



Investigation of geothermal potential of the Dahomey basin, Nigeria, through analysis of geomagnetic and geo-resistivity dataset

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ABSTRACT

Geothermal resource has attracted industrial and environmental interest in the last decades. However, the thermal condition of the Dahomey basin, which is instrumental to harnessing such geothermal resource, has remained largely unknown. The geothermal characterisation of Dahomey Basin was therefore undertaken to determine its thermal potential. The methodology involved analysis of the power spectra density of aeromagnetic data and interpretation of geo-resistivity data. The aeromagnetic dataset was divided into twenty one blocks with each block overlapping the adjacent blocks by 50%. Spectra peak, Curie depth, geothermal gradient, heat flow and temperature at depth were computed. Two-dimensional geo-resistivity profiling method was implemented to locate the top of the thermal aquifer. The results showed varied Curie depth (11–27 km) and heat flow (53–130 mW/m²) while the geothermal gradient ranges from 21 to 52 °C/km. Results of resistivity profiling showed the thermal aquifer to be sand (1–20 Ωm) whose top is located at 155 – 210 m depth. From the estimated geothermal parameters, two new geothermal prospect areas were identified. This study established that Dahomey basin is a thermally unstable basin with a very high potential for geothermal resource that is attributable to crustal thinning and possibly mantle dynamics.

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1. Introduction

Exploration of geothermal potential of a region via analysis of variations in Curie Point Depth (CPD) has attracted considerable interest in literature (Espinosa-Cardena and Campos-Enriquez 2008; Rajaram et al. 2009; Trifonova et al. 2009; Karastathis et al. 2011; El-Nabi 2012; Arnaiz-Rodríguez and Orihuela 2013; Bakak et al. 2015; Akbar and Fathianpour 2016; Bilim et al. 2016; Khojamli et al. 2016; Lichoro et al. 2019). Growing environmental consciousness over hydrocarbon consumption in the recent time has motivated awareness in environmental friendly energy supply such as geothermal energy. The Dahomey basin (Figure 1) is located in the southwest Nigeria where volcanicity and seismicity are rare. Several boreholes (Figure 2) located within this basin have however encountered thermal water. In spite of this observation, the knowledge of thermal condition of the basin which is instrumental to exploitation of such geothermal resource has remained poor till date.

Crustal rocks exhibit varying degrees of magnetism. This magnetic property can however be lost at certain depth known as the Curie depth point depth (CPD). Calculating the CPD and identifying regions where the crust is thin are considered vital in exploration of geothermal assets. Curie temperature has been established to range between 500°C and 600°C for the upper lithosphere (e.g. Nagata 1961; Okubo et al.

1985; Tsokas et al. 1998; Chiozzi et al. 2005; Manea and Manea 2011). At elevated temperature higher than Curie's, magnetic minerals change from ferromagnetic to paramagnetic condition. Diamagnetic and paramagnetic states do not contribute to the magnetic field of the earth (Nagata 1961; Chiozzi et al. 2005; Manea and Manea 2011). Information about CPD can be obtained through analysis of the low-frequency component of earth's magnetic data. The CPD-based data can therefore be employed to interpret the regional geothermal configuration of an area. Geothermal resources are beneficial in that they are localised, renewable and environmentally friendly. Exploration of these geothermal resources is conventionally based on direct and indirect methods.

Direct method entails measurement of temperature in boreholes via specialised equipment (e.g. Ebrahimi et al. 2019). However, this method is expensive and cumbersome due to paucity of thermally well-equilibrated boreholes. Direct method is also problematic in dealing with white noise. Indirect methods entail remote measurement of temperature with geophysical method such as magnetometry (Okubo and Matsunaga 1994). When compared to the direct method, indirect methods are cheap, fast and offer a wider coverage.

The utilisation of the magnetic method in Curie depth interpretation is predicated in the work of

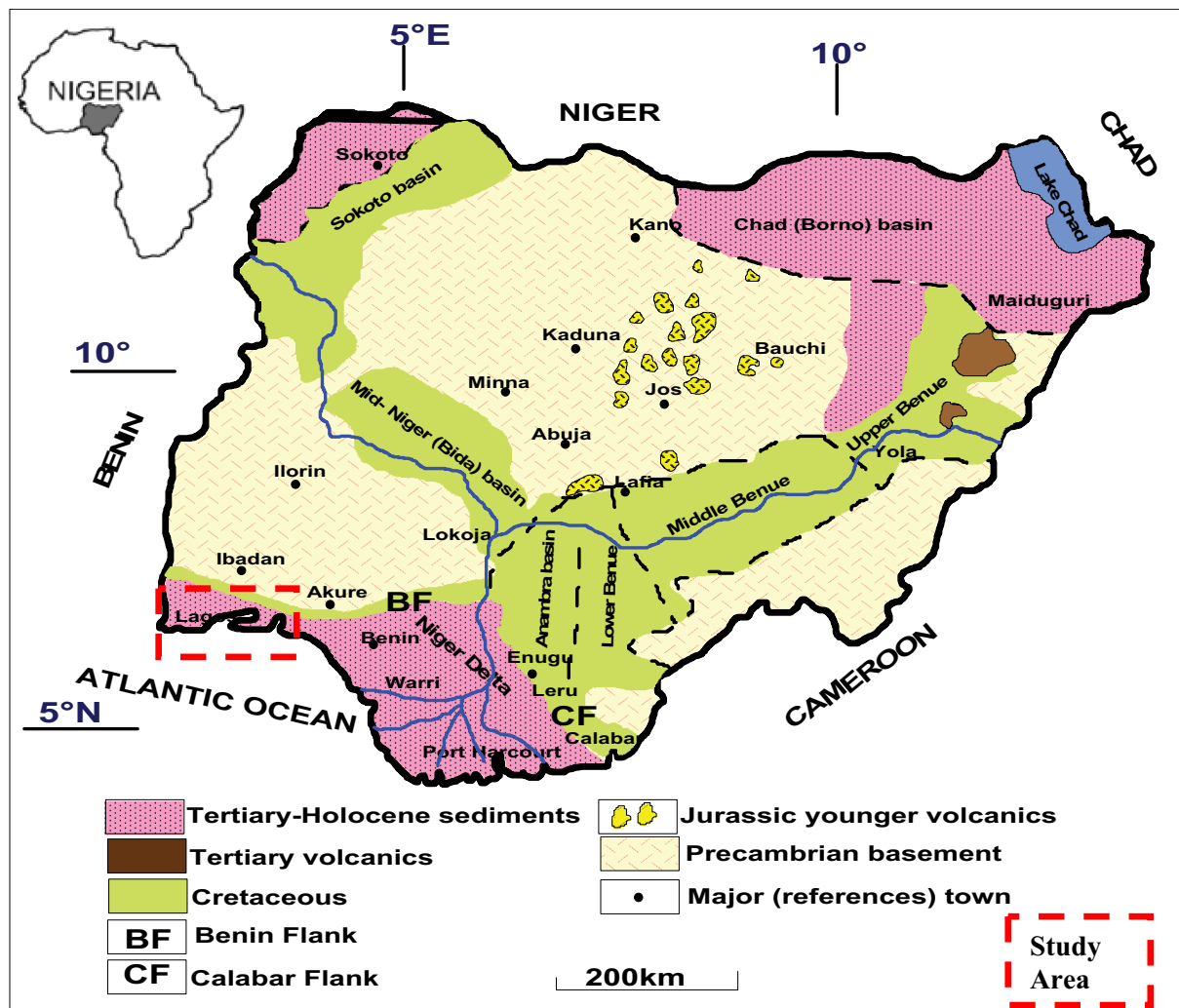


Figure 1. The study area located on Nigeria geological map (Modified After Obaje 2009).

Bhattacharyya (1966) and proliferated by several authors (e.g. Spector and Grant 1970; Mishra and Naidu 1974; Byerly and Stolt 1977; Connard et al. 1983; Blakely 1995; Hamdy and Khaled 2009).

(Tanaka et al. 1999; Salem et al. 2000; Saibi et al. 2015; Bilim et al. 2017; Lichoro et al. 2019) among others. These workers determined CPD through analysis of power spectrum of magnetic anomalies.

Power spectrum offers a reasonable connection linking the power spectrum of magnetic anomaly and the magnetic ensemble depth through Fourier conversion of the magnetic records to frequency domain. Ravat et al. (2007) discussed techniques that are typically used to estimate the CPD. They include centroid technique (Bhattacharya and Leu 1975; Okubo et al. 1985, 1989; Tanaka et al. 1999; Ibrahim et al. 2005), the spectral peak approach (Spector and Grant 1970; Shuey et al. 1977; Connard et al. 1983; Blakely 1995, 1995; Ross et al. 2006; Rajaram et al. 2009; Bilim et al. 2016), power law correction (Pilkington and Todoeschuck 1993; Maus and Dimri 1995; etc.). The spectral peak approach was employed in this study due to its capability to indicate the appropriateness of the data at hand for CPD analysis. It is

inappropriate to determine CPD if the signals in a data are significantly contributed from shallow sources (Rajaram et al. 2009). Interpretation of CPD is only achievable if the source bottom is measurable. In spectral peak method, a distinct spectral peak occurs when the extents of the chosen windows are sufficiently large and the source bottom is measurable (Bhattacharyya 1966; Salem et al. 2000). This approach enjoys simplicity in application, reproducibility, accuracy and production of geologically plausible information (Bilim et al. 2016). On the other hand, electrical resistivity disparity that exists between subsurface lithologies is often employed to differentiate between aquiferous and non permeable horizons (Dodds and Ivic 1988; Lashkarripour 2003; Delhay et al. 2019; Mandal et al. 2019). Accordingly, electrical resistivity method will prove a veritable tool in identification of the thermal aquifer in the study area.

CPD mapping to determine geothermal potential has been extensively employed in various regions across the globe (e.g. Shuey et al. 1977; Onwuemesi 1997; Stampolidis and Tsokas 2002; Li, 2011; Ates, et al., 2005; Dolmaz et al. 2005; Ibrahim et al. 2005; Tanaka and Ishikawa 2005; Chiozzi et al. 2005; Bilim,

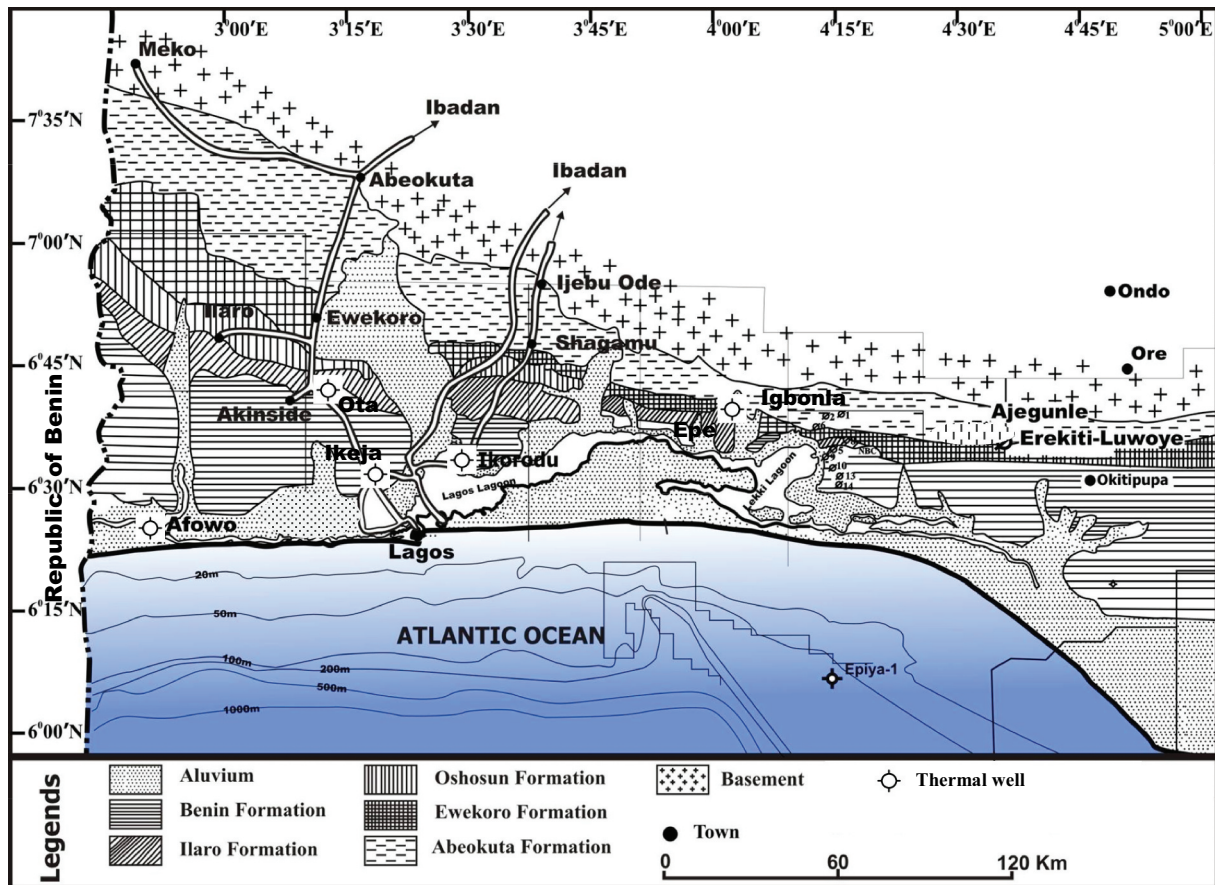


Figure 2. Geological map of Dahomey basin with locations of the thermal wells.

2011; Karastathis et al. 2011; Bektaş 2013; Obande et al. 2014) however, the knowledge of the Dahomey basin geothermal potential remains unknown due to paucity of exploration in that regard. In spite of existence of artesian thermal wells in the basin, little or no attention has been paid to the thermal conditions of this basin. Therefore, this work is committed to appraise the geothermal potential of the basin through application of spectral peak analysis of aeromagnetic and geo-resistivity data acquired from the study area.

2. Geology

The study district lies within the Dahomey basin (Figure 2) and has no basement rock outcrop. Comprehensive discussions of the geology of the Dahomey basin have been done by many researchers which include Jones and Hockey (1964); Ogbe (1972); Omatsola and Adegoke (1981); Frankl and Cordry (1967); Whiteman (1982); Adediran and Adegoke (1987).

The origin of the Dahomey basin has been linked with the parting of African and South American plates following the rifting that opened up the Atlantic in the Mesozoic. Basement fracturing happened in the Jurassic to Cretaceous which led to block faulting, disintegration and sagging of basement rocks (Adediran and Adegoke 1987). The

drift stage trailed in the Upper Cretaceous to Tertiary with episodes of block faulting which subsequently developed to horst and graben structure (Omatsola and Adegoke 1981). In the Santonian (late Cretaceous) there was occurrence of folding, tilting and block faulting probably related to relative slippages of two parts of the African plates (Burke et al. 1971). Existing boreholes located close to fault lines have encountered hot water and mineralisation such as sulphides as pyrites and galena, both of which are associated with faulting.

Oladele and Ayolabi (2014) estimated basement depth range of 0.183–6.3870 km in the basin while Coker and Ejedawe (1987) and Zaborski (1998) observed a consistent increase in basement depth in strike direction at the Nigeria – Benin Republic border in a manner typical of horst and graben structures. Cretaceous succession of Dahomey basin started with the Abeokuta Group, which is made up of Ise, Afowo and Araromi Formations.

2.1. Theory of Spectral Peak Analysis

The power spectrum of magnetic field anomaly (ΔT_{xy}) in Fourier domain was introduced by Blakely (1995) as:

$$\Phi_{\Delta T}(k_x, k_y) = \Phi_M(k_x, k_y) \cdot 4\pi^2 C_m^2 |\theta_m|^2 |\theta_f|^2 e^{-2|k|Z_t} \left[1 - e^{-|k|(Z_c - Z_t)}\right]^2 \quad (1)$$

This equation stands for 2-D power spectrum of potential field averaged within rings that share common origin. $\Phi_{\Delta T}$ and Φ_M are spectra density of the total field and magnetisation correspondingly. C_m is the proportionality constant while k is the wavenumber ($k^2 = k_x^2 + k_y^2$, k_x and k_y are the wave numbers in their respective directions), Z_t and Z_c are the respective depths to top and base of the source. The equation can be simplified by noting that all terms except $|\theta_m|^2$ and $|\theta_f|^2$ are radially symmetric and that radial averages of Θ_m and Θ_f are constants.

Transformation of Eq. (1) to 1-D power spectrum will give:

$$\Phi_{\Delta T}(|k|) = A \Phi_M(|k|) e^{-2|k|Z_t} \left[1 - e^{-|k|(Z_c - Z_t)}\right]^2 \quad (2)$$

Where A is a constant that depends on Θ_m and Θ_f and k is the wave number.

If the magnetisation ($M(x, y)$) is entirely arbitrary and unconnected, $\Phi_M(k_x, k_y)$ is an invariable Equation (2) can then be written as:

$$\Phi_{\Delta T}(|k|) = B e^{-2|k|Z_t} \left[1 - e^{-|k|(Z_c - Z_t)}\right]^2 \quad (3)$$

B is a constant. This equation represents 1-D spectrum of the field and is characterised by a maxima whose position is dependent on depth to top and base of the magnetic source and whose amplitude is a function of source magnetisation. Equation (3) becomes equation (4) when the natural logarithm of equation (3) is taken.

$$\ln \Phi_{\Delta T}(|k|) = \ln B - 2|k|Z_t + 2 \log \left[1 - e^{-|k|(Z_c - Z_t)}\right] \quad (4)$$

Where $(Z_c - Z_t)$ is the depth extent of the source. At less than 2 $(Z_c - Z_t)$, the curve of equation (4) behaves as a straight line whose gradient is the same as $-2Z_t$. According to Blakely (1995), depth to source top can subsequently be obtained from the spectrum of the observed magnetic field (ΔT_{xy}).

The spectral peak of infinitely deep prism occurs at wavenumber zero (Bhattacharyya 1966). Prism of a finite top and base will produce a peak (K_{\max}) in the power spectrum. When this happens, it signifies that the source base is measurable. The peak wave-number is associated with the depth base and top of the magnetic entity, Z_c and Z_t , by the following relation (Blakely 1995):

$$K_{\max} = \frac{\ln Z_c - \ln Z_t}{Z_c - Z_t} \quad (5)$$

The peak in the spectrum would be observed at wave-number $k > k_f$, where K_f is the fundamental wavenumber:

$$K_f = \frac{2\pi}{W} \quad (6)$$

3. Methodology

3.1. Aeromagnetic data

The data utilised for this study is part of 2005 aeromagnetic survey carried out by Fugro for Nigeria Geological Survey Agency. The data is characterised by Flight and tie line spacing of 500 m and 2 km in NW-SE and NE-SW respectively with 80 m terrain clearance. The International Geomagnetic Reference Field (IGRF) for year 2005 has been subtracted from the data.

To carry out the geothermal characterisation of the study area, the aeromagnetic data was transformed to the equator using inclination of -12.715° and declination of -2.763° signifying the geomagnetic parameters at the centre of the study area. Determination of depth to magnetic source bottom (Z_c) is usually difficult because the signals from the shallower (Z_t) parts typically saturate power spectral at all wavelengths. To minimise this difficulty, the reduced to the Equator (RTE) grid was upward continued (2 km) to accentuate spectral from bottom depth at the expense of shallow sources contributions before computing the power spectrum of the aeromagnetic grid. The smoothened grid was carved up into 21 overlapping blocks, each having 55×55 km dimension (Figure 3). This size was selected after a number of trials, being the grid at which obvious spectral peak occurs in the power spectra, signifying a detectable source bottom. Every block partly covers half of the flanking blocks.

3.2. Calculation of power spectrum

The spectral of each block was calculated using power spectrum algorithm of Oasis Montaj (Geosoft (Oasis Montaj) 2007).

3.3. Curie-Point Depth (CPD) Estimation

Spectral Peak approach of Blakely (1995) was adopted in the determination of CPD. This method involved determination of wavenumber of the spectral maximum (K_{\max}) and evaluation of depth to deep seated magnetised source top. Depths to top of sources (Z_t) were determined from the low wavenumber section of the power spectrum. The average CPD (Z_c) was computed through the use of K_{\max} and Z_t depth values for each sub-region using the equation by (Blakely 1995)

$$K_{\max} = \frac{\ln Z_c - \ln Z_t}{Z_c - Z_t} \quad (7)$$

Where \ln represents the natural logarithm, K_{\max} signifies spectral peak; Z_t and Z_c are mean depth to top and base (Curie point depth) of magnetic sources respectively. The resulting CPDs were gridded and contoured to produce the CPD map of the study area. It has been established that Curie point depth

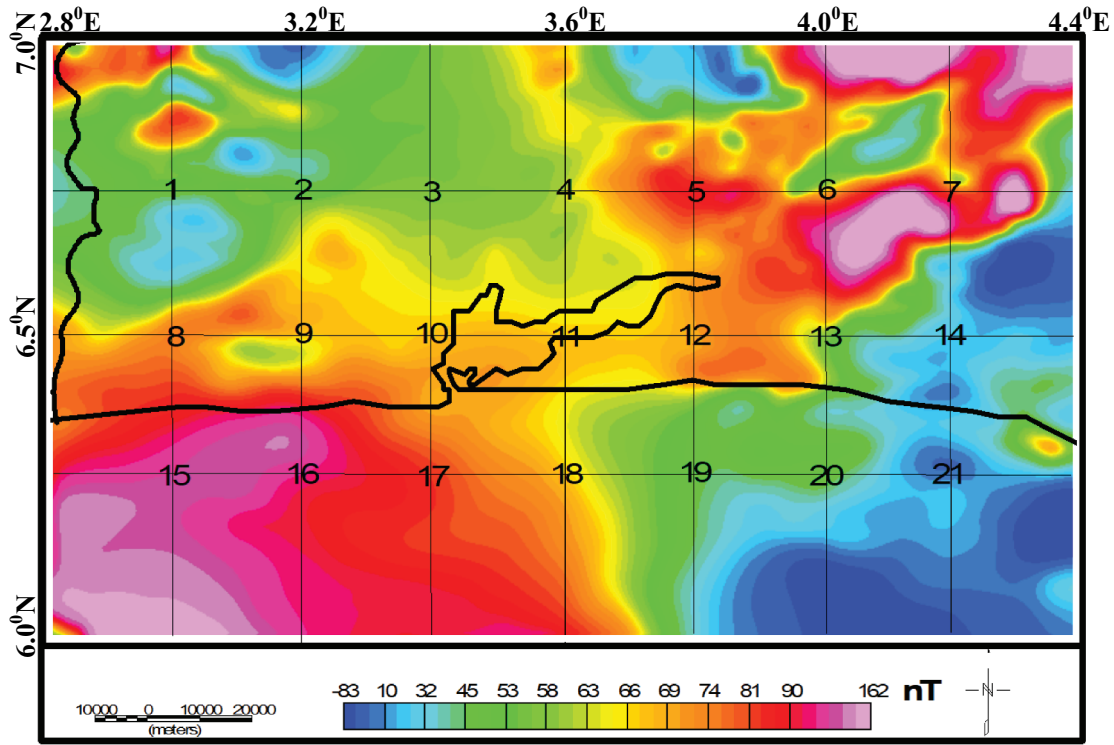


Figure 3. Reduced to the equator (RTE) map has been subtracted from the data). Blocks are indicated by the figure at its centre. The coastline (thick black line) has been superimposed.

that is shallower than 10 km are usually designated as geothermal anomalous zones (Tanaka et al. 1999). The vertical thermal gradient of the blocks were computed thermal supposing that magnetite (Curie temperature = 580°) is the dominant magnetic mineral in those rock.

$$\text{Vertical Thermal Gradient} \left(\frac{dT}{dZ} \right) = \frac{580^\circ\text{C}}{Z_C} \quad (8)$$

Determination of heat flow was accomplished through the heat equation of Fourier's law (Turcotte and Schubert 1982; Artemieva and Mooney 2001)

$$q = \left(\frac{dT}{dZ} \right) \quad (9)$$

Where q , k , T and Z are heat flow, coefficient of thermal conductivity, temperature and Curie depth respectively. The assumption in this equation is that temperature varies vertically and its gradient (dT/dZ) is constant. For this work, anomalous geothermal condition was attributed to values greater than 100 mW/m² (Jessop et al. 1976). Regions of high heat flow were demarcated as geothermal prospective zones. The thermal conductivity 2.1 Wm⁻¹⁰C⁻¹ was adopted as recommended by Tezcan and Turgay (1991).

Having assumed a linear temperature changes within the earth (Onwuemesi 1997), temperature at depth h of interest was computed as

$$T_h = mh + T_0 \quad (10)$$

Where T_h = temperature in °C at depth (h), m = geothermal gradient, T_0 = surface temperature (assume to be 27°C). Using the geothermal gradient of each block, the corresponding Temperature at 1 km depth was computed. The depth to bottom of magnetised source was interpreted as CPD.

3.4. Delineation of thermal aquifer

With a view to map depth to the top of thermal aquifer, geo-electric strata in the study area were imaged through geo-resistivity method. The geo-resistivity survey was carried out using a SuperSting R8-IP Resistivity metre comprising of 84 electrodes separated by 10 m spacing. Pole-Dipole configuration was employed for data acquisition because it provides a blend of good depth penetration and lateral resolution. The infinite electrode for the Pole-Dipole configuration was at over 1000 m from the first electrode. The raw data retrieved from the equipment was processed to generate the 2-D resistivity tomography. Interpretation of 2D resistivity tomography was integrated with available gamma ray logs.

4. Results and discussion

Figure 4 portrays samples of spectra of magnetic data of blocks (blocks 1, 5, 11 and 21) with their

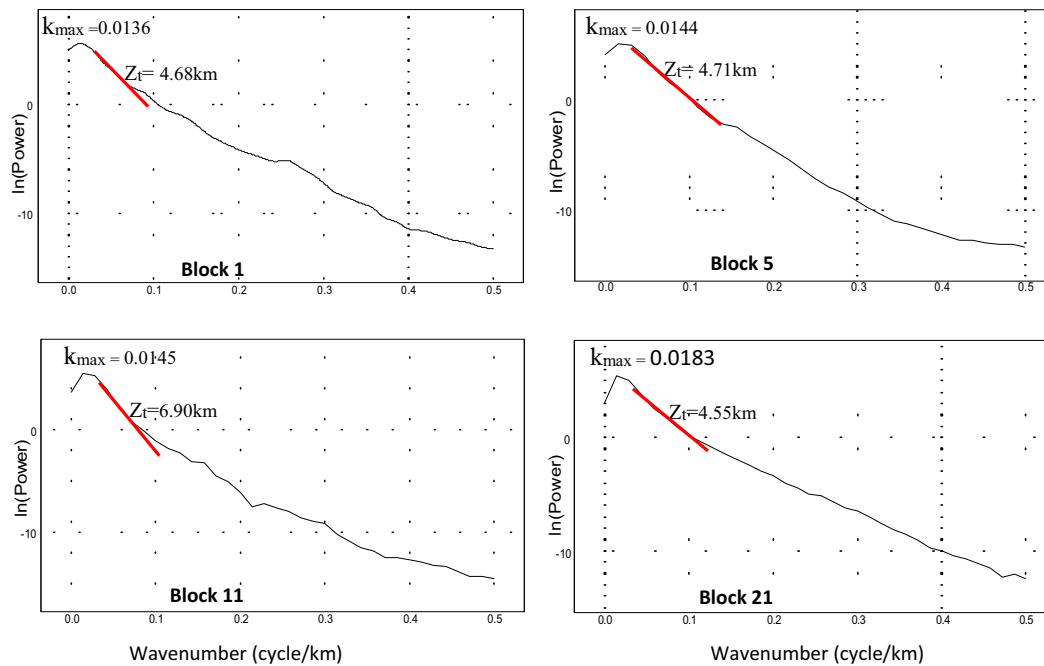


Figure 4. Radially Averaged Power Spectral of Representative Blocks 1, 5, 11 and 21 with their Corresponding Depth (Z_t) Estimates.

corresponding depth estimates. They show characteristic decay of power with increasing wavenumber. The frequency K_{max} , average source top depth (Z_t) and average source base depth (Z_c) for the study area are presented in Table 1.

4.1. Curie point depth and heat flow

The CPD map of the study area (Figure 5) reveals two shallow CPD zones: the central area of the Lagos metropolis (Ikeja and Ikorodu) and the other in the northeast of Igbonla. The two shallow CPD regions are named Lagos and Igbonla geothermal prospects respectively. This observation is most likely related to

a thin crust necessitated by rifting associated with the opening of the basin. The Lagos thermal dome in the centre of the study area gradually diminishes towards east and west directions. The thermal anomaly is about 50 km wide in E-W direction and extends outside the coverage of the map in N-S directions respectively. The map shows that the shallow CPD regions are strikingly correlated with thermal water wells (Figure 5) thus suggesting that the occurrence of geothermal water in those areas except the Afowo Well is genetically related to shallow Curie depth observed in the area.

The presence of thermal water in Afowo Well may be due to transmission of thermal water from the

Table 1. Spectra peak (K_{max}), average source top depth (Z_t) and average source base depth (Z_c), geothermal gradient (GRAD) and heat flow.

Block No	K_{max} (Cycle/km)	Z_t (km)	Z_c (CPD) (km)	GRAD ($^{\circ}\text{C}/\text{km}$)	HEAT FLOW (mW/m^2)
1	0.0136	4.68	24	25	61
2	0.0158	4.30	20	30	74
3	0.0153	6.25	16	36	90
4	0.0152	6.33	16	36	90
5	0.0144	4.71	21	27	68
6	0.0184	5.85	12	47	119
7	0.0179	6.01	13	46	115
8	0.0121	5.19	27	22	54
9	0.0140	5.76	20	29	73
10	0.0199	5.50	11	52	130
11	0.0145	6.90	16	35	88
12	0.0150	4.77	20	29	73
13	0.0079	7.20	19	30	74
14	0.0121	4.97	27	21	53
15	0.0129	5.21	24	24	60
16	0.0068	6.66	19	30	75
17	0.0152	5.74	17	34	84
18	0.0147	6.22	17	34	84
19	0.0145	6.25	18	33	82
20	0.0145	5.17	20	29	72
21	0.0183	4.55	15	39	98

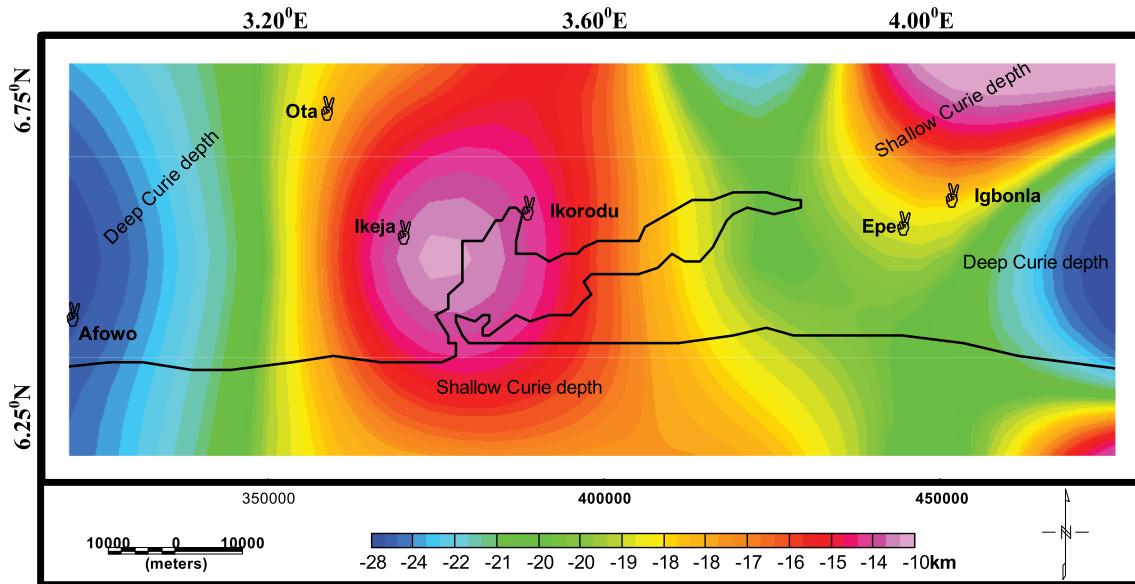


Figure 5. Curie depth map. The coastline and thermal well locations are superimposed on the map.

region of thermal dome to Afowo well by some of the fractures in the area.

The heat flow map with superimposed thermal well locations is shown in Figure 6. The heat flow generally varies from 53 to 130 mWm⁻² and shows the expected inverse relationship with the CPD. Yamano (1995) observed that shallow CPDs are in agreement with elevated heat flow. Elevated heat flow exists in Ikeja, Ikorodu and Igbonla areas in the central and northwestern regions respectively, while the western region is essentially characterised by low heat flow. The mean heat flow in the crust is about 60 mW/m² and records above 80–100 mW/m² point to thermal anomaly situation (Jessop et al. 1976). The thermal water wells are in striking association with areas of high heat flow except the Afowo well.

The chart of heat flows (q) versus Curie point depths (Z_c) values (Figure 7) shows hyperbolic relationship named Ayolad thermal equation: $Z_c = 0.002q^2 - 0.613q + 52.91$.

Zones of shallow CPD in the study area are associated with high temperature gradients (Figure 8) which is above the normal geothermal gradient of 30°C/km. Therefore, this district shows potential for possible exploitation of geothermal resource. The geothermal gradient is highest within the Lagos graben and is envisaged to have been strongly influenced by the thermal dome of the Lagos geothermal system. Appropriate geothermal gradient within the basin would have aided generation of hydrocarbon. Staplin (1977) noted that the range of hydrocarbon generation is from 65°C to 145°C for oil and up to 165° for gas respectively. This temperature range will be obtainable

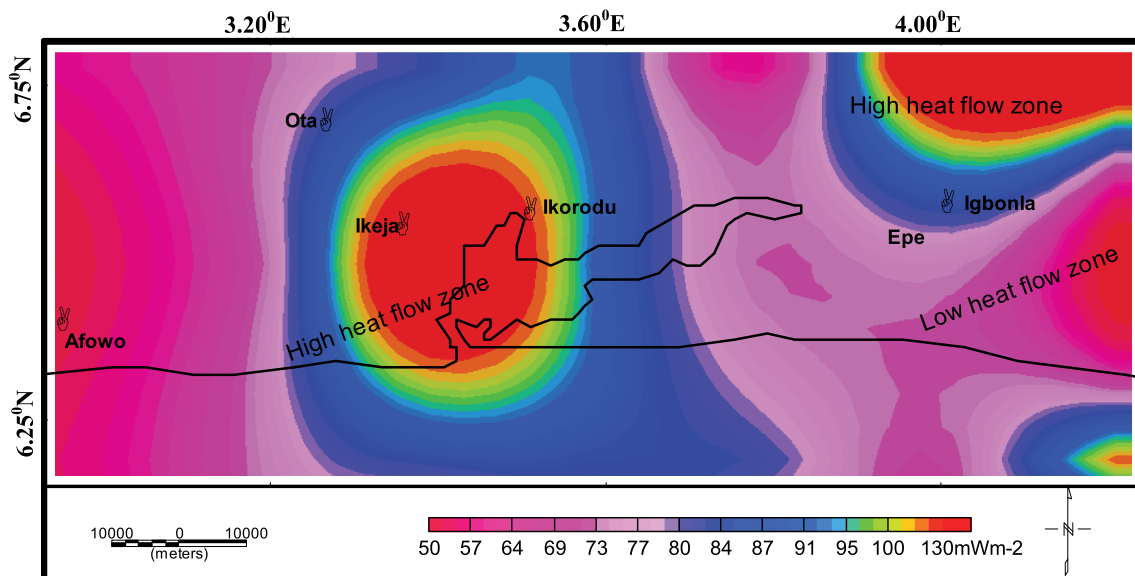


Figure 6. Heat flow map. The coastline and some thermal wells locations are superimposed on the map.

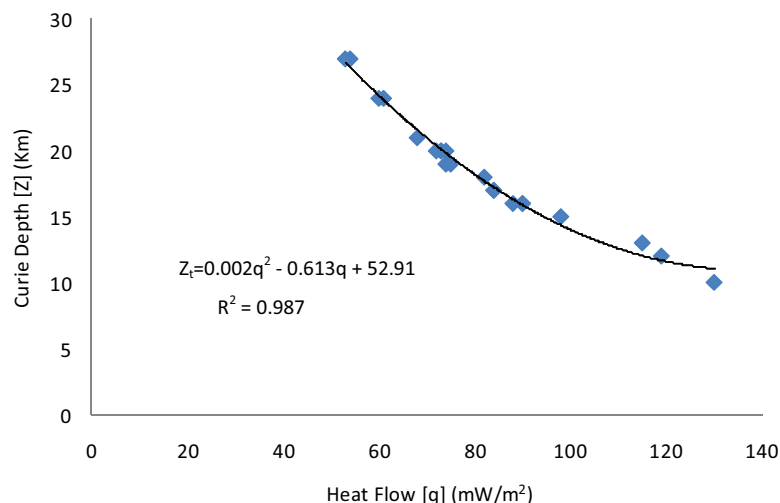


Figure 7. Curie depth versus heat flow plot for Dahomey Basin.

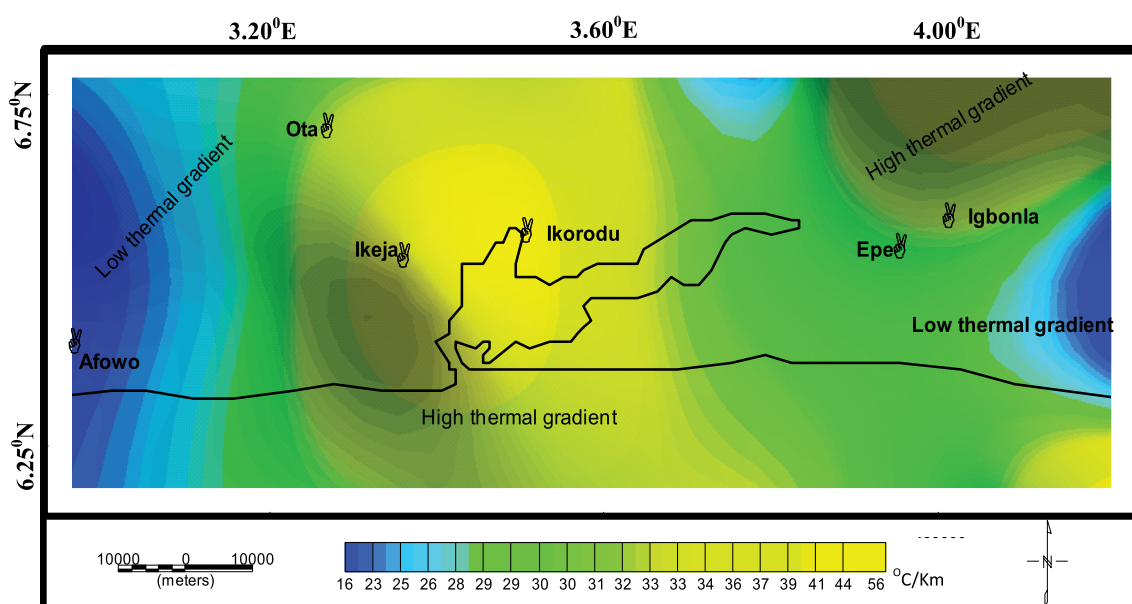


Figure 8. Simplified thermal gradient map of the study area.

at depth between 1 to 2.2 km. Given that the sediment thickness is up to 4 km within the Lagos graben, thermal maturity of the hydrocarbon source rock within the graben is therefore expected. This observation further laid credence to the high petroleum prospectivity of the Lagos graben. Figure 9 shows the map of estimated temperature at 2 km depth. At 2 km depth temperature varies from 72 to 134 °C. High temperature is observed at the locations of the earlier delineated thermal domes.

4.2. Geothermal prospect and structural system

Superposition of the principal structures mapped in the study area (Oladele et al. 2016) on the CPD map (Figure 10) shows that the fractures are cutting through the thermal anomalies. The structures which vary in length and directions will have significant

influence on hydrothermal reservoir and the existence of thermal water in areas far from any of the geothermal domes. The faults and fractures that cut across the Lagos and Igbonla geothermal prospects (Figure 10) will provide permeability and buoyancy around the geothermal prospects. Fluid flows in geothermal reservoirs occur mainly through fractures of various lengths and widths (Grant and West 1965). The presence of thermal water in Afowo Well may be due to transmission of water from the region of thermal dome to Afowo well by some of the fractures in the study area. These structures will generally increase the permeability of rocks and control the flow of fluid.

4.3. Correlation between depth map and CPD

Figure 11 shows the superposition of CPD map on the depth to basement map (Oladele and Ayolabi 2014).

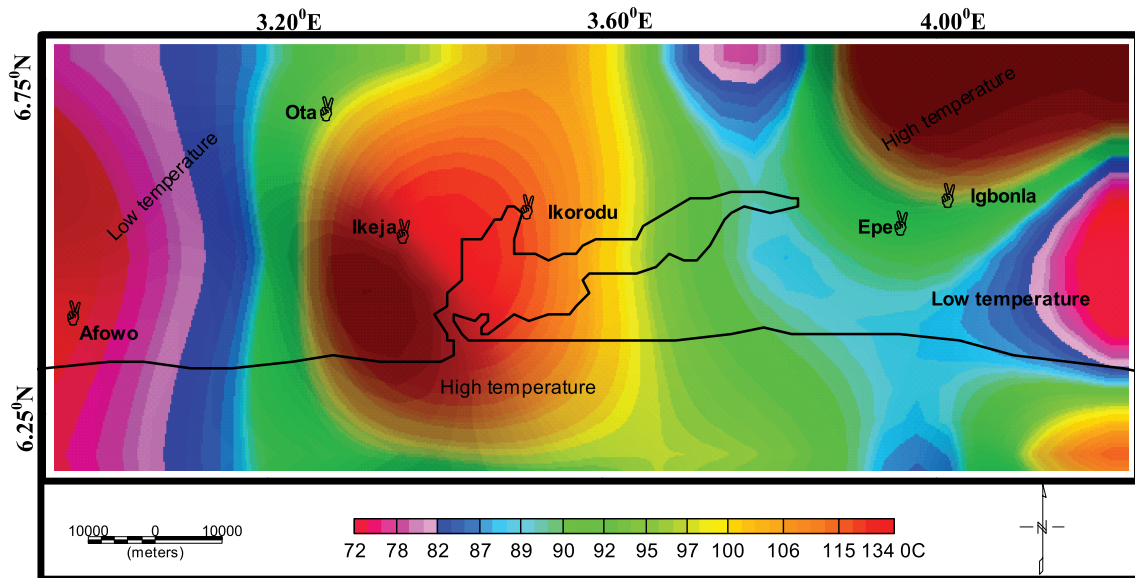


Figure 9. Map of temperature estimate at 2 km depth.

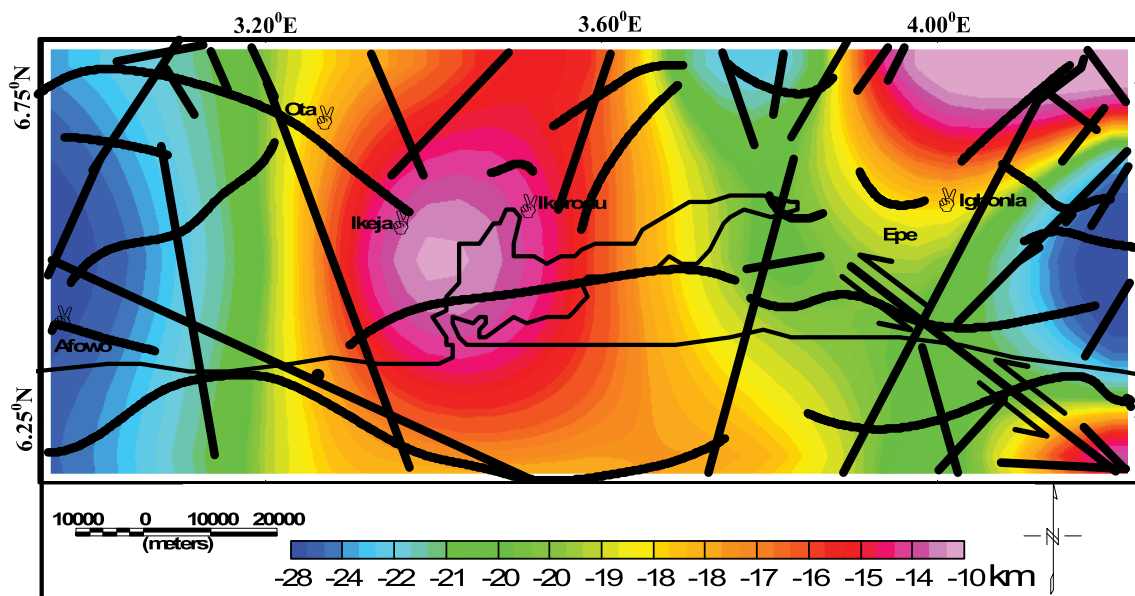


Figure 10. Superposition of structures (Oladele and Ayolabi 2014; Oladele et al. 2016) on the CPD map.

This map revealed that the central thermal dome is related to the area of basement depression known as Lagos graben. There is thick sediment (4 km) above this shallow CPD zone. This area has been tectonically inactive and in general has had no intrusive or extrusive activity. The association of shallow CPD with deepest parts of the basin suggests that there is genetic relationship between the thermal anomaly and the process that produced the Lagos graben (rifting). The present CPD may only reflect a stage in which the crust is reaching a geothermal stability.

4.4. Identification of thermal aquifer

2-D resistivity tomography result correlated with gamma ray logs (Figure 12) revealed the existence

of five layers: sandy clay (20–500 Ωm), clay/shale (10–40 Ωm), sand (150–1500 Ωm), limestone (20–80 Ωm) and sand (1–20 Ωm). The sand (1–20 Ωm) was interpreted as thermal aquifer due to its low gamma ray and low resistivity responses. The thermal aquifer horizon in these boreholes is regarded as Ise and Araromi formations which are members of the Abeokuta Group. The thermal aquifer is generally encountered at depth greater than 200 metres. Variation in depth to the top of the aquifer might have been caused by multiple dip slip faults that segmented the aquifer at different locations. The temperatures of the water produced from these aquifers ranged from 46–80 °C. In places, the groundwater occurs in artesian and sub-artesian conditions.

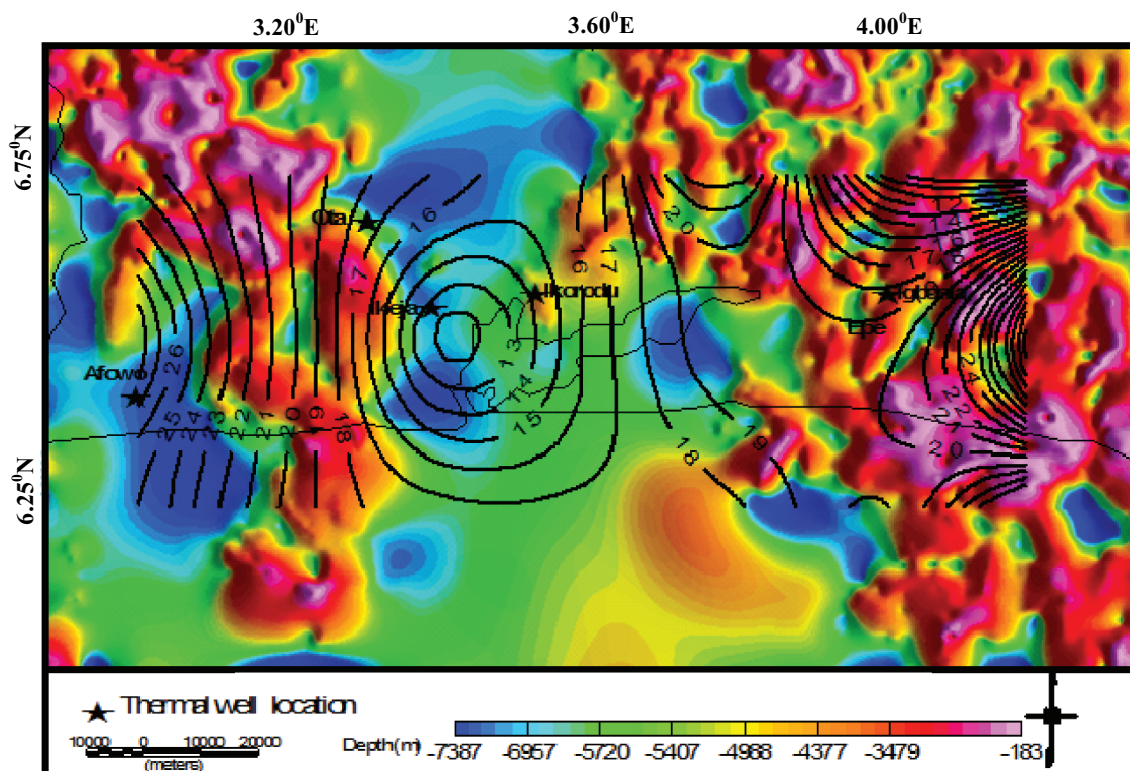


Figure 11. Superposition of depth to basement contours (Oladele and Ayolabi 2014) on CPD map.

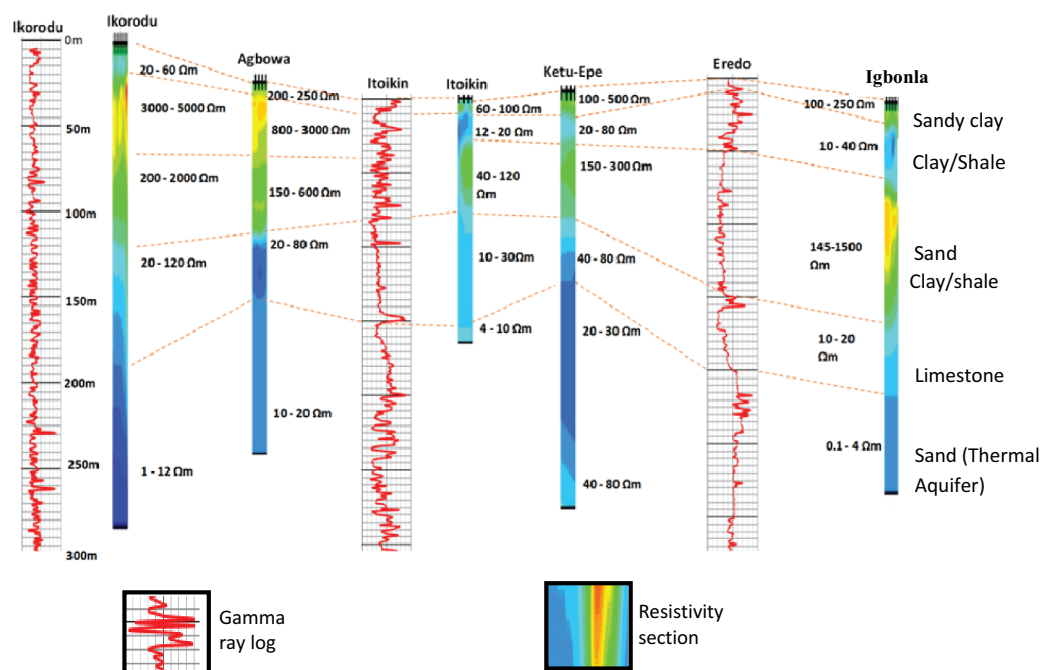


Figure 12. 2-D resistivity tomography, correlated with gamma ray logs, showing the geoelectric layers and the top of the thermal aquifer.

4.5. Origin of thermal water

The thermal domes are viewed to be associated with lithospheric stretching which leads to the thinning of the of the crust and consequently, rifting/block faulting and asthenospheric upwelling. The

uplifted asthenosphere brought hot magmatic melt unusually closer to the surface and this has been dissipating heat to the water hosted in overlying aquifers. Perforation of this deep aquifer will usually result in mining of thermal water as observed in deep thermal wells around Lagos.

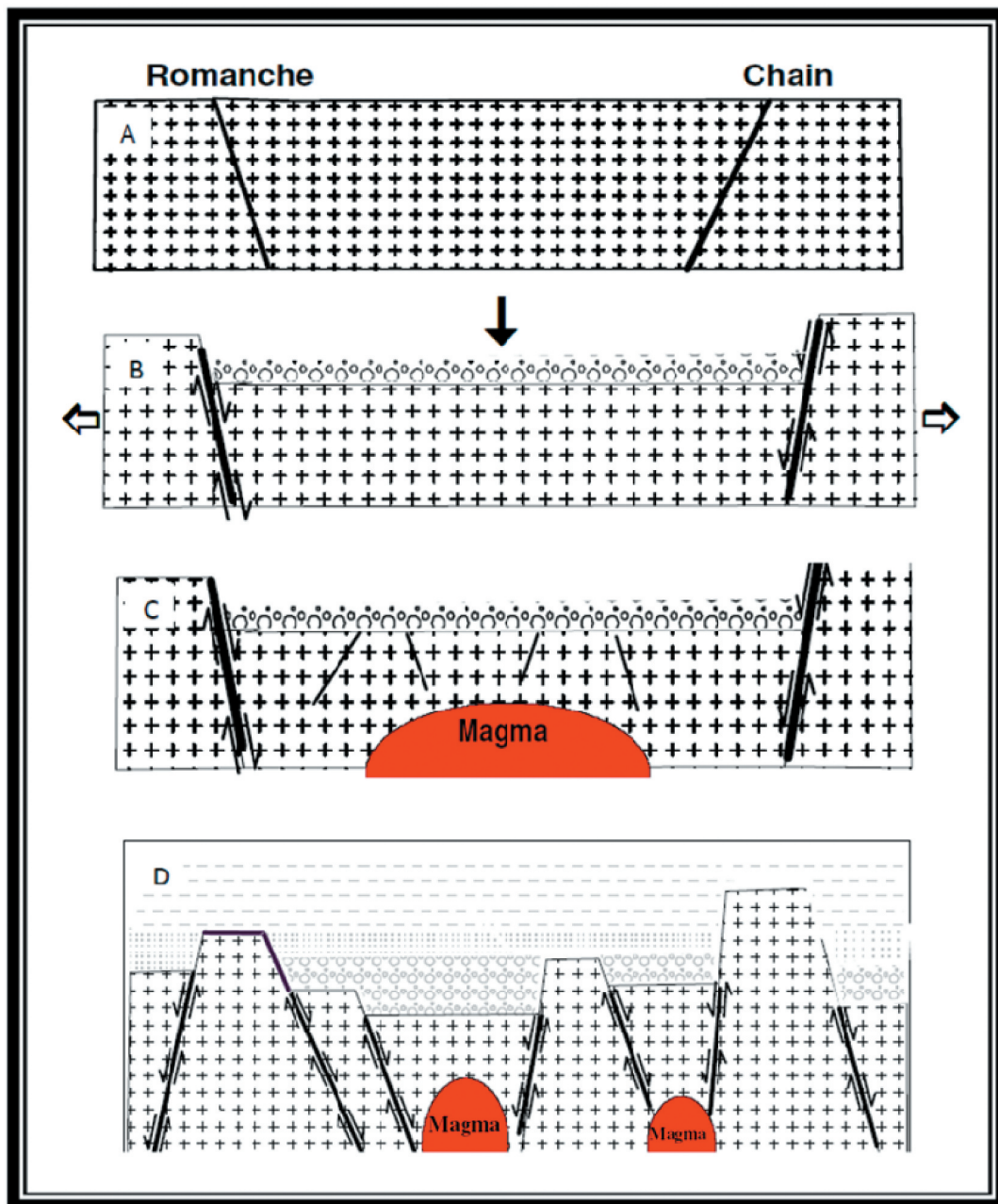


Figure 13. Conceptual model proposed to explain the occurrence of thermal water in Dahomey Basin. A: Opening of the Atlantic Ocean and formation of transform faults B: Lithospheric stretching/continental thinning C: Asthenospheric upwelling D: Faulting (rifting) and differential subsidence.

4.6. Geothermal model

On the basis of results earlier presented, the Dahomey Basin geothermal model (Figure 13) is hereby proposed. According to Mary and Mario (2004), hydrothermal systems generally consist of heat source, aquifer, recharge and conduits (e.g fractures and faults). The heat source in this basin is the mantle dome arching to shallow depth due to crustal thinning. The reservoir rock is the permeable conglomeratic Ise Formation which lies unconformably on the basement. The Afowo shale (impermeable cap rock) overlay the Ise Formation, thus making it a confined thermal aquifer. The water in the aquiferous Ise

Formation mine heat from the underlying heat source and circulate same. The aquifer is partly recharged by percolating meteoric water and water from the Atlantic Ocean. The delineated subsurface network of fractures and faults serve as the conduits for migration of hydrothermal fluid.

5. Conclusion

Occurrence of several thermal water producing boreholes around the Dahomey basin, southwestern Nigeria, remains largely uninvestigated to date. This study has produced the geothermal characteristics of

the Dahomey basin for the first time in literature. The reported characteristics are based on the spectral interpretation of aeromagnetic and geo-resistivity data. Dahomey basin is characterised with CPD values (11–27 km), geothermal gradient (21–52 °C/km) and heat flow (53–130 mW/m²). Two prospective geothermal anomalous regions are present in the basin: The Lagos anomalous region which coincides with Lagos graben and the Igbonla anomalous region which coincides with area characterised by interconnected system of deep faults running through the area. 2D resistivity results revealed geo-electric layers which include a thermal sand aquifer whose top lies at depth ranging from 155 to 200 m. The mechanism responsible for the elevated heat flow is believed to be associated with rift induced magmatism. The crust is possibly heated by hot materials transported upward by convection in response to thinning and sagging of the crust. This study therefore presents Dahomey basin as having great and exploitable potential for geothermal resource.

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Disclosure statement

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