RN-9

STUDY OF HYDROGEOLOGICAL PARAMETERS

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CONTENTS

LIST OF SYMBOLS

A	-	Cross-section area
a', b' and n_o	-	Constants
b	-	Thickness of the aquifer
C	-	Constant
d	-	Effective grain diameter
f_s	-	Grain shape factor
f_α	-	Porosity factor
g	-	Acceleration due to gravity
h	-	Pressure head
Δh	-	Contour interval
I	-	Hydraulic gradient
K	-	Hydraulic conductivity
K'	-	Constant
K_u	-	Unsaturated hydraulic conductivity
k	-	Permeability (intrinsic)
L	-	Length of closed contour
M	-	Specific surface of grains
m	-	Specific retention
n	-	Porosity
n_c	-	Specific yield of the capillary fringe
n'	-	Common specific yield
n_s	-	Superficial specific yield

Q	-	Rate of discharge
q	-	Rate of flow per unit area
R, r_w, r	-	Radius
Δr	-	Average distance between two closed contours
S	-	Storativity
S_y	-	Specific yield
S_o	-	Threshold saturation
S_s	-	Degree of saturation
s, s', s_w	-	Drawdown/recovery
T	-	Transmissivity
T'	-	Constant
t, t'	-	Time
U	-	Grain size uniformity of the rock
u	-	Non-dimensional time factor equal to $(r^2S/4Tt)$
V	-	Volume
v	-	Kinematic viscosity
$W(u)$	-	Theis well function
ρ	-	Fluid density
μ	-	Fluid viscosity
dh/dl	-	Hydraulic gradient

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SUMMARY

Frequently, the groundwater modeller faces with the problem of input parameters to the model. Hydrogeological parameters like hydraulic conductivity, transmissivity, storativity and specific yield form part of the model inputs. Estimation of these hydrogeological parameters through conventional pumping test methods is expensive and time consuming. Therefore, use of empirical formulae and graphical methods which are faster and cheaper may be given due consideration for approximate determination of the hydrogeological parameters for reconnaissance groundwater studies. In the present report it has been attempted to review and put together the important available graphical and empirical techniques for estimating specific yield, storativity, transmissivity and hydraulic conductivity of the water bearing formations. The tables, graphs and formulae that have been evolved through vast field and experimental data analysis have been presented at appropriate places in the report so that these tables etc. can be used directly for estimation the hydrogeological parameters with the help of available field data.

1.0 INTRODUCTION

Hydrogeological parameters like hydraulic conductivity transmissivity, storage coefficient, specific yield and porosity form part of the model input in most of the ground-water model studies. Estimation of these hydrogeological parameters through conventional pumping test methods is expensive and time consuming. Therefore, use of empirical formulae and graphical methods which are faster and cheaper may be given due consideration for approximate determination of these parameters. The required preliminary data can easily be acquired from the field studies and with the help of these preliminary and raw data one can estimate the approximate values of these hydrogeological parameters. In the present report it has been attempted to review and put together all the important available graphical and empirical techniques for estimating transmissivity, storage coefficient, specific yield and hydraulic conductivity. The tables, graphs and formulae have been presented at appropriate place in the report so that these tables etc. can be used directly for estimating the hydrogeological parameters.

1.1 Hydrogeological Characteristics of Common Type of Rocks

1.1.1 Metamorphic and igneous rocks

Solid fragments of fresh metamorphic unweathered rocks have porosities of 1 to 3 percent and very often less than 1 percent. The few pores that are present are small and generally not interconnected. Appreciable porosities and permeabilities, however, develop through jointing, faulting and weathering of the rocks. Fractures that are not associated with pronounced faults produce only a small increase in the overall porosity of rocks.

Well yields suggest that permeabilities produced by fracturing of unweathered rock within a few hundred meters of the surface generally range from 10^{-6} to 10^{-2} cm/sec. Owing to the single orientation of most water bearing fractures the permeability of the rock as a whole is strongly anisotropic Effects of weathering may extend more than 100 m into bed rock in regions of intense weathering. Depths of weathering from 2 to 15 m, are normally encountered. The weathered rocks at the surface consist of loose aggregates which have porosities in excess of 35 percent.

The average permeability of metamorphic and plutonic igneous rocks decrease rapidly with depth. This decrease can be attributed to the combined effect of the weight of overlying rock, short penetration of surface disturbances into the bed rock and the filling of pores by secondary mineralization.

In general, yields of wells are low in almost all metamorphic and plutonic rocks. Data from group of wells in different regions show average yields between 50 and 120 litres per minute (lpm). Deeply weathered rocks with substantial local recharge may have mean yields as high as 250 lpm. Differences in well yields usually reflect differences in topography, degree of weathering and orientation of secondary openings. Large yields will be obtained from rocks in moist climates than in dry climates, other factors being equal. This is probably owing to the fact that depths to water are generally less in moist regions, and the water will saturate the more permeable rocks near the surface. Also increased permeability will result from greater circulation of water due to accelerated near surface weathering and increased solution of materials along fractures.

Topography has been found to be an important indication of well yields in certain regions. Higher yields may be obtained from wells drilled and developed on flat and along or in vallyes and broad ravines developed by faulting. The lack of water on or near the steeper slopes can be explained by the fact that erosion has removed much of the weathered and more permeable rock. If rock exposures are quite numerous in the area of interest, detailed geological mapping will be highly useful in determining the extent and orientation of jointing and the location of faults, dikes, and geological contacts. In general, the most favourable water bearing zones are in marble or in dolomite which have been fractured by faulting and partly removed by solution.

The porosity of unfractured volcanic rocks varies from less than 1 percent in dense basalt to more than 85 percent in pumice. Both the permeability and porosity of volcanic rock tend to decrease slowly with geologic time. Some of this decrease is owing to compaction, but the filling of pores with secondary minerals is probably the most important cause of the decrease.

1.1.2 Sedimentary rocks

Shale, claystone, siltstone, and other fine-grained detrital rocks account for roughly 50 percent of all sedimentary rocks. Next in abundance are sandstones, then carbonate rocks, and finally several minor types including conglomerate, gypsum, chert, salt and diatomite. The minor types constitute less than 2 percent of all exposed sedimentary rocks.

Bed thicknesses most commonly range between a few centimeters to a several metres. Although alternating beds of Shale, limestone, and sandstone are characteristic of most sedimentary sequences, individual beds may be so thick that water wells within certain regions will penetrate only one rock type even though the wells are more than 100 m deep.

Siliceous Shale, some claystones, and most argillites will develop closely spaced joints if the rocks are near the surface. The joints and fractures may yield a few litres of water per minute to wells. Most commonly, however, the fine-grained rocks will be barriers to the movement of water. In areas of nearly horizontal strata, the fine-grained rocks serve as widespread confining beds for artesian systems.

The large pore space in many fine-grained sedimentary rocks provides storage for vast quantities of water. The water can be utilized by inducing slow drainage into aquifers by lowering the head in the aquifer.

Porosity of fine-grained sediments decreases with depth and to some extent with age. Newly deposited fine muds will have porosities between 50 and 90 percent. Compaction will force the pore water out of the fine material into adjacent permeable beds of sand so that porosities at depths of several hundred metres will be generally less than 50 percent. At depths of several thousand metres, the porosity will be most commonly less than 25 percent.

Porosity of sandstone ranges from less than 5 percent to a maximum of about 30 percent. The amount of pore space in an individual sample is a function of sorting, grain shape, packing and degree of cementation, of these variables cementation is the most important. Common cementing materials are clay minerals, calcite, dolomite, and quartz (silica). Clay miners may be present as original constituents or as products of diagenesis. Rocks cemented with clay are not usually as firm as other sandstones. The porosity of clay cemented sandstones tend to be quite high because the clay itself has considerable porosity.

Limestone and dolomite, the two common carbonate rocks, originate from a large number of different sedimentary deposits such as inorganically precipitated limey muds, shell fragments talus deposits, calcite sand, and accumulations of the remains of small planktonic organisms. Original porosity is relatively high in most young limestones. Permeability may range from less than 10^{-6} cm/sec for partly cemented coarse breccia. Intermediate values of 10^{-6} to 5×10^{-4} cm/sec are more common, for limestones having some primary porosity.

Carbonate rocks with extensive solution channels or fractures primarily developed in one direction will have bulk permeabilities that will be strongly anisotropic. The direction of groundwater flow cannot , therefore, be predicted by simply drawing orthogonal lines to the groundwater level contours. Tracer techniques are of great help in such studies.

1.1.3 Unconsolidated earth materials

The hydrogeological investigations indicate that unconsolidated deposits are the most important sources of water. The reasons for their importance are: (i) appreciably greater hydraulic conductivities of these deposits than other earth materials, (ii) larger volume of pore space than consolidated rock, (iii) ease of drilling resulting in rapid and relatively inexpensive development, (iv) proximity of ground water levels to surface, and (v) recharging of sediments during peak surface flows, by flooding, by through -

bank flow, and artificially by irrigation and water spreading.

Generally, aquifers consists of unconsolidated or partly consolidated gravel and sand. They are located in abandoned or burried valleys, in plains and in intermontane valleys. Some are of a limited area, others may extend over large areas. Their thickness may also vary from several metres to several hundred metres.

From the standpoint of ground water production the most important aquifers are alluvial formations comprising of sand and gravel deposits. The vertical succession of most alluvial channel deposits is coarse to fine. The relative thickness of representative units depends on the sediments carried.

2.0 REVIEW

2.1 Specific Yield

A part of the water is retained in interstices largely by the forces of molecular attraction, adhesion, and cohesion. The walls of interstices retain a film of water by adhesion and some water is retained in detached interstices. The amount of water retained varies directly as the aggregate surface of the interstices and indirectly as the size of the interstices; thus retention is greatest in rocks having small interstices (Ex. clay). The amount of water retained in interstices also depends on the time of drainage; on the temperature and mineral composition of groundwater which affect its surface tension, viscosity, and specific gravity.

The storage term for unconfined aquifer is known as the 'Specific yield' (S_y). The specific yield of a rock or soil is the ratio of the volume of water which the rock or soil, after being saturated, will yield by gravity drainage to the total volume of the saturated rock or soil (Johnson, 1972). The definition implies that gravity drainage is complete. It is equal to porosity minus specific retention. Figure 1 illustrates the concept schematically.

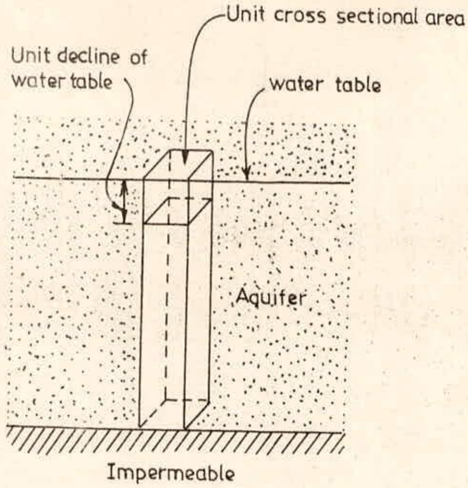

FIGURE 1 - SCHEMATIC REPRESENTATION OF STORATIVITY OF UNCONFINED AQUIFER

The idea of specific yield is best visualised with reference to the saturated - unsaturated interaction it represents. Figure 2 shows the water-table position and the vertical profile of moisture content vs.depth in the unsaturated zone at two times, t_1 and t_2. The shaded area represents the volume of water released from storage in a column of unit cross-section. If the water-table drop represents the specific yield. The usual range of specific yield in unconfined aquifer is 0.1 to 0.30. The higher values reflect the fact that release from storage in unconfined aquifers represent an actual dewatering of the soil pores, whereas releases from storage in confined aquifers represent only the secondary effects of water expansion and aquifer compaction caused by change in the fluid pressure. The favourable storage properties of unconfined aquifers make them more efficient for exploitation by wells. When compared to confined aquifers, the same yield can be achieved with smaller head changes over less extensive areas.

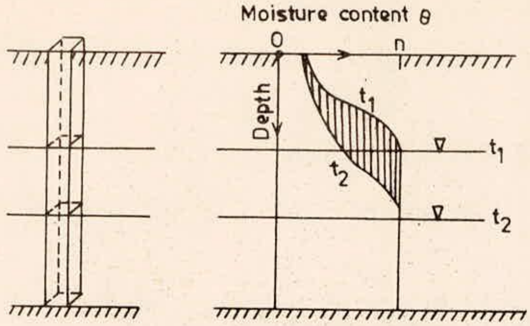

FIGURE 2 - CONCEPT OF SPECIFIC YIELD VIEWED IN TERMS OF THE UNSATURATED MOISTURE PROFILES ABOVE THE WATER-TABLE

Nosova (1962) studied the drainage process in aquiferrous sands by stages and he proved that the instantaneous value of the specific yield varies widely with time and therefore cannot be considered as a constant of the area studied. On this basis he proposed to use several parameters:

a) The specific yield (n_o); is a final characteristic of the gravity drainage; it represents the difference between the water saturation capacity (equal to porosity n) and the specific retention m, ($n_o = n - m$). Since the drainage is an unsteady flow which cannot be achieved in due time the computed value of the specific yield is lower than its maximum value n_o.

b) The specific yield of the capillary fring (n_c); represents the difference between the saturation capacity and the capillary capacity and is considered as a mean value along the height of the sample ($n_c = n - \omega$). Under actual conditions, this parameter characterizes the average drainage state of the rock in the area of capillary ascension due to the existence of the depression surface.

c) The common specific yield (n'); characterizes the draining process in time and is given by the ratio between the volume of water stored from the beginning of the drainage process upto the considered time and the volume of drained ground. Consequently, the value of this parameter depends on the time of drainage.

d) The superficial specific yield (n_s); is defined as the ratio between the released water volume during a certain time period and the drained rock volume. It is to be remarked that in hydrodynamic analysis of the draining process, the superficial specific yield is quite important (Gheorghe, 1978). If the draining is considered as a process which takes place by isolated stages, i.e. as uniform regime, it appears that coefficient n_s depends on the decreasing velocity of the water current surface, expressed by the hydraulic gradient (I). For 'I' equal to 0.001 to 0.02 the approximate value of the superficial specific yield is supposed by the following semi-empirical relation (Nosova, 1962),

$$n_s = n_o \; (1 - I)$$

For very small hydraulic gradients ($I < 0.001$) the coefficient n_s can be considered as a constant, practically equal to the specific yield n_o. Nosova offers some data concerning the relation between superficial specific yield n_s and grain size uniformity of the rock U:

For,

$I = 0.1 - 0.01 \quad n_s = (0.7\text{-}0.8).n_o$, for sands with $U<5$

$n_s = (0.6\text{-}0.7).n_o$, for sands with $U>5$

$I = 0.01 - 0.001 \quad n_s = (0.8\text{-}0.9).n_o$, for sands with $U<5$

$n_s = (0.7\text{-}0.8).n_o$, for sands with $U>5$

2.2 Methods of Estimating Specific Yield

The specific yield of the soil in the zone of water-table fluctuation must be determined in order to estimate the water that is stored due to an increment of rise in the water-table during the period of recharge, as well as the water that can be obtained for each stages of lowering of the water-table. The specific yield also is an important parameter for groundwater balance studies. Various norms have been fixed from time to time by different agencies to consider the value of specific yield. But, it has been practically observed that these norms don't suit at all places. The estimates of specific yield which might be slightly more or less than the average could be obtained from the following methods:

2.2.1 Laboratoy methods

i) Sample saturation and drainage: Columns of saturated aquifer material are drained by gravity. The volume of the aquifer material drained and the volume of the water are determined. The volume of water yielded can be measured directly or can be computed from the porosity and the moisture content after drainage. Then the

$$\text{Specific yield (\%)} = \frac{\text{Volume of water drained}}{\text{Total volume of the sample}} \times 100.$$

Table 1 shows the specific yield values determined by drainage method for U.P. area (UPIRI).

Table 1 - Specific Yield as Determined by Sample Saturation and Drainage

Sl.No.	District	No.of samples	Average specific yield (%)
1.	Saharanpur	23	19.40
2.	Meerut	24	21.49
3.	Muzaffarnagar	06	20.67
4.	Gaziabad	16	21.65
5.	Bulandshahar	20	20.81
		Mean	20.81

ii) Mechanical analysis of the aquifer material: The effective size or median diameter of the aquifer material can be determined using mechanical analysis of the sample. The approximate specific retention can be determined with the help of available standard plot of specific retention vs. effective size or median diameter as shown in Figure 3.

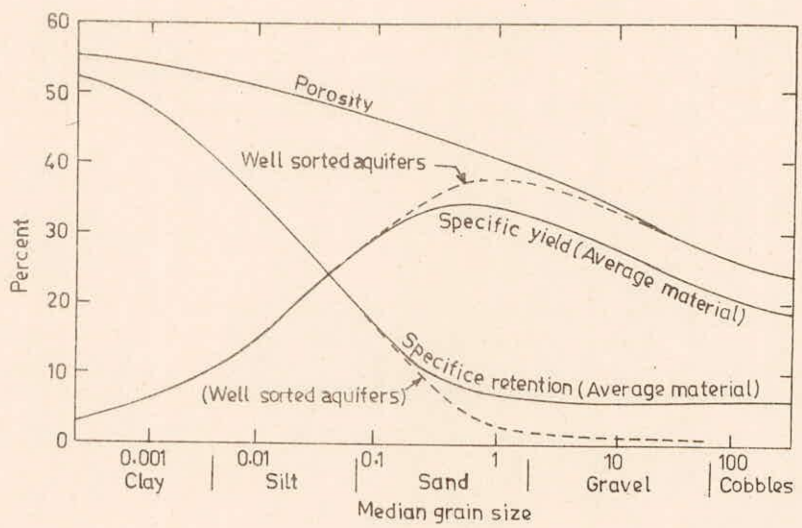

FIGURE 3 - RELATION BETWEEN MEDIAN GRAIN SIZE AND WATER-STORAGE PROPERTIES OF ALLUVIUM FROM LARGE VALLEYS

Similarly the porosity can be determined for the known median diameter. The specific yield can then be given as,

Spefici yield = Porosity - Specific retention.

The representative specific yield values for different rock types are given in Table 2.

Table 2 - Representative Values of Specific Yield (After Johnson, 1967)

Material	Specific yield percent
Gravel, Coarse	23
Gravel, Medium	24
Gravel, fine	25
Sand, coarse	27
Sand, medium	28
Sand, fine	23
Silt	8
Clay	3
Sandstone, fine grained	21
Sandstone, medium grained	27
Limestone	14
Dune sand	38
Loess	18
Peat	44
Schist	26
Siltstone	12
Till,predominantly silt	6
Till, predominantly sand	16
Till, predominantly gravel	16
Tuff	21

iii) Centrifuge moisture equivalent: The experimental works have indicated that for some medium textured materials, the moisture equivalent approximates specific retention. Therefore, specific yield is estimated indirectly by centrifuging to measure moisture equivalent.

2.2.2 Field methods

i) Sampling after lowering of water table: Pumping is done in an aquifer, and after appreciable lowering of the water table, aquifer samples are taken from the zone immediately above the capillary fringe. The moisture content and porosity of this sample are then determined to estimate the specific yield;

Specific yield = Porosity - Specific retention.

ii) Pumping method: A known volume of water is pumped out and the volume of the sediment that is being drained is noted by observing the depth of water-table lowered. The specific yield is then calculated as;

$$\text{Specific yield (\%)} = \frac{\text{Volume of water pumped out}}{\text{Volume of sediment drained}} \times 100.$$

iii) Recharge method: This is the reverse of the pumping method mentioned above. The specific yield is computed by knowing the volume of sediments saturated by a known (measured) amount of seepage from one or more streams.

iv) Neutron logging: Neutron logs, especially when properly calibrated in terms of moisture contents, helps to determine specific yield.

v) Pumping test method: The specific yield values may be computed from pumping tests and results obtained from detailed work done by different agencies may be adopted for similar hydrogeological situation. As a guide the following specific yield values for different types of geological formations in the zone of water level fluctuation may be adopted (Government of India, 1984).

1. Sandy alluvial area	12 - 18%
2. Valley fills	10 - 14%
3. Silty/clayey alluvial area	5 - 12%
4. Granites	2 - 4%
5. Basalts	1 - 3%
6. Laterites	2 - 4%
7. Weathered phyllites, Shales, Schists and associated rocks	1 - 3%
8. Sandstone	1 - 8%
9. Limestone	3%
10. Highly Korstified limestone	7%

The specific yield values have been worked out by CGWB and others for different formations in the hard rock regions. Table 3 shows the representative specific yield value for hard rock aquifers in India (Saxena, 1985).

2.3 Coefficient of Storage

The storage property of a confined aquifer was given quantitative significance by Theis (1935) who introduced the coefficient of storage 'S' in his classic equation. The current version of Theis definition of the storage coefficient is; the volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head.

The storage coefficient of most confined aquifers range from about 0.00001 to 0.001 and is about 10^{-6} per foot (3.3×10^{-6} per metre) of aquifer thickness. In contrast, the specific yield of most unconfined aquifers ranges from about 0.1 to 0.3.

2.3.1 Estimation of storage coefficient

The analysis of pumping tests play an important role in determining the value of storage coefficient. However, in examining well logs of confined aquifers or measuring sections of exposed rocks that dip down beneath confining bed to become confined aquifers, the storage coefficient may

Table 3 - Specific Yield for Aquifers

Sl.No.	Area/Formation	Specific yield(%) Range	Average	Method of Estimation
1	2	3	4	5
1.	Pre-Cambrian Crystalline Rocks			
i)	Weathered granite in A.P.	0.5-4.0	2.2	Long duration pumping test with 9 observation wells
ii)	Granite schists in A.P. and Karnataka	1.0-4.0	-	Long duration pump tests
iii)	Granite and gneiss in A.P.	1.0-14.0	4.0	Test on dugwells with observation wells or pumped well analysis on 12 wells
		2.0-5.2	2.9	By hydrographic analysis of wells
iv)	Khondalite in A.P.	1.7-3.2	2.5	By hydrographic analysis of wells
v)	Granite and gneisses in Maharashtra	1.0-8.0	-	Not known
vi)	Granite and gneisses in Rajasthan	0.4-2.5	0.9	Hydrographic analysis of wells
2.	Pre-Cambrian Consolidated Sedimentary Rocks			
i)	Schists in Delhi	0.2-2.1	1.0	Hydrographic analysis of wells
ii)	Lime Stones and shales in A.P.	0.8-5.5	3.0	Hydrographic analysis of wells
iii)	Highly Karstified limestone in Rajasthan	N.A.	7.0	Pump tests
iv)	Sandstones in Rajasthan	0.5-4.0	1.6	Hydrographic analysis of wells

Table 3 (Continued)

1	2	3	4	5
	v) Sandstones in U.P. and M.P.	0.65-5.8	3.6	Hydrographic analysis of wells
3.	Gondwana Aquifers			
	i) Sand stone in Maharashtra	4.0-8.0	6.0	Pump test
	ii) Sand stone in Rajasthan	1.6-5.7	3.0	Hydrographic analysis of wells
	iii) Sand stone in Kutch Gujarat	0.4-3.9	3.0	Hydrographic analysis of wells
4.	Deccan Traps			
	i) Sinaman Project Area in Maharashtra			
	a) Weathered MassiveBasalt	1.8-2.5		
	b) Weathered Vesicular Basalt	1.6-2.22		
	c) Fractured MassiveBasalt	1.6-2.2		
	d) Fractured Vesicular Basalt	2.0-3.0		
	d) Massive/Basalt	Almost nil		
	e) Vesicular Basalt	Insignificant		
	ii) a) Weathered Trap in A.P. and Karnataka	2.0		
	b) VEsicular Trap	1.0		

Table 3 (continued)

1	2	3	4	5
iii)	a) Yellow clay (Intertrappean) in U.P. and M.P.	0.007-0.67	0.34	
	b) Weathered Basalt in M.P.	0.2-16.0	3.0	
	c) Massive Basalt in M.P.	0.3-2.2	1.75	

be estimated from the following rule of thumb relationship:

Thickness of aquifer (ft)	Storage coefficient(S)	Specific Storage
1	10^{-6}	
10	10^{-5}	
100	10^{-4}	10^{-6}
1000	10^{-3}	

Values presented in the abovetable have been suggested without giving due consideration for porosity or for compressibility of the aquifer.

2.4 Hydraulic Conductivity

Hydraulic conductivity of a porous medium is the volume of water at the existing kinematic viscosity that will move in unit time under a unit hydraulic gradient through a unit area measured at right angles to the direction of flow.

The hydraulic conductivity is a parameter reflecting both the intrinsic rock permeability and the physical properties of the water. It represents as a whole, the facility of the water circulation within an aquifer. It is condiered that for gasless fresh water at low temperatures hydraulic conductivity is a constant. In actual cases however it was remarked that the value of this parameter depends on the direction of the water flow (horizontal, vertical, ascending or descending) in the rocks as well as on the value of the hydraulic gradient (Gheorghe, 1978).

The hydraulic conductivity of a soil or rock depends on a variety of physical factors, including porosity, particle size and distribution, shape of particles, arrangement of particles, and other factors (Le Grand, et al, 1971; Rumer, 1962). In general, for unconsolidated porous media, hydraulic conductivity varies with particle size (Figure 4); clay materials exhibit low values of hydraulic conductivity, whereas sands and gravels display high values.

An interesting illustration of the variation of hydraulic conductivity with particle size was reported by Todd (1980) and is shown in Figure 5. Here conductivities were measured for two uniformly sieved sands. These two sands were then mixed in varying proportions, and the corresponding hydraulic conductivities were again determined. Results show that any mixture of the two sands displays a conductivity less than linearly interpolated value. The physical explanation lies in the fact that the smaller grains occupy a larger fraction of the space around larger grains than do uniform grains of either size.

Table 4 contains representative hydraulic conductivities for a variety of geologic materials. It should be noted that these are averages of many measurements; clearly, a range of values exists for each rock type depending on factors such as weathering , fracturing, solution channels, and depth of burrial. Magnitudes of hydraulic conductivity for various classes of unconsolidated rocks are shown in Table 5. Tables 6 and 7 give the ranges of hydraulic conductivity of

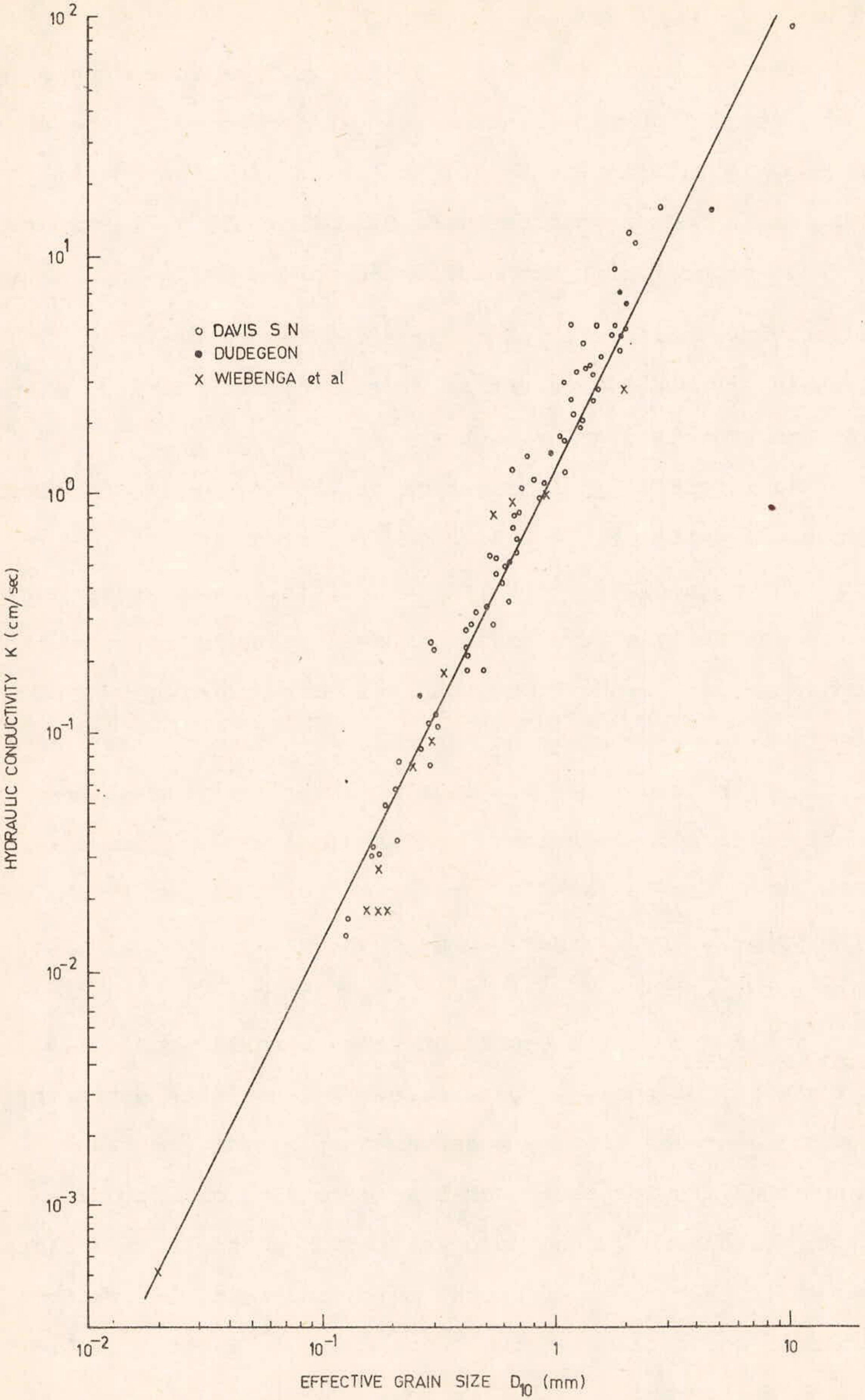

FIGURE 4 - HYDRAULIC CONDUCTIVITY AS A FUNCTION OF EFFECTIVE GRAIN SIZE

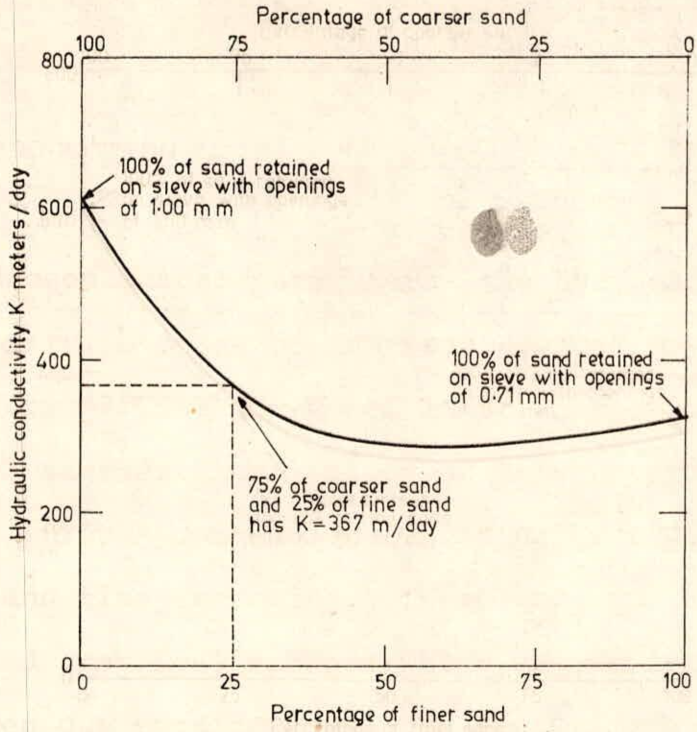

FIGURE 5 - HYDRAULIC CONDUCTIVITY OF VARIOUS PROPORTIONS OF TWO UNIFORM SANDS

unconsolidated/semi-consolidated and consolidated/fractured rocks respectively. These average values of the hydralic conductivity are based on the studies carried out by the Central Ground Water Board (CGWB).

The hydraulic conductivity values determined by laboratory method for selected sites in alluvium of U.P. are given in Table 8. The values are the averages of several samples.

Table 4 - Representative Values of Hydraulic Conductivity (after Morris and Johnson)

Material	Hydraulic conductivity m/day	Type of measurement
Gravel, coarse	150	R
Gravel, medium	270	R
Gravel, fine	450	R
Sand, coarse	45	R
Sand, medium	12	R
Sand ,fine	2.5	R
Silt	0.08	H
Clay	0.0002	H
Sandstone, fine-grained	0.2	V
Sandstone,medium-grained	3.1	V
Limestone	0.94	V
Dolomite	0.001	V
Dune sand	20	V
Loess	0.08	V
Peat	5.7	V
Schist	0.2	V
Slate	0.00008	V
Till,predominantly sand	0.49	R
Till,predominantly gravel	30	R
Tuff	0.2	V
Basalt	0.01	V
Gabbro,weathered	0.2	V

Table 4 (Continued)

Granite, weathered	1.4	V

where, H is horizontal hydraulic conductivity, R is a repacked sample, and V is verticle hydraulic conductivity.

Table 5 - Hydraulic Conductivities for Various Classes of Geologic Materials (after Bureau of Reclamation)

Hydraulic conductivity, metres/day

10^4	10^3	10^2	10^1	10^0	10^{-1}	10^{-2}	10^{-3}	10^{-4}	10^{-5}

Relative hydraulic conductivity

Very high	High	Moderate	Low	Very low

REPRESENTATIVE MATERIALS

Unconsolidated deposits

Clean gravel -	Clean sand and sand and gravel -	Fine sand -	Silt, clay, and mixtures of sand, silt and clay -	Massive clay

Consolidated Rocks

Vesicular and scoriaceous basalt and cavernous limestone and dolomite -	Clean sand stone and fractured igneous & metamorphic rocks -	Laminated sand stone, shale, mudstone -	Massive igneous and metamorphic rocks

Table 6 - Range of Hydraulic Conductivity of Unconsolidated to semi-consolidated Formations

Nature of Aquifer Material	Range of hydraulic Conductivity (m/day)
Gravel:	
Coarse	50 - 100
Medium	40 - 50
Fine	30 - 40
Sand:	
Gravel to very coarse	40 - 50
Very coarse	30 - 40
Very coarse to coarse	25 - 30
Coarse	20 - 25
Coarse to medium	10 - 20
Medium	5 - 10
Medium to fine	3 - 5
Fine	1 - 3
Loam	0.1-0.5
Clay:	
Clay	Less than or equal to 0.001

Table 7 - Range of Hydraulic Conductivity of Consolidated and Fractured Rocks (Based on studies carried out in hardrock area projects, by Central Groundwater Board. These are average values only)

Nature of Aquifer Material	Range of Hydraulic conductivity (m/day)
A. Granites/Gneisses etc	
1. Highly weathered granites with alluvium	25 to 20
2. Weathered and fractured	20 to 10
3. Partly weathered and fractured	10 to 5
4. Relatively fresh and fractured	5
B. Metamorphic Rocks (Schists Phyllites etc.)	
a. Highly weathered and fractured	Less than 1
b. Weathered and fractured	1 to 5
c. Relatively fresh and fractured	5 to 10
C. Basalt and Other Associated Formations:	
1. Voggey laterite	less than 5
2. Clayey laterite	less than 1
3. Weathered basalt	less than 1
4. Vesicular basalt	1 to 5
D. Limestones (Non-Cavernous)	more than 1
E. Sandstone (Shaly)	more than 1

Table 8 - Hydraulic Conductivity Values for Selected Sites in U.P.

District	Depth range below ground (m)	Average hydraulic conductivity (m/day)
1. Saharanpur	2 - 10	16.10
2. Meerut	0 - 23	32.43
3. Muzaffarnagar	-	18.16
4. Gaziabad	3 - 24	7.16
5. Bulandshahr	2 - 14	9.71
6. Aligarh	-	15.20
7. Bijnor	-	10.60

2.5 Intrinsic Permeability

Intrinsic permeability of a rock or a soil is a measure of its ability to transmit fluid, such as water, under a hydropotential gradient. Many earlier workers found that the intrinsic permeability is approximately proportional to the square of the mean grain diameter, i.e.

$$k = Cd^2 \qquad (L^2)$$

where,

k = intrinsic permeability,

C = a dimensionless constant depends upon porosity range and distribution of particle size, shape of grains, and other factors, and

d = the mean grain diameter or effective grain diameter.

The intrinsic permeability is a property of the medium alone and is independent of the nature or properties of the fluid, the U.S. Geological Survey is adopting the term 'intrinsic permeability', which is not to be confused with hydraulic conductivity as the latter includes the properties of natural ground water. Intrinsic permeability may be expressed as

$$k = -\frac{qv}{g(dh/dl)} \qquad (L^2)$$

where,

k = intrinsic permeability,

q = rate of flow per unit area (Q/A),

v = kinematic viscosity,

g = acceleration due to gravity, and

dh/dl = gradient, or change in head per unit length of flow.

The intrinsic permeability k and the hydraulic conductivity K are related to each other by equation

$$K = k\,\rho g/\mu$$

where,

ρ = fluid density,

μ = viscosity of the fluid, and

g = acceleration due to gravity.

Because of the analogy between flow of fluids through porous media and the flow of a viscous fluid through a thin tube of radius R, in which case the value of k is $R^2/8$, it seems reasonable to expect that the hydraulic conductivity K will be proportional to the second power of some characteristic pore size. This is confirmed by experimental evidence (Verruijt,1982). Both theoretical and experimental investigations have been carried out to establish a formula predicting the value of the permeability. The most familiar equation is that of Kozeny Carman

$$k = Cd^2 \frac{n^3}{(1-n)^2} \qquad \ldots(1)$$

where, n is the porosity of the soil and d is some mean particle size. A convenient definition for d can be made in terms of the specific surface M(the total area of the wetted surface per unit volume of the solid material), namely,

$$d = 6/M \qquad \ldots(2)$$

The factor 6 has been introduced so that for a packing of equal spheres the value of d corresponds to the diameter of the spheres. The value of the constant C in equation (1) corresponding to the definition in equation (2) for d, and best fitting the experimental data is of the order of magnitude of C approximately equal to 1/180. It should be realised that the equations given above can at best give a rough idea of the value of the permeability because factors such as the angularity of the particles are ignored. Therefore, the above equation give a first estimation of the permeability of a soil of which a grain size analysis, but no sample, is available. The circumstance that the permeability k is independent of the fluid, in contrast to the hydraulic conductivity, implies that for problems involving two fluids such as oil and water or fresh and salt water the Darcy's law should be formulated in terms of the permeability (Verruijt, 1982).

The numerical values of the permeability k and the hydraulic conductivity K (for fresh groundwater) are given below for certain soils (Verruijt, 1982).

Material	k(m^2)	K(m/sec)
Clay	10^{-17} to 10^{-15}	10^{-10} to 10^{-8}
Silt	10^{-15} to 10^{-13}	10^{-8} to 10^{-6}
Sand	10^{-12} to 10^{-10}	10^{-5} to 10^{-3}
Gravel	10^{-9} to 10^{-8}	10^{-2} to 10^{-1}

2.6 Methods of Determination of Hydraulic Conductivity and permeability

Hydraulic conductivity in saturated zones can be determined by a variety of techniques, including calculation from formulae, laboratory methods, tracer tests, auger hole tests, and pumping tests of wells.

2.6.1 Formulae

Numerous investigators have studied the relationship of permeability or hydraulic conductivity to the properties of porous media. Several formulae have resulted based on analytic or experimental work. Most permeability formulae have the general form

$$k = Cd^2$$

$$k = f_s f_\alpha d^2$$

where, C is a dimensionless coefficient, f_s is a grain (or pore) shape factor, f_α is a porosity factor, and d is characteristic grain diameter (Fair, et al, 1933; Krumbein, et al, 1943; and Masch, et al, 1966). Few formulae give reliable estimates of results because of the difficulty of including all possible variables in porous media. For an ideal medium, such as an assemblage of spheres of uniform diameter, hydraulic conductivity can be accurately evaluated from known porosity and packing conditions.

Because of the problems inherent in formulae, other techniques for determining hydraulic conductivity are preferable.

2.6.2 Laboratory methods

In the laboratory, hydraulic conductivity can be determined by a permeameter, in which flow is maintained through a small sample of material and measurements of flow rate and head loss are made (Wenzel, 1942). The constant head and falling head types of permeameters are widely employed and simple to operate.

Permeameter results may bear little relation to actual field hydraulic conductivities. Undisturbed samples of unconsolidated material are difficult to obtain, while disturbed samples experience changes in porosity, packing, and grain orientation, which modify hydraulic conductivities. Also one or even several samples from an aquifer may not represent the overall hydraulic conductivity of an aquifer. Variations of several orders of magnitude frequently occur for different depths and locations in an aquifer.

2.6.3 Field methods

i) Tracer tests: Field determination of hydraulic conductivity can be made by measuring the time interval for a tracer to travel between two observation wells or test holes (Cedergren,1977). For the tracer a dye, such as sodium fluorescein, or a salt, such as calcium chloride, is convenient, inexpensive, easy to detect, and safe. Figure 6 shows the cross-section of a portion of an unconfined aquifer with ground water flowing from hole A towards hole B. The tracer is injected as a slug in hole A after which samples of water are taken from hole B to determine the time passage of the tracer. Because the tracer flows through the aquifer with the average interstitial velocity V_a, then

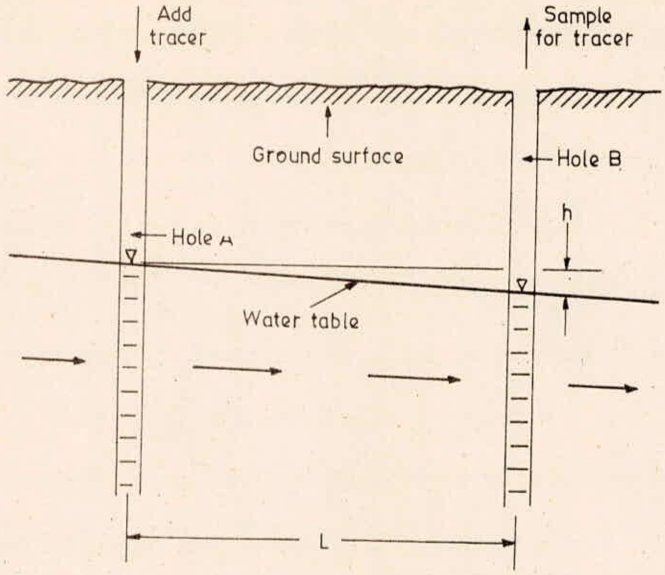

FIGURE 6 - CROSS-SECTION OF AN UNCONFINED AQUIFER ILLUSTRATING A TRACER TEST FOR DETERMINING HYDRAULIC CONDUCTIVITY

$$V_a = \frac{K}{n} \left(\frac{h}{L} \right)$$

where, K is hydraulic conductivity, n is porosity, h and L are shown in Figure 6, V_a is also given by

$$V_a = L/t$$

where, t is the travel time interval of the tracer between the holes. Equating these and solving for K yields,

$$K = \frac{nL^2}{ht}$$

Although this procedure is simple in principle, results are only approximate because of serious limitations in the field such as:

a. The holes need to be close together; otherwise, the travel time interval can be excessively long.

b. Unless the flow direction is accurately known, the tracer may miss the downstream hole entirely. Multiple sampling holes can help, but these add to the cost and complexity of conducting the test.

c. If the aquifer is stratified with layers having differing hydraulic conductivities, the first arrival of the tracer will result in a conductivity considerably larger than the average for the aquifer.

ii) Point dilution method: An alternate tracer technique, which has been successfully applied under field conditions, is the point dilution method (Drost, et al 1968; Halevy, et al, 1967 and Intl. Assoc ....1972). Here a tracer is introduced into an observation well and thoroughly mixed with the contained water. Thereafter, as water flows into and from the well, repeated measurements of tracer concentration are made. Analysis of the resulting dilution curve defines the groundwater velocity. By measuring the water table gradient and applying Darcy's law, a localized estimate of the hydraulic conductivity can be made and the direction of groundwater flow can be known (Rumer, et al, 1966).

iii) Auger hole tests: The auger hole method involves the measurement of the change in water level subsequent to a rapid removal of a volume of water from an unlined cylindrical hole. If the soil is loose, a screen may be necessary to maintain the hole. The method is relatively simple and is most adaptable to shallow water table conditions. The value of K obtained is essentially that for a horizontal direction in the immediate vicinity of the hole. Figure 7 illustrates an auger hole and the dimensions required for calculation.

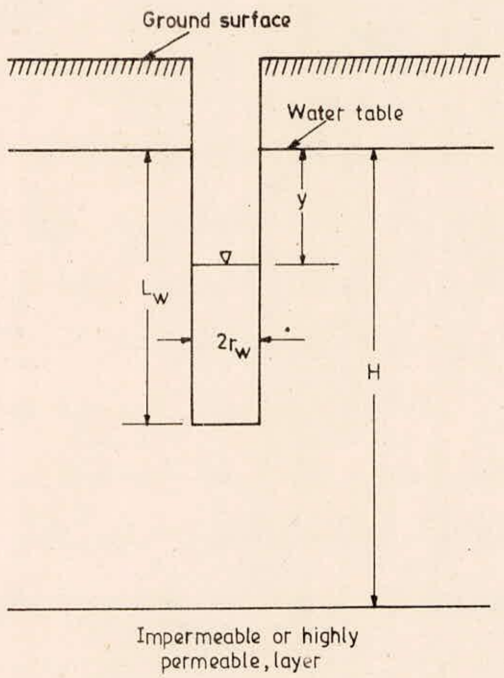

FIGURE 7 - DIAGRAM OF AN AUGER HOLE AND DIMENSIONS FOR DETERMINING HYDRAULIC CONDUCTIVITY

It can be shown (Boast, et al, 1971) that hydraulic conductivity is given by

$$K = \frac{C}{864} \left(\frac{dy}{dt}\right)$$

where, dy/dt is the measured rate of rise in cm/sec and the factor 864 yields K values in m/day. The factor C is a dimensionless constant listed in table 9 and governed by the variables shown in figure 7. Note that the tabulated values cover the following conditions below the hole: a shallow impermeable layer, an infinite homogeneous stratum and a shallow highly permeable (gravel) layer.

Several other techniques similar to the auger hole test have been developed in which water level changes are measured after an essentially instantaneous removal or addition of a volume of water. With a small-diameter pipe driven into the ground, K can be found by the piezometer, or tube method (Van Schilfgaarde, 1970). For wells in confined aquifers, the slug method can be employed (Cooper, et al, 1967; Lohman, 1972). Here a known volume of water is suddenly injected or removed from a well after which the decline or recovery of the water level is measured in the ensuing minutes. Where a pump is not available to conduct a pumping test on a well, the slug method serves as an alternate approach.

iv) Hooghoudt method

Assumptions: a) water table near the well does not fall

b) flow through the well is laminar/ horizontal, and

c) flow through the bottom of the well is vertical.

Table 9-Values of the Factor C for the Auger Hole Test to Determine Hydraulic Conductivity
(After Boast and Kirkham)

		$(H-L_w)/L_w$ for Impermeable Layer								$H-L_w$	$(H-L_w)/L_w$ for Infinitely Permeable Layer			
L_w/r_w	y/L_w	0	0.05	0.1	0.2	0.5	1	2	5	∞	5	2	1	0.5
1	1	447	423	404	375	323	286	264	255	254	252	241	213	166
	0.75	469	450	434	408	360	324	303	292	291	289	278	248	198
	0.5	555	537	522	497	449	411	386	380	379	377	359	324	264
2	1	186	176	167	154	134	123	118	116	115	115	113	106	91
	0.75	196	187	180	168	149	138	133	131	131	130	128	121	106
	0.5	234	225	218	207	188	175	169	167	167	166	164	156	139
5	1	51.9	48.6	46.2	42.8	38.7	36.9	36.1		35.8		35.5	34.6	32.4
	0.75	54.8	52.0	49.9	46.8	42.8	41.0	40.2		40.0		39.6	38.6	36.3
	0.5	66.1	63.4	61.3	58.1	53.9	51.9	51.0		50.7		50.3	49.2	46.6
10	1	18.1	16.9	16.1	15.1	14.1	13.6	13.4		13.4		13.3	13.1	12.6
	0.75	19.1	18.1	17.4	16.5	15.5	15.0	14.8		14.8		14.7	14.5	14.0
	0.5	23.3	22.3	21.5	20.6	19.5	19.0	18.8		18.7		18.6	18.4	17.8

Table 9 (Continued)

20	1	5.91	5.53	5.30	5.06	4.81	4.70	4.66	4.64	4.62	4.58	4.46
	0.75	6.27	5.94	5.73	5.50	5.25	5.15	5.10	5.08	5.07	5.02	4.89
	0.5	7.67	7.34	7.12	6.88	6.60	6.48	6.43	6.41	6.39	6.34	6.19
50	1	1.25	1.18	1.14	1.11	1.07	1.05		1.04		1.03	1.02
	0.75	1.33	1.27	1.23	1.20	1.16	1.14		1.13		1.12	1.11
	0.5	1.64	1.57	1.54	1.50	1.46	1.44		1.43		1.42	1.39
100	1	0.37	0.35	0.34	0.34	0.33	0.32		0.32		0.32	0.31
	0.75	0.40	0.38	0.37	0.36	0.35	0.35		0.35		0.34	0.34
	0.5	0.49	0.47	0.46	0.45	0.44	0.44		0.44		0.43	0.43

Procedure: a) The water level in the well/borehole is allowed to be stabilised for sometime.

b) The entire storage in the well/hole is pumped out quickly (care should be taken to see that no drawdown is created, only storage has to be bailed out.

c) The pumping is then stopped and the recovery measurements are taken at close intervals.

The hydraulic conductivity is given by

$$K = \frac{2.3\ r_w C}{(2b + r_w)t'} \log \frac{s}{s'}$$

$$C = \frac{r_w b}{0.19}$$

where,

r_w = radius of the well in metres

b = saturated thickness of unconfined aquifer in metres

t' = time of recovery in days

s = total drawdown in metres

s' = residual drawdown in metres.

In case where the borehole is drilled down to the impervious layers (i.e. there is no flow from the bottom of the well) the flow of water is horizontal through the wells only. The K is given by

$$K = \frac{2.3\ r_w C}{2t'b} \log \frac{s}{s'}$$

Figure 8 shows an example of estimation of hydraulic conductivity using Hooghoudt formula, in Vedavati river basin, India.

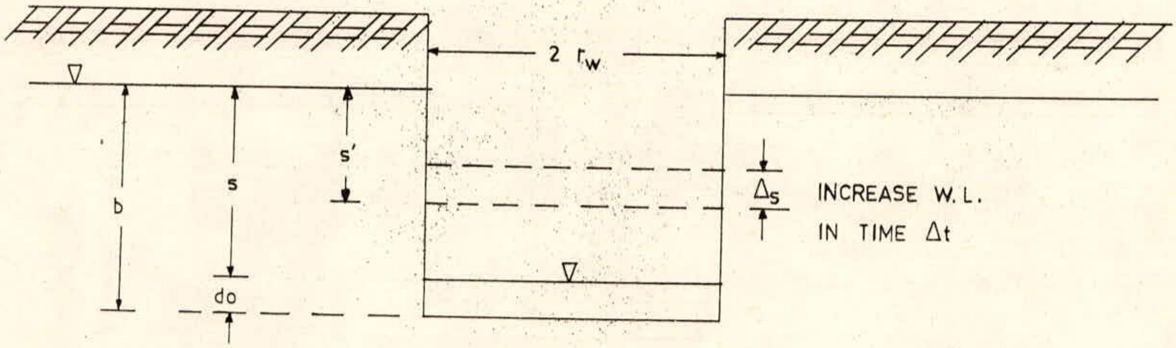

FIGURE 8 - DEFINITION SKETCH FOR PERMEABILITY TEST BY HOOGHOUDT FORMULA

EXAMPLE: Haraganadona water table well
edavati Project, C.G.W.B.

r_w = 0.0756 metres

b = 9.44 metres

s = 6.477 metres

s' = 5.272 metres

t = 100 metres

$$K = \frac{2.3\ r_w C}{2t'b} \log \frac{s}{s'} ;\ \text{where } C = \frac{r_w b}{0.19}$$

$$= \frac{2.3 \times 0.0756 \times 0.0756 \times 9.44 \times 1440}{2 \times 100 \times 9.44 \times 0.19} \log \frac{6.477}{5.272}$$

$$= 0.045 \text{ m/day.}$$

v) Pumping Tests of Wells: The most reliable method for estimating aquifer hydraulic conductivity is by pumping tests of wells. Based on observations of water levels near pumping wells, an integrated K value over a sizable aquifer section can be obtained. Because the aquifer is not disturbed, the reliability of such determinations is superior to laboratory methods. The pump test methods and computations are described in detail by Todd (1980), Kruseman and De Ridder (1976).

2.7 Unsaturated Hydraulic Conductivity

Unsaturated flow in the zone of aeration can be analyzed by Darcy's law, however, the hydraulic conductivity K_u for unsaturated soil is a function of the water content or the negative pressure head. Because part of the pore space is filled with air, the available cross-section area available for water flow is reduced; consequently, K_u is always less than the conductivity of saturated soil.

Although there are hysteresis effects present in the relation of K_u with water content and negative pressure, approximations based on empirical evidence can be stated. Water content data fit the form (Irmay, 1954),

$$\frac{K_u}{K} = \left(\frac{S_s - S_o}{1 - S_o} \right)^3$$

where, S_s is the degree of saturation and S_o is the threshold saturation, the saturation corresponding to that part of the voids filled with non-moving water held primarily by capillary forces. The above equation is plotted in Figure 9, note that K_u ranges from zero at $S_s = S_o$ to K at $S_s = 1$, which is saturation.

For hydraulic conductivity and negative pressure, S-shaped relations as indicated in Figure 10 are generally applicable (Vachaud, et al, 1971; Van, 1970). These can be approximated by a step function or by,

$$\frac{K_u}{K} = \frac{a'}{\frac{a'}{b'} (-h)^{n_o} + a'}$$

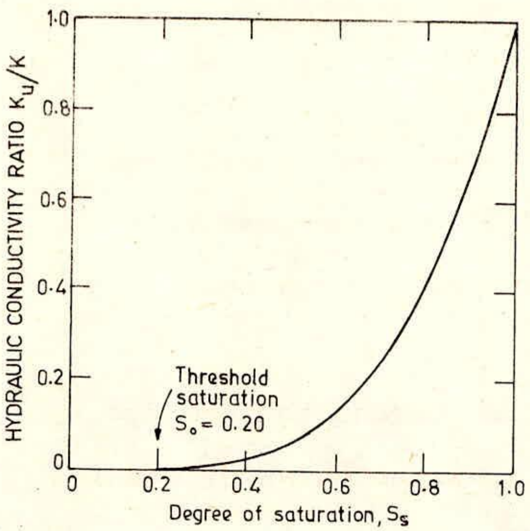

FIGURE 9 - RATIO OF UNSATURATED TO SATURATED HYDRAULIC CONDUCTIVITY AS A FUNCTION OF SATURATION

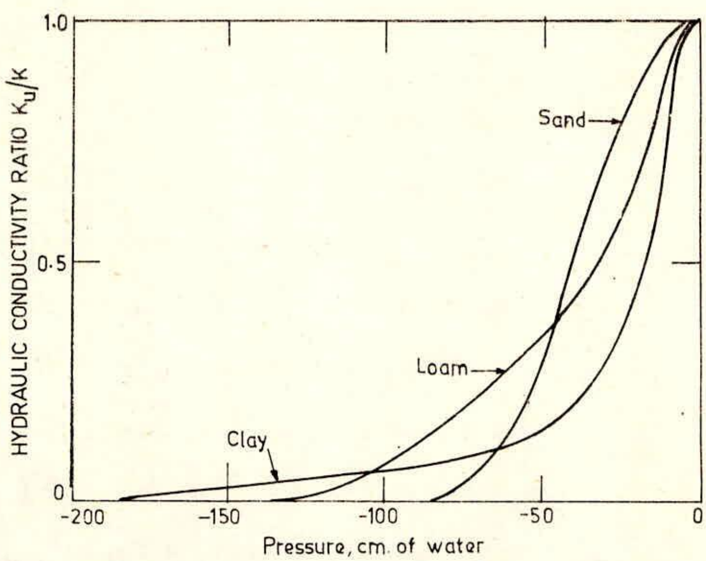

FIGURE 10 - MEDIAN RELATIONSHIP BETWEEN HYDRAULIC CONDUCTIVITY AND SOIL WATER PRESSURE (TENSION)

where, a',b' and n_o are constants that vary with particle sizes of unconsolidated material and h is the pressure head measured in centimetres. It can be seen that when h=0 which occurs at atmospheric pressure, $K_u = K$. Orders of magnitude of the constants in the above equation for different soils are as follows(Bouwer, 1964):

Material	a'	b'	n_o
Medium Sands	5×10^9	10^7	5
Fine sands, sandy loams	5×10^6	10^5	3
Loams and clays	5×10^3	5×10^3	2

2.8 Transmissivity

The aquifer parameters like transmissivity (T) and (S) are usually determined by aquifer test that is by observing the performance of aquifer in response to a long period of pumping at a given rate.

In many cases, especially during reconnaissance type of ground water investigations and for water balance studies it may not be practical or feasible to constrauct test wells and conduct the time consuming aquifer tests for estimation of hydrogeological parameters. Also, some of the modern quantitative techniques such as those for which electric analog models or mathematical models are contemplated, a sufficiently large number of T and S values are required.

In all such cases, quick and approximate methods may have to be resorted to, for the determination of hydrogeological parameters. These properties can be estimated with reasonable accuracy by some of the indirect methods based on analysis of water level fluctuations, specific capacity data of wells, and well logs etc.

2.8.1 General relationship between transmissivity and specific capacity

In many ground water investigations, especially those of reconnaissance type the specific capacities of wells provide the only basis for estimating the transmissivity of the aquifer. Generally speaking, high specific capacities indicate an aquifer having a high transmissivity. However, a precise correlation between the specific capacities of wells and the T values of the aquifers has not yet been established.

The specific capacity of a well cannot exactly determine the transmissivity of the aquifer in the vicinty of the well because, the yield of the well per metre of drawdown is also a function of other factors such as the diameter of the well, the depth to which the well extends into the aquifer, the type and amount of perforation in the well casing, and the extent to which the well has been developed. However, estimates of T that are based on the specific capacities of wells are reasonably reliable and could be made without the elaborate tests necessary for precise determinations. Therefore a formula expressing the theoretically exact relationship between the specific capacity of a well and the transmissivity of the aquifer which the well taps would be highly useful in the making of reconnaissance ground water studies provided the theoretical formula could be empirically modified for prevailing field conditions.

2.8.2 Methods of determination of transmissivity

2.8.2.1 Specific capacity method

i) Walton's method: The theoretical specific capacity of a well discharging at a constant rate in a homogenous, isotropic, nonleaky artesian aquifer of infinite areal extent is given by the following expression (Walton, 1962),

$$\frac{Q}{s_w} = \frac{4 \pi T}{2.30 \log_{10} (2.25 \ Tt/r_w^2 \ S)}$$

where,

- s_w = drawdown in a 100 percent efficient pumped well in metres,
- r_w = radius of the pumped well in metres,
- Q/s_w = specific capacity in m^3/day per metre of drawdown,
- Q = rate of discharge in m^3/day,
- T = transmissivity in m^2/day,
- S = dimensionless storage coefficient, and
- t = time after pumping started in days.

The above equation assumes that : 1) the production well has full penetration and the well is uncased in the entire depth of aquifer, 2) the well loss is negligible, and 3) the effective radius of the production well has not been affected during drilling and development of the production well and is equal to the nominal radius of the production well. The storativity can be estimated either from well log data or from study of water level data. As the specific capacity varies with the logarithm of 1/S, large error in assumed storage value results in comparatively small error in transmissivity estimated using the above relation.

The relationship between the specific capacity and transmissivity for artesian and water table conditions are shown in Figure 11 through 16 (Walton, 1962). These graphs can be used to obtain rough estimates of the transmissivity from specific capacity data provided approximate value of storage coefficient is known.

As seen from the specific capacity equation the value Q/s_w varies with the logarithm of $1/r_w^2$. Therefore, large increase in the radius of a well result in comparatively small changes in the specific capacity values (Figure 17). Specific capacity decreases with the period of pumping as shwon in Figure 18, because the drawdown continuously increases with time as the cone of influence of the well expands till the steady state conditions are arrived at. For this reason it is important to state the duration of the pumping period at which a specific capacity is computed.

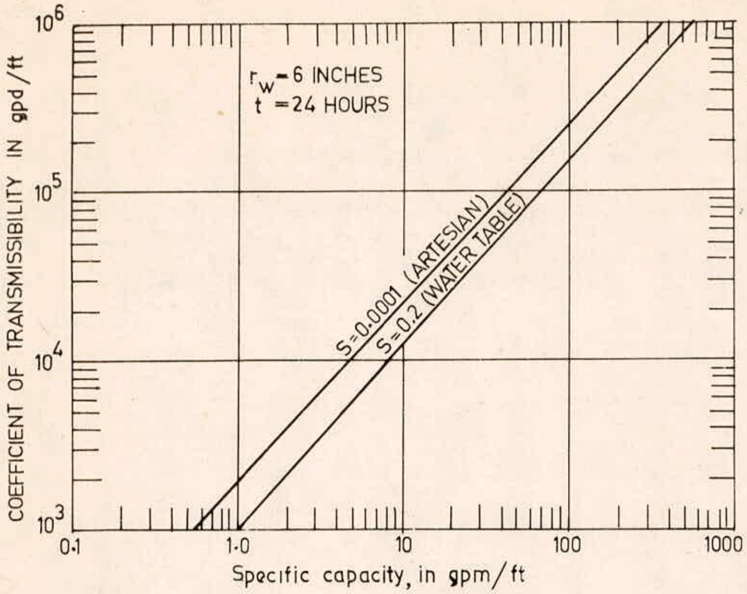

FIGURE 11

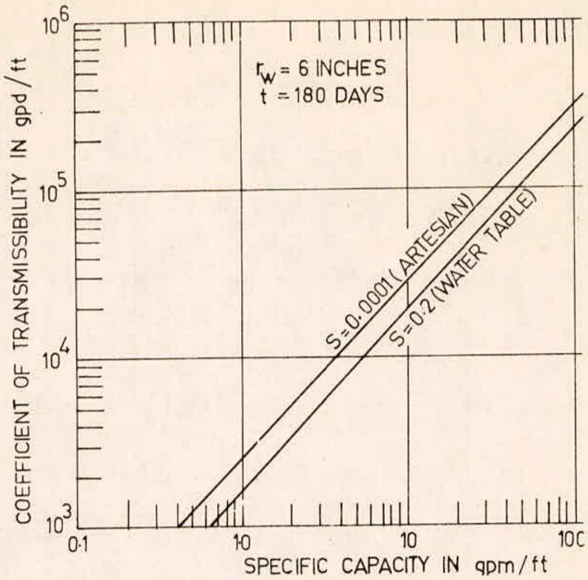

FIGURE 12

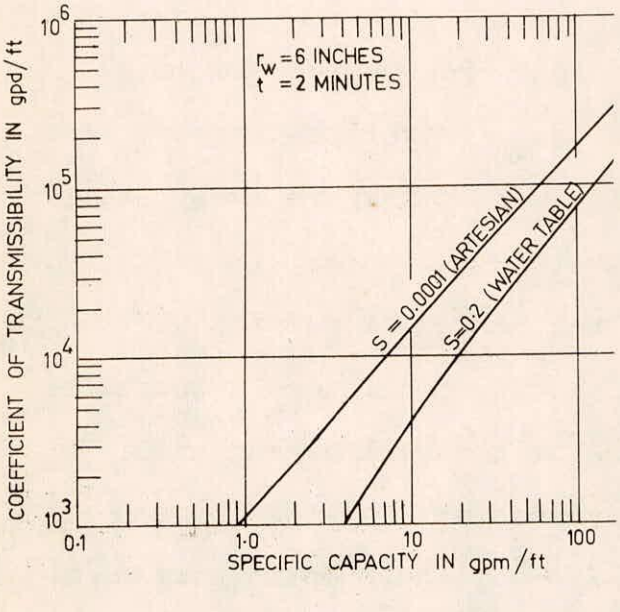

FIGURE 13

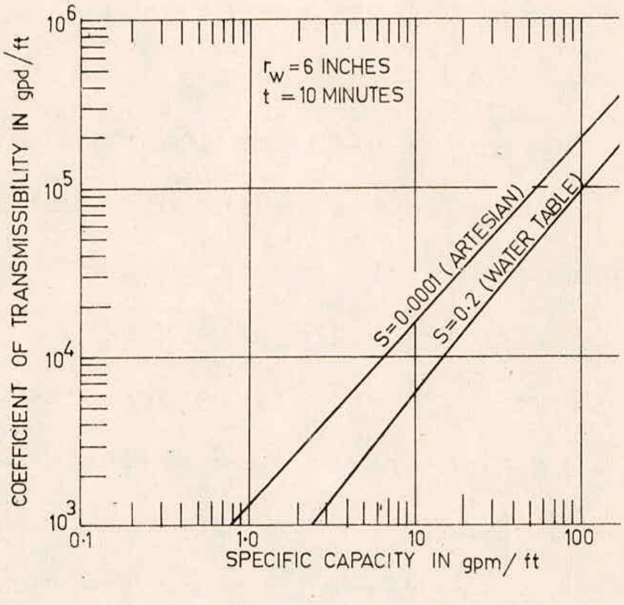

FIGURE 14

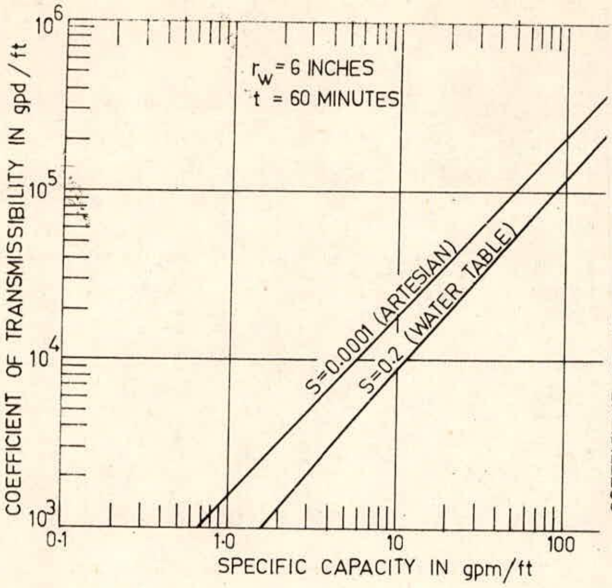

FIGURE 15

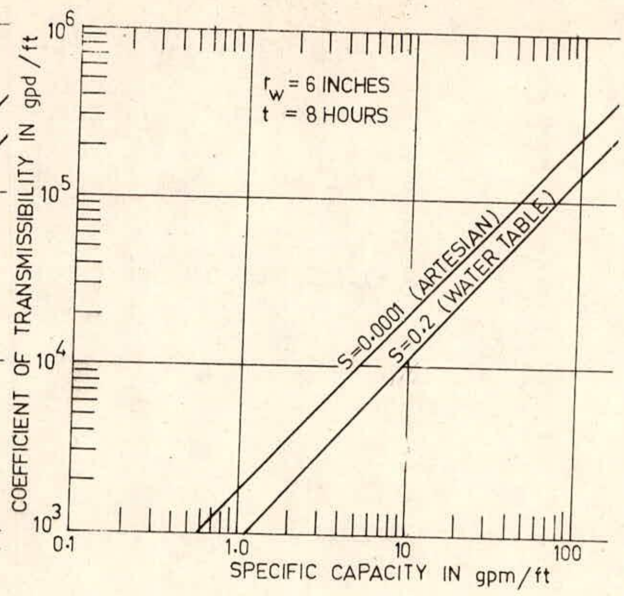

FIGURE 16

FIGURE 11 THROUGH 16 - GRAPHS OF SPECIFIC CAPACITY VS. COEFFICIENT OF TRANSMISSIBILITY FOR DIFFERENT PUMPING PERIODS (FROM WALTON, 1962)

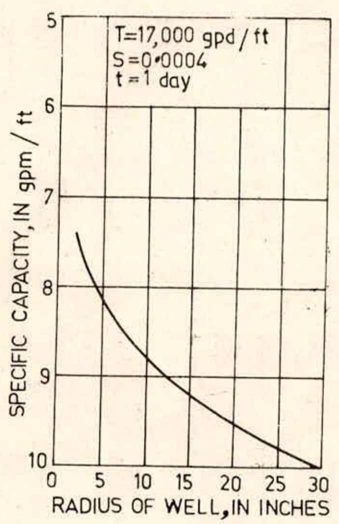

FIGURE 17- GRAPH OF SPECIFIC CAPACITY VS. WELL RADIUS

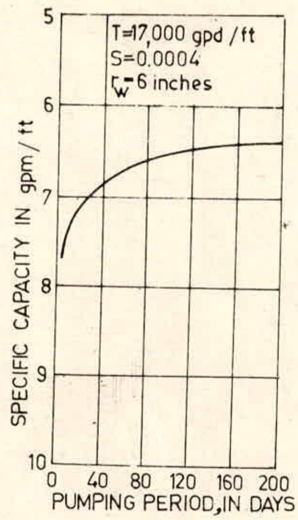

FIGURE 18 - GRAPH OF SPECIFIC CAPACITY VS. PUMPING PERIOD

ii) Theis method: Theis (1963) has proposed the following method for estimating transmissivity of a water-table aquifer using specific capacity data. The Theis equation for $\frac{r^2 S}{4Tt} < 0.01$ can be written with negligible error as follows:

$$T = \frac{114.6\ Q}{s} \left(-0.5772 - \log_e \left(\frac{1.87 r^2 S}{Tt} \right) \right)$$

Considering the fact that a fairely productive water table aquifer will have S and T values near to 0.2 and 100,000 gpd/ft (1242 m^2/day) respectively the above equation has been rewritten as

$$T = \frac{114.6\ Q}{s} \left(-0.5772 - \log_e \left(\frac{1.87 r^2 (0.2)}{100,000} \frac{100,000}{0.2} \frac{S}{T} \frac{1}{t} \right) \right)$$

$$= \frac{-66\ Q}{s} + \frac{264\ Q}{s} \left(-\log_{10}(3.74 r^2 . 10^{-6}) - \log_{10} 5S + \right.$$

$$\left. \log_{10} (T \times 10^{-5}) + \log_{10} t \right)$$

or

$$T - \frac{264\ Q}{s} \log_{10} (T \times 10^{-5}) = \frac{Q}{s} \left(-66 - 264 \log_{10} (3.74 r^2 . 10^{-6}) \right.$$

$$\left. -264 \log_{10} 5S + 264 \log_{10} t \right)$$

The terms $T - \frac{264\ Q}{s} \log_{10} (T \times 10^{-5})$ have been designated as T' and the terms $-66 - 264 \log_{10}(3.74 r^2 . 10^{-6})$ have been designated as K'. The equation finally takes the form

$$T' = \frac{Q}{s} \; (K' - 264 \log_{10} 5S + 264 \log_{10} t).$$

This formula indicates the importance of both the storage coefficient and the duration of pumping when the coefficient of transmissivity is estimated from a single measurement of drawdown. The variation of T' with $\frac{Q}{s}$ for different values of T are shown in Figure 19. The values of K' computed for selected values of r are given below:

r (ft)	r (m)	K
0.25	0.0762	1,684
0.50	0.1524	1,524
1.00	0.3048	1,367
5.00	1.524	996
10.00	3.048	838
20.00	6.096	680
30.00	9.144	588
40.00	12.192	521
50.00	15.24	469

If, S=0.2, the influence of S term in the final expression of T' is zero. However, if S = 0.1, the value of 264 $\log_{10} 5S$ = 80 which is about 8 percent of the constant K', for r = 5 ft. If S = 0.3, the S term is about -4.5 percent of K' for r = 5 ft. Thus if S is unknown the error in T' for a water - table aquifer for which S ranges from 0.1 to 0.3 probably will be smaller than the errors inherent in the method. Although the correction for the duration of pumping also is comparatively small, Theis has suggested that the duration of pumping should be taken into account. For an artesian aquifer, S is very small and the S term correction will be large and the formula should not be used. For known values of Q/s and t,T' can be estimated neglecting the correction for S term. Once T' is known the corresponding T can be estimated from Figure 19. Thus, within the limits of the idealized assumption, the coefficient of transmissivity of a water-table aquifer apparently can be computed without great error from a single measurement of drawdown in an observation well which is at a short distance from apumped well even if the coefficient of storage is not known.

iii) Brown's method: Brown (1963 has modified Theis equation and has suggested a similar method as that of Theis for estimating the transmissivity of an artesian aquifer from the specific capacity data. A formula and set of constants for artesian conditions have been derived using an assumed storativity of 2×10^{-4} and transmissivity of 1242 m^2/day (100,000 gpd/ft) in a manner similar to that of Theis' derivation. For the assumed values of T and S the expression for T' has been obtained as:

$$T' = \frac{Q}{s} (K' - 264 \log_{10} (5S \times 10^{-3}) + 264 \log_{10} t)$$

where,

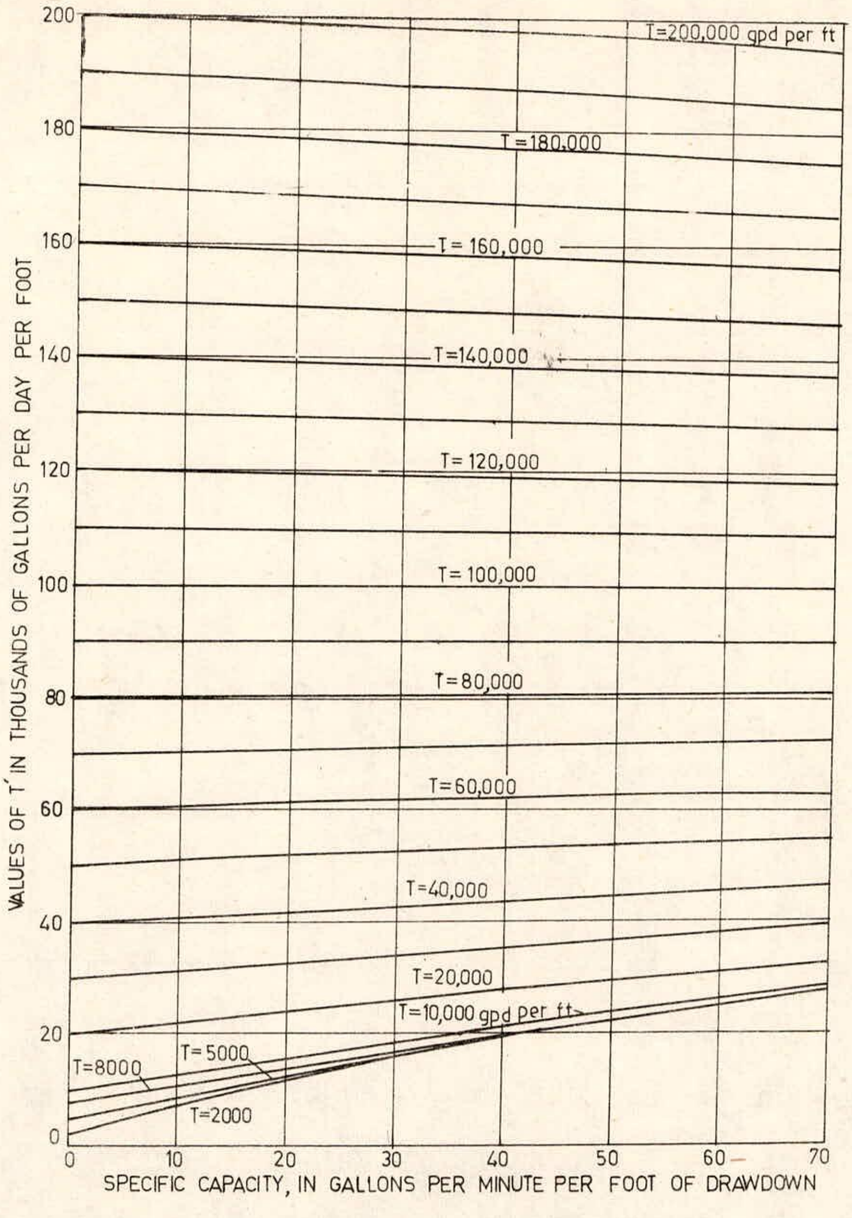

FIGURE 19 - DIAGRAM FOR ESTIMATING THE TRANSMISSIVITY OF AN AQUIFER FROM THE SPECIFIC CAPACITY OF A WELL

$$K' = -66 - 264 \log_{10} (3.73r^2 \times 10^{-9}).$$

Values of K', computed for selected values of r, are given below:

r (ft)	r (m)	K'
0.25	0.0762	2,477
0.50	0.1524	2,318
1.00	0.3048	2,159
5.00	1.524	1,790
10	3.048	1,633
20	6.096	1,472
30	9.144	1,379
40	12.192	1,313
50	15.24	1,262

If the value of S is as large as 2×10^{-3} (10 times the assumed value) the effect will be to decrease K' for r = 5 ft. by nearly 15 percent. Conversely, if S is as low as 2×10^{-5} (one tenth of the assumed value) the effect will be to increase K' by nearly 15 percent. Once T' is known by neglecting the contribution of S term, the transmissivity value can be known from Figure 19.

iv) Meyer's Method : Meyer (1963) gave the relationships between specific capacity at the end of 1 day, the transmissvity and storage coefficientfor various well diameter which are shown graphically (Figure 20). This plot can be used to determine the approximate transmissivity of an aquifer if the specific capacities of wells at the end of one day are the only available data. The figure can also be used to determine the approximate specific capacity of a well which is to be drilled into an aquifer for which T and S are known.

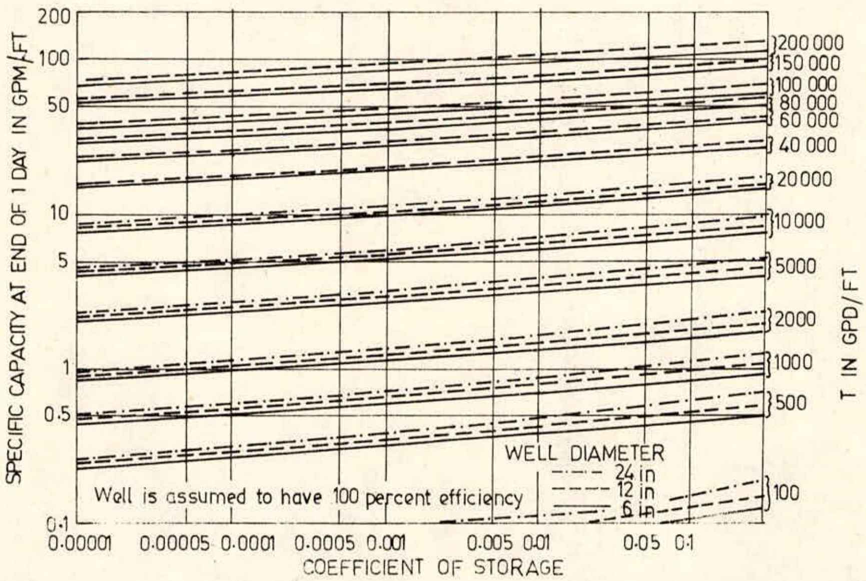

FIGURE 20 - GRAPH SHOWING RELATION OF WELL DIAMETER, SPECIFIC CAPACITY COEFFICIENT OF TRANSMISSIBILITY AND STORAGE

Although determinations made from Figure 20 may not be exact the graph serves as a measure for approximation. The limitation in the method is that it assumes the well to be 100 percent efficient which is practically not feasible.

v) Logan's method: According to Logan (1964) the Theim equation for a confined aquifer can be written as,

$$T = \frac{Q}{2\pi(s_{m1} - s_{m2})} \ln\left(\frac{r_2}{r_1}\right)$$

or

$$T = \frac{2.30\, Q \log(r_{max}/r_w)}{2\pi s_m}$$

where,

s_{m1} and s_{m2} are the drawdown at distances r_1 and r_2 respectively,

r_w is the radius of the pumped well (m),

$r_{max.}$ is the radius of influence (m), and

s_m is the maximum drawdown in the pumped well (m).

The accuracy in estimating transmissivity with this equation depends on the accuracy of measurements of s_m (which includes well losses) and on the accuracy of the ratio of r_{max}/r_w. The ratio can't be determined without the use of observation wells. However, although the variations in r_{max} and r_w may be substantial, the variation in the logarithm of their ratio is much smaller. Therefore, assuming average conditions of radii, a value of 3.33 for the log ratio may be taken as a rough approximation.

Substituting this value of 3.33 in the above equation for T, the equation becomes

$$T = \frac{1.22\, Q}{s_m}$$

This equation may also be applied in unconfined aquifer, replacing s_m by s'_m where

$$s'_m = (s_m - s_m^2/2b),$$

b is the saturated thickness of the aquifer.

vi) Hurr's method: Hurr's (1966) method is based on Theis non-equilibrium equation and can be used for calculation of T from a single drawdown observation provided the storage coefficient value could be assumed with reasonable accuracy. The Theis equation is written as

$$W(u) = \frac{4\pi Ts}{Q}$$

Hurr demonstrated that the multiplication of both sides of the above equation by u where $u = r^2S/4Tt$, results in the disappearance of T from the right hand term i.e.

$$uW(u) = \frac{4\pi Ts}{Q} \times \frac{r^2S}{4Tt}$$

$$= \frac{\pi r^2 s}{t} \times \frac{S}{Q}$$

A graph relating the value of u and uW(u) is given in Figure 21. The procedure to compute T includes the following:

i) Calculate the value of uW(u) for an assumed value of S and measured values of r,t,s and Q.

ii) Obtain from Figure 21 the corresponding value of u.

iii) Substitute the values of u,r,t and S into the equation

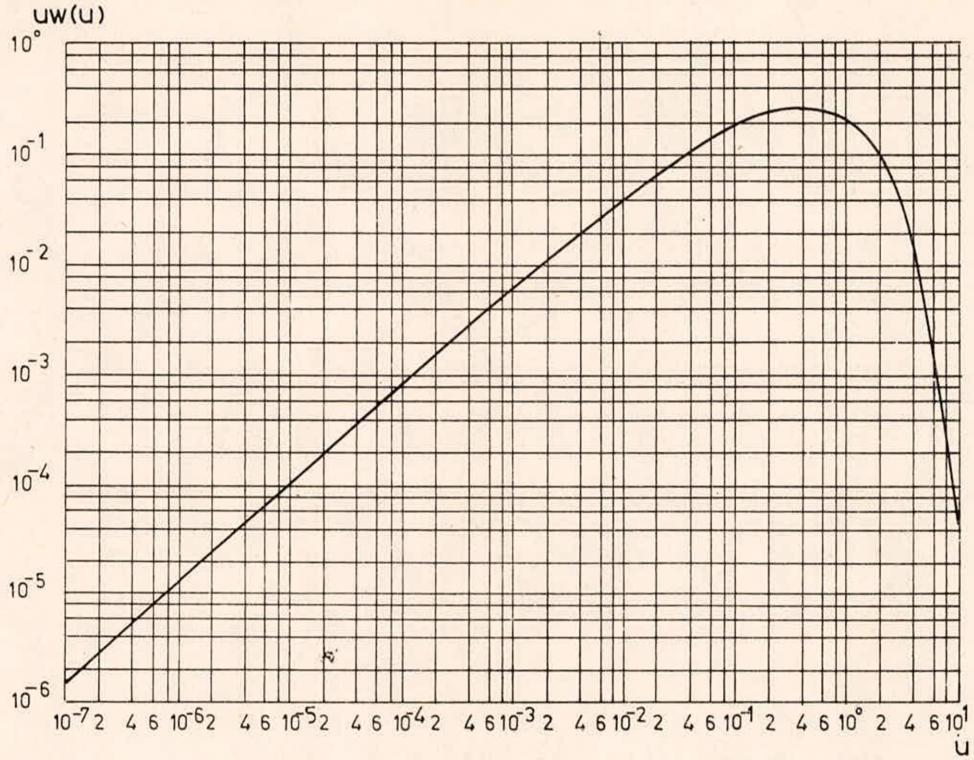

FIGURE 21 - TYPE CURVE FOR CORRESPONDING VALUES OF u and uW(u)

$u = r^2S/4Tt$ and calculate T.

vii) Closed contour method: A water-level contour map containing closed contours around a well or group of wells of known discharge rate may be used to determine the T of an aquifer under steady flow conditions. The Darcy's equation can be written as

$$Q = - KA \frac{\Delta h}{\Delta r}$$

$$= - TL \frac{\Delta h}{\Delta r}$$

where,

q = rate of discharge,

K = hydraulic conductivity,

A = cross-sectional area,

L = length of the closed contour,

Δh = contour interval, and

Δr = average distance between two closed contours.

For any two concentric closed contours of length L_1 and L_2 the above equation can be written as

$$T = \frac{2Q}{(L_1 + L_2)\frac{\Delta h}{\Delta r}} \qquad (L^2T^{-1})$$

Remark: The irregularity of the shape and spacing of the contours, the density and accuracy of the water level data, and the accuracy to which Q is known, control the accuracy of the T value.

viii) Bailer Method: Skibitzke (1958) proposed a method for determining the transmissivity from the recovery of water level in a well that has been bailed. The following equation is applicable at any given point on the recovery curve.

$$T = \frac{V}{4\pi s' t e^{(r_w^2 S/4Tt)}}$$

where,

s' = residual drawdown in metres,

V = volume of water removed in one bailing cycle in m^3,

t = length of time since bailing stopped in days, and

r_w = effective radius of the well in metres.

For small values of r_w and large values of t the term in brackets in the above equation approaches zero. Therefore for large values of t the equation reduces to

$$T = \frac{V}{4\pi s' t}$$

ix) Well log: Based on the observations of well log (drill cuttings) sample descriptions, it is possible to assign the value of coefficient of permeability (K) to each bed of known thickness (b).

The value of T is given by

$$T = K_1 b_1 + K_2 b_2 + K_3 b_3 + \ldots\ldots\ldots\ldots\ldots + K_n b_n.$$

3.0 REMARKS

A review of the important methods for evaluating the hydrogeological parameters commonly used in water balance and groundwater model studies has been carried out. It is seen that the commonly available field data like grain size distribution, well logs, thickness of the aquifer and specific capacity of the wells can be effectively used for estimating hydrogeological parameters such as transmissivity, coefficient of permeability and specific yield with reasonable accuracy. As the aquifer test and other field experiments are costly and time consuming it is preferred to use the existing empirical formulae and graphical methods to estimate the hydrogeological parameters for reconnaissance groundwater studies. The transmissivity can be estimated with reasonable accuracy by Theis and Brown's methods if the specific capacity of a well at the end of one day - pumping is known. The error involved in these methods which arises due to the assumption of storage coefficient value lies between $\pm$ 15 percent.

The Theis and Brown's methods should be developed for the duration of pumping commonly practised in an agricultural land. Formula similar to that of Theis and Brown are not available for wells having storage and are needs to be developed. Specific capacity contours for an area can be prepared to ascertain the potential zones of groundwater development.

REFERENCES

1. Boast, C.W. and D. Kirkham (1971), 'Auger hole seepage Theory' Soil Sci.Soc.of Amer,Proc., V.35,pp 365-373.

2. Brown,R.H. (1963), 'Estimating the transmissibility of an artesian aquifer from the specific capacity of a well', in Bentall Ray,compiler,Methods of determining permeability,transmibility and drawdown; US Geol.Survey.Water-Supply paper 1536-1, pp.336-338.

2. Bouwer, H.(1964), 'Measuring horizontal and vertical hydraulic conductivity of soil jwith the double-table method'. Proc. Soil Science Soc. of Amer, Vol.28,pp.19-23.

3. CGWB,(1982),'Evaluation of aquifer parameters',Manual,Central Ground Water Board,Min. of Irrigation,Govt of India.

4. Cedergren, H.R.(1977), 'Seepage , drainage, and flownets' 2nd edition.John Wiley and Sons N.Y. 534 pp.

5. Cooper, H.H.,Jr.,Bredehoaft,J.D.,and Papadopulos, I.S. (1967), 'Response of a finite - diameter well to an instantaneous charge of water', Water Res., Research, Vol.3, No.1.pp.263-269.

6. Drost, W., et al.,(1968),'Point dilution methods of investigating ground water flow by means of radioisotopes', Water Res Research, V.4.pp.125-146.

7. Fair, G.M., and L.P. Hatch, (1933), 'Fundamental factors governing the stream-line flow of water through sand', Jr. Amer.Water Works,Associ., V.25, pp. 1551-1565.

8. Gheorghe,A.(1978), 'Processing and synthesis of hydrogeological data'. ABACUS Press, pp.390.

9. Halevy, E., et al.,(1967),'Borehole dilution techniques: A critical review',Isotopes in hydrology.Intl.Atomic Energy Agency, Vienna,pp.531-564.

10. International Association of Hydraulic Research, (1972), 'Fundamentals of transport phenomena in porous media',Elsevier, Amsterdam,pp.392.

11. Irmay,S.(1954),'On the hydraulic conductivity of unsaturated soils'.Trans.Amer.Geophysical Union,V.35,pp 463-467.

12. Johnson,(1972),'USGS redefines some ground water terms The Johnson drillers Journ al, Sept-Dec.,pp.4-7.

13. Kruseman, G.P. and N.A. DeRidder, (1976),'Analysis and evaluation of pumping test data', Bull.11,Intl.Inst.for Land Reclamation and Improvement, Wageningen, The Netherlands.

14. Krumbein, W.C. and G.D.Monk, (1943),'Permeability as a function of the size parameters of unconsolidated sands', Tran. Amer.Inst.Min.and Met.Engrs.,V.151,pp.153-163.

15. Lohman,S.W.(1972),'Groundwater Hydraulics', USGS Prof.paper 704,pp.70.

16. Logan, J.(1964),'Estimating transmissibility from routin production tests of waterwells' Groundwater, V.2.,No.1,pp.35-37

17. LeGrand,H.E.,and V.T. Stringfield,(1971),'Development and distributiuon of permeability in carbonate aquifers , Water Res.Research.,V.7,pp.1284-1294.

18. Masch,F.D., and K.J. Denny(1966), 'Grain size ditribution and its effect on the permeability of unconsolidated sands' Water Res.Research V.2,pp.665-667.

19. Meyer, R.R.(1963),'A chart relating well diameter,specific capacity, and the coefficients of transmissibility and and storage',in Bentall,Ray,compiler, Methods of determining permeability, transmissibility and drawdown; U.S.Geol.Survey Water Supply paper 1536-I,pp.338-340.

20. Nosova, O.N.(1962), 'Rascet Vodootdaci pescianih gruntor', Gosenergoizdat, Moskva Leningrad.

21. Rumer, R.R. and P.A. Drinker (1966), 'Resistance to laminar flow through porous media', Jour.Hydraulics Div., Amer.Soc.Civil Engrs.V.92,No.Hy5,pp.155-163.

22. Rumer, R.R.(1962),'Longitudinal dispersion in steady and unsteady flow', Jour.Hydraulics Div.,Amer.Soc.Civil Engrs., V.88,No.Hy4,pp.147-172.

23. Saxena, R.C., and G.K. Rawat (1983),'A critical study of hydraulic conductivity and specific yield in various parts of Uttar Pradesh by auger hole method . Proc. seminar in ground water development- A perspective for year 2000 A.D. IWRS, Roorkee,Vol.I,pp.179-186.

24. Theis, C.V. (1935),'The relation between the lowering of the piezometric surface and the rate and duration of discharge of a well using ground water storage', Am.Geophycs Union Trans., V.16,pp.519-524.

25. Theis, C.V.(1963),'Estimating the transmissibility of a water table aquifer from the specific capacity of a well', in Bentall, Ray,compiler, Methods of determining permeability transmissibility and drawdown; U.S. Geol. Survey Water Supply Paper -1536-I, pp.332-336.

26. Todd, D.K. (1980), 'Groundwater hydrology' second edition, John Wiley and Sons. N.Y. pp.527.

27. Verruijt, A.(1982), 'Theory of groundwater flow', second edition. The Mcmillan Press Ltd.,London,pp.143.

28. Van Schilfgaarde, J.(ed), (1979), 'Drainage for agriculture', Agronomy Monograph No.17, Amer. Soc. Agronomy, Madison,Wis., pp.700.

29. Vanchaud, G.,and J.L. Thony, (1971), 'Hystereresis during infiltration and redistributed in a soil column at different water contents', Water Res.Research,V.7, pp.111-127.

30. Wenzel, L.K.(1942),"Methods for determining permeability of water bearing materials, with special reference to discharging well methods", with a section on direct laboratory methods and a bibliography on permeability and laminar flow by V.C.Fisshel; U.S.Geol. Survey Water Supply paper, pp.887-192.