Regional estimation of Curie-point depths and succeeding geothermal parameters from recently acquired high-resolution aeromagnetic data of the entire Bida Basin, north-central Nigeria

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Abstract. A regional estimation of Curie-point depths (CPDs) and succeeding geothermal gradients and subsurface crustal heat flow has been carried out from the spectral centroid analysis of the recently acquired high-resolution aeromagnetic (HRAM) data of the entire Bida Basin in north-central Nigeria. The HRAM data were divided into 28 overlapping blocks, and each block was analysed to obtain depths to the top, centroid, and bottom of the magnetic sources. The depth values were then used to assess the CPD, geothermal gradient, and subsurface crustal heat flow in the basin. The result shows that the CPD varies between 15.57 and 29.62 km with an average of 21.65 km, the geothermal gradient varies between 19.58 and 37.25 °C km⁻¹ with an average of 27.25 °C km⁻¹, and the crustal heat flow varies between 48.41 and 93.12 mW m⁻² with an average of 68.80 mW m⁻². Geodynamic processes are mainly controlled by the thermal structure of the Earth's crust; therefore this study is important for appraisal of the geo-processes, rheology, and understanding of the heat flow variations in the Bida Basin, north-central Nigeria.

1 Introduction

This study aims at quantitative estimation of regional Curie-point depth (CPD) and succeeding geothermal structures, namely geothermal gradients and subsurface crustal heat flow anomalies, in the whole of Bida Basin in north-central Nigeria using the spectral centroid analysis of the recently acquired high-resolution airborne magnetic (HRAM) data. The HRAM surveys were carried out by Fugro Airborne Survey Limited for the Nigerian Geological Survey Agency (NGSA) between 2004 and 2009. Acquisition, processing, and compilation of the HRAM data were jointly financed by the Federal Government of Nigeria and the World Bank as part of the Sustainable Management for Mineral Resources Project (SMMRP) in Nigeria.

Several studies have shown that regional magnetic data can be used extensively to determine the thermal structure of the Earth's crust in various geologic environments (Spec-

tor and Grant, 1970; Bhattacharyya and Leu, 1975, 1977; Byerly and Stolt, 1977; Blakely and Hassanzadeh, 1981; Okubo et al., 1985, 2003; Blakely, 1988, 1995; Maus et al., 1997; Tanaka et al., 1999; Chiozzi et al., 2005; Eppelbaum and Pilchin, 2006; Ross et al., 2006; Ravat et al., 2007; Trifonova et al., 2009; Gabriel et al., 2011, 2012; Bansal et al., 2011, 2013, 2016; Nabi, 2012; Hsieh et al., 2014; Nwankwo and Shehu, 2015; etc.). For example, dominant magnetic minerals in the Earth's crust pass from ferromagnetic to paramagnetic state at temperature, commonly called Curie-point temperature (CPT). Magnetite (Fe₃O₄) is the most common magnetic material in igneous rocks and has an approximate CPT value of 580 °C (Stacey, 1977). At temperature above CPT, the thermal agitation causes the spontaneous alignment of the various domains in the mineral to be destroyed (or randomized) to the extent that the ferromagnetic minerals become totally paramagnetic (Langel and Hinze, 1998).

Figure 1. Geological map of Nigeria showing location of the Bida Basin (after Nwankwo and Shehu, 2009).

Therefore CPD, which is defined as the depth at which CPT is reached within the subsurface, can be considered as an index of depth to the bottom of magnetic sources (DBMS) and can consequently be calculated from geomagnetic anomalies (Bansal et al., 2011, 2013; Hsieh et al., 2014). However, in some circumstances DBMS can be caused by contrasts in lithology instead of CPT and may not necessarily coincide with CPT in detail (Bansal et al., 2011; Trifonova et al., 2009). For instance, Trifonova et al. (2009) opined that even if the spectral method provides a good estimate of DBMS there is no assurance that it represents the CPD. They reasoned that a variety of geologic reasons exist for truncated magnetic sources that are unrelated to crustal temperatures; for example, a sequence of relatively non-magnetic sediments below young volcanic material may limit the depth of magnetic sources regardless of the CPT, and another reason is the variety of magnetic minerals like titanomagnetite ($\text{Fe}_{3-x}\text{Ti}_x\text{O}_4$). Titanomagnetite is the most important iron oxide in crustal magnetic sources; it has a CPT that is strongly influenced by the amount of titanium and ranges from 150 to 580 °C. In some geologic environments, alloys of iron with CPTs in excess of 620 °C may be significant contributors to magnetic anomalies. In spite of these limitations, many studies (Tanaka et al., 1999; Trifonova et al., 2009; Bansal et al., 2011; Hsieh et al., 2014; etc.) have reasonably used DBMS as an estimate of CPD and therefore serve as a proxy for temperature at depth. Again, Trifonova et al. (2009) pointed out that several studies have identified low-titanium titanomagnetite as the dominant magnetic phase, and CPTs at these depths are estimated to be between 575 and 600 °C. This confirms the estimated value of 580 °C by Stacey (1977) as the case in this study. Another important justification is that DBMS/CPD estimations can similarly be used to complement geothermal data in re-

gions where deep boreholes are unavailable (Chapman and Furlong, 1992; Ross et al., 2006; Bansal et al., 2011, 2013).

The Bida Basin is the least studied of all Nigeria's inland frontier basins. To date, the basin has no information on seismicity, no exploratory wells have penetrated its sequences, and deep crustal data are limited (Obaje et al., 2015). Therefore, this present work is expected to contribute immensely to a better understanding of the geothermal structures and geodynamic processes in the entire Bida Basin in north-central Nigeria.

2 Location and geology of the study area

The Bida Basin (also known as the Middle Niger Basin or Nupe Basin) is an elongated NW–SE-trending depression perpendicular to the main axis of the Benue Basin of Nigeria. The entire basin (Fig. 1) is bounded by latitudes 8°00' N and 10°30' N and longitudes 4°30' E and 7°30' E, and it covers an area of approximately 90 750 km².

The geology of the Bida Basin is believed to be a gentle down-warped shallow trough filled with Campanian–Maastrichtian marine to fluvial strata. The strata are also believed to be more than 300 m thick (Adeleye, 1976). Those with marine affinity, the limestones, often form cappings (under variable thickness of laterites) to the means of the basin. Some form prominent intermediate breaks of slope along the mesa walls. The buried basement complex probably has a high relief, and the sedimentary formations have been shown to be about 2 km thick (Obaje et al., 2015) with a constituted post-tectonic molasse facies and thin marine strata, which are all unfolded.

The basin might also be regarded as the north-western extension of the Anambra Basin, which is found in the south-east, both of which were major depocentres during the sec-

Figure 2. Residual total magnetic intensity map of the entire Bida Basin with superimposed federal survey half-degree sheets and showing major towns flown over. A constant TMI value of 33 000 nT was removed.

Figure 3. Sketch of procedure for achieving overlapping blocks.

ond major sedimentary cycle of southern Nigeria in the Upper Cretaceous (Obaje, 2009).

The stratigraphy of the basin consists of mainly Patti, Lokoja, and Agbaja formations (Ikumbur et al., 2013). According to Akande et al. (2005), Patti Formation is the only stratigraphic unit containing carbonaceous shale in the basin and is sandwiched between the older Campanian–Maastrichtian Lokoja Formation – which contains conglomerates, sandstones, and claystones – and the younger Agbaja Formation, which comprises mostly ironstones.

Tectonically, rift hypothesis has been proposed as the possible origin and evolution of the basin (Kogbe et al., 1983). Some researchers have inferred that the rifting in the basin started in the Upper Cretaceous and that the process started initially from the Benue Trough in the early Cretaceous and

eventually spread to other rifted neighbouring basins in the Cretaceous period (Kogbe et al., 1983).

3 Calculation of Curie-point depth, geothermal gradient, and heat flow

Calculations of DBMS, which is commonly construed as the Curie-point depth, are made using diverse methods (Spector and Grant, 1970; Shuey et al., 1977; Bhattacharyya and Leu, 1975, 1977; Okubo et al., 1985; Tanaka et al., 1999; Maus and Dimri, 1996; Finn and Ravat, 2004; Ravat et al., 2007; Bansal et al., 2013), namely the spectral-peak method (Spector and Grant, 1970; Shuey et al., 1977), the centroid method (Bhattacharyya and Leu, 1975, 1977; Okubo et al., 1985; Tanaka et al., 1999), the scaling spectral or power-law correction method (Maus and Dimri, 1996), forward modelling of the spectral-peak method (Finn and Ravat, 2004; Ravat et al., 2007), and the modified-centroid fractal method (Bansal et al., 2011; Nwankwo, 2015), among others. The conventional centroid method is used in this work for the reason that it gives better estimates with fewer depth errors compared to earlier methods (Ravat et al., 2007).

The mathematical models of the centroid method are based on the examination of the shape of isolated magnetic anomalies introduced by Bhattacharyya and Leu (1975, 1977) and the study of the statistical properties of magnetic ensembles by Spector and Grant (1970). Blakely (1995) subsequently introduced power spectral density of total magnetic field, $\phi_{\Delta T}(k_x, k_y)$, as

Figure 4. Azimuthally averaged power spectral plots for blocks 4 and 19.

$$\phi_{\Delta T}(k_x, k_y) = \phi_M(k_x, k_y) \cdot 4\pi^2 C_M^2 |\theta_M|^2 |\theta_f|^2 e^{-2|k|Z_t} (1 - e^{-2|k|(Z_b - Z_t)})^2, \quad (1)$$

where k_x and k_y are wave numbers in x and y direction, $\phi_M(k_x, k_y)$ is the power spectra of the magnetization, C_M is a constant, θ_M and θ_f are factors for magnetization direction and geomagnetic field direction, and Z_b and Z_t are depths to bottom and top of magnetic layer respectively.

If the layer's magnetization, $M(x, y)$, is a random function of x, y , it implies that $\phi_M(k_x, k_y)$ is a constant, and therefore the azimuthally averaged power spectrum, $\phi(|k|)$, would be given as

$$\phi(|k|) = A e^{-2|k|Z_t} (1 - e^{-2|k|(Z_b - Z_t)})^2. \quad (2)$$

The depth to the top of the magnetic source is therefore derived from the slope of the high-wave-number portion of the power spectrum as

$$\ln(P(k)^{1/2}) = A - |k|Z_t, \quad (3)$$

where $P(k)$ is the azimuthally averaged power spectrum, k is the wave number ($2\pi \text{ km}^{-1}$), A is a constant, and Z_t is the depth to the top of magnetic sources.

The centroid depth of magnetic sources can also be calculated from the low-wave-number portion of the wave-number-scaled power spectrum as (Tanaka et al., 1999)

$$\ln(P(k)^{1/2}/k) = B - |k|Z_o, \quad (4)$$

where B is a constant and Z_o is the centroid depth of magnetic sources.

The depth to the bottom of the magnetic source (Z_b) can subsequently be obtained from the relation (Okubo et al., 1985)

$$Z_b = 2Z_o - Z_t. \quad (5)$$

Using the depth to the bottom of magnetic sources (Z_b), the geothermal gradient (dT/dz) can be estimated as (Tanaka et al., 1999; Ross et al., 2006)

$$\left(\frac{dT}{dz}\right) = \left(\frac{\theta_c}{Z_b}\right), \quad (6)$$

where θ_c is the Curie temperature.

Next, using Z_b and dT/dz , the heat flow (q_z) can similarly be estimated as (Okubo et al., 1985)

$$q_z = -\sigma \left(\frac{\theta_c}{Z_b}\right) = -\sigma \left(\frac{dT}{dz}\right), \quad (7)$$

where σ is thermal conductivity. Thermal conductivity of $2.5 \text{ W m}^{-1} \text{ }^\circ\text{C}^{-1}$ as the average for igneous rocks and a Curie temperature of $580 \text{ }^\circ\text{C}$ (Stacey, 1977; Trifonova et al., 2009) are used as standard in this work.

4 Data acquisition and analysis

The regional airborne magnetic surveys over the entire Bida Basin and adjoining areas were carried out using 3 Scintrex

Figure 5. CPD map of the study area.

Figure 6. West African Rift System (WARS) and Central African Rift System (CARS) (after Heine et al., 2013). Red circle indicates position of the Bida Basin with extensions of inferred major fracture zones – St Paul, Romanche, and Chain drawn with broken blue lines. These extensions had been suggested by Ajakaiye et al. (1991).

CS-3 cesium vapour magnetometers with a data recording interval of 0.1 s by means of fixed-wing aircrafts. The aircrafts were flown at mean terrain clearance of 80 m with 500 m line spacing and nominal tie-line spacing of 2 km. The flight line and tie-line trends were 135 and 45° respectively. The resulting magnetic data were published as digital half-degree HRAM intensity maps by the NGSA.

Forty-three new HRAM maps (sheet number 117–122, 138–143, 159–166, 180–187, 203–208, and 224–229) on a scale of 1 : 100 000, covering a total area of 130 075 km², were used in this work. All data, covering the entire Bida Basin and adjoining areas, were procured from NGSA as a composite residual total magnetic field intensity (RTMI)

map (Fig. 2). Regional correction, which was based on the International Geomagnetic Reference Field (IGRF – 11) derived to spherical harmonic degree 13, was carried out by the NGSA prior to the publication of the map.

The composite residual map was then divided into 28 overlapping square blocks (Fig. 3), for the purpose of spectral analysis. Each block covers a square area of $200 \text{ km} \times 200 \text{ km}$. Since magnetic source bodies having bases deeper than $\text{window length}/2\pi$ may not be properly resolved by the spectral method (Shuey et al., 1977), a window length of 200 km is found to be suitable in this study.

Figure 7. Geothermal gradient map of the study area.

Figure 8. Heat flow map of the study area.

5 Results and discussions

Azimuthally averaged power and wave-number-scaled power spectra for each of the 28 overlapping blocks were calculated and used to estimate the DBMS, which of course serve as a proxy for CPDs. As typical examples, blocks 5 and 20 are shown in Fig. 4. The right-hand side of the figure shows the slope of the high-wave-number portion of the spectra, which leads to the estimation of the depth to the top of magnetic sources (Z_t), while the left-hand side shows the slope of the lower-wave-number part of the wave-number-scaled spectra, which leads to the estimation of centroid depth (Z_o).

The estimated results are shown in Table 1. The table shows that the estimated CPD varies from 15.57 to 29.62 km with an average of 21.65 km. The CPD isotherm map for the Bida Basin is consequently shown in Fig. 5. CPD varies

greatly with different geological settings (Tanaka et al., 1999; Salk et al., 2005). Tanaka et al. (1999), after a compilation of CPD results from several researchers across the globe, inferred that volcanic, tectonic, and associated geodynamic environments have CPD shallower than 10 km, while CPDs ranging between 15 and 25 km are as a result of island arcs and ridges, and deeper than 25 km in plateaus and trenches. Figure 5 also shows that the CPD values in the Bida Basin trend mostly in the NE–SW direction, with the shallowest portion (less than 15 km) in the north-western part of the basin; the CPD extends and deepens both north-eastward and south-eastward in the basin, with its deepest depth (about 32 km) in the north-eastern parts.

Previous studies have shown that two major regional fault lines (namely St Paul and Romanche) are likely to have traversed the basin; these are believed to be extensions of the

Table 1. Estimated Curie-point depths and succeeding geothermal parameters.

Blocks	Long (° E) ^a	Lat (° N) ^a	Depth to the top Z_t (km)	Centroid depth Z_o (km)	Depth to bottom Z_b (km) ^b	Geothermal gradient (°C km ⁻¹)	Heat flow (mW m ⁻²)
1	5.00	10.50	1.67 ± 0.01	11.90 ± 0.05	22.13 ± 0.100	26.21 ± 0.119	65.52 ± 0.119
2	5.50	10.50	0.84 ± 0.03	12.80 ± 0.01	24.76 ± 0.036	23.43 ± 0.034	58.57 ± 0.034
3	6.00	10.50	1.22 ± 0.11	10.40 ± 0.13	19.58 ± 0.282	29.62 ± 0.427	74.06 ± 0.427
4	6.50	10.50	0.78 ± 0.05	15.20 ± 0.02	29.62 ± 0.064	19.58 ± 0.042	48.95 ± 0.042
5	4.50	10.00	1.76 ± 0.08	9.81 ± 0.11	17.86 ± 0.234	32.47 ± 0.426	81.19 ± 0.426
6	5.00	10.00	1.34 ± 0.01	11.04 ± 0.05	20.74 ± 0.100	27.96 ± 0.135	69.91 ± 0.135
7	5.50	10.00	0.93 ± 0.02	9.65 ± 0.06	18.37 ± 0.122	31.57 ± 0.209	78.93 ± 0.209
8	6.00	10.00	1.49 ± 0.12	9.38 ± 0.07	17.27 ± 0.184	33.58 ± 0.359	83.96 ± 0.359
9	6.50	10.00	0.86 ± 0.11	10.10 ± 0.02	19.34 ± 0.117	29.99 ± 0.182	74.97 ± 0.182
10	4.50	9.50	0.90 ± 0.03	9.74 ± 0.12	18.58 ± 0.242	31.22 ± 0.406	78.04 ± 0.406
11	5.00	9.50	1.00 ± 0.01	8.88 ± 0.07	16.76 ± 0.140	34.61 ± 0.290	86.52 ± 0.290
12	5.50	9.50	1.47 ± 0.04	8.52 ± 0.03	15.57 ± 0.072	37.25 ± 0.173	93.12 ± 0.173
13	6.00	9.50	1.43 ± 0.13	9.91 ± 0.05	18.39 ± 0.164	31.54 ± 0.281	78.85 ± 0.281
14	6.50	9.50	0.77 ± 0.05	10.20 ± 0.02	19.63 ± 0.064	29.55 ± 0.096	73.87 ± 0.096
15	7.00	9.50	0.95 ± 0.03	11.98 ± 0.10	23.01 ± 0.202	25.21 ± 0.222	63.02 ± 0.222
16	7.50	9.50	1.45 ± 0.05	13.10 ± 0.13	24.75 ± 0.265	23.43 ± 0.251	58.56 ± 0.251
17	5.50	9.00	1.40 ± 0.01	15.50 ± 0.09	29.60 ± 0.180	19.59 ± 0.119	48.99 ± 0.119
18	6.00	9.00	0.98 ± 0.02	14.60 ± 0.04	28.22 ± 0.082	20.55 ± 0.060	51.38 ± 0.060
19	6.50	9.00	0.89 ± 0.08	13.40 ± 0.07	25.91 ± 0.161	22.39 ± 0.139	55.96 ± 0.139
20	7.00	9.00	0.97 ± 0.05	11.89 ± 0.02	22.81 ± 0.064	25.42 ± 0.071	63.57 ± 0.071
21	7.50	9.00	1.44 ± 0.11	10.32 ± 0.04	19.20 ± 0.136	30.21 ± 0.214	75.52 ± 0.214
22	5.50	8.50	1.46 ± 0.01	12.00 ± 0.11	22.54 ± 0.220	25.73 ± 0.251	64.33 ± 0.251
23	6.00	8.50	1.66 ± 0.03	11.10 ± 0.03	20.54 ± 0.067	28.24 ± 0.092	70.59 ± 0.092
24	6.50	8.50	0.97 ± 0.05	10.50 ± 0.10	20.03 ± 0.206	28.96 ± 0.298	72.39 ± 0.298
25	7.00	8.50	1.98 ± 0.01	11.87 ± 0.07	21.76 ± 0.140	26.65 ± 0.172	66.64 ± 0.172
26	7.50	8.50	2.00 ± 0.01	12.50 ± 0.09	23.00 ± 0.180	25.22 ± 0.198	63.04 ± 0.198
27	7.00	8.00	0.69 ± 0.04	12.6 ± 0.04	24.51 ± 0.089	23.66 ± 0.086	59.16 ± 0.086
28	7.50	8.00	0.72 ± 0.13	11.2 ± 0.05	21.68 ± 0.164	26.75 ± 0.202	66.88 ± 0.202
Average			1.22 ± 0.05	11.43 ± 0.06	21.65 ± 0.146	27.57 ± 0.198	68.80 ± 0.198

^a Centre of the blocks; ^b corresponds to Curie-point depth.

onshore lineaments in West Africa, which are part of the major weakness in the crust that predates the opening of the Atlantic Ocean, and were reactivated in the early stages of continental rifting (Fig. 6) (Ajakaiye et al., 1991; Buser, 1966). Thus, the relatively low CPD values over the central portion of the basin may be fairly consistent with probable positions of St Paul and Romanche palaeofracture zones.

Table 1 similarly discloses that the geothermal gradients in the basin vary between 19.58 and 37.25 °C km⁻¹ with an average of 27.25 °C km⁻¹, while the crustal heat flow varies between 48.41 and 93.12 mW m⁻² with an average of 68.80 mW m⁻². Contour maps for the geothermal gradient and heat flow are shown in Figs. 7 and 8 respectively. The geothermal gradient map also exhibits mostly NE–SW trending. The observed major trends are similar to the regional trending faults in the basin. The lowest values for the geothermal gradient were found in the south-western portion of the basin. The north-westward trend of gradient increase was found to result in a maximum value of 42 °C km⁻¹ in

the north-western part. The minimum heat flow value required for considerable generation of geothermal energy is approximately 60 mW m⁻², whereas values ranging from 80 to 100 mW m⁻² and above indicate anomalous geothermal conditions (Jessop et al., 1976). Crustal heat flow in the basin also exhibits NE–SW trending, while the derived amounts increase from the central portion towards the north-west, with maximum values above 90 mW m⁻² observed in the north-central portion. This portion signifies an anomalous crustal thermal state and, therefore, is recommended for further investigations.

6 Conclusion

The newly acquired high-resolution aeromagnetic anomaly data over the Bida Basin, north-central Nigeria, have been analysed to estimate the Curie-point depths, geothermal gradients, and near-surface crustal heat flow. The result shows that the CPD varies between 15.57 and 29.62 km

with an average of 21.65 km, the geothermal gradient varies between 19.58 and 37.25 °C km⁻¹ with an average of 27.25 °C km⁻¹, and the crustal heat flow varies between 48.41 and 93.12 mW m⁻² with an average of 68.80 mW m⁻².

Regions are observed in the basin with shallow Curie-point depths (below 15 km) and corresponding high heat flows (above 80 mW m⁻²), thus suggesting anomalous geothermal conditions (Jessop et al., 1976). Hence, further detailed studies are recommended in such regions. Finally, oftentimes, direct crustal temperature measurements may not be too feasible for regional studies; hence, the derived geothermal gradients suffice for the entire basin. Moreover, geodynamic processes are mainly controlled by the thermal structure of the Earth's crust; therefore this study is anticipated to contribute significantly to the quantitative appraisal of the geoprocesses, rheology, and understanding of the heat flux variations in the Bida Basin in north-central Nigeria.

Data availability. The data used are not publicly accessible. However, the data could be obtained from the Abuja office of the Nigerian Geological Survey Agency, citing High Resolution Aeromagnetic sheet numbers 117–122, 138–143, 159–166, 180–187, 203–208, and 224–229.

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Competing interests. The authors declare that they have no conflict of interest.

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