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Model analysis of hydraulic conductivity of an aquifer, December 1968

H. Y. Fang
R. D. Varrin

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Groundwater and Seepage

MODEL ANALYSIS OF HYDRAULIC CONDUCTIVITY OF AN AQUIFER

by

H.Y. Fang
Robert D. Varrin

Fritz Engineering Laboratory Report No. 341.2
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Model Analysis of Hydraulic Conductivity of an Aquifer

H. Y. Fang\(^1\) and Robert D. Varrin\(^2\)

SYNOPSIS

This paper presents a study of steady and nonsteady radial well flow using a horizontal viscous-flow model. The fundamental behavior of ground-water movement during pumping tests is demonstrated. The validity of the classical equilibrium and nonequilibrium well flow equations for computing the hydraulic conductivity of an aquifer is discussed.

The analysis demonstrates that there is a close agreement between flow quantities predicted by the existing theories and those obtained in the model. It is suggested that using the nonequilibrium equation for analyzing pumping test data is the more logical and time-saving method.

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1. Director, Geotechnical Engineering Division, Fritz Engineering Laboratory, Lehigh University
2. Director, Water Resources Center, University of Delaware
INTRODUCTION

Hydraulic conductivity (or coefficient of permeability), K, is one of the most important engineering properties of soil. It is a measure of the capacity of soil to transmit water. In order to reliably determine hydraulic conductivity in the field, pumping tests are commonly used.

Two types of equations are generally used to analyze field pumping test data. One is the Thiem equilibrium equation (Thiem, 1906), and the other is the Theis nonequilibrium equation (Theis, 1935). The equilibrium equation can be used to determine the hydraulic conductivity of an aquifer if the rate of discharge of a pumped well is known, and if the drawdown in the observation wells at various known distances from the pumped well after the cone of depression has been stabilized is established. The nonequilibrium equation permits this determination when the rate of discharge of a pumped well is known, and when the drawdown as a function of time is determined for one or more observation wells at given distances from the pumped well.

The equilibrium equation has been widely applied for hydraulic conductivity determinations in the field of soil mechanics. In ground-water hydrology, the nonequilibrium equation is commonly used. However, the validity of these
two equations has not yet been fully verified by laboratory studies.

Among the various types of analog models which are used in the laboratory for studying ground-water movement are the sand model (Hall, 1935), the membrane analogy (Hansen, 1935), the viscous-flow analogy (Todd, 1954), and the electrical analogy (Zee, Peterson and Bock, 1957). In the study of radial well-flow phenomena, the viscous-flow model appears to be highly promising (Santing, 1957; DeWiest, 1965; Varrin and Fang, 1967).

Recently, a new type of horizontal viscous-flow model with "infinite" areal extent has been developed by DeWiest (1966) and Varrin and Fang (1967). A conformal mapping technique has been applied in the model design in order to consider the "infinite" extent of an aquifer. This type of model has been found to be a useful device in analyzing many types of ground-water flow problems, especially for analyzing steady or nonsteady radial well flows.

The major objectives of this study were: to demonstrate the fundamental behavior of ground-water movement during a pumping test; to verify the validity of the classical equilibrium and nonequilibrium well-flow equations; to compare the test results on the basis of various curve fitting methods which have been applied to the Theis nonequilibrium solution; and to attempt to clarify the differences in hydraulic conductivity units and nomenclature as used by soils engineers and ground-water hydrologists.
HYDRAULIC CONDUCTIVITY AND TRANSMISSIBILITY

Various names have been given to K, including effective permeability (Muskat, 1937), coefficient of permeability (Terzaghi, 1943), seepage coefficient (Polubarinova-Kochina, 1962), and hydraulic conductivity (DeWiest, 1965; Wit, 1966). In the field of soil mechanics, the term coefficient of permeability is commonly adopted. In ground-water hydrology the term hydraulic conductivity is widely used. In this paper, hydraulic conductivity is used throughout.

The hydraulic conductivity of an aquifer is influenced by the properties of the fluid in the aquifer. The fluid influence may be expressed by the ratio of its unit weight to its dynamic viscosity. P. C. Nutting (1930) was among the first to recommend the term physical permeability, $k$, for which

$$k = K \frac{\mu}{\gamma} \quad [L^2]$$

where $K = \text{hydraulic conductivity}, [L/T]$

$\mu = \text{dynamic viscosity of fluid}, [FT/L^2] \text{ or } [M/LT]$

$\gamma = \text{unit weight of fluid}, [M/L^2T^2]$

Other names have also been used for $k$, such as transmission constant (Muskat, 1937), specific permeability (Todd, 1959), and intrinsic permeability (DeWiest, 1965). The term intrinsic permeability is used throughout this text.
In the field of soil mechanics, the centimeter-gram-second system of units is commonly used for hydraulic conductivity. In the field of ground-water hydrology, the foot-gallon-day system is used. For example, the U. S. Geological Survey used the meinzer unit as a measure of hydraulic conductivity (Wenzel, 1942). The meinzer unit is defined as the flow of water in gallons per day through a cross-section of aquifer 1 ft. thick and 1 mile wide under an hydraulic gradient of 1 ft. per mile at field temperature. The unit of intrinsic permeability, \( k \), is usually extremely small, so that the darcy has been adopted as a more practical unit. Conversion factors for the meinzer and darcy units are shown in the Appendix.

In 1935 Theis introduced the term coefficient of transmissibility, \( T \), which is expressed as the rate of flow in gallons per day through a 1 ft. wide vertical strip of the aquifer under a hydraulic gradient of 1 ft. per ft. at the prevailing water temperature. The relationship between hydraulic conductivity, \( K \), and the coefficient of transmissibility, \( T \), is as follows:

\[
K = \frac{T}{b} \quad (2)
\]

where \( b \) is the thickness of confined aquifer and has the dimensions of length \([L]\).
STEADY FLOW - EQUILIBRIUM EQUATION

The equilibrium equation was developed by Gunter Thiem of Germany in 1906 for the determination of hydraulic conductivity, $K$. The equation was based on the following assumptions: (a) that the aquifer is homogeneous and isotropic with respect to hydraulic conductivity, and of infinite areal extent; (b) that the hydraulic conductivity is independent of time; (c) that the flow is laminar and steady; (d) that the discharging well penetrates and receives water from the entire thickness of the permeable, water-bearing stratum; and (e) that the well is pumped continuously at a constant rate until the flow of water to the well is stabilized.

Using plan polar coordinates with the well as the origin, the radial flow equation for a well completely penetrating a confined aquifer is given by (see Figure 1):

$$Q_w = A v = 2\pi rbK \frac{dh}{dr} \quad (3)$$

where $Q_w =$ well discharge, $[L^3/T]$  
$A =$ cross-sectional area perpendicular to flow  
$= 2\pi rb, [L^2]$  
$b =$ thickness of confined aquifer, $[L]$  
$r =$ radial distance to any point from axis of well, $[L]$  
$v =$ flow velocity, $[L/T]$  
$h =$ head at any point in the aquifer at time, $t$, $[L]$  
$dh/dr = i =$ hydraulic gradient, $[\text{dimensionless}].$
Separation of variables gives the following differential equation:

\[ dh = \frac{Q_w}{2\pi bK} \frac{dr}{r} \]  

(4)

The boundary conditions for Eq. 4 are at the well \( h = h_w \) and \( r = r_w \), and at the edge of the area of well influence \( h = H \), and \( r = R \). Integrating Eq. 4 between limits as indicated:

\[ \int_{h_w}^{H} dh = \int_{r_w}^{R} \frac{Q_w}{2\pi bK} \frac{dr}{r} \]

\[ H - h_w = \frac{Q_w}{2\pi bK} \ln \frac{R}{r_w} \]  

(5)

Thus, the hydraulic conductivity, \( K \), can be calculated as

\[ K = \frac{Q_w}{2\pi b} \ln \left( \frac{R}{r_w} \right) / (H - h_w) \]  

(6)

Since any two points will define the drawdown curve, Eq. 6 can be written in terms of drawdowns measured in two observation wells. For this case, the equation for hydraulic conductivity, \( K \), becomes:

\[ K = \frac{Q_w}{2b} \ln \left( \frac{r_2}{r_1} \right) \frac{1}{s_1 - s_2} \]  

(7)

where \( s_1 = H - h_1 \)

\[ s_2 = H - h_2 \]
r₁ and r₂ refer to the radial distance from axis of the pumped well to observation wells 1 and 2 respectively, and s₁ and s₂ refer to the drawdown at observation wells 1 and 2, respectively.

NONSTEADY FLOW - NONEQUILIBRIUM EQUATION

The partial differential equation in plane polar co-ordinates governing nonsteady well-flow in an incompressible confined aquifer of uniform thickness is:

\[
\frac{3^2 h}{3r^2} + \frac{1}{r} \frac{3h}{3r} = \frac{S}{T} \frac{3h}{3t}
\]

where S = storage coefficient, * [dimensionless]

\( t = \text{time since the flow started, [L]} \)

The terms h, r, and T have been defined previously.

Theis (1935) obtained a solution for the above equation based on the analogy between ground-water flow and heat conduction. By assuming that the well is replaced by a mathematical sink of constant strength, that after pumping begins,

*Storage coefficient is defined as the volume of water that an aquifer releases or takes into storage per unit surface area of the aquifer per unit change in the component of head normal to that surface (Theis, 1935).
h approaches \( H \) and \( r \) approaches \( \infty \), the solution is:

\[
s = H - h = \frac{Q_w}{4\pi T} \int_0^\infty \frac{e^{-u}}{u} \, du
\]

where \( u = \frac{r^2 S}{4Tt} \)

Eq. 9 is known as the nonequilibrium of Theis equation. This equation permits determination of the aquifer constants: storage coefficient, \( S \); transmissibility, \( T \); and hydraulic conductivity, \( K \).

The exponential integral in Eq. 9 has been assigned the symbol \( W(u) \) which is called the "well function of \( u \)." Eq. 9 may be somewhat too complicated for engineering purposes, but several investigators have developed approximate solutions. Included amongst these are the Theis method of superposition, the Theis recovery method and the Cooper and Jacob method.

Following a description of the experimental procedure used in the model study, each of the above approximate solutions is illustrated using experimental data obtained from the model. The results are compared with those obtained from the Thiem equilibrium solution.

MODEL STUDY

Description of the Horizontal Viscous-Flow Model

The first horizontal viscous-flow model was developed by H. S. Hele-Shaw in 1897-1899. Since then the model has
been refined by many investigators (Todd, 1959; Bear, 1960; Sternberg and Scott, 1964; DeWiest, 1965), and shown to be useful for analysis of almost any two-dimensional groundwater flow problem, whether steady or not. In 1956 this type of model was further improved to include three-dimensional flow (Bear and Kruysse, 1956; Santing, 1957). Recently, by using a conformal mapping technique to take fully into consideration the infinite extent of an ideal aquifer, a further improvement has been made on the model (DeWiest, 1966; Varrin and Fang, 1967). In this paper, only the horizontal type of viscous-flow model is discussed.

Since the complete description of the horizontal viscous-flow model has been reported by Varrin and Fang (1967), a brief description of this model is presented here:

The horizontal viscous-flow model consists of two closely spaced parallel plates and may simulate a portion of an aquifer, either phreatic or confined. The interspace between the plates represents the aquifer, with the value of the hydraulic conductivity being influenced by the width of the interspace as well as the properties of the permeant and the temperature of the test. Storage capacity is achieved by means of vertical tubes or vessels on top of the upper plate each of which is connected to the interspace.

In this study, the interspace between the plates was 1 mm. and the diameter of each storage vessel was 1 cm. The model was divided into interior and exterior regions. The
interior region was 1 meter by 1 meter and represented a prototype aquifer 10,000 by 10,000 meters. The exterior region represented an aquifer of infinite areal extent. Of course, it is physically impossible to model exactly this condition, but the technique of conformal mapping can be used to extend the model aquifer a considerable distance with a modest increase in size. Thus, the resulting model tends to approach the condition of an aquifer of infinite areal extent. Details of the conformal mapping technique are described by Nehari (1952) and DeWiest (1966). The interior region and the exterior region of the aquifer are shown in Figure 2. The plan view of both regions is shown in Figure 3.

The time scale for the model containing 96% aqueous glycerine as the fluid was determined by dimensional analysis to be 1:10,900. Therefore, as an example, one hour of laboratory testing simulates 15 months of field testing.

For the laboratory nonequilibrium pumping test to demonstrate the fundamental behavior of ground-water movement, the experimental procedure adopted was as follows:

1. Aquifer and storage vessels were filled with glycerine to a certain level. The head at each storage vessel was recorded.

2. At the same time, glycerine was added at a constant rate to the exterior part of the aquifer (cloverleaf shape). This is important, because glycerine flows from the exterior part of the
aquifer to the interior part and a constant head must be maintained at "infinity".

3. Once the pumping test started, the head distribution in each of the storage vessels was recorded as a function of time. Millimeter scales attached to the storage vessels facilitated the reading of the head changes.

4. The discharge from the pumping well and the temperature of the glycerine used in the model were recorded.

5. The pumping test was continued until a near steady state condition was reached.

Several tests were performed for various discharges, $Q_w$. Results of these tests are shown in Figures 5 and 6. The variation of drawdown with distance from the pumped well and with time is shown in Figure 5. It is indicated that the drawdown decreases as distance from the pumped well increases. Also, as time since pumping started increases, the drawdown increases. Figure 6 shows the linear relationship between the drawdown and the discharge for a given well for all values of time until the boundary of the aquifer is reached.

**Computation of Hydraulic Conductivity Model**

From the Navier-Stokes equation (Harr, 1962; DeWiest, 1965) for flow between plates, the equation for the
transmissibility, \( T \), of the model is found to be

\[
T = \frac{1}{12} \frac{g}{\nu} b^3
\]  

(10)

From Eq. 2:

\[
K = \frac{T}{b} = \frac{1}{12} \frac{g}{\nu} b^2
\]  

(11)

where \( g \) = acceleration of gravity = 981 cm/sec\(^2\)

= kinematic viscosity of the fluid,

= 5.29 cm\(^2\)/sec for 96% aqueous glycerine at 20°C

\( b \) = thickness of the interspace between the plates,

= 1 mm

Substituting these known values in Eq. 11, gives

\[
K = 15.4 \times 10^{-2} \text{ cm/sec} \quad \text{(at 20°C)}
\]

Eq. 11 indicates that the hydraulic conductivity value is influenced by the viscosity of the fluid. The viscosity of the fluid is inversely proportional to the temperature (see Figure 7). Therefore, the hydraulic conductivity value is dependent upon the temperature. The relationship between hydraulic conductivity and the temperature is shown in Figure 8. Since the model scaling is based on glycerine at 20°C, a correction should be made if the temperature during the test varied.

**Thiem's Equilibrium Equation**

From Eq. 7:

\[
K = \frac{Q_w}{2\pi b} \ln \left( \frac{r_1}{r_2} \right) \frac{1}{s_1 - s_2}
\]
where \( Q_w \) = discharge from pumping well = 0.18 cm/sec.

\[ b = 1 \text{ mm}; \ r_1 = 7.06 \text{ cm}; \ r_2 = 21.02 \text{ cm}. \]

\[ s_1 = 5.6 \text{ cm}; \text{ and } s_2 = 3.8 \text{ cm}. \] (From Figure 9)

Therefore, \( K = 17.3 \times 10^{-2} \text{ cm/sec} \) (at 23°C)

Theis' Nonequilibrium Equation

A. Theis Method of Superposition (1935)

A logarithmic plot of \( W(u) \) versus \( u \), known as a "type curve" is prepared. Theoretical values of \( W(u) \) for a wide range of \( u \) have been prepared by Wenzel (1942) and may be found in any standard ground-water text book. Values of drawdown, \( s \), are plotted against values of \( r^2/t \) on logarithmic paper of the same size as the type curve. The observed data curve is superimposed on the type curve. An arbitrary point is selected on the coincident segment, and the coordinates of this matching point are recorded (see Figure 10). With values of \( W(u), u, s, \) and \( r^2/t \) thus determined, \( T \) and \( K \) can be obtained from Eq. 9.

For \( u = 0.07; \ W(u) = 1; \ s = 1.05 \text{ cm}. \)

\[ t = 100 \text{ sec}; \text{ and } Q = 0.22 \text{ cm}^3/\text{sec}. \]

Eq. 9 yields:

\[ T = 16.6 \times 10^{-3} \text{ cm}^2/\text{sec}. \]

\[ K = 16.6 \times 10^{-2} \text{ cm/sec}. \]

B. Theis Recovery Method (Theis, 1935; Wenzel, 1942)

If a well is pumped at a constant rate and then shut down, the head will recover from its lowest value at time
when pumping is stopped to attain value $h'$ at time $t'$ from the time of shutdown. If $H$ is the initial value of the head before pumping started, then $H - h' = s'$ is called the residual drawdown. If a semilog plot of $s'$ versus $t/t'$ is prepared, then $T$ may be calculated by the following equation:

$$ T = \frac{2.30 Q}{4\pi \Delta s'} $$  \hspace{1cm} (12)

where $\Delta s'$ = slope of the semilog plot

From Figure 11, $\Delta s' = 2.38$ cm.

$$ Q = 0.22 \text{ cm}^3/\text{sec.} $$

Therefore, $T = 17.5 \times 10^{-3} \text{ cm}^2/\text{sec.}$

$$ K = 17.5 \times 10^{-2} \text{ cm/sec.} \hspace{0.5cm} (at \hspace{0.2cm} 23^\circ\text{C}) $$

C. Cooper and Jacob Method

A simplified form of the Theis equation was developed by Cooper and Jacob (Cooper and Jacob, 1946; Jacob, 1950). The equation for computing $T$ and $K$ is as follows:

$$ T = \frac{2.30 Q}{4\pi \Delta s} \hspace{1cm} (13) $$

From Figure 12, $Q = 0.22 \text{ cm}^3/\text{sec.}$$

$$ \Delta s = 2.4 \text{ cm.} $$

Therefore, $T = 16.7 \times 10^{-3} \text{ cm}^2/\text{sec.}$

$$ K = 16.7 \times 10^{-2} \text{ cm/sec.} \hspace{0.5cm} (at \hspace{0.2cm} 23^\circ\text{C}) $$

A summary of hydraulic conductivity values computed from the above methods are tabulated as follows:
<table>
<thead>
<tr>
<th>Method</th>
<th>Hydraulic Conductivity cm/sec.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Model</td>
<td></td>
</tr>
<tr>
<td>Thiem equilibrium equation</td>
<td>$15.4 \times 10^{-2}$ at $20^\circ$C</td>
</tr>
<tr>
<td>Nonequilibrium equation</td>
<td></td>
</tr>
<tr>
<td>Theis method</td>
<td>$17.3 \times 10^{-2}$ at $23^\circ$C</td>
</tr>
<tr>
<td>Theis recovery method</td>
<td>$16.6 \times 10^{-2}$ at $23^\circ$C</td>
</tr>
<tr>
<td>Cooper and Jacob method</td>
<td>$17.5 \times 10^{-2}$ at $23^\circ$C</td>
</tr>
<tr>
<td>Cooper and Jacob method</td>
<td>$16.7 \times 10^{-2}$ at $23^\circ$C</td>
</tr>
</tbody>
</table>

From the model analysis, within the experimental limits, the equilibrium and nonequilibrium equation yielded the same hydraulic conductivity. However, with the equilibrium equation, the straight line portion of drawdown curve should be used.

**SUMMARY AND CONCLUSIONS**

The model analysis of hydraulic conductivity of an aquifer can be summarized as follows:

1. A new type of horizontal viscous-flow model with "infinite" areal extent has been found to be a useful device in analyzing steady and non-steady radial well flows.

2. The principal advantages of the analog model are:
   a. A time scale in the model whereby 1 second in the laboratory represented 3 hours in nature.
b. The simulation of a very large homogeneous and isotropic aquifer (24,000 meters by 24,000 meters).

3. From a model pumping test, the following relationships were shown:
   a. Drawdown decreased as distance from the pumped well increased.
   b. As time since pumping started increased, the drawdown increased until equilibrium was reached.
   c. A straight-line relationship existed between the drawdown and the discharge for a given well for all values of time until the boundary of the model aquifer was reached.

4. The Thiem equilibrium equation and various methods based on the Theis nonequilibrium equation were used to determine the hydraulic conductivity of the model aquifer. Within the experimental limits, the equilibrium and nonequilibrium equations yielded the same hydraulic conductivity.

5. The duration of the laboratory pump test required to establish equilibrium was about two hours or the equivalent of about 3 years in nature. In contrast, hydraulic conductivity was determined from the nonequilibrium equation after a few
minutes in the laboratory (several days in nature). The advantage of calculating hydraulic conductivity from the nonequilibrium methods was easily demonstrated.
ACKNOWLEDGEMENT

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Professor T. J. Hirst provided constructive criticisms and numerous suggestions.
REFERENCES


APPENDIX
Conversion Factors

From Darcy's Law

\[ Q = K \frac{A}{L} h \]

where
- \( Q \) = discharge, \([L^3/T]\)
- \( K \) = hydraulic conductivity, \([L/T]\)
- \( A \) = cross-sectional area, \([L^2]\)
- \( h/L \) = hydraulic gradient, \([\text{dimensionless}]\)

or

\[ Q = K A \frac{d h}{d L} \] expressed in general terms

**Hydraulic Conductivity**

\[ K = \frac{Q}{A \left( \frac{d h}{d L} \right)} \] \hspace{1cm} (A)

and

**Intrinsic Permeability**

\[ \bar{K} = K \frac{\mu}{\gamma} = \mu \frac{Q/A}{\gamma \left( \frac{d h}{d L} \right)} \] \hspace{1cm} (B)

From equation (B), the darcy is defined as:

1 darcy = \( \frac{1 \text{ centipoise} \times 1 \text{ cm}^3/\text{sec.}/\text{cm}^2}{1 \text{ atmosphere}/1 \text{ cm}} \)

= \( 0.987 \times 10^{-6} \text{ cm}^2 = 1.062 \times 10^{-11} \text{ ft}^2 \).

The meinzer is defined as:

1 meinzer = \( 5.5 \times 10^{-2} \text{ gal/day} \) (for water at 60°F).
Figure 1 Steady Flow to a Well in a Confined Aquifer
Figure 2 Relationship of Interior and Exterior Domains - Horizontal Viscous-Flow Model
Figure 3 Plan View of Model (Left portion only used)
Figure 4  Horizontal Viscous-Flow Model
OBSERVATION WELLS

Pumped Well
Q = 0.18 cm³/sec.

Distances from Pumping Well

r₁ = 7.06 cm
r₆ = 21.02 cm
r₁₀ = 35.40 cm
r₁₃ = 45.60 cm
r₁₅ = 63.70 cm

Drawdown S, cm

DISTANCE FROM PUMPING WELL, cm

Figure 5 Drawdown vs. Distance from Pumping Well
Figure 6 Drawdown vs. Discharge
Figure 7  Viscosity versus Temperature for Aqueous Glycerol
Figure 8  Effect of Temperature on Hydraulic Conductivity
Figure 9: Thiem's Equilibrium Method

Drawdown in Centimeters

\[ Q = 0.18 \text{ cm}^3/\text{sec.} \]
\[ T = 17.3 \times 10^{-3} \text{ cm}^2/\text{sec.} \text{ at } 23^\circ C \]

Observation Well No. 1
\( r_1 = 7.06 \text{ cm} \)

Aquifer No. 1

Observation Well No. 2
\( r_2 = 21.02 \text{ cm} \)

Approx. Time to Reach "Near Steady State"

Time After Pumping Started in Seconds

\[ S_1 = 5.6 \text{ cm} \]
\[ S_2 = 3.8 \text{ cm} \]
Figure 10 Theis' Nonequilibrium Method

AQUIFER NO. 1

\[
\begin{align*}
Q &= 0.22 \text{ cm}^3/\text{sec.} \\
T &= 16.6 \times 10^{-3} \text{ cm}^2/\text{sec. AT 23 °C} \\
K &= 16.6 \times 10^{-2} \text{ cm/sec.}
\end{align*}
\]

\[
\begin{align*}
\text{MATCH POINT} \\
s &= 1.05 \text{ cm} \\
t &= 100 \text{ sec.} \\
W(u) &= 1 \\
u &= 0.07
\end{align*}
\]
Figure 11 Theis' Recovery Method

\[ Q = 0.22 \text{ cm}^3/\text{sec.} \]
\[ T = 17.5 \times 10^{-3} \text{ cm}^2/\text{sec. AT } 25 \degree \text{C} \]
\[ K = 17.5 \times 10^{-2} \text{ cm/sec.} \]
Figure 12 Jacob's Modified Nonequilibrium Method

\[ Q = 0.22 \text{ cm}^3/\text{sec.} \]

\[ T = 16.7 \times 10^{-3} \text{ cm}^2/\text{sec. AT 23 °C} \]

\[ K = 16.7 \times 10^{-2} \text{ cm/sec.} \]

\[ \Delta s = 2.4 \text{ cm} \]

\[ \Delta \log t = 1 \]