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Conception of a high-resolution saturated model for the test site Pirna

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Declaration

I hereby declare that this is an independent research work. The thesis or any part of it has not been submitted to any institution or organization for review or publication purposes. Information derived from work of others has been indicated and its resources are listed in the references.

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ABSTRACT

The field site of Pirna has been used for many application and development of field exploration techniques. Based on the data obtained from previous investigations and long-term water level measurements, a 3D groundwater flow model was constructed to represent the local hydrogeological condition and flow process. The model was developed using MODFLOW-2005 for the period from June to October in the year of 2014. Interpolation methods have been applied to distribute hydraulic conductivity, elevation of each layer and initial head. Boundary conditions were set as the specific-head type for both Elbe river and Northern boundary. The model was calibrated using trial and error method by comparison of the head difference between the observation and simulation. The most important criteria for checking the calibrated model include Root-Mean-Square (RMS) error, Absolute Mean (MA) error, and percent discrepancy of water budget. The calibration was confirmed with RMS less than 10.6 cm, MA less than 9.6 cm, and percent discrepancy of water budget less than -0.05 for the entire time period. The Elbe river and groundwater are hydraulic connected, the groundwater can either gaining water from Elbe river or losing water to Elbe river. The resulting water table configuration can reflect the flow process within the aquifer and its relationship to the river stage. The modeling result showed the success of stratigraphic representation and could be used as a long-term assessment of the field with continued measurements.

Objectives

1. Collect and evaluate data to characterize the hydrogeological condition of test site.

2. Conceptualization of the groundwater and river regime based on the available data.

3. Numerical model set up represent the field site.

4. Draw conclusion of the hydrogeological condition.

5. Evaluation of interaction between groundwater and Elbe river.

6. Recommendation for future investigation.

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1 Introduction

1.1 Overview and scope of work

Groundwater is one of the most broadly distributed water resource that can be viewed as a long-term natural ‘reservoir’ used for agriculture, municipal, and environmental purposes. Groundwater can be fed by recharge processes of precipitation or surface water infiltration, and can discharge as baseflow to rivers or be extracted from wells.

Getting a good understanding of the local hydrogeological condition and flow processes plays a significant role in application and development of field exploration techniques. The study area of Pirna, located along the Elbe river, has been conducted for a number of field tests, including Direct Push Slug Test and Direct Push Injection Logging carried out by Dietze and Dietrich in 2006 to evaluate to vertical characteristics in hydraulic conductivity of the sediments, the tracer tests with Fluorescein and other particles for estimation of the flow direction and velocity, and sieve analysis of hydraulic conductivity of sediments in the filed etc. At present, a network of observation wells with diameter of 1 inch has been installed. Regular measurement of water table and the close-by Elbe river was taken for a long period. Based on the available data， a groundwater model could be constructed for delineation of the field.

Nowadays, groundwater modeling has been widely used to simulate and predict aquifer conditions, which proved to be an effective and convenient tool to investigate the complicated subsurface conditions. A model is simulation of the groundwater system that summarizes the groundwater problems and current understanding of the field site information. Nevertheless, application of groundwater model requires extensive information for data input and model calibration, the more sufficient and accurate of the geologic and hydrogeological data of field site, the more reliable of the simulation and prediction.

Groundwater and surface have long been considered separate entities, and have been investigated individually. The interactions between surface water and groundwater have gained more attentions in the recent decades. Flow flux between groundwater and surface water may influence the water quality of either of these two hydrological parts. Generally, the interaction between groundwater and surface water mainly falls into two types: groundwater flows through the streambed into the stream, and stream water infiltrates through the sediments into the groundwater. The flow direction is a function of the hydraulic head difference of the two systems. Precipitation events can sometimes alter groundwater table and stream stages, and therefore change the flow exchange direction.

The best method to investigate the subsurface properties is to measure the sediments directly, but in reality, it is almost impossible to get data of every points we are interested. Therefore, to analyze sediments at selected locations and then interpolation or extrapolation of the variations based on the locations where measurements were performed could be a useful approach in understanding the test site and form a data set that can be input to the model.

This study is focused on a small part of the aquifer with known water table, river stage, hydraulic conductivity distribution and precipitation. The finite-difference flow model code Modflow-2005 was selected for investigation of the filed site, which is used to simulate the three-dimensional transient ground flow. Main task of the constructing the model include: having a preliminary understanding of the hydrogeological properties of the local conditions and local flow process, evaluating interaction of aquifer and Elbe river, especially influence from Elbe on groundwater level dynamics.

The model was build up with a relatively fine spatial grid and small time step. The rectangular cell size was uniformly set with length of 2 m, and daily time step was applied. The model was constructed for a period of June to October in 2014, which is 151 days in total. During this period, 110 groundwater level measurements were taken for the observation wells approximately on a daily basis. Data used for construction of the model is mainly based previous investigation and field measurements.

1.2 Study area

The field site Pirna is situated adjacent to the Elbe river, in the Institute of Waste Management and Contaminated Site Treatment. The meadow field stretches from the main institute building to the right riverbank is suitable for many hydrogeological investigation at test-field scale.

Figure 1 : Study area of Pirna

In this research, study area is approximately 2.52 ha with 140m in width and 180m in length. Gauss-Krüger coordinates was used to define the location of observation wells, model domain range from 553872.84 m to 563412.86 m longitude east, and from 1543100.28 m to 1553630.30 m latitude north. Distance between the observation wells to the riverbank varies from 70 m to 150 m.

Elbe river originates from the Czech Republic and discharges into the North Sea near the city of Hamburg, drains nearly 100,000 km^2 catchment area within Germany. Pirna is characterized by a relatively temperate climate with warm summers and cold winters. Annual precipitation is roughly 800mm during which precipitation falls almost throughout the year. The annual average air temperature is 9 °C, with the coldest days mostly being measured in January and the hottest temperatures generally being measured in July.

The lithological units and the aquifer dip slightly toward the riverbank. The aquifer is formed of interbedded strata of medium sand to coarse gravel which are followed by fine sands with changing thicknesses. (Dietze & Dietrich, 2012) The basement is formed of marine sediment rocks consisting of sand- and mudstone of Upper Cretaceous age.

2 Literature review

2.1 Interaction between groundwater and streams

Groundwater and surface water have long been considered separate entities, and have been investigated individually. As development of water resources increases, it has been noted that development of either of the resources have impact on the quantity and quality of the other. Surface water commonly is hydraulically connected to ground water, but the interactions are difficult to observe and measure (Winter et al, 1998). The groundwater and surface water interactions are usually very complex, which is highly relevant with the local or regional climatology, landform, geology, and hydrogeological properties (Sophocleous, 2002).

The interaction between groundwater and streams generally takes place in three types: streams gain water from ground water through the streambed (gaining stream), streams infiltrate water to ground water (losing stream), or they can gain in some reaches and infiltrate in other reaches. For these hydraulically connected stream-aquifer systems, the resulting exchange flux is a function of the difference between the river water level and groundwater head. When the water table is high than the streams stage, steams obtaining water from the aquifer, conversely, when the water table is lower than the stream stage, streams losing water to aquifer. Sophocleous (2002) rank the most significant factors that affect the stream-aquifer interactions, including the streambed clogging, stream partial penetration, and aquifer heterogeneity. Streams commonly contain a silt layer in their beds that can reduce conductance between the stream and the aquifer (Guzha, 2008). Bank storage is also type of interaction that takes place in nearly all streams when a rapid rise in stream stage that causes water to move from the stream into the streambanks, which is usually caused by heavy precipitation, rapid snowmelt etc. Under the condition where an unsaturated zone exists between the streambed and water table, as shown in Figure2 D, the stream and groundwater may be hydraulic disconnected (Winter, 1998).

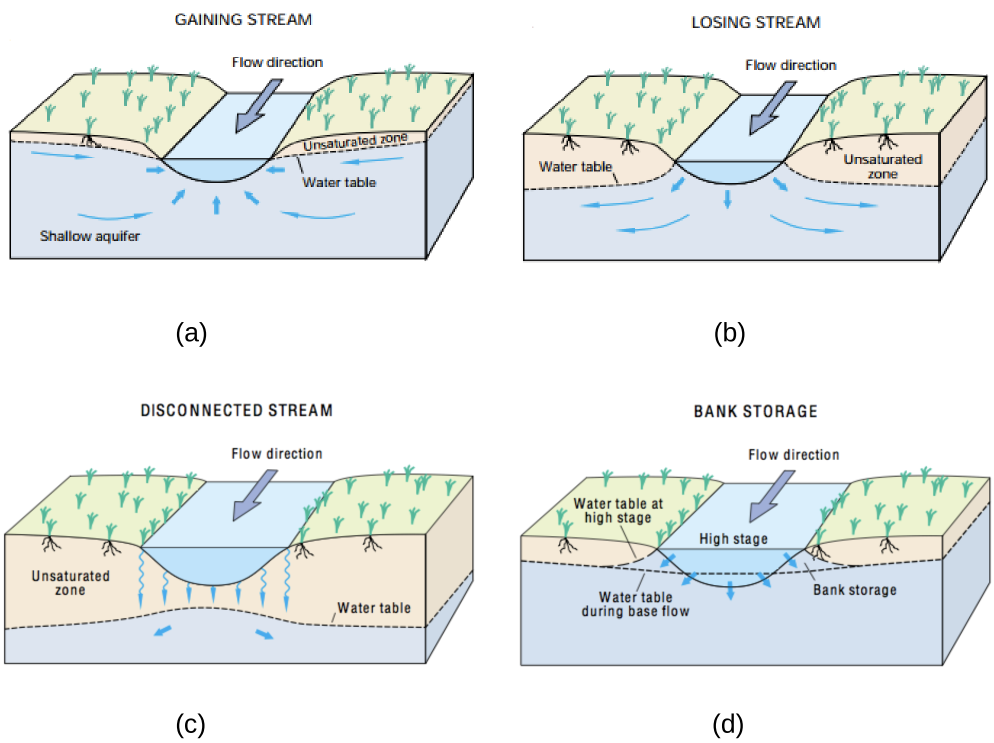


Figure 2 : Types of interaction between groundwater and stream (source: Winter et al 1998)

2.2 Model selection

The groundwater models are designed to represent the real groundwater system, which can be viewed as the best tool to conceptualize the complicated hydrogeological condition in groundwater basin (Himmelsbach, 2001). Application of groundwater models can help to improve understanding of the complicated aquifer conditions, and the results of modeling may have many practical purposes, including quantify overall cumulative groundwater budget, redefine and optimize the exploration drilling and field investigation, give useful information for the water management. In general, groundwater models have been used to solve four general types of problems, including the ground-water flow, heat flow, solute transport, and aquifer deformation. Among them, groundwater flow modeling have been proved able to estimate the regional steady-state flow in aquifer system, corresponding regional variations in hydraulic head induced by change in recharge or discharge of aquifer, and quantify the surface water and groundwater interaction. Specifically, the groundwater flow models solve for the distribution of heads, and solute transport models handle the concentration of solute, which is affected by the chemical reaction, dispersion and advection. Compared with analytical approaches, the numerical models have the advantage of being more reliable and flexible for describing the hydrogeological conditions, due to its ability to introduce the variability of properties of the groundwater system.

Groundwater flow models have been used for a long history and have many forms, and typically need the solution of partial equations. The Finite difference method and the finite element method are most frequently used for solving the groundwater equations in numerical models. In the early stages, finite difference method was widely accepted by the groundwater community due to its relatively simple concept and efficient computation. The finite element method became broadly used because of its ability to represent the irregular aquifer geometries more accurately (Pinder, 2003; Igboekwe & Achi, 2011).

In terms of the spatial dimensions, the groundwater model can be classified as two-dimensional areal (x-y), two dimensional profile (x-z), quasi three-dimensional (layered model) and three-dimensional (Thangarajan, 2004) . Whereas the two-dimensional areal and quasi three-dimensional models usually use the aquifer viewpoint, and the two-dimensional profile and three-dimensional ones mainly use the flow system viewpoint. Since the groundwater can flow both horizontally and laterally within the groundwater system, and the steams and groundwater exchange water both horizontally and vertically, flow dynamic are inherently three-dimensional. Therefore, simulation and analysis of the three-dimensional reality is necessary for getting better understanding of the variable aquifer properties and the stream-aquifer interaction.

MODFLOW is a three-dimensional block-centered finite-difference ground-water model that has been used for more than 30 years since the first version was published and has experienced a number of upgrade during this period, which made it become a versatile finite difference groundwater flow model that has been applied to abundant researches. It can simulate steady and non-steady flow in an irregularly shaped flow system in which aquifer layers can be confined, unconfined, or a convertible mode. The MODFLOW-2005 is the latest version, one of its major advantages is it organized into modular structure that allows it to be easily modified to adapt the code for a specific application.

The groundwater flow equation is the mathematical form that is used to describe the flow of groundwater through an aquifer (Anderson & Woessner, 1992). The governing equation is derived mathematically combining a water balance equation with Darcy's law. As a result, a matrix of equations could be formed to describe the flow through the individual model cells. The three-dimensional groundwater flow of constant density through porous medium can be described by the partial differential equations as follows (Harhaugh, 2005):

$$\frac{\partial}{\partial x}\left(K_{xx}\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_{yy}\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_{zz}\frac{\partial h}{\partial z}\right)+W=S_s\frac{\partial h}{\partial t} \qquad \text{(Equation 2.1)}$$

Where,

K_{xx}, K_{yy}, and K_{zz} are values of hydraulic conductivity along the x, y, and z coordinate axes (L/T);

h is the potentiometric head (L);

W is a volumetric flux per unit volume representing sources and/or sinks of water (T-1);

S_s is the specific storage of the porous material (L-1); and

T is time (T).

MODFLOW allows the comparison between simulated values, mainly contains two types: Head observation and Flow observation. The head observation compare the difference in head over time at a cell, whereas the flow observations describe a feature over the reach represented by the RIVER package.

Application of MODFLOW is often through a Graphical user interface (GUI). ModelMuse is such a GUI to generate 3-dimensional finite difference model and creates the flow and transport input files for, for example, PHAST and the input files for MODFLOW. It is designed to provide users with a graphical, interactive approach for building a groundwater model, and can uses objects (points, lines, and polygons) and formulas to define the spatial distribution of aquifer properties in a convenient fashion (Winston, 2009; Banta, 2011). Furthermore, the program is relatively easy to understand and freely public available.

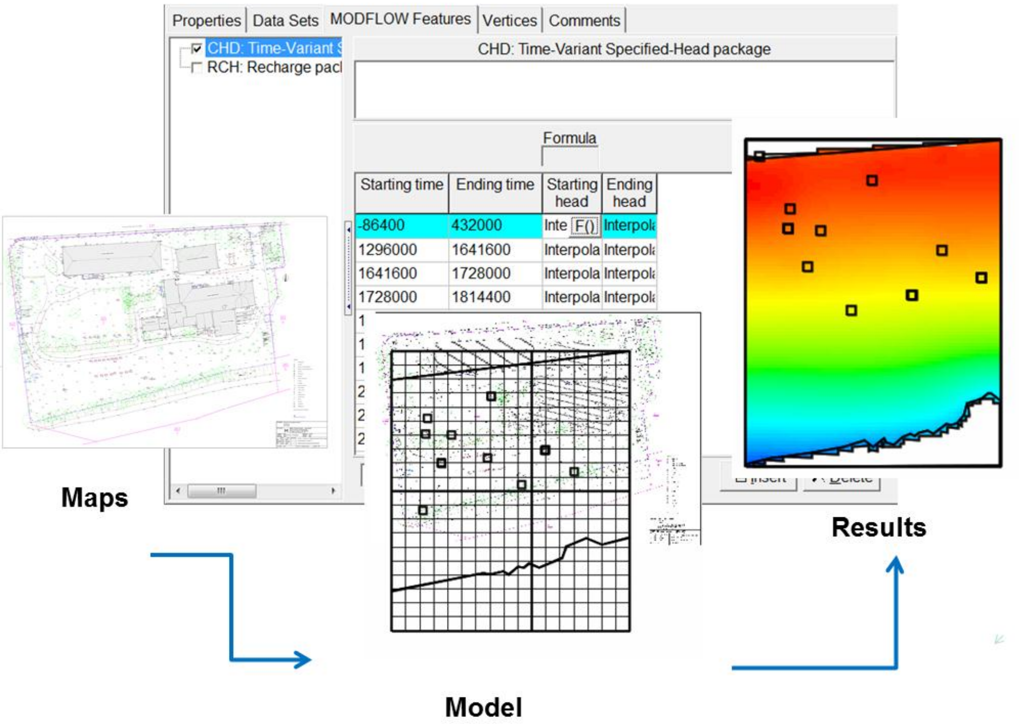


Figure 3: Model setup through Modelmuse

3 Data collection and evaluation

The test field of Pirna, occupies a large green area to the Elbe river bank, offers great opportunity for hydrogeological investigations at a test-field scale. During the past years, a number of field test have been applied for various study purposes. Several techniques are haven been applied at the site to investigate the aquifer. Data collected from various sources used in this research include:

Hydrogeological data: K information, Specific yield, Specific storage
Water level data: Groundwater table, Elbe river stage
Climate data: Precipitation

The figure below shows location of each investigation approach.

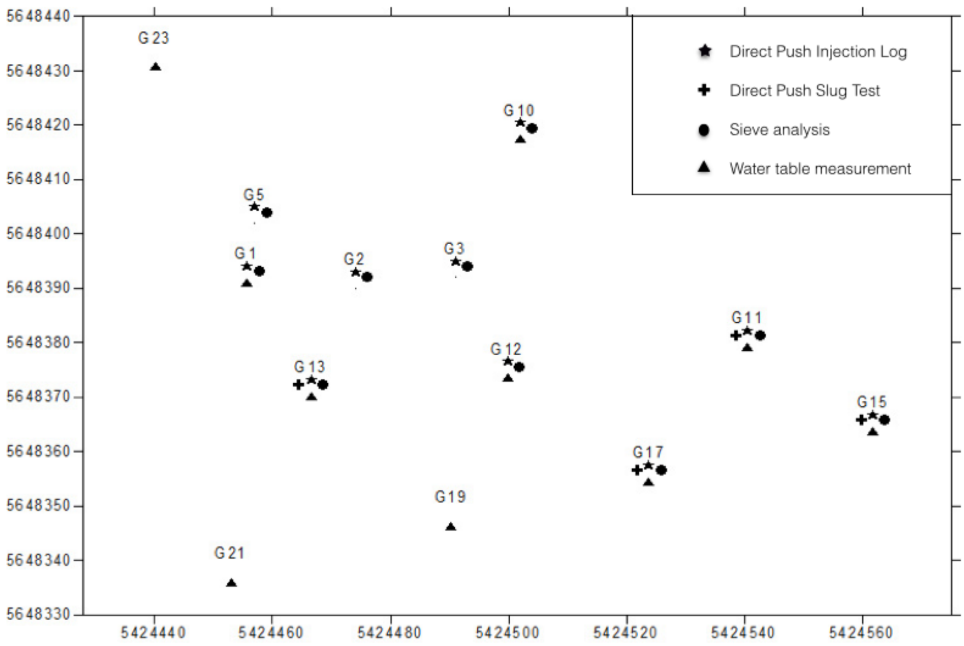


Figure 4: Locations of measurements

3.1 Hydrogeological data

3.1.1 DPST - Direct Injection Slug Test

Information of high spatial resolution of hydraulic conductivity is essential to characterize the groundwater dynamics at contaminated site or as input data of groundwater modeling. (Dietze & Dietrich, 2012) Single-well hydraulic test can serve as a useful approach in getting high-resolution K information. However, installation of wells normally need vast amount of funding. Therefore, it is usually difficult to obtain sufficient information at a site to evaluate influence of the spatial distribution in hydraulic conductivity on many hydrogeological issues (Sellwood, 2005).

During the past two decades, direct-push techniques have been widely applied for characterizing shallow unconsolidated formations. The direct push technology can rapidly advance small diameter tools into subsurface by using hydraulic ram supplemented with the weight of the vehicle or high frequency percussion hammers. Dietrich (2008) and other hydrogeologists have proved that the slug test combined with DP method can serve as an effective approach to get reliable and detailed information about K value in unconsolidated formations, and several modification have been made to improve the efficiency and accuracy. The major working steps of the Direct Push Slug Test involves development of the well screen, compressing the air in the hollow rods above water table, and then suddenly depressurizing and recording the head change with a pressure transducer. Afterwards, the recoding data could be transformed to absolute hydraulic conductivity using appropriate interpretation methods (Lessoff, 2010).

In the field of Pirna, in order to get detailed information of the vertical hydraulic conductivity distribution, a total number of 51 DPSTs were performed previously by Dietze and Dietrich at four locations (G11, G13, G15 and G17). DPSTs were performed using 1.5 inch diameter drilling rods, a Screen Point Groundwater sampler system with a length of 0.3m and a slug test kit. Each DPST consist of three test series of repeated measurements with input pressure given as 20, 40, and 60 mbar. The slug tests were measured from bottom to top due to the sand and gravel formation in the aquifer. At each test interval, the rod string and the screen were checked to avoid clogging of the screen. Depths of DPST are determined in intervals with constant Direct Push Injection Log results, which benefits the transformation of Kr values into K estimates. Due to the depths of water table to the land surface are different at these four locations, the measured depths of the tests are also different. The K value measured from the DPST at the four locations varies from 1.85E-5 m/s to 8.99E-3 m/s (Dietze & Dietrich, 2012).

3.1.2 DPIL-Direct Push Injection Log

The Direct push injection logger (DIPL), was developed to obtain rapid information about vertical variations in K in shallow unconsolidated settings, which uses a small diameter tool with a short screen (45mm in test site Pirna) that is attached to the lower end of a pipe string and advanced into the subsurface with hammer assisted DP technology. As the tool is advanced, water is continually injected through the screen at a relatively high rate to prevent clogging of the screen. Advancement is stopped at the desired depth intervals, water pressure in the injection tubing is measured at different injection rates with a pressure transducer and flow controller on the surface. Then a relative hydraulic conductivity can be calculated based on the injection rate, corresponding pressure and the system resistance correction parameters as show in equation 3.1. (Dietze & Dietrich, 2012)

$$K_{relative} = f(Q, \Delta P, S) \qquad \text{(Equation 3.1)}$$

Where，

Q = injection rate

ΔP = injection pressure

S – system parameter

The DPIL was performed on 10 different locations at the test site，at a depth from 0.3 m to 14.6 m below the land surface until reached the bed rock，with a depth interval of 0.3 m. At each depth interval, measurements with one high and one relatively low flow rates were performed for quality control, and resulting pressure was measured. Accordingly, two series of relative K value were generated, Kr1 and Kr2. Due to the hydrogeological properties is same at the respective depth, the resulting figure of the two series indicated a similar behavior. The Figure 5 bellow shows the DIPL sounding of at location G5 and G13. Variation of Kr at different depth interval can be observed which can provide valuable information of sediment formation. Due to the difference of the land surface elevation and the heterogeneities in the porous media, Kr at different locations vary to some extent but generally have the similar behavior. The low Kr took place at 4 to 7.5 m at G15 and 5 to 8.5 m at G13, where silt is contained. Afterwards the Kr rises to more than 1000 l/h*bar, which indicate the high permeability composited of sand and gravel. Kr variation at G15 generally shows 5 change intervals, that is, from 0 to 4 m, 4 to 7.5 m, 7.5 m to 9.75 m, 9.75 m to 11.89 m and 11.89 m to 13.1 m. Within each interval, Kr fluctuates in a small scope and has similar properties. Afterwards, by analyzing and summarizing the variation intervals, layer discretization could be determined at each location.

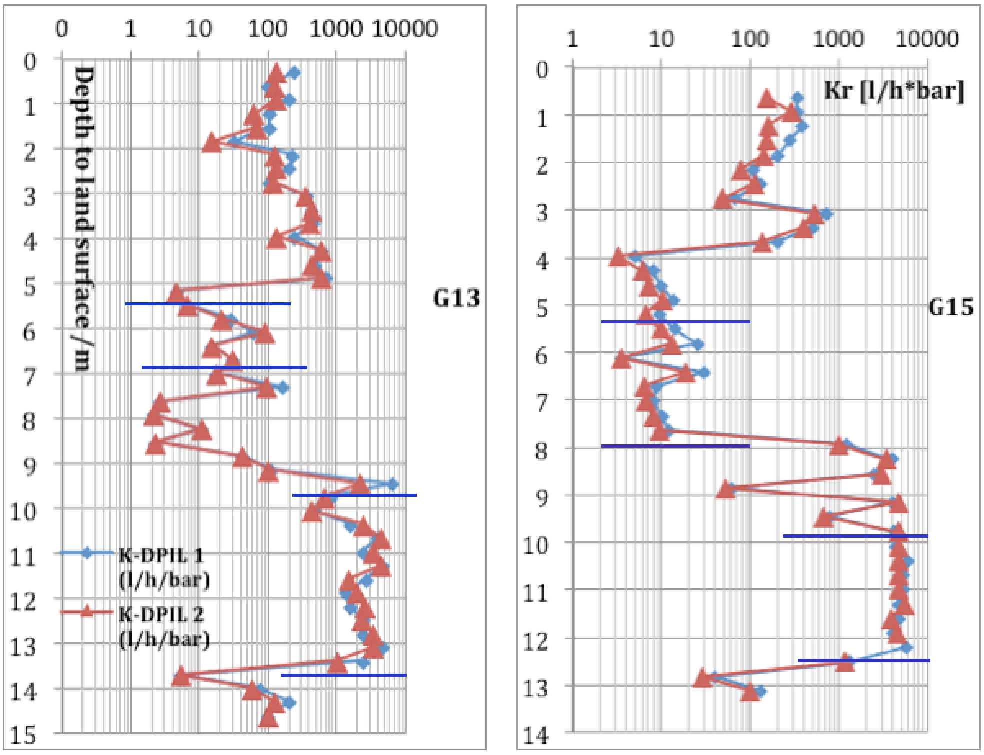


Figure 5：Kr distrbution in vertical direction at G13 and G15 (Source: Dietze & Dietrich 2012)

However, the K value, obtained from the DIPL is relative hydraulic conductivity, without calibration by appropriate methods, can just provide qualitatively subsurface facies. Dietrich et al. (2008) proved the ability of K values obtained from DPST to transform Kr values into semi-quantitative K through regression analysis. Thereby the results of slug tests obtained at four different locations were used for transformation (Dietze & Dietrich, 2012). Figure 6 shows the best agreement of calibration with a large R^2 equal to 0.92 indicates the strong correlation of the K_{DPST} and K_{DPIL}. The regression equation of K obtained from slug tests at four different locations and the corresponding DPIL relative K value is

$$y = 4.7E\text{-}05x^{5.1E\text{-}01} \qquad \text{(Equation3.2)}$$

For other locations where DPIL were performed without DPST measurements, this regression equation was used to transform the relative K value. Although transformation K value in such a simple method will result in errors, it still served as an effective way to obtain the detailed vertical absolute K value in at different locations.

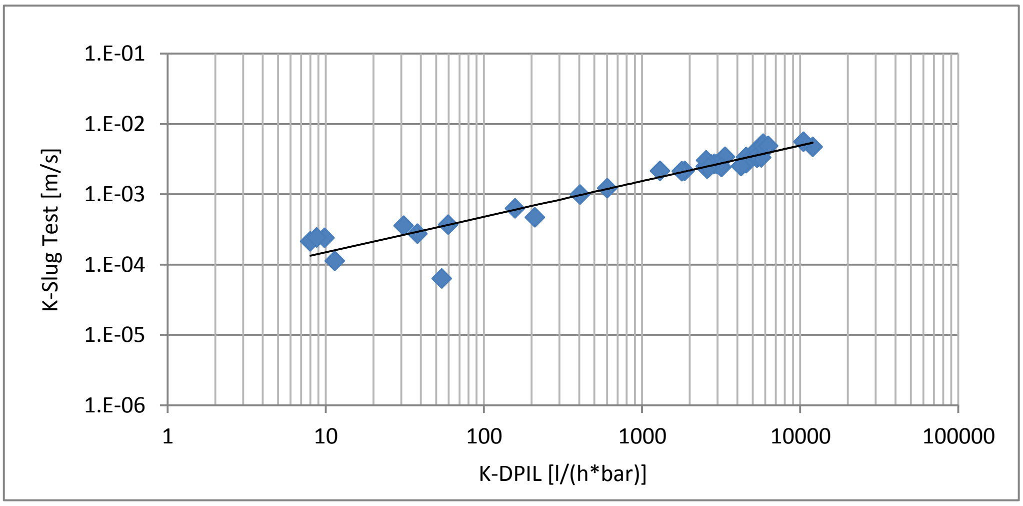


Figure 6：Correlation K value DPST-DPIL (source: Dietze & Dietrich 2012)

3.1.3 Sieve analysis and electric conductivity

Locations of sieve analysis where soil samples were taken shows in the Figure 4. Soil samples analyzed using sieve analysis were mainly for the unsaturated zone, where is less important to the model. Hydraulic conductivity distribution of all samples from 9 locations varies from 2.54E-7 m/s to 1.53E-2 m/s. While electric conductivity measurements were performed for a number of wells, there is, however, no effective approach to transform these results into the values that can be input into model. Therefore, the result of EC measurements can be used as a good qualitative indication of the sediments' properties.

3.2 Water level measurement

3.2.1 Groundwater table measurement

A network of wells with 1 inch diameter was installed in the field for various study purposes. Among them, groundwater table in 11 wells were measured. As show in the Figure 4. The water level was measured since November of 2008, with only two data sets. In the year of 2009, the data was collected nearly once a week from May to November. A total of 20 data sets were monitored from May to July in the year of 2010. Groundwater level was not measured in the next three years. Data became more sufficient in 2014, since June the water level was measured mostly on a daily basis. The model was built up using data of 2014, and data from 2009 and 2010 was used for model validation.

The ground table data is obtained in our test site using the electric tape. This instrument consists of a spool of dual conductor wire, with a probe attached to the end and an indicator. Once the probe lowered into well contact with the water, the circuit is completed and then an audible buzzer can be detected. The E-tape is marked every 5mm, thereby gives an accuracy of half a centimeter.

Figure 7: Instrument for water table measurement and 1 inch observation well at G1

According to the monitored data, the depth from water table to land surface varies from 4.8 m to 10 m at different locations. As the wells get closer to the river, a decreased thickness of unsaturated zone could be measured. Observed data gives information that groundwater level could vary up to 5 cm among the wells, the water level could rise up to 16 cm one day. Generally, groundwater flow directions are assessed according to the groundwater level elevation contour maps. Water level elevations are plotted on the base map and interpolation between monitoring points are made to get contours of equally distributed elevation intervals. Figure 8 shows the contour maps of the field site on the data 2nd of June, the northern region has higher water table than the southern, thus water comes from groundwater to the river which is in the south of this field.

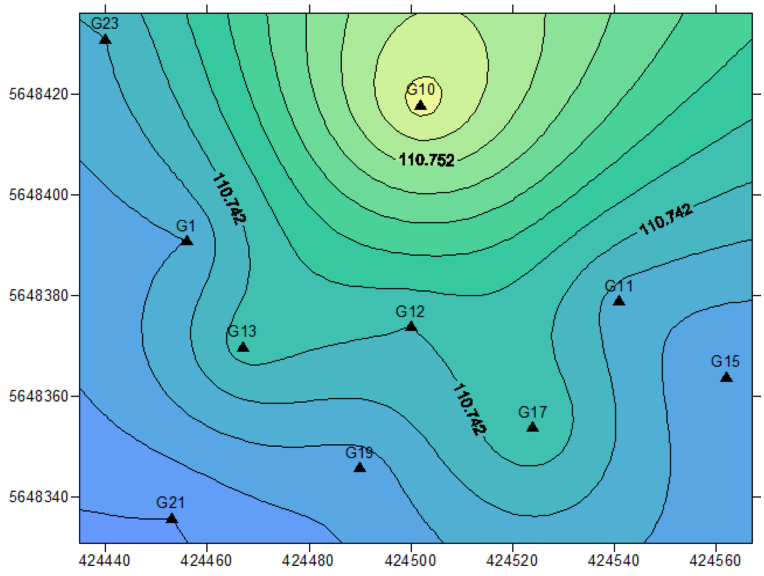


Figure 8：Contour map of the water table on the data 2nd of June

The precipitation events may have influence on the groundwater level, as shown in the Figure 9. Higher water table can be caused by dense and storm precipitation events, lower water table might due to rainfall scarcity and high rates evaporation. Nevertheless, the amount of precipitation and the duration it need to infiltrate into the aquifer is determined by a number of factors. As can be seen from the Figure 9, while water table may differ a few centimeters at different locations, the water table among these wells exhibits an identical variation trend.

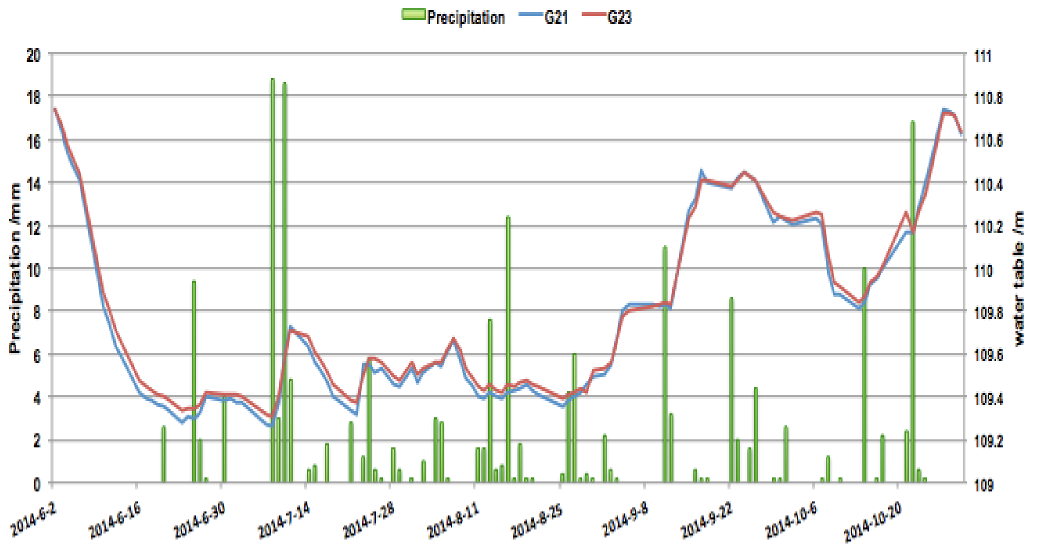


Figure 9：Water table fluctuation at two locations and precipitation variation

3.2.2 Elbe river stage

The water level data of Elbe river at Pirna station is quite sufficient due to it was monitored every 15 minutes, which is 96 data in total for each day. The data is public available on the website, which is provided by the German Federal Waterways and Shipping Administration (WSA). Average value of the 96 data was used as the daily water level and also used to define the specific-head boundary. The measuring point is close to the WSA, which is approximately 500m upstream from our filed site. The hydraulic gradient was estimated based on the river stage and distance among three contiguous stations along the Elbe river, the calculated gradient is $2.5E10^{-4}$ m/m. Therefore, the actual water stage along the site is approximately 10cm less than the measured data.

The groundwater observation indicates a strong and rapid response of fluctuation of the river stage. Compared with groundwater table, river stage shows slightly faster variation in relatively larger range. Precipitation can have a direct impact on river stage, but its impact on water table is more or less delayed.

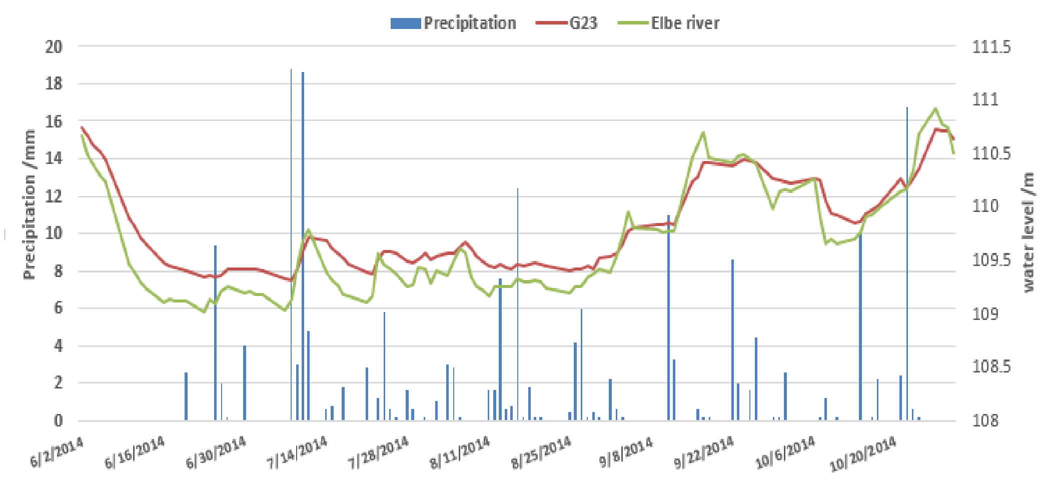


Figure 10: Water table and river stage fluctuation, and relationship to daily precipitation.

4 Conceptual model

The conceptual model summarizes information of the entire system in a physical visualization, which can be subsequently interpreted in mathematical models by using the appropriate parameters and equations (Reimann, 2012). The model simulates the groundwater system by introducing important physical features that affects the flow, and by introducing hydraulic properties that represent the porous media. The groundwater flow model was mainly constructed to provide an understanding of hydraulic head distribution and water budgets. The general conception of the model is that the water flows into the system as recharge from precipitation and flow out of the system as stream flow. Simplification is essential due to its not feasible to get a complete construction of the real system. The basic conceptual model should properly describe the site including the boundary of the domain to be modeled, sources and sinks, discretization of layers with sediments properties. Anderson and Woessner (2002) indicated the three steps in formulating a conceptual model, which involves defining hydrostratigraphic units, preparing the water budget and defining the flow system.

The simulation period was defined as one steady state continued with 131 transient stress periods. Time step was set at daily, which is 151 time steps in total. The geological boundary of the modeled area was determined by the field site map and morphology of the close-by Elbe river. A network of finite difference grid was then built cover an area of 2.52 ha with uniformly distributed grid spacing of 2 m by 2 m. A total of 70 rows, 90 columns, five layers and 31500 cells were generated to cover the whole area. The model unit is set as meter for Length and second for Time.

The model was simulated for the year of 2014 from 2nd of June to 30th of October where groundwater level was regularly observed.

4.1 Initial conditions

Initial conditions refer to the head distribution over the entire model domain at the beginning of the simulation and can be seen as a boundary condition in time. It plays a significant role in modeling transient flow problems. These initial heads act as the starting distribution for the numerical calculations, which accelerates the convergence for transient problems. A steady-state stress period can be used to determine the initial head distribution for the subsequent transient stress periods (Markstrom 2008, Anderson 1992). In MODFLOW, the initial head is always used as the initial estimate. In general, the initial estimate were supposed to have no effect on the solution to the steady-state flow equation, but it may influence the number of iterations steps needed to obtain the solution (Harhaugh, 2005).

Generally, measured head distribution is used for initial conditions. Field groundwater levels from the first day of regular measurement in the year of 2014 were used to describe the initial heads, and were defined to be identical throughout the entire five layers. Due to the fact there are 11 observation wells in the study area, head information of the rest areas can be obtained by interpolation methods. With application of computer software Surfer, an interpolated water table of appropriate data setting for each cell could be generated using the Kriging method. Kriging is a geostatistical interpolation method that is frequently used to estimate the water levels and transmissivity. It is often chosen for the reason that it is the best linear unbiased estimator which using the adjacent known values to make estimates and it considers the spatial structure of the variable and also provides the confidence of interval of estimation. (Thangarajan, 2004)

4.2 Boundary conditions

Boundary conditions in the model define the locations and the manner that water flow in and out of the active model domain. The natural hydrogeological boundaries should be used as the boundaries of the model if possible (Anderson & Woessner, 1992). However, in practice, it is usually impossible to locate the boundaries of the model at the actual physical boundaries of the entire system. Some researches, such as information acquisition of field site, investigation of contaminant transportation, and delineation of recharge areas for pumping wells, are focused on a relatively small parts of an aquifer system, which is usually too small to incorporate distant aquifer boundaries. (Leake & Lawson, 1998; Igboekwe & Achi, 2011) .

Correct selection of boundary condition is very important for model design. Hydrogeological boundaries is classified into three types of mathematical conditions:

Table 1: Types of hydrogeological boundary

Boundary type and name		*FormalName*	*Mathematical designation*
Type 1	Specified head	Dirichlet	h(x,y,z,t)=constant
Type 2	Specified flow	Neumann	dh(x,y,z,t)/dn=constant
Type 3	Head-dependent flow	Cauchy	dh/dn+ch =constant (where c is constant)

When the boundary is a river, the head along the boundary will vary spatially. Generally, specific head conditions are preferred than specific flow conditions, due to head measurements are usually easier than flow measurement and it is helpful in achieving calibration. Under some situations, such as the head has possibility to change during the simulation while the flows into the system can be constant, then the selection of specific flow may be advisable. The specific flow condition is either simulated as recharge well or a pumping well at a specific rate. Under Head-dependent flow boundary conditions, flux is calculated for variable head and assigned accordingly. The RIVER package in Modelmuse is designed as the third type condition. In a head dependent boundary, the model calculates the difference in head between the river node and corresponding model node where boundary is defined. Then, the head difference is multiply by the specified conductance to calculate the water flux that flows into or out of the aquifer. RIVER boundary is advisable to use when the river partially penetrates a layer and can either gain water or infiltrate water to the aquifer (Thangarajan, 2004; Winston, 2009).

In our case, the daily river stage is known, whereas the conductance is unknown. Therefore, specific-head boundary conditions were set for both the upper boundary and Elbe river. Beside this, groundwater levels at the site react very fast on changes in the river stages and therefore, show a direct connection of the river to the groundwater. It is assumed that there is 4cm difference of the highest and lowest river stages in the model domain, in order to improve the accuracy of simulation, the river stages are input using a linear interpolation function.

The Northern boundary is set in the Northern part of the model domain pass through location of G23 and approximately parallel with the position of Elbe river as shown in Figure 11. The reason to set boundary using data of G23 is because it located in the northernmost part among the wells and there is no other field test applied in this well during the simulation period. Initially, data from the official observation well Obs 50496161, where daily data is available, were considered to set as the boundary head. Whereas the distance from the field to the well is too far away, and the hydrogeological condition of subsurface varies spatially, it is arbitrary to use the value just after simple interpolation.

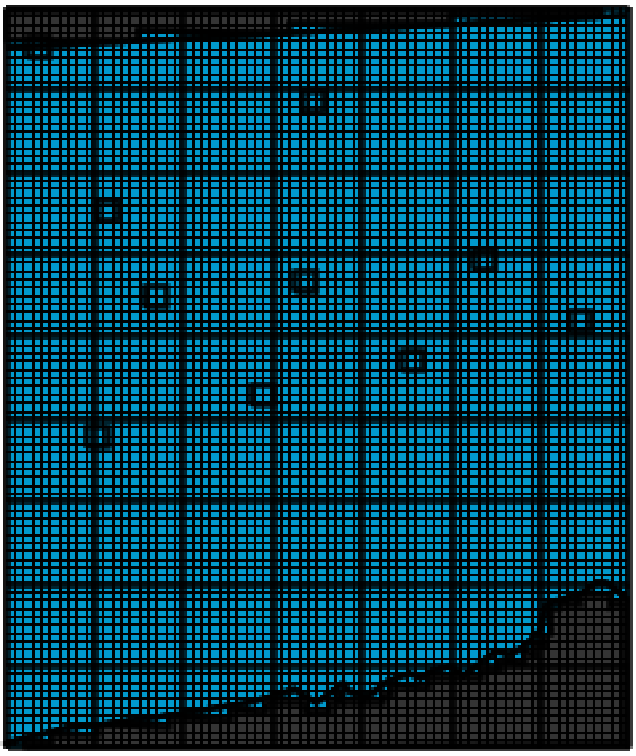


Figure 11: Location of boundaries and inactive cells in the model

As shown in Figure 11, the blue cells are active cells and grey cells are inactive cells. Inactive cells are assigned in the outer part of the boundary conditions, which indicate the type of cells where no flow into or out of the cell occurs during the entire simulation，thereby flows from or into the cell and hydraulic heads are not calculated. Relatively Kz value was set to the uppermost 3 layers of river boundary and the Northern boundary in the same position, in order to accommodate execution of the model. Water level of Elbe river and groundwater mainly distributed in the second or third layer in the model. Due to the fluctuation of river stage and water table at boundary, the uppermost layers have the risk to be dry. By assigning a high Kz value equal to 1, head distribution in the uppermost layers become sensitive to variation of water level and might result in a higher range of variation, and then can distribute the resulting head to horizontal direction.

4.3 Recharge

Recharge is the water that flows into the saturated zone, varies spatially and temporally and is affected by a variety of factors, making it very difficult to quantify in the water budget (Dripps & Bradbury, 2007; Bradley, 1996). In Modelmuse, the Recharge (RCH) Package is used to add terms to the flow equation to calculate areal recharge to the ground-water system within the model domain (Harhaugh, 2005).

Recharge from precipitation is usually a major source of water to the groundwater system. This amount of water flow into the aquifer is calculated based on the formula:

$$Pr=\alpha P \qquad \text{(Equation 4.1)}$$

Where α is the infiltration coefficient

This coefficient is affect by a number of factors, such as the vegetated land, soil moisture, hydraulic conductivity of sediments, porosity etc.

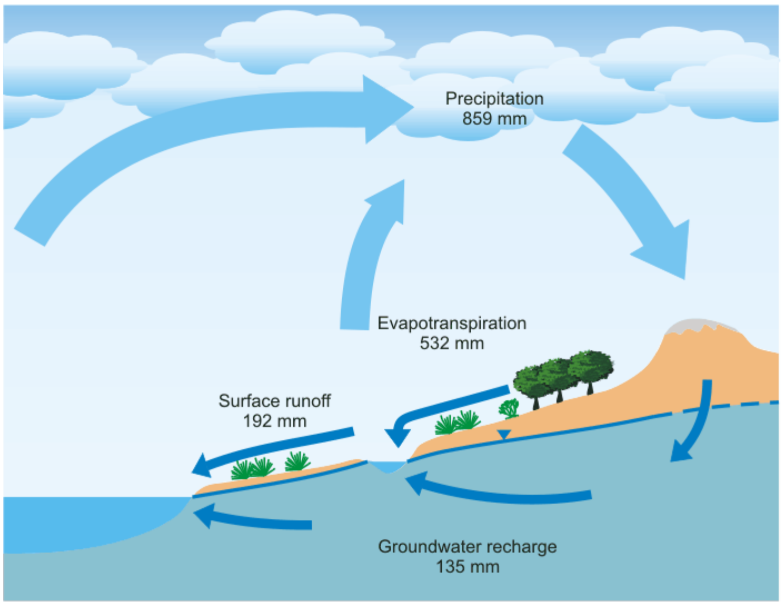


Figure 12：Water cycle with data of the average water balance for Germany 1961 - 1990 (Source: BGR)

Figure 12 gives information of water cycle of average water balance for Germany in a general and simple way, in which the groundwater recharge accounts for nearly 16% and the evaporation accounts for roughly 62%. In consideration of the aquifer in this field is shallow and has a medium permeability, the infiltration coefficient is assumed to be 0.2 in initial model. Recharge obtained from precipitation is set uniformly distributed over the model domain for the active uppermost layers, in the case of cells in first layer are dry. Recharge rate was estimated under two scenarios:

One is to calculate an average value for the entire model period. Precipitation accumulates 267 mm during the model period, therefore, the recharge rate can be calculated as 4.0E-9 m/s. The other approach is to use the daily precipitation data multiply by the infiltration coefficient, which could reflect fluctuation of water table response to the daily rainfall. However, water need time infiltrating into the aquifer, and duration is unknown in our field, therefore we assumed that water can infiltrate into the aquifer within one day. Small rainfall events between long no-rainfall days are neglected, due to this amount of water can be either intercept by plants or evaporated into air. In the calibration stage, recharge values can be adjusted if simulated water table levels are too high or too low. Both of the two approaches are simplified approaches to assume the recharge, which are required to be improved in the future work.

4.4 Discharge

Discharge of an aquifer normally occurs in the form of surface runoff, groundwater abstraction, and evaporation and river baseflow. The field is surrounded by meadow and dense residential area, therefore we assumed there is no abstraction well. The possible pumping wells are quite far away from the field site thus could not have much effect and beside this, the pumping information is unknown. Therefore, the groundwater abstraction is out of our consideration. Discharge caused by evaporation is assumed by subtracting this part from precipitation in the way of giving a low recharge coefficient. Under natural conditions, due to the hydraulic head of aquifer is usually higher than the river, the river gaining water from the aquifer. However, the river could also losing water to the aquifer during flood seasons when the water level is higher than the hydraulic head of the aquifer. To assess the exchange between groundwater and the adjacent river is one of main task of this study.

4.5 Model grid and layers discretization

In finite difference models, the modeled area is divided into square or rectangular grid regions called cells. In general, size of finite-difference cells is based on the spatial resolution required for providing model results that can properly describe the purpose of building the model (Markstrom 2008). Determination of time step and construction of grid are vital steps in setting up a model because the precision of space and time discretization can strongly influence the modeling results. Theoretically, small cell size and time steps is desirable for the reason that the numerical representation could better approximates the partial difference equation (Anderson & Woessner, 1992).

Within each cell there is a node at which head and concentration are calculated. The network of cells and nodes constitutes the grid. The grid is required to be set superimposed on the map of the area to be modeled.

In our test site, the model has a rectangular geometry with the grid network cover an area of 140m by 180m. A total of 70 columns and 90 rows were generated to cover the area composites of 6300 cells in each layer. The model cell is 2m by

2m and evenly spaced throughout the model domain. The fine cells are required for properly including small-scale tracer tests into the model, which is used for evaluating the flow transport process.

In Modflow, the layers can either be set in a uniform form or have variable thickness among the cells. Five hydrogeological layers were defined to represent the stratigraphy, which is obtained by analyzing the variation of hydraulic conductivity based on Direct Push Injection Logging measurements. Elevations of the top and bottom of individual layers were specified on the basis of surface topography and stratigraphic data. MODFLOW uses a structured grid which requires each layer to be continuously set over the entire model domain. Under the situation where the vertical flow component has a strong effect on the model result, it is important to have fine grid in the vertical direction in order to decrease the error in head(Philip Brunner, 2010).

Discretization of layer is based on the variation of K in vertical direction. By summarizing and analyzing the DIPL results, five variation interval could be identified at each wells where DPIL were performed. The variation interval is used to identify hydrostratigraphic units as well as to define layer thicknesses. With known depth of the variation interval and land surface elevation, herein we can get the depth of each interval to the land surface. By subtract these depths from the known surface elevation (m.a.s.l) at each location, top and bottom elevations of each layer are obtained. Thereafter, elevation of each layer is imported as point data layer-by-layer, and interpolated using the build-in triangulation interpolation to distribute the spatial discretization of layer.

Table 2：Elevation of land surface and bottom elevation of each layer (unit: m.a.s.l)

Layer	G1	G2	G5	G10	G11	G13	G15	G17
Land surface	119.478	119	119.34	119.277	116.939	116.963	116.076	115.709
1st layer	112.468	111.69	112.03	112.527	112.369	111.783	112.116	110.829
2nd layer	108.808	108.03	108.67	109.027	108.409	109.343	108.156	108.089
3rd layer	107.288	106.81	107.15	107.527	105.939	107.513	106.326	105.649
4th layer	105.768	105.59	105.93	106.027	104.449	103.253	103.886	104.129
5th layer	105.628	104.98	105.32	105.427	103.529	102.333	102.976	102.299

4.6 Time discretization

MODLOW allows the option of discretizing the simulation period into blocks of time of various lengths known as stress period. The use of stress periods offers the opportunity of changing some of the parameters or stresses in boundary condition packages (Anderson & Woessner, 1992). Stress periods can be set either as steady state or transient mode. Stress period of steady state take no account of changes in storage, whereas a transient stress period considers influence of changes in storage when calculating groundwater heads (Markstrom 2008). A combination of steady state and transient stress period can be applied; therefore, a simulation can start with a steady-state stress period and continue with transient stress periods (Harhaugh, 2005; Igboekwe & Achi, 2011).

Time steps are used to refine the stress period into smaller increments that may facilitate solution convergence or to improve accuracy of calculating groundwater heads distribution. The accuracy of solution can be examined by increasing the number of time steps for a stress period until ground-water heads at the end of the stress period become constant with increasing time steps(Markstrom 2008). Iterative solution methods are used to solve for the heads for each time step.

Time step is set on a daily basis in our model, due to the groundwater level was monitored generally on a daily basis，which is 151 time steps in total. Simulation of model using daily data can reflect the importance of the accuracy and timing of field data of precipitation events (Bradley, 1996; Dietze & Dietrich, 2012).

4.7 Data input

4.7.1 Hydraulic conductivity

Formation of aquifer in this field consists of interbedded strata of medium sand to coarse gravel that are followed by fine sands with changing thicknesses. The basement consists of marine sediment rocks formed of sand- and mudstone of Upper Cretaceous age (Dietze & Dietrich, 2012). Based on the result of previous DPIL and DPST measurements, the transformed K value varies from 1.2E-4 to 3.8E-3 m/s. Table 3 shows the transformed K value for each layer, which are averaged value of each previously defined variation interval (Figure 5) based on DPIL measurements. However, there are only 8 points with available data in this field, in order to get the overall information of the model domain, interpolation approaches were applied to get overall information.

Table 3：Average K value at different locations in each layer (unit: m/s)

Layer	G1	G2	G5	G10	G11	G13	G15	G17
1	7.9E-04	8.1E-04	7.7E-04	6.7E-04	8.0E-04	7.8E-04	7.6E-04	5.4E-04
2	4.9E-04	7.9E-04	1.5E-04	6.7E-04	1.2E-04	2.9E-04	1.5E-04	2.6E-04
3	6.5E-04	2.6E-03	2.7E-03	2.2E-04	1.3E-03	2.5E-04	2.4E-03	3.8E-03
4	2.7E-03	3.1E-03	3.2E-03	2.6E-03	3.4E-03	2.5E-03	3.4E-03	3.8E-03
5	1.5E-03	1.6E-03	9.4E-04	1.6E-03	2.3E-04	4.4E-04	2.1E-03	1.7E-04

Two possible interpolation approaches is considered, one is to interpolate data layer by layer，the other is 3D interpolation for the entire domain. 3D interpolation was not used because its complexity and the demanding user efforts. Furthermore, for most groundwater models that handle simple geological structures, complex interpolation methods are not required (Walther, 2012). Therefore, data of each layer is pre-processed with Surfer using the Kriging method before imported as points into Modelmuse. In surfer, user can define the grid line geometry by adjusting the coordinate, grid spacing and number of lines. Consequently, a set of data could be generated that matches the grid network in the model. Thereby K value could be given to the each corresponding cell, after importing the data sets layer by layer, the K value is defined for the entire model domain. Figure 13 shows the K value distribution of the 3rd layer and bottom layer. Compared with other layers, these two layers shows more obvious difference in hydraulic conductivity, range from 6.5E-4 m/s to 3.8E-3 m/s.

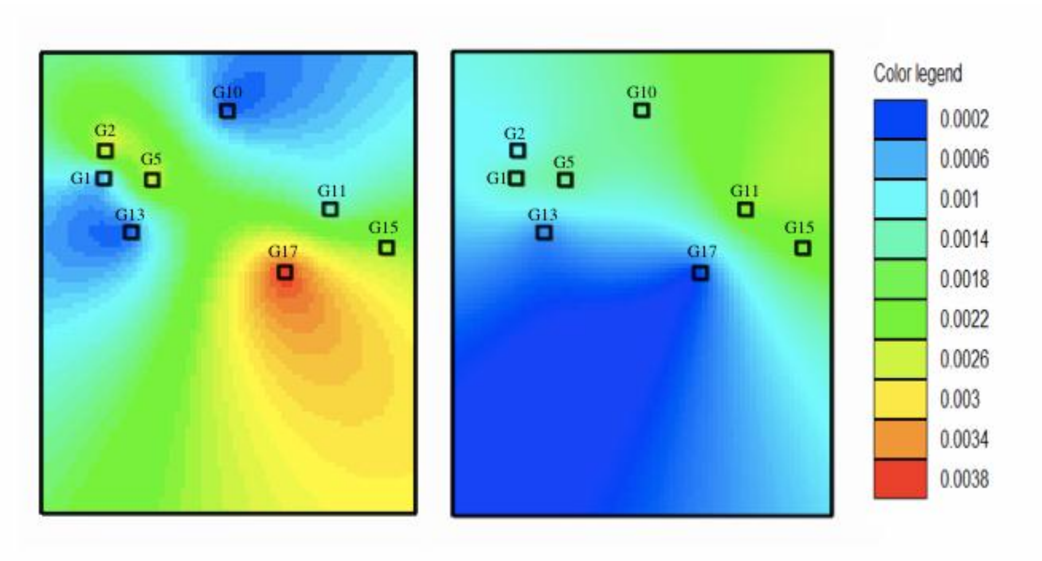


Figure 13: Hydraulic conductivity distribution in the middle and bottom layer using Kriging interpolation (unit: m/s)

Assignment of either the initial head value or the K value through interpolation is an effective and feasible method, however, the inherent limitation of each types of interpretation has to be aware. As can be seen form the Figure 13, the points with data are mainly distributed in the central domain, but there is no data for the Southern and Northeastern region. K value distribution in these large areas are only controlled a single or two points, which will definitely lead to errors and the data generated for these parts are less reliable. Linear interpolation is the simplest and direct interpolation method usually used for low-density data. However, linear triangulation method in Surfer 9 can just interpolate the area where data is available, no data can be generated beyond the area. Density and locations of the points are vital for the accuracy of interpolation. Further investigations could be considered at the regions where data is scarce.

4.7.2 Specific storage and specific yield

Specific yield and specific storage set for the model in the value 0.2 and 1E-5, respectively, since the aquifer is unconfined. Fetter (1980) listed the specific yield for different sediment textures, specific storage usually falls in the range of 1E-3 to 1E-5.

Table 4: Specific Yields in Percent (Source: Fetter 1980)

Material	Maximum	Minimum	Average
Clay	5	0	2
Sandy clay	12	3	7
Silt	19	3	18
Fine sand	28	10	21
Medium sand	32	15	26
Coarse sand	35	20	27
Gravelly sand	35	20	25
Fine gravel	35	21	25
Medium gravel	26	13	23
Coarse gravel	26	12	22

4.8 Dry Cells

Occurrence of dry cells is one of the most frustrating problems when running a MODFLOW model. The reason for a cell become dry is the calculated head in the cell is below the bottom of cell. Under this situation, MODFLOW automatically change the status of the cell into inactive. One of the major problems associate with a cells become dry and inactive is the external water could not be received in the cells. In some cases, coarse vertical discretization of aquifer is used to avoid the dry cells (Philip Brunner, 2010).

It is possible to rewet a cell Modflow-2005 when the water head of the neighboring cells is higher than the threshold defined by users. However, application of rewetting could leads to numerical instability (Doherty, 2001).

By activate Wetting in MODFLOW Options, the dry cell will be rewetted using appropriate wetting factor and, equation. In our model, water table always below the bottom of the first layer and for certain stress periods，below second layer bottom. While the first layer is the unsaturated zone in general, occurrence of dry cells will not affect the operation of model and modeling result. Whereas rewet of the second layer is necessary because the water level fluctuates, when the water level rises above the bottom of this layer, head is required to be calculated, otherwise will lead to error in the modeling. If head of the wetted cell falls below bottom of layer again, the cell will immediately became dry. It is possible for cells to cycle between wet and dry (Reimann, 2012; Sophocleous, 2002). Below is the equation used in our model for rewetting:

$$h = BOT+WETFCT(THRESH) \qquad \text{(Equation 4.2)}$$

Where，

h = the head in the newly wetted cell

BOT is the elevation of the bottom of the newly wetted cell

WETFCT is a user specified wetting factor, range between 0 and 1

4.9 Summary of Pirna MODFLOW Model

Table 5: Summary of Pirna MODFLOW Model

Number	*Item*	*Details*
1	MODFLOW Version	MODFLOW-2005
2	GUI	ModelMuse
3	Grid cell size	2m
4	Rows	90
5	Columns	70
6	Layers	5
7	Cells per layer	6300
8	Active cells per layer	5652
9	Maximum Elevation	119.478
10	Minimum Elevation	102.333
11	Model Simulation Type	Transient
12	Stress Periods	109
13	Time step	152
14	Stress period duration	4 months
15	Length of Simulation	152 days
16	Programme control package	Basic (BAS)
17	Solver Package	PCG
18	Internal Flow Package	Layer Property Flow (LPF)
19	Boundary Package	CHD, RCH
20	Observations	Head observations

5 Calibration and Validation

5.1 Calibration of model

The aim of the calibration is to obtain a reasonable fit between the observed data and model simulation by adjusting the parameters such as the K value, specific yield, specific storage etc. Repeated data adjustments and software excursion is needed until an acceptable match between the observed and simulated groundwater level is achieved. The manually trial and error procedure are commonly used approach for calibration, which are very time consuming. Furthermore, we could not get information on the degree of uncertainty in the final parameter selection and could not ensure the statistically best solution (Anderson & Woessner, 1992). Calibration techniques such as UCODE and PEST also can be applied, however, the inherent limitations in the program need to be understood and accommodated (Reimann, 2012).

Determination of appropriate criterion is usually depend on the objective of model simulation and the available data to compare the model output variables, which might be the water levels, mass balance and flow rates. In our case, the calibration process is carried out based on the comparison of observed groundwater head and calculated groundwater head. The criteria to decide whether the match is acceptable are subjective, despite the goal of minimizing the difference of the observed value and computed heads. Statistic approach of Mean Absolute Error (MA) and Root Mean Square error (RMS), and NormalizedRMS error (NRMS) are commonly used to check the result of calibrated model.

The MA error is the mean of absolute value of the measured and simulated heads, which measures the average absolute residual value defined by the equation below:

$$\text{Mean absolute error} = \frac{1}{n}\sum_{i=1}^{n}\left|\left(h_m - h_s\right)_i\right| \qquad \text{(Equation 5.1)}$$

The RMS error is the average of the squared difference in measured and calculated heads， defined to be the square root of the arithmetic mean of the squares of the original values :

$$\text{Root-mean-squared error} = \left[\frac{1}{n}\sum_{i=1}^{n}\left(h_m - h_s\right)^2_i\right]^{1/2} \qquad \text{(Equation 5.2)}$$

NormalizedRMS error is expressed as a percentage and is a more representative measure of the goodness of fit than the standard RMS as it accounts for the scale of the potential range of the data values.

$$NormalizedRMS = \frac{RMS}{(h_m)_{max} - (h_m)_{min}} \qquad \text{(Equation 5.3)}$$

where,

h_m is the measured head,
h_s is the simulated head.

Calibration of the model is confirmed using RMS error, MA error, NRMS error and water budget percent discrepancy through adjustment of hydraulic properties and recharge rate. In general, the errors in the model are randomly distributed among the selected observation wells. By application of HOB package, a table of the observed head, simulated head and absolute difference are generated after each model execution for each observation well and stress period. Because of the groundwater level, Eble river stage and precipitation vary with the time, the head distribution also differs with stress period. The simulated heads are computed by interpolating from the nearest cells to the position of the observation. Head observations can extend through specified layers, thereby cells of layers incorporated of the observation can be used for computing the simulated head.

Based on the data obtained from water table measurement in the field site, data of 5 wells (G10, G11, G12, G13, G21) with measurement dates in accordance with G23 were selected for the Head Observation.

Due to the K value was input as points data for each layer, adjustment of K value is achieved by percentage change of K value. The model was run several times with K value increment of 5 to 20 percent and decrement of 5 to 20 percent. After each execution of model, the head differences of observed head and simulated at each time step vary to some extent, some results are improved while others worsens from a millimeters to a few centimeters. Thereafter, determination of the good agreement is a function of the RMS error. Table 6 shows the percentage change of K value in each layer which result in the good agreement of observed and simulated head.

Table 6: Percentage change of K value in calibrated model

Layer	Variation in percent (%)
1	100
2	95
3	80
4	110
5	120

Calibration of recharge is quite subjective due to data insufficiency. By adjustment of infiltration coefficient, from 20% to 25%, which is recharge rate change from 4E-9 to 5E-9，a slightly better agreement is achieved. Under both of the two scenarios, using average annual precipitation or using corresponding daily precipitation, a few millimeters to a few centimeters change of the head difference can be observed whereas the calculated RMS errors are almost identical.

A scatter plot of observed and simulated head is a simple and straightforward approaches to show the results of calibrated fit. Fairly good visual comparison between the measured and calculated heads can be observed at each observation well. As shown in Figure 14, observed head and simulated head are distributed along the line, simulated head are generally smaller than the observed head. Large deviation might be caused by the measurement gap on the weekends that water levels may have a relatively large variation or some measurement errors.

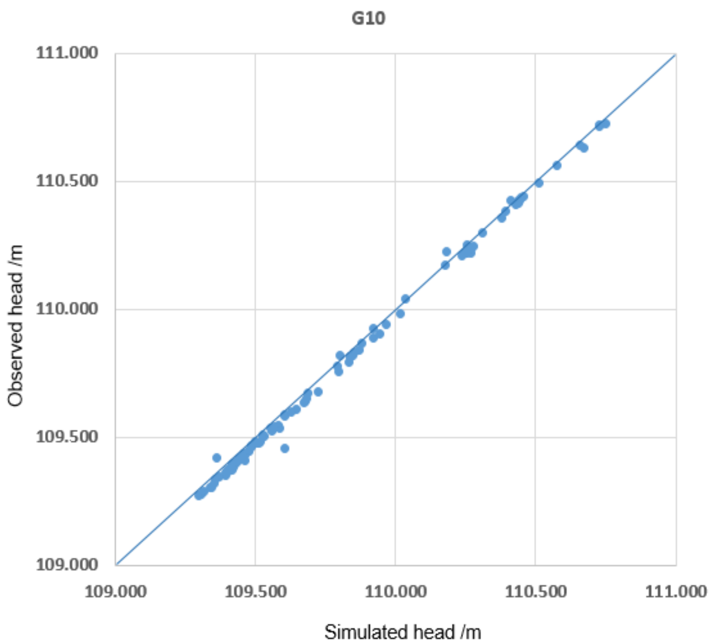


Figure 14: Scatter plot of observed head versus simulated head at G10

The calibration criteria of RMS error and MAE in this model are controlled less than 10.6 cm and 9.6 cm over the entire stress period. The Normalized RMS error is less than 7.2% within the 10% norm. Among them, observation well of G10 has the best fit while the G21 has the worst. The possible reason could be the description of hydrogeological conditions at G10 in the model could well fit the real condition, whereas the hydraulic conductivity at G21 are interpolated which is more deviate from the actual conditions.

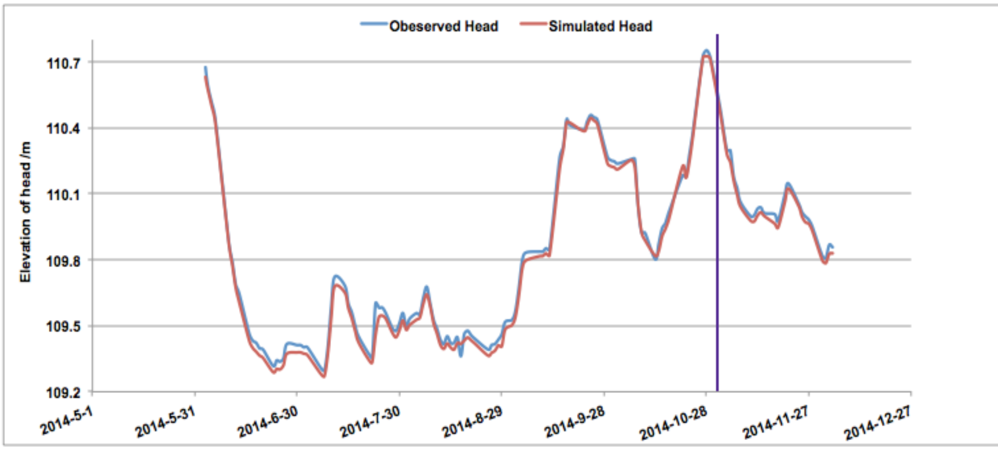
Table 7: Calculated MA error, RMS error and Normalized RMS error at different locations

Location	RMS /m	NRMS	MA /m
G10	0.031	2.1E-02	0.027
G11	0.264	1.8E-01	0.080
G13	0.078	5.3E-02	0.071
G12	0.073	4.9E-02	0.067
G21	0.106	7.2E-02	0.096

5.2 Validation of model

Simple validation of model performed using data from 2009, 2010 and 2014. Due to water table measurements were randomly taken in 2009 relative with a big time gaps, the model fails to converge thus could not be used for validation. Part of the measurements in 2010 was performed on a daily basis, a reasonable fit of the head difference is achieved with RMS error range from 0.06 m to 0.28 m, and MA error range from 0.05 m to 0.25 m at different observation points.

The model was set up using data from June to October in 2014, data of the following 34 days were used for model validation. An excellent fit can be observed for this time period, which indicates the model is capable of simulating the field in following days with measurements as data is used for G23 to define the Northern boundary. Figure 15 below shows the observed and simulated head difference of both the modeled period and validated period at G10 and G21.



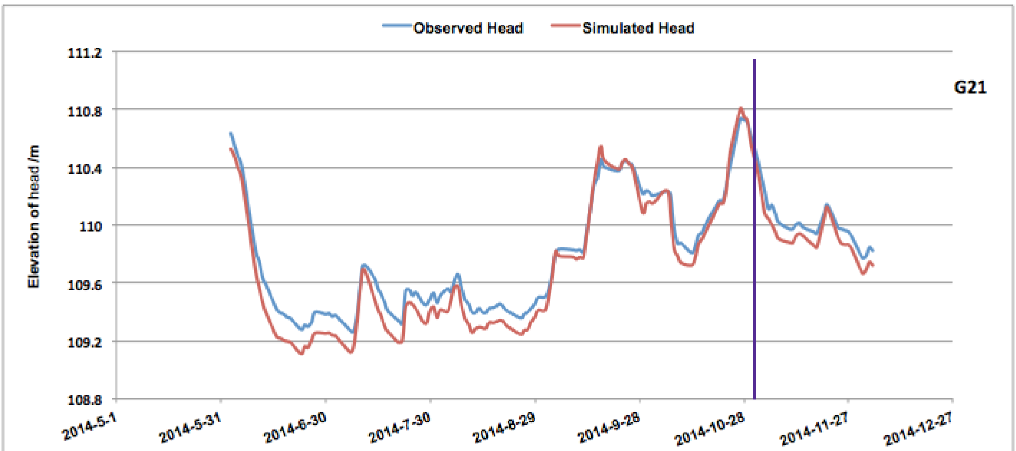


Figure 15: Observed and simulated head difference in modeled period (left) and validated period (right) at G10 and G21

The reasonable match of calibration and validation could indicate the model is capable of accurately representing the groundwater system and the result of simulation is reliable.

5.3 Percent discrepancy of Water budget

The water budget is a summary of all the inflows and outflows that take place in the groundwater system. The model calculate the water budget for the whole model and then check the acceptability of the solution, and provide information of the amount and rate of water sources or sink into this region. Moreover, the water budget is calculated independently of the equation solution process, and in this sense may provide independent evidence of a valid solution. Percent discrepancy, is calculated by the following formula:

$$D = \frac{100(IN - OUT)}{(In + OUT)/2} \quad \text{(Equation 5.4)}$$

where

D is the percent discrepancy,

IN is the total inflow to the system,

OUT is the total outflow.

Figure 16 shows the calculated percent discrepancy after execution of calibrated model, using the PCG (Preconditioned Conjugate Gradient) package under maximum absolute value change in head equal to 1E-6. The maximum value took place at time step 113 equal to -0.05. The small percent discrepancy throughout the time steps also indicate that the model could accurately simulated the groundwater system.

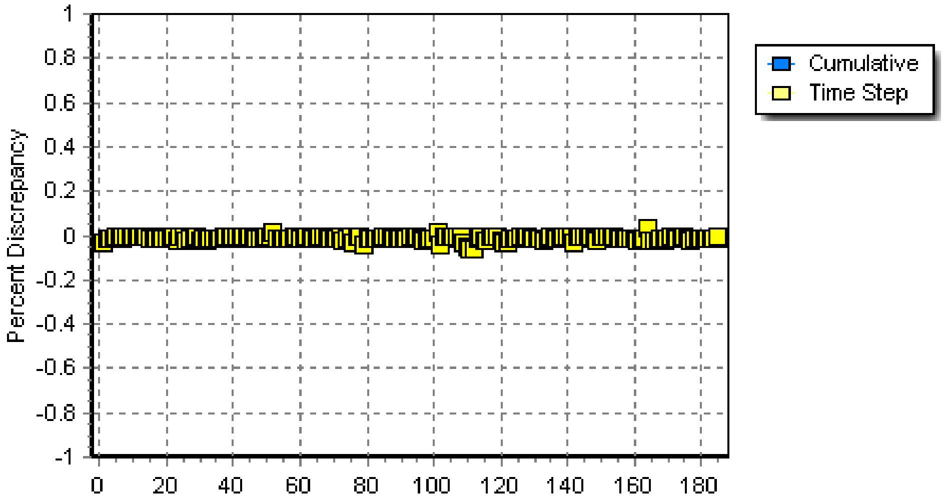


Figure 16: Percent discrepancy of water budget in Pirna model

6 Results and conclusion

6.1 Interpretation of the results

The resulting water table configuration reflects the balance between horizontal and lateral water flow through the field, and its relationship to river stage. The aquifer and Elbe river are hydraulic connected, when the water table higher than the river stage, river gains water from the groundwater, while when the water table lower than the river stage, river losing water to the aquifer. Figure 17 shows typical groundwater head distribution and its relationship with the river stage fluctuation at different stress period of 38, 42 and 72. Head distribution in the aquifer varies from day to day as well as the river stage. The figures below exhibits typical situations of the interaction between groundwater and Elbe river. In most cases during the model period, the groundwater losing water to the river as Figure 17a and Figure 17b shows, however, the exchange direction could be reversed (Figure 18) due to rapid rise in river stage which might be caused by heavy precipitation from upstream region. When the water comes from the river to the aquifer, a slightly larger gradient could be detect in the region close to the river, as water flows further, the gradient gets smaller and water exchange towards to be balance, which might affected by the slope of the bedrock and the amount of water that can flows into the aquifer.

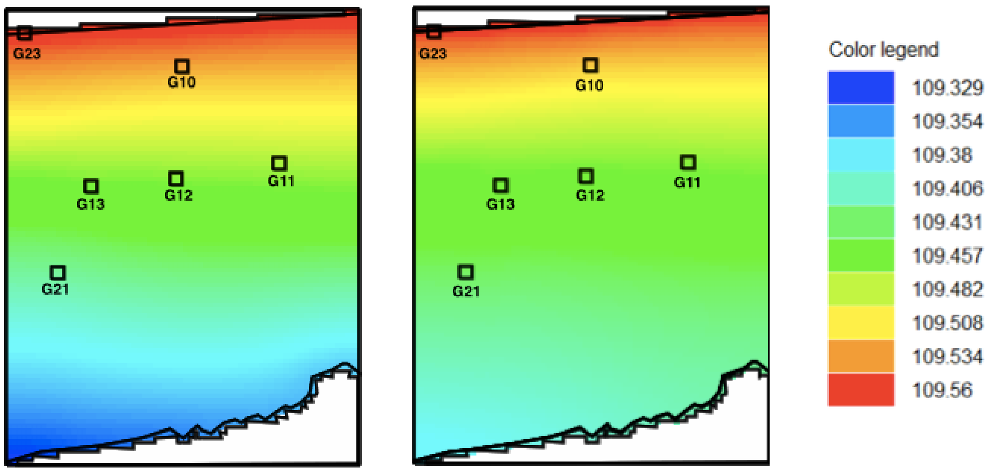


Figure 17: Head distribution of aquifer and river in the model domain in stress period 38 and 42 (gaining river, unit: m.a.s.l.)

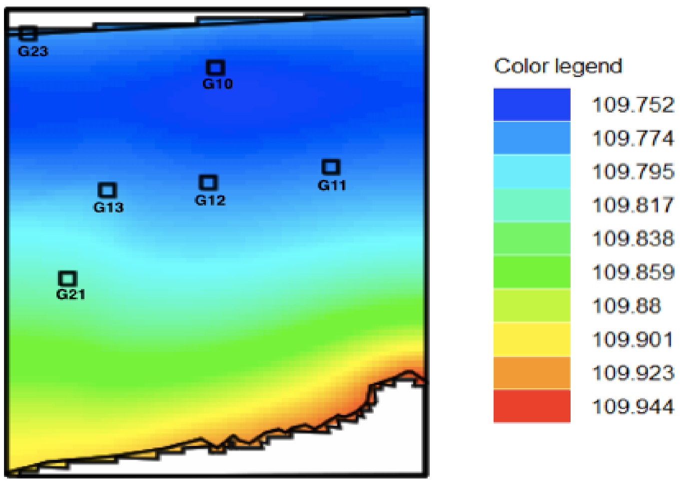


Figure 18: Head distribution within the model domain in stress period 72 (losing river,unit: m.a.s.l.)

Figure 19 indicates that the simulated head of groundwater and river stage are highly hydraulic connected in a similar variation behavior. Among them, G21 has a relatively small difference with the river stage whereas the G10 shows a big difference due to its longer distance to the riverbank. The good agreement shows the possible reason that the sediment compositions of the interface between the aquifer and river has a relatively high hydraulic conductivity that allows a relatively high water exchange rate. Furthermore, the uniformity variation behavior also confirms application of the Specific-head boundary for the river and aquifer is feasible and reliable.

Figure 19: Simulated head at five wells and river stage for the entire stress period.

6.2 Uncertain analysis

There are three major categories of uncertainty exist in this model: uncertainty generated during the measurements, uncertainty caused by interpolation and uncertainty cause by the model. The inherent uncertainty exist during measurements is often inevitable. For example, different colleagues may give readings of 0.5 cm difference of groundwater table due to their different standard towards accuracy of the tape. The accuracy of interpolation method is highly associated with the data density and distribution. In the filed of Pirna, there are only 8 wells with data, which will definitely lower the reliability of the interpolated data. Furthermore, groundwater models are designed to simulate the real groundwater system, but will not represent the real situation exactly.

6.3 Conclusion

In this work, a 3D groundwater flow model was constructed for the field site of Pirna, which could successfully represent the local hydrogeological conditions and reveal interaction between the aquifer and Elbe river. The aquifer and river are hydraulic connected and the aquifer could respond fast to the variation on the river stages. Hydraulic conductivity distribution in the field is obtained by transform the DPIL Kr value from DPST K value with regression analysis, which indicates that the aquifer mainly consists of sand and gravel. However, there is some limitations due to the data insufficiency. Therefore， further investigations can be performed for collecting more detailed information, such as the infiltration coefficient and specific yield of the field site, and conductance of Elbe river etc. The trial-and-error method was used for calibration, and the RMS error was controlled less than 0.106 m as well as MA error less than 0.096 m. Data from other years and November of 2014 was used for model validation, good fits were achieved and indicating that the model is capable of representing the field for continued simulation.

Outlook

1. Proper validation of the model could be performed in further steps with a long-term observation. Ucode or Pest could be involved to assist calibration and for sensitivity analysis of parameters.

2. More observation wells could be installed in the vicinity of river and in the northern part of the field, which is important for increasing the accuracy of interpolation and more appropriately set the boundary condition.

3. Due to the distribution of permeability and sediments thickness of the river bed could strongly affect the water exchange between groundwater and river, investigations of the conductance of the river could be very beneficial to get a better understanding of the flow processes, thereby the river boundary could be set as the third boundary which might be more appropriate to simulate the river.

4. A series of tracer tests with Fluorescein and particle tracer have been conducted in G12 to evaluate the flow direction and velocity of the aquifer. Construction of a transport model would be beneficial to get a better understanding of the flow processes together with the current model.

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