal. Aus diesen Daten und den Positionen der Elektroden ergeben sich mit den Geometriefaktoren die Messwerte als scheinbare spezifische Widerstände (Amplituden) und deren Phasen. Diese lassen sich zur Abschätzung der Eigenschaften des Profils als Pseudosektion auftragen.

Das physikalische Problem der spektralen induzierten Polarisierung und der Geoelektrik lässt sich für den Fall anisotroper komplexer Widerstände in x-, y- und z-Richtung in Form der Poisson-Gleichung

$$\partial_x(\sigma_x \partial_x \phi) + \partial_y(\sigma_y \partial_y \phi) + \partial_z(\sigma_z \partial_z \phi) + I \delta(x - x_s) \delta(y - y_s) \delta(z - z_s) = 0 \quad (1)$$

mit dem Potential $\phi = \phi(x, y, z)$, dem Strom I und der Diracschen Delta-Funktion δ schreiben. Die Anisotropie beschränke sich dabei auf die Diagonalelemente des Leitfähigkeitstensors

$$\underline{\underline{\sigma}} = \begin{pmatrix} \sigma_x & 0 & 0 \\ 0 & \sigma_y & 0 \\ 0 & 0 & \sigma_z \end{pmatrix}. \quad (2)$$

Die Stromeinspeisung findet an den Koordinaten (x_s, y_s, z_s) statt.

2.2 Modellierung

Für die Modellierung wird das oben aufgeführte Problem der Potentialverteilung durch einen 2D-Finite-Elemente-Algorithmus gelöst (vgl. [2]). Die Ergebnisse sind Potentialverläufe und zugehörige Phasen an jedem Gitterpunkt. Aus diesen Informationen und den Messelektrodenpositionen lassen sich damit die zum Modell gehörigen synthetischen Messdaten - Amplituden und Phasen der scheinbaren spezifischen Widerstände - berechnen und in Pseudosektionen auftragen. Das hier benutzte Finite-Elemente-Programm CRMod ist von einer bereits bestehenden Version für die Modellierung isotroper spezifischer Widerstände und Phasen ([2]) um die Möglichkeit der Modellierung von Anisotropie erweitert worden.

Die Poisson-Gleichung 1 wird unter der Annahme eines konstanten Leitfähigkeitstensors in y-Richtung durch die Fourier-Transformation in den Wellenzahlbereich mit

$$\begin{aligned} F(k) &= \int_{-\infty}^{\infty} f(y) e^{iky} dy \\ &= \int_{-\infty}^{\infty} f(y) \cos(ky) dy + i \int_{-\infty}^{\infty} f(y) \sin(ky) dy \end{aligned}$$

und der Wellenzahl k transformiert. Bei Reduktion auf Funktionen, die in y-Richtung zum Nullpunkt symmetrisch sind, fällt das zum Nullpunkt $y = 0$ antisymmetrische Teilintegral weg, so dass sich die Fourier-Kosinus-Transformation

$$F(k) = \int_{-\infty}^{\infty} f(y) \cos(ky) dy \quad (3)$$

ergibt. Das Ergebnis der Transformation der Gleichung 1 in den Wellenzahlbereich entlang der y-Achse ist dann

$$\partial_x(\sigma_x \partial_x \tilde{\phi}) - \sigma_y k^2 \tilde{\phi} + \partial_z(\sigma_z \tilde{\phi}) + \frac{I}{2} \delta(x - x_0) \delta(z - z_0) = 0 \quad (4)$$

mit $\tilde{\phi} = \tilde{\phi}(x, k, z)$. Diese Gleichung wird von dem hier verwendeten Finite-Elemente-Programm auf einem zweidimensionalen Gitter in x - und z -Richtung gelöst. Aus den erhaltenen Werten für $\tilde{\phi}$, bzw. nach Fourier-Kosinus-Rücktransformation ϕ lassen sich mit den Geometriefaktoren der Elektrodenkonfiguration die synthetischen Messwerte für Amplitude und Phase der scheinbaren spezifischen Widerstände angeben.

2.3 Bewertung der Genauigkeit

Das Programm muss auf seine Gültigkeit und Genauigkeit in Bezug sowohl auf analytisch berechenbare Probleme als auch auf verschiedene, mit dem bisherigen Programm zum Modellieren isotroper spezifischer Widerstandsverteilungen berechnete Modelle geprüft werden. Für einen homogenen Halbraum mit spezifischem Widerstand ρ_h und Phase ϕ_h erwartet man eine Pseudosektion mit dem konstanten Wert $\rho = \rho_h$ des scheinbaren spezifischen Widerstands und $\phi = \phi_h$ für die scheinbare Phase. Die Messwerte werden mit einer simulierten Dipol-Dipol-Anordnung mit Elektrodenabstand 1 m berechnet. Die Abweichungen der Modellierungsergebnisse zum analytischen Ergebnis betragen weniger als 10 % in den scheinbaren spezifischen Widerständen und weniger als 1 % in den scheinbaren Phasen. Ein homogener Halbraum mit anisotropen spezifischen Widerständen in Horizontal- und Vertikalrichtung, $\rho_{horizontal}$ und $\rho_{vertikal}$, weist nach analytischen Berechnungen einen scheinbaren spezifischen Widerstand von

$$\rho = \sqrt{\rho_{hori.} \cdot \rho_{vert.}} \quad (5)$$

(nach z.B. [5, S. 95]) auf. In Abb. 1 ist ein Modellierungsergebnis für diesen Fall dargestellt. Der spezifische Widerstand des Halbraums beträgt 100 Ωm in horizontaler Richtung (x- und y-Richtung) und 25 Ωm in z-Richtung. Dies entspricht einem scheinbaren spezifischen Widerstand von 50 Ωm in der Pseudosektion. Die Phase des Halbraums beträgt gleichmäßig (homogen und isotrop) -5 mrad. Die Abweichung des Modellierungsergebnisses zu diesem analytischen Wert beträgt an keiner Stelle mehr als 3 % sowohl für den scheinbaren spezifischen Widerstand als auch für die scheinbare Phase.

3 Modellparameter

Unter der Annahme einer horizontalen Wechsellagerung unterschiedlich leitfähiger Schichten wurde ein Modell mit isotropen spezifischen Widerständen erstellt und mit einem entsprechenden Modell mit anisotropen spezifischen Widerständen verglichen. Zur Abschätzung der Stärke der Anomalien wird ein Hintergrundmodell mit homogenem und

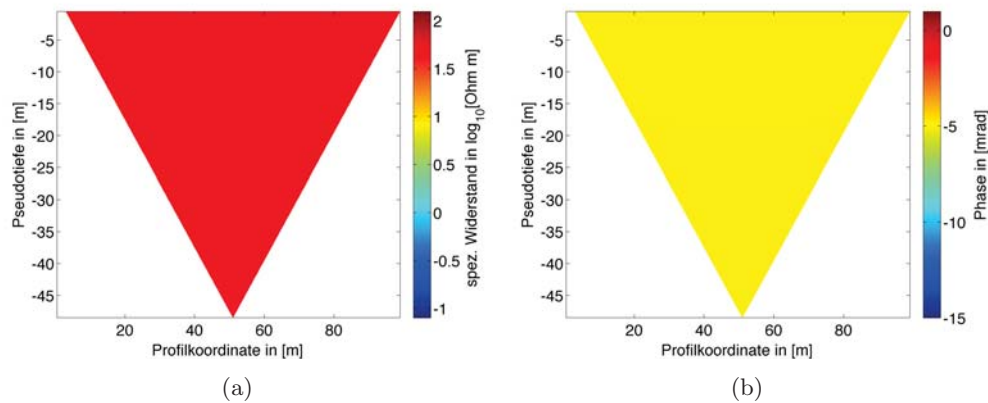


Abbildung 1: Dipol-Dipol-Pseudosektion von spezifischem Widerstand und Phase eines homogenen Halbraums mit den anisotropen spezifischen Widerständen (Amplituden) $100 \Omega\text{m}$ in x- und y-Richtung und $25 \Omega\text{m}$ in z-Richtung und der isotropen Phase -5 mrad .

isotropem spezifischen Widerstand angegeben. Abb. 2 zeigt die Diskretisierung des Modellraumes in 233 Zellen in x-Richtung und 114 Zellen in z-Richtung mit variablem Gitterlinienabstand in den Außenbereichen und konstantem Gitterlinienabstand von $0,5 \text{ m}$ in einem Bereich von 0 m bis 100 m in x-Richtung.

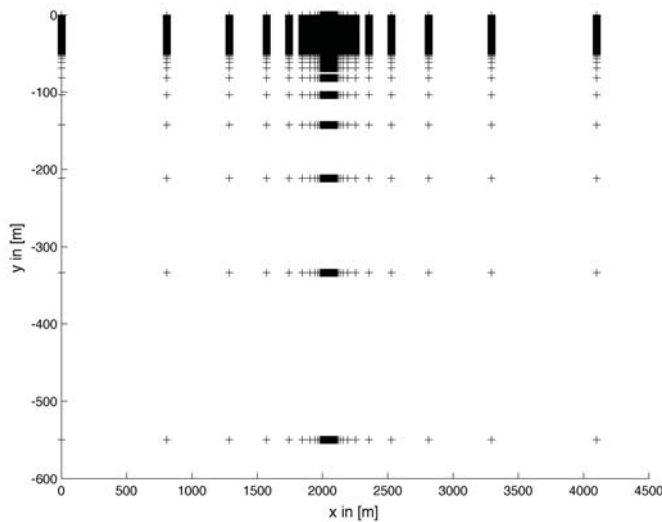


Abbildung 2: Gittermodell des Profils mit 233×114 Gitterzellen und einem Gitterlinienabstand von $0,5 \text{ m}$ im linearen Bereich.

Für das Hintergrundmodell wurde wie im vorherigen Abschnitt ein homogener und

isotroper spezifischer Widerstand mit Amplitude $100 \Omega m$ und Phase -5 mrad gewählt.

3.1 Wechsellagerungen mit unterschiedlichen spezifischen Widerständen

Im ersten Beispiel wird eine Wechsellagerung von unterschiedlich resistiven, isotropen Schichten ($100 \Omega m$ und $1 \Omega m$) in einem Hintergrund mit spezifischem Widerstand von $100 \Omega m$ betrachtet. Die Phase des spezifischen Widerstands beträgt überall -5 mrad . Eine Skizze der Anordnung findet sich in Abb. 3(a). Die Schichtung beginnt 1 m unter

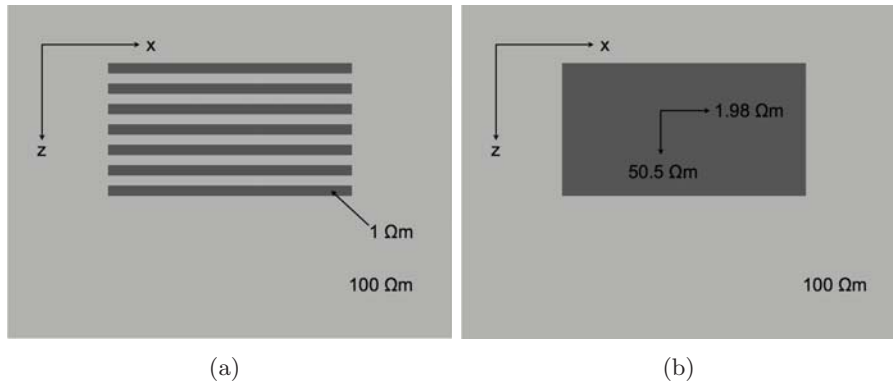


Abbildung 3: Schema des Untergrundmodells: a) vertikale Schichtung mit den spezifischen Widerständen $100 \Omega m$ in hellgrau und $1 \Omega m$ in dunkelgrau. b) Vertikale Schichtung mit spezifischem Widerstand $1,98 \Omega m$ in x- und y- sowie $50,5 \Omega m$ in z-Richtung. Der spezifische Hintergrundwiderstand beträgt $100 \Omega m$ in x-, y- und z-Richtung. Die Phase beträgt überall -5 mrad .

der Oberfläche und reicht bis zu einer Tiefe von $7,5 \text{ m}$. Die Ausdehnung in x-Richtung beträgt 40 m .

Die gleiche Schichtung wird in Abb. 3(b) durch einen Block mit anisotropem spezifischen Widerstand beschrieben. Die anisotropen spezifischen Widerstandswerte berechnen sich nach

$$\begin{aligned} \rho &= R \frac{A}{l} \\ \Leftrightarrow R &= \rho \frac{l}{A} \end{aligned} \quad (6)$$

mit dem Widerstand R eines Blocks der Länge l und der Querschnittsfläche A . Bei Reihenschaltung von zwei Schichten mit den Widerständen R_1 und R_2 mit je der Querschnittsfläche A und der Länge l ergibt sich aus dem Gesamtwiderstand R , der die Querschnittsfläche A und die Länge $2 \cdot l$ hat:

$$\begin{aligned} R_{\text{Serie}} &= R_1 + R_2 \\ \Leftrightarrow \rho_{\text{Serie}} \frac{2 \cdot l}{A} &= \rho_1 \frac{l}{A} + \rho_2 \frac{l}{A} \\ \Leftrightarrow \rho_{\text{Serie}} &= \frac{\rho_1 + \rho_2}{2}. \end{aligned} \quad (7)$$

Bei Parallelschaltung von zwei Schichten gleicher Querschnittsfläche A und gleicher Länge l und den Widerständen R_1 und R_2 ergibt sich der Gesamtwiderstand R , der die Querschnittsfläche $2 \cdot A$ und die Länge l hat:

$$\begin{aligned} \frac{1}{R_{\text{Serie}}} &= \frac{1}{R_1} + \frac{1}{R_2} \\ \Leftrightarrow \frac{2 \cdot A}{l} \frac{1}{\rho_{\text{Serie}}} &= \frac{A}{l} \rho_1 + \frac{A}{l} \rho_2 \\ \Leftrightarrow \rho_{\text{Parallel}} &= 2 \cdot \frac{1}{\frac{1}{\rho_1} + \frac{1}{\rho_2}}. \end{aligned} \quad (8)$$

Mit diesen Beziehungen folgt für die äquivalenten anisotropen spezifischen Widerstände $50,5 \, \Omega\text{m}$ in x- und $1,98 \, \Omega\text{m}$ in z-Richtung. In y-Richtung beträgt der spezifische Widerstand wie in x-Richtung $1,98 \, \Omega\text{m}$. Die Phasen der spezifischen Widerstände betragen homogen und isotrop $-5 \, \text{mrad}$. Die Modellierungsergebnisse sind als Pseudosektionen einer Dipol-Dipol-Anordnung in den Abb. 4(a) bis 5(b) jeweils für Betrag und Phase des spezifischen Widerstands dargestellt.

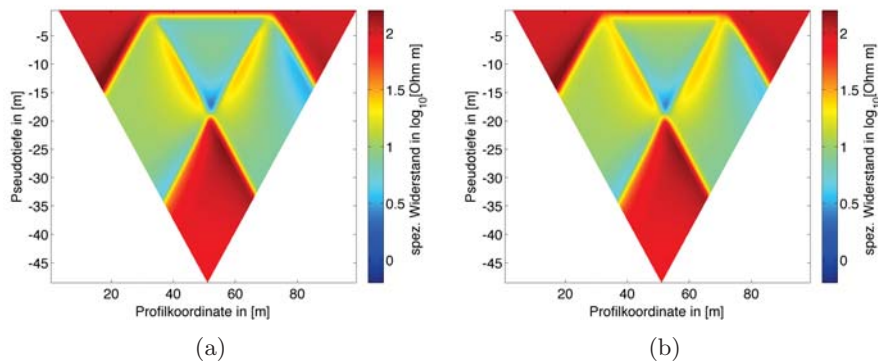


Abbildung 4: Pseudosektion des Betrags des scheinbaren spezifischen Widerstands. a) Vertikale Wechsellagerung isotroper Schichten mit $1 \, \Omega\text{m}$ und $100 \, \Omega\text{m}$. b) Anisotrope Beschreibung der Wechsellagerung durch spez. Widerstände $50,5 \, \Omega\text{m}$ in x- und $1,98 \, \Omega\text{m}$ in y- und z-Richtung. Die Phase beträgt in beiden Modellen überall $-5 \, \text{mrad}$.

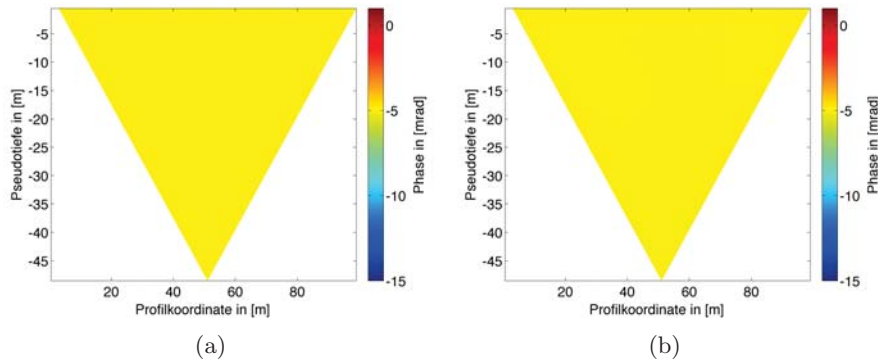


Abbildung 5: Pseudosektion der scheinbaren Phase. a) Vertikale Wechsellagerung isotroper Schichten mit $1 \Omega\text{m}$ und $100 \Omega\text{m}$. b) Anisotrope Beschreibung der Wechsellagerung durch spez. Widerstände $50,5 \Omega\text{m}$ in x- und $1,98 \Omega\text{m}$ in y- und z-Richtung. Die Phase beträgt in beiden Modellen überall -5 mrad .

Beide Modelle - isotropes mit Schichten und anisotropes mit Block - zeigen nahezu gleiche Ergebnisse in den Pseudosektionen. Sehr deutlich ist die durch die Schichtlagerung bzw. durch den Block anisotroper spezifischer Widerstände erzeugte Anomalie sichtbar. Der maximale Wert für den scheinbaren spezifischen Widerstand ρ_a ist etwa $100 \Omega\text{m}$, was dem Hintergrundmodell entspricht. Der minimale Wert für ρ_a beträgt etwa $10^{0,5} \Omega\text{m} = 3 \Omega\text{m}$. Der erwartete minimale scheinbare spezifische Widerstand beträgt im Fall des anisotropen Modells gemäß Gleichung 5 etwa $10 \Omega\text{m}$. Dieser Wert wird im Bereich des Blocks und darunter erreicht und sogar unterschritten. Der bis in $7,5 \text{ m}$ Tiefe ausgedehnte Block führt einen "Schatten" mit, der in Form eines nach unten gerichteten Dreiecks bis in ca. 20 m Tiefe reicht. An beiden Flanken des Dreiecks zeigen schräge "Schatten" nach außen, die ebenfalls einen relativ niedrigen spezifischen Widerstand von etwa $10 \Omega\text{m}$ aufweisen. Die scheinbaren Phasen betragen gleichmäßig -5 mrad , entsprechend den Vorgaben der beiden Modelle.

3.2 Wechsellagerungen mit unterschiedlichen Phasen des spezifischen Widerstandes

In diesem Abschnitt soll die Schichtlagerung mit Schichten unterschiedlicher Phasen betrachtet werden. Dazu wird das Schichtmodell aus Abb. 3(a) für Schichten mit isotropen Phasen und das Blockmodell aus Abb. 3(b) als entsprechendes Modell mit anisotropen Phasen betrachtet. Die spezifischen Widerstände ρ sind in allen Schichten gleich und entsprechend auch im anisotropen Fall in allen Richtungen gleich. Die Reihenschaltung von Elementen mit unterschiedlichen Phasen - also komplexen Impedanzen $X = R(\cos \phi_1 + i \sin \phi_1)$ - ergibt sich mit Realteil $\Re(X)$ und Imaginärteil $\Im(X)$ zu

$$\begin{aligned} X_{\text{Serie}} &= X_1 + X_2 \\ &= \Re(X_1 + X_2) + i\Im(X_1 + X_2) \\ &= (R \cos \phi_1 + R \cos \phi_2) + i(R \sin \phi_1 + R \sin \phi_2). \end{aligned} \quad (9)$$

Für kleine Winkel von ϕ_1 und ϕ_2 kann man die Kosinus-Terme über eine Taylor-Entwicklung in erster Ordnung um 0 mit

$$\cos \phi \approx 1 \quad (10)$$

und die Sinus-Terme mit

$$\sin \phi \approx \phi \quad (11)$$

annähern. Es ergibt sich dann für die Reihenschaltung

$$X_{Serie} \approx 2 \cdot R + iR(\phi_1 + \phi_2) \quad (12)$$

Diese Beziehung lässt sich zur Berechnung des komplexen spezifischen Widerstandes heranziehen. Zwei Elemente mit den Impedanzen X_1 und X_2 und je der Querschnittsfläche A und der Länge l ergeben in Reihenschaltung eine Impedanz X mit der gleichen Querschnittsfläche und der doppelten Länge $2 \cdot l$:

$$\begin{aligned} \rho &= X \frac{A}{l} \\ &= X_{Serie} \frac{A}{2l} \\ &\approx (R + iR \frac{\phi_1 + \phi_2}{2}) \frac{A}{l} \\ &\approx R(\cos \frac{\phi_1 + \phi_2}{2} + i \sin \frac{\phi_1 + \phi_2}{2}) \frac{A}{l}. \end{aligned} \quad (13)$$

Die Reihenschaltung zweier gleich mächtiger Schichten mit gleichem spezifischen Widerstand und unterschiedlicher Phase bedeutet folglich näherungsweise eine Mittelung der Phasen, also

$$\phi_{Serie} \approx \frac{\phi_1 + \phi_2}{2}. \quad (14)$$

Die Parallelschaltung unterschiedlicher Phasen bei gleichem Betrag des spezifischen Widerstands ergibt sich mit den obigen Näherungen und $\phi \ll 1$ zu

$$\begin{aligned} X_{Parallel} &= \frac{1}{\frac{1}{X_1} + \frac{1}{X_2}} \\ &\approx R \frac{1}{\frac{1}{1+i\phi_1} + \frac{1}{1+i\phi_2}} \\ &\approx R(\frac{1}{2} + \frac{i}{2}(\frac{\phi_1 + \phi_2}{2})). \end{aligned} \quad (15)$$

Werden zwei Elemente mit gleichem spezifischen Widerstand und unterschiedlicher Phase und je der Querschnittsfläche A und der Länge l parallel geschaltet, ergibt sich die Impedanz $X_{Parallel}$ mit der doppelten Querschnittsfläche $2 \cdot A$ und der gleichen Länge.

Der komplexe spezifische Widerstand ergibt sich damit zu

$$\begin{aligned}
 \rho &= X \frac{A}{l} \\
 &= X_{Parallel} \frac{2A}{l} \\
 &\approx R \left(\frac{1}{2} + \frac{i}{2} \left(\frac{\phi_1 + \phi_2}{2} \right) \right) \frac{2A}{l} \\
 &= R \left(1 + i \left(\frac{\phi_1 + \phi_2}{2} \right) \right) \frac{A}{l} \\
 &\approx R \left(\cos \frac{\phi_1 + \phi_2}{2} + i \sin \frac{\phi_1 + \phi_2}{2} \right) \frac{A}{l},
 \end{aligned} \tag{16}$$

und für die Phasen wie schon bei der Reihenschaltung näherungsweise:

$$\phi_{Parallel} \approx \frac{\phi_1 + \phi_2}{2}. \tag{17}$$

Dieses Ergebnis macht klar, dass bei konstantem spezifischen Widerstand eine Wechselagerung von Schichten unterschiedlicher Phasen wohl in ihrer Stärke im Mittel, nicht jedoch in ihrer Richtung erkannt werden kann. Nur in Verbindung mit unterschiedlich spezifischen Widerständen der einzelnen Schichten können auch die Phasenwerte der abwechselnden Schichten unterschieden werden.

Ein Schichtmodell mit den Phasen der einzelnen Schichten von -15 mrad bzw. -5 mrad (Hintergrund) hat demnach die anisotropen Phasen -10 Ωm in x-, y- und z-Richtung. Die Ergebnisse sind in den Abb. 6(a) bis 7(b) als scheinbare spezifische Widerstände und scheinbare Phasen in Pseudosektionen dargestellt.

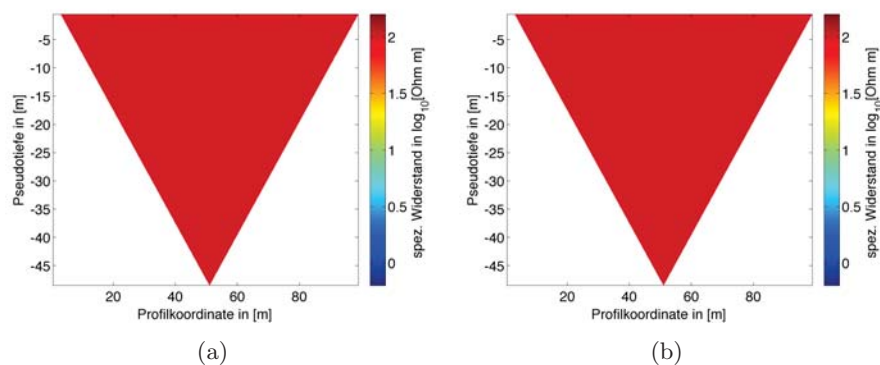


Abbildung 6: Dipol-Dipol-Pseudosektion der Amplitude des scheinbaren spezifischen Widerstands. a) Vertikale Wechsellagerung isotroper Schichten mit -15 mrad und -5 mrad. b) Anisotrope Beschreibung der Wechsellagerung durch Phasen -10 mrad in x-, y- und z-Richtung

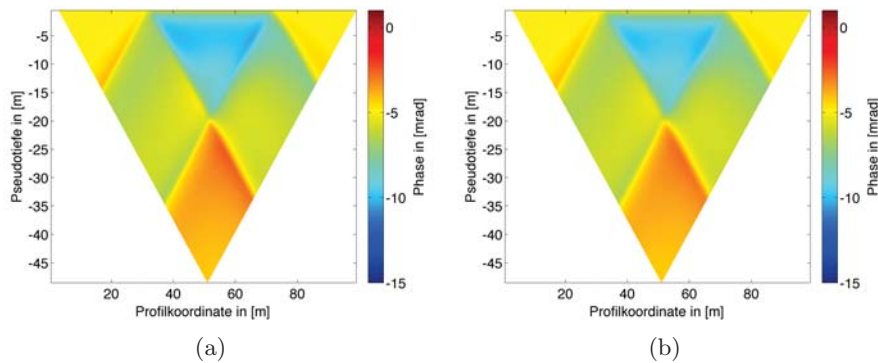


Abbildung 7: Dipol-Dipol-Pseudosektion der scheinbaren Phase. a) Vertikale Wechsellagerung isotroper Schichten mit -15 mrad und -5 mrad. b) Anisotrope Beschreibung der Wechsellagerung durch Phasen -10 mrad in x-, y- und z-Richtung

Wie erwartet zeigt sich in den Pseudosektionen des scheinbaren spezifischen Widerstandes keine Variation mit der eingestellten Phase. Dadurch wird der in Gl. 14 und 17 aufgestellte Zusammenhang bekräftigt. Die Pseudosektionen der Phasen weisen wie die Pseudosektionen im Abschnitt 3.1 starke "Schatten" unterhalb der unteren Störkörperringrenze auf. Auch die seitlichen Flanken sind sichtbar.

4 Schlussfolgerungen

Es ist mit dem auf anisotrope spezifische Widerstände und Phasen erweiterten Programm CRMod nun möglich, Effekte von Anisotropie, z.B. bei Wechsellagerungen von Sedimenten, zu berücksichtigen. Die vorliegenden Ergebnisse zeigen in den möglichen Vergleichen zu analytischen Berechnungen die Gültigkeit des Programms. Es wurde hierbei insbesondere Wert auf die Vorschrift zur Berechnung des spezifischen Widerstands im Fall eines homogenen Halbraums mit anisotropen spezifischen Widerständen (vgl. Gleichung 5) gelegt. Im Zusammenhang mit der Wechsellagerung von Phasen wurde mittels der Gleichungen 14 und 17 gezeigt, dass horizontale und vertikale Wechsellagerungen gleicher Mächtigkeit und gleichen spezifischen Widerstands nicht unterschieden werden können.

Im Zusammenhang mit dieser Arbeit ist eine Messkampagne geplant, die auf stark anisotropem Untergrund stattfinden soll. Die Messdaten sollen mit entsprechenden Modellen mit anisotropen spezifischen Widerständen des vorliegenden Modellierungsprogramms erklärt werden. Zusätzlich zu dieser Anwendung besteht die Möglichkeit, auch das auf dem Modellierungsprogramm CRMod basierende Inversionsprogramm CRTomo ([2]) auf Anisotropie zu erweitern.

5 Danksagungen

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From forward modelling of MT phases over 90° towards 2D anisotropic inversion

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Abstract

Within the framework of the German - South African geo-scientific research initiative Inkaba yeAfrica several magnetotelluric (MT) field experiments were conducted along the Agulhas-Karoo Transect in South Africa. This transect crosses several continental collision zones between the Cape Fold Belt, the Namaqua Natal Mobile Belt and Kaapvaal Craton. Along the profile we can identify areas ($> 10\text{ km}$) with phases over 90°. This phenomenon usually occurs in presence of electrical anisotropy. Due to the dense site spacing we are able to observe this behaviour consistently at several sites.

In this presentation we focus on the profile section between Prince Albert and Mosselbay. With isotropic 2D inversion we are able to explain most features in the MT data but not the abnormal phase behavior. With several anisotropic forward modelling studies we have tested the influence of anisotropy parameters on the MT responses. In a first step we use simple 2D models with embedded zones of electrical anisotropy to get a basic understanding of anisotropic responses. In a second step isotropic 2D inversion results serve as background models in which we included anisotropic zones, e.g. to fit the abnormal phase curves. These resolution tests are necessary and important for the future development of a 2D inversion with spatially constraint anisotropy.

1 Introduction

For a 2D geoelectric model with a finite system of homogeneous, but generally anisotropic blocks the electrical conductivity is a tensor instead of a scalar quantity. It is symmetric and positive-definite. Due to its symmetry the conductivity tensor $\hat{\sigma}$ within each layer of model can be diagonalized and expressed by three principal conductivities σ_1 , σ_2 , σ_3 and a rotation matrix, which can be decomposed into three elementary Euler's rotations α_S , α_D , α_L respectively.

$$\begin{aligned}\hat{\sigma} &= \begin{pmatrix} \sigma_{xx} & \sigma_{xy} & \sigma_{xz} \\ \sigma_{yx} & \sigma_{yy} & \sigma_{yz} \\ \sigma_{zx} & \sigma_{zy} & \sigma_{zz} \end{pmatrix} \\ &= R_z(-\alpha_S)R_x(-\alpha_D)R_z(-\alpha_L) \begin{pmatrix} \sigma_1 & 0 & 0 \\ 0 & \sigma_2 & 0 \\ 0 & 0 & \sigma_3 \end{pmatrix} R_z(\alpha_L)R_x(\alpha_D)R_z(\alpha_S)\end{aligned}$$

where R_x and R_z are elementary rotation matrices around the coordinate axis x and z , respectively. The angles α_S , α_D , α_L are typically called anisotropy strike, dip and slant, respectively.

In this work we investigate the feasibility of use of 2D forward anisotropy modelling to simulate phases over 90° which we observed in field data from the MT survey in South Africa. Using the 2D anisotropy forward modelling algorithm of Pek and Verner (1997) we vary the model parameters in

order to gain a basic understanding of the MT transfer function in presence of anisotropy. First we will give a brief introduction of the survey area and then we will show the results of 2D isotropic inversion and 2D anisotropic forward modelling.

2 Survey area

Within the framework of the German - South African geo-scientific research initiative Inkaba yeAfrica four magnetotelluric (MT) field experiments were conducted along the Agulhas-Karoo Transect in South Africa (Weckmann et al., 2007, Stankiewicz et al., 2008). These transects cross several continental collision zones and their respective units, such as the Cape Fold Belt, the Namaqua Natal Mobile Belt and the Kaapvaal Craton. The MT profile on which we focus in this paper is located in

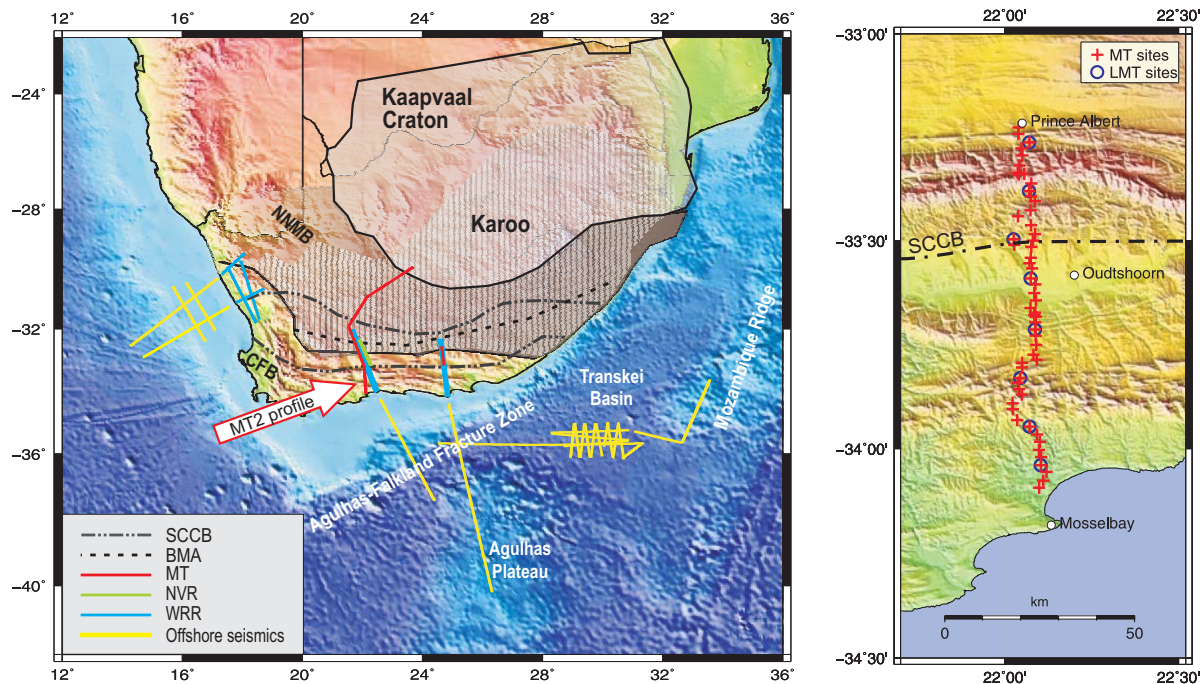


Figure 1: Left: Map of the Agulhas-Karoo Geoscience Transect with MT Profiles (red lines) across Cape Fold Belt, the Namaqua Natal Mobile Belt and the Kaapvaal Craton. Right: The profile MT2 contains 46 broad band MT sites and 8 broad band / long period MT sites. It extends 120 km from Prince Albert in the North to Mosselbay in the South and covers the entire Cape Fold Belt.

the Cape Fold Belt. In total, 54 MT sites were deployed along this 120 km long profile MT 2 (Fig. 1). It extends from Prince Albert in the North to Mosselbay in the South and covers the entire Cape Fold Belt (CFB), its inliers, the Oudtshoorn and the Kaaimans Basins, the Swartberg and the Outeniqua Mountain ranges and several major thrusts and faults.

3 Data and isotropic 2D inversion

Along the profile MT2 we acquired 5-component MT data at all stations in a period range from 0.001s to 1000s using GPS synchronized S.P.A.M. MkIII (Ritter et al., 1998) and CASTLE broadband instrument. Metronix MFS06/06 induction coil magnetometer and non-polarizable Ag/AgCl telluric electrodes were used to record natural magnetic and electric field variations. The data were processed with the EMERALD software package (Ritter et al., 1998) using both robust single site and remote reference techniques. At some sites, for which a suitable reference site was not available and the data were affected by cultural noise (extensive farming), we applied the frequency domain selection scheme after Weckmann et al. (2005) to improve data quality.

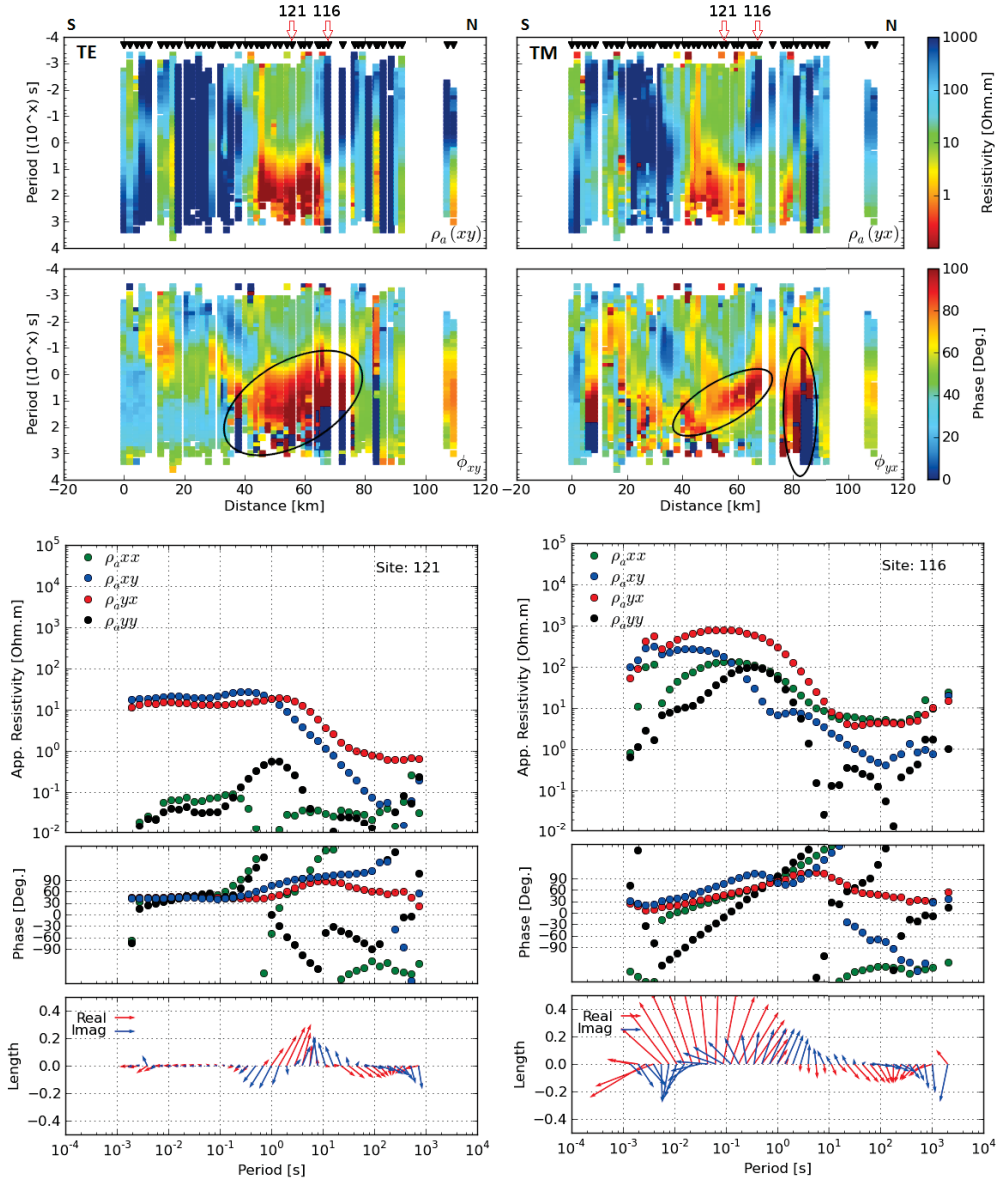


Figure 2: Pseudo-sections of TE and TM mode apparent resistivity and phase (x direction pointing west). We can observe phase values over 90 degrees in the middle of the profile in both TE and TM component (marked with black ellipses in the upper panel). Two exemplary sites (site 121 and site 116) are displayed as apparent resistivity, phase and induction arrows (Wiese convention) over period. The red arrows in the pseudo-sections show the location of the selected sites. We can identify that the phases, (especially in the TE component) leave the first quadrant at a minimum period of 10 s, but typically at longer periods of 100-1000 s.

After data processing, we applied different strike and dimensionality analyses, e.g. the ellipticity analysis after Becken & Burkhardt (2004), obtaining an electromagnetic strike direction of -90° , i.e. East-West direction. This is in general compatible with the tectonic grain of the CFB. Furthermore, most of the data seem to be compatible with a 2D interpretation approach. However, in a limited area we observe impedance phases exceeding 90 degrees at long periods. They typically appear in both, TE and TM mode, and persist in either component even if we rotate the coordinate system (Fig. 2).

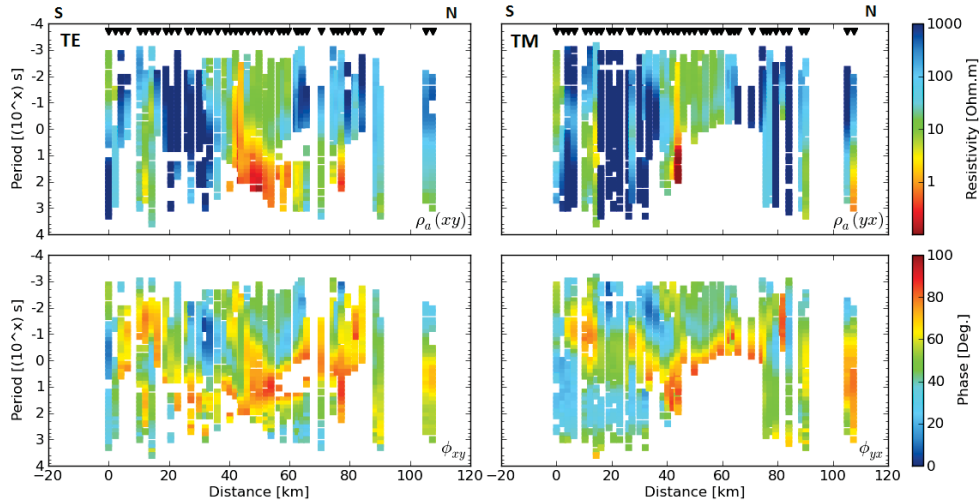


Figure 3: TE and TM mode data used for isotropic 2D inversion. Prior to inversion they were rotated into a common coordinate system according to the geoelectric strike direction of $\approx 90^\circ$ (obtained by strike and phase tensor analysis). Data with phases over 90° are discarded from the data base.

For an isotropic 2D inversion approach we use RLM2D (Rodi & Mackie, 2001, implemented in the WinGlink software package). Before starting a 2D inversion we have to make sure that large phase values over 90° are excluded as they cannot be fitted by a 2D inversion approach. Figure 3 shows the data which were finally used for 2D inversion. Figure 4 displays our preferred 2D conductivity image of the upper 35 km together with a section of the geological map after Hälbig et al. (1993). The inversion was started from a homogeneous half-space of $100\Omega m$ with $\tau = 10$ using TE and TM component, with intermediate inclusion of the vertical magnetic transfer function. Within the framework of this work, we refrain from interpreting conductivity anomalies in a geological context, but focus on the area where phases over 90° occur. One of the most prominent conductivity anomalies is a triangular shaped, highly conductive structure in the middle of the profile. In this area we also observe phases leaving the quadrant. This conductivity anomaly is required by the data which is compatible with a 2D interpretation approach; however, we believe that the phases over 90° are caused by some electrically anisotropic structures in this area (in Fig. 4 between black dashed lines). We should also note that the deep part of this area only possesses a very limited resolution (in Fig. 4 marked with semitransparent mask) because most of the data which relates to this part are excluded in order to satisfy the 2D isotropic procedure (see Fig. 3). Similar phase behaviour was explained with crustal anisotropy by Weckmann et al. (2003) and Heise & Pous (2003), where data within old continental collision zones in Namibia and on the Iberian Peninsula, respectively. In both cases the data could be fit by using a shallow and a deeper electrically anisotropic zone with different anisotropy strike.

4 Anisotropic forward modelling

Isotropic 2D inversion is adequate to explain the data in most parts along the profile but it is very unsatisfactory not being able to include phases greater than 90 degrees and thus neglect a substantial amount of data. In order to develop a 2D inversion with spatially constraint anisotropy, resolution tests and synthetic modelling studies are necessary. In a first step towards the constraint anisotropic

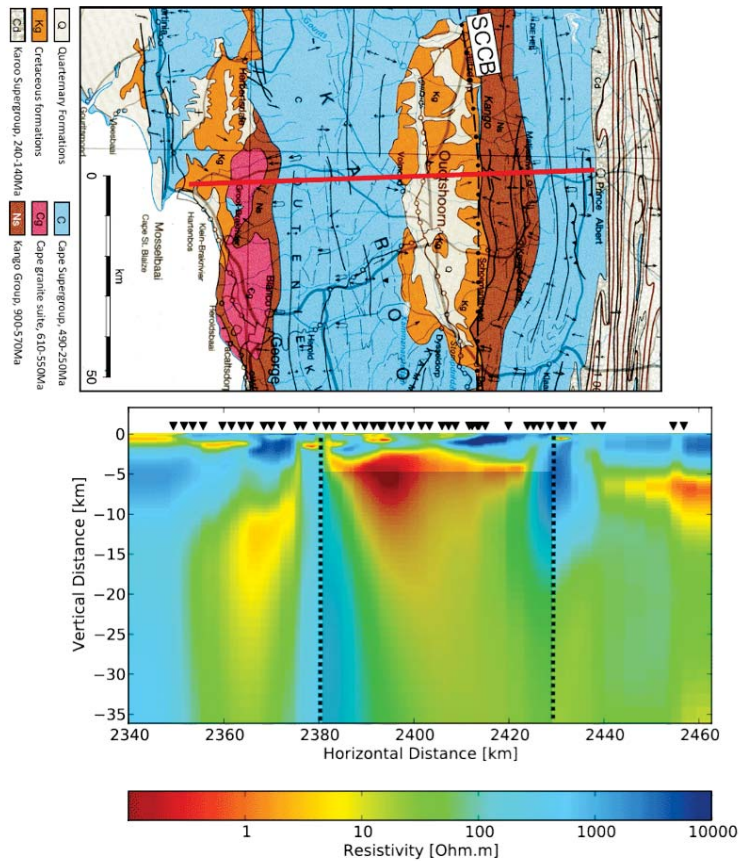


Figure 4: 2D inversion model together with a section from the geological map after Hälbig et al. (1993). The area where we observe phases over 90° is located where the inversion puts a triangular shaped high conductivity anomaly (area between black dashed lines). Zones of limited resolution are also in this area and marked with a semitransparent mask.

inversion we included zones with electrical anisotropy in a 2D isotropic model and calculated its forward responses which aim to understand the anisotropy effect.

4.1 Lateral extension

Pek & Verner (1997) have suggested that a combination of two azimuthal anisotropies with anisotropy strikes perpendicular to each other could produce phases that exceed 90° . Based on the suggestion we study at first a combination of two different anisotropic blocks. The initial model (Fig. 5, left) consists of a 300m isotropic surface layer with a resistivity of $30\Omega m$ and an anisotropic block starting at a depth of 300m embedded in a medium of $100\Omega m$. The principal resistivities of the block are $\sigma_1/\sigma_2/\sigma_3 = 50/0.5/50\Omega m$ and the anisotropy strike α_S is 120° . The block is underlain by an isotropic layer with a resistivity of $15\Omega m$. Beneath the isotropic layer a second anisotropic block with $\sigma_1/\sigma_2/\sigma_3 = 30/0.3/30\Omega m$ and $\alpha_S = 30^\circ$ (perpendicular to α_S of the first block) is embedded in an isotropic half-space with $100\Omega m$. The second, deeper anisotropic block has a lateral extension of 15km in the first (Fig. 5, left upper panel), and 80km in the second (Fig. 5, left lower panel) model, respectively.

The forward responses are displayed in figure 5 (right) as apparent resistivities and phases in xy and yx component, respectively. Comparing the responses of both models we see that the phases of yx component (Φ_{yx}) for the second model (Fig. 5, left lower panel) at sites above the first anisotropy block leave the quadrant at a period of $\approx 100s$ (Fig. 5, right lower panel), while they are smaller than 90° (Fig. 5, right upper panel) for the first model (Fig. 5, left upper panel) with the narrower deep anisotropic block.

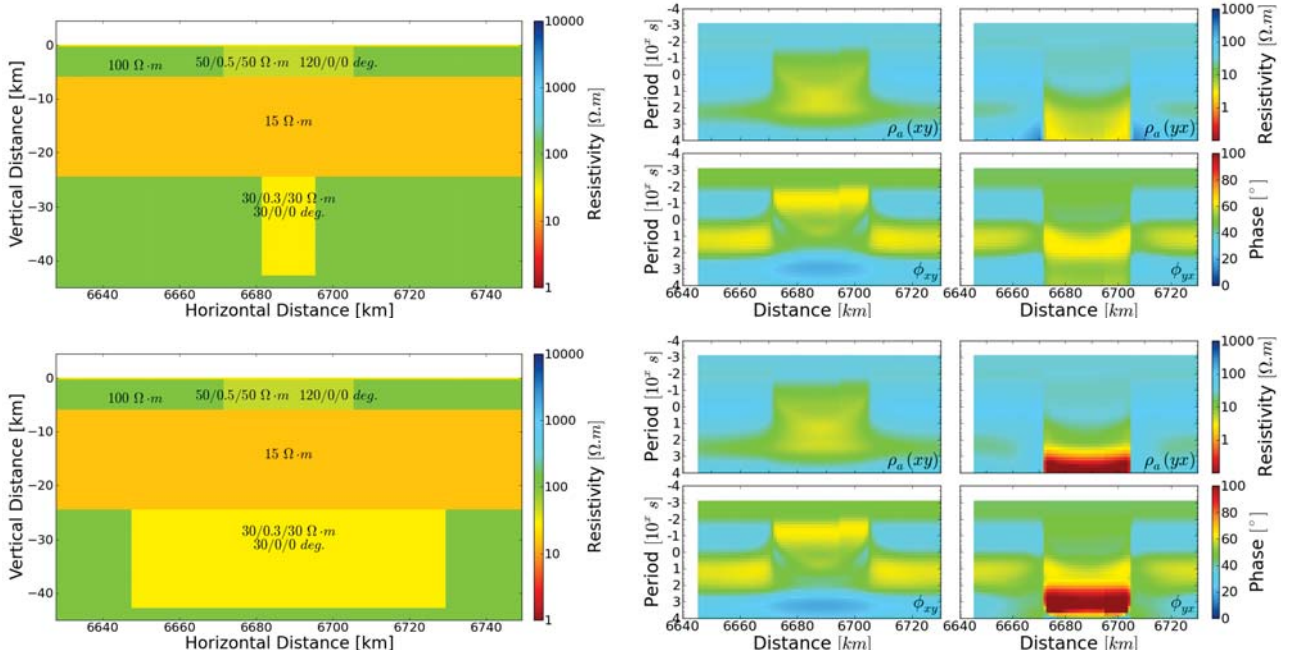


Figure 5: Models (left) and their forward responses (right). The models differ only in lateral extension of anisotropy block. The lateral extension of the second block is about 15 km in the first model (left upper panel) and 80 km (left lower panel) in the second model. The major difference in the responses appears in the yx component. Phases over 90° occur if the lateral extension of the deeper anisotropy block is greater than it of the shallower anisotropy block.

4.2 Depth

In a second step, we used the same anisotropy parameters as described above. The initial model consists of a 300 m isotropic surface layer and a isotropic half-space with resistivity of $100 \Omega m$. Two anisotropy blocks are embedded in the half-space. The first block started at depth of 300 m and shaped as a trapezoid. In this attempt we vary the depth of the second block. In the first model (Fig. 6, left upper panel) the second block started at a depth of 1.8 km and in the second model (Fig. 6, left lower panel) started at a depth of 6.8 km.

In the forward responses (Fig. 6, right) we see that the phases Φ_{yx} for the second model become larger than 90° at periods $> 10 s$ at sites above the first block, while they for the first model are all smaller than 90° . In this attempt the second anisotropy block has always sufficiently greater lateral extension than the first anisotropy block, but we only see the phase anomalies by model with anisotropy block in adequate depth. Besides the lateral extension of the anisotropy block, also the depth is one of those key conditions under which phase anomalies appear.

4.3 Rotation angle

In a third test we use a similar model as described above for step two. The model contains a 300 m isotropic surface layer and an isotropic half-space with resistivity of $100 \Omega m$. Two anisotropic blocks are embedded in the half-space. The first block starts directly beneath the surface layer and the second block in a depth of 6 km. They have the principal resistivities $\sigma_1/\sigma_2/\sigma_3 = 50/0.5/50 \Omega m$ and $\sigma_1/\sigma_2/\sigma_3 = 30/0.3/30 \Omega m$, respectively. In this attempt we vary the anisotropy strike angle α_{S1} and α_{S2} for both blocks.

The forward response for the model with $\alpha_{S1} = \alpha_{S2} = 0^\circ$ is displayed in the right upper panel of figure 7. The phases of the yx component approach 90° for long periods. For $\alpha_{S1} = 60^\circ$ and $\alpha_{S2} = 120^\circ$ (Fig. 7, left lower panel) the yx component phases become larger than 90° for periods $> 10 s$, while they stay below 90° for $\alpha_{S1} = 90^\circ$ and $\alpha_{S2} = 120^\circ$ (Fig. 7, right lower panel). Comparing the three examples, we see that model with different combination of strike angle in both shallow and

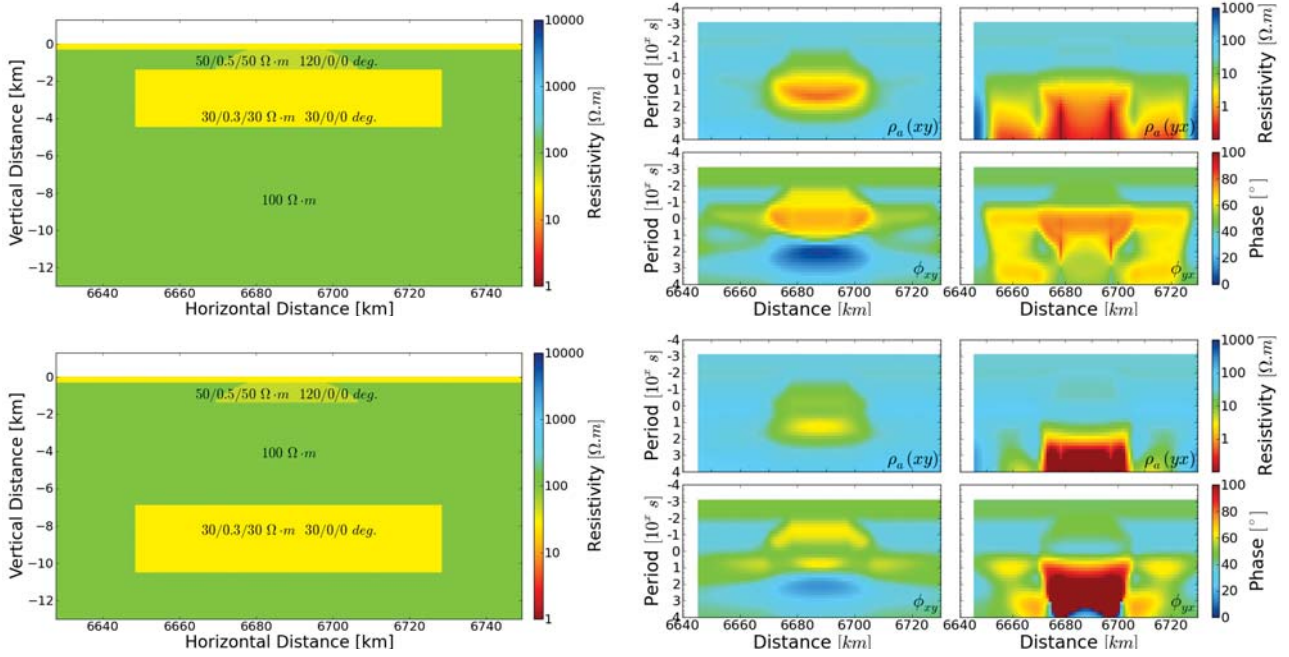


Figure 6: Models (left) and their forward responses (right). The models differ by depth of the second anisotropy block. In the first model (left upper panel) the second block starts in a depth of 1.8 km and in the second model (left lower panel) the block starts in a depth of 6.8 km . The phases of yx component for the model with the deeper anisotropic block become larger than 90° at periods $> 10\text{ s}$ at sites above the first block. For the model with the shallower block all phases are smaller than 90° .

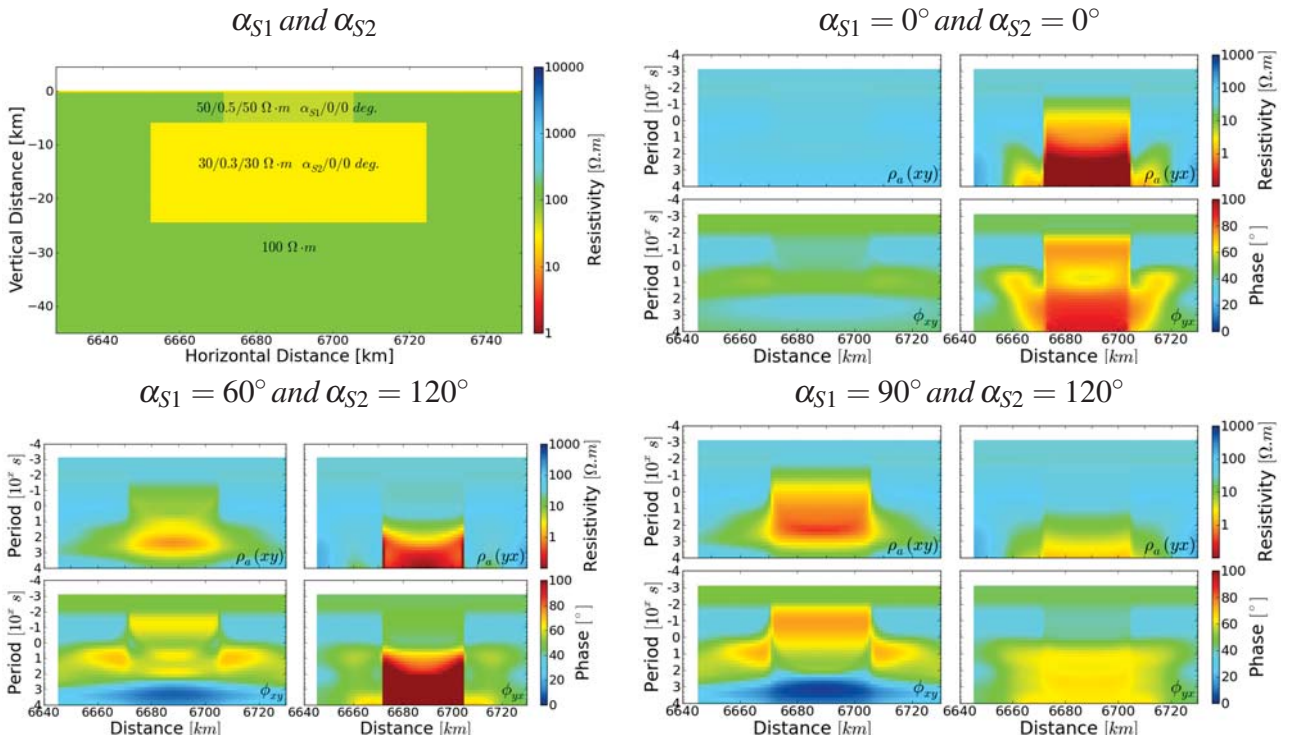


Figure 7: Model (left upper panel) and its forward responses with different anisotropy strike angles. The model contains the same surface layer, background medium and resistivities for both anisotropic blocks as the models in fig. 6. We vary the strike angles α_{S1} and α_{S2} for the shallow and the deep block (left upper panel). The forward responses are displayed as apparent resistivities and phases in xy and yx component.

deep blocks produce forward responses with pronounced difference. This can be observed not only in yx component ($\rho_a(yx)$ and Φ_{yx}) but also in xy component ($\rho_a(xy)$ and Φ_{xy}). A possible explanation is that a change of strike angle forces the current flow to change its preferred direction from the shallow to the deep block. According to this fact we may conclude that the anisotropy strike angle influences forward responses significantly and phase greater than 90° will appear if the strike angle of both blocks differ by an adequate amount.

5 Conclusions

Phases greater than 90° can only be modelled by a combination of a deep and a shallow anisotropic block. We varied several model parameters to study the changes in forward responses. For our models we can conclude that phases out of the first quadrant occur when: (i) the anisotropy ratio ($\sigma_{max}/\sigma_{min}$) is high (for both blocks); (ii) the deep anisotropic block has a much larger lateral extension than the shallow block; (iii) the deep anisotropic block is located in greater depth; (iv) the angles of anisotropy of both blocks differ by a considerable amount. In our models the difference should be at least 45° so that the preferred direction of current flow changes significantly from the shallow to the deep block.

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Substitute models for static shift in 2D

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Introduction

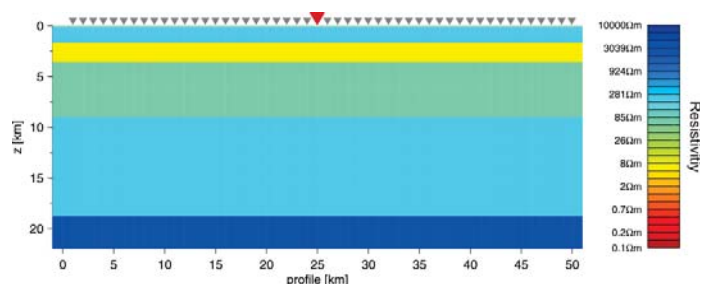
MT practitioners often down-weight apparent resistivity TE mode data prior to 2D inversion to avoid problems with static shift, obviously assuming that static shift of the TM mode is handled automatically by the inversion. Static shift is caused by conservation of charges at local conductivity discontinuities which are small with respect to the inductive scale length.

Here we present a class of shallow conductivity anomalies which can produce significant up- or downwards shifted TM mode apparent resistivity curves (static shift in the TE mode cannot be simulated with 2D modeling). We examine how this static shift is reproduced by 2D inversion and show that the results are strongly influenced by grid design and regularization. We conclude that modern 2D inversion packages are not optimized to handle static shift. Our results also indicate that it is no good reason to assume that static shift cancels out on average.

2D substitute structures for static shift

As 2D models only allow for conductivity changes in one of the horizontal directions, static shift can only be accounted for in one component (TM) and re-produced by forward modeling.

Fig. 1 (right): Starting from a 1D layered half-space (LH) model (Fig. 1a) 2D inhomogeneities were added to the resistivity model just below/at the surface (Figs 1b-g, left panel). TE and TM model responses were calculated for frequencies between 10^3 Hz and 10^{-3} Hz directly above the center of the inhomogeneities. The 2D forward computations are compared to the 1D LH results indicated by crosses (Figs 1b-g, right panel).



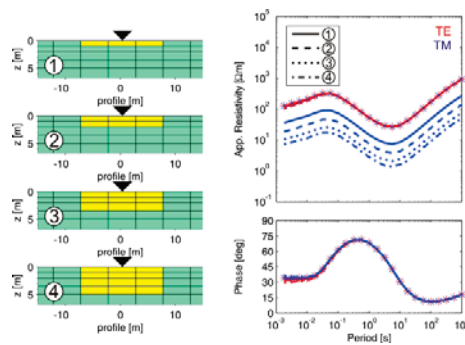


Fig. 1b (left): A 15 m wide 10 Ωm conductor with a thickness varying from 1 m to 5 m causes downward shift of the TM resistivity curves. The shift increases in dependence of the block thickness, i.e. the extent of the horizontal resistivity contrast. The amount of static shift which can be generated by extending the conductive inset downwards is limited. At some point the structure is becoming inductively effective.

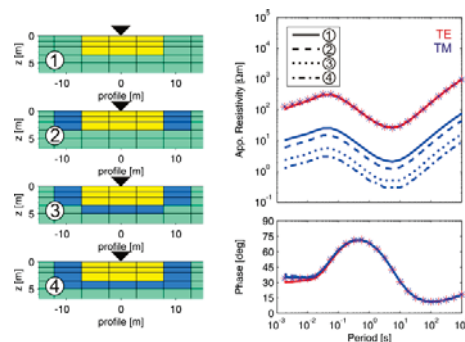


Fig. 1d (left): Enhancing the conductivity contrast by padding the conductor with resistive ($10^3 \Omega\text{m}$) cells TM downward shift is increased. Structures of this type can be observed in inversion models below static shift affected sites when static shift is not taken into account during inversion (cf. pane view in Fig. 2c (2)).

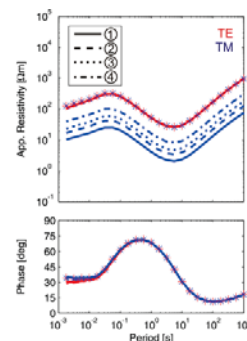
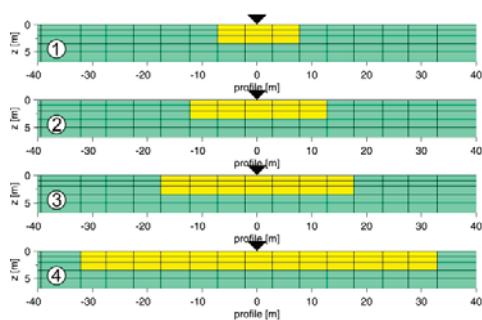


Fig. 1c (left): For a fixed thickness of 3 m the lateral dimensions of the 10 Ωm -block were altered between 15 m and 65 m. With increasing block width, the static shift effect becomes smaller as the lateral resistivity contrast which is responsible for the distortion of the electrical field moves away from the site location.

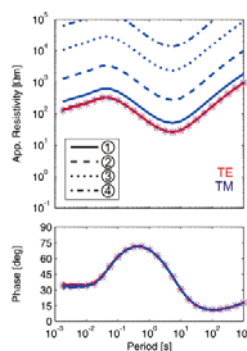
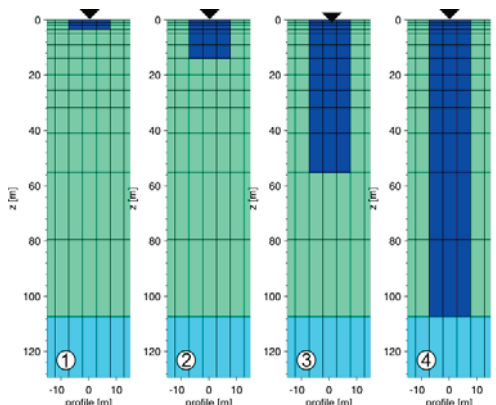


Fig. 1e (left): Placing a highly resistive block ($10^4 \Omega\text{m}$) with a width of 15 m and varying thicknesses between 5 m and 110 m beneath the site results in upward shift of the TM mode. Static shift increases with increasing vertical block dimensions.

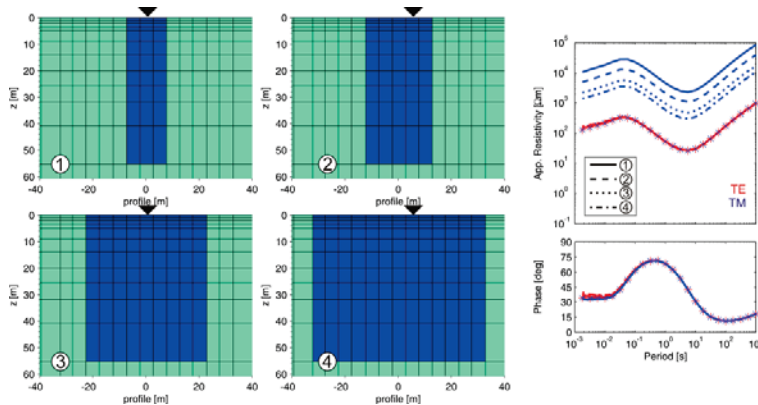


Fig. 1f (left): For a fixed thickness of 55m the lateral extent of the $10^4 \Omega\text{m}$ -block was altered between 15 m and 65 m. As seen for the $10 \Omega\text{m}$ -block, the static shift effect becomes smaller with increasing block width as the resistivity contrast distorting the electrical field moves away from the site location.

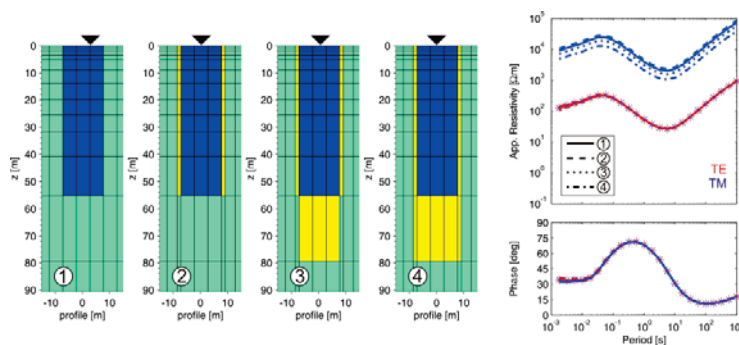


Fig. 1g (left): Padding the resistive block with a thin column of conductive ($10 \Omega\text{m}$) cells increases the upward shift of the TM resistivity curve slightly. The horizontal padding cells must be very small horizontally to prevent inductive effects.

As expected, the TE resistivity curves and the phase curves of both, TM and TE, modes are mostly unaffected by these very small scale structures.

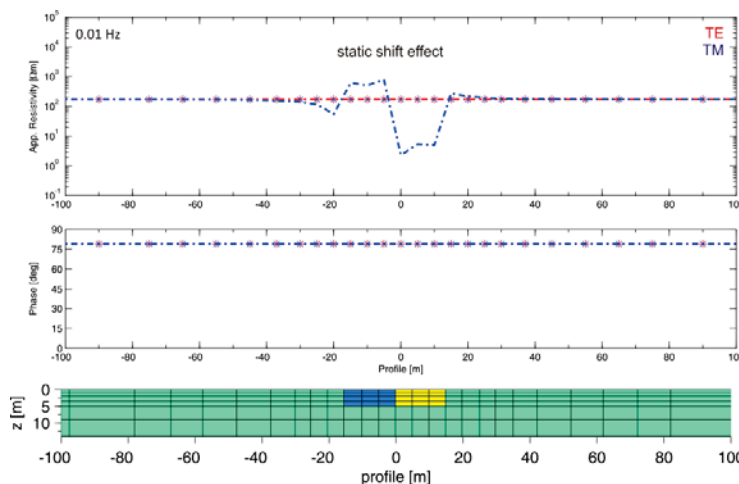


Fig. 1h (left): Distortion of the TM apparent resistivities is due to static shift as the skin depth exceeds 15 km for $\rho=10 \Omega\text{m}$ and $f=0.01 \text{ Hz}$. The downward shifting effect above the good conductor is significantly stronger than the upward shift above the resistive block, although resistivity contrasts of the two blocks to the background are the same. Static shift values along this profile do not sum up to zero. The figures above indicate that structures causing upwards shift have to have a larger vertical extent to produce the same amount of shift.

So, it is at least questionable if a zero sum/average assumption for static shift values, which is often applied in inversion schemes, is appropriate.

2D Inversion

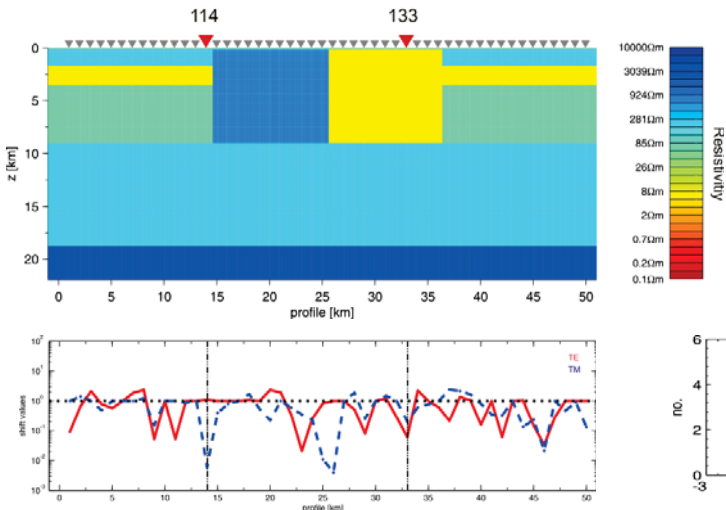


Fig. 2a (left): A 2D resistivity model was created by adding a low and a high resistive block to the layered half-space model (cf. Fig. 1a). Data were calculated for 50 sites along a 50km long profile with 1 km site spacing adding 2.5% Gaussian noise.

Fig. 2b (above): *Right:* Synthetic shift values c were generated obeying a modified logarithmic Gamma-distribution: $\log_{10}(c) = -\Gamma(0.5, 2) + 0.5$ and $\rho_{\text{shift}} = c\rho$. The maximum of the modified distribution is at 10^0 , the expected value is $10^{-0.5}$. *Left:* Shift was applied randomly to 2/3 of the sites, independently for TE and TM. The left panel shows the applied shift values for each site along the profile.

Fig. 2c (below): Inversion results for un-shifted (1) and shifted (2)-(4) data for 100 iterations starting from a 100 Ω m homogeneous half space, applying a uniform smoothing with $\tau=10$. Error floors were 2% (0.6°) for the phases and 5% and 500% for TM and TE apparent resistivities, respectively.

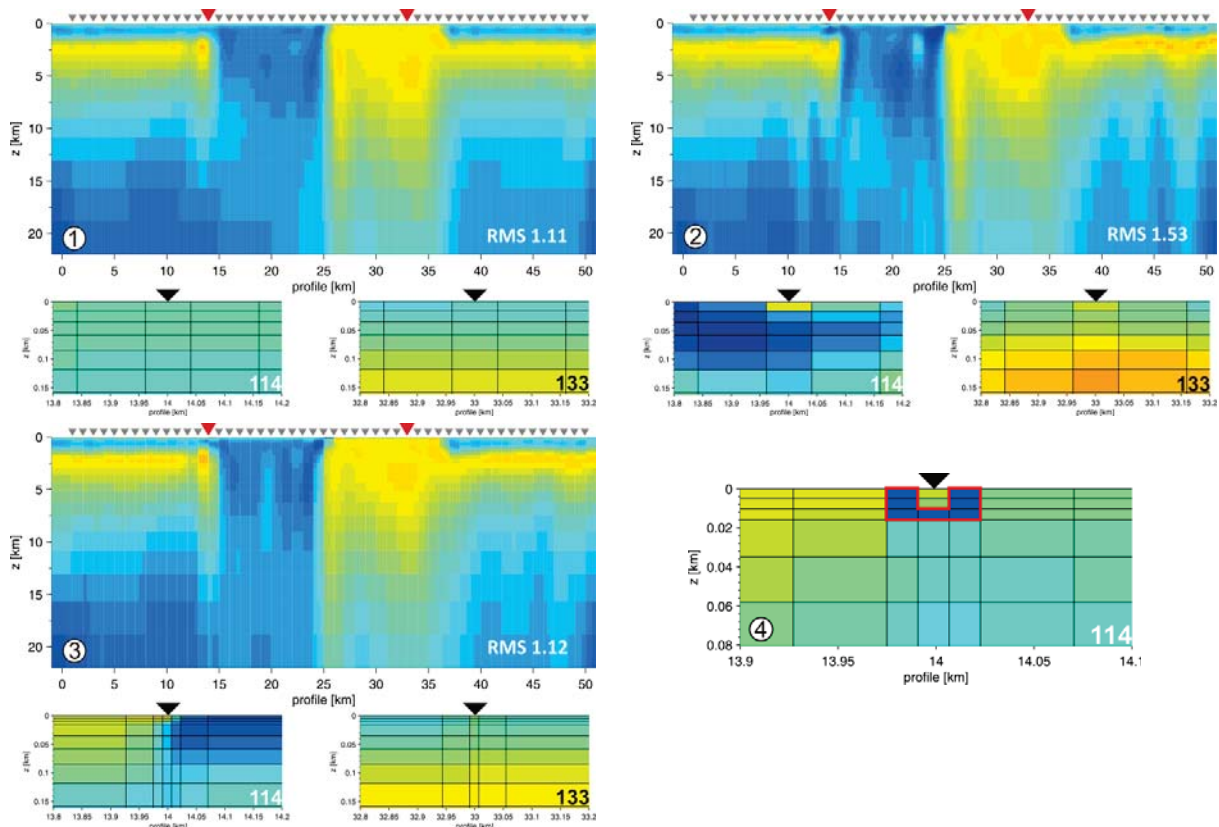
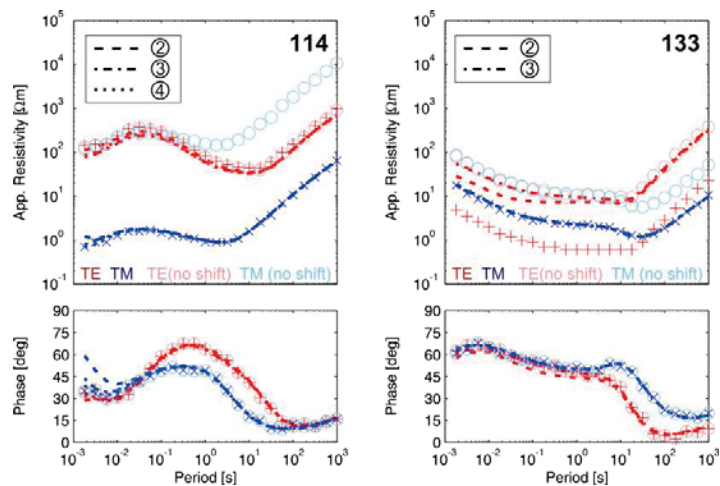


Fig. 2c (above, continued): The model grid was created setting the column widths to 0.5 times of the minimum skin depth $p_{\min}$ for (1) and (2), and was refined to $0.1 \cdot p_{\min}$ for (3) and (4) (cf. pane views).

(2) & (3): As smoothing (regularization) is working against sharp resistivity contrasts, the inversion introduces complex and wide-stretched structures below the sites affected by static shift. The refined grid in (3) reduces smoothing lengths for uniform regularization. As a consequence, introduced compensatory structures appear simpler or diminish.

(4) Inversion result after introducing a so-called “tear zone” (feature of WinGLink) at site 114 (red outline). Tear zone inversion does not penalize sharp conductivity contrasts at the outline of the tear zone. Now, the inversion can properly model static shift. The model can account for TM static shift within this zone while obtaining surrounding resistivity values close to the original ones.

Fig. 2d (right) shows un-shifted data, shifted data as symbols and the inversion results (cf. Fig. 2c) for sites 114 and 133 as lines. For periods longer than 10^{-2} s shifted data are fitted well by all inversions. For shorter periods, the inversion struggles to fit the data, especially at site 114 where TM resistivities were shifted downwards by two decades. TM phase responses below 10^{-2} s deviating from forward data clearly show the inductive effects of the near surface structures. Static shift effects decrease with the refined grid (3) and introduction of a tear zone (4).



Summary

Static shift for the TM mode can easily be produced by a range of simple structures at surface being introduced to the regional resistivity models as substitutes for natural structures causing static shift in field data.

To account for TM static shift within 2D inversion, model grids have to be very fine in the vicinity of sites with column widths and thicknesses much smaller than the induction volume of the highest frequency to be analyzed.

Smoothing routinely applied in minimum structure inversions does not allow for the sharp conductivity contrasts required to produce sensible static shift. It would therefore be desirable to be able to apply different smoothing factors to a static shift compensating top layer and the remaining model.

Die Übergangsimpedanz einer kapazitiv angekoppelten Elektrode

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Vorwort von A. Hördt

Die in diesem Artikel vorgestellte Theorie beruht auf einer unveröffentlichten Arbeit von Prof. Peter Weidelt, der am 1.7.2009 unerwartet verstorben ist. Die Arbeit ist ein typisches Beispiel für seine außerordentliche Hilfsbereitschaft, für die er unter Kollegen in aller Welt bekannt war. Im Jahr 2007 hatte ich ihn um Unterstützung bei der Berechnung der Übergangsimpedanz einer kapazitiven Elektrode gebeten. Einige Wochen später hatte er die vollständige Lösung hergeleitet und in ein Programm umgesetzt. Er selbst hat die Aufgabe als nicht allzu schwierig oder wichtig empfunden, und hätte die Ergebnisse vermutlich nie veröffentlicht.

1 Zusammenfassung

Es wird eine Möglichkeit vorgestellt, die Impedanz einer Kreisscheibe zu berechnen, die an einen Halbraum mit endlicher Leitfähigkeit angekoppelt ist. Basierend auf den Maxwellgleichungen wird eine Integralgleichung für die Ladungsdichte q auf der Kreisscheibe aufgestellt. Mit der Randbedingung, dass das Potential auf der Scheibe konstant ist, kann die Ladungsdichte q der zu angehobenen Platte stetig ausgerechnet und damit die Gesamtladung bestimmt werden. Es zeigt sich, dass der Übergang von aufliegender Funktion des Abstandes verläuft. Zudem nimmt die Impedanz als Funktion des Abstandes zu, d.h. die Ankopplung einer kapazitiv angekoppelten Elektrode kann niemals geringer sein, als die einer aufliegenden mit gleicher Fläche. Je nach Modellparametern kann die Impedanz jedoch innerhalb weniger Nanometer über Größenordnungen variieren, so dass es dennoch günstig sein kann, die Elektrode zu isolieren, um starke Schwankungen der Impedanz zu vermeiden.

2 Einleitung

Bei geoelektrischen Messungen über sehr schlecht leitendem Untergrund kann es von Vorteil sein, den Strom kapazitiv anzukoppeln (Kuras et al., 2006). Dabei wird ein hochfrequentes Wechselfeld an eine Elektrode angelegt, die keinen direkten Kontakt mit dem Untergrund hat. Eine genaue Berechnung der Übergangsimpedanz einer solchen Elektrode ist notwendig, um zu beurteilen, unter welchen Bedingungen eine kapazitive Ankopplung einer galvanischen überlegen ist. Für die praktische Umsetzung ist die Frage von Bedeutung, ob es günstig ist, eine kapazitive Elektrode zu isolieren und einen galvanischen Kontakt auszuschließen. Die Standardformeln für die Ankopplung kapazitiver Elektroden sind allerdings nur für hohe Leitfähigkeiten des Untergrundes gültig. Hördt (2007) hat analytische Gleichungen für eine Kugelelektrode im sphärisch geschichteten Vollraum hergeleitet und diskutiert. Hier wird nun eine Lösung für eine zylindrische Scheibe über einem Halbraum hergeleitet.

3 Modellbeschreibung

Abbildung 1 illustriert das zu lösende Problem. Zu berechnen ist, welcher Strom I bei einer vorgegebenen Spannung U zwischen zwei zylindrischen Scheiben über einem homogenen Halbraum fließt.

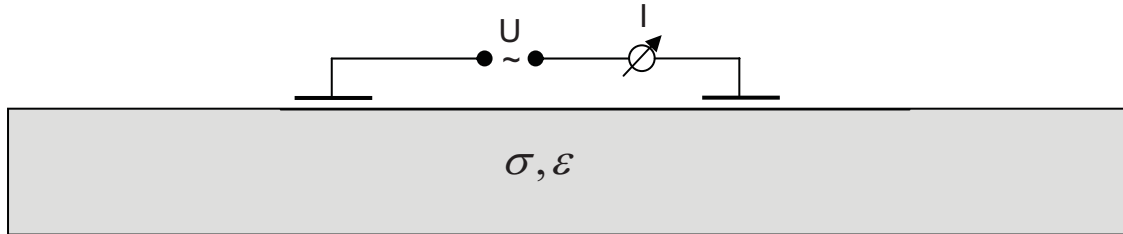


Abbildung 1: Skizze der kapazitiv angekoppelten Stromquelle: Zwei zylindrische Scheiben sind kapazitiv an einen homogenen Halbraum mit Leitfähigkeit σ und Permittivität ϵ angekoppelt, es fließt ein Strom I bei vorgegebener Spannung U .

Wenn der Untergrund ein idealer Leiter ist, gilt bei kleinem Abstand zwischen Elektrode und Untergrund für die komplexe Übergangsimpedanz (Smythe, 1968):

$$Z = \frac{1}{i\omega C} \quad (1)$$

wobei ω die Kreisfrequenz ist und C die Kapazität, mit

$$C = \frac{A\epsilon_0}{d} \quad (2)$$

Dabei ist A die Fläche der Elektrode, d der Abstand zum Untergrund, und ϵ_0 die Dielektrizitätszahl des Vakuums zwischen den Elektroden. Diese Formeln stellen allerdings eine Näherung für einen idealen Leiter dar. Je nach Frequenz und Leitfähigkeit gilt die Näherung nicht mehr, und die elektrischen Parameter des Untergrundes müssen explizit berücksichtigt werden. Dies erfordert eine vollständige Lösung der Maxwellgleichungen.

Die geometrischen Parameter des Modells sind in Abbildung 2 illustriert. Die ideal leitende Kreisscheibe mit Radius a befindet sich im Abstand d über dem Untergrund. Der Mittelpunkt der Scheibe ist der Ursprung des zylindrischen Koordinatensystems. Es wird angenommen, dass alle zeitlichen Variationen harmonisch sind, die zeitliche Abhängigkeit sich also durch $e^{i\omega t}$ darstellen lässt. Die Leitfähigkeit des Untergrundes lässt sich dann als komplexe Größe darstellen: $\bar{\sigma}_1 = \sigma_1 + i\omega\epsilon_1$, wobei σ_1 die Gleichstromleitfähigkeit ist. Das Medium zwischen Untergrund und Scheibe hat die rein imaginäre Leitfähigkeit $\bar{\sigma}_0 = i\omega\epsilon_0$.

Wenn an die Scheibe ein Potential U angelegt wird, stellt sich eine radialsymmetrische Ladungsverteilung $q(r)$ ein. Die Gesamtladung Q erhält man dann aus

$$Q = 2\pi \int_0^a q(s) ds \quad (3)$$

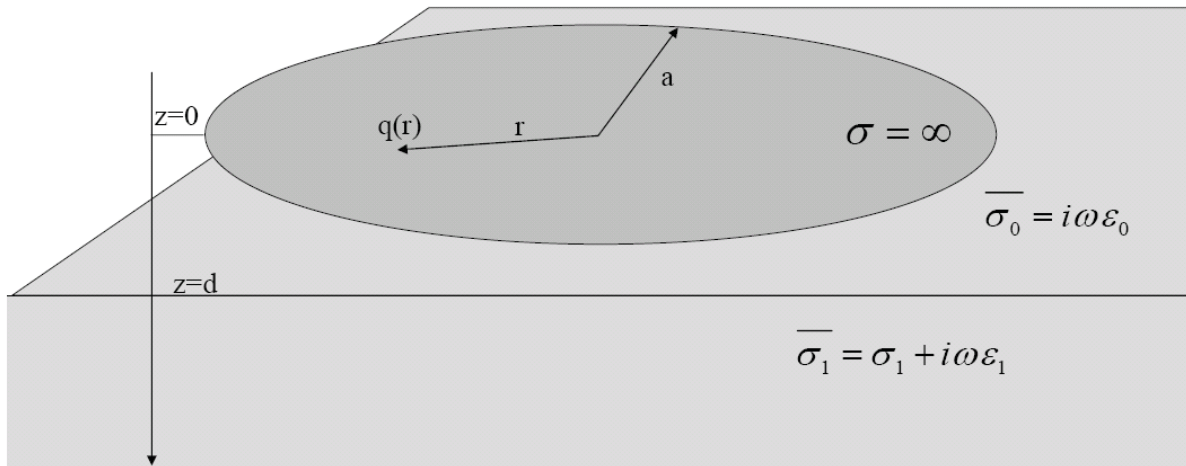


Abbildung 2: Parameter des Modelles der ideal leitenden Kreisscheibe über homogenem Halbraum.

Es muss also die Flächenladungsdichte $q(r)$ gefunden werden, aus der dann Q und mit

$$C = \frac{Q}{U} \quad (4)$$

die komplexe Kapazität, bzw. mit Gl. (1) die komplexe Impedanz der Kreisscheibe berechnet werden kann.

Aus der Forderung, dass die Leitfähigkeit auf der Scheibe unendlich ist, ergibt sich, dass das elektrische Feld auf der Scheibe verschwindet. Die Lösungsstrategie besteht darin, eine Gleichung für das elektrische Feld im gesamten Raum als Funktion von $q(r)$ zu berechnen und dann $q(r)$ so zu bestimmen, dass die Radialkomponente E_r auf der Scheibe verschwindet.

4 Theorie

4.1 Grundgleichungen

Um das elektrische Feld zu berechnen, müssen die Maxwellgleichungen in Zylinderkoordinaten gelöst werden. Aus dem Amperegesetz folgt unter der Berücksichtigung der harmonischen Zeitabhängigkeit:

$$\sigma \underline{E} + \frac{\partial \epsilon \underline{E}}{\partial t} = (\sigma + i\omega\epsilon) \underline{E} = \overline{\sigma} \underline{E} = \text{rot} \underline{H} \quad (5)$$

mit dem elektrischen Feld $\underline{E}$ und dem Magnetfeld $\underline{H}$. Die Leitfähigkeit und die Dielektrizitätszahl hängen von z ab. Aus Symmetriegründen besitzt das elektrische Feld nur eine Vertikal- und eine Radialkomponente, die nur von z und r abhängen. Das Magnetfeld besitzt nur eine von r und z abhängige φ -Komponente.

Damit folgt aus Gl. (5):

$$\begin{aligned}\bar{\sigma} E_r &= -\frac{\partial H_\varphi}{\partial z} \\ \bar{\sigma} E_z &= \frac{1}{r} \frac{\partial(r H_\varphi)}{\partial r}\end{aligned}\quad (6a,b)$$

Aus dem Induktionsgesetz:

$$-\frac{\partial \underline{B}}{\partial t} = \text{rot} \underline{E} \quad (7)$$

folgt:

$$-i\omega\mu_0 H_\varphi = \frac{\partial E_z}{\partial r} - \frac{\partial E_r}{\partial z} \quad (8)$$

Durch Einsetzen von (6a) und (6b) in (8) erhält man eine entkoppelte Gleichung für H_φ :

$$\partial_r \left(\frac{1}{r} \partial_r (r H_\varphi) \right) + \bar{\sigma} \partial_z \left(\frac{1}{\bar{\sigma}} \partial_z H_\varphi \right) = i\omega\mu_0 \bar{\sigma} H_\varphi \quad (9)$$

wobei die Schreibweise $\partial_r = \frac{\partial}{\partial r}$ angewandt wurde.

Der erste Term in Gl. 9 legt eine Trennung der Variablen mit Hilfe der Besselfunktion 1. Ordnung mit der Wellenzahl u nahe:

$$H_\varphi(r, z) = f(z, u) \cdot J_1(ur) \quad (10)$$

denn die Besselfunktion erfüllt die Bessel'sche Differentialgleichung:

$$\frac{\partial_r J_1(ur)}{r} + \partial_r^2 J_1(ur) - \frac{1}{r^2} J_1(ur) = -u^2 J_1(ur) \quad (11)$$

so dass mit (10) folgt:

$$\partial_r \left(\frac{1}{r} \partial_r (r H_\varphi) \right) = -u^2 H_\varphi \quad (12)$$

Bei dieser Strategie besteht eine Analogie zur Fouriertransformation, bei welcher z.B. eine zweifache zeitliche Ableitung nach Transformation in den Frequenzbereich mit der Funktion $e^{i\omega t}$ zu einer Multiplikation mit $-\omega^2$ führt. Entsprechend führt hier der Differentialoperator im 1. Term in Gl. (9) zu einer Multiplikation mit $-u^2$. Die Lösung von (9) lässt sich folglich darstellen als:

$$H_\varphi(r, z) = \int_0^\infty f(z, u) J_1(ur) du \quad (13)$$

Die Wahl der Besselfunktion 1. Ordnung ergibt sich daraus, dass aus Symmetriegründen H_φ bei $r=0$ verschwinden muss, was von J_1 erfüllt wird, aber nicht von J_0 . Aus dem Amperegesetz (6a und 6b) ergeben sich damit direkt die Gleichungen für die Komponenten des elektrischen Feldes:

$$E_r(r, z) = \frac{-1}{\bar{\sigma}} \int_0^{\infty} \frac{\partial f(z, u)}{\partial z} J_1(ur) du \quad (14)$$

$$E_z(r, z) = \frac{1}{\bar{\sigma}} \int_0^{\infty} u f(z, u) J_0(ur) du \quad (15)$$

Zur Bestimmung von $f(z, u)$ wird zunächst Gl. (13) herangezogen. Durch Einsetzen in (9) sieht man unter Benutzung von (12), dass die Funktion $f(z, u)$ die Gleichung

$$\bar{\sigma} \partial_z \left(\frac{1}{\bar{\sigma}} \partial_z f \right) = \alpha^2 f \quad (16)$$

erfüllen muss, mit

$$\alpha^2 = i\omega\mu_0\bar{\sigma} + u^2 \quad (17)$$

4.2 Bestimmung von $f(z, u)$

Gleichung (14) muss unter der Berücksichtigung der Randbedingungen an den Schichtgrenzen und der Quellbedingung gelöst werden. Die Quelle wird durch einen Sprung der Vertikalkomponente des elektrischen Feldes an der Scheibe beschrieben:

$$[E_z(r)]_{-}^{+} = \frac{q(r)}{\varepsilon_0} \quad (18)$$

wobei $[]_{-}^{+}$ die Differenz der Werte unmittelbar oberhalb und unterhalb der Scheibe, also den Sprung des Funktionswertes, kennzeichnet.

$$[E]_{-}^{+} = E(z = 0^+) - E(z = 0^-) \quad (18a)$$

Um aus (18) eine Bedingung für $f(r, z)$ abzuleiten, transformieren wir sie in den Wellenzahlbereich mittels Gl. (15), aus der folgt:

$$[E_z(r, z)]_{-}^{+} = \frac{1}{\bar{\sigma}} \int_0^{\infty} u [f(z, u)]_{-}^{+} J_0(ur) du = \begin{cases} \frac{q(r)}{\varepsilon_0}, & 0 \leq r \leq a \\ 0, & r > a \end{cases} \quad (19)$$

Die Ladungsdichte $q(r)$ lässt sich mittels der Hankeltransformation in den Wellenzahlbereich transformieren:

$$\tilde{q}(u) = \int_0^{\infty} u q(r) J_0(ur) dr \quad (20)$$

Mit $\tilde{q}(u)$ als der Ladungsdichte im Wellenzahlbereich.

Dementsprechend ist $q(r)$ darstellbar als:

$$q(r) = \int_0^{\infty} u \tilde{q}(u) J_0(ur) du \quad (21)$$

Durch Einsetzen von (21) in (19) wird klar, dass

$$[f(z,u)]_+^- = \frac{\bar{\sigma}_0}{\varepsilon_0} \tilde{q}(u) \quad (22)$$

d.h. die im Raumbereich formulierte Randbedingung (18) transformiert sich gemäß (22) in den Wellenzahlbereich.

Nach der Formulierung der Randbedingungen wird nun die Differenzialgleichung für f gelöst. Aus der Form von Gleichung (16) ergibt sich direkt, dass die Lösungen Funktionen der Form $e^{\pm \alpha z}$ sind, wobei berücksichtigt werden muss, dass α im unteren Halbraum und in der Luft unterschiedlich sind. Daraus ergibt sich folgender Ansatz für die Funktion $f(z,u)$:

$$f(z,u) = \begin{cases} -Ae^{+\alpha_0 z} + Be^{+\alpha_0 z}, & z < 0 \\ +Ae^{-\alpha_0 z} + Be^{+\alpha_0 z}, & 0 < z < d \\ Ce^{-\alpha_1 z}, & z > d \end{cases} \quad (23a,b,c)$$

wobei

$$\alpha_0^2 = i\omega\mu_0\bar{\sigma}_0 + u^2 \quad \text{und} \quad \alpha_1^2 = i\omega\mu_0\bar{\sigma}_1 + u^2 \quad (24a,b)$$

die Parameter in Luft und im unteren Halbraum sind.

Die Vorzeichen der Exponenten in den Ansatzfunktionen in (23) ergeben sich jeweils aus der Forderung, dass $f(z,u)$ im Unendlichen sowohl für $z < 0$, als auch $z > 0$ verschwinden muss.

Die Randbedingungen an Schichtgrenzen ergeben sich aus den Stetigkeitsbedingungen für die elektromagnetischen Felder. Da die Horizontalkomponenten von E und H an Schichtgrenzen stetig sind, folgt mittels (13) und (14), dass f und $\frac{1}{\bar{\sigma}} \frac{\partial f}{\partial z}$ stetig sein müssen.

Damit und mit der Sprungbedingung für f (Gl. (22)) lassen sich nun die Konstanten A, B und C in Gl. (23) berechnen. Beispielsweise folgt aus (23a,b), dass

$$\begin{aligned} f(z=0)_- &= B - A \\ f(z=0)_+ &= A + B \end{aligned} \quad (25a,b)$$

und damit für den Sprung von f bei $z=0$:

$$[f(z,u)]_+^- = f(z=0^+) - f(z=0^-) = 2A \quad (26)$$

und mit (22) erhält man hieraus für A :

$$A = \frac{\bar{\sigma}_0}{2\varepsilon_0} \tilde{q}(u) \quad (27)$$

Die Bestimmung von B und C aus den Stetigkeitsbedingungen erfolgt analog, ist aber komplexer und wird hier nicht im Einzelnen ausgeführt. Man erhält folgendes Endergebnis für f :

$$f(z,u) = \frac{\bar{\sigma}_0}{2\varepsilon_0} \tilde{q}(u) \begin{cases} e^{+\alpha_0 z} + R(u)e^{\alpha_0(z-2d)}, & z < 0 \\ e^{-\alpha_0 z} + R(u)e^{\alpha_0(z-2d)}, & 0 < z < d \\ T(u)e^{-\alpha_1(z-d)}e^{-\alpha_0 d}, & z > d \end{cases} \quad (28a,b,c)$$

mit Reflektions- und Transmissionsfaktor

$$R(u) = \frac{\alpha_0(u)\bar{\sigma}_1 - \alpha_1(u)\bar{\sigma}_0}{\alpha_0(u)\bar{\sigma}_1 + \alpha_1(u)\bar{\sigma}_0} \quad (29)$$

$$T(u) = \frac{2\alpha_0(u)\bar{\sigma}_1}{\alpha_0(u)\bar{\sigma}_1 + \alpha_1(u)\bar{\sigma}_0} \quad (30)$$

Mit (28-30) und (13-15) lassen sich dann alle Feldkomponenten berechnen, wobei für die Bestimmung der Ladungsdichte nur die Gleichung für E_r im Bereich zwischen Halbraum und Scheibe benötigt wird. Im Folgenden wird die Näherung $\alpha_0 = u$ verwendet. Im Vakuum ist nämlich

$$\alpha_0^2 = u^2 + \omega^2 \mu_0 \varepsilon_0 = u^2 + \frac{\omega^2}{c^2} \approx u^2 \quad (24c)$$

Die relevanten Wellenzahlen u ergeben sich aus der Dimension des Messsystemes, welches bei den hier berücksichtigten Frequenzen in jedem Fall klein ist gegen die elektromagnetische Wellenlänge im Vakuum. Es ist also $u \gg \omega/c$ und damit der 1. Term auf der rechten Seite vernachlässigbar.

Die Gleichung für E_r ergibt sich aus (28b) und (14), wobei die Ableitung von f nach z einen Faktor u unter dem Integral ergibt:

$$E_r(r, z) = \frac{1}{2\varepsilon_0} \int_0^\infty u \tilde{q}(u) [e^{-uz} - R(u)e^{u(z-2d)}] J_1(ur) du \quad 0 \leq z \leq d \quad (31)$$

4.3 Bestimmung der Ladungsdichteverteilung

Zur Bestimmung der Ladungsdichteverteilung wird wie oben erläutert gefordert, dass die Radialkomponente des elektrischen Feldes auf der Scheibe verschwindet. Benötigt wird also das Feld bei $z=0$:

$$E_r(r, z=0) = \frac{1}{2\varepsilon_0} \int_0^\infty u \tilde{q}(u) [1 - R(u)e^{-2ud}] J_1(ur) du \quad (32)$$

Die Idee ist nun, eine Ladungsverteilung $q(r)$ zu suchen, so dass E_r überall verschwindet. Damit lässt sich Gleichung (32) im Prinzip in ein Gleichungssystem überführen. Setzt man die Transformation der Ladungsdichte (Gl. 20) in (32) ein, so erhält man:

$$\frac{1}{2\varepsilon_0} \int_0^a \int_0^\infty s q(s) J_0(s) u [1 - R(u)e^{-2ud}] J_1(ur) du ds = 0 \quad (33)$$

Dabei wurde die neue Variable s eingeführt, die ebenso wie r von 0 bis a läuft, allerdings verschieden von r sein muss, da ja das elektrische Feld bei r von der gesamten Ladungsdichteverteilung abhängt. Definiert man nun die Funktion:

$$F(s, r) = \int_0^\infty u [1 - R(u)e^{-2ud}] J_0(us) J_1(ur) du \quad (34)$$

so kann man Gl. (33) schreiben als:

$$\frac{1}{2\varepsilon_0} \int_0^a s q(s) F(s, r) ds = 0 \quad (35)$$

Die Funktion $F(s,r)$ lässt sich nach Gl. (34) für jedes s und r direkt durch Integration über u berechnen. Die Scheibe wird also geeignet diskretisiert, und mit den diskreten Werten für s und r erhält man eine Matrix $\underline{\underline{F}}$. Gleichung 35 wird dann zur linearen Gleichung

$$\underline{\underline{F}} \cdot \underline{q} = \underline{0} \quad (36)$$

in der die Einträge von $\underline{q}$ die unbekannte Ladungsdichteverteilung darstellen. Durch Gleichung (36) ist $\underline{q}$ allerdings noch nicht vollständig bestimmt, da sie die triviale Lösung $\underline{q} = \underline{0}$ besitzt.

Um eine endliche Ladungsdichte zu erhalten, benötigt man eine Gleichung für das Potential der Scheibe. Eine Möglichkeit besteht darin, ein aus 2 identischen Scheiben bestehendes System zu betrachten (Abb. 3).

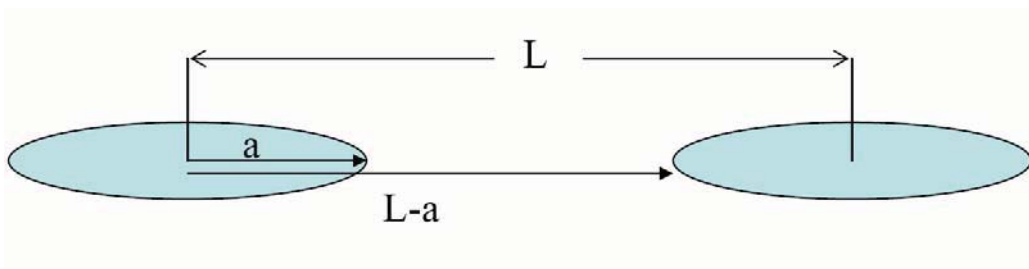


Abbildung 3: Geometrie des aus zwei identischen Scheiben aufgebauten Sendesystemes. Der Abstand der Mittelpunkte ist L , der Radius der Scheiben a .

Die Integration von E_r entlang der Verbindungslinie der beiden Scheibenmittelpunkte ergibt die angelegte Spannung U . Da $E_r=0$ auf den Scheiben, gilt:

$$U = \int_a^{L-a} [E_r(r,0) + E_r(L-r,0)] dr = 2 \int_a^{L-a} E_r(r,0) dr \quad (37)$$

$E_r(r,0)$ wurde in Gl. (32) berechnet und lässt sich direkt einsetzen. Die Integration über r kann man ausführen, und man erhält:

$$\begin{aligned} U &= 2 \int_a^{L-a} \frac{1}{2\epsilon_0} \int_0^\infty u \tilde{q}(u) [1 - R(u)e^{-2ud}] J_1(ur) du dr \\ &= \frac{1}{\epsilon_0} \int_0^\infty u \tilde{q}(u) (1 - R(u)e^{-2ud}) \left[\int_a^{L-a} J_1(ur) dr \right] du \\ &= \frac{1}{\epsilon_0} \int_0^\infty \tilde{q}(u) (1 - R(u)e^{-2ud}) [J_0(ua) - J_0(u(L-a))] du \end{aligned} \quad (38)$$

Um eine explizite Diskretisierung definieren zu können, wird die Ladungsdichte wieder mittels Gl. (20) im Raumbereich aufgeschrieben. Fasst man alle Terme mit u zusammen, so erhält man:

$$\begin{aligned} U &= \frac{1}{\epsilon_0} \int_0^a \int_0^a s q(s) J_0(us) ds (1 - R(u)e^{-2ud}) [J_0(ua) - J_0(u(L-a))] du \\ &= \frac{1}{\epsilon_0} \int_0^a s q(s) G(s) ds \end{aligned} \quad (39)$$

mit

$$G(s) = \int_0^{\infty} (1 - R(u)e^{-2ud}) J_0(us) [J_0(ua) - J_0(u(L-a))] du \quad (40)$$

Gleichungen (39) und (40) werden wiederum durch Diskretisierung der Scheibe in eine Gleichung der Form

$$\underline{q} \cdot \underline{G} = U \quad (41)$$

umgesetzt.

Gleichung (41) ist die inhomogene Bedingung, die nötig ist, um die triviale Lösung $\underline{q} = \underline{0}$ auszuschließen und eine vom Potential abhängige Ladungsdichte zu erhalten. Man wählt also ein beliebiges U und erhält aus der Lösung von (36) unter der Bedingung (41) eine Ladungsdichteverteilung $q(r)$. Die Gesamtladung erhält man mit Gl. (3), und die Kapazität aus Gl. (4).

Für ein aus zwei Scheiben bestehendes System gilt die Lösung nur in hinreichendem Abstand der Scheiben zueinander, da die gegenseitige Beeinflussung der Ladung auf den beiden Scheiben hier nicht berücksichtigt wurde. Eine quantitative Abschätzung dieses Effektes wurde bisher nicht vorgenommen. Betrachtungen mit halbkugelförmigen Elektroden im Gleichstromfall legen nahe, dass die vernachlässigten Terme von der Ordnung $(a/L)^4$ sind, so dass die Näherung schon bei geringen Elektrodenabständen gut funktioniert.

Die berechnete Kapazität C der Elektrode ist komplex, da der Reflektionsfaktor (Gl. 29) komplex ist. Aus ihr lässt sich mit

$$Z = \frac{1}{i\omega C}$$

die komplexe Impedanz der Elektrode, bzw. deren Betrag und Phase berechnen.

5 Ergebnisse

Um die Genauigkeit der Lösung zu bewerten, wurden die Ergebnisse mit einer analytischen Lösung für eine Kugelelektrode verglichen, die von Hördt (2007) vorgestellt wurde. Abbildung 4 zeigt die Geometrie und die Parametrisierung. Für die mittlere Schale mit Radius r_l wird hier Vakuum gewählt werden, mit $\sigma_l = 0$ und $\epsilon_l = \epsilon_0$ (Permittivität des Vakuums).

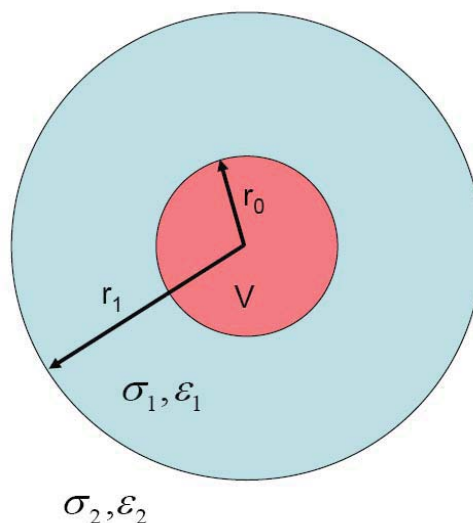


Abbildung 4: Geometrie der Kugelelektrode. Eine ideal leitende Elektrode mit Potential V und Radius r_0 ist in eine Kugelschale mit Radius r_l eingebettet.

Die Impedanz der Elektrode ergibt sich dann zu:

$$Z = \frac{1}{4\pi r_0 \sigma_1^*} \left(\frac{1 - \frac{\sigma_2^*}{\sigma_1^*} + \frac{\sigma_2^*}{\sigma_1^*} \frac{r_1}{r_0}}{\frac{\sigma_2^*}{\sigma_1^*} \frac{r_1}{r_0}} \right) \quad (42)$$

wobei

$$\sigma^* = \sigma + i\omega\varepsilon \quad (43)$$

In Abbildung 5 wird der Betrag der Impedanz der Scheibenelektrode mit der einer Kugelelektrode verglichen. Die Radien wurden so gewählt, dass die Oberfläche der Kugel gleich der einseitigen Fläche der Scheibe ist. Bei der Kugelelektrode wurde für die mittlere Schale die Werte für Vakuum eingesetzt, der Abstand entspricht der Dicke der mittleren Schale.

Für mittlere Abstände stimmen die Impedanzen der Kugelelektrode und der Scheibenelektrode überein, bei großen und kleinen Abständen sind die Impedanzen unterschiedlich, was auch so zu erwarten ist. Der Grenzwert großer Abstände entspricht der Impedanz der Elektrode im Vakuum. Die Impedanzen beider Elektroden in Abb. 5 konvergieren gegen den jeweiligen Grenzwert.

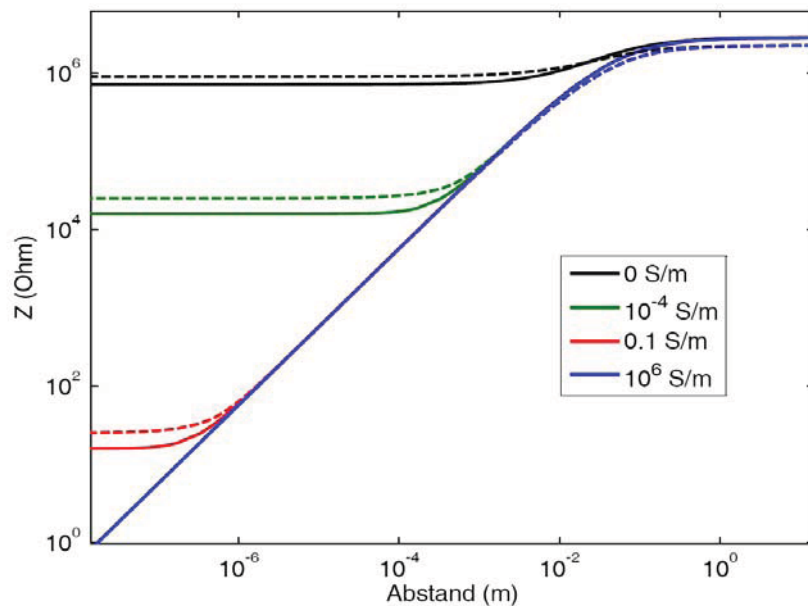


Abbildung 5: Impedanzbetrag der kapazitiven Elektrode gegen Abstand der Elektrode zum Medium für verschiedene Leitfähigkeiten des Mediums. Durchgezogene Linie: Kugelelektrode mit Radius 0.05m. Gestrichelte Linie: Scheibenelektrode mit Radius 0.1m. Die Frequenz ist $f=10$ kHz, die relative Permittivität des Untergrundes ist in beiden Fällen $\varepsilon_r=4$.

Für die Kugel mit Radius r_0 beträgt dieser (Smythe, 1968):

$$Z = \frac{1}{4\pi i\omega \varepsilon_0 r_0} \quad (44)$$

und für eine Scheibe mit Radius a :

$$Z = \frac{1}{8i\omega \varepsilon_0 a} \quad (45)$$

Wenn $a=2r_0$, wie für die Rechnung aus Abb. 5 der Fall, entspricht die Gesamtfläche der Kugel der einseitigen Fläche der Scheibe. Dann ist die Vakuumimpedanz der Scheibe etwas kleiner, weil im Vakuum beide Seiten der Scheibe gleich wirksam sind.

Für kleine Abstände ist der Vergleich von der Leitfähigkeit des Untergrundes abhängig. Bei moderaten Leitfähigkeiten (rote und grüne Kurve) konvergiert der Impedanzbetrag gegen den Grenzwert der galvanischen Ankopplung, also den Fall, dass die Scheibe aufliegt, bzw. die Kugel direkten Kontakt zum leitenden Medium hat. Der Grenzwert ist für die Kugel

$$Z = \frac{1}{4\pi\sigma r_0} \quad (46)$$

und für die kreisförmige Scheibe

$$Z = \frac{1}{4\sigma a} \quad (47)$$

wobei σ die Gleichstromleitfähigkeit ist.

In diesem Fall liefert die Scheibe bei gleicher Oberfläche (mit $a=2r_0$) die etwas größere Impedanz. Eine Gleichheit ist hier nicht zu erwarten, da nicht nur die Fläche wichtig ist, sondern auch die Stromverteilung im Medium. Bei sehr hohen Leitfähigkeiten in der Größenordnung von Metallen (blaue Kurve) wird der Grenzwert auch für extrem kleine Abstände noch nicht erreicht. Bei $\sigma=0$ (schwarze Kurve) sind nur die Verschiebungsströme (mit $\varepsilon_r=4$) zur Ankopplung wirksam. Die Scheibe hat eine etwas höhere Impedanz verglichen mit der Kugel.

Insgesamt verhalten sich die Kurven wie erwartet. Im mittleren Abstandsbereich ist nur die Ankopplungsfläche relevant und die Impedanzen von Kugel und Scheibe stimmen überein. Bei sehr großen und sehr kleinen Abständen werden jeweils die analytischen Grenzwerte erreicht, die für Kugel und Scheibe unterschiedlich sind.

Der Verlauf der Impedanz in Abbildung 5 hat zwei wichtige Konsequenzen. Zum Einen nimmt die Impedanz stetig als Funktion der Höhe zu. Dies schließt den Grenzwert einer galvanisch gekoppelten Elektrode ein. Das bedeutet, dass eine kapazitiv gekoppelte Elektrode niemals eine geringere Impedanz haben kann, als eine galvanisch gekoppelte, aufliegende. Zum Anderen ist der Übergang von der aufliegenden zur abgehobenen Elektrode stetig. Dies könnte der Intuition widersprechen; man würde evtl. einen Sprung der Impedanz zwischen aufliegender und abgehobener Impedanz erwarten.

Um das beobachtete Verhalten zu verstehen, ist in Abb. 6 die Impedanz in Real- und Imaginärteil aufgespalten dargestellt, mit anderem Abstandsbereich. Der Realteil (Mitte) der Impedanz ist für sehr große und sehr kleine Leitfähigkeiten verschwindend gering bzw. identisch null und wird nicht mit dargestellt. Bei moderaten Leitfähigkeiten (grüne und rote Kurve) wird die Impedanz durch den Realteil dominiert. Bei extrem niedriger und bei extrem hoher Leitfähigkeit überwiegt der Imaginärteil.

Dieses Verhalten lässt sich verstehen, wenn man die Gesamtimpedanz näherungsweise als Reihenschaltung zwischen einem kapazitiven Anteil und einem Erdungswiderstand betrachtet. Der kapazitive Anteil Z_c entspricht der Überwindung des Raumes zwischen Boden und Luft und ist rein imaginär, der Erdungswiderstand Z_e ist komplex und entspricht dem Ankopplungswiderstand der aufliegenden Platte. Man kann daher näherungsweise schreiben:

$$Z = Z_c + Z_e = \frac{d}{i\omega\epsilon_0\pi a^2} + \frac{1}{4a(\sigma + i\omega(\epsilon_0 + \epsilon_1))} \quad (48)$$

Zunächst einmal erklärt das Modell, warum der Übergang zwischen aufliegender und abgehobener Platte stetig ist und kein Sprung auftritt. Wenn der Abstand $d \rightarrow 0$, beginnt der 2. Term zu dominieren und „übernimmt“ stetig die Ankopplung. Allerdings ist in der Abbildung zu erkennen, dass der Übergang bei extrem kleinen Abständen erfolgt, die sogar unterhalb eines Atomdurchmessers liegen können. In der Praxis wird der stetige Übergang in der Regel also nicht zu sehen sein, sondern als Sprung erscheinen.

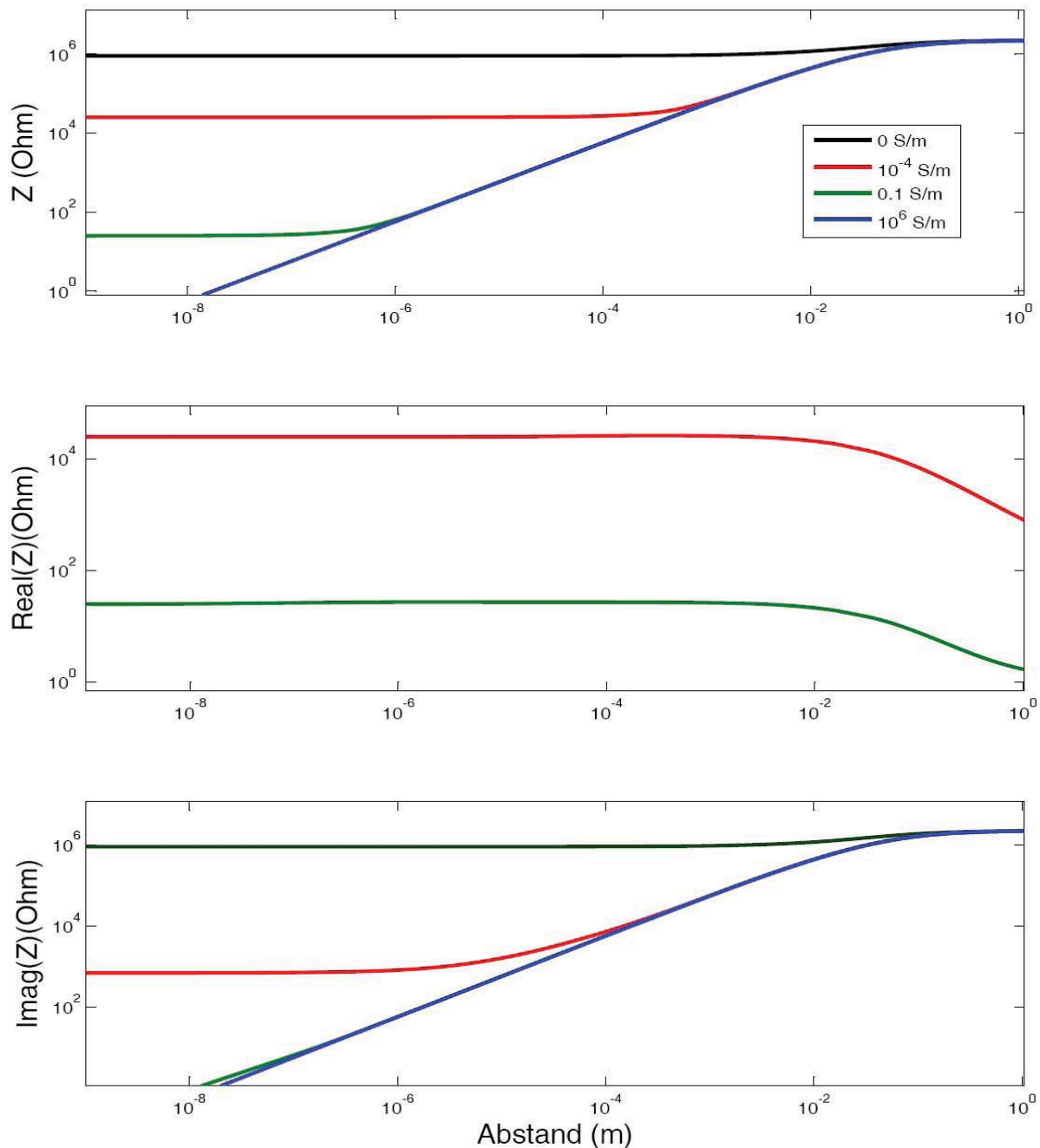


Abbildung 6: Betrag der Impedanz (oben), Realteil (mitte) und Imaginärteil (unten) für die Scheibe für dasselbe Modell wie in Abb. 5. Der Realteil der Impedanz (mitte) für $\sigma=0$ ist identisch null und nicht darstellbar, der Realteil für die sehr gut leitende Platte ist extrem klein und wird nicht dargestellt.

Das Modell aus Gl. (48) erklärt auch die Zerlegung der Impedanz in Real- und Imaginärteil. Bei sehr kleiner Leitfähigkeit und kleinem Abstand dominiert der Imaginärteil der Impedanz, bei moderater und hoher Leitfähigkeit dominiert der Realteil. Bei sehr hoher Leitfähigkeit wird der Abstand, in dem der 1. Term vernachlässigbar ist, in der Abbildung nicht erreicht und es dominiert immer noch der rein imaginäre 1. Term die Gesamtimpedanz.

6 Schlussfolgerungen

Es wurde ein Programm entwickelt, um die Impedanz einer kreisförmigen Scheibe über einem homogenen Halbraum zu berechnen. Das Verfahren beruht auf einer vollständigen Lösung der Maxwellgleichungen. Es verwendet eine Näherung, bei der angenommen wird, dass der Radius der Scheibe klein ist gegen die elektromagnetische Wellenlänge. Die Näherung verliert bei sehr hohen Frequenzen ihre Gültigkeit, ist jedoch beispielsweise für einen Radius 0.1m und eine Frequenz von 1 MHz vollkommen unproblematisch. Das Programm wurde durch Vergleich mit einer analytischen Lösung für eine Kugelelektrode und durch analytische Grenzwerte für große und kleine Abstände verifiziert.

Der Übergang zur Impedanz einer aufliegenden Elektrode erfolgt stetig als Funktion des Abstandes zwischen Elektrode und Halbraum. Die Impedanz der aufliegenden Scheibe ist immer kleiner als die der angehobenen. Dies bedeutet, dass die Übergangsimpedanz nicht durch eine kapazitive Elektrode gegenüber einer gleich gebauten, galvanischen verringert werden kann. Der galvanische Anteil der Impedanz ist auch bei kapazitiver Ankopplung mit enthalten.

In der Praxis kann es dennoch sinnvoll sein, kapazitive Elektroden statt galvanische zu verwenden, also den galvanischen Kontakt durch Isolierung oder Anheben zu verhindern. Bei aufliegenden Elektroden ist die Impedanz sehr schwer zu kontrollieren, da die Auflagefläche stark variieren kann. Dies kann dazu führen, dass die Impedanz um Größenordnungen schwankt, da sie sehr stark als Funktion des Abstandes variiert und die aufliegenden Teile die Ankopplung dominieren können. Eine Isolierung stabilisiert die Impedanz und verhindert starke Schwankungen, was für die Messung von Vorteil sein kann.

7 Referenzen

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Eine Methode gegen auslaufende Ag/AgCl//KCl(H_2O) Elektroden

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Einleitung

Im Rahmen der RiftLink Forschergruppe wurden in den Jahren 2007 und 2008 im Westen Ugandas langperiodische magnetotellurische (LMT) Messungen zur Erkundung der Kruste und des oberen Erdmantels durchgeführt. Für die LMT Messungen wurden Dreikomponenten Fluxgate Magnetometer (GEOMAG), Ag/AgCl//KCl(H_2O) Elektroden und GEOLORE (Roßberg 2007) Datenlogger verwendet (vgl.: Abbildung 1).

Die Ag/AgCl//KCl(H_2O) Elektrode selbst ist in einem mit einer gesättigten KCl Lösung gefüllten, abgeschlossenen PVC Behälter installiert. Die Keramik Fritten in der Behälterwand lassen einen Ladungstransport zu, verhindern aber das Auslaufen der KCl Lösung, die den Ladungstransport im Behälter zur Ag/AgCl//KCl(H_2O) Elektrode ermöglicht.

Bei Stationskontrollen während des Stationsaufbaus 2007 wurde festgestellt, dass die KCl Lösung der Elektroden ausgelaufen war. Dies äußerte sich zuerst in nicht induzierten Spannungsschwankungen zwischen je zwei Elektroden bei der Spannungskontrolle mit einem Multimeter und bestätigte sich, nachdem Ausgraben der Elektroden. Die Kontrolle der aufgezeichneten Zeitreihen zeigte, dass die Elektroden teilweise bereits nach drei bis fünf Tagen

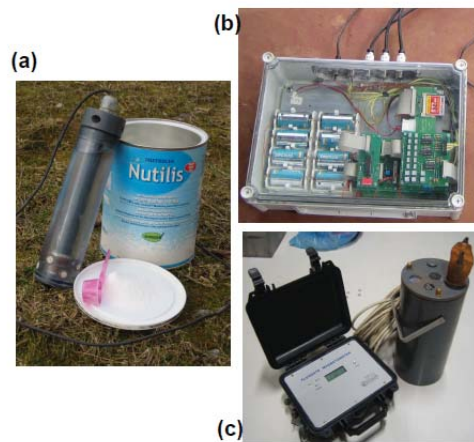


Abbildung 1: Messgeräte

(a) Ag/AgCl//KCl(H_2O) Elektrode mit Dickschichtpulver Nutilis, (b) Datenlogger GEOLORE, (c) Fluxgate Magnetometer (GEOMAG)

ausgelaufen waren und den Kontakt zum Boden verloren hatten. Die Elektroden wurden daher bei jeder weiteren Kontrolle ausgegraben und neu befüllt.

Die Ursache des Auslaufens liegt neben den heißen und trockenen äußeren Bedingungen im Messgebiet bei mechanischen Problemen bei der Herstellung des Elektrodengehäuses. Die Keramik Fritten wurden auf Grund ihres geringen Durchmessers von $5mm$ nicht mit dem Gehäuse verklebt. Dadurch sollte verhindert werden, dass Klebstoff vom Rand in die Poren der Keramik eindringt und den bereits geringen Durchmesser und damit die elektrisch leitende Kontaktfläche zwischen Elektrode und Boden weiter verkleinert. Die Keramik Fritten wurden daher lediglich in zum Durchmesser der Keramik passende Löcher das Gehäuse eingepresst. Die erforderliche Dichtheit wurde dadurch aber nicht erreicht, was unter anderem an den Toleranzen des Durchmessers der Keramik Fritten von $\pm 0.5mm$ lag.

Bei der Suche nach einer Lösung für dieses Problem sind wir auf Nutilis (Abbildung 1a) gestoßen. Nutilis ist ein Dickungspulver auf der Basis von Algen und wird im medizinischen Bereich in Kliniken bei Patienten mit Schluckproblemen eingesetzt. Flüssigkeiten werden mit Nutilis eingedickt, um von Patienten leichter geschluckt werden zu können. Nutilis ist in der Apotheke frei erhältlich und laut Packungsbeilage kann eine "honig"-artige Konsistenz von Flüssigkeiten erzeugt werden.

Für eine Elektrode mit einer Füllmenge von 300ml KCl Lösung wurde nach einigen Tests zwei Portionen des beiliegenden Messlöffels verwendet, um die KCl Lösung leicht zu einer Art Gel einzudicken. Der hohe Salzgehalt des Gels verhindert eventuelles Faulen der organischen Substanzen. Bei einem Test vor dem Institut, bei dem eine herkömmliche Elektrode mit KCl Lösung, und eine Gel Elektrode vergraben wurde, war der Flüssigkeitsverlust der herkömmlichen Elektrode in 30 Tagen deutlich, der der Gel-Elektrode gering. Die leicht veränderte Konsistenz der Salzlösung reicht aus, dass diese nicht mehr durch die undichten Stellen zwischen Keramik Fritten und Elektroden Gehäuse läuft.

Einfluss auf die Messergebnisse

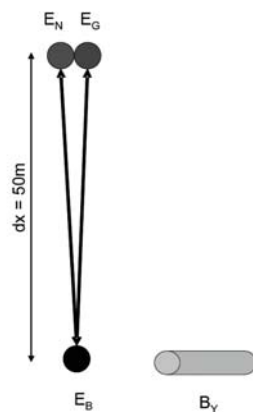


Abbildung 2: Gel-Elektroden Test: Schematischer Aufbau

Spannungsmessung zwischen: Gel-Elektrode (E_G) und Basis Elektrode (E_B), normalen Elektrode (E_N) und Basis Elektrode (E_B) mit einer Distanz in Nordrichtung $dx = 50m$.

Zusätzliche Messung der Magnetfeldkomponente B_Y .

Im Rahmen einer Diplomarbeit im Vogelsberg (Löwer 2009) wurde der Einfluss der Gel Elektroden auf die Messergebnisse gegenüber herkömmlichen Elektroden getestet. Der Testaufbau ist in Abbildung 2 dargestellt. Es wurde eine Gel Elektrode (E_G) und eine herkömmliche, Elektrode (E_N) in einem Loch vergraben und die Spannungen gegen eine gemeinsame Basis Elektrode (E_B) in 50m Entfernung in Nord-Süd Richtung mit der ADU07 (METRONIX) gemessen. Daraus wurden die elektrischen Felder E_{Gel} für die Gel Elektrode und E_{norm} für die herkömmliche Elektrode berechnet. Zusätzlich ist die Ost West Komponente (B_Y) des Magnetischen Feldes mit einer Induktionsspule aufgezeichnet worden. E_{Gel} , E_{norm} und B_Y kennzeichnen die gemessenen Felder im Frequenzbereich. In den folgenden Energiedichtespektren wurden die von künstlichem Noise gestörten Spektrallinien $16^{2/3}Hz$, $50Hz$ und deren Multiplen wegen ihrer großen Amplituden zur besseren Vergleichbarkeit der natürlichen Signale ausgeblendet. Die Frequenzachse ist zur Übersichtlichkeit linear dargestellt. Abbildung 3c zeigt das Energiedichtespektrum der Zeitreihe der Gel Elektrode (schwarz, durchgezogen) und der herkömmlichen Elektrode (rot, gestrichelt). Über den gesamten dargestellten Frequenzbereich zeigt sich eine hervorragende Übereinstimmung der Energiedichte beider Messungen. Die blaue Kurve stellt die Differenz zwischen der roten und der schwarzen Kurve dar und zeigt, dass der Unterschied zwischen den Spektren der herkömmlichen Elektrode und der Gel Elektrode nahezu zwei Dekaden kleiner als die Spektren selbst und damit sehr gering ist. In Abbildung 3d ist ein Ausschnitt aus Abbildung 3c von $0 - 250Hz$ zu sehen und zeigt neben den Spektren der E-Feld Messung noch zusätzlich in blau das Spektrum der B-Feld Messung und die frappierende Ähnlichkeit zwischen dem Energieinhalt der beiden Felder im Frequenzbereich.

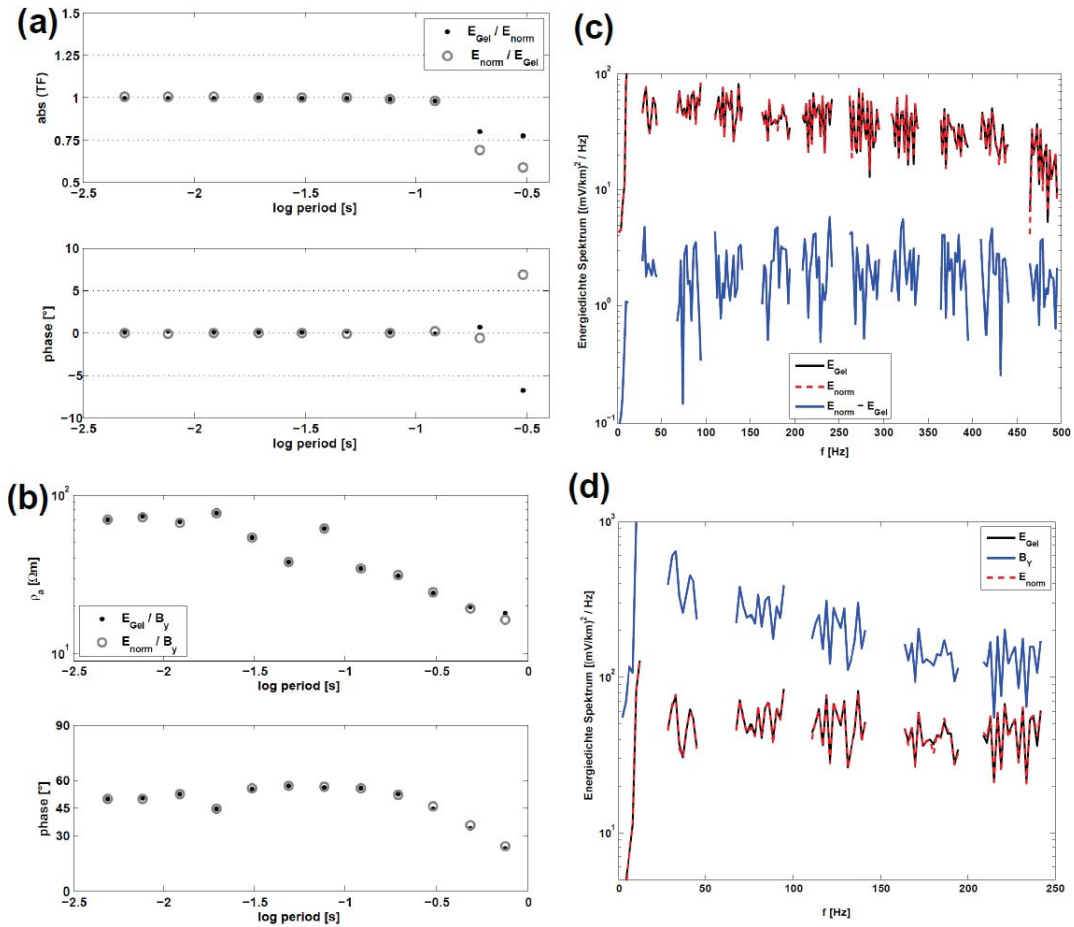


Abbildung 3: Spektren und Übertragungsfunktionen aus Gel-Elektroden Testmessungen
 (a) Absolutbetrag (oben) und Phase (unten) von univariaten Übertragungsfunktionen zwischen E_{Gel}/E_{norm} (schwarze Punkte) und E_{norm}/E_{Gel} (graue Kreise). (b) Absolutbetrag (oben) und Phase (unten) von univariaten Übertragungsfunktionen zwischen E_{Gel}/B_Y (schwarze Punkte) und E_{norm}/B_Y (graue Kreise), (c) Energiedichtespektren der Felder E_{Gel} (schwarz), E_{norm} (rot) und Differenz beider Spektren (blau) gegenüber der Frequenz aufgetragen. (d) Ausschnitt der Spektren von 0 – 250 Hz aus (a) mit Spektrum der Magnetfeldaufzeichnung (blau).

Zwischen E_{Gel} und E_{norm} wurden univariate Übertragungsfunktionen im Frequenzbereich berechnet und deren Betrag und Phase gegenüber der Periode aufgetragen:

$$T_{xy} = \frac{\langle XY^* \rangle}{\langle |Y| \rangle} \quad (1)$$

wobei X, Y stellvertretend E_{Gel} , E_{norm} , B_Y sein kann. Abbildung 3a zeigt die Übertragungsfunktionen, die einerseits aus dem Ansatz $E_{Gel} = T_{xy}E_{norm}$ (schwarze Punkte) und andererseits aus $E_{norm} = T_{xy}E_{Gel}$ (graue Kreise) berechnet wurden. Die Absolutbeträge

der schwarzen und grauen Kurve liegen bei eins und die Phasen bei Null von 204Hz bis 8Hz . Für den länger periodischen Bereich wird die Übereinstimmung schlechter, was auf schwache natürliche Anregung in diesem Frequenzbereich (dead band) zurückgeführt werden kann und auch in der Arbeit von Löwer (2009) ähnliche Probleme verursacht hat.

Univariate Übertragungsfunktionen aus den Ansätzen $E_{Gel} = T_{xy}B_y$ und $E_{norm} = T_{xy}B_y$ zwischen elektrischen- und magnetischen Feldern wurden berechnet und in Abbildung 3b als ρ_a - und Phasen Werte gegenüber der Periode dargestellt. Die schwarze Kurve zeigt die Übertragungsfunktion E_{Gel}/B_Y und die graue Kurve E_{norm}/B_Y . Beide Kurven zeigen nur geringe Abweichungen von einander über den berechneten Frequenzbereich von 204Hz bis 1Hz und zeigen dass das Gel keinen Störeinfluss im Vergleich zu herkömmlichen Elektroden auf die Messergebnisse hat.

Fazit

Die Grund für das Auslaufen der Frankfurter Elektroden wird auf mechanische Probleme zurückgeführt. Die hier vorgestellte Methode zur Abdichtung von $\text{Ag}/\text{AgCl}/\text{KCl}(\text{H}_2\text{O})$ Elektroden stellt eine preiswerte Alternative zum Bau bzw. Kauf neuer Elektroden dar. Das Gel in den Elektroden verhindert das Auslaufen und die Elektroden sind auf lange Sicht (2 Monate sind getestet) stabil. Das Gel beeinflusst die Aufzeichnungen der Elektroden nicht.

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Attitude Algorithm Utilised in Mobile Geophysical Measuring Systems

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1 Introduction

Accurate real-time tracking of orientation or attitude of rigid bodies has wide applications in robotics, aerospace, on-land and underwater vehicles, automotive industry, and virtual reality. A quite recent application in geophysics is the analysis of the attitude effect of geophysical sensors such as coil arrays or magnetometers attached to mobile measuring platforms e.g. helicopters and fixed wing airplanes and the correction with respect to the Earth's reference coordinate frame (*Reid et al, 2003, Yin and Fraser, 2004*). Another field of application is wireline logging with a borehole tool. While measuring, the tool spins about its vertical body axis due to the torsion force of the logging cable, which results in a disorientation of geophysical sensors.

In the borehole industry, it is essential to accurately monitor and guide the direction of the drill bit. It is also necessary for an oil rig to log the location of its boreholes at a regular frequency such that the oil rig can be properly monitored. To determine the location of a drill bit in a borehole, it is necessary to know the position and the attitude, which includes the vertical orientation and the North direction.

In prior art systems, a magnetometer is used to determine the magnetic field direction from which the direction of North is approximated. Triple component magnetometers were widely used to sense the tool rotation during a log run. However orientation via magnetometers is subject to restrictions. The employing of magnetometers in boreholes drilled into strongly magnetized formations such as oceanic crust, volcanoes and ore deposits, make it very difficult to utilize the Earth's magnetic field as a reference (Stevelling et al, 2003). First, these systems must make corrections for magnetic interference and use of magnetic materials for the drill pipe. Second, systems that rely only on magnetometers to determine North can suffer accuracy degradation due to the Earth's magnetic field variations. Third, a magnetometer alone cannot give an unequivocal measurement of a set of sensor attitude. Measurements made with such a sensor define the angle between the Earth's magnetic field and a particular axis of the sensor. However, this axis can lie anywhere on the surface of a cone of semi-angle equal to that angle about the magnetic vector. Hence, an additional measurement is required to determine attitude with respect to another fixed reference frame.

In order to determine the direction of a spinning borehole tool another type of sensor technology must be utilised to provide orientation that is independent from the Earth's magnetic field as a reference (Galliot et al, 2004, Stoll and Leven, 2002).

The problem to be solved consists of a way to sense the attitude of a borehole tool, which are free to rotate about any direction. A good compromise between accuracy and sensor size are the fibre optical gyros. These devices measure the turning rate about the sensor axis in the tool frame. In this paper an algorithm is presented that approaches the attitude determination based on this discrete data.

2 Principle of Fibre Optical Gyro (FOG)

Gyroscopes are used in various applications to sense the angular rate of turn about some defined axis. The most basic and the original form of gyroscopes make use of the inertial properties of a wheel, or rotor, spinning at high speed. The wheel tends to maintain the direction of its spinning axis in space due to conservation of the angular momentum vector. These devices are susceptible to damage from shock and vibration, exhibit cross-axis acceleration sensitivity and, for the lower cost versions, have reliability problems.

Another type is the vibratory gyroscope. The basic principle of operation of such sensors is that the vibratory motion of part of the instrument creates an oscillatory linear velocity. If the sensor is rotated about an axis orthogonal to this velocity, a Coriolis acceleration is induced. The acceleration modifies the motion of the vibrating element which is an indication of the magnitude of the applied rotation. However, this type of sensor tends to produce biases in the region of $1^\circ/\text{s}$.

Optical gyroscopes use an interferometer to sense angular motion. An optical gyroscope, laser or fibre, measures the interference pattern generated by two light beams, traveling in opposite directions within a mirrored ring or fibre loop, in order to detect very small changes in motion. This type of gyroscope can be subdivided into fibre optical gyro (FOG), ring laser gyro (RLG), and ring resonator gyro (PARR).

In order to keep track on the orientation of the geophysical sensors the utilisation of fibre optical gyroscopes are now common in navigation and replace prior art systems like mechanical gyros. Important attributes of this new technology are no moving parts, high reliability, stable performance and low costs. Many of its components are based on proven technology from the fiber optical telecommunications industry. Using optical gyros, inertial navigation is accomplished by integrating the output of a set of angular rate sensors to compute the attitude.

In **figure 1** a FOG manufactured by LITEF is shown exemplary. The gyro operates at an optical wavelength of 820 nanometers, with a 110 meter coil of elliptical-core polarization maintaining fiber. The low coherence reduces unwanted interference between waves reflected from the fusion splices used to join the components.

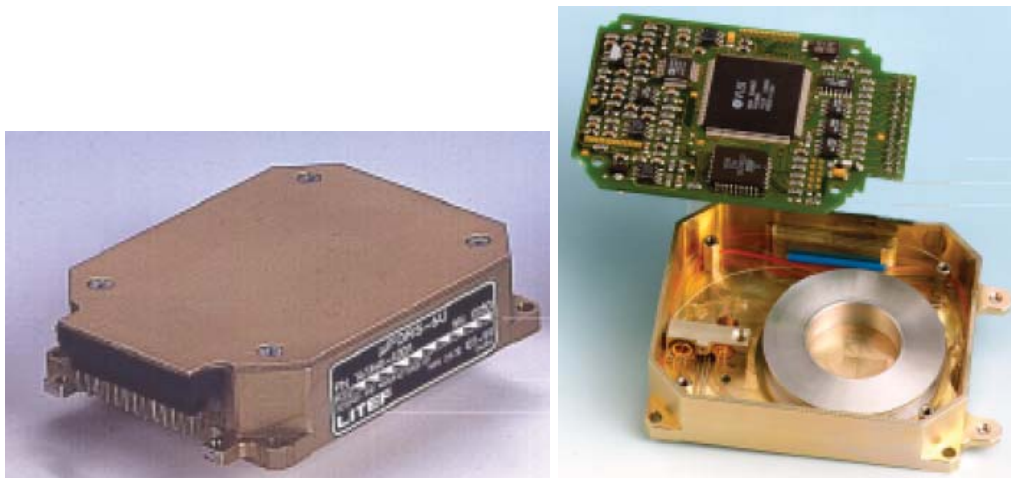


Fig. 1: Example of the μ FORS -family (LITEF, Freiburg) (here μ FORS-6U)

3 Attitude Computation

Inertial orientation tracking of borehole tools is based upon the same methods and algorithm as those used for aircrafts, ships, and missiles. There is a large quantity of technical literature that describes the fundamentals of inertial navigation technology in great detail, e.g., Grewal et al. (2001), v. Hinüber (1993), Savage (1998a, b), Stovall (1997). If angular rates of a tool are measured constantly with depth, the orientation of the sensors of the tool with respect to inertial space can be determined by applying a suitable coordinate transformation.

A transformation from one coordinate frame to another can be carried out as three successive rotations about different axes. The transformation matrix is given by the product of these three separate transformations as follows:

$$C_b^n = C_3 \cdot C_2 \cdot C_1 \quad (1)$$

Various mathematical representations can be used to define the attitude of a body with respect to a coordinate reference frame. One of the most common ways of parameterizing the transformation matrix is by use of direction cosine matrix (DCM). Other methods of describing rotations are the use of Euler Angles and Quaternions.

The DCM is a 3x3 matrix, the columns of which represent unit vectors in body axes projected along the reference axes. Matrix C_b^n describes the transformation from coordinate frame "b" to the frame "n" and is written here in component form as follows:

$$C_b^n = \begin{pmatrix} c_{11} & c_{12} & c_{13} \\ c_{21} & c_{22} & c_{23} \\ c_{31} & c_{32} & c_{33} \end{pmatrix} \quad (2)$$

The element in the i^{th} row and the j^{th} column represent the cosine of the angle between the i^{th} axis of the reference frame “n” and the j^{th} axis of the body frame “b”.

When designing a navigation system it is necessary to relate the information from the sensors to a navigation coordinate. The coordinate choice for borehole applications is a geographic or navigation frame “n” with axes {N, E, D}, (North, East and Down). The inertial measurement unit is mounted on the borehole tool constituting a new frame. This is called the body frame “b”, and has axes {X, Y, Z}. This frame will be in rotation with respect to the geographic frame. The velocity of this rotation is measured by three orthogonal gyros. The transformation matrix that relates “b” and “n” coordinate frames propagates with time in accordance with the following equation:

$$\dot{C}_b^n = C_b^n \Omega_{nb}^b, \text{ where } \Omega_{nb}^b = \begin{pmatrix} 0 & -\omega_z & \omega_y \\ \omega_z & 0 & -\omega_x \\ -\omega_y & \omega_x & 0 \end{pmatrix} \quad (3)$$

Ω_{nb}^b is the skew symmetric matrix formed from the elements of the vector $\omega_b^n = [\omega_x, \omega_y, \omega_z]^T$, which represents the turn rate of the body between the i^{th} -frame and the $(i+1)^{\text{th}}$ -frame as measured by the gyroscopes in the body frame. In real time applications the integration is implemented with the following approximation

$$C_b^n(t + \delta t) = C_b^n(t) \cdot A(t) \quad (4)$$

where $A(t)$ is the DCM which relates the b-frame at time t to the b-frame at time $t + \delta t$. For small angle rotations, $A(t)$ may be written as follows:

$$A(t) = [I + \delta\varphi] \quad (5)$$

where I is the 3×3 identity matrix and

$$\delta\varphi = \begin{pmatrix} 0 & -\delta\gamma & \delta\beta \\ \delta\gamma & 0 & -\delta\alpha \\ -\delta\beta & \delta\alpha & 0 \end{pmatrix} \quad (6)$$

in which $\delta\alpha$, $\delta\beta$ and $\delta\gamma$ are small rotation angles through which the body-frame has rotated over the time interval δt about its yaw, pitch and roll axes respectively. If the limit δt approaches zero, small angle approximations are valid and the order of the rotation becomes unimportant.

Hence, the minimum sampling time of the gyros is important for obtaining the transformation matrix with reasonable accuracy. This interval will be a function of the severity of manoeuvres expected from the borehole tool. In many applications the rotation velocity expected is usually less than 25 degrees/sec. With a sampling time of 100 Hz the maximum angle variation will be less than 0.25 degree satisfying the small angle approximation. Rotations with faster dynamics will require smaller sampling time to compute the transformation matrix appropriately.

In order to update the DCM C_b^n , it is necessary to solve a matrix differential equation (3). Over a single computer cycle, from time t_k to t_{k+1} , the solution of the equation may be written as:

$$C_{k+1} = C_k \cdot \exp \int_{t_k}^{t_{k+1}} \Omega dt \quad (7)$$

Provided that the orientation of the turn rate vector ω remains fixed in space over the update interval, we may define:

$$\int_{t_k}^{t_{k+1}} \Omega dt = [\sigma \times] \quad (8)$$

and

$$C_{k+1} = C_k \cdot \exp[\sigma \times] = C_k \cdot A_k \quad (9)$$

Where C_k represents the direction cosine matrix which relates body to reference axes at the k^{th} computer cycle, and A_k the direction cosine matrix which transforms a vector from body co-ordinates at the k^{th} computer cycle to body co-ordinates at the $(k+1)^{\text{th}}$ computer cycle.

The variable σ is an angle vector with direction and magnitude such that a rotation of the body frame about σ through an angle equal to the magnitude of σ will rotate the body frame from its orientation at computer cycle k to its position at the computer cycle $k+1$. The components of σ are denoted by σ_x , σ_y , and σ_z and its magnitude given by:

$$\sigma = \sqrt{\sigma_x^2 + \sigma_y^2 + \sigma_z^2} \quad (10)$$

and

$$\sigma \times = \begin{pmatrix} 0 & -\sigma_z & \sigma_y \\ \sigma_z & 0 & -\sigma_x \\ -\sigma_y & \sigma_x & 0 \end{pmatrix} \quad (11)$$

Expanding the matrix exponential function in a power series gives:

$$A_k = I + [\sigma \times] + \frac{[\sigma \times]^2}{2!} + \frac{[\sigma \times]^3}{3!} + \frac{[\sigma \times]^4}{4!} + \dots \quad (12)$$

where

$$[\sigma \times]^2 = \begin{pmatrix} -(\sigma_y^2 + \sigma_z^2) & \sigma_x \sigma_y & \sigma_x \sigma_z \\ \sigma_x \sigma_y & -(\sigma_x^2 + \sigma_z^2) & \sigma_y \sigma_z \\ \sigma_x \sigma_z & \sigma_y \sigma_z & -(\sigma_x^2 + \sigma_y^2) \end{pmatrix} \quad (13)$$

$$[\sigma \times]^3 = -\sigma^2 \cdot [\sigma \times] \quad (14)$$

$$[\sigma \times]^4 = -\sigma^2 \cdot [\sigma \times]^2 \quad (15)$$

Thus we may write

$$\begin{aligned} A_k &= I + [\sigma \times] + \frac{[\sigma \times]^2}{2!} - \sigma^2 \frac{[\sigma \times]}{3!} - \sigma^2 \frac{[\sigma \times]^2}{4!} + \dots \\ &= I + \left(1 - \frac{\sigma^2}{3!} + \frac{\sigma^4}{5!} - \dots\right) \cdot [\sigma \times] + \left(\frac{1}{2!} - \frac{\sigma^2}{4!} + \frac{\sigma^4}{6!} - \dots\right) \cdot [\sigma \times]^2 \end{aligned} \quad (16)$$

and which may be written as follows:

$$A_k = I + \frac{\sin \sigma}{\sigma} [\sigma \times] + \frac{(1 - \cos \sigma)}{\sigma^2} [\sigma \times]^2 \quad (17)$$

Equation (17) is input in equation (9) that updates the frame at time t to the frame at time $t+\Delta t$. To compute the attitude of the tool, first the measured rotation rates are input in equation 17. Then, beginning with $C_0 = I$ the rotation matrices C_{k+1} for each computer cycle $k+1$ are successively computed by equation (9). The tool attitude $\vec{r}_k^n$ in the external frame at the k^{th} cycle results to

$$\vec{r}_k^n = C_k \cdot \vec{r}_0^n \quad (18)$$

with the starting orientation $\vec{r}_0^n$.

4 Attitude Computation allied to the Göttingen Borehole Magnetometer (GBM)

The Göttingen Borehole Magnetometer (GBM) contains a triple of μ FORS-36M sensors, which are attached rigidly to the borehole tool. The rotation axes of the sensors are aligned with the body axes of the tool, and thereby sense the rotation rate about the vertical axis and the tilting movement about the two horizontal axes. Angular rates are output with a sample rate of 1 second. The salient specifications of the μ FORS-36m are the maximum input rotation rate of $\pm 720^\circ/\text{s}$ that allows sensing a maximum rotation rate of 2 revolutions per second with a resolution of $9 \cdot 10^{-5}^\circ$. Moreover, the dimensions of the sensor (53mm x 58mm x 19mm) are small enough to fit to the inner diameter of 65mm of the GBM. This is important to reduce the overall diameter of the tool and make it applicable in narrower boreholes. The bias drift is $< 6^\circ/\text{h}$ (1σ) and is reduced to $< 3^\circ/\text{h}$ at working temperature of 30°C to 50°C . The maximum operating temperature range is -40°C to $+75^\circ\text{C}$.

To check the performance of the attitude computation, the GBM was brought to the Gauss Haus at the Geophysical Institute of the University of Göttingen where it was fixed to the ceiling and hung vertically down. Then it was exposed to different kinds of movements like tilting and rotation about its vertical axis or combined movements. This test is visualized in two video clips, [video 1](#) and [video 2](#). Here, the tool attitude with the corresponding turning rate and the magnetic field data are shown in three components. The maximum rotation rate was $150^\circ/\text{s}$. In this case the rotation angles are no longer small, which resulted in an erroneous representation of the attitude of the tool. **Figure 2** displays the movement of the GBM with respect to the {N, E, D}, (North, East and Down) coordinate frame

Figure 2

Acknowledgments

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Joint Interpretation of Magnetotelluric and Seismic Models for Exploration of the Gross Schoenebeck Geothermal Site

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Introduction

Due to the non-uniqueness of the inverse problem, the interpretation of geophysical models in terms of geological units is not always straightforward. It is common to use a combination of different geophysical methods to obtain the distribution of independent physical properties over the area of interest in order to discriminate between the different lithologies or geologic units. This kind of studies is usually limited to qualitative comparisons of the different models, which may – or may not – support a relation between the parameters in certain areas. Quantitative approaches are in general based on empirical relations between physical parameters, which often are not of universal applicability.

The magnetotelluric (MT) and seismic methods resolve the physical parameters electric resistivity (ρ) and (seismic) velocity (V_p , V_s) respectively with similar spatial resolution and are often used in combination to derive earth models. By looking at both resistivity and velocity simultaneously, we can keep the strengths of both methods while avoiding their weaknesses. The problem with a joint interpretation is that there is no unique universal law linking electrical and acoustic properties. While electrical resistivity in deep sedimentary basins is mostly sensitive to the pore geometry and contents, seismic velocity is mostly imaging rock matrix properties. However, with a statistical analysis of the distributions of both resistivity and velocity, we can find certain areas of the models space where a particular relation between the physical parameters holds locally, thus allowing us to characterize this region as a particular lithology. In the present work, we use a statistical analysis, as described by Bedrosian et al. (2007) in order to correlate two independently obtained models of the Groß Schönebeck geothermal test site in the Northeast German Basin.

Methodology description

The methodology used in the present paper was described by Bedrosian et al. (2007) and is based on a probabilistic approach developed by Bosch (1999), in a sense that diverse geophysical parameters are represented as a probability density function (pdf) in the joint parameter space. The coincident velocity and resistivity models are first interpolated onto a common grid. Therefore, a joint parameter space is built, where each point in the modelled area is associated with a velocity – resistivity pair. By plotting one parameter against the other in a cross-plot and including the error estimates we can then construct a joint pdf in the parameter space. The areas of enhanced probability can be identified with classes represented by a certain range of values in both resistivity and velocity. By mapping back these classes onto the spatial domain they can be related to certain lithologies and/or geological units.

In the present work, the resistivity model was interpolated onto the seismic mesh, given that it is uniformly spaced and finer than the magnetotelluric mesh. An inverse distance weighted interpolation scheme was used, which forms estimates from a weighted average of many samples found within a pre-defined area around the point, with decreasing weights with distance.

Each element of this distribution can be interpreted as the outcome of a process defined by a probability density function (pdf). Assuming normal error distribution and independence of the data, the joint pdf is expressed as the sum of the individual pdfs (pdf_i) for each data point,

according to the following expression, with the errors of the resistivity and velocity ($\delta \log(\rho_i)$ and $\delta V_{p,i}$) estimated from the sensitivity matrix and the hit count distribution respectively:

$$pdf(\rho, v_p) = \frac{1}{n} \sum_{i=1}^n pdf_i(\rho, v_p)$$

$$pdf_i(\rho, v_p) = \frac{1}{\sqrt{2\pi\delta \log(\rho_i)\delta V_{p,i}}} \times \exp - \frac{1}{2} \left[\frac{(\log(\rho) - \log(\rho_i))^2}{\delta \log(\rho_i)^2} + \frac{(V_p - V_{p,i})^2}{\delta V_{p,i}^2} \right]$$

The different classes are identified as zones of enhanced probability in the joint pdf cross-plot. Assuming that the geological units are characterized by uniform physical properties normally distributed, each class is defined by a mean point and a covariance matrix (representing the ~60 per cent confidence interval ellipse for the peak) in the joint parameter space.

Geophysical models

The Groß Schönebeck low enthalpy geothermal site, with the well doublet GrSk 3/90 and GrSk 4/05, is located in the Northeast German Basin (NEGB). MT data was collected along a 40 km-long profile centred on the well doublet. The profile consists of 55 stations with a site spacing of 400 m in the central part (close to the borehole) of the profile, increasing to 800 m towards both profile ends. The period range of the observations was 0,001 to 1000 s. This profile is spatially coincident with the seismic tomography profile and most of the stations were located at the same places as the seismic shot points. At all sites, we recorded horizontal electric and magnetic field components and the vertical magnetic field.

The resistivity model for the MT profile (Figure 1a) shows a shallow conductive layer extending from the surface down to depths of about 4 km, with an antiformal-type shape below the central part of the profile. At a depth range of 4-5 km two conductive bodies are found, separated by a region of moderate conductivity. According to the seismic tomography, which shows high velocity values for depths greater than 4 km, a resistive basement was introduced *a priori* in the resistivity model (Muñoz et al., 2010).

A 40 km long seismic profile was measured coincident with the MT experiment (Figure 1b). The objective was to derive a regional 2-D seismic model, which can be combined with the electrical conductivity model from the MT data analysis to study the potential reservoir layers and overlying sediments. The experimental setup was designed to provide data suitable for refraction tomography. 45 explosion shots were fired from 20 m deep boreholes with charge sizes of 30 kg. The shot spacing was 800 m on average. The recording instrumentation consisted of 4.5 Hz 3-component geophones. These were deployed as a 40 km long receiver spread with spacings of 200 m. Each shot was recorded by all receivers.

The velocity model (Figure 1b) can be divided into three major sequences: The upper section (depth range 0-2 km) is characterized by low velocities (2-3.5 km/s) and a strong increase of velocity with depth. The section between 2 and 4 km depth shows velocities between 4 and 4.5 km/s and is characterized by a strong topography on top which is related with salt mobility. The third, deepest section is bounded on top by a subhorizontal interface at 4.2 km depth and reveals velocities of more than 5 km/s (Bauer et al., 2010).

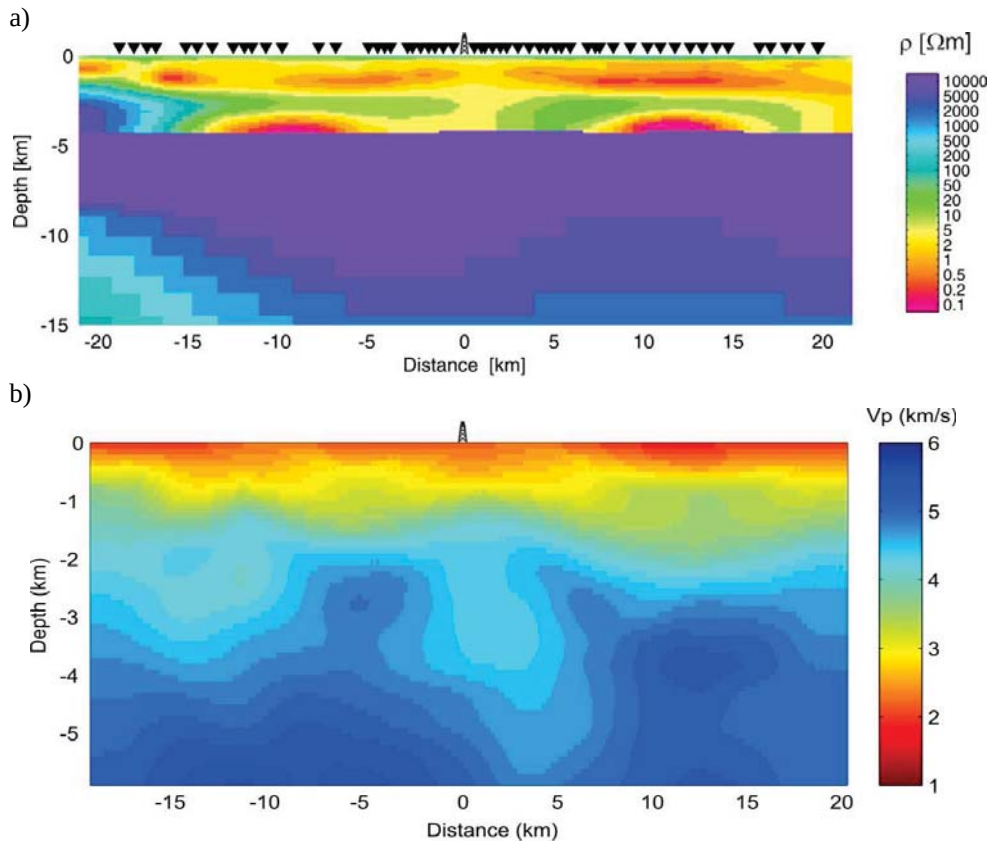


Figure 1: Electrical resistivity model obtained from inversion of the magnetotelluric data using a priori information from the seismic velocity model for the deeper part of the model (> 5 km) (a) and seismic velocity model obtained from (V_p) travel time inversion (b).

Joint analysis

In the cross-plot of the probability density function (Figure 2a) we can identify five more or less clear peaks, or areas of enhanced probability with respect to the neighbouring region. The coloured ellipses in Figure 2a represent the ~ 60 per cent confidence intervals of the Gaussian peaks best fitting the pdf. The clusters were mapped back to the cross section providing a depth distribution of the classes along the profile (Fig. 2c). In order to interpret the nature of these litho-types, the model is superimposed on the stratigraphy derived from pre-existing reflection seismic data and borehole information (Moeck et al., 2008).

Class 1, the shallowest, is characterised by low velocity (1.8 – 2.7 km/s) and moderate resistivity (5 – 70 Ωm) and comprises of unconsolidated sediments. Class 2, with higher velocities (2.7 – 3.9 km/s) and lower resistivities (0.5 – 3.5 Ωm) encompasses weak or soft rocks with high porosity, which are more conductive because they provide storage for a greater volumes of fluids. Class 3 coincides with successions of Middle Triassic to Lower Permian. Significantly these successions represent harder brittle rock of limestone and sandstone as indicated by increasingly higher velocities (4 – 5 km/s) and resistivities (2 – 15 Ωm). This class includes also thick salt rock layers (Zechstein) which, however, yield no significant variation of resistivity or velocity within the class. Class 4, the deepest one, is characterized by the highest velocity and resistivity values (4.7 – 5.5 km/s and around 3000 – 30000 Ωm). It represents the basin floor, comprising volcanic rock, quartzite and slate. This

class reflects the fact that the magnetotelluric model used seismic *a priori* information to introduce a resistive basin floor. Obviously this causes the very high correlation between velocity and resistivity of class 4 in Fig. 2b. Class 5 is characterized by high velocities (4.7 – 5.5 km/s) but extremely low resistivities (0.1 – 0.7 Ωm).

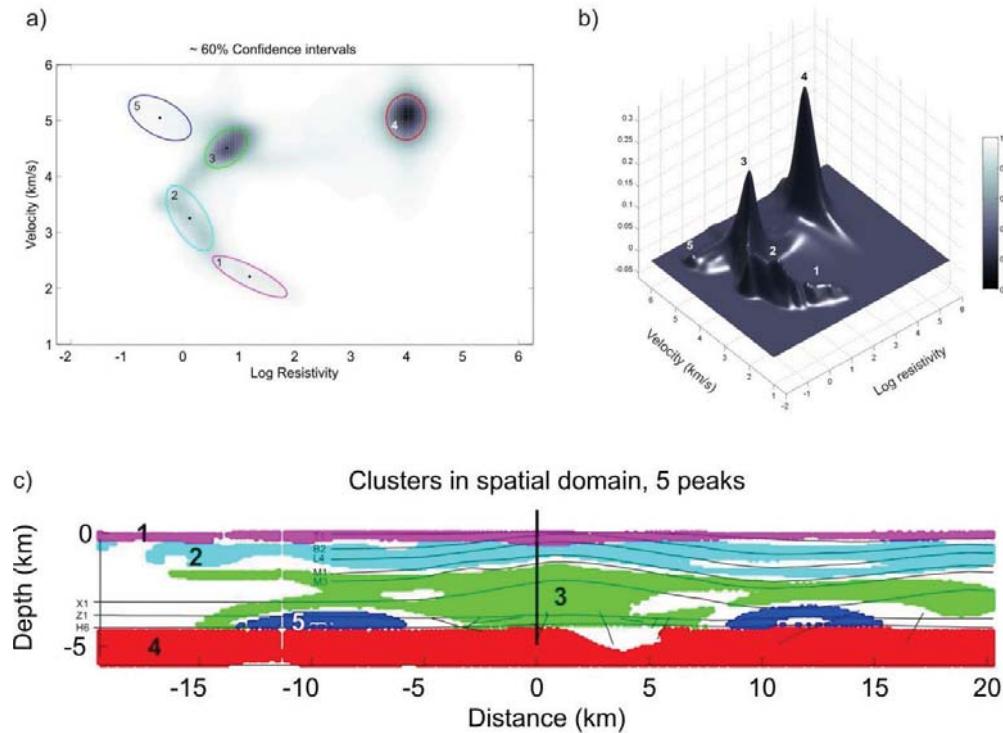


Figure 2: Cross-plot of the probability density function in gray-scale plan view (a) and three-dimensional view (b). Spatial distribution of classes (c). Colours correspond with the ellipses in (a) defining the class boundaries.

It is remarkable that Class 5 is restricted to salt lows where presumably anhydrites of Upper Permian age remain after salt movement. These anhydrites have a brittle behaviour and are expected to be highly fractured. In this case, an estimation of the resistivity by using Archie's Law (Archie, 1942) for fracture-controlled porosity and assuming a formation fluid salinity of 260 g/l (Giese et al., 2001), and a reasonable range of porosities and temperatures (15% and 130°C) the modelled resistivities of 0.1 – 0.7 Ωm can be explained. High velocities can be explained by the high density of anhydrite (2.9 g/cm³). The classes are summarized in the Table 1, below.

Class	ρ (Ωm)	V_p (km/s)	Lithology	Stratigraphic unit(s)
1	5 – 70	1.8 – 2.7	Unconsolidated sediments (sand, gravel clay)	Cenozoic
2	0.5 – 3.5	2.7 – 3.9	Clay, marl, marly limestone, limestone, sandstone	Cretaceous – Upper Triassic
3	2 – 15	4 – 5	Brittle limestone, sandstone, evaporites	Middle Triassic – Lower Permian
4	$3 \cdot 10^3$ – $3 \cdot 10^4$	4.7 – 5.5	Volcanic rock, quartzite, slate	Lower Permian - Carboniferous
5	0.1 – 0.7	4.7 – 5.5	Anhydrite	Lower Permian

Table 1: Resistivity and P velocity of the classes from figure 9. Also included are lithology and stratigraphic units.

Conclusions

MT and seismic data were used to derive independent 2-D models of the electrical conductivity and the seismic P velocity around the geothermal research well GrSk 3/90. The resulting models were combined in a statistical analysis to determine correlating features in both models. The classification method used in this analysis revealed 5 distinct litho-types which show up as separate clusters in the underlying geophysical parameter space of Pa and VP.

This study demonstrates the concert of MT, seismic and structural geologic models. Therefore the combination of different geophysical methods (MT and seismic) combined with structural geological information revealed information which was previously unknown and which could not be determined with the individual methods. Clearly, this approach is applicable in other areas too and could represent a promising new approach for geothermal exploration.

Acknowledgements

This work was funded within the 6th Framework Program of the European Union (I-GET Project, Contract n° 518378). The instruments for the geophysical experiments were provided by the Geophysical Instrument Pool Potsdam (GIPP). We wish to thank Paul Bedrosian for making his statistical analysis code available to us.

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Magnetotelluric measurements to explore for deeper structures of the Tendaho geothermal field, Afar, NE Ethiopia

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Introduction

In regions with high heat flow, like at volcanically active plate margins, high total thermodynamic energy is accumulated in the so called high enthalpy resources. The East African Rift System is one of the privileged areas to yield numerous sites for potential geothermal energy extraction including power generation. Along the Ethiopian part of the rift high temperature geothermal resources are associated with zones of quaternary tectonic and magmatic activity. Including the Afar depression at least 120 independent geothermal systems have been identified since the 1970s (UNEP 1973) about 24 of them are judged to have high enthalpy potential. To explore for these resources up to a pre feasibility stage a range of geoscientific methods is used in a defined sequence starting with regional reviews and remote sensing followed by geologic, hydrologic, geochemical and geophysical surveys.

The applied geophysical methods usually comprise temperature measurements (gradient boreholes), seismology, magnetics and resistivity methods, including Magnetotellurics (MT). Geothermal surface manifestations like hot springs, fumaroles, geysers and the associated geological and geochemical settings are indicating the presence of a geothermal reservoir.

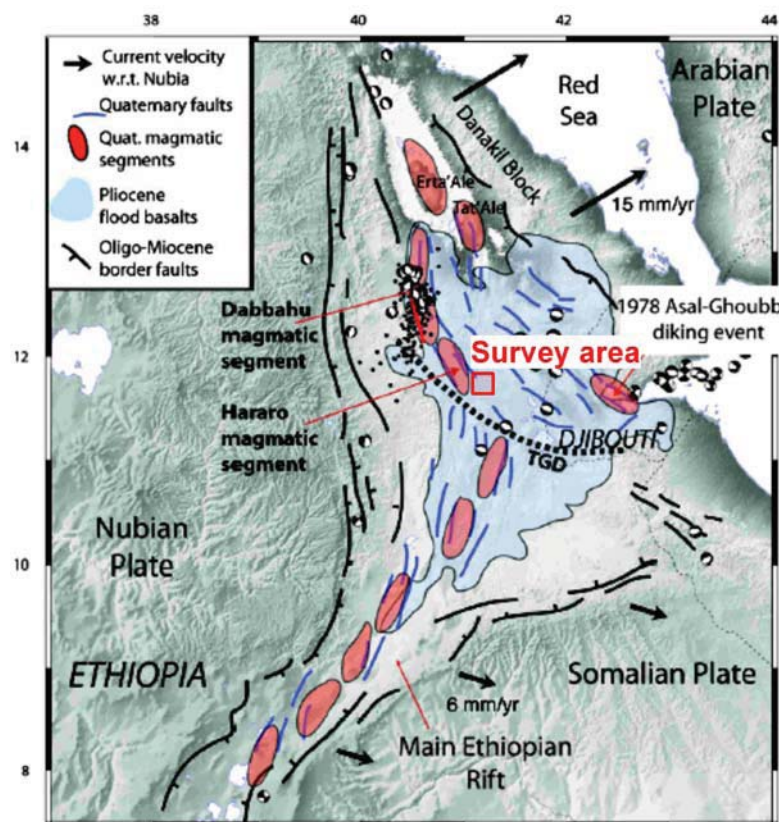


Figure 1: Survey area at the SE end of the Hararo and Dabbahu magmatic segments. TGD = Tendaho-Goba'ad Discontinuity (after Ebinger et al. 2008).

Particularly resistivity methods may be used for delineating the lateral and depth extensions of such potential reservoirs. The MT method is frequently used for this purpose since it easily covers the necessary exploration depth down to approximately 10 km.

In the following paper MT data and interpretation is presented, showing the already known shallow reservoir of the Tendaho geothermal field and its so far unknown deep structure, possibly feeding the shallow reservoir.

Survey area and tectonic setting

The Tendaho geothermal field is located in the Afar depression (NE Ethiopia), at or very close to the assumed triple junction formed by the Red Sea, Gulf of Aden and East African

rift arms. The extension of the Manda-Hararo axial rift zone in its south-easterly strike direction ends up at the Tendaho geothermal system (figure 1). The Red Sea and Aden rifts

are characterised by oceanic crust. The Afar triple junction is therefore a zone where thinned continental crust of the Main Ethiopian rift joins with crust of oceanic character (Barberi et al. 1972).

Within this area of active extensional tectonics several geothermal manifestations can be observed. Quaternary faulting and volcanism formed chains of fissural flows, basaltic cones and stratovolcanoes, usually accompanied by shallow seismicity and positive gravity anomalies. These active magmatic segments are comparable to slow spreading mid oceanic ridge segments and are thought to form new igneous crust by dyke like intrusions as demonstrated in the recent Dabbahu rift events (since 2005 up to now) just 80 km NW of the survey area. A previously mapped magmatic segment within the Manda-Hararo rift zone had been ruptured. Earthquake recordings indicated a 60 km long and about 8 m wide dyke intrusion (Ebinger et al. 2008, figures 8 & 9).

Tendaho geothermal field

The Tendaho geothermal field has been investigated in detail since the 1980s. Geological, geochemical as well as magnetic, seismological and resistivity data (DC soundings) were acquired by an extensive Ethiopian-Italian cooperation (Aquater, 1996). Based on this results, six exploratory wells had been drilled, three of them hitting a production horizon at approx. 300 m depth yielding steam temperatures above 250 °C (figure 2). Additional gravity and magnetic data had been acquired recently by Lemma and Hailu (2006) to delineate fractured zones which could serve as potential pathways for hydrothermal fluid flow from a deeper reservoir.

To further investigate the proposed deep reservoir and/or heat source, deep reaching resistivity methods had to be applied. Due to the generally low resistivities the penetration depths of the DC soundings were limited to a few hundred meters. To extend this depth of exploration the MT method has been applied by BGR in collaboration with the Geological Survey of Ethiopia (GSE) during a field survey in 2007 (figure 3). To reach the desired exploration depth of about 5 km a frequency range from



Figure 2: Production test of well TD5 (at well head: $T > 250\text{ °C}$, $p > 18\text{ bar}$). The well is located inside the MT survey area.

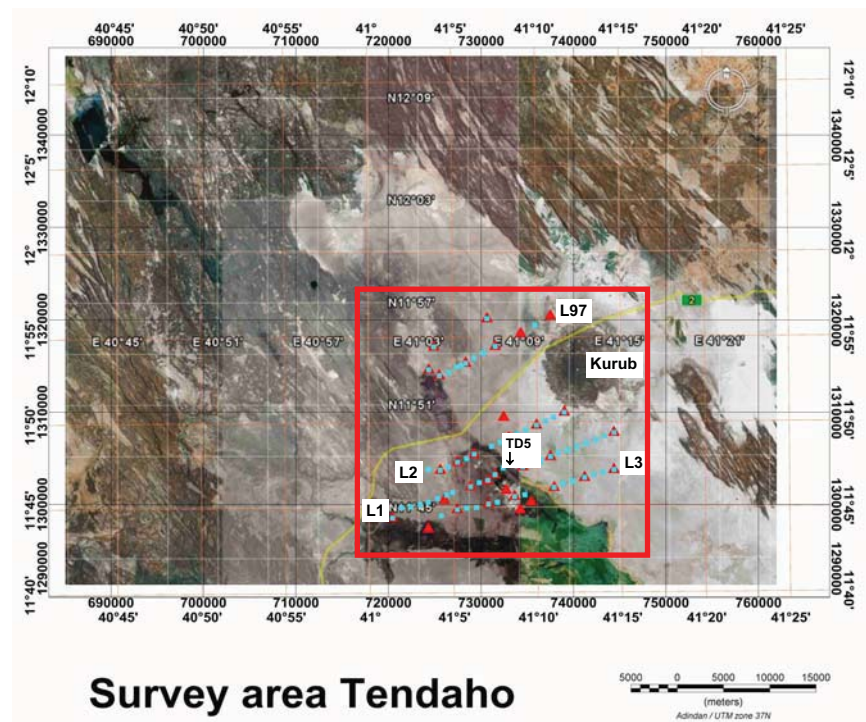


Figure 3: Survey area Tendaho geothermal field (red frame) projected onto satellite images (google earth). Red triangles = MT stations, blue squares = TEM stations. Line nos. L1, L2, L3 and L97 refer to MT lines in SE-NW bearing. TD5 = productive exploratory well. Yellow line = main road to Djibouti.

10 kHz to 0.01 Hz (100 s) has been acquired.

MT resistivity signature

Due to alluvial and lacustrine infill of the Tendaho graben apparent resistivity curves show generally values below 10 Ohm*m, even at high frequencies. A typical sounding curve is shown in figure 4.

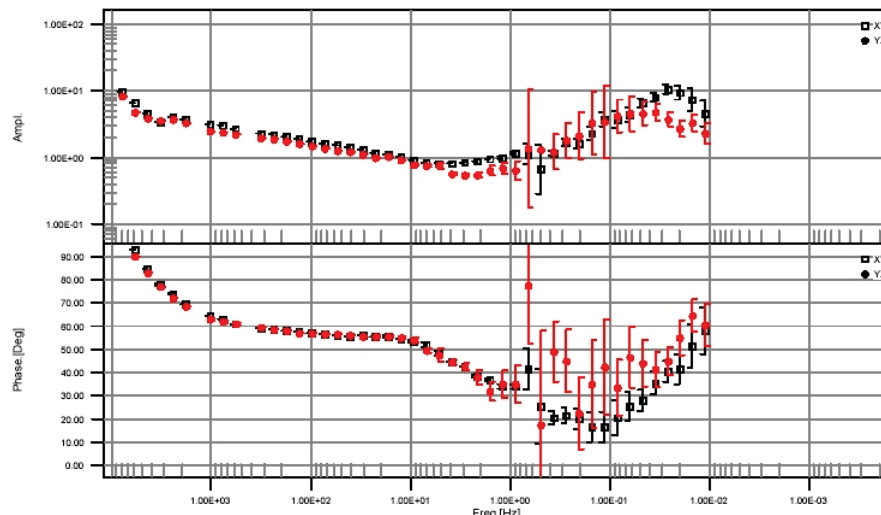


Figure 4: Typical sounding curve. Apparent resistivities (top) and phases (bottom). Due to low signal strength in the dead band from about 1 to 0.1 Hz unstable estimation especially of phase values, resulting in increased error bars.

The minimum is reached at approximately 4 Hz with apparent resistivities just below 1 Ohm*m. With decreasing frequency the apparent resistivity rises again. Source signal was generally quite low, especially within the *dead band* region ranging from approximately 1 to 0.1 Hz (1 to 10 sec.). Nonetheless 16 hrs of recording time proved to be sufficient to yield acceptable estimates of the impedance tensor, using mainly single site processing.

This apparent resistivity pattern, which is also reflected in the resistivity models (see below), is indicative for high temperature geothermal reservoirs (figure 5, see also e.g. Kalberkamp 2007) where the resistivity low is interpreted as the clay cap while the increasing resistivities below the clay cap point to the core of the reservoir and represent a possible drilling target.

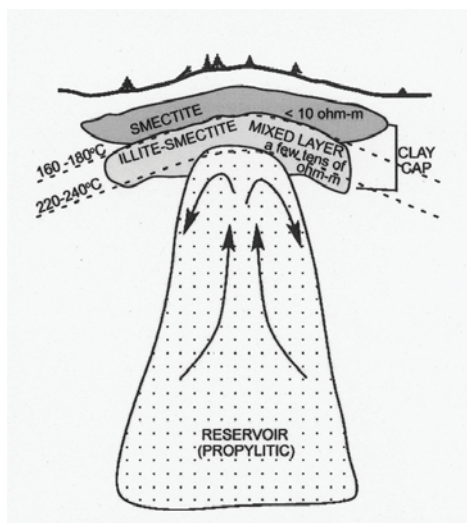


Figure 5: Schema of a generalised geothermal system. The smectite cap formed exhibits resistivities in the range of 2 Ohm*m, the mixed layer around 10 Ohm*m (modified after Johnston et al. 1992).

In the 2d inverted resistivity sections, exemplarily shown for profile L1 in figure 6, this resistivity pattern can be seen quite clearly. Although resistivities are generally low, they show signatures typical for hydrothermal alteration halos (see e.g. Johnston et al. 1992, Kalberkamp 2007) and (partly) molten magma intrusions at depth below 4 km. Taking the Dabbahu rift events into account it seems to be likely that the heat source for the geothermal reservoir is formed by magma intrusions along dyke like fracture zones as it is suggested by our interpretation of the MT data. Areas with high resistivities (>300 Ohm*m) may be associated with basalts from the Afar *Stratoid* Series, constituting the borders of the Tendaho graben structure.

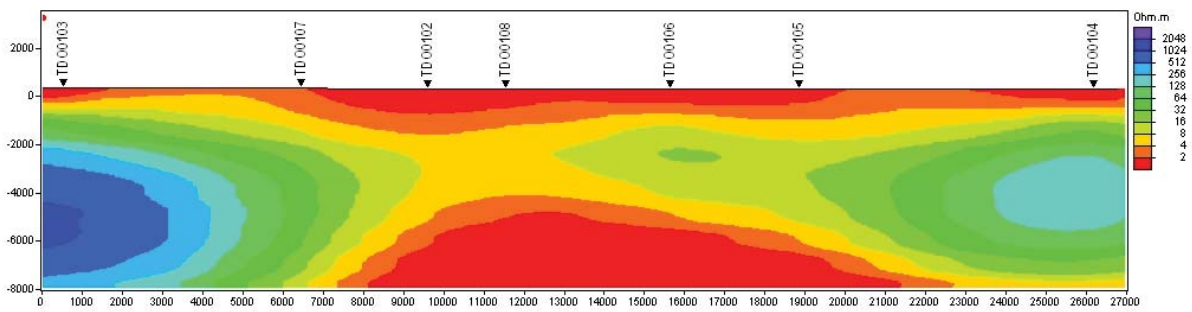


Figure 6: 2 d inverted resistivity section (L1). Lateral extension is 27 km; vertical depth covers 8 km.

Interpretation and recommendations

As has been shown for profile L1 exemplarily the magnetotelluric soundings show in their 2-dimensional inverted resistivity sections regions with resistivities as low as 2 Ohm*m, especially in near surface layers along profiles L1, 2, and 3 as well as at greater depth (below 5 km) in profiles P1, 3, and 97. When using the 2d MT sections to compile resistivity maps for constant elevations each, thus based on the inverted resistivities, we get a general view of the lateral resistivity structure as presented in figure 7 for selected elevations.

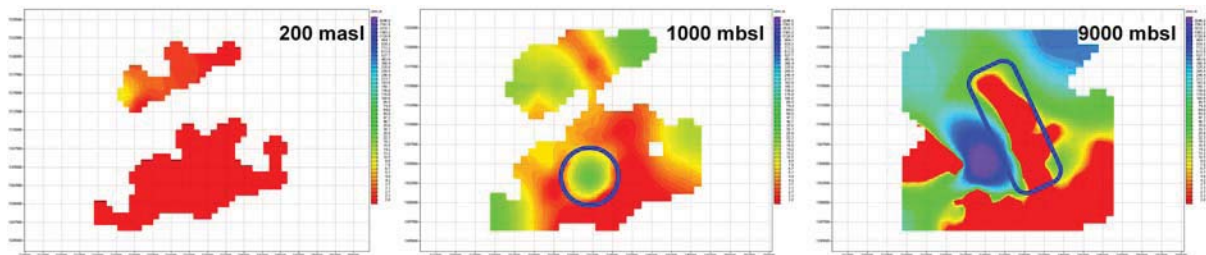


Figure 7: Resistivity maps of survey area as covered by red frame in figure 3, resistivity range 2 (red) – 2048 (blue) Ohm*m. *Left:* at 200 masl (150 m below surface). Shallow low resistive layer (red) due to sedimentary infill. The sediments are up to 1 km thick. *Centre:* at 1000 mbsl (1350 m below surface). Slight increase of resistivity (> 10 Ohm*m) possibly due to mixed layer clays and advancement towards deeper reservoir (blue circle). *Right:* at 9000 mbsl (9350 m below surface). Resistivity drop below 2 Ohm*m along NW-SE trending feature (partly molten magma dyke?). This may form the deep heat source feeding the shallow geothermal reservoir.

At greater depth of around 7 km a resistivity anomaly below 2 Ohm*m appears elongating in NW-SE direction. This direction coincides with the strike direction of the Tendaho graben and rift structures further to the NW, up to the Dabbahu rift and Boina vent, where lava ascended along a fault system, almost reaching the surface (figure 8). Therefore it seems likely, that the low resistive structures at greater depth are caused by lava filled fracture zones. (figure 7 *right*).

Ebinger et al. (2008) have presented a working model of the Dabbahu magmatic



Figure 8: Volcanic vent, created during 2006 eruption event (Photo by J. Rowland).

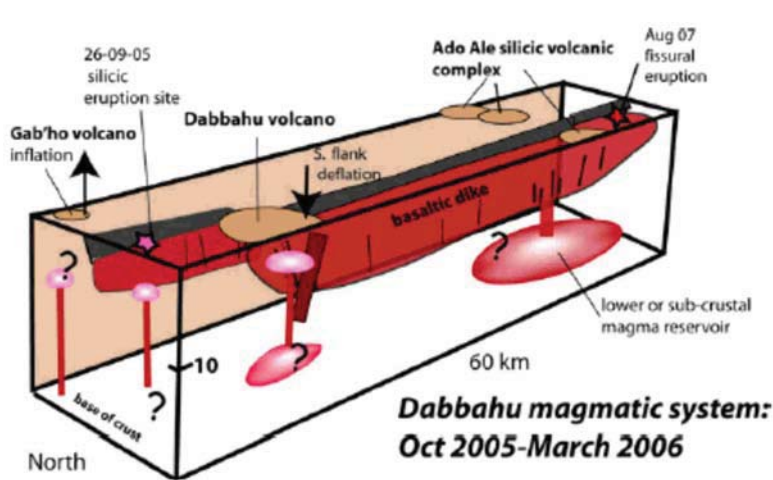


Figure 9: Working model of the Dabbahu magmatic system as presented by Ebinger et al. (2008). Pink ellipses = shallow magma chambers, connected to assumed lower crust/upper mantle feeding zones.

system (figure 9) based mainly on seismological, radar interferometric and structural data. They suggest that lower crustal/upper mantle source zones are feeding the basaltic dyke and shallow magma chambers at a few kilometres depth.

Magnetotelluric measurements by the Afar Rift Consortium / University of Edinburgh (Scotland, U.K.) are ongoing in that area and results thereof may help to refine the interpretation of the Tendaho geothermal field data.

With the current data therefore the following preliminary conclusions may be drawn:

- The deep heat source feeding the currently drilled reservoir at 300 m depth is most likely (partly) molten magma along NW-SE trending dykes or faults. In the Tendaho geothermal field these structures may be as shallow as 4 km depth.
- A deep reservoir may be expected below 1300 m.
- An up flow zone could be present at the S end of the survey area indicated by low resistivities throughout the acquired depth range.

To identify the proposed up flow zone at the S end of the current survey area it is recommended to apply additional TEM soundings which could enhance the resolution at shallow depth.

Additional MT sounding profiles are recommended further towards the Hararo and Dabbahu magmatic segments NW of the survey area. Since rift events including ascending magma are evident further to the NW it may be assumed that high temperature reservoirs could be reached at comparatively shallow depth there. First MT results from the research work within the Afar Rift Consortium do support this assumption (Desissa et al., 2009).

Also additional MT soundings beyond the Awash river up to the geothermal manifestations of Alalobad in the south would be helpful to establish either the extension of one large reservoir or the existence of a separate second geothermal system.

Acknowledgements

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GEOTHERM (www.bgr.de/geotherm/) is a technical cooperation programme to promote the use of geothermal energy in partner countries, implemented by the BGR on behalf of the German Federal Ministry for Economic Cooperation and Development (BMZ) under contract no. 2002.2061.6.

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Electromagnetic Monitoring of CO₂ Storage in Deep Saline Aquifers - Numerical Simulations and Laboratory Experiments

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1 Introduction

The knowledge of petrophysical parameters and their contrasts is crucial to reliably monitoring CO₂ storage processes. The electrical conductivity appears to be a sensitive indicator for a resistive gaseous or supercritical CO₂ phase replacing a conductive pore fluid in a porous medium like a saline sandstone aquifer. However, detailed knowledge on the influence of supercritical CO₂ on the electrical resistivity of a formation is not sufficiently available yet.

Therefore, we have carried out laboratory experiments to predict the contrast in electrical resistivity due to the presence of CO₂ in an initially water-saturated sand sample resembling the petrophysical situation typical for a reservoir. Furthermore, the expected parameter contrasts were estimated according to empirical equations and numerical simulations (Fig. 1).

On a middle- to long-term perspective, we aim at developing an electromagnetic monitoring technique using a borehole transient electromagnetic sensor which offers a unique opportunity to generate enhanced sensitivity at depth with respect to detecting migrating CO₂ in a reservoir. This work is therefore integrated into national CCS research programs with an interdisciplinary variety of partners.

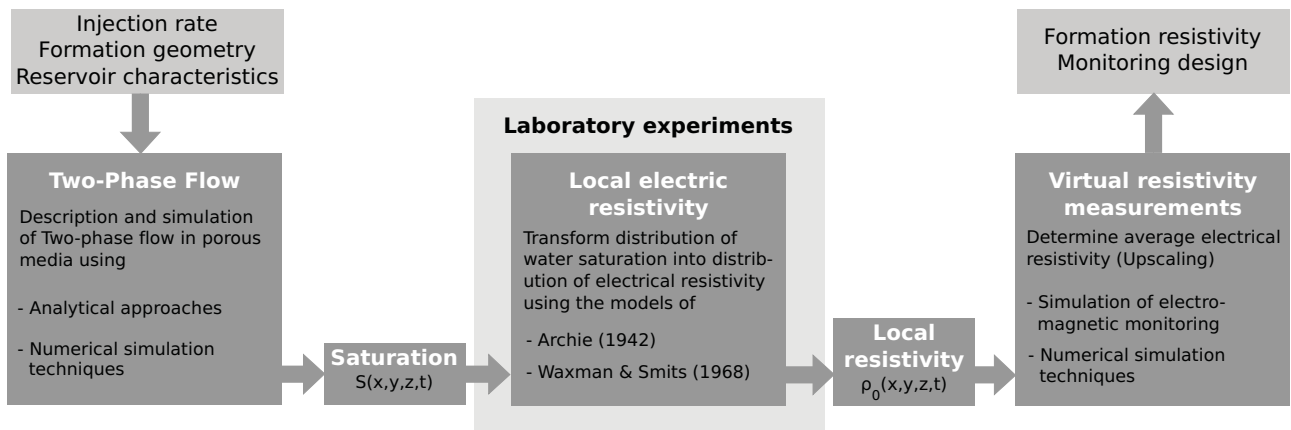


Figure 1: Workflow of the numerical simulation steps necessary for the transformation of water saturation into electrical resistivity.

2 Theory

Two-phase flow is governed by two equations simultaneously enforcing continuity of the water (w) and the CO₂ (co2) phase flow (Busch et al. 1993). Both equations are linked by retention curves $S_e(p_c)$ and relative permeabilities $k_r(S)$, such that

$$\begin{aligned} \nabla \cdot d^w \left[-\frac{\kappa k_r^w}{\eta^w} (\nabla p^w + d^w g \nabla D) \right] &= -\Phi \frac{\partial (d^w S_e^w)}{\partial t} + d^w w_0 \\ \nabla \cdot d^{co2} \left[-\frac{\kappa k_r^{co2}}{\eta^{co2}} (\nabla p^{co2} + d^{co2} g \nabla D) \right] &= -\Phi \frac{\partial (d^{co2} S_e^{co2})}{\partial t} + d^{co2} w_0 \end{aligned} \quad (1)$$

with d as density, κ as intrinsic permeability, η as viscosity, Φ as porosity and g as gravitational acceleration. The link between both phases has been established by experimental data or parameterization,

e.g., by van Genuchten (1980) or Mualem (1976) (with effective saturation S_e , capillary pressure height H_c and parameters α , n , m , L):

$$S_e^w = \frac{1}{(1 + |\alpha H_c|^n)^m} \quad S_e^{co2} = 1 - S_e^w \quad (2)$$

$$k_r^w = (S_e^w)^L \cdot \left[1 - \left(1 - (S_e^w)^{1/m} \right)^m \right]^2 \quad k_r^{co2} = (1 - S_e^w)^L \cdot \left(1 - (S_e^w)^{1/m} \right)^{2m} \quad (3)$$

The resulting distribution of water saturation $S(x, y, z, t)$ can be transformed into electrical formation resistivity ρ using empirical resistivity models, e.g., Archie's law (Archie 1942):

$$\rho = a_1 \cdot \Phi^{-a_2} \cdot S^{-a_3} \cdot \frac{1}{\sigma_w} \quad (4)$$

with σ_w denoting pore water conductivity. Archie's empirical parameters a_1 , a_2 and a_3 strongly depend on rock formation characteristics and can be combined with porosity and water saturation yielding the formation factor F , such that

$$\rho = F \cdot \rho_w. \quad (5)$$

For sands and sandstones with considerable clay content Waxman and Smits (1968) expanded Archie's law to account for both electrolytic conductance and interfacial conductance:

$$\rho = \frac{F^*}{S^n} \cdot \left(\sigma_w + \frac{B \cdot Q_v}{S} \right)^{-1} \quad (6)$$

with a unique formation factor F^* , the counterion mobility B describing the weak dependance of the interface conductivity on pore water salinity and the shalyness parameter Q_v being the cation exchange capacity normalized to the pore volume.

3 Numerical simulation studies

The process of CO₂ sequestration as well as the indication of a successful storage by observation of changes in electrical resistivity has been simulated following the steps indicated by the workflow shown in Fig. 1.

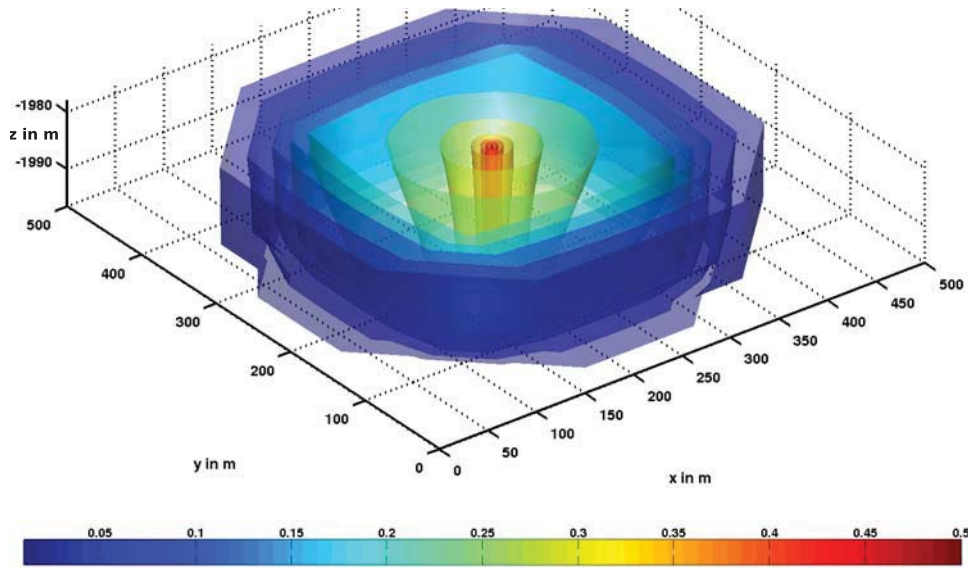


Figure 2: CO₂ saturation after 6 months of constant injection of 1000 m³ per day into a 25 m thick reservoir. The injection well is located at the geometric center. These results were obtained using the FD simulation code Mod2PhaseThermo.

We have incorporated data that is generally available for a CCS site such as the formation geometry, reservoir parameters, e.g. porosity, intrinsic permeability, and injection rates to simulate the injection process. The main result of the numerical simulation studies is an estimate of the temporal changes in the spatial distribution of brine saturation $S(x, y, z, t)$ within a finite computational domain.

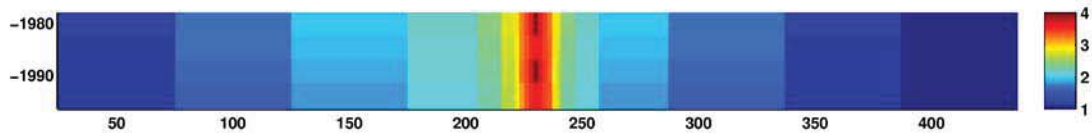


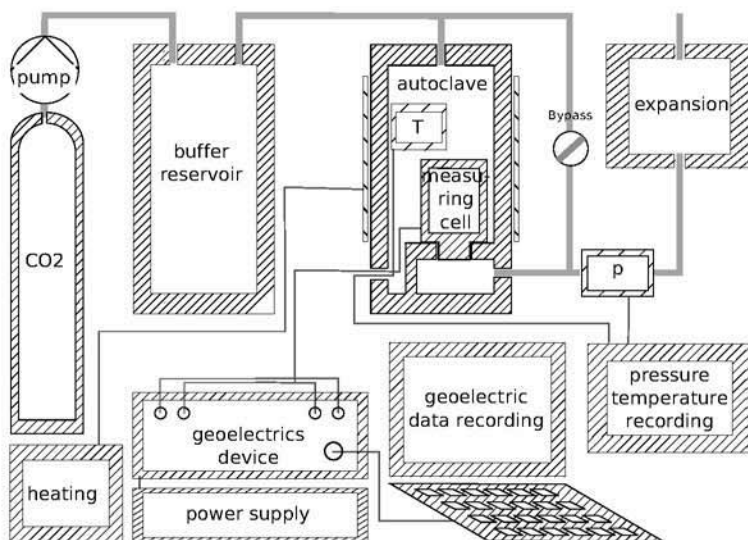
Figure 3: Vertical cross-section through the aquifer indicating the increase in formation resistivity derived from the saturation illustrated in Fig. 2. The transformation was done using eq. (4) and a background resistivity of $25 \Omega\text{m}$ has been considered.

These saturation estimates have later been transformed into local electrical resistivities by means of simple empirical resistivity models. The resulting distribution of local rock resistivity can be used as an input for a numerical simulation of electromagnetic monitoring measurements.

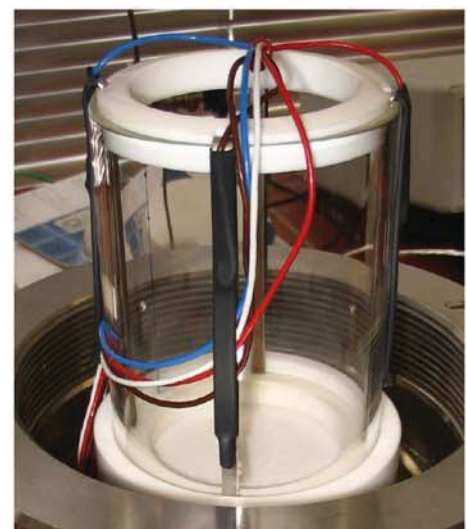
Numerical simulation studies of the injection process have been carried out using the Finite Difference (FD) code Mod2PhaseThermo (Häfner and Boy 2009) as well as the Finite Elements (FE) package COMSOL Multiphysics. Using COMSOL, we were able to model the CO_2 sequestration by a two-phase flow according to eqs. (1)-(3). Under the assumption of negligible compressibility, pressure or temperature induced density and viscosity variations, and chemical interactions between rock matrix, pore water and CO_2 , we have observed a maximum CO_2 saturation of 80 % in the vicinity of the injection well.

The FD code Mod2PhaseThermo simulates multiphase-flow taking into account its thermodynamics and the dissolution of CO_2 in the pore water. The result of such a simulation is shown in Fig. 2. CO_2 saturations of up to 50 % near the injection well can be observed.

Formation resistivities were derived from the observed saturation estimates according to eq. (4) and (6). An increase in resistivity by a factor of 2 to 10 obtained by the COMSOL model and 2 to 4.5 obtained by Mod2PhaseThermo could be predicted, respectively (Fig. 3). In addition to the observed lateral saturation changes, different results have been obtained depending on flow, resistivity model and reservoir characteristics.



(a) Block diagram of the laboratory experiment.



(b) Measuring cell prior to being filled with water saturated sand.

Figure 4: Experimental set-up for the simulation of CO_2 sequestration on a laboratory scale.

4 Laboratory experiments

To validate the results obtained by the numerical studies we have carried out petrophysical experiments on a laboratory scale. For this purpose, an experimental setup has been developed and implemented to carry out a laboratory simulation of the sequestration process (Fig. 4, left). Our intention was to investigate the behavior of supercritical CO_2 under in-situ conditions. Furthermore, we were interested in the validity of the empirical parameters used in eq. (4).

The central element of the apparatus schematically outlined in (Fig. 4) is a measuring cell (Fig. 4, right) inserted into an autoclave allowing for monitoring the apparent electrical resistivity of a sand sample at in-situ pressures and temperatures. In the laboratory assembly, CO_2 was forced to flow through the initially water-saturated sand sample, gradually replacing the pore water. During this process, the electrical resistance of the cell has been recorded with a sampling interval of 10 s. Pressures up to 130 bar were applied at temperatures up to 40°C .

From a practical point of view it might be argued that sand has different petrophysical properties than compact sandstone. However, since sands and sandstones are similar in terms of, e.g., mineral content, grain distribution, and pore space geometry, their electrical properties underlie the same physical mechanisms and can therefore be described by the same petrophysical models (Archie 1942, Schön 1996). Hence, despite substituting sandstone with sand we still expect sufficiently accurate and representative results.

5 Results

A first experiment could prove that pure CO_2 does not show any relevant electrical conductivity at pressures up to 130 bar. During further experiments, CO_2 was lead through an initially water-saturated sand sample resulting in an increase of the electrical resistivity of the sample by a factor of 27 to 33 at pressures up to 50 bar. The remaining electrical conductivity was dominated by a residual water content of 14 to 18 % (Fig. 5).

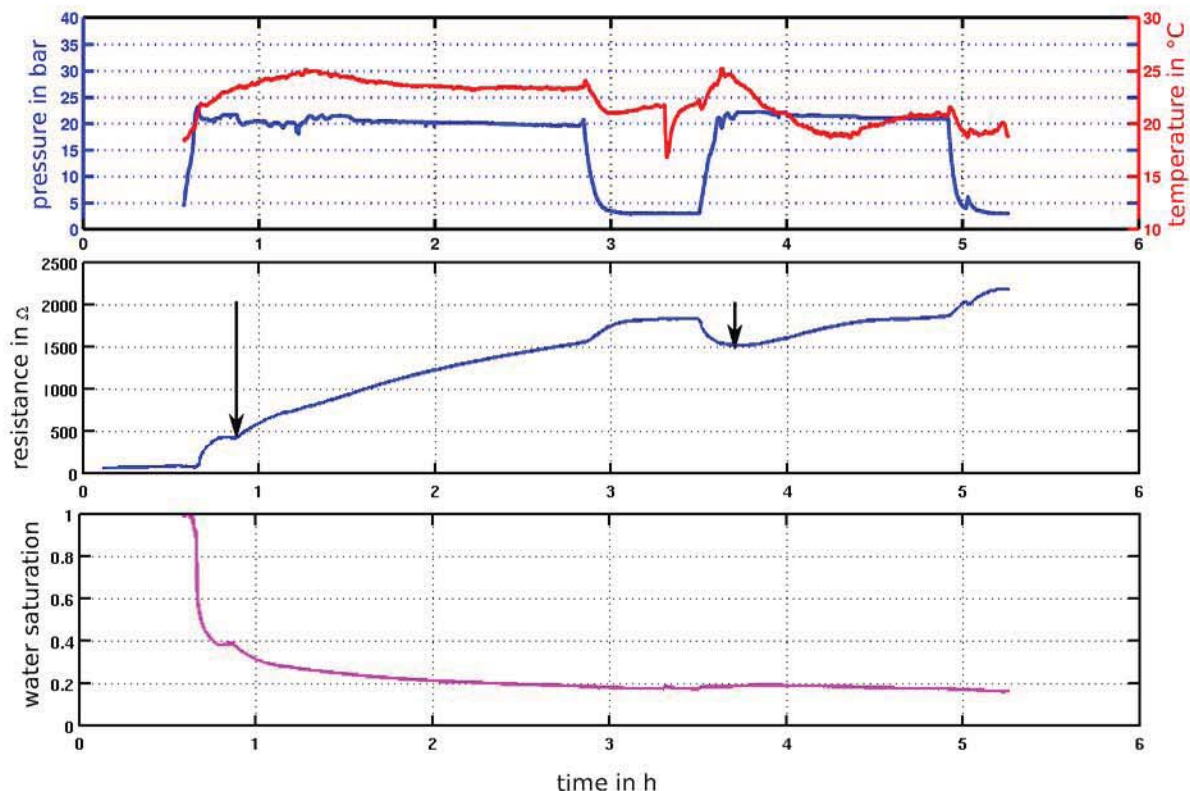


Figure 5: Electrical resistance and water saturation derived according to equation (4) of a sand probe obtained by a flow experiment carried out at a projected maximum pressure of 20 bar and temperature of 25°C . The arrows indicate start of CO_2 injection.

Final experiments have demonstrated that an increase in resistivity may also be achieved under supercritical conditions. However, the sample was not homogeneously infiltrated by the CO₂ and preferential flow paths were built up during the experiment. The pore water has been pressed out only partially, leaving behind tube-shaped flow channels and a very heterogeneous water distribution with average residual water contents of more than 70 %.

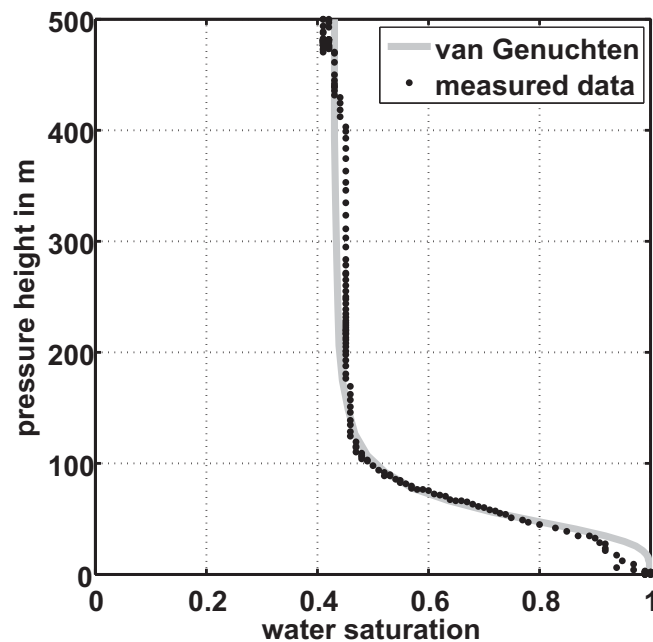


Figure 6: Relation between CO₂ pressure and water saturation derived from data obtained by the initial pressure build-up for an experiment with a projected maximum pressure of 50 bar. The reference data have been calculated according to van Genuchten (1980) with $\alpha = 0.02$ and $n = 3.9$ (cf. eqs. (2)-(3)).

The complications are due to the density of supercritical CO₂ which is large compared to the density of CO₂ in its gaseous state. Consequently, the current experimental assembly causes the high density to result in low flow velocities within the measuring cell. In addition to the flow channels, this effect prevents an effective replacement of the pore water by supercritical CO₂.

The experiments have shown that theoretical pressure-saturation relations can generally be verified in practice using our set-up (Fig. 6). All experimental data sufficiently agree with Archie's law. However, we have to carry out further investigations on the effects of CO₂ dissolving in the pore water. There are indications that this has an important impact on the pore water resistivity depending on pressure, temperature, and brine salinity.

6 Conclusions

Numerical simulation studies and laboratory experiments show that the electrical resistivity of a fluid-saturated porous medium is highly sensitive to the presence of CO₂. Geo-electromagnetic methods are therefore considered as a promising approach for monitoring CO₂ storage.

Still, the laboratory set-up has to be improved further to provide reliable results at high pressures. Subject to these prerequisites, well-founded simulations of electromagnetic monitoring scenarios can be carried out.

Further laboratory experiments will also aim at quantifying the influence of dissolved CO₂ on pore water and formation resistivity. A feasibility study could show whether electromagnetic methods are able to monitor, e.g., an expanding plume of CO₂-rich formation water during CO₂ injection. Finally, the applicability of the laboratory assembly for measuring reliable retention curves of unconsolidated sedimentary rocks will be further tested in the near future.

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Erkundung eines Aquifers unter dem Mittelmeer vor der israelischen Küste mit LOTEM.

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Abstract

Im Rahmen dieses BMBF-geförderten Projektes ² kommt die Long-Offset Transient Elektromagnetik (LOTEM) Methode zum ersten Mal in mariner Umgebung zur Erkundung von Grundwasseraquifern zum Einsatz. Hauptziel des Projektes ist die Detektion der Süßwassergrenze unter dem Mittelmeer an der Küste Israels. Zu diesem Zweck wurden Hardware-Modifikationen durchgeführt und bereits in einer ersten Testmessung in Israel erprobt. Des weiteren werden die Interpretationen der ersten Testmessung sowie die Modellierungsergebnisse des Küstenabschnitts vorgestellt, welche zur Planung der Hauptmessung wichtig sind.

Einleitung

Die Bedeutung von Offshore-Süßwasserleitern für das Grundwassermanagement nahm in den letzten Jahren stark zu. Das Vorhandensein dieser Aquifere, die sich bis zu mehreren Kilometern unter dem Meeresboden erstrecken können, wurde in der Fachliteratur beschrieben: z.B. an der Küste von Guyana [Arad, 1983] oder vor den Niederlanden [Groen et al., 2005]. Der israelische Küstenaquifer ist eine der Hauptgrundwasserressourcen des Landes (vgl. Fig. 1). Aufgrund der intensiven Nutzung verschlechtert sich allerdings die Wasserqualität zunehmend. Die Gründe hierfür liegen sowohl in der anthropogenen Verschmutzung vom Land aus, als auch im Eindringen von Salzwasser.

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Figure 1: Die großen Aquifere der Region.

Geologie und Fragestellung

Der Küstenaquifer besteht hauptsächlich aus kalkhaltigem Sandstein und Sanden und ist selbst noch in vier Sub-Aquifere unterteilt (vgl. Fig.2). Die oberen zwei grundwasserführenden Schichten werden vom Land aus anthropogen und vom Meer aus durch Salzwassereindringen verunreinigt. Bisher wurde davon ausgegangen, daß die unteren Sub-Aquifere stellenweise vom Meerwasser getrennt sind und dort keine Salzwasserintrusion stattfindet. Neuere Untersuchungen ([Kafri, U., Goldman, M., 2006],[Yechieli et. al, 2009])

hingegen deuten jedoch auf das Fortsetzen des Aquifers unter dem Meer hin. So zeigen z.B. TDEM-Messungen auf Land, nahe der Küste, einen schlechteren elektrischen Leiter 10 (Ωm) eingebettet zwischen zwei guten elektrischen Leitern ($\sim 2 \Omega m$). Diese Schicht, in der Tiefe der unteren Sub-Aquifere, wird als Grundwasserführenden Schicht interpretiert (vgl. Fig.3). Die Fragestellung an die LOTEM-Methode ist nun 1. die Prüfung der Existenz des unteren Sub-Aquifers in etwa 100m Tiefe und 2. dessen Ausbreitung unter dem Mittelmeer.

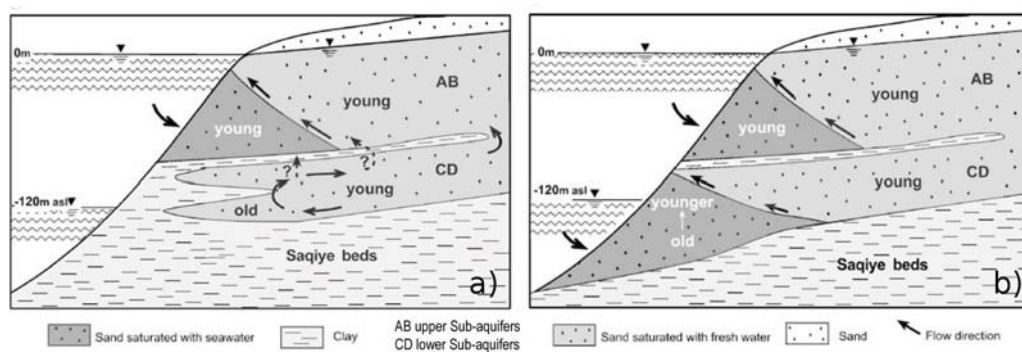


Figure 2: Möglicher Verlauf des unteren Subaquifers an der Küste [Yechieli et. al, 2009]: a) Der untere Subaquifer (CD) hat keine Verbindung zum Meer, b): Das Salzwasser dringt in den unteren Aquifer ein.

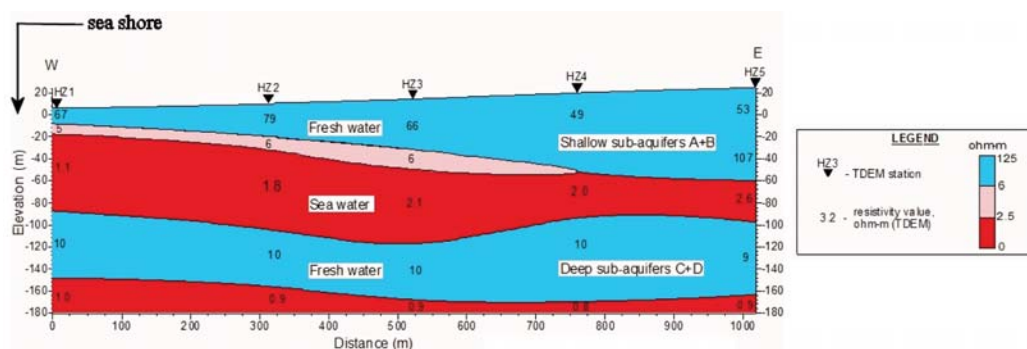


Figure 3: Onshore SHOTEM-Ergebnisse [Kafri, U., Goldman, M., 2006].

Hardwaremodifikationen

Das Kölner LOTEM-Equipement ist für Landmessungen konzipiert. Deswegen waren einige Hardware-Modifikationen für die Off-shoremessung nötig. Als Sender konnte aus logistischen Gründen nur der kleinere Kölner Sender ³ verwendet werden. Da dieser bei verwendeten Stromstärken von $\sim 9A$ zuviel Wärme entwickelt, war es nötig eine zusätzliche Wasserkühlung einzubauen. Für die Einspeisung des Sendesignals wurden neue Elektroden entwickelt (vgl. Fig.4), deren Form, durch Maximierung des angekoppelten Wasservolumens, elektrochemische Vorgänge minimiert. Zusätzlich kann das Sendesignal Stromgeregelt werden, um einen konstanten Stromverlauf zu erreichen. In Fig.5 sind Sendesignale gezeigt. Links wird mit konstanter Spannung gesendet. Bei sich verändernder Ankopplung ergibt sich kein konstanter Stromverlauf. Deswegen wurde der Sendestrom nachträglich auf einen konstan-

ten Wert heruntergeregelt (Fig.5 rechts).

Die verwendeten Magnetfeldsensoren ⁴ wurden in Druck- und Salzwasserbeständige Behälter verpackt (Fig.6).

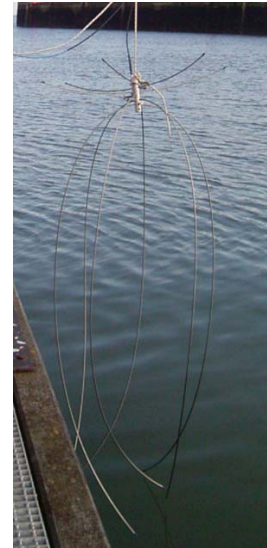


Figure 4: *Neuentwickelte Elektroden für die Einspeisung des Sendesignals.*

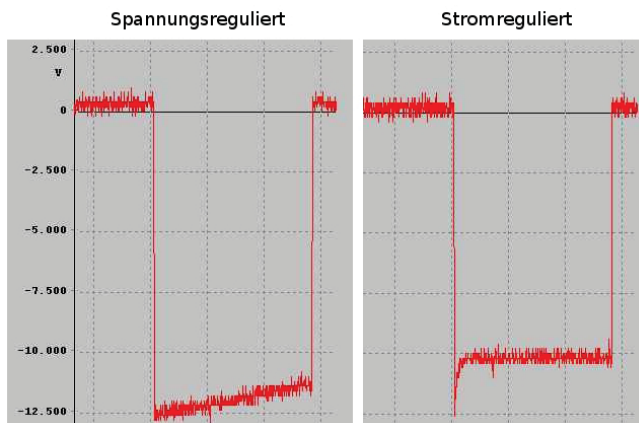


Figure 5: *Aufgezeichnete Sendesignale bei einem Test in Wilhelmshaven.*



Figure 6: *Seewasserfestes Gehäuse für die Magnetfeld-Sensoren.*

³NT-20 von Zonge Engineering & Research Organization, Inc., Tucson AZ, USA

⁴TEM/3-Spulen von Zonge Engineering & Research Organization, Inc., Tucson AZ, USA

Ergebnisse Messung 2008

Anfang November 2008 wurden die ersten Messung vor Ort durchgeführt. Hierbei wurden verschiedene Messkonfigurationen getestet:

- BroadSide: Sender an der Küstenlinie, Empfänger an Land. Offset 200m. vgl. Fig.7: Tx2 und Rx3.
- BroadSide: Sender an der Küstenlinie, Empfänger im Wasser. Offset 300m. vgl. Fig.7: Tx2 und Rx2.
- BroadSide: Sender im Wasser, Empfänger an Land. Offset 500m. vgl. Fig.7: Tx1 und Rx1.

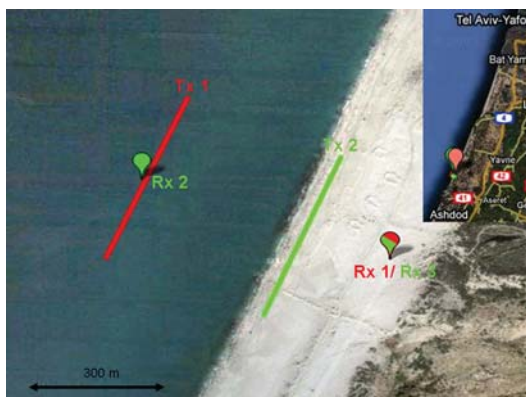


Figure 7: *Setups der Messung 2008.*

Das Setup “Sender an der Küstenlinie, Empfänger an Land” ist das einzige, welches annnehmbar 1D interpretiert werden konnte. Als Beispiel ist hier die dH_z/dt - Komponente gezeigt. Das Endmodell einer Marquardt-Inversion ist in Fig.9, die zugehörige Datenanpassung in Fig.8 zu sehen. Das Startmodell wurde anhand der Vorinformationen und dem Ergebnis einer Occam-Inversion gewählt ⁵.

⁵Vgl. M.v.Papen, B.Tezkan: “On the analysis of LOTEM time series from Israel and preliminary 1D inversion of data.”, Poster EMTF 2009.

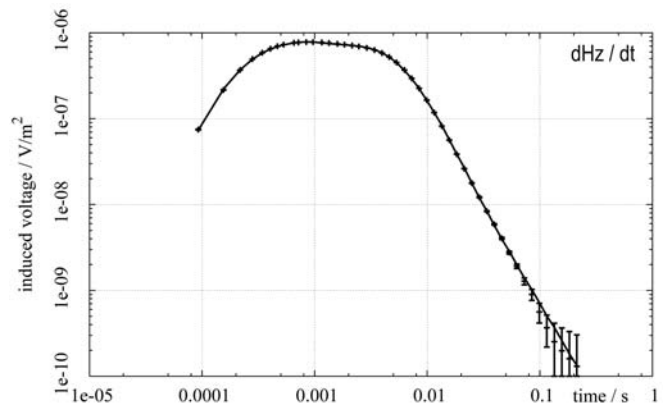


Figure 8: *Datenanpassung des “besten” Modells (rot) in Fig. 9.*

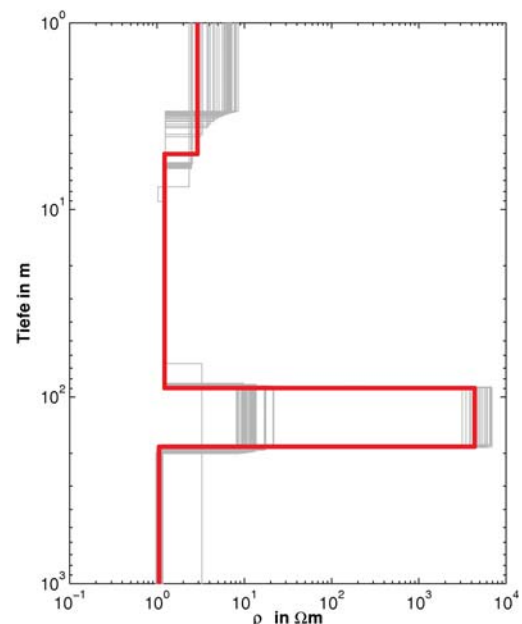


Figure 9: *1D-Inversionsergebnis mit Äquivalenzmodellen.*

Das Ergebnis ist gut mit den SHOTEM-Ergebnissen (Fig. 3) vergleichbar. Der gesuchte Aquifer ist deutlich in einer Tiefe ab ca. 90m zu erkennen. Sowohl die Schichtdicke mit ca. 100m als auch viele Äquivalenzmodelle mit einem Widerstand von $\sim 10 \Omega m$ sind vergleichbar und glaubwürdig.

Die anderen beiden Messkonfigurationen

können nicht als eindimensionales Problem angesehen und somit auch nicht 1D ausgewertet werden. Die gemessenen Transienten wurden mit synthetischen Daten verglichen. Das 2D Leitfähigkeitsmodell ist in Fig. 11 dargestellt. Mit diesem synthetischen Modell war eine qualitative Anpassung möglich. Exemplarisch ist in Fig.10 die Datenanpassung für die Ex-Komponente gezeigt.

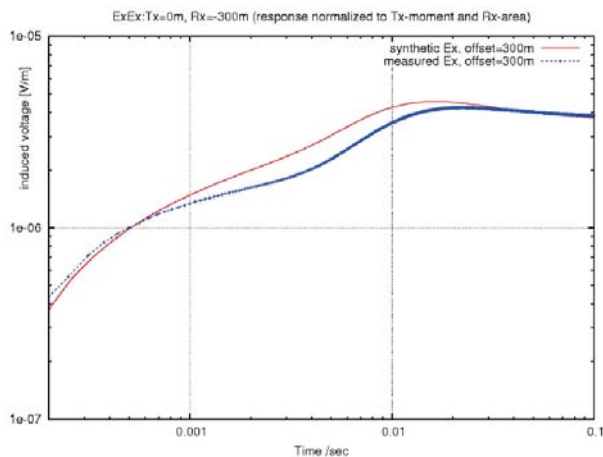


Figure 10: *Setup Tx2 - Rx3 (vgl.Fig.7): Anpassung der Messdaten der Ex-Komponente mit modellierten Daten.*

Vorab-Modellierungen

Für die Messung im November 2009 sind die Empfängerlokationen von Bedeutung. Eine Frage hierbei ist u.a. ob das Target auch mit einem Sender auf dem Meer und Empfängern auf dem Land auflösbar ist. Diese Konfiguration ist logistisch relativ einfach durchführbar. Es wurde also ein 2D-Küstenmodell erstellt, mit welchem synthetische Daten mittels SLDMEM3T [Druskin, Knizhnermann, 1988] erzeugt wurde. Als Datenbeispiel sind in Fig.12

die Ex-Komponenten dargestellt. Hierfür werden im Modell (Fig.11) Daten mit Targetlayer und einmal ohne Targetlayer unter dem Meeresboden verglichen. “Ohne Targetlayer” bedeutet in diesem Fall, daß der Targetlayer 300m im Landesinneren endet (nicht dargestellt). Die darüberliegende Schicht ($1.2 \Omega m$, gelb) wurde nach unten erweitert.

Modelliert wurden verschiedene Senderpositionen in unterschiedlichen Entfernungen zur Küstenlinie. Im Beispiel beträgt die Entfernung 1500m. Betrachtet werden nun 2 Empfängerpositionen: einmal 275m von der Küstenlinie entfernt auf dem Meeresboden und einmal 95m im Landesinneren. Modelliert wurde die BroadSide Konfiguration (Senderichtung und Profilrichtung senkrecht zueinander) und die InLine Konfiguration (Senderrichtung und Profilrichtung auf einer Linie). Die System-Antwort wurde für diese Modellierungen nicht berücksichtigt.

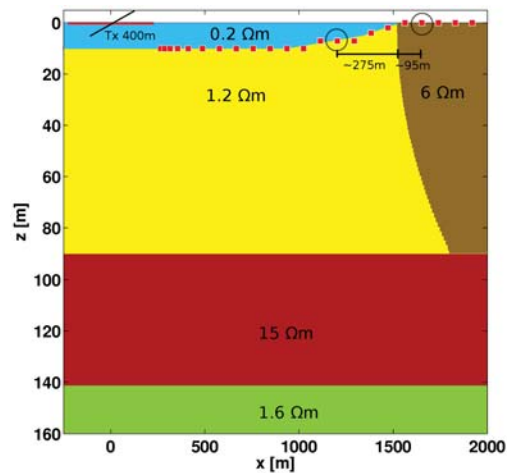


Figure 11: *Verwendetes Küstenmodell. Das gesuchte Target ist die $15 \Omega m$ -Schicht (rot). Blau = Wasser.*

Sowohl bei der InLine als auch bei der BroadSide Konfiguration ist in dieser Komponente der Targetlayer klar auflösbar: die Transienten unterscheiden sich deutlich in den Amplituden von den Transienten ohne Targetlayer. Bei der BroadSide Konfiguration verschiebt sich der auftretende Vorzeichenwechsel durch das Target zu späten Zeiten. Ein Vorzeichenwechsel ist bei der Auswertung immer schwer interpretierbar.

Aufgrund der Unterscheidbarkeit der Transienten sind Empfänger auf dem Land für diese Fragestellung sinnvoll.

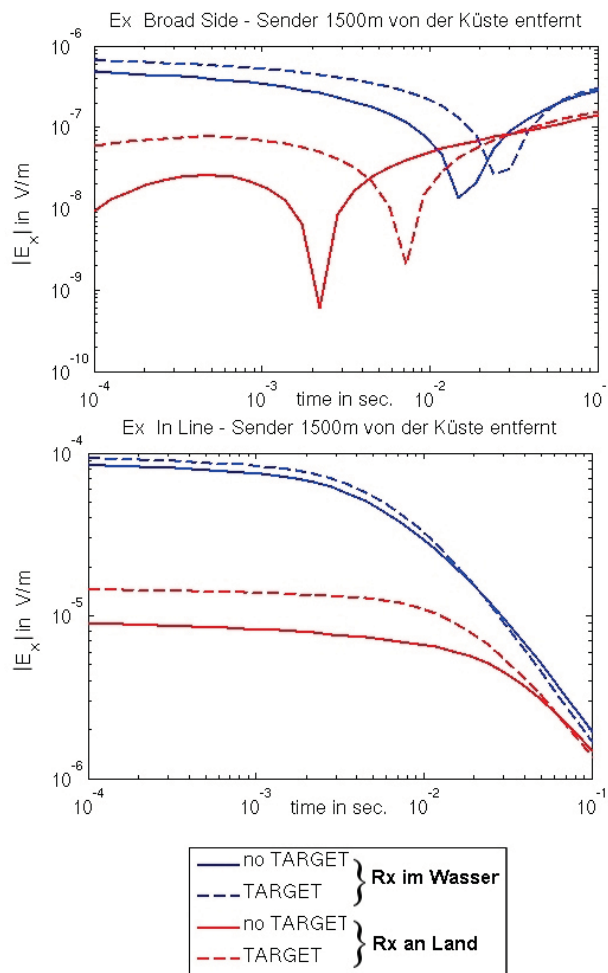


Figure 12: Modellierte Transienten der Ex-Komponente an den Positionen, die in Fig.11 als Kreise markiert sind.

Zusammenfassung und Ausblick

Aufgrund der Ergebnisse und der Erfahrungen der ersten Messung (2008) und der Modellierungen sind die Setups für die erste Hauptmessung im November 2009 festgelegt worden. Die Messung ist inzwischen durchgeführt worden. Es wurden die Konfigurationen InLine und BroadSide gemessen. Der Sender befand sich dabei immer auf dem Meer in verschiedenen Abständen zur Küste. Pro Senderposition gab es mehrere Empfängerpositionen, sowohl auf dem Wasser als auch auf Land.

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On the analysis of LOTEM time series from Israel and the preliminary 1D inversion of data

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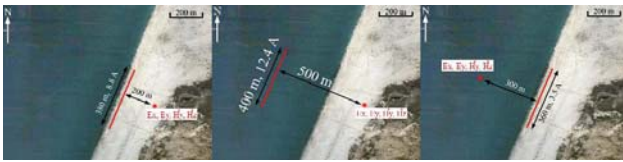


Abbildung 1: Overview of measurement setups (A,B,C from left to right)

1. Data Processing of TEM time series from the first Israel survey

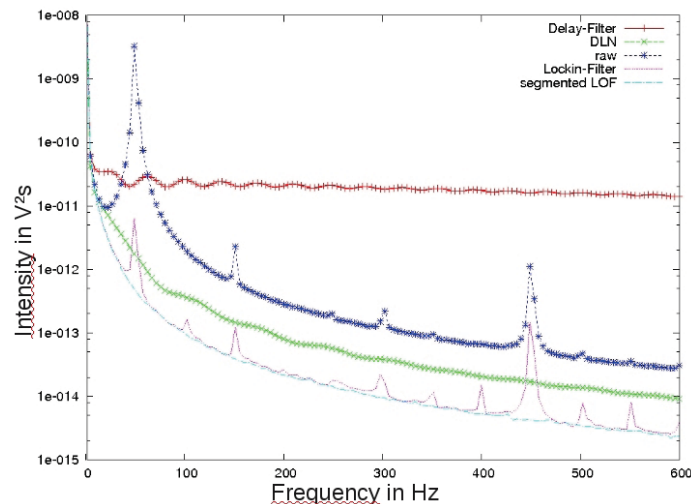


Abbildung 2: Power spectrum of Ex at station A/B before (blue asterisks) and after application of different filters

The measurement took place at the coast near Ashdod, Israel (fig.1), where the task is the detection of fresh groundwater bodies within the mediterranean submarine aquifers [3]. For that matter a 380 m long dipole emitting in a 50% duty cycle was used, located on the shoreline for stations A and C and at -300m offshore on the seabottom for station B. Receiver stations recorded four components of the EM field ($E_x, E_y, \dot{H}_y, \dot{H}_z$) with Summit receivers. Receivers at station C were set up 5 m beneath the water on the sea bottom.

- The measured transients consist of 4096 data points

(DP) with a sample rate of 1/16 ms and 256 DP as onset. Switch time of Tx is $t_0 = 1500ms$.

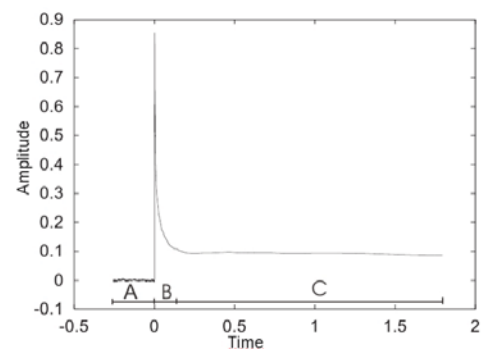


Abbildung 3: Parts of transient [5]

Noise measurements showed a dominant 50 Hz noise together with multiples (fig.2). The average noise power for each component is: $P_{N,E_x} = 0.2V^2$, $P_{N,E_y} = 0.02V^2$, $P_{N,\dot{H}_y} = 4.8V^2$, $P_{N,\dot{H}_z} = 0.2V^2$ (measured on land). Due to relative short offsets of 200-500 m between receiver and transmitter the signal is - except for E_y component - clearly visible in the raw transients.

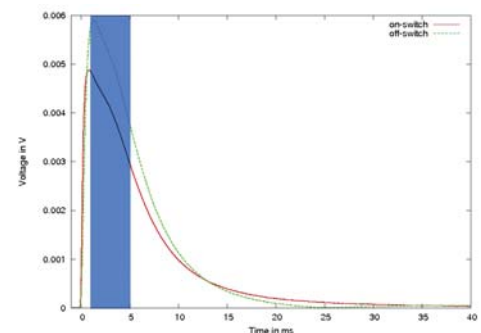


Abbildung 4: Stacked transients of only on-switches and only off-switches

2. Application of a segmented lockin filter

First step of the processing was to filter out the 50 Hz noise. This was done using a segmented lockin filter (LOF)

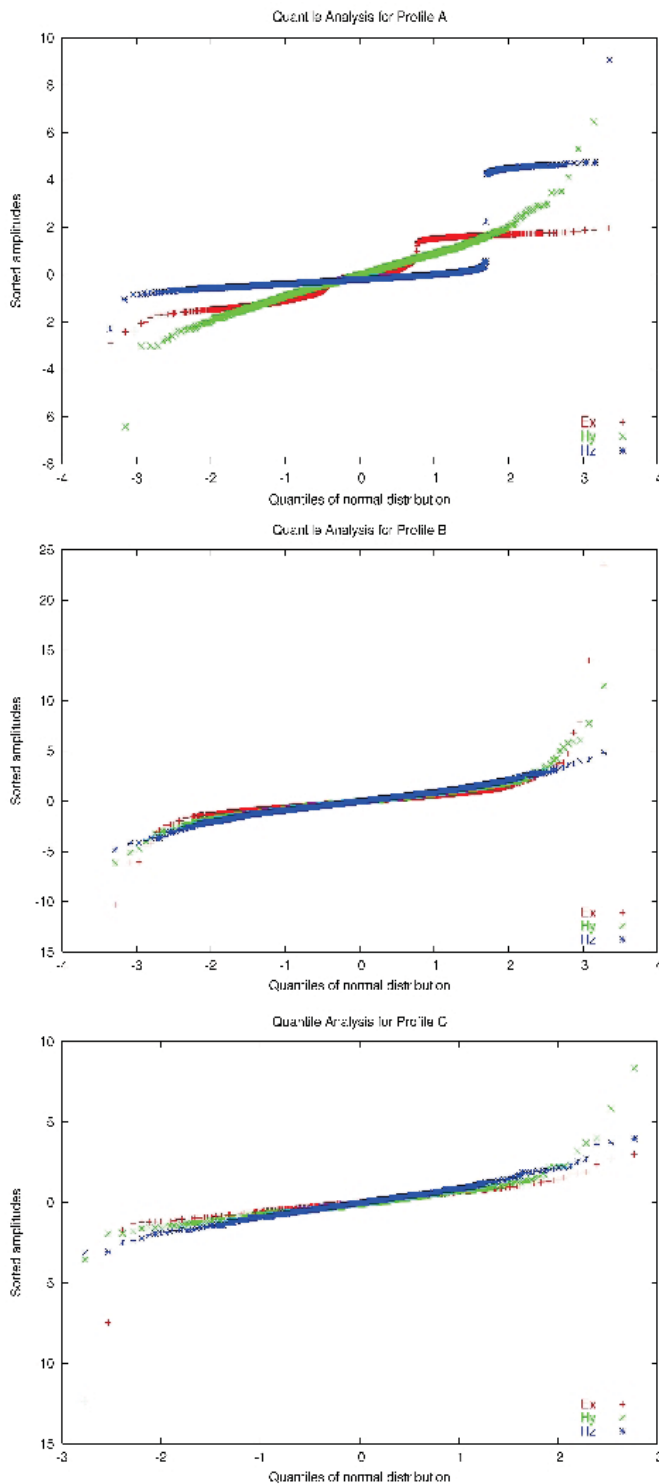


Abbildung 5: Quantiles of normal distribution for stations A,B,C from top to bottom

[1]. It fits harmonics in segments of 60 ms to the data and subtracts them. Amplitude and phase are computed with Fourier series expansion. Regarding the high dynamic of the transient at the time of the switch (e.g. switch on in part B of fig.3) the fitting is being done in parts A and C only [5]. In addition a 3rd degree polynomial is fitted in part C to simulate the descent of the transient. Other filters like Delay Lower Nyquist frequency (DLN), Delay filter or (not segmented) Lockin-Filter [1, 5] remove the noise less efficient for this dataset (fig.2).

- Application of the filter reduced the noise power, computed at the end of the time series, to $P_{N,Ex} = 0.2nV^2$, $P_{N,Hy} = 0.2V^2$ and $P_{N,Hx} = 1.4nV^2$.
- The DC bias is removed by levelling the data in the onset after filtering.

Due to the short onset and the resulting error in determining the DC bias the inverted conductivities will exhibit a slight shift [4]. Higher accuracy can be achieved by averaging over a complete transmitter period.

3. Analysis of data distribution

The data measured with 50% duty cycle consists of two different sets of transients. One set is created when switching the current off and another when switching on. Because the given transmitter system shows a much shorter ramp time for off-switches, these transients are generally less broad and have a higher peak than those created by switch-on (fig.4). Cluster analysis [2] in the blue area of fig.4 resulted in two equally populated groups identified as on and off switches by cross checking the Ex component.

The distribution of the data in terms of probability can be visualized with quantiles of gauss distribution (fig.5). The mean values of all transients in a certain time segment are sorted by ascending value and are then plotted against an axis, which is scaled in such a way that normal distributed data will show as a line of slope 1. This has been done for each component over a 62 ms window at the end of each switch-off time series, which have been filtered and levelled beforehand.

The top graph shows that the distribution of Hz (and - to a smaller extent - Ex) at station A is dominated by a step. However, cluster analysis could not sort out these time series. The other stations show a much better result with normal distributed data for all components.

4. Stacking and 1D Occam inversion

After the data is analysed and if it needs no further editing (like correction of the right switch moment) it is ready to be stacked and smoothed.

- The data was stacked selectively omitting the highest and lowest 5% of the values per DP.

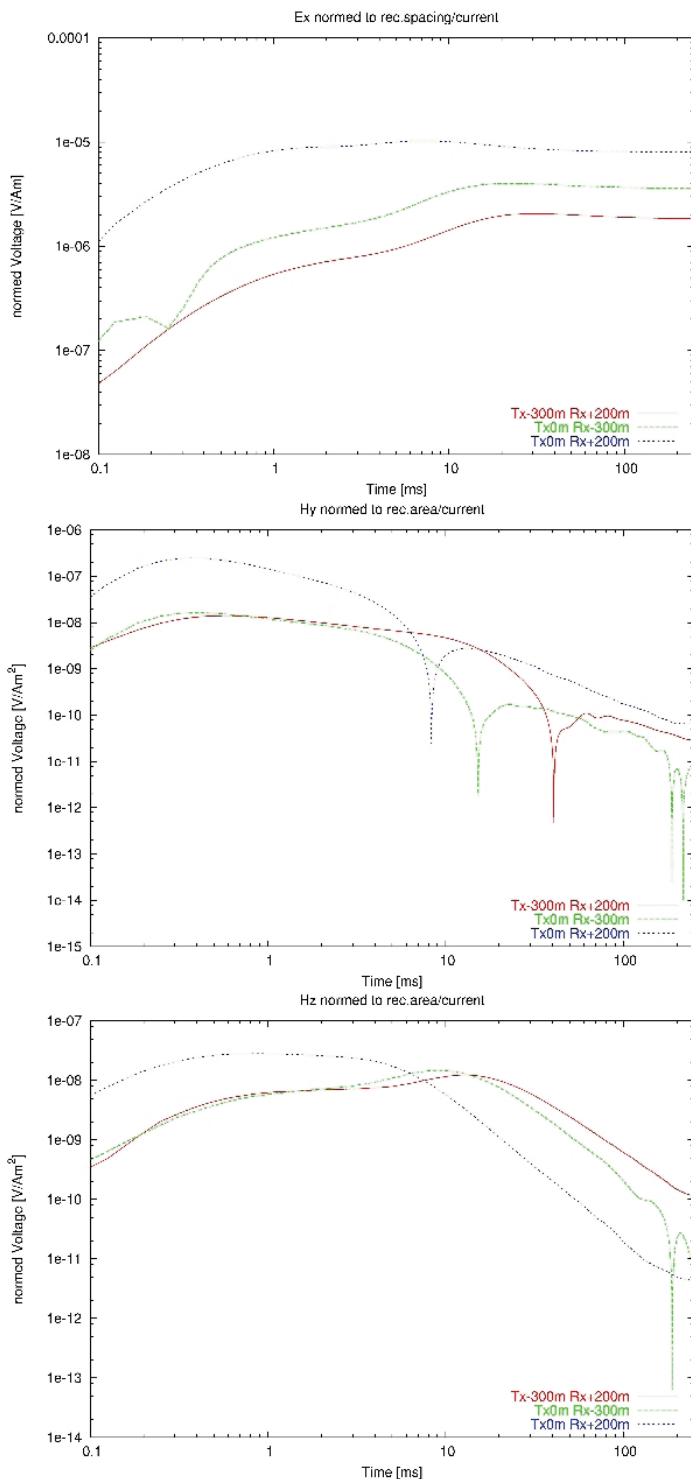


Abbildung 6: Normalized stacked transients sorted by component

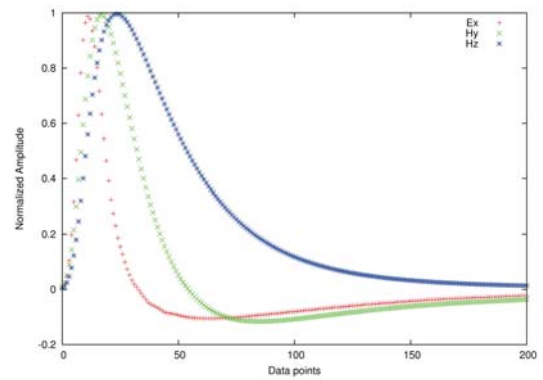


Abbildung 7: Normalized system responses measured broadside

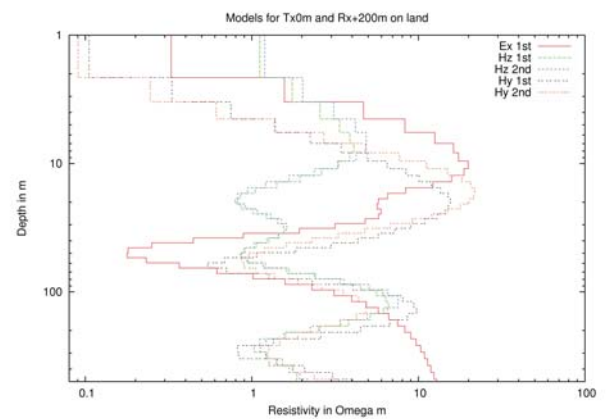


Abbildung 8: Result of Occam inversion of data at station A for 1st and 2nd order roughness

- Afterwards it was smoothed with a dynamic Hanning window function.

Fig.6 shows the normalized transients of the stations sorted by components. Field setup is according to fig. 1. Ex shows a noticeable “bump” in early times at receiver station C, which could be a 2D effect of the water-land interface. More on interpretation of this data can be found in *Lippert et al., 2010*.

System responses measured on land are shown in fig.7 and are used for convolution in forward calculation. Fig.8 shows the Occam inversion results for station A. All components show a good conductor at about 80 m depth between two more resistive layers at approximately 10 m and 150 m depth, which represent the sought fresh water bodies. Hz could resolve a second good conductor, which is not visible in Hy and only indicated in Ex. These results are in good agreement with the SHOTEM measurements of this area.

5. Future research

Some system responses measured broadside include negative values. This is supposed to be a geometric phenomenon related to the 'air wave'. Future measurements of the system response will therefore be conducted in the laboratory.

For the geometries of stations B and C there are at the moment no inversion codes available, therefore they will be interpreted using 2D forward modelling. Joint inversions using different LOTEM components will be realized.

6. References

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Grundwasserkontamination bei Roorkee/Indien: 2D Joint Inversion von Radiomagnetotellurik und Gleichstromgeo- elektrik Daten

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In der landwirtschaftlich geprägten Region um Roorkee in Nordindien ist der Einsatz von Düngemittel sowie die Bewässerung der Felder mit Abwasser gängige Praxis, wobei das Abwasser durch offene, teilweise unbefestigte Kanäle transportiert wird. Die umliegende Landbevölkerung bezieht ihr Frischwasser oftmals aus Brunnen, die aus einem oberflächennahen Aquifer gespeist werden. Eine Gefährdung dieser oberflächennahen Aquifersysteme durch den praktizierten Umgang mit Abwasser und Düngemittel ist nicht auszuschließen.

Im Rahmen der hier durchgeführten Untersuchungen wurden ein landwirtschaftlich genutztes und stark mit Abwasser bewässertes Gebiet nahe einer Mülldeponie und ein als kontaminationsfrei angenommenes Referenzprofil mit den Methoden der Gleichstromgeoelektrik (DC) und der Radiomagnetotellurik (RMT) vermessen.

Die einzelnen Profile wurden neben Benutzung von gängiger 1D und 2D Auswertesoftware mittels eines neu entwickelten 2D Joint Inversionsprogrammes invertiert und interpretiert.

Verglichen mit den Ergebnissen des Referenzprofils zeigen die Ergebnisse auf dem kontaminierten Gebiet nahe der Deponie deutlich erhöhte Leitfähigkeitswerte bis zu einer Tiefe von ca. 15 m. Der Tiefenbereich von 5 m bis ungefähr 15 m konnte mit dem oberflächennahen Aquifer identifiziert werden. Darüber hinaus ergibt sich eine Abnahme der Leitfähigkeitswerte mit dem Abstand von der Mülldeponie und den Abwasserauslässen.

Durch die Anwendung einer 2D Joint Inversion konnten die Datensätze beider Methoden durch einheitliche Untergrundmodelle angepasst werden.

Einleitung

Roorkee befindet sich in den Vorläufern des Himalaja am rechten Ufer des Solani Flusses im Distrikt Haridwar des nordindischen Staates Uttharakand (Geographische Breite: 29°50' bis 29°56' N, Geographische Länge: 77°48' bis 77°56' E). Der Solani ist ein Nebenfluss des Ganges. Der Ganges ist in dem Gebiet kanalisiert und teilt Roorkee in „Old“- und „New“-Roorkee. Die Stadt hat ca. 100000 Einwohner, mit einer jedoch wachsenden Bevölkerung. Die Region ist stark landwirtschaftlich geprägt, wobei sich dort vermehrt Industrie entwickelt.

Die Untersuchungen fanden in Zusammenarbeit mit dem Indian Institute of Technology in Roorkee/Nordindien (IIT-Roorkee) im Rahmen eines deutsch-indischen Partnerprojektes statt. Das Hauptanliegen dieses Projektes besteht in der Anwendung geophysikalischer Methoden zur Abschätzung der Gefährdung der Aquifersysteme der Region um Roorkee.

In der landwirtschaftlich geprägten Region um Roorkee ist der Einsatz von Düngemittel sowie die Bewässerung der Felder mit Abwasser gängige Praxis, wobei das Abwasser durch offene, teilweise nicht befestigte, Kanäle transportiert wird. Gleichzeitig bezieht die umliegende Landbevölkerung ihr Frischwasser oftmals aus Brunnen, die aus oberflächennahen Aquiferen gespeist werden. Eine Gefährdung dieser oberflächennahen Aquifersysteme durch den praktizierten Umgang mit Abwasser und Düngemittel ist nicht auszuschließen.

Ein weiteres Risiko geht von Mülldeponien aus, die

zu Grundwasserkontamination in deren unmittelbaren Umgebung führen können und deren schädlicher Einfluss auf die Aquifersysteme der Region ebenfalls nicht auszuschließen ist. Die Grundwasserproblematik wird, durch den Frischwasserbedarf der zunehmenden Bevölkerung und der wachsenden Industrie, verschärft.

Geophysikalische Erkundungsmethoden haben sich in ihrer Anwendung auf hydrogeologische Fragestellungen, sowie zur Untersuchung von Altlasten mehrfach bewährt ([Tezkan, 1999], [Recher, 2002] und [Seher, 2005]). Insbesondere eignet sich die Kombination von DC und RMT, da sich beide Methoden in ihrem Erkundungstiefenbereich und ihrer Sensitivität gegenüber leitfähigen Strukturen ergänzen.

Im Rahmen des Projektes wurde das umliegende Gebiet einer Mülldeponie nahe dem Dorf Saliyar mit diesen Methoden profilweise vermessen. Seit 1975 wird dieses Gebiet zur landwirtschaftlichen Nutzung an Kleinbauern verpachtet und intensiv mit Abwasser bewässert, welches von Roorkee durch Rohrleitungen auf die Felder transportiert wird.

In Abbildung 1 ist eine Übersichtskarte der Region dargestellt. Das Messgebiet bei Saliyar befindet sich ca. 6 km nordwestlich von Roorkee auf der rechten Uferseite des Solani. Das ungeklärte Abwasser der Stadt wird von der Mahingram Pumpstation über eine Kanalleitung zur Mülldeponie geleitet. Das vermessene Referenzgebiet befindet sich auf der linken Uferseite des Solani und ist ca 10 km von Saliyar entfernt. Das ganze Gelände hat ein leichtes Gefälle in Richtung Südosten.

Der Einfluss von Abwassernutzung auf die Grund-

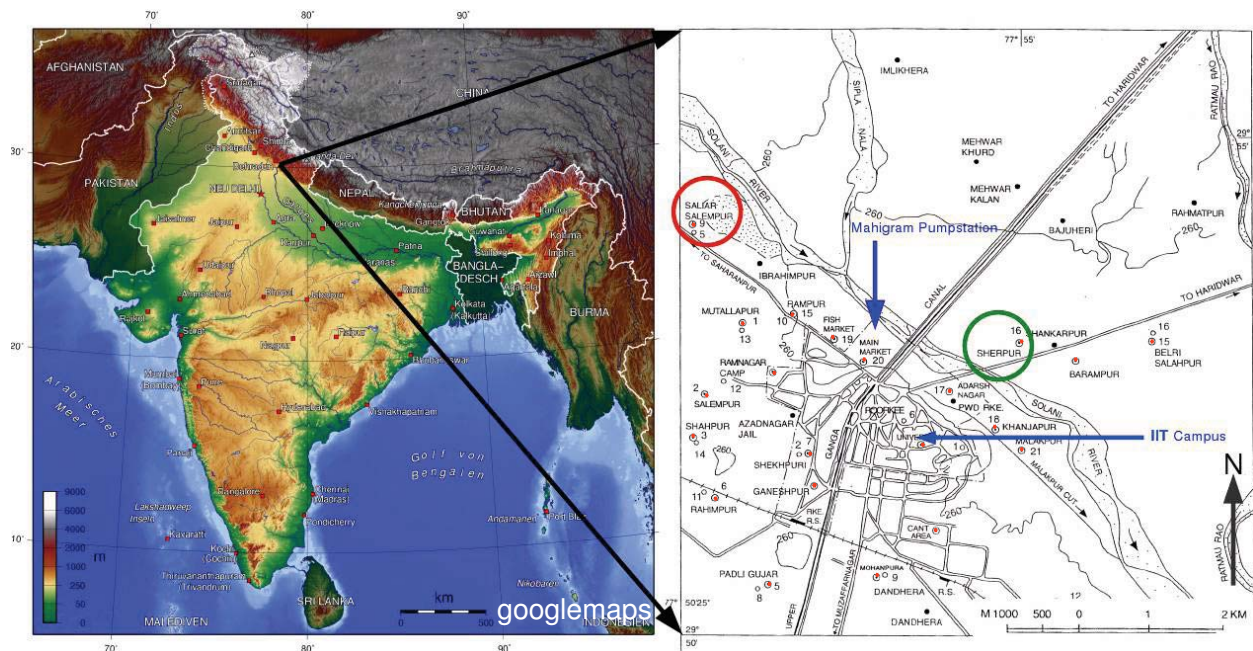


Abbildung 1: Übersicht der Region um Roorkee. Das Messgebiet Saliyar ist rot, das Referenzgebiet grün umrandet. Weitere wichtige Plätze sind farblich hervorgehoben und beschriftet. Brunnen, die überwacht und untersucht wurden sind als weiße Kringel mit roten Punkten dargestellt (Rechte Abbildung: [Singhal et al., 2003]).

wasserqualität wurde bisher überwiegend unter Berücksichtigung von hydrochemischen Daten untersucht. Die chemischen und bakteriellen Analysen des Grundwassers zeigen bisher nur erhöhte Werte für die Proben aus der unmittelbaren Umgebung von Saliyar, wo sich die Mülldeponie befindet. Singhal et al. [2003] schließen eine Gefährdung der Aquifersysteme weiter flussabwärts in Richtung Roorkee auf Grund der angenommenen Grundwasserfließrichtung. Sudha et al. [2010] haben bereits in den Gebieten nördlich von Roorkee, in Saliyar und in Khanjapur, eine Vielzahl von DC-Profilen und „Time Domain Electromagnetic“ (TEM) Stationen vermessen, um den Einfluss des Abwassers auf das oberflächennahe Aquifer zu detektieren und die Ausmaße des kontaminierten Bereichs zu bestimmen.

Ziel dieser Arbeit ist es ebenfalls, den Einfluss der Kontamination auf die Leitfähigkeitsverteilung der oberen 20 – 25 m zu bestimmen. Hierzu wurden die auf dem kontaminierten Gebiet vermessenen Profile mit dem Referenzprofil verglichen, auf dem keine Kontamination erwartet wird. Durch die Vielzahl der vermessenen Profile soll auch eine Aussage über die laterale Verbreitung des kontaminierten Bereichs, sowie die Ausbreitung von Kontaminationsfahnen getroffen werden.

Zusätzlich wurden die Datensätze beider Methoden mittels des 2D Joint Inversionsprogrammes RMTDC2D von Candansayar und Tezkan [2008] einzeln und gemeinsam invertiert. Es soll dabei untersucht werden, inwiefern die Anwendung der 2D Joint Inversion verlässlichere und einheitlichere Untergrundmodelle zu liefern vermag.

Geologie des Messgebiet

Der indische Subkontinent lässt sich geologisch in drei Bereiche unterteilen: den Himalaja im Norden, die Gangesebene und die indische Halbinsel im Süden. Die Gangesebene stellt dabei das Entwässerungsbecken der großen indischen Flüsse Ganges, Indus, Brahmaputra und Yamuna dar. Das Messgebiet befindet sich im nördlichen Teil der Gangesebene (Abbildung 1). Da es sich um Schwemmland des Ganges und des Solani handelt, weist der Boden überwiegend ungehärtete alluviale Sedimente jüngeren Alters mit einer einfachen eindimensionalen Schichtung auf. Diese Sedimente bestehen aus abwechselnden Schichten von Ton, Sand, Kunkur und in manchen Gebieten Kies, wobei Kunkur ein nodulares Kalziumkarbonat ist, welches sich häufig in Semi-ariden Regionen bildet. Lithologische Diagramme bei [Singhal et al.,

Bohrlochinformation		DC Ergebnisse	
Lithologie	z (m)	ρ (Ω m)	z (m)
Sandiger Lehm	0-4	>100	0-7
Sand	4-32	40-100	7-38
Ton + Kies	32-45	30-50	38-45
Sand	45-105	—	—

Tabelle 1: Bohrlochinformationen verglichen mit den DC Ergebnissen in Sherpur von Sudha et al. [2010].

2003] zeigen eine oberflächennahe Schicht aus sehr feinem bis sandigem Lehm von 3 – 6 m Mächtigkeit. Die darunterliegende Sandschicht in 3 – 27 m Tiefe bezeichnet den ersten unbegrenzten, oberflächennahen Aquifer. Darunter befindet sich eine nichtpermeable Ton-Kies-Schicht von ca. 13 m Mächtigkeit

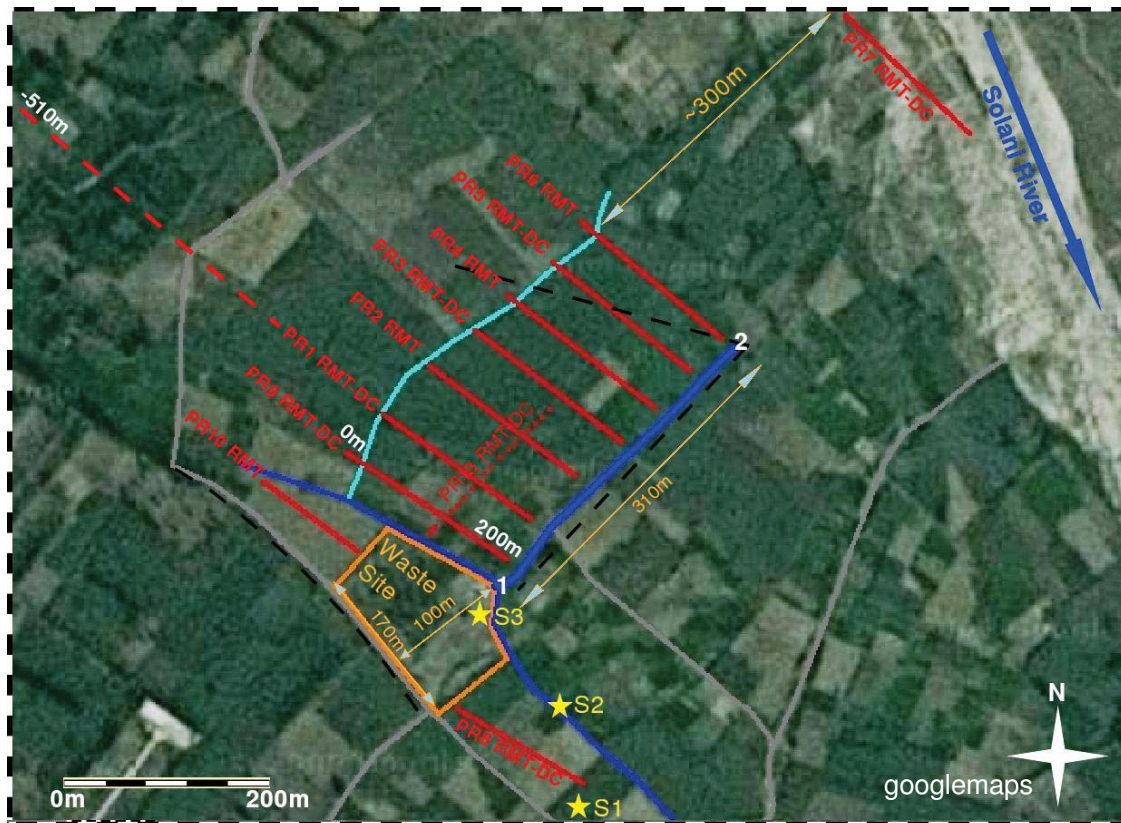


Abbildung 2: Die vermessenen Profile sind rot dargestellt und tragen die Bezeichnung der Profilnummer (PR1 bis PR13) sowie den Zusatz RMT oder RMT-DC, je nachdem ob nur RMT oder zusätzlich DC gemessen wurde. (Ab-)Wasserkanäle sind blau und Hochspannungsleitungen schwarz gestrichelt dargestellt. Der Hauptabwasserkanal verläuft von 1 nach 2. Auslässe für das Abwasser befinden sich entlang des Hauptabwasserkanals ca. alle 50m. An den Orten S1, S2 und S3 wurden Wasserproben entnommen.

und trennt den oberen Grundwasserleiter von dem tieferen, der überwiegend aus Sand und Kies besteht. In der Tabelle 1 sind die DC-Inversionsergebnisse von *Sudha et al.* [2010] mit den Bohrlochdaten verglichen. Die Bohrlochdaten wurden von einem Litholog auf dem Referenzprofil bei Sherpur abgeleitet und zeigen gute Übereinstimmung mit den DC-Ergebnissen.

Messung bei Roorkee

Insgesamt wurden 13 RMT und 8 DC Profile bei einer zweiwöchigen Kampagne im Februar 2009 vermessen (Abbildung 2). Zu dieser Jahreszeit war das Messgebiet nur teilweise kultiviert und daher gut zugänglich. Die Profile sind, ausgehend von der Mülldeponie am südlichen Bildrand bis zum Flussbett des Solani, überwiegend parallel angeordnet. Der Abstand der Profile betrug ca. 50 m, je nach verfügbarem Platz. Damit wurde eine ca. $200 \times 600 \text{ m}^2$ große Fläche abgedeckt. Das ganze Gebiet nordwestlich vom Hauptabwasserkanal hat ein leichtes Gefälle in Richtung Nordosten zum Solani hin. Die Mülldeponie im unteren Bildteil umfasst eine Fläche von $170 \times 100 \text{ m}^2$. Eine Messung direkt auf der Mülldeponie war nicht möglich, deswegen wurden Profil 8 und Profil 10 seitlich positioniert um einen etwaigen Einfluss der Mülldeponie durch z.B. Sickerwasser zu ermitteln. Der Hauptabwasserkanal verläuft von der Mülldepo-

nie nordöstlich in Richtung des Solani. Ungefähr alle 50 m befinden sich Wasserauslässe, um die Felder zu fluten. Der kleinere offene Kanal (hellblau dargestellt) im linken Bildbereich verläuft parallel zum großen und wurde von den Profilen gekreuzt. Der Bereich oberhalb von Profil 6 war stark bewachsen und konnte nicht vermessen werden. Das oberste Profil (Profil 7) verläuft im trockenen Flussbett des Solani und ist ca. 600 m von der Kontaminationsquelle entfernt, weswegen hier eine Abnahme der Kontamination zu erwarten ist. Der Bereich zwischen Profil 3 und Profil 6 war stark bewässert, konnte aber dennoch vermessen werden. Profil 1, welches im Folgenden explizit vorgestellt wird, wurde am ersten Tag mit einer Länge von 710 m und einem Stationsabstand von 10 – 50 m vermessen, um den Einfluss der Kontamination mit der Entfernung von der Quelle abzuschätzen. Die RMT-Messung wurde mit dem RMT-F Gerät der Universität zu Köln durchgeführt. Es arbeitet im erweiterten Frequenzbereich von 10 kHz bis 1 MHz und ermöglicht dadurch ein verbessertes Auflösungsvermögen oberflächennaher Schichten [Wiebe, 2007]. Der Messpunktabstand für die RMT-Stationen betrug bei allen anderen Profilen 10 m und die Anzahl in der Regel 21 Stationen. Gemessen wurden beide horizontalen Komponenten des elektrischen und des magnetischen Feldes zur Bestimmung der spezifischen Widerstands- und Phasenwerte der TE- und

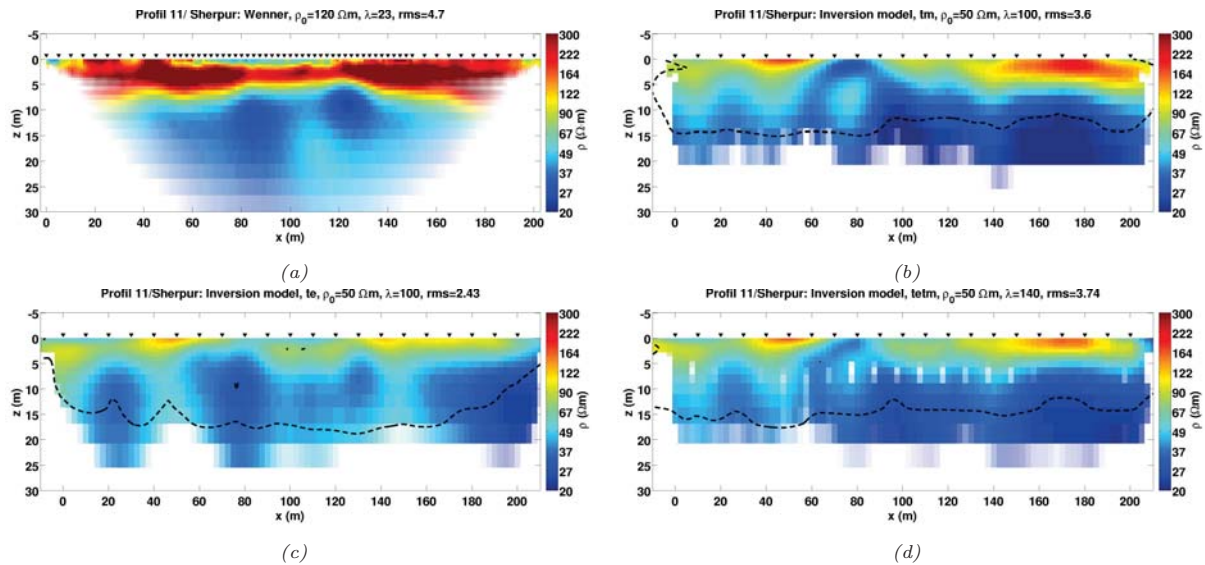


Abbildung 3: Inversionsmodelle des Referenzprofils 11 bei Sherpur der Daten der Wennerauslage (a), der Daten der TM-Mode (b), der der TE-Mode (c), sowie beider Datensätze (TE+TM) gemeinsam invertiert (d). Die RMT Stationen sind als schwarze Dreiecke dargestellt. Die 0.2 Isolinie des DOI ist als schwarz gestrichelte Linie geplottet und stellt die maximale Erkundungstiefe dar.

der TM-Mode. Für die DC wurde eine ABEM Terameter SAS 1000 benutzt. Als Messauslage wurde eine Wenner und eine Schlumberger „Short-Long“-Auslage gewählt. Durch den geringeren Elektrodenabstand im mittleren Bereich der Auslage ist eine oberflächennah bessere Auflösung gewährleistet.

Eine geologische Streichrichtung zur Ausrichtung der Profile lässt sich anhand der Kontaminationsausbreitung nur ungenügend festlegen. Zu erwarten wäre eine Ausbreitung der Kontamination von den Abwasserauslässen Richtung Nordwesten oder auch eine Verbreitung der Kontamination von der Mülldeponie ausgehend.

Demzufolge und in Anbetracht der nahezu eindimensionalen Struktur des Untergrundes wurden die Profile anhand der verfügbaren Sender für die RMT und der Verfügbarkeit von Platz ausgerichtet.

Inversionsergebnisse Referenzprofils 11 bei Sherpur

In den Abbildungen 3 sind die Inversionsmodelle der einzelnen 2D Inversionen für die Wennerauslage und für die RMT Daten mit DC2DINVRES [Günther, 2004] bzw. RUND2_NLCG2_FAST (RUND2) [Rodi und Mackie, 2001] dargestellt. Die nicht-sensitiven Bereiche sind in den Inversionsergebnissen ausgeblendet und der „Depth Of Investigation“-Index (DOI) nach Oldenburg und Li [1998] ist zur Abschätzung der maximalen Erkundungstiefe in den RMT-Modellen dargestellt.

Das DC Inversionsergebnis zeigt wie erwartet eine hochohmige Deckschicht mit einem spezifischen Widerstand von $\rho > 200 \Omega\text{m}$ und einer Mächtigkeit von 5 – 7 m, welche als sandige Lehmschicht identifiziert werden kann. Im Bereich von 0 – 60 m war das Profil sehr trocken, da dieser Bereich nicht als landwirtschaftliche Nutzfläche diente. Der Bereich von 60–

120 m war kultiviert und wurde wahrscheinlich regelmäßig künstlich bewässert, wodurch sich der erniedrigte Widerstand der Deckschicht bis zu einer Tiefe von ungefähr 2 m erklären lässt. Der Bereich danach zeigt wieder eine durchgehende hochohmige Deckschicht, obwohl dieser ebenfalls kultiviert war.

Entsprechend den lithologischen Informationen aus Tabelle 1 und den DC Ergebnissen von Sudha et al. [2010] zeigt das Inversionsmodell unter der Deckschicht eine leitfähige Schicht mit einem spezifischen Widerstand von 30 – 50 Ωm , welches das oberflächennahe Aquifer, also eine wassergesättigte Sandschicht, darstellt und als die aufzulösende Zielstruktur aufgefasst wird.

Die Inversionsmodelle der Daten der TM-Mode, der TE-Mode und der Joint Inversion beider Datensätze zeigen qualitativ eine identische Leitfähigkeitsstruktur. Auch die Abnahme des Widerstands im mittleren Teil des Profils ist identisch zum DC Inversionsmodell.

Die hochohmige Deckschicht wird allerdings mit einem spezifischen Widerstand von 100 – 200 Ωm rekonstruiert, welcher verglichen mit dem DC-Ergebnis zu gering ist. Die darunterliegende Zielstruktur wird wiederum mit einem spezifischen Widerstand von 20 – 40 Ωm aufgelöst und ist vergleichbar mit dem DC-Ergebnis. Zusätzlich wurde ein zweites RMT Profil in ca. 200 m Entfernung von Profil 11 vermessen. Die geologische Struktur ist hier identisch und der spezifische Widerstand der Zielstruktur wird ebenfalls mit einem Wert von 20 – 40 Ωm gut aufgelöst Yogeshwar [2010].

Eine Erklärung für den verminderten Widerstand der Deckschicht ist die geringe Sensitivität der RMT für hochohmige Bereiche. Desweiteren könnte ein statischer Versatz der ρ_a -Daten den geringen Widerstand der Deckschicht erklären. Oberflächennahe Inhomogenitäten in der Nähe der Empfängerelektroden kön-

nen z.B. einen zusätzlichen Versatz auf die ρ_a -Daten bewirken.

Die Unterkante des Aquifers wird in den Inversionsmodellen nicht aufgelöst. Zum einen hängt das mit der maximalen Erkundungstiefe zusammen, zum anderen ergibt sich wahrscheinlich kein entscheidender Leitfähigkeitskontrast zwischen dem oberflächennahen Aquifer und der darunterliegenden nicht-permeablen Ton-Kies-Schicht, deren spezifischer Widerstand ebenfalls ungefähr 40 – 50 Ωm beträgt.

Um die maximale Erkundungstiefe abzuschätzen wurde der DOI nach Oldenburg und Li [1998] berechnet. Für die Wennerauslage ergibt sich bei [Yogeshwar, 2010] eine maximale Erkundungstiefe von 20 m bis 25 m für den mittleren Bereich des Profils.

Für die RMT-Modelle ist der DOI als schwarz gestrichelte Linie in den Modellen dargestellt. Für das Inversionsmodell der Daten der TM-Mode ist die Erkundungstiefe ungefähr 15 – 20 m, wobei im Bereich von $x = 100 - 200$ m der DOI, auf Grund der erhöhten Leitfähigkeit, etwas abnimmt. Für die beiden anderen RMT-Inversionsmodelle in Abbildung 3 ergibt sich eine ähnliche Erkundungstiefe von ca. 15 m.

Inversionsergebnisse des kontaminierten Profils 1 bei Saliyar

Das Profil 1 befindet sich in ca 100 m Entfernung zur Mülldeponie und wurde am ersten Messtag vermessen (Abbildung 2). Es war nicht bewässert und kaum kultiviert. Benutzt wurde eine Wenner Long-Auslage mit einer Länge von 120 m, im Gegensatz zu den anderen DC Profilen, die ausschließlich eine Profillänge von 200 m haben. Das RMT-Profil wurde bis zu einer Entfernung von 710 m vom Hauptabwasserkanal (Abbildung 2) fortgesetzt um den Einfluss der Kontamination mit dem Abstand zur Quelle zu bestimmen, wobei der Stationsabstand im Bereich von $x = 0 - 200$ m zehn Meter betrug und dann bis zu 50 m vergrößert wurde. Ausgewertet wurden nur die Daten der TE-Mode, da die der TM-Mode viele Senderlücken und teilweise negative Phasenwerte aufwiesen.

In den Abbildungen 4(a) und 4(b) sind die Inversionsmodelle der einzelnen 2D Inversionen für die Wennerauslage und für die Daten der TE- Mode dargestellt. Das DC-Modell zeigt eine einfache, eindimensionale geologische Schichtung, wobei die Deckschicht im Vergleich zum Referenzprofil in Abbildung 3 wesentlich weniger mächtig ist und einen geringeren Widerstand von ca. 30 Ωm aufweist. Eine mögliche Ursache für die insgesamt weniger hochohmige Deckschicht mit ca. 2-3 m Mächtigkeit ist wahrscheinlich die sonst häufige Bewässerung mit Abwasser. Im rechten Bereich nahe des Hauptabwasserkanals befindet sich ein hochohmiger Bereich der Deckschicht, welcher von einer Erdaufschüttung herrühren könnte, und der in dem RMT Inversionsmodell identisch rekonstruiert ist. Darunter folgt das kontaminierte Aquifer, bzw., verglichen mit dem Ergebnis für das Referenzprofil, ein Bereich stark erhöhter Leitfähigkeit bis zu einer Tiefe von ungefähr 12 m. Dies ist im RMT-

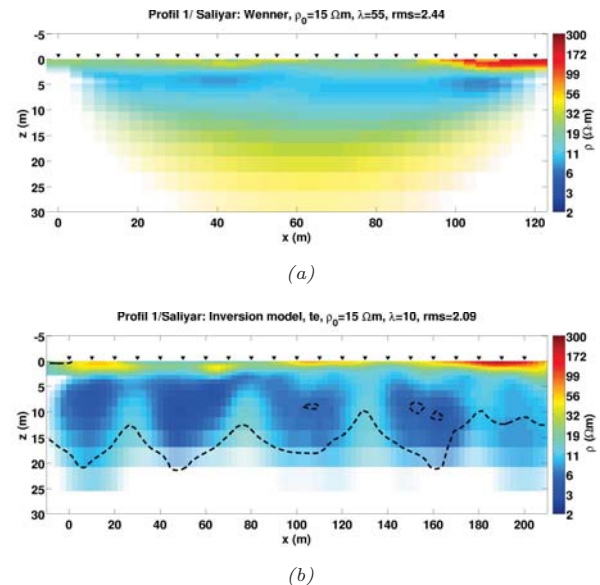


Abbildung 4: Inversionsergebnis des Profils 1 bei Saliyar der Daten der Wennerauslage (a) und der Daten der TE-Mode (b). Die RMT Stationen sind als schwarze Dreiecke dargestellt, der DOI als schwarz gestrichelte Linie.

Profil ebenfalls identisch wiedergegeben. Der spezifische Widerstand fällt im DC-Modell mit 6 – 15 Ωm etwas höher aus als der des RMT-Modells. Die Anpassung beider Datensätze ist mit einem $RMS < 2.5$ gut.

Die maximale Erkundungstiefe für das DC-Modell ergibt sich mit dem DOI-Index bei Yogeshwar [2010] zu 15 – 20 m im mittleren Bereich des Profils. Die schwarze Linie in Abbildung 4 stellt die maximale Erkundungstiefe von 10 – 15 m für das RMT-Modell dar. Mit dieser Erkundungstiefe für die RMT wird die Unterkante des kontaminierten Bereichs nur mit der DC aufgelöst.

Auf den vermessenen Profilen in Saliyar wurde mit dem Programm EMUPLUS [Wiebe, 2007] eine 1D RMT-DC und eine 1D RMT-DC Joint Inversion im jeweiligen Mittelpunkt des Profils durchgeführt um die Leitfähigkeit des kontaminierten Aquifers genauer zu bestimmen. Die 1D Marquardt Inversionen für Profil 1 (Abbildung 5) entsprechen sehr gut den 2D Ergebnissen im Mittelpunkt des Profils. Das Modell der Geoelektrik zeigt einen Dreischichtfall mit gut aufgelösten Schichtgrenzen und spezifischen Widerständen, außer für den der obersten Schicht. Die äquivalenten Modelle in Abbildung 5 unterscheiden sich also kaum. Ebenso zeigt das Ergebnis für die RMT-Daten nahezu identische äquivalente Modelle, was auf eine gute Auflösung beider Schichten schließen lässt. Beide Inversionsergebnisse stimmen gut überein, wobei die RMT die Unterkante des Aquifers nicht auflösen kann.

Das Modell der 1D Joint Inversion passt die Datensätze beider Methoden mit einem RMS von 1.65 gut an und löst die Unterkante der zweiten Schicht durch Hinzunahme der DC Daten auf. Der Widerstand des

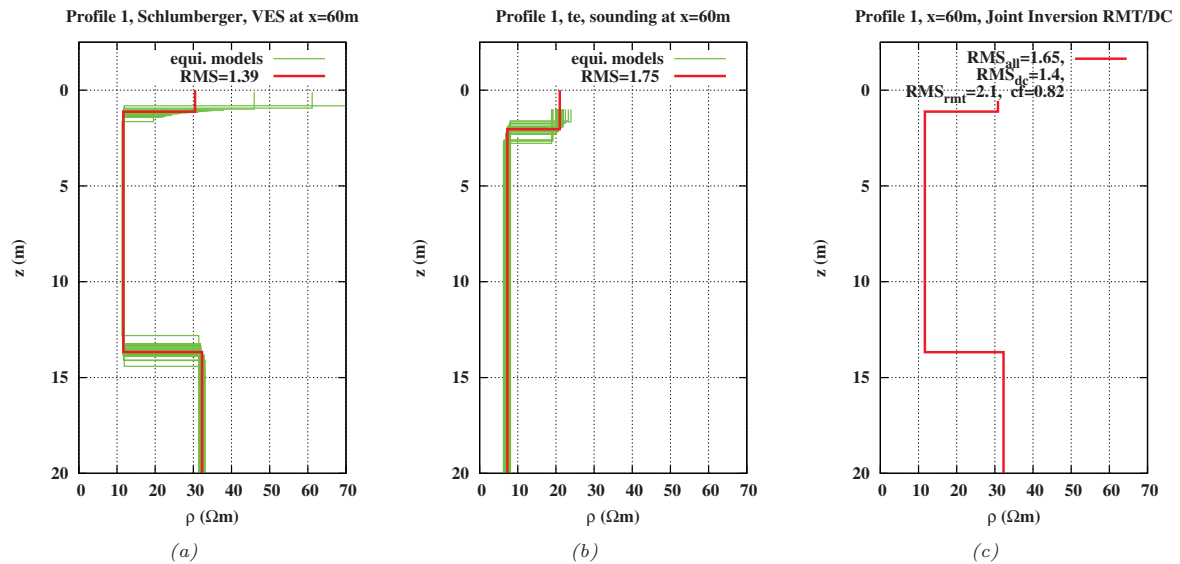


Abbildung 5: 1D Marquardt Inversionsmodelle für die Sondierungskurve der Schlumbergerauslage (a) und der Daten der TE-Mode (b), sowie der Joint Inversion beider Datensätze (c) bei $x = 60\text{ m}$ in der Mitte des Profils 1 bei Saliyar. Aufgetragen ist der spezifische Widerstand gegen die Tiefe. Äquivalente Modelle sind grün und der Mittelwert der äquivalenten Modelle sind rot dargestellt.

kontaminierten Bereichs ist mit $\rho \approx 10\text{ }\Omega\text{m}$ ziemlich genau bestimmt. Da weitaus mehr DC-Daten vorhanden sind, stellt das Inversionsergebnis zwar einen Informationsgewinn für die RMT dar, entspricht aber ziemlich genau dem 1D Modell der DC Inversion. Nur oberflächennah liefern die hochfrequenten RMT-Daten Zusatzinformation zur Verbesserung der Auflösung der obersten Schicht.

2d Joint Inversion

In der Abbildung 6 sind die Inversionsmodelle der Wenner Long Auslage, der Daten der TE-Mode und das Ergebnis der 2D Joint Inversion beider Datensätze mit RMTDC2D dargestellt. Die Leitfähigkeitstruktur der Endmodelle ist gut vergleichbar mit den Inversionsmodellen der einzelnen 2D Inversionen mit DC2DINVRES und RUND2.

Die Modelle der einzelnen Inversionen sind bis zu einer Tiefe von ungefähr 10 m gut vergleichbar, wobei die Anpassung der RMT-Daten mit einem RMS von 1.9 besser ist als die der DC-Daten. Der kontaminierte Bereich wird im RMT Inversionsmodell etwas leitfähiger rekonstruiert. Eine Aussage über die Unterkante des Aquifers lässt sich auf Grund der mangelnden Eindringtiefe der RMT nicht treffen, wohingegen die Unterkante im DC-Modell aufgelöst wird. Die 2D Joint Inversion in Abbildung 6 passt beide Datensätze mit einem RMS von 5.0 zufriedenstellend an. Die Unterkante und der Widerstand des Aquifers sind gut aufgelöst.

Die mit dem mittleren Tiefenbereich korrespondierenden DC-Daten in Abbildung 7(b) sind über die ganze Profillänge relativ gleichmäßig,

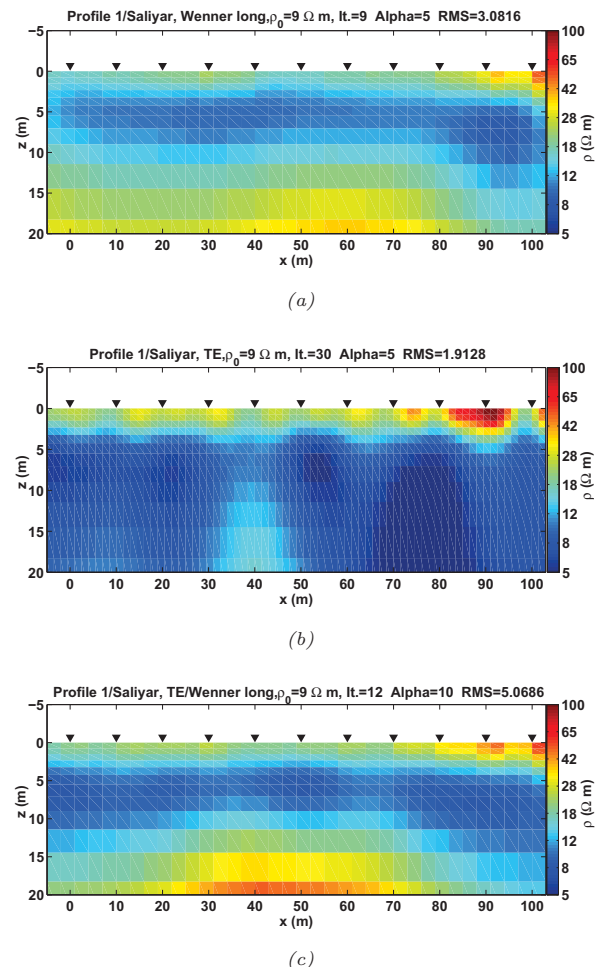


Abbildung 6: Inversionsergebnis der Wenner Long-Auslage (a), der TE-Mode (b) und der 2D Joint Inversion (c) des Profils 1. Die RMT-Stationen sind als schwarze Dreiecke dargestellt.

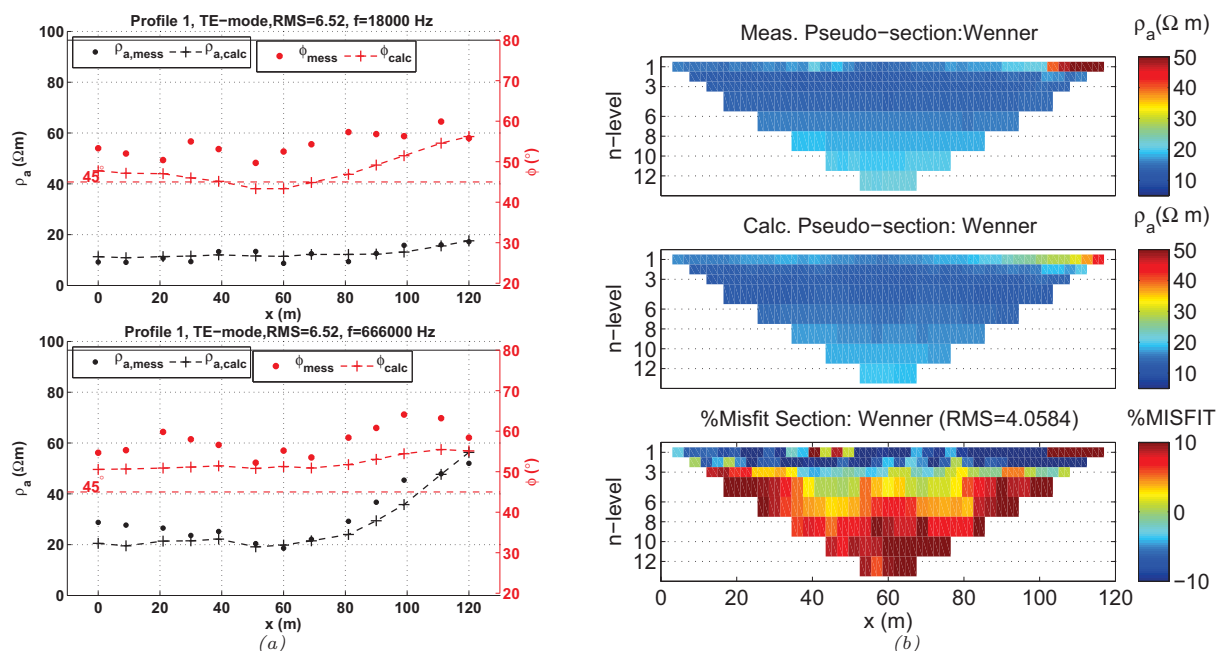


Abbildung 7: Daten und Datenanpassung der Daten der TE-Mode (a) und der Wenner Long Auslage (b) für das Ergebnis der 2D Joint Inversion mit RMTDC2D. Die gemessenen ρ_a und ϕ Werte sind als schwarze, bzw. rote Punkte und die Modellantwort als schwarze, bzw. rote Kreuze für $f = 18$ kHz, 666 kHz gegen die Station aufgetragen. Die DC-Daten und die Modellantwort sind als Pseudosektionen dargestellt, wobei die untere Abbildung (b) die relative Differenz beider zeigt.

wohingegen die RMT Daten der 18 kHz Frequenz in Abbildung 7(a) und das Inversionsmodell der einzelnen 2D RMT Inversion in Abbildung 6(b) auch Bereiche erhöhter Leitfähigkeit zeigen. Bei der 2D Joint Inversion werden die RMT-Daten daher nicht so gut wie die DC-Daten angepasst. Die Modellantwort der RMT ist durch die Regularisierung und dem Einfluss der insgesamt gleichmäßigeren DC-Daten sehr glatt. Der RMS für die RMT-Daten ergibt sich zu 6.5. Die Hinzunahme der RMT-Daten führt zu einer besseren Bestimmung der Deckschicht vor allem im etwas hochohmigeren Bereich von $x = 80 - 100$ m. Die DC-Daten in diesem Bereich sind nicht so gut angepasst, wie in der Darstellung der relativen Differenz der Daten und der Modellantwort in Abbildung 7(b) erkennbar ist. Eine Verringerung des Glättungsparameters passt diese zwar besser an, führt allerdings zu Inversionsartefakten für die tieferen Bereiche. Qualitativ zeigen die Auflösungsmatrizen der 2D Joint Inversion bei Yogeshwar [2010] eine verbesserte Auflösung im Vergleich zur RMT-Einzelinversion, wobei die Randbereiche im Vergleich zur DC-Einzelinversion besser aufgelöst sind, da die DC dort keine Tiefeninformation liefert. Die maximale Erkundungstiefe wird durch die DC-Daten bestimmt, weswegen keine erhöhte Sensitivität der tieferen Strukturen durch Hinzunahme der RMT-Daten erreicht wird.

Abschätzung der Kontaminationsverbreitung

Um die lateralen Ausmaße des kontaminierten Bereichs abzuschätzen, wurden insgesamt 9 parallele RMT und 5 parallele DC Profile auf einem ca 200×600 m² großen Gebiet vermessen (Abbildung 2). Die

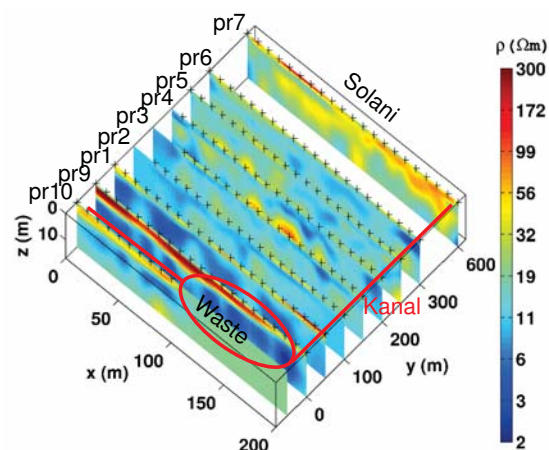


Abbildung 8: Ansicht aller RMT-Profile von der Mülldeponie bis zum Solani. Der Hauptabwasserkanal verläuft von der Mülldeponie in Richtung des Solani und ist rot dargestellt.

RMT-Profile sind in Abbildung 8 als Flächenschnitte der xz-Ebene dargestellt. Die Profile verlaufen parallel zwischen der Mülldeponie und dem Solani. Die RMT-Modelle liefern auf Grund der teilweise mäßigen Datenqualität nicht so glatte und gleichmäßige Modelle wie die DC-Modelle bei Yogeshwar [2010]. In der Darstellung 8 ist eine leichte Abnahme der Leitfähigkeit zum Solani hin zu beobachten. Das Profil 7 am Solani in Abbildung 8 zeigt insgesamt hochohmige Bereiche im Inversionsmodell für die RMT. Die Inversionsergebnisse der DC-Daten sind bei Yo-

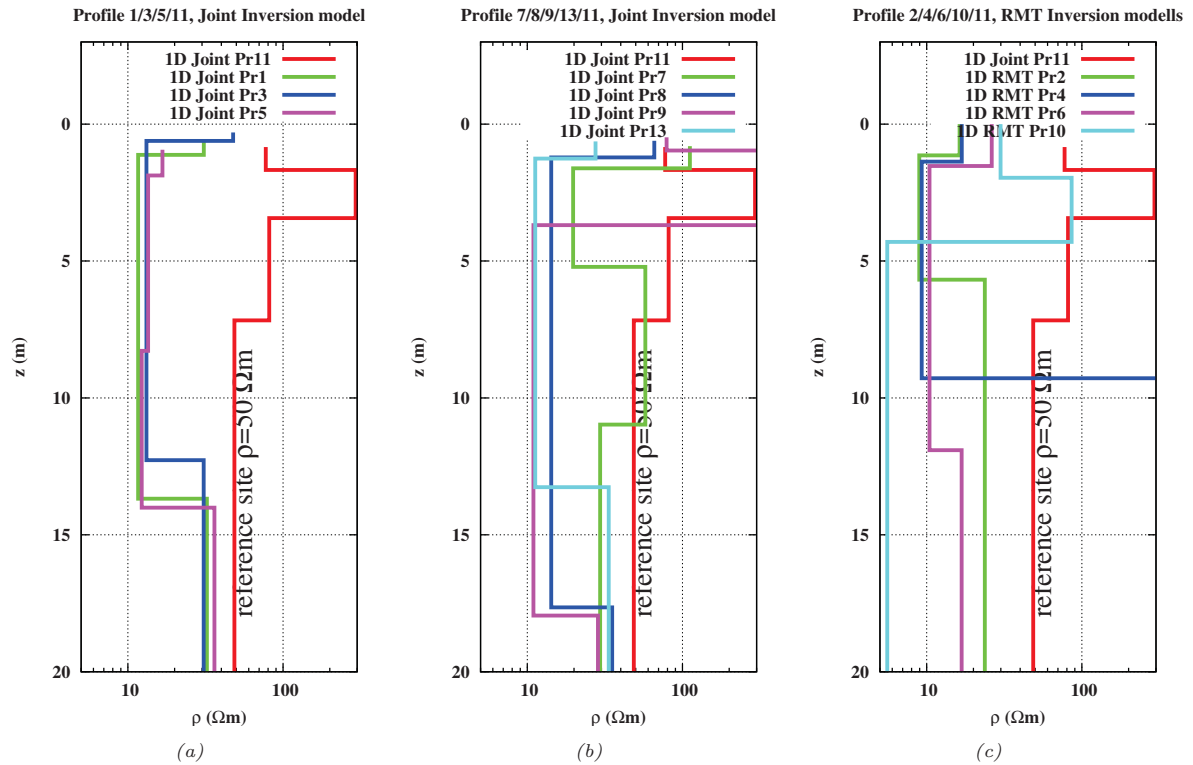


Abbildung 10: 1D Marquardt Joint Inversionsmodelle der DC- und RMT-Daten im jeweiligen Mittelpunkt der Auslage für Profil 1,3,5 (a) und Profil 7,8,9,13 (b), sowie einzelne RMT 1D Inversionsmodelle für Profil 2,4,6,10 (c). Das 1D Joint Inversionsmodell des Referenzprofils (Profil 11) ist rot dargestellt.

geshwar [2010] mit denen der RMT sehr gut vergleichbar, wobei die Unterkante des Aquifers jedoch mit der DC aufgelöst wird. Um eine genauere Vorstellung

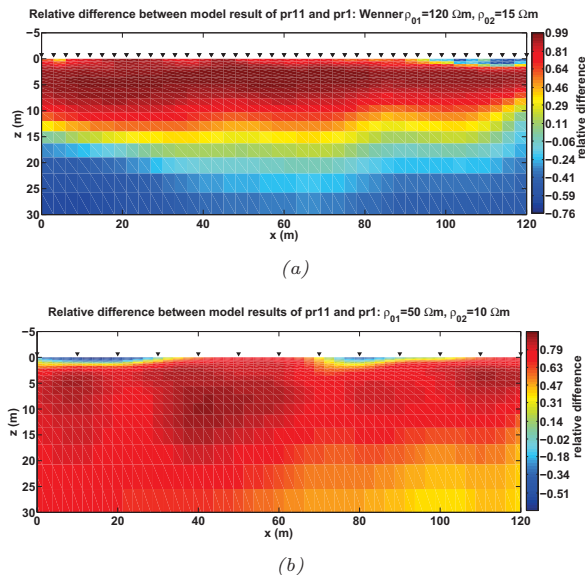


Abbildung 9: Dargestellt ist die relative Differenz der RMT- (a) und der DC-Inversionsmodelle (a) des Referenzprofils 11 und des kontaminierten Profils 1 (Relative Differenz $> 0 \Rightarrow \rho_{ref} > \rho_{pr1}$).

des Leitfähigkeitsunterschiedes zwischen dem Inversionsmodell des Referenzprofils und den Profilen auf dem kontaminierten Bereich zu erhalten, wurde für

jedes Profil die relative Differenz beider Modelle berechnet. Die relative Differenz des Profils 1 mit dem Referenzprofil 11 in Abbildung 9(a) zeigt für die Geoelektrik über die gesamte Auslagenlänge einen wesentlich geringeren spezifischen Widerstand. In einer Tiefe von ca. 15 m befindet sich die Unterkante des kontaminierten Aquifers und die Widerstände werden vergleichbar, bzw. der des Profils 1 wird größer. Für die RMT zeigt die Abbildung 9(b) ein identisches Ergebnis. Auch hier ist der spezifische Widerstand des Inversionsmodells für das Referenzprofil weitaus höher. Bei beiden Ergebnissen zeigen nur kleine oberflächennahe Bereiche der Deckschicht ein anderes Verhalten. In der Darstellung der relativen Differenz bei Yogeshwar [2010] zeigen die vermessenen Profile auf dem kontaminierten Gebiet in Saliyar im Vergleich mit dem Referenzprofil ausschließlich erhöhte Leitfähigkeitswerte bis zu einer Tiefe von ca. 15–20 m, bis auf das Profil 7, welches sich am Ufer des Solani befand und am weitesten von der Kontaminationsquelle entfernt war.

In Abbildung 10(a) und 10(b) sind die 1D Joint Inversionsmodelle der Daten aller Profilmittelpunkte in Saliyar mit dem des Referenzprofils verglichen. Insgesamt zeigen alle Profile unterhalb einer etwas hochohmigeren Deckschicht einen spezifischen Widerstand von $\rho \approx 10 \Omega\text{m}$ für das kontaminierte Aquifer, wiederum gefolgt von einer Schicht mit einem erhöhten spezifischen Widerstand. Das in grün geplottete Modell im rechten Teil der Abbildung 10 entstammt dem Pro-

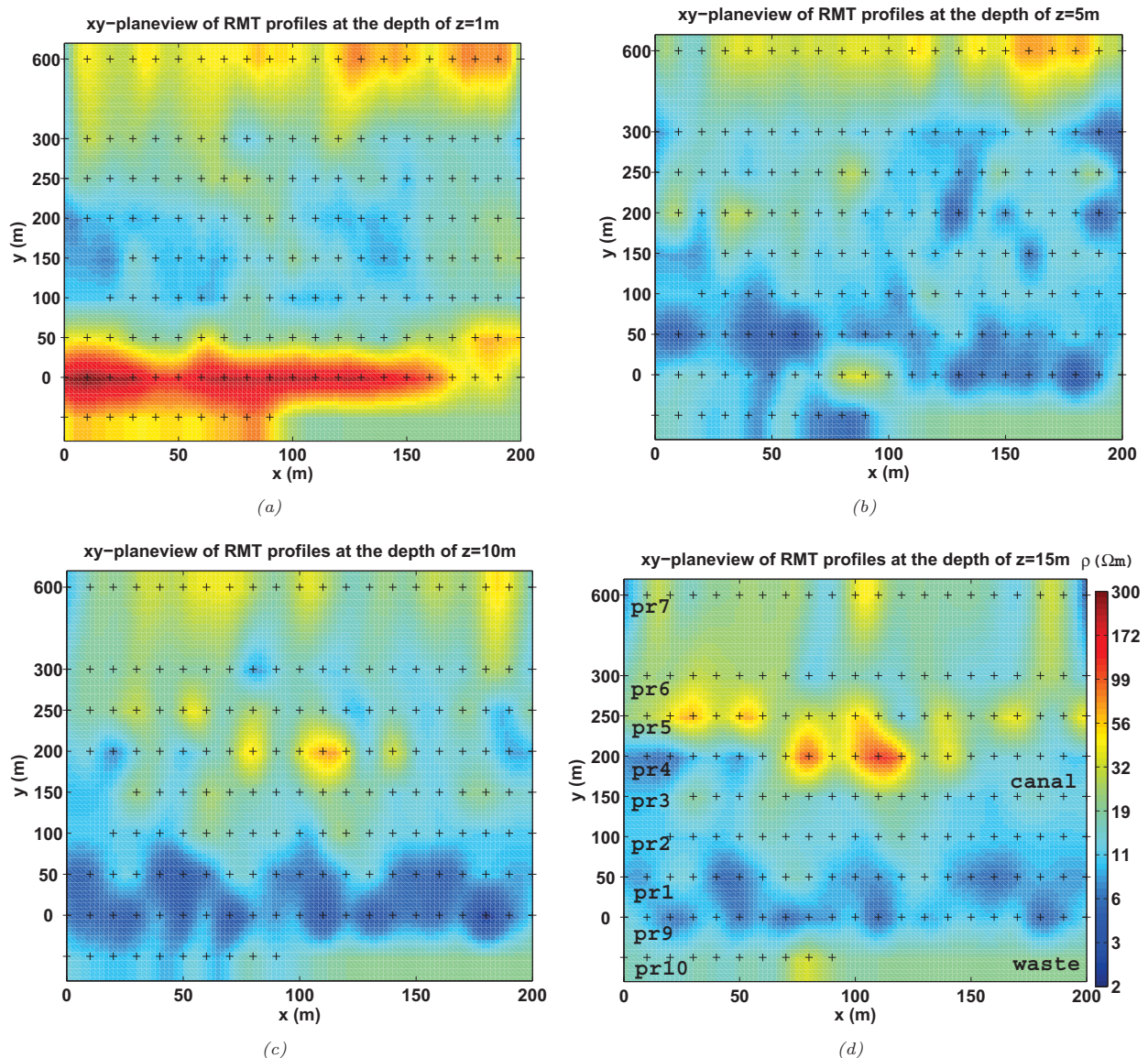


Abbildung 11: xy -Schnitte der RMT Inversionsmodelle von RUND2 für verschiedene Tiefen: 1 m (a), 5 m (b), 10 m (c) und 15 m (d). Die Kreuze bezeichnen die RMT-Stationen.

fil 7 am Solani und zeigt eine etwas andere Struktur mit höheren Widerständen, wie schon in Abbildung 8 ersichtlich war. Die Modelle der 1D RMT Inversionen in Abbildung 10(c) zeigen ebenfalls geringere spezifische Widerstände für das Aquifer. Das Ergebnis von Profil 4 in Abbildung 5(c) zeigt den Beginn einer hochohmigen Schicht in ca. 8 m Tiefe, die auch im 2D Modell der TM-Mode zu finden ist, aber nicht in dem der TE-Mode, und auf mangelnde Datenqualität oder die Streichrichtung zurückzuführen ist [Yoges-hwar, 2010]. Die Ergebnisse der einzelnen 2D Inversionsmodelle wurden für beide Methoden linear auf ein dreidimensionales Gitter interpoliert. Um eine Aussage über die laterale Verbreitung der Kontamination mit der Tiefe zu treffen, sind in Abbildung 11, am Beispiel der RMT, Flächenschnitte in der xy -Ebene für vier verschiedene Tiefen, $z=1\text{ m}$, 5 m , 10 m und 15 m , dargestellt.

In einer Tiefe von einem Meter erkennt einen sehr hochohmigen Bereich direkt in der Umgebung der

Mülldeponie. Das gesamte Profil 10, Profil 9 und die ersten 50 m direkt am Kanal von Profil 1 zeigen diese hochohmige Deckschicht. Dies könnte von einer Erdaufschüttung im Zusammenhang mit der Mülldeponie herrühren. Genauer ist jedoch nicht bekannt. Für $z = 5\text{ m}$ zeigt die Flächendarstellung spezifischen Widerstände um die $10\ \Omega\text{m}$ im Bereich von $y = 0\text{ m}$ bis 300 m . Die Leitfähigkeit ist in der Nähe der Mülldeponie erhöht und nimmt dann in Richtung des Solani etwas ab. Das Profil 7 am Solani ist wie bereits gesagt über den gesamten Tiefenbereich hochohmiger als die anderen Profile.

In einer Tiefe von 10 m und 15 m lässt die Kontamination etwas nach und es zeichnet sich die Unterkante des Aquifers ab, wobei der Bereich nahe der Mülldeponie im Vergleich zum Rest leitfähiger ist. Dies lässt wieder auf einen Einfluss der Mülldeponie schließen, vor allem da die Leitfähigkeit der Profile in der unmittelbaren Nähe der Deponie bis zu einer größeren Tiefe erhöht ist. So ist für Profil 9 in 15 m keine Un-

terkante des kontaminierten Aquifers erkennbar. Die DC-Ergebnisse bei *Yogeshwar* [2010] bestätigen die der RMT, wobei die Unterkante des kontaminierten Bereichs sich in 15 m wesentlich deutlicher bei den DC-Ergebnissen abzeichnet.

Eine solche flächenhafte Darstellung ist auf Grund der Interpolation zwischen den Profilen kritisch zu bewerten. Zwischen dem letzten Profil und dem Ufer des Solani lässt sich keine verlässliche Aussage treffen, da sich die Entfernung auf ca. 300 m beläuft. Problematisch sind auch Inversionsartefakte, die bei der Interpolation zwischen den Profilen überakzentuiert werden. Trotzdem eignet sich die Flächendarstellung zur Visualisierung der lateralen Leitfähigkeitsverteilung und insbesondere, um ein räumliches Bild des Untergrundes zu gewinnen.

Diskussion und Ausblick

Die Inversionsmodelle, der bei der Mülldeponie aufgenommenen Datensätze, haben in der Regel einander ähnliche Strukturen. Unterhalb einer etwas hochohmigeren Deckschicht zeigen diese, im Vergleich zu dem Inversionsmodell des Referenzprofils, Werte des spezifischen Widerstands kleiner als 10 Ωm bis zu einer Tiefe von ungefähr 15 m. Dieser Bereich konnte mit dem oberflächennahen Grundwasserleiter identifiziert werden und der im Vergleich geringere Wert des spezifischen Widerstands kann auf den Einfluss des Abwassers zurückgeführt werden.

Aus der flächenhaften Darstellung der Inversionsmodelle konnte die laterale Verbreitung der Kontamination abgeleitet werden. Erkennbar ist eine Zunahme des spezifischen Widerstands mit der Entfernung von der Mülldeponie, so dass diese vermutlich auch als Kontaminationsquelle angenommen werden kann. Auch reichen die Werte verringerter spezifischer Widerstände nahe der Deponie vergleichsweise tief.

Darüber hinaus sind die spezifischen Widerstände der RMT-Inversionsmodelle in der Nähe des Hauptabwasserkanals, wo sich die Wasserauslässe befinden, ebenfalls verringert. In ungefähr 600 m Entfernung von der Mülldeponie im Flussbett des Solani sind die Werte des spezifischen Widerstands wiederum deutlich höher und vergleichbar mit denen des Referenzprofils.

Insgesamt lassen sich die Mülldeponie und das Abwasser als Kontaminationsquellen bestätigen und es kann eine Systematik der Kontaminationsverbreitung mit dem Abstand von den Quellen abgeleitet werden. Das Programm RMTDC2D [*Candansayar und Tezkan*, 2008] lieferte vergleichbare Inversionsmodelle wie die einzelnen Inversionen der Datensätze mit DC2DINVRES [*Günther*, 2004] und RUND2 [*Rodi und Mackie*, 2001]. Insbesondere die Inversionsmodelle der einzelnen Inversionen mit RMTDC2D waren den Inversionsmodellen der anderen Programmen sehr ähnlich.

Bei Inversionsmodellen, die in ihrer Struktur starke Unterschiede zeigten, war die Joint Inversion schwierig, da beide Datensätze nicht gleichzeitig anzupassen waren. Trotzdem konnte mit der Jointinversion

ein einheitliches Modell beider Datensätze gefunden werden, was eine nützliche Hilfe bei der Interpretation der Ergebnisse darstellt.

Vor allem konnte die Unterkante des kontaminierten Bereichs durch Hinzunahme der DC-Daten aufgelöst werden und gleichzeitig lieferten die RMT-Daten Zusatzinformation an den Rändern der Profile, wo mit der DC keine Tiefenaussage zu treffen möglich war. Die Joint Inversion führte unter diesem Gesichtspunkt zu verbesserten Endmodellen.

Problematisch gestaltete sich allerdings die Auswirkung des Regularisierungsparameters. Bei einem zu geringen Wert ergaben sich stark überstrukturierte Inversionsmodelle und Inversionsartefakte.

Schwierig war auch die Bewertung der Inversionsmodelle anhand der Auflösungsmatrizen. Die Auflösungsmatrizen der Joint Inversion waren mit denen der einzelnen Inversionen von RMTDC2D nicht vergleichbar und eine quantitative Aussage über das verbesserte Auflösungsvermögen der Joint Inversion konnte daher nicht getroffen werden.

Im Rahmen des DFG-DST Projektes, welches diese Arbeit ermöglicht hat, ist eine weitere Messung auf dem selben Gebiet vorgesehen. Hierbei soll die „Transient Electromagnetic“ Methode (TEM) zum Einsatz kommen. Mit dieser Methode besteht die Möglichkeit tiefer liegende Strukturen aufzulösen. Wünschenswert wäre mit dieser Methode den Einfluss der Kontamination auf den tiefer liegenden Grundwasserleiter zu untersuchen, da sich die hier vorliegende Arbeit auf die Erkundung des oberen Grundwasserleiter konzentriert hat.

Des Weiteren eignen sich die verwendeten Methoden gut für die Erkundung von kontaminierten Böden und könnten gerade in Indien vermehrt zum Einsatz kommen um damit zur Verbesserung der Grundwasserversorgung beizutragen.

Interessant wäre auch die Einbeziehung der geophysikalischen Ergebnisse in eine hydrogeologische Interpretation, um die Auswirkung der Kontamination auf größeren räumlichen und auch zeitlichen Skalen abzuschätzen.

Die 2D Joint Inversion stellt ein zusätzliches Hilfsmittel zur Interpretation von RMT- und DC-Inversionsmodellen dar. Die einzeln erhaltenen Inversionsmodelle lassen sich durch die Hinzunahme der Joint Inversionsmodelle leichter qualitativ und quantitativ bewerten. Eine weitere Anwendung dieses Programmes ist daher wünschenswert.

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Site Effect Assessment in the Mygdonian Basin (EUROSEISTEST area, Northern Greece) using RMT and TEM Soundings

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Abstract

During the project “Euroseistest Volvi-Thessaloniki”, a strong-motion test site (EUROSEISTEST) for Engineering Seismology was installed in the Mygdonian Basin between the two lakes Volvi and Lagada ca. 45 km northeast of Thessaloniki (Northern Greece). The basin itself is a neotectonic graben structure (5 km wide) with increased seismic activity along distinct normal fault patterns. Fluvio-terrestrial and lacustrine sediments (approximately 350-400 m thick) are overlying the basement consisting of gneiss with schist. To improve the seismic wave propagation model it is vital to know about site effects, e.g. the geotectonic properties of the area such as the top-of-basement, vertical tectonic boundaries (faults and basement fracturation) and the geothermal regime. Therefore, we carried out near surface EM studies to understand the distribution of the active faulting and the top of basement structure of this particular area.

The RMT (Radiomagnetotelluric) and TEM (Transient electromagnetic) measurements were carried out on three profiles and thirty sounding. The inverted RMT and TEM data show generally a four layer model. The layers are indicated as metamorphic and sediment rocks, which are in detail: marly silty sand with gravel ($> 100 \Omega\text{m}$), marly silty sand with clay (50 - 100 Ωm), sandy clay (30 - 50 Ωm) and silty sand (10 - 30 Ωm). Due to the high resistivity of the top layer, the skin depths of the RMT soundings are around 35 m. The TEM data gives detail information of the lower structure down to a depth of 200 m. According to our analysis, a normal fault next to the Euroseistest could be located having a strike direction of N 60 E. The joint interpretation of RMT and TEM data proves to be an effective tool to investigate complex geology structures.

Introduction

This study refers to the Thessaloniki area, which has recently been affected by the 1978 destructive earthquake sequence [Papazachos and Papazachou, 1997]. It has been well established that the strong ground motion of such a seismic event causes irregularly distributed modification to the local geology [Tranos and Mountrakis, 1998], [Raptakis, et. al., 1999]. Different geophysical methods have been applied in this area [Thanassoulas, et.al.,1987], [Raptakis,et.al.,1999], [Savvaidis,et.al.,1999], [Tranos, et.al, 2003], [Gurk, et.al., 2007], however detail information of fault structures has not been verified so far. Ambient noise measurements from the area east of the Euroseistest experiment give strong implication for a complex 3-D tectonic setting. Therefore we carried out near surface EM studies to understand the distribution of the active faulting and the top of basement structure of this particular area.

Joint TEM and RMT inversion has been successfully applied to geological and engineering problems in the past [Tezkan et.al., 1995], [Schwinn, 1999], [Steuer, 2002]. The RMT method has low penetration depth and therefore gives information of the surface layers, whereas the TEM method gives detail information of the lower structure of the investigated area. Hence, a joint interpretation of RMT and TEM data will produce a good resolution of resistivities and thicknesses, in shallow and in deeper parts of the subsurface.

Test Area and Geological Setting

The investigated sites are located between the Lagada and the Volvi Lake in Northern Greece. The area is now covered with alluvial deposits as shown in (Fig.1). The local geological mapping in this area was done by the Geology Institute the University of Cologne [Maith, 2009]. The Mygdonian system corresponds to various types of sediments that were deposited in a quaternary graben structure. Due to its hydrothermal activity, the top was intensively covered with travertine/tufa deposits, now mostly removed by erosion [Jongmans, et.al. 1998]. The local geological map of the area is shown in Fig.1. We find four major units: metamorphic basement formed by schist and gneisses in the deeper depths, lower terrace deposit was deposited on the top of the basement metamorphic, lacustrine and deltaic sediments including conglomerate, gravel and sand. Fans comprise of soil and silt are located between lower terrace deposit and the Holocene deposit is composed of sand, silt and clay which are deposited on the top of these sediments. To obtain detail information of the geology in this area, TEM and RMT data were observed on all various types of the sediments (Fig.1).

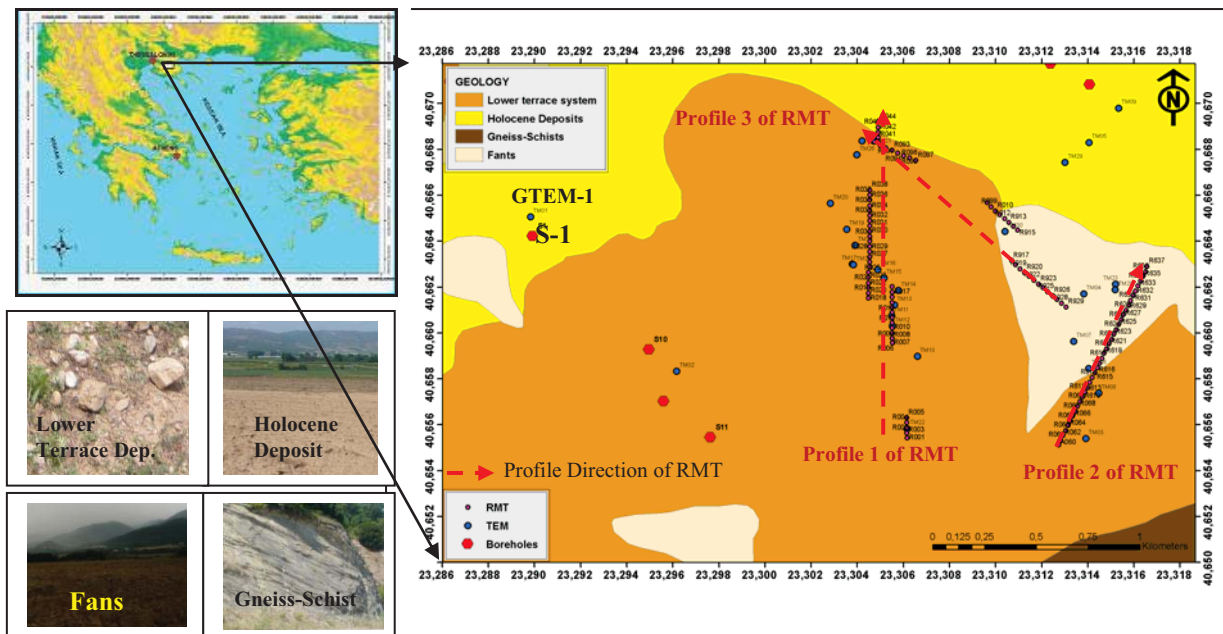


Fig. 1. Site Location, geology of the study area, location of RMT station (●), TEM station (●) and location of boreholes (■) are also displayed in the figure GTEM-1 at the location of TEM-1 sounding.

RMT Measurements and Interpretation

The RMT measurements were carried out on three profiles as indicated in Figure 1. Parameter of the survey design is listed in Table 1. The RMT of profile 1 is along 1600 m with direction N 0 S, whereas the direction of profile 2 is located N60 E, which are parallel with transmitters. The direction of profile 3 in this area is N60W with length of 1400 m, which are perpendicular with transmitter. Due to partial inaccessibility of the area the profiles could not be set up in a straight line. The RMT-F system consists out of two magnetic sensors (induction coils, 30 cm length), a preamplifier for the

electrical channels, two electrical antennae and a recorder. It is a four channel instrument (E_x , E_y , H_x , H_y) with the capability of estimating the full impedance tensor [Tezkan and Saraev, 2008].

Table1. Parameter of RMT survey design

RMT Profile	Length [m]	Site distance [m]	Orientation	Transmitter azimuth
Profile 1	1600	25	N0°S	N60°E
Profile 2	1000	25	N60°E	N60°E
Profile 3	1400	25	N60°W	N60°E

A special feature is the way how the electric field is sampled. Instead of grounded dipoles, the device uses symmetrical electrical dipoles (e.g. two arms of 20 m length) that are capacitively coupled to the ground. The system records time series in two bands: D2 band (10-100 kHz) and D4 band (100 kHz—1 MHz). The coherency level is important to prevent signals with high noise level, generally we used for this study a coherency value of 0.8.

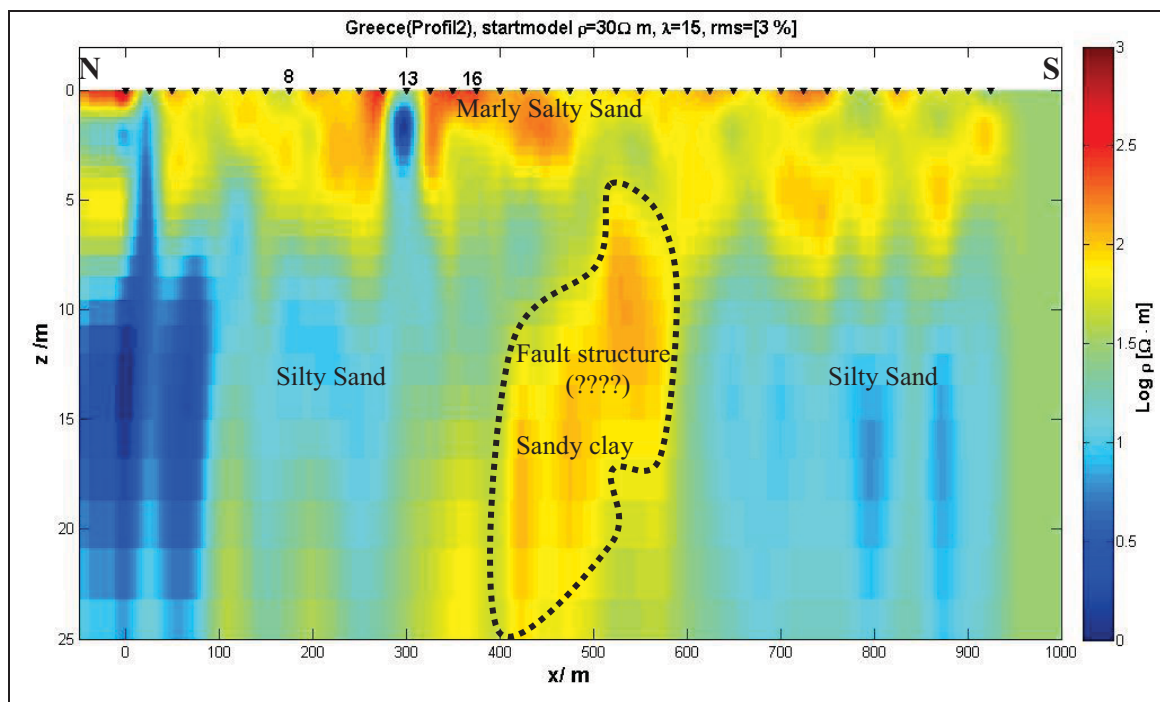


Fig 2. The RMT 2-D inversion model of profile 2 indicates metamorphic rock (marly salty sand) in a depth of 0 -5 m. Stations 1 – 16 and stations 25 – 38 show a very low resistivity ($<50 \Omega\text{m}$) zone at depth (5 m – 25 m) which is interpreted as sedimentary (silty sand). Whereas station 17-24 indicate the fault structure (----) as sandy clay.

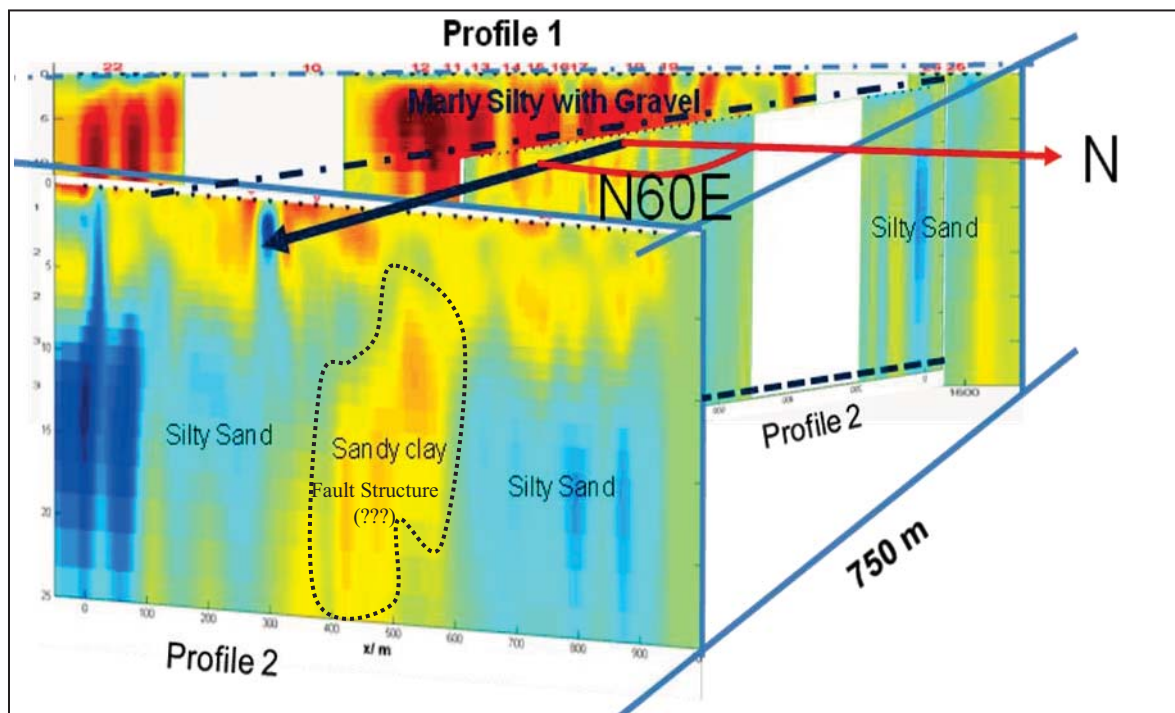


Fig.3. The Fault structure direction (black arrow) derived from 2-D RMT inversions

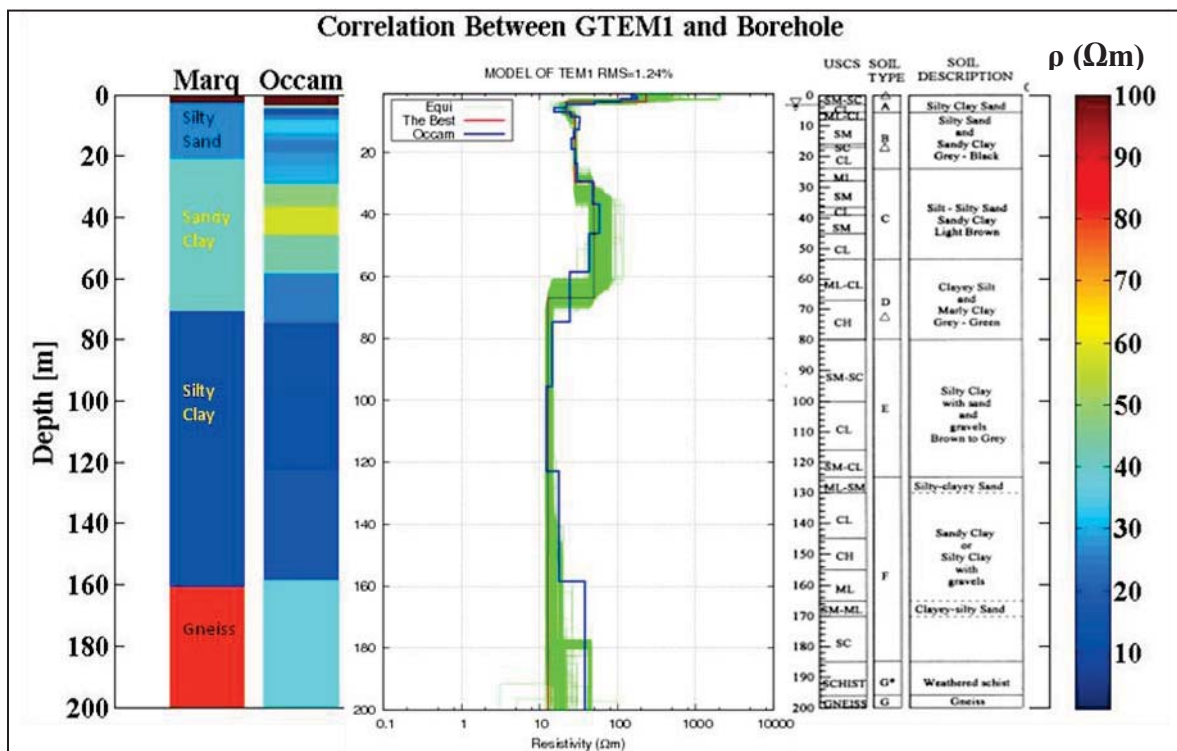


Fig.4. Correlation between the 1-D inversion of the TEM data (GTEM1) at the reference site and borehole S-1 information

In order to calibrate the geophysical measurements with the lithology, TEM data at our reference site was correlated with borehole data in S-1 (Fig.1). This analysis allows identifying characteristic strata based on the electrical conductivity. The modeling of RMT data was done by using 2-D inversion techniques (Fig. 2 and Fig.3). The 2-D inversion was performed with the 2-D Mackie code [Mackie, et.al, 1997], the 1-D inversion was done with the Marquardt [Cerv and Pek, 1979] and Occam algorithm [Constable, 1978]. Prior to any 2-D modeling, the penetration depth have been estimated using the ρ^* (z^*) transformation [Schmucker, 1979]. Penetration depths are found to be around 35 m.

As an example, we explain the RMT model of profile 2 in detail. This model (Fig.2) shows high resistivities on the top layer (more than 100 Ω m) which correlates to the marly silty with clay of the borehole data (Fig.4). For stations 1 -16 and stations 25-38 there is a low resistive structure beneath the surface layer. Between stations 17-24 there is a high resistive structure with the same resistivities as the surface layer. This region is interpreted as a fault structure (Fig.2), which is filled with sedimentary rock (sandy clay).

The data quality and the consistency between ρ_a and ϕ could be checked in the apparent resistivity curves, which are shown in Fig.5. RMT station 8 on profile 2 show that the resistivity values decrease from highest resistivity (> 80 Ω m) to low resistivity (30 Ω m) with phase values less than 50 °. The sounding curves of station 8 and 13 (figure 5) and station 16 (figure 5) on profile 2 shows that the phases exceeds value more than 45° indicating a low resistive (sediment) structure at larger depths, meanwhile the good conductive metamorphic series is indicated by the phase values lower than 45. Fig. 6 shows a comparison between measured and calculated data at selected frequencies (79 kHz, 387 kHz and 784 kHz). Measured and synthetic data fit well together keeping in mind the geological complexity and inhomogeneous in this survey area, however some misfits which are associated as 3-D effect.

The direction of the fault can be constructed using the 2-D inversion models of profile 1-3 as shown in Fig.3. The direction of the fault structure is found to be N60E.

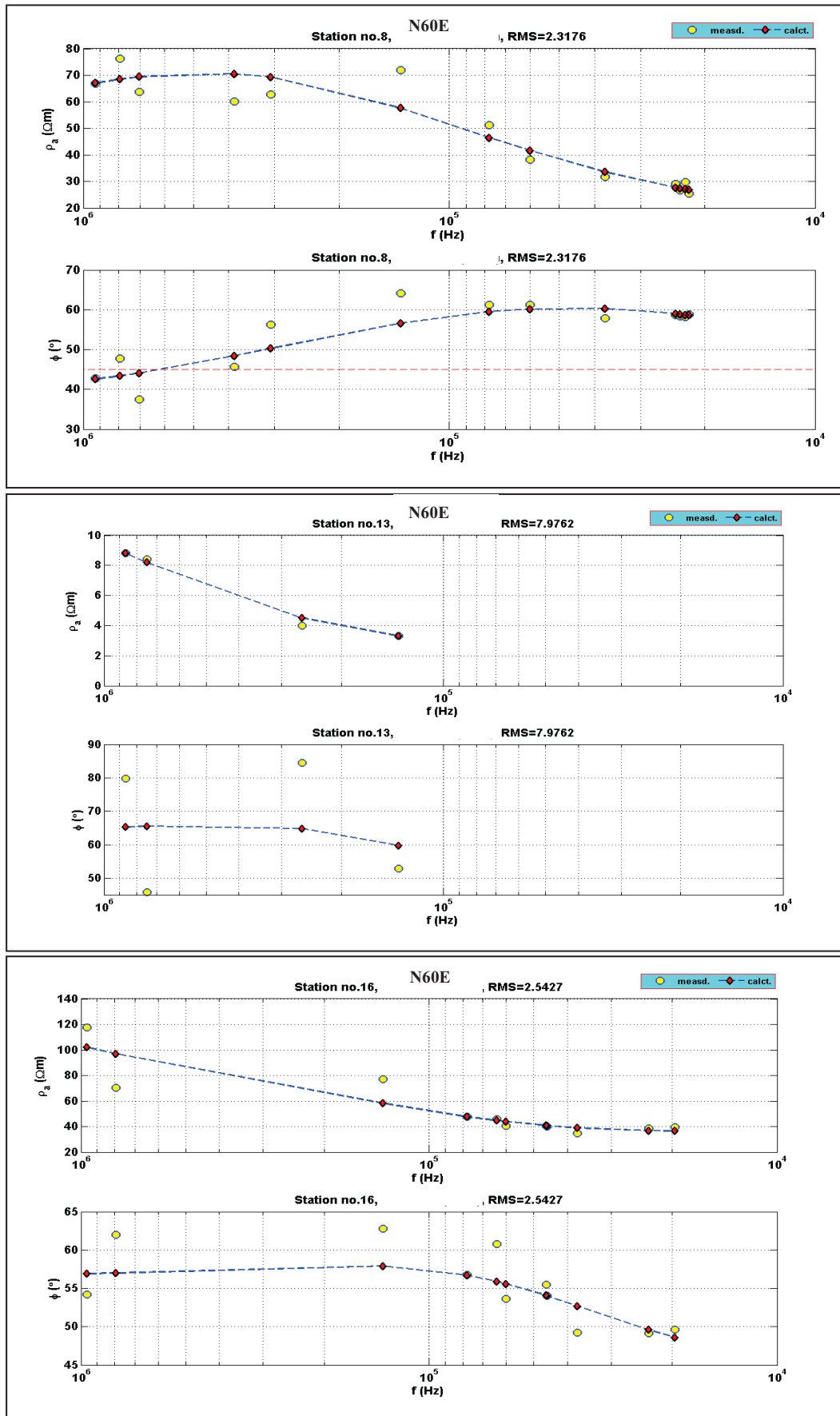


Fig.5. Comparison of measured and calculated RMT data of stations 8, 13 and 16 on profile 2

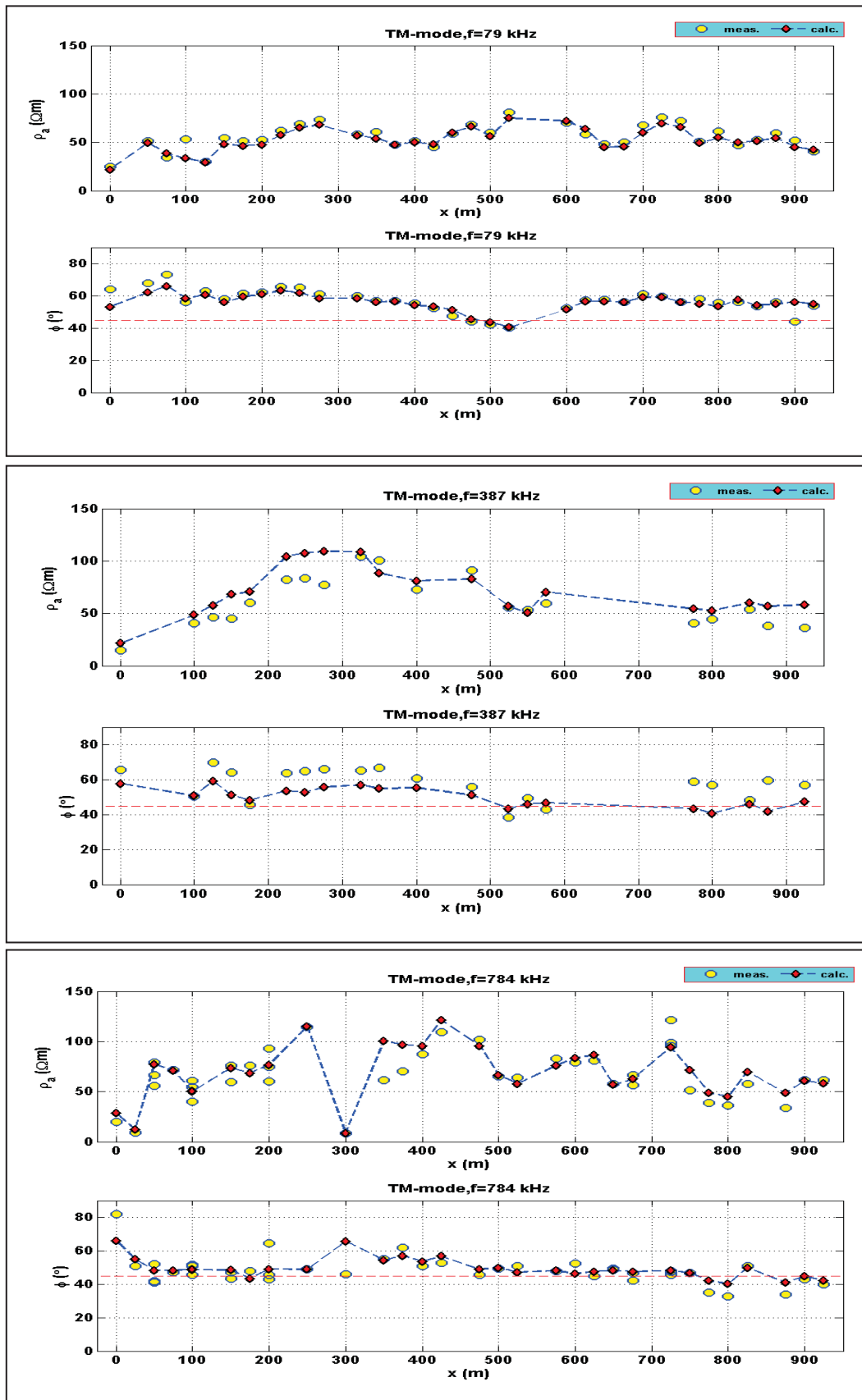


Fig.6. Comparison of measured and calculated RMT data at frequencies 79, 387 and 784 kHz

TEM soundings

In order to overcome the resolution problem of TEM soundings for near surface structures, we carried out RMT soundings at each TEM location to use this additional information in the following joint inversion. The distance between the TEM stations is depending on the accessibility of the area. Thirty loop-loop measurements (Tx: 50 m x 50 m and Rx: 10 m x 10 m) were performed during this campaign. TEM data were obtained using the Zonge NT 20 transmitter and the Zonge GDP32 receiver.

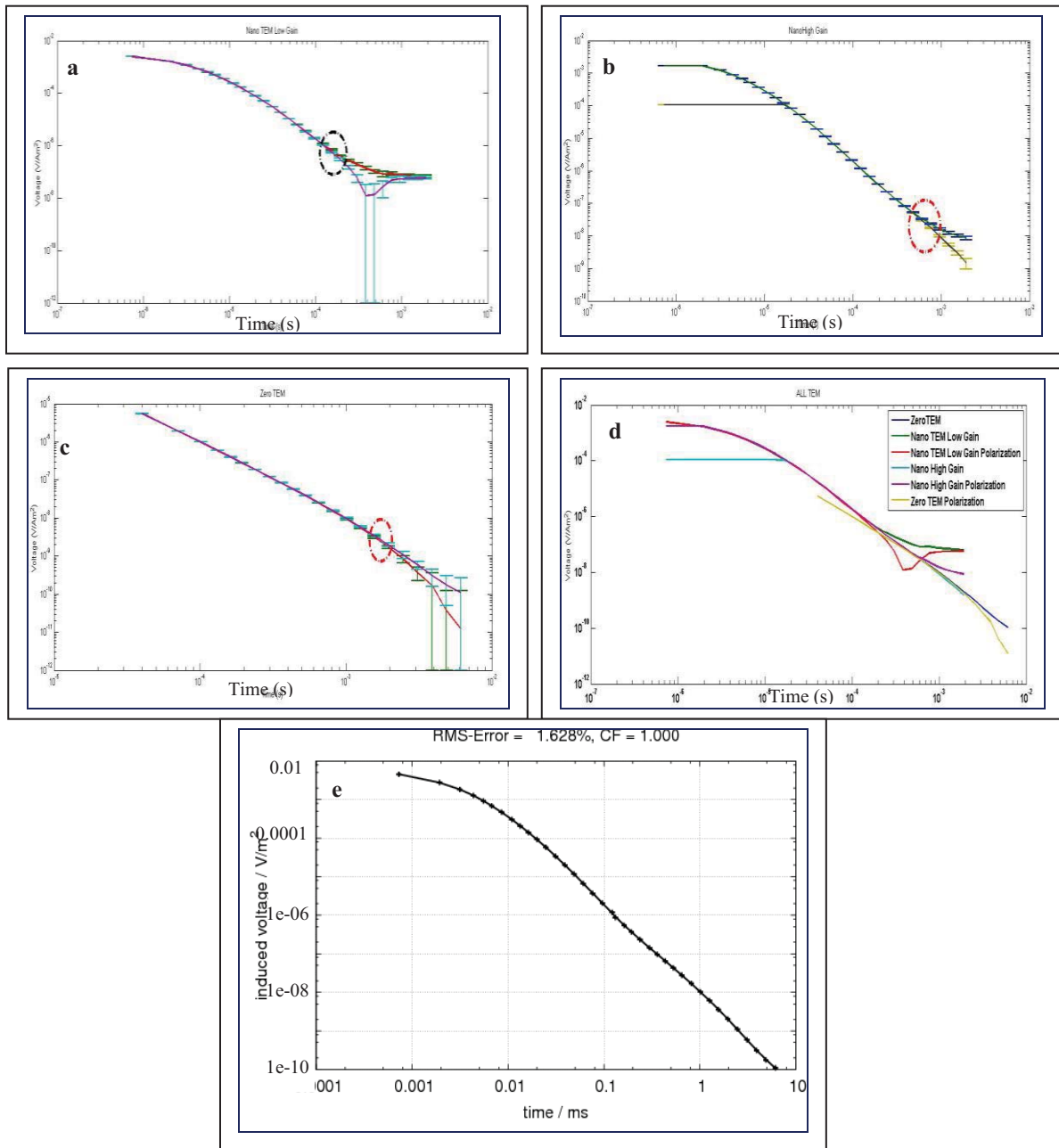


Fig.7. The segmentation recording scheme for the in-loop Zonge manufactured TEM system; low-gain Nano TEM (a), high-gain Nano TEM (b), automatic-gain Zero TEM (c), combination of all-transient (d), TEM sounding of station 1 was processed with deconvolution and inversion (e). The circle indicate unreliable decay curves for very late times that we address to be an A/D conversion problem of the Zonge device.

The system allows to measure TEM data in two distinct modes, according to different investigation depths: Nano TEM with low and high gain and Zero TEM measurements (Fig.7). Both modes give unreliable decay curves for very late times that we address to be an A/D conversion problem of the Zonge device. To cope with this problem we simply changed polarity of the receiver loop. Consequently a time consuming data acquisition procedure was required.

The first measurement for shallow investigation was done in NANO TEM low gain mode. This mode uses low currents (about 0.3 A) which is switched off quickly. Moreover, the ramp time were taken from 0.3 μ s to 2.5 ms with manual low-gain settings. Whereas, Nano TEM high-gain uses a higher transmitter current (about 3 A), which gives a higher penetration depth with a saturation for earlier time. Nano TEM low-gain and high-gain uses 12 Volt power supply. To obtain higher penetration depths, Zero TEM measurements were carried out from 31 μ s to 6 ms time window with accompanying automatic-gain settings. It uses a relatively high transmitter current (10 A) at 24 Volts and 50-55 μ s turn-off ramp time.

The penetration depth of TEM methods depends on the time after the transmitters current is switched off [Parasnis, 1986]. The diffusion process of the transient electromagnetic induction field can be visualized using the smoke ring concept [Nabighian, 1979]. Due to the low conductivity (metamorphic rock), the maximal penetration depth (δ_T) of TEM sounding is approximately 200-300 meter. The next step is a deconvolution of the data that was done by the EADEC algorithm (Lange, 2002) followed by a 1-D inversion with EMUPLUS (e.g. Scholl, 2005) (Fig.7 (e)). The 1-D interpretation of the TEM data also shows the lateral boundary of the fault structure at stations 12- 17 on profile 1 (Fig.8).

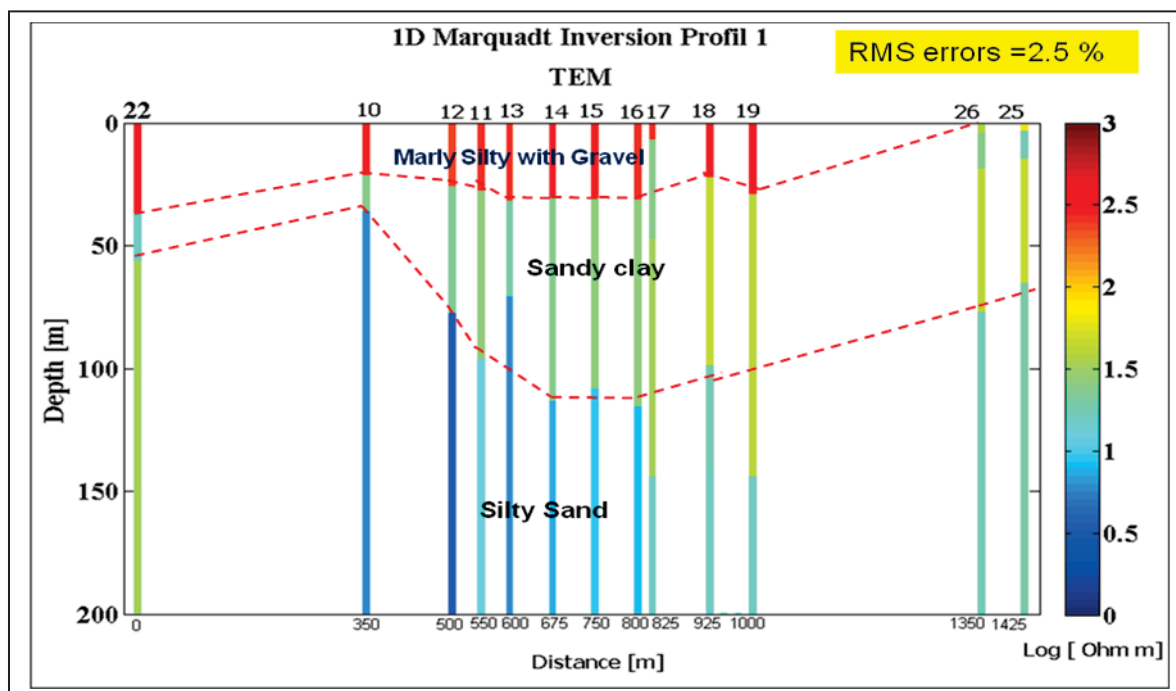


Fig.8. One-dimensional TEM model on profile 1

Joint Interpretation of RMT and TEM data

A joint interpretation can be used to get a better resolution of the model parameter and this will enhance the possibility to detect the active fault structure. Generally we conducted RMT soundings at each TEM site location so that we can estimate a jointly inverted model at each TEM location (Fig.9).

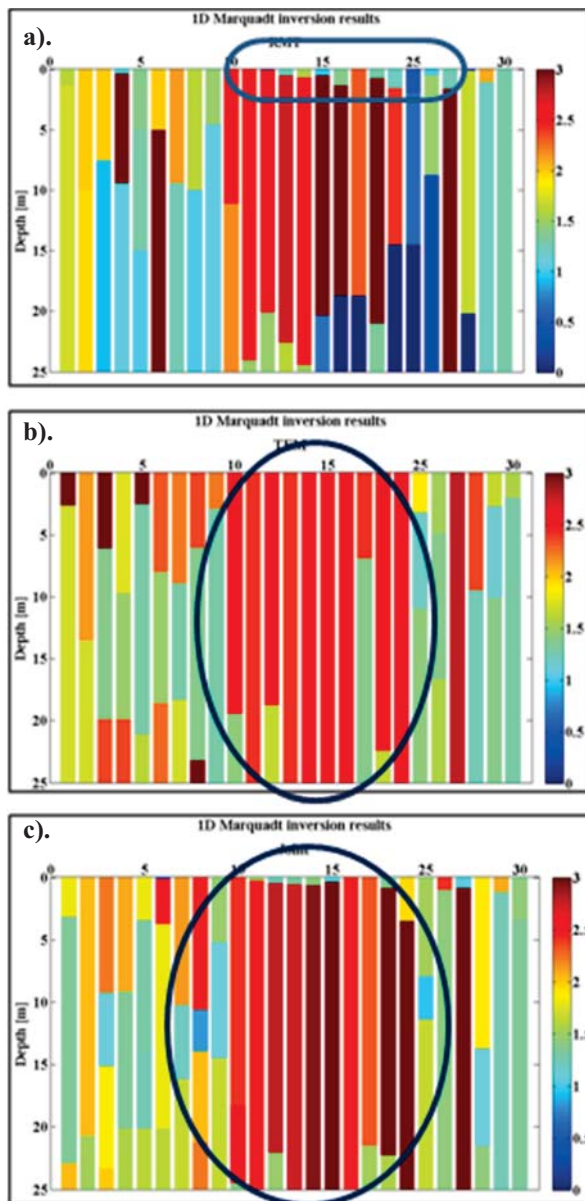


Table2. The importance parameter of station 10

	RMT	Imp	TEM	Imp	Joint	Imp
$\rho_1[\Omega m]$	322.46	0.99	367.98	0.57	247.99	0.99
$\rho_2[\Omega m]$	158.17	0.99	367.98	0.88	344.23	0.27
$h_1[m]$	5.56	0.97	9.75	0.97	9.15	0.84
RMS[%]	4.90		3.10		3.20	

Table 3. Model resistivities obtained from the RMT and TEM data for the selected stations

Station	RMT	TEM	Joint
10	322.46	367.98	344.23
11	332.20	273.45	303.06
12	450.73	372.20	616.36
13	562.36	271.88	531.33
14	363.93	420.46	865.49
17	199.91	260.49	208.14
26	4.47	23.48	26.93
25	28.71	15.83	32.98

Fig. 9. (a) One-dimensional model section derived by RMT data (b). One-dimensional model section of TEM data (c) Joint interpretation of TEM and RMT data. The comparison between TEM, RMT and Joint on station 10, 11,12,14,17 and 25 are shown with ellipses.

The 2-D model of the RMT data (Fig.2) gives information about the top structure of this area, but the deeper structures and the fault system cannot be resolved in detail solely based on RMT soundings. On the other hand, TEM shows a good resolution of deeper structure. The joint interpretation/inversion process was done using the Marquardt algorithm. The interpretation of the 1-D RMT model between station 13 – 25 resolves of the top layer (Fig 9 (a)) and the TEM 1-D model (Fig.9b) shows the bottom of boundary layers of fault structure (Fig.8) in this area more distinctly, however the top layers of TEM stations 13-25 could not be resolved well because the top layer has high resistivity value ($> 80 \Omega \text{ m}$) addressed as metamorphic rock from our reference borehole.

Joint inversion can increase the number of model parameter that includes some of which the methods cannot resolve separately (Fig.9c). The importance parameters were calculated for the model resistivities and thickness [Jupp and Vozoff, 1975] [Tezkan et.al., 1995], [Schwinn, 1999], which are plotted in Table 2. It shows the parameter (ρ_1, ρ_2, h_1) as a result of the individual single/joint inversions. Values between 0 (unimportant) and 1 (important) are calculated. From this analysis we conclude that the resistivity values and thicknesses of the fault structure (Fig.9) is well resolved at station 10, 11,12,14,17 and 25; however, at station 13, it is hardly resolved (importance 0.65). The resistivity value at station 26 is weakly resolved with importance parameter 0.12. The model resistivities of the second layer (between 5-20 m) obtained from 1-D models of TEM and RMT data are in good agreement with the joint inversion models (Table 3).

Conclusion

The Inversion of RMT and TEM data indicates a normal fault structure with a strike direction of approximately N 60 E. The RMT and TEM models generally show four layers that can be separated according to their resistivity values. We address them as to be metamorphic and sediment rocks, which are marly silty sand with gravel ($> 100 \Omega \text{ m}$), marly silty sand with clay (50 - 100 $\Omega \text{ m}$), sandy clay (30 – 50 $\Omega \text{ m}$) and silty sand (10-30 $\Omega \text{ m}$). The RMT data was interpreted using 2-D inversions technique that results in a good fitting between observed and calculated data. Due to the high resistivity of the top layer, the skin depths of the RMT soundings are around 35 m. The TEM data gives detail information of the lower structure down to a depth of 200 m.

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Die deutsche Nordseeküste im Fokus von aeroelektromagnetischen Untersuchungen

Teilgebiete Langeoog mit Wattenmeer und Elbemündung

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1. Einleitung

Hubschrauberelektromagnetische (HEM) Untersuchungen bieten ein großes Potential für die flächen-deckende Kartierung der Sedimente der ersten hundert Meter des Untergrundes (Siemon *et al.*, 2009). Sie sind für hydrogeologische Fragestellungen von großer Bedeutung, da mit Hilfe des spezifischen Widerstands die Verteilung sandiger und tonhaltiger Sedimente im Untergrund sowie Versalzungs-zonen und Süßwasserbereiche ermittelt werden können.

In den Jahren 2008 und 2009 wurden in Kooperation von LIAG und BGR im Rahmen des LIAG-Projektes zur „Flächenhaften Befliegung“ und des „D-AERO“-Projektes der BGR gemeinsam aero-geophysikalische Erkundungen zu Salz-/Süßwasserfragestellungen in insgesamt fünf Messgebieten im Bereich der deutschen Nordseeküste durchgeführt (Wiederhold *et al.* 2008, Steuer *et al.* 2009). Das Messgebiet Langeoog umfaßt die Ostfriesischen Inseln Langeoog und Spiekeroog und das an-grenzende Wattenmeer, das Messgebiet Glückstadt umfaßt den Bereich der Elbemündung nordwest-lich von Hamburg (Abb. 1).

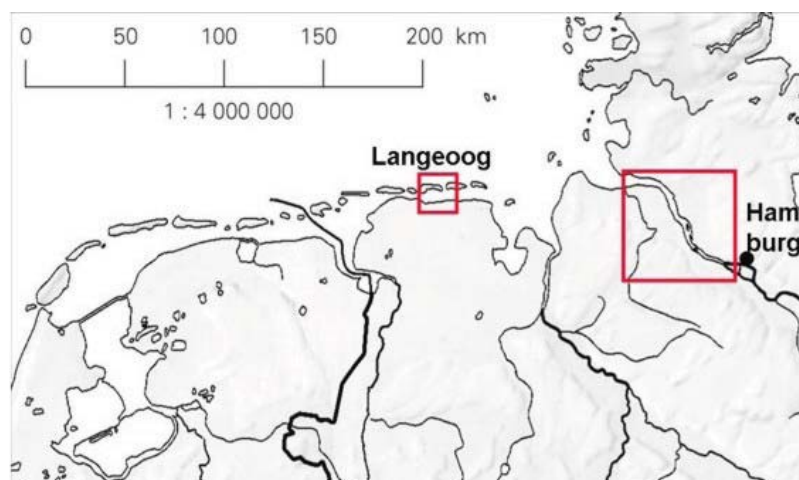


Abbildung 1: Lage der Messgebiete Langeoog und Glückstadt (nordwestlich von Hamburg).

Für die Insel Langeoog waren die Lage und die Ausdehnung der Süßwasserlinse zu ermitteln. Im küstennahen Bereich erhoffte man im Wattenmeer mutmaßliche Süßwasseraustritte zu finden. Im Bereich der Elbemündung sollten die grundwasserführenden Schichten und mögliche Versalzungszone kartiert werden.

Die Daten dienen als Grundlage für die Planung und Arbeit in vielfältigen ökonomischen und ökologischen Bereichen, wie z.B. Raumplanungen und der Entwicklung von Wassernutzungs- und Wasserschutzkonzepten. Sie werden über das Fachinformationssystem (FIS) Geophysik des LIAG nutzbar sein.

Das eingesetzte Messsystem wird bei *Steuer et al. (2010)* beschrieben. Die Karten des scheinbaren spezifischen Widerstands ρ_a und der Schwerpunktstiefe z^* (*Siemon, 2009*) für die verschiedenen Messfrequenzen geben einen ersten Überblick über die Leitfähigkeitsstrukturen in den Messgebieten. Die Erkundungstiefe nimmt dabei mit abnehmender Frequenz und Leitfähigkeit zu. Die HEM-Messungen liefern die Verteilung der elektrischen Leitfähigkeit bis maximal 150 m Tiefe. Dabei können Salzwasser und Süßwasser oder ton- und sandhaltige Sedimente unterschieden werden. Hier dargestellt sind erste Ergebnisse in Form von Karten auf dem Bearbeitungsstand nach der Grundprozessierung, bei der Korrekturfaktoren und Filtereinstellungen für jeden Flug gleich angesetzt werden. Die darauf folgende Feinprozessierung behandelt jeden Flug und jede Frequenz individuell. Die Prozessierung der Daten beinhaltet: Filterung, Sprungkorrektur, Nullniveauekorrektur, Feinjustierung der Kalibrierfaktoren und Berechnung der Halbraumparameter.

2. Messgebiet Langeoog

Langeoog und Spiekeroog sind zwei der Ostfriesischen Inseln im Wattenmeer der Nordsee, die sich wenige Kilometer vor der Küste befinden. Sie bestehen aus quartären Sedimenten wie Sanden, Tonen und Schluffen (Abb. 3). Durch Versickerung der Niederschläge in den Dünengürteln werden dort Grundwasserreservoirs mit Süßwasser aufgefüllt, die die Versorgung der Inseln mit Frischwasser sicherstellen. Weil Süßwasser ein geringeres spezifisches Gewicht als das versalzte Grundwasser hat, schwimmt es als „Linse“ auf dem Salzwasser. Verändert sich der Wasserstand des dichteren Salzwassers, ändert sich das Druckniveau und entsprechend auch Stand und Ausdehnung der Süßwasserlinse. Durch Stürme können Salzwassereinträge bis an den Rand des inneren Dünengürtels der Inseln gelangen und dort versickern. Solche Ereignisse, aber auch der erhöhte Wasserbedarf für den Tourismus in den Sommermonaten gefährden das Süßwasserreservoir. Noch erfolgt die gesamte Trinkwasserversorgung mit insoleigenem Grundwasser, welches in den Wintermonaten allein durch Regenfälle wieder aufgefüllt wird. Unterhalb wasserundurchlässiger Bodenschichten kann Süßwasser darüber hinaus vom Festland aus unterschiedlich weit in das Wattenmeer vordringen, um dann unter gewissen Bedingungen an die Oberfläche zu treten. Die Befliegungen wurden bei Niedrigwasser durchgeführt, um vergleichbare Bedingungen für jeden Flug zu haben und insbesondere auch, um den Untergrund des Wattenmeeres ohne die bei Flut vorhandene Meerwasserschicht besser erkun-

den zu können. Aufgrund der Berücksichtigung von Naturschutzauflagen konnte die aerogeophysikalische Erkundung nur in den Wintermonaten durchgeführt werden. Das Messgebiet umfasst die gesamte Insel Langeoog, den westlichen Teil der Insel Spiekeroog und das Wattenmeer zwischen Langeoog und dem Festland (Abb. 2). Einen Überblick über die Messkampagne gibt Tabelle 1.

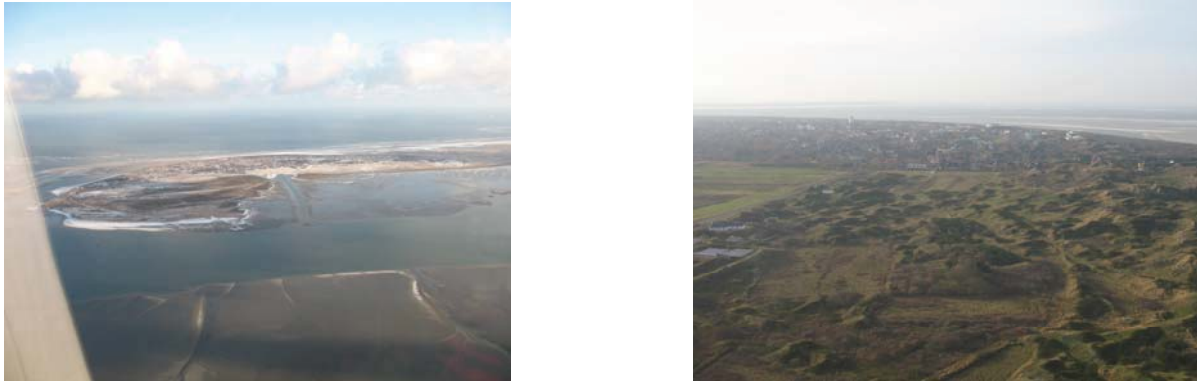


Abbildung 2: Insel Spiekeroog mit Wattenmeer während der Befliegung im März 2009 und westlicher Teil von Langeoog mit der Dünenlandschaft, unter der sich Süßwasserlinsen verbergen (Fotos: W. Voß).

Tabelle 1:

Größe des Messgebietes:	259 km ²
Gesamtprofillänge:	1200 km
Linienabstand:	67 Mess-Linien mit 250 m (N-S), 7 Kontroll-Linien mit 2000 m (W-O)
Messzeitraum:	Februar und November 2008, Februar und März 2009; nur bei Niedrigwasser

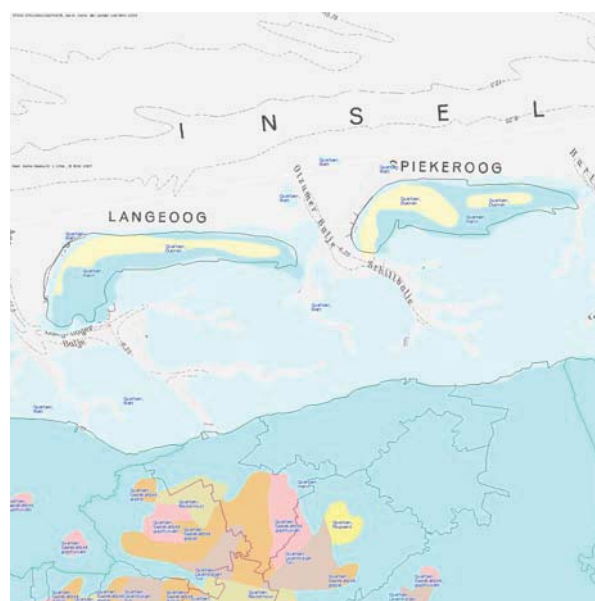


Abbildung 3: Geologische Karte des Messgebietes aus dem FIS Geophysik des L/AG.

Abbildung 4 zeigt die vorläufigen Karten des scheinbaren spezifischen Widerstands (mit ρ_a bis maximal 2000 Ωm) der verschiedenen Messfrequenzen (133400 Hz, 41520 Hz, 8390 Hz, 5345 Hz, 1823 Hz und 387 Hz) für das Messgebiet. Jeder Frequenz wird dabei zu den einzelnen Messpunkten eine Schwerpunkstiefe z^* zugeordnet, die in Abbildung 5 dargestellt ist. Für die höchsten Frequenzen sind dies einige wenige Meter unter der Erdoberfläche, die tiefsten Frequenzen erreichen eine Erkundungstiefe bis etwa 100 m. Diese Erkundungstiefen sind neben der Frequenz auch von den spezifischen Widerständen der Untergrundstrukturen abhängig und können somit von Messpunkt zu Messpunkt variieren. Die Abfolge einer Auswahl von Karten des scheinbaren spezifischen Widerstands in einer 3D-Darstellung gibt daher eine Abbildung der Leitfähigkeitsstrukturen mit Bezug zur Tiefe nur bedingt wieder (Abb. 7). Die tiefste Frequenz bildet die Basis der Süßwasserlinse ab.

Deutlich sind die den Süßwasserlinsen zugeordneten Strukturen auf der Insel und die mit salzhaltigem Nordseewasser gefüllten Priele zu erkennen. Im Bereich vor der Festlandküste kann man Gebiete mit für Süßwasser typischen Werten der scheinbaren Widerstände erkennen. Hier ist zu klären, ob dies Austritten von Süßwasser im Watt entspricht.

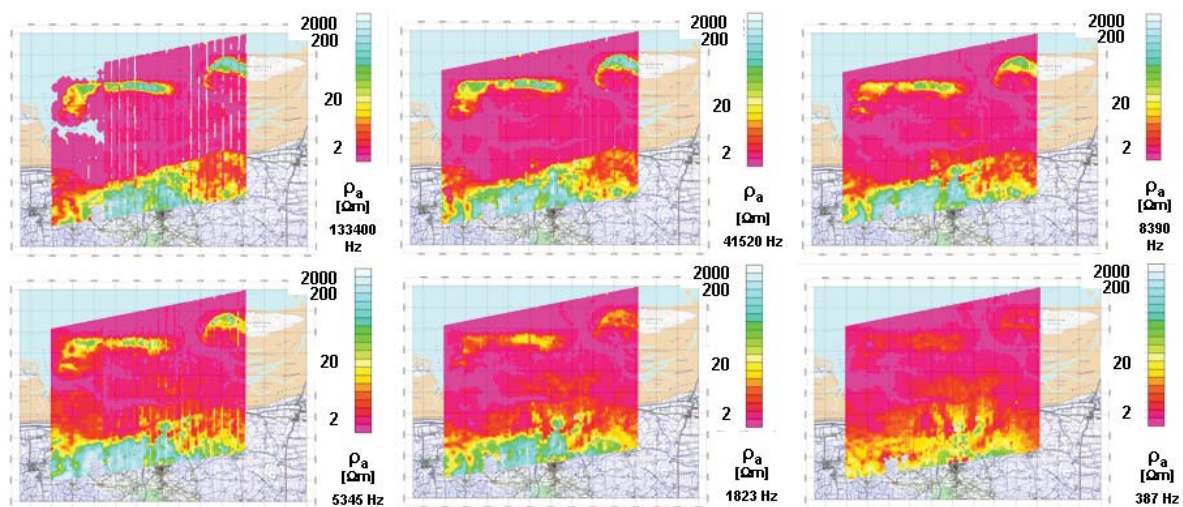


Abbildung 4: Vorläufige Karten des scheinbaren spezifischen Widerstands ρ_a für die Messfrequenzen 133400 Hz, 41520 Hz, 8390 Hz, 5345 Hz, 1823 Hz und 387 Hz (ρ_a bis maximal 2000 Ωm).

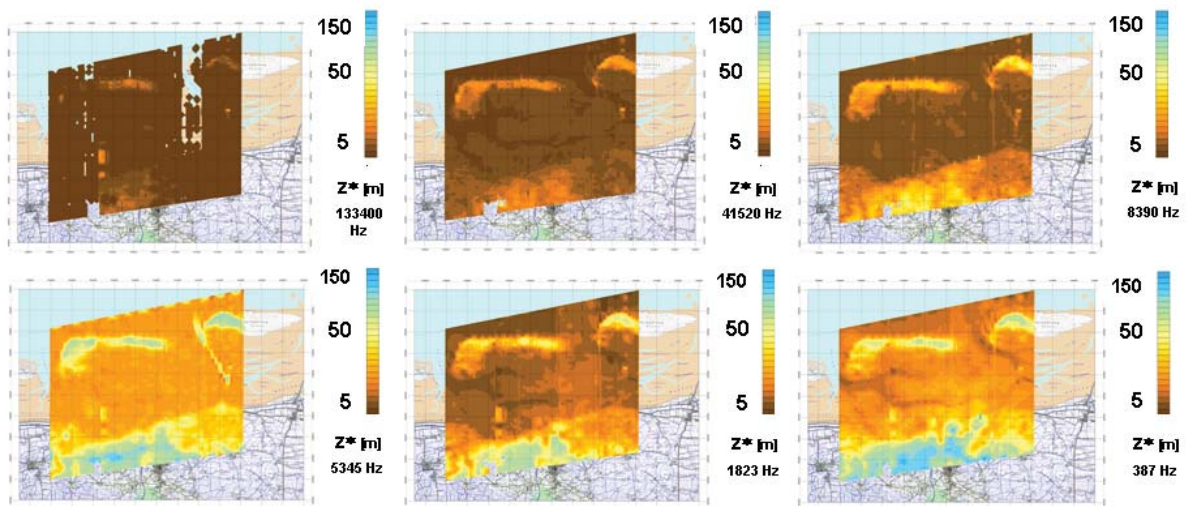


Abbildung 5: Vorläufige Karten der Schwerpunkstiefe z^* für die Messfrequenzen 133400 Hz, 41520 Hz, 8390 Hz, 5345 Hz, 1823 Hz und 387 Hz.

Benutzt man einen anderen Farbkeil für die Darstellung der scheinbaren spezifischen Widerstände (mit ρ_a bis maximal 200 Ωm), kann man beispielsweise die Priele noch viel deutlicher erkennen, siehe Abbildung 6. Auch die mutmaßlichen Austritte von Süßwasser im Bereich vor der Festlandküste sind wesentlich besser aufgelöst.

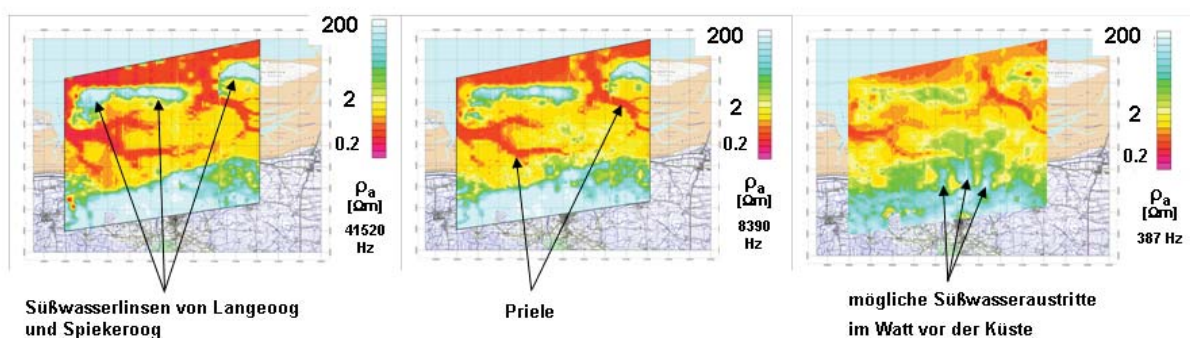


Abbildung 6: Vorläufige Karten des scheinbaren spezifischen Widerstands ρ_a für drei ausgewählte Frequenzen (41520 Hz, 8390 Hz und 387 Hz), dargestellt mit einem anderen Farbkeil (ρ_a bis maximal 200 Ωm).

In Abbildung 8 wird die topographische Karte mittels des digitalen Höhenmodells (*digital elevation model*, DEM) aus dem FIS Geophysik des LIAG gezeigt. Dabei wurden die Daten des Höhenmodells auf die Fluglinien übertragen. Die topographische Karte kann auch aus GPS- und Laseraltimeterdaten des Hubschraubermesssystems abgeleitet werden.

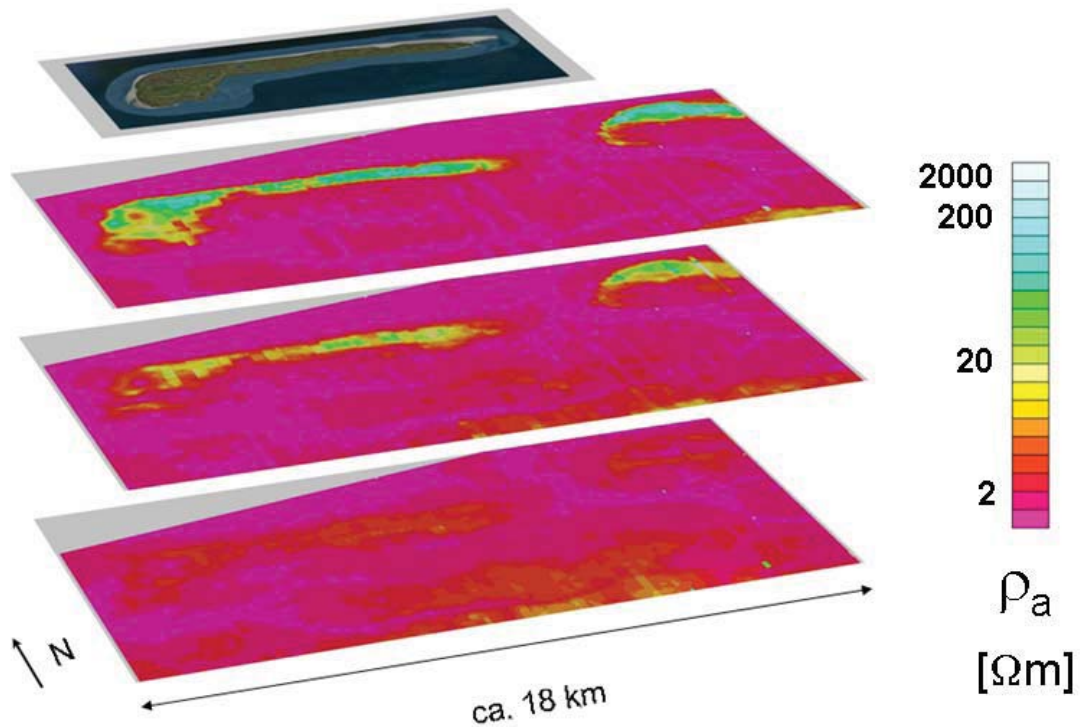


Abbildung 7: 3D-Darstellung der Karten des scheinbaren spezifischen Widerstands ρ_a für die Frequenzen 41520 Hz, 8350 Hz und 387 Hz. Bild aus Google-Earth, September 2009.

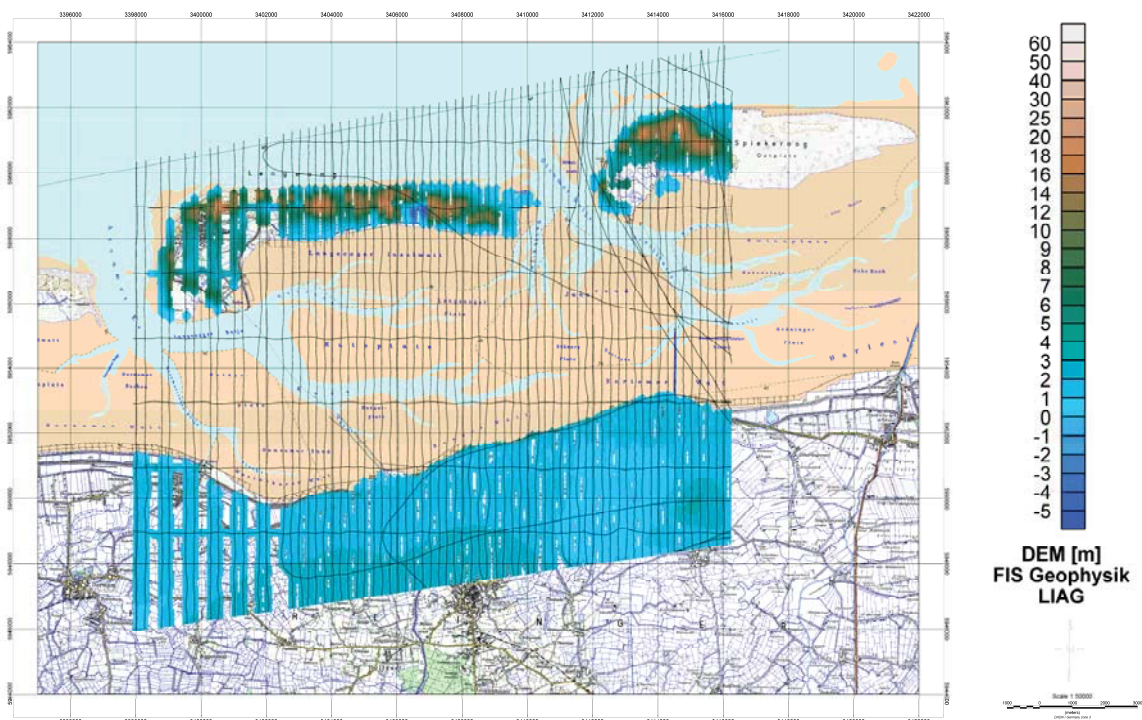


Abbildung 8: Topographische Karte mittels des digitalen Höhenmodells (*digital elevation model*, DEM) aus dem FIS Geophysik des LIAG. Dabei wurden die Daten des Höhenmodells auf die Fluglinien übertragen.

3. Messgebiet Glückstadt

Das Messgebiet Glückstadt umfasst große Teile des Elbemündungsbereichs (Abb. 9). Teile der Ergebnisse der HEM-Befliegung werden in das Projekt *KLIMZUG-NORD* (www.klimzug-nord.de) einfließen, bei dem unter anderem die Auswirkungen des Klimawandels auf das Ästuar der Elbe erforscht werden. Es ist davon auszugehen, dass sich der Elbwasserstand in diesem Gebiet erhöht, was eine Intrusion von Brackwasser in den Süßwasseraquifer nach sich ziehen würde. Mit hydraulischen Strömungsmodellen können solche Fragestellungen untersucht werden. Bereits vorhandene geologische Informationen sollen mit Hilfe von elektrischen Leitfähigkeitsmodellen der HEM flächenhaft zu einem geologischen Strukturmodell ergänzt werden. Dieses wird mit hydraulischen Parametern belegt und dient damit als Grundlage für eine hydraulische Modellierung.



Abbildung 9: Flugplan für das Messgebiet Glückstadt, welches große Teile des Elbemündungsbereichs umfasst.

Einen Überblick über den Umfang der Messkampagne gibt Tabelle 2.

Tabelle 2:

Größe des Messgebietes:	1949 km ²
Gesamtprofillänge:	8500 km
Linienabstand:	209 Mess-Linien mit 250 m (W-O), 19 Kontroll-Linien mit 2500 m (N-S)
Messzeitraum:	Juli und Oktober 2008, März-Mai 2009

Aus den GPS- und den Laseraltimeterdaten des Hubschraubermesssystems kann ein digitales Höhenmodell abgeleitet werden (Abb. 10). Hier wird besonders der Verlauf der Geest (Höhenzug bestehend aus pleistozänen Sandablagerungen von Moränen) und der Marsch (flaches holozänes Schwemmland) deutlich.

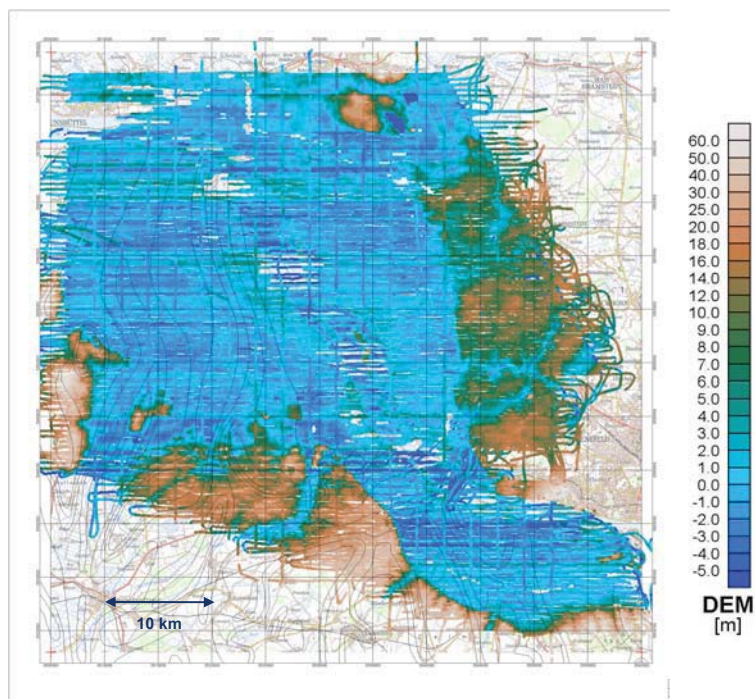


Abbildung 10: Digitales Höhenmodell des Elbemündungsgebietes, abgeleitet aus GPS- und Laseraltimeterdaten des Hubschraubermesssystems.

Erste Ergebnisse in Form von vorläufigen Karten des scheinbaren spezifischen Widerstands und der Schwerpunkstiefe sind in Abbildung 11 dargestellt. Die Datenlücken sind teilweise auf Hochspannungsleitungen und entlang der Elbe auf Radarstationen bzw. Sperrzonen um die Atomkraftwerke herum zurückzuführen.

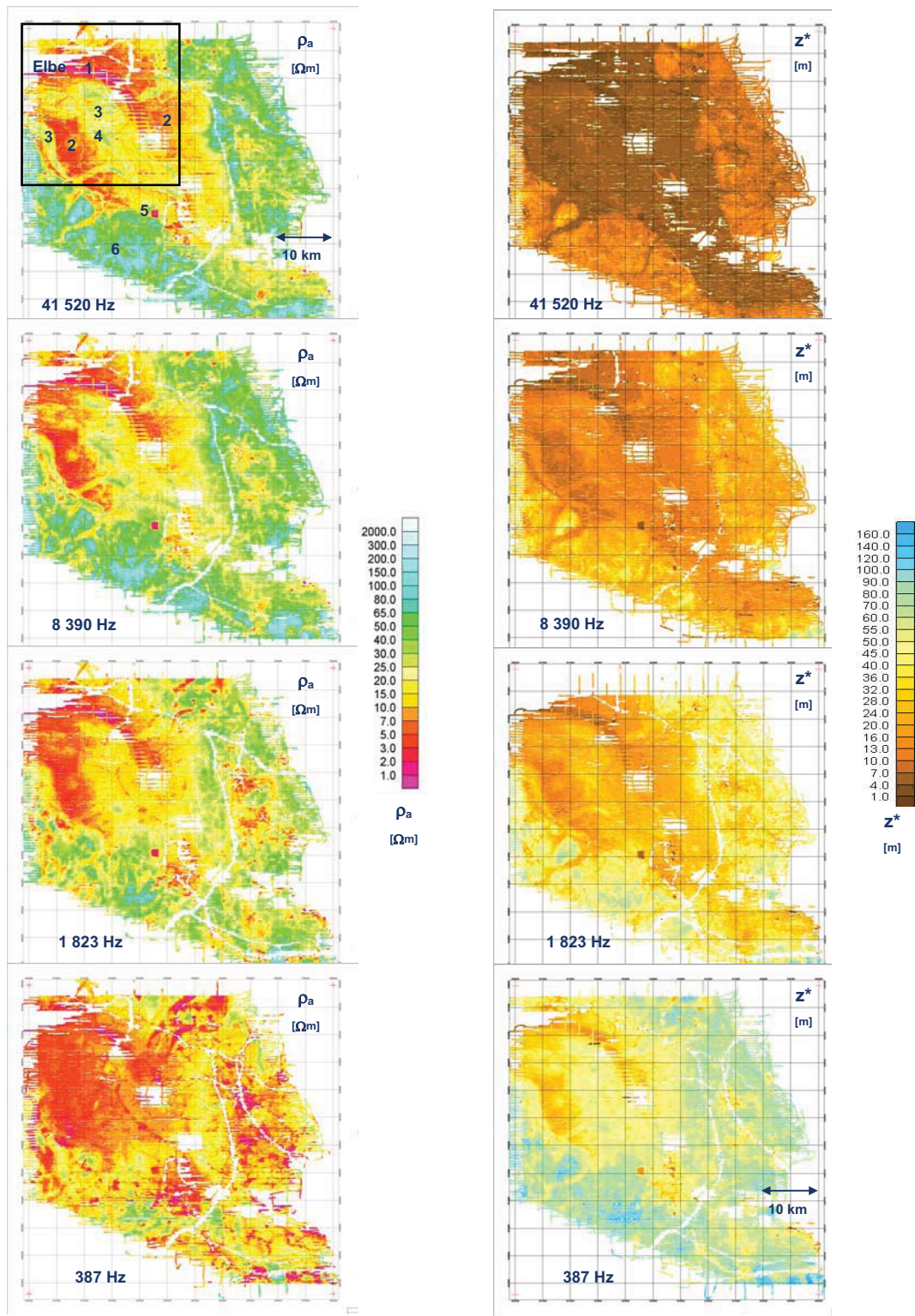


Abbildung 11: Vorläufige Karten des scheinbaren spezifischen Widerstands ρ_a und der Schwerpunkstiefe z^* für vier ausgewählte Messfrequenzen.

Mit Hilfe der geologischen Karte vom Mündungsbereich der Elbe (Abb. 12a) können z.B. in der Karte des scheinbaren spezifischen Widerstands zu der Frequenz 41 kHz (Abb. 12b) folgende Leitfähigkeitsstrukturen identifiziert werden: 1. Brackwasser der Elbe, 2. Perimarine Ablagerungen: Ton, schluffig, 3. Brackische Ablagerungen: Ton bis Schluff, feinsandig, 4. Moor: Torfe, 5. Rotschlammdeponie: Eisen- und Titanoxide, 6. Glazifluviale Ablagerungen: Sand und Kies.

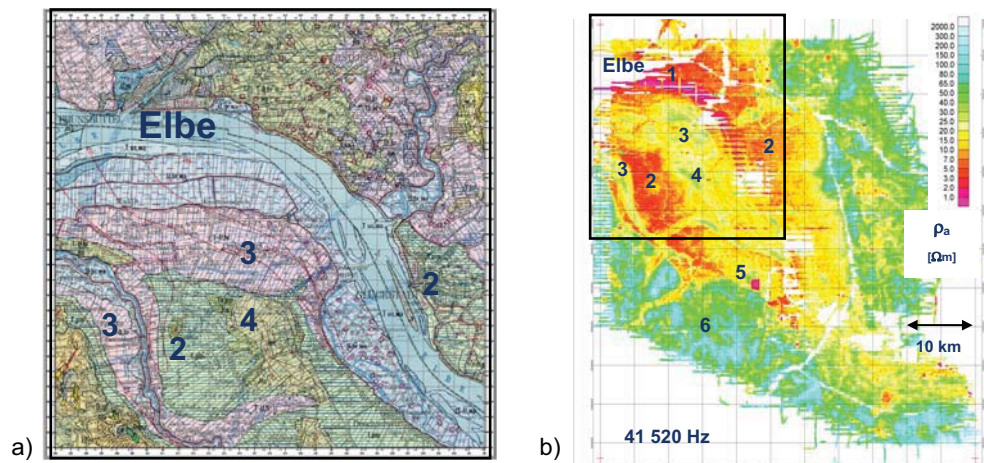


Abbildung 12: a) Geologische Karte und b) Karte des scheinbaren spezifischen Widerstands der Frequenz 41 kHz. Im Mündungsbereich der Elbe können folgende Leitfähigkeitsstrukturen identifiziert werden: 1. Brackwasser der Elbe, 2. Perimarine Ablagerungen: Ton, schluffig, 3. Brackische Ablagerungen: Ton bis Schluff, feinsandig, 4. Moor: Torfe, 5. Rotschlammdeponie: Eisen- und Titanoxide, 6. Glazifluviale Ablagerungen: Sand und Kies.

4. Zusammenfassung und Ausblick

In beiden Messgebieten wurde die HEM zur Erkundung von Grundwasserstrukturen erfolgreich eingesetzt. Die hier gezeigten Karten dokumentieren die vorläufigen Ergebnisse. Nach Abschluss der Prozessierung werden die HEM-Daten in Schichtmodelle des spezifischen Widerstands invertiert.

Auf den Ostfriesischen Inseln Langeoog und Spiekeroog konnten die Lage und Ausdehnung der Süßwasserlinsen erfasst werden. Daneben zeigten sich insbesondere in den Daten der tiefsten Messfrequenz Bereiche veränderter scheinbarer spezifischer Widerstände im Wattenmeer vor der Festlandküste, die die Vermutung nahe legen, dass hier Süßwasser austritt.

Im Gebiet Glückstadt werden als Fragestellung mögliche Intrusionen von Brackwasser in den Süßwasseraquifer untersucht. Die bereits vorhandenen geologischen Informationen werden mit Hilfe der HEM-Modelle flächenhaft zu einem geologischen Strukturmodell ergänzt. Nach einer Belegung mit hydraulischen Parametern kann es damit als Grundlage für eine Interpretation mit Strömungsmodellen dienen.

Danksagung

Die hier vorgestellten Teilgebiete des *BGR/LIAG*-Projektes wurden in Zusammenarbeit mit dem *Institut für Chemie und Biologie des Meeres (ICBM)*, der *Universität Oldenburg* sowie der *Behörde für Stadtentwicklung und Umwelt Hamburg* durchgeführt.

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<http://www.fis-geophysik.de/>

The German North Sea Coast in Focus of Airborne Electromagnetic Investigations: The Freshwater Lenses of Borkum

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Abstract

Helicopter-borne electromagnetic (HEM) measurements were conducted over the North Sea island of Borkum to determine the size of the freshwater lenses. Additionally, geoelectrical measurements were carried out at locations where data of 13–17 years old *Schlumberger* soundings exist. HEM successfully revealed the lateral extension as well as the thickness of the freshwater lenses of the island of Borkum. At several locations the depth of the freshwater/saltwater boundary determined by HEM was confirmed by the results of the Schlumberger soundings. No significant changes of the freshwater/saltwater boundary were found at the locations where old and new *Schlumberger* soundings exist. Regarding the results of direct-push and borehole measurements, the HEM related resistivity structure could be determined more precisely and additional to the freshwater/saltwater boundary thin clay layers were revealed. Based on the HEM inversion results the volume of the freshwater resource was estimated.

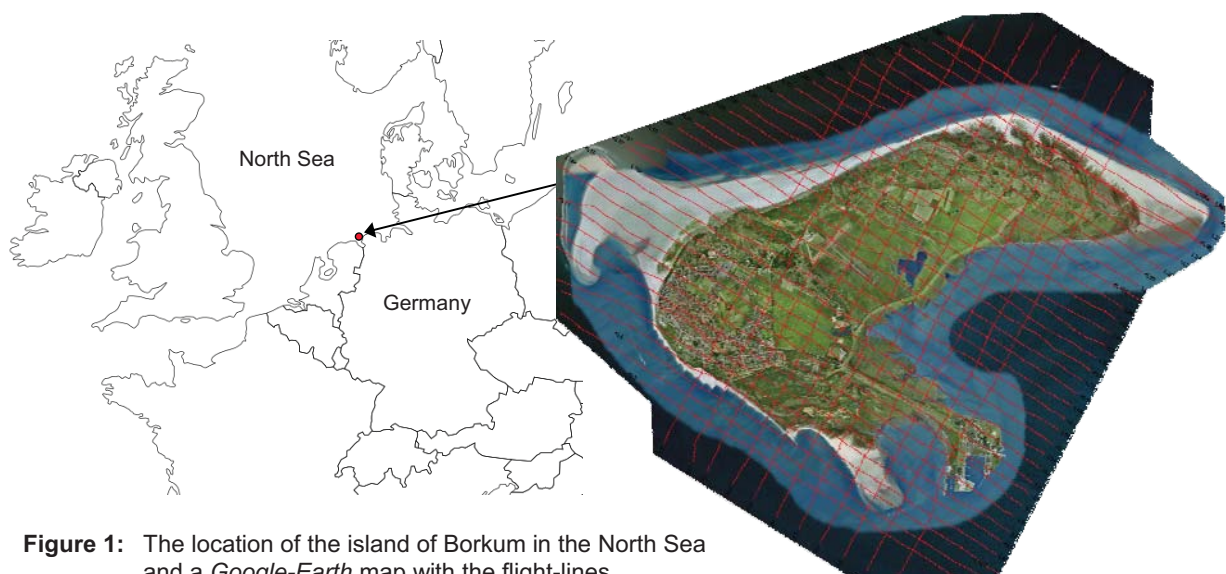


Figure 1: The location of the island of Borkum in the North Sea and a Google-Earth map with the flight-lines.

Introduction

Since 2008, the German North Sea Coast has been in the focus of airborne geophysical investigations carried out by the *Federal Institute for Geosciences and Natural Resources (BGR)* and the *Leibniz Institute for Applied Geophysics (LIAG)*. Especially electromagnetics is the most versatile of the airborne geophysical methods and widely applied in hydrogeological investigations as the measurements respond to both lithologic and water-chemistry variations. The applications comprise geologic mapping and aquifer structure, delineation of soil and groundwater salinization, saltwater intrusion into coastal aquifers etc.

Here, we present the HEM results for the North Sea island of Borkum, which was the first of the five HEM project areas investigated in 2008-2009, and compare them with DC *Schlumberger* soundings, lithological logs and direct-push measurements. These data will be used for a hydraulic modelling of the freshwater lenses of the island of Borkum in the *CLIWAT* project (<http://cliwat.eu>, *EU Interreg IVB North Sea Region*). Additionally we show an estimation of the volume of the freshwater lenses calculated by *Siemon et al. (2009b)*.

Other HEM results for the Wadden Sea, the North Sea islands of Langeoog and Spiekeroog, and the Elbe estuary are shown by *Schaumann et al. (2010)* in this volume. *Wiederhold et al. (2009)* presented the results of two *SkyTEM* surveys at the North Sea island of Föhr and the saltdome Bad Segeberg, which were conducted in 2008 by *SkyTEM Aps* by order of *LIAG*.

HEM Survey

The Borkum survey covers an area of 88 km² and was flown within two days in March 2008. The flight-line spacing was 250 m and the tie-line spacing was 500 m, totalling to about 412 line-kilometres (Figure 1). The *BGR* helicopter-borne geophysical system simultaneously uses frequency-domain EM, magnetics and radiometrics (Figure 2).

The HEM system, a *RESOLVE* bird manufactured by *Fugro Airborne Surveys*, operates at six frequencies ranging from 386 Hz to 133 kHz.

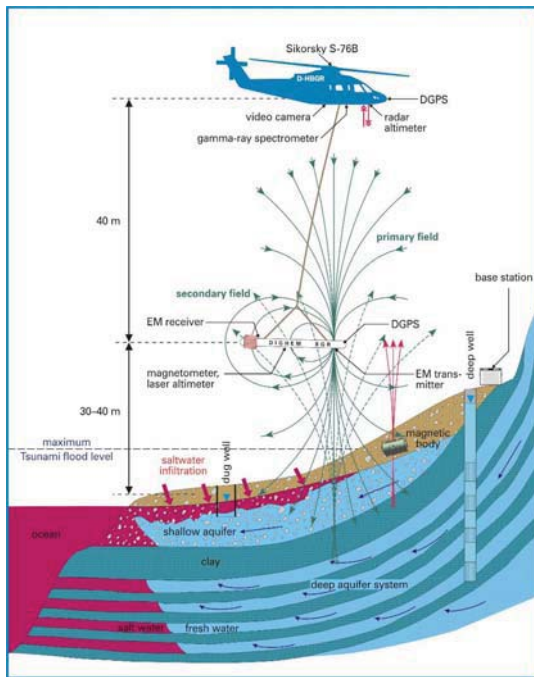


Figure 2: BGR helicopter-borne geophysical system:

Electromagnetics:

RESOLVE (Fugro Airborne Surveys)

Frequencies: 386, 1 823, 5 495, 8 338, 41 430 and 133 300 Hz

Coil separations: ~8 m

Coil orientation: 5× horizontal coplanar
1× vertical coaxial

Investigation depth: 1–70 m (at Borkum)

Magnetics:

Cs magnetometer

Radiometrics:

256 channel gamma-ray spectrometer

Navigation/Positioning:

GPS / radar and laser altimeter

Bird altitude: ~34 m

Flight speed: ~140 km/h

Sampling distance: ~4 m

The investigation depth increases with decreasing frequency. The EM fields of the lowest frequency penetrate the freshwater lenses nearly completely as seen in the apparent resistivity maps (Figure 3). Here, higher resistivities are represented by green and blue colours (freshwater) and lower resistivities by red colours (saltwater). The apparent resistivity maps, which are based on the half-space approximation, provide a first insight into the subsurface conductivity distribution.

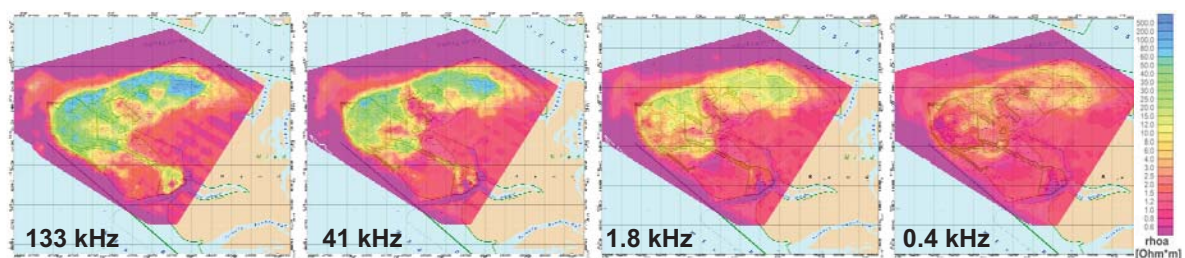


Figure 3: Apparent resistivity maps: The investigation depth increases with decreasing frequency. The freshwater lenses – represented by higher resistivities (green & blue) – are nearly completely penetrated by the EM fields of lowest frequency.

The HEM data were also inverted using *Marquardt-Levenberg* 1-D inversion technique based on a layered half-space model (*Siemon et al., 2009a*). Starting models required were derived automatically from the apparent resistivities vs. centroid depth sounding curves. Resistivity maps at selected depths and stitched together vertical resistivity sections were derived from these 1-D inversion models. The lateral extensions as well as the thicknesses of the freshwater lenses were mapped successfully with HEM, as shown in the resistivity maps and the vertical resistivity sections of Figure 4.

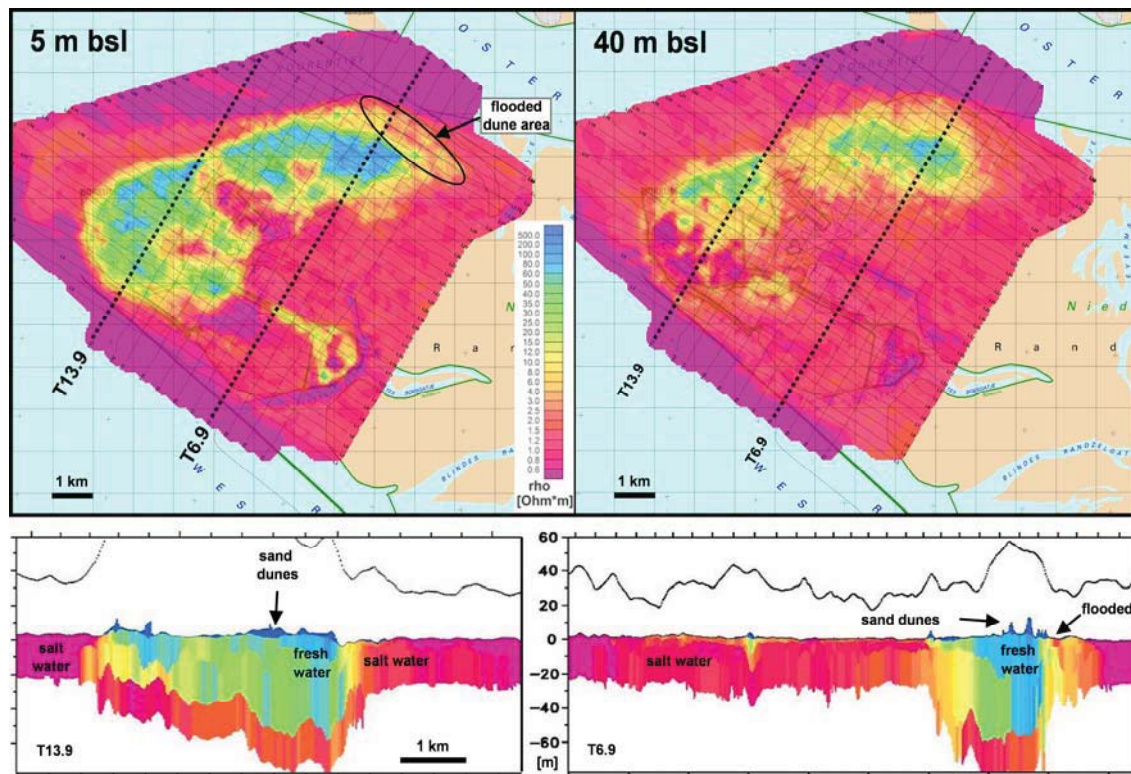


Figure 4: Resistivity maps at 5 and 40 m bsl and vertical resistivity sections along the tie lines T13.9 and T6.9.

The first layer represents the resistive dry dune sands above the groundwater table (dark blue) and the bottom layer the conductive saltwater-filled sediments (red) below the freshwater lenses (blue). The two layers in-between are associated with sediments filled with fresh or brackish water. The freshwater/saltwater boundary was detected down to about 60 m depth.

DC survey: Comparison of measurements 16 years apart

In 2008, 36 *Schlumberger* soundings were carried out at sites already measured between 1991 and 1995 (Worzyk, 1995a and 1995b).

All the sounding curves at Borkum have a Q-type shape with three or more layers (e.g. Figure 5). Higher resistivities are due to dry and freshwater filled sediments, whereas the lower resistivities at the end of the sounding curves are due to saltwater-filled sediments underlying the freshwater lenses.

In 1992, the freshwater filled sediments were associated with resistivities between 70 and 110 Ωm in the water catchment area

Ostland; these resistivities were almost reproduced in 2008. The underlying saltwater-filled sediments showed resistivities below 10 Ωm , but these values were not determined exactly. Taking equivalent models into consideration the thickness of the freshwater lens can be determined with an accuracy of about 6–8 m.

No significant changes of the freshwater/saltwater boundary were found at the locations where old and new *Schlumberger* soundings exist.

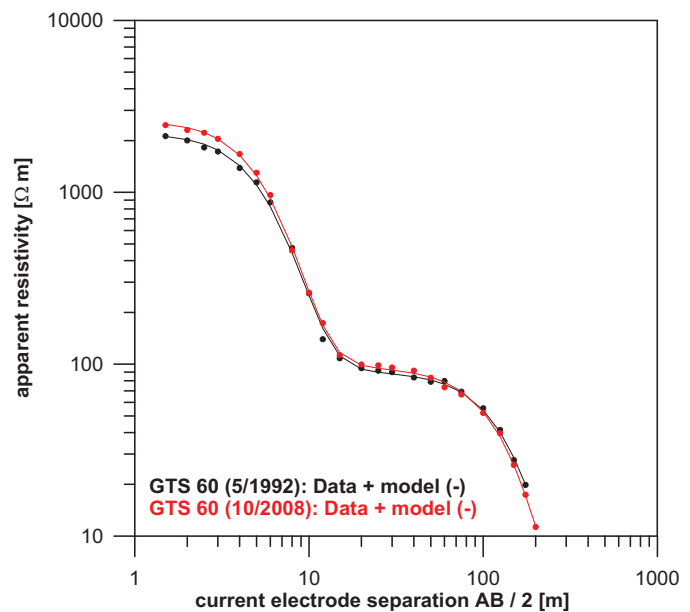


Figure 5: DC sounding curve of site GTS 60.

Combined analysis

At several locations the depth of the freshwater/saltwater boundary determined by HEM was confirmed by the results of the *Schlumberger* soundings (e.g., Figure 6).

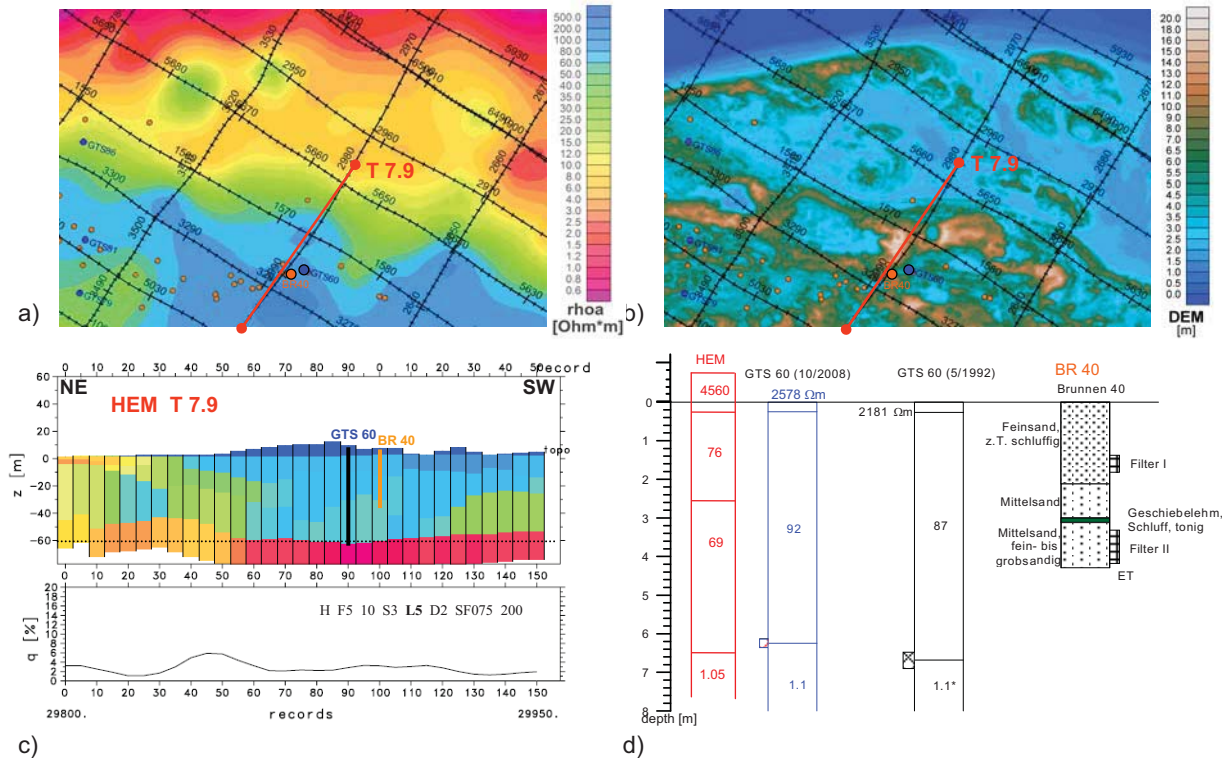
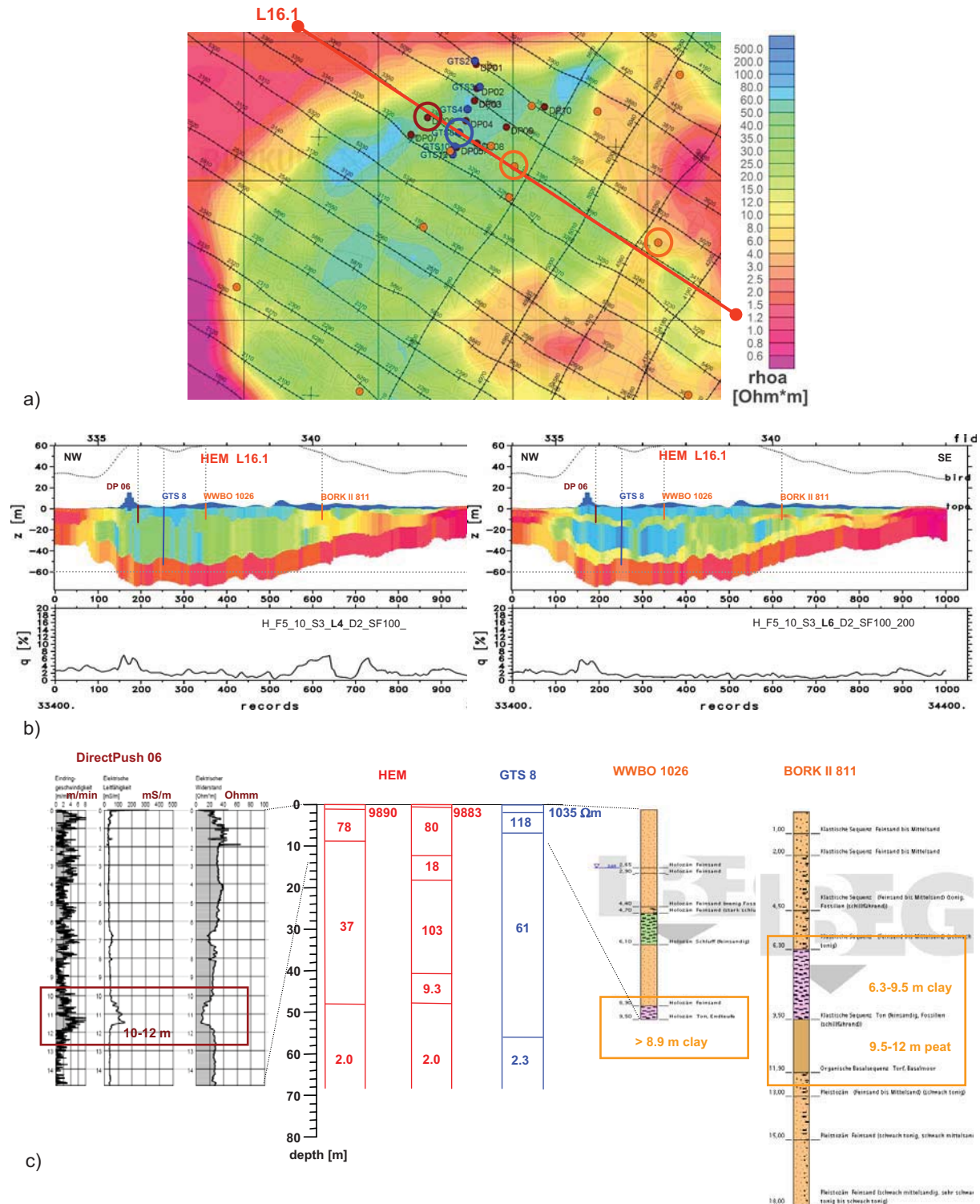


Figure 6: a) Resistivity map (5 mbsl) with the location of a part of flight-line T7.9 (red line), the lithological log BR40 (orange circle) and the DC site GTS 60 (blue circle). b) Digital elevation model revealed from laser scan data of NLWKN. c) Vertical resistivity section T7.9 as result of a 5-layer 1-D inversion (every 5th model is shown). d) Comparison of 1-D inversion results of HEM and DC reveals the freshwater/saltwater boundary at a similar depth of about 60 mbsl.

The elevation map in Figure 6b shows the location of the *Schlumberger* sounding GTS 60 (blue circle), the drilling BR 40 (orange circle) and a part of flight-line T 7.9 (red line). The flight-line crosses a dune and the *Schlumberger* sounding is located about 80 m apart from the flight-line in a dune valley. The vertical resistivity section of this part in Figure 6c shows that the saltwater/freshwater boundary was determined by the HEM 1-D inversions at about 60 m depth. This is confirmed by the 1-D inversion result of the *Schlumberger* sounding of 2008 which has not changed significantly since 1992 (Figure 6d).

Regarding the results of *direct-push* (Winter, 2008) and borehole measurements, the HEM related resistivity structure could be determined more precisely. A more detailed 6-layer approach revealed additionally to the freshwater/saltwater boundary thin conductive layers at about 10 m depth, not seen in the 4-layer models of the standard inversion procedure (Figure 7b). Partly, they were identified as clay layers by lithological logs; and also the result of a *direct-push* measurement shows an increasing conductivity at 10 m depth (Figure 7c). The depth of the freshwater/saltwater boundary remains nearly constant and has a layer with little higher resistivity above, which could be interpreted as transition

zone of brackish water. The data misfit q [%] of the 6-layer inversion is smaller than the misfit of the 4-layer inversion, especially in the south-eastern part of the section.



Estimation of the aquifer thickness

The aquifer thickness was estimated using *Archie's law* with the following approach: The resistivity of freshwater is assumed to be higher than $5.6 \Omega\text{m}$ (180 mS/m) and thereby the resistivity of freshwater saturated sediment with 33% porosity and a formation factor of 4 should be above $23 \Omega\text{m}$ (*Siemon et al., 2009b*). Based on that, the accumulated layer thickness with resistivities of $30\text{--}500 \Omega\text{m}$ is defined as aquifer thickness and is shown in Figure 8. From that, the freshwater volume was estimated to $175 \times 10^6 \text{ m}^3$.

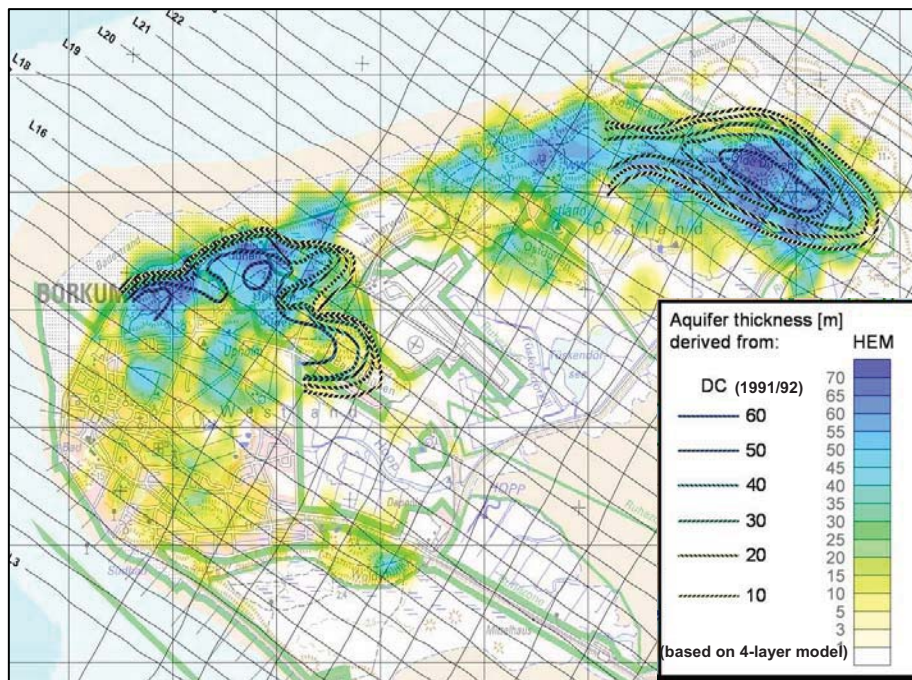


Figure 8: Aquifer thickness derived from DC and HEM.

Conclusions

At several locations the depth of the freshwater/saltwater boundary as derived by the HEM inversion was confirmed by *Schlumberger* soundings. The DC measurements were carried out at the same sites as 16 years before, no significant changes of the freshwater/saltwater boundary were found.

Regarding *direct-push* and borehole data, the HEM inversion provided more information about resistivity structure. A 6-layer approach revealed additionally to the freshwater/saltwater structure thin clay layers.

Based on the HEM inversion results the volume of the freshwater resource was estimated.

Acknowledgments

The HEM survey was funded by *LIAG/BGR* project “Airborne Geophysical Investigation of the German North Sea Coastal Area” (*Wiederhold et al., 2008*):

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http://www.bgr.bund.de/cln_101/nn_328750/DE/Themen/GG_Geophysik/Aerogeophysik/Projektbeitraege/D-AERO/deutschlandweite_aerogeophysik_befliegung_D-AERO.html

The lithological logs were taken from the *LBEG Map Server* (<http://www.geophysics-database.de/>) of the *Niedersächsisches Landesamt für Bergbau, Energie und Geologie (LBEG)*.

The direct-push results were provided by the *Ingenieurbüro für Hydrogeologie, Sedimentologie und Wasserwirtschaft (HSW)* (*Winter, 2008*).

The digital elevation model was provided by the *Niedersächsischer Landesbetrieb für Wasserwirtschaft, Küsten- und Naturschutz (NLWKN)*, Forschungsstelle Küste, An der Mühle 5, 26548 Norderney.

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Some notes on bathymetric effects in marine magnetotellurics, motivated by an amphibious experiment at the South Chilean margin

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1 Introduction

Whilst the onshore magnetotelluric method is established as a useful method for imaging electrical conductivity structures in the deep earth's interior, electromagnetic investigation in marine environment has attracted until recently much less attention, although the oceans cover the most part of the Earth's surface. Beside of logistical challenges, high technical demands on the instruments and involved higher costs, physical conditions in marine environments make the acquisition and interpretation of offshore data by far more difficult than onshore data. The broad period range of usable signals utilized in the terrestrial MT is substantially limited on the seafloor. The ionospheric and magnetospheric primary source field recorded on the seafloor is at short periods strongly attenuated by the covering highly conductive ocean layer and at long periods contaminated by motionally induced secondary electromagnetic fields. Additionally, attention has to be paid to the shape of the ocean bottom, which can massively distort the electromagnetic fields particularly in coast region.

2 Decay of electromagnetic fields in the ocean

The decay of the electromagnetic amplitude in a high conductive homogeneous half space like ocean water ($\rho = 0.3\Omega m$) and a less conductive medium like earth ($\rho = 100\Omega m$) shows Fig. 1. In a homogeneous half space without boundary in vertical direction the field behaviour in both media can be estimated according to the simple formula used for the calculation of the skin depth

$$\vec{F} = \vec{F}_0 e^{-kz}, \quad (1)$$

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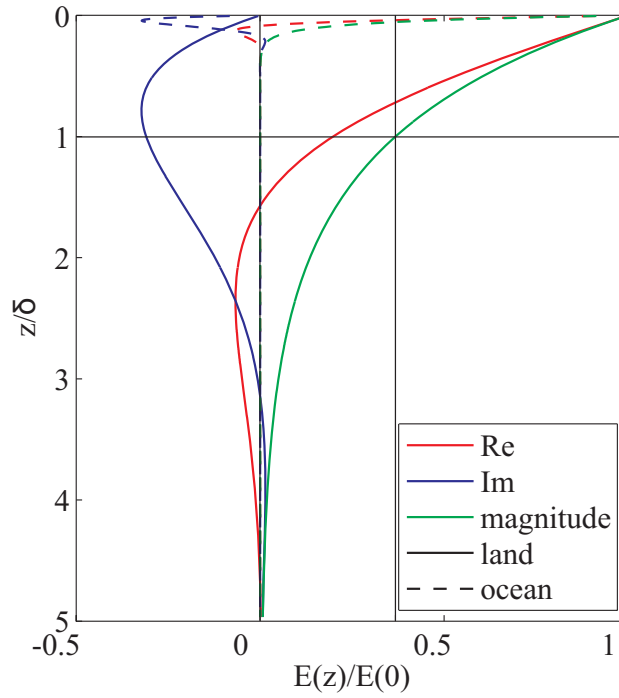


Figure 1: Decay of the of magnitude of electromagnetic field $\vec{F}$ (green) and its real (red) and imaginary part (blue) as function of depth in a highly conductive (i.e. ocean water, dashed line) and a less conductive homogeneous half space (solid line). The decay is scaled to the surface value $\vec{F}_0$.

where $k = \sqrt{i\mu\sigma\omega}$. At a depth of 4000 m, which corresponds to the average ocean depth, the amplitude of a field propagating in the ocean at a period of 10 s is attenuated hundredfold, i.e. the field on the ocean bottom would be approximately 1% of the surface value.

This simple relationship becomes more complicated when taking into account real conditions with a highly conductive ocean layer which is limited downward by a relatively resistive basement. Analog to the N -layered half space the solution has to be upgraded by a further term according to the reflection on the ocean bottom at depth h . The solution for the x -component of the electric field in the first and second layer is

$$E_{1,x} = E_{11}e^{-k_1z} + E_{12}e^{k_1z} \quad (2)$$

$$E_{2,x} = E_{22}e^{-k_2z} \quad (3)$$

where E_{11} is the amplitude of the downgoing and E_{12} of the upgoing wave in the first medium and E_{22} the amplitude of the downgoing wave in the second medium. For the y -component of the magnetic field follows from *Faraday's law* in a 1-D case

$$B_{1,y} = -\frac{1}{i\omega} \frac{\partial E_{1x}}{\partial z} = \frac{k_1}{i\omega} (E_{11}e^{-k_1z} - E_{12}e^{k_1z}) \quad (4)$$

$$B_{2,y} = -\frac{1}{i\omega} \frac{\partial E_{2x}}{\partial z} = \frac{k_2}{i\omega} E_{22}e^{-k_2z}. \quad (5)$$

A similar relation can be derived for the pair E_y and B_x . Using the continuity of the fields at $z = h$ resulting from the boundary conditions

$$E_{1,x}(h) = E_{2,x}(h), \quad B_{1,y}(h) = B_{2,y}(h) \quad (6)$$

and introducing the admittances

$$\zeta_1 = \frac{k_1}{i\omega} = \sqrt{\frac{i\omega\mu\sigma_1}{i\omega}} = \sqrt{\frac{\mu\sigma_1}{i\omega}} \quad \text{and} \quad \zeta_2 = \frac{k_2}{i\omega} = \sqrt{\frac{\mu\sigma_2}{i\omega}} \quad (7)$$

the ratio of amplitudes of the downgoing and upgoing fields in the ocean is

$$\frac{E_{12}}{E_{11}} = \frac{\zeta_1 - \zeta_2}{\zeta_1 + \zeta_2} e^{-2k_1 h} = \zeta_{12} e^{-2k_1 h}, \quad (8)$$

where $\zeta_{12} = \frac{\zeta_1 - \zeta_2}{\zeta_1 + \zeta_2}$ is the reflection coefficient. Its value is estimated according to eq. 7 by the conductivity contrast at the boundary between layers. For instance, in case of sharp contrast like ocean water ($\sigma_1 = 3.2 \text{ S/m}$) and basement ($\sigma_2 = 0.001 \text{ S/m}$) $\zeta_{12} = 0.965$. Note, that using as solution in eq. 2 and 3 the magnetic field instead of electric field, the reflection coefficient has to be expressed by reciprocal impedances instead of admittance and yields: $\eta_{12} = \frac{\eta_1 - \eta_2}{\eta_1 + \eta_2}$, with $\eta_{1,2} = \frac{k_1}{\mu\sigma_{1,2}} = \sqrt{\frac{i\omega}{\mu\sigma_{1,2}}}$.

Combining the equations 8 and 2 the electric field at depth $z \leq h$ becomes

$$E_{1,x}(z) = E_{11} (e^{-k_1 z} + \zeta_{12} e^{-2k_1 h} e^{k_1 z}) \quad (9)$$

and at the surface $z = 0$

$$E_{1,x}(0) = E_{11} (1 + \zeta_{12} e^{-2k_1 h}). \quad (10)$$

The ratio of the equations 9 and 10 gives the attenuation of the electric field in the ocean for $0 \leq z \leq h$ normalized in terms of the total field at the surface

$$|V_E| = \frac{E_{1,x}(z)}{E_{1,x}(0)} = \frac{e^{-k_1 z} + \zeta_{12} e^{-2k_1 h}}{1 + \zeta_{12} e^{-2k_1 h}} = \frac{1 + \zeta_{12} e^{-2k_1 (h-z)}}{1 + \zeta_{12} e^{-2k_1 h}} e^{-k_1 z} \quad (11)$$

This applies accordingly for the ratio of magnetic fields

$$B_{1,y}(z) = B_{11} \zeta_1 (e^{-k_1 z} - \zeta_{12} e^{-2k_1 h} e^{k_1 z}) \quad (12)$$

$$B_{1,y}(0) = B_{11} \zeta_1 (1 - \zeta_{12} e^{-2k_1 h}) \quad (13)$$

$$|V_B| = \frac{B_{1,y}(z)}{B_{1,y}(0)} = \frac{e^{-k_1 z} - \zeta_{12} e^{-2k_1 h}}{1 - \zeta_{12} e^{-2k_1 h}} = \frac{1 - \zeta_{12} e^{-2k_1 (h-z)}}{1 - \zeta_{12} e^{-2k_1 h}} e^{-k_1 z}. \quad (14)$$

The behaviour of the electric and magnetic fields as function of depth in a 5 km deep ocean for two selected and representative periods (100 s and 10000 s) and realistic conductivities in the ocean ($\sigma_1 = 3.2 \text{ S/m}$) and basement ($\sigma_1 = 0.001 \text{ S/m}$) is illustrated in Fig. 2(left). The fields are normalized to their surface magnitude according to relations

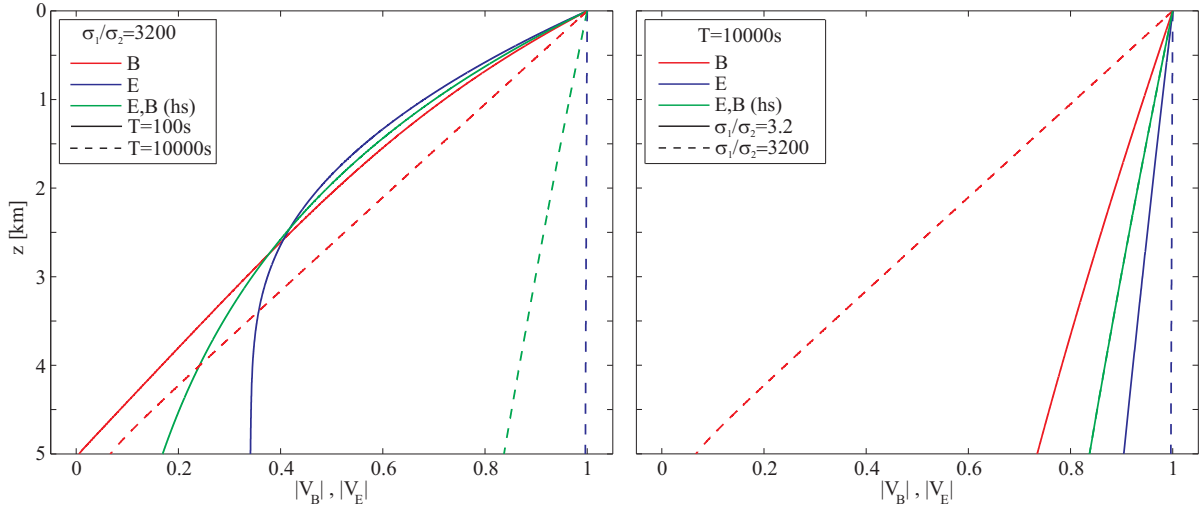


Figure 2: The decay of electromagnetic fields in an ocean shows a deviation from simple half space attenuation due to boundary conditions on the seafloor and depends beside period length and ocean depth also on resistivity contrast between ocean water and basement. Attenuation of electric (blue) and magnetic (red) fields for two periods $T_1 = 100 \text{ s}$ (solid line) and $T_1 = 10000 \text{ s}$ (dashed line) in a 5 km deep ocean with a sea water resistivity of $0.3 \Omega m$ and seafloor resistivity of $1000 \Omega m$ (right). Attenuation of the fields at a period of 10000 s for two different conductivities of the basement (0.001 S/m and 1 S/m) (left). See also (Brasse, 2009).

11 and 14. They show different decay in dependence of period. At short periods (solid line) both the electric (blue) and magnetic (red) fields experience strong attenuation and the curve shape is determined by exponential decay. Large differences in attenuation are observed at long periods (dashed line). While the electric field penetrates the ocean layer from surface to the seafloor nearly unchanged, the magnetic field is attenuated about fifteen-fold and reaches on the ocean bottom just about 7% of its surface value. However, both fields would behave identically in an infinitely extended ocean, that would be regarded as a homogeneous half space.

At first sight of the equations 11 and 14 one would think that the behaviour of the fields depends rather on the ocean thickness than on period. However, actually both is true. The decay of the fields is determined by both period T and thickness h of the ocean layer. That becomes evident considering the exponential and crucial term in the equations as function of skin depth according to the relations of skin depth $\delta = \sqrt{\frac{2}{\omega\sigma\mu}}$ and complex wave number $k = (1 + i)\sqrt{\frac{\omega\sigma\mu}{2}}$:

$$e^{-2k_1 h} = e^{-\frac{2h}{\delta_1}(i+1)}. \quad (15)$$

The quantity $\frac{h}{\delta_1}$ is the electrical layer thickness and via $\delta \sim \sqrt{T}$ also directly associated with the period length of the penetrating wave (McNeill and Labson, 1986). The fields behave identically if the exponent remains constant, regardless if high frequency fields propagate in a shallow ocean or low frequency fields in a deep ocean. The behaviour of the fields replicates, similar to a self-similarity principle by stimulating the exponent

$\frac{h}{\delta_1}$, i.e. by appropriate and simultaneous increasing or decreasing the frequency and the ocean thickness and thus creating either electrical thick $\frac{h}{\delta_1} \gg 1$ or thin conditions $\frac{h}{\delta_1} \ll 1$. In real world an electrical thick layer is equal to a high frequency field where the thickness of the ocean is much larger than the skin depth $\frac{h}{\delta_1} \gg 1$ and both $e^{-2k_1 h}$ and $e^{-2k_1(h-z)} \rightarrow 0$ (Brasse, 2009). Thus the effect of the basement on the fields is only marginal and the behaviour in a thick electrical layer is governed by exponential decay

$$\frac{E_{1,x}(z)}{E_{1,x}(0)} \approx e^{-k_1 z} \quad \text{respectively} \quad \frac{B_{1,y}(z)}{B_{1,y}(0)} \approx e^{-k_1 z}. \quad (16)$$

The decay of both fields becomes more alike the more the ocean depth or the frequency increases until finally approaching the decay curve of a homogeneous half space.

For low frequencies or for an electrically thin ocean, where the skin depth is significantly greater than ocean depth, $\frac{h}{\delta_1} \ll 1$, the exponential terms can be expanded into a power series

$$e^{\mp cx} \approx 1 \mp cx \quad (c \in \mathbb{C}) \quad (17)$$

Assuming additionally a non-conductive basement so that $\zeta_{12} = 1$ and substituting this approximations into eq. 11 it yields for an electric field

$$\frac{E_{1,x}(z)}{E_{1,x}(0)} = \frac{1 + e^{-2k_1(h-z)}}{1 + e^{-2k_1 h}} e^{-k_1 z} = \frac{1 - k_1(h-z)}{1 - k_1 h} (1 - k_1 z) \quad (18)$$

In the ocean layer at depth $z = h$

$$\frac{E_{1,x}(z)}{E_{1,x}(0)} = \frac{1}{1 - k_1 h} (1 - k_1 z) = 1 \quad (19)$$

and thus the electric field remains constant throughout the ocean as shown in Fig. 2. In this case the current density $J_x = \sigma E_x$ is constant as well.

Another behavior shows the magnetic field in an electrical thin layer. Assuming again an infinite conductivity contrast, $\zeta_{12} = 1$ and the equation 14 yields at $z = h$

$$\frac{B_{1,y}(z)}{B_{1,y}(0)} = \frac{1 - e^{-2k_1(h-z)}}{1 - e^{-2k_1 h}} e^{-k_1 z} = \frac{2k_1(h-z)}{2k_1 h} (1 - k_1 z) = \frac{(h-z)}{h} (1 - k_1 z) = 0 \quad (20)$$

Taking into account a lower conductivity contrast between ocean and basement, the reflection coefficient $\zeta_{12} < 1$ and the calculated attenuation values of the fields don't correspond exactly to the observed decay in Fig. 2(right). Especially the decay of the magnetic field doesn't converge to zero as to be expected from Fig. 2. The attenuation is smaller the larger the wave period and less the conductivity contrast. The first can be considered in terms of skin depth and thus as an effect of electrical thin ocean layer. The second is related to the current distribution in the ocean and the seafloor. According to Ohm's law, $\vec{j} = \sigma \vec{E}$ the current density is much higher in the conductive ocean than in the resistive basement. In the theoretical case of infinitely resistive basement it

would imply that the currents only flow in the ocean, inducing a secondary magnetic field which removes entirely the opposite primary field in agreement with the above mentioned result. In real conditions with high but not infinite basement resistivity the currents diffuse also to the seafloor, so that the the secondary magnetic field of the diminished currents in the ocean can in fact reduce but not completely cancel the primary magnetic field. Thus the ratio $\frac{B_{1,y}(z)}{B_{1,y}(0)}$ doesn't become zero rather $0 < \frac{B_{1,y}(z)}{B_{1,y}(0)} < 1$ depending on ζ_{12} and electrical conditions of the ocean layer. In other words the reflection coefficient regularizes the conditions for distribution of electrical currents in sea bottom and estimates in this manner the damping behaviour of the fields (Chave et al., 1991).

The decay of the electric field in the ocean is against intuition, and negligible compared to the magnetic field. In the theoretical case with infinite conductivity contrast, ($\zeta_{12} = 1$) the ocean represents for long periods an electrical thin layer that the electric field passes almost intact. In real condition mentioned above ζ_{12} is still high and the low frequency field decays gently, in terms of an electrical thin layer. That would change assuming a layer of highly conductive sediments on the ocean bottom. Since the layer becomes thicker, the electric field decays more strongly, the magnetic field in contrast much less since a part of currents now flows in the sediments. Thus both fields would approximate the curve of attenuation in a homogeneous half space, what for short periods corresponds to exponential decay Fig. 2(right).

Similar effects would be observed considering the fields at a fixed point z_1 , within the ocean ($0 < z_1 < h_1$) as a function of ocean depth. As the ocean depth increases from h_1 to h_2 at a constant period both fields behave differently at a fixed point z_1 : the electric field becomes smaller and the magnetic field larger. Again, in deeper ocean currents flow through a broader region, the density becomes smaller as well as the compensating effect of the secondary opposite magnetic field at depth h_1 and the magnetic field increases compared to flat ocean. On the other hand the electric field experiences more attenuation due to the increased electrical layer.

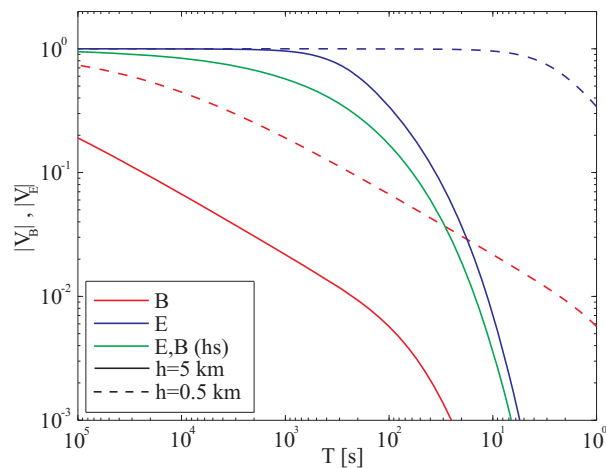


Figure 3: Magnitude of B- and E-fields at the seafloor in 0.5 km depth and 5 km as function of period. In comparison the behavior of the fields in depth of 5 km in a downward unlimited ocean (homogeneous half space).

It is also instructive to consider the electric and magnetic fields in the ocean as function of periods. Fig. 3 illustrates graphically the ratio $|V_B| = \frac{B_{1,y}(z)}{B_{1,y}(0)}$ and $|V_E| = \frac{E_{1,x}(z)}{E_{1,x}(0)}$ for various thicknesses of the ocean layer. It becomes clear immediately that in the real ocean environment attenuation of the fields differs fundamentally over a total period range. While the electric field remains nearly unchanged in electric thin layer ($\frac{h}{\delta_1} \ll 1$), i.e. when the skin depth exceeds the ocean thickness, the magnetic field experiences marked attenuation over the total period range, up to about hundredfold of the electric field attenuation. As expected, the range of periods over which the fields decay exponentially increases, when the depth of the ocean increases or the resistivity of the basement diminishes. In the limit case if ($\zeta_{12} = 0$) and the currents flow through an unlimited broad region the curves approximate the decay of a homogeneous half space.

Note that in case of reverse conditions, i.e. the upper layer is more resistive than the lower layer – like may be the case on land – the fields behave reversely as well: the electric field decays quickly while the magnetic field passes the layer almost unchanged.

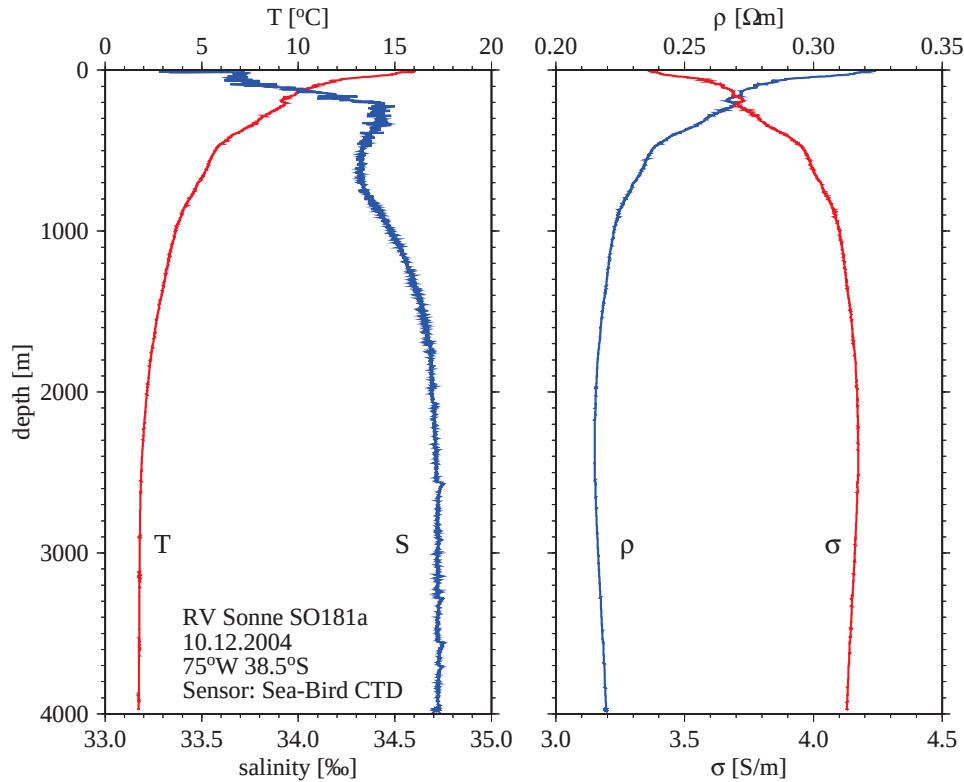


Figure 4: Logs of temperature and salinity (left), electrical conductivity, resistivity (right) measured during RV Sonne leg SO-181.

The resistivity values of the ocean assumed in prior theoretical consideration was motivated by a CTD log performed in 2004 close to the South Chilean continental margin during RV Sonne leg SO-181 at depths until 4000 m (Fig. 4). The knowledge of water conductivity is also essential for subsequent modeling where the ocean has to be incorporated as an a priori structure in the finite difference mesh. The conductivity in

ocean depends on pressure, salinity and temperature. While the effect of pressure and salinity is marginally, the temperature variations in ocean change resistivity of oceanic water from $0.16 \Omega m$ at $30^\circ C$ to $0.33 \Omega m$ at $1.0^\circ C$. The estimated resistivity values about $0.3 \Omega m$ corresponds very well to literature values (Filloux, 1987)

3 Offshore magnetotelluric and magnetic transfer functions in presence of bathymetry

The discussions about magnetotellurics on the seafloor are often dominated by attenuation of electromagnetic fields by high conductive ocean and secondary induction by sea tides. Fewer studies refer to distorting effects on electromagnetic fields caused by changes in seafloor bathymetry in presence of overlaying high conductive environment (Constable et al., 2009).

For a model incorporating a homogeneous half space with resistivity of $100 \Omega m$ and conductive ocean with bathymetry that is observed at the South Chilean continental margin (Fig. 5, top), magnetotelluric transfer function were calculated with the algorithm of Mackie et al. (1994). At least two clear findings can be derived from the model

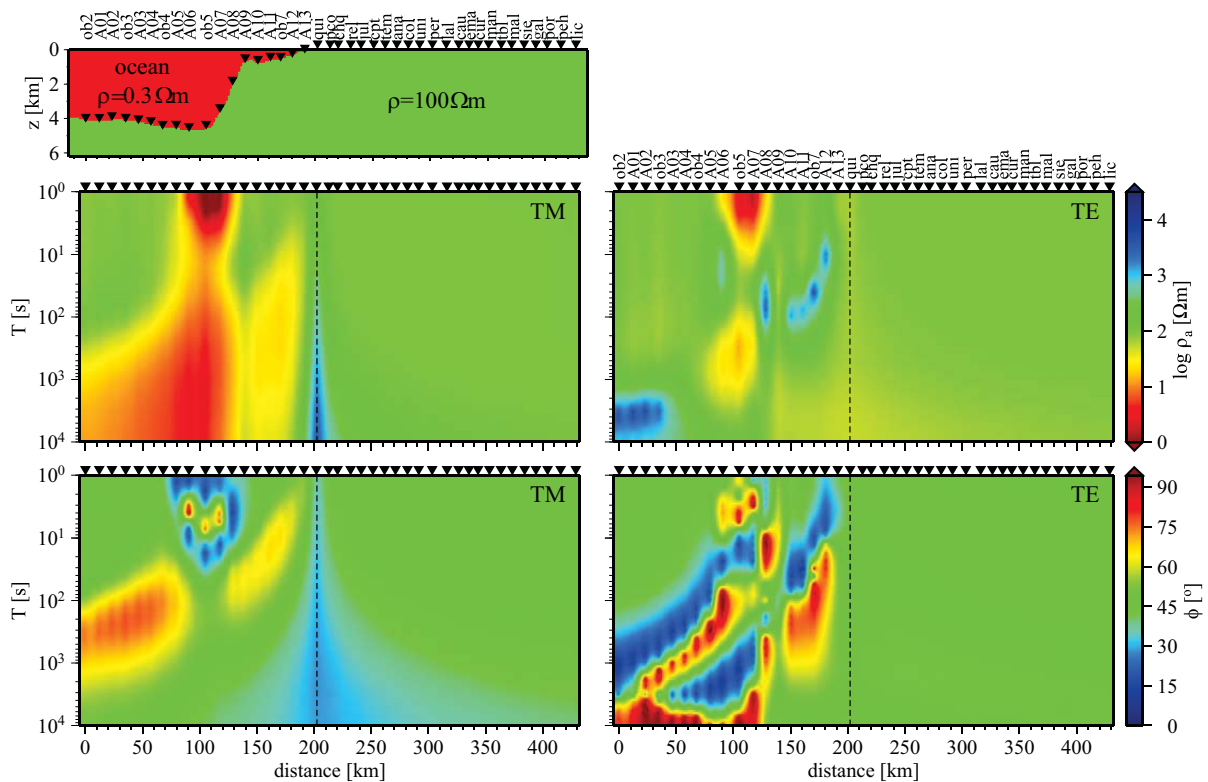


Figure 5: Responses of a synthetic amphibious model shown on the top. Middle: resistivity of TM and TE mode. Bottom: phase of TM and TE mode. Dashed line separates the profile into offshore and onshore part.

responses (Fig. 5, middle and bottom). Firstly the ocean has rather a marginal effect on the onshore magnetotelluric transfer function (right of the dashed line). Actually only in TM mode of onshore stations close to the ocean slightly decreasing phases and enhanced apparent resistivity on the edge between offshore- and onshore profile marked by a dashed line, can be observed. Secondly the bathymetry produces dramatic anomalous effects in the ocean, which can be observed in both modes as well as in phases and resistivities. The resistivity in TE mode rises and falls dramatically producing cusps in the image and phases even exceed -180° and 180° .

Similar anomalous features can be observed in the magnetic transfer functions (Fig. 6). Slight changes in the shape of the seafloor, adumbrated by a dotted line, are overdrawn reproduced by dramatic variations in the magnitude of the induction arrows (note the smooth bathymetry change between stations A09 and A13 and their huge impact on the magnitude of the induction arrows). On the other hand, the ocean effect on onshore station is limited to long-periods and to near-coastal region, again.

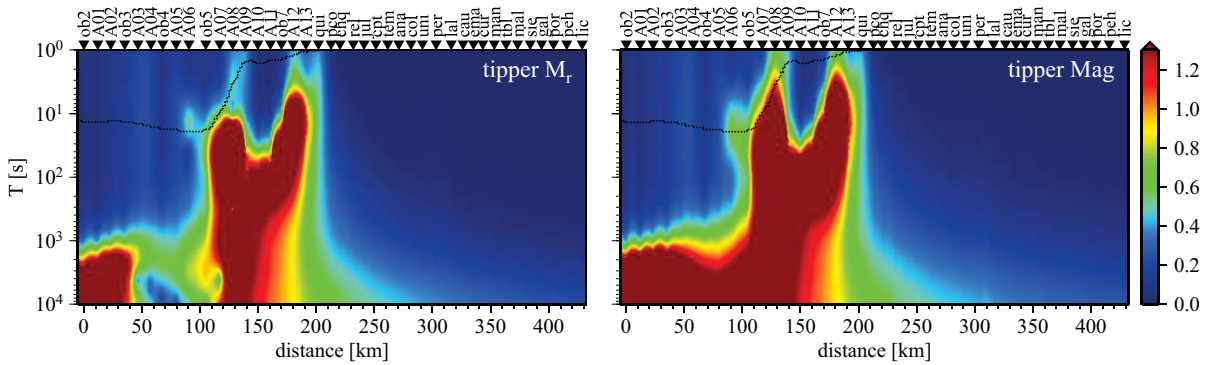


Figure 6: Real part (left) and the magnitude (right) of the tipper calculated for synthetic and amphibious model shown on the top of figure 5. The dotted line indicates the shape of the bathymetry of the model.

Comparing these synthetic responses with the real induction vectors observed at the only station at the continental slope, ob7, shows that the bathymetry indeed affects the measured data, as shown in Fig. 7. The huge induction vectors (over 1) at middle and long periods correspond roughly with magnitude values calculated from the simple amphibious model. For short periods (until 100 s) a comparison is impossible due to

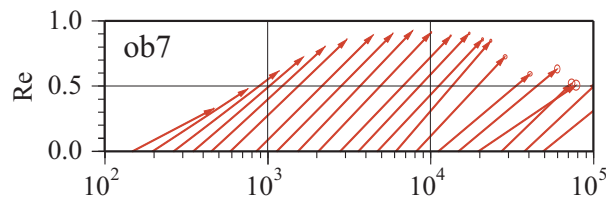


Figure 7: Real part of the observed induction vectors at station ob7, deployed on the continental margin off Southern Chile. Induction vectors observed land-side are presented by Brasse (2009).

unavailable data (note the different period range of the images).

The anomalous responses can be elucidated considering electric and magnetic fields in presence of changing seafloor shape. An enhanced concentration of electrical currents at the continental slope produces anomalous strong vertical magnetic field, that is known also on the land-side as coast effect (e.g., Fischer, 1979). Due to the high sea water conductivity the induced electric currents concentrate primarily in the ocean avoiding the much more resistive subsurface and generate a secondary magnetic field which compensates the primary opposite field, so that the magnetic field diminish considerably or disappears completely in case of insulating substrate as mentioned in 2 and shown in Fig. 2. At the continental margin, where the seafloor shallows towards land, the density of electric currents flowing in the ocean along the coast (i.e. in the TE mode) and above (instead below) measurement points increase inducing an anomalous and opposite magnetic field. This opposite, secondary field becomes predominant on the ocean bottom, where the primary field is strongly damped whereas the electric field is by the coast effect only marginally affected, and causes jumps in the phase (Weidelt, 1994). Moreover the resistivities in TE mode rise steeply generating upward cusps and the tipper gets very large as presented in Fig. 6 and 5.

The periods at which the effects occur depend on position of the probe in relation to the slope and corresponds to lateral distance to which a transfer function is sensitive to bathymetry. However, at high frequencies the field is generally unaffected by the bathymetry and the results are MT-like because the fields still behave like plane waves. At low frequencies, the currents are below the ocean. At the middle frequencies, the currents are in the ocean and this causes the responses looking like those due to infinite line currents right above the station. Note that if the MT responses were recorded on the top of the sea, the responses would be perfectly fine as expected in an usual 2-D case, since all the currents pass below the sounding surface instead above like in the presented marine model. This behavior is overcome or reduced by introducing a very conductive layer below the ocean bottom (e.g., sediments), which has to be at least several kilometers thick. However, the assumption of an oceanic crust being entirely well conductive seems unrealistic; this is even more the case for the continental crust below the slope: every aquiclude would prohibit the intrusion of sea water into deeper layers. Another class of models makes the effect described above disappear too: the association of the upper part of the downgoing plate with a good conductor, in accordance with standard models of subduction.

4 Summary

Offshore magnetotellurics is an useful method for exploration of seafloor and ocean-continent subduction zones, where fluids and melts are known to control the subduction process. A combined on- and offshore magnetotelluric transect across subduction zone is assumed to be able to resolve high conductive structures associated with dehydration processes, water migration and melts building.

However special conditions on the ocean floor not allow to interpret and to deal

with the marine records in the same manner as with onshore data. The presence of bathymetry distorts the magnetic and magnetotelluric transfer functions, particularly in the TE mode. Even a gently changing seafloor shape generates an enhanced concentrations of electric currents flowing above instead below measurement station and inducing on the ocean bottom a predominant anomalous opposite magnetic field. This results in phases exceeding the quadrant and cusps in the apparent resistivity.

The high conductive sea water causes a strong attenuation of electric and magnetic field at short periods. Towards long periods the decay differs clearly for both fields. The electric field penetrates the ocean layer from surface to the seafloor, counterintuitive, nearly unchanged, while the magnetic field experiences a strong decay and reaches the ocean bottom just with a fraction of its surface value. Moreover the decay depends strongly from the resistivity contrast between ocean and seafloor. Reducing the resistivity of the basement the electrical thickness of layer increases and both fields approximate a field decay like in a homogeneous half space.

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A permanent array of magnetotelluric stations located at the South American subduction zone in Northern Chile.

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Introduction

Monitoring the dynamic behavior of an active deep subduction system is focus of the Integrated Plate Boundary Observatory Chile (IPOC), a permanent array of combined geophysical and geodetic stations in Northern Chile which is operated since 2006 by the GFZ German Research Centre for Geosciences. Magnetotelluric (MT) data is gathered at seven out of a total of eleven observation sites.

The MT set up consists of three component long period fluxgate magnetometers (GeoMagnet) and non-polarizing Ag/AgCl electrodes to measure all three components of the magnetic field and both horizontal components of the electric field. The signals of the electromagnetic fields are continuously sampled at a rate of 20 Hz and at four sites transferred via satellite link to the GFZ in Germany. The objective of the project is to monitor and analyze electromagnetic data to decipher possible changes in the subsurface resistivity distribution, e.g. as a consequence of large scale fluid relocation.

Here, we present vertical magnetic transfer functions as time series over a time span of more than two years for the period range from 10^{-1} to 10^4 seconds¹. These vertical magnetic transfer functions are sensitive to lateral changes of electric conductivity in the subsurface. Some components of these transfer functions show frequency dependent variations with a periodicity of roughly one year. These effects are observed at all sites of the array.

Due to the extreme dry ground of the Atacama desert continuous monitoring the electric field is difficult. Contact resistances are on the order of M Ω and electrolyte is leaking. Different types of electrodes are currently being tested.

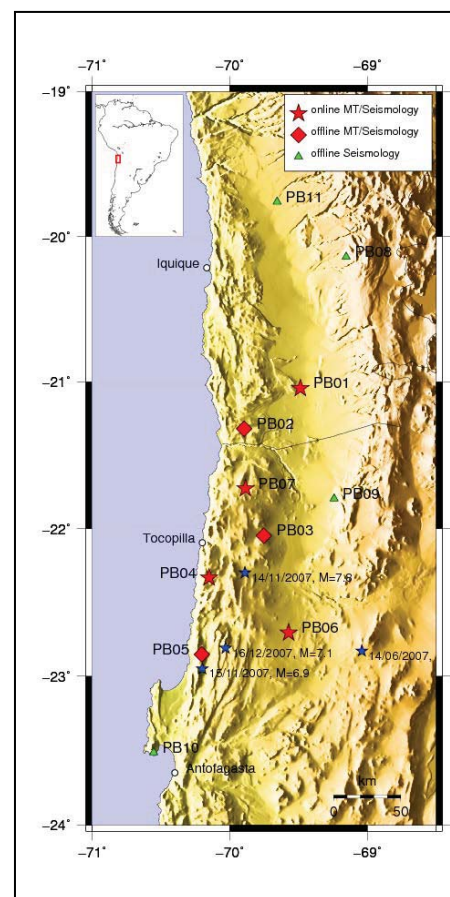


Figure 1: The IPOC-MT array in Northern Chile showing sites PB01 to PB07 (red symbols). Blue stars indicate major earthquakes.

¹Periods shorter than 10 seconds are not considered since this is the shortest period fluxgate magnetometers are able to resolve.

Periodic variations in the MT monitoring data

To identify changes in the subsurface we examine time series of vertical magnetic transfer functions $T_x(\omega)$ and $T_y(\omega)$:

$$B_z(\omega) = T_x(\omega) B_x(\omega) + T_y(\omega) B_y(\omega).$$

These quantities are sensitive to lateral changes of electric conductivity in the subsurface.

The value of each day and component of the transfer function time series is obtained by processing magnetic field time series of 3 days (using the day before and after). Subsequently the 3 day time window is forwarded by one day and the procedure is repeated until the entire time series is processed. The geomagnetic transfer functions were obtained using the robust processing described in Ritter et al. (1998), Weckmann et al. (2005), and Krings (2007).

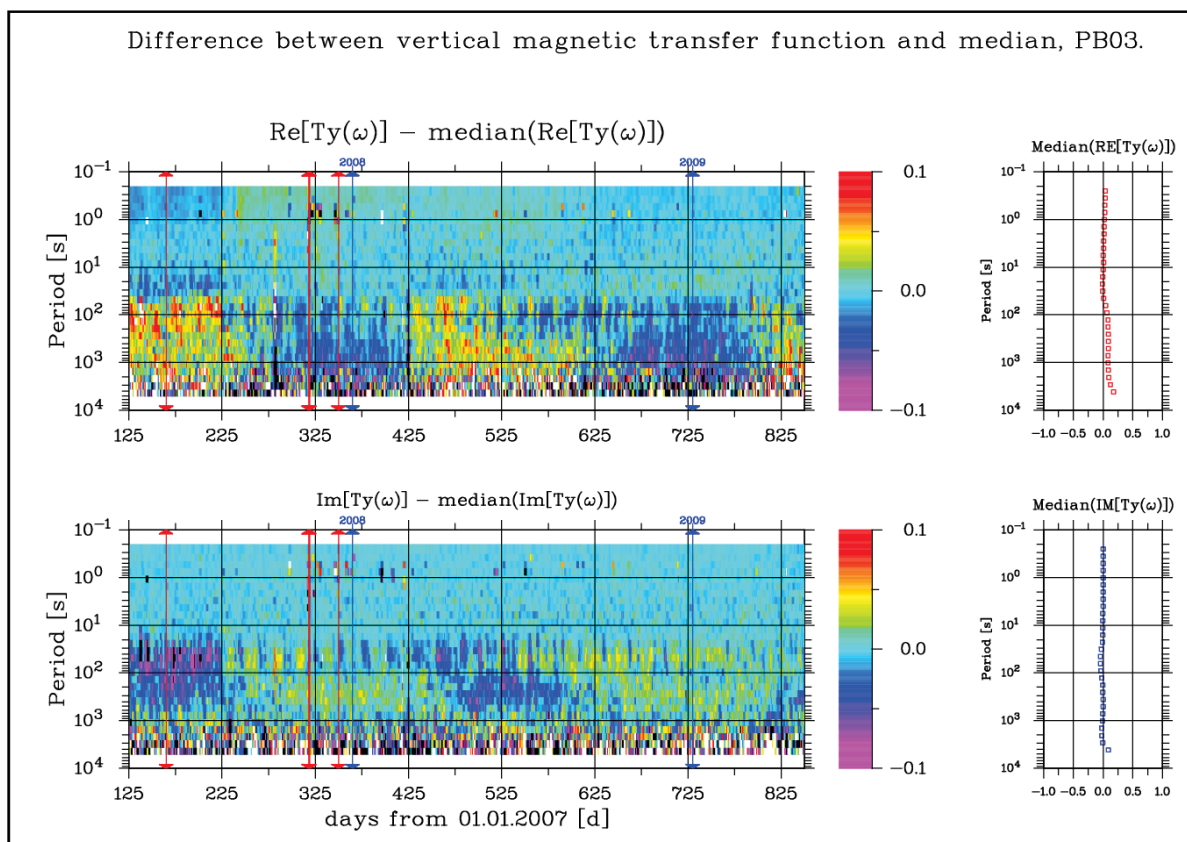


Figure 2: Time series showing the differences between daily values of T_y and the median of the entire data section of 730 days at each period for site PB03. For monitoring purposes it is desirable to amplify variations in the vertical magnetic transfer function components. Therefore, the median is calculated for each period and component and subtracted from the value of each day. The median of each frequency band is plotted on the right hand side. The median is more robust against outliers than the arithmetic average. Vertical red lines indicate major earthquakes (see Fig. 1), vertical blue lines indicate turn of the year. Interestingly, the ocean effect, which would be indicated by a large positive value of the median of $\text{Re}[T_y]$, is rather small at site PB03.

The time series of the real and imaginary parts of the T_y (east-west) component at PB03 show a remarkable frequency dependent feature (Fig. 2). At periods between approximately 40 and 1000 s the observed $\text{Re}[T_y]$ values show a continuous variation of the amplitudes with a periodicity of roughly one year. The amplitudes vary in the range ± 0.1 and reach maximum values in the austral

winter before the winter solstice. In the austral summer, the minimum occurs before the summer solstice. Similar effects can be observed consistently at all sites of the array (Fig. 3).

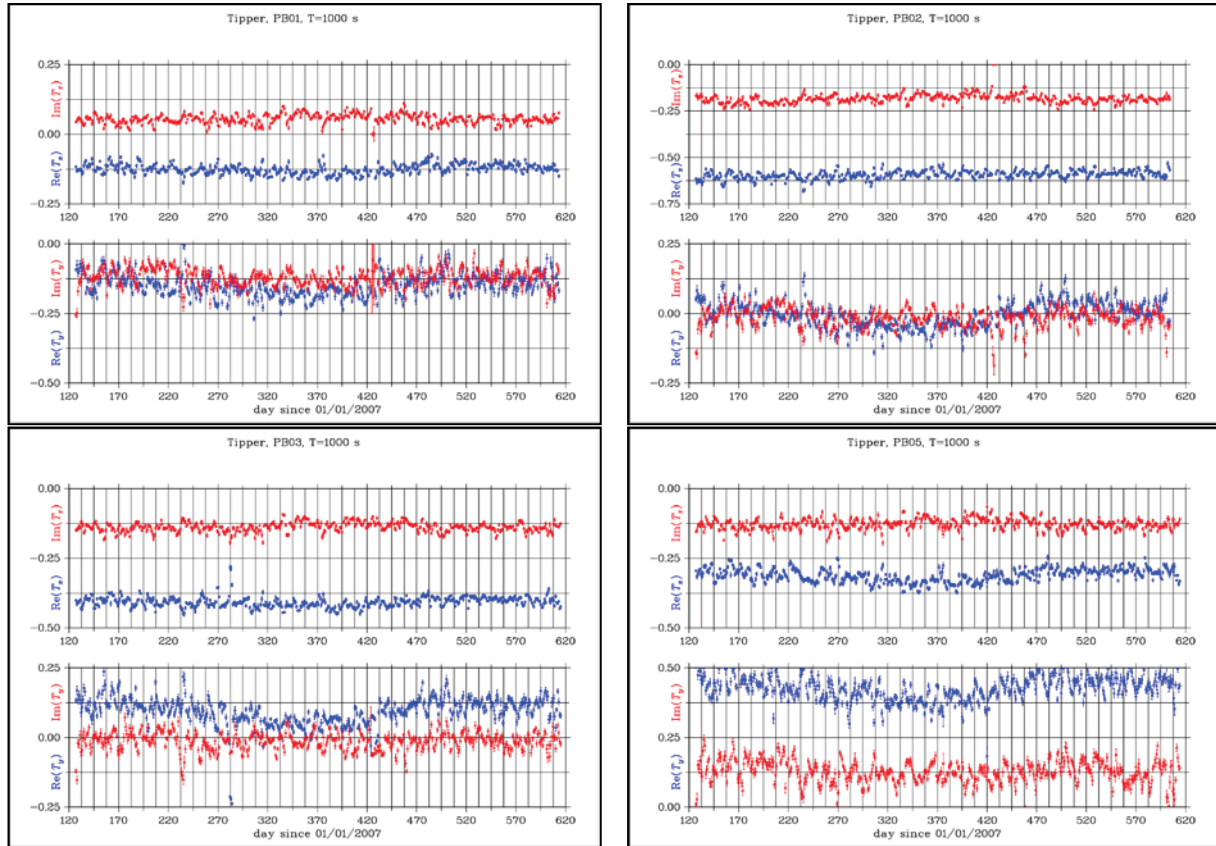


Figure 3: Time series of real (blue) and imaginary (red) parts of vertical magnetic transfer functions T_x and T_y of around 600 days at a period of 1000 seconds of sites PB01, PB02, PB03 and PB05.

Possible causes for the periodic variations of T_y

Possible causes for this periodic variation of the T_y component could be source field inhomogeneities. The sources for the MT-method are the natural variations of the earth's magnetic field which are assumed to be far away from the observer so that electromagnetic waves penetrate

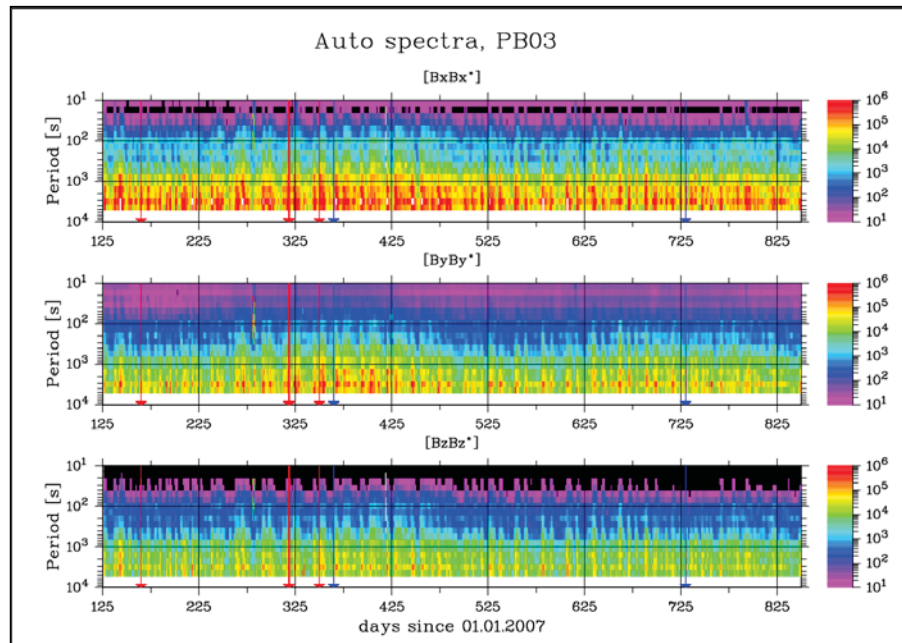


Figure 4: Time series of the magnetic auto spectra of PB03, extracted from the 3-day-average processing described in section 2. The $ByBy^*$ spectra show the same variations with a periodicity of one year as T_y .

the subsurface as plane waves.

To examine if the variations in the time series of vertical magnetic transfer functions are caused by seasonal alterations of source fields we plotted the time series of magnetic auto spectra of PB03 (Fig. 7) for the same time window as in Figure 2. These magnetic auto spectra show a clear correlation between the B_y -auto spectrum (east-west direction) and the T_y data (Fig. 8). This would support the idea that the results are influenced by source field effects.

Sq variations and equatorial electrojet

There exist several effects on the geomagnetic field which have a wide spectrum of variations (Onwumechili, 1997). Annual variations of the geomagnetic solar quiet daily variation (Sq) show the same periodicity as the measured variations of the magnetic auto spectra at PB03. The geomagnetic Sq field has a variation with a dominant periodicity of 1 day. Solar radiation generates ionized molecules in parts of the ionosphere (around 80 to 300 km height) which produces variable charged particles and, consequently, conducting air. Additionally, the solar radiation causes thermo-tidal winds which move the conducting ionosphere through the geomagnetic field. The result is a system of electric currents depending on the position of the sun. Horizontal and vertical components of Sq tend to be predominantly semi-annual in the zone of Equatorial Electrojet (EEJ) but predominantly annual at other latitudes (Onwumechili, 1997).

The IPOC MT array (21 - 23° south) in Northern Chile is located near the tropic of Capricorn (23° 26' south). This means, it is influenced by the Equatorial Electrojet (EEJ), a varying electric current flowing eastward in the ionosphere at a height of 100 - 130 km. Over South America the EEJ is warped to the south because it follows the magnetic dip equator (Fig. 8). The amplitudes are aligned to the north and to the south by return currents (Lühr et al., 2004) with peak values at latitudes some 5° away from the magnetic dip equator.

Annual Sq variations are most likely the cause of the variations of the Y-component of the vertical magnetic transfer functions at the MT-array in Northern Chile.

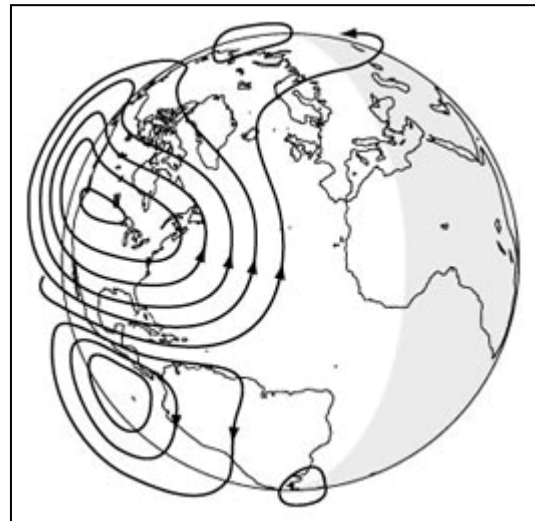


Figure 5: Exemplary electric currents in the ionosphere cause Sq variations in the northern summer.

http://geomag.usgs.gov/images/ionospheric_current.jpg (10.02.2010)

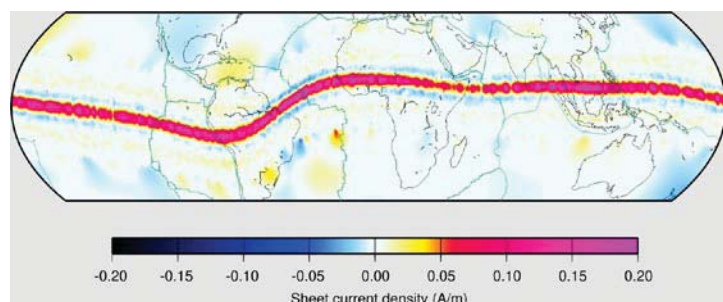


Figure 6: Electrojet current densities inferred from 2600 passes of the CHAMP satellite over the magnetic equator between 11:00 and 13:00 local time.

http://www.geomag.us/info/equatorial_electrojet.html (10.02.2010)

Non-periodic variations in the MT monitoring data

No periodic long term variations can be identified in the Tx component of the vertical magnetic transfer function time series (Fig. 7). A continuous variation at a period of approximately 20 s can be observed in the Re[Tx] component. It has a high amplitude variation between -0.2 and +0.2.

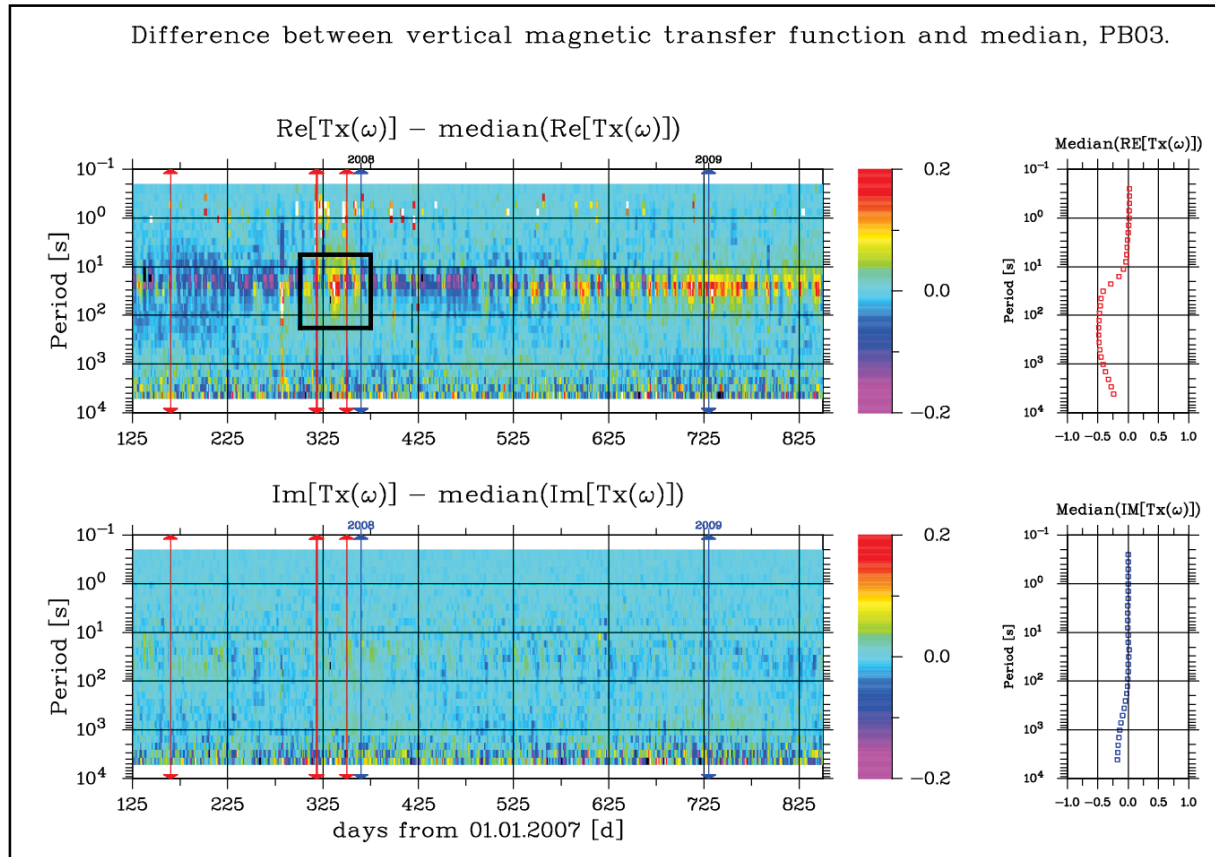


Figure 7: Time series of differences between daily values of Tx and median of the entire data section at each period at site PB03. Vertical red lines indicate major earthquakes, vertical blue lines indicate turn of the year. Values of median versus period are plotted on the right hand side.

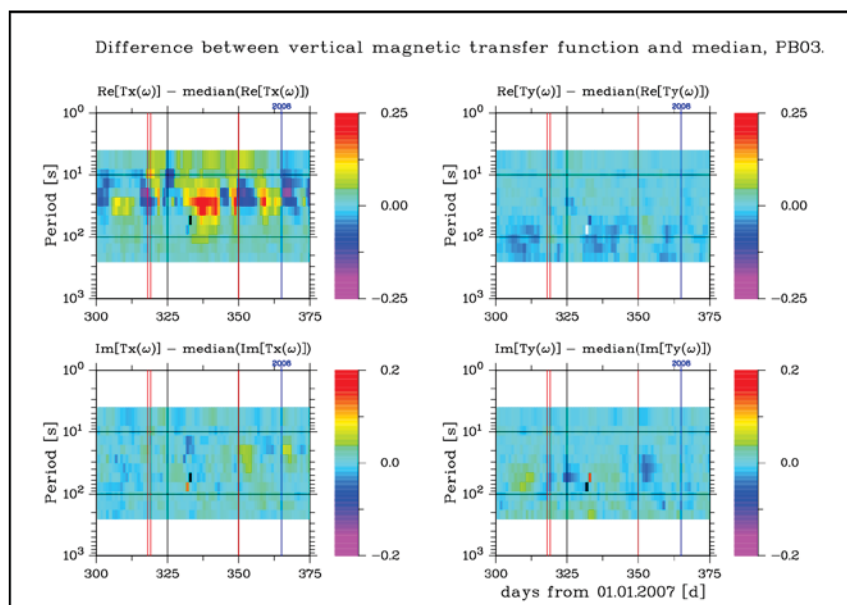


Figure 8: Time series of differences between daily values of vertical magnetic transfer functions and median at site PB03 versus period. Length and period range correspond to the black box in figure 6. Vertical red lines indicate major earthquakes (Fig. 1), vertical blue lines indicate turn of the year.

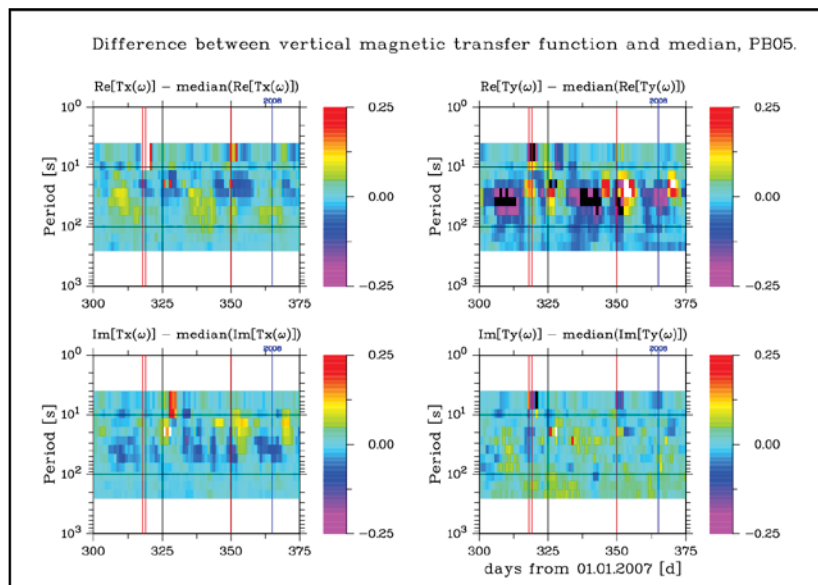


Figure 9: Time series of differences between daily values of vertical magnetic transfer functions and median at site PB05 versus period. Length and period range correspond to the black box of Fig. 6. Vertical red lines indicate major earthquakes (Fig. 1), vertical blue lines indicate turn of the year.

Note that the curve of the median, which represents the average transfer function, varies smoothly and consistently in this period range.

A remarkable feature can be observed between two major earthquakes in December 2007 (Fig. 7, black box). The area of interest shows a sudden rise in amplitudes up to ± 0.25 (Fig. 8), which corresponds to 100% of the median value of Tx (Fig. 7, right hand side). A similar but less pronounced effect is observed at site PB05 (Fig. 9).

More modeling is necessary to quantify the scale and location of structural changes of the electrical conductivity in the subsurface which could explain these non-periodic variations in the vertical magnetic transfer function time series. Calculation and modeling of inter station transfer function time series should also give more detailed information about changes in the conductivity structure of the underground.

Acknowledgements

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Discussion on backarc mantle melting in the central Andean subduction zone, based on results of magnetotelluric studies

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Abstract

Long-period magnetotelluric measurements in the Bolivian orocline revealed a large conductor in the backarc upper mantle of the central Andean subduction zone at 18°S. Data analysis and interpretation by 2-D inversion has already been described by Brasse & Eydam (2008a) who interpreted this conductor as an image of partial melt triggered by fluid influx from the subducting slab.

Remarkable are its high conductivities ranging up to above 1 S/m and implying high melting rates of more than 6vol%. However, regarding the distribution of intermediate depths seismicity which marks dehydration reactions in the subducting slab, one has to note the large distance of more than 60 km between fluid source and the assumed fluid-triggered partial melting of mantle peridotite. How do actual concepts of subduction zone dynamics fit to this resistivity image?

There are at least two approaches which are both followed by in the subsequent discussion: first is to study the mobility of deep fluids in subduction zones, second is to study the mobility of the slab itself.

Introduction

Ocean-continent subduction zones are earth regions of major raw material recyclings. Hot magmas erupt nearby where cold oceanic lithosphere subducts. This apparently contradictory observation is explainable by consumption of deep slab-derived fluids for flux melting of the peridotitic mantle in the asthenospheric wedge:

Hydrous minerals in the oceanic lithosphere are progressively dehydrating during subduction. The hereby released water may trigger partial melting of the overlying asthenospheric wedge by lowering mantle solidus generally at depths below 80 km. Melts and fluids ascent with some intermediate storage at density contrasts at the Moho or intra-crustal boundaries and give rise to arc and backarc volcanism. Due to their

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enhanced conductivities, saline fluids and (hydrous) melt should principally be detectable by deep electromagnetic sounding methods, as was demonstrated, e.g., in the Cascadia/Juan de Fuca subduction zones (Kurtz *et al.*, 1986; Wannamaker *et al.*, 1989; Varentsov *et al.*, 1996).

Hydrous phases in the oceanic crust relevant for sub-arc water release are amphibole and lawsonite. Amphibole is abundant in the basaltic oceanic crust and decays only slightly pressure-dependent at 65-80 km depths, marking the transition from blueschist to eclogite facies. In contrast, lawsonite can be stable up to high pressures of 8.4 GPa ($z \approx 250$ km) and dehydrates when temperatures exceed 700-800°C. Mantle water release is mainly controlled by serpentinite dehydration between 500-700°C. Serpentinite can be stable up to high pressures of 6 GPa where it transforms through a water-conserving reaction to the hydrous DHMS (*Dense Hydrous Magnesium Silicat*)-phase A.

The amount of released water thus strongly depends on the thermal structure as well as on subduction geometry. Young and slowly subducting slabs may dehydrate completely whereas old and rapidly subducting lithospheres may import large amounts of mantle water into the deeper mantle, making them crucial for the global water cycle.

The definitive amount of deep fluids in subduction zones is still uncertain because of the dependency of dehydrations from slab heating via mantle convection which in turn, strongly depends on water content in the wedge, besides other mostly poorly constrained parameters.

Geological setting

The central Andean subduction zone

Along 7000 km length the oceanic Nazca plate subducts beneath the South American continental margin where the Andes evolved as a Neogene volcanic chain embedded in a compressive backarc tectonic setting. Holocene volcanism is separated in four active segments, the austral, southern, central and northern volcanic zone (fig. 1).

Characteristic for the whole subduction system are low slab dips of less than 30° and inactive volcanism in regions where aseismic ridges subduct. Those hot oceanic lithospheres probably flatten at main depths of sub-arc water release due to a retard of dehydration and thus density increase of the slab which may lead to 'horizontal subduction' over several hundreds of kilometres (Gutscher *et al.*, 2000). Therefore, slab pull and buoyancy forces should be quite equilibrated probably making the slab mobile in geological time scales.

The Bolivian orocline (13-28°S) encompasses the central volcanic zone (CVZ) where the oldest part of the Nazca plate (55 Ma) subducts obliquely with an angle of N77°E and with a current velocity of 6.5 cm/a (Klotz *et al.*, 2006). Convergence rate has continuously slowed down from high 15 cm/a since the breakup of the oceanic Farallon plate in Nazca and Cocos plate 26 Ma ago (Somoza, 1998).

During the highly compressive Miocene regime, stresses are mainly relaxed by extreme

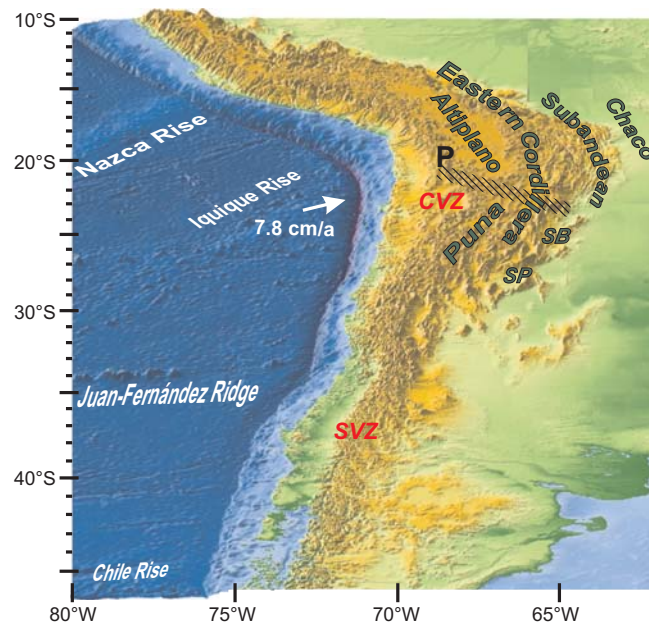


Figure 1: South American subduction zone with structural units of the central volcanic zone (CVZ) backarc and foreland. Hachures: Altiplano-Puna transition between 22-24°S; SB, SP - thick-skinned folding and thrusting in the Santa-Bárbara system and Sierras Pampeanas; P - Pica-Gap of Holocene volcanism (19.2-20.7°S) where Iquique Rise subducts (Wörner *et al.*, 2000); SVZ - southern volcanic zone. Convergence rate after Somoza (1998).

crustal shortening, thickening and uplift of the entire Andean arc and backarc region (e.g., Allmendinger *et al.*, 1997; Scheuber *et al.*, 1994; Elger *et al.*, 2005). As a consequence the Altiplano-Puna high plateau developed, encompassing the highest volcanoes of the world (e.g., 6500 m for Sajama volcano) in the Western Cordillera to the west, rough mountain ranges in the Eastern Cordillera at the eastern boundary and the intra-montaneous drained Altiplano and Puna basins with thick Miocene stratas in-between.

The up to 1800 km long and 350 km wide Altiplano-Puna plateau is the worldwide greatest high plateau in a subduction context. Crustal thickness is enormous (up to 75 km; e.g., Wigger *et al.*, 1994) and decreases beyond the plateau extension to normal values for continental lithosphere of 35-40 km. Likewise volcanism in the CVZ is highly crustally contaminated against more island-arc denoted lavas in the SVZ.

Altiplano and Puna form an unique tectonomorphic entity within the orogen although their morphology differs. While the Altiplano in the north is rather flat and tectonically quiet, the Puna in the south is pervaded by many deep reaching, active horst- and graben structures. Average elevations are around 3700 m in the Altiplano and 4500 m in the Puna.

Actual compressive tectonics are mainly restricted to the Andean foreland, where westvergent thin-skinned thrusting in the Subandean to the north change to thick-skinned thrusting in the Santa Bárbara system and Sierras Pampeanas to the south (Allmendinger *et al.*, 1997). At Altiplano latitudes, the comparatively fast westward subduction of the Brazilian Shield enhances the compressive backarc regime.

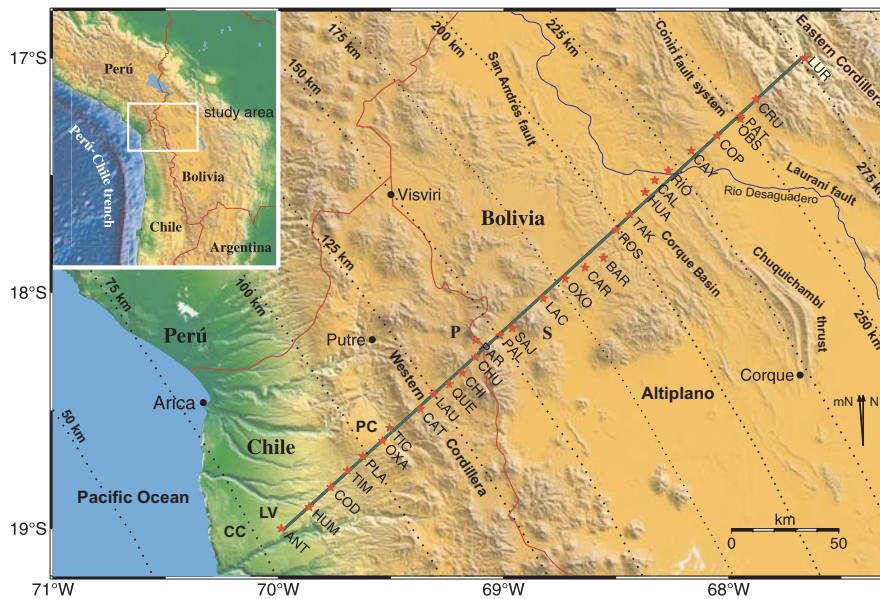


Figure 2: Survey area with locations of the 30 MT sites in the Bolivian orocline. Profile trends roughly perpendicular to the strike of main morphological units and to the contour lines of the Wadati-Benioff zone (dotted lines). CC denotes Coastal Cordillera, LV Longitudinal Valley and PC Precordillera; P, S are Parinacota, Sajama volcanoes; mN is magnetic north.

The survey area

The magnetotelluric profile transverses the central Bolivian orocline from the Coastal Cordillera in northernmost Chile, to the active volcanic arc and the Altiplano in central Bolivia and ends in the Eastern Cordillera. The profile follows a general trend of N48°S which is roughly perpendicular to main structural units as well as to contour lines of the Wadati-Benioff zone (fig. 2).

Main morphological units in the forearc are paralleling the trench and correspond to an ancient magmatic arc which shifted eastward through time, recordable since the Jurassic in the Coastal Cordillera, afterwards in the Longitudinal Valley, in the Precordillera and until the Neogene in the Western Cordillera (Scheuber *et al.*, 1994). The exposure of the forearc crust to periods of extensive and long-lived magmatism gave rise to the formation of deep, long and wide fault zones (sub-)paralleling the trench among which some are still active like the Precordillera Fault System at 19.2-21°S. Here, large copper deposits evidence strong fluid circulation and mineralisation after the eastward shift of the arc 32 Ma ago.

In the closer survey area, the forearc is incised by deep, E-W running valleys which open to the ocean and may have served as Miocene drainage systems. Some of them may be associated with active faults (Wörner *et al.*, 2000). A system of west vergent, steeply dipping thrust faults in the eastern Precordillera, the West-Vergent-Thrust-System, marks the transition to the high plateau and is thought to be the location of major plateau uplift (Muñoz & Charrier, 1996).

The smooth monoclinical western slope of the high plateau is locally shaped by huge land slides, e.g., the Oxaya collapes 12 Ma ago (Wörner *et al.*, 2002). In the West-

ern Cordillera, Plio-Pleistocene to recent, andesitic to rhyodacitic stratovolcanoes are build on thick layers of ignimbrites which are widespread throughout the entire fore-arc and are dated between 22-19 Ma (Oxaya Ignimbrites) and to 2.7 Ma (Lauca-Pérez Ignimbrite) (Wörner *et al.*, 2000).

The Altiplano can be separated into the volcanic Mauri region to the west and in huge sedimentary basins easterly. The San-Andrés fault marks the boundary in-between and limits the Arequipa terrane to the east, a component of the Arequipa-Antofalla massif which comprises up to 2 Ga old rocks and underpins the volcanic arc probably along the whole plateau (see in James & Sacks, 1999). The most prominent feature in the Altiplano is the 80 km wide Corque basin, a deep asymmetric sedimentary basin with thick and folded tertiary strata, exceeding 10 km thickness (Hérail *et al.*, 1997) and evidencing orogeny dynamics during Miocene's pronounced compressive regime.

To the east the basin is controlled by the Chuquichambi thrust system. The Coniri-Laurani fault system marks the transition to the Eastern Cordillera, the rough eastern flank of the plateau. Farther to the east the Interandean Zone and Main Andean Thrust separate the Altiplano plateau from the actual deformation front in the Subandean, with thin-skinned folding and thrusting holding the record of shortening rates within the study area (Allmendinger *et al.*, 1997).

After the Oligocene eastward shift of the magmatic arc, Neogene volcanism first relived in the Eastern Cordillera 27 Ma ago and afterwards encompassed the whole plateau region. Neogene volcanism of the Altiplano plateau is composed of three main units: Pliocene to recent stratovolcanoes are erupting in the main arc and Miocene to recent backarc calderas and mafic monogenetic pulses are distributed in the arc and backarc.

Intensive ignimbritic volcanism erupted 10-4 Ma ago at the Altiplano-Puna transition between 21-24°S and formed one of the greatest ignimbrite provinces of the earth, the Altiplano-Puna-Volcanic-Complex. Recent ignimbrite eruptions are older than 1 Ma. The high water, phenocrystal and silicate content signify long storage in huge crustal magma chambers with pronounced fluidal circulation.

Holocene volcanic activity is less than in the Puna where the lithosphere probably is 50 km thinner (Whitman *et al.*, 1996). Volcanism in the closer survey area is categorised from dormant (Parinacota) to solfataras state with intensive fumarolic activity at Guallatiri volcano. The profile volcano Parinacota may possess a stable magma chamber with only minor magmatic input.

Glance on data analysis and 2-D inversion

We measured horizontal components of the electromagnetic field as well as the vertical magnetic field for long periods between 10 and 20 000 s. Data analysis and its interpretation by 2-D inversion has already been described by Brasse & Eydam (2008a) and will not be replicated in this article.

Some important results are just mentioned: We could not resolve subduction-related features near the slab, like fluid curtains or fluid and melt pathways, due to significant

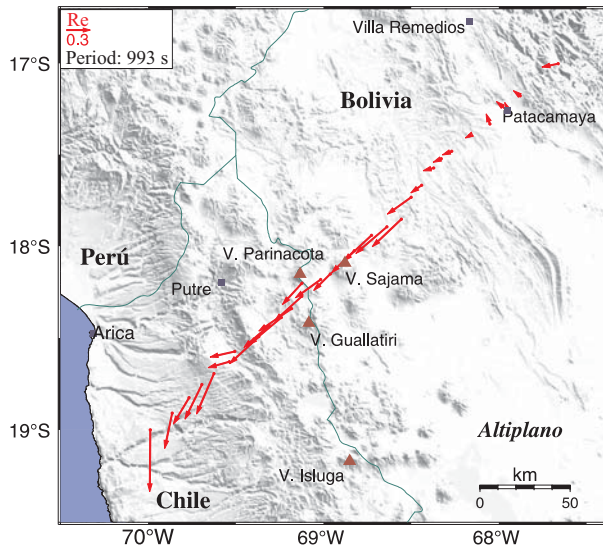


Figure 3: Real part induction arrows in Wiese convention for one chosen period demonstrating the remarkable good alignment of arrows in profile direction. Arrows in the forearc are deflected; they are pointing parallel to the coast.

3-D effects in the forearc. This can easily be seen in the behavior of real part induction arrows pointing parallel to the coast (fig. 3) which is also observed in other coastal regions in Chile so that one might call them 'Chile arrows'. Brasse *et al.* (2008b) showed that they can be explained by anisotropy in the forearc crust. Seven forearc station data were thus discarded from further interpretation by 2-D inversion.

Induction arrows as well as the analysis of impedance data, which reveals phase-sensitive skew values below 0.3 for plateau sites, suggest fairly good electrical 2-D approximation for the plateau region. We observe a remarkable good alignment of middle to long period real part induction arrows in profile direction, as seen in figure 3. Correspondingly, electrical strike derived by impedance data is fairly perpendicular to the profile. Real arrows disappear in the central Altiplano where resistivities are significantly reduced for all sites and rotation angles and follow a well marked downward trend for longer periods indicating assumably well conducting features at greater depths.

2-D inversions were performed with the inversion program of Rodi & Mackie (2001). Accounting for the consistency of induction arrows in the Altiplano, tipper weight was set high by assigning a small (absolute) error floor of 0.02. Error floors for resistivities and phases were set to 20% and 5% in order to overcome the static shift problem. Reliable models are achieved after numerous experiments including tests of starting models, specifying the regularisation term and further resolution tests (see Eydam, 2008).

Discussion on backarc mantle melting

The final model, shown in figure 4, is obtained by jointly inverting tippers, TE and TM mode apparent resistivities and phases. Static effects were accounted for by using the program internal adjustment. Fair regularisation parameters lie around 20. Model RMS is 1.80, with larger misfits at the electrically less 2-D plateau borders (see Brasse & Eydam, 2008a).

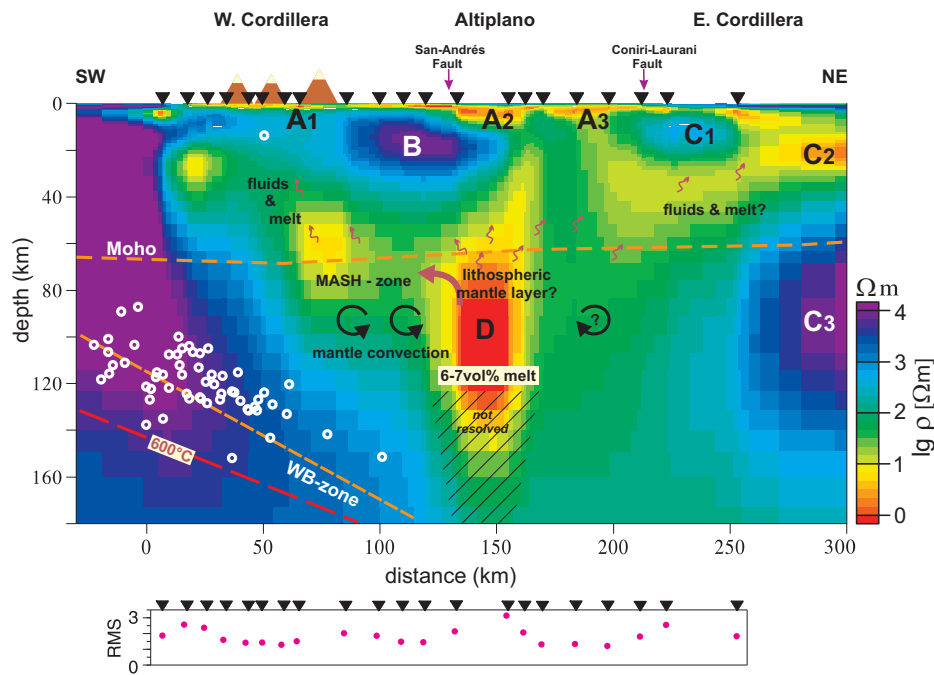


Figure 4: Final resistivity model with suspectable fluid & melt dynamics - Ai: Corque and minor basins, B: Arequipa block(?), Ci: structures beneath the Eastern Cordillera, D: backarc mantle magma reservoir plus rise of fluids and melts. Circles indicate location of mayor earthquakes ($M > 4.5$, Engdahl & Villaseñor, 2002). WB denotes Wadati-Benioff zone. MASH-zone stands for melting, assimilation, storage and homogenisation zone of arc magmas. Below: RMS for the inverted station data.

The large upper mantle conductor D is interpreted as an image of partial melts triggered by fluid influx from the slab. Conductivities range up to above 1 S/m and are definitely required by data down to 115 km depth, particularly to fit the downward trends in long period app. resistivities; they are consistent with data down to slab depths. This implies high melting rates of more than 6vol% (Eydam, 2008).

Lithospheric mantle material should stabilise magma and trap hot fluids which otherwise would rapidly trigger wide scale crustal melting like this is the case in the southern Altiplano (e.g., ANCORP Working Group, 2003). This is consistent with seismic data from Dorbath & Granet (1996) who resolve lithospheric characteristics just beneath the Altiplano Moho.

The source region of subduction related magmatism is offset from the arc by almost 100 km, implying fluid and melt storage near the Moho (in so called MASH-zones) and non-vertical rise of only few fluidal and molten material towards the volcanoes. This accords to low volcanic productivity in the closer survey area, while lavas are bearing highly modified mantle signatures (Wörner *et al.*, 1992). Magma and fluid lateral motion should be enforced by internal convections.

This 'scenario' implies a widely hydrated crust and upper mantle beneath the high plateau which is consistent with other geophysical anomalies, like high heat flow densities (e.g., Springer & Förster, 1998), neagative Bouguer anomalies (Tassara *et al.*, 2006), slow seismic velocities in a thick and highly absorbing crust (e.g., Beck & Zandt, 2002)

	water import	water release in 80-180 km depth			
component of oceanic plate	per m ² Nazca plate [kg]	in percentage of imported water	per subducted meter [kg]	per Ma [10 ⁶ kg] ²⁾	in neogene cycle; ³⁾ WC = VA [10 ⁶ kg]
sediments	1 700	20	340	32	832
crust	170 000	55	93 500	8 883	230 958
mantle	250 000 ¹⁾ 1 000 000 ²⁾	60	150 000 ¹⁾ 600 000 ²⁾	14 250 ¹⁾ 57 000 ²⁾	370 500 ¹⁾ 1 482 000 ²⁾
total	421 700 1 171 700		243 840 693 840	23 165 65 915	602 290 1 713 790

Table 1: Water budget for main dehydration depths in the central Andean subduction zone calculated by using 1) a standardised mantle hydration model after Rüpke *et al.* (2004), 2) a mantle hydration model deduced from local data after Ranero & Sallarès (2004) and 3) by using averaged subduction rates after Somoza (1998). WC - Western Cordillera, VA - main volcanic arc.

and a strong mantle signature of geothermal fluids (Hoke *et al.*, 1994).

Estimation of arc and backarc water release

A critical mind might state that the ability of water import is probably not sufficient to nurture such wide scale melting of mantle material. Table 1 shows the water balance for slab dewatering at main dehydration depths between 80-180 km, deduced from intermediate depths microseismicity published in David (2007) and seen in figure 7.

For the calculation we used hydration estimates after Rüpke *et al.* (2004) modified by local data (fig. 5). Mantle hydration is poorly definable and depends on depth and density of bending related faults as well as on the thermal structure of the oceanic lithosphere. Therefore estimates based on regional seismic and gravimetric data evaluated by Ranero & Sallarès (2004) are four times higher than those assumed by Rüpke *et al.* (2004) who standardised hydration levels with regards to the age of the lithosphere. Water balances are presented for both models.

The presented water releases are based on results of chemo-thermo-mechanical modellings performed by Rüpke *et al.* (2004) who solved for slab dehydration during sub-

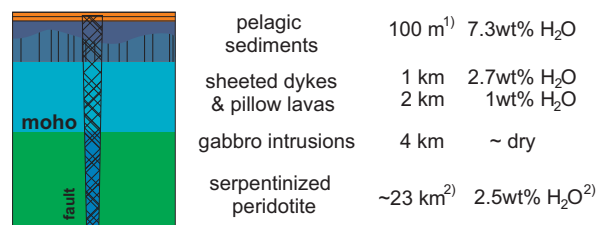


Figure 5: Oceanic lithosphere with estimated water content in sediments, crust and mantle after Rüpke *et al.* (2004) modified by local data: 1) the thickness of sedimentary deposits after von Huene *et al.* (1999) and 2) mantle hydration after Ranero & Sallarès (2004).

duction regarding latent heat consumption and neglecting shear heating¹ (fig. 6).

Thereafter, one column of one square metre Nazca lithosphere releases 700 t water at main dehydration depths (80-180 km) which sums up to 1.7 Gt water regarding the entire Neogene cycle of Nazca plate subduction starting 26 Ma ago (Sempere *et al.*, 1990). Using the molar volume of water at 120 km depths, this mass fills a cross-section of more than 1500 km² which finally lies in the dimension of the mantle magma reservoir.

Phases of oroclinal bending might have been important for the water budget, because bending related faults should have opened parallel as well as perpendicular to the trench so that large amounts of water could have been fixed in the mantle.

Slab-derived water in the mantle wedge should thus be sufficient to trigger a wide scale melting of mantle material, well noticing the presumably prior hydration of the Western Cordillera as pre-Neogene backarc. If the residual water, retained in the oceanic lithosphere, is released as well, the cross-section will expand for additional 170 km² in case of complete crustal dehydration resp. 880 km² for complete mantle

dehydration which points out the role of oceanic mantle for subduction zone dynamics.

The double seismic layer shown in figure 7 allows allocating the events to stability limits of mayor hydrous minerals in the oceanic lithosphere which can be used to predict slab temperatures at depths.

Crustal seismicity stops around 140 km depth, whereas seismicity in the oceanic mantle is ongoing until the stability limit of serpentine in 180 km. Lawsonite is the only hydrous mineral in the crust which is stable down to 140 km depth. Lawsonite decay starts at the hot interplate boundary at shallower depths and proceeds during subduction to deeper slab layers where temperatures at 140 km depth should exceed 730°C. Serpentine as the main hydrous mantle phase should still be present at 180 km depth where it transforms through

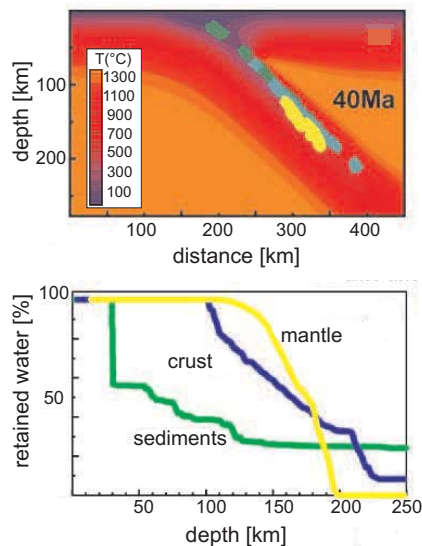


Figure 6: Dehydration of oceanic lithosphere during subduction, presented above) as main depths' water release of sediments, crust and mantle and below) as percentage of retained water respectively; after Rüpke *et al.* (2004).

a water-conserving reaction to mantle phase A (Eydam, 2008).

These estimates correspond to a fairly warm subduction system although the Nazca plate is quite old (55 Ma). Shear heating at the interplate boundary may elevate temperatures in the oceanic crust, because lubricates like sediments are sparse in the

¹Modifying the mantle hydration model to higher estimates after Ranero & Sallarès (2004) will change the modelling results to even larger water releases than calculated in table 1, considering that dehydrations are mostly exothermic reactions.

central Andean subduction zone as a consequence of its arid climate (Lamb & Davis, 2003). Thus more water than listed in table 1 may be released into the mantle wedge.

Dehydrations finish below 200 km depth, suggesting that slab-derived fluids supply partial melting of mantle peridotite more than 60 km laterally away from their source region. Might this be possible?

Discussion on fluid mobility

There are several models describing fluid mobility in subduction zones like fluid channeling along cracks opened by dehydration induced hydrofracturing, lateral transports in the coupled mantle wedge, porous flows and thermally induced convections in the asthenospheric wedge.

At considered depths hydrofractured faults shouldn't reach far into the mantle and a coupled mantle wedge should be unrealistically thick to explain the requested transportation lengths. Porous flows start when interstitial fluids or melts interconnect depending on temperature, fluid salinity and melt viscosity. Deep slab-derived fluids are highly saline (Scambelluri & Philippot, 2001) and temperatures below 140 km depth definitively exceed 620°C (see above), the minimum temperature of pure water interconnection in an olivine matrix (Mibe *et al.*, 1999).

Water enhances mantle convection via lowering the peridotitic solidus in the mantle wedge. This effect may be more pronounced than expected so far, considering the very slow reaction kinetics of olivine melting in laboratory experiments (Grove *et al.*, 2006). The correction of peridotitic solidus to much lower temperatures, e.g. to 860°C at 100 km depth, implies that first water-enriched melts may be generated just above the slab. The melt fraction should be minor due to fast segregation of the low-viscous melts into lower and hotter mantle regions.

Additionally, water import in the central Bolivian orocline might have been temporarily elevated in phases of its oroclinal bending leading to a significant reduction of mantle viscosity and enabling effective material transports via thermal convections, like it is modelled for some general case studies by Gerya *et al.* (2006). Saline fluids and melts may be transported far into the hotter backarc mantle where melt fractions rise and segregation rates decrease and fluids and melts become detectable by magnetotellurics.

Models of mantle convection deal with numerous low constrained parameters, like the degree of mantle hydration, shear heating at the interplate boundary, latent heat produced by mineral reactions or advective heat transports by fluids and melts. Therefore mantle rheology is only poorly definable and viscosity values may differ over several magnitudes.

Water release at intermediate depths plays thus an important role in subduction zone dynamics. Currie & Hyndman (2006) even postulate that hot backarc mantles are a fundamental characteristic of an ocean-continent subduction zone. And indeed, other magnetotelluric images from the central Andean backarc mantle seem to concur with this: Schwarz & Krüger (1997) modelled 0.02 S/m conductive mantle features at the

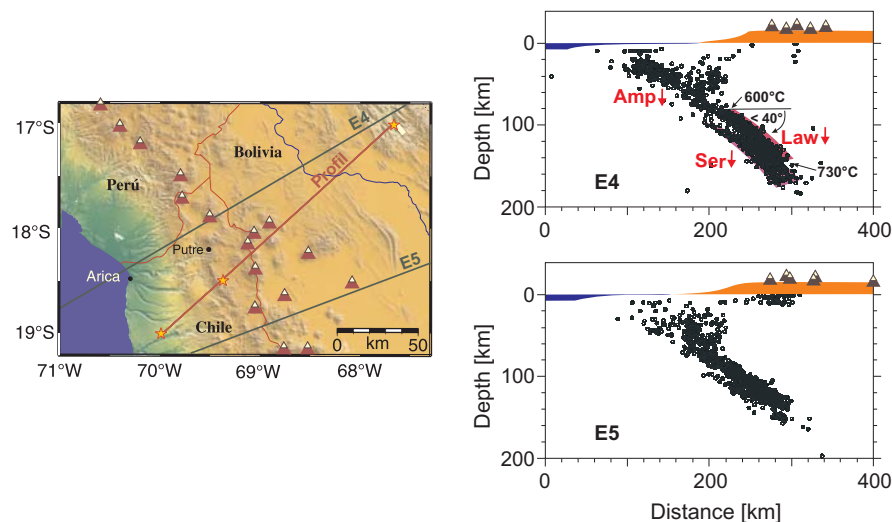


Figure 7: Local microseismic events ($M > 1$) after David (2007) with map of projection lines E4, E5 and MT profile. Above: Events allocated to decay temperatures of hydrous minerals, this are in the crust: Amp - amphibole and Law - lawsonite and in the oceanic mantle: Ser - serpentine. Note the double seismic layer.

transition to the Puna around 23°S and northwards at $20\text{--}21^{\circ}\text{S}$ upper mantle material seems to be quite conductive as well with values over 0.05 S/m (Lezaeta, 2001), though electromagnetic fields are severely attenuated by a crustal high conductivity zone here (Brasse *et al.*, 2002).

Discussion on the mobility of the slab itself

The Nazca plate subducts along the whole margin with small dip angles between $25\text{--}30^{\circ}$ or even horizontally in greater depths at some segments (so called 'flat subduction') where volcanism above is inactive.

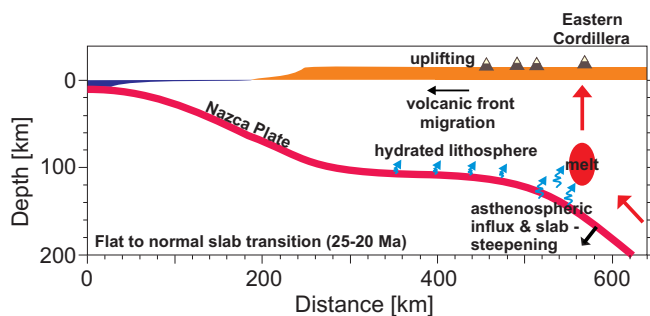


Figure 8: 'Mobile slab model' for the central Andes after James & Sacks (1999). Due to retarded dehydration at intermediate depths, slab flattens and dewateres sparsely, insufficient to melt but to hydrate the base of the overlying lithosphere. Slab compacts and steepens again when hot asthenospheric material influx and triggers effective dehydrations; magmatism and volcanism above develop.

Thus, slab pull and buoyancy forces are fairly equilibrated and the slab may react sensible to density changes due to retarded dehydrations; it gets mobile in geological time scales.

Such 'chemically controlled slab mobility' was already discussed by James & Sacks (1999) and applied on central Andean history by supposing a period of flat subduction 35-25 Ma ago, basically due to absent lava dates for this period (fig. 8).

Remarkable is the slight slab steepening in the central orocline (fig. 7).

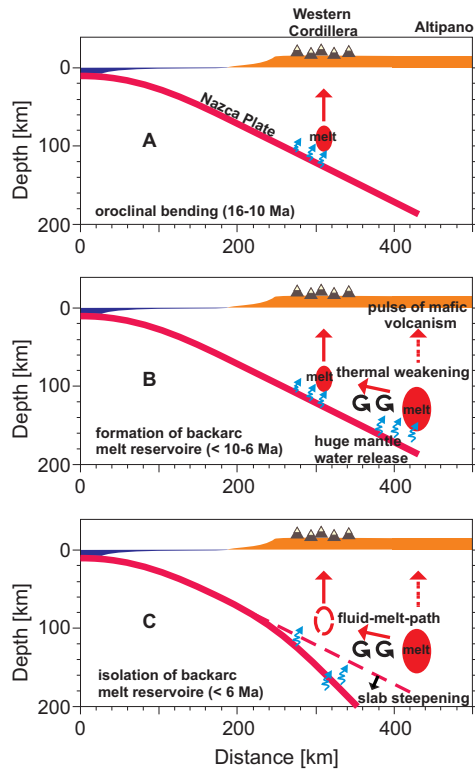


Figure 9: Attempt to reconstruct Neogene Nazca plate subduction in the Bolivian orocline:

A) Increased hydration of oceanic Nazca plate mantle due to significant oroclinal bending.

B) 2-4 Ma later, highly hydrated mantle reaches dehydration depths of serpentine (100-180 km) where the released fluids trigger high-grade partial melting in the backarc. Backarc melts rise vertically as well as non-vertically to the arc via internal convections.

C) Slab steepens during dehydration. The voluminous backarc reservoir persists but is maintained by less fluids. Volcanic productivity decreases.

Tracing the regression line of shallow Wadati-Benioff events ($z < 100$ km) to greater depths, this line cuts the mantle conductor at dehydration depths of serpentine ($z = 180$ km), the most important hydrous mantle mineral which may bound huge amounts of water. Therefore backarc peridotitic melting may have been triggered just above the slab by enormous fluid influx whereby the slab was compacting and steepening slowly to its actual position (fig. 9). Today, the magma reservoir should be maintained by much less fluidal material and probably cools down.

This scenario implies a prior period of extensive oceanic mantle hydration which might have occurred during phases of oroclinal bending. Bending dates are hard to fix, probably block rotations in the closer survey area give a hint, they were dated at 12 Ma (Wörner *et al.*, 2000).

Using an average for ancient subduction rates of 10 cm/a, evaluated by Somoza (1998), and a constant slab dip of 25° , those huge amounts of fixed water will have started to deliberate after 2.4 Ma of subduction ($z \approx 100$ km) and will have reached depths of high-grade partial backarc melting ($z \approx 180$ km) further 1.9 Ma later. Surficial expression of high-grade mantle melting are Pliocene to recent pulses of mafic andesites erupting in the Bolivian Altiplano backarc (Davidson & de Silva, 1994). Furthermore backarc mantle magma might have been transported via internal convections to the sub-arc region and have also erupted at the main volcanic arc. However, the basal input into the deep magma system of the recent strato-volcanoes is only minor (Wörner *et al.*, 2000).

Conclusion

The huge backarc mantle conductor imaged by 2-D inversion of magnetotelluric data from the central Bolivian orocline inspired reflections about fluid and melt dynamics in the central Andean subduction zone. Among the two approaches considered here to explain the remarkable offset between intermediate depths microseismic events and the imaged magma reservoir, sub-arc and -backarc water release plays a key role for subduction zone dynamics.

In the central Bolivian orocline huge amounts of water might have been released at main slab dehydration depths some million years after oroclinal bending occurred where faults should have opened not only parallel but also perpendicular to the trench allowing deep water infiltration and fixation in the oceanic mantle. This implies on the one hand that mantle viscosity is definitely lowered and leads to significant material transports due to thermal convections, whereby fluids and melts may be widespreaded in the arc and backarc mantle. On the other hand, enhanced slab dehydration and slab compacting may have been followed by slab steepening and the imaged mantle melting process isn't supplied by much fluidal material anymore and therefore probably cools down.

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Thin Sheet Conductance Models from Geomagnetic Induction Data: Application to Induction Anomalies at the Transition from the Bohemian Massif to the West Carpathians

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Abstract

Thin sheet approximation of Earth's conductive structures made it possible to quantitatively estimate effects of lateral conductivity variations in the Earth long before full 3-D electromagnetic modeling was practicable. The thin sheet approach is still useful when induction data with limited vertical resolution are to be interpreted on a surface. It especially refers to collections of long period induction arrows across large areas and geological units. For this purpose, inverse procedures, both linearized and stochastic, for a conductance distribution in a thin sheet have been suggested recently. We present a stochastic Monte Carlo inversion of geomagnetic induction data based on the bayesian formulation of the inverse problem. An example of the inversion of practical induction data from the transition zone between the Bohemian Massif and the West Carpathians suggests that an SW-NE anomalous induction zone observed above the eastern slopes of the Bohemian Massif admits explanation in terms of a phantom effect due to the superposition of fields of the strong SE-NW Carpathian conductivity anomaly to the east with NW-SE to W-E trending conductivity zones to the west that conform with the fault pattern of the eastern Bohemian Massif.

1 Introduction

Regional geoelectrical information on the contact zone between the Variscan Bohemian Massif and the Alpine Western Carpathians is mainly available from long-period geomagnetic transfer functions. Magnetotelluric data and broadband electromagnetic induction experiments are scarce in the region, mainly because of high level of the civilization noise all over the area, and also because of the lack of instrumentation in the past. Only recently, magnetotelluric experiments have been carried out in some subareas of the region, mainly for commercial targets (e.g., Vozar, 2005).

Long-period geomagnetic induction data covering a period band of about one decade in the range of thousands of seconds cannot provide detailed geoelectrical information on structures beneath the region of interest. They are, however, suitable characteristics to be used to model large-scale horizontal conductivity distribution in the Earth's crust or lithosphere, and thus to indicate regional lateral conductivity anomalies that are responsible for the observed induction pattern over the area under study. With data at periods of the order of thousands of seconds, with penetration depths starting at several tens of km for standard conditions of the continental lithosphere, a quasi 3-D thin sheet approximation of crustal conductivity structures is a reasonable induction model (e.g., Vasseur and Weidelt, 1977). The thin sheet approximation largely reduces the computational demands of the modeling procedure as compared with a full 3-D treatment and, moreover, bypasses some intrinsic difficulties of dealing with the geomagnetic induction data alone, especially their low sensitivity with respect to the normal layered background of the model.

Recently, two methods of inversion of geomagnetic induction data for conductances in a thin sheet have been developed, one based on the non-linear conjugate gradient technique (Wang and Lilley, 2002) and the other on a Bayesian approach with Markov chain Monte Carlo (MCMC) method used for a stochastic sampling process (Grandis, 2002). In this contribution, we are using the latter approach to analyze conductance distributions which are compatible with the geomagnetic induction data in the West Carpathians region and at its transition to the Bohemian Massif on the territories of the Czech Republic, Slovakia and Poland.

More specifically, the data base of this study are induction response data at 150 field stations that were published earlier (Prašný and Pěšáček, 1991) covering the Bohemian Massif (BM), the Brunovistulicum (BV), and the West Carpathian region (WCP) and re-analyzed in (Pěšáček and Prašný, 1996). Spatial distribution of both the in-phase and the out-of phase induction vectors, contour maps of individual transfer functions (TF) and contour maps of anomalous

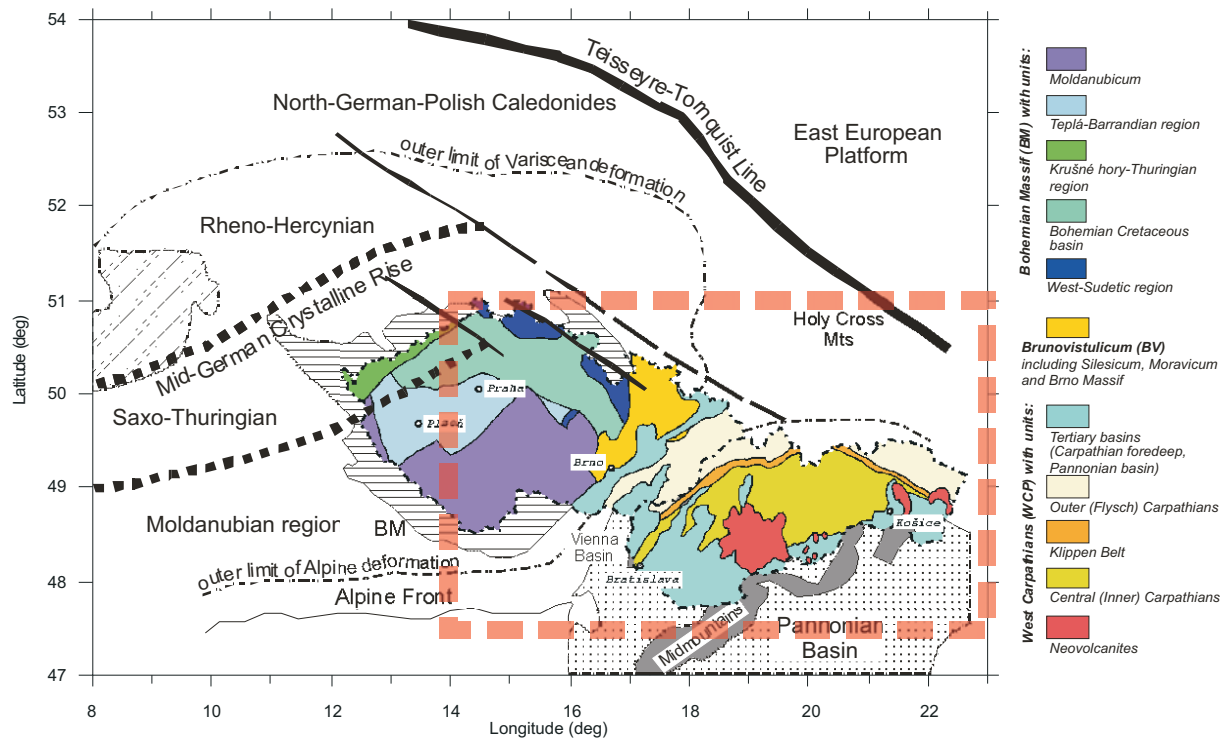


Figure 1: Geological scheme of the region of Central Europe relevant to our modeling study. The dashed-line rectangle shows the region in which the inversion for a laterally variable conductance distribution in a thin sheet is carried out.

vertical field that were generated from the TF's by the hypothetical field of different polarizations and systems of internal anomalous currents indicate the existence of two zones of anomalous induction at the eastern margin of the BM and near the boundary of the Carpathian plate (Kováčik *et al.*, 1997). The previous analyses, and the recent one performed by a new approach to imaging the induction data of the Wiese vectors at almost 1800 localities covering mainly the Central European area (Wybraniec *et al.*, 1999) suggest that these anomalies might be connected with the North-German-Polish anomalous zone, representing an important part of the Trans-European Suture Zone (TESZ). The analysis of certain models of electrical conductivity distribution is performed to fit the anomalous features of the induction response data over the Central European area, specifically zones of anomalous induction in the eastern margin of the BM, across the entire block of the BV and near the margin of the Carpathian tectonic plate.

The structure of the article is as follows: In Section 2, we give a short overview of the principal geological units of the region under study. Section 3 briefly summarizes the geoelectrical features of the region previously inferred from the geomagnetic induction data in the region. In Section 4, we formulate the thin sheet model used in the inversion. The principles of the Bayesian inversion of the geomagnetic induction data for the conductance distribution in the thin sheet are presented in Section 5. Section 6 then summarizes outputs of the Bayesian inversion for the induction data over the BM/BV/WCP region with indications on possible correlations of the conductance model with regional geological structures.

2 Geological context of the induction studies

Fundamental elements of the central European geological structure that are relevant to our modeling experiment are schematically displayed in Fig. 1 together with the major geological units over the Czech and Slovak Republics covered by induction response data involved in the modeling process.

(i) The Tornquist-Teisseyre tectonic zone (TTZ) constitutes one substantial part of a number of fault zones and sutures found within the Trans-European Suture Zone (TESZ) that represents the most important geological boundary in Europe separating mobile Phanerozoic western terranes (Meso-Europe) from the Precambrian east European Craton. It is as clearly defined in the deep lithosphere as in the upper crust, Moho depths increase across this zone from 30 km beneath Variscan Europe to 45 km beneath the East European Craton. In contrast to the relatively cold eastern craton relatively high heat flow characterises Western Europe.

(ii) The Bohemian Massif (BM) represents the easternmost consolidated block of the Variscan branch of the European Hercynides (Meso-Europe) that builds up Bohemia and the western part of Moravia. The major elements of

the BM are several SW-NE trending zones separated usually by deep-seated faults, which are reflected in the results of the deep seismic profiling, in the gravity, geomagnetic and heat flow maps (Suk *et al.*, 1984). The fault structures are essentially parallel with the boundaries of individual Hercynian zones. The intersection of these zones with the second order system of NW-SE trending faults is responsible for the complicated block structure of the BM.

(iii) The Carpathians belong to the young Tertiary Alpine-orogenic belt and they constitute the NE branch of the Alpides. Their boundary with the Eastern Alps runs along the Danube Valley. The northern boundary with the East European Platform is defined by the erosive margin of the flysch nappes. Our model covers the West Carpathians (WCP) and we distinguish here the inner (central) Carpathians, the Klippen zone, the outer (Flysch) Carpathians and the Carpathian foredeep (Fig. 1, dashed-line rectangle).

(iv) The Brno unit, termed recently the Brunovistulicum (BV) (Dudek, 1980; Suk, 1995) lies in between those previously mentioned structural elements. The BV unit is assumed to be an independent geological structure forming the Precambrian (Cadomian) basement (Palaeo-Europe) of both the eastern part of the Hercynides (Variscides) of the BM and the Alpides, i.e. of the WCP in Moravia. Recent geophysical and geological data have shown that the BV and the whole Moravian block occupy a highly independent position and form a separate geological unit belonging probably to the Fenno-Sarmatian Platform (Dudek, 1980).

3 Geoelectrical data and features of the region

3.1 The data

The experimental data were collected in 1970's and 1980's in a series of field experiments organized in co-operation with Czech, Slovak and Polish colleagues. The field measurements consisted in analogue recordings of magnetic transient variations at altogether 150 temporary field sites over an area of about $500 \times 250 \text{ km}^2$. They were analysed in terms of induction arrows (Wiese, 1962). At each station, in-phase and out-of phase induction arrows have been estimated for periods in the range of 1200 to 5840 s. They correspond to the real and imaginary components of single-station transfer functions between the horizontal and vertical components of the transient magnetic field at the station. A sample of the real and imaginary induction arrows at a particular period of $T = 3860 \text{ s}$ is presented in Fig. 2.

3.2 Major conductivity anomalies

As the magnetic Z -component is known to be highly sensitive to laterally inhomogeneous distribution of the internal electrical conductivity, maps of these vectors provide us with a view of the changing anomalous behaviour of the Z -variation as function of frequency and location. Reversals of the arrows distinguish zones of anomalous induction, which often mark important geological features such as contacts between blocks with different geological histories of development, zones of past and recent tectonic activities, collision zones and etc. A qualitative analysis of the maps of induction arrows led to evidence two major conductivity anomalies in the area under study. Quantitative modeling allowed to characterize the conductivity distribution in the lithosphere that accounts for the observed induction arrows.

The Carpathian anomaly, WCA The induction vectors across the WCP region show a clear perpendicular orientation with respect to a general trend of the anomalous induction zone localized along the external margin of the Carpathians Mts. chain (see WCA in Fig. 2). They show almost a perfect 180° reversals of their azimuths above the anomalous induction zone. In the WCP, modules of the induction vectors situated to the south from the zero line are by about 25-50% larger than the corresponding induction vectors located to the north of the anomaly. The Carpathian

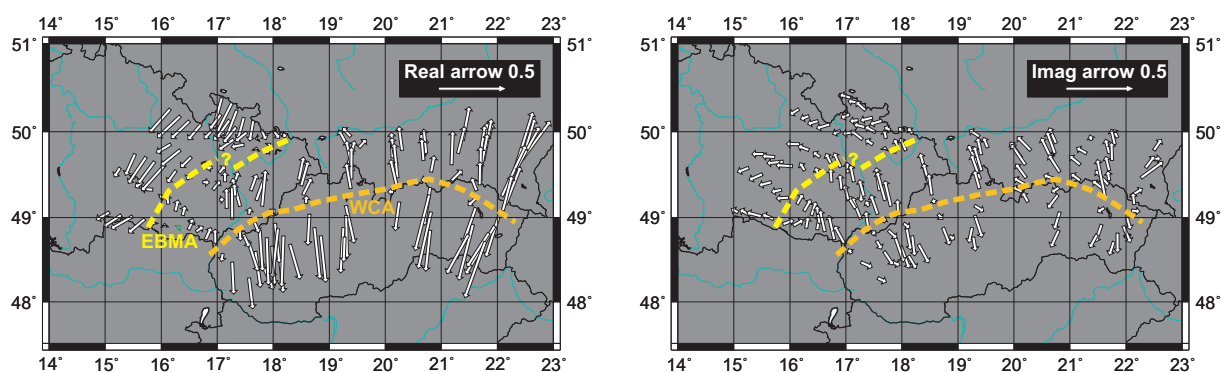


Figure 2: Sample of experimental real and imaginary induction arrows in the BM/BV/WCP region for the period of 3860 s, with two main regional conductivity anomalies indicated, the West Carpathian conductivity anomaly (WCA) and the conductivity anomaly on the eastern margin of the Bohemian Massif (EBMA).

geoelectrical anomaly is constrained by the presence of high-conductivity rock series at depth of 10-25 km in different segments of the orogen. Different geological models explaining the presence of highly conductive rocks or solutions as sources of the anomalies have been suggested. The role of altered and/or fractured rocks, saturated with hot mineral waters, as well as the proximity of anomaly sources to boundaries of different crustal blocks, including those between the Carpathians and adjacent platforms, have been taken into account (Čadim and Pospíšil, 1984; Jankowski *et al.*, 1984; 1991; 2008; Hvozdára and Vozár, 2004). Other authors discuss possible connection between the Carpathian anomaly and the presence of metamorphosed coal bearing Carboniferous strata beneath the orogen or relation to the graphitized rocks occurring close to the boundaries of crustal blocks (Glover and Vine, 1992; Glover and Vine, 1995; Žytko, 1997).

The EBMA anomaly The induction vectors on the eastern margin of the BM are distinguished by the vectors oriented predominantly parallel to the general SW-NE trend of the anomalous zone. Also the reversal of the azimuths at the anomalous zone is rather poorly developed, only sudden changes of azimuths are observed at individual profiles. These facts are clear indications of a 3-D character of the conductivity distribution. The EBMA has been generally attributed to processes in the subduction of the eastern margin of the BM beneath the WCP plate, but precise induction processes for generating the peculiar 3-D features of this anomaly are still unknown (e.g. Kovčík *et al.*, 2005). Speculating on the physical/geological sources of the anomaly may thus be premature.

3.3 Results from numerical modeling

From the equivalent current systems (Červ *et al.*, 1997) it was concluded that the source depth of the induction anomalies can be about 18 km in the WCP region and about 10 km in the EBM/BV. These estimates are suggesting the source of the anomalies at shallower depth than those obtained previously by separating the magnetic field variations into internal and external parts (Pěčová and Praus, 1996) and applying the line current approximation (Jankowski *et al.*, 1985).

2-D models for induction vectors along the profiles crossing the WCP are summarized in (Jankowski *et al.*, 1985). The models featured anomalous bodies with a cross-section $\times$ conductivity parameter of the order of 10^7 to 10^8 Sm, with the top of the bodies at depths beneath 12-15 km. In (Jankowski *et al.*, 1991) the 2-D modeling was used for simultaneous modeling of the induction vectors and apparent resistivities, collected in a series of sites in the Polish section of the Carpathian Foredeep. In the latter models, the source of the WCA was situated at shallower depths, less than about 10 km, and hypothesized to be related to deep sediments of the Carpathian Foredeep. The 2-D inversion on the Carpathian data was firstly used in (Červ and Pek, 1981). The geoelectrical structural model along the DSS profile No VI, crossing the BM, BV and WCP, based on the MT and MV results was presented in (Červ *et al.*, 1984).

3.4 Depth of the asthenosphere

The electrical asthenosphere, if present, is an additional structural feature that can affect the induction data especially at long periods. In the model derived from the P-wave residuals for the Bohemian Massif the depth of the lithosphere-asthenosphere transition zone are between 90 and 140 km (Babuška *et al.*, 1988). From MT sounding a layer of increased electrical conductivity attributable to the asthenosphere was interpreted at depth between 100 and 150 km (Červ *et al.*, 1984).

The most likely depth of the conducting layer in the upper mantle in the Pannonian Basin region are between 60 and 85 km in the central part of the depression. The depths seem to increase towards the flanks of the basin to about 100 km (Čadim, 1976). In some parts (Rába-Rónava tectonic line) the thickness of the lithosphere reduces to even less than 60 km (Čadim, 1988). The regions of lithosphere thinning penetrate from the Pannonian Basin into inner parts of the WCP in several promontories. In the thinned part of the lithosphere in the West Carpathians the lithosphere-asthenosphere transition zone is at the depth 90-120 km (Červ *et al.*, 1984).

A gradual increase of the lithosphere thickness to 140-180 km occurs in the Outer Carpathians and farther towards the margin of the East European Platform (Praus *et al.*, 1990). In the Alps the depth of the asthenosphere varies from 100 to 200 km (Praus *et al.*, 1990).

4 A possible thin sheet model

In the subsequent inversion for a laterally non-uniform conductance, we will use a thin sheet model formally defined as follows:

Let us consider a model consisting in a heterogeneous thin sheet at the surface of or embedded in a 1-D medium, hereafter called normal model. The normal model is defined by the conductivities $\sigma_n(\mathbf{r})$ at any point $\mathbf{r}$. In the thin sheet, the actual conductivity $\sigma(\mathbf{r})$ differs from the normal one in a domain of interest Ω . Inside Ω , $\sigma_a(\mathbf{r})$ denotes the

difference $\sigma(\mathbf{r}) - \sigma_n(\mathbf{r})$: $\sigma_a(\mathbf{r})$ is the anomalous conductivity that is zero outside Ω . Solutions of the forward and inverse problems have already been published for such a model. They are briefly recalled in the next section.

With regard to the experimental geomagnetic TF's available, several aspects of the thin sheet model must be considered. First, the validity of the thin sheet hypothesis must be assessed. We assume that the anomalous electrical targets are localized mainly in the crust. Penetration depths for typical continental crustal sections should not be less than about 50 km in the period range of thousands of seconds (corresponds to the resistivity of 10 Ωm and period of 1000 s). Thus, the anomalous zone of 20-30 km below the Earth's surface can be reasonably (though, may be, not completely unquestionably for highly conductive anomalous zones) approximated by a thin sheet. According to Bruton (1994), the size of the square tiles a used to discretize the anomalous subdomain Ω of the sheet should meet the condition $a\omega\mu S_{\max} \ll 1$. For the most extreme parameters in our models, say $T_{\min} = 1200$ s and $S_{\max} = 5000$ siemens, the size of the tiles should be much less than about 30 km. We typically use tiles with $a = 20$ to 25 km in our models, which meets the Bruton's condition for most of the model situations.

The second aspect is the depth of the thin sheet. As we are not able to employ a multiple sheet model in the inversion at present, we use a single thin sheet that integrates all the conductivity anomalies from the surface down to the crust. Thus, the sheet is situated on the surface of the Earth in our experiments. This may result in increased misfit, especially in imaginary induction arrows, if the anomalous current flow deep in the crust.

The third aspect is the embedding normal structure. Due to large differences of the depth vs. resistivity sections of the two main region of our study, the BM and the WCP with the adjacent Pannonian Basin, we cannot suggest any single 1-D normal model down to the asthenospheric depths for the whole region. Therefore, we simplified the normal model into a two-layer structure with a poor conductor resistivity of several hundreds of Ωm up to 1000 Ωm underlaid by a more conductive asthenosphere with resistivity within the range of 100-500 Ωm . The effect of the topography of the asthenospheric layer was checked independently by a 3-D modeling experiment with seismic data (Praus *et al.*, 1990) taken to approximate the top of the asthenosphere. The checks showed that the effect of the asthenosphere is negligible in induction arrows for periods of the order of thousands of seconds unless the resistivity of the asthenosphere is less than about 10 Ωm .

5 Bayesian Monte Carlo Markov Chain thin sheet inversion

We present in this section the Bayesian Monte Carlo Markov Chain (MCMC) method we used to solve the inverse problem. For the sake of self-completeness, let us first briefly recall basic notions concerning Bayesian inversion and Markov chains behaviour. The readers are referred to Roussignol *et al.* (1993), Menvielle and Roussignol (1995), and Grandis *et al.* (1999; 2002), and references therein for more details.

5.1 The Bayesian approach

Let the a priori knowledge be the information available on the model before processing the data, and the a posteriori knowledge the information available after processing the data. A priori and a posteriori distributions account in a probabilistic way for the a priori and a posteriori knowledge respectively. In the Bayesian context, solving the inverse problem thus comes down to determining the a posteriori knowledge by updating the a priori knowledge with the information gleaned from the data (Box et Tiao, 1973; Berger, 1985; Press et al., 1989).

Solving the inverse problem first requires the direct problem to be solved. Let F be the direct problem function which enables computation of the observations $\mathbf{d}$ for a model $\mathbf{m}$. Assuming that the error $\delta\mathbf{d}$ is only related to the data acquisition, it becomes

$$\mathbf{d} = F(\mathbf{m}) + \delta\mathbf{d}$$

The a priori knowledge is given by a probability distribution function (*pdf*) $P_0(\mathbf{M} = \mathbf{m} | \mathbf{M} \in \mathcal{M})$ defined on the set of possible models $\mathcal{M}$. The a posteriori probability for the parameter vector $\mathbf{M}$ to take the value $\mathbf{m}$ given the observations $\mathbf{d}$ is given by Bayes' formula (Bayes, 1763; also see, e.g., Bolstad, 2004, for a more modern treatment). Noting $P(\mathbf{M} = \mathbf{m} | \mathbf{D} = \mathbf{d}, \mathbf{M} \in \mathcal{M})$ the conditional probability of $\mathbf{M}$ given $\mathbf{D}$, and assuming that the error $\delta\mathbf{d}$ is gaussian with standard deviation τ , it becomes

$$P(\mathbf{M} = \mathbf{m} | \mathbf{D} = \mathbf{d}, \mathbf{M} \in \mathcal{M}) = \frac{\exp \left[-\frac{\|F(\mathbf{m}) - \mathbf{d}\|^2}{2\tau^2} \right] P_0(\mathbf{m})}{\sum_{\mathbf{m} \in \mathcal{M}} \exp \left[-\frac{\|F(\mathbf{m}) - \mathbf{d}\|^2}{2\tau^2} \right] P_0(\mathbf{m})} \quad (1)$$

The value $\mathbf{d}$ of the random vector $\mathbf{D}$ corresponds to the observed data. Since $\mathbf{d}$ is not modified during the inversion process, we will from now on omit $\mathbf{D}$ in the expression of the probability distributions and denote the a posteriori *pdf* $P(\mathbf{M} = \mathbf{m} | \mathbf{D} = \mathbf{d}, \mathbf{M} \in \mathcal{M})$ simply by $\Pi(\mathbf{m})$.

The normalization constant, which appears in the denominator of eq. (1) is a sum over the set of possible models, the dimension of which is very large: for instance, it is equal to $M \times L$ for models with M parameters that can each take on L different values. It is actually very difficult, if not impossible to estimate directly. We choose to use a Monte Carlo Markov Chain (MCMC) simulation method to achieve this estimation.

5.2 Markov chains

In order to estimate the *pdf* $\Pi(\mathbf{m})$, let us define a Markov Chain on the set $\mathcal{M}$ of possible models that have $\Pi(\mathbf{m})$ as invariant probability. The marginal *pdf*'s of $\Pi(\mathbf{m})$ will be estimated by empirical averages from simulations of the Markov Chain.

Consider a system which can take a certain number of states and evolves at random with time. At a given time, the state of the system can be described as a random variable $\mathbf{M}$, the values of which, $\mathbf{m}$, belong to $\mathcal{M}$. The system is a Markovian process if, at any time, its future evolution depends only on its present state. It means that a Markovian system depends only on its past through its present state. Markov chains are a particular class of Markovian processes such that (i) the set $\mathcal{M}$ is finite or countable, and (ii) the successive times for the evolution are denoted by integers. A series $\{\mathbf{M}(n), n = 0, 1, 2, \dots, N\}$ of random variables with values in a finite or countable space is then a Markov chain if the state of the system at the time $n + 1$, $\mathbf{M}(n + 1)$, depends only on its past through the state of the system at time n , $\mathbf{M}(n)$. The behaviour of a Markov chain is defined by the set of its transition probabilities,

$$P_n(\mathbf{m}, \mathbf{m}') = P(\mathbf{M}(n + 1) = \mathbf{m}' | \mathbf{M}(n) = \mathbf{m}),$$

where $\mathbf{m}$ and $\mathbf{m}'$ belong to $\mathcal{M}$. If (i) the transition probabilities do not depend on time n , (ii) $\mathcal{M}$ is finite, and (iii) each possible state can be reached from any other, the chain is a homogeneous ergodic aperiodic Markov chain, and there exists one, and only one probability density function Π on $\mathcal{M}$ which is invariant for the Markov chain. When n increases towards infinity, the behavior of ergodic Markov chains is such that the average fraction of time at which the chain is at a state $\mathbf{m}$ ($\mathbf{m} \in \mathcal{M}$) tends towards the invariant probability of $\mathbf{m}$, i.e. $\Pi(\mathbf{m})$.

Methods for Bayesian inversion with Markov chains has been proposed for the 1-D magnetotelluric (Grandis *et al.*, 1999) and DC problems (Schott *et al.*, 1999) with the a priori of smooth variation of resistivity with depth, and for the thin sheet approximation (Roussignol *et al.* 1993; Grandis, 1994; Grandis *et al.*, 2002). The present analysis of crustal conductivity over the transition from the Bohemian Massif to the West Carpathians relies on the inversion of induction arrows using the latter method, which is briefly described in the following sections.

5.3 The thin sheet forward problem

Let us consider a model consisting of a heterogeneous thin sheet at the surface of or embedded in a 1-D medium, specified in detail in the first two paragraphs of Section 4 above. Let $\mathbf{E}(\mathbf{r}, \omega, \sigma)$ and $\mathbf{H}(\mathbf{r}, \omega, \sigma)$ be the time Fourier transforms of the electric and magnetic field at point $\mathbf{r}$ and circular frequency ω , $\omega = 2\pi/T$, T period, for the actual conductivity structure σ . Using the Green kernel method, we can rewrite basic equations of electromagnetism as (Weidelt, 1975)

$$\mathbf{E}(\mathbf{r}, \omega, \sigma) = \mathbf{E}_n(\mathbf{r}, \omega) - i\omega\mu \int_{\Omega} \sigma_a(\mathbf{r}') \mathbb{G}(\mathbf{r}, \mathbf{r}', \omega) \mathbf{E}(\mathbf{r}', \omega, \sigma(\mathbf{r}')) d^3\mathbf{r}', \quad (2)$$

$$\mathbf{H}(\mathbf{r}, \omega, \sigma) = \mathbf{H}_n(\mathbf{r}, \omega) + \int_{\Omega} \sigma_a(\mathbf{r}') \text{curl}(\mathbb{G}(\mathbf{r}, \mathbf{r}', \omega)) \mathbf{E}(\mathbf{r}', \omega, \sigma(\mathbf{r}')) d^3\mathbf{r}', \quad (3)$$

at any point $\mathbf{r}$ and any frequency ω . The quantity μ is the magnetic permeability of the vacuum in our models. $\mathbb{G}(\mathbf{r}, \mathbf{r}', \omega)$ is a 3×3 complex matrix, named Green kernel, and represents the Fourier transform at frequency ω of the electric field created at point $\mathbf{r}$ by a unit dipole $\delta(\mathbf{r}')$ located at point $\mathbf{r}'$ (Morse and Feshbach, 1953), $\text{curl}(\mathbb{G}(\mathbf{r}, \mathbf{r}', \omega))$ is a 3×3 complex matrix obtained by taking the curl at point $\mathbf{r}$ of the field of the column vectors of the matrix $\mathbb{G}(\mathbf{r}, \mathbf{r}', \omega)$.

A numerical solution of the forward problem has been first proposed by Vasseur and Weidelt (1977) for a superficial thin sheet. The Vasseur and Weidelt's (1977) approach can be extended to a thin sheet at any depth below the surface. The code was made operational by Tarits (1989). These solutions are based upon a digitization of Ω in K small square cells P_k , $k = 1, \dots, K$, in which conductivities, Green kernels, electric and magnetic fields are nearly constant. Let $S(P_k)$ and $S_n(P_k)$ be the actual and normal integrated conductivities (i.e. conductances) of the cell P_k , and $S_{ak} = S(P_k) - S_n(P_k)$, $k = 1, \dots, K$, the anomalous conductance of P_k . Then basic equations become, using the

same notations for the sake of simplicity,

$$\mathbf{E}(\mathbf{r}, \omega, S) = \mathbf{E}_n(\mathbf{r}, \omega) - i\omega\mu \sum_k S_{ak} \mathbb{G}(\mathbf{r}, P_k, \omega) \mathbf{E}(P_k, \omega, S(\mathbf{r}')) |P_k|, \quad (4)$$

$$\mathbf{H}(\mathbf{r}, \omega, S) = \mathbf{H}_n(\mathbf{r}, \omega) + \sum_k S_{ak} \text{curl}(\mathbb{G}(\mathbf{r}, P_k, \omega)) \mathbf{E}(P_k, \omega, S(\mathbf{r}')) |P_k| \quad (5)$$

at any point $\mathbf{r}$ and any frequency ω .

Let us consider the linear system made up of equation (4) written at each point P_k of Ω , $k = 1, \dots, K$. Given the distribution of anomalous conductances S_{ak} , $k = 1, \dots, K$, the quantities $\mathbf{E}(P_k, \omega, S(\mathbf{r}'))$ are solutions of this system. For each frequency, the system has $6 \times K$ equations and $6 \times K$ real unknowns in the general case. Solving this system allows us to compute the electric field $\mathbf{E}(\mathbf{r}, \omega, S(\mathbf{r}))$ at any point $\mathbf{r}$ and frequency ω . The magnetic field $\mathbf{H}(\mathbf{r}, \omega, S(\mathbf{r}))$ is then derived using (5) written at any point $\mathbf{r}$ and frequency ω .

5.4 The inverse problem

The structure of the embedding 1-D model, i.e., $S_n(\mathbf{r})$, the depth of the heterogeneous thin sheet, the domain Ω , and the cells P_k are first defined, and remain fixed during the inversion. Their values are to be deduced from previous investigations in the studied area (geology, other geophysical methods, ...) and from laboratory measurements of rock conductivity. The conductances S_k , $k = 1, \dots, K$, are the parameters, and solving the inverse problem consists in estimating their a posteriori *pdf* given the data $\mathbf{d} = (d_{ij})$, $i = 1, \dots, I$, $j = 1, \dots, J$, and the a priori *pdf* P_0 .

The data $\mathbf{d}$ are actually estimates of a function $\mathbf{T}[\mathbf{E}(\mathbf{r}, \omega, S(\mathbf{r})), \mathbf{H}(\mathbf{r}, \omega, S(\mathbf{r}))]$ at points $\mathbf{r}_i$, $i = 1, \dots, I$, of the surface and frequencies ω_j , $j = 1, \dots, J$, plus an experimental noise $\delta\mathbf{d}$,

$$d_{ij} = T[\mathbf{E}(\mathbf{r}_i, \omega_j, S(\mathbf{r})), \mathbf{H}(\mathbf{r}_i, \omega_j, S(\mathbf{r}))] + \delta d_{ij}. \quad (6)$$

The δd_{ij} , and therefore the d_{ij} are assumed to be independent gaussian random variables with zero mean value and standard deviation τ .

In situations for which no particular information is available, we choose as a priori distribution the product of a uniform *pdf* on each layer. In this case P_0 is a uniform distribution over all the possible models and where the different parameters are independent. P_0 is digitised over a set of conductance values, hereafter called possible conductance values, the choice of which depends on the a priori knowledge of the considered medium. There is no constraint on this choice, but it is clear that the larger this number the better the determination of the a posteriori distribution, but the greater the computer time. The possible conductance values may depend on the cell. We will only consider here situations with the same number L of possible values S_l , $l = 1, \dots, L$, for each cell. The a posteriori distribution will accordingly be expressed as the a posteriori probability of these possible conductance values.

The a posteriori *pdf* of the parameters S is estimated by means of a Markov chain. Let an algorithm that considers each cell P_k successively and updates the value of the parameter for this cell with a transition probability equal to the conditional probability distribution of S_l , $l = 1, \dots, L$, given the actual values of the parameters for the other cells, the data $\mathbf{d}$ and the a priori *pdf* $P_0(S)$. The sequence of images thus obtained after each scanning of the whole set of K cells is a homogeneous Markov chain, because the probability of an image depends only on the previous image, and does not change with step n . This process is a Markov chain on $\mathcal{M}$ called the Gibbs sampler (Robert, 1996). It is ergodic, and its invariant probability Π is the a posteriori probability is the conditional probability $\Pi(S)$ of the parameters given the data $\mathbf{d}$ and the a priori *pdf* $P_0(S)$ (Roussignol *et al.*, 1993; Roussignol and Menvielle, 1995; Robert, 1996; Grandis *et al.*, 1999; 2002).

In the present case, the required $L \times K$ solutions of the forward problem that are necessary for computing the transition probability of the Markov chain (in the present case, the K conditional *pdf*'s corresponding to the whole set of cells) requires a very long CPU time. In order to limit the CPU time, the forward problem is not fully solved at each scanning, and the electric field is estimated by the Markov chain. The Markov chain is then a two-dimensional one, which estimates a quantity (the conductance) distributed over a finite set of real values and another one (the electric field) distributed over a continuous set of real vectors.

The asymptotic behavior of such a Markov chain has first been studied empirically by Jouanne (1991), then Grandis (1994) for situations representatives of those encountered in geophysics. Using synthetic models, Grandis (1994) evidenced that the empirical average of the transition *pdf* of this Markov chain tends towards an invariant *pdf* that corresponds to the a posteriori *pdf* of the parameters, provided the estimate of the electric field at each scanning is precise enough for the transition probabilities to be reasonably estimated. Later on, Touijar (1994) gave the first demonstration that such a chain converges under given conditions. The physical interpretation of Touijar's mathematical results suggests that the empirical results established by Jouanne (1991) and Grandis (1994) are likely to be valid for most of geophysical situations (Menvielle and Roussignol, 1995).

6 The inversion

6.1 The description of the model and data

In accord with the above discussion, we chose the following particular parameters for the thin sheet model used in the MCMC inversion of the long-period geomagnetic induction data over the BM/BV/WCP region:

(i) The anomalous domain of the superficial thin sheet is divided into $21(\text{N}) \times 29(\text{E})$ square cells, each $25 \times 25 \text{ km}^2$ in size, covering thus a region of about $500(\text{N}) \times 750(\text{E}) \text{ km}^2$. This gives in total 609 variable parameters (cell conductances) for the inversion.

(ii) The normal model was composed of a two-layer background model with the first layer of $100 \text{ km}/1000 \Omega\text{m}$ and a uniform basement of $300 \Omega\text{m}$. The basement was used to simulate the asthenospheric layer. The topography of the asthenosphere was neglected. The layered model was overlain by a uniform infinite thin sheet with the conductance of 300 siemens.

(iii) For the inversion purposes, the 150 experimental geomagnetic TF's available were interpolated into the centers of the sheet cells. As reliable data variances were not available for the experimental TF's, we assumed the data to be contaminated with Gaussian errors with the standard deviation of 0.03.

(iv) For the MCMC run, data for only one period, $T = 3860 \text{ s}$, were used to keep the computation times in reasonable limits. Fit of the data and model arrows for further periods available were only checked by direct modeling tests.

6.2 The MCMC procedure, convergence, output models, post-processing

For the MCMC procedure (Gibbs sampler), the conductance of each of the 609 variable cells could take on one of 12 predefined values from the interval of 3 to 5000 siemens. The Markov chains were typically running for 1000 cycles of the Gibbs sampler, which took about 10 hours of the CPU time on a standard PC station (Intel Xeon 3 GHz, 1 MB RAM). Repeated MCMC runs were carried out to simulate parallel chains and check the convergence of the MCMC process. First 500 cycles of each of the MCMC chains were discarded as a burn-in period during which the chain approaches a vicinity of the solution and stabilizes. In our runs, the chains did not change substantially after the burn-in period, and simulated parallel chains converged to similar conductance distributions for the same cells. From this rough diagnostics, we can conclude that the chains converged and can provide reasonable estimates of the means of the cell conductances within the selected model class $\mathcal{M}$. As the number of MCMC iteration cycles is relatively small in our experiments, it is not realistic to obtain reasonably accurate estimates of higher moments of the aposteriori *pdf*'s for the cell conductances. In fact, only highly qualitative conclusions may be made as regards the uncertainties of the variables.

The MCMC procedure gives a series of models which should be probabilistically distributed in accord with the experimental data. We obtain a whole histogram of conductance values for each cell, which shows how much likely are the particular conductance values in the cell considered. After long enough iteration process, the histograms approximate the marginal aposteriori probabilities of the discrete conductance values in the individual sheet cells. We can estimate the mean values of the cell conductances by integrating over the histogram, as well as the most probable cell conductances (maximum aposteriori probability, or MAP estimates) by taking the most frequent value from the histogram.

Fig. 3 displays sheet models over the BM/BV/WCP region set together from the mean values of the conductance in the individual sheet cells and from the MAP estimates of the cell conductances. These models do not represent the average or MAP models from the chain, which would both require integration of the multi-dimensional empirical posterior *pdf* over the whole parameter space. The presented models are put together from estimates based on individual marginal *pdf*'s for the individual sheet cells. It is clear that these models may differ considerably from each other, especially in regions where the cell conductances are poorly constrained by the data. Nevertheless, the models in Fig. 3 represent simple integrated information on the conductance structure projected from a 3-D probabilistic (histogram) image of the cell conductances provided by the whole underlying Markov chain. Fit of the model data from the 'average' model in Fig. 3 (left) and the experimental induction arrows is shown in Fig. 4 for the inverted period of 3860 s. Model vs. experiment fit for longer periods (not shown) is very similar, larger discrepancies are observed for the period $T = 1200 \text{ s}$ with model induction arrows substantially smaller than the experimental data. This is most likely due to a misspecified depth of the sheet in our model (superficial sheet).

Qualitatively, we can also estimate the uncertainty of the conductance by simply observing the 'flatness' or 'peakiness' of the histograms for the cell conductances. In Fig. 5 (left), we show the 'average' model from Fig. 3 (left) modified by keeping in it only those cells for which 90% of the cell conductance values in the MCMC histograms fall within an interval of a width of $2/3$ of a decade. All remaining cells, with flatter conductance histograms, have been discarded. The cells in Fig. 5 (left) thus indicate regions of the model in which the conductance is reasonably well constrained by the data. More precise, quantitative uncertainty estimation would be perhaps possible from longer MCMC

chains which would allow us to integrate directly for variance-covariance functions estimates for the parameters.

Because of high computational demands, thin sheet models for the MCMC inversion are mostly designed with only a coarse tiling. In the model post-processing stage, spatial smoothing often results in a more realistic image of the conductance distribution. The smoothed image has to be checked by an additional direct modeling test. We show a smoothed version of the 'average' conductance model from Fig. 3 (left) in the right-hand side panel of Fig. 5.

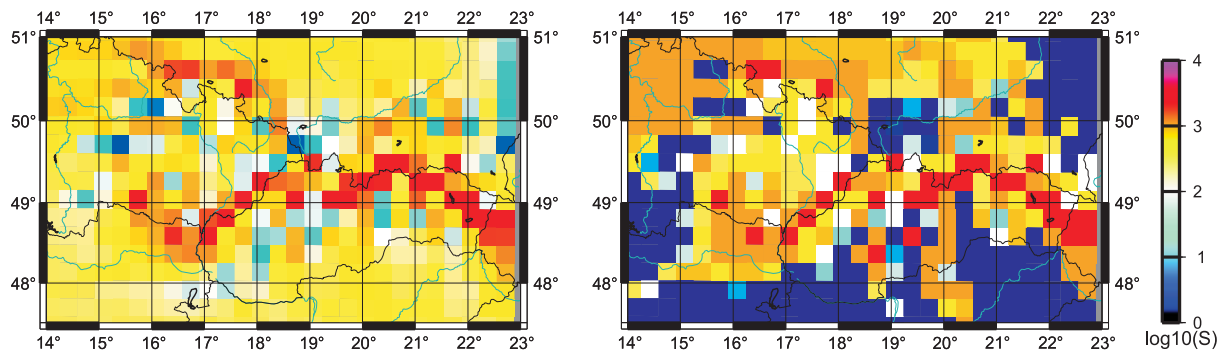


Figure 3: Left: Conductance model over the BM/BV/WCP region designed from average cell conductances obtained from the stabilized part of the Markov chain. Right: Conductance model designed from maximum a posteriori conductances in the individual sheet cells. The color scale corresponds to the logarithm of the conductances. Period of inverted data was 3860 s.

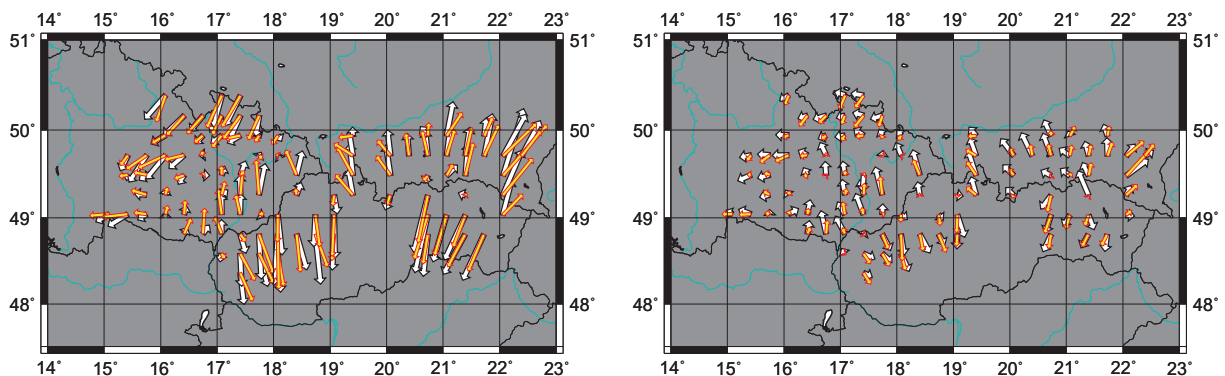


Figure 4: Fit of the real (left) and imaginary (right) induction arrows for the period of 3860 s. White arrows show the input data obtained by interpolating the experimental TF's into the centers of the sheet cells. Orange-yellow arrows were generated by the thin sheet model designed from average cell conductivities from the MCMC chain (model to the left in Fig. 3).

7 Tectonic correlation

Interpretation of narrow-band long-period geomagnetic data by the MCMC sampling can only provide a large-scale image of the conductance distribution which conforms with the observed TF's. The conductance pattern restored from the induction arrows collected above the eastern slopes of the BM and its transition to the WCP gives a geologically plausible image of the region. The Carpathian conductivity anomaly is reconstructed in detail, suggesting possible weakening/local interruption at the crossing with the Central Slovakia fault zone (near 19.2°E in Fig. 5 right) where other geophysical fields and geological indicators show discontinuous features on a presumably strike-slip fault (e.g., Kováčik and Hložek, 1993). Though in coarse mesh, the high conductance anomaly well fits the position of a low resistivity layer from the magnetotelluric data collected in the easternmost sector of the model, close 22°E (Vozar, 2005).

The induction anomaly at the eastern margin of the Bohemian Massif seems to have a more complex explanation than that of a quasi-linear conductor along the SW-NE zone of a rapid change of direction of the induction arrows (see Fig. 2). Two features may be observed in the model in Fig. 5 (right), which were already discussed earlier by Kováčik *et al.* (2005). First, it is a conductive zone in the NE of the BM, which may generate induction arrows directed towards the SW, as clearly observed all across the area to the west of the EBMA. Second, it is an alteration of conductive and resistive E-W zones in the eastern part of the BM. Though still not clear how much of this effect

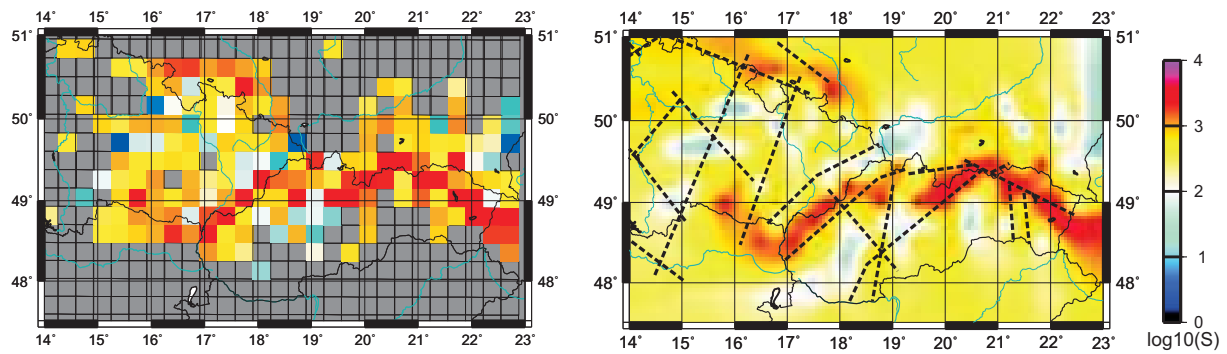


Figure 5: Left: Model from Fig 3 (left) with only those cells of the original ‘average’ model shown that have 90% of all conductance values within 2/3 of a decade. Cells with flat histograms are eliminated. Right: Spatially smoothed ‘average’ conductance model. The dashed lines indicate main regional fault zones.

may be caused by the profile arrangement of the data, this quasi-anisotropic domain is actually required by the data. It helps in feeding the induction process all across the Moravia region and in keeping the relatively large modules of the induction arrows in that region. If these induction sources could be verified, the regional induction in the eastern part of the BM would be mainly driven by the NW-SE to W-E pattern of tectonic zones (Elbe fault zone, Sudety faults, Pořč-Hronov fault zone, Odra fault zone, etc.) rather than by the SW-NE zones which conform with the tectonics of the transition to the WCP.

8 Conclusion

From the experience accumulated with the MCMC inversion we can conclude that for moderately sized thin sheet models (tens of cells in each direction), the MCMC inverse procedure is practicable with standard computer facilities. Speeding-up the forward solutions by using a 2-D Markov chain derived from the integral equation numerical approach by Vasseur and Weidelt (1977) is essential for the algorithm.

Though computationally still demanding, the advantage of the MCMC is that it provides a probabilistic output for the inverse solution. Parameters can be assessed with respect to their values and uncertainty ranges, though true uncertainties (variance-covariance matrices) are hard to obtain in problems with demanding direct solutions. Effective visualization of complete probabilistic outputs from the MCMC algorithm may still be one of its weak points.

The conductance pattern restored by the MCMC sampling from the induction arrows collected above the eastern slopes of the BM and its transition to the WCP gives a geologically plausible image of the region. The Carpathian conductivity anomaly is reconstructed in detail. The induction anomaly suggested at the eastern margin of the BM admits an alternative explanation in terms of a NW-SE to W-E conductance patterns, hypothetically coinciding with the fault pattern of the eastern Bohemian Massif. A detailed verification of the structural hypotheses in this region is, however, hardly possible by employing the long period geomagnetic induction data alone.

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Subsurface Conductivity Obtained from DC Railway Signal Propagation with a Dipole Model

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Abstract

In the presented study, an attempt is made to model the propagation behavior of signals emitted by a Polish DC railway line with a number of grounded horizontal electric dipoles on the surface of a homogeneous half-space. The signals were measured on a magnetotelluric profile perpendicular to the railway line mainly in its near-field and transition zone, and they were separated from the natural electromagnetic variations by means of a reference site. Fitting the DC signals to the dipole model yields the conductivity of the homogeneous half-space. Its value for the vertical and for the perpendicular horizontal magnetic component is confirmed by the usual 2D MT model obtained from the profile.

Introduction

The electrified part of the Polish railway network (fig. 1) is run by DC current. This makes magnetotelluric (MT) studies in this country difficult. On the other hand, the “disturbing” (in terms of MT) signals emitted by this system propagate according to a dipole model (Oettinger et al. 2001 and citations therein) which contains, like MT, the electrical conductivity of the subsurface as a parameter. So, after designing and performing a MT field experiment comprehending railway signals in an appropriate way, it should be possible to obtain information about the subsurface conductivity by both the dipole model and the MT approach. By this idea the present study has been motivated.

The methodology of this study lies somewhere in-between two other domains encountered in geophysics. One is certainly the electromagnetic methods of the controlled-source near-field branch. The difference to our case is that the source is more under control there, but in return not free of experimental effort. The second related domain is techniques for modeling the propagation behavior of disturbances from DC railways and cables, applied especially to investigate their influence on magnetic measurements run at observatories (e.g. Pirjola et al. 2007, Lowes 2009, Maule et al. 2009). However, the induction character of this propagation including phenomena like frequency-dependent damping, phase shift, and the electric conductivity as a propagation parameter is almost completely omitted there in contrast to the approach presented here.

In the following, there will be considered the data processing including the separation of the natural electromagnetic variations from the railway signals and the estimation of the transfer functions between different stations. The dipole model will be described and the data be inverted according to it. The conductivity result will be compared to the output of a usual 2D MT model.

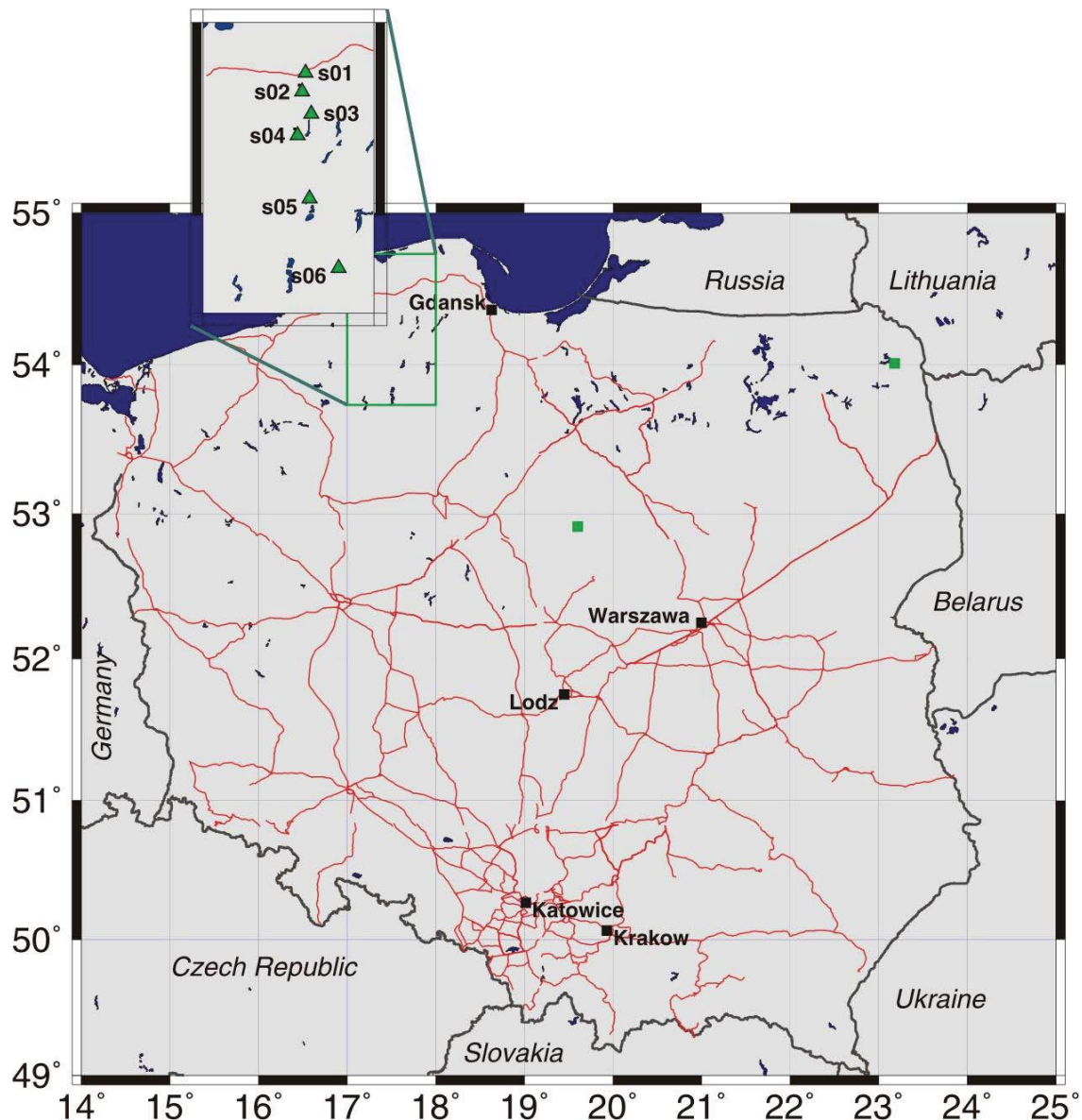


Fig. 1 The electrified part of the Polish railway network (red lines) with a profile of MT sites for investigation of one line (zoomed-in part). Green squares denote positions of reference stations.

Measurement and data processing

As shown in fig. 1, a profile of six MT sites has been installed perpendicular to an isolated railway line. The distances of sites to the line were 0.8, 7, 16, 25, 50, and 75 km. Two remote reference sites were set up at relatively noise-free places, and the whole array was running synchronously. The obtained data were processed with codes by Neska 2006, but not only for MT purposes. By means of its correlation to the reference sites, the natural part of the electromagnetic

signal could be removed from the measured data using a method by Larsen et al. 1996 (fig. 2). In this way a dataset for analysis of the railway signals was established (cf. fig. 3). It has to be emphasized that it can be problematic to work with such “deduced” data, since distortions introduced to it during the separation, e.g. due to uncorrelated noise in the reference site, can bias the subsequent results. It proved useful to separate the data of the station closest to the railway which plays a special role in the following with one reference site and the rest with the second one.

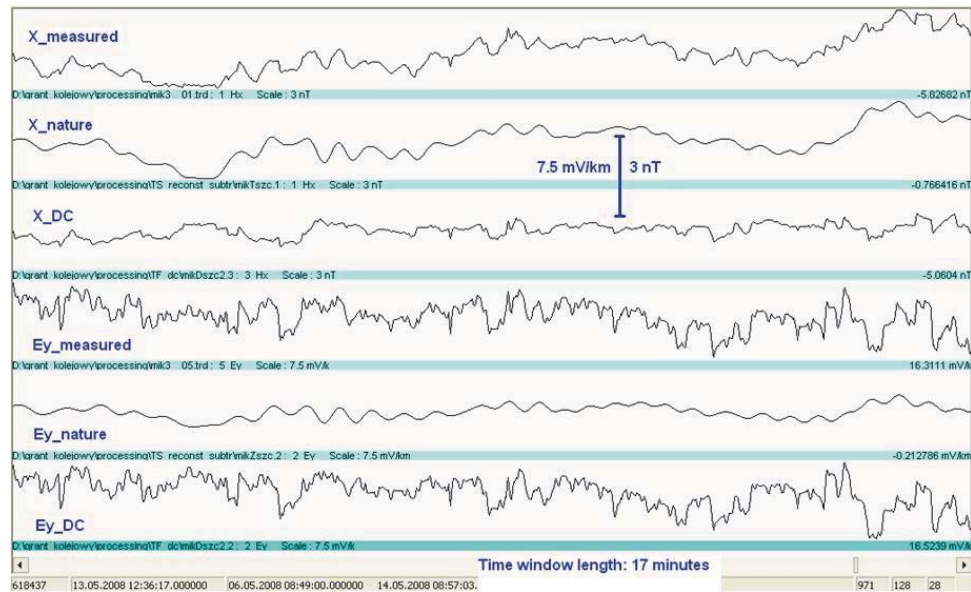


Fig. 2 Separation of time series into a “natural” and a “railway” part (site s02, see fig. 1).

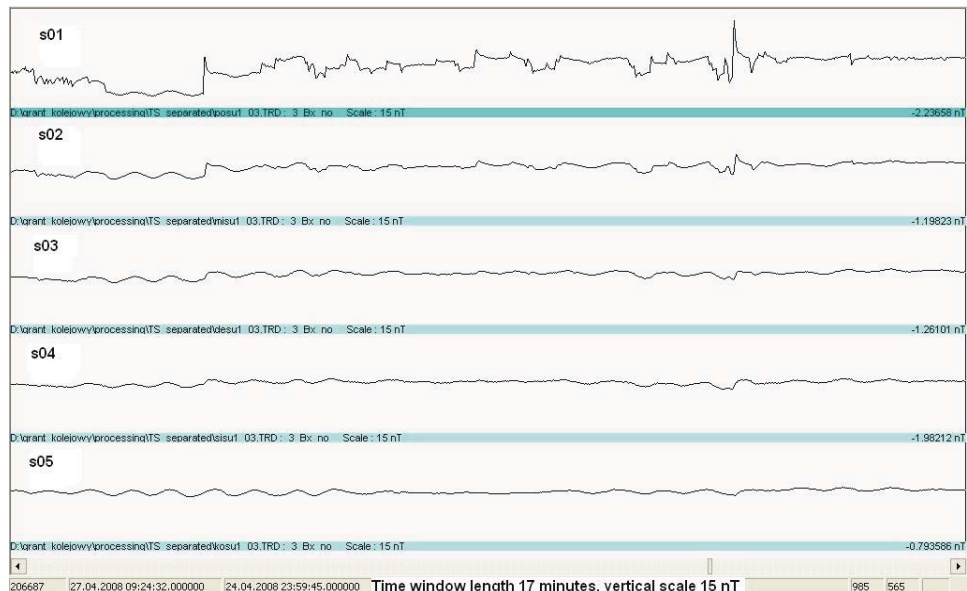


Fig. 3 The propagation/decay of railway signals (component Bx) over the profile. Note that a rest of natural variations remained in the time series in spite of separation.

All following considerations refer to “railway data” separated from natural signals in the way indicated above.

The formulas for the dipole response are given in frequency domain, so our calculation has to take place there. Furthermore, in these formulas there occur factors like current I and dipole length L (see table 1), i.e. technical parameters of railway traction which we cannot measure immediately. So it appeared reasonable to remove these factors by normalizing each field component to that of the railway-nearest site which represents the strongest, most pronounced, or least attenuated railway signal, respectively.

To perform this normalization in a convenient way, the algorithm for calculation of the inter-station transfer function or Horizontal Magnetic Tensor (HMT) was used, which played an essential role during the separation already. The site nearest to the railway takes the function of the reference, i.e. its data are input channels, and the data of the remaining sites are the output channels. So B_x and B_y data of “local” and “reference” sites lead to normalized values for both horizontal magnetic components, E_x and E_y to those for the electric components (even if the analogy to the HMT is overstretched here since it refers to magnetic data only, the formalism is the same), and B_z and some other, uncorrelated channel (e.g. of one of the real, off-profile references, just to fit the requirement of the HMT formalism that there must be two input channels) to the normalized value for the vertical magnetic component. This description remains a bit vague because it cannot be recommended for imitation for the following reason:

For this study, it has been noticed too late that this approach is not quite correct. The problem is that the HMT formalism bases on bivariate statistics (i.e. there are two independent variables or input channels, respectively), whereas the railway provides only one independent source polarization (pers. comm. K. Nowożyński). A similar problem occurs in Controlled-Source MT, where only one scalar transfer function instead of the usual 2×2 impedance tensor can be obtained if only one transmitter is used (pers. comm. M. Becken). So it has to be expected that our normalized values are influenced in an unfavorable way. This impact is expected to be very small in the case of B_z , since the warranted uncorrelatedness of the second input channel should make the difference between the univariate and the bivariate result vanish. This will be confirmed in the following section (cf. fig. 4: better result for B_z). However, the direction of the railway line under investigation is almost East-West (fig. 1), so its signal has a polarization almost coinciding with one of the directions we measure and calculate in with the MT or HMT approach. So there is hope that the error introduced due to this wrong statistic approach is somehow bearable if the diagonal elements of the “HMT” are used.

Finally, the normalized values are displayed over period and distance in 3D-plots (red lines in fig. 4). The decay of amplitudes with distance as well as their period-dependency (stronger damping at low periods) becomes clearly visible for B_x and B_z components. B_x has partly unsystematic and, especially at long periods, rather high values that make the impression to be distortions, maybe due to incomplete separation or the wrong statistic approach mentioned.

Hence the possibility to include data errors to the business was provided, since they have an important meaning as weights during the inversion.

The dipole model and its application

The railway line was simulated by a chain of horizontal grounded electric dipoles following its (not constant) bearing within a distance of some dozens of km around the profile. The distance between single dipole centers was 250 m. The solutions for the electromagnetic field components for the quasi-static approximation are taken from Zonge & Hughes 1987 and hold for one dipole on the surface of a homogeneous half-space (see table 1). These values are transformed from cylindrical to Cartesian coordinates to match the measuring system of MT and summarized over all dipoles subsequently. Then the values for each component and station were divided by the value of the corresponding component of the station closest to the railway. Since we look only for one model parameter (i.e. the conductivity σ), the inversion consists just of testing the whole model space and selecting the value with minimum RMS. The relevant software related to the dipole model was created by K. Nowożyński.

The components behaving best in our study were Bx and Bz. The inversion of Bx alone gave a resistivity of 5 Ωm , that of Bz alone 14 Ωm , and the joint inversion 10 Ωm . The model response for the latter is shown in fig. 4 as blue lines.

Table 1

Solution for horizontal grounded electric dipole on HH surface according to Zonge & Hughes, 1987	
$E_r = \frac{IL\cos\phi}{2\pi\sigma r^3} [1 + e^{-ikr}(1 + ikr)]$	
$E_\phi = \frac{IL\sin\phi}{2\pi\sigma r^3} [2 - e^{-ikr}(1 + ikr)]$	
$E_z = \frac{i\mu_0\omega IL\cos\phi}{2\pi r} \left[I_1\left(\frac{ikr}{2}\right) K_1\left(\frac{ikr}{2}\right) \right]$	
$H_r = \frac{IL\sin\phi}{2\pi r^2} \left\{ 3I_1\left(\frac{ikr}{2}\right) K_1\left(\frac{ikr}{2}\right) + \frac{ikr}{2} \left[I_1\left(\frac{ikr}{2}\right) K_0\left(\frac{ikr}{2}\right) - I_0\left(\frac{ikr}{2}\right) K_1\left(\frac{ikr}{2}\right) \right] \right\}$	
$H_\phi = -\frac{IL\cos\phi}{2\pi r^2} \left[I_1\left(\frac{ikr}{2}\right) K_1\left(\frac{ikr}{2}\right) \right]$	
$H_z = -\frac{3IL\sin\phi}{2\pi k^2 r^4} \left[1 - e^{-ikr}(1 + ikr - \frac{1}{3}k^2 r^2) \right]$	
Where $I/K_0/1$ - modified Bessel functions and k - wave number $= (1 - i)\sqrt{\frac{\omega\mu_0\sigma}{2}}$	

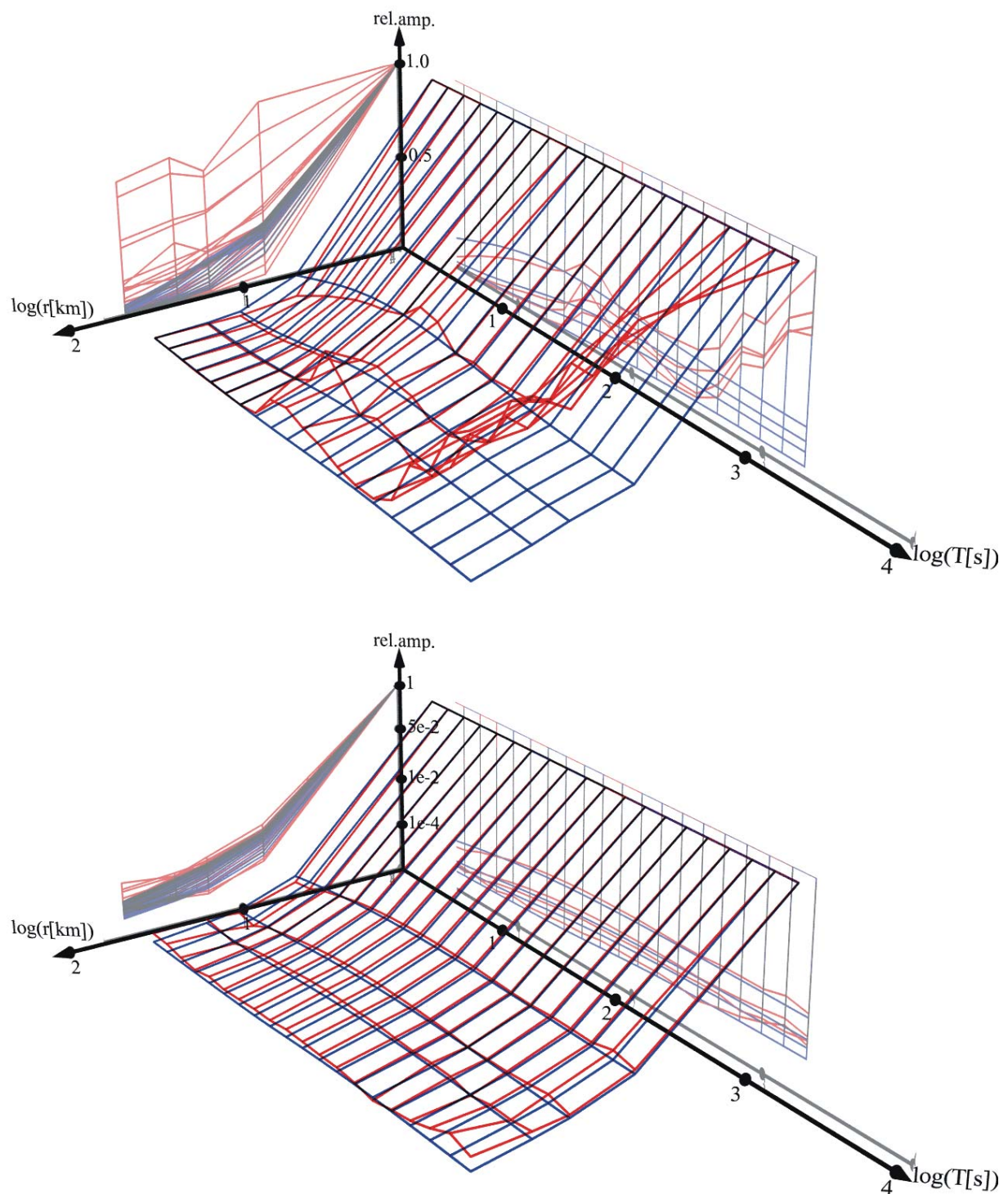


Fig. 4 Railway signal data (red) and dipole model response for a 10 Ωm homogeneous half-space (blue) over period and distance. Upper part for Bx, lower part for Bz component amplitude. The values on the vertical axis are normalized to that of site s01 nearest to the railway line (cf. fig. 1).

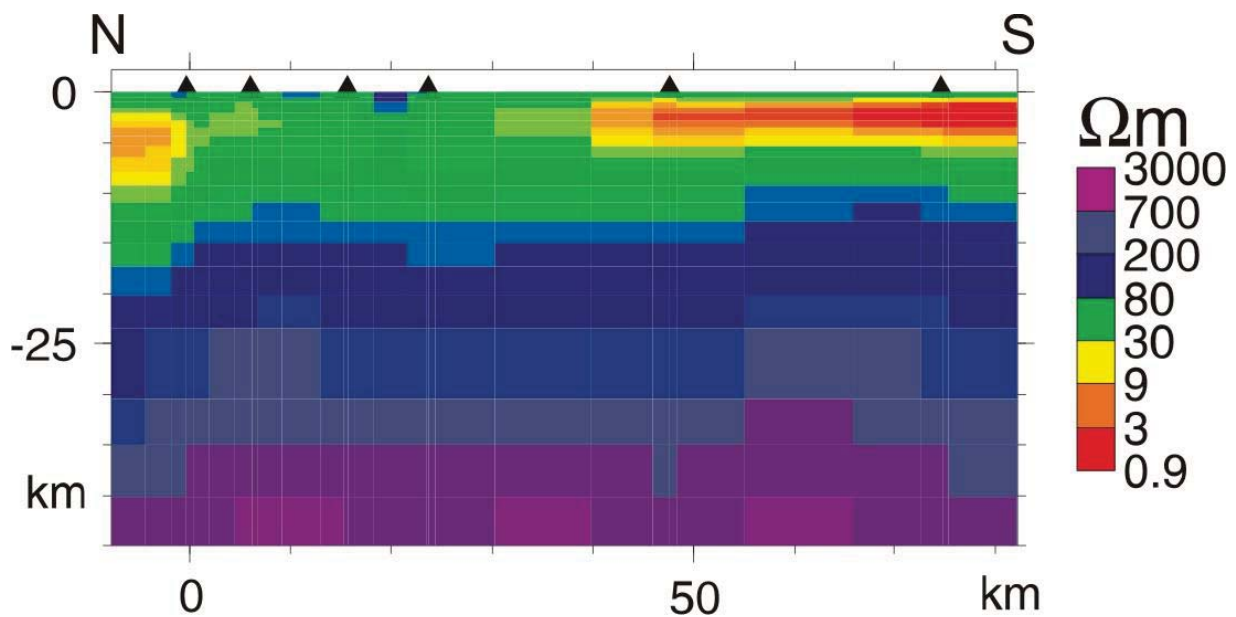


Fig. 5 2D MT model (TE and TM mode) of the profile on fig. 1 obtained with the REBOCC code (Siripunvaraporn & Egbert 2000). The shallow cells beneath the northernmost sites have resistivity values between 9 and 80 Ωm .

Discussion and conclusions

The MT model yields resistivities of 9-80 Ωm beneath the northernmost sites close to the railway (fig. 5). This is consistent with the dipole model result, so there is evidence that the dipole approach is not unreasonable.

Nevertheless, there are unexplained features in this model and one has to be critical with it. Not shown here is the behavior of phases and of electric components, where a similarity between data and model response is much harder to find for reasons still unclear.

However, taking into account first, the similarity of data and model response (fig. 4), and second, the compatibility of MT and dipole modeling results in spite of some quite principal problems in this implementation (i.e. input data biased during separation, errors due to application of bivariate instead of univariate statistics, limitation to the oversimplified case of a homogeneous half-space), there can be stated that the dipole model is not only a valid approach to model the propagation of DC railway signals, but even a relatively robust one.

Acknowledgements

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Magnetotelluric investigation of the Sorgenfrei-Tornquist Zone and the NE German Basin

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Abstract

The Sorgenfrei-Tornquist Zone (STZ) is the northwestern branch of the Trans-European Suture Zone (TESZ). The TESZ runs along more than 2000 km through the North Sea, South Scandinavia, the Baltic Sea, and Poland into the Black Sea. It divides old Precambrian lithosphere in the northeast from younger, Caledonian and Variscan one in the southwest. Data, dimensionality analysis and 2D models of a long-period magnetotelluric (LMT) profile crossing this prominent tectonic border in South Scandinavia are presented and a possible interpretation of conductivity anomalies below Rügen Island and south of Strelasund Basin is given. Furthermore these models map the saline aquifer of the Northeast German Basin. The "North German Conductivity Anomaly" is perhaps mainly due to the effect of the basin edges.

Introduction

The EMTESZ project (Electromagnetic Study of the Trans-European Suture Zone) was a multinational research project to study the electrical conductivity of the Trans-European Suture Zone (TESZ), which is one of the largest tectonic boundaries in Europe. It separates the East European Platform from the Paleozoic mobile belt of central and western Europe and is traced from the Black Sea through Poland and Southern Scandinavia into the North Sea (GEE (1996); PHARAOH (1999)). In the northeast the margin is marked by the lineaments of the Sorgenfrei-Tornquist Zone (STZ) and the Teisseyere-Tornquist Zone (TTZ, which runs in a SE-NW direction through Poland and the Baltic Sea, fig. 1). Several seismic refraction and reflection experiments have been carried out (BABEL and BASIN 9601 in Northeast Germany, POLONAISE and the LT surveys in Poland). Major results include a sedimentary thickness in the Northeast German Basin and below the TTZ of 4 to more than 11 km. Also sharp lateral boundaries, a Moho from 32-35 km in the SW to 40-45 km in the NE and a reflector below the STZ/TTZ at depth of 50-55 km are known. Most of the earlier EMTESZ measurements were carried out in 2003 to 2005 (e.g., BRASSE ET AL. (2006), ERNST ET AL. (2008)). Profiles MVB and MVS were conducted in 2006 and 2009.

The long period magnetotelluric profile shown here is one of these EMTESZ profiles. The direction of this profile was chosen because of earlier measurements by the Bundesanstalt für Geowissenschaften und Rohstoffe (BGR) in the 1990ies. Due to tipper data along this profile and of site MAT deployed in 2005 on Rügen Island (MT working group Free University of Berlin) a direction

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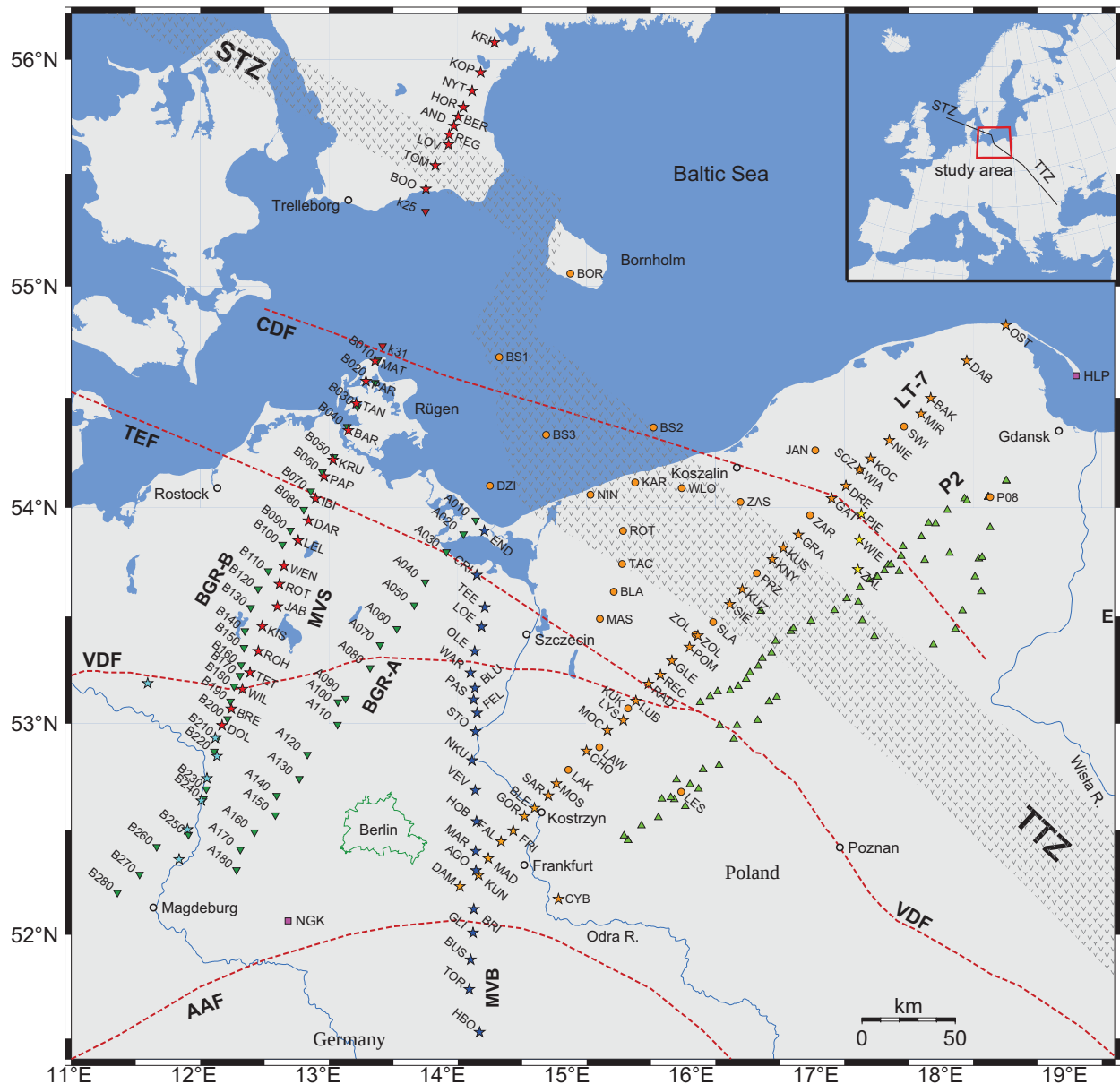


Figure 1: Measured profiles of the EMTESZ project and profile B from the Bundesanstalt für Geowissenschaften und Rohstoffe (BGR-B). Location of MT sites in Poland, Germany, Sweden and Bornholm (Denmark); STZ-Sorgenfrei-Tornquist-Zone; TTZ-Tornquist-Teisseyre-Zone; TEF-Trans European Fault; CDF-Caledonian Deformation Front; VDF-Variscan Deformation Front; AAF-America Avalonia Fault.

perpendicular to the assumed strike-direction of the Sorgenfrei-Tornquist Zone (see fig. 1) was chosen. It runs over about 370 km with a site spacing from 6 to 12 km.

Data examples

Stable transfer functions result from remote reference analysis according to EGBERT & BOOKER (1986). The remote data is from the observatories of Belsk and Niemegk and simultaneous measuring sites. In figure 2 three representative data examples for sites of profile MVS are shown.

Site ROT, which is located in the southern part of the profile, shows the characteristic curves for the North German Basin. Apparent resistivities at short periods are very low until 100 s (1 to 3 Ωm), pointing to a very good conductor near the surface. With increasing periods app. resistivity is growing, accompanied by a splitting of impedance components Z_{xy} and Z_{yx} , reflecting

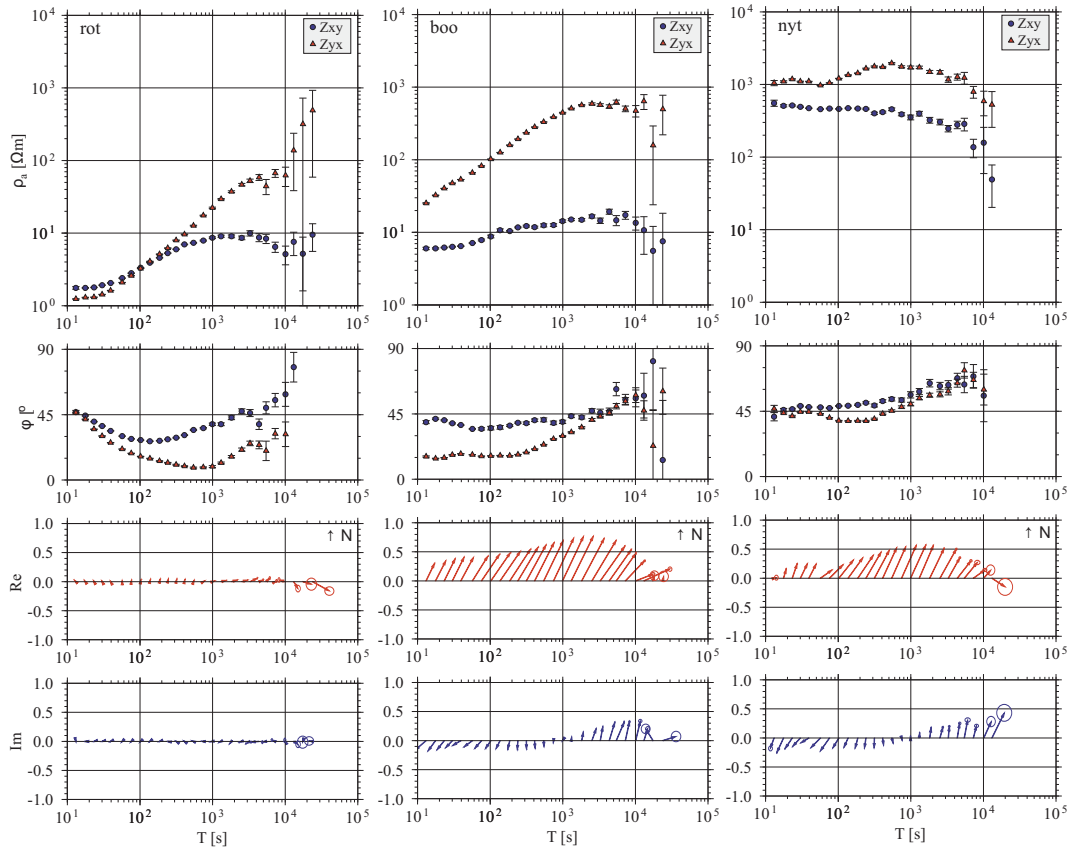


Figure 2: Examples of transfer functions for sites of the profile (ρ_a -apparent resistivity, ϕ -Phase and Re-Real part, Im- imaginary part of induction arrows as function of period). Site ROT as an example for the southern part of the profile shows low resistivity in short periods, which shows the thick sedimentary covers of North German Basin (ρ_a at low periods around 1-3 Ωm), for longer periods it shows higher resistivity. BOO is the site at the most southern part of the swedish part and shows the margin of the sedimentary basin. NYT is an example for the northern part of the profile in Southscandinavia. There the profile reaches the crystalline basement (ρ_a has higher values even for low periods).

multidimensional structures at greater depths. Induction arrows are very small. For sites in the southern part of the profile a change of direction for induction arrows is observed. Site BOO is located at the edge of the East European Craton. Here real parts of induction arrows are very large (up to an absolute value greater than 0.8) and are perpendicular to the strike direction of the Sorgenfrei-Tornquist Zone. They point to a very good conductor in southeastern direction. Apparent resistivities for this site are even for short periods much higher than in the North German part of the profile and increase rapidly for large periods, which points to the resistive basement of the East European Craton. NYT is an example for the northernmost part of the profile and shows high apparent resistivity even for the lowest periods (about 1 000 Ωm), reflecting the crystalline basement of Baltica.

Magnetic transfer functions

Induction arrows at long periods increase from south to the north, hinting at well-conductive structures in the south, i.e., below the Baltic Sea. The apparent resistivities ρ_a in the north of the profile are relatively high and get lower at the southern sites, which agrees with the geological structure. The high resistivities in the north are caused by the crystalline basement and the low resistivities in the south by the sedimentary cover.

Figure 3 shows a map of the real part of induction arrows at a period 1820 s for the EMTESZ profiles LT-7, MVB and the discussed profile MVS, according to Wiese-Convention (WIESE, 1962).

For profile MVS arrows are mainly perpendicular to the NW-SE striking direction. With values greater than 0.8, the unusually large induction arrows in the northern part of the profile indicate a good conductor below Rügen Island. Real induction arrows at the southernmost station of Sweden, the sites in the Baltic Sea and at the northernmost point of Rügen Island are largest.

The "flip-around" of induction arrows in Northern Germany and Poland was already recognized in the initial days of electromagnetic deep sounding. The underlying cause – a large conductivity anomaly at depth – was termed the North German-Polish Conductivity Anomaly (e.g., SCHMUCKER (1959); UNTIEDT (1970); JANKOWSKI (1967)). Until today, however, the depth extent of this anomaly is controversially discussed and models range from an upper mantle high conductivity zone to the simple effect of the basin edges. The magnitude of the inductive effect at the basin margins and their rapid decrease seem to favor the second explanation, which will also become evident from 2-D inversion (see later).

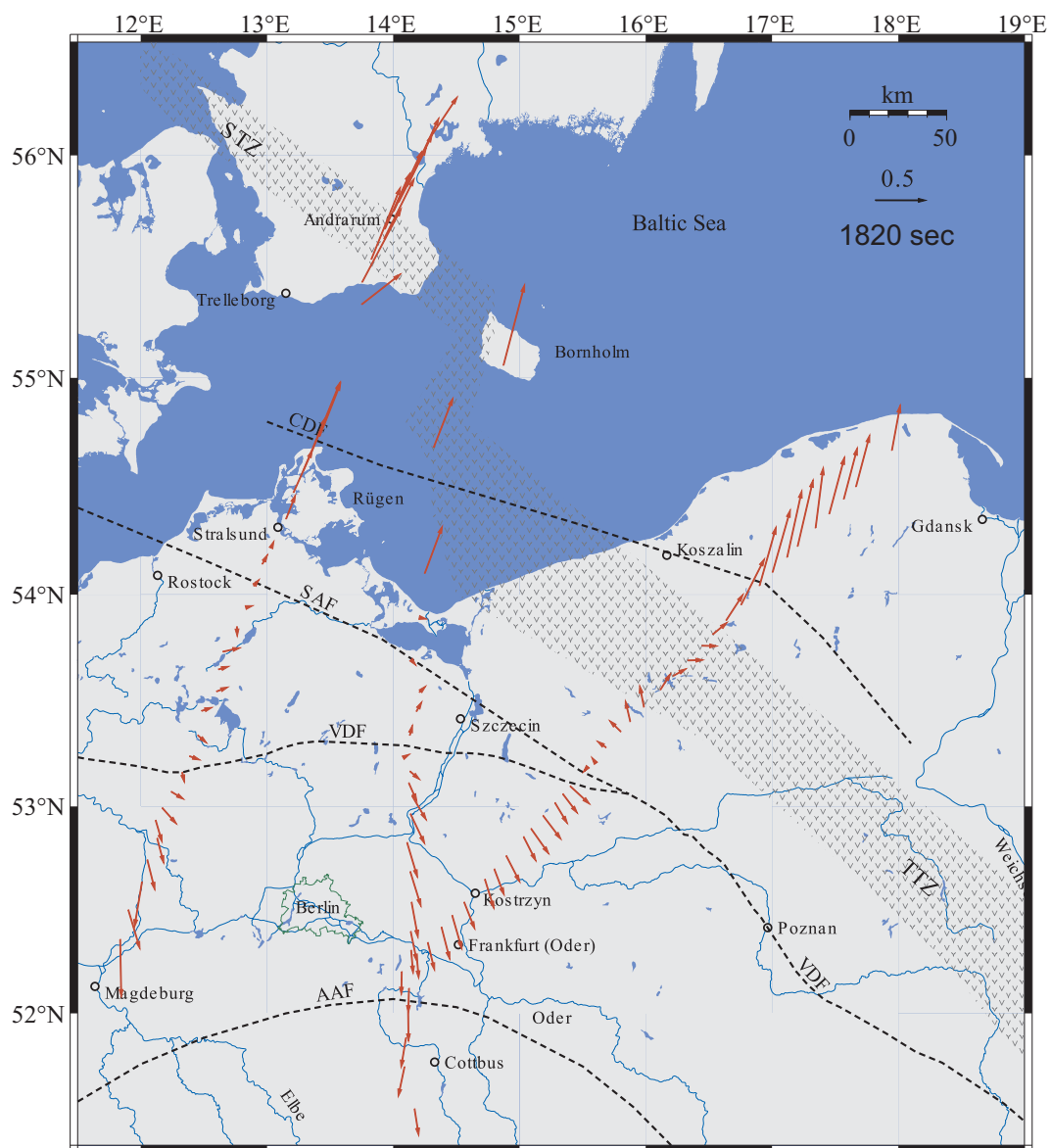


Figure 3: Induction arrows for profiles MVS, MVB and LT-7 (real part) at a period of 1820 s. In South Sweden and north of Rügen Island induction arrows are very large. They obviously result from the edge of the Baltic Shield (STZ). The change of direction of the real parts in the center of the NE German and Polish Basins shows the effect of the so-called "North German-Polish Conductivity Anomaly". It can be seen in all profiles. Profile MVS shows this effect (sign reversal of real part) approximately at the location of the Stralsund-Anklam-Fault (SAF). In the northern part of profile MVS the induction arrows are very large. The highest absolute value with more than 0.8 is reached in the southern part of Sweden and 0.7 in the northernmost part of Rügen Island.

Dimensionality analysis

Strike angles were calculated for single and multi sites after SMITH (1995), which allows to sort out bad data with a weighting matrix (see figure 4). The strike angle for profile MVS has a strong variation especially in the northern part; the average for the profile is around N67°W.

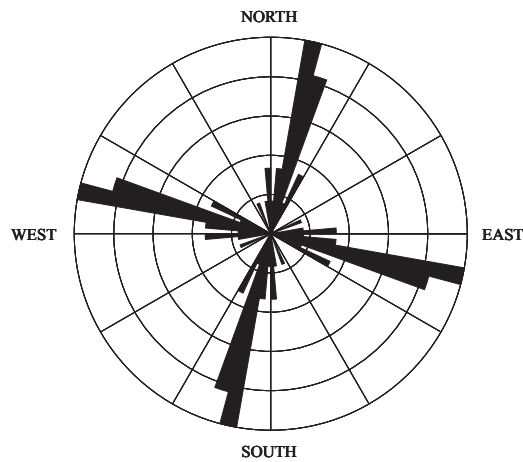


Figure 4: Average strike angle after SMITH (1995) for profile MVS: N67°W.

Figure 5 shows the phase tensor (CALDWELL ET AL., 2004) for each site of the profile for two different periods. The orientation of the ellipses point to the strike direction of the conductive structures. The northwest-pointing direction of the main axis of the ellipse corresponds with the calculated strike direction.

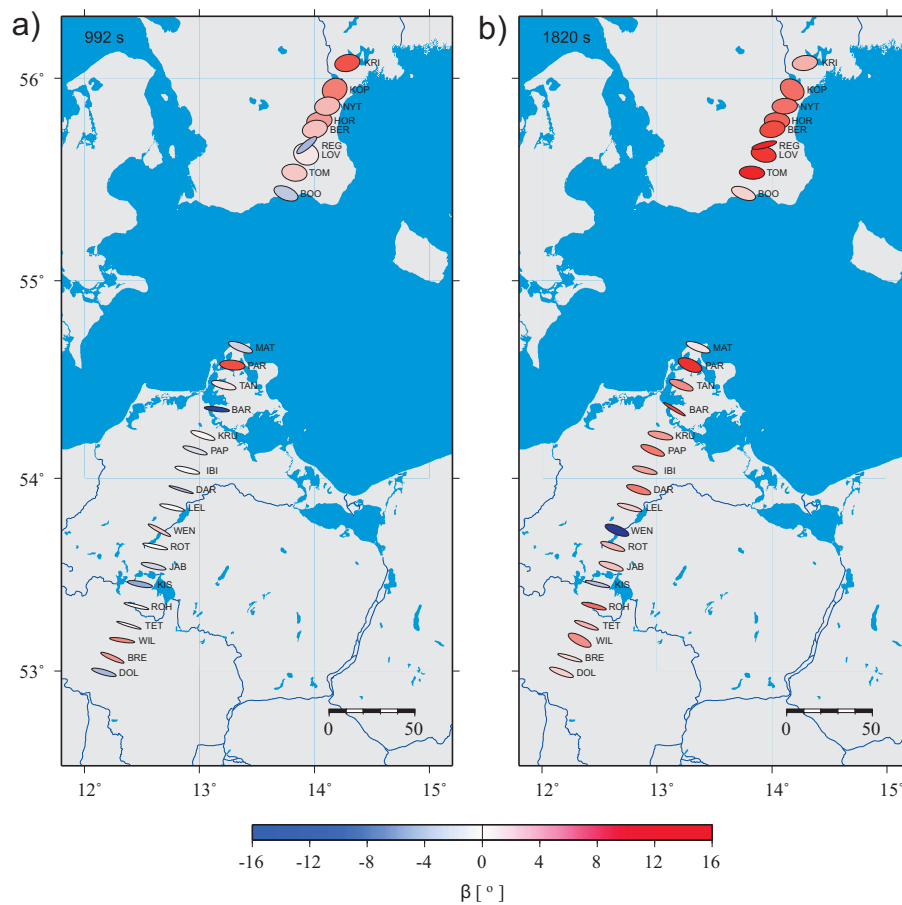


Figure 5: Phase tensor plot for periods of 992 s and 1820 s calculated after CALDWELL ET AL. (2004).

Skew angle β is small throughout the study area with the exception of the northernmost sites in Sweden. Summarizing, it suffices to carry out a 2-D modeling of the data.

2-D modeling

All presented 2-D inversion models were calculated with the CG-FD inversion algorithm of RODI & MACKIE (2001). First the data were rotated by -67° . A homogeneous halfspace of $100 \Omega\text{m}$ and a fine grid for the start model was chosen. The smoothing factor was set to $\tau=15$, which is the best trade-off between model roughness and data misfit (HANSEN & O'LEARY, 1993). For the inversion of the 2D-models in fig. 6 all components (TE-, TM-mode and tipper) were used. In these models all resolved structures can be seen; these models have the best fitting model response to the data (see examples in fig. 7). For a better overview of the modeled features, the profile intersecting geological structures in the respective locations are marked. Furthermore, the location of the basement is shown. Figure 6a shows the result of inversion of TE-, TM-mode and tipper data after 200 iterations. The resulting RMS is 2.27. The model shown in figure 6b also results from inversion of all components (TE-, TM-mode and tipper), furthermore a horizontal weighting function was used ($\beta=1$). The resulting RMS of this inversion after 200 iterations is with 2.67 worse then the RMS of the model without weighting function, but the model is more consistent with geological assumptions (HOFFMANN & FRANKE, 2008).

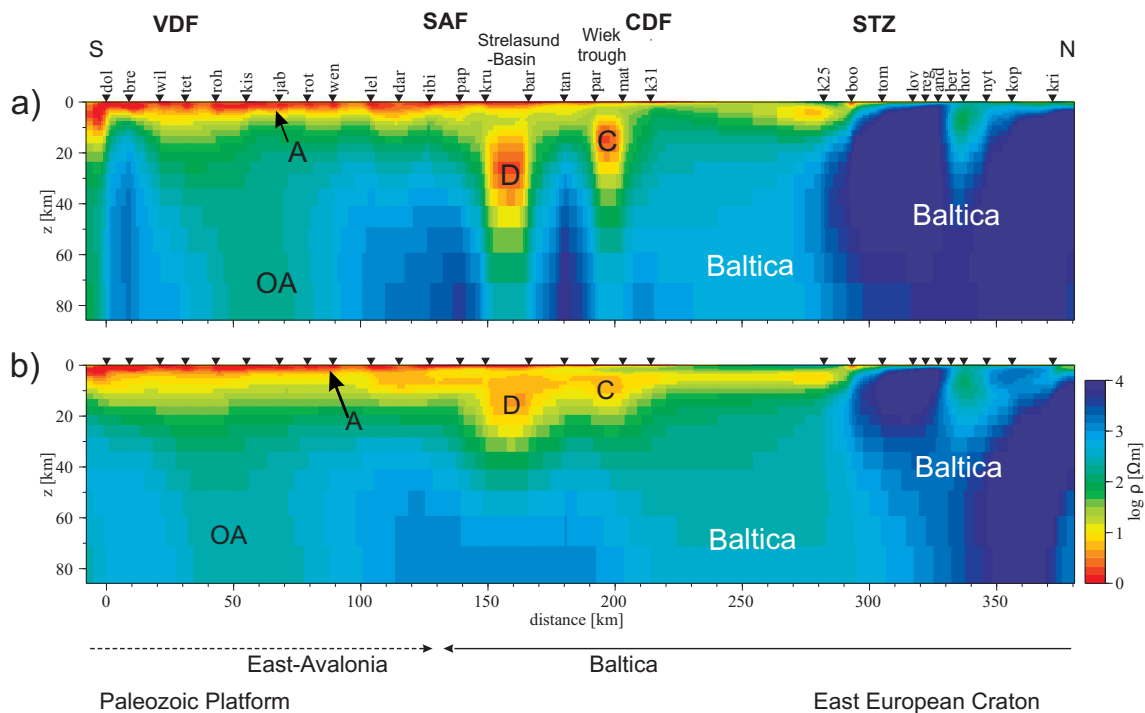


Figure 6: 2-D models for profile MVS inverted by applying Rodi and Mackie's algorithm starting from a homogeneous halfspace. a) For this model all components were inverted (TE-, TM-mode and tipper data). The resulting RMS after 200 iterations was 2.27. b) For inversion of this model all components were used (TE-, TM-mode and Tipper). A horizontal weighting function was used. Resulting RMS after 200 iterations was 2.67. For both models the smoothing factor was set to $\tau=15$. Letters sign resolved conductivity structures: A-saline aquifer in Northeast Germany, Baltica-basement of the Baltic Continent, OA-basement, associated with the basement of the East-Avalonian plate. Also the assumed location of plates and fault zones is marked.

In both models one can see the underlying plate of Baltica in the north as a poor conductor, the sediments in Northeast Germany as a very good conductor (structure A) and two conductivity anomalies below Rügen Island (structure C) and south of Stralsund (structure D) in depths from 8 to 30 km. Figure 7 shows examples of model response of the inversion shown in figure 6a.

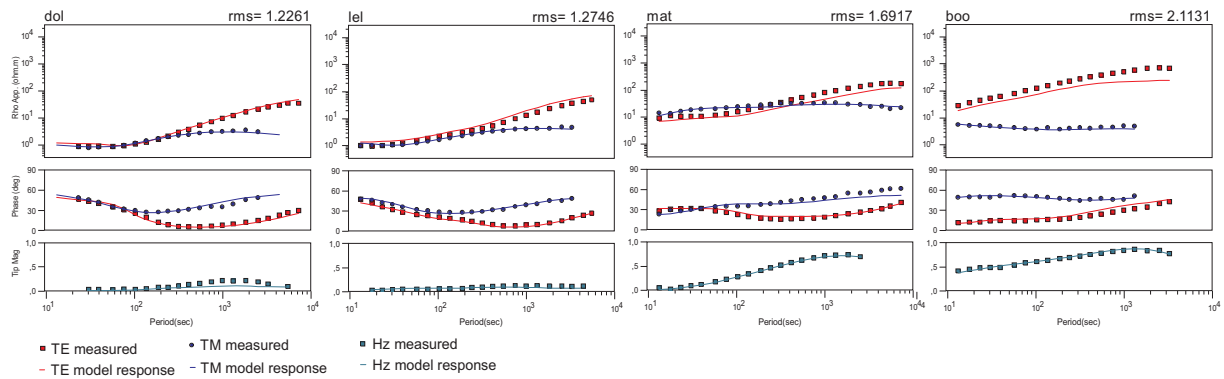


Figure 7: Examples of model responses for the 2-D inversion result as shown in fig. 6a.

Geological Interpretation

Structure A seen in the models of fig. 6 is characteristic for the North German Basin; it represents the saline aquifer and is already known from previous measurements. This structure is resolved in all inversions of this profile and could be seen in all components (TE-, TM-mode and tipper) and with its vertical elongation and extension to depths of 3 to 4 km it is consistent with simulations of MAGRI ET AL. (2007) and models of the BGR-profiles in HOFFMANN ET AL. (1997). This structure is resolved as a surface conductor in all other sections of the EMTESZ project (e.g., BRASSE ET AL. (2006), ERNST ET AL. (2008)) and in the models of HOUPPT (2008) as well. It is caused by huge deposits of Zechstein salt, which is dissolved by deeper ground water. Through thermal and hydraulic transport mechanisms it reaches in some cases up to the surface (MAGRI ET AL., 2005). In some parts of the North German Basin salinities up to 350 g/l are observed (HOTH ET AL., 1997). The conductivity of such layers depends on the salinity of the fluid, the size and connection of the pore spaces. High salt contents and pore sizes of sediment layers in North Germany may cause conductivities well above 1 S/m.

Structure "Baltica" is also resolved in all inversions of profile MVS - a very poor conductor in the north of the profile. It represents the crystalline Precambrian basement of Baltica. This basement of Baltica is with an age of 1.9 billion years one of the oldest rocks in Europe. Its high resistivity is due to the strong compression of rocks at depth with almost no interconnected fluid inclusions left.

Extension and value of the conductivity of the good conductor below Rügen Island (structure C) can also be regarded as assured due to its shallow depth and the sensitivity tests carried out (SCHÄFER, 2010). Its boundary to the north is indicated by the size of the induction arrows at MAT, the northernmost station on Rügen Island. Structure C is located at about 8 to 20 km depth. This good conductor is depicted in all inversions as a structure separated from the surface conductor (structure A).

Also structure D, with a depth of 10 to 30 km is resolved in all models of the joint inversion of TE-, TM-mode and tipper, and also proofed by sensitivity tests as a separate structure from the surface conductor. Conductors in these depths were explained with the occurrence of highly-carbonated paleozoic black shales, the so-called Scandinavian alum shales (e.g., HOFFMANN ET AL. (1998); HENGESBACH (2006)).

These alum shales crop out at the Andrarum quarry in the southern Swedish province of Scania, near MT sites AND and REG, where they have been mined since centuries. In order to test the hypothesis of high conductivity, several DC-geoelectric array measurements were conducted in this area during a student field trip of the Free University of Berlin (FIELD REPORT SCANIA,

2006). Fig. 8 displays an example. It clearly shows the black shale (blue colors) as an intermediate conductor only, with resistivities in the range of 100 to 150 Ωm . Nevertheless, this does not exclude high conductivities of deeply buried shales; at or near the surface, the relatively high resistivity may be due to strong weathering, destroying the conductive paths, which are otherwise continuous in deeper layers.

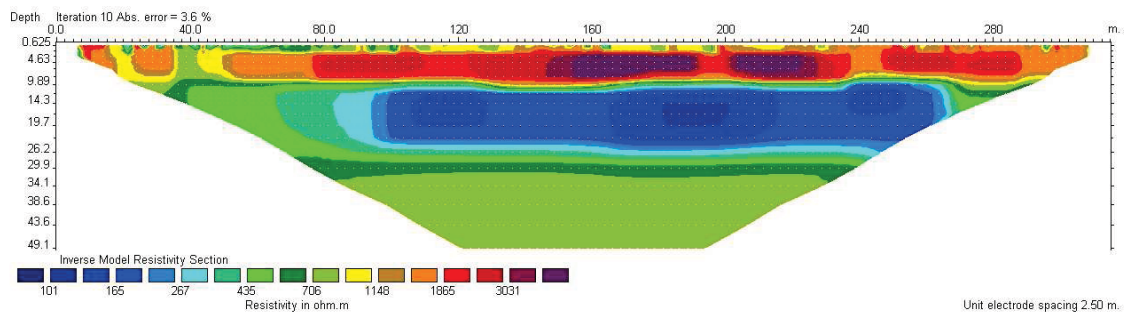


Figure 8: Geoelectric section near the Andrarum black shale quarry (South Sweden). Note the reverse color scale.

From laboratory studies and drilling black or alum shale at greater depths and without weathering it is obtained with conductivities of less than 1 Ωm (DUBA ET AL., 1988). Black shale has been investigated by drilling in the well of G14 (north of Rügen Island) and Rügen 5, where it reaches thicknesses of 50 to 70 m. In the well Pröttlin I further south, black shales were found, too, but could not be fully intersected. In addition, studies demonstrated that the necessary pore size for an electrolytic conduction mechanism not exists in deeper sediment layers (JÖDICKE, 1991). This would exclude the interpretation of structure D as a deep sedimentary trough.

Thus, the conductive "layer" in the model of figure 6b can be seen as a reference to alum shale occurrence. However, this layer can be even more conductive because of hydrothermal waters, which are often rising in geological fault zones. Under high pressure and special temperature conditions in highly-carbonated black shales also graphite-structures may form, according to estimates from DUBA ET AL. (1988) and RAAB ET AL. (1998). With a layer of mostly continuous black shale or graphite surface, a relatively low-friction thrust faulting with little deformation of the plate boundaries in the collision of Eastern Avalonia and Baltica at the time of the Ordovician could be explained.

Furthermore, RAAB ET AL. (1998) could demonstrate in laboratory experiments that pyrite inclusions (which often occur in black shale layers) may convert to much more conductive pyrrhotite at temperature and pressure conditions as encountered in the middle crust. Pyrrhotite usually occurs in dendritic form and increases conductivity even through small volumes.

Model experiments and sensitivity tests show that a 70 m thick alum shale layer at such depths is not sufficient, even if modeled with bulk resistivities as low 0.1 Ωm . Another possible explanation is the occurrence of intrusive bodies, which rose in the time of Rotliegend (Upper Carboniferous to Middle Permian). Their contact aureoles usually consist of highly conductive material (such as ilmenite or even graphite), and can extend over larger areas. So the high conductive structures C and D might be explained by a combination of highly conductive layers of black shales and such dykes. Motivation for this hypothesis emerged from the well Greifswald I, where intrusive granite bodies from the Rotliegend with apophyses could be found. This granite body is also known as Südrügen Pluton (HOTH ET AL., 1993) and is possibly extending to the south of Rügen Island.

In fact, the combination of these two interpretations, seems to be the best explanation of the high conductivities of structures C and D at depth. Thus, it is very likely a combination of graphitized black shale and intrusive bodies (e.g., Südrügen Pluton) with the accompanying conductive material and the associated apophyses. This interpretation would explain the very good conductivity in such depths and the resolution of the structures as an almost vertical body in the mid crust. Unfortunately no better resolution of these structures is possible due to the strong shielding effect of the saline aquifer.

The good conductor of model 6b ends in the south of the Stralsund Anklam Fault (SAF). The trend of the SAF, which is also indicated by magnetic transfer functions and their graphic presentation in fig. 3 fits very well to the results of HOUPP (2008) and ERNST ET AL. (2008), which were obtained further east. According to these models and the induction arrows shown in fig. 3, the SAF appears to be a fundamental crustal boundary of Northeastern Europe which can be traced to the southeast of Szczecin in Poland.

Figure 9 shows a comparison with an interpretation of the BASIN 9601 project and 2 prolonging profiles in the Baltic Sea from MCCANN & KRAWCZYK (2001). There we observe a good correlation with underlying Baltica and the Zechstein base (strong reflector), and the well-conducting sediments in the upper layers of the Northeast German Basin.

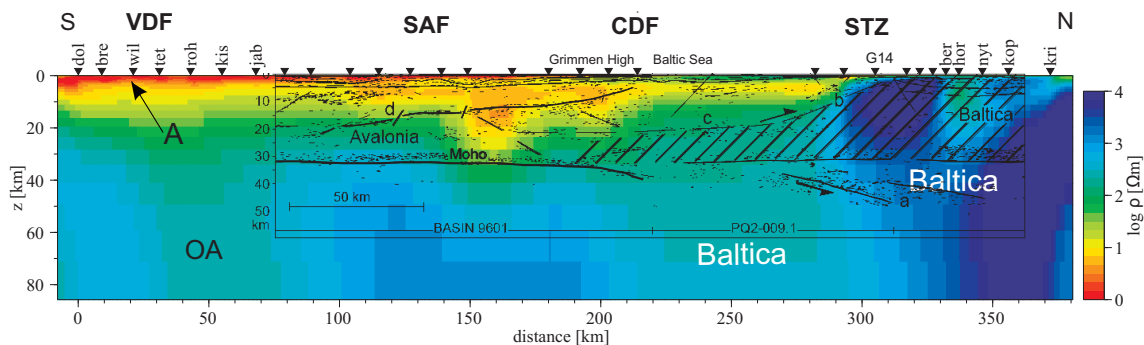


Figure 9: 2-D models for profile MVS with an overlaid interpretation of the seismic reflection/refraction profile BASIN 9601 and 2 prolonging profiles in the Baltic Sea from MCCANN & KRAWCZYK (2001).

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Magnetotelluric data from the Tien Shan and Pamir continental collision zones, Central Asia

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1 Introduction

We present magnetotelluric (MT) data obtained within the framework of the multi-disciplinary Tien Shan - Pamir Geodynamic Program (TIPAGE). The dynamics of the Tien Shan and Pamir orogenic belts are dominated by the collision of the Indian and Eurasian continental plates. With the geophysical components, we intend to image the deepest active intra-continental subduction zones on Earth (the N-dipping Hindu Kush and the S-dipping Pamir zones) and to establish how the highest strain over the shortest distance that is manifested in the India-Asia collision zone is accommodated structurally. The Tien Shan-Pamirs mountain knot forms the northwestern corner of the India-Asia

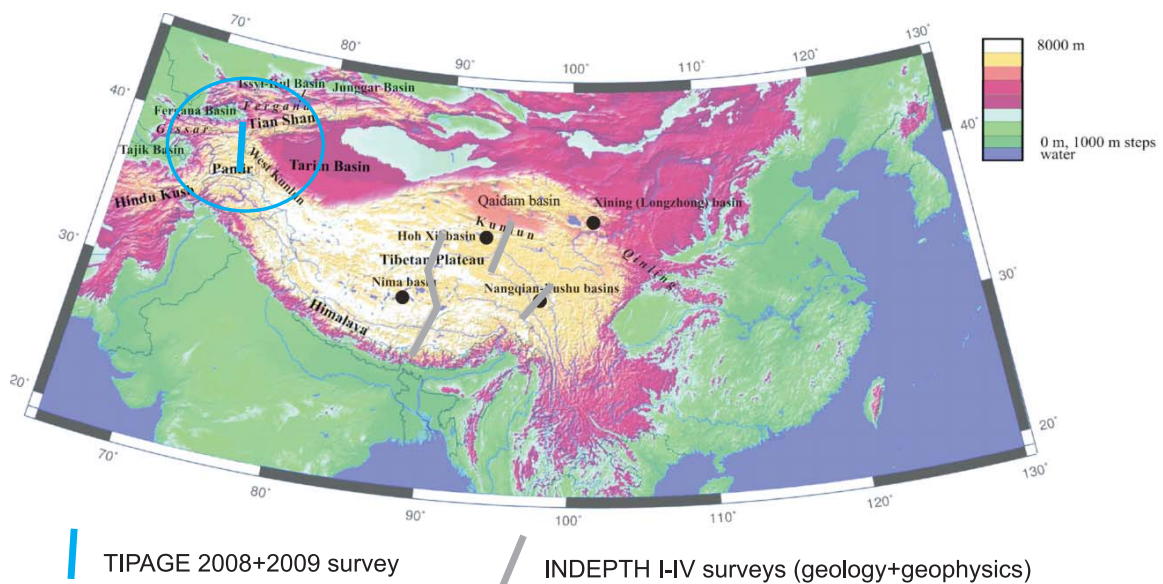


Figure 1: Central and eastern Asia orogens. TIPAGE and INDEPTH surveys are marked on the map. TIPAGE investigation area lies inside the blue circle.

collision zone and the Pamir-Tibet Plateau (Fig. 1). Note that N - S shortening in the Pamir and western Tien Shan is absorbed in less than 50% of the distance than further east. Basins denote areas little affected by intra-continental shortening. The Pamir excels over the adjacent Tibet by including the most active areas of intermediate-depth seismicity in the world and by far the most active one not associated with oceanic

subduction, three coevally active, thick-skinned, high-strain contractional belts and four distinct intra-continental magmatic belts.

2 Measurements

The MT data were recorded in summer 2008 at 80 stations in the Pamir mountain ranges in Tajikistan and in summer 2009 at 98 stations in the Pamir and the southern Tien Shan in Kyrgyzstan and Tajikistan (Fig. 2). A typical spacing was approximately 2 km between BB-only sites and 14 km for the combined BB+LMT sites. The stations form an approximately 340 km long profile from Osh in Kyrgyzstan via Sarytash, the Kyrgyz-Tajik border, Karakul and Murgab to Zorkul in southern Tajikistan.

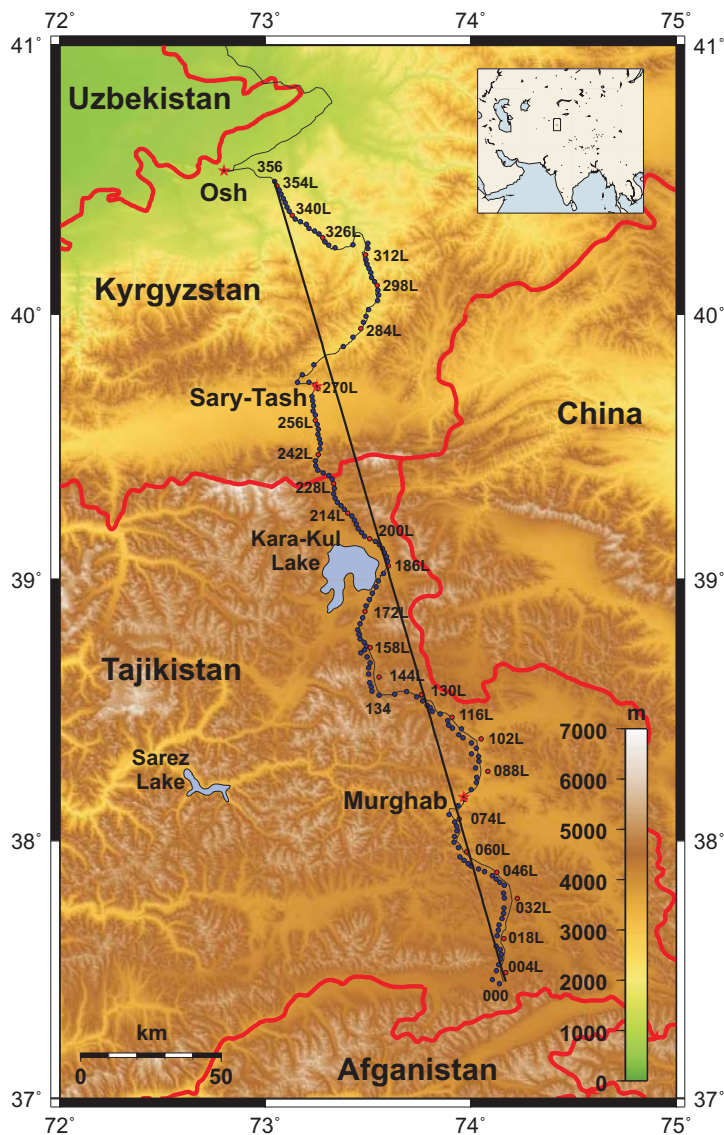


Figure 2: Topographic map with stations and profile of the TIPAGE survey. Numbers of stations with broad band and long period measurements (L) are marked. The station alignment follows the a single road in that remote area with rough terrain.

3 Preliminary Data Analysis

Data examples

Some representative graphs of apparent resistivities and phases are shown in Fig. 3. The curves were obtained after a recording time of three days, using robust single site processing. Varying shapes of the curves indicate considerable changes in the underlying conductivity structure. The distances between the shown stations are approximately 50 km. For station locations, see Fig. 2. The overall data quality is superb, especially in the very remote southern parts of the profile.

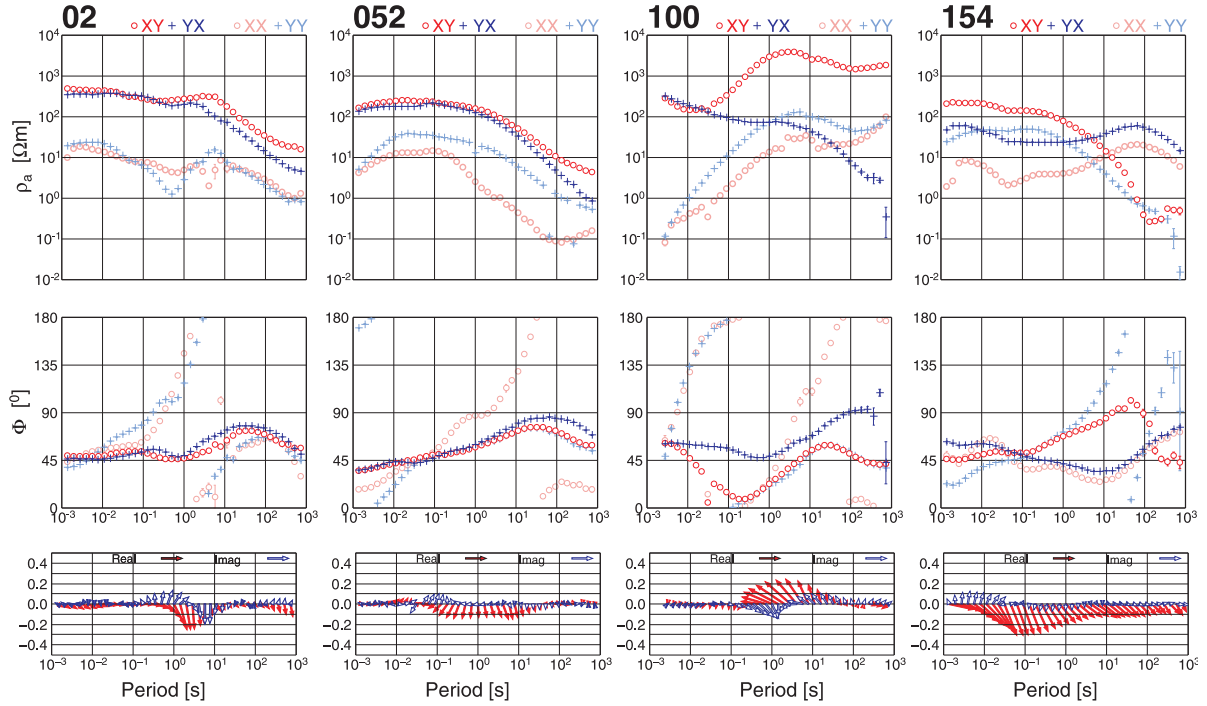


Figure 3: Exemplary apparent resistivity and phase curves of four broad-band stations from the Pamir. For station locations, see Fig. 2.

Pseudosections

Apparent resistivities and impedance phases of all TIPAGE stations are plotted as pseudo sections in Fig. 4. In the southern part of the profile, the TE and TM modes apparent resistivities show values below $10 \Omega m$ at higher periods and high phase values reaching 90° . There are more low-resistivity features further north at middle to high periods, which are more present in the TE mode. This hints at a deep conducting structure in the southern profile part and a complex distribution of conducting features in the northern segment of the profile.

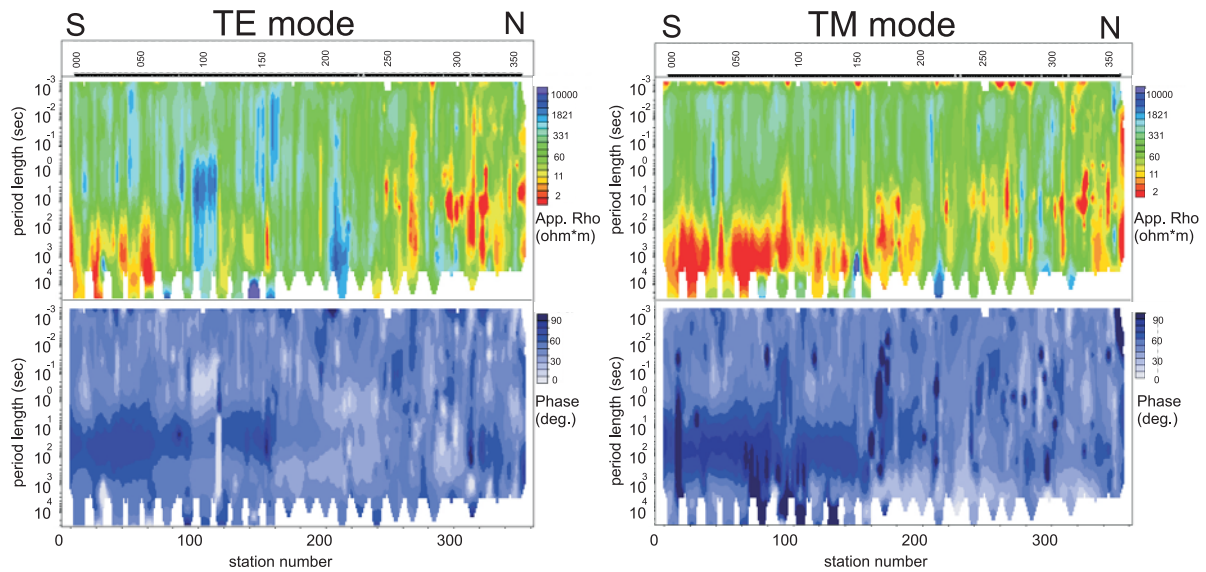


Figure 4: Pseudosections of apparent resistivity and phase data, in TE and TM modes. In the left part of each pseudosection the southern profile stations are plotted, in the right part the northern stations. For stations location, see Fig. 2.

Induction arrows

The presence or absence of lateral variation in conductivity can be inferred from induction arrow maps. In Fig. 5 induction arrows are plotted using the Wiese convention for periods of 1, 32, 128, and 1024 s. The arrow distribution indicates several 'reversals', which suggest the presence of elongated good conductors below the profile. Altogether the induction arrow plots exhibit a complex distribution of the subsurface conductivity structure.

Strike analysis

Regional strike analysis (Becken & Burkhard, 2003) of sites recorded in the Pamir and Tien Shan is displayed in Fig. 6. The analysis of surface sensitive higher frequencies (Fig. 6 left) shows a E - W (N - S) strike direction. This is consistent with the predominant E - W distribution of geological structures. The lower frequencies (Fig. 6 right) reveal a pronounced geoelectric strike of approximately -17° to -20° . This strike direction may be in agreement with structures of the India-Asian collision zone.

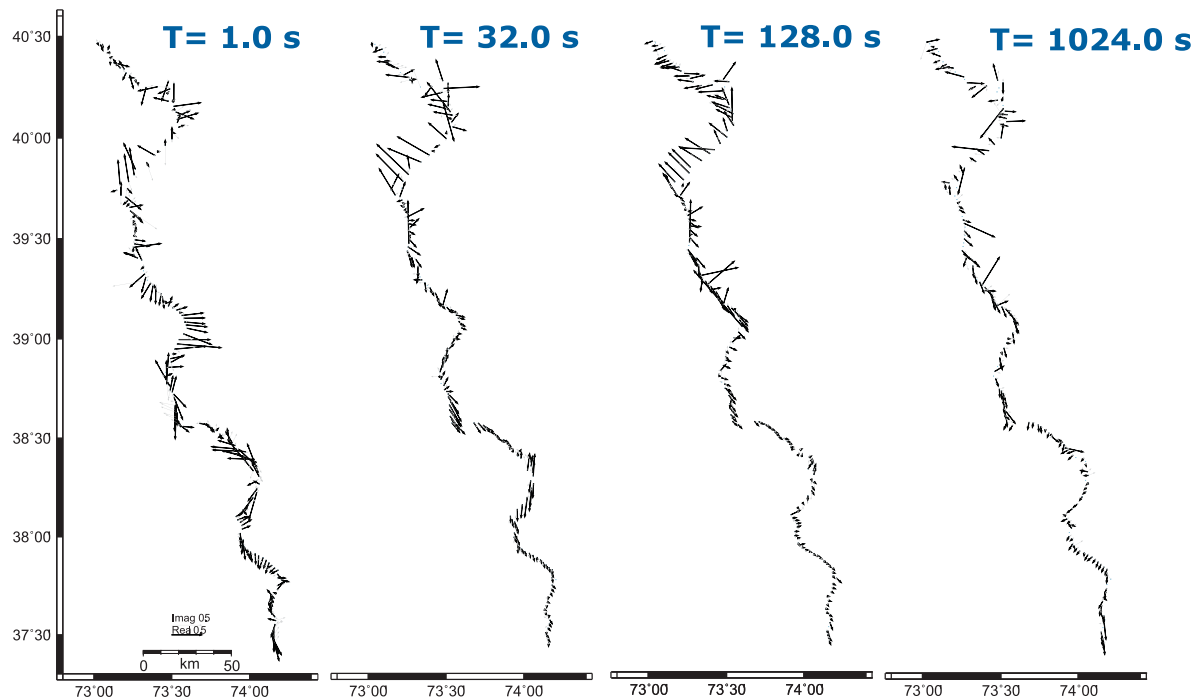


Figure 5: Maps of induction vectors (Wiese convention) for periods of 1, 32, 128, and 1024 s. Each arrow starts at one station location, compare Fig. 2.

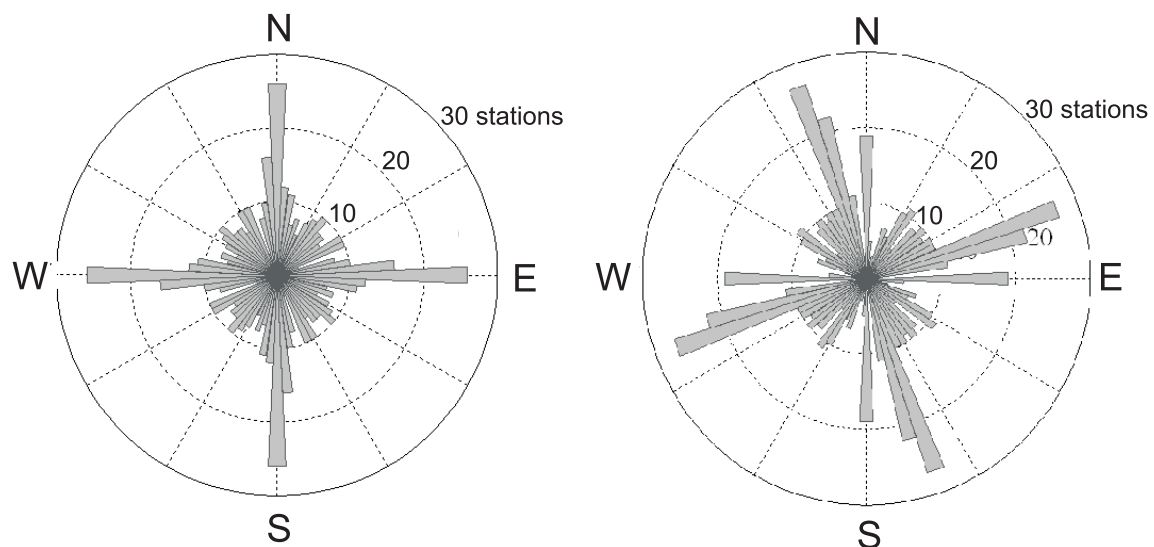


Figure 6: Regional strike analysis (Becken & Burkhard, 2003) of all sites recorded in the years 2008 and 2009. Left for the period range 0.001 s - 10 s, right for the period range 10 s - 10000 s.



Figure 7: The Kyrgyz/Russian - Tajik - German MT team of 2008 (left image) and 2009 (right image). The Research station of the Russian Academy of Sciences in Bishkek provided most of the field logistics, including 4WD vans and 4WD trucks.



Figure 8: The amazing hospitality of local people during the field campaigns was a memorable experience.

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Audiomagnetotellurik im Hohen Vogelsberg

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15. April 2010

Einleitung

Der Vogelsberg ist mit einer Fläche von 2100km² das größte geschlossene Vulkangebiet Mitteleuropas und Teil einer im Tertiär aktiven vulkanischen Zone, die sich von der Eifel bis nach Schlesien erstreckt. Vor 18 Millionen Jahren setzte die Förderung der Schmelzen aus über 100km Tiefe ein und war ca. vier Millionen Jahre lang aktiv. Heute stellen die vulkanischen Gesteine des Vogelsberges einen der bedeutendsten Grundwasserspeicher Hessens dar.

Im Jahr 2007 wurde die Forschungsbohrung Siechenhausen-Eschwald in der Nähe des Hoherodskopfes abgeteuft. In dieser Bohrung wurden ab einer Teufe von fast 200m Gesteine erbohrt, die auf einen endogenen Lavadom hinweisen.

Um die laterale Ausdehnung dieses Körpers genauer zu untersuchen, wurde 2008 durch das Hessische Landesamt für Umwelt und Geologie (HLUG), das Leibniz-Institut für Angewandte Geophysik (LIAG) und die Universität Frankfurt am Main geophysikalische Messungen mit den Methoden Geoelektrik und Magnetotellurik durchgeführt.

Ergänzend dazu wurden im November 2008 audiomagnetotellurische (AMT) Messungen an 15 Standorten durchgeführt. Neben der Erkundung der lateralen Ausdehnung des Störkörpers sollen die Messungen der Geoelektrik verifiziert werden.

Stationsübersicht

Das Messgebiet (Abb. 1) mit einer Größe von 3x4km² liegt rund um den Rehberg, ca. 10 km östlich der Stadt Schotten. Im Zentrum des Gebietes wurde die Forschungsbohrung Siechenhausen-Eschwald abgeteuft, das ausgewählte Gebiet liegt also direkt über dem vermuteten endogenen Lavadom. Da die Messungen auch als Tiefenergänzung zur großskaligen Geoelektrik von AGRICOLA ET AL (2009) dienen, wurden viele der Stationsstandorte gewählt, an denen bereits die Datenlogger der Geoelektrikmesskampagne standen.

Wegen der gewünschten Zieltiefe von 1500m wurde ein Stationsabstand von ca. 500 Metern angestrebt, um ein aussagekräftiges Messraster zu erhalten. Die Aufzeichnung der elektromagnetischen Felder erfolgte mit der Metronix ADU07 mit einer Aufzeichnungsdauer von bis zu 12 Stunden. Die aufgenommenen Daten wurden im Bereich von 10Hz bis 2500Hz ausgewertet.

Für die Messung der horizontalen elektrischen Felder werden mit einer gesättigten Kaliumchlorid-Wasser-Lösung befüllte Sonden verwendet, in denen sich potentialfreie Silber-Silberchloridelektroden befinden. Die Felder werden aus Spannungsdifferenzen eines Elektrodenpaares berechnet, eine Mindestauslagenlänge von 10 Metern zwischen den Elektroden wurde eingehalten.

Die zeitlichen Variationen der magnetischen Felder wurden an jeder Station in alle 3 Raumrichtungen den Induktionsspulen (Metronix MFS-05) aufgezeichnet.

Um eine Tiefenaufklärung von 100m - 1500m zu erreichen, genügt es, eine hochfrequente (HF)- Messung mit einer Abtastrate von 16384Hz und eine Niederfrequente (L1)- Messung mit einer Abtastrate von 2048Hz aufzunehmen. Da sich anthropogene Störquellen in unmittelbarer Nähe befanden, wurden lange Zeitreihen angestrebt. Als Minimum wurden 5 Minuten im HF- und 60 Minuten im L1-Bereich gewählt.

Datenbearbeitung

Die gemessenen Signale im Bereich des Vogelsberges sind stark von anthropogenen Störsender geprägt. Im Folgenden wird die Datenbearbeitung zur Reduktion des Einflusses der Störquellen beschrieben.

Harmonische Störsignale

Harmonische Störsignale können im Energiedichtespektrum gut erkannt werden, da sie sich dort als scharfer Peak darstellen (Abb. 2 (b)). Deutlich erkennbar sind die $16\frac{2}{3}$ Hz Störfrequenzen der Bahn sowie die 50Hz Netzstromfrequenz und deren Multiple. Na-

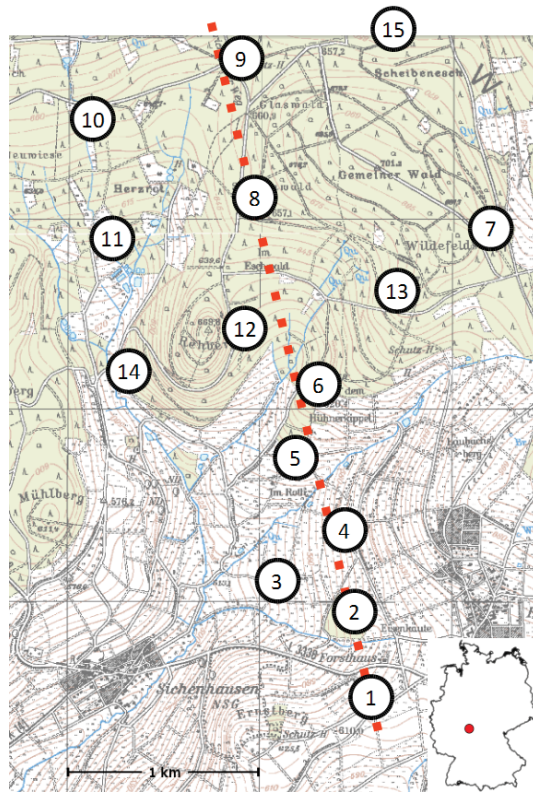


Abbildung 1: Übersicht der Stationsstandorte mit Profil der Auswertung (rote Linie)
Kartengrundlage: TK25 Hessische Verwaltung für Bodenmanagement und Geoinformationen

türliche Signale werden von diesen Störfrequenzen überlagert und sind visuell schwer erkennbar (Abb. 2 (a)).

Um eine visuelle Datenauswahl zu ermöglichen, werden im Frequenzbereich die stark gestörten Frequenzbänder eliminiert, indem die Energie des entsprechenden Bandes gleich Null gesetzt wird. Durch eine Rücktransformation in den Zeitbereich wird eine Zeitreihe erzeugt, die nahezu frei vom Einfluss harmonischer Störer ist.

Dieses Verfahren kann ohne negative Beeinflussung der Qualität der Übertragungsfunktionen durchgeführt werden, da diese über einen bivariaten Ansatz

$$\begin{pmatrix} E_x \\ E_y \end{pmatrix} = \begin{pmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{pmatrix} \begin{pmatrix} B_x \\ B_y \end{pmatrix}$$

im Frequenzbereich außerhalb der gestörten Frequenzbänder berechnet werden.

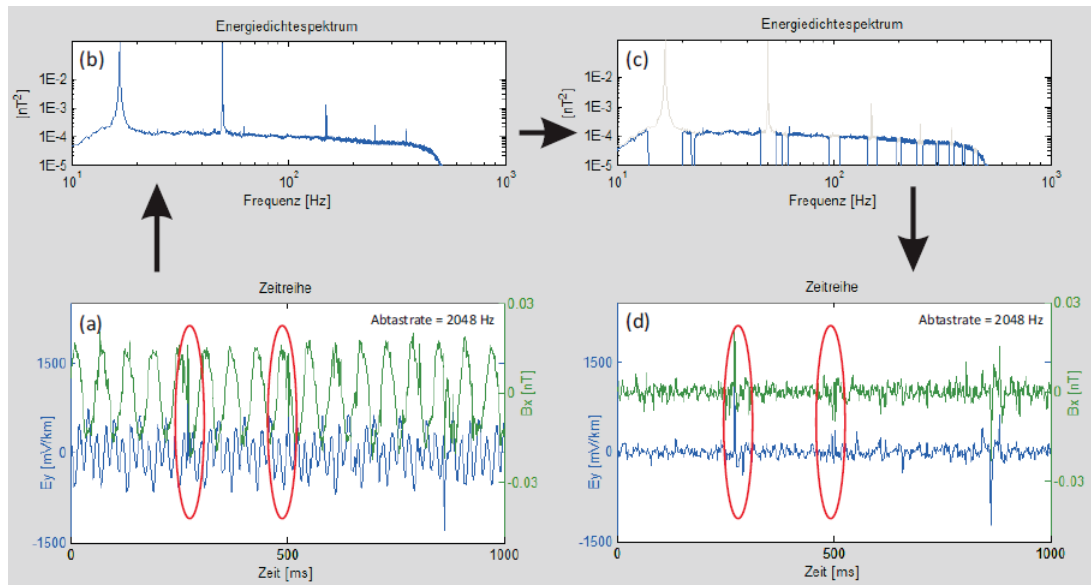


Abbildung 2: Bearbeitung harmonischer Störsignale am Beispiel der magnetischen Nordkomponente (Bx) und der elektrischen Ostkomponente (Ey) der Station 09. Das Energiedichtespektrum ist nur für den Bx-Kanal gezeigt.

Gepulste Störsignale

Gepulste Störsender können im Energiedichtespektrum im Allgemeinen nicht erkannt werden, da es sich um Deltapulse handelt, die beim Überführen in den Frequenzbereich über das komplette Spektrum verschmieren. Durch die Elimination der harmonischen Störsignale können die gepulsten Störsignale in der Rücktransformierten Zeitreihe sichtbar gemacht und erkannt werden. Nach dem Auswählen der Positionen störsenderunbehafteter Zeitbereiche können aus den entsprechenden Bereichen der Originalzeitreihe Übertragungsfunktionen berechnet werden, die weder von harmonischen noch von gepulsten Störsignalen beeinflusst werden.

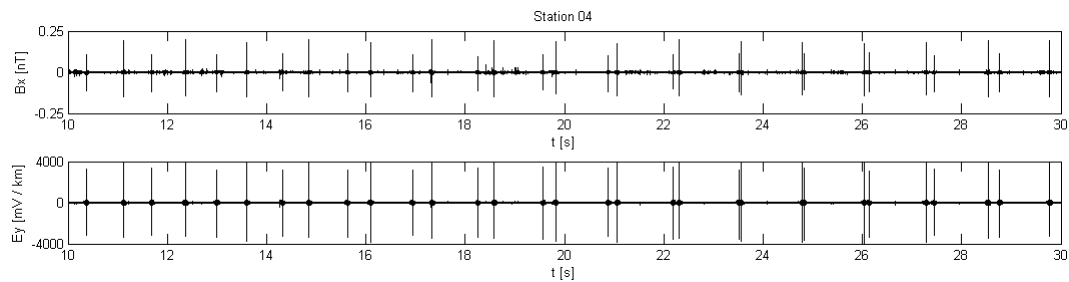


Abbildung 3: Gepulste Störsender mit Periode von ca. 1 Sekunde am Beispiel der Bx- und Ey-Komponente von Station 04.

Ergebnisse

Impedanztensor

Am Beispiel der Rhoa- und Phasenkurve des Impedanztensors der Station 1 (Abb. 4) ist eine deutliche Verbesserung der Datenqualität nach der Elimination der Störsignale zu erkennen. Die manuelle Datenauswahl brachte bei dieser Station gerade bei den hohen Frequenzen (150Hz - 3000Hz) eine deutliche Verbesserung. Vergleicht man die Ergebnisse der Station 1 mit den Rhoa- und Phasenkurven anderer Stationen (Abb. 5), so erkennt man, dass die Übertragungsfunktionen rund um den Störkörper nicht übereinstimmen, z.B. ist bei Station 11 die Phasenaufspaltung bei hohen Frequenzen stärker ausgeprägt und die xy-Komponente der Phase liegt unter der yx-Komponente, bei Station 05 ist es umgekehrt. Das Aufspalten der beiden Komponenten in der Phase deutet auf eine laterale Änderung der Leitfähigkeit hin.

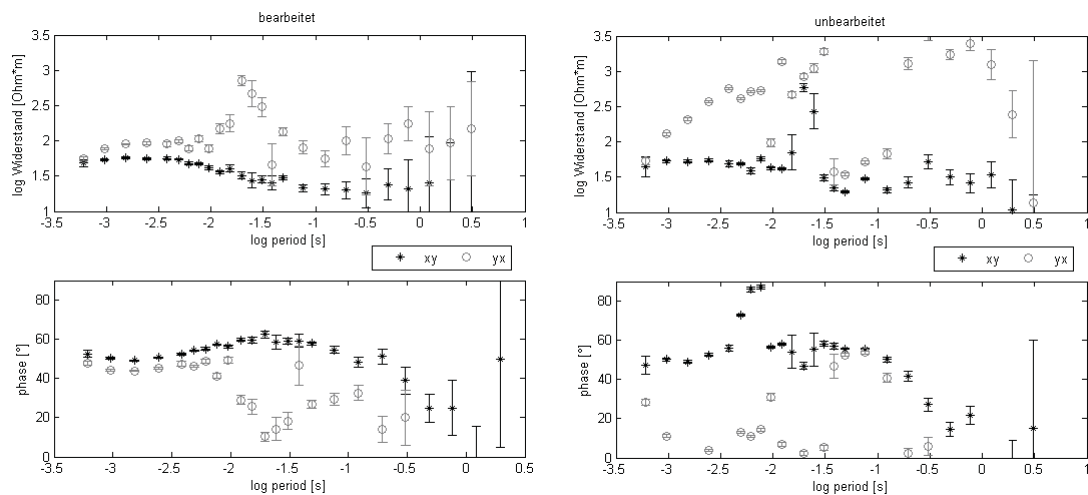


Abbildung 4: Ergebnisse der Datenauswertung am Beispiel der Station 1: Vergleich zwischen Elimination von harmonischen und gepulsten Störsendern (a) und standard robustes Auswerteverfahren (b).

Das Abfallen der Phasen beider Komponenten ab einer Periode von 0.1 Sekunden wird auf die geringe Anregungsenergie in diesem Bereich zurückgeführt. Da keine lang-periodischen Daten aufgenommen wurden, wurden die Impedanztensoren in der 1D-Modellierung nur bis ca. 5Hz betrachtet.

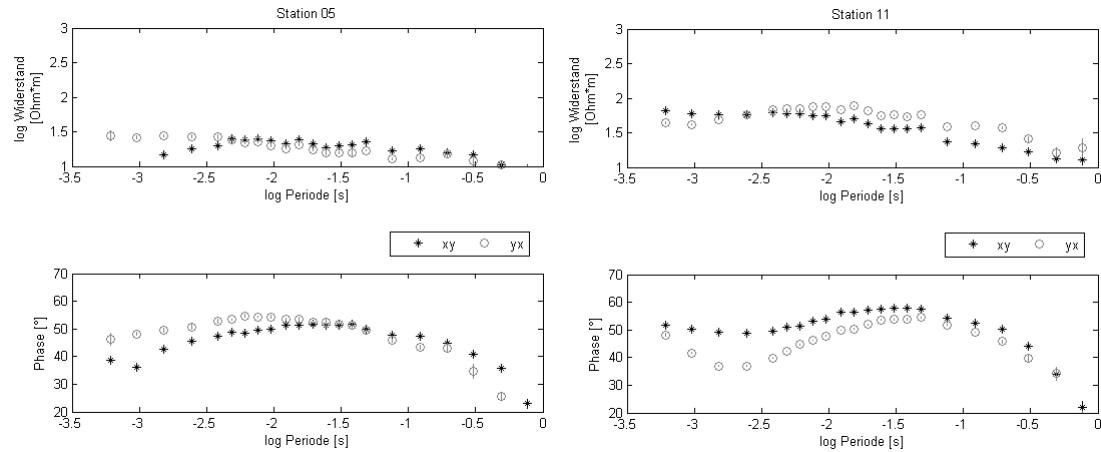


Abbildung 5: Rhoa- und Phasenkurve der Stationen 5 (links) und 11 (rechts).

1D-Modelle

Die Daten aller Stationen wurden zunächst mit dem WAIT-Algorithmus 1D-interpretiert. Als Datengrundlage wurde, unter Annahme einer 1D-Näherung, die BERDICHEVSKI-Invariante $Z_B = 0.5 \cdot (Z_{xy} - Z_{yx})$ verwendet. Die verschiedenen, voneinander unabhängigen

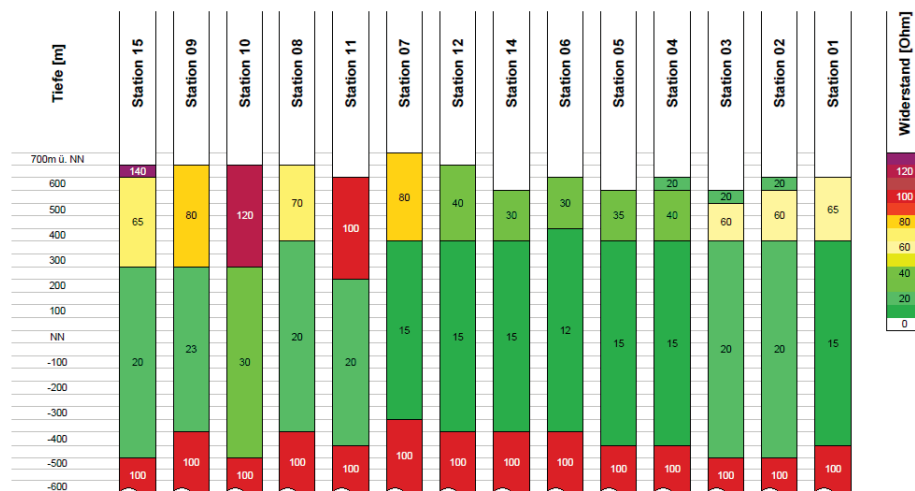


Abbildung 6: Grafische Darstellung der 1D-Modellierung projiziert auf ein Nord-Süd Profil.

berechneten 1D-Modelle (Abb. 6) zeigen, unter Berücksichtigung der Topographie, eine gut leitende, ca. 1km mächtige Schicht ab einer Tiefe von 350m ü. NN. Die Modelle unterscheiden sich hauptsächlich in der oberen Schicht. Es fällt ein gut leitender Bereich in der Mitte des Messgebietes von Station 12 bis Station 04 auf.

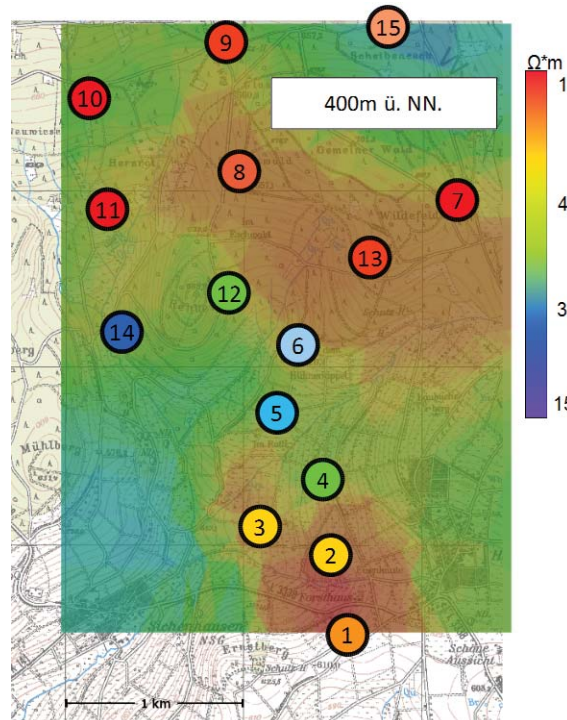


Abbildung 7: Vergleich AMT 1D-Interpretation mit Inversionsergebnissen der Geoelektrik, mit Stationsnummern.
Kartengrundlage: TK25 Hessische Verwaltung für Bodenmanagement und Geoinformationen

Um einen Vergleich zur großskaligen Geoelektrik (AGRICOLA ET AL (2009)) zu schaffen, wurde ein Tiefenschnitt der 1D-Modelle bei 400 m ü. NN. über den entsprechenden Tiefenschnitt der 3D-Geoelektrikinversion gelegt (Abb. 7). Die Werte der AMT-Modelle werden durch die Hintergrundfarbe der entsprechenden Stationsnummern dargestellt, die Legende des spezifischen Widerstands gilt sowohl für die AMT als auch für die Geoelektrik Ergebnisse.

Beide Messmethoden zeigen, bis auf leichte Abweichungen im Norden, die selben elektrischen Leitfähigkeitsstrukturen. In der Mitte des Messgebietes ergeben die Modellierungen beider Methoden einen besser leitenden Bereich. Die Abweichungen zwischen den Ergebnissen der audiomagnetotellurischen Daten und denen der Tiefengeoelektrik lassen sich wie folgt erklären:

Ein großer Unterschied der beiden Darstellungen ist, dass die AMT Modellierung einen gemittelten spezifischen Widerstand für einen Körper von 300 Metern Mächtigkeit zeigt, die Inversion der Geoelektrik hingegen die Widerstandsverteilung in einer festen Tiefe von ca. 250 Meter.

Die starke Abweichung am nördlichen Rand des Messgebietes (Station 15) wird darauf zurück geführt, dass die Geoelektrikdaten diesen Bereich schlecht auflösen und durch das Inversionsprogramm interpoliert wurden. Da die AMT-Daten eine Tiefensondierung ermöglichen ist dieser Bereich von der AMT besser aufgelöst.

3D-Modell

Auf Basis der durch die AMT-1D- und Geoelektrik-3D Modellierung gewonnenen Erkenntnisse wurde ein AMT-3D-Modell vorgegeben und die Modellantwort mit der Software X3D berechnet. Die verwendeten Modellparameter sind in Abb. 8(unten) darge-

stellt. Der Hintergrundwiderstand der Deckschicht (0-25m) liegt bei $45\Omega\text{m}$. Die weißen Felder haben den jeweiligen Hintergrundwiderstand von $45\Omega\text{m}$ in der ersten (25-375m) und $35\Omega\text{m}$ in der zweiten Schicht (375-725m). Darunter folgt ein homogener Halbraum mit $1000\Omega\text{m}$.

Die Phasentensorellipsen (nach CALDWELL ET AL 2004) wurden von allen Stationen auf ein in Nord-Süd-Richtung verlaufendes Tiefenprofil projiziert (Abb. 8). Um die Messdaten mit der 3D-Modellierung erklären zu können, mussten neben den beiden aus der 1D-Modellierung bekannten Störkörpern noch ein schlechter Leiter ($1000\Omega\text{m}$) in ca. 400m Tiefe eingefügt werden, der sich von Nord nach Süd durch das komplette Messgebiet erstreckt, in Ost-West-Richtung aber nur 120 Meter breit ist.

Dieser Störkörper stellt eine Verbindung zwischen den beiden oberflächennahen hochohmigen Bereichen dar. Diese wird auch aufgrund der geologischen Gegebenheiten vermutet. Es ist zu sehen, dass die Phasen der gemessenen Daten in der Profilmitte in etwa den Phasen dieses Modells entspricht. Außerdem werden in diesem Bereich die Abplattungen der Ellipsen besser angedeutet. Auch ein Anstieg der niedrigen Phasen der Zielfrequenzen mit längeren Periode ist gut zu erkennen. Ein Anstieg der Phasen im Süden des Profils bei einer Periode von ca. 2 Sekunden, der im Modell vom kleineren südlichen Störkörper hervorgerufen wird, ist ebenfalls in den Daten zu sehen. Die südlichsten Stationen 01, 02 und 04 wurden aufgrund ihrer stark gestörten elektrischen Ost-West-Komponente nicht in den Vergleich mit dem 3D-Modell mit einbezogen.

Fazit

Durch das Eliminieren der harmonischen Störsignale wird die visuelle Datenauswahl erheblich vereinfacht. Bei gepulsten Störsignalen führt die manuelle Auswahl der für die Analyse herangezogenen Datenbereiche zu einer starken Qualitätsverbesserung bei den berechneten Übertragungsfunktionen.

Es gibt eine gute Übereinstimmung zwischen den Ergebnissen der 1D- und 3D-Modellierung der audiomagnetotellurischen Messdaten und 3D-Inversionsrechnung der großskaligen Geoelektrik bis zu einigen hundert Metern Tiefe. Ein Leitfähigkeitsunterschied, der durch einen endogenen Lavadom interpretiert werden kann, ist mit der AMT erkennbar. Die laterale Ausdehnung kann nicht gezeigt werden, da die Randbereiche der Leitfähigkeitsanomalien nicht vollständig erfasst wurden.

Die AMT kann zur Verifikation und Tiefenergänzung großskaliger Geoelektrikmessungen herangezogen werden.

Quellen

- Agricola, T., Günther, T., Junge, A., Kracht, M., Nesbor, D. (2009): *Dem Trachyt-dom im Hohen Vogelsberg mit großskaliger Tiefengeoelektrik auf der Spur*, DGG Jahrestagung 2009 Kiel, Abstract, Vortrag
- Caldwell, C., Bibby, T. G., Brown, H. M. (2004): *The magnetotelluric phase tensor*, Geophys. J. Int., 2004 (158):457-469

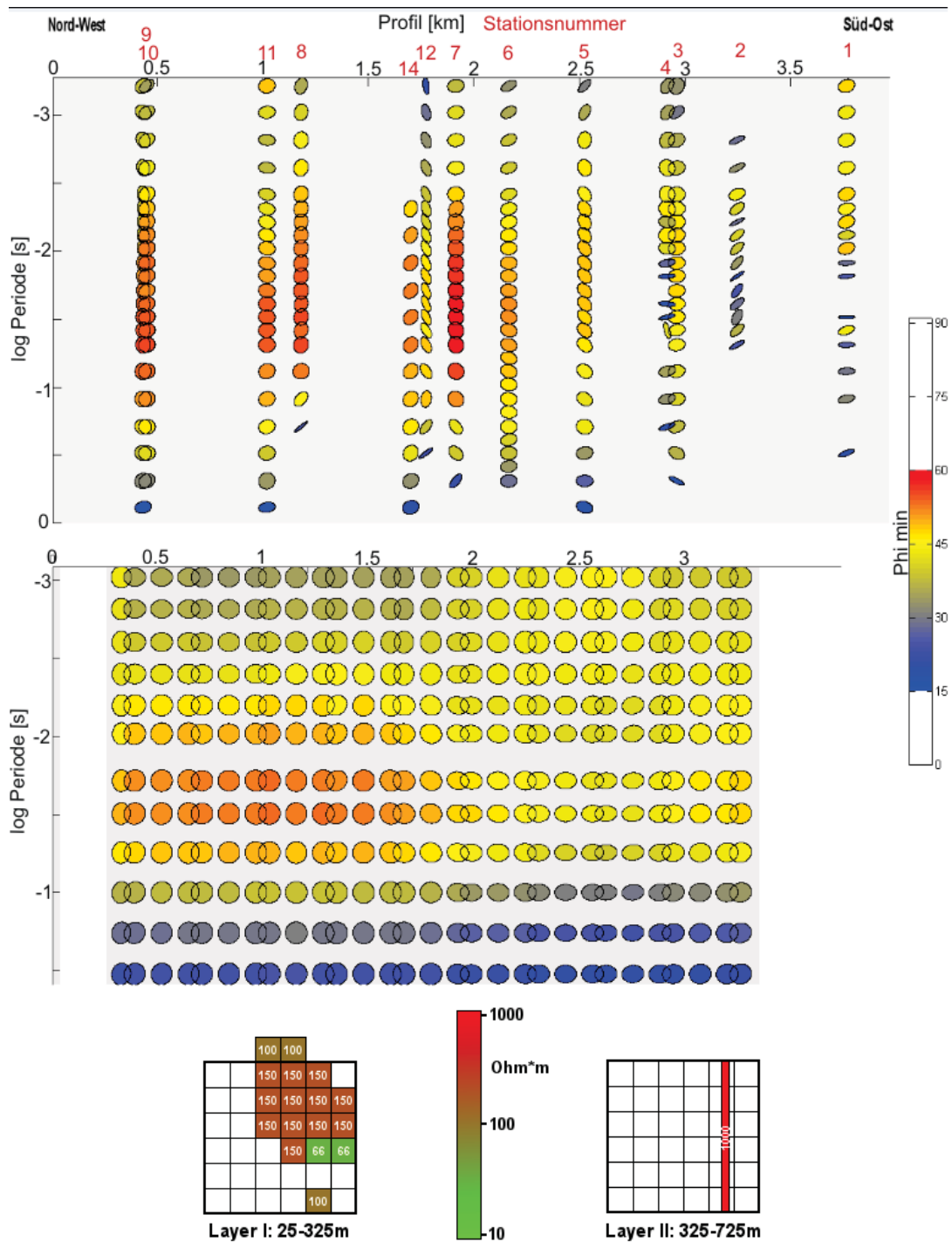


Abbildung 8: Vergleich der minimalen Phasen der Phasentensorellipsen des 3D-Modells (mitte) mit denen der gemessenen Daten (oben) und Darstellung der Modellparameter (unten).

In our session "What I always wanted to know ..." we had the question

How would you choose the period for LOTEM?

Carsten Scholl continued thinking about it even after the colloquium and send an email with some additional thoughts. Here is his email:

Well, in the discussion I said that it is not necessary to wait until the signal is decayed into the noise level before switching again, which is totally true. Actually, this would be to some extent a circular argument, because the noise level depends on the number of stacks, and the number of stacks depends on the period.

So, one should check before with synthetic models, which time range is required to resolve a certain feature. Note, the time range to RESOLVE a feature, is not the same (and typically significantly larger) than the time range up to the point where deviations between the "target" and "no target" curves appear. I recommend doing 1D inversions with reasonable noise estimates for resolution studies... . Further, it is advisable to increase this time, because the true background resistivity might deviate from the synthetic und thus delay the features representing the target.

The time saved by using a higher period and thus collecting the required number of stacks faster can be used by measuring at more sites.

Often, in academic LOTEM campaigns, the target depth is not well defined but you'd like to get as deep as possible. Further, setting up LOTEM stations is tedious and the number of stations is limited anyways because of limited equipment. Often, the time spent on deploying and packing up the sensors exceeds the actual measurement time. So, in this case, there is not really the option to measure more sites on the same day (well, I always envision a roll-along scheme, where you start to measure at one station while building up the next site. When the final site is set up, the first site is redeployed, while the TX is still running...).

In this case, I recommend to take some generous period the first day. In the evening the data for the day should be processed to see what the latest usable time is (this should be done with either a switch-off E-field or a magnetic component). For the subsequent days, the period should be set to something slightly longer (say, a factor of 2) longer than this time.

Another aspect of this question: You should check, whether your time domain forward solver actually can handle higher switching times by taking into account previous transients. Otherwise, it might be better to stick to a more conservative period as well.