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Estimation of Hydraulic Conductivity Using Geoelectrical and Infiltrometer Observations

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ABSTRACT

Hydraulic conductivity (*K*) as a parameter in surface and subsurface water interaction is an important study to research. Field observations using geoelectrics with the Schlumberger configuration and using infiltrometers with double ring were chosen as methods to estimate the (*K*) which aims to recognize the characteristics of the relationship between (*K*) obtained from different observation results. The estimated (*K*) obtained from infiltrometer observations are quite significant compared to geoelectric observations which range from 2.715×10^{-7} m/s to 6.132×10^{-7} m/s, while geoelectrical values range from 1.965×10^{-8} m/s to 3.896×10^{-9} m/s. In this study, the soil conditions in geoelectric observations were carried out in an unsaturated state and infiltrometer observations were in a saturated state. This soil condition is used as one of the reasons for interpreting the research results in this study, that the hydraulic conductivity in unsaturated soil conditions decreases compared to saturated soil.

INTRODUCTION

The hydrological cycle, as the primary focus in hydrology, is presented as a system with various processes occurring within it. Infiltration, as one of the processes in the hydrological cycle, becomes a crucial phenomenon in the interaction between surface water and groundwater. The infiltration process is influenced by the land cover above it and the conditions of the land. When considering the physical properties of the soil, these influencing factors comprise soil porosity, grain size, and hydraulic conductivity, interpreted as the soil's capacity to facilitate fluid flow. The flow entering the hydrological system in the form of precipitation can flow into rivers either on the surface as overland flow (surface runoff) or as subsurface flow after infiltrating into the ground (Cherry et al. 1979). According to Guymon (1994), the unsaturated zone in the hydrological cycle plays a role in channeling water that falls or pools on the surface into the ground or temporarily storing it near the surface for plant use. Cherry et al. (1979) describe unsaturated flow as a multiphase flow through porous media, involving both air and water phases. The flow through these porous media is regulated by hydraulic conductivity and soil permeability coefficients. Numerous studies have been conducted to investigate the interaction between surface water and groundwater in various case studies, employing diverse methods. In an integrated study, utilizing electrical resistivity tomography and infiltration methods to delineate the characteristics and potential of the unsaturated zone in crystalline rocks (Warsi et al. 2019), they asserted that water falling on the surface traverses the unsaturated zone, recognized as an active region controlling the flux of water between the surface and groundwater. The geoelectric method was applied to a lithologically complex porous aquifer in the Anthemountas Basin, Northern Greece, by Kazakis et al. (2016) to estimate the hydraulic properties of the aquifer using Archie's Law and Kozeny's equation. A similar approach was also employed by Niwas et al. (2012) in the Ruhtral aquifer in Germany, utilizing the cementation factor (*m*) and alpha factor (*m*) parameters to calculated porosity in estimating the hydraulic parameters of an aquifer. The application of geoelectric methods can also be utilized to estimate groundwater infiltration, as demonstrated by (Hossain et al. 2021). They computed hydraulic conductivity values using Archie's equation, as employed by (Niwas et al. 2012). However, the commonly used electrical resistivity equations do not apply to silty and clayey soils. Therefore, to calculate porosity, cementation factors were computed based on research conducted by (Choo et al. 2016). The estimated hydraulic conductivity values were used as parameters to calculate infiltration rates using the Green-Ampt model. In his book, Briaud (2013) describes the sealed double-ring infiltrometer test method developed by Trautwein & Danil in 1994 to measure hydraulic conductivity at shallow depths in soils above the groundwater table. The movement of surface water and atmospheric moisture to the groundwater system is significantly controlled by the hydraulic characteristics of the vadose zone above it. The movement of soil moisture through the unsaturated zone plays a crucial role in hydrological processes. According to (Zou et al. 2023), hydraulic conductivity is a dominant hydraulic parameter that governs the flow characteristics in the unsaturated zone, making its determination fundamental for groundwater dynamics characterization and prediction.

Based on a comprehensive review of various literature, this study is designed to identify the relationship between hydraulic conductivity values from different observations, including geoelectric and infiltrometer measurements. Geoelectric observations yield resistivity values used to estimate hydraulic conductivity, considering hydraulic parameters such as porosity, grain size diameter, and material formation factors comprising cementation and alpha factors. Double-ring infiltrometer observations provide infiltration rates and capacities using the Horton model to quantify them into hydraulic conductivity values.

STUDY AREA

The research area is located within the Faculty of Engineering University of Indonesia, in the city of Depok, West Java (Fig. 1). The study region falls within the Jakarta groundwater basin, as defined by the Ministry of Energy and Mineral Resources Regulation number 2/2017 regarding Indonesian Groundwater Basins. According to the Medium-Term Investment Program Plan (RPIJM) for the city of Depok from 2015 to 2019, the research area is predominantly characterized by the Alluvial Fan Rock Unit, consisting of deposits such as clay, sand and gravel, and conglomerate (Fig. 2). The average rainfall over the last ten years from the FT UI Rainfall Station is recorded at 121.12 mm. Based on Fig. 1 of the observation area, there is an orange line representing the observation span for the geoelectrical survey along 80 meters. The length of this span does not indicate any special explanation but is merely due to the limitation of the land at the observation site. The green point represents the midpoint of the geoelectrical observation span, and the two blue points, each located 10 meters from the midpoint, are taken as representative points for the infiltrometer observation. Several observation points have been made within the area delineated by the red line; however, only the points and locations indicated in the figure are explained and selected for analysis in this study.

In all discussions, the flow rate through porous media is regulated by hydraulic conductivity and soil permeability coefficients. Fitts (2013) explains that the physical properties of water and the distribution of pore space determine the amount of water stored in a specific volume and how easily water can move through the material. The physical properties of water consist of the mass density of water (ρ_w) valued at 1000 kg/m^3, and viscosity (μ) of 0.0014 kg/(s.m). Physical properties of porous media, such as porosity (φ) and grain size (d) are determining factors for hydraulic conductivity. Soil test in the laboratory was conducted to obtain soil index

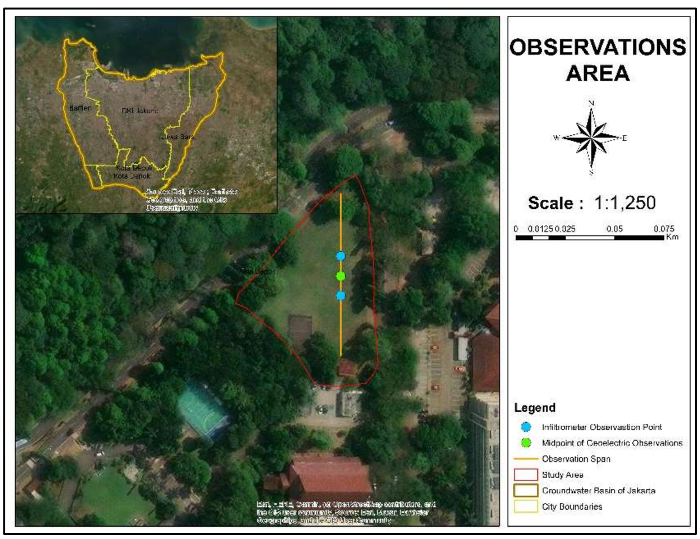

Fig. 1: Observation area.

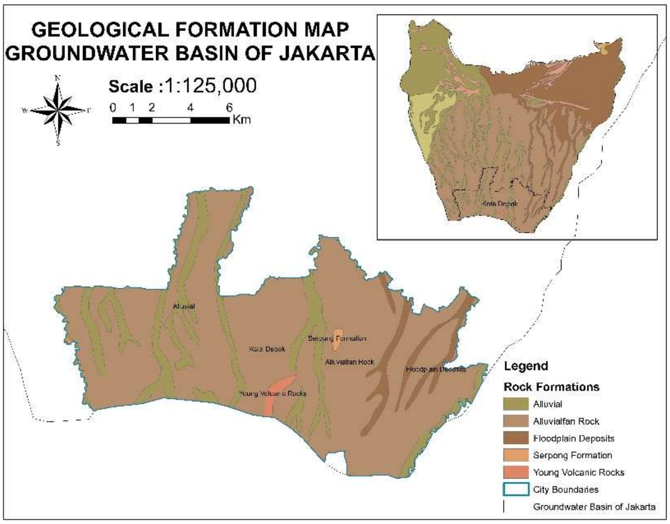

Fig. 2: Geological formation of the study area.

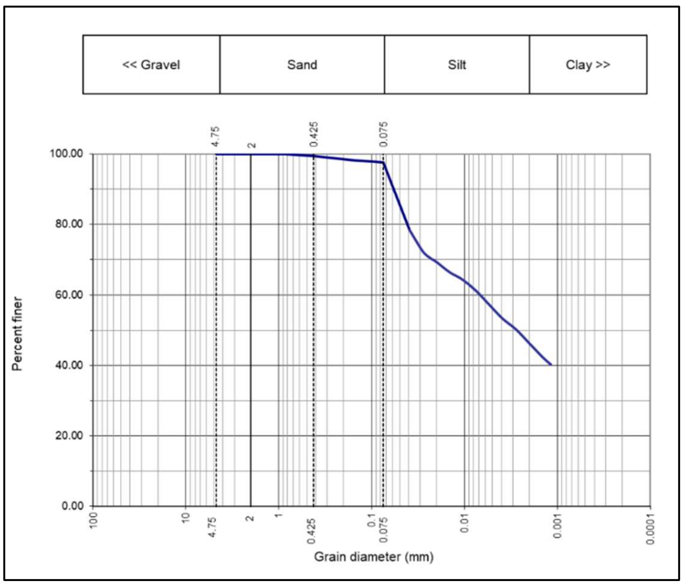

Fig. 3: Grain size distribution curve.

property data, which consists of water content at 46.19%, as well as the results of hydrometer testing analysis indicating 2.5% sand, 52.5% silt, and 46% clay, with grain size distribution depicted in Fig. 3. The visual description of the soil is silty with high plasticity (MH).

MATERIALS AND METHODS

The estimation of hydraulic conductivity is obtained by analyzing field observations, which include geoelectric measurements and infiltrometer measurements. Fetter (1994) explains in his book that the hydraulic conductivity value of soil material can be measured in the laboratory using a permeameter, commonly known as permeability testing. Therefore, in this study laboratory permeability testing is conducted as a control value to estimate the hydraulic conductivity from geoelectrical and infiltrometer observations. The procedure for laboratory permeability testing starts with collecting samples from the observation site, specifically at the midpoint of the observation span indicated by the green point in Fig. 1. Samples are taken using the hand boring method at depths of 0.3-0.5 meters and 1.7-2.0 meters to obtain undisturbed soil. Once the soil samples are collected and ready for permeability testing, a series of permeameter testing equipment is prepared and checked to ensure they are operational. Remove the soil sample from the sampling tube and place it into the permeability test tube. Measure the height of the soil sample for testing. The falling head method will be used for this test, and the calculations to determine the hydraulic conductivity value will be based on the provided explanations by Briaud (2013) and Fetter (1994). In this stage of the permeability test, the soil is saturated for approximately 5 hours. Then an initial reading is taken to determine the initial water level, followed by another reading approximately 16 hours later to measure the final water level. Permeability testing for soil samples at depths of 0.3 to 0.5 meters yielded a hydraulic conductivity value of 8.33×10^{-8} m/s, whereas at depths of 1.5 to 2 meters, it was 2.67×10^{-10} m/s. The hydraulic conductivity values derived from laboratory permeability testing are also utilized to aid in interpreting the hydraulic conductivity characteristics observed through geoelectrical and infiltrometer methods.

Geoelectric Observations

Geoelectric observations were conducted in the open green field within the Faculty of Engineering, Universitas Indonesia, with a span length of 80 meters using the Schlumberger configuration. This aimed to acquire a series of vertical electrical sounding (VES) data, consisting of potential difference and electrical current values, which would be used as parameters for calculating electrical resistivity. Measurements were carried out using the IRES T300F 1D instrument with a reading precision of 0.001 mV. Measurements are carried out in relatively flat regions, spanning from 0° N to 180° S. Electrodes must be installed directly in contact with the ground and positioned in a linear alignment. The electrode arrangement resembling the Schlumberger configuration is depicted in the accompanying diagram (Fig. 4). Electrodes M and N function as potential electrodes, while electrodes A and B serve as current electrodes.

The vertical electrical sounding (VES) data obtained from observations, consisting of potential difference (ΔV) and electrical current (I) values, were analyzed to obtain resistivity values (ρ_a). This involved the preliminary calculation of the geometric factor ($K_{schlumberger}$) for the Schlumberger configuration, expressed as follows (Kirsch, 2009):

$$K_{Schlumberger} = \pi.\ \frac{(AB/2)^2 - (MN/2)^2}{MN} \qquad ...(1)$$

$$\rho_a = K_{schlumberger} \frac{\Delta V}{I} \qquad ...(2)$$

Three sets of VES data were obtained from three separate geoelectric observations conducted at different times. These sets of VES data were then interpreted using Progress v.3.0 software, following the data processing guidelines for 1D resistivity interpretation with Progress software as outlined by Setiadi (2015).

Analysis of Resistivity Values and Hydraulic Parameters

The resistivity values used in estimating hydraulic conductivity are obtained from the interpretation results of the VES data. Kirsch (2009) in his book states, that the electrical resistivity of most minerals is high (except for: clay, metal ores, and graphite), and electrical current primarily flows through pore water. According to the famous Archie's law, the resistivity of a water-saturated clay-free material is defined as follows:

$$\rho_{Aquifer} = \rho_{water} \,.\, F \qquad ...(3)$$

Where $\rho_{aquifer}$ represents the specific resistivity of saturated sand, ρ_{water} is defined as the resistivity of pore water, and F is a formation factor that combines all material properties affecting the flow of electrical current, such as porosity, pore shape, and cementation factor, expressed as follows:

$$F = a\,\varphi^{-m} \qquad ...(4)$$

The constant represents the influence of mineral grains on electrical current. If the mineral grains are perfect insulators, then a equals 1. The value of a will decrease

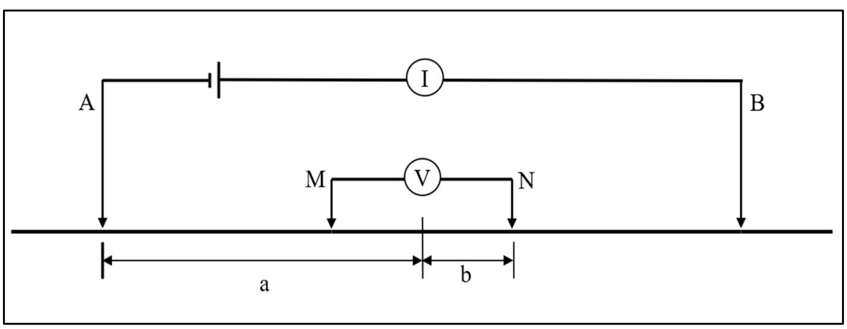

Fig. 4: Schlumberger configuration schematic.

if mineral grains contribute to electrical conductivity to a certain degree. Typical values for a and m according to Schön (1996), as provided by Worthington (1993), for sand are 1 and 1.3, and for sandstone are 0.7 and 1.9. Choo et al. (2016) researched to modify the cementation factor values applicable to clay and silt soil types, expressed by the following equation:

$$m = m_{sand} \,.\, (1 - VF_c) + m_{clay} \,.VF_c \qquad ...(5)$$

Where m_{sand} and m_{clay} are the cementation factors for pure sand and clay respectively and VF_c is the fraction of clay. The values for pure sand and pure clay are 1.55 and 2.11, and the tortuosity parameter (a) is assumed to be 1 because it is related to the length of the current flow path, which is nearly the same for unconsolidated sediments as explained by (Niwas et al. 2012). Thus, the porosity value can be calculated using the modified mathematical formula as follows:

$$\varphi = \sqrt[m]{\frac{\rho_w . a}{\rho}} \qquad ...(6)$$

Thus, the obtained porosity values can be utilized to calculate the coefficient of permeability, as stated by (Kazakis 2016), as follows:

$$k = \frac{d^2}{180} \frac{\varphi^3}{(1-\varphi)^2} \qquad ...(7)$$

Freeze et al. (1979) defined hydraulic conductivity as the proportionality constant in Darcy's law, which is a function of porous media and fluid. Hydraulic conductivity (K) is influenced by the permeability coefficient (k), gravitational acceleration (g), fluid density (ρ), and viscosity (m) of fluid expressed as follows:

$$K = \frac{k\, \rho g}{\mu} \qquad ...(8)$$

Two models for estimating hydraulic conductivity using Archie's Law are explained in this study. The first model, represented by equation …(3), where the value of F is obtained from the graph depicting the relationship between the formation factor F and grain size by (TNO 1976), expressed in the following equation by Kirsch (2009) in his book:

$$F = 1{,}26\, y^{-1{,}20} \qquad ...(9)$$

$$y = 0{,}149 \log M + 0.331 \qquad ...(10)$$

With M representing the grain size in micrometers, the grain size diameter ranges from 0.001 mm to 0.039 mm as obtained from laboratory soil test.

Infiltrometer Observations and Horton Model Infiltration Rate Analysis

Infiltration observations to determine the surface water infiltration into the soil were conducted at the same location and span as the geoelectric observations. Infiltrometer observations were carried out to obtain data on water table decline (Dh) over time intervals (Dt). The infiltrometer used was a double-ring infiltrometer with a ring height of 20 cm, inner diameter of 15 cm, and outer diameter of 30 cm. The observation procedure for the infiltrometer is based on SNI 7752:2012 regarding the method for measuring soil infiltration rates using a double-ring infiltrometer, which refers to ASTM D 3385-88, Standard Test Method for Infiltration Rate of Soils in Field Using Double-Ring Infiltrometer. The infiltrometer observation points can be seen in Fig. 1, the process begins by preparing the ground for placing a double-ring infiltrometer. The rings are then uniformly inserted into the soil to a depth of 2 to 3 cm using a rubber mallet. Before this, a measuring tool with a vertical ruler is fixed to both parts of the ring to facilitate water height readings. Water is introduced into the ring initially at a height of 15 cm. Subsequently, readings and recording of water level reductions are conducted at intervals of 1 minute for the first 10 minutes, 2 minutes up to 30 minutes, and then at 5, 10, to 15-minute intervals. Measurements cease once the water level reduction stabilizes over time. The observation data, represented by the decline in the water table (h), is quantified into the volume (V) of water that infiltrates into the soil by multiplying the water table height (h) with the inner ring area (A). Subsequently, it is presented in the form of a graph illustrating cumulative infiltration, which depicts the relationship between the infiltration volume (F) and the accumulated time (t). Infiltrometer measurements serve as a reference for estimating the infiltration rate using the Horton model. To quantify the infiltration rate with the Horton model, several Horton parameters such as initial infiltration rate (f_0), final infiltration rate (f_c) and recession contant (k) are required. These three Horton parameters will be calculated using the Solver program in Microsoft Excel, allowing the Horton model infiltration rate (f) to be calculated based on the following equation (Wanielista et al. 1990):

$$f(t) = f_c + (f_0 - f_c) \,.\, e^{-k.t} \qquad ...(11)$$

The total volume of infiltrate using Horton's equation is determined by integrating the area under the curve, or:

$$F = \int_0^t f(t)\, dt = f_c\, t + \frac{(f_0 - f_c)}{K} (1 - e^{-K.t}) \qquad ...(12)$$

As explained by (Briaud 2013) based on Darcy's law, which describes the flow through soil, hydraulic conductivity is defined as follows:

$$K = \frac{\frac{V_f}{A}}{\frac{\Delta h_t}{\Delta_z}} \qquad ...(13)$$

Where, V_f represents the volume of water infiltrating into the soil at time t, A is the area of the infiltrometer ring, Dh_t is the vertical distance from the bottom layer to the water level at the outer ring, and D_z is the thickness of the layer. The values of h_t and z are assumed to be the same, based on the water height entering the soil, resulting in the quotient Dh_t and D_z being equal to 1. The relationship between the volume of infiltrated water and the area of the infiltrometer ring is then translated into the value of the Horton infiltration rate in meters per hour, which has been calculated previously.

RESULTS

Geoelectrical Observation Result and Data Interpretation

Geoelectrical observations using the Schlumberger configuration yield varying results at each electrode spacing, as do the obtained resistivity values. For VES 1 data, the smallest resistivity value is 0.094 ohm.m at an electrode spacing of up to 5 m from the 80 m span, and the highest resistivity value is 182.126 ohm.m at a distance of 16 m from the span. The smallest resistivity value for VES 2 data is 1.056 ohm.m at a 5 m electrode spacing from the span, and 161.541 ohm.m at 30 m from the span, while for VES 3 data, resistivity values of 3.606 ohm.m at 60 m electrode spacing from the span, and 347.199 ohm.m at 40 m from the span were obtained. The calculation results of resistivity values vary significantly based on field observation data, which consist of potential difference and electric current data at each electrode spacing arranged according to the configuration used. Three sets of VES data from observations are summarized in Table 1.

Theoretically, soil and rock layers have resistivity values that are highly influenced by the composition of minerals contained within them. The magnitude of resistivity can be influenced by the porosity, and permeability of the material, and detached sedimentary rocks typically have lower resistivity compared to consolidated sedimentary rocks. Resistivity values typically depict and define the type of soil and rock. Electrical resistivity observations can provide insights into the depth of soil layers and rock formations across the span of measurements, through the interpretation of VES (Vertical Electrical Sounding) data. Interpretation is carried out using Progress v.3.0 software by constructing a parameter model comprising layer depths and their corresponding resistivity values, utilizing iterative trial-and-error methods to achieve a close fit between observed and

Table 1: VES data interpretation result.

VES 1		VES 2		VES 3	
Depth (m)	ρ**A** (ohm.m)	Depth (m)	ρ**A** (ohm.m)	Depth (m)	ρ**A** (ohm.m)
0	1.49	0	0.83	0	2.95
0.36	0.05	0.35	1.93	0.35	7.48
1.75	166.56	0.59	1.58	0.55	110.05
5.79	2.61	1.76	2.52	1.7	172.59
10.78	10.01	3.5	35.67	3.41	169.01
25.76	130.05	6.8	3.71	7.4	402.62

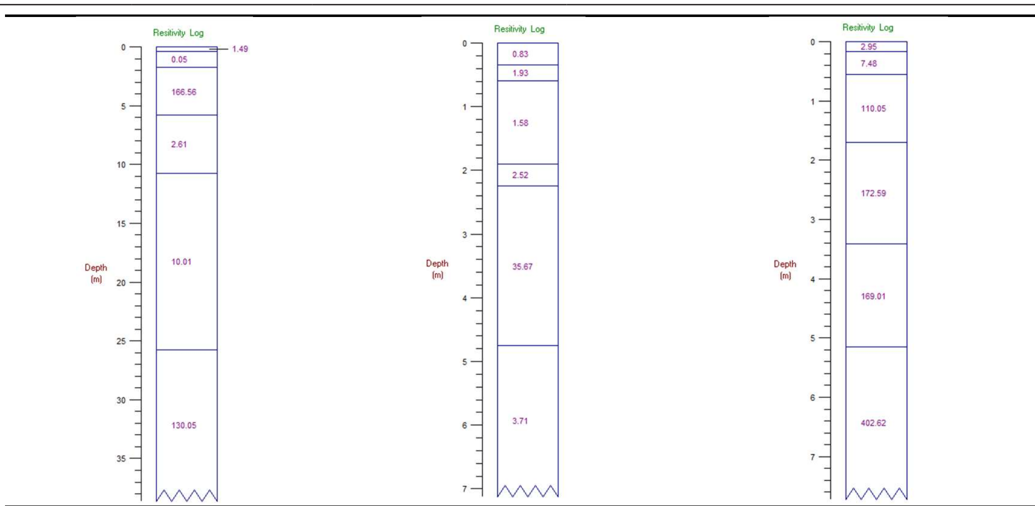

Fig. 5: Interpretation data with layer thickness and resistivity values.

interpreted values (Setiadi 2015). Interpretation results from three sets of VES data collected at the same observation point with an 80-meter span are presented in Table 1 and Fig. 5.

There are a total of 6 soil/rock layers with different thicknesses and resistivity values from the three VES data sets. The interpretation results of VES data 1 can estimate up to a depth of 35 meters with a resistivity value of 130.05 ohm.m. 7 meters with a resistivity value of 3.71 ohm.m for VES data 2, and 7 meters with 402.62 ohm.m for VES data 3. There are two layers with relatively similar depths but different resistivity values. For a layer depth of 0.3 m, the resistivity values obtained from VES data 1, 2, and 3 respectively are 0.05 ohm.m; 1.93 ohm.m; and 7.48 ohm.m. Meanwhile, at a depth of 1.7 m the resistivity values are 166.56 ohm.m; 1.52 ohm.m; and 172.59 ohm.m.

Estimation of Hydraulic Conductivity Values Based on Resistivity Values

According to equation …(3), an important hydraulic parameter to determine first, which is the formation factor (F) combining all material properties affecting electric current flow, such as the values of alpha (a), porosity (□), and cementation factor (m), formulated in equation …(4). Since the Archie equation …(3), is intended for clayey soil materials, it is necessary to reconsider the modified value of mmm based on equation …(5) by Choo et al. (2016). With m_{sand} and m_{clay} values of 1.55 and 2.11 respectively, and VF_c at 46% from laboratory soil sample testing, the value of is calculated as 1.808. Meanwhile, the alpha parameter is assumed to be 1. Two models will be explained in this study to determine the necessary hydraulic parameters. For Model 1, as described by Kirsch (2009) in his book, some formation values are related to the grain size of a material. Therefore, the value of F is obtained from a graph of the formation factor versus grain size from The Netherlands Organisation (TNO) using equations …(9) and …(10). For Model 2, in the research by Hossain et al. 2022 and based on previous studies, the range of pore water resistivity ($\square_w$) values is summarized based on the aquifer resistivity ($\square_a$). The ($\square_a$) value is obtained from observations in this study, and then interpolation of the existing data is done to determine the ($\square_w$) value.

In this study, porosity (φ) emerges as a highly influential parameter for estimating hydraulic conductivity. The interpretation outcomes are based on consistent relative depths employed in this analysis, specifically at 0 m, 0.3

Table 2: The results of hydraulic conductivity estimation model 1.

	Depth (m)	ρA (ohm.m)	d (m)	F	φ	ρ_w (ohm.m)	k (cm^2)	K (m/s)
VES 1	0	1.49	0.000001	4.749	0.422	0.314	1.2558E-15	8.7997E-09
	0.36	0.05	0.000001	4.749	0.422	0.011	1.2558E-15	8.7997E-09
	1.75	166.56	0.000001	4.749	0.422	35.075	1.2558E-15	8.7997E-09
VES 2	0	0.83	0.000001	4.749	0.422	0.175	1.2558E-15	8.7997E-09
	0.35	1.93	0.000001	4.749	0.422	0.406	1.2558E-15	8.7997E-09
	1.76	2.52	0.000001	4.749	0.422	0.531	1.2558E-15	8.7997E-09
VES 3	0	2.95	0.000001	4.749	0.422	0.621	1.2558E-15	8.7997E-09
	0.35	7.48	0.000001	4.749	0.422	1.575	1.2558E-15	8.7997E-09
	1.7	172.59	0.000001	4.749	0.422	36.344	1.2558E-15	8.7997E-09

Table 3: The results of hydraulic conductivity estimation model 2.

	Depth (m)	ρA (ohm.m)	m	a	ρ_w (ohm.m)	φ	F	d (m)	k (cm^2)	K (m/s)
VES 1	0	1.49	1.808	1	0.162	0.293	9.184	0.000001	2.805E-16	1.965E-09
	0.36	0.05	1.808	1	0.005	0.293	9.184	0.000001	2.805E-16	1.965E-09
	1.75	166.56	1.808	1	24.817	0.349	6.711	0.000001	5.560E-16	3.896E-09
VES 2	0	0.83	1.808	1	0.090	0.293	9.184	0.000001	5.560E-16	3.896E-09
	0.35	1.93	1.808	1	0.210	0.293	9.184	0.000001	2.805E-16	1.965E-09
	1.76	2.52	1.808	1	0.274	0.349	6.711	0.000001	2.805E-16	1.965E-09
VES 3	0	2.95	1.808	1	0.321	0.293	9.184	0.000001	2.805E-16	1.965E-09
	0.35	7.48	1.808	1	2.144	0.501	3.488	0.000001	2.805E-16	1.965E-09
	1.7	172.59	1.808	1	25.716	0.349	6.711	0.000001	5.560E-16	3.896E-09

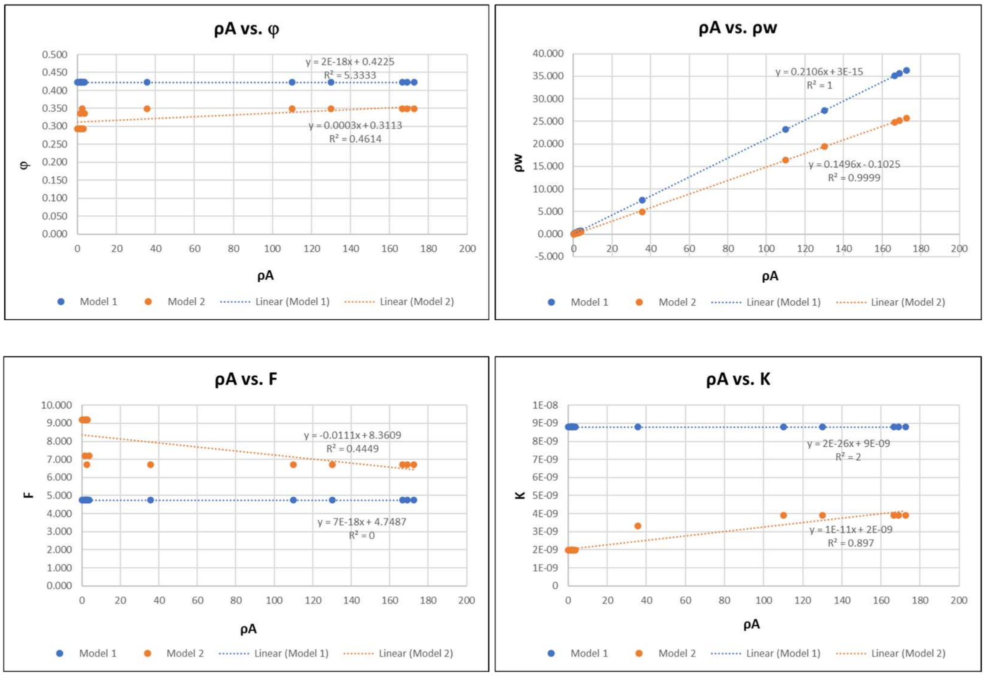

Fig. 6: The data plot of resistivity value with hydraulic parameters.

m, and 1.7 m with a grain size of 0.001 mm. The results of hydraulic conductivity estimation for both models are presented in Tables 2 and 3.

In the first model, the porosity (φ) remains constant because the estimation of the formation factor (*F*) is based on the same grain size diameter, with the value of (*F*) varying depending on changes in the grain size of the material (*d*). In contrast, in the second model, the porosity varies significantly even for the same grain size. From the estimation results, it is essential to understand the relationship between the hydraulic parameters and the hydraulic conductivity values. This relationship can be visualized through the data distribution and graphical plots presented in Fig. 6. Based on the analysis using two modeling approaches to estimate hydraulic conductivity values from resistivity measurements obtained from geoelectrical field observations, the formation factor significantly impacts the determination of (*K*) values.

Tables 4 and 5 summarize the estimated hydraulic conductivity values for different grain sizes.

The selection of the model parameters can be observed from the data plots of resistivity values against various hydraulic parameters, such as porosity (□), pore water resistivity (□$_w$) formation factor (*F*), and hydraulic conductivity (*K*). Based on the obtained R^2 values, Model 2 is preferred, as it has a better R^2 value compared to Model 1 and aligns with Fitts (2013) theory on the properties of porous media, where porosity represents a small portion of pore space within the volume of soil material components. Porosity is defined as the ratio of the volume of voids, consisting of air and water volumes, to the total volume of soil material. The total volume comprises the volumes of air, water, and solid material. In this study, it is assumed that there are no changes to the total volume. Changes in soil conditions occur across dry, unsaturated, and saturated conditions based on the three phases composing the soil material. Soil is considered unsaturated when the pore volume is partially filled with water and partially with air, while saturated soil indicates air voids being fully occupied by water. At a saturation degree of 100%, the volume of void equals the volume of water, whereas a saturation degree of 0% implies the volume of void equals the volume of air. These conditions do not alter the volume of the void hence, the porosity value remains constant.

Estimation of Hydraulic Conductivity Values Through Soil Infiltration

The infiltrometer observation started at $t = 0$ *hr* and continued until t = 1,667 *hr*, with intervals of 1 minute for the first 10 minutes, 2 minutes for the subsequent minutes up to 30 minutes, 5 minutes for the following minutes up to 60 minutes, and 10 minutes for the next minutes up to 100 minutes (SNI 7752:2012). The results of the infiltrometer observation include the actual volume of water infiltrated into the soil in the field and the infiltration volume analyzed

Table 4: The results of hydraulic conductivity estimation model 1 based on grain size.

	Depth (m)	ρA (obs) (ohm.m)	Grain size (mm)				
			0.001	0.005	0.011	0.020	0.039
VES 1	0	1.49	8.7997E-09	5.1962E-07	3.6925E-06	1.6176E-05	8.3711E-05
K (m/s)	0.36	0.05	8.7997E-09	5.1962E-07	3.6925E-06	1.6176E-05	8.3711E-05
	1.75	166.56	8.7997E-09	5.1962E-07	3.6925E-06	1.6176E-05	8.3711E-05
VES 2	0	0.83	8.7997E-09	5.1962E-07	3.6925E-06	1.6176E-05	8.3711E-05
K (m/s)	0.35	1.93	8.7997E-09	5.1962E-07	3.6925E-06	1.6176E-05	8.3711E-05
	1.76	2.52	8.7997E-09	5.1962E-07	3.6925E-06	1.6176E-05	8.3711E-05
VES 3	0	2.95	8.7997E-09	5.1962E-07	3.6925E-06	1.6176E-05	8.3711E-05
K (m/s)	0.35	7.48	8.7997E-09	5.1962E-07	3.6925E-06	1.6176E-05	8.3711E-05
	1.70	172.59	8.7997E-09	5.1962E-07	3.6925E-06	1.6176E-05	8.3711E-05

Table 5: The results of hydraulic conductivity estimation model 2 based on grain size.

	Depth (m)	ρA (obs) (ohm.m)	Grain size (mm)				
			0.001	0.005	0.011	0.020	0.039
VES 1	0	1.49	1.9654E-09	4.9136E-08	2.3782E-07	7.8618E-07	2.9894E-06
K (m/s)	0.36	0.05	1.9654E-09	4.9136E-08	2.3782E-07	7.8618E-07	2.9894E-06
	1.75	166.56	3.8961E-09	4.9134E-07	2.3781E-06	7.8615E-06	2.9893E-05
VES 2	0	0.83	1.9654E-09	4.9136E-08	2.3782E-07	7.8618E-07	2.9894E-06
K (m/s)	0.35	1.93	1.9654E-09	4.9136E-08	2.3782E-07	7.8618E-07	2.9894E-06
	1.76	2.52	1.9654E-09	8.3059E-08	4.0201E-07	1.3290E-06	5.0533E-06
VES 3	0	2.95	1.9654E-09	4.9136E-08	2.3782E-07	7.8618E-07	2.9894E-06
K (m/s)	0.35	7.48	1.9654E-08	4.9136E-08	2.3782E-07	7.8618E-07	2.9894E-06
	1.70	172.59	3.8961E-09	8.3059E-08	4.0201E-07	1.3290E-06	5.0533E-06

using the empirical Horton model, presented in Table 6 and Fig. 7.

The total volume of infiltrate obtained using the Horton model is considered satisfactory because the Sum of Squared Errors (SSE) from the observation and modeling results is very small and approaches zero. Additionally, the agreement between the observation and modeling results can be observed from the graph, indicating a close match. The calculation steps outlined above are used to analyze the infiltration rate, which is then utilized to estimate the hydraulic conductivity, presented in Table 7.

DISCUSSION

The estimated values of hydraulic conductivity from geoelectrical observations, infiltrometer measurements, and laboratory permeability tests are summarized in

The estimated hydraulic conductivity values obtained are quite significant compared to geoelectrical observations and permeability tests. This is also due to soil density and soil unsaturation, resulting in higher hydraulic conductivity values in infiltrometer observations, indicating a greater ability to rapidly transmit fluids. All observations were

Table 6: The actual infiltrometer observation result and Horton's Model.

		t (hr)	Δh (m)	F (Obs) (m^3)	F (Horton) (m^3)	SSE
Measurement 1	Point 1	1.667	0.190	0.0034	0.0039	0.0000002991
	Point 2	1.667	0.257	0.0045	0.0058	0.0000014630
Measurement 2	Point 1	1.667	0.305	0.0054	0.0053	0.0000000017
	Point 2	1.667	0.156	0.0028	0.0029	0.0000000125
Measurement 3	Point 1	1.667	0.524	0.0093	0.0093	0.0000000001
	Point 2	1.667	0.295	0.0052	0.0053	0.0000000116

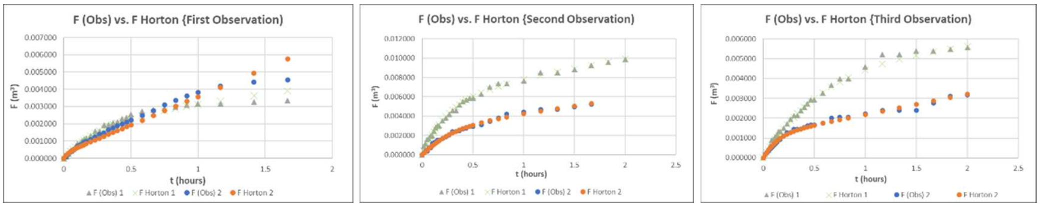

Fig. 7: Scatter plot of actual infiltration and Horton's Model.

Table 7: Infiltration rate Horton's model and hydraulic conductivity.

		t (hr)	**Δh** (m)	f (m/jam)	K (m/s)
Measurement 1	Point 1	1.667	0.190	0.00114	3.1724E-07
	Point 2	1.667	0.257	0.00327	9.0910E-07
Measurement 2	Point 1	1.667	0.305	0.00093	2.5759E-07
	Point 2	1.667	0.156	0.00103	2.8546E-07
Measurement 3	Point 1	1.667	0.524	0.00206	5.7187E-07
	Point 2	1.667	0.295	0.00153	4.2434E-07

Table 8: The hydraulic conductivity values from three observations.

	Geolistrik		Infiltrometer		Permeability Lab	
	Depth (m)	K (m/s)	Depth (m)	K (m/s)	Depth (m)	K (m/s)
Obs 1	0.35	1.965E-09	0.1	6.132E-07	0.3	8.330E-08
	1.76	1.965E-09			1.7	2.669E-10
Obs 2	0.36	1.965E-09	0.1	2.715E-07		
	1.75	3.896E-09				
Obs 3	0.35	1.965E-08	0.1	4.981E-07		
	1.7	3.896E-09				

recorded with timestamps as seen in the figure, along with weather conditions at the time of observation. For geoelectrical observation on December 4th, rainfall was recorded on the previous days totaling 29.8 mm. For infiltrometer observations, the weather tends to be hot with rainfall occurring before the second infiltrometer observation, amounting to 38.8 mm, precisely on October 25th. And the distribution of this data can be seen from the image in Fig. 8. The estimation of hydraulic conductivity values based on geoelectrical observations was conducted at depths of 0.3 m and 1.7 m, as well as permeability testing in the laboratory. In the infiltrometer observation, it is assumed that water seeps vertically into the soil, so the depth of the layer used is assumed to be 0.1 m.

The estimated hydraulic conductivity values obtained are quite significant compared to geoelectrical observations

Table 8: The hydraulic conductivity values from three observations.

	Geolistrik		Infiltrometer		Permeability Lab	
	Depth (m)	K (m/s)	Depth (m)	K (m/s)	Depth (m)	K (m/s)
Obs 1	0.35	1.965E-09	0.1	6.132E-07	0.3	8.330E-08
	1.76	1.965E-09			1.7	2.669E-10
Obs 2	0.36	1.965E-09	0.1	2.715E-07		
	1.75	3.896E-09				
Obs 3	0.35	1.965E-08	0.1	4.981E-07		
	1.7	3.896E-09				

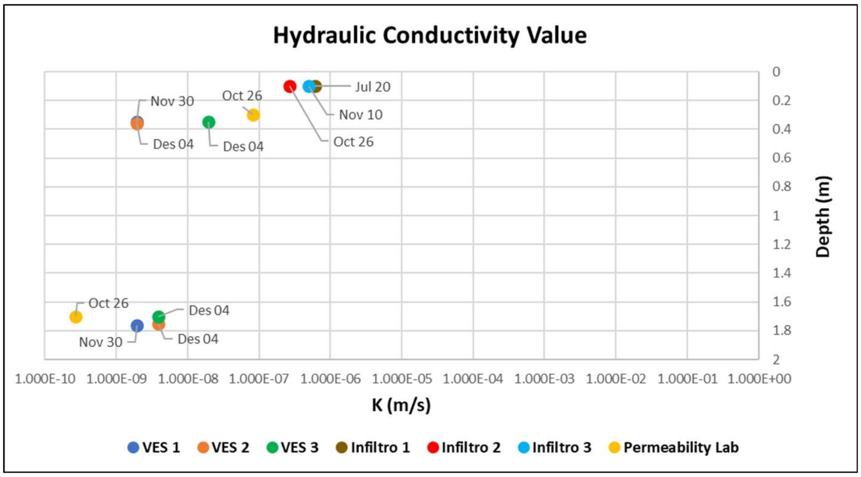

Fig. 8: Plot data of estimated hydraulic conductivity values from different observations.

and permeability tests. This is also due to soil density and soil unsaturation, resulting in higher hydraulic conductivity values in infiltrometer observations, indicating a greater ability to rapidly transmit fluids. All observations were recorded with timestamps as seen in the figure, along with weather conditions at the time of observation. For geoelectrical observation on December 4th, rainfall was recorded on the previous days totaling 29.8 mm. For infiltrometer observations, the weather tends to be hot with rainfall occurring before the second infiltrometer observation, amounting to 38.8 mm, precisely on October 25th.

Based on the observations, the hydraulic conductivity values from laboratory permeability tests, which are used as control values for estimates in this study, are the most accurate: 8.330×10^{-8} m/s at a depth of 0.3 meters and 2.669×10^{-10} m/s at a depth of 1.7 meters. The hydraulic conductivity values estimated from infiltrometer observations are significantly different from those obtained from geoelectrical observations and permeability tests. Geoelectrical observations yielded values in the range of 10^{-8} to 10^{-9}, while the average results from infiltrometer observations were in the range of 10^{-7}. It can be logically concluded that at a depth of 1.7 meters, the soil is more saturated compared to a depth of 0.3 meters. This difference could be because surface soil is more frequently affected by climatic and weather changes, such as exposure to sunlight, wind, and rain, which influence soil moisture and water content. Thus, it can be inferred that unsaturated soil, or soil with less water content, has a higher ability to quickly transmit fluids (water), resulting in higher surface conductivity values. However, other factors must also be considered. Specifically, the soil conditions during the observations were different: the geoelectrical observations were conducted on unsaturated soil, while the infiltrometer and laboratory permeability tests were done on saturated soil. This supports the estimation results, based on the theory explained by Briaud (2013), regarding hydraulic conductivity values for saturated and unsaturated soils.

Briaud (2013) explains that one fundamental observation about water flow in unsaturated soil is that hydraulic conductivity decreases compared to saturated soil. When the soil becomes drier, there is more space for water to flow. However, in reality, this is not the case because air occupies those voids and cannot easily escape, so water can only flow through the remaining water in the soil. The degree of saturation significantly affects hydraulic conductivity values. When the degree of saturation decreases, water content also decreases, and water tension in the soil increases. This is the basic theory of the soil-water retention curve (SWRC). Furthermore, previous research shows the relationship between hydraulic conductivity and water tension. Hydraulic conductivity values depend on water tension; when water tension increases, the amount of water in the soil decreases, making it more difficult for water to infiltrate the soil. Therefore, hydraulic conductivity is lower in unsaturated soil.

CONCLUSION

In this study, both geoelectrical and infiltrometer observations were conducted to estimate hydraulic conductivity values. Based on the estimation results, the geoelectrical observations provided values that closely approximated the laboratory permeability test results. This is attributed to several determining parameters analyzed through assumptions and deductions based on theory and field conditions, such as the formation factor (F) consisting of porosity (□), cementation (m), alpha factor (a), and pore water value. Although geoelectrical observations are generally more cost-effective and faster than drilling and can infer depth and soil layer thickness, upon comparing the results of all three observations as described and depicted, the estimation of hydraulic conductivity values using the infiltrometer observation method was deemed more effective. This is because the values obtained from the infiltrometer observation method aligned well with the theoretical framework and field conditions. However, it is important to note that infiltrometer observation estimation is limited to surface measurements only and lacks depth information.

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