

Determination of Aquifer Properties Using a Variable Pumping Rate and Heterogeneous Lithological Approach: A Case Study of Jujuti Village, Bardhaman

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Abstract :

Experimental solutions for constant-rate pumping tests are widely used to infer aquifer properties. This thesis explores the simulation and analysis of variable-rate pumping tests to better characterize aquifer properties under heterogeneous lithological conditions. This thesis implements a methodology that approximates the time-varying pumping record as a series of segments with linearly varying pumping rates. This approach is validated using an analytical solution for a varying pumping rate. Furthermore, our methodology is applied to analyze synthetic test data and compare the results with those obtained from the experimental test, which represents rate variations as a series of constant-rate test. To achieve these goals, the study quantifies the hydraulic properties of the aquifer, including hydraulic conductivity, transmissivity, storativity, and porosity. These properties are determined through a combination of field measurements and laboratory analyses. A notable finding from the litholog data is the transmissivity value of 2430.96 m²/day, which closely aligns with the transmissivity observed in Well OBS 1, where the flow rate ranges from 60 to 78 m³/hr, resulting in a transmissivity value of 2634 m²/day. This discrepancy leads to the application of a correction factor of 1.084 (2634/2431). This case study focuses on Jujuti Village in Bardhaman, aiming to provide a comprehensive understanding of the aquifer's behavior under variable pumping conditions and heterogeneous lithology.

Keywords: Transmissivity, Aquifer Properties, Pumping Test, Lithology, Damodar

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CHAPTER -1: INTRODUCTION

1.1. Back ground of study:

Groundwater is the water present below the earth's surface and is a vast resource of water.

Almost 22 percent of water is below the surface land in the form of groundwater. Groundwater is important as it is used for water supply in rural and urban areas. It is also often used for municipal, industrial and agricultural use by building and operating extraction wells.

Groundwater accounts for approximately 30% of all the freshwater on earth, which is second only to the polar ice 69%, and two orders of magnitude greater than the amount of freshwater in lakes and rivers 0.3%. Despite its enormous importance, this natural resource is often poorly understood, and consequently undervalued, mismanaged and even abused. In the context of growing water scarcity in many parts of the world, the vast potential of groundwater and the need to manage it carefully can no longer be overlooked.

Groundwater is available in some quantity almost everywhere. The depletion of groundwater levels is not a new story in Tunisia, but one that has continued for at least a decade, whilst annual water demand has increased for agricultural and industrial activities. Added to the depletion of groundwater resources, the arid climate aggravates the situation. Even though these resources are scarce, the hydrogeological system characterization and the artificial recharging of aquifers might help to alleviate the problem to some extent. The practice of artificial recharging is increasingly emerging as a powerful tool in water resources management (Ma and Spalding, 1997). By constructing suitable types of artificial recharge structures, groundwater resources can be augmented (Bas-a ~ gao ~ glu and Marino, 1999 ~). Many agencies have produced papers and reports on the depletion of groundwater levels in Tunisia and the need for artificial recharge of the groundwater aquifers (Besbes et al., 1978; Mamou and Kassah, 2000). Surface and sub-surface hydrological features such as lithology, geological structure, drainage density, groundwater flow and boundary conditions of the aquifer system play an important role in groundwater replenishment. But using conventional methods, it is not an easy task to study the hydrogeological basin parameters to identify the suitable sites for artificial recharge, since many controlling parameters must be independently derived and integrated, which involves additional cost, time and manpower. Groundwater recharge mapping depends on several variables and needs a computationally consistent manipulation of georeferenced information at different spatial scales. A groundwater model is a simplified depiction of a groundwater system which splits the space and/ or time into discrete sections. The governing equations and boundary conditions can be specified to vary over space and time, which enables a more realistic representation of a groundwater system. Numerical models are usually solved by a computer and are usually more computationally demanding than analytical models (Kumar, 2019). The groundwater modeling is an important structural tool for explaining many groundwater associated problems (Chakraborty et. al., 2019). A groundwater model provides a quantitative framework for synthesizing field information and for conceptualizing hydrogeological purposes (Biroju et. al., 2020). Besides, once they have been established to replicate the past behavior, they can project the outcome of

future groundwater behavior and allow the consideration of alternative management methods as a whole (Kumar, 2019). The groundwater model generally consists of two components- a conceptual model that need an understanding of the hydrological process and the other one a mathematical model, which is based on the mathematical equations (Baalousha et. al., 2008) is widely used as the calculation program broadly because of its simple methods, modular program structure and separate package to resolve special hydro-geologic problems (McDonald et. al., 1988). Significant advances in regional groundwater flow modeling have been driven by the demand to predict regional impacts of human inferences on groundwater systems and associated environment (Zhou et. al., 2011).

1.2. Research Gap:

Based on the literature review and the provided information, the research gap can be identified as the limited exploration of variable-rate pumping tests for aquifer characterization, especially in heterogeneous lithological conditions. While constant-rate pumping tests are commonly used, there is a lack of comprehensive methodologies and analyses for variable-rate tests. Additionally, there is a need to establish a more direct relationship between transmissivity values derived from lithological data and those obtained from drawdown versus time curves during variable-rate pumping tests. Further research is needed to develop and validate methodologies for approximating time-varying pumping records and to quantify hydraulic properties accurately under such conditions. This gap highlights the importance of improving aquifer characterization techniques, particularly in areas with heterogeneous lithology, to enhance groundwater management strategies.

1.3. Research Objectives:

The objective of this thesis is to determine the aquifer properties of Jujuti Village, Bardhaman, using a novel approach that incorporates variable pumping rates and accounts for heterogeneous lithological conditions. Specifically, the research aims to:

- Develop and validate a methodology for approximating time-varying pumping records as a series of segments with linearly varying pumping rates.
- Apply this methodology to analyze synthetic test data and compare the results with those from traditional constant-rate pumping tests.
- Quantify the hydraulic properties of the aquifer, including hydraulic conductivity, transmissivity, storativity, and porosity, through a combination of field measurements and laboratory analyses.
- Establishing the relationship between transmissivity values derived from lithological data and those obtained from drawdown versus time curves during pumping tests through analysis.
- Provide a comprehensive understanding of the aquifer's behavior under variable pumping conditions and heterogeneous lithology, contributing to improved aquifer characterization and management strategies.

1.4. Brief description of the study Area:

The location of the pumping test is in Jujuti village of Galsi-II block of Purba Bardhaman district, West Bengal (Figure 1). Purba Bardhaman receives an average annual rainfall of around 1400 mm while

having 75% of it pouring down during the months of June to September. The district has a tropical hot and humid climate. The mean temperature of the hottest month May reaches up to 40° C while the mean temperature of the coldest month January reaches close to 16° C.

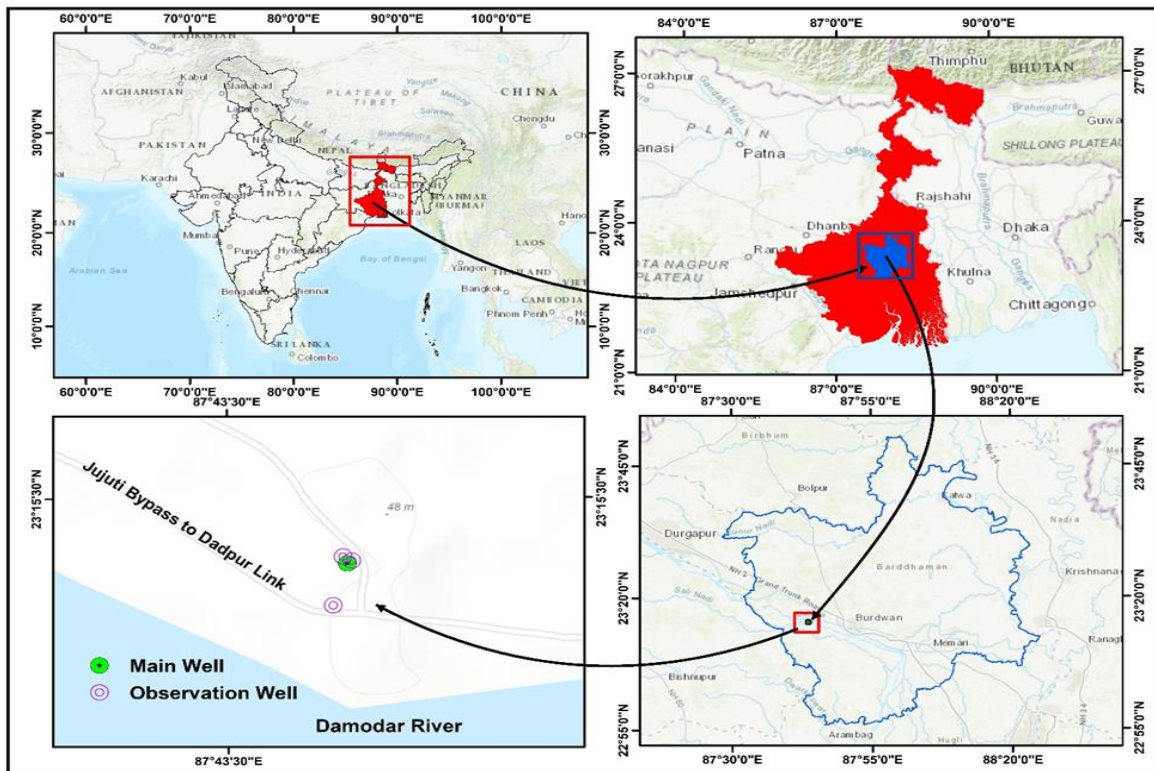


Figure 1 Location map of the study area

The Galsi-II block has a projected population of nearly 1.7 lakhs with a large portion dependent on agriculture. The land use land cover map of the block (Figure 2) shows agriculture covers 63.02% of the area with 21.49% and 11.86% areas covered by vegetation and human settlement, respectively. The rock type found in the block area is primarily of alternating clay, silt and sand layers. The lithology can be designated as upper mature deltaic plain and para deltaic flood surface.

1.5. Overall Methodology:

The research methodology for determining aquifer properties in Jujuti Village, Bardhaman, involves several key steps. Firstly, a comprehensive literature review is conducted to understand existing methodologies and theories related to aquifer properties determination, variable pumping rate analysis, and heterogeneous lithological conditions. Data collection involves the acquisition of hydrogeological data, including lithological data from boreholes and pumping test data. The experiment involves conducting variable pumping rate tests in the study area, with careful consideration of the lithological heterogeneity. Data analysis includes developing and validating a methodology for approximating time-

varying pumping records as a series of segments with linearly varying pumping rates. This methodology is then applied to analyze synthetic test data and compare the results with those from traditional constant-rate pumping tests. The hydraulic properties of the aquifer, such as hydraulic conductivity, transmissivity, storativity, and porosity, are quantified through field measurements and laboratory analyses. Additionally, the relationship between transmissivity values derived from lithological data and those obtained from drawdown versus time curves during pumping tests is established through analysis.

1.6. Proposed Research Outcome:

The research conducted in this thesis will contribute significantly to the understanding of aquifer properties under variable pumping conditions and heterogeneous lithological settings. The proposed outcomes are as follows:

- A novel methodology will be developed and validated for approximating time-varying pumping records as a series of segments with linearly varying pumping rates. This methodology will provide a more accurate representation of aquifer behavior compared to traditional constant-rate pumping tests.
- The developed methodology will be applied to analyze synthetic test data, and the results will be compared with those obtained from traditional constant-rate pumping tests. This comparison will demonstrate the effectiveness and reliability of the new approach in characterizing aquifer properties.
- The research will quantify the hydraulic properties of the aquifer, including hydraulic conductivity, transmissivity, storativity, and porosity, through a combination of field measurements and laboratory analyses. These quantifications will enhance our understanding of the aquifer's behavior under variable pumping conditions.
- The study will establish the relationship between transmissivity values derived from lithological data and those obtained from drawdown versus time curves during pumping tests. This relationship will provide valuable insights into the impact of lithology on aquifer properties.
- By providing a comprehensive understanding of aquifer behavior under variable pumping conditions and heterogeneous lithology, this research will contribute to improved aquifer characterization and management strategies. These strategies will be crucial for sustainable groundwater management in Jujuti Village, Bardhaman, and similar settings.

Overall, the proposed outcomes of this thesis will advance the field of hydrogeology by offering a more accurate and comprehensive approach to aquifer characterization under variable pumping conditions and heterogeneous lithological settings.

1.7 Outline Of Thesis:

The thesis consists of six chapters. Chapter 1 start with the discussion on background work and description of the industries under my study. Chapter 2 contains a literature review. Chapter 3 conceptualization behind the study. Chapter 4 illustrates the research methodology. Chapter 5 describes the results and discussions. The conclusion has been given in Chapter 6. References are given at the end of the dissertation

CHAPTER – 2:LITERATURE REVIEW

In order to have a better and comprehensive understanding of the present work on “**Determination of Aquifer Properties Using a Variable Pumping Rate and Heterogeneous Lithological Approach: A Case Study of Jujuti Village, Bardhaman**”, this section gives a detailed account/ description of the already accomplished research works/ case studies/ experimental and field works/review papers in this domain, that have been referred to/ cited in this present study.

- **Allan D. Randall et al.(2020)** states procedures for undertaking and analyzing recovery from aquifer tests of 13 to 132 seconds (described in reports cited herein) were applied to 51 domestic drilled wells that penetrated bedrock outside major valleys in the part of the Appalachian Plateau of New York drained by the Susquehanna River. Transmissivities calculated from these tests ranged over three orders of magnitude in both the Catskill-Cattaraugus lithofacies (shales, mudstones, siltstones, medium to coarse sandstones, pebbly sandstones) and the Chemung-Hamilton lithofacies (shales, mudstones, siltstones, fine to medium sandstones). Median transmissivity values were 0.000425 foot squared per second (36.7 feet squared per day) in the Catskill-Cattaraugus lithofacies and 0.00055 foot squared per second (47.5 feet squared per day) in the Chemung-Hamilton lithofacies. The distributions of transmissivity values within the two lithofacies were likewise similar. The range and median values of transmissivity were also nearly the same on lower and midlevel hillsides and were only slightly greater on a few upper hillsides. Transmissivities estimated from such easily arranged and analyzed tests may be appropriate for estimating groundwater flux under the small gradients that prevail under natural conditions, but not under larger drawdowns and steeper gradients near clusters of domestic wells. Four of the 51 wells tested were also pumped for 10 to 32 minutes; analysis by the Theis recovery method yielded transmissivities consistent with the brief tests for 2 wells, but 7 to 9 times smaller for 2 wells. Transmissivity values estimated by the PICKING model were not significantly different from values estimated by a automated application of the Picking method (PPC-Recovery) at a probability of 95 percent. Transmissivities calculated by either method from data for time intervals of 120 seconds or less may be of limited practical value because they apply only to a small volume of bedrock close to the pumped well.
- **Lawal Abdullateef et al. (2021)** The increasing demand for water occasioned by harsh climatic conditions, population explosion, and increasing urbanization has necessitated more attention and reliance on groundwater resources, particularly in water-limited regions. Thus, judicious management of available groundwater resources becomes crucial to meet the freshwater requirements in such zones. In this study, remote sensing and geographic information system were deployed to delineate groundwater recharge zones in semi-arid geological transition

zones of Bauchi, northeastern Nigeria. Seven thematic layers comprising elevation, slope, land use/land cover, drainage density, lithology, magnetic lineament density, and hydraulic head were integrated based on their degree of influence in groundwater recharge. Normalization of the weights was achieved through the analytical hierarchical process (AHP), while the overall integration of the thematic maps was actualized by overlay analysis in ArcGIS 10.3.1 environment to generate a final groundwater recharge zone map of the area. The resulting map was delineated into three different zones: low, moderate, and high, occupying 20% (627 km²), 77% (2352 km²), and 3% (80 km²) respectively. The overall assessment of recharge zones revealed that a high percentage of the southwestern part of the study area is characterized dominantly by poor groundwater recharge potential, attributable to high elevations and impervious rock outcrops with associated steep slopes, thus limited infiltration owing to the high velocity of runoff. The northwestern and parts of the southwestern zones, underlain by migmatite/gneiss, and partly granites, however, exhibited medium-low recharge potential, owing to the occurrence of considerably lower altitudes and occupied by thicker regoliths and underlying rocks of medium-high lineament densities. The sandstones revealed predominantly medium-high recharge potential, attributable to high permeability, lower hydraulic heads, and relatively flat geomorphology which enhances meteoric recharge and base flow processes.

- **Dana Khider Mawlood et al. (2019)** states In case of Production or (single) well pumping tests the Storativities are overestimated, it depends on radial distance between effective well bore radius in single well test. The objective of the paper proposed to drive the diffusion equation to find out the effective radial distance. Applying AQTISOLVE and Aquifer win 32 software via Jacob method for well testing, we can obtain the Storativity value and the Transmissivity using specific capacity data to check the Transmissivity value. The advantage of this method reducing the cost by having single well test (production well) without observation well. Justification the parameters of an aquifer is required because turbulent velocity and the screen of well (aquifer losses and well losses). Application of single well test lead to understand aquifer layer and the management of reservoir.
- **Ch. Suryanarayana et al. (2018)** Now a days Water scarcity in urban areas is a common problem in many cities of India, and Visakhapatnam, a fast growing industrial city on the east coast of India, is no exception. Increasing urban population, industrial expansion and shrinking surface-water sources have widened the gap between the demand and supply, resulting in groundwater depletion and saline water intrusion along the coastal region. MODFLOW is a widely used numerical groundwater flow model but requires realistic estimation of field inputs in order to contribute effectively to recommendations for proper management actions. The present study focuses on computing the spatial and temporal variations of model inputs such as

pumping and recharge rates using the field data collected from various organizations. The developed PMWIN MODFLOW model provides insight into the present and future trends in the variation of groundwater levels. Observation wells data are used in the model calibration to fix the aquifer parameters through the parameter estimation algorithm PEST. Models are performed for four projected scenarios with different rates of pumping and recharge values. Results indicate the importance of improving the recharge capability of potential areas, to sustain the aquifer's capacity to cope with stresses on groundwater resources. The model results are useful to fix optimum pumping limits in the study area for sustainable groundwater management and will help to prevent disastrous impacts on groundwater potential in the future.

- **Sasmita Sahoo et al.(2016)** Here states the Process-based groundwater models are useful to understand complex aquifer systems and make predictions about their response to hydrological changes. A conceptual model for evaluating responses to environmental changes is presented, considering the hydrogeologic framework, flow processes, aquifer hydraulic properties, boundary conditions, and sources and sinks of the groundwater system. Based on this conceptual model, a quasi-three-dimensional transient groundwater flow model was designed using MODFLOW to simulate the groundwater system of Mahanadi River delta, eastern India. The model was constructed in the context of an upper unconfined aquifer and lower confined aquifer, separated by an aquitard. Hydraulic heads of 13 shallow wells and 11 deep wells were used to calibrate transient groundwater conditions during 1997–2006, followed by validation (2007–2011). The aquifer and aquitard hydraulic properties were obtained by pumping tests and were calibrated along with the rainfall recharge. The statistical and graphical performance indicators suggested a reasonably good simulation of groundwater flow over the study area. Sensitivity analysis revealed that groundwater level is most sensitive to the hydraulic conductivities of both the aquifers, followed by vertical hydraulic conductivity of the confining layer. The calibrated model was then employed to explore groundwater-flow dynamics in response to changes in pumping and recharge conditions. The simulation results indicate that pumping has a substantial effect on the confined aquifer flow regime as compared to the unconfined aquifer. The results and insights from this study have important implications for other regional groundwater modeling studies, especially in multi-layered aquifer systems.
- **Md Bayzidul Islam et al.(2017)** States The water resources that supply most of the megacities in the world are under increased pressure because of land transformation, population growth, rapid urbanization, and climate-change impacts. Dhaka, in Bangladesh, is one of the largest of 22 growing megacities in the world, and it depends on mainly groundwater for all kinds of water needs. The regional groundwater-flow model MODFLOW-2005 was used to simulate the interaction between aquifers and rivers in steady-state and transient conditions during the period

1981–2013, to assess the impact of development and climate change on the regional groundwater resources. Detailed hydrostratigraphic units are described according to 150 lithology logs, and a three-dimensional model of the upper 400 m of the Greater Dhaka area was constructed. The results explain how the total abstraction (2.9 million m³/d) in the Dhaka megacity, which has caused regional cones of depression, is balanced by recharge and induced river leakage. The simulated outcome shows the general trend of groundwater flow in the sedimentary Holocene aquifers under a variety of hydrogeological conditions, which will assist in the future development of a rational and sustainable management approach.

- **Amah E. A et al. (2016)** Five boreholes namely IkotEdemOdo (Ak3), IkotOyom (Ak7), IkotMbakara (Ak9), AkwaObioInwang (Ak10), and IkotEkpo (Ak11) within the Akpabuyo Local Government Area of Cross River State were subjected to a number of pumping tests: step drawdown, constant discharge and recovery tests to provide some preliminary estimation of hydraulic parameters for the study area. The results indicate that transmissivity T , hydraulic conductivity k , and specific capacity SC , ranged from 485.0 m²/d to 1346.0 m²/d, 9.7 m/d to 27.9 m/d, 0.02 m³/d/m to 346. m³/d/m respectively. Mean static water level (SWL), saturated thickness of the aquifer (b) and borehole drilled depths (BDD) were 30.29 m, 48.0 m and 64.8 m, respectively. The litho-logs of the boreholes confirm that the estimated hydraulic parameters were obtained from unconfined gravelly sandy aquifers underlain by mostly sandy clay (aquitard). It is hereby recommended that an observation well should be included in a future research to enable the computation of storativity (S). Since pumping test is expensive, the use of electrical resistivity techniques should be considered for the estimation of aquifer hydraulic parameters in the area.
- **G. Tamma Rao et al. (2011)** Describe a three-dimensional steady-state finite difference groundwater flow model is used to quantify the groundwater fluxes and analyze the subsurface hydrodynamics in the basaltic terrain by giving particular emphasis to the well field that supplies domestic, agricultural, and industrial needs. The alluvial aquifer of the Ghatprabha River comprises shallow tertiary sediment deposits underlain by peninsular gneissic complex of Archean age, located in the central–eastern part of the Karnataka in southern India. Integrated hydrochemical, geophysical, and hydrogeological investigations have been helped in the conceptualization of groundwater flow model. Hydrochemical study has revealed that groundwater chemistry mainly controlled by silicate weathering in the study area. Higher concentration of TDS and NO₃-N are observed, due to domestic, agriculture, and local anthropogenic activities are directed into the groundwater, which would have increased the concentration of the ions in the water. Groundwater flow model is calibrated using head observations from 23 wells. The calibrated model is used to forecast groundwater flow pattern,

and anthropogenic contamination migration under different scenarios. The result indicates that the groundwater flows regionally towards the south of catchment area and the migration of contamination would be reached in the nearby well field in less than 10 years time. The findings of these studies are of strong relevance to addressing the groundwater pollution due to indiscriminate disposal practices of hazardous waste in areas located within the phreatic aquifer. This study has laid the foundation for developing detailed predictive groundwater model, which can be readily used for groundwater management practices.

- **Musa et al. (2015)** Describe Vertical Electrical Sounding (VES) with Schlumberger configuration were conducted at river Matan Fada, Argungu area of Kebbi State, North-western Nigeria. The Area is overlain by Gwandu Formation which consists of massive clays, clay grits, sandstones, mudstones and siltstones. A total of forty (40) VES was carried out in the study area with maximum electrode spacing of 200m. The primary parameters 'r' and 'h' and Dar-Zarrock parameters T, S, and λ were determined by curve matching and computer iteration modeling, using (Winrest) software in the sedimentary formation of the study area to demarcate ground water potential recharge zones. Maximum of five geoelectrical layers of different lithology were delineated from the VES data. The groundwater potential zones were then classified with respect to the prospect of finding groundwater in the study area as poor, moderate and good zones. Low resistivity, coupled with good aquifer thickness are found at the middle part of the area that constitute the productive water bearing zone referred to as good groundwater potential recharges aquifer.
- **Hsien-Tsung Lin et al. (2013)** States that Agricultural, aquaculture, industrial and domestic activities have placed enormous demands for water, which sometimes results in the over-pumping and associated continually declining groundwater levels. This in turn has led to land subsidence and soil salination. Therefore, it's important to understand the local pumping activities or the pumping rates in order to implement appropriate water management. The distribution of pumping rates varies spatially and temporally due to the availability of surface water and seasonality. In addition, to have correct estimate of the pumping rates, both the hydrology and geology should be considered. SWAT and MODFLOW are employed and run separately to acquire certain hydrologic components such as the recharge, boundary flow and change of aquifer storage in multi-aquifers. The water balance method (WBM) is then adopted to estimate pumping rates with these components. To validate the proposed model the results of WBM and the official records are compared. Besides, in view of the serious land subsidence occurred, artificial recharge is regarded as an effective tool to alleviate and mitigate the subsidence. Nevertheless, the location of conducting artificial recharge needs to be identified first. The potential recharge zones are assessed based on the simulated recharge rates from SWAT and the spatial distribution of hydrological characteristics of the unconfined aquifer. Ultimately,

an optimal recharge zone will be suggested. The proposed methodology is proved capable of estimating the pumping rates and locating the potential recharge zone.

- J. P. Bloomfield et al.** Describe the role of lithology in influencing basin form and function is explored empirically by investigating correlations between a range of catchment variables, where the spatial unit of analysis is not surface catchments but lithologically coherent groundwater units. Using the Thames basin, UK, as a case study, nine groundwater units have been identified. Values for 11 hydrological and geomorphological variables, including rainfall, drainage density, Baseflow Index, aquifer porosity, storage coefficient and log-hydraulic conductivity, aquifer and drainage elevation, river incision, and hypsometric integral have been estimated for each of the groundwater units in the basin, and Pearson correlation coefficients calculated for all pairs of variables. Seven of the correlation coefficients are found to be significant at a confidence level of >99%. Negative correlations between drainage density and log aquifer hydraulic conductivity, and between drainage density and river incision, and positive correlations between log-hydraulic conductivity and river incision, log-hydraulic conductivity and Baseflow Index, and between Baseflow Index and river incision are inferred to have consistent causal explanations. For example, incision of rivers into aquifers leads to relative increases in hydraulic gradients in the vicinity of rivers which, in turn, promotes the development of secondary porosity increasing both aquifer hydraulic conductivity and, hence, Baseflow Index. The implication of this interpretation is that the geomorphological evolution of basins is intimately linked to the evolution of hydraulic conductivity of the underlying aquifers. This is consistent with, and supports the notion of a coupled complexly evolving surface water-groundwater system.
- Sandra K. Richard et al. (2015)** This paper presents that in the absence of well-documented pumping tests, we investigate whether a reliable estimate of transmissivity T can be obtained using historical records of specific capacity data in granular and fractured-rock aquifers. Transmissivity values are calculated from the specific capacity data Q/s in several hundred wells located in the Saguenay-Lac-Saint-Jean region (Canada), with an iterative method applied to the Cooper-Jacob equation. The results are compared with short- and long-duration pumping tests performed throughout the region. We demonstrate that values of transmissivity compare well between the different types of tests for each lithology, when the scale of the test is similar. Therefore, using historical information easily increases the number of transmissivity values that can be estimated over a region. These values can be integrated in regional numerical models. Moreover, this study shows that $T = 4.48(Q/s)^{1.15}$ with $R^2 = 0.66$ for granular aquifers. This new empirical relationship can be used in similar environments.

- Gouri Sankar Bhunia et al. (2020)** With the increasing demands for water due to growing population, urban expansion, agricultural development and groundwater resources are gaining much attention, particularly in Gopiballavpur block, Paschim Medinipur district (West Bengal, India). Present study used cohesive approach of remote sensing and geographic information system to deliver an appropriate platform for convergent study of multi-disciplinary data and decision building for artificial groundwater recharge. Thematic maps were generated from Landsat8-Operational land Imager (OLI) and Shuttle Radar Topographic Mission (SRTM), and aquifer parameter thematic layers were organized from conventional field data. The weighted linear combination method was used to determine the weights of various themes and their classes, namely 'very good', 'good', 'poor' and 'very poor' for identifying recharge potential zone. It has been concluded that 14.58% area has very good groundwater recharge potentiality of the study area. The area having 'good', 'moderate' and 'low' recharge potential is about 32.77%, 39.67% and 12.98%, respectively.
- Izrar Ahmed et al. (2009)** States that groundwater is a major source of water for agricultural and domestic requirements in western Uttar Pradesh. Due to increasing agricultural requirements the abstraction of groundwater has increased manifold in the last two-to-three decades. The quaternary alluvium hosts the aquifer in the region. The study area forms a part of Yamuna–Krishni interfluvium. Although the area hosts potential aquifers these have been adversely affected by poor management. For effective groundwater management of a basin it is essential that a careful water balance study should be carried out. Keeping this in mind groundwater flow modelling was attempted to simulate the behaviour of the flow system and evaluate the water balance. The groundwater flow modelling was carried out. The horizontal flows, seepage losses from unlined canals, recharge from rainfall and irrigation return flows were applied using different boundary packages available in Visual MODFLOW
- Peter M. Meier et al. (1998)** In this study most pumping tests are interpreted using the classical Theis assumption of large-scale homogeneity with various corrections to account for early time behavior of drawdown curves. When drawdowns are plotted versus log time, late time data often delineate a straight line, which is consistent with Jacob's approximation of Theis's solution but may seem surprising in view of the heterogeneity of natural media. The aim of our work is to show that Jacob's method leads to a good approximation of the effective transmissivity of heterogeneous media when constrained to late time data. A review of several multiwell pumping tests demonstrates that when drawdown curves from each observation well are interpreted separately, they produce very similar transmissivity T estimates. However, the corresponding estimates for storativity span a broad range. This behavior is verified numerically for several models of formation heterogeneity. A very significant finding of the numerical investigation is that T values

estimated using simulated drawdown from individual observation wells are all very close to the effective Value for parallel flow. This was observed even in nonmulti Gaussian T fields, where high T zones are well connected and where the effective T is larger than the geometric average of point values. This implies that Jacob's method can be used for estimating effective T values in many, if not most, formations.

- J. J. Jiao et al. (1997)** Here the concepts of two-way coordinates and one-way coordinates are used to describe the different characteristics of two key aquifer parameters, transmissivity and storativity, under constant-rate pumping conditions. A two-way coordinate is such that the conditions at a given location are influenced by changes in conditions on either side of that location; a one-way coordinate is such that the conditions at a given location are influenced by changes in conditions on only one side of that location. Results from sensitivity analysis indicate that storativity has the characteristics of two-way coordinates, but transmissivity has the characteristics of one-way coordinates, i.e., its information can be transferred mainly from upstream to downstream. An upstream observation well can produce information on storativity both upstream and downstream, but it can produce little information on transmissivity downstream. These characteristics of the aquifer parameters have important implications on pumping-test designs and interpretation. For example, to estimate the parameters of an anomalous zone in an aquifer, an observation well should be located downstream but near the zone. It should not be placed upstream if the parameters downstream are to be estimated. An observation well which can provide adequate information for estimating storativity may not provide adequate information for estimating transmissivity, and vice versa. The aquifer area represented by estimated storativity may be different from that represented by estimated transmissivity.
- Jwan Sabah Mustafa et al. (2021)** The study presented a method for estimating aquifer storage coefficient from recovery test data meanwhile, it is hard to determine especially for single well test cases. The method involves an equation based on the principle of Jacob's modified Theise equation by measuring residual drawdown data versus recovery time in the production well. The main objectives of this study is to determine the reliability of this equation in the selected area and to evaluate and compare the results with the pumping test results as well. The method is starting by estimating the Aquifer parameters (Transmissivity and Storativity) which involves plotting residual drawdown as a function of time, later substituting the obtained test values in an appropriate transient groundwater flow equations based on the aquifer characteristics and types to find the parameter values. Finally, the study concluded that the proposed equation is reliable for determine the Storativity of the aquifer for the single well test, and also can be used as a check for the parameters that obtained from pumping test results and thus the method is verified the objective of the study.

- M.S. BEDINGER et al(2009)**The Sonoran region of California lies west of the Colorado River and adjoins the Mojave Desert on the west, Death Valley on the northwest, and the Salton trough on the south. The region is arid with annual precipitation ranging from less than 80 millimeters to as great as 250 millimeters in one mountain range; annual free-surface evaporation is as great as 2,500 millimeters. The characteristic basin and range topography of the region was caused by a mid-Tertiary period of intense crustal extension, accompanied by volcanic eruptions, clastic sedimentation, faulting, and tilting. Potential host media for isolation of high-level radioactive waste include granite and other coarse-grained plutonic rocks, ash-flow tuff, and basalt and basaltic andesite lava flows. Thick sections of the unsaturated zone in basin fill, intrusive, and volcanic rocks appear to have potential as host media. The region is bordered on the west by areas of relatively greater Quaternary faulting, vertical crustal uplift, and seismicity. The region has a few areas of Quaternary volcanic activity. Geothermal heat flows of 2.5 heat-flow units or greater and one earthquake of magnitude 6-7 have been recorded. The region includes topographically closed basins as well as basins that drain to the Colorado River. Dry lakes and playas occupy the closed basins. Ground-water recharge and surface runoff are small because of the small amount of precipitation and great potential evaporation. Natural ground-water discharge is by evaporation in the basin playas and by underflow to the Colorado River. Dissolved-solids concentration of ground water generally is less than 500 milligrams per liter, and much of it is of the sodium bicarbonate type. Ground water is saline in many of the playas, and chloride or sulfate is the predominant anion. Small tonnages of ore have been produced from numerous precious and fewer base-metal deposits.
- D. Bourdet et al.(2003)**Experience with a large number of well tests has shown that the analysis of some field tests using currently available interpretation models is not satisfactory. Marked departure from homogeneous behavior is evident and none of the available heterogeneous models, such as the double porosity models, yields a convincing interpretation. This is particularly apparent when the pressure derivative behavior is examined. In this paper, a new analytical solution is presented that describes the pressure response of a well intercepting a layered reservoir with crossflow. The model is designed to describe a system of alternating beds of relatively high permeability contrast. The interpretation models currently in use (homogeneous reservoir, two layers without crossflow, double porosity reservoir) are shown to be limiting forms of the new solution which is therefore a more general description of transient pressure responses. The layered reservoir behavior is discussed in terms of the pressure and the derivative of pressure. The limits of applicability of the homogeneous and the double porosity solutions for analysis of tests in layered formations are clearly defined. The new model is used to analyze an actual field case and a good match is obtained when no other solution is found to be applicable. In addition to the usual well and reservoir parameters (total permeability thickness product, skin and wellbore storage constant),

analysis yields quantitative information on the storativity and permeability of the contrasting layers.

- **JOHN D. BREDEHOEFT et al.(2005)**A method for testing formations of very low permeability is presented. The method is based on an analytical solution that describes the decay of a head change caused by pressurizing the volume of water stored in a shut-in well. Type curves prepared from this solution are matched with observed data to determine the hydraulic properties of the formation tested. The test is similar to the conventional slug test; however, its much shorter duration makes the testing of extremely tight formations feasible.
- **Thomas E. Reilly et al.(2009)**Cylindrical (axisymmetric) flow to a well is an important specialized topic of ground-water hydraulics and has been applied by many investigators to determine aquifer properties and determine heads and flows in the vicinity of the well. A recent modification to the U.S. Geological Survey Modular Three-Dimensional Finite-Difference Ground-Water Flow Model provides the opportunity to simulate axisymmetric flow to a well. The theory involves the conceptualization of a system of concentric shells that are capable of reproducing the large variations in gradient in the vicinity of the well by decreasing their area in the direction of the well. The computer program presented serves as a preprocessor to the U.S. Geological Survey model by creating the input data file needed to implement the axisymmetric conceptualization. Data input requirements to this preprocessor are described and a comparison with a known analytical solution indicates that the model functions appropriately.
- **Ochuko Anomohanran et al.(2014)**The aim of this study was to determine the aquifer parameters in Erho, Nigeria. This was carried out by drilling three test wells and three observation wells in the study location. Water was pumped from the test wells at a constant rate while the drawdown in the observation wells was determined as a function of time. The data obtained were subjected to graphical and analytical evaluation using the Cooper-Jacob method. The result revealed that the mean values of the transmissivity, storage coefficient and specific capacity of the aquifer layer are 6.15×10^{-2} m²/min, 9.3×10^{-4} and 0.33 m²/min respectively. Analysis of the results indicated that the aquifer will be very productive and able to supply sufficient quantity of water for the people. The result also revealed that the aquifer is confined and therefore dependable to provide good quality water for the people living in the area.

- H. H. Cooper et al.(1995)**The capacities of a water-bearing formation to transmit water under a hydraulic gradient and to yield water from storage when the water table or artesian pressure declines, are generally expressed, respectively, in terms of a coefficient of transmissibility and a coefficient of storage. Determinations of these two constants are almost always involved in quantitative studies of ground-water problems.C. V. THEIS [1935, see "References" at end of paper] gave an equation, adapted from the solution of the analogous problem in heat conduction, for computing the non-steady drawdown accompanying the radial flow of water to a well of constant discharge. This equation has been used successfully many times for determining coefficients of transmissibility and storage from observed drawdowns. As it involves a transcendental function known as the exponential integral and two unknown coefficients, one of which occurs both in the argument and as a divisor of the function, the coefficients cannot be determined directly. However, they may be determined by a graphical method devised by THEIS and described by JACOB [1940,p. 582] and WENZEL [1942, pp. 88-89]. This method requires the use of a "type curve," on which the observed data are superimposed to determine the coefficients. Later, WENZEL and GREENLEE [1944] gave a generalization of THEIS' graphical method by which the coefficients may be determined from tests of one or more discharging wells operated at changing rates. This method requires the computation of a special type curve for each observation of drawdown used. It is without doubt a worth-while contribution to the quantitative techniques of groundwater hydraulics, but in tests that involve more than a very few discharging wells or a very few changes in the rates of discharge, the computation of the special type curves is necessarily so laborious as to make the method difficult to apply.The present paper gives a simple straight-line graphical method for accomplishing the same purposes as the methods developed by THEIS and by WENZELand GREENLEE. Type curves are not required. The writers believe that the straight-line method, where applicable, has decided advantages, in ease of application and interpretation, over the other graphical methods. However, as the method will not be applicable in some cases, it is expected to supplement, rather than supersede, the other methods. The method is designed especially for artesian conditions, but it may be applied successfully to tests of non-artesian aquifers under favorable circumstances.
- Keith J. Halford et al.(2006)**Interpretation of single-well tests with the Cooper-Jacob method remains more reasonable than most alternatives. Drawdowns from 628 simulated single-well tests where transmissivity was specified were interpreted withthe Cooper-Jacob straight-line method to estimate transmissivity. Error and bias as a function of vertical anisotropy, partial penetration, specific yield, and interpretive technique were investigated for transmissivities thatranged from 10 to 10,000 m²/d. Cooper-Jacob transmissivity estimates in confined aquifers were affected minimally by partial penetration, vertical anisotropy, or analyst. Cooper-Jacob transmissivity estimates of simulatedunconfined aquifers averaged twice the known values. Transmissivity estimates of unconfined aquifers were notimproved by interpreting results with

an unconfined aquifer solution. Judicious interpretation of late-time data consistently improved estimates where transmissivity exceeded 250 m²/d in unconfined aquifers.

- **Faisal Ahammed et al.(2013)**This paper deals with a study on rainfall characteristics of Dhaka, the capital city of Bangladesh for the period of 1953 to 2009. Data were collected from Bangladesh Meteorological Department in January 2011 and found 2.84% missing data. Descriptive statistical analysis was conducted on annual rainfall, annual daily and monthly maximum rainfall. We applied Gumbel distribution function to estimate return periods of extreme rainfall events and found that annual daily maximum rainfall equal or greater than 425 mm had a return period of 100 years. Normal distribution function was adopted to forecast rainfall variability due to global climate change and found that annual daily maximum rainfall equal or greater than 200 mm might occur in any 12 years during the period of 2010 to 2066. The outcomes of this paper can be used in better understanding rainfall patterns.
- **Md. Kawser Ahmed et al.(2016)**A long-term (1948 to 2012) trend of precipitation (annual, pre-monsoon, monsoon, and post-monsoon seasons) in Bangladesh was analyzed in different regions using both parametric and nonparametric approaches. Moreover, the possible teleconnections of precipitation (annual and monsoon) variability with El Niño/Southern Oscillation (ENSO) episode and Indian Ocean Dipole (IOD) were investigated using both average and individual (both positive and negative) values of ENSO index and IOD. Our findings suggested that for annual precipitation, a significant increasing monotonic trend was found in whole Bangladesh (4.87 mm/year), its western region (5.82 mm/year) including Rangpur (9.41 mm/year) and Khulna (4.95 mm/year), and Sylhet (10.12 mm/year) and Barisal (6.94 mm/year) from eastern region. In pre-monsoon, only Rangpur (2.88 mm/year) showed significant increasing trend, while in monsoon, whole Bangladesh (3.04 mm/year), Sylhet (7.17 mm/year), and Barisal (6.94 mm/year) showed similar trend. In post-monsoon, there was no significant trend. Our results also revealed that the precipitation (annual or monsoon) of whole Bangladesh and almost all of the spatial regions did not show any significant correlation with ENSO events, whereas the average IOD values showed significant correlation only in monsoon precipitation of western region. The individual positive IODs showed significant correlation in whole Bangladesh, western region, and its two divisions (Rajshahi and Khulna). So, in the context of Bangladesh climate, IOD has the more teleconnection to precipitation than that of ENSO. Our findings indicate that the co-occurrence of ENSO and IOD events may suppress their influence on each other.

- **William Graham Burgess et al.(2010)**Dhaka, capital of Bangladesh, is a megacity dependant on groundwater for the majority of its water supply, pumped from 420 boreholes distributed throughout the city. Recharge to the groundwater aquifer is insufficient to balance abstraction, groundwater levels are in decline and water quality is compromised by seepage from areas of urban and industrial contamination and leakage from polluted rivers. The strategic importance of the naturally high-quality groundwater resource demands a targeted programme of aquifer protection. Repeated surveys at over 100 boreholes demonstrate the patterns and trends of deteriorating groundwater quality. Groundwater modelling and environmental isotope distributions have been used independently to evaluate the significance of potential sources of pollution. Both approaches identify the polluted River Buriganga as the main threat to groundwater quality, indicating priorities for monitoring and aquifer protection.
- **Mohammad A. Hoque et al.(2007)**Large abstraction by water-wells has been causing a linear to exponential drop in groundwater level and substantial aquifer dewatering in Dhaka, Bangladesh. The city is almost entirely dependent on groundwater, which occurs beneath the area in an unconsolidated PlioPleistocene sandy aquifer. Analysis shows that the pattern of water-level change largely replicates the patterns of change in the rate of groundwater abstraction. Contribution of the aquifer storage to the abstraction is estimated to be more than 15% in the year 2002. This abstraction has caused a sharp drop in water level throughout the city and turned into two cones of depression in the water level. Upper parts of the aquifer are already dewatered throughout the area, with the exception of part of the northeast and southeast corner of the city. It is calculated that about 41 million cubic metres (MCM) of the aquifer dewatered by the year 1988, which increased to 2,272 MCM in the year 2002. Water-level decline may increase non-linearly due to limiting vertical recharge in areas where the aquifer is dewatered and may severely threaten the sustainability of the aquifer.
- **M. A. Hoque et al.(2014)**Dhaka, the capital of Bangladesh, is home to a population of 15 million people, whose water supply is 85% drawn from groundwater in aquifers that underlie the city. Values of Cl/Br >500 are common in groundwater beneath western Dhaka in areas <3 km from the river, and in rivers and sewers around and within the city. The study shows that groundwater beneath western Dhaka is strongly influenced by infiltration of effluent from leaking sewers and unsewered sanitation, and by river-bank infiltration from the TuragBuriganga river system which bounds the western limit of the city. River-bank infiltration from other rivers around Dhaka is minor. Values of Cl/Br and Cl concentrations reveal that 23 % of wells sampled in Dhaka are influenced by saline connate water in amounts up to 1%. This residual natural salinity compromises the use of electrical conductivity of groundwater as a method for defining

pathways of recharge by contaminated surface waters. Concentrations of As, B, Ba, Cd, Cu, F, Ni, NO₃, Pb, Sb, Se and U in groundwater samples are less than WHO health-based guideline values for drinking water.

CHAPTER – 3 FUNDAMENTALS OF GROUND WATER FLOW AND WELL HYDRAULICS

3.1. INTRODUCTION:

The underground water found in fully saturated soils and geologic formations below the water table is typically referred to as groundwater. Groundwater is not just an asset. It is a crucial aspect of the natural world; it causes issues with the ecosystem and, occasionally, provides a means for ecological remedies. It is a component of the hydrologic cycle, and knowing how it functions within this cycle is needed to support integrated assessments in the evaluation of watershed resources, and in the evaluation of environmental contamination on a regional scale. When it comes to engineering, groundwater helps to such geotechnical issues including land subsidence and slope stability. Groundwater is also essential to comprehending a variety of several different geological processes.

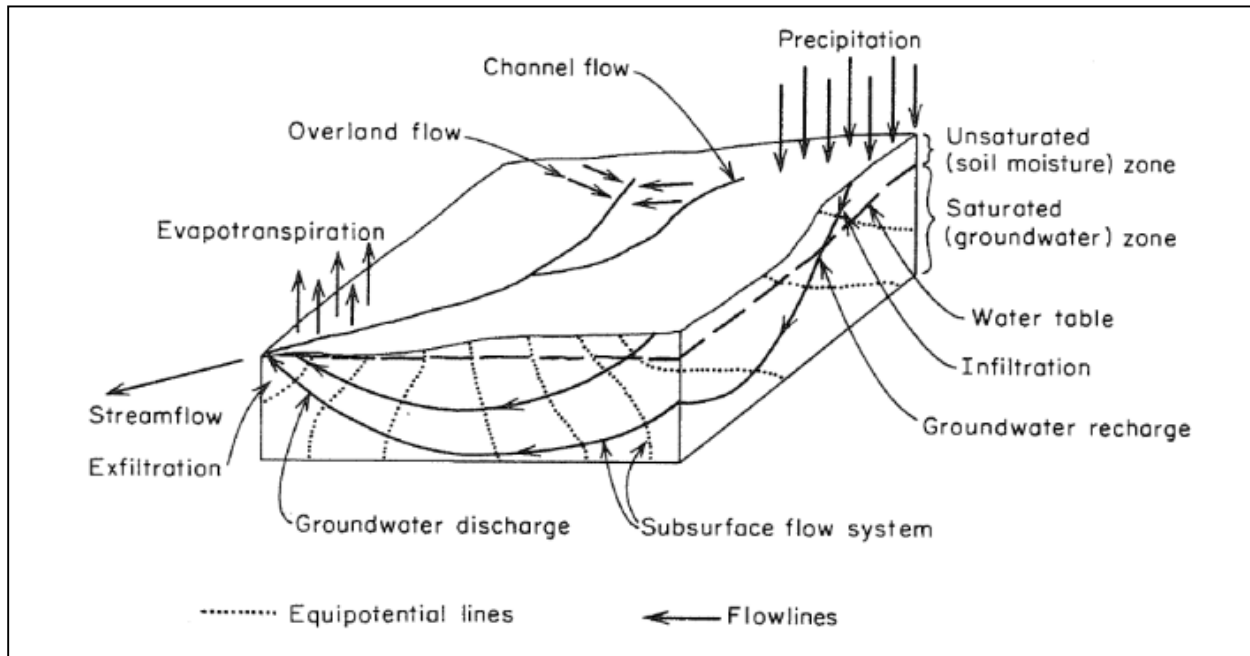


Figure 2 Schematic representation of the Hydrologic Cycle (Freeze & Cherry, 1979)

3.2. Types of groundwater:

Connate water refers to the groundwater that is trapped within the pore spaces of sedimentary rocks during their formation. As layers of sediment accumulate over time, pressure increases, squeezing water out of the sediment. However, some water remains trapped within the rock's pores, becoming connate water. This water is distinct from other groundwater sources because it is essentially "fossil water," having been isolated from the hydrological cycle for millions of years. Connate water is typically saline and can be found in reservoirs deep underground. It's an important consideration in various geological

and hydrological studies, particularly in the exploration and extraction of oil and natural gas, as well as in understanding groundwater dynamics.

3.3. Groundwater and the Hydrologic Cycle:

The water cycle describes where water is on Earth and how it moves. Water is stored in the atmosphere, on the land surface, and below the ground. It can be a liquid, a solid, or a gas. Liquid water can be fresh or saline (salty). Water moves between the places it is stored. Water moves at large scales, through watersheds, the atmosphere, and below the Earth's surface. Water moves at very small scales too. It is in us, plants, and other organisms. Human activities impact the water cycle, affecting where water is stored, how it moves, and how clean it is.

Oceans store 96% of all water on Earth. Ocean water is saline, meaning it's salty. On land, saline water is stored in saline lakes. The rest of the water on Earth is fresh water. Fresh water is stored in liquid form in freshwater lakes, artificial reservoirs, rivers, and wetlands. Water is stored in solid, frozen form in ice sheets and glaciers, and in snowpack at high elevations or near Earth's poles. Water vapor is a gas and is stored as atmospheric moisture over the ocean and land. In the soil, frozen water is stored as permafrost and liquid water is stored as soil moisture. Deeper below ground, liquid water is stored as groundwater in aquifers. Water in groundwater aquifers is found within cracks and pores in the rock.

As it moves, water can change form between liquid, solid, and gas. Circulation mixes water in the oceans and transports water vapor in the atmosphere. Water moves between the atmosphere and the surface through evaporation, evapotranspiration, and precipitation. Water moves across the surface through snowmelt, runoff, and streamflow. Water moves into the ground through infiltration and groundwater recharge. Underground, groundwater flows within aquifers. Groundwater can return to the surface through natural discharge into rivers, the ocean, and from springs.

Climate change is actively affecting the water cycle. It is impacting water quantity and timing. Precipitation patterns are changing. The frequency, intensity, and length of extreme weather events, like floods or droughts, are also changing. Ocean sea levels are rising, leading to coastal flooding. Climate change is also impacting water quality. It is causing ocean acidification which damages the shells and skeletons of many marine organisms. Climate change increases the likelihood and intensity of wildfires, which introduces unwanted pollutants from soot and ash into nearby lakes and streams.

3.4. Vertical Distribution of Groundwater:

3.4.1. Saturated zone:

This zone, also known as groundwater zone, is the space in which all the pores of the soil are filled with water under hydrostatic pressure. The water table forms its upper limit and marks a free surface, i.e., a surface having atmospheric pressure.

3.4.2 Zone of aeration:

In this zone the soil pores are only partially saturated with water. The space between the land surface and the water table marks the extent of this zone. The zone of aeration has three subzones:

3.4.3 Soil Water Zone:

This lies close to the ground surface in the major root band of the vegetation from which the water is lost to the atmosphere by evapotranspiration.

3.4.4 Intermediate Vadose Zone:

This lies between the soil water zone and the capillary fringe. The thickness of the zone of aeration and its constituent subzones depend upon the soil texture and moisture content and vary from region to region. The soil moisture in the zone of aeration is of importance in agricultural practice and irrigation.

3.4.5 Capillary Fringe:

In this the water is held by capillary action. This zone extends from the water table upwards to the limit of the capillary rise.

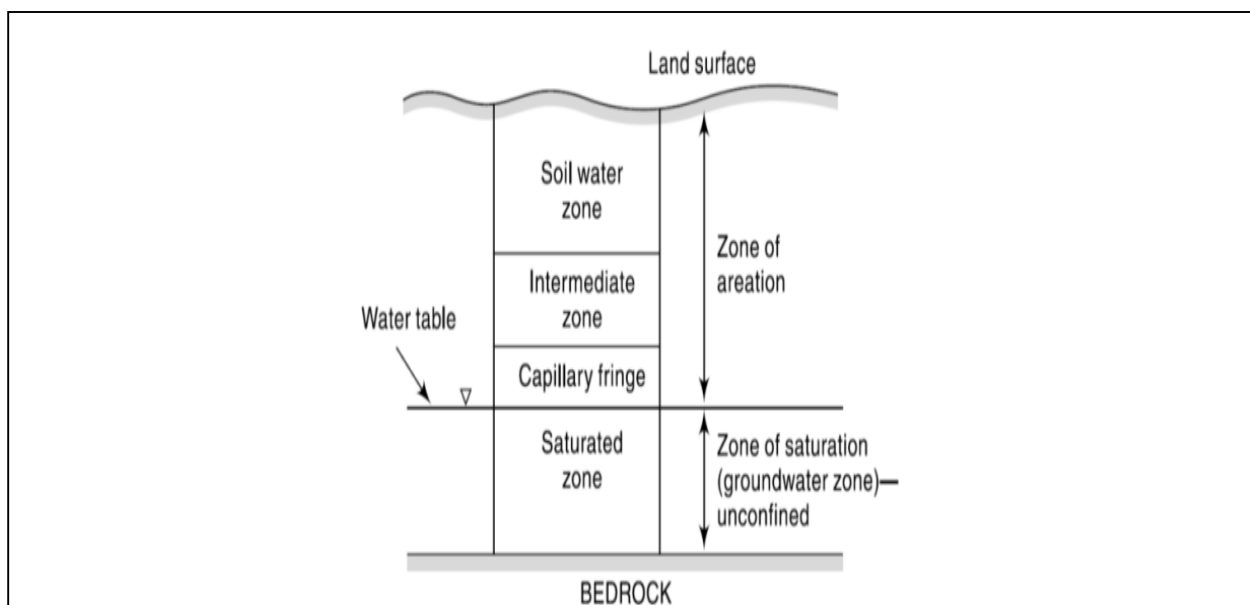


Figure 3 Divisions of subsurface water (Subramanya, 2008)

3.5. Saturated Formation

All earth materials, from soils to rocks have pore spaces. Although these pores are completely saturated with water below the water table, from the groundwater utilization aspect only such material through which water moves easily and hence can be extracted with ease are significant. On this basis the saturated formations are classified into four categories.

3.5.1. Aquifer:

An aquifer is a saturated formation of earth material which not only stores water but yields it in sufficient quantity. Thus an aquifer transmits water relatively easily due to its high permeability. Unconsolidated deposits of sand and gravel form good aquifers.

3.5.2. Aquitard:

It is a formation through which only seepage is possible and thus the yield is insignificant compared to an aquifer. It is partly permeable. A sandy clay unit is an example of aquitard. Through an aquitard appreciable quantities of water may leak to an aquifer below it.

3.5.3. Aquiclude:

It is a geological formation which is essentially impermeable to the flow of water. It may be considered as closed to water movement even though it may contain large amounts of water due to its high porosity. Clay is an example of an aquiclude.

3.5.4. Aquifuge:

It is a geological formation which is neither porous nor permeable. There are no interconnected openings and hence it cannot transmit water. Massive compact rock without any fractures is an aquifuge.

The definitions of aquifer, aquitard and aquiclude as above are relative. A formation which may be considered as an aquifer at a place where water is at a premium (e.g. arid zones) may be classified as an aquitard or even aquiclude in an area where plenty of water is available.

The availability of groundwater from an aquifer at a place depends upon the rates of withdrawal and replenishment (recharge). Aquifers play the roles of both a transmission conduit and a storage. Aquifers are classified as unconfined aquifers and confined aquifers on the basis of their occurrence and field situation. An unconfined aquifer (also known as water table aquifer) is one in which a free water surface, i.e. a water table exists. Only the saturated zone of this aquifer is of importance in groundwater studies. Recharge of this aquifer takes place through infiltration of precipitation from the ground surface. A well driven into an unconfined aquifer will indicate a static water level corresponding to the water table level at that location.

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On the other hand, a confined aquifer (also known as artesian aquifer) is an aquifer which is confined between two impervious beds such as aquicludes or aquifuges. Recharge of this aquifer takes place only in the area where it is exposed at the ground surface. The water in the confined aquifer will be under pressure and hence the piezometric level will be much higher than the top level of the aquifer. At some locations, the piezometric level can attain a level higher than the land surface and a well driven into the aquifer at such a location will flow freely without the aid of any pump. A confined aquifer is called a leaky aquifer if either or both of its confining beds are aquitards.

Whether a stream feeds water into groundwater storage or the other way around depends on the location of the water table in relation to the water level in the stream. During times of low stream flows, if the stream's bed is below the groundwater table, the water surface may drop below the general the groundwater and the level of the water table both influence the stream's flow. These streams that are receiving Effluent streams are groundwater flows. These include streams and rivers that are perennial.

If, however, the water table is below the bed of the stream, the stream-water percolates to the groundwater storage and a hump is formed in the groundwater table. Such streams which contribute to the groundwater are known as influent streams. Intermittent rivers and streams which go dry during long periods of dry spell (i.e., no rain periods) are of this kind.

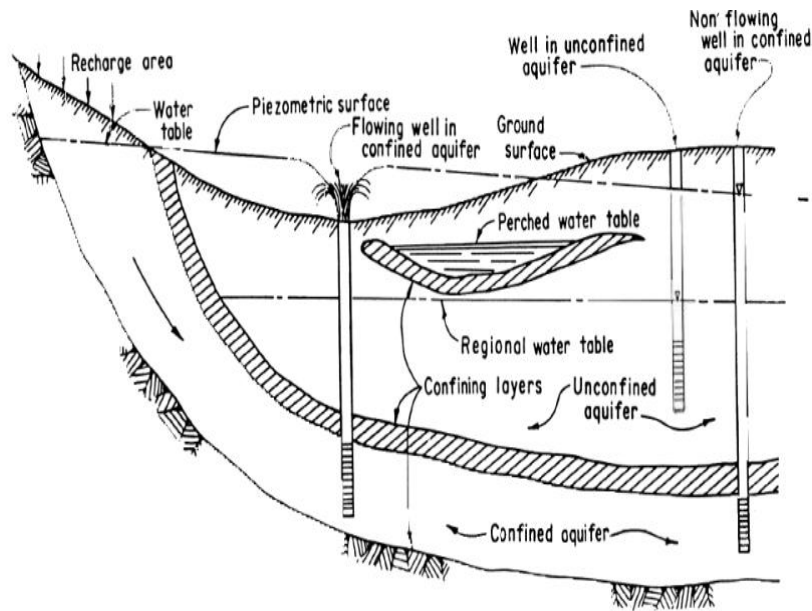


Figure 4Types of aquifers (U.S. Bureau of Reclamation, 1995)

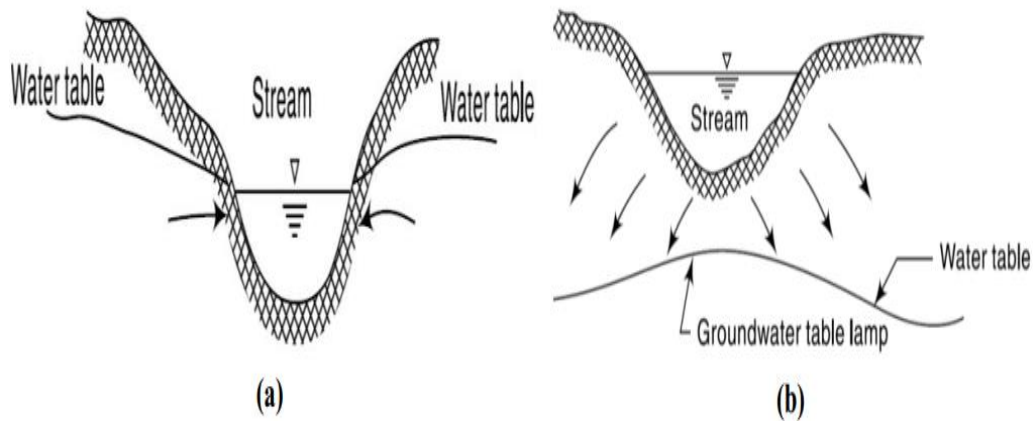


Figure 5Effluent Stream (b) Influent Stream (Subramanya, 2008)

3.6. Aquifer Properties:

3.6.1. POROSITY:

Aquifer porosity refers to the volume percentage of pore space within a rock or sediment that can contain water. It's a critical parameter in hydrogeology because it determines how much water a

formation can store and how easily water can flow through it. Porosity (ϕ) is commonly expressed as a fraction or percentage.

The equation to calculate porosity is:

$$\phi = V_v / V_t \quad \dots\dots (3.1)$$

Where: ϕ is the porosity (dimensionless)

V_v is the volume of void space (pore space) within the rock or sediment (cubic meters or cubic feet)

V_t is the total volume of the rock or sediment (cubic meters or cubic feet)

If you have access to bulk density (ρ_b) and grain density (ρ_g) of the material, you can also use the equation:

$$\phi = (\rho_g - \rho_b) / (\rho_g - \rho_w) \quad \dots\dots\dots (3.2)$$

Where: ρ_w is the density of water (typically around 1000 kg/m³ or 62.4 lb/ft³)

This equation is based on the assumption that the solid grains making up the rock have a uniform density. It's important to note that porosity is affected by factors such as grain size, sorting, and packing arrangement of grains within the rock. So, in some cases, more complex equations or empirical methods may be used to estimate porosity accurately.

Primary porosity comprises the original interstices created when a rock or soil was formed in its present state. In soil and sedimentary rocks the primary interstices are the spaces between grains or pebbles. In intrusive igneous rocks the few primary interstices result from cooling and crystallization. Extrusive igneous rocks may have large openings and high porosity resulting from the expansion of gas, but the openings may or may not be connected. Metamorphism of igneous or sedimentary rocks generally reduces the primary porosity and may virtually obliterate it.

Fractures such as joints, faults, and openings along planes of bedding or schistosity in consolidated rocks having low primary porosity and permeability may afford appreciable secondary porosity. In some rocks such secondary porosity affords the only means for the storage and movement of ground water. Solution of carbonate rocks such as limestone or dolomite by water containing dissolved carbon dioxide takes place mainly along joints and bedding planes and may greatly increase the secondary porosity. Similarly, solution of gypsum or anhydrite by water alone may greatly increase the secondary porosity.

3.6.2. Specific Yield and Specific Retention:

Specific yield and specific retention are two important parameters used in hydrogeology to describe how water is stored and moves within an aquifer.

Specific Yield (S_y): Specific yield refers to the ratio of the volume of water that an aquifer will yield by gravity drainage to the total volume of the aquifer. It represents the portion of water that an aquifer releases under the influence of gravity and is typically expressed as a percentage or decimal fraction.

The equation to calculate specific yield is:

$$S_y = V_w / V_t \quad \dots\dots\dots(3.3)$$

Where:

S_y is the specific yield (dimensionless)

V_w is the volume of water that drains out of the aquifer due to gravity (cubic meters or cubic feet)

V_t is the total volume of the aquifer (cubic meters or cubic feet)

Specific Retention (S_r): Specific retention refers to the volume of water retained in an aquifer per unit volume of aquifer material. It represents the portion of water that is retained within the pore spaces of the aquifer material against the pull of gravity and is typically expressed as a percentage or decimal fraction.

The equation to calculate specific retention is:

$$S_r = V_r / V_t$$

Where:

S_r is the specific retention (dimensionless)

V_r is the volume of water retained within the aquifer material (cubic meters or cubic feet)

V_t is the total volume of the aquifer (cubic meters or cubic feet)

It's important to note that specific yield and specific retention are often determined through laboratory tests or estimated using empirical relationships based on the characteristics of the aquifer material, such as porosity, grain size distribution, and hydraulic conductivity. These parameters play a crucial role in understanding groundwater storage, flow, and availability within aquifer systems.

While porosity gives a measure of the water-storage capability of a formation, not all the water held in the pores is available for extraction by pumping or draining by gravity. The pores hold back some water by molecular attraction and surface tension. The actual volume of water that can be extracted by the force of gravity from a unit volume of aquifer material is known as the specific yield .

3.6.3. Storage Coefficient or Storativity:

Storage coefficient, also known as storativity, is a dimensionless parameter used in hydrogeology to describe the ability of an aquifer to store water per unit decline or rise in hydraulic head. It represents the volume of water that a unit volume of aquifer material releases or takes in under a unit change in hydraulic head. Storativity takes into account both specific yield (which accounts for gravity drainage) and specific storage (which accounts for compression and expansion of the aquifer material).

The equation to calculate storativity (S) is:

$$S = S_y + S_s \quad \dots\dots\dots (3.4)$$

Where:

S is the storage coefficient or storativity (dimensionless)

S_y is the specific yield (dimensionless)

S_s is the specific storage (dimensionless)

Specific storage (S_s) represents the compressibility of the aquifer material and is typically expressed in units of inverse length (e.g., 1/m or 1/ft).

The relationship between storativity, specific yield, and specific storage can be complex and depends on factors such as the porosity, compressibility, and hydraulic properties of the aquifer material.

Storativity is an important parameter in groundwater flow modeling and aquifer analysis, as it influences the response of aquifers to changes in groundwater levels, such as those caused by pumping or recharge events. It helps hydro geologists understand how much water can be extracted from or stored in an aquifer under various conditions.

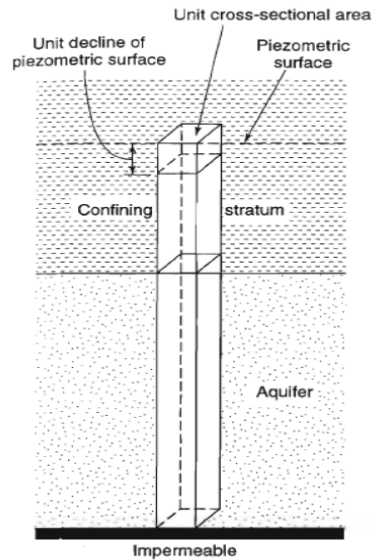


Figure 6 Illustrative sketches for defining storage coefficient of a confined aquifer

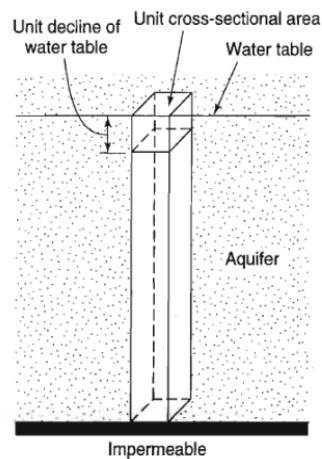


Figure 7 Illustrative sketches for defining storage coefficient of an unconfined aquifer

3.6.4. Specific Storage:

Specific storage (S_s) is a parameter used in hydrogeology to describe the ability of an aquifer material to release or take in water per unit volume per unit change in hydraulic head. It represents the compressibility of the aquifer material. Specific storage is particularly important in unconfined aquifers or where changes in hydraulic head lead to significant changes in the volume of water stored in the aquifer.

The equation to calculate specific storage is:

$$S_s = (1/n) * (\Delta V / \Delta h) \quad (3.5)$$

Where:

S_s is the specific storage (units of 1/length, often expressed in m^{-1} or ft^{-1})

n is the porosity of the aquifer material (dimensionless, ranging from 0 to 1)

ΔV is the change in volume of water in the aquifer (cubic meters or cubic feet)

Δh is the change in hydraulic head (meters or feet)

Specific storage quantifies how much the volume of water stored in the aquifer changes with a change in hydraulic head. It is a measure of the compressibility of the aquifer material, with higher values indicating greater compressibility.

3.6.5. Specific Storage:

Specific storage is a crucial parameter in groundwater flow modeling, especially when analyzing transient or time-varying conditions. It influences the rate at which water flows into or out of the aquifer in response to changes in pumping rates, recharge rates, or other hydrological events.

3.7. Darcy's Law:

Darcy's Law is a fundamental principle in hydrogeology and fluid dynamics that describes the flow of fluids (usually water) through porous media, such as soil or rock. It's named after the French engineer Henry Darcy who first formulated it in the 19th century.

The law states that the rate of flow of water through a porous medium is directly proportional to the hydraulic gradient and the cross-sectional area through which the water flows, and inversely proportional to the length of the flow path and the fluid's viscosity. Mathematically, it can be expressed as:

$$Q = -k * (A/L) * (h_2 - h_1) \quad \dots\dots\dots(3.6)$$

Where:

Q is the volumetric flow rate (volume of fluid passing through a given cross-sectional area per unit time),

k is the hydraulic conductivity of the porous medium (a measure of its ability to transmit water), typically measured in units of velocity (e.g., meters per second or feet per day),

A is the cross-sectional area through which the fluid flows (orthogonal to the flow direction),

L is the length of the flow path,

h_1 and h_2 are the hydraulic heads (elevation of the water surface relative to a reference point) at points 1 and 2 along the flow path, respectively.

In this equation, the hydraulic gradient $(h_2 - h_1)/L$ represents the change in hydraulic head per unit length of flow path, and $-k$ accounts for the resistance to flow posed by the porous medium.

Darcy's Law is applicable under certain conditions, such as when the flow is steady, the fluid is incompressible, and the porous medium is homogeneous and isotropic. Despite its simplifications, it serves as a foundational concept in groundwater hydrology and is widely used in various engineering applications, including groundwater modeling, contaminant transport, and petroleum reservoir engineering.

3.8. Coefficient of Permeability:

The coefficient of permeability, often denoted as k or K , is a fundamental property of porous media that quantifies its ability to transmit fluids. It's a key parameter in Darcy's Law, as mentioned earlier.

The coefficient of permeability is defined as the rate of flow of fluid through a unit cross-sectional area of the porous medium under a unit hydraulic gradient. Mathematically, it can be expressed as:

$$k = \frac{Q \cdot L}{A \cdot \Delta h} \quad \dots\dots\dots(3.8)$$

Where:

k is the coefficient of permeability (often in units of velocity, such as m/s or cm/s),

Q is the volumetric flow rate of fluid passing through the medium (m^3/s or cm^3/s),

L is the length of the flow path (m or cm),

A is the cross-sectional area perpendicular to the flow direction (m^2 or cm^2),

Δh is the difference in hydraulic head (m or cm) between two points along the flow path.

This equation essentially states that the coefficient of permeability is the proportionality constant relating the flow rate of fluid to the hydraulic gradient, cross-sectional area, and length of the flow path. A higher coefficient of permeability indicates greater ease of fluid flow through the porous medium.

The coefficient of permeability is influenced by various factors including the properties of the fluid (such as viscosity), the properties of the porous medium (such as porosity and grain size distribution), and the geometry of the medium. It's typically determined experimentally through laboratory tests such

as constant head or falling head permeability tests, where the flow rate through a sample of the porous medium is measured under controlled conditions.

The coefficient of permeability is a crucial parameter in fields like hydrogeology, civil engineering, and soil mechanics, as it governs the movement of fluids through soils, rocks, and other porous materials. Understanding this property is essential for various applications, including groundwater flow modeling, design of drainage systems, and assessment of contaminant transport in aquifers.

3.9. Transmissibility:

In hydrogeology, the transmissibility of water, often denoted as TT , refers to the ability of an aquifer to transmit water through it under a unit hydraulic gradient. It's a measure of the aquifer's hydraulic conductivity in combination with its thickness. Transmissibility is particularly relevant when analyzing the flow of groundwater through porous media.

The equation for transmissibility is:

$$T = k \cdot b \quad \dots\dots\dots(3.9)$$

Where:

T is the transmissibility (often in units of velocity times thickness, such as m^2/s or ft^2/day),

k is the hydraulic conductivity of the aquifer (m/s or ft/day),

b is the thickness of the aquifer through which water is flowing (m or ft).

This equation indicates that transmissibility is the product of hydraulic conductivity and aquifer thickness. It essentially represents the capacity of the aquifer to transmit groundwater flow per unit width of the aquifer. A higher transmissibility value indicates a greater ability of the aquifer to transmit water.

Transmissibility is a critical parameter in groundwater hydrology, as it influences the rate and direction of groundwater flow within an aquifer. It's used extensively in groundwater modeling to simulate and predict groundwater flow patterns, drawdown effects from pumping wells, and contaminant transport.

Understanding transmissibility is essential for water resource management, environmental protection, and various engineering applications related to groundwater, such as well field design, aquifer storage and recovery projects, and groundwater remediation strategies.

3.10. The coefficient of permeability and intrinsic permeability values can vary widely depending on the type of porous media involved. Here are some typical ranges for these parameters for various porous media:

Sands and Gravels:

Coefficient of Permeability: 10^{-2} to 10^{-4} m/s

Intrinsic Permeability: 10^{-10} to 10^{-8} m²

Clayey Soils:

Coefficient of Permeability: 10^{-6} to 10^{-10} m/s

Intrinsic Permeability: 10^{-14} to 10^{-10} m²

Silty Soils:

Coefficient of Permeability: 10^{-5} to 10^{-8} m/s

Intrinsic Permeability: 10^{-13} to 10^{-10} m²

Limestones:

Coefficient of Permeability: 10^{-5} to 10^{-2} m/s

Intrinsic Permeability: 10^{-12} to 10^{-8} m²

Sandstones:

Coefficient of Permeability: 10^{-4} to 10^{-11} m/s

Intrinsic Permeability: 10^{-11} to 10^{-7} m²

Fractured Rocks (such as granite):

Coefficient of Permeability: 10^{-5} to 10^{-2} m/s

Intrinsic Permeability: 10^{-11} to 10^{-7} m²

These values are approximate and can vary depending on factors such as porosity, grain size distribution, degree of consolidation, and the presence of fractures or other structural features. They are typically determined through laboratory tests or estimated based on empirical relationships and geological characteristics of the porous media.

3.11. Stratification of aquifers:

The stratification of aquifers refers to the occurrence of multiple layers of porous materials with different hydraulic properties stacked on top of each other beneath the Earth's surface. These layers, known as aquifer units or aquifer systems, may have varying permeability, porosity, thickness, and other hydraulic characteristics.

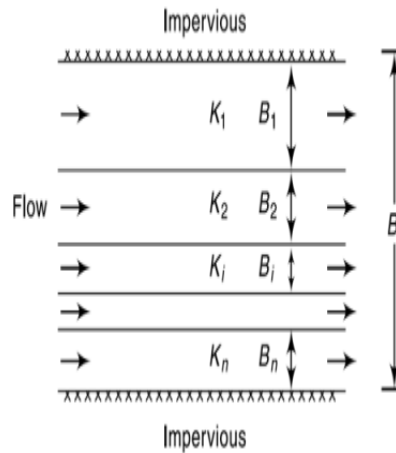


Figure 8 Flow parallel to the stratification (Subramanya, 2008)

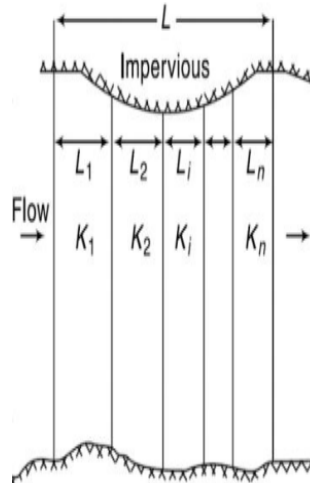


Figure 9 Flow normal to the stratification (Subramanya, 2008)

In this diagram, we have three layers representing different types of aquifers stacked vertically. The top layer represents an unconfined aquifer, followed by a semi-confined aquifer in the middle, and a confined aquifer at the bottom.

Equation-wise, we can represent the flow through a stratified aquifer system using Darcy's Law, modified to account for the characteristics of each individual layer. Let's denote the hydraulic conductivity, thickness, and hydraulic head of each layer as k_i , b_i , and h_i respectively, where i represents the layer number.

Then, the flow through each layer can be described by:

$$Q_i = -k_i \cdot A_i \cdot ((h_{i+1}) - (h_i)) / L_i \quad \text{.....(3.10)}$$

Where:

Q_i is the volumetric flow rate through layer i

k_i is the hydraulic conductivity of layer i

A_i is the cross-sectional area of layer i

h_{i+1} and h_i are the hydraulic heads at the boundaries of layer i ,

L_i is the thickness of layer i .

The total flow through the stratified aquifer system can then be obtained by summing the flow through each layer:

$$Q_{\text{total}} = \sum_i Q_i$$

This equation accounts for the varying properties of each layer and how they contribute to the overall flow within the aquifer system.

Understanding the stratification of aquifers is crucial for groundwater management and resource evaluation, as it affects the movement and distribution of groundwater, as well as the recharge and discharge processes within the aquifer system.

3.12. Anisotropy of Hydraulic Conductivity:

Anisotropy of hydraulic conductivity refers to the directional variation in the ability of a porous medium to transmit fluid. In other words, the hydraulic conductivity of the medium is different in different directions. This phenomenon is commonly encountered in naturally occurring geological formations where the sedimentary or structural features lead to preferential flow paths along certain directions.

Mathematically, anisotropy of hydraulic conductivity can be represented using a tensorial approach, where the hydraulic conductivity tensor K describes the hydraulic conductivity in three dimensions. For an isotropic medium (where hydraulic conductivity is the same in all directions), this tensor simplifies to a scalar value. However, for anisotropic media, the tensor has distinct components corresponding to different directions.

In a Cartesian coordinate system, the hydraulic conductivity tensor K can be represented as:

Where:

$$K = \begin{pmatrix} K_{xx} & K_{xy} & K_{xz} \\ K_{yx} & K_{yy} & K_{yz} \\ K_{zx} & K_{zy} & K_{zz} \end{pmatrix} \quad \dots\dots\dots(3.11)$$

K_{xx} , K_{yy} , and K_{zz} represent the hydraulic conductivities in the x , y , and z directions respectively.

$K_{xy}=K_{yx}=K_{yx}$, $K_{xz}=K_{zx}=K_{zx}$, and $K_{yz}=K_{zy}=K_{zy}$ represent the cross-components of hydraulic conductivity, indicating how the conductivity changes when moving in a combination of directions.

Anisotropy can also be quantified using a parameter called the anisotropy ratio (λ), which is defined as the ratio of the maximum hydraulic conductivity to the minimum hydraulic conductivity:

$$\lambda = \frac{K_{\max}}{K_{\min}} \quad \lambda = \frac{K_{\min}}{K_{\max}}$$

Where $K_{\max}$ and $K_{\min}$ are the maximum and minimum hydraulic conductivities respectively.

The anisotropy of hydraulic conductivity has significant implications for groundwater flow and contaminant transport in subsurface systems. Understanding and characterizing anisotropy is crucial for accurately modeling and predicting groundwater flow patterns and the migration of contaminants in heterogeneous geological formations.

3.13. Compressibility of aquifers:

In confined aquifers, the total overburden pressure at any point is borne by the combined action of the pore water pressure and inter-granular pressure. The compressibility of the aquifer and also that of the pore water causes a readjustment of these pressures whenever there is a change in storage and thus have an important bearing in the storage characteristics of the aquifer.

- i. The elemental volume is constrained in the lateral directions and undergoes change of length in the z -direction only, i.e., ΔA is constant.
- ii. The pore water is compressible.
- iii. The solid grains of the aquifer are incompressible but the pore structure is compressible.

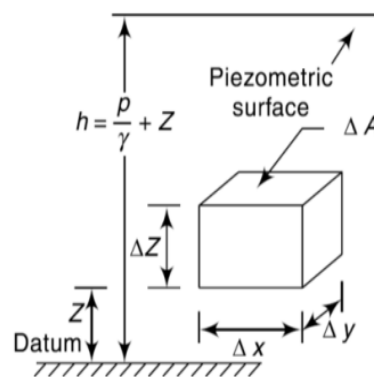


Fig 1 Elemental volume of a compressible aquifer (Subramanya, 2008)

The reciprocal of the bulk modulus of elasticity of a substance is indeed a measure of its compressibility. Bulk modulus of elasticity (often denoted as K) is a measure of a

substance's resistance to compression under uniform pressure. It quantifies how much a material will compress under a given pressure.

The bulk modulus of elasticity K is defined as the ratio of hydrostatic pressure change (dp) to the resulting fractional volume change (dV/V):

$$K = -(dp)/(dV/V)$$

The reciprocal of the bulk modulus of elasticity K gives us the compressibility (C) of the substance:

$$C = 1/K$$

For water, the bulk modulus of elasticity is approximately $(2.2 \times 10)^9$ N/m². Therefore, the compressibility of water can be calculated as:

$$C = 1/(2.2 \times 10)^9 \text{ m}^2/\text{N}$$

$$\frac{d(\Delta M)}{\rho \Delta V} = n \beta dp + \alpha (1 - n) dp + n \alpha dp = (n \beta + \alpha) dp = \gamma (n \beta + \alpha) dh = S_s dh \quad \dots (3.12)$$

This gives us a measure of how much volume change water undergoes per unit change in pressure. In other words, it tells us how easily water can be compressed. The compressibility of water is relatively low compared to many other fluids, indicating that it is not easily compressed under typical environmental conditions.

3.14. Groundwater Movement:

Groundwater movement, a vital component of the Earth's hydrological cycle, entails the flow of water through porous geological formations beneath the surface. Governed by physical laws such as Darcy's Law, groundwater migrates through interconnected pore spaces within aquifers, driven by hydraulic gradients and influenced by aquifer properties like porosity and permeability. This slow but persistent flow sustains ecosystems, supplies water for agriculture, industry, and human consumption, and serves as a buffer against droughts. Additionally, fracture flow, occurring through geological faults or fractures, contributes to more rapid groundwater transport in certain settings. Understanding groundwater movement is essential for managing water resources, assessing groundwater availability, and safeguarding against contamination. Through a combination of field studies, numerical modeling, and remote sensing techniques, hydrogeologists strive to unravel the complex dynamics of groundwater flow, ensuring its sustainable use and protection for future generations.

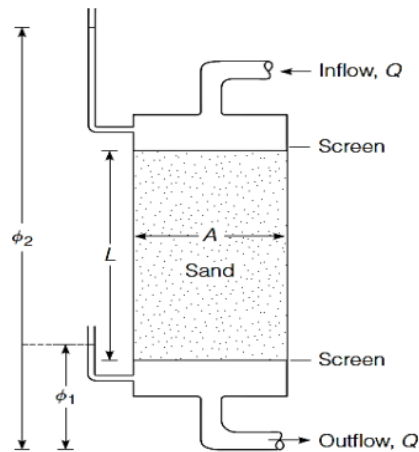


Figure 10 Schematic diagram of Darcy's experimental setup (Chin, 2013)

3.15. Limitations of Darcy's Law:

3.15.1. Homogeneous and Isotropic Media:

Darcy's Law assumes that the porous medium through which groundwater flows is homogeneous (uniform properties) and isotropic (uniform in all directions). This limitation restricts its applicability to situations where aquifers exhibit significant heterogeneity or anisotropy.

3.15.2. Steady-State Flow:

Darcy's Law is most accurate for steady-state flow conditions, where flow rates and hydraulic gradients remain constant over time. In unsteady flow situations, such as during recharge events or pump tests, the law may not accurately represent transient flow behavior.

3.15.3. Linear Flow Assumption:

Darcy's Law assumes that flow is linear, meaning that the relationship between hydraulic gradient and flow rate is constant. In reality, flow in some aquifers may exhibit non-linear behavior, especially under high gradients or turbulent flow conditions.

3.15.4. Neglects Gravitational Effects:

Darcy's Law does not account for the effects of gravity on groundwater flow, which can be significant in situations where vertical gradients are pronounced, such as in deep aquifers or near hillsides.

3.15.5. Neglects Density Variations:

Darcy's Law assumes that groundwater density remains constant, neglecting variations due to temperature, salinity, or chemical composition. In situations where these variations are significant, such as in saline intrusion or thermal convection, the law may not accurately predict flow behavior.

3.15.6. Neglects Chemical Reactions:

Darcy's Law does not consider the effects of chemical reactions on groundwater flow, such as dissolution, precipitation, or redox reactions. In aquifers with reactive minerals or contaminant plumes, these reactions can influence flow patterns and alter hydraulic properties over time.

The equation for Darcy's Law is:

$$Q = -k(A/L)(h_2 - h_1) \quad \dots\dots\dots(3.13)$$

Where:

Q is the volumetric flow rate,

k is the hydraulic conductivity of the porous medium,

A is the cross-sectional area through which flow occurs,

L is the length of the flow path,

h_1 and h_2 are the hydraulic heads at two points along the flow path.

Understanding these limitations is crucial for interpreting groundwater flow data and for applying Darcy's Law appropriately in hydrogeological studies.

Darcy's law indicates a linear relationship between the flow rate and the gradient in the piezometric head measured in the direction of flow. This linear relationship is the same as for laminar flow in pipes under low Reynolds number (Hagen-Poiseuille flow) and is symptomatic of the dominance of viscous forces as water flows through the pore spaces. Flow in porous media is typically laminar under natural conditions. As the Reynolds number increases, viscous forces become less dominant, and the flow rate deviates from being linearly proportional to the piezometric head gradient, and the relationship becomes nonlinear.

3.16. General 3-D Flow Equation in a porous medium:

Consider the control volume as shown in the figure the net mass inflow rate to the control volume is given by,

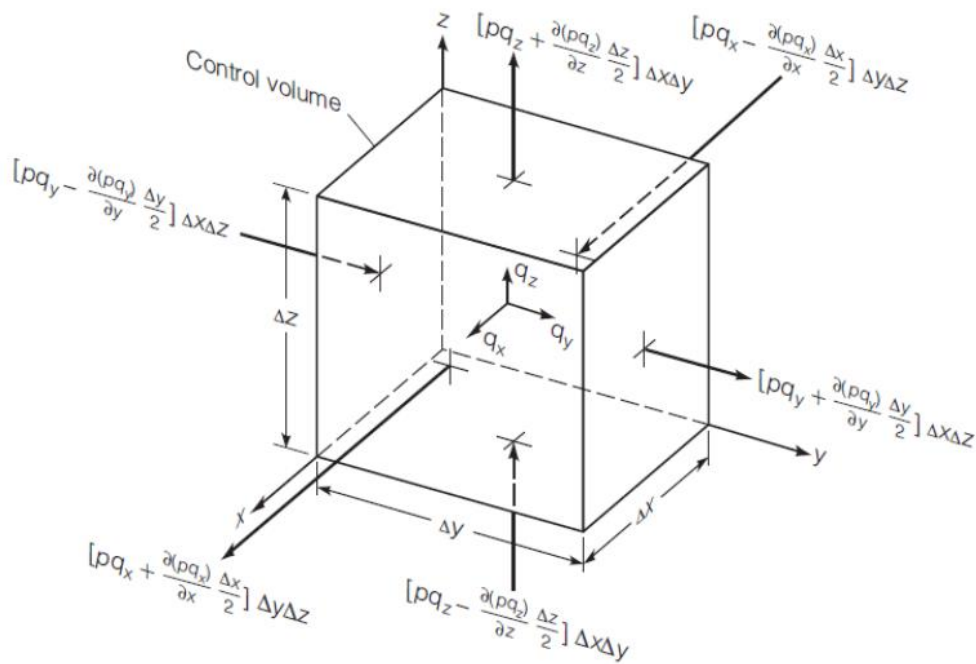


Figure 11Control volume in a porous medium (Chin, 2013)

$$\begin{aligned}
 \text{Net mass inflow rate} = & \left[(\rho q_x) - \frac{\partial(\rho q_x)}{\partial x} \frac{\Delta x}{2} \right] \Delta y \Delta z - \left[(\rho q_x) + \frac{\partial(\rho q_x)}{\partial x} \frac{\Delta x}{2} \right] \Delta y \Delta z \\
 & \left[(\rho q_y) - \frac{\partial(\rho q_y)}{\partial y} \frac{\Delta y}{2} \right] \Delta x \Delta z - \left[(\rho q_y) + \frac{\partial(\rho q_y)}{\partial y} \frac{\Delta y}{2} \right] \Delta x \Delta z \\
 & \left[(\rho q_z) - \frac{\partial(\rho q_z)}{\partial z} \frac{\Delta z}{2} \right] \Delta x \Delta y - \left[(\rho q_z) + \frac{\partial(\rho q_z)}{\partial z} \frac{\Delta z}{2} \right] \Delta x \Delta y
 \end{aligned}$$

To derive a general equation applicable to both non-homogeneous and isotropic/anisotropic formations, we start with the fundamental principles of fluid mechanics and adapt them to accommodate spatial variations in hydraulic properties. Let's outline the steps:

Conservation of Mass (Continuity Equation): Begin with the conservation of mass equation, which represents the balance between inflow and outflow within a control volume. In vector form, the continuity equation for groundwater flow can be expressed as:

$$\nabla \cdot \mathbf{q} + (\partial S / \partial t) = 0$$

Where:

$\mathbf{q}$ is the Darcy velocity vector,

S is the aquifer storage coefficient,

$\nabla \cdot$ is the divergence operator,

$\partial/\partial t$ is the partial derivative with respect to time.

Generalized Darcy's Law: Modify Darcy's Law to accommodate spatial variability in hydraulic conductivity. Introduce a tensorial form of hydraulic conductivity K to account for anisotropy and heterogeneity:

$$q = -K \cdot \nabla h$$

Where K is the hydraulic conductivity tensor.

Substitution and Simplification: Substituting the generalized Darcy's Law into the continuity equation yields:

$$\nabla \cdot (-K \cdot \nabla h) + \partial S / \partial t = 0$$

This equation accounts for spatial variations in hydraulic conductivity and hydraulic head, allowing for the consideration of both non-homogeneous and anisotropic formations.

Additional Assumptions and Boundary Conditions: Apply additional assumptions and boundary conditions as necessary, considering the specific characteristics of the aquifer system. These may include assumptions of isotropy or anisotropy, homogeneity or heterogeneity, steady-state or transient flow conditions, and various types of boundary conditions.

Solution and Analysis: Solve the resulting partial differential equation numerically using appropriate methods such as finite difference, finite element, or finite volume techniques. Solutions provide insights into groundwater flow patterns, hydraulic gradients, and aquifer behavior in non-homogeneous and anisotropic formations.

By incorporating a tensorial form of hydraulic conductivity and accounting for spatial variations in hydraulic properties, the derived equation can provide a comprehensive framework for modeling groundwater flow in a wide range of geological settings, including non-homogeneous and isotropic/anisotropic formations.

3.17. 2-D Approximations in unconfined aquifers:

In case of unconfined aquifers, the general flow equation, can be simplified by using an approximation first suggested by Dupuit (1863). The Dupuit approximation is that the equipotential lines can be assumed to be vertical, making the streamlines horizontal, and the head gradients constant over the depth, in which case

$$\begin{aligned} q_x &= -K_{xx} \frac{\partial \phi}{\partial x} \approx -K_{xx} \frac{\partial h}{\partial x} \\ q_y &= -K_{yy} \frac{\partial \phi}{\partial y} \approx -K_{yy} \frac{\partial h}{\partial y} \end{aligned} \quad \dots(3.14)$$

where, h is the thickness of the saturated zone. The Dupuit approximation is equivalent to assuming that the vertical pressure distribution in the aquifer is hydrostatic, leading to two-dimensional flow in the horizontal plane.

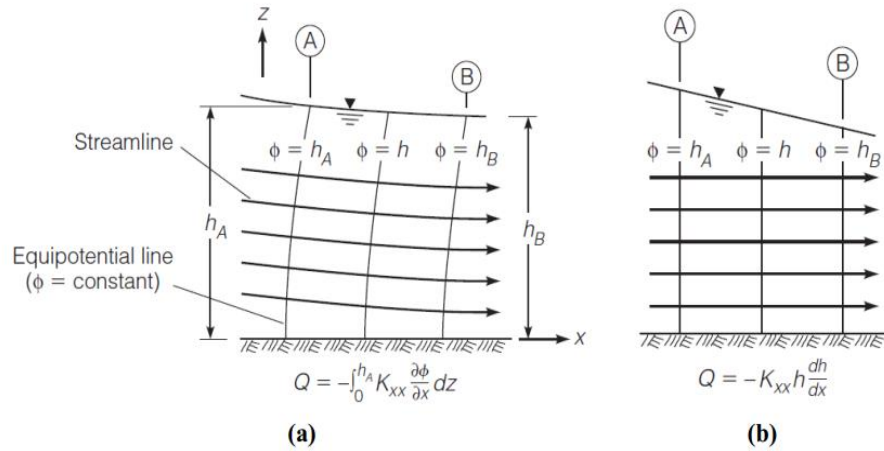


Figure 12(a) Actual case (b) Dupuit approximation (Chin, 2013)

Dupuit's approximation is a simplified analytical approach used to analyze steady-state groundwater flow in unconfined aquifers. It assumes that the flow is two-dimensional, horizontal, and occurs in an unconfined aquifer with a flat, horizontal free water table. The method simplifies the governing equations to make them solvable with relatively simple mathematical techniques. Here's an elaboration of the Dupuit's formula along with its equation:

Assumptions:

Steady-state flow: Dupuit's method assumes that the groundwater flow regime is steady, meaning that hydraulic heads do not change with time.

Horizontal flow: The method assumes that groundwater flow occurs predominantly in the horizontal direction, with negligible vertical flow components.

Homogeneous aquifer: Dupuit's approximation assumes that the aquifer properties (hydraulic conductivity, porosity) are uniform throughout the entire flow domain.

Constant recharge: It assumes a constant recharge rate across the aquifer, typically as a result of infiltration from precipitation or surface water bodies.

Governing Equation: Dupuit's formula is derived from the Laplace equation under the assumption of horizontal flow in a confined aquifer. It can be expressed as:

$$Q = -k \cdot b \cdot dh/dx$$

Where:

Q is the flow rate per unit width (m^2/s),

k is the hydraulic conductivity of the aquifer (m/s),

b is the thickness of the saturated zone (m),

dh/dx is the hydraulic gradient (m/m) along the direction of groundwater flow.

Interpretation: Dupuit's formula suggests that the flow rate per unit width is proportional to the product of hydraulic conductivity, thickness of the saturated zone, and the hydraulic gradient. It implies that groundwater flows from areas of higher hydraulic head to areas of lower hydraulic head, and the rate of flow is influenced by the permeability of the aquifer and the slope of the water table.

Applications: Dupuit's formula is particularly useful for preliminary analysis of groundwater flow in unconfined aquifers, where more complex numerical models may be unnecessary. It provides insights into flow patterns, groundwater velocities, and the influence of aquifer properties on flow behavior.

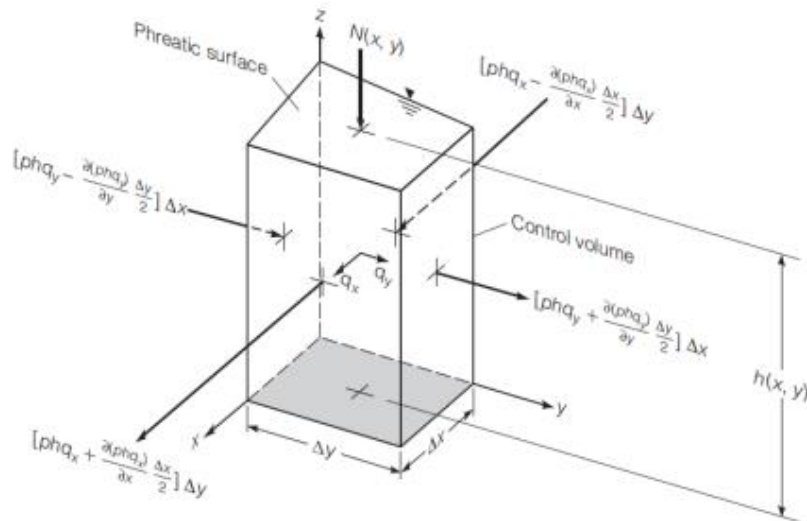


Figure 13 Control volume in an unconfined aquifer (Chin, 2013)

The general 2-D flow equation in an anisotropic and non-homogenous unconfined aquifer is given by,

$$\frac{\partial}{\partial x} \left(K_{xx} h \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left(K_{yy} h \frac{\partial h}{\partial y} \right) + N = S_y \frac{\partial h}{\partial t} \quad \text{.....(3.15)}$$

Limitations: Dupuit's approximation may not accurately represent flow in heterogeneous aquifers, areas with significant vertical flow components, or situations with variable recharge rates. It also neglects the effects of aquifer boundaries, such as rivers or impermeable layers, which can influence flow patterns. Therefore, it's essential to consider the assumptions and limitations of Dupuit's formula when applying it in practice

3.18. 2-D Approximations in confined aquifers:

Consider the case of a confined aquifer of thickness b illustrated,

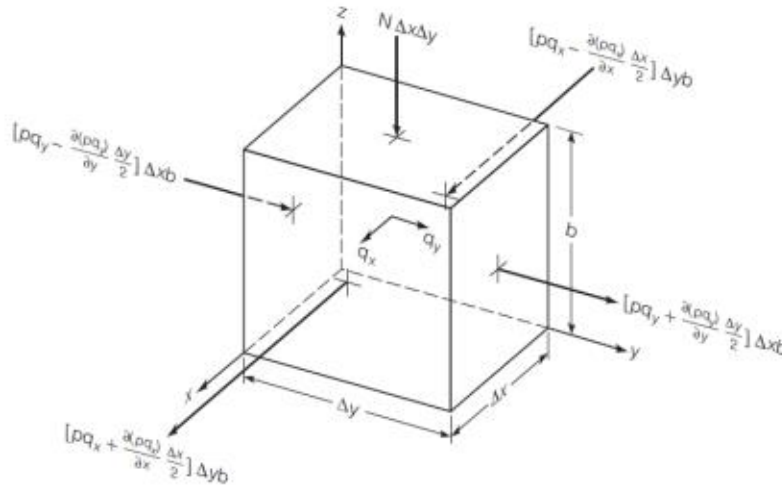


Figure 14 Control volume in a confined aquifer (Chin, 2013)

The general 2-D flow equation in an anisotropic and non-homogenous confined aquifer is given by

$$\frac{\partial}{\partial x} \left(K_{xx} \frac{\partial \phi}{\partial x} \right) + \frac{\partial}{\partial y} \left(K_{yy} \frac{\partial \phi}{\partial y} \right) + \frac{N}{b} = \frac{S}{b} \frac{\partial \phi}{\partial t} \quad \text{.....(3.16)}$$

3.19. Steady-State Well Hydraulics:

Steady-State Well Hydraulics: Steady-state well hydraulics refers to the condition where the rate of water withdrawal from a well remains constant over time, and the groundwater system reaches a stable equilibrium. This condition is often assumed in groundwater modeling and analysis to simplify calculations.

Pre-pumping Water Level: The pre-pumping water level, also known as the initial groundwater level or static water level, is the natural level of groundwater in an aquifer before any pumping activity occurs. It represents the equilibrium condition where the rate of recharge equals the rate of discharge.

Pumping Water Level: The pumping water level is the groundwater level observed in a well or pumping borehole during pumping. It is typically lower than the pre-pumping water level due to the extraction of water from the aquifer.

Drawdown: Drawdown is the decrease in groundwater level observed in a well or observation point as a result of pumping. It represents the difference between the pre-pumping water level and the pumping water level.

Cone of Depression: The cone of depression is the conical-shaped depression formed in the water table or potentiometric surface around a pumping well due to drawdown. It represents the extent of influence of the pumping well on the surrounding groundwater flow regime.

Radius of Influence: The radius of influence, also known as the capture radius or radius of influence, is the distance from a pumping well within which the drawdown caused by pumping is significant. It defines the area from which the pumping well extracts groundwater.

Unsteady Flow: Unsteady flow refers to groundwater flow conditions where hydraulic head or groundwater levels change with time. This typically occurs during pumping or recharge events, when the groundwater system is not in equilibrium.

Steady Flow: Steady flow refers to groundwater flow conditions where hydraulic head or groundwater levels remain constant over time. This occurs when the rate of water extraction or recharge remains constant, and the groundwater system reaches a stable equilibrium.

Recuperation or Recovery: Recuperation or recovery is the process by which groundwater levels return to their pre-pumping levels after pumping ceases. It represents the natural recharge of the aquifer and the restoration of groundwater levels following pumping activity.

Understanding these terms is essential for assessing groundwater resources, designing well systems, and managing groundwater extraction sustainably. They provide insights into the behavior of aquifers and the effects of pumping on groundwater levels and flow patterns.

3.20. Well in a confined aquifer:

The general 2-D flow equation in an anisotropic and non-homogenous confined aquifer is given by

$$\frac{\partial}{\partial x} \left(K_{xx} \frac{\partial \phi}{\partial x} \right) + \frac{\partial}{\partial y} \left(K_{yy} \frac{\partial \phi}{\partial y} \right) + \frac{N}{b} = \frac{S}{b} \frac{\partial \phi}{\partial t} \quad \text{.....(3.17)}$$

When the effective hydraulic conductivity of the aquifer is homogeneous as well as isotropic in the horizontal plane become,

$$K \frac{\partial^2 \phi}{\partial x^2} + K \frac{\partial^2 \phi}{\partial y^2} + \frac{N}{b} = \frac{S}{b} \frac{\partial \phi}{\partial t}$$

or, $\frac{\partial^2 \phi}{\partial x^2} + \frac{\partial^2 \phi}{\partial y^2} + \frac{N}{K b} = \frac{S}{K b} \frac{\partial \phi}{\partial t}$ (3.18)

In polar coordinate system,

$$\frac{\partial^2 \phi}{\partial r^2} + \frac{1}{r} \frac{\partial \phi}{\partial r} + \frac{1}{r^2} \frac{\partial^2 \phi}{\partial \theta^2} + \frac{N}{K b} = \frac{S}{K b} \frac{\partial \phi}{\partial t}$$

.....(3.19)

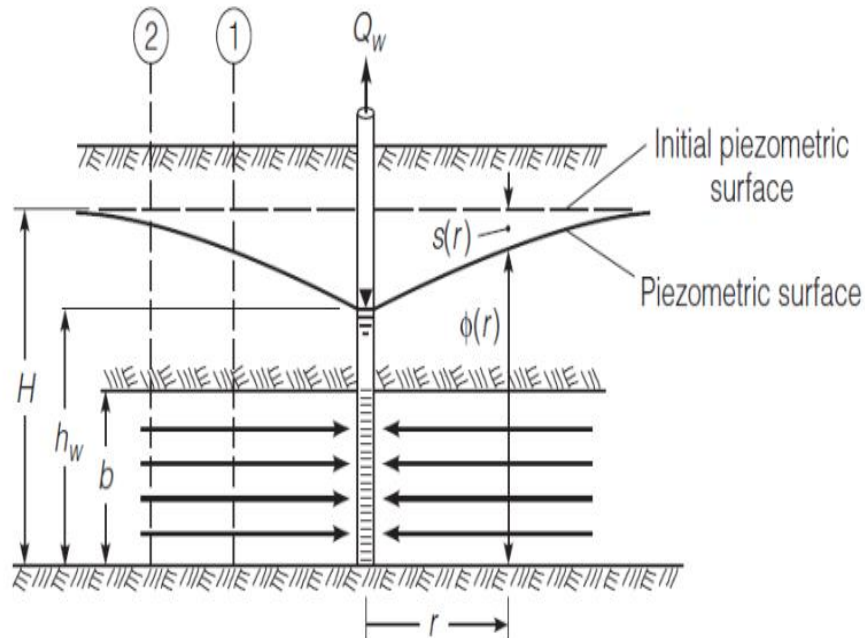


Figure 15 Fully penetrating well in a confined aquifer (Chin, 2013)

The Thiem equation is a fundamental solution for steady-state radial flow to a well in a confined aquifer. It is derived from Dupuit's equation under certain assumptions and boundary conditions. Let's establish the Thiem equation for a general 2-D flow equation in an anisotropic and non-homogeneous confined aquifer from Dupuit's equation:

Dupuit Equation: Dupuit's equation for radial flow to a well in a confined aquifer can be expressed as:

$$Q = -2\pi T \cdot h \cdot (dh/dr)$$

Where:

Q is the discharge rate from the well (volume per unit time),

T is the transmissivity of the aquifer (length squared per time),

h is the hydraulic head (length) at radial distance r from the well,

(dh/dr) is the hydraulic gradient (dimensionless) with respect to radial distance.

Thiem Equation: The Thiem equation is derived by assuming steady-state radial flow to a well in a confined aquifer. It simplifies Dupuit's equation under the assumption of horizontal flow and uniform properties:

$$Q = 2\pi T (h_1 - h_2) / \ln(r_2/r_1)$$

Where:

Q is the discharge rate from the well (volume per unit time),

T is the transmissivity of the aquifer (length squared per time),

h_1 is the hydraulic head (length) at radial distance r_1 from the well,

h_2 is the hydraulic head (length) at radial distance r_2 from the well,

$\ln(r_2/r_1)$ is the natural logarithm of the ratio of radial distances.

Extension to Anisotropic and Non-Homogeneous Aquifer: To extend the Thiem equation to an anisotropic and non-homogeneous aquifer, we replace the transmissivity T with the appropriate tensorial form of transmissivity T , which accounts for the anisotropy of the aquifer. Additionally, hydraulic heads h_1 and h_2 and radial distances r_1 and r_2 should be considered with appropriate directional components to reflect the anisotropic nature of the aquifer.

Boundary Conditions: Boundary conditions for the Thiem equation typically include specifying the hydraulic head at the well (pumping rate), as well as boundary conditions at the outer boundary of the aquifer system (e.g., no flow or specified head).

By establishing the Thiem equation from Dupuit's equation and extending it to an anisotropic and non-homogeneous confined aquifer, we can analyze radial flow to a well under more complex hydrogeological conditions. This equation provides valuable insights into well performance and groundwater extraction in real-world aquifer systems.

3.21. Well in an unconfined aquifer:

In an unconfined aquifer, the distribution of piezometric head for 2-D flows by Dupuit approximation, which can be simplified as follows considering an isotropic and homogenous state of the aquifer.

The general 2-D flow equation in an anisotropic and non-homogenous confined aquifer is given by,

$$\frac{\partial}{\partial x} \left(K_{xx} h \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left(K_{yy} h \frac{\partial h}{\partial y} \right) + N = S_y \frac{\partial h}{\partial t} \quad \dots(3.20)$$

In cases where the hydraulic conductivity of the aquifer is homogeneous, describes the distribution of saturated thickness h

$$K_{xx} \frac{\partial}{\partial x} \left(h \frac{\partial h}{\partial x} \right) + K_{yy} \frac{\partial}{\partial y} \left(h \frac{\partial h}{\partial y} \right) + N = S_y \frac{\partial h}{\partial t} \quad \dots(3.21)$$

Therefore, this simplifies to,

$$K_{xx} \frac{\partial^2 h^2}{\partial x^2} + K_{yy} \frac{\partial^2 h^2}{\partial y^2} + 2N = 2S_y \frac{\partial h}{\partial t}$$

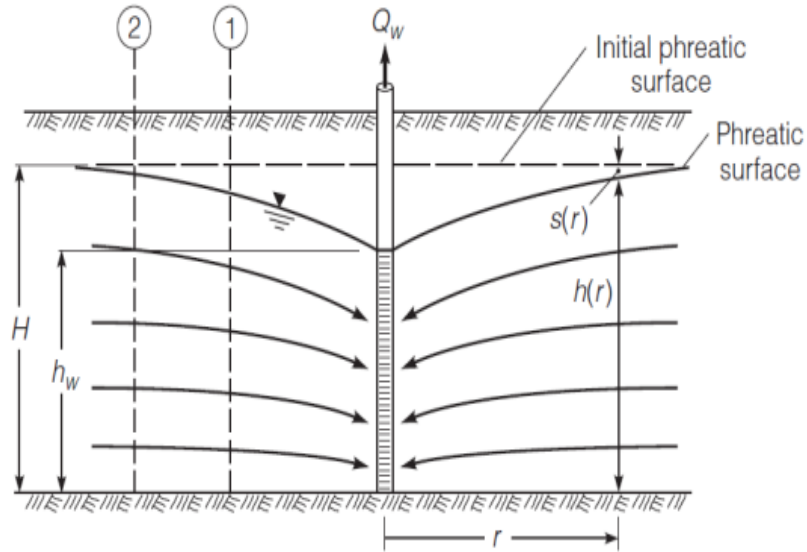


Figure 16 Fully penetrating well in an unconfined aquifer (Chin, 2013)

The expression for the head distribution induced by a fully penetrating pumping well in an unconfined aquifer is given by:

$$h(r) = \sqrt{h_0^2 + \frac{Q}{4\pi T} W(u)} \quad \dots(3.22)$$

Where:

$h(r)$ is the hydraulic head at radial distance r from the pumping well,

h_0 is the initial hydraulic head (before pumping begins),

Q is the pumping rate (volume per unit time),

T is the transmissivity of the aquifer (length squared per time),

$W(u)$ is the well function, which depends on the dimensionless variable u ,

u is defined as $u = (r^2 S) / (4 T t)$

S is the storativity (dimensionless), representing the storage coefficient of the aquifer,

t is the time since pumping began.

This expression represents the distribution of hydraulic head around the pumping well in the unconfined aquifer. It shows how the hydraulic head varies with radial distance from the well and over time as pumping continues. The term inside the square root represents the drawdown induced by pumping, and the initial hydraulic head h_0 is added to this drawdown to obtain the total head distribution.

boundary condition, the head distribution is given by,

$$h(r) = \sqrt{h_w^2 + (H - h_w)^2 \frac{\ln(\frac{r}{r_w})}{\ln(\frac{R}{r_w})}} \quad \dots(3.23)$$

3.22. Well in a leaky confined aquifer:

In isotropic and homogeneous semiconfined (leaky) aquifers, the distribution of piezometric head for 2-D flows is given by Eq.

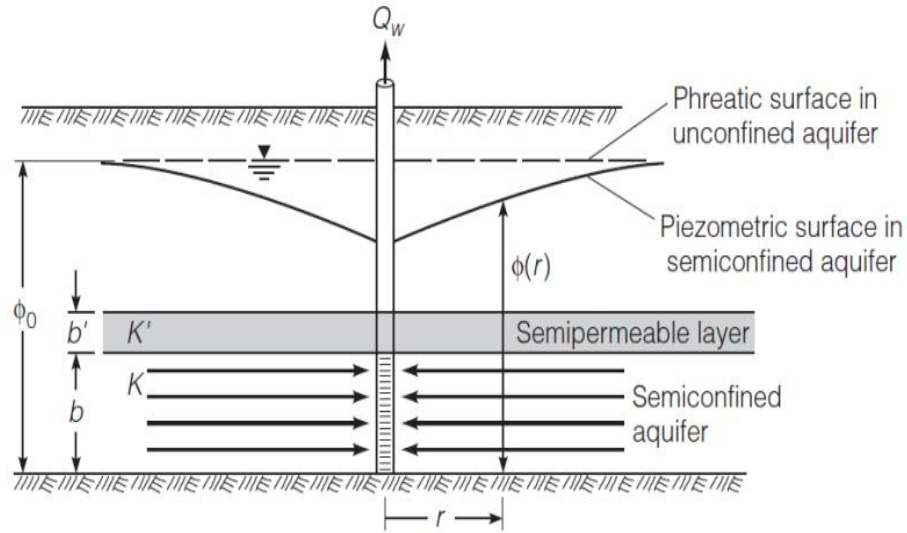


Figure 17 Fully penetrating well in a semi-confined aquifer (Chin, 2013)

If a well fully penetrates the semiconfined aquifer and withdraws water at a constant rate uniformly over the aquifer thickness,

$$\frac{d^2\phi}{dr^2} + \frac{1}{r} \frac{d\phi}{dr} + \frac{N}{Kb} = 0 \quad \text{.....(3.24)}$$

Leakage into the semiconfined aquifer can be calculated by applying Darcy's law across the semipermeable layer, in which case,

$$N = -K' \frac{\phi - \phi_0}{b'} \quad (3.25)$$

the piezometric head above the semipermeable layer, assumed to be constant.

Bessel's equation is a second-order ordinary differential equation that arises in many areas of mathematical physics, including problems involving wave propagation, heat conduction, and fluid dynamics. It is named after the German mathematician Friedrich Bessel, who studied these equations extensively in the early 19th century.

The standard form of Bessel's equation of order ν is:

$$x^2 \frac{d^2 y}{dx^2} + x \frac{dy}{dx} + (x^2 - \nu^2) y = 0 \quad \text{.....(3.26)}$$

Where:

y is the unknown function to be solved for,

x is the independent variable,

dy/dx and d^2y/dx^2 are the first and second derivatives of y with respect to x , respectively, and

ν is a constant parameter known as the order of the Bessel equation.

The solutions to Bessel's equation are called Bessel functions, denoted as $J_\nu(x)$ or $Y_\nu(x)$, depending on whether they are regular ($J_\nu(x)$) or irregular ($Y_\nu(x)$) at the origin. The behavior of these functions depends on the order ν .

Bessel functions are widely used in various fields of physics and engineering, particularly in problems with circular or cylindrical symmetry, such as electromagnetic wave propagation in circular waveguides, heat conduction in cylindrical objects, and fluid flow in cylindrical pipes.

Bessel's equation and its solutions play a crucial role in understanding the behavior of physical phenomena governed by cylindrical symmetry, providing elegant mathematical tools to solve such problems.

In this case, there are two aquifer properties: transmissivity, and leakage factor. These aquifer properties can be estimated using drawdown measurements at two locations.

Denoting the drawdown measurements by s ,

$$s(r) = \frac{Q_w}{2\pi T} \frac{K_0\left(\frac{r}{\lambda}\right)}{\left(\frac{r_w}{\lambda}\right) K_1\left(\frac{r_w}{\lambda}\right)} \quad \dots(3.27)$$

Eq. (3.27) can be used to estimate the leakage factor from the drawdown measurements. This leakage factor can subsequently be used to calculate the transmissivity of the semiconfined aquifer.

The derived leakage factor can be used to classify the **leakiness** of the semi-confining layer using the classification system in that was originally suggested by **Sen (1995)**. Aquifers with leakage factors exceeding 10,000 m can be treated as confined aquifers. The equations derived in this section assume that the flow is horizontal in the aquifer and vertical in the semi-confining (leaky) layer. Errors introduced by these assumptions are generally less than 5%, provided that the hydraulic conductivity of the aquifer is more than two orders of magnitude greater than that of the semi-confining layer (**Neuman and Witherspoon, 1969**).

3.23. Theis Method of Solution:

The drawdown is given by

The Theis method is a mathematical solution used in hydrogeology to analyze the behavior of groundwater flow to a pumping well in an unconfined aquifer. It was developed by Joseph H. Theis in 1935 and is widely used for estimating drawdown (the lowering of the water table) in aquifers due to pumping.

The Theis equation describes the drawdown of groundwater in response to pumping from a well in an unconfined aquifer. It is derived from the principles of groundwater flow and assumes radial flow towards the pumping well. The basic equation is:

$$s(u) = \left(\frac{Q_w}{4\pi T} \right) W(u)$$

The *Boltzman variable* can be rearranged and written as

$$\frac{r^2}{t} = \left(\frac{4T}{S} \right) u \quad \dots(3.28)$$

Where:

s is the drawdown at the pumping well,

Q is the pumping rate (in volume per unit time),

T is the transmissivity of the aquifer,

$W(u)$ is the well function, which is a mathematical function that depends on a dimensionless variable uu that represents time and distance.

The Theis solution assumes certain simplifying assumptions, including steady-state pumping, homogeneous and isotropic aquifer properties, and absence of boundaries that might influence flow. Despite these limitations, the Theis method is widely used because it provides a reasonably accurate estimation of drawdown in many practical situations.

The Theis method is often used in groundwater modeling and aquifer pumping tests to estimate parameters such as transmissivity and hydraulic conductivity of the aquifer. It's an essential tool for hydrogeologists and engineers involved in groundwater management and water resource development.

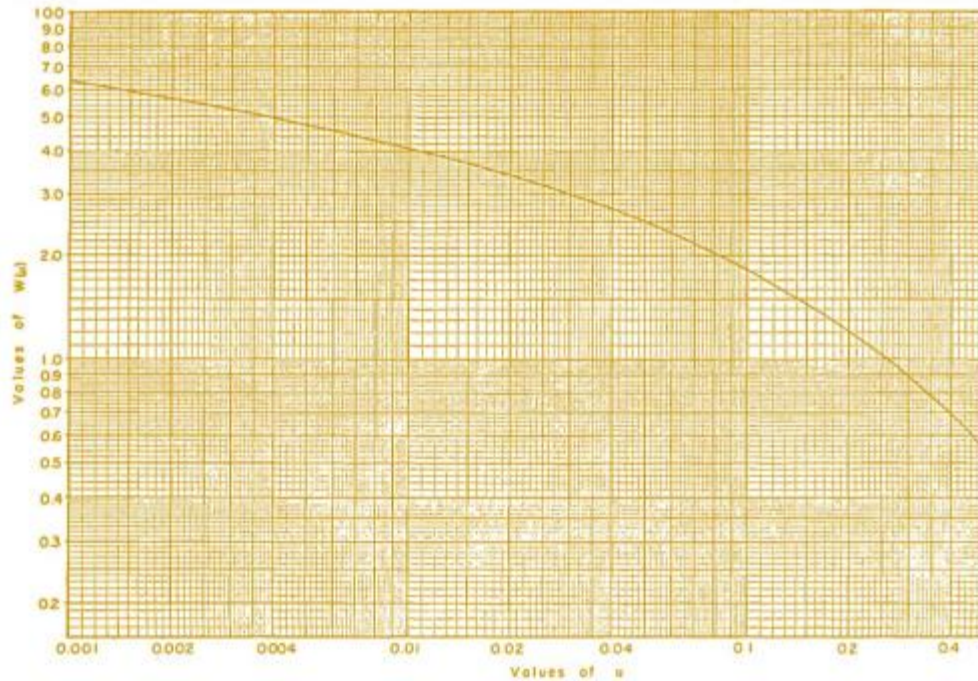


Figure 18Plot of $W(u)$ versus u known as a type curve

Table 5.3.: *Values of $W(u)$ for the different values of u*

u	1.0	2.0	3.0	4.0	5.0	6.0	7.0	8.0	9.0
$\times 1$	0.2194	0.1489	0.0130	0.0038	0.0011	0.0004	0.0001	0.0000	0.0000
$\times 10^{-1}$	1.8229	1.2227	0.9057	0.7024	0.5598	0.4544	0.3738	0.3106	0.2602
$\times 10^{-2}$	4.0379	3.3547	2.9591	2.6813	2.4679	2.2953	2.1508	2.0269	1.9187
$\times 10^{-3}$	6.3315	5.6394	5.2349	4.9482	4.7261	4.5448	4.3916	4.2591	4.1423
$\times 10^{-4}$	8.6332	7.9401	7.5348	7.2472	7.0242	6.8420	6.6879	6.5545	6.4368
$\times 10^{-5}$	10.9357	10.2428	9.8372	9.5494	9.3264	9.1440	8.9899	8.8564	8.7386

Table 1Values of $W(u)$ & u

3.24. Type curve:

In hydrogeology, a type curve is a standardized graphical representation used to analyze and interpret data from pumping tests conducted in aquifers. The type curve method is often employed in conjunction with solutions like Theis or Cooper-Jacob to estimate aquifer parameters.

It works:

Data Collection:

During a pumping test, data such as drawdown (change in water level) at the pumping well versus time is collected.

3.26.1 Normalization:

The drawdown data is then normalized using appropriate dimensionless parameters. For example, in the case of Theis solution, the drawdown is divided by the pumping rate and the aquifer transmissivity to normalize the data.

3.26.2. Plotting:

The normalized data is plotted on a graph. On the same graph, a pre-calculated type curve is also plotted. This type curve represents the expected behavior of drawdown versus time for a specific set of aquifer parameters.

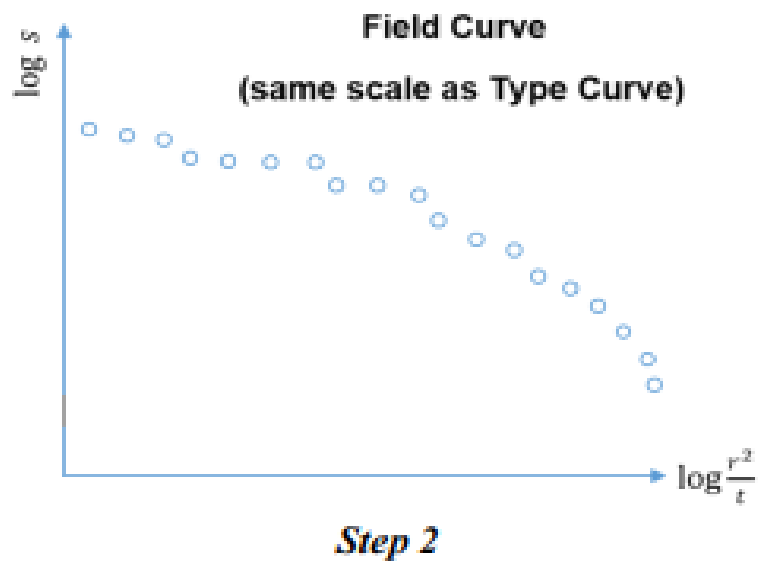
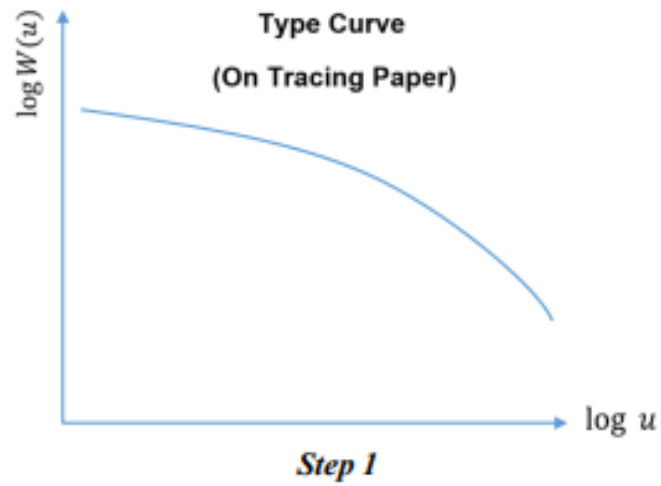
3.26.3. Comparison:

By comparing the normalized field data to the type curve, hydrogeologists can estimate aquifer parameters such as transmissivity or hydraulic conductivity. If the field data matches the type curve closely, it suggests that the aquifer parameters are similar to those represented by the type curve.

3.26.4. Iterative Process:

Often, the matching process involves adjusting the aquifer parameters until the field data closely matches the type curve. This iterative process helps in refining the estimates of aquifer properties.

Type curves provide a convenient way to analyze pumping test data and estimate aquifer parameters without needing complex mathematical calculations for each dataset. They are widely used in hydrogeology for their simplicity and efficiency in interpreting pumping test results.



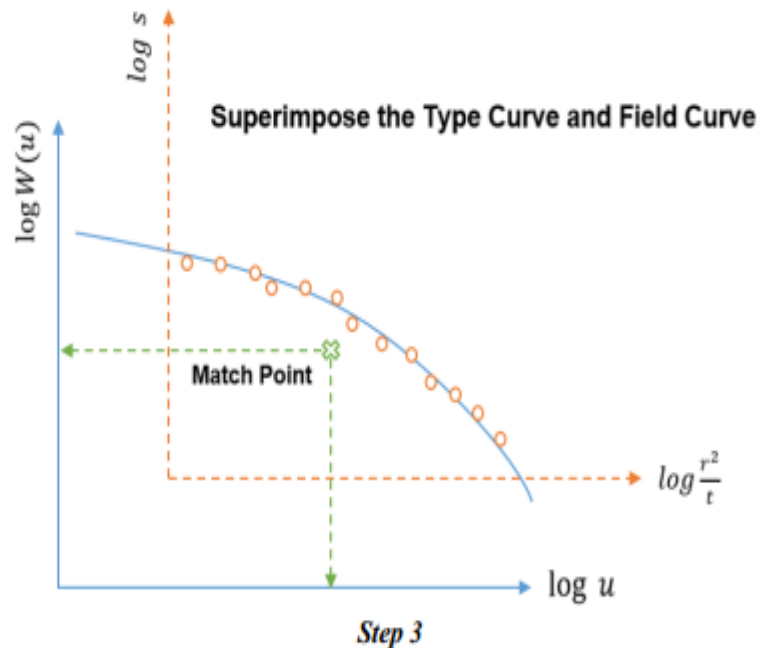


Figure 19 Theis Method for solution of the non-equilibrium equations

3.25. Cooper-Jacob Method of Solution:

The Cooper-Jacob method provides a systematic approach to solving the nonlinear flow equations governing groundwater movement. By discretizing the domain and iteratively solving the linearized equations, it allows for the approximation of complex groundwater flow scenarios. This method is particularly valuable in practical groundwater management and engineering applications where accurate predictions of groundwater behavior are essential.

Top of Form

The groundwater flow equation, typically expressed as Darcy's law combined with the continuity equation, represents the fundamental equation in groundwater hydrodynamics:

$$\nabla \cdot (K \nabla h) = S$$

Where:

h is the hydraulic head,

K is the hydraulic conductivity tensor,

S is the source/sink term.

Method Steps:

Discretization: The domain is divided into discrete elements (e.g., finite difference cells or finite elements) to approximate the continuous flow equations.

Linearization: Nonlinear terms in the equations are linearized using an iterative approach.

Iteration: The linearized equations are solved iteratively until convergence is achieved.

Update: Hydraulic head values are updated based on the iterative solution.

Convergence Criterion: The iteration continues until the difference between successive iterations falls below a predefined tolerance level.

The utility of the series expression for $W(u)$ given by Eq. is that for small values of (u) is,

$$W(u) = -0.5772157 - \ln u$$

$$\text{Therefore, } s(u) = \frac{Q_w}{4\pi T} [-0.5772157 - \ln u]$$

$$\text{or, } s(u) = \frac{Q_w}{4\pi T} \left[-0.5772157 - \ln \left(\frac{r^2 S}{4Tt} \right) \right]$$

$$\text{or, } s(u) = \frac{2.303 Q_w}{4\pi T} \log \left(\frac{2.25 T t}{r^2 S} \right)$$

.....(3.29)

This approximation was first applied to the analysis of groundwater flow by **Cooper and Jacob (1946)** and is commonly referred to as the **Cooper-Jacob approximation**. For $u < 0.03$ and the error is less than 1%, and for $u = 0.1$ the error is about 5%. Typical applications of the **Cooper-Jacob approximation** are limited to $u < 0.05$.

Application of the **Cooper-Jacob approximation** in cases where there is a constant pumping rate, usually consists of the following steps:

Step 1: Plot s versus (r^2/t) from measurements taken at one or more monitoring wells; different values of r can be used.

Step 2: Fit a straight line to the data corresponding to $u < 0.05$.

Step 3: Determine a and b from the equation of the best-fit line.

Step 4: Solve for T and S using Eq. (1.116a.) and Eq. (1.116b.).

Application:

The Cooper-Jacob method in groundwater hydrodynamics finds application in various groundwater modeling scenarios, including:

Aquifer Simulation: Modeling flow and transport in aquifers with heterogeneous properties.

Well Hydraulics: Analyzing the behavior of pumping wells and their impact on groundwater levels.

Contaminant Transport: Simulating the movement of contaminants through groundwater systems.

Managed Aquifer Recharge: Optimizing strategies for injecting water into aquifers for storage and later recovery.

3.26. Time-Drawdown Method:

Time is plotted along the logarithmic X-axis and drawdown is plotted along the linear Y-axis. Transmissivity and Storativity are calculated as follows:

$$T = \frac{2.303 Q_w}{4 \pi \Delta s} \log \left(\frac{t_2}{t_1} \right)$$

$$S = \frac{2.25 T t_0}{r^2}$$

where, Δs is the amount of drawdown

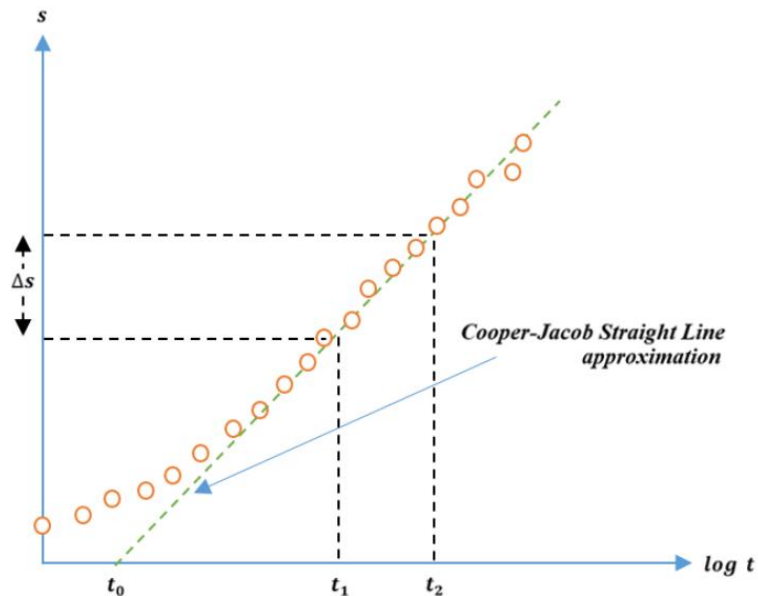


Figure 20 Cooper-Jacob Approximation using Time-Drawdown Method

3.27. Distance-Drawdown Method:

The observation well distance is plotted along the logarithmic X-axis and drawdown is plotted along the linear Y-axis. Transmissivity and Storativity are calculated as follow

$$T = \frac{2.303 Q_w}{4 \pi \Delta s} \log \left(\frac{r_2}{r_1} \right)$$

$$S = \frac{2.25 T t}{r_0^2} \quad \dots(3.30)$$

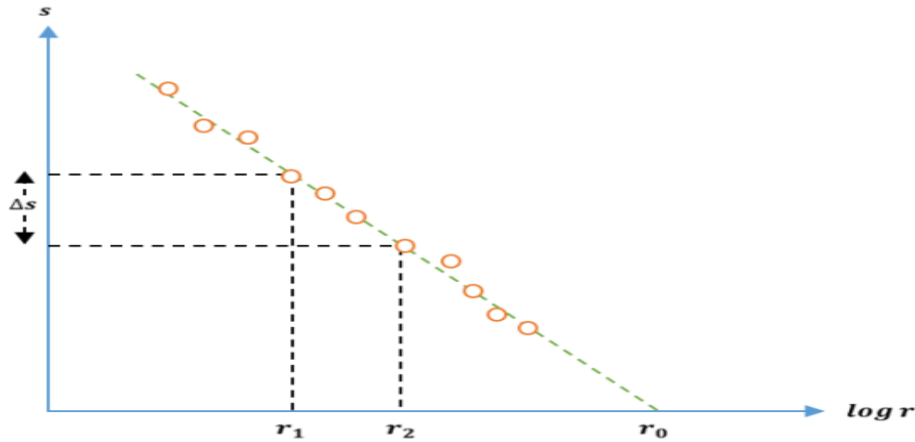


Figure 21 Cooper-Jacob Approximation using Distance-Drawdown Method

Therefore, under the restriction that $S \ll 2H$, the governing equation and associated initial and boundary conditions are exactly the same as the governing equation and associated initial and boundary conditions that describe unsteady drawdown in a confined aquifer, with the only exception being that the storage coefficient, S in the confined aquifer case is replaced by the specific yield, in the unconfined aquifer case.

The aquifer response in pumping tests carried out in unconfined aquifers frequently varies for small and large times. More specifically, compared to large-time data, the storage coefficient that is derived from small-time data may be substantially smaller. The usual explanation for this phenomena is that gravity drains from the pores above the water table does not rise instantly, and the first water release is solely connected to the compressibility of the aquifer's material and water. Gradually, gravity drainage starts to contribute much to aquifer recharging, resulting in a yield that is noticeably delayed. This is seen in aquifer materials with finer grain sizes, and it could take weeks for them to empty completely (Gorelick et al., 1993).

Neuman (1975; 1973; 1972) derived the following solution for a pumping well in an unconfined aquifer with delayed yield,

$$s = \frac{Q_w}{4 \pi T} W(u, u_y, \Gamma)$$

where, $W(u, u_y, \Gamma)$ is called the *unconfined well function*

Neuman's well function equation, also known as the Theis well function, is a fundamental solution derived by Charles Vernon Theis in 1935. It's used extensively in hydrogeology to describe the drawdown at a pumping well in an unconfined or confined aquifer. Neuman further extended and refined Theis's work, and his formulation is often referred to as Neuman's well function.

The equation for Neuman's well function, denoted as $W(u)$, is expressed as:

$$W(u) = \int_u^\infty \frac{e^{-s}}{s} ds$$

Where:

u is a dimensionless parameter related to time and aquifer properties, defined as $u = (r^2 S) / (4 T t)$,

r is the radial distance from the pumping well to the observation point,

S is the storage coefficient of the aquifer,

T is the transmissivity of the aquifer, and

t is time.

Neuman's well function provides a solution to the Theis equation, which describes transient groundwater flow in a confined or unconfined aquifer under steady-state pumping conditions. It represents the integral of the exponential integral function, which accounts for the effects of pumping on drawdown over time and space.

The well function $W(u)$ is typically tabulated or calculated numerically for various values of u . It serves as a fundamental tool for analyzing pumping tests and designing pumping schemes in hydrogeological studies. By using Neuman's well function, hydrogeologists can estimate drawdown at observation points for different pumping rates and aquifer properties without needing to solve the full differential equation governing groundwater flow.

Neuman's well function remains a cornerstone of groundwater hydrogeology and is widely used in practice to analyze and interpret data from pumping tests, groundwater modeling studies, and aquifer characterization efforts.

The equation derived by Neuman typically takes the form of a type curve, which relates dimensionless parameters representing aquifer properties, pumping rate, and time to the dimensionless drawdown at the pumping well. The general form of the equation can be expressed as:

$$s = W(u) - 2 \cdot (Q / 4\pi T) \cdot t \quad \dots\dots\dots (3.31)$$

Where:

s is the drawdown at the pumping well,

$W(u)$ is a dimensionless function representing the delayed yield effect,

Q is the pumping rate,

T is the transmissivity of the aquifer,

t is time, and

u is a dimensionless parameter related to time and aquifer properties.

The function $W(u)$ encapsulates the delayed drainage effect and depends on the dimensionless parameter uu , which accounts for the time-dependent behavior of the aquifer's response to pumping.

This solution allows hydrogeologists and engineers to estimate drawdown at a pumping well for given aquifer conditions and pumping rates without needing to solve the full differential equation governing groundwater flow. The type curve approach provides a convenient tool for analyzing pumping tests and designing pumping schemes in unconfined aquifers.

Neuman's contributions have had a profound impact on groundwater hydrology and have been instrumental in advancing the understanding of unconfined aquifer behavior under pumping stress. His work remains foundational in the field of hydrogeology and continues to be widely used and cited in groundwater studies.

Table 3 Values of $W(u)$ for unconfined aquifers

$\frac{1}{u}$	Γ								
	0.001	0.01	0.06	0.2	0.6	1.0	2.0	4.0	6.0
4.0×10^{-1}	0.0234	0.0240	0.0230	0.0217	0.0188	0.0170	0.0138	0.00933	0.00639
8.0×10^{-1}	0.144	0.140	0.131	0.119	0.0988	0.0849	0.0603	0.0317	0.0174
1.4×10^0	0.358	0.345	0.318	0.279	0.217	0.175	0.107	0.0445	0.0210
2.4×10^0	0.662	0.633	0.570	0.483	0.343	0.256	0.133	0.0476	0.0214
4.0×10^0	1.02	1.46	0.849	0.688	0.438	0.300	0.140	0.0478	0.0215
8.0×10^0	1.57	1.88	1.23	0.918	0.497	0.317	0.141	0.0478	0.0215
1.4×10^1	2.05	2.27	1.51	1.03	0.507	0.317	0.141	0.0478	0.0215
2.4×10^1	2.52	2.61	1.73	1.07	0.507	0.317	0.141	0.0478	0.0215
4.0×10^1	2.97	3.00	1.85	1.08	0.507	0.317	0.141	0.0478	0.0215
8.0×10^1	3.56	3.23	1.92	1.08	0.507	0.317	0.141	0.0478	0.0215
1.4×10^2	4.01	3.37	1.93	1.08	0.507	0.317	0.141	0.0478	0.0215
2.4×10^2	4.42	3.43	1.94	1.08	0.507	0.317	0.141	0.0478	0.0215
4.0×10^2	4.77	3.45	1.94	1.08	0.507	0.317	0.141	0.0478	0.0215
8.0×10^2	5.16	3.46	1.94	1.08	0.507	0.317	0.141	0.0478	0.0215

Table 2 Values of W(u) for unconfined aquifers

$\frac{1}{u_y}$	Γ								
	0.001	0.01	0.06	0.2	0.6	1.0	2.0	4.0	6.0
4.0×10^{-1}	5.62	3.46	1.96	1.13	0.589	0.421	0.248	0.149	0.113
8.0×10^{-1}	5.62	3.46	1.98	1.18	0.667	0.506	0.357	0.266	0.231
1.4×10^0	5.63	3.47	2.01	1.24	0.780	0.642	0.517	0.445	0.419
2.4×10^0	5.63	3.49	2.06	1.35	0.954	0.850	0.763	0.718	0.703
4.0×10^0	5.63	3.51	2.13	1.50	1.200	1.13	1.08	1.06	1.05
8.0×10^0	5.64	3.56	2.31	1.85	1.68	1.65	1.63	1.63	1.63
1.4×10^1	5.65	3.63	2.55	2.23	2.15	2.14	2.14	2.14	2.14
2.4×10^1	5.67	3.74	2.86	2.68	2.65	2.65	2.64	2.64	2.64
4.0×10^1	5.70	3.90	3.24	3.15	3.14	3.14	3.14	3.14	3.14
8.0×10^1	5.76	4.22	3.85	3.82	3.82	3.82	3.82	3.82	3.82
1.4×10^2	5.85	4.58	4.38	4.37	4.37	4.37	4.37	4.37	4.37
2.4×10^2	5.99	5.00	4.91	4.91	4.91	4.91	4.91	4.91	4.91
4.0×10^2	6.16	5.46	5.42	5.42	5.42	5.42	5.42	5.42	5.42
8.0×10^2	6.47	6.11	6.11	6.11	6.11	6.11	6.11	6.11	6.11
1.4×10^3	6.67	6.67	6.67	6.67	6.67	6.67	6.67	6.67	6.67
2.4×10^3	7.21	7.21	7.21	7.21	7.21	7.21	7.21	7.21	7.21
4.0×10^3	7.72	7.72	7.72	7.72	7.72	7.72	7.72	7.72	7.72
8.0×10^3	8.41	8.41	8.41	8.41	8.41	8.41	8.41	8.41	8.41
1.4×10^4	8.97	8.97	8.97	8.97	8.97	8.97	8.97	8.97	8.97

**Source: David A. Chin (2013)*

These **Type A** and **Type B curves** are tabulated in Tables and can be used to determine the aquifer parameters based on small-time and large-time pumping data using the relations

$$s = \frac{Q_w}{4\pi T} W(u, \Gamma), \text{ for small time}$$

$$s = \frac{Q_w}{4\pi T} W(u_y, \Gamma), \text{ for large time}$$

The application of above equation is illustrated in Fig. below, where **Type A curves** merge into the Theis curve shown on the left, and **Type B curves** merge into the Theis curve shown on the right.

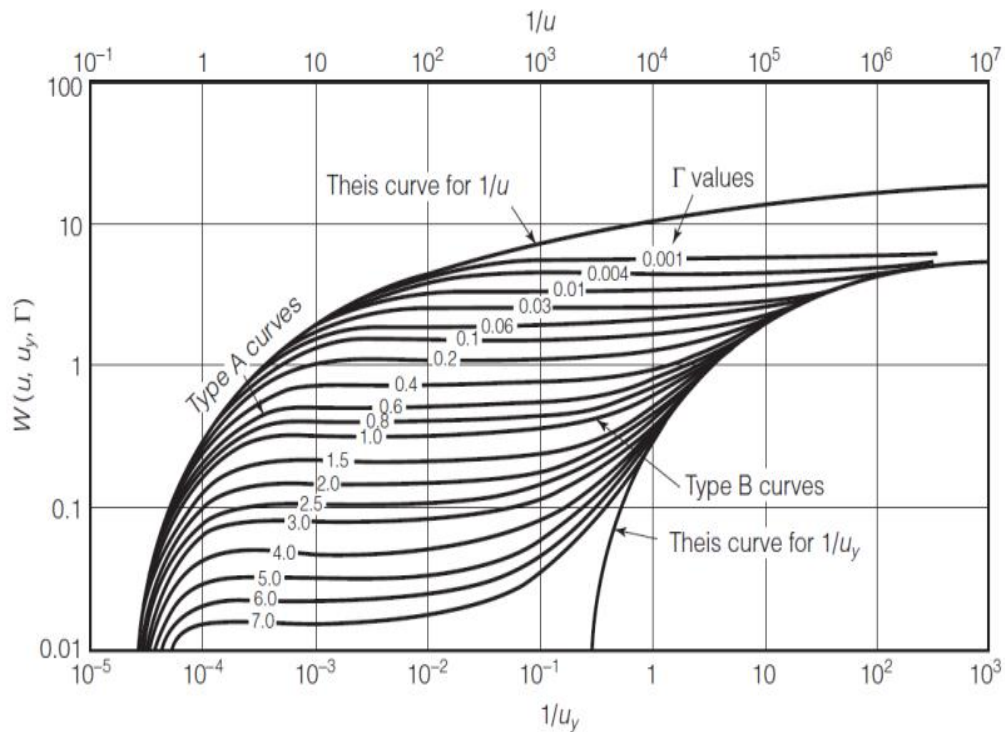


Figure 22 Unconfined aquifer well functions

In unconfined aquifers, the time-drawdown curve usually consists of three separate segments, as illustrates. The first part starts as soon as pumping starts, when water extracted from the aquifer is connected to the matrix's compressibility and subsurface water. These emissions are parameterized by the aquifer's storage coefficient. The second descent section depicts the transitional phase where the expansion of the Due to gravity drainage and the time-drawdown slope, the cone of depression shrinks. curve dips, indicating recharging. The time-drawdown curves in the third section follow the Theis solution with gravity drainage that is based on a given yield. The latter situation might originate from multiple minutes to several days following the start of pumping. Pump-drawdown information gathered over the previous.

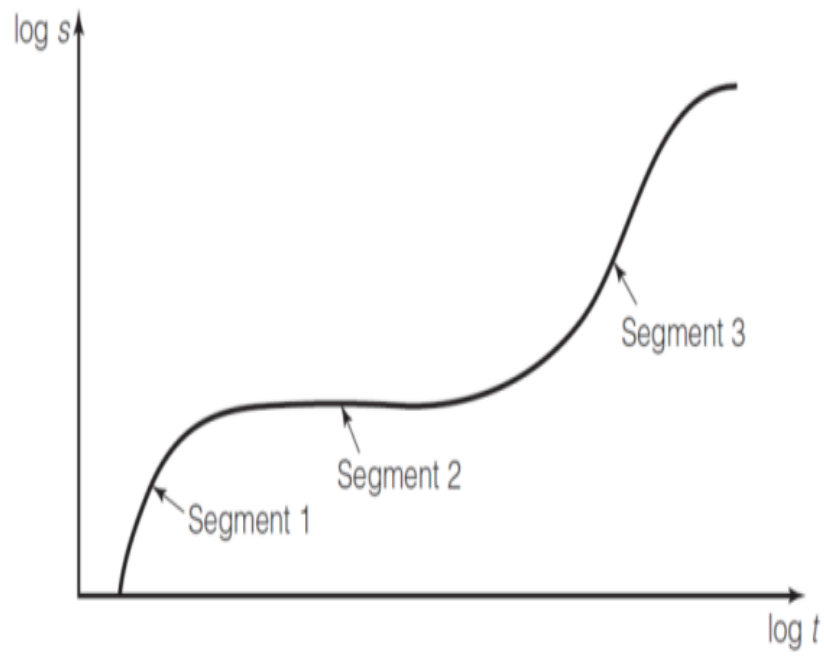
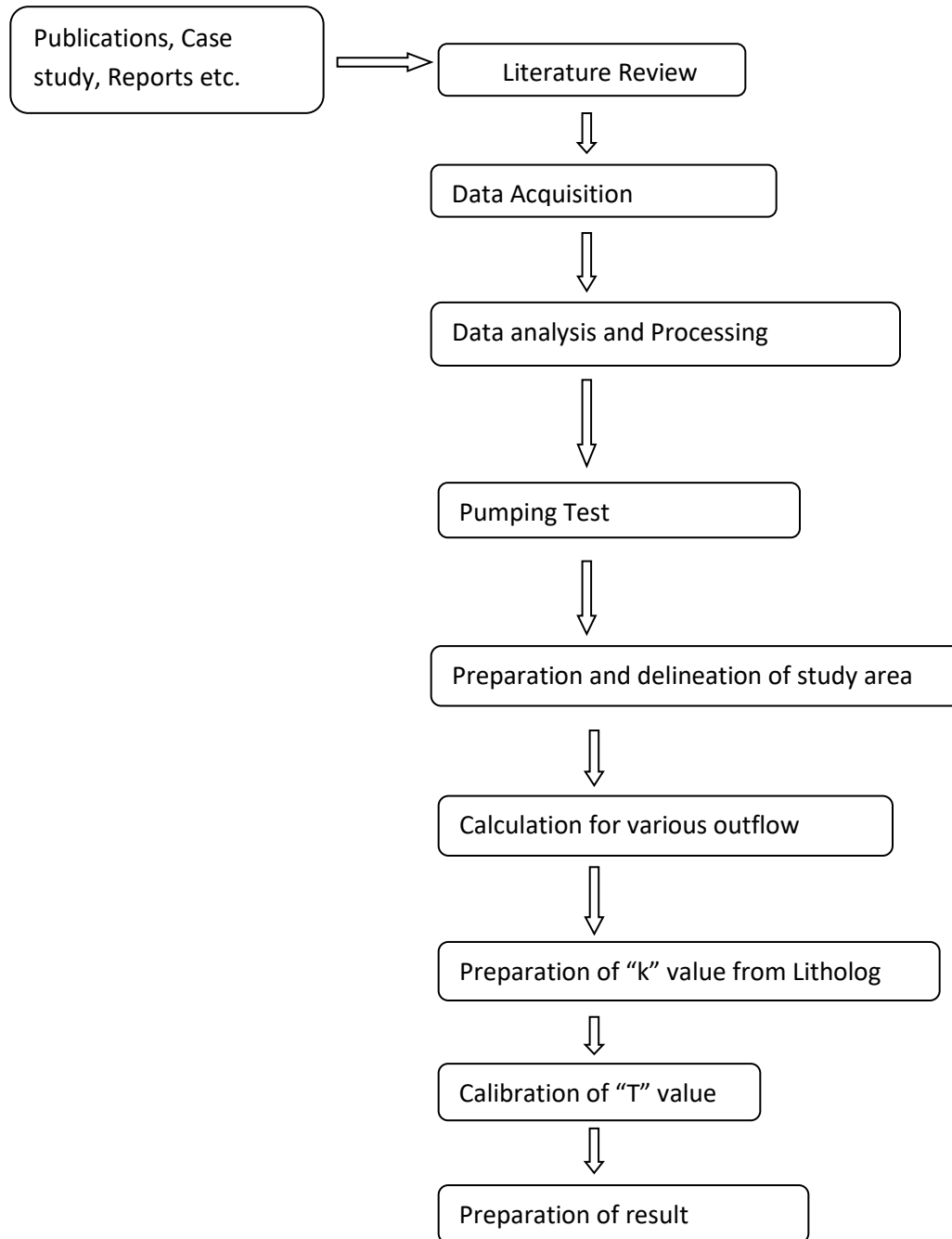


Figure 23 Three segments of time-drawdown curve in unconfined aquifers

CHAPTER – 4: METHODOLOGY

4.1. The overall research methodology adopted for the present work is depicted by the flowchart



4.2. Literature review & data acquisition and analyses:

At the initial stage, a comprehensive review of the literature already available in the form of scientific publications/ case studies/ review works/ reports/ book chapters etc., have been accomplished, to have an over view of the basics of the works that have already carried out related to the present research study.

Proceeding further, the pumping test data has been collected to determine and characterise the important hydraulic parameters of the aquifer for the study area. In addition to this, the piezometric surface data for the study area has also been obtained. A combination of the above data forms the secondary data set for this present study, and is crucial for the initial development of a groundwater model, which will actually approximate and conceptualise the groundwater flow of the study area to carry out future predictions with an acceptable range of accuracy. Moreover, the rainfall distribution, temperature, evapotranspiration etc. data have also been obtained for the study area for the complete simulation of the groundwater flow in the region.

4.3. Methodology of hydrological investigation:

The study aims to assess the viability of implementing a hydrogeological investigation and designing a sustainable production well to meet water demand for the water supply scheme considering intake at Jujuti for Zone-I (Hitta) and Zone-II (Khano) water Supply scheme for a total demand of $(820+596=1416 \text{ KLD} \approx 1500 \text{ KLD})$. This study focuses on the following aspects: reconnaissance survey, Vertical Electrical Sounding (VES) Survey, slim boring operations, pumping tests, recuperation test, pumping well discharge monitoring and calibration, and water quality testing. Special attention is given to addressing the concerns related to sand mining in the study area situated along the rain fed Damodar river. The field study has been carried out during the dry period from 14th December 2023 to 14th January 2024. The study begins with a detailed reconnaissance survey of the area to identify potential locations for the pumping well. This survey involves the collection of existing hydrological, geological and hydrogeological data, topographic maps, and satellite imageries to evaluate the suitability of the area for groundwater extraction. To obtain accurate subsurface information slim boring operations are conducted in five selected cross-sections. Fifteen (15) slim bores were drilled at selected intervals to determine the geological formations, lithology, and groundwater levels at different varying (up to around 45m to 50 m) depths at the study area. Vertical Electrical Sounding (VES) survey was also carried out to estimate overall aquifer properties in the study area. The data collected from the slim bores and VES survey provides valuable insights into the aquifer characteristics and helps to identify suitable locations for the pumping well. Accordingly, one main pumping well and three numbers of observation wells were installed at the **Melatala Ground** of **Jujuti** maintaining an equal depth of 45 m. A pumping test was conducted to estimate the sustainable yield of the aquifer. Water was pumped continuously from the installed main pumping well until the steady-state conditions were achieved. From this pumping well yield test, it was observed that the steady state condition was attained after 45 hours of continuous pumping from the main well. During the pumping test, groundwater levels in four wells were monitored continuously using temperature-level-conductivity (TLC) meters and discharge rates were monitored using an ultrasonic flow meter. It allows for the determination of the aquifer's characteristics

such as transmissivity and storage coefficient. Discharge from the main well was measured using an ultrasonic flowmeter and calibrated using the volumetric method. From these two measuring systems, a flow rate of 78 m³/h from the main well was accurately determined. It provides essential data for assessing the efficiency of the system. At the end of the pumping test, when the pump was stopped, immediately the recuperation test was started. After the pump shutdown, the groundwater level in the wells was monitored to determine the recovery time and estimate the transmissivity of the pumping well as well as observation wells. This groundwater level data was measured to determine the sustainable yield and ensure the long-term viability of the water supply of the radial collector well. Based on the hydrogeological investigation data, the sustainable yield was estimated and subsequently.

4.4. SURVEY AND INVESTIGATION:

Initially, a reconnaissance survey was done to examine the general character of the area to determine the most feasible preliminary water intake location, and this was followed by the collection of necessary information from different sources. The reconnaissance survey was conducted to cover the following aspects.

- Identification of study locations
- Assessment of field condition for the detailed hydrogeological investigation
- Study of topographical survey sheets,
- GPS data collection for GIS based map preparation.

4.5. LITHOLOGICAL SURVEY:

In this section, an important geophysical investigation has been carried out to ascertain the subsurface strata in the study location. Along the bank of River Damodar, slim bores of 50 mm diameter up to 45 m depth were targeted to drill in the selected locations. The borehole module of Starter 5.0 software has been used to visualise the borehole lithology created from drilling logs. The depth of the drilled slim bores ranges from 5 to 45 m. Lithologs obtained in **slim bores** to show vertical transitions for the present study along with their lateral persistence.

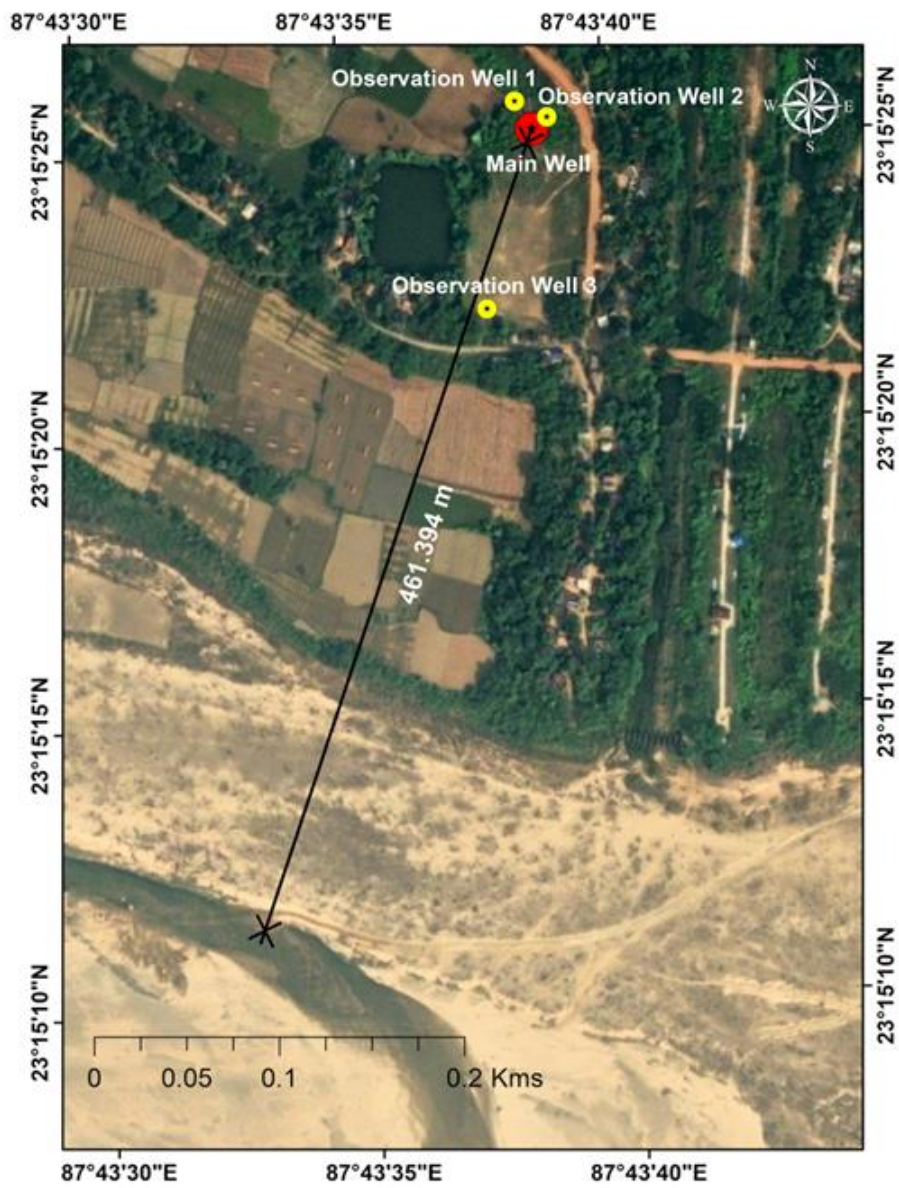


Figure 24 Location of the slim bores and pumping well/main well and three observation wells.



Photograph 1 Analysis of grain-size distribution in the Laboratory of the School of Water Resources Engineering, Jadavpur University

CHAPTER – 5: RESULT AND DISCUSSION

5.1. Determination of Hydraulic Parameters using Drawdown Data for Unconfined Aquifer:

The response of an unconfined aquifer to pumping is complicated. Once a cone of depression forms, it naturally decreases the aquifer thickness and transmissivity because the upper boundary of the aquifer is the water table. Also, the way in which water comes out of storage in the aquifer changes with time. At an early time, when the water from the main well is first pumped out, water is released from storage due to compression of the matrix and expansion of the water. This response is the same as Theis behaviour that applies to confined aquifers. Thus, early-time versus drawdown data, plotted on a semi-log graph paper, follow the Theis-type curve. The storativity values would be comparable to those for confined aquifers, 10^{-4} or 10^{-5} .

As pumping continues, water comes out from the slow gravity drainage of water from pores as the piezometric water table falls near the well. The pattern of drawdown depends on the vertical and horizontal hydraulic conductivity and the thickness of the aquifer. Once this delayed drainage begins, drawdown data starts to deviate from the Theis curve. The drawdown is less than expected, resembling the pattern of drawdown in a leaky aquifer. Eventually, the contribution of water comes when delayed drainage ceases. Flow in the aquifer is mainly radial, and the drawdown versus time data plot again shows a Theis-type curve pattern.

The specific yield (S_y) is the ratio of the volume of water that drains from a rock or sediment by gravity to the volume of the rock or soil. The storativity of the aquifer now is the same as the specific yield.

Calculation of drawdown by correcting the estimate for a confined aquifer was done using equation 1.

$$s' = s - \frac{s^2}{2b}$$

where, s is the drawdown for the unconfined aquifer and s' is the drawdown for the equivalent confined aquifer.

The following methodology has been adopted to determine the hydraulic parameters using the Distance Drawdown Method by taking the following steps:

- Calculate the corrected drawdown using equation (1)
- Plot corrected drawdowns against r on semi-log graph paper
- Approximate the curves versus $\log(t)$ with a straight line
- Determine the hydraulic parameters using the following equations (2-4):

a) Transmissivity (T)

$$T = \frac{2.3Q}{4\pi\Delta s} \quad \text{.....(5.1)}$$

b) Specific Yield /Storativity (S)

$$S_y = \frac{2.25Tt_{0y}}{r^2} \quad \text{.....(5.2)}$$

c) Radius of Influence

$$Q = \frac{2\pi T s_d}{\ln \ln \left(\frac{R}{r} \right)} \quad \text{.....(5.3)}$$

For observation wells 1, 2 and 3, values of drawdown with varying time are plotted on semi-log graph paper as shown in Figure 18, Figure 19, and Figure 20, respectively. The plots reflect the three types of drawdown responses that we described in the introduction section. At an early time, the aquifer behaves as though it is confined with data falling along a straight line. At some intermediate time, it deviates from this straight line due to delayed gravity drainage. At a late time, data again follow a straight line as the aquifer exhibits a Theis-type response with S equivalent to S_y .

The transmissivity and specific yield of the aquifer are determined by fitting a straight line through the late-time data. The early-time response provides another estimate of transmissivity, and a confined storativity value for the aquifer can be obtained.

5.2. Determination of Aquifer Parameters :

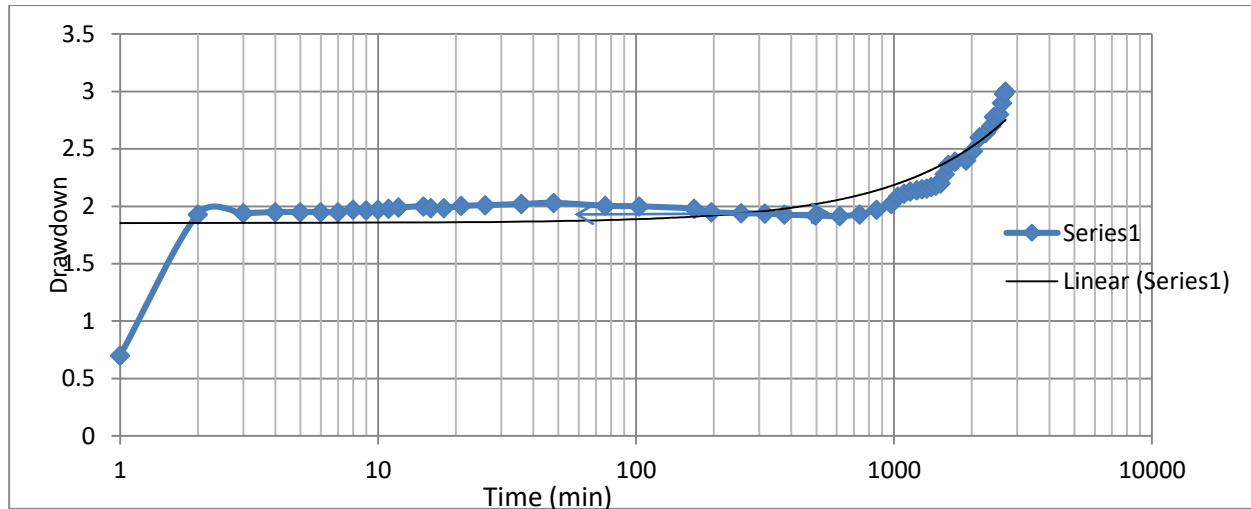
For the study area, Pumping rate, $Q = 78 \text{ m}^3/\text{hr}$.

Table 3 Initial Data recorded just before starting the Pumping Test

Well Name	Distance from the main well	Duration of monitoring	Depth of water from Datum / Static Water table	MSL of bed level	Discharge of pump
	(m)	(hr)	(m)	(m)	(m ³ /hr)
Main Well	--	45	2.98	37.12	78
OBS-1	10	45	0.42	37.40	--
OBS-2	20	45	0.34	37.03	--
OBS-3	98	45	1.1	38.25	--

For the main well,

For the study area, Pumping rate, $Q = 78 \text{ m}^3/\text{hr}$.



Observation **Well-1** is 20 m away from the pumping well.

Straight-line fitted through the data as shown from the basis for the calculation of hydraulic parameters.

For $Q = 78$,

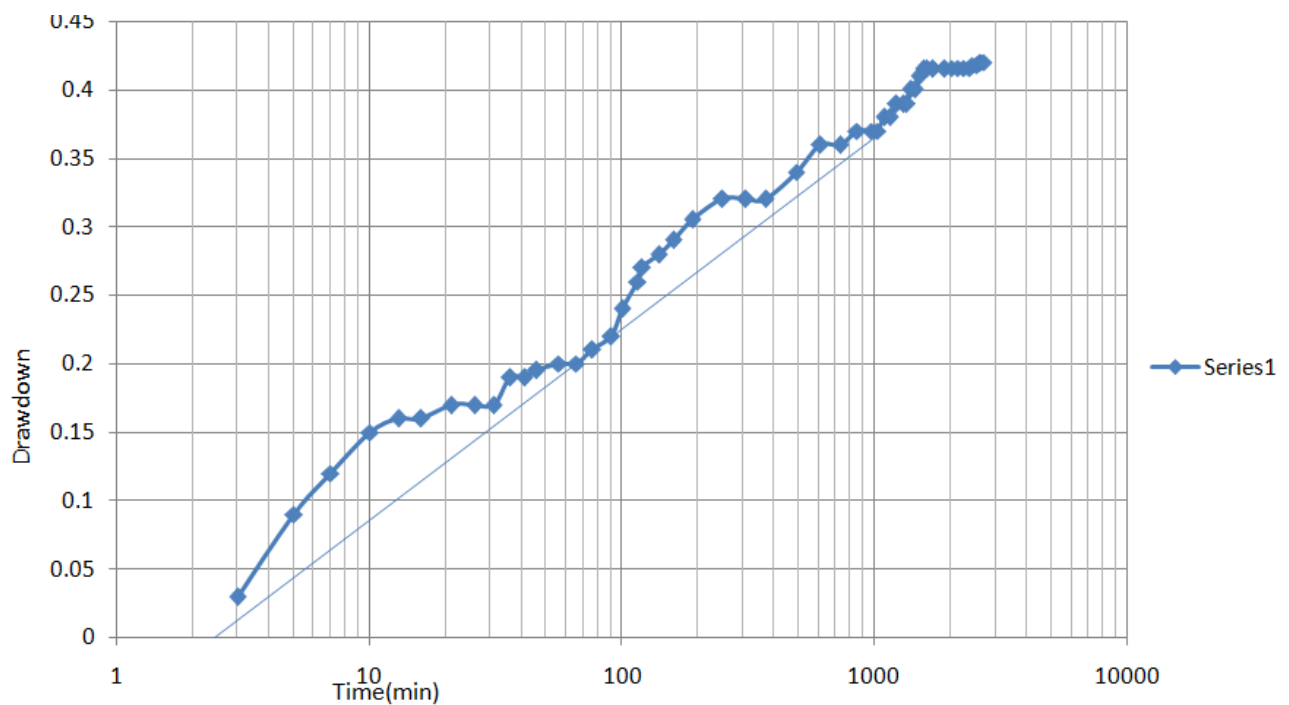


Figure 25 Drawdown (s) versus time (t) data for observation well-1

The transmissivity (T) of the aquifer has been determined using equation,

$$T = \frac{2.3Q}{4\pi\Delta s} = \frac{2.3 \times (78 \times 24)}{4\pi(0.13)} = 2634.42 \text{ m}^2/\text{day} \approx 2634 \text{ m}^2/\text{day}$$

For $Q = 60$,

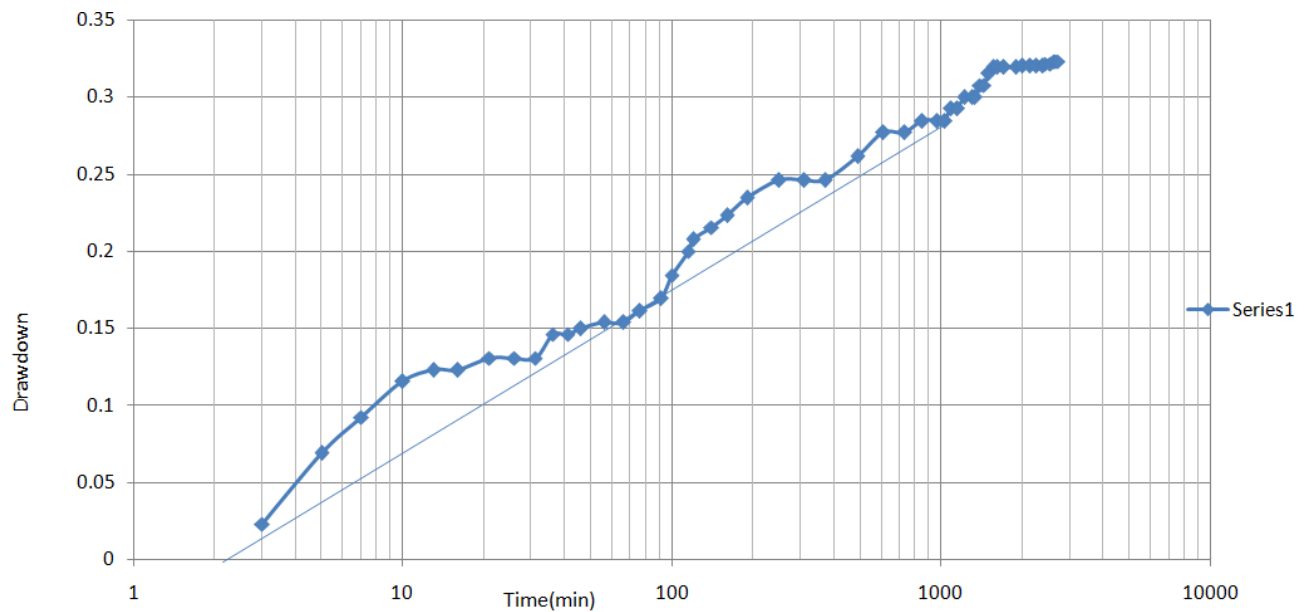


Figure 26 Drawdown (s) versus time (t) data for observation well-1

The transmissivity (T) of the aquifer has been determined using equation,

$$T = \frac{2.3Q}{4\pi\Delta s} = \frac{2.3 \times (60 \times 24)}{4\pi(0.1)} = 2634.42 \text{ m}^2/\text{day} \approx 2634 \text{ m}^2/\text{day}$$

For $Q = 100$,

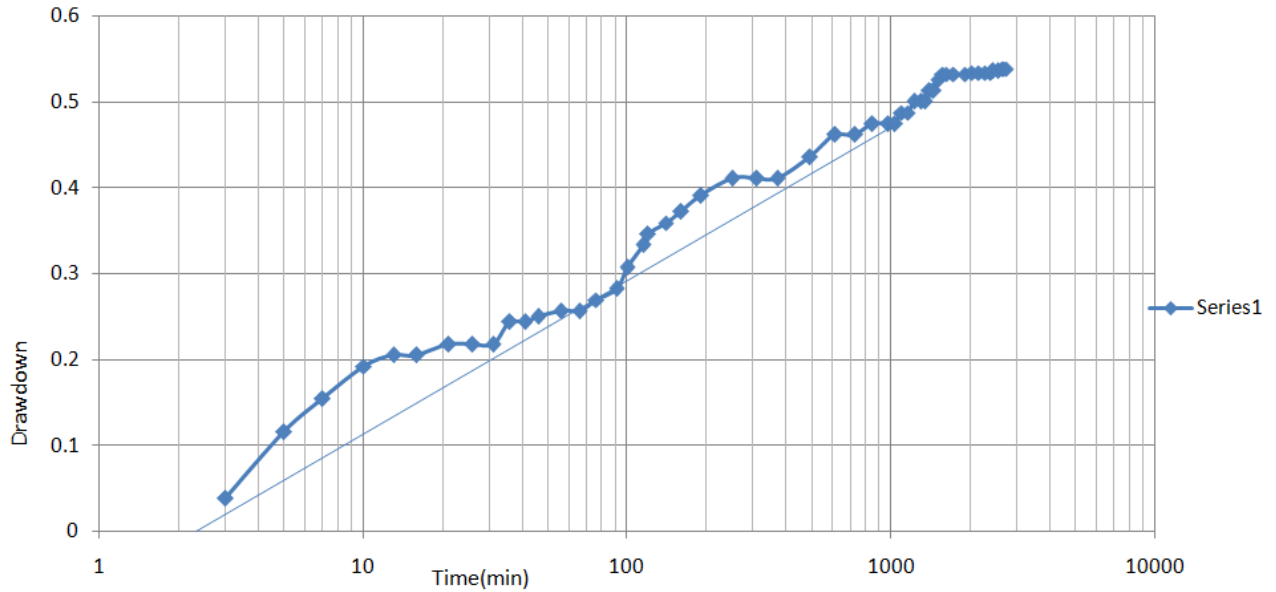


Figure 27 Drawdown (s) versus time (t) data for observation well-1

The transmissivity (T) of the aquifer has been determined using equation,

$$T = \frac{2.3Q}{4\pi\Delta s} = \frac{2.3 \times (100 \times 24)}{4\pi(0.17)} = 2582.77 \text{ m}^2/\text{day} \approx 2583 \text{ m}^2/\text{day}$$

For $Q = 120$,

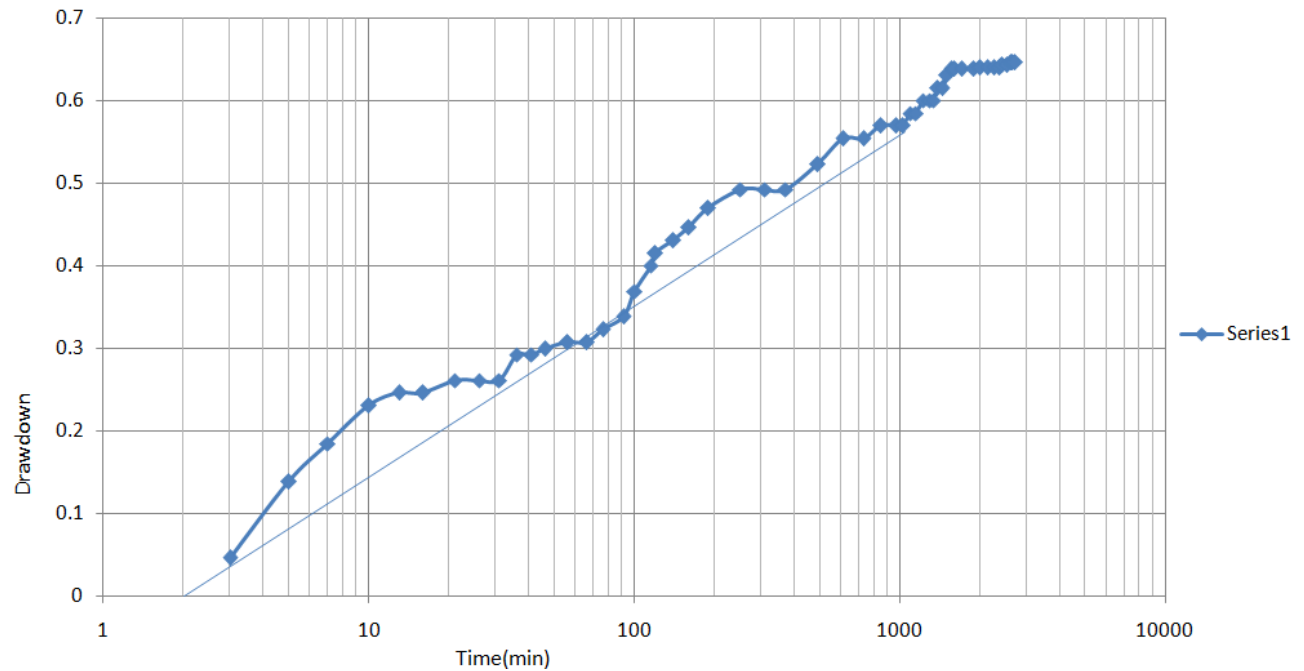
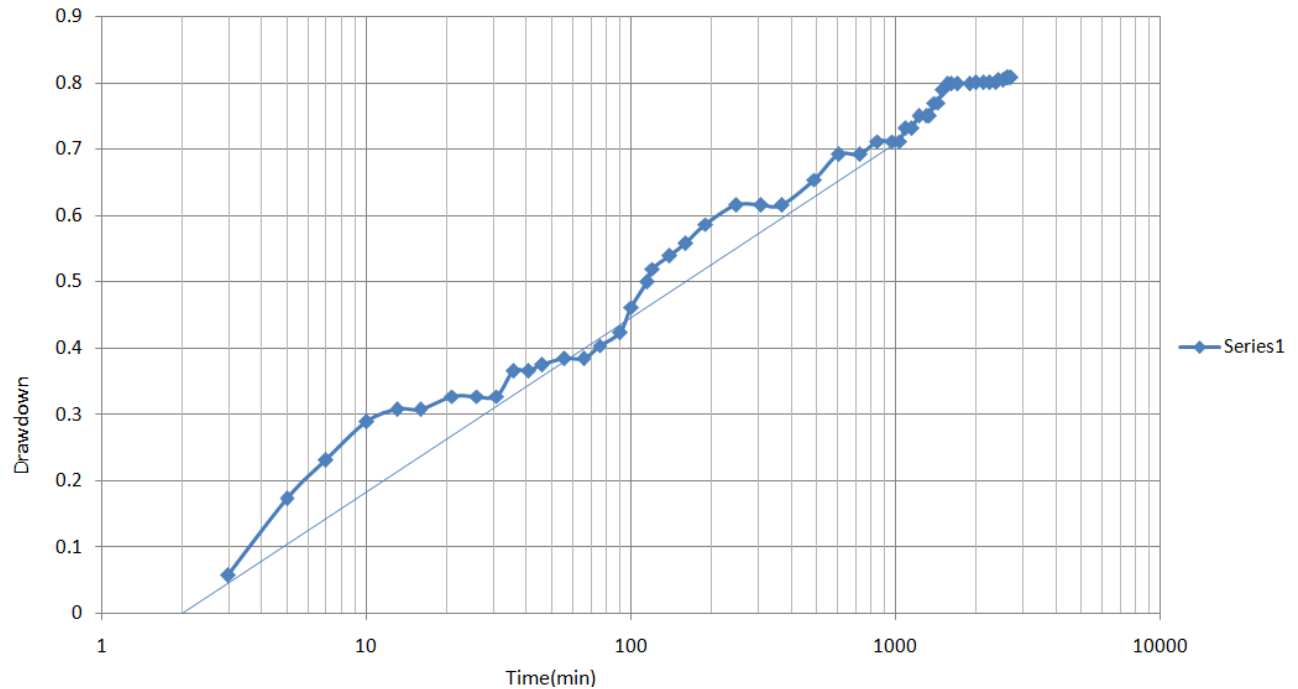


Figure 28 Drawdown (s) versus time (t) data for observation well-1

The transmissivity (T) of the aquifer has been determined using equation,

$$T = \frac{2.3Q}{4\pi\Delta s} = \frac{2.3 \times (120 \times 24)}{4\pi(0.2)} = 2634.42 \text{ m}^2/\text{day} \approx 2634 \text{ m}^2/\text{day}$$

For $Q = 150$,



Drawdown (s) versus time (t) data for observation well-1

The transmissivity (T) of the aquifer has been determined using equation,

$$T = \frac{2.3Q}{4\pi\Delta s} = \frac{2.3 \times (150 \times 24)}{4\pi(0.25)} = 2634.42 \text{ m}^2/\text{day} \approx 2634 \text{ m}^2/\text{day}$$

Average value of “ T ” in case of OBS -1 is,

$$T_{\text{avg}} = (2634 + 2634 + 2583 + 2634 + 2634)/5 = 2624 \text{ m}^2/\text{day}$$

For observation Well-2,

Straight-line fitted through the data as shown from the basis for the calculation of hydraulic parameters.

For $Q = 78$,

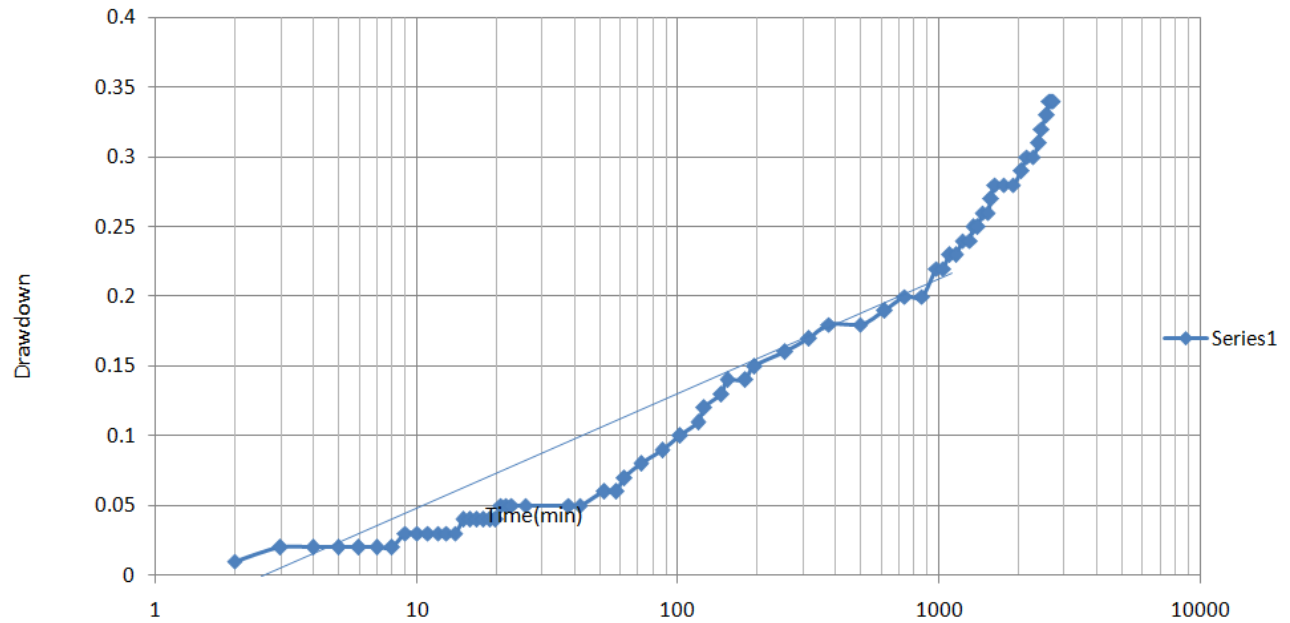


Figure 29 Drawdown (s) versus time (t) data for observation well-2

The transmissivity (T) of the aquifer has been determined using equation,

$$T = \frac{2.3Q}{4\pi\Delta s} = \frac{2.3 \times (78 \times 24)}{4\pi(0.12)} = 2853.96 \text{ m}^2/\text{day} \approx 2854 \text{ m}^2/\text{day}$$

For $Q = 60$,

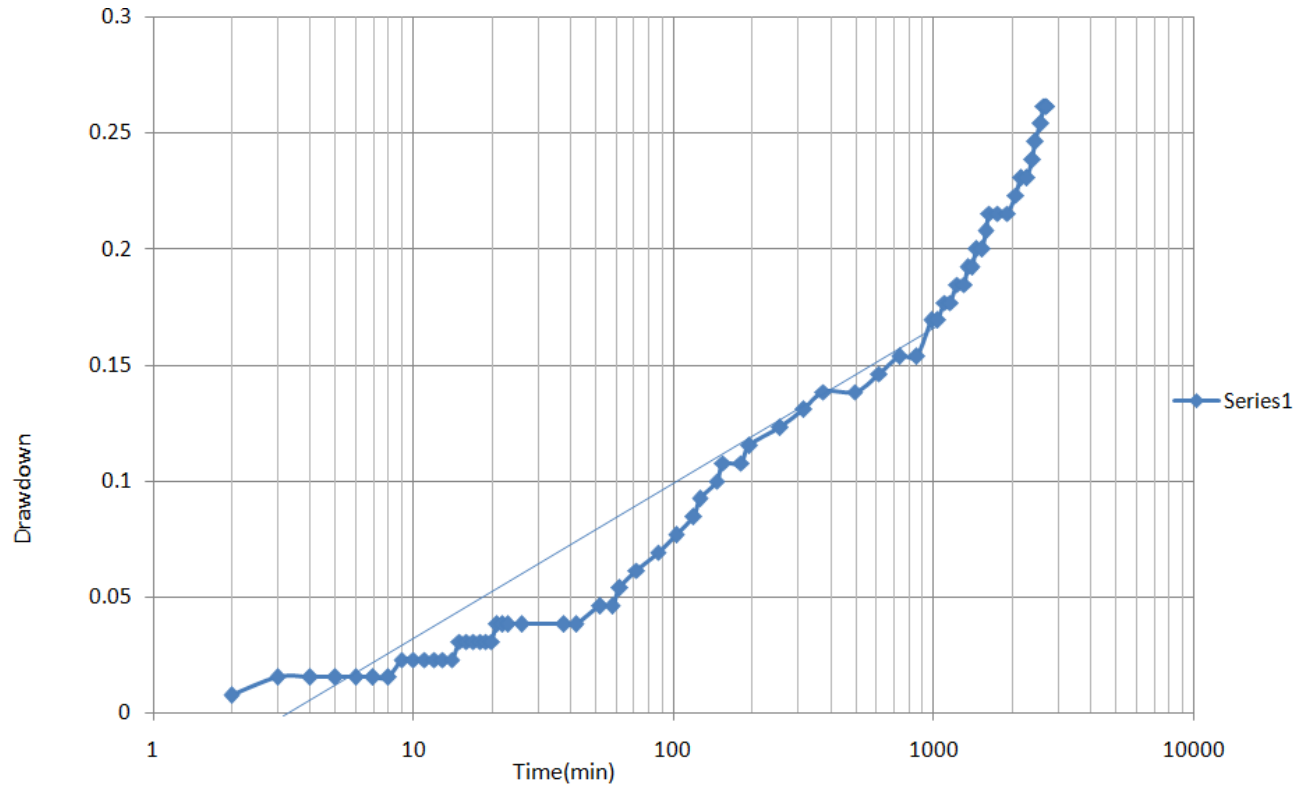


Figure 30 Drawdown (s) versus time (t) data for observation well-2

The transmissivity (T) of the aquifer has been determined using equation,

$$T = \frac{2.3Q}{4\pi\Delta s} = \frac{2.3 \times (60 \times 24)}{4\pi(0.93)} = 2832.71 \text{ m}^2/\text{day} \approx 2833 \text{ m}^2/\text{day}$$

For $Q = 100$,

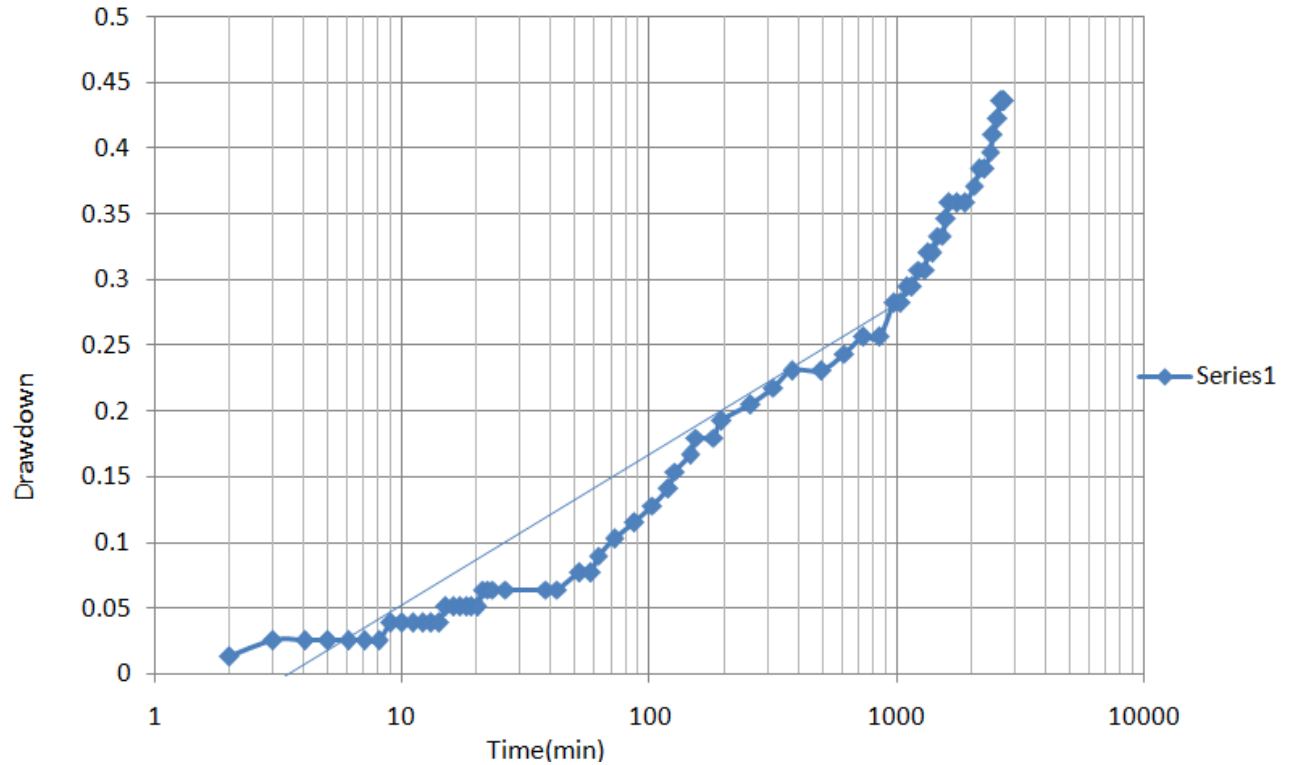


Figure 31 Drawdown (s) versus time (t) data for observation well-2

The transmissivity (T) of the aquifer has been determined using equation,

$$T = \frac{2.3Q}{4\pi\Delta s} = \frac{2.3 \times (100 \times 24)}{4\pi(0.154)} = 2851.11 \text{ m}^2/\text{day} \approx 2851 \text{ m}^2/\text{day}$$

For $Q = 120$,

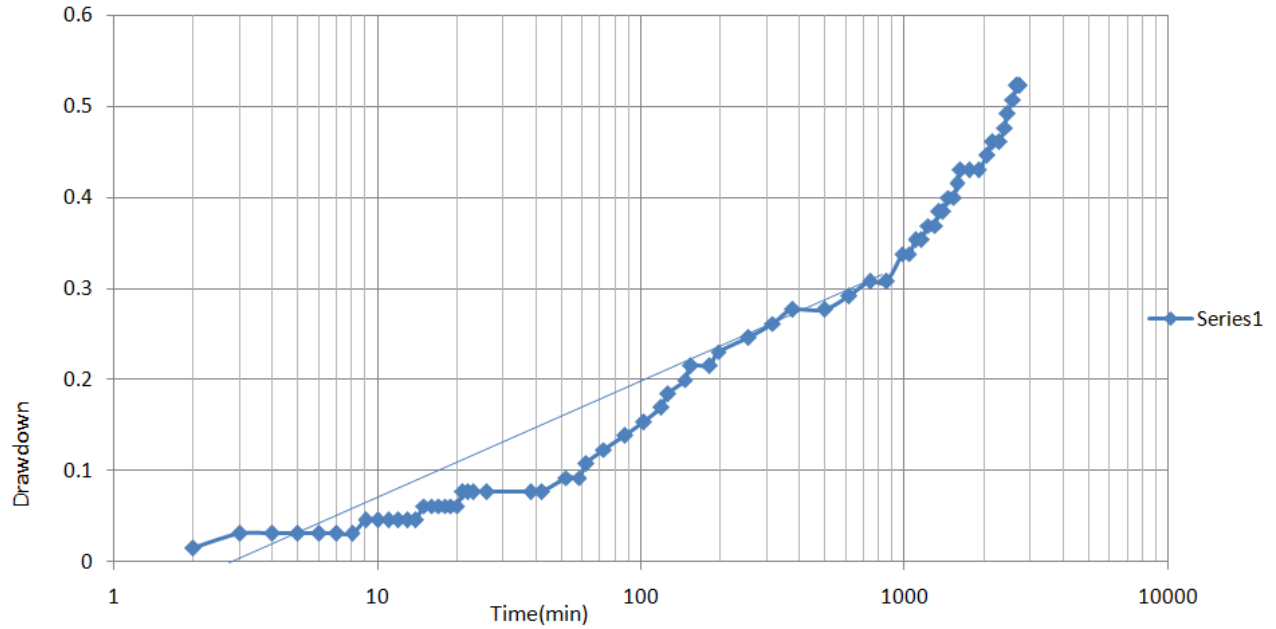


Figure 32 Drawdown (s) versus time (t) data for observation well-2

The transmissivity (T) of the aquifer has been determined using equation,

$$T = \frac{2.3Q}{4\pi\Delta s} = \frac{2.3 \times (120 \times 24)}{4\pi(0.184)} = 2863.5 \text{ m}^2/\text{day} \approx 2864 \text{ m}^2/\text{day}$$

For $Q = 150$,

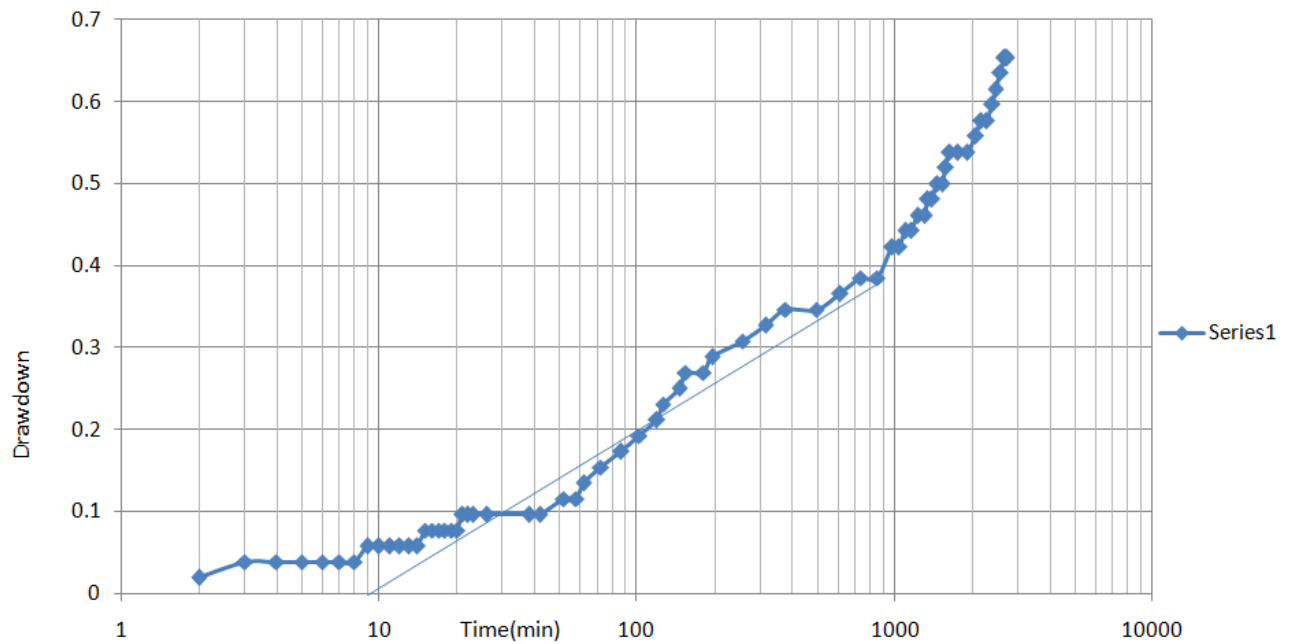


Figure 33 Drawdown (s) versus time (t) data for observation well-2

The transmissivity (T) of the aquifer has been determined using equation,

$$T = \frac{2.3Q}{4\pi\Delta s} = \frac{2.3 \times (150 \times 24)}{4\pi(0.23)} = 2863.5 \text{ m}^2/\text{day} \approx 2864 \text{ m}^2/\text{day}$$

Average value of “ T ” in case of OBS -2 is,

$$T_{\text{avg}} = (2854 + 2833 + 2851 + 2864 + 2864) / 5 = 2853 \text{ m}^2/\text{day}$$

For observation Well-3,

Straight-line fitted through the data as shown from the basis for the calculation of hydraulic parameters.

For $Q = 78$,

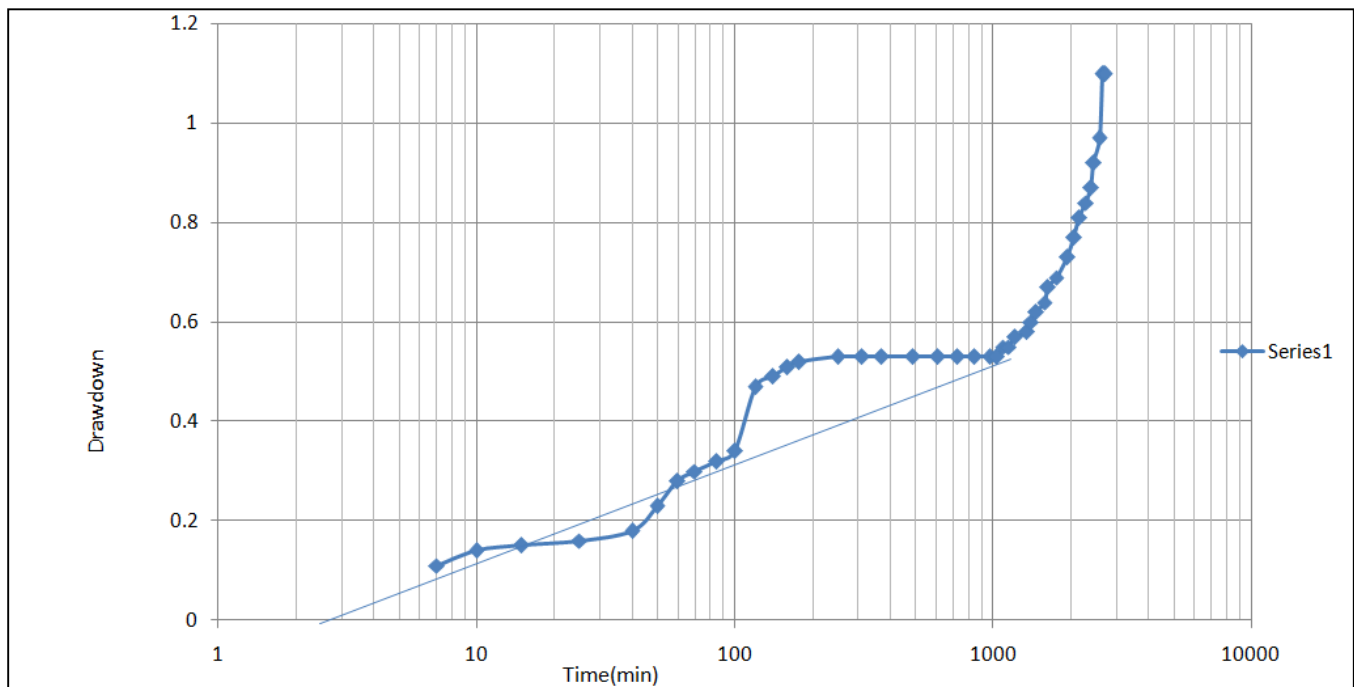


Figure 34 Drawdown (s) versus time (t) data for observation well-3

The transmissivity (T) of the aquifer has been determined using equation,

$$T = \frac{2.3Q}{4\pi\Delta s} = \frac{2.3 \times (78 \times 24)}{4\pi(0.19)} = 1802.5 \text{ m}^2/\text{day} \approx 1803 \text{ m}^2/\text{day}$$

For $Q = 60$,

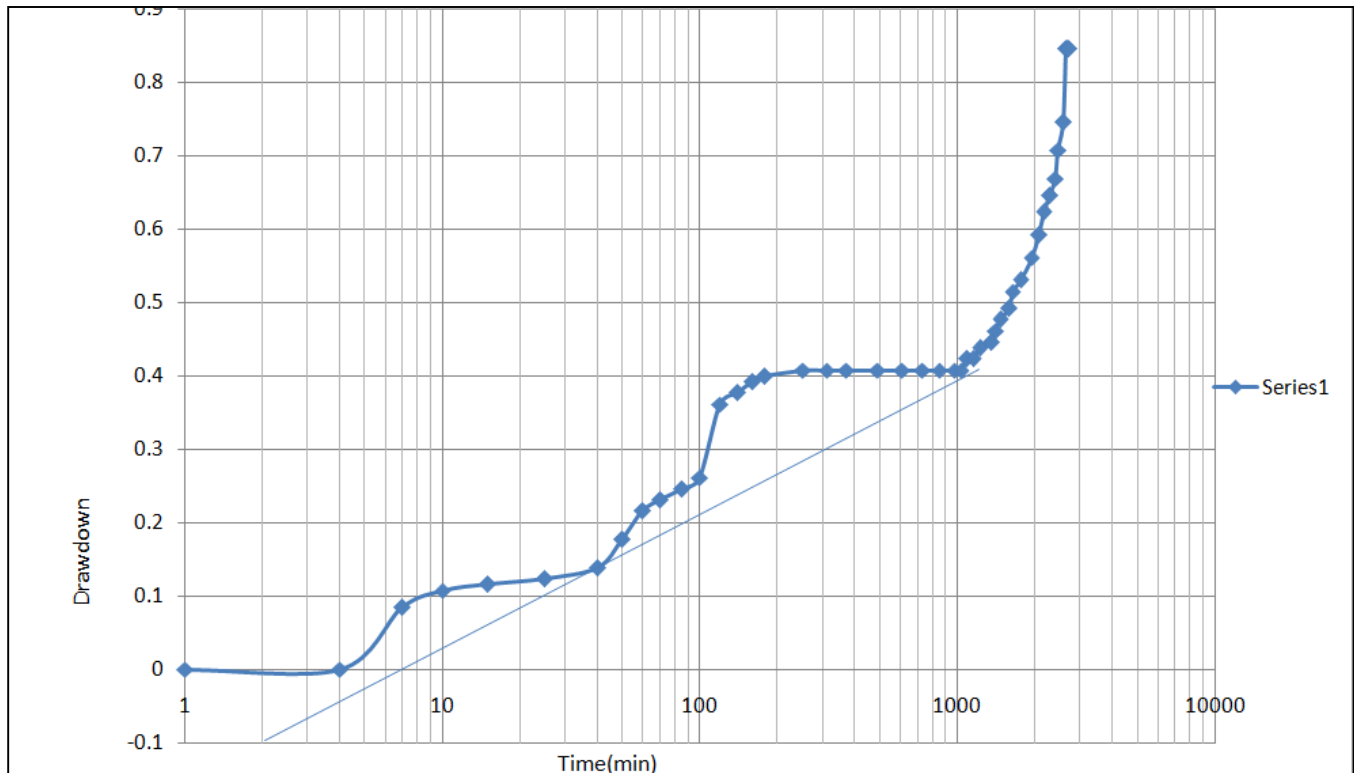


Figure 35 Drawdown (s) versus time (t) data for observation well-3

The transmissivity (T) of the aquifer has been determined using equation,

$$T = \frac{2.3Q}{4\pi\Delta s} = \frac{2.3 \times (60 \times 24)}{4\pi(0.16)} = 1646.51 \text{ m}^2/\text{day} \approx 1647 \text{ m}^2/\text{day}$$

For $Q = 100$,

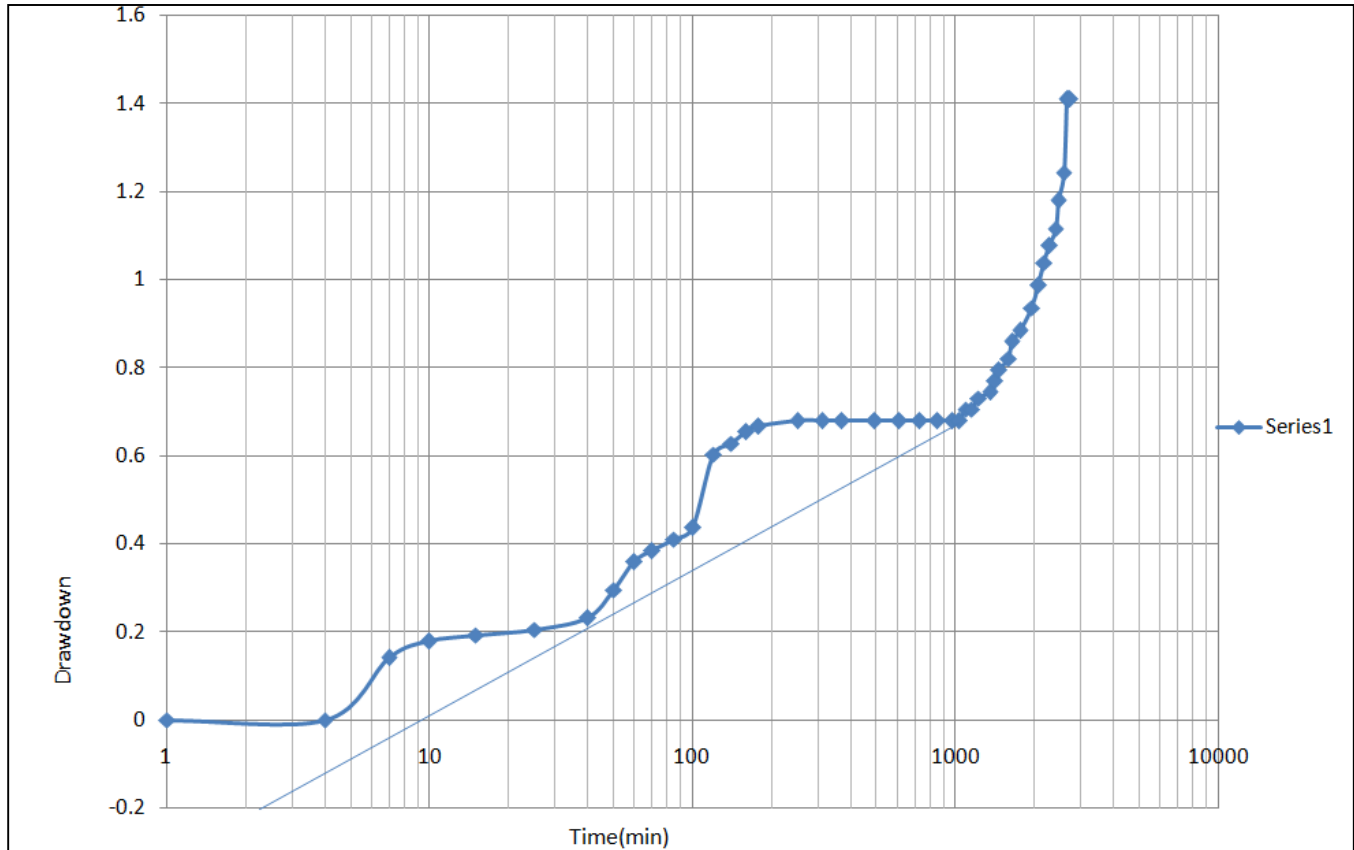


Figure 36 Drawdown (s) versus time (t) data for observation well-3

The transmissivity (T) of the aquifer has been determined using equation,

$$T = \frac{2.3Q}{4\pi\Delta s} = \frac{2.3 \times (100 \times 24)}{4\pi(0.24)} = 1829.46 \text{ m}^2/\text{day} \approx 1829 \text{ m}^2/\text{day}$$

For $Q = 120$,

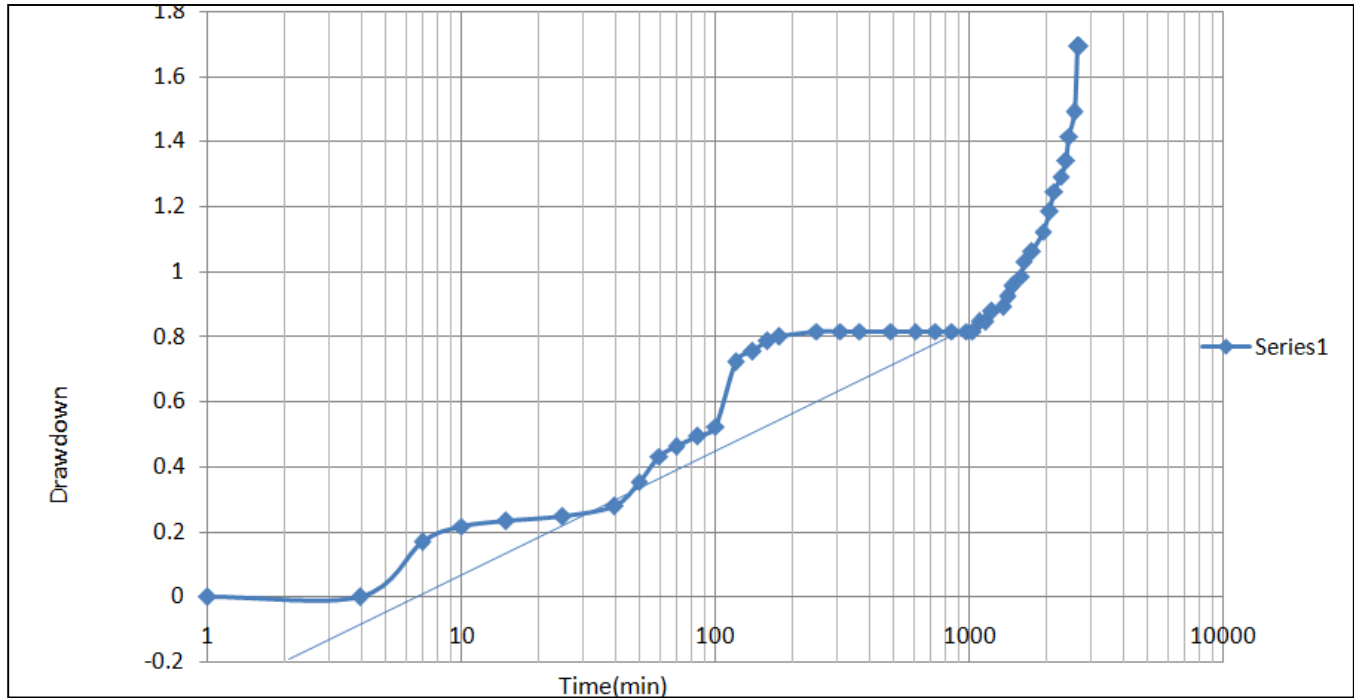


Figure 37 Drawdown (s) versus time (t) data for observation well-3

The transmissivity (T) of the aquifer has been determined using equation,

$$T = \frac{2.3Q}{4\pi\Delta s} = \frac{2.3 \times (120 \times 24)}{4\pi(0.29)} = 1816.84 \text{ m}^2/\text{day} \approx 1817 \text{ m}^2/\text{day}$$

For $Q = 150$,

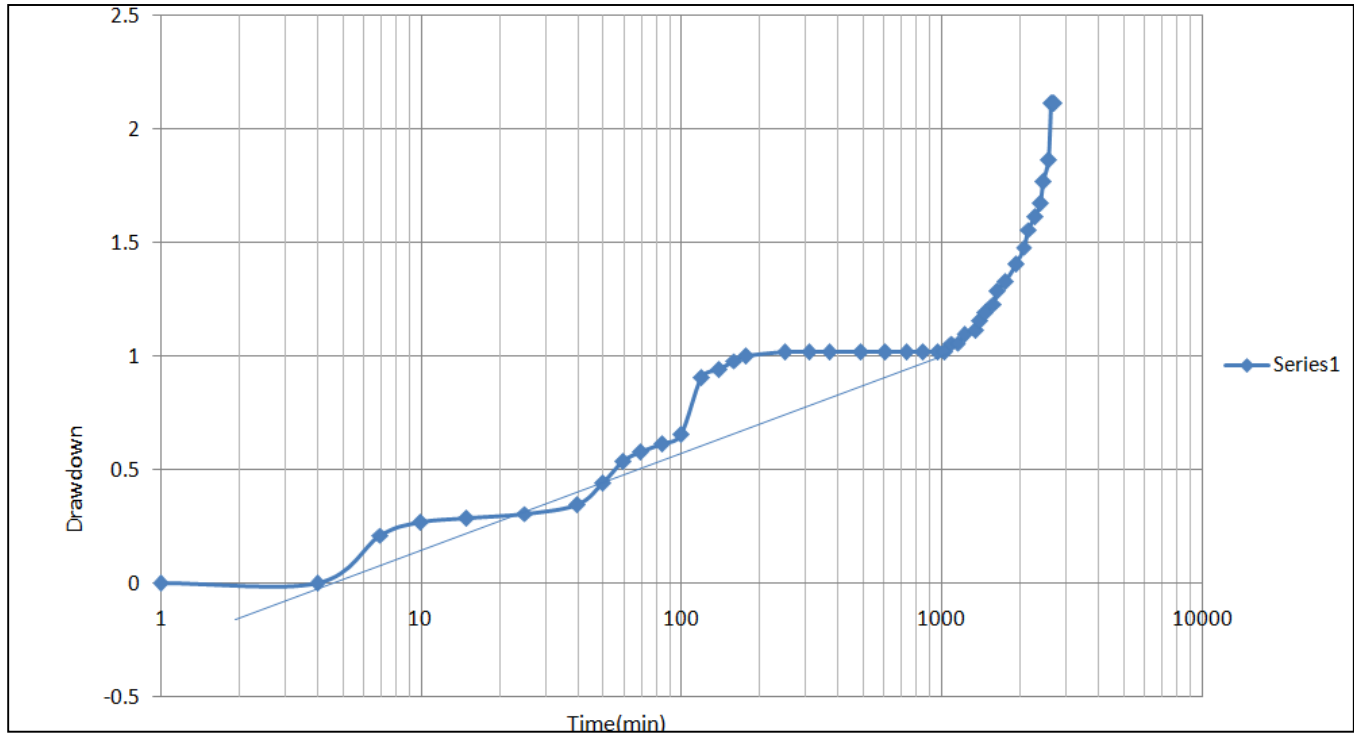


Figure 38 Drawdown (s) versus time (t) data for observation well-3

The transmissivity (T) of the aquifer has been determined using equation,

$$T = \frac{2.3Q}{4\pi\Delta s} = \frac{2.3 \times (150 \times 24)}{4\pi(0.35)} = 1881.73 \text{ m}^2/\text{day} \approx 1882 \text{ m}^2/\text{day}$$

Average value of “ T ” in case of OBS -3 is,

$$T_{\text{avg}} = (1803 + 1647 + 1829 + 1817 + 1882)/5 = 1796 \text{ m}^2/\text{day}$$

5.3. Calculation of Storativity and Radius of influence:

The confined storativity (S) has been determined using equation

$$S = \frac{2.25Tt_{0s}}{r^2} = \frac{2.25 \times 4759 \times 2.5}{10^2 \times 60 \times 24} = 0.1030$$

The **Radius of Influence** (R) has been determined by the following equation

$$Q = \frac{2\pi T s_d}{\ln \ln \left(\frac{R}{r} \right)} = \frac{2 \times \pi \times 2634 \times 0.42}{\ln \ln \left(\frac{R}{10} \right)}$$

$$R \approx 410 \text{ m}$$

The calculation for the rest of the observation wells for constant discharge and different distances from the pumping well is the same as above.

5.4. Results at a Glance:

Well Name	Distance from Main Well	Storativity	Discharge	Transmissivity	Maximum Drawdown
	(m)		(m ³ /hr)	(m ² /day)	(m)
Main Well (338 m from the left bank)	-	-	78		2.98
OBS-1	10	0.103006	-	2634	0.42
OBS-2	20	0.098668	-	2853	0.34
OBS-3	98	0.223179	-	1796	1.1
Average Transmissivity = 2427 m ² /day					
Critical Value of Transmissivity=1796 m ² /day					

All the distances of the observation wells were measured with respect to the respective main wells. The diameter of the main well was 200 mm and the depth of boring was 45 m. The diameter of the three observation wells was 50 mm each and their depth of boring was 45 m.

5.5. Recuperation Test:

At the end of the pumping test when the pump is stopped, the water level in the pumping well and three observation wells will begin to rise. This is referred to as recovery of groundwater level, while measurements of drawdown below the original static water level (prior to pumping) during this recovery period are known as residual drawdowns. In the present study, the maximum drawdown in the main well was measured 3.0 m and the residual drawdown was measured 0.0 m which suggests 100% recovery.

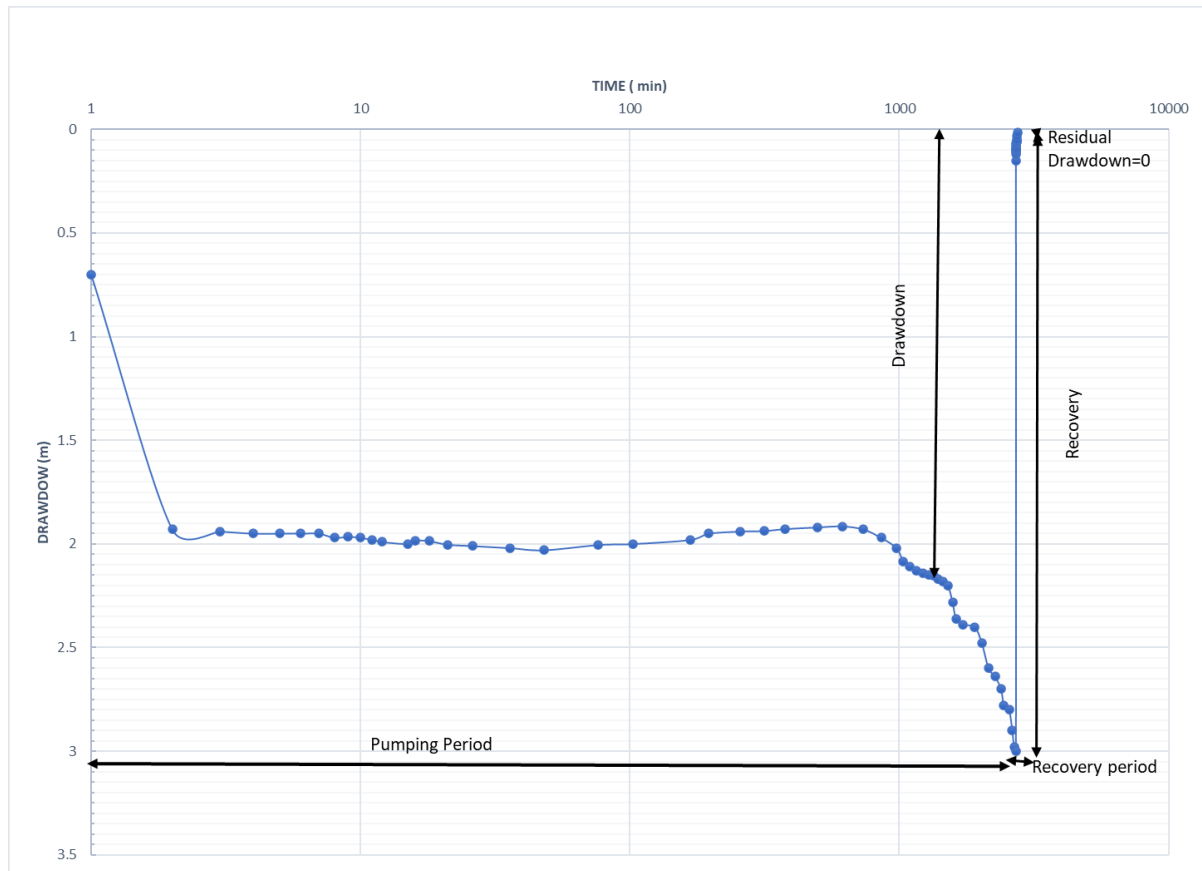


Figure 39 Drawdown versus time curve representing recuperation data for Main well

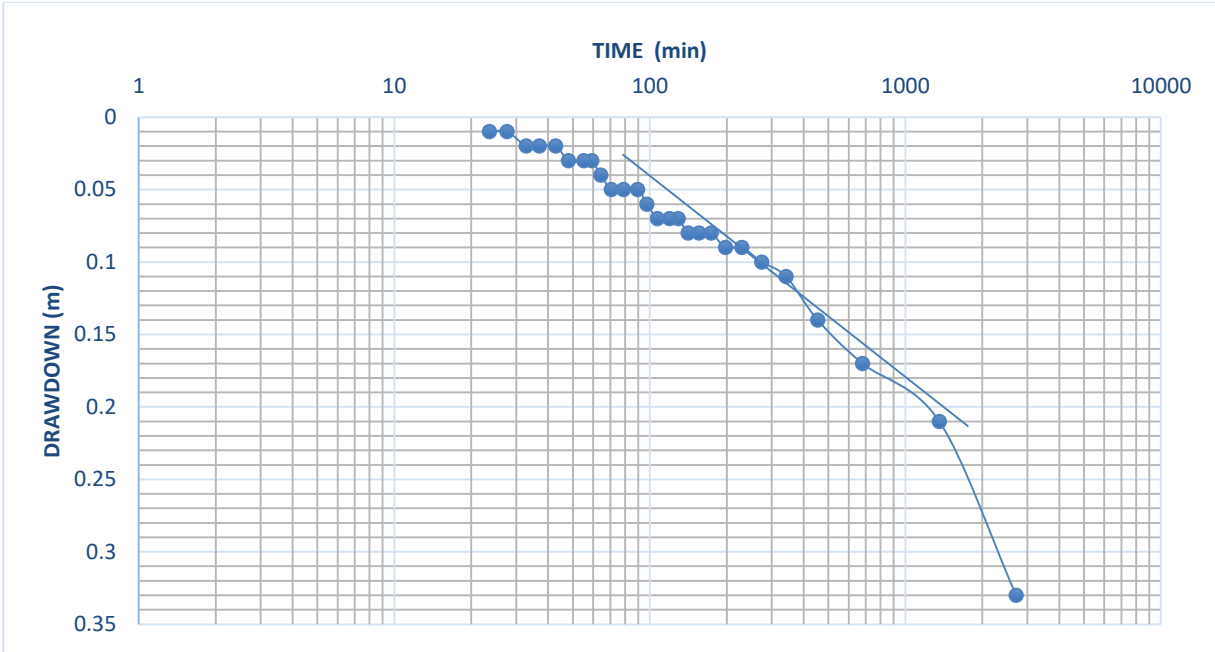


Figure 40 Time drawdown curve, Theis recovery for observation well-1.

The values of S' versus t/t' are plotted on semi-log paper and using the following equation, transmissivity and storage coefficient have been determined.

$$T = \frac{2.3Q}{4\pi\Delta S'} = \frac{2.3 \times 78 \times 24}{4\pi(0.17 - 0.06)} = 3116 \text{ m}^2/\text{d}$$

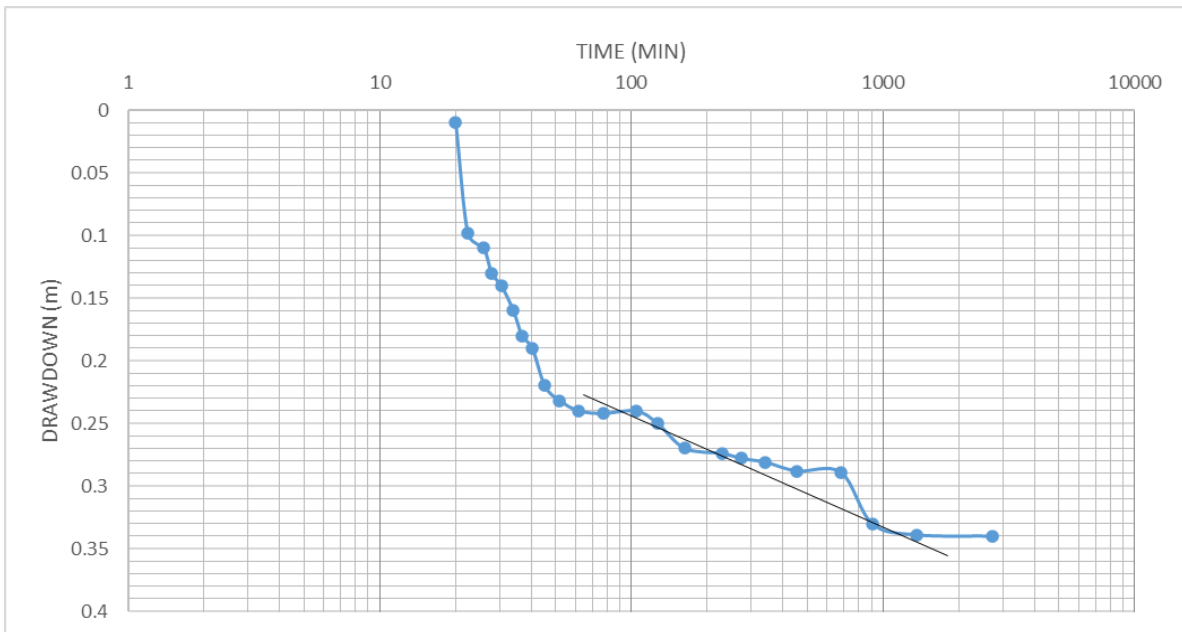


Figure 41 Time drawdown curve, Theis recovery for observation well-2.

5.6. Result at glance:

The table presents a detailed analysis of various parameters related to groundwater flow and aquifer characteristics at OBS 1 well and OBS 2. The data includes different flow rates (Q) measured in cubic meters per hour (m³/hr) and cubic meters per day (m³/day), alongside corresponding values of transmissivity (measured in m²/day), storativity (s), and the radius of influence (measured in meters).

Analysis of the Data

Flow Rates (Q):

The flow rates range from 60 m³/hr to 150 m³/hr, which translates to 1440 m³/day to 3600 m³/day. This variation in flow rates allows for the examination of how different pumping rates affect the aquifer's properties.

Transmissivity:

The transmissivity values observed are relatively consistent across different flow rates, varying slightly from 2634 m²/day to 2662 m²/day. This consistency indicates a stable aquifer system with a relatively uniform ability to transmit water through its porous medium.

Storativity:

The storativity values are also quite consistent, ranging from 0.1 to 0.103. Storativity represents the capacity of the aquifer to store water. These values suggest a moderate storage capacity, which is crucial for understanding the aquifer's response to pumping and recharge.

Radius of Influence:

The radius of influence increases slightly with higher flow rates, from 404 meters at 60 m³/hr to 419 meters at 150 m³/hr. The radius of influence indicates the distance from the well at which the pumping effect is noticeable. This gradual increase suggests that higher pumping rates extend the area affected by the well, potentially impacting a larger volume of the aquifer.

	Flow (Q) (m ³ /hr)	Flow (Q) (m ³ /day)	Transmissivity (m ² /day)	Storativity (s)	Radius of Influence (m)
OBS 1	60	1440	2634	0.103	404
	78	1872	2634	0.102	408
	100	2400	2642	0.1	411
	120	2880	2647	0.103	416
	150	3600	2662	0.103	419

OBS 2	Flow (Q) (m ³ /hr)	Flow (Q) (m ³ /day)	Transmissivity (m ² /day)	Storativity (s)	Radius of Influence (m)
	60	1440	2832	0.027	496
	78	1872	2853	0.028	516
	100	2400	2855	0.027	532
	120	2880	2861	0.028	536
	150	3600	2864	0.028	541

Table 4 All Data at a glance

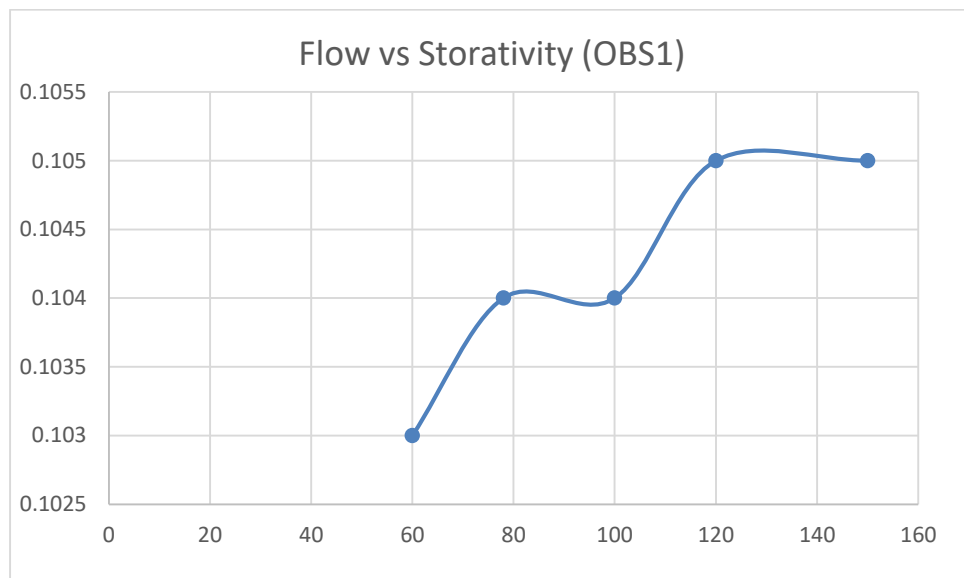


Figure 42 Graph of flow rate(m³/day) vs storativity

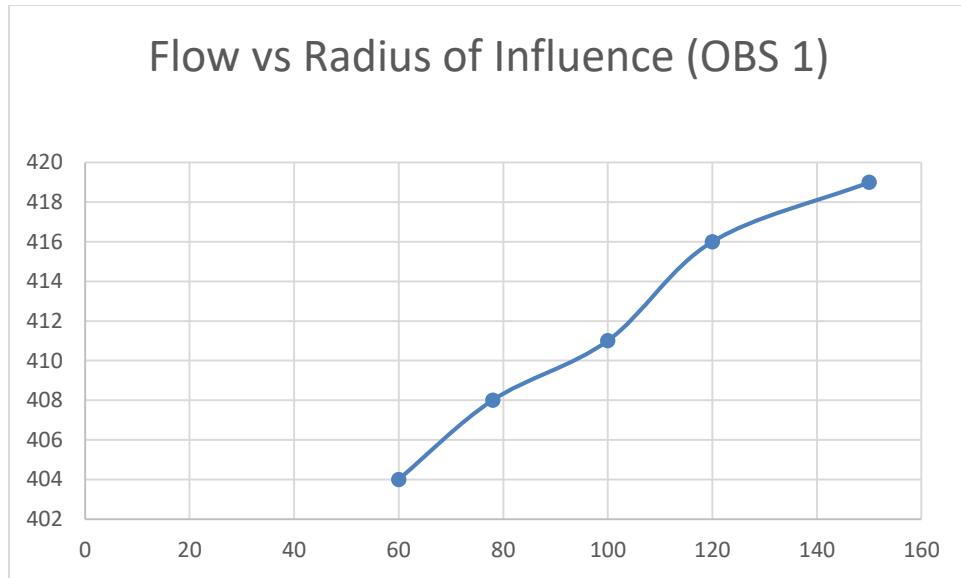


Figure 43 Graph of flow rate(m³/day) vs Radius of influence

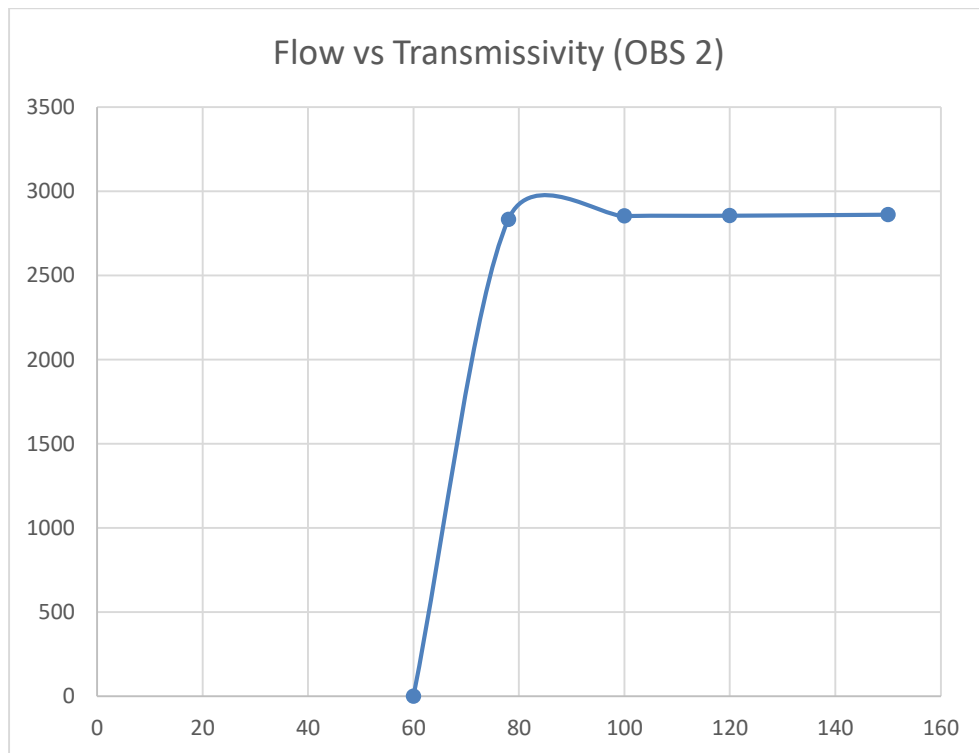


Figure 44 Graph of flow rate(m³/day) vs Transmissivity

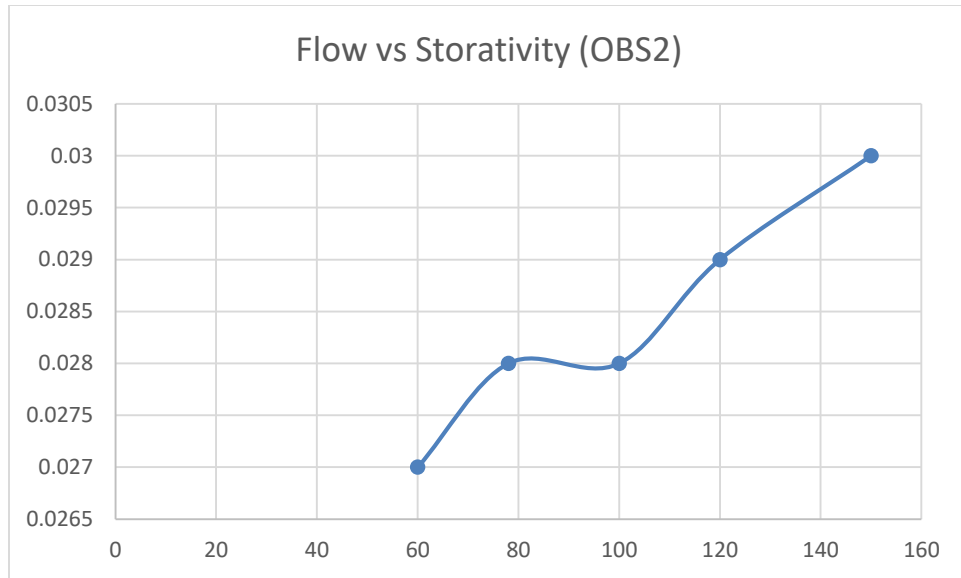


Figure 45 Graph of flow rate(m³/day) vs storativity

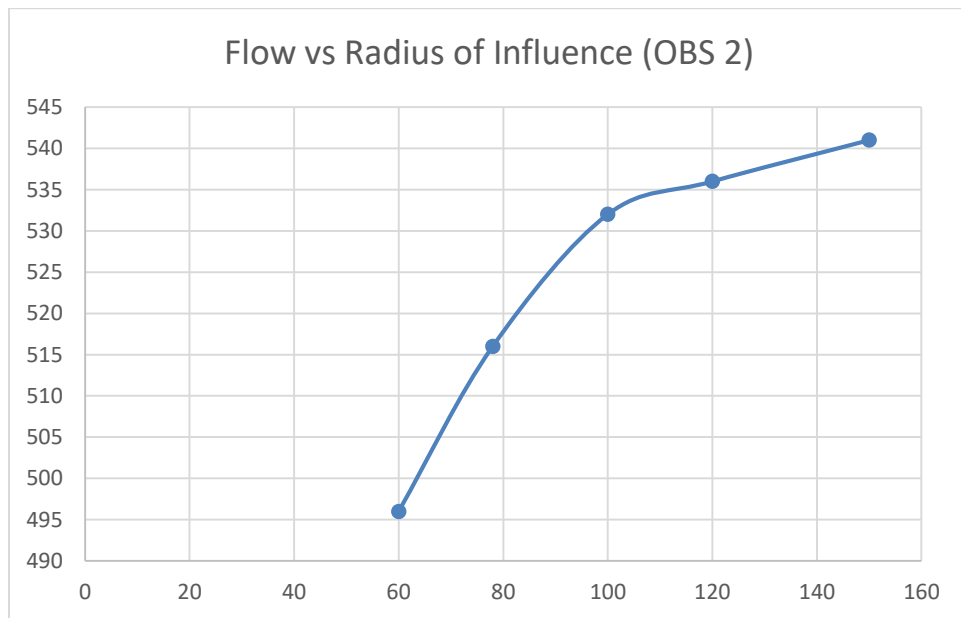


Figure 41 Graph of flow rate(m³/day) vs Radius of influence

5.7. Mechanical analysis data:

A mechanical analysis is done with the bore well log material and from that the material size is determined.

Table 5 Mechanical Analysis of Main Well (depth 18 m to 30 m)

Sieve Size (mm)	Weight Retained (gm)	Cumulative Weight Retained (gm)	% of Cum. Wt. Retained	% Finer
11.2	0	0	0.000	100.000
4.75	0.24	0.24	0.024	99.976
3.35	0	0.24	0.024	99.976
2.3	3.76	4	0.401	99.599
1.7	34.37	38.37	3.883	96.117
0.85	289.94	328.31	32.915	67.085
0.5	422.23	750.54	75.946	24.054
0.3	232.56	983.1	99.478	0.522
0.12	2.06	985.16	99.686	0.314
0.09	0.29	985.45	99.716	0.284
Pan	2.81	988.26	100.000	0.000
CALCULATION				
$D_{30} = 0.55 \text{ mm}$, $D_{50} = 0.68 \text{ mm}$, $D_{60} = 0.8 \text{ mm}$, $D_{10} = 0.4 \text{ mm}$				
Coefficient of Uniformity, $C_u = D_{60}/D_{10} = 2$				
Coefficient of Curvature, $C_c = D_{30}^2/(D_{60} \times D_{10}) = 0.95$				

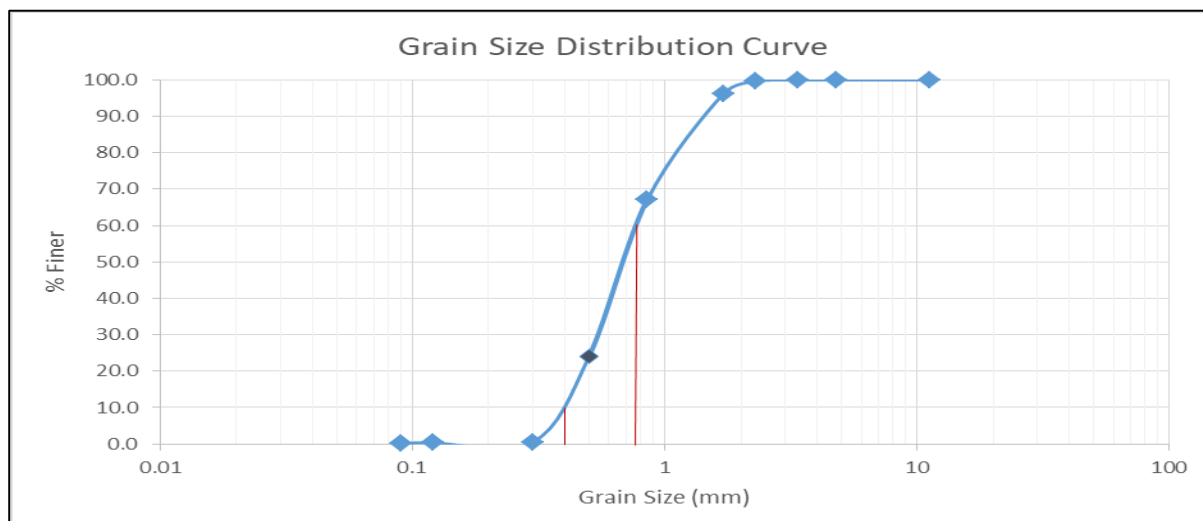


Figure 46 IS sieve aperture dimension (mm) vs Cumulative % passing as per Sieve analysis for the main well (depth 18 m to 30 m)

Table 6 Mechanical Analysis of Main Well (depth 31 m to 42 m)

Sieve Size (mm)	Weight Retained (gms)	Cumulative Weight Retained (gms)	% of Cum. Wt. Retained	% Finer
11.2	0	0	0.000	100.000
4.75	0	0	0.000	100.000
3.35	0	0	0.000	100.000
2.3	0	0	0.000	100.000
1.7	1.53	1.53	0.152	99.848
0.85	138.34	139.87	13.938	86.062
0.5	498.71	638.58	63.634	36.366
0.3	346.47	985.05	98.159	1.841
0.12	0.86	985.91	98.245	1.755
0.09	0	985.91	98.245	1.755
Pan	17.61	1003.52	100.000	0.000
CALCULATION				
$D_{30} = 0.46 \text{ mm}$, $D_{50} = .5 \text{ mm}$, $D_{60} = 0.63 \text{ mm}$, $D_{10} = 0.36 \text{ mm}$				
Coefficient of Uniformity, $C_u = D_{60}/D_{10} = 1.92$				
Coefficient of Curvature, $C_c = D_{30}^2/(D_{60} \times D_{10}) = 0.926$				

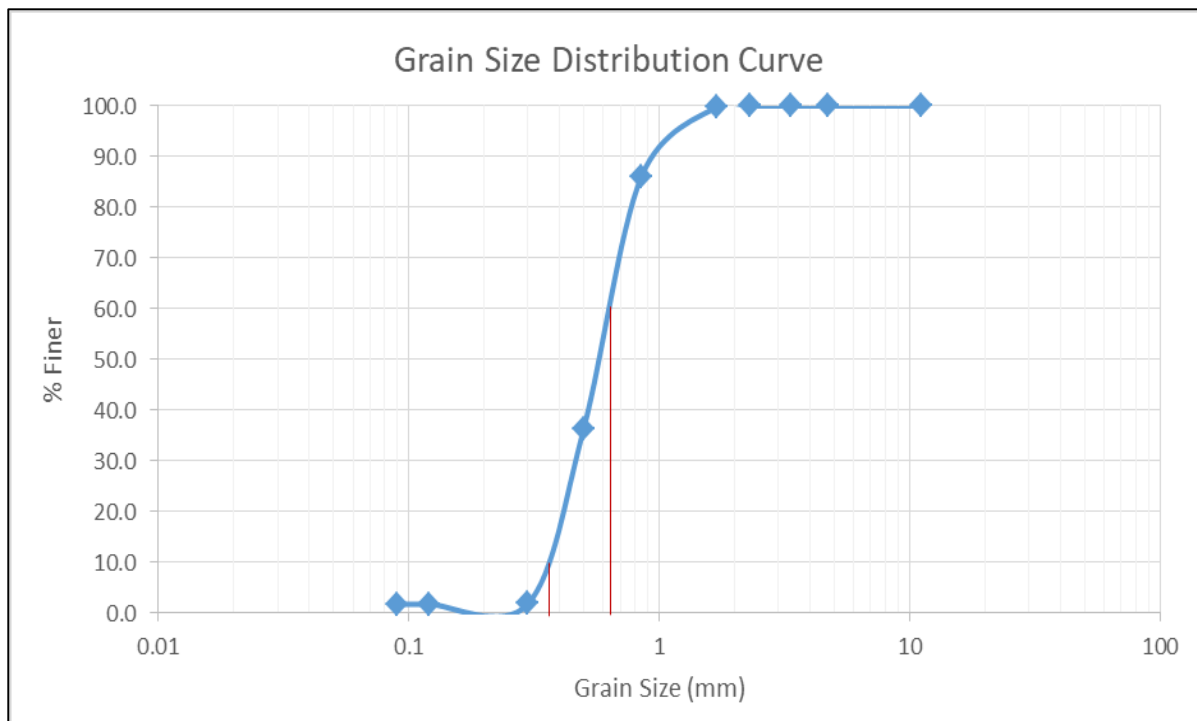


Figure 47 IS sieve aperture dimension (mm) versus cumulative % passing as per Sieve analysis for Main Well (depth 31 m to 42 m)

Table 7 Mechanical Analysis of Observation Well-1 (depth 20 m to 42 m)

Sieve Size (mm)	Weight Retained (gms)	Cumulative Weight Retained (gms)	% of Cum. Wt. Retained	% Finer
11.2	0	0	0.000	100.000
4.75	11.76	11.76	1.179	98.821
3.35	13	24.76	2.482	97.518
2.3	2.2	26.96	2.703	97.297
1.7	43.65	70.61	7.244	92.756
0.85	231.3	301.91	30.975	69.025
0.5	363.1	665.01	68.227	31.773
0.3	306.1	971.11	99.632	0.368
0.12	0.33	971.44	99.666	0.334
0.09	0.13	971.57	99.679	0.321
Pan	3.13	974.7	100.000	0.000
CALCULATION				
$D_{30} = 0.5 \text{ mm}$, $D_{50} = 0.67 \text{ mm}$, $D_{60} = 0.765 \text{ mm}$, $D_{10} = 0.36 \text{ mm}$				
Coefficient of Uniformity, $C_u = D_{60}/D_{10} = 2.12$				
Coefficient of Curvature, $C_c = D_{30}^2/(D_{60} \times D_{10}) = 0.864$				

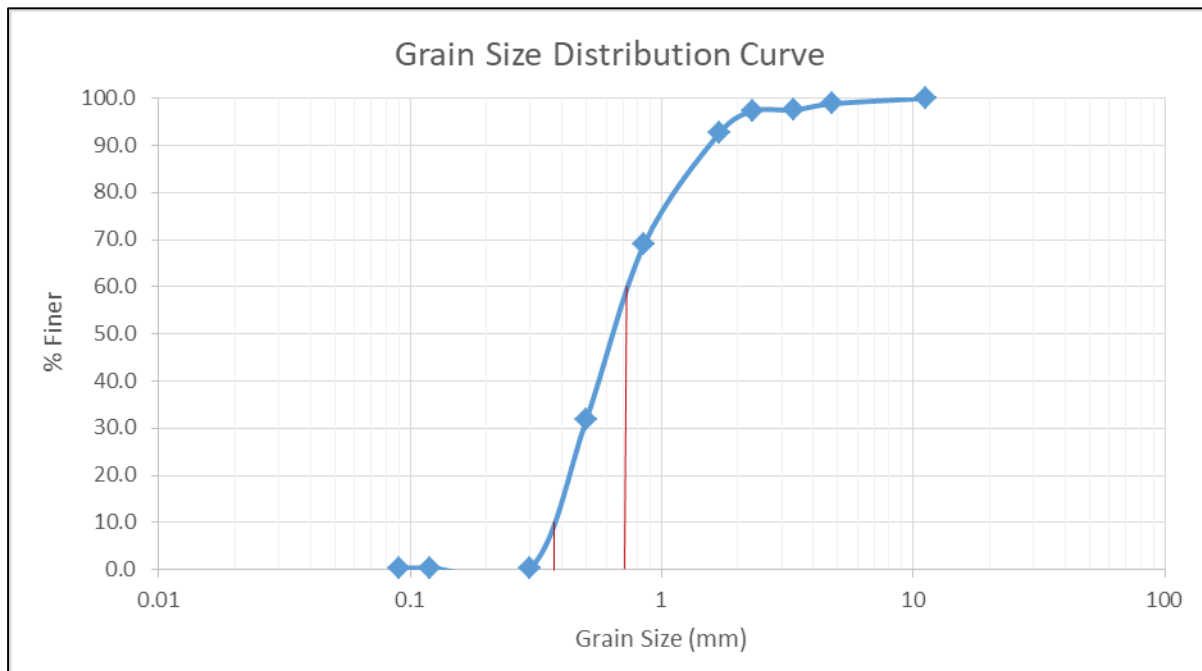


Figure 48 IS sieve aperture dimension (mm) versus cumulative % passing as per sieve analysis for observation well-1 (depth 20 m to 42 m)

Table 8 Mechanical Analysis of Observation Well-2 (depth 20 m to 42 m)

Sieve Size (mm)	Weight Retained (gms)	Cumulative Weight Retained (gms)	% of Cum. Wt. Retained	% Finer
11.2	0	0	0.000	100.000
4.75	13.99	13.99	1.403	98.597
3.35	20.81	34.8	3.489	96.511
2.3	36.67	71.47	7.165	92.835
1.7	54.53	126	12.647	87.353
0.85	217.43	343.43	34.472	65.528
0.5	415.65	759.08	76.193	23.807
0.3	234.27	993.35	99.708	0.292
0.12	0.02	993.37	99.710	0.290
0.09	0.07	993.44	99.717	0.283
Pan	2.82	996.26	100.000	0.000
CALCULATION				
$D_{30} = 0.55 \text{ mm}$, $D_{50} = 0.72 \text{ mm}$, $D_{60} = 0.8 \text{ mm}$, $D_{10} = 0.38 \text{ mm}$				
Coefficient of Uniformity, $C_u = D_{60}/D_{10} = 2.1$				
Coefficient of Curvature, $C_c = D_{30}^2/(D_{60} \times D_{10}) = 0.99$				

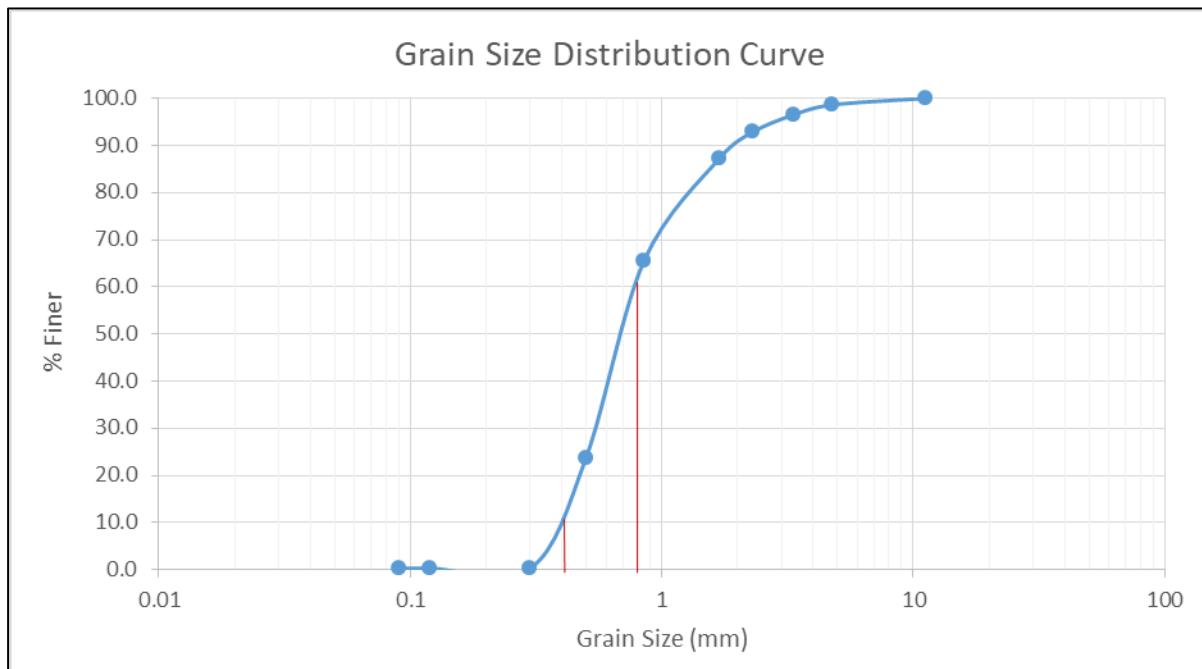


Figure 49 IS sieve aperture dimension (mm) versus cumulative % passing as per Sieve analysis for observation well-2 (depth 20 m to 42 m)

Table 9 Mechanical Analysis of Observation Well-3 (depth 20 m to 42 m)

Sieve Size (mm)	Weight Retained (gms)	Cumulative Weight Retained (gms)	% of Cum. Wt. Retained	% Finer
11.2	0	0	0.000	100.000
4.75	6.17	6.17	0.619	99.381
3.35	9.56	15.73	1.577	98.423
2.3	19.31	35.04	3.513	96.487
1.7	33.75	68.79	6.976	93.024
0.85	219.04	287.83	29.189	70.811
0.5	379.68	667.51	67.692	32.308
0.3	313.75	981.26	99.509	0.491
0.12	0.48	981.74	99.558	0.442
0.09	0.15	981.89	99.573	0.427
Pan	4.21	986.1	100.000	0.000
CALCULATION				
$D_{30} = 0.485 \text{ mm}$, $D_{50} = 0.66 \text{ mm}$, $D_{60} = 0.75 \text{ mm}$, $D_{10} = 0.36 \text{ mm}$				
Coefficient of Uniformity, $C_u = D_{60}/D_{10} = 2.1$				
Coefficient of Curvature, $C_c = D_{30}^2/(D_{60} \times D_{10}) = 0.924$				

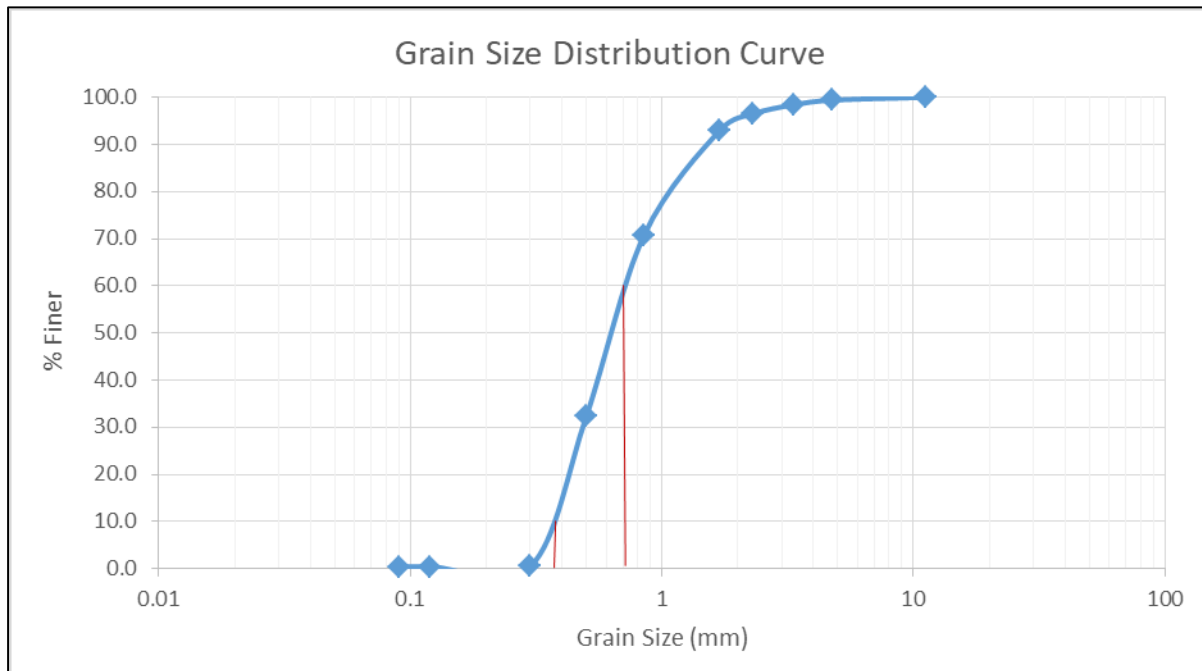


Figure 50 IS sieve aperture dimension (mm) versus cumulative % passing as per Sieve analysis for observation well-3 (depth 20 m to 42 m)

Based on the grain size distribution data, it is observed that the Coefficient of Uniformity(C_u) mostly varies from 1.9 to 2.1. Based on the analysis, an artificial gravel pack around the pumping well is required as the Coefficient of Uniformity is found less than 3.0. The gravel pack materials should have a Coefficient of Uniformity less than 2.5. For determination of the gravel size, D_{30} can be considered and around 4 to 6 times of D_{30} can be selected for artificial gravel packing.

5.8. Determination of “K” value from litholog:

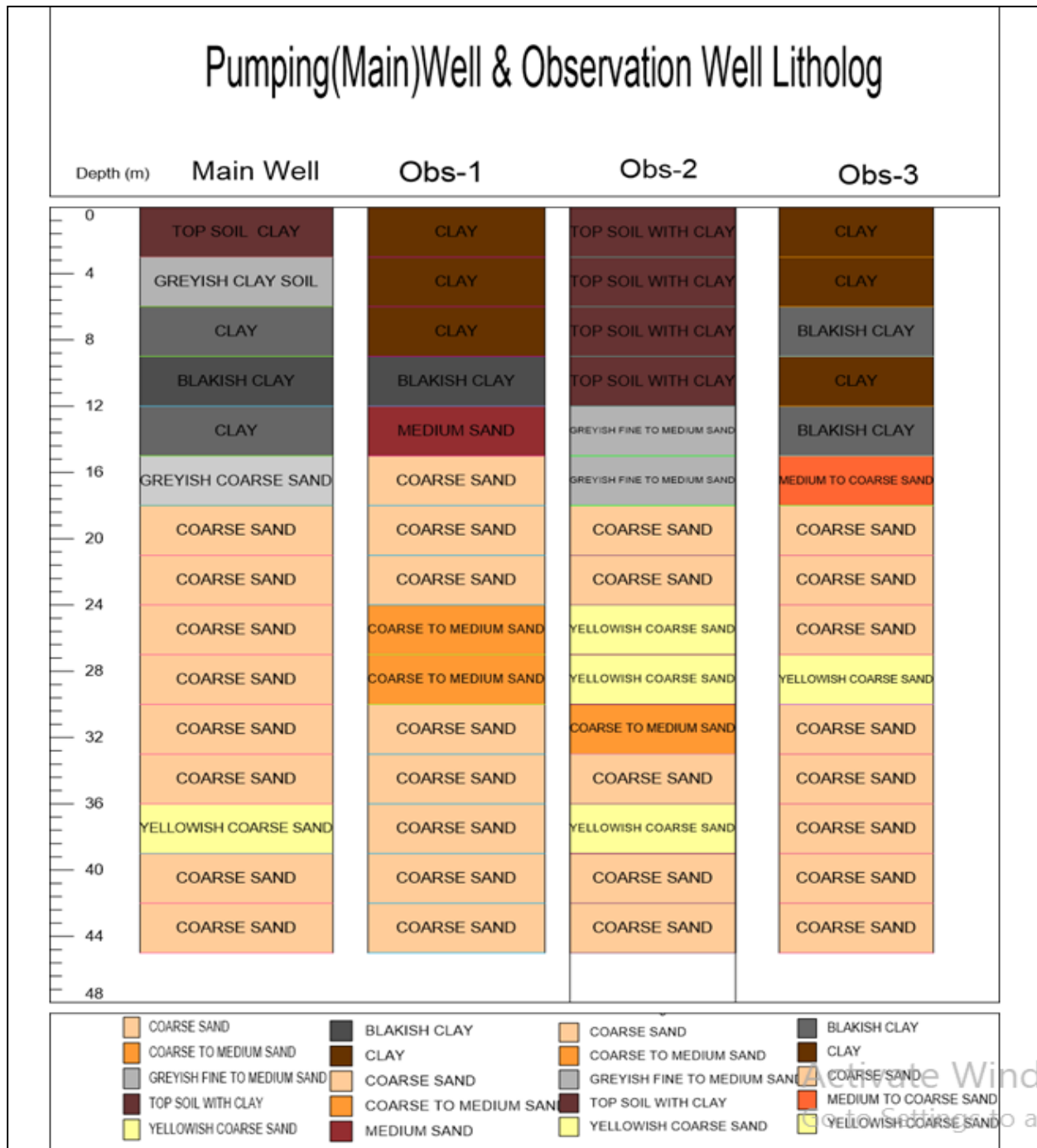


Figure 51Lithological details of the pumping well and three observation wells (OBS-1, OBS-2 and OBS-3).

The table provides a detailed stratigraphic profile of soil types at various depths. From 0 to 12 meters, the soil is predominantly top soil clay. Below this, from 12 to 18 meters, the soil transitions into greyish fine to medium sand. As the depth increases from 18 to 27 meters, the soil type changes to coarse sand and then to yellowish coarse sand. Between 27 to 30 meters, the soil becomes coarse to medium sand. Finally, from 30 meters to the bottom of the profile at 45 meters, the soil consists of coarse sand, with a brief section of yellowish coarse sand at 36 to 39 meters. This profile indicates a gradual change from clay at the surface to various types of sand at greater depths.

Table 10Main well lithlog

DEPTH (m)	TYPE OF SOIL
0 to 3	TOP SOIL CLAY
3 to 6	TOP SOIL CLAY
6 to 9	TOP SOIL CLAY
9 to 12	TOP SOIL CLAY
12 to 15	GREYISH FINE TO MEDIUM SAND
15 to 18	GREYISH FINE TO MEDIUM SAND
18 to 21	COARSE SAND
21 to 24	COARSE SAND
24 to 27	YELLOWISH COARSE SAND
27 to 30	YELLOWISH COARSE SAND
30 to 33	COARSE TO MEDIUM SAND
33 to 36	COARSE SAND
36 to 39	YELLOWISH COARSE SAND
39 to 42	COARSE SAND
42 to 45	COARSE SAND

Table 11 Observation well-1 litholog data:

DEPTH (m)	TYPE OF SOIL
0 to 3	TOP SOIL CLAY
3 to 6	GREYISH CLAY SOIL
6 to 9	CLAY
9 to 12	BLAKISH CLAY
12 to 15	CLAY
15 to 18	GREYISH COARSE SAND
18 to 21	COARSE SAND
21 to 24	COARSE SAND
24 to 27	COARSE SAND
27 to 30	COARSE SAND
30 to 33	COARSE SAND
33 to 36	COARSE SAND
36 to 39	YELLOWISH COARSE SAND
39 to 42	COARSE SAND

Table 12 Observation well-2 litholog data:

DEPTH (m)	TYPE OF SOIL
0 to 3	CLAY
3 to 6	CLAY
6 to 9	CLAY
9 to 12	BLAKISH CLAY
12 to 15	MEDIUM SAND
15 to 18	COARSE SAND
18 to 21	COARSE SAND
21 to 24	COARSE SAND
24 to 27	COARSE TO MEDIUM SAND
27 to 30	COARSE TO MEDIUM SAND
30 to 33	COARSE SAND
33 to 36	COARSE SAND
36 to 39	COARSE SAND
39 to 42	COARSE SAND

Table 13 Observation well-3 litholog data:

DEPTH (m)	TYPE OF SOIL
0 to 3	CLAY
3 to 6	CLAY
6 to 9	BLAKISH CLAY
9 to 12	CLAY
12 to 15	BLAKISH CLAY
15 to 18	MEDIUM TO COARSE SAND
18 to 21	COARSE SAND
21 to 24	COARSE SAND
24 to 27	COARSE SAND
27 to 30	YELLOWISH COARSE SAND
30 to 33	COARSE SAND
33 to 36	COARSE SAND
36 to 39	COARSE SAND
39 to 42	COARSE SAND

Calculating transmissivity (T) from litholog data typically involves using the hydraulic conductivity (K) of the geological layers and the thickness (b) of each layer. The transmissivity is given by the formula:

$$T = \sum (K_i \cdot b_i)$$

Calculating the “T” value from the Equation,

$$T = K \cdot B \quad \dots\dots\dots(5.4)$$

Where,

“T”(m²/day) represent Transmissivity.

“K” (m/day) represents Co- Efficient of permeability.

“B” (m) represents the depth.

Table 14Final value of Transmissivity from Litholog given below:

	K(m/day)	B(m)	T(m²/day)
Main Well	57.88	33	1910.04
OBS 1	57.88	27	1562.76
OBS 2	57.88	30	1736.4
OBS 2	57.88	27	1562.76

5.10. Comparison between “T” value from pumping test and litholog:

5.10.1. Evaluation of “T” value from pumping test:

In the given dataset, transmissivity (T) values for three observation points (OBS 1, OBS 2, and OBS 3) are analyzed at different pumping rates (Q). For OBS 1, the transmissivity remains constant at 2634 for most pumping rates except at Q=100, where it slightly drops to 2583. The average transmissivity (T_{avg}) for OBS 1 is calculated to be 2634. In OBS 2, transmissivity values range between 2832 and 2864 across the different pumping rates, resulting in an average transmissivity of 2853. For OBS 3, transmissivity varies more significantly, ranging from 1647 at Q=60 to 1882 at Q=150, with an average transmissivity of 1793. These calculations demonstrate how transmissivity values can vary with different pumping rates and highlight the importance of averaging to obtain a representative value for transmissivity in hydrogeological studies.

In case of OBS 1:

For Q= 60, Transmissivity= 2634

For Q= 78, Transmissivity=2634

For Q= 100, Transmissivity=2583

For Q= 120, Transmissivity=2634

For Q= 150, Transmissivity=2634

$$T_{avg} = (2634+2634+2583+2634+2634)/5 = 2634$$

In case of OBS 2:

For Q= 60, Transmissivity= 2832

For Q= 78, Transmissivity= 2853

For Q= 100, Transmissivity=2851

For Q= 120, Transmissivity=2864

For Q= 150, Transmissivity=2864

$$T_{avg} = (2832+2853+2851+2864+2864)/5 = 2853$$

In case of OBS 3:

For Q= 60, Transmissivity= 1647

For Q= 78, Transmissivity= 1803

For Q= 100, Transmissivity=1817

For Q= 120, Transmissivity=1817

For Q= 150, Transmissivity=1882

$$T_{avg} = (1647+1803+1817+1817+1882)/5 = 1793$$

Table 15 Result in tabular form:

	“T” value from pumping test ($T_{pumping}$)	“T” value from litholog data ($T_{litholog}$)	Comparison Factor ($T_{pumping}$)/($T_{litholog}$)
OBS 1	2634	1562.76	1.68
OBS 2	2853	1736.4	1.5
OBS 3	1793	1562.76	1.14

The table compares transmissivity (T) values obtained from pumping tests (T_{pumping}) with those derived from litholog data (T_{litholog}) for three observation points. The comparison factor is calculated as the ratio of to ($T_{\text{pumping}} / T_{\text{litholog}}$) .

1. Observation 1:

- T_{pumping} : 2634
- T_{litholog} : 1562.76
- Comparison Factor: 1.68

2. Observation 2:

- T_{pumping} : 2853
- T_{litholog} : 1736.4
- Comparison Factor: 1.5

3. Observation 3:

- T_{pumping} : 1793
- T_{litholog} : 1562.76
- Comparison Factor: 1.14

Summary:

The transmissivity values obtained from pumping tests are consistently higher than those derived from litholog data. The comparison factors indicate the magnitude of this difference, with values ranging from 1.14 to 1.68, suggesting that pumping test results are 14% to 68% higher than litholog estimates.

From the litholog data $T = 1562.76 \text{ m}^2/\text{day}$. Which is near about the value of Transmissivity OBS 1 well when the flow is 60 to 78 m^3/hr . is $2634 \text{ m}^2/\text{day}$.

So The correction factor is, $(2634/1562.76)=1.68$

So, In this zone ,

$$T_{\text{pumping}} = (1.68 * T_{\text{litholog}})$$

CHAPTER 6: CONCLUSION AND RECOMENDATION

6.1. Conclusion:

Based on the Lithological and Hydro geological investigation pertaining to the feasibility study, design considerations and site constraints, for the selected study area along the Damodar River near Jujuti village under the Galsi-II block of Bardhaman district, the following recommendations are furnished below.

- Based on the average transmissivity of 2634 m²/day and other aquifer parameters, it is feasible to withdraw 1500 KLD of water from above mention study location using a tube well.
- The pumping test results indicate that the aquifer has sufficient transmissivity to support the desired water withdrawal rate.
- The maximum drawdown observed during the pumping test was found to be within the acceptable limit.
- The groundwater recovery rate, as indicated by the recuperation, exceeded 100% within 2.5 hours, signifying favorable conditions for water replenishment after pumping ceases.
- During the construction, the artificial gravel packing should be ensured.
- The transmissivity values from the pumping tests (T_{pumping}) varied when compared to the litholog-derived values (T_{litholog}). For the three observations, the comparison factors ($T_{\text{pumping}} / T_{\text{litholog}}$) ranged from 1.14 to 1.68, indicating that the pumping test values can be both higher and lower than the litholog values. This variability underscores the importance of using multiple methods to accurately characterize aquifer properties.

The average transmissivity derived from litholog data (1620.64 m²/day) is reasonably close to the values obtained from the pumping tests. This consistency supports the reliability of using litholog data for preliminary assessments of aquifer properties, although pumping tests provide a more dynamic and direct measurement.

6.2. Recommendations:

- **Enhanced Monitoring:** Regular monitoring of the aquifer parameters should be conducted to ensure sustainable water withdrawal. This includes periodic pumping tests and continuous recording of water levels to track any changes in aquifer behavior.
- **Artificial Gravel Packing:** During well construction, artificial gravel packing should be implemented to enhance the efficiency of water extraction and protect the well from siltation and collapse.
- **Integrated Approach:** Utilize both litholog data and pumping test results for a comprehensive assessment of aquifer properties. This integrated approach provides a more robust understanding of the aquifer dynamics and helps in making informed decisions for groundwater management.

- **Sustainable Withdrawal Rates:** Adhere to the recommended water withdrawal rate of 1500 KLD to prevent over-extraction, which could lead to adverse effects such as land subsidence or depletion of the aquifer.
- **Future Studies:** Further studies should be conducted to explore the impact of seasonal variations on aquifer properties and to refine the methodologies for better accuracy in transmissivity and storativity estimation.

By following these recommendations, the sustainable management and utilization of groundwater resources in the study area can be ensured, thereby supporting the long-term water needs of the local population and ecosystems.

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