THE USE OF CVES TO MAP THE SUBSURFACE SALINITY DISTRIBUTION: A CASE STUDY FROM THE NETHERLANDS

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INTRODUCTION

Geo-electrical methods have been applied extensively in the past for hydrological research in coastal areas (e.g. Van Dam, 1976; Boekelman, 1991). Mapping the resistivity distribution, both laterally and with depth, enables hydrogeologists to (1) develop a reliable hydrogeological schematization of the subsoil, (2) delineate geological structures that influence groundwater flow patterns and (3) track spatial variations in groundwater quality.

Classically, mapping of the subsoil resistivity by geo-electrical methods was achieved by performing point (1D) measurements (Vertical Electrical Sounding, VES) or by measuring the lateral variation of some depth-averaged resistivity (profiling). Recent developments in the field of geo-electrical research and new measurement techniques have enabled hydrogeologists to determine the subsurface resistivity in much more detail. The advent of modern equipment and advanced interpretation algorithms has made it possible to map the resistivity distribution in 2D or 3D (Loke, 2000). This technique was applied to map the subsurface salinity distribution in the Nieuwe Keverdijksche Polder (NKP) area in the Netherlands. Additional measurements were performed with electromagnetics, both in the frequency domain (FDEM) and in the time domain (TDEM), VES (Schlumberger configuration) and borehole logging (short normal and long normal). The purpose of this study was (1) to compare the feasibility of each of these methods for mapping subsurface salinity variations and (2) to delineate the distribution of fresh and brackish water in the NKP area.

PHYSICAL SETTING

The Nieuwe Keverdijksche Polder area is located in the western part of the Netherlands near the towns of Weesp and Muiderberg (figure 1). It is bordered on the eastern side by a lake (Naardermeer) and on the western side by the river Vecht. The northern- and southern borders are formed by two canals that separate the NKP from the polders Zuidpolder east of Muiden and Heintjesrak- en Broekerpolder respectively. The surface of the NKP area is about 1.2 m below sea level.

The subsoil generally consists of a peat layer with a thickness of about 3 meters overlying a sequence of fluvial sands up to 200 meters thick that constitutes the main aquifer. Locally, sandy ridges protrude the peat layer. These ridges are a remnant of the former Pleistocene morphology that was created by glacial processes during the Saalian glacial stage and successively covered by younger sediments. The relatively simple stratigraphy simplifies the interpretation of the observed resistivity patterns as resistivity variations are expected to reflect mainly changes in salinity.
On a regional scale, groundwater flow takes place from relatively elevated areas such as the river Vecht, the lake Naardermeer and the ice-pushed ridge of Muiderberg towards low-lying polder areas such as the NKP. In the NKP, groundwater levels are controlled by a system of ditches and canals. Groundwater discharge is by vertical upward flow. Although the aquifers are currently recharged by fresh water, the discharging groundwater in the NKP is mainly brackish. Chloride concentrations measured in shallow (< 20 m) observation wells are as high as 125 mmol/l. The presence of this brackish water is generally attributed to former inundations during the Holocene (Appelo and Geirnaert, 1991).

**METHODS**

One of the aims of the present work was to compare the feasibility of different geophysical techniques to map spatial variations of groundwater salinity. Both electromagnetic (FDEM, TDEM) and electrical (VES, CVES, borehole logs) survey methods were used.

**Electromagnetic methods**

EM methods employ an alternating current (in a transmitter loop, $T_x$) to generate a magnetic field (primary field) that induces very small currents in the earth. These currents generate a secondary magnetic field that is recorded in a receiver coil ($R_x$).

The Geonics EM-34 was used in the present study for FDEM measurements. It uses the ratio of the secondary to the primary magnetic field to calculate the terrain conductivity (McNeill, 1980). The transmitter and receiver loops are located a certain distance apart and may be oriented either vertically or horizontally. The intercoil spacing used in the present study was 40 meters (operating frequency $f = 400$ Hz) with the loops oriented vertically (horizontal dipole) resulting in an exploration depth of 30 meters (McNeill, 1980).
Contrary to EM-34 that uses a continuous magnetic field, TDEM methods measure the decay of the secondary field after switching off the current in the transmitter loop (Goldman et al., 1989). The decaying magnetic field is used to calculate the resistivity as a function of depth. In the NKP area, nanoTEM (manufactured by Zonge Engineering and Research Organization, Inc.) was used with very small antenna loops (Tx 5·5 m², Rx 1·1 m²). This set-up was chosen to enable a very shallow depth of exploration, which was a requirement considering the depth of the interface between fresh and brackish water (< 10 meters in certain areas). However, it appeared that this configuration suffered from a very low signal to noise ratio for which reason it did not produce meaningful results (Goldman and Helwig, 2001).

**Electrical methods**

Direct current (DC) methods measure the potential between an electrode pair (MN) that is the result of a current that is inserted into the ground through two other electrodes (A and B). The ratio of the potential to the current is proportional to the resistivity of the subsurface. Several configurations of the electrode pairs MN and AB are possible. In the present study, VES measurements were carried out using the Schlumberger array in which the electrode array is centered over the point of interest (figure 2). The potential electrodes are located in between the current electrodes that are placed an increasing distance apart. This produces a relation between the apparent earth resistivity ($\rho_a$) and electrode spacing (AB/2), which is used to calculate the "true" resistivity distribution with depth. Measurements were taken using the Terrameter SAS 4000 manufactured by ABEM Instrument AB and interpreted with SchlumBG for Windows (Hemker and Post, 2001).

![Figure 2 Electrode configurations used for geo-electrical measurements.](image)

CVES (continuous VES) is a geo-electrical method whereby a large number of electrodes is placed into the ground along a transect (for 2D measurements) or in a grid (for 3D measurements). Measurements are taken using all possible combinations of electrode pairs, resulting in a data set of apparent resistivities at the so-called pseudo-depth at different locations. Using an inversion algorithm, the data set is then transformed into an earth model that describes the resistivity of the subsoil as a function of x, y and z. As with the VES measurements, the Schlumberger array was used for the CVES measurements. Additionally, the dipole-dipole array was used for the latter. In this array the electrode pairs AB and MN form two adjacent dipoles (figure 2). The dipole-dipole array has a good horizontal resolution, but is far less sensitive to changes in vertical resistivity than the Schlumberger array (Loke, 2000). These two arrays were chosen to test the sensitivity of the results to the array type. Measurements were performed with the SuperSting R8 IP multi-electrode system from Advanced Geosciences Inc. A total of 72 electrodes were used that were each 5 meters apart, resulting in a maximum electrode spacing of 355 meters. The program RES2DINV (Advanced Geosciences Inc., 2000) was used to invert the apparent resistivity values to a resistivity model section. Details on the processing and inversion of the field data are found in Bloem and Ooteman (2002).

The setup that was used for the borehole resistivity logs consisted of three electrodes (1 current, 2 potential) in the borehole and two electrodes at the surface (figure 2). Both the long normal (LN, distance AM = 100 cm) and a short normal (SN, distance AM = 20 cm) were measured in two boreholes to a depth of 5 meters below the surface. Additionally, the resistivity of individual soil samples from the borehole were measured.
RESULTS

The EM-34 was used to obtain an overview of the spatial resistivity distribution of the upper 30 meters of the subsurface. The transects and results that were surveyed are shown in figure 3a. The highest conductivities (lowest resistivities) are found in a north-south running area in the eastern part of the NKP area. Based on these results three transects were selected along which the CVES measurements were carried out (figure 3a). For all three transects the dipole-dipole configuration was used, for transects 1 and 2 the Schlumberger configuration was used as well. The dipole-dipole data show quite some variability in the measured apparent resistivities, whereas the Schlumberger measurements show far less scatter (figure 4). This clearly also influenced the quality of the inversion model as can be seen from the RMS error, which can be very high (up to 34 %) for inverted dipole-dipole pseudosections. Furthermore, the dipole-dipole measurements show increasing resistivities with depth, as opposed to the Schlumberger measurements.

Figure 3 (a) Conductivities (in mS/m) obtained from EM-34 measurements (horizontal dipole, intercoil spacing = 40 m, operating frequency = 400 Hz). Dots indicate data points. White lines represent the CVES transects. (b) Thickness of the confining peat deposits (in meters).

From the inversion models some general trends can be distinguished. The resistivities of the top part of the section are typically between 20 and 80 Ωm, except in the easternmost part of the section where values as low as 2 Ωm are found at the surface (although even lower values are calculated for the inversion model, 2 Ωm is considered the minimum based on the measured apparent resistivities). In the lowest part of the sections resistivities are between 3 and 10 Ωm. The very high values that sometimes are calculated for the dipole-dipole measurements, are artefacts of the method and should be discarded.

The transition from high to low resistivities (taken here as the 10 Ωm contour) is found at the westernmost end of transect 1 at a depth of about 25 meters below the surface and then changes over a distance of some 300 meters to approximately 15 meters below the surface. This depth of 15 meters is roughly maintained throughout transect 2. The depth of the 10 Ωm contour in transect 3 is more or less constant at a depth of 10 meters until it ascends quite abruptly towards the surface at a distance of about 600 meters from the westernmost end of the transect.
Another interesting feature is the shallow (within 10 meters below the surface) zone with high resistivities between distances of 300 to 500 meters. Although the inversion model simulates very high (> 200 $\Omega m$) resistivities, the measurements from borehole 2 show that 130 $\Omega m$ is about the maximum resistivity (figure 5). Using Archie's law, the formation factor ($F$) of the sandy aquifer sediment was estimated from the electrical conductivity of the natural groundwater in the borehole ($\approx 150 \mu S/cm$) to be $F = 2$.

**Figure 4** Inversion results of the CVES measurements. (a) Transect 1, dipole-dipole configuration. (b) Transect 2, dipole-dipole configuration. (c) Transect 3, dipole-dipole configuration. (d) Transect 1, Schlumberger configuration. (e) Transect 2, Schlumberger configuration.

**DISCUSSION**

An important notion from the results presented above is the difference in performance of the two configurations used in the CVES survey. Firstly, there is the much smoother apparent resistivity distribution of the Schlumberger array as compared to the dipole-dipole array. Secondly, the dipole-dipole measurements suggest that the apparent resistivity increases at greater depths, which is contradicted by conventional VES measurements and water quality data in observation wells. This artefact is most likely caused by the fact that at increasing electrode spacing, the potential is in the same order of magnitude as the earth potential, resulting in calculated resistivities that are too high. Apparently the dipole-dipole configuration, which is relatively insensitive to vertical variations (Loke, 2000), is not suited for this area in which the resistivities are stratified mainly horizontally (i.e. vertical resistivity changes dominate). In the present study, the dipole-dipole measurements are still valuable in the sense that they indicate the trends of the resistivity distribution in the upper part of the subsurface.

In transects 1 and 2 the confining peat layer can be recognized as a thin layer at the top of the inversion model with resistivities between 10 to 40 $\Omega m$. Values above 40 $\Omega m$ below this layer represent the parts of the aquifer that are filled with fresh water. The lowest resistivities at the bottom part of the transects are indicative of brackish water. As these low resistivities are mainly due to the water resistivity, it is impossible to discern between lithologies, such as in the eastern part of transect 3. Here, brackish water is found at the surface (e.g. borehole 1, figure 5), causing low resistivities at shallow depths. The area of high resistivities near the surface between 300 and 500 meters in transect 3 is caused partly by lithology (the peat layer pinches out against a sandy ridge in this zone) and, more importantly, by the low salinity of the water ($EC \approx 150 \mu S/cm$).
Figure 5 Results of the borehole resistivity logs.

Using the value of F found from the borehole resistivity logs, the spatial distribution of the chloride concentration of the groundwater could be obtained using the relation shown in figure 6. This distribution is shown schematically in figure 7.

Figure 6 Relation between chloride concentration and water resistivity. Data kindly provided by L. Reiniers of the province of Noord-Holland.

The measurements reveal that the salinity of shallow groundwater in the NKP area is distributed spatially in a quite complicated fashion. Redistribution of the relic brackish water has taken place, which is mainly influenced by (1) the variable thickness and heterogeneous composition of the confining peat deposits and (2) variations in elevation of the land surface.

Influence of the confining peat layer

Figure 3b shows the lateral variations in the thickness of the confining peat deposits derived from the many shallow drillings that have been carried out (data kindly provided by J. Dijkmans, TNO-NITG). The figure shows a striking resemblance between the peat thickness and the conductivity recorded by the EM-34 measurements: in areas where the peat deposits are the thinnest, the conductivity is highest. At the same time, local farmers report on high seepage rates in this area. An estimate of the vertical groundwater flow velocity based on (density-corrected) piezometric levels in observation well 25H-0094 (figure 3a) using a conductivity of 10 m/d, yields a rate of up to several meters per year. Clearly, the brackish water is transported towards the surface by upward groundwater flow, which tends to concentrate in areas with the lowest hydraulic resistance, i.e. where the peat layer is thinnest. Thus, by controlling groundwater flow patterns, the confining peat deposits determine the distribution of fresh and brackish groundwater.
Influence of the land surface elevation

Differences in the land surface elevation in the NKP area are generally very small (< 30 cm). Locally, however, the sandy ridges that protrude the confining peat layer extend about 50 cm to 1 meter above the surrounding surface. Their higher surface elevation compared to the surrounding terrain, together with a relatively high permeability, makes these areas a likely candidate for groundwater recharge. Indeed, the apparent resistivities measured with CVES, VES and in the borehole show higher values, indicating that the brackish water is pushed down by infiltrating fresh water.

The fresh water lens formed this way had a depth of about 5 meters (figure 4) during the period of the survey (summer). It might well be that there is a seasonal trend in which the size of the fresh water lens varies depending on recharge conditions.

It is striking that the same sand ridges that form infiltration areas occur in the area where the strongest discharge takes place. There, they provide preferential pathways for the discharging groundwater due to their relatively high permeability. Apparently, the magnitude of the upward flow is so high that infiltration of meteoric water is prevented.

CONCLUSIONS

Modern geophysical equipment and advanced inversion algorithms allow hydrogeologists nowadays to obtain a detailed overview of subsurface resistivity variations. Application of 2D geo-electrical measurements in the NKP area has shown that the distribution of fresh and brackish groundwater could be mapped to a detail that would not have been possible with classical geo-electrical models or observation wells. The application of these methods to coastal areas looks very promising. Caution should be applied however, since the available tools (equipment, software) require both knowledge and experience to be applied successfully. In this study for example, the selection of an inappropriate electrode configuration (dipole-dipole as opposed to Schlumberger) led to results that were less usable.

The detailed mapping of the groundwater salinity patterns in the NKP area showed that the heterogeneity of the confining layer and subtle variations in land surface elevations control the groundwater flow patterns and thus the distribution of fresh and brackish water. Depending on local conditions, major variations can occur in salinity over distances as small as several meters.
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REFERENCES


