Mansoura Engineering Journal

Volume 34 | Issue 2 Article 2

11-26-2020

# Estimation the Hydraulic Parameters of the Fractured Aquifer by Using Spring Hydrograph Analysis, Lange Bramke Basin, Germany.

Mohammed Gad
*Hydrology Division, Desert Research Center, Cairo, Egypt*, drmohamedgad@yahoo.com



## Recommended Citation

Gad, Mohammed (2020) "Estimation the Hydraulic Parameters of the Fractured Aquifer by Using Spring Hydrograph Analysis, Lange Bramke Basin, Germany.," *Mansoura Engineering Journal*: Vol. 34 : Iss. 2 , Article 2.
Available at: https://doi.org/10.21608/bfemu.2020.125594

# Estimation the hydraulic parameters of the fractured aquifer by using spring hydrograph analysis, Lange Bramke basin, Germany

# تقدير المعاملات الهيدروليكيه للخزان المتشقق من تحليل هيدروجراف تصرف العين بحوض لانج برامكه بألمانيا الإتحادية

M. I. GAD
Hydrology Division, Desert Research Center, Cairo, Egypt
e-mail: drmohamedgad@yahoo.com

الخلاصة

إن من أهم خصائص الخزان الجوفي المتشقق هو التغاير الشديد بين المعاملات الهيدروليكية للنطاق الصخري المتشقق والغير متشقق. وفي هذا البحث تم تقدير المعاملات الهيدروليكية للنطاقين سالفا الذكر بخزان الحجر الرملي المتشقق والتابع للعصر الديفوني المبكر بحوض لانج برامكة بألمانيا الإتحادية بطريقة غير تقليدية. وتعتمد منهجية البحث على تطوير النموذج الإعتباري لرارابوغ لسريان المياه الجوفية بالخزانات المتشققة، واستنباط العلاقة بين منحنيات الإنحسار بهيدروجراف تصرف عين متدفقة بالخزان المتشقق ومعامل الإنتشار الهيدروليكي له ومعامل الناقلية. وقد تم الاعتماد على البيانات الحقلية لتصرف العين لفترة زمنية محددة خلال أعوام ٢٠٠١و٢٠٠٢و ٢٠٠٤ (١٢يونيو-٣١ أغسطس) في إنشاء هيدروجراف تصرف العين بالخزان المتشقق ، بجانب قياس مناسيب المياه الجوفية ببئرين ممثلين للنطاق المتشقق والغير متشقق لنفس الفترة عام 2001. كما تم معايرة نتائج تقدير المعاملات الهيدروليكية للخزان المتشقق بهذه الطريقة المبتكرة بقيم المعاملات الهيدروليكية الناتجة من تحليل بيانات أربع تجارب ضخ ونزح تم إجراؤها أثناء هذه الدراسة. وايضاً تمت المقارنة بين نتائج هذا البحث ونتائج تقدير المعاملات الهيدروليكية لنفس الخزان بطرق مختلفة لباحثين آخرين كطريقة الحقن بالصبغات والنظائر المشعة.

وأوضحت نتائج البحث أن معامل الإنتشار الهيدروليكي المقدر لسريان المياه الجوفية خلال النطاق الصخري المتشقق تراوح من ٠,٠٢٤ إلى ١,٢٩٨ م$^2$/ث، بينما تراوح من ٠,٠١٦ إلى ٠,٢١١ م$^2$/ث لسريان المياه الجوفية بالإنتشار خلال النطاق الصخري الغير متشقق. كما تراوح معامل الناقلية للنطاق الصخري المتشقق والمشتق من تحليل ميل منحنيات الإنحسار لهيدروجراف تصرف العين من ٠,٠٢٤ إلى ١,٢٩٨ م$^2$/ث، بينما تراوحت قيمته من $3.2x10^{-4}$ إلى $42.2x10^{-4}$ م$^2$/ث وذلك بالنطاق الصخري الغير متشقق. وهي نتائج تتوافق إلى حد معقول مع الحد الأدنى لنتائج تقدير المعاملات الهيدروليكية للخزان المتشقق من تحليل تجارب الضخ والنزح والتي تراوحت بين $0.17x10^{-4}$ و $4.8x10^{-4}$ م$^2$/ث وكذلك نتائج طرق الحقن والنظائر البينية المشعة بينما تتفاوت في الحد الأعلى.

وبناءً عليه يوصى باستخدام هذه الطريقة المبتكرة والغير مكلفة إقتصادياً- بعد معايرتها بالطرق الحقلية المختلفة- في تقدير المعاملات الهيدروليكية للخزانات المتشققة والتي لها نفس الظروف الهيدرولوجية للخزان المتشقق محل الدراسة.

## ABSTRACT

One characteristic of a fractured aquifer is the large contrast between the hydraulic parameters of fracture zone and that of the surrounding rock matrix. In this paper a trial to determine the aquifer constants of the Early Devonian sandstone fractured aquifer in Lange Bramke Basin, Germany was carried out applying Rorabaugh's conceptual model. This model was modified and used in estimating the hydraulic parameters of the conduit and diffuse flow systems from discharge hydrograph analysis. Daily discharge from a spring in the aquifer under consideration was continuously monitored for the period 12 June-31 Aug during 2001, 2002 and 2004. These records were used in recession hydrograph analysis. The identification different slopes in the recession hydrograph proceed to isolate each slope and to derive the ratio between transmissivity T and specific yield Sy (aquifer diffusivities T/Sy) from each slope. Two pumping and two slug tests were carried out during this study to check the estimated hydraulic parameters of the diffuse flow system. The results showed that the

Accepted June 15, 2009.

T/Sy ratios derived from the recession hydrograph analysis ranged from 0.024 to 1.298 $m^2$/sec with an average value of 0.438 $m^2$/sec for conduit flow system, from 0.164 to 0.345 $m^2$/sec with an average value of 0.222 $m^2$/sec for intermediate-flow system and from 0.016 to 0.211 $m^2$/sec with average value of 0.107 $m^2$/sec for diffuse flow system. Assuming Sy value for pure conduit equals 1.0, the resulted T value (0.438 $m^2$/sec) was in order-of magnitude with T value obtained by tracer test (0.98 $m^2$/sec). In addition, the T values for the rock matrix derived from analysis of the recession hydrograph ranged from $3.2\times10^{-4}$ to $42.2\times10^{-4}$ $m^2$/sec which is comparable with the range calculated from the aquifer tests (from $0.17\times10^{-4}$ to $4.8\times10^{-4}$ $m^2$/sec). Since conduit systems typically display considerable spatial variability, it was concluded that the derived T values for the conduit flow system may need to be established at a local scale, while the values for the diffuse flow system may be applicable at a regional scale. As a result, the expensive isotope and/or tracer techniques in fractured rocks may be restricted to calibrate the more practicable and, according to the study, reliable analytical approaches for estimation the aquifer constants in fractured rocks.

*KEY WORDS:* Hydrogeology, Fractured rocks, Lange Bramke Basin, Aquifer diffusivity, Flow type, Hydrograph analysis, Aquifer tests.

## INTRODUCTION

Groundwater studies in fractured terrains are, inherently difficult owing to the dichotomy of flow characteristics between the rock matrix and conduit systems (Dreiss, 1982; Lakey and Krothe, 1996; and White, 1999). Also, fractured aquifers show temporal and spatial variation and abrupt discontinuities between rock matrix and conduit-dominated flow regimes. The variable nature of hydraulic properties in these aquifers makes it difficult to quantify and predict the movement of groundwater through and/or between different parts of the aquifer. The characterization of meaningful hydraulic parameters in these aquifers is further complicated by inadequate techniques of data collection and/or incorrect interpretation of results. Field methods, such as aquifer tests and slug tests, which were developed to determine hydraulic properties for diffuse flow systems, may not provide accurate estimates for the conduit system. Even in the rare case where a conduit is either penetrated by or is next to the well being tested, the spatial variation in conduit aperture size and the potential that conduits may not be interconnected limits the applicability of test results to the area immediately around the well. As a result, hydraulic parameters derived from methods relying on well data typically do not provide accurate measurements of conduit properties (Király, 1975; and Padilla et al., 1994).

In contrast to groundwater moving toward most wells, most springs show the response of the entire fractured system to areal recharge events (Király, 1975; Dreiss, 1989; Sauter, 1992; Teutsch, 1992; Padilla et al., 1994; Lakey and Krothe, 1996). In well-developed fractured terrains, hydrographs from springs often show rapid responses to rain events followed by two or more distinct portions of a long recession limb. When quantifying hydraulic properties from hydrographs such as this, the interpretation is that water flowing through portions of the fractured aquifer with different conductivities produces the distinctive shape of the hydrograph. The fastest response is thought to be derived from flow through the conduits, and the slowest is thought to be from the diffuse portion of the continuum (Milanović, 1976, Piotr et al 1999 and Thies 2007). This method is based upon a Darcian theory of flow, which may not be valid for higher flow rates through the aquifer. Despite this theoretical limitation, this method and other similar hydrograph methods have been successfully applied to several fractured systems (Atkinson, 1977; Milanović, 1981; Sauter, 1992; Padilla et al., 1994; Shevenell, 1996, Baedke, 1998, Baedke and Krothe, 2001, Sahuquillo and Go´mez-Herna´ndez 2003). Other researchers have successfully explained the distinctive shapes of storm hydrographs in fractured aquifers in terms of water flowing through conduits with constrictions or debris-filled passages (Vineyard, 1958; Bonacci and Bojanić, 1991; Halihan et al., 1998). Regardless of the technique used to quantify flow through a fractured aquifer, basic assumptions regarding the spatial and temporal distribution of flow characteristics and the conduit/basin geometry relative to aquifer inputs and outputs need to be made. Since data concerning the shape and dimensions of the conduit system are not available for this study, it is prefer to apply the hydrograph analysis technique to attempt to derive meaningful hydraulic parameters of the Early Devonian sandstone fractured aquifer at Lange Bramke Basin.

***Location of Lange Bramke basin***

The Lange Bramke Basin is located in the western Harz Mountains, Lower Saxony, and 7 km south of Goslar (Figure 1). The area covers 0.76 $km^2$ with elevations from 538–700 m a.m.s.l., the average slope is 12.5° with quite homogeneous hillslopes. The climate is cool and wet with a mean annual precipitation of 1235 mm (1949–2005), which are almost equally distributed between summer and winter. The average snow cover season is 100 days (1949–85, Herrmann et al. 1989). The basin is drained in north-eastern direction by the Lange Bramke River.

***Geomorphology and geology of the study area***

Geomorphological and hydrological information are listed in Table 1. The characteristic stream discharges represent the investigation period 1981–2005 and are provided by Harzwasserwerke (2005).

The geology of Lange Bramke basin is simple. The Upper Harz Mountains Devonian Anticline is integrated in the structure of the upper Harz Mountains. It was folded in the Variscan time with a complex system of normal and reverse faults (Figure 2). These faults strike to NW-SE, intersect the folded structure into several segments and build zones of mylonite and disruption. The Lange Bramke basin runs in this disturbed anticline structure with predominant WNW-ESE joint directions. The core area of the anticline is marked by the floodplain of the creek.

***Hydrogeology of Lange Bramke basin***

The geohydrological units in Lange Bramke basin can be subdivided into three compartments (Figure 3a and 3b). The first unit is the unsaturated soil zone consisting of skeleton-rich, residual weathering and allochthonic materials of Pleistocene age. The solifluidal material has an average thickness of 3.5 m (Scheelen 1985). Main soil types are Cambisol podsols, covering about 40% of the catchment area, partially with pseudogleyzation. Podsol cambisols, by 20%, with vertical macropore systems, which enhance draining (Piotr et al 1999), Podsols, by 20% and Podsolic cambisols, by 10% (Deutschmann 1987). On the remaining 10% catchment area, different forms of gley soils are developed. Soil textures are mainly made up by silt, sandy loam (-60 cm) and medium silty sand (-120 cm). From the hydrological point of view the well developed macro pore systems of the cambisol podsols are important. Due to a high skeleton content and root channels, the water infiltration rate is very high. Sprinkling experiments with water amounts of 90 mm/h for several hours created no overland flow (Thies 2007). The typical average infiltration rate under saturated conditions is about 4 mm/min (Ueberschär 1988). Interflow was not detected at any time under these conditions in Lange Bramke.

The second unit is the fractured-rock groundwater system. Situated in the Upper

Harz Mountains Devonian Anticline, the Lange Bramke Basin is characterized by a Lower Devonian rock series (Ober-Ems). The investigation site is disturbed and dissected by numerous joint and fault systems (Figure 3a). Several major faults cross one another with partial fillings of quartz and argillaceous materials. The unsaturated upper zone of the fractured rock groundwater system is partially weathered and strongly dissected (Figure 3b). This allows seepage through the interface between the unsaturated soil and the fractured rock. Hardrock geology is mainly composed of quarzites, quarzitic sandstones and sandy shales (Hinze 1971). Joint water in the piezometer field is partly covered by an aquiclude (quartzitic shale), therefore the aquifer can be considered as semi-confined. Matrix porosity ranges from 1.5–3.2% and matrix permeability between $3.4\times10^{-8}$ and $2.1\times10^{-7}$ cm/sec. So, less disturbed zones act as barriers with respect to the fast water movement.

The third geohydrologic unit is the porous groundwater reservoir which is used by the river course and consists of gravel, including pebbles and some boulders (Figure 3b). Total volume of the porous groundwater reservoir is about $0.5 \times 10^{6}$ $m^3$ with an average thickness of about 10 m. On the basis of hydraulic assessment of the bedrock, the fractured and faulted zones were characterized as fast flow drain lines (Maloszewski et al. 1999). Transmissivity (T) assumed that the hydraulic continuum (125 $m^3$) of the piezomter field is $5.2 \times 10^{-5}$ $m^2$/sec (calculated from pumping tests, Herrmann et al. 1989). In addition, the study aquifer is controlled to a large degree by the nature of the fractured-rock groundwater systems. Rain and surface water percolate into the fractured and faulted zones, which act as fast-flow channels and moves vertically and laterally by conduit and diffuse flow. Major cross faults are assumed to be the dominant transport pathways which recharge the underlying third hydrogeologic unit. Evidence for this assumption is based on the fact that drilling fluids and rock flour appeared in Lange Bramke Creek several hours after drilling the borehole HKLU, which penetrated a major cross fault, had commenced (Herrmann et al 1989).

A combined hydrologic, environmental-isotope ($^3H$ and $^{18}O$) and NA and Uranine tracer study in Lange Bramke basin indicated that the mean tracer age of water leaving the Basin is about 2.6 years, 1.2 years of which correspond to flow in the saturated fractured part of the system. This leads to regional hydraulic conductivity of $3\times10^{-7}$ m/sec (Zuber and Motyka, 1994) while the hydraulic conductivity of the conduit is $1.5\times10^{-2}$ m/sec (Piotr et al 1999). In addition, the results of dye tracer test carried out by Müller and Wötzel 1996 shows a range of dominant flow velocity from $0.312\times10^{-5}$ to $1.17\times10^{-5}$ m/sec and from $0.57\times10^{-5}$ to $9.58\times10^{-5}$ m/sec, when NA tracer test was used (Thies 2007). Applying a mean distance of saturated flow in the fracture system of 100m, a laboratory measurement of matrix porosity (0.02), and mean hydraulic gradient of 0.2, this leads to hydraulic conductivity values similar to dominant flow velocity. Table 2 shows the hydraulic parameters of the different aquifers in Lange Bramke basin.

## MATERIAL AND METHODS

Daily discharge from spring A3 representing the groundwater flow through the major faults in the Early Devonian sandstone fractured aquifer in Lange Bramke drainage Basin was continuously monitored for the same 81 days at the hydrologic years 2001, 2002 and 2004 with weirs that were calibrated for a range of flow conditions with a rating curve (Table 3 and Figure 4). In addition, the daily groundwater level fluctuations during the year 2001 in wells HKLQ and HKLU, representing both conduit and diffuse flow systems respectively, was recorded (Figure 5). These data were obtained with pressure transducers and programmable data recorders. Precipitation data were also continuously recorded for the same periods with a tipping bucket style rain gauge and a data recorder that was programmed to provide hourly totals of rainfall.

The basis for quantifying hydraulic parameters from recession hydrographs comes from the approach proposed by (Rorabaugh 1964). Rorabaugh's approach is based on the analytical solution for the discharge to a perfectly connected and fully penetrating river in a homogeneous aquifer subject to an instantaneous uniform recharge, which is analogous to conduit flow system, and is given by the expression (Rorabaugh 1964):,

$$q = 2T\,(h_0/L)(e^{-\alpha t} + e^{-9\alpha t} + r^{-25\alpha t} + \ldots..) \quad (1)$$

With

$$\alpha = T\pi^2 / 4SL^2 \quad (2)$$

where q ($L^2T^{-1}$) is the groundwater discharge to the stream per unit of stream length; T ($L^2T^{-1}$) is the aquifer transmissivity, $h_0$ (L) is the rise in water level due to the instantaneous recharge, L (L) is the width of the aquifer perpendicular to the river, S (-) is the storage coefficient, and t (T) is time. Since, after a short critical time has passed, only the first term of the infinite sum of decreasing exponentials is significant, α, the recession coefficient and relates to the rate of release of water, can be identified with the slope of the hydrograph drawn in semilogarithmic paper, and from it, the aquifer diffusivity T/S, provided that width L is known, can be derived.

Accordingly, the most steeply sloped recession coefficient (α) may represent release of water from the conduit end-member of the fracture continuum, while the shallowest sloped coefficient may correspond to water released from the diffuse end-member. Atkinson (1977) rearranged Equation (2) to estimate the hydraulic parameters of an unconfined aquifer from base flow recession curves:

$$\log\left(\frac{Q_1}{Q_2}\right) = \frac{T}{S_y}(t_1 - t_2)\frac{1.07}{L^2} \quad (3)$$

Where, Sy is specific yield, $Q_1$ is the spring discharge at time $t_1$ ($L^3T^{-1}$), and $Q_2$ is the spring discharge at time $t_2$ ($L^3T^{-1}$). In order to establish a value for the variable L in a well-developed fractured terrain, where the boundaries of the drainage divide can change spatially and temporally under different aquifer conditions, detailed tracing experiments are required. Otherwise, it is common practice to assign each discharge location a value of L that is measured from the discharge point to the topographically defined groundwater drainage divide (Teutsch, 1992; Shevenell, 1996). Equation (3) can be rearranged so that the aquifer diffusivity (T/Sy) for each segment of the recession curve can be calculated once a value of L has been established and the slope of each segment of the recession curve is known,

$$\frac{T}{S_1} = \frac{\log(\frac{Q_1}{Q_2})}{(t_2 - t_1)} x \left(\frac{L^2}{1.07}\right) \quad (4)$$

The slope of each segment of the hydrograph recession curve was initially identified by visually picking breaks in slope on the recession hydrograph. The slope was then estimated by averaging the difference between every two successive daily records as log values giving a best fit slope values through the data. The best fit slope values at 2004 was chosen for deriving T/Sy ratios from recession hydrograph analysis as the aquifer tests were carried out in the same interval (Figure 6). Eisenlohr et al. (1997) used numerical modeling to produce simulated spring hydrographs from a combination of high- and low-permeability elements which support, theoretically, the applied method in this study.

In addition, two pumping and slug tests were carried out during this study (June 2004) to calibrate the aquifer diffusivity values derived from the recession hydrograph analysis method. The pumping test was carried out in well HKLQ which has total depth 36m and screen filter depth 13-15m. The pumping rate was 5E-4 $m^3$/sec for long duration of 11040sec. The observed drawdown and residual drawdown were recorded in four observation wells using Campbell Scientific's PST systems which can be networked to collect data simultaneously over a large area. Two observation wells, (HKLA & HKLB), are far from the pumped well by 5.4m while the others, (HKLD & HKLE), far distance of 7.4m. The results of pumping test analysis are shown in Figure 7 applying the following equation (Theis 1935).

$$h_{(t,r)} = V \cdot e - \left( \frac{r^2 \cdot S}{(4 \cdot \pi \cdot T \cdot t) \cdot (4 \cdot \pi \cdot t)} \right) \qquad (5)$$

Where h (t, r) is the drawdown (L), V is the volume (L), e is the effective porosity (-), r is the well radius (L), S is the storativity (-), T is the aquifer transmissivity ($L^2T^{-1}$) and t is the time since pumping started (T).

On the other hand, the over damped slug and bail tests were carried out in the same well (HKLQ). The techniques like the straight-line methods of Hvorslev (1951) and Bouwer and Rice (1976) were used in the processing and analysis of slug-test response data. In this technique the pressure transducer readings were plotted versus the time since some arbitrary starting point. The time at which the test began and the pressure head corresponding to static conditions were identified from this plot. The static pressure head and the test start time were then subtracted from the head and time records, respectively, to obtain a plot of the deviation of the pressure head from static conditions versus the time since test initiation. The type curves were superimposed on a plot of the test data. The dimensionless time axis was expanded or contracted until a reasonable match was obtained between type curve and the test data (Figure 8). Match points were then determined by reading the corresponding times from the real and dimensionless time axes. The radial hydraulic conductivity (K) was estimated by substituting the well-construction parameters, the c value, and the match-point ratio into the appropriate following equations ( Hvorslev 1951 and Bouwer and Rice 1976):

$$for\ L_w < h \quad \ln\left(\frac{R_e}{R}\right) = \left[ \frac{1.1}{\ln\left(\frac{L_w}{R}\right)} + \frac{A + B \ln\left(\frac{(h - L_w)}{R}\right)}{\frac{L_e}{R}} \right]^{-1} \qquad (6)$$

$$for\ L_w = h \quad \ln\left(\frac{R_e}{R}\right) = \left[ \frac{1.1}{\ln\left(\frac{L_e}{R}\right)} + \frac{C}{\frac{L_e}{R}} \right]^{-1} \qquad (7)$$

$$K = \frac{r^2 \ln\left(\frac{L_e}{R_e}\right) \ln\left(\frac{h_1}{h_2}\right)}{2 L_e (t_2 - t_1)} \qquad (8)$$

Where K is the hydraulic conductivity ($LT^{-1}$), $y_0$ and $y_t$ are the water level in time zero and time t respectively (L), $r_w$ is bore hole radius (L), c is the configuration constant (-).

## RESULTS AND DISCUSSIONS

Three rain events caused noticeable discharge response in spring A3 during the chosen period of records in 2001, 2002 and 2004 (Figure 4). The well hydrographs show the same trend for both the conduit (Well HKLU hydrograph) and diffuse flow system (Well HKLQ hydrograph) as shown in Figure 5. In these three rain events, three components of flow are identifiable, inasmuch as each component has a distinctly different slope on the discharge hydrographs (Figure 6). The spring hydrographs show nearly identical timing of responses and duration of time that each segment tends to dominate the hydrograph. However, the magnitude of response to recharge (precipitation) is annually different having greater discharge after each rain event during the 81 days records at 2004.

The steeply sloped segment of the recession hydrograph immediately following peak discharge represents flow that is predominately through conduits. The water in the conduits is quickly discharged at the spring and dominance of the hydrograph by the conduit system ceases. This portion of the hydrograph is followed by an intermediate, less steeply inclined segment, which could potentially represent either a mixing of water moving through the conduits and the arrival of a pulse of rain-derived groundwater moving through the rock matrix or a source of water coming from an third reservoir, such as smaller-scale fractures and fissures. Regardless of the interpretation of the intermediate segment, this portion of the hydrograph gradually dissipates, and spring discharge becomes dominated by groundwater released from the pores of the rock matrix. This last component of flow is seen on the recession hydrograph as the line segment with the lowest slope. A summary of the data presented in the following discussion is tabulated in Table 1.

Using Eq. 4 and the value of L (750m), T/Sy ratios of this portion of the hydrograph for spring A3 range from 0.024 to 1.298 $m^2$/sec, with an average value of 0.438 $m^2$/sec. By definition, Sy is the ratio of the volume of water drained from storage per unit surface area of an aquifer per unit decline in the water table (Freeze and Cherry, 1979). In a pure conduit the volume of water released from the conduit is equal to the surface area of the open conduit multiplied by the decline of head in the conduit. It is commonly assumed that the maximum value of Sy for the conduit is equal to 1.0 (Sauter, 1992 and Shevenell, 1996). This assumption is made since conceptually this portion of the hydrograph represents discharge from only the water directly entering and exiting the pure conduit and not from the matrix of the aquifer. Therefore it is assumed that the value of Sy for the conduit has reached the maximum value that it can and should be equal 1.0. Similarly, as the aquifer begins to make a proportionally larger contribution to the discharge hydrograph, the value of Sy will diminish until all discharge at the spring is attributed to water released from the diffuse flow system. Assuming that Sy equal to 1.0, this results in the transmissivity of the conduit system feeding the spring being equal to the respective T/Sy ratios (Table 4).

Knowing that the Early Devonian sandstone fractured aquifer averages 65.5 m in thickness (Table 2), the effective hydraulic conductivity K calculated from the tracer tests (0.015 m/sec, Piotr et al. 1999) can be used to provide an estimate of the T for the conduit system at spring A3 as 0.98 $m^2$/sec. This procedure overestimates the T of the conduit systems since the entire 65.5 m thickness of the studied aquifer is not likely to be fractured throughout the aquifer and the porosity of the conduit is probably not equal to 1.0 as was previously assumed. Given these limitations, the tracer test derived T appears to provide an order-of-magnitude estimate that is in the range of T derived from the hydrograph analysis method. As evidenced by the magnitudes of response to recharge and the differences in the values of T for the conduit systems of spring A3, the hydraulic parameters for the conduit system are only locally accurate around the spring. Therefore, the hydraulic parameters of the conduit system are only accurate at a relatively local scale.

Interpretations of the intermediate portion of the recession curve are more difficult than either the conduit or diffuse flow systems. Maloszewski and Zuber (1990) and Piotr et al. (1999) suggest that there may only be two reservoirs of water discharged by the springs. Furthermore, the modeling work of Eisenlohr et al. (1997) suggests that it is possible to observe three distinct segments on a recession hydrograph with only two distinct reservoirs (conduit and diffuse) of flow. With this interpretation of the data, the method of hydrograph analysis produces meaningless results for the intermediate flow system. This is because the method only produces meaningful estimates of T/Sy when applied to a physical reservoir of water. In this interpretation of the data a physical reservoir does not exist for the intermediate segment of the hydrograph. An alternate interpretation of the data could be made if the intermediate segment truly represents flow from a separate reservoir of water, such as flow from small fractures or fissures. Following the same approach described above, the T/Sy ratios calculated for the intermediate-flow system gives a range of values for spring A3 from 0.164 to 0.345 $m^2$/sec with an average value of 0.222 $m^2$/sec (Table 5). Since no data exist that assess either T or Sy of this hypothetical reservoir. only ratios of T/Sy can be established under these conditions.

In the other side. the diffuse flow component of the system dominated the hydrographs during eight periods. Recession hydrograph analysis for spring A3 suggests that the ranges of T/Sy for these twelve periods are 0.016 and 0.211 $m^2$/sec with average value of 0.107 $m^2$/sec (Table 5).

On the basis of observations that the water table was depressed during the aquifer tests by 0.3 and 0.45m in the observation wells HKLB and HKLE respectively, and that only very low pumping rates were capable of being sustained during the test (0.0005 $m^3$/sec), it was suggested that the wells used for these tests were screened in the diffuse portion of the aquifer. The results of these aquifer tests show that the transmissivity values (T) of the diffuse system estimated from the pumping test analysis is less than that estimated from both bail and slug tests (Table 5). This may attribute to the fact that slug tests in formations of high hydraulic conductivity (K) are often affected by mechanisms that are ignored in models developed for tests in less permeable formations (e.g., Hvorslev, 1951 and Bouwer and Rice, 1976). Given the complexity of slug-induced flow in a well screened in a highly permeable aquifer, it is difficult to account for all contributions to the effective length of the water column term. As a general, T values range from $0.17 \times 10^{-4}$ to $4.8 \times 10^{-4}$ $m^2$/sec with average value of $1.9 \times 10^{-4}$ $m^2$/sec (Table 6).

Applying this range of T values, Sy can be calculated from the ratio T/Sy derived from the flow recession hydrograph analysis for the diffuse system. It is possible to substitute the value of Sy from the aquifer tests (0.02 as mentioned before) and calculate the result T value (Table 7). Regardless of which value is calculated, the process of complimenting the hydrograph derived ratios of T/Sy with aquifer test data for the diffuse flow system provides a means of comparison for the results of the two methods. The values of T derived from the flow recession hydrograph analysis ($3.2 \times 10^{-4}$ to $42.2 \times 10^{-4}$ $m^2$/sec) are more or less comparable to the values of T determined by Uranine tracer test ($2.04 \times 10^{-4}$ to $7.7 \times 10^{-4}$ $m^2$/sec, Müller and Wötzel 1996), NA tracer tests ($3.73 \times 10^{-4}$ to $62.7 \times 10^{-4}$ $m^2$/sec, Thies 2007), and aquifer tests ($0.17 \times 10^{-4}$ to $4.8 \times 10^{-4}$ $m^2$/sec). The difference between these T values may attribute to the technique by which these tests were performed. This led to the suggestion that the hydrograph method may be an effective tool for determining hydraulic properties of the diffuse flow system in this highly fractured study area. Additionally, given that the average thickness of the Early Devonian fractured sandstone aquifer unit is 65.5 m, $3 \times 10^{-7}$ m/sec for regional hydraulic conductivity K of this unit (Piotr et al 1999), T value reaches $0.196 \times 10^{-4}$ which is comparable with the T value derived from pumping test analysis.

As is commonly done with aquifer test data, it may be possible to apply the values of hydraulic parameters derived from the flow recession hydrograph analysis to regional or sub-regional portions of the diffuse system in highly fractured aquifers.

## CONCLUSIONS

Spring A3 at the Early Devonian fractured sandstone aquifer unit, Lange Bramke Basin, Germany, has been continuously monitored for discharge. Rain event recession hydrographs show distinct segments that represent flow derived from conduits, rock matrix, and a mixture of water from the conduits and rock matrix or, alternatively, water from small fractures. When plotted on a log discharge versus time graph, the slope of each segment of the recession curve and the distance of the spring from the drainage divide can be used to calculate ratios of T/Sy for the three portions of the fractured continuum identified in this study. Assuming a maximum value of Sy as 1.0 for the conduit system allows calculating a T of the conduit system. Owing to the location of spring A3 in the center of the Basin, a single distance (750m) from the spring to the drainage divide was used to calculate the average values of T/Sy ratios. Incorporation of the Sy value for the conduit results in a range of T values of 0.024 –1.298 $m^2$ /sec for spring A3 with average value of 0.438 $m^2$/sec. This value may agree with T value obtained by tracer test of the spring A3 conduit network (0.98 $m^2$/sec). Since conduit systems typically display considerable spatial variability, T values for this part of the continuum need to be established at a local scale. Although an intermediate flow segment is partially identifiable on the hydrographs, the necessity that this segment represents reservoir of water is questionable. If the intermediate segment truly results from a mixing of the two other reservoirs of water and not a separate reservoir the calculated ratios of T/Sy for this portion of the hydrograph are meaningless. However, for fractured systems with a significant contribution from a third reservoir of water, such as small fissures, then meaningful ratios of T/Sy can be calculated for this portion of the fracture continuum. Such an analysis of the available data would result in a range of T/Sy values for this case from 0.164 to 0.345 $m^2$/sec. Aquifer tests show values of T for the rock matrix ranging from $0.17\times10^{-4}$ to $4.8\times10^{-4}$ $m^2$/sec with average value of $1.9\times10^{-4}$ $m^2$/sec and matrix porosity of 0.02. Assuming an average value of Sy equals to matrix porosity (0.02), T values from analysis of the hydrographs can be calculated. The results show that T values from analysis of the recession hydrograph ($3.2\times10^{-4}$ to $42.2\times10^{-4}$ $m^2$/sec) is in general range with the values estimated from the pumping tests. It is likely that the values of hydraulic parameters derived from the flow recession hydrograph analysis are representative of regional or sub-regional portions of the diffuse system in the study area. As a result, the expensive isotope and/or tracer techniques in fractured rocks may be restricted to calibrate the more practicable and, according to the study, reliable analytical approaches for estimation of aquifer constants.

**Acknowledgments:** This work was started during my scientific mission stay at Geoecology Institute, Technical University, Braunschweig, Germany and completed in Egypt. Assistance from Prof. Dr. A. Hermann and Prof. Dr. M. Schöniger and R. Thies, Geoecology Institute is gratefully acknowledged. I'm indebted to them for their help in performing aquifer tests and providing the required data. The work is supported by Egyptian Scholarship Administration and Desert Research Center, Egypt. I appreciate the insightful comments and suggestions of the editor and anonymous reviewers, which greatly improved this manuscript.

## REFERENCES

Atkinson, T. C., 1997. Diffuse flow and conduit flow in a limestone terrain in the Mendip Hills, Somerset, J. Hydrol., 19, 323–349.

Baedke, S. J., 1998. Hydrologic and geochemical assessment of the porous media and karstic flow regimes of the Big Clifty-Beech Creek aquifer (Ammunition Burning Grounds, Naval Surface Warfare Center, Crane, Indiana), Ph.D. dissertation, Indiana University, Bloomington.

Baedke, S. J. and Krothe, N. C., 2001. Derivation of effective hydraulic parameters of a karst aquifer from discharge hydrograph analysis, Water Resources Research, Vol. 37, No. 1, pp 13-19.

Bonacci, O., and D. Bojanic', D., 1991. Rhythmic karst springs, Hydrol. Sci. J., 36(1), 35–47.

Bouwer, H., and R.C. Rice, 1976. A slug test for determining hydraulic conductivity of unconfined aquifers with completely or partially penetrating wells. Water Resour. Res. 12, no. 3: 423-428.

Deutschmann,G.1987.Bodenhydrologische Eigenschaften der Waldstandorte der Oberharzer Untersuchungsgebiete auf Grundlage der forstlichen Standortskartierung. Diploma Thesis. Institute of Geography, TU-Braunschweig.

Dreiss, S. J., 1982. Linear kernels for karst aquifers, Water Resour. Res., 18 (4), 865–876.

Dreiss, S. J., 1989. Regional scale transport in a karst aquifer, 1, Component separation of spring flow hydrographs, Water Resour. Res., 25(1), 117–125, 1989.

Eisenlohr, L., L. Kira'ly, M. Bouzelboudjen, and Y. Rossier, 1997. Numerical simulation as a tool for checking the interpretation of karst spring hydrographs, J. Hydrol., 193, 306–315.

Freeze, R. A., and J. A. Cherry, 1979. Groundwater, 604 pp., Prentice-Hall, Englewood Cliffs, N. J.

Halihan, T., C. M. Wicks, and J. F. Engeln, 1998. Physical response of a karst drainage Basin to flood pulses: Example of the Devil's Icebox case system (Missouri, USA), J. Hydrol., 208, 24–36.

Harzwasserwerke 2005, Harzwasserwerke GmbH, Jahresliste 2005 des Pegels Lange Bramke. Bibliography

Herrmann A., Koll J., Leibundgut Ch., Maloszewski P., Rau R., Rauert W., Schoniger M., and Stichler W, 1989. Wasserumsatz in einem kleinen Einzugsgebiet im palaozoischen Mittelgebirge. Eine hydrologische Systemanalyse mittels Umweltisotopen als Tracer (Turnover of water in a small catchment area of Paleozoic rock. A hydrological system analysis by using environmental isotopes as tracers). Landschaftsokologie und Umweltforschung 17:1-305.

Hinze, C. 1971. Erläuterungen zur geologischen Karte 1:25000. Blatt Clausthal-Zellerfeld (4128), Nds. Landesamt für Bodenforschung, Hannover.

Hvorslev, M. J., 1951. Time lag and soil permeability in ground-water observations. U.S. Army Corps of Engrs. Waterways Exper. Sta. Bull no. 36.

Kira'ly, L., 1975. Report on the current knowledge in the domain of physical properties of karstic rocks, Hydrogeology of Karstic Terrains, in Int. Union of Geol. Sci., Ser. B, vol. 3, pp. 53–67.

Lakey, B., and N. C. Krothe, 1996. Stable isotopic variation of storm discharge from a perennial karst spring, Water Resour. Res., 32(3), 721–731.

Maloszewski P., and Zuber A., 1990. Mathematical modeling of tracer behaviour in short-term experiments in fractured rocks. Water Res., Res., 26: 1517-1528.

Milanovic', P. T., 1976. Water regime in deep karst: Case study of the Ombla Spring drainage area, In: Yevjevich, V.

(Ed), Karst Hydrology and Water Resources: Proceedings of the US-Yugoslavian Symposium, Dubrovnik, 1975, vol. 1, edited by V. Yeujevich, pp. 165–191, Water Resour. Publ., Littleton, Colo.

Milanovic´, P. T., 1981. Karst Hydrogeology, 434 pp., Water Resour. Publ., Littleton, Colo.

Müller, A. and Wötzel, K., 1996. Durchführung eines Feldtracerexperiments und numerische Auswertung mit dem Programmsystem FEFLOW zum Verhalten von Fluoreszenztracern in einem geometrisch begrenzten, paläozoischen Festgesteinsbereich (Lange Bramke, Ober-harz). Study work, Institute of Geography, TU-Braunschweig.

Padilla, A., A. Pulio-Bosch, and A. Mangin, 1994. Relative importance of baseflow and quickflow from hydrographs of karst springs. Ground Water, 32(2), 267–277.

Piotr M., Herrmann A., and Zuber A., 1999. Interpretation of tracer tests performed in fractured rock of the Lange Bramke Basin, Germany. Hydrogeology Journal, 7:209-218.

Rorabaugh, M. I., 1964. Estimating changes in bank storage as ground-water contribution to streamflow, Int. Assoc. Sci. Hydrol. Publ., 63, 432– 441.

Sahuquillo, A. and Go´mez-Herna´ndez, J. J., 2003. Comment on "Derivation of effective hydraulic parameters of a carst aquifer from discharge hydrograph analysis" by S. J. Baedke and N. C. Krothe, Water Resources Research, V., 39, No. 6, 1152, doi:10.1029/2002WR001472, 1-3 P.

Sauter, M., 1992. Assessment of hydraulic conductivity in a karst aquifer at local and regional scale, in Ground Water Manage., 10, 39–57.

Scheelen, A. 1985. Refraktionsseismische Messungen im Einzugsgebiet Lange Bramke(Oberharz) zur Entwicklung eines Speichermodells. Diplomathesis, Institut of Geography, TU-Braunschweig.

Shevenell, L., 1996. Analysis of well hydrographs in a karst aquifer: Estimates of specific yields and continuum transmissivities, J. Hydrol., 174, 331–355.

Teutsch, G., 1992. Groundwater modeling in karst terranes: Scale effects, data acquisition and field validation, Ground Water Manage., 10, 17–35.

Theis, C. V., 1935. The relation between the lowering of the piezometric surface and the rate and duration of discharge of a well using groundwater storaagt: Am. Geophys. Union Trans., Vol. 16, pp. 510-524.

Thies, R. 2007. Determination of runoff generation processes in a fissured rock aquifer Lange Bramke catchment, Harz Mountains, Diploma Thesis. Institute of Geography, TU-Braunschweig. 94 P

Ueberschär, R. 1988. Infiltration undWasserbewegunginskelettreichen Böden auf Hangstandorten im Untersuchungsgebiet der Langen Bramke (Oberharz). Diploma Thesis. Institute of Geography, TU-Braunschweig.

Vineyard, J., 1958. The reservoir theory of spring flow, NSS Bull., 20, 46–53.

White, W. B., 1999. Conceptual models for karstic aquifers, in Karst Modeling, Spec. Publ., vol. 5, pp. 11–16, Karst Waters Inst., Charles Town, W. Va.

Zuber A., and Motyka J., 1994. Matrix porosity as the most important parameter for transport at large scales. J Hydrol 167:19-46.

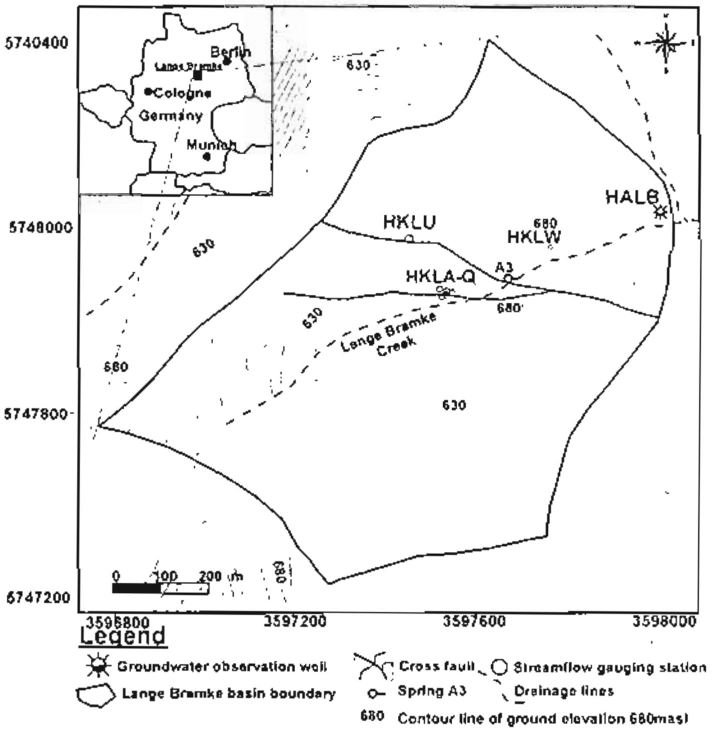


*Figure 1. Location map of the present 22 observation wells in experimental Lange Bramke basin, Germany*

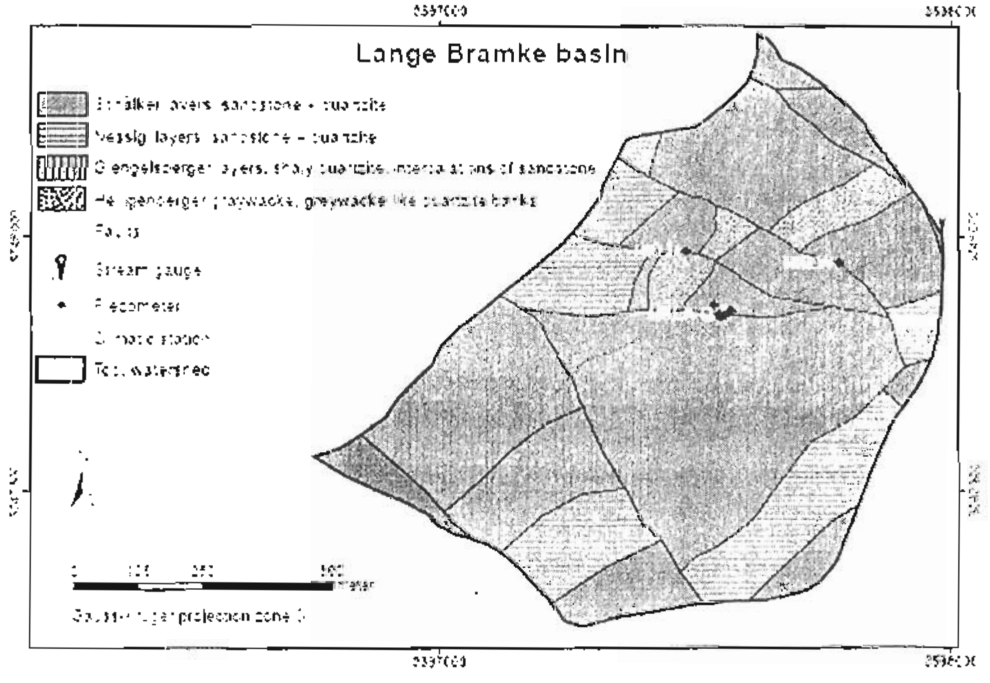


*Figure 2: Geological map of the basin (Geologic map 1.25000 sheet 4128 Clausthal-Zellerfelf and Hinze 1971 and modified by Herrmann et al. 1989)*

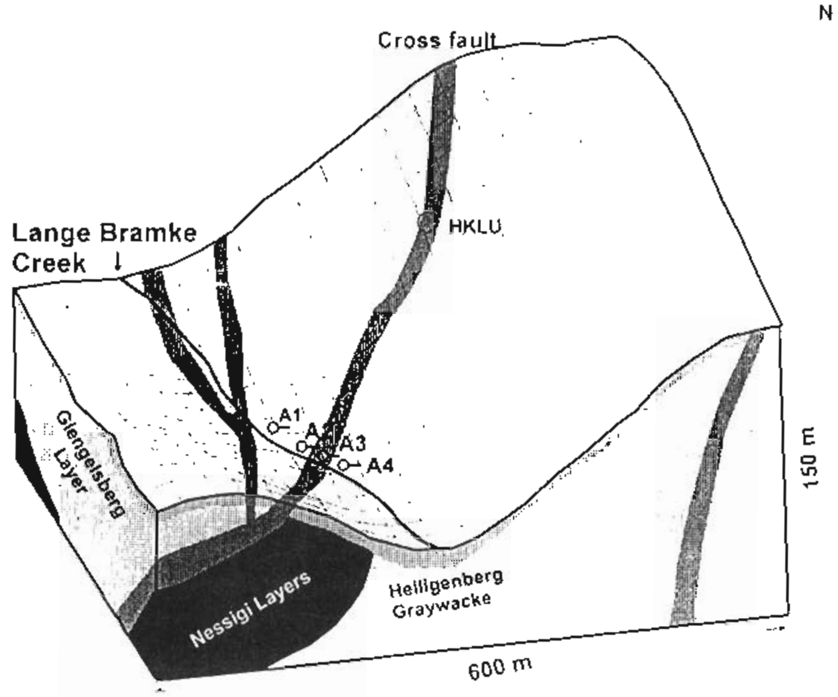


*Figure 3a: 3D view of the Lange Bramke Basin showing geology, major faults and experimental sites (after Piotr et al 1999)*

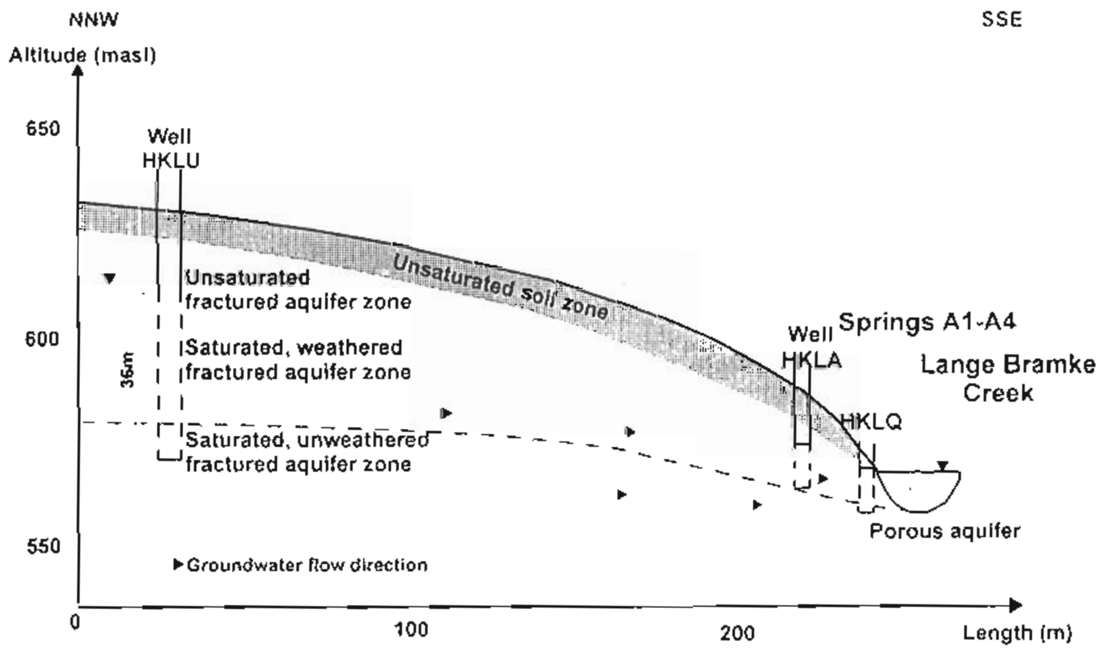


*Figure 3b. Simplified hydrogeological cross-section along the fault line between the well HKLU and spring A3 (after Piotr et al 1999)*

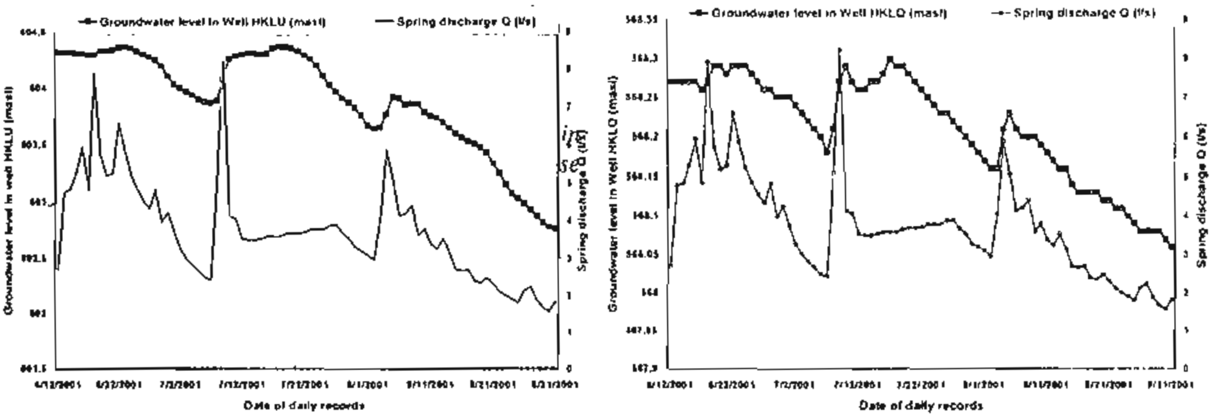


*Figure 4: The 81 day discharge hydrographs for spring A3 during the period 12June till 31 August at the years 2001, 2002 and 2004 in Lange Bramke Basin.*

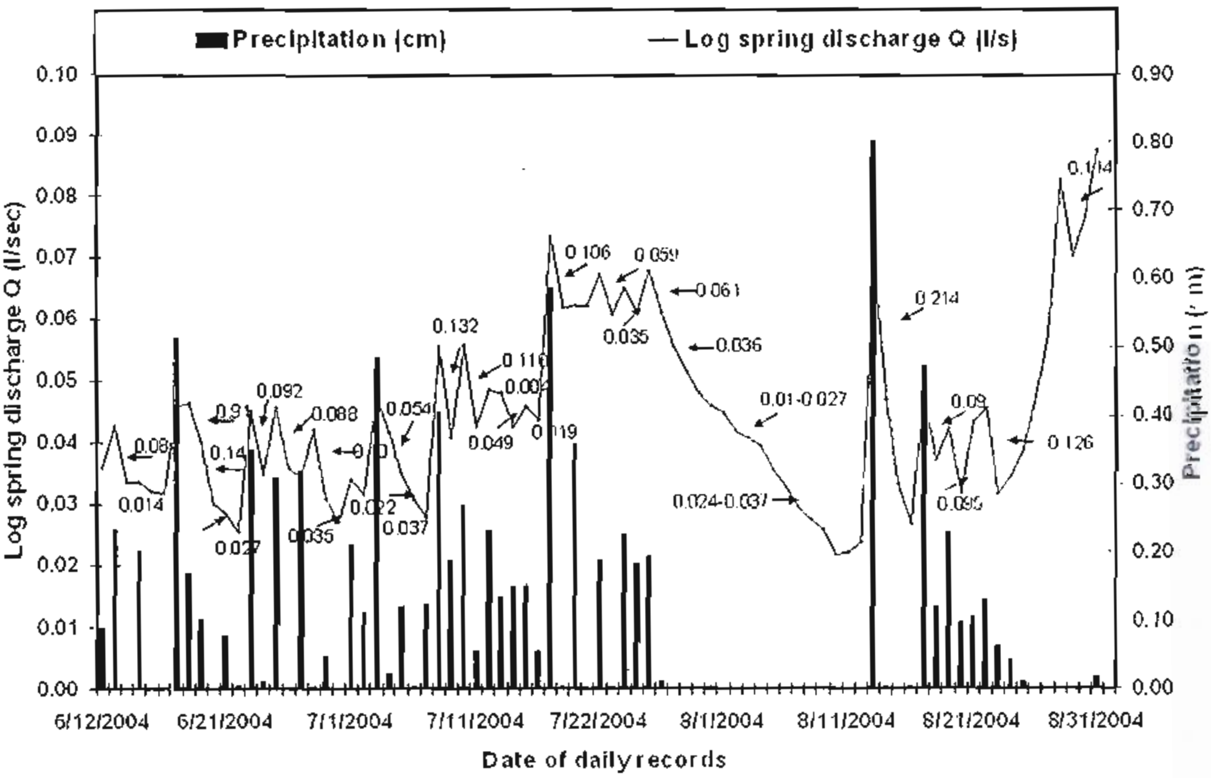


*Figure 5: The groundwater level fluctuations in the experimental wells penetrating conduit (Well HKLU) and diffuse (Well HKLQ) flow systems in Lange Bramke Basin during 2001*

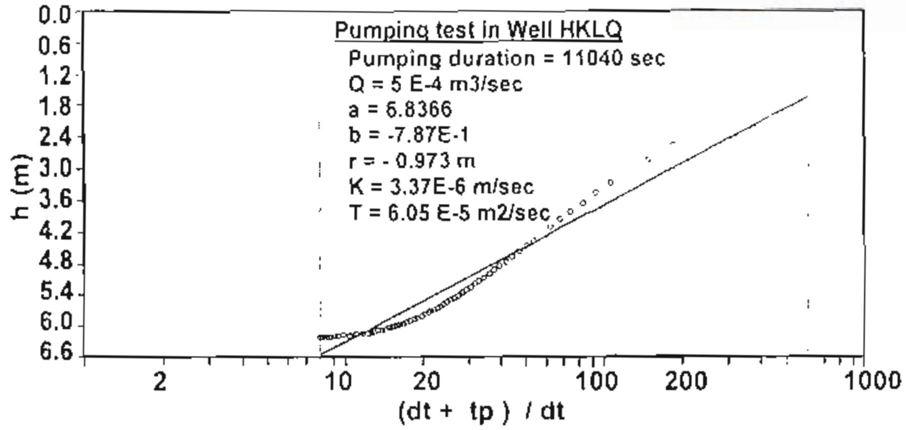


*Figure 6: The results of pumping test analysis in well HKLQ applying Theis 1935 method (diffuse flow system)*

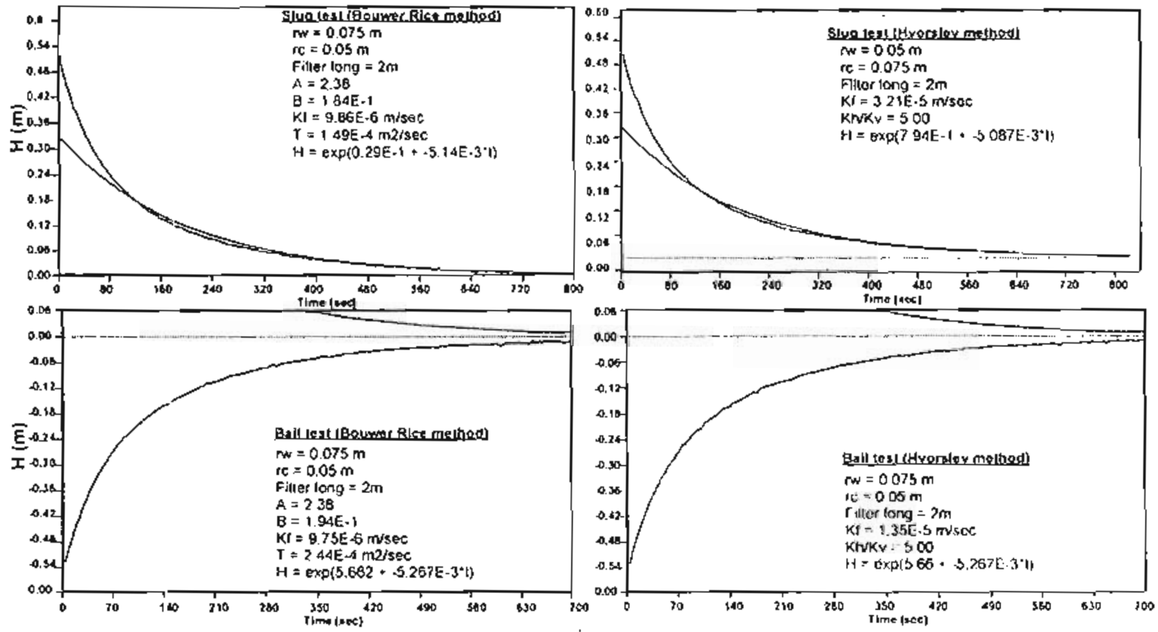


*Figure 7. The results of slug and bail tests analysis applying Hvorslev 1951 and Bouwer and Rice 1976 straight-line methods (diffuse flow system)*

*Table 1: Geomorphological and hydrological characteristics of Lange Bramke Basin (after Herrmann et al. 1989 and Harzwasserwerke 2005).*

| Geomorphological characteristics | | Unit |
|---|---|---|
| Area size | 0.78 | $km^2$ |
| Minimum elevation | 537 | Masl |
| Maximum elevation | 700 | Masl |
| Average elevation | 660 | Masl |
| Forested area | 90 | % |
| Hydrological characteristics | | |
| Minimum low flow discharge | 0.07 | l/sec |
| Average low flow discharge | 1.6 | l/sec |
| Average discharge | 14.4 | l/sec |
| Average flood discharge | 221.9 | l/sec |
| Maximum flood discharge | 676 | l/sec |
| Maximum stream length to watershed | 1.32 | Km |
| *Average slope of main stream* | 11 | % |

*Table 2: Parameters of subsurface main reservoirs in Lange Bramke Basin (after Herrmann et al. 1989)*

| Parameter | Soil reservoir | Porous reservoir | Fractured reservoir |
|---|---|---|---|
| Area ($10^6$ $m^2$) | 0.71 | 0.05 | 0.76 |
| Volume ($10^6$ $m^3$) | 2.5 | 0.5 | 50 |
| Average aquifer thickness (m) | (3.5) | 10 | 65.5 |
| Volume of water ($10^6$ $m^3$) | 0.5 | 0.16 | 0.6 |
| Volume of mobile water ($10^6$ $m^3$) | ? | 32 | 1.6 |
| Total porosity (%) | 20 | 32 | 1.6 |
| Matrix porosity (%) | ? | -- | 0.5 |
| Fissure porosity (%) | ? | -- | 1.1 |
| *Effective porosity (%)* | ? | 32 | 1.1 |

*Table 3: The daily spring discharge and precipitation records during the same 81 days in the study area*

| Date of record | Records of 2001 | | | Records of 2002 | | | Records of 2004 | | |
|---|---|---|---|---|---|---|---|---|---|
| | P (cm) | Log Q (l/sec) | T/Sy in $m^2$/sec | P (cm) | Log Q (l/sec) | T/Sy in $m^2$/sec | P(cm) | Log Q (l/s) | T/Sy in $m^2$/sec |
| 12-Jun | 0.2 | 0.431 | ---- | 2.93 | 0.957 | 0.345 | 0.01 | 0.323 | ---- |
| 13-Jun | 0 | 0.679 | ---- | 1.17 | 0.9 | 0.666 | 0.026 | 0.385 | 0.503 |
| 14-Jun | 0 | 0.684 | ---- | 0.14 | 0.79 | 0.009 | 0 | 0.302 | ---- |
| 15-Jun | 13.4 | 0.723 | ---- | 0.86 | 0.789 | 0.271 | 0.023 | 0.302 | 0.082 |
| 16-Jun | 0 | 0.774 | 0.541 | 0.02 | 0.744 | 0.536 | 0 | 0.288 | 0.016 |
| 17-Jun | 2.8 | 0.685 | ---- | 0 | 0.656 | 0.556 | 0 | 0.286 | ---- |
| 18-Jun | 20.9 | 0.898 | 0.856 | 0 | 0.565 | ---- | 0.057 | 0.411 | ---- |
| 19-Jun | 0 | 0.757 | 0.257 | 0 | 0.603 | ---- | 0.019 | 0.418 | 0.342 |
| 20-Jun | 0.2 | 0.715 | ---- | 0.32 | 0.658 | ---- | 0.011 | 0.361 | 0.551 |
| 21-Jun | 0.2 | 0.722 | ---- | 0.01 | 0.663 | 0.253 | 0 | 0.271 | 0.083 |
| 22-Jun | 10.9 | 0.82 | 0.309 | 0 | 0.621 | 0.096 | 0.009 | 0.257 | 0.164 |
| 23-Jun | 0.8 | 0.769 | 0.31 | 0.16 | 0.605 | 0.169 | 8.00E-04 | 0.23 | ---- |
| 24-Jun | 0 | 0.718 | 0.194 | 0 | 0.577 | 0.61 | 0.039 | 0.406 | 0.558 |
| 25-Jun | 0 | 0.686 | 0.169 | 0 | 0.477 | 0.347 | 0.001 | 0.314 | ---- |
| 26-Jun | 0 | 0.658 | 0.131 | 0 | 0.42 | ---- | 0.035 | 0.411 | 0.536 |
| 27-Jun | 4.9 | 0.636 | ---- | 0.1 | 0.453 | ---- | 0 | 0.323 | 0.094 |
| 28-Jun | 0.2 | 0.684 | 0.524 | 0.77 | 0.549 | 0.201 | 0.035 | 0.308 | ---- |
| 29-Jun | 0 | 0.598 | ---- | 0.13 | 0.516 | 0.781 | 4.00E-04 | 0.38 | 0.616 |
| 30-Jun | 7.6 | 0.627 | 0.345 | 0 | 0.387 | ---- | 0.005 | 0.278 | 0.211 |
| 1-Jul | 0.1 | 0.571 | 0.348 | 1.02 | 0.486 | ---- | 0 | 0.243 | ---- |
| 2-Jul | 0 | 0.513 | 0.219 | 0.42 | 0.513 | 0.468 | 0.023 | 0.306 | 0.133 |
| 3-Jul | 0 | 0.477 | 0.163 | 0.41 | 0.436 | ---- | 0.013 | 0.284 | ---- |
| 4-Jul | 0 | 0.45 | 0.154 | 1.16 | 0.558 | 1.179 | 0.054 | 0.43 | 0.33 |
| 5-Jul | 0 | 0.425 | 0.185 | 0.01 | 0.364 | 0.856 | 0.003 | 0.375 | 0.364 |
| 6-Jul | 0 | 0.394 | 0.065 | 0.05 | 0.223 | ---- | 0.013 | 0.315 | 0.227 |
| 7-Jul | 5 | 0.384 | ---- | 0.07 | 0.228 | 0.993 | 4.00E-04 | 0.278 | 0.159 |
| 8-Jul | 20.9 | 0.707 | ---- | 0 | 0.064 | 0.391 | 0.014 | 0.252 | ---- |
| 9-Jul | 25.7 | 0.914 | 1.807 | 1.16 | 0 | ---- | 0.045 | 0.499 | 0.801 |
| 10-Jul | 0 | 0.617 | 0.058 | 1.74 | 0.476 | 0.036 | 0.021 | 0.367 | ---- |
| 11-Jul | 4.4 | 0.607 | 0.37 | 0.75 | 0.47 | 1.232 | 0.03 | 0.501 | 0.72 |
| 12-Jul | 1.4 | 0.547 | 0.023 | 0.01 | 0.267 | 0.607 | 0.006 | 0.383 | ---- |
| 13-Jul | 2.6 | 0.543 | 0 | 0.02 | 0.167 | ---- | 0.026 | 0.437 | 0.024 |

| | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|
| 14-Aug | 0 | 0.5 | 0.406 | 0 | 1.935 | 1.191 | 4.00E-04 | 0.423 | 0.75 |
| 15-Aug | 0 | 0.433 | 0.039 | 0 | 1.739 | 0.94 | 0 | 0.299 | 0.345 |
| 16-Aug | 6.9 | 0.427 | --- | 0 | 1.584 | 0.758 | 4.00E-04 | 0.243 | --- |
| 17-Aug | 0.1 | 0.433 | 0.332 | 0 | 1.459 | 0.482 | 0.053 | 0.426 | 0.547 |
| 18-Aug | 0.2 | 0.378 | 0.045 | 0.72 | 1.38 | 0.741 | 0.013 | 0.336 | --- |
| 19-Aug | 0 | 0.371 | --- | 0 | 1.258 | 0.522 | 0.025 | 0.382 | 0.576 |
| 20-Aug | 2.7 | 0.394 | 0.176 | 0 | 1.172 | 0.109 | 0.011 | 0.287 | --- |
| 21-Aug | 0 | 0.365 | 0.238 | 0.84 | 1.154 | --- | 0.012 | 0.392 | --- |
| 22-Aug | 0 | 0.326 | 0.154 | 2.94 | 1.268 | 0.399 | 0.015 | 0.412 | 0.768 |
| 23-Aug | 0 | 0.301 | 0.122 | 1.39 | 1.202 | 0.371 | 0.007 | 0.285 | --- |
| 24-Aug | 0 | 0.281 | 0.157 | 0 | 1.141 | --- | 0.005 | 0.311 | --- |
| 25-Aug | 1.2 | 0.255 | --- | 0 | 1.146 | 0.112 | 0.001 | 0.35 | --- |
| 26-Aug | 4.4 | 0.33 | --- | 0.01 | 1.127 | 0.139 | 0 | 0.428 | --- |
| 27-Aug | 6.1 | 0.35 | 0.463 | 0 | 1.104 | 0.209 | 0 | 0.517 | --- |
| 28-Aug | 0 | 0.274 | 0.313 | 0 | 1.07 | 0.272 | 0 | 0.746 | 0.695 |
| 29-Aug | 0 | 0.223 | 0.18 | 0 | 1.025 | 0.27 | 0 | 0.631 | --- |
| 30-Aug | 0.3 | 0.193 | --- | 0 | 0.981 | 0.291 | 0 | 0.69 | --- |
| 31-Aug | 2.7 | 0.26 | 0.291 | 0 | 0.933 | 0.279 | 0.002 | 0.788 | 0.131 |

*Table 4: The T/Sy ratios derived from recession hydrograph analysis for the conduit flow system*

| Date of record | T/Sy ratios ($m^2$/sec) | Date of record | T/Sy ratios ($m^2$/sec) | Date of record | T/Sy ratios ($m^2$/sec) |
|---|---|---|---|---|---|
| 13-Jun | 0.503 | 9-Jul | 0.801 | 28-Jul | 0.218 |
| 15-Jun | 0.082 | 11-Jul | 0.720 | 29-Jul | 0.185 |
| 19-Jun | 0.342 | 13-Jul | 0.024 | 30-Jul | 0.115 |
| 20-Jun | 0.551 | 14-Jul | 0.301 | 13-Aug | 1.298 |
| 24-Jun | 0.558 | 16-Jul | 0.115 | 14-Aug | 0.750 |
| 26-Jun | 0.536 | 18-Jul | 0.644 | 17-Aug | 0.547 |
| 29-Jun | 0.616 | 22-Jul | 0.359 | 19-Aug | 0.576 |
| 2-Jul | 0.133 | 24-Jul | 0.212 | 22-Aug | 0.768 |
| 4-Jul | 0.330 | 26-Jul | 0.369 | 28-Aug | 0.695 |
| 5-Jul | 0.364 | 27-Jul | 0.302 | 31-Aug | 0.131 |
| Average value of aquifer diffusivity (T/Sy ratio) | | | | | 0.438 |

*Table 5. The T/Sy ratios derived from recession hydrograph analysis for mixed and the diffuse flow system*

| Mixed flow (Intermediate stage) | | Diffuse flow system (rock matrix) | | | |
|---|---|---|---|---|---|
| Date of record | T/Sy ratios ($m^2$/sec) | Date of record | T/Sy ratios ($m^2$/sec) | Date of record | T/Sy ratios ($m^2$/sec) |
| 22-Jun | 0.164 | 16-Jun | 0.016 | 2-Aug | 0.053 |
| 1-Aug | 0.166 | 21-Jun | 0.083 | 3-Aug | 0.079 |
| 4-Aug | 0.210 | 27-Jun | 0.094 | 5-Aug | 0.149 |
| 9-Aug | 0.225 | 30-Jun | 0.211 | 6-Aug | 0.181 |
| 6-Jul | 0.227 | 7-Jul | 0.159 | 7-Aug | 0.109 |
| 15-Aug | 0.345 | 31-Jul | 0.059 | 8-Aug | 0.095 |
| Average | 0.222 | | 0.107 | | |

*Table 6: Hydraulic parameters of Early Devonian sandstone fractured aquifer at Lange Bramke Basin resulted from aquifer test analysis (Diffuse flow system)*

| Kind of test | Method of analysis | K (m/sec) | T ($m^2$/sec) |
|---|---|---|---|
| Pumping | Theis (1935) | 3.4 E-6 | 0.51 E-4 |
| Pumping | Theis (1935) | 1.2 E-6 | 0.17 E-4 |
| Bail | Hvorslev (1951) | 1.4 E-5 | 2.0 E-4 |
| Bail | Bouwer and Rice (1976) | 9.8 E-5 | 2.4 E-4 |
| Slug | Hvorslev (1951) | 1.3 E-5 | 2.0 E-4 |
| Slug | Hvorslev (1951) | 3.2 E-5 | 4.8 E-4 |
| Slug | Bouwer and Rice (1976) | 1.0 E-5 | 1.5 E-4 |
| *Average* | | 2.4 E-5 | 1.9 E-4 |

*Table 7: Summary of the hydraulic parameters derived from the hydrograph method and aquifer tests in Lange Bramke Basin.*

| Aquifer portion and kind of test | T/Sy ($m^2$/sec) | Sy (--) | T ($m^2$/sec) | K (m/sec) |
|---|---|---|---|---|
| Conduit | | | | |
| Spring hydrograph analysis | 0.024 - 1.298 | 1 | 0.024 - 1.298 | --- |
| Dye trace [a] | NA | NA | 0.98 [a] | 0.015 [a] |
| Diffuse | | | | |
| Spring hydrograph analysis | 0.016 - 0.211 | 0.02 | $3.2\times10^{-4} - 42.2\times10^{-4}$ | $3\times10^{-7}$ |
| Uranine tracer test [b] | NA | NA | $2.04\times10^{-4} - 7.7\times10^{-4}$ [b] | $0.312\times10^{-5} - 1.17\times10^{-5}$ [b] |
| NA Tracer test [c] | NA | NA | $3.73\times10^{-4} - 62.7\times10^{-4}$ [c] | $0.57\times10^{-5} - 9.58\times10^{-5}$ [c] |
| Aquifer tests | --- | 0.02 | $0.17\times10^{-4} - 4.8\times10^{-4}$ | $0.12\times10^{-5} - 9.8\times10^{-5}$ |
| Intermediate | | | | |
| Spring hydrograph analysis | 0.164 - 0.345 | NA | NA | NA |

a= Data from Piotr et al (1999), b = Data from Müller and Wötzel (1996) and c = Data from Thies (2007)