Journal of Applied Geophysics 84 (2012) 77–85

Contents lists available at SciVerse ScienceDirect

Journal of Applied Geophysics

journal homepage: www.elsevier.com/locate/jappgeo

# Equation estimation of porosity and hydraulic conductivity of Ruhrtal aquifer in Germany using near surface geophysics

Sri Niwas [a,*], Muhammed Celik [b]

[a] Department of Earth Sciences, Indian Institute of Technology Roorkee, Roorkee-247667, India
[b] Institute of Geology, Geophysics and Mineralogy, Ruhr University, Bochum, Germany

ARTICLE INFO

*Article history:*
Received 20 April 2012
Accepted 4 June 2012
Available online 13 June 2012

*Keywords:*
Porosity
Hydraulic conductivity
Vertical electrical sounding
Hydrogeophysics

ABSTRACT

Vertical Electrical Sounding (VES) data are used to estimate the porosity and the hydraulic conductivity of the Ruhrtal aquifer in western Germany in addition to mapping the aquifer therein. Two theoretical methods based respectively on Kozeny–Archie equations and Ohm's–Darcy's laws are used for better confidence level in estimated values from VES data. Estimated hydraulic conductivity values from VES data and those determined from pumping test are strongly correlated.



## 1. Introduction

Near surface geophysical exploration methods such as geoelectrics and seismic refraction have attracted interest of engineers and hydrogeologists for exploring and delimiting aquifers with success for several decades. The basic equations for geoelectrical exploration are developed assuming that the medium is porous, the matrix is generally an insulator and electrical currents flows through the water present in the pore spaces. The aquifer's electrical resistivity is mainly influenced by porosity and fluid resistivity in the pores. The geoelectrical data recorded on the surface therefore contain useful information about the aquifer which can be interpreted by experienced geophysicists for hydrogeological studies.

In an ideal setting, the physical condition controlling the electric current flow i.e. tortuosity and porosity also controls the flow of the water in a porous media. Using this analogy a large number of empirical equations are reported in the literature that correlate electrical resistivity to hydraulic conductivity. These empirical equations imply a somewhat puzzling message as both direct and inverse relationships between hydraulic conductivity and electrical resistivity are reported (Mazac et al., 1985, 1990; Purvance and Andricevic, 2000). The empirical equations have generally very limited applicability in comparison to more broad relationships implied by rigorous theoretical derivation which, however, hold in idealized conditions. For example, experimental and analytical work on clean sandstone support a relationship between electrical resistivity and hydraulic permeability, which is a property of a porous rock regarding any fluid flow through the pore spaces (not just water), and having dimension of area (Croft, 1971). Starting from Archie's (1942, 1950) equation for electrical resistivity ($\rho$):

$$\rho = a\rho_{\mathrm{w}}\varphi^{-m} \tag{1}$$

and using Kozeny (1953) equation for intrinsic permeability ($k_{\mathrm{f}}$):

$$k_{\mathrm{f}} = \frac{d^2}{180}\frac{\varphi^3}{(1-\varphi)^2} \tag{2}$$

where

| | |
|---|---|
| $a$ | Electrical tortuosity parameter(Lynch, 1964) [−] |
| $\rho_{\mathrm{w}}$ | Resistivity of groundwater [Ohm m] |
| $\varphi$ | Porosity of aquifer [−] |
| $m$ | Cementation factor[−], see Table 1 for values |
| $d$ | Grain size [m], |

we see that, $k_f$, can effectively be computed using surface geoelectrical measurements.

Archie (1942) observed that "from a study of many group of data, $m$ has been found to range between 1.8 and 2.0 for consolidated sandstones. For clean unconsolidated sands packed in the laboratory, the value of $m$ appears to be about 1.3." Furthermore for sandstone, a range of values for $m$ used in log analysis are given in Table 1 (Doveton, 1986).

Heigold et al. (1979) combined Darcy's equation and Archie's equation to relate intrinsic permeability and porosity with limited success. As pointed out by Lima and Niwas (2000) hydraulic conductivity, $K$ (in m/s) is a more meaningful parameter which depends on

* Corresponding author. Tel.: +91 1332 285570; fax: +91 1332 273560.
*E-mail address:* srsnpfes@iitr.ernet.in (S. Niwas).

0926-9851/$ – see front matter © 2012 Elsevier B.V. All rights reserved.
doi:10.1016/j.jappgeo.2012.06.001

**Table 1**
Values of $m$ (Doveton, 1986).

| Formation type | $m$ value [−] |
|---|---|
| Unconsolidated sand | 1.3 |
| Very slightly cemented sandstone | 1.4–1.5 |
| Slightly cemented sandstone | 1.5–1.7 |
| Moderately cemented sandstone | 1.8–1.9 |
| Highly cemented sandstone | 2.0–2.2 |

both the type of formation and the fluid properties contained in it. To this end Nutting's equation (Hubert, 1940) relates $k_f$ to $K$ as,

$$K = \frac{\delta_w g}{\mu} k_f \tag{3}$$

where

$\delta_w$ Water density [1000 kg/m$^3$]
$g$ Acceleration due to gravity [9.81 m/s$^2$]
$\mu$ Water dynamic viscosity [0.0014 kg/m s].

Finally, works such as that of Rucker (2009), where fully coupled resistivity-flow models are developed, aim to estimate hydraulic properties by combining the physical phenomena of electrical current and water flow through porous media. However, the work still requires constitutive relations, such as Archie's equation, to convert the jointly influencing parameter (i.e. moisture) from one set of physical models to the other. Notwithstanding, the effort should be to obtain a more physically supported quantitative relation instead of an empirical one.

Using more basic Ohm's law of current flow and Darcy's law for horizontal fluid flow in a medium Niwas and Singhal (1981, 1985) derived two analytical equations,

$$T = \alpha S; \quad \alpha = K\rho \tag{4}$$

and

$$T = \beta R \quad \beta = K/\rho \tag{5}$$

respectively representing inverse and direct relationship between electrical resistivity and hydraulic conductivity,where

$T$ Hydraulic transmissivity [m$^2$/s]
$R = d\rho$ Transverse resistance [Ohm]
$S = d/\rho$ Longitudinal conductance [S],
$d$ [m] thickness of the aquifer;
$\alpha, \beta$ constant of proportionality.

Analyzing these two equations further, Niwas et al. (2011) successfully solved the contradiction between direct and inverse relationship of electrical resistivity and hydraulic conductivity. Their analysis included data from Krauthausen test site in Germany fitted to analytical geoelectrical modeling results. They concluded that Eq. (4) exists in case of highly resistive basement ($S$- dominant aquifer where electrical currents tends to flow horizontally) and Eq. (5) exists in case of highly conducting basement ($R$-dominant aquifer where electrical currents tend to flow vertically).

In this study, geoelectric data are acquired to estimate the hydraulic conductivity values of the Ruhr aquifer. It is proposed to use both theoretical approaches discussed above for the estimation of hydraulic conductivity so that merits and limitations of each method are highlighted. Method I uses Eqs. (1), (2) and (3) to calculate the porosity and the hydraulic conductivity from resistivity data whereas Method II uses Eqs. (4) or (5) depending on electrical nature of the basement. For validation of the two methods, hydraulic conductivity and porosity are acquired through the installation of wells and conducting pumping tests.

Estimated porosity and the hydraulic conductivity from geoelectric measurements are the desired values attributed at a point and nearby. This study establishes a correlation between the hydraulic conductivity values obtained by pumping test and from surface geoelectrical measurements. Additionally, the cross plotting of the geoelectric estimated hydraulic parameters from Method-I and Method-II enhances the reliability of the results.

## 2. Site description

### 2.1. Study area

The study was carried out in Ruhrtal located south of the Ruhr University in Bochum, Germany. The study area is on the bank of the south-west portion of Kemnader lake and covers an area of 199016 m$^2$ (Fig. 1). Our work was conducted just upstream from a hydroelectric dam. The region was chosen for the hydrophysical study because the geological and hydrogeological characteristics of the area were known. For example there were several existing wells along with well logs providing lithological description that were advantageous to the present study. Additionally for this study, satellite images taken from Google earth (Google, 2011) were georeferenced in ArcGis to UTM coordinate system.

### 2.2. Geology of the study area

During the building of the Kemnader dam the surface had been filled with backfill (Holocene). The fill, composed of silt and gravel, had a thickness from 1 m up to 6 m. There was no filling where the river Ruhr begins to flow. The fill was underlain by a 0.2 m to 3.5 m silt layer of Holocene age. The Pleistocene aged Niederterrasse formation consisted of a gravel-sand bed that filled under the silt layer. It was the aquifer layer with a high hydraulic conductivity and the thickness of the aquifer ranges from 4 m to 6.5 m. The aquifer layer was underlain by the faulted bed rock composed of claystone, siltstone and sandstone beds. The depth of the Carboniferous (Carbon.) aged bedrock ranges from 6.5 m up to 12.5 m (Hahne and Schmidt, 1982).

### 2.3. Hydrogeology

Over the past several decades, pumping tests have been performed in the Ruhrtal aquifer. Table 3 lists the values for six available wells (W4, W5, W6, W7, W8 and W9). The pumping test on the observation well W8 was carried out during the field course “Markierungsversuch” at the Ruhr University Bochum (Stemke et al., 2009). The pumping tests of the rest of the observation wells were performed by Ruhr University, Bochum in September 1979 (Obermann and Diegelmann, 1980). We show the hydraulic conductivity results of the pumping tests using three different methods: Thiem (stationary), Jacob (non-stationary) and Jacob recovery (non-stationary). For final analysis of the tests, we used an average hydraulic conductivity obtained from three methods.

## 3. Geoelectrical measurements

The aim of the study is to explore the gravel-sand aquifer layer to estimate the hydraulic parameters from Vertical Electrical Sounding (VES) measurements. The Schlumberger array was chosen due to its better lateral resolution. For the present work, we used the ABEM Terra-meter SAS 300C with a maximum half-current electrode separation $\left(\overline{AB}/2\right)$ of 250 m. Because of the natural boundaries on the field (trees, fence, way) the maximum separation $\left(\overline{AB}/2\right)$ of some soundings were less than 250 m. However, all spread were sufficient

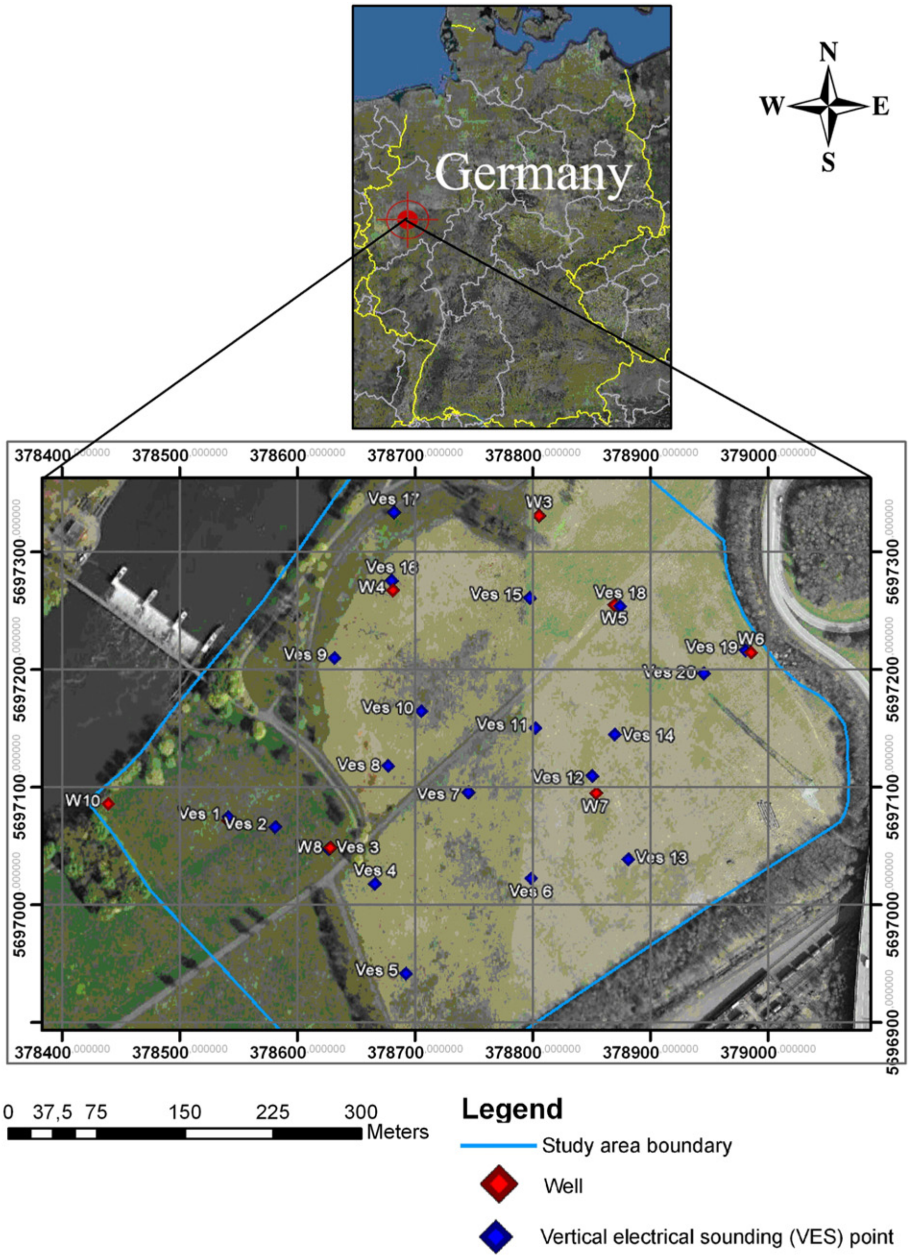


**Fig. 1.** Study area map and location of VES points along with existing wells (UTM coordinate system, 32 N Zone).

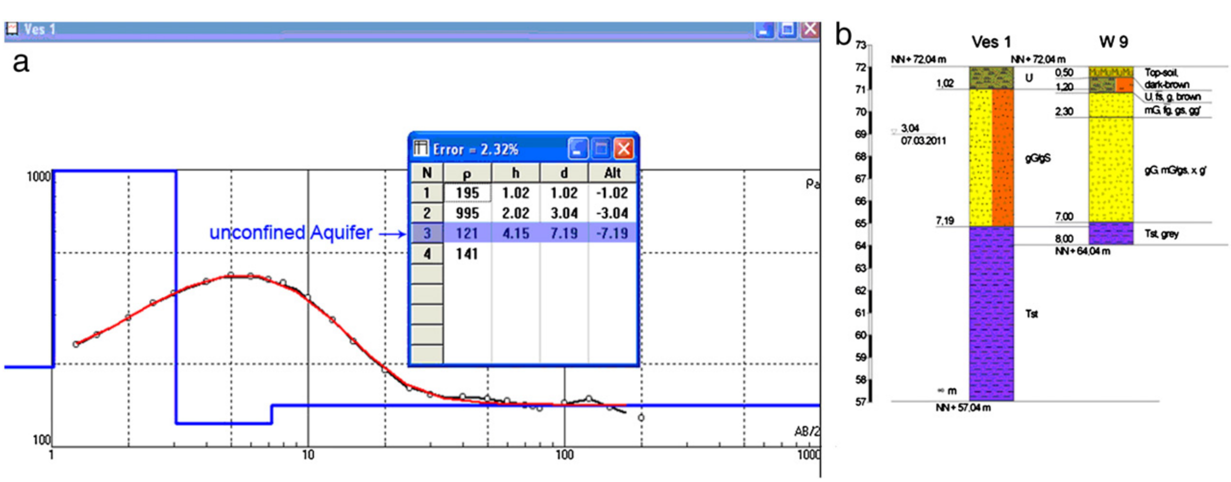


**Fig. 2.** a and b. Interpretation of VES 1 showing correlation between observation well W9 and VES1(soil identification according to DIN 4022).

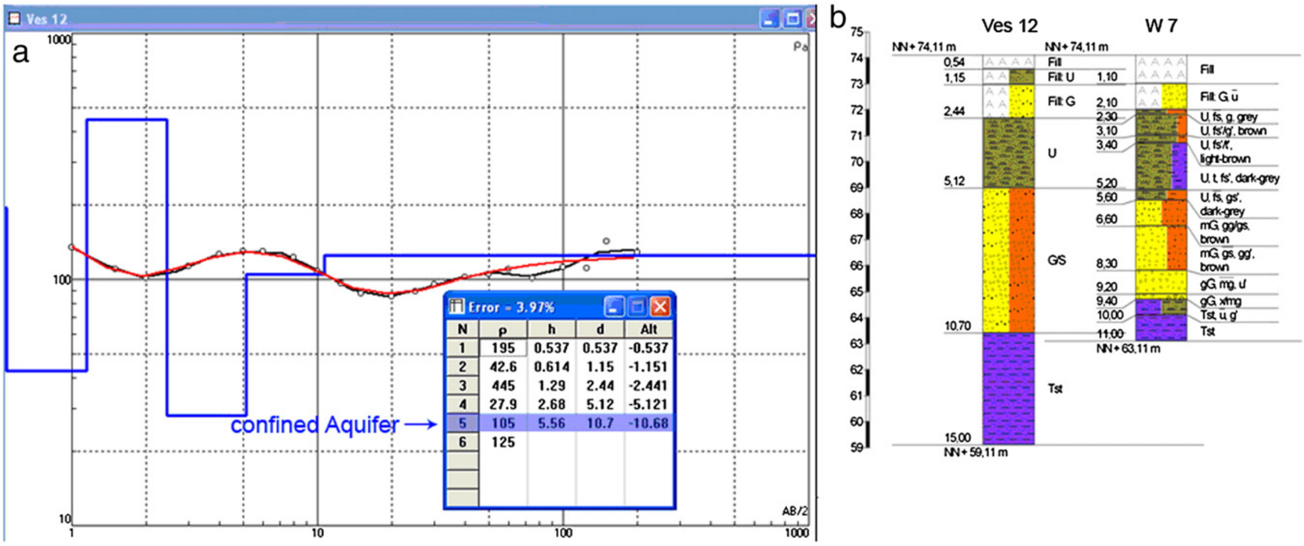


**Fig. 3.** a and b. Interpretation of VES 12 showing correlation between observations well W7 and VES12 (soil identification according to DIN 4022).

for the designed depth in view of the concept of 'Depth of Investigation (DI)' determined by the position of the current electrodes (A, B) and the measuring electrodes (M, N) and not by the current penetration or current distribution alone (Roy and Apparao, 1971). Roy and Apparao (1971) defined DI as that depth at which a thin horizontal layer of earth contributes the maximum amount to the total measured signal at the earth surface. Using basic law of physics Roy and Apparao (1971) mathematically derived ' Depth of Investigation Characteristic' (DIC) as contribution of a thin layer of thickness, dz, buried in a homogeneous earth of resistivity, $\rho$, at a depth , $z$, energized by a current of strength, $I$, for a four electrode array (AMNB) given by,

$$\mathrm{DIC} = \frac{\rho I}{4\pi^2}\mathrm{d}z\left[\frac{8\pi z}{(\mathrm{AM}^2+4z^2)^{3/2}} - \frac{8\pi z}{(\mathrm{BM}^2+4z^2)^{3/2}} - \frac{8\pi z}{(AN^2+4z^2)^{3/2}} + \frac{8\pi z}{(BN^2+4z^2)^{3/2}}\right]. \quad (6)$$

By taking AM = BN = 0.45AB and MN = 0.1 AB for Schlumberger configuration and normalizing Eq. (6) by the total response of homogeneous earth for this particular array given by, $\rho I/2.475\pi\mathrm{AB}$, Eq. (6) is reduced as Normalized DIC (NDIC) as,

$$\mathrm{NDIC} = \left[\frac{1}{\left\{(0.45\mathrm{AB})^2+4z^2\right\}^{3/2}} - \frac{1}{\left\{(0.55\mathrm{AB})^2+4z^2\right\}^{3/2}}\right]. \quad (7)$$

The maximum of NDIC is computed by Roy and Apparao (1971) as DI lies at 0.125 AB; $\overline{\mathrm{AB}} = 8^*$ DI. However, Kirsch (2006) gave a different estimate as $\overline{AB} \sim 5^*$DI. Special attention was also paid to the measurements to ensure they were conducted in a straight line. The VES electrode positions were installed uniformly along the line and measurements were taken near each of the existing wells for correlation. Altogether 20 VES were recorded and Fig. 1 shows the layout of both wells and VES points. A number of bad values caused reiteration of the acquisition process to improve the quality of data.

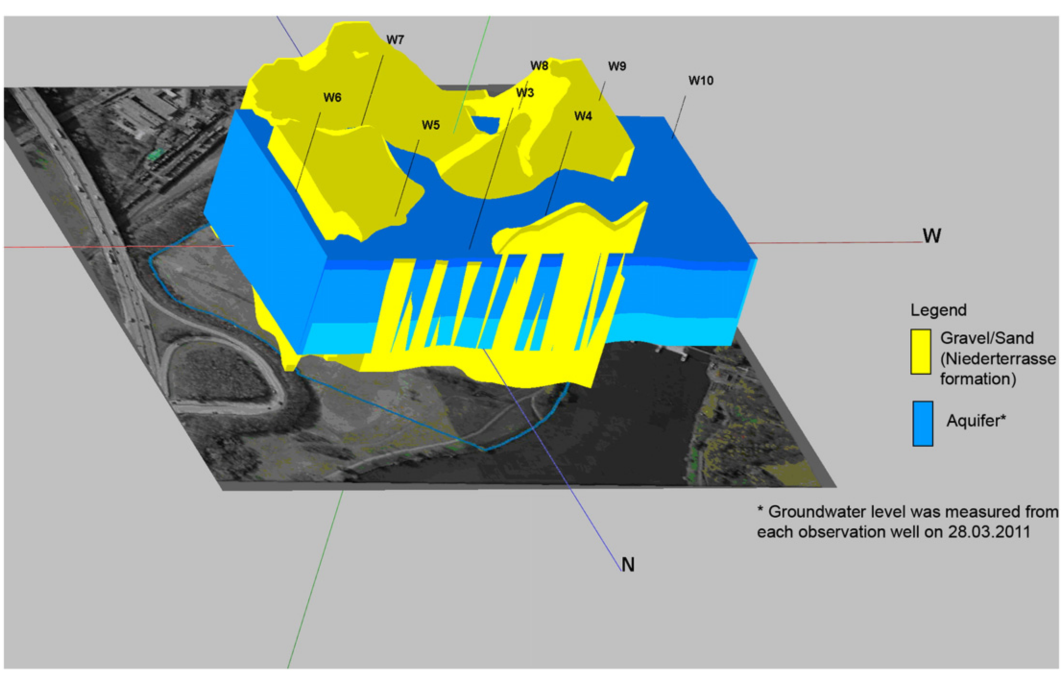


**Fig. 4.** Thickness of the gravel/sand layer determined from VES data.

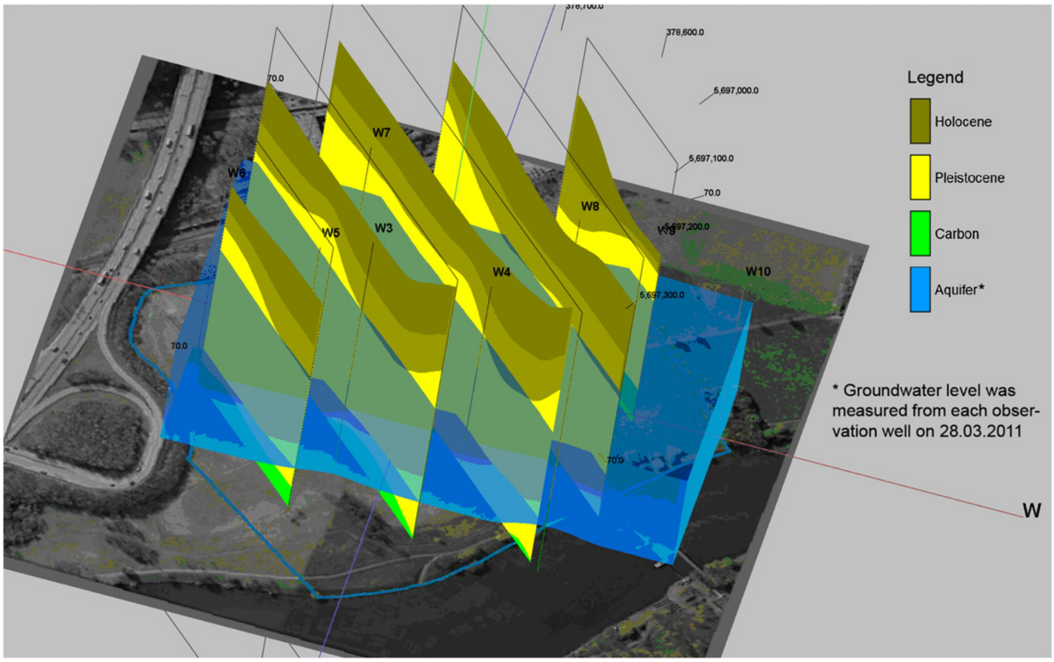


**Fig. 5.** Combination of saturated gravel/sand layer profile sections determined from VES data.

The computer interpretation software "IPI2win" was used for data inversion. Figs. 2 and 3 show examples of output from the software, including a graphical display of data, estimated geoelectrical properties, and calculated fitting error between modeled and measured data. The VES lines were near wells W9 and W7 and Figs. 2b and 3b show the comparison between the litho-logs and geoelectrical parameters. For the inversion, we took care to constrain the inversion to the well data to help with the correlation. Finally, we took all of the georeferenced resistivity values and rendered a three-dimensional depiction of the hydrogeology. Fig. 4 shows the location of the aquifer and water table relative to the Niederterrasse formation. Fig. 5 shows slices through the domain to highlight all geological layers.

### 3.1. Method I

The inverted resistivity data were used to estimate the porosity from Eq. (1) using literature values of respective parameters for an unconsolidated gravel-sand. Unconsolidated sediments are characterized by relatively low values of $m$ (between 1.1 and 1.3) and parameter $a \approx 1$ (Keller, 1989; Schön, 2004), and we chose to use $a = 1$ and $m = 1.3$. The exponent $m$ correlates with the sphericity $P$ of the sediment grains and follows an equation, $m = 2.9–1.8P$ (Atkins and Smith, 1961; Jackson et al., 1978; Schön, 2004). The computed porosity was used to estimate the hydraulic conductivity of the aquifer layer using Eqs. (2) and (3). Average grain size used in Eq. (2) was the $d_{50}$ value which was obtained through sieve analysis. For the analysis, two soil samples from the Niederterrase formation were taken at different depth (6.7–7 $m$ and 2.5–2.8 $m$). The average grain size value was estimated as 0.01 $m$ in the laboratory (Obermann and Diegelmann, 1980).

Table 2 shows the results of the calculated porosity and hydraulic conductivity of the aquifer layer through Method-I. The groundwater resistivity map was made by interpolating between the groundwater resistivity values of observation well as measured by Stemke (2010). The groundwater resistivity at VES points was determined using the interpolated groundwater resistivity ($\rho_w$) map. Accordingly the VES 9 results in Table 2 be shown in red rectangle. The fitting error of this point has a low value but the average grain size of 0.01 m is unrealistically large for this point. Therefore VES 9 is excluded from the interpretation and estimation process. The relationship between water resistivity and interpreted aquifer resistivity is shown

**Table 2**
Estimated porosity and hydraulic conductivity (Method I).

| Points | $\rho_w$ [Ohm m] | $\rho$ [Ohm m] | $\varphi$ [−] | $K$ [m/s] × $10^{-2}$ |
|---|---|---|---|---|
| Ves 1 | 17 | 121 | 0.22 | 6.9 |
| Ves 2 | 17 | 241 | 0.13 | 1.1 |
| Ves 3 | 16 | 155 | 0.17 | 3.0 |
| Ves 4 | 14 | 87 | 0.25 | 10.0 |
| Ves 5 | 12 | 69 | 0.26 | 13.0 |
| Ves 6 | 11 | 157 | 0.13 | 1.1 |
| Ves 7 | 14 | 256 | 0.11 | 0.6 |
| Ves 8 | 17 | 142 | 0.20 | 4.5 |
| Ves 9 | 29 | 104 | 0.37 | 52.0 |
| Ves 10 | 18 | 165 | 0.18 | 3.5 |
| Ves 11 | 13 | 216 | 0.12 | 0.8 |
| Ves 12 | 10 | 105 | 0.16 | 2.4 |
| Ves 13 | 10 | 65 | 0.24 | 8.9 |
| Ves 14 | 10.5 | 87 | 0.20 | 4.6 |
| Ves 15 | 19 | 235 | 0.14 | 1.6 |
| Ves 16 | 34 | 439 | 0.14 | 1.4 |
| Ves 17 | 27 | 284 | 0.16 | 2.4 |
| Ves 18 | 12 | 118 | 0.17 | 2.9 |
| Ves 19 | 7 | 103 | 0.13 | 1.0 |
| Ves 20 | 9 | 45 | 0.29 | 19.0 |

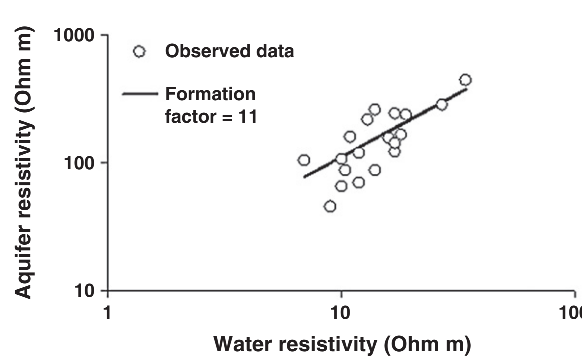


**Fig. 6.** Validation of Archie equation in the study area.

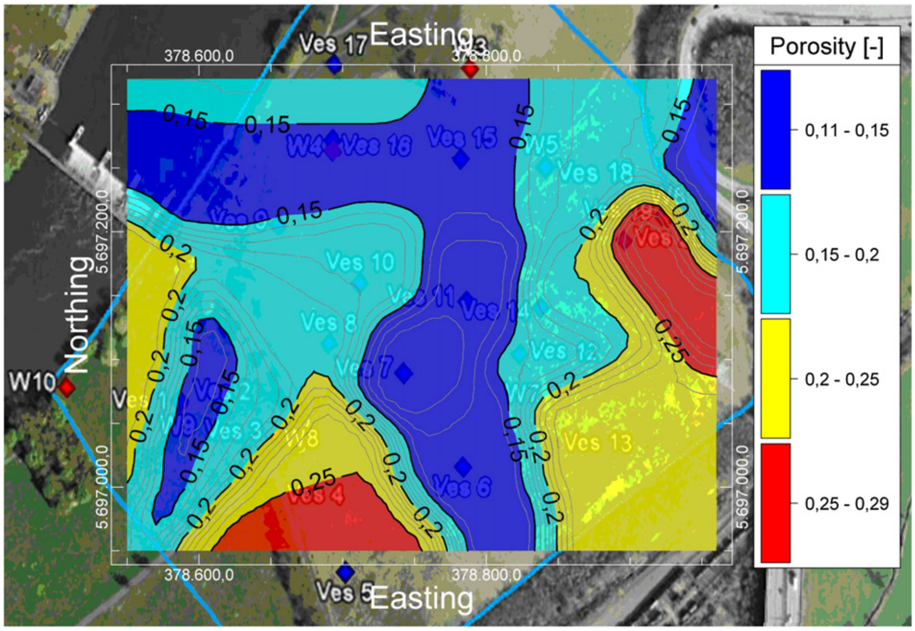


**Fig. 7.** Porosity map of the aquifer layer from resistivity data.

in Fig. 6. We calculated average Percent Error (PE) using a general formula as,

$$\left(\frac{1}{N}\sqrt{\sum_{i=1}^{N}\left(\frac{\mathrm{Obs}_i-\mathrm{Comp}_i}{\mathrm{Obs}_i}\right)^2}\right)\times 100,$$

where $\mathrm{Obs}_i$ is the observed data at $i$th observation point, $\mathrm{Comp}_i$ is the equation computed value of the same data at the same point and being $N$ the number of observation points. Based on the reasonable PE value (≈10.5%), the methodology qualifies for porosity estimation because the resistivity is a product of electrical current flow through the pore space and not along the grain surface. Fig. 6 also shows the formation factor, calculated by the ratio of

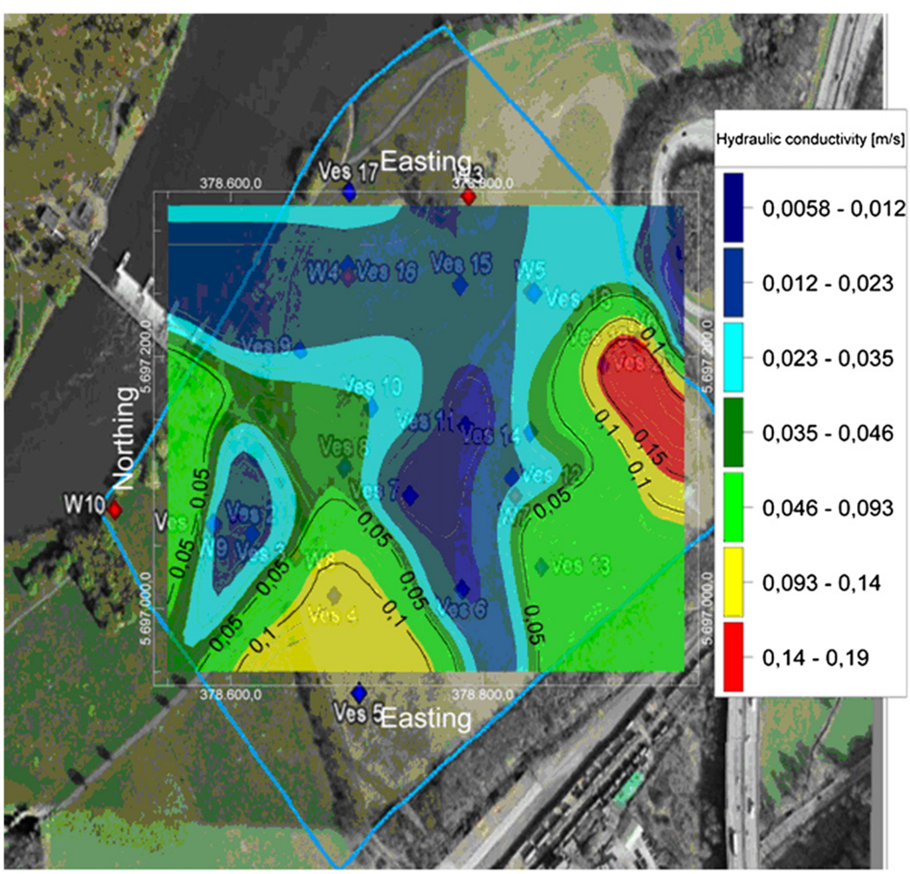


**Fig. 8.** The hydraulic conductivity map of Ruhrtal aquifer.

**Table 3**
Results of pumping test of each observation well and the hydraulic conductivity from resistivity data (all values in $\times 10^{-2}$).

| Well | Stationary (Thiem) $[m/s] \times 10^{-2}$ | non-stationary (Jacob) $[m/s] \times 10^{-2}$ | non-stationary (Wiederanstieg-Jacob) $[m/s] \times 10^{-2}$ | Average, $[m/s] \times 10^{-2}$ | From Method-I $[m/s] \times 10^{-2}$ |
|---|---|---|---|---|---|
| W4 | 2.4 | 1.3 | 1.8 | 1.83 | 1.44 |
| W5 | 1.4 | 2.5 | 2.2 | 2.03 | 2.91 |
| W6 | 3.5 | 1.5 | 1.9 | 2.30 | 3.40 |
| W7 | 1.3 | 2.2 | 2.9 | 2.13 | 2.45 |
| W8 | 2.8 | – | – | 2.80 | 2.60 |
| W9 | 1.5 | 1.9 | – | 1.85 | 3.30 |
| | 1.9 | 2.1 | – | – | – |

bulk resistivity, $\rho$, and water resistivity, $\rho_w$, the results imply a clean formation.

Fig. 7 shows the porosity map of Kemnader area using the geoelectrical data. It should be clarified that the minimum and maximum porosity values, calculated from the VES analysis, are likely due to the basic principles of averaging and are not the result of any spatial interpolations algorithms or measurement error. In the Schlumberger array, the potential differences are measured between $\overline{MN}$ electrodes and the apparent resistivity is calculated from these potential differences. By convention, the calculated apparent resistivity is attributed to the middle point of MN electrodes, but the 'depth of investigation' is determined by relation between $\overline{AB}$ and DI as demonstrated in Eq. (7) (Kirsch, 2006; Roy and Apparao, 1971). The larger the $\overline{AB}$ electrode spacing, the deeper is the sounding. Based on the information from the depth sounding, both resistivity and layer thickness are determined. However, the resistivity value of each layer is an average value, constructed from all of the smaller scaled heterogeneities within that layer. Thus, the calculated porosity value of an aquifer, using an average resistivity, results in an averaged porosity value. Therefore it is possible to measure, for example on VES 5 a porosity of 0.26 and ~50 m away on VES 6 porosity of 0.14 because the volume over which averaging occurs for each VES point is different.

The hydraulic conductivity was computed through Eqs. (2) and (3) and the final values are provided in Table 2. Fig. 8 shows the hydraulic conductivity map of the study area. In general, the Ruhrtal aquifer in the study area has a high hydraulic conductivity value, but the northern part of the area has a lower hydraulic conductivity than the southern part. When the hydraulic conductivity data from Method I is compared to that obtained from pumping tests (Table 3), we observed a worthy comparison between the independently derived data sets (Fig. 9). In particular, the comparison on well W4 and W7 between the average value of pumping test result and the hydraulic conductivity value from the resistivity data is quite good (overall average PE between estimated values and average pump test values $\approx$ 15%). There is a significant comparison (average PE $\approx$ 8%) between the values using the non-stationary method and the estimated hydraulic conductivity from resistivity data. For this analysis using Method I, we excluded well W9 due to low confidence in data. The reason of the bad correlation between the pumping test result and VES measurements on W9 were likely due to the wrong measured average groundwater resistivity on W9 and the strong variation of the hydraulic conductivity from VES point near the well because of heterogeneity.

**Table 4**
Results of the calculated hydraulic parameters (Method II).

| Points | $d$ [m] | Well | $\rho$ [Ohm m] | K_obs $[m/s] \times 10^{-2}$ | $S = d/\rho$, [mho] | K_com $[m/s] \times 10^{-2}$ | $T = 4\,S$, $[m^2/s] \times 10^{-2}$ |
|---|---|---|---|---|---|---|---|
| Ves 1 | 4.50 | W9 | 121 | 1.85 | 0.037 | 3.3 | 14.9 |
| Ves 2 | 6.54 | – | 241 | – | 0.027 | 1.7 | 10.9 |
| Ves 3 | 3.67 | W8 | 155 | 2.80 | 0.024 | 2.6 | 9.5 |
| Ves 4 | 7.90 | – | 87 | – | 0.091 | 4.6 | 36.3 |
| Ves 5 | 4.58 | – | 69 | – | 0.066 | 5.8 | 26.5 |
| Ves 6 | 6.12 | – | 157 | – | 0.039 | 2.5 | 15.6 |
| Ves 7 | 7.9 | – | 256 | – | 0.031 | 1.6 | 12.3 |
| Ves 8 | 10.0 | – | 142 | – | 0.070 | 2.8 | 28.2 |
| Ves 9 | 9.11 | – | 104 | – | 0.087 | 3.8 | 35.0 |
| Ves 10 | 9.10 | – | 165 | – | 0.053 | 2.4 | 22.1 |
| Ves 11 | 5.45 | – | 216 | – | 0.025 | 1.9 | 10.1 |
| Ves 12 | 5.56 | W7 | 105 | 2.13 | 0.053 | 2.5 | 21.2 |
| Ves 13 | 6.74 | – | 65 | – | 0.100 | 6.1 | 41.5 |
| Ves 14 | 11.3 | – | 87 | – | 0.130 | 4.6 | 52.1 |
| Ves 15 | 5.43 | – | 235 | – | 0.023 | 1.7 | 9.2 |
| Ves 16 | 4.49 | W4 | 439 | 1.83 | 0.010 | 1.4 | 4.1 |
| Ves 17 | 10.2 | – | 284 | – | 0.036 | 1.4 | 14.4 |
| Ves 18 | 5.87 | W5 | 118 | 2.03 | 0.050 | 2.9 | 19.9 |
| Ves 19 | 4.71 | W6 | 103 | 2.30 | 0.046 | 3.4 | 18.3 |
| Ves 20 | 3.20 | – | 45 | – | 0.071 | 8.9 | 28.4 |

## 3.2. Method II

One of the essential requirements of Method II for estimating hydraulic parameter from surface geoelectrical measurement is that the hydraulic conductivity from at least one or more points in the area must be known before hand. The hydraulic conductivity is used to ascertain the value of constant of proportionality. Table 3 provides the hydraulic conductivity values based on pumping test at 6 points giving an average value of $\alpha = 4$ for Eq. (4), as the electrical nature of bedrock is resistive in comparison to aquifer. Table 4 gives the relevant hydraulic parameters computed for Method II, with $K = 4/\rho$ and $T = 4\,S$.

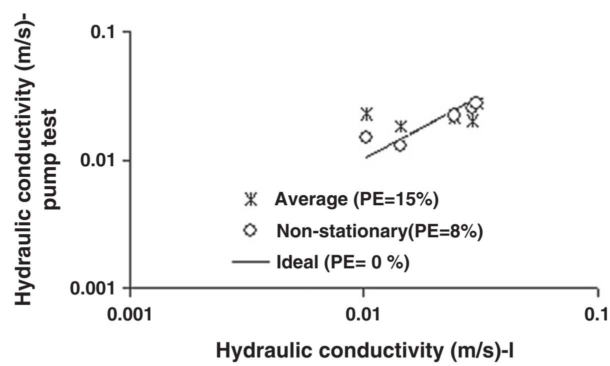


**Fig. 9.** Cross-plot between estimated hydraulic conductivity from resistivity data through Method I and that obtained from pump test.

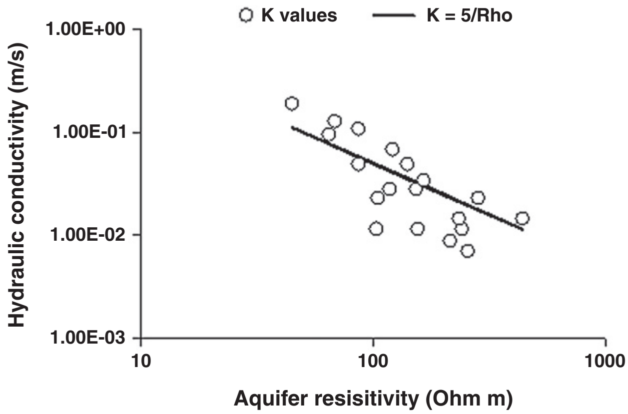


**Fig. 10.** Inverse relationship between aquifer resistivity and hydraulic conductivity for Ruhrtal aquifer.

### 3.3. Comparison between Method I and Method II

The estimated hydraulic conductivity from Method I was compared to that from Method II in order to provide an independent verification of the hydraulic properties between the two methods. First the calculated hydraulic conductivity was plotted as a function of aquifer resistivity and the results are presented in Fig. 10. There is an inverse relationship between the two, which aligned with the development of theoretical equations for a resistive basement (Niwas et al., 2011). From these data, we obtained the average value of $\alpha = 5$ (average PE $\approx$ 11%) while taking into account the 19 valid estimated hydraulic conductivity values. Finally, the two sets of values for hydraulic conductivity using Method I and Method II were cross plotted to observe the relationship between them. Fig. 11 shows the cross plot (actual and ideal) with satisfactory result (average PE $\approx$ 20%) keeping in mind the heterogeneous nature of formations in the area.

## 4. Discussion and conclusion

On the basis of the results of the estimated porosity with minimum of 11% and maximum 29% and the estimated hydraulic conductivity with minimum $7.6 \times 10^{-3}$ m/s and maximum $1.9 \times 10^{-1}$ m/s (in comparison to pump test determined range of $10^{-2}$ m/s) from 20 resistivity points given in Table 2, a realistic porosity-hydraulic conductivity model for Ruhrtal aquifer is obtained that strengthens the validity of determining the hydraulic parameters by using surface geophysical measurements. A realistic model is possible on the basis of reliable values of water resistivity and low mathematical fitting error of each VES measurements. The observed data show an expected structure of the correlation between hydraulic conductivity and porosity. The presence of the groundwater decreases the resistivity of the layer if the background resistivity (unsaturated soil) of the layer is higher than in a saturated situation. Therefore the increasing resistivity for the same material means that the porosity, which is also saturated, decreases. Eventually the increase in resistivity lowers the hydraulic conductivity (Fig. 10).

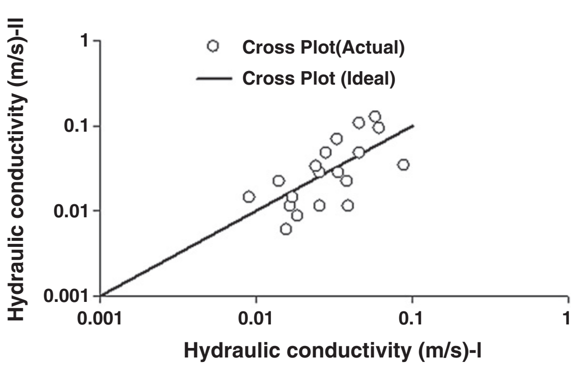


**Fig. 11.** Cross plot between estimated hydraulic conductivity from resistivity data using Method I and Method II.

The theoretical Method I and Method II, of hydraulic conductivity estimation from surface geoelectrical measurements are strongly correlated with themselves and with the available pump test values. As mentioned earlier Method I combines Archie (1942) and Kozeny (1953) through porosity. Thus the success of the method largely depends on porosity estimation using Archie's equation. However, this equation can only effectively be used in case of the formation having no appreciable amount of clay and the values for water resistivity are available from other sources. On the other hand Method II combines more basic laws of Darcy and Ohm through the cross sectional area perpendicular to the flow direction. In this case the hydraulic conductivity at one or more points of VES location and the electrical nature (conducting or resistive) of the bedrock must be available. Method II may be used even when the aquifer material is clayey. In the present area necessary conditions for both the methods are satisfied.

## Acknowledgements

We are thankful to Prof. Stefan Wohnich, Chair Professor of Geology of Ruhr University for providing all help in conducting fieldwork and completion of the project. We are highly indebted to two reviewers, Dr. Dale Rucker and one anonymous, for their constructive review that has highly improved the presentation of the paper.

## References

Archie, G., 1942. The electrical resistivity log as an aid in determining some reservoir characteristics. Petroleum Transactions of the American Institute of Mineralogical and Metallurgical Engineers 146, 54–62.

Archie, G., 1950. Introduction to petrophysics of reservoir rocks. AAPG Bulletin 34, 943–961.

Atkins, E.R., Smith, G.H., 1961. The significance of particle shape in formation resistivity factor porosity relationships. Journal of Petroleum Technology 13, 285–291.

Croft, M.G., 1971. Method of calculating permeability from electric log. U.S. Geological Survey Professional Paper 750, 265–269.

Doveton, J.H., 1986. Log Analysis of Subsurface Geology. Wiley, New York.

Google, 2011. Google EarthRetrieved from , http://www.google.com/earth 2011.

Hahne, C., Schmidt, R., 1982. Die Geologie des Niederrheinisch-Westfälischen Steinkohlengebietes. Verlag Glückauf.

Heigold, P.C., Gilkeson, R.H., Cartwright, K., Reed, P.C., 1979. Aquifer transmissivity from surficial electrical methods. Groundwater 17, 330–345.

Hubert, M.K., 1940. The theory of groundwater motions. Journal of Geology 48, 785–944.

Jackson, P.D., Taylor, D., Smith, P.M., 1978. Standard resistivity–porosity–particle shape relationships for marine sands. Geophysics 43, 1250–1268.

Keller, G.V., 1989. Electrical properties. In: Carmichael, R.S. (Ed.), Practical Handbook of Physical Properties of Rocks and Minerals. CRC Press, Boca Raton.

Kirsch, R., 2006. Groundwater Geophysics: A Tool for Hydrogeology. Springer 3540293833.

Kozeny, J., 1953. Hydraulics. Springer, Vienna.

Lima, O.A.L., Niwas, Sri, 2000. Estimation of hydraulic parameters of shaly sandstone aquifers from geoelectrical measurements. Journal of Hydrology 235, 12–26.

Lynch, E.J., 1964. Formation Evaluation. Harper and Row Publishers Inc., New York.

Mazac, O., Kelly, W.E., Landa, I., 1985. A hydrogeophysical model for relation between electrical and hydraulic properties of aquifers. Journal of Hydrology 79, 1–19.

Mazac, O., Kelly, W.E., Landa, I., Venhodova, D., 1990. Determination of hydraulic conductivities by surface geoelectrical methods. In: Ward, S.H. (Ed.), Geoelectrical and environmental geophysics: Environmental and groundwater applications, v. 2, pp. 125–131.

Niwas, Sri, Singhal, D.C., 1981. Estimation of aquifer transmissivity from Dar-Zarrouk parameters in porous media. Journal of Hydrology 50, 393–399.

Niwas, Sri, Singhal, D.C., 1985. Aquifer transmissivity of porous media from resistivity data. Journal of Hydrology 82, 143–153.

Niwas, Sri, Tezkan, B., Israil, M., 2011. Aquifer hydraulic conductivity estimation from surface geoelectrical measurements for Krauthausen test site, Germany. Hydrogeology Journal 19, 307–315.

Obermann, P., Diegelmann, B., 1980. Hydrogeologische Untersuchung am Kemnader See. Ruhr Universität Bochum, Bochum.

Purvance, D.T., Andricevic, R., 2000. On the electric-hydraulic conductivity correlation in aquifers. Water Resources Research 36, 2905–2913.
Roy, A., Apparao, A., 1971. Depth of investigation in direct current methods. Geophysics 36, 943–959.
Rucker, D.F., 2009. A coupled electrical resistivity-infiltration model for wetting front evaluation. Vadose Zone Journal 8, 383–388.
Schön, J.H., 2004. Physical Properties of Rocks. Elsevier, Amsterdam.
Stemke, M., 2010. *Labor und Feldversuche zur Desorptionscharakteristik von Fluoreszenzfarbstoffen im Talgrundwasserleiter der Ruhr*. Ruhr University, Bochum, Germany.
Stemke, M., Wiesner, B., Celik, M., 2009. Auswertung des Markierungsversuches am Kemnader See. Ruhr University, Bochum, Germany.