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Investigation of Hydrogeologic Properties using Resistivity Data in Parts of Delta State, Nigeria

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ABSTRACT
The main purpose of this study was to determine the hydrogeological conditions and to identify the groundwater potential zone within Benin Formation in Agbor-Obi,Ika South local government area , Delta State. The hydrogeologic properties were investigated using the electrical resistivity method within parts of Delta state in order to appraise the geohydraulic properties of the aquifer units. The electrical resistivity technique employed the Schlumberger electrode configuration with vertical electrical sounding (VES) within a maximum current electrode separation of 300 m. Interpretation of the VES result revealed five to six geo-electric layers with values of their resistivity, thickness and depth, and lithology delineated are mostly topsoil, medium grained sand, fine-medium grained sand and gravelly sand. The geo-electric indices (bulk aquifer resistivity and thickness) were used to estimate the geohydraulic parameters. Effective porosity (28.36 to 31.96 %), hydraulic conductivity (0.1567 to 0.1752 m/s), transmissivity (2.76 to 6.23 m^2/s), storativity (0.0540 to 0.1092), and hydraulic diffusivity (53.23 to 58.38 m/s) aid in appraising the groundwater repositories. The contour distribution of the geohydrodynamic parameters indicate their variations across the study area. These parameters are essential elements that are paramount in groundwater resource management and conservation. The result obtained from this study has identified the subsurface aquifer and given insight into the groundwater potential as it affects economic growth and social cohesion.

Keywords:
Aquifer,
Geo-electric layers,
Resistivity,
Transmissivity,
VES.

INTRODUCTION
Groundwater is part of the water cycle located beneath the earth's surface in pores and crevices of rocks and soil, the location present challenges for quantifying and management groundwater resources compared with surface water. The amount of pores and crevices in the subsurface soil and rocks and their interconnectivity controls ease of movement of groundwater through the subsurface. In the coastal region where the depth to the water table is not very deep creates potential risk to the groundwater resources (Ibuot et al., 2019; Omeje et al., 2021). The knowledge of geohydraulic parameters will provide useful information that will help in abstraction and management of the coastal groundwater resources. Geohydraulic parameters such as porosity, transmissivity, hydraulic conductivity are important parameters in groundwater exploration and exploitation; as such there is need for prior knowledge of the subsurface hydro-geological conditions before drilling of a borehole/well. It has been difficult to convince individuals especially in the coastal region to undertake geophysical survey before drilling and the refusal by some people to use the available geophysical report in areas that have been surveyed. This has resulted in incessant wildcat drilling and failed boreholes (Ibuot et al., 2013; George et al., 2014). Acquisition and interpretation of resistivity data has provided information about the subsurface geologic strata and the characteristics of the earth materials (Niwas and Singhal, 1981; George et al., 2015a; George et al., 2015b; Lashkaripour and Nakhaei, 2005; Obiora et al., 2016; Ekanem et al., 2019). A good understanding of the aquifer properties is essential in groundwater exploration as it helps in determination of the depth to the hydro-geologic units, groundwater flow, and quality of the groundwater resources.
Aquifer properties such as permeability, porosity, resistivity, layer thickness and aquifer yield control groundwater flow, availability, quality and potential (Uwa et al., 2018; George 2020; Ibuot et al., 2022). The aquifer hydro-geologic parameters can be estimated from the measured field parameters. These parameters

vary spatially due to heterogeneity of the geology of an area. The variation of resistivity in the subsurface is caused partly by the seepage/leakage of fluid (water) from the surface and the resistivity response, which depends on the flow of the leaked fluid.

Surficial electrical resistivity measurement employing vertical electrical Sounding is a non intrusive technique, which is important in a geophysical survey to investigate the subsurface hydrogeologic units. It establishes the relationships between geohydraulic and geoelectric parameters such as hydraulic conductivity and transmissivity. This paper employs the electrical resistivity data to estimate the geohydraulic properties of the subsurface aquifer units, which will enhance the determination of the aquifer potential of the study area.

Location and Geology of the study area

The study area is located in Agbor-Obi in Ika south local government area of Delta State, Nigeria (Figure 1). It lies within latitudes $6^0$10'N and $6^0$20'N and longitudes $6^0$10'E and $6^0$20'E. The study area is located in an equatorial climatic region that is characterised by two major seasons. The seasons are the rainy season (March–October) and dry season (November–February) (Martínez et al., 2008; George et al., 2010). Geologically, the town is situated within the Benin Formation, which is a sedimentary rock formation that dates back to the Early to Late Eocene Epochs of the Paleogene Period, approximately 56-34 million years ago. The Benin Formation is composed of sandstones, shales, and mudstones that were deposited in a shallow marine environment. The formation is part of the Niger Delta Basin, which is a major sedimentary basin that extends across several West African countries (Reijers et al., 1997; Nganje et al., 2007). In Agbor-Obi specifically, the geology is characterized by a mix of hilly and flat terrain, with some areas of exposed rock formations. In terms of mineral resources, Delta State is known to have deposits of oil and gas, as well as solid minerals such as limestone, kaolin, clay, silica, and sand. However, it is not clear whether Agbor-Obi specifically has any significant mineral deposits.

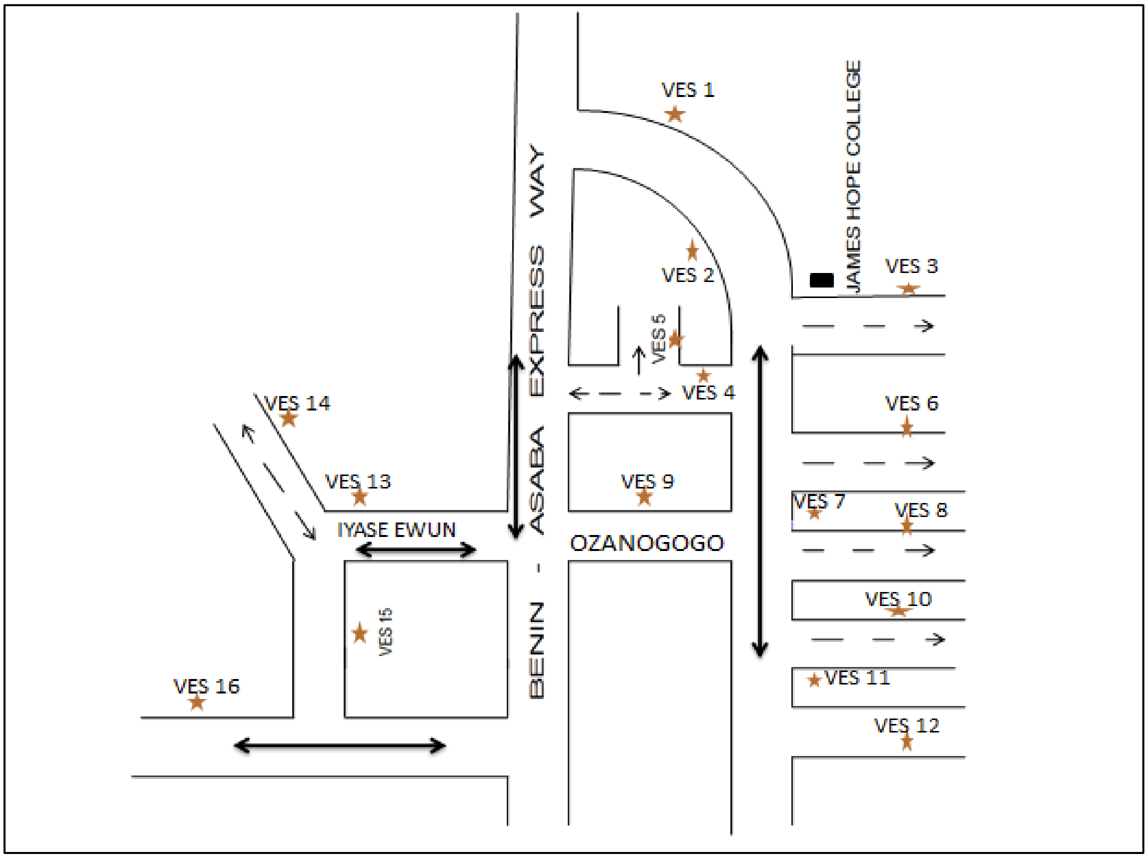

Figure 1: Map showing the studied area and VES locations

MATERIALS AND METHODS

The exploration for groundwater is affected by inadequate knowledge of the hydraulic properties of the subsurface aquifer. These properties play major role in the quantification of the subsurface hydrogeological units (George et al., 2015a). Hydraulic conductivity (K) is a property that characterises the hydraulic behaviour of an aquiferous layer and control the ease with which groundwater flow through an aquifer. The value of K is significant as it can be use as tools for hydrogeological modelling. In estimating the hydraulic conductivity, equation 1 Kozeny-Carmen-Bear's model equation (1927; 1956; 1972) was employed. Hydraulic conductivity, porosity and other site-dependent parameters are related by equation 1.

$$K = \left(\frac{\delta_w \cdot g}{\mu_d}\right) \cdot \left(\frac{{d_m}^2}{180}\right) \cdot \left(\frac{\phi^3}{(1-\phi)^2}\right) \quad (1)$$

Where δ_w is density of water (1000 kg/m^3), d_m is the mean grain size determined by direct measurement using vernier calliper and micrometer screw guage as 0.00035 m, μ_d is the dynamic viscosity of water given as 0.0014 kg/ms (Fetters 1994) and g is the acceleration due to gravity.

Porosity (ϕ) is the fraction of the volume of open space (pore space) in soils/rocks in relation to the total soil/rocks volume. As the drainable pore space, it can be calculated using equation 2. According to George et al. (2018) and Oguama et al. (2020), porosity is a rock property that determines aquifer productivity and depends on the grain composition of rocks, the way the rocks are formed, and the pressure to which rocks are exposed.

$$\phi = 36.51{\rho_a}^{-0.031} \quad (2)$$

where ρ_a is the aquifer resistivity and ϕ in %.

Transmissivity (Tr) determine the flow rate of groundwater through a saturated aquifer layer under a hydraulic gradient. The relationships between transmissivity and Dar-Zarrouk parameters according to Niwas and Singhal, 1981 are expressed in equation 5. The product of the estimated K values and the aquifer thickness (h) gives the values of transmissivity (equation 3).

$$Tr = K\sigma T = \frac{KS}{\sigma} = Kh \quad (3)$$

Where T and S are the transverse resistance and longitudinal conductance respectively. Storativity also referred to as the storage coefficient is the amount of groundwater released from or taken into storage with respect to the change in water level (head) and surface area of the aquifer. The value of storativity depends on the aquifer type (unconfined or confined). According to Anomohanram et al. (2020), it is the groundwater bearing capacity of aquifer. The storativity of an unconfined aquifer varies from 0.01 to 0.3 while that of a confined aquifer varies from 1×10^{-5} to 1×10^{-3}. To estimate the storativity for an unconfined aquifer, we use equation 4 proposed by Hamil and Bell (1986) and Guideal et al. (2011).

$$S = 3 \times 10^{-3} h \quad (4)$$

Where h is the thickness of the aquifer in meters (m).

This property describes the hydraulic properties of the aquifer unit, and it measures the diffusion speed of pressure disturbances in groundwater repositories. The combination of aquifer transmissivity and storativity gives a formation parameter called diffusivity (Hiscock 2005). It is the ratio of aquifer transmissivity in (m^2/day) to aquifer storativity as given in equation 5;

$$D = \frac{Tr}{S} \quad (5)$$

Where Tr is aquifer transmissivity, and S is aquifer storativity.

The study employs Schlumberger electrode configuration with vertical electrical sounding (VES) for ten soundings using the IGIS Resistivity meter model SSR-MP-ATS. The half current electrodes spread $\left(\frac{AB}{2}\right)$ and half potential electrodes spread $\left(\frac{MN}{2}\right)$ ranged from 1.0 – 400.0m and 0.25 – 20.0m respectively. Equation 6 was used to calculate the apparent resistivity (ρ_a).

$$\rho_a = G\frac{V}{I} = G.R_a \quad (6)$$

where G is the geometric factor which depends on the electrodes arrangements. R_a is the apparent resistance measured on the field. The G is expressed in the relation below;

$$G = \pi.\left[\frac{\left(\frac{AB}{2}\right)^2 - \left(\frac{MN}{2}\right)^2}{MN}\right]$$

The data were reduce to 1-D geological models utilising the manual and computer modelling techniques (Zohdy, 1965, Zohdy et al., 1974). The computed apparent resistivities were plotted against $\frac{AB}{2}$ on bi-logarithmic graphs and the curves obtained were smoothened in order to eliminate the effects of lateral heterogeneities and other forms of noisy signatures (Chakravarthi et al., 2007, Akpan et al., 2006). The curves were curve matched using master curves and charts according to Orellana and Mooney, 1966. The values of the apparent resistivity were inputted into computer software program (WinResist) for the computer modelling which generates a set of geoelectric curves (Figures 2-3) from where the values of resistivity, thickness and depth of each geoelectric layer were obtained. The depths during the inversion process were constraint using borehole lithologic log by fixing layer thicknesses and depths while allowing the resistivities to vary (Lowrie, 1997; Batayneh, 2009).This reduces the ambiguities in the interpretation stage and enhances the reliability and quality assurance of the modelled results. The curves obtained from VES 1 and 2 show a goodness of fit (Figures 2 and 3) and the inverted results over half space. The curves show a wide variation in values of resistivity, thicknesses and depths between and within the subsurface layers penetrated by current.

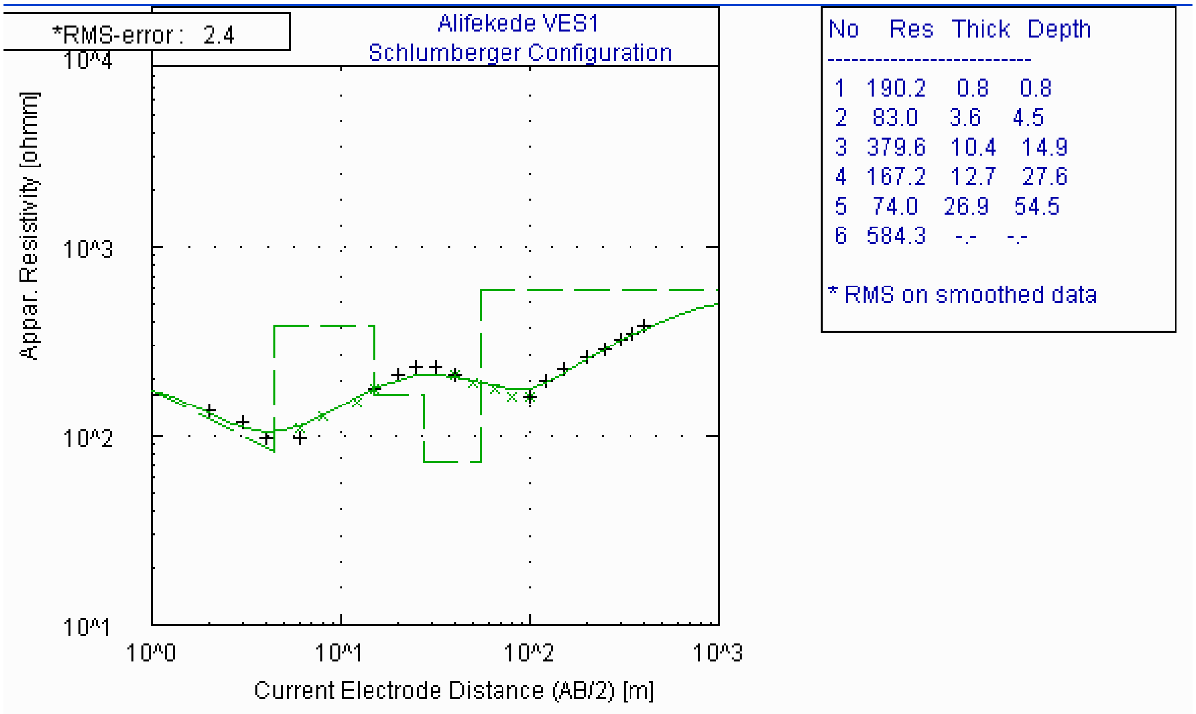

Figure 2: Typical Sounding Curve for Agbor-Obi VES 1

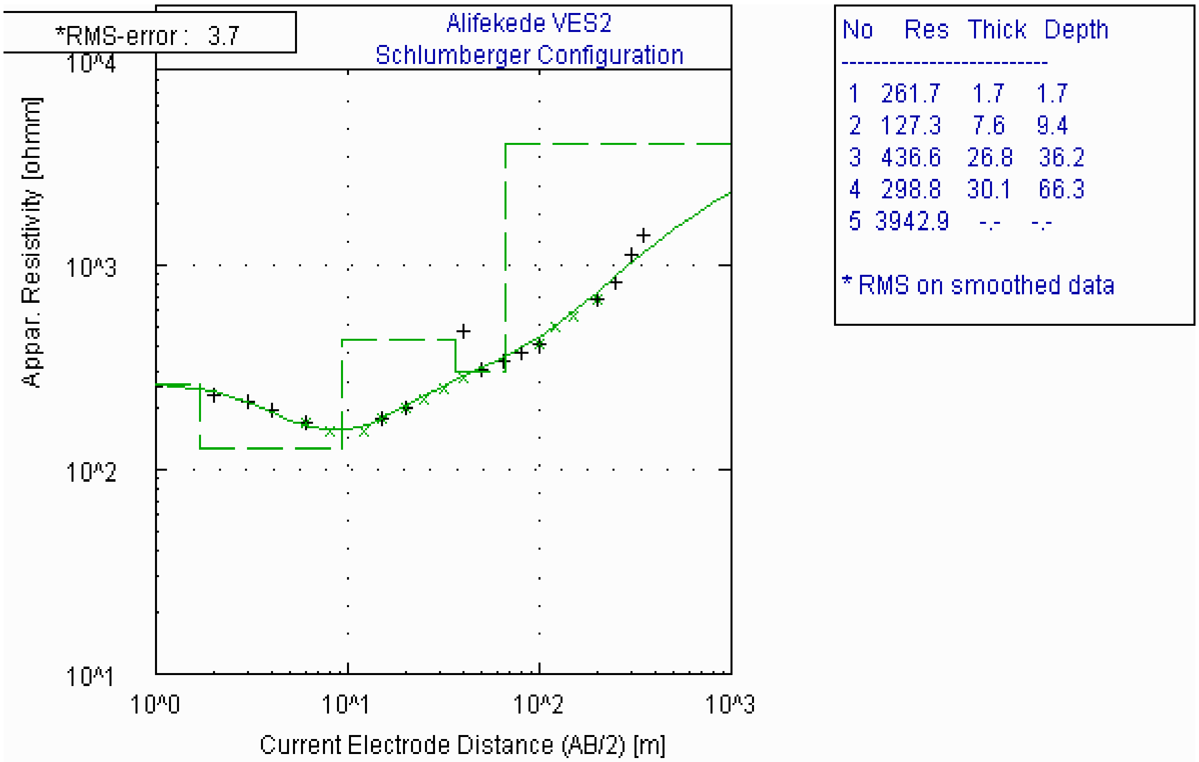

Figure 3: Typical Sounding Curve for Agbor-Obi VES 2

RESULTS AND DISCUSSION

The VES results show that the study area comprises of five to six subsurface geoelectric layers penetrated within the maximum current electrode separation employed (Table 1). The resistivity of the first layer ranges from 88.4 to 1022.5 Ωm, indicating that the layer consists more of sand of varying contents of clay and is generally unsaturated. This layer enhances percolation of rain water into the underlying geologic units. The thickness and depth of the top layer varies between 0.7 and 1.7 m. The second layer with a resistivity values ranging from 72.9 to 434.7 Ωm which consist of medium resistivity materials, which may be attributed to the presence of medium-coarse brownish sand intercalation with clay. The conductivity of this layer is generally less compared to the overlying layer. The thickness and depth of the second layer ranges from 3.6 to 10.2 m and 3.9 to 11.5 m respectively. The third layer is characterized with resistivity values ranging from 37.4 to 2644.7 Ωm with the thickness and depth ranging from 8.6 to 37.1 m and 12.4 to 43.1 m respectively. Underlain the third layer is the fourth layer which is a more resistive layer than the overlain layers with resistivity values ranging from 141.6 to 3500.1 Ωm, This layer was interpreted as fine to medium grained sand and was delineated as the aquifer unit except in VES 1 and 12. The thickness and depth of this layer ranged from 11.8 to 36.4 m and 24.2 to 68.3 m. The fifth layer has resistivity values ranging from 74.0 to 4509.4 Ωm, while the thickness and depth were undefined except in VES 1 and 12. The sixth layer was delineated in VES 1 and 12 with resistivity values ranging from 584.3 and 610.0 Ωm. The lateral variations in resistivity can be attributed to the complex nature of the subsurface geology and lateral lithological discontinuities in the study area (George et al., 2015b; Ibuot et al., 2019; Ekanem et al., 2020).

The bulk aquifer resistivity and thickness within the maximum current electrode separation employed (Table 2) were observed to range from 74.0 to 3500.1 Ωm and 17.3 – 36.4 m.

Table 1: Summary of VES results

VES No.	Longitude (^{O}E)	Latitude (^{O}N)	Layer Resistivity (Ωm)						Layer thickness (m)					Layer depth (m)					Curves types
			ρ_1	ρ_2	ρ_3	ρ_4	ρ_5	ρ_6	h_1	h_2	h_3	h_4	h_5	d_1	d_2	d_3	d_4	d_5	
1	6.1667	6.1667	190.2	83.0	379.6	167.2	74.0	584.3	0.8	3.6	10.4	12.7	26.9	0.8	4.5	14.9	27.6	54.5	HKQH
2	6.1878	6.3221	261.7	127.3	436.6	298.8	3942.9	--	1.7	7.6	26.8	30.1	--	1.7	9.4	36.2	66.3	--	HKH
3	6.1990	6.2155	648.9	216.9	742.3	1941.7	3849.8	--	0.9	7.6	14.2	17.3	--	0.9	8.5	22.7	40.0	--	HKH
4	6.2121	6.2466	209.8	72.9	370.9	464.1	555.7	--	1.4	4.5	37.1	25.2	--	1.4	6.0	43.1	68.3	--	KHA
5	6.1768	6.1769	286.4	142.1	337.7	668.4	4509.4	--	1.1	5.1	19.8	19.0	--	1.1	6.2	26.0	45.0	--	HAA
6	6.2897	6.2455	359.3	250.6	359.8	1075.2	2199.3		1.0	7.8	13.2	19.2	--	1.0	8.8	21.9	41.1	--	HAA
7	6.4568	6.3243	651.9	308.9	230.9	731.4	2099.3	--	1.2	9.3	18.9	19.0	--	1.2	10.5	29.4	48.4	--	QHA
8	6.3212	6.2787	126.7	266.5	136.4	345.2	1243.6	--	0.7	4.3	26.8	31.2	--	0.7	5.0	31.8	63.0	--	KHA
9	6.2899	6.3435	141.0	279.4	130.0	360.8	520.5	--	1.0	7.2	14.3	18.9	--	1.0	8.2	22.5	41.3	--	KHA
10	6.4789	6.2144	88.4	183.0	98.1	205.3	457.8	--	1.1	9.9	16.2	19.8	--	1.1	11.1	27.3	47.1	--	KHA
11	6.4267	6.1898	131.2	272.5	581.8	141.6	2496.0	--	1.2	6.7	19.7	36.4	--	1.2	7.9	27.6	64.0	--	KHA
12	6.3561	6.2456	63.6	224.8	37.4	358.0	605.6	610.0	0.8	3.1	8.6	11.8	18.0	0.8	3.9	12.4	24.2	42.2	KHAA
13	6.4566	6.3257	150.3	403.2	197.2	702.0	3432.5	--	1.4	10.2	32.4	20.0	--	1.4	11.5	43.9	63.9	--	KHA
14	6.5213	6.2984	1022.5	434.7	2644.7	3500.1	1162.4	--	1.2	6.2	13.3	21.9	--	1.2	7.3	20.7	42.5	--	HAK
15	6.3333	6.3533	163.0	428.1	209.8	700.8	211.7	--	1.1	9.6`	19.6	18.4	--	1.1	7.3	20.7	42.5	--	KHA
16	6.5037	6.2682	403.3	234.5	261.9	533.8	1699.1	--	0.7	9.3	19.1	19.4	--	0.7	10.7	29.1	48.5	--	HAA

Table 2: Estimated geohydraulic properties

VES No.	Long. (^{O}E)	Lat. (^{O}N)	Aquifer resistivity (Ωm)	Aquifer thickness (m)	Porosity	Hydraulic conductivity	Transmissivity	Storativity	Diffusivity
1	6.1667	6.1667	74.0	26.9	31.96	0.1752	4.71	0.0807	58.38
2	6.1878	6.3221	298.8	30.1	30.61	0.1682	5.06	0.0903	56.07
3	6.1990	6.2155	1941.7	17.3	28.88	0.1594	2.76	0.0519	53.12
4	6.2121	6.2466	464.1	25.2	30.19	0.1661	4.19	0.0756	55.36
5	6.1768	6.1769	668.4	19.0	29.85	0.1644	3.12	0.0570	54.78
6	6.2897	6.2455	1075.2	19.2	29.41	0.1621	3.11	0.0576	54.04
7	6.4568	6.3243	731.4	19.0	29.77	0.1639	3.12	0.0570	54.64
8	6.3212	6.2787	345.2	31.2	30.47	0.1675	5.23	0.0936	55.84
9	6.2899	6.3435	360.8	18.9	30.43	0.1673	3.16	0.0567	55.77
10	6.4789	6.2144	205.3	19.8	30.96	0.1701	3.37	0.0594	56.68
11	6.4267	6.1898	141.6	36.4	31.32	0.1719	6.23	0.1092	57.30
12	6.3561	6.2456	605.6	18.0	29.94	0.1648	2.97	0.0540	54.94
13	6.4566	6.3257	702.0	20.0	29.81	0.1641	3.28	0.0600	54.70
14	6.5213	6.2984	3500.1	21.9	28.36	0.1567	3.43	0.0657	52.23
15	6.3333	6.3533	700.8	18.4	29.81	0.1641	3.02	0.0552	54.71
16	6.5037	6.2682	533.8	19.4	30.06	0.1654	3.21	0.0582	55.14

The observed variation in resistivity in the aquifer layer can be attributed to the density, shape, size, pore size and porosity of the aquifer units. The aquifer geohydraulic parameters estimated using the combinations of aquifer resistivity, thickness include aquifer porosity (ϕ), hydraulic conductivity (K), transmissivity (Tr), storativity and diffusivity as shown in Table 2. The contour maps (Figures 4 and 5) show the variation of aquifer resistivity and thickness. The variation of resistivity in the aquifer layer may be attributed to the density, shape, size, pore size and porosity of the aquifer units. Figure 4 shows high aquifer resistivity in the northeastern part of the study area while low resistivity spread across the study area. Figure 5 show high values of aquifer thickness in the southeastern part of the study area.

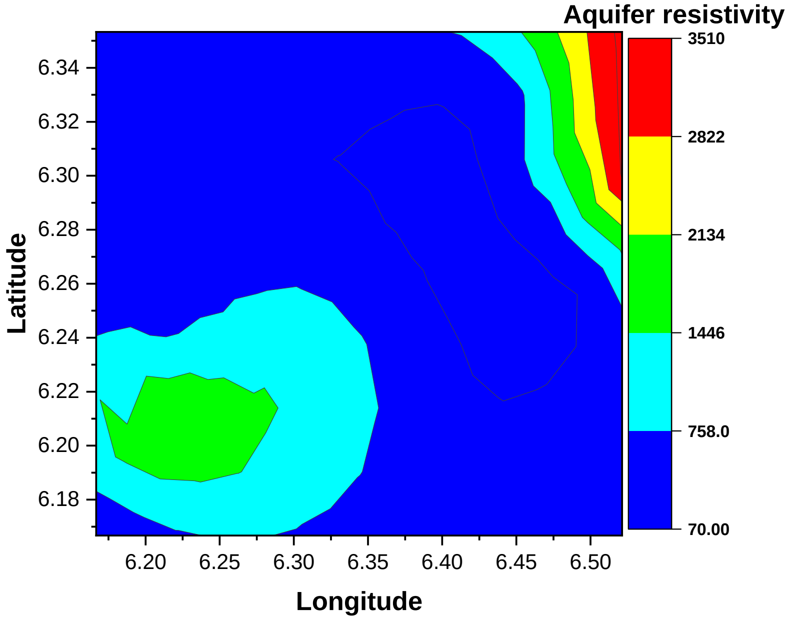

Figure 4: Contour map of aquifer resistivity

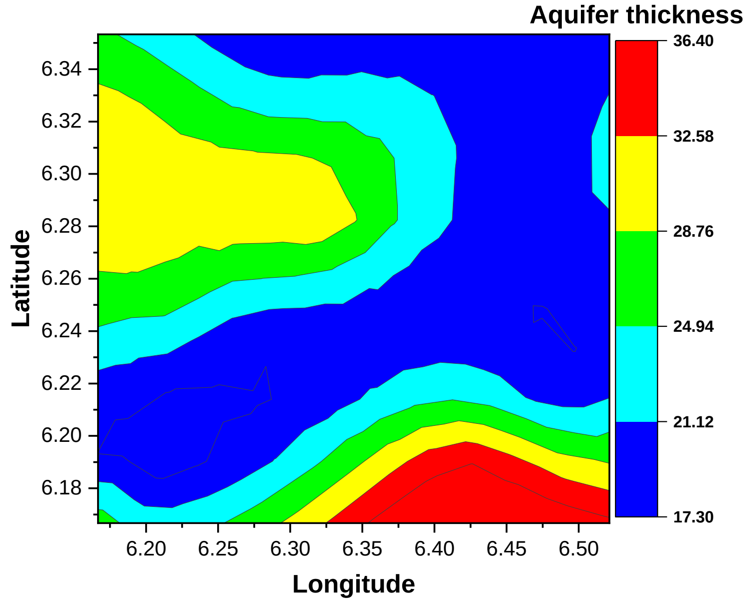

Figure 5: Contour map of aquifer thickness

The aquifer porosity ranged from 28.88 to 31.96 %, this indicates that the aquifer layer is composed of fine-coarse grain sand. The distribution of porosity (Figure 6) shows high porosity in the southern part of the study area, and this corresponds to area with low resistivity. This may be due to the high argillite–sand mixing ratio which reduces pore–matrix ratios in aquifers (Ibanga and George, 2016). Based on the porosity values, the lithology of the aquifer layer may be interpreted as sand, and sandstone (Roscoe, 1990). The hydraulic conductivity which measures the saturated soil's ability to transmit groundwater when subjected to a hydraulic gradient has values ranging from 0.1567 to 0.1752 m/day. The variation of hydraulic conductivity in Figure 7 show high values of hydraulic conductivity in southern part of the study area and decreases towards other parts of the study area. It may be inferred that high k_h values indicate areas with permeable subsurface materials through which water can easily pass and these areas may be lithologically classified as fine-coarse grained sand, and sandstone dominated area.

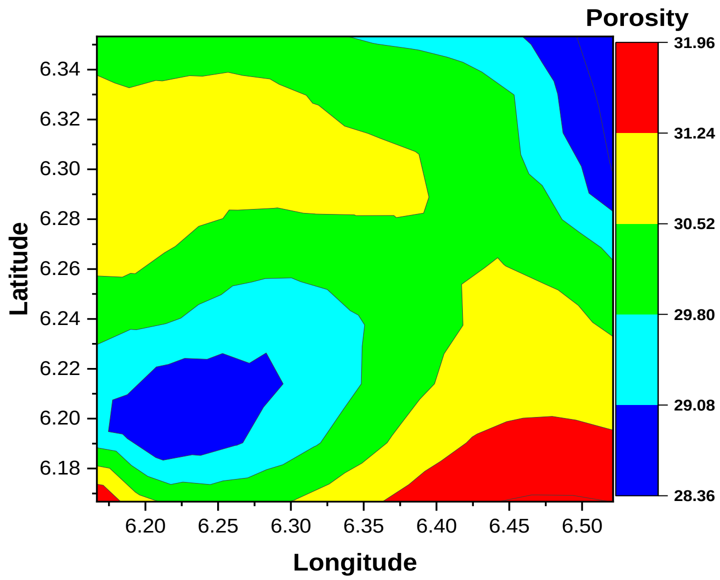

Figure 6: Contour map of aquifer porosity

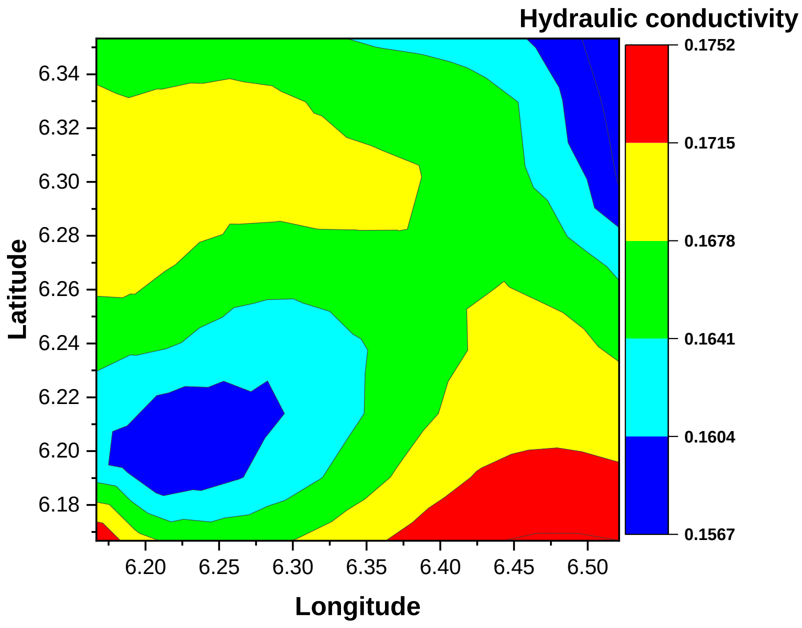

Figure 7: Contour map of aquifer hydraulic conductivity

The aquifer layer is also characterized by transmissivity ranging from 2.76 to 6.23 m^2/s Figure 8 shows the spatial variation of transmissivity with high values observed in the southeastern part of the study area corresponding to zone with low resistivity, high porosity and high hydraulic conductivity. This indicates good communication pore channels, high groundwater potential and the presence of materials that are highly permeable to fluid movement (Obiora et al., 2015). The high transmissivity values indicate high transmissivity magnitude, thus reflecting prolific aquifer repositories (Offodile 1983).

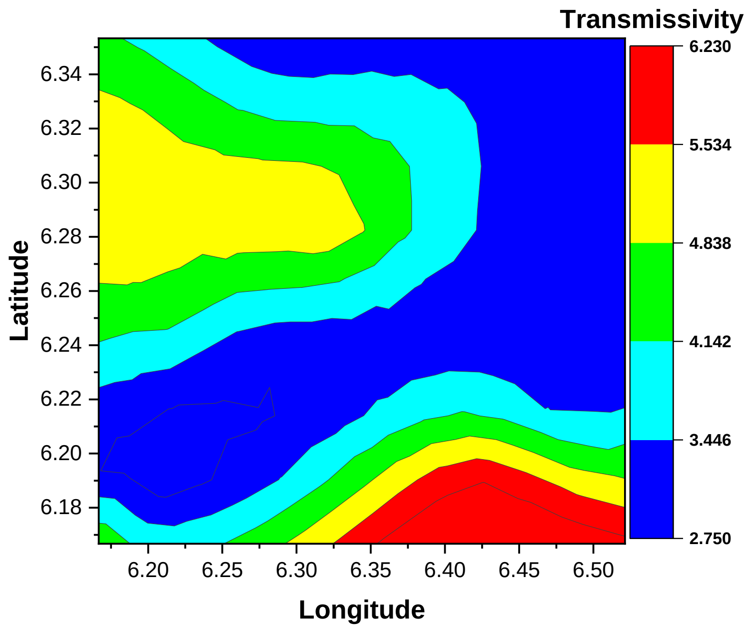

Figure 8: Contour map of aquifer transmissivity

The values of storativity range from 0.0519 to 0.1092 which indicates an unconfined aquifer according to the range (0.01 to 0.3) given by Lohman (1972), and the result of this study agrees with Lohman (1972) range, this reveals that the aquifer in the study area is unconfined. The distribution of storativity (Figure 9) reveals that the southeastern part has high storativity values which is similar to that of thickness. The similarity observed in Figures 5 and 9 shows increase in thickness leads to a corresponding increase in storativity. The values of hydraulic diffusivity range from 52.23 to 58.38, Figure 10 is a contour map showing the variation of diffusivity with high values observed in the southern part of the study area. It is observed that hydraulic diffusivity and hydraulic conductivity have similar trend as reveal in the contour maps, increase in hydraulic conductivity leads to an increase in hydraulic diffusivity.

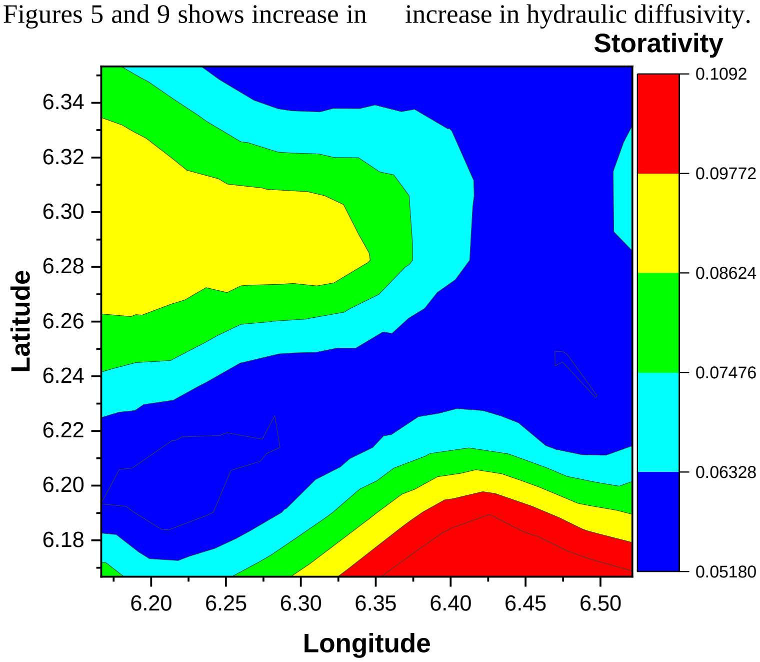

Figure 9: Contour map of aquifer storativity

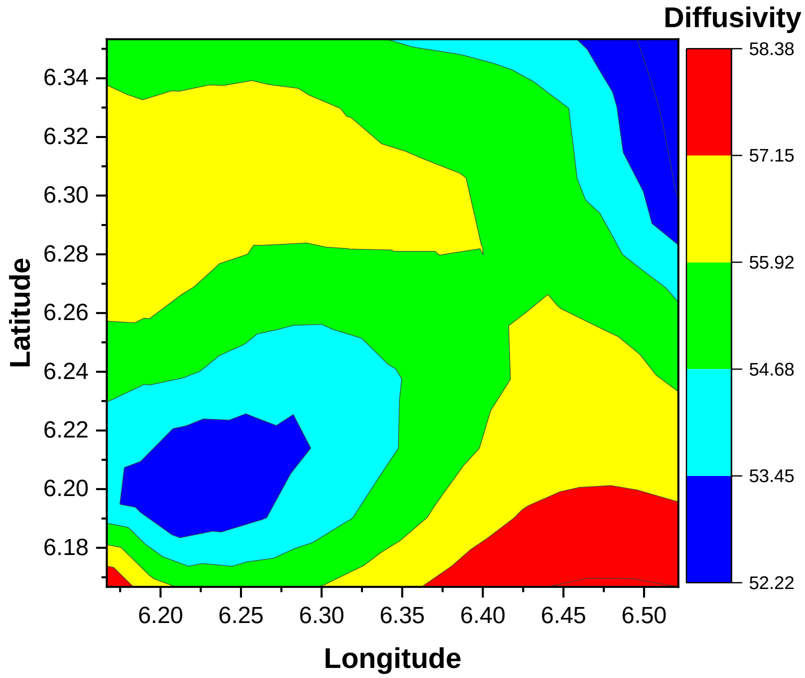

Figure 10: Contour map of aquifer diffusivity

CONCLUSION

This paper demonstrates the importance of using the VES data in characterizing the aquifer repositories in terms of geohydraulic parameters. The results from the study delineate the aquifer units to be unconfined and show wide variations of the measured and estimated geohydraulic parameters across the study area as a result of inhomogeneity of arenaceous geological water repositories. The results revealed that the northeastern part of the study area has high bulk aquifer, while porosity, transmissivity, hydraulic conductivity and storativity are high in the southeastern part of the study area and demonstrates the study area as having prolific aquifer repositories. The quantitative estimated parameters proved significant and helpful in understanding the geohydraulic response of the subsurface aquifer and enhance a better understanding of the geohydrodynamics of study area. The distribution of these parameters across the study area reflects lithological discontinuities of the subsurface. The results showed good prospect for groundwater exploration and management.

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