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Evaluation and Analysis of Geoelectric Parameters of Abeokuta South Local Government Area, Ogun State, South West Nigeria

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Abstract

Geostatistics is an invaluable tool that can be used to characterize spatial or temporal phenomena. In both mining and petroleum industries, it is successfully applied to solve cases where decisions concerning expensive operations are based on interpretations from sparse data located in space. It involves data analysis and spatial continuity modeling to establish quantitative measure of spatial correlation to be used for subsequent estimation and simulation. In this work, the data analysis aspect of geostatistics is applied in which geostatistical analysis of geoelectric parameters of Abeokuta South Local Government Area was carried out using data obtained from twenty–eight (28) Vertical Electrical Soundings (VES) conducted over the area using the Schlumberger configuration. Basic parameters analyzed include resistivity and thickness of topsoil, resistivity and thickness of aquifer layer, resistivity of bedrock, overburden thickness, longitudinal unit conductance, hydraulic conductivity and transmissivity. High transmissivity values (0.05×10^4 - 2.14×10^4 m^2s^{-1}) and low ranges of values of the longitudinal unit conductance (0.0015 – 0.0521 S), and hydraulic conductivity (15.21×10^{-3} - 231.9×10^{-3} ms^{-1}) were recorded. The low values of longitudinal unit conductance and hydraulic conductivity could be attributed to the clay content in the aquifers which affect the porosity of the layer. Resistivity of weathered layer has a low range of values between 186.60Ωm and 252 Ωm, with a high standard deviation value of 371.81Ωm. The range of values of layer parameters determine to a significant extent the yield or prolific nature of boreholes. Weathered and fractured horizons which constitute the aquifer zones underlying VES stations have hence been identified in the study area. Analysis showed that the parameters are generally positively skewed so also is the kurtosis generally positive indicating fair distribution about the mean; an indication that there is a fair distribution of groundwater resources within the area and fair all-year-round availability of groundwater supply.

Keywords: Geoelectric parameters, geostatistical analysis, VES, Schlumberger configuration, weathered and fractured horizons.

1.0 Introduction

Water is a natural and essential resource to life on earth. Water is naturally bounteous in proportion with its quality of transformation through perennial hydrogeological evaporation, condensation and precipitation [1]. Surface water, which mostly occurs as rivers, is the most common source of water supply in rural areas; it is however subjected to pollution. Most of the rivers in Nigeria are highly polluted, the pollutants being introduced by man through industrial and waste generation activities. As demand for water resources increases and the variety of pollutants becomes more diverse [2], there is an increasing conflict between the use of rivers for water supply and 'sewers' for disposal of industrial and domestic effluent.

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Hence, ground water exploration is the only alternative to surface water in order to overcome the variety of pollutants. Although, the quantity and disposition of ground water depends on the geological characteristics of the host rock formation; the search for ground water is faced with lots of uncertainties. To minimize or avoid failures altogether, it is pertinent that the right exploration techniques are utilized in the delineation of subsurface water-bearing formations. In crystalline basement rocks such as are found in the study area, electrical resistivity method is usually found most suitable in the study of underground structures for the identification of water bearing layers [3].

2.0 The Study Area

The study area is located in Abeokuta, between latitudes 7°12.334' and 7°20.904'N, and longitudes 3°32.962' and 3°45.911'E (Figs. 1 and 2). It lies within the South-Western part of the Nigerian Precambrian basement complex, which occupy approximately 50% of the surface area of the country, as part of the Pan African crystalline shield. The dominant rock types in the study area are characterized by various rock types ranging from granite, granitic gneiss and associated rock suites; common metamorphic rocks encountered in all the major outcrops in the study area include pegmatites, gneiss, schist, quartzite and amphiboles. Generally, wells from quartzite areas produce more water than wells from other rock types. This is because their transmissivites and permeability are higher due to the presence of fissures and quartz veins [4].

3.0 Basic Principles of Electrical Resistivity Method

The earth is assumed to be an ideally isotropic and homogeneous medium through which electrical current can flow. In developing an equation expressing the potential about a single point source current, Ohms law is adopted. In its application to the solid earth, the law considered the resistance of a cylindrical earth material of length L and cross-sectional area, A.

The resistance of the material is expressed as:

$$\boldsymbol{R}=\frac{\rho L}{A} \qquad (1)$$

where $\boldsymbol{\rho}$ is called the resistivity of the material.

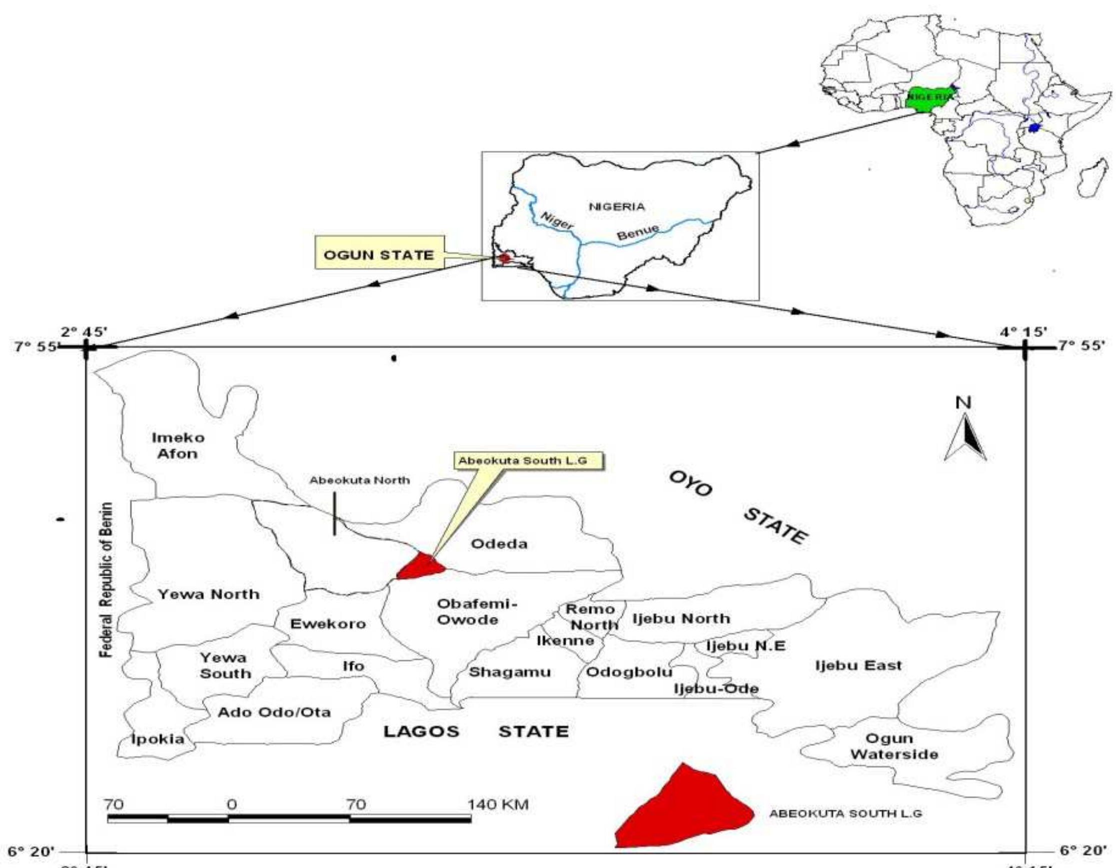


Fig. 1: Inset Map showing the Study Area in Ogun State (using Esri Data/Nigeria political Information in Arcview GIS 3.2A Environment)

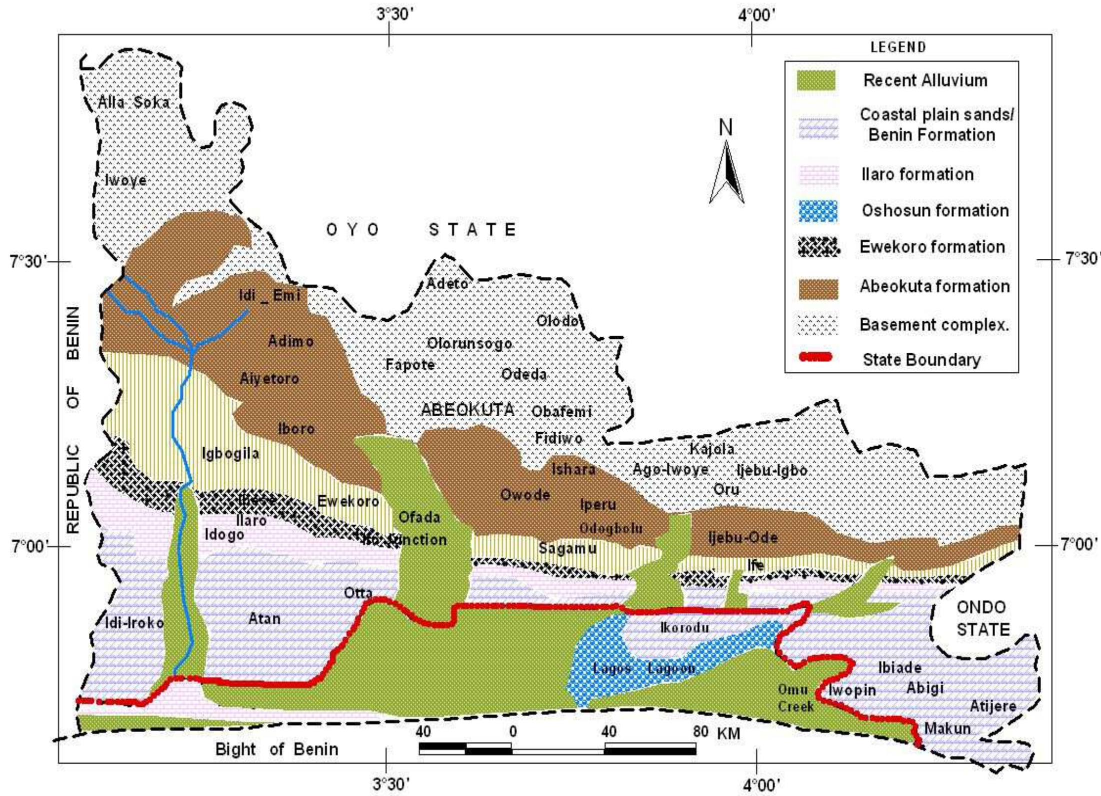


Fig. 2: Map of Ogun State showing the Geology of the Study Area (Modified after [5, 6])

If a potential difference ΔV**,** is maintained across the ends of the material, then from ohms law, the current I flowing in the earth material is $I = \frac{\Delta V}{R}$. But $R = \frac{\Delta V}{I}$, hence

$\frac{\Delta V}{I} = \frac{\rho L}{A}$, that is,

$$\rho = \frac{\Delta V A}{L I} \qquad (2)$$

Equation (2) gives the resistivity of a homogeneous and isotropic medium provided the geometry is simple such as cube, cylinder and parallel pipe [7].

If a current *I* is passed into the ground through an electrode, the potential difference across an hemispherical shell of radius *r* is given as

$$\int \Delta V = \frac{\int I\rho dr}{2\pi r}$$

$$V = \frac{I\rho}{2\pi r} \qquad (3)$$

Equation (3) is vital in the development of the resistivity method.

ρ in equation (3) is the true resistivity of the earth medium. However, if the medium is inhomogeneous, then ρ becomes the apparent resistivity.

3.1 Field Layout of the Schlumberger Electrode Configuration

Figure 3 shows the field arrangement of the Schlumberger four electrode configuration. A and B are current electrodes while M and N are potential electrodes.

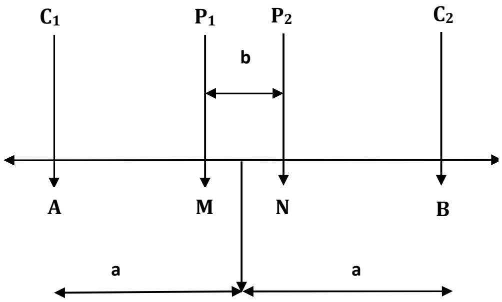


Fig. 3: Field Arrangement of the VES Schlumberger 4-Electrode Configuration

In general, for the Schlumberger array,

$\boldsymbol{\rho = G\,R}$ where $\boldsymbol{G}$ is the geometric factor of the electrode array given as

$$\mathbf{G} = \boldsymbol{\pi}\left(\frac{\mathbf{a}^2}{\mathbf{b}} - \frac{\mathbf{b}}{\mathbf{4}}\right) \quad (4)$$

$$\boldsymbol{\rho}_{\mathbf{a}} = \mathbf{GR} = \boldsymbol{\pi}\mathbf{R}\left(\frac{\mathbf{a}^2}{\mathbf{b}} - \frac{\mathbf{b}}{\mathbf{4}}\right) \quad (5)$$

where **a** is half array length; **b** is the minimum spacing between the potential electrodes, and $\boldsymbol{\rho}$ is the true resistivity when the ground is homogeneous and isotropic but when the ground is not homogenous (that is, heterogeneous) $\boldsymbol{\rho}$ becomes the apparent resistivity **($\boldsymbol{\rho}_{\mathbf{a}}$)**.

Layering and fracturing often affect resistivity measured in any environment, thus; there is no uniformity in the flow of electric current [7]**.**

4.0 Methodology

In locating and delineating subsurface water resources in hard rock terrains, several geophysical techniques comes in handy [8]; they provide useful information on the subsurface geology of a region without the large cost of an extensive drilling programme. According to [9], the electrical resistivity method, of all the surface geophysical methods, has been most widely applied in groundwater exploration studies. This is so because it is capable of clarifying the subsurface structure, delineating subsurface geology for engineering construction, delineating groundwater zones and is inexpensive [10 - 13].

The electrical resistivity method can be best employed to estimate the thickness of overburden and weathered/fractured zones with reasonable accuracy [4, 14]. For this work, electrical resistivity method was used because of its numerous advantages over other methods in groundwater investigation. Twenty-eight (28) stations were probed in the study area (Fig. 4) using Vertical Electrical Sounding (VES) with Schlumberger electrode configuration (Fig. 3).

Geostatistics is a set of statistical estimation tools involving quantities which vary in space. It is used extensively in almost all branches of hydrosciences, geography, epidemiology, commerce, military, planning (logistics), and the development of efficient spatial networks [15]. Geostatistical algorithms are incorporated in many places, including geographic information systems (GIS) and statistical environment. As a collection of all statistical and probabilistic methods applied to sciences of earth, geoscientists often face interpolation and estimation problems when analyzing sparse data from field observations [16]. Geostatistics is therefore an invaluable tool that can be used to characterize spatial or temporal phenomena. In both mining and petroleum industries, geostatistics is successfully applied to solve cases where decisions concerning expensive operations are based on interpretations from sparse data located in space. Geostatistics has since been extended to many other fields in or related to the earth sciences such as hydrogeology, hydrology, meteorology, oceanography, geochemistry, geography, soil sciences, forestry, and landscape ecology. Geostatistics therefore involve data analysis and spatial continuity modeling to establish quantitative measure of spatial correlation to be used for subsequent estimation and simulation [17]. Tools used in geostatistics are

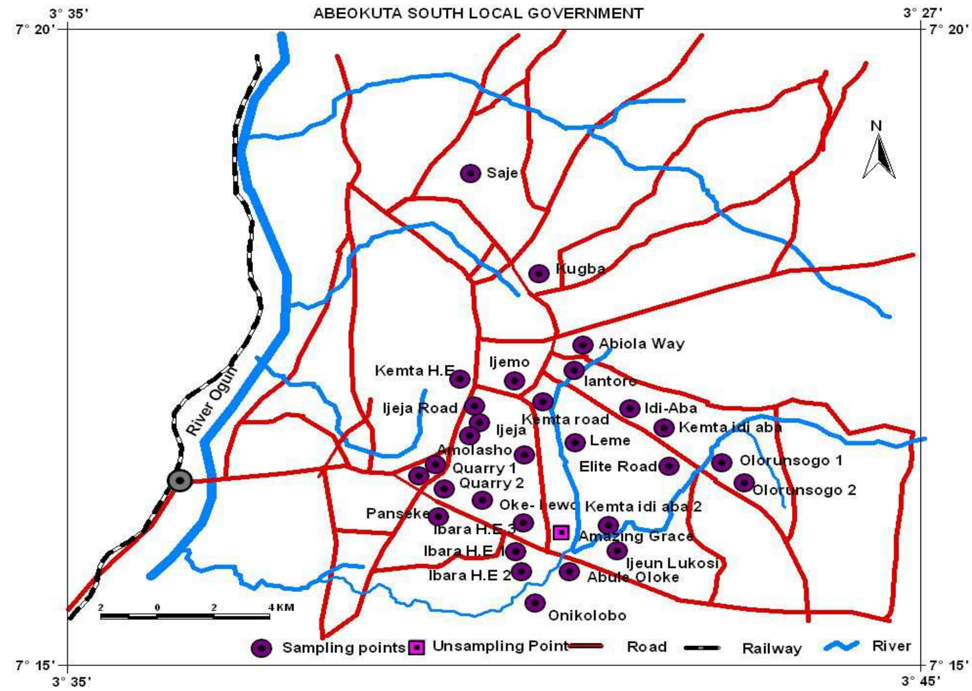


Fig. 4: Location Map of the Study Area showing the Sounding and Sampling Points

Regionalized Variable Theory, covariance function, mean, semi-variance, variogram, kriging, range, among others [18]. In this work, the data analysis aspect of geostatistics is applied.

Classical statistics is generally devoted to the analysis and interpretation of uncertainties caused by limited sampling of a property under study. Geostatistics however deviate from classic statistics in that it is not tied to a population distribution model that it assumes, for example, all samples of a population are normally distributed independent from one another. Earth science data mostly such as rock properties, contaminant concentrations, hydraulic conductivities, permeability among others, do not satisfy these assumptions as they can be highly skewed and/ or possess spatial correlation that is, data values from locations that are closer together tend to be more similar than data values from locations that are farther apart. To most geologists, the fact that closely spaced samples tend to be similar is not surprising since such samples are influenced by similar physical and chemical depositional/transport process [19].

5.0 Estimation of Geoelectric Parameters

Anisotropy coefficient is a measure of inhomogeneity of a medium [20]; it increases linearly with increase in groundwater yield. In theories about stratified conductors, certain parameters are fundamentally important both in the interpretation and understanding of the geoelectrical model consisting of stratified conductors. These parameters are related to different combinations of the thickness and resistivity of each geoelectric layers in the model [14]. Considering geo-electrical section with a unit cross-sectional area, the combination of the thickness and resistivity of the geoelectric layers into single variables; the Dar-Zarouk parameters of Transverse unit resistance (R) and Longitudinal unit conductance (S), can be used as a basis for the evaluation of aquifer properties such as transmissivity (T) and coefficient of Anisotropy.

For a horizontal, homogenous and isotropic layer, the Dar-Zarouk parameters of transverse unit resistance and longitudinal unit conductance are given in [10, 14] as:

$$H = \sum_{i=1}^{n} h_i \qquad (6)$$

The longitudinal conductance S is given as

$$S_i = \sum_{i=1}^{n} \frac{h_i}{p_i} \qquad (7)$$

where H is the summation of thickness while the transverse unit resistance R is given as

$$R_i = \sum_{i=1}^{n} h_i p_i \qquad (8)$$

From equation (6) and (7) the longitudinal resistivity is

$$\rho_l = \frac{H}{S} = \frac{\sum h_i}{\sum h_i/p_i} \qquad (9)$$

From equation (6) and (8) the transverse resistivity is

$$\rho_t = \frac{R}{H} = \frac{\sum h_i p_i}{\sum h_i} \quad (10)$$

The Anisotropic coefficient $(\lambda) = \sqrt{\frac{\rho_t}{\rho_l}}$ (11)

$$(\lambda) = \sqrt{\frac{T}{H} \cdot \frac{S}{H}}. \quad (12)$$

T and S are known as Dar-Zarrouk parameters.
For an isotropic medium $\rho_t = \rho_l$ such that $\lambda = 1$
For an anisotropic medium $\rho_t > \rho_l$ such that $\lambda > 1$
Equation (12) is used for layered rocks such as sedimentary rocks and it is also applicable to basement complex rocks that shows layered structure [21].
The aquifer transmissivity (T) is expressed as the product of the hydraulic conductivity (K) and layer thickness (h), that is,

$$T = Kh \quad (13)$$

For clean saturated aquifers whose natural fluid characteristics are fairly constant (that is, no appreciable impact on the general ground water quality by surface contaminants loads), the hydraulic conductivity is proportional to the resistivity of the aquifer. This implies that in the absence of a pumping test data, the aquifer hydraulic conductivity K can be approximated to the true resistivity of the aquifer derived from geoelectric investigation [22]. Therefore,

$$T = Kh = \rho h \quad (14)$$

The product of the resistivity to its thickness is the transverse resistance (R), which is numerically equal to the transmissivity (T). Hence,

$$T = R = \sum_{i=1}^{n} h_i p_i \quad (15)$$

The longitudinal conductance (S) gives a measure of the impermeability of a confining clay/shale layer. Such layers have low hydraulic conductivity (K) and low resistivity.

5.1 Geostatistical Analyses of the Geoelectrical Parameters

The population mean of a random variable y is defined as the mean of all possible values of y and is denoted by **μ**. The mean is also referred to as the expected value of y, or **E(y).** If the density f**(y)** is known, the mean can sometimes be found but if **f(y)** is unknown, the population mean **μ** will ordinarily remain unknown unless it has been established in the past experience with a stable population. If a large random sample from the population represented by **f(y)** is available, it is highly probable that the mean of the sample is close to **μ**. The sample mean of a random sample of n observations in each sounded location y_i $(y_1, y_2, \ldots, y_n)$ is given by the ordinary arithmetic average.

$$\bar{y} = \frac{1}{n} \sum_{i=1}^{n} y_1 \quad (16)$$

Generally, $\bar{y}$ will never be equal to **μ**; by this we mean that the probability is zero that a sample will ever arise in which $\bar{y}$ is exactly equal to **μ.** However, $\bar{y}$ is considered a good estimation for **μ** because **E(y)** = **μ** and **var(**$\bar{y}$) = **μ** hence

$$\mathbf{var}(\bar{y}) = \sigma^2/n \quad (17)$$

where σ^2 is the variance of y.
In other words, $\bar{y}$ is an unbiased estimator of **μ** and has a smaller variance than a single observation y. the notation **E(y)** indicates the mean of all possible values of $\bar{y}$; that is, conceptually, every value at each sounded point is obtained from the entire formation in the hydrogeological environment of the study area, the average of all these aquifer parameters are calculated.
If every y in the entire population is multiplied by constant **a,** the expected value is also multiplied by a:

$$\mathbf{E(ay) = aE(y) = a\mu} \quad (18)$$

The sample mean has a similar property. If $z_i = ay_i$ for i=1, 2, ……………, n, then

$$\hat{z} = a\bar{y} \quad (19)$$

The variance of the population is defined as

$$\mathbf{var(y)} = \sigma^2 = \mathbf{E}(\mathbf{y} - \mu)^2 \quad (20)$$

Equation (20) is the average squared deviation from the mean and is thus an indication of the extent to which the values of y (geoelectric properties) are spread, distributed, dispersed and scattered in the groundwater formation of the study area. It can be shown that

$$\sigma^2 = \mathbf{E}(y^2) - \mu^2 \quad (21)$$

The sample variance is defined as

$$S^2 = \frac{\sum_{i=1}^{n}(y_i - \bar{y})}{n-1} \quad (22)$$

Equation (22) can be expressed further to be equal to

$$S^2 = \sum_{i=1}^{n} \frac{y_i{}^2 - n\bar{y}^2}{n-1} \tag{23}$$

The sample variance S^2 is generally never equal to the population variance σ^2 (the probability of such an occurrence is zero, but it is an unbiased estimator for σ^2) where the notations $\mathbf{E}(S^2)$ indicates that the mean of all possible sample variances. The square root of either the population variance or sample variance is called the standard deviation.

Skewness is a measure of asymmetry (or departure from symmetry) of a distribution. For a distribution in each sounded location y_i (y_1, y_2, ….., y_n) having corresponding frequencies f_i (f_1, f_2, ……. f_n) and a mean of $\bar{y}$ and a standard deviation σ, we define skewness α_3 as :

$$\alpha_3 = \frac{1}{N\sigma^3} \sum_{i=1}^{1} f_i (y_i - \bar{y})^3 \tag{24}$$

If a distribution is symmetrical about the mean, the skewness is equal to zero; because to every positive value in (y_i- $\bar{y}$), there will be an equal and negative value. If these negative values are cubed they would remain negative and consequently the sum of all the values would certainly be zero. In other words, $\alpha_3 = 0$, if there is symmetry about the mean. Therefore any data in which α_3 is positive is said to have a positive skewness, and any data in which α_3 is negative is said to have negative skewness [23].

Kurtosis is the measure of peakedness of a distribution. For a distribution in each sounded location y_i (y_1, y_2, ….., y_n) discrete members having corresponding frequencies f_i (f_1, f_2, ……. f_n) and if the mean and standard deviation $\bar{y}$ **and** σ respectively , then the quantity α_4 , called kurtosis is defined as:

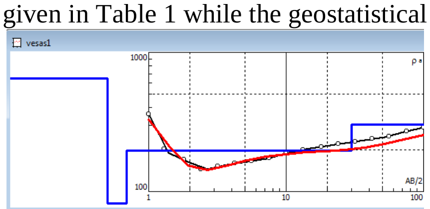
$$\alpha_4 = \frac{1}{N\sigma^4} \sum_{i=1}^{1} f_i (y_i - \bar{y})^4 \tag{25}$$

For a normal curve $\alpha_4 = 3$, and we say a normal curve has zero kurtosis [23].

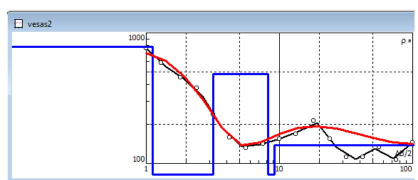
Given that $\sqrt{\beta_1}$ and β_2 are population skewness and kurtosis parameters respectively, in a normal distribution, $\sqrt{\beta_1}$ = 0, and β_2 =3. If $\sqrt{\beta_1}$ < 0, the skewness is negative, and if $\sqrt{\beta_1}$ > 0 the skewness is positive. If β_2 < 3, kurtosis is negative, and if β_2>3, there is positive kurtosis.

6.0 Results and Discussion

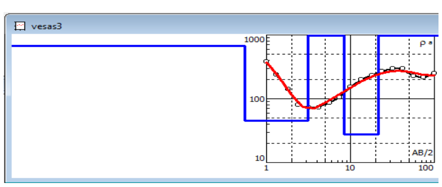
Calculations and geostatistical analysis of the geoelectrical parameters were carried out on the twenty-eight Schlumberger VES data obtained from the study locations. Basic parameters analyzed include resisitivity and thickness of topsoil, resistivity and thickness of aquifer layer, bedrock relief, overburden thickness, longitudinal unit conductance, hydraulic conductivity, and transmissivity. Figure 5 shows typical VES profiles obtained over some stations investigated while figure 6 shows typical geoelectric sections obtained over the same stations. Result of the calculations of the geoelectric parameters are given in Table 1 while the geostatistical analyses of the parameters are given in Table 2.



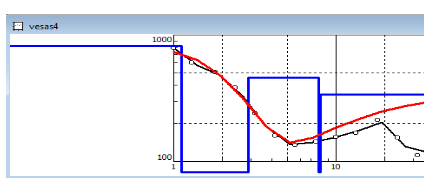
VESAS1



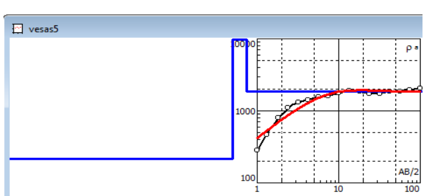
VESAS2



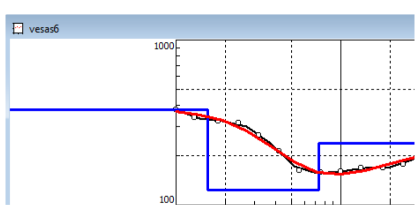
VESAS3



VESAS4



VESAS5



VESAS6

Figure 5: VES Plots over AS1 – AS6

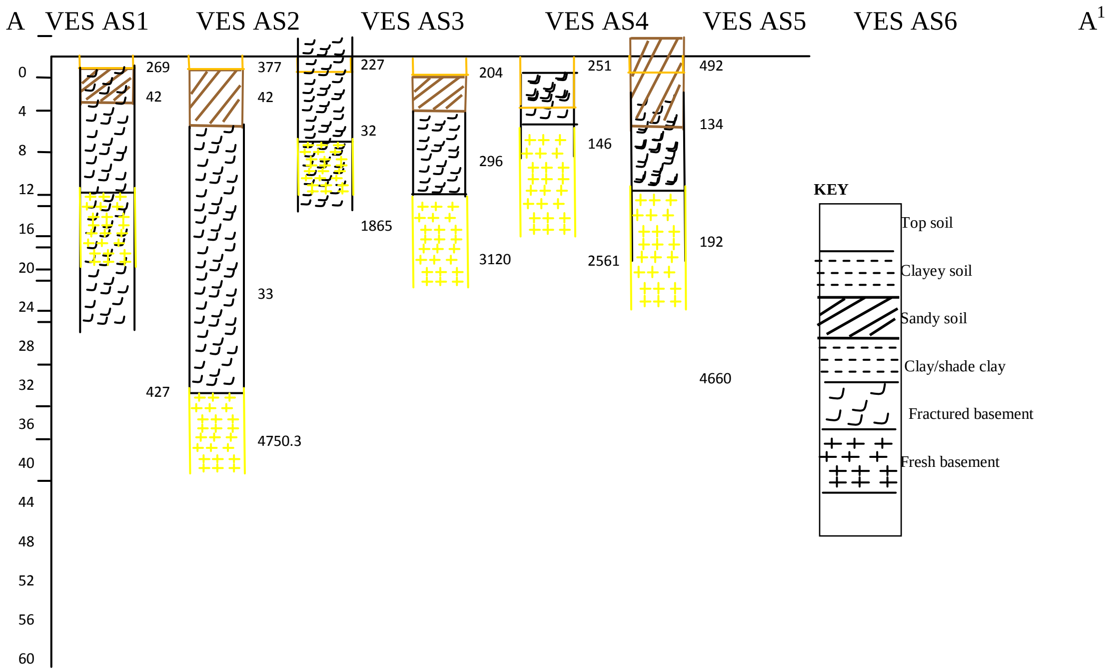


Figure 6: Geoelectric section beneath VES AS1- AS6

Table 1: Geoelectric Parameters Analysed over the Study Area

VES Station	Top Layer Resistivity (Ωm)	Top Layer Thickness (m)	Weathered Layer Resistivity (Ωm)	Bedrock Relief (m)	Aquifer Thickness (m)	Overburden Thickness (m)	Longitudinal Resistance (Ω)	Coefficient of Anisotropy (λ)	Hydraulic Conductivity (m/s)	Transmissivity $\times 10^{-5}$ (m^2/s)	Longitudinal Conductance (S)
VES AS1	208.5	1.0	321.00	61	12.70	14.45	1389.42	0.9990	78.46×10^{-3}	0.41×10^{4}	0.0104
VES AS2	377.2	1.1	33.50	25	19.60	15.60	1053.89	0.9998	47.86×10^{-3}	0.07×10^{4}	0.0334
VES AS3	226.7	1.0	32.30	125	11.40	2.00	2123.29	1.0002	80.75×10^{-3}	0.04×10^{4}	0.0058
VES AS4	204.0	0.7	880.40	71	6.90	3.65	4591.84	0.9999	144.3×10^{-3}	0.61×10^{4}	0.0029
VES AS5	251.3	1.0	221.70	57	14.90	27.70	8457.37	1.0000	67.18×10^{-3}	0.33×10^{4}	0.0072
VES AS6	491.6	1.5	192.10	11	21.50	11.00	1284.58	0.9996	46.84×10^{-3}	0.41×10^{4}	0.0253
VES AS7	233.0	1.1	271.00	149	1.70	4.10	1213.12	1.0004	54.20×10^{-3}	0.05×10^{4}	0.0031
VES AS8	735.8	1.3	450.90	138	12.33	8.50	2137.72	0.9999	80.52×10^{-3}	0.56×10^{4}	0.0097
VES AS9	918.7	0.5	122.20	16	63.80	12.33	3185.36	0.9998	15.67×10^{-3}	0.78×10^{4}	0.0239
VESAS10	1095.2	4.9	513.40	114	41.60	4.90	2421.88	0.9999	23.99×10^{-3}	2.14×10^{4}	0.0192
VESAS11	1559.5	1.9	198.90	119	26.70	28.70	1805.03	0.1880	37.53×10^{-3}	0.53×10^{4}	0.0159
VESAS12	790.9	1.3	360.10	134	13.90	13.90	1198.28	0.9980	72.02×10^{-3}	0.50×10^{4}	0.0116
VESAS13	821.8	1.1	145.60	80	24.07	3.60	3561.22	0.9986	41.60×10^{-3}	0.35×10^{4}	0.0098
VESAS14	159.3	1.4	152.08	133	6.76	5.45	459.02	1.0008	15.21×10^{-3}	0.10×10^{4}	0.0268
VESAS15	334.2	0.6	122.80	59	31.90	5.10	2746.38	1.0008	31.49×10^{-3}	0.39×10^{4}	0.0138
VESAS16	88.0	1.1	208.71	81	4.17	3.50	2020.73	0.9976	231.9×10^{-3}	0.09×10^{4}	0.0082
VESAS17	567.8	0.5	139.80	74	6.50	2.80	1882.00	0.9989	155.3×10^{-3}	0.09×10^{4}	0.0044
VESAS18	229.7	1.8	36.42	73	9.00	4.87	492.21	1.0008	121.4×10^{-3}	0.03×10^{4}	0.0521
VESAS19	488.7	2.7	55.60	72	10.80	2.70	9000.00	1.0095	92.67×10^{-3}	0.06×10^{4}	0.0015
VESAS20	566.7	1.5	60.40	27	28.10	2.50	2069.93	0.9985	35.53×10^{-3}	0.17×10^{4}	0.0143
VESAS21	1769.8	0.9	45.30	26	12.87	2.90	1069.93	0.9828	75.50×10^{-3}	0.06×10^{4}	0.0146
VESAS22	45.9	1.0	187.50	50	13.70	2.60	568.30	1.0020	72.12×10^{-3}	0.26×10^{4}	0.0265
VESAS23	232.8	1.3	75.36	39	8.76	6.30	1612.95	0.9994	125.6×10^{-3}	0.06×10^{4}	0.0139
VESAS24	448.4	1.3	224.57	50	14.37	8.05	660.82	1.0003	70.18×10^{-3}	0.32×10^{4}	0.0342
VESAS25	122.6	0.7	64.70	29	19.70	2.90	2037.50	1.0003	49.77×10^{-3}	0.13×10^{4}	0.0080
VESAS26	224.2	2.2	39.30	26	13.50	2.80	2032.25	1.0067	78.60×10^{-3}	0.05×10^{4}	0.0062
VESAS27	1382.3	0.5	39.30	21	5.50	7.10	2032.25	1.0067	196.5×10^{-3}	0.02×10^{4}	0.0062
VESAS28	180.0	2.5	184.00	42	9.84	12.70	5769.23	1.0066	102.2×10^{-3}	0.18×10^{4}	0.0039

Table 2: Results of Geostatistical Analysis

AQUIFER PARAMETERS	RANGE	MEAN	MEDIAN	MODE	STANDARD DEVIATION	VARIANCE	SKEWNESS	KURTOSIS
TOPSOIL RESISTIVITY (Ωm)	45.9 – 10769.8	1002.60	412.80	45.90	2487.29	6186622	+5.019	+1.290
TOPSOIL THICKNESS (m)	0.5 - 4.9	1.32	1.10	1.00	0.88	0.78	+2.74	+8.487
AQUIFER RESIST (Ωm)	32 - 880	252.82	148.84	39.30	371.81	138239.40	+3.60	+6.356
AQUIFER THICKNESS (m)	62.10	16.66	13.19	1.70	12.94	167.34	+2.13	+5.789
OVERBURDEN THICKNESS (m)	1.7 - 63.8	9.78	5.00	0.90	13.27	176.00	+3.21	+3.240
BEDROCK RELIEF (m)	16 - 149	71.64	66.23	15.67	41.27	1702.82	+0.44	-0.905
LONGITUDINAL CONDUCTANCE (Siemens)	0.0015 - 0.0521	0.12	0.01	0.01	0.02	0.00	+2.34	+2.374
ANISOTROPY COEFFICIENT	0.1880 - 1.0095	0.94	1.00	1.00	0.24	0.06	-3.59	+27.947
TRANSMISSIVITY (m^2/s)	$0.05 \times 10^4 - 2.14 \times 10^4$	3150	1750	600	4157.06	17281111	+3.34	+14.065
HYDRAULIC CONDUCTIVITY (m/s)	$15.21 \times 10^{-3} - 231.9 \times 10^{-3}$	0.08	0.07	0.00	0.05	0.003	+1.20	-1.416

6.1 Resistivity and Thickness of Topsoil

The value of topsoil resistivity ranges between 45.9 – 10769.8(Ωm). The ranges of resistivity values suggest dissimilarities in the composition of materials constituting the topsoil in the study area. The mean value of topsoil thickness is 1.32m. The low thickness value of the top soil due to the corresponding thinness of the layer is as a result of relatively high resistivity mean value recorded in the first layers.

6.2 Resistivity and Thickness of Weathered Layer

The resistivity of weathered layer has a low range of values of between 32 - 880Ωm with an average value of 252Ωm and a low standard deviation value 371.81Ωm. The low range in value of the layer determines to a significant extent the yield of boreholes [24]. The resistivity of the weathered layer suggests lithology in the suite of sandy clay, clayey sand and shale/clay to be widespread. Hydro-geologically, the weathered layer is relevant in groundwater prospecting when it is thick enough, above minimum thickness of 10m suggested in [25], the layer could support hand dug well [26]. In this study area, the average value is 16.6m with a variance of 167.34m which is relatively high and measures the degree of clustering of the data around the mean.

6.3 Bedrock Relief and Bedrock Resistivity

The range of values of bedrock relief is between 16m and 149m and a mean value of 71.64m in the study area. In consonance with bedrock resistivity, high values are recorded in almost all the study locations. According to [27], the resistivity values that exceed 1000Ωm are of fresh bedrock and if less, the bedrock is fractured and saturated with fresh water. Therefore, it could be stated that areas with low bedrock resistivities as well as low bedrock reliefs are the areas of appreciable prospects in terms of groundwater abstraction.

6.4 Longitudinal Unit Conductance (S)

Longitudinal Unit Conductance is calculated using equation (7). According to [7], conductance is related to clay content which increases the porosity of a layer but decreases its permeability. Since permeability decreases with an increase in conductance, the overburden might therefore have absorption and retention capacity. In the study area, the conductance has a range of values between 0.0015 and 0.0521 with an average value of 0.0120 Siemens; it is very low in most of the locations. Many of the locations have low to moderate clay content, thus a decreased porosity of the layer but an increase in its permeability. The longitudinal unit conductance is positively skewed with accompanied positive kurtosis.

6.5 Hydraulic Conductivity (K)

Using the relation K = 95.5x10^{-9} □$^{1.195}$ where □ is the resistivity of the porous layer in Ωm [10], hydraulic conductivity values were found to vary from one aquifer to the other; it ranges from 15.21×10^{-3} m/s - 231.9×10^{-3} m/s, with a mean value of 0.08×10^{-3} m/s.

In general, these values are low except in a few locations where K is moderately high (VES 16 and VES 27) as shown in Table 1. The low values recorded could be attributed to the clay content in the aquifers and the fact that the degree of hydraulic conductivity between the fractures is low. The mean value of K is very low (0.08×10^{-3} m/s), and K is positively skewed with value 1.20 as a result of low departure from the symmetry. The scatter is also small with a very low value of standard deviation 0.05. Also, the negative kurtosis recorded here signified that K is flatter or less peaked than the normal distribution. The less peaked of the hydraulic conductivity in overall geoelectric parameters analyzed revealed that the hydraulic conductivity in the hydrogeological environment of the study area is considerably or moderately high; this probably is attributed to the clay content in the aquifers of the study area thereby lowering the ease or rate of conductivity between the fractured and weathered layer.

6.6 Coefficient of Anisotropy (λ)

In Abeokuta South LGA, the range of coefficient of Anisotropy (λ) value is between 0.1880 and 1.0095 with mean value of 0.94 and has a mode of 1.00 (Table 2). Generally, the coefficient of Anisotropy (λ) is 1 and does not exceed 2 in most of the geological conditions [14]. Compact rock at shallow depth increases the coefficient of Anisotropy [21]. Coefficient of Anisotropy value (λ) greater than 2 indicates very hard porphyritic granite gneiss and garnet biotite gneiss terrain of the geological environment while areas whose values are less than 1.5 are considered as potential aquifers for groundwater exploitation. Alluvial aquifers, fractured zones and valley fills with anisotropy value around 1 are recognized as good groundwater potential zones [28]. It is noteworthy that in all the study areas, coefficient of Anisotropy is around 1 and is negatively skewed with low value of Standard deviation (SD) depicting a low degree of distortion or dispersion from the normal value.

6.7 Transmissivity (T)

It is the rate at which water flows through a vertical strip of the aquifer of unit width and extending to full saturated thickness under hydraulic gradient 1.00. Using $T = Kh$, where K is the coefficient of conductivity (m/s), and h is the aquifer thickness (m), the estimates of T obtained in the study area shows that the value ranges from 0.05×10^4 - 2.14×10^4 m^2/s.

In general, it is very high in many of the locations; this may be due to the weathered nature of the basement rock. The average value recorded is 3150m^2/s with a relatively high standard deviation value of 4157.06. It is positively skewed with value 3.34. Lower value of 64.6 m^2/s was reported in typical low yield basement area of Oke-badan Estate area of Akobo in Oyo State, Southwestern Nigeria [4].

A listing of obtained values of the parameters over the entire study area are as earlier shown in table 1, while results of geostatistical analyses using Statistical software packages are as earlier given in Table 2.

7.0 Conclusion

Hitherto, to the best of the authors' knowledge, there has been no previous data of study of the hydraulic characteristics of Abeokuta South LGA. Based on the electrical resistivity survey conducted in the study area, this study has, through geostatistical analysis of data acquired, evaluated the hydraulic characteristics of Abeokuta South LGA. The principal hydraulic characteristics determined for this investigation are longitudinal unit conductance, horizontal hydraulic conductivity and transmissivity. The longitudinal unit conductance value obtained in this study, for Abeokuta South LGA, is between 0.0015 and 0.0521 S, the hydraulic conductivity value is between 0.44×10^{-5} and 27.92×10^{-5}ms^{-1}, while the transmissivity value is between 15.21×10^{-3} m/s and 231.9×10^{-3} m/s. The delineated zones of high transmissivities are suggested for installation of monitoring wells for unconfined aquifer; this is because high transmissivities equally suggest that aquifer materials in that area are permeable to fluid movement within the aquifer system, which may possibly enhance migration and circulation of contaminants in the groundwater system. The low range of values recorded for the longitudinal unit conductance and hydraulic conductivity could be attributed to the clay or partly shale content in the aquifers which affect the porosity of the layer. Weathered and fractured horizons, which constitute the aquifer zones underlying the VES stations, have been delineated in the study area. It is recommended that areas where overburden layer is relatively thick and has favourably low resistivity values in the study area should be considered for groundwater development.

8.0 References

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