5 Electromagnetic methods – frequency domain

5.1 Airborne techniques

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5.1.1 Introduction

Modern frequency domain airborne electromagnetic (AEM) systems utilise small transmitter and receiver coils having a diameter of about half a metre. The transmitter signal, the primary magnetic field, is generated by sinusoidal current flow through the transmitter coil at a discrete frequency. As the primary magnetic field is very close to a dipole field at some distance from the transmitter coil, it can be regarded as a field of a magnetic dipole sitting in the centre of the transmitter coil and having an axis perpendicular to the area of the coil. The oscillating primary magnetic field induces eddy currents in the subsurface. These currents, in turn, generate the secondary magnetic field which is dependent on the underground conductivity distribution. The secondary magnetic field is picked up by the receiver coil and related to the primary magnetic field expected at the centre of the receiver coil. As the secondary field is very small with respect to the primary field, the primary field is generally bucked out and the relative secondary field is measured in parts per million (ppm). Due to the induction process within the earth, there is a small phase shift between the primary and secondary field, i.e., the relative secondary magnetic field is a complex quantity. The orientation of the transmitter coil is horizontal (VMD: vertical magnetic dipole) or vertical (HMD: horizontal magnetic dipole) and the receiver coil is oriented in a maximum coupled position, resulting in horizontal coplanar, vertical coplanar, or vertical coaxial coil systems.
5.1.2 Theory

**Calculation of secondary field values**

The secondary magnetic field for a stratified subsurface caused by an oscillating (frequency $f$) magnetic dipole source in the air is calculated using well-known formulae (e.g. Wait 1982, Ward and Hohmann 1988). They are based on Maxwell’s equations and solve the homogeneous induction equation in the earth for the electromagnetic field vector $\mathbf{F}$

$$\frac{d^2 F}{dz^2} = v^2 F, \quad v^2 = \lambda^2 + i\omega\mu_\sigma - \omega^2 \mu_e$$

(5.1)

assuming a homogeneous and isotropic resistivity $\rho$, which is the reciprocal of the conductivity $\sigma$, $\omega = 2\pi f$ is the angular frequency, $\lambda$ is the wave number, and $i = (-1)^{\frac{1}{2}}$ is the imaginary unit. Magnetic effects and displacement currents are normally neglected, i.e., the magnetic permeability $\mu$ is set to that of free space, $\mu = \mu_0 = 4\pi \times 10^{-7}$ Vs/Am, and the dielectric permittivity $\epsilon$ is assumed to be far less than $\sigma/\omega$ yielding a propagation factor $v^2 = \lambda^2 - i\omega\mu_0\sigma$. The inhomogeneous induction equation containing the source term has to be solved in a non-conductive environment (air) and both solutions are combined at the earth’s surface.

For a horizontal-coplanar coil pair with a coil separation $r$ and at an altitude $h$ above the surface, the relative secondary magnetic field $Z$ is given by (e.g. Wait 1982)

$$Z = r^{3-h} \int_0^\infty R_1(f, \lambda, \rho(z)) \lambda^2 e^{-2\lambda h} J_0(\lambda r) d\lambda$$

(5.2)

where $R_1$ is the complex reflection factor containing the underground vertical resistivity distribution $\rho(z)$ with $z$ pointing vertically downwards, and $J_0$ and $J_1$ are Bessel functions of first kind and zero or first order, respectively, which can be approximated by

$$J_0(x) \approx 1 - \frac{x^2}{4} + \frac{x^4}{64} - ...$$

(5.3)

$$J_1(x) \approx \frac{x}{2} - \frac{x^3}{16} + \frac{x^5}{384} - ...$$
Similar formulae are valid for vertical-coplanar
\[ Y = r^2 \int_0^\infty [R_1(f,\lambda,\rho(z))\lambda e^{-2\lambda h} J_1(\lambda r)] d\lambda \] (5.4)
and vertical-coaxial
\[ X = r^2 \int_0^\infty [R_1(f,\lambda,\rho(z))\lambda e^{-2\lambda h} \frac{J_1(\lambda r) - \lambda J_0(\lambda r)}{2}] d\lambda \] (5.5)
coil configurations. As long as \( r < 0.3h \), the horizontal secondary field values perpendicular (Eq. 5.4) and along (Eq. 5.5) the transmitter-receiver direction can be approximated by \( Y \approx Z/2 \) and \( X \approx -Z/4 \), respectively (Mundry 1984). Thus, only \( Z \) is regarded in the following.

The reflection factor \( R_1 \) is derived for the model of a n-layered half-space by a recurrence formula, see Frischknecht (1967) or Mundry (1984):
\[
R_{j-1} = \frac{K_{j-1} + R_j e^{-2j\gamma}}{1 + K_{j-1} R_j}, \quad R_{n-1} = K_{n-1}, \quad j = n - 1, \ldots, 2
\] (5.6)
where \( \rho_j \) and \( t_j \) are the model parameters resistivity and thickness of the \( j^{th} \) layer (\( j = 1 \): air layer, i.e., \( \rho_1 \) infinite, and \( t_1 = h \), sensor altitude; \( j = n \): substratum, i.e., \( t_n \) infinite). Substituting \( k = \lambda h \) and setting \( \gamma = r/h \) in Eq. 5.2 yield to
\[
Z = \gamma^3 \int_0^\infty [R_1(f,k,\rho(z))k^2 e^{-2k} J_0(\lambda k)] dk
\] (5.7)
The complex integrals in Eqs. 5.2, 5.4, 5.5, and 5.7 can be evaluated numerically using fast Hankel transforms (Anderson 1989, Johanson and Sørenson 1979). A very fast calculation of the integral in Eq. 5.7 is realised by using a Laplace transform (Fluche and Sengpiel 1997). In this case, the coil separation has to be sufficiently smaller than the sensor altitude in order to approximate the Bessel function \( J_0 \) by the first term(s) of Eq. 5.3.

The reflection factor for a homogeneous half-space model (\( n = 2 \), \( \rho_2 = \rho \), \( t_2 \) infinite) can be derived from Eqs. 5.6 and 5.7 using the substitution \( k = \lambda h \) (cf. Mundry 1984)
\[
R_1 = \frac{v_1 - v_2}{v_1 + v_2}, \quad v_1 = k, \quad v_2 = \sqrt{k^2 + i2\delta^2} = v,
\] (5.8)
where \( \delta = \frac{h}{p} \), \( p = \frac{2\rho}{\omega\mu_0} \).
and Eq. 5.7 reduces for ratios $\gamma < 0.3$, i.e., $J_0 \approx 1$ (cf. Eq. 5.3), to

$$Z = \gamma^3 \int_0^\infty \frac{k - \nu}{k + \nu} k^2 e^{-2k} dk = \gamma^3 Z'$$  \hspace{1cm} (5.9)$$

It is obvious from Eqs. 5.8 and 5.9 that the transformed secondary field $Z'$ depends only on the ratio $\delta = h/p$. Therefore, two curves are sufficient to describe the transformed secondary field $Z'$ for all combinations of half-space resistivity, system frequency, and sensor altitude. Fig. 5.1 shows the components of the complex function $Z'$ as in-phase $R'$ (real part of $Z'$) and quadrature $Q'$ (imaginary part of $Z'$) as well as amplitude $A' = (R'^2 + Q'^2)^{1/2} = A/\gamma^3$ and ratio $\varepsilon = Q'/R' = Q/R$.

![Fig. 5.1. In-phase R' and quadrature Q' (left) and Amplitude A' and phase ratio $\varepsilon$ (right) of the transformed secondary field $Z'$ for arbitrary half-space resistivity, system frequency, and sensor altitude on a log-log scale](image)

If the ratio $\gamma = r/h$ is greater than 0.3, the integral of Eq. 5.7 including $J_0$ has to be evaluated in order to calculate $Z'$, because the argument $(k\gamma)$ of the Bessel function $J_0$ gains a non-negligible importance. Instead of a single pair of curves, two arrays have to be calculated for the various ratios of $\gamma$ (Fluche et al. 1998).

In case of lateral resistivity changes, a numerical calculation of the secondary field is necessary, e.g. Avdeev et al. (1998), Newmann and Alumbaugh (1995), Stuntebeck (2003), or Xiong and Tripp (1995). Analytical solutions only exist for simple geometries, e.g. a conducting sphere or cylinder. An overview is given by Ward and Hohmann (1988).
Inversion of secondary field values

Generally, the (measured) secondary field data \((R, Q)\) is inverted into resistivity using two principal models: the homogeneous half-space model and the layered half-space model (Fig. 5.2). While the homogeneous half-space inversion uses single frequency data, i.e., the inversion is done individually for each of the frequencies used, the multi-layer (or one-dimensional, 1D) inversion is able to take the data of all frequencies available into account.

The resulting parameter of the half-space inversion is the apparent resistivity (or half-space resistivity) \(\rho_a\), which is the inverse of the apparent conductivity. Due to the skin-effect (high frequency currents are flowing on top of a perfect conductor) the plane-wave apparent skin depth

\[
\delta = \frac{2\rho_a}{\omega \mu_0} \approx 503.3 \sqrt{\frac{\rho_a}{f}}
\]  

(5.10)

of the AEM fields increase with decreasing frequency \(f\) and increasing half-space resistivity \(\rho_a\). Therefore, the apparent resistivities derived from high-frequency AEM data describe the shallower parts of the conducting subsurface and the low-frequency ones the deeper parts.

Any of the field components shown in Fig. 5.1 can be used to calculate the apparent resistivity, if the distance between the AEM sensor and the top of the half-space is known. Unfortunately, the dependency of the secondary field on the half-space resistivity is highly non-linear. Thus, the inversion is not straightforward and the apparent resistivities have to be derived by the use of look-up tables, curve fitting or iterative inversion procedures. A single-component inversion, however, has the disadvantage that the inphase component \(R'\) is very small for low frequencies and high resistivities, i.e., \(\delta < 0.1\) (low induction mode), and nearly constant for high frequencies and low resistivities, i.e., \(\delta > 10\) (strong induction mode), and the quadrature component \(Q'\) is not unique (cf. Fig. 5.1). On the other hand, the utilisation of the amplitude \(A'\) or the ratio \(\varepsilon\) yields to inaccurate half-space resistivities for a half-space with overburden (Siemon 2001). Using the sensor altitude as an input parameter for the inversion, a further strong disadvantage occurs: The sensor altitude, which is measured in field surveys by laser or radar altimeters, may be affected by trees or buildings (see Fig. 5.3). Therefore, the calculation of the apparent resistivity from both the amplitude \(A'\) and the ratio \(\varepsilon\) (or inphase \(R'\) and quadrature \(Q'\)) is not only more accurate but yields also the apparent distance \(D_a\) of the AEM system to the top of the conducting half-space, i.e., a pseudo-layer half-space model is taken into account (Fraser 1978). Siemon (2001) has published a very fast and accurate approach for the calculation of both
half-space parameters ($\rho_a$ and $D_a$) from the curves $A'(\delta)$ and $\varepsilon(\delta)$ (cf. Fig. 5.1) approximated by polynomials.

**Fig. 5.2.** AEM data inversion: I) homogeneous half-space model, II) layered half-space model (for a five-frequency data set)

The apparent distance $D_a$ can be greater or smaller than the measured sensor altitude $h$: The difference of both, which is called apparent depth

$$d_a = D_a - h$$  \hspace{1cm} (5.11)

is positive in case of a resistive cover (including air) as it is the case in Figs. 5.3 and 5.4; otherwise a conductive cover exists above a more resistive substratum. If the apparent distance equals the measured sensor altitude, no resistivity change with depth is supposed. This is the only case where all approaches for calculating the apparent resistivity will yield identical results.

From the apparent resistivity and the apparent depth, a third parameter can be derived: The centroid depth (Fig. 5.3)

$$z^* = d_a + \rho_a/2$$  \hspace{1cm} (5.12)
is a measure of the mean penetration of the induced underground currents (Siemon 2001). Each set of half-space parameters is obtained individually for each of the AEM frequencies at each of the measured sites.

**Fig. 5.3.** Graphical display of apparent distance $D_a$, apparent depth $d_a$, centroid depth $z^*$, sensor altitude $h$, sensor elevation $h_{GPS}$ and topographic elevation topo. In case of buildings or trees, the sensor altitude $h_v$, and thus, the apparent depth $d_{av}$ and the centroid depth $z^*_v$ differ from their correct values, but their associated elevations in m a.m.s.l. (metre above mean sea level) are correct.

The model parameters of the 1D inversion (cf. Fig. 5.2) are the resistivities $\rho$ and thicknesses $t$ of the model layers (the thickness of the underlying half-space is assumed to be infinite). There are several procedures for the inversion of AEM data available (e.g. Qian et al. 1997, Fluche and Sengpiel 1997, Beard and Nyquist 1998, Ahl 2003, or Huang and Fraser 2003) which are often adapted from algorithms developed for ground EM data. We use a Marquardt inversion procedure which requires a starting model. Due to the huge number of inversion models to be calculated in an airborne survey, it is not feasible to optimise the starting model at each of the models sites. Therefore, an automatic generation of starting models is necessary, e.g. on the basis of apparent resistivity vs. centroid depth values (Fig. 5.4). The standard model contains as many layers as frequencies used plus a highly resistive cover layer. The layer resistivities are set equal to the apparent resistivities, the layer boundaries are chosen as the logarith-
mic mean of each two neighbouring centroid depth values. The thickness of the top layer is derived from the apparent depth $d_a$ of the highest frequency used for the inversion. If this apparent depth value is less than a given minimum depth value, the minimum depth value (e.g. 1 m) is used.

![Fig. 5.4. Sketch on the derivation of the starting model from five-frequency half-space parameters apparent resistivity $\rho_a$ vs. centroid depth $z^*$ and the apparent depth $d_a$ of the highest frequency](image)

The inversion procedure is stopped when a given threshold is reached. This threshold is defined as the differential fit of the modelled data to the measured HEM data. We normally use a 10% threshold; i.e., the inversion stops when the enhancement of the fit is less than 10%. This standard starting model requires more model parameters than data values are available, i.e., an underdetermined equation system has to be solved. This is feasible because the inversion procedure is constraint and searches the smoothest model fitting the data (Sengpiel and Siemon 2000).

3D AEM inversion procedures (e.g. Liu et al. 1991, Sasaki 2001) are not only scarce but also very intensive in computing time and storage. In practice, 3D modelling is only necessary when strong lateral resistivity changes occur on a local scale. Due to the limited footprint of AEM systems (Beamish 2003), it is mostly adequate to invert the AEM data using an 1D inversion procedure (Sengpiel and Siemon 1998).

### 5.1.3 Systems

Far more helicopter than fixed-wing systems are used in airborne frequency-domain surveys. A summary of AEM systems is listed in Table 5.1.
Helicopter systems use a towed rigid-boom AEM system. On a fixed-wing system, the transmitter and receiver coils are mounted at the wing tips. While the helicopter-borne electromagnetic (HEM) system is towed at a sufficiently long distance below the helicopter, the fixed-wing electromagnetic (FEM) system has to cope with interactions with the aircraft.

Table 5.1. Helicopter (HEM) and fixed-wing (FEM) frequency-domain systems (#F: no. of frequencies f, #×: no. of coils, coil orientation: hor: horizontal, vert: vertical, copl: coplanar, coax: coaxial, r: coil separation)

<table>
<thead>
<tr>
<th>Method</th>
<th>System</th>
<th>Properties</th>
<th>www link</th>
</tr>
</thead>
<tbody>
<tr>
<td>HEM</td>
<td>AWI</td>
<td>2F, 2× hor copl, r = 2.1 / 2.8 m, f = 3.7 / 112 kHz</td>
<td>awi-bremerhaven.de</td>
</tr>
<tr>
<td>HEM</td>
<td>Impulse</td>
<td>6F, 3× hor copl / 3× vert coax, r = 6.5 m, f = 870 Hz – 23 kHz</td>
<td>aeroquestsurveys.com</td>
</tr>
<tr>
<td>HEM</td>
<td>Humming-bird</td>
<td>5F, 3× hor copl / 2× vert coax, r = 4.7 m, f = 880 Hz - 35 kHz</td>
<td>geotechairborne-surveys.com</td>
</tr>
<tr>
<td>HEM</td>
<td>GEM2-A</td>
<td>6F, 1× hor copl, r = 5.1 m, f = 300 Hz - 48 kHz</td>
<td>geophex.com</td>
</tr>
<tr>
<td>HEM</td>
<td>Dighem</td>
<td>5F, 5× hor copl, r = 6.3 / 8 m, f = 400 Hz - 56 kHz</td>
<td>fugroairborne.com</td>
</tr>
<tr>
<td>HEM</td>
<td>RESOLVE</td>
<td>6F, 5× hor copl / 1× vert coax, r = 7.9-9 m, f = 380 Hz - 101 kHz</td>
<td>fugroairborne.com</td>
</tr>
<tr>
<td>HEM</td>
<td>Dighem-BGR</td>
<td>5F, 5× hor copl, r = 6.7 m, f = 385 Hz - 195 kHz</td>
<td>bgr.de</td>
</tr>
<tr>
<td>FEM</td>
<td>GSF-95</td>
<td>2F, 2× vert copl, S = 21.4 m, F = 3.1 / 14.4 kHz</td>
<td>gsf.fi\aerogeo</td>
</tr>
<tr>
<td>FEM</td>
<td>Hawk</td>
<td>1 - 10F, 1× hor copl / 1× vert copl, r = wing span, f = 200 Hz – 12.5 / 25 kHz</td>
<td>geotechairborne-surveys.com</td>
</tr>
</tbody>
</table>

Commonly, several geophysical methods are used simultaneously in an airborne survey. A typical helicopter-borne geophysical system operated by the German Federal Institute for Geosciences and Natural Resources (BGR) is shown in Fig. 5.5. It includes geophysical sensors that collect
five-frequency electromagnetic, magnetic, and gamma-ray spectrometry data, as well as altimeters and positioning systems.

Fig. 5.5. BGR’s helicopter-borne geophysical system: Electromagnetic, magnetic, GPS and laser altimeter sensors are housed by a “bird”, a cigar-shaped 9 m long tube, which is kept at about 30–40 m above ground level. The gamma-ray spectrometer, additional altimeters and the navigation system are installed into the helicopter. The base station records the time varying parameters diurnal magnetic variations and air pressure history. The sampling rate is 10 Hz except for the spectrometer (1 Hz), which provides sampling distances of about 4 m and 40 m, respectively, taking an average flight velocity of 140 km/h into account.
5.1.4 Data Processing

The receiver coils of a frequency-domain AEM system measure the induced voltages of the secondary magnetic fields at specific frequencies. These voltages have to be converted to relative values with respect to the primary fields at the receivers. These conversions are done using special calibration coils which produce definite signals in the measured AEM data. Based on these well-known calibration signals, the measured secondary field voltages are transformed into ppm (parts per million) values. The unit ppm is adequate to display the tiny secondary AEM fields.

Due to the induction process within the conducting earth, phase shifts occur between primary and secondary fields, i.e., the secondary field is a complex quantity having inphase and quadrature components. As a consequence, a phase adjustment has to take place at the beginning of each survey flight, e.g., using the well-defined signals of the calibration coils.

Calibration and phase adjustment are best performed above a highly resistive subsurface or in the air at high flight altitude. From the formulae for calculating the secondary fields, it is evident that these fields are strongly dependant on the sensor altitude, even for a homogeneous subsurface (cf. Eq. 5.9). Flight altitudes of several hundred meters (e.g., 350 m for a common HEM system) are sufficient to drop down the signal of the secondary field below the system noise level. This effect is not only used for accurate calibrations and phase adjustments but also for determining the zero levels of the AEM data. Remaining signals due to insufficiently bucked-out primary fields, coupling effects with the aircraft, or (thermal) system drift are generally detected at high flight altitude several times during a survey flight.

These basic values measured at reference points are used to shift the AEM data with respects to their zero levels (cf. Valleau 2000). This procedure enables the elimination of a long-term, quasi-linear drift; short-term variations caused by e.g. varying air temperatures due to alternating sensor elevations, however, cannot be determined successfully by this procedure.

Therefore, additional reference points – also along the profiles at normal survey flight altitude – may be determined where the secondary fields are small but not negligible. At these locations, the estimated half-space parameters are used to calculate the expected secondary field values which then serve as local reference levels.

A standard airborne survey consists of a number of parallel profile lines covering the entire survey area and several tie-lines which should be flown perpendicular to the profile lines. At the cross-over points, the AEM data from profile and tie-lines are compared and statistically analysed to correct the AEM line data for remaining zero-level errors. Due to the altitude de-
pendency of the AEM data, the tie-line levelling and further AEM levelling procedures (e.g. Huang and Fraser 1999) are normally applied to half-space parameters (apparent resistivity and apparent depth) which are less affected by the changes of the sensor altitude.

Noise from external sources (e.g. from radio transmitters or power lines) should be eliminated from the AEM data by appropriate filtering or interpolation procedures. Induction effects from buildings and other electrical installations or effects from strongly magnetised underground sources should not be erased from the data during the first step of data processing. These effects appear particularly on a low-frequency resistivity map as conductive or resistive features outlining the locations of man-made installations or strongly magnetised sources, respectively. If necessary, these effects can be cancelled out after a thorough inspection, and the data may be interpolated in case of small data gaps and smoothly varying resistivities.

5.1.5 Presentation

The AEM results are generally presented as maps and vertical resistivity sections (VRS). The maps display, e.g. the half-space parameters apparent resistivity and centroid depth or the resistivities derived from the 1D inversion results for certain model layers or at several depth or elevation levels. An example is shown in Fig. 5.6.

The VRS - also based on the 1D inversion results - are produced along the survey lines. The vertical sections are constructed by placing the resistivity models for every sounding point along a survey profile next to each other using the topographic relief as base line in metre above mean sea level (m a.m.s.l.). The altitude of the HEM bird, the misfit of the inversion, and the HEM data are plotted above the resistivity models (Fig. 5.7).

Example

The area between the coastal towns of Cuxhaven and Bremerhaven, Germany, was surveyed in 2000/2001 using HEM (Siemon et al. 2004) in order to map glacial meltwater channels and saltwater intrusions from the estuaries of the Elbe and Weser rivers. The HEM data set serves as a base to revise and upgrade the groundwater model of the entire survey area (Fig. 5.6) and to assess the groundwater potential of the area in view of the increasing water consumption used by industry and tourism.

The morphology of the survey area is described by wet marshlands just above sea level and smooth sand ridges called “Geest” with elevations of ten to thirty metres above sea level. Marshlands run more or less parallel to
the banks of the Elbe and Weser rivers and along the shore line, whereas the Geest ridge forms the centre part of the area under consideration. Minor settlements are spread all over the survey area with major centres in the north and south (cities of Cuxhaven and Bremerhaven). Roads, railway tracks, power lines, fences and many other infrastructural networks are existent. All these are potential sources of noise resulting in reduced quality of the airborne geophysical signals.

Figure 5.6 depicts the lateral variation of apparent resistivity for the frequency of 1830 Hz. The central area between the marshlands in the west and north-east is characterised by elevated resistivity values ($\rho_a > 65 \, \Omega\text{m}$, light grey) associated with freshwater saturated sands. A few linear, north-east to north-west striking conductive features ($\rho_a = 10-65 \, \Omega\text{m}$, grey) can be ascribed to channels incised by glacial meltwater during Pleistocene glacial regression epochs. The channels were subsequently filled with coarse sands and gravels in the bottom and mostly silt and clayish materials in the upper parts forming ideal freshwater aquifers. The total thickness of the channel fillings is approximately 300 m (Kuster and Meyer 1979) covered by several ten metres thick clay layers, referred to as cover or lid clays which are covered by about 30-60 m thick Pleistocene sediments.

The conductive response of the lid clays indicates the existence of buried meltwater channels. The lowest apparent resistivity values $\rho_a < 2 \, \Omega\text{m}$, very dark grey) occur offshore clearly due to the presence of seawater and in the north-east and west $\rho_a = 2-7 \, \Omega\text{m}$, dark grey) where saltwater intrudes several kilometres inland. Noteworthy, elevated resistivity values $\rho_a = 20-50 \, \Omega\text{m}$, grey) are mapped offshore close to the north-western tip of the mainland. A freshwater aquifer extending seawards across the shore line is the hydrogeological source of this resistivity high.

The resistivity section in Fig. 5.7 shows the results of the 1D inversion of four-frequency HEM data collected along the survey line 219.1. This line runs WNW–ESE and is approximately half way between the cities of Cuxhaven and Bremerhaven (cf. Fig 5.6). From the top to the bottom, Fig. 5.7 displays the inphase (R) and quadrature (Q) HEM data, the resistivity section, and the relative misfit of the 1D inversion. The line above the resistivity models indicates the sensor (bird) elevation in metres above mean sea level which is the difference between the barometric altimeter record and the effective cable length of about 40 m.

The ground elevation is obtained as the difference between the sensor elevation and the radar/laser altitude of the bird above ground level. As radar/laser signals are frequently distorted by the tree canopy, the ground elevation may be too high in wooded areas (cf. Fig. 5.3).
Fig. 5.6. Example for an apparent resistivity map of a coastal area in NW Germany. The survey was flown in two parts: the dashed line marks the boundary of the Cuxhaven (2000) and Bremerhaven (2001) survey areas. The solid line shows the location of profile 219.1 (see Fig. 5.7)
The resistivity models are plotted downward from the ground elevation line. Therefore, the top layer of the inversion models includes both trees and resistive cover layers. The clear drop in resistivity from the highly resistive top layer to the medium resistive underlying model layer is obviously caused by the groundwater table. Between profile-km 10 and 14, a pair of meltwater channels (buried valleys) has been identified due to the conductive properties of the lid clays (cf. B2.1). It can also be noted that the penetration depth of the HEM system is reduced within these meltwater channels because of the lid clays, i.e., the induced eddy currents may not have reached the bottom of the buried valley. The occurrence and the lateral extent of the buried valleys have reliably been located by the HEM system (see Fig. 5.6) as was confirmed by comparison with existing boreholes (Siemon and Binot 2002) and results of ground geophysics (Gabriel et al. 2003).

Fig. 5.7. Example for a vertical resistivity section (VRS): From the top to the bottom, the inphase and out-of-phase (quadrature) values of the HEM data (in ppm) for the four frequencies \(f_1 - f_4\), the 1D resistivity models (in \(\Omega\)m) using the topographic relief as base line in meters above mean sea level (m a.m.s.l.), and the misfit of the inversion \(q\) (in %) are displayed. The altitude of the HEM bird is plotted above the resistivity models.
5.1.6 Discussion and Recommendations

Airborne electromagnetics is a very useful method for surveying large areas in order to support hydrogeological investigations. Due to the dependency of the geophysical parameter electrical conductivity from both the mineralization of the groundwater and the clay content, information about water quality and aquifer characteristics, respectively, can be derived from AEM data. The results, however, are sometimes ambiguous: A clayey aquitard in a freshwater environment and a brackish, sandy aquifer are associated with similar conductivities. As a consequence, additional information, e.g. from drillings, are required for a solid hydrogeological interpretation of the AEM data.

Frequency-domain electromagnetic measurements are suitable for high-resolution surveys as long as the targets are seated not deeper than 100 m. For deeper targets ground-based or airborne time-domain measurements are more suitable. Helicopter-borne multi-frequency systems are widely used in groundwater explorations due to their high-resolving properties and their applicability even in rough terrain. Fixed-wing systems are applicable for reconnaissance surveys in a flat terrain because these systems outrange helicopter-borne systems and they are less expensive, but they are less flexible and have less-resolving properties. Frequency-domain systems using natural (e.g. audio-frequency magnetic (AFMAG) systems) sources are not very practicable for detailed groundwater surveys.

5.2 Ground based techniques

Reinhard Kirsch

Electromagnetic methods enable the fast mapping of highly conducting underground structures. Comparable to the electromagnetic airborne methods, soundings are also possible when several frequencies are used for the measurements. The theory of induction and the use of secondary fields for resistivity determination are discussed in details in the previous section. The application of electromagnetic methods for the detection of fracture zone aquifers is illustrated in Chap. 13.

5.2.1 Slingram and ground conductivity meters

Originally, transmitter and receiver coils with horizontal orientations are used for exploration purposes leading to the name Horizontal Loop Elec-
tromagnetic (HLEM) systems or the Swedish name slingram. Both coils are operated at a fixed distance during the survey. Depth penetration can be controlled by the frequency or the coil separation, both are often coupled. High frequency systems specially designed for mapping of shallow underground structures (e.g. GEONICS EM 31 and EM 38) often combine transmitter and receiver coil in one instrument, so one-man operation is possible. Operation is also possible with vertical coils. Some instruments, e.g. MAXMIN (APEX) are not restricted to coplanar coil orientations. Unlike geoelectrical methods, no galvanic contact to the ground is required; therefore measurements on sealed terrains are possible.

Normally, the magnetic component of the superposition of primary and secondary field is measured. The measured field is split into the inphase and outphase (=quadrature, 90° phase shift) component with respect to the primary field. Both components are recorded. A typical response of the inphase and quadrature signal to a steeply dipping and highly conducting fracture zone is shown in Fig. 5.8. Examples for the use of HLEM measurements for fracture zone detection are given in Chap. 13, Figs. 13.11, 13.12, 13.16, 13.18, and 13.19.

![Diagram](image)

**Fig. 5.8.** left: Slingram response over a highly conductive fracture zone. The offset at the quadrature response is due to the conductivity of overburden (after McNeill 1990), right: influence of a good conductive layer on the Slingram response (after Grissemann and Ludwig 1986). Examples showing the influence of dipping layers and conducting sheets are shown in Chap. 13, Figs. 13.15 and 13.17.
The depth of penetration is characterised by the skin-depth, where the amplitude of the primary electromagnetic wave is reduced by the factor $e$ compared to the transmitted wave. Skin depth $\delta$ depends on signal frequency $\omega$, ground conductivity $\sigma$ in mS/m, and magnetic permeability $\mu$:

$$\delta = \sqrt{\frac{2}{\sigma \cdot \omega \cdot \mu}}$$ \hspace{1cm} (5.13)

which leads to a rule of thumb for the assessment of skin depth:

$$\delta = 503 \cdot \frac{\rho}{\sqrt{f}}$$ \hspace{1cm} (5.14)

skin depth $\delta$ in m, specific resistivity $\rho$ of the ground in $\Omega$m, and frequency $f$ in s$^{-1}$.

If the coil space is smaller than the skin depth, the term operation at low induction numbers is used (McNeill 1990). Then, the inphase component of the signal becomes very small and the quadrature component is directly related to the conductivity of the ground. This enables a direct reading of the mean ground conductivity. Examples for ground conductivity measurements for the mapping of a waste dump are shown in Chap. 17 Figs. 17.1 and 17.2.

However, for an interpretation of the so obtained conductivities it should be kept in mind that the induced current flow is mainly horizontal. If electrical anisotropy due to alternating layering of good and poor conductors (e.g. clay, sand) occur, then current flow is mainly in the good conducting layers resulting in high measured conductivities (Seidel 1997). Therefore, ground resistivity determination by electromagnetic methods can result in lower resistivities than obtained by VES measurements.

For a layered ground the measured conductivity is a weighted mean of the conductivities of the layers in which currents were induced. This is similar, but not identical to the apparent resistivity obtained by VES measurements. Data inversion to obtain the conductivity depth structure is possible, if measurements were done with different frequencies or coil orientations. For EM-34 and EM-31 instruments, a simple method to calculate the layered ground response is given by McNeill (1980).

From a sensitivity function $R(Z)$ the contribution of the layers below the depth $Z$ (normalised by the coil separation) can be calculated (Fig. 5.9). As an example, for a coil separation of 10 m the depth range below 10 m ($Z = 1$) contributes to the measured apparent conductivity by 42%, whereas the contribution of the depth range below 20 m ($Z = 2$) is only 25%.
The sensitivity function $R(Z)$ can be calculated after McNeill (1980) for horizontal loops by:

$$R(Z) = \frac{1}{\sqrt{4 \cdot Z^2 + 1}}$$ (5.15)

and for vertical loops by:

$$R(Z) = \sqrt{4 \cdot Z^2 + 1 - 2 \cdot Z}$$ (5.16)

For a layered ground consisting of $n$ layers with conductivities $\sigma_i$ and normalized depths $Z_i$ to the bottom of layer $i$, the measured apparent conductivity $\sigma_a$ is given by:

$$\rho_a = \sigma_1 \cdot [1 - R(Z_1)] + \sigma_2 \cdot [R(Z_2) - R(Z_1)] + \ldots + \sigma_n \cdot R(Z_{n-1})$$ (5.17)

This can be used for the planning of field surveys, if the depth ranges of the target layers are known. Also a rough data interpretation is possible. However, it must be kept in mind that the resolution of these measurements for highly resistive layers is poor.
5.2.2 VLF, VLF-R, and RMT

**VLF (very low frequency)**

VLF techniques use the signal of very long wavelength radio transmitters. These transmitters are world-wide distributed, their signal is mainly used for marine navigation and communication. The transmitter frequencies are in the range of 15 – 25 kHz with transmitter powers in the range of 100 – 1000 kW.

The transmitted electromagnetic wave consists of electric field components $E_x$ and $E_z$ and a magnetic component $H_y$ (Fig. 5.10). The electric component $E_x$ leads to current flow in the ground, especially in elongated conductive underground structures like fracture zones stretching roughly in the direction to the transmitter. These currents induce an additional magnetic field component $H_Z$ which results in a tilted magnetic field vector in the vicinity of the good conductor with a crossover at the center of the good conductor. The tilt angle is measured by a radio receiver with the antenna tilted to the direction of maximum received signal.

![Fig. 5.10. Left: field components from a remote VLF transmitter and electric current flow in the ground, right: electric field component along the profile indicated in the left picture (after McNeill 1990)](image)

Tilt angle data can be plotted as profiles where the crossovers indicate narrow conducting zones. Examples for the use of VLF-techniques for fracture zone detection are given in Chap. 13, Figs. 13.20, 13.21, 13.22, and 23. However, if the data density is sufficiently large to produce a contour map, the crossover should be converted to peak values to give a clear
picture on the map. This can be done by applying an averaging filter to the data, e.g. the Fraser filter which converts a sequence of equidistant data $M_1,...,M_4$ into:

$$F_1 = (M_3 + M_4) - (M_1 - M_2)$$  \hspace{1cm} (5.18)

$F_1$ is plotted at the center of this data sequence. The application of Fraser-filtering is demonstrated by Hutchinson and Barta (2002) and in Chap. 13 (Fig. 13.22).

**VLF-resistivity and radio magnetotelluric (RMT)**

From the ratio of the horizontal components of the electric and the magnetic field the apparent resistivity of the ground can be determined by (McNeill 1980):

$$\rho_a = \frac{1}{\mu \omega} \left( \frac{|E|}{|H|} \right)^2$$  \hspace{1cm} (5.19)

$E =$ Amplitude of the horizontal electric field, V/m  
$H =$ Amplitude of the horizontal magnetic field, A/m  
$\mu =$ magnetic permeability of free space, $\mu = 4\pi \times 10^{-7}$ H/m  
$\omega = 2\pi f$, $f =$ frequency, Hz

$H$ is measured by an antenna and $E$ is measured by an electrode pair orientated towards the radio transmitter (Fig. 5.11). The electrode spacing is in the range of 1 – 5 m. Radio transmitters are chosen in the VLF-frequency range (VLF-R) or in the radio-frequency range 10 – 300 kHz (RMT). As the penetration depth of the radio signal depends on the frequency, a VLF-R or RMT multi-frequency data set can be inverted to the resistivity depth structure. For inversion, the apparent resistivity $\rho_a(f)$ and the phase-shift $\Phi(f)$ (between $E$ and $H$) is used. Examples for RMT-measurements on waste dumps are given by Hördt et al. (2000).
Fig. 5.11. Schematic principle of radiomagnetotelluric sounding (after Bosch and Gurk 2000)

5.3 References


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