HYDROGEOLOGICAL PARAMETER DISTRIBUTION ESTIMATION BY GEOSTATISTICAL METHODS IN REGIONAL GROUNDWATER MODELING IN THE UPPER CENTRAL PLAIN, THAILAND

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ABSTRACT

Groundwater is playing an important role in water extraction and utilizations especially in Upper Central Plain, Thailand. The farmers traditionally relied on rain and flood water for crops but the amount needed for rice cultivation was not proper in dry years and used groundwater as a supplement especially in the dry year. In this study, regional groundwater model is developed with small grid size (2km x 2km) and improved by adding more well data, estimating transmissivity from well data and applying the geostatistical method to estimate the hydraulic conductivity distribution. The groundwater model in the Upper Central Plain was calibrated (1993-1999) and verified (2000-2005) using peizometric heads observed. The proposed parameter estimation and its distribution were proved to workable under limited available well pump test data and can improve the peizometric head simulation when applied to the regional groundwater model in the study area.

Keywords: Transmissivity, Kriging method, Hydraulic conductivity, Groundwater Model


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1. INTRODUCTION

Upper Central Plain is the most important area for Thailand’s economy. It is also the most agriculturally productive area. Demand for water in this area far exceeds locally available supply. The amount needed for rice cultivation was not proper in dry years and used groundwater as supplement [3]. The local farmers depended on both surface water and groundwater sources especially in the dry season from October to April [10]. There is a need to evaluate the groundwater potential and understand surface water and groundwater
interactions in order to manage both surface water and groundwater properly [4]. Groundwater system used a groundwater model for conjunctive use patterns investigation in the Upper Central Plain of Thailand [5]. The proper parameter estimation from short pumping data and pumping test data was applied for the groundwater modeling [8]. In this area, there are some studies on the flow budget, conjunctive use and climate change impact, e.g., the impact study on groundwater using regional groundwater model created the hydraulic conductivity by zoning approach with the grid size of 10 km x 10 km [11, 12].

1.1. Purpose of study
In order to improve the regional groundwater modeling in the study area, in this study, regional groundwater model is developed with smaller grid size (2km x 2km) and improved by adding more well data to estimate transmissivity and applying the geostatistical methods to estimate the hydraulic conductivity distribution.

1.2. Study area
Upper Central Plain is located in the Northern part of Chao Phraya Plain covering the areas of Sukhothai, Phitsanulok, Kamphangphet, Pichit, and Nakornsawan Provinces. Total area is 47,986km². Average height is approximately 40-60 meters above mean sea level. The main rivers in the study area are the Yom River (West) and the Nan River (East) which are parallel flow from North to South (Figure 1a). The average annual rainfall is between 900 to 1336 mm/year with more than 81% of the annual rain falling during the rainy season from April to September, and less than 19% of the annual rain falling during the dry season from October to March.

![Figure 1](image1.jpg)  
**Figure 1** Upper Central Plain Basin (1a) and Model grid design (1b)

2. MATERIALS & METHODS
Hydrogeological approaches were employed to estimate the values of aquifer parameters such as specific capacity ($S_c$), transmissivity ($T$), and hydraulic conductivity ($K$) from pumping test. The aquifer system for this study defined unconfined layer between 40-100m. The 3D block-centered grid model representing the groundwater basin has a uniform grid size of 2 km x 2 km, resulting in a model grid of 152 rows and 93 columns, total number of nodes are...
14136 elements in the layer. There are more than 16597 active cells, covering a model area of 47,986km$^2$. The grid is aligned N-S and E-W (Figure 1b).

The study used the bore logs data of 259 observations wells (Figure 2a) to estimate the hydraulic conductivity distribution from transmissivity and verify with data with pumping test (7 wells) data (Figure 2b) and these values are used as initial values for parameter optimization in groundwater modeling. The distribution of hydraulic conductivity, by using geostatistical methods, are used to interpolate the groundwater level based on observed wells by comparing three methods (Inverse Distance Weight, Natural Neighbour and Kriging). A steady-state model was calibrated and a transient model was constructed using the calculated heads from the steady-state model as initial conditions.

Figure 2 Maps showing the aquifer characteristics (Source; Dept. of Groundwater Resources) and well locations; a) yellow spots mean short pumping wells (259 wells), b) Red spots mean full pumping test wells (7 wells).

2.1. Aquifer characteristics

The high terrace deposits, the low terrace deposits and recent flood plain deposits are the main hydrogeological characteristics of this area, while the western and eastern areas were consolidated aquifers, composed of granite and volcanic rocks. The aquifer system in this study was defined by vertically as a two-layer aquifer, whereby the thickness of the upper, semi-confined layer varies between 40-100m and lower, confined layer between 100-200m [3] (Figure 3).
2.2. Equation used

2.2.1. Aquifer Transmissivity and Specific Capacity

The theoretical relationship between specific capacity and transmissivity is linear on a log scale. Several authors have also developed empirical or observed relationship between specific capacity and transmissivity \([1, 2, 6, 14, 7, and 4]\). Because of its abundance and cost effectiveness, specific capacity is used to estimate the transmissivity of an aquifer to be used as initial values of parameter optimization. An empirical relationship for the study area is determined by linear regression of the log-transformed information. The hydraulic properties of the aquifer, namely, transmissivity and specific capacity, were estimated from pumping tests by using this equation and are expressed as:

\[
T = a(S_c)b \quad \text{(1)}
\]

\[
S_c = \frac{Q}{\Delta s \cdot I} \quad \text{(2)}
\]

where, \(Q\) is discharge \((m^3/h)\); \(\Delta s\) is drawdown (m); \(I\) is screen length of the well (m); \(T\) is transmissivity (long pumping test); \(S_c\) is specific capacity \((m^3/h)\); and \(a, b\) are constants.

2.2.2. Hydraulic Conductivity

The hydraulic conductivities are generated from the aquifer data and optimized the hydraulic conductivities in steady state with known observed peizometric. The hydraulic conductivity is defined as transmissivity \((T)\) divided by the saturated thickness of the aquifer. The estimating of hydrological parameter distribution, in particular, hydraulic conductivity was examined in geostatistical frameworks (Inverse Distance Weight, Natural Neighbour and Kriging) and selected the appropriate methods. Then, the regional groundwater model was computed with best geostatistical interpolation, i.e., Kriging method.

2.2.3. Kriging Method

Kriging interpolation is used in calculating estimates of the surface at each grid node. Errors in this method are independency from variable and dependent to spatial location and it cause to predict the best location sampling is possible. Variogram relationship based on the measured points is as follows:

\[
\gamma(h) = \frac{1}{2n(h)} \sum_{i=1}^{n(h)} [z(x + h) - z(x)]^2 \quad \text{(3)}
\]

where, \(\gamma(h)\) is the variogram for a distance (lag) \(h\) between observations \(z(x)\) and \(z(x+h)\); \(n(h)\) is the number of pairs of observations which are at distance \(h\); \(z(x)\) is the observed variable. \(z(x+h)\): the observed variable is the \(h\) distance from \(z(x)\) and variogram \(\gamma(h)\).
2.2.4. Groundwater Modeling

Groundwater model is used to predict aquifer response, in terms of head (ground water level) and fluxes into and out of an aquifer, to natural and human induced stresses. The three-dimensional movement of ground water of constant density through porous earth material may be described by the partial differential equation.

\[
\frac{\partial}{\partial x} [K_{xx} \frac{\partial h}{\partial x}] + \frac{\partial}{\partial y} [K_{yy} \frac{\partial h}{\partial y}] + \frac{\partial}{\partial z} [K_{zz} \frac{\partial h}{\partial z}] + W = S_s \frac{\partial h}{\partial t} \tag{4}
\]

where, \(K_{xx}, K_{yy}\) and \(K_{zz}\) are the values of hydraulic conductivity along the x, y, and z coordinate axes and are function of space. \(h\) is the potentiometric head (hydraulic head). \(W\) is a volumetric flux per unit volume representing sources and/or sinks of water. \(S_s\) is the specific storage of the porous material and is function of space and \(t\) is time.

3. RESULTS OF MODEL APPLICATION AND DISCUSSION

3.1. Estimation of Transmissivity from Pumping Test

3.1.1. Relationship of transmissivity \((T)\) and Specific Capacity \((S_c)\) from Full Pumping Test Data

The transmissivity value from long pumping test data is useful to develop a relationship between specific capacity and transmissivity. The empirical relation between transmissivity \((T)\) and specific capacity \((S_c)\) was obtained from the data set long pump test (7 well data) utilizing log-log regression analysis. The hydraulic parameters are estimated from equation 1 with \(a = 0.46, b = 1.01\) (Figure 4).

![Figure 4 Relationship of T and Sc from Pumping test (m²/h)](image)

3.1.2. Relationship of Transmissivity \((T')\) and Specific Capacity \((S_c)\) from Short Pumping Well Data

To justify the determination of transmissivity based on specific capacity data, it is assumed that transmissivity is linearly proportional to the specific capacity of a well [13]. Analytical methods for relating transmissivity to specific capacity involve using equation 1 and 2 that are based on the theory of groundwater flow. 259 pairs of transmissivity and specific capacity values calculated a best-fit line by utilizing log-log regression analyses from well data records with short pump data. It is useful to develop a relationship between transmissivity \((T')\) and specific capacity. Specific capacity is used to indicate the productivity of the well, and is defined as the discharge of a well divided by the drawdown. The derived constants \(a\) and \(b\) from full pumping data (7 well data) are substituted into equation 1 and estimated the transmissivity \((T)\) for short well pumping data. The estimated transmissivity \((T)\) is divided by...
aquifer depth (L) and find the transmissivity (T') to check the slope of well data with pumping test data (Figure 5). The derived constants a and b are 1.62 and 0.42 respectively.

![Relationship of T' and S_c from well data(m²/h)](image)

**Figure 5** Relationship of T' and S_c from well data records (with short pumping data)

It can be seen that the graph from Figure 4 (from full pumping test) and Figure 5 (from short pumping well data) showed a comparable slope which means that the slope derived from the short pumping well data go along well with full pumping test data and transmissivity value can be estimated by using the relationships and to be used as initial values in the parameter optimization of groundwater modeling stage.

### 3.2. Estimation of Hydraulic Conductivity Distribution

Distribution of hydraulic conductivity estimated using various methods of geostatistics in the study area. Hence Inverse Distance Weighting (IDW), Natural Neighbour (NN) and Kriging with steady-state and choose the good method based on their errors. A result of Kriging method is better than other methods. According to good performance, Kriging method of geostatistics is selected to interpolate hydraulic conductivity in the groundwater model (Table 1). This method can calculate the estimated value and estimation accuracy. Before interpolating a scatter points set using Kriging option, a model variogram is defined involving in constructing an experimental variogram and then construct a model variogram. (Figure 6)

The parameters for the Gaussian model are using $h = 5,000$ to represent lag distance, $a = 46,846$ to represent (practical) range, and $c = 30.268$. The distribution of hydraulic conductivity field conformed to hydrogeological map using Kriging method. (Fig 7a) The hydraulic conductivity value in the river is from 20 m/d to 30m/d while in mountains are from 10m/d to 20m/d. Then, hydraulic conductivity distribution pattern is similar to the pattern of hydrogeological map (Figure 7b).

**Table 1** Comparison among methods for hydraulic conductivity distribution

<table>
<thead>
<tr>
<th>No.</th>
<th>Estimation methods</th>
<th>IDW</th>
<th>NN</th>
<th>Kriging</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Minimum</td>
<td>-35.55</td>
<td>-37.07</td>
<td>-30.26</td>
</tr>
<tr>
<td>2</td>
<td>Maximum</td>
<td>44.39</td>
<td>47.10</td>
<td>38.16</td>
</tr>
<tr>
<td>3</td>
<td>Mean error</td>
<td>-0.52</td>
<td>0.09</td>
<td>-0.26</td>
</tr>
<tr>
<td>4</td>
<td>Mean abs. error</td>
<td>8.99</td>
<td>9.15</td>
<td>8.31</td>
</tr>
<tr>
<td>5</td>
<td>Root mean squared error</td>
<td>12.42</td>
<td>12.79</td>
<td>10.62</td>
</tr>
<tr>
<td>6</td>
<td>Nash-Sutcliffe coefficient</td>
<td>0.13</td>
<td>0.08</td>
<td>0.80</td>
</tr>
<tr>
<td>7</td>
<td>Performance</td>
<td>2</td>
<td>3</td>
<td>1</td>
</tr>
</tbody>
</table>
3.3. Application to Regional Groundwater Model

A steady-state model was calibrated with data from predevelopment time and a transient model was constructed using the calculated heads from the steady-state model as initial conditions. The simulation time is divided into a steady-state period and a transient period. The steady-state period is 1993 in which no pumping was simulated. The transient period from 1993 to 2005 was divided into annual stress period for which pumping rates were defined. The model recalibrated with more well data, new parameter estimated from this study and compared with observed data.

3.3.1. Steady-state

The hydraulic parameters estimated from Equation (1) (with a=1.62b, b=0.42) and 259 well data were input to the model as initial values and by using optimization scheme; the computed heads were compared with the observed data. Results of recalibration model show that simulation values were closed with observed data compared with the former and pumping test in Equation (1) for hydrological parameters estimation in steady state. The model was
calibrated in the steady state and, after optimization, gave the good performance when compared with the observed data. (Figure 8)

Figure 8 Comparison of computed and observed peizometric heads in the steady-state

3.4. Model Calibration and Verification

Groundwater flow model (MODFLOW) was used to simulate groundwater flow conditions in the study area during the period 1993-2005. River water level, recharge rate and boundary condition were used (flow in is +257,576m$^3$/d; flow out is -111,000m$^3$/d) from previous study [15]. Observed groundwater level and estimated hydraulic conductivity were used as input data in this study. The boundary conditions and model parameters are computed and assigned to the proper cells. Boundary is used to identify specific head boundaries and gave the heads at boundaries.

3.4.1. Transient state

Verifications from 2000 to 2005 were preceded in the model area to calibrate conductance and verified the groundwater flow. Similar with the result in the steady-state, the computed peizometric head values in transient state, gave the good performance when compared with the observed data (Figure 9). The error from this study is smaller than the previous study [15]. The total error summary in both states and previous study are compared and shown in Table 2.

Figure 9 Comparison of computed and observed peizometric heads in transient state
Table 2 Error summary of calibration results and verification result in both states compared with the previous study results

<table>
<thead>
<tr>
<th>No.</th>
<th>Error (m)</th>
<th>Steady-state</th>
<th>Transient state</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>Maximum</td>
<td>9.74</td>
<td>3.13</td>
</tr>
<tr>
<td>3</td>
<td>Mean error (ME)</td>
<td>0.43</td>
<td>0.62</td>
</tr>
<tr>
<td>4</td>
<td>Mean absolute error (MAE)</td>
<td>2.80</td>
<td>1.94</td>
</tr>
<tr>
<td>5</td>
<td>Root mean squared error (RMSE)</td>
<td>3.42</td>
<td>2.09</td>
</tr>
<tr>
<td>6</td>
<td>Nash-Sutcliffe coefficient (NSE)</td>
<td>0.92</td>
<td>0.98</td>
</tr>
</tbody>
</table>

4. CONCLUSIONS

The introduction of smaller grid size in regional groundwater model with more hydraulic conductivities data from more pumping well data and hydraulic conductivity distribution via Kriging methods improved the regional groundwater model simulation in the study area. This study found that there is a good relationship of transmissivity and specific capacity from both short pumping well data and long pumping test data. Then the hydraulic conductivity can be estimated from (short) pumping data from well data records, due to limited pumping data, and used as initial values for parameter optimization in groundwater modeling. Kriging method can be well applied to interpolate hydraulic conductivity distribution. The hydraulic conductivity in the plain is found to be higher than in the mountainous area which is corresponded with hydrogeological characteristics. The Nash-Sutcliffe coefficient in steady-state mode is 0.98 m and in transient state is 0.94 m for calibration and 0.93 m for verification. In summary, the root mean square calibration error is 3.42 m in steady-state mode and 4.53 m in transient mode. The root mean square error of verification model is 4.61 m. The estimated hydraulic conductivities can reflect the fluctuation of water potential in the study area and gave the better distribution. Hence, the proposed parameter estimation and its distribution were proved to workable under limited available well pump test data and can improve the piezometric head simulation when applied to the regional groundwater model in the study area.

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