A PRACTICAL GUIDE FOR ESTIMATING RECHARGE FROM WATER TABLE HYDROGRAPHS

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ABSTRACT

A water table level in an unconfined aquifer fluctuates due to recharge, discharge and horizontal groundwater flow. A water table hydrograph represents the fluctuations of the water table due to these processes with respect to time. Therefore by analysing a hydrograph, it is possible to estimate net recharge to a water table aquifer. This technical memorandum presents the general guidelines for installation of a piezometer and construction of a water table hydrograph. Methods to estimate recharge from a single water table hydrograph and from hydrographs for a network of piezometers are presented. This memorandum is intended for field practitioners.
Introduction

An unconfined aquifer or a watertable aquifer, is an aquifer in which a watertable forms the upper boundary. The change in watertable level at a certain position over time can be plotted on a graph; this is called a watertable hydrograph. A typical watertable hydrograph is presented in Fig. 1.

Watertable hydrographs are generally smoother than surface water hydrographs. A watertable hydrograph reflects the net result of recharge and discharge, to and from a watertable. It also aggregates the effect of spatial variability on water movement processes within the soil profile. Watertable level in an unconfined aquifer fluctuates due to recharge, discharge and horizontal groundwater flow. When horizontal groundwater flow is negligible, the recharge or discharge at a location can be estimated from a watertable hydrograph from the same location. When horizontal groundwater flow is significant, watertable hydrographs from a network of piezometers are required to estimate recharge from a location.

The magnitude of watertable fluctuation due to recharge or discharge will depend on the specific yield of the watertable aquifer. The specific yield is defined as the depth of water that an unconfined aquifer releases from storage per unit decline in the watertable level. Depending on the texture, bulk density and pore structure of the aquifer material, it varies from 0.01 to 0.3.
Figure 1 A watertable hydrograph monitored under an onion crop

Measuring changes in watertable level

A piezometer or an observation well can be used to monitor watertable fluctuations at a particular location. A piezometer is a vertical tube slotted only at the point of intake of water within the aquifer. An observation well is also a vertical tube, but slotted throughout its length. Either an observation well or a piezometer can be used to monitor the fluctuation of the watertable. However, when an observation well is used in an irrigation bay, care must be
taken to prevent direct entry of irrigation water into the well. Furthermore, if a less permeable layer exists between the soil surface and the watertable aquifer, the observation well must penetrate the less permeable layer fully and enter the watertable. Otherwise, the observation well reading will reflect the perched watertable level above the less permeable layer. It is difficult to generalise the radius of influence of a piezometer. Ideally, the watertable behaviour monitored by a piezometer should be representative of the soil, aquifer material and the land use practices in the area of interest. This is essential especially when groundwater flow is ignored and recharge is estimated from a single piezometer.

**Construction and installation of a piezometer**

A piezometer can be constructed from a PVC water pipe (e.g. 80 mm Class 6) of required length. The bottom of the pipe may be capped if the aquifer material is of a swelling nature. The lower part of the PVC pipe (e.g. 300 mm) is then slotted at an interval of 25 mm. This facilitates free entry of groundwater into the pipe. The water level in the piezometer can be monitored manually or with a depth sensor and a data logger. Installation of a piezometer can be achieved as follows:

1. Bore a hole to the required depth with a jarrod auger of 120 mm diameter.

2. Insert the piezometer into the hole and backfill the gap between the pipe and the side of the hole, until the slotted area is covered with pea gravel of 5 mm diameter.
3. Lay approximately 50 mm of sand on top of the pea gravel. Steps 3 and 4 will prevent the slot being contaminated by bentonite from above.

5. Lay approximately 100 mm of bentonite on top of the pit sand.

6. Fill the remaining depth with a 1:3 bentonite : oven dried clay mixture. Steps 5 and 6 will prevent water flowing from upper layers into the slot.

7. Install a depth sensor into the piezometer.

Various components of a piezometer are presented in Fig. 2.

**Determining specific yield**

Traditionally, the specific yield of a watertable aquifer is determined during pump tests (Freeze and Cherry 1979). These values will reflect the specific yield of the aquifer layer at which water is drawn by the pump. The specific yield can also be estimated from the texture of the aquifer material (Davis and De Wiest 1966). When the watertable is near the soil surface (e.g. within 2 m of the soil surface), the specific yield of the soil layers near the surface need to be determined. The specific yield of shallow soil layers can be determined either from a field experiment or from a lysimeter experiment.
Fig. 2. Components of a piezometer.
Field experiment:

A field experiment to determine the specific yield can be set up in the following way:

1. Construct a bay large enough to minimise the variability of soil physical properties. A bay of 10 m X 10 m surrounded by a 1 metre wide buffer is appropriate in general circumstances.

2. Install a piezometer long enough to enter the existing watertable and include a depth sensor and a logger to continuously monitor the watertable fluctuation.

3. Install multiple (e.g. 3) neutron access tubes within the bay to monitor the change in soil water content. The location of neutron access tubes within the bay should reflect any soil variability present within the bay.

4. Cover with a plastic sheet to prevent evaporation from the bay and the buffer zone during the experiment.

5. Flood the bay and buffer zone until the watertable is brought to, or near, the soil surface. If the hydraulic conductivity of the surface layer is lower than the hydraulic conductivity of the deeper layers, the piezometer level will not reach the soil surface. Under such circumstances, addition of water into the bay should be stopped when the piezometer level stops rising.

6. Monitor the change in volumetric soil water content due to drop in watertable level with time. If the watertable drops rapidly, daily measurements are recommended.
A typical layout of the experimental site is given in Fig. 3.

Figure 3 A layout of an experimental plot to determine specific yield

Fig. 4 shows the change in watertable depth when monitored continuously following flooding in Beelbangera clay. Since the bay and the buffer zone were
covered to prevent evapotranspiration, the reduction in watertable level is due to horizontal groundwater flow and *leakage* to deeper aquifers. Changes in volumetric soil-water content of various layers, from their saturated values for the same soil, are presented in Fig. 5. The soil water content of a layer decreases due to decline in watertable level. Therefore, the maximum change in volumetric soil water content of a layer from its saturated level is equal to its specific yield.

![Graph showing change in depth to watertable over time](image)

*Fig. 4. Change in depth to watertable in a Beelbangera clay.*
Fig. 5. Change in volumetric soil water content from saturated level of soil layers at 0.1 m and 0.3 m of Beelbangera clay. (Note the change becomes asymptotic with time).

The primary advantage of this method is that a relatively large surface area is used, therefore, the effect of soil variability is minimised. Using this method, the volume of water drained from a layer is estimated from its saturated soil water content and the soil water content after draining, rather than directly measuring the drainage water. This method is therefore an indirect one to measure specific yield. As a result, the error associated with neutron probe calibration will influence the estimated specific yield. This method is also restricted to determining the specific yield between the depth to which the watertable can be raised due to flooding and to the depth to which the watertable will fall due to groundwater flow and leakage. The watertable
cannot be brought to the surface if the existing watertable is deep and the infiltration rate is less than the groundwater flow and leakage.

**Lysimeter experiment:**

The following procedure could be used to conduct a lysimeter experiment to measure the specific yield.

1. Obtain a soil core from a representative soil site. To minimise the effect of soil variability, multiple cores are recommended.
2. Install drainpipes into the side of the core at 0.1 m depth increments from the surface.
3. Attach a mariotte tank to the base of each core to establish and maintain a static watertable at desired depth. Raise the watertable level in the core by adding water through the mariotte system until it reaches the soil surface.
4. Cover the soil surface with plastic to prevent evaporation and turn off the tap at the base connecting lysimeter to mariotte tank.
5. Open the top drainpipe along the side of the core and collect the drainage water. Open the second drainpipe only after the first one has stopped draining, and collect drainage water from the second layer. Repeat this procedure until water is drained through all the drainpipes.
6. Estimate the specific yield of each layer by dividing the volume of water drained from the layer by the cross-sectional area of the lysimeter core and the thickness of the layer.
\[ S_y = \frac{\text{Volume of water drained (m}^3\text{)}}{\text{Cross sectional area (m}^2\text{)} \times \text{Layer thickness (m)}} \]  \hspace{1cm} (1)

A typical lysimeter set up to determine the specific yield is presented in Fig. 6. A similar procedure could be adopted to determine specific yield from smaller soil cores in a laboratory. Since the specific yield is measured directly by measuring the drainage, the lysimeter method is more accurate than the field method. However, there is less confidence that the results will be representative of the larger area of interest because of soil variability. In addition the method is restricted to shallower depths (e.g. 1.5 m) due to limitations in the size of the core obtainable.

Miriams (1992) measured the specific yield of Beelbangera Clay in three lysimeter cores. The measured values at a depth of 0.2-0.3 m, were 0.002, 0.01 and 0.06. Drainable porosity measured by the field method for the same layer was 0.03 (Peters et al. 1992).

**Estimating recharge from a single watertable hydrograph**

Since the rate of horizontal groundwater flow in an unconfined aquifer is generally small (say 0.05 m d\(^{-1}\)), a single hydrograph will be adequate to estimate recharge following an irrigation event. Under such circumstances, the recharge (or discharge) during a particular period of time to the watertable aquifer could be estimated by,
Fig. 6. A lysimeter set up to determine specific yield.
\[ \text{Recharge} = \Delta \text{watertable level} \times \text{Specific yield} \] (2)

In soils with cracks, a sharp rise in watertable level may be observed during, and immediately after, irrigation. This is due to bypass flow through the soil cracks. However, such a rise may disappear in a short period of time due to water movement into the soil matrix.

In addition to recharge or discharge mechanisms, the piezometer level observed in a watertable aquifer, in an irrigated environment, is influenced by diurnal and seasonal changes in atmospheric pressure, and by air entrapment during rain or irrigation.

**Atmospheric pressure and piezometer level**

Changes in atmospheric pressure are not always transmitted to the watertable immediately, because the air permeability of the unsaturated zone is generally small. However, such changes are immediately transmitted to the water level in the piezometer. Therefore, when atmospheric pressure increases, the piezometer level will be lower than the watertable level. Similarly a decrease in the atmospheric pressure may cause the piezometer level to rise. A 0.1 kPa change in barometric pressure can cause the piezometer level to temporarily deviate 0.14 m from the true watertable level (Bouwer 1978).
Atmospheric pressure and watertable level

Changes in atmospheric pressure can also lead to a change in watertable level (Peck 1960; Turk 1975). When the atmospheric pressure increases, the soil air is compressed and reduced in volume. Water from the watertable will flow upwards to compensate for the 'vacuum' created. Thus the watertable will fall when the atmospheric pressure is increased. The movement of water from the watertable when the atmospheric pressure is increased could be explained by soil water potential theory as well. When the soil air pressure is increased, the soil water potential should also be increased for the soil-air-water interface to be in an equilibrium. At higher soil water potentials, the soil matrix will hold more water; this 'extra' water will flow from the free watertable as capillary upflow.

Change in soil air pressure could also be caused by compression of entrapped air during flood irrigation. Thus, the change in watertable level during irrigation is a net effect of recharge to the watertable and soil air entrapment. The increase in soil air pressure due to entrapment, in a soil with well developed macropores, will be negligible.

Estimating recharge from a network of piezometers

When the horizontal groundwater flow is significant over a period of time and space, it is feasible to determine recharge to a watertable aquifer if the watertable hydrographs and the aquifer properties (transmissivity and specific yield) are available. In conventional groundwater modelling, the aquifer
response (change in watertable level) is simulated for a specified recharge. The procedure adopted to estimate recharge from change in watertable level and aquifer parameters is known as inverse modelling. Review of various inverse modelling techniques to determine recharge from watertable hydrographs is found in Yeh (1986). One such technique using linear programming is described below.

In this technique, the region of interest is initially subdivided into a number of cells based on the variability of the soil, hydrogeology and land use practices. Subsequently, the recharge during a particular time step in a cell is determined by predicting the watertable level at the end of the time step \( h_{p_{ij}}^{n+1} \) using the present observed level \( h_{o_{ij}}^{n} \) and aquifer properties, and by minimising the difference between \( h_{p_{ij}}^{n+1} \) and the observed level at the future time step \( h_{o_{ij}}^{n+1} \). Hence the objective function of the inverse model is to minimise

\[
    z = \sum_{i=1}^{J} \sum_{j=1}^{J} d_{ij}^+ + d_{ij}^-
\]  

(3)

where,

- \( z \) : objective variable (L)
- \( d^+ \) : positive over achievement variable (L)
- \( d^- \) : positive under achievement variable (L)
- \( i \) : number of rows in grid (i=1,2,...,I)
- \( j \) : number of columns in grid (j=1,2,...,J)

The over and under achievement variables in eq. 3 are defined as
\[ h_{p_{ij}}^{n+1} - d_{ij}^* + d_{ij}^- = h_{p_{ij}}^{n+1} \]  \hspace{1cm} (4)

where,

\( h \) : watertable level above AHD (L)

This procedure is referred to as goal programming in the literature, where the objective is to minimise the sum of deviations of predicted values from the corresponding 'target' values.

In addition to eq. 3, the inverse model should include an equation (as a constraint) to account for groundwater flow and recharge. Groundwater flow and recharge to a watertable aquifer, if the change in watertable level is small compared to the thickness of the aquifer, can be represented by the linearised Boussinesq's equation (eq. 5).

\[ \frac{\partial^2 h}{\partial x^2} + \frac{\partial^2 h}{\partial y^2} = \frac{S}{T} \frac{\partial h}{\partial t} - \frac{R(x, y, t)}{T} \]  \hspace{1cm} (5)

where,

\( x, y \) : cartesian coordinates (L)

\( S \) : storage coefficient

\( T \) : transmissivity (L²/T)

\( t \) : time (T)

\( R \) : recharge (+) / discharge (-)

A two dimensional finite difference form of equation 5 is given in equation 6.
\[ h_{ij}^{n-1} = \frac{1}{[(abS/4T\Delta t)+a]} \left[ a h_{ij}^{n-1} + ab \frac{S}{4T\Delta t} h_{ij}^n + (1-a)(h_{ij}^n - h_{ij}^n) + ab \frac{R_y^n}{4T} \right] \] (6)

where,

a : horizontal grid size of cell ij (L)
b : vertical grid size of cell ij (L)
\( \alpha \) : implicit/explicit parameter

\( h^- \) : average watertable level (L)

The average level of a cell ij is defined as the average watertable level of the four (front and back, side by side) surrounding cells of ij. Since equation 4 uses the Crank-Nicolson approximation (\( \alpha = 0.5 \)), it presumes that the best values of the spatial derivatives (eq. 5) lie half way between time n and n+1. This approximation is preferred, because the watertable level in a cell continues to change during any specified period. Prathapar and Erskine (1990, 1991) reported a case study using the above methodology.

**Accuracy of recharge estimates**

The linear programming procedure is an inverse procedure to determine the recharge. In general, inverse problems are considered ill-posed (Yeh, 1986). This is especially true for transmissivity determination and recharge determination in a confined aquifer (Dietrich *et. al.* 1988). Accuracy of recharge estimate is improved by using filtered piezometric data and adopting a plausibility criterion by imposing realistic limit on the estimates (Allison and Peck, 1985).
Another source of error in the above procedure is due to truncation of the Taylor series approximation of the transient flow equation. Groundwater simulation models generally use an implicit formulation of the flow equation and iterate until a convergence criterion is met during each time step. This prevents the truncation error propagating from time step to time step. Although this procedure adopted a Crank-Nicolson approximation of the flow equation, which is second order correct, it did not adopt an iterative solution because recharge is determined for each time step separately. Hence, the truncation error associated with higher orders could not propagate. Furthermore, the truncation error associated with higher orders for a single time step is generally negligible. The assumptions used with respect to cell sizes, aquifer properties, conditions in the boundary and buffer cells could also contribute to errors.

GLOSSARY

Aquifer  A saturated porous medium which is capable of releasing water.
Capillary upflow  Process of water moving into the root zone from the subsoil.
Deep percolation  Process of water leaving the root zone into the subsoil.
Discharge  Process of water moving from a watertable into the sub-soil.
Infiltration  Process of water entering a soil surface.
Leakage  Process of water moving from an aquifer through a less permeable layer into a deeper aquifer.
Recharge  Process of water reaching a watertable.
REFERENCES


