PROPERTIES OF AQUIFER MATERIALS

INTRODUCTION

The amount of water to be soaked by the earth depends upon the nature of rocks or sediments. On the basis of rocks or sediments bear to the percolation of water, these may be divided into two types–pervious and impervious rocks. The pervious structure has no connection with porous nature of rock. Many rocks that are massive and crystalline, have so many joints that may become pervious. On the other hand, many porous rocks have their pore spaces so small that water cannot pass through them freely. For the purposes of groundwater, the rocks or sediments that can hold, transit and yield water are called aquifers and accumulation of groundwater in a particular region will depend upon the presence of aquifers materials.

The most common aquifer materials are unconsolidated sands and gravel, which occur in alluvial valleys, old streambeds covered by fine deposits, coastal plains, dunes and glaciers deposits. Sandstones are also good aquifers. Cavernous limestone with solution channels, caves, underground streams and other karst development can also be high yielding aquifers. Other sedimentary rocks–shale, solid limestone etc., normally do not make good aquifers. Small yield may be possible where these rocks are highly fractured. Same is true for granite, gneiss, and other crystalline or metamorphic rocks. Basalt, lava, and other materials of volcanic origin can make excellent aquifers if they are sufficient porous (Mahajan, 1989). Four types of aquifers based on the permeability of the covering layers have been considered–(a) unconfined aquifer, (b) confined aquifer, (c) semi-unconfined aquifer and (d) semi-confined aquifer. The permeability and water quality of individual aquifer may differ considerably. It is worth mentioning that a particular aquifer at one place may be a confined aquifer while at another place it may behave as an unconfined aquifer where the water level falls below the base of the overlying confining layer. Similarly, at one particular place an aquifer may change from confined to unconfined character with time.

Most of the aquifers in the study area are unconfined aquifers of which the water table serves as the upper surface of saturation and also known as free phreatic or non-artesian aquifer. There is no clay or other restricting material at the top of the groundwater, so the groundwater levels are free to rise or fall. The top of a phreatic aquifer is the water table, which is the place where groundwater pressure is equal to atmospheric pressure. The water table height corresponds to equilibrium of water level in a well. They may extent from a few metres or less to hundreds of metres or more. The principal source of groundwater in unconfined aquifers is precipitation that has infiltrated into the soil above the aquifers, either directly as it fall on the ground or indirectly via surface runoff and seepage from stream and lakes. Some of the places of the study area, the wells tapped from the confined or artesian aquifer which may be flowing or non-flowing wells. The well known examples of flowing wells are in terai plains at a shallow depth, but the non-flowing wells are mainly observed at a greater depth of exploratory wells.

4.1 TYPE OF AQUIFER PROPERTIES

The yield of the well is probably the most important single item of ultimate interest. Particularly in most cases; the yield will determine whether or not the well is a success or failure in the eyes of the investigation drilling staff, well owner and every one etc. The individual factors that effect the yield of screened wells may be grouped as–(a) natural characteristics of the water bearing sand, (b) elements of well design and (c) the water methods used in the construction and the development of the well. But in this chapter, the only, natural characteristics of the water bearing sand or the fundamental properties of aquifer materials are discussed here in details.

An aquifer performs two important functions–(a) storage function and (b) conduit function. It stores water serving as reservoir and transmits water like a pipeline. The openings or pores in a water bearing formation serve both as storage

LOCATION OF 12 EXPLORATORY TUBEWELL SITES AND 6 HYDROGRAPH STATIONS

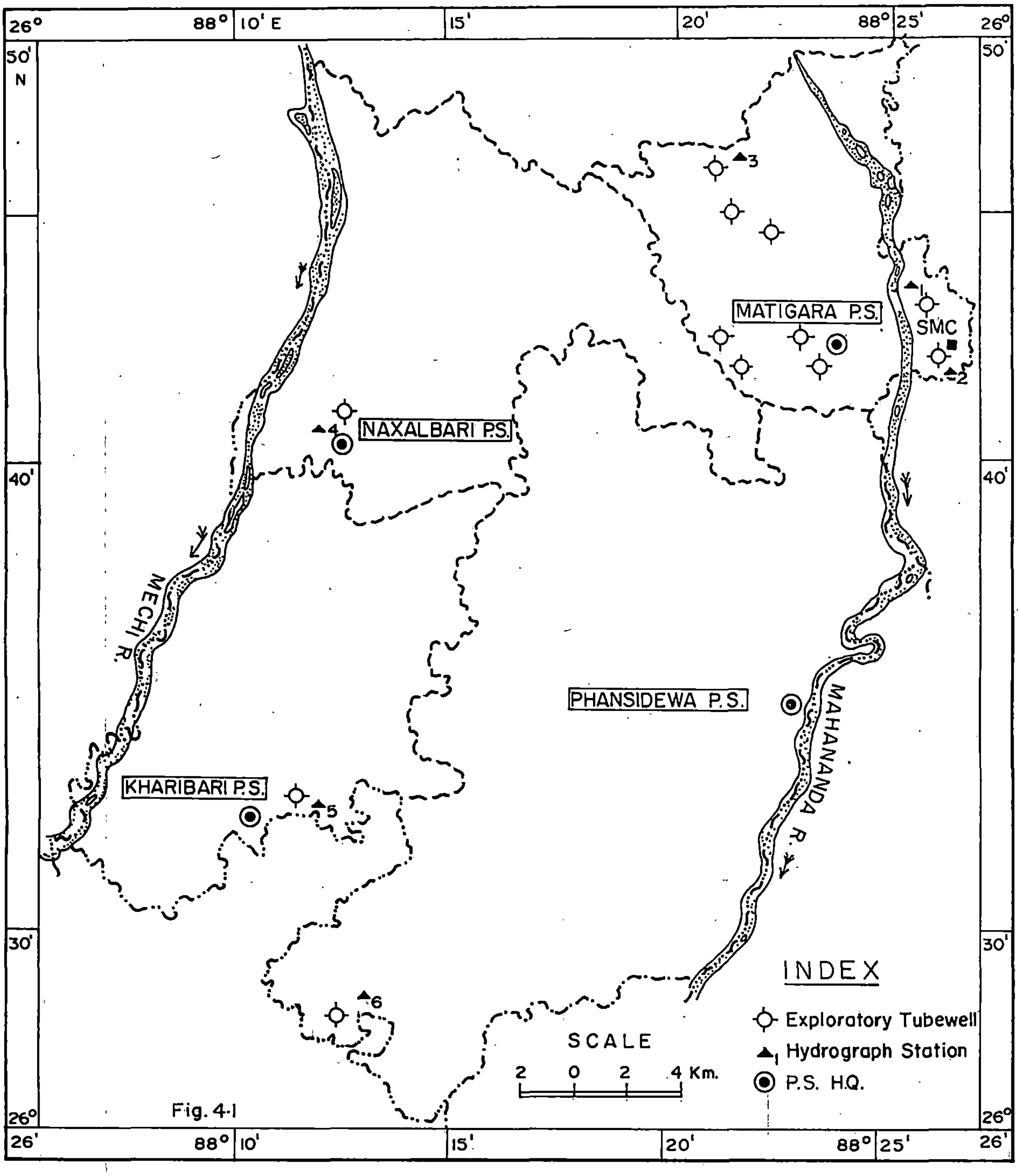

Fig. 4·1

spaces and as a network of conduits. The opening in sub-surface geologic formations are of three categories–(a) opening between individual particles as in sand and gravel, (b) crevices, joints or fractures in weathered rocks and (c) solution channels and cavern limestone and openings in basalt due to evolution of gases in lava. Porosity and specific yield are related to its storage function and conduit function is dealt with only permeability.

In the course of well-boring in Siliguri Subdivision of Darjiling district, the sediments that are penetrated, consist of incoherent materials for example clay, sand, shingle and mixture of these three. Sometimes the beds are well assorted gravels and sands are also struck by well which are treated as "good aquifers" but in most cases the sediments, that are penetrated, are unassorted which clearly show the dishevelled nature of the streams, which were active during the time of deposition of these sediments. Rock formations and the openings are the result of primary geologic processes which form the rocks and secondary processes that modify them either increasing or decreasing their porosity and permeability. Geologic processes acting on rocks may change or induce entirely new hydrogeological properties which control the movement of underground water. The hydrogeologic characteristics of sedimentary and alluvial deposits may be inherited from the parent formation. For example, weathering of sandstone may produce abundant sands which, under classifying process of nature, may be redeposited as aquifers.

In the present investigation, the hydrological processes of the aquifer materials of the alluvial tract as well as the piedmont deposits of the Terai region of Darjiling district have been determined with the help of grain size distribution. For this purpose, the core samples from 12 exploratory tubewells have been collected from different places at different depth which are shown in Fig.–4.1. A brief description of the mechanical analysis of grain size distribution of collected tube-well samples are as follows.

4.2 GRAIN SIZE DISTRIBUTION

Grain size distribution is an inherent characteristic of the soil material that

Table-4.1: Results of mechanical analysis of sand samples from the selected exploratory boreholes.

Sl No	Location	Depth ranges in m.	Cumulative weight percent, retained (mm).								
			16	4	2	0.84	0.59	0.42	0.21	0.149	0.125
01	Ashrampara, SMC	31.71 to 71.34	8.01	13.10	19.83	32.71	51.53	80.46	89.08	95.49	100.00
02	Pradhannagar, SMC	62.00 to 85.03	6.41	13.53	21.46	35.12	53.15	82.59	89.91	96.63	100.00
03	Bengdubi, Matigra	89.70 to 120.18	8.72	19.43	32.27	57.31	72.64	85.08	91.54	97.51	100.00
04	Matigara hat-I, Matigara	112.95 to 155.80	10.35	33.12	51.54	67.03	72.68	77.22	89.54	95.82	100.00
05	Matigara hat-II, Matigara	90.60 to 119.60	7.70	23.50	34.00	46.69	50.26	56.89	82.07	93.95	100.00
06	NBU, Matigara	73.00 to 93.40	6.45	18.64	39.08	52.79	56.23	60.83	81.12	94.48	100.00
07	New Chamta T. E., Matigara	42.50 to 72.50	8.24	18.61	34.64	55.53	68.12	83.09	93.27	97.87	100.00
08	Salbari Bazar, Matigara	39.65 to 87.80	10.76	26.27	33.04	55.51	74.45	85.11	92.23	96.72	100.00
09	ERS, Sukna, Matigara	106.68 to 124.67	6.58	21.82	32.10	53.21	74.73	83.84	91.18	96.91	100.00
10	Naxalbari Hos., Naxalbari	175.56 to 199.34	7.98	18.44	32.84	54.42	67.71	83.63	92.98	97.02	100.00
11	Kharibari 3rd Site, Kharibari	38.41 to 70.12	3.71	8.95	17.35	35.79	51.72	84.33	93.64	97.81	100.00
12	Bidhannagar, Phansidewa	70.43 to 107.32	5.05	12.37	24.64	43.32	59.58	83.10	93.70	99.00	100.00

distinguishes different types of soils and plays an important role in shaping their behaviour for groundwater studies. The graphical representation of the grain size analysis of soil based upon the concept that the grain size distribution curve is continuous function of grain size which is determined by the geological processes of operating to form a soil. In order to determine its gradation the soil is sieved through a stack of sieves of progressively smaller mesh sizes. Many countries have developed standards about such sieves. In this experiment, soil classification

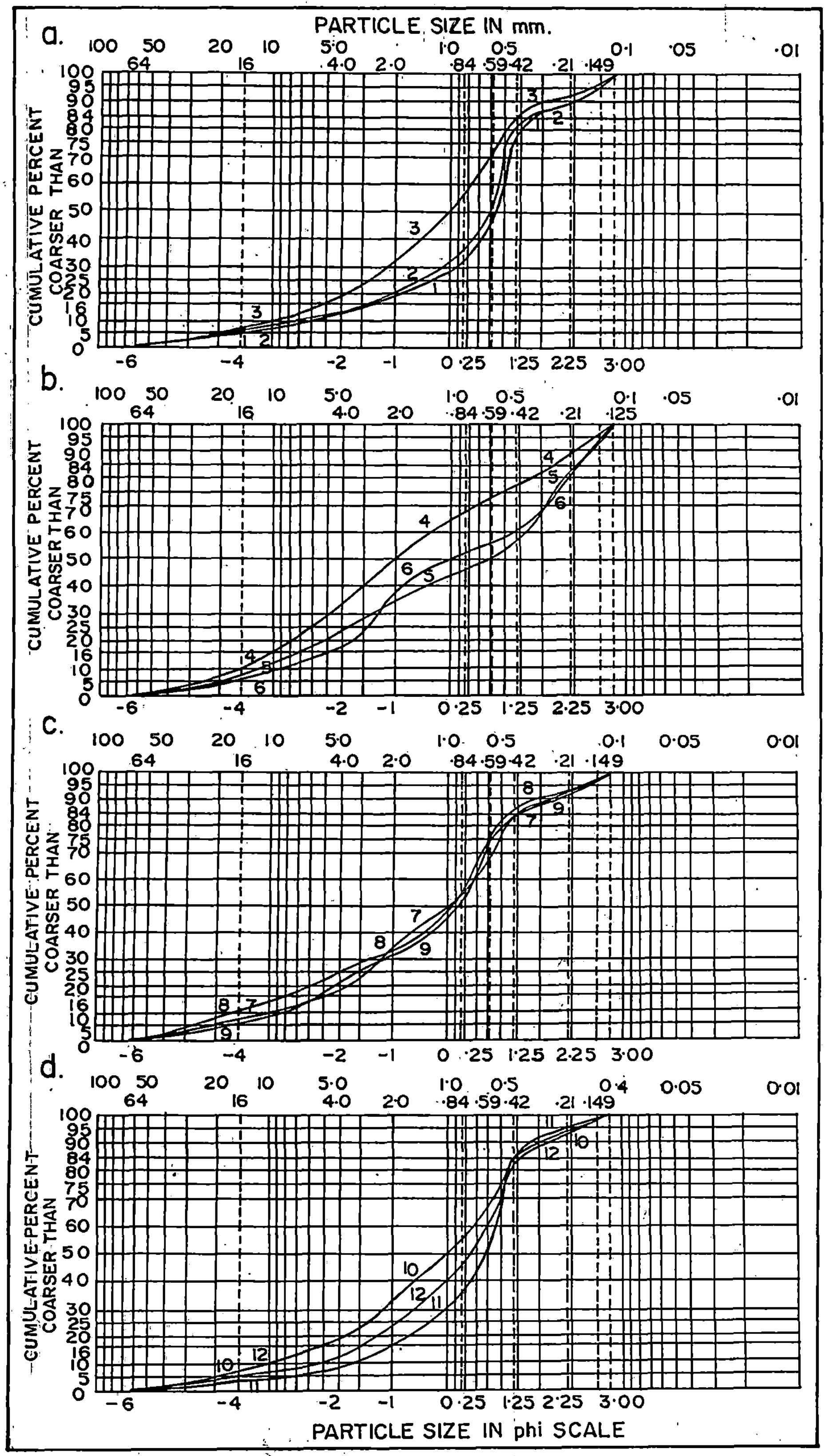

Fig.-4·2: GRAIN SIZE DISTRIBUTION CURVES OF SELECTED WELL LOG SAMPLES (1,2,3 etc. indicate the well log samples number) according to Indian Standards Institution (ISI) and Massachusetts Institute of Technology (MIT) have been considered. To carry out a sieve analysis, a representative sample of aquifer material is drawn.

Table-4.1 give the results of mechanical analysis of the tube-wells bore log sample taken only from the aquifer tapped zones or layers. The results have also been graphically exhibited in Fig.–4.2, which is important for the study and comparison of the results. The height of any curve, in these figures, at any point is the percent of material finer than the sizes indicated at the bottom of the diagram. The lines representing the diameters are spaced according to the logarithms of the diameters of the particles, thus greatly facilitating a comparison of different materials. On the basis of mechanical analysis of hydrological properties of the aquifer materials of different depths of bore logs of the tube wells from the ground surface in the Terai region of Darjiling District have been determined. Ill-sorted sediments are indicated by rectangular blocks of nearly equal size in the histogram and by a flat line in the simple frequency curve and in the cumulative frequency curve. Well-sorted materials are indicated by one or two adjacent over sized blocks in the histogram by a single well defined peak in the simple frequency curve, and by nearly vertical curve in the cumulative frequency curve. Cumulative curves are useful for obtaining the effective size of the granular materials, degree of sorting and uniformity of grain-size distribution. Some statistical values useful in groundwater hydrology is shown in Tables –4.2 to 4.4.

Table–4.2 shows the different values of different statistical parameters. First of all, the standard deviation (σ_1) measures the sorting or uniformity of the particle size distribution. In the analysed samples, the value of standard deviation ranges from 1.37 φ to 2.46 φ which indicate the nature of sorting, that is poorly to very poorly sorted grains are dominated in the study area. Skewness (Sk_1) value of the analysed samples indicates that more than 90 percent samples are coarsely to very coarsely skewed which ranges from -0.15 to 0.48, but only two sample sites located at Matigara Hat and adjoining areas are nearly symmetrical arrangement of the grains which are ranges from +0.07 to –0.09. The values of Kurtosis are normally show that more than 50 percent of the analysed samples belonged to Leptokartic which ranges from 1.16 to 2.09. The samples only located at Matigara

Hat and its surrounding areas have the characteristics of platy Kurtic. The analytical values of sphericity of pebbles of aquifers tapped formation of the analysed samples have been recognised by using the Sneed & Folk's (1958) sphericity formula as

$$\text{Sphericity} = \sqrt[3]{S^2 / LI}$$

where, L, I and S are the long, intermediate and short axes respectively of a grain.

The average value of sphericity of the study area ranges from 0.61 to 0.71 which normally indicate that the grains are mostly discoidal in shape having the characteristics of matured depositions. From the discussion, it is to be concluded that the analysed formation's sample are the admixture of gravels and sands and the presence of minimum percentage of clay and normally the formations are of matured in deposition, which over all indicate that the study area is very rich in groundwater exploitation from both in shallow and greater depth.

4.3 POROSITY

The porosity of a rock is its property of containing interstices. It is quantitatively expressed as the percentage of the total volume of the material occupied by open spaces, or that which is not occupied by solid rock material. In other words, the porosity may be designated by the formula —

$$P = 100 \times \frac{W}{V}$$

where P is the porosity by volume, W is the total volume of the pore spaces and V is the total volume of rock (Todd, 1959).

The porosity of a rock formation depends chiefly on the shape and arrangement of its constituent particles, the degree of assortment of its particles, the cementation and compacting to which it has been subjected since its deposition, the removal of mineral matter through solution by percolating water and the fracturing of rock, resulting in joints and other openings. Porosity does not depend on size of grains provided the grains are of equal in size.

The chief control over the porosity of a deposit is of variety in size grains, or the degree of assortment. A deposit of large grains of uniform size has a high porosity and a deposit of small grains of uniform size has equally high porosity but a deposit composed of mixture of grains of these two sizes has much lower porosity. Therefore, if it is possible to determine the degree of assortment or uniformity in size of grains, constituting a deposit, the porosity of it can be determined because of the fact that porosity depends on the degree of assortment of the grains. An arbitrary quantity of the 'Uniformity coefficient' is used for a simple quantitative expression of the degree of uniformity in size and is defined as the ratio of the diameter of grain that has 60 percent (by weight) of the sample finer than itself to the diameter of a grain that has 10 percent finer than itself (Meinzer, 1959). If all the grains of the sample were absolutely of the same size, the uniformity coefficient is to be 1; with most comparatively even-grained sample, the coefficient ranges from 2 to 3.

According to Hazen (1893) in regards to the relation of the uniformity coefficient to the porosity, the amount of open spaces depends on the shape and uniformity in size of the particles and is independent upon their absolute size. The materials which have the sharpest rise (Fig.– 4.2) indicates the greatest uniformity in size, have the greatest open space, while the sands having a more gradual rise pack more closely, the finer particles occupy the spaces between the larger ones or vice-versa thus greatly reducing the open spaces. It is therefore obvious that uniformity coefficient is an index to the porosity. The larger the coefficient the smaller the porosity. The arrangement of grains is also important. With all other factors remaining the same, the square packing gives the highest porosity to about 48 percent and the rhombic packing gives the lowest porosity of about 26 percent. The porosity can be anything between these two extremes depending upon the type of arrangement of the grains. Porosity is the same irrespective of the size with all other factors remaining the same. Regarding the shape of grains it is seen that the angularity tends to increase the porosity.

Hazen (1893) also determined the relation between uniformity coefficient

Table-4.2 : Different analytical values of grain size distribution in φ and some statistical parameters of the tested samples.

Grain sizes & parameters	Tested samples number											
	01	02	03	04	05	06	07	08	09	10	11	12
φ5	-4.71	-4.29	-4.19	-4.48	-4.23	-4.00	-4.58	-4.81	-4.19	-4.44	-2.99	-4.06
φ10	-4.00	-3.76	-3.33	-3.43	-2.90	-2.53	-3.03	-4.12	-2.83	-3.47	-1.73	-2.50
φ16	-3.40	-2.90	-2.20	-2.37	-1.50	-1.70	-2.27	-3.03	-2.43	-2.33	-1.00	-1.63
φ25	-2.67	-1.90	-1.53	-1.53	-0.34	-0.56	-1.43	-2.00	-1.63	-1.43	-0.31	-0.88
φ50	-1.06	0.66	0.03	-0.06	0.75	0.72	0.00	0.13	0.22	0.00	0.75	0.44
φ60	-0.31	1.44	1.25	0.41	0.91	0.84	0.47	0.41	0.53	0.50	0.97	0.75
φ75	1.00	1.94	2.03	0.88	1.03	1.13	1.00	0.78	0.81	1.00	1.22	1.06
φ84	1.94	2.36	2.39	1.25	1.53	1.44	1.34	1.19	1.44	1.28	1.25	1.38
φ90	2.39	2.61	2.61	1.75	2.36	2.36	2.00	1.50	2.12	1.78	1.56	1.98
φ95	2.69	2.86	2.81	2.58	2.75	2.75	2.58	2.20	2.69	2.42	2.31	2.50
1st quartile (d_{75})	1.00	1.94	2.03	0.88	1.03	1.13	1.00	0.78	0.81	1.00	1.22	1.06
3rd quartile (d_{25})	-2.67	-1.90	-1.53	-1.53	-0.34	-0.56	-1.43	-2.00	-1.63	-1.43	-0.31	-0.88
Median (d_{50})	-1.06	0.66	0.03	-0.06	0.75	0.72	0.00	0.13	0.22	0.00	0.75	0.44
Std. Dev. (σ_1)	2.46	2.40	2.21	1.97	1.81	1.80	1.99	2.16	2.00	1.94	1.37	1.75
Skewness (Sk_1)	0.07	-0.37	-0.09	-0.26	-0.46	-0.47	-0.27	-0.42	-0.15	-0.29	-0.48	-0.37
Kurtosis (Kg)	0.83	0.76	0.81	1.20	2.09	1.64	1.21	1.09	1.16	1.16	1.42	1.39

and the porosity for a large number of samples on the basis of which the porosity of soils and water bearing materials has been determined in the present investigation. The porosity of soils of the alluvial tract of the study area is 38 percent (Tables– 4.2 to 4.4). But the porosity of good aquifer material from where the water is taped on, both in piedmont as well as the alluvial deposits, whose specific yields is much higher than its specific retention and which can freely yield to well, is about 39 and 41 percent respectively.

The determination of porosity of the taped aquifer materials of the Terai belt comprises of a number of terraces and piedmont deposits have been sanguine when the explorations for groundwater were carried out in the terai area under Indo-U.S. Technical co-operation mission programme guided by Banerjee (1969) in 1959-61 and also by the Geologists of GSI and Hydrogeologists of CGWB of Eastern Region, who determine the value of it to be about 39 to 41 percent which fairly agree well with the value found above for the water-bearing materials. Therefore, finally, it can be said that a good water-bearing formation in the terai area of Darjiling district has a high porosity.

Table-4.3 : Hydrological properties of soils of the selected exploratory wells at an average depth of 1 meter from the ground surface.

Properties	Tested samples number											
	01	02	03	04	05	06	07	08	09	10	11	12
Effective grain size mm	.005	.007	.008	.007	.003	.008	.005	.005	.009	.026	.009	.039
Uniformity coefficient	6.5	4.5	3.5	3.4	6.2	3.8	7.2	4.3	6.8	8.5	5.5	6.6
Specific gravity	2.6	2.5	2.7	2.5	2.4	2.6	2.7	2.6	2.6	2.7	2.6	2.5
Porosity	38.0	43.0	42.0	39.0	43.0	39.0	42.0	40.0	41.0	43.0	42.0	43.0
Specific yield (S_y) (% by vol)	16.0	18.5	17.3	15.8	16.6	18.0	19.2	18.2	18.5	17.9	18.3	19.0
Sp. retention (S_r) (% by vol)	22.0	24.5	24.7	23.2	26.4	21.0	22.8	21.8	22.5	25.1	23.7	24.0
Moisture holding capacity (% by wt.)	59.5	62.2	58.8	61.4	60.6	53.7	57.4	55.3	56.8	63.1	54.4	59.8
Wilting coef. (% by wt.)	15.2	14.5	13.8	15.8	13.2	12.6	14.1	12.9	14.3	13.2	12.7	12.9
Hygroscopic coef. (% by wt.)	10.2	8.5	8.2	9.5	8.6	7.9	8.3	8.5	10.1	9.1	8.2	7.9
Water available for growth (% by wt.)	44.3	47.7	45.0	45.6	47.4	41.1	43.3	42.4	42.5	49.9	41.7	46.9

4.4 SPECIFIC YIELD AND SPECIFIC RETENTION

Below the water table or the zone of saturation, where the rock interstices are completely filled with water, all these water can not be drained by gravity or by pumping from wells, as a portion of the water is held in the void spaces by molecular and surface tension forces. The volume of water, expressed as a percentage of the total volume of the saturated aquifer, that can be drained by gravity is called Specific Yield (S_y) and the volume of water retained by molecular and surface tension forces, against the force of gravity, expressed as a percentage of the total volume of the saturated aquifer, is called the Specific retention (S_r) and correspond to field capacity. The specific yield and specific retention of a soil when added together are equal to its porosity (P) i.e., $P = S_y + S_r$.

The quantity of water yielded to wells from a body of saturated soil depends upon its specific yield and not on its porosity. Some of the aquifer materials such as clay, silt etc. have high porosity and are capable of holding large quantities of water, but yield only a small part of it in a reasonable length of time, consequently these materials usually are unproductive and insignificant for water

Table-4.4 : Hydrogeological properties of the tapped aquifer materials collected from the selected exploratory wells at different depth from b.g.l.

Properties	Tested samples number											
	01	02	03	04	05	06	07	08	09	10	11	12
Effective grain size (d_{10})	16.0 mm	14.0 mm	10.0 mm	11.5 mm	7.5 mm	6.0 mm	10.0 mm	17.5 mm	8.5 mm	11.0 mm	3.2 mm	5.6 m m
Effective screen size (d_{60})	1.37 mm	0.37 mm	0.44 mm	0.77 mm	0.52 mm	0.55 mm	0.74 mm	0.76 mm	0.69 mm	0.62 mm	0.53 mm	0.59 mm
Sorting coef. $\sqrt{(d_{75}/d_{25})}$	0.28	0.26	0.29	0.44	0.58	0.56	0.43	0.38	0.44	0.43	0.59	0.50
Uniformity coef. (d_{60}/d_{10})	0.09	0.03	0.04	0.06	0.07	0.09	0.07	0.04	0.08	0.05	0.16	0.11
Specific gravity	2.6	2.6	2.5	2.6	2.4	2.5	2.7	2.5	2.5	2.7	2.6	2.7
Porosity	41.0	42.0	40.0	42.0	39.0	40.0	41.0	39.0	40.0	41.0	40.0	42.0
Specific yield (S_y) (% by vol.)	31.0	32.5	31.3	31.6	32.0	30.7	31.5	30.0	32.0	30.0	29.8	31.2
Sp. Retention (S_r) (% by vol.)	10.0	9.5	8.7	10.4	7.0	9.3	9.5	9.0	7.0	11.0	10.2	10.8

supply, whereas a compact and fractured rock may contain much less water and yet yield abundantly. Unlike porosity, specific yield and specific retention of a rock formation is dependent on variables such as time of drainage, temperature, mineral composition of the water and various soil physical characteristics such as texture. Specific yield decreases and specific retention increases as the size of the grains decreases. Briggs and Shantz (1912) have given the following statements regarding the relationship between the moisture-holding capacity and texture of aquifer materials.

"Soil texture has been used for the quantitative description of soils more extensively than any other physical property, and unfortunately it has been one of the most difficult to interpret from the standpoint of moisture retentiveness. Texture is quantitatively expressed by means of mechanical analysis, which shows the composition of soils when the particles are separated into groups according to size. The accuracy with which the texture of the soil can be expressed by this means is dependent on the number of groups into which the particles are separated. The use of mechanical analysis as a basis for determining the moisture retentiveness of a soil is further complicated by the fact that the soils having a high clay content will show great differences in the amount of colloidal material, which greatly affects the moisture retentiveness".

Soils of the investigated area of Darjiling district do not contain an appreciable amount of clay and hence the determination of texture of soils with the help of mechanical analysis can be successfully achieved. The problem does not arise at all for aquifer materials due to the presence of very less amount of clay mixed with them. For interpreting mechanical analysis of aquifer materials with respect to their water-retaining capacities, the following equation has been developed by Briggs and Shantz (1912) based on the study of a large number of samples.

$$\text{Moisture-Holding Capacity} = (0.03\ \text{Sand} + 1.35\ \text{Silt} + 1.65\ \text{Clay}) + 21.$$

In this equation, 'moisture-holding capacity' is the 'specific retention'. Actually 'moisture-holding capacity' in the following case has been determined by using low columns of material and hence is greater than the true specific retention. The porosity minus this quantity would give the value of the specific yield which therefore, will be slight less than the true specific yield of the material. In the above equation, the term 'Sand' refers to the percentage by weight of dry sample of particles between 2 and 0.05 mm in diameter; 'Silt' refers to the percentage by weight of dry sample of particles between 0.05 and 0.005 mm and 'Clay' refers to the percentage by weight of dry sample of particles smaller than 0.005 mm in dia.

The formation, whose specific yield is much higher than the specific retention can be designated as a good water-bearing formation which can freely yield to wells. Fine textured materials not only yield less water than coarse textured materials, but they also yield it more slowly. For most water-bearing materials, the water drained out by rapid lowering of the water table in the immediate vicinity of a heavily pumped well is doubtless considerable less than the total that would be yielded by long term draining, but the draining that accompanies the annual fluctuation of the water in virtually complete.

A large number of field and laboratory methods have been devised for the measurement of specific yield. Although most field methods determine specific yield directly, most laboratory methods determine specific retention and specific yield is found indirectly by subtracting the retention from the porosity. But all these methods have limitations. Thus, laboratory samples may be disturbed, in field test it is difficult to control and measure variables and estimates lack of accuracy. Out of the seven methods proposed by Meinzer (1932), the most effective Recharge method and Mechanical analysis method have been used in the present case for determining the specific yield.

a) **Recharge Method** : It consists of observing the seepage losses from streams or canals and observing the corresponding rise in the water table. Converse to the pumping method the rise in the water table can be observed during injection of water into a recharge well. The specific yield (S_y) is obtained as a ratio of the volume of water recharged to the volume of the material by the formula :

$$S_y = 100x\frac{R_p}{r}$$

Where, R_p is the rainfall recharge to the water table and **r** the corresponding water table rise. Moreover, R_p is calculated by Rao (Rao, K., 1970):

$$R_p = 0.35\,(R - 600)$$

where, R_p is the net recharge to the groundwater and R is the rainfall in mm.

By taking a normal rainfall year (wet season only) and the amount of rainfall is substituting in the formula, the value of R_p will be determined and rise of water table in the study area in an average is 3 to 4 metres. In this case R_p is 958 mm and the corresponding value of r is 3 metres. The specific yield is :

$$S_y = 100 x \frac{958}{3000} = 31.92 \text{percent}$$

b) **Mechanical Analysis method :** Making mechanical analysis and determination of property and estimating there from the specific retention by using the Briggs and Shantz equation for Moisture–holding capacity, and finally subtracting the value of Specific retention from the porosity, the value of specific yield is determined. By using this formula , the specific yield value for good water bearing material for ground water supply is ranges between 30-32 percent, which is equal to value determined by the Recharge method. The soils of the study area of the Darjiling district have an average specific yield of 18 percent and an average specific retention of 21 percent revealing thereby most of the water in the soil zone is retained by the soils and is used for crop growth. But the specific yield of good water-bearing formation of the terai plain collected from the only taped aquifer area is approximately to 32 percent and the specific retention is 8 percent (Tables- 4.2 to 4.4).

The specific yield (S_y) of water bearing materials, determined for the north-eastern part of the West Bengal by the Geologists of GSI, Hydrogeologist of CGWB of Eastern region and Geologists and Engineers of SWID of Darjiling district by using pumping method, is about 30 percent which seems to be slight lower than the value of calculated by the said methods.

The specific yield values reported are, in general, representatives of the ultimate amount of time for drainage. From the values, it observes that the highest specific yields tend to be around the medium and coarse texture which is due to the fact that these sands are normally more uniform in grain size distribution. In the ultimate analysis one should recognise that the specific yield and specific retention are essentially dependent on the size, shape and distribution of the voids which in turn are governed by textural characteristics.

4.5 PERMEABILITY

The property of water bearing formation which is related to its pipe line or conduit function is called permeability. It is also a measure of the capacity of the porous medium to transmit water under pressure. Rocks that lack porosity do not transmit water and are said to be impermeable. But clay having maximum pores and some vesicular lavas lacking interconnected pores may be treated as impermeable. Hence, porosity is therefore not a measure of permeability. But coarse sand and gravels, which are as porous as silts, are many times more permeable by virtue of possessing larger pore spaces which allow water to move freely through them. Rocks that transmit water uniformly in all directions are said to be isotropic. But the rocks which transmit water more freely in one direction are called anisotropic. In general, the permeability in the horizontal direction parallel to the bedding plane of sedimentary rocks is more uniform than in the perpendicular direction. The movement of water from one point to another in the material takes place whenever a difference in pressure of head occurs between two points. The rate of movement of groundwater depends on the permeability of rock through which it is flowing and on the hydraulic gradient which provides force to move.

The basic law for flow of fluids through porous media was established by Darcy (1856), who states that the velocity of water flowing through a porous medium is directly proportional to the hydraulic gradient, and is expressed as :

$$v = p x \frac{h}{l} \qquad \text{..................(1)}$$

where, v = velocity of water through a column of permeable material, p = Constant of proportionality,

h = difference in head at the ends of the column,

l = length of the column.

The constant p in the equation (1) is commonly known as the coefficient of permeability. Its value varies with materials. Thus, under similar hydraulic gradients, the quantity of ground water flowing in a unit time through cross sectional area will be greater in the more permeable materials as compared with the less permeable ones. Darcy's law may then be expressed as

$$Q = P\,I\,A \quad \ldots\ldots\ldots\ldots\ldots\ldots (2)$$

Where,

Q = quantity of water discharged in a unit of time

P = coefficient of permeability

I = hydraulic gradient (the difference in head h divided by length of flow l)

A = cross- sectional area through which water percolates.

Thus, coefficient of permeability can also be expressed in terms of the rate of flow of ground water in a unit of time through a unit cross–sectional area of the material at a hydraulic gradient of 100 percent. In Meinzer's definition (Stearns,1928), the coefficient of permeability (P_m) is expressed as "the rate of flow of groundwater in gallons per day through a cross-sectional area of one square foot under a hydraulic gradient of 100 percent and a temperature of 60^0 F."

Generally, in groundwater investigations, permeability is expressed in terms of 'Field Coefficient of Permeability' which is abbreviated as P_f and is defined as "the number of gallons of water a day that percolates under prevailing conditions through each km. of water- bearing bed under investigation (measured at right angle to the direction of flow) for each meter of thickness of bed and for each foot per mile of hydraulic gradient". The permeability value depends in general upon the degree of sorting and upon arrangement of size of the particles, it is usually low for clay and other fine–textured or tightly cemented materials and high for coarse, clean gravel. Most productive water bearing materials have coefficient of permeability of 100 or above and usually above 1000. A large number of laboratory and field methods have been devised for the measurement of permeability of a taped aquifer materials. The different methods may be grouped as: (a) direct laboratory methods, (b) indirect laboratory methods, (c) field velocity methods and (d) field discharge methods. However, none of these methods could be used here due to lack of sufficient facilities. Instead, the permeability of the study area has been tried to find out and computed by adopting the "grain size distribution analysis method" as well as the following procedure applied by the different hydrogeologist and groundwater engineers.

The velocity of groundwater percolation can be determined from Hazen's formula (1893):

$$V = C x \frac{h}{l} x d^2 \ (0.70 + 0.03\,t) \ldots\ldots\ldots\ldots (3)$$

Where, V = velocity in meters per day in a solid column of same area as that of the sand.

C =constant which varies with the compactness and uniformity of sand.

h = head loss (measured in any unit)

l = distance of percolation (measured in same unit as head)

d = effective size of sand grains in mm.

t = temperature of water in degrees centigrade.

The equation (3) clearly indicates the rate of flow of ground water and therefore, the quantity of water transmitted by a sand column depends not only upon its length and head as explained earlier by Darcy (1856), but varies greatly with the effective size of the sand grains. Temperature also counts because viscocity of water and hence its percolation rate change with temperature variations. The above formula indicates that, provided the other conditions are same, the velocity of percolation of groundwater in a region will be greater in summer season than in winter season. Further, the velocity V in the equation (3) is not the actual rate of percolation. In fact, the effective or actual velocity V by the ratio V/P_e, where P_e is the effective porosity or specific yield of sand. Hazen (1893) at first stated that the constant C in equation (3) approximates 1000 and that his formula is applicable to sands of uniformity coefficient below 5 and effective size ranging from 0.10 to 3.0 mm. Subsequently, Hazen (1893) found that the value of C varies with uniformity coefficient, shape, degree of assortment and compactness of sand, and hence it may be as high as 1200 for very uniform and perfectly clean sand and as low as 400 for closely packed and dirty sand. The value of C will thus decrease with increase in uniformity coefficient.

The next attempt to find the quantity of groundwater percolation from the equation as —

$$Q = P_e x V_e \ldots \ldots \ldots \ldots \ldots \ldots (4)$$

where, Q = quantity of groundwater percolating through the medium in a unit time.

P_e = effective porosity or specific yield.

V_e = v/ P_e, then equation no. (4) may be expressed as

$Q = v \ldots \ldots \ldots \ldots \ldots (5)$

The value of coefficient of permeability(P_m (in Meinzer's units) is obtained by substituting the value of Q, computed from the equation (4), and of hydraulic gradient I in

$$P_m = \frac{Q}{I} \ldots \ldots \ldots \ldots \ldots (6)$$

The following values are obtained for the water-bearing materials in terai region of Darjiling district :

$$C = 4500, h/l = 1/2000, d = 2.8, t = 22°C, P_e = 32\%$$

Substituting the values in equation (3), we have —

$V_e = 4500 \times (1/2000) \times (2.8)^2 (0.70 + 0.03 \times 22)$

$\therefore$ $V_e = 23.9904$ m / day.

$\therefore$ $Q = P_e \times V_e = 23.9904 \times 32/100 = 7.676928$ m³/ day.

$\therefore$ $P_m = Q/ (h/l) = 7.676928 \times 2000 = 15353.856$ m³/ day.

Taylor (1936), by using Thiem's equilibrium formula (discharging well method) conducted three experiments out of which equilibrium condition is obtained in one of them. From these experiment, Taylor estimated the value of coefficient of permeability to be 4 x 10^{-3} cubic meter per second which seems to be much lower than the value calculated (Eqn-6). Further more, Singh et. al

(1939), determined the value of coefficient of permeability for Punjab sands to be 4×10^{-4} cubic meter per second which seems to be much smaller than the value calculated above. Characteristics of some common formation materials are given in Table – 4.5.

Table : 4.5 : Characteristics of some common formation materials.

Formation materials	Porosity percent	Specific yield percent	Permeability lpd/m²
Clay	45–55	1–10	0.05-100 (For Silt also)
Sand	35–40	10–30	$5 \times 10^3 - 15 \times 10^4$
Gravel	30–40	15–30	$5 \times 10^4 - 7.5 \times 10^5$
Sand and gravel	20–35	15–25	$10^3 - 2.5 \times 10^5$
Sandstone	10–20	5–15	$5 - 2.5 \times 10^3$
Shale	1–10	0.5–5	$10^{-5} - 0.1$
Limestone	1–10	0.5–5	–

Source : *Raghunath, H. M. (1987).*

4.6 MOISTURE CONTENT OF SOIL

When water is added to uniform soil and allowed to seep downward, a condition of distribution is reached at which the moisture content at all depth is approximately the same (Shaw, 1927). Moisture content of soil is also known as 'field capacity' and may be defined as the percentage of volume of water retained against gravity to total volume of material (Shaw, 1927), or as defined by the soil scientist, the percentage by weight of water retained against gravity to the weight of the sample when dry. Water content in the soils may be chiefly of three types :

a) **Available Moisture** : A portion of soil moisture is available for plant growth and hence it is subdivided by the agricultural scientists into 'available moisture' which is available to plant and 'unavailable moisture', that is unavailable to plants. The available moisture is also termed 'moisture available for growth' and equals field capacity less the amount of water held in the soil at the wilting point. Therefore to estimate the available moisture, a term 'wilting coefficient' (Briggs & Shantz, 1912) is defined which is the ratio of the weight of water in the soil when the leaves of the plants undergo permanent wilting to the weight of the soil when dry. The following equation has been developed by Briggs & Shantz (1912) for determining the wilting coefficient—

Wilting coefficient = (moisture-holding capacity — 21)/ 2.90
= 0.01 Sand + 0.12 Silt + 0.57 Clay

In this equation, moisture-holding capacity, sand, silt, and clay have the same meaning as explained before at the time of determining the specific yield and specific retention. In the study area of the terai region, the soils are having an average wilting coefficient of 12 percent by weight.

b) **Hygroscopic water** : After permanent wilting has taken place, the soil still contains some water. This remaining water is so firmly held that it is unavailable for plant growth. A part of the water that plants are unable to utilize for growth can be removed by evaporation, but some water remains in soil even after it has been fully exposed to evaporation. This last remnant is called 'hygroscopic water' (Meinzer, 1959). The term 'hygroscopic coefficient' is used to express quantitatively the capacity of a soil for holding hygroscopic water and is defined as the percentage of water in soil which in a dry condition, has been brought into a saturated atmosphere and kept in that atmosphere at a constant temperature until it has absorbed all the atmospheric water vapour that is capable of absorbing. Briggs & Shantz (1912) developed the following equation to determine the 'Hygroscopic coefficient'.

Hygroscopic coefficient = Wilting coefficient x 0.68
= (moisture-holding capacity—21)/ 2.90
= 0.007 Sand + 0.082 Silt + 0.39 Clay.

In this formula, Sand, Silt, and Clay have the same meaning as explained earlier at the time of determining the specific yield and specific retention. The 'hygroscopic coefficient' of the soils of the investigated area is approximate to 10 percent by weight.

It is observed that 'moisture-holding capacity' minus the 'wilting coefficient' gives measure of the 'water available for growth' which for the soils of the study area of the terai plain of Darjiling district is approximate to 46 percent by weight (Table–4.3) Therefore it can be inferred, on the basis of the hydrogeologic properties of soils in the terai area that the soils are of rich quality and fertile in nature because of the fact that most of the water retained in the soil is available to plant growth. From the 'mechanical analysis' of the samples, it is seen that the depth from the ground surface of a good water-bearing layer in the terai region of Darjiling district varies from place to place. From the bore hole log of the study area, it is seen that a good water yielding strata is met at a depth of 118.90 meters. From the well logs it is observed that the depth and thickness of tapped aquifer materials are variable from area to area. Thus it can be said, in the study area, a good water yielding strata is expected to meet at an average depth between 73.16–100.33 m from the ground surface and having an average thickness of the aquifer is 27 meters.

c) **Unavailable Water** : Water is retained in the soil against the force of surface evaporation and against the absorptive power of plant roots. Unavailable water probably includes water held by complex physical and physico-chemical forces.

4.7 STORAGE CAPACITY

The storage capacity of groundwater reservoir, which can freely yield water, in the terai region of Darjiling district is estimated. As stated above, a good water yielding strata has an average thickness of 27 m and an average porosity of about 41 percent. Hence, the storage capacity of the good groundwater reservoir is—

$$\text{Storage Capacity} = (41 \times 27 \times 1000 \times 1000)/100 \text{ m}^3/\text{km}^2$$
$$= 1{,}10{,}70{,}000 \text{ m}^3/\text{km}^2$$

This calculated value of storage capacity simply indicates that in the terai region of the Darjiling district, the groundwater reservoir is having an enormous storage capacity.

CONCLUSION

The foregoing discussion in this chapter gives the results of mechanical analyses of the deep tubewell samples which show an uniformity condition all over the study area. It has been observed that the porosity of soils and water-bearing materials are respectively 41.0% and 40.5%. The specific yield, specific retention and coefficient of permeability of good water-bearing formations in the area are successively 31%, 9.5% and 0.005 m^3/ minute. The soils in the alluvial tract of the study area are very rich in agricultural development and its wilting coefficient is 12% by weight and hygroscopic coefficient is 10% by weight and 45% by weight of water availability for growth of plants have been determined. The good water-bearing formation in the piedmont as well as alluvial plains of the investigated area is expected to meet an economic depth that is between 79.20 m and 124.53 m below the ground surface and is having an average thickness of 105 meters. The storage capacity of the aquifers, which can freely yield water in the study area is about 1,10,70,000 m^3/ km^2, which shows an enormous storage capacity of groundwater.

But these enormous groundwater reservoir is unconfined and occurs under water table conditions, at least upto the depths reached by the deepest well that is 219.15 m in the study area. For the better understanding of this groundwater position before the drilling, water table is the only indicator of the hydraulic conditions prevailing in the zone of saturation, which have been discussed elaborately in the subsequent Chapter.