Estimation of Watershed-scale Hydraulic Conductivity for Two Watershed Sites using GFLOW

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ABSTRACT

For hydrologic and water quality studies, proper estimation of the hydraulic conductivity of the study site is very important. The hydraulic conductivity values determined in the laboratory are usually lower than those observed in the field. The hydraulic conductivity increases with measurement scale. This increase with larger scale is the result of spatial heterogeneities and is described as scaling-up of hydraulic conductivity. Field and laboratory experiments to determine hydraulic conductivity values for large areas are expensive and time consuming. Modeling may be a practical option to estimate hydraulic conductivity when the study area is large. GFLOW, which is an analytical element model, was used to estimate the hydraulic conductivity values for two watershed sites in Illinois, namely the Big Ditch watershed and the Upper Embarras River watershed. For each site, heads in shallow observation wells and stream discharge were used to calibrate the model. The calibrated hydraulic conductivity values for the Big Ditch and Upper Embarras River watersheds were 4.05E-04 and 4.86E-04 m/s, respectively. For watershed-scale studies, the hydraulic conductivity values estimated by the model might be acceptable.

Keywords: GFLOW, model calibration, hydraulic conductivity, measurement scale, USA.

1. INTRODUCTION

The hydraulic conductivity is a very important hydrologic parameter which influences the groundwater movements to a great extent. Soil hydraulic properties are usually measured in the laboratory using representative soil samples from the study area. Since the hydraulic properties exhibit large variations within a spatial domain, a large number of soil samples is required to characterize the hydraulic properties of the study area. Field and laboratory methods for the estimation of hydraulic properties are complex and time-consuming (Rawls *et al.*, 1982; Sepaskhah and Ataee, 2004; Parasuraman *et al.*, 2006). Spatial variability analysis of the hydraulic conductivity involves a large number of soil data which is not easy to collect. Direct measurement of hydraulic conductivity does not appear to be generally feasible because of the high cost, dynamic nature and substantial short-range variation of the parameter in the field (McKenzie and Jacquier, 1997).

Another important fact is that the hydraulic conductivity values estimated in the laboratory are lower than *in situ* observations (Zecharias and Brutsaert, 1988). Scale effects (increase in the value with increasing scale of measurement) on hydraulic conductivity have been reported in the literature. The median hydraulic conductivity increases with measurement scale (Guimerà *et al.*, 1995). The increase in hydraulic conductivity with larger scale is the result of spatial heterogeneities (Rovey II, 1998) and was described as scaling-up of the hydraulic conductivity (Desbarats, 1992). The high value of hydraulic conductivity in the shallow geologic material might also be due to the presence of macropores, such as desiccation cracks, root channels and worm holes (Mehnert *et al*., 2005). There are numerous examples in the literature which reported high hydraulic conductivity values for unconfined and confined aquifers. Table 1 shows some of the high hydraulic conductivity values from different studies.

Table 1. Watershed-scale hydraulic conductivity values from different studies.

Reference	Calibrated hydraulic conductivity (m/s)
Sloan, 2000	9.26E-04
Barlow *et al.*, 2003	7.06E-04
Mehnert *et al.*, 2005	1.32E-04
ISWS, 2003	1.23E-03
Roadcap and Wilson, 2001	9.88E-04
Modica and Buxton, 1998	7.06E-04
Rodriguez *et al.*, 2005	3.99E-03
Goswami and Kalita, 2009	5.52E-04

Models may be useful in estimating the hydraulic conductivity that represents the entire area under study. The advantage of using such a predictive model is that it provides a means for predicting reliably and rapidly the best estimate possible of the representative value from limited *in situ* measurements (Sepaskhah and Ataee, 2004). There are numerous models available to calibrate watershed-scale hydraulic conductivity. In this study, a steady-state model was used to show how a simple model could be conveniently used to estimate the hydraulic conductivity for a relatively large area. The model is based on an analytical element model called GFLOW developed by Haitjema (1995). GFLOW allows one to develop conceptual models of groundwater flow based on steady-state water elevations, such as mean water levels in streams, lakes, and wells.

2. MATERIALS AND METHODS

2.1 Site Description

The study was conducted in two watersheds, namely the Big Ditch (DB) watershed and the Upper Embarras River (UER) watershed in Illinois. The BD watershed is predominantly an intermorainal landscape. It also includes portions of the Rantoul Moraine in the south-southeast part of the watershed and the Illiana Morainic system in the northeast part of the watershed. This moraine system suggests that a significant amount of glacial melt-water flowed through the watershed. Sand and gravel deposits associated with these streams are found within the shallow subsurface. The unconsolidated geologic deposits throughout the watershed consist of a sequence of glacial and post-glacial deposits that range in thickness from 79 to 122 m and overlie Mississippian and Devonian bedrock. The highest land surface elevation peaks at 252 m above mean sea level (MSL) in the northernmost end of the watershed. The lowest elevation in the watershed is 212 m, at the outlet of the watershed (Mehnert *et al.*, 2005). The surficial soils in the Big Ditch watershed are predominantly silt loams and silty clay loams. The five most common soils, Drummer silty clay loam, Raub silt loam, Elliott silt loam, Parr silt loam, and Ashkum silty clay loam, cover approximately 82% of the watershed. Most soils in this watershed are considered somewhat poorly to poorly drained and have moderate to high organic matter content (USDA-SCS, 1982).

The Embarras River originates near Urbana-Champaign, IL, and the UER watershed encompasses an area of 48,173 ha. Soils of this area developed from Wisconsinan till that supported primarily prairie vegetation. Drummer silty clay loams and Flanagan-Catlin are dominant soil types in the UER watershed (David *et al.* 1997).

2.2 The GFLOW Model

GFLOW is a highly efficient groundwater flow modeling system based on the analytic element method. The analytical element method does not require discretization of a groundwater flow domain by grids. In GFLOW, only the surface water features in the domain are discretized, and entered into the model as input data. Each of the stream or lake sections is represented by the analytic elements. The comprehensive solution to a complex, regional groundwater flow problem is obtained by superposition of all analytic elements in the model (Haitjema, 1995). Since the model does not have grids, the heads and flow can be computed anywhere in the model domain without nodal averaging (Juckem and Hunt, 2007). It simulates steady-state flow in a single heterogeneous aquifer using the Dupuit-Forchheimer assumptions (Reddi, 2003). It is particularly suitable for simulating regional horizontal flow (Yager and Neville, 2002). GFLOW has powerful elements like line-sinks with bottom resistance, drains, wells, recharge and domains with different hydraulic conductivity values. Specialized analytic elements may be used for special features, such as drains, cracks, slurry walls, etc. (Haitjema, 1986; Haitjema, 1995; Scientific Software Group, 2007).

For this study, all the major streams and their branches in the selected areas within the two sites were delineated in base maps (Figures 1 and 2). The next step was to assign head elevations to those streams in the base maps. This was carried out with the help of 7.5 minute USGS topographic maps (ISGS, 2008) and Digital Elevation Models (DEMs) for the two sites. The stream head elevations can be determined from the 7.5 minute USGS topographic maps (Haitjema Software, 2001) and DEMs (Johnson and Paquin, 2007). The model determines the water table contours for the entire site based on these elevation data and other hydrologic inputs like hydraulic conductivity, porosity, recharge rate, and aquifer thickness.

2.3 Calibration Procedure

Stream flow is known to be a sensitive parameter for defining steady-state groundwater flow (Mitchell-Bruker and Haitjema, 1996). Therefore, this was considered in the model calibration. The measured heads at the wells can be compared with the simulated heads for model calibrations (Ireson *et al*., 2006). Therefore, in this study, both stream flow and well heads were used to calibrate the model for both sites.

The BD and UER watersheds are subsurface (tile)-drained, but the days chosen for the model calibrations were such that the tile drains were not flowing on those days (23 June, 2003 and 19 October, 2005 for the BD and the UER sites, respectively). In other words, on those days, the water table level was below the tile-drain elevations for the respective watersheds. Flow rates in a stream section and well (denoted by well 1 for the BD site, and well 1-3 for the UER site in Figures 1 and 2, respectively) head data near that stream were available for each watershed. These were the data collected for another study at the two sites. Additionally, data from wells maintained by the Illinois State Geological Survey (for the BD site, denoted by wells 2-7 in Figure 1) and the State Water Survey, Illinois (for the UER site, denoted by well 4 in Figure 2) were used for model calibration. Various combinations of hydraulic conductivity, recharge, and aquifer thickness were used for the model calibration. A porosity of 0.2 was used from the literature (Mehnert et al., 2005) in the model. The objective of the calibration process was to find the hydraulic conductivity value that would result in a good agreement between the observed and simulated data (stream flow and well heads).

For each site, all the streams were input as far-field features except for the stream for which the flow rate was known at a single point denoted by a red triangle (Figures 1 and 2). This particular stream was input as a near-field feature. For the near-field feature, the depth, and width of the stream need to be incorporated in addition to the stream head elevations. For the far-field feature, only the stream head elevations for the stream need to be incorporated. GFLOW determines the head across the aquifer and stream flow rates in near-field streams. In Figures 1 and 2, the dotted lines are the lines of equal heads (water table elevation). The water table elevations (in meter) are marked along with the lines. GFLOW has the option to select the area within the model where the user wants to apply recharge. Recharge was applied to the area within the red rectangle (Figures 1 and 2).

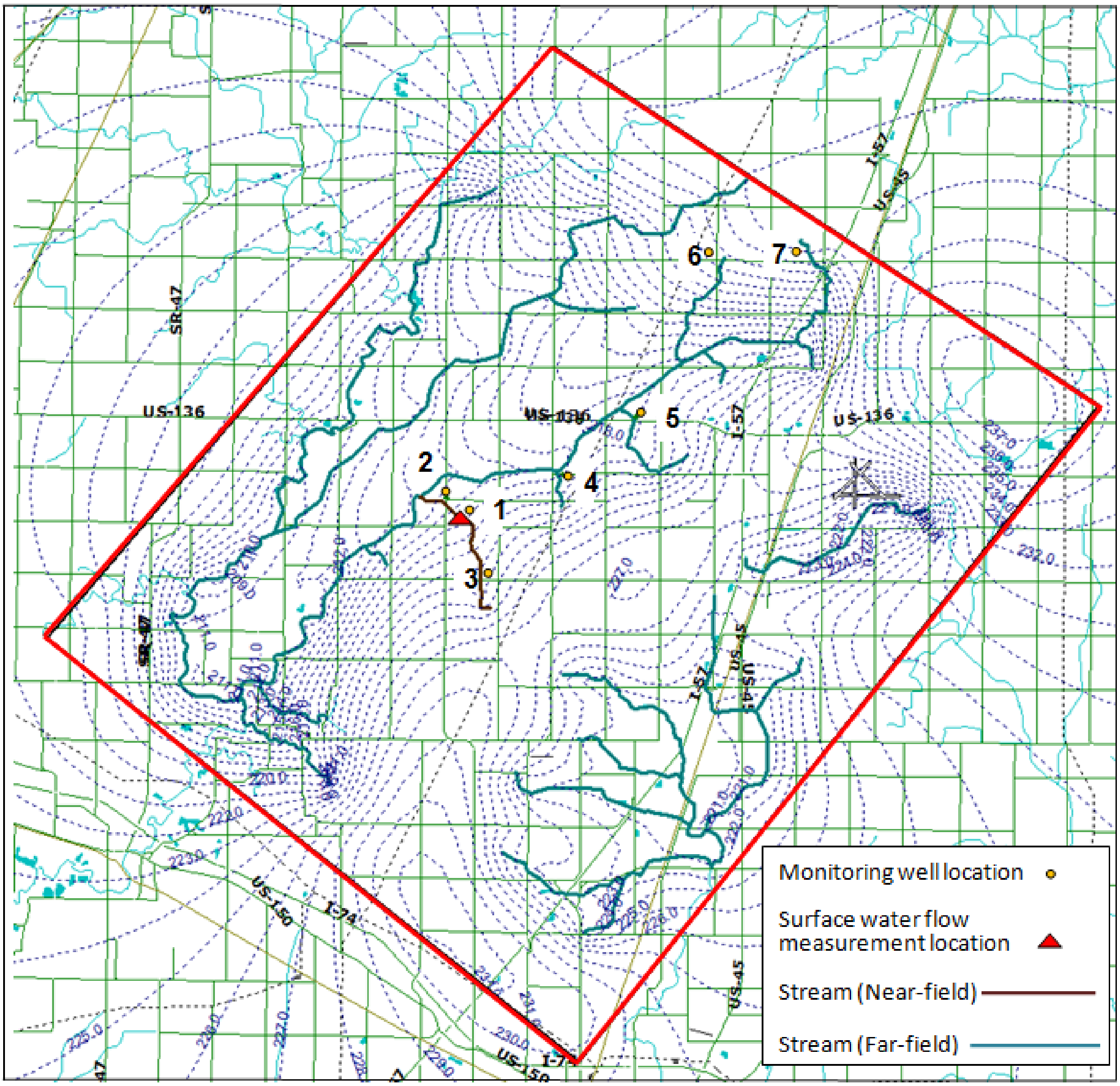

Figure 1. GFLOW calibration of hydraulic conductivity for the Big Ditch site.

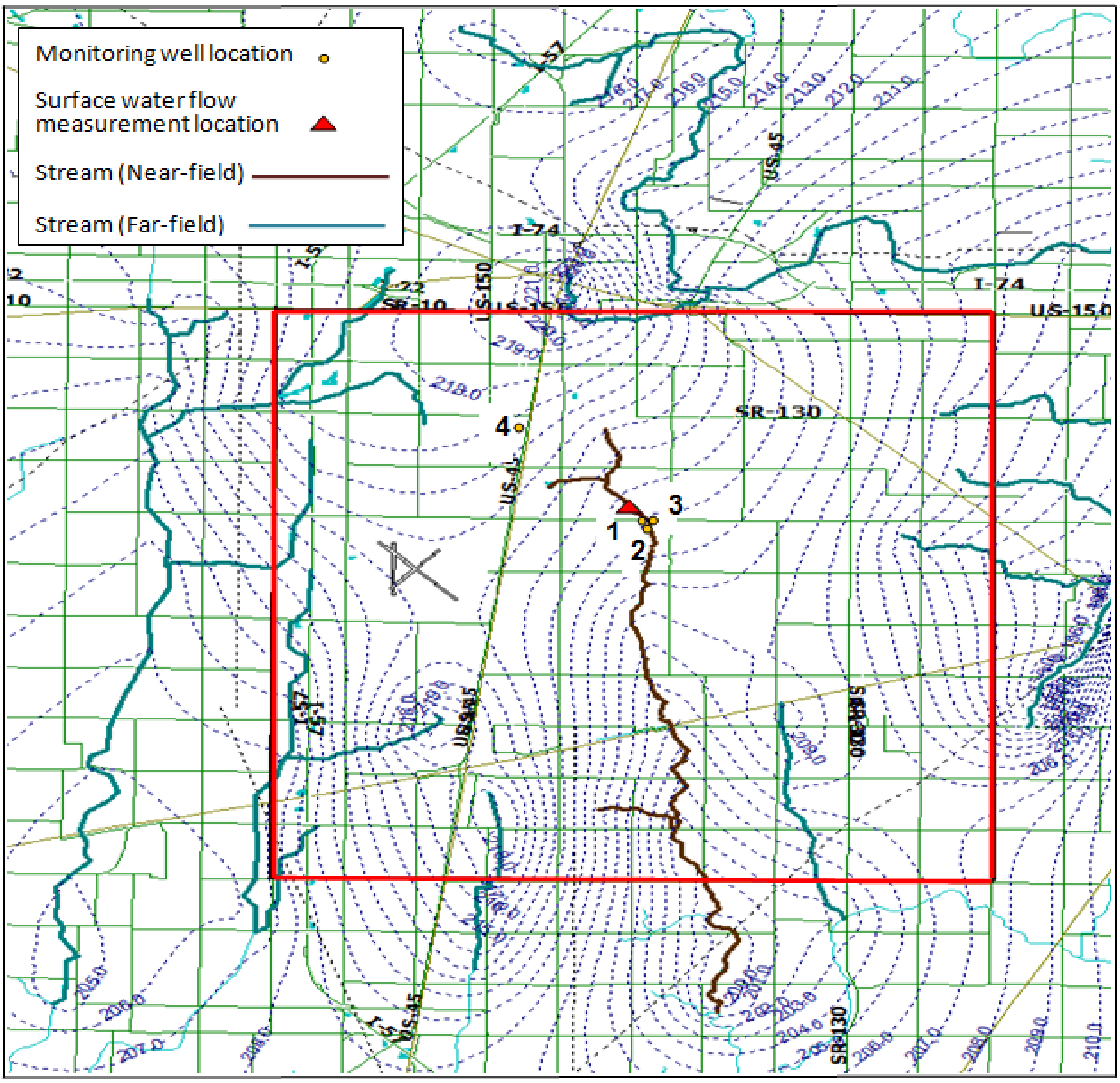

Figure 2. GFLOW calibration of hydraulic conductivity for the Upper Embarras River site.

3. MODELING RESULTS AND DISCUSSIONS

Table 2 shows the calibrated hydraulic conductivity, and other parameters for which there was a good agreement between the measured and simulated data (stream flow and well heads) for the two sites. Tables 3 and 4 show the measured and simulated heads for the two sites. The measured and simulated flow rates at the monitoring site in the near-field stream at the BD site were 0.039 and 0.036 m^3/s, respectively. For the UER site, the measured and simulated stream flow rates in the near-field stream were 0.024 and 0.022 m^3/s, respectively. The calibrated hydraulic conductivity values for the BD and UER sites were 4.05E-4 and 4.86E-04 m/s, respectively.

For the BD site, the hydraulic conductivity calibrated by GFLOW was larger than the values determined using slug tests. Mehnert *et al.* (2005) found median hydraulic conductivity value for the BD site using slug tests to be 2.9E-06 m/s. ISWS (2003) and Mehnert *et al.* (2005) reported higher hydraulic conductivity values from model calibrations for the BD site (1.32E-04 and 1.23E-03 m/s, respectively). The hydraulic conductivity for the BD site calibrated by GFLOW (4.05E-04 m/s) was within these two values mentioned in ISWS (2003) and Mehnert *et al.* (2005). Sanderson (1998) found the average hydraulic conductivity at four observation wells at the Embarras River Valley at Jasper County to be 5.68E-04 m/s by an aquifer test. The hydraulic conductivity for the UER watershed calibrated by GFLOW (4.86E-04 m/s) was comparable with this value.

Table 2. Calibrated model parameters for the big Ditch and the Upper Embarras River sites.

Parameter	Big Ditch site	Upper Embarras River site
Aquifer thickness (m)	6.0	7.0
Recharge (mm/d)	0.3	0.1
Porosity	0.2	0.2
Hydraulic conductivity (m/s)	4.05E-04	4.86E-04

Table 3. Measured and calibrated well heads for the Big Ditch watershed site.

Well	Measured head (m)	Simulated head (m)
Well 1	217.88	217.73
Well 2	215.66	216.30
Well 3	221.33	221.74
Well 4	217.39	217.82
Well 5	217.68	218.25
Well 6	229.63	229.32
Well 7	232.36	232.20

Table 4. Measured and calibrated well heads for the Upper Embarras River watershed site.

Well	Measured head (m)	Simulated head (m)
Well 1	211.21	211.09
Well 2	210.85	210.50
Well 3	210.60	210.44
Well 4	216.97	216.64

4. CONCLUSIONS

GFLOW was used to calibrate the hydraulic conductivity values for two watershed sites considering the fact that with increased scale, the watershed-scale hydraulic conductivity also increases due to spatial heterogeneity. For hydrologic and water quality studies, a good estimate of the hydrologic conductivity is necessary. Field and laboratory experiments to determine hydraulic conductivity for a large area are expensive and time consuming. Therefore, modeling might be a good option to estimate the hydraulic conductivity for a larger area.

Acknowledgement. Appreciation is extended to the Illinois State Geological Survey and the Illinois State Water Survey for providing data relating to wells maintained by the respective agencies.

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