ESTIMATION OF POROSITY AND HYDRAULIC CONDUCTIVITY OF SHALLOW QUATERNARY ALLUVIAL AQUIFER IN YENAGOA, SOUTHERN NIGERIA, USING GEOELECTRICAL MEASUREMENTS

Okiongbo, K. S.*[1], and Soronnadi-Ononiwu, G. C.[2]
[1]Department of Physics, Geophysics Research Group, Niger Delta University, Wilberforce Island, Bayelsa State, Nigeria
[2]Department of Geology, Niger Delta University, Wilberforce Island, Bayelsa State, Nigeria
*Correspondence Email address: okenlani@yahoo.com Phone: +2348037682197
(Received: 21st April, 2015; Accepted: 18th June, 2015)

ABSTRACT

In this study, two theoretical methods based respectively on Archie-Kozeny equations and Ohm's-Darcy's laws were used to determine porosity and hydraulic conductivity of shallow aquifer in Yenagoa, Southern Nigeria. Fourteen Vertical Electrical Soundings (VES) using the Schlumberger configuration were carried out within the Quaternary Alluvial deposit with maximum electrode spacing (AB/2) of 200 m. The calculated apparent resistivity data were inverted using the Interpex 1×1D computer software. The results indicated that the porosity (ϕ) and hydraulic conductivity values in the aquifer varied from 0.08 – 0.29 and 1.14×10^{-4} - 87.4×10^{-4}m/s respectively. Correlation of estimated hydraulic conductivity values from the VES data and those conventionally determined by pump test showed a fairly good correlation.

Keywords: Hydraulic Conductivity, Porosity, Formation Factor, Geoelectric Sounding, Yenagoa, Niger Delta Basin

INTRODUCTION

In order to evolve a pragmatic and scientific planning for the management of groundwater resources in an area, one needs to quantify the characteristic hydrogeologic parameters. Estimation of aquifer hydraulic characteristics involves assessing primarily the physical properties controlling groundwater flow and transport. This includes the determination of transmissivity (T), storativity (S), porosity (ϕ) and hydraulic conductivity (K). Knowledge of the water transmitting properties e.g hydraulic conductivity (K) and its variations in porous media such as unconsolidated sediments is of vital importance in various aspects of geologic and geotechnical investigation and management of groundwater resources as it underpins any understanding of fluid flow in sedimentary basins (Hermanrud, 1993). In many applications, such as the modelling of fluid flow in compacting sedimentary basins (Hermanrud, 1993; Dewhurst *et al.*, 1999), evaluation of basin slope stability (Dugan and Flemings, 2000) and as well as for the development, management, and protection of groundwater resources (Masch and Denny, 1996; Boadu, 2000), hydraulic conductivity and porosity are indispensible input parameters.

Determination of hydraulic characteristics such as hydraulic conductivity, transmissivity and storativity are best obtained through standard techniques such as well tests, permeameter measurements and grain-size analysis. De Lima and Niwas (2000) observed that due to physical and conceptual constraints underlying such tests, the results are non-unique and represent macroscopic averages over large volumes of the pore medium. Additionally, conventional methods such as pump tests are time consuming and costly. In Nigeria, lack of funds has prohibited systematic pumping test activities and analysis, as a result there is paucity of these data for use in successful groundwater development and management practices

However, geoelectrical soundings, which are extensively used for the location of aquifers (Keller and Frischknecht, 1966; Zhody, 1989), can also be used reliably for determining the hydraulic parameters of an aquifer. This is because the physical conditions (tortuosity and porosity) controlling the electric current flow also likewise control the lateral flow of the water in porous media. Exploiting this similarity, Soupios *et al.* (2007) used geophysical methods in combination with pumping tests to estimate aquifer parameters in the Keritis Basin in Chania (Crete – Greece). Massoud *et al.* (2010) used geophysical measurements to estimate hydraulic parameters in the Upper Cretaceous aquifer, central Sinai, Egypt. Asfahani (2012) used Vertical Electrical Sounding technique to estimate transmissivity in the Semi-arid Khanasser Valley region of Syria. Earlier studies carried out in the present study area are on the geochemistry of groundwater (Amadi *et al.,* 1987), prospecting for groundwater resources (Okiongbo and Ogobiri, 2011) and groundwater quality with respect to drinking and agricultural purposes (Okiongbo and Douglas, 2013). These studies focused mainly on the groundwater potential and quality without any attempt to determine the water transmitting properties of the alluvial aquifer.

In this study, porosity and hydraulic conductivity of the shallow alluvial aquifer were determined using surface geoelectrical measurements. This technique is rapid and reliable, and the results can be used to improve the quality of groundwater flow simulation models.

Description of Study Area and Geology

The study area (Yenagoa and environs) is part of the Coastal Sedimentary Basin of Southern Nigeria. It is bounded by Latitudes 04° 23.3′ - 04° 38.2′ North and Longitudes 006° 05′ - 006° 40′ East (Fig. 1). It is a typical deltaic plain with essentially flat topography drained and criss-crossed by a network of rivers, creeks and oxbow lakes e.g Kolo Creek, Epie Creek, Yenagoa and Nun River, etc.

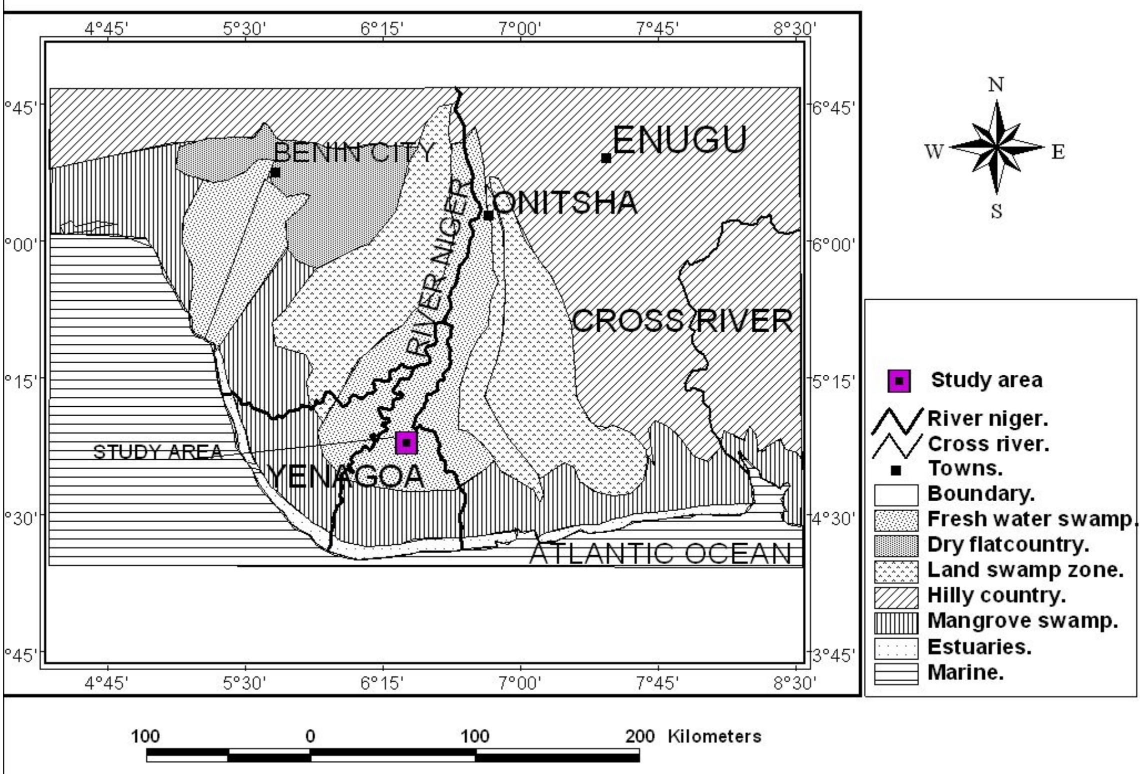

Fig.1: Map of Niger Delta Showing the Study Area (after Akpokodje, 1986)

They all form a network which empties into the Atlantic Ocean through the Nun River Estuary. The low-lying alluvial plains are characterised by vegetation consisting of various trees, including palm trees and a variety of shrubs. The study area has a tropical climate with two distinct seasons, wet (April – October) and dry (November – March). Land use within the area is primarily agricultural. The fertile land produces abundant yields of corn, cassava, plantain etc and provides excellent pasture for cattle. In addition to farming, other local industries include sand and gravel mining.

Water is abstracted from the shallow aquifer. This same aquifer provides municipal and industrial water to the area, as well as domestic water to local inhabitants. Okiongbo and Douglas (2013) reported that the cations and anions analysed from groundwater samples in the area are low and well within the standard specified for drinking and other purposes (WHO, 2004), except iron in about 21% of the samples where the concentrations exceed the permissible limit.

The general geology of the Niger Delta consists of various types of Quaternary deposits overlying thick Tertiary sandy and clayey deltaic deposits. Three main subsurface lithostratigraphic units have been identified (Short and Stauble, 1967) in the Niger Delta. From bottom to top they are Akata, Agbada and Benin Formations. Detailed studies of the Quaternary deposits of the Niger Delta by Allen (1965) revealed that the sediments were deposited under the influence of fluctuating Pleistocene eustatic sea levels. The Coastal Plain Sands (Benin Formation) fluvial in origin have been identified as the main regional and most important freshwater water bearing aquifer in the study area. Groundwater in the Coastal Plain Sands occurs mainly under phreatic (unconfined) conditions. The sediments of the Coastal Plain Sands which were deposited during the Late Tertiary – Early Quaternary period is about 2100 m thick (Abam, 1999) and consists of massive lenticular, unconsolidated coarse to medium–fine grained sands, while gravel and pebbles are minor components with localized clay/shale interbedding. The sands are generally moderately sorted, poorly cemented, and angular in shape (Mbonu *et al.,* 1991). Thin clay horizons and lenses create discontinuities in the vertical and lateral continuity of the porous sands and gravel. This condition results in the presence of local perched aquifers. Groundwater recharge in the aquifer occurs mainly through direct infiltration of rainfall. The unconfined aquifer has high water yielding potential (Amajor, 1991). In the study area, the water table is about 3-4 m during the dry season. During the wet season, the water table rises considerably, in some cases, to the ground surface.

MATERIALS AND METHOD

In this study, fourteen (14) Vertical Electrical Soundings using Schlumberger configuration with a maximum current electrode half-spacing (AB/2) ranging between 100 - 200 m were carried out (Fig. 2). The Schlumberger array was chosen due to its better lateral resolution. Resistivity measurements were carried out using an SAS 1000 Abem Resistivity meter.

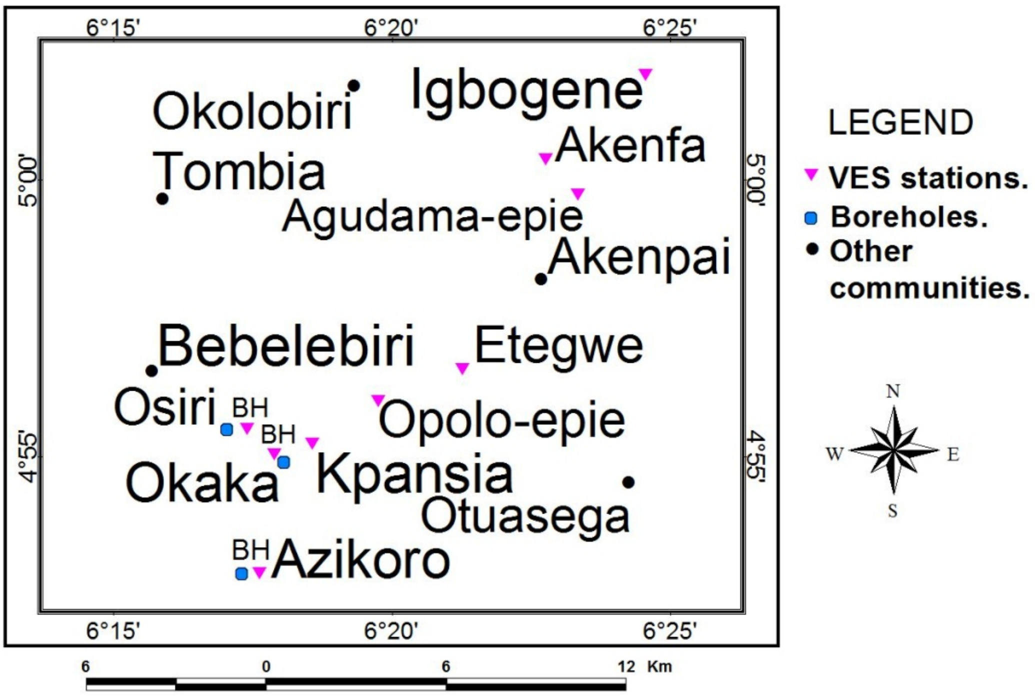

Fig.2: Map of the Study Area Showing VES and Borehole Locations

All VES stations were located close to pre-existing water boreholes. Groundwater samples collected from boreholes, were analysed for electrical conductivity (EC) for the determination of water resistivity. Conventional pump tests were carried out on boreholes at Okaka, Azikoro and Osiri close to VES stations 1, 3 and 11 to determine the hydraulic conductivity for the purpose of comparison with the computed values from the VES data. The pump test involved the measurement of the fall in water level with respect to time. The tests were performed using submersible pumps. The pumping test data were interpreted using Jacob's straight line method (Fetter, 1994) following the formula:

$$T = \frac{2.3Q}{4\pi\Delta s}, \; K = \frac{T}{H} \tag{1}$$

Where T is the transmissivity in m^2/s, Q is the rate of discharge in m^3/s, Δs is the slope in m, K is the hydraulic conductivity in m/s, and H is the saturated thickness in metre obtained from the VES data. The hydraulic conductivity (K) and transmissivity (T) values based on the pump test data at the three sites are shown in Table 1.

The calculated apparent resistivity data were inverted using Interpex, a 1-D inversion software. All depths were constrained with lithological log from the nearest borehole. The model aquifer parameters (thickness and resistivity) for each VES station are shown in Table 2. In order to determine the aquifer porosity at each VES station, Archie's equation was used (Archie, 1942). Archie's law relates the bulk resistivity of a fully saturated granular medium to its porosity and the resistivity of the fluid within the pores according to equation :

$$\rho_b = \alpha\rho_w\phi^{-m} \tag{2}$$

where ρ_b is the bulk resistivity, ρ_w is the fluid resistivity, ϕ is the porosity of the medium, and the dimensionless coefficients α and m (cementation factor) depend on the rock type. The ratio $\frac{\rho_b}{\rho_w}$ is called the apparent formation factor F_a. The application of Archie's law is only valuable for clay-free, clean, sand but fails in predicting the porosity in the case of unclean, clayey and shaly sands (Worthington, 1993). The Quaternary alluvial aquifer under study is characterised by a mixture of sand and clay. In such a case, the apparent formation factor F_a, must be corrected by making a slight modification on the Archie's law in order to take into consideration the presence of clay. The corrected formation factor is denoted as F_c. Substituting the corrected formation factor F_c for F_a, the porosity can be expressed as:

$$\phi = \exp\left[\frac{1}{m}LN(\alpha) + \frac{1}{m}LN\left(\frac{1}{F_c}\right)\right] \tag{3}$$

To estimate the porosity using Eq. (3), literature values of m and α were used for an unconsolidated gravel-sand. Unconsolidated sediments are characterised by relatively low values of m (between 1.1 and 1.5) and parameters $\alpha \approx 1.0$ (Schon, 2004). In this study, we assumed $\alpha = 0.9$ for alluvial aquifer and $m = 1.5$ for Quaternary deposits (Tizro *et al.,* 2012). The model of Waxman and Smith (1968) which relates the apparent and corrected formation factors, F_a and F_c, and takes the clay effects into consideration is used in this study. It can be written as (Worthington, 1993):

$$F_a = F_C\left(1 + BQ_V\rho_w\right)^{-1} \tag{4}$$

where BQ_V term is related to the effects of surface conduction, mainly due to clay particles. In case surface conduction effects are non-existent, the apparent formation factor becomes equal to the corrected formation factor (F_C). This equation was arranged to obtain a linear relationship between $\frac{1}{F_a}$ and ρ_w as:

$$\frac{1}{F_a} = \frac{1}{F_C} + \left(\frac{BQ_V}{F_C}\right)\rho_w \tag{5}$$

where $\frac{1}{F_c}$ is the intercept of the straight line and $\frac{BQ_V}{F_C}$ represents the gradient (Huntley, 1986; Worthington, 1993). Thus, by plotting $\frac{1}{F_a}$ versus pore fluid resistivity ρ_w, we obtain a value for the corrected formation factor, which was subsequently used to estimate porosity using eq. (3). To follow the above approach, we used bulk resistivities (ρ_b), as obtained from the 1-D resistivity inversion and the measured pore fluid resistivities, (ρ_w), obtained using the wells nearest to the VES locations. These values were used to calculate the apparent formation factor $F_a = \frac{\rho_b}{\rho_w}$

of the saturated aquifer. Figure 3 shows a plot of water resistivity as a function of $1/F_a$ where two trends of correlation relationship have been established as follows:

Trend 1 has the following relationship:

$$\frac{1}{F_a} = 0.002\rho_w + 0.108$$

Trend 2 has the following relationship:

$$\frac{1}{F_a} = 0.001\rho_w + 0.024$$

The two trends are as a result of the inhomogeneity of the sedimentary formation and each trend consists of data points with similar subsurface characteristics.

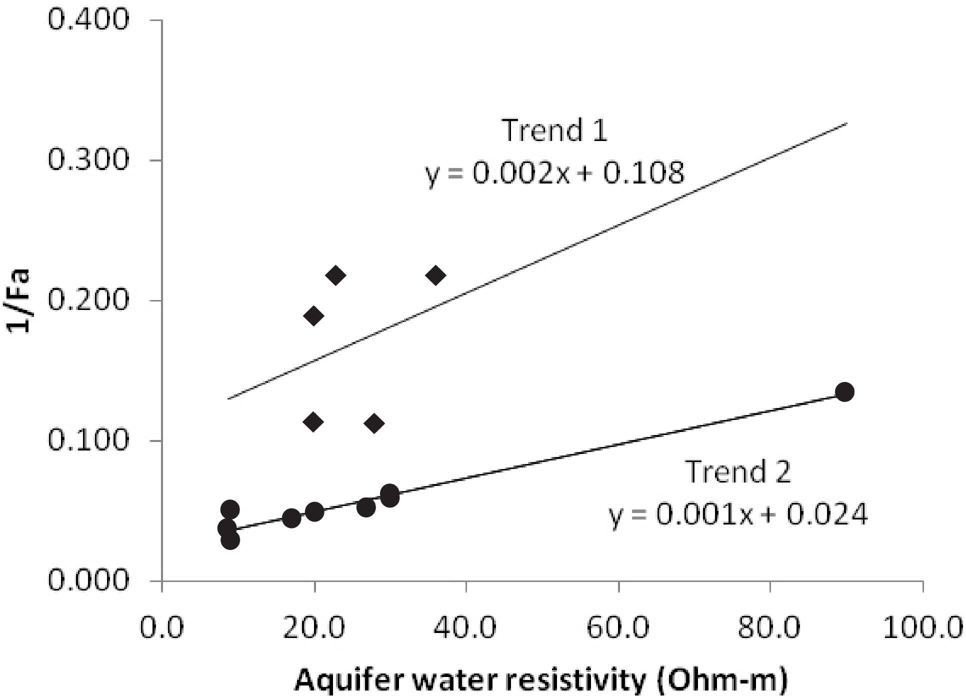

Fig. 3: Plot of 1/Fa versus Aquifer Water Resistivity used to Evaluate F_c

The corrected F_c were computed from these trends and therefore are site dependent. Having determined the porosity, the hydraulic conductivity was determined using the Kozeny-Carman-Bear equation given by Bear (1972) and Domenico and Schwartz (1990) as:

$$K = \left[\frac{\delta_w g}{\mu}\right]\left[\frac{d^2}{180}\right]\left[\frac{\phi^3}{(1-\phi)^2}\right] \quad (6)$$

where d is the sand grain size, δ_w is the fluid density, μ is the dynamic viscosity, and g is the gravitational acceleration. The average grain size (d) used in Eq. (6) was obtained through grain size analysis using the sieve technique carried out on aquifer sand samples in the study area. The average grain size was estimated as 0.0006 m. Following Fetter (1994), the values of δ_w and μ were taken to be 1000 kg/m^3 and 0.0014 kg/ms respectively.

The transverse unit resistance (T_R) and the longitudinal conductance (L_C) (both are Dar-Zarrouk parameters) are defined as follows:

$$T_R = \rho \times h \quad (7)$$

$$L_C = \frac{h}{\rho} \quad (8)$$

where h is the thickness of the aquifer and ϱ is the resistivity of the aquifer.

The groundwater flow through an aquifer is governed by the transmissivity T, which is expressed as:

$$T = Kh \quad (9)$$

Dividing equation 9 by equation 7

$$\frac{T}{T_R} = \frac{Kh}{\rho h}$$

$$T = (K\sigma)T_R \quad (10)$$

Dividing equation 9 by 8

$$\frac{T}{L_C} = K\rho$$

$$T = (K\rho)L_C$$

$$T = \alpha L_C \tag{11}$$

Where

$$\alpha = K\rho$$

Hence,

$$K = \frac{\alpha}{\rho} \tag{12}$$

Where K is the hydraulic conductivity, h is the aquifer thickness at the location of the VES sounding, ρ is the resistivity of the aquifer, L_C is the longitudinal conductance and T_R is transverse resistance. Niwas *et al.* (2011) reported that Eq. (10) exists in case of highly conducting basement (transverse resistance (R) dominant aquifer where electrical currents tend to flow vertically) and Eq. (11) exists in case of highly resistive basement (longitudinal conductance (L_C) dominant aquifer where electrical currents tend to flow horizontally). Equation 11 was chosen for the determination of the transmissivity. This is because the aquifer in the present case exhibits H-type sounding curves at all stations in the study area. So, it was more appropriate to estimate the hydraulic conductivity and transmissivity variations in the aquifer by means of the longitudinal conductance (Eq. 11) rather than the transverse resistance (Eq. 10). Two methods were used in the estimation of hydraulic conductivity. Method I uses Eqs. (3), (6) to calculate the porosity and hydraulic conductivity from resistivity data whereas Method II uses Eq. (12) to calculate the hydraulic conductivity and transmissivity was determined using equation 11. In method – II, the hydraulic conductivity from the Osiri borehole, close to VES 11 was chosen as a reference well and was used to ascertain the value of the constant of proportionality (α) expressed as $\alpha = K\rho$. The hydraulic conductivity (K) from pump test in the reference well (Osiri borehole) was 3.5×10^{-4}m/s, and the aquifer resistivity from the nearby VES was 165.0 Ωm. The value of α obtained was 0.058. Table 3 gives the relevant hydraulic parameters computed from Method – II, with $K = 0.058/\rho$ and $T = 0.058L_C$.

RESULTS AND DISCUSSION

Table 2 shows the model values of the thicknesses and resistivities obtained from the interpretation of the VES data. Figure 4 shows the correlation of the results obtained from VES11 and the lithological information from the nearest borehole. Based on the correlation, three major subsurface layers are recognised. The first geoelectric layer extends from the ground surface to a depth of 1.5 m with layer resistivity of 46 Ωm. The second geoelectric layer extends to a depth of 11.1 m, with resistivity value of 13 Ωm. This layer corresponds to clay.

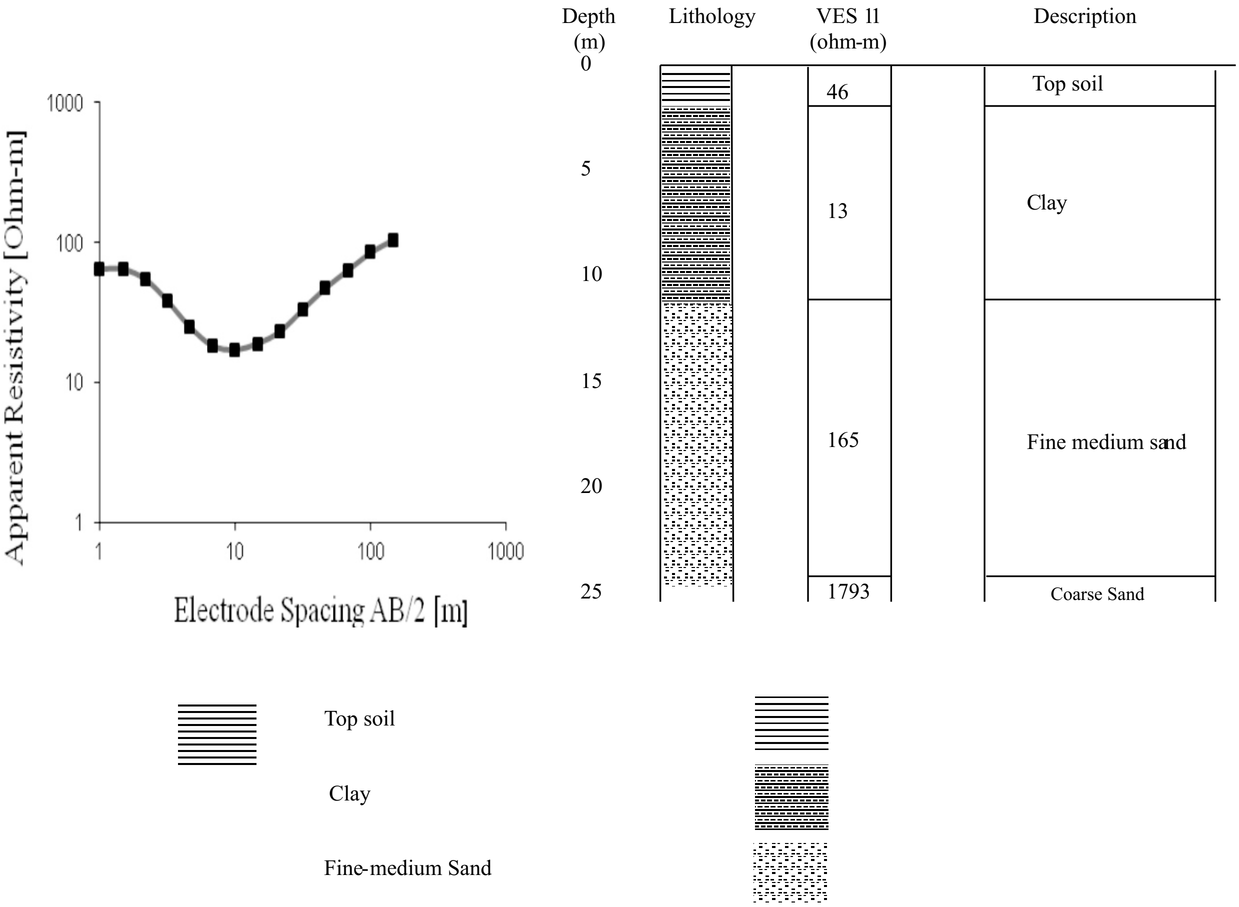

Fig. 4 (a) VES 11 Curve and (b) Correlation of the VES 11 Interpretation Results with the Lithology of the nearest Borehole.

Table 1: Results of Pumping Test for Three Existing Boreholes in the Study Area

S/N	Well name	Hydraulic Conductivity (m/s) $\times 10^{-4}$	Transmissivity (m^2/s) $\times 10^{-3}$
1	Okaka	1.2	5.7
2	Azikoro	3.2	31.2
3	Osiri	3.5	4.8

Table 2: Geoelectric Parameters Obtained from the VES data

VES	Layer Thickness (m)				Layer Resistivity (ohm-m)				
No	1	2	3	4	1	2	3	4	5
1	2.1	2.5	47.5	-	5	24	510	2945	-
2	1.3	25.5	22.8	-	222	72	105	293	-
3	0.5	4.7	97.6	-	57	31	230	126	-
4	1.4	13.2	25.6		72	15	664	149	-
5	0.7	5.7	17.9	-	146	57	304	1413	-
6	0.5	6.3	81.2	-	240	47	376	375	-
7	2.0	5.8	51.2	-	51	10	404	155	-
8	0.8	5.9	18.9	-	145	59	178	1413	-
9	1.1	11.1	29.4	-	214	477	110	1204	-
10	3.2	35.2	30	-	57	175	39	414	-
11	1.5	9.6	13.7	-	46	13	165	1793	-
12	1.2	4.4	24.3	-	18	56	136	86	-
13	2.8	5.6	21.7	-	36	131	250	75	-
14	0.6	1.5	8.3	17.7	208	115	511	70	748

The third geoelectric layer extends to a depth of about 24.8 m and has a resistivity of 165 Ωm. This layer corresponds to fine-medium grained sand and is interpreted to be the fresh water saturated alluvial aquifer. Table 3 shows the results of the calculated porosity and hydraulic conductivity of the aquifer layer.

Table 3: Estimated Porosity and Hydraulic Conductivity (Method I)

VES No	h (m)	ρ (Ωm)	ϱ_w	1/Fa	1/Fc	Φ	K (m/s) × 10^{-4}
1	47.5	509.5	30.0	0.059	0.029	0.09	12.6
2	20.2	105.6	23.0	0.218	0.172	0.29	6.9
3	97.6	230.2	8.6	0.037	0.028	0.09	12.6
4	25.6	663.5	90.0	0.136	0.046	0.12	31.9
5	17.9	303.6	9.0	0.03	0.021	0.08	0.09
6	81.2	376.0	17.0	0.045	0.028	0.09	12.6
7	51.2	404.1	20.0	0.049	0.029	0.09	12.6
8	47.4	178.2	9.0	0.051	0.042	0.11	24.0
9	11.4	477.7	30.0	0.063	0.033	0.10	17.6
10	30.3	175.3	20.0	0.114	0.074	0.16	82.9
11	13.7	165.0	36.0	0.218	0.146	0.26	4.6
12	30.4	106.0	20.0	0.189	0.149	0.26	4.6
13	28.1	250.3	28.0	0.112	0.056	0.14	53.0
14	8.3	511.0	27.0	0.053	0.026	0.08	0.09

Ha – aquifer thickness from VES data, ϱ_w – pore water resistivity, K – hydraulic conductivity,

Table 4: Results of the Calculated Hydraulic Parameters (Method II)

VES No	h (m)	ϱ (Ωm)	K obs [m/s] × 10^{-4}	L_C=h/ϱ [mho]	Com [m/s] × 10^{-4}	T=0.053L_C, [m²/s] × 10^{-4}
1	47.5	509.5	1.2	0.093	1.14	53.9
2	20.2	105.6		0.191	5.49	110.8
3	97.6	230.2	3.2	0.413	2.52	239.5
4	25.6	663.5		0.039	87.4	22.6
5	17.9	303.6		0.059	1.91	34.2
6	81.2	376.0		0.216	1.51	125.3
7	51.2	404.1		0.127	1.44	73.7
8	47.4	178.2		0.090	3.25	52.2
9	11.4	477.7		0.024	1.21	13.9
10	30.3	175.3		0.173	3.31	100.3
11	13.7	165.0	3.5	0.061	3.52	35.4
12	30.4	106.0		0.287	5.47	166.5
13	28.1	250.3		0.112	2.32	65.0
14	8.3	511.0		0.016	1.14	9.3

h – aquifer thickness from VES data, ϱ – aquifer resistivity, L_C – longitudinal conductance, K_obs – hydraulic conductivity from pump test data, K_com – computed hydraulic conductivity from VES data, T – transmissivity

The porosity values vary between 0.08 – 0.29. It is important to mention that the resistivity values obtained from the VES analysis are average values. Thus, the calculated porosity values of the aquifer using the average resistivity values could also be considered as average porosity values. The porosity values range obtained in this study compare fairly well with the porosity range of 0.20 – 0.35 reported by (Todd, 1980) for granular aquifers. The hydraulic conductivity computed using Kozeny-Carman-Bear equation (Eq. 6) varies between a minimum of 0.09 × 10^{-4}m/s and a maximum of 82.9 × 10^{-4}m/s (Table 3). The calculated hydraulic conductivity values were plotted as a function of aquifer resistivity (Fig. 5). Figure 5 shows an inverse relationship between the two parameters. Niwas *et al.* (2011) reported that both direct and inverse relation exists between hydraulic conductivity and electrical resistivity of an aquifer and that an inverse relation exists if the basement layer is resistive.

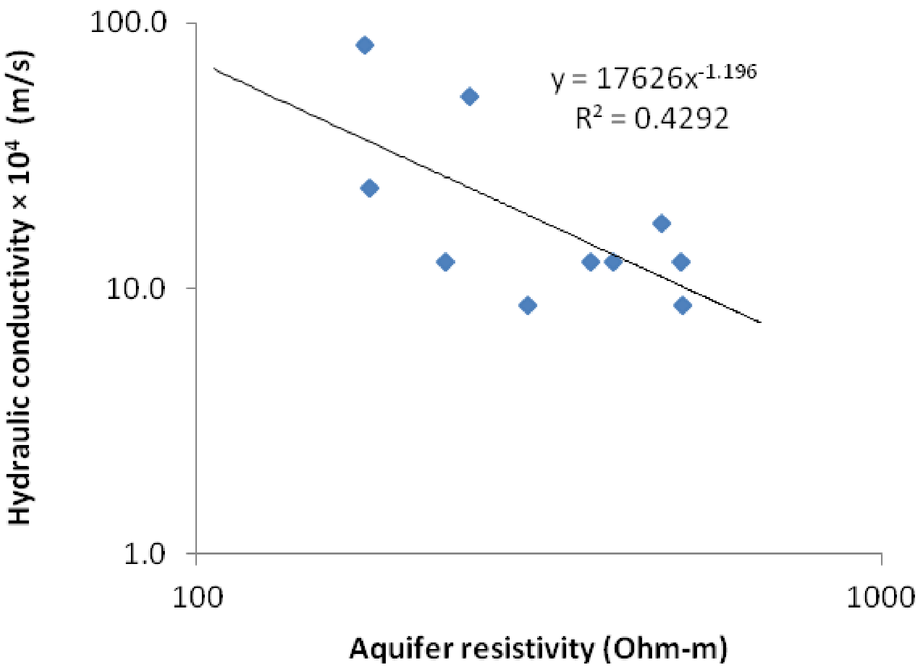

Fig. 5. Relationship Between Aquifer Resistivity and Hydraulic Conductivity

The results of the three pumping tests carried out to determine the hydraulic conductivity of the aquifer are presented in Table 1. The locations of these pumping test sites are Okaka Housing Estate (close to the location of VES 1), Azikoro (close to the location of VES 3), and Osiri (close to the location of VES 11), where the hydraulic conductivity was 1.2×10^{-4} m/s, 3.2×10^{-4} m/s, and 3.5×10^{-4} m/s respectively (Table 1). In order to check the accuracy of the computed hydraulic conductivity values, the three hydraulic conductivity values determined from the pump test (the minimum acceptable number of points from statistical point of view) were plotted against the values obtained in the same location by resistivity method. The regression line fitted to these data indicates a good relation (Fig. 6). These estimations affirm that surface geoelectrical data give useful hydrogeological information.

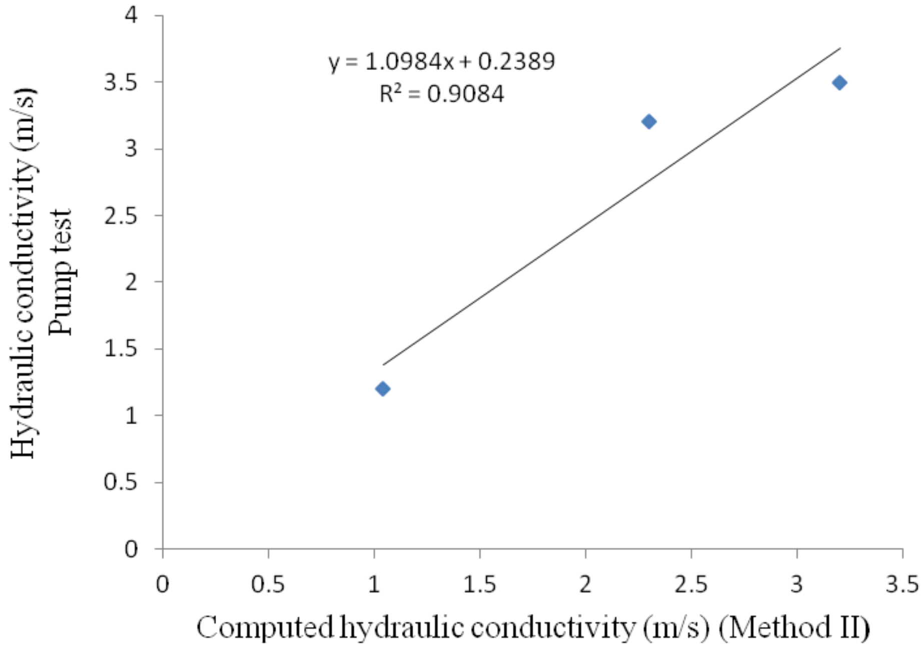

Fig. 6: Relation Between Field Hhydraulic Conductivity and Computed Hydraulic Conductivity

Also, the hydraulic conductivity values estimated from the VES data and those obtained from the pump test are within the range 10^{-5} – 10^{-2}m/s which is the characteristic range of sand and gravel aquifer (Kallergis, 1999). The estimated hydraulic conductivity values from Method II show better correlation with the pump test hydraulic conductivity values. The estimated transmissivity values using the VES data (Table 4) vary between a minimum of 9.3 × $10^{-4}$$m^2$/s at VES 14, and a maximum of 239.5 × $10^{-4}$$m^2$/s at VES 3. The estimated transmissivity values of the geological formation in the study area shows a wide range due to the inhomogeneity of the sedimentary formation. The high transmissivity values recorded in the study area are consistent with the finding that the Quaternary aquifer is composed of unconsolidated fine-medium-coarse sand (Mbonu *et al.,* 1991).

CONCLUSION

Schlumberger Vertical Electrical Soundings carried out within Yenagoa and environs were used to determine the alluvial aquifer hydraulic parameters. Method I combined modified Archie's equation to determine the porosity, while Kozeny-Craman-Bear equation was used to determine hydraulic conductivity. In Method II, hydraulic conductivity for at least one VES location must be known. Estimated porosities and hydraulic conductivity values vary from 0.08 – 0.29 and 1.14 × 10^{-4} – 87.4 × 10^{-4}m/s respectively. The estimated hydraulic conductivity values using Method II show significant correlation with the available hydraulic conductivity values obtained from the analysis of pump test data. The good correlation of the hydraulic conductivity values estimated from method II using the VES measurements and those obtained from pump tests is a good indication of the reliability and applicability of the method.

ACKNOWLEDGEMENTS

The authors are immensely grateful to the Postgraduate Exploration Geophysics students in the Department of Physics for assisting in the VES data acquisition and Mr Eleazer Ogulu who assisted in carrying out the pump test. We also thank the two anonymous reviewers for critically reviewing the original manuscript and for giving relevant suggestions for improvement.

REFERENCES

Abam, T. K. S. 1999. Dynamics and quality of water resources in the Niger Delta. Impacts of urban growth on surface and groundwater quality. *Preceedings of IUGG Symposium, Birmingham, Publication* no.

259.

Allen, J.R.L. 1965. Late Quaternary Niger Delta and adjacent areas: sedimentary environments and lithofacies. *American Association of petroleum Geologists Bulletin,* 49, 549 – 600.

Akpokodje, E. G. 1986. A method of reducing the cement content of two stabilized Niger Delta soils. Quarterly Journal of Eng. Geol. London, 19, 359 – 363.

Amadi, P. A.; Ofoegbu, C.O. and Morrison, T. 1987. Hydrogeochemical assessment of groundwater quality in parts of the Niger Delta, Nigeria. *Environmental Geology and Water Science.* 14, 195-202.

Amajor, L. C. 1991. Aquifers in the Benin Formation (Miocene – Recent), Eastern Niger Delta , Nigeria. Lithostratigraphy, Hydraulics and water quality. *Environmental Geology & Water Science*, 17, pp.85 – 101.

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.

Asfahani, J. 2012. Quaternary aquifer transmissivity derived from Vertical Electrical Sounding measurements in the Semi-arid Khanasser Valley Region, Syria. *Acta Geophysica,* 60, 1143-1158.

Bear, J. 1972. *Dyanamic of fluids in Porous media.* Elsevier, New York.

Boadu, F. K. 2000. Hydraulic conductivity of soils from grain size distribution: new models. *J. Geotech. Geoenvironmental Eng.,* 126, 739 – 746.

Dewhurst, D. N.; Yang, Y. and Aplin, C. A. 1999. Permeability and fluid flow in natural Mudstones. In Aplin, A.C., Fleet, A. J., and Macquaker, J.H.S. (Eds), *Muds and Mudstones: Physical and Fluid flow properties.* Geological Society, London, Special Publications, 158, 23-43.

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

Domenico, P.A. and Schwartz, F.W. 1990. *Physical and Chemical Hydrology*, 2nd ed, John Wiley and Sons, New York.

Dugan, B. and Fleming, P.B. 2000. Overpressure and fluid flow in the New Jersey Continental slope: Implications for slope stability, failure and cold seeps. *Science*, 289, 288-291.

Fetter, C.W. 1994. *Applied Hydrogeology*, 3rd ed, Prentice-Hall, Inc., New Jessey. Hermanrud, C. 1993. *Basin modelling techniques – An overview in Basin modelling advances and applications.* Norwegian Petroleum Society Special Publication, Dore A.G. (Ed), Elsevier, New York, 3, 1-34.

Huntley, D. 1986. Relations between permeability and electrical resistivity in granular aquifers. Groundwater, 24, 466-474.

Kallergis, G. 1999. *Applied Environmental Hydrogeology*, Technical Chamber of Greece, Athens.

Keller, G. V. and Frischknecht, F.C. 1966. *Electrical method in geophysical prospecting.* Pergamon Press, Oxford.

Massoud, U., Santos, F., Khalil, M. A., Taha, A., and Abbas, A. M. 2010. Estimation of aquifer hydraulic parameters from surface geophysical measurements: a case study of the Upper Cretaceous aquifer, central Sinai, Eqypt. *Hydrogeology Journal*, 18, 699-710.

Mbonu, P.D.C.; Ebeniro, J.O.; Ofoegbu, C. O. and Ekine, A.S. 1991. Geoelectric sounding for the determination of aquifer characteristics in parts of Umuahia area of Nigeria. *Geophysics,* 56, 284-291.

Masch, F. D. and Denny, K. J. 1996. Grain size distribution and its effects on the hydraulic conductivity of unconsolidated sands. *Water Resources Research*, .2, 665 – 677.

Niwas, S.; Tezkan, B. and Israil, M. 2011. Aquifer hydraulic conductivity estimation from surface geoelectrical measurements for Krauthausen test site, Germany. *Hydrogeology Journal*, 19, 307-315.

Okiongbo, K.S. and Ogobiri, G. 2011. Geoelectric investigation of groundwater resources in parts of Bayelsa State, Nigeria. *Research Journal of Environmental and Earth Sciences,* 3, 620 – 624.

Okiongbo, K.S. and Douglas, R. 2013.

Hydrogeochemical analysis and evaluation of groundwater quality in Yenagoa city and environs, Southern Nigeria. *Ife Journal of Science,* 15, 209-222.

Schon, J. H. 2004. *Physical properties of Rocks*. Elsevier, Amsterdam.

Short, K.C. and Stauble, J. 1967. Outline of the geology of the Niger Delta. Bull. *AAPG*., 51, 761- 779.

Soupios, P. M., Kouli, M., Vallianatos, F., Vafidis, A., and Stavroulakis, G. 2007. Estimation of aquifer hydraulic parameters from surficial geophysical methods: A case study of Keritis Basin in Chania (Crete-Greece). *Journal of Hydrology*, 338, 122-131.

Tizro, A. T.; Voudouris, K. and Basami, Y. 2012. Estimation of porosity and specific yield by application of geoelectrical method – A case study in western Iran. *Journal of Hydrology,* 454-455, 160-172.

Todd, T. D. 1980. *Groundwater Hydrology*, 2nd ed, Wiley and Sons, New York.

Waxman, M. H. and Smith, L. J. M. 1968. Electrical conductivities in oil bearing sands. *Journal of the Society of Petroleum Engineers*, 8, 107-122.

World Health Organisation (WHO) 2004. *Guidelines for drinking water quality*: incorporating 1st and 2nd vol. 1 recommendations 3rd ed., Geneva.

Worthington, P. F. 1993. The uses and abuses of the Archie equations: The formation – porosity relationship. *Journal of Applied Geophysics*, 30, 215-228.

Zohdy, A.A.R. 1989. A new method for automatic interpretation of Schlumberger and Wenner sounding curves. *Geophysics*, 5, .245-252.