Critical Thinking in Hydrogeologic Test Interpretation
Additional materials #2

Guidance materials for pumping test interpreters

Christopher J. Neville
S.S. Papadopulos & Associates, Inc.
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Checklist for the Execution of Pumping Tests

Based on the notes of David Dahlstrom, Barr Engineering
Christopher J. Neville, S.S. Papadopulos & Associates, Inc.
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1. What to do before you travel to the site

1. Design your test.

2. Provide for accurate measurements of elapsed time, distances between wells, drawdown, and pumping rate.

   Reliable analyses of the data requires accurate measurements of elapsed time, distances between wells, drawdown, and pumping rate:
   - Record time with a synchronized digital watches and a stopwatch;
   - Measure distances twice;
   - Make a manual water level measurement every time you download a transducer; and
   - Measure pumping rates frequently and, if necessary, adjust the rates carefully.

3. Select transducers that cover the range of you anticipate.

   A 10 psi transducer can measure up to 23.1 feet (7.04 m) of drawdown. If it is placed more than 23.1 feet below static water level, it will be off-scale and will register a constant reading. If a 10 psi transducer is placed more than 46.2 feet below static water level (twice the operating range) the transducer can be ruined. It is best to place the transducer at the midpoint of its range for maximum accuracy and to allow for rising water levels during the test. The total depth of a well equals the water level below top of casing plus the transducer output with the transducer at the bottom of the well.

4. Make sure that your water level tapes, transducers and dataloggers are working properly.

5. Know whether your transducers are vented or non-vented.

2. What to do when you arrive at the site

1. Check the calibration of your transducers in the field.

   Make sure you know how to check the calibration of a transducer.

2. Perform pretest monitoring to determine the magnitude of water level variations in the monitoring network due to external influences.

   To yield useful data the aquifer test must induce drawdowns of greater magnitude than the variations caused be external forces. Identify and quantify the external causes and, if possible, filter out their influence. It may be possible to time the aquifer test such that the external forces are either constant or not active. Some examples are: pumping in other wells; water level fluctuations in surface water bodies; barometric pressure changes; trains; and infiltration.

3. Locate the generator away from the wellhead.

   If possible, have the generator located away from the wellhead to reduce the noise you have to work around.

3. Establish the pumping requirements

1. Make sure the pump intake is deep enough to allow adequate drawdown.

2. Induce the maximum stress possible on the aquifer, but avoid drawing the water level below the top of a confined aquifer, and avoid pumping sand.

   The bigger the pump the better. An oversized pump that is throttled back will generally runs at a higher rate than a pump running near capacity. A fully open valve is a horrifying sight when the pumping rate is declining.

3. Conduct a step test before the constant-rate test to identify a sustainable rate.

4. If you conduct a step test or start the pump to make sure it runs, allow the aquifer to stabilize before beginning the constant-rate test.

5. Choose pumping rates that can be monitored easily.

   Have a stop watch available and estimate the range of acceptable times to pump a given volume of water. For example, if the target discharge rate is 30 gpm, at 28.5 gpm it will take 10 minutes and 32 seconds to pump 300 gallons, and at 31.5 gpm it will take 9 minutes and 32 seconds to pump 300 gallons. To be considered a constant-rate test, the pumping rate should not vary by more than 5 percent.

4. Measuring flow rates

1. Measure the flow rate carefully.

   If you are going to deliberately change the pumping rate, make sure that you can measure the rate accurately throughout the test.

   Measure the pumping rate with a totalizing flow meter, and control the rate with a valve and an in-line flowmeter. Install the valve downstream of the flow meter to reduce the turbulence through the meter. Have the meter and valve located close enough together so one person can adjust the valve and readily read the meter.

2. If a well does not produce sand, measure the pumping rate with a totalizing flow meter and control the rate with a valve and an in-line flowmeter.

3. For wells that produces sand, consider alternative approaches for measuring the pumping rate that may include orifice tube, weir box, measuring the profile of the falling water stream, or filling a container of known volume.

   Totalizing flow meters work wonderfully unless the well produces "too much" sand. "Too much" sand is enough to jam the flow meter. This may not be much at all. Alternative measurement techniques include: orifice tube, weir box, measuring the profile of the falling water stream, or filling a container of known volume. These methods allow accurate instantaneous flow rate measurements but do not allow calculation of the cumulative average flow rate for the entire pumping period.

4. Make sure that there is a backflow-preventer (check valve) on the discharge pipe.

   A check valve on the discharge pipe above the submersible pump will prevent backflow when the pump is shut off. If the discharge line discharges to a tank or surface water body, make sure that an air break exists so a siphon is not created when the pump is shut off.

5. Measuring water levels

1. Know where the bottom of a well is before you install a transducer.

   Don't let the mud generally occupying the bottom of the well clog the ports on the transducer cap.

2. When setting transducers in wells, be extremely careful not to cut the transducer cable on the top of the riser.

   If water gets inside the cable the transducer can be ruined. Use wire rope thimbles or duct tape to make a collar for the cable where it contacts the riser pipe and other sharp edges.

3. Be cautious when measuring the water level in a pumping well.

   Know where the pump intake is and do not lower the water level indicator to that level. Never put the pressure transducer below or opposite the pump intake. Avoid placing a water level tape or transducer in a well with a pump installed in it that does not contain a drop pipe. A drop pipe will prevent the water level monitoring device from becoming snagged on the pump wires or other objects in the well.

4. Loosen well seals and vent the covers before taking any measurements and leave them loose throughout the testing period.

   Airtight well seals and non-vented dust covers impede water level changes if the water level is above the well screen. Loosen seals and vent the covers before taking any measurements and leave them loose throughout the testing period.

5. Set the references on the dataloggers such that the output is equal to the manually measured water level.

   Datalogger record water level changes relative to a reference value defined by the user. Set the references on the dataloggers such that the output is equal to the manually measured water level. Set the datalogger to record in length expressed relative the top of casing. Keep a log of manual measurements versus datalogger outputs. Dataloggers often drift during the early stages of a test, due to cable straightening and temperature equilibration. Be prepared to reset the reference if the datalogger drifts during the test. If the reference is reset while the test is underway, the previously collected data are automatically adjusted. If the data logger output continues to drift, check for other causes: slipping of the transducer; failing of the transducer; or kinking of the tape of the manual device.

6. Avoid immersing dataloggers.

   Not all dataloggers are watertight. For some vented transducer designs, there is a vent on the instrument panel to keep the transducer in equilibrium with barometric pressure. The box can be exposed to rain, sleet, and snow but cannot be immersed.

7. Allow a few seconds for a datalogger to wake up if you are doing delayed starts.

   If you are using delayed starts and stepping a test without stopping the previous step, you will not be able to synchronize the clocks of the datalogger(s). Read the time output from the datalogger to determine how much drift has occurred and allow for this drift when starting the next step of the test.

8. Increment the test number before adjusting the transducer parameters.

   Changes to the transducer parameters are retroactive to the previous test if the test number is not incremented.

9. Avoid shocks.

   Shocks erase data. During cold weather, make sure static electricity does not build up between you and the datalogger. When downloading data, touch the cable and your hands to the screws on the top of the data logger before connecting the cable.

10. Back up downloaded data. Back up backed-up downloaded data.

11. Always have enough manual measurements to suffice as a data set for analysis in case the datalogger fails, the data are lost, etc.

12. Assign the nightshift to the groundwater modellers.

6. What to do after the test

1. Measure recoveries for as long as you can.
2. Secure your transducers.

   Transducers that are accessible are frequently stolen or vandalized.
3. Plot the pumping rate versus time.
4. Plot drawdowns versus time for all of the observation wells.

Checklist for the Interpretation of Pumping Tests

Christopher J. Neville
S.S. Papadopulos & Associates, Inc.
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1. Have the methods used to measure flow rates and water levels been reported clearly?

2. Is the pumping history reported in detail?

3. Have unreliable measurements of drawdown been identified and eliminated from the data set?

4. Have the water level data been filtered, so that the reported drawdowns are due only to the effects of pumping?

5. Have the data been assembled on a composite plot and responses from non-representative observation wells been identified?

6. Are the methods of analysis consistent with the understanding of the hydrologic and geologic settings?

7. How do site conditions compare with the idealizations of the analysis?
   As a start, how do site conditions compare with the assumptions underlying the Theis analysis?
   - Darcy's law is valid.
   - The aquifer is horizontal and has constant thickness.
   - The aquifer is infinite in areal extent.
   - The hydraulic conductivity is uniform.
   - The hydraulic conductivity is isotropic.
   - The aquifer is perfectly confined along its top and bottom by impermeable strata.
   - The head in the pumped aquifer always remains above the top of the aquifer.
   - The release of water from storage is governing by linear constitutive relations with properties that remain constant through time.
   - There is a single pumping well.
   - The pumping well has an infinitesimal diameter.
   - The pumping well penetrates the full thickness of the aquifer.
   - There are no inertial effects in the wellbore.
   - The pumping rate is constant during the duration of the test.
   - Observations wells have infinitesimal diameters.
   - Observation wells penetrate the full thickness of the aquifer.
   - The hydraulic head is known everywhere prior to the start of the test.
   - There are no variations in water levels through time that are not due to pumping.

8. Do the ranges of times and distances over which the data are matched satisfy the limits of applicability of the analyses?

   Example 1: Cooper-Jacob approximation $u = \frac{r^2 S}{4Tt} < 0.1$

   - Time-drawdown: $t > 10 \frac{r^2 S}{4T}$
   - Distance-drawdown: $r < \left(0.1 \frac{4Tt}{S} \right)^{1/2}$
   - Composite: $\frac{t}{r^2} > 10 \frac{S}{4T}$

   Example 2: Wellbore storage

   $$t > 250 \frac{r_c^2}{T}$$

9. Does the reporting of the parameter estimates reflect a clear understanding of the difference between accuracy and resolution?

10. Are the parameter estimates derived from complex models consistent with the preliminary estimates derived from simpler conceptual models? [They don't have to be consistent if we have a good explanation for significant discrepancies.]

11. Are the parameter estimates consistent with underlying restrictions in the analyses?

    Example: Leaky aquifer analysis -- $K' < 0.01K$

12. Do the parameter estimates lie within physically realistic bounds. For example, does the storage coefficient fall within the likely limits $10^{-6} < S < 0.2$?

13. Are the parameter estimates consistent with the hydrostratigraphic conceptualization?

14. Are the parameter estimates consistent with other estimates that are available:

    - Literature values of hydraulic conductivity;
    - Estimates of hydraulic conductivity derived from the analysis of grain size distributions;
    - Reconnaissance-level estimates of transmissivity from specific capacity values from regional compilations of water well records; and
    - Single-well tests (slug tests, step tests)?

Diagnostic responses to pumping

Christopher J. Neville
S.S. Papadopulos & Associates, Inc.
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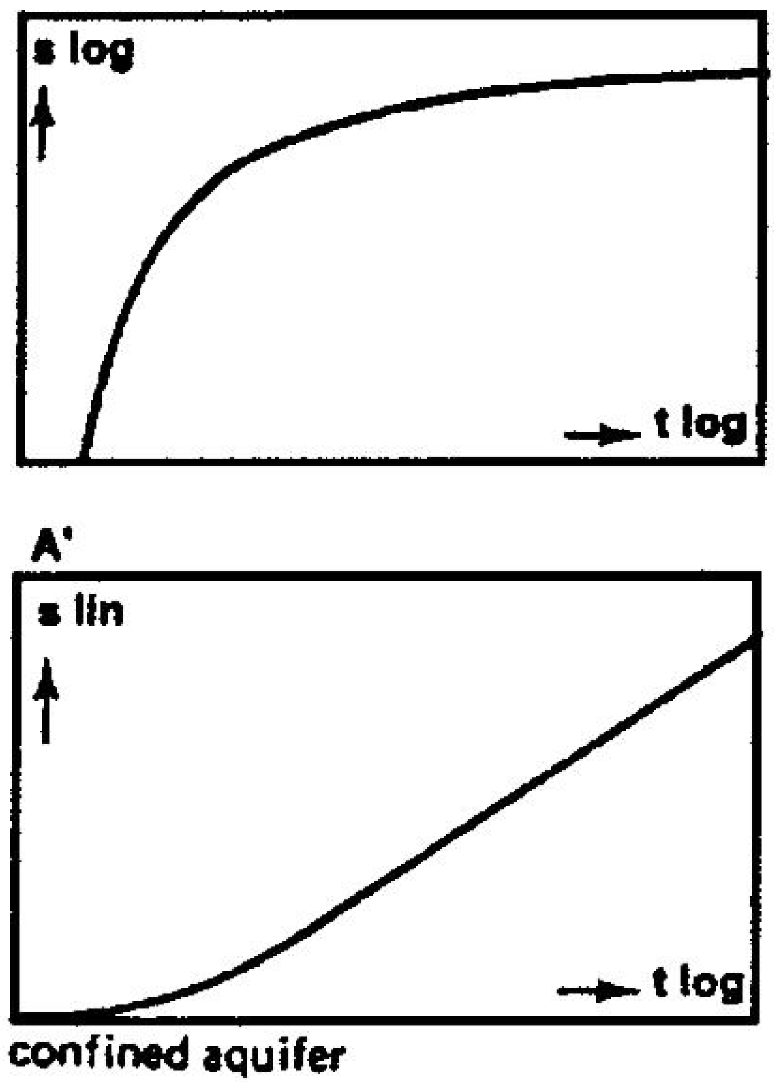
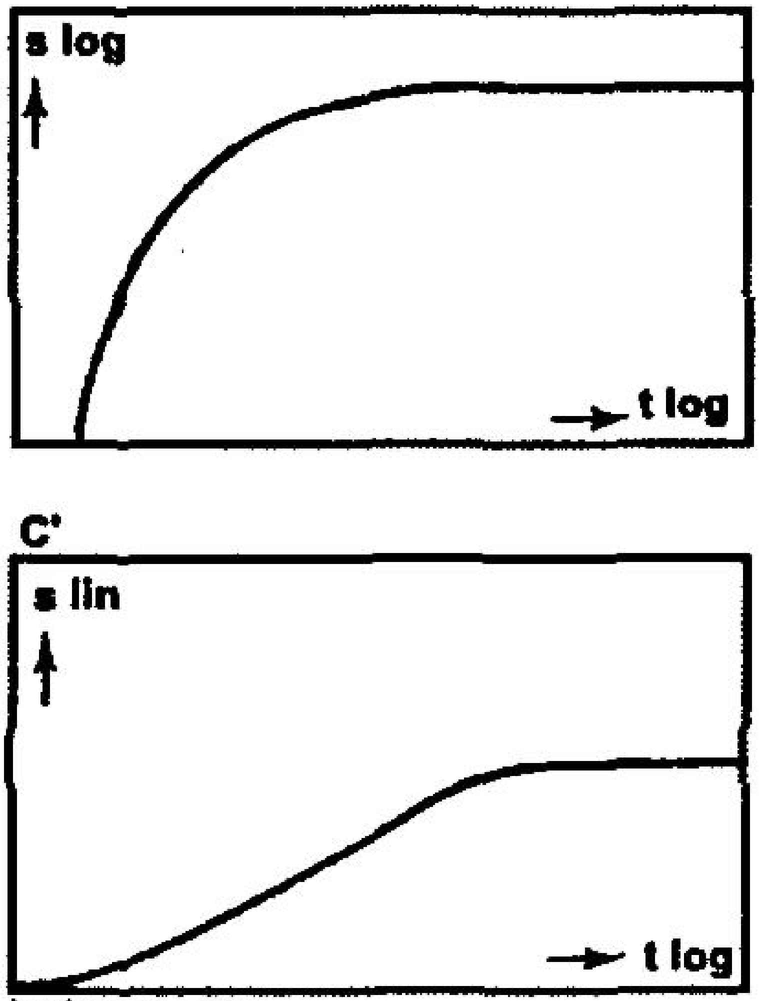
1. Ideal confined aquifer
2. Ideal leaky aquifer
3. Ideal unconfined aquifer
4. Ideal confined aquifer with a linear recharge boundary
5. Ideal confined aquifer with a linear barrier boundary
6. Ideal confined aquifer: Partially penetrating pumping well
7. Ideal confined aquifer: Wellbore storage
8. Confined double porosity aquifer
9. Confined flow to a vertical fracture
10. Pumping well in a fractured dike

1. Ideal confined aquifer

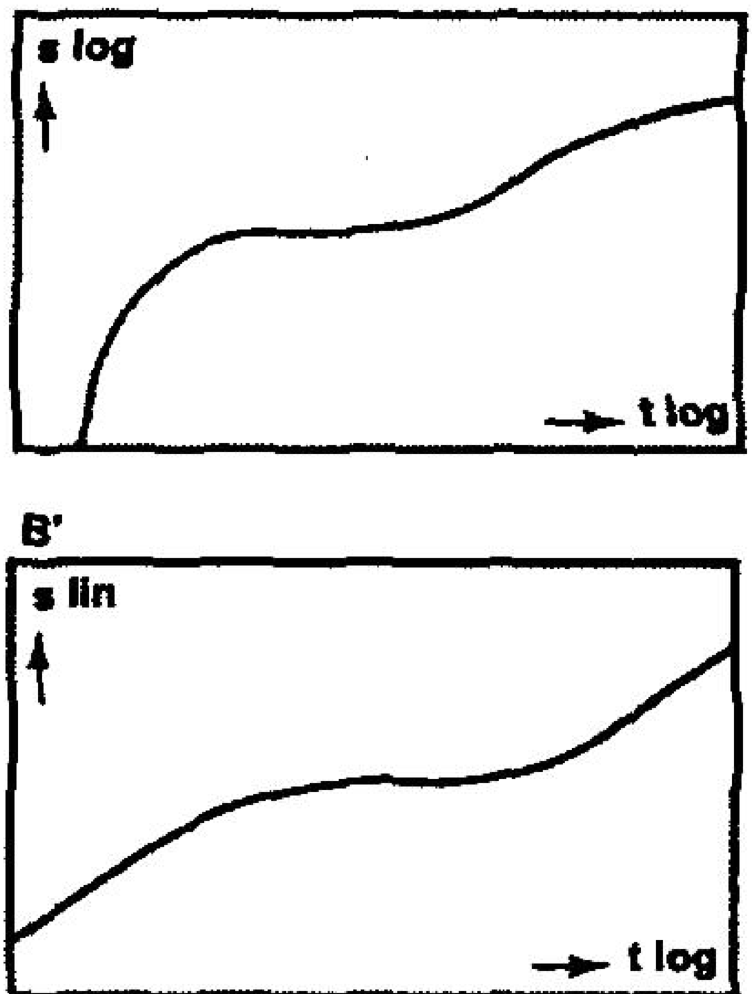


2. Ideal leaky aquifer



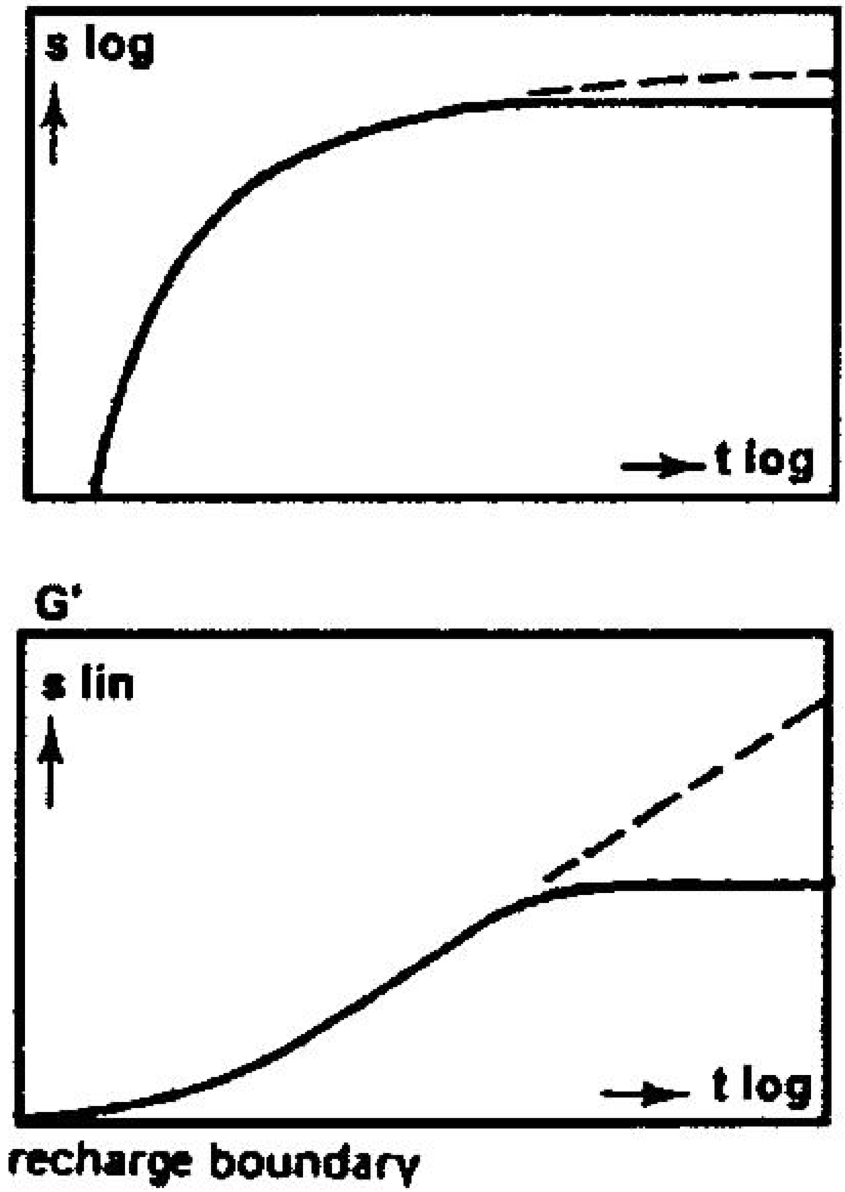
leaky aquifer

3. Ideal unconfined aquifer

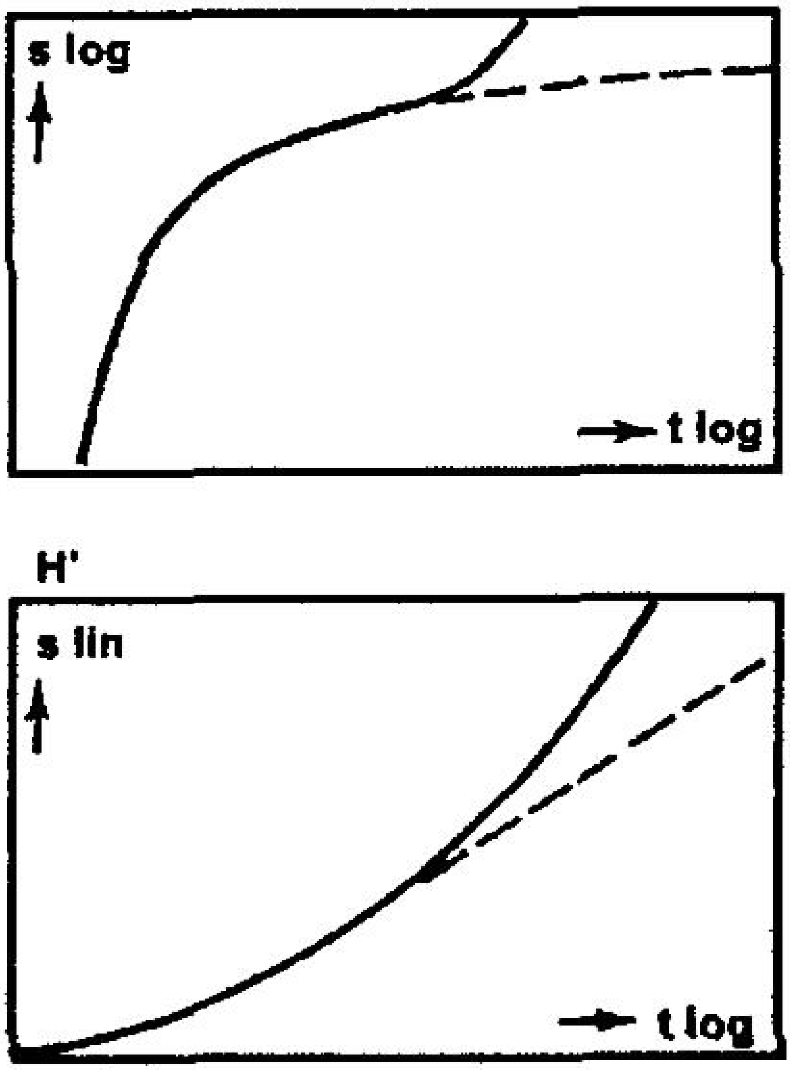


unconfined aquifer, delayed yield

4. Ideal confined aquifer with a linear recharge boundary

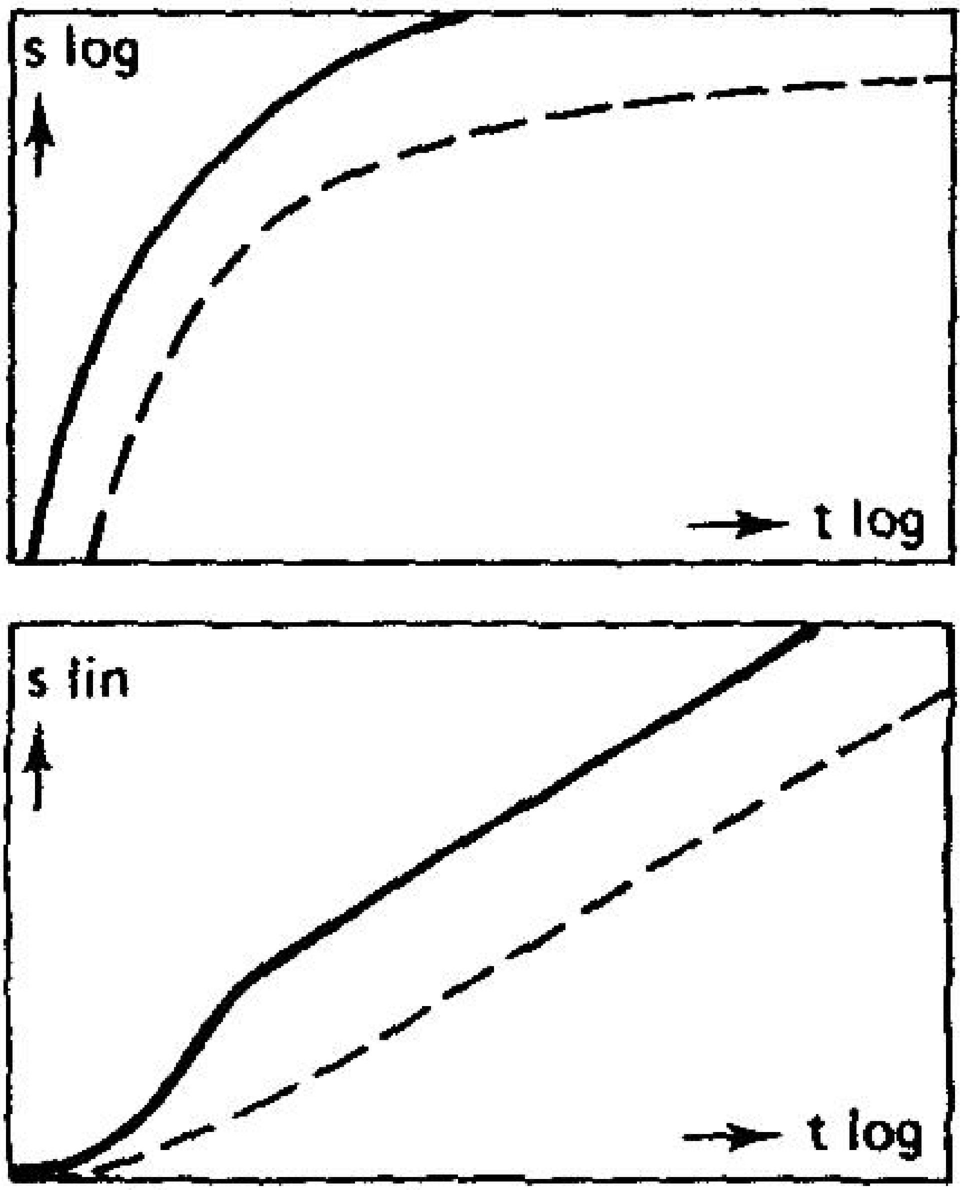


5. Ideal confined aquifer with a linear barrier boundary



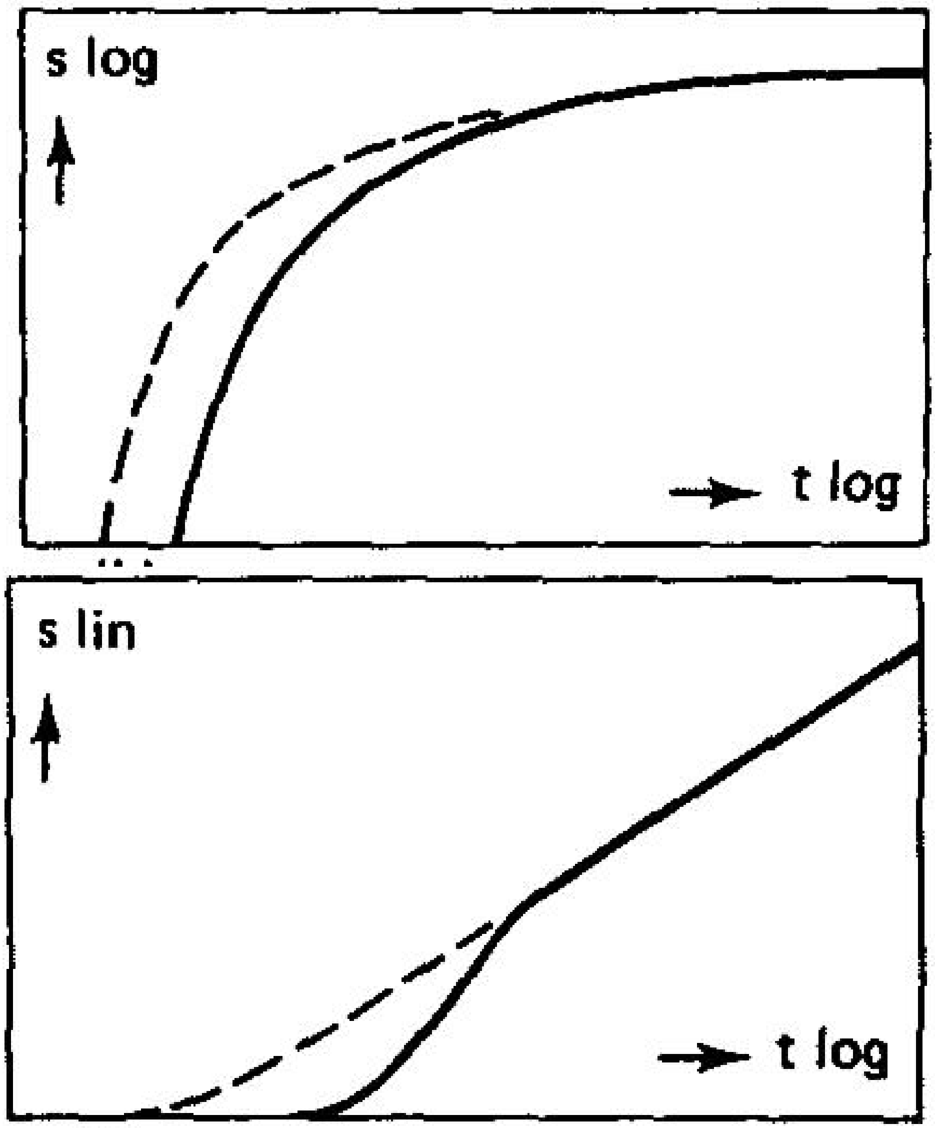
barrier boundary

6. Ideal confined aquifer:
Partially penetrating pumping well



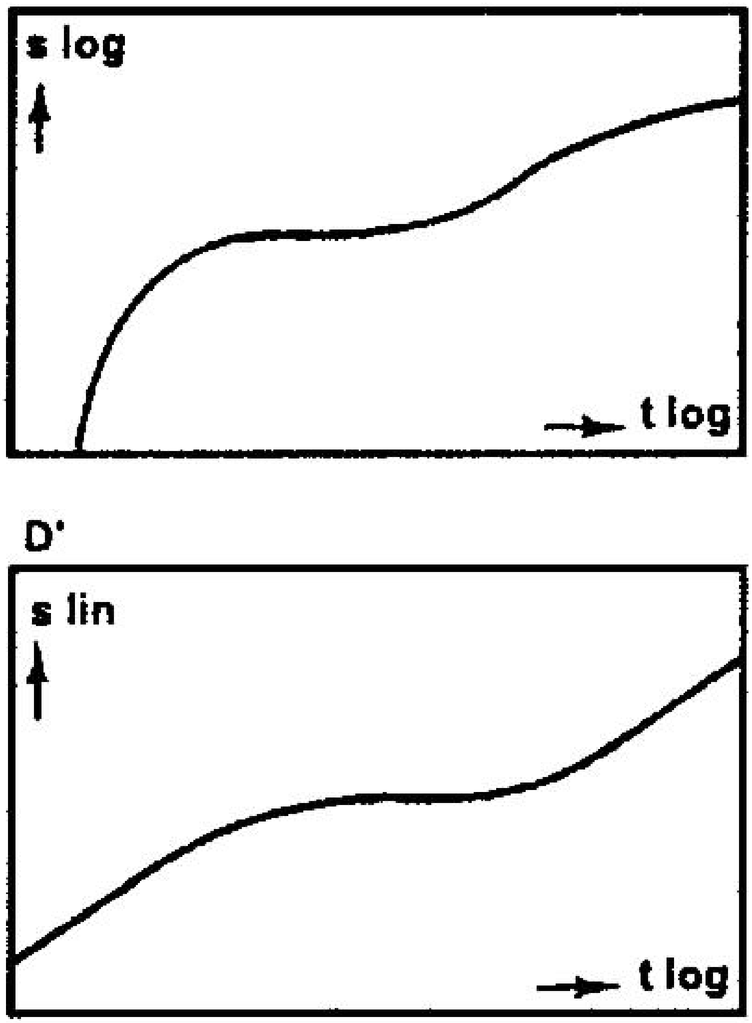
partial penetration

7. Ideal confined aquifer:
Wellbore storage



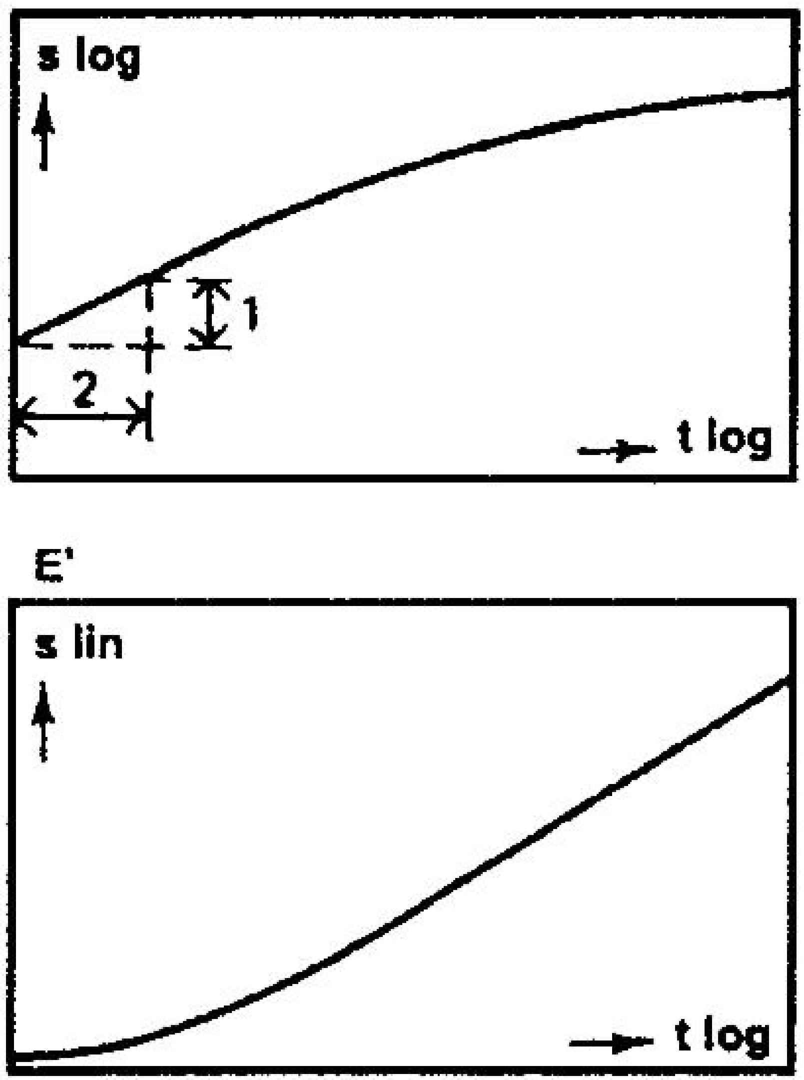
well-bore storage

8. Confined double porosity aquifer



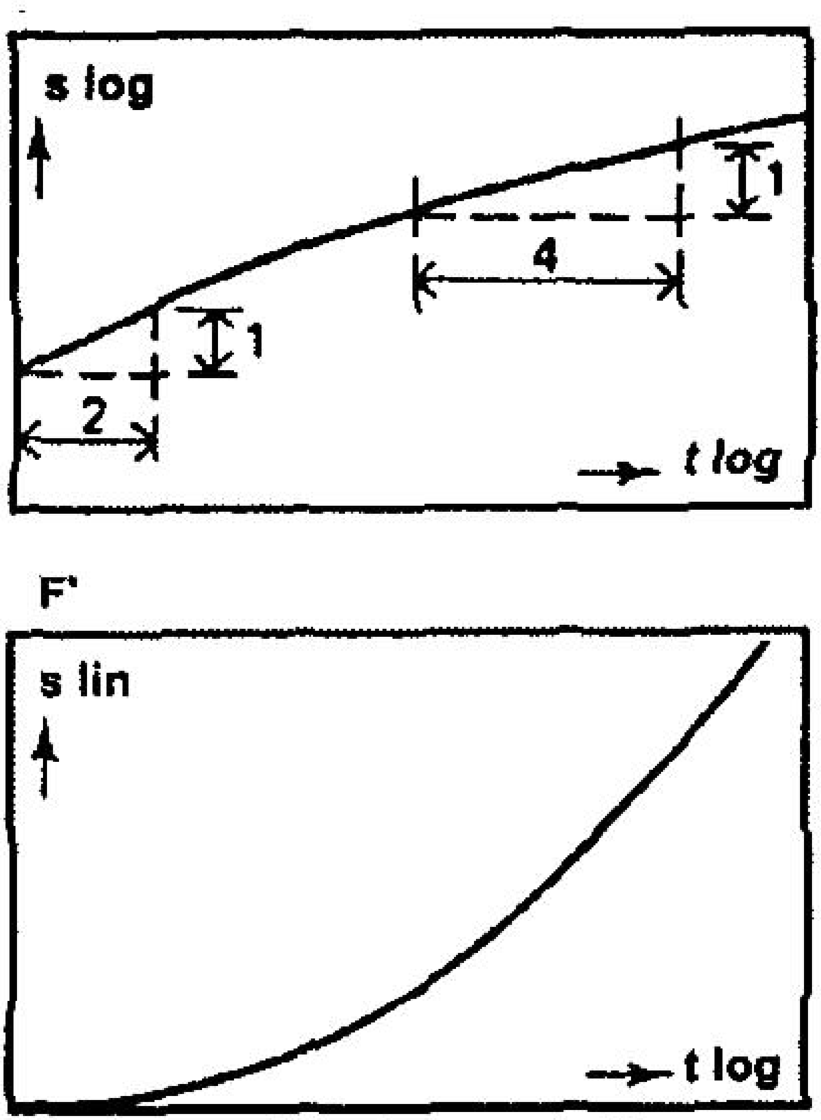
confined fractured aquifer
(double porosity type)

9. Confined flow to a vertical fracture



pumped well in single plane, vertical fracture

10. Pumping well in a fractured dike



pumped well in fractured dike

Literature values of hydraulic conductivity

Christopher J. Neville
S.S. Papadopulos & Associates, Inc.
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Introduction

Tabulations of hydraulic conductivities in the literature provide a useful reality check in pumping test interpretations. Significant discrepancies between hydraulic conductivity estimates derived from *in situ* tests and ranges cited in textbooks are a good indicator that either we have done something wrong in our analyses, or there is something unusual at our site. Although we can hardly expect a value of hydraulic conductivity that is picked out of a textbook to provide a representative estimate of hydraulic conductivity at a particular site, there are some compelling reasons why these tabulations should be consulted.

- Guidelines on hydraulic conductivity can be very handy when carrying out an early desktop study. In these applications, sometimes all we need is a "ballpark" estimate of hydraulic conductivity.

- Tabulations of typical hydraulic conductivities offer us the simplest and quickest way to attach values to what is otherwise only qualitative information. Frequently, the only information we have about part of a study area may be geologic descriptions in drilling logs.

- Textbooks values may be all that are available to assign to particular units. Values of hydraulic conductivity for some units may not be available, as their characterization may be outside the immediate scope of a project. In my experience as a hydrogeologist, quantitative data on fine-grained units is scarce. Good data on these fine-grained units are difficult to obtain, and wells are generally not installed in them.

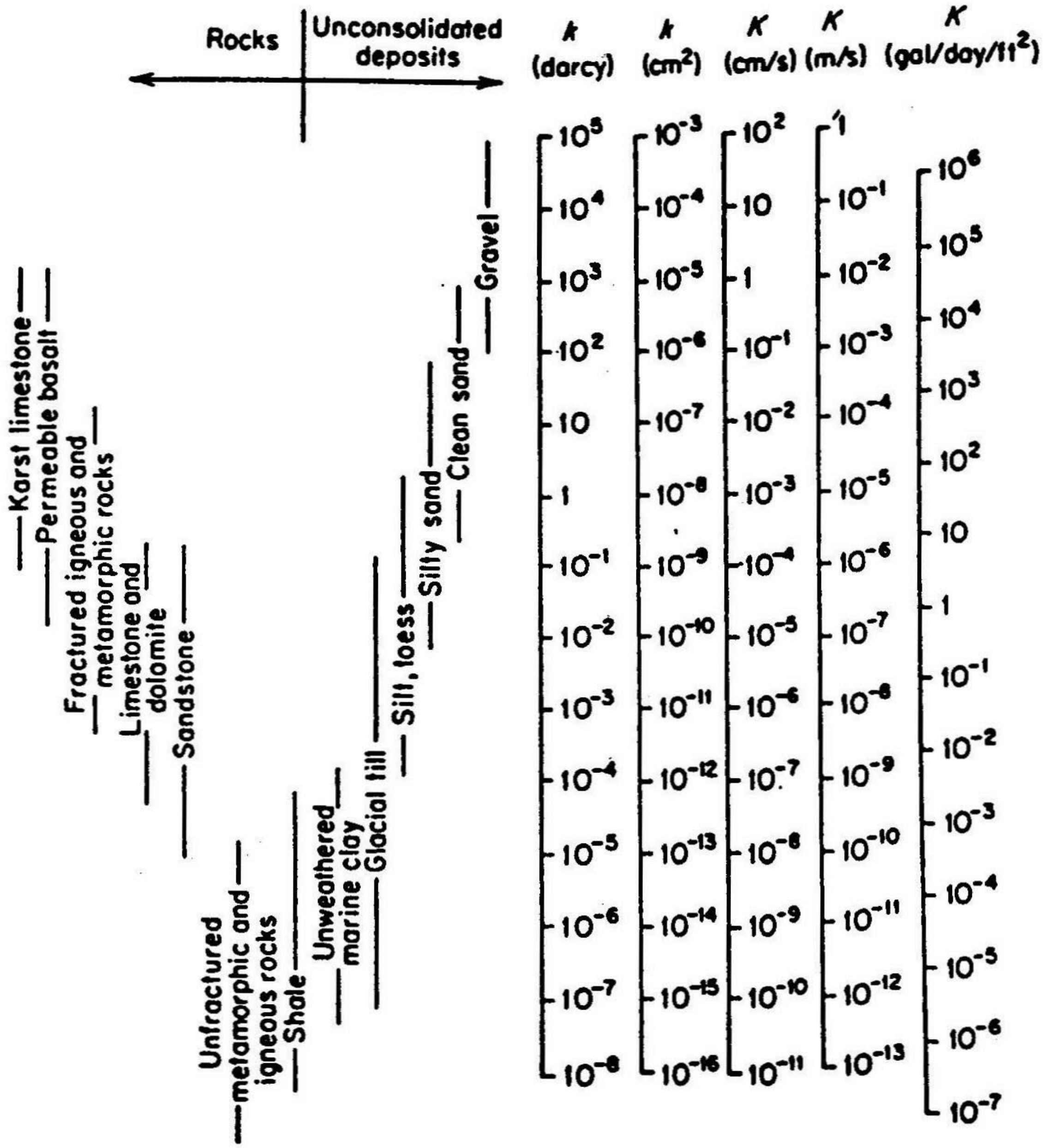
- The guidance documents can help us identify what is the likely range of hydraulic conductivity for a material sharing that geological description.

Typical values of hydraulic conductivity are presented in many references in the hydrogeology and geotechnical engineering literatures. We have assembled eight compilations from various sources.

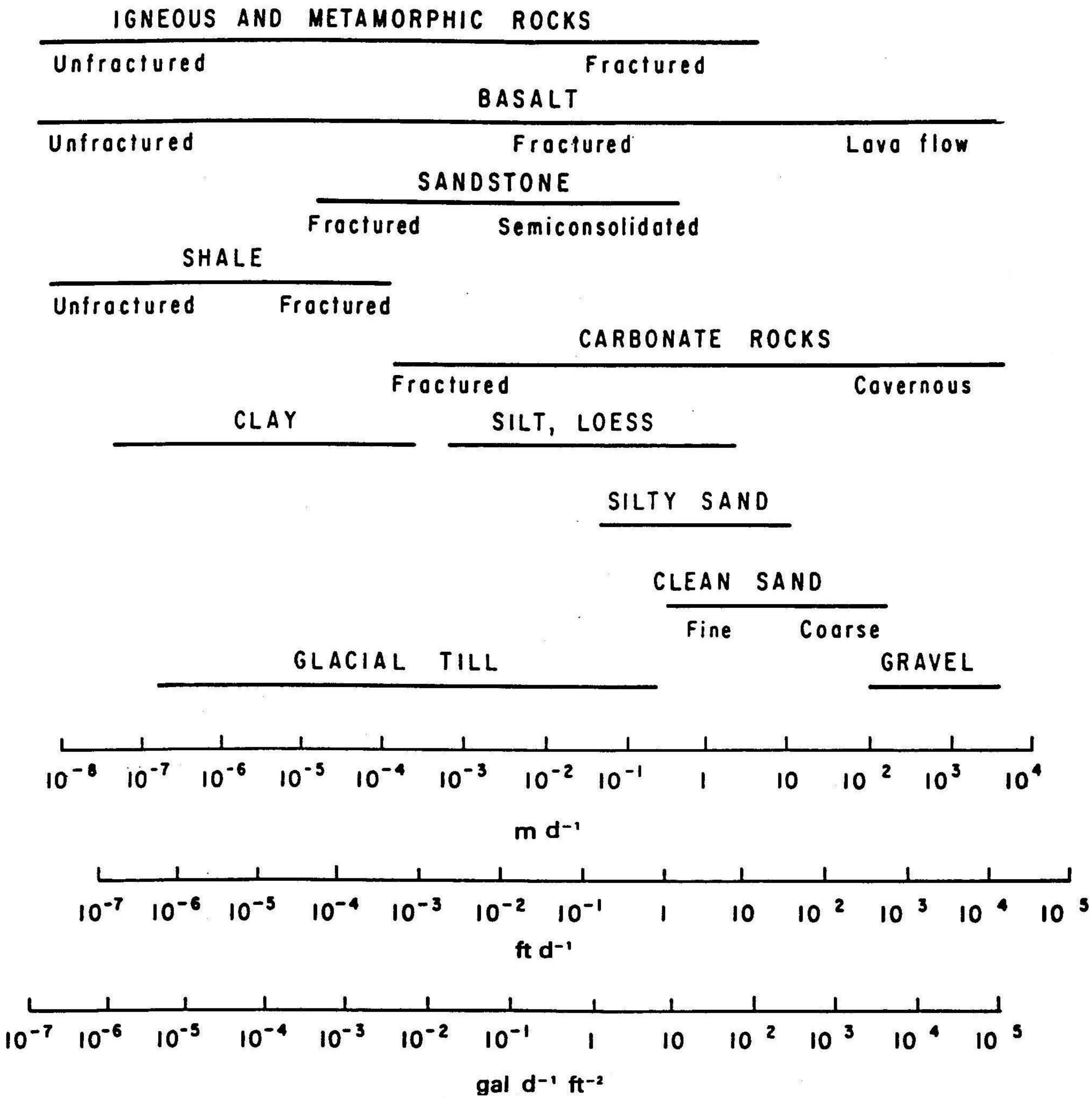
1. Bear, Zaslavsky and Irmay (1968)

<table>
<tr><td>$-\log_{10}$ K(cm/sec)</td><td>−2</td><td>−1</td><td>0</td><td>1</td><td>2</td><td>3</td><td>4</td><td>5</td><td>6</td><td>7</td><td>8</td><td>9</td><td>10</td><td>11</td></tr>
<tr><td>Permeability</td><td colspan="4">Pervious</td><td colspan="4">Semipervious</td><td colspan="6">Impervious</td></tr>
<tr><td>Aquifer</td><td colspan="5">Good</td><td colspan="4">Poor</td><td colspan="5">None</td></tr>
<tr><td rowspan="2">Soils</td><td colspan="2">Clean gravel</td><td colspan="3">Clean sand or sand and gravel</td><td colspan="4">Very fine sand, silt, loess, loam, solonetz</td><td colspan="5"></td></tr>
<tr><td colspan="4"></td><td colspan="2">Peat</td><td colspan="3">Stratified clay</td><td colspan="5">Unweathered clay</td></tr>
<tr><td>Rocks</td><td colspan="4"></td><td colspan="3">Oil rocks</td><td colspan="2">Sandstone</td><td colspan="2">Good limestone, dolomite</td><td colspan="3">Breccia, granite</td></tr>
<tr><td>$-\log_{10}$ k(cm^2)</td><td>3</td><td>4</td><td>5</td><td>6</td><td>7</td><td>8</td><td>9</td><td>10</td><td>11</td><td>12</td><td>13</td><td>14</td><td>15</td><td>16</td></tr>
<tr><td>$\log_{10} k$(md)</td><td>8</td><td>7</td><td>6</td><td>5</td><td>4</td><td>3</td><td>2</td><td>1</td><td>0</td><td>−1</td><td>−2</td><td>−3</td><td>−4</td><td>−5</td></tr>
</table>

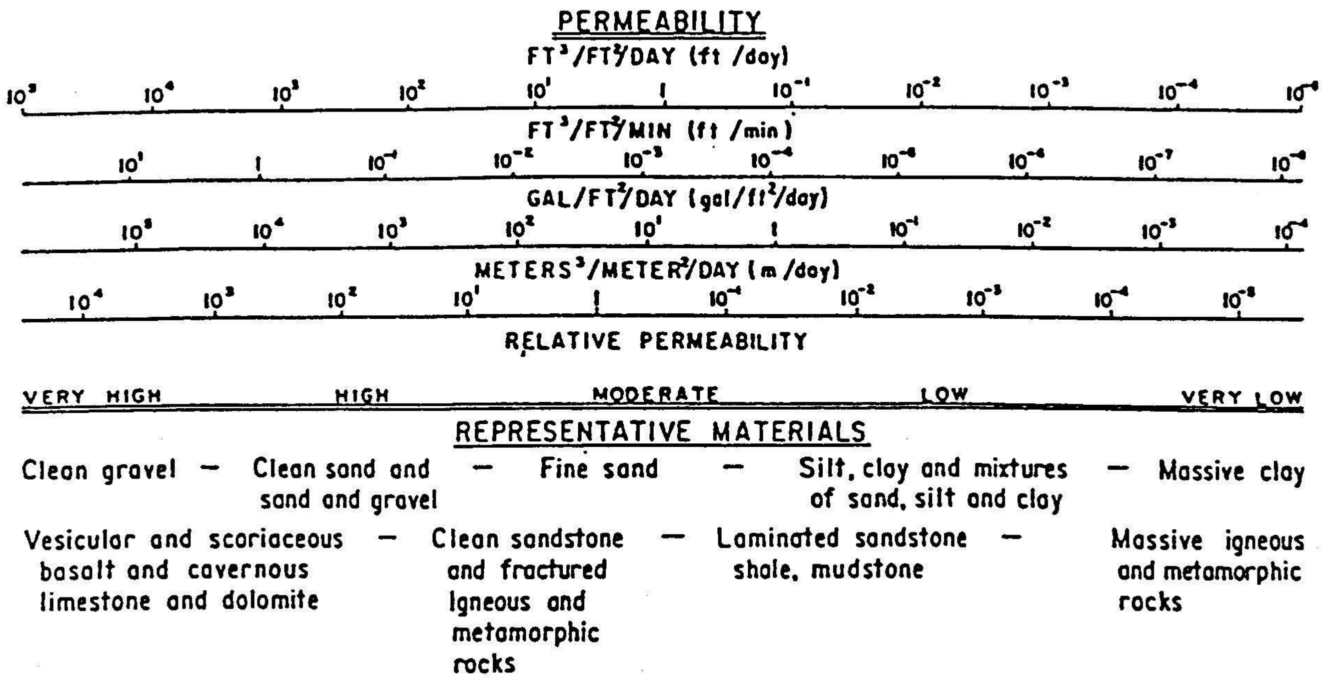
2. Freeze and Cherry (1979)



3. Heath (1983)


IGNEOUS AND METAMORPHIC ROCKS
Unfractured
Fractured
BASALT
Unfractured
Fractured
Lava flow
SANDSTONE
Fractured
Semiconsolidated
SHALE
Unfractured
Fractured
CARBONATE ROCKS
Fractured
Cavernous
CLAY
SILT, LOESS
SILTY SAND
CLEAN SAND
Fine
Coarse
GLACIAL TILL
GRAVEL
10⁻⁸ 10⁻⁷ 10⁻⁶ 10⁻⁵ 10⁻⁴ 10⁻³ 10⁻² 10⁻¹ 1 10 10² 10³ 10⁴
m d⁻¹
10⁻⁷ 10⁻⁶ 10⁻⁵ 10⁻⁴ 10⁻³ 10⁻² 10⁻¹ 1 10 10² 10³ 10⁴ 10⁵
ft d⁻¹
10⁻⁷ 10⁻⁶ 10⁻⁵ 10⁻⁴ 10⁻³ 10⁻² 10⁻¹ 1 10 10² 10³ 10⁴ 10⁵
gal d⁻¹ ft⁻²

4. U.S. Bureau of Reclamation (1995)



5. Younger (2007)

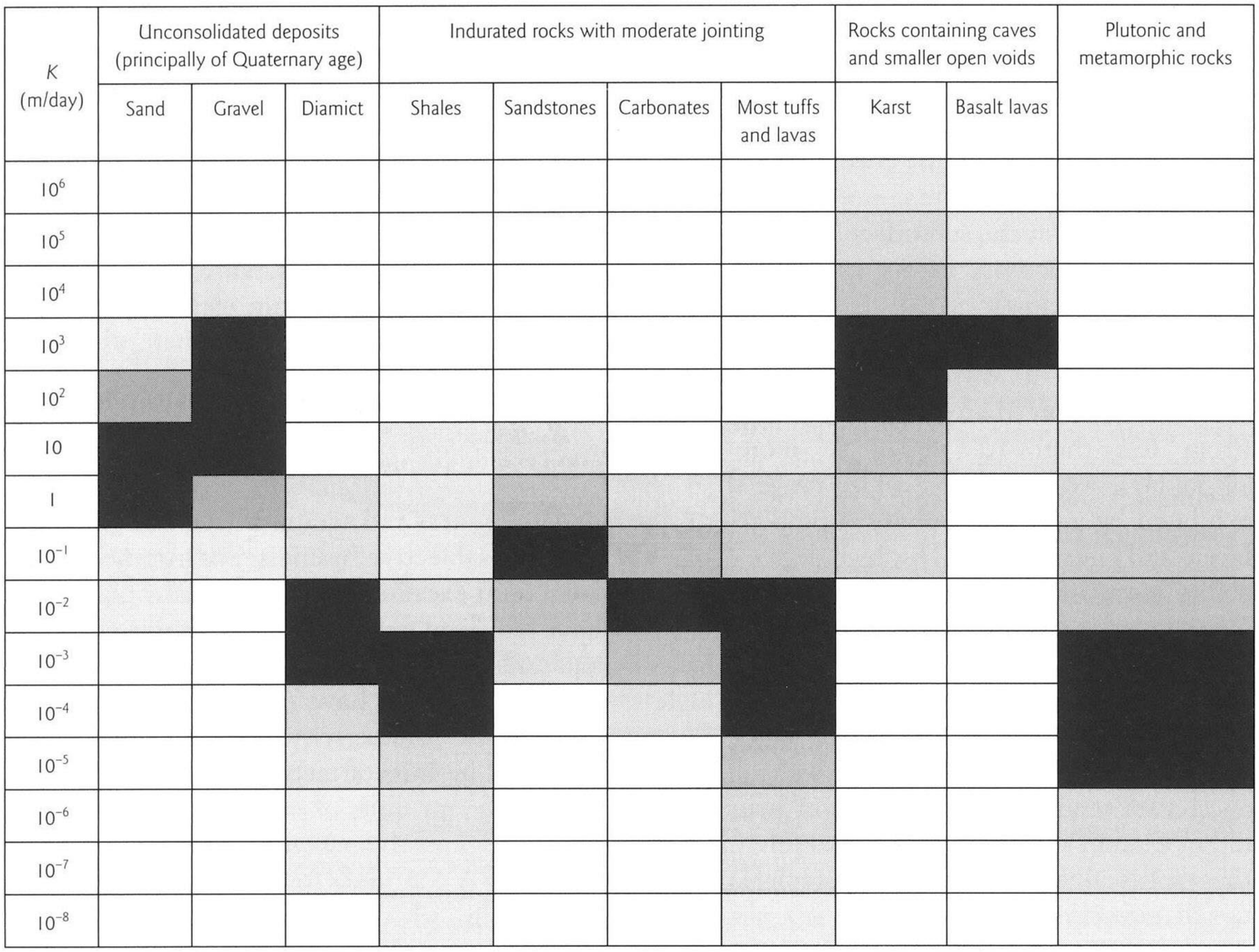
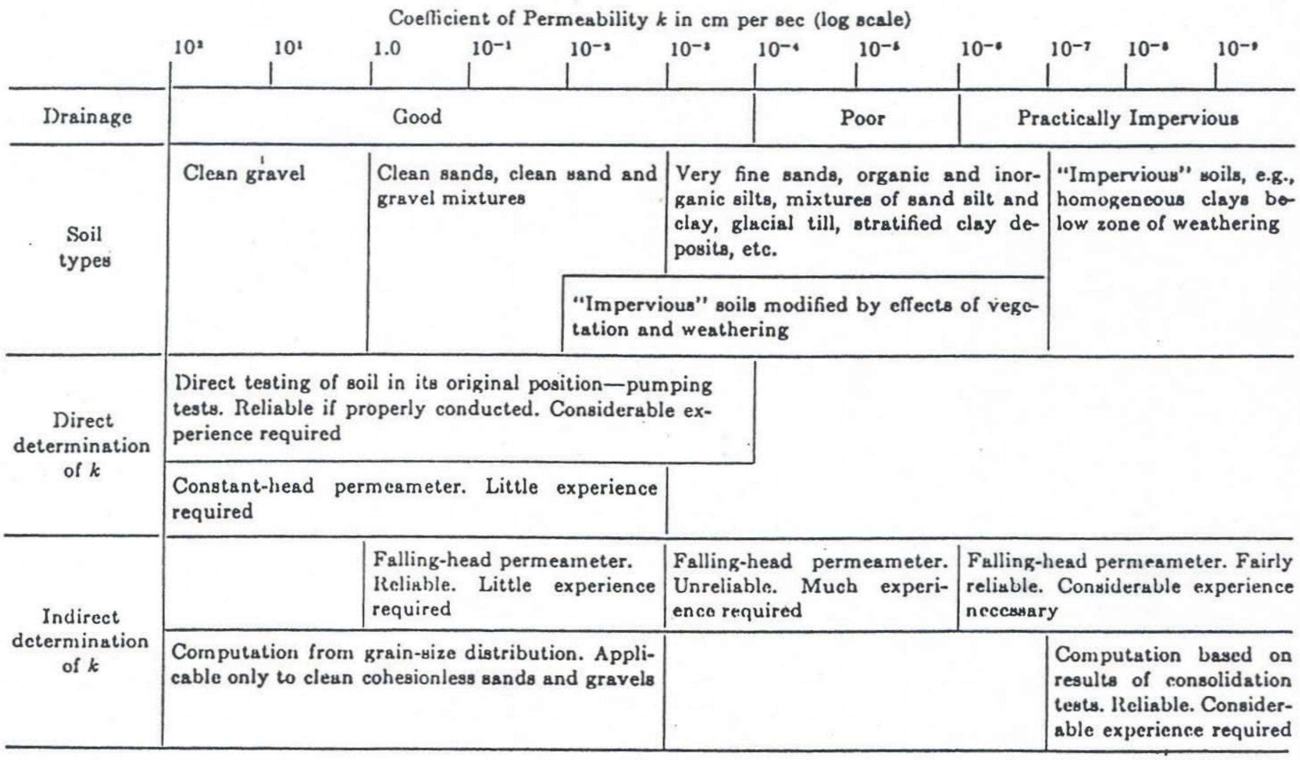


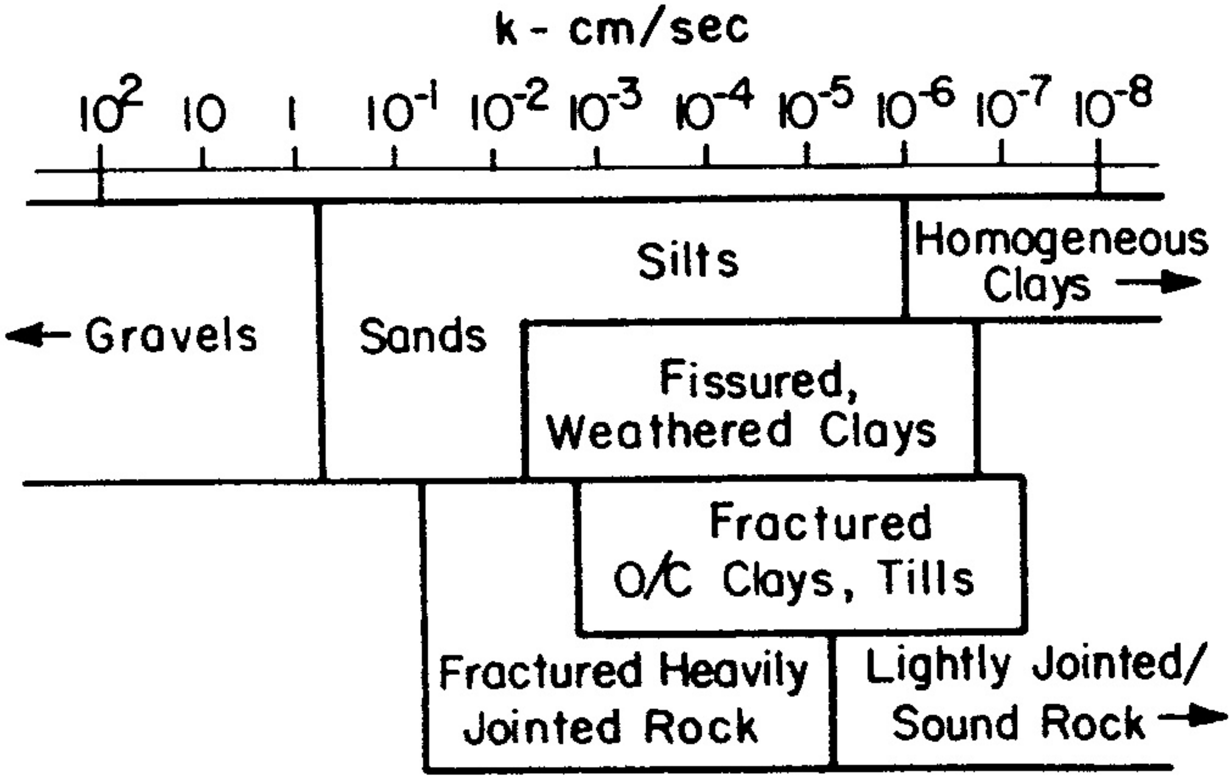
Fig. 3.5 Ranges of hydraulic conductivities encountered in various rock types. The darker the shading, the more common are values in that range for the rocks indicated. "Diamict" refers to deposits in which large clasts are set within a finer-grained matrix. "Carbonates" includes limestones and dolostones. "Karst" refers to classic cave-bearing limestones, and also similarly weathered dolostones and evaporates (especially gypsum). The "Basalt lavas" referred to here are those which contain lava-tubes. The term "plutonic and metamorphic rocks" refers to granites, gneiss, schist, etc. The data upon which this figure was based were collated from datasets for all of North America, summarized by Back et al. (1988).

6. Terzaghi and Peck (1967) [after Casagrande and Fadum]



After A. Casagrande and R. E. Fadum

7. Milligan (1975)



8. Hoek and Bray (1981)

	PERMEABILITY COEFFICIENTS FOR TYPICAL ROCKS AND SOILS			
	k - cm/sec	*Intact rock*	*Fractured rock*	*Soil*
Practically impermeable	10^{-10}	Slate		Homogeneous clay below zone of weathering
	10^{-9}	Dolomite		
	10^{-8}	Granite		
	10^{-7}			
Low discharge poor drainage	10^{-6}			Very fine sands, organic and inorganic silts, mixtures of sand and clay, glacial till, stratified clay deposits
	10^{-5}	Limestone, Sandstone	Clay-filled joints	
	10^{-4}			
	10^{-3}			
High discharge free drainage	10^{-2}		Jointed rock	
	10^{-1}			Clean sand, clean sand and gravel mixtures
	1.0		Open-jointed rock	
	10^{1}			
	10^{2}		Heavily fractured rock	Clean gravel

9. References

Bear, J., D. Zaslavsky, and S. Irmay, 1968: Physical Principles of Water Percolation and Seepage, UNESCO, Paris, 465 p.

Freeze, R.A., and J.A. Cherry, 1979: **Groundwater**, Prentice-Hall, Inc., Englewood Cliffs, New Jersey.

Heath, R.C., 1983: Basic Ground-Water Hydrology, U.S. Geological Survey Water-Supply Paper 2220, United States Geological Survey, Alexandria, Virginia.

Hoek, E., and J.W. Bray, 1981: **Rock Slope Engineering**, 3rd edition, Institution of Mining and Metallurgy, London, U.K.

Milligan, V., 1975: Field measurement of permeability in soil and rock, in *Proceedings of the Conference on In-Situ Measurement of Soil Properties*, North Carolina State University, Raleigh, North Carolina, vol. 2, pp. 3-36.

Terzaghi, K., and R.B. Peck, 1967: **Soil Mechanics in Engineering Practice**, Second edition, John Wiley & Sons, New York, New York.

U.S. Bureau of Reclamation, 1995: Ground Water Manual, Second edition, U.S. Department of the Interior, Bureau of Reclamation, Washington, DC.

Younger, P.L., 2007: **Groundwater in the Environment**, Blackwell Publishing, Malden, Massachusetts, pp. 17-20.

Guidance regarding transmissivity

Christopher J. Neville
S.S. Papadopulos & Associates, Inc.
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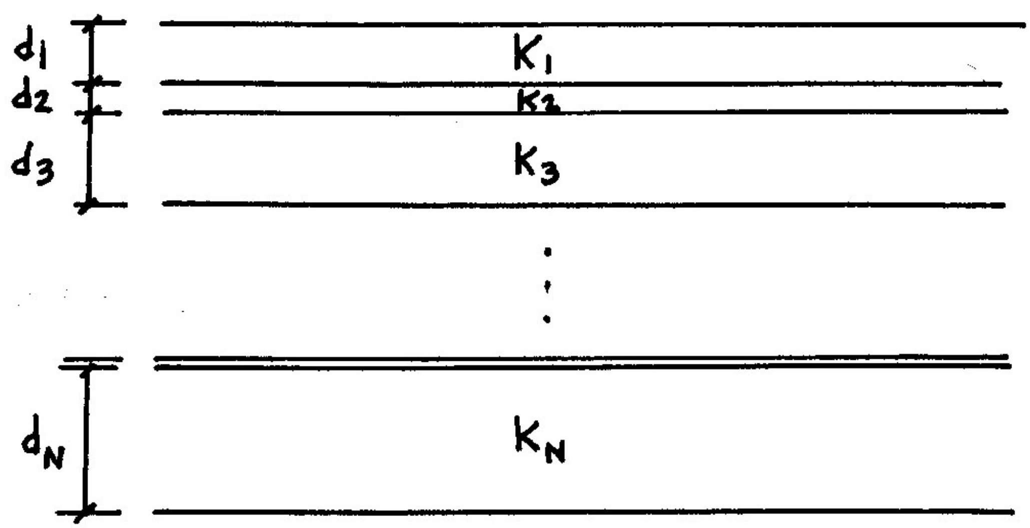
1. Definitions

The transmissivity of a hydrostratigraphic unit is defined as the product of its horizontal hydraulic conductivity and thickness:

$$T = K_H \times B$$

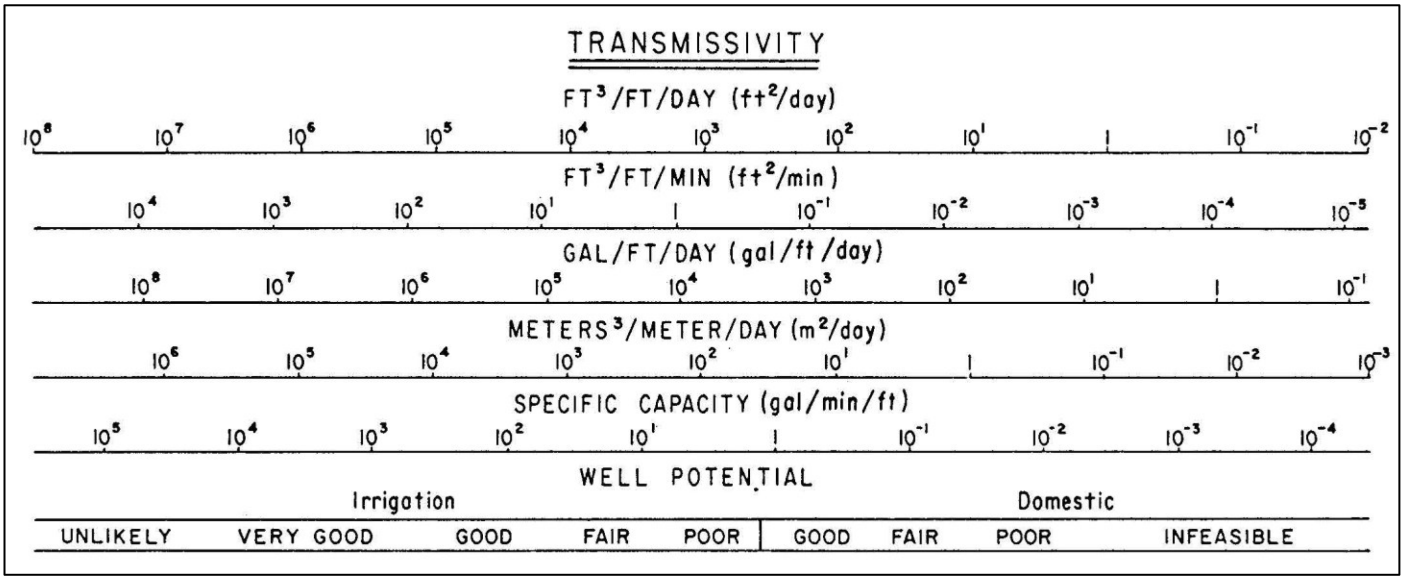
For a unit that consists of strata with different properties, the transmissivity of the unit is the cumulative transmissivity of the strata:

$$T = \sum_{n=1}^{N} K_{Hn} \times d_n$$



2. Relation between transmissivity and well potential

The United States Bureau of Reclamation (USBR) **Ground Water Manual** (2nd edition, 1995) includes a useful chart for interpreting transmissivity estimates in terms of their capacity to support pumping.



The relation between the transmissivity and specific capacity is not indicated explicitly on the USBR chart. The important reference **Groundwater and Wells** (Driscoll, 1986; p. 1021) includes the following approximation for a confined aquifer:

$$T\ [\mathrm{gpd/ft}] = 2000\, SC\ \left[\frac{\mathrm{gpm}}{\mathrm{ft}}\right]$$

Converting to the units adopted for the USBR chart:

$$T\ [\mathrm{ft}^2/\mathrm{d}] = 2000\, SC\ \left[\frac{\mathrm{gpm}}{\mathrm{ft}}\right] \left|\frac{\mathrm{ft}^3}{7.4805\ \mathrm{gal}}\right| \left|\frac{1440\ \mathrm{min}}{\mathrm{day}}\right| = 250\, SC\ \left[\frac{\mathrm{gpm}}{\mathrm{ft}}\right]$$

Example calculations:

- $SC = 10^{-4}$ gpm/ft → $T = 2.5\times10^{-2}$ ft²/day √
- $SC = 1$ gpm/ft → $T = 2.5\times10^{2}$ ft²/day √
- $SC = 10^{4}$ gpm/ft → $T = 2.5\times10^{6}$ ft²/day √

For consistent units, the relation reduces to $\boldsymbol{T = 1.4\, SC}$.

3. Classification of transmissivity (Krasny, 1993)

Krasny (1993) presents an alternative table that can also be used to assess the magnitudes of transmissivity estimates.

Table 1. Classification of Transmissivity Magnitude

Coefficient of transmissivity (m^2/d)	Class of transmissivity magnitude	Designation of transmissivity magnitude	Comparative regional parameters approximately corresponding to the coefficient of transmissivity: Nonlogarithmic — Specific capacity q in l/s m	Logarithmic — Index Y	Ground-water supply potential	Very approximate expected discharge in l/s of a single well at 5 m drawdown
	I	Very high			Withdrawals of great regional importance	> 50
1,000			10	7.0		
	II	High			Withdrawals of lesser regional importance	5 -50
100			1	6.0		
	III	Intermediate			Withdrawals for local water supply (small communities, plants, etc.)	0.5 - 5
10			0.1	5.0		
	IV	Low			Smaller withdrawals for local water supply (private consumption, etc.)	0.05 - 0.5
1			0.01	4.0		
	V	Very low			Withdrawals for local water supply with limited consumption	0.005- 0.05
0.1			0.001	3.0		
	VI	Imperceptible			Sources for local water supply are difficult (if possible) to ensure	< 0.005

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United States Bureau of Reclamation, 1995: **Ground Water Manual**, 2nd edition, U.S. Department of the Interior, Bureau of Reclamation, Washington, DC[Figure 5-4, p. 139].

Literature values of specific storage and storativity

Christopher J. Neville
S.S. Papadopulos & Associates, Inc.
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Overview

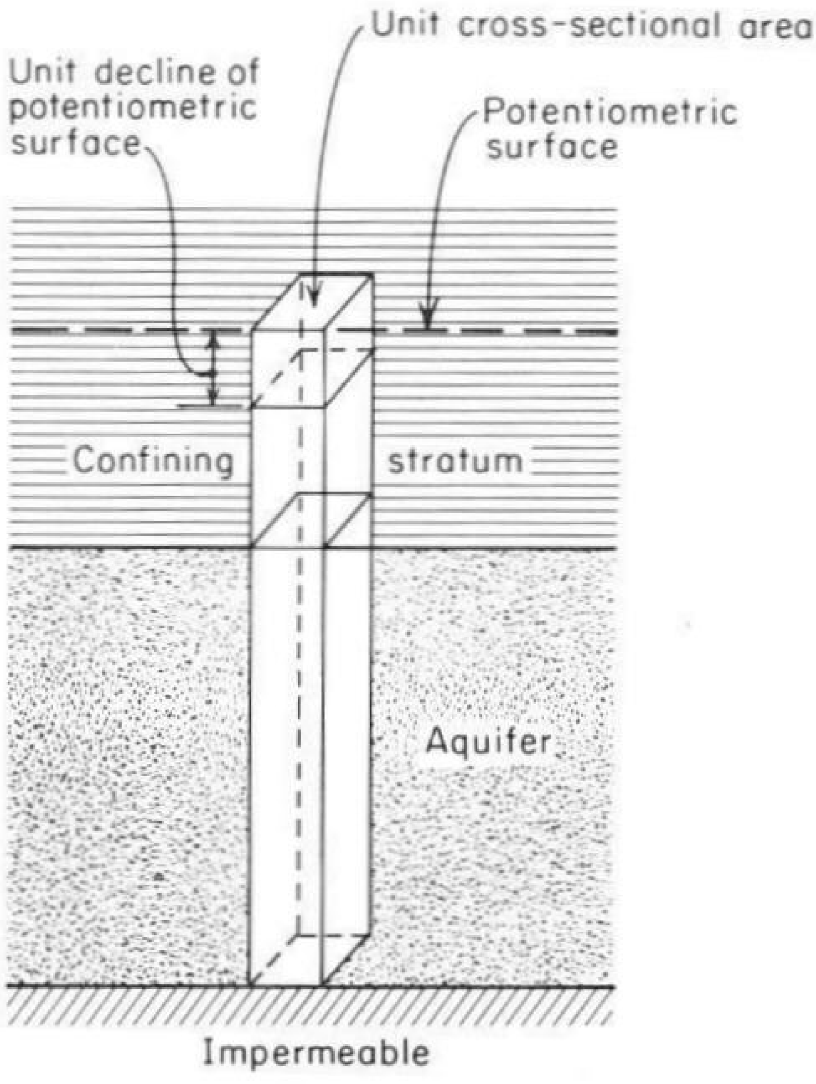
It is generally difficult to quantify with much certainty the specific storage from a pumping test in a confined aquifer. However, the estimates derived from the interpretation of pumping tests may have important diagnostic value. In general, we can say that if the estimated specific storage/storativity is unrealistic, it is likely that the interpretation has missed an aspect of site conditions. As just one example, "strange" estimates of the specific storage/storativity may be obtained if the well penetrates a relatively small portion of the thickness of an aquifer but the drawdown data are interpreted with a model that assumes full penetration. In these notes we present information to better inform our expectations regarding physically realistic specific storage/storativity values. The notes are divided into four main sections:

1. Definitions of specific storage and storativity;
2. Calculation of the specific storage and storativity from fundamental quantities; and
3. Rules-of-thumb for specifying specific storage and storativity values for aquifer materials; and
4. Specification of specific storage values for fine-grained materials.

1. Definitions of specific storage and storativity

The *specific storage, S_s,* of a saturated aquifer is defined as the volume of water that a unit volume of aquifer releases from storage for a unit decline in hydraulic head (Freeze and Cherry, 1979; p. 58).

The *storativity* of a saturated aquifer, also referred to as the *confined storage coefficient*, *S*, is defined as the volume of water that an aquifer releases from storage per unit surface area of aquifer per unit decline in the hydraulic head (Freeze and Cherry, 1979; p. 60). The storativity is given by the product of the specific storage and the aquifer thickness:



2. Calculation of the specific storage and storativity from fundamental quantities

The stored water is released from compaction (consolidation) of the aquifer and expansion of the water.

Jacob (1940; p. 576) derived the following expression for the specific storage:

$$S_s = \rho_w g(\alpha + n\beta) \tag{1}$$

Here ρ_w is the density of water, g is the acceleration due to gravity, α is the compressibility of the aquifer, β is the compressibility of water, and n is the porosity of the sediments.

SI units:

$[\rho]$ = kg/m^3;
[g] = m/s^2;
$[\alpha]$ = m^2/N (or Pa^{-1}, or m-s^2/kg);
$[\beta]$ = m^2/N (or Pa^{-1}, or m-s^2/kg); and
$[n]$ = dimensionless.

The specific storage has units of L^{-1}; for SI units that is m^{-1}.

The *storativity* is calculated as the product of the specific storage and the aquifer thickness:

$$\begin{aligned} S &= S_s B \\ &= \rho_w g(\alpha + n\beta)B \end{aligned} \tag{2}$$

Here B is the aquifer thickness. The storativity is dimensionless.

Typical values for the properties of water

ρ_w = 1000 kg/m^3; and
β = 4.4×10^{-10} m-s^2/kg.

Typical values of the compressibilities of aquifers

Younger (1993) presents "typical" order-of-magnitude values of the compressibility of aquifer materials. These values are reproduced below.

Material	**α (m-s^2/kg)**
Clay	10^{-6}
Silt, fine sand	10^{-7}
Medium sand	10^{-8}
Coarse sand	10^{-9}
Gravel	10^{-10}
Intact rock	10^{-11}

The porosity of sand and gravel aquifers generally ranges from between about 0.10 and 0.40 (Freeze and Cherry, 1979). Using these values, the specific storage for a range of compressibilities and porosities is plotted in Figure 1. As shown in Figure 1, the specific storage depends only weakly on the porosity, and is essentially independent of porosity for aquifer compressibilities greater than 10^{-9} m-s^2/kg.

Example calculations

Material	**Typical α (m-s^2/kg)**	**Typical *n***	**Likely value of S_s (m^{-1})**
Clay	10^{-6}	0.40	1×10^{-2}
Sand	10^{-8}	0.35	1×10^{-4}
Gravel	10^{-10}	0.25	2×10^{-6}

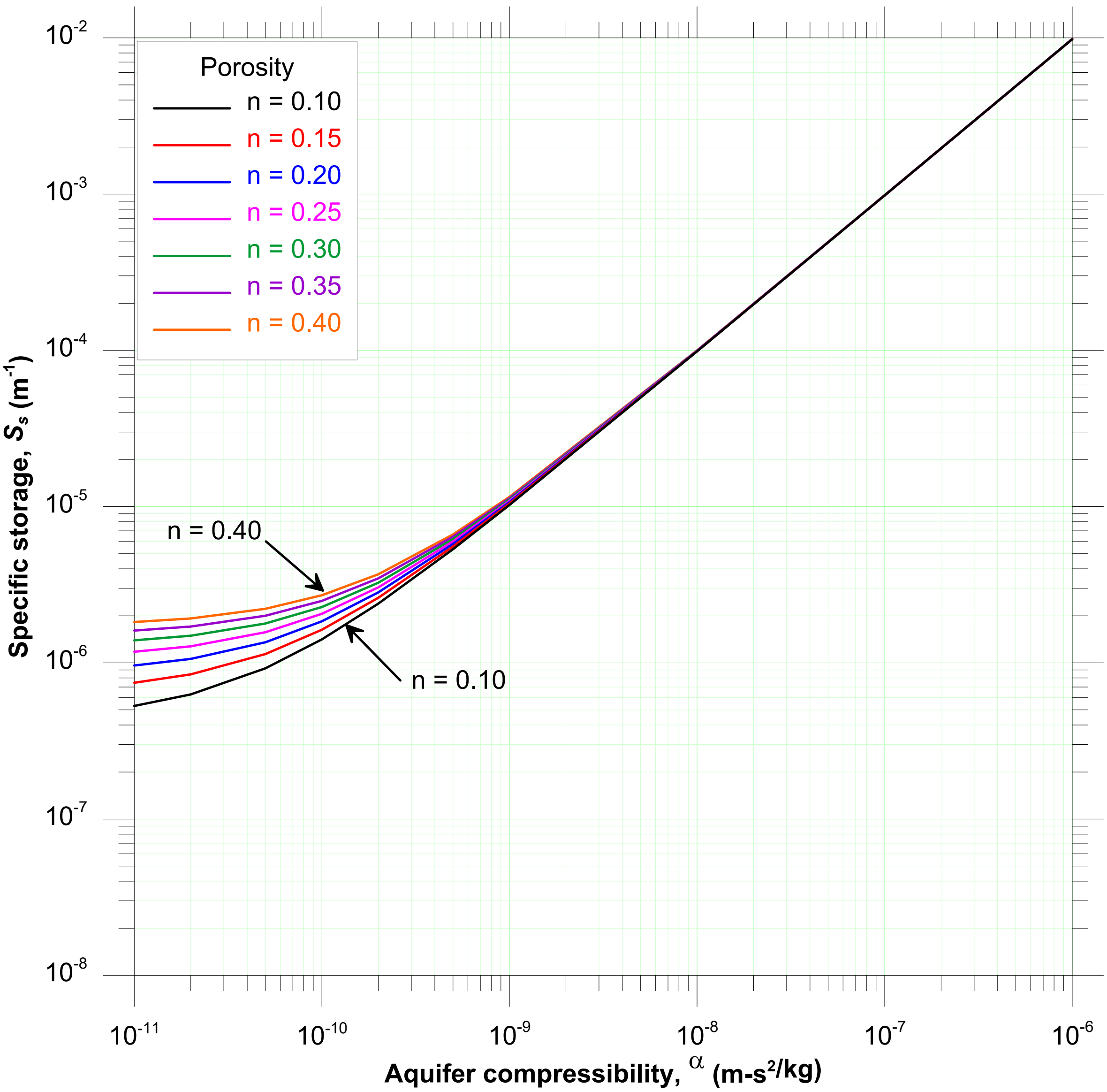


Figure 1. Specific storage for a range of aquifer compressibilities and porosities

3. Rules-of-thumb for specifying specific storage and storativity values for aquifer materials

Rules-of-thumb values of the specific storage are listed below.

Source	Suggested value
Lohman (1972; p. 8)	$S_s \approx 10^{-6}\text{ft}^{-1}\ (3 \times 10^{-6} m^{-1})$
Boonstra (1989; p. 17)	$S_s \approx 10^{-6}$ to $10^{-4}\ \text{m}^{-1}$

- Lohman (1972; p. 8) suggests that a typical value of S_s is 10^{-6} ft^{-1} (3.3×10^{-6} m^{-1}). This corresponds to a soil skeleton compressibility of between 10^{-11} and 10^{-10} m-s^2/kg, which is in the range that Younger (1993) provides for gravel.

- Boonstra (1989; p. 17) suggest that typical values of S_s range from 10^{-6} to 10^{-4} m^{-1}. Values of S_s ranging from 10^{-6} to 10^{-4} (again either ft^{-1} or m^{-1}) correspond to aquifer compressibilities of between 10^{-11} and 10^{-8} m-s^2/kg. Referring to the values of aquifer compressibility listed in Section 2 (Younger, 1993), this range corresponds to the suggested range between gravel and medium sand.

Reported typical ranges for the storativity are listed below. These ranges can be used to check estimates derived from the interpretation of pumping tests.

Source	Suggested value
Lohman (1972; p. 8)	$S \approx 10^{-5}$ to 10^{-3}
Boonstra (1989; p. 17)	$S \approx 5 \times 10^{-5}$ to 5×10^{-3}
Todd (1980; p. 46)	$S \approx 3 \times 10^{-6}\ B$

4. Specifying specific storage values for aquitard materials

The rule-of-thumb values are appropriate for coarse-grained aquifer materials. The values are likely too low to be representative of fine-grained aquitard materials. Batu (1998) has suggested the following representative values of specific storage for clayey materials:

- Plastic clay: 2.5×10^{-3} m^{-1} to 2.0×10^{-2} m^{-1};
- Stiff clay: 1.3×10^{-3} m^{-1} to 2.6×10^{-3} m^{-1}; and
- Medium hard clay: 9.2×10^{-4} m^{-1} to 1.3×10^{-3} m^{-1}.

Konikow and Neuzil (2007) developed a chart indicating the likely range of specific storage for aquitard materials. Their chart is reproduced in Figure 2. Konikow and Neuzil suggest that it is a "generalized relation" for normally consolidated and overconsolidated clayey confining layers. The chart synthesizes data from several sources (Domenico and Mifflin, 1965; Skempton, 1970; Cripps and Taylor, 1981; Tellam and Lloyd, 1981; Burland, 1990; and Neuzil, 1993).

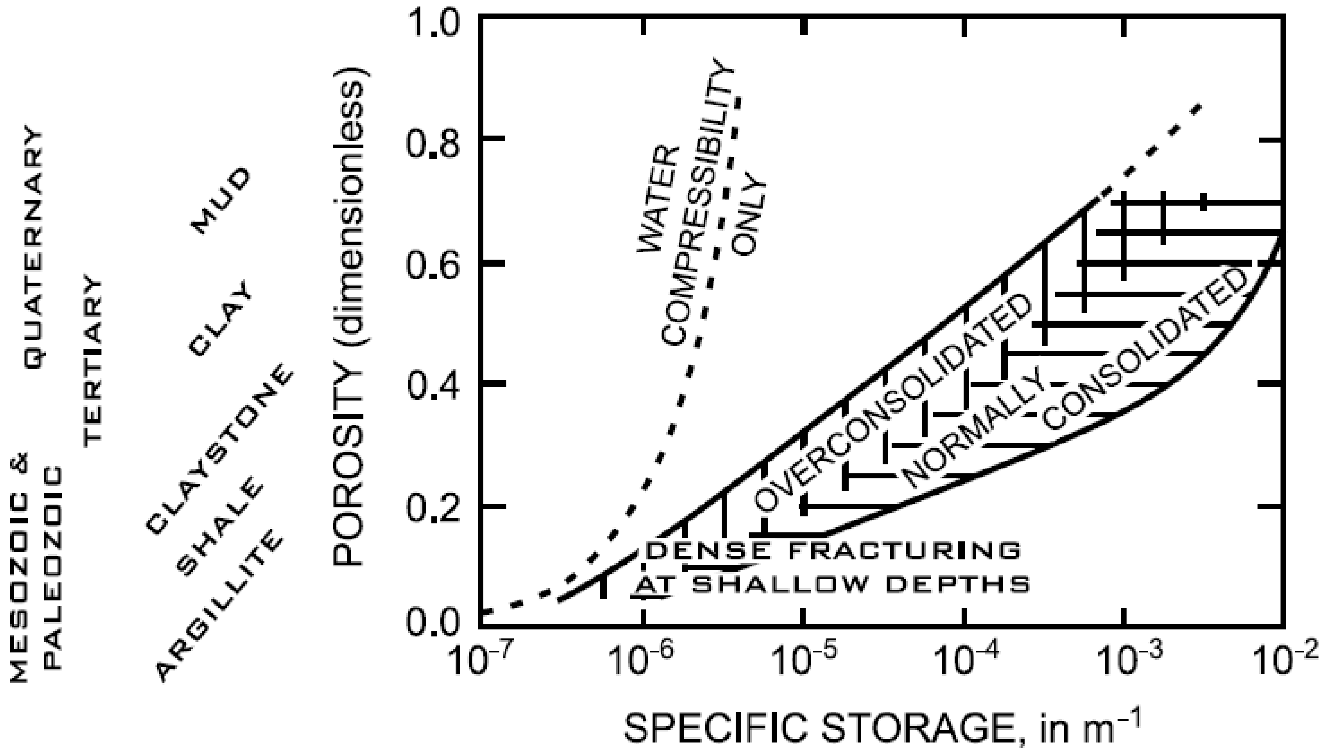


Figure 2. Specific storage for fine-grained materials
Reproduced from Konikow and Neuzil (2007; Figure 5)

The “water compressibility only” curve shown in Figure 2 refers to the specific storage if the porous medium was incompressible. For this case, the specific storage is:

$$S_s = \rho_w g n \beta$$

If the porous medium is incompressible and the porosity is equal to 0.2, the specific storage would be 8.6×10^{-7} m^{-1}.

Some values of specific storage of clay till that have been reported in the literature are tabulated below. The literature values are consistent with the “back-of-the-envelope range of between 10^{-4} to 10^{-2} m^{-1}.

Location	S_s (m^{-1})	Reference	Notes
Grand Forks area, ND	1.0×10^{-4} to 1.2×10^{-3}	Shaver (1998)	Consolidation tests; 107 samples
“Manitoba till”	9.9×10^{-3}	Grisak and Cherry (1975)	Consolidation tests; mean of 34 samples
WRNE, Manitoba	8.2×10^{-3} to 1.6×10^{-2}	Grisak and Cherry (1975)	Consolidation tests; 8 samples
“Saskatchewan till”	1.1×10^{-2}	Grisak and Cherry (1975)	Consolidation tests; mean of 24 samples
Dalmeny, Saskatchewan	1.1×10^{-4} to 1.6×10^{-4}	Keller et al. (1986)	Consolidation tests; 3 samples
Warman, Saskatchewan	2.1×10^{-4} to 3.3×10^{-4}	Keller et al. (1989)	Consolidation tests; 3 samples
“Alberta till”	1.0×10^{-2}	Grisak and Cherry (1975)	Consolidation tests; mean of 27 samples
Pine Coulee, Alberta	1.0×10^{-3}	Smerdon et al. (2005)	Model calibration

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Representative values of the specific yield of unconfined aquifers

Christopher J. Neville
S.S. Papadopulos & Associates, Inc.
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Overview

The specific yield is a fundamental parameter in the interpretation of pumping tests conducted in unconfined aquifers. It is generally difficult to obtain reliable estimates of the specific yield from short-term pumping tests, and an indication of the reason for this difficulty is included in these notes. Although estimates of the specific yield inferred from the interpretation of pumping tests may not be reliable, the estimates may still have diagnostic value. In general, we can say that if the estimated specific yield is unrealistic, it is likely that the interpretation has missed an aspect of site conditions. These notes have been assembled to better inform our expectations regarding physically realistic specific yield values. The notes are divided into five main sections:

1. Definition of specific yield;
2. Calculation of the specific yield from fundamental quantities;
3. Relation between specific yield and median grain size;
4. Tabulation of values of specific yield;
5. Assessment of estimates of specific yield obtained from pumping tests conducted in unconfined aquifers

1. Definition of specific yield

The *specific yield* is defined as the volume of water that an unconfined aquifer releases from storage per unit surface area of aquifer per unit decline in the water table. The specific yield is the unconfined counterpart of the storativity and like storativity it is dimensionless.

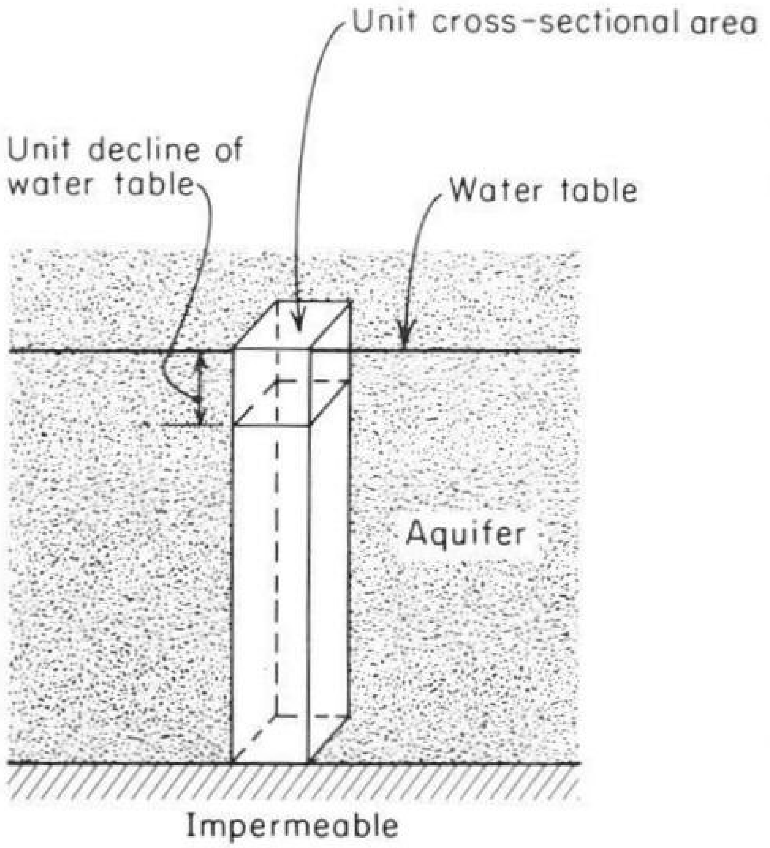

Figure 1. Definition sketch for the specific yield

2. Calculation of the specific yield from fundamental quantities

As shown in Figure 2, the specific yield corresponds to the difference between the saturated water content (porosity), ϕ, and the residual water content, θ_r:

$$S_y = \phi - \theta_r \tag{1}$$

The specific yield is also referred to as the *drainable porosity* and in some references as an *effective porosity*.

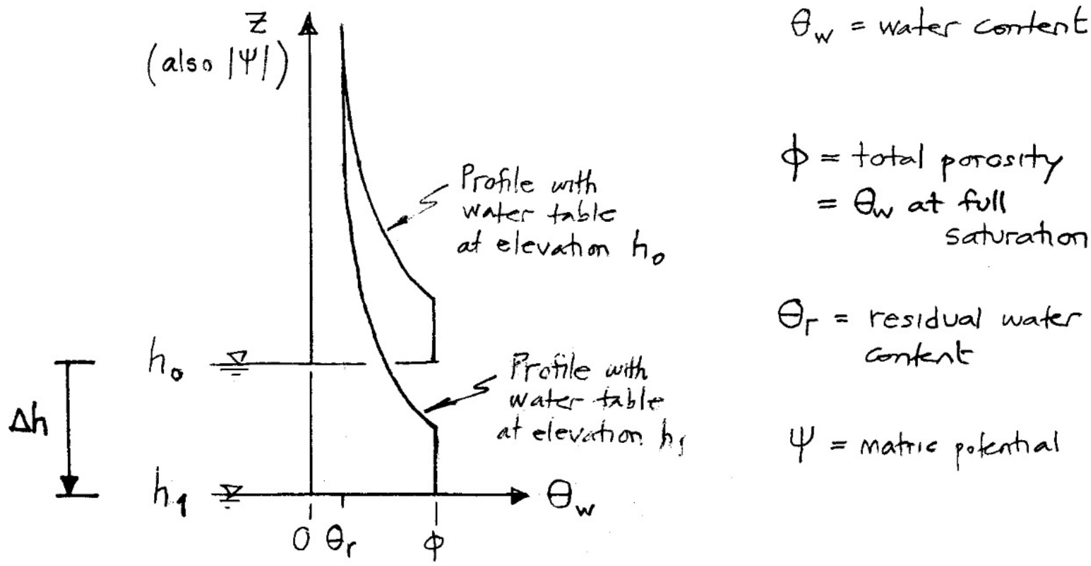

Figure 2. Changes in soil moisture profiles with a declining water table

3. Relation between specific yield and median grain size

Relations between the specific yield and the soil characteristics, expressed as median grain size, are presented in several sources. These relations can be used to obtain first-cut estimates of the specific yield. Relations presented in Stephens and others (1988) are reproduced in Figure 3. Other versions of this figure are presented in Johnson (1967), Davis and DeWiest (1966) and Davis (1969).

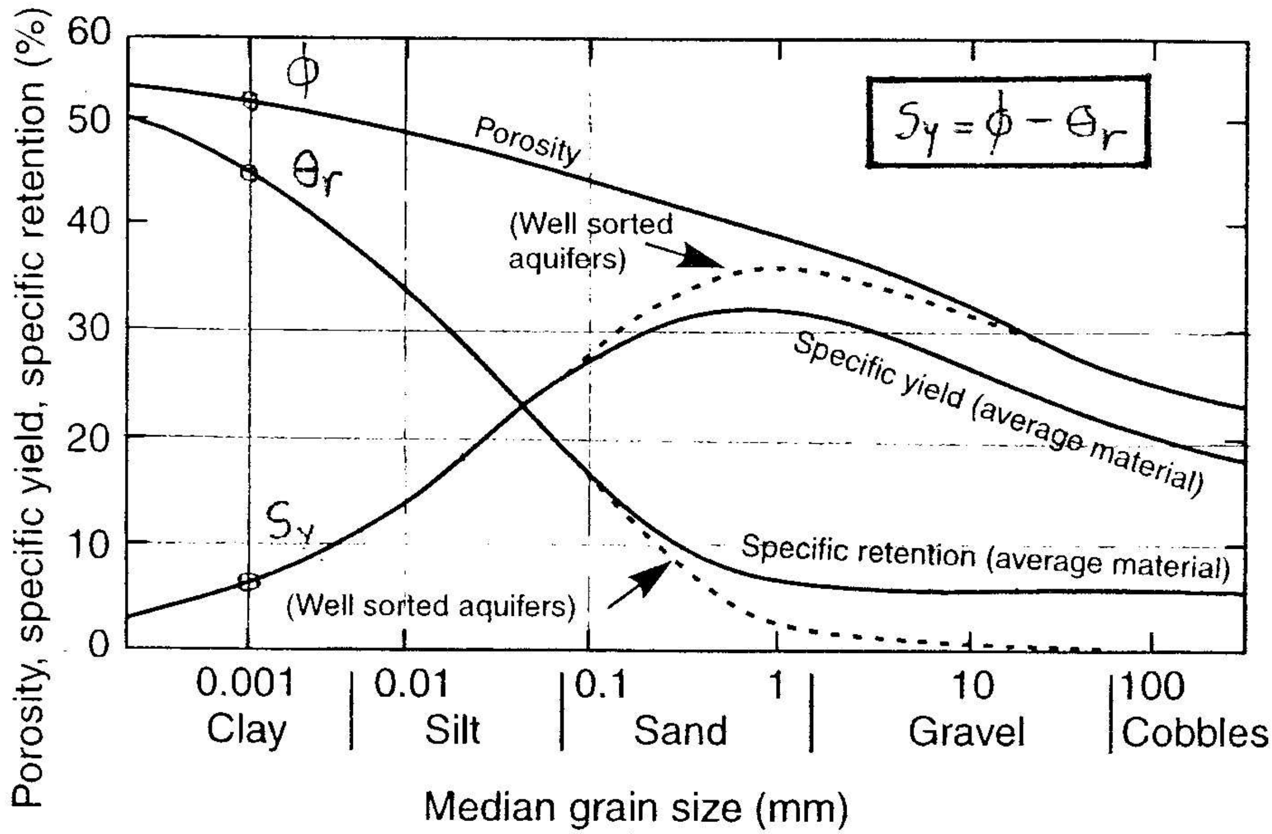

Figure 3. Relation between specific yield and material properties
(Stephens and others, 1988)

4. Tabulation of values of specific yield

There are several excellent compilations of specific yield values. A tabulation of typical values of specific yield assembled by Spitz and Moreno (1996) is reproduced below. These values are obtained from long-term drainage experiments conducted on laboratory-scale columns, as described in Prill and others (1965).

Material	Specific Yield in [l]	Reference
Clay	0.01 to 0.18	A
Gravel, fine	0.13 to 0.40	A
Gravel, medium	0.17 to 0.44	A
Gravel, coarse	0.18 to 0.43	A
Limestone	0.00 to 0.36	A
Loess	0.14 to 0.22	A
Sand, fine	0.01 to 0.46	A
Sand, medium	0.16 to 0.46	A
Sand, coarse	0.18 to 0.43	A
Sand, Eolian	0.32 to 0.47	A
Sandstone, fine	0.02 to 0.40	A
Sandstone, medium	0.12 to 0.41	A
Schist	0.22 to 0.33	A
Schist, weathered	0.06 to 0.21	B
Silt	0.01 to 0.39	A
Siltstone	0.01 to 0.33	A
Tuff	0.02 to 0.47	A

Sources of the tabulated values:

(A) Morris, D. A., and A. I. Johnson, 1967. Summary of hydrological and physical properties of rock and soil materials as analyzed by the Hydrologic Laboratory of the U.S. Geological Survey. USGS Water Supply Paper 1839-D.

(B) Stewart, J. W., 1964. Infiltration and permeability of weathered crystalline rocks, Georgia Nuclear Laboratory, Dawson County, Georgia. USGS Bull. 1133-D, p. 57.

5. Assessment of estimates of specific yield obtained from pumping tests conducted in unconfined aquifers

It is assumed implicitly in Equation (1) that the specific yield corresponds to complete drainage. In reality, some time is required for sediments to drain under gravity. Time-dependent gravity drainage is illustrated with the results of a numerical experiment (Aschenbrenner, 1996). In the experiment shown in Figure 4, the water level in a column of fine sand is dropped instantaneously by 1 m. The sand has a porosity of 44% and the residual water content is 10%. As shown in the figure, even after 100 days, the specific yield is only about 28%, with a very slow approach to its ultimate value of 0.34.

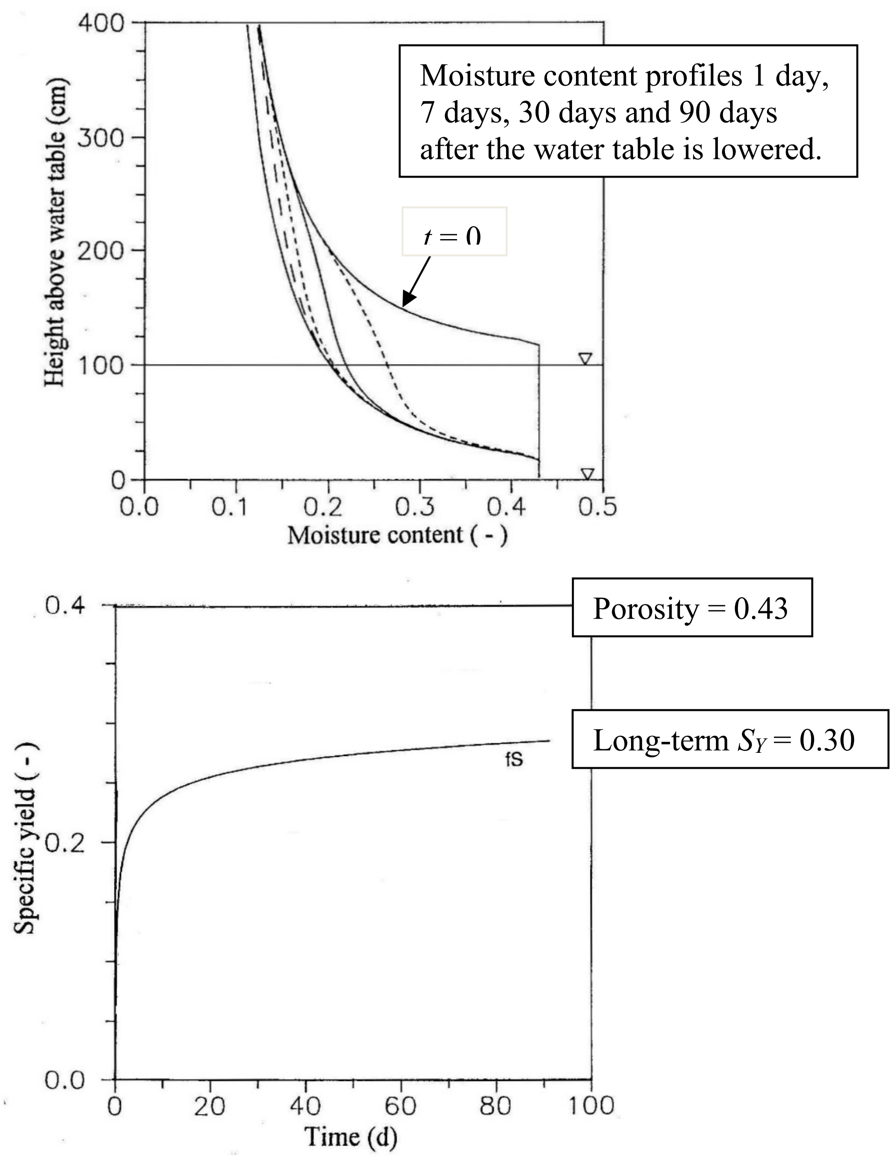

Figure 4. Gravity drainage of fine sand

In theory, since the specific yield is a parameter in the analyses of pumping tests conducted in unconfined aquifers, it should be possible to estimate the specific yield from a pumping test. The simulation results shown in Figure 4 provide insights into the likely reliability of such estimates. As suggested in the figure, gravity drainage may proceed slowly. The decline in the rate of drainage shown in the Aschenbrenner (1996) simulation results are attributed to the decreasing hydraulic conductivity under progressively lower moisture contents. During laboratory column tests Prill and others (1965) observed the rate at which water drained from the pores at the water table diminished through time.

Until relatively recently, the available analytical solutions for pumping tests in unconfined aquifers been developed with the assumption that drainage at the water table occurs instantaneously. Since the duration of typical pumping tests may be brief relative to the time required for complete drainage to occur, estimates of the specific yield derived from pumping tests can be unrealistically small, as suggested in Nwankwor and others (1992) and Akindunni and Gillham (1992). The evolution of the estimated specific yield during a pumping test conducted at the Borden aquifer is shown in Figure 5.

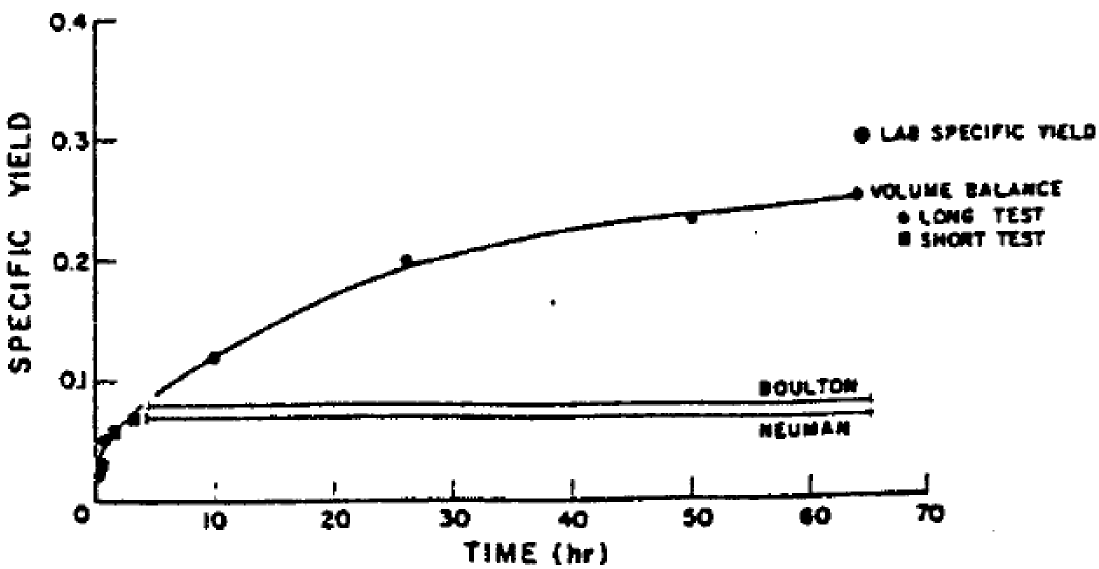

Figure 5. Evolution of the estimate of the specific yield during a pumping test (Reproduced from Nwankwor et al., 1984; Figure 7)

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