NRIAG Journal of Astronomy and Geophysics 6 (2017) 422–433

Contents lists available at ScienceDirect

NRIAG Journal of Astronomy and Geophysics

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

Full length article

Assessment of spatial distrilbution of porosity and aquifer geohydraulic parameters in parts of the Tertiary – Quaternary hydrogeoresource of south-eastern Nigeria

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ARTICLE INFO

Article history:
Received 24 May 2017
Revised 9 August 2017
Accepted 8 September 2017
Available online 6 October 2017

Keywords:
Porosity
hydraulic parameters
resistivity
Coastal Plain Sands
Niger Delta
Nigeria

ABSTRACT

An integrated attempt exploring information deduced from extensive surface resistivity study in three Local Government Areas of Akwa Ibom State, Nigeria and data from hydrogeological sources obtained from water boreholes have been explored to economically estimate porosity and coefficient of permeability/hydraulic conductivity in parts of the clastic Tertiary – Quaternary sediments of the Niger Delta region. Generally, these parameters are predominantly estimated from empirical analysis of core samples and pumping test data generated from boreholes in the laboratory. However, this analysis is not only costly and time consuming, but also limited in areal coverage. The chosen technique employs surface resistivity data, core samples and pumping test data in order to estimate porosity and aquifer hydraulic parameters (transverse resistance, hydraulic conductivity and transmissivity). In correlating the two sets of results, Porosity and hydraulic conductivity were observed to be more elevated near the riverbanks. Empirical models utilising Archie's, Waxman-Smits and Kozeny-Carman Bear relations were employed characterising the formation parameters with wonderfully deduced good fits. The effect of surface conduction occasioned by clay usually disregarded or ignored in Archie's model was estimated to be 2.58 $\times 10^{-5}$ Siemens. This conductance can be used as a corrective factor to the conduction values obtained from Archie's equation. Interpretation aided measures such as graphs, mathematical models and maps which geared towards realistic conclusions and interrelationship between the porosity and other aquifer parameters were generated. The values of the hydraulic conductivity estimated from Waxman-Smits model was approximately 9.6×10^{-5}m/s everywhere. This revelation indicates that there is no pronounced change in the quality of the saturating fluid and the geological formations that serve as aquifers even though the porosities were varying. The deciphered parameter relations can be used to estimate geohydraulic parameters in other locations with little or no borehole data.

1. Introduction

The arcuate Niger Delta region located in the Gulf of Guinea is the mainstay of Nigeria's economy. The industries located in the region contribute greatly to economic boom of the states within the region. The region is one of the ten most prominent wetland and marine ecosystems in the world. The occupants of the region suffer immensely from harmful industrial effluents and other solid wastes of chemical, biochemical, microbiological and mechanical sources (Egboka and Uma, 1986; Kadafa, 2012). The prominent human activities which have negative impact on the Niger Delta environment include industrial (hydrocarbon exploration and exploitation, noise pollution, gas flaring, oil spillage, waste dis-

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Peer review under responsibility of National Research Institute of Astronomy and Geophysics.

Production and hosting by Elsevier

https://doi.org/10.1016/j.nrjag.2017.09.001
2090-9977/

posal), removal of backshore vegetation, construction of barges and other coastal control works, dredging of river, agricultural (excessive and uncontrolled application of pesticides, herbicides and inorganic fertilizers) practices (George et al., 2013). Others human activates that degrade the region include municipal waste disposal, urbanisation and mining (Ezeigbo and Ezeanyim, 2006; George et al., 2017a, 2017b). Consequently, environmental problems such as coastal and riverbank erosion, flooding, land degradation, loss of biodiversity, loss of soil fertility and deforestation are now predominant in the area (UNEP, 2011; Kadafa, 2012; include municipal waste disposal, urbanisation and mining (Ezeigbo and Ezeanyim, 2006; George et al., 2017a, 2017b). The water resources of the area which the settlers of the region used for their domestic, agricultural, industrial and social needs are now degraded by both natural and anthropogenic sources (UNEP, 2011, Saliha et al., 2017). In recent times, criminal activities involving illegal pipeline vandalism, crude oil theft and refining have compounded the environmental problems in the region. The quality of hydrogeoresource in many parts of the region is highly degraded (Okiongbo and Akpofure, 2012, George et al., 2014). This calls for urgent need to understand the pollutant flow-path in order to design appropriate mitigation, remediation and protection strategies especially now that such plans are in the drawing board by governmental, environmental and other international partners. The design of appropriate groundwater management strategies in any geologic environment hinges on the nature of subsurface hydrogeological materials whose properties (physical and chemical) and spatial distribution constitute the intent of all hydrogeological and hydrogeophysical investigations (Kallergis, 1999; Batte et al., 2008; Akpan et al., 2009). The basic aquifer properties needed for effective description and mitigation of hydrological and other hydrogeological problems have been deciphered by George et al. (2015a) and Soupios et al. (2007) to include fluid transmissivity, porosity, Dar zarouk parameters, aquifer thickness and hydraulic conductivity. These geohydraulic parameters are predominantly employed for hydrogeoresource evaluations, contaminant distribution, flow modelling, storage and productive capacity of the groundwater repositories, protective strength and in the design and construction of safe engineering structures (Yadav and Abolfazli, 1998; Aristodemou and Thomas-Betts, 2000; Batte et al., 2010; Gemail et al., 2011; Okiongbo and Akpofure, 2012, George et al., 2015a, 2017b). Information on the quality, nature and spread of these parameters are usually embedded in field data and can be accurately extracted if the data are available.

Generally, these geohydraulic parameters are often determined from pumping and/or slug tests in boreholes and/or analysis of core samples using laboratory procedures. These techniques utilise point data are comparatively slow and expensive. The measured parameters show some degree of spatial variability, which leads to inappropriate generalization for different geologic units exhibiting various geohydraulic properties (Kalinski et al., 1993). These, traditional approaches are invasive and cause significant disturbances to the natural environment. In view of this, environmentally friendly, less expensive and faster techniques such as ground based geophysical techniques have been developed and they are frequently employed in estimating these geohydraulic parameters (Slater, 2007; Khalil and Santos, 2009, George et al. 2015a). The availability of many geophysical equipment and inversion algorithms have even made the geophysical techniques in estimating aquifer parameters to be conveniently mapped in more than two dimensions (2-D) irrespective of the scale of variability. Aquifer parameters estimated only from surface geophysical data alone have been reported to be fraught with errors even when the conditions for data acquisition are favourable. This effect counteracts the well known documented advantages of geophysical techniques. The inherent errors can be grouped according to their originating sources to include inversion errors arising from smoothness constraints and other parameter resolution problems; the fidelity level of the field data and the extent of the validity of assumptions employed in the petrophysical relations that link the measured variables with the hydraulic parameters. In order to avoid these drawbacks, geophysical methods gearing towards estimating aquifer parameters in a satisfactory manner are integrated with geologic and hydrological samples to bring about hybrid techniques. The hybrid technique involes simultaneous analysis of geophysical and hydrological data (Niwas and Singhal, 1981, 1985; Mbonu et al., 1991; Khalil and Abd-Alla, 2005; Ekwe et al., 2006; Soupios et al., 2007; Chandra et al., 2008); theoretically based petrophysical models like the Waxman-Smits (1968) and the use of empirically and semi-empirically based hydrogeological and geophysical relations (Mazac et al., 1978; Domenico and Schwartz, 1990; Worthington, 1993; Frohlich, 1994). The later technique utilises information inferred from petrophysical properties of the groundwater repositories and the fidelity of the estimated parameters hinges on several simplifying assumptions such as type of aquifer, nature of fluid flow, well storage under field condition and pumping rate (Worthington, 1976; Kelly, 1977; Heigold et al., 1979; Kosiniski and Kelly, 1981; Mazac et al., 1985; Frohlich et al., 1996).

The Kozeny-Carman-Bear relation (Domenico and Schwartz, 1990) is a well known member of the family of such empirically and semi-empirically based relations commonly employed to estimate hydraulic conductivity in areas where formation factor, the ratio of bulk to fluid resistivities can be obtained. The porosity information needed in the relation is sourced from Archie's law (Archie, 1942), which relates formation factor to its porosity, cementation factor and other formation properties.

Groundwater repositories have electrical properties inferred from geophysical techniques. These properties have good degree of correlations with aquifer geohydraulic properties extracted from either pumping tests in boreholes or directly from laboratory measurements. This observation beefs up the research and makes room for direct estimation of aquifer parameters from geophysical data like electrical resistivity since both the geoelectrical and geohydraulic properties convey reliable and valuable information on pore space geometry and aquifer heterogeneity (Mazac et al., 1985; Börner et al., 1996; Christensen and Sorensen, 1998; Aristodemou and Thomas-Betts, 2000; Slater, 2007; Soupios et al., 2007; Börner, 2009; Niwas et al., 2011; George et al., 2015a). In view of this, correlation of borehole dependent aquifer geohydraulic parameter distribution information with alike information deduced from the analyses of surface resistivity data gives good hydraulic conductivity and other groundwater repository relations (Yadav and Abolfazli, 1998). These relations can be explored optimally to estimate aquifer characteristics in other locations within the vicinity (Robinson and Metternicht, 2006; Batte et al., 2010).

Despite the fact that conventional approach employing surface electrical resistivity derived parameters is dominating many literatures at both international and local levels, other classical surface geophysical techniques like the self potential (SP), seismic refraction, ground penetration radar (GPR), time and frequency domain electromagnetic techniques, induced polarisation and so on in aquifer parameter quantification and spread mapping are also significantly gaining ground (Corwin, 1990; Slater, 2007; Kirsch, 2009; Jouniaux et al., 2009). The conversion of geophysical data, usually obtained by indirect procedures into its corresponding geological facies is often observed and as pointed out by Ibanga and George (2016), the absence of a functional relations linking the observed formation parameters with the grain size distribution and other facies properties of the formation relating to permeability directly has been a major factor debilitating against direct extraction of any needed input. Even with this, recent innovations in geophysical methodologies such as surface nuclear magnetic resonance (SNMR) technique and spectral induced polarisation (SIP) or enhancements in interpretational techniques employed in the analyses and interpretation of classical geophysical data have high prospects for upsetting the ambiguities imposed by direct information extraction (Yaramanci et al. 1999; Sailhac et al., 2004; Vereecken et al., 2004; Jouniaux et al., 2009; Kirsch, 2009; Daigle and Dugan, 2011; Günther and Müller-Petke, 2012; Ikard et al., 2012; Jouniaux and Ishido, 2012; Kulessa et al., 2012). The use of SNMR to groundwater investigation has been innovative as in various investigations including direct detection, identification and quantification of the free-water content in a saturated formation and estimation of the geohydraulic properties (Yaramanci et al. 1999). Enhancements in the extraction of information from coupled geophysical data tremendously reduced the uncertainties associated with the direct conversion of geophysical information to hydrologic properties and other associated deductions (Hinnell et al., 2010).

Subsurface hydrogeological and hydrogeophysical analyses connecting to generation of information about water availability and usability require information about formation porosity distribution which is related to measurable properties of the formation like electrical resistivity (George et al., 2010b). Formation Porosity is a property whose spatial variability is viewed to be dependent upon numerous factors such as mineralogical composition, clay content, cementation, pore water saturation, density, tortuosity, hydraulic conductivity, formation factor, resistivity of the pore fluid and the rock matrix (Jackson et al., 1978; George et al., 2015a, 2015b). Transverse resistance, longitudinal conductance and transmissivity in general, are significantly influenced by formation porosity (Robinson and Metternicht, 2006; Soupios et al., 2007). The complexity of the real relations between porosity of formation and other geohydraulic properties is the enormous task which is often resolved by intense geophysical analyses (Cosentino, 2001).

In this survey, integrated approach involving information inferred from the analyses of surface resistivity data, pumping test data from boreholes and direct laboratory measurement from core samples in assessing the spatial distribution of porosity, estimating hydraulic conductivity and other dynamic properties of the Coastal Plain Sand (CPS) aquifers in some parts of the Niger Delta area of Southern Nigeria has been the primary objective.

1.1. Estimation of porosity from intrinsic formation factor

The basic relation of several rocks between bulk electrical resistivity ρ, fractional porosity ϕ and specific resistivity of the saturating fluid ρ_w, is based on Eq. (1) observed by Archie (1942)

$$\rho = \alpha \cdot \rho_w \cdot \phi^{-m} \tag{1}$$

where α is a parameter depending on formation whose value is often assumed to be unity for perfectly insulating rock matrix. Although in some geologic units, the actual value is found to vary minimally from the assumed unity value. Generally, the constant α reflects the ease with which the mineral grains allow the flow electric current through it (Aristodemou and Thomas-Betts, 2000; Niwas and de Lima, 2003; Slater, 2007; Soupios et al., 2007; Kirsch, 2009). The formation constant m has multiple interpretations such as shape cementation factor, porosity exponent, pore-shape factor and grain- shape (Khalil and Santos, 2009). The values of m depend on factors such as insolating properties of the cementation, mineral composition, pore shape geometry and extent of compaction (Timur, 1968; Ransom, 1984; Khalil and Santos, 2009). The other factors that m depends are anisotropy and overburden pressure. Some typical ranges of α and m for values of geologic formations have been documented in many literatures such as Hill and Milburn (1956), Carothers (1968), Porter and Carothers (1970), Gomez-Rivero (1977), Schön (1983, 1996), Worthington (1993), Friedman (2005), Khalil and Santos (2009), and Kirsch (2009) and great variations have been noticed.

Ideally, saturated clean sand environment characterised by mineral grains assumed to be perfectly insulating leads to the concept of intrinsic formation factor, F_i which is defined as $\frac{\rho_o}{\rho_w}$ (ρ_o is the resistivity of the saturated rock). ϕ relates to F_i and other formation geohydraulic parameters as

$$\phi = e^{\frac{1}{m}\ln(\alpha)+\frac{1}{m}\ln\left(\frac{1}{F_i}\right)} \tag{2}$$

In as much as changes in pore geometry are functions of α and m, there is need for site determination of their values for conformity with the prevailing subsurface conditions at each location. On the basis of this, Archie's law is not universal due to several factors that influence the resistivity inferred from it. These factors are the existence of clay minerals within the pore spaces and the extant sources of internal surface conductivity such as clay content of the formation and the presence of clay minerals within the pores (Keller and Frischknecht, 1966). Besides, faults and fractures in the formation, can either reduce the bulk resistivity predicted from Archie's law when these structures contain conductive fluids like water or augment the bulk resistivity of the formation for fractures and faults that are filled with insulating materials like air that do not allow for free flow of electrical current through them (Benkabbour et al., 2004). Again, geologic unit resistivities estimated from the law may be realisable from locations characterised by poorly conducting fluids such as chlorinated dense non-aqueous phase liquids (DNAPLs) (for instance trichloroethene) which contaminate the environment of saturated aquifer (Vinegar and Waxman, 1984; Worthington, 1993; Tait et al., 2004; Rivett and Clark, 2007; Chambers et al., 2010), slightly saturated aquifer (Börner et al., 1996; Martys, 1999) and fresh water aquifer (Alger 1966; Huntley, 1987). In such surface and subsurface geologic conditions, more corrective terms, which account for the additional sources of conductivity is opted for. A chunk of such models are presently in use and most of them fall into either shale-fraction or cation-exchange model derived practically from the parallel conduction process (Patnode and Wyllie, 1950; Winsauer and McCardel, 1953; Waxman and Smits, 1968; Clavier et al., 1984; Frohlich and Parke, 1989; Sen et al., 1998).

The Waxman-Smits model referred to as a modified version of Archie's law is a common choice when examining the flow of current in aquifer systems. In this case, original version of Archie's law is violated due to additional sources of subsurface conductivity (Vinegar and Waxman, 1984; Worthington, 1993; Soupios et al., 2007). According to Worthington (1993) and George et al. (2015b), the respective apparent and intrinsic formation factors, (F_a (the ratio of bulk resistivity to fluid resistivity) and F_i (the same ratio after taking into account the shale effects)) are related according to Eq. (3) as

$$F_a = \frac{F_i}{1 + BQ_v\rho_w} \tag{3}$$

where B and Q_v reflect respectively the equivalent conductance of the sodium clay-exchange cation as a function of formation water conductivity and cation exchange capacity of the rock unit per volume. These properties reveal valuable information associated with subsurface condition of flow of electric current. Expression in Eq. (3) can be transformed into a linear form as

$$\frac{1}{F_a} = \frac{1}{F_i} + \left(\frac{BQ_v}{F_i}\right)\rho_w \tag{4}$$

According to Eq. (4), F_i can be deduced from the inverse of the intercept along the $\frac{1}{F_a}$ axis when a plotting $\frac{1}{F_a}$ against ρ_w while BQ_v can be estimated from the slope-intercept relation of the same plot. Thus plotting $\frac{1}{F_a}$ against ρ_w, F_i can be obtained through the estimation of the porosity of a clay-free medium in Eq. (2). This method requires the estimation of variations of ρ_o with depth in the formation from I-D resistivity inversion and ρ_w from water wells closed to the VES stations (Aristodemou and Thomas-Betts, 2000; Slater, 2007; Soupios et al., 2007). These two sets of data allow for estimation of F_a of a water saturated repository. It is worthwhile to note that $BQ_v\rho_w$ term vanishes when F_a and F_i are equal (George et al., 2015a). That is

$$BQ_v\rho_w = 0 \tag{5}$$

Since water resistivity, ρ_w must be greater than 0, the BQ_v term must vanish when the above condition is fulfilled. Hence, Eq. (4) it is clear that the major source of error is coming from wrong estimation of F_a.

It is also worthwhile to note that factors control current flow and conductivity distribution in the soil are not fixed but variable. In the light of this, measured and calculated formation geohydraulic properties cannot be seen as absolute but relative. Consequently, it is only possible and unique to make relative conclusion about the formation properties associated to an area (Vinegar and Waxman, 1984).

2. Location and geology of the study area

Judging from Fig. 1B, mapped area falls between longitudes 7°30′E and 820′E and latitudes 4°30′N and 5°30′N in the Gulf of Guinea, southern Nigeria. The site was to cover three Local Government Areas (LGAs) comprising Nsit Ubium, Eket and Onna (Fig. 1D) in Akwa Ibom State (Fig. 1C) of southern Nigeria. The study area is within in an equatorial climatic region characterised by dry and wet seasons which are the rainy season (March – October) and dry season (November – February) (Evans et al., 2010; George et al., 2010a, 2015a, 2015b). The area is negatively affected by the current global climatic changes which cause a shifts in both the upper and lower divides of these climate (Martínez et al. 2008; Rapti-Caputo 2010; Riddell et al. 2010; Wagner and Zeckhauser, 2011; Farauta, et al., 2012).

The area under study is within a Tertiary – Quaternary CPS equally called Benin Formation (BF) and Alluvial environments in the Niger Delta of southern Nigeria (Fig. 1B). This Formation which covers about 80% overlies the parallic Agbada Formation of in the area. The clastic sediments of the BF comprise the interfringing units of lacustrine and loose fluvial sands, clays and lignite, pebbles, and layers of varying thicknesses. The alluvial units have characteristic beach sands, tidal and lagoonal sediments (Reijers and Petters, 1987; Reijers et al., 1997; Baker Hughes INTEQ, 1999; Nganje et al., 2007) commonly found in the river banks as well as southern parts. The CPS has overlay of thin lateritic overburden materials characterised by unequal thicknesses spread at different locations. However, along the shorelines, BF has massive exposure. The CPS constitutes the major hydrological repositories in the area. This formation is unevenly sorted (fine-medium-coarse) aranaceous materials which alternate with layers of thin clay horizons and lenses and lignite streaks at various positions. The thin clay-shale units cut the lateral and vertical regions of the aranaceous groundwater repository. This builds up systems of many confined and open aquifers flowing southwardly into the Atlantic Ocean within the study area (Edet and Worden, 2009).

3. Field studies

Ground based electrical techniques have been applied in the investigation of various environmentally, geotechnically and geologically related problems for search of solutions to the problems. Generally, the electrical resistivity technique aims at injecting electrical current into a pair of electrodes current electrodes. The effect of injected current that passes through the earth is monitored through the streaming potential difference measured between another pair of electrodes known as potential electrodes (Zhdanov and Keller, 1994).

Many electrode arrangements and field procedures for electrical resistivity abound (Telford et al., 1990) However, the Schlumberger array utilising vertical electrical sounding (VES) field procedure was chosen to assess the vertical and lateral variations of subsurface electrical conductivity with depth. Twenty three VES investigations were conducted across the study area with a SAS1000 model of ABEM terrameter. In communities with good access paths/roads, the current electrode separations were up to 800 m. This was designed to make sure that depths above 150 m were probably sampled when taking into consideration the possibility of depth penetration between 0.25 and 0.5 of current electrode separations (Roy and Elliot, 1981; Singh, 2005). The equivalent receiving electrode separation (MN) ranged from a minimum of 0.5 m at AB = 2 m to a maximum of 40 m at AB = 800 m. In all the electrode locations, the separation between the potential electrodes did not go beyond one fifth of the separation of the current electrodes (Gowd, 2004).

The analyses of geophysical data were constrained by drilling information for effective interpretations. The drilling exercise started as soon as the geophysical reports were submitted to the Akwa Ibom State Millennium Development Goal (AKSMDG, 2011) who sponsored the borehole drilling projects. Some of the boreholes were cited near the VES locations while those with no access path to spread the cables at the vicinity of the boreholes were separated by about 150 m. Cuttings from drilling were logged geologically in all the 23 locations and samples were analysed in the laboratory. The boreholes were cased with 75 mm high pressure PVC casing materials. The casings at various depths were slotted (Fig. 2B) and the slotted region of the well annulus was packed with gravel to enable good delivery of water from the aquifer unit to the borehole as well as checking for ingress of sediments into the borehole. Back flow of water at the surface into the well was achieved by mixing of sand and cement to grout the boreholes. The development of the wells was accompanied by pumping test which took place at the same time. A 24 h pumping period and a 24 hour recovery data collection was opted. All the procedures and analytical techniques adopted in the analyses of the pumping test detailed in AKSMDG (2011). The in-situ measurements of electrical conductivity from water samples were performed in all the borehole locations employing WTW LF91 conductivity meter.

According to API (1960), distilled water was used to pre-wash the core samples to ensure removal of traces of clay or argillaceous materials deemed to have come from the coring operations. Thereafter, the samples were inserted into a vacuum desiccator and evacuation was done at a pressure of 0.3 mBar for a period of 60 min (Emerson, 1969). De-ionized/de-aerated distilled water was carefully put into the desiccator until samples were completely covered and this was followed by soaking for a period of 24 h. The soaking was to diffuse out the trace of soluble contaminants and salt into the water medium.

The drying of cleaned samples in a temperature controlled oven at 105 °C for 16 h followed to check for any noticeable irreversible alterations in the composition of the samples (Emerson, 1969; Galehouse, 1971). Again, the oven dried samples were subjected

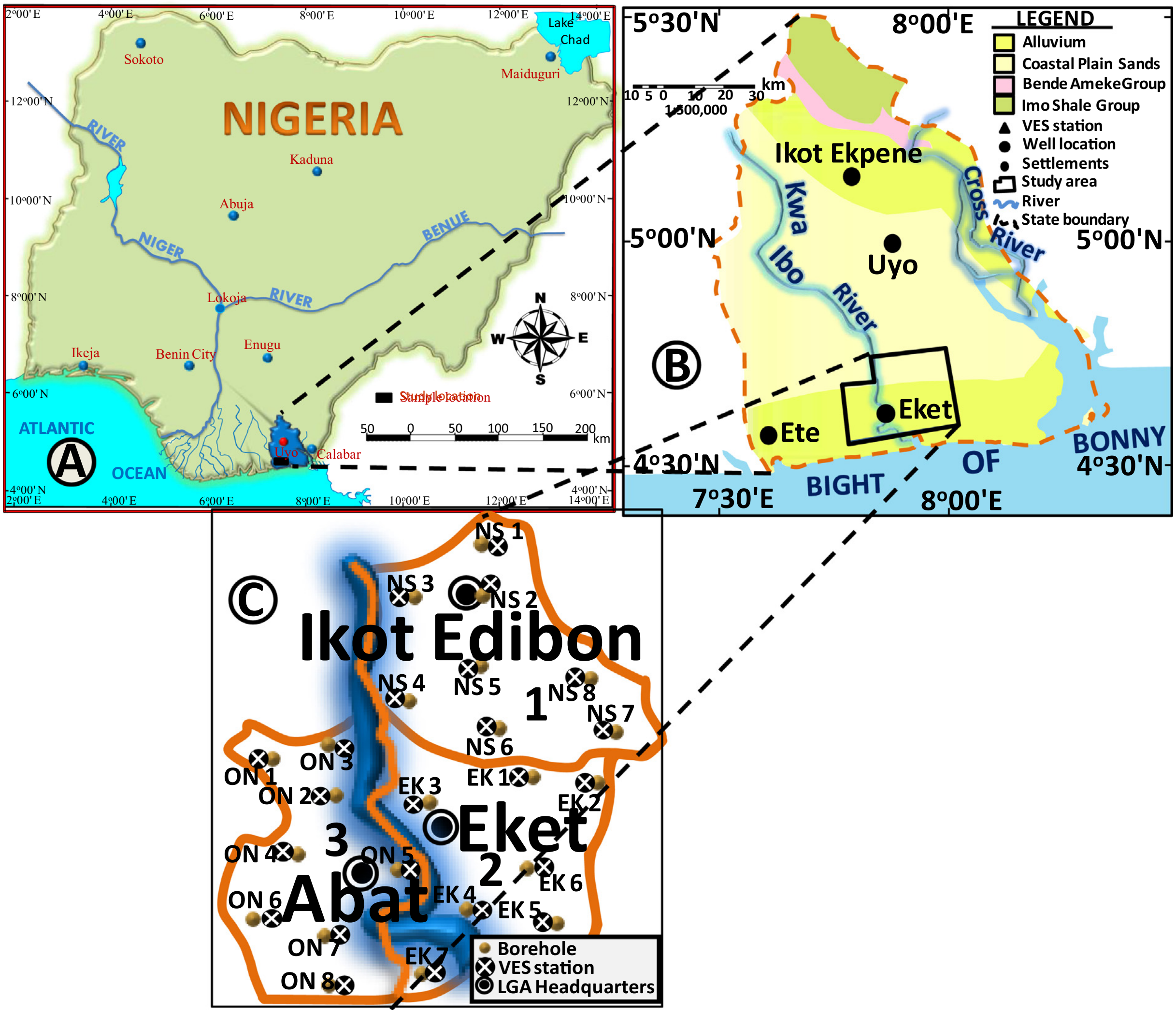

Fig. 1. Political map of Nigeria (A) showing location of Akwa Ibom State (B). Generalised geological map of Akwa Ibom State showing the location of the three investigated LGAs (Nsit Ubium (1), Eket (2) and Onna (3)).

to cooling at normal air temperature in a dessicator. The cool and dry samples were weighed using an electronic weighing balance five times and the mean calculated and recorded as W_d. The samples were again soaked with distilled water that was boiled for half an hour in a vacuum pressure of 0.3 mBar for 18 h. The weight of the wet samples was also measured five times and the mean calculated and recorded as W_w.

From the measurements, effective porosity (ϕ) of the samples was calculated using Eq. (6)

$$\phi = 100 \cdot \left(\frac{W_w - W_d}{V} \right) \% \tag{6}$$

where V is volume of the samples measured from cylindrical mold.

Though other methods abound, hydraulic conductivity was also estimated from the samples using the constant head permeameter (CHP) method (Freeze and Cherry, 1979; Frohlich and Urish, 2002; Travelletti et al., 2011). The core samples were inserted into a cylindrical mold enclosed between two porous plates of length L and cross sectional area A and a constant differential head h was set up across the samples. Water drains the medium cylinder from the bottom and was collected as overflow after passing upward through the sample. From Darcy's law, the hydraulic conductivity k relates to these measurable parameters according to Eq. (7)

$$k = \frac{QL}{Ah} \tag{7}$$

where $Q = \frac{V}{t}$ is the volume rate of flow and V is the volume of water collected in a time t.

The k which is equivalent to the flow rate of water through a unit cross sectional area under a unit hydraulic gradient was estimated empirically from the core samples through CHP method. The procedures are detailed in Freeze and Cherry (1979) and Obianwu et al. (2011). In the CHP, the head in which water flows through the specimen remains fixed throughout the experiment.

4. Data analyses, results and discussion

Manual and computer modelling techniques were deployed in the reduction of the 1-D VES data (Zohdy, 1965; Zohdy et al., 1974; George et al., 2016a, 2016b) to geological equivalents. The manual procedure considers the plotting of the computed apparent resistivities on bi-logarithmic graph. The generated plots where necessary, were smoothed in order to remove the effects of lateral heterogeneities and other signatures related to noisy (Bhattacharya and Patra, 1968; Chakravarthi et al. 2007). The smoothening was performed by either averaging the two readings at the cross over points, or deleting any outlier at the cross over

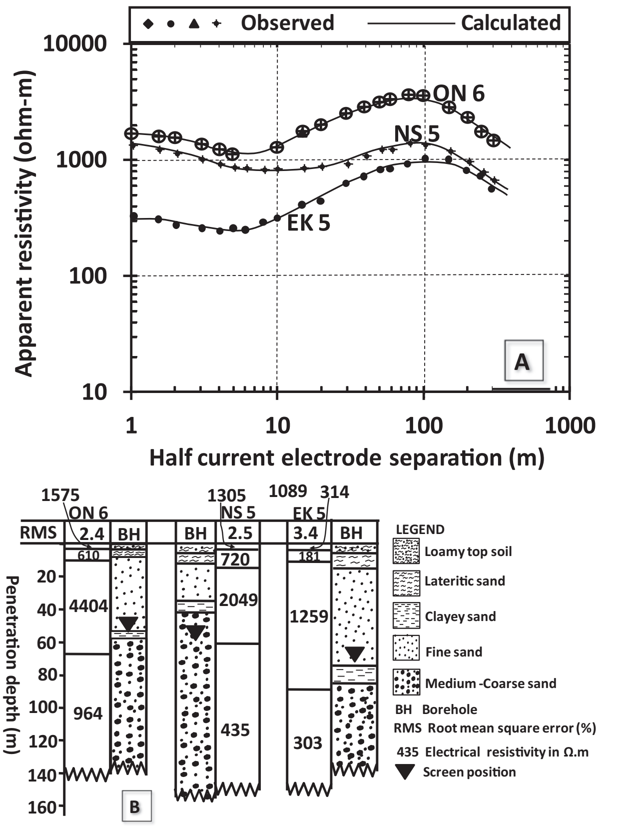

Fig. 2. Samples of modelled VES curves (A) and (B) correlation between VES derived 1-D subsurface models and borehole lithologs

points that did not conform to the dominant trend of the curve. Again, data that stood out as outliers in the prevalent curve trend which caused serious increase in root mean square error (RMSE) during the modelling phase of the work were also removed. Generally, the outliers constitute less than 2% of the total data generated in each VES station. As the measurement was taken over ten data per decade, expunging such noisy data did not change the trend of the sounding curve. Some of the noisy loaded data are viewed to have come from electrical signatures of the thin clay beds that have been suppressed from the over-and underlying thick sandy aquifers (Gurunadha Rao et al., 2011; Sabet, 1975). The discontinuity observed afterward in the smoothened curves was attributed to vertical change in electrical conductivity with depth. Preliminary deductions of primary layer characteristics from smoothened curves were done through the traditional partial curve marching technique. According to Orellana and Mooney (1966), partial curve matching technique employing master curves and charts was used. VES modelling using RESIST software (Vender Velpen, 1988) enabled the automated approximation of the initial resistivity model from the observed data to improve on the preliminary inferred results using the iterative inversion technique. The software employs the initial layer properties to carry out some calculations and finally generates a theoretical curve. The theoretical curves were compared with the field data curve and depending on the fit, a root-mean square error (RMSE) was generated. Due to inherent problem of equivalence (Van Overmeeren, 1989), quantitative interpretation of geoelectrical sounding data is always difficult. However, this challenge is overcome by using borehole data to constrain all depths in order to minimise the choice of equivalent geologic models. According to Batayneh (2009), this is achieved by fixing layer thicknesses and depths while allowing the resistivities to vary. The predominant observed minima and maxima on the smoothened VES curves were utilised as the starting number of layers (or models) over a half space for the data inversion process. The display in Fig. 2A shows some of the modelled VES curves observed while Fig. 2B gives the correlation between the nearby borehole lithology and interpreted results. As observed a good fit was seen between the borehole lithology data and the inverted results over half space in many VES stations except in a few stations (4 in particular), where correlation was not well fitted. These alterations were viewed to have originated from the failure of the 1-D interpretation to the assumption of the shallow subsurface of the half space to be horizontally spread (Gurunadha Rao et al., 2011). Taking a clue from Soupios et al. (2007), noticeable distortions can be realizable when the 1-D subsurface assumption is applied in resolving a 2-D (or 3-D) problems. Interestingly, White et al.(1988) opined that most sounding results do not always show a perfect match with drilling results, instead depth estimates from most sounding results sometimes fall within 25% of the drilling depths.

Using Waxman-Smits model (Eq. (4)), a $\frac{1}{F_a} - \rho_w$ plot estimated respectively, F_i and BQv values as 32.3 and 2.58×10^{-5} Siemens. The estimated F_i was used to compute the mean or intrinsic porosity using Eq. (2). Longitudinal unit conductance (S) and transverse unit resistance (TR) (Dar-Zarrouk parameters) (Maillet, 1947; Singh et al., 2004; Soupios et al., 2007; Batte et al., 2010; Gemail et al., 2011) were calculated from the bulk resistivities of the saturated sand medium and thicknesses. According to Macdonald et al. (1999), the usual problem of non-uniqueness associated with the interpretation of electrical resistivity results through Dar-Zarrouk parameters in estimating transmissivity and other aquifer hydraulic parameters was minimised and this was a valuable advantage in the validity of the result. Transmissivity and the Dar-Zarrouk parameters are bulk parameters and do not depend on independent measurements of bed thickness, resistivity or permeability. On this note, S was calculated in two phases. In the first phase, S_1 was computed for the saturated aquifer by dividing the observed saturated layer thickness (h) obtained from the modelling and interpretation of sounding curve by the corresponding bulk resistivity of the same saturated layer according to Eq. (8) as

$$S_1 = \frac{h}{\rho_o} = h \cdot \sigma_o \tag{8}$$

where σ_o is the electrical conductivity (inverse of resistivity) of the saturated aquifer in Siemens. In the second phase, S_2 was computed from TR by multiplying h with ρ_o according to the equation

$$TR = h \cdot \rho_o = \frac{h}{\sigma_o} \tag{9}$$

Finally, the transmissivity (T in m^2/s), the ability or ease of the aquifer to transmit fluid across its whole length using Eq. (10) as

$$T = k \cdot h \tag{10}$$

Results obtained are shown in Table 1. The computed values of TR ($\Omega \cdot m^2$) were plotted against T (m^2/s) as shown in Fig. 4. Results show that TR and T are related by an empirical relationship of the form

$$TR = 143.25T + 35,610 \tag{11}$$

The distribution of the T and ρ values and shown in Table 1 demonstrate that there is no simple relation between them as opined by Singh (2005). This revelation is contrary to the exponential relations that Singh (2005) proposed between T and ρ for some coastal hydrogological repositories. Based on Khalil and Santos (2009) the relationship between (T) and (ρ) is non-unique and therefore, no generalised, simple and easily predictable T–ρ relations in aquifers may be realisable. On the other hand, TR and T

Table 1
Summary of aquifer parameters.

Name of LGA	VES station code	Longitude (E)	Latitude (N)	Thickness (m)	Resistivity (Ω m)	Estimated porosity (%)	K_{obs} (m/day)	K_{est} (m/day)	K_{obs} σ ($\times 10^{-3}$) (m^{-1}/day^{-1})	T ($\times 10^{-3}$) (m^3/1 day)	TR ($\times 10^3$) (Ω m^2)	S ($\times 10^{-3}$) (Siemens)
Nsit	Ubium	NS1				7°59′25.04′′			4°48′31.64″	65.4	3920.0	17.4
	8.2	7.0	2.09	536.78	125.00	16.68						
	NS2	7°58′58.73″	4°46′47.64″	57.7	525.7	16.4	8.6	5.9	16.36	496.22	92.40	109.26
	NS3	7°55′20.86″	4°46′28.78″	89.2	393.9	8.1	7.9	4.9	19.95	701.11	82.00	226.45
	NS4	7°55′13.40″	4°42′50.90″	50.9	924.2	10.2	7.8	13.7	8.44	397.02	47.04	55.07
	NS5	7°58′21.25″	4°43′54.70″	51.2	2049.4	18.0	8.1	6.8	3.95	414.72	104.93	24.98
	NS6	7°59′02.47″	4°41′46.93″	60.6	4940.9	39.5	8.2	12.5	1.66	496.92	129.00	12.27
	NS7	8°03′48.06″	4°41′43.19″	112.3	1913.0	19.1	8.3	8.6	4.34	932.09	114.00	58.70
NS8	8°02′55.49″	4°43′32.12″	62.2	1726.0	20.1	8.4	17.6	4.87	522.48	107.36	36.04	
Onna	ON1	7°49′33.31″″	4°40′39.40″	45.2	40.0	2.8	7.8	9.4	195.00	352.56	117.00	1130.00
	ON2	7°52′09.26″	4°39′14.90″	125.8	3364.6	36.6	8.5	18.2	2.51	1063.01	118.00	19.32
	ON3	7°52′33.71″	4°41′03.84″	12.0	6583.8	40.1	8.4	7.0	1.28	100.80	126.00	6.38
	ON4	7°50′27.78″	4°37′03.40″	56.7	3575.6	28.3	8.0	10.2	2.18	453.60	108.00	15.43
	ON5	7°55′15.24″	4°36′35.21″	75.4	3094.4	36.0	81	13.3	2.12	610.74	133.00	24.37
	ON6	7^{4}59′16.36″	4°34′50.02″	64.2	4404.3	29.5	7.8	14.7	1.78	503.33	148.00	14.58
	ON7	7°52′18.59″	4°34′10.45″	67.4	4614.6	40.0	7.8	18.6	1.69	5325.72	116.00	14.61
	ON8	7°52′29.96″	4°32′17.77″	91.2	1274.6	39.1	8.5	16.4	6.67	775.20	116.24	71.55
Eket	EK1	8°00′21.38″	4°39′52.38″	81.8	9025.6	37.9	8.6	16.2	4.27	706.75	146.00	40.40
	EK2	8°03′12.42″	4°39′39.31″	94.5	126.7	5.6	7.8	13.1	61.80	739.94	92.50	743.86
	EK3	7°56′02.25″	4°34′54.20″	30.6	251.4	9.5	8.2	7.0	32.62	250.92	76.00	121.72
	EK4	7°58′19.38″	4°35′06.79″	27.6	479.4	7.3	8.5	4.0	17.73	234.60	53.00	57.57
	EK5	8°01′23.48″	4°34′36.77″	50.9	1258.4	32.3	8.2	16.8	6.52	417.38	64.05	40.45
	EK6	8°00′49.54″	4°36′37.08″	23.4	3711.5	27.5	7.8	8.1	2.11	183.22	86.85	6.30
	EK7	7°56′19.07″	4°33′00.86″	82.7	42.7	4.8	8.2	2.4	192.04	678.14	65.00	1936.77

were noticed to relate linearly. The linearly dependent relation between TR and T has also been observed by Soupios et al. (2007), Batte et al. (2010) and Ibanga and George (2016).

The k values determined using the conventional methods of CHP and pumping test were correlated with k values inferred from porosity estimates from formation factor using modified Archie's law. In Table 1, the observed values of S presupposes that the aquifers in the area are poorly protected since all values of S are below1 Siemens, which is a benchmark for assessing aquifer protection proposed by Henriet (1976). The observed results are in agreement with what is expected in a highly porous sandy formation with low clay-sand thickness ratio. As a confirmation, Okiongbo and Akpofure (2012) observed similar poor aquifer protective strengths in the Bayelsa State axis of the same CPS of Nigeria. Braga et al. (2006) employed values of S estimated from 1-D resistivity data acquired in Canoas, Brazil to assess aquifer protective capacity.

The total longitudinal unit conductance values were employed in assessing the vulnerability or protective capacity of the hydrogeorsource. The earth layer naturally filters the percolating fluid. The ability of the earth to retard and filter percolating fluid is a dependent on the protective strength of the earth medium. Aquifer overburden overlying layer protection is depends on its hydraulic conductivity. High clay content which impedes fluid movement is basically characterised by low resistivity values, low hydraulic conductivities and thus low longitudinal unit conductance. Consequently, sandy aquifers in the study area are characterised by low protective overburden covering materials, which make them vulnerable to contamination from microbiological, chemical, biochemical and other sources of groundwater contamination in the oil rich Niger Delta (George et al., 2017c). The UNEP (2011) opined that the area has widespread contamination of the water resources in the study area. Based on the CHP result, k was noticed to range from 2.82×10^{-5} m/s to 2.15×10^{-4} m/s ($\bar{k} = 9.50 \times 10^{-5}$ m/s and standard deviation (δ) of 3.20×10^{-6} m/s). The results realised from pumping test technique vary from 9.00×10^{-5} m/s to 1.00 $\times 10^{-4}$ m/s ($\bar{k} = 1.27 \times 10^{-4}$ m/s, and standard deviation $\delta = 0.57$ $\times 10^{-4}$ m/s). However, the k values inferred from the analysis of core samples range from 2.80×10^{-5} to 2.15×10^{-4} m/s (or 2.144 – 18.59 m/day) and these results were not significantly different from k values obtained from the CHP method and the Kozeny-Carman-Bear model (Eq. (12) Table 1). According to Niwas and Singhal (1981), the theoretically computed k value of 1.00×10^{-4} m/s fairly correlate with these sets of results for identical geologic units. Again, the values of k calculated from the Kozeny-Caman-Bear's equation (Eq. (12)) reflect a fairly constant value of approximately 0.96×10^{-4} m/s which is in consonance with k values experimentally determined from the Constant flow head permeameter (CHP) technique, pumping test procedure and the theoretically computed value available in documented literatures for similar geologic formations (Edet and Okereke, 2002), Bouwer (1978) and Freeze and Cherry (1979) opined that sandy aquifers have k values ranging from 10^{-7} m/s to 10^{-5} m/s. Through fractional porosities, k value for the aranaceous formation was regressed to be 9.6×10^{-5} m/s (or 8.3 m/day). The result is in good agreement with the theoretical value of 1.00×10^{-4} m/s (or 8.64 m/day) given by Niwas and Singhal (1981) for pure for aranaceous aquifer formations. The close range noticed in k values suggests that the aquifer medium is fairly homogeneous and anisotropic and probably free from major contaminants like iron and saltwater that are common problems in some parts of the Niger Delta coastline (Edet and Okereke, 2001a, 200b; Manheim et al., 2004; Wilson et al., 2005; Kirsch, 2009; Islami, 2011; Gemail et al., 2011; George et al., 2016b). This again corroborates with the findings of Ajayi and Umoh (1998), Edet (1993) and Edet and Okereke (2001a, 2001b) who viewed the low levels of concentrations of chlorine, salinity and total dissolved solids (TDS), density and Br/Cl ratio in water samples from the coastal parts of the Niger Delta region to be a reflection of the absence of ingress of seawater into coastal environments. The absence of classical seawater contaminants such as saltwater and ferruginous substances in samples of water can be linked to the enormous volume of water discharged into the Atlantic Ocean from the unconfined to partially confined CPS aquifers (Ajayi and Umoh, 1998; Edet and Okereke, 2001a, 2001b). It has been pointed out by Edet and Okereke (2001b) that the high hydraulic gradient created from the high volume of water that is being discharged into the ocean have pushed the sea water-fresh water interface deep into the ocean and this causes the interface to be shifted seaward into the ocean. The spatial spread of $k_{obs}\sigma_o$ values (Table 1) is another supportive evidence of the seemingly homogeneous distribution of the sandy aquifers across the entire area with no significant change in the composition of the pore fluid content. Singh et al. (2004) used regression analysis of D-Z parameters to attribute outliers from the line of best fit to be reflection of variations in pore fluid compositions in a coastal environment further concluded that the use of D-Z parameters in the analysis of aquifer characteristics can furnish a simplified technique for resolving any ambiguity in discriminating between contaminated (saline) and freshwater aquifers in coastal environments. Frohlich and Urish (2002) deciphered the difference in the observed values of k to be due to the existence of heterogeneities in glacial till aquifers. The findings of this study have also agreed with the computed values of k by Edet and Okereke (2002) in the Calabar axis of the same.

The Kozeny-Carman-Bear's equation establishes a quantitative relation that connects k with ϕ according to Eq. (12)

$$k = \left(\frac{\delta_w g}{\mu}\right)\cdot\left(\frac{d_m^2}{180}\right)\cdot\left[\frac{\phi^3}{(1-\phi)^2}\right] \tag{12}$$

where d_m is the mean grain size which was determined in this study by direct measurement using Vernier Callipers and micrometer screw gauge to be 0.000477 m, δ_w is the density of water (taken as 1000 kg/m^3) and μ is the dynamic viscosity of water which according to (Fetters, 1994) can be taken to be 0.0014 kg/m s. The observed porosity values for each of the sites investigated are given in Table 1. The computed/observed porosity values were contoured using ordinary and surface plots (Fig. 3) procedures to examine the spatial spread across the mapped area. The results of porosity - formation factor plots for clayey and clay-free cases realised are given in Fig 5. Regression techniques were employed to deduce a quantitative relation between porosity and formation. The observed porosity was found to relate to the formation factor in a clayey medium according to the power model given in Eq. (13).

$$F = 1.097\phi^{-2.011} \tag{13}$$

The regressed power model in Eq. (13) with an inverse of $\phi = 1.048F^{-0.497}$ agrees fairly well with Archie's law for idealised clean sand formation. As observed in Eq. (13), the cementation factor (m) is 2.011 while tortuosity (α), which depends on the lithology is 1.097. As a matter of fact, the observed values of m and α fit into the spread of ranges established by previous researchers such as Hill and Milburn (1956), Carothers (1968), and Porter and Carothers (1970), in sandstone environment. The porosity of the pre-washed samples, which technically has been stripped of its argillaceous constituents was observed to have been reduced by an average of about 10% and a graphical plot of the assumed clay-free samples against the estimated formation factor generated a different power model shown in Eq. (14)

$$\phi = 0.122F^{-0.24} \tag{14}$$

The model for clay-free condition shown in Eq. (14) with an inverse of $F = 1.582F^{-4.16}$ (Archie's format) indicates that m is 4.16 while α is 1.582. Fig. 5 shows that the porosity of the water saturated sand is highly affected by the presence of clayey and shaly materials in the samples. The influence of argillaceous constituents on the formation factor is also significant because their values change significantly when the samples were stripped of their argillaceous constituents. Zhdanov (2009) consented to the fact that the geometric form of the cross section of the pore along the path of current flow seems to exert strong influence on the parameters α and m. Since the actual pore structural system is viewed to be very complicated in nature due to the well known poorly sorted nature of the Benin Formation, the observed regressed values of α and m for clayey and clay free sand has high disparity due to heterogeneously cemented Tertiary sands medium.

The computed porosities were observed to vary from 0.028 (2.8%) to 0.41 (41.0%) ($\overline{\varphi}$ = 23.2%) and its spatial distribution in the study area is shown in Fig. 3. Higher porosity values were observed in the south-western parts of the study area especially along the bank of the Kwa River where k values are also higher. The low porosity values were prominently found in the sandy silt enriched areas. The tabulated ranges of different porosity values for a wide range of geologic formations has been provided by McWorter and Sunada (1977). This observation synchronises well with the expected near shore deposition pattern. According to Richmond and Sallenger (1984) and Dyer (1986), in normal near shore deposition patterns, coarser and denser materials are usually deposited near the shore especially when a low energy-transporting agent was involved in the transport of the sediments while finer and less dense materials are usually transported further

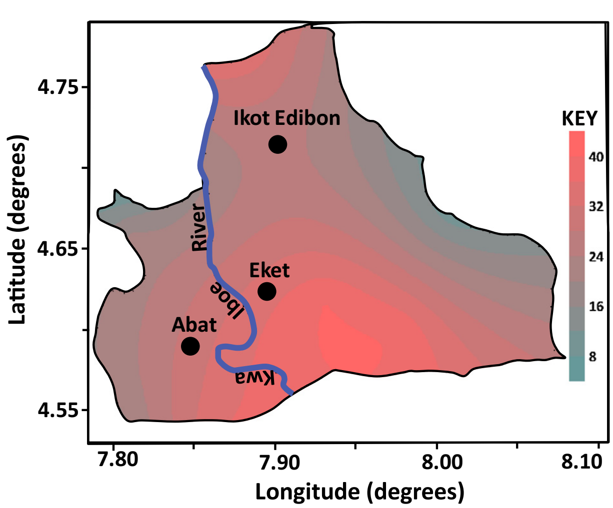

Fig. 3. Distribution of effective porosity in the study area

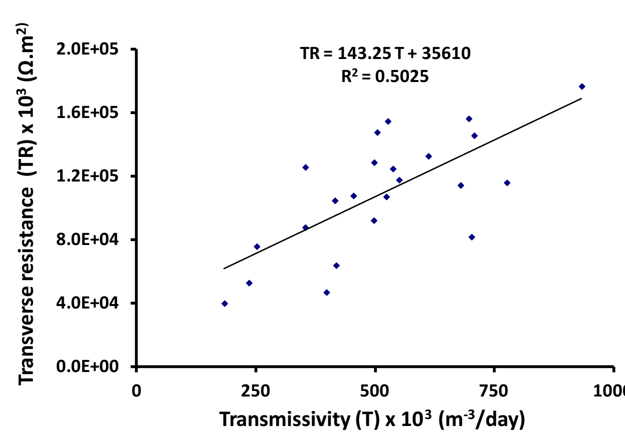

Fig. 4. Graph of estimated transverse resistance against transmissivity in the study area

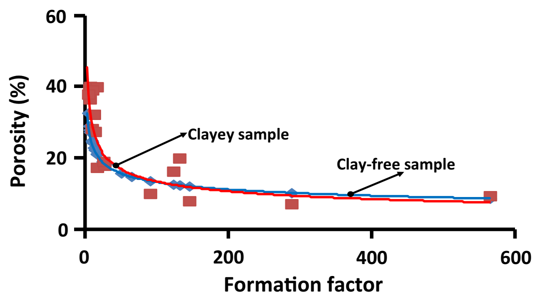

Fig. 5. A plot of porosity against the formation factor for clayey samples (red) and clay free samples (blue). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

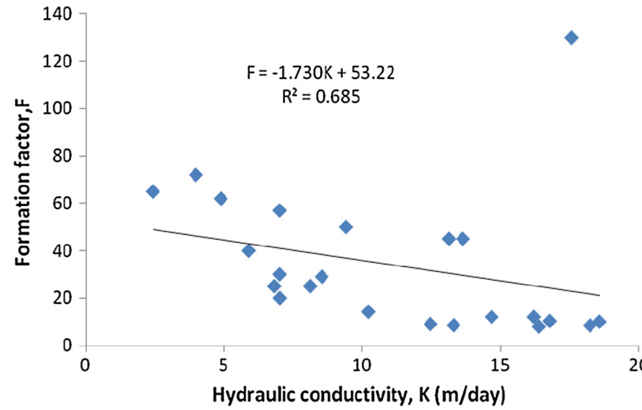

Fig. 6. A graph of estimated formation factor against hydraulic conductivity in the study area

seaward (offshore). Garg (1978), Metwaly et al. (2006) and Batayneh (2009) viewed the increase in k with ϕ to be due to increase in individual pore size and as well as increase in the flow area. The high porosity locations seem to agree with areas characterised by coarse sediments and other denser materials transported from the adjoining areas by low energy surface runoff water from heavy rains. These materials are usually deposited at the bank of the Kwa River while the finer constituents are transported further into the water. Hence along the riverbank, the quantity of clay and other argillaceous materials that can reduce the effective porosity is not significant. Interestingly, k values and other aquifer parameters (TR, S, T, ϕ) observed in the vicinity of the rivers and other regions with alluvial deposits were observed to be slightly higher than values observed in the other regions. The noticeable increase in porosity along the bank of the river does not necessarily translate into an increase in the amount of interstitial water because it has been generally observed that as the percentage of reservoir water increases, permeability, porosity and grain size decreases. However, the seemingly high values of the aquifer geohydraulic parameters (k, TR, S, T and ϕ) suggest that contaminant can be easily transported from one part of the aquifer to the other. In Jones and Buford (1951) and Batayneh (2009) observations from their investigations, the grain size increases with formation factor and intrinsic permeability of the aquifer. Going by Cosentino (2001) obsessions, aquifers whose grain sizes are basically small, the pore sizes will be correspondingly small. This causes capillary pressure to increase thereby resulting in more water being retained in their pores spaces. A comparative assessment of the relationship between k values observed from the CHP and the formation factors obtained from the twenty-three wells is shown in Fig. 6. The figure reveals the inverse relation that exists between formation factor and hydraulic conductivity, thereby confirming that earth materials with low voids though they are less permeable, can accommodate more fluid than high grain size aquifers.

5. Conclusion

The study demonstrates how the applicability of estimates of porosity and aquifer hydraulic properties can be determined economically from surface resistivity data in sites where insufficient information from boreholes exists. The values of porosity and other geohydraulic parameters estimated from inexpensive geophysical method correlate fairly well with the values obtained from the conventional methods in similar geologic units. Thus the availability of a few expensive pumping test data and core samples, which are limited along with economically available surface geophysical method, can serve as a very useful pair in economical and quantitative estimation of aquifer geohydraulic parameters.

At the riverbanks the deposition of coarser and denser materials transported by low energy runoff water high porosity and hydraulic conductivities are more predominant than in the adjoining areas. Although increase in porosity and hydraulic conductivity does not necessarily reflect increase in pore water, the findings will be very valuable in future groundwater flow modelling studies in the area. Besides, these findings can be useful in the positioning exploratory wells. Moreover, significant achievements in planning for appropriate prevention and remediation schemes for groundwater and other contaminated sites such as the Niger Delta region where hydrocarbon exploration and exploitation exercises have elevated the level of contamination of groundwater resources can be achieved (UNEP, 2011).

The established empirical relations connecting porosity and other geohydraulic properties of a saturated formation have been generated and these model equations derived could be used in modelling the groundwater repositories for effective monitoring and managements. Since porosity and hydraulic conductivity are important parameters in oil exploration/exploitation, the models realised in this study could also be considered as input parameters during well development and appraisal in the study.

The findings show that the aquifers are highly vulnerable to contamination from the various sources of contamination including hydrocarbon that abounds. In all the study locations, the only natural protective mechanism for the coastal plain sand aquifers falls below the level of protection. Hence, this calls for closer monitoring of the aquifer in the area where oil spills and other adverse effects of oil exploration and exploitation activities are high. The CPS of the Niger Delta region of Nigeria is a fairly homogeneous geologic formation in terms of spatial spread of the various lithostratigraphic units, groundwater dynamics and quality variation. In view of this, it is expected that the findings made in the study locations will not significantly change in the other locations except where the groundwater quality is seriously degraded by industry related activities (UNEP, 2011).

Acknowledgments

Development Goals (MDG) initiative of the Akwa Ibom State Government who provided us with data is greatly acknowledged.

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