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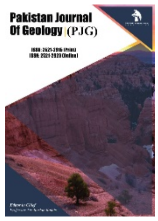

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RESEARCH ARTICLE

ASSESSMENT OF AQUIFER POTENTIAL OF THE VEA CATCHMENT IN GHANA USING PUMPING TEST TECHNIQUE

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ABSTRACT

Groundwater is an important source of drinking water for many people in sub-Saharan Africa due to its generally suitable quality. The Vea catchment of north-east Ghana is well known for its high reliance on groundwater for domestic use. The major livelihood of the indigenes is agriculture, which is severely affected by the seasonal long dry spell during harmattan. Therefore, the purpose of this study is to compute the aquifer characteristics to aid assessment of the storage potential for other uses such as irrigation in the catchment using the pumping test method. Pumping and recovery test data on 77 boreholes are used to determine the aquifer properties. The results indicated that transmissivity (T) and specific capacity (Sc) were within the ranges of 0.42–60.8 m^2/d and 2–70 m^3/d/m, respectively. The T and Sc are linearly related with a coefficient of determination, R^2, of 0.85. The area has a specific yield of about 0.06% and a storage coefficient in the order of 10^{-7}–10^{-3}. The shallow aquifer system (below 50 m deep) and the fractured aquifer system have mean safe yields of 138 m^3/d and 345 m^3/d, respectively. Thus, the groundwater potential can support other uses such as small-scale irrigation apart from drinking. The results of the study could be used as a guide for managing groundwater resources in arid and semi-arid areas. Through careful planning of groundwater withdrawals for irrigation and other uses for the socioeconomic development of the people, this will help improve water security resilience in the catchment.

KEYWORDS

Aquifer potential, Pumping test, Vea Catchment, Ghana.

1. INTRODUCTION

In recent times, the rise in surface water quality deterioration in sub-Saharan Africa due to impacts of land use practices like artisanal mining, agriculture, etc. has made groundwater the alternative water source for many people due to its generally suitable quality. In view of this, groundwater needs to be sustainably managed to meet the various water demands for domestic, livestock, industrial and irrigation use. This calls for aquifer potential assessment and management planning scheme to support the demand and supply of groundwater. According to a study, groundwater potential is the capacity of a terrain to yield enough groundwater based on series of indicators (Díaz-Alcaide et al., 2019). The indicators may be direct like aquifer properties (e.g. yield, transmissivity, storage coefficient, specific capacity, etc.) or indirect like degree of weathering, slope, lineament, geology, rainfall, soil type, etc. These indicators provide several approaches for groundwater potential assessment including: (1) groundwater potential mapping using remote sensing and the Geographic Information System (GIS), (2) geophysical techniques, (3) modelling techniques and (4) field evaluation using pumping test techniques.

Some researchers in their use of the remote sensing techniques, identified the basic factors influencing groundwater potential to include geology, lineaments, landforms, soil, land use/land cover, rainfall, drainage density, and slope (Díaz-Alcaide et al., 2019). They found the approach very useful for regional studies but less effective in local settings (Manap et al., 2014). Likewise, the accuracy of the groundwater potential map developed for the remote sensing technique depends on the number of thematic layers of factors deployed in the mapping. For instance, a group researchers reported an accuracy of 84.78% in the case of using eight factors (i.e.: elevation, slope, curvature, river, lineament, geology, soil, and land use) whereas obtained 73.66% with five factors (i.e. rainfall, lithology, drainage density, lineament density, and slope) (Manap et al., 2014; Rahmati et al., 2015).

The GIS application tool has been used for mapping groundwater potential zones with input data from the catastrophe (or DRASTIC) approach, machine learning approach and analytic hierarchy approach (Ahmed et al., 2015; Kalantar et al., 2019; Rahmati et al., 2015). These techniques assign indices to selected factors and rank them based on their significance to groundwater potential assessment. According to the accuracy of these techniques largely depends on the selection of appropriate factors influencing groundwater storage and availability of reliable data on the selected factors (Ahmed et al., 2015). Even though the GIS approach is cost effective, robust, and applied by most researchers in recent times, there is some degree of subjectivity since the normalized indices are mostly based on expert judgement (Ahmed et al., 2015). Also, the output of the GIS approach only provides qualitative description of the groundwater potential and lack evidence-based results, which may undermine the objectivity to reduce the quality assurance of the findings.

The geophysical survey technique is non-invasive, but it operates from the ground surface without direct access to the groundwater system. It explores for the targeted physical parameter contrast (e.g. resistivity, conductivity, susceptibility, etc.), which is interpreted in relation to the subsurface anomaly for inference. According to assessing aquifer potential with the geophysical parameters is not absolute because these parameters

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directly depend on the subsurface structures and geological formation, which vary locally and is characterized by possible ambiguities in the interpretation of the target anomaly (Hasan et al., 2018). The field evaluation of groundwater potential using pumping test approach may be labor-intensive, expensive, and time-consuming as expressed by most authors (Manap et al., 2014; Ahmed et al., 2015; Rahmati et al., 2015; Hasan et al., 2018; Kalantar et al., 2019; Díaz-Alcaide et al., 2019).

However, this approach provides objective and precise information about the groundwater characteristic and behavior of aquifer to better inform the decision makers and groundwater managers to better manage the groundwater system. A group researchers affirmed that groundwater mapping technique cannot provide optimal groundwater assessment unless it complements field studies such as pumping test method (Díaz-Alcaide et al., 2019). Thus, pumping test method has always been a reliable means to estimate aquifer properties and determine groundwater flow direction, especially, if it is conducted over long duration. The aim of this study is, therefore, to evaluate the aquifer storage potential, groundwater flow characteristics and establish a relationship among the aquifer characteristics using pumping test method in the Vea catchment of northern Ghana. The indigenous people in the catchment largely depend on small dams and streambed dugouts as their main sources of water for livestock and small-scale irrigation of vegetable farms during prolonged dry seasons. Thus, the study is necessary to ascertain the groundwater potential of the catchment for sustainable utilization of the groundwater resource. This will aid improve water security resilience in the catchment through effective planning of groundwater withdrawals for irrigation and other uses for socio-economic development of the people.

2. Study Area

Vea catchment, located within longitudes 0°44′0″ W – 1°00′00″ W and latitudes 10°43′00″ N – 11°01′00″ N, is a transboundary basin in Ghana with less than 8% in Burkina-Faso (*Fig. 1*). The catchment is within the Sahelian climatic region (semi-arid) with low and erratic unimodal rainfall of about 970 mm/yr with 87% accumulated from June to October and about 2540 mm/yr potential evapotranspiration at a maximum temperature of about 42°C in March and April (Barry et al., 2005). It experiences prolonged dry season between November and May (Mul et al., 2015). The population of the area was approximately 1,046,545 as of 2010 with a growth rate of 1.2% in Ghana (Ghana Statistical Service [GSS], 2012) and 300 as of 2015 statistics in Burkina-Faso (Beal et al., 2015). The major occupation of the indigenous people is crop farming and animal rearing. The study area is a sub-catchment covering an area of 315 km^2, which is less than 1% of the White Volta Basin. The main river in the area is *Yarigatanga River* and its tributaries, which have been impounded to construct a dam (capacity = 17.3 Mm^3) in the *Vea* community (Koffi et al., 2017). Land use in the basin is mainly for cereal cropping, grazing and settlement (Dickson and Benneh, 1988; Adongo et al., 2014; Mul et al., 2015).

The basement rock of the catchment is mainly granite and granodiorite with basalt and sandstone as minor rocks (Figure 1) trending in the NE-SW direction (Kesse, 1985; Dapaah-Siakwan and Gyau-Boakye, 2000). Even though the granite is massive and intrusive with interlocking texture by nature, its weathered subsurface and fractures provides considerable groundwater storage potential. The basalt at the fringes of the northeastern and southeastern boundaries are associated with vesicles, which have great water storage potential and if interconnected with fractures. It occurs as volcanoclastic (or lava) flow overlying buried formation of high groundwater potential. The sandstone conduit running in the SW-NE direction at the southern part of the catchment serves as a water receptacle to support groundwater development and its potential is greatly enhanced by the presence of conglomerates, which tends to have larger grain sizes with highly permeable pores. The numerous joints in the granitoid and the existence of dikes and outcrops provides a favourable environment for water percolation into the subsurface to support groundwater development (Darko and Krasny, 2003; Martin and van de Giesen, 2005; Barry et al., 2010).

The groundwater system within the K-feldspar-rich granitoid located in the middle belt of the catchment is associated with topaz compounds causing leachate of excess fluoride ions usually above the World Health Organization (WHO) recommended guideline value of 1.5 mg/L into the groundwater system, which makes it unsuitable for drinking unless treated (Kesavulu, 1993; Edmunds and Smedley, 2005). This condition is evidenced in the prevalent cases of fluorosis in children living in the area after drinking water over a long period. The weathered portion of the subsurface consists of laterite, clay, sand, and gravel soil deposit, which vary locally in thickness. The large clayey deposit in the form of hardpan at various locations tend to confine the aquifer at shallow depth.

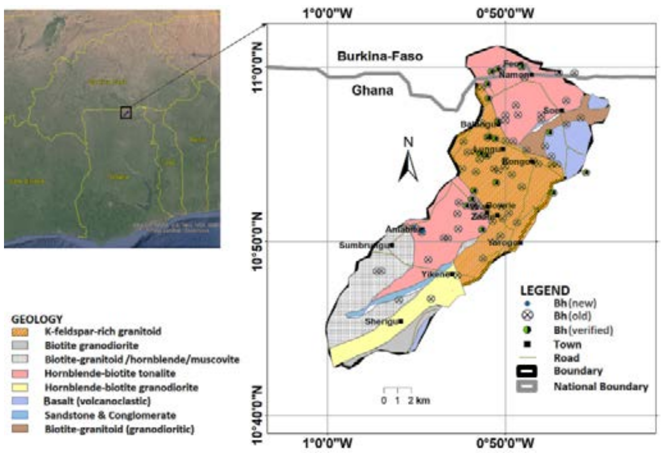

Figure 1: Location and geology map of the study area showing borehole locations

3. Methodology

Out of the seventy-seven (77) boreholes used for the study, ten (10) boreholes were newly drilled while the remaining sixty-seven (67) were existing boreholes with comprehensive information obtained from Community Water and Sanitation Agency (CWSA). Step-drawdown test was carried out on the newly drilled boreholes to estimate their optimum discharge rates for the constant discharge test as proposed by Summa, among others (Summa, 2010). Also, for quality control purposes, 19 boreholes (Table 1) were selected from the existing boreholes and pumping tests conducted on them for verification. The static water levels and the depth of the boreholes were measured before the test. Where siltation was suspected to be high, the pump was installed at shallow depth to protect the pump impeller from damage.

During the constant-discharge pumping test on the wells, the static water levels and the depth of the boreholes were measured and recorded prior to setting-up the test equipment. The test equipment (two 100 m uPVC hose coupling, dip-meters, 9 m and 30 m transducers, 5KVA Diesel Generator Set and high capacity Grundfos SP14A-17 submersible pump) was setup with lead hose of over 100 m to discharge the water to minimize return flow. Six boreholes were earmarked for observation during the pumping and recovery period. The pump was installed within 10 m from the bottom of the well except boreholes with significantly high siltation cases, where enough room was allowed to protect the pump and normalize its operation. During the test, water level drops were recorded at specific time intervals until the end of the test. At pump shut-down, residual drawdowns were measured during the recovery period until 85-96% recovery was attained for each borehole. The drawdown was computed, and the data was analysed to obtain the specific capacity, transmissivity, hydraulic conductivity, and storage coefficient of the aquifer system of the catchment. The recovery rate (R) after an hour of pumping was computed using Eqn. (1) to aid evaluate the aquifer's sustainability to withstand long duration of pumping.

$$R = \left(\frac{\text{Water recovery within 60 min}}{\text{Maximum drawdown}}\right) x100\% \quad (1)$$

The transmissivity was determined using the Theis modified equation given by the Eqn. (2) (Cooper and Jacob, 1946; Mawlood and Mustafa, 2016). The transmissivity was also determined using the Theis recovery method from the residual-drawdown data during recovery period for quality control (Theis, 1935; Willmann et al., 2007). Storage Coefficient (S) is a dimensionless physical quantity indicating the volume of water an aquifer can potentially release per unit aquifer storage area per unit decline in hydraulic head (Freeze and Cherry, 1979; Wang and Anderson, 1982). It is the vertically integrated specific storage value assuming homogeneous and compressible aquifer and water for a confined aquifer system and it is also approximately equal to specific yield for unconfined aquifer system. The coefficient can be calculated from the modified non-equilibrium approach provided observation wells are available in the pumping test. It is worth noted that the residual-drawdown plot cannot be used to determine the storage coefficient, even though it is only applicable in the determination of transmissivity (Driscoll, 1986). In this study, the storage coefficient is computed using Eqn. (3), which is based on the modified non-equilibrium equation (Cooper and Jacob, 1946; Mawlood and Mustafa, 2016). In the computation of the storage coefficient (S), the time (t_o) at zero drawdown was obtained when the tangent to the curve was extended to intercept the time-axis (see Appendix).

Table 1: Information on Verified Boreholes in The Study Area

Bh	Town	Long	Lat	Elev (m)	Depth (m)	SWL (m)	WS (m)	b (m)	PS (m)	Q (m^3/d)
1	Gowrie	-0.8552	10.8447	184	56.00	7.64	32.00	30.36	50.0	86.40
2	Feo	-0.8179	11.0003	241	50.00	9.23	31.00	27.77	40.0	93.60
3	Vea	-0.8648	10.8737	201	51.53	7.81	35.00	33.19	50.0	86.40
4	Vea	-0.8699	10.8674	196	32.93	3.29	38.00	40.71	25.0	144.00
5	Vea	-0.8628	10.8817	201	31.80	6.76	35.00	40.24	28.0	93.60
6	Adaboya	-0.7919	10.9380	217	54.88	4.27	30.00	46.73	50.0	115.20
7	Adaboya	-0.7886	10.8798	215	46.25	0.67	14.00	43.33	40.0	132.48
8	Adaboya	-0.7518	10.8984	229	51.53	3.36	31.00	33.64	20.0	96.48
9	Balungu	-0.8403	10.9483	238	49.00	3.27	20.00	25.73	43.0	172.80
10	Balungu	-0.8485	10.9338	230	45.87	8.62	35.00	38.38	40.0	172.80
11	Balungu	-0.8426	10.9319	231	23.50	5.29	20.00	48.71	20.0	86.40
12	Lungu	-0.8428	10.8934	201	44.37	6.75	29.35	22.60	24.0	120.00
13	Lungu	-0.8419	10.8897	199	41.72	9.70	17.00	19.30	36.0	93.60
14	Lungu	-0.8514	10.9159	211	41.65	2.06	32.00	41.94	38.4	86.40
15	Lungu	-0.8568	10.9176	218	55.60	8.91	42.00	48.09	40.0	72.00
16	Namoo	-0.8493	10.9704	238	34.50	5.68	30.00	39.32	24.0	100.80
17	Namoo	-0.8501	10.9841	238	44.85	5.39	30.00	39.61	40.0	172.80
18	Namoo	-0.8463	10.9959	238	41.65	8.01	20.00	20.99	32.0	172.80
19	Namoo	-0.8407	10.9982	239	55.60	5.03	44.00	44.97	50.0	103.68

(SWL = static water level; WS = water strike; b = saturated thickness; PS = pump setting; Q = pumping rate)

$$T = \frac{0.183Q}{\Delta s} \quad (2)$$

$$\boldsymbol{S} = \frac{2.25\boldsymbol{T}t_o}{r^2} \quad (3)$$

Where t_o is the time representing the intercept at zero drawdown, r is the distance between pumping and observation wells, Q is the discharge rate and Δs is the change in drawdown per log-cycle. The hydraulic conductivity (K) was then calculated using results of the transmissivity with the aid of aquifer thickness encountered at each borehole location (see Appendix). Also, specific capacity was computed for each borehole using Eqn. (4) and a relationship between the specific capacity and transmissivity was investigated through regression analysis to support future estimation of T values from Sc.

$$Sc = \frac{Q}{s_{max}} \quad (4)$$

Where **Q** is discharge and $\mathbf{s_{max}}$ is the maximum drawdown.

The specific yield (Sy) is expressed as the ratio of the volume of water drained (or abstracted) to the volume of the dewatered aquifer material, was computed using Eqn. (5). Assuming no recharge effect, negligible head loss and fully screen aquifer in the pumping well, used Darcy's principle to develop a relationship to computing the volume (V) of the dewatered aquifer material given as Eqn. (6), which was used in this study to compute Sy (Ramsahoye and Lang, 1993). The percentage of Sy in the storage coefficient (S) can be evaluated from Eqn. (7) and (8) to quantify the level of confinement of the aquifer system.

$$Sy = \frac{Qt}{V}x100\% \quad (5)$$

$$V = \frac{Qr^2 exp\left(\frac{4\pi Ts}{Q}\right)}{4T} \quad (6)$$

Where all the symbols have the same meanings defined in the preceding equations.

Safe yield of a productive borehole is the discharge rate to maintain dynamic equilibrium at the major water strike of the borehole (Helweg et al., 1991; van Tonder et al., 2001). Thus, equating the specific capacity at the dynamic water level to that at the minor (shallow) and/or the major (fractured) water strike provides the estimate for the safe yield (Q_s), which was estimated using Eqn. (9).

$$Q_s = Q_{obs}\left(\frac{s_s}{s_{obs}}\right) \quad (7)$$

Where Q_s and Q_{obs} are the discharges with drawdown to the water strike (s_s) and the observed maximum drawdown (s_{obs}) respectively. Groundwater flow occurs from higher hydraulic head to lower head in a direction perpendicular to the line joining points of equal water table in the subsurface. Thus, the static water level elevations from all the boreholes in the study were contoured to obtain the potential flow direction of the groundwater and aid delineate the recharge and discharge zones within the catchment.

4. Results and Discussion

4.1 Analysis of Specific Capacity, Transmissivity and Hydraulic Conductivity

The catchment's specific capacity ranges largely between 1.0 and 70.0 m^3/d/m (Fig. 2**a**), which is comparable to Agyekum and Kankam's findings of 1.04 - 56.7 m^3/d/m in the granitoid of north-western Ghana (Agyekum and Kankam's, 2011). The exceptionally high specific capacity occurs in the north-western corridor and the central portion of the catchment. The specific capacity values less than 1.0 m^3/d/m covers relatively small and isolated portions of the catchment. The transmissivity values of the study area varied between 0.42 and 60.8 m^2/d and may be classified as very low-to-intermediate according (Krasny, 1993). This result is close to the 0.37-44.5 m^2/d obtained in the north-western Ghana (Agyekum and Kankam, 2011). Specifically, per Krasny, 68% of the study area has low transmissivity range while 14% and 18% of the area have intermediate and very low transmissivity respectively (Krasny, 1993).

The low-lying mid-section of the catchment (around Vea and Gowrie) have the lowest transmissivity values (< 1 m^2/d) as indicated in Figure 2b, even though water storage potential is high. Notably, aquifer transmissivity of over 124 m^2/d can supply water in commercial quantity and below 12.4 m^2/d is only suitable for domestic application (Driscoll, 1986). According to transmissivity values greater than 100 m^2/d are considered good in hard rock terrains (Kumar et al., 2016). The transmissivity obtained marks the true values of the formation as it was determined from the pumping wells rather than the observation wells. Gomo, who used aquifer models with MODFLOW, verified that transmissivity varies exponentially with distance from the pumping well (Gomo, 2019).

The hydraulic conductivity (K) of the catchment ranges from 2.4 x 10^{-7} to 5.1 x 10^{-4} m/s, and the average value is in the order of 10^{-5} m/s (~ 3.7 x 10^{-5} m/s). It is low at the downstream and the northern section of the area with average value of 6.43 x 10^{-7} m/s while high at the mid-stream around the dam in the mid portion of the catchment at an average value of 2.15 x 10^{-4} m/s. Figure 2c indicates rapid hydraulic conductivity at the mid-section of the catchment around the location of the Vea dam whereas the degree of conductivity reduces to moderate outwards to the northern and southern portions of the catchment using the United State Development of Agriculture (USDA) criteria (USDA, 2004).

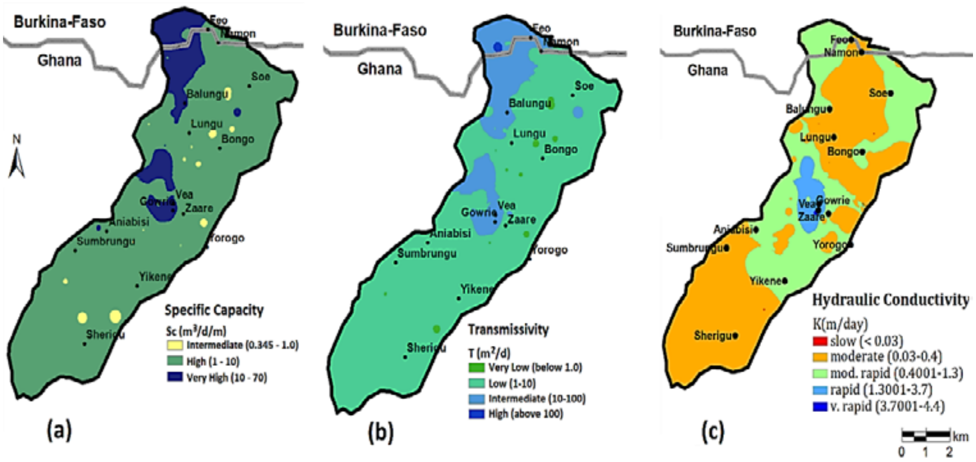

Figure 2: Spatial distribution of (a) specific capacity, (b) transmissivity and (c) hydraulic conductivity in the Vea catchment classified based on IDW interpolation technique

4.2 Relationship between Transmissivity (T) and Specific Capacity (Sc)

Specific Capacity (Sc) directly depends on measured field values and any error associated with its computation is limited to instrument used or human error (Mace, 2001). Thus, Sc is more often easy to compute as compared to transmissivity, which depends on change in drawdown per log cycle determined graphically based on Cooper-Jacob straight-line log-normal analysis (Razack and Huntley, 1991). Figure 3a shows a scatter plot of Sc against T values for the study area, which produced linear relation between the two parameters expressed in Eqn. (8) with coefficient of determination, R^2, of 0.85.

$$\mathbf{T} = 1.264\mathbf{Sc} \quad (8)$$

Assessment of the derived relation conducted with 19 borehole datasets in the catchment is shown in Figure 3b and indicated a reliable model with R^2 of 0.81. Also, the relationship performed effectively well when it was compared with existing relationships by given respectively as Eqns. (9) and (10) (Mace, 1997; Huntley et al., 1992).

$$\mathbf{T} = 0.76\mathbf{Sc}^{1.08} \quad (9)$$

$$\mathbf{T} = 0.29\mathbf{Sc}^{1.18} \quad (10)$$

A boxplot also revealed statistically good performing relationship as compared with the models against the observed transmissivity data (Mace, 1997; Huntley et al., 1992). Figure 3c shows descriptive statistical distribution of the numerical data for the various models, which has similar median mark located close to the twenty-fifth percentile, which also indicates positive skewness of about 2.738. About 50% of the transmissivity values as computed by the various models produced quite low T values close to the minimum value. The standard error of the distribution from the study is ±0.176. Therefore, the mean and median of the distribution are 14.974 ±0.176 m²/d and 9.488 m²/d respectively. The interquartile range of the distribution is quite broad compared to that of distributions, which reduced the number of supposed outliers in the dataset as compared to the various models (Figure 3c) (Mace, 1997; Huntley et al., 1992).

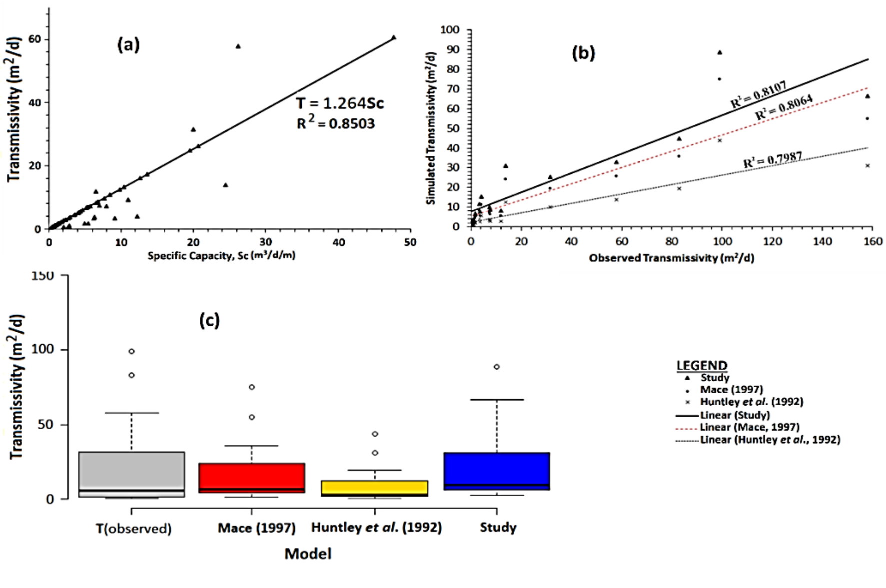

Figure 3: (**a**) Derived T and Sc relation for the study area, (**b**) calculated values against T values obtained using the T-Sc model for this study in comparison and (**c**) boxplot to compare the central tendencies of transmissivity values generated by the various models (Mace, 1997; Huntley et al., 1992)

4.3 Evaluation of Storage Coefficient, Specific Yield and Safe Yield

The spatial distribution of the storage coefficients determined in the study are shown in *Fig. 4***b**. Generally, it ranges from 7.85×10^{-7} to 1.32×10^{-3} with a mean value of 3.82×10^{-4}. According to the average storage coefficient value of 3.82×10^{-4} is associated with clay (hardpan), which is dominant in the subsurface weathered formation of the catchment (Verruijt, 2018). The other soil types are sandy (dense) of the order of 10^{-5}, gravel (dense) of the order of 10^{-6} and fractured rock of the order of 10^{-7} as classified (Verruijt, 2018). The storage coefficient value helps to estimate the total volume of water abstracted from or stored in an aquifer per a given change in water level. Thus, for the study area, the volume of water (V) required to cause a unit rise in head is equal to the storage coefficient (3.82×10^{-4}) multiplied by the area of the catchment (3.15×10^{8} m²) multiplied by the change in head (1 m), which is 0.12 Mm³. Generally, for any change in head (**Δh**) of water table, the volume of water in storage is given as Eqn. (11) for the Vea catchment. According to Community Water and Sanitation Agency (CWSA) criteria, the minimum water consumption per person per day is 20 L and the projected population of the inhabitants in the catchment is currently about 1,207,929 people, in accordance with a population growth rate of 1.2%. Thus, the minimum total domestic water demand per day is estimated at 0.0242 Mm³, which is only 20% of the unit change in water storage (0.12 Mm^3). Hence, the 80% storage provides enormous potential for other uses, such as irrigation, among others. An increase in water storage head (**h**) improves the water availability and groundwater potential for sustainable use.

$$V = 1.2 \times 10^5 \Delta h \quad (11)$$

The computed specific yield of the weathered subsurface aquifer system ranges from a low of 1.345E-5% within the clayey soil to a high of 2.522E-2% within the gravel and sandy soil with an average specific yield of 6.0E-2% (Table *2*). Its spatial distribution in *Fig. 4***a** shows that Sy is mainly below average within the catchment. The Sy values compare well with the findings of Saha and Agrawal (2006) of 1.9E-3 – 1.73E-2%.

Table 2: Computed Specific Yield and Storage Coefficient of The Aquifer System in The Vea Catchment

OW	s(m)	r(m)	t(hour)	T(m^2/d)	Q(m^3/d)	V(km^3)	Sy (%)	S
2	2.576	5.6	72.483	61.488	120.96	0.21653	1.687E-04	1.838E-05
3	2.116	6.3	72.492	58.252	120.96	0.00752	4.861E-03	4.128E-04
4	2.334	5.5	72.492	55.339	120.96	0.01113	3.282E-03	3.430E-04
8	4.662	3.6	73.225	52.341	155.52	3.52738	1.345E-05	9.390E-07
9	3.541	5.3	73.225	38.827	155.52	0.00188	2.522E-02	1.620E-03
10	3.447	7.4	73.225	44.961	155.52	0.01302	3.644E-03	2.310E-04
MEAN					6.199E-03			

(OW = observation well; s = drawdown; t = time; Q = discharge; *V* = volume of dewatered aquifer material)

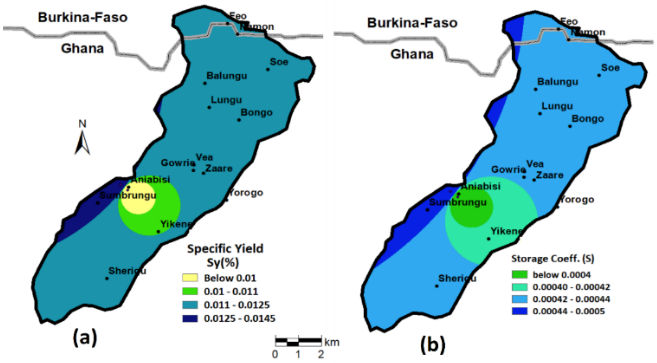

Figure 4: Spatial distribution of (**a**) specific yield and (**b**) storage coefficient of Vea catchment

In the study area, water strike occurs at two main locations defining the weathered (shallow) from 14 to 49 m below ground level (*bgl*) with an average of 28 m *bgl* and the fractured (deeper) aquifer systems, which ranges from 29 to 60 m *bgl* with an average of 42 m *bgl*. Based on the observed discharge ranging 17 - 353 m^3/d (mean = 104 m^3/d) and the corresponding drawdown range 3 – 42 m (mean = 20 m) measured during the pumping test, the safe yield of the shallow aquifer system (below 50 m deep well) in the study area, computed from Eqn. (9), ranges from 4 to 1517 m^3/d with an average of 138 m^3/d while that for the fractured aquifer system (above 50 m deep well) ranges from 15 to 4007 m^3/d with an average of 345 m^3/d. These average yields are classified as high yielding potential by for yields above 240 m^3/d (Manap et al., 2013). In the catchment, the distribution of safe yield is generally moderate within the shallow aquifer, but it is high-to-very high in the deep aquifer (Figure 5). The aquifer potential decreases from the western towards the eastern corridor of the catchment as depicted by Figure 5. A group researchers estimated the safe yield for 26 boreholes in the Kassena-Nankana District of the Upper East Region of Ghana, which shares the western boundary of the study area to range from 15 to 572 m^3/d with an average of 115 m^3/d, which compares well with the results of the study (Figure 5) (Anim-Gyampo et al., 2012).

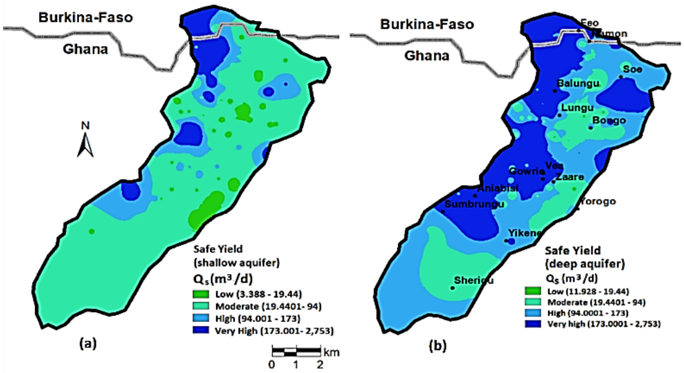

Figure 5: Spatial distribution of safe yield for (**a**) shallow and (**b**) deep aquifers in the Vea catchment

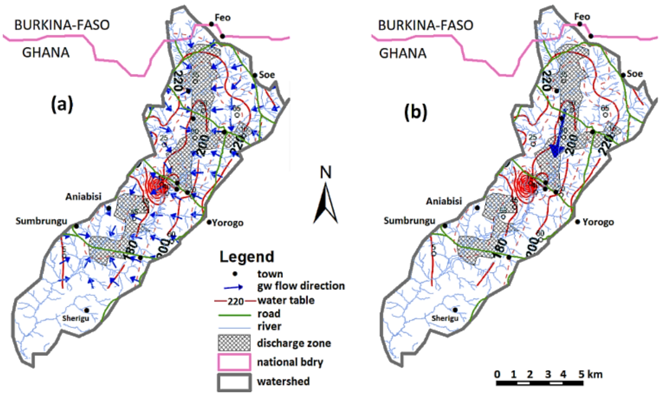

Figure 6: Piezometric map of Vea catchment showing groundwater flow direction at (**a**) local and (**b**) regional levels

4.4 Groundwater Flow Direction

Generally, the water table elevation of the study area varies from 165 to 384 m above sea level (*asl*) with an average elevation of 209 m *asl*. Figure 6 shows the local and regional groundwater flow directions. The local flow directions revealed the discharge zones mapped at the central portion of the catchment, which coincided with the major drainage channel of the main river (Figure 6a). The recharge zones, however, coincided with the groundwater divide at the catchment boundary with higher elevation compared with the discharge zones. The regional groundwater flow occurred in the south-western direction of the catchment (Figure 6b). Knowledge of the groundwater dynamics is a useful tool to support aquifer protection against possible contamination and groundwater governance.

5. Conclusions

The study has successfully evaluated the transmissivity (and hydraulic conductivity), specific capacity, storage coefficient, specific yield and safe yield from pumping test data to assess the groundwater potential of the Vea catchment. The results showed that the transmissivity of the catchment varies between 0.42 and 60.8 m^2/d with a mean of 14.974 ±0.176 m^2/d, and can be expressed as a linear function of specific capacity with coefficient of determination, R^2, of 0.85. The hydraulic conductivity of the downstream and the upstream sections of the catchment is averagely 6.43 x 10^{-7} m/s while that of the mid-stream is averagely 2.15 x 10^{-4} m/s. The storage coefficient varies in the order of 10^{-7} – 10^{-3} with an average of 3.82 x 10^{-4}, which mostly depicts a confined aquifer system in the catchment. The weathered aquifer system, representing only 12.7% of the entire aquifer system, has an average specific yield of about 0.06%. The safe yields vary between a mean of 138 m^3/d for the shallow aquifer system (below 50 m deep) and a mean of 345 m^3/d for the fractured aquifer system in the catchment.

Thus, the groundwater potential is classified as moderate for the shallow and high-to-very high for the deep aquifers of the catchment. The spatial distribution of the safe yield revealed that high groundwater potential is associated with the deep aquifer system, implying that the high groundwater potential in the catchment is mainly controlled by secondary porosity. Generally, the catchment has a reasonably good aquifer system with discharge zones coinciding with the main stream channel while recharge zones coincide with the groundwater divide. Thus, any contamination at the recharge zones can cause widespread pollution and affect all those depending on the groundwater system directly or indirectly; hence needs to be protected.

The study has provided detailed information on the characteristics of the aquifer system in the Vea catchment to aid understand the groundwater potential, which can be used as a guide to sustainably manage the groundwater resource. The knowledge generated by this study could also be applied in similar geological terrains in arid and semi-arid climatic areas. Future studies could focus on the groundwater quality appraisal of the catchment for comprehensive decision-making on the suitability of groundwater resource for drinking, irrigation and other domestic purposes.

Appendix

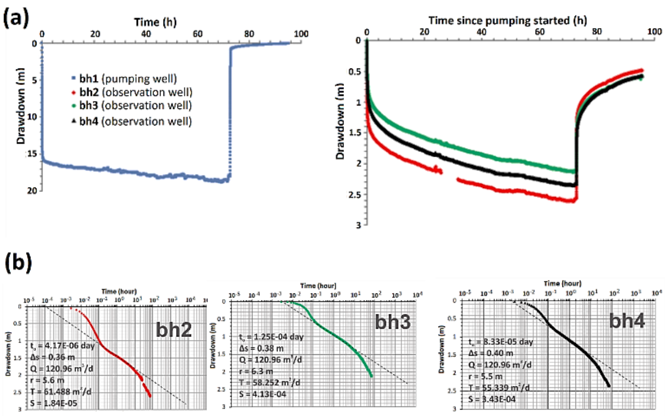

Figure A.1: Pumping test on bh1 with observation wells (bh2, bh3, and bh4) showing (**a**) time-drawdown plot and (**b**) graphical analysis for aquifer properties (Discontinuity of curve indicates missing data due to quality control check on logger).

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Data Availability Statement

Essential data incorporated in the article and any other data will be made available on request.

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