Using solver to determine vertical groundwater velocities by temperature variations, Purdue University, Indiana, USA

Miguel A. Arriaga · Darrell I. Leap

Abstract Vertical groundwater velocities can be estimated based on directions of groundwater thermal gradients. Temperature-depth profiles were obtained from 12 monitoring wells at 3 different times of the year (Fall, Winter, and Spring) in West Lafayette, Indiana (USA) mainly on the Purdue University campus. Microsoft Excel Solver was used to match plots of groundwater temperature distribution in the wells with published type curves in order to find a dimensionless parameter β, from which vertical groundwater velocities were obtained. The vertical groundwater velocities found in the monitoring wells ranged from 0.92 to 4.53 cm/yr. Clay-rich aquitards presented greater vertical groundwater velocities than outwash aquifers. The highest groundwater velocities occurred in the Spring while the lowest were during the Winter. This method was found to be especially useful in glacially-derived materials with varying hydraulic conductivities for estimating vertical groundwater velocities in-situ.

Résumé Les vitesses verticales des eaux souterraines peuvent être estimées d'après les directions du gradient thermique. On a mesuré la variation de la température avec la profondeur dans douze forages de monitoring dont la plupart se trouvent dans le campus de l'Université de Purdue-Indiana, États -Unies. Les mesures ont été effectuées pendant trois périodes différentes: l'automne, l'hiver et le printemps. Le Solver de Microsoft Excel a été utilisé pour caler les graphiques de la température mesurés dans les forages sur les abaques publiées affin d'obtenir un paramètre non dimensionnel, soit ?Ò?z?n d'où on peut obtenir les vitesses verticales de l'eau. Les valeurs des vitesses verticales se rangent entre 0.92 cm/an et 4.53 cm/an. Dans les aquidard argileux les vitesses verticales présentent des valeurs plus grandes par rapport aux ceux obtenues dans les aquifères cantonnés dans des dépôts glaciaires. Les plus grandes vitesses ont été mesurées pendant le printemps tendis que les valeurs les plus basses apparaissent pendant l'hiver. Cete méthode est surtout utile pour déterminer in situ les vitesses verticales dans les matériaux glaciaires avec des conductivités hydrauliques variables.

Resumen Las velocidades verticales de aguas subterráneas pueden ser calculadas en base a direcciones de gradientes térmicos de aguas subterráneas. Perfiles de profundidades y temperaturas de aguas subterráneas fueron obtenidos de doce pozos de monitoreo durante diferentes épocas del año (otoño, invierno, y primavera) en la ciudad de West Lafayette, Indiana (Estados Unidos de América), principalmente en el campus de la Universidad de Purdue. Microsoft Excel Solver fue usado para igualar la distribución de las temperaturas de aguas subterráneas de los pozos con un conjunto de curvas ya existentes para encontrar un parámetro β sin ninguna dimensión, del cual se generaron velocidades verticales de aguas subterráneas. Las velocidades verticales de aguas subterráneas obtenidas en los pozos de monitoreo variaron de 0.92 cm/yr a 4.53 cm/yr. Los acuitardos arcillosos presentaron velocidades verticales de aguas subterráneas más altas que los acuiferos compuestos de arenas y gravas. Las velocidades de aguas subterráneas más altas ocurrieron durante la primavera, mientras que las más bajas se presentaron en el invierno. El método usado en esta investigación resultó útil para determinar velocidades verticales de aguas subterráneas in-situ, particularmente en materiales glaciales de distintas conductividades hidráulicas.

Keywords Groundwater velocity · Groundwater flow · Groundwater temperature · Groundwater recharge · Thermal conditions

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Introduction

The quantification and identification of mechanisms for replenishing aquifers comprise important aspects of ef-

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fective water resource management. Aquifers are replenished by precipitation, interbasin flow, interaquifer leakage, and induced stream infiltration. After recharge reaches the saturated zone, aquifer permeabilities and hydraulic heads control its magnitude and flow direction. In certain areas, windows of recharge and temperature gradients can help in showing velocities of vertical movement in these areas. Even though vertical groundwater velocities can be obtained in several ways from hydrologic and hydrogeologic data, this research has to do with the feasibility of determining vertical groundwater velocities throughout various geologic strata by using Microsoft Excel Solver based on groundwater thermal gradients obtained from monitoring wells located mainly on the Purdue University Campus.

Van Orstrand (1934) recognized the influence of moving groundwater on heat flux from the Earth. Because the natural heat-flux density from the Earth is usually small, the upward thermal gradient within the Earth is affected by groundwater movement (Bredehoeft and Papadopulos 1965). Stallman (1960) showed the basic equations for the simultaneous transfer of heat and water within the Earth and suggested that groundwater temperature measurements might provide a way of measuring rates of groundwater movement. Bredehoeft and Papadopulos (1965) derived type curves based on Stallman's equation for determining one-dimensional groundwater flux by fitting temperature data in wells.

Improvements by Stallman (1967), Sorey (1971), Cartwright (1979), Boyle and Saleem (1979), and Taniguchi (1993) confirmed that vertically moving groundwater transports heat by advection and causes curvature in the Earth's thermal profile. Additionally, Lu and Ge (1996) presented a method to evaluate the temperature effects of horizontal groundwater flow in semi-confining layers. Dimensionless plots of the temperature distribution in wells can be matched with published type curves to obtain solutions for vertical groundwater velocities, if the thermal conductivity of the solid-fluid complex is known or can be estimated. Equipment is now available for measuring temperatures in wells to within ± 0.01°C. Where temperatures in the well are in equilibrium with the surrounding aquifer or aquitard, it is possible to determine vertical groundwater flow from thermal profiles.

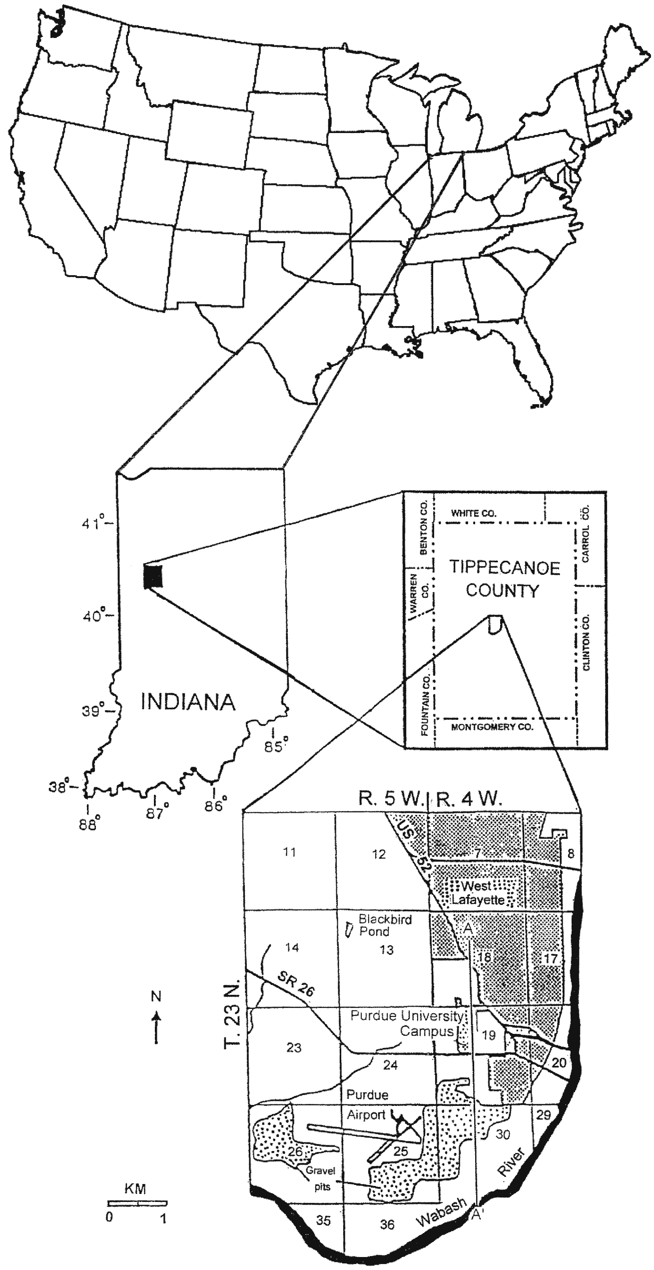

Fig. 1 Location of the study area

Study Area

The study area is located in the United States of America, specifically in the west-central part of the state of Indiana. Indiana is a state located in the Midwestern region of the country (Fig. 1). The area of study covers the northwestern part of Tippecanoe County, mainly the city of West Lafayette, which is where Purdue University is situated. The study area is about 140 square kilometers, bounded on the east and the south by the Wabash River. Purdue is located on the southern edge of West Lafayette (Fig. 1). Most of the rest of the area consists of agricultural and forested land and small housing developments.

The geology of this area has been described in several reports (Pohlmann 1987). The whole area of study is located within a major preglacial bedrock valley known as the Teays-Mahomet Valley (Bruns et al. 1985) and it features a thick sequence of glacially derived unconsolidated deposits that overlie bedrock. Figure 2 shows a geologic cross section through the Wabash Valley that runs N/S. The most predominant deposits are coarse-grained outwash sands and gravels, and fine-grained silty clay tills. The lithology of the area is quite complex because of repetitive retreats and advances of two different ice lobes between Kansan and the late-Wisconsinan time. The uppermost bedrock within the local study area are

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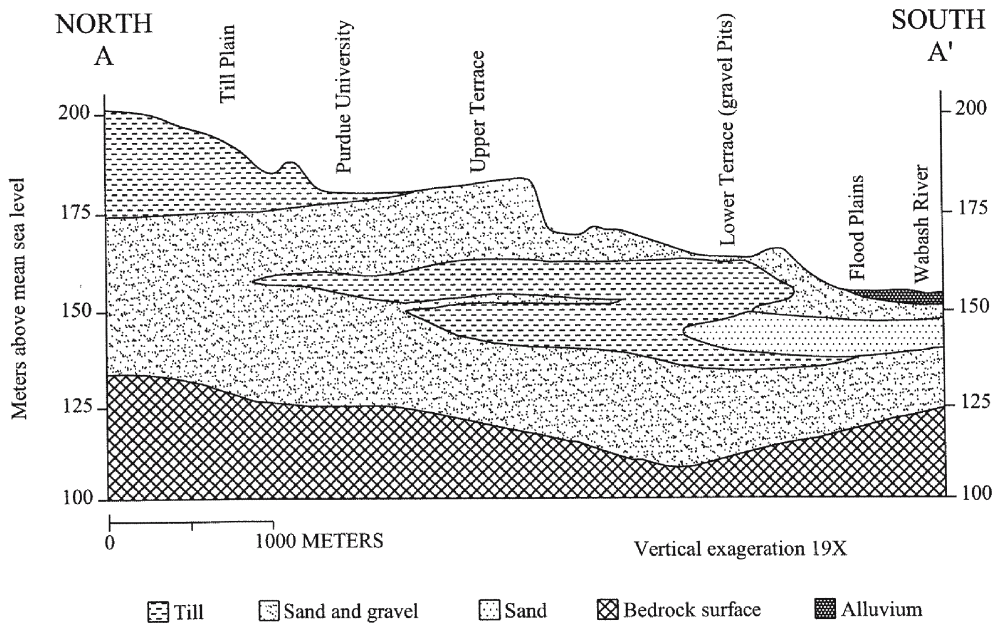

Fig. 2 Geologic cross section through the Wabash Valley (Pohlmann 1987)

primarily Devonian and Mississippian age New Albany Shale, characterized by dark, and organic-rich pyritic shale and mudstone.

The principal source of groundwater for West Lafayette is located in the thick pre-Wisconsinan and Wisconsinan sand and gravel outwash, which directly overlies the bedrock. The primary local sources of recharge to the outwash aquifer in the area of study are vertical infiltration through overlying deposits due to precipitation, and lateral underflow from aquifers to the north and west. Even though lateral underflow is the major source, vertical infiltration is a very important factor from the standpoint of potential contamination from the surface. Based on the geology of deposits between the outwash aquifer and the land surface, the major mechanisms of vertical flow to the outwash aquifer are: slow minimal infiltration through thick, low permeability till and faster infiltration through high permeability outwash sands and gravels (Pohlmann 1987).

Background

The general differential equation for 3-D simultaneous transient heat and fluid flow through isotropic, homogeneous, and fully saturated porous medium was developed by (Stallman 1960):

$$\frac{\partial^2 T}{\partial x^2}+\frac{\partial^2 T}{\partial y^2}+\frac{\partial^2 T}{\partial z^2} - \frac{c_w p_w}{k}\left[\frac{\partial(v_x T)}{\partial x}+\frac{\partial(v_y T)}{\partial y}+\frac{\partial(v_z T)}{\partial z}\right] = \frac{cp}{k}\frac{\partial T}{\partial t} \tag{1}$$

where

T= temperature at any point in time t;
c= specific heat of solid-fluid complex;
p= density of solid-fluid complex;
k= thermal conductivity;
v_x, v_y, v_z= components of fluid velocity in the x, y, and z direction;
c_w= specific heat of groundwater;
p_w= density of groundwater;
x, y, z= Cartesian coordinates;
t= time.

For the steady-state case, $\frac{\partial T}{\partial t} = 0$ (the right hand side of the equation goes to zero).

If the flow of heat and groundwater is one-dimensional (vertical), and assuming steady state with both z and v_z being positive downward, the differential equation is reduced to:

$$\frac{\partial^2 \mathrm{T}}{\partial \mathrm{z}^2} - \frac{\mathrm{c_w p_w v_z}}{\mathrm{k}}\left(\frac{\partial \mathrm{T}}{\partial \mathrm{z}}\right) = 0 \tag{2}$$

Bredehoeft and Papadopulos (1965) considered the vertical steady anisothermal groundwater flow of an aquifer system. They derived an analytical solution of the previous equation with the following boundary conditions:

$$\mathrm{T_z} = \mathrm{T_0 at z} = 0 \tag{3}$$

$$\mathrm{T_z} = \mathrm{T_L at z} = \mathrm{L} \tag{4}$$

where

T_z= temperature at any depth;

T_o= temperature at z = 0 (uppermost temperature measurement);

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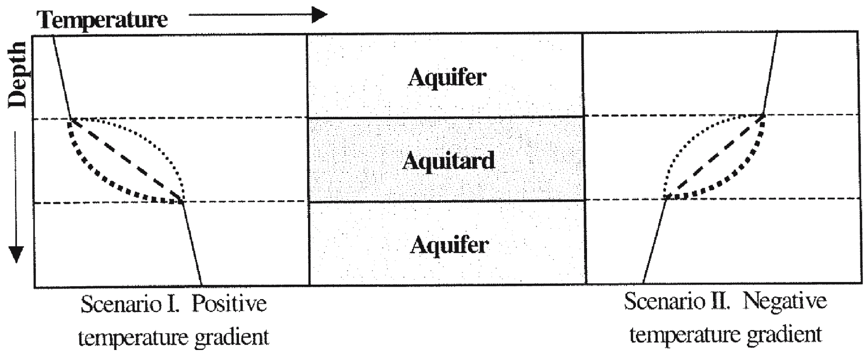
Fig. 3 Idealized temperature-depth profiles that indicate the direction of vertical leakage through an aquitard: upward, downward, and no flow



T_L= temperature at z = L (lowermost temperature measurement);
L = vertical distance over which temperatures are being observed;
v_z= vertical groundwater velocity.

After solving Eq. (2) and applying boundary conditions (3) and (4), the solution to that differential equation is:

$$\frac{T_Z - T_o}{T_L - T_o} = f(\beta, z/L) = \frac{e^{\beta(z/L)} - 1}{e^{\beta} - 1} \tag{5}$$

where

$$\beta = \frac{c_w p_w v_z L}{k} \tag{6}$$

β is a dimensionless parameter that can be positive or negative depending on whether v_z is downward or upward respectively. Bredehoeft and Papadopulos (1965) provided values of the function f (β, z / L) for a practical range of parameters and a set of type curves containing arithmetic plots for different values of β.

Ratios $(T_z - T_o) / (T_L - T_o)$ calculated from measured temperature data are plotted against the depth factor z/L at exactly the same scale as the type curves. The values are superimposed on the type curve set keeping the coordinate axes in coincidence. β is obtained from the type curve that best matches the field data curve. β is also related to the curvature of a temperature-depth profile at several depths that range from z = 0 to z = L. The thickness L is less than or equal to the thickness of the layer through which groundwater is leaking vertically. After obtaining β, it is possible to determine the groundwater velocity, which is calculated from the following relation:

$$v_z = \frac{k\beta}{c_w p_w L} \tag{7}$$

The rest of the parameters in this equation are known and v_z provides the vertical groundwater velocity. It is important to mention that under isotropic and steady-state conditions where groundwater flow does not occur (β = 0), the thermal gradient is linear with depth. However, when groundwater movement occurs, the thermal profile curves and the plot of $[e^{\beta\ (z/L)} - 1] / [e^{\beta} - 1]$ against z/L is convex upward or downward, depending on the direction of groundwater movement. The thermal profile curvature increases with groundwater velocity because v_z is proportional to β.

The thermal conductivity of soils is a function of the thermal properties of the solid materials, soil texture, pore size distribution, water content, and the temperature of the medium. The typical value for the thermal conductivity of water-saturated clays is k = 8.4 X 10^{-3} J / cm sec °C (Birch et al. 1942). The average value for the thermal conductivity of water-saturated sand and gravel outwash aquifers is k = 1.68 X 10^{-2} J / cm sec °C and for water-saturated silts is k = 8.4 X 10^{-3} J / cm sec °C (Wade 1994).

When water passes from one stratum to another with a different hydraulic conductivity, the direction of the flow path will change (Fetter 1994). Flowlines follow high permeability formations as conduits, and traverse low-permeability formations by the shortest route. In aquifer-aquitard systems with permeability contrasts of two orders of magnitude or more, flowlines tend to become almost horizontal in the aquifers and almost vertical in the aquitards (Freeze and Cherry 1979).

Boyle and Saleem (1979) presented two different scenarios related to this case. Scenario I represented conditions of positive temperature gradients while scenario II represented conditions of negative temperature gradients (Fig. 3). The variations in slopes of linear segments in a given profile correspond to changes in the thermal properties associated with various lithologies found in the system. Both scenarios contain three different possibilities for groundwater flow within the aquitard: (a) upward, (b) downward, and (c) no vertical flow. The linear segments (c) within the aquitard indicate transfer of heat by conduction only (no heat transfer is accomplished by migrating groundwater). On the other hand, the curvilinear segments (a) and (b) indicate vertical transfer of heat by upward and downward flowing groundwater, respectively.

Boyle and Saleem (1979) used a computer procedure to get values of β from the temperature-depth data observed at one-meter spacings over a specified depth interval L. The temperature values were used to compute $(T_z - T_o) / (T_L - T_o)$ as z changes from zero to L. These ratios were compared to the theoretical values of f (β, z/L) computed from the right-hand side of Eq. (5) for $0 \leq z \leq L$ and for specified values of T_o, T_L, L, and β.

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The program determined the optimum value of β by assigning an initial value to β and then adjusting it until the objective function F(β) was minimized (Saleem 1970), where

$$F(\beta) = \sum_{z=0}^{z=L} \left[\frac{T_z - T_o}{T_L - T_o} - \frac{e^{\beta(z/L)} - 1}{e^{\beta} - 1} \right]^2 \tag{8}$$

The optimum value of β was the one that yielded the minimum value of F(β), for each value of L. It is important to remember that the system resolution is 0.01°C for each temperature measurement. Since the solution represented by (5) is in the form of a temperature-difference ratio, the accuracy of the ratios and the resulting value of β is determined by the resolution.

Thermal Stability of Water Columns

Thermal stability of the water column within the well is another critical point in the application of the main equation used for this research. Groundwater temperatures obtained from the wells and used in the solution of Eq. (5) are valid only if the groundwater temperatures inside of the well represent the temperatures of the fluid-porous medium complex adjacent to the well. This is most likely to occur when groundwater within the well is thermally stable. If heat transfer within the well is accomplished through convection, the resulting temperature fluctuations will provide thermal disturbances large enough to invalidate the groundwater temperatures. Thus, convective transfer of heat in the fluid column must be of negligible magnitude.

Diment (1967), Gretener (1967), and Sammel (1968) have been studying problems encountered in obtaining representative groundwater temperatures from wells. Krige (1939) developed an expression for the critical gradient G_c of a fluid-filled column:

$$G_c = \frac{ga\theta}{c} + \frac{C\mu\alpha}{gar^4} \tag{9}$$

where

g = acceleration due to gravity (cm / sec^2);
θ = absolute temperature (°K);
a = volume coefficient of thermal expansion (1 / °C);
α = thermal diffusivity (cm^2 / sec);
μ = kinematic viscosity of the fluid (cm^2 / sec);
r = radius of the column (cm);
c = specific heat of groundwater at constant pressure (ergs / gm °C);
C = a constant, which is equal to 216 cgs units;
G_c = critical thermal gradient in groundwater (°K / cm).

Sammel (1968) defined the critical gradient G_c as the rate of temperature change with depth at which convective flow is incipient. A case in which temperature increases with depth, i.e., positive temperature gradient, is thermally unstable when the actual temperature gradient exceeds the critical gradient. In situations where negative gradients exist, G_c cannot be exceeded and thermal stability is maintained. By using Krige's formula, critical thermal gradients were obtained from all monitoring wells used in this research, and compared to their actual thermal gradients. Each thermal gradient is defined as:

$$GT = \frac{(T_{Z2} - T_{Z1})}{(Z_2 - Z_1)} \tag{10}$$

where

GT = thermal gradient (°C/m);
T_{Z2} = groundwater temperature at depth Z_2, deeper than Z_1 (°C);
T_{Z1} = groundwater temperature at depth Z_1 (°C);
Z_2 = a certain depth within the geological stratum, deeper than Z_1 (meters);
Z_1 = a certain depth within the geological stratum (meters).

If the thermal gradient exceeds the critical gradient, then that means the groundwater temperature inside of the well is not representative of the temperature of the water-porous medium system next to the well.

Solver

Microsoft Excel Solver is an optimization modeling system that combines the functions of a graphical user interface (GUI), an algebraic modeling language, and optimizers for linear, nonlinear, and integer programs. Each function is integrated into the host spreadsheet program. Optimization begins with an ordinary spreadsheet model. The spreadsheet's formula language functions as the algebraic language used to define the model. Through the Solver's GUI, the user specifies an objective and constraints. Solver then analyzes the complete optimization model and produces the matrix form required by the optimizers (Lasdon et al. 1998).

Microsoft Excel Solver employs the Generalized Reduced Gradient (GRG2) Algorithm for optimizing nonlinear problems and uses the solution values to update the model spreadsheet. Solver uses iterative numerical methods that involve plugging in trial values for the adjustable cells and observing the results calculated by the constraint cells and the optimum cell. Each trial is called an iteration. Because a pure trial and error approach would take such a long time (specially for problems involving many adjustable cells and constraints), Microsoft Excel Solver performs extensive analyses of the observed outputs and their rates of changes as the inputs are varied, to guide the selection of new trial values.

In a typical problem, the constraints and the optimum cell are functions of (that is, they depend on) the adjustable cells. The first derivative of a function measures its rate of change as the input is varied. When there are several values entered, the function has several partial

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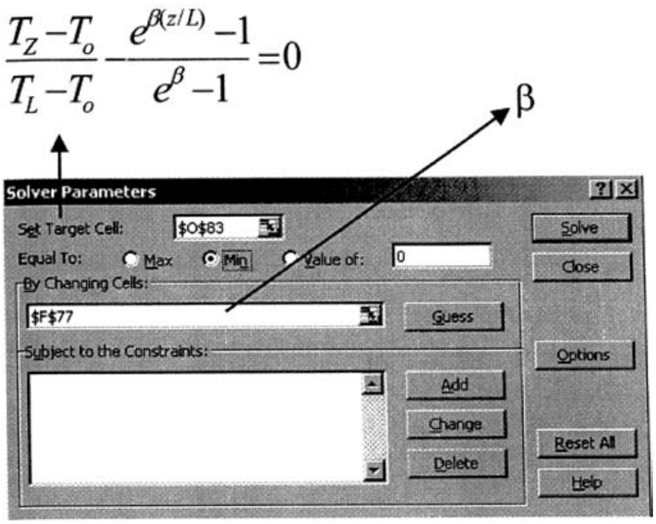

Fig. 4 The Solver Parameters dialog box

derivatives measuring its rate of change with respect to each of the input values; together, the partial derivatives form a vector called the gradient of the function. Derivatives and gradients play a crucial role in iterative methods in Microsoft Excel Solver. They provide clues as to how the adjustable cells should be varied.

The main equation used to determine the groundwater velocity is the following:

$$\frac{T_Z - T_o}{T_L - T_o} - \frac{e^{\beta(z/L)} - 1}{e^{\beta} - 1} = 0 \qquad (11)$$

In order to use Microsoft Excel Solver, the Solver add-in must be downloaded into the system. Typically, this feature is not installed by default when Excel is first setup on the hard disk. To add this facility to the Tools menu, it is necessary to select the menu option Tools → Add-Ins and then choose the Solver Add-In checkbox. Add-ins are Excel worksheets that have been saved as Microsoft Excel Add-Ins (.xla).

Solver provides a less complex method for finding the values needed to obtain the desired results because it has the ability to change the value in multiple cells. Once Solver is used, the Solver Parameters dialog is displayed as shown in Fig. 4. In the Solver Dialog Box, there are several functions: set target cell, by changing cells, subject to constraints, options, and solve. Set target cell specifies the solving target value. This cell must contain a formula and it can be selected by either typing the cell reference in the field or selecting the Collapse Dialog button and clicking on the cell. The target cell used is the left-hand side of Eq. (11). Once the target cell is specified, the value of the cell is indicated by either setting it to the maximum possible size (Max), minimum size (Min), or a specific value. According to Eq. (8), the target cell is minimized to zero.

By changing cells indicates the cells whose values Solver needs to modify in order to obtain the specified results for the target cell. It is possible to specify a maximum of 200 cells. Solver can also automatically select the cells based upon the cells referenced in the target cell formula by selecting the Guess button. The cell that is being constantly modified until the target cell is minimized to zero is the cell that contains the β value.

Subject to constraints creates any limitations applied to the changing cells or the target cell. No constraints were specified for any of the cells. Options refine the process by which Solver performs a series of internal calculations to derive values for cells that produce the desired result. Solver attempts to find the results that most closely match the target value and places the results directly in the appropriate cells in the worksheet.

Well Selection

The selection of wells was accomplished based on well location and well log availability. The Indiana Department of Natural Resources (IDEM), Division of Water, Groundwater Section provided information about the existing wells in West Lafayette. The well logs were obtained from Pohlmann (1987), Weinreb (1987), and Harvey (1990). The 12 monitoring wells were selected based on the following criteria: spatially distributed throughout the study area, various depths, not equipped with pumps, screened through different geologic formations, good accessibility, and geologic log or driller's log on record. Figure 5 shows a detailed map of the location of the wells.

Results and Discussion

The Solinst Reelogger (a portable, reel-mounted configuration using a flat tape marked at every meter and millimeter) with a temperature probe sensor attached on one end was used for measuring the groundwater temperatures. The instrument was fitted with a semiconductor thermistor capable of reading temperatures with an accuracy of 1×10^{-2} °C. The temperature-sensing unit was carefully lowered into the well in order to minimize a possible disruption of the actual thermal conditions. In the top 5 m of each monitoring well, groundwater temperatures were recorded at every 1-m interval; below this depth, temperatures were logged with intervals of 2 m. Groundwater temperatures were taken at each depth after the temperature probe was stabilized.

Changes in groundwater temperatures with depth were monitored and recorded in each of the monitoring wells at three different times of the year (between September 2001 and April 2002) in order to obtain seasonal vertical groundwater velocities. The temperature-depth profiles provided continuous temperature information for the entire water column of each well. Once groundwater temperatures were collected, groundwater thermal gradients were analyzed and compared to their critical thermal gradients. Figure 6 shows charts for monitoring wells PE-2 and PE-3 where thermal gradients did not exceed

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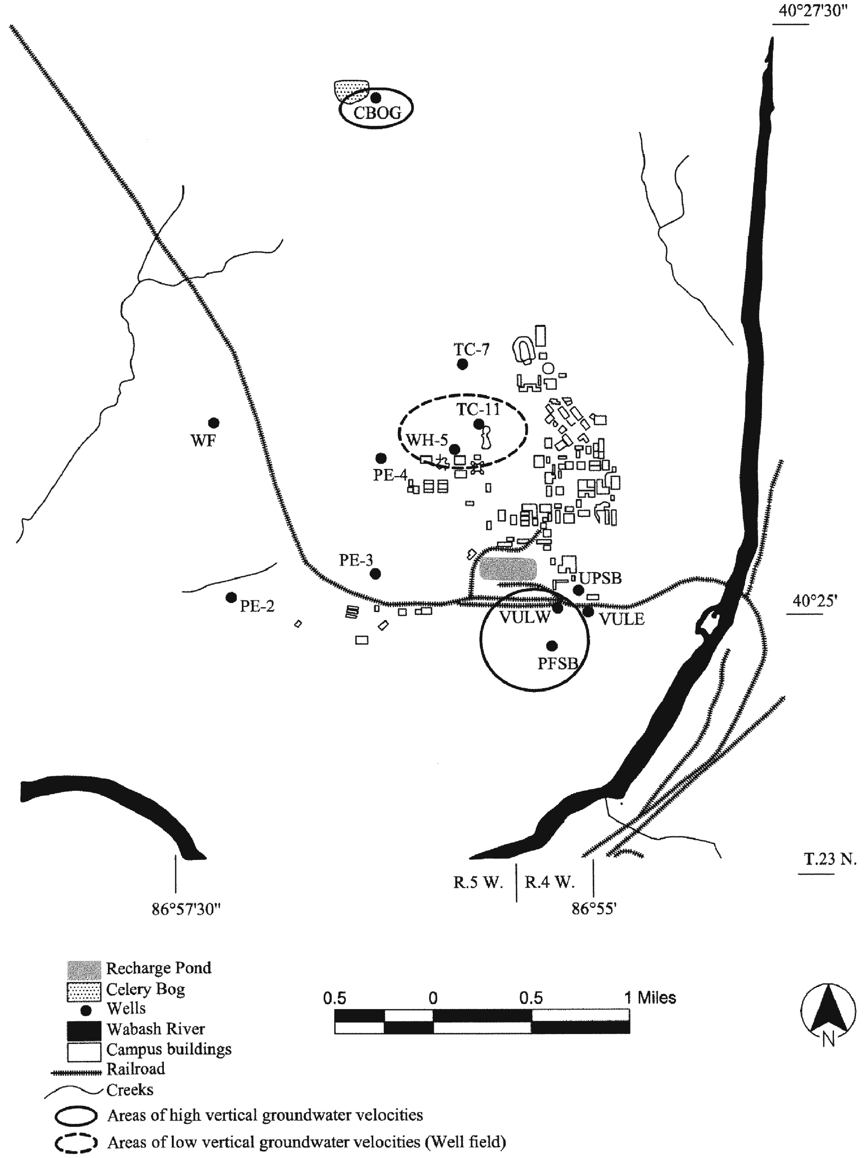
Fig. 5 Well location map



their respective critical gradients. Complete and detailed groundwater thermal gradient analyses in each monitoring well were performed and none of the calculated ther-mal gradients were above their critical gradients, mean-ing groundwater temperatures were representative of the water-porous medium systems adjacent to the wells (Arriaga 2002).

After thermal gradients were calculated, Microsoft Excel Solver computed the β values. The general trend of β values, which depends on both v_z and L, was considered to be indicative of the vertical component of the groundwater flow through the aquitard or aquifer; the larger the β value, the larger the vertical groundwater velocity. Positive β values represent downward leakage while negative values correspond to upward movement. Table 1 summarizes the seasonal β values calculated for each monitoring well. All of the wells presented positive β values, meaning vertical groundwater velocities are moving downward.

For validation purposes, the groundwater temperatures used to calculate the β values obtained from Solver were superimposed and matched on the set of type curves of the function f (β, z / L). By comparing the β values obtained from the curve matching to the Solver β values, it is possible to check the accuracy of Solver in determining β values. Figure 7 shows type curve matching for some geological strata in monitoring wells PE-2 and PE-3. Extensive type curve matching analyses were performed in all of the monitoring wells and the β values turned out to be exactly the same as the ones derived from Solver (Arriaga 2002).

Once Solver computed the β values, groundwater velocities were calculated by using Eq. (7). Table 2 summarizes the seasonal vertical groundwater velocities found for each corresponding β value. Overall, the average vertical groundwater velocity (including aquifers and aquitards) found in the 12 monitoring wells during the Fall, Winter, and Spring was 1.91 cm/year. The outwash

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Table 1 Summary of seasonal β values

Well ID	General hydrogeologic setting	Fall	Winter	Spring	Average
PE-2	Upper Aquifer	0.245	0.245	0.34	0.277
	Aquitard	1.101	1.048	1.262	1.137
	Lower Aquifer (Upper Part)	2.142	1.482	2.284	1.969
	Lower Aquifer (Lower Part)	1.184	0.779	1.227	1.063
PE-3	Upper Aquitard	1.059	0.995	1.205	1.086
	Upper Aquifer	1.952	1.908	2.233	2.031
	Lower Aquitard	1.351	1.243	1.243	1.279
	Lower Aquifer	1.283	1.016	1.162	1.154
PE-4	Aquifer (Upper Part)	1.436	1.444	1.791	1.557
	Aquifer (Lower Part)	1.356	1.376	1.707	1.48
TC-11	Aquifer	2.238	–	–	2.238
WH-5	Aquifer	2.07	2.123	2.474	2.222
TC-7	Aquitard	2.49	2.49	2.49	2.49
WF	Aquifer	0.367	0.34	0.426	0.378
CBOG	Aquitard	1.854	1.854	2.156	1.955
	Aquifer	2.694	2.694	2.683	2.69
VULW	Aquifer	0.548	0.487	0.548	0.528
PFSB	Aquifer	1.952	1.974	2.41	2.112
VULE	Aquifer (Upper Part)	1.874	1.649	2.055	1.859
	Aquifer (Lower Part)	0.951	0.9	1.05	0.967
UPSB	Aquitard	2.717	2.514	2.88	2.704
	Aquifer	1.42	1.348	1.499	1.422

Note: – β value not obtained due to well abandonment

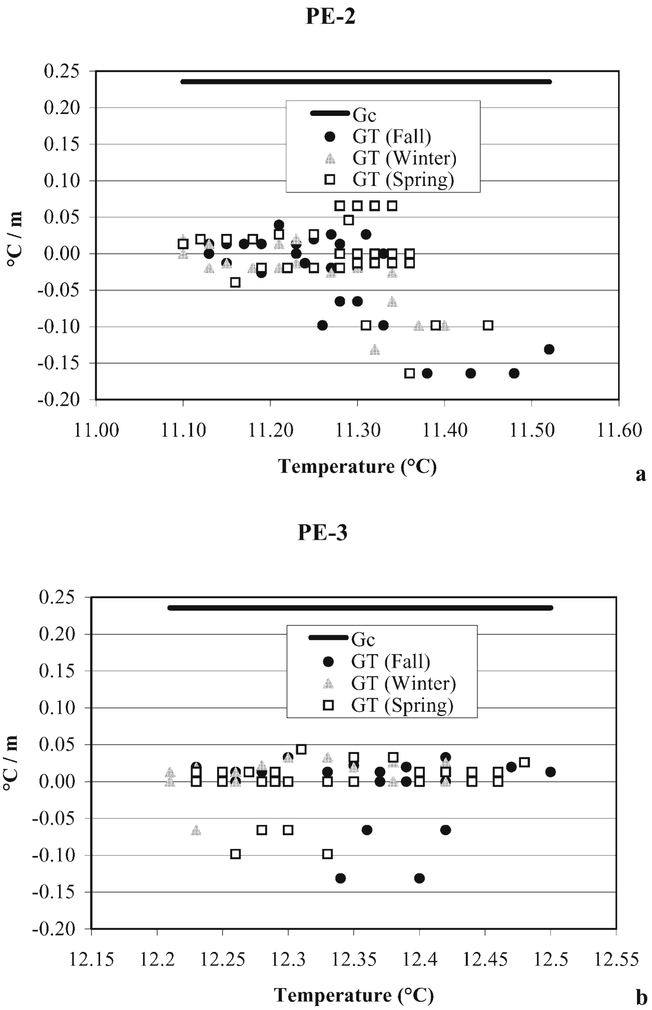

Fig. 6a,b Critical gradients for monitoring wells: (**a**) PE-2, and (**b**) PE-3

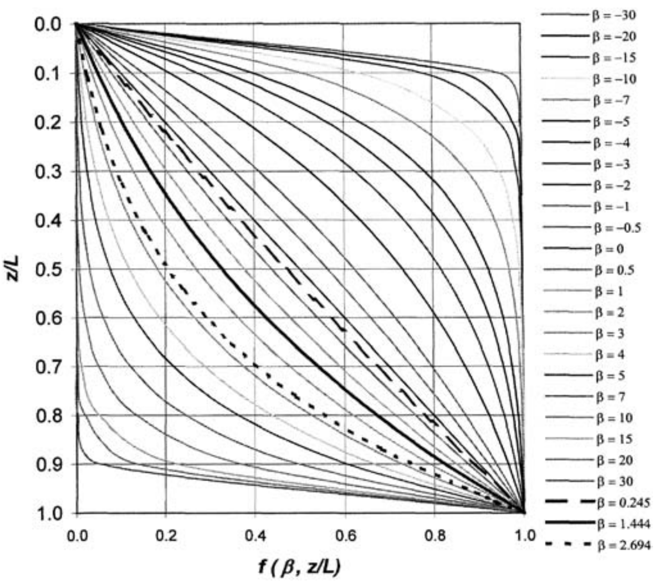

Fig. 7 Type curve matching of the function f(β, z/l) with different β values

aquifers had an average vertical groundwater velocity of 1.83 cm/year. The minimum vertical groundwater velocity in the aquifers was 0.92 cm/year and obtained during the Winter in well PE-2, whereas the maximum vertical groundwater velocity was 4.53 cm/year, which was recorded during the Spring in well VULW. The aquitards had an average vertical groundwater velocity of 2.08 cm/year. The minimum vertical groundwater velocity in the aquitards was 1.49 cm/year and obtained during the Winter in well UPSB, whereas the maximum vertical groundwater velocity was 2.97 cm/year, which was recorded during the Spring in well CBOG.

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Table 2 Summary of seasonal vertical groundwater velocities

Well ID	General hydrogeologic setting	Fall (cm/yr)	Winter (cm/yr)	Spring (cm/yr)	Average (cm/yr)
PE-2	Upper Aquifer	2.53	2.54	2.81	2.63
	Aquitard	2.28	2.17	2.61	2.35
	Lower Aquifer (Upper Part)	1.47	1.02	1.58	1.36
	Lower Aquifer (Lower Part)	1.40	0.92	1.45	1.26
PE-3	Upper Aquitard	2.43	2.29	2.77	2.77
	Upper Aquifer	2.02	1.97	2.31	2.10
	Lower Aquitard	1.86	1.71	1.71	1.76
	Lower Aquifer	1.44	1.14	1.30	1.29
PE-4	Aquifer (Upper Part)	1.19	1.17	1.45	1.27
	Aquifer (Lower Part)	1.12	1.14	1.41	1.22
TC-11	Aquifer	1.34	–	–	1.34
WH-5	Aquifer	1.22	1.22	1.42	1.29
TC-7	Aquitard	1.78	1.78	1.78	1.78
WF	Aquifer	1.27	1.17	1.47	1.30
CBOG	Aquitard	2.56	2.56	2.97	2.70
	Aquifer	2.42	2.42	2.41	2.42
VULW	Aquifer	3.78	4.03	4.53	4.11
PFSB	Aquifer	2.89	2.82	3.32	3.01
VULE	Aquifer (Upper Part)	1.55	1.36	1.68	1.53
	Aquifer (Lower Part)	1.41	1.33	1.55	1.43
UPSB	Aquitard	1.56	1.49	1.66	1.57
	Aquifer	1.37	1.30	1.44	1.37

Note: – Groundwater velocity not obtained due to well abandonment

Vertical groundwater velocities were greater in the Spring, whereas the lowest ones were obtained during the Winter. Overall, the aquitards had greater vertical groundwater velocities than the aquifers, which makes sense because according to the law of tangents flow in clay-rich aquitards tends to be mostly vertical while flow in outwash aquifers is predominantly horizontal (Fetter 1994). The groundwater velocities obtained in this research provided reasonable values in agreement with those of Arihood (1982) and Daniels et al. (1991).

Two sources of recharge were detected across the study area: (1) the gravel pit recharge pond and (2) the Celery Bog. Wells located close to these recharge source areas presented the highest vertical groundwater velocities (see Fig. 5). On the other hand, wells located next to the Purdue University well field (PUWF) had the lowest vertical groundwater velocities (see Fig. 5). The PUWF is located in the area south of monitoring well TC-11 and east of well WH-5 and is operated by the Purdue University water works. This well field is considered a sink because it draws great amounts of groundwater from its nine high-capacity water wells (housed in six well houses), each capable of pumping in excess of 1,000 gallons per minute

According to Bense and Kooi (2004), the seasonal temperature variations for shallow depths can be explained partially by the seasonal fluctuations of temperatures at the ground surface. Seasonal temperature-depth profiles were obtained from each of the wells (Arriaga 2002). Figure 8 shows the seasonal temperature-depth profiles for wells PE-2 and PE-3. Variations in groundwater temperatures occurred mainly in the upper part of the zone of saturation (close to the static water level), where temperatures were most subject to changes due to weather conditions. However, after reaching some depth, temperatures from one season to the next were very similar. This depth varied from well to well, but on average groundwater temperatures became constant three to seven meters below the static water level.

Summary and Conclusions

Groundwater temperature data was obtained from 12 monitoring wells located in the Purdue University Campus at three different times of the year, in Fall 2001, Winter 2001, and Spring 2002. The selection of these wells was based on well location and well log availability. These monitoring wells provided estimates of vertical groundwater velocities through various geologic strata in the zone of saturation based on groundwater thermal gradients. Groundwater temperatures in the wells were shown to be thermally stable because the groundwater thermal gradients did not exceed their critical gradients. The resulting temperature fluctuations did not produce thermal disturbances large enough to invalidate the assumption that the groundwater temperatures inside the wells represented the temperature of the fluid-porous medium complex adjacent to the well.

Microsoft Excel Solver provided accurate results to optimize the nonlinear equation used in this research and to obtain solutions for vertical groundwater velocities. Because a pure trial and error approach would take such a long time, Solver performed extensive analyses of the observed outputs to guide the selection of new trial values and to find a final solution. Dimensionless plots of groundwater temperature distribution were matched with published type curves to validate the β values obtained from Solver. Vertical groundwater velocities ranged from 0.92 cm/year to 4.53 cm/year. The lowest velocities were

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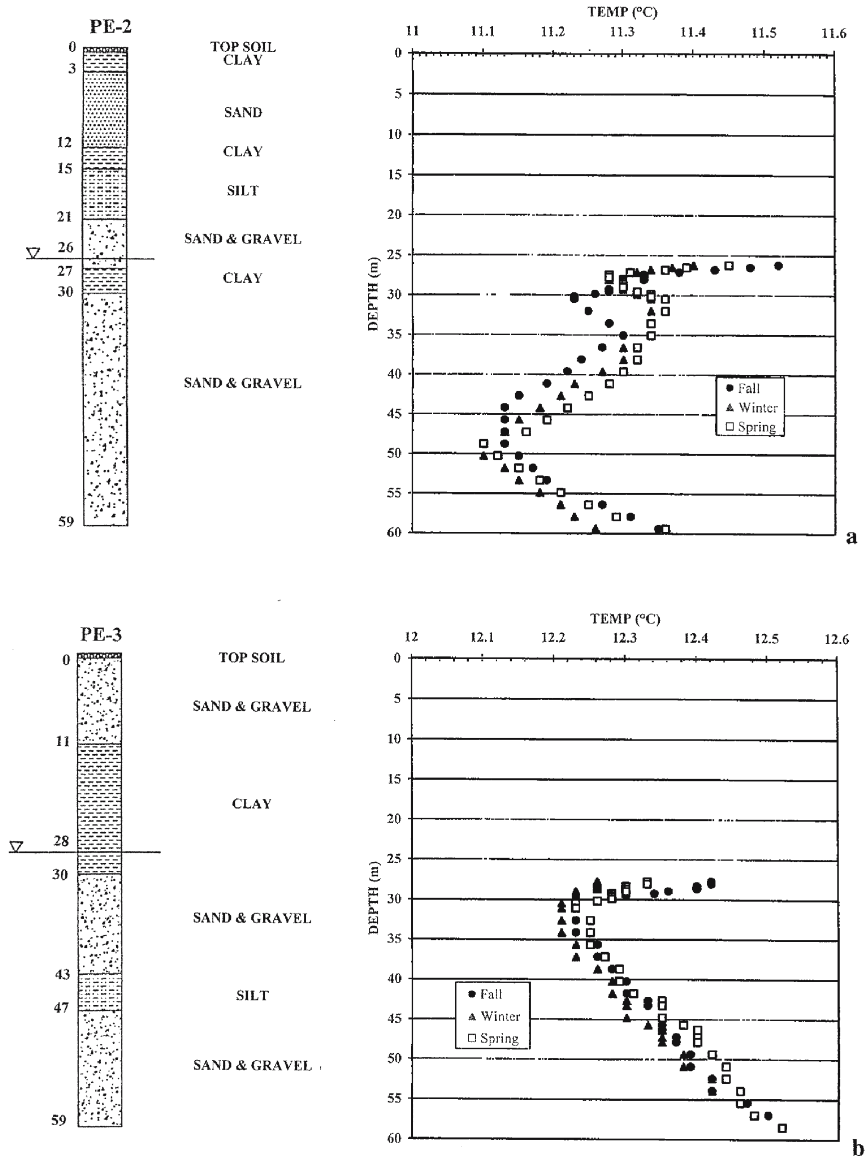

Fig. 8a,b Seasonal temperature-depth profiles for monitoring wells: (**a**) PE-2, and (**b**) PE-3

detected during the Winter and the highest occurred in the Spring.

Clay-rich tills presented greater vertical groundwater velocities than outwash aquifers, which is reasonable because according to the law of tangents when groundwater moving downward from a medium with high hydraulic conductivity (outwash aquifer) passes through a zone of low hydraulic conductivity (clay aquitard), flow tends to remain vertical. Monitoring wells located near the recharge source areas (the gravel pit recharge pond and the Celery Bog) had the highest vertical groundwater velocities. By contrast, wells located next to the Purdue University well field (one of the major pumping center in West Lafayette) presented the lowest vertical groundwater velocities.

This groundwater temperature profile technique is not intended to supplant other applicable methods such as numerical modeling of water-budget data in regional studies because the temperature profiles are very local to extrapolate to larger scale. The results of field investigations described in this research show that under favorable conditions, the temperature distribution in monitoring wells can be used in conjunction with Solver to determine vertical groundwater velocities throughout various glacially derived materials.

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